Metamorphic and Igneous Rocks Along the Northwest Border Zone of the Idaho Batholith

By ANNA HIETANEN

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Metasomatic Metamorphism in Western Clearwater County Idaho

By ANNA HIETANEN

METAMORPHIC AND IGNEOUS ROCKS ALONG THE NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

GEOLOGICAL SURVEY PROFESSIONAL PAPER 344-A

Petrologic study of the inner contact aureole of the Idaho batholith

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METAMORPHIC AND IGNEOUS ROCKS ALONG THE NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

METASOMATIC METAMORPHISM IN WESTERN CLEARWATER COUNTY, IDAHO

By Anna Hietanen

ABSTRACT

The area studied comprises about 600 square miles in the western part of Clearwater County, Idaho, along the eastern margin of the Columbia River plateau. The plateau is underlain by almost horizontal flows of Columbia River basalt. In the westernmost part the Clearwater River and its tributaries have eroded deep canyons through this basalt cover, exposing the older metamorphic and plutonic rocks low along the canyon walls. In the eastern part of the area the Clearwater Mountains rise about 2,000 feet above the basalt cover, exposing the rocks of the Precambrian Belt series and those belonging to the Idaho batholith.

The metamorphic rocks exposed along the canyons in the western part of the area could not be correlated with any of the known formations of the Belt series and are referred to as the metasedimentary rocks near Orofino. These rocks as well as the rocks of the Belt series near the batholith are so highly metamorphosed and recrystallized that the original sedimentary structures and color variations which are typical of the Belt series in northern Idaho and Montana are lacking.

In some parts of the area the metamorphic rocks are metasomatically altered to the extent that they megascopically resemble the plutonic rocks—quartz diorite and tonalite. Many small bodies of intrusive rock have wide contact aureoles of coarse-grained plagioclase-rich gneiss which grades gradually to the surrounding metasedimentary rock. Some small bodies of quartz diorite and tonalite, measuring from half a mile to several miles in length, consist entirely of a gneissic variety in which sedimentary relict textures and structures are apparent. In other localities hornblende and biotite together, and calcic plagioclase alone, have segregated into separate small masses. These segregations vary in shape and size; some are long and thin schlieren, some are lens shaped, ranging from 5 cm to 100 m in length, and still others are round masses with a diameter of 1 to 10 m.

Chemical analyses show that iron, magnesium, calcium, aluminum, and sodium were the principal elements introduced into, and silicon and potassium removed from, the metasomatically altered parts of the metasedimentary rocks. Mineralogically, this exchange of material is shown by the replacement of quartz by minerals such as hornblende, biotite, and plagioclase. Near Orofino these minerals were deposited together into the metasedimentary rocks, converting them to quartz diorite and tonalite, whereas near Dent there was a differentiation of femic and salic constituents during the metasomatism; the femic constituents segregated into small masses and the calcic plagioclase either replaced quartz in the surrounding schist or segregated into pods and veins. The minor elements—mainly titanium, zirconium, barium, strontium, vanadium, phosphorus and lanthanum—were concentrated into mafic metasomatic rocks, forming accessory minerals such as ilmenite, zircon, apatite, and allanite or entering into the silicate structures.

Near Orofino the preservation of the structures of the metasedimentary rocks during the metasomatism suggests that the introduction of elements took place during the deformation, but in the parts of the area that are closer to the batholith—in the Headquarters quadrangle and also near Dent—the obliteration of the folded structure of the metasedimentary rocks by the development of hornblende and plagioclase proves that the introduction of elements was later than the deformation. Because the differentiation of the femic and salic constituents occurred only during this later phase, it is concluded that during the early phase—during the deformation—solutions that carried the elements were pushed mechanically into the country rock, whereas in static conditions after the deformation the elements migrated by diffusion in the pore liquid, the rate of migration of various ions giving rise to the differentiation.

PART 1. INTRODUCTION AND MAJOR GEOLOGIC FEATURES

PURPOSE AND SCOPE OF THE INVESTIGATION

This report covers a field and laboratory investigation of the rocks in an area in Idaho along the northwestern border zone of the Idaho batholith. The goal of the work was to determine the metamorphic relations between the Idaho batholith and bordering rocks. The information presented in this report concerns the fundamental problems of the exchange of material during metamorphism and the origin of certain plutonic rocks.

It was recognized long ago that the country rocks around many of the migmatized and granitized areas are enriched in minerals like garnet, cordierite, and biotite. Wegmann (Wegmann and Kranck, 1931; Wegmann, 1935) suggested that this enrichment is due to the migration of magnesium, iron, and sometimes also aluminum from the migmatized area to its frontal zone and gave the name "Mg-front" to this zone. Wegmann suggested that the "Mg-front" moves to the country rock ahead of the migmatite front, which in turn moves ahead of a rising granite. He found evidence for his conception in the deeply eroded Precambrian territory in Finland. There, numerous granitic intrusive bodies are surrounded by migmatites, the nonmigmatized metasedimentary rocks being rare. One would scarcely expect to find such a front in younger mountain chains where most of the granite bodies are discordant and only small-scale granitization is found here and there around them.

In two papers, Doris Reynolds (1946, 1947) suggested that during granitization the elements iron, magnesium, and calcium (cafemic constituents) move to the country rocks ahead of the zone of granitization. The aureole in which these elements are enriched is called a "basic front." It is a zone of desilication and basification. Potassium and one or more of the minor elements titanium, phosphorus, and manganese are usually introduced into the same zone. Because Reynolds' theory included the suggestion that the migration of ions takes place by solid diffusion, it was strongly opposed by a great majority of petrologists. Eskola (1932) has suggested that granitic magma is formed by squeezing out the lowest melting materials from partially refused rocks and the latest crystallizing part from the partially solidified basic rocks in the deep parts of geosynclines. It is clear that the residual material left behind in such extraction of granitic material from geosynclinal sediments would be rich in iron, magnesium, and calcium, thus in the elements which Reynolds found around granitized areas.

This paper will show that in the western part of Clearwater County, Idaho, the schist, gneiss, and quartzite around the Idaho batholith locally contain concentrations of minerals like hornblende, biotite, and andesine. Since these minerals are rare elsewhere in the strata, their presence seems, therefore, to require an explanation. In general there are four possibilities for such occurrences:

1. The original sediment was heterogeneous and contained locally more iron, magnesium, and calcium, perhaps in the form of dolomite. During the metamorphism the dolomite would react with the silica of surrounding sediment to form hornblende. The excess of calcium could be combined with aluminum and silicon to form calcic plagioclase. An extraordinarily large amount of biotite in some localities would be more difficult to understand.

2. The rocks rich in hornblende, biotite, and andesine could be of igneous origin. Some of the andesine- and hornblende-bearing rocks resemble igneous rocks in their appearance, and in places they form discordant contacts with the metasedimentary rocks. However, a major part of these minerals occur as sporadic large grains in the schist and gneiss.

3. Hornblende, biotite, and andesine may have crystallized as a result of a metasomatic introduction of iron, magnesium, and calcium into mainly aluminous strata.

4. The country rocks of the batholith were locally enriched in iron, magnesium, and calcium because of segregation of quartzo-feldspathic material to form the granitic batholith.

The evidence gathered in the field and in the laboratory seems to favor the third and fourth possibility combined. The thin-section studies show that the newly formed minerals replace mainly quartz and thus are of metasomatic nature. There is no indication, however, that the introduced elements moved by solid diffusion. On the contrary, the frequent occurrence of the newly formed minerals parallel to the cracks and joints—also the cross joints—suggests that the migration of elements took place in the liquids filling the cracks. The ultimate origin of the introduced elements probably is the excess not needed for the formation of magmas by partial melting of a part of the crust.

ARRANGEMENT OF THE REPORT

This introduction is followed by brief sections that describe the geography and general features of geology of the area as a whole.

Thereafter, because parts of the area differ from one another markedly in many aspects of geology, they are described separately and in detail in three major sections. The report ends with a section that discusses further and summarizes the theories and details of metamorphism and metasomatism.

LOCATION AND EXTENT OF THE AREA

Clearwater County is in the south-central part of the Idaho "panhandle" about 60 miles south of the Coeur d'Alene silver-mining district. This county extends from the eastern margin of the Columbia River plateau over the Clearwater Mountains to the border of Montana. The area investigated comprises about 600 square miles in the western part of Clearwater County along the northwestern border zone of the Idaho batholith (fig. 1).

In this report the area as a whole is divided into three parts or districts which, as explained above, are given separate and detailed treatment. These parts are as follows:

The Headquarters quadrangle includes about 207 square miles between 46°30′ and 46°45′ north latitude and between 115°45′ and 116° west longitude (pls. 1 and 2). Immediately adjoining it on the west is the smaller but geologically similar district herein referred to as the "Big Island area." It comprises about 45 square miles in the eastern part of plate 3.

In this report the name "Orofino district" is applied to an area of 300 square miles around the town of Orofino in the westernmost part of Clearwater County (pl. 1). This is what was called by Johnson (1947) the Orofino region, a name that earlier had been used by Anderson (1930) for a large area, which included the whole of Clearwater County and a part of Idaho County.

The third district, termed here "the Dent area," comprises an area of about 85 square miles around Dent along the North Fork of the Clearwater River (western part of pl. 3). It joins the Big Island area in the west and the Orofino district in the south.

LOCALITY AND SPECIMEN NUMBERS

Localities are identified by specimen numbers on the maps. The following index gives the section, township, and range for each locality mentioned in the text.

Number	*Township (north)*	*Range*	*Section*
11	38	1E	24
12	38	2E	19
13	38	2E	20
20	36	1E	6
25	37	1E	33
29	37	1E	31
32	36	1W	1
33	36	1W	2
42	35	2E	14
43	35	2E	14
56	37	1E	13
106	37	1E	33
109	37	1E	33
119	38	2E	20
140	38	3E	14
145	36	2E	17
146	36	2E	17
147	36	2E	17
150	36	1W	5
153	36	1W	5
160	36	1W	4
164	36	1W	4
168	36	1W	11
170	36	1W	2
178	36	1W	11
179	36	1W	2
180	36	1W	2
190	38	3E	30
196	38	2E	26
197	38	2E	27
208	40	4E	15
213	40	4E	21
215	40	4E	28
217	40	4E	33
222	39	4E	4
228	39	4E	8
234	39	4E	8
246	39	4E	8
247	39	4E	8
248	39	4E	8
249	39	4E	8
252	39	4E	18
254	39	4E	18
264	39	3E	24
268	39	3E	24
271	39	3E	24
274	39	4E	30
279	39	4E	30
287	37	5E	34
289	37	5E	5
292	38	5E	9
299	38	3E	12
301	38	3E	12
322	40	4E	28
331	39	5E	20
354	36	3E	12
360	36	3E	12
365	37	4E	9
373	37	4E	9
379	38	2E	27
405	37	2E	32
418	37	1E	26
428	37	1E	26
430	37	1E	23
435	37	1E	23
443	37	1E	14
448	37	1E	13
471	38	1E	35
508	37	1E	13
509	36	1E	6
533	36	2E	6
550	38	2E	21
555	38	2E	27
568	37	1E	35
705	38	4E	25
706	38	4E	26
708	38	4E	26
727	39	4E	15
729	39	4E	26
739	38	5E	22
743	38	4E	6
758	38	8E	18
767	38	1E	24
770	37	1E	1
805	36	3E	3
813	36	2E	11
837	37	5E	33
912	41	5E	29
971	42	5E	16
1161	37	1E	11
1162	37	1E	27
1164	37	1E	17
1449	38	2E	18

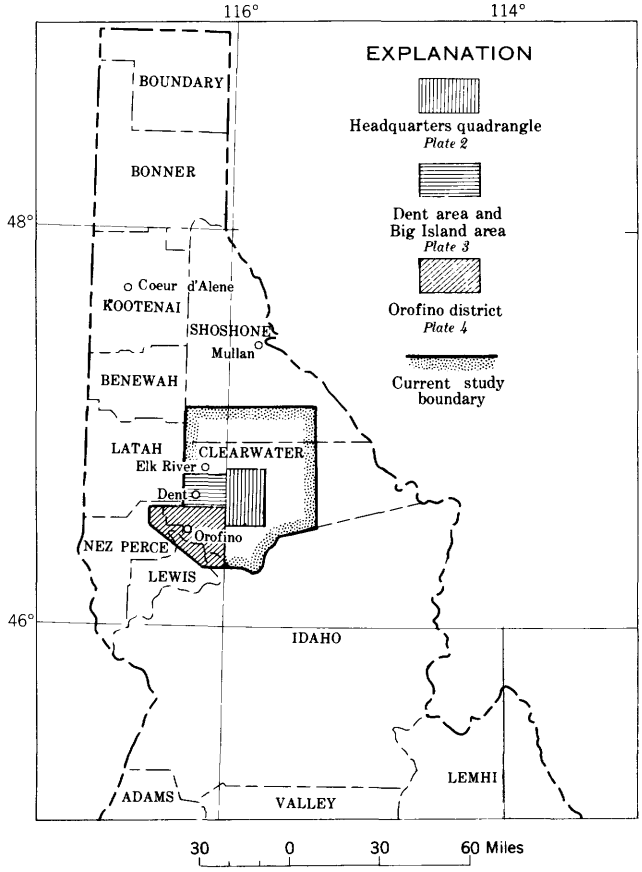

FIGURE 1.—Index map of Idaho showing the area of this report and the area of current study.

EARLIER INVESTIGATIONS IN THE AREA

Reconnaissance work in the area was done by Anderson (1930), who distinguished the following four rock types: gneissic rocks near the Idaho batholith, rocks of the Belt series, granitic rocks of the Idaho batholith, and the Columbia River basalt. According to Anderson, the gneissic rocks include metamorphosed rocks of the Belt series (schist and quartzite) and also injection gneisses in the Headquarters quadrangle and the rocks of the so-called Orofino series near Orofino. He recognized the petrologic difference between the Belt series farther north and the gneissic rocks near Orofino and suggested that the Orofino series is the lowest part of the Prichard formation of the Belt series not exposed elsewhere. Ross and Forrester (1947) followed his reconnaissance in their compilation of the geologic map of the State of Idaho. Only the most southern tongue of basalt in the Headquarters quadrangle is indicated on the map by Anderson (1930) and on the geologic map of the State of Idaho (Ross and Forrester, 1947).

Johnson (1947) divided the pre-Miocene rocks of the Orofino region into three major types according to their constituent minerals: biotite-hornblende-quartz diorite, hornblende-quartz diorite, and biotite-quartz diorite. He gave a petrographic description of each rock type and concluded that all these rocks were originally siliceous sediments that were changed into diorites by hydrothermal solutions from the Idaho batholith.

PRESENT FIELD AND LABORATORY STUDIES

The field studies for this work were begun during the summer of 1946 and were continued during the summers from 1950 through 1953. In the summer of 1946 parts of the canyons near Orofino were mapped. During the summer of 1950, the canyon of the North Fork of the Clearwater River from the mouth of the Little North Fork to Dent was studied. George Makela and Jane Whitner assisted me in this work. Travel down the river was in part by rubber raft and in part by foot along the trail following the river. After the fresh outcrops were studied in the fairly continuous section along the river, fieldwork was extended to higher altitudes where the rocks are deeply weathered and outcrops discontinuous. Virginia Walker assisted me during the summer of 1951, and Dorothy Rainsford and Cynthia Wilkin assisted during the field season of 1952.

Aerial photographs on a scale of 2 inches to the mile were used in the field, and the information was transferred to base maps compiled at a scale of 1: 31,680 by the U.S. Forest Service on the basis of aerial photographs. These base maps, also on a scale of 2 inches to the mile, show the drainage in detail, most of the ridges, and a few roads. In the Headquarters quadrangle the topographic map on a scale of 1 inch to the mile was used in addition to the aerial photographs.

The petrographic studies were begun during the winter of 1946–47 in the School of Mineral Sciences at Stanford University, Calif., and continued during the winters of 1950 through 1953 in the U.S. Geological Survey.

Some of the determinations of the plagioclase were made by using the universal stage and Nikitin's curves and some by measuring the indices of refraction. In many cases both methods were applied. All indices of refraction refer to values as measured with the D-line of the sodium light.

GEOGRAPHY

ACCESSIBILITY

Orofino is reached most easily from Lewiston, which is 45 miles to the west. These towns are connected by State Highway 9 and by a railroad, both of which follow the canyon of the Clearwater River. From Orofino the road continues through Greer, Weippe, and Pierce to Headquarters. Another road, also from the west through Orofino, follows Whiskey Creek and is graveled up to Grangemont, beyond which it is a narrow dirt road that is passable only during dry season. The wooded parts of the area are traversed by many logging roads and fire roads which are passable during the dry season. Many of the trails have been opened to serve as dry-season jeep roads. A country road leading from Orofino to Elk River passes through Dent and forms the only connection of this vicinity to the towns nearby.

TOPOGRAPHY

Two distinctly different topographic units can be distinguished in this area: the Columbia River plateau with canyons and the Clearwater Mountains.

The plateau extends from the west over the Orofino and Dent districts to the western part of the Headquarters quadrangle. The plateau rises slightly toward the Clearwater Mountains in the east; near Orofino it is about 2,900 feet, and in the Headquarters quadrangle about 3,000 feet. The Clearwater River and its tributaries have eroded canyons, 2 to 4 miles wide, into the plateau. The altitude of the river near Orofino is 800 feet above sea level. The canyons are about 2,000 feet deep and their walls are rather steep along the narrow part. In some of the wider places are terraces, most of them less than a quarter of a square mile in area. Toward the east the canyons get narrower and are steep in their lowest part. In many places along the river and along some creeks in the Headquarters quadrangle the outcrops rise vertically from the water. About 10 miles east of Orofino a double-crested hill, Huckleberry Butte, rises more than 1,000 feet above the plateau. The next mountains toward the east, Whiskey Butte, Buck Butte, Bald Mountain, and Democrat Mountain in the southern part of the Headquarters quadrangle and John Lewis Lookout in the northwestern part (pl. 2), form the western border of the Clearwater Mountains, which in the Headquarters quadrangle rise from 3,500 to 5,000 feet above sea level. Remnants of the eastern margin of the plateau form level areas around and between these mountains. The Clearwater River and its tributaries have eroded canyons deep into this marginal plateau area, raising the total relief in places to 3,000 feet.

DRAINAGE

The area is drained westward to the Clearwater River. The North and South Forks of this river join 4 miles northwest of Orofino. The vicinity of Dent and most of the Headquarters quadrangle are drained to the North Fork of the Clearwater River. The extreme southern part of this quadrangle and the eastern part of the Orofino district are drained to Orofino

Creek which joins the South Fork of the Clearwater River in Orofino.

VEGETATION

The mountain area and the eastern part of the plateau westward to Dent and Huckleberry Butte are heavily timbered. The largest stand of white pine in North America is around the North Fork of the Clearwater River and its tributaries near Headquarters. Yellow pine, white fir, cedar, spruce, hemlock, and tamarack also are common. The underbrush is thick, especially along the lower canyon walls and in burned-over areas.

In the western part of the area, near Orofino, only the canyon walls are slightly timbered; the plateau and many of the terraces along the canyons are farmed.

MAJOR FEATURES OF GENERAL GEOLOGY

The rocks of the area can be grouped into four major units: metasedimentary rocks, plutonic rocks, dike rocks, and Columbia River basalt, listed from oldest to youngest.

The plateau is underlain by the Columbia River basalt of Miocene age; the older metasedimentary rocks and plutonic rocks are exposed along the lower parts of the canyons that are eroded through the basalt cover and also in the mountains that rise above the basalt. Most of the Orofino district and the vicinity of Dent are covered by the Columbia River basalt. Best sections of metamorphic rocks in this western part of the mapped area are along Orofino Creek and along the North Fork of the Clearwater River north of Bruces Eddy and near Dent (pl. 1). In the eastern part, in the Headquarters quadrangle (pl. 2) about one-third of the terrain is underlain by the Columbia River basalt, one-third by the rocks of the Idaho batholith, and the rest by metamorphic rocks. As the metasedimentary and plutonic rocks are discussed in detail in later sections, only some general features are mentioned here.

METASEDIMENTARY ROCKS

The oldest rocks in the area are sedimentary in origin. They are strongly folded and faulted and have undergone a high-temperature metamorphism. Moreover, in places their composition and texture have been changed to such extent as to resemble those of plutonic rocks. All these changes make it difficult if not impossible to correlate these rocks with the sedimentary formations in the regions nearby.

The petrographically different rock units were studied and mapped in detail in the hope that enough information could be gathered about the stratigraphy and lithology of the metasedimetary rocks to make correlation possible. In the Headquarters quadrangle and in an adjoining part of the area to the west the metasedimentary rocks consist of schists, gneisses, and quartzites that can be correlated with the rocks of the Precambrian Belt series, which cover wide terrains in northern Idaho and western Montana.

The sequence near Dent resembles that of the Headquarters quadrangle and may belong approximately to the same stratigraphic unit.

Near Orofino, only remnants of the original metasedimentary sequence occur among the plutonic rocks and among the rocks that are highly changed in their composition. These metasedimentary rocks differ stratigraphically and petrologically from those near Headquarters. The main petrologic differences are the occurrence of limestone (marble) and the prevalence of plagioclase-rich gneisses near Orofino. Because of the lack of exposures these rocks could not be correlated with certainty with any of the formations of the Belt series and are referred to as the metamorphic rocks near Orofino.

The modified metasedimentary rocks have undergone an extensive introduction of new material. It was hoped that a detailed petrographic and chemical study of this introduction would lead to better understanding of metasomatic processes, would define the relations between the modified metasedimentary rocks and the plutonic rocks, and would throw light on the genesis of the plutonic rocks.

PLUTONIC ROCKS

The coarse-grained rocks with granitoid texture and usually rich in feldspars are called plutonic rocks regardless of their origin.

The plutonic rocks are mainly quartz diorite and tonalite. The name tonalite is used in this paper for the plutonic rocks that are more siliceous and contain less hornblende than the quartz diorite. Plagioclase in the tonalite is commonly more sodic than that in the quartz diorite. However, all gradations between quartz diorite and tonalite are found. The tonalite differs from trondhjemite mainly because its plagioclase is more calcic (commonly An_{30-35}).

The quartz diorite exposed in the eastern part of the Headquarters quadrangle is a part of the border zone of the Idaho batholith (pl. 2). The satellitic intrusions near the batholith are petrographically and texturally much like the main intrusive body. The long narrow bodies of quartz diorite and tonalite near Orofino differ from the plutonic rocks in the Headquarters quadrangle mainly because of their strongly gneissic texture. The transition from one type to the other can be traced through the river canyon near Dent.

A few small bodies of dunite, hornblendite, and some larger bodies of gabbro were found near Orofino. True granite with abundant potassium feldspar is completely lacking in the whole area.

DIKE ROCKS

The metasedimentary and plutonic rocks are cut by numerous porphyritic dikes of dioritic and granitic composition. Many of these dikes that range from a few meters to 100 m or more in thickness were emplaced parallel to fault zones and to joints. The granite porphyry dikes are especially abundant in the quartz diorite in the Headquarters quadrangle south and east of Dull Axe Mountain and near Big Island. Several dioritic dikes were found north of Silver Butte.

GRANITE PORPHYRY DIKES

Most of the dikes of granitic composition contain abundant large phenocrysts of quartz and plagioclase. The groundmass of the small dikes, which range from a few meters to about 10 m in width, and of the border zones of the larger dikes is fine grained and dark to medium gray. The centers of the larger dikes are coaser grained and light gray. The phenocrysts measure from 2 to 15 mm in diameter and constitute from 5 to 50 percent of the rock.

The plagioclase phenocrysts are zoned and show complex twinning. Many of them actually consist of two or three grains. The quartz phenocrysts are either rounded or have simple pyramidal habits typical of β quartz. Some of the hornblende and biotite occur as phenocrysts of smaller size than those of quartz and plagioclase. Also there are aggregates of light-greenish chlorite which have the shape of augite crystals; these are apparently pseudomorphs after augite. Most of these aggregates of chlorite are surrounded by small crystals of hornblende.

The groundmass of the dikes consists of plagioclase, orthoclase, quartz, hornblende, and biotite. Apatite, sphene, and magnetite are common accessories. In the coarser grained centers of the dikes some of the minerals in the groundmass occur as euhedral grains of medium size, which give the groundmass itself a porphyritic texture. This suggests that there were two major phases of crystallization of phenocrysts and then the final consolidation of the fine-grained groundmass.

A specimen of a granite porphyry dike about a mile south of Big Island was analyzed chemically (table 1). This dike contains abundant quartz and plagioclase phenocrysts of medium size, many of which are surrounded by fine granophyric intergrowth of quartz and feldspars. Orthoclase forms about 10 percent of the rock and appears only in the groundmass. Biotite is partly altered to chlorite that includes some leucoxene. The amount of potassium in this dike is less than that of the normal granite but somewhat more than that of the tonalites. The normative amount of orthoclase is closer to that of the tonalite than to that of a normal granite. The high content of anorthite also brings the composition close to that of tonalites.

Most dikes near Orofino consist of fine-grained greenish-weathering rock in which quartz phenocrysts can be seen with a hand lens. These dikes are probably chemically and mineralogically similar to the fine-grained granite porphyry dikes in the Headquarters quadrangle.

A fine-grained light-gray rock exposed on the plateau north of Ahsahka (loc. 1164, pl. 4) also is granitic in composition but its mineralogy is different. The main constituents are sanidine, albite, green biotite, and hornblende. Sanidine occurs in small phenocrysts and also in the groundmass with albite. Albite crystals are lath shaped and well oriented. Tiny hornblende prisms and biotite flakes, many of them altered to chlorite, are the dark minerals. Sphene, apatite, and muscovite occur as additional constituents in this rhyolitic rock.

TABLE 1.—*Chemical composition, norms, and Niggli values of granite porphyry No. 301, Big Island area (pl. 3)*

[Marietta Corbin, analyst]

Constituent		Analysis		Norm			Niggli values	
Conventional symbol	Symbol rearranged for cation percent	Weight percent	Cation percent	Mineral	CIPW	Molecular		
SiO_2	--------	65.66	61.35	Q	18.44	17.23	si	253.0
Al_2O_3	$AlO_{3/2}$	15.93	17.54	Or	14.96	15.10	ti	1.42
Fe_2O_3	$FeO_3/_2$	.41	.29	Ab	34.76	37.25	al	36.20
FeO	--------	2.79	2.18	An	17.52	17.60	fm	26.70
MnO	--------	.06	.05	C	--------	.03	c	15.45
MgO	--------	2.82	3.92	En	7.01	7.84	alk	21.65
CaO	--------	3.73	3.73	Fs	4.09	3.50	qz	66.40
Na_2O	$NaO_1/_2$	4.11	7.45	Ap	.37	.32	k	.27
K_2O	$KO_1/_2$	2.53	3.02	Il	.93	.68	mg	.61
TiO_2	--------	.49	.34	Mt	.60	.43	c/fm	.58
P_2O_5	$PO_5/_2$	.15	.12	Cc	.02	.02		
CO_2	--------	.01	.01	H_2O	1.25	--------		
H_2O+	--------	1.13	(3.52)					
H_2O-	--------	.12	--------					
Total	--------	99.94	100.00	--------	99.95	100.00	--------	
	O		162.04					
	OH		7.04					
	Total anions	--------	169.08					

LAMPROPHYRES

Generally the lamprophyres in the western part of the area are more mafic than those in the eastern part. Near Orofino gabbroic composition is prevailing, whereas in the Headquarters quadrangle the dioritic dikes are more common.

The gabbroic dikes are dark and fine grained.

Brownish hornblende occurs as phenocrysts, and small green hornblende prisms oriented at random are the main constituent in the groundmass. Light-colored minerals in the groundmass are plagioclase in small lath-shaped crystals and some small quartz crystals. A few round quartz phenocrysts including radial hornblende needles along their surfaces were found in the horizontal dike along Orofino Creek about a mile west of Lime Mountain.

A dark dike exposed in a road cut about half a mile north of Greer contains considerably less hornblende and more quartz than the dike along Orofino Creek. Thin sections show that small augite crystals and some biotite occur with hornblende. Plagioclase is zoned and twinned. Quartz phenocrysts are surrounded by tiny radial augite needles. The mineralogy of this dike suggests that it is dioritic in composition.

In the eastern part of the area two texturally different types of dioritic dikes are common: porphyritic dikes with plagioclase and hornblende phenocrysts and equigranular dikes with needlelike hornblende oriented at random.

The phenocrysts in the porphyritic dikes are euhedral plagioclase and hornblende and measure from 1 to 6 mm in diameter. The plagioclase is zoned and twinned mainly according to the Carlsbad law. The groundmass consists of fine-grained plagioclase, quartz, hornblende, and biotite. Sphene, magnetite, and apatite are the common accessory minerals.

The dikes with needlelike hornblende contain the same minerals as the porphyritic dikes. There are a few euhedral plagioclase and hornblende phenocrysts but most of the minerals occur as grains of medium size. About one-half of the hornblende needles are pale green, and the others are brownish or have a brownish center. The larger phenocrysts of hornblende have brown spots of biotite or are partly altered to chlorite.

COLUMBIA RIVER BASALT

Most of the terrain in the western part of the mapped area is underlain by the Columbia River basalt; the older rocks are exposed only along the lower parts of the canyons and in some hills that rise above the basalt cover. In places the basalt also covers the bottom of the river canyon, proving that the relief of the pre-Miocene erosion surface was greater than the present one. In the Headquarters quadrangle the basalt covers the southwestern part of the quadrangle and more than 20 square miles of the northwestern part. These parts are easily recognized by their low relief.

No extensive study was made of the basalt. The thickness of the basalt cover exposed in the western part of the mapped area ranges from 300 to 600 m, thinning toward the east. In the southwestern part of the Headquarters quadrangle and in the surroundings of John Lewis Lookout, it is about 300 m, several flows being exposed along Orofino Creek. Only one flow occurs near Hollywood and Jaype and this flow has a conspicuous columnar structure. Apparently the various occurrences shown on the map of the Headquarters quadrangle are remnants of a larger flow or series of flows which once covered most of the quadrangle.

Near Orofino and Dent several thick horizontal flows are exposed along the canyon walls on the top of the metasedimentary and plutonic rocks. The more resistant parts of some flows stand out along the canyon walls and have prominent columnar jointing.

The basalt is dark fine- to medium-grained rock in which slender plagioclase laths can be detected with the unaided eye. Small round vesicles occur in many outcrops, especially in the upper part of the flows. Microscopic study shows that the basalt in the various parts of the mapped area is much alike. About 40 percent of the typical basalt consists of lath-shaped twinned labradorite (An_{55}) crystals oriented at random or forming an ophitic texture. The interstices are filled by augite (about 35 percent) and greenish or brownish glass with some tiny crystallites. Long laths of ilmenite, grains of magnetite, and a few small crystals of olivine altered to iddingsite are common additional constituents. In some localities, as north of Ahsahka, round grains of magnetite abound.

A very light, weathered and brecciated rock which is hard to break with the hammer occurs on the plateau in the northwestern part of the region where only a narrow strip of basalt is marked on the map (pl. 3). Abundant quartz crystals can be seen with the naked eye. Late weathering products fill the honeycombed cavities in this rock. Thin section shows that this rock is brecciated glass which is partly divitrified. The index of refraction, $n=1.560\pm0.002$, indicates that it is basaltic in composition and contains about 50 percent silica (George, 1924, p. 365). Quartz in this rock occurs as angular fragments that show undulatory extinction and tiny dust like inclusions, just as does quartz in the coarse-grained quartzite exposed half a mile north of this locality. These boulders probably represent the bottom of the first flow that picked up fragments of quartz from the underlying quartzite.

PART 2. THE HEADQUARTERS QUADRANGLE AND THE BIG ISLAND AREA

Each of the three rock units, metasedimentary rocks (Belt series), plutonic rocks, and Columbia River basalt, covers about one-third of the Headquarters quadrangle (pl. 2). The largest continuous area of metasedimentary rocks lies southwest of Headquarters, and another large area is exposed in the northwestern part of the quadrangle. The rocks of the Idaho batholith are exposed in the eastern part.

BELT SERIES

The metasedimentary rocks are continuous with the Belt series in Latah and Shoshone Counties, where they are far less metamorphosed than in the Headquarters quadrangle. In Shoshone County the Belt series consists of a series of interbedded quartzites, argillites, shales, sandstones, and limestones that have been divided on the basis of lithology into formations. Each formational division, Prichard, Burke, Revett, St. Regis, Wallace, and Striped Peak, has a characteristic color and distinctive relict structures, such as ripple marks, mud cracks, raindrops, and "molar-tooth" (Gibson, 1941, p. 375). In the Headquarters quadrangle, recrystallization of the constituent minerals has completely obliterated the characteristic colors and relict structures, and numerous faults interrupt the stratigraphic sequence. Therefore the correlation must be based mainly on lithologic composition and on stratigraphic sequence as determined in individual sections and on the basis of mapping.

For descriptive purposes the metasedimentary rocks are subdivided into the following units, each of which is found in one or in several stratigraphic units: (1) sillimanite-garnet schist, (2) coarse-grained muscovite-biotite schist, (3) banded biotite-plagioclase schist and gneiss with muscovite-sillimanite laminae, (4) white granular quartzite and biotite-plagioclase gneiss, (5) diopside-plagioclase gneiss with quartzitic layers, (6) thin-bedded diopside-plagioclase gneiss with thin biotite-bearing layers and biotite quartzite, and (7) coarse-grained thick-bedded quartzite.

The lithologic character, mineral content, and chemical composition of these rock types and the sequences found in individual sections are described first and then an attempt will be made to correlate these rocks with their less metamorphosed equivalents in Shoshone County.

Sillimanite-garnet schist interlayered with biotite-plagioclase schist is the main rock unit among the metasedimentary rocks in the Headquarters quadrangle. It is very resistant to weathering, forming the highest mountain tops in the area. Two or more units of diopside-plagioclase gneiss with quartzitic layers are interbedded with the schist. This schist and the gneiss units interbedded with it cover much of the central and northern parts of the quadrangle and the lower parts of the canyon walls in the vicinity of Big Island (pl. 3); the other metasedimentary rocks occur only locally. The coarse-grained thick-bedded quartzite is exposed in the northeastern corner of the Headquarters quadrangle and the thin-bedded biotite quartzite, biotite gneiss, and diopside-plagioclase gneiss crop out just south of it.

LITHOLOGY AND PETROGRAPHY

SCHIST

In most outcrops of schist the bedding is distinct because of variation in the amounts of the major constituents: quartz, plagioclase, biotite, muscovite, sillimanite, and garnet. Thick layers of coarse-grained schist rich in biotite, sillimanite, and garnet alternate with thinner layers of medium-grained biotite-plagioclase schist or gneiss. In the biotite-plagioclase schist the individual beds are 2 to 5 cm thick and are separated by thin biotite-muscovite-sillimanite laminae. In most outcrops the schist breaks along these laminae, showing shiny bronze-colored beddings planes.

The schist that occurs below the lower diopside-plagioclase gneiss unit and also the schist above the upper gneiss unit contain less sillimanite and garnet and more muscovite than the schist between the two gneiss units. Layers of this muscovite-biotite schist are exposed in the northwestern part of the quadrangle, west of John Lewis Lookout where they stratigraphically overlie a biotite-plagioclase schist with muscovite laminae, which is exposed along the river west of the lookout. Where sillimanite and garnet are abundant, they are indicated on the map (pls. 2 and 3).

SILLIMANITE-GARNET SCHIST

The sillimanite-garnet schist is coarse grained, gray or brownish, and commonly rich in micaceous minerals. Homogeneous beds ranging from 20 cm to 2 m in thickness alternate with thin-bedded or laminated beds which are 1 to 2 m thick. In these laminated beds, layers ranging from 2 to 3 mm in thickness are separated by paper-thin laminae composed of biotite and sillimanite or of biotite, muscovite, and sillimanite. A thin lamination is common in shaly layers in the Wallace formation in Shoshone County, but it is not certain that it would be preserved in rock that is as strongly deformed and thoroughly recrystallized as is the coarse-grained sillimanite-garnet schist of Headquarters quadrangle. It is more plausible that the laminae in the schist are a result of segregation of micaceous minerals during the recrystallization. Large mica flakes and sillimanite needles crystallized along the slip planes, which are usually parallel to the bedding in this area. The schist breaks parallel to these planes, showing glittering surfaces with white sillimanite clusters (fig. 2).

Microscopic study shows that layers between the mica-sillimanite laminae consist mainly of quartz and small flakes of biotite (fig. 3). Garnet is generally abundant, and in some places plagioclase occurs as an additional constituent. Ilmenite, magnetite, and zircon are the accessories. The sillimanite is either fibrous or in clusters of fine needles. Some of the sillimanite has been altered to sericite, garnet to antigorite along the cracks, and biotite to chlorite. The more schistose layers contain larger needles and prisms of sillimanite distributed fairly evenly throughout the rock (fig. 4).

Chemical composition in weight percentages and in ionic percentages for a sample (loc. 247) of the sillimanite-garnet schist is shown in table 2. The ionic percentages were calculated from the weight percentages of the chemical analyses by use of the method suggested by Barth (1952) and Eskola (1954). Molecular norm was calculated on the basis of the ionic percentages by use of the method suggested first by Niggli (1936) and modified by Barth (1952) and Eskola (1954). The CIPW norm, or weight norm, is given in table 2 for comparison. The numerical values for the molecular norm are close to those for the weight norm. Thus the molecular norm (the calculation of which requires much less time) can be substituted for the weight norm, and the new values can still be compared with the CIPW norm values given in the literature. Barth (1952) and Eskola (1954) have pointed out that the ionic percentage and molecular norm have a great advantage over the weight percentage and the weight norm when metasomatic replacements are considered.

Because of the heterogeneity of the rocks the amounts of the major constituents vary within short distances, and the mode measured in the thin section does not give a true picture of the mineral content of the sample analyzed. The measured mode was therefore corrected to correspond to the chemical analyses, and the method of calculation used is essentially the same as used earlier by the author (Hietanen, 1947, p. 1035–1038). For this purpose the chemical composition of the constituent minerals was needed.

The composition of these minerals can be computed from the chemical analyses of those rock samples that contain only two or three minerals; for instance, the composition of biotite was calculated from the analysis of biotite quartzite. For the samples from the schist

FIGURE 2.—Clusters of sillimanite (white) in the sillimanite-garnet schist measure from 2 to 3 cm in diameter. North Fork of the Clearwater River (loc. 222, pl. 1).

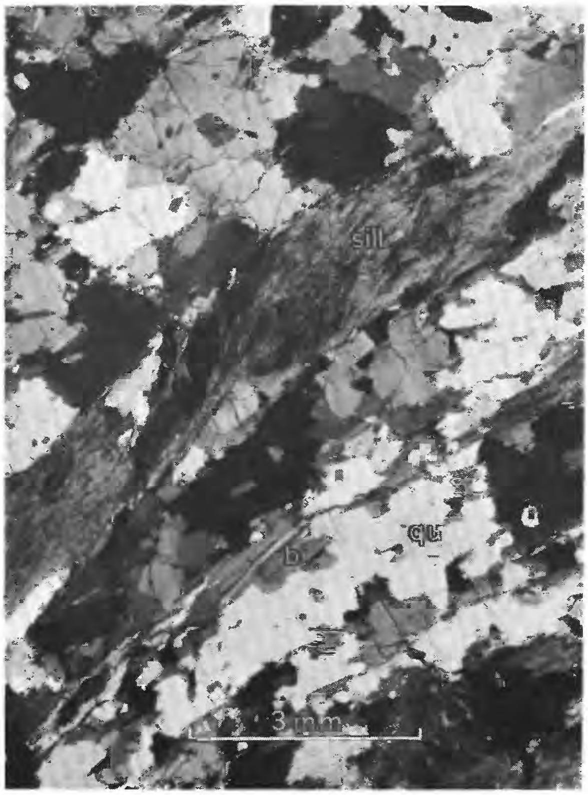

FIGURE 3.—Photomicrograph showing lamination in the sillimanite-garnet schist along the North Fork of the Clearwater River (loc. 271, pl. 1). Sillimanite (sill) is fibrous and partly altered to sericite. bi=biotite; qu=quartz. Crossed nicols.

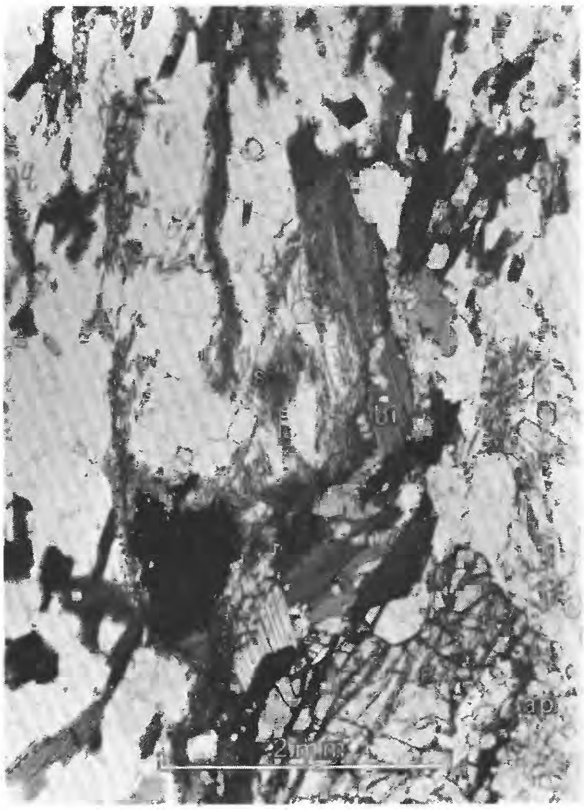

FIGURE 4.—Photomicrograph of a typical sillimanite-garnet schist. Note the needle-like sillimanite (sill) and abundant round grains of apatite (ap); g=garnet; bi=biotite. Locality 217 along the North Fork of the Clearwater River (pl. 1). Plane polarized light.

nearby the Fe:Mg ratio of this calculated analysis was corrected to correspond to the slightly different optical properties of biotite in the schist. The composition of hornblende can be calculated from the analysis of hornblende gneiss. For the other hornblende-bearing samples the Fe:Mg ratio of this calculated analysis can be adjusted—using Winchell's graphs—to correspond to the optical properties of hornblende in these other samples. Theoretical formulas were used for the minor constituents and for the feldspars. Consequently more microcline than actually present may occur in the list of minerals, and the conclusion presented in a later section of this paper is thus by no means exaggerated. Compositional water ($+H_2O$) indicates the amount of water left over in the calculation of mineral content.

More than 50 percent of sample 247 is quartz. Feldspar is absent, and biotite is more abundant than muscovite. The garnet occurs as small round grains (fig. 5); some of the sillimanite is altered to sericite.

COARSE-GRAINED MUSCOVITE-BIOTITE SCHIST

The coarse-grained muscovite-biotite schist contains numerous large flakes of muscovite, which are well oriented parallel to the bedding. Garnet and sillimanite are usually present in small amounts. This schist is petrographically very similar to the major part of the schist of the Prichard formation in southern Shoshone County. Petrologically similar layers of schist occur in every formation and therefore the schist cannot be used to identify the formations.

BANDED BIOTITE-PLAGIOCLASE SCHIST AND GNEISS WITH MUSCOVITE-SILLIMANITE LAMINAE

Layers of banded schist are interbedded with the homogeneous coarse-grained sillimanite-garnet schist. The banded schist consists of thin layers of biotite-plagioclase gneiss separated by laminae that are composed of muscovite and sillimanite (fig. 6). The thickness of the feldspathic layers ranges from 1 to 5 cm and that of the micaceous laminae from 1 to 3 mm. Some micaceous schist layers ranging from 2 to 30 cm in thickness are also interbedded. Most plagioclase is untwinned; only a few grains show albite twinning. As a rule, muscovite is more abundant in the laminated

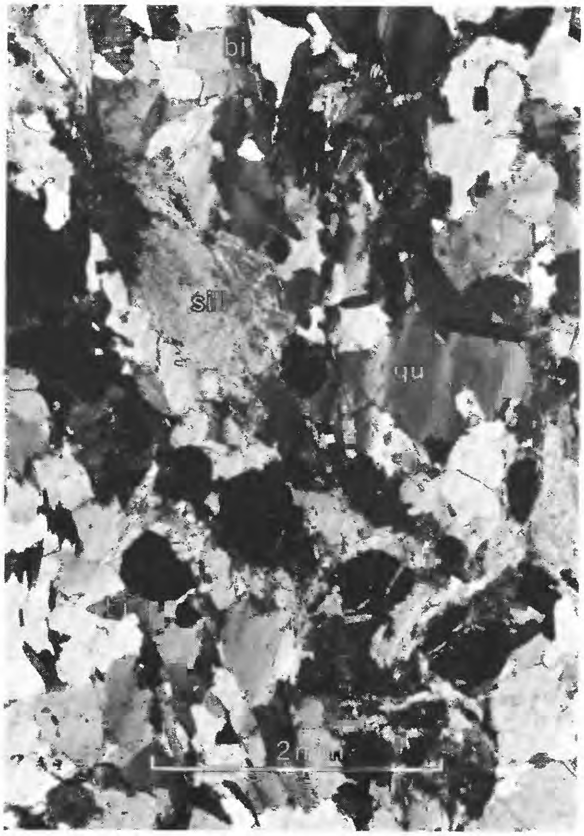

FIGURE 5.—Photomicrograph of a thick homogeneous layer of sillimanite-garnet schist. Garnet is distributed throughout the layer in fairly small round grains (black). Fibrous sillimanite (sill) occurs in small clusters which are surrounded by muscovite. Quartz (qu) shows strong undulatory extinction. Biotite (bi) occurs in clusters and in individual flakes of medium size. Locality 247 (pl. 2) along the North Fork of the Clearwater River. Crossed nicols.

TABLE 2.—*Chemical composition, ionic percentage, norms, and minerals of metasedimentary rocks of the Headquarters quadrangle and vicinity*

[Ruth Holzinger, analyst, U.S. Geological Survey]

Locality and specimen No.	246	247	248	252	254	289
Rock type	Banded biotite-plagioclase schist with muscovite-sillimanite laminae.	Sillimanite-garnet schist.	Biotite gneiss	Biotite gneiss	White granular quartzite.	Diopside-plagioclase gneiss.
Location	North Fork of Clearwater River.	North Fork of Clearwater River.	North Fork of Clearwater River.	North Fork of Clearwater River.	North Fork of Clearwater River.	Snake Creek.

Chemical composition and ionic percentage

Constituent		Weight percent	Cation percent	Weight percent	Cation percent	Weight percent	Cation percent	Weight percent	Cation percent	Weight percent	Cation percent	Weight percent	Cation percent
Conventional symbol	Symbol rearranged for cation percent												
SiO_2		62.93	60.94	71.61	71.73	74.19	71.07	76.94	71.67	90.60	87.52	63.22	60.19
Al_2O_3	$AlO_{3/2}$	18.94	21.61	12.30	14.52	12.55	14.16	10.97	12.04	5.60	6.38	11.15	12.51
Fe_2O_3	$FeO_{3/2}$	1.04	.76	.85	.64	.31	.22	.12	.08	.04	.03	.24	.17
FeO		6.68	5.41	6.92	5.80	2.59	2.08	.93	.72	.13	.10	3.07	2.44
MnO		.11	.09	.18	.15	.02	.02	.01	.01	.00	.00	.05	.04
MgO		1.84	2.65	1.57	2.34	1.51	2.16	3.31	4.59	.15	.21	5.31	7.53
CaO		.68	.70	.15	.16	2.63	2.70	.28	.29	.50	.52	14.87	15.16
Na_2O	$NaO_{1/2}$	1.40	2.63	.30	.58	2.78	5.16	4.42	7.98	2.52	4.73	.76	1.41
K_2O	$KO_{1/2}$	3.70	4.57	2.68	3.43	1.50	1.83	1.92	2.28	.26	.32	.12	.15
TiO_2		.80	.58	.81	.61	.74	.53	.41	.29	.19	.14	.44	.32
P_2O_5	$PO_{5/2}$	.05	.05	.03	.02	.07	.06	.06	.04	.05	.04	.09	.07
CO_2		.01	.01	.02	.02	.01	.01	.01	.01	.01	.01	.01	.01
H_2O+		1.83	(5.91)	2.26	(7.54)	.82	(2.62)	.48	(1.30)	.03	(.10)	.71	(2.25)
H_2O-		.15		.09		.05		.06		.03		.26	
Total		100.16	100.00	99.77	100.00	99.77	100.00	99.92	100.00	100.11	100.00	100.30	100.00
	O		163.27		170.43		172.78		171.66		188.31		163.93
	OH		11.82		15.08		5.24		2.60		.20		4.50
	Total anions		175.09		185.51		178.02		174.26		188.51		168.43

Norm

Mineral	CIPW	Molecular	CIPW	Molecular	CIPW	Molecular	CIPW	Molecular	CIPW	Molecular	CIPW	Molecular
Q	31.92	30.93	52.06	52.12	43.12	41.30	38.10	35.48	73.80	71.28	26.67	25.39
Or	21.87	22.85	15.86	17.15	8.85	9.15	11.35	11.40	1.56	1.60	.72	.75
Ab	11.85	13.15	2.52	2.90	23.49	25.80	37.38	39.90	21.34	23.65	6.45	7.05
An	2.95	3.05	.47	.55	12.57	12.95	1.00	1.05	2.08	2.20	26.64	27.37
C	11.55	13.19	8.74	10.29	1.75	1.99	1.25	1.36	.40	.45		
Wo											19.42	19.12
En	4.58	5.30	3.90	4.68	3.76	4.32	8.24	9.18	.37	.42	13.22	15.06
Fs	10.30	9.08	11.00	10.04	3.34	2.92	.95	.80	.00	.00	4.80	4.16
Ap	.13	.13	.07	.05	.16	.16	.13	.11	.13	.11	.20	.19
Il	1.52	1.16	1.53	1.22	1.40	1.06	.77	.58	.23	.16	.83	.64
Ru									.07	.06		
Mt	1.50	1.14	1.23	.96	.44	.33	.16	.12	.07	.05	.35	.25
Cc	.02	.02	.04	.04	.02	.02	.02	.02	.02	.02	.02	.02
H_2O	1.98		2.35		.87		.54		.06		.97	
Total	100.17	100.00	99.77	100.00	99.77	100.00	99.89	100.00	100.13	100.00	100.29	100.00

Mineral content calculated on basis of chemical analyses

Quartz	32.8		54.2		46.11		43.90		73.76		28.75	
Plagioclase	13.2				36.24		38.60		23.90		32.47	
(Anorthite content)	(20)				(35)		(3)		(11)		(85)	
Microcline	1.0				1.30		1.60		1.07			
Biotite	19.6		17.5		12.34		14.33		1.00			
Muscovite	18.6		13.9		2.27		1.00					
Chlorite			1.6									
Sillimanite	6.5		4.1									
Garnet	8.0		7.7									
Diopside											36.92	
Sphene											1.00	
Apatite	.1		.1		.15		.12		.12		.19	
Ilmenite			.6		1.00							
Rutile									.15			
Calcite					.02		.02		.02			
Subtotal	99.8		99.7		99.43		99.57		100.02		99.33	
$+H_2O$	.3				.34		.35		.09		.97	
Total	100.1		99.7		99.77		99.92		100.11		100.30	

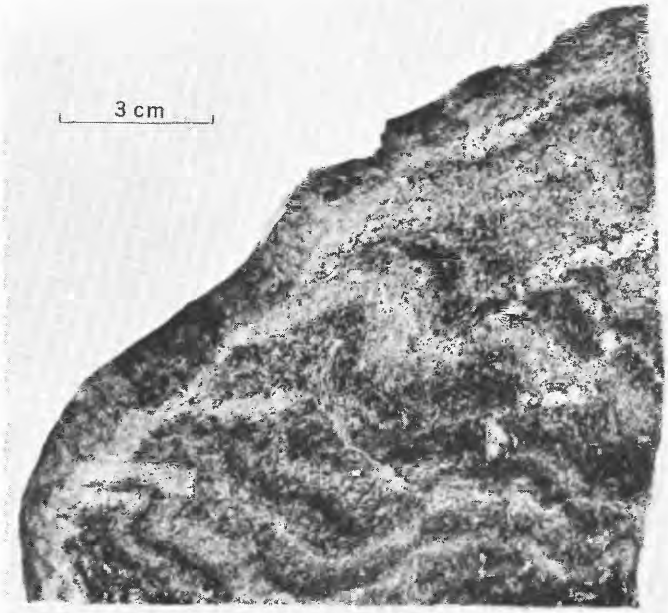

FIGURE 6.—A polished specimen of folded and banded biotite-plagioclase schist with muscovite-sillimanite laminae. The polished surface was cut perpendicular to the axis of small folds (parallel to *ac* plane). The dark layers are rich in sillimanite, muscovite, and biotite, and the light layers consist of quartz and small biotite flakes. Note an irregular pegmatitic veinlet on the right. Snake Creek (loc. 708, pl. 2).

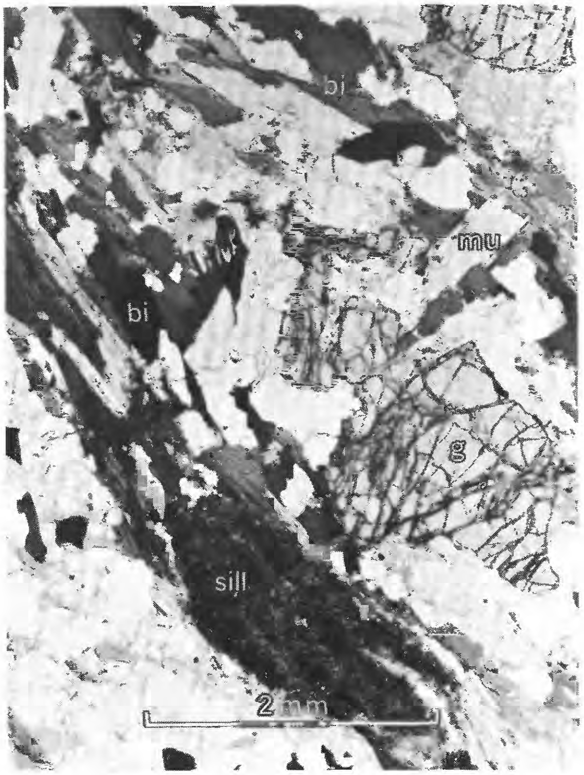

FIGURE 7.—Photomicrograph of the banded biotite-plagioclase schist with muscovite-sillimanite laminae. Along the North Fork of the Clearwater River (loc. 246, pl. 2). mu=muscovite; bi=biotite; sill=sillimanite; g=garnet. Plane polarized light.

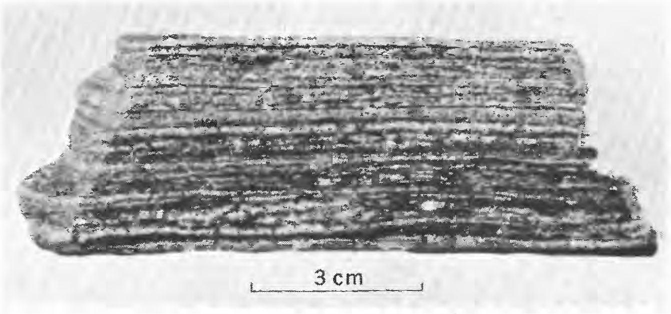

FIGURE 8.—A polished hand specimen of a thin-laminated gneissic layer in the schist. The dark laminae consist of sillimanite and biotite. The light layers are quartz, plagioclase, and some garnet. Snake Creek (loc. 706, pl. 2).

schist and gneiss than in the more homogeneous coarse-grained sillimanite-garnet schist (fig. 7). In a part of the laminated schist and gneiss, the individual layers are only from 0.5 to 2 mm thick. A polished hand specimen of this type of gneiss is shown in figure 8. The dark laminae consist of biotite and sillimanite, and the light-colored layers consist of quartz, plagioclase, and some garnet. The thin-bedded and laminated schist and gneiss occur next to the diopside-plagioclase gneiss in localities 706, 727, 729, and 739 (pl. 2).

The chemical and mineralogical composition of banded biotite-plagioclase schist is shown in table 2 as No. 246. The essential difference between this rock and the schist of No. 247 is in the abundance of plagioclase in the banded schist.

QUARTZITE AND GNEISS

Three petrographically different rock units are interbedded with the schists: (1) white granular quartzite interbedded with biotite quartzite and bordered by biotite-plagioclase gneiss, (2) diopside-plagioclase gneiss with quartzitic layers, and (3) thin-bedded diopside-plagioclase gneiss with biotite-bearing layers and biotite quartzite.

WHITE GRANULAR QUARTZITE AND BIOTITE-PLAGIOCLASE GNEISS

White granular quartzite grading to biotite quartzite and biotite-plagioclase gneiss is exposed in several localities along the North Fork of the Clearwater River at Big Island and at localities about 3½, 4½, and 6 miles north from it. In the upper part of the white granular quartzite 3½ miles north of Big Island (west of Riverview Lookout) the layers of pure quartzite, ranging from 1 to 3 cm in thickness, are intercalated with layers of biottite-plagioclase gneiss and biotite quartzite that range from 1 to 5 cm in thickness. In the middle part of this quartzite unit the thicknesses of the corresponding layers are 10 to 20 cm and 3 to 4 cm, and in the lowest part 10 to 50 cm and 3 to 5 cm. The estimated thickness of the white granular quartzite in this locality (loc. 268, pl. 1) is about 300 m. Toward the top and bottom the number and thickness of biotite-bearing layers increase and the quartzite grades to thin-bedded biotite quartzite with plagioclase-bearing layers. In some outcrops, as at Big Island, thin layers of diopside-plagioclase gneiss are interbedded with biotite quartzite. Comparison of the lithology of these quartzite units with that of the section at Surveyors Ridge (northeast corner of pl. 1) suggests that the white granular quartzite is equivalent to the lowest part of the Wallace formation. In the section along the North Fork of the Clearwater River the banded biotite-plagioclase schist with muscovite-sillimanite laminae and the coarse-grained sillimanite-garnet schist are stratigraphically above the white granular quartzite and are equivalent to the upper part of the Wallace formation on Surveyors Ridge (staurolite-garnet schist).

The white granular quartzite consists of a mosaic of irregularly shaped quartz grains with some interstitial plagioclase and very little biotite (fig. 9). In layers that are rich in plagioclase, the plagioclase has crystallized to form large grains around the small round quartz grains (fig. 10), but in the quartz-rich beds quartz is recrystallized as large grains enclosing round plagioclase grains (fig. 11). The biotite-plagioclase gneiss contains considerably more biotite and plagioclase than the white granular quartzite. Large biotite flakes form thin laminae between the layers (fig. 12), which consist of about 50 percent quartz, 35 percent plagioclase, 10 percent biotite, and some muscovite. Ilmenite, rutile, and zircon are the common accessories.

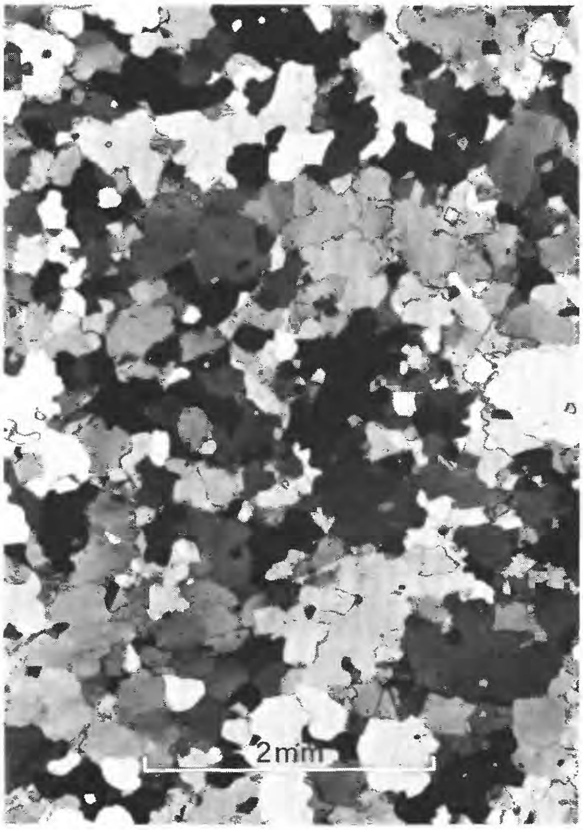

FIGURE 9.—Photomicrograph of a layer in white granular quartzite. Quartz grains are somewhat rounded or have irregular borders. The plagioclase is twinned and occurs as small grains between the quartz grains. Along the North Fork of the Clearwater River (loc. 254, pl. 1). Crossed nicols.

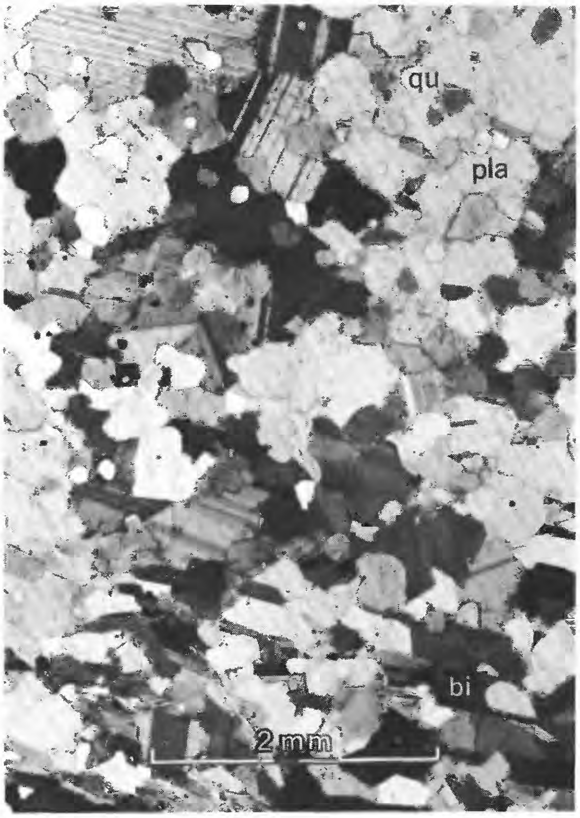

FIGURE 10.—Plagioclase-rich layer in white granular quartzite. The large grains of plagioclase (pla) enclose small round quartz grains (qu). bi=biotite. Along the North Fork of the Clearwater River (loc. 234, pl. 2). Crossed nicols.

The chemical and mineralogical composition of samples from these various gneiss and quartzite layers are shown in table 2 under Nos. 248, 252, 254.

DIOPSIDE-PLAGIOCLASE GNEISS WITH QUARTZITIC LAYERS

Each of the two or more gneiss-quartzite units that are interbedded with the coarse-grained sillimanite-garnet schist in the Headquarters quadrangle and vicinity is about 200 m thick and consists of coarse- to medium-grained light-green diopside-plagioclase-quartz rock. The variation in the amounts of the three main constituents—diopside, plagioclase, and quartz—

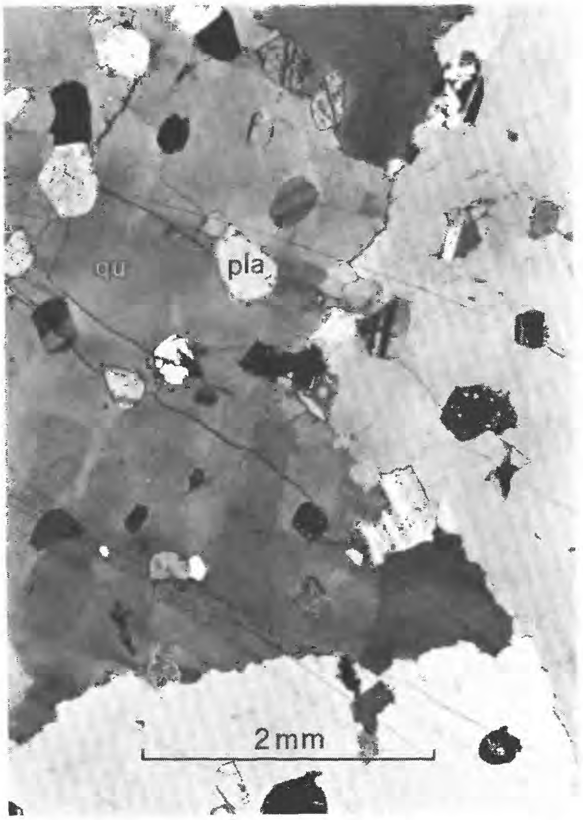

FIGURE 11.—Coarse-grained layer of white granular quartzite consists of large quartz (qu) grains which enclose small round plagioclase (pla) grains. Along the North Fork of the Clearwater River (loc. 268, pl. 1). Crossed nicols.

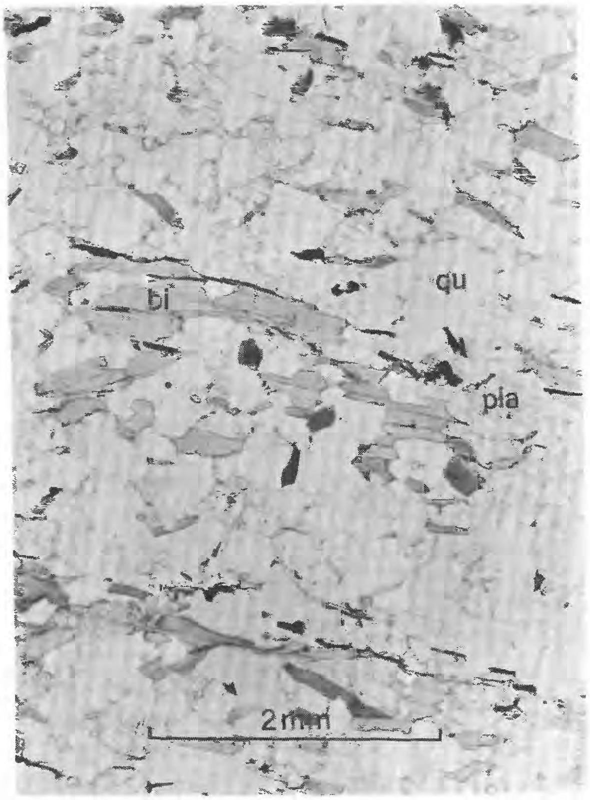

FIGURE 12.—Biotite-plagioclase gneiss layer next to the white granular quartzite. Biotite (bi) and scales of ilmenite (black) occur in thin layers separating the plagioclase-(pla) and quartz-(qu) rich layers. Along the North Fork of the Clearwater River (loc. 248, pl. 2). Plane polarized light.

gives rise on a small scale to a distinct banding, and on a larger scale to an irregular alternation of quartzitic and gneissic layers. The gneissic layers consist of about equal amounts of plagioclase, diopside, and quartz, whereas in the quartzitic layers the amount of quartz is larger; the pure quartzite layers, however, are rare. Locally hornblende is abundant instead of diopside and quartz (fig. 13). Potassium feldspar is generally very scarce or lacking. Abundant potassium feldspar was found only in a small outcrop near the contact of quartz diorite about 1 mile south of Headquarters.

The two diopside-plagioclase gneiss units in the southwestern part of the quadrangle have different textures. The lower of these two is exposed just north of the Columbia River basalt in the southern part of the quadrangle. It is a fine grained, dense, very hard, distinctly banded rock that contains abundant quartz and some biotite-bearing layers. The banding parallels the bedding and is due to variations in amount of the three main constituents—quartz, plagioclase, and diopside. In contrast to this dense texture of the lower unit the upper unit that occurs east of Whiskey Butte and extends along Snake Creek, is coarse to medium grained and has a sandy or gritty feeling if weathered. Under the miscoscope the texture of this upper unit appears equigranular and granoblastic, but in the lower unit many of the quartz grains have recrystallized as larger irregular-shaped grains and there is far less plagioclase.

Diopside occurs in subhedral grains that show $\alpha=1.684\pm0.001$, $\beta=1.691\pm0.001$ and $\gamma=1.709\pm0.001$ in specimen 289 (table 2). The plagioclase in the same specimen is bytownite (An_{85}); sphene and a few small grains of apatite are the accessories. The simple mineral composition of this rock allows fairly accurate calculation of the chemical composition of the diopside. In the calculation all aluminum was combined with alkalies to form plagioclase (An_{85}). Oxides to form 1 percent of sphene and 0.19 percent of apatite were subtracted, and the rest was calculated as quartz and diopside. The result (table 3) suggests that the atomic ratio Fe : Mg = 1 : 4 in this diopside.

Abundant light reddish brown grossularite, a mineral found in only a few places in the quadrangle, occurs in a calcareous layer exposed in a railroad cut near the southern border of the quadrangle (loc. 287, pl. 2) and on Dull Axe Mountain in the northeastern part of

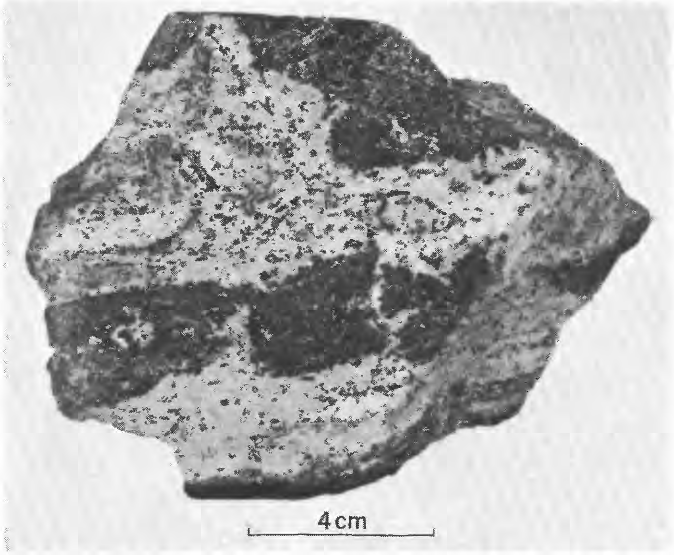

FIGURE 13.—A polished hand specimen of diopside-plagioclase gneiss with irregular areas rich in hornblende (dark). Crawford mine (loc. 373, pl. 2).

TABLE 3.—*Composition of diopside calculated from the rock analysis of specimen No. 289 (table 2 and pl. 2)*

Constituent	*Weight percent*	*Molecular equivalent*	*Number of atoms*		
SiO_2	53.79	8956	Si	2.00	
FeO	6.85	953	Fe	.21	1.00
MnO	.13	18	Mn	.00	
MgO	14.18	3517	Mg	.79	
CaO	25.05	4467	Ca	1.00	
			O	6.00	
Total	100.00				

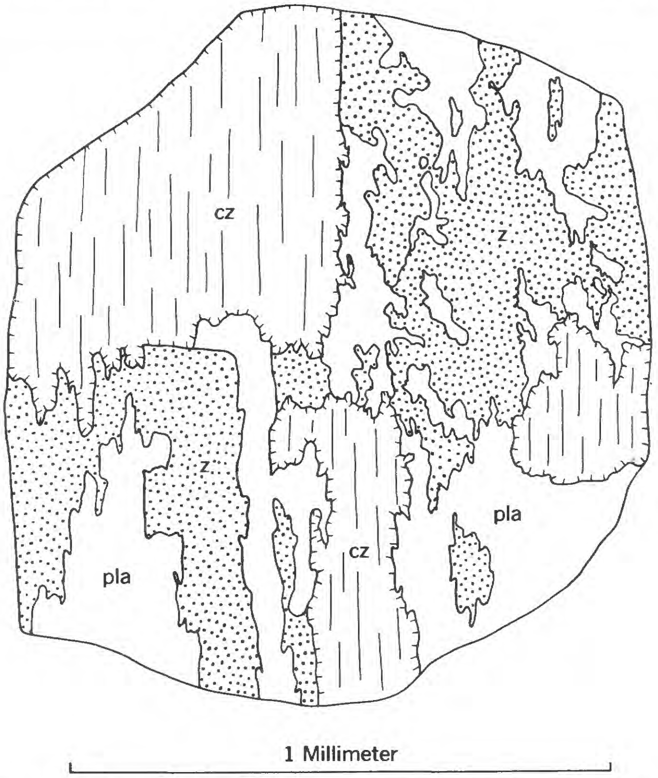

FIGURE 14.—Camera lucida drawing showing the relations among plagioclase (pla), zoisite (z), and clinozoisite (cz) in the calcium-magnesium-aluminum silicate rock at Dull Axe Mountain in the northeastern part of the Headquarters quadrangle (pl. 2).

the quadrangle. In rocks at both localities considerable amount of calcite occurs between the other minerals and as inclusions in grossularite and clinozoisite. The rock at the railroad cut is medium grained; that on Dull Axe Mountain is coarse grained and contains grayish-green diopside grains that range from 1 to 4 cm in diameter. The garnet crystals are of about the same size and show shiny clean crystal faces in hand specimens. The index of refraction of the grossularite on Dull Axe Mountain is $n=1.751\pm0.001$. The indices of refraction of the diopside, $\alpha=1.712\pm0.001$, $\beta=1.720\pm0.001$, $\gamma=1.734\pm0.001$, indicate that this diopside is considerably richer in iron than the diopside in the common diopside-plagioclase gneiss (compare specimen 289, table 2). Zoisite with $\beta=1.712\pm0.001$ and clinozoisite are abundant and show relations that suggest that they replace plagioclase and calcite (fig. 14). Probably anorthite crystallized first and was replaced by zoisite and clinozoisite under the influence of hydrous solutions. Grossularite was formed around diopside, calcite, quartz, and clinozoisite and was therefore the last mineral to crystallize. At all localities where grossularite was studied it is associated with epidote minerals and calcite.

At some localities west and north of the Headquarters quadrangle, also scapolite and phlogopite occur with other calcium-magnesium-aluminum silicates. The occurrences in the Headquarters quadrangle may have crystallized from the calcareous or dolomitic layers as a result of an introduction of silicon, and perhaps magnesium from the quartz dioritic magma, but elsewhere in the area under investigation similar coarse-grained rocks rich in diopside, grossularite, plagioclase, and epidote minerals and generally containing quartz, calcite, scapolite, and phlogopite occur in definite layers interbedded with diopside gneiss and quartzite. They are called lime-silicate rocks in this report and considered to be a result of more or less isochemical recrystallization of calcareous layers that contained appreciable amounts of magnesium and aluminum (dolomitic shales). It is clear that the same minerals would crystallize in thin limestone layers interbedded with shales and argillites because of a short-distance migration of silicon, magnesium, and aluminum. Quartz and calcite are common constituents in many layers and there is every transition from the lime-silicate rock to quartzite or to marble with only a small amount of diopside and grossularite.

A layer of peculiar epidote-chlorite rock was found near Big Island (loc. 743, pl. 3). Green crystals of epidote ranging from 2 to 6 mm in length, oriented at random, and showing $\gamma=1.740\pm0.001$, are the major

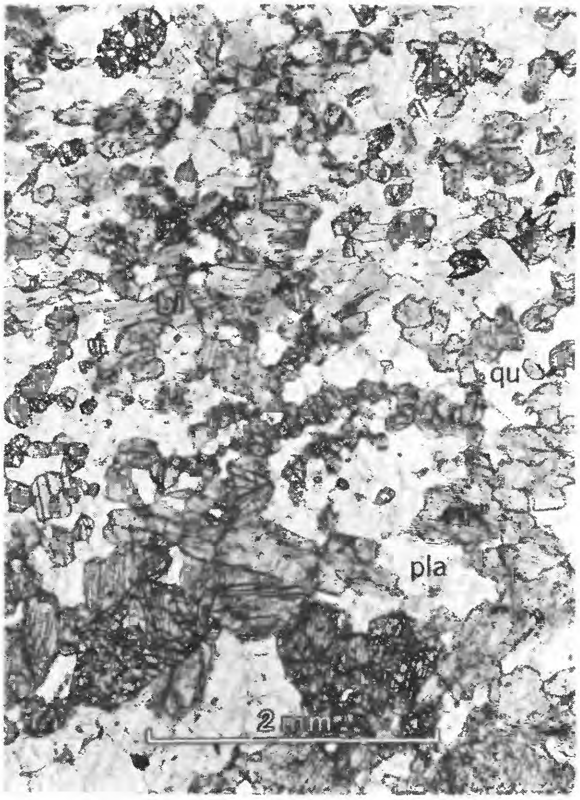

FIGURE 15.—Photomicrograph of thin-bedded diopside-plagioclase gneiss with biotite-bearing layers. The biotite-bearing layers are, as a rule, finer grained than the layers of diopside-plagioclase gneiss; qu=quartz; pla=plagioclase; di=diopside; bi=biotite. Cougar Creek (loc. 331, pl. 2). Plane polarized light.

constituents of this layer. The interstices between the epidote crystals are filled by vermicular aggregates of green chlorite which show $\gamma=1.610\pm0.001$, $\gamma-\alpha=0.010$, and a weak pleochroism (Z=pale green, X=colorless), suggesting that this mineral is prochlorite.

THIN-BEDDED DIOPSIDE-PLAGIOCLASE GNEISS WITH BIOTITE-BEARING LAYERS AND BIOTITE QUARTZITE

Thin-bedded diopside-plagioclase gneiss with biotite-bearing layers and biotite quartzite are exposed in the north central part of the Headquarters quadrangle southwest of Casey Cabin. These rocks are complexly folded and faulted and intruded by small bodies of diorite and tonalite. They are bordered to the east by the Idaho batholith, and many of the schist and quartzite layers are strongly migmatized. A sequence of laminated sillimanite-biotite schist, diopside-plagioclase gneiss, thin-bedded diopside-plagioclase gneiss with biotite-bearing layers, and thin-bedded biotite quartzite is exposed in several localities in this part of the quadrangle.

The diopside-plagioclase gneiss with biotite-bearing layers is thin bedded and locally contains hornblende in addition to quartz, plagioclase, microcline, and diopside. The alternation of thin beds of fine-grained diopside-plagioclase gneiss and biotite gneiss is characteristic of this rock unit. The diopside-bearing layers are light green and consist of quartz, plagioclase, and diopside. The layers of dark-brownish biotite gneiss contrast strongly with these light-green beds. The thickness of the individual beds ranges from 1 to 3 cm, the thickness of 2 cm being most common.

Thin sections show that the biotite-bearing layers consist of quartz, diopside, biotite, plagioclase, and orthoclase. In the light-green layers biotite is lacking and the rock is coarser grained (fig. 15). Sphene is a common accessory in all layers.

The thin-bedded diopside-plagioclase gneiss with biotite-bearing layers exposed along Cougar Creek just northeast of locality 331 (pl. 2) is overlain by a very thin bedded biotite quartzite. This quartzite consists of thin layers of light-gray fine-grained quartzite that are separated by dark laminae rich in biotite. The thickness of the light-colored layers ranges from 2 to 15 mm, whereas the dark laminae are 1 to 3 mm thick. A few thick layers of medium-grained biotite-bearing quartzite are interbedded with the thin-bedded biotite quartzite.

About one-third of the light minerals in the thin-bedded biotite quartzite is plagioclase (An_{35-38}) and two-thirds quartz. Both minerals occur as small polygonal grains. Small flakes of biotite are evenly distributed through the light-colored layers. In the dark laminae the biotite is recrystallized as large flakes and concentrated parallel to the bedding planes. Small grains of zircon and apatite are the accessories.

The beds of medium-grained biotite quartzite range from 5 to 50 cm in thickness and are very light gray and well foliated. The biotite flakes range from 2 to 3 mm in their longest dimension and are evenly distributed throughout the bed. This quartzite weathers readily to quartz sand.

Layers of muscovite-biotite schist are interbedded with thin-bedded quartzite. This schist consists of quartz, biotite, muscovite, and occasional sillimanite. It resembles the layers of muscovite-biotite schist in the Wallace formation. Some layers are laminated: thin layers of biotite quartzite are separated by muscovite laminae. The lithologic character of the quartzitic layers is more variable than that of the schists, and the quartzite layers are therefore more useful in correlating the formations.

REVETT QUARTZITE

A coarse-grained thick-bedded quartzite body more than 3 square miles in area is exposed in the northeastern corner of the quadrangle. Its individual beds are from 20 to 100 cm thick. They are separated by thin muscovite laminae. Most beds are very coarse grained and contain only a few plagioclase grains and muscovite or biotite flakes. Some beds contain a considerable amount of sillimanite. The color ranges from white to bluish or reddish white.

The total thickness of this quartzite body could not be measured, but the structure of surrounding rocks suggests that the thickness corresponds to that of the Revett quartzite in Shoshone County where it is 500 m (Umpleby and Jones, 1923).

Microscopic study shows the quartz as deformed grains with strain shadows. The additional constituents—plagioclase, biotite, muscovite, and sillimanite—occur between the quartz grains. Rutile and sphene are the accessories.

CORRELATION OF THE BELT SERIES

The metasedimentary rocks can be most closely correlated with rocks described by Calkins and Jones (1911) and Umpleby and Jones (1923) from an area between the St. Joe River and the Little North Fork of the Clearwater River. This area is about 35 miles north of the northern border of the Headquarters quadrangle, and the rocks of the Belt series there are much less metamorphosed. The geologic map by Calkins and Jones shows, on Surveyors Ridge east of the Little North Fork of the Clearwater River (pl. 1) a stratigraphic sequence of the Belt series from the Prichard formation to the Wallace formation. I visited this area in 1954, studied samples of all formations, and prepared the section shown in table 4. The correlation of the metasedimentary rocks of the Headquarters quadrangle was based mainly on the comparison of their lithology with that of this section on Surveyors Ridge (table 4). A paper by Wagner (1949) gives additional information about the lithology of the Burke, Revett, St. Regis, Wallace, and Striped Peak formations on the south slope of the St. Joe Mountains. Wagner subdivided the Wallace formation into five members (1 to 5 in table 4). Comparison of the thickness and lithology of the section on Surveyors Ridge with those of these five members suggests that only a lower part of the Wallace formation (1 to 3 in Wagner's subdivision) is exposed there. The thickness and lithology of the sequence of sillimanite-garnet schist, diopside-plagioclase gneiss, and garnet-mica schist in the Headquarters quadrangle suggest that these rocks are metamorphosed equivalents of the three upper members of the Wallace formation in the St. Joe Mountains. According to Wagner's subdivision (1949, p. 12–13) non-calcareous shale with interbedded dolomitic sandstone and sandy shale form three upper members of the Wallace formation. In a higher temperature and pressure these rocks would be recrystallized as schist and diopside-bearing gneiss. The sillimanite-garnet schist and diopside-plagioclase gneiss can also be correlated with the rocks that were ascribed to the upper part of the Wallace formation by Ransome and Calkins (1908) and by Calkins and Jones (1911, p. 77–80). The white granular quartzite with thinner layers of biotite quartzite that is exposed in the canyon of the North Fork of the Clearwater River west of Riverview Lookout and on the bottom of the river at Big Island is underlain by dark biotite-rich quartzite and overlain by thin-bedded biotite quartzite (table 5). This sequence is similar to the lower part of the Wallace formation on Surveyors Ridge. The pure coarse-grained thick-bedded quartzite that is exposed in the northernmost part of the Headquarters quadrangle is most likely equivalent to the Revett quartzite of Shoshone County.

TABLE 4.—*Generalized sections of the Belt series on Surveyors Ridge and in the St. Joe Mountains*

St. Joe Mountains (Wagner, 1949)				Surveyors Ridge		
Formation		Thickness (feet)	Lithology	Estimated thickness		Lithology
				Feet	Meters	
Wallace	5	500	Black thin-laminated noncalcareous shale.			
	4	550	Sandy calcareous or dolomitic shale.			
	3	850	Black thin-laminated noncalcareous shale.	900	274	Staurolite-kyanite schist. Banded biotite quartzite and biotite-muscovite-garnet schist with muscovite laminae.
				60	18	Very thin laminated quartzitic muscovite-biotite schist.
	2	600	Calcareous sandstone and shale.	1,200	366	Thin-bedded biotite quartzite with light-gray to white granular layers; in places rich in biotite and scapolite.
	1	2000	Impure quartzite, sandstone, limestone, and shale.	1,500	457	White granular quartzite with thin biotite-bearing layers.
St. Regis		600 to 800	Green and purple banded, thin-bedded quartzite and siliceous shale.	800	244	Biotite-muscovite-garnet schist with thin quartzitic beds. Homogeneous fine-grained biotite quartzite.
Revett		800	Thick- bedded white quartzite.	700	213	Coarse-grained thick-bedded quartzite; contains some muscovite and biotite.
Burke		Not known.	Micaceous and quartzitic schist and schistose quartzite.	600	183	Gray granular thin-bedded quartzite with biotite-rich layers and some biotite schist layers.

A very thin-bedded biotite quartzite south of Casey Cabin dips under the Revett quartzite and is probably faulted against it. A diopside-plagioclase gneiss interbedded with white granular quartzite, biotite quartzite

TABLE 5.—*Section of metasedimentary rocks near Big Island*

Formation	Correlation to subdivision in table 4	Estimated thickness		Lithology
		Feet	Meters	
Middle(?) part of the Wallace.	3(?)			Coarse-grained muscovite-biotite schist.
				Banded biotite-plagioclase gneiss with muscovite-sillimanite laminae.
		65	20	Thin-bedded biotite quartzite.
Fault				
Lower part of the Wallace.	2	65	20	Thin-bedded diopside-plagioclase gneiss with layers of biotite quartzite.
		100	30	Biotite schist.
		100	30	Diopside-plagioclase gneiss and quartzite.
	1	330	100	White granular quartzite.

TABLE 6.—*Section of metasedimentary rocks along Cougar Creek*

Formation	Correlation to subdivision in table 4	Estimated thickness		Lithology
		Feet	Meters	
Lower part of the Wallace (?)	2	100	30	Thin-bedded biotite quartzite.
		65	20	Diopside-plagioclase gneiss.
		65	20	Biotite quartzite.
		33	10	Diopside-plagioclase gneiss.
		650	200	Biotite schist.
		100	30	Diopside-plagioclase gneiss.
	1	200	60	Medium-grained light-colored quartzite.

and biotite gneiss occurs southwest of it (table 6). It is not certain whether the biotite quartzite is equivalent to the Burke formation or to a part of the Wallace formation which covers the main part of the Headquarters quadrangle. A similar sequence of white granular quartzite interbedded with diopside-plagioclase gneiss and overlain by thin-bedded biotite quartzite is exposed along Calhoun Creek about 1 mile west of the Clearwater Timber Protective Association (C.T.P.A.) headquarters (pl. 2). There this sequence, probably equivalent to the lower part of the Wallace formation, is underlain by a thick layer of sillimanite-garnet schist which is considered to be equivalent to the upper part of the Wallace formation. A fault separates the schist from the quartzite west of Summit Lookout.

THICKNESS OF FORMATIONS

Because of a discontinuity of outcrops, intense folding, and faulting, the thickness of the formations could not be measured. A section of the upper part of the Wallace formation in the deep gorge of Reeds Creek shows that there are two major units of diopside-plagioclase gneiss, each about 200 m thick, interbedded with the coarse-grained garnet-sillimanite schist. The estimated thicknesses of similar units elsewhere are less, partly because of faulting and partly because of lateral change in mineralogy of the layers. The schist unit between the gneiss units is about 200 m thick and about 150 m of schist is exposed above the second gneiss unit. These estimated thicknesses correspond to the thicknesses of the upper sandy and shaly members of the Wallace formation given by Wagner (1949, p. 12, 13). The thickness of the white granular quartzite member north of Big Island—estimated at 300 m—is thinner than the white granular quartzite on Surveyors Ridge. The total estimated thickness of the sequence that is thought to be the metamorphosed equivalent of the Wallace formation is thus 1,050 m.

CHEMICAL COMPOSITION AND MINERAL ASSOCIATIONS

The chemical analyses in table 2 (pl. A-11) show that the schist layers have a composition typical of argillites. The percentage of aluminum is high, and there is more iron than magnesium. There is a considerable amount of potassium, but it is all contained in biotite and muscovite. There is scarcely any potassium feldspar.

The schist layers in the Headquarters quadrangle show considerable amounts of corundum in their norm. In the biotite gneiss and quartzite the excess of aluminum is only slight, and abundant wollastonite appears in the norm of the diopside-plagioclase gneiss. These layers were originally dolomitic sands with a rather high content of aluminum, resulting in the crystallization of a considerable amount of plagioclase. Potassium feldspar is rare or completely lacking, although it is fairly common in the Wallace formation in the areas farther north. The lack of it in the schists, gneiss, and quartzites in the Headquarters quadrangle is due either to the facies change during sedimentation or to the removal of potassium during metamorphism. If one considers that there has been an extensive metasomatic exchange of material in a wide area northwest of the Idaho batholith, as described in later sections, it seems more probable that a part of the potassium was removed from the metasedimentary rocks near the contact of the Idaho batholith rather than that there was a facies change during sedimentation over a distance of about 15 miles around the batholith.

The mineral associations, sillimanite-garnet-biotite-muscovite, in the schist and the occurrence of diopside in the quartzite and the gneiss indicate that temperature and pressure during metamorphism corresponded to those prevailing during recrystallization to of the amphibolite facies. A part of the sillimanite was altered to sericite at a later time. Locally, biotite and garnet show some alteration to chlorite.

METASOMATIC CHANGES

The chemical and mineralogical composition of the schist, gneiss, and quartzite deviates locally from that commonly found in the strata. The change in the composition takes place parallel to the beds as well as across them. In the field this is especially noticeable in the diopside-plagioclase quartzite and gneiss, where hornblende replaces diopside and quartz. Accumulations of biotite in the schist are found near the contacts of tonalite.

METASOMATIC HORNBLENDE IN DIOPSIDE-PLAGIOCLASE GNEISS

Some local parts of the diopside-plagioclase gneiss contain hornblende either instead of diopside or in addition to this mineral. Together these parts may form about 15 percent of the total amount of the gneiss exposed in the Headquarters quadrangle and vicinity. The distribution of hornblende crystals in the diopside-plagioclase gneiss is either irregular (fig. 13) or parallel to the bedding (fig. 16). Commonly, patches measuring from a few centimeters to 20 m in diameter consist mainly of hornblende and plagioclase, whereas sporadic grains of hornblende appear in the surrounding gneiss. In many outcrops the sporadic grains as well as the hornblende in the patches form large holoblasts, which are conspicuous against the light-green fine-grained host rock (fig. 17, 1 and 2). In many outcrops the hornblende-bearing patches cut abruptly across the bedding.

The thin-section studies show that there is no diopside and very little quartz in the spots where hornblende occurs; the amount of plagioclase may be the same or larger in the hornblende-bearing part of the strata. A camera lucida drawing (fig. 18) of a thin section made from the same specimen of the hornblende-bearing rock as shown in figure 17 shows details of the texture of a hornblende-bearing spot. In the diopside-plagioclase layer, next to this spot, quartz occurs as large grains that include small plagioclase and diopside grains. Where hornblende crystals were developed, the small plagioclase grains remain as inclusions in the large euhedral hornblende holoblasts but quartz and diopside are absent. This suggests that both of these minerals were replaced by hornblende. Peculiarly, the plagioclase grains seem to be unattacked and remain in their original positions.

Late hornblende, growing parallel to the bedding or on both sides of the quartz-plagioclase veinlets, is common in all diopside-plagioclase gneiss (fig. 19). The crystallization of hornblende parallel to the bedding could be explained by the variation in the composition of the original sediment, but the lower specimen in figure 19 shows that the hornblende was crystallized later as a result of introduction of elements in solutions moving along a fissure. Abundant euhedral apatite was crystallized in the dikelet with hornblende and plagioclase. Large anhedral plagioclase grains also crystallized late in the gneiss. The plagioclase and hornblende grains enclose numerous small rounded quartz grains. Small grains of zircon, sphene, and magnetite occur as accessories.

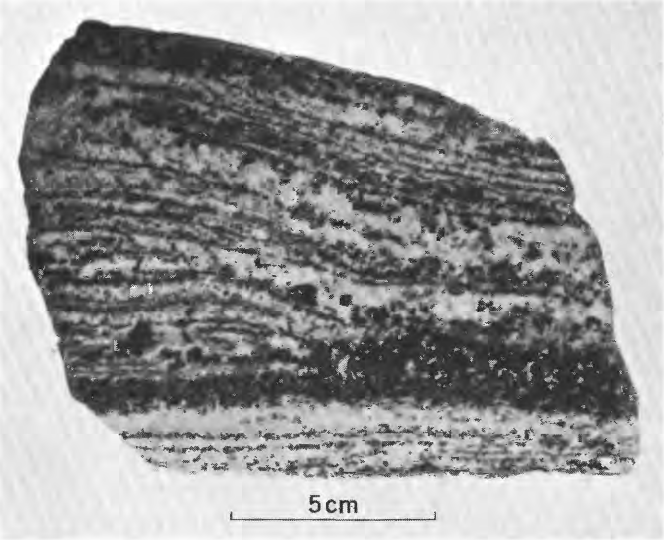

FIGURE 16.—Hornblende parallel to the bedding in the diopside-plagioclase gneiss along Snake Creek (loc. 705, pl. 2).

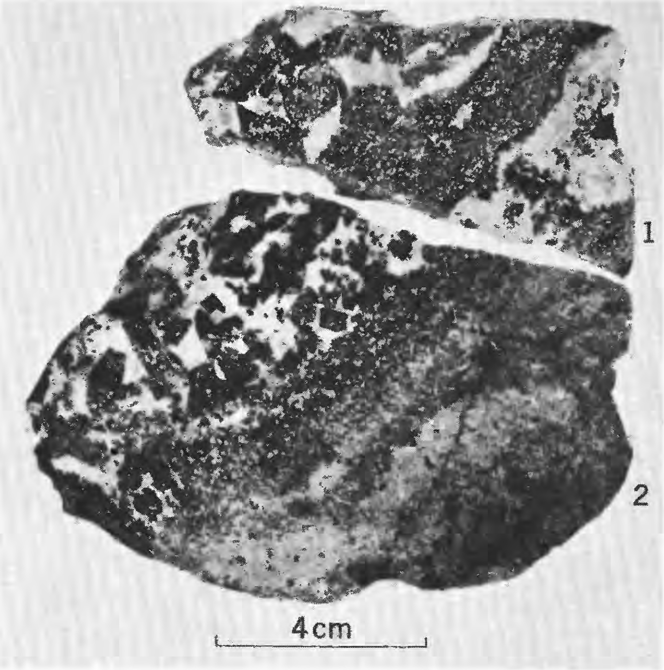

FIGURE 17.—Holoblasts of euhedral hornblende locally replace quartz and diopside in the diopside-plagioclase gneiss. The upper specimen (No. 1) is from locality 365 west of Bald Mountain (pl. 2). The lower specimen (No. 2) is from locality 140 along the North Fork of the Clearwater River about 6 miles southwest of Big Island (pl. 3).

In some localities, larger portions of diopside-plagioclase gneiss have recrystallized as coarse-grained hornblende-plagioclase rock. A body of such rock

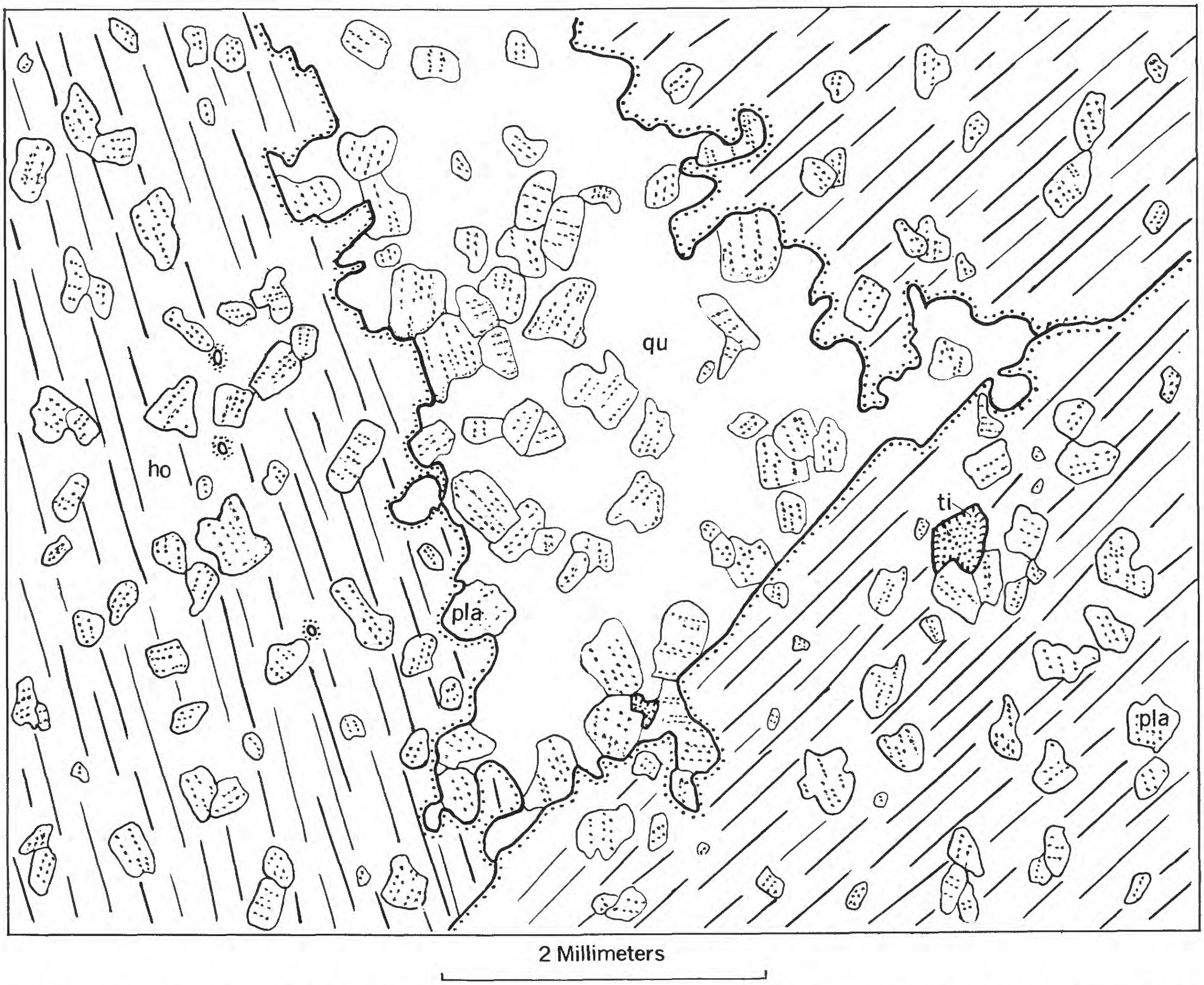

FIGURE 18.—Camera lucida drawing of a diopside-plagioclase gneiss west of Bald Mountain (loc. 365, pl. 2). Late hornblende (ho) has replaced quartz and diopside but includes rounded plagioclase grains (pla) which in the unaltered rock are included in quartz (qu); ti=sphene. Photograph of this specimen is shown in upper figure (1) on figure 17.

about 10 m thick is exposed along the North Fork of the Clearwater River about 3 miles southwest of Big Island (loc. 140, pl. 3). Hornblende in this rock occurs as conspicuous euhedral holoblasts including numerous small round grains of quartz and plagioclase (fig. 20) resembling the sporadic hornblende crystals in specimen 365 (fig. 17, upper part). The contact between the hornblende-plagioclase rock and the diopside-plagioclase gneiss cuts irregularly across the bedding. Abundant small grains of hornblende occur along the contact which is about 3 cm in width (fig. 17, lower part). Chemically, these hornblende-plagioclase rocks are close to quartz diorite except for less potassium because of lack of biotite. Table 7 shows the chemical and mineralogical composition of the hornblende-plagioclase rock No. 140. This rock contains about 30 percent quartz, which is the same amount as in most of the unaltered diopside-plagioclase gneiss. Only a small addition of water may have been needed to change this diopside-plagioclase gneiss to the coarse hornblende-plagioclase rock.

To find out whether there was any change in the Fe:Mg ratio during the replacement of diopside by hornblende, the indices of refraction of these minerals were determined in several specimens (table 8). The molecular percentages of the iron compound for diopside and hornblende in various specimens were determined on the basis of these measured indices and Winchell's diagrams (1951). To check the accuracy of this method, the iron content of the hornblende in the hornblende-plagioclase rock No. 140 was determined also by calculating the composition of hornblende (table 9) from the rock analysis in table 7. The iron content of diopside in sample 289 was found by

TABLE 7.—*Chemical composition, ionic percentage, norms, and minerals of the hornblende-plagioclase rock No. 140 (pl. 3)*

[Harry M. Hyman, analyst]

Constituent				Norm			Minerals	
Conventional symbol	Symbol rearranged for cation percent	Weight percent	Cation percent	Mineral	CIPW	Molecular		
SiO_2		65.49	62.14	Q	26.72	25.35	Quartz	30.8
Al_2O_3	$AlO_{3/2}$	9.66	10.80	Or	2.11	2.15	Plagioclase	25.9
							(Anorthite content)	(37)
Fe_2O_3	$FeO_{3/2}$	.45	.32	Ab	16.67	18.15	Orthoclase	1.9
FeO		5.13	4.07	An	16.46	16.85	Hornblende	39.5
MnO		.08	.06	Wo	9.46	9.30	Apatite	.4
MgO		6.87	9.71	En	17.11	19.42	Sphene	1.3
CaO		8.11	8.24	Fs	8.13	7.02		
Na_2O	$NaO_{1/2}$	1.97	3.63	Ap	.37	.34		
K_2O	$KO_{1/2}$	.36	.43	Il	1.23	.92		
TiO_2		.65	.46	Mt	.65	.48		
P_2O_5	$PO_{5/2}$	.15	.13	Cc	.02	.02		
CO_2		.01	.01	H_2O	.82			
H_2O+		.79	(2.50)					
H_2O-		.03						
Total		99.75	100.00		99.75	100.00		99.8
	O		163.84					
	OH		5.00					
	Total anions		168.84					

calculating the composition of diopside (table 3) from the chemical analysis of the diopside-plagioclase gneiss No. 289. Comparison of the iron content of the diopside in the parent rock (column 1, table 8, diopside-plagioclase gneiss) with the iron content of the hornblende in the metasomatically altered part of the same rock (columns 2 and 3, table 8) shows a considerably higher content of iron in hornblende in two specimens (Nos. 140 and 289), a little higher content of iron in hornblende in specimen 365, and a somewhat lower one in specimen 837. The results seem contradictory and need some explanation in order to be evaluated correctly. In specimen No. 837 the amphibole is actinolite that occurs as small granoblastic grains similar to those of the diopside, being thus of metamorphic origin. The low content of iron in this metamorphic hornblende is in accordance with Ramberg's suggestion (1952a) that in metamorphic rocks crystallized in equilibrium the Mg:Fe ratio increases in the series orthosilicate → pyrosilicate → metasilicate → double-chain silicate → phyllosilicate → tectosilicate, but a substitution of aluminum for silicon in the structure tends to decrease the Mg:Fe ratio.

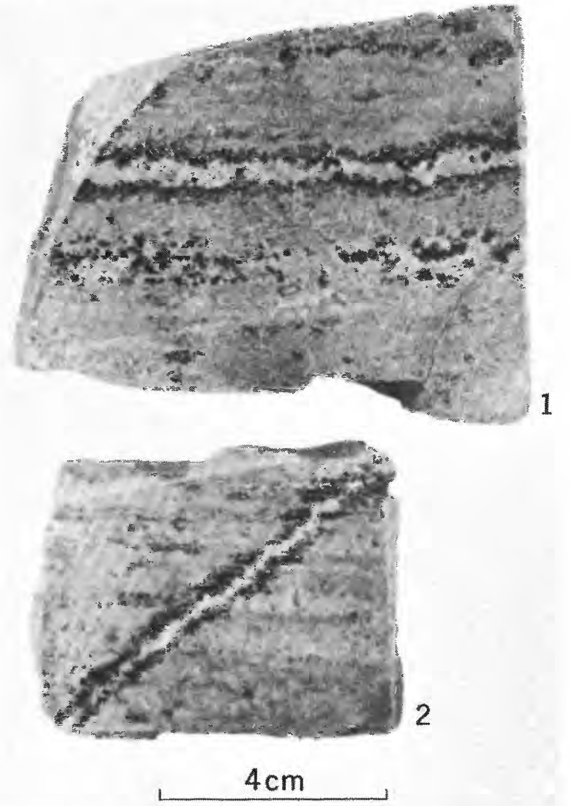

FIGURE 19.—Polished hand specimens of diopside-plagioclase gneiss with secondary hornblende and plagioclase: No. 1, parallel to the bedding; No. 2, along a crack. Locality 208 along the North Fork of the Clearwater River (pl. 1).

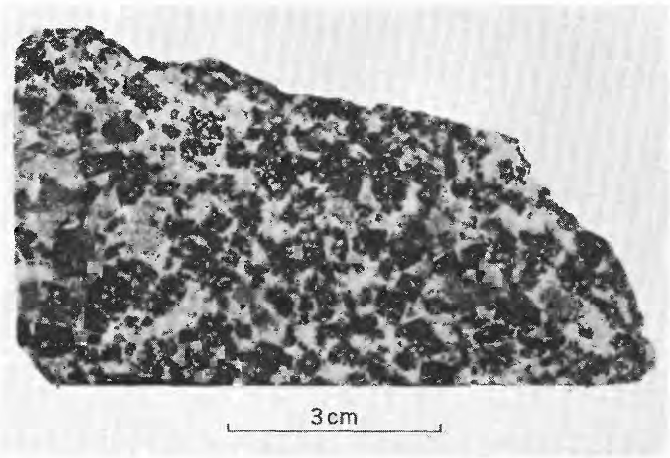

FIGURE 20.—Metasomatic hornblende-plagioclase rock along the river 3 miles southwest of Big Island (loc. 140, pl. 3). The white inclusions in the dark euhedral hornblende holoblasts are quartz and plagioclase.

Because the hornblende in this specimen (No. 837) is not derived from replacement of diopside, this sample cannot be used for determining whether there was change in the Fe:Mg ratio. In the other three specimens the hornblende is secondary, and only they should be considered. Among these three, Nos. 140 and 289 indicate an appreciable increase of iron during the crystallization of hornblende, whereas the iron content in No. 365 remained practically the same. This suggests that during the crystallization of secondary hornblende some iron was introduced locally.

TABLE 8.—*Optical properties of diopside and hornblende from gneissic layers of the Wallace formation at different localities*

Optical properties	1 Diopside in the diopside-plagioclase gneiss	2 Hornblende in diopside-plagioclase gneiss	3 Hornblende in the plagioclase-hornblende rock [1]
From loc. 140			
α	1.679±0.001	1.638±0.001	1.633±0.001
β	1.687±0.001	1.654±0.001	1.640±0.001
γ	1.705±0.001	1.660±0.001	1.657±0.001
$Z \wedge c$	38°	23°	23°
Mol percent Fe	20	30	30
From loc. 365			
α	1.677±0.001	1.627±0.001	
β	1.685±0.001		
γ	1.702±0.001	1.647±0.001	
$Z \wedge c$	40°	19°	
Mol percent Fe	18	20	
From loc. 289			
α	1.684±0.001	1.633±0.001	
β	1.691±0 001	1.640±0.001	
γ	1.709±0.001	1.657±0.001	
$Z \wedge c$	44°		
Mol percent Fe	21	30	
From loc. 837			
α	1.675±0.001	[2] 1.612±0.001	
β	1.683±0.001	1.625±0.001	
γ	1.701±0.001	1.640±0.001	
$Z \wedge c$	44°	25°	
Mol percent Fe	16	13	

[1] A portion of the diopside-plagioclase gneiss at locality 140 has recrystallized as plagioclase-hornblende rock (left side of the lower specimen in figure 17; specimen in figure 20 is from this part of the rock).

[2] This amphibole is actinolite.

TABLE 9.—*Chemical composition of hornblende calculated from the rock analysis of specimen No. 140, table 7*

Constituent	*Weight percent*	*Molecular equivalent*	*Number of atoms*		
SiO_2	51.49	8573	Si	7.38	8.00
Al_2O_3	4.28	420	Al	.62	
FeO	11.46	1595	Al	.10	5.15
MnO	.16	23	Fe	1.37	
MgO	16.76	4157	Mn	.02	
TiO_2	.75	94	Mg	3.58	
CaO	13.00	2318	Ti	.08	
H_2O	2.10	1166	Ca	2.00	
			H	2.01	
Total	100.00		O	24.00	

HORNBLENDE AND PLAGIOCLASE IN THE SCHIST

In some localities late hornblende is crystallized in schist as large anhedral holoblasts. Figure 21 shows this kind of hornblende in fine-grained biotite gneiss near its contact with diorite about 1 mile south of Big Island (loc. 299, pl. 3). The biotite gneiss is a layer in biotite schist and averages about 50 cm in thickness. It is one of the few layers observed that contain potassium feldspar in addition to quartz and plagioclase, which are the typical light-colored minerals in the metamorphosed sedimentary rocks as well as in the intrusive rocks of the Big Island area and the Headquarters quadrangle.

FIGURE 21.—Large secondary hornblende holoblasts in a fine-grained biotite gneiss near a contact with diorite. Locality 299 along the river, 1 mile south of Big Island (pl. 3).

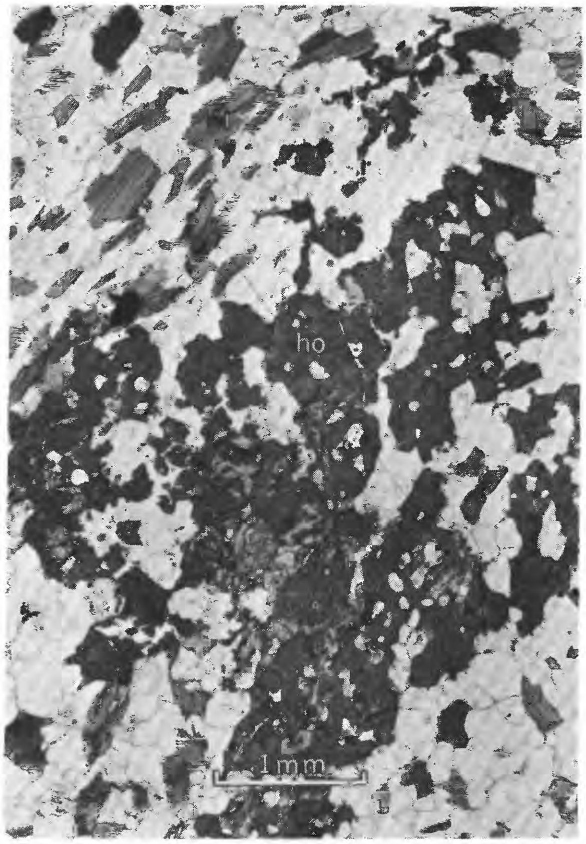

FIGURE 22.—Photomicrograph of the rock shown in figure 21. The light-colored minerals are quartz, microcline, and plagioclase; bi=biotite; ho=hornblende. Ordinary light.

Study under the microscope suggests that the hornblende has replaced the biotite (fig. 22)—the only dark mineral in the unaltered part of the gneiss— as well as part of the light-colored constituents. The parallel orientation of biotite in the immediate vicinity of the hornblende holoblasts is disturbed. Sphene, apatite, and zircon are the accessories in the hornblende-bearing part of the gneiss. Generally there is more microcline in the vicinity of the large hornblende holoblasts than in the unaltered biotite gneiss. This suggests that the iron and magnesium of the biotite joined to form hornblende with the introduced elements and that the potassium crystallized as microcline. The large amount of hornblende suggests that considerable amounts of iron, magnesium, calcium, and aluminum were introduced. The decrease of quartz shows that mainly silicon was removed. The nearness of the quartz diorite body suggests that the crystallization of the hornblende is connected with the emplacement of the quartz diorite. However, its crystallization is not a contact phenomenon in a strict sense, because unaltered layers of schist occur between the quartz diorite and the hornblende-bearing schist.

A few small sill-like bodies of amphibolite occur in the schist; they commonly are adjacent to tonalite. The amphibolite bodies are from 2 to 10 m thick and 5 to 100 m or more long. Only the largest of them are indicated on the maps (pls. 2 and 3). The amphibolite is medium grained and dark and has a gneissic texture. Dark minerals make up from 50 to 80 percent of the rock. About 70 percent of the light-colored minerals are quartz, and only 30 percent plagioclase. The dark minerals are hornblende and biotite. Magnetite, apatite, sphene, and zircon occur as accesories. The texture is lepidoblastic. In some bodies (as No. 249) the hornblende is a common dark-green, strongly pleochroic variety, but in the others (as No. 274) it is a pale-green variety with a very weak pleochroism. The pale-green hornblende has a larger grain size and contains small round inclusions of quartz which give an impression that the hornblende is a late mineral. Biotite is strongly pleochroic, brown or reddish brown, and well alined parallel to the platy structure of the country rocks. Plagioclase grains are small and anhedral and include some sericite. The cracks in them are filled by green antigorite. Small grains of quartz are found with plagioclase.

The mode of occurrence of quartz, plagioclase, and biotite indicates that they are the primary constituents whereas the hornblende crystallized late. It seems, therefore, that these amphibolite bodies are a result of a late development of abundant hornblende in the biotite-plagioclase schist.

As a rule the tonalite bodies are larger than those consisting of amphibolite. The largest bodies are shown on the maps; small bodies are common in areas where symbols for plagioclase occur (pl. 3). These small bodies range from a few centimeters to several hundred meters in length. Their contacts are gradational or interfingering. The tonalite contains abundant large round plagioclase grains and interstitial quartz and biotite. The texture of the tonalite is gneissic, the biotite flakes being subparallel to the platy structure of the country rocks. In many localities, large round plagioclase grains occur in the schist next to the tonalite and the amphibolite. In locality 228 (pl. 2) these round plagioclase grains contain abundant small round inclusions of quartz and a few small biotite flakes. The main part of the biotite occurs between the round plagioclase grains (fig. 23). The biotite flakes are not parallel to the foliation as they are in the schist but are oriented at random in a manner suggesting that the late growth of the large plagioclase grains pushed them into interstices.

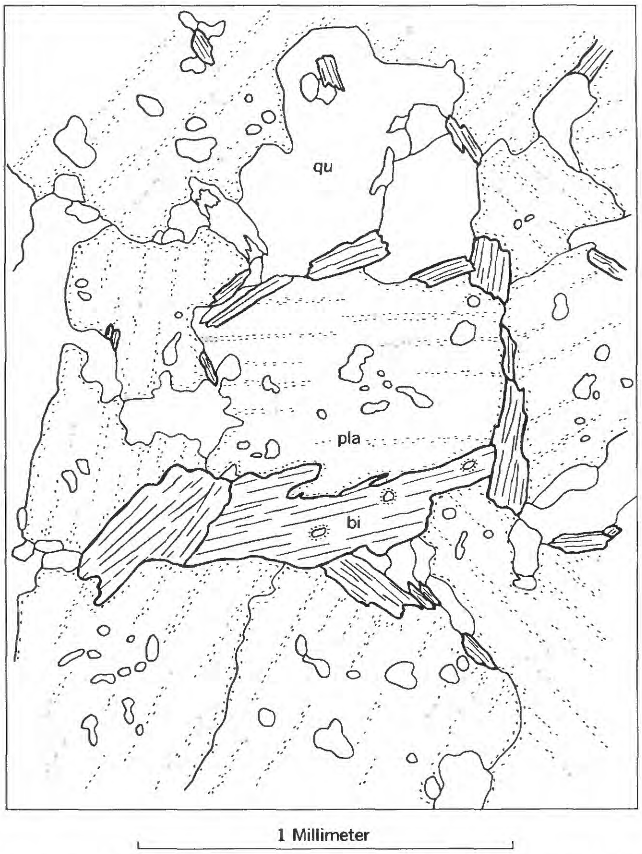

FIGURE 23.—Camera lucida drawing of a gneissic tonalite (No. 228) from the North Fork of the Clearwater River (pl. 2); pla=plagioclase; bi=biotite; qu=quartz. Small round inclusions in plagioclase are quartz.

The common occurrence of tonalite next to the amphibolite suggests that these two rock types in this area are genetically related. The late crystallization of hornblende in the amphibolite suggests that iron, magnesium, calcium, and aluminum were introduced after the recrystallization of quartz, plagioclase, and biotite in the schist. The alinement of the hornblende parallel to the platy structure indicates that the same

FIGURE 24.—Photomicrograph of a specimen from a small dioritic body along the North Fork of the Clearwater River (loc. 215, pl. 1). Note the abundant small inclusions of biotite (bi) in the large round plagioclase grains (pla). ho=hornblende. Crossed nicols.

stresses that caused the orientation of biotite were still operating during the crystallization of hornblende. Because the feldspathization (tonalitization) tends to obscure the platy structure, the large plagioclase grains in the tonalite crystallized during a somewhat later phase. The lesser amount of quartz in the feldspathized schist than in an unaltered schist shows that quartz was replaced by plagioclase.

Diorite exposed for about 20 m between small satellitic bodies of gabbro and tonalite along the North Fork of the Clearwater River (loc. 215, in pl. 1) has a peculiar texture. It contains numerous large round plagioclase grains embedded among hornblende and biotite grains of medium size. The thin sections show that these round plagioclase grains enclose numerous small biotite flakes (fig. 24) and that some strained quartz occurs as interstitial mineral. A fine-grained biotite schist is exposed just north of this locality (No. 215). This biotite schist contains abundant quartz and small biotite flakes (about 20 percent) but no hornblende. In the diorite the small biotite inclusions in plagioclase and some strained quartz between the other minerals suggest that a similar fine-grained gneiss formerly occupied the place where diorite now occurs. The large rounded grains of plagioclase, the hornblende, and a part of the biotite crystallized late, replacing mainly quartz and including the small biotite flakes. The difference in the mineralogic composition of the diorite and the parent rock suggests an addition of iron, magnesium, calcium, and aluminum and removal of silicon during the replacement.

SEGREGATION OF BIOTITE IN THE SCHIST

The biotite-muscovite schist in a road cut about 0.2 mile west of the junction to Crawford mine along the road from Bald Mountain to Whiskey Butte is exceptionally rich in biotite. Four small bodies of medium-grained tonalite ranging from 2 to 5 m in thickness occur in the schist, which is tightly folded in this locality. Schist, several meters thick, next to the tonalite bodies contains about 60 percent biotite and 40 percent quartz. The most extensive layer rich in biotite measures about 8 m in thickness; some thinner layers are almost pure biotite. The westernmost contact zone contains abundant hornblende in addition to biotite.

Many of the inclusions of schist and of biotite-plagioclase gneiss in the quartz diorite, which is shown on the map in the eastern part of the Headquarters quadrangle, are enriched in biotite. The centers of the inclusions have generally preserved the original composition. The biotite has segregated into lenticles or layers along the borders or along certain beds.

The indices of refraction of biotite from such an inclusion near Alan Siding (loc. 292, pl. 2) are $\beta=\gamma=1.642\pm0.002$, which are those of common biotite.

The segregation of biotite is apparently associated with the emplacement of tonalite and quartz diorite. These plutonic rocks contain less biotite than the schist that originally occupied their present places. The emplacement of the plutonic rock apparently was not solely a mechanical but partly a chemical process during which the original metasedimentary rock was completely digested. During the introduction of the material that formed the plutonic rock, the excess of iron, magnesium, and potassium migrated into the country rock, forming biotite. The amount of the biotite added to the schist near the tonalite bodies corresponds approximately to the amount that was removed from the place now occupied by tonalite. Thus, it seems that the elements for the biotite originated in the schist that was digested by the plutonic rock nearby.

SERICITIZATION OF THE SILLIMANITE

In many places a part of the sillimanite in the schist is sericitized. This alteration requires an addition of potassium. A small part of this potassium may originate in the schist itself, (the schist shows chloritization of biotite) but the main part was probably introduced from the layers nearby from which potassium was removed during the Fe-Mg-Ca-Al metasomatism.

STRUCTURE

The metasedimentary rocks are strongly folded and faulted. In addition to the main folding, the schist layers show a wrinkling of bedding planes around the lineation.

BEDDING

In most outcrops of the quartzite the bedding is distinct, but in the schist it is less clear. In the biotite quartzite the bedding is pronounced because of a variation in the amounts of biotite, quartz, and plagioclase in the alternating layers. Some of the diopside-plagioclase gneiss is recrystallized to the extent that all the original structures have been obliterated. The muscovite-sillimanite laminae in the laminated schist are parallel to the bedding and make this structure easy to identify. In many outcrops laminated schist is interbedded with the coarse-grained sillimanite-garnet schist, but in other outcrops the schist seems massive or shows only foliation. However, bedding is indicated by repetition of laminated layers and massive layers.

FOLIATION

In outcrops where the alternation of thin quartzitic and micaceous layers makes the bedding distinct, foliation parallels the bedding. In thin sections the biotite and muscovite flakes are subparallel to the bedding plane. In the thick schist layers, which usually show a coarse wrinkling of the bedding plane, some of the mica flakes are parallel to this wrinkled plane and some are parallel to the axial plane of the tiny folds (fig. 25). All micaceous minerals are recrystallized, and petrofabric diagrams would show good girdles around the axes of the tiny folds.

In some localities where abundant small folds occur, the micaceous minerals are oriented parallel to the axial planes of the folds (as at loc. 246 along the river and at some localities along Orofino Creek).

FOLDING

The schist and quartzite are strongly folded. Small folds were observed in the road cuts and along the river at the localities where fold axes are marked on the map. Large folds become apparent when the strike and dip of the bedding are measured and plotted on the map. There are two sets of folds: the axes of the large folds plunge mainly 10° to 50° E. or SE. and those of the small folds 20° to 50° NE. The angle between the direction of these two sets of fold axes ranges from 60° to 80°. Wrinkling of the bedding plane around the axes of small folds is common and appears as a pronounced lineation in the schist layers. The variation of the angle between these two sets of fold axes suggests that the two sets of folding are independent of each other and are due to two acts of deformation. The orientation of the micaceous minerals parallel to the folded and wrinkled bedding plane and locally parallel to the axial plane of the folds indicates that the recrystallization took place during deformation.

FIGURE 25.—Small folds in a sillimanite-biotite schist at Bear Butte. Some sillimanite and biotite are parallel to the axial plane of the folds but most of them are parallel to the bedding. Ordinary light.

LINEATION

At some localities all rock types show good lineation, but at neighboring localities no lineation can be detected in the field. In the micaceous layers the lineation appears to be a wrinkling of the bedding plane. The direction of the lineation parallels the axes of the small folds and the lineation is actually a second fold axis.

FAULTING

Only a few large faults are exposed, but several are indicated by a juxtaposition of the formations in the outcrops nearby and abrupt changes of the rock types parallel to the bedding. A sharp fault in a laminated schist is exposed along Cougar Creek south of

Casey Cabin. This fault strikes N. 25° E. and dips steeply to the west. The schist layers on each side of the fault are at an angle of 80°. The structure of the formations north of this locality indicates that this fault continues for more than 1 mile at least. Another fault parallel to this one is exposed along the North Fork of Silver Creek about 1 mile west of Casey Cabin. A discontinuity of the formations near Lightning Points suggests that a fault occurs there also.

Near Big Island northwest-southeast faulting has brought the quartzite—presumably of the lower part of the Wallace formation—into juxtaposition with banded biotite-plagioclase schist of the Wallace formation. Another northwest-southeast fault cuts off the diopside quartzite west of Bald Mountain. Small northwest-southeast faults were observed in several other localities.

In addition to these two dominant fault directions—northwest and north-northeast—there are several north-south faults and small faults oriented at random.

PLUTONIC ROCKS OF THE IDAHO BATHOLITH AND SATELLITIC INTRUSIONS

In the Headquarters quadrangle the plutonic rocks are quartz diorite and tonalite. The largest body of quartz diorite—it covers the eastern part of the quadrangle—is a part of the Idaho batholith. The satellitic small bodies consist either of a similar quartz diorite or of a gneissic more siliceous rock which is here called tonalite.

PETROGRAPHIC DESCRIPTION

QUARTZ DIORITE

The quartz diorite is a coarse-grained rock in which the major constituents—plagioclase, quartz, hornblende, and biotite—are easily identified by the naked eye. The dark minerals are in strong contrast to the white or grayish mixture of quartz and plagioclase.

In thin sections the plagioclase (An_{38}) appears subhedral to anhedral and shows complex twinnings (fig. 26). The quartz grains are smaller and occur between the plagioclase grains. Hornblende is strongly pleochroic with α (light yellowish green) $=1.662\pm0.001$, β (green) $=1.679\pm0.001$, and γ (dark bluish green) $=1.683\pm0.001$. These indices of refraction suggest that the hornblende in the quartz diorite is much richer in iron than the hornblende in the metasomatic hornblende-plagioclase rocks (see table 8, column 3). Biotite is dark brown and has $\beta=\gamma=1.650\pm0.001$. Large grains of hornblende and biotite are scattered, but the smaller ones are clustered. Anhedral magnetite usually are in these clusters of dark minerals. Apatite occurs in tiny rounded prisms. The chemical and mineralogical composition of quartz diorite specimen No. 292 is shown in table 10.

FIGURE 26.—Photomicrograph of typical quartz diorite in the border zone of the Idaho batholith. Quartz (qu) occurs as small grains between the larger twinned plagioclase crystals (pla). Hornblende (ho) includes ilmenite-magnetite (black). Alan Siding (loc. 292, pl. 2). Crossed nicols.

The satellitic bodies of quartz diorite are chemically and mineralogically similar to the largest body that is a part of the border zone of the batholith. The largest satellitic body is exposed around the mouth of Reeds Creek south of Big Island (pl. 3) and measures about 1½ miles in diameter. The quartz diorite in this body is megascopically similar to that of the batholith, but the structure in most bodies is more gneissic than in the batholith and the amount of biotite is generally larger. The chemical and mineralogical composition of a sample from a satellitic body just north of the Headquarters quadrangle is shown in table 10 as No. 322.

Thin sections show that the texture of the quartz diorite in many satellitic bodies differs from that of the quartz diorite in the batholith. Plagioclase in the satellitic bodies shows a strong tendency to form large round grains, and the other minerals—quartz, biotite, and hornblende—tend to fill the interstices (fig. 27). Thus their texture has features that are characteristic of the feldspathized schist in this area and near Dent (fig. 80b). The similarity of the mode of occurrence of the plagioclase in the feldspathized schist and in the small satellitic bodies of quartz diorite suggests

TABLE 10.—*Chemical composition, ionic percentage, norms, minerals, and Niggli values of plutonic rocks of the Headquarters quadrangle and vicinity*

[Ruth Holzinger, analyst, U.S. Geological Survey]

Locality and specimen No.	292	322	190	379	213
Rock type	Quartz diorite of Idaho batholith.	Quartz diorite of a satellite.	Tonalite	Tonalite	Tonalite.
Location	Alan Siding	North Fork of the Clearwater River, mouth of Elkberry Creek.	North Fork of the Clearwater River, 4 miles east of Dent.	Dent	North Fork of the Clearwater River, 8 miles north of Big Island.

Chemical composition and ionic percentage

Constituent: Conventional symbol	Constituent: Symbol rearranged for cation percent	292 Weight percent	292 Cation percent	322 Weight percent	322 Cation percent	190 Weight percent	190 Cation percent	379 Weight percent	379 Cation percent	213 Weight percent	213 Cation percent
SiO_2		59. 16	55. 40	60. 35	56. 48	65. 42	60. 78	65. 61	60. 86	75. 52	72. 21
Al_2O_3	$A1O_{3/2}$	18. 15	20. 02	18. 22	20. 10	18. 90	20. 69	18. 78	20. 52	12. 58	14. 17
Fe_2O_3	$FeO_{3/2}$	1. 57	1. 10	1. 27	. 89	. 33	: 24	. 29	. 20	. 45	. 32
FeO		4. 08	3. 19	3. 62	2. 83	2. 28	1. 77	1. 99	1. 55	2. 12	1. 69
MnO		. 10	. 08	. 08	. 06	. 03	. 02	. 03	. 02	. 02	. 02
MgO		3. 13	4. 36	2. 84	3. 96	. 93	1. 29	. 97	1. 34	1. 28	1. 83
CaO		6. 47	6. 49	5. 95	5. 97	4. 87	4. 84	4. 75	4. 71	2. 68	2. 75
Na_2O	$NaO_{1/2}$	3. 71	6. 73	4. 10	7. 44	4. 62	8. 31	4. 79	8. 61	2. 77	5. 14
K_2O	$KO_{1/2}$	1. 47	1. 75	1. 28	1. 53	1. 34	1. 59	1. 51	1. 78	1. 26	1. 54
TiO_2		. 85	. 60	. 73	. 51	. 49	. 34	. 40	. 28	. 43	. 31
P_2O_5	$PO_{5/2}$	. 30	. 24	. 27	. 21	. 15	. 12	. 16	. 12	. 03	. 02
CO_2		. 03	. 04	. 02	. 02	. 01	. 01	. 01	. 01	. 00	. 00
H_2O^+		. 79	(2. 46)	1. 09	(3. 40)	. 44	(1. 36)	. 59	(1. 83)	. 63	(2 01)
H_2O^-		. 05		. 08		. 02		. 06		. 12	
Total		99. 86	100. 00	99. 90	100. 00	99. 83	100. 00	99. 94	100. 00	99. 89	100. 00
	O		160. 27		159. 93		165. 47		164. 67		174. 44
	OH		4. 92		6. 80		2. 72		3. 66		4. 02
	Total anions		165. 19		166. 73		168. 19		168. 33		178. 46

Norm

Mineral	292 CIPW	292 Molecular	322 CIPW	322 Molecular	190 CIPW	190 Molecular	379 CIPW	379 Molecular	213 CIPW	213 Molecular
Q	12. 38	11. 66	13. 38	12. 52	20. 62	19. 20	19. 53	18. 16	45. 71	43. 66
Or	8. 68	8. 75	7. 57	7. 65	7. 90	7. 95	8. 90	8. 90	7. 46	7. 70
Ab	31. 35	33. 65	34. 66	37. 20	39. 06	41. 55	40. 53	43. 05	23. 44	25. 70
An	28. 53	28. 85	27. 53	27. 83	23. 17	23. 15	22. 53	22. 50	13. 13	13. 60
C					1. 34	1. 53	1. 00	1. 13	1. 84	2. 05
Wo	. 67	. 56	. 12	. 06						
En	7. 79	8. 72	7. 07	7. 92	2. 32	2. 58	2. 42	2. 68	3. 19	3. 66
Fs	4. 99	4. 24	4. 54	3. 86	3. 15	2. 66	2. 81	2. 38	2. 85	2. 48
Ap	. 70	. 64	. 64	. 56	. 37	. 32	. 37	. 32	. 07	. 05
Il	1. 61	1. 20	1. 39	1. 02	. 93	. 68	. 76	. 56	. 82	. 62
Mt	2. 27	1. 65	1. 85	1. 34	. 49	. 36	. 42	. 30	. 65	. 48
Cc	. 07	0. 8	. 04	. 04	. 02	. 02	. 02	. 02		
H_2O	. 84		1. 17		. 46		. 64		. 75	
Total	99. 88	100. 00	99. 96	100. 00	99. 83	100. 00	99. 93	100. 00	99. 91	100. 00

Mineral content calculated on basis of chemical analysis

	292		322		190		379		213	
Quartz	17. 0		17. 4		22. 9		22. 3		48. 2	
Plagioclase	54. 2		55. 3		61. 0		58. 7		36. 7	
(anorthite content)	(38)		(36)		(37)		(32)		(36)	
Orthoclase					1. 3		2. 2			
Muscovite					3. 3		3. 3		2. 9	
Biotite	14. 7		14. 7		9. 6		10. 1		11. 7	
Chlorite					. 3		. 2			
Hornblende	11. 9		9. 0							
Epidote			2. 2		. 7		2. 6			
Apatite	. 7		. 6		. 3		. 4		. 1	
Sphene	. 6						. 2			
Ilmenite					. 4					
Magnetite	. 6		. 6						. 1	
Calcite	. 1		. 1							
Total	99. 8		99. 9		99. 8		100. 0		99. 7	

Niggli values

	292		322		190		379		213	
si		181. 50		199. 20		258. 50		260. 70		423. 0
ti		. 96		1. 81		1. 45		1. 20		1. 82
qz		23. 90		35. 64		74. 02		71. 34		244. 56
al		33. 88		35. 52		44. 03		44. 09		41. 60
fm		29. 69		27. 44		14. 15		13. 34		22. 68
c		22. 03		21. 15		20. 70		20. 23		16. 11
alk		14. 40		15. 89		21. 12		22. 34		19. 61
k		. 21		. 17		. 16		. 17		. 13
mg		. 50		. 51		. 39		. 43		. 47
c/fm		. 74		. 77		1. 46		1. 53		. 71

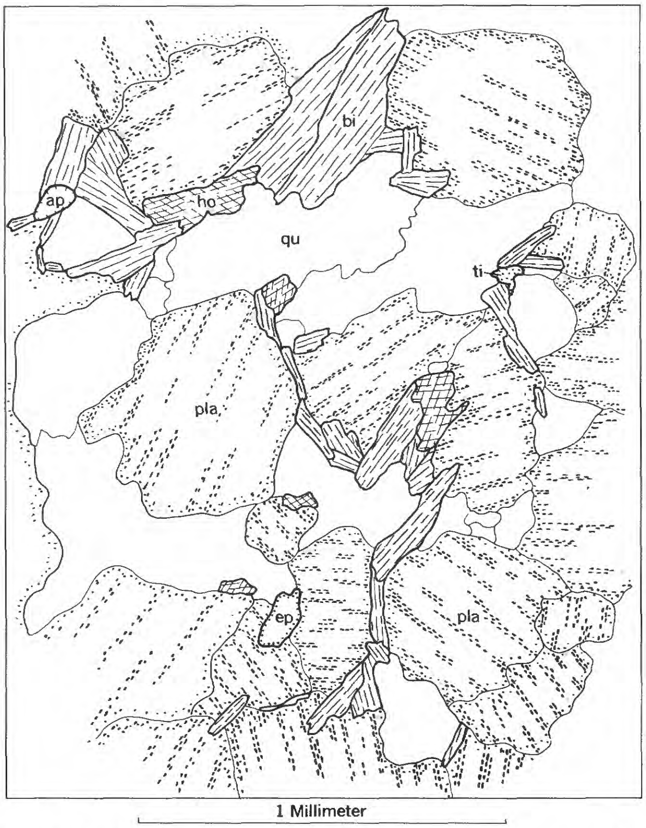

FIGURE 27.—Camera lucida drawing of quartz diorite in a satellitic body (No. 322) along the North Fork of the Clearwater River (pl. 1); pla=plagioclase; qu=quartz; bi=biotite; ho=hornblende; ap=apatite; ep=epidote; ti=sphene.

that these small bodies may have been formed in part by replacement and that some minerals (mainly biotite) in the country rock—the schist—were recrystallized, remaining between the newly formed plagioclase grains.

TONALITE

Small bodies of tonalite occur in the schist; most range from 5 to 300 m in thickness. The texture and mineralogy vary in individual bodies. Some are fine-grained bluish-gray rocks that contain only biotite as a dark constituent; some are very light in color, medium grained, and contain abundant quartz. Others are gneissic; these resemble the satellitic bodies of quartz diorite but their color is lighter.

FINE-GRAINED GRAY TONALITE

A small body of fine-grained medium-gray tonalite is exposed along the North Fork of the Clearwater River 2¼ miles north of Big Island (loc. 279, pl. 3). Some dikelike bodies of fine-grained tonalite occur along Silver Creek, south of Casey Cabin.

Plagioclase (An_{37}) is the main constituent of the tonalite at locality 279. It occurs as subhedral and polygonal grains of medium size and is twinned according to albite and Carlsbad laws. It constitutes about 74 percent of the rock. The other minerals are quartz, biotite, muscovite, apatite, epidote, magnetite, and garnet. The amount of quartz is about 15 percent and that of biotite about 10 percent. Quartz occurs in small round grains between the plagioclase grains and also as inclusions in them. Some small muscovite flakes are with the biotite. Both micaceous minerals show a random orientation. Apatite occurs as tiny prisms and as small round grains.

FIGURE 28.—Photomicrograph of intrusive tonalite from the riverside between Big Island and Dent (loc. 190, pl. 3). Plagioclase (pla) grains are subhedral, strongly twinned, and zoned. Quartz (qu) occurs in small grains between the plagioclase grains. Small flakes of biotite (bi) are the only dark constituent. Crossed nicols.

A chemical analysis was made of a similar fine-grained gray tonalite (fig. 28) from the riverside between Dent and Big Island (loc. 190 in pl. 3). The analysis (No. 190, table 10) shows that the amount of potassium in the tonalite is much lower that that of sodium. Mineralogically this is reflected in the small amount of potassium feldspar and the large amount of plagioclase. The Niggli value *k* is lower and *c* higher than the same values for the granite.

MEDIUM-GRAINED LIGHT-COLORED TONALITE

In locality 279 the fine-grained tonalite is cut by a medium-grained tonalite that is similar in composition

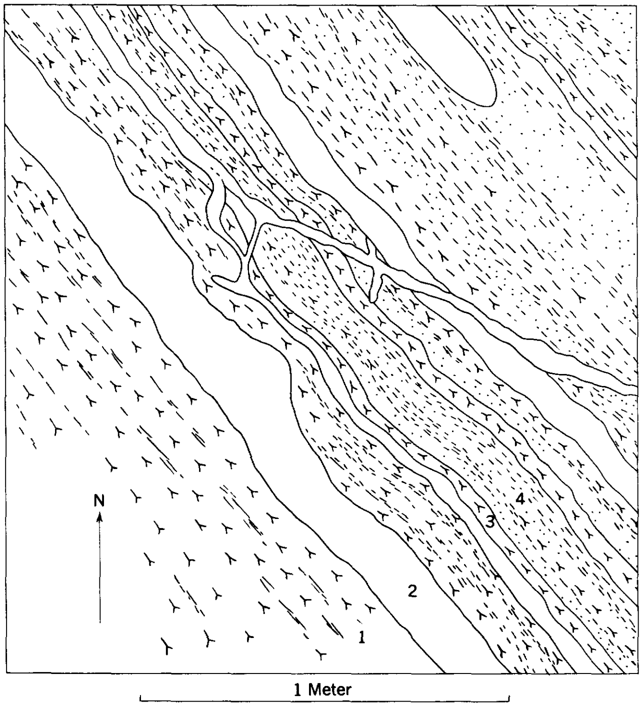

FIGURE 29.—Tonalitic (3) and pegmatitic (2) veins in banded biotite-plagioclase schist (4). Horizontal surface along the North Fork of the Clearwater River (loc. 234, pl. 2); 1 = biotite schlieren in gneissic tonalite.

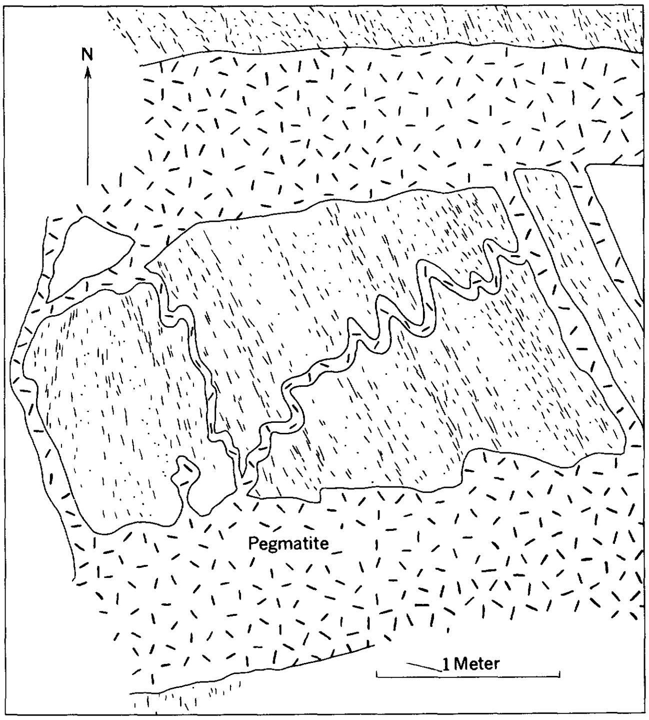

FIGURE 30.—Pegmatitic veins cutting the biotite-plagioclase gneiss discordantly in locality 264 along the North Fork of the Clearwater River (pl. 1). Horizontal surface. The bedding in the gneiss is vertical.

except that it contains more quartz and less biotite than the fine-grained variety. A few allanite grains were observed in a thin section of the medium-grained tonalite just south of locality 279.

A larger body of light-colored medium-grained tonalite occurs near Dent, and a sample of this body was analyzed chemically. The analysis (No. 379, table 10) shows that this tonalite is chemically and mineralogically very similar to the fine-grained variety.

GNEISSIC TONALITE

Many of the gneissic border zones of the satellitic bodies of quartz diorite contain a smaller amount of dark minerals and more quartz than the main part of the body. Plagioclase is oligoclase and in many places biotite is the only dark constituent. Thus these border zones are tonalitic in composition. In addition, there are several small bodies of similar gneissic tonalite in the schist, and the border zone of the main body of quartz diorite near Hollywood consists of gneissic tonalite.

The gneissic tonalite is heterogeneous due to an irregular distribution of biotite. It grades into feldspathized schist and contains inclusions of schist and quartzite. In many places plagioclase occurs as round grains that are larger than the other minerals in the rock and resemble the plagioclase in the feldspathized schist. The chemical analysis of gneissic tonalite enclosed in biotite quartzite (No. 213, table 10) shows an exceptionally large amount of quartz.

PEGMATITES

Narrow veins of pegmatite are exceedingly common in the schists and quartzites. The thickness of these veins ranges from a few millimeters to several meters. The grain size varies, the narrow veins are medium grained and the larger ones coarse grained. Most veins parallel the bedding, but many cut the rocks discordantly. Figure 29 shows pegmatitic veins in a banded biotite-plagioclase schist at locality 234 along the North Fork of the Clearwater River near the contact of a pegmatitic tonalite. The pegmatitic tonalite and tonalitic veins in the schist consist of quartz and plagioclase, whereas the granite pegmatite veins contain potassium feldspar also. The tonalitic veins are cut by the granite pegmatite and are therefore older. The larger veins—granitic pegmatite in composition—at locality 234 are parallel or subparallel to the structure of the older rock. At some other localities the larger veins cut the schist and biotite quartzite layers discordantly, and the smaller veins either parallel the layers or form meandering veinlets (fig. 30). Microscopic study of sample 264 from the large veins in figure 30 shows that the amount of plagioclase (An_{22}) in this pegmatite is much larger than that of potassium feldspar (microcline). The rock contains about 35 percent quartz in large strained grains; biotite is the main additional constituent. Only a few small flakes of muscovite occur with the biotite. Zircon and epidote are the accessories. Most pegmatites in the Headquarters quadrangle are similar to this plagioclase-biotite pegmatite. The amount of potassium feldspar and biotite in them varies locally, the smaller veins generally containing less biotite.

CHEMICAL COMPOSITION

The most striking feature in the chemical composition of the plutonic rocks is the scarcity of potassium, even in the most silicic members. This is well shown by a low Niggli value, $k=0.17$, which is a value common for the trondhjemitic rocks. The tonalites differ from the rocks of the trondhjemitic suite, however, in other respects. They are richer in calcium and magnesium, and the plagioclase is more anorthitic than in a typical trondhjemite. The tonalite is distinguished from quartz diorite by a higher percentage of quartz and much lower percentage or lack of hornblende. Scarcity of potassium is characteristic also of the pegmatites of this area, where they are plagioclase-biotite rocks with no or only a little potassium feldspar and very little muscovite. In this respect they are in striking contrast to the pegmatites in the main part of the Idaho batholith and in the areas north and south of the Headquarters quadrangle where pegmatite veins carry abundant potassium feldspar and muscovite in minable quantities (Stoll, 1950).

Thus the chemical composition of all the plutonic rocks in the area investigated shows the rocks to have the same characteristics; all belong to the tonalitic suite. True granites are lacking and there is no quartz monzonite exposed. The lack or scarcity of potassium feldspar along the border zone of a granitic or quartz monzonitic batholith and in small intrusive bodies about it is common in many other areas. According to Waters (1938), the border facies of the Chelan batholith in Washington State is chemically much lower in potassium than the central part. The chemical analyses given by him (p. 769) show that this border facies is very similar to the quartz dioritic satellites in the Headquarters quadrangle. According to Waters, the center of the Chelan batholith is somewhat younger than the border zone. In Idaho the rocks of the quartz monzonite series cut the quartz dioritic border facies discordantly in many localities east of the Headquarters quadrangle; they were emplaced while the major part of the border zone was solid. Thus the rocks of the tonalite series represent an earlier phase of intrusion, whereas the younger intrusive rocks are granitic or quartz monzonitic and contain abundant potassium feldspar.

STRUCTURE OF THE PLUTONIC ROCKS

CONTACT RELATIONS

On a large scale the contact between the quartz dioritic border zone of the Idaho batholith and the metasedimentary rocks follows the folded bedding of the latter. The same relation is true of the contacts between the satellitic intrusive bodies and the metasedimentary rocks. However, in individual outcrops crosscutting relations can be observed. The contact is parallel to the bedding for long distances, but locally it cuts abruptly across the bedding and forms a zigzag line; apophyses extend into the country rock parallel to the bedding (fig. 31). Remnants of the country rock that seems to have preserved their orientation are common near the contacts and mark the continuation of the structure of the country rock in the quartz diorite. The grain size of the intrusive rock is just as coarse

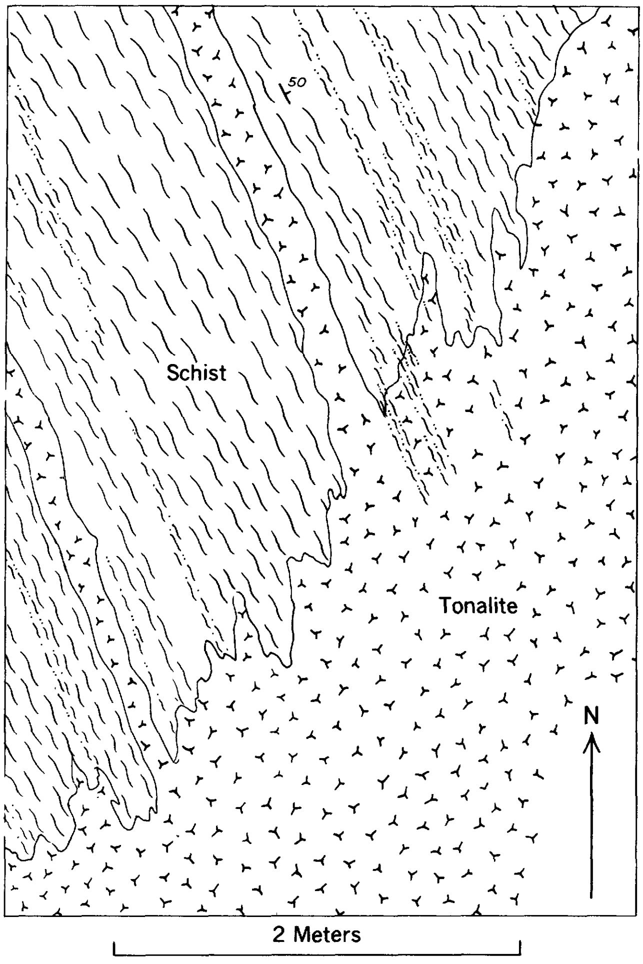

FIGURE 31.—Contact of tonalitic satellite and schist at the mouth of Silver Creek (just north of loc. 228, pl. 2). Horizontal surface. The planar structure in the schist dips 50° NE.

near the contact as it is in the main part of the intrusive body.

A wavy contact parallel to the bedding is extremely common; in such bodies metasomatically altered rocks occur in the contact zone. This type of contact with a diopside-plagioclase gneiss is exposed 1 mile west of Casey Meadows, in the northwest corner of T. 38 N., R. 5 E. (pl. 2). The diopside-plagioclase gneiss dips at an angle of 25° to 50° under the quartz diorite. Pegmatitic rock occurs along the contact; it includes abundant remnants of diopside-plagioclase gneiss and many individual large hornblende crystals. Hornblende has been developed also in the diopside-plagioclase gneiss as a result of hydrous solutions infiltrating the inclusions and the country rock.

Abundant pegmatite occurs in the contact zone exposed along the southern border of the Headquarters quadrangle. There hornblende has replaced a part of the diopside and quartz in the diopside-plagioclase gneiss and quartzite, and it abounds also in biotite gneiss layers. Small masses of plagioclase, rimmed by hornblende, are common in the diopside-plagioclase gneiss and quartzite. Crystallization of grossularite, zoisite, and clinozoisite in calcite-bearing layers near the contact is found at only a few localities in this area, and it is not clear whether the crystallization of these silicates is due to a regional metamorphism or to a later contact metamorphism.

At some localities the schist near the contact is enriched in biotite, and at other localities it contains abundant pegmatitic veins. The thin-bedded biotite quartzite layers contain more pegmatitic veinlets than do the other rocks, and in some places regular migmatitic veining has been developed parallel to the bedding. The apophyses of quartz diorite generally are parallel to the bedding, but at some localities discordant relations were observed. The contacts of these apophyses across the bedding are mostly zigsag lines similar to those of the large bodies (fig. 31). Narrow pegmatitic veinlets occur in the country rock near the contact of the apophyses.

That type of contact migmatite which would suggest the rising of the batholith in a manner similar to salt domes—diapir structures by Wegmann (1930)—is completely lacking in the Headquarters quadrangle. The structure of the country rock was not disturbed dynamically, nor is there any evidence of protoclastic structures. The migmatitization is local and is due to the infiltration and segregation of pegmatitic material parallel to the bedding. All contacts studied in detail are replacement contacts over a short distance. The remnants of the country rocks in the quartz diorite seem to have preserved their original position. The contact line is irregular, and abundant apophyses truncate the wallrock. The wallrocks are locally enriched in biotite, hornblende, and plagioclase; that is, in the constituents of the quartz diorite. Grain size is of the same magnitude in the intrusive rock and in the wallrock. The structure of the wallrocks shows that they were recrystallized to the amphibolite facies simultaneously with the folding. The discordant contact relations indicate that the quartz diorite was emplaced after the folding. The metasomatic effects—the development of hornblende and plagioclase in the wallrocks—are also post-folding and most likely connected with the intrusion of the quartz diorite.

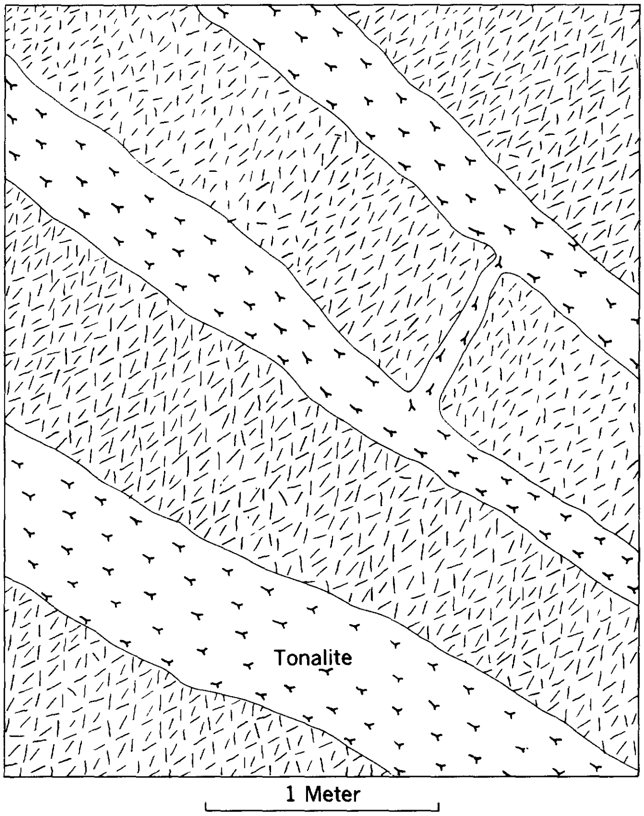

FIGURE 32.—Tonalite dikes in quartz diorite along Orofino Creek. Vertical railroad cut faces south. The planar structure in the dikes is parallel to the walls.

Fine-grained and medium-grained tonalite cut the metasedimentary rocks discordantly and occur as dikes in the quartz diorite (fig. 32), thus being somewhat younger than the quartz diorite.

PLANAR AND LINEAR STRUCTURES

Planar and linear structures become visible because of the parallel orientation of biotite and hornblende. Planar structure, parallel or subparallel to the bedding of the metasedimentary rocks, is well developed along those contacts that parallel the bedding of these country rocks. Most outcrops along the discordant contacts and several of those farther from the contacts are massive or show only linear structure. When the symbols for the planar structure are ploted on a map, they seem to parallel the contact of the batholith with a few exceptions. Where the exceptions occur, the planar structure seems to be folded and inclusions of country rock parallel to the planar structure are common.

Near the walls the linear structure is usually parallel to the axis of the folding or parallel to the lineation of the wallrock, and even within the mass the linear structure still seems to parallel these two major structural directions. This parallelism, together with the orientation of the planar structure parallel to the folded country rock, may indicate that the level now exposed is close to the roof or that the border zone of the batholith is only a thin sheet which followed a shallow syncline in the northern part and a gentle folding of the country rocks in the southern part of the quadrangle.

INCLUSIONS

Several large inclusions of country rocks are exposed and shown on the map (pl. 2) within the largest body of quartz diorite. These inclusions are metasomatically altered in a manner similar to that of the wallrock. The inclusions of schist and biotite quartzite are locally enriched in biotite, and the inclusions of diopside-plagioclase gneiss contain abundant secondary hornblende. At many localities, the grain size of the metasomatically altered rock is much coarser than the grain size of a normal country rock. On Dull Axe Mountain the grain size of diopside, plagioclase, and grossularite in the lime-silicate rock is much larger than the grain size of these minerals elsewhere. The orientation of bedding in many small inclusions near the contact suggests that these inclusions have preserved their orientation and position. Thus the movement of the magma near the contact must have been negligible. An inclusion of coarse-grained schistose quartzite with flakes of scattered biotite and grains of plagioclase is exposed in a road cut half a mile north of Headquarters. The lineation in this quartzite inclusion is parallel to the major fold axes of the country rocks. A large inclusion of thin-bedded biotite quartzite with ptygmatic veinlets is exposed in a road cut 0.7 mile north of Alan Siding, and another is exposed about 1 mile southeast of Silver Butte; the latter inclusion is in alignment with layers of thin-bedded biotite quartzite in the country rocks west of this locality.

In contrast to the seemingly undisturbed position of inclusions in the quartz diorite, the small inclusions in the tonalite bodies between Dent and Big Island have been rotated from their original position; the bedding and foliation of the inclusions there show a random orientation.

ORIGIN OF THE PLUTONIC ROCKS

The contact relations, structure, and texture of the plutonic rocks suggest that some are typical igneous rocks and that others are a result of development of hornblende and plagioclase in schist and quartzite. The largest body of quartz diorite and some of the tonalite bodies nearby are representatives of the igneous variety; many of the satellitic bodies of gneissic quartz diorite and tonalite are of metasomatic origin. The evidence shows that rock of quartz dioritic and tonalitic composition can be derived either from the schist or from the quartzite by addition of calcium, iron, magnesium, and aluminum and by subsequent removal of silicon. The chemistry and mechanism of the transfer of these elements are discussed in the concluding sections.

SUMMARY OF THE GEOLOGY OF THE HEADQUARTERS QUADRANGLE AND THE BIG ISLAND AREA

One of the most striking features of the rocks of the Headquarters quadrangle is the scarcity of potassium feldspar. It is completely lacking in the border zone of the Idaho batholith and very scarce in the satellitic bodies. It is scarce in the metamorphosed rocks of the Belt series, even though the equivalents of these same rocks in the areas farther north contain a considerable amount of potassium feldspar.

In the Belt series within the Headquarters quadrangle the amount of quartz also is less than in areas farther north. For example, in the Coeur d'Alene district the Wallace formation has been described as consisting mainly of quartzite with some calcareous and shaly layers. In the Headquarters quadrangle, schist is the main rock type and the amount of quartzite is subordinate. Only a few of these quartzite layers contain as much as 80 percent quartz; most contain only 50 to 70 percent.

The highly metamorphosed zone around the batholith extends more than 20 miles north and northwest, from the northern border of the Headquarters quadrangle. Beyond that, the grade of metamorphism is considerably less. Staurolite, garnet, and kyanite schists have been reported from the St. Joe-Clearwater divide (Calkins and Jones, 1911, p. 80) about 25 miles north of the Headquarters quadrangle, but about 10 miles north of that locality, near the St. Joe River and toward the north, the equivalents of schists and gneisses are shales and sandstones (Umpleby and Jones, 1923).

Conclusions regarding the folding, metamorphism, and intrusion can be summarized as follows:

1. The country rock of the Idaho batholith was metamorphosed to the sillimanite-muscovite subfacies of the amphibolite facies contemporaneously with the folding.
2. Emplacement of quartz diorite is postfolding and later than the major phase of metamorphism.
3. Metasomatic development of hornblende and plagioclase is postfolding and genetically connected with the emplacement of the quartz diorite.

This relation between intrusion and metamorphism shows that the place of the Idaho batholith was a heat center also before the intrusion of quartz diorite took place. The following explanations can be offered:

1. It is possible that the quartz diorite now exposed has completely digested the earlier intrusive rocks that caused the metamorphism.
2. Metamorphism was caused by orogenic movements, and the quartz diorite is a result of a high-temperature metamorphism and partial melting of the metasedimentary rocks.

The position of the Idaho batholith in the junction of two arcuate segments of Nevada folding (Eardley, 1951, p. 310) favors both of these possibilities. Openings are easily formed during the folding in such junctions, and these openings may be filled by molten magma from below. On the other hand, an extreme metamorphism is common in such junctions, and this metamorphism may give rise to the segregation of granitic material. Thus the batholith itself could be a result of the strong deformation accompanied by ultrametamorphism and formation of magma. This problem is discussed further in the concluding sections.

PART 3. THE OROFINO DISTRICT

The Orofino district offers excellent material for the study of the transformation of metasedimentary rocks to plutonic rocks. On the basis of the mapping done during this study, about 50 percent or less of the pre-Miocene rocks in the Orofino district are classed as remnants of the bedded metasedimentary unit that was called the Orofino series by Anderson (1930). The remaining rocks are of igneous aspect and in this report are called plutonic rocks. The transition from the metasedimentary rocks to plutonic rocks can be seen in many localities in the field and studied in the laboratory. The transition zones, which are irregular in width, may either parallel the bedding or cut across it. Much of the metasedimentary rock chemically and mineralogically resembles plutonic rocks because of partial dioritization or tonalitization. These modified metasedimentary rocks form the major part of the Orofino series and are discussed later in this paper under the heading "Metasomatic changes in the metasedimentary rocks near Orofino."

METASEDIMENTARY ROCKS

The metasedimentary rocks—the oldest rocks in the district—are exposed along the canyons of the Clearwater River and its tributaries. The best and most continuous exposures are along Orofino Creek and along the North Fork of the Clearwater River between Bruces Eddy and the mouth of Canyon Creek (pl. 4). Elsewhere in the area, plutonic rocks and metasomatically altered rocks interrupt the continuity and obscure the stratigraphic sequence. The metasedimentary rocks consist of (1) marble, (2) quartzite, diopside gneiss, and calcium magnesium-aluminum silicate rock (lime-silicate rock), (3) biotite schist and gneiss with thin hornblende-bearing layers or diopside- and scapolite-bearing layers, (4) hornblende gneiss, hornblende-biotite gneiss, and biotite gneiss, (5) amphibolite, (6) kyanite-garnet amphibolite, and (7) biotite-garnet schist.

STRATIGRAPHY

Only fragmentary information has been obtained about the stratigraphy of the metasedimentary series. The series contains several marble layers, most of them discontinuous (pl. 4), and many quartzite layers near or next to the marble layers. The main rocks between these horizons of marble and quartzite are layered hornblende-biotite-plagioclase and biotite-plagioclase gneisses. Amphibolite is either interbedded with gneisses or adjacent to the marble and quartzite. Biotite-garnet schist is exposed only along Orofino Creek about 1 mile east of the mouth of Whiskey Creek.

SECTION ALONG OROFINO CREEK

Because of the discontinuity of the outcrops, folding, faulting, and occurrence of plutonic rocks, it was impossible to determine the complete stratigraphic sequence of the layers or the thickness of the various members of the formation. The best available section (table 11) is exposed along the railroad following Orofino Creek east of the mouth of Whiskey Creek, and although discontinuous, faulted, and deformed by folding, it gives a clear picture of the variety of material involved. The main rock type in this section is thin-bedded and fine-grained biotite-hornblende gneiss that contains either resistant layers of diopside-garnet-scapolite rock 1 to 4 cm thick or layers of biotite schist 3 to 20 cm thick. The thickness of the biotite-hornblende gneiss layers ranges from 4 to 30 cm. In

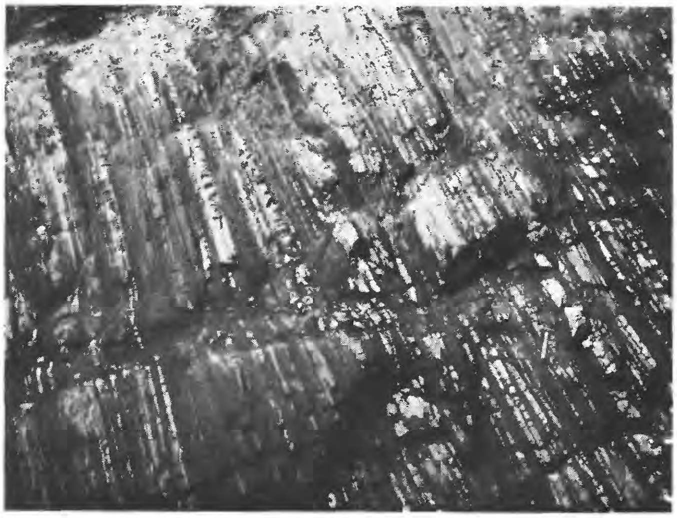

FIGURE 33.—Thin-bedded biotite schist (dark layers) with harder layers rich in quartz, diopside, scapolite, and garnet (light layers). This photograph was taken immediately after a rain. The soft, partly weathered schist layers were still wet and were therefore much darker than the quartzitic layers. The bedding is scarcely noticeable in dry weather. Railroad cut half a mile east of the mouth of Whiskey Creek. Scale is shown by a hammer.

places diopside-garnet-scapolite layers are interbedded with thin-bedded biotite schist (fig. 33). Three marble layers ranging from 10 to 30 m in thickness occur at regular intervals in this section.

It is not clear whether part of the repetition of marble layers is due to folding and faulting. The beds strike almost due north in this section and dip 45° to 65° E. Small folds are overturned to the west. Metamorphism and metasomatism have obliterated the sedimentary relict structures of the individual beds to the extent that it is impossible to tell whether a part of the section is overturned. Much of the repetition of the layers in the Orofino series elsewhere also is due to the isoclinal folding and faulting.

Another fairly continuous section along Orofino Creek is exposed near Lime Mountain (pl. 4). Two thicker marble beds—one of them dolomitic—occur in the center of this section and are separated by thin-bedded light bluish gray biotite-bearing quartzite. The upper marble bed is overlain by a layer of sillimanite-biotite gneiss a few meters thick, and above this gneiss is thin-bedded biotite gneiss that contains schistose laminae and thin hornblende-bearing layers. Thin-bedded fine-grained biotite gneiss and biotite quartzite also are dominant beneath the marble layer. Two thick garnet amphibolite layers occur in this part of the formation. Farther to the east fairly homogeneous garnet amphibolite is exposed for more than 1 mile; it is overlain by thin-bedded biotite schist with hornblende-bearing layers and quartzite layers.

Boulders of coarse-grained bluish-white pure quartzite with a glassy appearance occur about a quarter of a mile east of the mouth of Cedar Creek. The stratigraphic relation of this quartzite to the thin-bedded and laminated biotite gneiss east of it is not known because they are separated by tonalite. A similar coarse-grained white quartzite with scattered large biotite flakes is exposed in the two small areas of metamorphic rocks that rise through the basalt cover near Huckleberry Butte. There the coarse-grained quartzite is underlain by biotite-garnet-sillimanite schist. Only plutonic rocks are exposed under the coarse-grained quartzite along Orofino Creek, and it is impossible to tell whether these two quartzite occurrences belong to the same layer.

TABLE 11.—*Section along Orofino Creek east of the mouth of Whiskey Creek*

Lithology	Thickness	
	Feet	Meters
Hornblende-biotite gneiss with biotite schist layers and marble layers	260	79
Marble	100	30
Not exposed	300	91
Thin-bedded biotite schist with diopside-, hornblende-, and scapolite-bearing layers	100	30
Amphibolite	160	49
Thin-bedded biotite schist with diopside-, hornblende-, and scapolite-bearing layers	240	73
Marble	60	18
Not exposed, pegmatite boulders	130	40
Biotite gneiss with some hornblende- and garnet-bearing layers	35	11
Not exposed	60	18
Pegmatite	20	6
Amphibolite	40	12
Biotite-garnet schist with lenticles rich in hornblende	70	21
Quartzite	3	1
Fault	----	------
Hornblende-biotite gneiss with garnet	100	30
Pegmatite	6	2
Not exposed	20	6
Basalt boulders (float from basalt exposed on higher part of the canyon wall)	25	8
Thin-bedded biotite-garnet schist with diopside-, hornblende-, and scapolite-bearing layers	35	11
Basalt boulders	15	5
Biotite-garnet schist	15	5
Hornblende-biotite gneiss with biotite schist layers	20	6
Amphibolite	40	12
Hornblende-biotite gneiss with some garnet	110	34
Not exposed	90	27
Marble	50	15
Thin-bedded biotite-hornblende gneiss with diopside- and scapolite-bearing layers	160	49
Not exposed	100	30
Hornblende-biotite gneiss with biotite schist layers	80	24
Faults	----	------

SECTION ALONG THE NORTH FORK OF THE CLEARWATER RIVER

A fairly continuous section is exposed between Bruces Eddy and the mouth of Canyon Creek. Two layers of marble, one about 8 m thick and the other 80 m thick, occur toward the bottom in this section. These layers

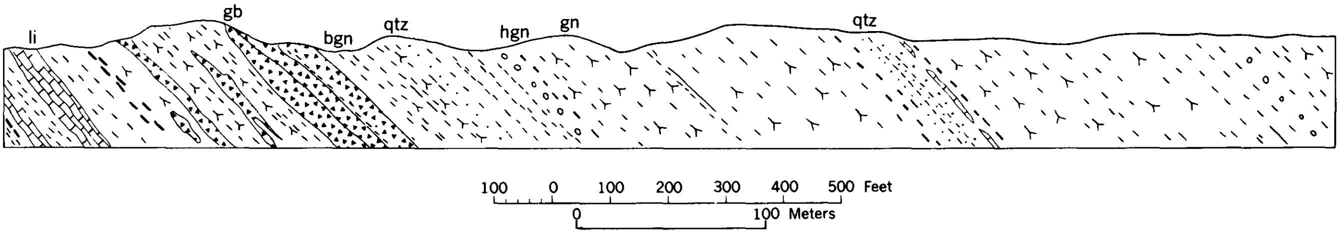

FIGURE 34.—Exposure in a road cut which faces east near Bruces Eddy. li=marble (at bottom of section); qtz=quartzite; bgn=biotite gneiss; hgn=hornblende gneiss; gn=tonalitic gneiss; gb=gabbroic rock. Vertical scale is the same as the horizontal one.

seem to continue eastward as layers interbedded with thin-bedded quartzite about a mile north of Orofino. To the west the position of the marble at the higher altitude on the western canyon wall is occupied by amphibolite and tonalitic rock. At Bruces Eddy, quartzite underlies the marble. The section above the marble, as shown in figure 34, consists of thick-bedded medium- to fine-grained biotite-gneiss alternating with a thin-bedded biotite gneiss that contains thin quartzite layers, amphibole-bearing layers, and micaceous layers. A few diopside- and scapolite-bearing layers, about 50 cm thick, are interbedded with the thin-bedded quartzite. A thin-bedded diopside amphibolite is exposed in the center of this section (pl. 4) and above it is a biotite schist and thick-bedded biotite-plagioclase gneiss with hornblende-bearing layers.

On the northern canyon wall of the North Fork of the Clearwater River south of Freeman Creek, a coarse-grained quartzite with a glassy and milky appearance is interbedded with diopside gneiss, quartzite, and sillimanite-biotite schist. Some beds of the coarse-grained quartzite contain abundant long crystals of sillimanite. A fault separates these beds from the North Fork section to the south. Toward the north they are bordered by quartz diorite. This sequence of coarse-grained quartzite, diopside gneiss, quartzite, and schist is comparable to a similar rock series near Dent and may belong to the Wallace formation of the Belt series. The section opposite the mouth of Canyon Creek is stratigraphically above the coarse-grained quartzite, provided that the quartz diorite does not occupy a fault. This section consists of thin-bedded biotite quartzite with thin plagioclase-bearing layers. Two diopside gneiss layers with calcite lenticles, about 30 cm thick, are exposed in the lowest part next to the tonalite that forms a silicic border zone of the diorite.

SECTION NORTH OF AHSAHKA

The lower part of the section, exposed along an old logging road 1½ miles north of Ahsahka, consists of thin-bedded biotite quartzite with plagioclase gneiss layers. The upper part is thin-bedded biotite quartzite interbedded with thin-bedded diopside-plagioclase gneiss resembling part of the Wallace formation in an area about 15 miles north of Orofino. A fine- to medium-grained biotite gneiss with small pink garnet crystals is exposed on both sides of this quartzite-gneiss sequence.

SUMMARY OF THE STRATIGRAPHIC SEQUENCE

The main rock type in the sections that do not show signs of metasomatic transformation is thin-bedded biotite gneiss or biotite-hornblende gneiss; both may contain thin micaceous laminae, calcium-magnesium silicate-bearing layers, or layers of hornblende gneiss. Interbedded with the gneisses are: (1) sequences consisting of quartzite, diopside gneiss, calcium-magnesium-aluminum silicate rock, and marble; (2) thick units of biotite-plagioclase gneiss interbedded with biotite-hornblende-plagioclase gneiss; and (3) amphibolite. A sequence of thin-bedded quartzite, diopside gneiss, marble, and amphibolite is exposed northwest of Orofino. This sequence is overlain by thick beds of biotite-plagioclase gneiss and hornblende-plagioclase gneiss that are exposed along the South Fork of the Clearwater River at the mouth of Orofino Creek.

A similar sequence of thin-bedded quartzite, diopside gneiss, marble and amphibolite is exposed just east of the mouth of Orofino Creek. Strikes and dips of beds suggest that folding is isoclinal and that this is the same sequence as that exposed north of Orofino. The same sequence is repeated also 1 mile east of Orofino where it is overlain by a thick section of thin-bedded hornblende-biotite gneiss. The section east of the mouth of Whiskey Creek and that near Lime Mountain contain similar sequences and are probably structural repetitions.

STRUCTURE

BEDDING

All metasedimentary rocks of the Orofino district show a distinct bedding. The thickness of the beds varies; the biotite-plagioclase gneiss near Orofino and Ahsahka is fairly thick bedded; the thickness of individual beds may range from 10 to 50 cm or more. Most rocks in the section given in table 11 and those near the mouth of Canyon Creek along the North Fork of the Clearwater River are thin bedded; the thickness of individual beds ranges from 1 to 20 cm. The diopside amphibolite and quartzite southwest of the mouth of Canyon Creek (loc. 443 and 448, pl. 4) are also thin bedded.

FOLDS

Small folds can be seen in road and railroad cuts. Along the North Fork of the Clearwater River, fold axes strike N. 60° to 80° W. and plunge gently to the northwest. Folds are overturned to the south and southwest. North of the town of Orofino, bedded fine-grained quartzite and intercalated marble beds show a large open drag fold; in horizontal section the northern section has moved eastward in relation to the southern section. The north-south strike of the bedding along Orofino Creek is due to this drag. The axes of the folds exposed in this area also strike in a northerly direction; presumably the strata were curved from the east-west direction to a north-south position after the major folding. The strikes and dips measured along Orofino Creek indicate that the rocks there are isoclinally folded. The repetition of the marble-quartzite sequence is due mainly to this folding.

FOLIATION

Near Orofino the foliation is parallel to the bedding in all outcrops where bedding is visible. The foliation is due to alinement of biotite flakes and hornblende prisms and is emphasized by parallel white veinlets. On the east side of Big Canyon Creek near Peck, a transecting cleavage parallel to the axial plane of small folds was observed. The outcrops there consist of fine-grained dark amphibolite (metabasalt), which is intensely folded; the axial planes strike N. 60° W. and dip 20° NE.

LINEATION

Lineation is visible in most outcrops. It is either a wrinkling of the bedding plane or a parallel orientation of pencil-shaped minerals such as hornblende and biotite. The direction of the lineation varies. In a few outcrops it seems to parallel the fold axis, but in most localities it makes an angle of 60° to 85° with the fold axis in the *ab* plane, differing thus from the direction of the *a* axis of the folds.

PETROGRAPHY

The main types of metasedimentary rocks that do not show evidence of metasomatic addition of elements are described in detail in this section to give the reader a better understanding of the wide variety of sedimentary rocks that were present in the Orofino district before metamorphism.

MARBLE

Limestone is recrystallized to coarse-grained white, or more rarely, light bluish gray marble in which the individual grains range from 0.5 to 5 mm in diameter. Most beds are pure calcite with only a few small grains of pyrite or occasional tremolite, diopside, and phlogopite. One of the beds at Lime Mountain, however, is dolomitic.

Tremolite occurs in the marble in the quarry south of Orofino in groups of fine white needles between the calcite grains near the contact of biotite schist and pegmatite. The indices of refraction $\alpha=1.600\pm0.001$, $\beta=1.613\pm0.002$, $\gamma=1.625\pm0.001$, $Z\wedge c=20^\circ$ and $-2V$ about 70° indicate, according to Winchell (1951), that this tremolite is an almost pure magnesium end member of the tremolite-actinolite series.

Phlogopite in the same rock shows $\gamma=1.573\pm0.001$ and 2V about 3°. Phlogopite in the limestone at Bruces Eddy has about the same index of refraction: $\gamma=1.576\pm0.001$, but in marble east of the mouth of Whiskey Creek, phlogopite has a considerably lower index of refraction, $\gamma=1.530\pm0.001$, indicating a lower iron content in this mineral.

White calcite grains in marble northeast of Orofino show $\omega=1.659\pm0.001$, and the pinkish grains in limestone south of Orofino show $\omega=1.659\pm0.001$; both types of calcite are thus pure calcium carbonate.

Sphene, a common accessory mineral in the marble, occurs usually in small quantities as small yellow crystals, but in some localities, for example at Lime Mountain, crystals measuring as much as 1 cm in diameter abound. These crystals are pyramidal, with well-developed (001) and (111) faces. Some large crystals have two sets of twinning lamellae at an angle of 71°. One set of twinning lamellae is parallel to the cleavage.

A sandy gray limestone layer occurs on the south side of Orofino Creek opposite the mouth of Whiskey Creek. This limestone contains round and oval plagioclase grains which give it a pebbly appearance. The rounded plagioclase grains, which range from 0.5 to 10 mm in diameter, are embedded in calcite and include small grains of calcite. A part of one larger pebble is fine grained. Probably all rounded plagioclase grains were originally fine-grained pebbles consisting of a rock rich in plagioclase and were deposited with the calcareous material. The large individual plagioclase grains crystallized from these pebbles, preserved their original shape and enclosed the calcite contained in the pebbles. The upper part of this limestone contains about 30 percent plagioclase (An_{32}), a few diopside grains (which are surrounded by muscovite), and some hornblende, pyrite, magnetite, sphene, and graphite.

QUARTZITE, DIOPSIDE GNEISS, AND LIME-SILICATE ROCK

Although quartzite commonly occurs next to limestone, in some places these rocks are separated by a thin layer of amphibolite. The quartzites can be divided in terms of their constituent minerals into biotite quartzite and diopside quartzite. Both are thin bedded and medium grained. The amount of biotite and diopside varies from layer to layer and the color varies accordingly. The pure quartzite layers are light gray or bluish gray, and the biotite-rich layers are dark gray. At the mouth of Jim Ford Creek, hard fine-grained quartzite beds are interbedded with fine-grained biotite schist and many of these quartzite beds form boudins. The diopside quartzite is green, the color intensifying as the amount of diopside increases and the rock grades to diopside gneiss.

The diopside quartzite and diopside gneiss of the Orofino district do not have the granular texture that typifies the diopside-bearing gneiss and quartzites east and north of the Orofino region. Grain boundaries in the diopside gneiss and diopside quartzite of the Orofino series are irregular and most diopside is concentrated in thin layers (fig. 35). Most layers rich in diopside also contain abundant epidote minerals (usually epidote and zoisite, but also clinozoisite), plagioclase (An_{37-40}), garnet, and scapolite, and therefore resemble the lime-silicate rocks in composition. The amount of plagioclase, like that of the other constituents, varies. Some layers are almost pure quartz, some contain about 30 percent quartz, 30 percent plagioclase, and the rest diopside and epidote minerals. In several layers the amount of garnet is considerable; for example, just north of Bruces Eddy about 50 percent of the rock consists of light brownish red garnet crystals that include grains of quartz, diopside, and epidote minerals.

On a hill south of the two marble layers near Bruces Eddy (loc. 568, pl. 4), euhedral garnet crystals, 5 cm in diameter, are embedded in quartz. The garnet is light reddish brown and has an index of refraction of $n=1.761\pm0.001$. The index of refraction together with the unit-cell length[1] of $a_o = 11.885\pm0.006$Å indicate that it is mainly grossularite. Garnet also is abundant in a layer of lime-silicate rock 1 mile north of the marble at Bruces Eddy (loc. 430). Diopside in this layer shows $\alpha=1.680\pm0.001$, $\beta=1.685\pm0.001$, $\gamma=1.704\pm0.001$, and 2V about 60°. According to Hess' diagram (1949), this diopside contains about 20 percent hedenbergite. Zoisite in the same rock shows $\alpha=1.702\pm0.001$, $\beta=1.703\pm0.001$, $\gamma=1.706\pm0.001$, which according to Winchell (1951) indicates only a small amount of iron, if any. If it is assumed that the present mineralogical composition of this layer reflects its original composition, it can be concluded that this layer was a carbonate-bearing shale—the carbonate being calcite and dolomite with some siderite.

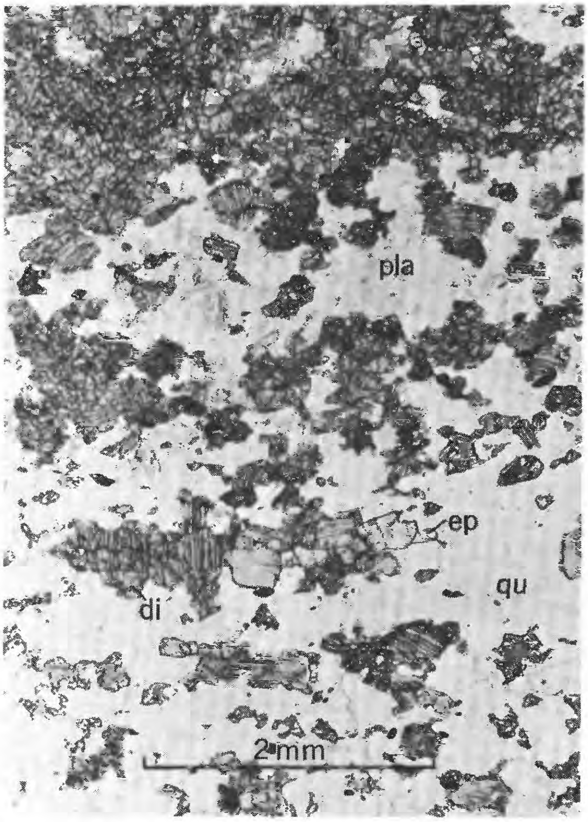

FIGURE 35.—Photomicrograph of diopside-plagioclase gneiss. Epidote (ep) occurs in layers rich in diopside (di). The light minerals are quartz (qu) and plagioclase (pla). Locality 109 north of Ahsahka (pl. 4). Plane polarized light.

The diopside and epidote minerals in all the other specimens studied seem to be poor in iron. The indices of refraction of the diopside are low and the zoisite is the dominant epidote mineral. Epidote with $\alpha=1.716\pm0.001$, $\beta=1.720\pm0.001$, $\gamma=1.725\pm0.001$ and clinozoisite with anomalous blue interference colors occur in several localities, together with zoisite.

The amount of plagioclase (An_{37}) ranges from a few grains to about 50 percent. Scapolite is lacking in most diopside gneiss layers, but generally abounds in those layers which contain abundant grossularite. Scapolite from such lime-silicate rock about 1 mile east of the mouth of Whiskey Creek shows $\epsilon=1.549\pm0.001$ and $\omega=1.584\pm0.001$ containing, according to Winchell's diagram, 60 percent meionite (Winchell, 1951).

Sphene occurs in small wedge-shaped crystals which measure from 0.1 to 1 mm in the longest dimension. Rounded elongated grains of apatite 0.1 to 0.3 mm in diameter are common but much less so than the crystals of sphene. Graphite abounds in some layers. For ex-

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[1] Determination by F. A. Hildebrand, U.S. Geological Survey.

ample, in many diopside gneiss layers in the limestone quarry 2 miles south of Orofino numerous small flakes of graphite are between small angular plagioclase and quartz grains. In some localities, hornblende occurs with the diopside, and is probably an alteration product. It occurs as green spots in the diopside and rims the diopside grains in many specimens collected along Orofino Creek.

The chemical composition of a layer in diopside gneiss (No. 109) is shown in table 12. The layer sampled is rich in epidote minerals and diopside and contains nearly equal amounts of plagioclase and quartz. In its composition and mineralogy this rock resembles the diopside-plagioclase gneiss of the Wallace formation, east and north of the Orofino district (compare table 2, No. 289), except that the diopside-plagioclase gneiss of the Wallace formation does not contain epidote minerals and generally has more plagioclase.

Iron oxide and iron sulfide occur in small vugs in a quartzite layer and along the contact between the quartzite and an overlying marble layer exposed in a railroad cut 4 miles east of Orofino. The small, irregular-shaped vugs, which are filled by black, brown, yellow, and white powdery products of weathering of iron ore minerals are flattened parallel to the bedding of the quartzite and elongated parallel to the lineation. Most of the overlying carbonate layer has been converted to a lime-silicate rock with abundant diopside and scapolite. Small tabular scapolite crystals also line the walls of the vugs next to the lime-silicate rock. It may be significant that scapolite has been found in many biotite gneiss layers of the Wallace formation north of the mapped area (see table 3) but not in any other formation in the Belt series.

Biotite quartzite is exposed in two sections; one is opposite the mouth of Canyon Creek and the other north of Ahsahka. In both localities biotite quartzite is thin bedded and contains layers rich in plagioclase. The amount of biotite varies from one layer to the next. In layers rich in quartz, biotite occurs in thin laminae (1 to 3 mm thick) separating the individual quartzite beds, which range from 1 to 4 cm in thickness. In the plagioclase-bearing layers many of which have a composition of biotite-plagioclase gneiss, bedding becomes apparent because of alternation of thin layers consisting either of pure quartz, or of quartz and plagioclase, or of quartz, plagioclase, and biotite.

Microscopic study shows that quartz grains are strongly strained and elongated parallel to the bedding. Biotite flakes occur between the quartz grains; they are small and fairly well oriented parallel to the bedding. In some layers, quartz grains are clustered or form thin layers that are separated by paper-thin sheets of biotite flakes. The clusters are elongated and give an impression that they once were larger grains and were granulated during deformation and recrystallization. Individual large rounded plagioclase grains occur with quartz, and in some layers they form about 60 percent of the rock. Muscovite occurs as individual large flakes, groups of small flakes, and fibers. Magnetite, apatite, and sphene are common accessories. A few small grains of allanite surrounded by epidote and small rounded zircon crystals with pleochroic halos occur in some beds. Small pink round garnet crystals are common in many plagioclase-bearing layers.

The layer of sillimanite quartzite south of Freeman Creek is strongly deformed and has a distinct lineation, marked by parallel orientation of sillimanite needles and drawn-out muscovite. The coarse-grained quartzite beneath the sillimanite-bearing layer resembles the Revett quartzite in the Headquarters quadrangle.

THIN-BEDDED BIOTITE GNEISS AND BIOTITE SCHIST WITH HORNBLENDE-BEARING OR WITH DIOPSIDE- AND SCAPOLITE-BEARING LAYERS

The layers of biotite gneiss and biotite schist are medium to fine grained and weather more readily than the layers that contain calcium-magnesium silicate minerals (fig. 33). Minerals in the biotite gneiss layers are quartz, plagioclase, biotite, and magnetite. A few hornblende crystals occur with biotite. The schistose layers contain more biotite and less plagioclase than the gneissic layers.

Most layers rich in calcium-magnesium silicate minerals contain more quartz and less plagioclase than the biotite gneiss layers. Diopside and hornblende may occur together or in separate thin layers. Thin-section study shows that the hornblende that occurs with diopside is an alteration product of diopside. It rims the diopside grains or occurs as small irregular-shaped grains either next to the diopside or parallel to the cracks and cleavage in the diopside. It is strongly pleochroic with Z=blue-green, Y=green, X=pale. The hornblende not associated with diopside occurs as large independent crystals and does not show the pronounced bluish-green color typical of the secondary hornblende. Its pleochroism is Z=green, Y=green, X=light green.

Epidote and zoisite form either fine-grained aggregates that seem gray under the microscope or large grains that include numerous tiny round quartz grains. Near Dent epidote minerals of this type were found to be alteration products after plagioclase.

Scapolite is more abundant in the diopside-bearing layers than in the hornblende-bearing layers. Probably more scapolite crystallized in the beds richer in calcium; it was also noted, however, that where

TABLE 12.—*Chemical composition, ionic percentage, norm, and minerals of metasedimentary rocks near Orofino*

[Marietta Corbin, analyst, U.S. Geological Survey]

Locality and specimen No.		109		106		145		146		147	
Rock type		Epidote-rich layer in diopside-plagioclase gneiss.		Hornblende gneiss		Biotite-hornblende gneiss.		Biotite gneiss		Light-colored layer in biotite gneiss.	
Location		One mile north of Ahsahka		One mile north of Ahsahka		Railroad tunnel south of Orofino		Railroad tunnel south of Orofino		Railroad tunnel south of Orofino	
Chemical composition and ionic percentage											
Constituent											
Conventional symbol	Symbol rearranged for cation percent	Weight percent	Cation percent	Weight percent	Cation percent	Weight percent	Cation percent	Weight percent	Cation percent	Weight percent	Cation percent
SiO_2		55.05	52.62	70.53	66.61	67.16	62.64	74.37	69.90	79.32	73.83
Al_2O_3	$AlO_{3/2}$	13.25	14.93	13.39	14.91	15.52	17.05	13.57	15.02	12.31	13.51
Fe_2O_3	$FeO_{3/2}$	2.79	2.01	1.75	1.25	1.24	.87	.59	.42	.16	.11
FeO		2.49	1.99	2.57	2.03	2.62	2.05	1.60	1.26	.42	.33
MnO		.09	.07	.14	.12	.08	.06	.04	.03	.01	.01
MgO		4.62	6.58	1.63	2.29	1.42	1.97	.84	1.17	.23	.32
CaO		19.52	19.98	3.94	3.98	3.91	3.90	2.77	2.79	1.24	1.24
Na_2O	$NaO_{1/2}$	.49	.91	4.22	7.73	4.60	8.31	4.21	7.67	5.50	9.92
K_2O	$KO_{1/2}$	.07	.08	.52	.62	2.23	2.66	1.19	1.42	.51	.60
TiO_2		.66	.47	.50	.35	.50	.35	.31	.22	.16	.11
P_2O_5	$PO_{5/2}$	.19	.15	.13	.10	.17	.14	.12	.09	.01	.01
CO_2		.16	.21	.01	.01	.00		.01	.01	.01	.01
H_2O+		.71	(2.26)	.59	(1.86)	.39	(1.22)	.36	(1.13)	.13	(.40)
H_2O-		.07		.10		.08		.03		.02	
Total		100.16	100.00	100.02	100.00	99.92	100.00	100.01	100.00	100.03	100.00
	O		159.23		169.16		165.45		172.31		175.11
	OH		4.52		3.72		2.44		2.26		.80
	Total anions		163.75		172.88		167.89		174.57		175.91
Norm											
Mineral		CIPW	Molecular	CIPW	Molecular	CIPW	Molecular	CIPW	Molecular	CIPW	Molecular
Q		16.71	16.00	32.83	31.01	21.12	19.72	37.57	35.34	42.29	39.35
Or		.39	.40	3.06	3.10	13.19	13.30	7.01	7.10	3.01	3.00
Ab		4.14	4.55	35.70	38.65	38.90	41.55	35.60	38.35	46.51	49.60
An		33.76	34.85	16.07	16.40	15.10	15.20	13.02	13.15	6.01	6.05
C								.59	.67	.52	.57
Wo		25.44	25.10	1.10	1.04	1.37	1.26				
En		11.51	13.16	4.06	4.58	3.53	3.94	2.09	2.34	.57	.64
Fs		1.36	1.16	2.72	2.36	3.11	2.66	2.02	1.72	.40	.36
Ap		.43	.40	.30	.27	.40	.37	.27	.24	.03	.03
Il		1.24	.94	.94	.70	.94	.70	.59	.44	.30	.22
Mt		4.05	3.02	2.55	1.87	1.81	1.30	.86	.63	.23	.16
Cc		.36	.42	.02	.02			.02	.02	.02	.02
H_2O		.78		.69		.47		.39		.15	
Total		100.17	100.00	100.04	100.00	99.94	100.00	100.03	100.00	100.04	100.00
Mineral content calculated on basis of chemical analysis											
Quartz		16.52		34.56		22.4		39.48		42.57	
Plagioclase		9.05		47.30		49.8		47.00		52.67	
(anorthite content)		(38)		(27)		(22)		(24)		(11)	
Microcline				2.36		10.0		2.96		1.47	
Biotite						6.0		8.58		2.15	
Muscovite										.85	
Hornblende				12.00		9.2					
Diopside		34.34									
Epidote		36.97				1.2		1.52			
Sphene		1.57				.9					
Apatite		.45		.32		.3		.25		.03	
Ilmenite				.89							
Magnetite		1.00		1.87				.10			
Hematite				.43							
Rutile								.01		.09	
Calcite		.36						.02		.02	
Subtotal		100.26		99.73		99.8		99.92		99.85	
$+H_2O$		[1].10		.30				.08		.07	
Total		100.16		100.03		99.8		100.00		99.92	

[1] This value is—Na_2O, rather than $+H_2O$.

diopside is altered to hornblende, the scapolite is altered to sericite. Scapolite is a common mineral in small lenses of lime-silicate rock that occur in thin-bedded biotite quartzite along Orofino Creek, 11 miles east of Orofino (loc. 805, pl. 4). Diopside in these lenses is surrounded by bluish-green hornblende.

As a rule, the plagioclase (An_{37}) occurs in grains three times as large as the average quartz grain in the same rock. Garnet in irregularly shaped grains, small rounded crystals of sphene, and magnetite are the common additional constituents.

HORNBLENDE GNEISS, HORNBLENDE-BIOTITE GNEISS, AND BIOTITE GNEISS

The gneisses are medium-grained rocks in which variation in amounts of the four main constituents—quartz, plagioclase, biotite, and hornblende—have produced a distinct banding. The contrast in color and composition between bands varies; in some localities fairly homogeneous biotite gneiss layers, 10 to 30 cm thick, alternate with hornblende-bearing layers, both rock types containing about an equal amount of dark minerals. In contrast to this type, in some layers the amount of dark constituents in individual thin beds ranges from negligible to about 20 percent; this causes a distinct white to dark-gray banding. The thickness of the bands ranges from 1 to 10 cm.

Thicker beds of homogeneous hornblende-biotite gneiss, which are interlayered with biotite gneiss or biotite schist, may contain either small epidote grains or small pink garnet crystals. Plagioclase (An_{26}) occurs either as rounded grains two or three times as large as the quartz grains or as polygonal grains the same size as the quartz grains. The layers of hornblende-biotite gneiss with large rounded plagioclase grains commonly contain epidote and grade into biotite gneiss; those with granoblastic texture contain garnet and grade into amphibolite.

The hornblende, which is a bluish-green variety, occurs as small grains with biotite between the plagioclase grains. Its pleochroism is Z=blue-green or green, Y=green, X=pale yellow green. The indices of refraction $\alpha'=1.661\pm0.001$ and $\gamma'=1.683\pm0.001$ are close to those of the bluish-green hornblende in gabbro and diorite in this region. All bluish-green hornblende is an alteration product of diopside, whereas the green hornblende in the gneiss most likely crystallized during the metamorphism from a marly clay matrix that cemented the quartz and plagioclase grains in the original marly sandstone.

Biotite in the hornblende-biotite gneiss is a common strongly pleochroic brown variety with $\beta=\gamma=1.653\pm0.001$. Epidote is slightly pleochroic (greenish yellow to colorless) and occurs as small subhedral grains with hornblende and biotite. Sphene commonly occurs as small rounded wedges, but in some layers it surrounds grains of ilmenite-magnetite. No zircon was found in the thin sections studied.

The hornblende-biotite gneiss grades either to hornblende gneiss or to biotite gneiss. These gneisses resemble the hornblende-biotite gneiss except that each contains only one dark constituent.

A layer of fine-grained hornblende gneiss interbedded with biotite gneiss 1 mile north of Ahsahka was analyzed chemically (table 12, No. 106). The major constituents of this layer are plagioclase (An_{27}), in round grains that measure 0.25 to 0.5 mm in diameter, and quartz. The only dark constituent is hornblende and it occurs as small grains between the rounded plagioclase grains. The amount of microcline is less than 3 percent.

Specimens of three succeeding bands in distinctly banded hornblende-biotite gneiss south of Orofino (fig. 36A, loc. 147) were analyzed (table 12, Nos. 145, 146, and 147). Specimen No. 145 which is from the darkest band in this gneiss contains more iron, magnesium, calcium, and potassium and less silicon than the adjacent layers. Accordingly the thin sections show less quartz and more dark constituents and microcline. The plagioclase in this darkest band occurs as round or oval grains that are 1 mm in diameter and have fringed borders. The interstices between the plagioclase grains are filled by quartz grains that range from 0.1 to 0.5 mm in diameter, by small grains of microcline, hornblende, and biotite, and by tiny grains of epidote. The accessory minerals are apatite, sphene, and a few tiny crystals of zircon. Myrmekite is seen along the borders of some plagioclase grains and between microcline and plagioclase.

Specimen No. 146 is from a medium gray layer adjacent to the darkest band. Biotite, the only dark constituent in this medium gray layer, is concentrated in thin laminae parallel the bedding (figs. 36*B* and *C*). Quartz is more abundant than in the darkest band, but the amounts of plagioclase are about equal (table 12). The texture of the medium-gray layer is similar to that of the darkest band; plagioclase occurs in large round or oval grains and the other minerals fill the interstices. This layer also contains some epidote and a few small grains of apatite and zircon.

Specimen No. 147 is from a layer that is lighest in color. Table 12 shows that this layer contains more plagioclase and quartz and far less dark minerals than the adjacent layers. The amount of plagioclase is larger than that of quartz. Biotite, the only dark constituent, forms about 2 percent of this layer.

In all layers the biotite is of a dark reddish brown

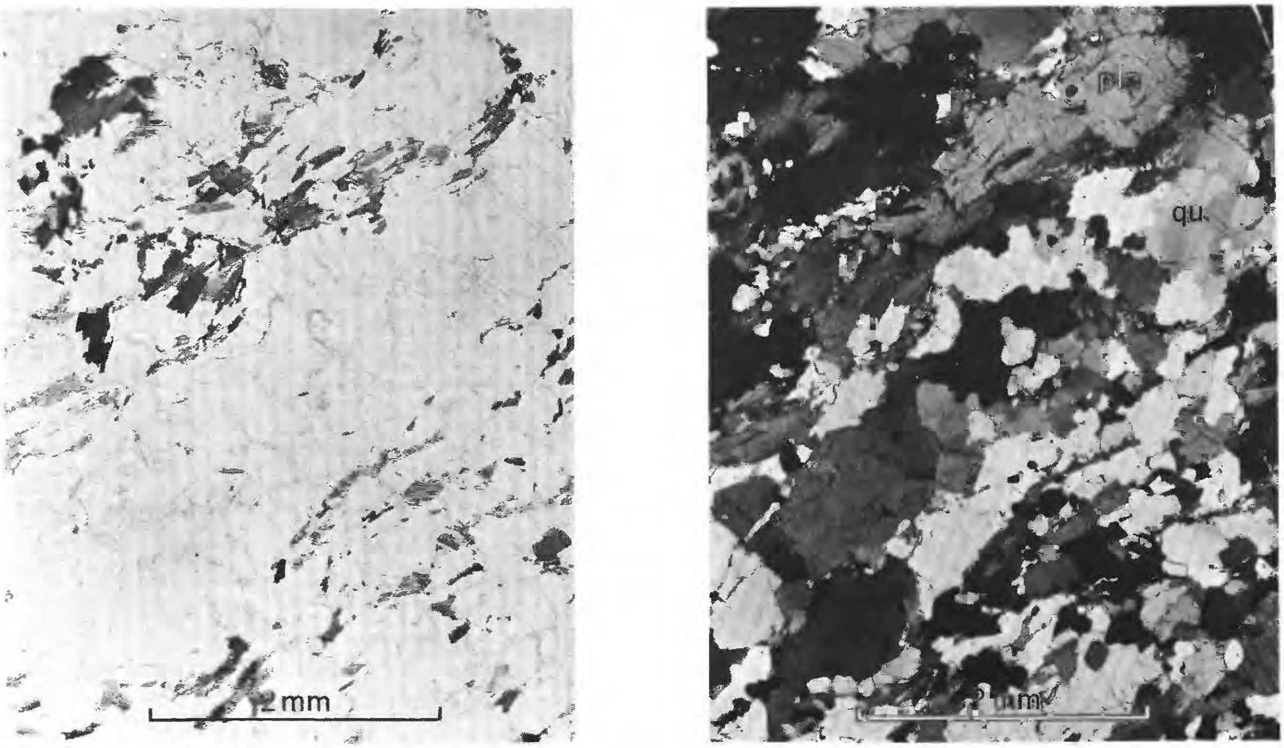
A. Dark bands contain hornblende and biotite; in the lighter colored bands biotite is the only dark constituent. Veinlets consist of quartz and plagioclase. (Loc. 147, pl. 4.)

iron-rich variety. The approximate chemical composition of the biotite in specimen No. 146 can be calculated from the rock analysis of the biotite gneiss (No. 146). Because biotite is virtually the only dark constituent in this rock, all magnesium oxide (0.4 percent of the rock) and remainder of iron oxides after subtraction of magnetite should be contained within it. The calculation indicates that the biotite contains about 10 percent magnesium oxide and about 18 percent ferrous oxide.

AMPHIBOLITE

The hornblende gneiss locally grades into darker layers that are richer in hornblende and poorer in quartz. Many of these dark amphibolite layers contain small garnet crystals; some are interbedded with green layers rich in diopside. Kyanite and garnet occur with aluminous anthophyllite (gedrite) in one locality.

The amphibolite layers are generally thicker and more homogeneous that the hornblende-biotite gneiss layers. Many layers are near limestone beds for example, in the town of Orofino a layer of dark amphibolite is exposed just west of a limestone layer. Toward the south the limestone thins out and its place is taken by amphibolite which locally contains abundant epidote.



B. Photomicrograph of biotite gneiss (No. 146). Banding is caused by the variation in the amount of biotite (dark). Plane polarized light.

C. Same thin section as shown in *B*, with crossed nicols. Quartz (qu) and plagioclase (pla) occur in anhedral or rounded grains of small to medium size.

FIGURE 36.—BANDING IN HORNBLENDE-BIOTITE GNEISS ALONG THE CLEARWATER RIVER SOUTH OF OROFINO

The major constituents of amphibolite are plagioclase (An_{50}) and hornblende; quartz occurs in smaller amounts and epidote is a common additional constituent. Abundant sphene in oval or in wedge-shaped crystals and some magnetite and apatite are the accessory minerals. In many layers abundant epidote occurs as polygonal grains of medium size. The association of plagioclase (An_{50}) with epidote in these layers indicates that the temperature and pressure during metamorphism corresponded to those of the amphibolite facies. Texture of most amphibolite is granoblastic or lepidoblastic.

DIOPSIDE AMPHIBOLITE

Some of the amphibolite at Bruces Eddy and outcrops 2 miles to the north contain thin light-green layers that are rich in diopside. The rock at Bruces Eddy is fine to medium grained and the individual layers range from 1 to 10 mm in thickness, but at locality 443—2 miles to the north—the rock is coarse grained and the layers are somewhat thicker. Layers of dark biotite schist ranging from 10 to 100 cm in thickness are interbedded with the diopside amphibolite in this locality.

Thin sections show that epidote is a common minor constituent in diopside amphibolite. Plagioclase (An_{55}) occurs as polygonal grains of medium size and is more abundant in the diopside-bearing layers than in the hornblende-bearing layers. Many of the hornblende-rich layers also contain diopside. Spots of hornblende occur within many large diopside grains, and some smaller grains are rimmed by hornblende; thus also in this rock some of the hornblende is an alteration product of diopside. In locality 443 plagioclase is partly altered to sericite and to epidote minerals—zoisite, clinozoisite, and epidote. Sphene, garnet, and apatite occur as accessories.

GARNET AMPHIBOLITE

Many layers of amphibolite near Orofino contain small red garnet crystals and some biotite flakes. These garnet-bearing amphibolites are medium grained and grade to garnetiferous biotite gneiss, representing calcium-rich layers in the gneiss formation. The garnet amphibolite exposed for a mile along Orofino Creek east of Lime Mountain, 12½ to 13½ miles east of Orofino differs from the layers near Orofino in several aspects. It is coarser grained and darker and occurs as a larger massive body. It contains lenticular inclusions of diopside-plagioclase gneiss which in composition are similar to the diopside-plagioclase gneiss just east of the garnet amphibolite (figs. 37*A* and *B*). The minerals in the garnet amphibolite are plagioclase (An_{55}), hornblende, and garnet, with some biotite, apatite, sphene, and ilmenite. Garnet occurs in round or subhedral grains that measure about half a centimeter in diameter and contain small rounded inclusions of quartz and sphene (fig. 37*B*). The index of refraction of the garnet is $n=1.783\pm0.002$. Its chemical composition is shown in table 13. The garnet contains a considerable amount of calcium and an X-ray study made by C. L. Christ and F. A. Hildebrand of the U.S. Geological Survey shows that two phases are present in the garnet, as might be expected if the original rock was a carbonate-bearing layer in which the grossularite crystallized first and was later in part replaced by a pyrope-almandite garnet. This supports the view that the garnet amphibolite was derived from a carbonate-bearing layer by a metasomatic addition of iron and magnesium. There is an exceptionally large amount of apatite also in some small remnants of diopside-

TABLE 13.—*Chemical analysis of garnet from garnet amphibolite (No. 354) on Orofino Creek*

[Harry M. Hyman, analyst, U.S. Geological Survey]

Constituent	*Weight percent*	*Molecular equivalent*	*Composition*	
SiO_2	38.40	6394	Pyrope	22.4
Al_2O_3	21.43	2102	Almandite	59.0
Fe_2O_3	.75	47	Spessartite	2.4
FeO	26.73	3721	Grossularite	14.0
MnO	1.09	154	Andradite	2.2
MgO	5.70	1414		
CaO	5.72	1020		
Na_2O	.02			
K_2O	.03			
TiO_2	.24			
H_2O+	.00			
H_2O-	.02			
Total	100.13			100.0

TABLE 14.—*Chemical composition, ionic percentage, norms, and minerals of garnet amphibolite specimen No. 360 from Orofino Creek*

[Marietta Corbin, analyst, U.S. Geological Survey]

Constituent		Weight percent	Cation percent	Norm			Minerals	
Conventional symbol	Symbol rearranged for for cation percent			Mineral	CIPW	Molecular		
SiO_2	------	49.56	46.03	Q	------	------		
Al_2O_3	$AlO_{3/2}$	19.40	21.23	Or	3.34	3.35	Plagioclase (An_{40}).	53.7
Fe_2O_3	$FeO_{3/2}$	1.01	.70	Ab	30.25	32.20	Hornblende	30.2
FeO	------	7.78	6.04	An	35.21	35.30	Garnet	10.0
MnO	------	.16	.13	Wo	4.97	4.74	Biotite	4.0
MgO	------	5.89	8.15	En	3.24	3.64	Apatite	.2
CaO	------	9.82	9.77	Fs	2.80	2.40	Sphene	1.3
Na_2O	$NaO_{1/2}$	3.58	6.44	Fo	8.01	9.49	Calcite	.4
K_2O	$KO_{1/2}$	.56	.67	Fa	7.46	6.10	Magnetite.	.3
TiO_2	------	.79	.55	Ap	.20	.19		
P_2O_5	$PO_{5/2}$	.08	.07	Il	1.50	1.10		
CO_2	------	.17	.22	Mt	1.46	1.05		
H_2O+	------	1.21	------	Cc	.39	.44		
H_2O-	------	.04	------	H_2O	1.25	------		
Total		100.05	100.00		100.08	100.00		100.1
O			150.56					
OH			7.50					
Total anions.			158.06					

plagioclase gneiss. Part of the phosphorus in this apatite may have been introduced metasomatically. Chemical analysis and minerals of the garnet amphibolite No. 360, are shown in table 14.

KYANITE-GARNET AMPHIBOLITE

A coarse-grained dark amphibole-rich rock is exposed in a railroad cut along Orofino Creek about 4½ miles east of Orofino (loc. 813). A thin-bedded biotite schist with diopside- and hornblende-bearing layers occurs above and beneath it.

Amphibole in this coarse-grained dark rock occurs in long prisms, and numerous red garnet crystals ranging from 1 to 6 mm in diameter are embedded between them. Pale-blue kyanite crystals average about 1 cm. in length and occur between the amphibole prisms or are included in them (fig. 38).

Microscopic study shows that the amphibole is orthorhombic, exhibits weak pleochroism: Z=Y=greenish, X=colorless, and has the following indices of refraction: $\alpha=1.649\pm0.001$, $\beta=1.656\pm0.001$, $\gamma=1.669\pm0.001$. These are the properties of anthophyllite with about 28 percent iron.

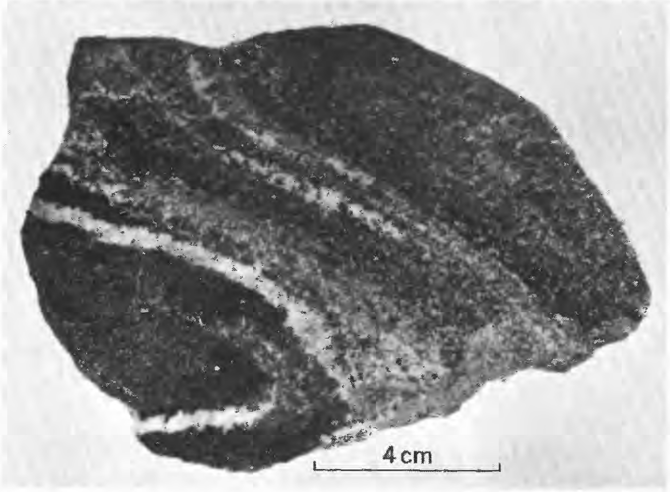
The pink garnet shows $n=1.775\pm0.001$, and the unit-cell length[2] $a_0=11.526\pm0.006$ A suggests that it belongs to the pyralspite group. The large garnet crystals have abundant inclusions of quartz. Kyanite crystals, a few small biotite flakes, staurolite grains, and numerous small round brown rutile grains all are included in the anthophyllite. These inclusions were obviously crystallized earlier and were enclosed by growing anthophyllite prisms. The light-colored constituents—quartz and plagioclase (An_{20})—fill the interstices between the anthophyllite prisms.

It is possible that a metasomatic addition of magnesium gave rise to the crystallization of this exceptional mineral association. In the schist just east of the kyanite-garnet amphibolite, accumulations of hornblende were noted, and farther east a feldspathized biotite-gneiss occurs. Both are indications of a metasomatic activity in this area.

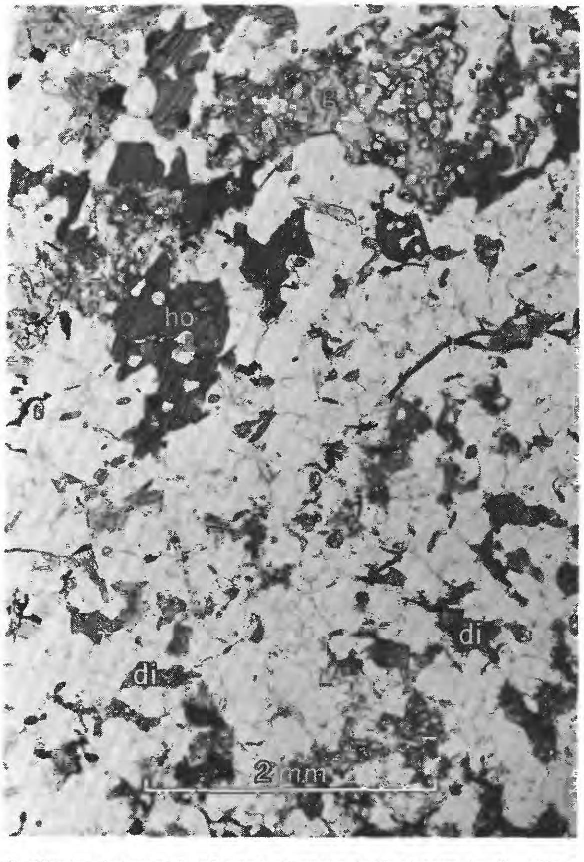
BIOTITE-GARNET SCHIST

The biotite-garnet schist is coarser grained and contains more biotite and less plagioclase than the biotite-gneiss. Sillimanite may occur as a minor constituent. Muscovite is far less common in the biotite-garnet schist in the Orofino district than in the schists in the Headquarters quadrangle. Garnet is pink or red and occurs in small crystals.

Microscopic study confirms that quartz is the main light-colored constituent and that plagioclase (An_{25})



A. Polished hand specimen (No. 354). White bands are plagioclase-quartz veinlets; gray band in the center contains diopside.



B. Photomicrograph of part of same rock shown in *A.* Garnet (g) occurs in rounded holoblasts that include quartz. Generally the hornblende crystals (ho) are larger than the diopside grains. The light minerals are quartz and plagioclase. Plane polarized light.

FIGURE 37.—SPECIMEN OF A FOLD AND RELICT BEDDING IN GARNET AMPHIBOLITE FROM THE SIDE OF OROFINO CREEK ABOUT 13 MILES EAST OF OROFINO

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[2] Determination by F. A. Hildebrand.

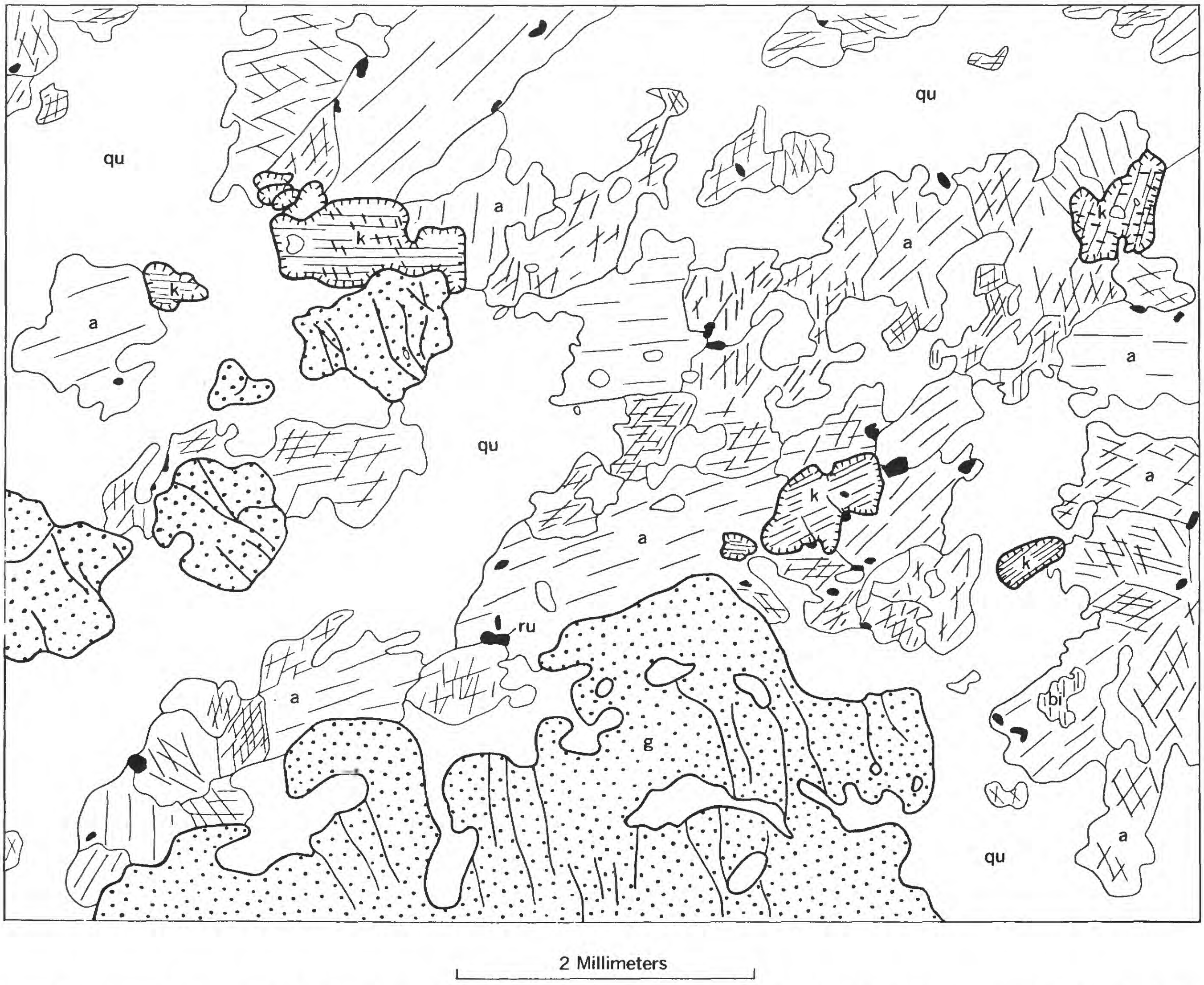


FIGURE 38.—Camera lucida drawing of kyanite-garnet amphibolite exposed along Orofino Creek (loc. 813, pl. 4). qu=quartz; a=anthophyllite; g=garnet; k=kyanite; bi=biotite; ru=rutile.

occurs in minor amounts. Apatite, magnetite and zircon are the accessories.

METASEDIMENTARY ROCKS NEAR PECK

The metamorphic rocks along the Clearwater River near Peck differ in appearance, texture, and mineralogy from the metasedimentary rocks near Orofino. Bedding is not so distinct near Peck as it is near Orofino. The major part of the metamorphic rocks consists of dark fine-grained biotite gneiss and biotite schist with some quartzitic layers and epidote-rich amphibolite layers. The biotite schist is rich in small flakes of biotite that are well oriented parallel to the foliation. Some layers show signs of shearing and contain larger angular grains of feldspar (fig. 39); in other layers the feldspar grains are rounded and small grains of quartz and biotite fill the interstices. Epidote and muscovite are common minor constituents in these layers.

The amphibolite layers consist of green hornblende, plagioclase (An_{15-19}), quartz, and epidote. The variation in the relative amount of these minerals gives rise to a layering in the rock. Some layers contain a considerable amount of chlorite. Epidote occurs usually in small prisms in the rounded plagioclase grains (fig. 40).

A section northeast of the town of Peck consists of epidote amphibolite with some light-colored muscovite-plagioclase (An_8) schist layers. The amphibolite contains large grains of plagioclase, and in many layers the amount of plagioclase is sufficient to give the rock a dioritic composition. The muscovite-plagioclase schist contains scattered grains of euhedral plagioclase (fig. 41). This is the only layered rock in the mapped

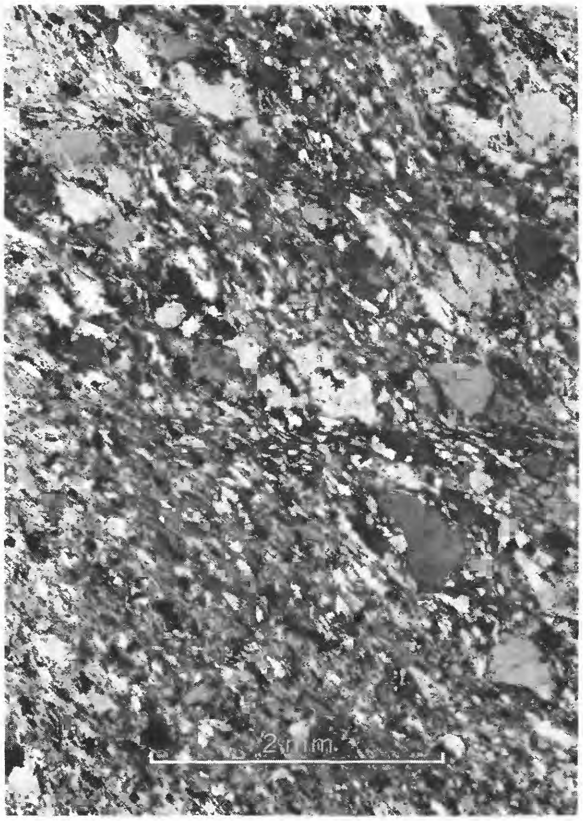


FIGURE 39.—Photomicrograph of sheared biotite schist specimen from near Peck. Large angular grains are plagioclase (loc. 179). Crossed nicols.

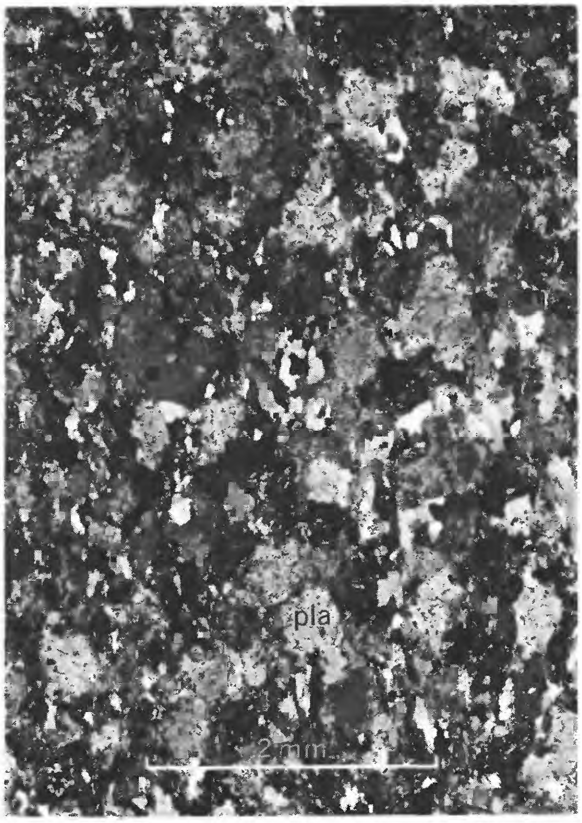


FIGURE 40.—Photomicrograph of amphibolite from near Peck (loc. 178, pl. 4). The round plagioclase grains (pla) contain small epidote inclusions. Small grains of hornblende and quartz fill the interstices between the plagioclase grains. Crossed nicols.

area in which plagioclase is euhedral. In metasedimentary rocks the early plagioclase occurs as round grains; the secondary plagioclase also is found as round grains but of larger size. In many igneous rocks, such as gabbros and dike rocks, the plagioclase is subhedral. Therefore it seems possible that the muscovite-plagioclase schist with euhedral plagioclase may contain tuffaceous material.

The folded layers of amphibolite exposed just north of Peck (loc. 168) are thinly laminated. Dark laminae contain abundant hornblende, and light-colored laminae are rich in epidote minerals and contain plagioclase, quartz, muscovite, chlorite, magnetite, apatite, and green spinel. The composition and mineral content of these layers are much like those of the gabbroic metamorphic rock north of this locality. It is possible that the laminated amphibolite is a metamorphosed basalt of basaltic tuff. The intercalated quartzitic layers support the view that this formation consists mainly of depositional material.

CHEMICAL COMPOSITION AND MINERAL ASSOCIATIONS

The mineral associations described in an earlier part show that the metasedimentary rocks were metamorphosed to the amphibolite facies (Eskola, 1914; Barth and others, 1939). All rocks of the Orofino district are poor in potassium feldspar and contain little if any muscovite. Most of the marble consists of a very pure calcite as is common in the areas of high-temperature metamorphism. The dolomite reacted with silica to form diopside or hornblende, but because the temperature during metamorphism was not high enough to form wollastonite, the calcite recrystallized and segregated to form lens-shaped bodies. The occurrence of many layers exceptionally rich in hornblende near the marble lenticles suggests that calcium and magnesium of the dolomitic limestone joined with aluminum and silicon of the clayey layers near the limestone to form hornblende. In dolomitic sand layers adjacent to limestone, diopside crystallized at the expense of dolomite and silica. The diopside and hornblende crystallized at the same temperature and pressure but from a chemically different material. In the equilibrium diagram (fig. 42), where the results of the chemical analyses are plotted, the points for Nos. 109 and 106 representing these two compositions fall into separate triangles. The epidote-rich layer in diopside quartzite (No. 109, table 12) lies in the triangle near the calcium corner. A possible

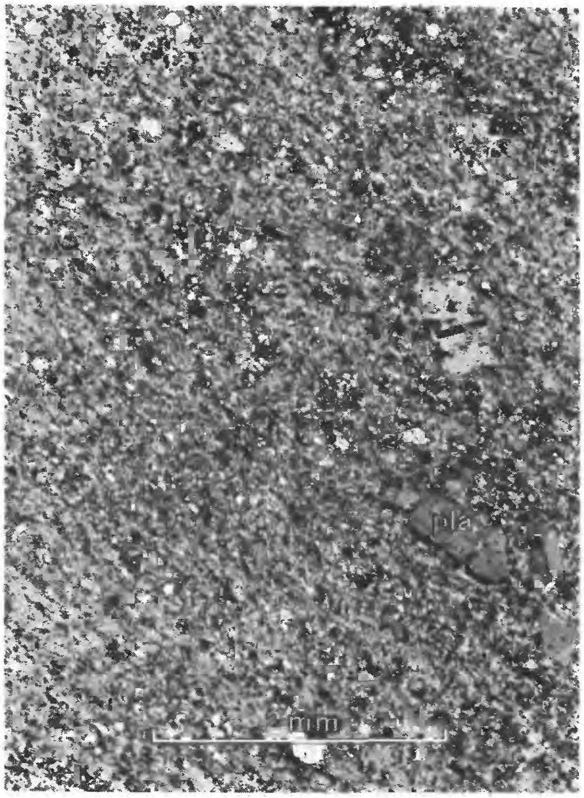


FIGURE 41.—Photomicrograph of euhedral plagioclase crystals (pla) in a fine-grained matrix of quartz, plagioclase, and muscovite. Locality 170 near Peck. Crossed nicols.

mineral association is diopside, grossularite, epidote, and anorthite. The garnet amphibolite (No. 360, table 14) and hornblende gneiss (No. 106, table 12) are in the anorthite-tremolite-anthophyllite triangle. The garnet of the garnet amphibolite is a mixture of almandine, pyrope, and grossularite. The point for the analysis of the biotite-hornblende gneiss, No. 145, is on the anorthite-biotite line. The analysis points for the biotite gneiss, Nos. 146 and 147, in which biotite is the only dark constituent, are very close to the same line.

Chemical composition of most layers in the Orofino series would be close to that of the layers shown in figure 36. There are no layers rich in aluminum silicates, as is typical of the schists of the Prichard and Wallace formations farther east and north of the Orofino region. For comparison, an analysis of kyanite-andalusite-sillimanite-cordierite gneiss (No. 912) from Boehls Butte quadrangle (Hietanen, 1956, table 4) and analyses of two typical layers in the schist (Nos. 246 and 247), two gneissic layers (Nos. 248 and 252), a quartzite (No. 254) and a diopside-plagioclase gneiss (No. 289) in the Headquarters quadrangle (table 2) were plotted in the same figure. Nos. 246 and 247 are layers in sillimanite-garnet schist, and No. 248 is a plagioclase-rich biotite gneiss. Figure 12 shows that the specimens 246, 247, and 912 differ strikingly from the typical layers of the metamorphic rocks in the Orofino district. The difference is due mainly to a larger amount of calcium and smaller amount of iron, magnesium, and aluminum in the rocks of the Orofino series. Only the biotite-plagioclase gneiss No. 248 has a composition close to that of the gneisses of the Orofino district. The composition of the diopside-plagioclase gneiss (No. 289) of the Headquarters quadrangle is similar to that of the corresponding layers near Orofino.

The occurrence of muscovite, chlorite, and epidote with plagioclase (An_{15-19}) near Peck suggests that the rocks there recrystallized at a somewhat lower temperature.

CORRELATION

The metasedimentary rocks near Orofino differ lithologically from the metasedimentary rocks in the areas east and north of the Orofino district. The metasedimentary formations to the east and north are metamorphosed equivalents of the Prichard, Burke, Revett, St. Regis, and Wallace formations of the Belt series in Latah and Shoshone Counties. The rocks of the Orofino district cannot be correlated with certainty with any of the formations in the south slope of the St. Joe Mountains described by Wagner (1949), in Latah County described by Tullis (1944), in Shoshone County described by Umpleby and Jones (1923), or in the Headquarters quadrangle described earlier in this paper. The thin-bedded scapolite- and diopside-bearing gneiss, as well as the thin-bedded diopside gneiss and biotite quartzite, resemble the thin-bedded diopside-plagioclase gneiss and biotite-quartzite in the Headquarters quadrangle. The sequence of thick-bedded biotite and hornblende gneisses, thin-bedded biotite-hornblende gneiss, biotite-quartzite, and diopside gneiss resembles the sequence found in the Wallace formation near Elk River (pl. 1). No limestone occurs near Elk River but limestone is common in the Wallace formation in Shoshone County. The major difference between the metasedimentary rocks near Orofino and the Belt series is an abundance of biotite and hornblende gneiss and a lack of aluminous layers near Orofino. There is a possibility that the biotite gneiss and biotite-hornblende gneiss were derived from shaly layers by an introduction of sodium and calcium at an early stage of metamorphism. This introduction would give rise to a development of abundant plagioclase. Yet, it is hard to believe that sodium and calcium would have been introduced selectively into some layers and not into the neighboring layers in the same strata. As has been

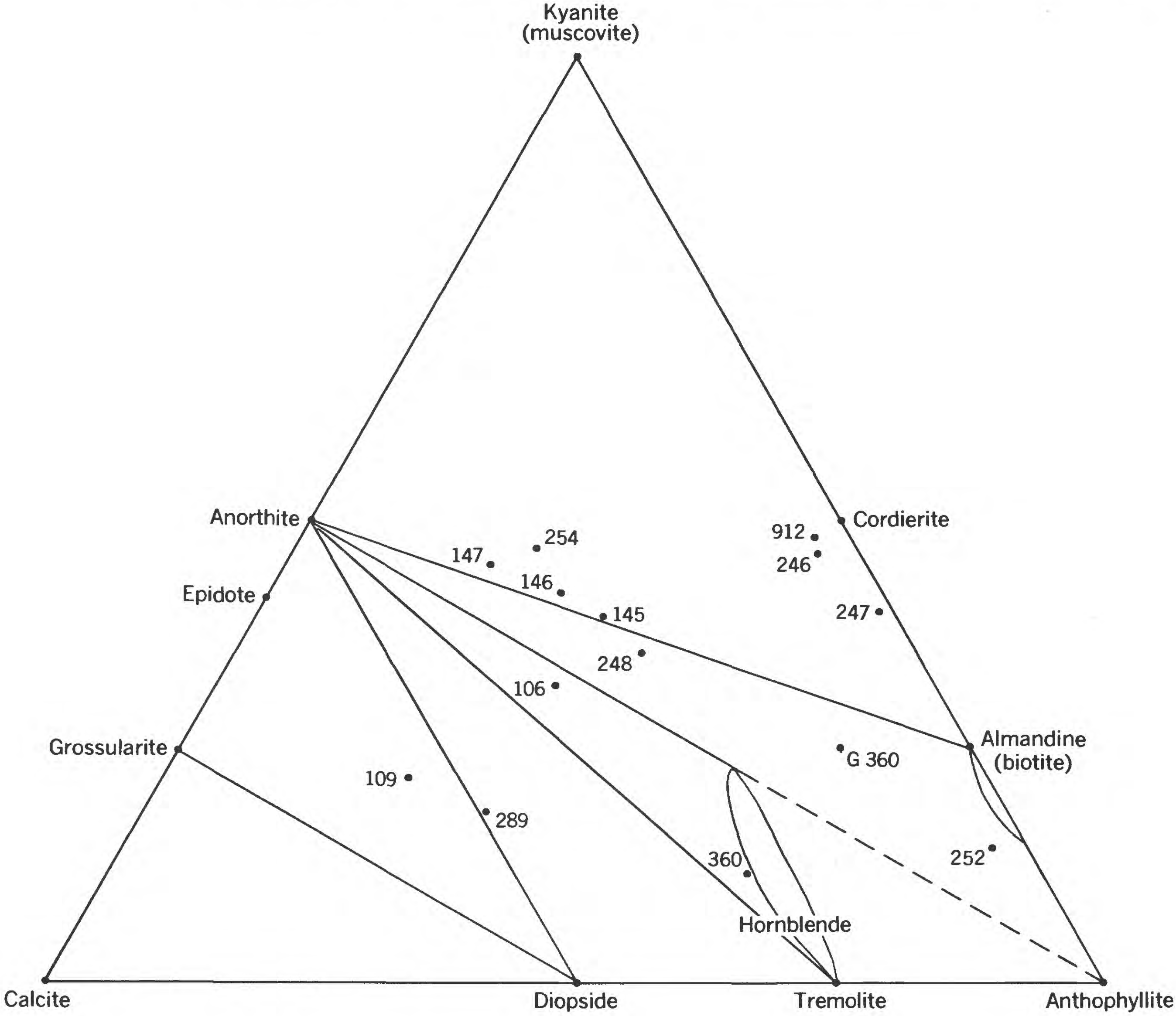


FIGURE 42.—Equilibrium diagram for metasedimentary rocks. Nos. 246, 247, 248, 252, 254 and 289 are from table 2, and 106, 109, 145, 146 and 147 from table 12. G 360 is garnet in garnet amphibolite No. 360 (tables 13 and 14). No. 912 is andalusite-kyanite-sillimanite-cordierite gneiss (Hietanen, 1956).

shown in the Headquarters quadrangle and will be shown in the Orofino district, a later metasomatism locally transformed all layers in the metasedimentary rocks to quartz diorite and tonalite. The irregular distribution of metasomatic minerals during this later phase also suggests the implausibility of selective introduction during an earlier phase.

Another explanation for the difference between the Orofino series and the rocks of the known Belt series would be a facies change during the deposition; this explanation seems unlikely because of the short distance involved. The first exposures of the Belt series with abundant aluminous layers are found along the river only 3 miles north of the Orofino district. Anderson (1930) earlier suggested that the Orofino series may be the lowest part of the Prichard formation not exposed elsewhere. Later he expressed the view that they could be much younger, probably Paleozoic or Mesozoic in age (Johnson, 1947, p. 491).

The lithologic character and stratigraphic sequence of the sillimanite-garnet schist, biotite schist, and coarse-grained quartzite near Huckleberry Butte closely resemble those of the schist of the Prichard and Burke formations and of the Revett quartzite exposed near Dent. Strikes and dips near Dent and at Huckleberry Butte suggest that these occurrences may belong to the same layer. If the coarse-grained quartzite along Orofino Creek, west of Lime Mountain, also is Revett quartzite, the sequence at Lime Mountain is equivalent to the lower part of the Wallace formation. The lithologic similarity of the section near Orofino and that at Lime Mountain would then suggest that all metasedimentary rocks in the Orofino district are equivalent to the lower part of the Wallace formation.

The metasedimentary rocks near Peck differ petrologically from those near Orofino and may belong to another formation. The presence of tuffaceous sedimentary material suggests a similarity to the Permian rocks exposed in small areas north and south of this locality (Ross and Forrester, 1947).

PLUTONIC ROCKS NEAR OROFINO

All rocks with the appearance and composition of igneous rocks but having igneous, metamorphic, or metasomatic origin are in this paper called plutonic rocks. On the basis of their composition the plutonic rocks of the Orofino region are divided as follows: (1) dunite, (2) hornblendite, (3) metagabbro and peridotite, (4) gabbro and norite, (5) quartz diorite, (6) hornblende pegmatite, (7) tonalite, and (8) pegmatite. Among the bodies of gabbro, diorite, and tonalite there are a few that seem intrusive, some that are products of the introduction of elements into metasedimentary rocks, and still others that probably are recrystallized metasedimentary rocks. Distinction among these three genetic varieties within each petrologic rock type is not clear; it is based mainly on studies of the structures, textures, and mineralogy in the field and in the laboratory. Mineralogically and chemically there is complete gradation from metasedimentary rocks to plutonic rocks.

PETROGRAPHIC DESCRIPTION

DUNITE

Small lenticular bodies of dunite occur in the gneissic and metasedimentary rocks. Johnson (1947, p. 492–494) has given a detailed petrographic description of three occurrences of dunite, and the reader is referred to his paper.

HORNBLENDITE

An occurrence of a very coarse-grained hornblendite in the railroad cut just west of Ahsahka has been described by Anderson (1933). Johnson (1947, p. 497–498) shows four occurrences on his map and describes them as dikelike bodies with gradational replacement contacts.

I studied in detail the hornblendite on the south side of the Clearwater River 3 miles west of Ahsahka (loc. 20, pl. 4). Two hornblendite bodies occur in a biotite-plagioclase gneiss near the contact of a gabbro (loc. 20, pl. 4). Contacts between the hornblendite and the gneiss are concordant or interfingering and small lenticles of hornblendite occur in the gneiss near the large bodies (fig. 43). Most of the hornblende is black and coarse grained but near the contact it is finer grained, foliated, and rich in biotite. A gneissic medium-grained light-gray tonalite is between the two hornblendite bodies. The tonalite is less foliated and richer in plagioclase than the biotite gneiss.

Thin-section study of the hornblendite shows that the angular areas between the large hornblende crystals are filled with plagioclase, quartz, and biotite. The texture in these areas is similar to that in the biotite gneiss (fig. 36*C*). The hornblende grains have rounded corners and include some small biotite flakes and rutile grains. Rutile in brown round grains and in tiny prisms and small grains of epidote occur in the interstitial rock between the hornblende grains and along the cracks in hornblende. The hornblende is green to bluish-green and its indices of refraction are $\alpha'=1.650 \pm 0.001$, $\gamma'=1.671 \pm 0.001$ and $Z \wedge c=23°$. The Mg:Fe ratio of 2.8 was calculated on the basis of the rock analysis (table 15, No. 20).

Chemical analysis of the hornblendite (No. 20, table 15) shows that this hornblendite is poor in iron and rich in magnesium, owing to a high Mg:Fe ratio in the hornblende.

METAGABBRO AND PERIDOTITE

The dark gneissic rocks exposed along the Clearwater River west of Peck (pl. 4) probably are part of a sill-like body in the gneissic metasedimentary rocks. Only the eastern contact of this body is exposed. In the west it disappears under the basalt. Most of the rock is a medium-grained mixture of hornblende, epidote, and zoisite. In some localities, large white rounded grains of plagioclase give this rock an appearance of coarse-grained gabbro. In other localities the rock is well foliated, and plagioclase forms small augen (fig. 44). The middle part of this metamorphosed sill consists of dark rock in which plagioclase grains are small and scarce.

Thin sections show that this dark rock is metamorphosed peridotite consisting of olivine, pyroxene, hornblende, serpentine, chlorite, and ilmenite-magnetite. The olivine grains are rimmed by reaction products (fig. 45) that consist of a bluish-green hornblende next to the olivine and serpentine and chlorite on outer zones. The augite is rimmed by hornblende and contains spots of this mineral. Zoisite with blue interference colors fills the space between olivine and augite. In some other specimens all olivine is altered to serpentine and chlorite. In the gabbroic parts of the sill hornblende, epidote, zoisite, plagioclase, and quartz are the main constituents. The hornblende is of the bluish-green variety which is common in this district as

FIGURE 43.—A hornblendite outcrop on a road cut along the Clearwater River about 2 miles west of Ahsahka. hb=hornblendite; gn=gneiss. Contact is indicated by a white line.

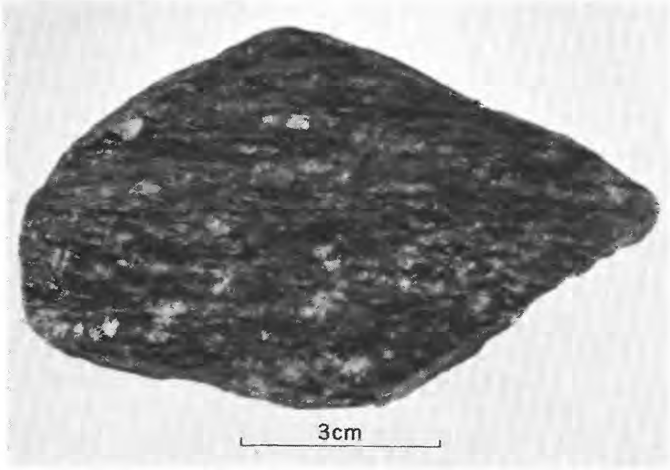


FIGURE 44.—Polished hand specimen of the metagabbro from near Harpers Bend (loc. 150, pl. 4). The light-colored augen consist of plagioclase and epidote.

an alteration product after pyroxene; it occurs as small individual prismatic grains or is clustered. In some specimens the plagioclase, which in the outcrops appears as large white grains, contains as much as 50 percent epidote. In some outcrops the major part of the white mineral is zoisite. The plagioclase which contains less epidote minerals includes quartz and hornblende.

In a few outcrops west of Harpers Bend the amount of plagioclase is so large that the rock can be called diorite. Epidote in this rock occurs as scattered small prisms in the plagioclase.

In their mineral content the sheared parts of the sill are similar to the gabbroic part, but the grain size in the sheared rock is smaller and the minerals are in parallel orientation. In the southernmost part of Harpers Bend (loc. 164) a light-colored layer, a few meters thick, occurs in metagabbro. This layer is sheared and consists of quartz and epidote (fig. 46) in equal amounts. Some other layers consist of about 75 percent epidote (specimen 153) and 25 percent quartz. It is possible that epidote and quartz were segregated along the shear zones during metamorphism and shearing. Another possibility is that calcic plagioclase was segregated from the gabbro magma—as were the olivine and pyroxene of the peridotitic parts—and that this plagioclase in turn was altered later to zoisite and epidote. In the dioritic parts the amount of epidote minerals is less because of the presence of secondary plagioclase. The composition of the new plagioclase

TABLE 15.—*Chemical composition, ionic percentage, norms, minerals, and Niggli values of plutonic rocks near Orofino*

[Marietta Corbin, analyst, U.S. Geological Survey]

Locality and specimen no.	20	32	180	42	43
Rock type	Hornblendite	Gabbro	Quartz diorite	Quartz diorite	Tonalite.
Location	Ahsahka	East of Peck	East of Peck	Greer	Greer.

Chemical composition and ionic percentage

Constituent: Conventional symbol	Constituent: Symbol rearranged for cation percent	20 Weight percent	20 Cation percent	32 Weight percent	32 Cation percent	180 Weight percent	180 Cation percent	42 Weight percent	42 Cation percent	43 Weight percent	43 Cation percent
SiO_2		46.16	43.47	52.60	49.33	57.30	53.49	64.05	60.25	68.73	63.89
Al_2O_3	$AlO_{3/2}$	12.48	13.85	18.52	20.47	17.64	19.40	17.24	19.11	17.20	18.84
Fe_2O_3	$FeO_{3/2}$	3.71	2.63	2.36	1.67	1.88	1.32	2.07	.73	1.30	.90
FeO		8.37	6.59	5.28	4.14	4.29	3.35	2.55	2.01	1.15	.89
MnO		.23	.18	.13	.10	.11	.08	.08	.06	.02	.02
MgO		12.48	17.50	6.14	8.58	4.26	5.93	2.06	2.89	.80	1.11
CaO		10.52	10.61	9.12	9.16	7.11	7.11	6.10	6.15	4.35	4.33
Na_2O	$NaO_{1/2}$	1.87	3.42	3.01	5.47	3.72	6.73	4.44	8.09	5.00	9.01
K_2O	$KO_{1/2}$	.65	.78	.46	.55	1.38	1.64	.17	.20	.55	.65
TiO_2		1.33	.94	.53	.35	.74	.52	.53	.36	.27	.19
P_2O_5	$PO_{5/2}$	.03	.02	.21	.17	.21	.17	.17	.13	.16	.12
CO_2		.01	.01	.01	.01	.10	.26	.02	.02	.04	.05
H_2O+		2.20	(6.91)	1.70	(5.32)	1.16	(3.61)	.54	(1.69)	.37	(1.14)
H_2O-		.09		.06		.06		.04		.05	
Total		100.13	100.00	100.13	100.00	99.96	100.00	100.06	100.00	99.99	100.00
	O		143.67		152.68		157.09		164.90		168.22
	OH		13.82		10.64		7.22		3.38		2.28
	Total anions		157.49		163.32		164.31		168.28		170.50

Norms

Mineral	20 CIPW	20 Molecular	32 CIPW	32 Molecular	180 CIPW	180 Molecular	42 CIPW	42 Molecular	43 CIPW	43 Molecular
Q			3.74	3.55	8.53	8.13	21.46	19.82	27.27	25.37
Or	3.84	3.90	2.73	2.75	8.12	8.20	1.00	1.00	3.23	3.25
Ab	15.83	17.10	25.48	27.35	31.46	33.65	37.54	40.45	42.31	45.05
An	23.72	24.13	35.65	36.12	27.37	27.58	26.61	27.05	20.41	20.40
C									.90	1.02
WO	11.78	11.48	3.45	3.28	2.51	2.08	1.06	1.00		
En	12.64	14.20	15.29	17.16	10.61	11.86	5.13	5.78	1.99	2.22
Fs	4.29	3.68	7.15	6.12	5.30	4.50	2.28	2.70	.63	.54
Fo	12.92	15.60								
Fa	4.83	4.02								
Ap	.06	.05	.50	.46	.50	.46	.40	.35	.37	.32
Il	2.52	1.88	.96	.70	1.40	1.04	.96	.72	.52	.38
Mt	5.37	3.94	3.43	2.50	2.73	1.98	3.01	1.09	1.88	1.35
Cc	.02	.02	.02	.02	.23	.52	.04	.04	.09	.10
H_2O	2.29		1.76		1.22		.58		.42	
Total	100.11	100.00	100.16	100.00	99.98	100.00	100.07	100.00	100.02	100.00

Minerals calculated on basis of chemical analysis

Mineral	20		32		180		42		43	
Quartz	2.3		7.8		13.3		22.1		28.3	
Plagioclase	15.2		42.4		51.4		55.4		58.0	
Anorthite content	(34)		(46)		(35)		(32)		(27)	
Microcline										
Biotite	1.0		1.0		12.6				3.0	
Muscovite			1.4						2.9	
Chlorite	1.0		1.0				2.2		2.1	
hornblende	79.0		38.2		17.9		14.6			
Epidote			6.3		3.1		3.5		4.2	
Sphene			.3							
Apatite	.1		.5		.5		.4		.3	
Ilmenite	1.0									
Magnetite			1.2		.6		1.8		1.0	
Rutile	.5				.1					
Calcite	.0		.0		.2		.0		.1	
Total	100.1		100.1		99.7		100.0		99.9	

Niggli values

	20		32		180		42		43	
si		93.2		133.0		168.0		228.8		296.0
ti		2.02		.96		1.63		1.36		.88
qz		24.84		.32		15.12		65.60		106.0
al		14.87		27.69		30.56		36.28		43.74
fm		57.80		39.34		33.77		24.50		13.58
c		22.82		24.80		22.45		23.42		20.18
alk		4.51		8.17		13.22		15.80		22.50
k		.186		.09		.196		.025		.067
mg		.65		.59		.55		.45		.37
c/fm		.395		.63		.66		.96		1.486

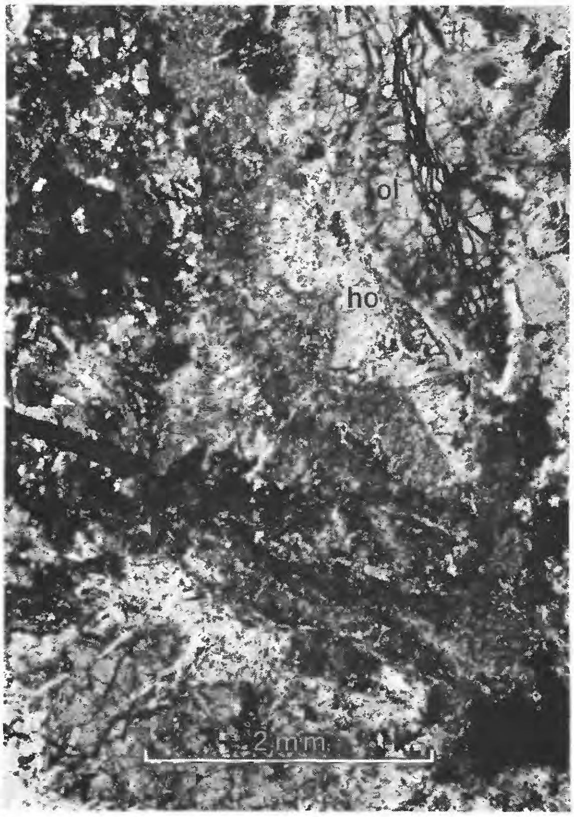


FIGURE 45.—Kelyphitic reaction rims around the olivine grains in peridotite from Harpers Bend (loc. 160, pl. 4). ol=olivine; ho=hornblende. Crossed nicols.

FIGURE 46.—Photomicrograph of a gneissic layer in metagabbro from Harpers Bend (loc. 164, pl. 4). qu=quartz; ep=epidote. Plane polarized light.

is An_{45}. The Na:Ca ratio in the new plagioclase and remaining epidote is higher than that of the mixture of zoisite and the early plagioclase, indicating that sodium was introduced during the recrystallization.

A dikelike body 20 m thick of metamorphosed diabase occurs in the metagabbro parallel to the cross jointing about 1 mile west of Harpers Bend (loc. 150). The mineral content of this metadiabase is similar to that of the metagabbro, but the texture of the two rocks is different. In the metadiabase, plagioclase (An_{47}) occurs in long laths, many of which measure 1 to 2 cm in length. Some of these plagioclase laths are broken, and some contain rows of other minerals, such as epidote, quartz, and hornblende (fig. 47). Hornblende also occurs as phenocrysts, but these phenocrysts are fewer and smaller. The groundmass consists of plagioclase, hornblende, epidote, biotite, ilmenite, and apatite.

The amphibolite just east of these basic rocks in Harpers Bend may represent a sheared and metamorphosed border zone of the sill. Another possibility is that it was either a basaltic lava flow or tuff. Toward the south, similar amphibolite rich in epidote minerals is intercalated with quartzite and muscovite-plagioclase (An_8) schist may contain tuffaceous material.

GABBRO AND NORITE

Much of the plutonic rock between Peck and Ahsahka is dark enough to be called gabbro. In most outcrops, hornblende is the main dark mineral and constitutes about 40 percent of the rock. It occurs as rounded grains or prisms or as clusters which are clearly visible against the white plagioclase. The largest bodies of gabbro are along the Clearwater River, between Ahsahka and Peck. The contacts between the gabbro and country rocks—dark biotite schist or hornblendite—are concordant. Locally within the bodies of gabbro the amount of dark minerals is less, and the rock has the composition of quartz diorite.

Microscopic study shows that plagioclase (An_{46}) occurs as subhedral long lath-shaped crystals that are oriented parallel to the lineation of the rock (fig. 48). Albite, carlsbad, and pericline twinning are common. Some of the hornblende occurs in large grains, many of which have the shape of augite. The hornblende in the centers of these grains is lighter in color and contains numerous small elongated quartz grains (fig. 49). In addition to these large spongelike grains there are abundant bluish-green needles of hornblende. Small needles are included in the plagioclase which, together

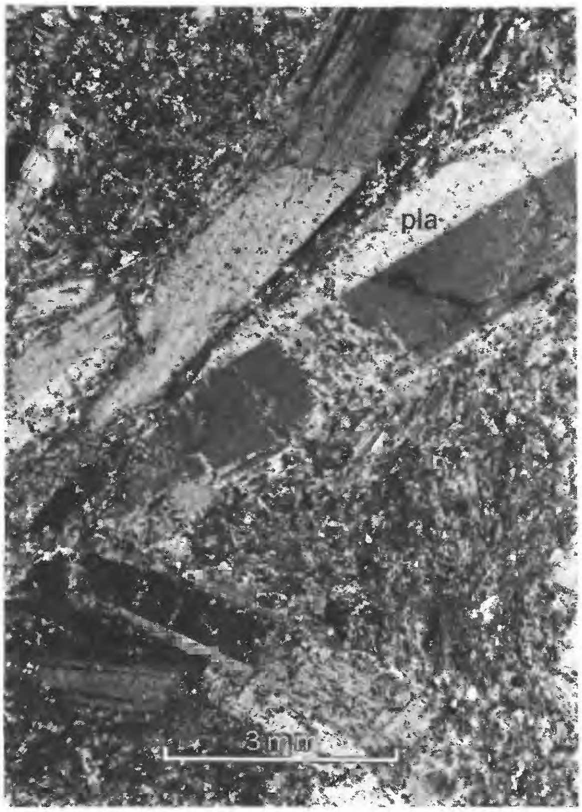


FIGURE 47.—Photomicrograph of sample from metamorphosed diabase dike containing long lath-shaped plagioclase phenocrysts (pla). The groundmass consists of plagioclase, hornblende, quartz, and epidote. Inclusions in the phenocrysts are hornblende and epidote. Just east of locality 150 along the Clearwater River (pl. 4). Crossed nicols.

FIGURE 48.—Photomicrograph of gabbro from the north side of the Clearwater River near Peck (loc. 32, pl. 4). The lath-shaped plagioclase crystals (pla) show complex twinnings and are well oriented parallel to the linear structure in the outcrop. Biotite is altered to chlorite (chl). ho=hornblende. Crossed nicols.

with small epidote grains, forms a texture that Sederholm called a weed texture. Larger grains of epidote, some quartz, magnetite, biotite, and chlorite are next to the large hornblende grains. In specimen 32 the indices of refraction of the blue-green hornblende are $\alpha=1.658\pm0.001$, $\beta=1.671\pm0.001$, and $\gamma=1.678\pm0.001$.

Chemical analyses, norms, minerals, and Niggli values for this gabbro (No. 32) are shown in table 15. The amount of Al_2O_3 is greater and that of MgO smaller than in normal gabbros, bringing this gabbro close to a diorite in composition.

The gabbro exposed just west of the hornblendite at locality 20 is exceptionally dark (loc. 509, pl. 4). Microscopic study shows that the dark constituents are hypersthene, augite, hornblende, and cummingtonite. Hypersthene with brownish-green serpentine along the cracks forms the centers of the dark grains. This hypersthene center is surrounded by fibrous and colorless cummingtonite which shows multiple twinning, bright interference colors, and $Z\wedge c=15°$. Cummingtonite in turn is rimmed by bluish-green hornblende (fig. 50). Iron oxides and sulfide are common in cummingtonite either as large grains or as tiny dustlike inclusions that in many grains are parallel to the former cleavage of the hypersthene.

In addition to the bluish-green hornblende that rims the pyroxene, brownish-green hornblende occurs as large grains that include small grains of altered hypersthene. The brownish-green hornblende is a primary mineral and, in accordance with Bowen's reaction series, it encloses the hypersthene that crystallized earlier from the gabbroic magma. Augite, like most of the hypersthene grains, is surrounded by a bluish hornblende as a result of a later reaction. Thus hypersthene, augite, and brownish hornblende are magmatic minerals, whereas cummingtonite and bluish-green hornblende are alteration products. Plagioclase (An_{45}) occurs in thick tabular crystals which show parallel orientation similar to that shown in figure 48.

The centers of some bluish-green hornblende crystals in this pyroxene gabbro (No. 509) contain abundant small grains of quartz similar to those in the hornblende in the gabbro (fig. 49). It is apparent that in the pyroxene gabbro at locality 509 this texture is due to the reaction: hypersthene→cummingtonite→hornblende, which completely altered some hypersthene grains into bluish-green hornblende. The excess of silica crystallized as quartz inclusions. A part of silica is freed also if diopside alters to hornblende. Probably all bluish-green hornblende with quartz inclusions in the gabbro at locality 32 also was originally pyroxene. Field relations, mineralogy, texture, and chemical composition of both gabbros are typical of those of normal intrusive gabbros.

A schistose gabbroic rock a mile north of Orofino is texturally quite different from the gabbros along the Clearwater River. It grades from a medium-grained foliated rock that contains large hornblendite inclusions to a medium- to fine-grained dioritic rock. Probably this gabbro was a sill-like body which was emplaced parallel to a fault zone and was strongly sheared. Many outcrops of fine-grained gabbro contain large scattered hornblende crystals, which may have been porphyroblasts in the original rock. The texture is granoblastic, and the large hornblende crystals are surrounded by smaller crystals as common in the granulated rocks. Hornblende and plagioclase are the major constituents. Ilmenite-magnetite, sphene, and apatite occur as accessories.

QUARTZ DIORITE

Quartz diorite is far more abundant than gabbro and the bodies are larger in size. Megascopically much of this rock resembles gabbro except for less dark constituents. The major constituents in quartz diorite are plagioclase (An_{30-40}), quartz, hornblende, and biotite. Dark constituents are clustered which gives an impression of a coarser grain size than it is. On slightly weathered surface plagioclase and quartz are white and contrast strongly against the black color of the hornblende and biotite. Texture varies; most bodies are gneissic but some show a hypidiomorphic texture. A good example of the hypidiomorphic variety is a quartz diorite east of Peck (loc. 33). However, only a part of this body has a hypidiomorphic texture similar to that of the intrusive gabbro at locality 32. In this hypidiomorphic part, the plagioclase forms tabular crystals that are well oriented, forming a linear structure parallel to the lineation in the country rocks just as the same mineral in the gabbro does. Most outcrops, however, show only a planar structure, resulting from a parallel orientation of biotite and hornblende, which gives the rock a gneissic appearance. The amount of biotite, quartz, and epidote is generally larger in this gneissic quartz diorite than in the quartz diorite with hypidiomorphic texture.

Thin sections show that the plagioclase in the gneissic quartz diorite occurs as rounded large grains between which there are small grains of quartz, biotite, and hornblende, (figs. 51*A* and *B*). Abundant epidote

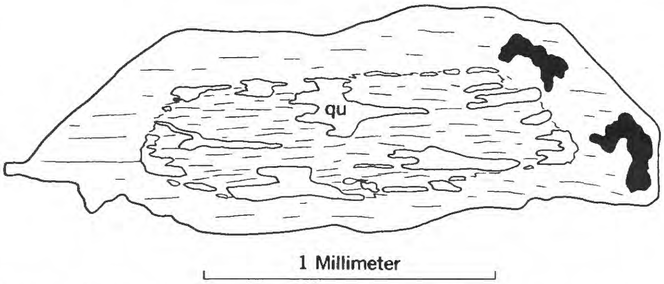


FIGURE 49.—Camera lucida drawing of a hornblende grain with elongated quartz (qu) inclusions in its center. Black is magnetite. From gabbro in locality 32 along the Clearwater River, near Peck (pl. 4).

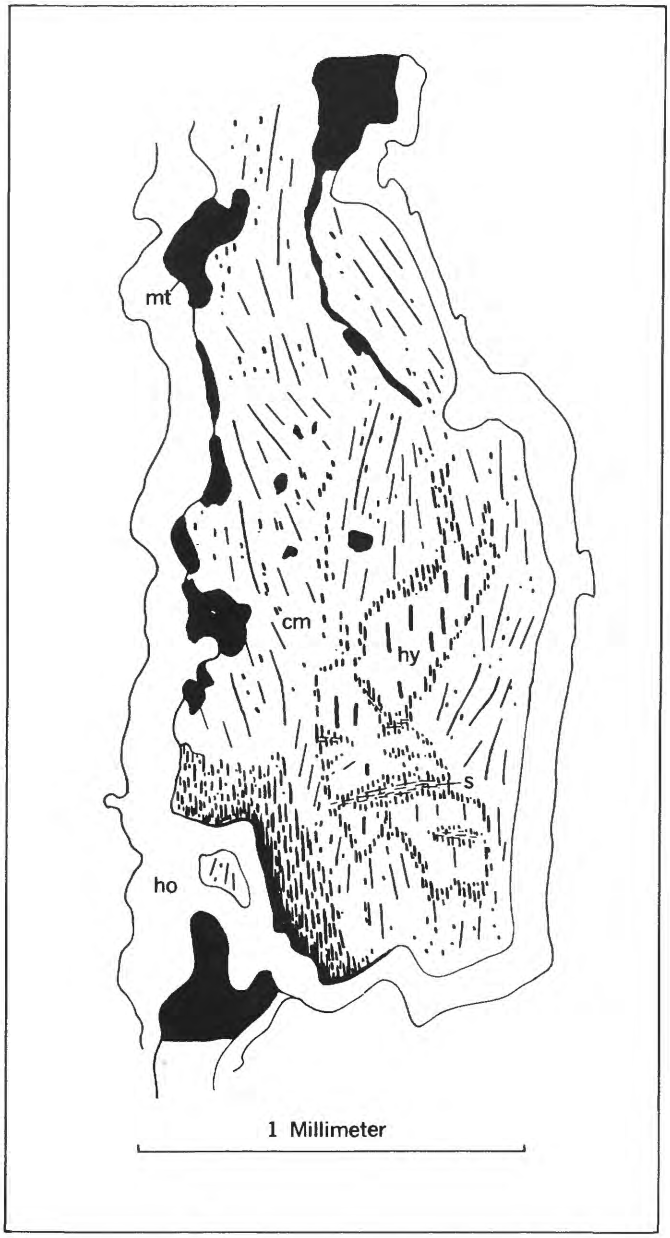
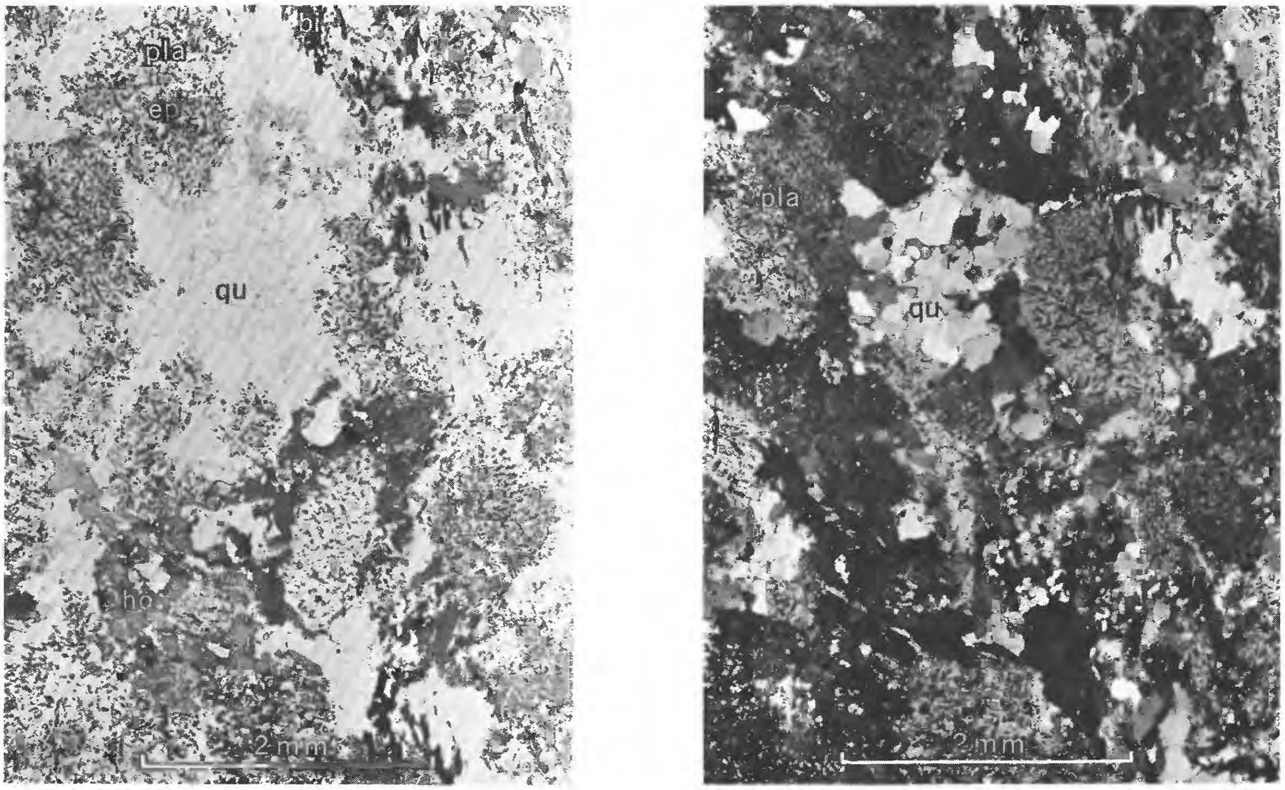


FIGURE 50.—Camera lucida drawing of hypersthene (hy) partly altered to cummingtonite (cm) and hornblende (ho). From norite at locality 509 along the Clearwater River (pl. 4). mt=magnetite; s=serpentine.



A. Plane polarized light. Plagioclase (pla) occurs in large rounded grains including numerous small epidote crystals (ep). Hornblende (ho) and biotite (bi) occur in clusters.

B. Crossed nicols. The abundant quartz (qu) between the plagioclase (pla) grains occurs in small grains.

FIGURE 51.—PHOTOMICROGRAPHS OF GNEISSIC QUARTZ DIORITE FROM THE SIDE OF THE CLEARWATER RIVER NEAR PECK (LOC. 33)

occurs as small prismatic grains in this plagioclase (An_{26-32}). Plagioclase in the quartz diorite with hypidiomorphic texture is An_{35}, thus richer in anorthite than the same mineral in the gneissic diorite. Hornblende in both varieties is bluish green to green, and abundant quartz is included in the centers of larger grains. Its indices of refraction are $\alpha'=1.658\pm0.001$, $\gamma'=1.679\pm0.001$. Biotite is light brown to brown with $\gamma=1.630\pm0.001$. Large flakes contain inclusions of ilmenite and small needles of rutile. Chlorite, magnetite, and apatite are the additional minor constituents. Comparison of the chemical analyses of the hypidiomorphic quartz diorite (No. 180) with that of the gabbro (No. 32, table 15) shows that the quartz diorite is richer in silicon and alkalies and poorer in iron, magnesium, and calcium than the gabbro. Mineralogically, this is reflected in the larger amount of more sodic plagioclase, quartz, and biotite and less hornblende in the quartz diorite.

The field relations show that the quartz diorite with hypidiomorphic texture occurs next to the gabbro and is a differentiate of the same magma. The linear arrangement of the lath-shaped plagioclase crystals may be due to a flowage during the emplacement. The planar structure in the gneissic quartz diorite is parallel to the foliation of the bordering schist and gneiss and probably is a relict structure. The large amount of biotite and quartz in this gneissic quartz diorite, its texture, and the parallelism of the planar structure suggest that this quartz diorite came to its present place by a metasomatic replacement of metasedimentary country rocks rather than by injection.

Thin sections of the gneissic quartz diorite show abundant granulated quartz between plagioclase grains (figs. 51*A* and *B*). A similar texture was found in some cross-cutting intrusive tonalite dikes about 30 miles northeast of Orofino along the Clearwater River. There, the granulation of quartz is clearly due to stresses operating during or after the solidification of the tonalite magma. The occurrence of the granulated quartz in the dikes shows that this texture alone cannot be used as a proof of a sedimentary origin of the quartz in the quartz diorite near Orofino as was assumed by Johnson (1947, p. 494, 504). This same texture also led him to believe that the older metasedimentary rocks in the Orofino region were fairly pure quartzites and that material for all biotite and most minerals other than quartz was introduced from the Idaho batholith.

The quartz diorite near Greer and that south of Peck are much alike and may belong to the same body. This quartz diorite, most of which is lighter in color than the quartz diorite between Ahsahka and Peck, also is less gneissic. Darker parts near Greer are gabbroic in composition; for example, the outcrops along the road from the bridge at Greer for half a mile north are dark enough to be called gabbro. A very coarse grained quartz diorite, in which hornblende is the only dark mineral, is found north of this gabbro.

Thin sections of the quartz diorite from near Greer show abundant small grains of quartz between the larger plagioclase (An$_{32}$) grains (fig. 52). Individual grains of epidote occur with the dark constituents, hornblende and biotite. The hornblende shows pleochroism: Z=blue-green, Y=green, and X=light yellowish-green, $Z\wedge c=22°$, and the indices of refraction $\alpha=1.658\pm0.001$, $\beta=1.671\pm0.001$, and $\gamma=1.679\pm0.001$. These properties are similar to those of the hornblende in gabbro (No. 32).

The quartz diorite south of Freeman Creek and that exposed on the plateau west of Bruces Eddy resemble the quartz diorite near Greer in texture and mineralogy. Under the microscope the banded gneissic quartz diorite between Bruces Eddy and Ahsahka shows a similar texture. However, this banded quartz diorite contains tonalitic layers and remnants of fine-grained biotite gneiss and biotite-hornblende gneiss. It grades over to the metasedimentary rocks in the west and includes a large remnant of the same rock series to the east.

TONALITE

Some of the coarse-grained gneissic rocks contain less hornblende and more quartz than the quartz diorite; plagioclase in them is more sodic. These rocks are called tonalite. In addition to the coarse-grained variety, a fine-grained gray slightly gneissic tonalite occurs opposite the mouth of Canyon Creek.

The largest body of tonalite is well exposed along the North Fork of the Clearwater River northeast of Ahsahka, along the north side of the South Fork of the Clearwater River east of Ahsahka, and along a road and railroad west of Ahsahka. East of Ahsahka it contains a considerable amount of hornblende and epidote. Gabbroic and dioritic lenticles are common there, especially toward the top of the ridge where the tonalite grades into a diorite.

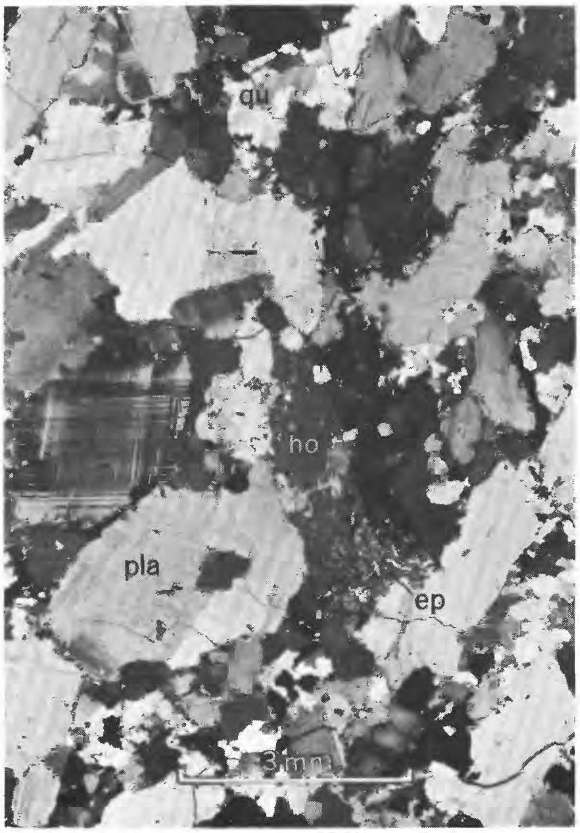
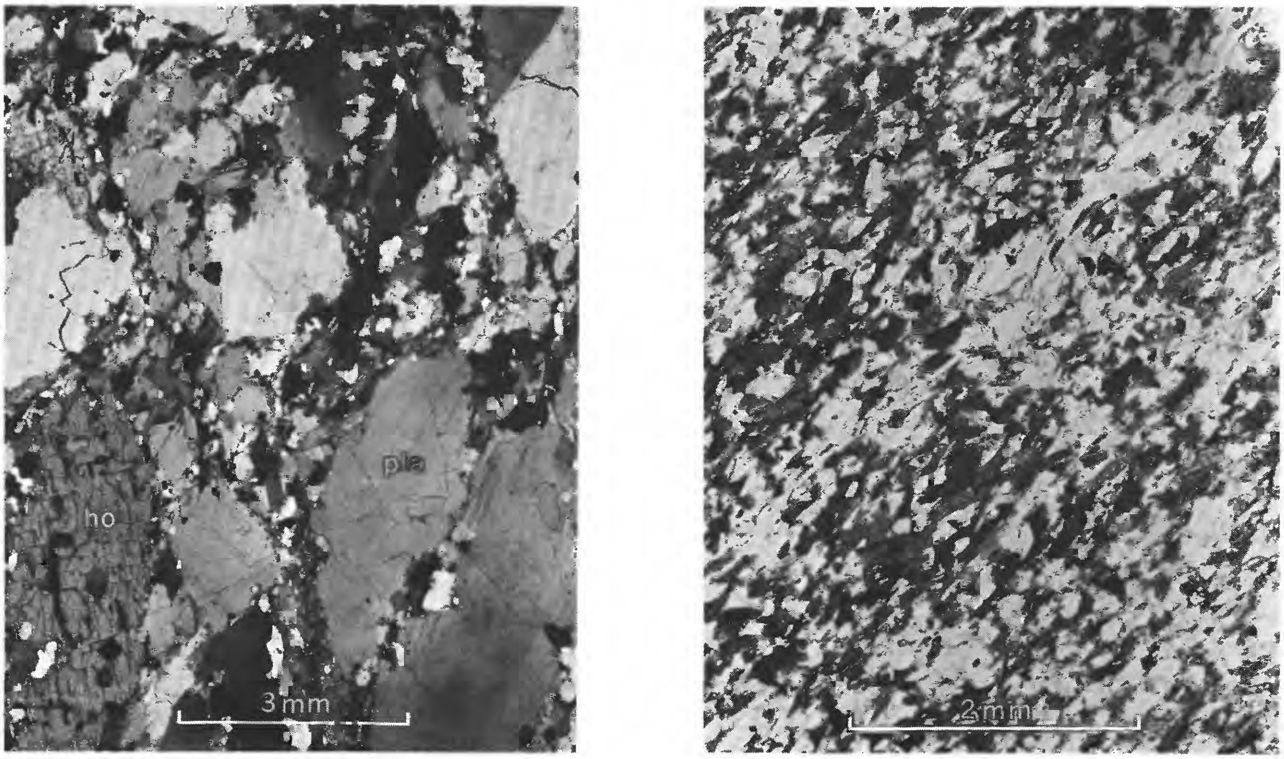


FIGURE 52.—Photomicrograph of quartz diorite from near Greer (loc. 42). The plagioclase (pla) grains are rounded or subhedral. Quartz (qu), hornblende (ho), epidote (ep), and chlorite occur in smaller grains between the plagioclase grains. Crossed nicols.

Thin-section study of the specimens taken along the road leading from Orofino to Ahsahka shows that the dark minerals are hornblende and biotite in nearly equal amounts or only biotite. Most of the biotite occurs in small laths, but a few larger flakes are seen with the hornblende. Hornblende grains are rounded, with irregular borders, and include a few round small quartz grains. Most of biotite and hornblende are in thin layers, which lends a strongly gneissic appearance to the rock. Large grains of epidote—many of them with allanite in their centers—occur with the dark minerals. Some epidote grains are surrounded by myrmekite-like intergrowth of plagioclase and epidote. Many plagioclase (An$_{30}$) grains are round or elongated with rounded corners and are larger than the other mineral grains; small grains or plagioclase (An$_{26-30}$) occur with quartz between the large grains. Small grains of sphene, leucoxene, and apatite are the accessories.

To the west this hornblende tonalite grades to a lighter colored rock in which hornblende is less abundant or is completely lacking. The plagioclase and hornblende in the light-colored tonalite occur as large oval grains embedded in a fine-grained gneissic rock consisting of quartz, plagioclase, and biotite (fig. 53*A*). Feldspathized inclusions of metasedimentary rocks are common in the light-colored tonalite and they seem to have preserved their original position. These in-



A. Photomicrograph of light-colored gneissic tonalite. Large oval plagioclase (pla) and hornblende (ho) grains are embedded in fine-grainned rock consisting of quartz, plagioclase, and biotite (loc. 25, pl. 4). Crossed nicols.

B. Photomicrograph of country rock just south of gneissic tonalite No. 25. Dark minerals are hornblende and biotite. Light-colored constituents, quartz and plagioclase, occur in small rounded grains similar to those in fine-grained part of the gneissic tonalite. Plane polarized light.

FIGURE 53.—LIGHT-COLORED GNEISSIC TONALITE NEAR AHSAHKA, AND COUNTRY ROCK JUST SOUTH OF THE TONALITE

clusions consist of fine-grained biotite gneiss or biotite-hornblende gneiss (fig. 53*B*) texturally and mineralogically similar to gneiss layers in the country rocks. In the tonalite the matrix between the large plagioclase grains also has a composition and texture similar to those of the gneisses. This similarity in texture and composition suggests that the tonalite was originally a fine-grained gneiss; the large plagioclase and hornblende grains were developed later and give the tonalite the appearance and composition of a plutonic rock. In many localities near contacts with biotite gneiss or biotite-hornblende gneiss the quartz diorite grades to tonalite. The small bodies of tonalite that are common everywhere in the metasedimentary rocks are much like these tonalitic contact zones. The structure of the metasedimentary rock is preserved, the bedding planes continuing undisturbed from the country rock to the tonalite at the ends of the bodies. The texture of the tonalite is similar to that shown in figure 53*A*. Mineralogically these tonalite bodies differ from the surrounding gneiss only by containing more plagioclase (An_{30}) and less quartz.

Part of the intrusive rock near Greer contains little or no hornblende, is finer grained and lighter in color than the diorite, and its texture is similar to that of the surrounding diorite (fig. 52). The chemical and mineralogical composition (No. 43, table 15) shows that this rock is tonalite.

The appearance of the fine-grained tonalite opposite the mouth of Canyon Creek is similar to that of the fine-grained gray intrusive tonalite exposed along the North Fork of the Clearwater River between Dent and Big Island (No. 190, table 10). Its occurrence as layers and dikelike bodies in the fine-grained garnet-biotite gneiss suggests an igneous origin. However, it has a layering parallel to the bedding of the metasedimentary rocks and it contains minerals such as garnet and epidote that are rare in the intrusive tonalite but are common constituent minerals in the metasedimentary biotite gneiss and biotite-hornblende gneiss. Thin sections show that this tonalite is more nearly equigranular than the coarse-grained gneissic tonalite described earlier. The grain size of quartz and plagioclase is almost the same. Plagioclase (An_{26}) grains are round and contain small square patches of antiperthitic potassium feldspar. Quartz grains are between the plagioclase grains and show undulatory extinction as is common in metamorphic rocks in this area. Many small biotite flakes are included in plagioclase and quartz. Small irregularly shaped grains of garnet and some muscovite, chlorite, and apatite are the additional constituents. The relative amounts of the constituent minerals in this fine-grained tonalite suggest that its chemical composition is close to that of the fine-grained intrusive tonalite No. 190. Texturally, however, these two tonalite bodies are quite different. The intrusive tonalite shows a hypidiomorphic texture with zoned subhedral plagioclase crystals, whereas the texture of the tonalite opposite the mouth of Canyon Creek is similar to that of the metamorphic rocks. It is possible that in this locality feldspathized gneiss was partially melted and squeezed as dikelike bodies into the country rocks.

PEGMATITE

GRANITE PEGMATITE

Pegmatite veins and dikes abound in the area. Many dikes are as much as 10 m wide and either parallel or cut the bedding of the metasedimentary rocks.

In some pegmatites, both feldspars, plagioclase and orthoclase, occur together, but many dikes in the quartz diorite contain only plagioclase. Muscovite and biotite are the common additional constituents. Garnet, magnetite, and sphene were found in several localities.

HORBLENDE PEGMATITE

Many outcrops of quartz diorite on the west side of the South Fork of the Clearwater River south of the bridge at Greer and those along the lower canyon wall on the east side of the river contain pipelike masses of hornblende pegmatite. These pegmatites are round or triangular in cross sections that are from 50 to 100 cm in diameter. The hornblende occurs as long crystals, some as much as 30 cm in length, that are perpendicular to the walls and thus form a radial pattern (fig. 54).

Thin-section study confirms that hornblende, plagioclase, and quartz are the main constituents of these pegmatites. Quartz occurs as small grains between large round plagioclase grains and along the fractures in the plagioclase. Near Greer tiny epidote prisms are included in the plagioclase and larger grains of epidote occur next to the hornblende, whereas near Peck about 50 percent of the plagioclase is altered to eipdote. Chlorite is common in both localities and occurs next to the hornblende. The hornblende is a bluish-green variety similar to that in the quartz diorite.

The mode of occurrence of the hornblende pegmatites suggest that they are segregations of the residual liquids of the quartz diorite.

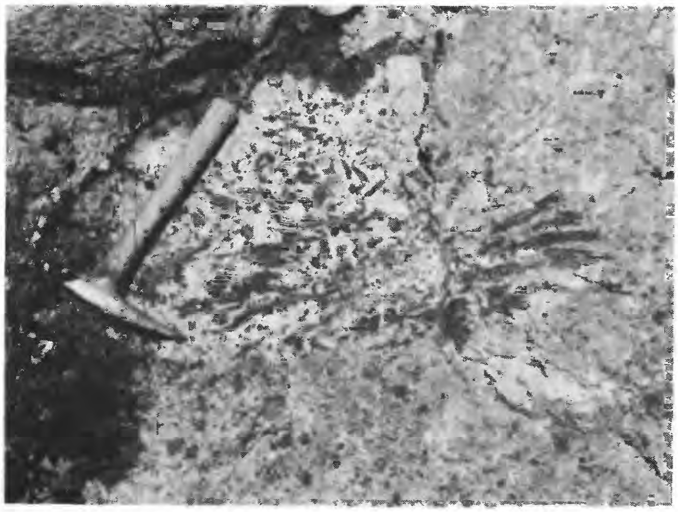


FIGURE 54.—A triangular cross section of a pipelike hornblende pegmatite in the quartz diorite near Greer. The dark radial crystals are hornblende. The hammer showing the scale is 32 cm long.

ALBITE-EPIDOTE DIKES

A fine-grained granular light-green rock fills the joints in gabbro exposed in a road cut along the Clearwater River about 2 miles west of Ahsahka. Study under the microscope shows that this rock consists of about 55 percent albite (An_5), 30 percent epidote, and 15 percent quartz. Apatite and sphene occur as accessories. The texture is granoblastic. Quartz occurs in small rounded grains; the albite grains are a little larger and are angular, with slightly fringed boundaries. Epidote has $\beta=1.740\pm0.005$, which indicates about 10 percent iron content. This albite-epidote rock fills several cross joints, longitudinal joints, and small cracks that branch off from the joints. In the gabbroic country rock the plagioclase contains inclusions of epidote, and small grains of quartz occur between the plagioclase grains. The composition of these dikes suggests that they are not intrusive but that the material in them sweated out from the gabbro during the opening of the joint system.

Joints that transect the quartz diorite nearby and contact zones between several large inclusions and the enclosing quartz diorite are filled by light-colored pegmatitic materials in which quartz and plagioclase are the major constituents. The plagioclase crystals are large and oriented with their long dimensions perpendicular to the walls of these small dikelike bodies. The walls are not smooth surfaces but irregular zigzag lines between the grains of the coarse-grained rocks. Small grains of epidote are included in the plagioclase within the quartz diorite and also in that within the dikelets; large recrystallized grains of epidote rim the walls of the dikelets. Quartz occurs in groups of small strained grains with fringed borders. A few large

Figure 55.—Photomicrograph of sample from the anorthosite dike on a road cut 5.2 miles south of Orofino. Labradorite grains are rounded and show complex twinnings. A few grains are surrounded by a narrow more albitic rim (a). Crossed nicols.

biotite flakes have grown across the dikelets. In a dikelet that occurs between coarse-grained quartz diorite and a fine-grained inclusion some large flakes of biotite extend from the quartz diorite across the dikelet and penetrate 5 to 7 mm into the inclusion. The inclusion consists of fine-grained quartz-dioritic rock. The planar structure of the inclusion forms an angle of about 80° with the planar structure of the host rock. The dikelet between the host rock and the inclusion cuts across the planar structure of the host rock but parallels the same structure in the inclusion. The larger biotite flakes that grow across the dikelet are parallel to the planar structure in the host rock but penetrate the inclusion across its planar structure.

The mineralogy and texture of the dikelets suggest that material in them was derived from the wallrocks and that some iron, magnesium, aluminum, and potassium migrated from the host rock across the dikelets into the inclusions. The orientation of the large biotite flakes parallel to the platy structure of the quartz diorite and across the dikelets supports this conclusion.

ANORTHOSITE DIKE

A coarse-grained anorthosite dike 2 m wide was exposed for a distance of 5 m at a road cut 5.2 miles south of Orofino when I visited the area in 1946. The dike resembled a common discordant pegmatite but consisted entirely of bluish-white plagioclase. Thin-section study showed that this plagioclase is labradorite (An_{52}) and that a few small grains of epidote and apatite occur as accessories. Labradorite grains are anhedral and strongly twinned (fig. 55). Some grains show narrow rims of more sodic plagioclase.

CHEMICAL COMPOSITION

The plutonic rocks in the Orofino district, like those in the Headquarters quadrangle, are poor in potassium and rich in calcium and sodium. In these respects their composition is close to that of the gneissic layers in the metasedimentary rocks. The ionic percentages of major elements, calculated from the weight percentages of the chemical analyses (tables 12 and 15), are plotted against the ionic percentage of silicon for metasedimentary and plutonic rocks (fig. 56). The analyses of quartz-diorite and tonalite from the border zone of the Idaho batholith and from the satellitic intrusions were plotted in the same graph for comparison (Nos. 292, 322, 190, and 379).

The tonalitic members of the plutonic series are chemically close to the hornblende-bearing gneiss (Nos. 106 and 145 in fig. 56). However, the content of aluminum, calcium, and sodium is somewhat higher, and that of silicon, iron, magnesium, and potassium lower. Mineralogically, this difference is shown by a higher content of plagioclase and less biotite and quartz in the plutonic rocks. The most silicic member, a silicic tonalite near Greer (No. 43 in fig. 56) contains 5 percent sodium oxide and only 0.5 percent potassium oxide, all of which is in muscovite and biotite. The Niggli values in table 15 and figure 57 show that the alkali ratio *k* for all plutonic rocks is smaller than 0.2, which is much lower than *k* in the normal granite series. However, the *k* as well as *mg* value is close to that in the normal quartz diorite and plagioclase granite.

The Niggli values *al*, *fm*, *c*, and *alk* were plotted against the *si* values (fig. 58). For comparison, the Niggli values of the plutonic rocks in the Headquarters quadrangle were plotted on the same graph. The graph shows that the silicic members of the plutonic rock series of the Orofino district are exceptionally high in aluminum and low in ferric constituents. Because the amount of calcium is higher than that in the normal granite, the plagioclase is andesine even in the most silicic members of the series.

The weight and molecular norms for the feldspars were calculated to 100 percent and plotted in a ternary diagram (fig. 59). The points for molecular norms (heavy symbols) are slightly toward the Ab corner

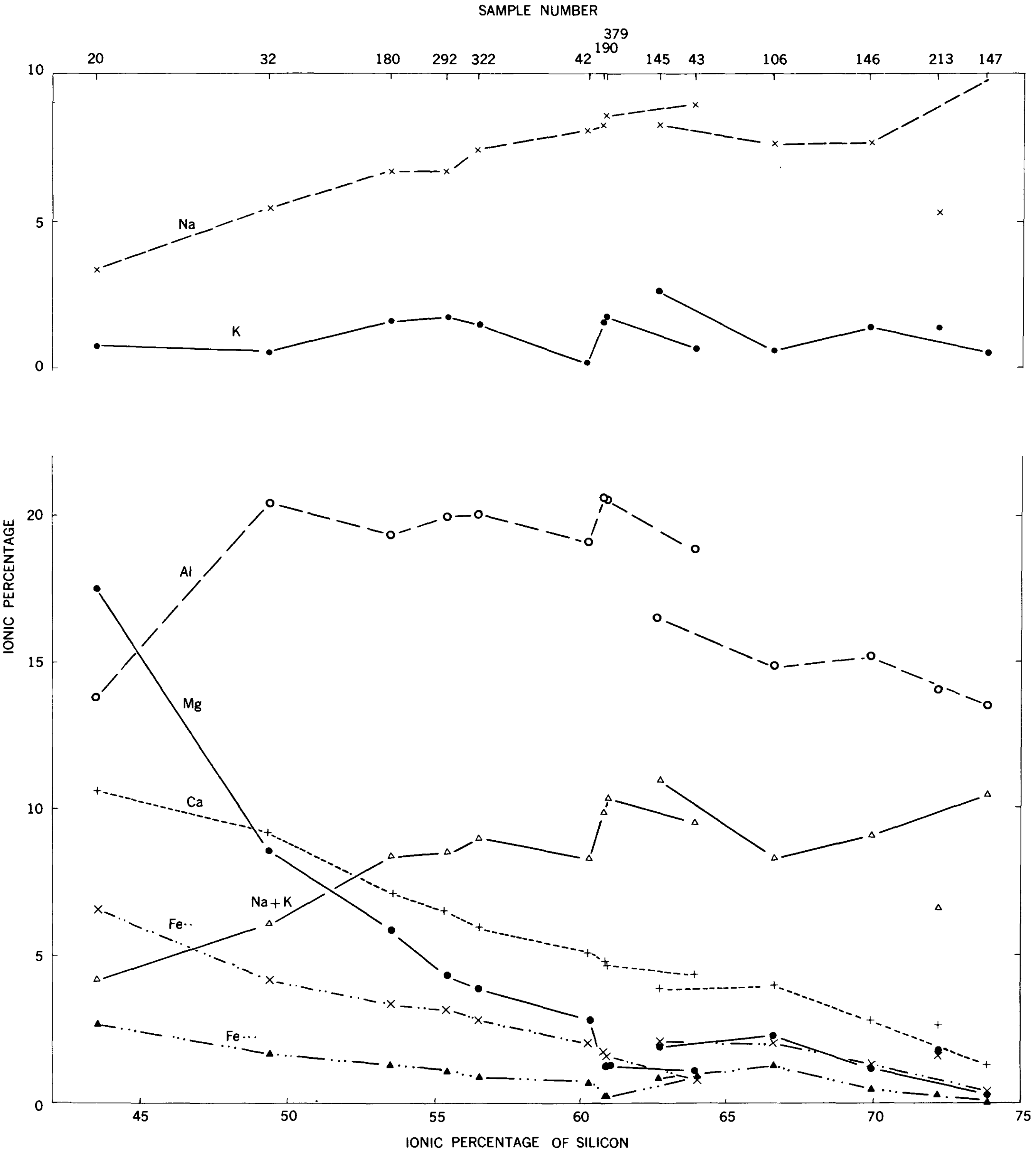


FIGURE 56.—Graph showing the amount of various elements in ionic percentages in metasedimentary and plutonic rocks of the Orofino district. Samples 145, 106, 146, and 147 are of metasedimentary rocks, and samples 20, 32, 180, 42, and 43 are of plutonic rocks. Samples of plutonic rocks in the Headquarters quadrangle and vicinity are added for comparison (Nos. 292, 322, 190, 379, and 213).

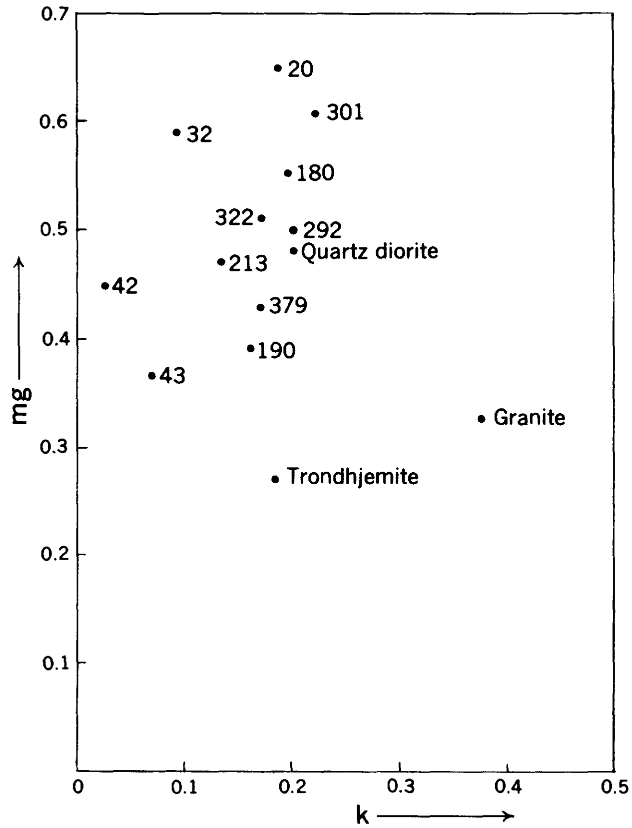


FIGURE 57.—Niggli *k-mg* diagram for the plutonic rocks of the Orofino district. For comparison values are added for rocks in the Headquarters quadrangle and vicinity (Nos. 301, 322, 292, 213, 379, and 190) and those for Niggli's (1923) normal quartz diorite, granite, and trondhjemite.

from the corresponding points for weight norms (light symbols). The normative feldspar, being poor in orthoclase, is close to the Ab–An line in the ternary diagram (fig. 59). In the quartz diorite (Nos. 180 and 292) the normative amount of Ab is as large as that of An, whereas the tonalite (Nos. 43, 190, and 379) contains more albite, and the gabbro (No. 32) more anorthite. This composition differs strikingly from that of the normative feldspar in the granite series. For comparison the normative feldspar in the granite at Bungalow, about 36 miles east of Orofino, was plotted on figure 59 (No. 758 from table 31). The point for this feldspar lies close to the Or–Ab line in the graph.

The relations between the weight and molecular norms for feldspar, quartz, and femic constituents is shown in figure 60. The points for molecular norms (heavy symbols) are slightly toward the feldspar corner from the corresponding points for weight norms (light symbols). For this diagram the molecular norms for enstatite and ferrosilite were recalculated as forsterite and fayalite so that the mafic members of the rock series could be plotted inside the same triangle. The excess quartz was added to the molecular norm of quartz. The total amount of normative feldspar i quartz diorite and tonalite is nearly equal. The amoun of quartz increases and that of femic constituents decreases regularly from gabbro to tonalite. In figure 6 the points for the intrusive rock series form a smoot curve. The amounts of normative quartz, feldspar and femic constituents of the metasedimentary rock were plotted for comparison on the same graph. Al points for the metasedimentary rocks except the on for the hornblende gneiss (No. 145) fall considerabl farther from the feldpsar corner than the points fo the plutonic rocks. The point for the granite at Bungalow (No. 758, table 31) is included for comparison This granite is considerably poorer in feldspars an richer in quartz than is the silicic tonalite. Thus th rocks of the tonalitic suite differ from all the othe rocks in the region, mainly because of their highe content of plagioclase (andesine).

STRUCTURE

Most of the plutonic rocks near Orofino show a well-developed planar or linear structure. Inclusions of th wallrocks are common.

PLANAR STRUCTURES

Planar structure in the plutonic rocks is parallel t the bedding of the metasedimentary rocks and in man places there is a distinct layering parallel to the sam plane. The planar structure is due to parallel arrangement of biotite and hornblende. In many localitie these minerals tend to form continuous thin layers which gives the rock a strongly gneissic appearance In the layered rock the amount of the dark constituents varies from one layer to the next. Th layers range in thickness from a few millimeters t 30 or 50 cm. The layers are less distinct than the are in the gneisses of clearly metasedimentary origin; in many localities, however, they follow the same pattern. Between Bruces Eddy and Ahsahka the layering is accentuated by occurrence of white plagioclase-quartz veinlets and dark hornblende-rich schlieren. I some outcrops light-colored tonalitic layers alternat with dioritic layers; in others the layering is mainl a variation in the grain size. In many tonalite bodie a distinct layering is caused by a variation in the number of the large oval plagioclase grains embedded in a finer grained mixture of quartz, plagioclase, and biotite.

LINEAR STRUCTURE

Linear structure commonly is caused by a parallel arrangement of hornblende and biotite, more rarely by a parallel arrangement of plagioclase. The latter was found only in a gabbro and quartz diorite northeast of

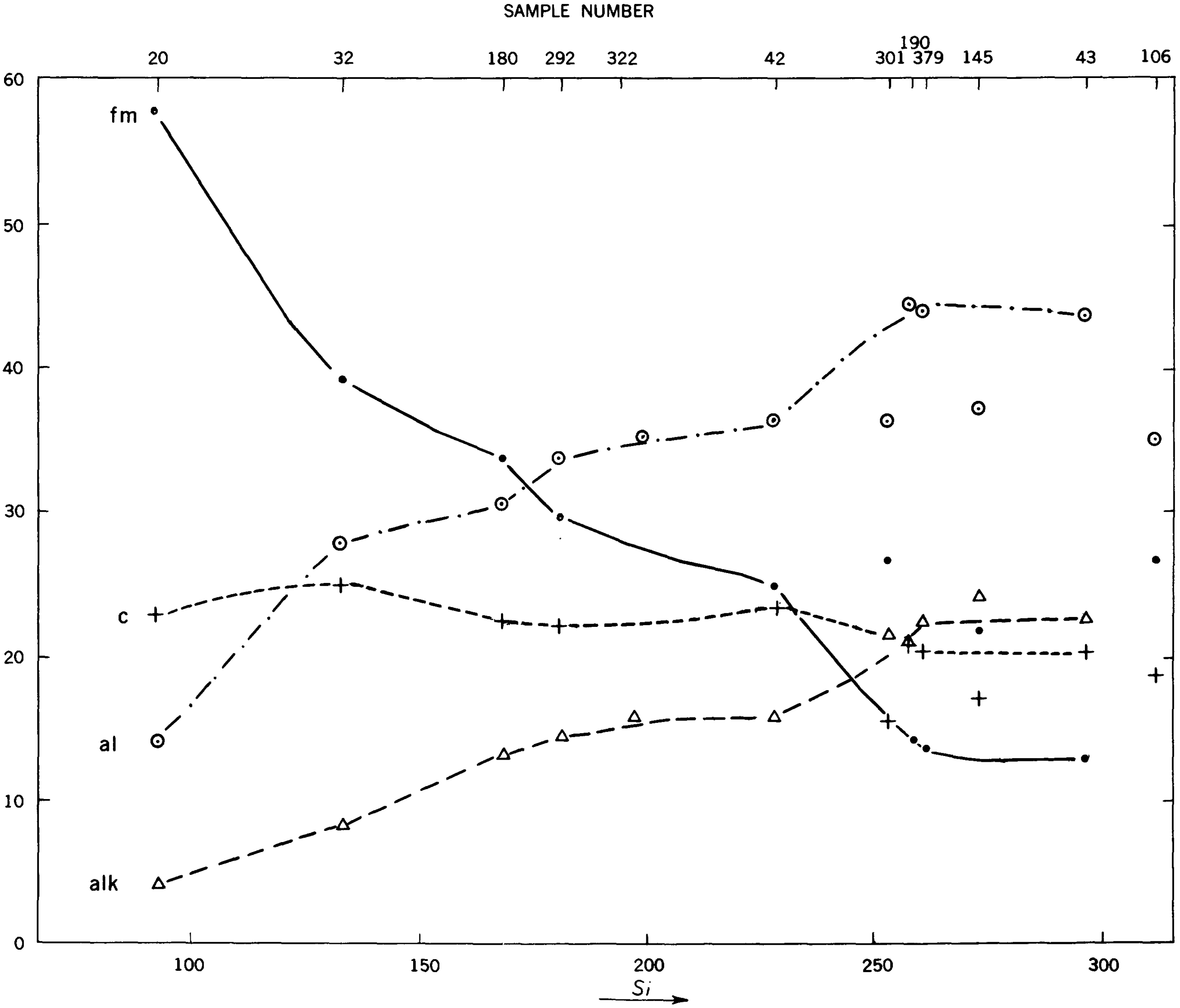


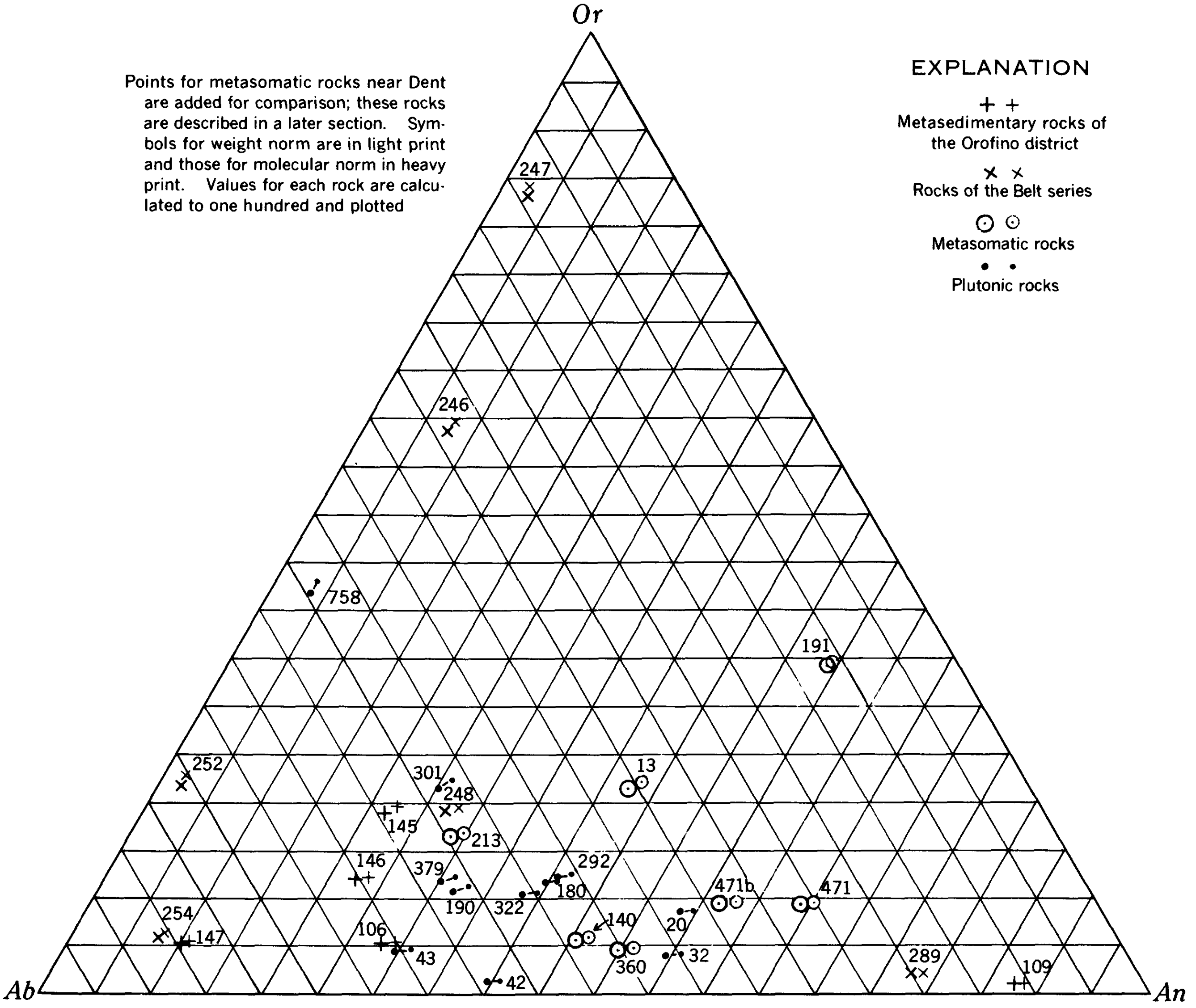
FIGURE 58.—Graph showing Niggli values for the plutonic rocks (Nos. 20, 32, 180, 42, and 43) and hornblende gneisses (Nos. 145 and 106) of the Orofino district. For comparison values are added for plutonic rocks from the Headquarters quadrangle and vicinity. (Nos. 292, 322, 301, 190, and 379).

Peck. Linear structure is always parallel to the lineation of the adjacent metasedimentary rocks.

INCLUSIONS

Remnants of metasedimentary rocks are abundant in the plutonic rocks. These inclusions are long thin sheets that occur parallel to the planar structure. They range from a few millimeters to several tens of meters in thickness. In the inclusions of schist or gneiss the small grains of quartz and plagioclase are the main constituents, but a few scattered large grains of plagioclase similar to those in the plutonic rock also occur. The contacts between the inclusions and the plutonic rocks are usually gradational over a few centimeters, but in some outcrops a biotite seam occurs along the contact.

Thin-section study of the gneissic tonalite shows numerous small spots of fine-grained biotite-plagioclase gneiss that probably are remnants of the older rocks, just as the larger inclusions are. Abundant epidote in some layers may be an indication of digested calcareous material. Remnants of a fairly pure quartzite were found in many localities in the quartz diorite west of Greer. In locality 405, about 3 miles north of Greer, a layer of a strongly sheared fine-grained biotite schist and biotite quartzite is included in the quartz diorite. Part of this schist is very rich in biotite. Large plagioclase and hornblende grains are scattered in some



Points for metasomatic rocks near Dent are added for comparison; these rocks are described in a later section. Symbols for weight norm are in light print and those for molecular norm in heavy print. Values for each rock are calculated to one hundred and plotted

FIGURE 59.—Ternary diagram showing the composition of the normative feldspar in the metasedimentary and plutonic rocks in the Orofino district, Headquarters quadrangle and vicinity. Sample numbers refer to tables 1, 2, 7, 10, 12, 14, 15, 18, 20, 22, and 31.

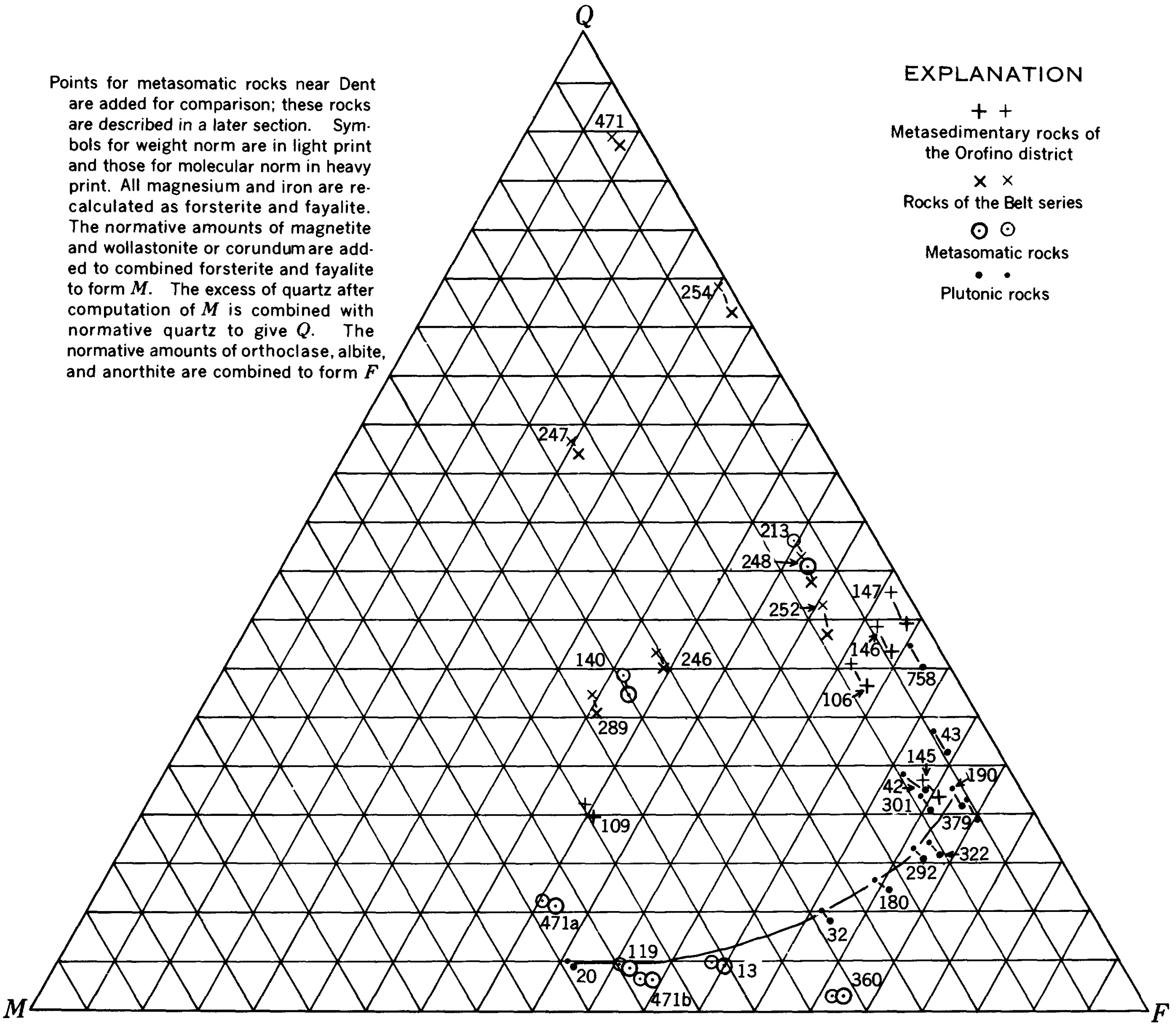
layers; in other layers biotite and hornblende are altered to chlorite.

The quartz diorite east of Greer contains large inclusions of a dark fine-grained diorite that megascopically appears much like the dark granoblastic diorite sill 1 mile north of Orofino.

CONTACTS BETWEEN THE PLUTONIC AND MESTASEDIMENTARY ROCKS

Contacts of the plutonic rocks with the metasedimentary rocks are either parallel to the bedding or interfinger across it. The contact zone parallel to the bedding is usually gradational over a short distance owing to the decrease in the amount of large plagioclase grains toward the metasedimentary rock. In some localities the intrusive rock is bordered by a fault and the country rock near the fault is strongly sheared (for instance loc. 509). The gradational contact zone is much wider where the contact crosses the bedding. Near the contacts of this type undisturbed inclusions of the metasedimentary rocks are numerous. Along the North Fork of the Clearwater River northeast of Ahsahka, small inclusions of finer grained rock rich in biotite mark the continuations of the biotite gneiss layers in the tonalite. Abundant epidote has crystallized in the contact zone between diopside gneiss and quartz diorite west of Bruces Eddy (loc. 1162). In this contact zone, which is more than 100 m wide, diopside is partly altered to hornblende and the rock is as coarse grained as the quartz diorite but contains abundant quartz and some diopside.

An eastward extension of the marble layer in the


Points for metasomatic rocks near Dent are added for comparison; these rocks are described in a later section. Symbols for weight norm are in light print and those for molecular norm in heavy print. All magnesium and iron are recalculated as forsterite and fayalite. The normative amounts of magnetite and wollastonite or corundum are added to combined forsterite and fayalite to form *M*. The excess of quartz after computation of *M* is combined with normative quartz to give *Q*. The normative amounts of orthoclase, albite, and anorthite are combined to form *F*

EXPLANATION

+ +
Metasedimentary rocks of the Orofino district

× ×
Rocks of the Belt series

⊙ ⊙
Metasomatic rocks

• •
Plutonic rocks

FIGURE 60.—Ternary diagram showing the amounts of normative quartz (Q), femic constituents (M), and feldspars (F) in the metasedimentary and plutonic rocks in the Orofino district, Headquarters quadrangle and vicinity. Sample numbers refer to tables as in figure 59.

quartz diorite east of Bruces Eddy is replaced by a dark fine-grained hornblende rock which grades to a coarse-grained quartz diorite. In some contact zones, as that 1 mile north of Orofino (loc. 533), skarn occurs next to the marble. In this locality, marble is bordered by gneissic granoblastic quartz diorite which occurs as a sill-like body along a fault. The eastern end of the marble layer is exposed between two road cuts at locality 533. Toward the east, within 2 to 4 m parallel to the bedding, the medium-grained pure-white marble grades to a pistachio-green diopside-epidote-scapolite skarn. About 5 meters farther east this skarn grades to a hornblende rock which in turn is partly feldspathized in the easternmost road cut. The feldspathized parts have quartz dioritic composition. The skarn, amphibole rock, and schistose quartz dioritic rock represent various stages in the dioritization of this limestone.

A contact zone between the metasedimentary rocks and gneissic tonalite is exposed along the upper part of the canyon wall northwest of Bruces Eddy. Limestone is exposed at the river, but its place on the upper part of the wall is occupied by schistose amphibolite. The augen gneiss and biotite gneiss that are exposed along the river grade to a tonalite on the higher elevation. On a small scale this relation contradicts the often expressed theory that the granitization would increase downward because of possible intrusion underneath. In the Orofino area the remnants of the metasedimentary rocks seem rather to be parts of large inclusions in the plutonic rocks. In the vertical plane perpendicular to the trend of the fold axis, many of these inclusions pinch out upward as well as downward. Thus the tonalitization of the metasedimentary rocks is not a simple contact phenomenon but rather a metasomatic process controlled by the distance from the batholith and factors other than the distance. Before these other factors are explained, the metasomatic changes of metasedimentary rocks will be discussed in detail.

METASOMATIC CHANGES IN THE METASEDIMENTARY ROCKS NEAR OROFINO

In places the metasedimentary rocks are transformed into coarse-grained rocks whose chemical composition differs somewhat from that of the ordinary gneiss. These coarse-grained parts of the strata contain large round grains of plagioclase or of plagioclase and hornblende embedded in medium- to fine-grained gneiss. In places the number of these large grains has increased to such an extent as to give an appearance of a plutonic rock to these parts of the strata. The structures—bedding and foliation—however, are well preserved and reveal their sedimentary origin. The total amount of plagioclase and bornblende in these parts is larger than it is in the surrounding gneiss. The distribution of these large grains is not confined to bedding planes as it should be if these grains crystallized from the sedimentary material. On the contrary, the feldspathized parts form irregularly shaped patches cutting also across the bedding. It is thought, therefore, that the large plagioclase and hornblende grains crystallized because of an introduction of iron, magnesium, calcium, aluminum, and sodium into metasedimentary rocks rich in quartz. The origin of biotite and of some plagioclase in the gneiss is uncertain. Besides the gneissic rocks, there are accumulations of dark minerals, hornblende and biotite, both of metasomatic origin.

GNEISSIC ROCKS OF METASOMATIC ORIGIN

AUGEN GNEISS

Several layers of augen gneiss are exposed along the North Fork of the Clearwater River north of Bruces Eddy. Remnants of older rock included into this gneiss are fine-grained biotite-plagioclase gneiss. Occasional large round plagioclase grains occur in these inclusions. Segregations of biotite are common between the inclusions and the gneiss (fig. 61*A*). The grain size in the gneiss is highly variable. The augen measure from 0.5 to 1.5 cm in diameter, rarely more. The rock between the augen is fine-grained biotite-plagioclase gneiss with scattered round plagioclase grains that measure 1 to 3 mm in diameter. It is impossible to tell whether these small round plagioclase grains crystallized from the material originally present in the sedimentary layer or from elements introduced at an early stage during the metamorphism. Many of the large round plagioclase grains include small quartz and biotite grains. The amount of quartz in the gneiss with large round plagioclase grains is less than in ordinary biotite-plagioclase gneiss. It is concluded, therefore, that the feldspathization was associated with an exchange of material during which silicon was removed and sodium, calcium, and aluminum were introduced. Probably some iron, magnesium, and potassium were also removed during the feldspathization, but these elements did not migrate far. They segregated along the border zones and formed small lenticles and layers of biotite. In places these lenticles were later surrounded by the invading front of feldspathization. If the parent rock—the finer grained gneiss—contained potassium feldspar, this mineral either was segregated into thin layers or bands or formed large oval microcline grains that crystallized with quartz and plagioclase to form coarse-grained light-colored bands. The garnet in the garnetiferous layers (for example, north of loc. 428) and epidote in some other layers (as at loc. 56) occur in these pegmatitic layers with the light-colored minerals (fig. 61*B*).

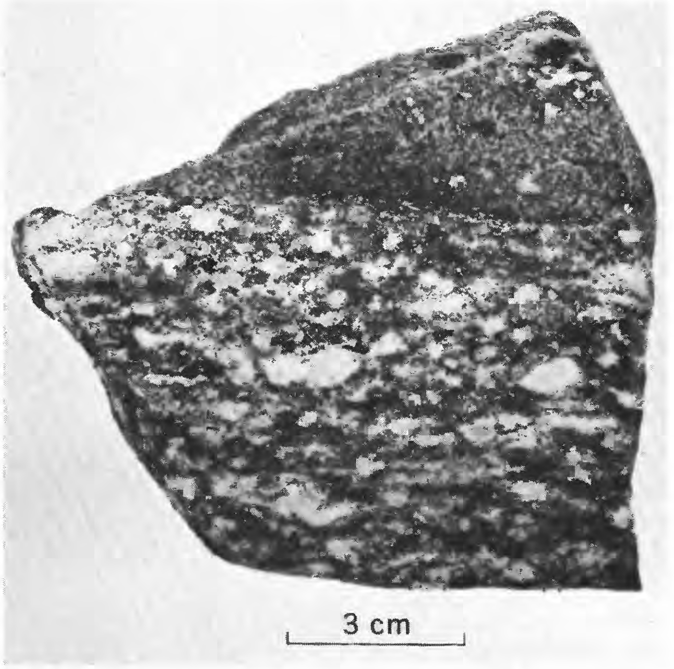
TONALITIC GNEISS

In places the biotite gneiss and hornblende gneiss contain more plagioclase than the ordinary varieties. These have an appearance similar to that of the gneissic tonalite except for less plagioclase. A closer study of such gneiss near Ahsahka reveals that a part of the plagioclase and hornblende (when present) in the gneiss also occur as larger rounded grains embedded in a finer grained gneissic rock (fig. 61*C*). Thin-section study shows that the rock between the large grains is similar to the biotite-plagioclase gneiss or biotite-bornblende gneiss nearby. Inclusions of these fine-grained rocks are common in the gneiss. It seems that this gneiss is a result of a late development of large plagioclase and hornblende grains in the metasedimentary rock. In contrast to the augen gneiss, the late plagioclase in the tonalitic gneiss occurs as somewhat smaller round grains; the number of these grains is large enough to give the gneiss the appearance of a plutonic rock.

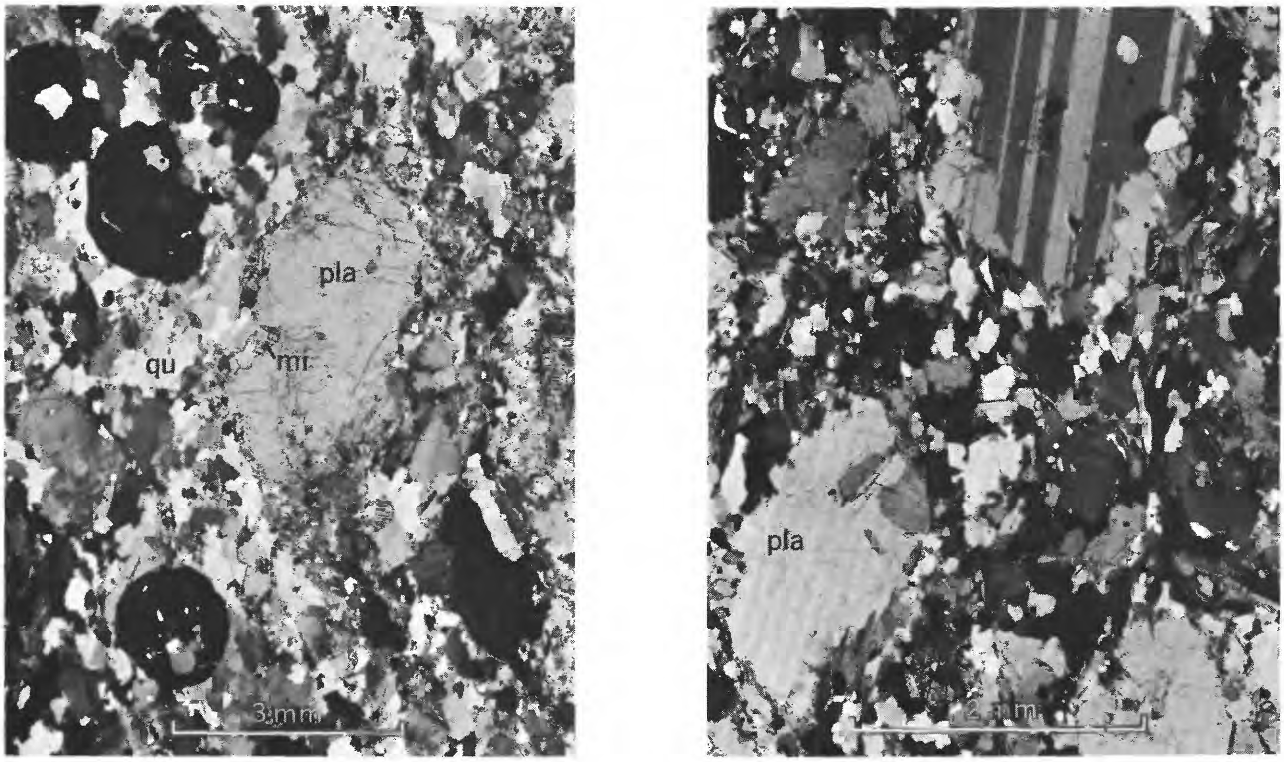
Thin sections show every gradation from the augen gneiss through tonalitic gneiss to gneissic tonalite. The principal difference between the tonalitic gneiss and gneissic tonalite is textural. In the tonalite a part of the plagioclase and hornblende occurs as large rounded grains, but the grain size of the minerals occurring between these large grains is larger than that in the tonalitic gneiss. This textural difference is probably due to more nearly complete recrystallization of the minerals of the parent rock in the tonalite. In contrast to these tonalites of probable metasomatic origin, some layers in the metasedimentary rocks contain abundant small plagioclase grains and are chemically identical with the tonalite. Only recrystallization would be needed to change these layers to tonalite.

DIORITIC GNEISS

A coarse-grained gneiss rich in hornblende and plagioclase has been formed from the fine-grained metasedimentary amphibolite and hornblende gneiss in a manner similar to that described for the tonalitic gneiss. Throughout this area every gradation from metasedimentary rocks to plutonic rocks can be found. For example, north of Ahsahka, hornblende gneiss layers are found to grade to dioritic gneiss and biotite gneiss layers to tonalitic gneiss when the strata are followed parallel to the bedding toward the east



A. Polished hand specimen of augen gneiss with dark fine-grained inclusion. Dark layer between the inclusion and the gneiss is biotite. About 1 mile north of Bruces Eddy (loc. 428, pl. 4).



B. Photomicrograph showing small round garnet crystals (black) in a microcline-bearing light-colored band in the coarse-grained gneiss. Just north of locality 428. pla= plagioclase; mi=microcline; qu=quartz. Crossed nicols.

C. Photomicrograph showing tonalitic gneiss that contains large round plagioclase (pla) crystals embedded in fine-grained biotite-plagioclase gneiss. From railroad cut 2 miles west of Ahsahka (loc. 29, pl. 4). Crossed nicols.

FIGURE 61.—SPECIMENS OF COARSE-GRAINED GNEISS FROM NORTH FORK OF THE CLEARWATER RIVER (*A*, *B*) AND TONALITIC GNEISS FROM WEST OF AHSAHKA (*C*)

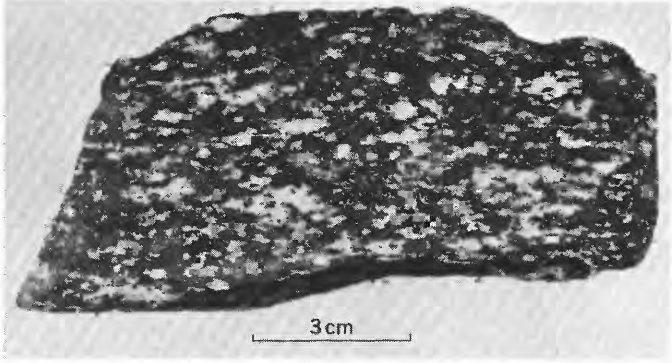


FIGURE 62.—Polished hand specimen of a gabbroic gneiss near Bruces Eddy (loc. 418, pl. 4). The light-colored minerals are quartz and plagioclase, the dark ones hornblende and biotite.

(pl. 4). Because some of the hornblende layers have a composition of quartz diorite, isochemical recrystallization would change them to a quartz diorite.

MASSES RICH IN FERROMAGNESIAN MINERALS

Segregations of hornblende and biotite were found in the metasedimentary rocks north of Bruces Eddy and east of the mouth of Whiskey Creek. North of Bruces Eddy lens-shaped bodies of ferromagnesian minerals from 0.5 to 5 m thick occur in banded biotite-plagioclase gneiss. Megascopically they resemble gabbroic gneiss (fig. 62) but contain abundant biotite. The biotite flakes are arranged along the *s* planes that are parallel to the bedding (and cleavage) in the surrounding gneiss. The rock cleaves easily along these planes.

The plagioclase grains in this gabbroic gneiss are segregated into long lenticles parallel to the foliation. Microscopic study shows that they are more sodic than plagioclase in a regular gabbro, being An_{32}, thus similar to the plagioclase in the surrounding gneiss. The hornblende is light green and only slightly pleochroic as is most of the late hornblende in this area. The biotite is reddish-brown and strongly pleochroic like the biotite in the surrounding gneiss. Some small round quartz grains are included in hornblende and plagioclase. Biotite flakes are much smaller than the hornblende grains and occur in thin laminae between layers of large polygonal grains of hornblende and plagioclase. The mode of occurrence of these bodies, their texture, and mineralogy differ strikingly from those of the calcareous layers and lenses in the metasedimentary rocks, and it was therefore concluded that these rock bodies are not metamorphosed calcareous concretions.

The texture and mineralogy of the gabbroic gneiss indicate that quartz and biotite are the early minerals and that hornblende and plagioclase crystallized late. The composition of the plagioclase suggests that the material in the plagioclase originated mainly in the parent rock, which was biotite-hornblende-plagioclase gneiss. The amount of hornblende is larger than that in any layer in the gneiss, and its distribution and mode of occurrence are more irregular. The abundance of the hornblende indicates that at least part of it crystallized because elements (magnesium, iron, and calcium) were introduced from outside sources. The gabbroic masses occur next to a fault zone. This fault zone apparently facilitated the introduction of solutions carrying the cafemic constituents. The scarcity of quartz in the gabbroic masses indicates that much of the silica was removed during the metasomatic replacement.

The contact relations, mineralogy, and texture of the hornblendite bodies near Ahsahka, and those east of the mouth of Whiskey Creek suggest that these bodies are metasomatic in origin. The quartz and biotite in these bodies are apparently relict minerals, and the large hornblende grains were crystallized because of introduction of iron, magnesium, and calcium.

Long sporadic hornblende prisms occur in fine-grained biotite-plagioclase gneiss near the contact of the quartz diorite south of Ahsahka. The hornblende prisms, 4 to 6 cm in length, transect the gneiss at random and are clearly younger than the minerals in the gneiss.

CONVERSION OF THE METASEDIMENTARY ROCKS TO THE PLUTONIC ROCKS NEAR OROFINO

The contact relations, textures, and mineralogy of the plutonic rocks suggest that some of these rocks are intrusive, others are metasomatic, and still others are products of isochemical recrystallization of metasedimentary gneiss. The textures of gabbro and quartz diorite north of Peck suggest an intrusive origin for these two rock bodies. The gradational relations between the metasedimentary gneisses and the gneissic tonalite and gneissic quartz diorite and comparison of the textures and chemical composition of all rocks concerned suggest that the gneissic tonalite and gneissic quartz diorite were formed from metasedimentary rocks by a partial replacement of quartz by plagioclase and hornblende.

As shown in figure 60 the normative amount of feldspar in the plutonic rocks is much larger than that in the metasedimentary rocks. The amount of normative femic constituents remains about the same if the gneissic rocks are converted to tonalite, but there is an increase also of the femic constituents if they are converted to diorite. The amount of normative quartz decreases in both cases. Only the amount of plagioclase increases, and the amount of quartz decreases if the metasedimentary rocks are converted to tonalite.

It is not known whether some elements were introduced into the metasedimentary rocks before tonalitization and dioritization. The texture of the biotite gneiss and biotite-hornblende gneiss, especially the mode of occurrence of the plagioclase in these rocks, resembles that of the metasomatic tonalite except that the grain size of the plagioclase is smaller. If there was an introduction of elements (feldspathization) at an early stage, only certain layers—the gneisses—were affected, whereas other layers—the quartzite and limestone—were unaffected. This kind of feldspathization does not seem very likely, because during a later phase in tonalitized areas all layers were equally converted to tonalite. Thus it seems more reasonable to assume that the gneissic layers originally contained some sedimentary plagioclase and quantities were only locally increased during the tonalitization. The increased portion crystallized as large grains replacing other minerals. A definite correlation of the metasedimentary rocks to their less metamorphosed equivalents farther north (?) or northeast (?) would help to provide the answer.

The same reasoning holds true for the origin of small biotite and hornblende grains in the gneisses. Both of these minerals in the fine-grained bedded gneisses are distributed in definite layers, suggesting that they crystallized from the sedimentary material rather than from the introduced elements. Distribution of minerals whose crystallization is a result of an introduction element is usually more irregular than that of the small biotite and hornblende grains in the gneisses. Within the metasedimentary formation the fine-grained gneisses of tonalitic and quartz dioritic composition occur as definite bedlike layers, which supports the suggestion that all minerals in them crystallized from the sedimentary material.

The composition of the amphibolite layers is similar to that of a hornblende gabbro, and presumably recrystallization alone could convert the amphibolite to gabbro. However, the amphibolite layers are usually converted either to diorite or to hornblende tonalite that contain less hornblende than the original amphibolite; this indicates that part of the iron and magnesium were removed. This part could be a source of the iron and magnesium that have been added to the quartzitic and limy layers nearby. Considering the whole area, considerable amounts of iron and magnesium must have been introduced from outside sources because the amount of sedimentary material rich in iron and magnesium is not sufficient to be a source for all iron and magnesium introduced into the biotite gneiss and quartzite during the dioritization.

The contact relations between the marble and plutonic rocks indicate that much of the calcium from the marble layers has been incorporated into the plutonic rocks. The fact that the contact rocks are rich in such minerals as grossularite, diopside, hornblende, and epidote indicates the presence of magnesium, iron, aluminum, and silicon, as well as calcium. In the contact zone the increase in the amount of hornblende toward the plutonic rock, and the decrease of the calcium-rich silicates—as diopside, grossularite, and epidote—indicates a loss of calcium in the outer zone. In the zone nearer the plutonic rock the hornblende rock has been converted to diorite and tonalite by the introduction of sodium and silicon and the loss of calcium, magnesium, and iron. The removed magnesium and iron probably reacted with the unaffected part of marble to form more calcium-rich silicates, thus broadening the silicated zone. At many localities all the limestone is converted into dark hornblende rock and amphibolite. Near the tonalite bodies the hornblende rock and amphibolite are feldspathized to form either diorite or tonalite. Abundant epidote is usually present in the plutonic rock formed in this manner. For example, between Orofino and Ahsahka (loc. 770) the layer in the tonalite that structurally corresponds to the southeastward extension of the silicated limestone No. 568 contains abundant epidote. Chemically this layer (No. 770) in the tonalite is richer in calcium than the neighboring layers, which shows that the products of tonalitization are not homogeneous.

MECHANISM OF THE TRANSFORMATION

The planar and linear structures of the metasedimentary rocks have been preserved during the metasomatism. The contacts between the metasedimentary and plutonic rocks are either conformable to bedding or interfingering across it. Only in a few localities do the newly crystallized minerals grow across the structures of the metasedimentary rocks. These structural relations suggest that the major part of metasomatic introduction of elements took place either earlier than the deformation or simultaneously with it.

The rocks produced by an exchange of elements contain abundant plagioclase and hornblende. The salic and femic constituents were probably introduced simultaneously and were deposited together in the country rock without notable differentiation. Such is to be expected during the injection of magmas under dynamic conditions. Probably small bodies of gabbroic and quartz dioritic magma were forced into the country rocks of the Idaho batholith during the folding. Solutions carrying elements that would yield bornblende, biotite, and andesine were moved bodily into the sedimentary formations along planes of least resistence:

bedding planes, cracks, joints, and other structural planes. Subsequently, the transfer of elements was chemical, as is demonstrated by the replacement textures. The femic constituents were deposited as large hornblende and biotite grains in the metasedimentary rock nearby, and the salic elements formed large round plagioclase grains in the same rock; all these newly formed minerals replace mainly quartz. Stresses in operation during the deposition caused an orientation of newly formed minerals and preserved the orientation of old minerals.

The solutions probably approached the composition of hornblende pegmatite. The occurrence of hornblende pegmatite as segregations in the quartz diorite suggests that the elements of hornblende and plagioclase stayed mobile to the latest stages in the crystallization of the quartz dioritic magma.

AGE OF THE PLUTONIC ROCKS NEAR OROFINO

Most of the plutonic rocks in the Orofino district are believed to be of metasomatic origin, and were probably formed during the emplacement of quartz dorite near Greer and Peck. These quartz diorite bodies may be connected with the petrographically similar quartz diorite along the border zone of the Idaho batholith. The rocks belonging to the quartz monzonite series cut the quartz diorite in the area about 35 miles east of Orofino and are therefore younger. Thus it seems that the metasomatism which converted the metasedimentary rocks to plutonic rocks near Orofino may have been contemporaneous with the earliest phase of the intrusion of the Idaho batholith. Yet, this study shows that there is a definite structural relation between the older metasedimentary rocks and younger plutonic rocks. Parallelism of the linear and platy structures and the identical orientation of minerals in both groups of rocks suggest that the plutonic rocks were formed under the same stresses that caused the orientation of minerals in the metasedimentary rocks. Consequently, it seems possible that the dioritization and tonalitization were contemporaneous with the latest phase of the folding which is considered to be of Nevadan age. These structural relations suggest that border zone and the satellitic intrusions are somewhat older than the main part of the batholith.

According to Ross (1928), the Idaho batholith is post-Triassic and pre-Tertiary, probably of Early Cretaceous age. Eardley (1951, p. 312) suggested a Late Cretaceous and early Tertiary (Laramide) age for the batholith. Langton (1935) came to the same conclusion in his work in the northeastern part of the Idaho batholith. Anderson (1952) considers that the Idaho batholith resulted from multiple intrusions that occurred from the end of Jurassic time to Late Cretaceous time. The lead-alpha age determinations by Larsen and others (1954) of five rocks from the Idaho batholith average 103 million years; this places the intrusion within late Early Cretaceous.

CONCLUSIONS CONCERNING THE METASOMATISM NEAR OROFINO

The Orofino district provides a good example of a transformation of metasedimetary rocks to plutonic rocks. The chemical composition of the metasedimentary rocks has undergone a gradual change during the transformation. The end product in the Orofino district is not a granite but a tonalite, a plutonic rock poor in potassium. The metasedimentary rocks are impoverished mainly in silicon and enriched in iron, magnesium, calcium, and aluminum. Mineralogically this change is seen in the replacement of quartz by andesine and hornblende.

Structures and textures of the rebuilt plutonic rocks suggest that they were subjected to stresses during the mineral growth. No differentiation occurred during the metasomatism; femic and salic minerals crystallized contemporaneously. It is thought, therefore, that the major part of the fluids carrying the elements of plagioclase and hornblende impregnated the metasedimentary rocks along the structural planes. From there on the mode of transport was chemical, as the replacement textures show. Because the linear structure of the plutonic rocks is parallel to the lineation (and second fold axes) of the metasedimetary rocks, the metasomatism seems to be contemporaneous with the folding, which is of Nevadan age, whereas in many localities in the Headquarters quadrangle and Big Island area the metasomatic development of hornblende and plagioclase in the schist and quartzite was found to be postdeformation. Therefore, the metasomatic introduction of elements that caused the formation of gneissic tonalite and quartz diorite near Orofino took place during an earlier phase than the introduction of the same elements (mainly iron, magnesium, calcium, aluminum, and sodium) in the Headquarters quadrangle and near Big Island.

PART 4. THE DENT AREA

The outcrops along the canyons near Dent offer exceptionally good material for the study of metasomatic changes in schist and quartzite. As has just been described, the metasomatic changes near Orofino took place contemporaneously with the deformation; whereas in the Headquarters quadrangle the development of hornblende in the diopside-plagioclase quartzite postdates the deformation. Near Orofino the resultant rocks are mainly quartz diorite and tonalite; whereas in the Headquarters quadrangle and in the vicinity of Big Island rocks rich in hornblende and biotite were formed. It is apparent that the same elements—iron, magnesium, calcium, aluminum, and sodium—were introduced in both parts of the area, but that there was a difference in time and manner of deposition of the introduced elements. The additional information obtained from the Dent area throws new light on many problems concerning this metasomatism northwest of the Idaho batholith.

The relation between the fold axes and lineation in the metasedimentary rocks near Orofino and in the Headquarters quadrangle suggests that there were two sets of folding. Good outcrops near Dent yielded more detailed information to confirm this.

METASEDIMENTARY ROCKS

The metamorphic rocks near Dent (pl. 3) resemble the schist and diopside-plagioclase gneiss and quartzite that are considered to be equivalent to the upper part of the Wallace formation in the vicinity of Big Island and in the Headquarters quadrangle. However, lime-silicate rock (originally a limestone), which was not found east of Dent, occurs near Dent between the mouths of Elk Creek and Dicks Creek, and thin-bedded biotite gneiss with hornblende-bearing layers is interbedded with the schist and diopside-plagioclase gneiss at Dent. These rocks resemble the thin-bedded gneisses near Orofino.

The metasedimentary series at Dent consists of coarse-grained pure quartzite, thin-bedded biotite quartzite and diopside-plagioclase gneiss and quartzite, sillimanite-garnet schist, diopside-grossularite-plagioclase rock (lime-silicate rock), and hornblende and biotite gneiss. The plutonic rocks are mainly quartz diorite and tonalite. A few small gabbroic bodies occur in the schist.

STRATIGRAPHIC SEQUENCE

Metasedimentary rocks retain their character only in small local areas. Most of these rocks are feldspathized to the extent that it is difficult to work out any stratigraphic sequence. In addition, numerous faults have disturbed the original structure. Between the mouths of the Elk Creek and Dicks Creek a rock series consisting of fairly pure quartzite, lime-silicate rock with minor marble lenticles, and garnet-sillimanite schist, is well exposed. A fault occurs between the lowest quartzite bed and a garnet-sillimanite schist south of it. On the upper part of the canyon wall another layer of similar light bluish white quartzite with granular beds is overlain by lime-silicate rock, above which is a garnet-sillimanite schist. South of the mouth of Dicks Creek the garnet-sillimanite schist is feldspathized and encloses minor tonalite bodies. A coarse-grained thick-bedded quartzite with a glassy appearance is exposed south of the schist and tonalite. Quartz diorite and tonalite interrupt the sequence south of this quartzite.

Along the North Fork of the Clearwater River at Dent a similar rock series is exposed. There, a coarse-grained quartzite is overlain by schist, quartzite, and lime-silicate rock. An overthrust fault separates the coarse-grained quartzite from the underlying sillimanite-garnet schist. The quartzite that is exposed about 2 miles north of the mouth of Elk Creek is probably a northwesterly extension of this lower coarse-grained quartzite layer. Like the quartzite southwest of Dent (loc. 471, pl. 3), the layer at Dent also is interbedded with garnet-sillimanite schist. The quartzite east of Elk Creek (2 miles north of Dent) and that south of the mouth of Meadow Creek in the gorge of Elk Creek are probably northward extensions of the upper quartzite layer that is overlain by lime-silicate rock and diopside-plagioclase gneiss and diopside quartzite. This sequence is similar to the lower part of the Wallace formation on Surveyors Ridge (table 4). Because of the faulting and interruption by plutonic rocks, it is impossible to measure the thicknesses of various layers. The estimated thicknesses of some layers are given in table 16. The thickness of the lower quartzite layer near Dent and its northwest extension 2 miles north of the mouth of Elk Creek are only about 50 m, but some of the quartzite in each locality may be missing because of faulting. The quartzite layer and intercalated lime-silicate rock between Dicks Creek and Elk Creek is about 200 m thick.

A diopside-plagioclase gneiss is exposed along the river trail south of Dent (loc. 197, pl. 3). The main rock type in this locality is a thin-bedded biotite quartzite with some hornblende-bearing layers. A similar thin-bedded biotite quartzite and diopside-plagioclase gneiss occur in the schist along the river east of Dent and also in the gorge of Elk Creek in the northernmost part of the area. In the easternmost part of the Dent area the schist is the main rock type. Some fairly thin

TABLE 16.—*Tentative correlation of two sections of metasedimentary rocks in the Dent area*

Formation	Correlation with subdivision in table 4	Lithology near Dicks Creek	Lithology along Elk Creek
Wallace	5	Not exposed	Biotite schist.
	4	Not exposed	Biotite quartzite and coarse grained quartzite with layers of diopside-plagioclase gneiss and quartzite.
	3	Sillimanite-garnet schist	Thin-bedded quartzitic biotite schist.
	1-2	Coarse - grained pure quartzite with layers of lime-silicate rock, about 200 m thick. Fault	Biotite quartzite, coarse-grained quartzite, biotite quartzite
St. Regis		Sillimanite-garnet schist, biotite schist.	Sillimanite-garnet schist. Fault
Revett		Coarse-grained quartzite 150 m thick.	Coarse-grained quartzite.
Prichard and Burke		Not exposed	Sillimanite-garnet schist.

layers of biotite quartzite and hornblende-biotite gneiss are intercalated with the schist.

STRUCTURE

The metasedimentary rocks are strongly folded and faulted. In addition they show a clear lineation and foliation.

BEDDING

The quartzite layers and some of the schist show a distinct bedding that is due to variation in amounts of the major constituents. In the coarse-grained quartzite the individual beds, ranging from 1 to 50 cm in thickness, are separated by thin biotite laminae. In thin-bedded biotite schist and gneiss with hornblende-bearing layers the bedding is pronounced; layers rich in hornblende and those rich in plagioclase alternate with biotite-bearing layers.

The bedding in the sillimanite-garnet schist can be detected only in some outcrops. The schist is strongly deformed, wrinkled, and metasomatically changed in its composition. Commonly there is only one planar structure, the foliation, and it is not always certain whether the bedding is parallel to this structure. However, the quartzitic or gneissic layers interbedded with the schist make it possible to map the bedding in detail. In the southwestern part of the area and also along the river canyon farther south the beds dip 40°–50° NE. Gentle dips were measured at many localities near Dent.

FOLDING

Several small folds are exposed along the river at Dent and on the road cuts northeast of the mouth of Dicks Creek and along Elk Creek. There are two sets of folds. Near Dent the axis of one set plunges gently to the east, but near the mouth of Dicks Creek it plunges to the west. The axis of the other set of folds, trending N. 20°–60° E. and plunging 20°–30° NE., is the direction of a strong lineation. The folds around the east-trending axis are overturned to the south and those around the second fold axes—the lineation—are overturned to the southeast. In the schist above the coarse-grained quartzite at Dent (loc. 550) the crests of the folds around the east-trending axis are broken along the overthrust planes.

FOLIATION

In the southern part of the area and in many localities where only flanks of larger folds are exposed, the foliation parallels the bedding. Only in a few localities a transecting cleavage parallel to the axial planes was observed. In the schist layers and in the biotite-rich quartzite along the river west of Dent, foliation is parallel to the axial planes of the overturned folds. In the coarse-grained quartzite at Dent closely spaced cleavage planes dip 65° SE. and are parallel to the lineation in the overlying schist layer.

LINEATION

Most outcrops of metasedimentary rocks show a strong lineation. In the quartzite near the mouth of Dicks Creek and at Dent this lineation is a parallel orientation of elongated biotite flakes. In the hornblende-bearing layers the long dimension of the hornblende parallels the lineation. In the schist this structure appears as a wrinkling of the bedding plane. The measurements of the lineation show that it is parallel to the axes of the second folding.

A detailed study of the lineation was made in a fold exposed in a quartzitic layer in the schist along the river west of Dent (loc. 12). The axis of this fold plunges 2° to 10° E. in the direction S. 70° E. A strong lineation appears on the flanks of this large fold which is domed. A careful study suggests that the lineation is a wrinkling of the bedding plane. There is no wrinkling on the round tight crest of the fold and the flanks seem rather straight on the *ac* plane of the fold (fig. 63). A few minor faults near the crest pinch off some layers.

PETROFABRIC STUDY

Two thin sections were cut of the fold in locality 12 (pl. 3), one from a flank parallel to the *bc* plane and the other from the crest parallel to the *ac* plane (fig. 64). The orientation of mica and quartz were measured in each section. On the flank of the fold the petrofabric diagrams show fairly good girdles around the lineation (fig. 65 and 66) which in this locality parallels the *a* axis of the fold. Therefore the girdles are *bc* girdles. The petrofabric diagrams of mica and quartz near the crest of the fold show broken girdles around the *b* axis (fig. 67 and 68). A composite diagram of the crest would give complete *ac* girdles. Thus the lineation shows characteristics of a folding both in megascopic microsocopic scale. Because its direction usually deviates 10° to 50° from the direction of the *a* axes of the major folds, it was concluded that it is a true second folding.

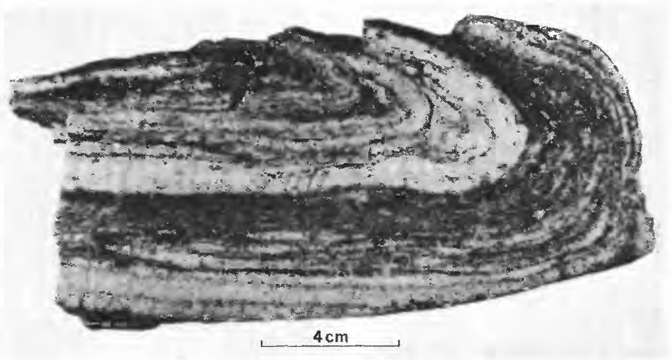


FIGURE 63.—A polished hand specimen showing *ac* plane of a fold west of Dent. (loc. 12, pl. 3).

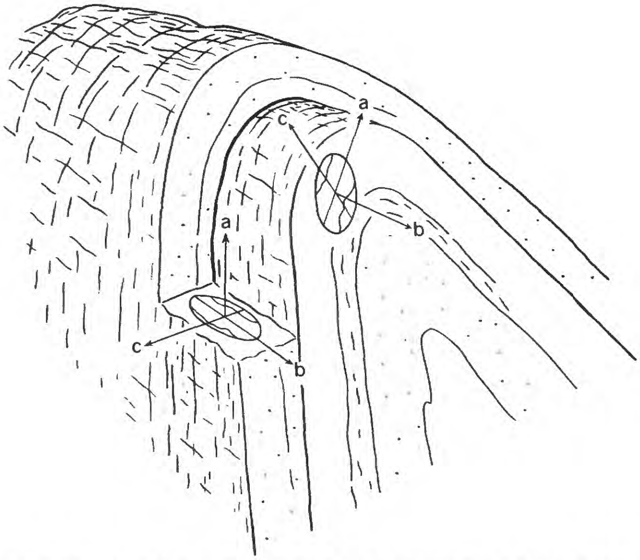


FIGURE 64.—A sketch of a fold near Dent showing the location of the two thin sections subjected to a petrofabric study.

RELATION BETWEEN THE TWO SETS OF FOLDING

The time of the second folding varies in relation to the major east-west folding. In many outcrops the major folding seems earlier but in others later. In an area about 10 miles north of Dent the second folding appears locally to be clearly later than the main folding. There the axes of large folds are horizontal and trend N. 70° W., and the axes of small folds plunge 10° SW. in the direction S. 70° W. The small folds are later and are overturned to the northwest. The crests of these folds are broken into overthrust faults which dip only about 10° SE. At locality 12, west of Dent, the large folds around the east-trending axis are overturned to the south and the overthrusting, which is a late movement, is to the south. Wrinkling of the flanks of these large folds is apparent, but there is no wrinkling on the crests of the folds. The straight flanks and round but fairly sharply turned crests are typical of all large folds in thin-bedded quartzite layers and give an impression that the rock was fairly competent when folded around the eastward-trending axis.

COMPARISON OF THE TWO SETS OF FOLD AXES WITH THE REGIONAL TRENDS

Eardley (1951) has analyzed the regional trends around the Idaho batholith and, according to his compilation, the trends of Nevadan orogeny form two arcuate segments. The Idaho batholith is just south of the junction of the arcuate segments which are convex westward. It seems that in the area studied there is a good parallelism, first, between the major fold axes and trends of the south end of the northern arcuate segment, and second, between the lineation (axis of the second folding) and trends of the north end of the southern arcuate segment. This parallelism suggests that both sets of folds were formed during the Nevadan orogeny. In the light of this interpretation it is easily understood that the second folding can be either later or earlier than the major folding or simultaneous with it. Both arcuate segments of Nevadan orogeny were folded simultaneously, but the release of stress took place intermittently through time. Therefore, in some localities major folds may have formed first, and later these folds were crumpled around the axis of second folding (lineation). In other places the beds first wrinkled around the axis of the southern arc (that is, parallel to the lineation) and were folded at a still later time around the axis of the northern arc (major fold axis).

FAULTS AND OVERTHRUSTS

Minor overthrust faults parallel to the axial planes of the major folds that are overturned to the south are common. Along the river at Dent (loc. 550) these thrust planes strike N. 80° W. and dip 10° N. (fig. 69*A*) but near the mouth of Dicks Creek the thrust fault between the schist and quartzite strikes N. 70° E. and dips 70° N.

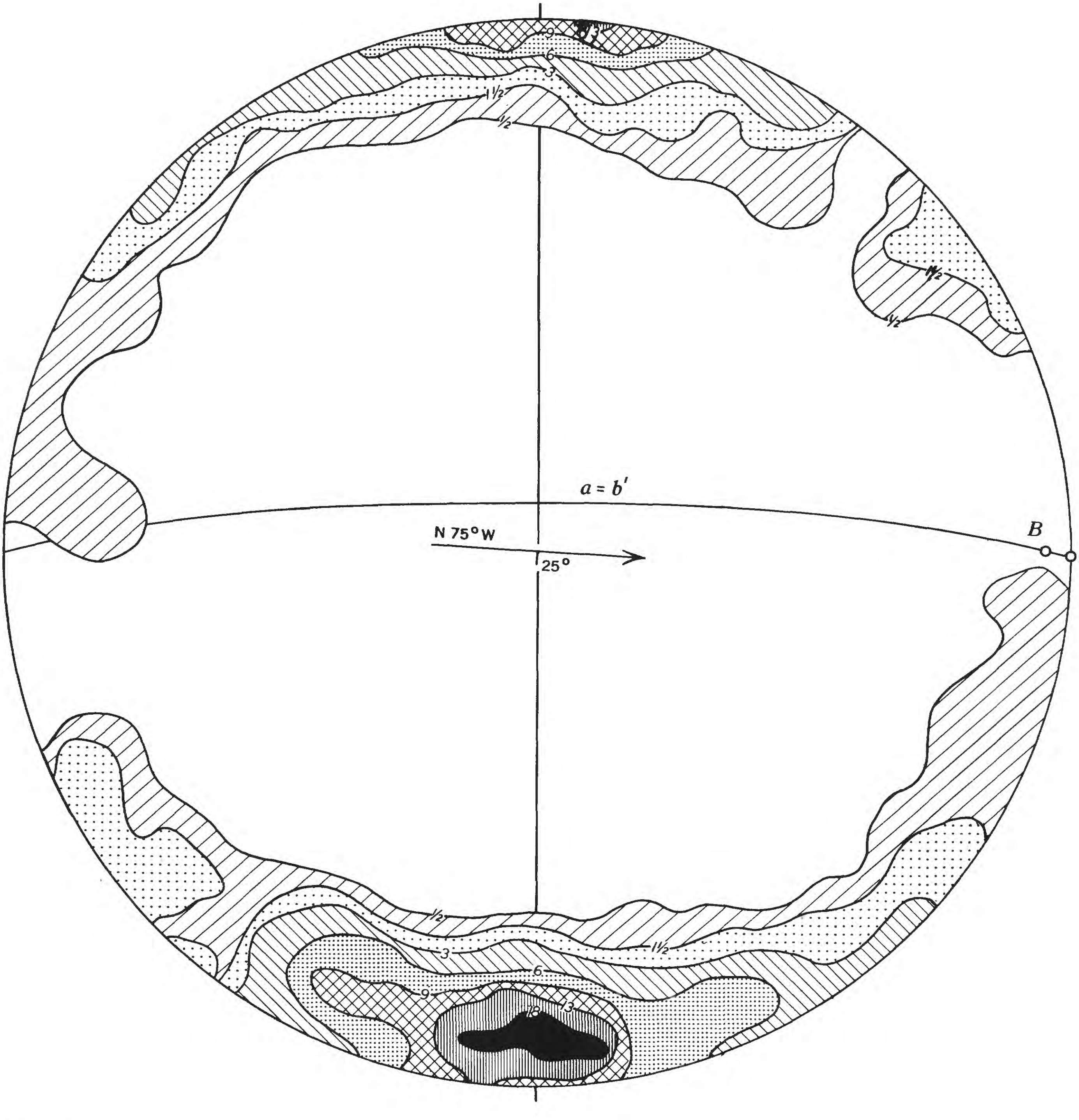


Figure 65.—Petrofabric diagram showing the orientation of mica cleavage on a flank of a fold in figure 64 (loc. 12); 216 biotite and chlorite flakes; contours: ½, 1½, 3, 6, 9, 13, and 18 percent.

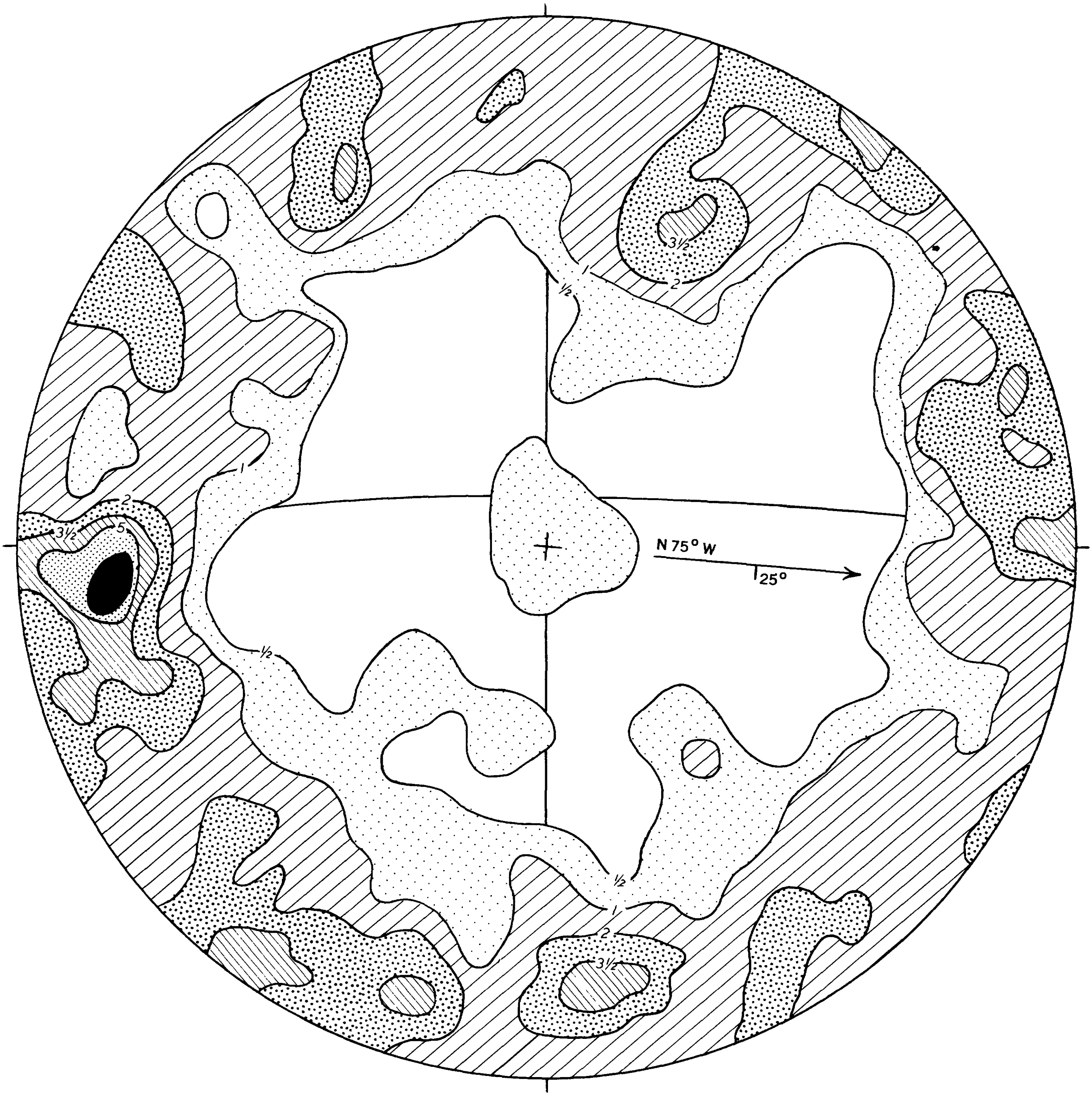


FIGURE 66.—Petrofabric diagram showing the orientation of *c* axes of quartz in the thin section of figure 65 (loc. 12). 325 quartz axes; contours: ½, 1, 2, 3½, and 5½ percent.

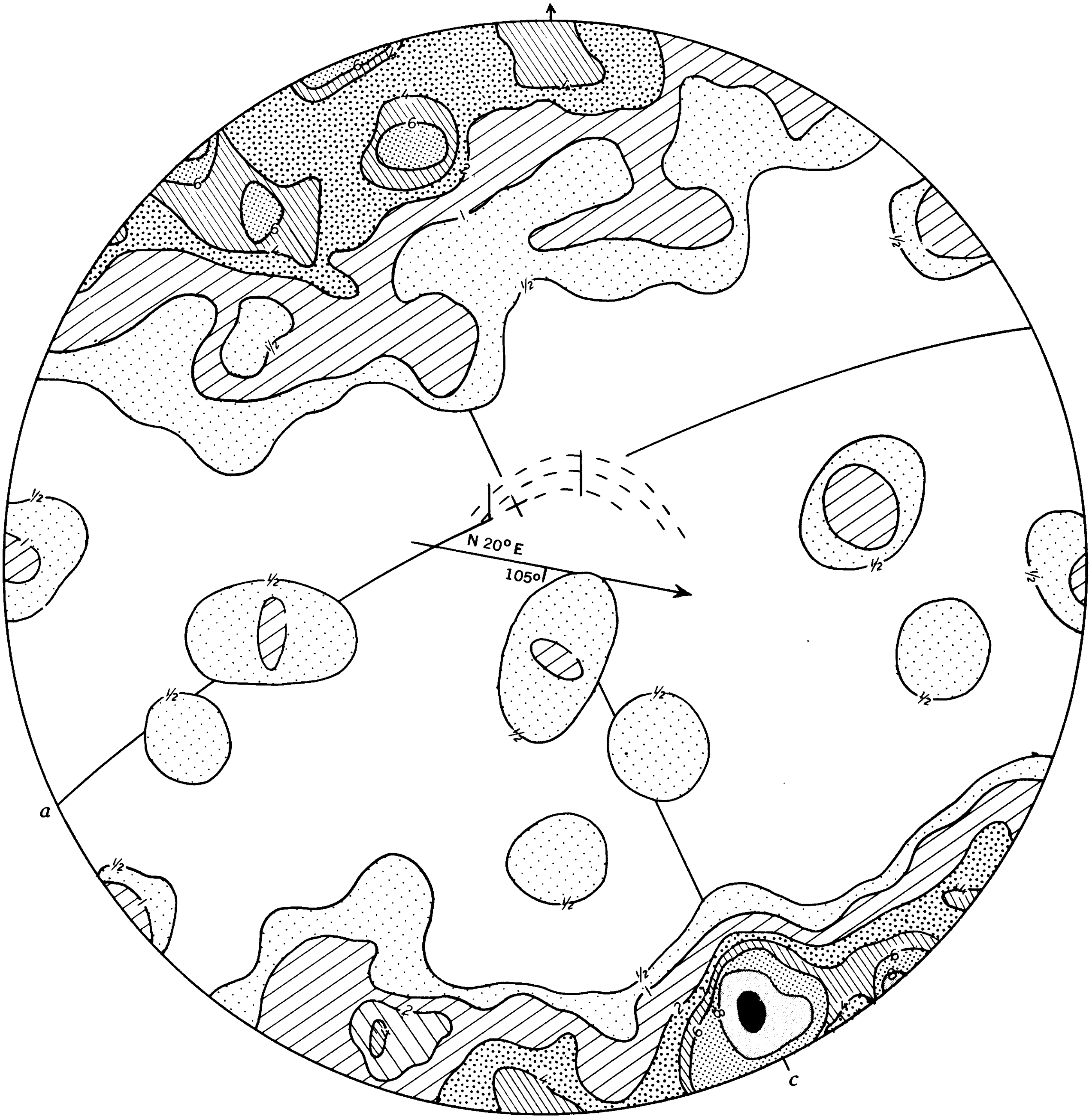


FIGURE 67.—Petrofabric diagram showing the orientation of mica on the crest of the fold in locality 12. 120 biotite flakes; contours: ½, 1, 2, 4, 6, and 8 percent.

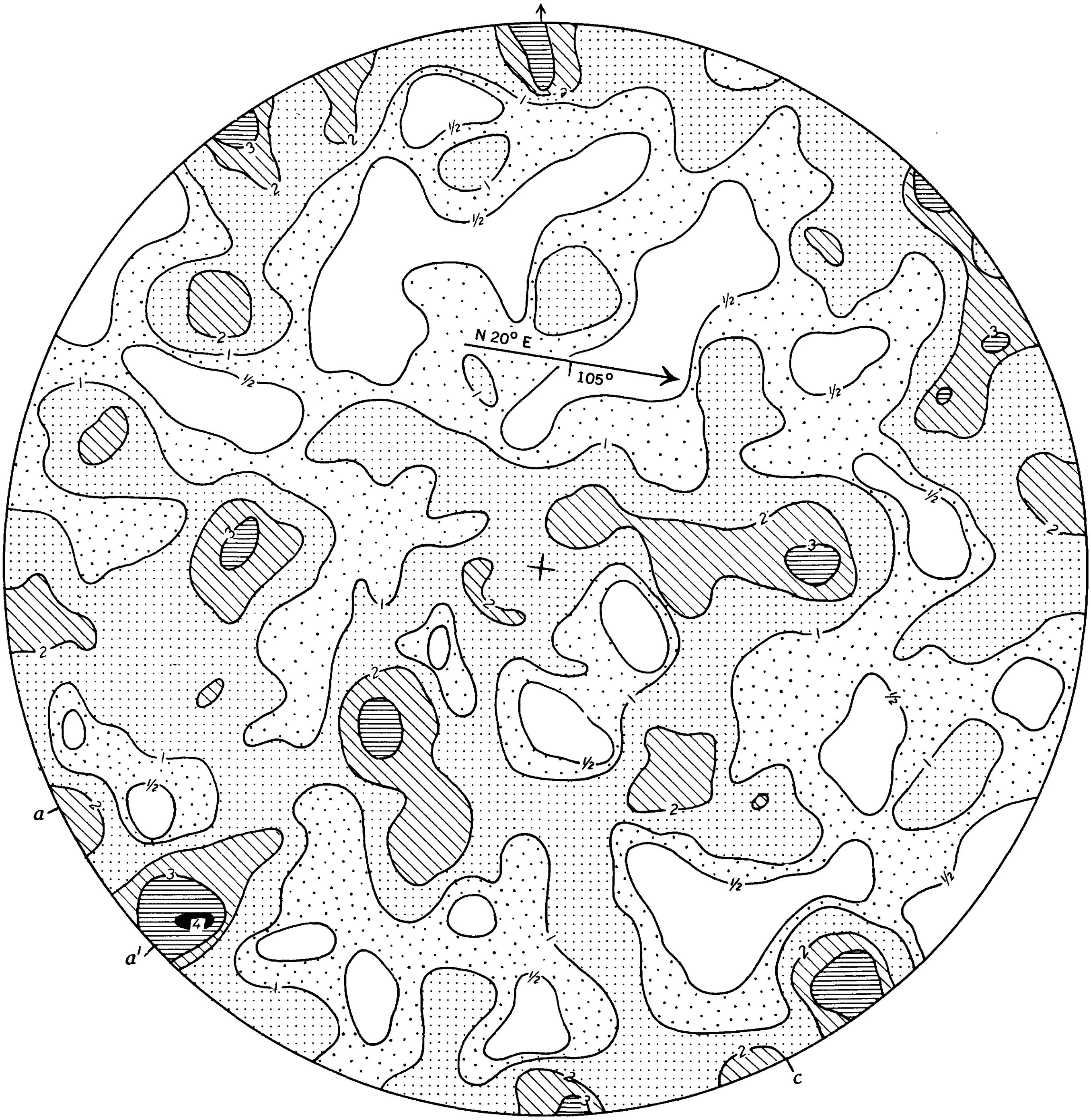


FIGURE 68.—Petrofabric diagram showing the orientation of quartz in the thin section of figure 67. 200 quartz axes; contours: ½, 1, 2, 3, and 4 percent.

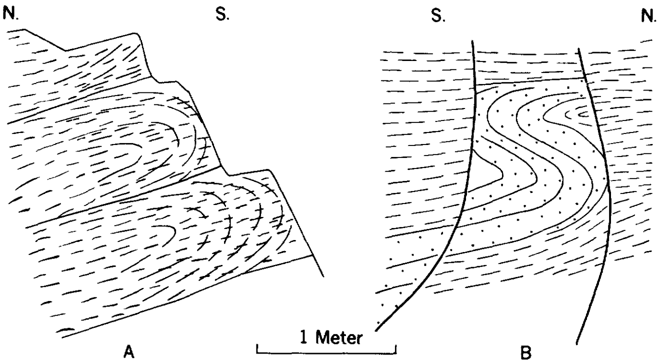


FIGURE 69.—*A*. Small overthrust faults parallel to the axial planes of the overturned folds near Dent (loc. 550, pl. 3). Vertical wall faces to the west. *B*. Underthrust in schist with a layer of white granular quartzite. Vertical road cut faces to the east. One mile north of the mouth of Elk Creek.

An underthrust to the south is exposed in a road cut about 0.5 mile north of the mouth of Elk Creek. An almost horizontal thrust plane cuts off a layer of white granular quartzite that is about 50 cm thick (fig. 69*B*) between two later vertical faults.

The north-south faults are steeply dipping and late as shown by a graben, 100 m wide, near the mouth of Elk Creek. This graben is filled by gravel and sand that show crossbedding. Study of microfossils in this sediment by E. S. Barghoorn of the U.S. Geological Survey did not reveal a definite age.

PETROGRAPHIC DESCRIPTION

COARSE-GRAINED QUARTZITE AND GRANULAR QUARTZITE

The coarse-grained quartzite is light bluish or grayish white, and contains individual biotite flakes that are either evenly distributed throughout the rock or form thin laminae parallel to the bedding planes. The biotite flakes which measure 1 to 3 mm in their longest dimension are well oriented parallel to the lineation in the bedding planes. The layer along the river at Dent and that exposed south of the mouth of Meadow Creek and east of Elk Creek contain more biotite, which is evenly distributed throughout most of the layers.

Microscopic study shows that the biotite, which is the only micaceous mineral in the coarse-grained quartzite, occurs in long thin flakes between the large quartz grains. The flakes are parallel to the bedding. Some small round or angular grains of plagioclase occur between the quartz grains or are included in them. Many of these plagioclase grains are partly altered to sericite and epidote minerals. The small quartz grains are elongated parallel to the bedding, but the large grains that form the main part of the rock are irregular in shape and show strong strain shadows. Some large grains in the coarse-grained quartzite at Dent (loc. 550) are angular, the boundaries being parallel to the bedding and to the fracturing that transects the bedding at an angle of 65°. These grains have elongated across the bedding. The quartzite exposed between Dicks Creek and Elk Creek is light-bluish white and in addition to quartz contains only very little biotite and plagioclase. Individual beds are thinner than those in the coarse-grained quartzite and some beds seem granular. The quartzite 2 miles north of Dent is thin bedded and contains abundant biotite that occurs as laminae parallel to the bedding.

BIOTITE QUARTZITE

The thin-bedded biotite quartzite is gray and fine grained. The thickness of individual beds is only 1 to 20 mm. The bedding is distinct because of the variation of the amount of biotite in individual layers. Microscopic study shows that the texture is granoblastic and equigranular, quartz showing a strong undulatory extinction. Most layers of biotite quartzite contain some plagioclase (An_{34-37}), but it is not clear whether it is a primary constituent because feldspathization of the schist and biotite quartzite is common in this area. Magnetite, zircon, allanite, and epidote are the minor constituents.

DIOPSIDE-PLAGIOCLASE GNEISS AND QUARTZITE

The general appearance and mineral composition of the diopside-plagioclase gneiss and quartzite intercalated with the biotite quartzite along the river near Dent and along Elk Creek are much like those of the fine-grained hard diopside-plagioclase gneiss near Big Island and in the southern part of the Headquarters quadrangle. The stratigraphic sequence in which the gneiss and quartzite occur, however, is more like that along Cougar Creek in the northern part of the Headquarters quadrangle where layers of thin-bedded biotite quartzite alternate with layers of diopside-plagioclase gneiss. (See table 6.)

SILLIMANITE-GARNET SCHIST

The major part of the sillimanite-garnet schist in all parts of the Dent area is coarse grained and contains abundant biotite, garnet, and sillimanite. In contrast to the sillimanite-garnet schist in the Headquarters quadrangle there is scarcely any muscovite. Layers of fine-grained plagioclase-biotite schist and biotite quartzite are interbedded with the schist.

The biotite is reddish brown and strongly pleochroic. The sillimanite occurs in long needles and prisms that in many places are clustered or form nodules. The garnet is red almandine in euhedral or rounded crystals that range from 2 to 20 mm in diameter. In the outcrops not affected by feldspathization—such as those on the south side of the river 3 miles southeast of Dent—plagioclase is scarce or even absent. Zircon was found as inclusions in biotite in all thin sections studied.

A paragonite type of mica was found in some small lenticles in the schist east of Dicks Creek. This mica occurs in clusters of fine fibers, is white, has a pearly luster, and shows $\gamma=1.642\pm0.001$. Determination by visual spectroscopic methods by Nola B. Sheffey of the U.S. Geological Survey showed that it contains sodium as a chief constituent and lesser amounts of lithium and calcium. No potassium was detected.

THIN-BEDDED HORNBLENDE AND BIOTITE GNEISS

Along the river at Dent and about 2 to 3 miles east of there, layers of thin-bedded gneiss are interbedded with the schist. In its mineral content and appearance this gneiss is similar to the thin-bedded hornblende-biotite gneiss near Orofino. However, at Dent the amount of plagioclase is less and most of the layers are finer grained.

LIME-SILICATE ROCK

The lime-silicate rock is coarse to medium grained and heterogeneous. The main constituents—plagioclase, quartz, calcite, diopside, epidote minerals, grossularite, hornblende, phlogopite, and scapolite—occur in large crystals that are commonly segregated into beds and lenticular bodies. The color of the rock varies according to the abundance of the constituent minerals. Beds and lenticles rich in diopside and epidote minerals are grayish green; those rich in grossularite are reddish brown. Phlogopite forms monomineralic segregations which are as much as half a meter thick and 2 to 3 m long. Scapolite occurs as white masses, usually with plagioclase. Small lenticles of pure coarse-grained calcite are found in some layers. Originally this rock was probably a shaly dolomitic limestone, and much of the segregation occurred during the metamorphism.

Thin-section study shows that plagioclase, diopside, scapolite, and sphene crysallized first. Large garnet crystals include these minerals and therefore are later. Zoisite and sodic plagioclase are the latest minerals in many lime-silicate layers. They accur as large grains that include other minerals. Intergrowths between zoisite and plagioclase are common.

The constituent minerals in a lime-silicate rock east of the mouth of Dicks Creek (loc. 11, fig. 70) were studied in detail. The diopside in the sample from locality 11 is grayish green and occurs as small grains and prisms as long as 2 cm among the quartz and plagioclase. Its optical properties, $\alpha=1.700\pm0.001$, $\beta=1.707\pm0.001$, $\gamma=1.725\pm0.001$, and $2V=61°$, indicate that it contains iron and magnesium in about equal amounts (Hess, 1949).

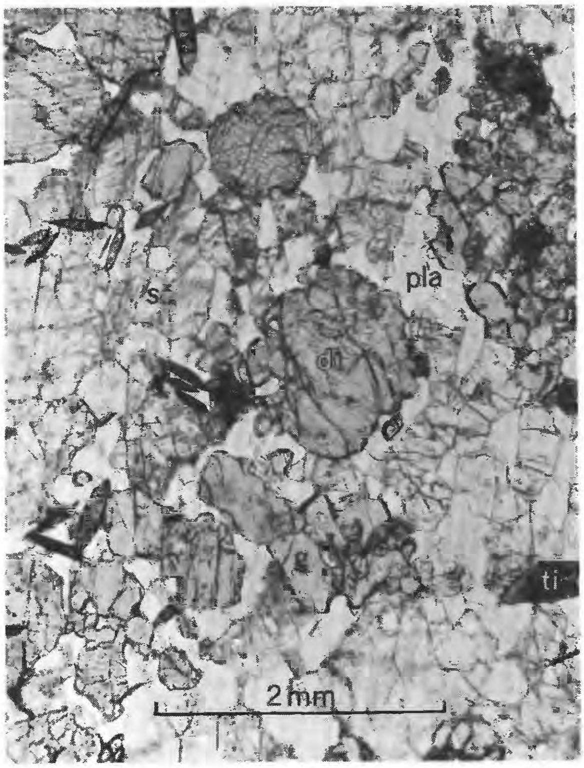


FIGURE 70.—Calcium-magnesium silicate rock with abundant scapolite (sc), diopside (di), and grossularite (g). pla=plagioclase; ti=sphene. Road cut along North Fork of Clearwater River half a mile east of the mouth of Dicks Creek (loc. 11, pl. 3). Plane polarized light.

Tremolite with $\alpha=1.604\pm0.001$, $\beta=1.619\pm0.001$, $\gamma=1.631\pm0.001$ and $Z\wedge c=20°$ was found in some specimens from locality 11. The phlogopite is very light brown to colorless under the microscope. It includes small crystals of zircon and shows $\gamma=1.577\pm0.001$ and $-2V=2°$. Specific gravity is 2.86. Calculation of the formula from the chemical analysis (table 17) gave the following result:

$$(K_{1.79}Na_{0.08}Ca_{0.01})_{(\Sigma=1.88)}(Mg_{5.41}Fe''_{0.35}Ti_{0.06}P_{0.02}Fe'''_{0.04}Al_{0.10})_{(\Sigma=5.98)}(Si_{5.99}Al_{2.01})_{(\Sigma=8.0)}(F_{1.19}(OH)_{2.69}O_{20.12})_{(\Sigma=24.0)}$$

This corresponds favorably with the theoretical formula:

$$K_2Mg_6(OH)_4Si_6Al_2O_{20}$$

in which Fe substitutes for a part of Mg and F for a part of OH.

The scapolite occurs either in small shiny prisms or in round or irregular-shaped grains among the other minerals (fig. 70). The indices of refraction $\omega=1.584\pm0.001$ and $\epsilon=1.552\pm0.001$ indicate, according to Winchell (1951), that the scapolite contains 62 percent meionite.

White zoisite is the most common of the epidote minerals, but clinozoisite, epidote, and allanite also are found. The indices of refraction of the zoisite are

$\alpha=1.700\pm0.001$, $\beta=1.702\pm0.001$, $\gamma=1.706\pm0.001$. Dispersion $v>\rho$ is strong and +2V small. Some calcite usually occurs with the zoisite.

Grossularite with $n=1.751\pm0.001$ is especially abundant next to the calcite lenticles, which are the only remnant of the originally fairly thick limestone. Euhedral crystals line the walls of the vugs from which calcite has weathered out.

TABLE 17.—*Chemical composition of phlogopite (No. 767) from lime-silicate rock east of the mouth of Dicks Creek (loc. 11)*

[Robert N. Eccher, analyst, U.S. Geological Survey]

Constituent	*Weight percent*	*Molecular equivalent*	*Number of atoms*		
SiO_2	42.19	7025	Si	5.99	8.00
Al_2O_3	12.60	1236	Al	2.01	
Fe_2O_3	.33	21	Al	.10	5.98
FeO	2.91	405	Fe‴	.04	
MnO	.02	3	Fe″	.35	
MgO	25.61	6352	Mn	.00	
CaO	[1].06	11	Mg	5.41	
Na_2O	.31	50	Ti	.06	
K_2O	9.91	1052	P	.02	
TiO_2	.61	76	Ca	.01	1.88
P_2O_5	.13	9	Na	.08	
F	2.66	1400	K	1.79	
H_2O+	2.84	1576	F	1.19	3.88
H_2O-	.33		OH	2.69	
			O	20.12	
Subtotal	100.51				
Less O	1.12				
Total	99.39				

[1] Determined spectrographically by A. A. Chodos.

CORRELATION

The metasedimentary rocks of the Dent area are structurally beneath the schist and diopside-plagioclase gneiss of the Headquarters quadrangle provided there is no major fault. The lithologic character of the formations (table 16) resembles, in many respects, that of a part of the Belt series on Surveyors Ridge (the upper part of the Prichard to the lower part of the Wallace formation in table 4). Much of the quartzite adjacent to the lime-silicate rock east of Dicks Creek is nearly pure quartz and has a granular texture. The grain size is coarser than that of the white granular quartzite of the Wallace formation on Surveyors Ridge, but this could be due to higher temperature and pressure during the metamorphism. The calcium-rich rocks near Dent occur as definite layers and lenticles; the calcareous material could have segregated from dolomitic and calcareous sands during the recrystallization. The schist that underlies the lime-silicate rock and quartzite is probably a metamorphosed equivalent of the St. Regis formation, and the coarse-grained quartzite with the glassy appearance (locs. 471 and 550, pl. 3) is most likely equivalent to the Revett quartzite. The schist under the Revett quartzite is considered to be equivalent to the upper part of the Prichard formation. Faulting may account for the absence of the Burke formation or its equivalent was not deposited in the Dent area.

PLUTONIC ROCKS

Two large bodies of quartz diorite, several bodies of gneissic tonalite, and a body of medium-grained light-colored massive tonalite are exposed in the Dent area. It is not clear whether any of the tonalite bodies with a gneissic structure should be considered to be intrusive, because most of similar tonalite in the Orofino area was found to be a product of feldspathization of biotite and hornblende gneiss. The genesis of the quartz diorite also is doubtful. A small body of gabbro occurs in the schist west of Dicks Creek.

PETROGRAPHIC DESCRIPTION

GABBRO

A few boulders of gabbro occur on a small meadow in the upper part of the northern canyon wall west of Dicks Creek. This gabbro is coarse grained and consists mainly of hornblende and plagioclase.

QUARTZ DIORITE

The quartz diorite in the southern part of the area along the river is coarse grained and gneissic. The dark minerals, hornblende and biotite, occur in clusters that are elongated parallel to the planar structure. In addition there are sporadic large angular hornblende crystals and inclusions of hornblendite. Remnants of medium-grained hornblende gneiss are common in the more gneissic parts of this quartz diorite. Microscopic study shows that a part of the plagioclase (An_{34}) in the quartz diorite occurs as large anhedral grains with fringed borders in a matrix of plagioclase, quartz, biotite, and hornblende with some epidote, sphene, apatite, and occasional microcline. The hornblende is strongly pleochroic with X=yellowish, Y=green, Z=dark green. The indices of refraction are $\alpha'=1.672\pm0.001$, $\gamma'=1.690\pm0.001$. Brown biotite with $\gamma=1.640$ occurs in small flakes with small round grains of quartz and plagioclase. Microcline when present is interstitial. Fairly large grains of epidote occur with hornblende. A few small grains of monazite are also present.

Smaller bodies of quartz diorite are medium grained and generally more gneissic, but in other respects they are similar to the large body.

GNEISSIC TONALITE

The largest body of gneissic tonalite is exposed along the old logging roads west of Elk Creek. This body consists of a medium-grained bluish-gray tonalite with abundant remnants of biotite-rich schist. The plagioclase (An_{34-37}) crystals are larger than those of the other constituents and generally show square or round cross sections on the rock surface. Quartz and the dark constituents, biotite and some hornblende, occur between the plagioclase crystals. Small round quartz grains are included in the plagioclase.

INTRUSIVE TONALITE

In contrast with the gneissic tonalite and quartz diorite the tonalite southeast of Dent (secs. 27 and 35, T. 38 N., R. 2 E.) shows all the characteristics of an intrusive body. It cuts the metamorphosed rocks discordantly and includes angular and rounded fragments of them. In several outcrops, random orientation of a platy structure—bedding or foliation—in these inclusions is apparent.

The minerals in this intrusive tonalite are plagioclase, quartz, and biotite, and some microcline, epidote, chlorite, rutile, sphene, zircon, and apatite; the texture is hypidiomorphic. Plagioclase occurs as long laths, blocky zoned crystals, and anhedral grains. Both simple and complex twins are common. Part of the quartz occurs as large round grains or groups of grains. The chemical composition of this tonalite (No. 379 pl. 3) has been discussed in connection with the other tonalites from the Headquarters quadrangle and vicinity (table 10).

STRUCTURE

PLANAR STRUCTURE

Most of the quartz diorite and gneissic tonalite show a planer structure due to the parallel orientation of biotite and hornblende, but this structure is rarely as pronounced as it is in the gneissic quartz diorite and tonalite of the Orofino area. The intrusive tonalite and a part of many other bodies are massive.

LINEAR STRUCTURE

Linear structure is less common than the planar structure. It could be measured in only a few outcrops and was there found to parallel the lineation in the metasedimentary rocks.

CONTACTS

Most contacts between the schist and the quartz diorite or gneissic tonalite tend to be concordant and gradational over a distance of about 1 to 10 m. In the schist near the contact, abundant large round or square grains of plagioclase are common, and the plutonic rock is generally richer in biotite along its margins than in the center of the body. The contact between the quartzite and the quartz dorite south of Dicks Creek is discussed in a later paragraph.

The contacts of the instrusive tonalite bodies with the country rocks are usually sharp and cut across the planar and linear structure of the latter. Concordant contacts were observed locally, but fragmentation of the wall rocks may occur also there. The coarse-grained light-colored tonalite at Dent has shattered its country rock on the north end of the body, but in the extreme southern end the contact is interfingering, showing locally concordant contacts and sending veinlets to the schist (fig. 71). At the round end of some "fingers" the schist is cut discordantly and the alinement of biotite flakes in the tonalite follows the contact.

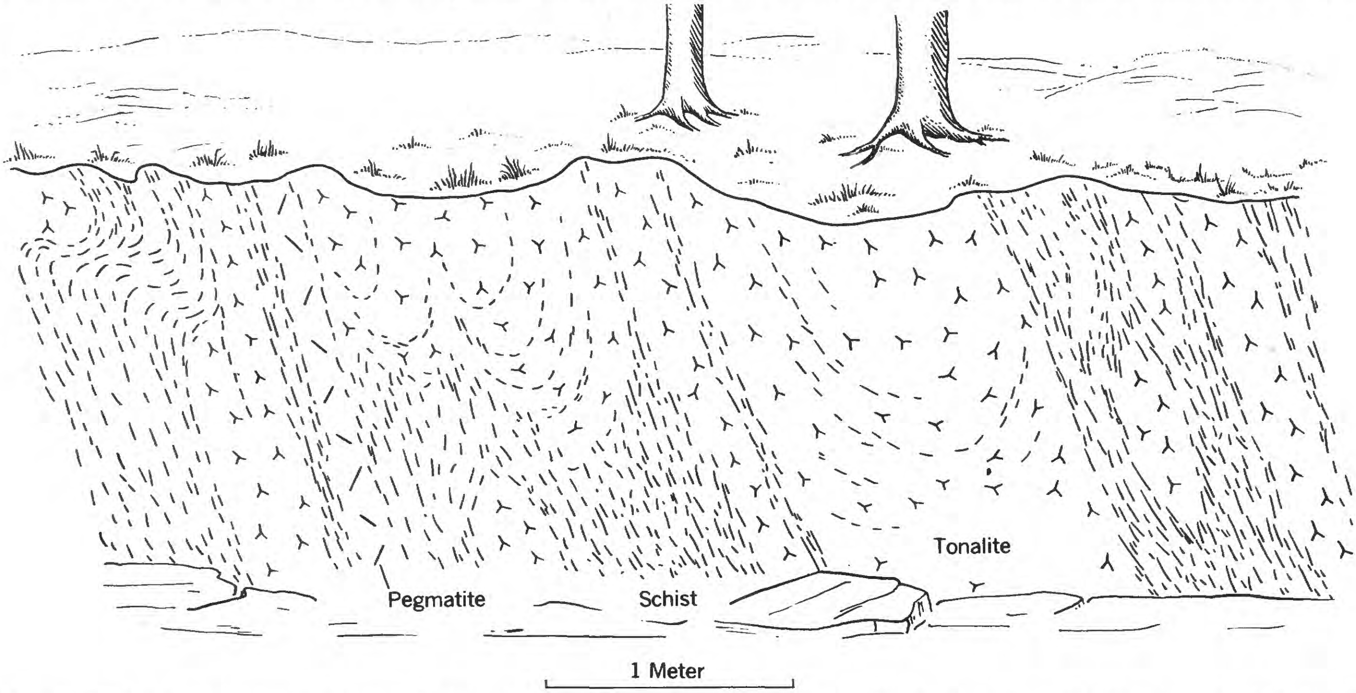


FIGURE 71.—Sketch of the contact between the intrusive tonalite and schist south of Dent. Note the curvature of a platy structure of the tonalite at the discordant ends of the lenses.

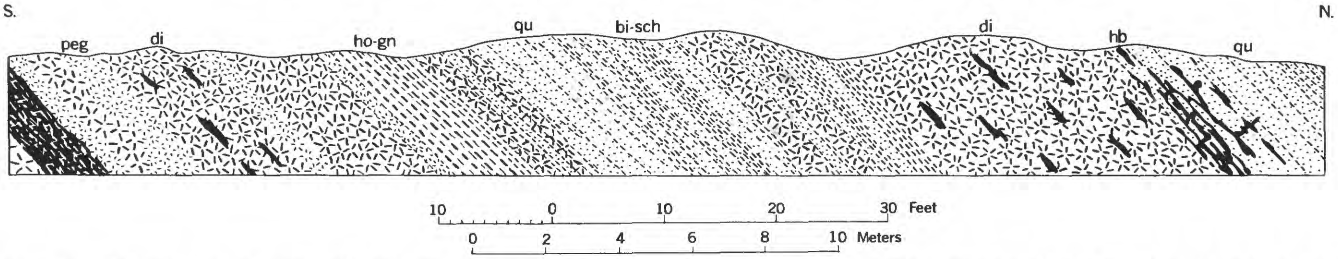


FIGURE 72.—The east face of a vertical road cut 1 mile south of the mouth of Dicks Creek shows metasomatic replacement of coarse-grained quartzite by hornblendite (at right), hb=hornblendite; di=diorite; ho-gn=hornblende gneiss; bi-sch=biotite schist; qu=quartzite; peg=pegmatite. Vertical scale is same as horizontal. (See fig. 73.)

Small pods of tonalite occur in the schist near the contact.

INCLUSIONS

Inclusions of the country rocks are common in all plutonic rocks, but there is a striking difference in their mode of occurence in the various rock types. In the quartz diorite and in the gneissic tonalite the inclusions are sheetlike bodies that are parallel to the planar structure and give an impression of being in their original position. In the intrusive tonalite the inclusions are angular fragments in which the planar structure—the bedding—is now oriented at random.

METASOMATIC CHANGES OF THE OLDER ROCKS

The mineralogy and chemical composition of the metasedimentary rocks change parallel to the bedding within short distances. This change is due to local replacement mainly of quartz by minerals such as hornblende, biotite, and calcic plagioclase. The composition of the resultant rock depends on the original composition of the parent rock, on the introduced elements, and on the extent of the exchange of material.

REPLACEMENT OF QUARTZ BY HORNBLENDE AND BIOTITE IN QUARTZITE SOUTH OF DICKS CREEK

Evidence of replacement of quartzite to form a mafic rock consisting mainly of hornblende, some biotite and very little plagioclase was observed in several localities in the Dent area. The best outcrop showing this type of replacement is a mile south of the mouth of Dicks Creek (loc. 471, pl. 3). Here a medium-grained diorite intrudes a thick-bedded coarse-grained quartzite that contains some layers of muscovite-biotite schist. Alternating layers of diorite, hornblende-biotite gneiss, quartzite, and schist are seen along the northern contact of this diorite (fig. 72). The replacement of quartzite by hornblende, biotite, and plagioclase occurs in the contact zone of the quartzite and the northernmost diorite lens (loc. 471).

The quartzite at locality 471 is coarse grained, appears glassy and bluish in most beds, and contains only a little albitic plagioclase and biotite. Thin biotite laminae separate the individual beds, the thickness of which ranges from 4 to 50 cm. The bedding dips about 45° N. Under the microscope the quartz appears strongly deformed, as shown by undulatory extinction and a preferred orientation. Small scattered plagioclase grains and biotite flakes are included in large quartz grains or occur between the grains. The plagioclase includes small zoisite and epidote crystals as alteration products. A few small rounded grains of zircon and euhedral grains of apatite are also included in the quartz.

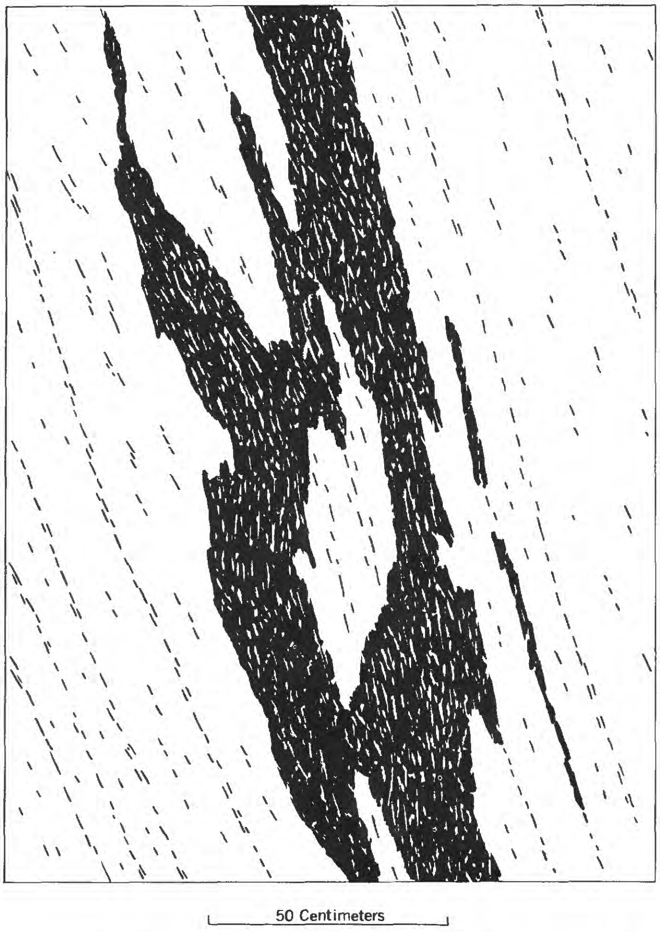


FIGURE 73.—Detailed drawing of the metasomatic replacement of quartzite (white) by hornblendite (black) along the contact of diorite and quartzite shown on the right side of figure 72.

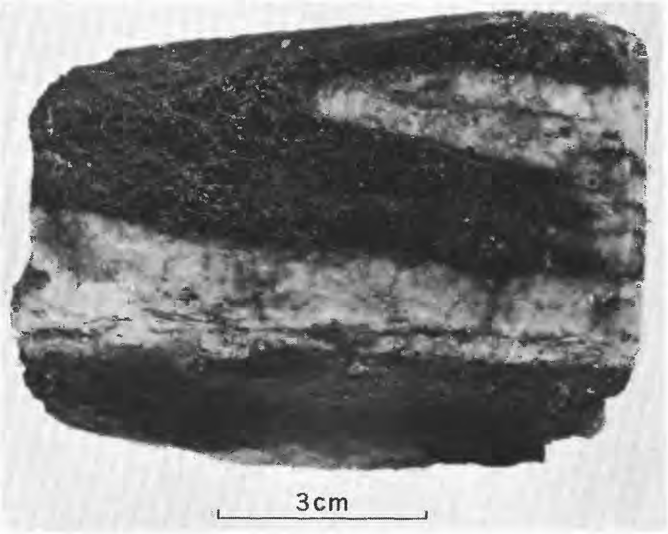
The quartz diorite at locality 471 is medium grained and has a planar structure parallel to the bedding of the quartzite. It consists of hornblende, plagioclase, some biotite, and accessories. It includes small lens-shaped bodies of hornblendite, which in places constitute as much as one-third of the rock. The minerals in the hornblendite inclusions are hornblende and some biotite, apatite, magnetite, and zircon. Discordant and concordant dikes of pegmatite cut the quartz diorite near the quartzite contact. Most of the quartz diorite is badly weathered.

Irregular masses, layers, and stringers of hornblendite occur in the quartzite on the northern contact zone of the northernmost quartz diorite body at locality 471.

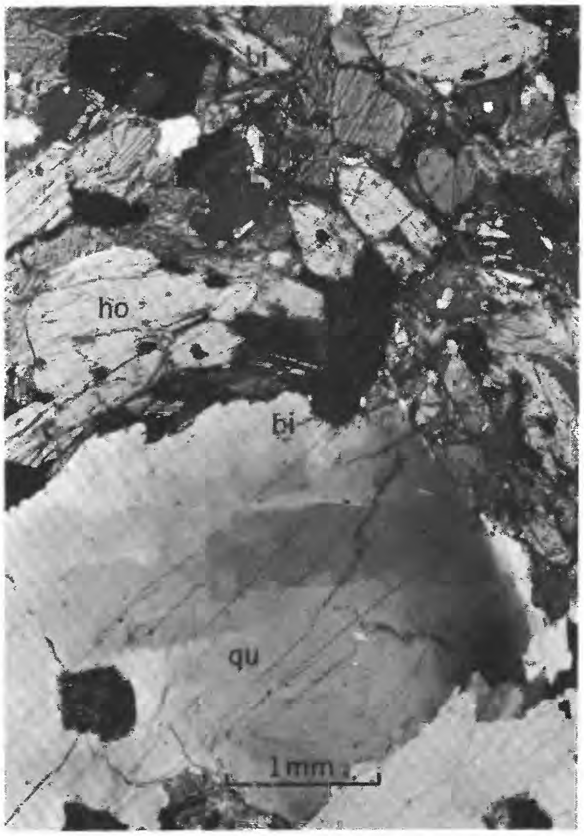
Most masses are elongated parallel to the bedding planes, and at their ends cut across it in an irregular manner. A cross joint filled with hornblendite for a distance of several tens of centimeters was found in the lower part of the road cut at locality 471. This hornblendite dikelet is about 3 cm thick, and it joins in its western end a main body which parallels the bedding. Several smaller stringers were found parallel to the fractures in the quartzite. The bedding planes—marked with rows of short lines in figure 73 and shown by parallel orientation of biotite flakes in figure 74*A*—continue undisturbed in the surroundings of hornblendite. The hornblendite bodies range in thickness from a few millimeters to 20 cm or more. The contacts between the hornblendite and quartzite seem sharp and straight when studied in the field. Thin sections show that actually the contact forms a zigzag line between large grains of quartz, biotite and hornblende (fig. 74 *B*).

Thin sections also show that the cracks in the quartzite bordering the hornblendite masses are filled with a brown material that contains individual biotite flakes. Magnetite occurs along the cracks and as subhedral grains in and between the quartz grains. Some individual hornblende grains and biotite flakes, which range in thickness from 0.25 to 0.5 mm and in length from 1 to 2 mm, separate the quartz grains or, more rarely, are included in the quartz grains. Hornblende with some biotite, magnetite, and apatite also forms small irregular patches in the quartzite.

The hornblendite consists chiefly of hornblende with a little biotite, magnetite, apatite, and zircon. The biotite flakes, which range from 1 to 5 mm in length, are along those surfaces that were originally structural planes in the quartzite. Hornblende grains are euhedral to subhedral and range from 1 to 3 mm in length. Relict quartz is frequently left between the grains. This quartz, like most of the quartz in the quartzite, still shows a strong undulatory extinction.



A. Polished hand specimen showing metasomatic replacement of coarse-grained quartzite by hornblendite. Road cut along the North Fork of the Clearwater River 1 mile south of the mouth of Dicks Creek (loc. 471, pl. 3).



B. Photomicrograph showing hornblende (ho) and biotite (bi) with some ilmenite (black) replacing coarse-grained quartzite (qu). Part of same specimen shown in *A*. Crossed nicols.

FIGURE 74.—METASOMATIC REPLACEMENT OF COARSE-GRAINED QUARTZITE

Small crystals of zircon with pleochroic halos are included in the hornblende and biotite. Zircon is more abundant in a narrow contact zone next to the quartzite than in the center of the hornblendite lenticles; small crystals occur also in quartzite. The centers of the thicker hornblendite masses contain enough plagioclase (An_{35}) to make this part of the rock gabbroic in composition.

The occurrence of the mafic material also parallel to the cross joints and fractures in the quartzite is strongly against the hypothesis that the hornblende rock would represent calcareous concretions or layers metamorphosed to amphibolite. As described in an earlier chapter, the calcareous layers and lenses of this region are metamorphosed to heterogeneous lime-silicate rocks that show quite different mineralogy and contact relations. The contacts of these calcareous beds and lenses with the surrounding quartzite are gradational, and abundant lime-silicate minerals are found also in the surrounding quartzite. No discordant structures between the lime-silicate rocks and surrounding quartzite were observed at any locality studied.

The hornblende rock and its mode of occurrence at locality 471 show a far greater resemblance to the rocks of igneous origin than to those of sedimentary origin. It is difficult, however, to explain where the quartz went from those small irregularly shaped parts of the strata that are now occupied by hornblendite. The material in the mafic lenses can well be of igneous origin, but it could not be injected forcibly into the quartzite, and parts of the quartzite could not have been carried away by the entering magma through narrow curved or nonexisting channels. It is, therefore, concluded that the emplacement of the hornblendite was a chemical rather than a physical process. The quartz in the place now occupied by hornblendite and gabbroic rock was apparently replaced by hornblende and plagioclase. The serious room problem is satisfactorily solved if it is considered that the emplacement was not a mechanical intrusion but was a metasomatic replacement during which the excess of silica was carried away in the solution.

Comparison of chemical analyses of a parent rock, the quartzite, and its metasomatic alteration product, the hornblendite, gives a quantitative picture of the extensive exchange of elements. Table 18 shows chemical analyses of the hornblendite (No. 471a) and the gabbroic center (No. 471b). The composition of the quartzite (No. 471) was calculated on the basis of a measured mode. Comparison of the analyses of quartzite and hornblendite shows that about 50 percent of the silicon was removed (column 4, table 18), and all the other oxides were introduced during replacement. Mineralogically, this means an introduction of components of hornblende, ilmenite, anorthite, and apatite and removal of quartz. When plagioclase is formed at the center of a body, sodium and aluminum are added and iron, titanium, and phosphorus are removed from the hornblendite. Ilmenite and apatite are thus concentrated toward the frontal zone of the basic rock.

Because replacement has taken place without change in volume, the exchange of substance in a unit volume should be shown. Barth (1948; 1952, p. 82) has pointed out that oxygen makes up about 94 percent by volume of most rocks and that, therefore, the number of oxygen atoms in a unit volume remains approximately constant. He has suggested for the standard of comparison a volume that contains exactly 160 oxygen ions and called it a standard cell of a rock. The number of cations in such a volume is approximately 100. Eskola (1954) suggested that 1-cation molecular percentage may substitute for Barth's standard cell. When the number of oxygen atoms bound to 100 cations is calculated, it will show an increase or decrease of oxygen atoms during the metasomatism. This is practical if the exchange of material is small or if there has been a change in volume. Near Dent the volume was not changed but instead the specific gravity of the rock increased considerably. (The specific gravity of the hornblendite is about 1.25 times that of the quartzite.) Table 19 shows the 1-cation molecular percentages and the corresponding number of anions in the quartize, hornblendite, and gabboric rock from locality 471. In the hornblendite (No. 471a), 147.11 oxygen atoms are bound to 100 cations, whereas the corresponding number in the quartzite is 194.25, thus considerably higher.

If we take the same volume, there are about 184 oxygen atoms in the hornblendite against the 194.25 oxygen atoms in the quartzite, thus only slightly fewer. Therefore, in this extreme case Barth's standard-cell calculation gives a better picture of the cation exchange during the metasomatism than does the ionic percentage alone. Columns 4 and 5 of table 19 show the decrease and increase of the number of cations in a standard cell of the rock during the replacement of the quartzite.

In the graphic representation (fig. 75) the numbers of ions of Fe, Mg, Ca, Al, K, Na, and Ti were plotted as ordinates and the numbers of corresponding Si ions as the abscissas. The graphs show that a considerable amount of Fe, Mg, Ca, Al, and some Ti, K and Na are introduced and Si removed when quartzite is replaced by hornblendite. The Ca, Al, K, and Na continue to increase, but parts of the Fe, Mg, and Ti are removed when hornblendite is transformed to a rock of gabbroic composition during a later phase of meta-

TABLE 18.—*Comparison of the chemical composition, norms, and minerals of quartzite, hornblendite, and gabbroic rock from 1 mile south of the mouth of Dicks Creek*

[Ruth H. Stokes, analyst, U.S. Geological Survey]

	1	2	3	4	5
Specimen no	471	471a	471b	Cations introduced (+) or removed (−) between Nos. 471 and 471a	Cations introduced (+) or removed (−) between Nos. 471a and 471b
Rock type	Quartzite (calculated from mode)	Hornblendite	Gabbroic rock		

Chemical composition

Constituent		Weight percent	Cation percent	Weight percent	Cation percent	Weight percent	Cation percent	Cation percent	Cation percent
Conventional symbol	Symbol rearranged for cation percent								
SiO_2		95. 15	94. 31	44. 78	43. 82	44. 45	42. 74	−50. 5	−1. 1
Al_2O_3	$AlO_{3/2}$	1. 81	2. 12	11. 13	12. 84	14. 52	16. 45	+10. 7	+3. 6
Fe_2O_3	$FeO_{3/2}$	. 18	. 13	3. 34	2. 46	3. 39	2. 45	+ 2. 3	. 0
FeO		. 87	. 72	14. 05	11. 49	12. 07	9. 70	+10. 8	−1. 8
MnO		. 01	. 01	. 28	. 23	. 22	. 18	+ . 2	. 0
MgO		. 55	. 81	8. 77	12. 78	8. 31	11. 90	+12. 0	− . 9
CaO		. 10	. 11	10. 08	10. 56	10. 29	10. 60	+10. 5	. 0
Na_2O	$NaO_{1/2}$	. 60	1. 15	1. 09	2. 07	1. 91	3. 56	+ . 9	+1. 5
K_2O	$KO_{1/2}$	. 41	. 51	. 59	. 74	. 84	1. 03	+ . 2	+ . 3
TiO_2		. 18	. 13	3. 63	2. 66	1. 75	1. 26	+ 2. 5	−1. 4
P_2O_5	$PO_{5/2}$		. 00	. 41	. 34	. 15	. 13	+ . 3	− . 2
CO_2				. 01	. 01	. 01	. 01		
H_2O+		. 15	(. 49)	1. 88	(6. 14)	1. 98	(6. 35)	(+5. 7)	(+ . 2)
H_2O-				. 16		. 23			
Total		100. 01	100. 00	100. 20	100. 00	100. 12	100. 01		
	O		194. 25		147. 11		145. 00		
	OH		. 98		12. 28		12. 70		
	Total anions		195. 23		159. 39		157. 70		

Norms

Mineral	C.I.P.W.	Molecular	C.I.P.W.	Molecular	C.I.P.W.	Molecular	Molecular	Molecular
Q	88. 53	87. 76					−87. 8	
Or	2. 39	2. 55	3. 50	3. 70	4. 95	5. 15	+ 1. 2	+ 1. 5
Ab	5. 09	5. 75	9. 23	10. 35	16. 15	17. 80	+ 4. 6	+ 7. 5
An	. 50	. 55	23. 72	25. 08	28. 56	29. 65	+24. 5	+ 4. 6
C	. 19	. 24					− . 2	
Wo			9. 82	9. 92	8. 94	8. 88	+ 9. 9	− 1. 0
En	. 89	1. 62	21. 39	25. 04	6. 51	7. 44	+23. 4	−17. 6
Fs	1. 79	1. 08	17. 20	15. 32	5. 30	4. 64	+14. 2	−10. 7
Fo			. 31	. 39	9. 93	12. 27	+ . 4	+11. 9
Fa			. 29	. 26	8. 95	7. 62	+ . 3	+ 7. 4
Ap			. 97	. 91	. 37	. 35	+ . 9	− . 6
Il	. 33	. 26	6. 87	5. 32	3. 32	2. 52	+ 5. 1	− 2. 8
Mt	. 25	. 19	4. 84	3. 69	4. 91	3. 67	+ 3. 5	. 0
Cc			. 02	. 02	. 02	. 02		
Subtotal	99. 96	100. 00	98. 16	100. 00	97. 91	100. 01		
$+H_2O$	. 15		2. 04		2. 21			
Total	100. 11		100. 20		100. 12			

Minerals calculated on basis of chemical analysis

Quartz	90. 0		9. 70		2. 60		−80. 3	−7. 1
Plagioclase	5. 0		1. 77		13. 50		− 3. 23	+ 11. 7
(Albite)	(4. 75)		(1. 01)		(7. 70)		(− 3. 74)	(+ 6. 7)
(Anorthite)	(. 25)		(. 76)		(5. 80)		(+ . 51)	(+5. 0)
Biotite	5. 0		6. 48		6. 48		+1. 48	. 0
Hornblende			74. 50		74. 32		+74. 50	− . 2
Ilmenite			5. 88		2. 36		+ 5. 88	−3. 5
Magnetite			. 88		. 49		+ . 88	− . 4
Apatite			. 97		. 35		+ . 97	− . 6
Total	100. 00		100. 18		100. 10			

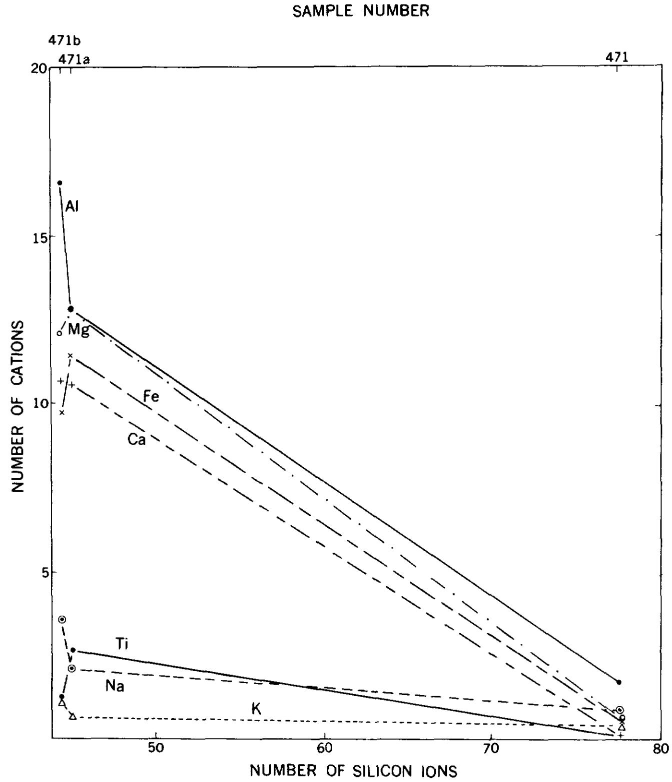


FIGURE 75.—Graph showing the number of cations in standard cells of quartzite (No. 471) and replacing hornblendite (No. 471a) and gabbroic rock (No. 471b).

TABLE 19.—*Comparison of the number of cations in a standard cell of quartzite, hornblendite, and gabbroic rock*

[See table 18 for analyses]

	1	2	3	4	5
Cation	471	471a	471b	Cations introduced (+) or removed (−) between nos. 471 and 471a	Cations introduced (+) or removed (−) between nos. 471a and 471b
	Number of cations in standard cell				
Si	77.3	44.0	43.4	−33.3	−0.6
Al	1.7	12.9	16.7	+11.2	+3.8
Fe‴	.1	2.5	2.5	+2.4	.0
Fe″	.6	11.5	9.8	+10.9	−1.7
Mn	.0	.2	.2	+.2	.0
Mg	.7	12.8	12.1	+12.1	−.7
Ca	.1	10.6	10.7	+10.5	+.1
Na	.9	2.1	3.6	+1.2	+1.5
K	.4	.7	1.1	+.3	+.4
Ti	.1	2.7	1.3	+2.6	−1.4
P	.0	.3	.1	+.3	−.2
C		.0	.0		
H_2O	(.4)	(6.2)	(6.4)	(+5.8)	(+.2)
Total	81.9	100.3	101.5		

somatism. If the introduction Na and Al continues, more plagioclase is crystallized and a dioritic rock is formed.

The main part of the quartz diorite just south of the coarse-grained quartzite may have been formed in this manner. The numerous lenses of hornblendite that occur as inclusions in this quartz diorite may represent remnants of a more basic rock that first replaced the quartzite. The sequence of crystallization was as follows: First, biotite and some magnetite, apatite, and possibly zircon, was crystallized, and immediately after it hornblende and more apatite and magnetite. This suggests that Fe, Mg, Al, K, P, and possibly Zr were introduced early. The second phase in the metasomatism gave rise to the development of plagioclase as individual grains or small groups of grains or as irregular stringers. First, a hornblendite was formed; then during the second phase a gabbroic rock. As more components of plagioclase were introduced, a quartz diorite or a tonalite was formed. The metasomatic biotite lies roughly parallel to the bedding and cleavage of the quartzite; also, many hornblende crystals have their longest dimensions parallel to the same planes, lending to the dioritic rock an irregular planar structure parallel to the bedding of the quartzite.

The coarse-grained quartz diorite in the southern part of the area shown on plate 3 is fairly heterogeneous. It contains large scattered hornblende crystals, hornblendite inclusions, and remnants of older sedimentary rocks—fine-grained quartz-plagioclase-biotite gneiss, and quartzite. Scattered medium-sized grains of hornblende were developed in the sedimentary inclusions. In places a lighter colored medium-grained hornblende-bearing gneissic granodiorite occurs as irregular veins in this coarse-grained quartz diorite. These granodioritic parts are also heterogeneous and contain remnants of metasedimentary rocks, resembling the partly granitized trondhjemites and kinzigites from the streets of Turku in Finland (Hietanen, 1947). The coarse-grained quartz diorite itself resembles the heterogeneous dioritic contact rocks of the granodiorite plutons in the Sierra Nevada (Hietanen, 1951, p. 592–594). There the material for the large hornblende crystals originated in the mafic country rocks. In the area under discussion the components of hornblende were probably introduced, because the older rocks are mainly siliceous metasediments with subordinate amounts of hornblende-bearing schist and gneiss. The scattered large hornblende crystals probably represent remnants of the earlier phase of metasomatism, the phase during which iron and magnesium were introduced, and which preceded the introduction of components of plagioclase. This view is supported by the occurrence of the hornblende inclusions in the same rock. The second phase of metasomatism, the introduction of alkalies, aluminum, and silicon, gave rise to further crystallization of andesine, changing the basic rock to hornblende-biotite-quartz diorite. Still later, coarse-grained granodiorite veins were formed by local introduction of potassium into this metasomatic quartz diorite. Thin-section study shows that the potassium feldspar that crystallized during the potassium metasomatism is microcline and occurs as small grains and films between the other minerals and occasionally as large holoblasts including the other minerals.

SEGREGATION OF HORNBLENDE, BIOTITE, AND CALCIC PLAGIOCLASE IN SCHIST

Numerous rather small dark lenticles consisting mainly of hornblende, biotite, and occasionally andesine or labradorite are included in biotite-garnet schist along the road cuts near Dent and along the North Fork of the Clearwater River between Big Island and the mouth of Elk Creek. Most of these lenticles range from 1 to 5 m in diameter. Locally the schist also contains many large rounded scattered grains of andesine or veins consisting of almost pure andesine, but it contains no quartz even though quartz is abundant in the unaltered parts of the same schist. Plagioclase in the ordinary schist is oligoclase, and it occurs only sparingly. It seems, therefore, that quartz in the schist is locally replaced by the newly crystallized minerals hornblende, biotite, and plagioclase and that this replacement occurred in all formations exposed near Dent.

MAFIC LENSES IN THE SCHIST

Lens-shaped bodies consisting of dark constituents are especially common in the schist near Dent. Figure 76 is a photomicrograph of such rock. The older rock in this area was probably a carbonate-bearing layer in a biotite-garnet-sillimanite schist similar in composition to the hornblende schist half a mile to the east. This hornblende schist is medium grained, has a lepidoblastic texture, and consists of quartz, plagioclase, hornblende, and biotite. Some quartz and calcite still occur between the biotite flakes and hornblende grains in the mafic lens, which is coarse-grained and in which minerals are oriented at random. Crystals of allanite, as much as 1 cm in diameter and small crystals of zircon are included in the hornblende and biotite. Both the allanite and the zircon crystals are surrounded by pleochroic halos. Abundant magnetite and ilmenite grains are included in the other minerals. The alteration of garnet to chlorite suggests a lowering of temperature toward the end of mineral development.

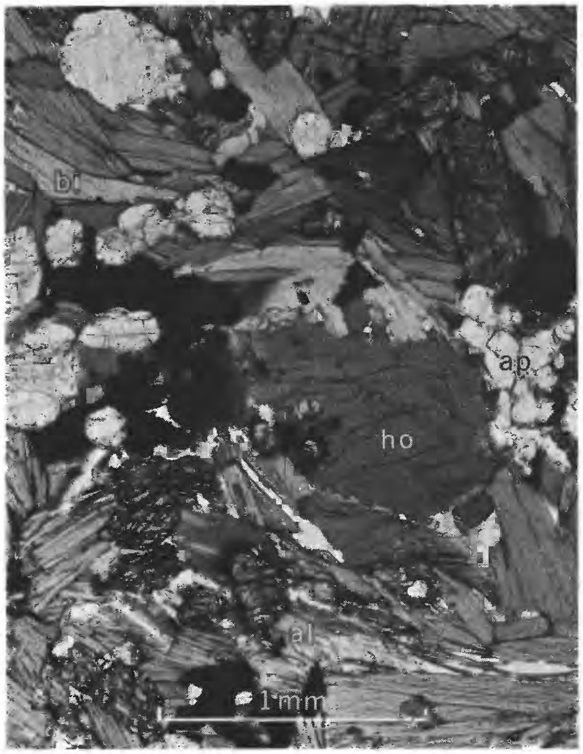


FIGURE 76.—Photomicrograph of a mafic lens from the feldspathized schist 1 mile east of the mouth of Elk Creek on the north side of the North Fork of the Clearwater River (loc. 119, pl. 3). Minerals: ho=hornblende; bi=biotite; ap=apatite; al=allanite; black=ilmenite. Plane polarized light.

TABLE 20.—*Chemical composition, ionic percentage, norms, and minerals of a mafic lens from the feldspathized schist along the North Fork of the Clearwater River (loc. 119)*

[Ruth H. Stokes, analyst, U.S. Geological Survey]

Constituent		Weight percent	Cation percent	Norm			Minerals	
Conventional symbols	Symbols rearranged for cation percent				CIPW	Molecular		
SiO_2		37.49	37.46	Or	14.41	15.55	Quartz	8.4
Al_2O_3	$AlO_{3/2}$	14.45	17.02	Ab	4.56	5.20	Biotite	27.3
Fe_2O_3	$FeO_{3/2}$	1.88	1.42	An	22.67	24.40	Hornblende	33.8
FeO		18.23	15.23	C	2.62	3.11		
MnO		.10	.08	En	7.48	9.08	Chlorite[1]	16.1
MgO		8.24	12.27	Fs	8.09	7.40	Ilmenite	9.9
CaO		6.04	6.46	Fo	9.12	11.59	Apatite	1.1
Na_2O	$NaO_{1/2}$	.54	1.04	Fa	10.72	9.45	Allanite	1.4
K_2O	$KO_{1/2}$	2.44	3.11	Ap	1.11	1.07	Calcite	1.5
TiO_2		6.15	4.60	Il	11.64	9.20	Zircon	.5
P_2O_5	$PO_{5/2}$	.47	.40	Mt	2.73	2.13		
CO_2		.67	.91	Cc	1.52	1.82		
H_2O+		3.20	(10.66)	H_2O	3.29			
H_2O-		.09						
Total		99.99	100.00		99.96	100.00		100.0
	O		140.06					
	OH		21.32					
	Total anions		161.38					

[1] Garnet→Chlorite.

The chemical and mineralogical composition of the mafic rock at locality 119 is shown in table 20. Comparison with the hornblendite no. 471a (table 18) shows that there is more biotite and less hornblende in the mafic rock (no. 119). Generally the ratio of biotite to hornblende in individual occurrences is highly variable.

Several lens-shaped and dikelike bodies of gabbroic aspect were found in the same area near the river and at a road cut north of locality 13 on plate 3. Thin-

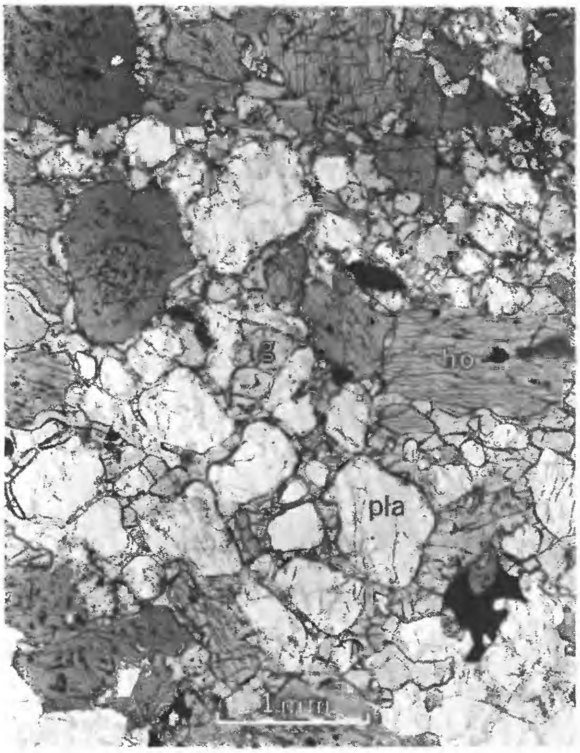


FIGURE 77.—Gabbroic rock with abundant garnet (g) between the other minerals, rounded plagioclase (white) and subhedral hornblende (ho). Ilmenite (black) occurs as grains among the other minerals and as small scaly inclusions in hornblende. One mile east of the mouth of Elk Creek along the North Fork of the Clearwater River (loc. 13, pl. 3). Plane polarized light.

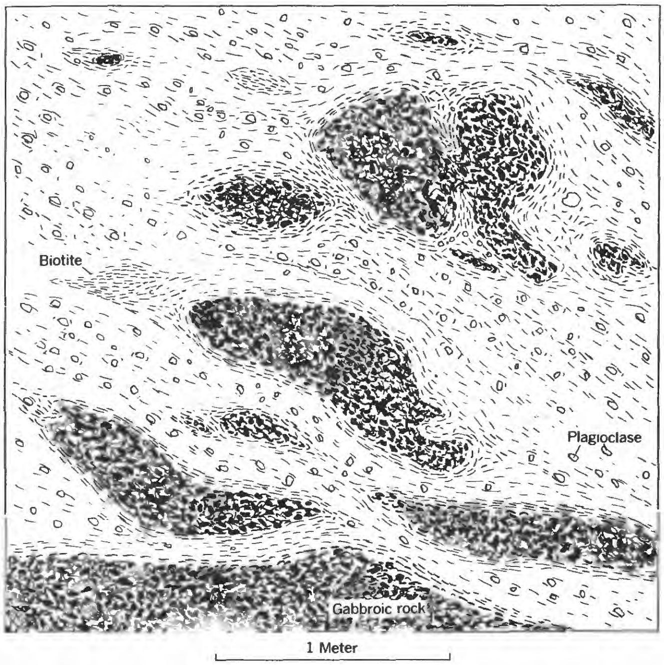


FIGURE 78.—Mafic lenses surrounded by biotite occur in a feldspathized schist on the south face of a road cut north of locality 13 (pl. 3).

section study of a lens exposed at the river shows that the rock contains numerous large garnet crystals and that the texture is crystalloblastic (fig. 77). Hornblende occurs as long subhedral prisms that include abundant small ilmenite scales. The plagioclase (An_{65}), which is more anorthitic than in other rocks of the area, occurs as rounded grains that contain sericite along the cracks. Parallel fractures transect all the feldspar grains but not the other minerals, giving an impression that the labradorite grains are older than the hornblende and garnet. Garnet fills the interstices between the plagioclase and hornblende. Actually garnet occurs as large holoblasts that include the rounded plagioclase grains and hornblende prisms. Biotite is scarce and occurs between the plagioclase and hornblende. Rounded magnetite grains and ilmenite scales are included in the hornblende. This texture is much like that of the lime-silicate rock, and it is possible that some of the lenses were dolomitic shale or calcareous concretions.

The occurrence of dikelike bodies of medium-grained garnetiferous gabbroic rock in a road cut just north of locality 119 indicates that some of the bodies may be igneous. A similar association of garnet, hornblende, and calcic plagioclase is common in the contact zones between quartz diorite and biotite-garnet schist on Benton Butte (pl. 1). In these contact zones however, elements that gave rise to the crystallization of hornblende and calcic plagioclase were introduced into the country rock, and the small garnet crystals which are abundant in the schist were recrystallized as large holoblasts. Comparison of the mafic lenses with these contact rocks suggests a metasomatic origin also for the mafic lenses near Dent. At both localities the host rock is a garnet-biotite schist, which does not contain calcareous concretions elsewhere in the area under investigation.

Small gabbroic lenses occur in a biotite-plagioclase schist on a road cut north of locality 13 (fig. 78). These lenses are enveloped by biotite layers 1 to 20 mm thick. The enveloping biotite is coarser grained than the biotite in the surrounding schist. Most of the gabbroic lenses are elongated parallel to the schistosity of the country rocks, but some irregularly shaped masses have their longest dimension at an angle to that structure and some appear as dikelike bodies. Plagioclase in the surrounding schist is andesine, An_{43}, and occurs as tabular to rounded large crystals, as secondary plagioclase in this area usually does. Some large grains are also found in the biotite accumulations at the ends of the gabbroic lenses. No good sample of schist that had not been feldspathized was found in any of the formations near Dent. Fairly unaltered sillimanite-garnet schist is exposed on the south side of the river southeast of Dent, along Dicks Creek, and along Elk Creek west and north of Dent. The schists at these localities are much alike and similar to the sillimanite-garnet schist described from the Headquarters quadrangle (No. 247 in table 2). The major constituents are quartz, biotite, muscovite, sillimanite, and garnet, and ilmenite and zircon occur as accessories. Layers that contain oligoclase are interbedded. These layers are mineralogically similar to the banded biotite-plagioclase schist with muscovite-sillimanite laminae in the Headquarters quadrangle (No. 246, table 2). No calcareous concretions were found in any of the unaltered schist.

The weathered schist on some road cuts near Dent is of the same general type as the unaltered schist just described, and in the areas farther north all sillimanite-garnet schist, regardless of the formation to which it belongs, is much alike. To illustrate the possible exchange of material during the crystallization of the mafic lenses in the schist, their chemical composition was therefore compared with the analysed specimens of the schist from the Headquarters quadrangle (Nos. 246 and 247 in table 2). The schist just west of Dent resembles more closely the coarse-grained sillimanite-garnet-schist No. 247, but because the plagioclase-bearing layers (No. 246) are intercalated with this coarse-grained schist west of Dent, the comparison was extended to include these layers in order not to exaggerate the increase in the amount of plagioclase.

Comparison of the chemical composition of the schist and the mafic lenses as deduced from their mineral composition, shows about 20 percent less SiO_2 and about 2 percent less K_2O in the mafic rock than in the schist. In the mafic rock the amounts of Al_2O_3, FeO, MgO, and CaO are considerably larger (4 to 9 percent). If the mafic lenses are considered to be formed by metasomatic replacement of the schist, a considerable amount of silicon must have been removed and mainly Fe, Mg, Ca, and Al introduced. The removed potassium probably migrated into surrounding schist and reacted with other elements to form biotite. Thus, here, as in the replacement of the coarse-grained quartzite by hornblendite, biotite may have formed a front that moved ahead of hornblende and calcic plagioclase.

The schist about a mile north of locality 13 also shows alterations. Quartz is very scarce; biotite and garnet have altered to chlorite. Abundant muscovite occurs either as large flakes parallel to the sheer planes or as small flakes that are oriented at random between these shear planes. Similar schist elsewhere in the region contains only a little muscovite but abundant sillimanite. Illemite, rutile, and zircon abound in this schist. Thus here K, Fe, Mg, Ti, and Zr were introduced and Si was removed. This zone is chemically comparable to the biotite-rich "front" in the metasomatism of the coarse-grained quartzite south of the mouth of Dicks Creek, even if it differs mineralogically from it.

Some other small bodies of mafic rocks 4 miles east of Dent (just west of loc. 190, pl. 3) contain large grains of poikilitic bytownite (An_{85}) and abundant recrystallized garnet that includes other minerals. Magnetite, apatite, and zircon abound also in these bodies, and rutile and ilmenite occur in small amounts.

ANDESINE IN THE SCHIST

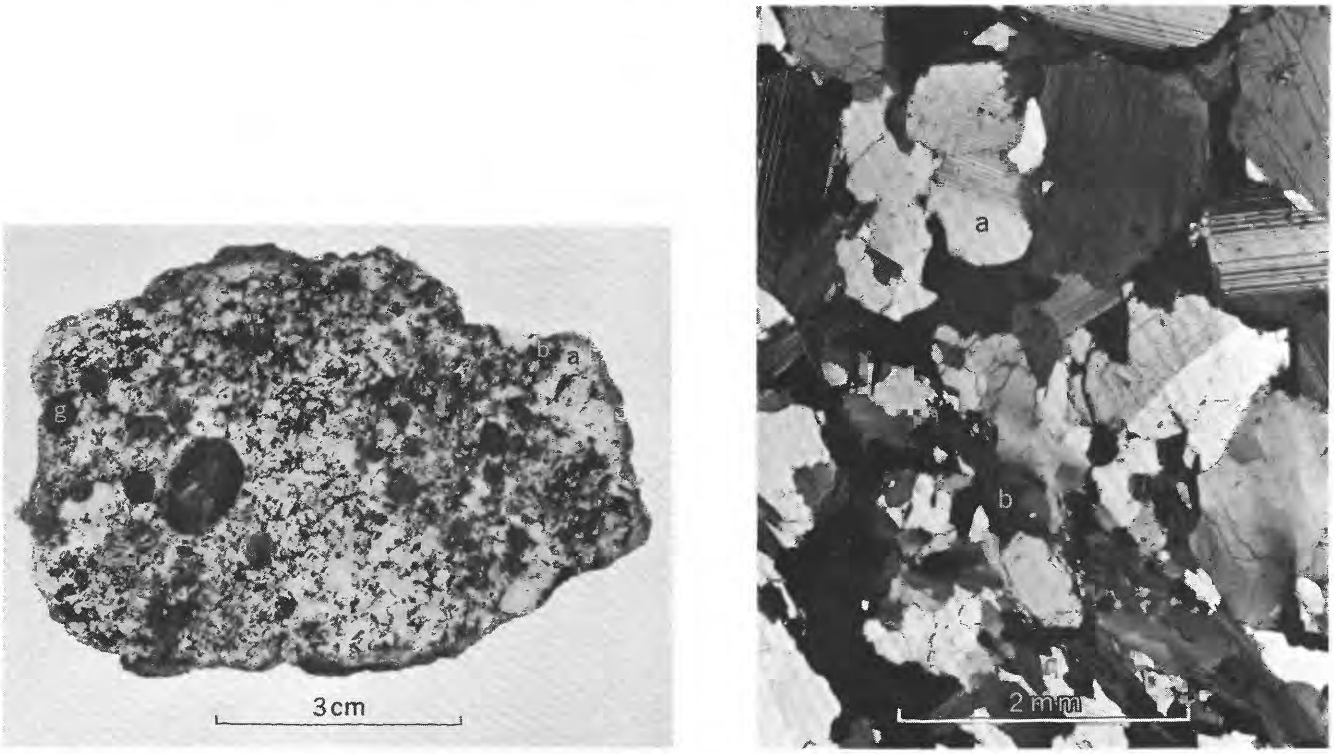
Much of the schist along the river canyon at Dent and toward the east—regardless of the formation to which it belongs—contains large white grains or clusters of smaller grains of andesine. In most outcrops the andesine grains are round, but in some localities, for instance northeast of the mouth of Elk Creek, they are square. Dark biotite flakes usually surround these white andesine grains which may reach 6 mm in diameter. The number of the andesine grains varies greatly; the outcrops about half a mile west of Dent contain from 40 to 60 percent andesine, whereas some parts of the same schist along the Elk Creek contain only a few grains of oligoclase. A similar relation was observed east of Dent. The outcrops on the northern canyon wall are rich in andesine, but those on the southern wall contain much less or locally none of this mineral. The part of the rock that is rich in andesine contains lenses and layers of schist in which only a few sporadic grains of andesine occur with quartz.

Study under the microscope shows that the additional constituents in the schist and in the andesine-bearing rock are the same. Biotite is the main dark constituent in the andesine-bearing rock also, and garnet abounds in some layers. Sillimanite, staurolite, apatite, brown rutile, and ilemenite are common minor constituents.

The texture of the andesine-bearing rock differs greatly from that of the schist. The salient textural features of the andesine-bearing rocks are the large size and round shape of the andesine grains. The dark biotite flakes surround the andesine grains and the minor constituents occur with biotite in the interstices (fig. 79). Where garnet abounds, it occurs as subhedral gains as much as 1 cm in diameter (fig. 80*A*). The inclusions of schist in the andesine rock contain abundant quartz and show a well-foliated texture contrasting with the texture of the andesine rock (fig. 80*B*). An abrupt change between the schist and andesine-bearing rock was observed in many outcrops (fig. 81). These observations proved that the distribution of the andesine was not confined to certain beds in the schist

FIGURE 79.—Camera lucida drawing of the plagioclase-biotite-garnet rock (No. 13) near Dent. Recrystallized biotite and the minor constituents occur between the large rounded plagioclase (An_{50}) grains. pla=plagioclase; bi=biotite; g=garnet; ky=kyanite; sill=sillimanite; st=staurolite; il=ilmenite; ru=rutile; ap=apatite; cal=calcite; zr=zircon; se=sericite. Location 13 on pl. 3.

but was rather irregular, cutting across the bedding and resembling an occurrence of igneous material. The mineral assemblage and the heterogeneity, however, prove that these rocks are not igneous. The minerals are much the same as in an unaltered schist except that quartz has been replaced by andesine. Also the unaltered portions and inclusions are still in their original positions, in contrast to the disturbed position of inclusions in intrusive tonalite. The only possibility seems to be, therefore, that the crystallization of andesine is due to the introduction of certain elements into the schist. The schistose and folded structure of the parent rock is usually obliterated where abundant late plagioclase crystallized (figs. 80*A*, *B* and 81). This shows that the metasomatism which gave rise to the crystallization of andesine was postfolding. In many outcrops all quartz has been replaced by andesine, and the other constituents—biotite, garnet, and sillimanite—were recrystallized. The recrystallized biotite flakes occur between round grains of plagioclase, being



A. Polished hand specimen of the metasomatic plagioclase (An_{50})-biotite-garnet rock from 1 mile east of the mouth of Elk Creek (loc. 13, pl. 3). All light-colored minerals are andesine (a); the dark round grains are garnet (g); biotite (b) occurs between the other minerals.

B. Photomicrograph showing secondary large round andesine grains in a fine-grained biotite schist along the North Fork of the Clearwater River near Dent (loc. 196, pl. 3). Large round plagioclase grains shown in top part of photomicrograph have crystallized late and disturbed the original foliated structure of the schist. a=andesine, b=biotite, q=quartz. Crossed nicols.

FIGURE 80.—SAMPLES OF FELDSPATHIZED SCHISTS.

oriented also across the foliated texture of the original schist (fig. 79). In the original schist the biotite plates are well oriented parallel to the planes of foliation between rows of elongated quartz grains (cf. figs. 3 and 4). Sillimanite in the metasomatized schist has recrystallized as larger prisms with a diagonal cleavage in their cross sections. Garnet in the feldspathized schist is mainly almandite, with 20 percent pyrope (table 21) and shows $n=1.804\pm0.001$.

TABLE 21.—*Chemical analysis of garnet from the plagioclase-biotite-garnet rock No. 13*

[Harry H. Hyman, analyst, U.S. Geological Survey]

Constituent	*Weight percent*	*Molecular equivalent*	*Composition*	
SiO_2	37.88	6307		
Al_2O_3	21.60	2119		
Fe_2O_3	1.10	69	Pyrope	20.4
FeO	31.68	4410	Almandite	72.3
MnO	.79	111	Spessartite	1.8
MgO	5.02	1245	Grossularite	2.1
CaO	1.88	335	Andradite	3.4
Na_2O	.00	----------		
K_2O	.02	----------		
TiO_2	.15	19		
H_2O+	.02			
H_2O-	.02			
Total	100.16			

The chemical and mineralogical composition of the metasomatic plagioclase-biotite-garnet rock No. 13 is shown in table 22. Comparison of this chemical analysis with that of two typical layers in ordinary biotite-garnet-sillimanite schist (table 2, Nos. 246, 247) suggests that considerable amounts of material were removed and introduced when the biotite-garnet-sillimanite schist was metasomatically altered to rock No. 13 (table 23). Changes in the amounts of silicon, aluminum, calcium, and sodium are due to the replacement of quartz by plagioclase. Larger amounts of titanium and iron are due to larger amounts of ilmenite and rutile. It is impossible to tell whether the main difference in the amounts of iron and magnesium are due to a replacement of muscovite by garnet or whether more garnet was originally present in the parent rock. All garnet in specimen 13 is recrystallized, and the amount of muscovite and garnet in the schist in general varies from one layer to the next. An exceptionally large amount of biotite and chlorite in the schist exposed 1 mile north of locality 13 indicates that iron and magnesium were introduced there, and one might expect

FIGURE 81.—The folded structure of the schist is obliterated in places by the metasomatic development of hornblende and plagioclase. Location 197, along the North Fork of the Clearwater River at Dent, plate 3.

TABLE 22.—*Chemical composition, ionic percentage, norms, and minerals of a plagioclase-biotite-garnet rock No. 13*

[Ruth Holzinger, analyst, U.S. Geological Survey]

Constituent		Weight percent	Cation percent	Norm			Minerals	
Conventional symbol	Symbol rearranged for cation percent				CIPW	Molecular		
SiO_2		43.24	41.13	Q	--------	--------	Plagioclase	39.8
Al_2O_3	$AlO_{3/2}$	23.77	26.65	Or	12.13	12.45	(Anorthite content)	(50)
Fe_2O_3	$FeO_{3/2}$	.88	.63	Ab	18.77	20.45	Biotite	21.5
FeO		12.76	10.15	An	23.36	24.00	Chlorite	4.5
MnO		.18	.14	C	9.34	10.47	Garnet	24.4
MgO		4.52	6.40	En	6.76	7.76	Sillimanite	4.7
CaO		5.63	5.74	Fs	11.53	10.04	Staurolite	1.7
Na_2O	$NaO_{1/2}$	2.22	4.09	Fo	3.15	3.78	Apatite	.1
K_2O	$KO_{1/2}$	2.05	2.49	Fa	5.95	4.89	Rutile	1.2
TiO_2		2.31	1.65	Ap	.03	.03	Ilmenite	.7
P_2O_5	$PO_{5/2}$	.02	.01	Il	4.37	3.30	Calcite	1.6
CO_2		.71	.92	Mt	1.27	.99		
H_2O+		1.67	(5.30)	Cc	1.61	1.84		
H_2O-		.18	--------	H_2O	1.85	--------		
Total		100.14	100.00		100.12	100.00		
	O		148.30					
	OH		10.60					
	Total anions		158.90					100.2

that these elements were also introduced into rock No. 13. If they were, this introduction preceded the introduction of sodium, calcium, and aluminum because plagioclase is the latest mineral.

The number of cations in standard cells of these rocks (table 23) are shown graphically in figure 82. The two layers in the unaltered schist, though fairly different mineralogically, show about the same cation contents except that more aluminum and less silicon occur in No. 246 than in No. 247. During the metasomatism mainly iron, magnesium, calcium, and aluminum were introduced and silicon removed. There was only a slight addition of sodium and removal of potassium.

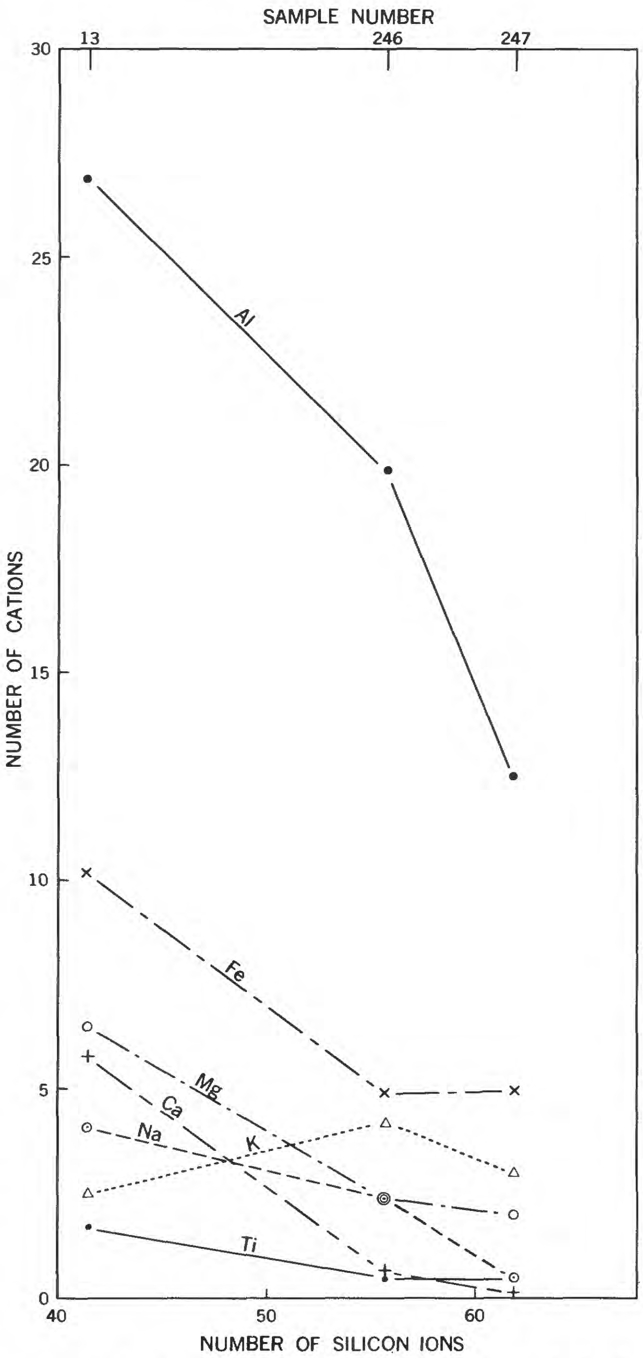


FIGURE 82.—Graph showing the number of cations in standard cells of schists (nos. 246 and 247) and metasomatic plagioclase-biotite-garnet rock (No. 13).

A large amount of secondary plagioclase (An_{40-45}) and some hornblende has crystallized in the schist between Dent and Big Island. Many of the outcrops

TABLE 23.—*Comparison of the number of cations in a standard cell of a plagioclase-biotite-garnet rock No. 13 and the possible parent rocks*

	1	2	3	4	5
Cation	246 (table 2)	247 (table 2)	13 (table 22)	Cations introduced (+) or removed (−) between Nos. 246 and 13	Cations introduced (+) or removed (−) between Nos. 247 and 13
	Number of cations in standard cell				
Si	55.7	61.9	41.4	−14.3	−20.5
Al	19.8	12.5	26.8	+7.0	+14.3
Fe'''	.7	.6	.6	−.1	.0
Fe''	4.9	5.0	10.2	+5.3	+5.2
Mn	.1	.1	.1	.0	.0
Mg	2.4	2.0	6.5	+4.1	+4.5
Ca	.6	.1	5.8	+5.2	+5.7
Na	2.4	.5	4.1	+1.7	+3.6
K	4.2	3.0	2.5	−1.7	−.5
Ti	.5	.5	1.7	+1.2	+1.2
C	.0	.0	.9	+.9	+.9
H_2O	(5.4)	(7.1)	(5.3)	−.1	−1.8
Total	91.3	86.2	100.6		

along the river give excellent material for studying the distribution of these minerals and the disappearance of earlier folded structure, as shown on figure 81. The oldest rock in this locality (loc. 197) is a fine-grained thin-bedded biotite schist tentatively correlated to the Wallace formation of the Belt series. This schist consists mainly of quartz and biotite, with some plagioclase. Many layers contain large muscovite flakes, some of which grow across the bedding and are later than the other minerals. Small rounded grains of zircon abound, and ilmenite-magnetite is common. In a few places, late plagioclase forms veins parallel to the bedding planes, and the resultant rock resembles veined gneiss (migmatite); elsewhere large plagioclase grains are evenly distributed throughout the rock, as seen on the right-hand side of figure 81. On the left, in the same illustration, plagioclase-quartz veinlets are parallel to the bedding in the schist. These veinlets are probably older than the large grains of plagioclase on the right. The original schist contains only a few grains of oligoclase among the quartz grains, whereas the feldspar in the metasomatic rock is calcic andesine, and quartz has disappeared. Figure 80*B* shows a photomicrograph of a small schist inclusion in similar metasomatic rock (No. 196). The alinement of biotite is easily seen in the schist, but is obliterated where the large round andesine grains replace the quartz. Medium-grained light-colored tonalite cuts the fine-grained schist and its metasomatic alteration products discordantly (fig. 83) and is younger than the development of calcic plagioclase.

ANORTHOSITIC VEINS

Plagioclase of the same composition (An_{50}) as that developed as individual grains in the schist also forms veins and dikes in the metasomatic rock No. 13. Under the microscope the plagioclase grains in the dikes appear elongated and somewhat rounded. Complex twinning is common, if not dominant. The minor constitutents are biotite, chlorite, sericite, sphene, ilmenite, magnetite, apatite, allanite, epidote, and zircon. The amount of accessories in these veins is less than that in the metasomatically altered schist (No. 13) and much less than that in the basic lenses (No. 119). This suggests that the main part of the P, Ce, and Zr were concentrated with the Fe and Mg ions during the metasomatism. Titanium forms ilmenite or rutile when associated with Fe and Mg, but it also forms sphene when Ca, Na, and Al are introduced. There is a notable amount of sphene in an anorthosite vein and in metasomatic diorite and tonalite bodies in the surrounding areas. Sphene is especially abundant in lime-silicate rock, where it frequently forms euhedral yellow crystals that measure 1 cm in diameter.

RELATION OF ANDESINE TO FEMIC CONSTITUENTS

At Dent the hornblende was crystallized earlier than the plagioclase—the same relation that was found south of Dicks Creek. However, in contrast with the formation of gabbroic or dioritic rocks near Dicks Creek, the dark constituents—hornblende and biotite—became segregated from the plagioclase near Dent. Hornblende and biotite with abundant magnetite, ilmenite, allanite, apatite, and zircon are segregated in small masses; the plagioclase (An_{50}) is either evenly distributed throughout the schist, replacing quartz, or is segregated in veins and small masses. Apparently the same solution that carried iron and magnesium also brought in calcium, but there was a differentiation during the metasomatism. A similar differentiation probably occurred in the country rock of anorthosite in the northernmost part of the area shown in plate 1.

The anorthosite bodies, three of which are about 10 miles long, show replacement textures, relicts of bedding and folding of the parent metasediment (interbedded limestone and aluminous shale?) and contain abundant metamorphic minerals like kyanite, andalusite, sillimanite, garnet, and staurolite (Hietanen, 1956). The schist next to the anorthosite contains abundant calcic plagioclase, and encloses segregations of biotite and bodies of garnet amphibolite. It seems that differentiation similar to that which took place on a small scale near Dent occurred on a larger scale in the country rocks of the anorthosite.

The dioritization described from the Orofino district is chemically and mineralogically similar to the basification in the Dent area. But in contrast to the Dent area the major part of the femic constituents in the Orofino area were deposited together with the plagio-

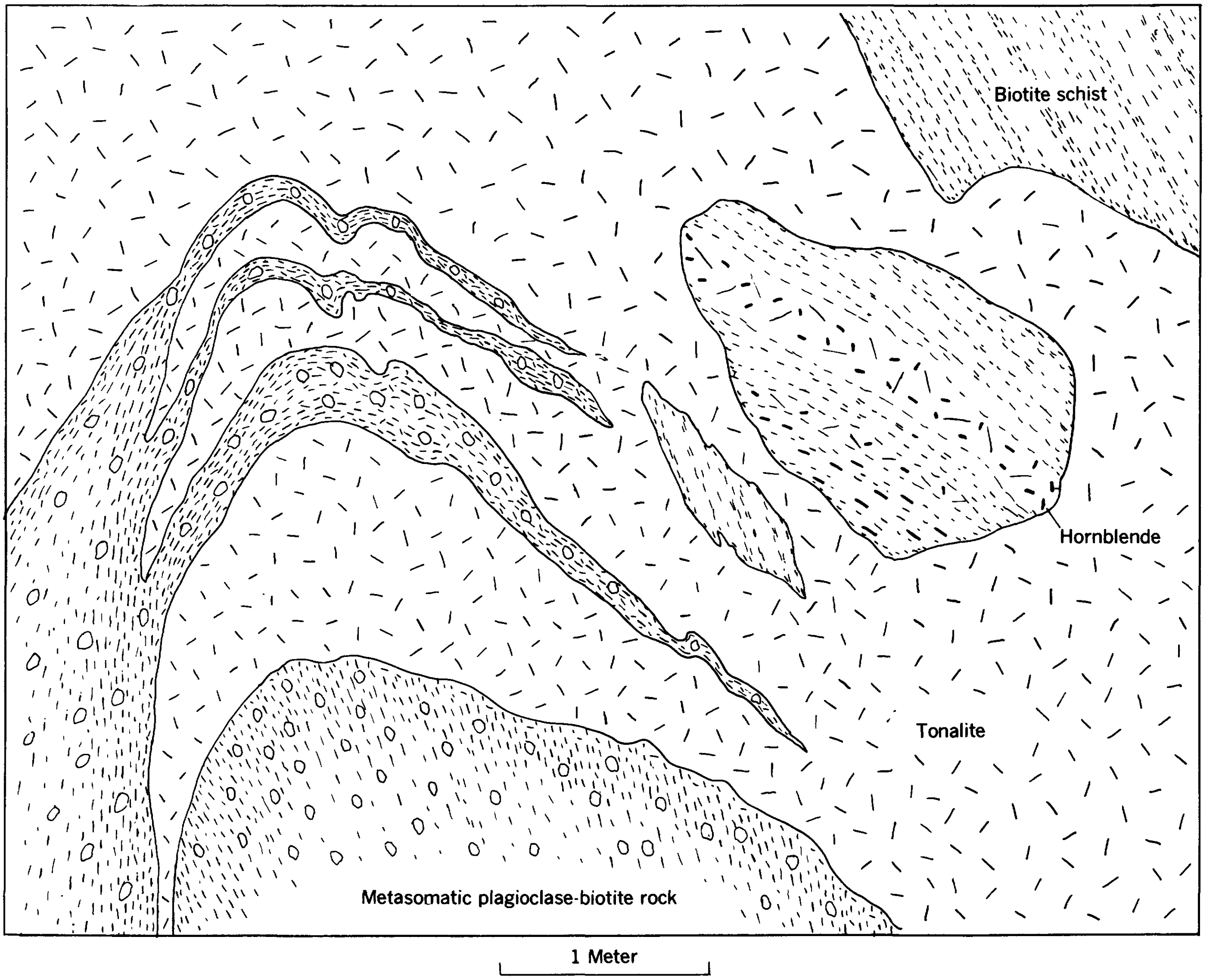


FIGURE 83.—Inclusions of country rocks—fine-grained biotite schist, and metasomatized schist—in an intrusive tonalite. From a vertical wall facing south along the North Fork of Clearwater River east of Dent (loc. 196, pl. 3). Note the displacement of xenoliths.

clase to form quartz diorite. The tonalitization near Orofino is comparable to the feldspathization near Dent with the exception that the plagioclase in the Orofino area is, as a rule, more sodic.

METASOMATIC CHANGES IN ROCKS RICH IN CALCIUM

The carbonate-bearing layers as a rule are exceptionally susceptible to mineralogic changes. Carbonates react readily with the solutions carrying silicon and other elements either from the neighboring silicate layers or from igneous sources. In this manner, skarn is formed from limestone and dolomite. Near Dent coarse-grained rock consisting of calcium-magnesium-aluminum silicates was formed from calcareous shale during the regional metamorphism. The dolomitic sand was crystallized as diopside-plagioclase gneiss during the same period. Later, during the second phase of introduction of elements, the phase when schist was feldspathized, new minerals were crystallized in these rocks. Many elements that migrate through the silicate rocks during metasomatic metamorphism are captured by the carbonate rocks. The mineralogy of the lime-silicate rock is, therefore, perhaps more indicative of the nature of the metasomatizing solution than is the mineralogy of the other rocks.

The major constituents of the lime-silicate rocks—quartz, calcite, diopside, scapolite, phlogopite, grossularite, sphene, and apatite—crystallized during an early phase and were later locally replaced by hornblende and more sodic plagioclase. The occurrence of scapolite, phlogopite, and apatite among the early minerals indicates that chlorine and some fluorine were present during the early phase of metamorphism. Because scapolite is common in many beds of the Wallace formation in the northern part of the area shown in plate 1, it seems most likely that chlorine was contained in the marine calcareous shale in which scapolite crystallized during the metamorphism. A late development of abundant hornblende suggests that some iron and magnesium were introduced from outside sources. It is possible that a part of these heavy cations were introduced as fluorides as has been suggested by Eskola (1914, and in Barth and others, 1939, p. 384). The lack of fluorine-bearing minerals in the schist and quartzite shows that these highly volatile elements were passed through the silicate rocks but were captured by the carbonate rocks. A part of the iron and magnesium is probably of sedimentary origin because carbonate rocks of the Wallace formation in the Coeur d'Alene district are known to consist mainly of dolomite and calcite with some siderite (Ransome and Calkins, 1908, p. 41).

The occurrence of hornblende in the lime-silicate rock is similar to that in the diopside-plagioclase gneiss southwest of Big Island and in the Headquarters quadrangle. The resultant rock consists mainly of hornblende and plagioclase with or without quartz. Locally the amount of plagioclase is insignificant, the metasomatic rock being hornblendite.

Evidence of the formation of quartz diorite from diopside-plagioclase gneiss was found along the North Fork of the Clearwater River near Dent, just west of locality 197. Thin beds of light-green diopside-plagioclase gneiss are here interbedded with the thin-bedded biotite-plagioclase quartzite. A medium-grained heterogeneous rock consisting of plagioclase and hornblende occurs as an irregular mass in the quartzite and the gneiss. The contact zone across the bedding is rich in hornblende, but segregations of almost pure plagioclase occur in the quartz diorite. The quartz diorite is much finer grained than the hornblende-plagioclase rock No. 140 southwest of Big Island, described earlier (p. A–20), but chemically and mineralogically these two rocks are alike and apparently have the same origin. In both rocks mainly iron, magesium, and sodium were introduced and calcium was removed. The changes in the amount of silicon and aluminum show local variation. The amount of silicon usually decreases but it may stay unchanged. The amount of aluminum may increase or decrease. The increase gives rise to a crystallization of abundant calcic plagioclase in place of quartz. If the amount of aluminum decreased, the resultant rock is enriched in hornblende, the amount of plagioclase staying approximately the same.

Comparison of the number of cations in a standard cell of the diopside-plagioclase gneiss (No. 289, table 2) and the metasomatic hornblende-plagioclase rock (No. 140, table 7) shows that Si, Fe, Mg, and Na are introduced and Al and Ca removed when diopside-plagioclase gneiss is transformed to hornblende-plagioclase rock (column 4, table 24, and fig. 84). Transformation of diopside-plagioclase gneiss to a more mafic rock such as garnet amphibolite requires introduction of Fe, Mg, Al, and Na and removal of Si and Ca (column 5, table 24, fig. 84).

On an old logging road along the east side of Dicks Creek an insignificant amount of sulfide minerals was found in the lime-silicate rock. These minerals include mainly pyrite, but some chalcopyrite also occurs. The sulfide crystals occur between diopside and plagioclase grains and along the cracks in these minerals and are thus crystallized late.

TABLE 24.—*Comparison of the number of cations in a standard cell of diopside-plagioclase gneiss, metasomatic hornblende-plagioclase rock, and garnet amphibolite*

Cation	1	2	3	4	5
	289 (table 2)	140 (table 7)	360 (table 14)	Cations introduced (+) or removed (−) between Nos. 289 and 140	Cations introduced (+) or removed (−) between Nos. 289 and 360
	Number of cations in standard cell				
Si	57.2	58.9	46.6	+1.7	−10.6
Al	11.9	10.2	21.5	−1.7	+9.6
Fe'''	.2	.3	.7	+.1	+.5
Fe''	2.3	3.9	6.1	+1.6	+3.8
Mn	.0	.1	.1	+.1	+.1
Mg	7.2	9.2	8.2	+2.0	+1.0
Ca	14.4	7.8	9.9	−6.6	−4.5
Na	1.3	3.4	6.5	+2.1	+5.2
K	.1	.4	.7	+.3	+.6
Ti	.3	.4	.6	+.1	+.3
P	.1	.1	.1	.0	.0
C	.0	.0	.2	------	+.2
H_2O	(2.1)	(2.4)	(3.8)	(+.3)	(+1.7)
Total	95.0	94.7	101.2		

Figure 85 shows a replacement of plagioclase by zoisite in a lime-silicate rock at Dent. Grossularite and diopside are abundant in this layer of lime-silicate rock, which is included in an intrusive tonalite. Plagioclase rich in anorthite crystallized first; and then, probably during the intrusion of the tonalite, the calcic plagioclase was replaced by zoisite and albite. This indicates a low temperature and introduction of hydrous solutions during the intrusion. A fine-grained grayish mixture of epidote minerals between the albite and zoisite grains represents the first stage of replacement of the plagioclase. The quartz inclusions in the newly crystallized zoisite grains are larger and more irregular in shape than the round inclusions in the original plagioclase. The excess of quartz after the alteration of plagioclase to zoisite increased the size of the quartz inclusions.

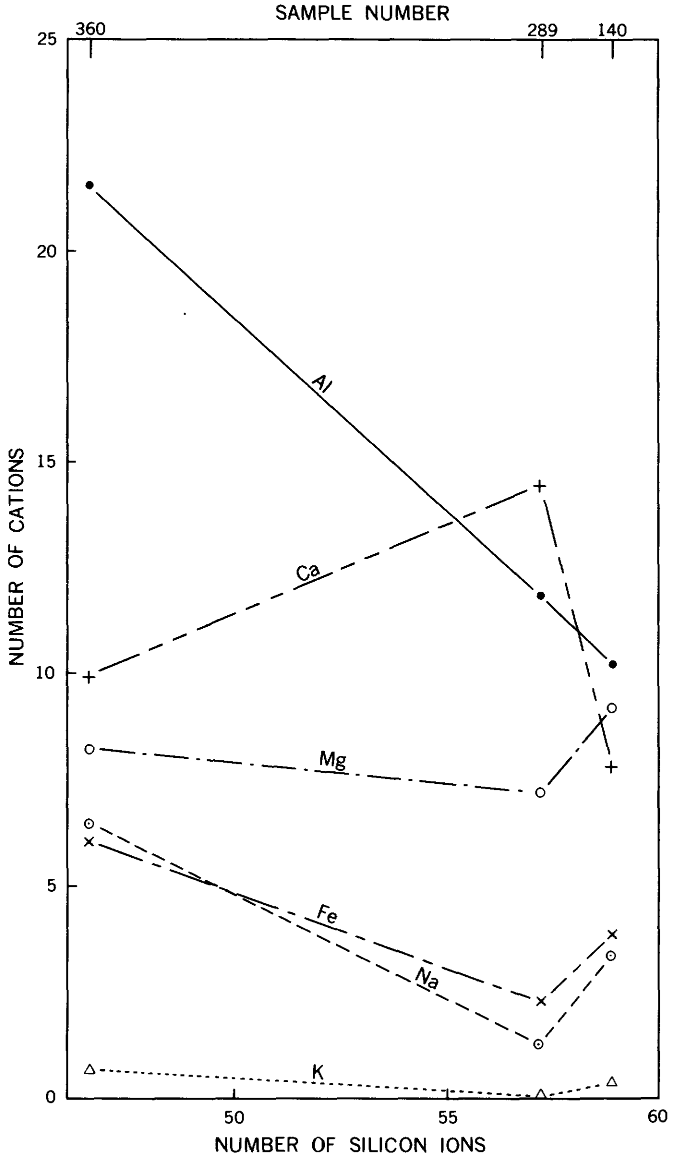


FIGURE 84.—Graph showing the number of cations in standard cells of diopside-plagioclase gneiss (No. 289) and metasomatic equivalents: hornblende-plagioclase rock (No. 140) and garnet amphibolite (No. 360).

INTRODUCTION OF ALKALIES

Most of the regionally metamorphosed rocks in the Dent area contain little or no potassium feldspar. On the contrary, the sodic plagioclase is common in varying amounts and it is often impossible to tell whether this mineral was originally present in the rock or was introduced later. However, a comparison of the textures and homogeneity of the rock is generally helpful. In the vicinity of Dent the secondary plagioclase ranges in composition from An_{25} to An_{50}. Reactions between the original sodic plagioclase and the introduced calcium, as well as the reactions between the secondary calcic plagioclase and the soda added later, probably caused crystallization of the plagioclase between the two extremes.

The metasomatic tonalite, which is common in the whole area shown in plate 1 (though only a few occurrences are shown on the map), is a result of development of oligoclase or andesine in schist and quartzite. Texturally these rocks differ strikingly from the intrusive tonalite, which in many localities cuts the metasomatic rocks discordantly (cf. figs. 28 and 53*A*).

In the metasomatic tonalite the platy structure is inherited from the parent rock. In many places a relict bedding is recognizable and remnants of the metasedimentary rock show the nature of the parent rock. In places the metasomatized rock changes gradually to an ordinary metasedimentary rock parallel to the bedding; in a few places the change is fairly abrupt. Also the contacts across the planar structure are gradational, and the schist outside the tonalite body contains scattered plagioclase. The scattered plagioclase is similar to the plagioclase in the tonalite. The grains are angular, round, or elongated and are larger than those of the primary plagioclase. In some localities larger augen may occur (fig. 61*A*). Inclusion of other minerals, as small quartz grains, biotite, and accessories, is common in the secondary plagioclase. Near Dent, west of the mouth of Elk Creek, migmatite was formed through a development of oligoclase along the planes of schistosity in the biotite-garnet schist. As a result of a continued crystallization of secondary andesine, small bodies of gray tonalite formed in the same area. Metasomatic quartz diorite with a more calcic plagioclase and abundant hornblende, as well as garnet amphibolite, are common in the aureoles of the tonalitized areas. Apparently these mafic rocks are complementary to the secondary plagioclase in the schist, just as the masses of ferromagnesian minerals and the plagioclase (An_{50}) in locality 13 are complementary.

Late albite with very little orthoclase was crystallized locally in banded plagioclase-biotite-muscovite schist about 5 miles north of Big Island (loc. 234, pl. 2). The albite occurs as larger grains in narrow bands or veinlets and includes small rounded quartz grains.

Potassium feldspar grains in specimens collected between localities 12 and 13 (pl. 3) are very conspicuous because they occur as large oval porphyroblasts, which have crystallized across the structures of the folded biotite-garnet schist (fig. 86). The porphyroblasts measure 5 to 8 cm in longest dimension and include numerous grains of the minerals of the host rock and thus crystallized late. The surrounding schist contains no potassium feldspar, and therefore this local occurrence is thought to be of a result of a late metasomatic addition of potassium. The porphyro-

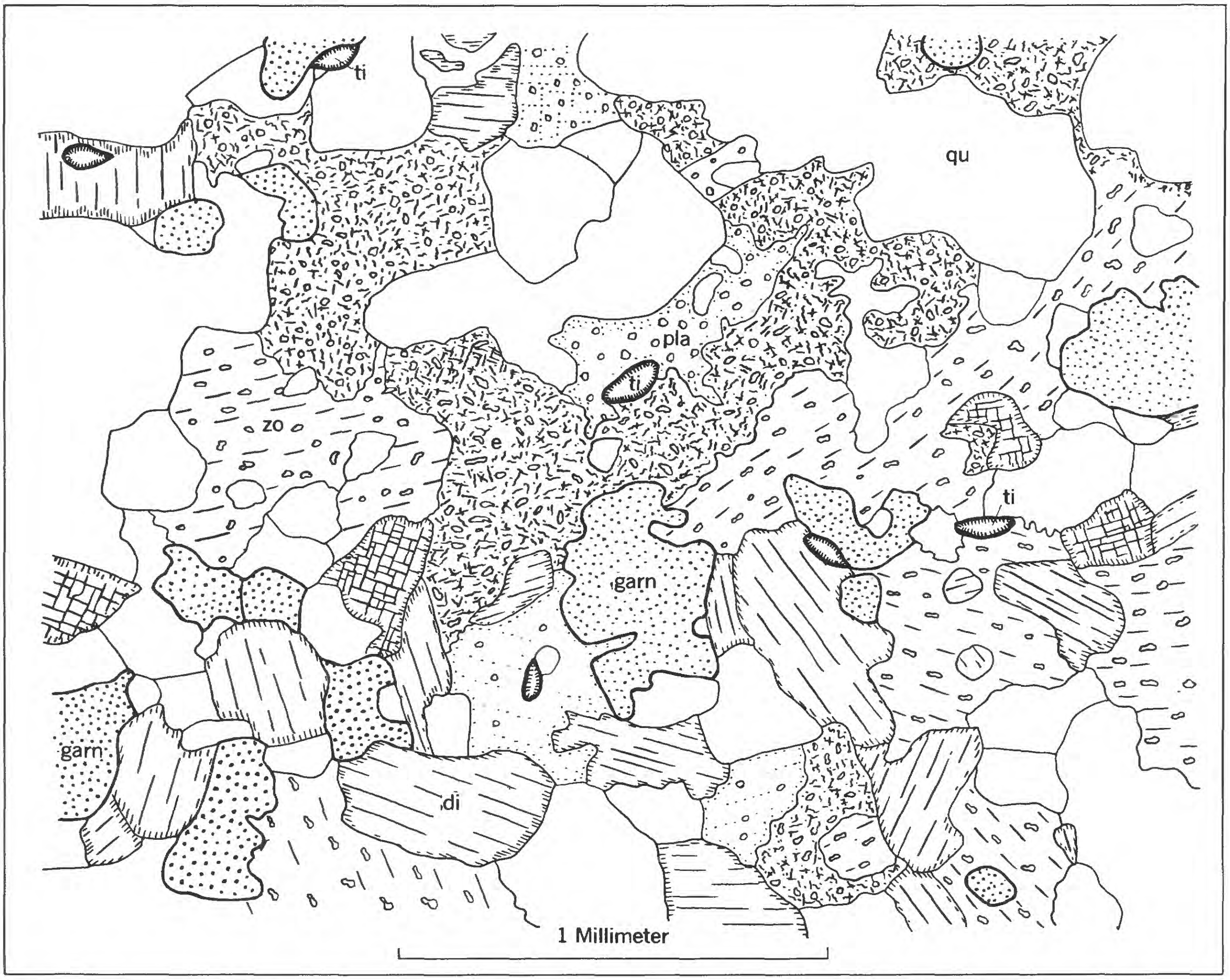


FIGURE 85.—Camera lucida drawing showing the replacement of plagioclase (pla) by zoisite (zo) in a diopside(di)-grossularite(garn)-plagioclase rock at Dent (loc. 555, pl. 17). Note the rounded quartz inclusions in plagioclase and zoisite and the first stage of replacement (e) between these minerals. qu=quartz; ti=sphene.

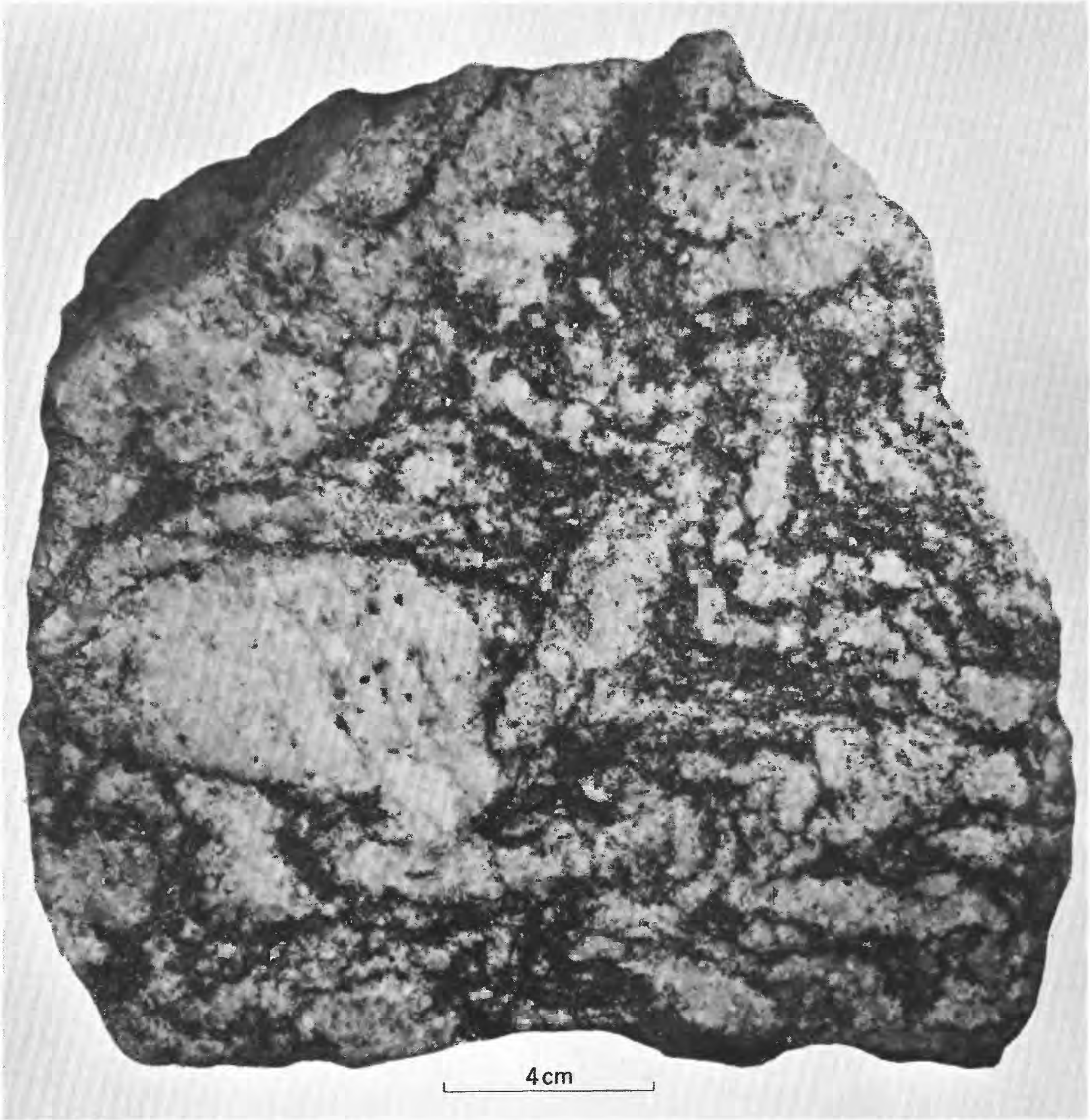


FIGURE 86.—Large round orthoclase grains in biotite-garnet schist about 1 mile east of the mouth of Elk Creek along North Fork of the Clearwater River (loc. 13, pl. 3).

blasts are perthitic orthoclase with $\alpha=1.518\pm0.001$, $\gamma=1.526\pm0.001$; some spots show a microcline grid structure. According to Laves (1952), orthoclase is not a stable form of $KAlSi_3O_8$ but represents an intermediate stage between sanidine and microcline. Sanidine is stable at high temperatures and microcline below 650° to 700° C. Smaller tabular crystals of orthoclase, 3 mm to 1 cm thick and 0.5 to 3 cm long, occur in irregular veinlets or as individual crystals alined parallel to the schistosity which gives a gneissic appearance to the rock. These crystals are twinned according to the Carlsbad law.

PART 5. EVIDENCE AND THEORIES

SUMMARY OF EVIDENCE BEARING ON THE ORIGIN OF ANDESINE AND HORNBLENDE

Three possible sources should be considered in explaining the origin of the andesine and hornblende in the schist, gneiss, and quartzite: sedimentary, igneous, and metasomatic. The evidence for and against each of these possibilities is summarized as follows:

I, A. Evidence for sedimentary origin.

The andesine- and hornblende-bearing rocks occur in the metasedimentary strata and contain many of the minerals of the surrounding schist even though in different proportions.

B. Evidence against sedimentary origin.

1. Distribution of andesine and hornblende in irregular patches and lenses which in many places cut sharply across the bedding (example fig. 81), and their occurrence parallel to the fissures and cross joints (fig. 19, p. A–81) and in dikelike bodies (p. A–85, 86).
2. Chemical evidence. The andesine and hornblende are foreign to the host rocks. Examples: Hornblende in the quartzite (p. A–82); andesine in the sillimanite-garnet schist (fig. 81); hornblende in the aluminous schist (p. A–85).

II, A. Evidence for igneous origin.

Both of the lines of evidence under I, B favor the hypothesis of igneous origin of andesine and hornblende-bearing rocks.

B. Evidence against igneous origin.

1. Mineral content. The minerals of the host rocks (schist and gneiss), such as garnet, biotite, and sillimanite, occur also in the hornblende- and andesine-bearing rocks.
2. Texture.
    a. Hornblende and andesine occur as large holoblasts including small grains of the minerals of the host rock, many of them still in their original position. Example: Hornblende in diopside-plagioclase gneiss (figs. 17, 18, 20); hornblende in biotite gneiss (figs. 21 and 22); andesine in biotite schist (figs. 23 and 24).
    b. Andesine occurs in large round grains embedded among the minerals of the schist (examples, figs. 79 and 80*B*) or among the minerals of the gneiss (figs. 53 *A*, 61*A*, 61*C*).
3. Heterogeneity of the rocks. The number of large andesine and hornblende grains varies locally, giving rise to every gradation from a schist or gneiss to a rock rich in andesine or hornblende, or in both of these minerals. This variation is observable between individual occurrences and also within various parts of the same occurrence. (This variation is separate from a gradational contact.)
4. Contacts. Near Orofino many contacts between the gneiss and the gneissic quartz diorite or tonalite are gradational so that the number of large oval andesine and hornblende grains decreases toward the country rock, the fine-grained gneiss.
5. Room problem becomes acute in places such as locality 471 where small hornblendite lenses occupy a part of the strata in quartzite without disturbing the structure (figs. 73 and 74*A*).
6. Inclusions. Along the canyons near Orofino, rows of small inclusions of fine-grained biotite or biotite-hornblende gneiss mark the continuation of the bedding of the country rock in the gneissic quartz diorite and tonalite (p. A–62). Small sheetlike inclusions of schist are common in the gneissic plutonic rocks near Orofino and in the andesine-bearing rocks in the western part of the area (p. A–61, 80). These inclusions seem to have preserved their original position (p. A–80).

III, A. Evidence for metasomatic origin.

1. Distribution of the andesine and hornblende (IB1).
2. Chemical composition of the andesine- and hornblende-bearing rocks (IB2).
3. Their mineral content (IIB1).
4. Texture (IIB2).
5. Contacts (IIB4).
6. Room problem (IIB5).
7. Inclusions (IIB6).

These relations can be explained satisfactorily only if it is assumed that large grains of andesine and hornblende crystallized later, replacing the earlier minerals of the schist, gneiss, and quartzite, and that some elements were introduced into and some removed from the parent rocks.

B. Evidence against metasomatic origin.

Difficulty in transportation of material in solid rocks. Even if the evidence for the metasomatic origin is conclusive, one tends to seek other explanations because it is difficult to understand the mechanism of such an extensive exchange of material.

Because the hypothesis of metasomatic origin explains best the relations between the andesine- and hornblende-bearing rocks and the surrounding metasedimentary rocks, it was adopted in explaining the occurrence of andesine, hornblende, and some of the biotite.

Comparison of the chemical composition of the parent rocks and the metasomatic rocks gives a picture of the extent of exchange of elements during the metasomatism.

PLACE OF METASOMATISM IN SEQUENCE OF GEOLOGIC EVENTS

The tentative sequence of geologic events in the area studied is presented in table 25, which also includes some additional information about the quartz monzonite series. This information is based mainly on data gathered during four years of work in the western part of Clearwater County. Relation of the newly formed minerals and rocks to the earlier textures and structures, contacts between the metamorphic and igneous-looking rocks, and texture in all rocks concerned were carefully studied.

The intrusive rocks near Dent and Orofino may be divided on the basis of structural criteria into five groups:

1. The earliest intrusive rock is a metamorphosed diabase exposed northwest of Peck, about 10 miles west of Orofino (pl. 4). This diabase is folded with the sedimentary rocks and is thus older than the folding. Recrystallization and development of large round plagioclase grains has locally increased the grain size and transformed the diabase to a gabbroic rock. A younger gabbro cuts this metamorphosed diabase discordantly at Peck.

2. The small intrusive bodies of norite, gabbro, and diorite near Orofino and farther to the west show planar or linear structures similar to those of the country rock, and the contacts are generally concordant. The more silicic rocks—hornblende-bearing tonalites—show the same characteristics.

3. In contrast with these concordant bodies near Orofino, the hornblende tonalite and quartz diorite at Dent, in the Headquarters quadrangle, and near Big Island locally show discordant contacts, and the alinement of minerals in them is less pronounced. They, as well as the quartz dioritic border zone of the batholith, were emplaced after the folding. The gabbro that cuts the metadiabase at Peck belongs probably to the same phase of intrusion as the small bodies of gabbro and diorite (group 2); yet it may be a northward extension of the quartz diorite near Greer, which in turn may be connected with the quartz dioritic border zone of the main batholith (group 3). Thus the intrusive rocks of groups 2 and 3 may not form two separate groups but were probably emplaced at the beginning and at the end of the same phase of intrusion.

4. The rocks metasomatized during the emplacement of quartz diorite and hornblende tonalite are cut discordantly by younger tonalite (fig. 83). The emplacement of these discordant silicic tonalite bodies continued over a considerable time interval, just as did the emplacement of quartz diorite and hornblende tonalite. Crosscutting relation between two types of silicic tonalite—a light-colored coarse-grained variety and a fine-grained gray one—was found west of locality 190 (pl. 3), where dikes of light-colored tonalite cut the gray fine-grained variety. In locality 379 (pl. 3) the fine-grained gray tonalite occurs as a wide dike with chilled borders. This dike also is cut by the light-colored coarse-grained tonalite. In the northern and eastern part of the area shown in plate 1, members of an intrusive series rich in potassium feldspars cut the metasomatic rocks discordantly. This series includes gabbro, diorite, quartz monzonite, granite, and aplite. The age relation between the younger intrusive tonalite

Table 25.—*Sequence of intrusion and metasomatism and their relation to structures*

Intrusion and metasomatism	*Relation to the structures*
1. Intrusion of the basic sills. Metadiabase and peridotite near Peck.	Folded: lineation is parallel to the axes of small folds, which trend N. 35°–60° E. and plunge 25°–50° NE.
2. Folding, intrusion, and metamorphism (Nevadan). Emplacement of small igneous bodies of norite, gabbro, and diorite. Metasomatism 1: Development of hornblende, biotite, plagioclase, and accessory minerals mainly in the wide contact aureoles of the intrusive bodies. Metasomatic hornblendite, gabbro, diorite, and tonalite near Orofino were formed.	Axes of large folds trend N. 70°–80° W. Small folds whose axes trend N. 35°–60° E. may be contemporary or earlier. Gabbro cuts the metadiabase discordantly near Peck. Both intrusive and metasomatic rocks show planar and linear structures similar to those in the metamorphic country rocks.
3. Emplacement of quartz diorite and tonalite near Greer and the quartz diorite of the border zone of the Idaho batholith. (Early Cretaceous). Metasomatism 2: Development of: (a) Biotite, hornblende, and accessory minerals. (b) Calcic plagioclase. Basic segregations, feldspathized schist, and anorthosite are result of differentiation during the metasomatism.	Metasomatism 2 has cut the folded structures discordantly and has obliterated the earlier structures and textures (fig. 81).
4. Emplacement of tonalite near Dent, quartz monzonite series near Canyon Ranger Station, and granite at Bungalow (late Early Cretaceous). Metasomatism 3: Development of: (a) Sodic plagioclase and potassium feldspar in schist and gneiss. (b) Feldspars in the contact aureoles of the rocks of the quartz monzonite series.	This tonalite cuts the metasomatized schist discordantly. Large orthoclase porphyroblasts grow across the textures in the metasomatized schist (fig. 86). Rocks of the quartz monzonite series cut the metasomatic anorthosite in southern Boehls Butte quadrangle.
5. Emplacement of dioritic dikes and granite porphyry dikes (post-Laramide).	Many of the dike rocks were emplaced along fault zones.
6. Extrusion of Columbia River basalt (Miocene).	Basalt flows are horizontal and overlie the other rocks.

and the rocks belonging to this quartz monzonite series remains uncertain because the two rock types were not found in contact.

5. Porphyritic dike rocks—ranging mainly from diorite to granite porphyry in composition—cut the other intrusive rocks discordantly and form the youngest intrusive group in the area.

Structural and petrologic studies show that at least three phases of metasomatism can be considered:

1. The first phase of metasomatism affected wide areas, but the structures of the parent rocks were preserved. Near Orofino this metasomatism was probably associated with the emplacement of small igneous bodies of gabbro and diorite that display mostly parallel contacts and the same parallel structures—foliation and lineation—as the country rocks. Thus, it seems that the igneous activity, the first phase of metasomatism, and the recrystallization were contemporaneous with the deformation. Many small bodies of gabbro and diorite near Orofino and farther to the west show a typical hypidiomorphic texture in the centers of the bodies, but replacement textures increase toward the contacts. The rock in these bodies grades from an intrusive gabbro and diorite through a metasomatic gabbro or diorite to a gneissic hornblende-bearing rock and further to a metamorphic parent rock—biotite-plagioclase gneiss or garnetiferous biotite schist. At the present stage of study it is impossible to tell whether material has also been added to these schists and gneisses.

2. The metasomatic introduction of elements that resulted in development of hornblende, biotite, and calcic plagioclase in the Dent area is later than the metamorphism and the first phase of metasomatism. Many of the earlier structures and textures are obliterated by this second metasomatism (fig. 81). The contacts in many places are sharp, and the development of new minerals is restricted to rather small areas; but in these affected areas the exchange of material tends to be far more extensive than it was during the first phase of metasomatism. The introduced minerals are the same, but the local variation in proportions gives rise on one extreme to mafic masses and on the other to anorthositic veins and dikelike bodies. In many places the earlier structures are only partly obliterated, and it is impossible to tell whether crystallization of metasomatic minerals in these rocks took place during the first or second phase of metasomatism. Probably there were not actually two separate phases, but rather the introduction of elements continued sporadically over a period of time, as did the igneous activity with which it is genetically connected.

Most rocks of tonalitic composition are gneissic in appearance and are either recrystallized biotite-plagioclase gneiss or products of metasomatic introduction of components of hornblende and plagioclase into schist and quartzite. As a rule the plutonic rocks near Orofino show a stronger planar and linear structure than those at Dent and in the Headquarters quadrangle. Similarly, the foliation and lineation in the metasomatic rocks tend to be less pronounced near Dent. At Dent and farther to the east the development of metasomatic hornblende, biotite, and calcic plagioclase is postfolding (fig. 81), as are the hornblende tonalite and quartz diorite of the same area. In the Headquarters quadrangle the intrusive quartz diorite along the northwestern border zone of the batholith has given rise to an extensive metasomatic development of hornblende, calcic plagioclase, and locally biotite in the country rock near their contacts. This contact metasomatism is postdeformation and is similar to the metasomatism of the Dent area. Some outcrops between Dent and Orofino show two types of metasomatic plagioclase (figs. 87 and 88). Figure 87 shows that the earlier plagioclase crystallized as large grains or segregations of small grains in a biotite quartzite and biotite gneiss. Remnants of this metasomatized rock as well as of early plagioclase remain as inclusions in the metasomatic quartz diorite. Figure 88 shows a pegmatite cutting the metasomatic quartz diorite discordantly. This pegmatite consists of a group of large plagioclase grains that were formed in a definite band. Biotite accumulated around the large round plagioclase grains.

3. The late development of sodic plagioclase and orthoclase was probably contemporaneous with the emplacement of silicic tonalites and the quartz monzonite.

DISTRIBUTION OF MINOR ELEMENTS

Quantitative spectrochemical analyses of minor elements were made by A.A. Chodos and Paul R. Barnett of the U.S. Geological Survey of all specimens analyzed chemically. The results are given in tables 26, 27, and 28. There are fairly high concentrations of zirconium, strontium, and barium in many of the specimens analyzed. Beryllium, molybdenum, lanthanum, and boron are scarce or completely lacking.

In the tables the mean concentration for each element is calculated for metasedimentary and plutonic rocks in the Headquarters quadrangle and near Orofino and for the metasomatic rocks near Dent. These means were plotted into histograms (figs. 89 and 90), which show that the metasomatic rocks have the highest concentration of the minor elements present in a noteworthy quantity. In the following pages the distribution of each element is discussed separately.

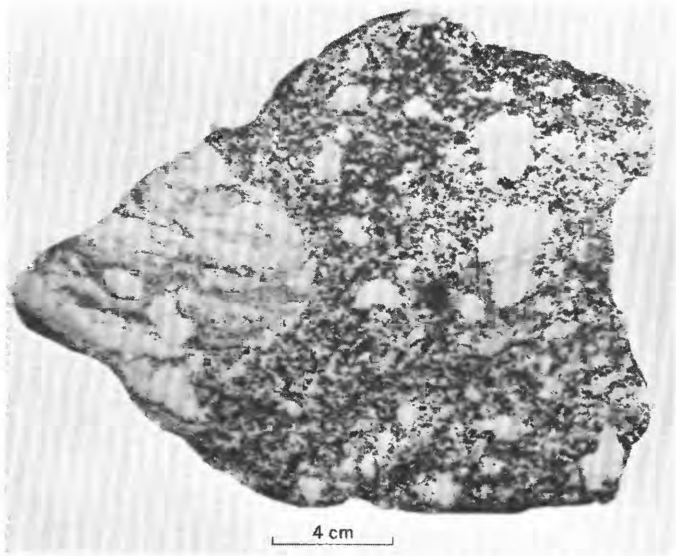


FIGURE 87.—Two generations of late plagioclase are seen in a specimen of quartz diorite about a mile north of Bruces Eddy (loc. 435, pl. 4). The earlier plagioclase was crystallized as large grains in biotite quartzite and gneiss, the remnants of which are seen at left. The large white grains in the diorite are remnants of the same early plagioclase. The small grains of plagioclase in the quartz diorite are later.

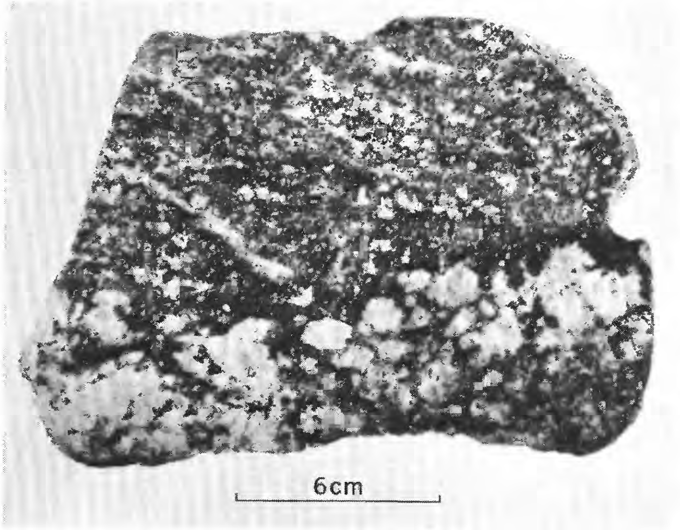


FIGURE 88.—Discordant "pegmatite vein" consists of large round plagioclase grains which crystallized late in a metasomatic quartz diorite parallel to a certain plane. The dark minerals of the vein area recrystallized and accumulated around the late plagioclase grains. Freeman Creek (loc. 508, pl. 4).

DISCUSSION OF INDIVIDUAL ELEMENTS

ZIRCONIUM

The mean value for the concentration of zirconium in the schist in the Headquarters quadrangle is about twice as high as the value reported by Shaw (1954, p. 1163) for the high-grade rocks in the pelitic Devonian Littleton formation of New Hampshire. In the border zone of the Idaho batholith and in the satellitic intru-

TABLE 26.—*Minor elements in metasedimentary and plutonic rocks of the Headquarters quadrangle*[1]

[In parts per million]

Specimen no.	Ni	Co	Cr	Cu	V	Ga	Sc	Ba	Sr	Be	Mo	Y	Yb	La[2]	Zr	B
Metasedimentary rocks																
246	40	20	100	8	100	30	40	900	300	3	----	100	10	----	400	4
247	30	10	100	6	90	20	30	500	10	----	5	100	10	100	400	4
248	20	30	80	6	70	20	10	500	500	1	----	60	4	----	400	--
252	40	---	40	5	40	20	7	70	40	20	----	500	30	----	400	--
254	---	---	8	4	5	---	---	80	20	----	----	40	2	----	400	--
289	20	9	30	5	30	10	20	300	700	1	----	80	5	----	200	--
Mean	25	11	59	6	56	17	18	392	262	4	1	147	10	17	367	1
Plutonic rocks																
292	10	10	40	10	80	20	6	500	900	----	----	----	----	----	100	3
322	20	20	80	4	100	30	6	600	1,000	----	----	40	2	----	200	--
190	---	4	----	2	20	20	---	600	1,000	----	----	----	----	----	200	--
379	5	5	8	4	40	20	---	500	1,000	----	----	----	----	----	100	--
Mean	9	10	32	5	60	23	3	550	975	----	----	10	1	----	150	1
Quartz porphyry																
301	40	10	100	8	50	20	---	600	700	----	----	----	----	----	70	--

[1] Looked for but not found: Pb, Ag, Au, Pt, W, Ge, Sn, As, Sb, Bi, Zn, Cd, Tl, In, Nb, Ta, Th, and U.

[2] Cerium-earth metals are reported as a group under La.

TABLE 27.—*Minor elements in metasedimentary and plutonic rocks near Orofino*[1]

[In parts per million]

Specimen No.	Ni	Co	Cr	Cu	V	Ga	Sc	Ba	Sr	Y	Yb	Zr	B
Metasedimentary rocks													
109	30	20	50	5	100	30	20	10	3,000	----	----	70	----
106	7	5	20	5	60	10	10	100	500	50	4	70	----
145	5	7	9	9	70	10	3	700	900	40	3	200	----
146	----	7	5	10	40	10	----	400	800	----	----	100	----
147	----	----	-----	2	10	10	----	80	100	40	3	200	----
360	30	40	60	5	200	20	30	200	1,000	----	----	30	5
Mean	12	13	24	6	80	15	11	248	1,050	22	2	112	1
Plutonic rocks													
20	100	80	200	1	500	30	100	200	600	100	9	70	5
32	80	40	200	50	200	20	20	200	1,000	----	----	30	----
180	50	30	100	60	200	20	20	500	1,000	----	----	80	5
42	8	10	10	10	70	20	5	60	1,000	----	----	100	----
43	----	----	-----	5	30	20	----	300	1,000	----	----	70	----
Mean	48	32	102	25	200	22	29	252	920	20	2	87	1

[1] Looked for but not found: Pb, Ag, Au, Pt, W, Ge, Sn, As, Sb, Bi, Zn, Cd, Tl, In, Nb, Ta, Th, U, Be, Mo, and La.

TABLE 28.—*Minor elements in metasomatic rocks near Dent and in granite No. 758*[1]

[In parts per million]

Specimen No.	Ni	Co	Cr	Cu	V	Ga	Sc	Ba	Sr	Be	Y	Yb	La[2]	Zr	B
Metasomatic rocks															
471a	100	50	400	30	600	20	50	100	300	----	100	10	----	3,000	--
471b	100	50	200	40	500	20	30	300	2,000	----	80	8	----	100	--
119	100	60	200	60	500	30	40	2,000	400	----	200	20	500	3,000	--
13	100	40	200	10	200	50	50	2,000	2,000	3	100	20	----	500	4
Mean	100	50	250	35	450	30	43	1,100	1,175	1	120	15	125	1,650	1
140	40	10	70	3	70	20	40	200	500	----	90	5	----	200	--
Granite															
758	----	----	1	4	5	10	2	50	10	9	100	10	5	200	--

[1] Looked for but not found: Pb, Ag, Au, Pt, W, Ge, Sn, As, Sb, Bi, Zn, Cd, Tl, In, Nb, Ta, Th, U and Mo.

[2] Cerium-earth metals are reported as a group under La.

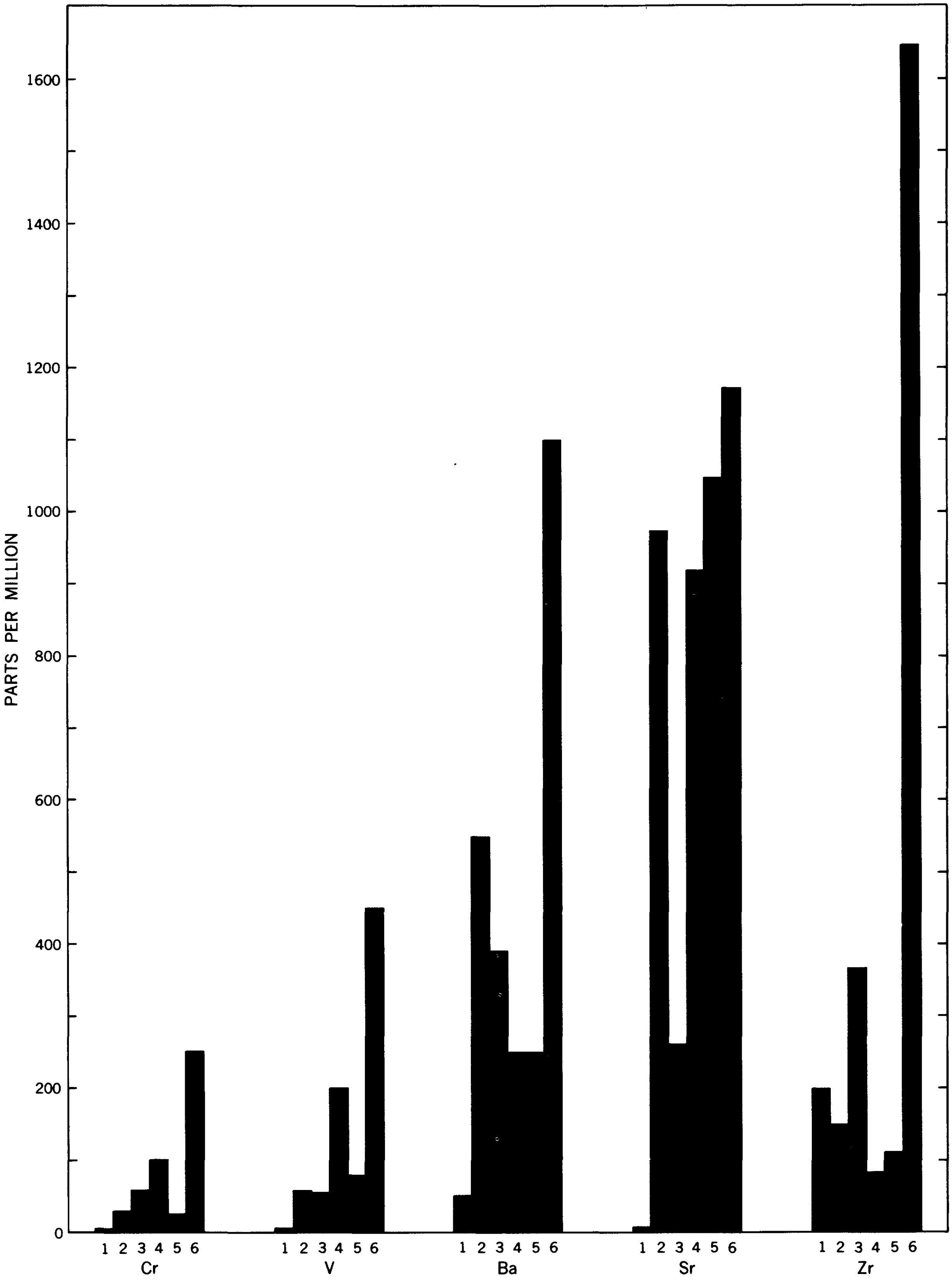
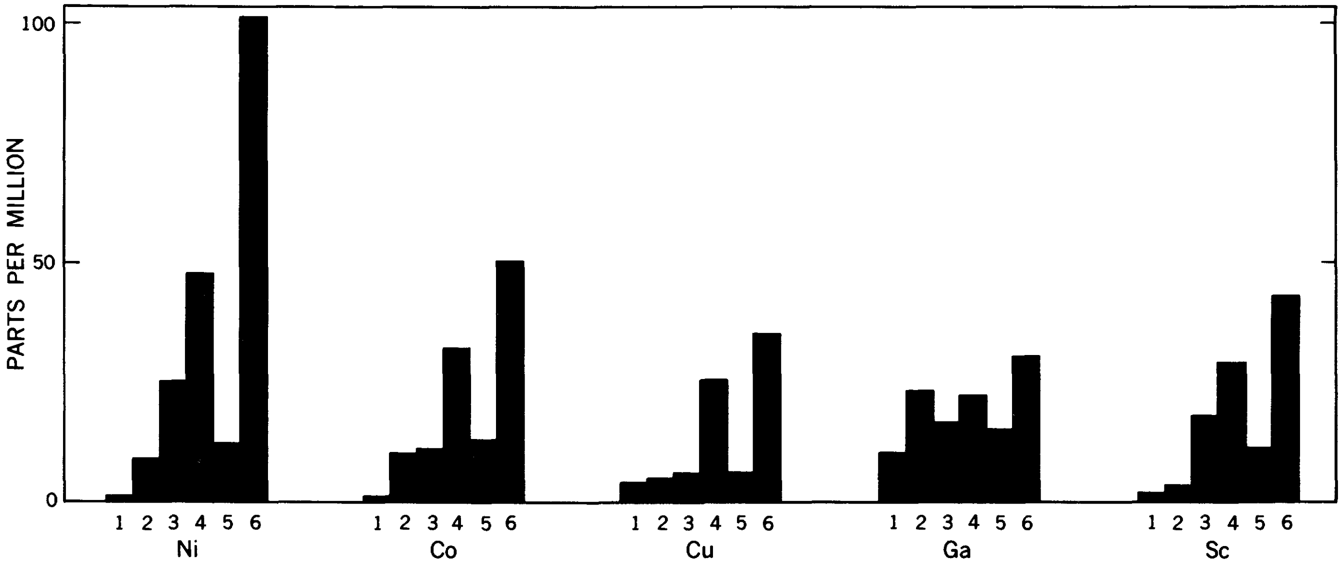


FIGURE 89.—Histogram showing the average amount of minor elements, Cr, V, Ba, Sr, Zr, in: 1. a granite (table 28), 2. plutonic rocks in Headquarters quadrangle and vicinity (table 26), 3. metasedimentary rocks in Headquarters quadrangle (table 26), 4. plutonic rocks near Orofino (table 27), 5. metasedimentary rocks near Orofino (table 27), and 6. metasomatic rocks near Dent (table 28).



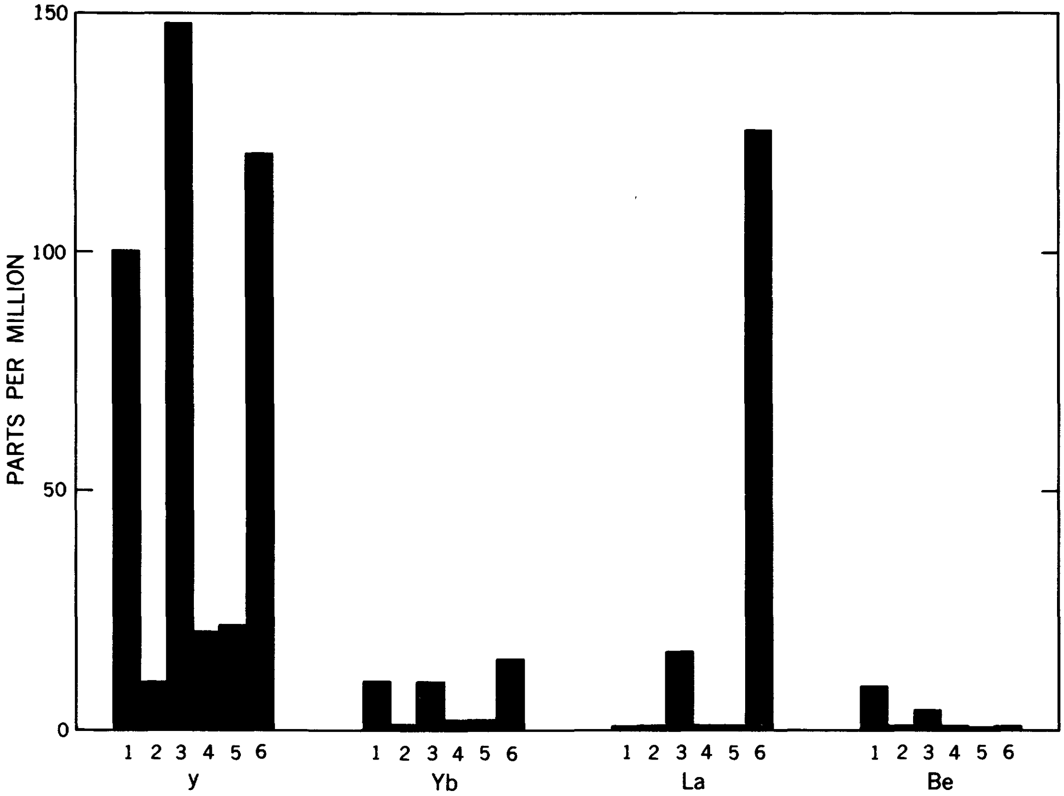


FIGURE 90.—A histogram showing the average amount of minor elements, Ni, Co, Cu, Ga, Sc, Y, Yb, La, Be, in: 1. a granite (table 28), 2. plutonic rocks in Headquarters quadrangle and vicinity (table 26), 3. metasedimentary rocks in Headquarters quadrangle (table 26), 4. plutonic rocks near Orofino (table 27), 5. metasedimentary rocks near Orofino (table 27), and 6. metasomatic rocks near Dent (table 28).

sions the amount of zirconium is 150 ppm, which is about the same amount as the average for quartz diorites as given by von Hevesy and Würstlin (1934a, p. 309). The tonalite and granite show lower concentration of zirconium than reported by these authors for the average granite. Near Orofino the corresponding concentrations are still lower.

The high concentration of zirconium in the metasomatic rocks near Dent forms a striking contrast with the low average values in the plutonic rocks and in the gneisses of the surrounding area. The specimens of basic rocks (Nos. 471a and 119) contain as much as 0.3 percent zirconium, which is about 10 times the amount found in the parent rocks and about 20 times as much as found by von Hevesy and Würstlin for gabbroic rocks. In rocks containing metasomatic plagioclase, the amount of zirconium is less (Nos. 471b and 13). Thus the zirconium was concentrated with iron and magnesium during the metasomatism.

STRONTIUM AND BARIUM

The concentration of strontium in all rocks near Orofino and in the plutonic rocks in the Headquarters quadrangle is higher than that of barium and much higher than values given by von Hevesy and Würstlin (1934b, p. 313) for average quartz diorite and slate. The plutonic rocks of the tonalitic suite contain about 0.1 percent strontium, which is about twice the amount of barium in the same rocks and a hundred times the amount of strontium found in the granite No. 758. Strontium is concentrated in the plagioclase-bearing layers in the metasedimentary rocks. The specimens (Nos. 246 and 248) that contain abundant plagioclase have 300 to 500 ppm strontium, whereas the corresponding value for the neighboring layers in the schist and quartzite is only 10 to 40 ppm. The highest concentration is found in diopside-plagioclase gneiss (Nos. 289 and 109), suggesting that the strontium there was concentrated with calcium carbonate during sedimentation. During metasomatism, strontium accompanies the plagioclase, as shown by the high concentration in specimens 471b and 13. In general, the rocks near Orofino contain more strontium than those in the Headquarters quadrangle.

The distribution of barium is in some respects similar to that of strontium, but there are notable differences, especially among the metasedimentary rocks. The layers rich in diopside are poor in barium, which suggests that only a small amount of barium was deposited with dolomite. The highest concentration in the Headquarters quadrangle is found in layers rich in biotite, muscovite, and plagioclase (No. 246). Specimen 247, which shows a very low concentration of strontium because of low content of plagioclase, is rich in barium. The barium here is contained in biotite, substituting for potassium. Spectrochemical analyses of biotite and phlogopite show from 0.1 to 0.2 percent barium but scarcely any strontium (table 29). The muscovite in this area contains both of these elements, the amount of barium varying considerably.

Near Orofino the highest concentration of barium is found in the gneiss No. 145, which contains 10 percent microcline. The barium content of the adjoining layer (No. 146) is much lower. Because this layer (No. 146) is rich in biotite but contains only 3 percent microcline, it was concluded that microcline is the chief carrier of barium.

Considerable barium is contained also in the diopside-plagioclase gneiss and in plagioclase-hornblende rock, which contain neither potassium feldspar nor biotite (Nos. 289 and 140). The possible carriers in these rocks are plagioclase, diopside, and hornblende. Plagioclase here may contain more barium than usually found in this mineral. Von Engelhardt (1936) reports a high concentration of barium in potassium feldspars and in biotite, but low concentration in plagioclase and in augite. In the diopside-plagioclase gneiss, diopside and plagioclase constitute 68 percent of the rock that carries 300 ppm barium. The highest concentration of barium among the metasomatic rocks is in Nos. 13 and 119, which contain 21 and 27 percent biotite, respectively. A part of the barium also in No. 13 is probably contained in plagioclase. Specimens 471a and 471b contain equal amounts of biotite, but the amount of plagioclase in 471b is 12 percent higher and the concentration of barium is 200 ppm higher.

TABLE 29.—*Minor elements in biotite, muscovite, and phlogopite*[1]

[In parts per million]

Specimen No.	Ni	Co	Cr	Cu	V	Ga	Sc	Ba	Sr	Zr	B	Li	Pb
912b	80	30	9	5	30	20	----	1,000	2	----	----	----	----
971c	40	50	4	6	30	30	----	2,000	------	----	----	100	----
971d	----	----	5	20	30	40	----	6,000	300	----	50	60	10
222	----	----	----	----	----	----	----	X00	X00	----	----	----	----
1449	----	----	----	----	----	----	----	X00	X00	----	----	----	----
767	20	9	200	5	70	20	6	1,000	------	500	----	----	----

[1] Looked for but not found: Ag, Au, Pt, W, Ge, Sn, As, Sb, Bi, Zn, Cd, Tl, In, Nb, Ta, Th, U, Be, Mo, Y, Yb, and La.

SOURCE OF SAMPLE:

912b	Biotite from kyanite-andalusite-sillimanite-cordierite gneiss, about 2 miles east of Boehls (loc. 912, pl.1). Chemical analysis of this biotite has been published earlier (Hietanen, 1956, table 2).
971c	Biotite from kyanite-andalusite schist, Goat Mountain (loc. 971, pl. 1).
971d	Muscovite from kyanite-andalusite schist, Goat Mountain (loc. 971, pl.1).
222	Muscovite from sillimanite-mica schist, along North Fork of Clearwater River, just north of Headquarters quadrangle (loc. 222, pl. 1).
1449	Muscovite from sillimanite-garnet schist, about 1½ miles north of the mouth of Elk Creek near Dent.
767	Phlogopite from lime-silicate rock near Dent (loc. 11, pl. 3).

VANADIUM

The mafic and intermediate plutonic rocks near Orofino have a higher vanadium content than the corresponding rocks in the Headquarters quadrangle. In both groups of plutonic rocks the amount of vanadium is lower than the average for igneous rocks as given by Lundegårdh (1946, p. 141). The metasedimentary rocks contain less vanadium than the average of high-grade pelitic rocks studied by Shaw (1954). The concentration of vanadium in the metasomatic rocks near Dent is notably higher, especially in the mafic rocks where it is probably due mainly to a large amount of ilmenite present. Specimen 119, for example, contains about 10 percent ilmenite and 471a about 6 percent, which is much higher than in any other rock in the area under discussion. The hornblendite No. 20, near Orofino, contains about as much vanadium as do the metasomatic rocks. Metasomatic origin for this rock is suggested on basis of its texture. In the micaceous rocks biotite and muscovite carry a part of the vanadium as shown in table 29.

CHROMIUM

The concentration of chromium is somewhat lower than that of vanadium, which it resembles in the pattern of distribution. The highest concentration occurs in the metasomatic rocks near Dent where 440 ppm was measured in specimen 471a. In the mafic and intermediate plutonic rocks near Orofino the abundance of chromium rates next, being much higher than that in the quartz diorite in the Headquarters quadrangle. The metasedimentary rocks, expecially those near Orofino, show low concentrations. The exceptions are the sillimanite-garnet schists in the Headquarters quadrangle, which contain more chromium than the quartz diorite in the same area. A high concentration of chromium was found in the phlogopite (No. 767) from a lime-silicate rock near Dent (table 29).

NICKEL AND COBALT

The pattern of distribution of nickel and cobalt is similar to that of chromium. The highest concentration is found in the metasomatic rocks near Dent and the second highest in the basic plutonic rocks near Orofino, the amount of nickel being higher than that of cobalt. The metasedimentary rocks in the Headquarters quadrangle contain more nickel than the plutonic rocks of the same area. The concentration of these elements in the plutonic rocks near Orofino is close to the concentration for diorite given by Goldschmidt (1937).

COPPER

Some copper was found in all specimens analyzed, the concentration being highest in the metasomatic rocks. In the plutonic rocks the concentration is lower than reported by Sandell and Goldich (1943) for the average igneous rocks.

THE RARE-EARTH METALS

Scandium.—Scandium is fairly evenly distributed in those rocks that contain hornblende or biotite in considerable amounts. This distribution is in accordance with the results obtained by Oftedal (1943) for the concentration of scandium in pyroxenes, hornblende, and biotite in igneous rocks. The metasomatic rocks and especially the bornblendite No. 20 are richer in this metal than the other rock types. No scandium was found in the metasedimentary rocks that consist mainly of quartz and plagioclase or in the silicic tonalite. Table 29 shows that two biotites and a muscovite analyzed from sillimanite-andalusite-kyanite schist in the Boehls Butte quadrangle (Hietanen, 1956) contain no scandium, and in the phlogopite near Dent the scandium content is only 6 ppm. According to Oftedahl, low concentrations are to be expected in high-temperature biotites. The garnet that abounds in the metasomatic rock No. 13 and in the garnet-sillimanite schists probably carries the main part of scandium. According to Eberhard (1908, p. 866), garnet is the chief carrier of scandium in mica schist.

Yttrium.—The highest concentration of yttrium, 0.05 percent, was found in the plagioclase-quartzite No. 252. The thin section shows several small yellowish grains of xenotime, YPO_4. The grains are elongated, have rounded corners, and show a cleavage parallel to the positive elongation. The rounded shape of the grains suggests that the xenotime was enriched in this particular bed during sedimentation. The concentration of yttrium in the adjoining layers of sillimanite-garnet schist is also fairly high. Among the metasomatic rocks the highest concentration is in specimen No. 119. No xenotime was found in this rock, but there are abundant allanite and zircon which may carry yttrium. Probably yttrium in the granite No. 758 is also carried by allanite and zircon, which occur in small crystals with pleochroic halos in biotite.

Ytterbium.—The concentration of ytterbium is about one-tenth of that of yttrium, their distributions being much alike.

Cerium-earth metals.—Cerium-earth metals are reported as a group under La in the tables. Notable amounts were found in only two specimens, Nos. 119 and 247. The possible carrier in the basic rock No. 119 is the allanite, which constitutes about 1.4 percent of this rock. Some of the cores of pleochroic halos in the sillimanite-garnet schist No. 247 may consist of monazite.

GALLIUM

The abundance of gallium follows closely the abundance of aluminum, as has been suggested by many authors, but the ratio gallium:aluminum does not remain constant. The highest concentration is found in the metasomatic plagioclase-biotite-garnet rock No. 13, which is richest in aluminum.

BORON

In accordance with the lack of tourmaline in the rocks, the content of boron is very low or is completely lacking. This is in striking contrast with the rocks of the Belt series in the Coeur d'Alene district where tourmaline is a common additional constituent.

SUMMARY OF THE DISTRIBUTION OF MINOR ELEMENTS

The trace elements were concentrated with the cafemic constituents during the metasomatism. Zirconium, barium, vanadium, nickel, and lanthanides show the highest concentrations in comparison with their abundance in the parent rocks or in the plutonic rocks of the region. The concentration of strontium in all plagioclase-bearing rocks is many times higher than the average value reported in the literature for the igneous rocks (Rankama and Sahama, 1950, p. 476). Only a few micaceous minerals were analyzed spectrochemically, but the results (table 29) suggest that much of the barium is contained in the mica structure. Some of the vanadium, gallium, copper, and chromium also go into micaceous minerals. Biotite and phlogopite contain some nickel and cobalt, whereas muscovite has a considerable amount of strontium.

The average concentrations calculated for titanium and phosphorus (table 30) on the basis of chemical analyses given earlier in this report show that titanium is strongly concentrated in the mafic metasomatic rocks. The concentration of phosphorus is not so conspicuous, but comparison of the amount of P_2O_5 in the metasomatic rock with that in the sillimanite-garnet schist shows that this element was also mobile during the metasomatism.

SOURCE OF INTRODUCED MATERIAL

The introduced material may be from either metamorphic or igneous sources. If the introduced elements were derived from the metasedimentary series, the total composition before and after the rearrangement should be the same. If the total composition was changed during the metamorphism, at least part of the elements came from outside sources.

SOURCE OF CAFEMIC CONSTITUENTS

Comparison of the chemical composition of the metasomatic rock with the composition of the parent rock shows that there has been considerable addition of iron, magnesium, calcium, and aluminum and loss of silicon in the schist and quartzite. Calcium was removed from the limestone layers and iron, magnesium, and aluminum were added. The total amount of calcium in the area probably increased, for the amount of limestone is insignificant. Large inclusions of the older rocks in the northwestern corner of the Idaho batholith consist of quartzites and schists. The main intrusive rock is quartz diorite, and it contains more iron, magnesium, calcium, and aluminum than the older sedimentary rocks. Therefore the possibility that the introduced material originated in the metamorphic rocks does not seem very likely.

TABLE 30.—*Average of the percentage of TiO_2 and P_2O_5 in the metasedimentary, plutonic, and metasomatic rocks*

[In weight percent]

Rock type	TiO_2	P_2O_5
Plutonic rocks in Headquarters quadrangle (table 10, Nos. 292, 322, 190, and 379)	0.62	0.22
Plutonic rocks near Orofino (table 15)	.68	.16
Metasedimentary rocks in Headquarters quadrangle (table 2)	.58	.06
Metasedimentary rocks near Orofino (tables 12 and 14)	.49	.12
Metasomatic rocks near Dent (table 18, Nos. 471a and 471b; table 20, No. 119; table 22, No. 13)	3.46	.26

Earlier authors (Anderson, 1933 and 1942; Johnson, 1947) have stated, without further discussion, that emanations from the Idaho batholith were responsible for the introduction of various elements into metasedimentary country rock in the Orofino region. Thus the metasomatism should be associated with the intrusion of the granite batholith, the first outcrops of which are 10 miles south (near Greer) and 15 miles east (near Headquarters) of the Dent area. (See pl. 1.) Between Dent and the first outcrops of the batholith there are thick layers of crystalline schist and quartzite in which there may be only a few signs of metasomatic introduction. Thus the relation between the batholith and the metasomatism seems not to be a simple contact phenomenon. The metamorphism that preceded the second period of metasomatism described from the Dent area decreases toward the north, away from the batholith, and is obviously associated with the formation of the batholith.

Metamorphism was contemporaneous with deformation, as shown in the orientation of the recrystallized biotite and quartz parallel to the foliation that generally parallels the bedding. Where the bedding and foliation planes are wrinkled around the lineation of the tectonite, the biotite flakes follow the wrinkled planes. They are not bent but recrystallized. In the biotite quartzite layers the scattered biotite flakes are well alined parallel to the lineation. In places the main folding may be later than the wrinkling and folding around the lineation. The orientation of the minerals through recrystallization, however, is due mainly to

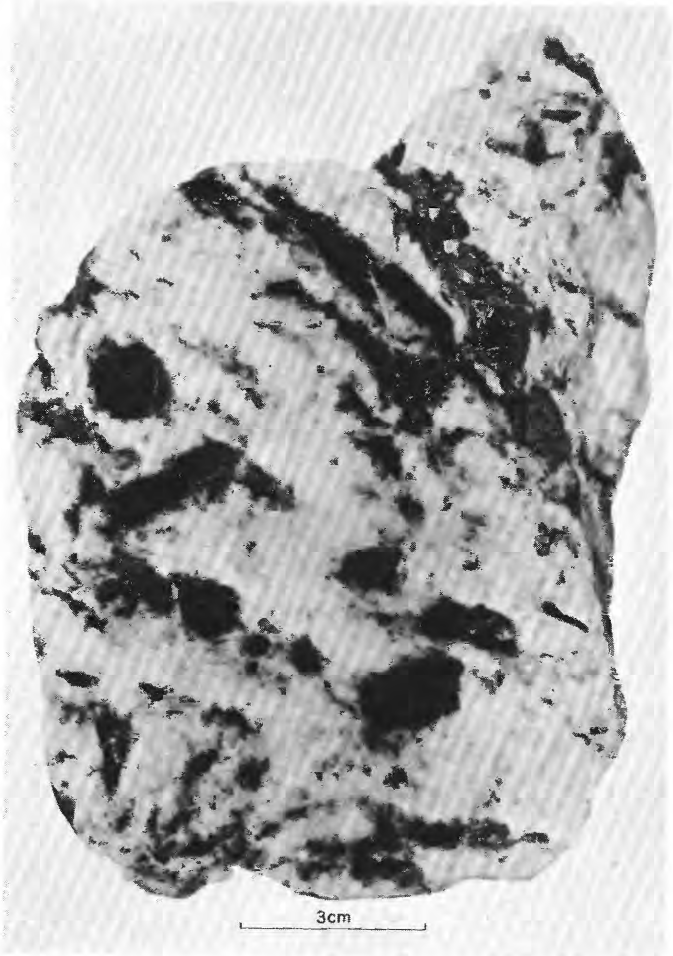


FIGURE 91.—Gabbro pegmatite with large hornblende crystals. Near Greer.

the act of deformation, which is responsible for the lineation.

Rearrangement of material probably began during the regional metamorphism. This must be judged on the basis of correlation and comparison of the chemical composition of the metamorphic rocks of this area with that of the known Belt series farther north. Thus both the structural and the metamorphic evolutions are fairly complex and represent sporadic release of energy through time. Insofar as the metasomatism is related to the emplacement of the batholith, one has to consider the age relations of deformations, intrusions, and metasomatic changes. The quartz diorite of the border zone of the batholith shows a planar, and in places also linear, structure, parallel to those of the country rock. The contacts are mostly concordant. Interfingering contacts are found where the intrusive body cuts across the beds of the older formations. Fragmentation is rare. The mode of emplacement of the quartz diorite must have been partly that of replacement and the time post-folding.

The satellitic bodies of norite, gabbro, and quartz diorite are probably older than the border zone of the batholith (intrusion phase 2, in table 25). The later phase of metasomatism may be associated with the latest phase of the intrusion of quartz diorite. This conception is supported by the occurrence of gabbro pegmatite and anorthosite dikes, which cut diorite, gabbro, and country rocks west and south of Orofino. The gabbro pegmatite dikes consist of hornblende crystals as much as 20 cm long, large plagioclase grains, and some biotite flakes (fig. 91). The anorthosite dikes consist of almost pure plagioclase (An_{52}) and occur parallel to the joints and shear planes. Thus the constituents of hornblende, biotite, and calcic plagioclase seem to have been highly mobile, and most likely they had the same source as the quartz diorite and gabbro in which they occur. The question of the original source of cafemic constitutents is therefore connected with the genesis of the igneous rocks.

SOURCE OF ALKALI FELDSPARS

In several localities the metasomatic rock is tonalite with fewer dark minerals than the quartz diorite and with more sodic plagioclase, usually An_{25-40}. The primary plagioclase in the gneiss is of the same composition, and it is therefore possible that the plagioclase in the metasomatic tonalite was derived from the older rocks. Thus the tonalitic layers in the quartz diorite may represent recrystallized biotite-plagioclase gneiss layers with only minor metasomatic changes. Yet when calcium is introduced, sodium is removed from the original sodic plagioclase in the metasedimentary rocks, during their transformation to quartz diorite. The material of the quartz-albite-muscovite dikes that fills joints in the metasomatic gneiss south of the coarse-grained metasomatic quartz diorite opposite the mouth of Canyon Creek (pl. 3) may have been derived from the elements removed from the sedimentary rocks during their transformation. Furthermore, the removed elements would cause regional metasomatic changes in the surrounding rocks, such as the crystallization of secondary quartz and sodic plagioclase, and would give rise to the metasomatic rocks of tonalitic and trondhjemitic composition.

The source of the introduced potassium was either the quartz monzonite magma or the metamorphic rocks. All rocks west of the batholith are exceptionally poor in potassium feldspar, but this mineral is fairly common in the Belt series farther north. The secondary microcline and orthoclase may thus include the potassium feldspar removed from the older rocks during their basification.

GENESIS OF QUARTZ DIORITE

The genesis of quartz diorite, like that of granite, can be explained in two ways. The quartz diorite can be either a product of crystallization differentiation of a magma or a result of metasomatic transformation and partial remelting of older sedimentary rocks. Among the supporters of a metasomatic origin of granitoid rocks the differences in the opinions about the mode of migration and sources of introduced material are striking. In the view of those who support solid diffusion, introduced ions are derived from the older solid rock by activation (Bugge, 1946; Ramberg, 1952b; Reynolds, 1946, 1947). According to this school of thought, the high content of femic minerals in the quartz dioritic border zone of the Idaho batholith should be due to a migration of cafemic constituents (calcium, iron, and magnesium) from the center of the batholith; that is, from the area now occupied by quartz monzonite and orthoclase-rich granite. Other students of the problem consider a liquid phase necessary, and the source of the introduced material is thought to be either a highly metamorphosed rock in which a differential resolution and partial melting have started, or a primary igneous magma (Eskola, 1932, 1950).

The structure and texture of the quartz diorite suggest an igneous origin. If its composition is due to the migration of elements, this migration must have taken place during an earlier phase and the metasomatic rocks formed in this manner were partially melted and mobilized. In the Orofino region it can be seen that rocks chemically and mineralogically identical with the intrusive quartz diorite and tonalite *can* be formed by metasomatic introduction of certain elements into the sedimentary rocks. If these metasomatized rocks were mobilized, as are many Precambrian palingenetic granites, or if they were recrystallized to such an extent as to obliterate the older structures, they would appear identical with the intrusive quartz diorite and tonalite. If a palingenetic magma were formed from them, it would behave like an intrusive magma and be capable of differentiation and intrusion. Thus, rising temperature and pressure would cause first metamorphism and migration of materials and finally a formation of mobile magmas. When these magmas or their differentiates intrude into higher levels, they bring heat and solutions, which in turn cause alterations of the country rocks similar to those effective in the generation of the magma. Thus, more and more sedimentary rocks are converted to plutonic rocks through the repeated metasomatic and palingenetic processes.

To test the idea that the quartz diorite and granite magmas were generated from the older sedimentary rocks by a rearrangement of materials at depth, the chemical compositions of the rocks concerned were compared. The mean composition of the older metamorphic rocks belonging to the Belt series was calculated on the basis of chemical analyses made of five representative layers in the formations west of Headquarters (analyses of specimen No. 246, 247, 248, 254, and 289 in table 2; the mineralogical composition of these layers is indicated in table 2). The mean composition was weighted in terms of estimated volume of each rock type.

If this mean composition (table 31) is compared with the composition of the quartz diorite, No. 292 (table 10), which is typical of the intrusive rocks in the region, we get the approximate total amounts of elements that should be removed from and introduced into the sedimentary strata to produce a quartz dioritic rock. Column 4 in table 31 shows the amounts of various oxides and ions and column 4 in table 32 the number of cations in a standard cell removed and introduced during the possible "dioritization." Comparison with tables 19 and 23 shows that qualitatively the exchange of elements during this dioritization would be similar to the basification now found around the quartz diorite. Homogenization and recrystallization of the contact zone would thus produce a rock whose composition would be close to that of the quartz diorite. Partial or total melting of the contact zone of the batholith would give rise to the formation of quartz dioritic magma.

Where the pure Revett quartzite occurs, more quartz should be removed and all the other oxides should be introduced. The middle part of the Wallace formation is considerably richer in calcium than the mean composition and could be a source of added calcium. Column 5 in table 32 shows that iron, magnesium, calcium and titanium are removed from the metasedimentary formations if they are converted to granite, No. 758. The removal of iron, magnesium, and calcium is also shown in figure 92. Thus the sedimentary rocks could be a source of these elements. The total amount of silicon may remain constant, as considerable silicon is present in the granite and in veinlets in the older rocks. It is noteworthy that alkalies must be added whether the older rocks are converted to quartz diorite or to granite (fig. 92). During the dioritization only sodium is introduced, but during the formation of granite considerable amounts of both sodium and potassium should be added.

A part of the potassium in the granitic rocks of the batholith probably originates in the metasedimetary rocks of the contact zone. As mentioned earlier the Belt series farther from the batholith contains a considerable amount of microcline, whereas in the zone next to the batholith, that is in the area under discus-

TABLE 31.—*Comparison of the chemical composition of quartz diorite, granite, and the weighted mean composition of the metasedimentary rocks*

Constituent		1		2		3		4	5	6
		292 (table 10, p. A-27)		758[1]		Mean composition of metasedimentary rocks from table 2		Oxides introduced (+) or removed (−) between the mean and diorite (No. 292)	Oxides introduced (+) or removed (−) between the mean and granite (No. 758)	Oxides introduced (+) or removed (−) between diorite (No. 292) and granite (No. 758)
		Quartz diorite		Granite						
Conventional Symbol	Symbol rearranged for cation percent	Weight percent	Cation percent	Weight percent	Cation percent	Weight percent	Cation percent	Weight percent		
SiO_2		59.16	55.40	77.43	72.55	72.5	69.65	−13.3	+4.9	+18.3
Al_2O_3	$AlO_{3/2}$	18.15	20.02	12.47	13.77	12.1	13.73	+6.1	+.4	−5.7
Fe_2O_3	$FeO_{3/2}$	1.57	1.10	.22	.16	.5	.36	+1.1	−.3	−1.4
FeO		4.08	3.19	.84	.66	3.9	3.21	+.2	−3.1	−3.2
MnO		.10	.08	.04	.03	.1	.14			
MgO		3.13	4.36	.13	.18	2.1	3.13	+1.0	−2.0	−3.0
CaO		6.47	6.49	.51	.51	3.8	3.97	+2.7	−3.3	−6.0
Na_2O	$NaO_{1/2}$	3.71	6.73	3.77	6.84	1.6	3.08	+2.1	+2.2	
K_2O	$KO_{1/2}$	1.47	1.75	4.37	5.22	1.7	2.16	−.2	+2.7	+2.9
TiO_2		.85	.60	.06	.04	.6	.43	+.2	−.5	−.8
P_2O_5	$PO_{5/2}$	.30	.24	.04	.03	.1	.13	+.2		−.3
CO_2		.03	.04	.01	.01	.0	.01			
H_2O+		.79	(2.46)	.19	(.60)	1.1	(3.63)	−.3	−.9	−.6
H_2O-		.05		.03		.1				
Total		99.86	100.00	100.11	100.00	100.2	100.00			
	O		160.27		172.97		171.07			
	OH		4.92		1.20		7.26			
	Total anions		165.19		174.17		178.33			

[1] Ruth H. Stokes, analyst.

TABLE 32.—*Comparison of the number of cations in a standard cell of quartz diorite, granite, and the weighted mean composition of the metasedimentary rocks*

Cations	1	2	3	4	5	6
	Diorite No. 292 (table 10)	Granite No. 758 (table 31)	Mean composition of metasedimentary rocks (table 31)	Cations introduced (+) or removed (−) between the mean and diorite	Cations introduced (+) or removed (−) between the mean and granite	Cations introduced (+) or removed (−) between diorite and granite
	Number of cations in standard cell					
Si	53.7	66.6	62.5	−8.8	+4.1	+12.9
Al	19.4	12.7	12.3	+7.1	+.4	−6.7
Fe‴	1.1	.1	.3	+.8	−.2	−1.0
Fe″	3.1	.6	2.9	+.2	−2.3	−2.5
Mn	.1	.0	.1	.0	−.1	−.1
Mg	4.2	.2	2.8	+1.4	−2.6	−4.0
Ca	6.3	.5	3.6	+2.7	−3.1	−5.8
Na	6.5	6.3	2.8	+3.7	+3.5	−.2
K	1.7	4.8	1.9	−.2	+2.9	+3.1
Ti	.6	.0	.4	+.2	−.4	−.6
P	.2	.0	.1	+.1	−.1	−.2
C	.0	.0	.0			
H_2O	(2.4)	(.6)	(3.3)	−.9	−2.7	−1.8
Total	96.9	91.8	89.7			

sion, this mineral is very scarce. Because of large amounts of granitic rocks, it seems necessary to assume that at least some alkalies from some source other than the older rock must have been added if the sedimentary formations were converted to quartz diorite and granite. This source was probably a primary igneous magma that was squeezed upward or into which the lower parts of the sedimentary formation were pressed during the mountain building. Heat and solutions from this primary magma may have started the metamorphic and metasomatic processes in the deeply buried sedimentary formations. These processes first formed metasomatic igneous-looking rocks and later generated new magmas by partial melting. Judging from the great thickness of the sedimentary formations that have been digested by the igneous rocks and from numerous large inclusions of these older rocks in the Idaho batholith, it is not likely that any rocks of the composition of the primary igneous magma would be exposed among the plutonic rocks of this region. Rather, all plutonic rocks are differentiates of secondary magmas that were generated at various depths. The difference in the chemical composition of earlier (quartz dioritic) and younger (quartz monzonitic) magmas could be explained as follows:

During the Nevadan folding, when the magmas were being generated (through combined processes of metamorphism, metasomatism, and finally partial melting of older rock), a basic zone around the granitic magma was formed at depth. The upward migration of magma started, and movement had to be through the basic roof. The magma was contaminated, and the palingenetic magma that was formed from the country rock (that is, from the basified zone) was added to it. Therefore, all magma that reached the level now exposed was more basic than the primary granite magma that was forming at depth. This magma had a composition of quartz diorite and solutions from it were responsible for the basification of the present contact aureole. Later differentiates of the primary magma, still retained at depth, migrated upward (rocks of the quartz monzonite series). The contamination was less because the differentiates were emplaced through the previously metamorphosed rocks at shallower depth. Typically, the metasomatic contact aureoles around the intrusive quartz monzonite are rather narrow compared with those around the intrusive quartz diorite.

The processes described in this paper may be common during a certain phase of intrusion in the contact aureoles of great batholiths even if no traces of them are preserved. As soon as enough melt is formed and the mobile stage reached, the palingenetic rocks behave like plutonic rocks of juvenile magmatic origin. They become intrusive and obliterate completely or in part their metasomatic aureoles and thus conceal or confuse the history of their formation. Many large plutons show a metasomatic (migmatitic) contact on one side of the pluton and an "intrusive contact" on the other side (Hietanen, 1943; 1951, p. 592). On the metasomatic side the magma was less mobile and hence did not shatter the metasomatic contact aureole.

MODE OF MIGRATION OF MATERIAL

Field observations and thin-section studies show that the newly formed minerals are concentrated along cracks, joints, fissures, cleavage, and other structural surfaces that may have served as mechanical passageways during the metasomatism. The migration parallel to the structures is conspicuously more extensive than that across the structures. This mode of occurrence suggests that the elements were introduced by the solutions in channels provided by the structures.

The anorthosite veins and dikelike bodies may represent larger channels along which the calcic plagioclase was supplied to the surrounding schist. The elongation of plagioclase grains parallel to the walls in these veins suggests a flowage in a channel. As a rule, the channels that offer the least resistance are used by the solutions advancing into the country rock. The occurrence of biotite in the interstices of the earlier minerals, and a brown coloration (probably due to iron compounds) of the grain boundaries in the quartzite next to the metasomatic basic rock suggest that in the rock itself intergranular film may have served as a channel. Probably in this rock the ions diffused faster in the aqueous solution filling the pores and subcapillary openings than they did through the solid crystals. The introduced minerals form small pods and lenses along and at the ends of these intergranular channels. The earlier minerals, which had became unstable in the presence of the newly introduced ions in the new temperature-pressure field, would form new stable compounds with the introduced ions. A good example of this reaction is the replacement of diopside by hornblende in the diopside-plagioclase gneiss during the metasomatism. This replacement indicates that the solutions carrying introduced elements were aqueous. Part of the escaped silica was precipitated along the fractures and fissures and was thus more likely in a solution than as a dispersed phase of ions.

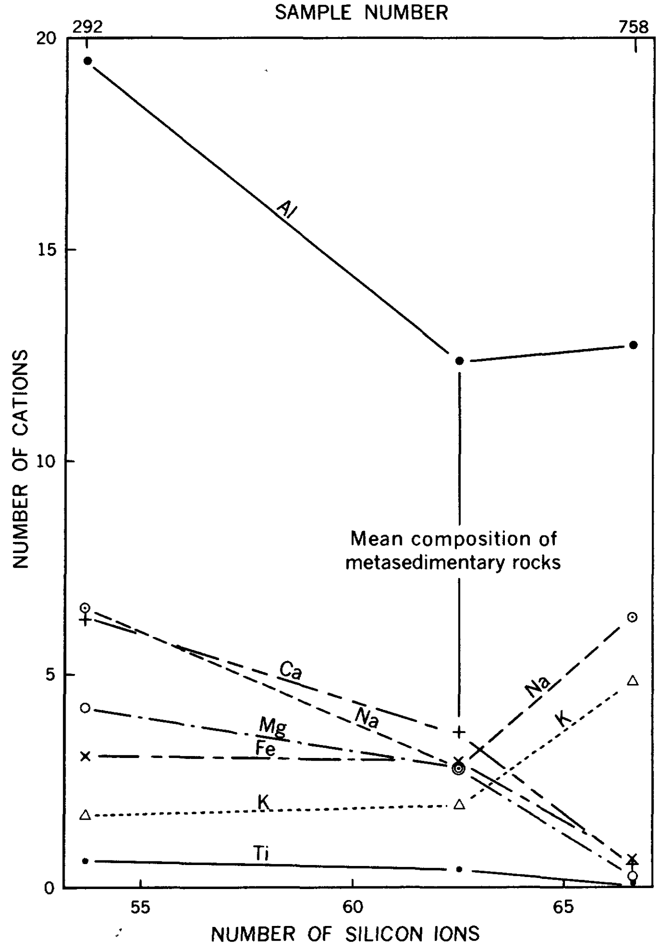


FIGURE 92.—Graph showing the number of cations in standard cells of a diorite and a granite, and weighted mean composition of mestasedimentary rocks in the Headquarters quadrangle.

The same solutions that supplied material for the abundant ferromagnesian minerals and andesine in the schist in areas north of Orofino gave rise to the formation of metasomatic quartz diorite and tonalite near Orofino. The difference in the mode of deposition of the introduced elements was created by the physical conditions: during the first phase of metasomatism the orogenic stresses and a pressure gradient helped the solutions to move into the country rock. The fluids were pressed through open channels such as joints and cracks. In the schist the planes of foliation and shear surfaces acted as main channels for introduction of elements. There was no differentiation of femic and salic constitutents on such a scale as occurred during the later phase. During this later phase of metasomatism the solutions filling the pores and subcapillary openings of the rocks that were metamorphosed during the earlier phase must have been fairly stationary or must have moved only slowly because the permeability was reduced. The principal mode of transport then was most likely the diffusion of ions in a stationary pore liquid, and the rate of diffusion of the various ions gave rise to the differentiation of femic and salic constituents.

My field observations are in contrast with those made by Holmes and Reynolds (1947) in the Dalradian series in Ireland and with those of Perrin and Roubault (1941) in the Alps of Savoy. They found that cracks, pores, and crystal boundaries were avoided during the replacement, and therefore they concluded that the metamorphic front advanced through diffusion in a solid rather than through solutions permeating the rock.

Near Dent and in the Headquarters quadrangle, diffusion through crystals was probably active where metasomatic minerals grew in earlier ones. The isolated secondary hornblende, for instance, may have been crystallized in a coarse-grained quartz whose undulous structure—due to deformation prior to metasomatism—is not in the least disturbed and shows no visible channels.

The rate of diffusion in a solid is not well known. The experimental work hitherto published suggests rates that are geologically insignificant. Jagitsch (1949) obtained low rates of diffusion for Na_2O in the system Na_2O-Al_2O_3-SiO_2. Jensen (1952) made a study of the solid diffusion of radioactive sodium in perthite and got the diffusion coefficients of 10^{-11} and 10^{-12} cm^2/sec at a temperature of 550° C, which is of the same magnitude as that of Rosenqvist (1949) for the diffusion of Ra^{++} in albite at a temperature near the melting point of albite. It is well known that the structure of the surface layer of the minerals differs from that of the adjacent inner layer and that this surface layer is capable of absorbing molecules and ions. Reactions between the ions absorbed by a surface layer may take place with considerable ease. Also the rate of diffusion along the surface is found to be faster than through the main body (Barrer, 1941; Jost, 1937; Hedwall, 1942). The intergranular film in the rocks is comparable with this surface layer, and it is possible that the diffusion along it would be greater than that through the grains. Mosaic structure develops in crystals under mechanical stress, and ions may migrate along these mosaic fissures more freely than through the close-packed parts of the same crystal. Thus it seems that the diffusion of ions through cracks, intergranular film, and other channels offering the least resistance may be a more effective mode of transport of ions than that through the crystals.

Water vapor has been found to accelerate the rate of diffusion. For instance, Bowen and Tuttle (1950) found that water and water vapor under pressure greatly facilitate the mobility of ions during the unmixing of alkali feldspars. Turner and Verhoogen (1951, p. 401) have suggested that even minute amounts of water greatly accelerate surface reactions and surface diffusion. Chao (1951) considers that ions are transported mainly by diffusion though a stationary pore liquid rather than by solid diffusion or by a moving liquid.

Quartz is generally considered one of the stablest minerals in the rocks; yet in certain physical and chemical conditions this mineral is easily replaced. According to experimental work by Kennedy (1950), solubility of quartz starts to rise rapidly with rising pressure at temperatures between 380° and 400° C. Extrapolation from his values suggests that the specific volume of water vapor in that temperature range approaches that of water at a pressure of about 4000 bars, a pressure that could prevail about 15 kilometers below the earth's surface. It is thus possible that the solutions which carried the heavy elements Mg, Fe, Ca, and Al were supercritical watery vapors moving away from the batholith.

The field observations and thin-section studies show that biotite moved ahead of the other minerals into the country rock. Large biotite flakes fill cracks and grow across contacts from the intrusive rock to the metamorphic rock. The advance of the biotite ahead of the other minerals may offer an alternative explanation for the seemingly easy replacement of quartz in the mapped area. It is well known to chemists that quartz is easily dissolved in alkaline solution. Thus, if the K and OH ions advanced ahead, they were capable of dissolving the quartz, an excess of which was then removed. The removed quartz was either deposited as veins in the surrounding rocks or was incorporated into the magma from which the metasomatizing solutions came. Biotite seams are extremely common along the contacts between the metasomatized rocks and the remnants of parent rocks (fig. 61A), around the inclusions of country rock in the tonalite, around the pegmatite pods and veins, and around the hornblende-rich masses, forming thus a frontal zone around the metasomatized areas. The hornblende and the main bulk of the accessories followed the biotite, and the calcic plagioclase was latest during this phase of metasomatism. The difference in time and rate of migration of ions probably caused a difference in the place of deposition and thus made the metasomatic differentiation possible.

This segregation of femic minerals and calcic plagioclase into separate lenticles can be compared with the growth of porphyroblasts. In both types of segregation just a few kinds of ions accumulate in certain spots to form one kind of mineral. In the monomineralic lenticles this segregation takes place only on a larger scale, so that instead of an individual crystal there is a group of crystals. The diffusion of ions in a stationary condensed pore liquid makes the selective segregation possible, whereas transportation of elements by moving liquids would scarcely permit such a differentiation, as occurs in the Dent area.

As a general conclusion, the mode of transfer of substances changes during the structural and metamorphic evolution. At early stages, during the deformation period, matter is transferred on a large scale by intrusion of magmas and by liquids and gases streaming through the rocks. The mechanical passageways play an important role during this stage. Later, when recrystallization has reduced the permeability of the rock, when deformational movements have ceased, and when the pressure gradient is too small to force the liquids through the pores, the diffusion in the stationary pore liquid becomes significant. Formation of dikelike monomineralic bodies would be possible when the late sporadic movements press the differentiated pore solutions into the country rock or when the fissures are filled by material extracted from the country rock. From the pore solution the ions may advance through the solid rock by solid diffusion. The rate of diffusion in a solid is determined by physical and chemical conditions in the rock, and diffusion probably becomes more effective at lower levels in the earth's crust and in the surroundings of large magma reservoirs.

TEMPERATURE DURING METASOMATISM

The temperature and pressure during the metasomatism described in this paper were below those in which granitic melt starts to form. The rocks were metamorphosed before the metasomatism to the sillimanite subfacies of the amphibolite facies, as shown by the common occurrence of assemblages sillimanite-almandite-biotite, plagioclase-diopside, calcite-grossularite-diopside, and plagioclase-hornblende-biotite, all with excess of quartz. As a rule the diopside is replaced by hornblende during the metasomatism. This replacement may indicate a lowering of temperature. However, garnet amphibolite—a typical rock in the sillimanite subfacies—was formed by the metasomatic introduction of hornblende and plagioclase into garnet-biotite schist. The common lack of hypersthene indicates that the temperature and pressure were not high enough for the appearance of typical mineral assemblages of the granulite facies. According to Barth (1952, p. 338 and 343) the temperature for the formation of the amphibolite facies is about 400° C and at about 500° C the mineral associations typical of this facies become unstable; these are lower values than the temperature limits given by Turner (1948, p. 285). The occurrence of sillimanite and muscovite together in many layers in the schist proves that the temperature during the metamorphism did not exceed 650° C (Yoder and Eugster, 1953, 1955).

Kullerud and Neumann (1953, p. 154–155) set the temperature range for granitization and metamorphism in amphibolite facies in the Rendalsvik area, Norway at 440° C±25° C at a depth between 3¾ and 11 km, using the iron content of sphalerite as a geological thermometer.

Locally, along the zones of strong shearing, muscovite and chlorite appear instead of biotite, and garnet is partly chloritized. Grains of brown rutile and crystals of zircon abound as accessories. There is an addition of sericite and decrease in the amount of quartz and sillimanite in some of these shear zones. This exchange means an increase of potassium and a decrease of silicon, as is common in the frontal zones of metasomatized areas near Dent, but the mineral assemblage indicates lower temperature-pressure conditions. Studies should be extended northward from the area shown in plate 1 to determine whether similar changes appear in the mineral assemblages in the zones farther from the batholith. The temperature during the deposition of the microcline along irregular replacement veins south of locality 471 must have been above the exsolution temperature because the centers of the microcline grains contain small perthitic albite blebs and stringers. Small albite grains contain myrmekitic quartz along their borders, but it is not clear whether this albite is earlier or later than the microcline. Two miles northeast of Bungalow orthoclase and sodic plagioclase replace quartz. This orthoclase is perthitic, suggesting a fairly high temperature during the metasomatism.

Conversion of the kyanite to andalusite and sillimanite and of sillimanite to kyanite in some metasomatized schists in the Boehls Butte quadrangle suggests that the temperature and pressure were close to those in which these three polymorphs of Al_2SiO_5 are stable (Hietanen 1956), perhaps close to 500° C.

RELATION OF METASOMATIC ROCKS TO THE BASIC FRONT

As described on the earlier pages the country rocks of the Idaho batholith are enriched in iron, magnesium, calcium, and aluminum. This enrichment occurred before the intrusion of the granitic batholith; it probably occurred during the intrusion of quartz diorite with which it is associated. It has been shown (p. A–108) that the same elements which were introduced to the country rocks should be removed from the metasedimentary rocks if they are converted to granite or to quartz diorite. As the lowest melting materials in the metasediments are quartz and feldspars the minerals rich in iron, magnesium, calcium, and aluminum will be left behind when a granitic melt starts to form. This was the first phase in the formation of the batholith and probably it continued over a considerable length of time. Later the temperature rose and also the country rocks that were enriched in cafemic constituents started to melt forming quartz dioritic magma. The excess of iron, magnesium, calcium, aluminum, and sodium now became mobile and migrated to country rocks of the quartz diorite forming the basic front. At a still later date granite magma, which had been retained at depth, invaded its roof in great quantities.

In some respects, the chemistry and mode of occurrence of the basified rocks in Idaho resemble those of basic front described by Reynolds (1947). In detail, however, many differences became evident in the course of this study; many of them due to a higher temperature and pressure. The temperature in the country rocks of the Idaho batholith during their metamorphism and metasomatism is thought to have been only slightly lower than the melting temperature of granite, perhaps close to 500° C. The transport of material is suggested to have been first by flowage in open channels and later by diffusion in stationary pore solution. During the first phase calcic plagioclase and femic minerals were deposited together, whereas during the later phase a differentiation of salic and femic constituents is common. No potassium was introduced during the Fe-Mg-Ca metasomatism. Rather, the potassium was removed, a part of it formed potassium-rich minerals (orthoclase and biotite) in the country rock, but another part probably migrated to the granitic batholith, where more alkalies were needed to form a granitic melt from the metasedimentary rocks.

The minor elements—mainly Ti, Zr, Ba, Sr, V, P, and La—were concentrated in the basic metasomatic rocks with Fe, Mg, Ca, and Al. During the differentiation, Ti, Zr, and V were concentrated with Fe and Mg, Sr with Ca, and Ba with K. Study of the minor elements shows that some minor elements that are rare in basic igneous rocks may be concentrated with iron and magnesium during the metasomatism. Therefore, a high concentration of such minor elements (for example, zircon) in some basic rocks may offer proof of the metasomatic origin of these rocks.

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Anorthosite and Associated Rocks in the Boehls Butte Quadrangle and Vicinity Idaho

By ANNA HIETANEN

METAMORPHIC AND IGNEOUS ROCKS ALONG THE NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

GEOLOGICAL SURVEY PROFESSIONAL PAPER 344-B

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METAMORPHIC AND IGNEOUS ROCKS ALONG THE NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

ANORTHOSITE AND ASSOCIATED ROCKS IN THE BOEHLS BUTTE QUADRANGLE AND VICINITY, IDAHO

By Anna Hietanen

ABSTRACT

Three large and several small bodies of anorthosite occur in the schist of the Prichard formation of the Precambrian Belt series in the Boehls Butte quadrangle and vicinity in the southern part of the Idaho panhandle. The Prichard formation, the oldest formation of the Belt series, consists mainly of garnet-mica schist with one or two quartzitic units in the middle of the formation. The lower quartzite unit of the Prichard formation grades laterally from white and gray quartzite to lime-silicate rock near the Little North Fork of the Clearwater River. The anorthosite occurs in the schist just under the lower quartzite unit. The composition, structure, and texture of the anorthosite are in no way affected by the change in the lithology of the overlying quartzite beds. As a rule, a thin layer of aluminum-rich schist occurs between the anorthosite and the overlying quartzite or lime-silicate rocks, and numerous thin, sheetlike inclusions of schist and aluminum-silicate minerals (kyanite, andalusite, and sillimanite) occur in the anorthosite. The schist under the anorthosite is coarse-grained mica schist with abundant kyanite, andalusite, sillimanite, garnet, and occasional staurolite. All schist next to the anorthosite bodies contains numerous large round grains of calcic plagioclase (An_{25-45}), and an amount of quartz far less than that in the schist farther from the anorthosite.

Locally the anorthosite contains thin, dark layers rich either in biotite and aluminum-silicates or in hornblende, or in places in all these minerals. These dark layers are parallel to the bedding of the country rocks; they are extremely common in the southernmost large body and occur locally in the other bodies. The central parts of the two northern large bodies consist of massive coarse-grained hornblende-bearing anorthosite in which sheetlike micaceous portions are found only in a few localities. Almost all of the anorthosite contains two types of plagioclase, large round grains of andesine (An_{45-48}) and small grains of bytownite (An_{85}), the latter occurring either between the andesine grains or as platelike inclusions in the andesine. Neither plagioclase is zoned. A few small bodies consisting almost exlusively of bytownite are interbedded with hornblende and biotite gneisses near the contacts of anorthosite. Several small bodies of garnet amphibolite occur in the schist near the anorthosite masses and some are included in the anorthosite.

The change from garnet-mica schist through aluminous schist rich in oligoclase and andesine to anorthosite could be due to a gradual change in the chemical composition of the sediment from which these various layers crystallized. However, where the contact zone was studied in detail, it was found that the change from garnet-mica schist to aluminous schist rich in plagioclase takes place paralled to as well as across the bedding and that the distribution of plagioclase is highly irregular. Moreover, the mineralogy, structure, and texture of the isochemically metamorphosed sedimentary layers rich in andesine differ considerably from those of the plagioclase-bearing schist that occurs next to the anorthosite. Yet, the inclusions, structure, texture, and mineralogy of the anorthosite indicate that much, if not all, of the anorthosite was formed by metamorphic and metasomatic processes rather than by intrusion.

The occurrence of two types of plagioclase, andesine and bytownite, in separate grains proves that the anorthosite in the Boehls Butte quadrangle is not a product of simple fractional crystallization of a magma. The layers of bytownite anorthosite interbedded with gneisses are most likely of sedimentary origin, perhaps metamorphosed shaly limestone. In the anorthosite that contains two types of plagioclase, andesine crystallized later as a result of sodium metasomatism. Thus all anorthosite could have been shaly limestone in which bytownite crystallized during the regional metamorphism and andesine formed later as a result of reaction between the early bytownite and introduced sodium. A small portion of coarse-grained anorthosite consists of labradorite with some tiny platelike inclusions of bytownite. Probably in this rock the reaction between the bytownite and the sodium-rich solutions proceeded further and produced an almost homogeneous labradorite anorthosite, thus nearly the variety which is common in the large Adirondack-type anorthosite masses.

The conclusions are supported by 7 new chemical analyses of minerals and by 20 quantitative chemical and spectrographic analyses of anorthosite, its country rocks, and associated metamorphic rocks.

INTRODUCTION

THE ANORTHOSITE PROBLEM

The very earliest investigators classed anorthosite and other monomineralic rocks with the sedimentary formations, but since the turn of the century it has been generally agreed that anorthosite is of igneous origin. This change in the development of thought is clearly indicated in Adams' report (1896) on the anorthosite near Montreal, Quebec. Of his fieldwork in 1891, he writes (in Selwyn, 1892, p. 40): "These [anorthosites], together with some of the associated gneisses and limestones, were formerly supposed to constitute a separate overlying series, to which the name Upper Laurentian was given. Their boundaries have, however, now been traced out, and their stratigraphical relations determined, and they have been found to be without doubt igneous rocks." Most anorthosites described since the time of Adams occur in plutonic surroundings and are undoubtedly of igneous origin. The associated rocks are usually either gabbro, granite, and syenite or rocks of the charnockite series. The crystallization differentiation of anorthosite from a normal basaltic magma has been proved in many sill-like stratiform intrusions in which anorthosite was formed near the roof of the sill above the gabbroic rock. Examples of such occurrences of anorthosite in North America are at Stillwater, Mont.; Bay of Islands, Newfoundland; Duluth (Winchell, 1899) and Pigeon Point (Grout, 1928), Minn.; and Preston, Conn. Similar stratiform sheets occur in Bushveld complex (Wagner, 1924) and in Sierra Leone in Africa.

The study of these sheetlike occurrences proved that gravitative crystallization differentiation in place, as suggested by Bowen (1917), gave rise to a separation and accumulation of labradorite crystals on the upper parts of a floored magma chamber. The field relations, for example, in the diabase sill in Pigeon Point, Minn. (Grout, 1928), leave no doubt that the anorthosite there was derived from a basaltic magma by crystallization differentiation. The anorthosite in Vancouver Island also occurs with intrusive gabbro (Clapp, 1917; Cooke, 1919). In many other areas, however, the large amount of anorthosite as compared with the known gabbroic rocks makes it difficult to adopt a similar explanation. Such large bodies are, for example, the Adirondack, New York (Buddington, 1939; Balk, 1930, 1931); St. Urbain, Quebec (Mawdsley, 1927); Morin district, Quebec (Adams, 1896, Osborne, 1949); Lower Romaine River area, Quebec (Retty, 1944); Saguenay district, Quebec (Dresser and Denis, 1944); St. Paul, Labrador; Nain, Labrador (Wheeler, 1942); and many others on the Labrador peninsula; Honeybrook quadrangle, Pennsylvania (Smith, 1922); Laramie Mountains, Wyoming (Fowler, 1930); San Gabriel Mountains, California (Miller, 1931); Bergen district, Norway (Kolderup, 1903, 1936); Egersund, Norway (Kolderup, 1914, Michot, 1939); Sogn district, Norway (Goldschmidt, 1916; Vogt, 1924); Volhynia, U.S.S.R. (Polkanov, 1937); and Angola, Africa (Mouta and O'Donnell, 1933). Many of these large bodies of anorthosite, with only a subordinate amount of gabbroic rocks in their surroundings, were earlier thought to represent crystallization products of a liquid anorthosite magma (Miller, 1914). It was pointed out by Bowen (1917), however, that the geologic evidence was against the hypothesis of a liquid anorthosite magma, because the surrounding rocks do not show signs of such high temperature as would be required by a molten plagioclase. Bowen suggested that the anorthosite was separated from a gabbroic magma by the separation of the femic constituents while the plagioclase crystals remained suspended. Later, when the liquid is lighter (composition of diorite-syenite) the plagioclase (now labradorite) would accumulate by sinking and give masses of anorthosite. The residual liquid would be syenitic in composition. The movements of such masses of anorthosite would give rise to protoclastic structures, which are common in many large anorthosite bodies. If anorthosite has this mode of origin, no truly intrusive anorthosite dikes should occur.

Bowen's suggestion was criticized by Cushing (1917) and Miller (1918), who pointed out that syenite in the Adirondack region is not a differentiation product from the same magma as the anorthosite, but is a distinctly younger intrusion. An opposing view was suggested also by Goldschmidt (1922), who, in his investigation of a Norwegian anorthosite, came to the conclusion that plagioclase crystals separated from the parent magma by rising. According to Goldschmidt, the anorthosite in Norway is a differentiate of the so-called mangerite magma that differs from the normal basaltic magma essentially because of its dryness. The scarcity of water would make the mangerite magma heavier than the normal basaltic magma and the labradorite crystals would accumulate by rising. Vogt (1924, p. 91) compared the specific gravities of the assumed magmas and the plagioclase in the anorthosite in Norway and concluded that the plagioclase crystals must have been sinking also in the dry magma. Suter (1922) suggested that anorthosite is universally a member of the mangerite magma. The scarcity of water (or OH) in this magma gives rise to a series of rocks in which minerals poor in OH are predominant. Hypersthene and potassium feldspar substitute for biotite in the granites and give rise to a rock series known as charnockites. In the more mafic rocks of this series hypersthene is a predominant dark mineral, but augite and biotite also may occur.

The possibility of a liquid anorthosite magma was raised again by Mawdsley (1927, p. 33) in his study of the St. Urbain area, Quebec. He found andesine anorthosite cutting the older labradorite anorthosite; this seemed "* * * to imply that the andesine anorthosite as a whole did exist in a liquid state * * *." He found no reason for supposing that labradorite anorthosite also came to its place in a liquid condition.

Balk (1930, 1931) made a detailed structural study of the Adirondack anorthosite and suggested that it was differentiated by filter-pressing from a magma that resembled diorite. This magma was squeezed upward as a crystal mush and formed a tilted large body and several small outliers. In the crystal mush the solid part was assumed to have been plagioclase crystal and the lubricating liquid to have been syenitic in composition. Because of the subordinate amount of gabbroic rocks exposed in the Adirondack region, Buddington (1939) assumed that the parent magma was gabbroic anorthosite. He expanded his hypothesis by suggesting that calcic labradorite may have differentiated from the basaltic magma in the lower part of the crust and formed an anorthosite shell above the peridotite shell (Buddington, 1936). Partial melting of such an anorthosite stratum would yield gabbroic anorthosite magma of Adirondack type. Earlier, Kolderup (1936, p. 292) had suggested that the anorthosite near Bergen, Norway, is a differentiation product of a magma, the composition of which must have been that of anorthosite-gabbro in the same area. Osborne (1949) defended the possibility of liquid anorthosite magma in his study of the western part of the Morin massive and claimed that the white andesine anorthosite there forms a dike, 5 miles thick, that served as a feeder along which anorthosite magma was brought to a higher level to form the stratiform Morin mass.

Barth (1952, p. 228) claimed that the anorthosite in the Egersund-Sogndal district in southern Norway shows no evidence of squeezing but instead shows features that cannot be explained by assuming a magmatic origin alone. He points out that "* * * these rocks have gone through complicated processes of metasomatic nature," and compares the problem of origin of anorthosite to that of the granite.

A small wedge-shaped occurrence of gneissic anorthosite in Rodil District, South Harris, Outer Hebrides, shows, according to Davidson (1943), features that are similar to those of the anorthosites in Norway. This anorthosite is metamorphosed and contains schlieren of metamorphosed gabbroic rocks, now amphibolites, garnet amphibolites, and eclogites.

Thus, the investigators seem to agree that the anorthosite in the stratiform sheets of Bushveld and Stillwater type and that in smaller sills (for example, Pigeon Point) is a product of crystallization differentiation of basaltic or gabbroic magma. The origin of the anorthosite bodies of the Adirondack type is more problematic and a matter of controversy. The earlier discussions are centered around the composition of the parent magma, mechanism of the differentiation, type of intrusion, and time relation between the differentiation and intrusion. Recently Michot has advanced a different approach; in several papers he has suggested that such great masses of anorthosite could have been formed in the catazone through magmatic and metasomatic processes (Michot, 1955, 1957). Synkinematic basic magmatism and assimilation of pelitic sediments could produce magmatic anorthosites; the anatectic and metasomatic processes, essentially the removal of ferromagnesian minerals from noritic and leuconoritic rocks, could give rise to residual anorthosite and enrichment of ferromagnesian minerals and ilmenitic iron ores.

A common principal feature of all these anorthosites described in the literature is their association with plutonic rocks. The closest approach to an effusive equivalent is a porphyritic dike rock, kenningite, named and described by Eckermann (1938) from the Nordingrå-Rödö region, Sweden. Eckermann assumed that the anorthosite is a product of the differentiation of a doleritic magma at constant silicon value and that a partial redissolving of plagioclase crystals took place in the Rödö region. No true effusive equivalents of anorthosite are found.

The anorthosite in Boehls Butte quadrangle, Idaho, differs strikingly in many respects from the anorthosites described in the literature. The mode of occurrence resembles in some respects that of the Adirondack type of anorthosite, but the masses in Idaho are much smaller. The floors of most bodies are exposed but contain little if any mafic rocks, and the border zones of the anorthosite are not more mafic than their centers. The most striking differences, however, were found in the mineral content, structure, and texture.

AREA STUDIED

The area studied comprises the Boehls Butte quadrangle and an area of about 120 square miles adjoining this quadrangle on the east and on the north. The Boehls Butte quadrangle lies in the southern part of the Idaho panhandle (fig. 1) and is bounded by 115°45′ and 116°00′ west longitudes and by 46°45′ and 47°00′ north latitudes. The northern part of the quadrangle lies along the western part of the Bitterroot Mountains, which in this area rise 5,000 and 6,000 feet above sea level. The North Fork of the Clearwater River and its tributaries have eroded steep canyons between the mountains.

Most of the area is heavily wooded with white pine, yellow pine, cedar, white fir, spruce, and hemlock. In the northernmost part, however, forest fires have destroyed the timber, and the highest mountains are vegetated mainly by grass, low brush, and scattered groups of young trees. In this part of the area the exposures are good and geology can be mapped with greater certainty than in the southern wooded part.

The area can be reached through Elk River or through Clarkia, both about 30 miles to the west. From Elk River a narrow dirt road leads to Boehls where it joins a dirt road leading from the North Fork of the Clearwater River to Goat Mountain. The road from Clarkia traverses a scenic mountain area and joins a road going from Goat Mountain to Avery, a town about 30 miles to

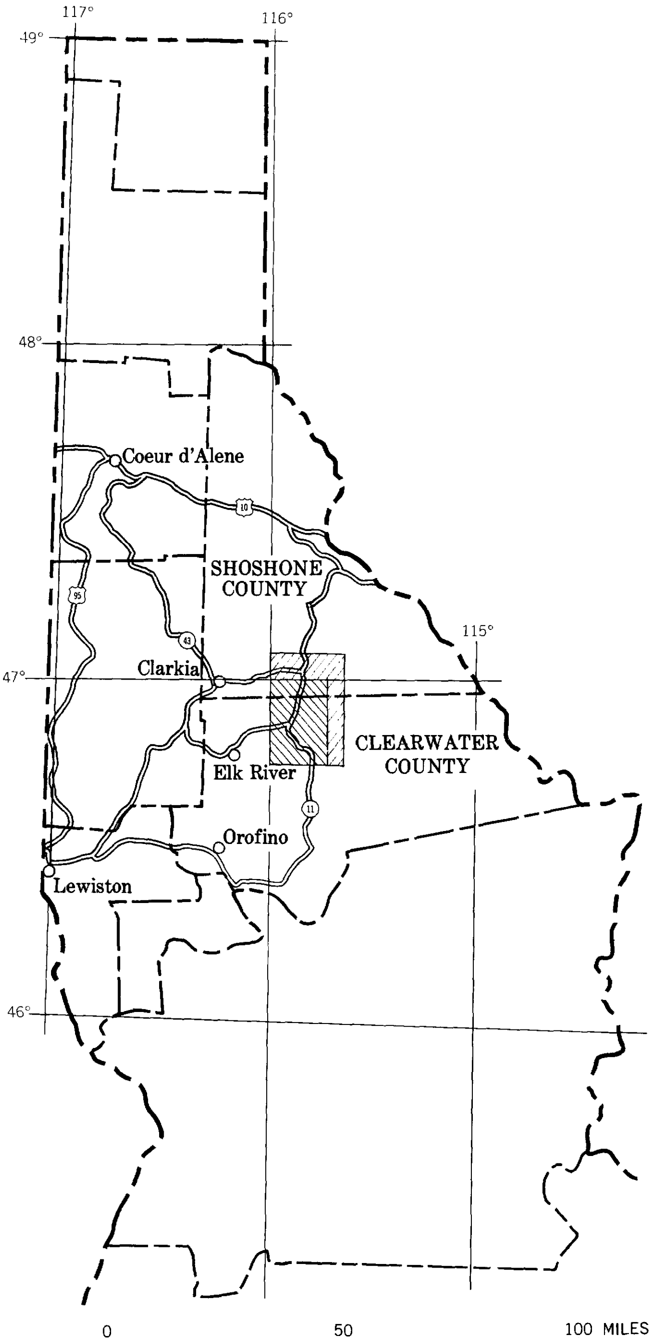

FIGURE 1.—Map of northern Idaho showing location of the Boehls Butte quadrangle and vicinity.

the north. During recent years, logging has been extended to the southern part of the map area. A new logging road that was constructed along the lower part of the Little North Fork of the Clearwater River and along Cedar Creek exposed parts of two large anorthosite bodies which, until then, were accessible only with difficulty.

PREVIOUS WORK

An occurrence of anorthosite in the vicinity of Goat Mountain was indicated on the map and mentioned in the text of Umpleby and Jones (1923, p. 10). On the "Geologic Map of the State of Idaho," compiled by Ross and Forrester (1947), the same occurrence is indicated as a Precambrian intrusive body.

FIELDWORK

Fieldwork for this report was done in the summers of 1951-55. The canyon of the North Fork of the Clearwater River was mapped the first summer. A rubber raft, used as a means of access to the area, was launched 4 miles upstream from the eastern border of the map area (pl. 1), at a point where a road leading to the Canyon Ranger Station crosses the river. Anorthosite is well exposed for 9 miles along the North Fork of the Clearwater River. In the summer of 1952 a trip from Boehls to Goat Mountain revealed a fairly large body of anorthosite near Stocking Meadows. A section through this body was exposed when a logging road along Cedar Creek was opened in the summer of 1953. Most of the mapping of the anorthosite bodies was done during the summers of 1953 and 1955. During the summer of 1953 the author was assisted in the field by Katherine Wagner and Patty Newbold and during 1955 by Alice Vrbsky and Frances Gilbert.

SPECIMEN AND LOCALITY NUMBERS

Localities from which specimens mentioned in the text were taken are shown with specimen numbers on the geologic map (pl. 1). Specimen numbers are also used to identify the chemical analyses of rocks. If two or more analyses are made from the same hand specimen and one or more minerals analyzed in addition to the rock, a letter (a, b, c) is added to the number to identify the mineral.

The following list gives township, range, and section for each locality:

Locality Number	*Township (North)*	*Range (East)*	*Sec.*
599	41	5	25
600	41	5	25
607	41	5	33
608	41	5	33
611	41	5	33
613	41	5	32
630	41	4	26
632	41	4	26
633	41	4	26
669	40	6	19
684	40	6	31
694	40	6	32
947	42	5	30
949	42	5	30
954	42	5	19
956	42	5	19
963	42	5	17
967	42	5	16
968	42	5	16
971	42	5	16
1014	42	5	9
1015	42	5	4
1016	42	5	3
1018	42	5	3
1021	43	5	31
1023	43	4	28
1154	42	5	1
1197	41	4	25
1210	41	4	1
1225	41	5	9

Locality Number	*Township (North)*	*Range (East)*	*Sec.*
1227	41	5	9
1234	41	5	10
1299	41	4	26
1306	41	5	23
1308	41	5	23
1321	42	5	3
1353	43	4	25
1363	43	5	31
1381	41	6	27
1386	42	4	4
1393	42	4	26
1418	42	5	34
1474	40	4	25
1484	39	4	1
1495	41	5	1
1681	42	4	2
1683	42	4	2
1690	42	4	12
1692	42	5	18
1744	42	4	2
1747	43	4	36
1749	43	4	36
1751	42	4	1
1753	42	5	33
1759	42	5	20

CLASSIFICATION OF THE ROCKS

The oldest rocks in the area are metamorphic rocks belonging to the Precambrian Belt series (pl. 1). In the northwestern and northeastern parts of the area, schist and quartzite are continuous with the Prichard formation, the lowest formation of the Belt series of Shoshone County. In the southern part of this county, Umpleby and Jones (1923) show some younger formations in addition to the Prichard formation. These were the Burke and Revett formations combined, and the Wallace formation, listed in order of deposition. St. Regis formation, which in the northern part of Shoshone County occurs between the Revett quartzite and the Wallace formation, was not shown in the southern part. All these formations are exposed in the northeastern corner of the map area (pl. 1), and their equivalents probably occur in the southern part of the Boehls Butte quadrangle where higher grade metamorphism has obliterated most of the sedimentary structures on which the formational subdivisions are based. The shaly layers were metamorphosed to schists that have approximately the same mineral assemblages regardless of the formation in which they occur. Therefore, the petrologically different rock units, rather than the formations, were mapped; specifically, the schist that overlies the thick beds of pure Revett quartzite and is probably equivalent to the St. Regis formation was mapped with the lowest schist unit of the Wallace formation, and the layers that may be equivalent to the Burke formation are included either in the upper part of the Prichard formation or in the lower part of the Revett quartzite.

Among the plutonic rocks two distinctly different groups can be distinguished: older plutonic rocks and young discordant intrusive bodies. The older plutonic rocks are gneissic and occur as small elongate conformable masses in the schist, whereas the younger intrusive bodies are massive or the structures in them are not conformable with the structure of the country rock. The older plutonic rocks range from serpentine and gabbro through quartz diorite to tonalite in composition. The younger intrusive bodies consist of gabbro or quartz monzonite or granite.

The anorthosite could not be included in any of the groups mentioned and is therefore discussed in a separate section. The anorthosite is older than the younger intrusive rocks, but its age relative to the older plutonic rocks remains uncertain. Amphibolite and garnet amphibolite occur as layers and lenslike bodies in the shist and quartzite and are most common in the Prichard formation.

BELT SERIES

CORRELATION

Correlation is mainly based on the work by Pardee (1911) and Umpleby and Jones (1923) in the southern part of the Shoshone County. The northernmost part of plate 1 overlaps their geologic maps, and the formations were first identified there and then traced south. Specifically, the quartzite and schist exposed along the northern boundary in the northwestern part of the area (on Lookout Mountain and to the southwest) were mapped as the Prichard formation by Pardee (1911). The descriptions by Wagner (1949), Ransome and Calkins (1908) were helpful especially in identifying various compositional units in the Wallace formation.

A normal depositional sequence—the Prichard, Burke, Revett, St. Regis, and Wallace formations listed in the order of deposition—is exposed in the northeastern corner of the area on the northeast side of the Little North Fork of the Clearwater River (table 1).

TABLE 1.—*Generalized section through the Belt series northeast of the Little North Fork of the Clearwater River along Spotted Louis Creek*

Formation and rock type	Estimated thickness (*meters*)	(*feet*)
Revett, coarse-grained pure quartzite	470	1, 550
Burke, schist and fine-grained biotite quartzite	200	660
Prichard:		
Schist	420	1, 390
Quartzite	360	1, 180
Schist	130	430
Amphibolite	70	230
Garnet-mica schist	80	260
Granodiorite and granite	200	660
Total	1, 930	6, 360

A coarse-grained thick-bedded fairly pure quartzite in this section is a part of the layers shown as the Revett and Burke formations by Umpleby and Jones (1923). This quartzite is easy to identify in the field (see petrographic description) and was used as a key during the mapping and correlating of the highly metamorphic schists, quartzites, and gneisses of the area studied with the Belt series exposed farther north. On plate 1, it is shown as the Revett quartzite but it may include beds equivalent to parts of the Burke formation. The formations beneath it are considered to be equivalents to the Burke and Prichard formations; those overlying it are equivalents to the St. Regis and Wallace formations.

In the northeastern corner of the map area, the Revett quartzite is underlain by a thin-bedded biotite quartzite and a coarse-grained mica schist in which one or two layers of fairly pure, medium- to coarse-grained granular quartzite are interbedded about in the middle of the section. The stratigraphic position of the thin-bedded biotite quartzite just beneath the massive beds of the pure coarse-grained Revett quartzite corresponds to that of the Burke formation in the Coeur d'Alene district, and the composition and stratigraphic position of the mica schist with one or two quartzitic units are those assigned to the Prichard formation. However, no contact could be placed between the beds of the Burke formation and the underlying beds of the Prichard formation because identical beds occur in each. The Prichard formation consists mainly of schist with less biotite quartzite, whereas the beds just under the Revett quartzite are mainly biotite quartzite with less schist.

The problem of distinguishing the petrologically similar beds of various formations is more complicated in the southern part of the Boehls Butte quadrangle where numerous faults interrupt the normal sequence and bring the Revett quartzite into contact with layers of schist and gneisses which may belong to any of the other four formations—Prichard, Burke, St. Regis, or Wallace.

In many conformable sections, the beds underlying the thick-bedded quartzite consist of coarse-grained garnet-biotite-muscovite schist with some thin beds of biotite quartzite. The contact between the massive quartzite and the schist is sharp, and the lower contact of the Revett quartzite was placed on the bottom of the massive pure quartzite beds. A layer of schist, about 20 m thick, is interbedded with the Revett quartzite in the lower part of this formation. The equivalent of the Burke formation could not be identified with certainty. It is possible that the schist just mentioned and the quartzite beneath it are equivalent to the Burke formation or that an upper part of this formation, which is transitional between the Revett quartzite and the shale of the Prichard formation in the Coeur d'Alene district, is in the Boehls Butte quadrangle included in the massive white quartzite and the lower beds in the schist underneath (schist of the Burke and Prichard formations on pl. 1). It is also possible that the equivalent of the Burke formation was not deposited in the southern part of the area or that it is not exposed because of faulting. Fault contacts are extremely common next to the Revett quartzite and it is impossible to tell how much of the section is missing. A definite correlation cannot be done until after mapping the belt, about 10 miles wide, that separates the low-grade metamorphic Belt series of the Coeur d'Alene district from their high-grade metamorphic equivalents near the Clearwater River.

A similar reasoning is used in mapping the schist of the St. Regis formation with the lower schist unit of the Wallace formation. The faulting is common also along the upper contact of the Revett quartzite, and in such places it is impossible to tell which schist overlies the quartzite.

Umpleby and Jones (1923, p. 8) state that in the southern part of Shoshone County in general the St. Regis formation could not be traced with certainty because of a perfect gradation between the lower beds of this formation and the underlying massive Revett quartzite and also because of a gradation between the upper beds of the St. Regis formation and the lowest shaly beds of the Wallace formation. These shaly beds (the lower part of them probably belongs to the St. Regis, and the upper part to the Wallace formation, as assigned by Umpleby and Jones) would correspond to the schist (Wss) that in the area of plate 1 normally overlies the thick-bedded coarse-grained quartzite that has been mapped as the Revett quartzite.

A heterogeneous sequence of white granular quartzite, thin-bedded biotite and diopside quartzite, diopside-plagioclase gneiss, and sillimanite-garnet schist with thin layers of biotite-plagioclase gneiss is considered to be equivalent to the Wallace formation. The dolomitic sand layers which are typical of this formation in areas farther north, were recrystallized as light-green diopside-plagioclase gneiss. Together with the interbedded biotite gneiss and biotite quartzite, the diopside-plagioclase gneiss forms a major portion of the quartzite-gneiss units (pl. 1) of the Wallace formation. Interbedded with the quartzite-gneiss units are at least two schist units, each several hundred meters thick. The stratigraphic sections correlate fairly well with those given by Wagner (1949) on the south slope of the St. Joe River, about 10 miles north of the northern boundary of plate 1.

The host rocks of the anorthosite bodies are separated by faults from rocks identified as equivalents of the Belt series. Petrologically the schist and the quartzite that form the host rock are similar to the schist and quartzite of the Prichard formation except that calcareous layers are interbedded with the quartzite in the central part. Because a petrologically similar quartzite with some calcareous layers is exposed near Pinchot Butte about 1,500 m below the upper quartzite unit on Widow Mountain, it is thought that the host rocks of the anorthosite belong to a lower part of the Prichard formation not exposed elsewhere in Shoshone County.

PRICHARD FORMATION

Most of the area is covered by the Prichard formation, a portion of which is a direct continuation of this formation from areas farther north. This portion is exposed in the northwestern and northeastern parts of the area, whereas the north-central part is covered by rocks that probably belong to lower units in the Prichard formation. The main rock type in the Prichard formation is a coarse-grained mica schist that contains layers rich either in quartz or in minerals such as biotite, muscovite, garnet, kyanite, andalusite, and sillimanite. One or two thick layers of pure quartzite are interbedded with the schist in the northwestern and northeastern parts of the area, and a quartzite layer with abundant calcareous material in the central part. These quartzitic parts of the formation are referred to as upper and lower quartzite units for distinction from the other quartzitic layers in the same formation. The Prichard formation includes numerous lenticular bodies and layers of amphibolite, a part of which is garnetiferous. The anorthosite bodies occur in the schist just beneath the lower quartzite unit.

STRATIGRAPHIC SEQUENCE

The lithology of the Prichard formation varies from place to place because it includes many lenticular petrologically different rock units. Most of these individual rock units show a similar lack of homogeneity on a small scale; for instance, the main rock type, a coarse-grained mica schist, includes chemically and mineralogically different layers and lens-shaped bodies. The minority rock types that are included in the schist—amphibolite, garnet amphibolite, lime-silicate rocks, phlogopite-plagioclase rocks, and quartzite—are more abundant in the lower part than in the upper part of the formation.

The lowest part of the Prichard formation is exposed along Floodwood Creek where thick layers of coarse-grained mica schist rich in aluminum silicates underlie the anorthosite and the lower quartzite unit. This schist dips gently under the large bodies of anorthosite, and small bodies are included in it. In the northern part of the area, the contacts of anorthosite follow the bedding toward the Orphan Point, where a fault separates the anorthosite and the aluminum-rich schist from the quartzite and garnet-mica schist of the upper part of the Prichard formation in the west. The schist west of this fault contains only locally some aluminum silicates, and it has much less plagioclase than the schist on the east side of the fault. The thickness of the lowest schist unit south of the mouth of Timber Creek is about 600 meters. Conformably on this schist rests anorthosite (500 to 2,000 m thick) and the lower quartzite unit with lime silicate layers (20 to 600 m thick). A thin layer of schist separates the anorthosite from the overlying quartzite unit. On Monumental Buttes and near Goat Mountain, this schist contains an exceptionally large amount of kyanite, andalusite, and sillimanite.

A great thickness of garnet-mica schist with only little if any aluminum silicates and plagioclase overlies the lower quartzite unit. This schist, which petrologically is similar to the schist in the westernmost part of the area, is well exposed in the 2,000 feet deep gorges of Foehl Creek and the Little North Fork of the Clearwater River near Brush Hill, Indian Dip, and Getaway Point. The bedding in the schist is gently dipping and the fold axes plunge to the east. The younger beds occur thus toward the east where north-trending steeply dipping faults separate this schist unit from the normal sequence of the Prichard, Burke, Revett, and Wallace formations.

The quartzite layers interbedded with the schist of the Prichard formation in the northeastern part of the area, such as those exposed between the various northward-trending faults on Stubtoe Peak, near Buzzard Roost, and on the northeast side of the Little North Fork, are petrologically similar to the quartzite on the west side of the anorthosite and its country rocks (on Lookout Mountain). This petrologic similarity supports the structural interpretation and correlation according to which these quartzite layers belong to the same (upper) unit.

The stratigraphic position of the quartzite layers exposed on the northeast side of the steep canyon of the Little North Fork of the Clearwater River leaves no doubt that this part of the formation is equivalent to the uppermost part of the Prichard formation, and that the quartzite there is the upper unit. In this section (table 1), gneissic granodiorite and granite are exposed on the bottom of the canyon along the river; on the top of this plutonic unit is a coarse-grained garnet-mica schist with one or two quartzite layers, together about 360 meters thick. These two layers are mapped together in this vicinity because the layer of schist that separates them is only 1 to 20 meters thick, but attains a thickness of about 60 meters east of the mapped area. The quartzite layers are medium grained, foliated, and contain some micas. The major part of the schist above the quartzite layers is similar to the schist under the quartzite, both being rich in micas. Toward the top of the formation, more fine-grained layers are interbedded with the coarse-grained schist. The stratigraphic position of the last 200 meters just under the coarse-grained Revett quartzite corresponds to that of the Burke formation (table 1), provided that no large fault occurs just under the Revett quartzite.

The schist on either side of the fault zone that borders the anorthosite block in the east is petrologically sim-

TABLE 2.—*Four generalized sections through various parts of the Prichard formation*

1. Section through upper quartzite unit

Rock types on a ridge south of Little Lost Lake	Estimated thickness (meters)	Estimated thickness (feet)
Gabbro	200	660
Coarse-grained gray quartzite	30	100
Anorthosite	30	100
Garnet-muscovite-biotite schist	30	100
Fault.		
Thin-bedded fine-grained quartzite	20	65
Coarse-grained gray quartzite	20	65
Coarse-grained white quartzite	30	100
Fault.		
Gray quartzite with garnets	20	65
Fault.		
Garnet-muscovite-biotite schist	20	65
Coarse-grained gray quartzite with drawn-out garnets	40	130
Coarse-grained white quartzite	30	100
Fault.		
Coarse-grained gray quartzite with drawn-out garnets	40	130
Coarse-grained white quartzite	50	165
Biotite schist	120	400
Amphibolite	5	17
Biotite schist	--------	-------
Total	685	2,262

2. Section through lower quartzite unit

Rock types near Monumental Buttes, lower quartzite unit	Estimated thickness (meters)	Estimated thickness (feet)
Coarse-grained garnet-muscovite-biotite schist.		
Thin-bedded fine-grained gray biotite quartzite	130	430
Coarse-grained white quartzite	30	100
Amphibolite	60	200
Fault.		
Thin-bedded fine-grained gray biotite quartzite	100	330
Coarse-grained white quartzite	35	115
Fault.		
Thin-bedded fine-grained gray quartzite	140	460
Coarse-grained white quartzite	40	130
Amphibolite	20	65
Kyanite-andalusite-garnet schist	60	200
Anorthosite.		
Total	615	2,030

3. Section through lower quartzite unit

Rock types at the mouth of Cedar Creek, lower quartzite unit	Estimated thickness (meters)	Estimated thickness (feet)
Garnet biotite gneiss.		
Actinolite- and graphite-bearing biotite quartzite	50	165
Lime-silicate rock	280	920
Amphibolite	60	200
Lime-silicate rock	160	530
Actinolite-bearing biotite quartzite	10	30
Lime-silicate rock	20	65
Kyanite-andalusite-sillimanite-biotite-muscovite-plagioclase rock	200	660
Anorthosite	50	165
Andesine-bearing mica schist	100	330
Anorthosite	--------	-------
Total	930	3,065

4. Section above anorthosite

Rock types on south slope of Smith Ridge, one mile east of section A–A′, average dip 30° N.	Estimated thickness (meters)	Estimated thickness (feet)
Crest of the syncline, elev. 4,900 ft.		
Garnet-biotite schist	220	730
Garnet amphibolite	130	430
Garnet-biotite schist	280	930
Garnet amphibolite	160	530
Coarse-grained quartzite with some biotite and muscovite	40	130
Garnet-biotite schist	15	50
Amphibolite	2	7
Garnet-biotite schist	20	65
Amphibolite	5	17
Fine-grained dark biotite-muscovite quartzite	55	180
Amphibolite	25	80
Biotite schist	15	50
Amphibolite	50	165
Biotite-muscovite-garnet schist	100	330
Kyanite-andalusite-sillimanite schist	100	330
Tremolite-quartz-plagioclase rock (no. 1306)	60	200
Granular white quartzite	30	100
Tremolite-quartz-plagioclase rock	30	100
Kyanite-andalusite-sillimanite schist	50	170
Actinolite-plagioclase rock	10	33
Diopside-plagioclase rock, pyrrhotite in the lower part	5	17
Kyanite-andalusite-sillimanite schist rich in biotite	10	30
Anorthosite	1,380	4,550
North Fork of the Clearwater River, elevation 1,500 feet	--------	-------
Total	2,792	9,224

ilar and only a minor displacement may have occurred along the faults east of Indian Dip and Getaway Point.

In the northwestern part of the area, the upper quartzite unit is composed of thick layers of white to light-gray medium- to coarse-grained quartzite overlain by fine-grained thin-bedded gray biotite-bearing quartzite. The best section is along the ridge south of Little Lost Lake (table 2, column 1). This section is interrupted by several faults, and the repetition of similar sequences in it is apparently due to the faulting. Allowing for this repetition, the section shows that coarse-grained white quartzite, 30 to 50 m thick, occurs in the lower part of the quartzite unit and is overlain by garnetiferous gray beds that show a strong lineation. The upper part consists of fine-grained thin-bedded gray biotite quartzite and is overlain by a coarse-grained muscovite-biotite-garnet schist similar to that which occurs under the upper quartzite unit.

The thickness of the schist exposed southwest of Widow Mountain under the upper quartzite unit is estimated to be 1,500 meters. Another layer of micaceous light-gray quartzite is exposed under this schist in the deep stream valleys north and west of Pinchot Butte. This quartzite is petrologically similar to the upper layers of the quartzite unit that in the north-central part overlies the anorthosite and may belong at the same stratigraphic horizon.

Near Little Lost Lake, a northward-trending vertical fault cuts off the upper quartzite unit. On the east side of this fault a fine-grained garnet-bearing mica schist forms an interfingering contact with the anorthosite and only a thin layer of quartzite, belonging to the lower unit is exposed between the anorthosite and the gabbro at Long Hike Rock. This quartzite layer becomes thicker toward the east and extends about 25 miles from Long Hike Rock in the northwest to Crescendo Peak in the southeast. The southward extension of this lower quartzite unit is exposed along the ridge south of Crescendo Peak and along the steep canyon of the Little North Fork of the Clearwater River where it trends again westward. An intrusive gabbro interrupts the stratigraphic sequence east of the mouth of Cedar Creek. Provided there is no great fault concealed by the gabbro and amphibolites east of it, the fine-grained biotite quartzite and the thick layers of lime-silicate rocks exposed at the mouth of Cedar Creek belong at the same stratigraphic horizon as the lower quartzite unit. If they do, the lithologic character of the lower quartzite unit of the Prichard formation changes considerably from the northwestern part of the map area toward the southeast.

Near Monumental Buttes a good section of white to gray quartzite is exposed along the road leading to Buzzard Roost. In this section (table 2, column 2) the white quartzite beds are overlain by fine-grained thin-bedded biotite quartzite layers. Two small masses of lime-silicate rock occur between the schist and quartzite on Monumental Buttes. One of them is exposed just on the northwest side of the North Butte, and boulders of the other were found on the trail between North Butte and South Butte. The one at North Butte is about 10 meters thick and is cut by intrusive fine-grained amphibolite. The thickness of the white quartzite near Monumental Buttes is about 35 m but it thins toward the northwest and southeast.

The section through the lower part of the Prichard formation east of Goat Mountain is much the same as that near Monumental Buttes. The lower quartzite unit continues farther to the southeast toward Crescendo Peak. The gray thin-bedded fine-grained layers of the section there are exposed along the road leading from Goat Mountain to Crescendo Peak and the white coarse-grained layers are exposed on the ridge that extends southward from Crescendo Peak. Diopside and actinolite were found in a few thin layers interbedded with thick beds of pure white to cream-colored quartzite between Goat Mountain and Crescendo Peak. Their occurrence indicates that some calcareous material was deposited with quartz sand in this part of the section. At the south end of the ridge that extends southward from Crescendo Peak, white coarse-grained quartzite is overlain by fine-grained gray biotite quartzite which contains several layers of fine-grained plagiolase-phlogopite rock and some coarser layers rich in plagioclase, scapolite, diopside, hornblende, and other lime-silicates. Toward the Little North Fork of the Clearwater River, the layers rich in calcic plagioclase and lime-silicates become more abundant.

A fairly abrupt change in composition occurs between the top of this ridge and the bottom of the canyon of the Little North Fork of the Clearwater River to the south. The section near the river dips 50° south, parallel to the slope of the steep canyon, and therefore only a few layers are exposed; but these layers are definitely richer in calcium than any of the layers exposed between Little Lost Lake and Crescendo Peak. At the bottom of the canyon, 2 miles east of the mouth of Cedar Creek, the main rock type is a fine-grained biotite quartzite. Several layers rich in lime-silicates and a few layers of coarse-grained marble, 1 to 5 cm thick, are interbedded with this quartzite.

Toward the west, at the mouth of Cedar Creek, dark-gray thin-bedded graphite-bearing biotite quartzite is exposed on the south side of the river and lime-silicate rocks on the north side, along the lower drainage of Cedar Creek, overlying there the kyanite-andalusite-

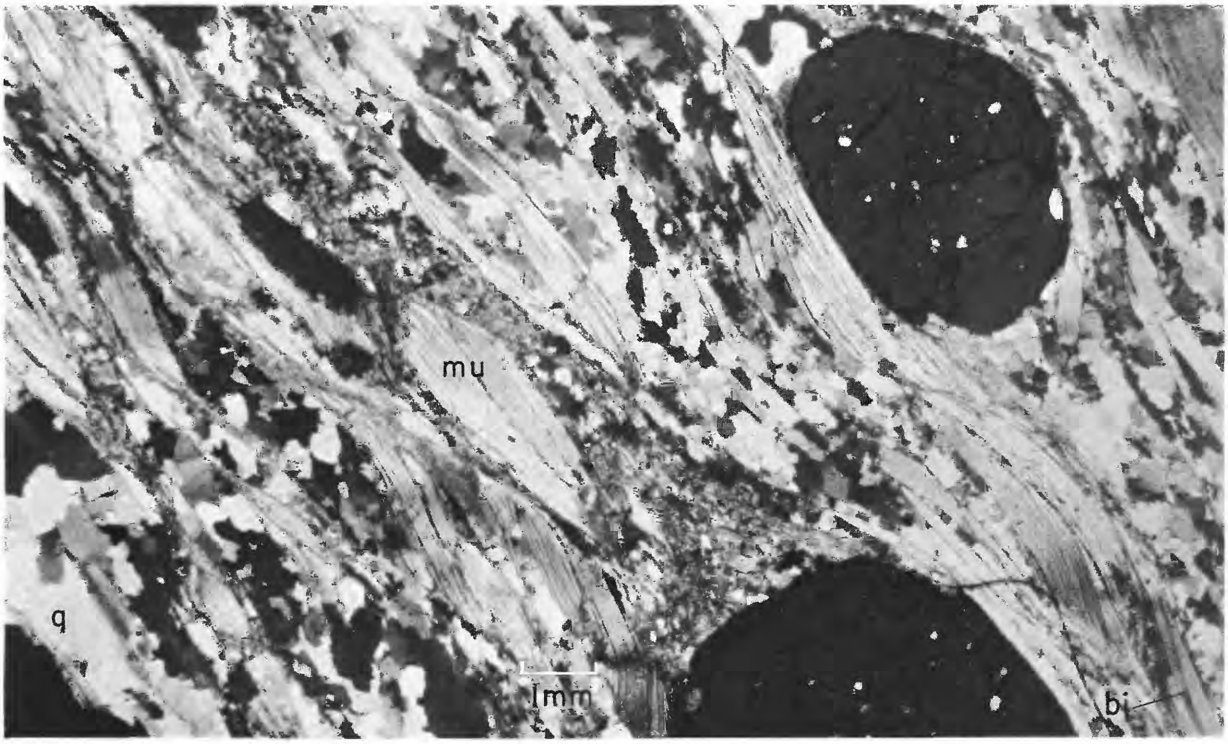

FIGURE 2.—Photomicrograph of a common type of schist in the Prichard formation. Contains numerous round garnet crystals embedded in laminated mica schist. The micaceous laminae contain large flakes of muscovite (mu) and small flakes of muscovite and small flakes of biotite (bi). Tiny quartz grains (q) occur in small lenticles and in thin layers. Loc. 1023, 1 mile west of Orphan Point. Crossed nicols.

sillimanite schist and anorthosite (table 2, column 3). The fine-grained gray beds of biotite quartzite on the south side are stratigraphically above the coarse-grained lime-silicate rocks. Thus, these calcareous layers apparently substitute for the coarse-grained white quartzite, which elsewhere is beneath the fine-grained gray beds of this quartzite unit (compare columns 2 and 3 in table 2). Farther to the west, between Cedar Creek and Stocking Meadow Lookout, the lime-silicate rocks are interbedded with layers of coarse-grained fairly pure quartzite and fine-grained biotite-rich quartzite which are similar to the corresponding quartzite layers between Big Talk Lake and Crescendo Peak and probably belong at the same stratigraphic horizon. Near Stocking Meadow Lookout diopside-plagioclase-phlogopite rock, tremolite quartzite, and a thick layer of pure quartzite are exposed along the road leading from Boehls to Goat Mountain and along the road leading to the lookout.

The fine-grained dark graphite-bearing quartzite extends from the mouth of Cedar Creek westward and forms steep outcrops along the Little North Fork of the Clearwater River. Three miles east of Boehls a layer of biotite gneiss with abundant large anthophyllite prisms and huge garnet crystals, a kyanite-andalusite-sillimanite-cordierite gneiss, and anorthosite are exposed under this quartzite.

Synclinal structure on Smith Ridge suggests that this rock series is the same as is exposed along the south slope of Smith Ridge (table 2, column 4). However, only two rather thin layers of quartzite, the upper one rich in biotite, are exposed above the kyanite-andalusite-sillimanite schist on the south slope of Smith Ridge. The uppermost part of the section there consists of coarse-grained garnetiferous biotite-muscovite schist interbedded with amphibolite and dark fine-grained biotite quartzite or biotite schist. The lowest 180 meters of the section consist of kyanite-andalusite-sillimanite schist interbedded with quartzitic layers rich in tremolite, actinolite, or diopside and containing a considerable amount of calcic plagioclase. This sequence overlies an anorthosite which locally shows a distinct layering. A similar aluminous schist with tremolite- or actinolite-bearing layers is exposed north and south of the anorthosite at Stocking Meadows where also it overlies a layered anorthosite.

PETROGRAPHY

The schist between the lower and upper quartzite units and that above the upper unit are much alike, consisting of a common garnet-mica schist, whereas the schist under the lower quartzite unit is exceptionally rich in aluminum silicates.

GARNET MICA SCHIST

The most common variety of schist in the middle and upper parts of the Prichard formation is coarse grained and contains abundant flakes of muscovite and biotite

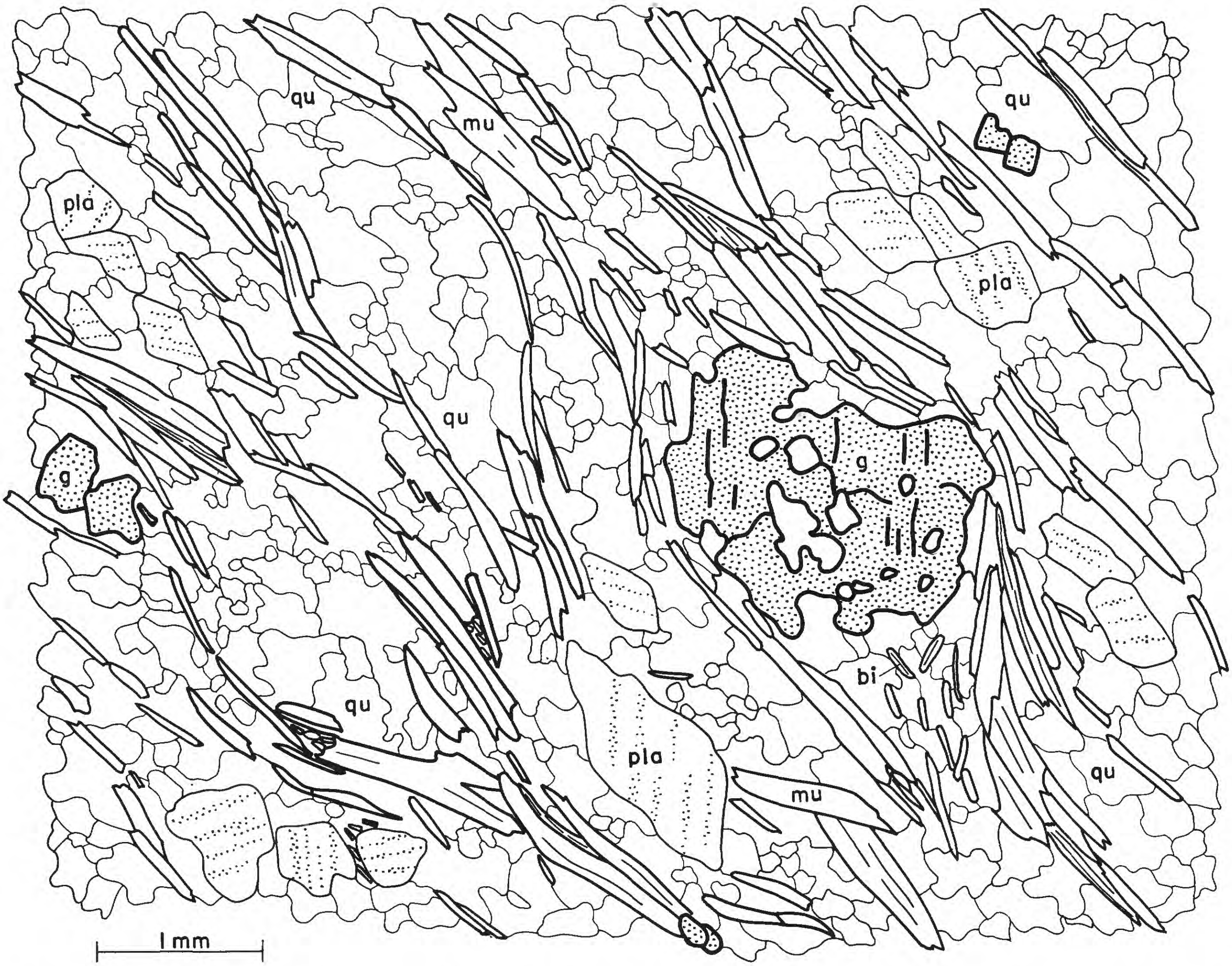

FIGURE 3.—A camera lucida drawing from a common type of garnet-mica schist in the Prichard formation shows subhedral to anhedral garnet (g), muscovite (mu), biotite (bi), quartz (qu), and plagioclase (pla). Anhedral quartz grains with irregular borders occur in small lens-shaped aggregates and in thin layers. Laminae of mica flakes separate these layers and curve around the aggregates of quartz and grains of garnet. Plagioclase (pla) occurs sparingly with quartz. One mile west of Orphan Point (loc. 1023)

2 to 3 mm in diameter, which lend a glittering bronze luster to the cleavage planes of weathered rock. The fresh rock is medium gray and hard. Most of the schist contains garnet crystals in varying sizes and amounts. In the upper part of the formation garnet crystals abound but are rather small, ranging from 2 to 5 mm in diameter. In the middle part of the formation, most garnet crystals are about 5 mm in diameter, but locally they measure as much as 15 cm in diameter; for example, near Twin Springs Creek, east of Freezeout Lookout, and along the Little North Fork of the Clearwater River about a mile southwest of the mouth of Cedar Creek.

Microscopic study shows that in the middle schist unit exposed in the extreme northwestern corner of the area, the muscovite is far more abundant than the biotite, whereas toward the south, near the southern border of Shoshone County, the amount of biotite usually exceeds that of muscovite. In the northwestern part of the area, the common mica schist is strongly sheared (fig. 2). Small lens-shaped mosaic aggregates and discontinuous layers that are 0.1 to 1 mm thick and that consist of small strained grains of quartz and larger grains of plagioclase are enveloped by large sheared flakes of muscovite that converge at either end of the aggregate. The muscovite flakes are bent, showing slight strain shadows under crossed nicols. Biotite occurs as flakes of medium size either interleaved with muscovite or as small flakes oriented at random along the shear zones and on small areas at either end of the large muscovite flakes. Muscovite flakes also envelope the garnet crystals which are slightly elongated parallel to the schistosity, and include small round grains of quartz and magnetite (fig. 3). Many garnet crystals show well-developed dodecahedral faces. Small flakes of biotite and grains of quartz fill the wedge-shaped areas at either end of elongated garnet crystals. These flakes of biotite butt against the garnet but muscovite flakes bend around the garnet and are thus earlier. The ends of some elongated garnet crystals are altered to chlorite. The chlorite flakes are subparallel to the grain boundary of the garnet, and thus about at right angles to the small flakes of biotite. Locally, abudant magnetite occurs as thin films along the cleavage in the muscovite and also as irregularly shaped grains between the other minerals. Ilmenite scales abound in some layers and tourmaline in others. Zircon occurs in subhedral grains that have rounded corners and that measure 0.1 to 0.4 mm in diameter.

As a rule, the layers poorer in mica are finer grained. In these layers the mica flakes of medium size are more evenly distributed throughout the rock and the quartz shows less tendency to occur in lens-shaped segregations. Locally, thin quartzitic layers are interbedded with the schist. A considerable amount of biotite occurs in these layers and there is every gradation from the quartzite to the schist. The chemical composition of a common type of mica schist calculated from a modal analysis is shown in table 5, locality 1023. This analysis shows it to be a metamorphosed equivalent of a common argillite.

The schist of the middle part of the Prichard formation that is exposed above the anorthosite in the central part of the area, on Smith Ridge, and along the Little North Fork of the Clearwater River, is much coarser grained and contains less muscovite than its equivalent in the northwestern part of the area. This change in texture and mineral content indicates an increase in the degree of metamorphism toward the south. In the schist on Smith Ridge thin layers of fine-grained biotite-plagioclase schist are interbedded with coarse-grained biotite-muscovite-plagioclase schist. Abundant ilmenite and zircon occur as accessory minerals in all layers. Plagioclase (An_{38}) forms about 30 percent of the rock and occurs as large- to medium-sized grains embedded in granulated quartz and micaceous minerals. In the mica-rich beds, micas occur in thin laminae between layers, 1 to 3 mm thick, that consist of plagioclase (An_{38}) and quartz. A similar thin lamination was found in some beds of the Prichard formation along Sawtooth Creek under the upper quartzite unit just east of the northeast corner of the mapped area. There the lamination seems to be a primary feature due to deposition, whereas on Smith Ridge its more irregular magnitude and distribution suggest that the depositional feature was accentuated during the metamorphism by segregation of micaceous minerals and quartzo-feldspathic material into separate thin layers. This suggestion is strengthened by the fact that at a lower altitude, along the Little North Fork of the Clearwater River, a similar segregation took a coarser form leading in places to formation of veined gneisses, in which the quartzo-feldspathic veins range from 2 mm to 10 cm in thickness.

In the gneiss exposed in a roadcut 1 mile west of the mouth of Cedar Creek, huge garnet crystals and long anthophyllite prisms were found in some intensely recrystallized layers. In one layer, the garnets are as much as 15 cm in diameter and most are enveloped by biotite and anthophyllite. At the same locality, in the lowest part of the roadcut, garnet-biotite-anthophyllite (gedrite) rock is exposed (fig. 4). In this rock the garnet crystals range from 0.2 to 3 cm in diameter and the anthophyllite prisms are 0.5 to 1 cm long. Some plagioclase and quartz are between the dark constituents (fig. 5); and rutile, magnetite, and apatite are the common accessories. The garnet and the amphibole were separated from the coarse-grained gneissic layer and studied in detail.

The amphibole prisms are 2 to 5 cm long and dark green in hand specimen. Under the microscope they show a parallel extinction and a weak pleochroism γ=light bluish green, β=light yellowish green, α=colorless. The indices of refraction measured in immersion liquids are $\alpha=1.652\pm0.001$, $\beta=1.659\pm0.001$, $\gamma=1.671\pm0.001$, $\gamma-\alpha=0.019$ (when the right liquid was found,

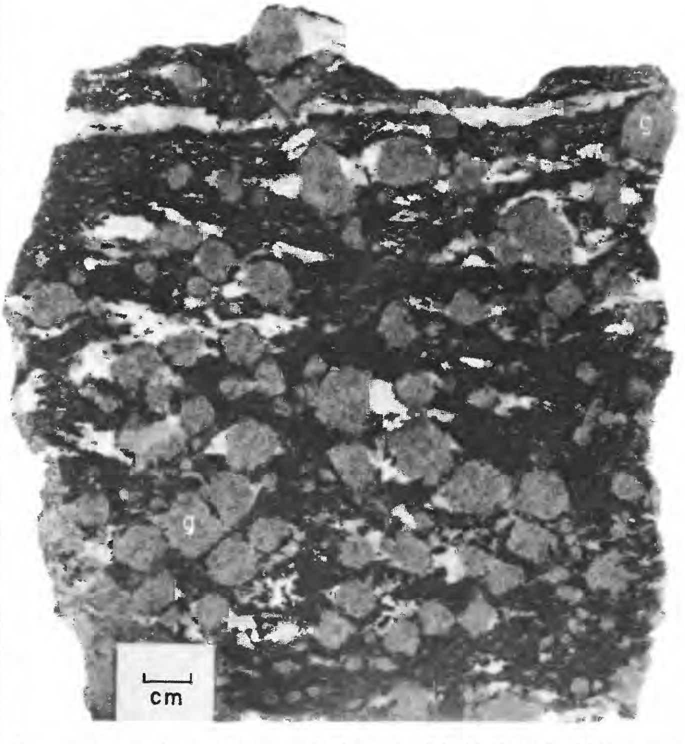

FIGURE 4.—A specimen of a coarse-grained garnet-gedrite-biotite rock with numerous round garnet crystals (g). The black parts are rich in gedrite and biotite; white is plagioclase and quartz. One mile west of the mouth of Cedar Creek (loc. 1227).

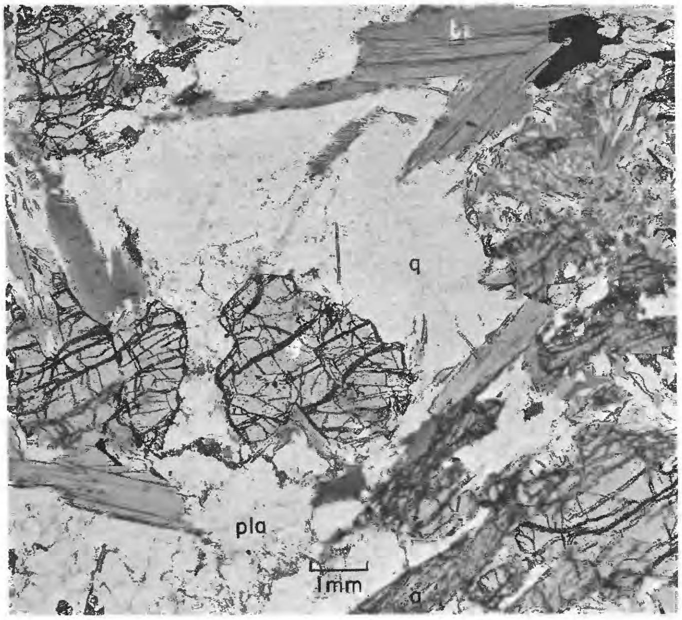

FIGURE 5.—Photomicrograph of a thin section made of the specimen shown in fig. 4. The constituent minerals are large garnets (g), euhedral to subhedral aluminous anthophyllite (a), quartz (q), plagioclase (pla), and biotite (bi). Plane-polarized light. (Loc. 1227.)

it was immediately measured by Abbé refractometer in order to avoid the influence of change of temperature). This mineral was separated by Richard P. Marquiss, U.S. Geological Survey, and a fraction with the specific gravity of 3.17±0.02 was analyzed chemically (table 3). The analyzed material contained a few tiny grains of quartz and rutile.

TABLE 3.—*Chemical analysis of an aluminous anthophyllite (gedrite) from garnet-biotite gneiss of the Prichard formation, 1 mile west of the mouth of Cedar Creek on Little North Fork of Clearwater River (loc. 1227)*

[Analyst, L. N. Tarrant, U.S. Geol. Survey]

Constituents	Weight percent	Molecular equivalent	Number of atoms		
SiO_2	46. 87	7804	Si	6. 74	8. 00
Al_2O_3	12. 11	1188	Al	1. 26	
Fe_2O_3	1. 04	65	Al	. 79	. 90
FeO	18. 20	2533	Fe^{+3}	. 11	
MnO	. 18	25	Fe^{+2}	2. 19	5. 91
MgO	17. 14	4251	Mn	. 02	
CaO	. 50	89	Mg	3. 67	
Na_2O	1. 08	174	P	. 00	
K_2O	. 10	11	Ti	. 03	
TiO_2	. 30	38	Ca	. 08	. 40
P_2O_5	. 01	1	Na	. 30	
F	. 10	53	K	. 02	
H_2O+	2. 19	1216	F	. 05	2. 15
H_2O-	. 12	--------	H	2. 10	
	99. 94	--------	O	24. 00	
Less O for F	. 04	--------			
Total	99. 90	--------			
Sp gr=3.17±0.02					

Calculation of the formula from the chemical analysis (table 3) gave the following result:

$$(H_{2.10}F_{0.05})_{\Sigma=2.15}(K_{0.02}Na_{0.30}Ca_{0.08})_{\Sigma=0.4}(Mg_{3.67}Fe^{+2}_{2.19}Mn^{+2}_{0.02}\\ Ti^{+4}_{0.03})_{\Sigma=5.91}(Al_{0.79}Fe^{+3}_{0.11})_{\Sigma=0.9}(Al_{1.26}Si_{6.74})_{\Sigma=8.0}O_{24}.$$

A considerable amount of aluminum is present and substitutes for silicon in the tetrahedral groups and for magnesium in the octahedral positions. The optical properties and the chemical composition are those typical of the aluminous anthophyllite (gedrite) with magnesium: iron=5:3 (Winchell and Winchell, 1951, p. 427).

The large garnet crystals are subhedral and contain numerous inclusions of quartz, magnetite, and biotite. The index of refraction of the garnet is $n=1.784\pm0.003$ and specific gravity is 3.80±0.02. The X-ray study made by F. Hildebrand of the U.S. Geological Survey shows that two phases of garnet are present, one having a cell dimension $a_0=11.576\pm0.006$A and the other $a_0=11.557\pm0.006$A, resembling in this respect the garnet in garnet amphibolite along Orofino Creek, about 20 miles south of Boehls (cf. Hietanen, 1962, table 13). The garnet was analyzed chemically (table 4, no. 1225) and the ratios of various compounds were calculated in molecular percents. The analyzed material contains about 1 percent impurities, mainly

TABLE 4.—*Chemical analyses and optical properties of garnets*

[Chemical analyses by Lucile N. Tarrant, U.S. Geol. Survey. Mineral separation by Richard P. Marquiss, U.S. Geol. Survey]

Locality No.	1225.		1353.	
Location	Little North Fork of Clearwater River.		Two miles east of Orphan Point.	
Rock type	Garnet-anthophyllite rock.		Anorthosite.	
Constituent	Weight percent	Molecular equivalent	Weight percent	Molecular equivalent
SiO_2	38. 30	6377	38. 34	6384
Al_2O_3	21. 33	2092	21. 40	2099
Fe_2O_3	. 32	20	. 17	11
FeO	27. 79	3868	28. 33	3943
MnO	. 84	118	. 48	68
MgO	4. 59	1138	3. 37	836
CaO	6. 36	1134	7. 65	1364
Na_2O	. 04	7	. 03	5
K_2O	. 02	2	. 03	3
TiO_2	. 21	26	. 11	14
P_2O_5	. 28	20	. 01	1
H_2O+	. 05	28	. 10	55
H_2O-	. 00	-----------	. 00	-----------
Total	100. 13	-----------	100. 02	-----------
	Composition		Composition	
Pyrope	18. 15		13. 45	
Almandite	61. 73		63. 36	
Spessartite	1. 87		1. 11	
Grossularite	17. 29		21. 55	
Andradite	. 96		. 53	
a_0	11. 576±0. 006 A			
a_0'	11. 557±0. 006 A		11. 611±0. 006 A	
n	1. 784±0. 003		1. 784±0. 003	
Sp gr	3. 80 ±0. 02		3. 93 ±0. 02	

quartz, anthophyllite, rutile, and magnetite. Comparison with the analyses of garnets published earlier shows that this garnet is similar to garnet found in eclogites (Eskola, 1921) and in anorthosite (Kemp, 1921).

Some layers in the schist contain a considerable amount of plagioclase (An_{15-30}) which occurs in small grains with quartz, its compounds being in all probability sedimentary in origin. The schist west of Orphan Point contains in places abundant plagioclase which, however, differs in its mode of occurrence from the plagioclase crystallized from sedimentary material. The plagioclase-bearing layers west of Orphan Point also contain some hornblende and are in places transformed to quartz dioritic or tonalitic gneiss. The composition and mode of occurrence of this gneiss resemble those of the gneissic quartz diorite and tonalite that occur south of the area under investigation and that were found to be metasomatic in origin (Hietanen, 1962).

SCHIST RICH IN ALUMINUM SILICATES

Most of the schist under the lower quartzite unit contains aluminum silicates and plagioclase (An_{25-40}) in variable amounts. A chemical analysis (table 5, No. 1690) shows that this schist is considerably richer in aluminum, calcium, and sodium and poorer in silicon and potassium than the normal garnet mica schist (no. 1023) in the upper part of the formation. Sillimanite is the most common aluminum silicate near Floodwood Creek and around the mouth of Timber Creek (fig. 6), but abundant kyanite and andalusite occur with the sillimanite near the anorthosite bodies (pl. 1). In many places schist in a zone 200 to 500 m wide next to the anorthosite bodies contains 20 to 40 percent kyanite and andalusite, and 1 to 5 percent sillimanite. Schist that is exceptionally rich in aluminum silicates is generally very coarse grained, the kyanite and andalusite crystals averaging 2 to 5 cm in length and the mica flakes 3 to 8 mm in diameter. In some occurrences, however, the grain size is somewhat smaller; for example, in the staurolite-bearing schist near Timber Creek and locally in the kyanite-andalusite-sillimanite schist on the north side of the North Fork of the Clearwater River east of Larson Cabin, the crystals of aluminum silicates are only 5 to 10 mm long and the individual mica flakes 1 to 4 mm in diameter. The largest kyanite crystals are on the southeast slope of Goat Mountain, where they range from 2 or 3 cm to 30 cm in length. The mineralogy of the kyanite-andalusite-sillimanite schist has been discussed in an earlier paper (Hietanen, 1956).

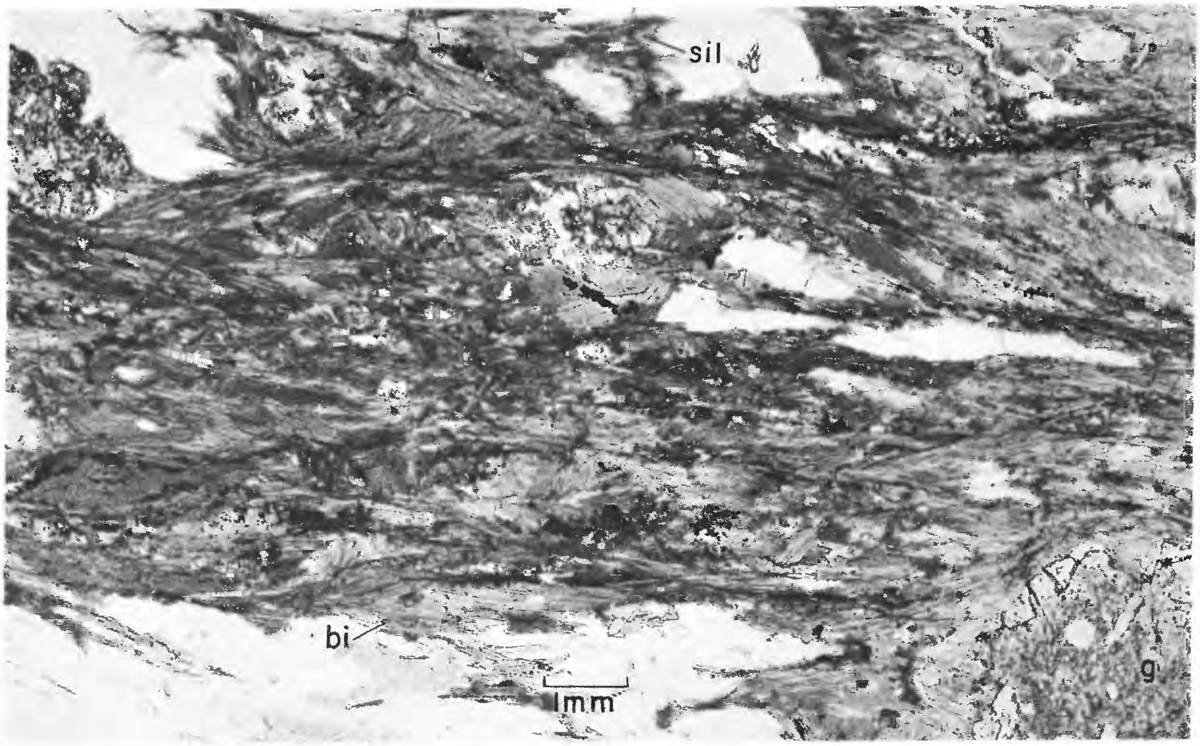

FIGURE 6.—Photomicrograph of a sillimanite-garnet-biotite schist southeast of Orphan Point (Loc. 1751 on pl. 1). A part of the biotite (bi) has been replaced by sillimanite (sil). Light-colored constituents are quartz and plagioclase. g, garnet. Plane-polarized light.

Staurolite is found only in the northwestern part of the area; toward the south its place is taken by aluminum silicates and garnet. The lowest part of the Prichard formation exposed around Floodwood Creek, Pinchot Creek, and Timber Creek generally contains some staurolite in addition to abundant sillimanite. The staurolite crystals are small, ranging mostly from 1 to 5 mm in length. Many small grains are included in biotite or in garnet, apparently forming armored relicts.

TABLE 5.—*Chemical analyses in weight and ionic percentage, molecular norms, and modes of rocks of the Prichard formation*

Locality and specimen No.	1023 [1]	1690 [2]	1234 [3]	1306 [3]	1308 [3]	1495 [3].
Rock type	Garnet-mica schist.	Sillimanite-garnet schist.	Lime-silicate rock.	Tremolite-plagioclase rock.	Diopside-plagioclase rock.	Plagioclase-phlogopite rock.
Location	One mile west of Orphan Point.	Mouth of Timber Creek.	Mouth of Cedar Creek.	South slope of Smith Ridge.	South slope of Smith Ridge.	Little North Fork Clearwater River.
Weight percent						
SiO_2	71.24	60.77	50.94	64.12	49.81	53.80
Al_2O_3	14.23	21.83	8.32	9.81	13.56	24.66
Fe_2O_3	.39	.89	.35	.24	.31	.45
FeO	6.13	6.39	2.76	2.36	4.06	.72
MnO	.14	.10	.07	.07	.12	.02
MgO	1.88	1.25	8.34	10.67	8.10	3.89
CaO	.51	1.88	21.22	9.78	21.88	6.61
Na_2O	.54	1.06	.76	.77	.59	4.69
K_2O	3.07	2.99	.57	.19	.06	2.75
TiO_2	.10	.78	.29	.35	.58	.24
P_2O_5		.06	.08	.09	.26	.05
CO_2		.01	3.65	.02	.07	.01
F						.11
S			.90			
C			1.67			.80
H_2O+	1.66	1.43	.35	1.22	.68	.98
H_2O-	.11	.16	.05	.12	.07	.17
Subtotal			100.32			99.95
Less O			.45			.05
Total	100.00	99.60	99.87	99.81	100.15	99.90
Cation percent						
SiO_2	69.87	59.02	48.06	60.33	46.56	49.12
$AlO_{3/2}$	16.45	24.98	9.25	10.88	14.93	26.53
$FeO_{3/2}$	.29	.65	.25	.17	.21	.31
FeO	5.02	5.19	2.18	1.77	3.17	.55
MnO	.12	.08	.06	.06	.10	.02
MgO	2.76	1.81	11.72	14.95	11.28	5.29
CaO	.53	1.95	21.45	9.86	21.90	6.47
$NaO_{1/2}$	1.03	1.99	1.39	1.40	1.07	8.30
$KO_{1/2}$	3.85	3.70	.68	.23	.07	3.20
TiO_2	.08	.57	.20	.25	.41	.16
$PO_{5/2}$		.05	.06	.07	.21	.04
CO_2		.01	4.70	.03	.09	.01
C			(7.88)			(3.65)
H_2O	(5.76)	(4.63)	(1.10)	(3.83)	(2.12)	(2.98)
	100.00	100.00	100.00	100.00	100.00	100.00
O	170.12	165.01	155.67	161.59	152.25	154.05
OH	11.52	9.26	2.20	7.66	4.24	5.96
F						.32
S			1.59			
Total anions	181.64	174.27	159.46	169.25	156.49	160.33
Molecular norm						
Q	46.49	32.04	8.78	24.65	0.74	
Or	19.25	18.50	3.40	1.15	.35	16.00
Ab	5.15	9.95	6.95	7.00	5.35	39.15
An	2.65	9.30	17.95	23.12	34.47	31.95
Ne						1.41
C	10.51	15.57				2.25
Wo			26.12	10.18	29.14	
En	5.52	3.62	23.44	29.90	22.56	
Fs	9.84	8.76	2.22	3.00	5.52	
Fo						7.94
Fa						.39
Ap		.13	.16	.19	.56	.11
Il	.16	1.14	.40	.50	.82	.32
Mt	.43	.97	.38	.25	.31	.46
Pr			2.39			
Cc		.02	9.40	.06	.18	.02
Total	100.00	100.00	101.59	100.00	100.00	100.00

See footnotes at end of table.

TABLE 5.—*Chemical analyses in weight and ionic percentage, molecular norms, and modes of rocks of the Prichard formation*—Continued

Locality and specimen No.	1023 [1]	1690 [2]	1234 [3]	1306 [3]	1308 [3]	1495 [3].
Rock type	Garnet-mica schist.	Sillimanite-garnet schist.	Lime-silicate rock.	Tremolite-plagioclase rock.	Diopside-plagioclase rock.	Plagioclase-phlogopite rock.
Location	One mile west of Orphan Point.	Mouth of Timber Creek.	Mouth of Cedar Creek.	South slope of Smith Ridge.	South slope of Smith Ridge.	Little North Fork Clearwater River.
Molecular mode calculated from cation percent [4] (measured)						
Quartz	(51.0)	29.56	8.78	23.90	0.39	
Plagioclase		An_{46} 18.50	An_{72} 24.90	An_{79} 21.20	An_{86} 40.17	An_{44} 73.5
Orthoclase		10.80	3.40	.45		7.7
Muscovite	(26.0)					
Biotite	(9.4)	14.89				22.6
Chlorite	(1.2)					
Garnet	(12.1)	8.56				
Diopside			51.78		57.56	
Tremolite				57.26		
Sillimanite		19.07				
Apatite		.13	.16	.19	.56	
Sphene				.54	.72	
Ilmenite		1.00	.40		.11	.3
Magnetite	(0.3)	.45	.38	.57	.31	.6
Pyrite			2.39			
Calcite			9.40	.06	.18	
Graphite			7.88			
Total	100.0	102.96	109.47	104.17	100.00	104.7

[1] Weight percent calculated from measured mode.

[2] Analyst, Faye H. Neuerburg, U.S. Geol. Survey.

[3] Analysts, M. Balazs and Lucile N. Tarrant, U.S. Geol. Survey.

[4] Molecular mode is received directly from cation percentage by simple regrouping of the cations and anions to form minerals found in the thin sections. The optical properties and chemical analyses of the minerals are used as a guide to their formula. The excess over 100 in totals shows the number of anions (OH, F, S, C) used in calculation.

Sillimanite is the most common aluminum silicate in this lowest part of the Prichard formation. It occurs as fine needles and as brownish nodules which show irregular parting and contain small brown patches of biotite. The nodules are commonly embedded in biotite, and abundant fine needles of sillimanite transect the biotite flakes. It is apparent that sillimanite crystallized from the biotite. The first sign of this replacement is a much lighter color of the biotite that contains sillimanite needles. Thin sections show every gradation from the biotite schist without sillimanite to sillimanite schist with remnants of biotite. Apparently the recrystallized sillimanite used the aluminum and silicon contained in biotite, but the other elements of biotite—potassium, magnesium and most of the iron—migrated to the country rocks to form more biotite there. Muscovite occurs with biotite, but its amount is less than in garnet-mica schist of the upper part of the formation. In some layers of sillimanite schist along Floodwood Creek it is scarce or absent.

Cordierite abounds in a few layers on Smith Ridge and along the Little North Fork of the Clearwater River, but is scarce in the northern part of the area. In the southern part some layers may contain more than 37 percent cordierite (Hietanen, 1956, table 3), raising the percentage of magnesium abnormally high for an aluminous sedimentary rock.

On the south slope of Smith Ridge the layer of kyanite-andalusite-sillimanite schist that is exposed just above the anorthosite is considerably richer in biotite than the other schist layers. Since biotite is the only ferromagnesian mineral present and it is a magnesium-rich variety, this layer is rich in magnesium. Biotite rather than cordierite crystallized in it because sufficient potassium was present. An exceptionally large amount of biotite in the schist near its contact with the anorthosite also occurs in several localities near Goat Mountain and along the Little North Fork of the Clearwater River. Locally the biotite was segregated into discontinuous layers and lens-shaped bodies that range from 5 to 50 cm in thickness. Biotite in such segregations occurs in flakes that range from 5 to 15 mm in diameter. Small flakes of greenish soft muscovite and some white grains of andesine usually are embedded between the biotite flakes.

Garnet in round to subhedral crystals, which are 2 to 10 mm in diameter, is common in the schist rich in aluminum silicates. In most layers the amount of this mineral ranges from 2 percent to about 20 percent. Some outcrops near the anorthosite bodies contain many large garnet crystals; for example, anhedral crystals that include abundant kyanite and quartz and measure as much as 12 cm in diameter constitute about 15 percent of the schist and plagioclase-rich contact rock south of Monumental Buttes. On the south slope of Goat Mountain sporadic crystals that range from 3 to 5 cm in diameter are sparsely distributed throughout a coarse-grained plagioclase-biotite-kyanite rock. These large garnet crystals are 24 to 30 cm apart and the matrix next to them is free from biotite, indicating that iron and magnesium from the surrounding matrix joined the garnet.

QUARTZITE UNITS

The lower part of the upper quartzite unit of the Prichard formation consists of thick beds of white to light-gray coarse- to medium-grained quartzite. Many of the white beds ranging from 10 to 30 cm in thickness are almost pure quartz and have only a few muscovite and graphite flakes along their bedding planes. The texture seems crudely granular through a hand lens, but an examination under the microscope shows that quartz grains are recrystallized and have sutured borders, and that small interstitial grains occur between the larger ones. In most beds the grains are elongate parallel to the lineation. Layers that contain more mica and some garnet are interbedded with these beds of pure quartz. In many localities, for example, on Widow Mountain, large red garnet crystals or groups of crystals abound and appear spectacular against the light-gray quartzite, because on the bedding planes they are drawn out to long spindles parallel to the lineation. Numerous large garnet crystals were also found in coarse-grained white to gray muscovite-bearing quartzite southwest of Buzzard Roost. The garnet crystals there are strongly flattened but appear almost equidimensional along the bedding planes. The quartz grains show a similar deformation. Most of the muscovite is concentrated into thin laminae, but some scattered grains that transect the grain boundaries occur in the beds of pure quartz.

The white quartzite on Stubtoe Peak is similar to that near Buzzard Roost except that no garnet was found in the beds exposed and that the quartz grains and muscovite flakes show a strong elongation parallel to the lineation. The quartzite on the northeast side of the Little North Fork is fairly coarse grained, has a granular texture as most of the quartzite in the Prichard formation, and contains a considerable amount of muscovite.

The dark-gray fine-grained biotite quartzite that overlies the white and light-gray quartzite on Widow Mountain is thin-bedded and contains biotite-rich laminae and layers of garnet-mica schist.

Many layers of the lower quartzite unit resemble the rock types of the upper unit. There are, however, enough differences in the appearance, sequence, and thickness of the individual layers to make distinction between the two units possible. In the northern part of the area, coarse-grained white granular quartzite of the lower unit is overlain by a much thicker layer of thin-bedded gray micaceous quartzite than that found in the upper unit, and the coarse-grained garnetiferous gray beds are missing. Toward the south, lime-silicate rocks substitute for the coarse-grained white quartzite of the lower unit in increasing quantitites.

Most beds of the white quartzite that forms the bottom of the lower quartzite unit in the vicinity of Monumental Buttes and Goat Mountain have a light-brownish hue and contain less muscovite than the white quartzite of the upper unit. Some biotite occurs in this quartzite near Monumental Buttes and its amount increases toward Goat Mountain and Crescendo Peak. The individual beds of fairly pure quartz range from 10 to 50 cm in thickness and are separated by micaceous layers 2 to 10 mm thick. The grains are 0.1 to 10 mm in diameter, have sutured borders, and strong strain shadows are seen under the microscope.

The thin-bedded dark-gray quartzite that overlies the white and coarse-grained light-gray quartzite layers of the lower unit is fine grained and contains a considerable amount of biotite. The thickness of the individual beds ranges from a few millimeters to several centimeters. Microscopic study shows that quartz in this fine-grained gray quartzite occurs in small round or polygonal grains, with biotite filling the interstices. Sphene, small grains of magnetite, graphite, and zircon are common accessories. Some beds south of Big Talk Lake contain scapolite, actinolite, and diopside. Scapolite occurs in small round grains with quartz, whereas actinolite and diopside form large holoblasts that include numerous round quartz grains. Under the microscope the actinolite is yellowish, shows no pleochroism, and has $\alpha=1.615\pm0.001$ and $\gamma=1.638\pm0.001$. Also the orthoclase forms large grains which contain many small round quartz inclusions.

The lime-silicate-bearing layers that were found about half a mile west and south of Crescendo Peak are rather thin (1–5 cm thick) and are interbedded with coarse-grained pure quartzite. A thin layer containing tremolite and diopside was also found in the lower quartzite unit 1½ miles south of Orphan Point. A layer of light-gray fine-grained quartzite with tremolite, zoisite, muscovite, and graphite was found interbedded in biotite-plagioclase schist on Buzzard Roost. The colorless tremolite prisms of this layer are oriented at random. Abundant graphite in small flakes and some plagioclase occur as additional constituents. The scarcity of lime-silicate minerals and scapolite in the northern part of the area suggests that very little calcareous material was deposited there with the quartz-rich sediment.

The quartzite at Stocking Meadow Lookout is fine to medium grained, light bluish gray, and contains some small biotite flakes. A thick layer of quartzite exposed along Twin Springs Creek west of Pinchot Butte and a layer south of Orphan Point are coarse to medium grained, light gray, well foliated, and contain some muscovite and biotite. Quartzite beds at higher altitudes on Pinchot Butte are dark-gray, fine-grained, thin-bedded, and rich in biotite. They are petrographically similar to the fine-grained gray beds above

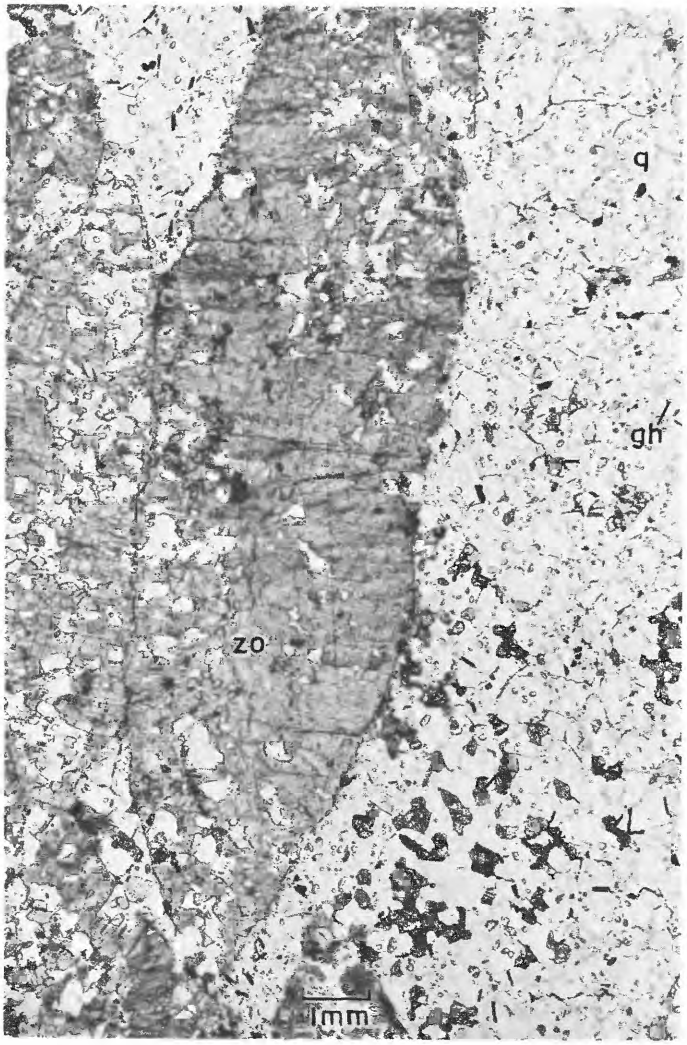

FIGURE 7.—Photomicrograph of large zoisite (zo) prisms in fine-grained graphite-bearing quartzite. Includes numerous small grains of quartz. gh, graphite; q, quartz. Location 1386, about 1¾ miles north-northwest of Pinchot Butte. Plane polarized light.

the white and gray quartzite near Monumental Buttes.

A layer exposed on the southwest side of Twin Springs Creek about 1½ miles southwest of Pinchot Butte consists of fine-grained gray quartzite overlain by a layer of coarse-grained white quartzite, 6 m thick. Both layers are schistose and contain biotite and muscovite. In the upper layer biotite flakes, 1 to 3 mm long, are evenly distributed throughout the rock. A layer of garnetiferous mica schist separates the two quartzite layers, and a thin layer of amphibolite is included in the lower quartzite layer. Quartzite exposed on Stony Creek is white to very light gray and contains some muscovite and scattered flakes of biotite.

Layers of biotite-bearing quartzite, only a few meters in thickness, were found at several localities. Some thin beds in these layers contain abundant diopside or actinolite; for example, a layer of quartzite exposed on the northern border of Clearwater County along the road leading from Boehls to Freezeout Lookout, another layer along Floodwood Creek, and still another exposed on the the ridge north of Pinchot Butte (pl. 1). A thin layer of quartzite occurs between anorthosite and amphibolite about a mile east of Orphan Point. This quartzite is fine to medium grained and well foliated and contains diopside and scapolite as additional constituents. Magnetite, apatite, and sphene occur as accessory minerals.

A zoisite-bearing quartzite with an exceptional texture was found on the ridge north of Pinchot Butte and another layer near Getaway Point. Both layers occur in the schist overlying the lower quartzite unit. Zoisite forms long, light-gray prisms, which are oriented at random in a fine-grained dark graphite-bearing quartzite (fig. 7). Study under the microscope shows that the zoisite prisms contain numerous round quartz inclusions, and that some diopside, hornblende, and garnet, occur as additional constituents in the layer north of Pinchot Butte and garnet in that near Getaway Point.

Small blocks of another rare type of rock were found east of the mouth of Salmon Creek on the north side of North Fork of the Clearwater River (loc. 1381). This rock is greenish blue-gray, fine grained, and weathers readily to a sandy material. About 75 percent of this rock is tourmaline ($\epsilon=1.620\pm0.005$ and $\omega=1.640 \pm0.005$) and about 25 percent quartz. A few tiny muscovite flakes are the only additional constituents.

LIME-SILICATE ROCKS

The rocks rich in lime-silicates are bluish green or green and coarse to medium grained. Minerals in most layers are oriented at random, but in some they show a subparallel orientation that lends a gneissic appearance to the rock. A thickness of about 470 m of lime-silicate rocks is well exposed along the lower drainage of Cedar Creek. Coarse-grained layers there are distinctly bedded, the dark-bluish or greenish beds alternating with the light-colored ones. The thickness of the individual beds ranges from 2 to 50 cm.

Microscopic study shows that the darker layers are rich in diopside or in actinolite, whereas the light-colored layers contain more plagioclase. Layers with a bluish tint are generally rich in diopside and those with a greenish tint rich in actinolite. Some layers consist of about 70 percent diopside with plagioclase, quartz, scapolite, and calcite in varying amounts. Some other layers are rich in quartz and contain plagioclase, diopside, and tremolite in lesser quantities. Tremolite in some of these quartzose layers forms large holoblasts that include numerous round small quartz and plagioclase grains. In some other layers quartz occurs as medium-sized rounded grains with the other minerals—

calcic plagioclase (An_{85}), diopside, and actinolite—filling the interstices. However, this texture is rare. Most layers are fully granoblastic, the dark minerals show a tendency to form larger crystals, and the plagioclase is more sodic.

Diopside in this common type of lime-silicate rock has the indices of refraction $\alpha=1.670\pm0.001$, $\beta=1.676\pm0.001$, $\gamma=1.697\pm0.001$ being thus rich in magnesium. Many sections parallel to the long dimension of the prisms show polysynthetic twinning. The actinolite has $\alpha'=1.634\pm0.001$ and $\gamma'=1.653\pm0.001$ which suggest that it contains about 24 percent iron end-member. The indices of refraction of scapolite, $\epsilon=1.551\pm0.001$, $\omega=1.581\pm0.001$, indicate it to be $Ma_{40}Me_{60}$ (Winchell and Winchell, 1951, p. 353). Several scapolite grains include rods of quartz, their texture resembling that of myrmekite. Plagioclase is oligoclase An_{25-30}. A few grains of uniaxial positive åkermannite with low interference colors were found in some sections.

A chemical analysis of a typical layer in the lime-silicate rock is shown in table 5, no. 1234. This layer is rich in calcium, contains a considerable amount of aluminum and magnesium, and thus has a composition of a shaly limestone. The aluminum is contained mainly in plagioclase which constitutes about 25 percent of the rock.

A lenticular mass rich in calcite was found interbedded with the lime-silicate rocks about a mile north of the mouth of Cedar Creek. Farther north several very hard light-gray layers are interbedded with actinolite rocks. These layers are rich in plagioclase (An_{80-94}) and quartz and contain diopside, grossularite, prehnite, clinozoisite, tremolite, graphite, pyrrhotite, magnetite, sphene, rutile, and hematite. Diopside has $\alpha=1.676\pm0.001$, $\beta=1.680\pm0.001$, $\gamma=1.702\pm0.001$, which indicate that it contains about 15 percent hedenbergite. Yellowish prehnite with $\alpha=1.611\pm0.001$, $\beta=1.621\pm0.001$, $\gamma=1.638\pm0.001$, $+2V\sim75°$ was found as cavity filling in several specimens. Many tabular crystals of prehnite are covered by fine needles of tremolite with $\gamma=1.633\pm0.001$ and $Z\wedge c=12°$. The index of refraction of grossularite varies around $n=1.74$. Scapolite and phlogopite are common in calcite-bearing layers. Wollastonite was found only in one thin-section studied.

Toward the east the sequence that is exposed along Cedar Creek is interrupted by an intrusive gabbro, but the lime-silicate rocks exposed east of the gabbro have the same stratigraphic position. However, these rocks are finer grained and are interbedded with dark-gray fine-grained biotite quartzite. A few layers, 3 to 5 cm thick, consisting of pure coarse-grained calcite are also interbedded. Many samples collected along the Little North Fork of the Clearwater River, 2 miles east of the mouth of Cedar Creek, contain abundant diopside in round or oval holoblasts which are 5 to 12 mm in diameter. The indices of refraction of the diopside $\alpha=1.667\pm0.001$, $\beta=1.674\pm0.001$ and $\gamma=1.696\pm0.001$ indicate that it is an almost pure magnesium member. The main part of the rock is fine-grained biotite-tremolite-quartz rock. Microscopic study shows that the biotite is very light brown and has $2V=0$ and the indices of refraction $\beta=1.590\pm0.001$ and $\gamma=1.591\pm0.001$, being thus a member of the phlogopite-eastonite series.

One of the samples collected from a plagioclase-phlogopite layer along the Little North Fork of the Clearwater River 2 miles east of the mouth of Cedar Creek contains oval aggregates of plagioclase surrounded by thin shells of orthoclase (fig. 8). On the polished surface many of these aggregates show an asymmetric concentric structure resembling that of brachiopods. Microscopic study shows that the rock is completely recrystallized to form a fine-grained granoblastic graphite-bearing plagioclase-phlogopite

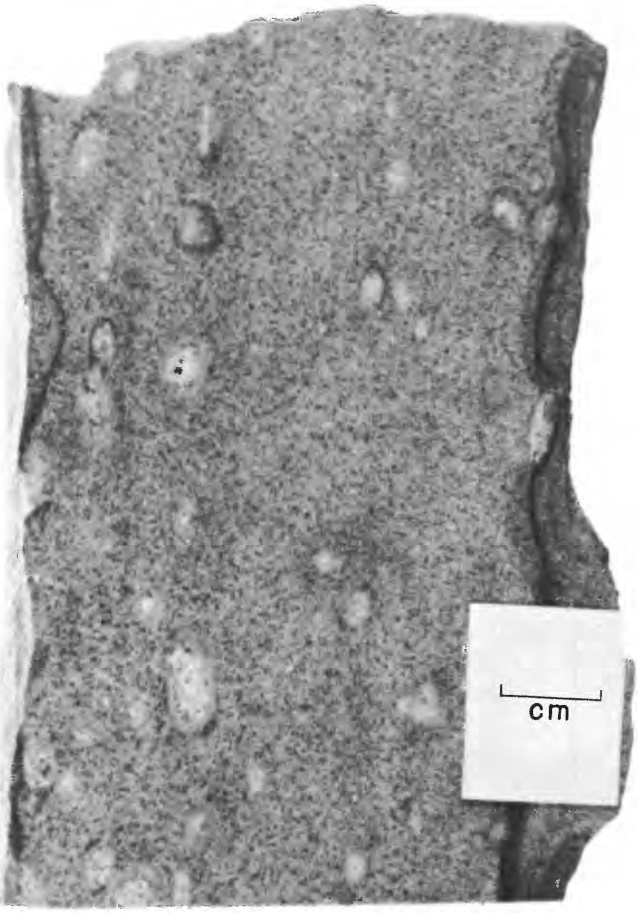

FIGURE 8.—A specimen of a fine-grained phlogopite-andesine rock with light-colored oval areas shelled by orthoclase. Loc. 1495 along the Little North Fork of the Clearwater River about 2 miles east of the mouth of Cedar Creek.

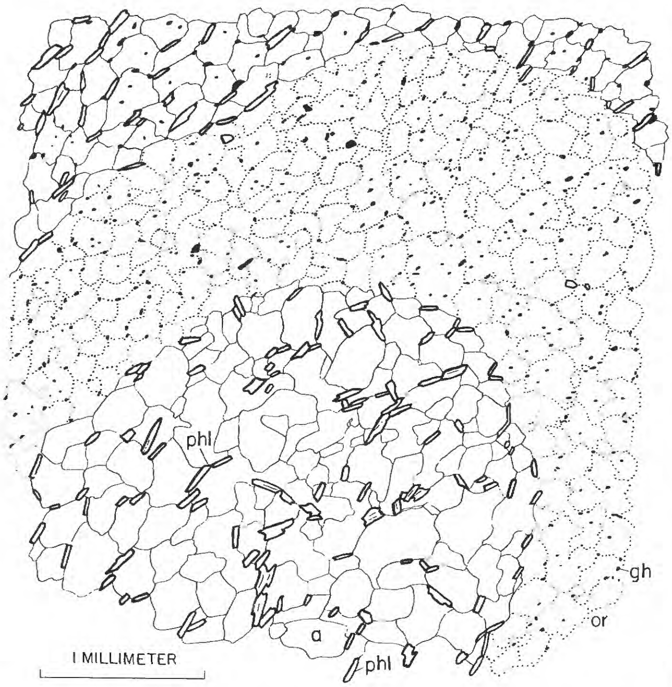

FIGURE 9.—A camera lucida drawing of an end of a nodule shelled by orthoclase from the phlogopite-andesine rock along the Little North Fork of the Clearwater River 2 miles east of the mouth of Cedar Creek (loc. 1495). The center of the nodule (lower center of the figure) is free of graphite that abounds elsewhere in the rock (small black spots). The orthoclase shell (or) is about 1 mm thick along the long sides of the oval nodules and about 1½ mm at the ends. a, andesine; phl, phlogopite; gh, graphite.

rock (fig. 9). The aggregates consist of the same minerals as the main part of the rock except that there is no graphite in them. The shells consist of small grains of orthoclase with $\alpha=1.519\pm0.001$, $\beta=1.523\pm0.001$, $\gamma=1.525\pm0.001$, and tiny flakes of graphite. There is no microstructure that would prove whether these shelled aggregates are remnants of fossils or products of some mechanism such as rolling during the sedimentation. The fact that the main part of the rock contains only a little orthoclase makes the orthoclase shells seem more problematic. Chemical analysis (table 5, no. 1495) shows that this layer is rich in aluminum and contains considerably more calcium and sodium and less silicon than the normal quartzite; magnesium dominates greatly over the iron.

Plagioclase (An_{48}) in this layer crystallized from sedimentary material. It occurs in small granoblastic grains and there are no signs of metasomatic action in any of the fine-grained quartzite layers with which this layer is interbedded. The fine-grained dark biotite quartzite layers are similar to the gray biotite-rich layers in the upper part of the lower quartzite unit of the Prichard formation elsewhere and they are continuous with the quartzite layers on Crescendo Peak. The thin amphibole-bearing layers and the marble layers that are interbedded with the biotite quartzite in this locality prove that some calcareous material was deposited with the shaly sand.

Layers rich in amphibole (hornblende, or tremolite, or actinolite) are interbedded with schist and quartzite on south slope of Smith Ridge and along the Little North Fork of the Clearwater River. From the Little North Fork this sequence extends westward to the area north of Boehls and to the vicinity of Stocking Meadow Lookout. On the south slope of Smith Ridge the tremolite- and actinolite-bearing rocks are interbedded with kyanite-andalusite-sillimanite schist and include only a few rather thin biotite quartzite layers. The hornblende-bearing layers are interbedded with biotite quartzite and schist; they contain about 50 percent hornblende and 50 percent quartz with only very little plagioclase. In fresh hand specimen the hornblende schist appears much like the normal amphibolite, which is rich in plagioclase, but on the weathered surface the plagioclase of the amphibolite is easily distinguished from the quartz of the hornblende schist because of its white weathering. The gradational and fine-grained rocks, however, can be identified only under the microscope. The hornblende-quartz schist probably represents metamorphosed dolomitic and shaly sand layers. No tuffaceous layers have been found in the less metamorphosed equivalents of the Belt series farther north.

Mineral content of the tremolite- and actinolite-bearing layers varies from place to place. Most are rich in actinolite and calcic plagioclase (An_{90}) and contain quartz in varying amounts. Pyrrhotite and graphite are common minor constituents. The iron-to-magnesium ratio in the amphibole varies considerably. Some layers contain light-green tremolite, whereas the others are rich in dark-green actinolite. The light-green tremolite was separated from specimen 1306; a pure fraction with specific gravity of 3.033–3.060 was analyzed chemically. The result (table 6) shows that it contains only about 8 percent iron end member. Calculation to the base of O=24 gives the following formula:

$$(H_{1.66}F_{0.30})_{\Sigma=1.96}\,(Ca_{1.94}\,Na_{0.13}\,K_{0.04})_{\Sigma=2.11}\,(Mg_{4.22}\,Fe^{+2}_{0.36}\,Mn^{+2}_{0.01}\,Ti^{+4}_{0.02}\,Fe^{+3}_{0.03}\,Al_{0.37})_{\Sigma=5.01}\,(Si_{7.51}\,Al_{0.49})_{\Sigma=8.0}\,O_{24}.$$

The optical properties of this tremolite are as follows: $\alpha=0.617\pm0.001$, $\beta=1.625\pm0.001$, $\gamma=1.637\pm0.001$, $\gamma-\alpha=0.020\pm0.001$, and $\gamma\wedge c=20°$. The chemical analysis of the host rock (table 5, no. 1306) shows a larger iron-to-magnesium ratio because of the presence of magnetite.

In addition to the amphibole-bearing layers there are a few lens-shaped bodies and discontinuous layers of light-green rocks in which diopside is a main constituent.

TABLE 6.—*Chemical analysis of tremolite from tremolite-plagioclase rock no. 1306, south slope of Smith Ridge*

[Analyst, M. Balazs, U.S. Geol. Survey]

Constituents	Weight percent	Molecular equivalent	Number of atoms			Number of atoms after correction of $H_2O=1.8$ percent	
SiO_2	54.15	9019	Si	7.63	8.00	7.51	8.00
Al_2O_3	5.27	518	Al	.37		.49	
Fe_2O_3	.32	20	Al	.51	.54	.37	.40
FeO	3.13	436	Fe^{+3}	.03		.03	
MnO	.09	13	Fe^{+2}	.37	4.69	.36	4.61
MgO	20.42	5065	Mn	.01		.01	
CaO	13.10	2336	Mg	4.29		4.22	
Na_2O	.50	81	Ti	.02		.02	
K_2O	.22	23	Ca	1.98	2.16	1.94	2.11
TiO_2	.22	28	Na	.14		.13	
F	.68	358	K	.04		.04	
H_2O+	.94	522	F	.32	1.20	.30	1.96
H_2O-	.03	--------	H	.88		1.66	
			O	24.00			
Subtotal	99.07	--------					
Less O for F	.29	--------					
Total	98.78	--------					

In some lacalities these layers are interbedded with the tremolite-bearing layers, as on the south slope of Smith Ridge; in other localities they are found in the schist next to the lower quartzite unit, as on Monumental Buttes. A chemical analysis of a diopside-rich layer on the south slope of Smith Ridge (table 5, no. 1308) shows that this layer contains considerably more calcium and aluminum than the tremolite-bearing layer (no. 1306). The difference in aluminum content is due to a larger amount of plagioclase (An_{80}) in the diopside-bearing layer and is not essential. Another difference appears in the iron-to-magnesium ratio. The only dark constituent in the diopside-bearing layer is diopside, the composition of which can be calculated from the rock analysis (table 5, no. 1308). The result $CaMg_{0.78}Fe_{0.22}Si_2O_6$ suggests that the iron-to-magnesium ratio is about 1:4, whereas in the tremolite of the neighboring layer it is 1:12. The optical properties of the diopside—$\alpha=1.675\pm0.001$, $\beta=1.681\pm0.001$, $\gamma=1.703\pm0.001$, $Z\wedge c=38°$—are in good accordance with the calculated chemical analysis. The small lenticles of lime-silicate rock on Monumental Buttes consist mainly of similar diopside. Very little plagioclase, sphene, magnetite, and locally some quartz occur as additional constituents.

RAVALLI GROUP

LITHOLOGY AND STRATIGRAPHIC SEQUENCE

For convenience of discussion, the rocks that are equivalent to the Burke, Revett, and St. Regis formations, are described here under the heading Ravalli group, the name applied to this part of the Belt series in areas where subdivisions are difficult (Ransome and Calkins, 1908, p. 26–28). A fairly complete conformable section through all three formations of the Ravalli group is exposed northeast of the Little North Fork of the Clearwater River (table 1). Thick beds of coarse-grained fairly-pure quartzite form the major part of this section, which is interrupted by several faults. The beds beneath the quartzite, which are those stratigraphically equivalent to the Burke formation, consist of medium-grained schist and thin-bedded fine-grained biotite quartzite. Similar beds occur in the Prichard formation, and without the stratigraphic guide, the massive Revett quartzite, it would be impossible to tell to which formation these beds belong. The beds that overlie the Revett quartzite conformably, which are those equivalent to the St. Regis formation, consist of garnetiferous muscovite-biotite schist and fine-grained biotite gneiss. A white granular quartzite, which forms the lowest beds of the Wallace formation, marks the upper contact of the St. Regis formation in the conformable sections of the northeastern part of the area.

Parts of all three formations of the Ravalli group—Burke, Revett, and St. Regis—are probably exposed in the southern part of the area, but only the Revett quartzite can be identified with certainty. The schist overlying the Revett quartzite is thought to be equivalent to the St. Regis formation, and the schist and impure quartzite between the coarse-grained garnet-mica schist of the Prichard formation and the first thick beds of the massive Revett quartzite equivalent to the Burke formation. However, the Burke and St. Regis formations can be identified only where found with the Revett quartzite. Without this stratigraphic guide their separation from the schist of the Prichard formation or from the schist of the Wallace formation is highly uncertain, if not impossible. The numerous faults have interrupted the stratigraphic sequence wherever the massive Revett quartzite occurs, and no good section of this part of the Belt series exists. The best section in the southern part of the Boehls Butte quadrangle is along the Little North Fork near its mouth and continues along the North Fork of the Clearwater River south of the mouth of the Little North Fork (table 7).

In this section (table 7) layers of fairly homogeneous schist and thin-bedded hornblende-bearing diopside-plagioclase gneiss underlie the Revett quartzite and probably represent the Burke and Wallace formations, the latter in a fault contact. The layers overlying the Revett quartzite consists of medium-grained homogeneous biotite schist with some thin-bedded biotite quartzite layers. Along Robinson Creek intensive faulting has obscured the sequence. However, a conformable contact between the underlying Burke formation and the Revett quartzite is exposed in some localities; for example, along a tributary of the North Fork of Robinson Creek northwest of logging Camp X. The beds that are stratigraphically equivalent to the Burke formation here consists of a heterogeneous schist with some hornblende gneiss layers.

TABLE 7.—*A generalized section along the North Fork and the Little North Fork of the Clearwater River south of the southernmost anorthosite mass*

Formation	*Rock type*	*Estimated thickness* (*meters*)	(*feet*)
	Anorthosite.		
	Fault.		
Wallace	Banded hornblende-biotite gneiss with diopside in some layers.	300	980
St. Regis	Biotite-plagioclase schist, homogeneous.	560	1,840
	Biotite-plagioclase gneiss	150	490
	Fault.		
Revett	Thick-bedded coarse-grained quartzite.	100	330
	Fault.		
St. Regis(?)	Garnet-mica schist, kyanite-bearing.	560	1,840
	Fault.		
	Thin-bedded gray quartzite	150	490
Revett	Thick-bedded coarse-grained pure quartzite with some sillimanite in the middle part.	140	460
	Sillimanite-mica schist	20	70
	Thick-bedded coarse grained pure quartzite.	140	460
Burke(?)	Garnet-sillimanite schist	50	160
Total		2,170	7,120

Elsewhere in the southern part of the Boehls Butte quadrangle, the beds that overlie the Revett quartzite and underlie the thin-bedded quartzite or schist that are equivalent to the Wallace formation have a varied lithologic character. In most localities the beds resting on the Revett quartzite consist of fairly homogeneous biotite schist and probably are equivalent to the St. Regis formation. It is not clear whether all diopside-plagioclase quartzite and gneiss which are interbedded with this schist belong to the St. Regis or to the lower part of the Wallace formation. North of the mouth of Robinson Creek a layer of thin-bedded biotite-plagioclase quartzite with diopside-bearing layers occurs just above the Revett quartzite and this thin-bedded quartzite is overlain by mica schist. In other localities along Robinson Creek the equivalent of St. Regis formation consists of mica schist with some sillimanite and garnet.

A white granular quartzite with some diopside is exposed in the southwest corner of the area on the ridge between the North Fork of the Clearwater River and Boathouse Creek and also along the ridge on the east side of the river opposite the mouth of Gold Creek. This quartzite is similar to the quartzite of the lowest part of the Wallace formation on Surveyors Ridge, and the schist underlying it may therefore be equivalent to the St. Regis formation and the schist overlying it equivalent to the Wallace formation.

Another locality where the Revett quartzite and probably rocks equivalent to the St. Regis formation are exposed is around Bertha Hill along the south central edge of the map. The Revett quartzite there is bordered by hornblende-diopside-plagioclase gneiss interbedded with coarse-grained mica schist, many layers of which contain sillimanite and garnet. Diopside-plagioclase gneiss is exposed just north of the Revett quartzite near Thunder Creek, but toward the east a biotite-muscovite schist overlies the quartzite. This schist is similar to the schist north of the mouth of Robinson Creek, but because the Wallace formation also contains layers of similar schist, it is impossible to tell whether the schist north of Bertha Hill is equivalent to the St. Regis or to the Wallace formation. The formation overlying the Revett quartzite east of Bertha Hill is composed of mica schist and of thin-bedded biotite- or diopside-bearing quartzite.

PETROGRAPHY

BURKE FORMATION

Layers that lie conformably just under the Revett quartzite and that are considered equivalent to the Burke formation are exposed in several localities. A layer of biotite schist which separates the Revett quartzite from the underlying hornblende-bearing diopside-plagioclase gneiss south of the mouth of Robinson Creek is coarse grained and contains impure quartzite layers. A similar schist is exposed under the Revett quartzite in many other localities near Robinson Creek. The major constituents in this schist are quartz and biotite; the muscovite and plagioclase occur in small quantities. Garnet- and sillimanite-bearing layers are not as common as in the schist overlying the Revett quartzite. The schist just under the Revett quartzite shown in the northeast corner of plate 1 is similar to the schist described above except that it contains more muscovite. Thin beds of fine-grained biotite-rich quartzite are interbedded with this muscovite-biotite schist lower in the section. The thin-bedded layers grade to a coarse-grained muscovite-biotite-garnet schist (most likely the top of the Prichard formation) which contains beds of fine- to medium-grained muscovite-biotite schist and quartzite.

REVETT QUARTZITE

Most of the coarse-grained massive quartzite whose stratigraphic position is equivalent to that of the Revett quartzite in the Coeur d'Alene district is white or bluish, but it also contains some reddish beds. The individual beds of massive pure quartzite range from 3 to 100 cm in thickness and are separated by paper-thin laminae of biotite and muscovite. In blocks of the massive Revett quartzite, glittering, fairly smooth fracture surfaces of individual large grains make this formation easy to identify. A few grains of plagioclase, rutile, and magnetite occur as minor constituents. Abundant sillimanite was found in reddish beds exposed just north of the mouth of Robinson Creek and on Bertha Hill. Some beds of biotite-sillimanite schist and medium-grained biotite-plagioclase quartzite are interbedded with the coarse-grained quartzite.

ST. REGIS FORMATION

The rocks that overlie conformably the Revett quartzite and are considered to be equivalent, to the St. Regis formation consist mainly of schist and thin-bedded quartzite. In the thin-bedded quartzite north of the mouth of Robinson Creek, gray biotite quartzite layers alternate with light-colored plagioclase-bearing layers. Some light-greenish layers containing diopside also are interbedded. The thickness of the individual beds ranges mostly from 2 mm to 3 cm. Paper-thin laminae of biotite are common between the individual layers. This quartzite grades over to a coarse-grained muscovite-biotite-garnet-sillimanite schist, which is very similar to the schist of the Wallace formation.

Biotite is the main micaceous mineral in the schist farther north, but a considerable amount of muscovite occurs in some layers. Plagioclase and quartz are the light-colored constituents; garnet, magnetite, and zircon appear in small quantities. The thick layer of homogeneous schist that overlies the Revett quartzite just north of the mouth of the Little North Fork of the Clearwater River is medium-grained and contains more plagioclase than the other layers. Biotite is the main micaceous mineral and muscovite occurs in small quantities. A few garnet crystals and magnetite grains are commonly present.

In the canyon of Breakfast Creek the schist above the Revett quartzite is rich in quartz and contains in addition to biotite a considerable amount of muscovite. The muscovite occurs as large flakes, many of which are bent. Elongated quartz grains are strained and have sutured borders or tiny grains occur between the larger ones. Biotite, intermingled with muscovite, forms thin laminae between layers (0.5 to 2 mm thick) that consist of quartz with some oligoclase.

WALLACE FORMATION

DISTRIBUTION AND LITHOLOGY

Sillimanite-garnet schist and interbedded diopside-plagioclase gneiss exposed in the southern part of the area are considered to be equivalent to the Wallace formation. The contact between the St. Regis formation and the overlying Wallace formation was not mapped because it could not be placed accurately, and some of the schist and the gneiss considered to be St. Regis may be equivalent to the Wallace formation. The lowest part of the Wallace formation contains thick beds of medium-grained biotite quartzite, thin-bedded biotite-plagioclase quartzite, white granular quartzite with some diopside, diopside-plagioclase gneiss with local development of hornblende, and garnet-biotite schist, all interbedded. This part of the formation is exposed in the southwestern part of the area along Elkberry Creek west of the logging Camp T, along the load leading from Bertha Hill to the river about a mile east of the junction to Armstrong Lookout, along the ridge between Grandad Creek and the head of Little Meadow Creek, along the ridge between the river and Boathouse Creek, and on the ridge between Little Silver Creek and Elkberry Creek. A layer of mica schist overlies the lowest unit and is in turn overlain by a thick unit of light-green diopside-plagioclase gneiss which locally contains abundant hornblende. Thus, the lower part of the Wallace formation is heterogeneous and contains layers that are petrographically very similar to the layers in the St. Regis formation and in the upper part of the Wallace formation. Near Benton Creek and in the area south of Township Butte two fairly thick layers of diopside-plagioclase gneiss are interbedded with the schist (pl. 1).

At the mouth of Benton Creek and along the Little North Fork of the Clearwater River a rather homogeneous biotite-plagioclase schist is underlain by a layer of banded hornblende-biotite-plagioclase gneiss. This gneiss is hetergeneous, consisting of white layers of quartz-plagioclase rock interbedded with dark layers rich in hornblende, or in hornblende and biotite, or in biotite. Small crystals of garnet occur in some layers. The thickness of individual layers ranges from ½ to 5 cm.

A similar gneissic layer continues toward the east along the North Fork of Benton Creek, and forms a sheared zone half a mile wide between the Prichard formation to the north and the Wallace formation to the south. On the basis of the interrupted stratigraphic sequence, it is evident that this gneiss zone represents one of the major faults in the area.

PETROGRAPHIC DESCRIPTION

QUARTZITE

Three types of quartzite occur in the lowest unit of the Wallace formation: medium-grained light-colored biotite quartzite, white granular quartzite, and thin-bedded biotite quartzite with or without plagioclase. Only a few discontinuous beds of medium-grained biotite quartzite are interbedded either with the schist or with the diopside-plagioclase gneiss. These layers weather readily to coarse quartz sand. The beds are thicker than those in the other types of quartzite, ranging from 1 to 5 m in thickness. Biotite flakes of medium size are evenly distributed throughout the rock.

Beds of white granular quartzite, ranging from 5 to 20 cm in thickness, are interbedded with thinner layers of biotite and biotite-plagioclase quartzite along the road leading from Bertha Hill to the river. Similar beds are interbedded with diopside-bearing quartzite and diopside-plagioclase gneiss in the southeastern part of the quadrangle. These beds are medium grained and consist of almost pure quartz with only a few biotite flakes or diopside grains. The thin-bedded layers are fine grained and contain biotite and plagioclase in varying amounts. They are interbedded either with the other quartzite layers or with the schist. Muscovite is a common additional constituent, and apatite and magnetite occur as accessories.

The thin-bedded biotite-plagioclase quartzite and diopside-plagioclase gneiss interbedded with the schist east of Bertha Hill are probably equivalent to the St. Regis or to the lower part of the Wallace formation. In the quartzite, alternating layers, 2 to 5 cm thick and rich in either quartz or plagioclase are separated by paper-thin laminae rich in biotite (fig. 10). A few green layers, 1 to 5 m thick, of diopside-plagioclase gneiss with some honblende (fig. 11) are interbedded with this quartzite.

A very thin bedded biotite quartzite is exposed under the diopside-hornblende gneiss along Gyppo Creek, 1½ miles north of the mouth of Elkberry Creek. Some of the thin-bedded layers, which consist mainly of quartz, plagioclase, and biotite, are very similar to the thin-bedded biotite quartzite interbedded with diopside quartzite in the Wallace formation along Couger Creek in Headquarters quadrangle (Hietanen, 1962). However, a major part of the formation near Gyppo Creek contains paper-thin laminae rich in sillimanite.

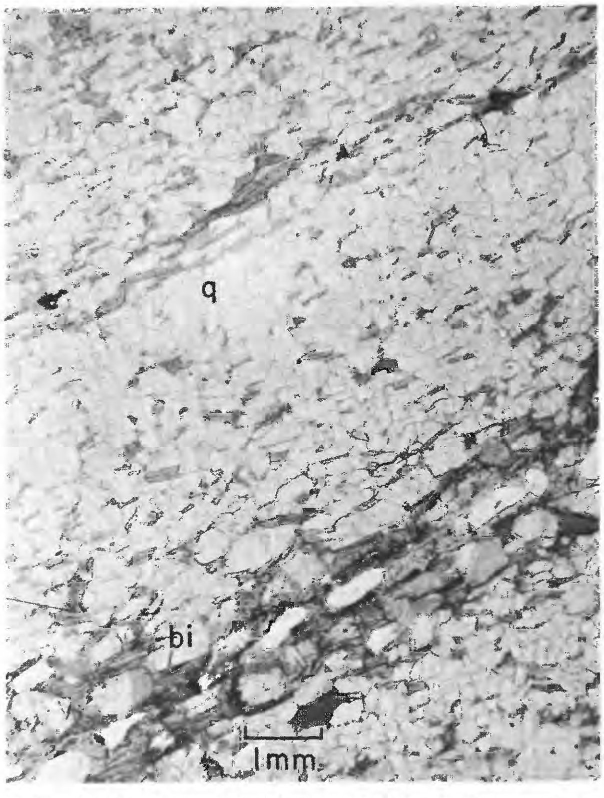

FIGURE 10.—Photomicrograph of thin-bedded biotite-plagioclase quartzite. The dark-colored thin layers contain abundant biotite (bi) in flakes of medium size. In the light-colored layers quartz (q) and plagioclase are the major constituents, with the biotite occurring only sparingly. Loc. 684 east of Bertha Hill. Plane-polarized light.

FIGURE 11.—A hand specimen of diopside-plagioclase gneiss with some hornblende (dark bands). The white bands consist mainly of plagioclase. Diopside is gray in the photograph. About half a mile east of Bertha Hill, loc. 694.

SCHIST AND THE OCCURRENCE OF MARGARITE IN IT

The schist interbedded with the quartzite in the lowest part of the Wallace formation is medium to coarse grained and contains abundant biotite and garnet, and muscovite in varying amounts. Sillimanite is not as common as it is in the upper part of the Wallace formation. In most outcrops the muscovite occurs as larger flakes than the biotite and forms thin laminae between the biotite-bearing layers. Garnet crystals are small and their number is far less than that in the upper part of the formation.

The schist units interbedded with the diopside-plagioclase gneiss in the upper part of the Wallace formation are similar to the corresponding layers in the Wallace formation in the Headquarters quadrangle (Hietanen, 1962), except that abundant light bluish-gray to white micaceous mineral occurs instead of sillimanite in the schist southwest of Township Butte. Large blocks of this micaceous material were found on the southern border of the quadrangle along a fire road

TABLE 8.—*Chemical analyses of margarite and margarite-sericite rock. Locality 1484, 1 mile southwest of Township Butte*

[Analyst: M. Balazs, U.S. Geol. Survey]

	Margarite							Margarite-sericite rock					
	Weight percent	Molecular equivalent	Number of atoms			Number of atoms after correction to H_2O=4.4 percent		Weight percent		Cation percent		Molecular norm	
SiO_2	32.59	5426	Si	4.43	8.00	4.28	8.00	SiO_2	37.38	SiO_2	35.14	Or	25.35
Al_2O_3	49.95	4900	Al	3.57		3.72		Al_2O_3	44.80	$AlO_{3/2}$	49.62	Ab	9.90
Fe_2O_3	.20	13	Al	4.43	4.45	4.00	4.02	Fe_2O_3	.18	$FeO_{3/2}$	.13	Ne	2.52
FeO	.35	49	Fe^{+3}	.02		.02		FeO	.33	FeO	.26	An	32.20
MnO	.00	00	Fe^{+2}	.04	.10	.04	.10	MnO	.00	MnO	.00	C	28.85
MgO	.25	62	Mn	.00		.00		MgO	.28	MgO	.39	Fo	.59
CaO	10.09	1799	Mg	.05		.05		CaO	6.43	CaO	6.48	Fa	.21
Na_2O	1.97	318	Ti	.01		.01		Na_2O	1.55	$NaO_{1/2}$	2.82	Ap	.05
K_2O	.69	73	Ca	1.47	2.11	1.42	2.04	K_2O	4.23	$KO_{1/2}$	5.07	Il	.12
TiO_2	.08	10	Na	.52		.50		TiO_2	.09	TiO_2	.06	Mt	.19
F	.39	205	K	.12		.12		P_2O_5	.03	$PO_{5/2}$	.02	Cc	.02
H_2O+	2.51	1393	F	.17	2.44	.16	4.01	CO_2	.01	CO_2	.01		
H_2O-	.05	------------	H	2.27		3.85		H_2O+	4.43	H_2O+	(13.88)		
			O	24.00		24.00		H_2O-	.07	H_2O-	(.22)		
Subtotal	99.12	------------						Total	99.81	Total	100.00	Total	100.00
								Molecular percent					
Less O for F	.16	------------											
								Margarite	64.6				
Total	98.96	------------						Muscovite	35.4				

leading from Township Butte southward to Lightning Point in the Headquarters quadrangle (loc. 1484, pl. 1). These blocks consist of light bluish-gray dense minerals that appear fibrous to a naked eye and resemble sillimanite in their mode of occurrence. Microscopic study shows that there are two major constituents, a white sericitic mica with an index of refraction $\gamma=1.593$ to 1.594 ± 0.001 and another micaceous mineral with indices of refraction $\alpha=1.629\pm0.001$ and $\gamma=1.639\pm0.001$. The latter mica was separated from the rock by means of heavy liquids. The fraction, which had specific gravities between 3.063 and 3.066 and contained some tiny inclusions of sericite, was analyzed chemically (table 8). This micaceous mineral is margarite. Its formula calculated from the chemical analysis is

$$(H_{3.85}F_{0.16})_{\Sigma=4.01}(Ca_{1.42}Na_{0.50}K_{0.12})_{\Sigma=2.04}$$
$$(Fe^{+2}_{0.04}Mg_{0.05}Ti_{0.01})_{\Sigma=0.01}(Al_{4.00}Fe^{+3}_{0.02})_{\Sigma=4.02}$$
$$(Si_{4.28}Al_{3.72})_{\Sigma=8.00}O_{24}.$$

Comparison with the theoretical formula $CaAl^2(OH)^2$ $Al_2(OH)_2O_{10}$ suggests that Na substitutes for a part of Ca and Al for a part of Si. The mode of occurrence suggests that it is an alteration product after sillimanite.

Chemical analysis of the margarite-sericite rock (table 8) from which margarite was separated shows considerably more potassium; this suggests that the other mica is mainly muscovite. Indeed, if all calcium is assumed to go to form margarite and the amount of margarite is calculated (64.6 molecular percent) and deducted from the rock analysis, the remainder gives a composition of muscovite with the following formula:

$$H_4(Na_{0.21}K_{1.79})_{\Sigma=2.00}(Mg_{0.06}Fe^{+2}_{0.03})_{\Sigma=0.09}(Al_{3.85}Fe^{+2}_{0.02})_{\Sigma=3.87}$$
$$(Si_{6.17}Al_{1.83})_{\Sigma=8.0}O_{24}.$$

In the schist south of locality 1484, margarite forms conspicuous white clusters that look fibrous in a hand specimen (fig. 12). They are, in all probability, an alteration product after sillimanite, which commonly occurs in similar clusters in other parts of the same formation. The schist rich in margarite is strongly sheared and finer grained than the common sillimanite-bearing schist. Quartz occurs in thin laminae which are separated by thin micaceous layers. Study under the

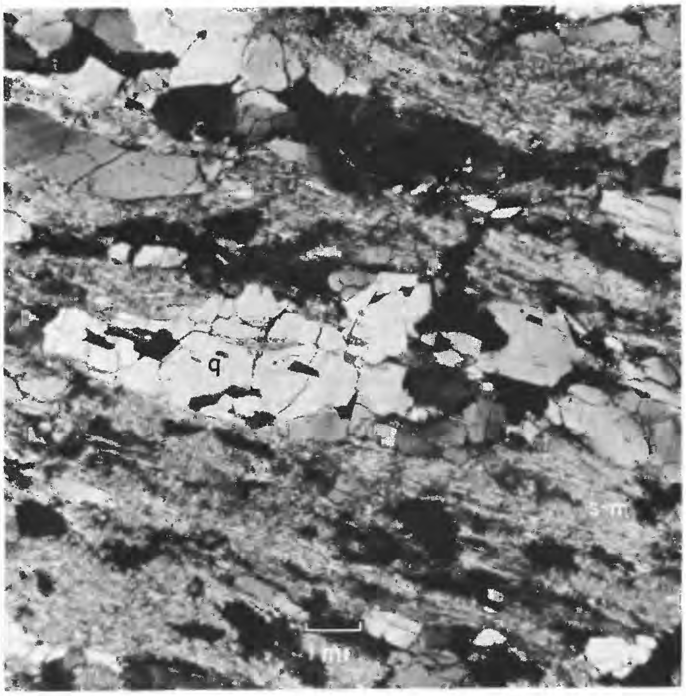

FIGURE 12.—Photomicrograph of sericite-margarite schist a mile southwest of Township Butte. Quartz (q) occurs in thin discontinuous layers that are separated by micaceous layers consisting of a fine-grained mixture of sericite and margarite (s-m). Crossed nicols.

microscope shows that sericite in these micaceous layers is intermingled with margarite and that hematite, including tiny prisms of rutile, occurs instead of biotite as flakes of medium size. The occurrence of margarite, sericite, and hematite instead of sillimanite and biotite indicates that considerably more calcium and less magnesium are present in this schist than in a normal schist of the Wallace formation. The alteration of margarite to sillimanite suggests that calcium was introduced locally into the schist. This introduction probably took place during a shearing because the micas are well oriented parallel to the shear planes. The magnesium of the biotite was removed when calcium was introduced, leaving only skeletons of biotite consisting of hematite. The potassium of the biotite probably reacted with sillimanite to form sericite.

The blocks of the margarite-sericite rock (no. 1484) were found next to a pegmatite dike. The introduction of calcium may have proceeded from the dikes to the country rock. These relations suggest that the margarite crystallized in the temperatures corresponding to those of the pegmatitic stage.

DIOPSIDE-PLAGIOCLASE GNEISS

Most layers of the diopside-plagioclase gneiss interbedded with the garnet-sillimanite schist in the Boehls Butte quadrangle contain more diopside and in places also more hornblende than the corresponding layers in the Headquarters quadrangle south of the present area. The name "gneiss" is used, even if a small portion of these layers consists of quartzite.

The diopside-plagioclase gneiss south of the mouth of Robinson Creek is thin bedded and contains a considerable amount of hornblende. The major constituents—quartz, diopside, and plagioclase—occur in equal amounts. The distribution of hornblende is highly irregular; it is concentrated to thin layers parallel to the bedding, or it is scattered as individual large grains throughout the rock, or it occurs on both sides of thin pegmatitic dikelets that cut the rock across the bedding. In its mode of occurrence it is thus similar to the metasomatic hornblende described earlier from the Headquarters quadrangle (Hietanen, 1962). The original sediment was a dolomite-bearing clayey sand that recrystallized to a fine-grained diopside-plagioclase gneiss during a regional metamorphism. Hornblende is later than the main constituents—quartz, plagioclase, and diopside. Adjacent to many small pegmatitic veins and dikes it crystallized as large grains replacing the diopside and including small grains of quartz and plagioclase. A somewhat higher content of iron in the hornblende-bearing parts of the rock suggests that some iron was added to the rock during the crystallization of hornblende.

Tremolite and epidote group minerals, especially clinozoisite, occur locally in the gneiss instead of diopside (for example, locality 1474). In specimen 669 clinozoisite with $\alpha=1.715\pm0.001$, $\beta=1.719\pm0.001$, $\gamma=1.724\pm0.001$ has replaced zoisite along the cracks and borders. The indices of refraction of the zoisite are lower than those of the clinozoisite: $\alpha=1.701\pm0.001$ and $\gamma=1.7085\pm0.001$. Also some epidote and diopside, the latter with $\gamma=1.715\pm0.001$, occur in the same rock.

METAMORPHIC FACIES OF THE BELT SERIES

The critical assemblages in the rocks of the Belt series in the southernmost part of the area are almandine-sillimanite-biotite with only a little muscovite in the pelitic layers, and diopside-hornblende in the calcareous layers. These mineral assemblages suggest recrystallization to the sillimanite-muscovite subfacies of the amphibolite facies. In the northern part more muscovite occurs, and the assemblage kyanite-andalusite-sillimanite is typical. The mutual relations among these three aluminum silicates (Hietanen, 1956) indicate that the temperature and pressure during the recrystallization fluctuated around the triple point at which all three are stable. Some staurolite occurs with the three aluminum silicates, and it seems that the upper limit of the stability field of staurolite is close to the temperature of the triple point. Because this temperature is also the lower limit of the stability field of sillimanite, it forms a natural boundary between the amphibolite and epidote-amphibolite facies as has also been suggested by Francis (1956). In the calcareous layers, tremolite and actinolite are as abundant as the diopside. This also suggests recrystallization in a somewhat lower temperature field than prevailed in the southernmost part of the area where tremolite is found only in layers that are exceptionally rich in magnesium and that did not contain enough calcium for the formation of diopside.

STRUCTURE OF THE BELT SERIES

The regional structures of the rocks of the Belt series have been discussed in another paper (Hietanen, 1961); therefore attention here is given only to some specific local features.

FAULTS

The anorthosite and its country rocks are separated by north- and east-southeast-trending faults from the upper schist and quartzite units of the Prichard formation. In the northwestern part of the area, two north-trending faults with a vertical displacement of east side up are exposed on the ridge southeast of Little Lost Lake. Near Orphan Point a north-trending fault between the anorthosite and schist was mapped on the basis of boulders, and a narrow gulley marks the extension of the same fault south of Orphan Point.

Near Pinchot Butte a fault occurs on either side of the mountain and is transected by east-northeast-trending transverse faults. Farther to the south the fault follows the gorge of Floodwood Creek and ends in an intrusive body of granite which conceals its junction with the east-southeast-trending fault zone that borders the anorthosite and its country rocks in the south.

In the northeastern part of the area several north-trending faults separate the normal sequence of the Belt series in the east from the anorthosite and its country rocks in the west. A southeast-trending fault was traced along Foehl Creek south of Buzzard Roost and another along Minnisaka Creek. These faults, together with north-trending faults, form the eastern border of the anorthosite block.

The uplift of the anorthosite block along this eastern border was probably much less than along the western border. The schist above the lower quartzite unit is petrologically similar to the schist under the upper quartzite unit on the east side of the fault zone. Both schists probably belong to the same unit, the thickness of which near Orphan Point is about 1,500 m.

In the western part of the area, the rocks of the anorthosite block are in the fault contact with the schist that occurs between the two quartzite units of the Prichard formation. If the anorthosite occurring at the 4,500-foot elevation on the southeast slope of Pinchot Butte has stratigraphically the same position as the large anorthosite bodies, and if the quartzite along Twin Springs Creek is equivalent to the lower quartzite unit that in the central part overlies the large anorthosite bodies, the amount of uplift in this vicinity is more than 400 m.

In the southern part of the area, the rocks of the Wallace formation are faulted against the anorthosite and its country rocks. Thus the vertical displacement there is much larger than along the north-trending faults, probably more than 1,000 m.

BEDDING

Bedding is distinct in quartzitic units and in some of the diopside-plagioclase gneiss; however, in most outcrops of the diopside-plagioclase gneiss, recrystallization has obliterated all structures ascribed to the equivalent formations in the Coeur d'Alene district. In the schist, bedding is visible where fine-grained quartzitic layers or biotite-plagioclase gneiss layers are interbedded. In small outcrops of coarse-grained homogeneous schist only one planar structure, the foliation, is visible, and bedding may or may not be parallel to it.

FOLDING

A large double-crested anticline in the Prichard formation in the vicinity of Monumental Buttes and Cedar Creek forms a major structure in the area. The axis of this fold plunges gently to the southeast. Another large anticline occurs along the North Fork of the Clearwater River, west of the mouth of Salmon Creek. The river valley there is an anticlinal valley, and the high ridges to the north and to the south (Smith Ridge and vicinity of Benton Butte) have synclinal structures. (See section on pl. 1.) Several other somewhat smaller folds are seen on the geologic map (pl. 1).

In most parts of the area, the fold axes plunge gently to the east or to the east-southeast, but at places horizontal or westward-plunging axes were observed. In the northern part of the area, the schist and quartzite are gently or moderately folded and the same formation covers large areas, whereas in the southern part the folding is more intense, the flanks of the folds in many localities are either steep or overturned, and a strong crenulation appears along the flanks.

The trend of the measurable fold axes varies greatly and in several localities two sets of folds are apparent. The major fold axis trends about S. 70° E. and the second axis makes an angle from 50° to 90° with it. The trend of the second fold axes is in either northeasterly or northwesterly direction coinciding with the regional trends of the Nevadan folding (Hietanen, 1961). Generally the second set of folds appears as a local crenulation and minor folding of the flanks of the major folds. In some localities, however, larger folds were formed around the northwestward-trending axes. Small overturned folds exposed along a logging road between Elkberry Creek and Gyppo Creek belong to this second set; their axes plunge 10° NW. and the axial planes dip 10° NE.

FOLIATION

Foliation parallels the bedding in the areas of gentle folding (for example, on Smith Ridge, near Indian Dip, and in many localities near Floodwood Creek) and along the flanks of steep folds. On the crests of the folds, transecting cleavage parallel to the axial plane of the folds is common (fig. 13). Thin sections show that on the crests of the folds in many thin-bedded quartzite layers some of the micaceous minerals are subparallel to the bedding plane and some subparallel to the axial plane. Thus, both of these structural planes served as glide planes during the various phases of deformation which was accompanied by recrystallization.

LINEATION

Lineation is a fine wrinkling of bedding planes, or elongation of mineral grains, or parallel orientation of pencil-shaped minerals, such as hornblende. It is clearly visible in most micaceous layers, but cannot be detected in many pure quartzite layers. However, many exceptions to this rule were found; some coarse-grained

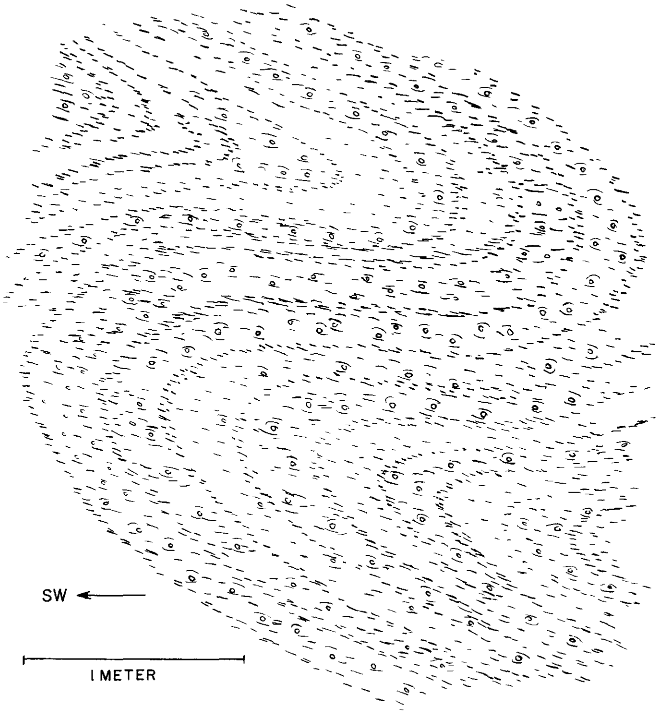

FIGURE 13.—A small fold overturned to south in garnet-mica schist of the Prichard formation about 1 mile west of Orphan Point (loc. 1023). The mica flakes are parallel to the axial plane of the folds and render to the rock a good foliation that transects the bedding on the crests of the folds. Vertical road cut facing southeast.

schist layers show only irregular crumbling and some pure quartzite layers in the Prichard formation may show a distinct lineation. The lineation in the quartzite is noticeable because of stretching of grains or because of a strong shearing. The stretching of grains is conspicuous in garnetiferous gray coarse- to medium-grained quartzite in which garnet crystals are drawn out parallel to the lineation. Near Lost Lake this lineation is parallel to the fold axis (*b* lineation). The axis of the wrinkling in the micaceous layers is either parallel to the major fold axis (*b* lineation) or parallel to the second fold axis (*b′* lineation).

IGNEOUS ROCKS

Several structurally and petrologically different groups of igneous rocks occur in the area. Coarse-grained gneissic hornblendite, gabbro, and quartz dioritic and tonalitic rocks form an older group and occur mostly as conformable bodies in the Belt series. This group is called the quartz dioritic suite, and the rocks that belong to it are, in their mode of occurrence, structure, and petrology, similar to the corresponding rocks in the Headquarters quadrangle and vicinity (Hietanen, 1962). The rocks of the quartz dioritic suite are therefore discussed in this paper only briefly.

In contrast to the quartz dioritic suite, medium-grained gabbro with fine-grained contacts and medium-grained reddish-gray granite cut the rocks of the Belt series and the anorthosite discordantly and are younger than the rocks of the quartz dioritic suite. The large body of quartz monzonite around Beaver Butte, just east of the Boehls Butte quadrangle, and that in the northernmost part of the area probably are of the same age as or somewhat older than the reddish-gray granite. The quartz monzonite and granite belong genetically to the same suite, which is here referred to as the quartz monzonitic suite. Porphyritic dikes, mainly of granitic and dioritic composition, are common both in the rocks of the Belt series and in the older intrusive series.

QUARTZ DIORITIC SUITE

SERPENTINE

A small body of serpentine occurs in the schist along the county road leading from Boehls to Elk River about 1 mile northwest of logging Camp X (western margin of pl. 1). Most of this body consists of fine-grained, dark greenish-gray serpentine with rust-colored specks, but a small portion contains radiating crystals of anthophyllite interleaved with chlorite.

Study under the microscope shows that the mesh of serpentine minerals is transected by numerous rows of tiny magnetite grains that originally crystallized along the crack of olivine. Some small specks of carbonate, talc, and a few flakes of chlorite occur as additional constituents. More chlorite and talc are in that portion of serpentine, in which anthophyllite prisms occur. Olivine with only narrow serpentinized cracks fills the interstices between the anthophyllite prisms in the fresh part of the anthophyllite-bearing outcrop. The cracks in the anthophyllite are filled by talc. Small grains of magnetite are along the cracks in the olivine and a few larger grains occur next to the anthophyllite.

The mineralogy and texture of the serpentine suggest that originally most of it was dunite and that only a small part contained anthophyllite in addition to olivine.

HORNBLENDITE

Small bodies of coarse-grained black hornblendite occur in the rocks of the Belt series and in the rocks of the quartz dioritic suite. These bodies consist of 80 to 90 percent hornblende and some biotite, plagioclase, and very little quartz. Hornblendite included in gabbro near Benton Butte probably represents accumulation of hornblende in the gabbro, but the small irregular bodies in the rocks of the Belt series were formed at least in part by replacement, their mode of occurrence being similar to the hornblendite and garnet-amphibolite in the Headquarters quadrangle and in the vicinity of Dent (Hietanen, 1962). Several small bodies, ranging from 2 to 10 m in diameter, were found in the Revett quartzite along Robinson Creek. These bodies are irregular in shape, and hornblende fills cracks and joints in the quartzite near the contact.

GABBRO

Several small bodies of gneissic medium-grained gabbro occur within quartz diorite and tonalite or in their vicinity. Thin sections show that the gabbro consists of about 55 percent hornblende, 40 percent plagioclase (An_{35}), and 5 percent quartz in small round grains, and that accessories are sphene, epidote, apatite, and magnetite. Hornblende and plagioclase occur in groups of small polygonal grains, giving an impression of granulation and recrystallization.

QUARTZ DIORITE

The quartz diorite exposed in the southeast corner of the mapped area (pl. 1) is a part of a large body in the northern part of the Idaho batholith. Its petrography and structure have been discussed in the earlier paper (Hietanen, 1962). Two large and several small bodies of quartz diorite are exposed in the southern part of the Boehls Butte quadrangle. Most of these bodies consist of heterogeneous hornblende-plagioclase rocks with a considerable amount of quartz. Gabbroic and hornblenditic portions are included in the quartz diorite on Benton Butte. Smaller inclusions consisting of hornblende-rich rock are common elsewhere. The petrography and structure of these bodies are similar to those of the satellitic bodies of the same composition in the Headquarters quadrangle.

TONALITE

The occurrence of tonalite is restricted to the southern part of the Boehls Butte quadrangle in the same area where quartz diorite occurs. The bodies are small, have a gneissic structure, and show gradational contacts toward the country rocks. Most are petrographically similar to the gneissic tonalite in the Headquarters quadrangle. Many are distinctly banded and grade over to a banded gneiss in which veins consist of plagioclase and quartz. Biotite is usually the only dark constituent in these tonalites; magnetite and zircon occur as accessories.

PLAGIOCLASE PEGMATITE

Heterogeneous medium-grained pegmatitic veins consisting mainly of quartz and plagioclase are common in the rocks of the Belt series in the map area. These pegmatites are similar to those in the Headquarters quadrangle.

A very coarse grained sheared plagioclase-biotite pegmatite, about 5 m thick, is exposed on the road cut about half a mile east of the mouth of the Little North Fork of the Clearwater River. This dike extends eastward and is exposed again along the road leading from the river to Bertha Hill. The pegmatite is strongly sheared, with biotite covering the irregular shear zones. The quartz has sutured borders and is strongly deformed. Large undeformed albite grains are embedded in the deformed quartz. Magnetite and zircon occur as accessories. The pegmatitic veins in the biotite-plagioclase schist north of the mouth of the Little North Fork of the Clearwater River consist of quartz and plagioclase. Plagioclase-biotite pegmatites were found also 0.3 mile south of the mouth of the Little North Fork of the Clearwater River and along the logging roads curving around the tributaries of Benton Creek about 3 miles northeast of Bertha Hill along the same structural zone as the two other occurrences.

QUARTZ MONZONITIC SUITE

Two large bodies of rocks of the quartz monzonitic suite are exposed in the area studied, one in the southeastern and the other in the northern part of the area. In addition to these two plutons, granitic stocks and dikes belonging to this suite occur at the mouth of Floodwood Creek, opposite Larsons Cabin, south of Stocking Meadow Lookout, and southeast of Orphan Point.

QUARTZ MONZONITE

Quartz monzonite around Beaver Butte is a medium-grained light-gray rock in which minerals are oriented at random. The dark constituents are hornblende and biotite or biotite alone. Most biotite flakes are small and occur in clusters. Plagioclase (An_{20-43}) forms euhedral to subhedral zoned crystals, many of which are included in large orthoclase grains (fig. 14). Quartz is abundant and occurs as irregularly shaped grains among the other minerals. Sphene, magnetite, apatite, and zircon are the common accessories.

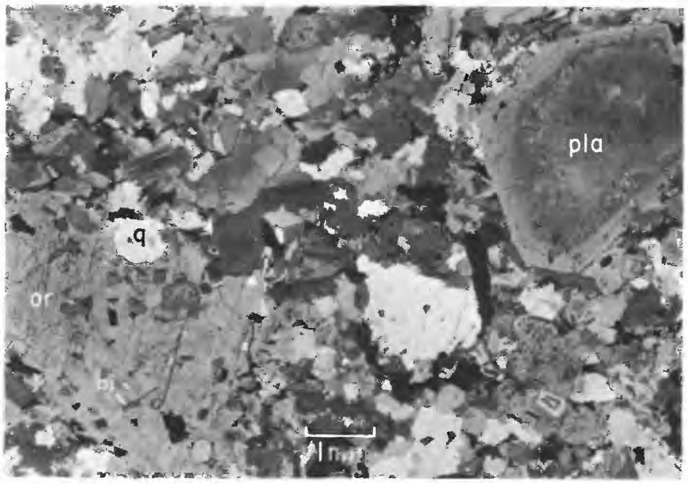

FIGURE 14.—Photomicrograph of quartz monzonite along the North Fork of the Clearwater River 2½ miles east of the mouth of Thompson Creek. Euhedral plagioclase (pla) occurs as phenocrysts in the medium-grained rock. The orthoclase grains (or) are anhedral; larger grains include small euhedral plagioclase, some quartz (q), and biotite (bi). Crossed nicols.

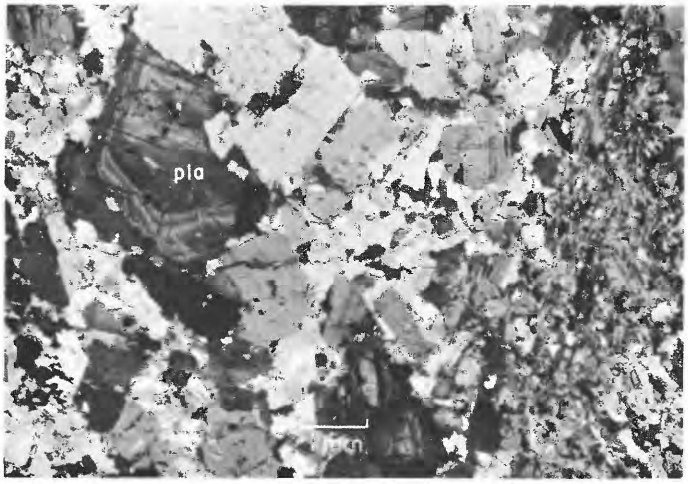

FIGURE 15.—Photomicrograph of quartz monzonite along the North Fork of the Clearwater River 1 mile east of the mouth of Thompson Creek. Fine-grained portion on the right is dioritic in composition. This fine-grained gneissoid rock contains veinlets of similar medium-grained quartz monzonite (on the left) as shown in figure 14. pla, plagioclase. Crossed nicols.

A small mass of fine-grained dark gneissoid rock of dioritic composition is included in the quartz monzonite body near its western border along the North Fork of the Clearwater River. Quartz monzonite forms narrow veinlets and small pods in this fine-grained rock. Thin sections made from the quartz monzonitic portions show abundant strained quartz between euhedral zoned plagioclase grains (fig. 15). Straining of quartz indicates a slight postcrystallization deformation.

Quartz monzonite in the northern part of the area is very similar to that near Beaver Butte. In the center of the pluton the minerals are oriented at random; the border zone, however, is gneissic. The border zone contains more biotite and less orthoclase than the main part of the pluton.

The quartz monzonite dikes on the east side of Breakfast Creek at Boehls have a gneissic structure parallel to the walls. Biotite clusters are drawn out into long shreds, and plagioclase tablets are oriented parallel to the planar structure. Under the microscope the borders of the mineral grains appear more irregular than those in the large intrusive bodies.

GRANITE

The granite of the stocks is gray and fine to medium grained. Gneissic structures occur near many contacts, but the centers of the stocks are massive. A stock south of Stocking Meadow Lookout and a part of two large bodies, one at the mouth of Floodwood Creek and the other southeast of Orphan Point, consist of coarser grained granite than the rest of the occurrences. The contact zones and locally also the centers of these three bodies show a well-developed planar structure parallel to the bedding of the country rocks. The minerals in the granite are quartz, plagioclase (An_{23-30}), orthoclase, biotite, and some muscovite. Magnetite, sphene, apatite, and zircon are the common accessories.

The granite exposed under the middle schist unit of the Prichard formation in the canyon of the Little North Fork of the Clearwater River in the northeast corner of the area is coarse to medium grained, gray, and heterogeneous. This granite includes abundant schlieren of schist and in many places grades over to schist. Pegmatitic veins parallel to the planar structure are common. These features suggest that much of the granite is a product of granitization of the schist, the amount of intrusive material being minor if not negligible.

The granite dikes are either equigranular or porphyritic. Phenocrysts in the porphyritic dikes are plagioclase and hornblende or biotite. The groundmass is medium grained and granoblastic, and consists of quartz, plagioclase, orthoclase, and some biotite. Many dikes are strongly sheared; in these the biotite occurs in long shreds parallel to the planar structure which is parallel to the walls of the dikes. Granite cuts anorthosite discordantly on the north side of the North Fork of the Clearwater River opposite Larson Cabin. This granite is fine grained, and tiny biotite flakes are evenly distributed throughout it. An exceptionally light-colored fairly coarse-grained granite occurs in a dike at the head of Benton Creek. In this granite, euhedral small albitic plagioclase crystals and large orthoclase crystals are included in quartz. Biotite is altered to chlorite and some calcite is present, both of which suggests a low temperature during the latest phase of crystallization of this dike.

ORTHOCLASE DIKE

A dike, about 2 m thick, and consisting of white medium-grained equigranular homogeneous rock, was found along the Little North Fork of the Clearwater River south of the junction of the road leading from Boehls to Elk River. This dike occurred near the contact of the anorthosite and its country rock, biotite-plagioclase gneiss, and cut the country rock discordantly. The dike was quarried out during the following spring when the road was widened. Megascopically this dike rock resembles the white anorthosite nearby except that no kaolin coating occurs on the weathered surface of the dike rock and there is a very slight pinkish hue to it. Study under the microscope showed that this dike consists of almost pure orthoclase with very little plagioclase, sericite, magnetite, tourmaline, rutile, and a few small myrmekitelike blebs of quartz. Some coarse perthitic plagioclase occurs in places. A chemical analysis (table 9) shows about 80 molecular percent orthoclase, 16 percent albite, and a little more than 1 percent anorthite. The analysis shows a considerable amount of barium—apparently contained in orthoclase.

TABLE 9.—*Chemical composition of orthoclase dike south of Boehls (loc. 633)*

[Analyst: Lois D. Trumbull, U.S. Geol. Survey]

Weight percent		*Cation percent*		*Molecular norm*	
SiO_2	64.86	SiO_2	59.89	Q	1.63
Al_2O_3	18.80	$AlO_{3/2}$	20.45	Or	80.60
Fe_2O_3	.01	$FeO_{3/2}$	.01	Ab	15.80
FeO	.07	FeO	.05	An	1.40
MnO	.00	MnO		C	.44
MgO	.00	MgO		Fs	.06
CaO	.30	CaO	.30	Ap	.03
BaO	.47	BaO	.17	Il	.02
Na_2O	1.77	$NaO_{1/2}$	3.16	Mt	.02
K_2O	13.55	$KO_{1/2}$	15.95		
TiO_2	.01	TiO_2	.01		
P_2O_5	.01	$PO_{5/2}$	.01		
CO_2	.00	CO_2			
H_2O+	.08	H_2O	[.24]		
H_2O-	.01				
			100.00		
Total	99.94	O	160.35		100.00
		OH	.48		
		Total	160.83		

Mode calculated from molecular norm	
Quartz	1.61
Orthoclase ($Or_{88}Ab_{12}$)	91.44
Plagioclase ($Ab_{78}An_{22}$)	6.36
Muscovite	.05
Tourmaline	.11
Apatite	.03
Ilmenite	.02
Magnetite	.04
Total	99.66

YOUNGER GABBRO SILLS

Four large sills of pyroxene gabbro intrude the rocks of the Belt series. Three of them occur near the northernmost anorthosite body and one just east of the anorthosite body along Cedar Creek. The centers of these sills consist of medium-grained, massive rock in which dark augite crystals and light-colored plagioclase (An_{74}) are clearly visible. Thin sections show that augite grains are rimmed by light-green hornblende and that spots of hornblende occur in the grains. Toward the borders and locally also in the centers of the sills all augite is altered to hornblende. Plagioclase occurs in clusters of small lath-shaped crystals between the large rounded augite crystals. A few reddish-brown biotite flakes, some magnetite, and apatite are the additional constituents. Locally abundant zoisite occurs in those parts of the sills where augite is altered to hornblende, as on the south side of Big Talk Lake. There is every gradation from this massive gabbro that occurs in the center to foliated amphibolite and to a fine-grained dark hornblende gabbro near the contacts of the sills.

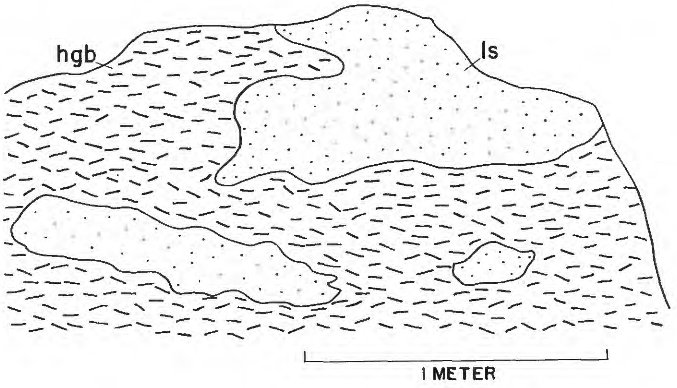

FIGURE 16.—Inclusions of lime-silicate rock (ls) in gneissic hornblende gabbro (hgb) on North Butte. Vertical wall facing north.

Most of the center of the sill at the mouth of Cedar Creek consists of similar massive gabbro, but the border zones and a part of the center are fine-grained amphibolite. Amphibolite exposed next to quartzite and to lime-silicate rock contains some garnet. This garnet amphibolite is similar to that which occurs as conformable small bodies in the schist, and its relation to the gabbro is not clear because no contact between the two is exposed.

In most localities the border zones of the gabbro sills consist of dark fine-grained rock which is either massive or is slightly schistose and in which hornblende is the only dark constituent. These border zones apparently represent chilled borders and indicate that the gabbro sills were emplaced at shallower depths than the other intrusive bodies in this area, and thus after the area had been uplifted and erosion had removed much of the cover. A sill-like body of fine-grained gabbro-amphibolite on Monumental Buttes is more deformed than the gabbro bodies described above. The main dark constituent is hornblende, and in most outcrops this mineral is well oriented parallel to the lineation. Only a minor part of the rock on North Butte appears massive. A discordant contact between this foliated gabbro-amphibolite and lime-silicate rock is exposed just northwest of North Butte. Rectangular fragments of lime-silicate rock in the gabbro-amphibolite in the contact zone (fig. 16) prove that the gabbro-amphibolite was emplaced by mechanical intrusion.

DIKES NOT DIRECTLY ASSOCIATED WITH THE PLUTONIC ROCKS

Most of the fault zones, many joints, and other structurally weak planes are locations for porphyritic dikes that range from dioritic to granitic in composition. These dikes transect all other rocks except the younger gabbro sills. As a rule the porphyritic dikes have fine-grained borders and coarse- to medium-grained centers.

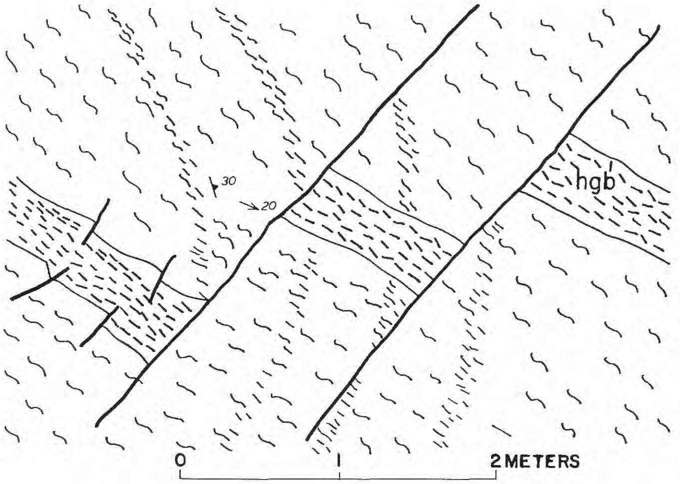

FIGURE 17.—A dike of gneissic hornblende gabbro (hgb) in plagioclase-rich biotite-hornblende gneiss. The dike is subparallel to the lineation that plunges 20° E. and transects the planar structure that strikes N. 20° W. and dips 30° NE. The faults (heavy lines) that have displaced the dike are vertical and strike N. 45° E. Horizontal surface about a mile north of Orphan Point.

The mineralogy and petrography of these dikes are similar to those of the porphyritic dikes in the Headquarters quadrangle and vicinity that have been described in an earlier paper (Hietanen, 1962).

In addition to granitic and dioritic dikes there are a few dikes of gabbroic composition. Some of these dikes have a mineralogy similar to that of the younger gabbro sills, but some consist of foliated hornblende gabbro in which the major constituents are plagioclase and hornblende with some quartz, sphene, epidote, and zircon. Foliation is parallel to the walls of the dikes that cut the planar structure of the country rock discordantly but is subparallel to the lineation of the country rock. North of Orphan Point a dike of this type is transected by small faults that indicate that the east side moved several meters toward the north-northeast along the vertical fault planes that strike N. 25° E. (fig. 17). This dike is thus older than the faults that trend in a north-northeasterly direction.

Dikes of basaltic composition occur along the southern contact of the anorthosite on the west side of the Little North Fork of the Clearwater River south of Boehls. These dikes consist of dark fine-grained rock that weathers to small round boulders. Study under the microscope shows that the major constituents are lath-shaped plagioclase (about 35 percent) and augite (about 55 percent). Brownish-green fine-grained material, in which some chlorite and ilmenite can be identified, fills the interstices.

A few small blocks of coarse-grained rock were found along the road traversing the middle anorthosite body (pl. 1), about 1 mile north of the junction to Stocking Meadows. These blocks consist of light-green prehnite and pink zoisite, thulite, with $\alpha=1.689\pm0.001$, $\beta=1.694\pm0.001$, $\gamma=1.703\pm0.001$, and $2V\approx45°$. Some tiny muscovite flakes occur as an additional constituent. Thulite has been found in several other localities in the United States (Schaller and Glass, 1942).

METASOMATIC ROCKS

AMPHIBOLITE

The amphibolite could not be conveniently included in any of the preceding groups of rocks and is therefore discussed separately. A part of amphibolite is probably of igneous origin, and some of the layers may be sedimentary, but many sill-like bodies show mineralogy and contact relations that suggest metasomatic replacement. Most of the amphibolite bodies contain abundant garnet and quartz and some biotite in addition to the regular constituents, plagioclase and hornblende. Some contain actinolite or anthophyllite instead of common hornblende. Most of the amphibolite sills occur in the surroundings of the anorthosite bodies and may be genetically connected. A few bodies appear to be discordant dikes.

MAFIC AMPHIBOLITE

Two of the occurrences of amphibolite consist of only ferromagnesian minerals. One of them is exposed along the road to Stocking Meadow Lookout about half a mile east of the lookout tower. This rock is dark greenish-gray and well foliated. It consists of actinolite, chlorite, and olivine in about equal amounts (fig. 18). The accessories are magnetite and green spinel.

FIGURE 18.—Photomicrograph of olivine- and chlorite-bearing amphibolite, half a mile east of the Stocking Meadow Lookout (loc. 1210). ol, olivine; ho, hornblende; chl, chlorite. Plane-polarized light.

The olivine is chrysolite ($Fo_{77}Fa_{23}$) with $\alpha=1.684 \pm 0.001$, $\beta=1.696\pm0.001$, $\gamma=1.717\pm0.003$, and $-2V$ =large. The optical properties of the actinolite are α (colorless)$=1.623\pm0.001$, γ(colorless)$=1.645\pm0.001$ and $Z\wedge c=16°$; these indicate that the actinolite contains about 16 molecular percent iron member.

This amphibolite may represent an ultramafic sill metamorphosed to the amphibolite facies. Olvine may be a relict mineral and chlorite crystallized because of lack of potassium.

The other occurrence of mafic amphibolite is in the anorthosite on the west slope of Goat Mountain about a mile west of the road. This amphibolite consists of hornblende and chlorite with some magnetite. The hornblende is dark green in hand specimen but shows only a very light coloring under the microscope. It is slightly pleochroic with X=colorless, Y=pale green, Z= pale bluish-green, and $Z\wedge c=17°$.

These amphibolite bodies are apparently metamorphosed ultramafic sills. Their mineralogy resembles that of the schistose hornblende-bearing gabbro-amphibolite found on and east of Monumental Buttes (specimen 1154) except that there is no plagioclase in the mafic amphibolite.

AMPHIBOLITE OF GABBROIC COMPOSITION

Some of the amphibolite bodies have a normal gabbroic composition and are probably metamorphosed intrusive sills petrologically similar to the gabbro-amphibolite on Monumental Buttes, but emplaced considerably earlier. These amphibolite bodies are dark and well foliated. The major constituents are hornblende and plagioclase, but in some bodies quartz and biotite occur in varying amounts. Every gradation from a biotite-muscovite schist to normal amphibolite can be found. In the normal amphibolite, the parallel orientation of hornblende needles makes the lineation conspicuous, whereas in the biotite-bearing occurrences, the foliation is more pronounced. The amount of quartz is highly variable. As a rule the biotite-bearing amphibolite contains more quartz than the normal amphibolite. Many of these biotite-bearing amphibolite bodies grade into biotite schist and may be metasomatic in origin. The mode of occurrence of hornblende and plagioclase in them resembles that of these minerals in the metasomatic quartz diorite and tonalite in the Headquarters quadrangle, only more hornblende occurs in the amphibolite.

GARNET AMPHIBOLITE

Garnet amphibolite occurs as sill-like bodies in the schist and in the anorthosite. Several bodies were found along the contacts of the anorthosite (pl. 1). The garnet amphibolite in the schist generally contains more biotite than that in the anorthosite. In most garnet amphibolite red garnet crystals are embedded in a dark fine- to medium-grained rock. A coarse-grained gneissic variety with abundant plagioclase was found locally in a sill next to the anorthosite along the upper drainage of Timber Creek 2 miles east of Orphan Point.

The mineralogy of the garnet amphibolite is fairly uniform. The major constituents are hornblende, garnet, plagioclase, and quartz; and common accessories are sphene and magnetite. Hornblende and plagioclase occur usually in equal amounts, but the amount of quartz and of garnet ranges from negligible to about 30 percent. Biotite is more abundant along the border zones than in the centers of the garnet amphibolite bodies.

Hornblende is of a common green variety with pleochroism X=light green, Y=green, Z=bluish green, and $Z\wedge c=20°$. Abundant round red garnet crystals are scattered throughout most amphibolite layers (fig. 19). Their diameter ranges from a few millimeters to about 10 cm, those with a diameter of ½ to 1 cm being most common. Euhedral garnet crystals showing dodecahedral faces and ranging from 3 to 4 cm in diameter were found in a few localities in the northwestern part of the area. Just west of the mapped area near Freezeout Lookout, these large crystals occur in a contact zone between garnet amphibolite and mica schist, together

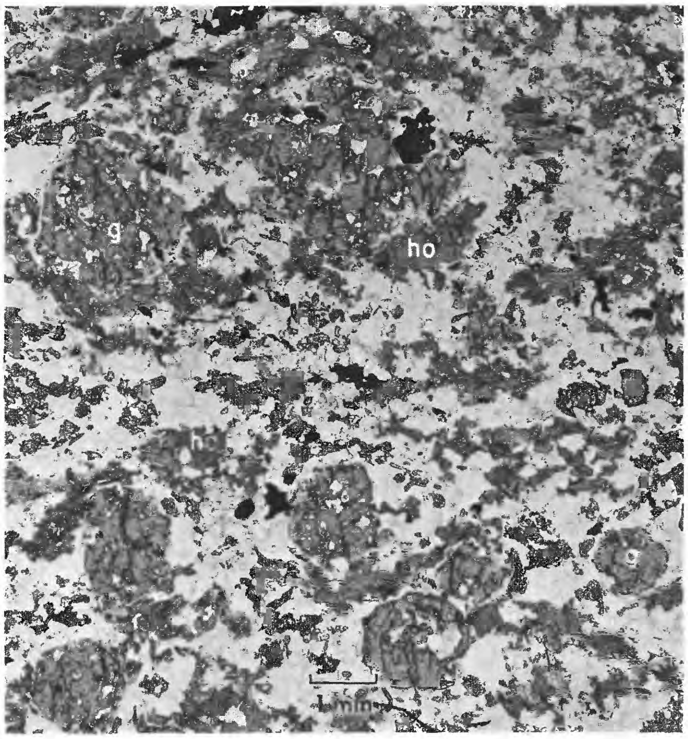

FIGURE 19.—Photomicrograph of garnet amphibolite on south slope of Goat Mountain (loc. 963). Garnet (g) crystals are round and contain inclusions of quartz. Hornblende (ho) prisms are oriented parallel to the foliation. Light-colored constituents are plagioclase and quartz; black is magnetite. Plane-polarized light.

with amphibole prisms, which are larger than those commonly found in the amphibole-bearing rocks of this region. All garnet crystals contain numerous inclusions of quartz, biotite, hornblende, and plagioclase. The index of refraction is $n=1.794\pm0.001$ to $n=1.797$ ±0.001, which, with a specific gravity of about 4.0, suggests that the garnet is almandite with about 20 percent pyrope.

Plagioclase is labradorite and occurs in anhedral grains that show polysynthetic twinning. In some occurrences, for example, in the garnet amphibolite 2 miles east of Orphan Point and in that 2 miles south of Stocking Meadow Lookout, spongelike grains of plagioclase include numerous small round quartz grains. Samples collected from these two localities contain more quartz than is normally present in the garnet amphibolite. Ilmenite grains, which are a common accessory in garnet amphibolite, are surrounded by sphene in these samples. Epidote in small grains occurs in some localities; for example, in locality 1018.

The texture of the garnet amphibolite is granoblastic or lepidoblastic, most of the hornblende prisms being parallel to the lineation. The grains of quartz are, as a rule, smaller than those of the plagioclase. A slight banding caused by accumulation of hornblende into thin discontinuous layers is common in the garnet amphibolite between Orphan Point and Goat Mountain.

Table 10 shows chemical analyses of a fine-grained quartz-bearing garnet amphibolite (no. 963) and of a coarse-grained garnet amphibolite (no. 1321). The higher content of silicon, iron, and magnesium, and a lesser quantity of calcium in no. 1321 is mineralogically seen in a larger amount of quartz and garnet and less plagioclase in this coarse-grained variety. The fine-grained variety is rich in hornblende and contains considerably more plagioclase than quartz (fig. 19).

TABLE 10.—*Chemical analyses of garnet amphibolite in weight and cation percentage, their molecular norms and measured modes*

[Analyst, Faye H. Neuerburg, U.S. Geol. Survey]

Locality and specimen No.		1321		963	
Location		Southwest of Monumental Buttes.		Three-fourths mile southwest of Goat Mountain	
Constituent					
Conventional symbol	*Symbol rearranged for cation percent*	*Weight percent*	*Cation percent*	*Weight percent*	*Cation percent*
SiO_2		59.02	57.48	57.96	57.04
Al_2O_3	$AlO_{3/2}$	13.24	15.20	13.80	16.01
Fe_2O_3	$FeO_{3/2}$	1.23	.90	1.21	.89
FeO		11.90	9.69	10.82	8.90
MnO		.34	.28	.49	.41
MgO		5.03	7.30	3.83	5.62
CaO		4.25	4.43	5.69	6.00
Na_2O	$NaO_{1/2}$	1.57	2.96	2.68	3.20
K_2O	$KO_{1/2}$	.19	.23	.28	.35
TiO_2		1.85	1.36	1.93	1.43
P_2O_5	$PO_{5/2}$	.18	.15	.19	.15
CO_2		.01	.01	.00	
H_2O+		1.03	(3.34)	.73	(2.39)
H_2O-		.09		.09	
Total		99.93	99.99	99.70	100.00
	O		162.18		162.98
	OH		6.69		4.79
	Total anions		168.87		167.77
Molecular norm					
	Q		24.11		21.83
	Or		1.15		1.75
	Ab		14.80		16.00
	An		20.85		28.75
	C		3.67		.96
	En		14.60		11.24
	Fs		16.32		14.88
	Ap		.40		.40
	Il		2.72		2.86
	Mt		1.35		1.33
	Cc		.02		
	Total		99.99		100.00
Measured mode in volume percentage					
	Quartz		24		22
	Plagioclase		25		32
	Hornblende		28		32
	Garnet		20		12
	Biotite		1		
	Magnetite		1		2
	Total		100		100

ANORTHOSITE

OCCURRENCE AND DISTRIBUTION

Three large and several small bodies of anorthosite occur in the lower part of the Prichard formation in the Boehls Butte quadrangle and vicinity. The exposed parts of all the bodies are oblong, the large bodies ranging from 6 to 12 miles in length and 1½ to 3 miles in width. Fifteen of the small bodies—all concordant layerlike or sill-like bodies—are shown on the map (pl. 1), but in addition there are numerous small aggregates and lens-shaped bodies ranging from 1 to 5 m or more in length. These small bodies are in the schist surrounding the three largest bodies. No anorthosite was found outside the area shown on plate 1. However, a feldspathized kyanite-garnet-bearing layer, similar to the country rock of the anorthosite in the Boehls Butte quadrangle, was found in the schist of the Prichard formation just west of this quadrangle 3 miles northwest of Pinchot Butte.

The southernmost anorthosite body is well exposed for a distance of more than 9 miles along the North Fork of the Clearwater River. The Little North Fork traverses the western end of this body. The anorthosite along Cedar Creek is well exposed along the logging roads that follow Cedar Creek and its tributaries, but poorly exposed elsewhere. Most of the center of the northernmost anorthosite body between Goat Mountain and Widow Mountain is also poorly exposed. However, the outcrops are sufficient for determination of the major structure, and the white powdery soil and numerous blocks of anorthosite are a good indication of the underlying anorthosite. The anorthosite weathers more readily than the schist and quartzite and forms level areas in this mountainous country where relief is more than 3,000 feet.

STRUCTURE

STRUCTURAL LOCATION AND THICKNESS OF THE ANORTHOSITE BODIES

The three largest bodies of anorthosite are situated along crests of various anticlines (section, pl. 1). For a length of about 8 miles west of the mouth of Salmon Creek, the North Fork of the Clearwater River forms an anticlinal valley. The anorthosite occupies the bottom and lower part of the slopes of this valley. On the north slope of the valley the layers of schist and quartzite overlying the anorthosite dip 30° to 40° N.; on the south slope the dip in the overlying schist is 20° to 50° S. The schist layers turn around the nose of this anticline at the mouth of Salmon Creek where the axis plunges 15° to 40° E. On the west end of the body, near Boehls, the axis plunges gently to the west or northwest, suggesting the structure to be a gentle dome along a major anticline. The thickness of this body could not be determined. The contact of the anorthosite and schist along the north side of the river follows the 2,750-foot contour line, but on the south side of the river the anorthosite is exposed up to an altitude of 3,500 feet; the bottom of the river is 1,500 feet above sea level and consists of anorthosite. If the contact is considered to have a dip of about 30° N., a thickness of about 1,380 meters is exposed. The two other bodies have a sheetlike structure on the nose of a double-crested anticline; this permits the estimation of thickness. Between Monumental Buttes and Goat Mountain and along Rocky Run Creek, the average dip of anorthosite at 30° E. suggests that the thickness of this body is about one-half of its width as shown on the map; thus more than 2 km. This thickness is about one-eighth of the length of this body. The body along Cedar Creek ranges from about ½ to 1 km in thickness in its eastern end. The folded structure makes estimation of the thickness elsewhere impossible. For the small lens-shaped bodies of anorthosite that occur parallel to the bedding in the schist, the relation of the thickness to the length is about the same as, or less than, that of the large bodies.

CONTACTS

The most common country rock around the anorthosite bodies is a coarse-grained mica schist. In many places, where quartzite is mapped next to the anorthosite, a layer of schist, 2 to 5 m thick, separates these two rock types. Most contacts of the anorthosite with the schist are gradational. The contact line on the map was drawn between the first layer of the anorthosite, which consists of about 85 percent calcic plagioclase, and the feldspathized schist, which in places contains as much as 60 percent andesine or oligoclase. The contact line between the anorthosite and the schist seems concordant when studied in the outcrops; however, the map (pl. 1) suggests that the northernmost body cuts its country rocks slightly discordantly. Near Monumental Buttes, the anorthosite is overlain by a thin layer of schist and a thick layer of quartzite, whereas about 4 miles to the west, near Lund Creek, only a thin layer of quartzite is exposed above the anorthosite and the place where the major part of the quartzite should occur seems to be occupied by the anorthosite. Furthermore, the schist 1½ miles east of Orphan Point overlies the anorthosite, whereas the same schist layers seem to dip under the anorthosite farther toward the southeast. At its western end, this anorthosite body interfingers on a large scale with a fine- to medium-grained mica schist. The outcrops in this vicinity are few and small, but a study of the soil and numerous small blocks permits the tracing of the contact fairly accurately.

Anorthosite and a feldspar-rich schist inter finger on a small scale along a logging road following the West Fork of Cedar Creek (fig. 20). At a distance of about 5 m from the contact the schist contains about 20 percent biotite as a dark constituent. The light minerals are plagioclase and some quartz. Toward the anorthosite, the amount of biotite becomes less and muscovite appears between the plagioclase grains. In a zone that extends from 1 to 5 m outside the contact the percentages of biotite and muscovite are about 5 and 10, whereas next to the contact they are 2 and 10 respectively. The anorthosite is a very pure white plagioclase rock with very little chlorite as an additional constituent.

Fault contacts between massive coarse-grained anorthosite and the rocks of the Prichard formation occur south of Orphan Point and about 2 miles sonth

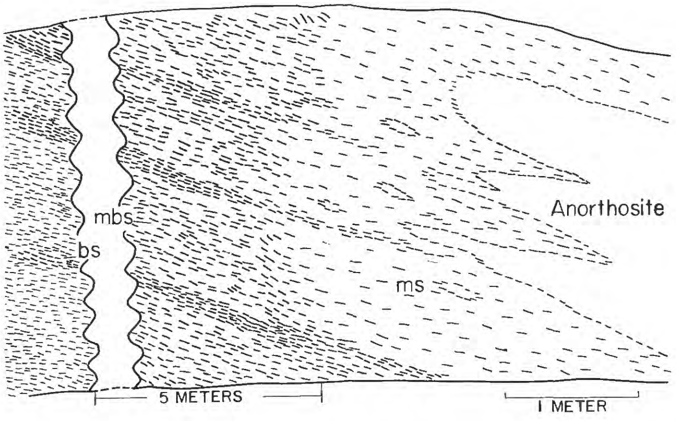

FIGURE 20.—Interfingering of the anorthosite and the schist along the West Fork of Cedar Creek. The rock (ms) next to the anorthosite consists of 88 percent plagioclase and 12 percent micas, mainly muscovite. In the next zone (mbs), 5 m wide, the amount of biotite is about 5 percent and that of muscovite 10 percent. Only biotite is found in the schist (bs) farther from the anorthosite and its amount is about 20 percent. Vertical roadcut facing south.

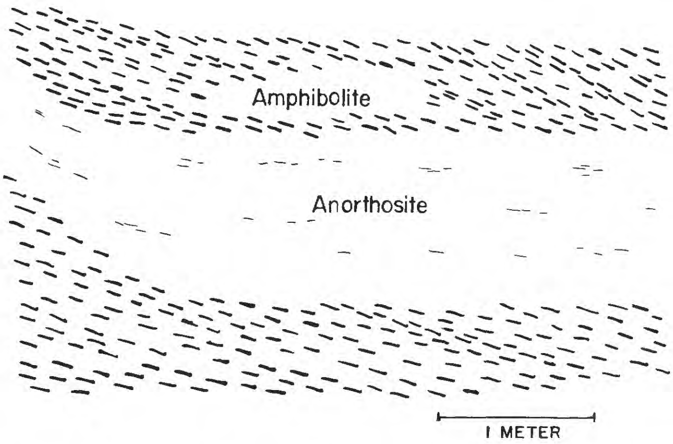

FIGURE 21.—A slightly discordant contact between the amphibolite and a small dikelike body of anorthosite west of the mouth of Cedar Creek. Vertical roadcut facing southwest.

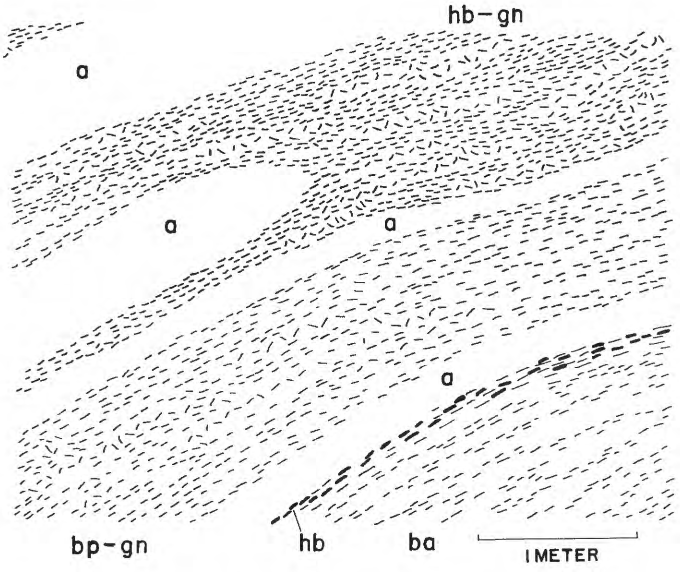

FIGURE 22.—Lens-shaped and layerlike bodies of anorthosite (a) in fine-grained biotite-hornblende gneiss (hb-gn) and in biotite-plagioclase gneiss (bp-gn). Vertical roadcut facing west about half a mile south of Boehls (loc. 1197). a, anorthosite; hb, coarse-grained hornblende-biotite rock; ba, biotite-rich anorthosite.

of Boehls. A strongly sheared plagioclase-rich rock suggesting a fault between schist and anorthosite was also found at the mouth of Lund Creek where a thin layer of schist is exposed between the anorthosite and the quartzite.

Most of the contacts between the amphibolite and anorthosite are concordant and fairly sharp. A slightly discordant contact between the anorthosite and the amphibolite was observed only near the mouth of Cedar Creek (fig. 21). Lens-shaped and layerlike bodies of anorthosite in a fine-grained biotite-hornblende gneiss are exposed along the logging road south of Boehls (fig. 22). The contact at the end of lenses is comparable with that at the end of the fingers in figure 20 except that no muscovite appears in this locality.

PLANAR STRUCTURES

About 50 percent of all anorthosite exposed is massive, but the other 50 percent locally shows a planar structure caused by an arrangement of dark constituents, biotite or hornblende, along discontinuous planes. The center and most of the western part of the northernmost body and a part of the body at Cedar Creek consist of massive coarse-grained anorthosite in which constituent minerals are distributed and oriented at random. In the southernmost large body, in which only about 30 percent of the exposures consist of massive anorthosite, planar structure is common along the borders and locally occurs in the center also. In all bodies exposed, planar structure generally is more prevalent along the borders than in the centers, but exceptions to this generalization are not rare; for example, the anorthosite between Goat Mountain and Widow Mountain locally preserves its massive texture up to the contact, and an excellent planar structure occurs in several outcrops in the centers of the two large southern bodies.

Differential weathering brings out another planar structure that is common in all small bodies and along borders of the large bodies. This structure is a thin lamination; the more deeply weathered bytownite laminae, 1 to 5 mm thick, alternate with the less weathered layers that consist mainly of andesine with some bytownite and are 2 to 10 mm thick. Also this structure is more common along the borders than in the centers of the bodies.

Most of the outcrops of anorthosite along the North Fork of the Clearwater River, along the Little North Fork, and along Cedar Creek show a beddinglike structure (fig. 23). This structure is caused by accumulation of constituent minerals like muscovite, biotite, kyanite, and hornblende into thin parallel or subparallel layers. In some localities along Cedar Creek, the biotite-bearing layers are paper thin and occur at regular intervals, giving a laminated structure to the anorthosite. The anorthosite between the biotite laminae is a white coarse-grained variety that consists of pure plagioclase. Along the North Fork of Clearwater River and along the Little North Fork, the layering is more irregular, and sheetlike inclusions of coarse-grained muscovite-biotite schist with the aluminum silicates—kyanite, andalusite, and sillimanite—are common. Small lenticles and scattered grains of plagioclase occur in the included schist. In some dark layers on both sides of the river west of Larson Cabin and along the road between Goat Mountain and Monumental Buttes hornblende is found with aluminum silicates. The

FIGURE 23.—Layering in anorthosite in a roadcut facing southeast on the north side of the North Fork of the Clearwater River just west of Larson Cabin. Thick layers that consist of almost pure plagioclase are less weathered than the thin hornblende bearing more schistose layers.

layering in these localities is very regular, the dark layers range from 2 to 5 cm in thickness and the plagioclase layers from 5 to 20 cm in thickness; the outcrops resemble bedded marble when seen from a distance (fig. 23).

In the northernmost anorthosite body biotite- and muscovite-bearing layers are found locally, but a layering caused by accumulation of hornblende into thin sheets is more common, especially in the thinner parts of the body south of Goat Mountain. The thickness of the hornblende-bearing layers ranges from 2 mm to 5 cm and that of the interbanded plagioclase layers from 1 to 10 cm. Some of the hornblende-bearing layers are thicker and resemble amphibolite. Coarse-grained massive anorthosite consisting of almost pure plagioclase occurs as layers 2 to 10 m thick with all layered varieties.

The planar structure is always parallel to the bedding of the surrounding rocks. This orientation is especially noticeable in the small bodies that consist mainly of fine- to medium-grained banded anorthosite.

FOLDING OF PLANAR STRUCTURE

Many outcrops of anorthosite along the North Fork of Clearwater River show round folds (fig. 24). Folding on a larger scale becomes evident when the orientation of planar structure is measured with a compass. It was possible to do such measuring only along the North Fork of Clearwater River, along the Little North Fork, and along Cedar Creek. The anorthosite between Goat Mountain and Widow Mountain is too poorly exposed to allow a detailed determination of structure. The numerous boulders, many of which measure 3 m and more in diameter, consist of massive anorthosite. The fold axis could be measured in several places in the two southern bodies and was found to be parallel to the fold axis in the surrounding schist and quartzite.

An intense minor wrinkling of a thin layering is common in some anorthosite blocks near the contact 1½ miles south of Orphan Point. No outcrops occur at this locality (no. 1744), and therefore the orientation of the axis of the wrinkling could not be determined. The neighboring schist shows a wrinkling around the fold axis, which plunges 15° E. Many good outcrops between this locality and Orphan Point consist of massive anorthosite.

LINEAR STRUCTURE

Linear structure is rare in the anorthosite. Only some of the contact zones of small bodies of banded anorthosite and some of the hornblende-bearing layers in the thin southern part of the northernmost large body

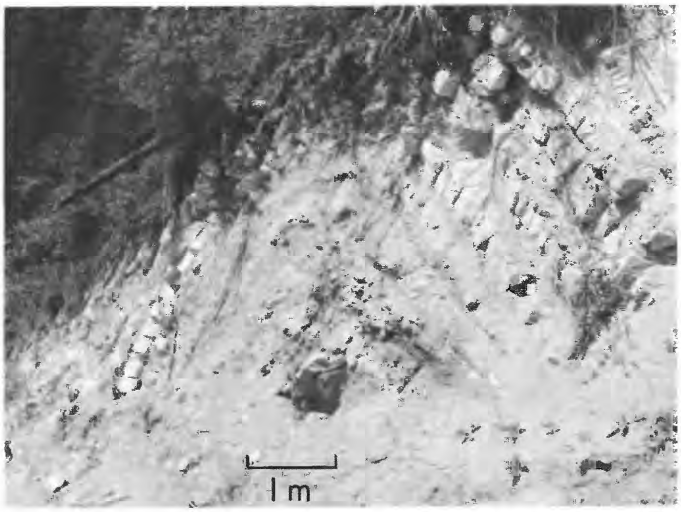

FIGURE 24.—A fold in the same layered anorthosite as shown in figure 23. The axis of folding is plunging gently toward the observer.

show a crude lineation. This structure is caused by a parallel orientation of the hornblende prisms and elongated flakes or groups of small flakes of biotite. The lineation is parallel to the corresponding structure in the surrounding country rock.

INCLUSIONS

Inclusions of schist are extremely common in the anorthosite that shows a beddinglike structure, but rare in the massive variety. Inclusions of amphibolite were found at only a few localities. Quartzite and lime-silicate rock are included in anorthosite between the forks of Cedar Creek.

The schist inclusions are oriented with their longest dimension parallel to the planar or linear structure of the anorthosite. Their size is highly variable; the sheetlike inclusions range from a few centimeters to 100 m or more in length. The smallest inclusions are only 1 to 2 mm thick, but the larger ones are 1 to 10 m thick. Generally all these inclusions are rich in aluminum silicates, and plagioclase is their main light constituent. Thus, their mineral content is very similar to that of the schist that occurs next to the anorthosite bodies.

A few large bodies of amphibolite are included in the three largest bodies of anorthosite as shown on the map (pl. 1). Small inclusions of amphibolite were found in an anorthosite boulder along the road at Goat Mountain.

PETROGRAPHIC DESCRIPTION

There is a considerable variation in the mineral content, texture, and chemical composition of the anorthosite in individual bodies. The three largest bodies are therefore described separately. The anorthosite in the small bodies differs from that in the large bodies in several respects.

ANORTHOSITE ALONG THE NORTH FORK OF THE CLEARWATER RIVER

The southernmost of the three large bodies of anorthosite lies along the North Fork and along the Little North Fork of the Clearwater River. Most outcrops consist of coarse-grained anorthosite that shows a beddinglike structure. A massive variety occurs only in some localities south of Boehls and along the river in the middle of the body, where it constitutes about one third of the entire body. The anorthosite with discontinuous layers rich in micaceous minerals has a coarsely granular texture owing to the roundness of the plagioclase grains. Small flakes of biotite and muscovite are common between the large plagioclase grains. The anorthosite weathers to chalky white powder, and the outcrops are white because they are beginning to weather. The fresh anorthosite is light bluish or bluish gray, rarely light brownish.

LIGHT-COLORED LAYERS

The white to light-grayish layers consist of about 97 percent plagioclase with very little quartz, muscovite, biotite, chlorite, sphene, magnetite, ilmenite, rutile, epidote, and occasional kyanite, andalusite, sillimanite, garnet, cordierite, hornblende, and orthoclase. The amount of these additional constituents rarely exceeds 3 percent except in the coarsely granular layers where the amount of micas is larger, ranging from 5 to 10 percent. Small masses of laumontite were found near Boehls; a zeolite, probably also laumontite, was found in one thin section studied (no. 611).

As a rule, two varieties of plagioclase, bytownite and andesine, are present in the anorthosite. The composition of twinned plagioclase was determined by universal stage and use of Nikitin's curves. The results were verified and the untwinned grains measured by immersion method. Bytownite—usually An_{85}—occurs in small grains between the large round andesine grains and also as numerous tiny platelike inclusions in the andesine. The composition of the andesine is An_{43} to An_{45}, but An_{32}, An_{36}, and An_{47} were each measured by universal stage in one specimen. Both types of plagioclase are twinned but not zoned (fig. 25). The only exception is found in a weak reversed zoning in the

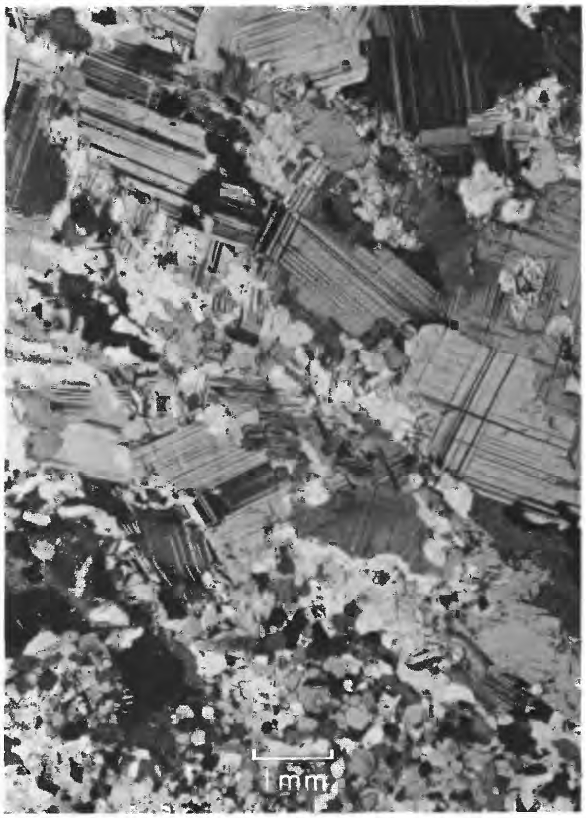

FIGURE 25. Photomicrograph of a light-colored layer in anorthosite similar to the one shown in figures 23 and 24. The large grains are andesine and the small ones bytownite. (Loc. 611)

FIGURE 26. Photomicrograph of anorthosite 1½ miles east of Larson Cabin on the north side of the North Fork of the Clearwater River. (Loc. 603.) Large twinned grains (a) are andesine. Inclusions and small grains (b) are bytownite. Crossed nicols.

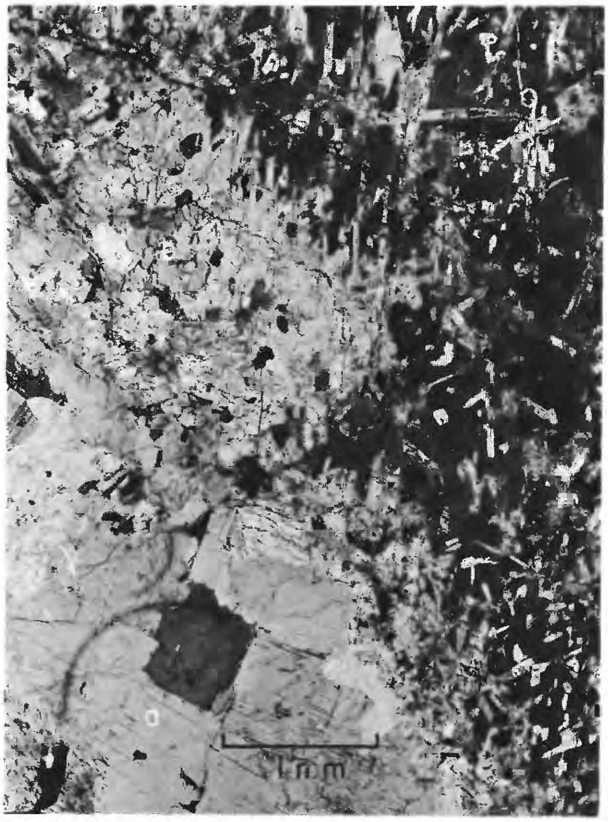

FIGURE 27. Photomicrograph of anorthosite 1¼ miles east of Larson Cabin on the south side of the North Fork of the Clearwater River. The large grains on the upper right and the lower left corner are andesine (a). Small grains between these two large grains and the inclusions in the upper right corner are bytownite (b). (Loc. 605)

andesine in specimen no. 600. The relation between the types of plagioclase is illustrated in the photomicrographs in Figures 26 and 27. In Fig. 26 the large grains (a) that show complex twinnings consist of An_{43}, and the platelike inclusions and small grains (b) consist of An_{85}. Larger bytownite grains show uneven extinction resembling the extinction in the spots occupied by groups of small grains. The long dimension of most bytownite inclusions in andesine is parallel to some crystallographic face of the host but some inclusions have a random shape orientation. Each group of inclusions in one individual andesine grain has a contemporaneous extinction and a common cleavage (010), which together indicate that the inclusions have the same crystallographic orientation in spite of a random shape orientation. The small bytownite grains between the andesine grains are oriented at random.

Figure 27 shows the contact between a large round andesine grain (a) and a group of small bytownite grains (b). The light-colored lamellae in the dark andesine are bytownite. Platelike extensions from the bytownite grains extend into the andesine along its border. The bytownite grains adjacent to the andesine grain are larger than those that commonly occur between the andesine grains, and they have an extinction that is parallel to the extinction of the bytownite lamellae in the andesine. A ghostlike mosaic texture, similar to the one found in many larger bytownite grains, is common in this bytownite adjacent to the andesine and suggests that several small grains recrystallized to form a larger grain.

In some thin sections made of layers that contain aluminum silicates small remnants of andalusite found in the centers of several bytownite inclusions suggest that at least a part of the bytownite crystallized replacing the andalusite. In these layers chlorite is the most common dark constituent and is probably an alteration product after biotite. The minor constituents occur as small grains between the plagioclase grains or are included in the andesine. The micaceous minerals are usually oriented parallel to the planar structure, but the small flakes included in the andesine are either oriented at random or parallel the cleavage of the andesine. Small light-colored grains of rutile are included parallel to the cleavage in the chlorite. Brown rutile occurs as larger grains with ilmenite or is included in ilmenite; (for example, specimen nos. 599 and 600). In some other specimens, as in no. 613, ilmenite is surrounded by sphene.

In some layers of anorthosite along the North Fork of the Clearwater River small scattered grains or groups of grains of kyanite, andalusite, and sillimanite occur as major additional constituents. Numerous bluish kyanite prisms ranging from 1 to 2 cm in length and showing $\alpha=1.1712\pm0.001$, $\beta=1.720\pm0.001$, and $\gamma=1.728\pm0.001$ were found southwest of Larson Cabin (loc. 607). In this locality chlorite, with some rutile and hornblende, occurs with kyanite (fig. 28). Chlorite is uniaxial and positive, and shows $\gamma=1.585\pm0.001$, suggesting it to be clinochlore. Hornblende grains are few and small and show pleochroism Z=blue green, Y=green, X=light yellowish green.

Near Boehls, anorthosite layers, light gray to white, and 10 to 100 cm thick, are interbedded with thin layers that contain biotite, cordierite, and garnet and with some others that contain hornblende as the dark constituent. Accumulations of tourmaline as blue radiating crystals and masses of tiny dark-green prisms were found on the east side of Breakfast Creek near the bridge at Boehls. The green prisms have $\epsilon=1.620$ to 1.623 ± 0.001, $\omega=1.649$ to $1.652+0.001$, and $\omega-\epsilon\ 0.029=\pm0.001$; these data suggest that this tourmaline consists of 30 percent iron compound, schorlite, and 70 percent magnesium compound, dravite. In the same locality euhedral to subhedral prisms of laumontite with $\alpha=1.504\pm0.001$, $\gamma=1.516\pm0.001$, and $Z\wedge c=44°$ form small accumulations and line the walls of vugs and cracks in the anorthosite.

FIGURE 28.—Photomicrograph of kyanite (ky) and chlorite (chl) in layered anorthosite half a mile southwest of Larson Cabin (loc. 607). pla =plagioclase. Plane-polarized light.

On the east side of the Little North Fork of the Clearwater River south of Boehls, layers of fine-grained biotite-hornblende gneiss with pegmatitic veinlets and lenticles alternate with layers and lens-shaped bodies of anorthosite (fig. 22). The anorthosite contains light greenish-gray fine-grained granular layers and bluish-gray gneissic hornblende-bearing layers. The fine-grained layers are granoblastic and consist of polygonal grains of plagioclase with some epidote and chlorite. The gneissic layers show a fine banding in weathered surface; pinkish laminae about 1 mm thick alternate with white laminae that are from 2 to 5 mm thick. Thin sections show that the pinkish laminae consist of small grains of bytownite that have been altered to kaolin along their cracks. Most of the cracks are subparallel to the layering. The white to grayish laminae consist of large grains of a more sodic plagioclase, which looks fresh and has no cracks. Hornblende prisms are oriented parallel to the layering. Some sericite, chlorite, epidote, sphene, and ilmenite-magnetite occur as additional constituents.

The sill-like body of coarse-grained anorthosite that occurs in amphibolite just west of the mouth of Cedar Creek (fig. 21) is almost pure plagioclase, mottled white and gray, and has only a few streaks of biotite, in part altered to chlorite.

Chemical analysis of a coarse-grained white layer on the south side of the North Fork of the Clearwater River south of Larson Cabin (table 11, no. 611) shows that these layers contain less than 1 percent combined iron and magnesium. The amounts of aluminum and calcium are high, and there is more sodium than potassium, as would be expected in normal anorthosite.

DARK LAYERS

The dark layers constitute 10 to 20 percent of the total amount of the layered anorthosite exposed. The minerals constituting the dark layers are the same as those of the light-colored layers, but their quantities vary greatly. Some of the dark layers are rich in biotite, some in biotite and muscovite, and still others in hornblende. Aluminum silicates, especially kyanite, are far more abundant in dark layers than in light-colored layers. Generally, the micaceous layers are more irregular in their shape and distribution than the

TABLE 11.—*Chemical analyses of anorthosite in weight and cation percentage, molecular norms, and modes*

Locality and Specimen no.	1747 [1]	949 [2]	1016 [2]	1015 [2]	611 [3]	608 [4]
Rock type	Gabbroic anorthosite.	Anorthosite	Anorthosite	Anorthosite	Anorthosite	Hornblende-kyanite-bearing layer in anorthosite.
Location	2 miles southeast of Orphan Point.	1.9 miles north of junction to Stocking Meadow Lookout.	South of Monumental Buttes.	South of Monumental Buttes.	North Fork of Clearwater River one-fourth mile southwest of Larson Cabin.	North Fork of Clearwater River one-fourth mile southwest of Larson Cabin.
Weight percent						
SiO_2	39.88	47.06	52.70	50.54	53.47	46.68
Al_2O_3	29.03	32.61	28.84	26.82	29.08	27.49
Fe_2O_3	2.14	.18	.03	.26	.22	1.04
FeO	2.34	.65	.50	2.60	.37	4.50
MnO	.05	.01	.01	.04	.01	.07
MgO	2.33	.28	.59	2.58	.13	5.70
CaO	20.59	16.03	11.73	11.67	11.63	7.98
Na_2O	.54	1.97	4.41	3.51	4.65	2.73
K_2O	.21	.29	.28	.40	.06	.53
TiO_2	.10	.05	.08	.09	.08	.12
P_2O_5	.01	.03	.01	.01	.04	.04
CO_2	.01	.00	.00	.01	.00	.00
H_2O+	2.05	1.00	.66	1.58	.46	2.66
H_2O-	.27	.07	.06	.06	.03	.41
Total	99.55	100.23	99.90	100.17	100.23	99.95
Cation percent						
SiO_2	37.26	43.57	48.11	46.67	48.59	43.71
$AlO_{3/2}$	31.97	35.58	31.02	29.19	31.14	30.33
$FeO_{3/2}$	1.50	.13	.02	.18	.15	.73
FeO	1.83	.50	.38	2.01	.28	3.52
MnO	.04	.01	.01	.03	.01	.06
MgO	3.24	.38	.80	3.55	.18	7.95
CaO	22.84	15.89	11.46	11.54	11.31	8.00
$NaO_{1/2}$	.98	3.54	7.80	6.28	8.19	4.96
$KO_{1/2}$	.25	.35	.33	.47	.07	.63
TiO_2	.07	.03	.06	.06	.05	.08
$PO_{5/2}$	.01	.02	.01	.01	.03	.03
CO_2	.01	.00		.01	.00	
H_2O	(3.19)	(3.09)	(2.01)	(4.86)	(0.25)	(8.30)
	100.00	100.00	100.00	100.00	100.00	100.00
O	150.30	156.45	157.64	153.21	159.96	148.27
OH	6.39	6.18	4.02	9.72	.50	16.60
Total anions	156.69	162.63	161.66	162.93	160.46	164.87
Molecular norm						
Q					0.95	
Or		1.75	1.65	2.35	.35	3.15
Ab		17.25	39.00	31.40	40.95	24.80
An	74.95	79.23	57.20	56.10	56.30	39.75
Ne	2.94	.27				
Kp	.75					
C	.76		.01		.36	8.84
Wo		.02		.58		
Cs	11.73					
En			.80	2.56	.36	15.76
Fs			.32	1.36	.32	6.22
Fo	4.86	.57	.60	3.40		.11
Fa	1.57	.60	.24	1.81		.04
Ap	.03	.05	.03	.03	.08	.08
Il	.14	.06	.12	.12	.10	.16
Mt	2.25	.20	.03	.27	.23	1.09
Cc	.02			.02		
Total	100.00	100.00	100.00	100.00	100.00	100.00
Molecular mode calculated from cation percent						
Quartz					1.17	0.42
Plagioclase		An$_{82}$ 95.33	An$_{60}$ 96.20	An$_{63}$ 80.00	An$_{58}$ 96.50	An$_{47}$ 35.50
Orthoclase		1.75	1.65	2.35	.35	2.35
Chlorite		1.81	2.93	5.29		2.17
Hornblende				14.33	1.23	48.17
Kyanite					.44	17.85
Epidote		1.31				
Apatite		.05	.03	.03	.08	
Ilmenite		.06		.12	.10	
Magnetite		.20	.03	.27	.28	.13
Calcite				.02		
Total		100.51	100.84	102.41	100.15	106.59

[1] Faye H. Neuerburg, analyst, U.S. Geol. Survey.
[2] Lois D. Trumbull, analyst, U.S. Geol. Survey.
[3] Ruth H. Stokes, analyst, U.S. Geol. Survey.
[4] Robert N. Eccher, analyst, U.S. Geol. Survey.

hornblende-bearing layers, and their mineral content varies more. In many outcrops of micaceous anorthosite, biotite-bearing layers, 1 to 3 cm thick, and muscovite-rich layers of the same thickness alternate with layers that consist of plagioclase, some biotite, and occasional chlorite. The plagioclase-rich layers are 5 to 20 cm thick and have an appearance similar to that of the feldspathized schist along the border zones of the anorthosite bodies. Only one kind of plagioclase (An_{45}) is found in some layers, but the others contain also bytownite. The andesine in all layers occurs as large rounded grains. Biotite, chlorite, kyanite, and andalusite are interstitial. Some tourmaline and apatite occur as accessories. The mica-rich layers contain abundant large flakes or segregations of large flakes of either biotite or muscovite, or of both. Along the Little North Fork of the Clearwater River opposite Boehls, a layer of biotite about 30 cm or more in thickness is exposed near the contact of anorthosite with feldspathized schist. This layer consists of about 90 percent biotite with only a few white grains and small clusters of plagioclase embedded between the dark large flakes. The biotite is greenish, and the index of refraction, $\gamma = 1.623 \pm 0.001$, indicates that this biotite is richer in iron than the greenish biotite in the cordierite gneiss nearby (Hietanen, 1956, table 2).

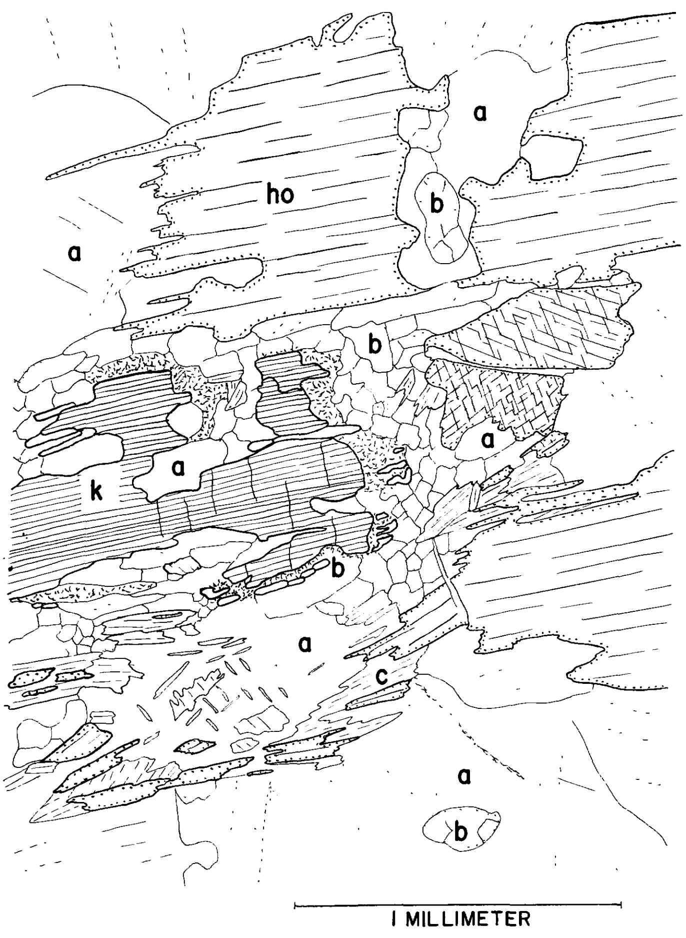

FIGURE 29.—A camera lucida drawing of a kyanite- and hornblende-bearing layer in anorthosite. Small grains of bytownite (b) occur between the kyanite (k) and the hornblende (ho). a, andesine; c, chlorite. Location 608, half a mile southwest of Larson Cabin.

Along the North Fork of the Clearwater River west of Larson Cabin, the dark hornblende-bearing layers, ranging from 2 to 10 cm in thickness, alternate with thicker layers that consist of coarse-grained white anorthosite. Differential weathering is typical of these outcrops, the hornblende-bearing layers being less resistant. The structure is more gneissic than that in the coarse-grained pure anorthosite, owing to the orientation of hornblende crystals subparallel to the planar structure.

An extraordinary petrologic feature of these dark layers is the occurrence together of kyanite and hornblende (fig. 29). The kyanite occurs in small bluish prisms, 1 to 2 cm, long showing $\alpha = 1.713 \pm 0.001$, $\beta = 1.722 \pm 0.001$, $\gamma = 1.728 \pm 0.001$, the indices commonly found for kyanite. The hornblende is dark green in hand specimen but very light greenish in thin section. It shows a weak pleochroism X=yellowish, Y=light green, Z=light bluish green. The indices of refraction are $\alpha = 1.640 \pm 0.001$, $\beta = 1.655 \pm 0.001$, $\gamma = 1.669 \pm 0.001$, $2V = 90°$, and $Z \wedge c = 15°$. The hornblende was separated by means of heavy liquids from specimen no. 608. A fraction having the specific gravity of 3.16 ± 0.02 and containing only a few tiny grains of magnetite and quartz was analyzed chemically (table 12). The result shows that this hornblende is of a common variety in which the percentages of iron, magnesium, and calcium are practically equal. The formula can be expressed as follows:

$$H_{2.17}(Ca_{1.70}Na_{0.43}K_{0.06})_{\Sigma=2.19}\\ (Mg_{2.55}Fe_{1.28}Mn_{0.02}Ti_{0.03}P_{0.01})_{\Sigma=3.89}\\ (Al_{1.04}Fe_{0.22})_{\Sigma=1.26}(Si_{6.3}Al_{1.7})_{\Sigma=8.0}O_{24}.$$

The percentage of Al_2O_3 is fairly high, as would be expected in a rock rich in aluminum.

Kyanite and hornblende are usually considered to be incompatible. Garnet or cordierite or both should appear instead of hornblende. In the dark layers under discussion the relations between the hornblende and kyanite fail to give definite evidence as to whether or not these two minerals crystallized in a stable equilibrium. Most of the hornblende is fairly fresh, although some alteration to chlorite was observed; and some of the kyanite crystals are in part rimmed by sericite. Both of these alterations however are common in the anorthosite and are due to a late hydrothermal activity. Two types of plagioclase, andesine (An_{43}) and bytownite

TABLE 12.—*Chemical analysis of hornblende (no. 608a) in anorthosite along the North Fork of Clearwater River one-fourth mile southwest of Larson Cabin* (loc. 608)

[Analyst, Robert N. Eccher, U.S. Geol. Survey]

Constituents	Weight percent	Molecular equivalent	Number of atoms		
SiO_2	43.70	7,276	Si	6.30	8.00
Al_2O_3	16.11	1,580	Al	1.70	
Fe_2O_3	2.02	126	Al	1.04	1.26
FeO	10.60	1,475	Fe^{+3}	.22	
MnO	.15	21	Mg	2.55	3.89
MgO	11.86	2,941	Fe^{+2}	1.28	
CaO	11.02	1,965	Mn	.02	
Na_2O	1.53	247	Ti	.03	
K_2O	.35	37	P	.01	
TiO_2	.27	34	Ca	1.70	2.19
P_2O_5	.07	5	Na	.43	
H_2O+	2.26	1,254	K	.06	
H_2O-	.10	--------	H	2.17	
			O	24.0	
	100.04	--------			

Sp gr 3.16±0.02

(An_{85}), also occur in these dark layers. The andesine grains are large and fresh, but the bytownite grains are small and have some cracks that are filled by chlorite and sericite. Commonly the bytownite occurs next to the kyanite. A few small grains of andalusite, biotite, rutile, and ilmenite occur as additional constituents.

A chemical analysis (table 11, no. 608) shows that these dark layers are considerably richer in iron and magnesium than the light-colored layers, but the amounts of aluminum, calcium, and sodium are smaller.

MASSIVE ANORTHOSITE

Coarse-grained massive anorthosite occurs locally east of Larson Cabin and south of Boehls. It resembles the light-colored layers in the layered variety and may be thicker layers of this light type. Along the North Fork of Clearwater River and south of Boehls the anorthosite is white or bluish gray. On the south side of the Little North Fork of Clearwater River just east of Boehls, a light brownish variety with exceptionally large plagioclase grains is exposed. The individual grains of plagioclase in this rock are as much as 3 cm in diameter; elsewhere they range only from 1 to 5 mm in diameter.

Thin sections show that two types of plagioclase occur also in the massive anorthosite; the amount of bytownite, however, is less than it is in the layered part of the rock. The outcrop of brownish very coarse grained massive anorthosite east of Boehls consists of labradorite with bluish-gray hornblende and some chlorite, biotite, and epidote. The labradorite grains are rounded, have interlocking borders, and contain numerous tiny platelike inclusions of more calcic plagioclase. As in the other varieties, all bytownite inclusions in each individual labradorite grain show a parallel extinction whose direction differs from that of the host mineral. The hornblende occurs in clusters of small elongated grains which in places have their longest dimensions perpendicular to the boundaries of the labradorite grains. Tiny biotite flakes are as scattered inclusions in labradorite. Chlorite is usually adjacent to the hornblende. The amount of dark constituents, hornblende and biotite ranges from 1 to 5 percent.

Along the North Fork of the Clearwater River, the most common dark constituent is biotite, which occurs in small lenticular clusters of small grains and constitutes 2 to 5 percent of the rock. These clusters are scattered throughout the rock and their longest dimension is parallel to the planar structure in the layered rock nearby. Near Boehls the most common dark constituent is hornblende, which shows a tendency to cluster into irregularly shaped dark spots. The additional constituents are the same as those that occur in the light-colored layers. A few grains of pyrrhotite surrounded by magnetite were found in some thin sections studied.

FINE-GRAINED GRANOBLASTIC VARIETY

The outcrops along the Little North Fork of the Clearwater River, from half a mile to a mile south of Boehls, consist of heterogeneous fine-grained granular white to light-gray to greenish rock, which locally grades into layered hornblende-bearing or micaceous gneiss and in which there is only one kind of plagioclase (An_{55} or An_{85}). The white portions of this rock are homogeneous, but the light-gray and greenish parts contain schlieren and spots of gray or greenish material. Thin-section study shows that the white portions consist of pure plagioclase and that the darker portions contain biotite or epidote, or, more rarely, hornblende. The texture is granoblastic, with plagioclase (An_{54-55}) occurring in polygonal grains, some of which show a weak reversed zoning. In the greenish parts of the rock the epidote is found in small grains between the plagioclase grains. On the west side of the Little North Fork south of Boehls, many fine-grained patches consist of a myrmekite-like intergrowth of bytownite (An_{85}) and epidote (as at loc. 632). Tiny grains of epidote and small hornblende prisms are scattered throughout this anorthosite, the outcrop of which contains shadowy remnants of diopside-plagioclase gneiss. In the light-gray anorthosite south of this locality, biotite occurs in varying quantities and forms small dark-gray spots and schlieren in a white to light-gray plagioclase (An_{55}) rock (fig. 30). The schlieren are only from 10 to 50 cm long and from 1 to 5 mm thick, and there is every transition from the biotite-plagioclase quartz rock in the center of the schlieren to about 99 percent plagioclase in the white portions of the outcrop.

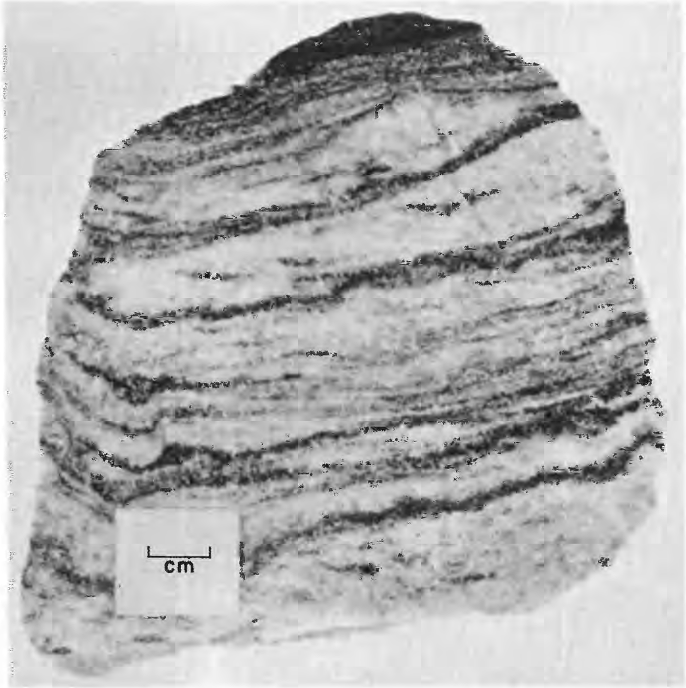

FIGURE 30.—Photomicrograph of dark biotite-rich schlieren in white granular anorthositic rock about a mile south of Boehls (loc. 1299). The dark layers consist of plagioclase, biotite, and quartz in about equal amounts. The white layers consist of 98 percent plagioclase (An_{45}) with some biotite and some small round grains of quartz included in plagioclase.

FIGURE 31.—Photomicrograph of phlogopite-andesine rock from the border zone of the anorthosite south of Boehls (loc. 630). phl, phlogopite; a, andesine. Crossed nicols.

A thick layer of white pure labradorite (An_{55})anorthosite with a sugary granular texture occurs just north of the granular anorthosite with ghostlike remnants of diopside-plagioclase gneiss. Only a few small quartz grains were found in this white portion of the rock. Farther to the north a plagioclase-rich gneiss, in which muscovite, phlogopite, and hornblende-bearing layers alternate, is included in anorthosite (fig. 31). Megascopically these layers resemble the gneissic and micaceous layers in the Belt series but contain far more plagioclase and less quartz. Moreover, no distinct bedding but only schlieren and ghostlike remnants of schistose and gneissic rocks occur in the anorthositic rock. All thin sections made of these layers of granoblastic plagioclase rock contain only one kind of plagioclase, mostly An_{55} but some An_{85}. These rocks are comparable with the bytownite rocks described in the following section.

ANORTHOSITE ALONG CEDAR CREEK

About 70 percent of the anorthosite in the middlemost large body that is exposed along Cedar Creek shows a layering or a banding, but the outcrops near the forks of Cedar Creek and locally in the north-central part of the body are massive and coarse grained. A very fine grained variety consisting of almost pure bytownite is exposed along the tributaries of the West Fork of Cedar Creek. Blocks of similar bytownite rock were found west of Cedar Creek along the road leading from Boehls to Goat Mountain.

LAYERED ANORTHOSITE

In the northern part of the body exposed along the tributaries of Cedar Creek, most of the anorthosite is fine to medium grained, white or light bluish gray, and contains dark streaks or thin layers rich in hornblende or rich in hornblende, chlorite, and biotite. These hornblende-bearing layers are 1 to 2 mm thick and discontinuous. Study under the microscope shows that they consist of andesine, bytownite, hornblende, and chlorite. In some specimens many long grains, or groups of grains of epidote, are included in chlorite parallel to its cleavage. Small grains of sphene, apatite, and magnetite occur as accessories.

The light-colored layers consist of andesine and bytownite. The andesine occurs as larger grains, most of which are elongated parallel to the layering; the bytownite occurs as small grains between the thin andesine layers and also forms platelike inclusions in andesine. On the weathered surface this thin layering is visible as fine striations because of the differential weathering of andesine and bytownite. This striation is similar to that found in the hornblende-bearing layers near Boehls. The layers consisting of bytownite

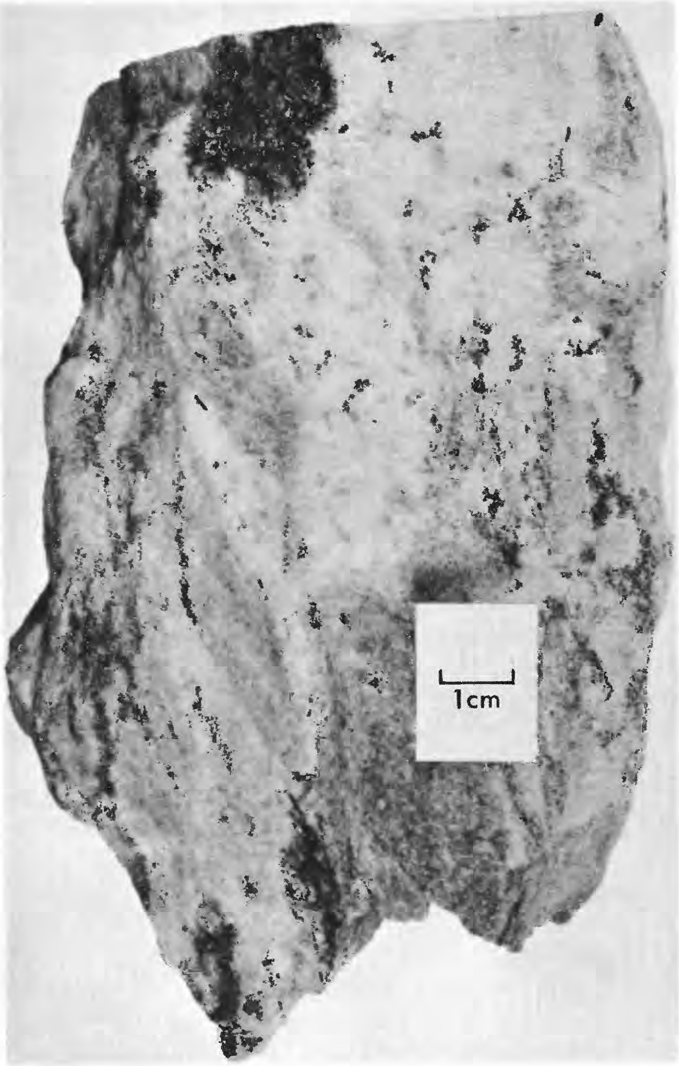

FIGURE 32.—Dark-colored remnants of biotite schist in granoblastic fine-grained border zone of anorthosite along the West Branch of Cedar Creek (loc. 1759).

are usually fractured and weather more readily than those consisting of andesine. In thin sections, andesine appears much fresher and shows no alteration, whereas many of the bytownite grains have sericite and clay minerals along their fractures.

The border zone of the tonguelike extension of anorthosite along the West Branch of Cedar Creek consists of granular andesine rock with irregular biotite-rich patches and schlieren (fig. 32). The country rock in this locality is a biotite schist with fine-grained biotite quartzite layers.

The anorthosite along Cedar Creek, about 1 mile northwest of the forks of the East Branch and West Branch, consists of a medium- to coarse-grained layered anorthosite in which the constituent minerals are the same as those in the northern part, but in which there is more biotite and less hornblende. In many outcrops the major part of the biotite occurs in thin laminae that separate the thicker layers consisting of light bluish-gray plagioclases. The thickness of the plagioclase layers ranges from 2 to 4 cm, and that of the biotite layers from ½ to 5 mm.

MASSIVE VARIETY

About half a mile northwest of the forks of the West and East Branches of Cedar Creek the layered anorthosite grades to a coarse-grained variety that is rich in hornblende and contains small dark clusters of biotite elongated parallel to the planar structure. Toward the south this variety grades over to a massive hornblende- and biotite-free light-greenish or grayish coarse-grained anorthosite in which epidote and some chlorite, ilmenite, and sphene are the only additional constituents. A similar coarse-grained massive anorthosite is exposed north of the layered anorthosite. Two types of plagioclase, andesine and bytownite, are the major constituents of this massive variety and the texture is similar to that in many other occurrences of coarse-grained anorthosite, with andesine occurring in large rounded grains and bytownite as small grains between the andesine grains and as platelike inclusions in andesine. In one of the thin sections studied, a few small andesine grains showed a weak reversed zoning. A few tiny orthoclase grains were found in another thin section.

South of the light-colored anorthosite, the amount of hornblende increases again, and the massive, very coarse grained anorthosite at the forks of Cedar Creek contains dark patches rich in hornblende (fig. 33). Andesine (An_{43}) grains in this rock are as much as 10 cm in diameter and include small grains of hornblende. A thin-banded anorthosite is found south of the coarse-grained variety, and farther south layers of this thin-banded anorthosite, 5 to 10 cm thick, alternate with andesine-rich schist. Some orthoclase is included in andesine in the banded anorthosite.

BYTOWNITE ANORTHOSITE

Blocks of very fine grained anorthosite along the road near the northern border of the anorthosite body along Cedar Creek consist of almost pure bytownite (An_{83}). Bytownite occurs in small twinned grains that are elongated parallel to a beddinglike structure. Some small grains of epidote, sphene, hornblende, and chlorite are as additional constituents. Chemical analysis (table 11, no. 949) shows that this fine-grained bytownite rock is richer in calcium and aluminum than the normal anorthosite. About half a mile east of this locality (loc. 949, pl. 1) a similar fine-grained bluish bytownite rock is exposed along a logging road between a white layered anorthosite and a coarse-grained massive variety. Toward the north, the massive

FIGURE 33.—Massive coarse-grained anorthosite at the forks of West Branch and East Branch of Cedar Creek. The dark constituent is mainly hornblende (loc. 1418).

variety contains patches of biotite and grades over to a biotite-plagioclase gneiss. Another layer of fine-grained bluish bytownite rock is found in this plagioclase-rich gneiss, which toward the north grades over to a thin-banded hornblende-bearing anorthosite. White anorthosite with thin discontinuous layers of biotite and muscovite occurs north of the hornblende-bearing variety and still farther north a feldspathized schist is found.

The measured mode of the bytownite rock exposed on the logging road shows 96 percent bytownite, 2 percent orthoclase, and 2 percent combined diopside and hornblende. A few small grains of quartz, chlorite, and magnetite also occur.

Blocks of hornblende-rich bytownite anorthosite were found along the road leading from Stocking Meadows to Goat Mountain (loc. 947) just north of a small occurrence of amphibolite. The hornblende in this anorthosite occurs as anhedral to subhedral blocky prisms that are evenly scattered throuhgout the rock and oriented parallel to a linear structure in the plane of foliation. Plagioclase (An_{80}) grains have sutured borders and contain numerous small round to irregularly shaped inclusions of quartz. Large quartz grains occur among the plagioclase grains. A considerable amount (about 1 percent) of sphene occurs in round clusters of small grains. Some ilmenite is included in a few of these clusters. Epidote and apatite are the additional constituents. The mineralogy of this gabbroic anorthosite is between that of amphibolite and the bytownite rock. Small blocks of a similar hornblende-rich rock were found 2½ miles east of this locality between the forks of Cedar Creek near locality 1753.

ANORTHOSITE AT GOAT MOUNTAIN AND VICINITY

The third large body of anorthosite, which occurs near Goat Mountain, mainly repeats the varieties found in the body exposed along Cedar Creek. However, the amount of the coarse-grained massive variety is much larger (about 60 percent), and a garnetiferous fine- or medium-grained variety was found along the southwestern contact. The center and the northwestern part of the body consist mainly of light-grayish massive coarse-grained hornblende-bearing andesine-bytownite rock, whereas the border zones have a gneissic structure or show a fine layering similar to the layering found locally in the northern part of the body along Cedar Creek. However, the banded and layered structure occurs in some localities in the center of the body.

Differential weathering, which is pronounced in many outcrops and boulders, shows the distribution of bytownite either into thin laminae or into interstitial areas between the large less weathered andesine or labradorite grains.

MASSIVE VARIETY

The coarse-grained anorthosite near Goat Mountain, west of Monumental Buttes, and near tributaries of Lund Creek, is light bluish gray in fresh outcrops and white on weathered surface. Most of this rock consists almost entirely of plagioclase, the amount of dark constituents being less than 3 percent (fig. 34).

Near Lund Creek the anorthosite is coarse to medium grained and consists of about 98 percent labradorite (An_{57}) with only minute inclusions of bytownite. Tiny grains of epidote and chlorite with or without hornblende and sparse small grains of plagioclase occur between the labradorite grains which range from ½ to 3 cm in diameter. A few small grains of ilmenite surrounded by sphene are the accessories. Two small grains of myrmekitelike intergrowth were found in one of the thin sections studied and some small grains of orthoclase in another.

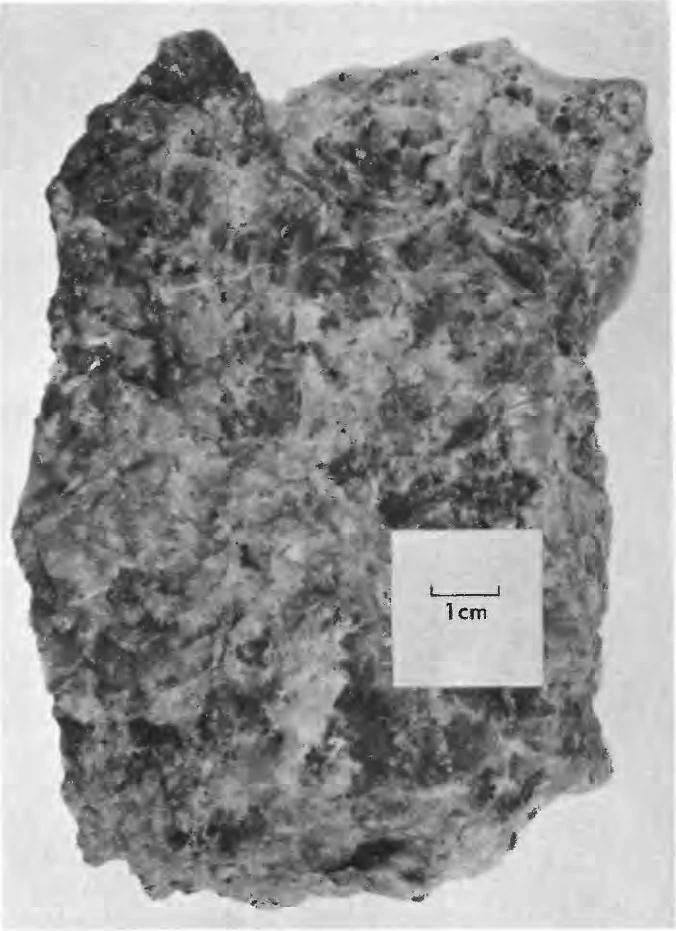

FIGURE 34.—A specimen of massive anorthosite near Lund Creek. About 98 percent of this rock is labradorite (An_{57}).

Near Monumental Buttes hornblende and chlorite are the common dark constituents, but in the majority of the outcrops the total amount of the dark constituents is less than 3 molecular percent (table 11, no. 1016). The amount of bytownite (An_{80-85}) is much larger than that near Lund Creek. The host plagioclase is more sodic, consisting of An_{47-55} as determined by the immersion method. The texture is similar to that shown in figure 27.

The exceptionally large size of the bytownite inclusions in specimen no. 1016 made possible the measurement of the optic orientation and the cleavage of the host andesine, its twinning lamellae, and the bytownite inclusions. The result of two such measurements is shown in figure 35. The pole of the basal pinacoid of the bytownite (An_{80}) inclusions, $(001)_3$ in diagram *a* of figure 35, makes an angle of 17° with the pole of the basal pinacoid of the host $(001)_{1,2}$, in the same diagram. The host consists of andesine (An_{45}) and is twinned according to the pericline law.

In diagram b of figure 35 cleavage (010) is common for the host and the inclusions (marked $(010)_{1,3}$), and the pole of the basal pinacoid of the host $(001)_{1,2}$, makes an angle of 16° with the pole of the basal pinacoid of the inclusions $(001)_3$. Both the andesine host and the bytownite inclusions are twinned according to the pericline law. In addition to the pericline twinning the host shows the albite twinning. Many of the twinning lamellae on the (001), face, or nearly so, seem to contine from the host to the inclusions (fig. 36) where they either turn parallel to the $(001)_3$ plane of the inclusion, or stop in the bytownite at a short distance from the contact of the two plagioclases. Some twinning lamellae go straight through a few small inclusions of bytownite. In some other grains measured, the poles of the (001) face of the host and the inclusions make an angle of 15° to 16°.

The tiny rounded inclusions in the larger bytownite inclusions and the groups of small bytownite grains (4 in fig. 36) included in andesine are oriented at random. The texture of the groups of small bytownite grains is similar to that of the bytownite which occurs between the large andesine grains. This similarity in texture suggests a common origin for all bytownite. The tiny inclusions and the irregularly shaped extensions from bytownite (3) into these groups of small grains (bytownite (4) on lower left corner of fig. 36) suggest that bytownite (4) represents an earlier phase that was, in part, replaced by bytownite (3). Thus, the bytownite (3) seems to represent recrystallized remnants of the fine-grained bytownite rock. The optic orientation of the bytownite (3) and the andesine suggest that the bytownite was reoriented and recrystallized during the formation of andesine.

Locally the massive anorthosite contains more hornblende than that described in the earlier paragraphs; the percentage of the dark constituents range from 3 to 20 percent. Hornblende and chlorite occur in clusters and form irregularly shaped dark patches. Under the microscope the hornblende is seen as subhedral prisms which are very light greenish with pleochroism X= colorless, Y=light green, Z=light bluish green; thus it is much like the hornblende in specimen no. 608 (table 12). The indices of refraction are $\alpha=1.646\pm0.001$, $\beta=1.657\pm0.001$, $\gamma=1.670\pm0.001$, $Z\wedge c=17°$, and 2V about 90°. Chlorite is light greenish with abnormal blue interference colors. Some epidote is included in chlorite. Sphene and ilmenite occur as accessories.

Chemical analysis of the hornblende-bearing anorthosite (table 11, no. 1015) shows that the composition is close to that of the light-colored variety except for some more iron and magnesium.

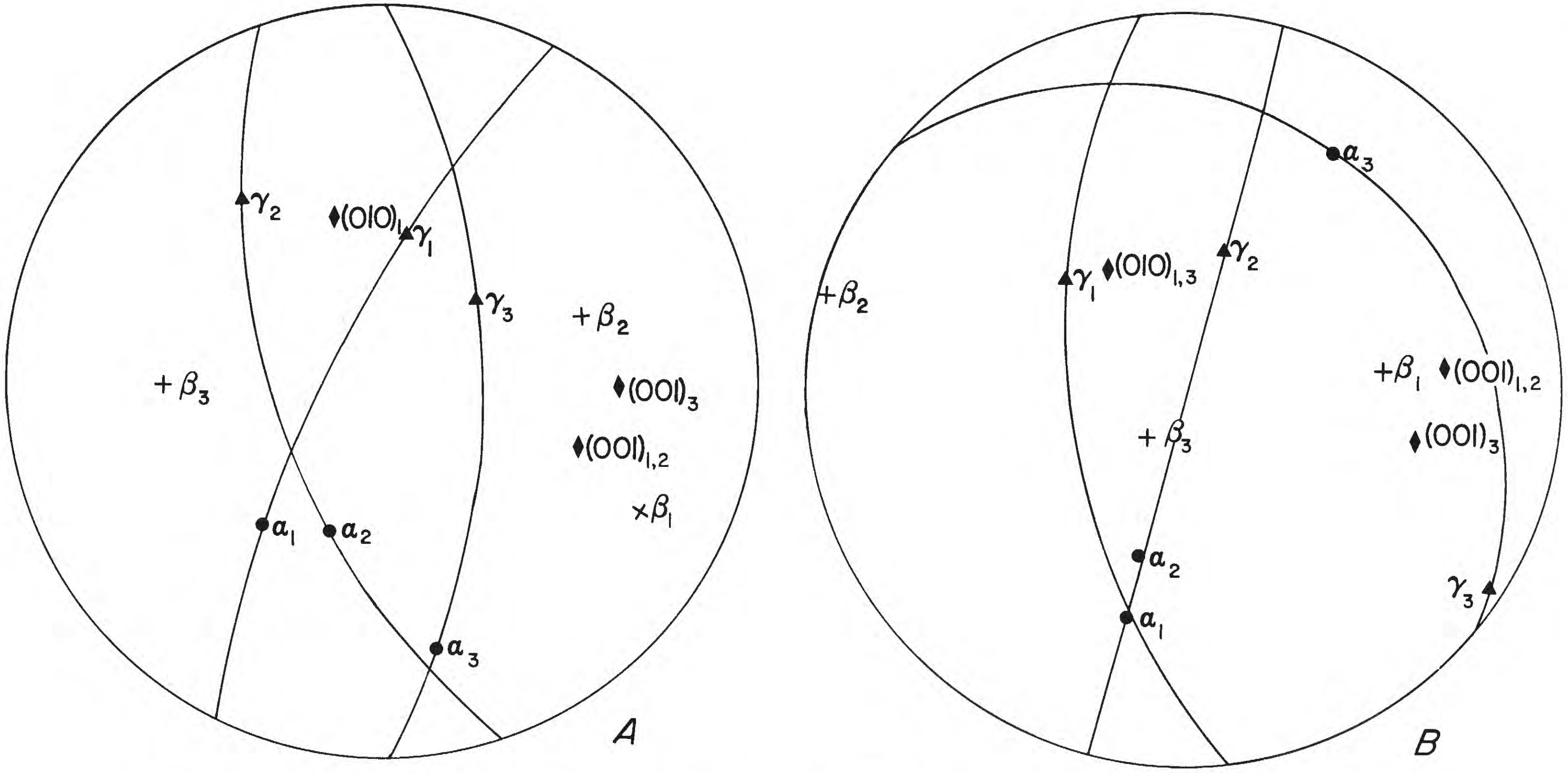

FIGURE 35.—Orientation of the inclusions of bytownite (3) in andesine (1) that shows pericline twinning lamellae (2). Subscripts refer to these three units. Where two subscripts appear, the face is common for two units; for example, $(001)_{1,2}$ is basal pinacoid for the host andesine (1) and its twinning lamellae (2). A and B are measurements of two separate andesine grains from the same thin section. Massive anorthosite southwest of Monumental Buttes (loc. 1016).

FIGURE 36.—Camera lucida drawing of bytownite inclusions (3) in the same andesine grain (1), the orientation of which is shown in figure 35. The pericline twinning lamellae are indicated by (2). The tiny inclusions of bytownite (4) are oriented at random. Massive anorthosite southwest of Monumental Buttes (loc. 1016).

HORNBLENDE-BEARING BANDED VARIETY

The part of the northernmost large anorthosite body that extends southward from Goat Mountain consists of banded plagioclase-hornblende rock, which resembles the thin-layered anorthosite along Cedar Creek. It is medium grained, has a gneissic structure, and contains from 2 to 20 percent hornblende. The hornblende prisms occur in paper-thin discontinuous laminae between layers of plagioclase 1 to 2 cm thick. A similar anorthosite occurs in many other localities; for example, near the garnet amphibolite body between Goat Mountain and Monumental Buttes, and locally along the borders and also in the center of the body. In localities 1021 and 1363 the banded anorthosite grades over to a dark gneissic variety that contains from 10 to 30 percent hornblende. Two types of plagioclase, andesine and bytownite, occur in both varieties, and the texture is similar to that of the thin-banded anorthosite exposed along the Little North Fork of the Clearwater River and along Cedar Creek. In many outcrops in the southern part of the body south of Goat Moutain the amount of hornblende varies, giving rise to a coarser banding that resembles differentiation layers in igneous rocks. Long layerlike bodies of amphibolite are included in anorthosite in this part of the body. Only the large ones are marked on the map, the others being from 10 to 50 cm thick. The amphibolite grades over to the anorthosite that contains discontinuous dark layers rich in hornblende. The rock in the two localities mentioned (nos. 1021 and 1363) may represent similar hornblende-rich layers. The occurrence of two kinds of plagioclase brings the mineral content of these layers close to that of the normal anorthosite; however, some quartz is common in them and the hornblende is a darker variety than is commonly found in the anorthosite. In these respects the mineralogy of these dark layers resembles that of the amphibolite.

A small outcrop along the road south of Goat Mountain (loc. 968) contains more hornblende than any of the anorthosite described earlier. It could be classified as a gabbroic anorthosite, but in addition to abundant hornblende it contains a few kyanite crystals and some chlorite, thus being mineralogically similar to the dark kyanite-bearing layers in the anorthosite along the Clearwater River.

Another locality where kyanite occurs in anorthosite rich in hornblende is south of Monumental Buttes just south of locality 1321. No outcrops were found in this locality but several small blocks contain layers, 5 to 10 cm thick, that are rich in hornblende, with $\gamma=1.666\pm0.001$ and $\alpha=1.643\pm0.001$, and contain numerous bluish prisms of kyanite (fig. 37). These prisms range from 1 to 5 cm in length and 1 to 5 mm in thickness and are oriented at random in the plane of foliation. The kyanite prisms have rounded corners, and they are shelled by a thin layer of alteration products. Among these, sericite could be identified, but most of the alteration products consist of fine-grained grayish material that has low indices of refraction and is dark under crossed nicols (probably clay minerals).

An intense wrinkling was found in a banded fine-grained light-colored border zone of the anorthosite west of Pinchot Creek (loc. 1744). In this rock paper-thin bytownite layers alternate with layers, ranging from 1 to 4 mm in thickness, that contain andesine and hornblende (fig. 38).

GARNETIFEROUS BORDER ZONE OF THE ANORTHOSITE

Much of the anorthosite in the northernmost large body near its southern edge east of Orphan Point is fine grained and locally shows a distinct planar structure (fig. 39). The plagioclase in this rock consists of labradorite (An_{52-57}), and the scarce dark constituents are hornblende, chlorite, and epidote. Layers rich in garnet and some other layers containing aluminum

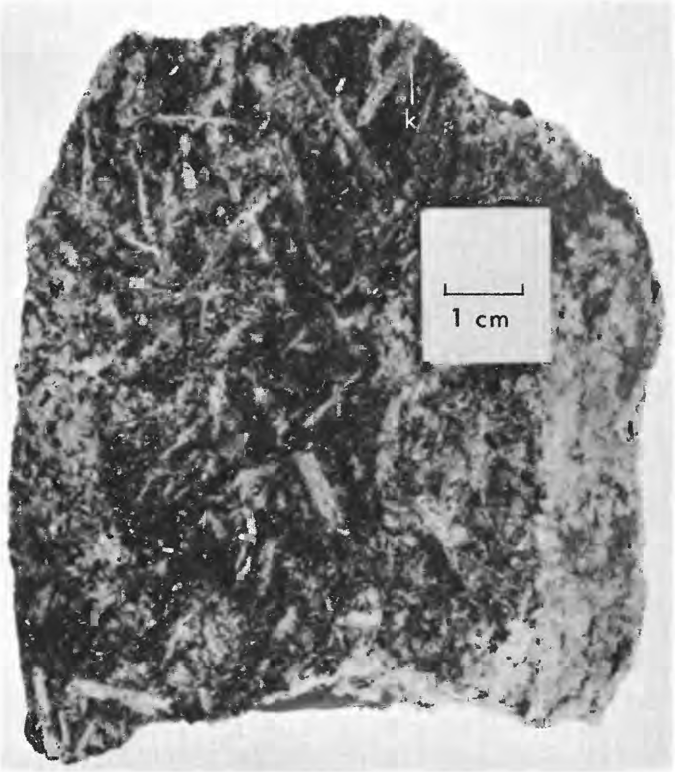

FIGURE 37.—Light-colored prisms of kyanite in a dark hornblende-rich layer in anorthosite south of Monumental Buttes, just south of location 1321.

silicates and biotite are interbedded with this fine-grained anorthositic rock.

Abundant garnet crystals ranging from 1 to 10 mm in diameter were found in the float along the road leading from Goat Mountain to Orphan Point. The blocks in which the garnet occurs consist of light-gray fine- to medium-grained plagioclase-rich rock that contains thin dark-colored micaceous layers. No outcrops were found in this locality, but the structure of the garnetiferous biotite schist and biotite quartzite exposed just west of this locality suggests that the garnetiferous rock occurs on the crest of a gentle anticline. The float of the garnetiferous and fine-grained anorthositic rock is structurally above this schist, and the massive anorthosite along Lund Creek and its tributaries is beneath the same schist. Neither exposures nor float, but light-colored soil suggesting an anorthositic bedrock was found just north of the garnet-bearing anorthositic float in the place where the structural continuation of the schist layer should occur. Several blocks of gray fine-grained plagioclase rock contain pink garnetiferous ellipsoidal lenses shelled by a darker red layer very rich in garnet. In some other blocks discontinuous layers of pink, greenish, and light-gray rock alternate in an irregular manner (fig. 40).

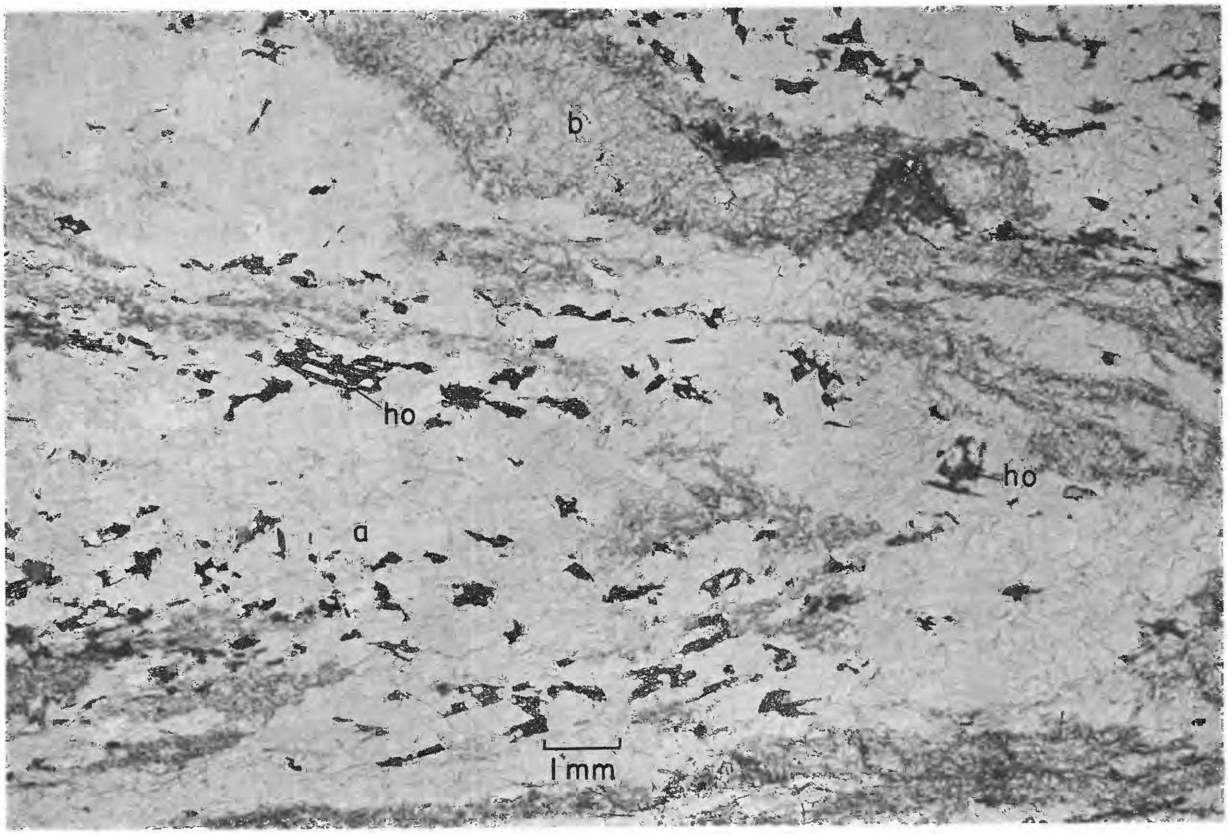

FIGURE 38.—A small fold in fine-grained layered anorthosite along the border south of Orphan Point (west of Pinchot Creek, loc. 1744). The grayish layers are bytownite (b). Hornblende (ho) occurs with andesine (a) in the light-colored layers. Plane-polarized light.

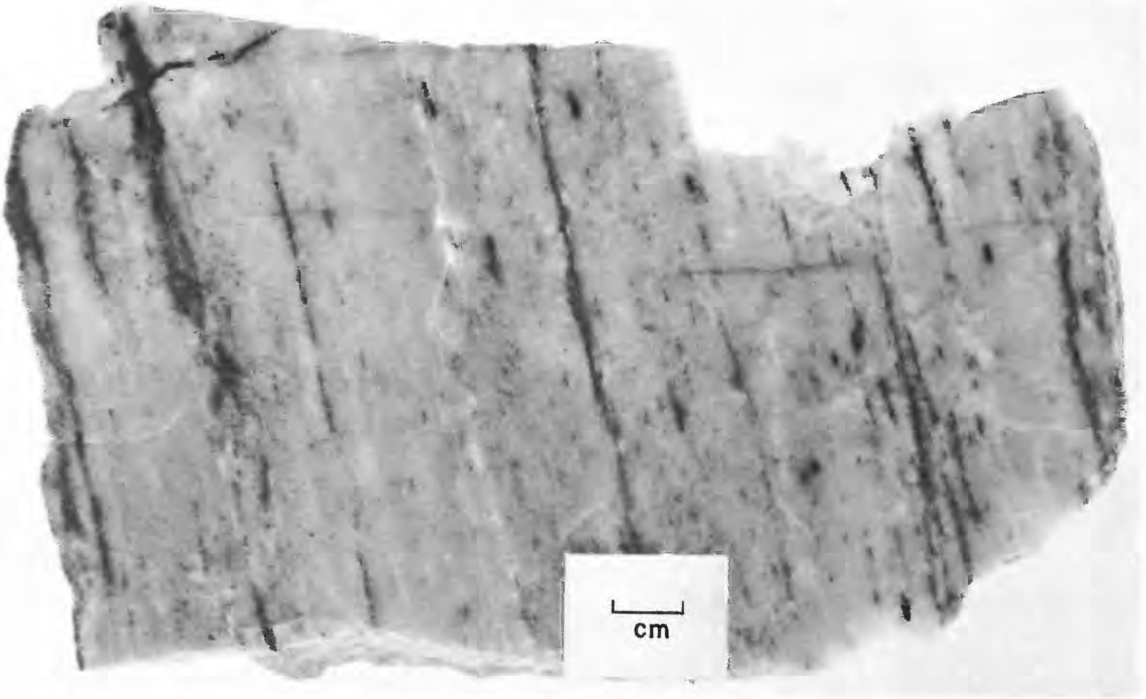

FIGURE 39.—A layered fine-grained anorthosite east of Orphan Point. The dark minerals are hornblende and chlorite; the gray part consists of labradorite (loc. 1353).

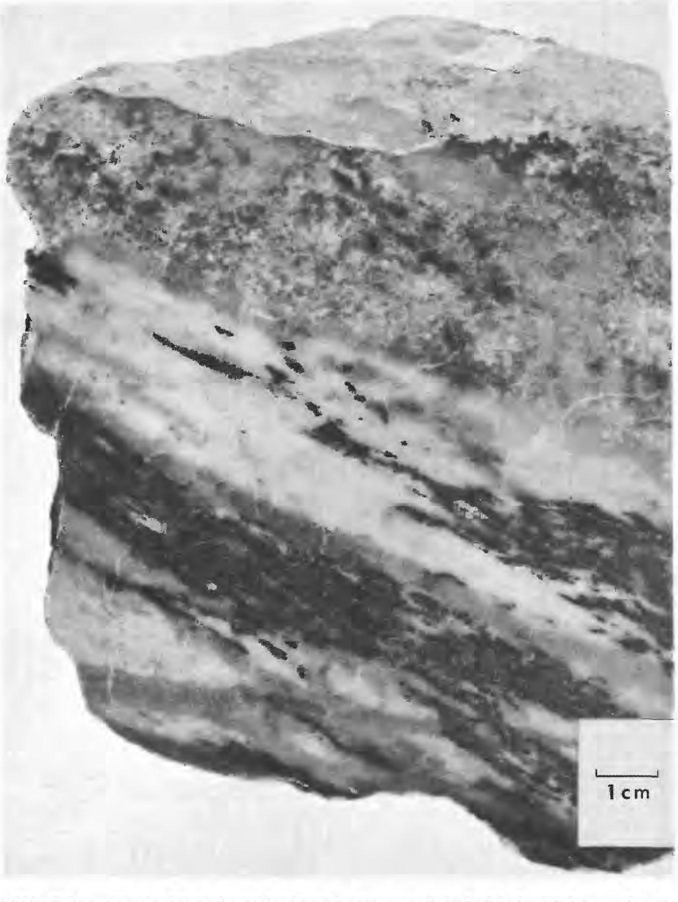

FIGURE 40.—Layers rich in hornblende and garnet are interbedded with layers shown in figure 39. The black portions are rich in hornblende. The dark gray portions are rich in garnet, and the light-colored parts consist of andesine.

Study under the microscope shows that andesine is the major constituent in the light-gray garnetiferous layers, which—in addition to abundant quartz, anadalusite, and biotite—contain some kyanite, sillimanite, muscovite, chlorite, and staurolite as minor constituents. Magnetite, ilmenite, apatite, and rutile occur as accessories. The micaceous minerals, small grains of quartz, and the sillimanite commonly surround the more or less oval or rounded grains of andesine and the crystals of garnet, andalusite, and kyanite; but they also occur as inclusions in andesine and garnet. The garnet tends to be euhedral with dodecahedral faces, and includes grains of quartz, biotite, and magnetite. The unit cell length was determined by F. A. Hildebrand and is $a_0=11.611\pm0.005$A. Index of refraction is $n=1.784\pm0.003$ and specific gravity of 3.93 ± 0.02. Chemical analysis (table 4, no. 1353) shows that this garnet consists of about 63.5 percent almandite, 13.5 percent pyrope, and 21.5 percent grossularite, it being similar to the garnet in the garnet-biotite-gedrite gneiss along the Little North Fork of the Clearwater River (table 4, no. 1225). The analyzed material contained some tiny inclusions of quartz. The garnet in garnet amphibolites contains a little more pyrope (22.5 percent) and less grossularite (14.0 percent) (Hietanen, 1962). For comparison the composition of a garnet from reaction rims in the Adirondack anorthosite was calculated from the analysis given by Kemp (1921, p. 41) with the following result: 61.5 percent almandite, 19.3 percent

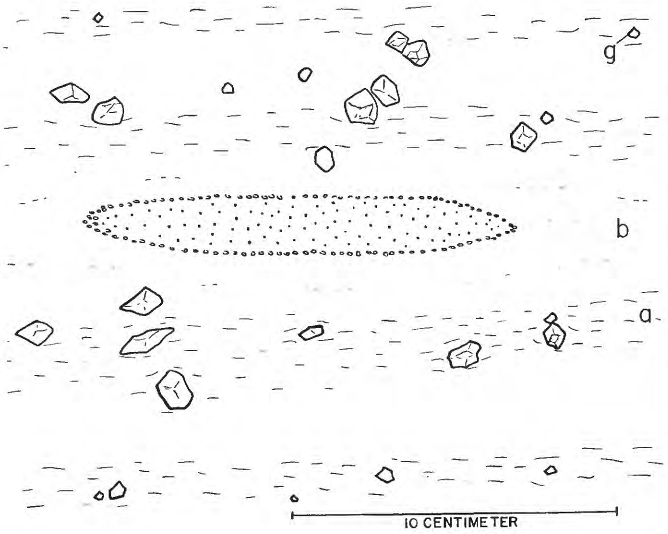

FIGURE 41.—A lens-shaped aggregate consisting of quartz, plagioclase, and garnet in garnetiferous anorthosite about 2 miles east of Orphan Point. *a*, layers consisting mainly of andesine and containing some biotite and bytownite. *b*, layers consisting mainly of bytownite with subordinate amount of andesine. Subhedral garnet crystals (*g*) are scattered throughout both layers.

pyrope, 17.2 percent grossularite, and 2.0 percent andradite. This garnet is chemically closest to the garnet in garnet-biotite-gedrite gneiss (table 4, no. 1225). All these garnets are in their composition close to those that occur in eclogites (compare Eskola, 1921).

The pink ellipsoidal lenses are embedded in a fine-grained gray plagioclase-rich rock that contains sporadic crystals of garnet and abundant quartz. The quartz in this calcic plagioclase rock occurs as elongated grains of medium size or as numerous tiny round inclusions. The sporadic crystals of garnet range from 3 to 10 mm in diameter. The pink lens-shaped aggregates, 2 cm thick and 15 to 20 cm long, resemble, in their mode of occurrence and shape, calcareous concretions (fig. 41) in metasedimentary rocks. Study under the microscope shows that they consist of small grains of calcic plagioclase, quartz, and garnet (fig. 42). The quartz grains are elongated parallel to the long dimension of the lenses and many of them are rimmed by tiny crystals of garnet. Garnet occurs also as numerous small inclusions in plagioclase, making the rock pink. The percentage of quartz in these pink lenses is larger than that in the host rock.

The irregular layering in the pink and green striped plagioclase rock is caused by the variation of the dark constituents. The pink layers contain numerous small garnet crystals, whereas the green layers contain zoisite and hornblende as their dark constituents. Zoisite in elongated grains with rounded edges is the major constituent in many light-gray layers.

The amount of quartz in the garnetiferous plagioclase rock is highly variable. Some layers consist of pure plagioclase, and some others of about 50 percent quartz and 50 percent plagioclase. Plagioclase is more calcic than it is in the anorthosite, ranging from An$_{80}$ to An$_{94}$. Most of the quartz is as individual elongated grains or tiny inclusions in plagioclase, but in some layers thin bands or rows of grains of quartz occur in a manner similar to late vein quartz. There is every gradation from this quartz-rich rock to a normal anorthosite in which quartz is absent and, on the other hand, to a plagioclase-rich schist with aluminum silicates.

ANORTHOSITE IN SMALL BODIES

The anorthosite in most of the small bodies is fine to medium grained and banded, and either micaceous minerals or hornblende occur in it very much in the same way that they occur in the thin-banded variety in the large bodies. In a few outcrops, however, a granoblastic variety with some diopside was found. The diopside-bearing anorthosite is heterogeneous and contains grayish fine-grained lenticular spots surrounded by a somewhat coarser, lighter colored rock in which hornblende grains are clearly visible. The grayish fine-grained part contains small irregularly shaped grains of diopside. The hornblende is darker green than that in a normal anorthosite, and a few grains of orthoclase were found in one of the thin sections studied (no. 1393). This variety is very similar to the fine-grained biotite- and epidote-bearing parts in the anorthosite along the Little North Fork of the Clearwater River about 1 mile south of the mouth of Breakfast Creek (loc. 1299).

The small bodies near the mouth of Pinchot Creek (locs. 1681, 1683) consist of foliated fine-grained anorthosite with clusters of biotite and sillimanite. These clusters are about 1 mm thick and range from ½ to 2 cm in length. In addition to plagioclase (An$_{54}$) this rock contains a few small grains of quartz, some biotite, and sillimanite. Small flakes of biotite with tiny flakes of muscovite and small needles of sillimanite are scattered throughout the rock. The center of the largest body (specimen 1683) consists of pure light bluish gray massive anorthosite. Thin-section study of this anorthosite shows that it consists of irregularly shaped grains of plagioclase (An$_{43}$) that include many small round quartz grains. Epidote, chlorite, magnetite, sphene are sparse.

Another small body of anorthosite between Timber Creek and Pinchot Creek (loc. 1749) consists of fine- to medium-grained finely banded anorthosite which is intensely wrinkled in places. The fine banding is due to the alternation of thin layers consisting either of bytownite or of andesine with some hornblende. The

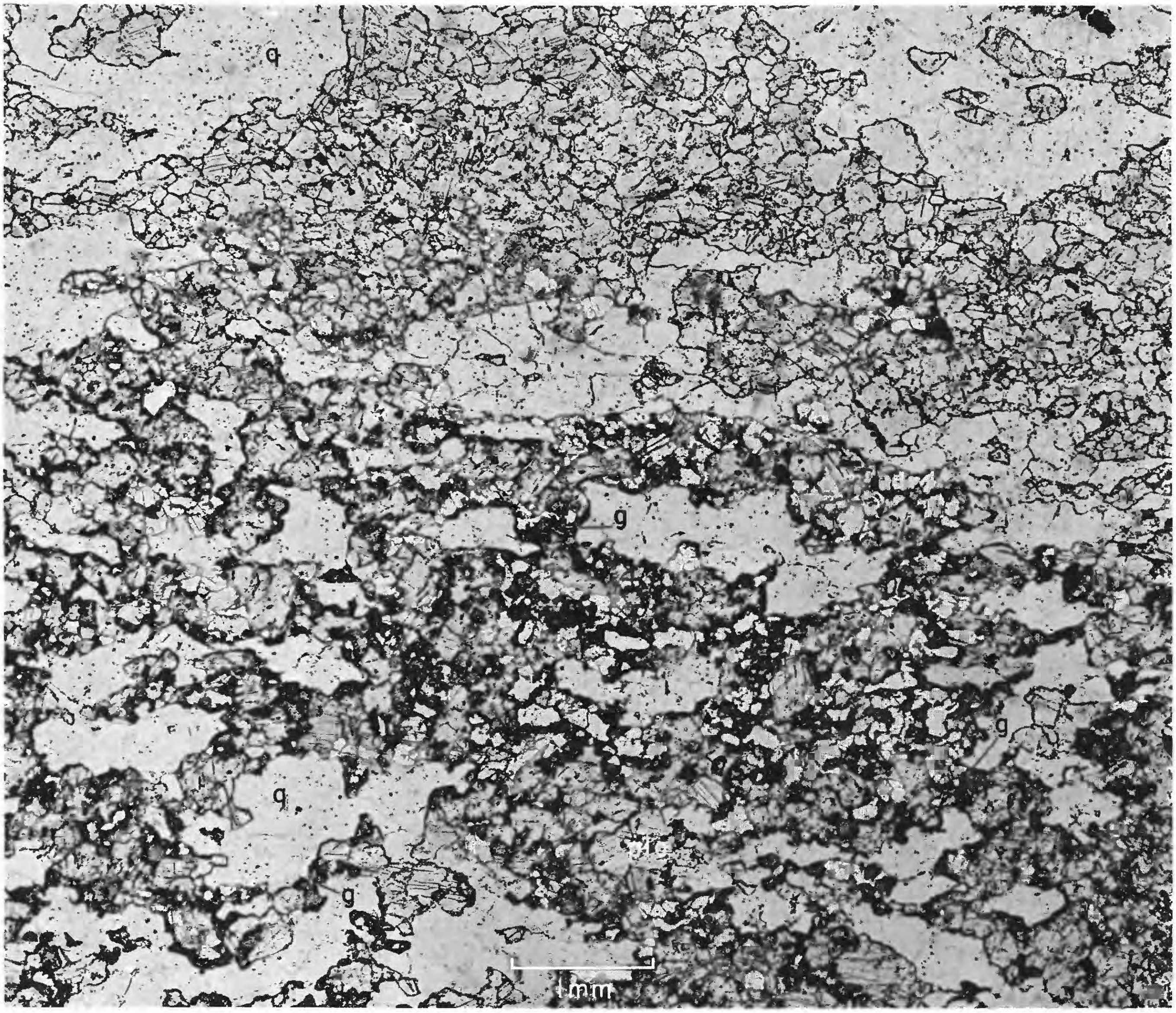

FIGURE 42.—Photomicrograph of a garnetiferous lens in calcic plagioclase-quartz rock. The rock in the upper part of the figure consists of quartz (*q*) and calcic plagioclase (*pla*). A part of an ellipsoidal lens is shown in the lower part of the figure where tiny garnet crystals (*g*) abound. Plane-polarized light.

bytownite layers are fine grained and about 1 mm thick, whereas the andesine layers are medium grained and 2 to 3 mm thick. In places about 25 percent of the rock consists of small prisms of hornblende oriented parallel to the bands.

Blocks of fine-grained garnetiferous pink and green banded anorthosite are strewn along the road south of Goat Mountain (loc. 956). In this rock layering is due to alternation of garnet-bearing layers (pink) with those containing diopside or hornblende (green). Most of the hornblende and diopside occur in separate thin layers, but some hornblende is with diopside. Abundant sphene in elongated grains or groups of grains, with ilmenite cores, is common in diopside-bearing layers. Hornblende-bearing layers contain spongelike plagioclase (An_{94}) grains with numerous round quartz inclusions. Large crystals of garnet include small grains of other minerals. Small garnet crystals occur between the quartz and plagioclase grains rimming many of the quartz grains in the pink layers.

Diopside-bearing anorthosite was also found at locality 1692 near its contact with amphibolite. Andesine in this anorthosite contains numerous tiny, round to oblong inclusions of quartz. Diopside is in clusters of irregularly shaped grains and include small plagioclase grains.

Remnants of a concordant sill-like body of dark hornblende-rich rock that can be classified as gabbroic anorthosite is folded with garnet mica schist on the ridge between Timber Creek and Pinchot Creek (loc. 1747). The schist in this locality is gently folded, the axis of folding plunging 15° east. Small remnants of gneissic plagioclase-hornblende rock, 3 to 7 cm thick, occur in small synclinal notches on the top of the schist on the highest part of an isolated outcrop. The anorthositic rock is gneissic and coarse grained. The mica schist in most parts of this outcrop is much finer grained and shows a lower grade of metamorphism than the schist in the outcrops nearby. However, the schist next to the anorthosite is recrystallized as a dark coarse-grained garnet-hornblende-chlorite rock. The thickness of this recrystallized contact zone is about 6 cm. The garnet crystals in it are 1 to 3 cm in diameter; the hornblende crystals are euhedral, and they decrease in number away from the contact. Small flakes of chlorite fill the interstices between the garnet and hornblende crystals. Quartz, which is a common constituent of the normal schist, is completely lacking. The mineral content indicates that this contact zone is greatly enriched in iron and magnesium and is impoverished in silicon and potassium.

Thin-section study of the gabbroic-anorthositic rock indicates that much of the anorthosite component of the plagioclase is altered to epidote minerals. Hornblende, which is the only dark constituent, is bluish green and contains numerous inclusions of plagioclase. The grains of medium size are clustered; thus the rock appears coarser grained than it really is. Epidote minerals also form clusters of small grains. Plagioclase (An_{60}) grains of medium size separate the hornblende crystals from the epidote clusters. The chemical composition (table 11, no. 1747), mineral content, and contact relations of this body differ strikingly from those of the other anorthosite bodies. It is possible that this thin body crystallized from a normal gabbro magma and that its anorthositic composition was attained by the migration of a part of the iron and magnesium to the country rock to form the garnet-hornblende-chlorite seam along the contact.

CHANGES IN THE COUNTRY ROCKS NEAR THE ANORTHOSITE

INCREASE OF PLAGIOCLASE AND ALUMINUM SILICATES TOWARD THE ANORTHOSITE

The composition and mineral content of the schist and quartzite near the anorthosite differ considerably from those found in the same rocks elsewhere. A gradual change in the amounts of the major constituents—quartz, aluminum silicates, micas, and plagioclase—is well demonstrated in several sections near Monumental Buttes, on the north side of the North Fork of the Clearwater River between Salmon Creek and Boehls Butte, and along Cedar Creek and its tributaries. In all these sections, the average amount of quartz decreases and that of the plagioclase, and commonly also that of the aluminum silicates, increase toward the anorthosite. In detail, the distribution of these various minerals in the schist around the anorthosite bodies is highly irregular, because some of the plagioclase and biotite and also some of the aluminum silicates are segregated to form monomineralic or highly concentrated clusters and masses.

Many good exposures on Monumental Buttes offer excellent material for a detailed study. A section from the anorthosite through schist to quartzite is exposed on the west slope of Monumental Buttes. The anorthosite in the center of the body is coarse grained and massive, but attains a gneissic thin-banded structure near the contact. Some sillimanite, muscovite, and chlorite appear as additional constituents in this banded fairly pure andesine-bytownite rock. The thickness of the banded zone ranges from 50 to 300 m. Abundant aluminum silicates—kyanite, andalusite, and sillimanite—appear in the schist next to the contact. The contact shows signs of stronger shearing than is apparent elsewhere in the formations. The sheared zone exposed is about 10 m thick, and the minerals in it are plagioclase, chlorite, kyanite, andalusite, and sillimanite. The schist outside of this sheared zone contains biotite and muscovite instead of chlorite, and some quartz appears among the andesine. A thick layer of aluninous schist is exposed on the uppermost part of the slope and on the top of the ridge. The trend is parallel to the ridge, and the same layer can be followed for 2 miles. Most of the schist is rich in plagioclase, which occurs in individual large grains and in clusters, but some of the schist contains only garnet and micas in addition to quartz. The change from the plagioclase-rich rock to this normal garnet-mica schist takes place across the bedding, as well as parallel to the bedding.

In detail the distribution of plagioclase in the plagioclase-rich parts also is rather irregular. The plagioclase occurs in individual scattered large round grains, in clusters, and small masses (fig. 43), all unevenly distributed throughout the schist. The roundness and whiteness of the sporadic andesine grains makes them conspicuous against the dark framework of the micaceous minerals. Where their number increases, the rock attains a coarsely granular texture, which is typical of micaceous layers of the anorthosite along the North Fork and the Little North Fork of the Clearwater River. The oblong masses of pure plagioclase in this plagioclase-mica schist range from a few centimeters to 10 m or more in length. These masses have a coarsely granular texture; some have the appearance of pegmatite, but the only constituent mineral in them is plagioclase,

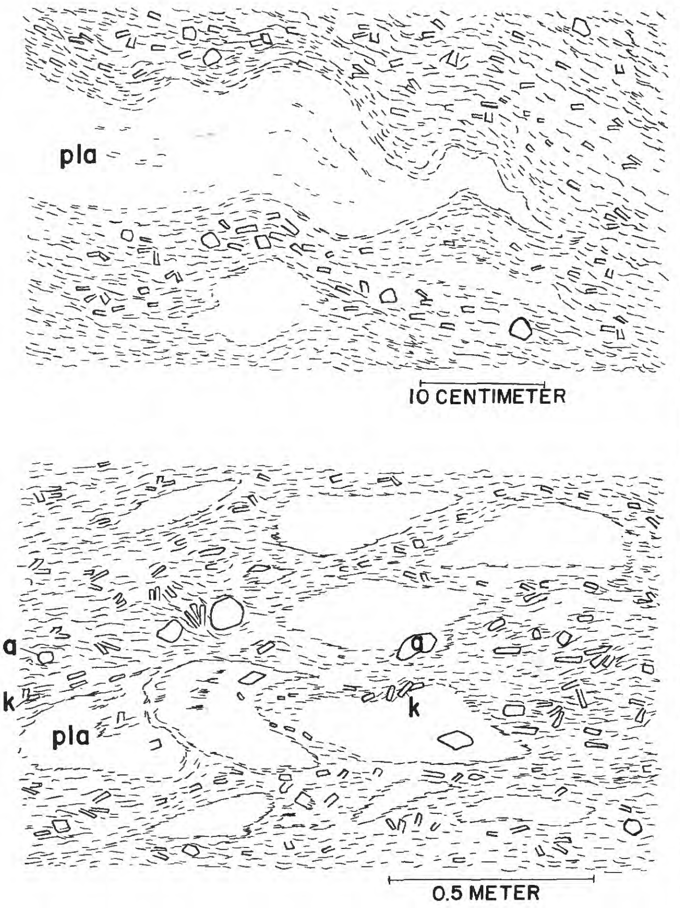

FIGURE 43.—Lens-shaped segregations of andesitic plagioclase (pla) in coarse-grained kyanite-mica schist rich in plagioclase. a, andalusite; k, kyanite. Horizontal surfaces on Monumental Buttes.

ranging in composition from An_{25} to An_{40}. More quartz instead of plagioclase occurs in the schist adjacent to the quartzite.

Much of the contact rock south of Monumental Buttes (south of loc. 1018) consists of fairly coarse-grained plagioclase-biotite rock with large kyanite crystals, as long as 10 cm, evenly distributed throughout the rock. These kyanite-rich layers alternate either with more micaceous layers or with fine-grained gneissic anorthosite that contains biotite. Large garnet crystals abound in some thicker mica-rich layers and also in many outcrops of plagioclase-biotite-kyanite rock; for example, just south of locality 1321, garnet crystals are 5 to 10 cm in diameter and include numerous small kyanite crystals.

The contact zones at Goat Mountain and along the North Fork of the Clearwater River consist of a plagioclase-rich kyanite-andalusite-sillimanite schist similar to that found on Monumental Buttes. Oligoclase-andesine is a major constituent in a zone whose thickness ranges from 50 to 300 m. On the south slope of Goat Mountain, huge kyanite crystals and large sporadic garnets are embedded in coarse-grained plagioclase-biotite rock that occurs next to the anorthosite. The chemical analysis of the plagioclase-biotite "groundmass" (table 13, no. 971) suggests that it is considerably richer in calcium and poorer in silicon than the normal schist. When the total composition of this rock is computed (no. 971a), the abundance of aluminum becomes apparent. The amount of iron is greater where large sporadic garnet crystals appear. Layers rich in biotite and aluminum silicates (table 13, no. 967), and containing small round grains of corundum, are interbedded with the plagioclase-rich schist. The biotite in these layers as well as in the plagioclase-rich parts of the rock is greenish and shows bright interference colors. The index of refraction is $\gamma = 1.610 \pm 0.001$. This mineral was separated and a clean fraction of a specific gravity between 2.809–2.910 was chemically analyzed. The result of chemical analysis (table 14) shows more than 21 percent alumina and a high percentage of magnesium oxide. The formula computed from the analysis is:

$$(H_{4.13}F_{0.01})_{\Sigma=4.14}(K_{1.62}Na_{0.18})_{\Sigma=1.8}(Mg_{3.55}Fe^{+2}_{0.96}Mn^{+2}_{0.02}Ti^{+4}_{0.03}Fe^{+3}_{0.13}Al_{1.05})_{\Sigma=5.74}(Si_{5.41}Al_{2.59})_{\Sigma=8.0}O_{24}.$$

This biotite is very similar to the green biotite in the cordierite gneiss on Smith Ridge (Hietanen, 1956, p. 7–8), showing an excessive amount of aluminum and a deficiency of potassium.

Muscovite abounds locally in the mica-rich layers but is rare in the plagioclase-rich rock. The habit and mode of occurrence of the muscovite in the rocks adjacent to the anorthosite differ from those of the muscovite in the normal schist of the Prichard formation. In the normal schist, muscovite forms large flakes (1 to 4 mm in diameter) that show all properties of normal muscovite, but in the schist adjacent to the anorthosite the muscovite occurs in groups of tiny, soft flakes that have a greasy feeling like talc when pressed between the fingers. The indices of refraction in specimen 971 are higher than those of the normal muscovite, $\beta = 1.594 \pm 0.001$, $\gamma = 1.600 \pm 0.001$. The alkali determination was done by L. N. Tarrant of the U.S. Geological Survey and shows 1.74 percent Na_2O and 8.67 percent K_2O. Quartzite and amphibolite adjacent to the plagioclase-rich schist do not contain secondary plagioclase in any locality studied.

Along the Little North Fork of the Clearwater River, on the north slope of Benton Butte, and along the tributaries of the West Fork of Cedar Creek, the contact rocks are more heterogeneous than those on Monumental Buttes and on Goat Mountain. The schist next to the contact contains abundant oligoclase-andesine, layerlike masses of biotite, and in places segregations of muscovite. Lens-shaped masses

TABLE 13.—*Chemical analyses, norms and modes of the schist next to the anorthosite*

Locality and specimen	630[1]	954[2]	967	971[2]	971a.
Rock type	Plagioclase-mica schist.	Biotite-plagioclase schist with aluminum silicates.	Biotite-plagioclase schist with cordierite and aluminum silicates.	Kyanite-bearing biotite-plagioclase rocks.[3]	Biotite-plagioclase rock 971 recalculated to include the large kyanite-andalusite prisms.
Location	Boehls	2½ miles northeast of Stocking Meadow Lookout.	South slope of Goat Mountain.	South slope of Goat Mountain.	South slope of Goat Mountain.
Weight percent					
SiO_2	51.65	56.00	46.04	56.47	52.55
Al_2O_3	26.42	27.16	29.20	25.44	32.97
Fe_2O_3	.15	.12	.73	.33	.33
FeO	1.89	2.67	4.35	1.44	1.16
MnO	.10	.03	.20	.04	.03
MgO	4.79	1.75	9.00	2.64	2.12
CaO	3.35	4.88	.68	4.12	3.30
Na_2O	4.41	4.34	2.99	6.73	5.40
K_2O	5.10	1.76	3.60	1.47	1.19
TiO_2	.12	.24	.24	.07	.06
P_2O_5	.01	.00	.00	.02	.00
CO_2	.01	.00	.01	.00	.00
H_2O+	1.85	.79	2.33	1.03	.83
H_2O-	.04	.10	.22	.07	.06
Total	99.89	99.84	99.59	99.87	100.00
Cation percent					
SiO_2	46.77	51.37	42.05	50.75	47.19
$AlO_{3/2}$	28.19	29.36	31.42	26.94	34.88
$FeO_{3/2}$	.10	.08	.51	.23	.23
FeO	1.43	2.05	3.32	1.08	.87
MnO	.08	.02	.15	.03	.02
MgO	6.46	2.39	12.24	3.54	2.84
CaO	3.25	4.79	.66	3.97	3.17
$NaO_{1/2}$	7.74	7.71	5.29	11.72	9.39
$KO_{1/2}$	5.88	2.06	4.19	1.68	1.36
TiO_2	.08	.17	.16	.05	.04
$PO_{t/2}$	.01			.01	.01
CO_2	.01		.01		
H_2O	(5.59)	(2.42)	(7.20)	(3.08)	(4.97)
	100.00	100.00	100.00	100.00	100.00
O	148.62	158.95	146.24	154.61	156.94
OH	11.18	4.84	14.19	6.17	4.97
Total anions	159.80	163.79	160.43	160.78	161.91
Molecular norm					
Q		8.23			5.07
Or	29.40	10.30	20.95	8.40	6.80
Ab	27.60	38.55	26.45	58.60	46.95
An	16.10	23.95	3.25	19.75	15.75
Ne	6.66				
C	8.13	10.01	20.64	5.64	17.83
En		4.78	14.92	1.28	5.68
Fs		3.72	3.72	.36	1.46
Fo	9.69		7.17	4.35	
Fa	2.07		1.80	1.14	
Ap	.03			.03	.03
Il	.16	.34	.32	.10	.08
Mt	.15	.12	.76	.35	.35
Cc	.02		.02		
Total	100.01	100.00	100.00	100.00	100.00
Molecular mode calculated from cation percent					
Quartz		8.41			
Plagioclase	An_{29} 54.80	An_{38} 62.50	An_{11} 29.70	An_{25} 78.35	An_{25} 60.59
Orthoclase		1.25			
Biotite	33.20	19.70	46.18	20.00	15.47
Muscovite	23.13				
Kyanite Andalusite Sillimanite		12.27	7.47	5.94	23.92
Cordierite			21.72		
Corundum			3.52		
Apatite	.03			.03	.02
Ilmenite			.20		
Magnetite			.76		
Calcite	.02		.02		
Total	111.18	104.13	109.57	104.32	100.00

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[1] Analyst, Ruth H. Stokes, U.S. Geol. Survey

[2] Analyst, Faye H. Neuerburg, U.S. Geol. Survey

[3] Large kyanite crystals, which constitute estimated 20 percent of the rock, were not included in this analysis.

TABLE 14.—*Chemical analysis of biotite in kyanite-andalusite-plagioclase-biotite rock (no. 971c) on the south slope of Goat Mountain*

[Analyst, L. N. Tarrant, U.S. Geol. Survey]

Constituents	Weight percent	Molecular equivalent	Number of atoms		
SiO_2	37.85	6302	Si	5.41	8.00
Al_2O_3	21.61	2120	Al	2.59	
Fe_2O_3	1.21	76	Al	1.05	5.74
FeO	8.04	1119	Fe^{+3}	.13	
MnO	.13	18	Fe^{+2}	.96	
MgO	16.66	4132	Mn^{+2}	.02	
TiO_2	.32	40	Mg	3.55	
CaO	.00	--------	Ti^{+4}	.03	
Na_2O	.63	102	Na	.18	1.80
K_2O	8.89	944	K	1.62	
F	.03	16	F	.01	4.14
H_2O+	4.33	2403	H	4.13	
H_2O-	.12	--------	O	24.00	
	99.82	--------			
Less O for F	.01	--------			
Total	99.81	--------			

Sp gr=2.809–2.910

of pure andesine in the plagioclase-rich schist are common and many of them are several meters long. The anorthosite adjacent to the contact shows an irregular planar structure due to segregation of micaceous minerals and aluminum silicates into thin discontinuous layers or small lenticular masses.

A contact between fine-grained biotite quartzite and anorthosite is exposed along the West Fork of Cedar Creek. The anorthosite near the contact is fine grained and granular; its texture resembles that of quartzite. Small patches of light-gray to dark-gray rock occur in this white granular anorthosite. Thin sections show that these darker parts are fine grained and contain numerous small flakes of biotite and some epidote. The white portions consist of polygonal grains of andesine, many of which show reversed zoning. In addition, there are a few grains of hornblende and epidote and several large grains of allanite. Biotite-rich schlieren in this rock contain quartz. Elsewhere the quartzite and the lime-silicate rocks preserve their grain size, texture, and mineralogy up to the contact. The layer of bytownite rock that occurs next to the coarse-grained anorthosite along the West Fork of Cedar Creek is fine grained, hard, and unaltered as is also the other layer of bytownite rock that is interbedded with hornblende-biotite schist. No andesine was found in either layer. It seems that the occurrence of micaceous minerals greatly facilitated the development of andesine in the country rocks.

Near Boehls a coarse-grained phlogopite-andesine rock is between the anorthosite and the biotite schist. Chemical analysis (table 13, no. 630) shows that this schist is rich in aluminum and contains a considerable amount of calcium and magnesium.

The schist near the southern contact of the anorthosite along Cedar Creek contains abundant plagioclase and aluminum silicates. A part of this schist is recrystallized as a coarse- to medium-grained homogeneous light-gray plagioclase-biotite rock in which numerous slender blue prisms of kyanite 1 to 5 cm long are oriented at random.

The occurrence of garnet in the anorthosite along the contact between Orphan Point and Goat Mountain probably is a contact phenomenon comparable to the occurrence of aluminum silicates. The garnetiferous ellipsoidal lenticles and the abundance of quartz indicate the presence of abundant sedimentary material in this zone.

Thus, the distribution of plagioclase is not in all contact zones confined to bedding planes but is in places rather irregular, resembling then the feldspathization around metasomatic quartz diorite and tonalite bodies in the Headquarters quadrangle and vicinity (Hietanen, 1962). It seems that at least a part of the plagioclase in the schist near the anorthosite bodies crystallized late as a result of introduction of calcium and sodium and removal of excess silicon. The exchange of elements involved in the mineralogic changes around the anorthosite bodies can be well demonstrated by the chemistry of a cross section through the schist on the east side of Floodwood Creek to the anorthosite at Goat Mountain. Most of the schist along Floodwood Creek is a coarse-grained garnet-mica schist with abundant quartz and only a little oligoclase (compare with table 5, no. 1023). Near the mouth of Timber Creek (loc. 1690) and toward the north (loc. 1751) sillimanite and locally some staurolite appear in the schist. Thin sections show that the sillimanite replaces the biotite. Small grains of oligoclase or andesine occur with quartz. The amount of plagioclase varies but usually stays smaller than that of quartz. In the schist at the mouth of Timber Creek about 20 percent of the light minerals are plagioclase and 30 percent quartz (table 5, no. 1690). Toward the east, the beds rich in plagioclase become more common and the total amount of plagioclase in them increases. The amount of brown sillimanite also increases, and locally some kyanite and colorless needles of sillimanite are present. Near the contact of the anorthosite about 80 percent of the light constituents are plagioclase. Many outcrops in this zone consist of plagioclase-biotite rock with abundant aluminum silicates. Layers rich in micas and others rich in garnet are interbedded.

The gradual changes in mineralogy around the anorthosite bodies take place in an orderly manner and some of the changes indicate an introduction and removal of elements. In the outer zone where biotite is replaced by sillimanite, aluminum was introduced and potassium removed. The first local appearances of andesine in this zone indicate that in addition to aluminum some calcium was introduced. There is a definite increase of the total amount of calcium and aluminum and decrease of silicon and potassium toward the anorthosite. In the zone next to the anorthosite, calcic plagioclase abounds and the amount of quartz is small; the aluminum silicates constitute in places about 40 percent of the rock, with the average being about 15 percent. It is possible that much of the change in the total composition is depositional and that local rearrangements took place during the formation of anorthosite.

GROWTH OF GRAIN SIZE TOWARD THE ANORTHOSITE

The grain size of the minerals in the schist around the anorthosite bodies grows with the increase of plagioclase and aluminum silicates. The average grain size of quartz and biotite in the normal schist of the Prichard formation outside of the contact aureole is 0.01 to 1 mm and that of the garnet 1 to 5 mm. In many localities in the feldspathized schist, the biotite flakes reach a diameter of 1 to 3 mm and the andesine 0.5 to 5 mm. Kyanite crystals, as much as 30 cm long, were found in the schist next to the anorthosite. The grain size of the plagioclase in the anorthosite next to such coarse-grained feldspathized schist is of about the same magnitude as that in the schist, and the texture is coarsely granular. Thus, the growth of the grain size and the increase of plagioclase and aluminum silicates around the anorthosite bodies seem to be closely connected. It seems that the anorthosite bodies were centers of recrystallization and also centers of introduction of elements.

CONTACTS WITH AMPHIBOLITE

The contact of anorthosite with garnet amphibolite is exposed along the road north of Goat Mountain Lookout (loc. 1014). Anorthosite in this vicinity is medium grained, and dikes of similar anorthosite transect the garnet amphibolite. The garnet amphibolite has a good platy and linear structure. Thin sections show that numerous small grains and prisms of epidote and zoisite have crystallized in the anorthosite next to the garnet amphibolite. A few clusters of bluish-green hornblende occur as dark constituents in the anorthosite near the contact as well as farther from it. The garnet amphibolite next to the contact is very similar to the rest of this rock. The contact is sharp and both rocks preserve their grain size near it.

Numerous subhedral zoisite crystals were found in garnet amphibolite next to the anorthosite in a few blocks south of Goat Mountain. On the surface, the zoisite has weathered out, leaving double wedge-shaped cavities. Tiny subhedral garnet crystals have formed along the contact, which is sharp and along which the rock cleaves easily.

Inclusions of amphibolite were found in some large blocks of anorthosite south of Goat Mountain. The contact between these inclusions and the anorthosite is sharp.

The contact between the pyroxene-bearing anorthosite and the amphibolite in locality 1692 is quite different. The amounts of plagioclase and diopside decrease and that of hornblende increases gradually toward the amphibolite over a distance of 2 to 3 cm. Both dark constituents, diopside and hornblende, occur together in the contact zone. The plagioclase in the amphibolite is similar to that in the anorthosite except that the grains in the amphibolite are larger and contain small inclusions of hornblende. No garnet or biotite occurs in this amphibolite. This contact resembles the contact found between the intrusive hornblende gabbro and the inclusions of anorthosite southeast of Goat Mountain. No outcrops were found, but large blocks of hornblende gabbro on the southeast slope of Goat Mountain contain inclusions of fine- to medium-grained anorthosite. The contact is gradational over about 4 cm with the amount of hornblende decreasing toward the anorthosite. Numerous tiny round grains of quartz are included in plagioclase as well as in hornblende. Both of these latter minerals have complex interlocking boundaries resembling sutured texture. A similar relation between the anorthosite and hornblende-rich rock was found in blocks north of the forks of Cedar Creek just south of locality 1753. It seems that these contacts represent a contact between the anorthosite and a younger gabbroic rock. The hornblende-rich portion in the anorthosite 2 miles northeast of Stocking Meadow Lookout (loc. 947) probably represents a similar gradational zone between the anorthosite and hornblende gabbro.

CHEMICAL COMPOSITION OF THE ANORTHOSITE AND ITS COUNTRY ROCKS

The analyses of the anorthosite and of the country rocks were plotted in the QMF diagram (fig. 44) that was prepared in the following way: F is the total amount of feldspars in the molecular norm. Total of iron and magnesium is calculated as orthosilicates, the excess of aluminum oxide and magnetite from the molecular norm is added to this total and the sum used for the M coordinate. Q is free quartz after computation of F and M. The sum of $Q+F+M=100$. In this ternary diagram the plots for the anorthosite and feldspar-rich country rocks are scattered close to the F corner. The rocks of the intrusive quartz monzonite series, which were added for comparison, form a smooth curve near the F corner (nos. 1063, 1073, 1072, 301,

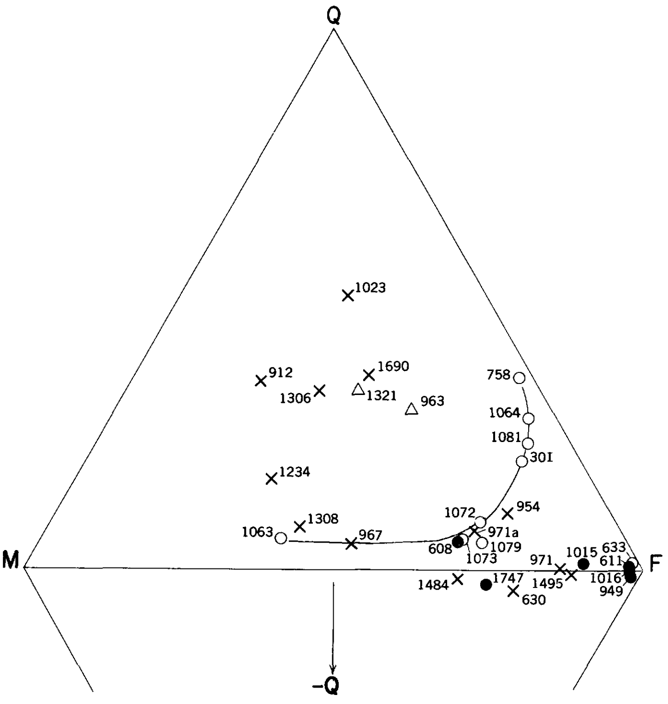

FIGURE 44.—QMF diagram for the anorthosite and its country rocks. (•), anorthosite; (X), country rocks; (Δ), amphibolite; (O), intrusive rocks. F, total amount of feldspars in the molecular norm. M, total of iron and magnesium calculated as orthosilicates to which is added corundum and magnetite from the molecular norm. Q, quartz in the molecular norm after calculation of M. Q+M+F=100. The analyses that are deficient in Q are plotted under the MF line. Numbers 1747, 949, 1016, 1015, 611, and 608 are anorthosites (table 11); nos. 630, 954, 967, 971, and 971a are schists next to the anorthosite (table 13); nos. 1023, 1690, 1234, 1306, and 1308 are rocks of the Prichard formation (table 5); no. 1484 is sericite-margarite schist (table 8); nos. 963 and 1321 are garnet amphibolite (table 10); no. 633 is orthoclase dike (table 9); no. 912 is cordierite gneiss (Hietanen, 1956, table 3); nos. 758, 1064, 1081, 301, 1072, 1073, 1079, and 1063 are rocks of the quartz monzonite series that were added for comparison.

1081, 1064, and 758). The plots for the metasedimentary rocks and amphibolites are scattered around the center of the diagram; these rocks have more iron-magnesium silicates and quartz and less feldspars than the anorthosite and its country rocks and the intrusive series. The plots for the country rocks of the anorthosite fall close to the curve for the igneous rocks and toward the F corner.

The ratio of the normative amounts of the three feldspars—orthoclase, albite, and anorthosite—is shown in figure 45. The plots for most varieties of anorthosite fall around An_{60}, but 80 percent of the normative feldspar in no. 949 and more than 90 percent in no. 1747 is anorthite. The plots for the rocks of the quartz monzonite series are connected with a curve for easier observation. The plots for most of the plagioclase-rich schist around the anorthosite fall close to this curve or toward the Ab corner. The position of the plots shows that the normative plagioclase of the country rocks is more albitic than that of the anorthosite. The normative feldspar in the lime-silicate rocks (nos. 1306 and 1308) is similar to that in the bytownite anorthosite (no. 949). The other anorthosites occupy a field between these lime-silicate rocks and the feldspathized schist that occurs next to the anorthosite.

In order to study the average amount of the major constituents in the anorthosite, its country rocks, and unaltered rocks of the Prichard formation, the averages of the ionic percentages in the rocks concerned were plotted in figure 46. In this figure *a* represents an average of cation percentages of 6 analyses of the anorthosite of table 11 and *c* the same average for 5 country rocks (table 13). These country rocks represent the composition of the schist in a zone as far as half a mile from the anorthosite. Number 1690 is a representative of a feldspathized schist about 1 mile from the anorthosite and no. 1023 is a typical unaltered schist of the Prichard formation. In the normal schist of the Prichard formation (no. 1023), 70 percent of the cations are silicon and 16 percent aluminum. In the feldspathized schist about a mile from the anorthosite the corresponding percentages are about 59 and 25. The average of the ionic percentage of silicon in the schist next to the anorthosite is 52.5 and in the anorthosite only 44.5. The corresponding percentages for aluminum are 28 and 31.5. Thus the amount of aluminum increases and that of silicon decreases considerably from the normal schist (no. 1023) through the feldspathized schist (no. 1690) and the country rock (c) to the anorthosite (a). Calcium and sodium show a considerable increase in the same direction, but the amounts of potassium and iron decrease.

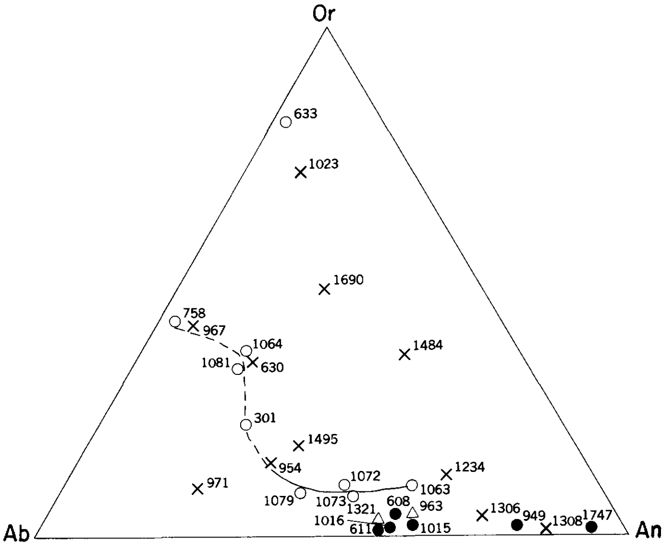

FIGURE 45.—Or-Ab-An diagram for the anorthosite and associated rocks. ●, anorthosite; x, country rocks; Δ, amphibolite; o, intrusive rocks. The amounts of orthoclase (Or), albite (Ab), and anorthite (An) from molecular norm was calculated to 100 and used for coordinates. The numbers refer to the same analyses as those in figure 44.

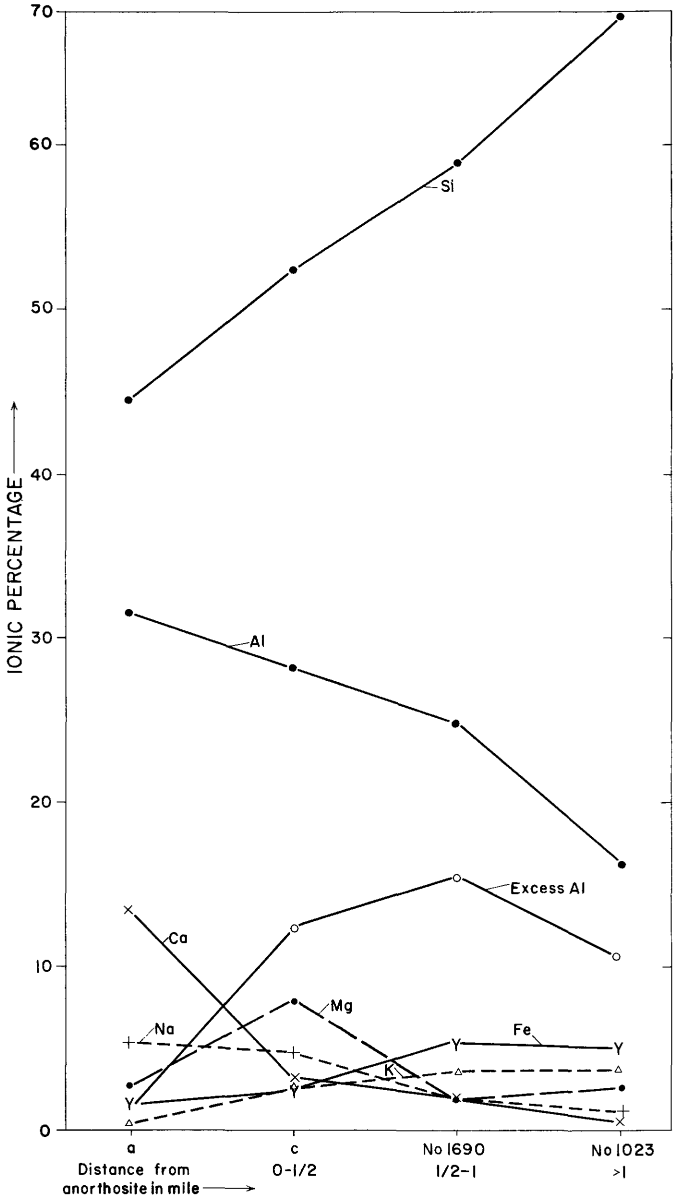

FIGURE 46.—Graph showing the average ionic percentages of silicon, aluminum, calcium, sodium, potassium, iron, and magnesium in anorthosite and its country rocks. The excess of aluminum after subtracting the amount combined with K, Na and Ca to form feldspars is also shown. a=average of the ionic percentages for anorthosites in table 11, c=average of the analyses of the country rocks in table 13, no. 1690 (table 5) represents a schist about a mile from the anorthosite, and no. 1023 (table 5) is a normal schist in the Prichard formation.

The feldspathized schist next to the anorthosite shows the highest concentration of magnesium. The average amount of this element in the anorthosite is slightly higher than that in the normal schist and that in the schist about a mile from the anorthosite (no. 1690).

The excess of aluminum after subtraction of the amount combined with K, Na, and Ca to form feldspars is largest in the schist about a mile from the anorthosite. Even though the total amount of aluminum is highest in the anorthosite, the excess is very slight because of the large amount of calcium and sodium present in this rock.

DISTRIBUTION OF MINOR ELEMENTS

Minor elements were determined spectrographically in those rocks and minerals which were analyzed chemically. In the anorthosite and its country rocks Ni, Co, Cr, Cu, V, Ga, Ba, and Sr were found (table 15); in the garnet amphibolite (table 16), and in various layers of the Prichard formation (table 17) Sc, Y, Yb, and Zr also are common. The two latter groups or rocks contain abundant garnet in which a part of these elements are contained (table 18). Also, tremolite and anthophyllite carry Sc, Y, Yb, and Zr. An appreciable amount of lithium was found in margarite. Beryllium, niobium, lanthanum, lead, and tin are scarce or completely lacking. Boron abounds in those schist layers which contain tourmaline. Ag, Au, Pt, Mo, W, Ge, As, Sb, Bi, Cd, In, Th, and U were looked for but not found.

To compare the average amounts of each element in garnet amphibolite, in anorthosite, and in its country rock, the mean concentration for each element in these three rock types was calculated and plotted in figures 47 and 48. Only a slight variation is seen in the amounts of Ni, Co, Cr, and Ga. In contrast to this Cu, V, Sc, Y, Zr, Ba, and Sr show definite concentrations in certain rock types.

Copper and vanadium are concentrated in the garnet amphibolite, with the anorthosite and its country rock containing only small amounts of these elements. In all rock types, the major part of vanadium is contained in garnet and in amphibole, but these minerals (nos. 1353 and 608a in table 18) in the anorthosite contain only about one-tenth of the amount that is contained in garnet and amphibole of the garnet amphibolite and the metasedimentary rocks (nos. 1225, 568, 1227, and 1306a in table 18). Most of the copper in the garnet amphibolite, however, is contained in chalcopyrite, which was found in some of the thin sections.

No scandium was found in anorthosite or in its plagioclase-rich country rock, whereas the various layers of the Prichard formation and the garnet amphibolite contain an appreciable amount of this element. Among the analyzed minerals (table 18) the garnet is a chief carrier of scandium; some of the amphiboles contain a moderate amount and the micas very little of it. It is noteworthy that the aluminous anthophyllite (nos. 813 and 1227) and the garnet occurring with it (nos. 813b and 1225) are rich in scandium, whereas the hornblende in the anorthosite (calcium-rich environment) and the calcium-rich garnet in skarn in an area nearby (31 miles southwest of Boehls, no. 568) show no trace of this element. Because the temperature of recrystal-

TABLE 15.—*Quantitative spectrographic analyses, in parts per million, of minor elements in anorthosite and its country rocks and in an orthoclase dike in the country rock*

[Analyst, Paul R. Barnett, U.S. Geol. Survey. Leaders indicate zero; Li was not determined]

Specimen	Ni	Co	Cr	Cu	V	Ga	Sc	Ba	Sr	Be	Y	Yb	La	Zr	B	Li	Pb
Anorthosite (see table 11)																	
1747	40	20	20	2	20	20	----	30	4000	----	----	----	----	----	150	----	100
949	10	----	4	50	20	7	----	90	1000	----	----	----	----	----	----	----	----
1016	6	----	5	10	8	6	----	100	800	----	----	----	----	----	----	----	----
1015	100	30	10	20	10	7	----	200	1000	----	----	----	----	----	----	----	----
611	----	----	3	8	8	10	----	70	4000	----	----	----	----	----	----	----	----
608 [1]	200	30	20	4	20	10	----	30	300	----	----	----	----	----	----	----	----
Average	60	13	10	16	14	10	----	90	1867	----	----	----	----	----	----	----	----
Country rocks (see table 13)																	
630	----	5	4	4	20	10	----	2000	3000	----	----	----	----	----	----	----	20
954	25	7	30	----	30	15	----	700	700	----	----	----	----	----	----	----	----
967	70	20	30	2	40	15	----	600	200	----	----	----	----	----	----	----	----
971	----	7	----	4	8	15	----	800	1000	----	----	----	----	----	----	----	----
Average	24	10	16	2	24	14	----	1025	975	----	----	----	----	----	----	----	5
Orthoclase dike (see table 9)																	
633	----	----	2	10	----	7	----	4000	400	----	----	----	----	60	----	----	----

[1] Analyst, A. A. Chodos, U.S. Geological Survey.

TABLE 16.—*Quantitative spectrographic analyses, in parts per million, of minor elements in garnet amphibolite (table 10)*

[Analyst, Paul R. Barnett, U.S. Geol. Survey. Leaders indicate zero; Li was not determined]

Specimen	Ni	Co	Cr	Cu	V	Ga	Sc	Ba	Sr	Be	Y	Yb	La	Zr	B	Li	Pb
963	40	40	30	200	300	20	40	30	300	----	70	5	----	150	----	----	----
1321	40	40	25	400	300	15	40	40	60	----	70	6	----	150	----	----	----
Average	40	40	27	300	300	18	40	35	180	----	70	5	----	150	----	----	----

TABLE 17.—*Spectrographic analyses of minor elements in the metasedimentary rocks* [1]

[Specimen 1690 analyzed quantitatively, and results are in parts per million; other specimens (see tables 5 and 8) analyzed semiquantitatively, and results are in percent, as explained in footnote. Analyst, Paul R. Barnett, U.S. Geol. Survey. Leaders indicate zero; Li was not determined]

Specimen	Ni	Co	Cr	Cu	V	Ga	Sc	Ba	Sr	Be	Y	Yb	La	Zr	B	Pb	Nb
1690	30	20	100	20	100	15	20	1000	100	----	50	4	----	150	500	----	----
1234	0.00x−	0.000x	0.00x	0.00x	0.00x	0.000x−	0.000x+	0.00x	0.00x+	----	0.00x−	0.000x	----	0.00x+	----	----	0.00x−
1306	.000x+	.000x	.00x	.000x+	.00x+	.000x+	.00x−	.000x−	.00x	----	.00x−	.000x	----	.00x+	----	----	.000x+
1308	.000x	.000x	.00x−	.00x+	.00x	.00x−	.00x−	.000x+	.0x−	----	.00x−	.000x	0.00x	.00x+	----	----	.000x+
1495	.000x	----	.00x−	.000x	.00x	.00x	.000x	.0x−	.0x−	----	----	----	----	.00x−	----	0.00x+	----
1484	.000x	----	.00x+	----	.00x+	.00x+	.000x	.0x	.0x−	0.000x	.000x+	.000x−	----	----	0.00x	.000x	----

[1] The concentations of the elements as determined by semiquantitative spectrographic analysis are bracketed into groups each of approximately one-third of the order of magnitude, x+ indicates the higher portion (10–5 percent), x of the middle portion (5–2 percent), and x− the lower portion (2 to 1 percent). Comparisons of this type of semiquantitative results with those obtained by quantitative methods either chemical or spectrographic, show that the assigned group includes the quantitative value about 60 percent of the time.

TABLE 18.—*Quantitative spectrographic analyses, in parts per million, of minor elements in amphiboles, micas, and garnet from the rocks of Boehls Butte quadrangle and vicinity*

[Analysts: A. A. Chodos, no. 608a; Paul R. Barnett, nos. 1306a, 971c, 1484a; Harry Bastron, nos. 813a, 1227, 568, 813b, 1225a, 1225b, 1353, 13, 354; all of the U.S. Geol. Survey. Leaders indicate zero; Li was not determined]

Analysis	Ni	Co	Cr	Cu	V	Ga	Sc	Ba	Sr	Be	Y	Yb	La	Zr	B	Li	Pb	Tl [1]	Sn	Zn
608a	300	100	3	10	20	10	----	5	20	----	----	----	----	----	----	----	----	----	----	----
1306a	3	----	70	3	150	2	15	150	15	----	15	2	----	30	----	----	15	(1,500)	----	----
813a	5	30	20	7	800	30	50	----	----	----	30	6	----	80	----	----	----	----	----	1,000
1227	100	40	200	9	300	40	60	20	----	----	50	8	----	100	----	----	----	----	100	----
971c	40	50	4	6	30	30	----	2,000	----	----	----	----	----	----	----	100	----	----	----	----
1484a	7	3	150	15	70	70	3	150	700	7	----	----	----	----	30	1,500	30	(1,500)	----	----
568	----	----	----	10	200	20	----	----	----	----	20	4	----	50	----	----	----	----	----	----
813b	----	30	9	20	100	10	300	----	----	----	200	20	----	80	----	----	----	----	----	200
1225a	----	20	80	20	200	----	100	----	----	----	200	30	----	100	----	----	----	(200)	----	----
1225b	----	20	100	80	100	10	60	30	----	----	100	20	----	200	----	----	----	(1,000)	----	----
1353	----	30	20	7	10	10	40	----	----	----	30	9	----	30	----	----	----	----	----	----
13	----	40	300	60	80	20	300	----	----	----	400	70	----	400	----	----	----	----	----	----
354	----	20	40	7	200	10	100	----	----	----	60	10	----	40	----	----	----	----	----	----

608a, Hornblende, table 12
1306a, Tremolite, table 6
813a, Aluminous anthophyllite (gedrite) from garnet-kyanite gedritite near Orofino (Hietanen, 1959)
1227, Aluminous anthophyllite (gedrite), table 3
971c, Biotite, table 14
1484a, Margarite, table 8
568, Grossularite from lime-silicate rock at Bruces Eddy near Orofino (Hietanen, 1962) [2]
813b, Garnet from garnet-kyanite gedritite near Orofino (Hietanen, 1959)
1225a, Garnet, table 4
1225b, Garnet, table 4
1353, Garnet, table 4
13, Garnet from garnet-biotite-plagioclase rock, Dent (Hietanen, 1962)
354, Garnet from garnet amphibolite, Orofino Creek (Hietanen, 1962)

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[1] Tl probably introduced by heavy-liquid separation procedures.

[2] Composition in molecular percentages is: grossularite, 85.74; andradite, 12.52; almandite, 1.10; pyrope, 0.46; and spessortite, 0.18

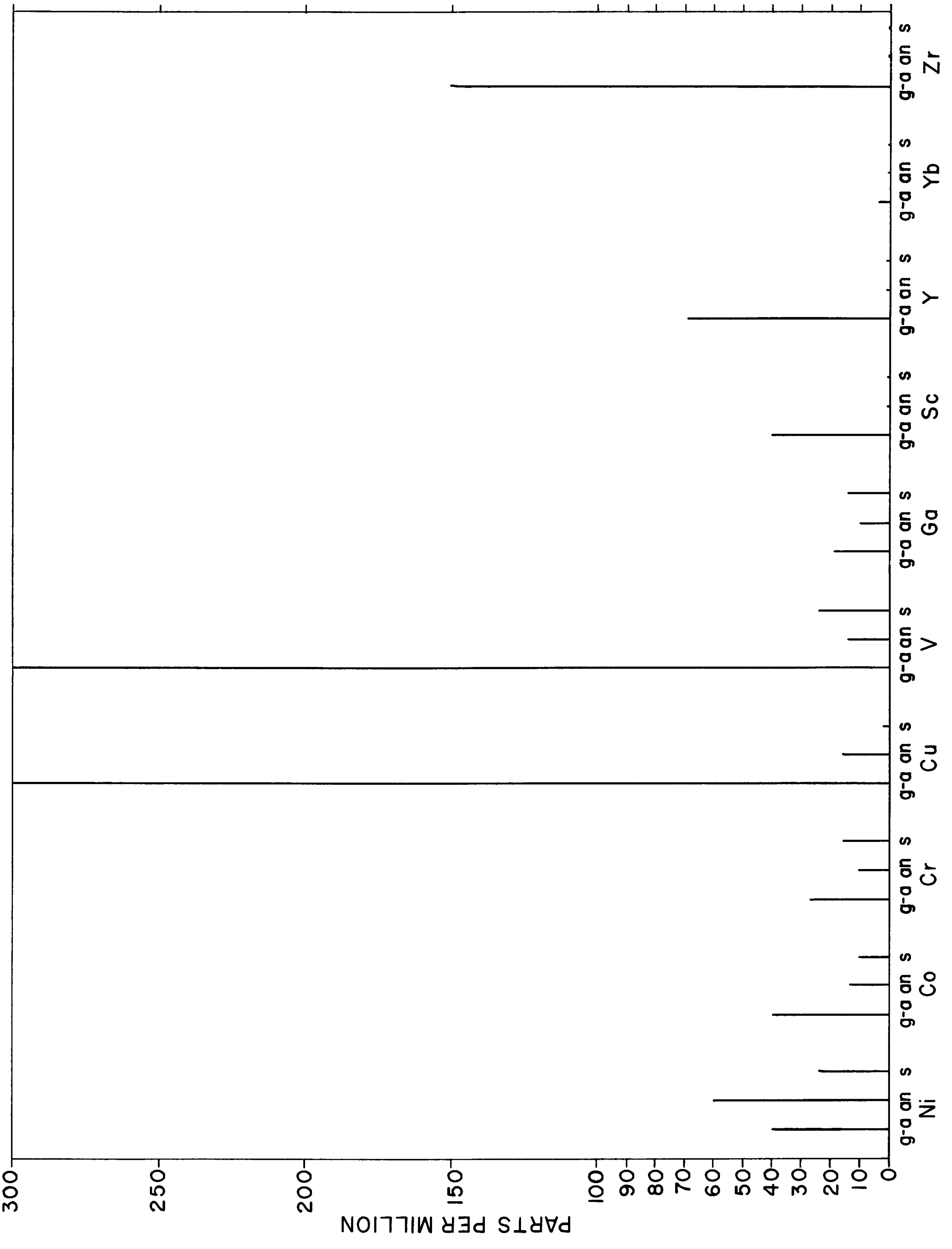

FIGURE 47.—Average concentration of Ni, Co, Cr, Cu, V, Ga, Sc, Y, Yb, and Zr in the garnet amphibolite (g–a), the anorthosite (an), and in the country rock (s).

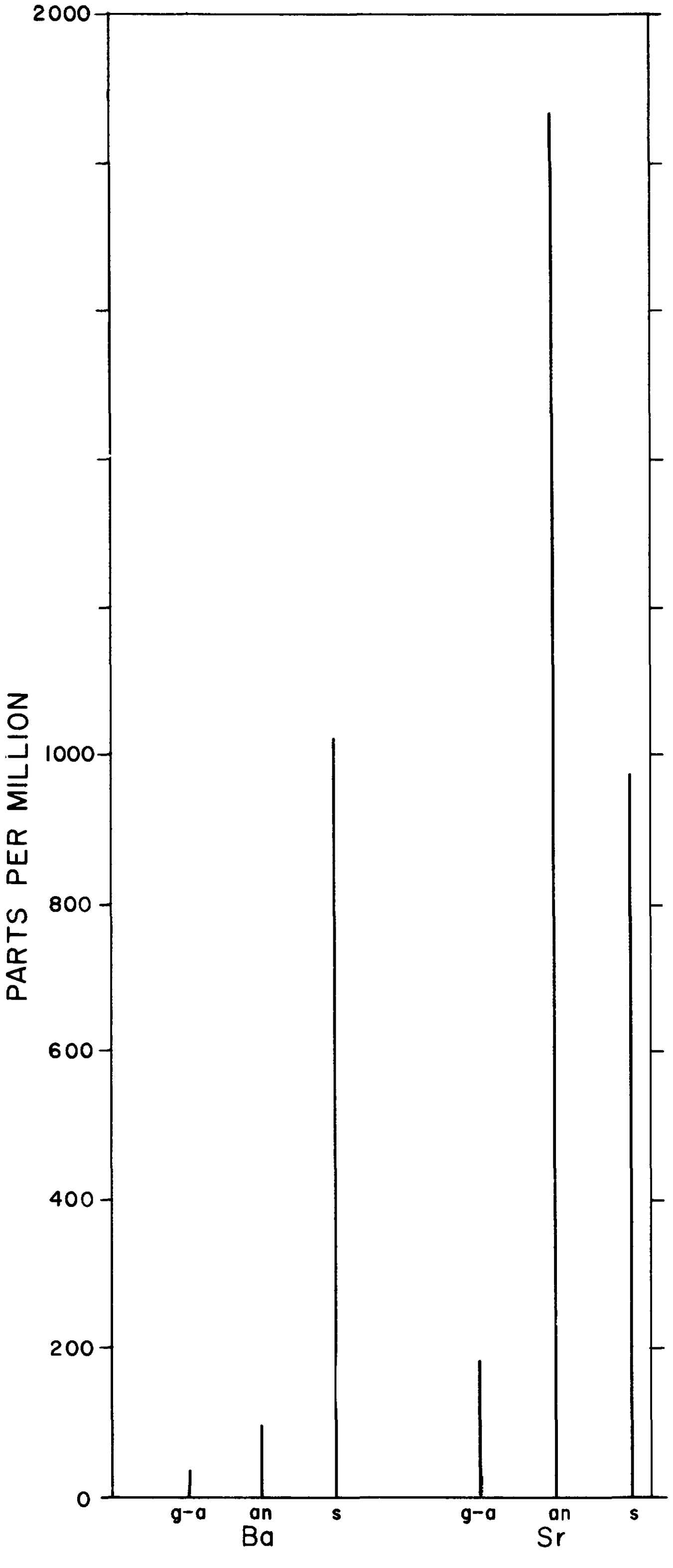

FIGURE 48.—Average concentration of barium and strontium in the garnet amphibolite (g-a), the anorthosite (an), and in the country rock (*s*).

lization was about the same for all rocks analyzed, this distribution suggests that the enrichment of scandium depends much on the chemical environment. It was earlier found that most micas of this region do not contain scandium (Hietanen, 1962); the present study confirms that the garnet is its chief carrier, the almandine of the garnet-biotite-plagioclase rock (no. 13) and the pyrope of the garnet gedritite being richest in this element.

The pattern of distribution of yttrium, ytterbium, and zirconium follows closely to that of the scandium. The highest concentrations are found in the garnet-amphibolite, and the anorthosite and its country rocks are void of these elements. Study of the minerals suggests that garnet and amphiboles are the chief carriers. However, a few small zircon crystals occasionally are found in the garnet amphibolite.

Barium and strontium are concentrated in anorthosite and in its country rocks, the other rocks of the area containing much less of these elements. In the anorthosite, the amount of strontium is about ten times higher than that of barium, whereas in the country rocks these two elements occur in equal amounts. The highest concentration of barium is found in the orthoclase dike,' the muscovite-rich schist (no. 630) being the second in order. It was found earlier that muscovite contains three times as much barium as the biotite in the same rock (Hietanen, 1962). Orthoclase in specimen no. 633 has 4,000 ppm barium, thus twice the amount contained in biotite (no. 971c). The calcium mica (the margarite no. 1484a) contained 700 ppm strontium and only 150 ppm barium. Thus barium is concentrated in the rocks and minerals rich in potassium, whereas strontium accompanies calcium and is contained mainly in plagioclase.

The distribution of the minor elements has a pattern similar to that found in the area south of Boehls Butte quadrangle (Hietanen, 1962). Strontium is concentrated with calcium in the plagioclase-rich rocks, as in anorthosite and its country rocks. Barium follows closely the distribution of minerals rich in potassium, such as orthoclase, muscovite, and biotite. High concentrations were found in the country rocks of the anorthosite.

Copper, vanadium, zirconium, yttrium, and ytterbium are concentrated with magnesium and iron in the garnet amphibolite that represents the basic rocks of the area and can be correlated with the accumulations of ferromagnesium minerals in the area south of the Boehls Butte quadrangle.

COMPARISON OF THE OCCURRENCES OF ANORTHOSITE

On the basis of the above descriptions of the anorthosite in the Boehls Butte quadrangle, it is evident that there are a few similarities but also several marked differences between the anorthosite in the Boehls Butte quadrangle and the occurrences of anorthosite described in the literature.

The mode of occurrence and the shape of anorthosite bodies in the Boehls Butte quadrangle are, in some respects, comparable with those of the Jay mass within the Adirondacks anorthosite area as described by Buddington (1939, p. 25). He writes: "The curved shape of the Jay mass, the foliation of the outer border, and the shape and foliation of the prong of the main anorthosite massif to the south will suggest that it is a folded sheet on the nose of a north-pitching anticline." The petrology of the anorthosite and that of the associated rocks in the Boehls Butte quadrangle, however, are quite different. The Adirondack anorthosite contains only one type of plagioclase, usually andesine or labradorite (Barth, 1930), and the mafic constituents in it tend to increase toward the contact zones. No aluminum silicates of such an extent as present in the Boehls Butte quadrangle have been reported from the Adirondacks or any other anorthosite area. Only corundum has been reported earlier in some rare hornblende-bearing anorthosites (Miller, 1899, p. 279–280, Rosenbusch and Osann, 1922, p. 243) and in oligoclase pegmatites (Barth, 1927, p. 98).

Pyroxenes are common mafic constituents in most anorthosites described in the literature, suggesting a dry environment (Suter, 1922; Chatterjee, 1929). In contrast to this, hornblende and chlorite are the regular dark constituents of the anorthosite in the Boehls Butte quadrangle, suggesting a wet environment. Biotite, which is rare in other anorthosites, is common in most of the anorthosite under discussion. No block structure or cataclastic textures that are common in the Adirondacks (Balk, 1931, p. 357–358) and in Nain, Labrador (Wheeler 1942, p. 619) were found in the anorthosite in the Boehls Butte quadrangle. Large crystals of plagioclase that are common in the Marcy anorthosite and in similar coarse-grained dark anorthosite of Morin Heights, near Montreal, are rare in the Boehls Butte quadrangle where the grain size and color of the coarse-grained massive anorthosite are closest to those of the light-gray, coarse-grained anorthosite at San Gabriel Mountains, California. On the other hand, the hornblende-bearing gneissic variety resembles the Whiteface facies in the Adirondack area except that garnet is rare in this type in Idaho and the layering is more pronounced.

A gradation to a gabbroic anorthosite, gabbro, and norite is common in most anorthosite areas described in the literature (Buddington, 1939; Harrison, 1944; Wheeler, 1942; Michot, 1955, 1957) but very rare in the area under discussion. No normal syenitic rocks, as are common elsewhere, are associated with anorthosite in the Boehls Butte quadrangle. The only potassium-rich rock found is the orthoclase dike (table 9) near Boehls. The associated intrusive rocks are quartz diorite and quartz monzonite, the latter being younger than the anorthosite.

In the Boehls Butte quadrangle, the floors of most anorthosite bodies are exposed, but the mafic rocks, amphibolite, and garnet amphibolite rarely occur along this lower contact zone. Rather, the associated mafic rocks occur as small concordant lenses in the country rocks above as well as below the anorthosite, or are included in the anorthosite bodies. These mafic bodies are richer in silicon and aluminum than the normal gabbros. Many bodies contain abundant biotite and garnet and grade over to garnet-mica schist. Most are concordant but a few dikelike discordant bodies were found. Some of the mafic bodies in anorthosite are older than the anorthosite; a few next to the anorthosite bodies seem contemporaneous or younger.

The distribution of mafic rocks and garnet amphibolite in the same general area where anorthosite occurs suggests a genetic relationship. They form the only group of basic rocks that can be complementary to the anorthosite. The fine-grained gabbro with diabasic texture occurs as sill-like bodies and is definitely younger than the anorthosite. In general, the mafic rocks of this area range in age from pre-anorthosite to post-anorthosite, only the youngest group being true gabbro. The volume of the mafic rocks exposed is subordinate and quite inadequate to give a gabbroic composition if added to the anorthosite. However, the amount of the mafic constituents is larger than the geologic map (pl. 1) shows, because some of the surrounding schist is enriched in ferromagnesian minerals such as garnet, biotite, and cordierite. The enrichment of the country rocks in aluminum silicates is, however, far more striking and one of the unique features of the anorthosite area in the Boehls Butte quadrangle.

The association of the anorthosite with metasedimentary rocks rich in aluminum is comparable to that of the anorthite rocks described briefly by Lacroix (1939) from Madagascar. These rocks, called "sakenites" by Lacroix, are found as layers as much as 12 m thick with paragneiss rich in sillimanite, cordierite, and almandine. The other associated rocks are pyroxenites and amphibolites of various kinds. The sakenites are bluish or greenish white, medium to coarse grained, saccharoidal, resembling marble in their appearance. Some samples show indistinct banding. Anorthite or some other plagioclase close to anorthite in composition is the main constituent; the additional minerals are spinel, sapphirine, corundum, and leucaugite rich in aluminum. Some light-colored phlogopite or chlorite occurs in places. In some layers corundum is a major constituent. The description is too brief to give a full picture of these rare types of calcic plagioclase rocks or of the associated rocks. Their mode of occurrence in rather thin layers in the metasedimentary strata (paragneiss) seems to suggest that they are of sedimentary origin and are probably so considered by French geologists.

GENESIS OF THE ANORTHOSITE

As pointed out in the introduction, the igneous origin of anorthosite is generally accepted, and the discussions are centered around the possible composition of the parent magma and around the mechanism of differentiation and intrusion or possible later modifications through metamorphic and metasomatic processes. The contact relations, structure, texture, and mineralogy of the anorthosite in the Boehls Butte quadrangle make an intrusive origin for these particular bodies questionable. Some of the layers in the metasedimentary rocks consist of about 90 percent calcic plagioclase (bytownite), giving rise to a thought of a possible sedimentary origin. On the other hand, the widespread enrichment of country rocks in elements like aluminum and calcium suggests that metasomatic processes played an important role in the formation of these anorthosite bodies. In the following discussion each possible explanation is summarized and weighed.

POSSIBILITY OF IGNEOUS ORIGIN

The massive coarse-grained anorthosite that is common in the central parts of the two northernmost large bodies and that also occurs locally elsewhere has the appearance of an intrusive rock. However, the presence of two plagioclases—andesine and bytownite—and their peculiar texture are difficult to explain on the basis of our present knowledge about the crystallization of the albite-anorthite series. It is apparent that anorthosite of this composition and texture could not be formed through a simple crystallization differentiation from a gabbroic or gabbroic-anorthositic magma as has been suggested for the anorthosite in general. Magmatic origin must be questioned unless it can be shown that in certain physico-chemical conditions andesine and bytownite will crystallize contemporaneously from a liquid and form separate grains, or that a later exsolution of labradoritic plagioclase would yield a mixture of andesine and bytownite. Moreover, the physicochemical conditions that prevailed during such crystallization or exsolution must be consonant with the geology of the area.

The occurrence of sillimanite and muscovite together in the schist and in the schistose inclusions in the anorthosite suggests that the temperature was under 650°C (Yoder and Eugster, 1955). There was enough water present to prevent the crystallization of pyroxenes. If it is assumed that the associated amphibolites are differentiates of the same magma as the anorthosite, the composition of such magma should have been similar to the composition of the anorthosite no. 1015 (table 11). It is clear that a magma of this composition could not be liquid under 650°C. The presence of abundant volatiles could have prevented freezing of the small amount of liquid needed to facilitate an intrusion of a mush of plagioclase crystals. However, there are no signs of such an extensive protoclasis as would be expected to result from an intrusion of fairly solid crystal mush. The small grains of bytownite between the larger grains of andesine resemble the granulated texture, but the composition of the two plagioclases refutes such a theory. If the granulated small grains crystallized from the residual liquid, they should be more sodic than the large crystals. The banded structure along the border zones of the anorthosite is an alternation of thin layers of andesine and bytownite and, therefore, cannot be considered to result from the granulation alone.

If it is assumed that a very large amount of volatiles were present to keep the plagioclase in the solution at a temperature close to 650°C, such a solution is no longer a magma, but rather resembles a pneumatolytic or hydrothermal solution, and a rock deposited from such a solution is not magmatic in its strict sense.

The other possibility would be a later exsolution of andesine and bytownite from an originally labradoritic plagioclase and a later low-temperature metamorphism. The mineral associations found now would be the result of a recrystallization of the primary constituents in a low-temperature field. If this were the case, one would expect to find some traces of the earlier high-temperature minerals; one should also find a similar exsolution in many other areas in the world where plutonic rocks have been subjected to a low-temperature metamorphism. However, there are no reports of such an exsolution. Moreover, it would be difficult to explain why parts of the border zones contain exclusively bytownite, whereas only andesine is found in the plagioclase-rich schist. The common occurrence of minerals that are typical of rocks of sedimentary origin (kyanite, andalusite, sillimanite, staurolite, biotite, garnet) and of the beddinglike structure in the anorthosite, the gradational contacts between the anorthosite and the schist, and the substitution of andesine for the quartz in the schist next to the anorthosite, are further features that refute the hypothesis of igneous origin of these anorthosite bodies.

The anorthosite was found to change over to a gabbroic anorthosite only in one locality and to amphibolite in two localities. One would expect to find such a change in many places if the anorthosite were a differentiate of a gabbroic-anorthositic magma.

POSSIBILITY OF ISOCHEMICAL METAMORPHIC ORIGIN

Many of the contact relations, the mineralogy, and the structure of the anorthosite resemble those found in the rocks of sedimentary origin. The distribution of the various rock types in the Belt series is lenticular, and it is possible that a plagioclase-rich sediment was desposited in the lower part of the Prichard formation. No such distinctly discordant contacts as described from the vicinity of Dent, about 20 miles southwest of Boehls (Hietanen, 1962), were found in the present area. Therefore, the possibility of sedimentary origin must be carefully considered. This possibility is strengthened by the occurrence of the lime-silicate rocks, some of them rich in bytownite, and andesine-phlogopite schist in the lower quartzite unit of the Prichard formation as described earlier in this report. Both of these rocks are rich in calcium and contain a considerable amount of aluminum (table 5, nos. 1234, 1495). The mode of occurrence, texture, and mineralogy of the two bytownite rock layers along the West Fork of Cedar Creek suggest a sedimentary rather than an igneous origin for the bytownite anorthosite. Paper-thin layers of bytownite interbedded with thin layers of quartzite and lime-silicate rock were found in two other localities. Some of these thin beds have a peculiar texture; bytownite occurs in grains of medium size and includes numerous tiny round quartz grains. This kind of texture is typical of metamorphic rocks. It is scarcely possible that the paper-thin laminae of bytownite in this rock were crystallized as a result of introduction of elements. The layering is more likely due to the alternation of dolomitic, limy, shaly, and sandy layers. The bytownite rocks may represent thicker layers of pure plagioclase in a similar sequence. As pointed out earlier, one of these bytownite layers is interbedded with fine-grained bedded hornblende-biotite schist, originally probably a calcareous clay.

It is more difficult to explain the occurrence of bytownite in the kyanite-andalusite-bearing anorthosite that contains inclusions of aluminous schist. If the bytownite in the anorthosite with excess of aluminum were of sedimentary origin, these sediments should have been calcareous shales exceptionally rich in aluminum and very poor in iron and magnesium. No such sediments are known to occur elsewhere in the Beltian strata, nor have descriptions of such sediments been found in the literature. The only reference that has some bearing on the subject is the short description by Lacroix (1939) of a corundum-bearing anorthite rock layer in a paragneiss in Madagascar. In the Idaho area the aluminous schist farther from the anorthosite bodies is poor in calcium, and most of the calcareous layers and masses in this schist are rich in iron and magnesium. Hence, during metamorphism the calcareous rocks gave rise to the formation of amphibolites and lime-silicate rocks rich in diopside or in amphiboles of the tremolite-actinolite series.

The general aspect and mineralogy of the lime-silicate rocks along Cedar Creek are very different from those of the anorthosite. The bedding in the lime-silicate rocks and in the phlogopite-andesine schist is definitely more distinct and more regular than the beddinglike structure in the anorthosite. The sedimentary phlogopite-andesine schist along the Little North Fork is very similar to the fine-grained thin-bedded gray biotite quartzite with which it is interbedded, whereas most of the anorthosite is coarse grained and the centers of the bodies are massive. The lime-silicate rocks along Cedar Creek have varied mineralogy, and none of the layers studied is similar to the anorthosite. The lime-silicate rocks contain abundant pyroxene, amphibole, and other silicates rich in lime; quartz is always present and constitutes about 30 percent of most layers. Graphite is a typical accessory mineral in the lime-silicate rock but is completly lacking in the anorthosite. Phlogopite is common in the lime-silicate rocks, but in the anorthosite the micaceous minerals are biotite and muscovite. Moreover, the lime-silicate rocks contain only one type of plagioclase, the composition of which varies from An_{25} to An_{85}, whereas the anorthosite has two types of plagioclase. The lime-silicate rocks do not contain aluminum silicates that are common in a part of the anorthosite. If the anorthosite represented isochemically metamorphosed limy layers, one would expect to find every gradation from the metasedimentary rocks rich in lime to the anorthosite, but this is not the case.

In many respects the mineralogy of the anorthosite shows a closer resemblance to the surrounding schist than to the lime-silicate rocks. Biotite, muscovite, and the aluminum silicates are common additional constituents in large parts of the anorthosite.

The occurrence of two plagioclases together in the anorthosite is another feature that would be difficult to understand if it is assumed that no migration of material were involved. One might suggest that andesine crystallized first including some calcite, aluminum-silicates, and quartz, and that during a later phase a higher temperature gave rise to the reaction between andesine and included minerals producing bytownite.

This theory includes the assumption that the sedimentation had accidentally produced a bulk composition that was just right for the formation of anorthosite. Thus, there should have been a sudden facies change from a normal sedimentary sequence of schist and quartzite to an anorthositic composition.

The gradation of the anorthosite to a plagioclase-rich shist along the concordant contacts can be interpreted to be depositional, but it is more difficult to explain the fairly abrupt interfingering contact between the anorthosite and the schist of the Prichard formation poor in feldspar east of Widow Mountain. In this locality the schist ocurring just west of the coarse-grained massive anorthosite is fine-grained biotite schist typical of those layers which elsewhere are found just above the quartzite units of the Prichard formation. This schist is separated by a fault from the upper quartzite unit at Widow Mountain and by a narrow zone of anorthosite from the lower quartzite unit in the north. In this locality the anorthosite seems to have a similar discordant contact with the schist layers of the Prichard formation as shown on a small scale in figure 20.

Between Widow Mountain and Lund Creek the anorthosite lies conformably under a thin layer of light-gray medium-grained quartzite that becomes thicker east of Lund Creek. In three localities between Lund Creek and Monumental Buttes the contact seems discordant, in each case moving down in the stratum. Near Monumental Buttes a layer of schist, about 100 m thick, separates the anorthosite from the quartzite. Thus, the northeastern contact of the anorthosite between Widow Mountain and Monumental Buttes cuts slightly discordantly from the beds of the major quartzite layer to the schist under the quartzite.

A similar slightly discordant contact can be traced between Stocking Meadow Lookout and Cedar Creek. The southwestern contact between the anorthosite and the schist between the heads of Pinchot Creek and Timber Creek also seems to cut across the beds; northeast of the head of Pinchot Creek the schist lies above the anorthosite that is exposed just north of it along the tributaries of Lund Creek but dips under the anorthosite toward the east.

A slightly discordant contact between amphibolite and a small body of anorthosite exposed along the logging road that follows the Little North Fork of the Clearwater River toward Cedar Creek has been described earlier (fig. 21). This anorthosite is a very coarse-grained pegmatitic variety and may represent a part of a dikelike body. Contacts of this kind are common in concordant intrusive bodies but would not be expected to occur between various beds of a sedimentary formation. Moreover, a discordant dike consisting of about 99 percent labradorite (An_{52}) was found south of the Boehls Butte quadrangle (Hietanen, 1962).

It would be difficult to understand many other field relations if it were assumed that no migration of material was involved but that the anorthosite is a product of isochemical metamorphism; for example, the massive coarse-grained variety is found next to a fine-grained schist and quartzite in many localities near Orphan Point. It would be difficult to understand how some beds (the anorthosite) recrystallized as a coarse-grained rock, losing all their original sedimentary structures, while the neighboring beds show only a moderate recrystallization and a well-preserved bedding. Thus the hypothesis of the sedimentary origin alone does not explain all the facts seen in the field.

POSSIBILITY OF METASOMATIC ORIGIN

It has been pointed out that the hypothesis of igneous origin and that of sedimentary origin each fails in certain respects. The mainly concordant and gradational contacts and the feldspathization of the country rocks resemble closely the field relations found around the metasomatic granite bodies with the exception that instead of alkali feldspars, the newly crystallized mineral in the area under discussion is andesine (An_{45-48}) in the anorthosite and oligoclase-andesine (An_{25-45}) in the country rocks.

In general, it is possible to find both types of contacts, discordant and concordant, between the metasomatic rocks and their host rocks. The distribution of andesine in the area south of the Boehls Butte quadrangle was found to be lenticular on a large scale, but when the contacts were studied in detail both concordant and discordant contacts were found. Also, the contact between a metasomatic rock and the host rock can be either sharp or gradational. The gradational contacts have been described from many granitized areas and are considered by many petrologists to be typical of metasomatic rocks. Sharp contacts were found in the vicinity of Dent (Hietanen, 1962).

The beddinglike structure that is pronounced in many localities and the sheetlike inclusions of schist in the anorthosite can be explained as relict structures. Both of these structures are parallel to the bedding of the country rocks. The distribution of the outcrops that contain remnants of the schist is rather irregular, and the schistose layers are seen to fade out gradually when followed parallel to the structure. The schistose layers in the anorthosite contain abundant aluminum silicates just as does the schist that surrounds the anorthosite bodies.

Abundant calcic plagioclase and hornblende were crystallized in the schists and gneisses south of the Boehls Butte quadrangle (Hietanen, 1962). This calcic plagioclase and hornblende replace mainly quartz in the schist, and quartz and diopside in the diopside-plagioclase gneiss. In some localities, the calcic plagioclase was found as dikes and veinlets cutting a gneissic diorite and a feldspathized schist. This shows that an anorthositic rock can be derived through an accumulation of calcic plagioclase during a metasomatic introduction of Mg, Fe, Ca, Al, and possible Na. These dikes and veinlets range only from a few centimeters to about 1 m in width; the question arises whether or not a similar segregation during the metasomatism could be responsible for the formation of bodies which are as much as 20 km in length and 2 km in thickness. Such a segregation would require an extensive exchange of material and a large-scale migration of cations such as Ca, Al, Fe, and Mg which are not so easily detached from the mineral structures and some of which are heavy (Ca, Fe). The question of the formation of the anorthosite is further complicated by the occurrence of two plagioclases.

Only that hypothesis which is capable of explaining all the facts seen in the field and found in the laboratory should be accepted in explaining the origin of the anorthosite. A closer study of the features suggesting metasomatic origin may contribute to the solution.

EVIDENCE OFFERED BY THE MINERALOGY OF THE ANORTHOSITE

First it is necessary to find out which of the constituent minerals of anorthosite crystallized late and are responsible for the features that suggest metasomatic origin. Two mineralogic features of the anorthosite, the occurrence of two types of plagioclase together and the abundance of the minerals of the country rocks in the anorthosite, are strong evidence of metasomatic processes in formation of the anorthosite. The two compositional varieties of plagioclase should react and form labradorite at metamorphic temperatures. Their case is comparable with those of the other pairs of incompatible minerals. Usually in such cases, one is older and preserved only because it was protected by a shell of some other composition, or because the rock was cooled so rapidly that there was no time for the reaction between the minerals to take place. In the rocks under discussion, the bytownite is not protected but occurs between the andesine grains and forms small inclusions in the andesine. The contact relations between the andesine and the interstitial bytownite (fig. 27) and those between the andesine and its inclusions (fig. 36) strongly suggest that bytownite is an earlier mineral than, and is partly replaced by, andesine. Peculiarly, the bytownite grains next to the andesine and most of the inclusions in the andesine seem to have partially recrystallized and reoriented, forming larger grains and groups of inclusions that have a parallel optic orientation. Thus, the andesine grains seem to have been centers of recrystallization representing a later phase than the crystallization of interstitial bytownite. Sodium must have been added during this phase because the newly formed mineral is considerably more sodic. Silica needed for the formation of andesine may have been present as quartz.

The coarse-grained anorthosite in the massive parts of the large bodies generally contains only tiny platelike inclusions of bytownite. More rarely, interstitial bytownite is found in this variety. Hornblende fills some of the interstices between the large labrodorite grains in this rock. It is possible that the reaction between the earlier bytownite and the added sodium has proceeded farther in these localities.

The common occurrence of the aluminum silicates—kyanite, andalusite, and sillimanite—and of garnet, staurolite, biotite, and muscovite in many layers in the anorthosite suggests that the anorthosite contains abundant sedimentary material. Kyanite was found with hornblende in some layers. Petrologists have pointed out the incompatibility of these two minerals. Their occurrence together in the anorthosite is easily explained if it is assumed that the kyanite originated in the schist that was gradually replaced by plagioclase to form anorthosite. Thus, the kyanite would be a relict mineral, whereas some of the elements that formed hornblende were introduced during the formation of anorthosite.

EVIDENCE OFFERED BY METASOMATIC CHANGES IN THE COUNTRY ROCKS

A closer study of the mineralogic changes in the schist around the anorthosite bodies suggests an extensive exchange of material in a zone whose width ranges from a few meters to several kilometers. The occurrence of numerous large grains and small masses of andesine, or of oligoclase-andesine instead of quartz in the schist around the anorthosite bodies is comparable with the feldspathization of the schist near Dent (Hietanen, 1962), and can be explained as a result of introduction of calcium and sodium into and removal of silicon from the schist.

The schist next to the three largest anorthosite bodies contains considerably more aluminum silicates than the same schist a little farther from the anorthosite. As described earlier by the author (Hietanen, 1956), all three modifications of Al_2SiO_5 occur together in this contact zone. Their occurrence together was attributed to a fluctuation of the temperature around the triple point where all three aluminum silicates can exist in equilibrium. In a wide area around the mouth of Timber Creek, thus in a zone a little farther from the contact of anorthosite, a replacement of biotite by the brown sillimanite was observed. The replacement of biotite by the sillimanite suggests that at least some of the aluminum was introduced and potassium removed from this zone. Plagioclase is less abundant and occurs as smaller grains in this outer zone than in the zone next to the large anorthosite bodies. Considerably more aluminum silicates occur in those parts of the stratum in which plagioclase abounds (fig. 46), suggesting a metasomatic addition of aluminum in these parts also.

EVIDENCE OF POSSIBLE PARENT ROCKS

Numerous remnants of the older rocks in the anorthosite give best evidence of rock types that originally occupied the parts of the area where anorthosite occurs now. In the southernmost body, most of the inclusions and ghostlike remnants of older rocks consist of an aluminum-rich schist similar to that found around the body. Amphibolite is included in the anorthosite north of Benton Butte, and shadowy remnants of quartzite and diopside-plagioclase gneiss south of Boehls. Layers and lenticles rich in bytownite were found along the contacts of the northernmost and the middle anorthosite bodies. In the upper parts of these two bodies, schist inclusions are common and some inclusions consisting of amphibolite, quartzite, and lime-silicate rock were found. Thus, it seems possible that most of the original rocks in the part of the area now occupied by the southernmost anorthosite body were of the same composition as those found in the surroundings of the anorthosite. Aluminum-rich schist may have been the major rock type and quartzitic and calcareous layers and masses were interbedded with it. This schist contained lenticular bodies of amphibolite as does the schist surrounding the anorthosite.

The fine-grained bytownite rock that was found as layerlike bodies in the northern part of the middle anorthosite probably was an earlier rock type in a much wider area than now exposed, and was transformed to the andesine-bytownite anorthosite by a later introduction of sodium and possibly silicon if no excess quartz was present. If this was the origin of the massive anorthosite, bytownite rock should have occupied large areas and a question about its origin forms a vital part of the problem concerning the ultimate origin of the anorthosite.

ORIGIN OF BYTOWNITE

The bytownite must have originated in one of these three ways: (1) bytownite is of igneous origin, crystallizing from a melt; (2) bytownite results from the metamorphism of sedimentary material; (3) bytownite is of metasomatic origin.

IGNEOUS

The first way would mean that the layers of bytownite rock would by metamorphosed fine-grained bytownite anorthosite and that most of the bytownite would be of igneous origin. The massive anorthosite would represent larger masses of igneous calcic anorthosite transformed to a more sodic variety by metasomatic introduction of sodium. The layered anorthosite with aluminum silicates would represent either thin layers of igneous anorthosite in the schist or metasomatic contact aureoles of the larger igneous masses. One could think that first bytownite crystallized in the country rocks in a manner similar to the alkali feldspars in the surroundings of many igneous granite bodies. During the introduction of sodium, andesine crystallized as a result of reaction between the sodium and this earlier bytownite. If the volume stayed unchanged, a part of the calcium was removed from the bytownite-bearing rocks. This calcium, together with sodium that was introduced and aluminum and silicon that were present in the schist, formed oligoclase-andesine in the schist next to the bytownite-bearing contact aureoles. The sodium metasomatism could have obliterated all traces of the high temperature that should have accompanied intrusion of bytownite anorthosite.

However, the gradational contacts and especially the thin micaceous laminae in many small anorthosite bodies near Timber Creek cast doubt on an igneous origin of bytownite. It is scarcely possible that the paper-thin laminae of micaceous material that should originate in the schist would be preserved during an igneous intrusion of bytownite anorthosite and remain as parallel inclusions in it. Rather these thin bodies are comparable with the aluminum silicate-bearing parts of the large bodies, in which the bytownite and andesine are either both metasomatic or else the bytownite is of sedimentary and the andesine of metasomatic origin.

METAMORPHIC

The bytownite rock contains about 16 percent CaO, more than 30 percent Al_2O_3, and only about 1 percent combined FeO and MgO. This composition is quite abnormal for a sediment. The content of calcium is too high for an argillaceous sediment and that of aluminum too high for a calcareous sediment. Therefore it seems highly questionable that bytownite rocks would be products of isochemical metamorphism of sediments. There is, however, another possibility. The lens-shaped garnetiferous bodies in the bytownite layers along the lower contact of the northernmost anorthosite body are, in all probability, metamorphosed calcareous concretions suggesting a sedimentary origin for these bytownite layers. The neighboring layers consist of andesine anorthosite rich in aluminum silicates and micas, the abundance of these minerals indicates that the original sediment in these layers was rich in aluminum. Thus the original sedimentary strata in this locality consisted of calcareous layers interbedded with clayey layers, both poor in iron and magnesium. During the metamorphism, bytownite crystallized in the calcareous layers from calcium, silicon, and sodium present in these layers and the aluminum provided by the neighboring clayey layers. A small fraction of calcium probably migrated in the opposite direction, forming bytownite in the aluminum-rich layers.

The two layers of the bytownite rock at Cedar Creek most likely have a similar modified sedimentary origin. Also here the aluminum probably was provided by the neighboring aluminum-rich sediments. The lack of aluminum silicates in the massive anorthosite can be considered as an indication of the prevalence of calcareous sediments in these parts of the area. However, the crystallization of the albite component of the andesine may have used some aluminum, according to the following equation:

$$\underset{\text{Kyanite}}{Al_2O_3 \cdot SiO_2} + \underset{\text{quartz}}{5SiO_2} + \underset{\text{sodium rich solutions}}{Na_2O} \longrightarrow \underset{\text{albite}}{Na_2O \cdot Al_2O_3 \cdot 6SiO_2}$$

Thus the older rock may have contained aluminous layers in all localities where micaceous layers are now common, even if no aluminum silicates are found. The pronounced layering in the anorthosite containing aluminum silicates and micas suggests a strong compositional variation that may be due to the original sediment. Comparison of analyses nos. 611 and 608 in table 11 shows that the main difference between the dark- and light-colored layers is a larger amount of iron and magnesium and a smaller amount of calcium and sodium in the dark layers. In the micaceous layers the amount of potassium would be higher (compare analysis no. 630, table 13). The content of $A1_20_3$ in the anorthosite shows remarkably little variation, being about 27 percent in the dark micaceous and in the hornblende- and kyanite-bearing layers, and about 29 percent in the pure plagioclase layers. The range in the variation of the calcium content is much larger. The bytownite rock has about 16 percent calcium oxide, the common anorthosite about 11 percent, and the dark-colored layers about 8 percent.

The occurrence of abundant banded hornblende-bearing anorthosite south of Goat Mountain is another feature that differs from the general aspect of the large bodies. This part of the body contains very little aluminum silicates, even if the country rocks are aluminous schists and a large sheetlike inclusion of schist occurs in the center. A sheetlike inclusion of amphibolite is above the schist inclusion, and the anorthosite above this amphibolite contains more hornblende than is found elsewhere. The field relations conform with the possibility of igneous origin, the basic rock occuring under the anorthosite and the anorthosite containing thin layers and lenticles of hornblende-rich rock.

The abundance of the hornblende in this gneissic anorthosite, however, can have another explanation; it may originate in the older amphibolite. The mode of occurrence of the hornblende is comparable with that of the biotite and aluminum silicates along the West Fork of Cedar Creek and along the North Fork of the Clearwater River. The only difference between the mode of occurrence of the hornblende and that of the micaceous minerals in the anorthosite is the greater regularity of the hornblende-bearing bands. This difference, however, may be a result of the more pronounced and straight platy and linear structures in the amphibolites as compared with the wavy bedding planes in the schist.

The mineralogy and texture of the hornblende-rich anorthosite at localities no. 947, 968, and 1753 are in accordance with the suggestion that the abundance of hornblende in some parts of the anorthosite may be due to the abundance of this mineral in a parent rock.

It is noteworthy that all inclusions show a metamorphism and metasomatic transformation similar to that of the country rocks. The lime-silicate rock included in the anorthosite between the forks of Cedar Creek is similar to some layers of lime-silicate rock south of the anorthosite. Abundant graphite occurs in the inclusion as well as in the main occurrence of lime-silicate rocks. No graphite, however, was found in the bytownite rock along the West Fork of Cedar Creek. The inclusions of quartzite are granoblastic and finer grained than the schist inclusions. They show feldspathization only in a fairly narrow zone (10 cm to 2 m) next to the anorthosite, just as the quartzite between Stocking Meadow Lookout and Cedar Creek does. In contrast to these fairly well-preserved inclusions of quartzite and lime-silicate rock, the inclusions of schist are strongly feldspathized. Very little quartz is found, but instead there are numerous sporadic large round grains and clusters of andesine in the inclusions just as is in the surrounding schist adjacent to the anorthosite. It seems that the schist was less resistant to the feldspathization than were the quartzite and the lime-silicate rock. Some layers of the lime-silicate rock contain about 30 percent plagioclase, but the elements to form this plagioclase were contained in the original sediment. The plagioclase occurs as interlocking grains of medium size with quartz and pyroxene. The texture is sutured or granoblastic, similar to the texture in the metamorphic rocks of this area in general.

This mode of occurrence and the texture of the primary plagioclase are in contrast to those of the large round grains of secondary andesitic plagioclase in the schist. In most of the bytownite rocks, the texture is granoblastic. Large grains of bytownite with tiny round inclusions of quartz were found in only one locality.

METASOMATIC

The third possibility includes an assumption that not only sodium but also calcium came from outside sources. Bytownite would have crystallized either from aluminum and silicon present in the sedimentary strata and calcium introduced later into it; or all elements to form bytownite were introduced. This mode of origin of bytownite could be correlated with the feldspathization of schist near Dent, about 20 miles south-southwest of the area under discussion. Near Dent, plagioclase (An_{50}) locally replaces quartz in a fairly homogeneous garnet-biotite-sillimanite schist. During the study of these rocks, the author came to the conclusion that the crystallization of andesine (An_{50}) was accompanied by an addition of calcium, sodium, and some other elements from outside sources, probably from the quartz dioritic magma that gave rise to the formation of the quartz dioritic border zone of the Idaho batholith. This conception was supported by the fact that the elements to form andesine and hornblende were found to have been mobile during the latest phase of crystallization of the quartz dioritic magma and to have caused an extensive dioritization and feldspathization of schist and quartzite. Only a few fairly thin veins of anorthosite were formed by segregation during this metasomatism. Near Dent the plagioclase of metasomatic origin has, as a rule, the same composition as the plagioclase in the quartz diorite (An_{35-50}) nearby. Secondary plagioclase around the granitic bodies is generally albite or oligoclase, thus in composition close to the plagioclase in the intrusive body that caused the metasomatism.

The intrusive rocks near the anorthosite bodies are mainly granite, quartz diorite, and quartz monzonite with plagioclase (An_{25-40}). The amphibolite bodies with labradoritic plagioclase would be better representatives of the intrusive rock that could be a source of bytownite than the rocks of the quartz diorite and quartz monzonite suite. Most of these amphibolite bodies, however, contain garnet, biotite, and quartz in large quantities, and are probably themselves products of metasomatic replacement. No suitable igneous source rock for the bytownite is thus exposed, but we might assume that a magma capable of sending large amounts of elements of bytownite could exist underneath the present erosion surface. This same hypothetical magma could have been a source for the iron and magnesium that accumulated around the anorthosite bodies to form garnet amphibolite. A later introduction of sodium would, as in the other cases, be responsible for the crystallization of andesine at the expense of earlier bytownite.

The ratio between the amounts of ferromagnesian minerals and bytownite in the anorthosite-amphibolite group would suggest that the composition of the hypothetical magma below the present surface should have been of gabbroic-anorthositic composition. Buddington (1939) suggested this composition for the parent magma of the anorthosites in general, and later (Buddington, 1943) suggested that this magma type would have a global extension just above the basaltic layer. Also Michot (1956, 1957) has suggested that an anorthositic layer would have a global extension underneath the granitic layer. According to him, the formation of the anorthositic magma would be a result of an assimilation of pelitic sediments by basaltic magma. This assimilation would reverse the order of the crystallization so that plagioclase would crystallize earlier than the ferromagnesian minerals. Large amounts of anorthosite would be formed through anatectic processes, principally by selective elimination of excess of ferromagnesian minerals from leuconoritic material. The upper limit of the front of such an anorthositization would represent the discontinuity found by seismic methods under the granitic layer of the continents. This discontinuity (called Sical by Michot for principal ingredients, Si, Ca, and Al) would be at the depth of about 20 km during the time of its formation and could be exposed in tectonically suitable places by a later erosion. At this depth, anorthosite would be formed contemporaneously through metasomatic, anatectic, and magmatic processes. The ferromagnesian minerals would be removed by selective resolution in eutectic ratios and would be precipitated around the residual anorthosite body, forming a basic front to which the ilmenitic iron ores would belong.

It is possible that in the area under discussion, the garnet amphibolite around the anorthosite bodies is a result of such a selective removal of ferromagnesian minerals from the area now occupied by anorthosite, but it seems less likely that the anorthosite itself is a modified igneous rock. The formation of the garnet amphibolite is essentially earlier than the crystallization of andesine and is probably connected with the formation of bytownite. The evidence in favor of a modified sedimentary origin of most of the bytownite rocks found in the present area is more weighty than that for such a modified igneous and metasomatic origin as suggested by Michot (1956, 1957) for the Egersund area and for anorthosite in general. Only the small occurrence of gabbroic anorthosite (no. 1747, table 11) that occurs in a fine-grained mica schist southeast of Orphan Point and is shelled by a layer of coarse-grained chlorite-hornblende-garnet rock may have attained its present composition by elimination of a part of the iron and magnesium from an igneous rock. Elsewhere some iron and magnesium may have been removed from the metasedimentary strata in which bytownite crystallized during the metamorphism. A more detailed discussion of the rearrangement of material around the anorthosite bodies will give a clearer picture of the processes leading to the formation of anorthosite in Idaho.

REARRANGEMENT OF MATERIAL AROUND THE ANORTHOSITE BODIES

There is no doubt that a considerable rearrangement of material was necessary during the formation of the anorthosite under discussion. Short-distance migration of calcium and aluminum was pointed out in the discussion of a sedimentary mode of origin. The occurrence of abundant aluminum silicates in many layers of the two southern bodies suggests a metasomatic origin at least for a part of the bytownite. The crystallization of bytownite in the aluminum-rich sediment would require an addition of calcium, the source of which could be a sedimentary rock nearby. Furthermore, the bytownite rock layers, such as those found along the West Fork of Cedar Creek, probably crystallized from calcareous shale, acquiring additional aluminum from the clayey layers nearby. These calcareous layers were poor in iron and magnesium, as shown by the scarcity of diopside and hornblende in all bytownite rocks; or else a part of the iron and magnesium was removed during the metamorphism. The reaction may have proceeded as follows:

$$\begin{array}{llll} 4\mathrm{CaCO_3}+\mathrm{NaAlSi_3O_8}+8\mathrm{SiO_2} & +4\mathrm{Al^2O^3} & \longrightarrow 5(\mathrm{Ab_{20}An_{80}}) & +4\mathrm{CO_2} \\ \underbrace{\text{Calcite }+\text{albite}\quad+\text{quartz}}_{\text{from calcareous shale}} & +\begin{array}[t]{l}\text{aluminum}\\ \text{from the}\\ \text{clayey}\\ \text{layers}\\ \text{nearby}\end{array} & \rightarrow \text{bytownite} & +\begin{array}[t]{r}\text{carbon}\\ \text{dioxide}\end{array} \end{array}$$

The total composition of the present anorthosite is about $Ab_{40}An_{60}$, one-third of its plagioclase consisting of bytownite (An_{80}) and two-thirds of andesine (An_{50}). Thus, two-thirds of the original bytownite may have been changed to andesine by a later sodium metasomatism according to the following reaction:

$$\begin{array}{llll} 5(\mathrm{Ab_{20}An_{80}})+4\mathrm{SiO_2} & +\mathrm{Na_2O} & \longrightarrow 6(\mathrm{Ab_{50}An_{50}}) & +\mathrm{CaO} \quad (2) \\ \underbrace{\text{bytownite }+\text{quartz}}_{\text{in metasediment}} & +\begin{array}[t]{l}\text{sodium-}\\ \text{rich solutions}\end{array} & \rightarrow \text{andesine} & +\begin{array}[t]{l}\text{calcium-}\\ \text{rich}\\ \text{solu-}\\ \text{tions}\end{array} \end{array}$$

The calcium released in this reaction probably was the source for the calcium introduced into aluminum-rich layers nearby where it reacted forming bytownite according to the following equation:

$$4\mathrm{Al_2O_3}+8\mathrm{SiO_2}+\mathrm{NaAlSi_3O_8}+4\mathrm{CaO}\longrightarrow 5(\mathrm{Ab_{20}An_{80}})$$

Further introduction of sodium would give rise to the formation of andesine at the expense of this bytownite according to reaction (2) and the resultant rock would have a composition and structure of the aluminum-rich border phase that is common in the southernmost anorthosite body and in the southern part of the middle body. The layering in these parts of the anorthosite would be a relict structure inherited from the original aluminum-rich sedimentary rock.

The calcium released during the formation of andesine in the layered border phase would migrate into the neighboring schist layers and be combined there with aluminum, silicon, and introduced sodium to form oligoclase-andesine. In this way, the calcium would gradually migrate from the calcareous layers and masses into the surrounding aluminum-rich layers, forming first bytownite and later, during the metasomatic introduction of sodium, andesine in the inner zone next to the anorthosite and oligoclase in the outer zones. Thus it seems that the introduction of sodium gave rise to the recrystallization and rearrangement of earlier minerals and gave to the anorthosite and its country rocks their present composition and texture.

The recrystallization and homogenization have proceeded farthest in the massive parts of the large anorthosite bodies. The heterogeneity becomes conspicuous toward the border zones because of two reasons: presence along the borders of more remnants of the original sedimentary rocks, and the regrouping of ions during the formation of the anorthosite. Owing to the regrouping and segregation, the salic and femic constituents of the border zone crystallized in different layers, giving rise on one hand to lens-shaped and layerlike bodies of pure plagioclase, and on the other hand, to segregations of biotite and hornblende with or without garnet. This differentiation may have started during the crystallization of bytownite, in which case it is comparable with the metamorphic differentiation as suggested by Eskola (1932). However, the major part of all rearrangements in this area seems to be a result of sodium metasomatism and the differentiation should be more properly termed as "metasomatic differentiation."

A tendency to a metasomatic differentiation on a large scale is clearly manifested in the country rocks in a zone, 1 to 3 miles wide, outside the anorthosite bodies. The extraordinarily large amount of aluminum silicates and plagioclase in the schist found only around the anorthosite bodies suggests a genetic relationship. The abundance of plagioclase in this contact zone can in part be tied to the introduction of sodium. It remains uncertain whether all of the aluminum was contained in the original sediment or whether some of it was introduced with the sodium. Regardless of its original source, the aluminum shows a definite concentration in the schist next to the anorthosite. Moreover, this concentration is not confined to beds but follows closely the same pattern as does the feldspathization. The fact that the same zone is impoverished in potassium and silicon suggests a considerable exchange of material. In the outermost parts of the contact aureole, where biotite is replaced by sillimanite, iron is precipitated as magnetite, but potassium and magnesium are removed from the system according to the following equation:

$$\begin{array}{l} H_4K_2Fe_3Mg_3Al_2Si_6O_{24} \longrightarrow Al_2SiO_5 + Fe_3O_4 + 5SiO_2 + \underbrace{K_2O+3MgO+H_2O+H_2}_{\text{removed in solution}} \\ \text{biotite} \longrightarrow \text{sillimanite}+\text{magnetite}+\text{quartz}+ \end{array}$$

In the normal regionally metamorphosed rocks the reaction leading to the formation of cordierite and potassium feldspar from the biotite and muscovite is common (Eskola, 1939, p. 354). In the area under discussion, cordierite is a common additional constituent, but potassium feldspar is absent in the schist. Crystallization of plagioclase instead of potassium feldspar suggests an exchange of alkalies subsequent to the reaction that led to the replacement of biotite and muscovite by sillimanite and cordierite. The orthoclase in the dike found near the anorthosite south of Boehls (table 9) may represent a segregation of the potassium that was removed from the surrounding feldspathized schist. The magnesium removed during the replacement of biotite by sillimanite was precipitated in the schist nearby, where it formed magnesium-rich biotite and hornblende.

The secondary plagioclase in the schist around the anorthosite bodies contains a considerable amount of calcium, and a question arises as to whether this calcium was originally present in the sediment or introduced with the sodium. Most of the schist in the Prichard formation contains very little calcium, and there may have been a metasomatic addition in the parts that carry plagioclase. This suggestion is strengthened by the occurrence of abundant calcium mica, margarite, instead of sillimanite in some of the schist of this area. The mode of occurrence and the appearance of the margarite are similar to those of sillimanite; so margarite may be a pseudomorph after sillimanite. The reaction,

$$2\ \text{sillimanite}+CaO+H_2O \longrightarrow \text{margarite},$$

requires an addition of calcium and suggests that the calcium was mobile during the recrystallization. The reaction sillimanite⟶margarite also requires an addition of water and indicates (together with its occurrence next to a pegmatitic dike) that the calcium may have been introduced in water solutions. When calcium and sodium were introduced into the schist, they would form andesine with aluminum silicates and quartz, according to the following reaction:

$$2Al_2SiO_5+CaO+Na_2O+6SiO_2 \longrightarrow 2NaAlSi_3O_8+ CaAl_2Si_2O_8.$$

This reaction would use much of the aluminum present in the schist, and the amount of aluminum silicates in the feldspathized schist should be less than it is in the normal schist; instead, the feldspathized parts of the schist contain more aluminum silicates than the normal schist in the same layer. Therefore it seems likely that at least a part of the aluminum in the contact zone was introduced metasomatically.

The source of this excess aluminum could be either the aluminum-rich layers of the sediments transformed to anorthosite or an outside source, the same as that of the introduced sodium.

TEMPERATURE AND PRESSURE DURING THE FORMATION OF ANORTHOSITE

On the basis of the above discussion, it is clear that in the formation of anorthosite two phases have to be considered: (1) metamorphism leading to the formation of bytownite in the anorthosite; (2) metasomatism leading to the crystallization of andesine in the same rock. The temperature-pressure conditions during the crystallization of bytownite in a shaly limestone with aluminum-rich clayey layers were those of amphibolite facies. The typical mineral assemblages were bytownite and diopside in limy layers and kyanite and biotite in clayey layers. Crystallization of kyanite instead of andalusite is in accordance with the conclusion that metamorphism took place during the period of folding.

The crystallization of andesine has obscured the foliation and bedding of the metamorphic rocks in many places; this obscuring indicates that andesine crystallized after the deformation. The inversion of kyanite to andalusite probably is due to a change from kinematic to static conditions. It is not necessary to assume any change in temperature during this inversion (Hietanen, 1956). Crystallization of hornblende instead of diopside during this phase may be due merely to the presence of water in the sodium-rich solutions that altered much of the bytownite to andesine. The occurrence of sillimanite with muscovite proves, according to the experiments made by Yoder and Eugster (1955), that the temperature did not exceed 650 °C. Michot (1956, 1957) places the upper limit of anorthositization (discontinuity sical) to the depth of 20 km and assumes temperatures of 660°C and higher. According to Adams (1924) the temperature at this depth would be between 450°C and 500°C, thus lower than estimated by Michot.

Barth (1956) has shown that the ratio of the distribution of the albite molecule between potassium feldspar and plagioclase crystallized in equilibrium may serve as a geologic thermometer. Using his method, one can estimate the temperature of formation of the orthoclase dike near Boehls (table 9). In this dike, plagioclase of composition $Ab_{78}An_{22}$ crystallized with orthoclase $Or_{88}Ab_{12}$, which gives the ratio of distribution k=0.15 and a temperature of about 420°C.

The occurrence of epidote with labradorite (An_{55}) in some parts of the anorthosite would suggest temperatures around 430°C (Ramberg, 1949).

The varying rate of homogenization of the two plagioclases (andesine and bytownite) in different parts of the anorthosite bodies suggests that the physico-chemical conditions were locally approaching those necessary for the formation of true anorthosite with only one kind of plagioclase, labradorite (An_{55-58}). It is also possible that the time factor plays an important role in this respect and that in all parts of the area, labradorite anorthosite would have been formed instead of andesine-bytownite rock provided that there was enough time for the reaction leading to the homogenization. In either case, the present area provides an example of formation of metamorphic-metasomatic anorthosite and of the upper limit of metasomatic anorthositization. The next step after the homogenization would be a partial remelting and mobilization that could result in an intrusion of anorthosite.

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Metamorphism of the Belt Series in the Elk River-Clarkia Area Idaho

By ANNA HIETANEN

METAMORPHIC AND IGNEOUS ROCKS ALONG THE NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

GEOLOGICAL SURVEY PROFESSIONAL PAPER 344-C

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METAMORPHIC AND IGNEOUS ROCKS ALONG THE NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

METAMORPHISM OF THE BELT SERIES IN THE ELK RIVER-CLARKIA AREA, IDAHO

By Anna Hietanen

ABSTRACT

The Elk River–Clarkia area lies from 10 to 35 miles northwest of the Idaho batholith in a zone of moderately to highly metamorphosed sedimentary rocks of the Precambrian Belt series. The Prichard formation, which is the oldest unit in the Belt series, covers the northeastern part of the area and consists mainly of garnet-mica schist with a minor amount of quartzite. It is separated by a fault zone from the younger formations. The massive Revett quartzite is fragmentarily exposed along and near this fault zone and is overlain by a fairly thin layer of schist of the St. Regis formation. The youngest formation, the Wallace, covers the western and southern part of the area. It consists of aluminous schist and interbedded calcareous gneiss and quartzite. The Burke formation, which should occur between the Prichard and Revett formations, could not be identified in this area.

Two large and a number of small bodies of igneous rocks occur in the metasedimentary series. One of the large bodies is quartz diorite and the other is quartz monzonite. The composition of the small bodies ranges from gabbro to quartz diorite, quartz monzonite, and granite.

The grade of metamorphism of the Belt series increases gradually from northwest to southeast parallel to the major structural trend and suggests that the metamorphism is related to the intrusion of the Idaho batholith, the nearest outcrops of which are exposed about 10 miles southeast of the area under discussion. The critical mineral assemblage in the schists in the northern part is almandine–muscovite–biotite–quartz; in the calcareous layers of the gneisses the common assemblage is hornblende–biotite–epidote minerals–albite–microcline–quartz with or without scapolite and calcite. These mineral assemblages are typical of the biotite–almandite subfacies of the epidote amphibolite facies. Toward the south, kyanite appears with staurolite in the aluminum-rich layers of the garnet-mica schist and diopside becomes stable in the calcareous layers of the gneissic units. Oligoclase instead of albite occurs with the epidote minerals. Microcline abounds in many thin biotite-rich layers interbedded with diopside and biotite gneisses. These mineral associations suggest metamorphic temperatures of the staurolite-kyanite subfacies of the epidote amphibolite facies.

The occurrence of sillimanite instead of staurolite and kyanite in the schist in the southern part of the area suggests the sillimanite-muscovite subfacies of the amphibolite facies. Grossularite occurs with calcite in a few calcareous layers, but the assemblage diopside–biotite–plagioclase (An_{30}) is most common. Metamorphic temperatures are estimated to rise from 380°C in the northern border to 570°C in the southern border.

Contrary to the normal distribution of the potassium feldspar in the regionally metamorphosed area, microcline is rare or lacking in the gneisses of the southernmost part—that is, in the zone of highest grade of metamorphism. This lack of microcline can be either depositional or due to a later partial removal of potassium from a contact zone of the Idaho batholith about 20 miles wide. In the light of the regional geology the latter possibility seems more likely and would suggest that potassium was mobilized in a wider contact aureole than the other elements (calcium, aluminum, iron, and magnesium). It was removed from the inner contact aureole, which is the zone of metasomatic enrichment of basic components (Hietanen, 1962), and probably also from the southern part of the area under discussion, which part belongs to an outer contact aureole of the batholith. No other traces of regional metasomatism could be found in this outer zone.

INTRODUCTION

This report is a petrologic and structural study of the metamorphic and igneous rocks in an area outside of the immediate contact zone of the Idaho batholith. The area lies from 10 to 35 miles northwest of the batholith (fig. 1). In this outer zone, metasedimentary rocks belonging to the Precambrian Belt series show a moderate to high degree of metamorphism, but they show little if any of the kind of profound metasomatism that was found in the zone next to the batholith, where quartz is locally replaced by andesine and hornblende (Hietanen, 1962). However, some of the gneissic units have a definite and regular lateral change in composition parallel to the bedding.

The Belt series in the Coeur d'Alene district and in western Montana has been described as "a monotonous sequence of sombre-colored, rather fine-grained clastic rocks" (Ross, 1956, p. 687) consisting mainly of great thicknesses of quartzite and containing only insignificant amounts of carbonates and other normal miogeosynclinal sediments. In contrast, the corresponding depositional units in the present area were metamorphosed to distinctive rock types of varied mineralogy and composition, containing abundant kyanite,

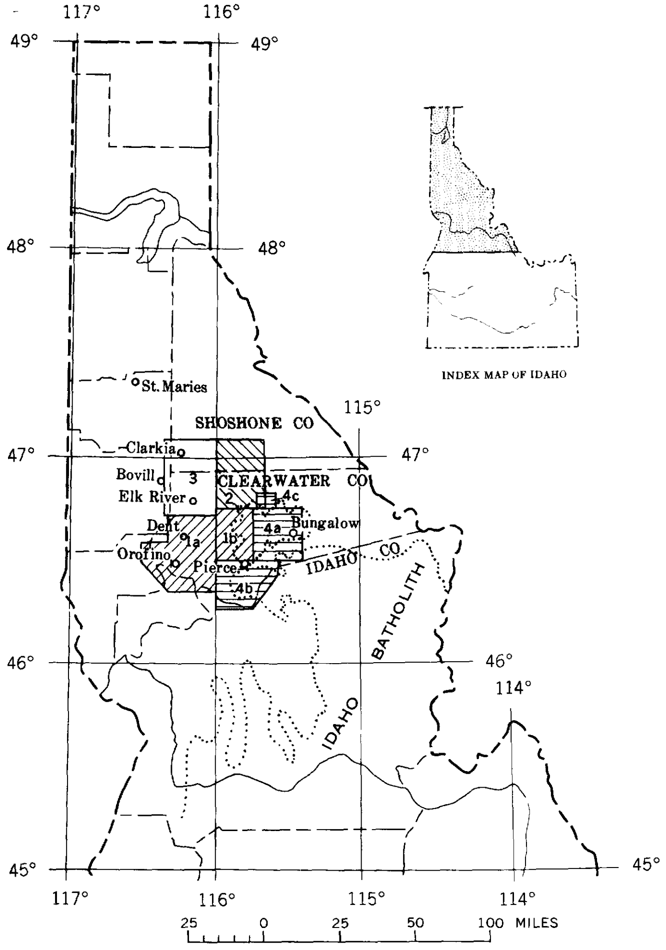


FIGURE 1.—Index map of northern Idaho. 1a, Orofino, Dent, and Big Island areas; 1b, Headquarters quadrangle (Hietanen, 1962); 2, Boehls Butte quadrangle and vicinity (Hietanen, 1963); 3, Elk River–Clarkia area, this report; 4a, Bungalow area; 4b, Pierce area; 4c, Beaver Creek area. The dotted line shows the outline of the northern part of the Idaho batholith.

sillimanite, garnet, biotite, and muscovite in the original shaly layers (now coarse-grained schist) and much hornblende, diopside, tremolite, biotite, and calcite in the dolomitic sand layers (now fine- to medium-grained gneiss). Only a few pure quartzite layers are interbedded with the schist and gneiss. In the gneissic units the variation in the amounts of the major constituents—quartz, plagioclase, microcline, diopside, hornblende, tremolite, and biotite—makes the layering distinct and gives rise to identifiable, fairly persistent subunits.

Distribution of potassium feldspar in some of these subunits shows a regular lateral variation and offers a special problem concerning its origin. Two possibilities are considered: that the variation in the amount of potassium feldspar is depositional, or that it is metasomatic. The field evidence favors the first possibility for the northern part of the area. Comparison with the earlier studies in the zone closest to the batholith (Hietanen, 1962) suggests that the lack of the potassium feldspar in the southern part may be due to the removal of potassium during metasomatic metamorphism.

AREA STUDIED

The area studied lies in the northwestern part of Clearwater County and the extreme southwestern part of Shoshone County, Idaho (fig. 1). It is bound by long 116°00′ and 116°20′ W. and lat 46°42′ and 47°05′ N. and comprises 420 square miles. Orofino, Dent, and big Island areas and Headquarters quadrangle, and Boehls Butte quadrangle and vicinity were previously mapped by the author (Hietanen, 1962, 1963, respectively).

The area lies just east of the Columbia River Plateau and is characterized by deep stream valleys which are separated by narrow divides. The southern part of the area is drained southward to the North Fork of the Clearwater River, and the northern part northward to the St. Joe River mainly by the St. Maries River and its tributaries. The altitude of the stream valleys near Elk River is 2,900 feet and near Clarkia is 2,850 feet. The highest mountains rise 6,000 feet above sea level. The local relief ranges from 1,500 feet in the western part to 3,000 feet in the eastern part.

A good gravel road connects Clarkia and St. Maries, a town 20 miles to the northwest. Another gravel road leads from Clarkia to Bovill, a small logging town 18 miles to the south-southwest. Elk River is connected to Bovill by a gravel road and a 20-mile-long dirt road leads from Elk River to Dent, on the North Fork of the Clearwater River 5 miles south of the area studied. Dent is in turn connected to Orofino and other points to the west. Narrow dirt roads lead eastward from Elk River and Clarkia to the Boehls Butte quadrangle. Many logging roads and dry season jeep roads (pl. 1) make the area fairly accessible. The mountain areas are crossed by a few trails, following mainly ridges and streams.

Most of the area is heavily timbered by white and yellow pine, fir, cedar, hemlock, spruce, and larch. Locally in the northeastern part, however, forest fires have destroyed virgin timber and a second growth of small- to medium-size trees has reforested only the drainage valleys and lower mountain slopes, the higher surfaces being covered by brush, grass, and scattered small trees. In all forested parts of the area the soil cover is thick and accordingly the outcrops are poor.

FIELDWORK

Previous work has been restricted to reconnaissance by Anderson (1930), who recognized the Belt series covering most of the area, a granitic stock north of Elk River, and the Columbia River basalt south and southwest of this town.

Fieldwork for this report was done intermittently during the summers of 1952 to 1955 and 1957. The author was assisted in the field by Dorothy Rainsford and Cynthia Wilkin in 1952, by Katharine Wagner and Patty Newbold in 1953, and by Alice Vrbsky and Frances Gilbert in 1955.

As the main purpose was to study metamorphism, detailed mapping was not attempted. Rather, in all forested parts of the area, where soil cover is thick and bedrock is rarely seen elsewhere except on the roadcuts, the fieldwork consisted mainly of study of the outcrops and float along the roads, shown on plate 1, and along additional logging roads that branch off from these. In the northeastern part of the area, where ridges are barren and logging roads follow streams, the mapping consisted of rather widely spaced ridge and stream traverses, easily identified on plate 1 by the concentration of structure symbols. Geologic relations between traverses are largely interpretive, and plate 1 is therefore to be regarded as detailed reconnaissance.

SPECIMEN AND LOCALITY NUMBERS

Specimen numbers are used as locality numbers and shown on the geologic map (pl. 1). The following gives township, range, and section for each locality.

Number	*Township*	*Range*	*Sec.*
255	T. 39 N.	R. 4 E.	18
266	T. 39 N.	R. 3 E.	24
557	T. 39 N.	R. 2 E.	24
560	T. 39 N.	R. 2 E.	24
566	T. 39 N.	R. 3 E.	9
567	T. 39 N.	R. 2 E.	12
1183	T. 40 N.	R. 2 E.	9
1190	T. 41 N.	R. 2 E.	33
1246	T. 41 N.	R. 2 E.	20
1248	T. 41 N.	R. 2 E.	20
1251	T. 41 N.	R. 2 E.	29
1260	T. 41 N.	R. 1 E.	25
1261	T. 41 N.	R. 1 E.	36
1262	T. 41 N.	R. 1 E.	35
1428	T. 41 N.	R. 2 E.	24
1429	T. 41 N.	R. 2 E.	12
1431	T. 41 N.	R. 2 E.	10
1433	T. 41 N.	R. 2 E.	8
1435	T. 41 N.	R. 2 E.	8
1437	T. 41 N.	R. 1 E.	14
1438	T. 40 N.	R. 1 E.	1
1441	T. 40 N.	R. 1 E.	12
1443	T. 40 N.	R. 3 E.	4
1444	T. 40 N.	R. 3 E.	5
1445	T. 41 N.	R. 3 E.	27
1468	T. 40 N.	R. 3 E.	14
1520	T. 41 N.	R. 3 E.	9
1521	T. 41 N.	R. 3 E.	9
1524	T. 42 N.	R. 3 E.	32
1525	T. 42 N.	R. 2 E.	14
1527	T. 42 N.	R. 2 E.	14
1528	T. 42 N.	R. 2 E.	14
1610	T. 42 N.	R. 2 E.	11
1613	T. 42 N.	R. 2 E.	11
1615	T. 42 N.	R. 2 E.	10
1616	T. 42 N.	R. 2 E.	10
1623	T. 42 N.	R. 2 E.	4
1628	T. 43 N.	R. 2 E.	33
1630	T. 43 N.	R. 2 E.	23
1677	T. 43 N.	R. 3 E.	27
1678	T. 42 N.	R. 3 E.	5
1789	T. 43 N.	R. 2 E.	[1] 4
1841	T. 43 N.	R. 2 E.	18
1842	T. 42 N.	R. 1 E.	3
1848	T. 42 N.	R. 1 E.	15
1851	T. 42 N.	R. 1 E.	[2] 9
1885	T. 42 N.	R. 1 E.	3
1886	T. 42 N.	R. 1 E.	3
1889	T. 42 N.	R. 1 E.	3
1904	T. 42 N.	R. 2 E.	5
1906	T. 43 N.	R. 2 E.	33
1908, 1909	T. 43 N.	R. 2 E.	23
1910	T. 43 N.	R. 2 E.	13
1912	T. 43 N.	R. 2 E.	14
1913	T. 43 N.	R. 2 E.	14
1915	T. 43 N.	R. 2 E.	23
1917	T. 42 N.	R. 2 E.	11
1919	T. 41 N.	R. 1 E.	2
1920	T. 41 N.	R. 1 E.	1
1940	T. 40 N.	R. 2 E.	31
1992	T. 39 N.	R. 3 E.	14
2005	T. 40 N.	R. 3 E.	20
2008	T. 40 N.	R. 3 E.	14
2012	T. 40 N.	R. 3 E.	5

[1] North of pl. 1.

[2] West of pl. 1.

CLASSIFICATION OF THE ROCKS

The rocks of the area arranged from oldest to youngest are grouped as follows:

1. Metasedimentary rocks of the Precambrian Belt series.
2. Plutonic outliers of the Idaho batholith, mainly quartz diorite, quartz monzonite, and granite of Cretaceous age.
3. Dikes and sills of gabbroic and granitic composition, of Tertiary(?) age.
4. Columbia River basalt of Miocene and Pliocene(?) age.

BELT SERIES

In the southern and eastern part of the Elk River–Clarkia area (pl. 1) the rocks equivalent to the Belt series are highly metamorphosed, but in the northwest the grade of metamorphism is somewhat lower, and in places sedimentary structures are preserved, such as crossbedding in the quartzitic layers of the Wallace formation. However, most of the sedimentary structures and colors that characterize the Belt series in the Coeur d'Alene district and elsewhere were obliterated during metamorphism, and the correlation must be based mainly on lithology. A sequence of thick beds of coarse-grained pure quartzite, equivalent to the Revett quartzite in lithology and stratigraphic position, was used as a key map unit. A layer of schist overlying this quartzite unit was mapped as St. Regis formation, and interbedded gneiss, quartzite, and schist above this schist were considered to be equivalent to the Wallace formation.

The Burke formation, which should underlie the Revett quartzite, could not be identified in the area. In the exposures found, the quartzite unit assigned to the Revett quartzite is separated by a fault from schist mapped as the Prichard formation. In some localities of poor exposures the quartzite unit seems to overlie schist conformably. Such a relation was found around Stony Butte and north of Breakfast Creek along the eastern edge of the area. In both localities, however, the underlying rock is a coarse-grained mica schist dissimilar to the Burke formation near Surveyors Ridge, where Umpleby and Jones (1923) mapped all the lowest five formations of the Belt series in a conformable sequence, and also dissimilar to the Burke formation in the Coeur d'Alene district (Ransome and Calkins, 1908). It seems, therefore, that either the Burke formation was not deposited here or the equivalent beds are included in the lowest part of the Revett quartzite or in the uppermost part of the Prichard formation.

PRICHARD FORMATION

DISTRIBUTION AND LITHOLOGY

Highly metamorphosed schist, gneiss, and quartzite exposed in the northeastern part of the area are continuous with the Prichard formation north of plate 1, as mapped by Pardee (1911) and by Umpleby and Jones (1923), who show the Prichard formation faulted against the Wallace formation to the west in the northeastern part of T. 43 N., R. 2 E. Several northward-trending faults and large faulted blocks of the Revett quartzite occur along the border of sections 12 and 13 of this township, just east of the southern end of the fault mapped by Umpleby and Jones. The fault zone that extends for about 5 miles southward from section 13 and then turns to a southeasterly direction separates the major area of schist of the Prichard formation from the overlying Revett quartzite and from the rocks equivalent to the St. Regis and the Wallace formations.

Small areas of coarse-grained mica schist that underlie conformably the Revett quartzite south of White Rock and north of Stony Butte and Breakfast Creek were mapped as the Prichard formation, because the schist is petrographically similar to the major part of the Prichard formation and no division could be made between it and the underlying beds. The major part of the rocks of the Prichard formation in the map area consists of coarse-grained garnet-mica schist with minor amount of quartzite and biotite-plagioclase gneiss.

Numerous lenses of amphibolite and garnet amphibolite occur in the schist and gneiss. Some lenses may be sills or products of metasomatic processes, but at least some are metamorphosed lenses of calcareous shale deposited as part of the Prichard formation.

A discontinuous layer of micaceous quartzite, as much as 100 m thick, is interbedded with the schist on the southwest side of Glover Creek.

STRATIGRAPHY

The schist and gneiss of the Prichard formation in the Elk River–Clarkia area lie beneath a thick quartzite unit, which, north of this area, occurs in the upper part of the Prichard formation. This quartzite is well exposed on Lookout Mountain, 2 miles east of the area (Pardee, 1911). Therefore, most of the schist and gneiss under discussion probably are equivalent to beds that in northern Shoshone County form the lower part of the Prichard formation. The micaceous quartzite in the syncline southwest of Glover Creek may be stratigraphically equivalent to the quartzite on Lookout Mountain. Owing to structural complexities the thickness of the exposed Prichard could not be measured; it is probably several hundred meters.

Layers and sill-like bodies of coarse-grained gneiss and small masses of garnet amphibolite are irregularly scattered through the schist.

PETROGRAPHY

SCHIST

Most of the schist of the Prichard formation is coarse to medium grained. Weathered surfaces are bronze colored; fresh surfaces, bluish gray. Numerous thin fine-grained quartzitic layers are interbedded with coarse-grained micaceous layers. The amount of the major constituents—quartz, plagioclase, biotite, muscovite, and garnet—vary markedly in the schist. Many plagioclase-rich layers contain hornblende.

The fine-grained layers are rich in quartz (60 to 80 percent) and contain some oligoclase; biotite is the main micaceous mineral. Red almandine garnet, tourmaline, magnetite, zircon, and rutile are minor constituents.

The coarse- to medium-grained layers contain 50 to 60 percent quartz and about 25 percent muscovite in large flakes; oligoclase is more abundant than in the fine-grained layers, constituting 1 to 20 percent of the rock. Garnet crystals range from 2 to 20 mm in diameter and constitute about 10 percent of the rock. They are subhedral to round or oval and include numerous small round quartz grains. Biotite is either interleaved with muscovite or in separate thin laminae made of flakes 1 to 2 mm in diameter. Magnetite, zircon, tourmaline, and corundum are accessories. The amount of biotite increases and that of muscovite decreases toward the south.

The coarse-grained garnet-mica schist around Stony Butte consists of quartz (30 to 50 percent), plagioclase (2 to 20 percent), muscovite (10 to 20 percent), biotite (about 10 percent), and garnet. In addition to these major constituents many layers contain 1 to 2 percent of kyanite and tiny subhedral crystals of staurolite. Apatite, tourmaline, and zircon occur as accessories. Shiny prisms of black tourmaline, 1 cm long and 4 mm thick, were found in some strongly sheared layers about a mile east of Stony Butte. This tourmaline is dravite with $\omega=1.638\pm0.002$.

In many places, fine-grained quartzitic layers, 1 to 10 cm thick, are interbedded with the schist. In some of these layers (for example, loc. 1677, southeast of Grandmother Mountain) are lenses as much as 2 m long which contain zoisite, indicating that the strata were slightly calcareous.

GNEISS

The schist unit contains many medium- to coarse-grained gneissic layers and lenses that grade locally to coarse-grained gneissic quartz diorite or to gneissic tonalite and gray granite. The gneisses have abundant oligoclase, very little if any muscovite, and much less quartz than the average schist. Veins and sill-like bodies of pegmatite are common. The major constituents of pegmatite are plagioclase and quartz; biotite, garnet and (or) hornblende are minor.

Near several small bodies of granite and quartz diorite north of Glover Creek and south of Grandmother Mountain gneiss makes up about 40 percent of the Prichard formation. Elsewhere the amount of this rock type is much less (about 15 percent).

Plagioclase (An_{25-30}) constitutes about 60 percent of the gneiss, quartz about 20 percent, and biotite 10 to 15 percent. The remaining 5 to 10 percent is hornblende, garnet, and epidote in varied amounts. The accessories are apatite, magnetite, zircon, and locally tourmaline. There is every gradation from these gneissic layers to a schist in which the amount of quartz is about 60 percent and in which plagioclase is scarce or lacking.

A very coarse grained gneiss, rich in plagioclase and containing large crystals of garnet and of blue kyanite, crops out just west of Freezeout Mountain. The garnets range from ½ to 3 cm in diameter; many are enveloped by biotite. This garnet-kyanite gneiss is similar to the plagioclase-rich country rock of the anorthosite in the Boehls Butte quadrangle (Hietanen, 1963).

About a mile west of Freezeout Mountain anthophyllite prisms, 2 to 6 mm long, occur in plagioclase-rich layers, about 20 cm thick, in the schist. The indices of refraction of the anthophyllite are $\alpha=1.659\pm0.001$, $\beta=1.670\pm0.002$, $\gamma=1.681\pm0.001$; 2V is large. These optical properties are close to those of the aluminous anthophyllite described by Rabbitt (1948) from the Cherry Creek area, Montana.

QUARTZITE

The quartzite interbedded with the schist of the Prichard formation on the southwest side of Glover Creek is a light-gray medium-grained rock in which shiny muscovite flakes, ¼ to 2 mm in diameter, are evenly scattered throughout most beds.

Most of the quartz grains are flattened in the plane of foliation. In this plane they are 0.2 to 0.3 mm in diameter and almost equidimensional. The orientation of the muscovite flakes suggests a weak northeastward-striking lineation; and about 80 percent of the quartz grains have their *c*-axes parallel to the same direction.

GARNET AMPHIBOLITE

Several elongate bodies of dark well-foliated amphibolite and garnet amphibolite occur in the schist of the Prichard formation. Some of them are probably sedimentary in origin. Many, however, are metamorphosed sills and dikes and some are metasomatic. Because all varieties are petrologically alike and the origin of individual bodies is uncertain, all the amphibolite and garnet amphibolite are described in the section on plutonic rocks.

REVETT QUARTZITE

DISTRIBUTION AND LITHOLOGY

A sequence of thick beds of coarse-grained nearly pure quartzite is assigned to the Revett quartzite. Easy to identify in the field, the unit was used as a key in mapping. In most places the lower beds of the Revett quartzite are faulted against the Prichard formation or the Wallace formation. Where conformable contacts are indicated, biotite schist, possibly equivalent to the Burke formation but inseparable from the schist of the Prichard formation, underlies the Revett quartzite.

The Revett quartzite crops out in and near the fault zone that passes through the area from the northern border to the south and southeast. Widest distribution is along Breakfast Creek in T. 40 N., R. 3 E., and T. 41 N., R. 3 E.

STRATIGRAPHIC SEQUENCE AND THICKNESS

The stratigraphic sequence is interrupted by numerous faults, many of which occur along the contact of the massive Revett quartzite or cut across its beds. Because of this faulting and the lack of outcrops, the complete section is not exposed and the thickness could not be measured.

Nearly complete sections of the Revett quartzite are exposed along Breakfast Creek and to the south. There the lowest beds are light bluish gray; they are overlain by thick beds of white quartzite, which grades to pink or light brownish red quartzite toward the top. The thickness of this section is estimated to be about 500 m, which is close to the thickness of the Revett quartzite in southern Shoshone County (Umpleby and Jones, 1923).

PETROGRAPHIC DESCRIPTION

The Revett quartzite is coarse grained and has splintery fracture. Individual grains in the splinters have shiny surfaces that resemble cleavage planes but are not as smooth. These shiny surfaces are typical only of quartzites of the Revett and are probably due to the deformation and orientation of quartz. They are not found in vein quartz, the blocks of which otherwise resemble the coarse-grained Revett quartzite. The quartzites of the Prichard and Wallace formations also are easily distinguished from the Revett quartzite because their quartz grains are small, rounded, and dull.

Beds of nearly pure quartzite, 5 cm to 1 m thick, are separated by paper-thin muscovite laminae. In a few places one or two layers of schist, 2 to 10 m thick, are interbedded with the coarse-grained quartzite. The beds of quartzite consist of 95 to 99 percent quartz and a few scattered grains of plagioclase, biotite, and muscovite. The plagioclase, biotite, and muscovite occur between irregularly shaped strained grains of quartz, as do a few grains of rutile, zircon, and magnetite. Some large grains of quartz show shadowy mosaic texture, which suggests that many tiny grains were compined to form each large grain. Parallel orientation of the *c*-axes and strong strain shadows indicate that recrystallization occurred during deformation.

ST. REGIS FORMATION

DISTRIBUTION AND LITHOLOGY

A layer of schist that separates the Revett quartzite from the overlying diopside- and hornblende-bearing gneiss and quartzite of the Wallace formation is assigned to the St. Regis formation. This division is applicable in the central part of the area, but locally in the southeastern part the schist equivalent to the St. Regis formation could not be identified with certainty, because the lowest gneiss-quartzite unit of the overlying Wallace formation thins out southeast of Elk Butte and appears higher in the series south of Elk Butte. The schist equivalent to the Wallace formation may therefore overlie the St. Regis formation in this southern part. Because these schists are petrographically alike and thus indistinguishable, a part of the schist north of Gold Butte mapped as the Wallace formation may be equivalent to the St. Regis formation. The estimated thickness of the schist of the St. Regis formation along Breakfast Creek is about 150 m, which is only about one-half the thickness exposed in Shoshone County (Umpleby and Jones, 1923).

PETROGRAPHY

The St. Regis formation is muscovite-biotite schist that contains up to 15 percent garnet. Specimens from the lower beds along Breakfast Creek are strongly sheared, with micas in thin shreds parallel to the foliation. Muscovite flakes are bent and show strain shadows under crossed nicols. Most of them are larger than the interleaved biotite flakes. Some of the largest biotite flakes are also strained but not as strongly as the muscovite.

The measured mode of typical schist from the lower part of the St. Regis formation is 60 percent quartz, 20 percent muscovite, 15 percent biotite, 4 percent plagioclase, and about 1 percent garnet, apatite, magnetite, and zircon. The upper part of the formation has more garnet and less quartz. In many localities medium-grained layers containing less muscovite and more biotite are interbedded with muscovite-rich schist. Many beds have large muscovite flakes in thick plates oriented at an angle of about 2° to 10° from the plane of foliation.

Beds of similar schist occur in the Prichard formation and in the Wallace formation; therefore, only the stratigraphic position of the schist immediately above the massive Revett quartzite and beneath the quartzite-gneiss unit of the Wallace formation indicates that the schist is equivalent to the St. Regis formation.

WALLACE FORMATION

DISTRIBUTION AND LITHOLOGY

The Wallace formation is the most widespread and varied formation in the area; schist, gneiss, and quartzite assigned to it underlie the western and southern two-thirds.

In the northern part of the area the Wallace has been subdivided into four map units:

Upper schist unit
Upper quartzite-gneiss unit
Lower schist unit
Lower quartzite-gneiss unit

Some thin but mappable gneiss layers are also interbedded in the lower schist unit. Comparison of this subdivision with that of Shenon and McConnel (1939, p. 5) in the Coeur d'Alene district suggests that the lower and upper schist units near Elk River and Clarkia are the metamorphosed equivalents of the thinly laminated, noncalcareous argillite of the Coeur d'Alene district (subdivisions 2 and 4 of Shenon and McConnel). Sequences corresponding to the limy quartzite and sandy, shaly, and arenaceous limestones near Coeur d'Alene were recrystallized as quartzites and as biotite, biotite-plagioclase, and diopside or hornblende gneisses near Elk River and Clarkia. These petrographically different rock types were not mapped separately, however, because of their irregular distribution and discontinuous exposure.

In the southeastern part of the area the Wallace formation has been subdivided into five units. The upper four units seem to be the same as those in the northern part though the quartzite-gneiss units are somewhat thinner and a third unit of schist underlies the lower quartzite-gneiss unit. This subdivision corresponds to that of Wagner (1949, p. 9), who distinguished two units of noncalcareous black shale, two units of calcareous shaly sandstone, each 500 to 800 feet thick, and a basal unit, 2,000 feet thick, consisting of impure quartzite, shale, and sandstone with occasional limestone layer, in the Wallace formation near St. Joe River, about 13 miles north of the present area.

A transition from the northern to the southern subdivision takes place near Elk Butte, where the lower quartzite-gneiss unit thins out southward and another quartzite-gneiss unit appears within the schist of the Wallace formation, apparently about 600 m above the base, so that the sequence consists of two quartzite-gneiss units and three schist units. Since the schist of the St. Regis formation is petrologically similar to the basal schist of the Wallace formation, no contact could be mapped between these two formations south of Elk Butte; thus all schist between the Revett quartzite and lower quartzite-gneiss unit of the Wallace formation was mapped as a part of the Wallace formation.

All the schist units are petrographically much alike. The major part of the schist contains either kyanite and staurolite or sillimanite in addition to quartz, biotite, muscovite, and garnet. The schist is coarse-grained in the central and southern part of the area, but fine to medium grained in the northwestern part. Some layers of thin-bedded biotite quartzite with muscovite and biotite laminae are interbedded with the schist.

The two quartzite-gneiss units are mainly of bedded diopside-plagioclase gneiss and quartzite with minor schist layers. Most of the quartzite is thin bedded and contains much biotite and (or) diopside. Thick layers of white granular quartzite with a little feldspar or diopside are interbedded with the diopside-plagioclase gneiss in the lower part of the formation.

The lower quartzite-gneiss unit is exposed within and just west of the fault zone that crosses the area from the northern border to the south and southeast. The upper quartzite-gneiss unit covers the westernmost part of the area. Thin-bedded biotite quartzite and diopside quartzite between the stocks of quartz monzonite and quartz diorite probably belong to the upper unit.

STRATIGRAPHY AND CORRELATION

No complete section of the Wallace formation is available; however, several good partial sections are exposed along logging roads.

The lower quartzite-gneiss unit is best exposed east and northeast of Clarkia along the St. Maries River and its tributaries. The lowermost beds consist of white granular quartzite, which are interbedded with some thin layers of biotite quartzite, hornblende-bearing quartzite, and hornblende gneiss. These beds (1*a*, table 1) cover a wide area along the forks of Merry Creek cropping out along the crest of a gentle anticline. Similar quartzite, but with diopside rather than hornblende, overlies the Revett quartzite and the schist of the St. Regis formation north of Isabella Creek.

Along the St. Maries River, the white granular quartzite is overlain by thin-bedded white and gray banded biotite quartzite with some schist layers (1*b*, table 2). Together these two subunits (1*a* and 1*b*) correspond in their thickness and stratigraphic position to the lowest member of the Wallace formation in Wagner's (1949) subdivision of this formation near the St. Joe River.

A green and white banded diopside-plagioclase gneiss with abundant tremolite, exposed on the north side of the St. Maries River at the mouth of Gold Center Creek, lies stratigraphically above the thin-bedded

TABLE 1.—*Generalized section of the lower quartzite-gneiss unit of the Wallace formation, exposed along Merry Creek in secs. 23 and 14, T. 43 N., R. 2 E.*

Subunit	Estimated thickness		Rock type
	(meters)	*(feet)*	
			Fault
1b	5	16	Light colored hornblende quartzite layers interbedded with dark hornblende-biotite gneiss layers.
			Fault
	10	33	Scapolite-bearing biotite gneiss with quartzitic layers.
	35	115	Hornblende-biotite gneiss.
			Diorite dike
	70	230	White granular quartzite with layers of biotite quartzite.
			Fault
	140	460	White granular quartzite (layers 10–30 cm thick) interbedded with biotite quartzite (layers 5–10 cm thick).
	70	230	Hornblende-biotite gneiss.
	70	230	Biotite quartzite, thin bedded.
	150	490	White granular quartzite (layers 5–20 cm thick) interbedded with gray biotite quartzite (layers 2–5 cm thick).
1a	250	820	White granular quartzite with very thin biotite-bearing layers. Bottom not exposed.
Total	800	2624	

TABLE 2.—*Generalized section of the lower quartzite-gneiss unit of the Wallace formation along the St. Maries River, secs. 11, 14, and 23, T. 42 N., R. 2 E.*

Subunit	Estimated thickness		Rock type
	(meters)	*(feet)*	
1c	200	655	Diopside-plagioclase gneiss with tremolite or actinolite.
1b	30	100	Thin-bedded (½–2 cm thick beds) biotite quartzite with diopside-hornblende-plagioclase quartzite layers.
	30	100	Banded dense dark quartzite with biotite or diopside and hornblende in alternating layers that are 2–10 mm thick.
	70	230	Thin-bedded biotite quartzite with layers, 5–10 mm thick, consisting of pure quartzite.
	150	490	Biotite-plagioclase schist, medium grained.
	30	100	Biotite-plagioclase quartzite interbedded with biotite quartzite and containing some schist layers; variation in the composition between the layers is only slight.
	70	230	Thin-bedded gray and white granular quartzite with quartz veins, 2–5 mm thick.
	20	65	Biotite-rich banded quartzite; light colored layers are few and thin, muscovite-biotite schist is interbedded.
	10	35	Thin-bedded biotite quartzite with hornblende in some dark layers.
1a	10	35	Medium-grained pure quartzite.
	10	35	Biotite schist.
	30	100	White quartzite with muscovite laminae, beds 5–15 cm thick.
	30	100	Thin-bedded granular pure quartzite with biotite-bearing layers.
	160	525	Granular coarse- to medium-grained pure quartzite.
	20	65	Very thin bedded quartzite, white pure quartzite layers and biotite quartzite layers alternating.
Total	870	2865	

biotite quartzite. This diopside-tremolite-plagioclase gneiss (1*c*) corresponds to member 2 in Wagner's subdivision.

In the extreme southeastern corner the lowest quartzite gneiss unit resembles the lower quartzite gneiss exposed along Merry Creek; in both sections the white granular quartzite is the major rock type. Two good sections of this quartzite, each about 250 m thick, are well exposed on the flanks of an anticline along the North Fork of the Clearwater River. In these sections individual beds of white granular quartzite range from 10 to 40 cm in thickness and are separated by beds of biotite gneiss 3 to 4 cm thick. The unit is underlain by thin-bedded biotite gneiss and biotite schist with quartzitic layers and is overlain by coarse-grained sillimanite garnet schist. The western flank of this anticline is exposed along the ridge between Falls Creek and Weitas Creek, where most beds contain much plagioclase and diopside, some beds contain amphiboles, and a few beds, potassium feldspar. The gneiss units exposed south of Green Mountain and east of Gold Butte consist mainly of diopside-plagioclase gneiss with some actinolite or hornblende resembling the corresponding rock types in the Boehls Butte quadrangle (Hietanen, 1963) and in the Headquarters quadrangle (Hietanen, 1962). Diopside quartzite and diopside-plagioclase gneiss exposed in the south-central part of the area underlie schist of the Wallace formation and most likely belong to the lower quartzite-gneiss unit. A small lenslike body of diopside gneiss interbedded with the schist southeast of Little Green Mountain is lithologically similar to the southwesterly dipping lower unit exposed just northeast of it and may be a part of this unit brought up along the crest of an anticline. However, a similar small body of diopside gneiss is interbedded with the schist about 2 miles to the north in the vicinity of Camp 43, and a third such body is about half a mile south of Green Mountain. It seems, therefore, that in the area south of Elk Butte two thinner quartzose sequences, which recrystallized to quartzite and gneiss, instead of one thick sequence were deposited in the lowest part of the Wallace formation. The geologic maps of the Headquarters quadrangle and of the area adjoining this quadrangle in the west also show several thin units rather than two thick units of diopside and biotite gneiss in the schist of the Wallace formation (Hietanen, 1962).

The lower quartzite-gneiss unit is overlain by garnetiferous mica schist, most of which contains either kyanite or staurolite or sillimanite. A good section through the lower schist unit is exposed along the St. Maries River east of Clarkia. The major part consists of fine- to medium-grained muscovite-biotite-garnet schist with some thin-bedded gray quartzitic layers. Most of the schist cleaves parallel to the bedding, but rocks in some outcrops have a fairly well developed transecting cleavage; all cleavage planes have light to medium- gray silvery surfaces peppered with small dark red garnets. Most of the schist is thinly laminated; micaceous layers, 1 to 10 mm thick, alternate with quartz-rich layers, which are 5 to 10 mm. Layers of homogeneous schist, 10 to 50 cm thick, and other layers consisting of biotite quartzite with muscovite laminae are interbedded with the laminated schist. Many outcrops of schist in the southern part of the area are crudely laminated with quartz-rich layers, 3 to 5 mm thick, separated by paper-thin mica laminae.

Folding and faulting makes estimation of the thickness of this schist unit difficult. A thickness of about 1,100 m is estimated near Elk Butte, where the lower quartzite-gneiss unit is only about 100 m thick. Together these would well correspond to the combined thickness of the three lowest members in Wagner's subdivision (3,450 ft). The thickness of the lower schist unit near Anthony Peak is at least 350 m (more than 1,000 ft), and thus a little thicker than that of the corresponding unit near the St. Joe River (850 ft according to Wagner).

The garnetiferous kyanite- and sillimanite-bearing mica schist is overlain by a second quartzite-gneiss unit consisting of interbedded quartzite, diopside- and amphibole-bearing gneiss, and fine-grained mica schist. This upper quartzite-gneiss unit is well exposed along the East Fork of Potlatch Creek and its tributaries, and on the lower slopes of Jackson Mountain. Folding and faulting have obscured the stratigraphic sequence. The lower schist unit and the upper quartzite-gneiss unit are in conformable contact near the forks of Mallory Creek and along the East Fork of Mallory Creek. The schist along the East Fork and the West Fork of Mallory Creek is coarse grained, rich in muscovite and biotite, and contains abundant pegmatitic veins and a few layers, about 1 m thick, of micaceous quartzite. Above this schist along Mallory Creek is a sequence of fine- to medium-grained distinctly bedded quartzites with layers rich in diopside and feldspars (table 3).

TABLE 3.—*Generalized section of the upper quartzite-gneiss unit of the Wallace formation along Mallory Creek and the East Fork of Potlatch Creek, west of the mouth of Mallory Creek.*

Subunit	*Estimated thickness* *(meters)*	*(feet)*	*Rock type*
3	(30)	(100)	Coarse-grained pure quartzite.
	(30)	(100)	White granular quartizte.
			Fault
1*b*	40	130	Biotite-quartzite, thin-bedded with some thick beds and some thin diopside-bearing layers.
2*c*	50	160	Diopside gneiss with thin biotite-bearing layers and white granular quartzite layers, 2–5 cm thick.
			Faults (dikes near Bloom Meadows).
2*b*	150	490	Diopside gneiss, includes thin-bedded gneiss that has biotite-rich layers near the top and bottom.
3	40	130	White granular quartzite with some diopside grains and diopside-gneiss layers.
	(180)	(600)	No exposures (faults)
4	20	65	Plagioclase quartzite in layers 2–5 cm thick, with some biotite-bearing layers, ½–3 cm thick (Mouth of Mallory Creek).
2*b*	150	490	Diopside gneiss.
3	(60)	(195)	White granular quartzite.
2*a*	(150)	(490)	Diopside gneiss.
			Fault
3	60	195	White granular quartzite; beds, 5–20 cm thick, are separated by thin biotite-rich layers.
2*a*	150	490	Diopside gneiss, with some thin biotite-bearing layers.
1*a*	40	130	Biotite quartzite and diopside gneiss, thin bedded.
Total exposed	970	3, 165	
Less thickness owing to repetition	700	2, 280	

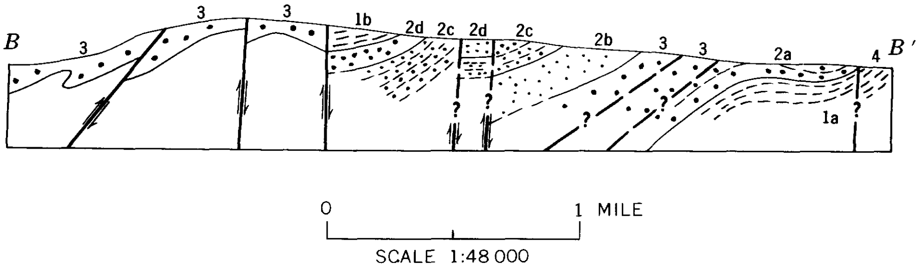


Figure 2.—A cross section through the upper quartzite-gneiss unit of the Wallace formation near Bloom Meadows (*B–B'* on pl. 1). *1a* and *1b* are thin-bedded biotite quartzite with thin layers of diopside gneiss. *2a*, *2b*, *2c* and *2d* are mainly diopside gneiss with thin biotite-bearing layers; *2a* and *2c* have more of these biotite-bearing layers than *2b* and *2d*. Layers of white granular quartzite are interbedded in *2c*. Subunit 3 consists of fairly pure granular quartzite that is coarse grained in the western part of the section.

The sequence has been divided into four lithologic subunits numbered on table 3. Structural repetition is shown by repetition of numbers. Depositional repetition of similar rock types is indicated by letters. The section is complicated by faulting similar to that in figure 2, which shows a cross section along line *B–B'* on plate 1. The stratigraphic column is further complicated by lateral changes in the amounts of the major constituents, quartz, diopside, plagioclase, and biotite. The white granular beds (No. 3) may laterally grade into biotite-, diopside-, or plagioclase-bearing layers (Nos. *2a*, *2b*, *2c*, or 4). Diopside gneiss grades laterally into thin-bedded biotite-bearing gneiss, making the distinction between the subunits 1 and 2 doubtful. The plagioclase quartzite (No. 4) may contain also diopside-rich layers, and parts of it are similar to the thin-bedded diopside-biotite gneiss of subunit No. 1. The individual layers in all the subunits range from a few millimeters to 20 cm or more. In the thin-bedded diopside-biotite gneiss the thicknesses of 1 to 5 cm are most common (fig. 3). In the majority of outcrops the biotite-rich layers are thinner than the plagioclase- and diopside-bearing layers; locally the micaceous laminae are paper thin between quartzitic or gneissic layers that contain very little or no mica. In some parts of the sequence, beds 1 to 5 m thick consist of thin layers, 1 to 2 cm thick, of biotite quartzite separated by mica-rich laminae. These biotite quartzite or biotite gneiss beds alternate either with plagioclase quartzite or with diopside-plagioclase gneiss beds of the same thickness. In some other parts of the unit, biotite schist instead of biotite quartzite and gneiss is interbedded with diopside-plagioclase gneiss. Thick beds of coarse-grained pure quartzite resembling the Revett quartzite are exposed next to the white granular quartzite in sec. 2, T. 40 N., R. 1 E. and in secs. 23 and 34, T. 41 N., R. 1 E. This coarse-grained quartzite may represent an upper part of the subunit 3 (top of table 3) in this vicinity and establish a structural repetition just east of the mouth of Bobs Creek and about a mile south of Bloom Meadows (cross section *B–B'*, pl. 1).

The thickness of the subunits in individual sections can be estimated, but detailed mapping of the subunits would be necessary to establish a complete stratigraphic column. However, this would be a difficult and time-consuming task, because the structure is complicated by many faults and by several sets of folds (Hietanen, 1961a), and because similar layers occur in several subunits. In addition, there is considerable lateral variation on a large scale, which makes the correlation of subunits of different parts of the area difficult if not impossible. For instance, most of the section of the upper quartzite-gneiss unit along Emerald Creek, west of Clarkia, consists of white granular quartzite. Only the uppermost 200 m of this section, which is about 1,300 m thick, consists of thin-bedded biotite quartzite. Some diopside-bearing layers are interbedded in the upper part of the white granular quartzite. The sequence exposed west of Cedar Butte near Clarkia contains more diopside gneiss and less white granular quartzite. Most of the same unit south of Shattuck Butte consists mainly of diopside gneiss and thin beds of biotite gneiss and biotite quartzite; the thickness of the white granular quartzite there is less than along the East Fork of Potlatch Creek and much less than on Emerald Creek, and many beds contain diopside. Near Elk River there is more diopside gneiss and much less white granular quartzite than along the East Fork of Potlatch Creek, but the quartzite is purer. Thus the thickness and sequence of the subunits in each individual section varies considerably. The average total thickness of the unit is probably about 1,000 m.

PETROGRAPHY

The two schist units of the Wallace formation are petrographically alike and are therefore discussed under one heading. The quartzite-gneiss units comprise four main intergradational rock types: (1) thin-

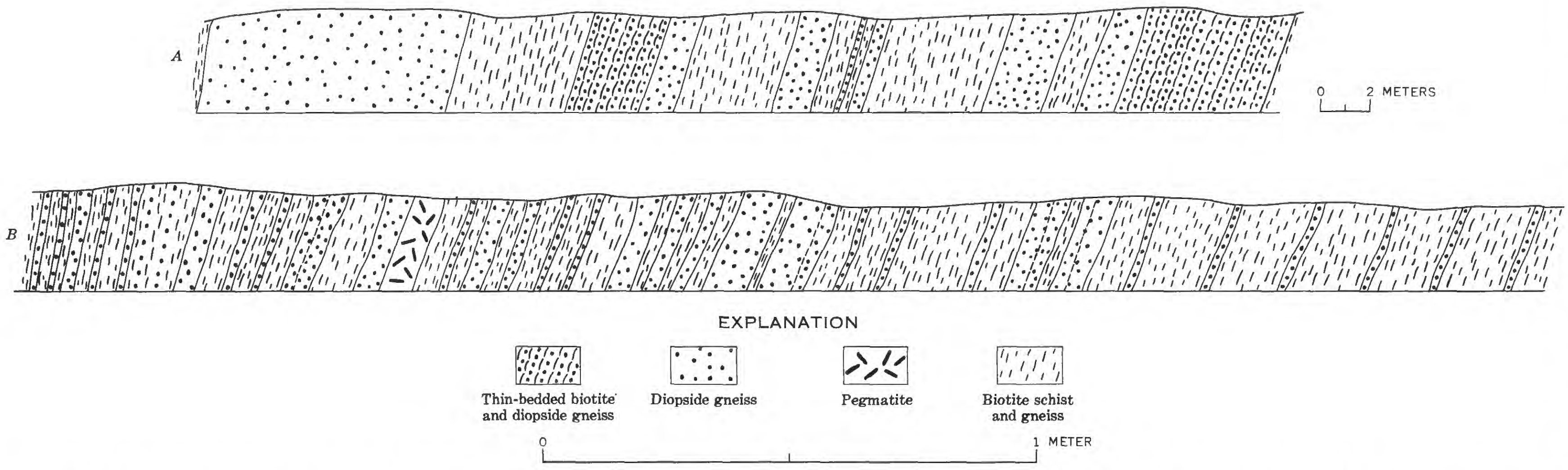


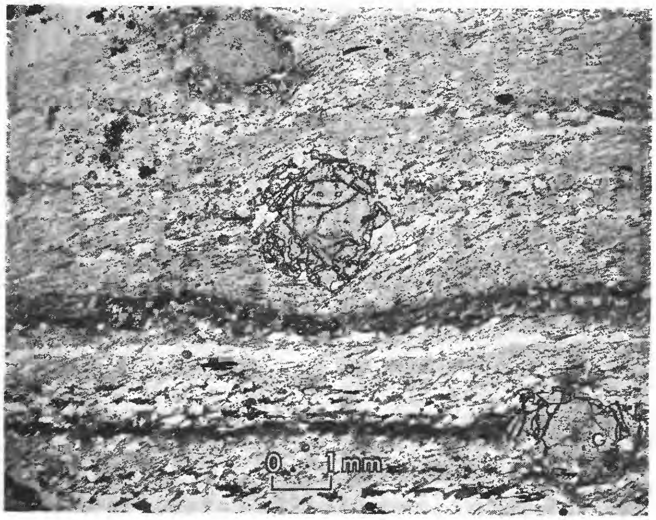
FIGURE 3.—Bedding in biotite and diopside gneiss along the head of Shattuck Creek, 2 miles east of Jackson Mountain. *A*, Beds of diopside gneiss, 0.2–10 meters thick, interbedded with biotite schist and thin-bedded gneiss; *B*, Detail of thin-bedded gneiss.

bedded biotite and diopside gneiss and quartzite, (2) diopside or hornblende gneiss, (3) white granular quartzite, and (4) thin-bedded biotite-plagioclase quartzite and gneiss.

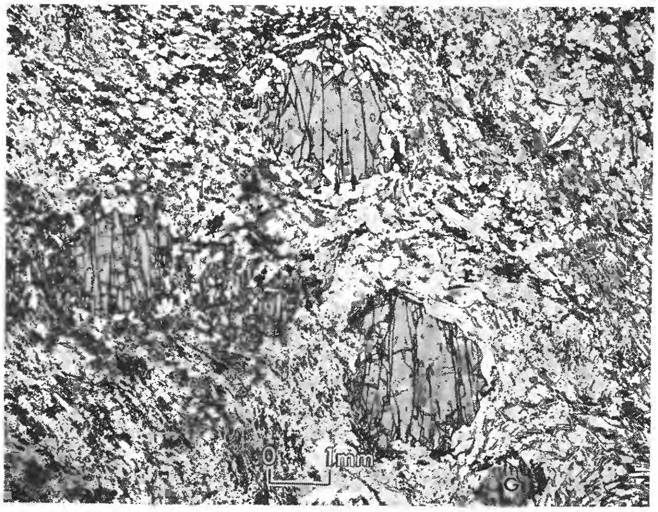
SCHIST

In the northwestern part of the area the schist is fine- to medium-grained thinly laminated muscovite-biotite-garnet schist (fig. 4*A*) with local development of kyanite (fig. 4*B*) and staurolite. Toward the south and southeast the grain size becomes larger, the thin lamination is locally obscured by more intense recrystallization, and sillimanite instead of kyanite is a prevalent aluminum silicate (fig. 5*A*). Interbedded with the schist are fine-grained quartz-rich layers, many of which contain muscovite laminae (fig. 5*B*).

The thicker layers in the schist consist of quartz-muscovite-biotite rock with numerous subhedral gar-

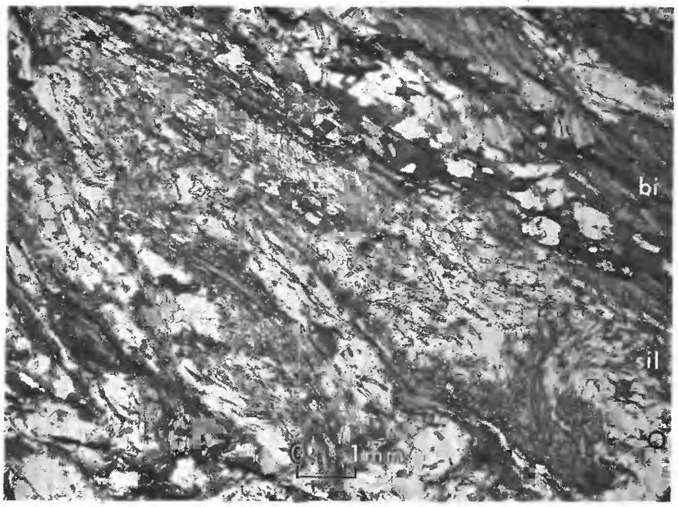


A. Laminated fine-grained garnet-mica schist from Bechtel Mountain (loc. 1848), 3 miles west-southwest of Clarkia. Dark bands are biotite laminae in the muscovite schist. Note that the euhedral cores in garnets (G) contain only a few tiny inclusions of quartz (white), whereas the rims contain abundant quartz. The orientation of muscovite flakes (light gray) shows the direction of cleavage. Plane polarized light.

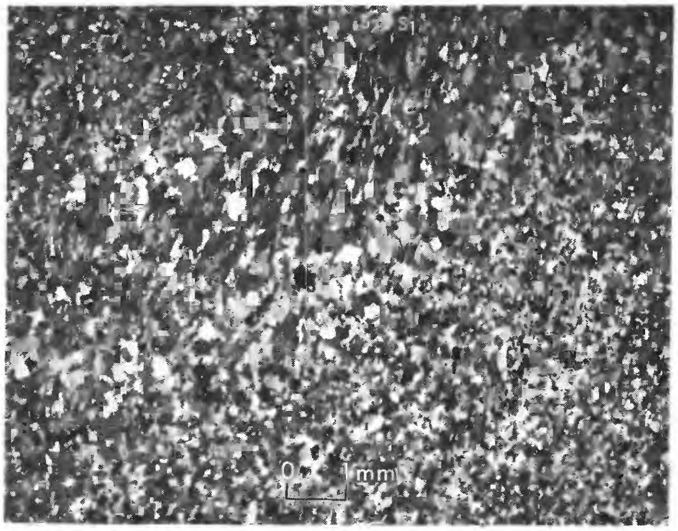


B. Garnet-kyanite schist, 1 mile west of Gold Center (loc. 1615). Small kyanite (K) crystals occurs in aggregate and nodules in this medium-grained rock. Dark mineral is biotite, light-colored minerals are muscovite and quartz; G-garnet. Plane-polarized light.

FIGURE 4.—PHOTOMICROGRAPHS OF SCHIST OF THE WALLACE FORMATION IN THE NORTHERN PART OF THE AREA.



A. Sillimanite-mica schist from Hemlock Butte in the central part of the area (loc. 1429). Biotite (bi) occurs more abundantly in certain thin layers which correspond to the laminae of figure 4*A* but are partly obscured by a strong deformation and recrystallization. Sillimanite (sil), quartz (Q). Plane-polarized light.



B. A quartzitic layer in garnet-mica schist about 1 mile west of Gold Center. The orientation of mica flakes (s_2 almost vertical) shows the plane of foliation which makes an angle from 50° to 80° with the bedding plane s_1 that crosses the photograph diagonally from lower left to the upper right corner. Crossed nicols.

FIGURE 5.—PHOTOMICROGRAPHS OF A SILLIMANITE SCHIST AND A QUARTZITE LAYER IN SCHIST OF THE WALLACE FORMATION.

nets and white nodules consisting of muscovite, kyanite, and a few tiny crystals of staurolite. Apatite, magnetite, tourmaline, sphene, and zircon occur as accessories.

The mode of a typical layer in the schist from about a mile west of Gold center is: 28 percent quartz, 31 percent muscovite, 19 percent biotite, 12 percent kyanite, 9 percent garnet, 1 percent staurolite and accessories. The quartz is in irregular grains 0.1 to 0.3 mm in diameter. Interleaved biotite and muscovite flakes, 0.2 to 1 mm in diameter, occur between the quartz grains. Clusters of larger flakes of muscovite are scattered through the schist or are next to nodules, 1 to 2 cm in diameter, of muscovite and small grains of kyanite (fig. 6). In some layers these nodules have nearly square cross sections, suggesting that they are pseudomorphs after andalusite.

Garnet crystals in the schist are subhedral, ½ to 2 mm in diameter, and contain many round quartz inclusions. The index of refraction $n=1,805\pm0.003$ and the unit cell dimension $a_0=11.582\pm0.006$ A were measured in a garnet from the schist along Marble Creek. According to the diagrams by Sriramadas (1957) these values indicate that the garnet is almandite with about 18 percent grossularite and 8 percent pyrope. Staurolite crystals, 3 to 7 mm long, occur in some layers. In other layers, garnet crystals are altered to chlorite and the nodules consist of muscovite without kyanite. Sphene is abundant along cracks where biotite is altered to chlorite. The titanium in this sphene was probably derived from the biotite. Apatite and tourmaline occur in tiny euhedral crystals about 0.1 mm and 2 to 5 mm long respectively.

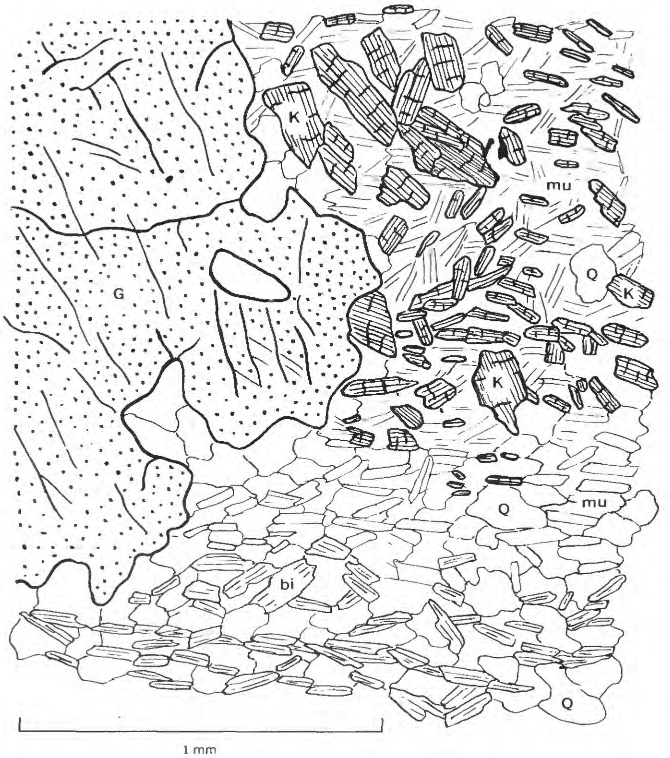


FIGURE 6.—Camera lucida drawing of a part of a nodule consisting of muscovite (mu) and kyanite (K) in garnet-kyanite schist along the St. Maries River, 1 mile west of Gold Center (loc. 1615). Biotite (bi) occurs only in the lower part of the figure. Garnet (G), quartz (Q).

In the quartzitic layers interbedded with the schist, the same minerals occur but in differing proportions. About 50 to 60 percent of the rock is quartz, and some plagioclase is common. The quartzitic layers also contain nodules of muscovite and kyanite, 1 to 2 mm long, but only in their mica-rich laminae.

The upper schist unit is very similar to the lower schist unit except that in some localities, for example on Bechtel Mountain, sillimanite is present instead of kyanite. On Bechtel Mountain, the schist is fine grained, thinly laminated and contains numerous garnet crystals, 1 to 5 mm in diameter (fig. 4*A*). Light-colored aggregates of sillimanite, 10 to 20 cm long and 1 to 2 cm wide, are randomly oriented along the bedding planes. The outlines of the aggregates resemble those of the kyanite on Anthony Peak, but the indices of refraction $\alpha=1.654\pm0.002$ and $\gamma=1.677\pm0.002$, indicate that they are sillimanite.

In typical schist from west of Clarkia, the micaceous laminae consist of tiny flakes of biotite, whereas the quartz-rich layers contain both muscovite and biotite. Garnet crystals have solid cores, but the rims have quartz inclusions as in fig. 4*A*. Garnets grow across the laminae and the micas around the garnets are only slightly bent. Cleavage makes an angle of about 35° with the bedding plane. The muscovite flakes in the quartzitic layers are parallel to this cleavage, whereas many of the biotite flakes in the mica-rich laminae are parallel to the bedding.

In some layers clusters of large muscovite plates are pseudomorphs after an orthorhombic(?) mineral, probably andalusite. The best exposure of such layers is 2 miles west of Cedar Butte, just west of the mapped area (loc. 1851). The clusters of muscovite range from 1.5 to 2 cm in diameter and have square outlines. The individual muscovite plates in the clusters are about 0.5 mm thick, whereas the flakes of muscovite in the main part of the schist are less than 0.1 mm thick.

South and southeast of Clarkia, the grain size of the schist gradually becomes larger. The schist between Anthony Peak and Hemlock Butte is medium grained and contains abundant biotite and muscovite. On Anthony Peak crystals of red garnet, 5 to 15 mm in diameter, and prism-shaped clusters, 5 to 10 cm long, of tiny crystals of kyanite are embedded in a fine-grained micaceous groundmass. Biotite and muscovite flakes, 1 to 3 mm long, are interleaved. Muscovite also is in larger flakes 4 to 7 mm in diameter, that are oriented across the bedding planes.

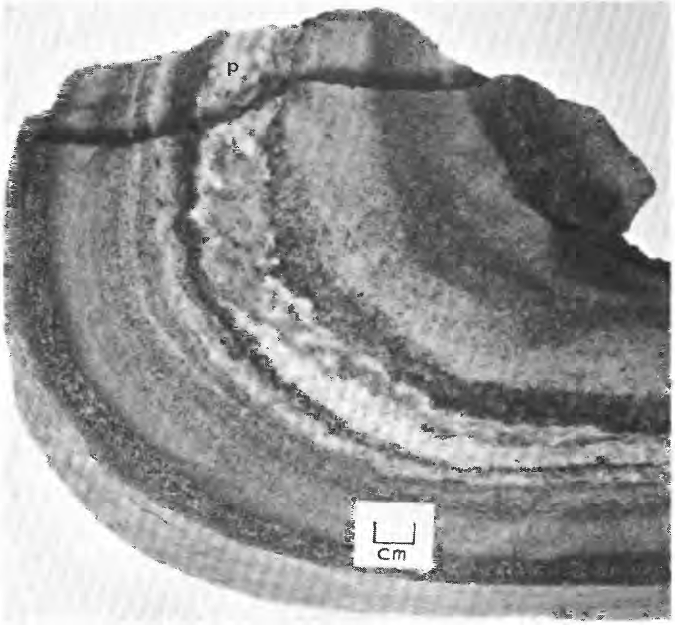
In the schist on Hemlock Butte, sillimanite abounds but kyanite crystals are fewer and smaller than those on Anthony Peak. Brown biotite, in which $\gamma=1.653 \pm 0.002$ and which is riddled with tiny inclusions of zircon, is the dominant mica, the flakes ranging from 2 to 4 mm in diameter. Many layers contain only a few muscovite plates, all oriented across the plane of foliation. Nodules are common; some consist of sillimanite, others of small grains of quartz and kyanite, and still others of pegmatitic material. Sillimanite in the nodules is light brown, whereas the needles of sillimanite within and between the quartz grains and mica flakes are colorless. Irregularly shaped grains of magnetite are enclosed in the centers of some nodules.

Still farther south, in the vicinity of Elk Butte, Jackson Mountain, and Little Greek Mountain, the schist contains sillimanite but no kyanite or staurolite. Here the schist is coarse grained, the quartz grains ranging from 0.5 to 7 mm in diameter. In most layers there is more biotite than muscovite. The biotite flakes are of medium size and parallel to the plane of foliation, which is commonly parallel to the bedding. Some muscovite is interleaved with biotite and the remainder occurs as large plates that make an angle of 10° to 40° with the plane of foliation. These large plates include small quartz grains; conversely many small mica flakes are included in larger quartz grains. Some muscovite forms clusters in which flakes of medium size are oriented at random. Sillimanite is also common in these clusters, which thus correspond chemically to the nodules of muscovite and kyanite near Clarkia. Most of the schist contains 2 to 10 percent of garnet in crystals ½ to 5 mm in diameter.

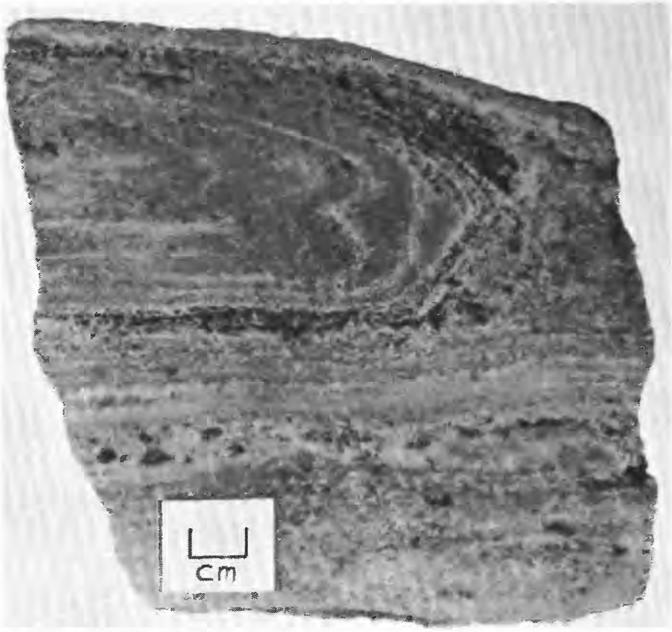
South of Little Green Mountain, a few layers, 20 to 50 cm thick, that consist of muscovite and a little sillimanite and quartz are interbedded with the garnet-sillimanite schist. In some places laminated layers rich in quartz are interbedded with the schist. These layers contain very little if any sillimanite or garnet, and the amount of quartz is 60 to 80 percent. The grain size ranges from ½ to 3 mm. Every gradation exists from these quartz-rich layers to layers rich in micas, garnet, and sillimanite.

THIN-BEDDED BIOTITE AND DIOPSIDE OR HORNBLENDE QUARTZITE AND GNEISS

Both quartzite-gneiss units contain thin-bedded strata in which fine-grained layers of gray biotite quartzite (fig. 7*A*) alternate with some other layers that consist either of quartz with some plagioclase or of quartz, plagioclase, microcline, and diopside in about equal amounts. Where diopside is present the layers are light green, whereas the feldspar-bearing and pure quartzite layers are white or light bluish gray. Most



A. A fold in thin-bedded biotite quartzite along West Stony Creek (loc. 1521). Pegmatitic vein (P) parallels the bedding. The dark layers are rich in biotite.



B. Diopside gneiss of the Wallace formation, with a thin layering and a small intraformational fold caused by sliding of sediments (see p. C34). The dark spots are cavities from which calcite has weathered. Along the east Fork of Potlatch Creek just west of Bloom Meadows (loc. 1262).

FIGURE 7.—THIN-BEDDED BIOTITE GNEISS AND DIOPSIDE GNEISS OF THE WALLACE FORMATION.

of the thin-bedded rocks in the lower quartzite-gneiss unit contain less diopside and more quartz than the thin-bedded layers in the upper quartzite-gneiss unit. The thin-bedded layers in the northern part of the area contain hornblende or actinolite instead of diopside.

The diopside-bearing layers in typical thin-bedded gneiss consist mainly of quartz, plagioclase, and diopside, with or without potassium feldspar and with a few small grains of biotite, sphene, apatite, and zircon. The light-colored constituents of most biotite-bearing layers include much potassium feldspar in addition to quartz and plagioclase. The texture of both diopside-bearing and biotite-bearing layers is granoblastic; the biotite flakes are parallel to the bedding (fig. 8) and the diopside occurs in small irregular-shaped grains. As a rule, light-green hornblende instead of diopside occurs in the corresponding layers north of Clarkia. The distribution of muscovite is highly irregular; most flakes occur in the thin micaceous laminae that are interbedded with biotite quartzite. Muscovite is rarely found in the thin biotite-bearing layers interbedded with diopside gneiss.

Many diopside- and hornblende-bearing beds in the northern and central part of the area contain scapolite and calcite. In some of these layers microcline is the main light-colored constituent; scapolite abounds and the amount of quartz is small. A measured mode of such layer in volume percents is: 47 percent diopside, 29 percent microcline, 18 percent scapolite, 3 percent quartz, 2 percent calcite, and 1 percent biotite.

In most of the other diopside- and hornblende-bearing beds in the northern and central part of the area, about half the feldspar is microcline and the other half albite (An_{5-10}), whereas in similar beds farther south the amount of microcline is negligible and the plagioclase ranges from andesine (An_{30}) to bytownite (An_{85}). In biotite-bearing layers also the plagioclase changes from albite in the northern part to oligoclase and andesine in the southern part of the area. The total amount of microcline also changes from north to south; it abounds in the northern part of the area but decreases southward, becoming negligible in most layers near the North Fork of the Clearwater River. Only a slight change occurs in the amounts of muscovite and biotite. The total amount of micas stays fairly constant but the amount of muscovite decreases and that of biotite increases toward the south.

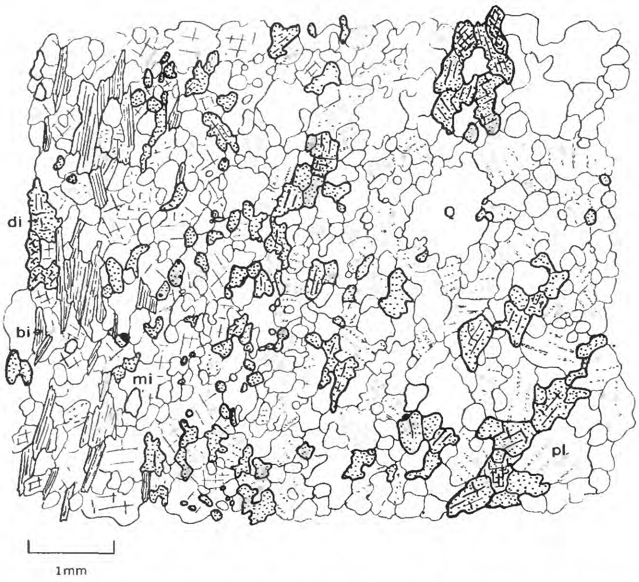


Figure 8.—Camera lucida drawing of a biotite-bearing layer in diopside gneiss along the East Fork of Potlatch Creek 1.8 miles east of the mouth of Mallory Creek (loc. 1246). Quartz (q), plagioclase (pl), microcline (mi), diopside (di), biotite (bi).

The texture of the thin-bedded quartzite and gneiss suggests more intense recrystallization toward the south. In the northern part of the area, quartz and feldspar grains are round or slightly irregular in shape and average 0.1 to 0.2 mm in diameter. In contrast, in the southern part of the area the grains are fairly irregular with sutured or wavy borders and their diameter ranges from 0.1 to 3 mm.

DIOPSIDE GNEISS

The diopside gneiss is banded, fine to medium grained, and grayish green. The thickness of the beds ranges from 5 cm to several meters, the thickness of 1 to 2 m being most common. The banding is due to the distribution of the constituent minerals, mainly to the abundance of diopside in laminae, 2 to 5 mm thick, and in part to the occurrence of calcite or biotite in certain thin laminae. Differential weathering is typical of the beds that contain calcite-rich laminae. Carbonate material, which may have been dolomitic, has weathered out, leaving cavities lined by hematite. In some layers diopside occurs in lens-shaped masses 5 cm thick and 10 to 20 cm long. These were probably dolomitic concretions in the original sedimentary rock. In the northernmost part of the area (for example, in loc. 1842) green hornblende and a little epidote, instead of diopside, crystallized in the presumed dolomitic layers.

In most layers of diopside gneiss the major constituents are quartz, plagioclase, microcline, and diopside, with or without scapolite, zoisite, clinozoisite, tremolite, hornblende, biotite, and calcite. The accessories are sphene, apatite, and magnetite. The hornblende that takes the place of diopside in the northernmost part of the area is pleochroic in pale green and light blue-green; in specimen 1910 it has $\alpha=1.630\pm0.001$, $\beta=1.644\pm0.001$, $\gamma=1.652\pm0.001$.

Light to dark-green prisms of tremolite and actinolite form the major portion of some layers near Clarkia. The amount of the major constituents shows local and regional variation. The local variation is visible as a more or less distinct banding (fig. 7*b*), caused mainly by distribution of dark constituents.

To study the local variation in detail, the mode was measured in typical specimens of fairly homogeneous diopside gneiss with differential weathering (table 4, column 1), in three adjacent thin layers of distinctly banded diopside gneiss (table 4, columns 2, 3, and 4, and fig. 9*a*) and in slightly banded diopside gneiss (table 4, column 5), all from the upper quartzite-gneiss unit exposed along Potlatch Creek.

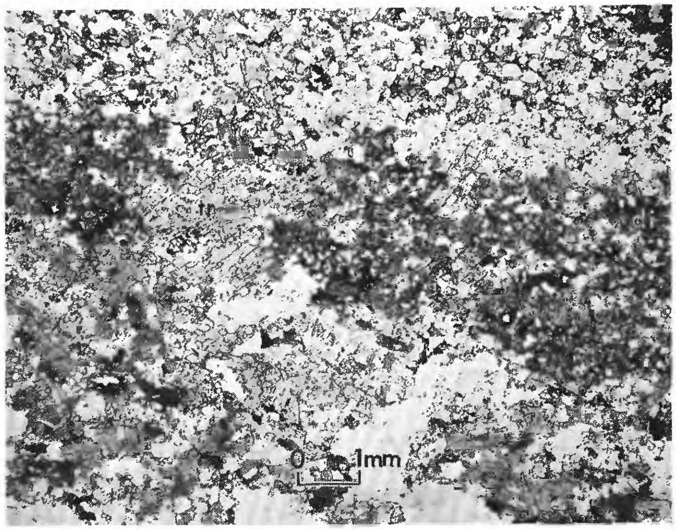
Comparison of the results (table 4, columns 1 to 5) shows an extensive variation in the amounts of the major constituents in the various layers of the diopside gneiss. The amount of diopside ranges from 16 to 49 percent, which gives rise to the visible banding. Variation in the amounts of the three light-colored constituents—quartz, plagioclase, and microcline—is even greater. Most layers contain 10 to 20 percent plagioclase but the amount of microcline varies from negligible to about 60 percent. Many layers rich in microcline contain only a small amount of quartz, whereas those poor in microcline are rich in quartz. Less plagioclase and microcline occur in layers rich in scapolite.

In many outcrops the layers, 1 to 3 cm thick and consisting of about 60 percent quartz, 30 percent scapolite, 10 percent diopside, and 10 percent calcite, alternate with other layers that range from 1 to 10 mm in thickness and consist of about 60 percent microcline, 30 percent quartz, 15 percent biotite, and 5 percent diopside. Abundant clinozoisite, which includes numerous tiny irregular-shaped grains of quartz, occurs with diopside in some scapolite-bearing quartz-rich layers.

Quartz-rich layers are prominent in an area south of Elk River where the gneiss has very little if any microcline (table 4, columns 6, 7, and 8). Most layers there are rich in quartz (table 4, column 6); many contain abundant plagioclase in addition to quartz and diopside (table 4, column 8). In some layers the $CaAlSi_2O_8$ component of plagioclase supplied material for zoisite (table 4, column 7).

To supplement the study of the distribution of potassium feldspar, volume percentages of the major constituents were estimated in a number of other gneissic layers collected from various parts of the lower and upper quartzite-gneiss units (tables 5 and 6). The results show a great variation in the amounts of major constituents of the diopside- and amphibole-bearing layers. Some of the samples of hornblende- and diopside-gneiss from the lower unit (locs. 1910, 1915, 1610, 1524, 1445, 2012, 1468, 2008, 2005, 1992) contain potassium feldspar in amounts ranging from 20 to 30 percent. Samples 1915 and 1524 are similar to numerous other layers in the northern part of the area. In the southern part only one specimen studied (No. 1992) has much potassium feldspar; in most layers this mineral is absent. The mineral content of the additional samples from the upper quartzite gneiss unit (table 6) confirms the finding that potassium feldspar is more abundant in the northern and central part of the area.

A. Biotite-rich laminae (b) in diopside gneiss. Along the East Fork of Potlatch Creek 1.8 miles east of the mouth of Mallory Creek (loc. 1246). Diopside (di), the light-colored minerals are quartz, plagioclase, and orthoclase. Plane-polarized light.



B. A layer of tremolite-diopside gneiss interbedded with garnet-mica schist on the ridge between Isabella and Breakfast Creeks (loc. 1444). Tremolite (tr), diopside (di). The light-colored minerals are mainly quartz and only about 10 percent bytownite. Plane-polarized light.

FIGURE 9.—PHOTOMICROGRAPHS OF BIOTITE LAMINAE AND TREMOLITE IN DIOPSIDE GNEISS IN THE WALLACE FORMATION.

TABLE 4.—*Measured mode in volume percentages of various layers of diopside gneiss along East Fork of Potlatch Creek and south Elk River.*

Location:	East Fork of Potlatch Creek						South of Elk River		North Fork of Clearwater River	
	1	2	3	4	5	6	7	8	9	
Rock:	Diopside gneiss with differential weathering	Quartz-rich layers, 1.5 cm thick, in banded diopside gneiss	Biotite-rich laminae, 2 mm thick, in banded diopside gneiss	Layer between 2 and 3, 1 mm thick	Average composition of slightly banded diopside gneiss	Quartz-rich layer, 1 cm thick, in diopside gneiss	Layer rich in zoisite, ½ to 1 cm thick	Diopside-plagioclase gneiss	Diopside-plagioclase gneiss	Range of the volume percentage of constituent minerals
Locality No.:	1261	1246	1246	1246	1251	560	560	557		
Quartz	5	52	20	13	13	76	31	31	30	5–76
Plagioclase	1	(An$_{28}$)23	(An$_{28}$)20	----	(An$_{28}$)14	----	(An$_{25}$) 8	(An$_{43-50}$)38	(An$_{85}$)34	0–38
Microcline	24	----	20	60	11	----	----	2	----	0–60
Scapolite	15	----	----	----	12	----	----	----	----	0–15
Zoisite	----	----	----	----	1	3	35	----	----	0–35
Diopside	49	23	16	27	36	20	10	28	36	10–49
Biotite	2	2	24	----	----	----	----	----	----	0–24
Calcite	4	----	----	----	13	----	15	----	----	0–15
Sphene	----	----	----	----	----	1	1	1	----	0–1
Total	100	100	100	100	100	100	100	100	100	----

TABLE 5.—*Estimated volume percentage of major constituents in typical layers of the lower quartzite-gneiss unit of the Wallace formation*

Locality	Rock type	Quartz	Plagioclase	Microcline	Biotite	Muscovite	Diopside	Hornblende	Tremolite or actinolite	Calcite	Scapolite	Epidote minerals	Remarks
1910	Hornblende gneiss	2	10	30	----	----	----	40	----	----	15	3	Interbedded with quartz-rich layers.
1913	Thin-bedded biotite-plagioclase gneiss.	45	(An$_8$) 35	5	10	5	----	----	----	----	----	----	----
1915	Thin-bedded hornblende gneiss with scapolite.	30	----	20	2	----	----	25	----	----	20	3	----
1908	Thin-bedded muscovite and biotite gneiss. a	55	35	----	10	----	----	----	----	----	----	----	Layer b makes up about 20 percent of the total rock.
	b	55	----	35	----	10	----	----	----	----	----	----	
1906	Thin-bedded biotite-plagioclase gneiss.	50	35	2	10	----	----	----	----	3	----	----	----
1904	Scapolite- and calcite-bearing biotite gneiss.	53	2	----	15	----	----	----	----	20	10	----	----
1917	Scapolite- and calcite-bearing biotite-hornblende gneiss.	40	----	10	10	----	----	10	----	10	20	----	----
1610	Tremolite gneiss	15	(An$_5$) 15	----	----	----	----	----	70	----	----	----	----
1527	Thin-bedded hornblende gneiss.	55	(An$_{30}$) 10	----	5	----	----	25	----	----	----	5	Zoisite.
1528	Plagioclase gneiss	55	(An$_6$) 35	----	1	9	----	----	----	----	----	----	----
1524	Thin-bedded diopside and hornblende gneiss.	20	20	20	----	----	20	20	----	----	----	----	Diopside and hornblende in separate thin layers.
1525	White granular quartzite	85	10	4	1	----	----	----	----	----	----	----	----
1520	Microcline-rich layer in thin-bedded biotite gneiss.	20	(An$_{30}$) 15	55	10	----	----	----	----	----	----	----	----
1445	Diopside-tremolite gneiss	10	(An$_{30}$) 15	----	2	----	20	----	30	13	----	----	----
2012	Diopside-tremolite gneiss	20	(An$_{28}$) 50	----	----	----	15	----	15	----	----	----	----
1468	Diopside-actinolite gneiss	20	(An$_{15}$) 45	----	5	----	15	----	15	----	----	----	----
2008	Diopside-actinolite gneiss	35	(An$_{12}$) 35	----	----	----	15	----	15	----	----	----	Diopside and actinolite in separate thin layers.
2005	Hornblende gneiss	32	(An$_{86}$) 38	----	----	----	5	25	----	----	----	----	----
1992	Diopside-actinolite gneiss	15	(An$_{36}$) 40	25	----	----	10	----	10	----	----	----	----

TABLE 6.—*Estimated volume percentage of major constituents in typical layers of the upper quartzite-gneiss unit of the Wallace formation*

Locality	Rock type	Quartz	Plagioclase	Microcline	Biotite	Diopside	Hornblende	Calcite	Scapolite	Epidote minerals	Clinozoisite
1885	Diopside-hornblende gneiss	40	(An$_{15}$) 30	20	--------	7	3	--------	--------	--------	--------
1886	Biotite gneiss	20	20	37	20	--------	2	--------	--------	1	--------
1889	Thin-bedded biotite gneiss	30	--------	55	15	--------	--------	--------	--------	--------	--------
1919	Diopside gneiss	52	(An$_{50}$) 8	10	--------	20	--------	--------	--------	10	--------
1920	Diopside-hornblende gneiss	35	(An$_{58}$) 35	--------	--------	10	20	--------	--------	--------	--------
1435	Thinly laminated diopside and biotite gneiss	20	(An$_{32}$) 20	--------	10	50	--------	--------	--------	--------	--------
1433	Diopside gneiss	35	(An$_{8}$) 50	--------	--------	15	--------	--------	--------	--------	--------
1437	Diopside gneiss	30	(An$_{22}$) 30	--------	--------	40	--------	--------	--------	--------	--------
1260	Thin-bedded diopside and biotite gneiss. a	60	--------	--------	--------	10	--------	10	20	--------	--------
	b	20	--------	60	15	5	--------	--------	--------	--------	--------
1262	Diopside gneiss	40	--------	--------	--------	30	--------	--------	20	10	--------
1438	Scapolite-biotite gneiss	30	25	--------	20	--------	--------	--------	25	--------	--------
1940	Diopside gneiss	20	40	10	--------	25	--------	--------	--------	5	--------

None of the diopside-plagioclase gneiss layers studied from the Boehls Butte quadrangle and from the area south of plate 1 contain potassium feldspar and certainly these layers correspond to the lower and upper quartzite-gneiss units of the present area (Hietanen, 1962, and 1963). In the Headquarters quadrangle (southeast of the present area) only one sample of diopside gneiss and another of thin-bedded biotite- and diopside-plagioclase gneiss are rich in potassium feldspar. Most of the diopside-plagioclase gneiss layers east and south of the present area consist of quartz, plagioclase, and diopside in about equal amounts. Thus there seems to be a definite decrease in the number of microcline-bearing layers and increase in plagioclase-bearing layers toward the south and southeast. Because the amount of micas stays fairly constant, this means a decrease in the total amount of potassium and increase in sodium toward the south and southeast parallel to the regional structural trend. Before it can be concluded whether this lateral change in composition of diopside and hornblende gneiss is due to initial differences in the original strata or to a later metasomatic exchange of alkalies during metamorphism, the textural relations and the composition of the adjoining layers must be considered.

The distribution of diopside shows a considerable lateral variation, which, however, is quite different from that of potassium feldspar. The total amount of diopside is fairly constant over the central and southern part of the area, but in the individual beds the content of diopside is highly variable. For instance, the gneiss layers east and west of Gold Butte contain abundant diopside, but the amount of this mineral decreases toward the south, and the corresponding layers along the North Fork of the Clearwater River consist exclusively of white granular quartzite with rather thin beds of biotite gneiss.

An exceptionally large amount of diopside (about 50 percent) occurs in some granular light green layers of thinly laminated diopside-biotite gneiss at the head of Mallory Creek (table 6, No. 1435). The paper-thin laminae that separate the diopside-rich layers, 2 to 10 mm thick, consist of light brown mica with γ=1.590 ±0.005 and small 2V; pleochroism is Z=light brown, X=very light tan. These optical properties suggest this mica to be closest to phlogopite.

In many gneissic layers in the northern half of the central part of the mapped area abundant tremolite or actinolite occurs instead of diopside (table 5, No. 1610) or with the diopside (locs. 1468, 2008, 2012, and fig. 9*b*). These tremolite- and actinolite-rich diopside gneisses are especially common near Gold Center but also occur along the ridge between Isabella Creek and Breakfast Creek and along Gold Creek.

All layers in the northernmost part of the area contain green hornblende instead of actinolite and diopside (table 5, Nos. 1910, 1915, 1917) or with the diopside (table 6, No. 1885). Clinozoisite and zoisite are common additional constituents, and plagioclase is albitic.

Some bytownite-bearing layers east of Gold Butte contain grossularite in anhedral crystals that measure 1 to 5 mm in diameter and include numerous tiny round quartz and plagioclase grains. Diopside grains in this rock are larger and more irregular in shape than those in the other layers.

The textures of three specimens of diopside gneiss from the upper quartzite-gneiss unit are shown on figures 10–12. Figure 10 is a camera lucida drawing from a banded diopside gneiss along the East Fork of Potlatch Creek, figure 11 shows a layer of diopside quartzite south of Elk River, and figure 12 is diopside-plagioclase gneiss along the North Fork of the Clearwater River. In all three specimens the texture is granoblastic and the grain size varies within fairly narrow limits. Diopside occurs as small rounded or angular grains or groups of grains between the light-colored constituents. Along the East Fork of Potlatch Creek and near the North Fork of the Clearwater River most of the quartz grains are small and round; many of them are included in the other minerals. In the layers exceptionally rich in quartz, such as No. 560 south of the town of Elk River, the quartz has recrystallized as irregularly shaped, large grains with slightly sutured borders. Most of the microcline and plagioclase occur as small rounded or irregularly shaped grains (figs. 10 and 12) with quartz, but a few larger grains include one or two, rarely more, of the small round quartz grains. These textures are similar to those of isochemically metamorphosed sediments. Specifically, no replacement textures that would indicate a late introduction of potassium were found near the East Fork of Potlatch Creek.

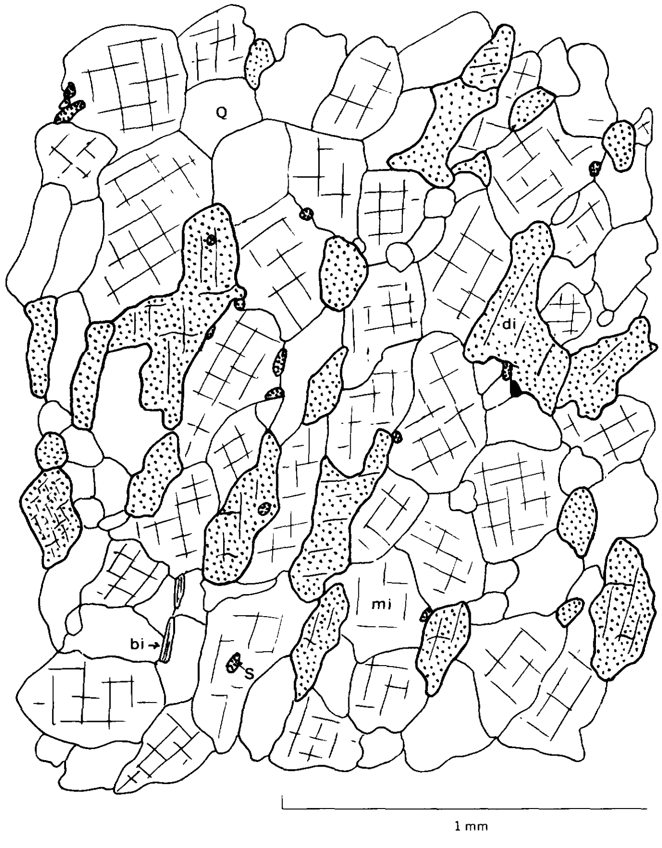


Figure 10.—Camera lucida drawing of diopside gneiss along the East Fork of Potlatch Creek, 1.7 miles east of the mouth of Mallory Creek (loc. 1248). Diopside (di) occurs as small rounded grains between larger quartz (Q) and microcline (mi) grains. Biotite (bi), sphene (S). Note abundance of microcline.

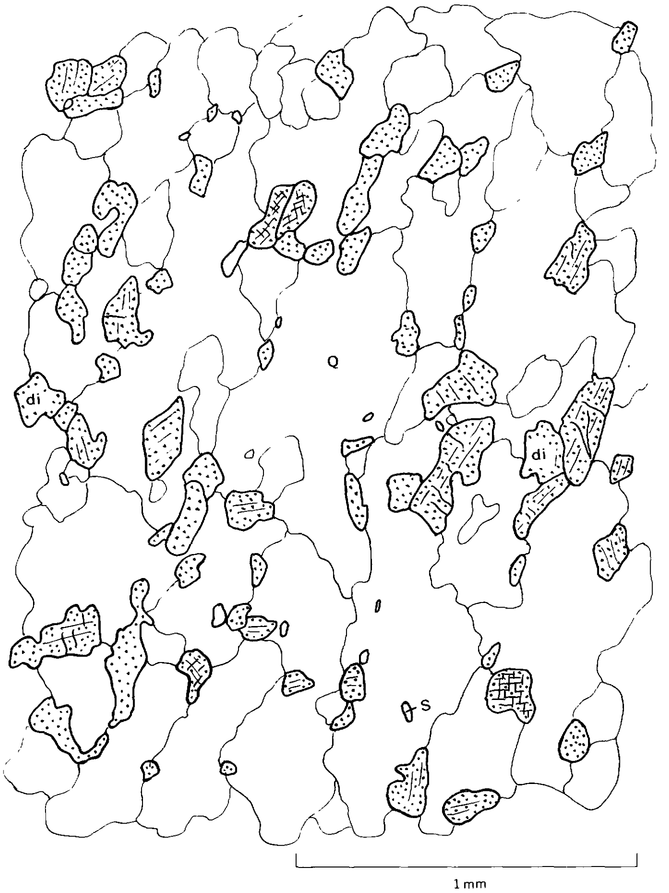


Figure 11.—Camera lucida drawing of diopside quartzite, 4.7 miles south of Elk River along the county road leading from Dent to Elk River. This rock lacks potassium feldspar. Quartz (Q), diopside (di), sphene (S).

The diopside-plagioclase gneiss near the North Fork of the Clearwater River has a granoblastic equigranular texture typical of isochemically metamorphosed sedimentary rocks. A part of these diopside-plagioclase gneisses, however, has undergone metasomatic changes that altered the diopside into hornblende and generally increased the amount of plagioclase and decreased that of quartz. These metasomatic changes took place after the major period of recrystallization, as has been earlier described from an area south and southeast of the Elk River–Clarkia area (Hietanen, 1962). Signs of similar metasomatism are rare in the southeastern part of the area.

WHITE GRANULAR QUARTZITE

The third kind of layer that can easily be identified in the field consists of white granular rocks fairly rich in quartz but locally containing large amounts of feldspars. The individual beds, which range from 2 to 30 cm in thickness, are separated by micaceous layers, a few millimeters to several centimeters thick. In some localities in the northern part of the area, light-gray micaceous quartzite layers are interbedded with the white granular quartzite. Near the North Fork of the Clearwater River thin layers of medium-grained biotite-plagioclase gneiss are interbedded with white granular quartzite representing the more highly metamorphosed equivalent of the gray layers.

In the upper quartzite-gneiss unit along the East Fork of Potlatch Creek and its tributaries scattered individual light-green grains of diopside or small groups of grain give a spotted appearance to many layers that are interbedded with white granular quartzite. In some outcrops the small grains of green diopside occur in greater numbers in layers, 1 or 2 cm thick, that separate the pure quartzite layers, which range from 5 to 15 cm in thickness.

In several localities light-colored layers of the upper quartzite-gneiss unit have relict crossbedding. This crossbedding becomes visible because the tiny grains of green diopside are in rows that make an angle with the overlying bed and join the underlying layer asymptotically. Crossbedding has been described from the sandy layers of the Wallace formation in the Coeur d'Alene district by Ransome and Calkins (1908, p. 41) as follows:

> The weathered surfaces are as a rule pure white * * * upon these surfaces, when they have been long exposed to the etching action of atmospheric water, a beautiful cross-bedded structure appears, of which no trace is visible upon fresh fractures

and the etched material was suggested to be mainly calcite. Near Elk River, crystallization of diopside in the corresponding layers indicates that the carbonate was mainly dolomite. Crossbedding is conspicuous also in fresh surfaces, because of the contrasting color of the white quartz-feldspar rock and the green diopside. Ransome and Calkins described the sandstones with crossbedding to be characteristic of the middle part of the Wallace formation. Near Elk River this structure appears in the middle part of the upper quartzite-gneiss unit above the beds of diopside gneiss. This horizon would thus correspond to the calcareous quartzite of the middle division in the threefold division by Ransome and Calkins (1908, p. 40).

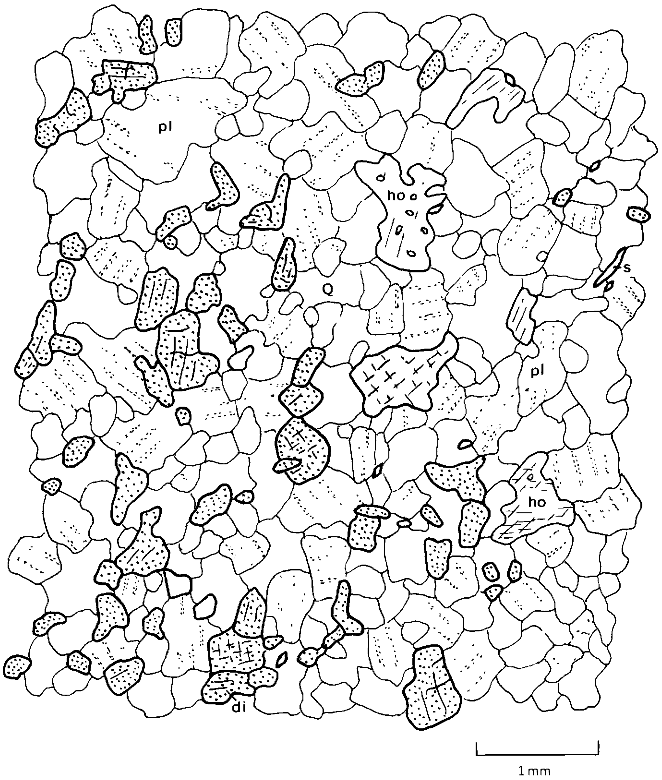
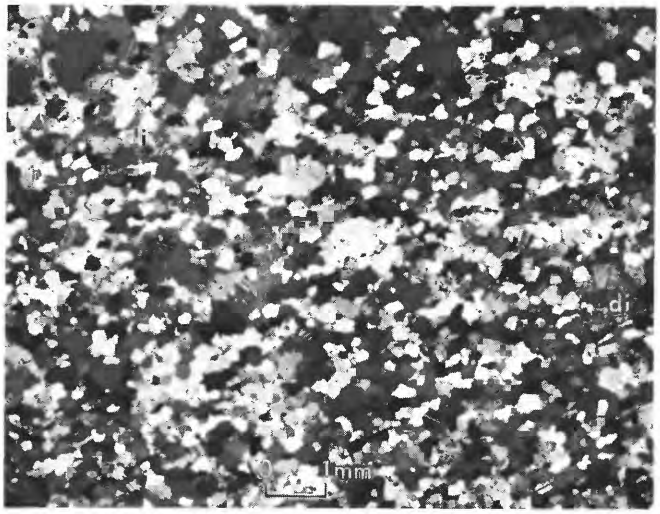


FIGURE 12.—Camera lucida drawing of diopside-plagioclase gneiss along the North Fork of the Clearwater River (in lower right corner of pl. 1). Quartz (Q), plagioclase (pl), Diopside (di), hornblende (ho), sphene (S). This rock has no potassium feldspar.

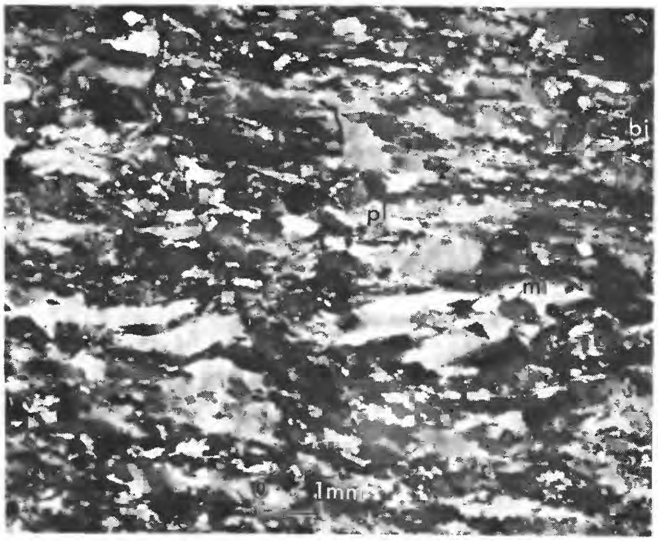
A microscopic study of the white granular quartzite reveals a considerable variation in composition between the individual layers. Most layers are rich in quartz, the percentage of this mineral ranging from 60 to 85 (table 5, No. 1525), but in some layers the amount of albitic plagioclase (An_{6-15}) may be considerable (about 25 to 55 percent, fig. 13*A*). A small amount (3 to 6 percent) of untwinned potassium feldspar is present in these layers. Small grains of brown rutile and zircon occur as accessories.

Muscovite with some biotite or diopside and actinolite are the additional constituents in many layers interbedded with the white granular quartzite; the amount of these minerals ranges from negligible to 20 percent. In contrast to the diopside gneiss, the diopside-bearing layers of the white granular quartzite contain very little potassium feldspar.

In many quartz-rich layers the quartz grains are much larger (1 to 2 mm in diameter) than the albite grains (0.1 to 0.2 mm in diameter) and the mica flakes (0.2 to 0.4 mm in diameter). The quartz grains are strongly deformed (fig. 13*B*) and have sutured borders, whereas the feldspar grains (albite and microcline) occur either rounded or as elongated grains between the quartz grains. In the layers rich in albite, quartz occurs in small round grains and albite fills the interstices. The mica flakes are included in the quartz grains or occur between them. The diopside occurs either as scattered individual grains with rounded corners or small groups of irregular-shaped grains very similar to those in the diopside gneiss. Microcline occurs either as scattered grains with albite or as large grains in certain



A. White granular quartzite containing albitic plagioclase with albite twinning. A few small grains of diopside (di) occur as additional constituents. Texture is granoblastic. This layer is interbedded with biotite quartzite along the east fork of Potlatch Creek about 0.5 mile north of Bloom Meadows. Crossed nicols.



B. A strongly deformed layer in white granular quartzite along the St. Maries River 1.1 miles south of Gold Center (loc. 1525). About 3 percent plagioclase in small round or elongated grains, 1 percent orthoclase in elongated grains, and 1 percent biotite in small flakes occur among the quartz grains. Plagioclase (pl), microcline (mi), biotite (bi). Crossed nicols.

FIGURE 13.—PHOTOMICROGRAPHS OF WHITE GRANULAR QUARTZITE IN THE WALLACE FORMATION.

thin layers. These large grains are elongated parallel to the bedding and include round quartz and albite grains.

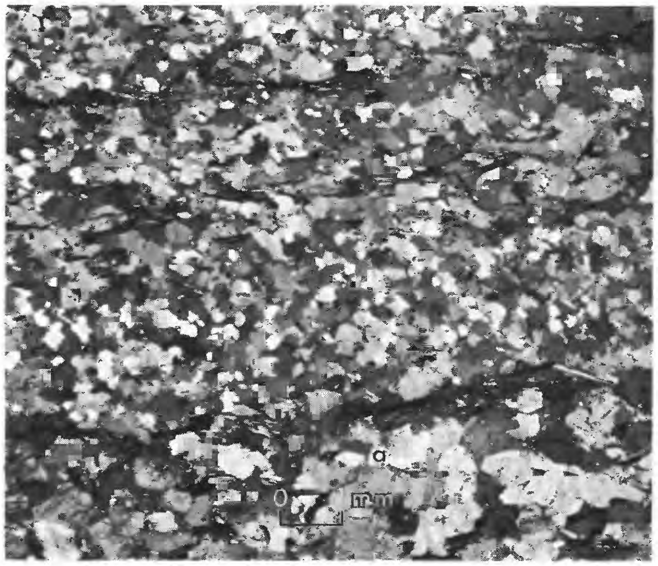
In its mean composition the white granular quartzite is fairly uniform throughout the mapped area. The plagioclase is albitic also in the southernmost part; in most layers it ranges from An_5 to An_{12} and constitutes from 10 to 50 percent of the rock.

THIN-BEDDED BIOTITE QUARTZITE AND GNEISS

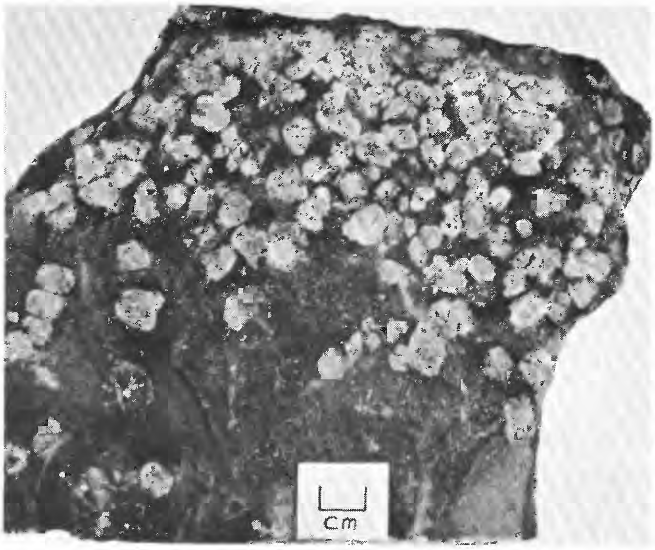
Thin-bedded layers of biotite- and plagioclase-bearing light- to medium-gray quartzite and gneiss form a large proportion of the lower quartzite-gneiss unit. The uppermost part of the upper quartzite-gneiss unit consists mainly of beds of white to light-gray plagioclase quartzite 2 to 10 cm thick separated by thin biotite-bearing layers; layers of thin-bedded gray quartzite rich in biotite, and thus similar to the thin-bedded biotite quartzite of the lower unit, are interbedded also in the upper unit.

The amounts of the four major constituents of the biotite gneiss are highly variable: quartz, 20 to 60 percent; plagioclase, 12 to 50 percent; potassium feldspar, 0 to 60 percent; and biotite, 2 to 25 percent. Numbers 1913, 1908, 1906, and 1520 on table 5 and Nos. 1886 and 1889 on table 6 are representative of this rock type. The dark layers are rich in biotite. The light layers are rich either in quartz and plagioclase or in quartz and potassium feldspar (fig. 14*A*). In some outcrops the plagioclase-rich layers form the major part of the rock and the potassium feldspar occurs only in some thin layers (2 to 3 mm thick) interbedded with thicker plagioclase-rich layers (5 to 10 mm thick). In some other outcrops, however, the ratio between the thicknesses of the plagioclase- and the potassium feldspar-rich layers is about one to four. Layers that contain both feldspars in about equal amounts are also interbedded. Muscovite occurs with biotite in some paper-thin mica-rich laminae that separate the layers (1 to 2 cm thick), consisting of light-colored quartz-feldspar rock with only a small amount of biotite.

The amount of potassium feldspar decreases toward the south; the composition of the biotite gneiss is in this respect comparable to that of the diopside gneiss. All specimens studied along the East Fork of Potlatch Creek have abundant potassium feldspar. In the vicinity of Elk River, the layers rich in potassium feldspar are few and thin, and near the North Fork of the Clearwater River only one contains potassium feldspar. This specimen is from a fine-grained biotite gneiss layer, 3 cm thick, interbedded with biotite schist, and contains about 40 percent quartz, 40 percent plagioclase, 30 percent microcline, and very little biotite. Microcline in this rock occurs as large grains that include many small round quartz grains. In the common biotite gneiss near the North Fork of the Clearwater River, plagioclase (An_{12}) forms about 65 percent of the rock, the amount of quartz is about 25 percent, and that of biotite, muscovite, and accessories combined, about 10 percent. These compositions suggest that the biotite gneiss is the metamorphosed equivalent of arkosic layers. Some of these layers were inter-



A. Thin-bedded biotite gneiss along Stony Creek, 3 miles east of Hemlock Butte (loc. 1520). Layer *a* consists of plagioclase with only a few tiny grains of quartz and flakes of biotite; layer *b* is rich in untwinned potassium feldspar and contains small grains of quartz (white). Most of the biotite occurs in the thin laminae (dark) and is partly altered to chlorite with inclusions of rutile and leucoxene. Crossed nicols.



B. Light-colored round scapolite crystals in dark fine-grained biotite gneiss at Incline Lookout (just north of the northern boundary of pl. 1).

FIGURE 14.—PHOTOMICROGRAPH OF BIOTITE GNEISS AND PHOTOGRAPH OF ROUND SCAPOLITE GRAINS IN BIOTITE GNEISS IN THE WALLACE FORMATION.

bedded with quartzites and dolomitic sandstones, but most of them represent transitional phases between the argillaceous layers and the quartzites or the dolomitic sandstones.

The texture of the biotite gneiss exposed in the northern part of the area differs from that of the gneiss in the southern part. In the northern part the gneiss is fine grained, the quartz occurs in small (0.1 to 0.2 mm in diameter) rounded grains and the feldspars fill the interstices. The small flakes of biotite are between the other minerals or grow through them; most of the biotite flakes are oriented parallel to the bedding, but a few make an angle of about 20° with this plane. Toward the south the grain size becomes larger. Near Elk River the quartz and feldspar grains range from 0.5 to 2 mm in diameter and the mica flakes from 1 to 2 mm in length. Near the North Fork of the Clearwater River the majority of the quartz and plagioclase grains are 1 to 2 mm in diameter and the mica flakes may be as much as 3 mm in diameter. As a rule, the mica flakes in the biotite-rich laminae have a diameter that is two or three times that of the biotite in the light-colored layers.

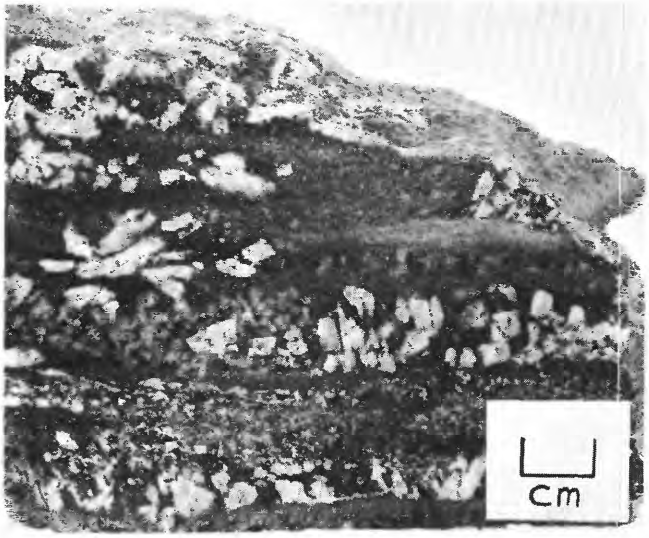
In some localities the biotite gneiss contains scapolite in large- to medium-size white crystals (for example locs. 1904, 1438). This mineral is more common in the northern part than in the southern part of the area; thick layers very rich in large round scapolite grains occur just north of the northern border (fig. 14*B*).

The scapolite grains in a layer 1 mile south of Bloom Meadows (fig. 15*A*) contain numerous small inclusions of quartz and a few grains of epidote (fig. 15*A*). Indices of refraction of scapolite in specimen 1438 are $\epsilon=1.547\pm0.002$, $\omega=1.575\pm0.002$ indicating that this scapolite is mizzonite with 55 percent meionite (Winchell and Winchell, 1951, p. 353). Biotite in this specimen is strongly pleochroic, with Y=Z=reddish brown and X=straw color; it has $\gamma=1.611\pm0.002$ and $2V=0°$. A considerable amount of plagioclase and potassium feldspar occur in this rock.

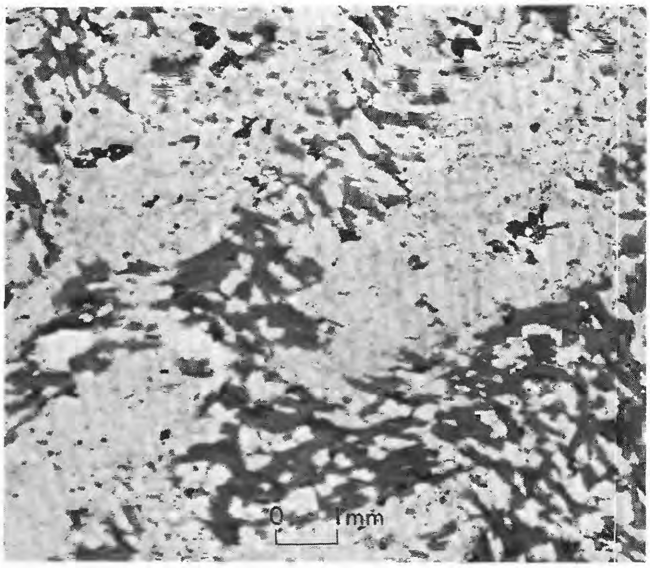
Along the north-central boundary of the area the scapolite is dipyre with about 45 percent meionite ($\epsilon=1.546\pm0.002$, $\omega=1.566\pm0.002$). Biotite is strongly pleochroic with Z=light brown, X=straw color, $\gamma=1.641\pm0.002$, and $-2V=3°$. Epidote, calcite, brown rutile, and zircon are the additional constituents in the scapolite-bearing layers.

SOME CONCLUSIONS BASED ON OPTICAL PROPERTIES OF MINERALS IN THE WALLACE FORMATION

To find out whether or not the composition of the constituent minerals such as biotite, diopside, and amphiboles varies, the optical properties of these minerals were determined in several specimens from various parts of the area. In many layers of the schist, gneiss, and quartzite, biotite is the only ferromagnesian mineral, and it can therefore give an approximate value for the iron-magnesium ratio in the rock. In the diopside gneiss, the diopside usually



A. A hand specimen of a scapolite-bearing biotite gneiss about one mile south of Bloom Meadows (loc. 1438). The white angular crystals are scapolite.



B. Photomicrograph of the rock shown in figure 15*A*. The scapolite grains are light colored and contain small round inclusions of quartz. The rock between the scapolite grains consists of quartz, plagioclase, orthoclase, and biotite. Crossed nicols.

FIGURE 15.—PHOTOGRAPH AND PHOTOMICROGRAPH OF SCAPOLITE-BEARING LAYER IN BIOTITE GNEISS OF THE WALLACE FORMATION.

includes all the iron and magnesium present and serves as a similar indicator.

The optical properties of biotites taken from different environments vary considerably but the biotites from petrographically similar layers are similar. All biotites are strongly pleochroic with Z=brilliant brown and X=straw color; the optic angles are close to zero. The γ index was measured in several specimens, and the values on table 7 show its variation in the biotites from various layers of the Wallace formation. The index of refraction of the biotite in the schist is slightly higher than that of the biotite in the quartzite and gneiss. This higher index of refraction suggests that the biotite in the schist is richer in iron. The biotite in the diopside and hornblende-bearing gneiss and in some of the scapolite-bearing layers has a considerably lower index of refraction, which suggests that the biotite in these layers is rich in magnesium. In all the biotites $-2V$ is small.

The values of γ on table 7 are arranged in geographic order, each column starting from the north and ending in the south. Comparison of the values within each column suggests that there is no regular geographic variation in the composition of biotite, even though in the schist the index tends to be higher in the southern part of the area, suggesting a composition richer in iron.

TABLE 7. *Index of refraction of biotite in various layers of the Wallace formation*

Schist		Biotite-quartzite and gneiss		Scapolite-bearing biotite gneiss		Diopside-biotite gneiss		Hornblende gneiss	
Locality No.	Index of refraction (γ)	Locality No.	Index of refraction (γ)	Locality No.	Index of refraction (γ)	Locality No.	Index of refraction (γ)	Locality No.	Index of refraction (γ)
1851	1. 637	1909	1. 633	1789	1. 641	1246	1. 610	1885	1. 606
1429	1. 643	1616	1. 637	1438	1. 611	1435	1. 590	1886	1. 607
1441	1. 650	1521	1. 630	-----	-----	1445	1. 587	1527	1. 625
566	1. 651	1183	1. 643	-----	-----	-----	-----	-----	-----
		255	1. 625	-----	-----	-----	-----	-----	-----
50–60 percent iron members		40–55 percent iron members							

Localities:

1851	2 miles west of Cedar Butte.
1429	Hemlock Butte.
1441	Jackson Mountain.
566	Little Green Mountain.
1909	Along Merry Creek
1616	3 miles east of Clarkia.
1521	Along West Stony Creek, 3 miles east of Hemlock Butte.
1183	2.7 miles east of Jackson Mountain.
255	North Fork of Clearwater River.
1787	Incline Lookout (just north of the mapped area).
1438	1.9 miles north of Jackson Mountain.
1246	Along East Fork of Potlatch Creek, about 1.6 miles west of Mallory Creek.
1435	Head of Mallory Creek.
1445	About 3 miles northeast of Elk Butte.
1885	West-northwest of Cedar Butte.
1886	West-northwest of Cedar Butte.
1527	Along St. Maries River, south of Gold Center.

The indices of refraction of hornblende, tremolite, actinolite, diopside, biotite, scapolite, and epidote minerals were measured in several specimens and are shown on table 8 in geographic order, the northernmost samples appearing on the top of the list. The estimated volume percentages of minerals for these samples are given on tables 5 and 6.

The indices of refraction of hornblende and diopside are fairly uniform suggesting that there is very little variation in the chemical composition of these minerals. The mean values for the hornblende are $\alpha=1.637$, $\gamma=1.656$, $\gamma\wedge c=23°$. Comparison with the optical properties of a large number of hornblendes given by Winchell (1945, p. 42–43) and Hallimond (1943) suggest that the average hornblende near Clarkia is optically close to the light-green hornblende first described by Beskow (1929, p. 73) and later referred to by Kulling (1933, p. 341, No. 14) from Gabbi, N. Storfjället, Swedish Lappland. Calculation of the formula from the chemical analysis of this hornblende gives the following result:

$$(Na_{0{,}79}K_{0{,}10}Ca_{1{,}83})_{\Sigma=2{,}72}(Mg_{2{,}89}F_{2}\overset{+3}{\,}_{0.22}Ti_{0{,}11}Fe^{+2}_{1.21}Al_{0{,}54})_{\Sigma=4{,}97}(Al_{1{,}22}Si_{6{,}78})_{\Sigma=8{,}0}(OH_{1{,}76}O_{22{,}24})_{\Sigma=24{,}0}.$$

or in the generalized form:

$$(O,OH,F)_2Na_{0-1}\ (Ca,Na)_2Mg_3(Mg,\ Fe,\ Ti^{+4},\ Fe^{+3},\ Al)_2(Al,Si)Si_7O_{22}$$

in which Mg/Fe=2.39.

TABLE 8.—*Optical properties of minerals in the amphibole- and diopside-bearing gneiss of the Wallace formation*

Locality No.	Rock type	Hornblende	Tremolite or actinolite	Diopside	Biotite	Scapolite	Clinozoisite or zoisite
1910	Hornblende gneiss	$\alpha=1.630$ $\beta=1.644$ $\gamma=1.652$ $\gamma\wedge c=23°$	--- ---	--- ---	--- ---	$\omega=1.564$ $\epsilon=1.546$	---
1885	Diopside-hornblende gneiss	$\alpha=1.633$ $\gamma=1.652$ $\gamma\wedge c=23°$	---	---	$\gamma=1.606$	---	---
1917	Scapolite- and calcite-bearing biotite-hornblende gneiss.	$\alpha=1.633$ $\gamma=1.652$	---	---	$\gamma=1.632$	$\omega=1.561$ $\epsilon=1.542$	---
1610	Tremolite gneiss	---	$\alpha=1.607$ $\beta=1.618$ $\gamma=1.631$	---	---	---	---
1919	Diopside gneiss	---	---	$\alpha=1.680$ $\beta=1.686$ $\gamma=1.705$	---	---	$\alpha=1.713$ $\beta=1.715$ $\gamma=1.721$
1920	Diopside-hornblende gneiss	$\alpha=1.640$ $\gamma=1.664$ $\gamma\wedge c=19°$	---	$\alpha=1.681$ $\gamma=1.705$	---	---	---
1437	Diopside gneiss	---	---	$\alpha=1.682$ $\beta=1.687$ $\gamma=1.707$	---	---	---
1527	Thin-bedded hornblende gneiss.	$\alpha=1.634$ $\gamma=1.656$	---	---	$\gamma=1.625$	---	$\alpha=1.693$ $\gamma=1.698$
1445	Diopside-tremolite gneiss	---	$\alpha=1.610$ $\gamma=1.634$	$\alpha=1.671$ $\gamma=1.696$	$\gamma=1.587$	---	---
2012	Diopside-tremolite gneiss	---	$\alpha=1.608$ $\gamma=1.632$	$\alpha=1.673$ $\beta=1.681$ $\gamma=1.701$	---	---	---
1468	Diopside-actinolite gneiss	---	$\alpha=1.618$ $\beta=1.634$ $\gamma=1.642$	$\alpha=1.674$ $\gamma=1.700$	---	---	---
2008	Diopside-actinolite gneiss	---	$\alpha=1.618$ $\gamma=1.642$	$\alpha=1.679$ $\gamma=1.704$	---	---	---
2005	Hornblende gneiss	$\alpha=1.631$ $\gamma=1.651$	---	$\alpha=1.682$ $\gamma=1.705$	---	---	---

In samples from the northernmost part of the area (for example No. 1910), the indices of refraction of hornblende are somewhat lower than the average values near Clarkia; in samples from the southern part they tend to be higher, which indicates that the magnesium-iron ratio ranges from about 2.4 to 4.0 (Winchell and Winchell, 1951).

The minerals of the tremolite–actinolite series occur in light- to dark-green prisms and show a slight variation in optical properties. The indices of refraction range as follows: $\alpha=1.607$ to 1.618; and $\gamma=1.631$ to 1,642, which indicates that the composition ranges from a very pure tremolite to an actinolite end member containing about 10 percent iron.

The diopside that occurs with tremolite shows slightly lower indices of refraction ($\alpha\approx1.672$) than the diopside of the tremolite-free layers ($\alpha\approx1.690$). In the actinolite-bearing layers the indices of refraction of diopside also tend to become higher with the increase of the indices of actinolite. This indicates an approximately equal distribution of the increase of iron between the diopside and the amphibole. The optical properties of the diopside show that its composition ranges from a fairly pure magnesium end member to a diopside containing about 15 percent of iron end member (Winchell and Winchell, 1951, p. 413). Thus all diopsides are rich in magnesium, which together with the composition of the other ferromagnesian minerals show that the original sediment was dolomitic with only a minor amount of iron. In the hornblende- and biotite-bearing layers, the iron content of the metasedimentary rock is somewhat higher—as indicated by a higher iron-magnesium ratio in hornblende (1:3) and in biotite (1:5 to 1:2).

In most layers of the quartzite-gneiss units the diopside, amphiboles, and biotite are the chief carriers of iron and magnesium and the iron-magnesium ratio of these minerals reflects the iron-magnesium ratio of the original sediment. From the study of the minerals it can be concluded that in the layers of diopside gneiss and hornblende gneiss, which is metamorphosed equivalents of dolomitic sandstones, the molecular iron-magnesium ratio is about 1:5 and 1:3, respectively. In the biotite-bearing layers (originally quartzitic and shaly layers) the same ratio varies from 1:4 to 1:1; and in the common mica schist, which contains about 10 percent almandite and 20 percent biotite, the iron-magnesium ratio is about 3:2. The schist layers are equivalent to the slates and blue-gray argillites of the Coeur d'Alene district.

The chemical analyses made of the corresponding layers of the Wallace formation in the Headquarters quadrangle (which lies southeast of the area under discussion) show the following iron-magnesium ratios: 1:3 in diopside-plagioclase gneiss, 1:1 in biotite gneiss, 2:1 and 5:2 in schist (Hietanen, 1962, table 2, Nos. 289, 248, 246, 247).

Another difference in the chemical composition between the mineralogically different layers of the Wallace formation is the variation in the content of potassium and aluminum as reflected in the distribution of biotite. Many horizons consist of remarkably thin beds of diopside gneiss alternating with biotite quartzite and biotite gneiss (fig. 3); this indicates an original rapid and sharp compositional variation. Growth of fairly large crystals of high-grade metamorphic minerals such as diopside, garnet, sillimanite, kyanite, and biotite has made this compositional variation strikingly visible.

METAMORPHIC FACIES OF THE BELT SERIES

The abundance of pelitic and calcareous material in the Wallace formation near Elk River and Clarkia makes this formation especially useful for the study of variation in the degree of metamorphism. The distribution of index minerals such as muscovite, biotite, garnet, staurolite, kyanite, sillimanite, epidote, hornblende, and diopside provides means of following the changes in the relative temperature-pressure field of the recrystallization. The various units of the Wallace formation extend over the whole area from the northwest corner to the southeast, allowing possible changes in the mineral assemblages of each unit to be followed over a distance of about 25 miles. The metamorphic grade increases slightly from northwest to southeast. The area can be divided into three parts, each with a characteristic metamorphic grade, as indicated by typical mineral assemblages in the schists and gneisses: (1) the northern part, north and northeast of Clarkia; (2) the central part, between Anthony Peak and Hemlock Butte; and (3) the southern part, south of Hemlock Butte.

The common mineral assemblages in various layers in the schist exposed in the northern part of the area (T. 43 N. across the map, pl. 1) are as follows:

1. (Andalusite) - almandite - biotite - muscovite-albite-zoisite-quartz.
1b. (Andalusite) - almandite - biotite - muscovite - quartz.
2. Muscovite - biotite - microcline - albite - clino - zoisite-quartz.
2a. Muscovite - biotite - albite - clinozoisite - quartz with or without scapolite.
2b. Muscovite - biotite - microcline - plagioclase (An$_8$)-quartz.

Quartzo-feldspathic layers have the mineral assemblage:

2c. Biotite - microcline - epidote - plagioclase (An$_{8-10}$)-quartz.

Epidote is stable with hornblende and albitic plagioclase in the garnet amphibolite with the following mineral assemblages:

3. Hornblende - almandite - biotite - epidote -plagioclase (An$_{10}$)-quartz.
3a. Hornblende - almandite - epidote - plagioclase (An$_{10}$) - quartz.

Hornblende-biotite gneiss which is common in the area has the following mineral assemblage:

3b. Biotite - hornblende - epidote - plagioclase (An$_{10-15}$) - quartz.

Mineral assemblages in the gneissic layers are as follows:

4a. Hornblende - biotite - microcline - calcite - quartz.
5. Hornblende - biotite - microcline - clinozoisite - plagioclase (An$_{5-15}$) - quartz.
5a. Hornblende-biotite-microcline-quartz.
6. Hornblende - microcline - calcite - clinozoisite - plagioclase (An$_{5-15}$) - quartz.
6a. Hornblende - microcline - (clinozoisite) - plagio - clase (An$_{5-15}$) - scapolite - quartz.
7a. Hornblende - biotite - calcite - scapolite - quartz.
7b. Biotite - calcite - epidote - albite - scapolite - quartz.
8a. Biotite-microcline-calcite-albite-quartz.
9a. Microcline-biotite-tremolite-(albite)-quartz.

These mineral assemblages indicate the conditions of the biotite-almandite subfacies of the epidote-amphibolite facies. Occurrence of andalusite would indicate a higher temperature subfacies.

In addition to aluminum- and calcium-rich minerals, abundant microcline, muscovite, and biotite occur in many layers and make it necessary to include potassium in graphic presentation. On the other hand, ferromagnesian minerals form an essential part of the rocks. Iron substitutes only for a small portion of magnesium in diopside, hornblende, and biotite, whereas the garnets are mainly iron-aluminum compounds. In the graphic presentation, iron and magnesium can be considered together and the number of the essential components is reduced to the following four: A= aluminum, C=calcium, F=ferrous iron+magnesium+manganese, and K=potassium. On figures 16–18, these four components are plotted at the corners of a compositional tetrahedron, within which the compositions of minerals are represented by appropriate atomic ratios. For example, the point for muscovite $(OH)_2KAl_3Si_3O_{10}$ divides the AK edge of the tetrahedron in the ratio 1 : 3, and that for microline in the ratio 1 : 1. The composition of biotite analyzed from the schist in the nearby Boehls Butte quadrangle (Hietanen, 1963) was used to find the point for the biotite. The points for biotite and phlogopite are on the AKF face and those for hornblende on the ACF face, which is the back face of the tetrahedron.

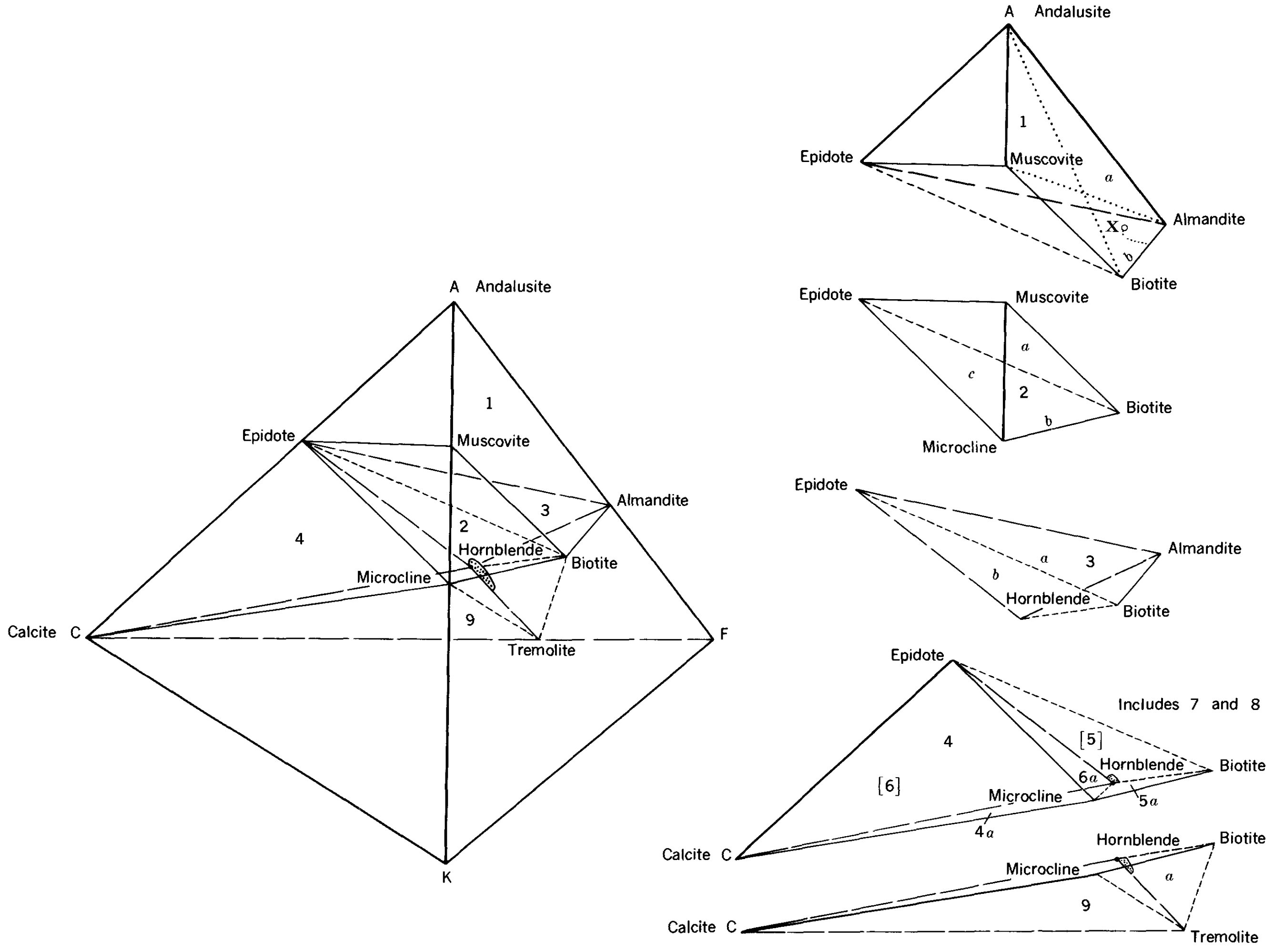


FIGURE 16.—ACFK diagram for the rocks of the biotite-almandite subfacies of the epidote-amphibolite facies. Aluminum (A), calcium (C), ferrous iron, magnesium, and manganese combined (F), potassium (K)—all in atomic ratios. The points for muscovite and microcline are on AK edge and the point for biotite on the AKF face of the transparent tetrahedron. The fields of stability for various mineral assemblages are determined by placing planes through points for minerals that can coexist. These planes divide the tetrahedron into pyramidal sections which have a triangular or quadrilateral base and a vertex in the point for epidote. The pyramidal sections are shown separately on the right. The lines on the back face (ACF) of the tetrahedron are broken lines, those going through the tetrahedron are stippled. Each pyramidal field of stability carries the number of the mineral assemblage mentioned in the text that is stable within it. The numbers with a letter (*a*, *b*, *c*) refer to the faces of these pyramidal sections. For instance, the mineral assemblage 3 is stable in all rocks whose composition would fall within the pyramid (3) that has the triangle almandite–biotite–hornblende as base and whose vertex is in the point epidote. The mineral assemblage 3*a* is stable in all rocks whose composition is on the face: hornblende–almandite–epidote.

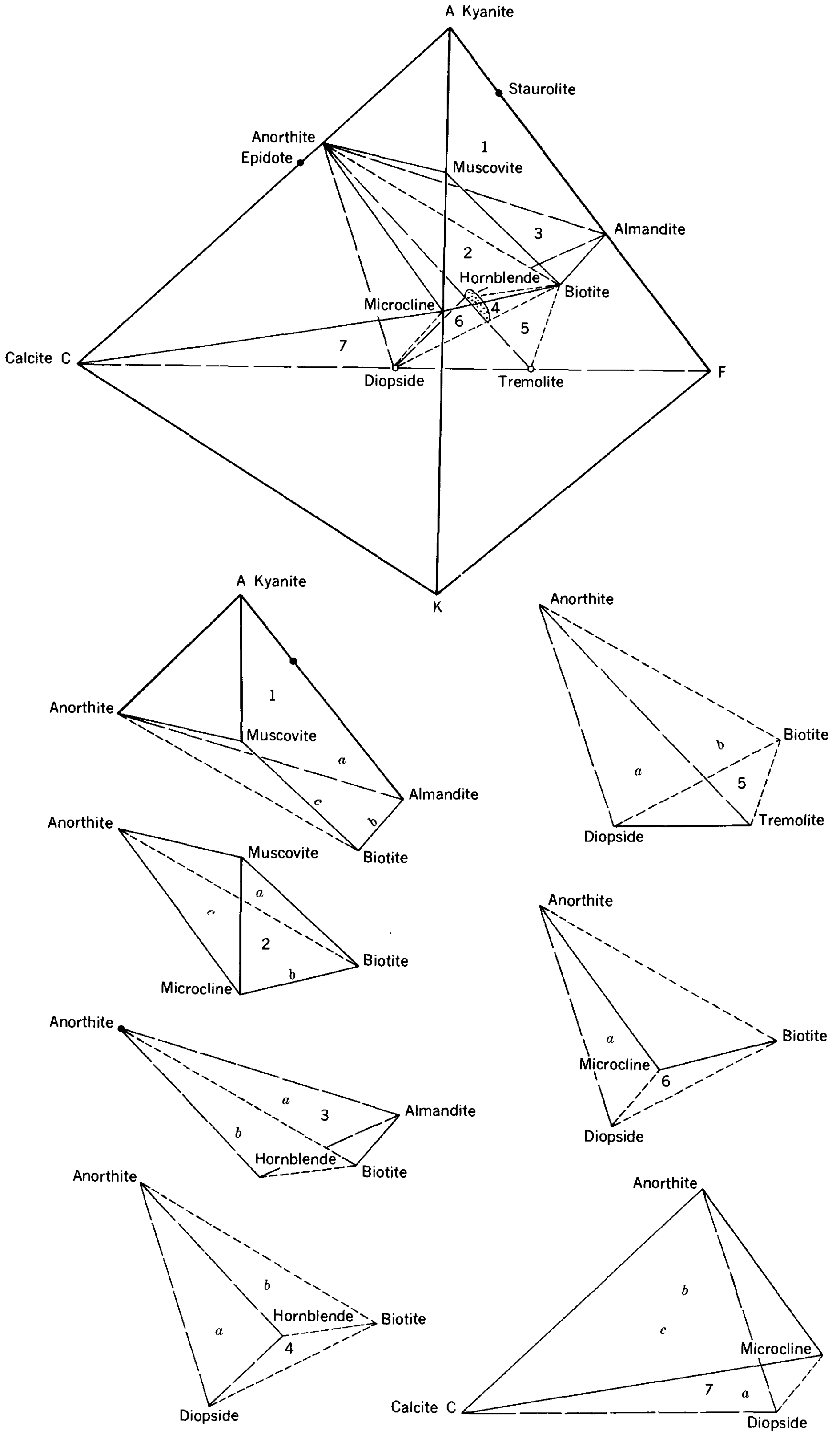


Figure 17.—ACFK diagram for the rocks of the staurolite-kyanite subfacies of the epidote-amphibolite facies. Aluminum (A), calcium (C), ferrous iron, magnesium, and manganese combined (F), potassium (K). The fields of stability for the mineral assemblages mentioned in the text (1–7) are constructed in the same way as in figure 16.

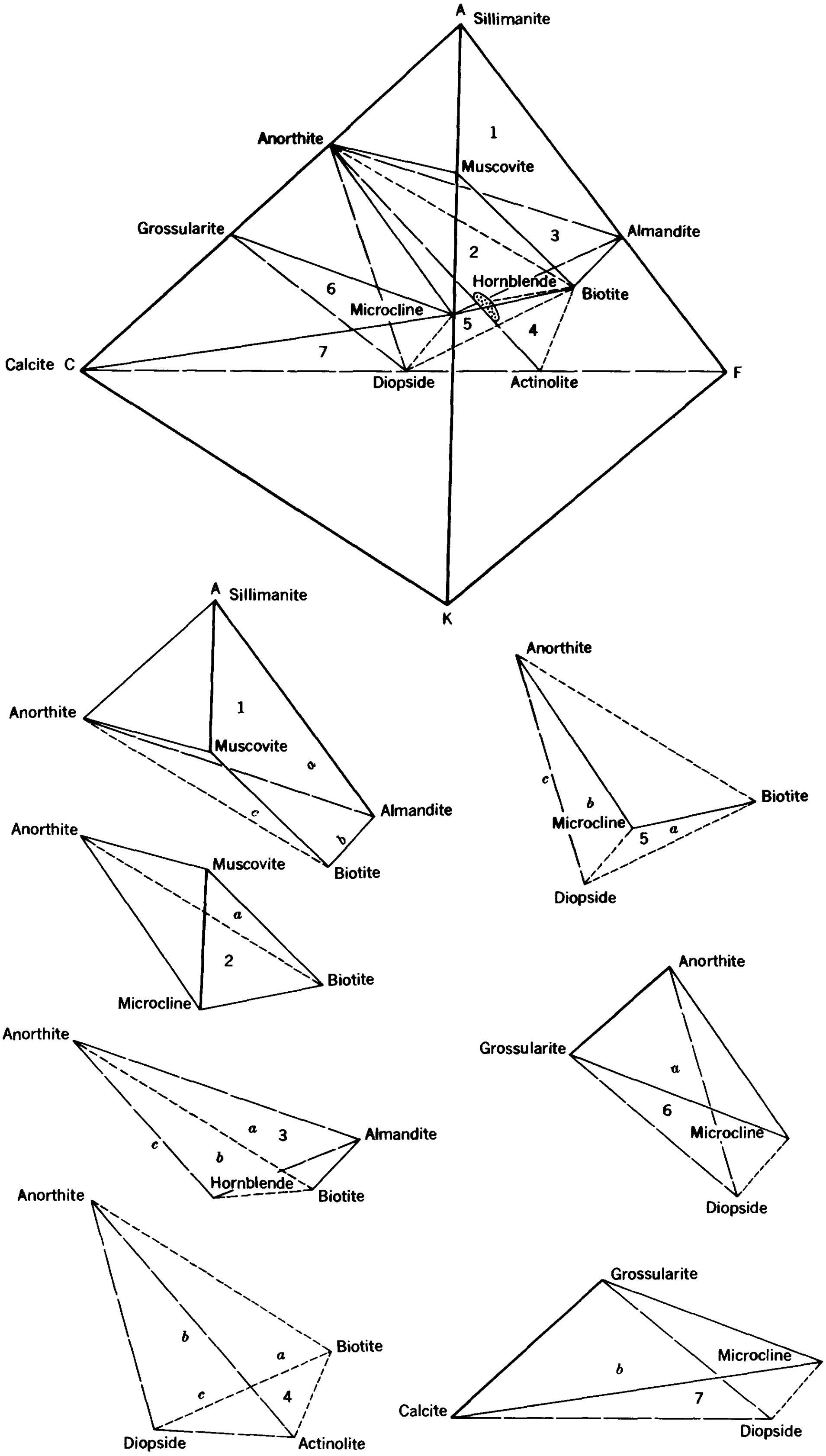


FIGURE 18.—ACFK diagram for the rocks of the sillimanite-muscovite subfacies of the amphibolite facies. Aluminum (A), calcium (C), combined ferrous iron, magnesium, and manganese (F), potassium (K). The fields of stability for the mineral assemblages mentioned in the text (1–7) are constructed in the same way as on figure 16.

If the index minerals of each assemblage are connected by lines, and planes constructed through each three or four points, these planes will intersect along the lines and divide the tetrahedron into pyramidal sections and subtetrahedra. Each of these subtetrahedra will include the bulk compositions of all rocks in which the mineral assemblage defined by the corners of the subtetrahedron is stable. The subtetrahedra and the pyramidal sections are keyed by numbers to the tetrahedron. The numbers with letters *a*, *b*, and *c* in the text refer to the triangular and quadrilateral faces of these sections. It is apparent that only a portion of possible mineral assemblages is represented by the thin sections studied. The plots for the bulk compositions of rocks with only three stable index minerals, such as the assemblage biotite-muscovite-epidote-(albite-quartz) would fall on the plane biotite–muscovite–epidote. The plots for the bulk compositions of rocks with only two index minerals will fall along the lines connecting the points for these two minerals.

The principle of this presentation is the same as used by Eskola (1915, 1939) and later by Turner (1948), Barth (1936, 1952), Thompson (1957), Fyfe, Turner and Verhoogen (1958), and many others, and the same theoretical considerations concerning the equilibrium assemblages are valid. The 4-phase assemblages in this tetrahedral projection form from any bulk composition that plots within subtetrahedra determined by the index minerals at the corners. The compositions at the corners respond to variation in P, T, and μH_2O during the recrystallization, giving different diagrams for various facies and subfacies (compare figs. 16 to 18).

The assemblages of more than four minerals are of particular interest. According to the phase rule, the number of minerals cannot exceed that of the components if the assemblage was formed in equilibrium conditions, or then the number of variants P, T, μH_2O is reduced to two (bivariant equilibrium) or to one (univariant equilibrium). Such an assemblage would be, for instance, No. 1 on figure 16: andalusite-almandite-biotite-muscovite-epidote-(albite-quartz) the bulk compositions of which would be within a four-sided pyramidal section. If the fourth component, calcium, is eliminated, the tetrahedron ACFK would be reduced to the triangle AKF and in this diagram the assemblage andalusite-muscovite-biotite-almandite has four phases instead of three. Similar cases have been noted by many other authors; Turner and Verhoogen (1951, p. 425) consider the common occurrence of the assemblages as a sign that they were formed in equilibrium and suggest that this situation may arise because the components in the graphic presentation are too few. Thompson (1957, p. 856–857) considers that these cases with too many phases include discontinuities in facies, and that these discontinuities represent conditions of either univariant or bivariant equilibrium—the latter when water is gained or lost.

In the Elk River–Clarkia area several assemblages with five phases were noted as shown with the pyramidal sections on figures 16 to 18. Some of these involve a change either of pressure or of temperature and some probably are combinations of normal trivariant systems. An example of the former type would be the assemblage andalusite-almandite-biotite-muscovite-epidote of the andalusite subfacies (fig. 16 for andalusite) and of the latter type the assemblage hornblende-biotite-microcline-calcite-epidote.

The field of stability for the andalusite-bearing rocks would be the four-sided pyramidal section illustrated on figure 16–1. By joining the opposite corners of the quadrilateral base this section can be divided into subtetrahedra in two ways (dotted lines on fig. 16–1). The assemblages, such as almandite-biotite-muscovite-epidot-(albite-quartz), crystallize in stable equilibrium from the bulk compositions that plot in the subtetrahedron determined by the four critical phases at the corners. In this case no andalusite appears and the almandite corner is joined with the muscovite corner. From some bulk compositions, such as X on figure 16, the above assemblage as well as the assemblage andalusite-almandite-biotite-epidote-(albite-quartz) can crystallize; if no muscovite appears the andalusite corner is joined with the biotite corner. From the bulk compositions that plot along the join between epidote and the intersection of almandite–muscovite and andalusite–biotite joins, all four critical phases at the corners of the quadrilateral base may crystallize with epidote, albite, and quartz. The ratio in the amounts of the two pairs at opposite corners of the quadrilateral base may vary to a certain extent, as determined by substitution of iron by magnesium in almandite and biotite. The reaction involved can be expressed as follows:

$$\begin{array}{lccc} & \text{1 biotite}+ & \text{4 andalusite}+ & \text{2 quartz}\rightleftharpoons \\ \text{Sp. gr-----} & 2.88 & 3.15 & 2.65 \\ \text{Mol. vol---} & \multicolumn{3}{c}{\underbrace{355.42+\quad 4\times51.43+2\times22.66}_{606.46}} \end{array}$$

$$\begin{array}{cc} \text{2 almandite}+ & \text{1 muscovite} \\ 4.12 & 2.90 \\ \multicolumn{2}{c}{\underbrace{2\times120.79+\quad 274.62}_{516.20}} \end{array}$$

This reaction takes place without gain or loss of water, is thus univariant, and involves a change in the topology of the projection. The right-hand side of this equation would have somewhat smaller molecular volume; thus a higher pressure might accelerate the crystallization of garnet and muscovite instead of biotite and andalusite. The pseudomorphs of muscovite after andalusite, west of Cedar Butte, could be a result of such a reaction. Alternately, a higher temperature would favor the formation of biotite and andalusite instead of garnet and muscovite. The above equation is valid only if the iron-magnesium ratio in the biotite is the same as that in the almandite. Generally biotite contains more magnesium and the excess of iron crystallizes as magnetite.

To visualize the fields of stability of the hornblende-, biotite-, and microcline-bearing assemblages in three dimensions, the ACFK tetrahedron was constructed from wire and paper. The critical minerals shown on figure 16 were plotted in it and connected with pieces of straight wire. It was apparent that with a slight variation in the aluminum content of hornblende and biotite, the calcite-biotite join can be in the same plane as the hornblende-microcline join, or above or below it. The common occurrence of the mineral assemblages hornblende-biotite-microcline-epidote, hornblende-biotite-microcline-calcite, and hornblende-biotite-calcite-scapolite—all with albite and quartz—in thin adjoining layers indicates that aluminum is distributed between hornblende and biotite in such a manner that the points for hornblende, biotite, microcline, and calcite will fall into the same plane (base of pyramid 4, fig. 16). In some layers these four minerals occur together with scapolite (for example, No. 1917, table 5). Theoretically either epidote or tremolite should be stable with these four minerals.

The mineral associations found in the thin sections suggest that, with a slight variation in the composition of the rock, 1 or 2 of the 4 minerals hornblende, biotite, microcline, and calcite can be missing and the rest of the minerals give assemblages which are numbered from 5 to 8 on p. 72. The field of stability for each of these assemblages would be a subtetrahedron within the 4-sided pyramid 4 or any of the faces of the subtetrahedrons and represent normal trivariant systems.

In the schist of the central part of the map area (T. 42 N., across the map) staurolite appears with kyanite. Large prism-shaped groups of small crystals of kyanite and a small amount of staurolite occur on Anthony Peak. In the vicinity of Stony Butte small crystals of kyanite and staurolite occur with large muscovite and biotite flakes. Toward the south the amount of kyanite decreases, staurolite disappears, and sillimanite appears in increasingly large quantities. On Hemlock Butte kyanite is rare, but sillimanite abounds. None of the thin sections of schist of the Wallace formation in T. 41 N. contain staurolite. Muscovite abounds in some layers, but its total amount definitely decreases toward the south. The common mineral assemblages in the schist in the central part (T. 42 N.) are (fig. 17):

1. Kyanite-staurolite-almandite-biotite-muscovite-plagioclase-quartz,
1*a*. kyanite-staurolite-almandite-biotite-muscovite-quartz,
1*b*. almandite-biotite-muscovite-quartz,
1*c*. almandite-biotite-plagioclase-quartz.

The following mineral assemblages are common in quartzofeldspathic layers:

2. Muscovite-biotite-microcline-plagioclase-(An_{15-20})-quartz,
2*a*. muscovite-biotite-plagioclase(An_{20})-quartz,
2*b*. muscovite-biotite-microcline-quartz,
2*c*. biotite-microcline-plagioclase(An_{20})-quartz.

Garnet amphibolite shows the following mineral assemblages:

3. Horneblende-almandite-biotite-plagioclase (An_{30})-quartz,
3*a*. hornblende-almandite-plagioclase(An_{30})-quartz,
3*b*. hornblende-biotite-plagioclase(An_{20-30})-quartz.

The calcium-rich layers in the quartzite-gneiss units of the Wallace formation in the central part of the area contain abundant diopside, but rarely epidote minerals. Typical mineral assemblages are as follows:

4. Diopside-hornblende-biotite-plagioclase(An_{20})-quartz.
4*a*. diopside-horneblende-plagioclase(An_{20})-quartz,
4*b*. diopside-biotite-plagioclase(An_{20})-quartz,
5. diopside-tremolite-biotite-plagioclase(An_{12-30})-quartz,
5*a*. diopside-tremolite-plagioclase(An_{12-30})-quartz,
5*b*. tremolite-biotite-plagioclase(An_{12-30})-quartz,
6. diopside-biotite-microline-plagioclase(An_{20})-quartz,
6*a*. diopside-microcline-plagioclase(An_{20})-quartz,
7. diopside-calcite-microcline-plagioclase(An_{20})-epidote-quartz,
7*a*. diopside-calcite-microcline-quartz,
7*b*. diopside-calcite-plagioclase(An_{20})-epidote-quartz,
7*c*. microcline-calcite-plagioclase-epidote-quartz.

These mineral assemblages are those commonly found in the staurolite-kyanite subfacies of the epidote-amphibolite facies. In the graphic presentation kyanite occurs in the A corner of the tetrahedron (fig. 17). The point for staurolite $HFe_2Al_9Si_4O_{24}$ divides the AF edge in ratio 2:9. Diopside is stable with tremolite and hornblende. Oligoclase instead of albite is common and epidote occurs with it. The mineral assemblages that are stable in the pyramidal sections of the tetrahedron are numbered from 1 to 7; those with a letter *a*, *b*, *c* represent mineral assemblages that are stable on the faces of the pyramidal sections.

Light-green hornblende occurs with diopside in some localities, and tremolite coexists with calcite and diopside, suggesting a middle-grade subfacies. The zone in which kyanite occurs with sillimanite is fairly narrow and is considered to represent a transition from the kyanite-almandite subfacies to the sillimanite-muscovite subfacies. Anthophyllite was found only in one locality about a mile west of Freezeout Mountain where it occurs with plagioclase and biotite.

In the southern part of the area (Tps. 39 and 40 N.) only sillimanite, but no kyanite nor staurolite, occurs in the schists, which suggests that the degree of metamorphism there is higher than in the central part. The grain size becomes larger toward the south and in many places pegmatitic veinlets, ranging from 3 to 20 mm in width, occur in the schist parallel to the bedding. The amount of biotite is larger and muscovite smaller than in the central part. Typical mineral assemblages in the schist are as follows:

1. Sillimanite-almandite-biotite-muscovite-plagioclase(An_{20-35})-quartz,
1*a*. sillimanite-almandite-biotite-muscovite-quartz,
1*b*. almandite - biotite - (muscovite) - plagioclase-(An_{20-35})-quartz,
2*a*. biotite-muscovite-plagioclase(An_{20-35})-quartz.

Garnet in the amphibolite is mainly almandite with some pyrope. The mineral assemblages are as follows:

3. Hornblende - garnet - biotite - plagioclase(An_{36})-quartz.
3*a*. hornblende-garnet-plagioclase(An_{36})-quartz,
3*b*. hornblende-garnet-plagioclase(An_{36})-quartz,
3*c*. horneblende-plagioclase(An_{30-36})-quartz.

The gneissic layers are somewhat coarser than the corresponding layers in the central part, and in many places the bedding is obscured by intense recrystallization. The most marked change in the mineralogy is the scarcity of potassium feldspar. Typical mineral assemblages are as follows:

4. Diopside-biotite-actinolite-plagioclase(An_{30-85})-quartz,
4*a*. biotite-actinolite-plagioclase(An_{35})-quartz,
4*b*. diopside-biotite-plagioclase(An_{30})-quartz,
4*c*. diopsite-actinolite-plagioclase(An_{12-30})-quartz,
5. diopside-biotite-microcline-plagioclase (An_{12-30})-quartz.
5*a*. diopside-biotite-microcline-quartz,
5*b*. diopside-microcline-plagioclase(An_{12-30})-quartz,
5*c*. diopside-plagioclase(An_{30-85})-quartz,
6*a*. diopside-grossularite-plagioclase(An_{86})-quartz,
7*b*. diopside-calcite-grossularite-quartz.

In the graphic presentation sillimanite occurs in the A corner of the tetrahedron (fig. 18). Grossularite is found in some calcium-rich layers, but microcline, though a stable compound in this subfacies, is usually missing. This suggests a compositional limitation of the gneissic layers close to the ACF face. Biotite is common in all other assemblages except in Nos. 5*c*, 6a, and 7*b*. Magnesium-rich layers have actinolite instead of tremolite; staurolite is missing. Epidote minerals appear only in hydrothermally altered parts of the rocks next to shear zones and to intrusive bodies. In every subfacies the composition of plagioclase varies to a certain extent according to the amount of calcium present in the original sediment.

Actinolite is present only in a few thin laminae and layer- and lens-shaped bodies in the diopside-plagioclase gneiss; its dark-green color contrasts strongly with the light grayish-green host rock. The mode of occurrence of actinolite indicates that it crystallized in the magnesium-rich portions of the original dolomitic sandstone; its existence reflects the chemical composition of the parent sediment rather than a local lower temperature-pressure field. Diopside crystallized where enough calcium was present, which, together with the mineral assemblages listed above, proves that in the southern part of the area the conditions during the metamorphism were those of the sillimanite-muscovite subfacies of the amphibolite facies.

As described above staurolite is always accompanied by kyanite, whereas staurolite-almandite rocks are well developed in the schist of the Wallace formation about 30 miles east of Clarkia, where they are interbedded with layers that contain kyanite in addition to staurolite and almandite. Only almandine, biotite, and muscovite occur in the specimens from the schist of the Prichard formation north of the kyanite-bearing rock, north of Freezeout Mountain and of White Rock. The lack of staurolite in this schist is probably due to an unsuitable chemical composition rather than to differences in temperature and pressure since staurolite appears again in schists east of Freezeout

Mountain. Dark prisms which may have been staurolite but are completely altered to sericite, biotite, and magnetite are present in the garnet-biotite-muscovite schist 4½ miles west of Clarkia, about 1 mile west of the map area.

Staurolite disappears with crystallization of sillimanite, but the stability range of kyanite overlaps that of sillimanite between White Rock and Hemlock Butte. These stability relations are shown on table 9 by arrows extending over the lines separating the fields of various subfacies.

Almandite, biotite, and muscovite are stable over the whole area, but their amount and grain size varies as indicated on table 9. Plagioclase becomes more calcic toward the south. Clinozoisite is the major epidote mineral in the northern part, but zoisite and epidote also occur. In the southern part, epidote is confined to hydrothermally altered portions of the gneisses.

Potassium feldspar appears in notable quantities only in two specimens from the southern part. One of these specimens, No. 1992, table 5, is from a layer interbedded with the white granular quartzite between Weitas and Falls Creeks; the other one (No. 266) is from a layer of migmatized biotite gneiss on the North Fork of the Clearwater River. Microcline occurs in both samples as large grains that include small round quartz grains. The abundance of microcline in the rocks of the epidote-amphibolite facies and its scarcity in the rocks of the amphibolite facies, is contrary to the normal distribution of this mineral in regionally metamorphosed areas. This abnormal distribution may be due either to a lateral variation in the composition of the original sediment or to a later metasomatic rearrangement of elements near intrusive bodies, as will be discussed on page C43.

STRUCTURE OF THE BELT SERIES

The structure on a regional scale is discussed in another report (Hietanen, 1961a). Attention here is called only to some local features.

BEDDING

In most rocks the bedding is distinct because of sharp differences in the mineral content of adjacent layers. This structural element is especially striking in the thin-bedded gneisses (fig. 3) where dark-gray biotite-bearing layers alternate with either light-green diopside-bearing or light-gray to white plagioclase-bearing layers. In the quartzites the variation in the amount of micaceous minerals and feldspars makes the bedding recognizable, but it is not nearly as striking as it is in the gneisses. The thickness of individual beds in the gneisses and quartzites has been given in connection with the stratigraphic and petrographic descriptions.

In the schists, the bedding is not as pronounced as in the gneisses; it is however, recognizable in most outcrops. The thickness of the individual beds ranges from a few centimeters to a meter; thicknesses between 10 and 30 cm are most common. Bedding is distinct in the localities where layers of biotite gneiss or quartzite are interbedded with the schist. The thickness of these beds is, as a rule, less than that of the micaceous layers, and ranges from 1 to 10 cm. However, thicker layers consisting of laminated biotite gneiss are interbedded with the schist. The laminae in these layers consist of large muscovite flakes.

FOLDING, AXIS OF FOLDING, AND LINEATION

In the northern part of the area the rocks of the Belt series are generally gently folded; the amplitudes of the folds are smaller than their wave length of 1 to 2 miles. Some folds, however, are strongly overturned, especially near fault zones. The flanks of folds are usually straight, as shown in the topography by many dip slopes east of Clarkia, and the crests are rather sharply bent. Toward the south the folding becomes more intense, and along the North Fork of the Clearwater River steeply dipping flanks of isoclinal folds are exposed. In the eastern part of the area the axes of most large folds plunge to southeast, and in the western part to the west-northwest; but there are many local deviations.

In addition to large folds there are folds of outcrop size or smaller, the axes of which coincide with the direction of lineation as discussed below. The small folds are more common in the southern part, and some migmatized parts of the strata are intricately folded. Migmatitic veins, formed by segregation of quartzofeldspathic material, parallel the bedding.

Lineation is well developed in schist and in micaceous layers in gneiss and quartzite. It is a minor wrinkling of bedding surfaces around an axis which makes an angle from 45° to 90° with the axes of major folds. In many localities tight folds of considerable size were developed around the axis of wrinkling, and these folds were folded a second time around the major axis.

An example of this type of folding is well exposed in an area 3 miles southeast of Elk River, where the axes of small overturned folds plunge 10° SW. (fig. 19), and the axes of large gentle folds are horizontal, striking eastward. The folding on this east-west axis is later, and a still later movement on the axis that plunges 10° southeasterly caused a doming of previously folded strata. Large folds on axes that strike

TABLE 9.—*Stability ranges of critical minerals in schists and calcareous gneisses, northwest of the Idaho batholith*

		(Andalusite)	*Almandite*	*Muscovite*	*Biotite*	*Quartz*	*Light-green hornblende*	*Tremolite*	*Clinozoisite*	*Plagioclase*	*Microcline*
Increase mainly in pressure	Biotite-almandite subfacies of the epidote-amphibolite facies in T. 43 N., R. 1 to 3 E. (T=380° to 400°C, P about 5,000 bars).	↓	│ amount │ crystals	│ amount	│ amount	│	│	│	│	An_{5-15} ↓	│
	Staurolite-kyanite subfacies of the epidote amphibolite facies in T. 42 N., R. 1 to 3 E. (T=400° to 470°C, P about 5,500 bars).	*Kyanite, staurolite* ↓ ↓	│ increases │ grow larger	│ decreases	│ increases	│ amount decreases	↓ *Diopside* │	↓	│	An_{10-20} ↓	↓
Increase mainly in temperature	Sillimanite-muscovite subfacies of the amphibolite facies in T. 39 to 40 N., R. 1 to 3 E. (T=470° to 580°C, P about 5,500 bars).	↓ *Sillimanite*	↓	↓	│	│	*Common hornblende* ↓ │ increases	*Actinolite* ↓	*(Epidote)* ↓	An_{20-35} increases ↓ amount	
	Quartz diorite melting point.				decreases ↓	↓	↓			An_{35-38} ↓	

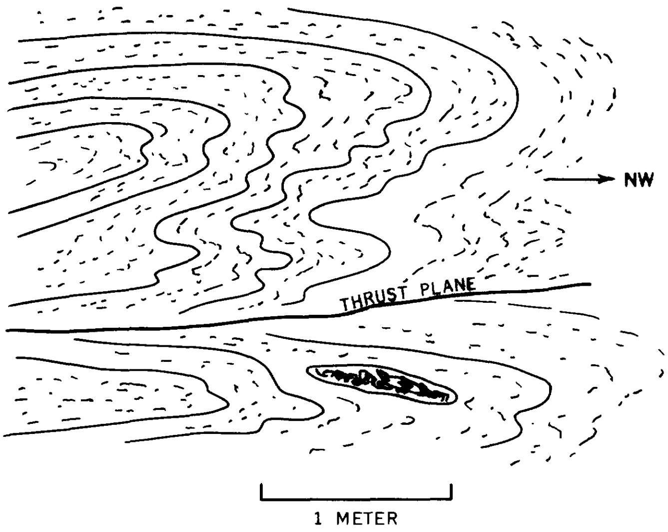


FIGURE 19.—Small folds overturned to the northwest in thin-bedded biotite-diopside-plagioclase gneiss just east of logging camp 43, 3 miles southeast of Elk River, a vertical road cut facing to northeast. Thrust plane dips gently to the southeast.

in the same direction as the lineation in nearby outcrops are seen 2 miles southeast of Anthony Peak in the north-central part of the area, around the mouth of Glover Creek, and just south of Breakfast Creek along the eastern edge of the map area.

In schist of the Prichard formation in the northeastern part of the area, lineation plunges either 15°W. or 20° E. because of gentle folding around the axis that plunges 10° S.E. (see for instance pl. 1, sec. 23, T. 43 N., R. 3 E.). Folds around the lineation are common, but here it is not clear whether they are earlier or later than the major folding on the southeasterly axis.

In most outcrops the direction of lineation deviates 50° to 85° from the direction of the fold axis. This relation, together with the development of fairly large folds around the lineation, suggests that the lineation represents an independent structural element, a second folding, which in some localities (fig. 19) is earlier and in some others later than the major folding.

The general direction of the major fold axes coincides with the regional trends of the south end of an arcuate segment of Nevadan folding that strikes southward in British Columbia and southeast north of the Idaho batholith (Eardley, 1951). The wrinkling and second folding around the northeast- or southwest-plunging axes is analogous to a similar structure south of the Elk River–Clarkia area (Hietanen, 1962). This structure parallels the regional trend of the north end of an arcuate segment of Nevadan folding that strikes north in the Klamath Mountains in California, northeast across Oregon and joins the northern segment just north of the Idaho batholith. The area under investigation lies within the junction of these two arcuate segments. Both segments were formed over the same period of time, but in any locality within the junction the folds on the axis of the northern arc may have been formed either earlier or later than those on the axis of the southern arc.

Small folds on both sets of axes are locally overturned, those on the westerly set to the north (fig. 20) and those on the southeasterly set to the southwest.

A structure resembling overturning but occurring only in certain beds of diopside gneiss is interpreted to be due to sliding of the original sediments under water before their consolidation (fig. 7*B*) because the beds just above and below the disturbed strata are not folded synchronously.

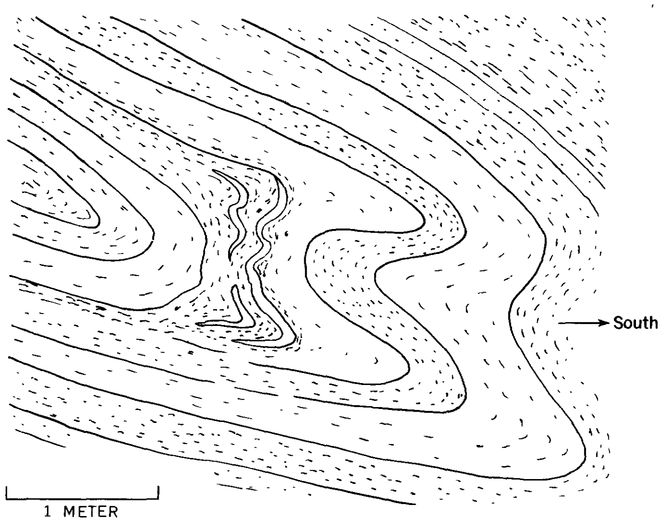


FIGURE 20.—A small syncline in a fine-grained gneissic layer in the schist of the Prichard formation along West Stony Creek at the mouth of Glover Creek. Road cut faces to the west. Overturning is to the north; the axial plane dips gently to the south.

FOLIATION

Foliation is well developed in all micaceous layers but is not discernible in layers consisting of diopside-plagioclase gneiss and quartzite. The direction of foliation and its relation to the bedding varies. Where the folding is gentle and the bedding is strongly developed as, for instance, in most of the thin-bedded biotite and diopside gneiss, the foliation is parallel to bedding. A transecting cleavage is developed in strongly folded parts of the strata and near many fault zones. It is pronounced in micaceous layers of the gneisses and in fine- to medium-grained layers of the schist, especially along the crests of folds. In general the transecting cleavage is more common in the northern and in the central part of the area than in the southernmost part, where the folding is more intense. This seems contradictory, but can be explained to be due to a different type of folding and to the difference in the plasticity of the material folded. On the flanks of the isoclinal folds along the North Fork of the Clearwater River, the foliation either parallels the bedding or deviates only 5° to 20° from it. In the intricately folded migmatized parts of the schist and gneiss there, the foliation as well as the quartzofeldspathic veins parallel the bedding. Segregation of the quartzofeldspathic material accentuated the bedding and encouraged bedding-plane slippage during folding.

FAULTS

A major north-trending zone of near vertical faults extends across the north-central part of the area and turns to a northwesterly trend in the central and east-central part. Movement along this fault zone has raised the rocks of the Prichard formation on the northeast with respect to those of the Revett, St. Regis, and Wallace formations on the west and southwest; the maximum displacement is about 400 m. The fault zone consists of a number of smaller faults, some of which are steep normal, the others vertical; most of them strike northerly but are connected by others that strike northwesterly. Together they form a zone that parallels the major structural trend.

Local faults and dikes parallel to this same direction occur in the central part of the area near Elk River and northwest of Clarkia. The other faults in the area strike either 5° N. to 10° E., or east to east-northeast. These faults have displaced various layers of the Wallace formation, and the vertical movement along them is far less than that along the major fault zone. The east-northeast-striking faults are perpendicular to the major structural trend, which forms an arch in the central part of the area. These faults are thus transverse faults formed perpendicular to the major direction of tension.

Small faults are very common in and next to the Revett quartzite. Some of these faults are parallel to the major fault directions, but others exhibit a rather irregular breakage. Most of them dip almost vertically (see plate 1, cross section *A–A′*).

A small overthrust fault cuts diopside-biotite gneiss of the Wallace formation northwest of Elk River (fig. 21). Thrusting at right angles to the bedding has brought thin-bedded biotite-diopside gneiss on top of thick-bedded diopside gneiss.

PLUTONIC ROCKS

Two large and several small intrusive bodies invade the rocks of the Belt series. The two larger bodies are stocks of quartz diorite and quartz monzonite; the small bodies are granite, quartz diorite, tonalite, or various gabbroic rocks. At least part of the amphibolite is igneous, but some small bodies may be metamorphic and metasomatic. Because it is impossible to distinguish the various types with adequate certainty, all amphibolite is discussed under the same heading.

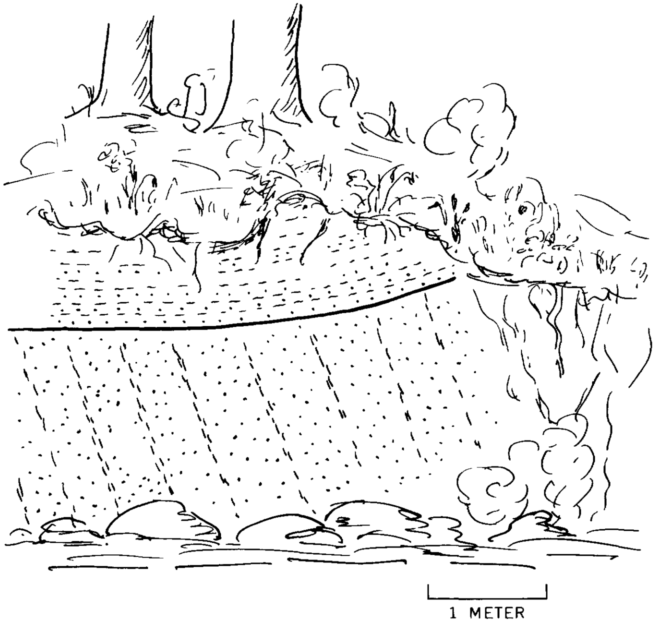


FIGURE 21.—A small overthrust to the south in diopside gneiss of the Wallace formation along the head of Shattuck Creek, 2 miles east of Jackson Mountain. Road cut faces to the west.

HORNBLENDE GABBRO, AMPHIBOLITE, AND GARNET AMPHIBOLITE

Only a few small bodies of hornblende gabbro were found in the area. One of them occurs in the quartzite of the Wallace formation in the northernmost part; another body is exposed between the diopside quartzite of the Wallace formation and the underlying schist 2 miles southeast of Green Mountain; and a third tiny body is about 1¼ miles southeast of Gold Butte. The gabbro is coarse to medium grained, massive or slightly gneissic and consists of plagioclase, hornblende, and some quartz. Biotite, sphene, and apatite are additional constituents.

Small sill-like bodies of coarse- to medium-grained well-foliated amphibolite are common in the northeastern part of the area and occur sparsely elsewhere. Most of these bodies are included in the Prichard formation or occur along the major fault zone between the Prichard and the Wallace formations. Most of the amphibolites are fine-grained intrusive gabbroic rocks which became schistose during deformations; some of the small bodies may represent dolomitic lenses in the sandy layers of the schist, and some others, especially those containing abundant garnet, were probably formed by metasomatic addition of iron and magnesium to garnet-mica schist. No amphibolite seems to have been derived from tuff. The distinction between the amphibolites of various origins is not clear because there seems to be a complete gradation from one group to another. It is assumed, however, that all dikelike bodies of amphibolite belong to the igneous group. Many of these bodies occur along fault zones, are strongly foliated, and rarely contain garnet. The major constituents are hornblende and plagioclase. Quartz is a common minor constituent; biotite is rare. Contacts between schist and amphibolite of intrusive origin are usually fairly sharp, but some amphibolites grade through gneissic gabbro to gneissic diorite (for example, at Gold Creek).

Some of the garnetiferous amphibolites contain abundant quartz, biotite, and garnet, and through a decrease in the content of hornblende and plagioclase grade into garnet-mica schist. Some coarse-grained amphibolites have large garnets (3 to 10 cm in diameter) enveloped by hornblende or anthophyllite and biotite. The exceptionally large grain size, high content of aluminum-bearing minerals and quartz, and the gradational contacts suggest that these amphibolites may have been formed by introduction of iron, magnesium, and calcium into garnet-mica schist. This type of amphibolite is common along the eastern border of the mapped area and is comparable in its mineralogy and texture to similar rocks in the Boehls Butte quadrangle (Hietanen, 1963).

QUARTZ DIORITE

A stock of quartz diorite 6 to 8 miles north of Elk River (south of Hemlock Butte) has a surface area of about 7 square miles. Two small masses of similar quartz diorite occur 0.3 to 1.7 miles west of the main body. The stock is poorly exposed, but on the basis of the distribution of residual boulders it is elongate parallel to the major structural trend. In the few outcrops there are no linear or planar structures other than joints. Most of the mineral grains are oriented at random even at the contacts.

A discordant intrusive contact between the quartz diorite and diopside gneiss is exposed at the northwest end of the stock. The quartz diorite cuts sharply across the beds of the diopside gneiss and includes fragments of the gneiss (fig. 22). Large quantities of contact minerals such as epidote occur in the gneiss next to the contact. The quartz diorite next to the contact is coarse to medium grained.

The quartz diorite is petrographically similar to the same rock type in the Headquarters quadrangle (Hietanen, 1962). The major constituents—plagioclase, quartz, hornblende, and biotite—are easily seen. The grain size ranges from 1 to 5 mm. The dark constituents tend to occur in small clusters which give a black and white spotted appearance to the rock.

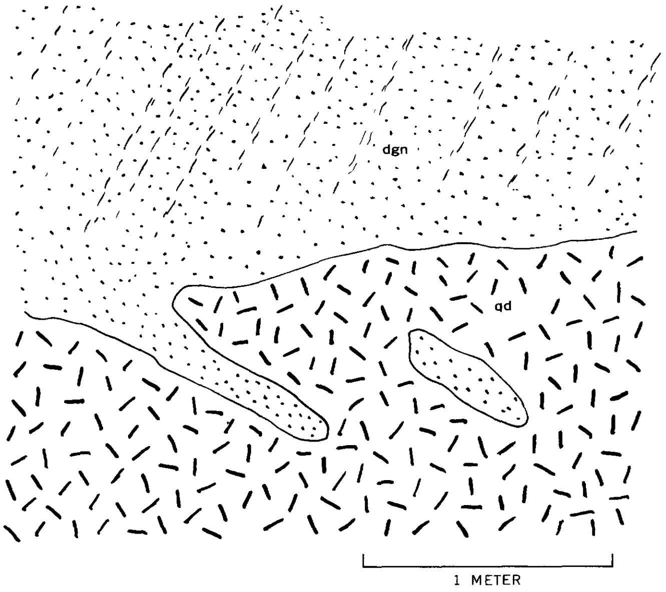


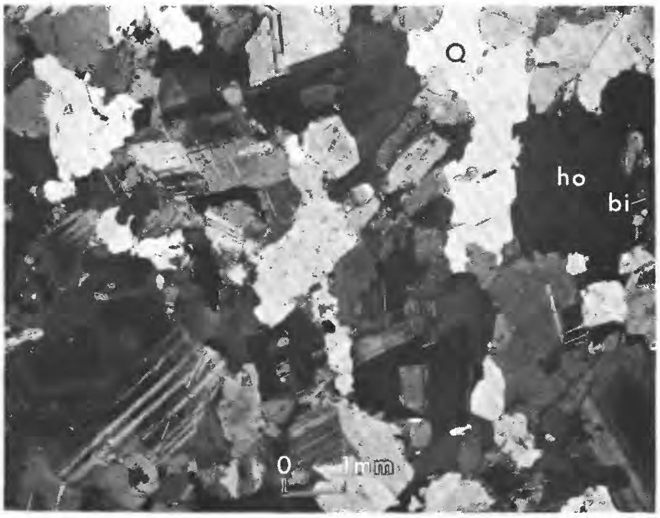
Figure 22.—A contact of quartz diorite and the diopside gneiss of the Wallace formation, along a small road on the west side of Elk Creek, 3½ miles north of Elk River. Diopside gneiss (dgn). Quartz diorite (qd).

In addition to the major constituents, the rock has small amounts of sphene, epidote, magnetite, pyrite, apatite, zircon, and chlorite. Plagioclase crystals are anhedral (fig. 23A) and show albite, Carlsbad, and complex twinnings. Biotite is partly altered to chlorite with tiny inclusions of rutile. Epidote occurs as irregular-shaped grains next to hornblende and biotite. The crystals of sphene are subhedral and range from 0.01 to 1 mm in diameter. Tiny apatite prisms have rounded corners.

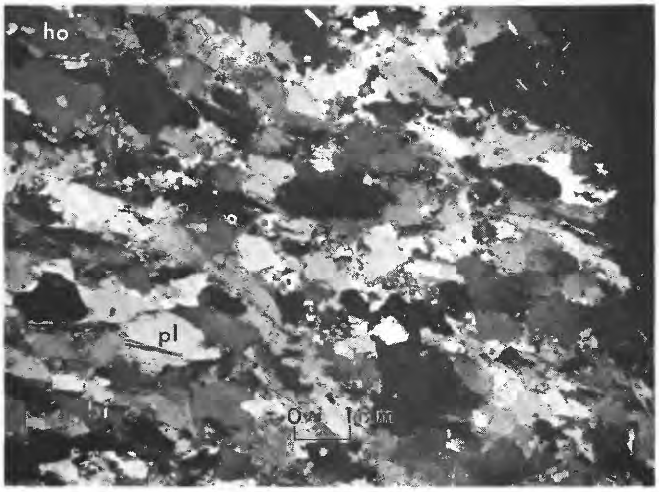
A small mass of quartz diorite 1.5 miles southeast of Grandmother Mountain and another mass at Gold Center Creek are gneissic. Some potassium feldspar and myrmekite occur with quartz and plagioclase in this rock. Small grains of epidote, apatite, and sphene occur in clusters of hornblende and biotite. The gneissic quartz diorite grades locally to an orthoclase-bearing, more silicic variety (fig. 23-*B*) in which the dark constituents are hornblende and biotite or biotite alone.

QUARTZ MONZONITE

A stock of quartz monzonite, about 11 square miles in area, is exposed just north of Elk River. It is elongate parallel to the regional trend and seems to butt



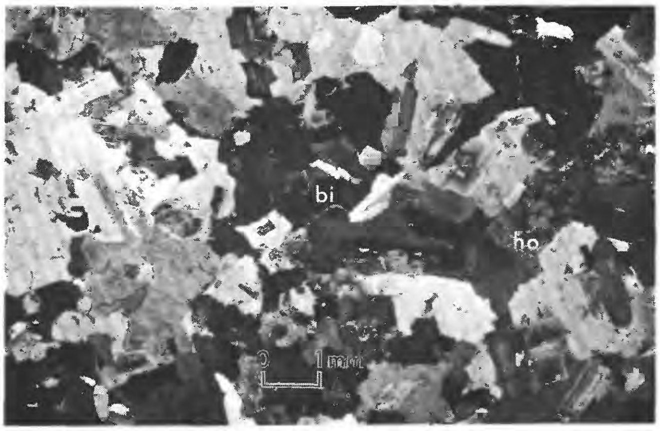
A. Quartz diorite, 1.7 miles south of Hemlock Butte (loc. 1428). Plagioclase is anhedral to subhedral and shows albite, Carlsbad, and complex twinnings. Quartz (Q), hornblende (ho), and biotite (bi) are the other major constituents. Crossed nicols.



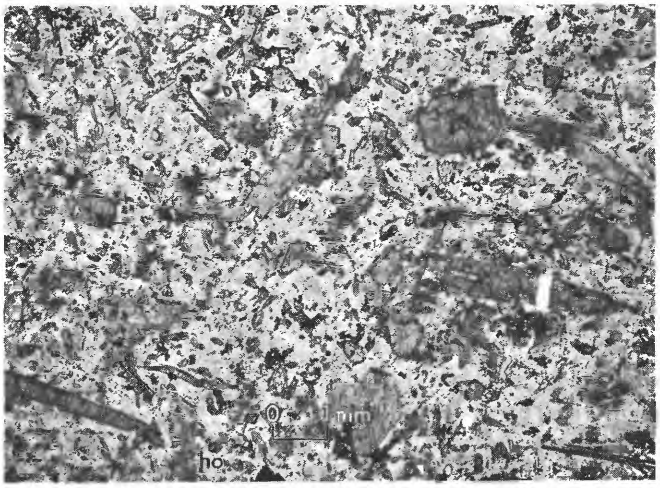
B. Granodioritic gneiss along Gold Center Creek at the mouth of Placer Creek (loc. 1678). Plagioclase (An$_{20}$) is the major light-colored constituent (pla); quartz and orthoclase occur as small grains between the large plagioclase grains. Hornblende (ho) and biotite (bi) are the dark minerals. Crossed nicols.

FIGURE 23.—PHOTOMICROGRAPHS OF QUARTZ DIORITE AND GRANODIORITE GNEISS.

against faults and porphyritic dikes on either end, but this is uncertain because neither end is exposed. Good outcrops occur along a small road leading from Elk Creek westward to the East Fork of Potlatch Creek and along another road going from Elk Creek about a mile eastward, toward Elk Butte. The rock in these outcrops is coarse grained and light gray, and has a distinct joint system (see pl. 1, 4 miles north of Elk River). In some outcrops the dark minerals are slightly oriented, but in the major part of the stock the minerals are unoriented. The dark constituents occur as small evenly distributed crystals.



A. Quartz monzonite from the west side of Elk Creek, 4½ miles north of Elk River. Euhedral plagioclase crystals of medium size are included in large anhedral orthoclase and quartz grains. The dark constituents are hornblende (ho) and biotite (bi). Crossed nicols.



B. Porphyritic hornblende gabbro (lamprophyre) dike along Merry Creek, 2 miles northeast of Clarkia (loc. 1628). Euhedral hornblende (ho) prisms occur as phenocrysts and the groundmass consists mainly of plagioclase and quartz. Plane polarized light.

FIGURE 24.—PHOTOMICROGRAPHS OF QUARTZ MONZONITE AND PORPHYRITIC HORNBLENDE GABBRO.

This rock has a texture and mineralogy typical of quartz monzonite (fig. 24A). Large anhedral orthoclase grains include many small subhedral to euhedral strongly zoned and twinned plagioclase (An$_{25-40}$) crystals. Orthoclase is perthitic and many grains include numerous tiny sericite flakes as an alteration product. The stocky plagioclase laths outside the orthoclase grains are larger than those included in the orthoclase, and their composition ranges from An$_{40}$ to An$_{28}$ from the centers of the crystals to their borders. Many small subhedral quartz grains are included in the orthoclase; one the other hand, large quartz grains include plagioclase. Hornblende and biotite fill the interstices between the stocky plagioclase laths that are not included in the orthoclase, but some subhedral grains are included in the orthoclase and quartz. Biotite is strongly pleochroic—Z=dark brown and X=light brown—and it is partly altered to light-green chlorite. Sphene, epidote, magnetite, and zircon occur as accessories.

A contact between the quartz monzonite and the biotite gneiss of the Wallace formation is exposed along an old logging road on the west side of Elk Creek about 3 miles north of Elk River. The biotite gneiss is brecciated and, instead of biotite, sericite and hematite occur as additional constituents next to the contact. A little farther south pegmatite and granite appear along the contact between the quartz monzonite and sillimanite-biotite-muscovite schist (fig. 25*A*). The granite cuts the schist discordantly and sends veins into the schist parallel to its foliation. Because of the lack of exposures it could not be ascertained whether the granite in figure 25*A* is a marginal phase of quartz monzonite or a separate intrusion.

A very light pinkish-gray granite occurs as a marginal facies between the quartz monzonite and the schist along the contact exposed about a mile north-northwest of Elk River (fig. 25*B*).

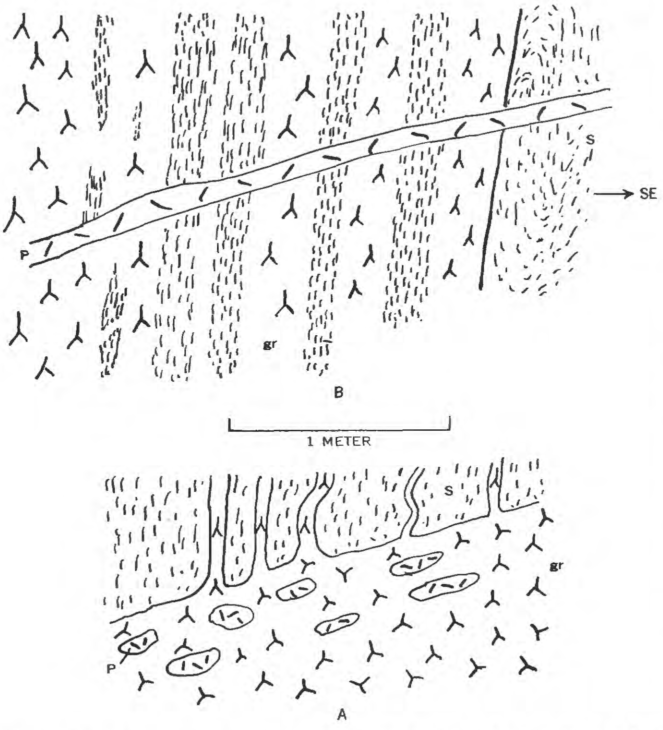


FIGURE 25.—*A*, A contact zone between schists (S) and a granite (gr) just outside of the large quartz monzonite stock. Granite includes pegmatitic portions (P). Along a small dirt road on the west side of Elk Creek about 3 miles north of Elk River. *B*, A contact between schist (S) and granite (gr) along a logging road a mile north-northwest of Elk River. This granite is a marginal facies of quartz monzonite. Pegmatite (P) cuts the schist and the granitic veins discordantly and is thus the youngest rock type.

GRANITE AND TONALITE

Several rather small masses of granite, all strongly elongate parallel to the regional trend of the metasedimentary rocks, are shown on the geologic map (pl. 1). Two types of granites, pink and light gray, have been distinguished. Both types are equigranular, coarse to medium grained, and locally gneissic. Gradation from quartz diorite to light-gray granite is common in the northeastern part of the area and locally along the southern margin of the largest quartz diorite stock.

The pink granite has a texture similar to that of the quartz monzonite; euhedral to subhedral, zoned and twinned plagioclase (An_{17-19}) crystals are included in quartz and orthoclase. The amounts of quartz and plagioclase each range from 30 to 40 percent and that of orthoclase from 15 to 30 percent. Biotite, partly altered to chlorite, is the only dark constituent, and its amount is 2 to 4 percent. Muscovite and epidote are the common minor constituents. Some of the feldspar grains contain tiny sericite inclusions, and biotite that is slightly altered to chlorite contains rutile needles.

In the light-gray granite the constituent minerals are the same as in the pink variety, but the texture is different. The plagioclase is subhedral. Myrmekite, which occurs in round grains next to the potassium feldspar, is common. Biotite and muscovite flakes show locally a crude parallel orientation.

The sill-like bodies of granite in the Prichard formation in the northeastern part of the area are more gneissic than those in the Wallace formation. There is every gradation from the gneissic quartz diorite to tonalite and to a gneissic granite in which the amount of potassium feldspar is much less than in the normal granite (about 5 percent). The tonalite is more silicic than the quartz diorite, its plagioclase (An_{20-26}) is more sodic and it lacks hornblende.

The tonalite in the two small bodies exposed along the eastern border of plate 1 resembles megascopically the gneissic granite and tonalite in the northeastern part. The amount of potassium feldspar in these bodies is 1 to 5 percent, and plagioclase (An_{20-26}) constitutes about 50 percent of the rock. The amount of quartz (about 35 percent) is larger than that in the quartz diorite, and biotite is the only dark constituent.

No direct evidence of the age relations between the pink and gray granite was found. If these rocks are correctly correlated with corresponding varieties in the Boehls Butte quadrangle, the pink granite is younger and is a differentiation product of the quartz monzonitic magma. The gradation of the quartz diorite to tonalite and gray granite indicates a common origin for these rocks. Thus, the same twofold division into potassium-rich and potassium-poor series that exists along the border zone of the Idaho batholith (Hietanen 1962) is carried over to these satellitic intrusions.

PEGMATITES

Two types of pegmatites occur in the area, older veins and dikes poor in potassium and younger ones rich in potassium feldspar and in muscovite. The older pegmatitic veins and dikes are cut discordantly by pink granite (fig. 26). On the other hand, younger pegmatites cut pink granite that occurs as separate small masses and as a marginal facies (fig. 25*B*).

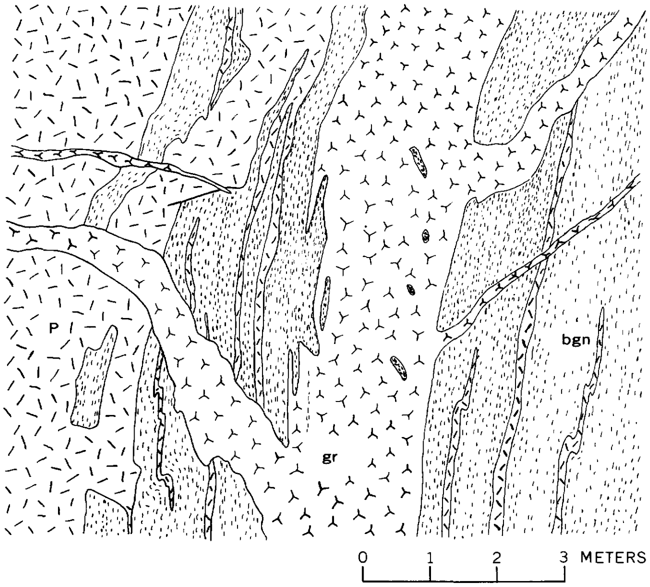


FIGURE 26.—Veins of medium-grained pinkish-gray granite (gr) and older pegmatite (P) in fine-grained biotite gneiss (bgn) of the Wallace formation along Elk Creek, 5½ miles north of Elk River.

The major constituents of the older pegmatites are quartz and plagioclase. Orthoclase, biotite, and muscovite occur in minor quantities.

The younger potassium-rich pegmatites occur as fairly thick dikes, are very coarse grained, and contain—in addition to quartz, orthoclase, and sodic plagioclase—muscovite plates as much as 5 cm in diameter. Some of the muscovite may be of commercial quality (Stoll, 1950, p. 14).

HYPABYSSAL ROCKS

Most dikes and sills occur along fault zones in the rocks of the Belt series and follow their major structural trends. The composition of the dike rocks ranges from diabase, pyroxene gabbro, and hornblende gabbro through tonalite to porphyritic granite. Only a few dikes are sheared, showing that they were emplaced before the faulting; most are post faulting and later than the intrusion of Cretaceous quartz monzonite.

DIABASE

Two long dikes of dark fine-grained diabase occur in the northwest corner of the mapped areas. These dikes resemble the nearby Columbia River basalt except that their grains are larger and they have no glass. The long lath-shaped plagioclase (An_{48}) crystals show albite, Carlsbad, and complex twinnings and a weak zoning. The pyroxene is greenish-gray augite in groups of subhedral grains. Quartz and granophyric intergrowth of quartz and feldspar fill the interstices between plagioclase laths. Irregular shaped grains of magnetite rimmed by bluish-green hornblende and some grains of green hornblende occur with the augite. Small prisms of apatite and a few scales of ilmenite are the accessories.

A very strongly brecciated diabase dike is exposed in a rock quarry about half a mile northeast of Clarkia on the north side of the St. Maries River. This locality is along the structural continuation of the dikes described above. The diabase breaks along the fracture surfaces that are covered by dark shiny serpentine minerals and white calcite. Included in this diabase are several lens-shaped pockets, layers, and stringers of medium-grained very light gray rock in which the constituent minerals are albite, quartz, calcite, epidote, and some biotite. Between these leucocratic portions and the dark diabase is a layer of medium-grained diabase, a few centimeters thick, in which the amount of light-colored constituents is larger than in the normal diabase. It is clear that the leucocratic portions represent segregations of salic material.

The diabase is very similar to that in the northwest corner, except that the augite crystals include a few grains of orthorhombic pyroxene with low interference colors and are surrounded by blue-green hornblende. Irregular fractures in the othorhombic pyroxene are filled with green alteration products, probably serpentine minerals.

Albite (An_5), the major constituent (about 70 percent) in the leucocratic portions, occurs in irregular-shaped or elongate grains, most of which show twinning. In some grains the twinning lamellae are discontinuous and the grains look checkered under crossed nicols. Calcite fills the interstices between the albite and quartz grains and also traverses the rock in thin stringers. Only a few grains of hornblende occur with biotite. Apatite, sphene, magnetite, and zircon are the accessories.

PYROXENE GABBRO

A sill-like body, 10 m thick and consisting of pyroxene-bearing gabbro, occurs along the East Fork of Potlatch Creek 1.5 miles east of the mouth of Mallory

Creek. This gabbro is dark gray and coarse grained; the white lath-shaped plagioclase crystals are unoriented and contrast strongly with the mesh of dark constituents.

Subhedral augite, brown hornblende, and alteration products—chlorite and epidote—fill most of the interstices between the plagioclase laths. A few of the interstices are filled by quartz. Ilmenite, magnetite, apatite, and allanite occur as minor constituents.

The plagioclase crystals range from 3 to 6 mm in length, are strongly zoned, and show complex twinnings. Many are filled by alteration products—epidote, sericite, and chlorite—to the extent that the twinning is obscured. Augite occurs next to brown hornblende or is surrounded by hornblende and chlorite. Some of the brown hornblende and augite are altered to bluish-green hornblende, but a greater part, about half, is altered to aggregates of chlorite and epidote. Small elongate grains of epidote and leucoxene are included in the chlorite and contain the calcium and titanium that were originally in augite and hornblende. Allanite occurs in rounded light-brown pleochroic zoned grains. Leucoxene containing magnetite lamellae, and euhedral to subhedral crystals of magnetite occur with the dark constituents. The leucoxene is an alteration product of ilmenite, which has lamellar intergrowths with magnetite. Small prisms of apatite are sparsely scattered throughout the rock.

A sill of similar pyroxene gabbro is exposed along the East Fork of Potlatch Creek, about 1.7 miles west of the area mapped (pl. 1). In this gabbro, granophyric intergrowth of quartz and feldspars fill some of the interstices.

PORPHYRITIC HORNBLENDE GABBRO

Medium-gray medium-grained dikes of porphyritic hornblende gabbro are common in the north-central part of the area. These dikes traverse the metamorphic rocks parallel to the major fault directions and are more common across the major structural trend than parallel to it.

The hornblende prisms are euhedral (fig. 24*B* and strongly pleochroic, with Z=brown and green spotted, Y=green, X=straw color. The centers of many prisms are chlorite. Some rutile and epidote occur with the chlorite.

The light-colored constituents are plagioclase and some quartz. In the fine-grained dikes plagioclase occurs as slender laths, whereas in the medium-grained rock plagioclase crystals are stocky (for example, loc. 1841). Epidote is abundant as an alteration product of plagioclase. Magnetite, sphene, and apatite are the accessories.

A sill-like body near the northern border of plate 1 (loc. 1912) consists of coarse-grained hornblende gabbro, in which euhedral to subhedral prisms (½–1½ cm long) of hornblende are oriented at random and plagioclase fills the interstices. The hornblende is a brown basaltic variety, and some chlorite and epidote occur next to and within the hornblende. Outlines of a few chlorite-epidote clusters suggest that they may be alteration products of augite. Lath-shaped albite crystals are twinned according to albite, Carlsbad, and complex laws. Many albite crystals have abundant epidote in their centers and additional epidote fills some of the interstices. Other interstices are filled by granophyric intergrowth of quartz and albite, or by quartz and orthoclase. Apatite forms slender prisms that traverse the grains. Skeletal grains of ilmenite accompanied by leucoxene are common. This gabbro resembles the pyroxene gabbro described earlier except that hornblende tends to be better formed and augite is lacking.

PLAGIOCLASE PORPHYRY

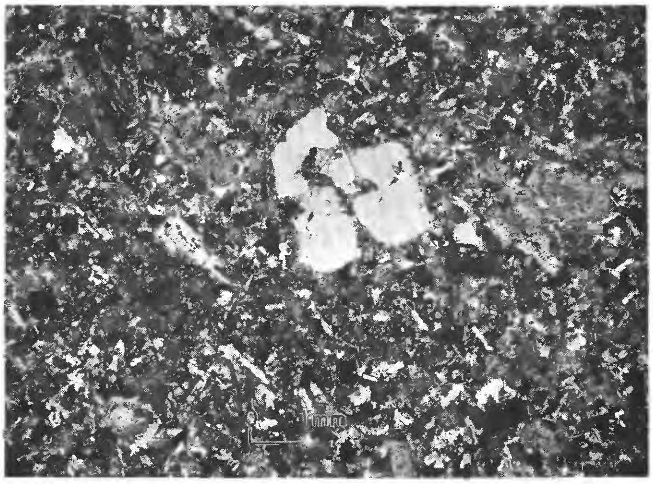
Plagioclase porphyry dikes with euhedral plagioclase and quartz phenocrysts in a dark-gray fine-grained groundmass are widespread (fig. 27*A*). The composition of these dikes is tonalitic. Plagioclase and quartz phenocrysts constitute about 10 percent of the rock; the groundmass consists mainly of small brown euhedral hornblende prisms and plagioclase laths. Orthoclase, chlorite, magnetite, sphene, and apatite are additional constituents. Small grains of epidote occur as an alteration product in plagioclase. The groundmass is mineralogically and texturally close to that of the porphyritic hornblende gabbro dikes.

Aphanitic dikes of tonalitic composition occur, for example in locality 567. These dikes are dark gray and contain only a few small phenocrysts of quartz and plagioclase. The groundmass consists of tiny lath-shaped plagioclase crystals and interstitial quartz, biotite, and hornblende. Magnetite, calcite, and a few tiny grains of epidote occur with the dark constituents.

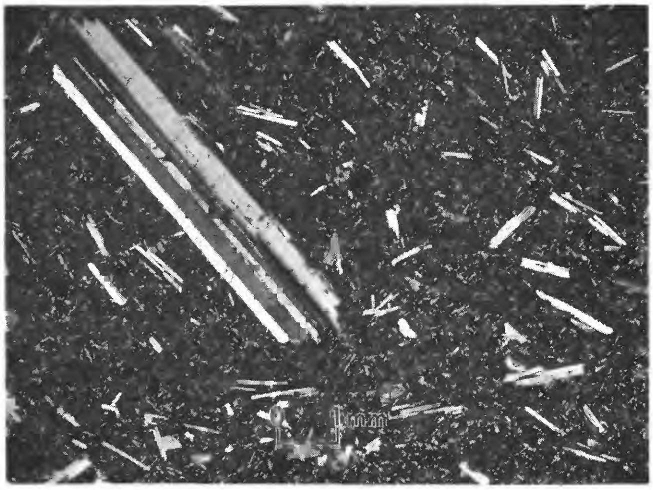
GRANITE PORPHYRY

Classed as granite porphyry are all light-colored silicic dikes in which the major constituents are quartz, plagioclase, and orthoclase. The amount of orthoclase in many is less than in common granite. The composition of these dikes is close to that of quartz monzonite except that the amount of orthoclase is less.

The common granite porphyry is a light-brown or light- to medium-gray rock with phenocrysts of quartz and plagioclase. The phenocrysts constitute from 10



A. Quartz and plagioclase phenocrysts in a dioritic dike along Merry Creek, 4 miles northeast of Clarkia (loc. 1630). The groundmass consists mainly of tiny euhedral hornblende prisms and plagioclase laths. Crossed nicols.



B. Columbia River basalt along Merry Creek, 1 mile northeast of Clarkia (loc. 1623). A few larger phenocrysts of labradorite occur in the porphyritic basalt; small phenocrysts of labradorite abound. Crossed nicols.

FIGURE 27.—PHOTOMICROGRAPHS OF PLAGIOCLASE PORPHYRY AND COLUMBIA RIVER BASALT.

to 50 percent of the rock; the individual grains range from 3 to 10 mm in diameter. In some outcrops phenocrysts of quartz are euhedral, with pyramidal faces; but in the other outcrops, both the quartz and the plagioclase phenocrysts are subhedral. In most dikes the amount of plagioclase phenocrysts exceeds that of quartz phenocrysts. Phenocrysts of hornblende and biotite, or only biotite, occur also, but always in subordinate amounts.

The fine grained groundmass consists of quartz, plagioclase, orthoclase, biotite, hornblende and some sphene, apatite, and zircon. In some dikes, chlorite, sericite, calcite, and epidote occur as alteration products. In the very light colored dikes, biotite is the only dark constituent. Alteration of biotite into chlorite with inclusions of epidote, sphene, and rutile is common.

A gray porphyritic dike that is exposed along Stony Creek about 2 miles west of the mouth of Glover Creek differs in many respects from the common type described above. This dike contains numerous phenocrysts of perthitic orthoclase but none of quartz. Groups of three or four anhedral pinkish-white orthoclase crystals and elongated groups and shreds of small dark hornblende and biotite grains contrast with the fine-grained steel-gray groundmass. All larger phenocrysts are orthoclase and only a few small phenocrysts are albitic plagioclase. The groundmass is sheared and consists of quartz, plagioclase, orthoclase, hornblende, and biotite.

Granite porphyry dikes, which are extremely common along the fault zones, were probably emplaced over a considerable time interval; some of the dikes are brecciated and the fractures are filled with pegmatite (fig. 28).

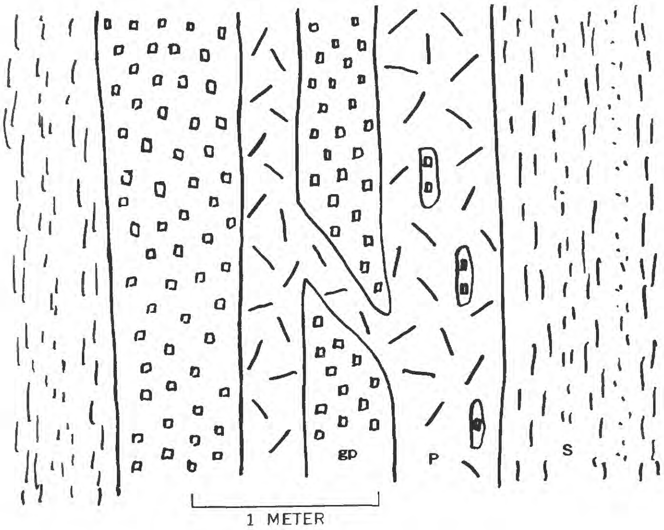


FIGURE 28.—Granite porphyry (gp) dike cut by pegmatite (P) in a schist (S) along a small dirt road leading from Elk Creek to the East Fork of Potlatch Creek, about 5 miles north of Elk River (loc. 1190, 1 mile west of Elk Creek).

COLUMBIA RIVER BASALT

The basalt in the southern part of the mapped area belongs to the eastern margin of the Columbia River plateau. The basalt in the northwestern part is the southernmost remnant of the basalt cover that extends along the St. Maries River valley from Coeur d'Alene Lake (28 miles to the northwest of the northwest corner of pl. 1) to Clarkia. Only one flow with columnar jointing is exposed in the river valley near Clarkia, whereas several flows, each ranging about 8 to 20 meters in thickness, occur along the canyons of Elk Creek and its tributaries in the southwest. The total thickness of the basalt in these canyons is as much as 300 meters.

The basalt is an aphanitic or fine-grained dark rock with abundant vesicles in the upper and lower parts of the flows. Columnar jointing is prominent in many flows, but some others break into irregular-shaped polygonal blocks.

Numerous slender laths of plagioclase oriented at random are visible in most outcrops. Thin sections show that the major constituents are labradorite and augite with interstitial brown glass and abundant magnetite and ilmenite. In the very fine grained basalt plagioclase phenocrysts are fewer (fig. 27*B*) and the groundmass in fine grained and porphyritic, consisting of augite and tiny plagioclase laths. In the coarser grained variety, plagioclase phenocrysts constitute 35 to 45 percent of the rock, 40 to 45 percent being augite and the rest interstitial brown glass, green celadonite, magnetite, ilmenite, and olivine.

GRANITIZATION AND MIGMATIZATION

Only a minor part of the schist and biotite gneiss next to the granitic bodies and to the largest quartz diorite stock has been transformed to gneissic rocks in which the amount of plagioclase and orthoclase is increased enough to give the contact zone a granitic or migmatitic appearance and composition. Most of the granite next to the quartz diorite stock north of Elk River (pl. 1) is gneissic, has gradational contacts with the schist, and was probably formed by granitization. Elsewhere the grantitized parts of the schist and biotite gneiss range only from a few meters to about 10 m in thickness and extend from ¼ to 2 miles in length.

Pegmatitic and granitic veins and dikes abound in metasedimentary rocks near some plutonic rocks, giving rise to rocks that in their general appearance resemble contact migmatites (for example, figs. 25 and 26). These veins are especially abundant in the biotite gneiss between the two largest intrusive bodies. There are at least two generations of pegmatitic veins and dikes: one earlier than the granite and one later. The granitic veins and dikes contain inclusions of schist and dark fine-grained dioritic rock. Minerals in the granite are oriented at random and the texture is hypidiomorphic. Thus these veins and dikes have textures and structures typical of igneous rocks and probably represent offshoots of the same magna from which pink granite belonging to the quartz monzonite series crystallized.

Granitized sillimanite-mica schist in the vicinity of the quartz diorite stock crops out along a dirt road leading from Hemlock Butte to the East Fork of Potlatch Creek, about 1 mile west of Hemlock Butte (loc. 1431). The granite is fine grained and pink, similar to that in dikes and veins (fig. 26). Small inclusions of kyanite-biotite schist in every stage of digestion are distributed throughout the small occurrence of this granite (fig. 29).

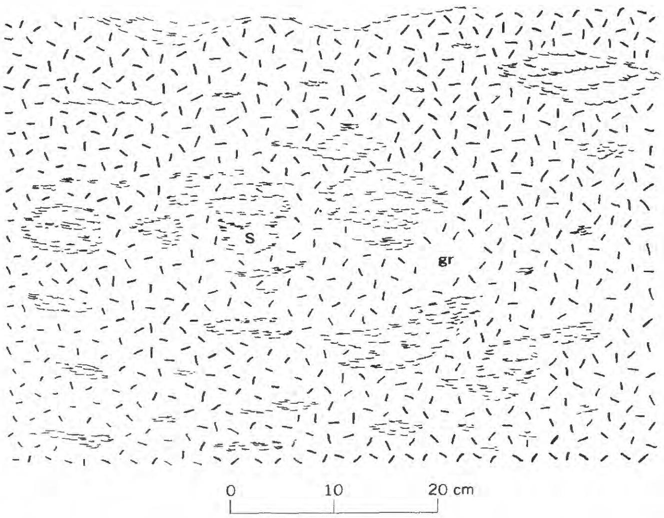


FIGURE 29.—Remnants of kyanite-sillimanite-biotite-muscovite schist (S) in fine-grained gray granite (gr), along a road leading from Hemlock Butte to the East Fork of Potlatch Creek, about 1 mile west from Hemlock Butte.

The granite along the southern contact of the quartz diorite stock is coarse grained, has a distinctly foliated texture, and contains biotite-rich schlieren and ghostlike remnants of biotite-muscovite schist, giving an impression that it was formed by granitization of the schist.

Much of the coarse-grained granodioritic gneiss in the northeastern part of the area is a result of secondary crystallization of plagioclase (An_{20}) and hornblende in the schist, in which the amount of quartz is therefore less than in the normal schist. These gneissic rocks abound in the southern part of T. 43 N., R. 3 E., south of Gold Center Creek, and in the northern part of T. 42 N., R. 3 E., north of Glover Creek, where small bodies of gray granite and quartz diorite occur. The granodioritic gneiss is a light-gray medium- to coarse-grained rock, made mainly of easily recognized plagioclase (An_{20}), quartz, hornblende, and biotite. In addition to these major constituents, there are interstitial orthoclase, anhedral grains of epidote, euhedral to subhedral crystals of sphene, and small prisms of apatite. A few small crystals of zircon are included in biotite and in hornblende. Hornblende occurs in small anhedral elongated grains that have their longest dimensions (*c* or *b* axis) parallel to the plane of foliation and show pleochroism Z=green, Y=light green, X=yellowish green. Biotite flakes are of the same size as the hornblende crystals (1–3 mm long) and oriented with their cleavage subparallel to the plane of foliation (fig. 23*B*). Tiny grains of hornblende and biotite occur next to the larger grains of these minerals and along the plane of foliation between the larger grains. The texture and mode of occurrence of these gneisses resemble those of the metasomatic tonalites described from the Orofino area (Hietanen, 1962); and most likely also these bodies are a result of introduction of the elements (sodium, calcium, iron, magnesium) that together with the silicon and aluminum of the metamorphic rocks formed feldspars and hornblende.

In some of the gneissic rocks, biotite is the only dark-colored constituent, and the composition is close to that of the granite. However, the amount of potassium feldspar is less than in the normal granite. These rocks grade to garnet-biotite-muscovite schist and were formed by development of abundant oligoclase and some orthoclase in the schist near the granitic and dioritic bodies.

DISTRIBUTION OF METAMORPHIC FACIES IN SPACE AND TIME

The sequence of crystallization of the three aluminum silicates—andalusite, kyanite, and sillimanite—in the schists around Clarkia yield some information about the distribution of the pressure and temperature fields during the recrystallization. The square cross sections of crystal aggregates of muscovite that are seen on the bedding planes of the schist in some localities west of Clarkia are most likely pseudomorphs after andalusite. Their occurrence would suggest that andalusite crystallized early, but became unstable. Also the nodules of kyanite and muscovite in the schists east and southeast of Clarkia and those of kyanite and sillimanite near Hemlock Butte could be pseudomorphs after andalusite. Because andalusite is stable at fairly high temperatures, but only at low pressures, it is likely that an increase in pressure was the major reason for its disappearance and change to kyanite.

West of Clarkia, sillimanite is found as a pseudomorph after kyanite, as for instance, on Bechtel Mountain. This change suggests a local rise in the temperature after the crystallization of kyanite. The local rise in temperature may be due to an intrusive body not exposed in the area of plate 1. Toward the south the sillimanite becomes prevalent around Hemlock Butte. During the metamorphism, the temperature and pressure near Clarkia moved from the stability field of andalusite, first to the stability field of kyanite then to the border line that separates the stability fields of kyanite and sillimanite (Hietanen, 1956, fig. 5 and 1961*b*, fig. 3). These inversions between the three aluminum silicates indicate that the gradual increase in the degree of metamorphism toward the south is not simple contact metamorphism along the border zone of the huge Idaho batholith, but is complicated by factors other than regular and gradual change in temperature. The order of recrystallization near Clarkia and the pseudomorphs after andalusite and kyanite suggest that there was first an increase in pressure and later an increase in temperature. The orientation of the micaceous minerals and the hornblende parallel to the planes that served as glide planes during the deformation suggests that the crystallization of these minerals took place in a field of stress, most likely during the deformation and under moderate pressure. The occurrence of kyanite crystals along the most prominent *s* plane (the bedding) suggests that the kyanite crystallized during this same phase under high pressure and approximately at the same temperature. Andalusite crystallized earlier at a somewhat lower pressure but probably at about the same (moderate) temperature. On the other hand, the crystallization of sillimanite at the expense of kyanite indicates a rise in temperature. Thus three major steps can be separated in the sequence of events near Clarkia:

1. Folding and recrystallization at moderate temperature and pressure. Crystallization of andalusite, muscovite, biotite, and almandite in the schists, and of hornblende, epidote, and plagioclase (An_{5-20}) in the calcareous layers, suggesting conditions of the epidote-amphibolite facies.
2. Increase in pressure and probably slight increase also in temperature to correspond to the conditions of the higher part of the epidote-amphibolite facies. Crystallization of kyanite at the expense of andalusite, formation of staurolite in the schists, and crystallization of diopside and plagioclase (An_{10-25}) instead of (or with) hornblende and epidote in the calcareous layers.
3. Increase in temperature to correspond to the lower part of the amphibolite facies. Crystallization of sillimanite at the expense of kyanite. Increase in size of minerals such as garnet, biotite, muscovite, diopside, plagioclase, and quartz.

These three phases of recrystallization can be correlated with the major geologic events in the area as follows:

In phase 1, the rocks of the Belt series were folded and metamorphosed during the Nevadan orogeny. Small intrusive bodies of gabbro and quartz diorite were emplaced during the folding. These bodies are more common in a zone closer to the Idaho batholith (Hietanen, 1962) than in the present area, which indicates that the place now occupied by the Idaho batholith was a center of heat and intrusive activity during Nevadan time.

The region around Clarkia lies about 30 miles northwest of the border of the Idaho batholith, and it would take some time (probably millions of years) for the heat to travel this distance in rocks 15 to 20 km below the surface, whereas the pressure would be transmitted practically instantaneously. It would, therefore, be only logical to assume that some high temperature (and antistress) minerals would crystallize after the main dynamic phase in the course of regional metamorphism. Near Clarkia andalusite was one of the early high-temperature (but antistress) minerals.

Phases 2 and 3 are most likely connected with the intrusion of the border zone of the Idaho batholith. Phase 3 represents the heat wave that lagged behind

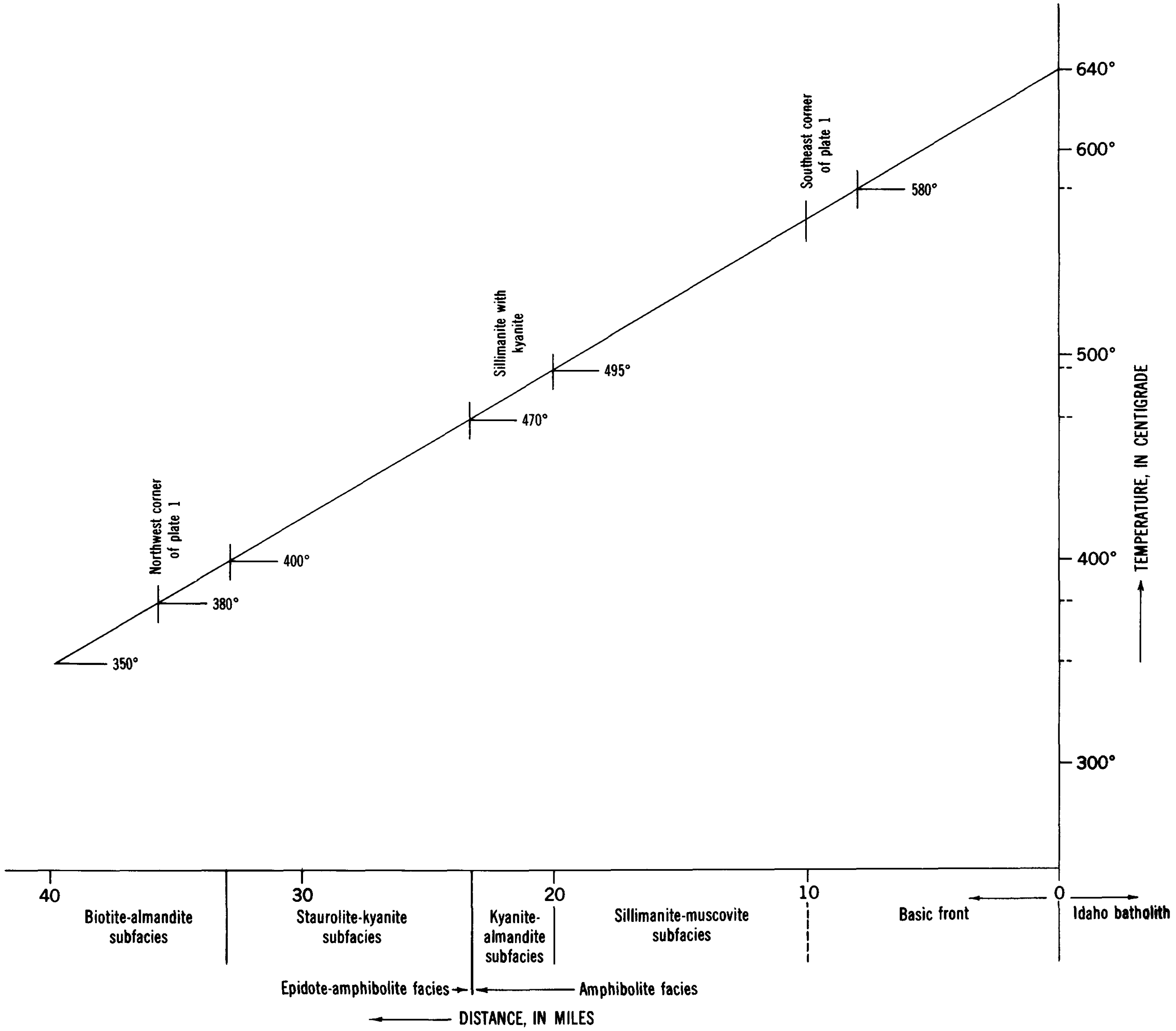


FIGURE 30.—Hypothetical curve for the temperature of recrystallization in various metamorphic facies. Abscissa is the horizontal distance in miles from the border of the Idaho batholith. The end temperatures are those assumed for the green schist facies and for the beginning of the melting of granite. See further explanation in the text.

the accelerated pressure. Again the interval between the times when the maximum pressure and the maximum temperature were reached near Clarkia may have been of the order of millions of years, long enough for the crystallization of kyanite before the temperature field of sillimanite was reached. In the course of recrystallization, the temperature and pressure approached the triple point of the three aluminum silicates but never reached it as they did in the neighboring Boehls Butte quadrangle (Hietanen, 1956). As long as the temperature and pressure of this triple point in the map area is not known, other mineral associations must be used in estimating the temperature and pressure of recrystallization.

In the high-temperature zones of the regionally metamorphosed pelitic schist, muscovite (+quartz) normally yields microcline and sillimanite; this reaction, together with the disappearance of kyanite, theoretically marks the lower limit of the sillimanite subfacies. In the Elk River-Clarkia area, the occurrence of muscovite persists into the sillimanite subfacies, and farther southward up to the border zone of the batholith. Its quantity, however, is greatly reduced in the zone next to the batholith.

The association of muscovite with sillimanite shows that the metamorphic temperatures in the southernmost part were lower than 650°C (Yoder and Eugster, 1955). Furthermore, the melting point of granite, which according to Tuttle and Bowen (1958, p. 122) is 640°C at depth of about 20 km, was not reached closer than 8 miles southeast of the southeastern corner of the Elk River-Clarkia area, perhaps still farther because the border zone of the batholith was emplaced by intrusion and the zone of remelting would have been either under the level exposed or farther toward the center of the batholith. The grade of metamorphism decreases northward, and the southern border of the green schist facies is about 10 miles north of Clarkia. If the temperature of this limit is assumed to be about 350°C, temperature at the contact of the batholith 640°C, and the increase of temperature toward the batholith more or less linear, the graph in figure 30 gives a rough estimate of the temperatures in each facies. The temperature would increase from 380° to 580°C from the northern border to the southern border. Staurolite would be stable with kyanite between 400° and 470°C. Sillimanite would start to crystallize at about 470°C but become the only stable form of Al_2SiO_5 at about 500°C. The actual temperature interval was most likely less and the curve should be flatter in this central part, or the zone where sillimanite and kyanite occur together should be narrower.

DISTRIBUTION OF POTASSIUM FELDSPAR

The abundance of the potassium feldspar in the gneissic layers in the northern and central part of the area and its scarcity in the corresponding layers in the southern part has three possible explanations: (1) distribution of potassium is due to original deposition, (2) the abundance of microcline in the northern part is due to metasomatic introduction of potassium, (3) the scarcity of microcline in the southern part is a result of removal of potassium during metasomatic metamorphism.

No evidence decisively favoring one of these possibilities has been found in the field or under the microscope. Moreover, an acceptable explanation has to agree not only with the facts in the present area, but also with facts and conclusions reached in neighboring areas. In the following discussion, the three possibilities are therefore weighed in the light of local and regional evidence.

1. Lateral variation in the composition of individual units is fairly common in the Belt series. For example, in the Boehls Butte quadrangle abundant calcareous material occurs in the Prichard formation in the stratum which, just north of this quadrangle, consists of fairly pure quartzite (Hietanen, 1963). This facies change takes place by gradual increase of the number and the thickness of the calcareous layers, and by decrease of quartzitic material. In the Elk River-Clarkia area, the change from gneisses rich in potassium feldspar to those in which this mineral is scarce or lacking takes place in a similar way; the number and thickness of the potassium-rich layers decrease from the vicinity of Elk Butte toward the south over a distance of about 10 miles. Most of the potassium-rich layers contain, in addition to microcline, abundant biotite or biotite and muscovite. In many outcrops of diopside or biotite gneiss the potassium-rich layers (1 to 3 mm thick) are well defined and persist over tens of meters. Microcline in these layers occurs as small irregular-shaped or rounded grains in association with quartz and micas. This mode of occurrence suggests that microcline crystallized from the clastic grains of the original sedimentary material. Abundance of micas in the microcline-rich layers indicates that these layers were originally clays rich in potassium. It is well known that clays are capable of absorbing potassium and that potassium-bearing authigenic clay minerals, such as illite, are formed early during weathering and sedimentation (Grim, 1953).

2. Extensive metasomatism is known in the area adjoining the Elk River-Clarkia area to the south and southeast (Hietanen, 1962). It involves introduction

of sodium, calcium, aluminum, iron, and magnesium and removal of potassium and silicon. The hypothesis that the potassium removed from the zone next to the Idaho batholith might have been deposited in the outer contact zone of the batholith is very entertaining, but the potassium-rich gneisses have many features that speak against such an interpretation. First, the textures and structures of the microcline-bearing gneisses indicate that all minerals in them crystallized more or less contemporaneously from the material present in each layer. There are none of the textures typical of the metasomatic minerals in the Boehls Butte quadrangle, in the Headquarters quadrangle, or near Dent. In these areas, the metasomatic minerals occur as large, round, or more rarely euhedral grains, include other minerals, and have a rather irregular distibution that cuts in many places acoss the bedding. In contrast, the potassium feldspar in the gneisses near Elk River and Clarkia is confined to certain thin beds and laminae, where it occurs in small irregularly-shaped grains in association with quartz, biotite, diopside, and plagioclase. The bedding is extremely well preserved and the thin (1–3 mm) potassium-rich beds persist over several tens or even hundreds of meters just as do the paper-thin micaceous laminae. Therefore it does not seem likely that potassium was introduced metasomatically; rather the occurrence of potassium feldspar is comparable to that of the other potassium-bearing minerals, biotite and muscovite, which crystallized within thin well-defined potassium-rich clayey layers interbedded with calcareous sandstone.

3. The third explanation is postulated upon the following concept: during sedimentation potassium was equally distributed over the whole area but that later, during metasomatic metamorphism, it was removed from the southern part of the area. This southern part is next to the inner contact zone of the Idaho batholith from which, as described in an earlier paper (Hietanen, 1962) potassium and silicon were removed and sodium, calcium, aluminum, iron, and magnesium were introduced. The textures and structures suggest that more intensive recrystallization took place in this southern part than in the northern part. In the southern part, grain size is larger and the beds are not as well defined. However, typical metasomatic textures and structures as defined under explanation 2 are rare. Large albitic plagioclase grains that contain numerous small round quartz inclusions and that could be secondary were found only in one specimen. This specimen was collected along the North Fork of Clearwater River in the extreme southeast corner of the mapped area, thus close to the metasomatized inner contact zone of the Idaho batholith. Elsewhere the light-colored constituents occur as irregularly-shaped grains of about equal size and the micas are well alined parallel to the bedding or parallel partly to the bedding and partly to the transecting cleavage. In a few localities, secondary hornblende occurs in diopside gneisses, indicating addition mainly of hydrous molecule. The number and thickness of quartz-rich layers in the diopside-plagioclase gneiss is larger than that in the corresponding gneisses next to the batholith. This change, too, can be either original or due to the removal of silicon from the zone next to the batholith. In many localities, bedding has been obscured by recrystallization; but there are no distinct signs of large-scale removal of potassium, and it must be concluded that if such removal occurred, it must have been during an early phase of recrystallization.

All diopside gneiss exposed in the southern part of the Elk River–Clarkia area belongs to the lower quartzite-gneiss unit of the Wallace formation. In the northern part, the white granular quartzite, which rarely contains potassium feldspar, is the major rock type of this unit. The biotite and hornblende gneisses bearing potassium feldspar are a minor portion of the rock unit. Southward, the amount of tremolite-actinolite and diopside increases, but fairly pure white granular beds crop out again along the North Fork of Clearwater River. It seems, therefore, that the diopside-plagioclase gneisses between Elk Butte and Little Green Mountain include beds that are dolomitic equivalents of the white granular quartzite. In this part of the area, layers rich in potassium feldspar probably are rare because the original sediment was poor in potassium. It is more difficult to understand why the biotite gneiss layers of the southern part are poor in potassium feldspar whereas the corresponding layers in the northern part contain a considerable amount of microcline.

In summary, the following facts support a sedimentary explanation for the distribution of potassium: (a) Potassium feldspar is more abundant in the upper than in the lower quartzite-gneiss unit. (b) Potassium feldspar occurs only in the biotite gneisses and biotite-bearing diopside gneisses. (c) Potassium minerals are confined to thin layers that persist over long distances, in a distribution comparable with that of the biotite laminae, which crystallized from sedimentary material. (d) Potassium feldspar occurs as irregular-shaped or rounded grains that are of the same size as quartz and plagioclase and lack replacement textures.

Because the textures and structures of the gneisses in the northern part of the Elk River–Clarkia area fail to indicate any metasomatic addition of potassium, it is concluded that microcline crystallized there at the expense of sedimentary material. It remains to decide whether or not potassium-rich layers were equally deposited in the southern part; if they were, the lack of potassium feldspar is presumably due to metasomatic removal of potassium.

The fact that potassium feldspar is rare in all rocks near the Idaho batholith, whereas layers rich in this mineral occur in increasing numbers in the corresponding layers outside the contact zone, supports the possibility of metasomatic removal and deserves careful consideration.

The changes in grade of metamorphism and in the distribution of potassium are regular over the whole area and are not related to the small intrusive bodies exposed in the central part. Because the grade of metamorphism increases fairly regularly toward the southeast, it was concluded that the metamorphism is connected with the intrusive activity of the main mass of the Idaho batholith, the nearest outcrops of which are exposed about 10 miles southeast of the North Fork of the Clearwater River. Irregularly shaped parts of all rocks within this 10-mile-wide inner contact zone show metasomatic alteration in which silicon and potassium were removed and sodium, calcium, aluminum, iron, and magnesium introduced (Hietanen, 1962).

The unaltered parts of the strata in this inner contact zone are mineralogically very similar to the corresponding layers in the southern part of the Elk River-Clarkia area. Also, potassium feldspar is scarce or lacking. The removal of potassium from the inner contact zone is connected with the late replacement of biotite by hornblende. Such a replacement is rare in the southern part of the Elk River–Clarkia area, which represents the outer contact zone of the Idaho batholith. It is possible, however, that a part of the potassium was removed from this outer contact zone during an early phase of metamorphism.

ELK RIVER-CLARKIA AREA AS A PART OF THE OUTER CONTACT ZONE OF THE IDAHO BATHOLITH

The rocks exposed in the northern part of the Elk River-Clarkia area represent isochemically metamorphosed equivalents of the Belt series in the Coeur d'Alene district. Changes in composition and local changes in texture occur from north to south. The major compositional difference between the rocks of the Belt series in the northern and southern parts is the lack of potassium feldspar in the southern part. The mineral assemblages of the metasedimentary rocks indicate a steady increase in metamorphic temperatures and pressures toward the south. In the northernmost part, recrystallization took place under conditions of the biotite-almandite subfacies of the epidote-amphibolite facies. In the central part, the mineral assemblages are those common in the staurolite-kyanite subfacies of the epidote-amphibolite facies; in the southern part, those of the sillimanite-muscovite subfacies of the amphibolite facies.

The structural continuation of the Belt series toward the south and the east from the Elk River–Clarkia area have been studied earlier (Hietanen, 1962, and 1963). Comparison of the mineral content and the texture of the schist and gneiss in the Elk River–Clarkia area with those of the corresponding layers in the Wallace and the Prichard formations south and east of this area indicates that the grade of metamorphism continues to increase toward the south and the east of the area shown in plate 1. A greater change, however, is caused by extensive metasomatic introduction of sodium, calcium, aluminum, iron, and magnesium in a zone next to the Idaho batholith (Hietanen, 1962) and by introduction of sodium and possibly aluminum into the schist in the Boehls Butte quadrangle and vicinity (Hietanen, 1963). Traces of similar metasomatic action appear in only a few localities in the southern and eastern parts of the Elk River–Clarkia area, where the rocks of the Belt series have suffered the same grade of metamorphism as those next to the batholith.

Granitization connected with the intrusive stocks has affected small local parts of the strata (about 1 percent), whereas most of the rocks of sedimentary origin were recrystallized without any determinable addition of material from outside sources (isochemical metamorphism). Most of the igneous stocks were emplaced by mechanical intrusion and the rocks that resulted from reaction between the igneous and the metasedimentary material form only a small proportion of the total amount of plutonic rocks.

The regional metamorphism is clearly connected with the history of the formation of the Idaho batholith as pointed out in the preceding chapter. The major rock types in the batholith are quartz diorite, granite, and quartz monzonite. The marginal zone consists of quartz diorite devoid of potassium feldspar. The center consists mainly of quartz monzonite rocks rich in orthoclase. The composition of the quartz diorite is similar to the mean composition of the surrounding metasedimentary rocks; but the central part is considerably richer in alkalies, especially in potassium.

If it is assumed that the granitoid rocks of the batholith are a result of granitization and partial remelting of the metasedimentary rocks which in composition were similar to those around the batholith, considerable amounts of iron (about 3 percent), magnesium (2 percent), and calcium (3 percent) should have been removed and mainly silicon (5 percent), sodium (2 percent), and potassium (3 percent) introduced into the place now occupied by the granitoid rocks (Hietanen, 1962, table 31). If the granitized metasedimentary rocks contained more potassium—resembling the strata that are exposed in the northern part of the Elk River–Clarkia area—the amount of silicon and potassium needed to transform the strata into granite would be somewhat less, perhaps only 3 and 2 percent respectively, but still qualitatively the same.

A part of the potassium and silicon needed for the formation of the quartz monzonitic batholith was most likely derived from the surrounding metasedimentary rocks, which were consequently impoverished in these elements and relatively enriched in calcium, iron, and magnesium. Later, during the intrusion of quartz diorite more potassium and silicon may have joined the magmas; cafemic constituents migrated in an opposite direction and became fixed in the marginal zone of the botholith and in the metasedimentary rocks along the contact as decribed earlier (Hietanen, 1962). As a result of this migration, a zone of enrichment of basic components, 10 to 20 miles wide, was formed northwest of the batholith. The southern part of the Elk River–Clarkia area adjoins this basified zone in the south. There was no large-scale addition of calcium, iron, or magnesium in this outer contact zone, but the lack of potassium feldspar is conspicuous. Considering the higher mobility of potassium in comparison with calcium, iron, and magnesium, it seems very likely that the potassium could have been mobilized and removed within a wider zone than the other elements. The direction of its migration would have been the same as in the inner contact zone—that is, toward the growing batholith. The general increase in the amount of plagioclase toward the batholith suggests that sodium replaced potassium along the outer contact zone.

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Idaho Batholith Near Pierce and Bungalow Clearwater County, Idaho

By ANNA HIETANEN

METAMORPHIC AND IGNEOUS ROCKS ALONG THE
NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

GEOLOGICAL SURVEY PROFESSIONAL PAPER 344-D

Petrologic study of igneous rocks in the northwestern corner of the Idaho batholith

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METAMORPHIC AND IGNEOUS ROCKS ALONG THE NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

IDAHO BATHOLITH NEAR PIERCE AND BUNGALOW, CLEARWATER COUNTY, IDAHO

By Anna Hietanen

ABSTRACT

The northwestern part of the Idaho batholith consists of several intrusive bodies of differing composition, separated by belts of metasedimentary rocks, by dike rocks, or by fault zones. On the basis of chemical analyses and thin-section studies, the intrusive rocks can be grouped into two suites: a quartz dioritic suite poor in potassium and a quartz monzonitic suite rich in potassium. In the quartz dioritic suite, quartz diorite is most common and forms bodies as large as 200 square miles (about 600 sq. km) in land-surface area; other members of the suite—hornblendite, gabbro, and tonalite—form small sill-like bodies. Among the rocks of the quartz monzonitic suite, quartz monzonite and granite occur in large bodies, 50–100 square miles (100–250 square km in land-surface area); the mafic varieties are in small bodies, dikes, and inclusions. Both suites are represented also by pegmatites and hypabyssal dike rocks.

Field relations, structures, and textures indicate that the quartz diorite and related rocks were emplaced earlier than the rocks of the quartz monzonitic suite. According to lead-alpha age determinations, both suites are of Early Cretaceous age (Larsen and others, 1958) and were thus emplaced during a relatively short time interval. Because the quartz diorite and related rocks are in the western part of the area and the quartz monzonite and granite are in the eastern part, it is evident that magmatic activity started in the west with the intrusion of rocks poor in potassium and later moved toward the east as the magmas became enriched in potassium.

Variation diagrams indicate a close compositional relationship between the rocks of both suites. Notable differences occur only in the distribution of potassium, calcium, and vanadium. Two possibilities of origin are considered: either the quartz dioritic rocks are early differentiates of the quartz monzonitic magma, or both magmas were formed directly through melting or parts of the crust. The latter mode of origin postulates that the difference in the chemical compositions of the two rock suites is due to the differences in material from which the magmas were formed, or that the differentiation took place during the formation of the magmas rather than during their crystallization.

INTRODUCTION

This work is a petrologic and structural study of the igneous and metamorphic rocks in the northernmost part of the Idaho batholith. Metamorphic rocks are discussed only briefly because similar rocks have been described in detail in earlier papers (Hietanen, 1962, 1963a, b). A petrographic description of quartz monzonite and related rocks in a stocklike mass just north of the area studied is included in this report for comparison with the quartz monzonite in the southern part of the area.

During the present study, it became evident that much of the area shown on earlier geologic maps as the northwestern part of the Idaho batholith (Anderson, 1930; Ross and Forrester, 1947), is actually underlain partly by prebatholitic metamorphic rocks and partly by small satellitic bodies of intrusive rocks, the composition of which has a rather wide range. It was necessary to map this area in considerable detail in order to establish accurately the north boundary of the batholith, to establish the relations between the prebatholitic metasedimentary rocks and the igneous rocks, and to examine the sequence of intrusion.

Chemical petrology is used to determine the magmatic relations between the various intrusions of the Idaho batholith and its outliers. Study of the effects of the intrusive bodies on the metasedimentary country rocks throws light on the physico-chemical conditions of the intrusion.

The area studied lies in the south-central part of Clearwater County in the southern part of the Idaho "panhandle". It joins the areas described in earlier reports (fig. 1). The Orofino, Dent, and Big Island areas, and the Headquarters quadrangle (Hietanen, 1962) adjoin the area of this report in the west; the Boehls Butte quadrangle (Hietanen, 1963a) and the Elk River-Clarkia area (Hietanen, 1963b) lie to the northwest. The northeastern part of the area (pl. 1) comprises about 300 square miles around Bungalow Ranger station and is called the Bungalow area in this report. It is bounded by long 115°25′ and 115°45′ W. and lat 46°30′ and 46°45′ N. The area adjoining the Bungalow area and the Headquarters quadrangle on the south is called the Pierce area (pl. 2). It includes two small logging towns, Weippe and Pierce, and is bounded by long 115°35′ and 116°00′ W. and lat 46°17′ and 46°30′ N.; it covers an area of about 200 square miles. Quartz monzonite just north of the Bungalow area covers an area of about 27 square miles and is well exposed along Beaver Creek (pl. 3).

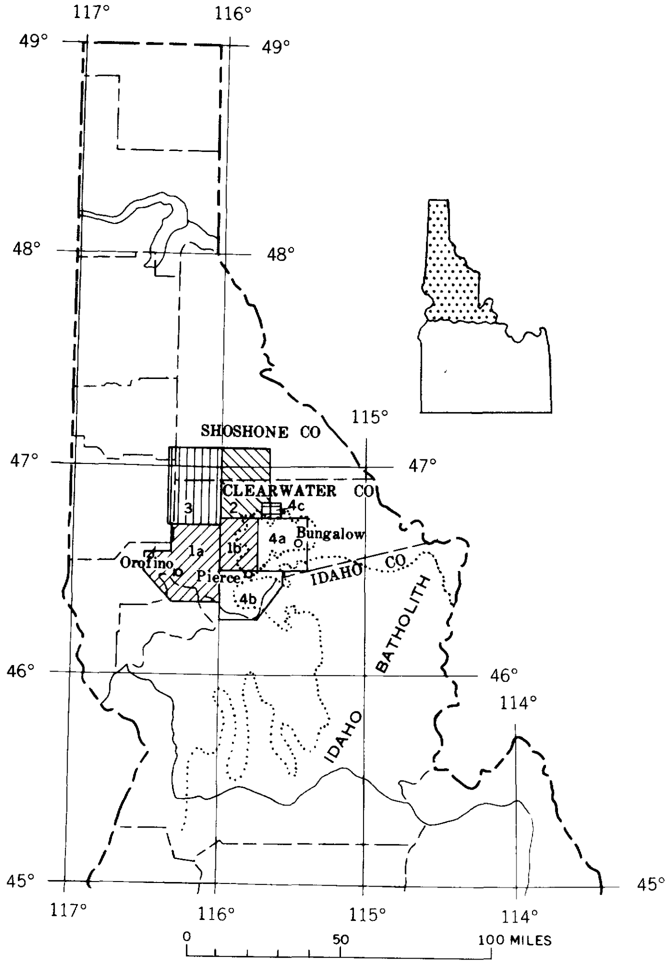

FIGURE 1.—Index map of northern Idaho: 1a, Orofino, Dent, and Big Island areas commbined; 1b, Headquarters quadrangle; 2, Boehls Butte quadrangle and vicinity; 3, Elk River-Clarkia area; 4a, Bungalow area; 4b, Pierce area; 4c, Beaver Creek area. The dotted line shows the outline of the northern part of the Idaho batholith.

The western part of the Pierce area (pl. 2) lies along the east margin of the Columbia River Plateau, which here is about 3,000 feet above sea level. The plateau is cut deeply by two large streams, Lolo and Orofino Creeks, the bottoms of which are about 1,200 feet above sea level.

The North Fork of the Clearwater River flows through the northeastern corner of the Bungalow area, where it turns from west to north and, farther downstream, to northeast. Most of the creeks are tributaries of this river. The southwestern part of the Pierce area is drained to the South Fork of the Clearwater River by Orofino, Jim Ford, and Lolo Creeks and their tributaries.

Part of the Columbia River Plateau in the mapped area is farming land and part is covered by a heavy growth of timber, which includes white and yellow pine, fir, cedar, hemlock, spruce, and larch. The forested mountain area is logged for white and yellow pine and cedar.

Most of the roads (pls. 1, 2) are gravel or dirt, but the road from Pierce and Weippe to Orofino and points west is paved. The gravel road from Pierce to Bungalow Ranger station is good, and many new logging roads have been constructed during recent years. These roads make the area easily accessible during the dry summer season, which usually lasts from the beginning of July to the end of August.

The field work on which this report is based was done intermittently during the summers of 1952, 1957, and 1958. In 1952, the author was assisted by Dorothy Rainsford and Cynthia Wilkin. The canyon of the swift North Fork of the Clearwater River north of Bungalow was mapped with the aid of a rubber raft.

LOCALITY AND SPECIMEN NUMBERS

Specimen numbers are used as locality numbers and are shown on plates 1 to 3. The following index gives the section, township, and range for each locality mentioned in the text:

No.	*Township*	*Range*	*Sec.*
574	40 N.	7 E.	5
578	40 N.	6 E.	24
719	39 N.	6 E.	21
758	38 N.	8 E.	20
818	35 N.	5 E.	23
827	35 N.	6 E.	20
830	35 N.	6 E.	5
831	36 N.	6 E.	32
848	39 N.	8 E.	6
1063	40 N.	6 E.	26
1064	40 N.	6 E.	26
1065	40 N.	6 E.	26
1072	40 N.	7 E.	7
1073	40 N.	7 E.	7
1079	40 N.	7 E.	31
1081	39 N.	7 E.	6
1085	39 N.	7 E.	18
1121	36 N.	7 E.	20
1125	36 N.	7 E.	19
1126	36 N.	7 E.	19
1139	38 N.	7 E.	33
1143	37 N.	6 E.	22
1171	38 N.	8 E.	22
1539	35 N.	5 E.	21
1880	34 N.	9 E.	[1]5
1944	38 N.	7 E.	35
1976	35 N.	6 E.	7
1983	38 N.	6 E.	31
1988	36 N.	6 E.	7

[1] East of the area of plate 2.

Localities 20, 32, 42, 43, 180, 190, 213, 292, 301, 322 and 379 are west of areas of plates 1 and 2 and are shown on the maps of Headquarters quadrangle and of Orofino, Dent, and Big Island areas (Hietanen, (1962).

MAJOR GEOLOGIC UNITS

The major geologic units (fig. 2), in the order of decreasing age are the Belt series (Precambrian), rocks of the Idaho batholith (Cretaceous), and the Columbia River basalt (Miocene and Pliocene(?)).

The Belt series is a thick sequence of sediments metamorphosed to diopside and biotite gneiss, schist, and quartzite. Equivalents to the Prichard, Revett, and St. Regis formations are exposed in various parts

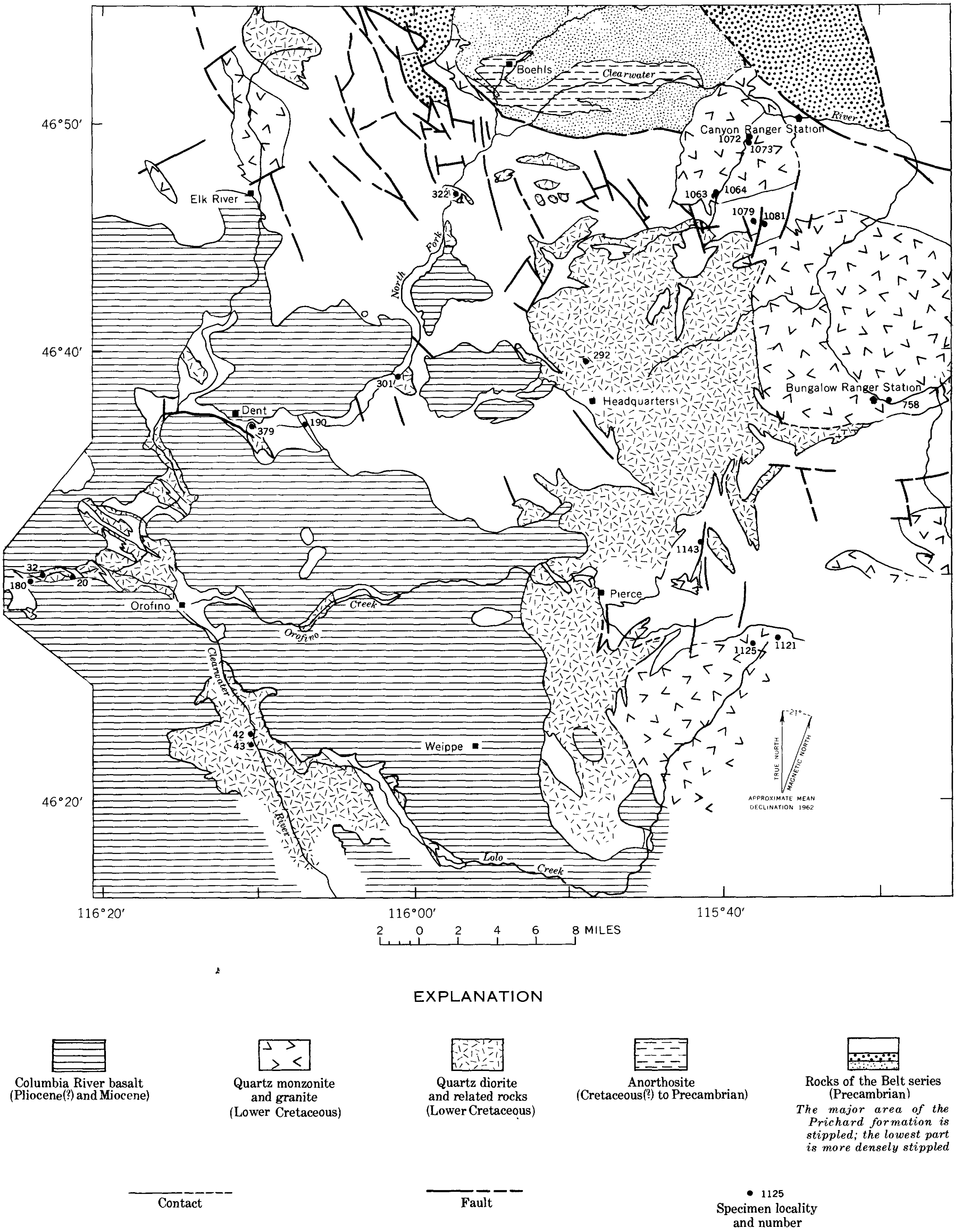

Figure 2.—A geologic sketch map showing the distribution of the igneous rocks along the northwest border zone of the Idaho batholith.

of the area (pls. 1, 2) but most of the rocks in the area are equivalent to the Wallace formation.

In the area mapped, the Idaho batholith is a composite batholith consisting of two larger and several small bodies of petrologically different plutonic rocks. These bodies are separated from each other by rocks of the Belt series, by dike rocks, or by sheared contact zones. The southernmost body, which is the largest, extends at least 150 miles south of the present area and consists of coarse-grained quartz monzonite. Quartz diorite forms one large and several small bodies. The largest body extends west of the mapped area and covers about 200 square miles. A large stocklike intrusion of granite covers the northeastern part of the area. Another stocklike mass occurs north of the area and consists of fine- to medium-grained quartz monzonite (pl. 3).

ANORTHOSITE, ITS COUNTRY ROCKS, AND THE OROFINO SERIES OF ANDERSON (1930)

The biotite gneiss exposed low on the Canyon walls of Lolo Creek in the southwestern part of the Pierce area is similar to the medium-grained biotite gneiss exposed 16 miles to the northwest, near Orofino and described in an earlier report (Hietanen, 1962).

The anorthosite shown on the western part of plate 3 and its country rocks have been described and discussed in an earlier report (Hietanen, 1963a). Description of metamorphic rocks in the eastern part of this plate will be included in a later report.

BELT SERIES

DISTRIBUTION

The northern contact of the intrusive rocks follows closely the north boundary of the Bungalow area (pl. 1), leaving only a narrow strip of metamorphic rocks in the northwestern corner. Most of the rocks of the Belt series are exposed south of Bungalow (pl. 1) and continue to the northeastern part of the Pierce area (pl. 2). Most of this terrane was formerly mapped as the Idaho batholith (Ross and Forrester, 1947). In addition to these occurrences several inclusions of rocks of the Belt series, ranging from ½ to 3 miles in length, are within the intrusive masses in the western and southern parts of the mapped area.

LITHOLOGY AND CORRELATION

Most of the rocks of the Belt series consist of diopside and biotite gneiss, interbedded with garnet-mica schist that locally contains abundant sillimanite. Pure quartzite beds are rare.

Comparison of the lithology and stratigraphic sequence of the metamorphic rocks with those of the Belt series northwest of the mapped area indicates that parts of the formations are equivalent to the Prichard, Revett, St. Regis, and Wallace formations. The Burke formation, which occurs between the Prichard formation and the Revett quartzite in Shoshone County, was not identified with certainty. The uppermost beds of the Prichard formation as herein mapped probably are equivalent to the Burke formation.

Coarse-grained fairly pure quartzite occurs along the northern boundary of the Bungalow area (pl. 1) and in the northern part of the Pierce area (pl. 2), east of Pierce. The quartzite in the Bungalow area continues northward into the Boehls Butte quadrangle, where it has been mapped as the Revett quartzite (Hietanen, 1963*a*). The quartzite south of Hemlock Butte and southwest of Dan Lee Lookout (southeast of Pierce) is petrographically and structurally similar to the Revett quartzite and occupies a corresponding stratigraphic position under a thick sequence of interbedded schist and gneiss and is therefore also mapped as Revett. A layer of schist that overlies the Revett quartzite is considered to be equivalent to the schist of the St. Regis formation near Elk River (Hietanen, 1963*b*). Between Hemlock Butte and Dan Lee Lookout (pl. 2) a layer of schist just north of the quartz monzonite body dips under the Revett quartzite and is mapped as the Prichard formation. The Burke formation, which in Shoshone County is transitional between the Prichard formation and the Revett quartzite and thus should underline the Revett quartzite, is not identified; either its equivalent was not deposited in the Pierce area or equivalent beds are here included with the adjacent formations. The contact between the Revett quartzite and the underlying schist (Prichard formation) is fairly sharp.

A thick sequence of diopside gneiss, biotite gneiss, and interbedded schist overlies the St. Regis formation; it is mapped as the Wallace formation, which elsewhere rests on the St. Regis formation. In this area the stratigraphic sequence within the Wallace is difficult to establish because of faulting and discontinuous exposures. The attitude of beds north of Hemlock Butte indicates that the lowest beds consist of thin-bedded biotite gneiss and diopside-plagioclase gneiss. This lower gneiss unit, which may be as much as 1800 feet thick, is overlain by a sillimanite-garnet-biotite schist with interbedded quartz-rich layers. The thickness of the schist unit could not be measured but is estimated to be about the same as that of the gneiss unit. Another unit of diopside-plagioclase gneiss crops out north of the schist and is in turn overlain by a second schist unit. This sequence is similar to that of the Wallace formation in the Headquarters quadrangle

(Hietanen, 1962) and near Elk River (Hietanen, 1963b).

Outcrops along French Creek show that the gneiss units consist of rock types that are similar to those of the quartzite-gneiss units near Elk River, about 30 miles to the northwest. At French Creek layers of thin-bedded biotite gneiss alternate with layers of banded diopside-plagioclase gneiss, which is the major rock type. In some layers, beds of biotite quartzite, 1 to 2 cm thick, are interbedded with diopside-plagioclase gneiss. A number of thin fairly pure quartzite layers that contain only a few mica flakes are interbedded with biotite gneiss and also with thin-bedded diopside-plagioclase gneiss.

PETROGRAPHY

PRICHARD FORMATION

The coarse-grained mica schist that dips under the Revett quartzite between Hemlock Butte and Dan Lee Lookout is mapped as the Prichard formation. This schist is poorly exposed along the south slope of a wooded ridge. The outcrops on the western end of the ridge south of Orofino Creek consist of bedded biotite gneiss and biotite-muscovite schist containing garnet and sillimanite. The beds of biotite gneiss range from 2 to 5 cm in thickness, are medium grained and consist of quartz, plagioclase, biotite, and some muscovite. The schist layers are coarse grained and rich in both micas. The amount of garnet and sillimanite ranges from 0 to about 10 weight percent. The southern part of this schist is granitized and grades into a coarse-grained quartz monzonitic rock which includes remnants of schist and gneiss.

REVETT QUARTZITE

The Revett quartzite along the northern boundary of the Bungalow area is well exposed along logging roads. Most of it is coarse grained and light bluish. The individual beds, 1 to 30 cm thick, are separated by paper-thin laminae of mica. About 85 to 95 percent of the rock is quartz; the rest is plagioclase, microcline, biotite, and muscovite. The quartz grains range from 0.5 to 2 mm in diameter, are strained, and have interlocking boundaries. Feldspars occur as isolated grains. The micas, especially muscovite, tend to form paper-thin laminae that separate thick layers of pure quartz. Laminae south of Sheep Mountain Creek have abundant sillimanite prisms, 3 to 10 mm long, whereas numerous small needles of sillimanite with nearly square cross sections and a few flakes of biotite and muscovite are scattered in the quartzose layers. In some beds biotite flakes, 1 to 2 mm long, are scattered throughout the rock.

Many of the pure quartzite beds near the northern boundary of the Bungalow area consist of polygonal grains from 0.01 to 0.04 mm in diameter. A few larger grains, about 0.2 mm in diameter, occur among the small grains, and some of these large grains have a relict mosaic structure, which suggests that large grains grew at the expense of small ones during recrystallization.

The Revett quartzite on Dan Lee Ridge southeast of Pierce is poorly exposed and was traced mainly by float. Most of the float is very pure coarse-grained quartzite with very little if any mica. Sillimanite is abundant about a mile west-southwest of Hemlock Butte. The outcrops south of Orofino Creek contain some biotite; those north and west of Dan Lee consist of bluish to white coarse-grained quartzite containing about 10 percent feldspar and about 2 percent biotite; all minor constituents occur in scattered grains.

ST. REGIS FORMATION

Schist overlying the Revett quartzite is considered to be equivalent to the St. Regis formation. This schist is thin bedded or laminated and contains layers of thin-bedded biotite quartzite and biotite gneiss. The major constituents are quartz, plagioclase, biotite, sillimanite, garnet, and muscovite. In the laminated schist, layers (2 to 3 mm thick) of quartz, plagioclase, and small flakes of biotite, are separated by paper-thin micaceous laminae. The amount of biotite exceeds that of muscovite, and in many layers, biotite is the only micaceous mineral. Sillimanite and garnet range from 0 to about 15 percent. In the layers of thin-bedded biotite quartzite and biotite gneiss, paper-thin biotite laminae separate the individual beds, which are 2 mm to 2 cm thick and rich in quartz or in quartz and oligoclase. Pegmatitic veins are common in the vicinity of Hemlock Butte and Dan Lee Lookout. These veins range in thickness from 1 to 50 cm and locally form as much as 60 percent of the rock. The composition and origin of these veins is dicussed in the chapter on the contacts of plutonic rocks.

WALLACE FORMATION

Most of the metamorphic rocks in the area are similar to the rock types mapped as the Wallace formation in the Headquarters quadrangle (Hietanen, 1962). The equivalent of this formation in the present area consists of at least two fairly thick units of diopside gneiss and schist. In each unit, layers of quartzite and biotite gneiss are interbedded.

SCHIST UNITS

The schist units consist mainly of biotite-garnet-sillimanite schist with interbedded layers of biotite gneiss. The schist is coarse grained and many outcrops contain abundant pegmatitic veins. The major constituents are quartz, plagioclase (An_{24}), and biotite. Garnet, sillimanite, and muscovite are in places abundant but can be sparse or lacking. Magnetite and zircon are the common accessories. The amount of quartz is usually much larger than that of oligoclase, but in some outcrops, especially in those near tonalitic bodies, the amount of oligoclase ranges from 40 to 50 percent. The amount of biotite exceeds that of muscovite or the biotite is the sole micaceous mineral. Garnet crystals range from a fraction of a millimeter to about 1 cm in diameter; the maximum amount is about 5 percent. Sillimanite, though commonly sparse or lacking, is as much as 15 percent in such outcrops as those along Hook Creek (sec. 35, T. 38 N., R. 7 E.), others about 1½ miles northeast of Hemlock Butte, and still others along Lolo Creek. Muscovite is sparse in the sillimanite-rich schist.

The biotite gneiss interbedded with the schist is medium grained and contains more oligoclase and less biotite than the schist. Garnet occurs sparingly and sillimanite only in micaceous laminae. Most of the gneiss is thin bedded and resembles the biotite gneiss of the St. Regis formation.

GNEISS UNITS

At least two units consisting of a fairly heterogeneous sequence of diopside-plagioclase gneiss, biotite gneiss, and quartzite are interbedded with the schist of the Wallace formation. Most of the gneiss and quartzite is thin bedded and the mineral content of the individual beds varies considerably. A few thick beds are of more homogeneous diopside-plagioclase gneiss.

In typical gneiss along Orogrande Creek, light-green diopside-bearing layers, 1 to 5 cm thick, are separated by biotite-bearing laminae or are interbedded with layers of plagioclase quartzite. The individual beds in plagioclase quartzite are 2 to 5 cm thick and are separated by thinner biotite-bearing or diopside-bearing layers. Megascopically this gneiss is very similar to parts of the quartzite-gneiss units near Elk River (Hietanen, 1963b), except that the grain size tends to be larger and pegmatitic veinlets are more common.

Thick beds of diopside-plagioclase gneiss are exposed in many localities along French and Orofino Creeks. These layers resemble thick beds of diopside-plagioclase gneiss in the Headquarters quadrangle (Hietanen, 1962). The major constituents, quartz, plagioclase (An_{29-30}), and diopside, occur about in equal amounts. Sphene and magnetite are the common accessories. Quartz and plagioclase occur as irregular-shaped grains with rounded corners. The size of the quartz grains in a common variety of diopside gneiss ranges from ¼ to 1 mm in diameter and that of the plagioclase grains from ½ to 2 mm. Diopside is light green with $\alpha=1.684\pm0.001$, $\beta=1.691\pm0.001$ and $\gamma=1.709\pm0.001$. It occurs in subhedral grains that generally measure 1 to 2 mm in diameter. In some outcrops, as for instance in locality 1988, about 2 miles east of Pierce, euhedral prisms, up to 5 cm long, are embedded in a coarse-grained matrix of quartz and plagioclase (An_{29}).

In most specimens microcline is sparse and appears in the form of small square antiperthitic inclusions in plagioclase. In specimen 1944, however, several layers, 2 to 3 mm thick, contain plagioclase and microcline in equal amounts. This specimen was collected along Hook Creek, 3 miles southwest of Bungalow, just south of the contact of diopside gneiss with quartz diorite. Plagioclase (An_{22}) in this rock is more sodic than that usually found in the diopside gneiss. Diopside is altered to antigorite along the cracks and cleavage planes, and very light green hornblende instead of diopside occurs in many layers, especially in those that contain microcline. Microcline occurs as polygonal grains with the plagioclase.

Actinolite with $\alpha=1.612\pm0.001$, $\beta=1.625\pm0.001$, $\gamma=1.640\pm0.001$ and $Z\wedge c=25°$ occurs in several layers in diopside-plagioclase gneiss along Orofino and Orogrande Creeks. The actinolite is darker green than the diopside and can be easily identified in the outcrops.

Layers of biotite gneiss that are interbedded with diopside-bearing gneisses are thin bedded, medium grained, and consist of quartz, oligoclase, and biotite. Muscovite is usually absent. Microcline occurs in some layers, especially in those that contain granitic veinlets. The very thin bedded biotite gneiss 1½ miles southwest of Musselshell and that just north of Musselshell (loc. 827) are examples of this type. In these rocks, paper-thin biotite laminae separate the light-colored layers that consists of quartz, oligoclase, and some microcline. More potassium feldspar occurs in the medium-grained granite-pegmatitic veinlets which are 0.5 to 1 cm wide and parallel the bedding. Grains in the gneiss range from 0.2 to 1 mm in diameter whereas those in the veinlets are 1 to 2 mm long. These veinlets are common near the contacts of the coarse-grained quartz monzonite and their origin is discussed later.

Beds of light bluish-gray, fairly pure quartzite are exposed in sections 10 and 22 of T. 37 N., R. 6 E. in the vicinity of Orogrande Creek. These beds range from 5 to 20 cm in thickness and are separated by micaceous laminae or by layers, about 1 cm thick, of biotite quartzite. Scattered grains of oligoclase constitute 5 to 15 percent of most layers. The quartz grains are rounded with interlocking borders. The texture and mineralogy of these beds resemble those of the white granular quartzite in the Elk River-Clarkia area (Hietanen, 1963b) even though the beds under discussion are more completely recrystallized and quartz is of translucent variety giving a glassy appearance to many beds.

STRUCTURE OF THE BELT SERIES

The metamorphic rocks trend mainly eastward, thus deviating considerably from the regional northwesterly trends common in the rocks of the Belt series to the northwest. Northwesterly trends can be recognized only in the extreme northwestern part of the Bungalow area (pl. 1). Local deviations from easterly trend occur east of Pierce where the strike of the Wallace formation swings north (pls. 1 and 2). The northerly trend seems to be localized in a zone about 4 miles wide extending northward through the quartz diorite in the western part of the Bungalow area where it is defined by a swarm of northward-trending dikes and inclusions. Farther toward the north several large faults occur in this zone, indicating that it had been a zone of weakness during deformation.

Attitudes of the beds and stratigraphic relations show that folding is isoclinal, with local overturning to the south. Dips between 45° and 85° are most common; beds are flat only on the crests of folds. In the zone of north-south trends east of Pierce the beds dip eastward, which indicates overturning to the west.

The axes of folds plunge mainly east or southeast, although some in the northwestern part of the Bungalow area and in the vicinity of Hemlock Butte plunge northwestward.

Lineation appears as a fine wrinkling and parallel orientation of elongated mica flakes along the flanks of the folds. It either is parallel to the fold axis or intersects it at an angle of 60° to 80°. Very strong lineation with local small-scale folding on it occurs in some outcrops near Hemlock Butte. This lineation plunges 60° to 65° north-northeast and makes an angle of about 50° with the northwestward-plunging fold axis.

The relation of the lineation to the fold axis is similar to that observed in the areas to the northwest (Hietanen 1961, 1962) and may be attributed to two acts of folding. The axes of the major folds trend mainly in an easterly or southeasterly direction, and those of the minor and somewhat later folds, in a northeasterly direction.

Generally the foliation in the schists and the gneisses is parallel to the bedding. A weak transecting foliation was observed at only a few localities in the quartzitic layers, as near the mouth of Crystal Creek, in the southwestern part of the Bungalow area. Strong fracturing parallel to a fault system is seen in the outcrops of Revett quartzite along Deadhorse Creek and west of Swanson Creek, near the north boundary of plate 1. The lack of mineral development along these fractures indicates that they are late. Their formation is probably related to the intrusion of the granite, as discussed later.

Several northward- or northwestward-trending and a few eastward-trending faults transect the metamorphic rocks or occur along their contacts with the igneous rocks. Most faults are steeply dipping with only minor displacement, mainly vertical; but a few normal faults such as those along Sheep Mountain Creek in the northwestern part of the Bungalow area also interrupt the stratigraphic sequence. These faults are parallel to the regional fault system; specifically, the northward-trending faults form the south end of a fault zone that is one of the major breaks in the area north of the Bungalow area (Hietanen, 1961, and 1963a).

METAMORPHISM

The development of sillimanite, garnet, biotite, and some muscovite in the metamorphosed equivalents of the shaly layers and diopside, actinolite, and plagioclase (An_{20-85}) in the calcareous layers of the Belt series indicates *P-T* conditions of the sillimanite-muscovite subfacies of the amphibolite facies during recrystallization. The orientation of the micaceous minerals, sillimanite, and actinolite parallel to the bedding, which was the dominant plane of slip during the folding, indicates that recrystallization took place contemporaneously with deformation.

In some diopside gneiss, secondary hornblende is developed. These hornblende crystals are larger than the other mineral grains and are oriented at random, many across the bedding. They are similar to the secondary hornblende described earlier from the Headquarters quadrangle (Hietanen, 1962). Their development is thought to be connected with the emplacement of the plutonic rocks.

In many outcrops east of Pierce, pegmatitic veins and secondary feldspars are abundant in the biotite gneiss and schist. Most of the veins are parallel to the bedding; only a few are transecting. As a rule the total amount of feldspars increases toward the intrusive rocks. In many places along Orofino Creek and near Dan Lee Lookout, there is complete gradation from metamorphic rocks to those with a plutonic appearance. All feldspathization is therefore thought to be connected with the emplacement of the plutonic rocks, discussed later.

PLUTONIC ROCKS

DISTRIBUTION AND CLASSIFICATION

Each petrologically different rock type forms one or several intrusive bodies, which are bordered by metamorphic rocks, dike rocks, or other plutonic rocks. The western part of the Bungalow area (pl. 1) and the central part of the Pierce area (pl. 2) are occupied by quartz diorite so coarse grained that the major constituents—plagioclase, quartz, hornblende, and biotite—are obvious. Tongues from this large body extend into the country rocks southwest of Bungalow Ranger Station and along Orogrande and French Creeks. Small bodies of similar quartz diorite crop out 2 miles northeast of French Mountain, 4 miles east of Bungalow, and along the north boundary of the Bungalow area.

About 200 square miles are underlain by quartz diorite; about 140 square miles are exposed in the mapped area, and 60 square miles in the Headquarters quadrangle to the west. Larsen and Schmidt (1958, p. 4) estimated the area of quartz diorite to be 1,000 square miles, assuming that considerable quartz diorite underlies the Columbia River basalt around Weippe. However, all outcrops of prebasaltic rocks in that vicinity are of metasedimentary rocks. Metasedimentary rocks are also exposed along the southern tributaries of Lolo Creek. Accordingly it is here assumed that only a few tens of square miles of quartz diorite is covered by basalt. The platy structure along the western border zone of the quartz diorite dips about 45° eastward (Hietanen, 1962); therefore it is not likely that this intrusion would extend farther to the west and be covered by only a thin layer of metasedimentary rocks.

Rocks that have a chemical composition intermediate between quartz diorite and trondhjemite are called tonalite in this report. The tonalite is more silicic and finer grained than the quartz diorite. Biotite is the only dark-colored constituent, and plagioclase is more sodic than the plagioclase of the quartz diorite. Tonalite occurs as small lenticular bodies next to the quartz diorite, or as separate dikes and other small intrusive bodies in the metamorphic rocks.

The quartz monzonite in the southeastern part of the mapped area is a light-gray, coarse-grained rock in which biotite is the only dark constituent and in which a part of the orthoclase occurs as phenocrysts. In the eastern part of the Pierce area the northern border zone of this quartz monzonite consists of light-gray medium-grained rock that megascopically appears very similar to the tonalite. About a mile south of the north boundary abundant large orthoclase crystals are embedded in this light-gray medium-grained rock, which with an increase in the number of orthoclase phenocrysts grades to a normal coarse-grained quartz monzonite. The light-gray medium-grained rock and the "groundmass" between the large orthoclase crystals have a composition which is intermediate between the tonalite and the main body of the quartz monzonite and is therefore called monzotonalite in this report. The common occurrence of porphyritic dikes of monzotonalitic composition in the country rocks prove that magma of this intermediate composition also existed. Coarse-grained quartz monzonite with anhedral to subhedral phenocrysts is a common variety along the south border of Clearwater County and forms a major rock type in the body that continues southward from the county line and covers about 2,000 square miles in Idaho County (fig. 1).

Coarse-grained true granite with about equal amounts of sodic plagioclase, quartz, and orthoclase forms a fairly large rectangular body of about 110 square miles in the northeastern part of the Bungalow area. It is accompanied by granophyric dikes of the same composition.

A few small occurrences of ultramafic rocks (not shown on the maps) and of hornblende gabbro were found within and near the plutonic rocks.

Pegmatitic veins and dikes[2] are common in all rock types. The pegmatites related to the quartz diorite are plagioclase-biotite pegmatites, but those related to the quartz monzonite and the granite contain abundant potassium feldspar.

STRUCTURE

The structural elements and their orientation in each individual body of plutonic rocks differ markedly from those of the other bodies. These differences are most striking between the quartz diorite and the granite. Tonalitic bodies exhibit two structural varieties; in one variety, structures are similar to those in quartz diorite, and in the other, similar to those in quartz monzonite.

SHAPE OF THE BODIES

The quartz diorite mass is elongate from north to south. Its exposed length is 32 miles and its width is 4 to 10 miles. The direction of elongation parallels a prominent fault zone, north of the Bungalow area, that is the locus of a stocklike intrusion of quartz monzonite (pl. 3).

The western contact of the quartz diorite as exposed in the Headquarters quadrangle (Hietanen, 1962) dips from 40° to 50° E. and generally parallels the bedding of the rocks of the Belt series. The eastern contact dips 50° to 60° E. or is nearly vertical. Thus the quartz diorite mass is a dikelike body, dipping eastward and intruded parallel to a fault zone.

Reconnaissance mapping in Idaho County indicates that the three occurrences of quartz monzonite along the southern border of Clearwater County probably

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[2] Veins and dikes are used as descriptive terms, veins referring to thinner irregular bodies, and dikes to thicker crosscutting sheetlike bodies.

belong to a large body of quartz monzonite that extends 15 miles southeastward to the vicinity of Lochsa River where it curves southward and continues at least 30 miles farther.

Reconnaissance mapping east of the Bungalow area shows that the outline of the granite mass at Bungalow is rectangular with rounded corners. The dimension in the easterly direction is 13 miles and that in the northerly direction 11 miles. The contacts of the mass are steep and irregular, typical of a pluglike intrusion. The western contact lines up with a northward-trending fault that extends northward from Deadhorse Mountain.

CONTACTS

The quartz diorite, the tonalite, and the quartz monzonite have both discordant and concordant contacts with the metamorphic rocks. Both types of contacts are sharp or gradational, but the latter are more common. The width of the gradational zone varies from a few centimeters to hundreds of meters, and along the discordant contacts the plutonic rock sends tonguelike extensions into the country rock. This type of tonguing relation is more common where the intrusive rock is in contact with schist or biotite gneiss than with diopside gneiss.

The contact rock between the intrusive rock and the schist or gneiss is medium-grained biotite gneiss in which bedding is far less distinct than in the metamorphic rock. The grain size and the amount of plagioclase increase toward the intrusive rock. Tonalitic veins and masses are common in the biotite gneiss and schist near the quartz diorite and tonalite exposed along Orofino Creek southeast of Pierce. In this vicinity abundant oligoclase has crystallized in metamorphic rocks along a mile-wide contact zone, changing about 40 percent of the biotite gneiss and schist to a medium-grained rock that has tonalitic composition but contains ghostlike remnants of metamorphic rocks. The schist north of this zone (northwest of Rosebud Creek) also contains plagioclase-rich portions, small masses of inhomogeneous tonalitic rock, and numerous pegmatitic veins.

Migmatite is exposed 2½ miles southeast of Hemlock Butte along a road leading to Weitas Ranger Station. The older rock in this vicinity is thin-bedded biotite gneiss with quartzitic layers, and the nearby intrusive rocks are quartz monzonite. The migmatite contains two types of quartzofeldspathic layerlike masses and veins: medium-grained tonalitic veins, 1 to 3 cm thick, and coarse-grained pegmatite veins, 2 to 10 cm thick. The tonalitic veins, consisting of quartz, plagioclase (An_{26}), and some small flakes of biotite, are separated by gneiss layers that range from 1 to 3 mm in thickness and are rich in biotite. The texture of the tonalitic veins is granoblastic, and there is every gradation from these veins to the pegmatitic veins that have orthoclase in addition to quartz and plagioclase. The grain size increases from 1 to 2 mm in the tonalitic veins to about 3 to 10 mm, or even more, in the pegmatitic veins. The tonalitic veins and most of the pegmatitic veins parallel the bedding, but some pegmatitic veins cut the bedding. Because the mineralogy of the tonalitic veins is similar to that of the quartzofeldspathic layers in the biotite gneiss, although the grain size is larger, it is believed that these veins crystallized at the expense of sedimentary material. The occurrence of orthoclase in the pegmatites, however, proves that at least a part of the material (potassium and perhaps some other elements) in them was introduced because the country rock does not contain potassium feldspar. The source of the introduced material was most likely the nearby quartz monzonite. This migmatite represents a typical contact migmatite in which the minerals of the parent rock recrystallized and differentiated sufficiently to increase the amount of vein material.

Parts of the north border zone of the large quartz monzonite body are exposed between Weitas Ranger Station and Lean-to Ridge lookout and also along Hemlock Creek. This border zone is heterogeneous, containing abundant pegmatite and remnants of older metasedimentary rocks. The outcrops along the road north of Lean-to Ridge as well as those south of Weitas Ranger Station indicate that the border zone here consists of 50 to 70 percent pegmatite, about 20 percent rocks of the Belt series, and about 20 to 30 percent hypabyssal dike rocks. The rocks of the Belt series are mainly biotite gneiss and diopside gneiss with less biotite quartzite and schist. There is every gradation of these rocks to feldspathized gneiss and to quartz monzonitic gneiss in which ghostlike remnants of metasedimentary rocks are common. The orientation of relict bedding in the remnants of the metasedimentary rocks indicates that the position of these remnants was not changed during the emplacement of the igneous rocks.

The quartz monzonite along Hemlock Creek is part of a tonguelike extension. The northern part contains abundant pegmatite and the southern part and the core are fairly homogeneous coarse- to medium-grained quartz monzonite.

The contact zone of the granite stock near Bungalow Ranger Station differs strikingly from that of the coarse-grained quartz monzonite. The contacts with the country rocks—quartz diorite and the rocks of the Belt series—are discordant. In many places, granite porphyry occurs between the coarse-grained granite and its country rock. In most localities, the granite porphyry intrudes the granite or is in fault contact with it. However, along the northern contact zone, between Deadhorse Lookout and the North Fork of the Clearwater River, the rock grades from granite porphyry through coarse-grained porphyritic granite to coarse-grained granite. The porphyritic granite is mineralogically and chemically similar to the coarse-grained granite. In the porphyritic variety, the euhedral to subhedral crystals of quartz and feldspars are embedded in a medium-grained groundmass consisting of quartz, plagioclase, orthoclase, and biotite. Granophyric intergrowth of quartz and feldspars is common around the large perthitic orthoclase crystals (fig. 3*A*).

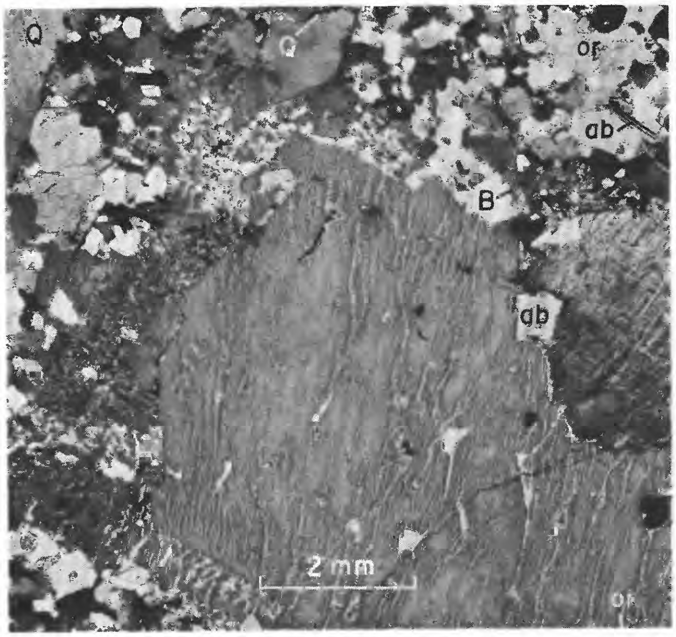
The porphyritic granite is well exposed along the North Fork of the Clearwater River half a mile northeast of the mouth of Governor Creek, where it grades to a coarse-grained granite toward the south and to a fine-grained granophyric granite (fig. 3*B*) toward the north, near the contact with the rocks of the Belt series. A similar granophyre occurs as a border facies of the granite porphyry west of Deadhorse Lookout.

Along the southern contact west of Bighorn Point, granite porphyry separates the coarse-grained granite from the rocks of the Belt series. This mass, lacking outcrops but mapped on the basis of float, is probably fairly thick; many dikelike bodies ranging from a few meters to about 30 meters in width occur elsewhere along this southern contact zone.

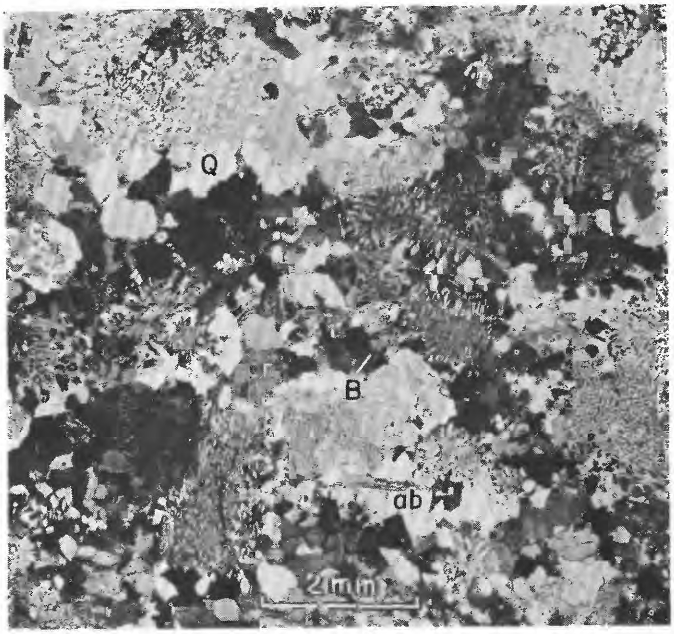
Near the mouth of Fischer Creek, 6 miles northeast of the mouth of Weitas Creek (east of the Bungalow area), granite porphyry occurs between coarse-grained granite and quartz diorite to the south. The western end of this quartz diorite mass is exposed along the North Fork of Clearwater River east of the mouth of Weitas Creek (pl. 1). Along Fischer Creek the granite itself, the quartz diorite, and also a gneissic medium-grained granite south of the quartz diorite are cut by many dikes of quartz porphyry.

Medium-grained equigranular pink granite forms a border facies of the coarse-grained granite west of Deadhorse Mountain. This granite is similar to a small mass of pink equigranular granite in the schist about 9 miles southwest of Deadhorse Mountain. The contact between the pink granite and the coarse-grained granite is not exposed. The two granites may well be separated by a fault that extends northward from Deadhorse Mountain.

Along Orogrande Creek about 1½ miles south of Bungalow, coarse-grained granite is in contact with gneissic quartz diorite. On a small scale, the contact



A, A porphyritic granite with large phenocrysts of perthitic orthoclase (or) and small subhedral phenocrysts of quartz (Q, only a part shown). Granophyric intergrowth of quartz and feldspar forms lacy rims around the orthoclase crystals; groundmass consists of quartz, orthoclase (or), ablite (ab), and a few small flakes of biotite (B). Locality 848, along the North Fork of the Clearwater River near the northern contact of the granite. Crossed nicols.



B, Fine-grained granophyric granite north of locality 848, along the river. Granophyre consists of quartz and orthoclase; quartz (Q), orthoclase (or), albite (ab), and biotite (B), are the constituents in the portion with a granitoid texture. Crossed nicols.

FIGURE 3.—PHOTOMICROGRAPHS OF THE BORDER ZONE OF THE GRANITE.

is discordant and the gray quartz diorite changes to a reddish orthoclase-rich medium-grained rock along the contact. The altered zone is about 1 meter thick and contains about 15 percent dark minerals. Thin sections show that the amounts of orthoclase, oligoclase, and quartz are about equal and that dark-brown biotite is the main dark constituent. Some hornblende is found with the biotite; and magnetite, apatite, and zircon are the accessories.

PLATY AND LINEAR STRUCTURES

Platy and linear structures are common in quartz diorite and related rocks, but rare in the rocks of the quartz monzonite series. Fairly well developed platy and linear structures occur in the border zone of the quartz diorite and in adjacent tonalite. Farther from the contact only the linear structure is visible; in the central parts of large bodies the minerals are oriented at random.

The foliation is due mainly to parallel orientation of biotite, but some of the hornblende is also oriented similarly. In places, both of these dark constituents form elongated clusters. The foliation is parallel to the contacts, which are parallel to the bedding of the country rocks. The linear structure is, as a rule, parallel to the fold axes of the metasedimentary rocks.

In contrast to the structures of quartz diorite and related tonalite, almost all the quartz monzonite and granite is massive. A weakly developed parallel orientation of biotite occurs rarely in the coarse-grained quartz monzonite. Minerals in the groundmass of the porphyritic quartz monzonite are oriented at random.

INCLUSIONS

Inclusions are common in quartz diorite, tonalite, and quartz monzonite but not in the granite near the Bungalow Ranger Station. Most inclusions are of rocks of the Belt series, and many, especially those consisting of schist and biotite gneiss, are feldspathized and migmatized.

A few gabbroic inclusions occur in quartz diorite. Minerals in these inclusions are the same as those in the quartz diorite except that plagioclase is more calcic and the amount of hornblende is 60 percent or more. Only large inclusions are shown on the maps and most of them consist of gneissic layers of the Wallace formation with a minor amount of schist. Many inclusions are elongate parallel to the bedding. Those which are not, as for instance those north and west of Browns Rock (pl. 1), occur on a ridgetop and their shape in plan is modified by erosion. Schist and some of the biotite gneiss in the inclusions contain abundant quartzofeldspathic veins and scattered grains of feldspar. The amount of quartzofeldspathic material in many inclusions is from 40 to 60 percent. Some parts of each inclusion, however, consist of unaltered rock that can be easily identified.

The secondary feldspar in the inclusions in the quartz diorite is mainly plagioclase (An_{22-28}). Potassium feldspar occurs only in the veins and in the pegmatitic dikes cutting the inclusions near contacts with quartz monzonite. The large elongate inclusion 2 miles west of Musselshell belongs to this group. This inclusion consists of very thin bedded biotite gneiss and schist similar to many layers in the Wallace formation. The schist contains numerous round plagioclase (An_{28}) grains which range from 3 to 6 mm in diameter. The gneiss contains two types of veins, fine-grained veins parallel to the bedding and medium- to coarse-grained ones which either parallel the bedding or cut across it. The fine-grained veins consist of quartz and oligoclase with very little biotite and occasionally some potassium feldspar. The medium- to coarse-grained veins are rich in orthoclase (about 45 percent) and contain about 20 percent oligoclase, 30 percent quartz, and as much as 5 percent biotite. Similar coarse-grained pegmatitic veins occur also in the schist. Large subhedral orthoclase crystals, similar to those in the quartz monzonite, appear in the schist next to a coarse-grained pegmatite (30 cm wide) about half a mile north of Musselshell Creek. The schist in this locality also contains abundant round plagioclase grains and is rich in biotite.

A small body of coarse-grained hornblende gabbro occurs in the center of this inclusion (loc. 818) and a fine- to medium-grained equigranular gray tonalite dike cuts the schist discordantly just south of the gabbro body . The gabbro consists of plagioclose (An_{55}), hornblende, and a little biotite.

Most of the inclusions in the quartz monzonite consists of thin-bedded biotite gneiss and diopside gneiss. These inclusions are migmatized and contain potassium feldspar and plagioclase in scattered grains, in rows of grains, and in quartzofeldspathic veins. A part of the potassium feldspar—especially small granoblastic grains in the thin-bedded biotite gneiss—is microcline, as shown by microscopic grid structure and by X-ray diffraction. The potassium feldspar in many veins is orthoclase, as indicated by X-ray diffraction. Large scattered grains show transition from orthoclase to microcline; X-ray diffraction of these crystals indicate a triclinicity of about 0.75 when determined by the method of Goldsmith and Laves (1954).

The granoblastic grains of medium size that occur with quartz and oligoclase are, in their mode of occurrence, similar to the potassium feldspar in the thin-bedded gneiss of the Wallace formation near Elk River and Clarkia. They most likely crystallized at the expense of sedimentary material. The orthoclase of the veins and the large scattered grains of potassium feldspar are secondary and crystallized from material supplied by the nearby igneous rock.

The diopside gneiss of the inclusions is much less granitized and migmatized than the schist and biotite gneiss. Common contact phenomena are alteration of diopside to green hornblende and development of secondary feldspar. Quartzofeldspathic veins and pegmatites are common but not as abundant as in the biotite gneiss and the schist.

JOINTS

Joints are prominent in the granite near Bungalow but are only locally developed in the other intrusive rocks.

The joint system in the granite is especially noticeable in the outcrops near Buckingham Point, where rugged peaks are bounded by northward-trending joints that dip about 70°W. This northward-trending joint system is also prominent north of the Bungalow Ranger Station, in the outcrops on the steep slopes of the North Fork of the Clearwater River. Westward-trending steeply dipping joints appear in several localities, but they are not as well developed as the northward-trending joints. In the northern part of the river valley the steeply dipping joints strike northwest; in places a northeastward-trending joint set is also present.

Nearly horizontal joints that are about perpendicular to the steeply dipping joints are prominent in the steep cliffs along the river, but are more difficult to detect elsewhere. Where measurement was possible, the dips range from 0° to 25°, those from 5° to 20° being most common.

The northwestward-trending steeply dipping joint system continues to the northwest border of the granite mass. Strong fracturing of the same trend affects the Revett quartzite to the northwest.

PETROGRAPHIC DESCRIPTION

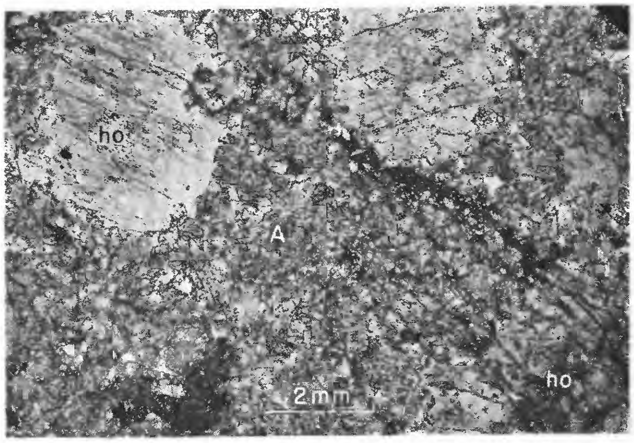
ULTRAMAFIC ROCKS

A hornblendite mass, a few meters thick, is included in quartz diorite about 2 miles east of logging camp 57. About 96 percent of this rock is hornblende, which appears dark in the hand specimen and pale green under the microscope. Most of the rest is greenish-brown biotite partly altered to chlorite. Small grains of magnetite occur as inclusions in hornblende.

Two sill-like bodies of pyroxenite crop out in the area. One is exposed on a road cut along Weitas Creek 1½ miles south of the creek mouth (loc. 1171). Large boulders from the other lie along the road 2 miles northwest of Brown Creek Lookout (loc. 1539).

Augite constitutes 60 to 70 percent of both masses. In the pyroxenite along Weitas Creek, dark round grains of hornblende ranging from 2 to 5 mm in diameter are embedded in grayish-green augite (fig. 4*A*).

A, Hornblende (dark)-bearing pyroxenite along Weitas Creek (loc. 1171).



B, Photomicrograph of the rock shown in *A*. Large round hornblende (ho) grains are embedded in medium-grained rock consisting of augite (A). Plane-polarized light.

FIGURE 4.—PHOTOGRAPH AND PHOTOMICROGRAPH OF PYROXENITE WITH HORNBLENDE CRYSTALS.

The augite is in grains 1 to 2 mm long that contain small patches of green hornblende (fig. 4*B*). The large hornblende grains are brownish green and have green borders. The rock also has small interstitial grains of untwinned plagioclase and magnetite.

In the mass northwest of Brown Creek Lookout, the hornblende grains are small and irregular. They are browner than those in the sill along Weitas Creek. The rock has 2 percent of interstitial quartz, rather than plagioclase.

HORNBLENDE GABBRO AND AMPHIBOLITE

Small bodies of hornblende gabbro—most from 100 to 600 m long—crop out within and near the quartz diorite. Only the largest of these are shown on plate 1; two are near Crystal Creek in section 34 of T. 38 N., R. 6 E.

Most of the gabbros are medium- to coarse-grained rocks, in which the minerals are the same as those in the quartz diorite, except that there is more hornblende and less quartz. Some of them, however, are very coarse grained and contain more plagioclase than the medium-grained variety.

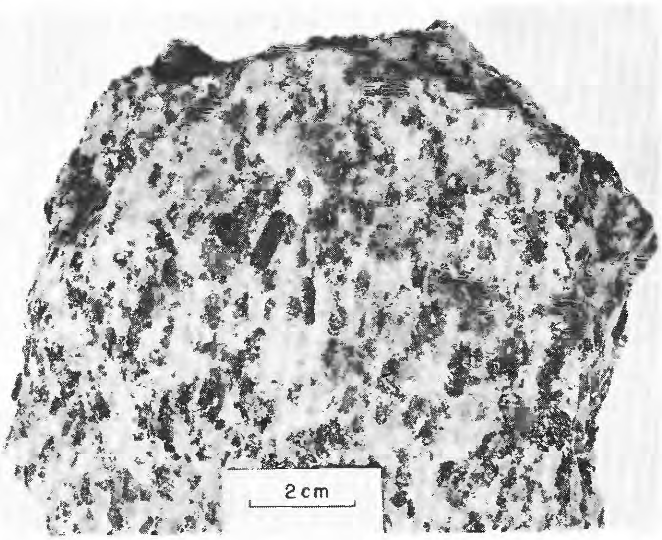
A few small bodies of medium-grained gneissic hornblende-plagioclase rock occur in the metamorphic rocks northwest of Hemlock Butte (pl. 2) and southeast of Larch Butte (pl. 1). The mineralogy of these amphibolites is similar to that of the hornblende gabbro and they may be old gabbroic intrusions that were deformed during the folding of the enclosing metasedimentary rocks.

The body of amphibolite that is about a mile north-northwest of Hemlock Butte is coarse grained and contains about 40 percent hornblende, 25 percent plagioclase (An_{65}), 20 percent quartz, and 15 percent garnet. The garnet crystals are rounded and range from 3 mm to 1 cm in diameter; many contain small quartz inclusions. Ilmenite-magnetite, in skeletonlike crystals and in groups of small elongate grains with round ends, constitutes 1 to 3 percent of this rock. Brown biotite is abundant in the border zone of this mass but sparse in the center. Other accessories are apatite, sphene, and zircon.

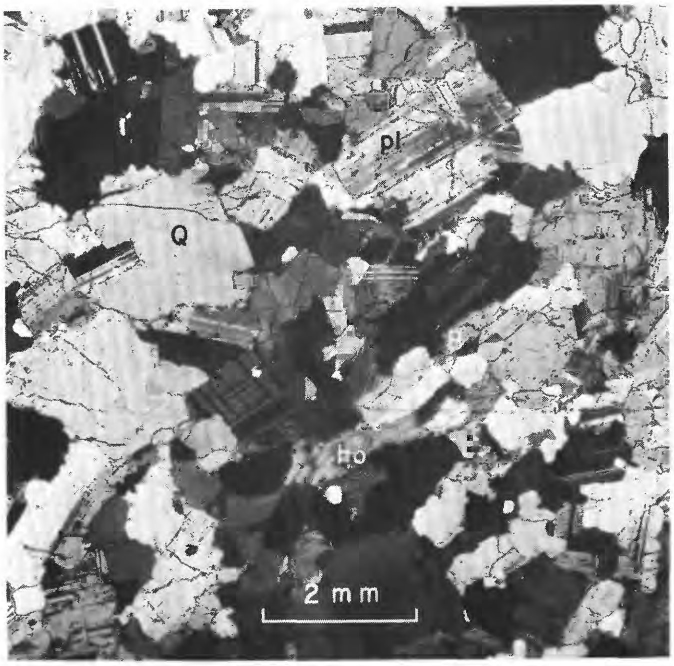
The mineralogy and mode of occurrence of the garnet amphibolite are similar to those of the garnet amphibolite near the anorthosite bodies in Boehls Butte quadrangle (Hietanen, 1963a), except that the amount of ilmenite-magnetite is larger near Hemlock Butte.

QUARTZ DIORITE

The quartz diorite is a coarse-grained rock in which the dark constituents, hornblende and biotite, are strongly contrasted against the light-colored mixture of quartz and plagioclase (fig. 5*A*). Part of the hornblende is in stubby prisms 5 to 15 mm long; smaller grains form clusters with biotite. Some of the central part of the largest body is somewhat coarser, with less biotite and less mineral orientation than the common variety. In many places near the contacts, variation in the distribution of biotite produces dark- and light-



A, A common type of quartz diorite a mile west of logging camp 57. The dark minerals, hornblende and biotite, show a parallel orientation. The light-colored constituents are quartz and plagioclase (loc. 1983).



B, Photomicrograph of the rock (no. 1983) shown in *A*. Unzoned plagioclase (An_{34}) occurs in elongated grains that show parallel orientation and complex twinning (pl). Quartz grains (Q) are anhedral and strained. Hornblende (ho) and biotite (B) are the dark constituents. Crossed nicols.

FIGURE 5.—PHOTOGRAPH AND PHOTOMICROGRAPH OF QUARTZ DIORITE.

colored portions. In a few localities, the quartz diorite grades to a light-gray gneissic tonalite in which biotite is the only dark constituent and the plagioclase is more sodic than that in the quartz diorite. Magnetite, apatite, and zircon are the common accessories.

The texture of the quartz diorite is hypidiomorphic granular (fig. 5*B*); hornblende and plagioclase are subhedral, whereas quartz fills the interstices. The grains of plagioclase (An_{36}–38) are 2 to 6 mm long. The rounded interstitial grains of quartz have diameters of 0.5 to 5 mm. The dark constituents are clustered; the individual grains in most common types are 3 to 10 mm long.

Biotite is reddish brown, strongly pleochroic, and has $\gamma = 1.651 \pm 0.002$; hornblende has α (pale green) $= 1.660 \pm 0.002$; β (green) $= 1.679 \pm 0.002$, and γ (blue green) $= 1.683 \pm 0.001$ measured in a specimen taken about a mile west of Camp 57 in the southwest part of the Bungalow area. Larsen and Schmidt (1958, table 9, nos. 2 and 3) give chemical analyses of biotite and hornblende in quartz diorite collected about 2 miles southeast of Camp 57. Calculation of the formula for this biotite gives the following result: $(K_{1.68}Na_{.09}Ca_{.19})_{\Sigma=1.96}$ $(Mg_{2.57}Fe_{2.24}Mn_{.02}Ti_{.46}P_{.02}Fe_{.25}Al_{.27})_{\Sigma=5.83}$ $(Al_{2.29}Si_{5.71})_{\Sigma=8.0}$ $(OH_{1.50}O_{22.50})_{\Sigma=24.0}$, in which Mg/Fe=1.

The formula calculated for the hornblende in the same rock is as follows: $(Na_{.28}K_{.14}Ca_{1.92})_{\Sigma=2.34}$ $(Mg_{2.29}Fe_{1.64}Mn_{.04}Ti_{.16}P_{.02}Fe_{.37}Al_{.61})_{\Sigma=5.13}(Al_{1.34}Si_{6.66})_{\Sigma=8.0}(OH_{1.46}O_{22.54})_{\Sigma=24.0}$ and Mg/Fe=1.4 (42 percent iron end member).

TONALITE

Two types of tonalite are distinguished in the area, gneissic and massive. The gneissic variety occurs in small lens-shaped bodies in the metasedimentary rocks and as a border facies of quartz diorite. The massive variety is in dikes transecting the metasedimentary rocks near the contacts of quartz diorite. The same types appear also west of the Bungalow area (Hietanen, 1962).

The major constituents of both varieties are plagioclase An_{30-34}, quartz, and biotite. The gneissic tonalite is medium grained, light to medium gray, and banded in an irregular manner. The banding is due to an irregular distribution of biotite, the amount of which ranges from 5 to 15 percent. The tonalite southeast of Pierce grades northward to feldspathized schist and gneiss by a gradual decrease in the amount of plagioclase and increase in the amount of biotite. Southward the tonalite grades to quartz diorite, and thus is a contact rock between the quartz diorite and metasedimentary rocks. Specimen L 227 of Larsen and Schmidt (1958) was collected from such a contact zone. Small bodies of similar gneissic tonalite are surrounded by metasedimentary rocks.

The massive tonalite is light bluish gray and fine to medium grained with a hypidiomorphic granular texture. In its mode of occurrence and mineralogy, it is similar to the intrusive tonalite near Dent (Hietanen, 1962).

QUARTZ MONZONITE AND RELATED ROCKS

Small exposures of quartz monzonite along the southeastern border of the Bungalow and Pierce areas are probably northward extensions of a large body. The westernmost occurrence contains in places large euhedral orthoclase phenocrysts. A stocklike body of quartz monzonite is exposed along Beaver Creek (pl. 3) just north of the Bungalow area.

STOCK AT BEAVER CREEK

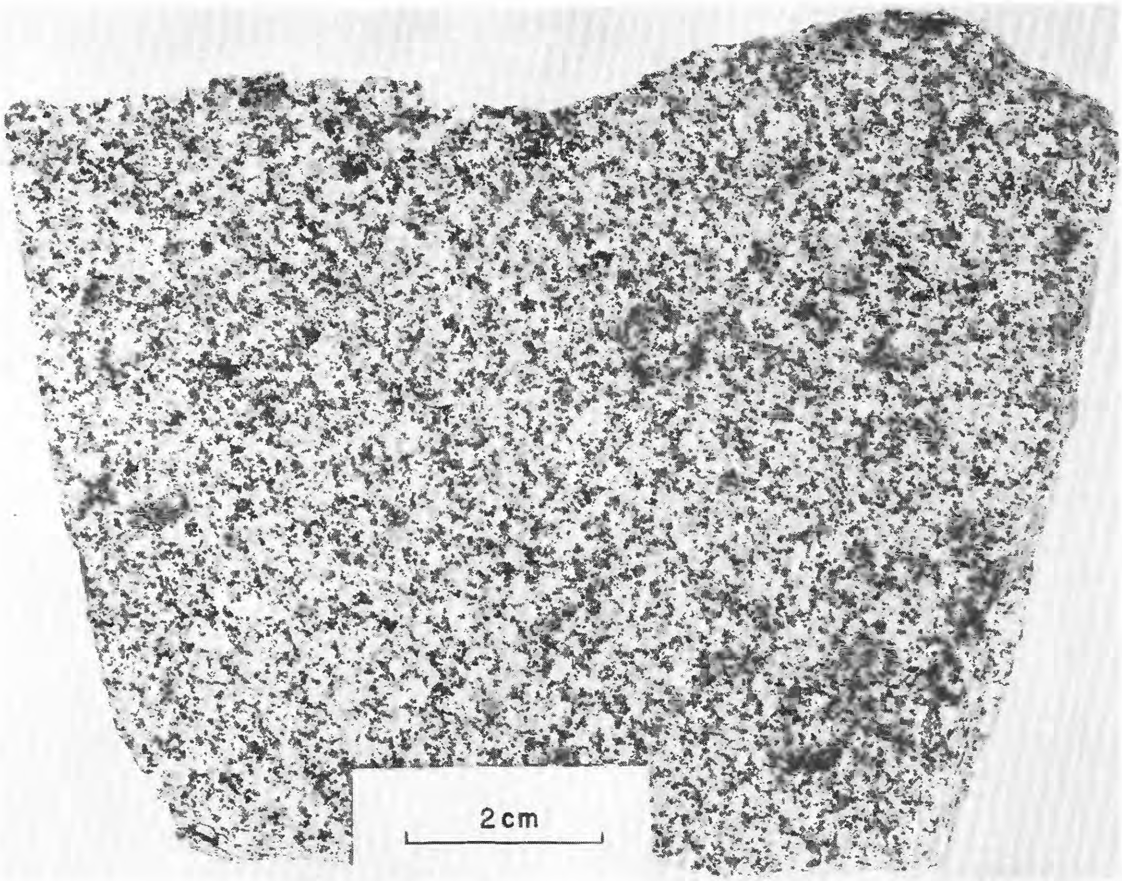
The main rock type in the outlier at Beaver Creek is light-gray or light-brownish-gray fine- to medium-grained quartz monzonite in which minerals are oriented at random (fig. 6*A*). It includes a few small lens-shaped bodies of olivine gabbro and several bodies of pyroxene gabbro, hornblende gabbro, and diorite.

Texture of the main rock type is typical of quartz monzonite; small euhedral to subhedral, strongly zoned and twinned plagioclase (An_{26-28}) crystals are included in large anhedral grains of quartz and orthoclase (fig. 6*B*). Plagioclase (An_{26-28}) grains of medium size occur with quartz between the large poikilitic quartz and orthoclase crystals. Near the northern border of the stock, along the North Fork of the Clearwater River, large anhedral zoned phenocrysts of plagioclase are embedded in the fairly fine grained quartz monzonite.

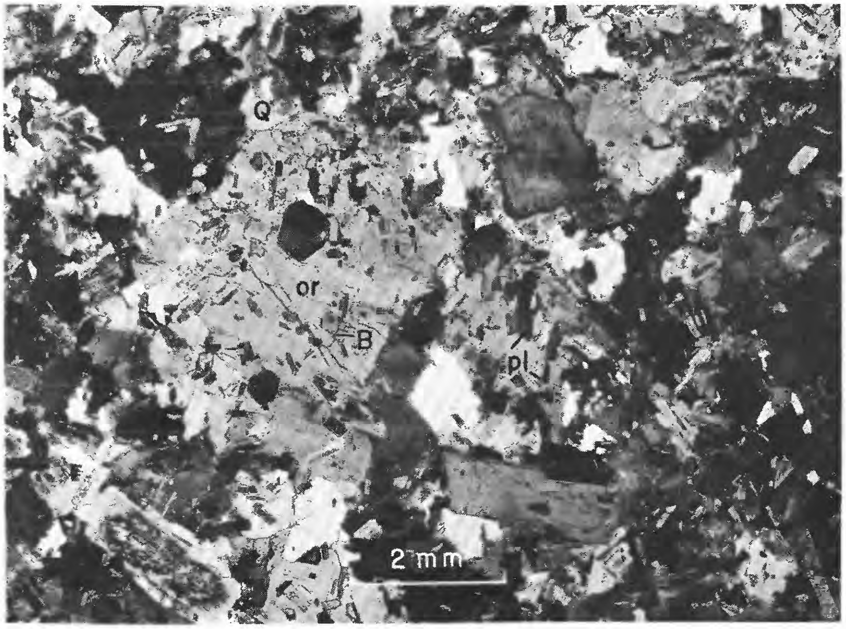
Biotite ($\gamma = 1.648 \pm 0.002$) is the main dark constituent; in addition, most of the rock has a little chlorite. Tiny muscovite laths are next to the biotite and in the centers of some plagioclase crystals. Small grains of magnetite, apatite, allanite, and zircon are the accessories. Much of the zircon is included in biotite. The pleochroic haloes around zircon crystals are darker and wider than those in other rocks of this area.

In places, quartz and orthoclase are granophyrically intergrown around subhedral orthoclase grains. The occurrence of numerous small lath-shaped zoned plagioclase crystals and the granophyric texture indicate fairly rapid cooling of the magma.

Chemical analyses (No. 1064 on table 1) shows that the quartz monzonite is rich in silicon and potassium. It is chemically close to a granite except that the amount of calcium is higher due to a higher anorthite content.



A, Fine-grained quartz monzonite along Beaver Creek (loc. 1064). Minerals are oriented at random.



B, Photomicrograph of the rock shown in *A.* Small plagioclase laths (pl) and biotite flakes (B) are included in anhedral grains of orthoclase (or) and quartz (Q). Crossed nicols.

FIGURE 6.—PHOTOGRAPH AND PHOTOMICROGRAPH OF QUARTZ MONZONITE ALONG BEAVER CREEK.

TABLE 1.—*Chemical composition in weight and ionic percentage, molecular norms, and modes of the plutonic rocks and dike rocks near Pierce and Bungalow*

[Analyses by the U.S. Geological Survey. Analysts: Lois D. Trumbull, Nos. 1063, 1064, 1073, 1072, 1079, 1081; Paula Montalto, Nos. 1121, 1125, 1143; Ruth H. Stokes, No. 758; Dorothy F. Powers, No. 1880]

	1063	1073	1072	1079	1143	1125	1121	1081	1064	1880	758	L 219
Rock type	Olivine gabbro	Pyroxene gabbro	Gabbro	Gabbroic dike	Monzotonalitic dike	Coarse-grained monzotonalitic groundmass	Monzotonalite	Quartz monzonitic dike	Quartz monzonite	Coarse-grained quartz monzonite	Granite	Granite
Locality	Beaver Creek 1.5 miles north of Camp 14	1.5 miles south of mouth of Beaver Creek	1.2 miles south of mouth of Beaver Creek	2.5 miles east of Camp 14	2 miles northeast of French Mountain	1.7 miles south of Hemlock Butte	2 miles southeast of Hemlock Butte	2.5 miles east of Camp 14	Beaver Creek 1.5 miles north of Camp 14	Lochsa River	1 mile east of Bungalow (Hietanen, 1962)	1 mile east of Bungalow (Larsen and Schmidt 1958)
Weight percent												
SiO_2	46.57	50.60	52.72	50.37	65.75	66.62	67.69	66.14	70.36	71.37	77.43	75.28
Al_2O_3	11.59	18.17	18.06	17.34	16.47	17.35	17.09	15.33	15.14	15.56	12.47	13.14
Fe_2O_3	1.28	1.60	1.65	3.04	.60	1.05	1.13	1.45	.51	.68	.22	.89
FeO	9.20	8.17	5.93	7.50	2.61	1.71	1.49	3.42	1.80	.92	.84	1.47
MnO	.15	.15	.12	.16	.05	.07	.04	.07	.04	.04	.04	.07
MgO	21.77	6.59	5.49	5.17	2.09	1.08	.85	.93	.93	.45	13	.27
CaO	5.09	8.20	8.70	7.41	4.08	3.05	3.37	2.82	2.55	1.91	.51	.89
Na_2O	1.39	3.09	3.18	3.80	4.15	5.24	4.82	3.60	3.38	4.41	3.77	3.50
K_2O	.66	.94	1.12	1.04	2.36	2.16	2.19	3.80	4.08	3.63	4.37	4.47
BaO	-----	-----	.07	.04	-----	-----	-----	.12	.09	-----	-----	.01
TiO_2	.34	1.39	1.31	1.79	.55	.33	.40	.77	.39	.25	.06	.14
P_2O_5	.10	.24	.49	.37	.15	.26	.14	.25	.14	.05	.04	.03
CO_2	.02	-----	-----	-----	.24	.01	.12	.01	-----	.02	.01	-----
H_2O+	1.63	.47	.91	1.37	.69	.67	.46	.78	.32	.32	.19	.08
H_2O-	.10	.14	.06	.32	.06	.25	.03	.25	.06	.07	.03	.07
Total	99.89	99.75	99.81	99.72	99.85	99.85	99.82	99.74	99.79	99.68	100.11	100.31
Cation percent												
SiO_2	42.04	47.20	49.36	47.75	61.31	61.91	62.94	62.80	66.10	66.59	72.55	70.54
$AlO_{3/2}$	12.33	19.97	19.92	19.37	18.10	19.00	18.72	17.15	16.76	17.11	13.77	14.51
$FeO_{3/2}$	.87	1.12	1.16	2.17	.43	.74	.79	1.04	.36	.48	.16	.63
FeO	6.94	6.37	4.64	5.94	2.03	1.33	1.16	2.74	1.42	.72	.66	1.15
MnO	.11	.12	.09	.13	.04	.06	.03	.06	.03	.04	.03	.06
MgO	29.27	9.15	7.66	7.30	2.90	1.49	1.18	1.32	1.30	.63	.18	.38
CaO	4.92	8.19	8.72	7.52	4.07	3.04	3.36	2.87	2.57	1.91	.51	.89
$NaO_{1/2}$	2.43	5.59	5.77	6.98	7.49	9.43	8.68	6.62	6.15	7.98	6.84	6.36
$KO_{1/2}$	.76	1.12	1.34	1.25	2.81	2.56	2.60	4.60	4.89	4.32	5.22	5.35
BaO	-----	-----	.03	.02	-----	-----	-----	.04	03	-----	-----	.01
TiO_2	.23	.98	.92	1.27	.39	.23	.28	.55	.28	.17	.04	.10
$PO_{5/2}$	.08	.19	.39	.30	.12	.20	.11	.20	.11	.03	.03	.02
CO_2	.02	-----	-----	-----	.31	.01	.15	.01	-----	.02	.01	-----
H_2O	(4.91)	(1.46)	(2.84)	(4.33)	(2.14)	(2.08)	(1.42)	(2.47)	(1.00)	(1.00)	(.60)	(.23)
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
O	142.51	154.19	155.02	151.79	164.17	163.60	166.23	164.67	168.59	168.45	172.97	171.89
OH	9.82	2.92	5.68	8.66	4.28	4.16	2.84	4.94	2.00	2.00	1.20	1.00
Total anions	152.33	157.11	160.70	160.45	168.45	167.76	169.07	169.61	170.59	170.45	174.17	172.89
Molecular norm												
Q	-----	-----	2.64	-----	18.93	18.26	21.34	20.95	25.85	25.03	34.72	32.48
Or	3.80	5.60	6.70	6.25	14.05	12.80	13.00	23.00	24.45	21.60	26.10	26.80
Ab	12.15	27.95	28.85	34.90	37.45	47.15	43.40	33.10	30.75	39.90	34.20	31.80
An	22.85	33.15	32.03	27.85	17.80	13.50	15.15	12.85	12.10	9.20	2.25	4.30
C	-----	-----	-----	-----	.68	1.61	1.38	.79	.88	1.13	.81	1.07
Wo	0.40	2.48	3.38	2.94	-----	-----	-----	-----	-----	-----	-----	-----
En	17.44	14.32	15.32	13.08	5.80	2.98	2.36	2.64	2.60	1.26	.36	.76
Fs	3.80	7.76	6.46	6.68	2.92	1.58	1.04	3.46	1.98	.70	1.14	1.60
Fo	30.82	2.99	-----	1.14	-----	-----	-----	-----	-----	-----	-----	-----
Fa	6.72	1.60	-----	.57	-----	-----	-----	-----	-----	-----	-----	-----
Ap	.21	.51	1.04	.80	.32	.53	.29	.53	.29	.08	.08	.05
Il	.46	1.96	1.84	2.54	.78	.46	.56	1.10	.56	.34	.08	.20
Mt	1.31	1.68	1.74	3.25	.65	1.11	1.18	1.56	.54	.72	.24	.94
Cc	.04	-----	-----	-----	.62	.02	.30	.02	-----	.04	.02	-----
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

TABLE 1.—Continued

	Measured mode [1]			Calculated molecular mode [2]								
Quartz			0.5	3.65	21.50	20.16	22.53	21.73	28.02	26.00	35.15	
Plagioclase	27.04	57.3	58.3	49.10	53.05	59.95	58.25	42.85	41.90	48.75	36.30	
(An content)	(60–65)	(47)	(45)	(29)	(29–30)	(22)	(26)	(23)	(27)	(18)	(6)	
Orthoclase					7.50	8.15	7.95	19.40	20.25	18.20	24.50	
Biotite	8.2	7.0	4.7	12.49	12.93	9.39	9.79	7.40	9.60	5.19	3.40	
Muscovite						1.80	1.35		.10	2.24	.90	
Hornblende	16.0	15.0	31.1	33.39	6.52			8.54				
Augite	19.2	19.2	3.8									
Olivine	29.0											
Allanite									.16			
Apatite			.8	.80	.33	.53	.29	.53	.30	.08	.08	
Magnetite	.2	1.5	.8	3.00		.96	.40	.78	.20	.36	.12	
Ilmenite				1.90		.26	.20	.32	.10	.14		
Rutile									.18			
Sphene							.15	.60				
Stilbite							.05					
Calcite										.04		
Total	100.0	100.0	100.0	104.33	101.83	101.20	100.96	102.15	100.81	.101.00	100.45	
Less calcium								.70				
Total								101.45				

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[1] The presence of several ferromagnesian minerals (augite, hornblende, biotite, and olivine) makes calculating the mode for the gabbros impossible.

[2] The mode was calculated from molecular norm by regrouping the oxides. In calculation the formulas for biotite and hornblende were modified from those determined from the analyses by Larsen and Schmidt (1958, table 9, nos. 2 and 3) for these minerals in tonalites of this area. The magnesium and the remaining part of iron, after forming ilmenite and magnetite, were divided between biotite and hornblende using arithmetic equations. The rest of the potassium and aluminum will determine the amount of orthoclase if no muscovite is present.

In many localities west of Beaver Creek, the quartz monzonite contains evenly scattered dark spots 1 to 2 cm in diameter. The centers of these spots are cordierite, which is altered to yellowish-green pinite and light-grayish-green chlorite along the cracks and borders (fig. 7). More chlorite borders the cordierite. Many grains of cordierite are completely altered to pinite in which the cracks are filled by chlorite. Chlorite that surrounds the grains of pinite forms radiating aggregates. Along the edges of these aggregates, brown biotite has crystallized at the expense of chlorite. The sequence of recrystallization is cordierite⟶pinite⟶chlorite⟶biotite. West of Beaver Creek, all cordierite is altered to pinite and chlorite or to fine-grained aggregates of chlorite and muscovite (fig. 8). These aggregates have irregular outlines and contain relict cracks filled by chlorite just as do the cordierite and pinite. The presence of every step of alteration of cordierite indicates that the cordierite is a relict mineral and not in equilibrium in this quartz monzonite.

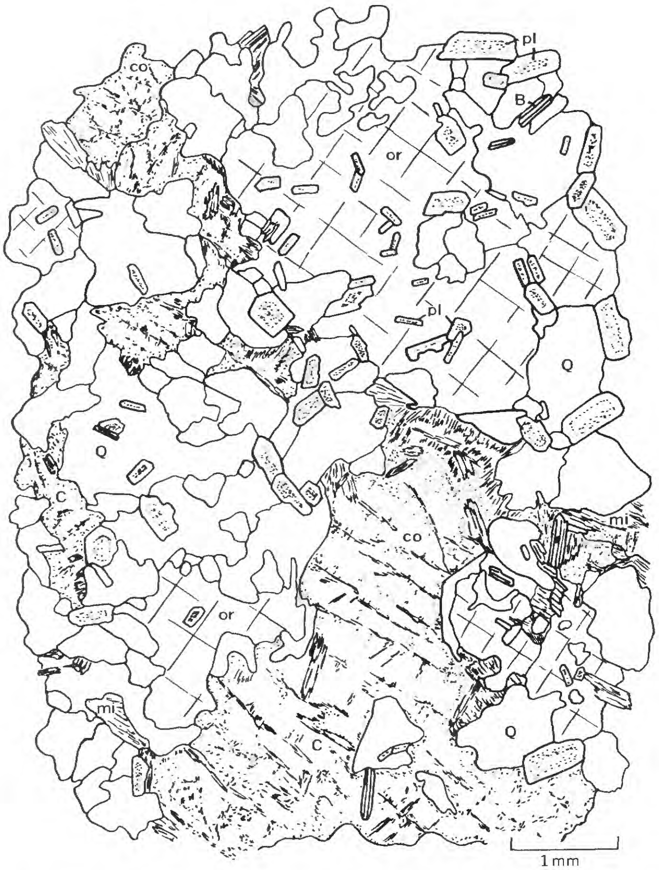


FIGURE 7.—Camera lucida drawing of partly altered cordierite (co) in the quartz monzonite along Beaver Creek (loc. 578). Orthoclase (or); quartz (Q); plagioclase (pl); muscovite (mi); biotite (B).

As cordierite is abundant in the metasedimentary country rocks to the west, it is quite probable that the cordierite spots represent remnants of inclusions of these rocks; however, cordierite is not confined to certain zones which might represent relict bedding. Rather, the distribution is irregular and, together with the hypidiomorphic texture of the quartz monzonite, indicates that the magma that picked up the fragments of cordierite-bearing country rock was moving and hot enough to melt or digest all other minerals of the fragments except the cordierite. The small lath-shaped strongly zoned plagioclase crystals were floating in this magma. During the final cooling period, the large anhedral crystals of quartz and feldspars crystallized and enclosed the preexisting small laths of plagioclase. During this phase, the cordierite altered to pinite and chlorite because of lower temperature and the presence of more water (in the residual solution). The temperature was still high enough for the crystallization of biotite at the expense of chlorite derived from cordierite and of potassium present in the magma surrounding the inclusions.

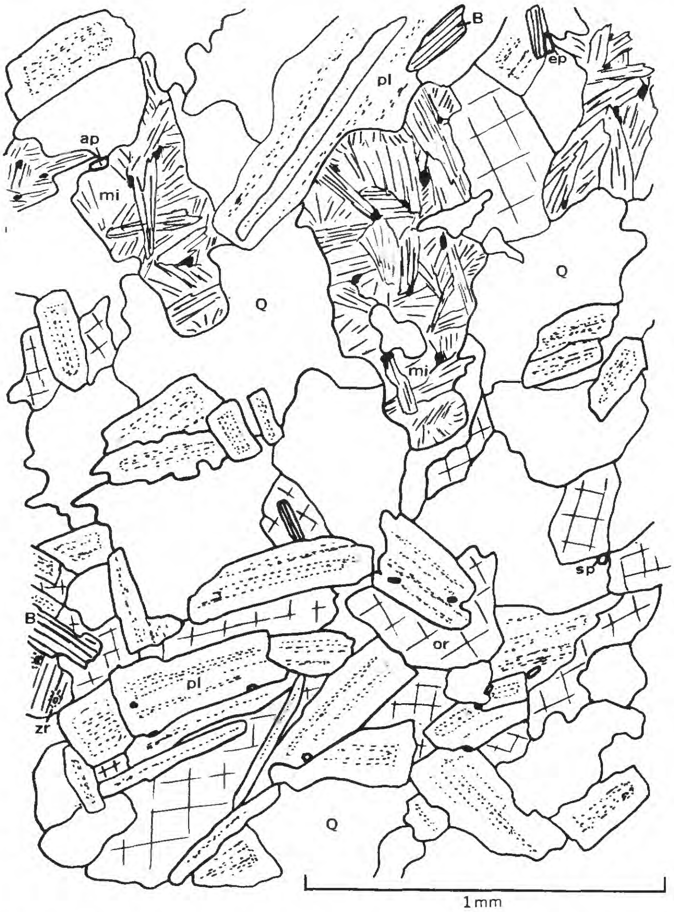


FIGURE 8.—Camera lucida drawing showing aggregates of muscovite and chlorite (mi), originally cordierite, in fine-grained quartz monzonite along Beaver Creek (loc. 1065). Quartz (Q); sphene (sp); epidote (ep); plagioclase (pl); orthoclase (or); biotite (B); apatite (ap); black, magnetite; zircon (zr).

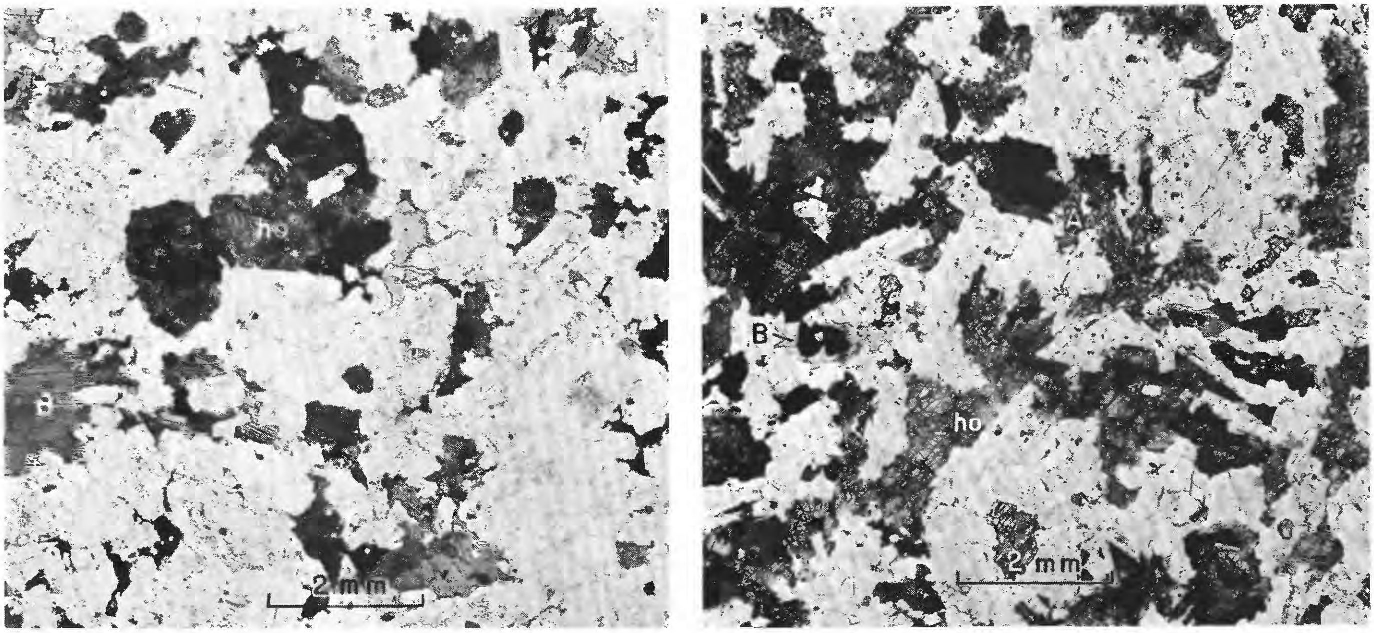
Dioritic rocks along the northern border zone of the stock are medium-grained dark-gray hornblende-biotite-plagioclase rocks with interstitial quartz. Plagioclase is subhedral, twinned, and strongly zoned, with centers An_{40-42}, and borders An_{27}. Sphene, magnetite, apatite, and zircon are the accessories. Small bodies of similar diorite crop out along the contacts of the stock (fig. 9*A*, *B*).

Road cuts in dioritic rocks along Beaver Creek expose a number of dark- to brownish-gray inclusions of coarse-grained olivine gabbro and pyroxene-hornblende gabbro. The inclusions of olivine gabbro range from 30 cm to 1 m in length and from 20 to 50 cm in thickness. They consist of about 30 percent plagioclase, 20 percent enstatite-hypersthene, 30 percent olivine, 14 percent hornblende, 6 percent biotite and a little pyrrhotite. Plagioclase (An_{60-65}) occurs as subhedral lath-shaped crystals which range from 1 to 3 mm in length, show complex twinning, and are oriented at random. Enstatite-hypersthene, brown hornblende, and biotite form large anhedral crystals, 3 to 5 mm in diameter and include round olivine crystals that range from 0.5 to 2 mm in diameter (fig. 10).

The olivine crystals are chrysolite with $\alpha=1.678\pm0.001$, $\beta=1.696\pm0.001$, $\gamma=1.717\pm0.002$. Enstatite-hypersthene has $\alpha=1.673\pm0.001$,$\gamma=1.691\pm0.001$; and $-2V$, large; it contains about 20 percent of the iron end member. Most of the hornblende is brown and shows $\gamma=1.677\pm0.002$; and $\gamma\wedge c=20°$; but parts of some grains are grayish green. Biotite is strongly pleochroic with γ=reddish brown and α=pale yellow. The optical properties, $\gamma=1.601\pm0.001$ and $-2V$=small, indicate that it is eastonite rich in magnesium. A chemical analysis of the olivine gabbro is given in table 1, No. 1063.

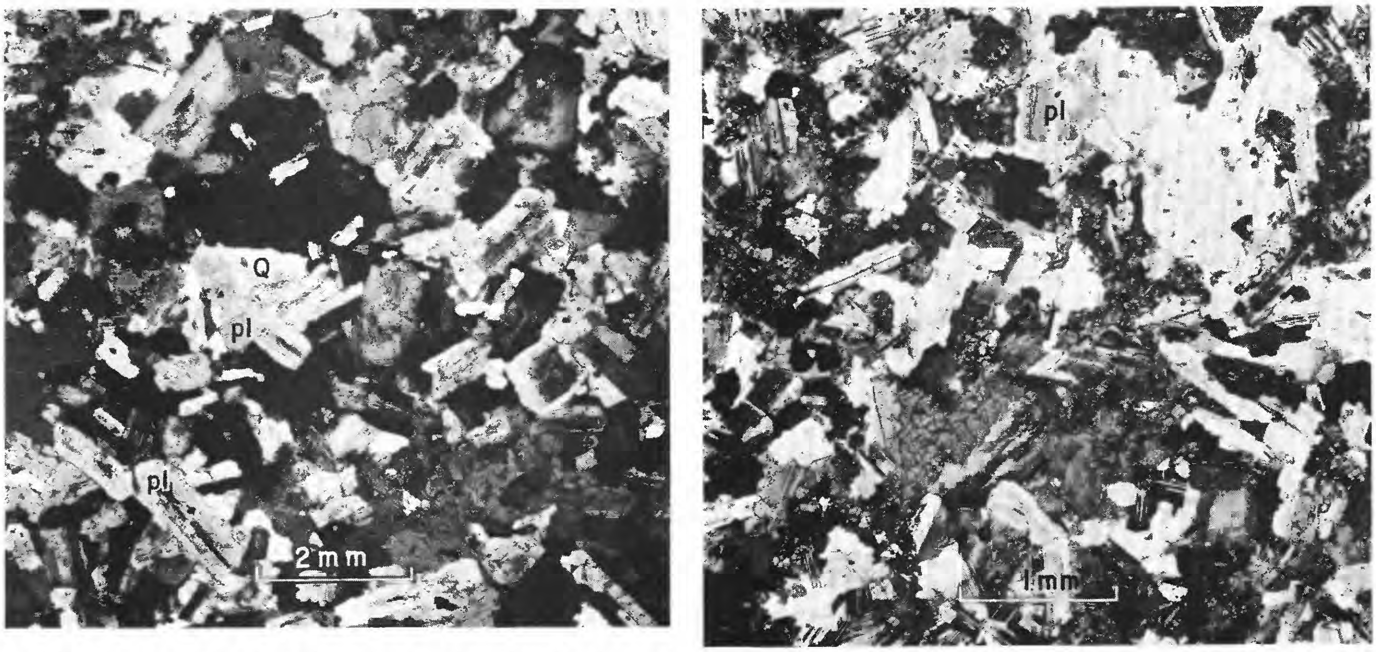
Loose blocks of fine- to medium-grained dark pyroxene gabbro occur for 1 mile along Beaver Creek in sec. 7, T. 40 N., R. 7 E. Light-gray quartz monzonite is exposed to the north and to the south of the gabbro. The minerals in the gabbro are plagioclase (An_{47}), hypersthene, clinohypersthene, hornblende, biotite, magnetite, and apatite. The texture of the gabbro is hypidiomorphic (fig. 9*C*, *D*). Plagioclase occurs in stubby laths, which range from 1 to 2 mm in length and are oriented at random. Dark constituents form anhedral grains that have irregular borders. Most pyroxene grains are surrounded by hornblende, and spots of hornblende are common in pyroxene. Clinohypersthene contains abundant small ilmenite lamellae oriented parallel to the cleavages. Biotite flakes, 0.5 to 1 mm in diameter, are clustered with other dark constituents and include rounded magnetite grains.

Hypersthene is pleochronic with α=reddish tan, β=pale tan, γ=light bluish green, and $-2V$=large. The indices of refraction, $\alpha=1.696\pm0.002$, $\gamma=1.771\pm0.002$, indicate that it contains about 37 percent iron end member (Winchell and Winchell, 1951, p. 406). Clinohypersthene shows $\gamma\wedge c=31°$, $+2V$ about 65°, and $\alpha=1.689\pm0.002$, $\gamma=1.709\pm0.002$—values which indicate according to Winchell's curves about 32 percent iron end member. Hornblende is brownish gren and the indices of refraction $\alpha=1.668\pm0.002$, $\gamma=1.690\pm0.002$, and $\gamma\wedge c=17°$ indicate it to be richer in iron than the hornblende in the olivine gabbro. Biotite is dark brown, with $\gamma=1.648\pm0.002$. Some of the gabbro is coarser grained and lighter in color (spec. 1072) than the common variety (spec. 1073). The lighter colored variety has a few small quartz grains between the plagioclase laths, and most of the dark minerals are green hornblende. All the pyroxene is surrounded by hornblende and shows spotty alteration to this mineral. Tiny inclusions of magnetite are common in hornblende. Chemical analyses of a common gabbro (spec. 1073)



A, Quartz diorite along Beaver Creek (loc. 574). Dark minerals are hornblende (ho) and biotite (Bi). Plane-polarized light.

C, Gabbro along Beaver Creek (loc. 1073). The main dark constituents are pyroxene (A) and hornblende (ho). Biotite (Bi) and magnetite (black) occur in small quantities. Plane-polarized light.



B, The field of figure *A* with crossed nicols. Plagioclase (pl) occurs in lath-shaped crystals that are twinned and zoned. Quartz (Q) is interstitial.

D, The field of figure *C* with crossed nicols. Plagioclase (pl) is subhedral, shows complex twinning, and is zoned.

FIGURE 9*A–D*.—PHOTOMICROGRAPHS OF QUARTZ DIORITE AND GABBRO OF THE QUARTZ MONZONITE SERIES.

and of a light-colored variety (spec. 1072) are reported on table 1. The light-colored variety is slightly more silicic and contains less iron and magnesium than the common variety.

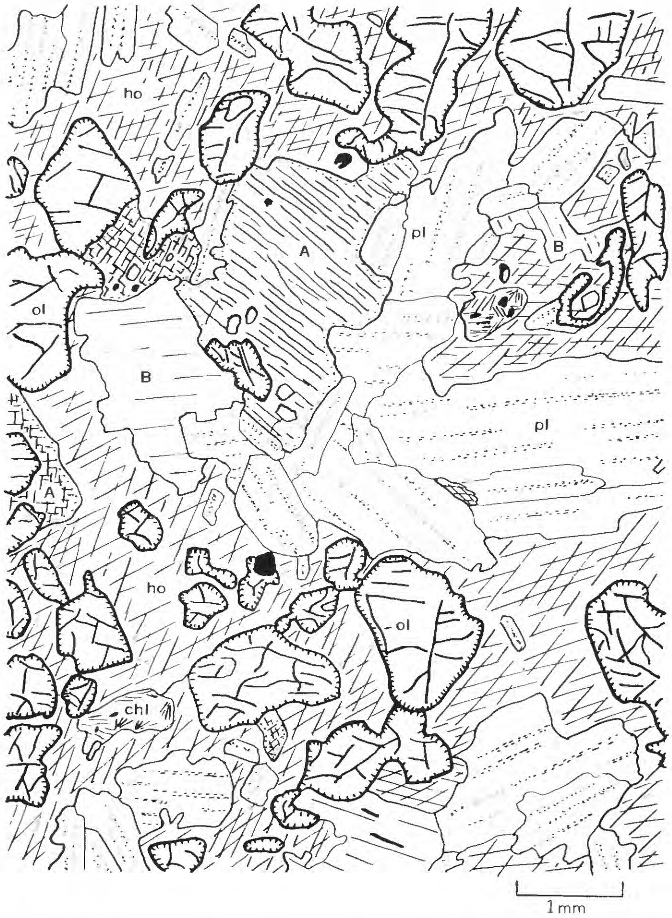


FIGURE 10.—Camera lucida drawing of olivine gabbro. From an inclusion in quartz monzonite along Beaver Creek (loc. 1063). Olivine (ol); augite (A); hornblende (ho); plagioclase (pl); chlorite (chl); biotite (B).

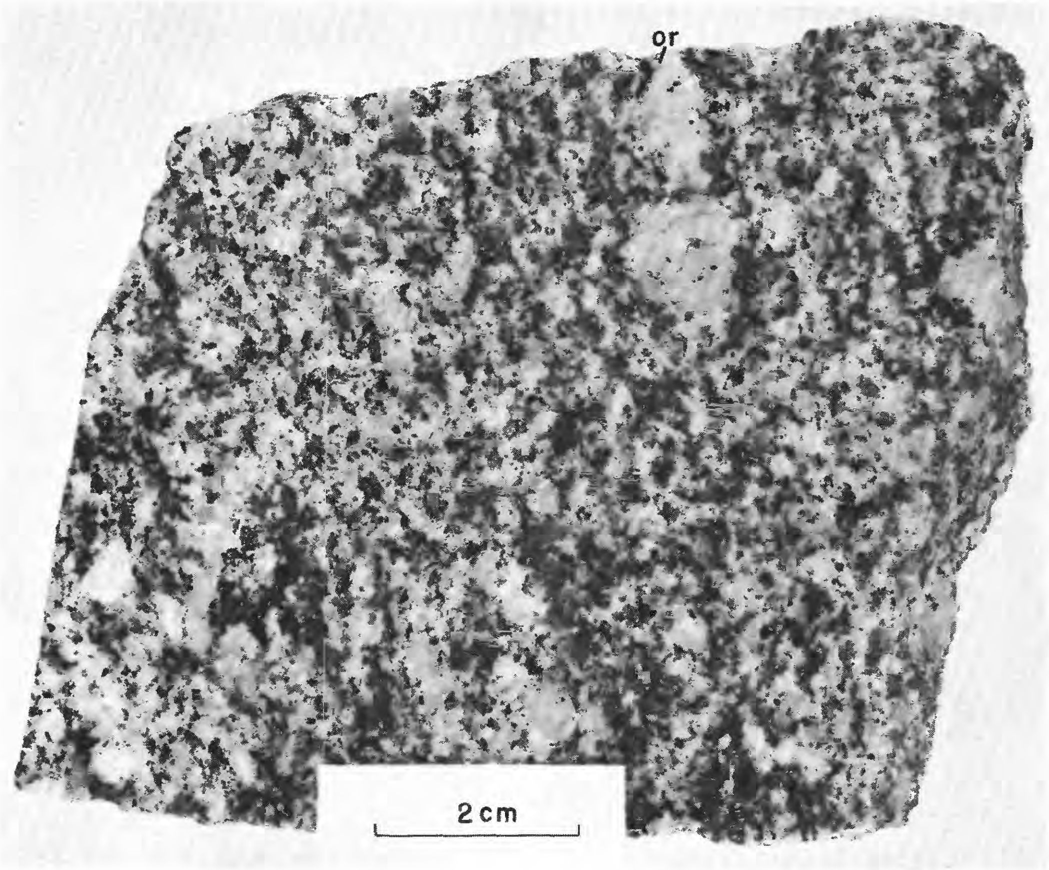
COARSE-GRAINED QUARTZ MONZONITE AND MONZOTONALITE

Most of the quartz monzonite in the southern part of the Pierce area is light gray, coarse to medium grained and contains orthoclase crystals that are larger in size than the other mineral grains (fig. 11*A*). The size, shape, and number of the large orthoclase crystals vary, so that the rock is rather inhomogenous. This inhomogeneity is increased by variation in the amount of interstitial orthoclase in the groundmass. At most places, the orthoclase crystals are anhedral to subhedral and range from 0.5 to 2 cm in diameter, but in a few localities euhedral phenocrysts as much as 5 cm in diameter abound. These large phenocrysts show well-developed (001), (010), (110), and (201) faces. They are more resistant to weathering than the rest of the rock, and can be picked by hand from decomposed outcrops, as for instance along Musselshell Creek (loc. 830), 2 miles south of Hemlock Butte (loc. 1126), and along Swede Creek (loc. 1976) near the western border of the mass. The average grain size of the groundmass ranges from 0.5 to 3 mm. In a few outcrops, however, the groundmass is coarse grained; many plagioclase grains range from 0.5 to 2 cm in diameter. The amount of orthoclase in the groundmass is about 10 to 25 percent. The amount of quartz is generally less than that of plagioclase (An_{26-28}) and the amount of biotite is 3 to 6 percent.

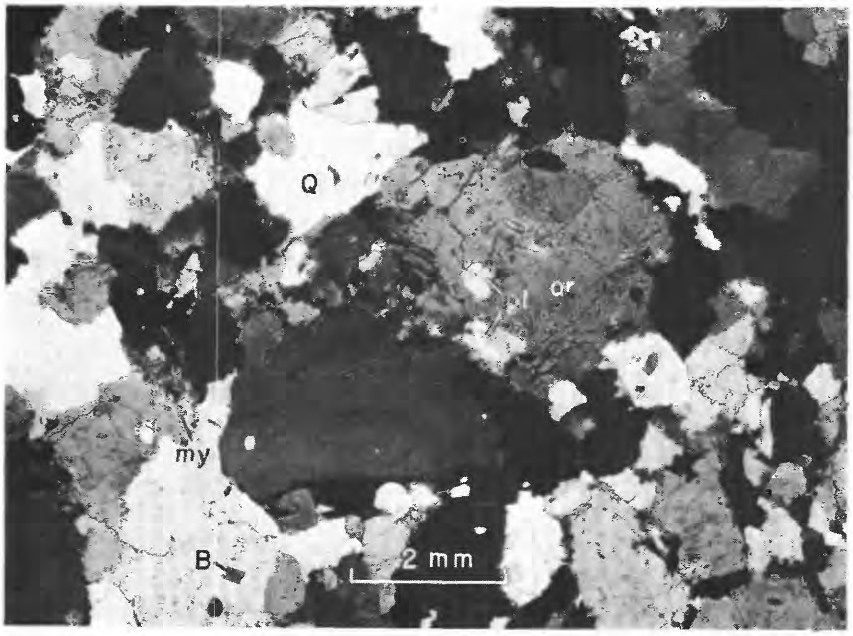
The texture of the quartz monzonite with orthoclase phenocrysts is very different from the texture of the quartz monzonite along Beaver Creek. The texture of the groundmass (fig. 11*B*) is similar to that found in the massive tonalite. Some of the plagioclase grains are subhedral and zoned, whereas others are anhedral and form granoblastic texture with irregularly shaped or rounded grains of quartz and flakes of biotite (fig. 12). The subhedral grains of plagioclase tend to be larger than the anhedral ones; they range from 3 to 4 mm in diameter, whereas the anhedral grains plagioclase are about 1 to 3 mm in diameter, as are the quartz grains. A few small grains of quartz are interstitial. Rarely, myrmekite grains appear, with rounded borders against orthoclase. Some rodlike inclusions of quartz in the myrmekite extend over the border into the orthoclase; some of the myrmekite was probably replaced by orthoclase in the latest episode of crystallization.

Biotite is the only dark constituent. It occurs in flakes, 1 to 2 mm long, between the quartz and plagioclase grains and has a subparallel orientation giving weak gneissic structure to some outcrops. The index of refraction $\gamma=1.649\pm0.002$ indicates that this biotite is similar to that in the quartz monzonite along Beaver Creek. These biotites are richer in iron than the biotite in the quartz diorite. A small amount of muscovite occurs with biotite and magnetite; apatite and zircon are the accessory minerals.

The orthoclase phenocrysts in a fine- to medium-grained light-gray groundmass are perthitic and contain only a few small inclusions of biotite, quartz, plagioclase (An_{28}), and zircon. The indices of refraction are $\alpha=1.520\pm0.001$, $\beta=1.523\pm0.001$, $\gamma=1.527\pm0.001$, and X-ray study shows a good orthoclase pattern. Some orthoclase phenocrysts in a fairly fine grained rock about 2 miles south of Hemlock Butte were separated by means of heavy liquids and analyzed chemically (table 2). They contain 16.77 molecular percent albite, 5.85 molecular percent barium feldspar, and only a small amount of anorthite. The high content of barium shows that orthoclase is a chief carrier of barium in the porphyritic quartz monzonite. This high content of barium makes it difficult to compare the com-



A, Medium-grained light-gray quartz monzonite with a few larger grains of orthoclase (or). Note crude parallel orientation of biotite.



B, Photomicrograph of the rock shown in figure *A.* Orthoclase (or) includes a few small plagioclase (pl) grains; some myrmekite (my) occurs between the orthoclase and bordering plagioclase. Quartz (Q); biotite (B).

FIGURE 11.—PHOTOGRAPH AND PHOTOMICROGRAPH OF QUARTZ MONZONITE ALONG MUSSELSHELL CREEK (loc. 831).

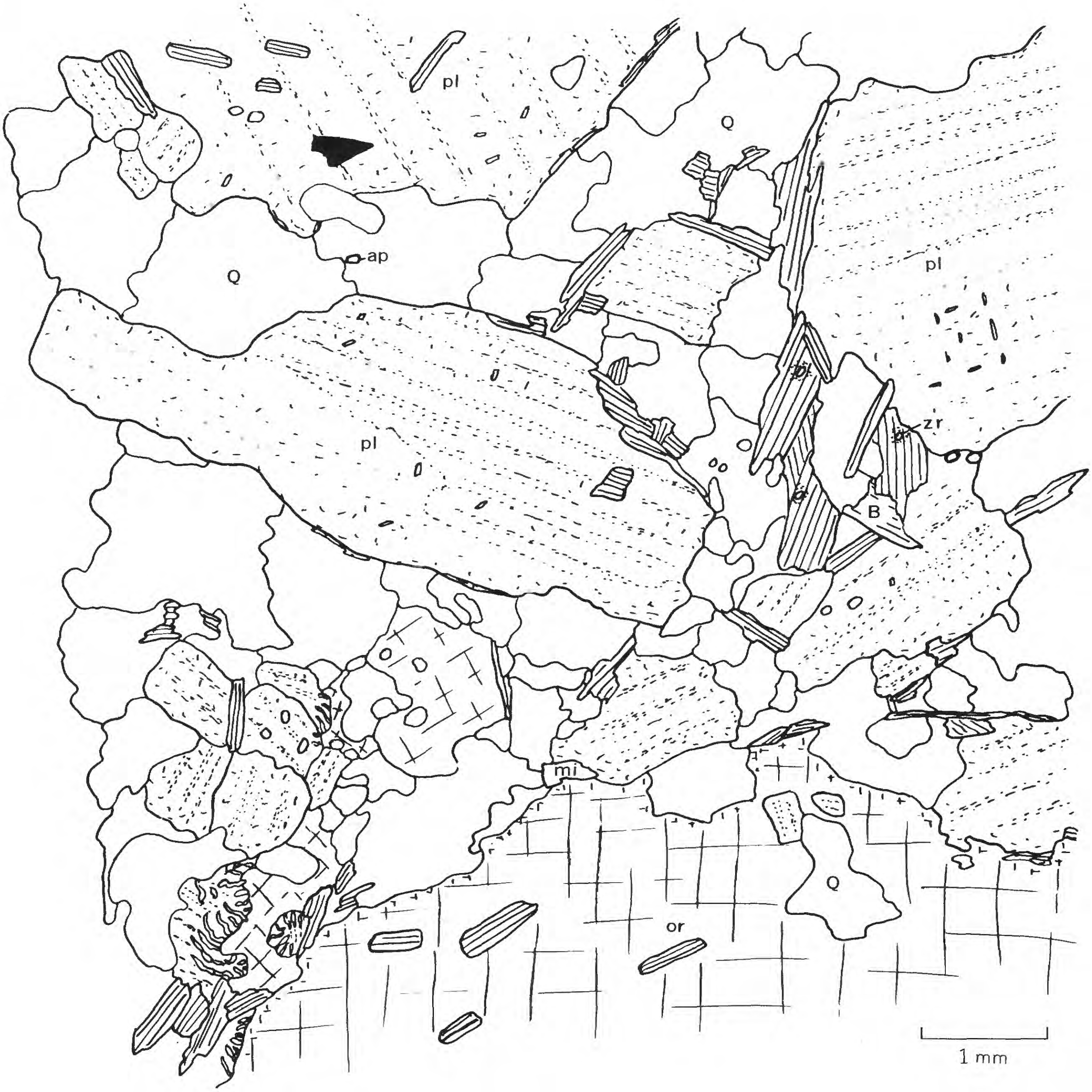


FIGURE 12.—Camera lucida drawing of coarse-grained quartz monzonite south of Hemlock Butte (loc. 1125). Quartz (Q); plagioclase (pl); orthoclase (or); biotite (B); apatite (ap); zircon (zr); black, magnetite; muscovite (mi).

position of these phenocrysts with the compositions used in the experimental work by Tuttle and Bowen (1958). The orthoclase-albite ratio of this analysis, together with the general amount of perthite as determined under the microscope, indicates that the quartz monzonite in the northern part of the Idaho batholith is type B of the subsolvus granites in the classification of Tuttle and Bowen.

Some specimens have scattered small interstitial groups of radiating crystals with low indices of refraction, negative elongation, and small extinction angle, probably stilbite.

The "groundmass" of the coarse-grained quartz monzonite (table 1, No. 1125) has a chemical composition between that of quartz monzonite and the tonalites analyzed earlier from the Dent area, about 20 miles west-northwest of Pierce (Hietanen, 1962, table 10). The total composition of this groundmass plus the phenocrysts has the composition of quartz monzonite. The origin of the phenocrysts and the sequence of crystallization are discussed in a later section.

The northern border zone south of Hemlock Butte consists of medium-grained, light-gray rock with only a small amount of interstitial orthoclase (fig. 13).

TABLE 2.—*Chemical composition of orthoclase phenocrysts from quartz monzonite 1.5 miles south of Hemlock Butte*

[Analyst: Dorothy F. Powers, U.S. Geol. Survey]

	Weight percent	Inclusions			Feldspar after subtraction of inclusions		Molecular proportions	Minerals (molecular percent)	
		Biotite	Kaolinite	Quartz	a	b = a, calculated to 100			
SiO_2	63. 77	0. 16	0. 74	0. 82	62. 05	64. 02	10654	Orthoclase	76. 81
TiO_2	. 01	. 01	------	------	------	------	------	Albite	16. 77
Al_2O_3	19. 04	. 07	. 63	------	18. 34	18. 92	1856	Celsian	5. 83
Fe_2O_3	. 01	. 01	------	------	------	------	------	Anorthite	. 59
FeO	. 05	. 05	------	------	------	------	------		
MnO	. 00	------	------	------	------	------	------		
MgO	. 03	. 03	------	------	------	------	------		
CaO	. 06	------	------	------	. 06	. 06	11		
BaO	1. 61	------	------	------	1. 61	1. 66	108		
Na_2O	1. 87	------	------	------	1. 87	1. 93	311		
K_2O	13. 03	. 03	------	------	13. 00	13. 41	1424		
P_2O_5	. 01	. 01	------	------	------	------	------		
H_2O+	. 10	. 01	. 11	------	------	------	------		
H_2O-	. 02	------	------	------	------	------	------		
Total	99. 61	. 38	1. 48	. 82	96. 93	100. 00	------		100. 00

FIGURE 13.—Camera lucida drawing of the potassium-poor border zone of quartz monzonite south of Hemlock Butte (loc. 1121). Quartz (Q); plagioclase (pl); biotite (B); muscovite (mi); orthoclase (or); zircon (zr); sphene (sp); epidote (ep); apatite (ap); black, magnetite.

Chemical analysis of this rock (table 1, No. 1121) shows that it contains more calcium and sodium and less silicon and potassium than the main part of the quartz monzonite. In its mineral content and composition this border facies is between quartz monzonite and the intrusive tonalite and is therefore called monzotonalite. The transition from the potassium-poor border zone into normal quartz monzonite takes place over about a hundred meters. In this transitional zone, large orthoclase crystals are embedded in a fine-grained groundmass, the composition of which is similar to that of the border facies. Southward, more interstitial orthoclase appears and phenocrysts there are smaller and less euhedral.

The quartz monzonite in the southern part of the Bungalow area is light gray, coarse grained, and slightly gneissic. The grains of orthoclase are somewhat larger than those of plagioclase, quartz, and biotite. The amount of orthoclase in specimens from south of Larch Butte is about 10 percent. Many outcrops elsewhere contain large subhedral orthoclase crystals, which together with the monzotonalitic groundmass give the average composition of common quartz monzonite. Most of the rocks mapped as quartz monzonite in the easternmost part of the Bungalow area are very coarse grained and grade to a pegmatitic quartz-feldspar rock with some muscovite.

GRANITE

The granite near Bungalow Ranger Station is a very light gray coarse-grained homogeneous rock in which the major constituents—quarts, feldspars, and biotite—are easily recognized (fig. 14*A*). The grains of quartz and orthoclase are 2 to 3 mm in diameter, whereas those of plagioclase are 0.5 to 2 mm long. Orthoclase is twinned according to the Carlsbad law, and contains perthite and tiny dustlike inclusions of an opaque mineral, probably hematite. Biotite, partly altered to chlorite, is in small dark-brown flakes or groups of

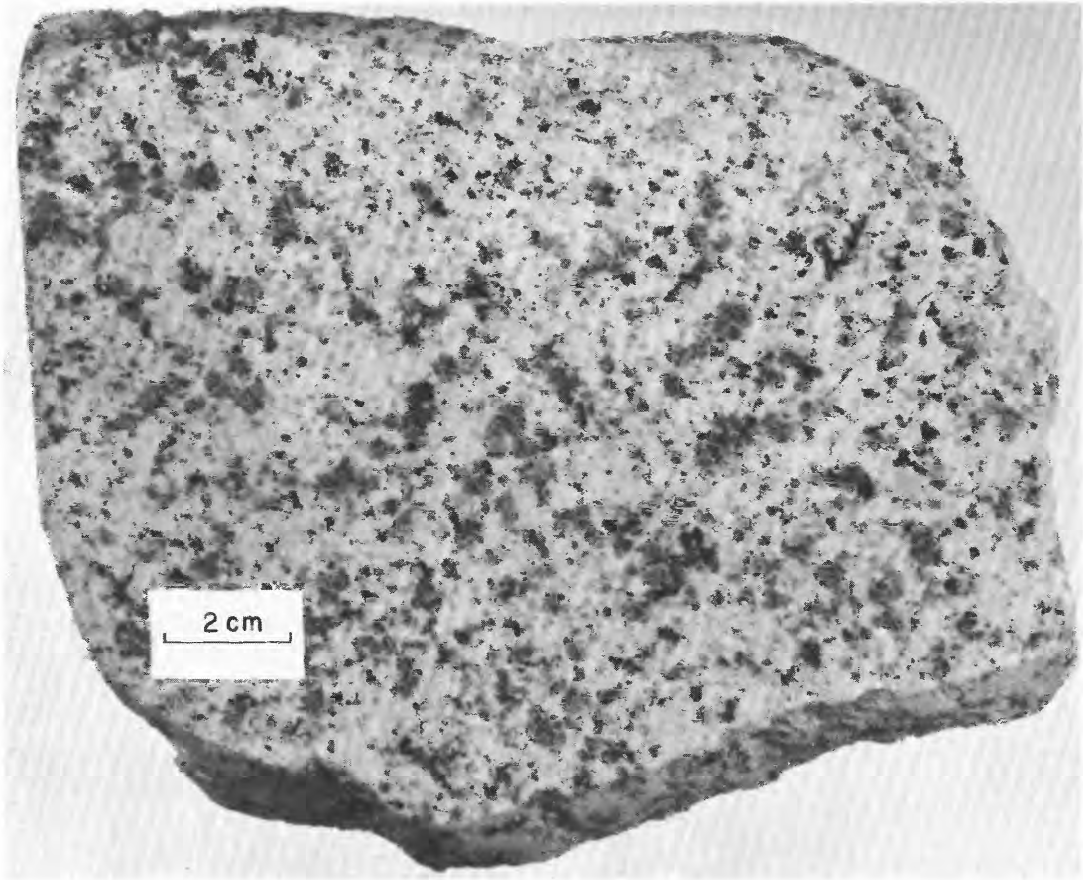


FIGURE 14*A*, Photograph of granite near Bungalow (loc. 758).
Coarse-grained granite in which quartz grains (gray) and biotite flakes (black) are easily seen.

flakes. A few grains of allanite and abundant tiny crystals of zircon with strongly pleochroic halos are included in biotite. Some muscovite occurs next to the biotite.

Two chemical analyses of this granite have been published and are shown in table 1 as Nos. 758 (Hietanen, 1962) and L 219 (Larsen and Schmidt, 1958) for comparison with the analyses of quartz monzonite. Both analyses of granite are much alike and show that the amounts of potassium and sodium are larger and those of calcium and magnesium smaller than the amounts of these elements in the quartz monzonites.

The outcrops just east of the Bungalow area between Pot Mountain and Bar Point are porphyritic coarse-grained granite with quartz and feldspar phenocrysts, 1 to 2 cm in diameter, embedded in a medium-grained groundmass of quartz, feldspar, and biotite.

The granite northeast of the mouth of Governor Creek and west of Deadhorse Lookout is coarse grained and porphyritic. It contains abundant euhedral to subhedral phenocrysts of orthoclase, 1 to 2 cm in diameter, and subhedral grains of quartz, about 1 cm in diameter. The phenocrysts of orthoclase contain about 20 percent albite as perthitic stringers and are surrounded by narrow lacy rims of granophyrically intergrown orthoclase and quartz (fig. 3*A*). All orthoclase in the rims is perthitic and has the same optical orientation as the host orthoclase. The groundmass is medium grained and consists of quartz, albitic plagioclase, orthoclase, and small flakes and needles of biotite. Some quartz occurs in small subhedral or round crystals included in plagioclase and orthoclase. Tiny prisms or apatite and small crystals of zircon with pleochroic halos are the accessories.

The pink medium-grained granite west of Deadhorse Mountain consists of about 60 percent orthoclase, 10 percent plagioclase, and 30 percent quartz, with a little biotite and magnetite. The orthoclase crystals are 1 to 2 mm long, subhedral, and include only about 1 percent perthitic stringers of albite. Many grains have Carlsbad twinning and all are clouded by tiny dustlike inclusions, similar to those in the coarse-grained granite and in the porphyritic variety.

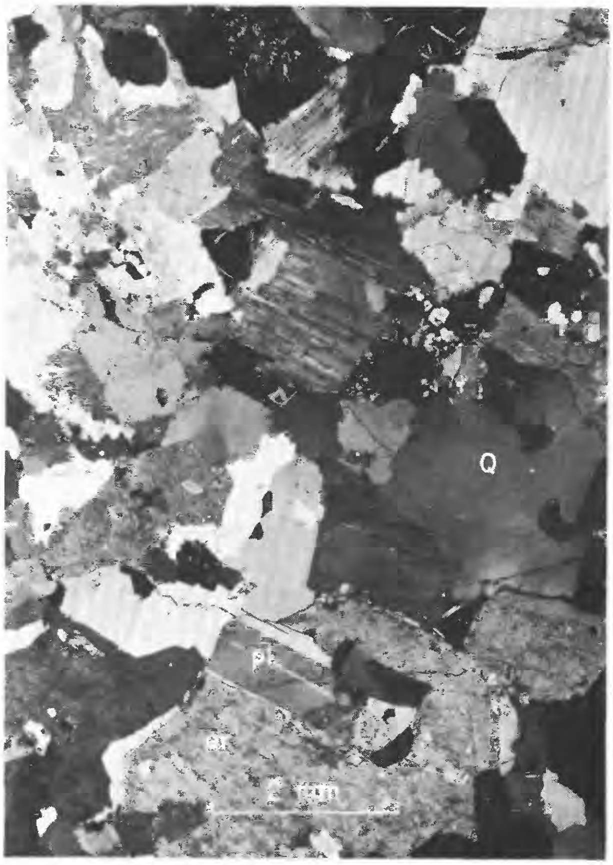


FIGURE 14*B*, Photomicrograph of the specimen of figure 14*A*. Orthoclase (or) is perthitic, includes a few grains of plagioclase (pl), and is clouded by tiny inclusions of hematite. Quartz (Q) is subhedral. Crossed nicols.

PEGMATITES

Two petrologically different types of pegmatite, plagioclase pegmatite and granite pegmatite, are common in the area. The major difference between them is in the amount of potassium-bearing minerals; the plagioclase pegmatite contains little or no potassium feldspar and muscovite, whereas the granite pegmatite is rich in these minerals.

PLAGIOCLASE PEGMATITE

Plagioclase pegmatite is common in the metamorphic rocks, in the quartz diorite, and in its tonalitic border facies, but is lacking in the quartz monzonite and granite. In the metamorphic rocks, pegmatite occurs in veins a few millimeters to a meter wide. The narrow veins are parallel to the bedding or more rarely parallel to the foliation, whereas the wide ones follow fractures and joints.

The mineralogy is fairly simple; either plagioclase (An_{25}) and quartz occur in equal amounts or the amount of plagioclase is larger (as much as 60 percent). Biotite is the only dark constituent in most of the pegmatite in the metamorphic rocks, whereas many dikes transecting the quartz diorite have hornblende also. Small amounts of muscovite, chlorite, sphene, zircon, and apatite are common.

Plagioclase pegmatite is especially common along the contact zones of the quartz diorite and its border facies, the gneissic tonalite. For instance, along Orofino Creek southeast of Pierce, much of the rocks mapped as tonalite consists of plagioclase pegmatite and there is every gradation from an equigranular homogeneous tonalite to an inhomogeneous coarse-grained plagioclase-quartz rock in which biotite occurs either as schlieren or as large plates, 0.5 to 2 cm in diameter.

The mode of occurrence of the plagioclase pegmatite indicates that it is a result of accumulation of quartzofeldspathic material along the contact zones, shear planes, joints, and other structural planes. Their mineralogy indicates that they are genetically connected with the formation of gneissic tonalite, which is considered to be a result of feldspathization of the country rocks near and along the contacts of quartz diorite. The processes leading to this feldspathization have been discussed earlier (Hietanen, 1962). Metamorphic differentiation in biotite gneiss can alone produce veinlets of plagioclase pegmatitic composition with little or no biotite.

GRANITE PEGMATITE

Granite pegmatite is more common near and in quartz monzonite and granite than in other settings. Most of these pegmatites are discordant, but concordant masses of varied thicknesses occur in the metamorphic rocks next to the coarse-grained quartz monzonite in the southeastern part of the Bungalow area. Large pegmatitic masses occur along many contacts of the coarse-grained quartz monzonite; this mode of occurrence is very similar to that of the plagioclase pegmatite along the contacts of quartz diorite. Every gradation from the pegmatite to a coarse-grained quartz monzonite occurs along Weitas Creek and in the area west of it. Pegmatite also occupies the contact between the quartz monzonite and the quartz diorite along Weaver Creek.

Pegmatite dikes, 10 to 20 cm wide, and masses up to 50 cm in diameter are common in the northeastern part of the granite pluton northeast of Bungalow. The metamorphic rocks near the quartz monzonite bodies have been migmatized and contain abundant veinlets, 1 to 10 mm thick, of granitic-pegmatitic composition. In addition to these rather fine grained narrow veinlets that parallel the bedding, there are veins and dikes, 20 cm to 1 mm thick, parallel to the bedding and the joint systems.

The minerals in fine- to medium-grained veinlets are quartz, plagioclase (An_{20}), orthoclase, and a few small flakes of biotite and muscovite. The granitic veins in the migmatized biotite gneiss are mineralogically similar to the light-colored layers of the thinly laminated biotite gneisses of the Wallace formation, except that the borders of the veins tend to be more irregular than the bedding planes of the laminated rocks. However, there is every gradation from a laminated rock to a migmatite, and it is often impossible to tell on the basis of field study whether the veins were formed by recrystallization of sedimentary material, or by introduction of granitic material, or perhaps by introduction only of potassium. Study of textures and mineralogy is helpful but not conclusive.

Since the composition and crystal structure of potassium feldspar depends on the temperature of crystallization, it might be possible in some cases to distinguish between the metamorphic and igneous feldspar either by determining the amount of albite in orthoclase (Tuttle and Bowen, 1958, p. 128–129) or by determining the triclinicity, that is the "degree of order" of Al and Si atoms in the potassium feldspar as suggested by Goldsmith and Laves (1954). The disordered (monoclinic) form crystallizes at high temperatures and the ordered (triclinic) form at low temperatures. In the intermediate types the triclinicity is a function of temperature of crystallization.

The X-ray powder diffraction data shows that the potassium feldspar in quartz monzonite in the area studied is orthoclase, but in the quartzofeldspathic laminae of the metasedimentary rocks it is microcline. The triclinicity should therefore offer means to distinguish between the igneous and metamorphic potassium feldspar in the veins. Several samples were examined; some appeared to contain orthoclase but many showed triclinicity of 60 to 70 percent. These intermediate types were further examined under the microscope and compared with those that contain only orthoclase or only microcline. Results of this study are as folows:

All veins near the plutonic rocks contain orthoclase. Most of these veins are medium- to coarse-grained pegmatite, but some resemble the quartzofeldspathic laminae. The latter are fine to medium grained and have granoblastic texture, as do the laminae; they can be distinguished from the laminae, however, because of their reddish color and more irregular shape. Many intermediate types also occur and it seems possible that heating of the sedimentary material to magmatic temperatures changed the triclinic metamorphic feldspar to monoclinic form.

All potassium feldspar farther from the immediate contact zone of plutonic rocks tends to be triclinic. For instance, large orthoclase crystals with Carlsbad twinning in schist next to a granite pegmatite show triclinicity of 70 percent. Components to form this orthoclase were introduced from the nearby pegmatite, which in turn represents the last crystallizing part of quartz monzonite magma. The high degree of triclinicity apparently is due to the low temperature of crystallization. This raises the question: Did all potassium feldspar with trinclinicity of 60 to 70 percent come from igneous sources and thus are most of the veins of igneous origin?

The presence or absence of perthitic lamellae offers additional evidence for the origin of potassium feldspar. All orthoclase in the quartz monzonite is perthitic, contains tiny inclusions of sericite, and is clouded by dust-like inclusions of opaque minerals. The orthoclase in the pegmatitic dikes and in the large veins consists of similar perthite, but the microcline in the gneissic layers of the Wallace formation farther from the batholith is nonperthitic and free of inclusions.

X-ray study of some of this nonperthitic clear potassium feldspar shows a good orthoclase pattern, and it seems possible that this orthoclase crystallized from the sedimentary material in magmatic temperatures. On the other hand, the perthitic clouded orthoclase next to the igneous bodies with similar orthoclase probably crystallized from material introduced from igneous sources.

HYPABYSSAL ROCKS

Hypabyssal dikes and sills transect all rocks of the other types except the Columbia River basalt. They are especially common along fault zones and along the contacts between the plutonic and metamorphic rocks or between two types of plutonic rocks, or they traverse the host rocks at random. Some dikes are cut by plutonic rocks, but many occur parallel to the joints in the plutonic rocks. In composition they range from gabbro through diorite to quartz monzonite and to granite. These structural relations suggest that some dikes may be older than the plutonic rocks but most are younger.

GRANOPHYRIC AND PORPHYRITIC DIKES OF GRANITIC COMPOSITION

Many of the dikes and sills are light- to medium-gray porphyritic rocks with easily recognized light-gray phenocrysts of quartz and white phenocrysts of feldspars. The number of phenocrysts and the ratio between quartz and feldspars among them vary widely. Some dikes have far more quartz than feldspar phenocrysts, whereas others have more feldspar phenocrysts. In some dikes most of the feldspar phenocrysts are orthoclase; in others, plagioclase. Where the orthoclase phenocrysts are more numerous, they are larger than the plagioclase phenocrysts; however, where the amounts of orthoclase and plagioclase phenocrysts are about equal, their sizes also tend to be equal. The phenocrysts commonly range from 1 mm to 1 cm in diameter.

The groundmass is fine to medium grained, 0.1 to 0.5 mm, and consists of quartz, orthoclase, plagioclase, and biotite. Granophyric intergrowth of quartz and feldspars is common in the groundmass, and also around the quartz phenocrysts (fig. 15*A*). In these granaphyric rims the orientation of a part of the quartz is parallel to the host phenocryst but another part forms groups belonging to small individual crystals. Biotite is in small flakes, more rarely as needlelike crystals. Magnetite and zircon are the accessories.

About a mile east of Browns Rock, this common type of granite porphyry grades to a fine-grained spherulitic rock with albite spherules, 0.5 to 1 cm in diameter, in a fine-grained groundmass of tiny polygonal grains of quartz, albite, and orthoclase, flakes of biotite, and small spherules of albite. The grains in this groundmass range from 0.1 to 0.2 mm in diameter.

Several of the granitic dikes along Swanson Creek are medium grained and almost equigranular. Plagioclase crystals are subhedral and smaller than the anhedral grains of quartz and orthoclase. Some quartz occurs in small round grains, many of which are included in orthoclase. Small flakes of biotite occur as the dark constituent. Some of these equigranular dikes have granite porphyry in their center, others are cut by porphyritic dikes.

Near Deadhorse Lookout, the common type of granite porphyry grades to a variety with as much as 60 percent of large phenocrysts. The phenocrysts, 0.5 to 1 cm in diameter, are quartz, orthoclase, and albite. The groundmass, medium- to fine-grained, consists of polygonal grains of quartz, orthoclase, albite, and biotite. Granophyric intergrowth of quartz and feldspars is common around the phenocrysts (fig. 15*B*). This coarse-grained granite porphyry is mineralogically similar to the porphyritic granite border facies of the coarse-grained granite between Deadhorse Lookout and the North Fork of the Clearwater River. Thus there seems to be a complete gradation from coarse-grained granite through porphyritic granite to granite porphyry and further to the spherulitic rock. The field relations and the similarity of composition and mineralogy indicate that all these granitic intrusive rocks solidified from the same magma, the different textures being due mainly to the rate of cooling.

Many of the joints in the coarse-grained granite at Bungalow are filled by dikes of fine- to medium-grained gray rock with scattered drusy vugs. The walls of the dikes are straight, contacts are sharp, and their grains are smaller toward the walls. The vugs are lined by euhedral crystals, 3 to 5 mm long, of smoky quartz and white slightly perthitic orthoclase. The quartz crystals have well-developed prism and rhomb faces. The orthoclase contains abundant small opaque inclusions similar to those in the main part of the granite.

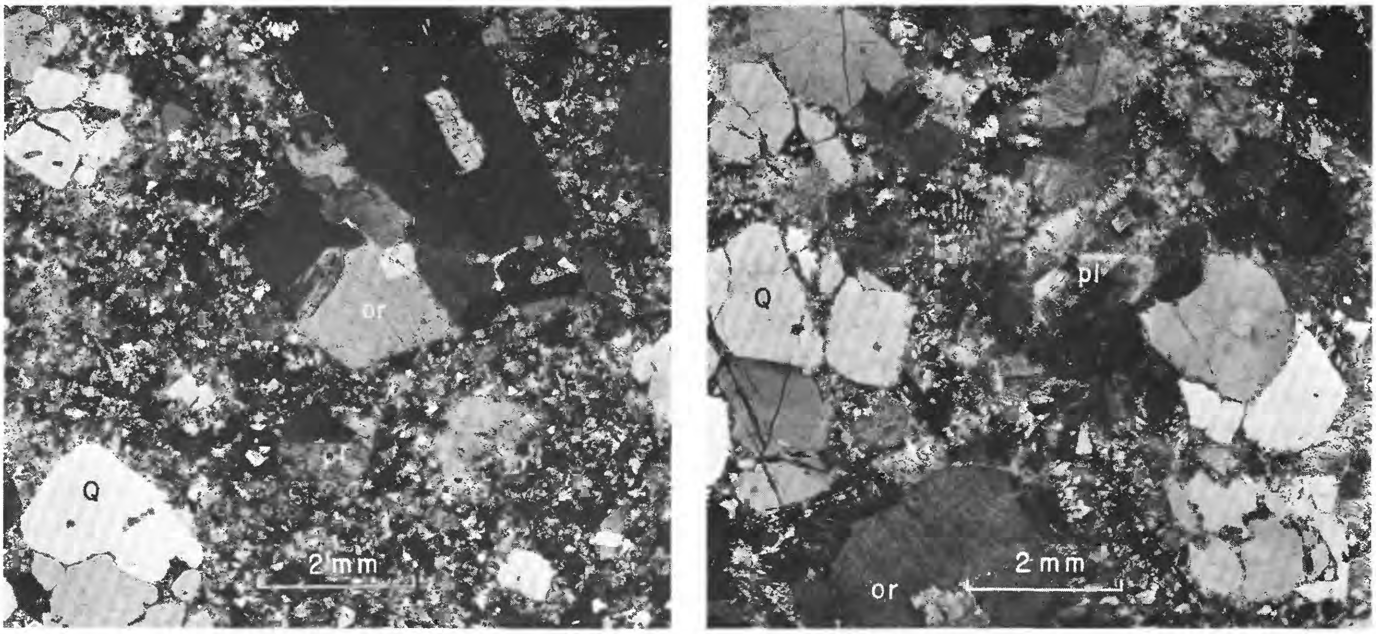
The minerals in these dikes are the same as those in the host granite, but the texture is different. A part of the quartz, plagioclase (An_8), and orthoclase occurs as subhedral or, occasionally, euhedral, crystals in a granophyric groundmass. Both plagioclase and orthoclase are intergrown with quartz in this groundmass. Some of the granophyre has crystallized around euhedral plagioclase and orthoclase grains. The central plagioclase crystals show square cross sections and polysynthetic twinning. Granophyre next to the central feldspar grains is finer grained than that between the grains. All the orthoclase is heavily clouded by tiny dustlike inclusions of opaque iron oxide similar to that in the orthoclase of the coarse-grained granite.

Greenish-brown biotite and chlorite, in amounts usually less than 3 percent, are the dark constituents. Most of them occur as long slender needlelike flakes; only a few small flakes are equidimensional. All the biotite flakes are symplectically intergrown with quartz. Zircon with dark pleochroic halos is included in biotite. The chlorite is deuterically altered biotite.

It is noteworthy that all the granitic dike rocks have perthitic orthoclase heavily clouded by tiny opaque inclusions. The orthoclase of the dikes has more perthitic albite and more dustlike inclusions than the orthoclase of the coarse-grained quartz monzonite. These foreign elements were originally in solid solution and their large amount indicates rapid crystallization at high magmatic temperature followed by exsolution. The granophyric groundmass indicates that the final solidification took place near the eutectic temperature, which, according to Tuttle and Bowen (1958, p. 122), is about 650°C at a depth of about 13 km, but higher if closer to the surface.

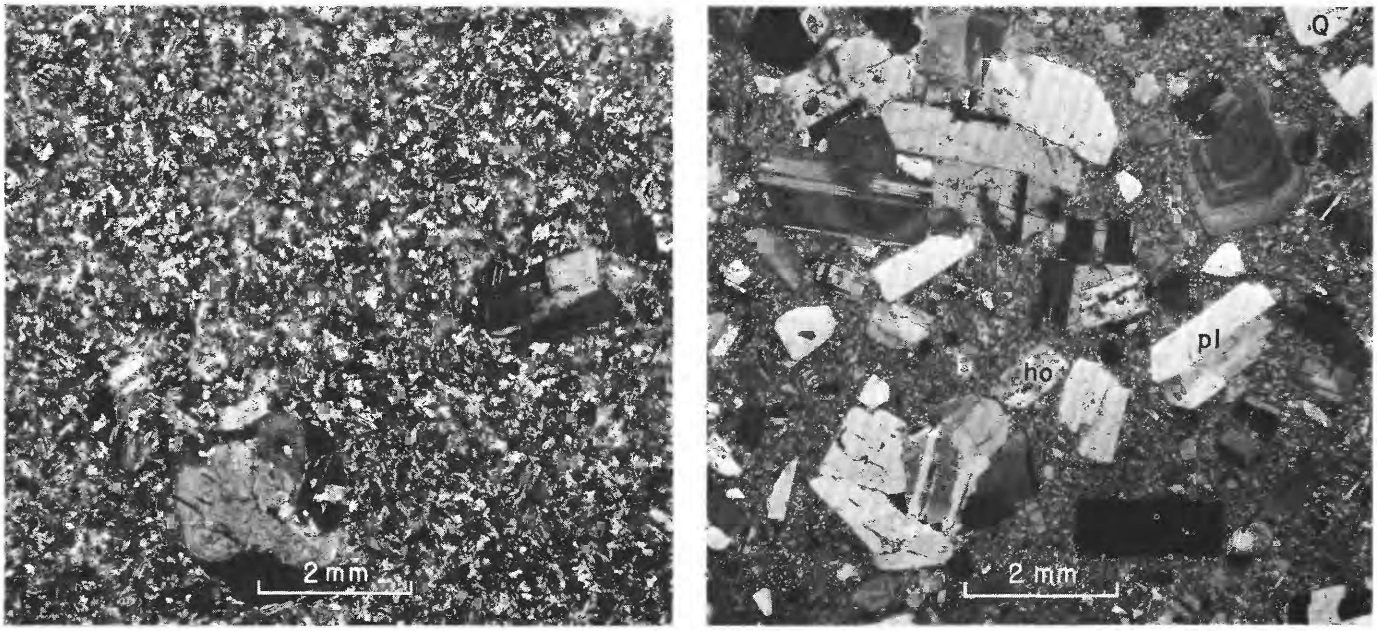
PORPHYRITIC DIKES OF QUARTZ MONZONITIC AND MONZOTONALITIC COMPOSITION

Dikes of intermediate composition vary extensively in their appearance and type of phenocrysts. Most quartz monzonitic dikes have small phenocrysts of plagioclase embedded in fine- to medium-grained dark or medium-gray groundmass of quartz, plagioclase, orthoclase, and biotite (fig. 15*C*). Some dikes, however, are fine grained and equigranular. In the dikes along Sheep Mountain Creek (loc. 1081) light-colored phenocrysts of plagioclase range from 0.5 to 2 mm in diameter, and the fine-grained dark-gray groundmass consists of



A, A common type of granite porphyry just east of Browns Rock (loc. 719). Phenocrysts are quartz (Q), orthoclase (or), and plagioclase (pl); the groundmass, which consists of these minerals plus a small amount of biotite, is fine to medium grained. Narrow rims of granophyric intergrowth of quartz and orthoclase are common around the quartz phenocrysts.

B, A coarse-grained granite porphyry near Deadhorse Lookout (loc. 1085). Larger phenocrysts are quartz (Q) and orthoclase (or). Plagioclase (pl) forms phenocrysts of medium size. Groundmass is granophyric and contains small flakes of biotite.



C, Quartz monzonitic dike 1½ miles north of Deadhorse Mountain (loc. 1081. Phenocrysts are plagioclase that occurs in tabular crystals or groups of crystals. Groundmass is fine grained and consists of lath-shaped crystals of twinned plagioclase and interstitial quartz, orthoclase, biotite, and hornblende.

D, Monzotonalitic dike 2 miles northeast of French Mountain (loc. 1143). Phenocrysts are plagioclase (pl), quartz (Q), biotite (B), and hornblende (ho). Groundmass is fine grained and contains orthoclase in addition to the minerals that occur as phenocrysts.

FIGURE 15.—PHOTOMICROGRAPHS OF HYPABYSSAL ROCKS.

tiny laths of plagioclase and intersititial grains of quartz, orthoclase, hornblende, and biotite. Sphene and magnetite are the accessory minerals. The chemical composition of these dikes is similar to that of the fine-grained quartz monzonite (table 1, No. 1081).

Dikes that have quartz as well as plagioclase phenocrysts (fig.15*D*) are common; they are similar to the granite porphyry dikes except that the number of quartz phenocrysts is smaller, orthoclase occurs only in the groundmass, the amount of plagioclase is more than 50 percent, and hornblende is common in addition to biotite. The chemical composition of these dikes is intermediate between that of the quartz monzonite and the tonalite (table 1, No. 1143); they contain more calcium and less potassium than the quartz monzonitic dikes and considerably less silicon and potassium than the normal granitic dikes. Because of their intermediate composition, these dikes are called monzotonalitic in this report.

The monzotonalitic dikes east and south of French Mountain are light to medium gray and coarse to medium grained. The phenocrysts, which make up 10 to 40 percent of the rock, consist of about 60 percent plagioclase, 10 percent quartz, 20 percent biotite, and 10 percent hornblende. The plagioclase is strongly zoned and twinned and many "phenocrysts" are made of groups of crystals 1 to 3 mm long. Biotite phenocrysts have hexagonal or rounded outlines and range from 0.2 to 1 mm in diameter. The groundmass is fine grained, granoblastic and makes up about 50 percent of the rock. Some of the biotite is altered to chlorite; sphene and magnetite are accessories. In certain much-altered dikes, abundant calcite fills the centers of hornblende crystals, and leucoxene and rutile are included in altered biotite flakes. A chemical analysis, No. 1143 in table 1, shows that the composition of typical monzotonalitic dike rock is close to that of the medium-grained monzotonalitic border zone of the coarse-grained quartz monzonite.

DIORITIC DIKES

Most dioritic dikes are fine grained, dark to medium gray and equigranular. The major constituents are plagioclase and hornblende; biotite and interstitial quartz occur in small quantities; and sphene, magnetite, and epidote are the accessories. Plagioclase occurs either in small lath-shaped crystals, as along Orogrande Creek (loc. 1139), or as equidimensional grains. The hornblende prisms are dark green, 0.1 to 1 mm long, and unoriented.

A medium-gray fine- to medium-grained dioritic dike cuts quartz diorite just east of the granite stock at Bungalow. About 60 percent of this rock is albite in equidimensional subhedral grains, 0.1 to 0.2 mm in diameter, that contain tiny inclusions of sericite and epidote minerals. Hornblende, which constitutes about 15 percent of the rock, is dark to blue green, and some of the grains have augite in their centers. The amount of quartz is about 15 percent. About 4 percent of the rock is orthoclase, both as small grains and in granophyric intergrowth. Biotite, chlorite, and magnetite occur in small quantities.

GABBROIC DIKES

The gabbroic dikes are fine grained and dark gray, resembling the dioritic dikes. The major constituents are plagioclase (45–55 percent), hornblende (15–35 percent), augite (10–20 percent), biotite (2–15 percent), and magnetite. Plagioclase (An_{44}) is in unoriented lath-shaped crystals in which cracks are filled by green hornblende. Most of the augite is altered to green hornblende that includes small grains and lamellae of ilmenite-magnetite. Hornblende is light green and occurs in aggregates and small prisms. Biotite is reddish brown and contains tiny inclusions of magnetite. Many small flakes are included in the aggregates of hornblende. Small grains of magnetite, irregular in shape, occur with hornblende and biotite. A chemical analysis of a gabbroic dike along Sheep Mountain Creek is No. 1079 in table 1; this dike rock is chemically and mineralogically similar to the gabbro along Beaver Creek and probably is genetically related to it.

GEOCHEMISTRY OF THE IGNEOUS ROCKS

The plutonic and dike rocks range from ultramafic pyroxenite and hornblendite through olivine gabbro, pyroxene and hornblende gabbro, diorite, and quartz diorite to tonalite, quartz monzonite, and granite. The field relations, mineralogy, and chemical composition suggest that there are two separate suites of intrusive rocks: a quartz dioritic suite that includes pyroxenite, hornblendite, hornblende gabbro, quartz diorite, and tonalite; and a quartz monzonitic suite that includes olivine and pyroxene gabbro, diorite, quartz monzonite, and granite. Minerals such as olivine, pyroxene, and orthoclase indicate that the rocks of the quartz monzonitic suite crystallized from a dry magma whereas the minerals of the rocks of the quartz dioritic suite indicate a wet environment in which hornblende and biotite crystallized instead of pyroxenes and orthoclase. Comparison of the chemical composition of the rocks and distribution of the minor elements within each series should help resolve the problem of origin and differentiation of the magmas. Did the two rock series crystallize from the same parent magma or were two separate magmas formed at different places and times?

VARIATION IN THE MAJOR AND MINOR ELEMENTS

Ten new analyses of the rocks belonging to the quartz monzonite series are given in table 1, along with two analyses published earlier (No. 758, Hietanen, 1962; No. L 219, Larsen and Schmidt, 1958) of the granite at Bungalow. Results of spectrographic analysis of minor elements in the same specimens are presented in table 3. Silica variation diagrams (fig. 16–20) are based on the data in these two tables (1, 3) and on the data published earlier on intrusive rocks just northwest of the present area (Hietanen, 1962, tables 1, 10, 15, 26, and 27). The major elements are plotted in ionic percentages. The minor elements are plotted in both weight percent, on the right, and in number of ions, corresponding to the ionic percentage of major elements, on the left.

Fairly smooth curves can be drawn by inspection through the plots for most elements. Some points, however, are considerably above or below the curves, so that two branches could have been drawn. Most of the plots for the rocks of the quartz diorite series fall along the same curves as those for the quartz monzonite series; in a few instances, the curves are notably different.

TABLE 3.—*Quantitative spectrographic analyses, in parts per million, of minor elements in the igneous rocks in the northwestern part of the Idaho batholith* [1]

[Analyst: Paul R. Barnett, U.S. Geological Survey]

Sample No.	Rock name	Ni	Co	Cr	Cu	V	Ga	Sc	Ba	Sr	Be	Y	Yb	La	Zr	B	Pb	Nb	Sn
Gabbroic rocks																			
1063	Olivine gabbro	600	100	300	50	100	6	6	500	600	0	0	0	0	30	0	0	0	0
1073	Pyroxene gabbro	90	50	200	80	100	20	20	1000	1000	0	80	4	0	100	0	0	0	0
1072	Gabbro	60	40	90	40	100	10	30	1000	2000	0	90	3	0	200	0	0	0	0
1079	Gabbroic dike	90	50	70	50	100	20	30	1000	1000	0	90	5	0	200	0	0	0	0
Mean		80	47	120	57	100	17	26	1000	1333	0	87	4	0	166	0	0	0	0
Monzotonalitic rocks																			
1143	Monzotonalitic dike	16	11	48	15	56	20	10	700	400	---	20	2	<100	110	<20	0	0	0
1125	Monzotonalite	4	4	14	3	68	20	10	540	660	---	30	3	<100	140	<20	0	0	0
1121	Monzotonalite	2	4	4	3	43	20	10	980	760	---	20	2	100	180	<20	0	0	0
Mean		7	6	33	7	56	20	10	740	606	---	23	2	100	143	20	0	0	0
Quartz monzonitic rocks																			
1081	Quartz monzonitic dike	0	10	3	10	40	10	10	2000	500	2	80	3	200	300	0	0	0	0
1064	Quartz monzonite	7	5	20	10	30	10	6	1000	400	2	50	1	100	200	0	20	0	0
1880	Quartz monzonite	0	0	0	2	20	22	0	1400	700	0	0	0	0	180	0	0	0	0
Mean		2	5	8	7	30	14	5	1467	533	1	43	1	100	227	0	7	0	0
758	Granite	0	0	1	4	5	10	2	50	10	9	100	10	0	200	0	40	20	10
Average of 3 tonalites (190, 379, 43 from tables 26 and 27 in Hietanen, 1962)		2	3	3	4	30	20	0	466	1000	---	0	0	0	123	-----	0	0	0
Average of 4 quartz diorites (180, 42, 292, 322 from tables 26 and 27 in Hietanen, 1962)		44	17	57	21	112	22	9	415	975	---	10	0	0	140	2	0	0	0

[1] Looked for but not found: Ag, As, Au, Bi, Cd, Ge, In, Mo, Pt, Sb, Ta, Th, Tl, U, V, W.

SODIUM, POTASSIUM, AND BARIUM

Variation curves for sodium, potassium, and barium are combined in figure 16. The amount of sodium increases steadily up to about 61 percent silicon, drops between 61 and 65 percent silicon and stays fairly constant at the silicic end. The amount of potassium is 1 to 1.5 percent at the mafic end, increases rapidly from 1.5 to 5 percent between 61 and 65 percent silicon, and at the silicic end, is again constant at about 5 percent.

A considerable scattering of the sodium and potassium points between 61 and 65 percent silicon reflects the branching of parental magmas into two series, one rich in potassium and the other poor in potassium and exceptionally rich in sodium. The latter series includes the tonalitic rocks (Nos. 42, 190, 379, 43) and the potassium-poor border zone of the quartz monzonitic batholith (Nos. 1125, 1121). Representatives of the potassium-rich rocks within this interval are dike rocks (Nos. 1143, 301, 1081) that may represent the composition of the quartz-monzonitic parent magma. The question arises whether or not the tonalites could be products of crystallization differentiation of the quartz

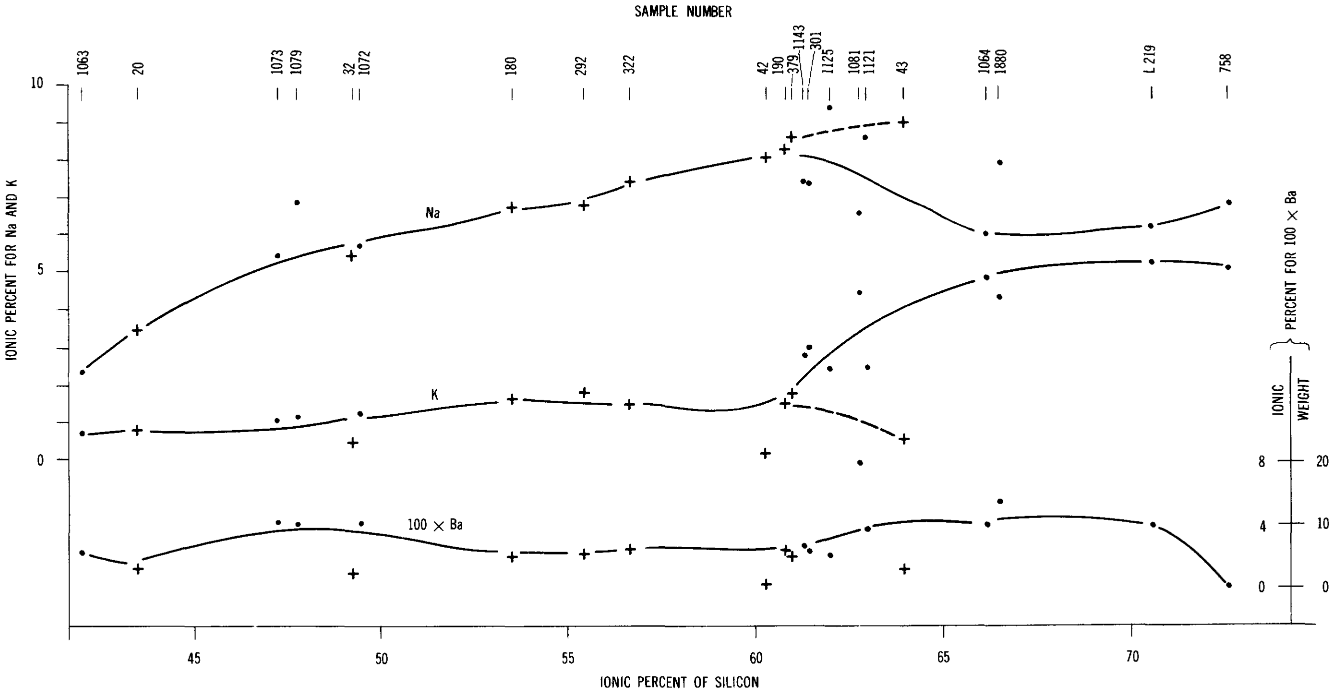


FIGURE 16.—Variation of sodium, potassium, and barium in the rocks of the Idaho batholith. Nos. 1063, 1073, 1072, 1079, 1143, 1125, 1121, 1081, 1064, 1880, and 758 are from tables 1 and 3. No. L 219 is from Larsen and Schmidt (1958). Added for comparison are igneous rocks from the area to the west (fig. 2; and Hietanen, 1962, tables 1, 10, 15, 26, and 27) as follows: No. 20 is hornblendite: No. 32 hornblende gabbro; Nos. 180, 292, 322, and 42, quartz diorite; Nos. 379, 190, and 43, tonalite; No. 301 is monzotonalitic dike rock called "granite porphyry" earlier. Localities for these specimens are shown on figure 2. Dots refer to the rocks of the quartz monzonite series and crosses to those of the quartz diorite series.

monzonitic magma, formed in a manner similar to that of the potassium-poor border zone. This question is considered later.

The barium curve is shaped like the potassium curve except that the content of barium drops abruptly at the extreme silicic end. According to Sen, Nockolds, and Allen (1959) potassium feldspar and biotite are the chief carriers of barium in the Southern California batholith. In Idaho, the amount of potassium feldspar stays constant in rocks that have more than 65 ionic percent of silicon and the amount of biotite in the most silicic member is only slightly less than in the intermediate rocks. It seems, therefore, that the barium entered the crystal structure early and that the residual magma from which the granite crystallized was depleted in this element.

CALCIUM AND STRONTIUM

The amount of calcium decreases steadily from hornblendite and gabbro to granite (fig. 17). A considerable drop occurs at the extreme mafic end between pyroxene gabbro and more mafic olivine-bearing gabbro in which the calcium content is about the same as in the quartz diorites. There is no systematic difference between the distribution of calcium in the quartz monzonite series and in the quartz diorite series.

The curve for strontium is similar to that for calcium. The amount increases from olivine gabbro to gabbro and then decreases steadily toward the silicic end. The ratio Sr×100/Ca increases from about 0.8 at the basic end to about 1 at the silicic end. The rocks of the Idaho batholith have more strontium than comparable rocks of the Southern California batholith (Nockolds and Allen, 1953) but the ratio Sr×100/Ca is approximately equal.

IRON, MAGNESIUM, NICKEL, COBALT, VANADIUM, SCANDIUM, AND CHROMIUM

Iron and magnesium increase steadily from the silicic end to the mafic end, the increase in magnesium, however, is much greater (fig. 18). No difference can be seen in the distribution of these elements between the rocks of the quartz monzonite and those of the quartz diorite series. The shape of the variation curve for cobalt is much like that for iron; the ratio Co×1000/Fe decreases slightly from the mafic end toward the silicic end where the amount of cobalt is below the analytical limit. The amount of nickel increases with increasing magnesium content; olivine in the most mafic member has the ratio Ni×1000/Mg=2. The same ratio in gabbros is about 1, and in tonalites and quartz monzonites about 0.5.

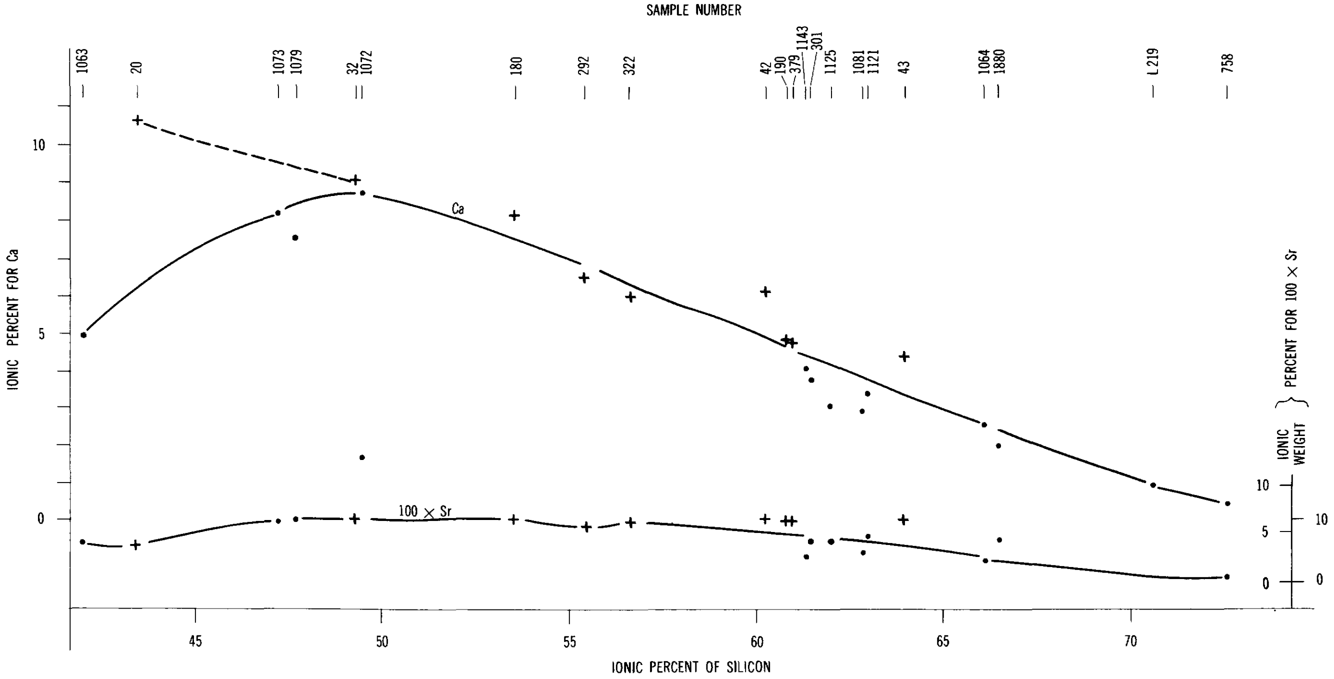


Figure 17.—Variation of calcium and strontium in the rocks of the Idaho batholith. Numbers refer to the same analyses as those in figure 16. Dots refer to rocks of the quartz monzonite series and crosses to rocks of the quartz diorite series.

The mafic members of the quartz diorite series, especially the hornblendite, have more vanadium than the rocks of the quartz monzonite series. In general, the amount of vanadium increases with increasing content of magnesium but drops abruptly at the extreme mafic end. Most of the vanadium in the Southern California batholith (Sen, Nockolds, and Allen, 1959) is in magnetite, augite, hornblende, and biotite, whereas in olivine the amount of vanadium is below the analytical limit. The olivine gabbro in Idaho similarly has little vanadium. Hornblendite (No. 20) has an extraordinarily high content of vanadium even though it contains only one percent ilmenite-magnetite (Hietanen, in 1962, table 15). The hornblende gabbro (No. 32) contains considerably more vanadium than the pyroxene gabbros (Nos. 1072, 1073, 1079). It seems, therefore, that hornblende is the chief carrier of vanadium in the rocks of the Idaho batholith.

The shape of the variation curve for scandium is similar to that for vanadium; the highest content is also in the hornblendite (No. 20). With the exception of this rock the points for scandium in the quartz monzonite series and in the quartz diorite series fall along the same curve.

The content of chromium and ferric iron increase toward the mafic end where the amount of ferric iron suddenly drops but that of chromium increases. Magnetite is the chief carrier of these elements and according to Sen, Nockolds, and Allen (1959) this mineral is richer in chromium when in mafic rocks.

ALUMINUM AND GALLIUM

The ionic percentage of aluminum increases rapidly from the mafic end to a silicon content of 47 percent (fig. 19). It stays constant between 47 and 61 percent of silicon and decreases toward the silicic end. The aluminum curves are similar for rocks of both series. The content of gallium increases slightly from the very mafic end to hornblendite, stays constantly low for most silicon contents, and declines toward the silicic end to approach there a value that is as low as that at the mafic end.

YTTRIUM AND ZIRCONIUM

In general the amounts of yttrium and zirconium are fairly small but variable; the points for zirconium especially are notably scattered around the curve (fig. 20); some of the gabbros and dike rocks contain considerably more of these elements than other rocks with about the same silicon content. The quartz diorite No. 322 and the tonalite No. 190 are rich in zirconium whereas the gabbro No. 32 and the quartz diorite No. 42 are poor in it. The average content of yttrium in the quartz diorite series is lower than that in the quartz monzonite series.

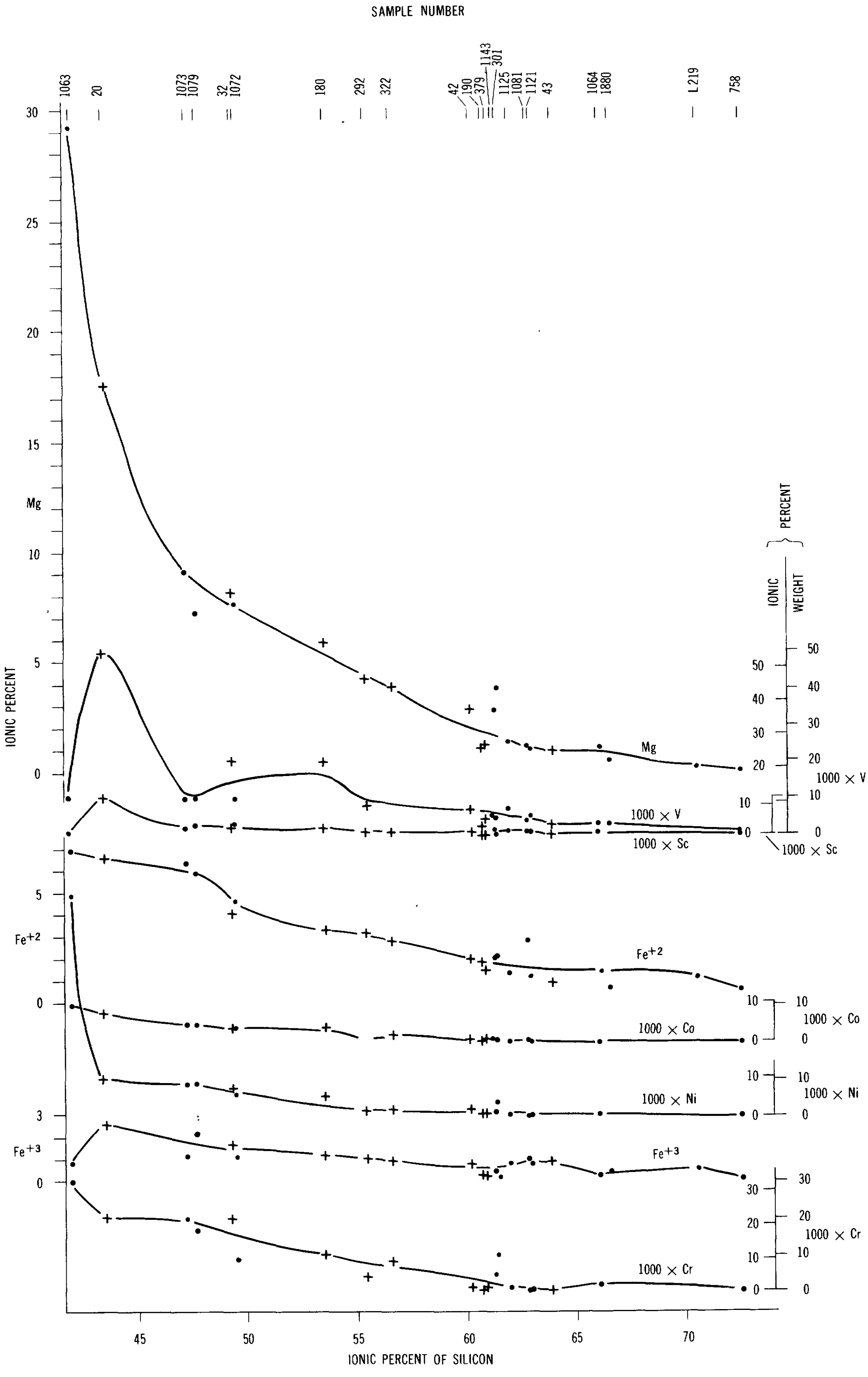


Figure 18.—Variation of iron, magnesium, nickel, cobalt, vanadium, scandium, and chromium in the rocks of the Idaho batholith. Numbers refer to the same analyses as those in figure 16. Dots refer to rocks of the quartz monzonite series and crosses to those of the quartz diorite series.

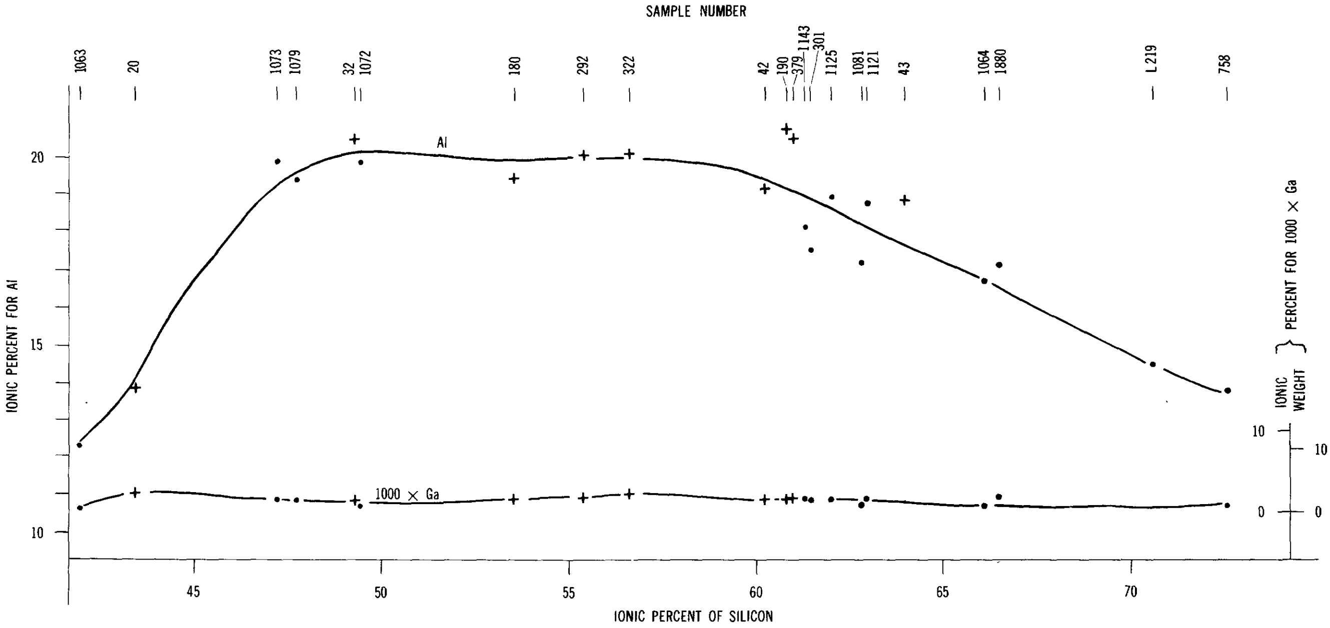


Figure 19.—Variation of aluminum and gallium in the rocks of the Idaho batholith. Numbers refer to the same analyses as those in figure 16. Dots refer to rocks of the quartz monzonite series and crosses to rocks of the quartz diorite series.

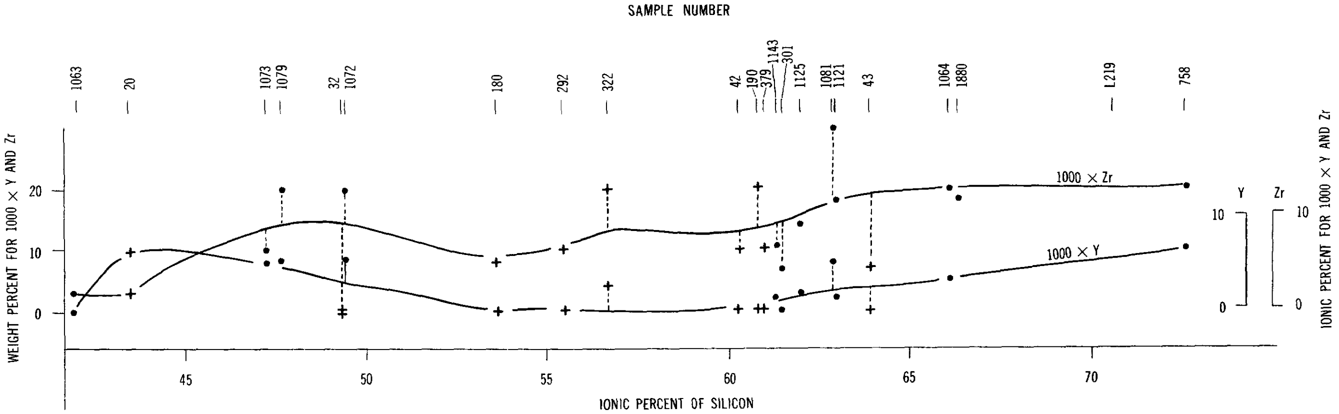


Figure 20.—Variation of yttrium and zirconium in the rocks of the Idaho batholith. Numbers refer to the same analyses as those in figure 16. Dots refer to rocks of the quartz monzonite series and crosses to rocks of the quartz diorite series.

MOLECULAR NORMS AND CLASSIFICATION OF IGNEOUS ROCKS

Molecular norms have been calculated by regrouping the ionic percentage following the method by Barth (1952). The ternary diagrams of figures 21 to 23 are based on these molecular norms.

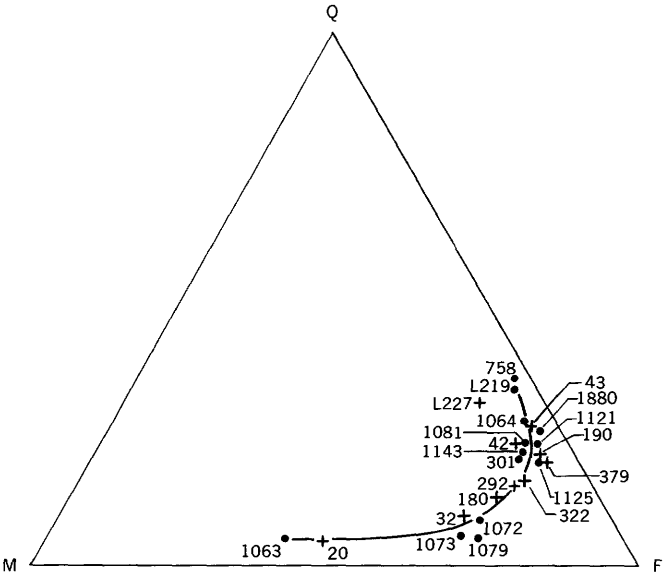


FIGURE 21.—MFQ diagram for the igneous rocks. F, total amount of feldspars in the molecular norm; M, total of iron and magnesium calculated as orthosilicates to which is added corundum and magnetite, all from molecular norm; Q, molecular norm of quartz after calculation of M. Q+M+F=100. Nos. 1063, 1073, 1072, 1079, 1143, 1125, 1121, 1081, 1064, 1880, 758, and L 219 are from table 1. No. L 227 is tonalite between Pierce and Bungalow (Larsen and Schmidt, 1958). Added for comparison are igneous rocks from the area to the west (fig. 2, and Hietanen, 1962, tables 1, 10, and 15) as follows: No. 20, hornblendite; No. 32, hornblende gabbro; Nos. 180, 292, 322, and 42 are quartz diorite; Nos. 379, 190, and 43, tonalite. No. 301 is monzotonalitic dike rock called "granite porphyry" earlier (Hietanen, 1962). Dots refer to rocks of the quartz monzonite series and crosses to those of the quartz diorite series.

The plots for rocks of the quartz diorite series in an area west of the Bungalow area are added for comparison. These analyses were made from rock types similar to those in the western part of the area under discussion (Hietanen, 1962).

In the MFQ diagram (fig. 21), F is the total amount of normative feldspar. M is the combined iron and magnesium calculated as orthosilicates, forsterite and fayalite, to which corundum or wollastonite and magnetite are added. Q is the amount of normative quartz after calculating M and Q+F+M=100. The plots for most analyses are alined along a curve close to the F corner; only one point, L 227, falls very far from this curve. According to the description by Larsen and Schmidt (1958), L 227 was collected from quartz diorite northeast of Pierce along the road leading to Bungalow, apparently near the contact with biotite

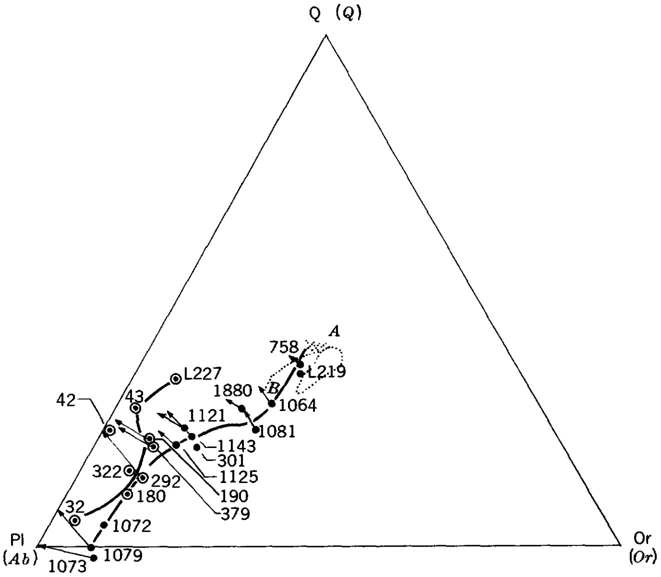


FIGURE 22.—Pl–Or–Q diagram from the igneous rocks. Molecular norms for plagioclase (Pl), orthoclase (Or), and quartz (Q) are recalculated to 100 and plotted (dots). The rocks that are deficient in Q are plotted under the Pl–Or line. The numbers refer to the same analyses as those in figure 21. The dots for the rocks of the quartz diorite series are encircled. The trend line for the rocks of the quartz diorite series (at left) and that for the rocks of the quartz monzonite series are drawn by inspection. Molecular modes for 9 rocks of the quartz monzonite series and for 3 rocks of the quartz diorite series, 2 tonalites (190 and 379), and 1 quartz diorite (292) are shown by the arrowheads connected with the dots that represent the molecular norm for the same rock. Superimposed with a stippled closed line is the area of highest concentration of granites and the positions of the "ternary" minimum for mixtures of Ab, Or, Q at various pressures of water vapor, 500 kg/cm² at A to 4000 kg/cm² at B after Tuttle and Bowen (1958, p. 75). The weight norms of two granites, Nos. 758 and L 219, are shown by stippled crosses that also are connected with the corresponding molecular norms. All stippled values refer to weight norm.

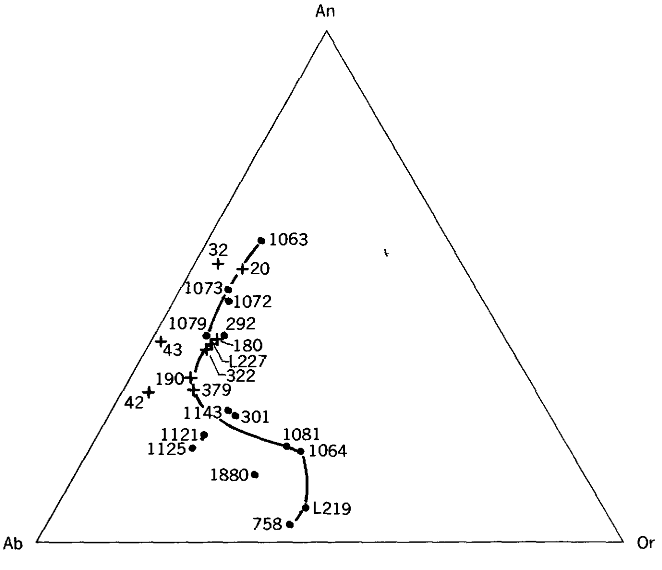


FIGURE 23.—Ab–Or–An diagram for the igneous rocks. Molecular norms for orthoclase (Or), albite (Ab), and anorthite (An) are recalculated to 100 and plotted on this ternary diagram. The numbers, dots, and crosses refer to the same analyses as those in figure 21.

quartzite of the Wallace formation. The large amount of quartz in this specimen as compared with the normal quartz diorite (No. 292) is probably due to digestion of metasedimentary country rocks rich in quartz.

Molecular norms and modes of quartz, plagioclase, and orthoclase are recalculated to 100 and plotted in figure 22. The plots for the rocks of the quartz diorite series (circled dots) are closer to the plagioclase corner than those for the rocks of the quartz monzonite series. A trend line drawn by inspection through them stays close to the Pl–Q side of the triangle. The dots for the rocks of the quartz monzonite series are scattered around a trend line that extends from the center of the diagram toward the plagioclase corner. The plots for the two granites (Nos. 758 and L 219) at the end of this trend line are in the minimum melting trough as determined by Tuttle and Bowen (1958). Plots for quartz monzonites are a short distance from the granite toward the plagioclase corner. The monzotonalites and monzotonalitic dikes form a distinct group between the quartz monzonites and tonalites. Plots for two typical quartz diorites, Nos. 292 and 180, are along the trend line for the quartz monzonite series, whereas at the basic and silicic ends the trend lines for these two rock series deviate considerably.

The composition of the normative feldspar is shown in figure 23, which was prepared by recalculating the sum of molecular norms Or, Ab, and An to 100. In this Ab–Or–An diagram, the plots for various different rock types are grouped along a zone that curves from near the center of Or–Ab line to near the center of Ab–An line. In the granite (Nos. 758 and L 219) the amount of An is less than 10 percent and the amounts of Or and Ab are about equal. In the quartz monzonite An is considerably larger (15–20 percent), and in the tonalites, An is larger and Or smaller than the corresponding values for quartz monzonite. The relative amount of An increases further in the quartz diorite, gabbro, and ultrabasic rocks.

The parameters shown in figures 21 and 23 provide the basis for a simple and easy classification of intrusive rocks. The MFQ diagram gives the values for M and Q in the various petrologic groups and the Ab–Or–An diagram defines the relationship of the feldspars. Some standard rock groups are defined on the basis of molecular norms in table 4, which also shows examples of each group, and the range of the amount of normative feldspars. The plots for the normative feldspar (fig. 23) in the border facies of the coarse-grained quartz monzonite (Nos. 1121 and 1125) are between the plots for the quartz monzonites and the tonalites. The common occurrence of porphyritic dikes of the same composition (Nos. 1143 and 301) proves that this composition existed as a magma and should be considered as a separate petrologic type. The name monzotonalite is used for these rocks because of their intermediate character between quartz monzonites and tonalites. The amounts of dark constituents and quartz in the analyzed examples of the three groups—quartz monzonite, monzotonalite, and tonalite—are about equal (M, 5–10; Q, 18–30), but the differences are in the proportions of orthoclase, albite, and anorthite. The rocks of the monzotonalite group have little orthoclase, whereas the quartz monzonite is rich in this mineral. The dike rocks of monzotonalitic composition resemble many porphyritic granite dikes but lack orthoclase phenocrysts and have much less normative orthoclase than the granitic dikes. No porphyritic dikes of tonalitic composition were found.

The intrusive tonalite looks like the fine-grained monzotonalitic border facies of the coarse-grained quartz monzonite batholith, but these rocks differ chemically and mineralogically. The tonalite contains less potassium and a little more calcium and therefore less orthoclase and more plagioclase (table 4).

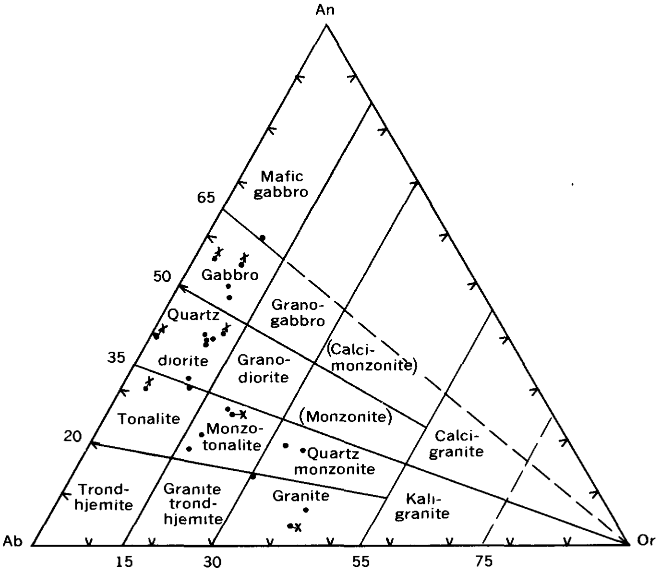
The intrusive tonalites are chemically and mineralogically similar to the medium-grained gneissic border facies of the quartz diorite and also similar to the small occurrences of gneissic tonalite formed through metasomatic processes (Hietanen, 1962); they may represent mobilized parts of such occurrences. The tonalites contain less femic constituents and more quartz than the quartz diorite, and the plagioclase in the tonalite is more albitic (table 4). Thus the rocks that are called tonalite in this report are intermediate between quartz diorite and the most silicic member of the potassium-poor intrusive rock series, the trondhjemite (Goldschmidt, 1916).

The gabbro associated with the quartz diorite contains only hornblende and biotite as dark constituents, whereas most gabbro belonging to the quartz monzonite series contains pyroxene as well. Olivine appears only in a few small mafic inclusions in the quartz monzonite. The hornblende and pyroxene gabbro commonly have a few grains of quartz. After all femic constituents are calculated as olivine, the amount of quartz in the molecular norm is 5 to 10 in the gabbros, and about 5 in the mafic gabbros. The normative amount of anorthite increases toward the mafic end of the series, but the amount of normative orthoclase stays around 10 percent of the total amount of feldspars.

Several of the subdivisions shown in table 4 are based mainly on the relative amounts of orthoclase, albite, and anorthite; therefore the Ab–Or–An diagram is best suited for the illustration of this classification. Figure 24 was prepared on the basis of figure 23 by dividing

TABLE 4.—*Classification of igneous rocks on basis of their molecular norms*

	Granite	Quartz monzonite	Monzotonalite	Tonalite	Quartz diorite	Gabbro	Mafic gabbro
Range of M and Q from figure 21							
M	5	5–10	5–10	4–10	8–20	20–40	40
Q	30–35	23–30	18–23	18–28	10–18	5–10	5
Normative feldspar content from figure 23							
Or	30–55	30–55	15–30	0–15	0–15	0–15	0–15
$\frac{An}{Ab+An}$	0–20	20–35	20–35	20–35	35–50	50–65	65
Examples of various groups and the range of their Or, Ab, and An in table 1							
	758 L 219	1064 1081	1143, 301 1121, 1125	190, 379 43	322, 292, 180 L 227, 42	32, 1072 1073, 1079	20, 1063
Or	25–30	20–25	12–15	3–10	1–10	3–7	3–4
Ab	30–35	30–35	35–47	40–45	34–40	27–35	12–17
An	2–5	10–15	13–18	20–25	27–30	28–36	22–24
Ratios between Or, Ab, and An in the rocks of table 1							
	Or>3 An Or≤Ab Ab>4 An	An<Or<3 An Or<Ab Ab>2 An	Or<An Or<½ Ab Ab>2 An	Or<½ An Or<¼ Ab Ab≃2 An	Or<⅓ An Or<¼ Ab Ab>An	Or<¼ An Or<¼ Ab Ab<An	Or<⅙ An Or<⅓ Ab Ab≃½ An



	Q	M
Granite	30-35	<5
Quartz monzonite	20-30	5-10
Monzotonalite	20-25	5-10
Tonalite	20-30	5-10
Quartz diorite	10-20	10-20
Gabbro	5-10	20-40
Mafic gabbro	<5	> 40

FIGURE 24.—Classification of coarse-grained calc-alkalic igneous rocks on the basis of their normative feldspar content. Ab, Or, and An refer to the molecular norms of albite, orthoclase, and anorthite. The plots for the weight norm would fall a short distance farther from the Ab corner, as shown by crosses for 2 gabbros, 2 quartz diorites, a tonalite, a monzotonalite, and a granite.

the field occupied by the intrusive rocks into subfields in such a manner that each subfield contains plots for a certain rock type. The composition of plagioclase along each tie line that joins a point along the Ab–An line to the orthoclase corner is constant. Thus the range of the anorthite content of the plagioclase in the rocks between two tie lines, such as trondhjemite, granite-trondhjemite, and granite, stays the same and only the amount of orthoclase increases from left to right.

For the next group—the tonalite, monzotonalite, and the quartz monzonite—the tie lines are drawn through An_{20} and An_{35}. Two of the analyzed tonalites are close to the upper line because they contain considerable epidote. The composition of plagioclase in most tonalites ranges from An_{28} to An_{34}. The dividing lines parallel to the Ab–An side give the range of normative orthoclase in the various rock types. Because some potassium is in biotite, the relative amount of potassium feldspar appearing in the rock is smaller than that shown in this diagram. For example, normal trondhjemites, tonalites, quartz diorites, and gabbros have less than 5 percent modal orthoclase. Granite trondhjemite, monzotonalite, and granodiorite are intermediate rocks that contain less orthoclase than normal granite, quartz monzonite, and monzonite. The amount of normative quartz in the trondhjemites and granites is about 35 percent; in the tonalite, monzotonalite, and quartz monzonite it is usually 20 to 25 percent, and decreases steadily through quartz diorite and gabbro to mafic gabbro. In an Ab–Or–An–Q tetrahedron, figure 24 would thus be close to a section that cuts the Ab–Or–Q face along the line 35 percent quartz, and it is tilted toward the An corner where it approaches the Ab–Or–An plane. The amount of mafic constituents increases from less than 5 percent in the granite to about 45 percent in the mafic gabbro and is constant in the groups of equal anorthite content, such as tonalite, monzotonalite, and quartz monzonite.

It appears that this norm classification is better suited to the present suite of rocks than a modal classification. In the modal classification, the tonalite and quartz diorite would fall into the same subdivision because a part of the calcium in the quartz diorite is combined with iron and magnesium to form hornblende, the amount of which is not shown in the Ab–Or–An diagram. All differences between these two rock types would thus be eliminated in a classification based on the modal composition of feldspars. This lack of difference in the modal feldspar content may be the reason why these two rock types are usually considered synonymous. The norm classification usefully brings out the considerably higher content of calcium, iron, and magnesium in the quartz diorite as compared with the tonalite.

SEQUENCE OF CRYSTALLIZATION IN IGNEOUS ROCKS

The textures of the igneous rocks indicate that the order of crystallization of the major constituents varied considerably. Typical examples can be found among the silicic rock types of the quartz monzonite series. In the medium-grained quartz monzonite along Beaver Creek, numerous small euhedral and lath-shaped plagioclase crystals are included in and apparently are older than large anhedral orthoclase crystals. In the dikes of the same composition, plagioclase forms phenocrysts and the groundmass is thus richer in quartz and orthoclase than is the total rock. In the Pl–Or–Q diagram, a plot for the groundmass would be toward the ternary eutectic from the plot for the dike rock No. 1081 (fig. 22), a normal crystallization trend for this magma type as experimentally demonstrated by Tuttle and Bowen (1958).

In the granite at Bunglow (No. 758) the quartz grains have better crystal forms than the feldspar grains. In dikes of the same composition, euhedral quartz phenocrysts abound and tend to be larger than the feldspar phenocrysts. Crystallization apparently started with quartz, and after the magma reached the eutectic composition, quartz, orthoclase, and plagioclase crystallized simultaneously. Composition of the phenocrysts plus the groundmass would lie toward the quartz corner from the eutectic.

In the monzotonalitic dikes plagioclase and quartz occur as phenocrysts; the plagioclase is more abundant and the orthoclase is only in the groundmass. Crystallization started with plagioclase, and continued by separation of plagioclase and quartz, enriching the residual magma, the groundmass, in the components of orthoclase. The plots for the groundmass would lie toward the eutectic (minimum melting trough) from the plots for the total composition of these dikes (fig. 22, Nos. 1143 and 301).

The cooling history of the coarse-grained quartz monzonite with large orthoclase phenocrysts is not as easy to follow. If the orthoclase crystals are true phenocrysts formed at an early stage in a liquid magma, the magma should have been rich in potassium; accordingly, the plots for this rock should lie toward the Or corner from the eutectic area of the Pl–Or–Q diagram. However, the plots for all quartz monzonites fall toward the Pl corner, which indicates that the magma was rich in sodium and calcium rather than in potassium. In such magma, crystallization should have started by separation of plagioclase as it did in the quartz monzonite along Beaver Creek. The textures in the two masses of quartz monzonite differ markedly; many euhedral plagioclase crystals are included in anhedral orthoclase along Beaver Creek, whereas the inclusions of plagioclase in large euhedral to subhedral orthoclase phenocrysts of the coarse-grained quartz monzonite are few and small. The "groundmass" between the large orthoclase phenocrysts contains only a small amount of interstitial orthoclase and is compositionally close to monzotonalite. Thus the relation between the composition of groundmass and that of the total rock is contrary to the normal sequence of crystallization, and the orthoclase crystals can scarcely be true phenocrysts. Two explanations are worth considering:

1. That crystallization started with separation of plagioclase, and was followed by quartz, and these two minerals continued to crystallize until the composition of the liquid part of the magma reached the eutectic region where orthoclase also was crystallizing. Slow cooling permitted partial recrystallization and rearrangement of material, so that the orthoclase crystals attained a large size, and the quartz and plagioclase formed a granoblastic groundmass.

2. That crystallization started with separation of plagioclase (An_{28}), was followed by quartz, and these two minerals continued to crystallize until the phase boundary of orthoclase plus quartz was reached. There the orthoclase phenocrysts formed fast and the composition of the remaining liquid moved along the boundary

line toward the eutectic. In the eutectic region, the rest of the quartz, plagioclase (An_{26}), and orthoclase crystallized.

In the first possibility, partial recrystallization during the very latest phase of crystallization would be responsible for the formation of large euhedral orthoclase crystals. In the second case, the separation of quartz and plagioclase would continue until the remaining magma would be enriched in orthoclase to such an extent that it would behave like a potassium-rich magma. The orthoclase crystals would thus be late phenocrysts, but formed in a liquid, not in solid rock.

The equal distribution of the orthoclase in the groundmass indicates that the phenocrysts did not grow in a solid or almost solid rock. Comparison with the textures common in the porphyritic rocks throws light on the order of crystallization of the porphyritic quartz monzonite. Late crystallizing minerals usually take the elements needed for their growth from the surrounding groundmass. This is well demonstrated in many metamorphic rocks in which ferromagnesian porphyroblasts rob iron from their surroundings. Staining of a specimen from the Cathedral Peak granite, Sierra Nevada, brings out a similar relation between feldspars. Subhedral phenocrysts of orthoclase in this specimen are surrounded by a zone about 1 to 2 cm wide almost devoid of orthoclase even though interstitial orthoclase is abundant elsewhere in the rock. These phenocrysts completed their growth during a late stage, when the other constituents—quartz, plagioclase, and biotite—were solid, and took from their surroundings all the potassium that would normally go to form interstitial orthoclase. The width of the potassium-free zone around each phenocryst—1 to 2 cm—gives the maximum distance of migration of potassium atoms during this late phase of crystallization.

The orthoclase phenocrysts in the coarse-grained quartz monzonite in Idaho are similar to those in the granite just described, but there is no potassium-free zone around them. This is most likely due to a slight difference in the order of crystallization. The crystallization of the normal porphyritic rock in Idaho may have proceeded as follows: The few small plagioclase crystals that are included in the orthoclase phenocrysts were first to crystallize. The orthoclase phenocrysts started to crystallize when the plagioclase grains were 1 to 2 mm long and continued to the end of the crystallization period, using most of the potassium. They did not, however, continue to grow after the rock was almost solid because they did not rob potassium from their surroundings. Rather, enough potassium remained in the melt to permit orthoclase to crystallize interstitially.

Crystallization of an exceptionally large amount of euhedral orthoclase phenocrysts in some outcrops about a mile from the border and the formation of the border zone poor in potassium are most likely results of crystallization differentiation of the quartz monzonitic magma. Perhaps crystallization started by separation of plagioclase, quartz, and biotite; the border zone, where cooling was fastest, solidified first. The crystals were attached to the border, and the potassium that remained in the liquid moved toward the center ahead of the solidifying border. In this way, a solid border zone poor in potassium was formed and the liquid magma next to it was enriched in potassium. As the temperature continued to drop, large euedral crystals of orthoclase started to form in this potassium-rich inner zone. In places the magma became undercooled and supersaturated in potassium, so that crystallization was rapid and a great number of large euhedral crystals were formed.

SEQUENCE AND MODE OF EMPLACEMENT OF PLUTONIC ROCKS

The sequence of emplacement of various igneous bodies can be determined on the basis of their contact relations, structure, and texture. The conformable bodies which show platy and linear structures parallel to the corresponding structures in the folded country rocks are considered to be earlier than the massive bodies with discordant contacts. The sequence is as follows: hornblende gabbro and amphibolite, quartz diorite, tonalite, plagioclase pegmatite, coarse-grained quartz monzonite, medium-grained quartz monzonite, granite, and granite pegmatite. The small occurrences of pyroxene-bearing gabbro and diorite that represent early differentiates of the quartz monzonite magma may have been emplaced shortly after the intrusion of quartz diorite.

The earliest members of the intrusive series are synkinematic and the latest ones postkinematic. Small bodies of mafic rocks, which were emplaced before the folding of the metasedimentary rocks, occur west of the Bungalow area. Some of the plutonic rocks of metasomatic origin have contact relations, structures, and textures similar to those of the early intrusive bodies. But they also have enough relict structures and textures to reveal their origin. All major intrusive bodies in the mapped area have contact relations and textures typical of true igneous rocks; they were at one time, at least in part, liquid. Some intrusive rocks, such as the quartz diorite, may have been derived from large quantities of country rocks, the influence of which is seen in the variable chemical composition, especially in the exceptionally large amount of quartz in those parts of the quartz diorite bodies near quartzite country rocks (for example, Nos. L227 and 43).

The contacts between the igneous rocks and the metasedimentary country rocks are in part conformable and in part discordant. They show that the magmas chose the easiest ways parallel to the preexisting structures—the bedding or fault zones—but also had enough force to break locally across the structures of the country rocks. The extent of migmatization and metasomatic alteration of the country rocks varies and gives some information about the change of physicochemical conditions during the intrusion period. The metasomatic alteration of local portions of country rocks west of the quartz diorite around Pierce has been described in an earlier paper (Hietanen, 1962) and attributed to the introduction mainly of calcium, iron, and magnesium, and removal of silicon and potassium from local parts of the contact aureole, which is 15 to 20 miles wide. Structural studies show that this metasomatic exchange of elements took place under the quartz diorite mass, which is 5 to 15 miles thick and is tilted about 45° to the east. Probably the hot solutions were trapped for a long time under the magma, which may have extended farther to the west above the present erosion surface.

The coarse-grained quartz monzonite in the southern part of the area (pl. 2) is bordered by a zone of contact migmatites about a mile wide. Feldspathization of the metasedimentary country rocks there is common, but where the contact is between quartz diorite and quartz monzonite, the composition of quartz diorite is little affected.

In contrast to these contact effects, the medium-grained quartz monzonite along Beaver Creek and the granite around Bungalow have fairly narrow contact aureoles. Contact migmatites are lacking and the feldspathized zone is usually only a few meters wide. This difference in the chemical activity around the igneous bodies is probably closely connected with the time of intrusion and the depth of burial during the intrusion. The porphyritic border facies of the granite at Bungalow indicates that this youngest body was emplaced at shallow depths, not very far from those of the hypabyssal dikes, and cooled rapidly. The quartz diorite, however, must have cooled more slowly to make the extensive migration of elements possible. The slower cooling indicates a thicker cover, which in turn accords with the idea that the quartz diorite was intruded well before the granite.

AGE OF THE ROCKS OF THE BATHOLITH

No absolute age determinations have been made of any of the intrusive bodies of this area. According to Larsen and others (1958) the average age of tonalites (including quartz diorites) of other parts of the Idaho batholith is 108 million years and that of quartz monzonites 102 million years. The relative ages fit the sequence of emplacement of the quartz diorite and quartz monzonite in the area under discussion. The period of intrusion, of course, started earlier and lasted longer than indicated by these two averages. Absolute age determinations of the oldest plutonic rock in the area, the quartz diorite in small bodies to the west, and of the youngest, the granite at Bungalow, would answer the question of range in age.

ORIGIN OF THE MAGMAS

As pointed out earlier, the igneous rocks can be grouped into two series, an older one poor in potassium and a younger one rich in potassium. Each series has representatives among the mafic, intermediate, and silicic rocks, even though the intermediate members are most common and form larger bodies than the mafic and silicic members. The following questions arise: What was the composition of the parent magma of each differentiation series? Could potassium-poor and potassium-rich magma types have been derived from the same parent magma, or must they have originated in separate magma chambers at different levels or different places in the crust?

The textures and structures of the quartz diorite indicate that this rock crystallized from a magma that was liquid at least in part. The gabbroic rocks associated with the quartz diorite consist of the same kind of minerals as the quartz diorite, but in different proportions. The hornblendite and gabbro inclusions in quartz diorite were formed during an early stage of crystallization. Small masses of hornblendite in the metasedimentary rocks near the quartz diorite are most likely products of introduction of iron and magnesium into the country rocks (Hietanen, 1962). Some of the quartz diorite and all of the gneissic tonalites are products of feldspathization of the country rocks along and near the contacts of the quartz diorite. The intrusive tonalite contains more silicon and sodium, and less calcium, iron, and magnesium than the quartz diorite. This tonalite can be either a silicic magmatic differentiate of the quartz dioritic magma or it may represent mobilized parts of gneissic (metasomatic) tonalite. Quartz diorite forms a major part of the rocks of this series, and it seems safe to assume that the magma from which most of the intrusive rocks of the quartz diorite series crystallized had a composition close to quartz diorite.

The rocks of the quartz monzonite series are more varied in composition than those of the quartz diorite series, and two of the rock types, quartz monzonite and granite, are found as fairly large bodies. That granite is an end member of the series fits its theoretical position in the area of minimum melting in the Pl–Or–Q diagram. The composition and textures indicate that the granite and quartz monzonite are derivatives of one magma, with a composition that may have been the mean composition of all rocks belonging to this series. To find this mean composition, the area of the Idaho batholith should be carefully mapped, the amount of each rock type estimated on the assumption that all the rock masses have depths proportional to their area, and the average composition computed. Larson and Schmidt (1958, table 7) made an estimation of this average composition on the basis of their reconnaissance work. Their result is close to the composition of the quartz monzonitic dikes in the present area (comp. table 1 No. 1081) except for a somewhat higher content of CaO and Na_2O and a lower content of K_2O in the mean. These differences become insignificant if it is considered that the area of quartz diorite around Pierce is about 200 square miles and not 1,000 square miles as was estimated by Larsen. A reasonable position is that some of the dikes crystallized from an undifferentiated magma and that these dikes represent the composition of the parent magma. Some of the stocklike bodies, such as those along Beaver Creek, include small bodies of more mafic differentiates. The computed mean composition of all intrusive rocks within and near the stock is close to the composition of the quartz monzonite dikes. This shows that the quartz monzonite and associated rocks along Beaver Creek are most likely offshoots of the same magma that formed the larger bodies of the batholith farther south. According to Ross (1936) this southern part is remarkably uniform in composition, consisting of somewhat calcic quartz monzonite.

The quartz monzonite in the stock along Beaver Creek has a fairly uniform composition. The amounts of major constituents—plagioclase, orthoclase, quartz, and biotite—are nearly constant throughout. In contrast, the amounts of the major constituents of the large body of quartz monzonite in the southern part of the area vary considerably. Along some of the border zones, the amount of orthoclase is very small, yet some of the outcrops about a mile form the contact contain numerous large euhedral orthoclase phenocrysts, which increase the amount of potassium to about 5 or 6 percent, compared with only 4 percent in the main part of the body.

The presence of the potassium-poor border zone of the quartz monzonite batholith (samples Nos. 1121, and 1125) shows that a rock poor in potassium feldspar can be formed by crystallization differentiation from a quartz monzonitic magma. It is possible that the rocks of the quartz diorite series were formed in a similar way and represent early crystallization differentiates of the quartz monzonitic magma. Moreover, only one set of variation diagrams can be drawn for most of the major and minor elements. Some of the curves branch, however, which indicates different behavior of certain elements in some parts of the series. Such elements are calcium and vanadium, which are enriched in the mafic members of the quartz diorite series, and potassium, which is considerably enriched in the silicic rocks of the quartz monzonite series. The general similarity of distribution of elements certainly indicates a close relationship, if not a common origin, of both series.

The field relations show that the large body of quartz diorite around Pierce and some small bodies of tonalite to the northwest were crystallized from mobile magmas. If these rocks crystallized from a quartz monzonitic magma, all of the residual liquid rich in potassium was removed before the final solidification. The largest body of quartz diorite could represent a potassium-poor border facies of the quartz monzonitic batholith but the dikes and sill-like bodies of tonalite indicate that a magma poor in potassium existed and that potassium separated from it before intrusion. One of the possible processes leading to such a separation would be early crystallization and accumulation of plagioclase, quartz, hornblende, and biotite to form a quartz dioritic or tonalitic rock, and later partial remelting of this rock to form a quartz dioritic or tonalitic magma. If a quartz dioritic magma was formed it would be capable of limited crystallization differentiation to produce hornblende gabbro, quartz diorite, and tonalite.

An alternative possibility is that the quartz dioritic magma formed directly through melting of a part of the local crust. If it is assumed that rocks of the Belt series, or older strata below it, were folded down to a temperature level above the melting point of quartz monzonite, first a granitic and then a quartz monzonitic melt would start to form. The first melt would contain the material necessary to crystallize quartz, potassium feldspar, plagioclase, and biotite in eutectic ratios. The composition of the later melt would be similar to that of the quartz monzonitic dikes (No. 1081). As the Belt series along the northwestern border zone of the Idaho batholith is composed of a sequence of common type of garnet-mica schist, biotite gneiss, diopside gneiss with occasional silicated dolomite, quartzite, and other products of metamorphism of normal miogeosynclinal sediments, the chemistry of the formation of magmas can be illustrated by using the average rock of the Belt series as an example of a possible source rock. It has been shown earlier (Hietanen, 1962) that the average rock of the Belt series along the northwestern border zone of the Idaho batholith is richer in calcium, magnesium, and iron, and poorer in potassium than the granitic and quartz monzonitic melt assumed to be forming at depth. At this stage, the rocks next to the reservoir of melt would therefore be enriched in the elements left behind as the melt was forming, such as calcium, magnesium, and iron, and impoverished in potassium and silicon, which were needed for the formation of the quartz monzonitic melt. Thus a zone of enrichment of basic components would envelop the area where the eutectic melt was forming. As shown earlier, the average composition of the basified zone in Headquarters quadrangle and farther west (Hietanen, 1962), is very close to that of the quartz diorite. Thus if the temperature was raised, or pressure released, the basified zone capping the melt may have melted partially and formed a quartz dioritic magma, which was squeezed up along a zone of weakness ahead of the major intrusion of quartz monzonite. This latter alternative would require only one more or less continuous major rise in the temperature, whereas the former alternative postulates a period of cooling and crystallization of quartz dioritic rock before a second heating, during which the quartz dioritic rock would have been remelted and intruded into the country rocks.

A general similarity of composition and distribution of minor elements could be a result of common origin of the magmas as remelted parts of the same formations, perhaps with added material from a basaltic magma. The few differences in the distribution of elements between the two rock series could be due to selective solution at temperatures close to 650°C and above. Because at the beginning of melting, the elements enter solution in eutectic ratios, the country rock must then have been enriched in elements not needed and impoverished in elements needed to form eutectic melt, the granitic magma. Thus the differentiation into two rock series, one rich in potassium and the other poor, could have occurred during the formation of the magmas rather than during their crystallization. During the cooling period, each type of magma may have formed a more or less complete crystallization differentiation series.

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Belt Series in the Region Around Snow Peak and Mallard Peak, Idaho

By ANNA HIETANEN

METAMORPHIC AND IGNEOUS ROCKS ALONG THE NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

GEOLOGICAL SURVEY PROFESSIONAL PAPER 344-E

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METAMORPHIC AND IGNEOUS ROCKS ALONG THE NORTHWEST BORDER ZONE OF THE IDAHO BATHOLITH

BELT SERIES IN THE REGION AROUND SNOW PEAK AND MALLARD PEAK, IDAHO

By Anna Hietanen

ABSTRACT

The Snow Peak and Mallard Peak areas provide a far clearer and more complete account of various geologic events in the northwestern contact aureole of the Idaho batholith than do the Headquarters, Boehls Butte, and Elk River-Clarkia areas studied previously (Hietanen 1962a, 1963a, b). The normal stratigraphic sequence of the five lowest formations of the Belt Series of Precambrian age is well exposed in many localities. This sequence allows correlation with known equivalents farther north with greater certainty than could be done previously. All metamorphic zones are represented in the mapped area described in this report starting with biotite zone in the north and followed successively by zones of garnet, staurolite, staurolite-kyanite, kyanite, kyanite-sillimanite, and sillimanite-muscovite toward the south. Two episodes of recrystallization are evident: a synkinematic and a later postkinematic one during which the temperature was somewhat higher.

Many of the structural features, such as the style of folding and orientation of foliation and cleavage, change with the grade of metamorphism and with the material folded. Structure is complicated by two to three sets of folds, whose axes intersect at angles of 60°–80°. The folds range in size from minute wrinkles to large folds having wave lengths of several miles. Several high-angle faults—most trending north, some trending northwest—interrupt the stratigraphic and structural continuity.

INTRODUCTION

This report is part of a series of petrologic and structural studies of the northwestern contact aureole of the Idaho batholith. The mapped area (pls. 1, 2) is a 12-mile-wide belt through the metamorphic contact aureole of the batholith and has several unique features that make it a key area in the study of stratigraphy, structure, and metamorphism of the Belt Series near the Idaho batholith. A normal stratigraphic sequence of the five lowest formations of the Belt Series is exposed at several localities, allowing much more certain correlation with the equivalents to the north than was previously possible. Many units can be traced for more than 20 miles from zones of low-grade to zones of high-grade metamorphism. The metamorphic history of the contact aureole can be studied in greater detail here than elsewhere because evidence of several episodes of recrystallization is well preserved. Various episodes can be dated in relation to deformation, and the relative changes of their pressure-temperature fields can be determined. Biotite, garnet, staurolite, kyanite, and sillimanite isograds show a uniform increase in the grade of metamorphism southward toward the batholith. Moreover, the area offers excellent opportunity for a study of the style of folding in layers of different bulk compositions under various metamorphic conditions.

The area studied comprises 400 square miles in the southeastern part of Shoshone County and in the north-central part of Clearwater County (fig. 1). It joins the Bungalow area (Hietanen, 1963c) to the south and the Boehls Butte quadrangle and vicinity (Hietanen, 1963a) to the west. The southern part of the mapped area (pl. 1), called the Mallard Peak area in this report, is bounded by long 115°25′ and 115°40′ W. and by lat 46°45′ and 47°00′ N. This area is covered by the following 7—½-minute topographic quadrangle sheets: the eastern part of Buzzard Roost, the Mallard Peak, the western part of Pole Mountain, the Sheep Mountain, and the western part of The Nub.

The northern part of the mapped area (pl. 2), which lies between the St. Joe River and lat 47°00′, is called the Snow Peak area after a prominent mountain that rises 5,000 feet above the canyon of the Little North Fork of the Clearwater River in the southwest. The easternmost part of this area is covered by the Simmons Peak 30-minute quadrangle map.

The topography is characterized by rugged peaks and high ridges separated by steep canyons of the North Fork of the Clearwater River, the St. Joe River, and their tributaries. The highest peaks stand nearly 7,000 feet above sea level; and the altitude of the North Fork of the Clearwater River at Canyon Ranger Station is 1,700 feet. Most of the area has a relief of 3,000–4,000 feet.

Exposures are good along the ridges and on south-facing slopes. North-facing slopes and most creek bottoms are covered by talus and soil. Much of the land surface in the northernmost and southernmost parts of the area and at lower altitudes elsewhere is timbered. During recent years, logging operations have been extended to the northern and southern parts of the area, and several new gravel and dirt roads have been constructed. These roads make parts of the area easily accessible; the rugged central part, however, is traversed only by a few pack trails.

Some preliminary studies were carried out in 1952, 1954, and 1955; most of the fieldwork was done during the summers of 1960–1963. The author was assisted in the field by Mary Lou Conant in 1960, by Barbara Voorhies in 1961, by Seena Nicolaisen in 1962, and by Penny Powell in 1963.

The oldest rocks in the mapped area (pls. 1, 2) are continuous with the Belt Series of Precambrian age to the north and northwest. The five oldest formations in the Belt Series—Prichard, Burke, Revett, St. Regis, and Wallace—have been mapped by previous authors between the Little North Fork of the Clearwater River and the St. Joe River and in the areas to the north (Pardee, 1911; Calkins and Jones, 1913; Umpleby and Jones, 1923; Wagner, 1949). North of the St. Joe River, the rocks of the Belt Series are very similar to those in the Coeur d'Alene area first described by Ransome and Calkins (1908). Toward the south, progressively more intense metamorphism has changed the sedimentary rocks to coarse-grained schist, quartzite, and gneiss, but the chemical composition of the lithologic units, as well as their sequence and thickness, roughly corresponds to that of the less metamorphosed equivalents.

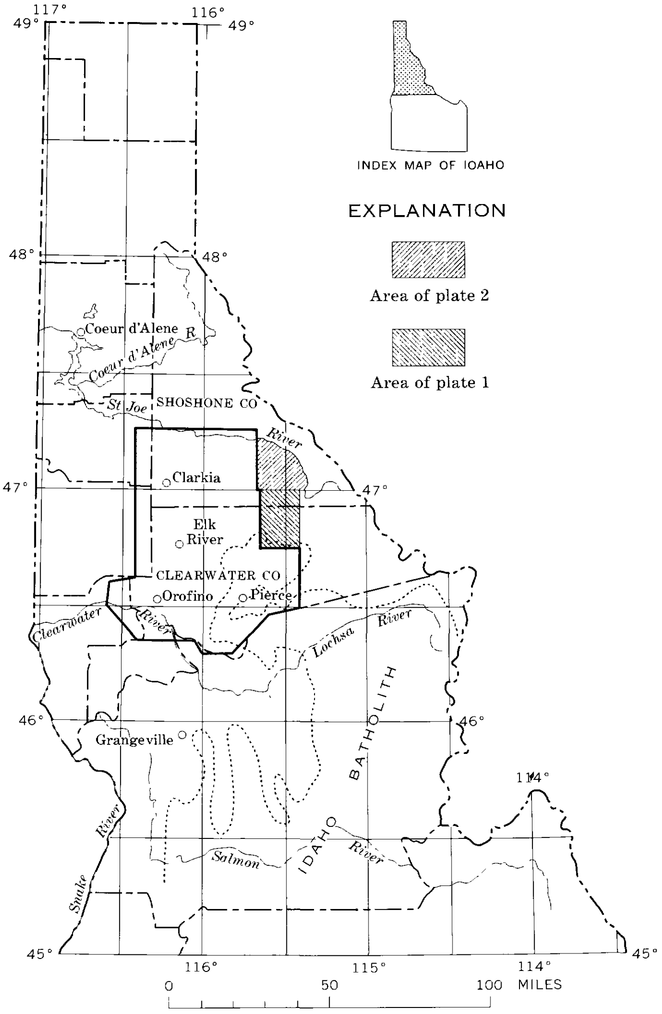


FIGURE 1.—Index map of northern Idaho. Outline of northern part of the Idaho batholith is shown by dotted line. Outline of the area described previously (Hietanen, 1962a, 1963a, b, c, 1967) is shown by heavy line.

The Belt Series is intruded by quartz dioritic, quartz monzonitic, and granitic rocks of Cretaceous age. Those rocks in the southernmost part of the mapped area are part of the northwestern border zone of the Idaho batholith. Dikes and sills of gabbroic, quartz monzonitic, and granitic compositions are later than the plutonic rocks. Silver-bearing galena and chalcopyrite occur in a quartz vein on the east side of the St. Joe River northeast of Red Ives.

METAMORPHIC ROCKS

STRATIGRAPHY AND CORRELATION

The area of plate 2 lies within the southeastern corner of the geologic map of Shoshone County by Umpleby and Jones (1923). They divided the Wallace Formation into three map units; they did not differentiate the St. Regis Formation, and they combined the quartzitic rocks of the Revett and Burke Formations into a single unit. The schist under these rocks is shown as the Prichard Formation.

For the present investigation the Wallace Formation has been divided into a schist unit and a quartzite unit which correspond closely to the upper two units of the Wallace Formation as mapped by Umpleby and Jones. The St. Regis Formation has been mapped separately and consists of a relatively thin group of interlayered schist and micaceous quartzite which lies above the Revett Formation and below the quartzite unit of the Wallace Formation. The Burke and Revett Formations have been mapped separately, and the Burke Formation has been divided into a schist unit and a quartzite unit. Schist of the Prichard Formation underlies the quartzite unit of the Burke Formation.

These units have been traced southward and are mapped separately as shown on plate 1, even though the progressive metamorphism has caused marked changes and obscured many characteristic features. Sedimentary structural features such as crossbedding, channeling, and mud cracks that are distinguishable near the St. Joe River are gradually obliterated by recrystallization southward; however, good marker units such as the thin-bedded quartzite of the Wallace Formation still remain distinctive and have been most helpful in tracing the formations southward. This quartzite contains thin argillitic layers, calcareous beds, and some dolomite which remain distinctive even though their mineralogic makeup is changed and the grain size has coarsened southward. This quartzitic zone is well exposed in the eastern part of the area of plate 1 and continues northward beyond the St. Joe River where the metamorphic effects are only slight.

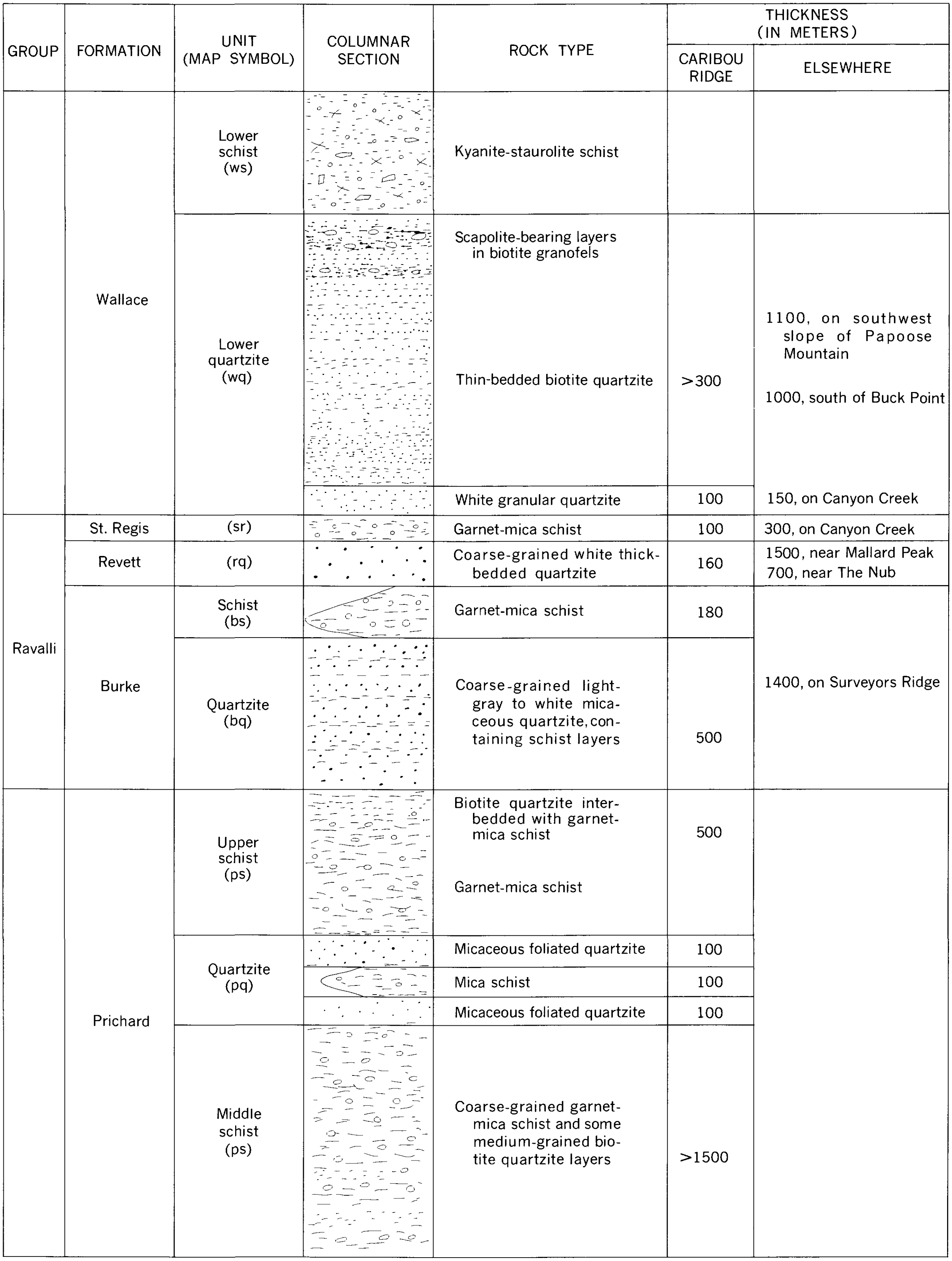
A sequence of coarse-grained quartzite and garnet-mica schist which makes up the Ravalli Group lies conformably under the quartzite unit of the Wallace Formation. The sequence has been differentiated into the St. Regis, Revett, and Burke Formations, units that are probably correlative to those mapped in the Coeur d'Alene district to the north. Because these formations are exposed only in zones of high-grade metamorphism, their general appearance is preserved over the whole area. Most of the quartzite is coarse grained and foliated, and the schist is mainly garnet-mica schist. The grain size changes only slightly and the index minerals, such as kyanite and sillimanite, that are typical of the schist of the Wallace and Prichard Formations are sparse and small in size or are absent. In contrast, the effects of progressive metamorphism on the overlying schist unit of the Wallace Formation are pronounced. Near the St. Joe River this unit has been changed to a fine-grained dark-gray muscovite-biotite granofels (a term coined by Goldsmith, 1959), but near Middle Sister, 3 miles to the south, it contains garnet, and south of East Sister and of Conrad Peak it grades to staurolite schist. Farther south, near Bathtub Mountain, Peggy Peak and Papoose Mountain, the schist is coarser grained and contains kyanite in addition to staurolite and garnet. The grain size becomes coarser south of Pole Mountain, and only kyanite occurs with garnet, muscovite, and biotite. Distribution, stratigraphic position, and thickness of the lithologic sequences of each formation are described separately in the following sections.

PRICHARD FORMATION

About 2,300 m (7,000 ft) of the Prichard Formation, which lies conformably below Burke quartzite, is exposed in the mapped area. About 1,500 m of schist is exposed in the eastern canyon wall of the Little North Fork between Sawtooth Creek (pl. 1) and Canyon Creek (pl. 2). Above this schist, two quartzite layers and an intervening schist layer are well exposed below Spotted Louis Point and along Canyon Creek. Each quartzite layer and each schist layer is about 100 m thick. The upper quartzite layer is overlain by another unit of the schist, 400–600 m thick. Near Black Mountain (pl. 1), two north-trending faults interrupt the normal sequence. The quartzite layers in this vicinity consist of coarse-grained slightly granular quartzite similar to the quartzite layers near the Little North Fork of the Clearwater River. The continuity between these two occurrences is interrupted by faulting near Larkins Peak and by second folding and erosion along the canyon of Isabella Creek. From Black Mountain the two quartzite layers continue southward to the north slope of the canyon of the North Fork of the Clearwater River between Lost Pete Creek and the Twin Creeks and are interrupted again by faulting and erosion. The intervening schist was not mapped in the southernmost part of the area. In addition to the quartzite units shown on the maps, thin layers of white to light-gray quartzite are interbedded with the schist just under and just above the major quartzite units. Some of these layers are discontinuous and form lens-shaped bodies, 1–10 m long and 10–20 cm thick. Such discontinuous layers are well exposed in the vicinity of Black Mountain. Just west of the area shown on plate 1, on the eastern canyon wall of the Little North Fork of the Clearwater River between Devils Club Creek and Minnesaka Creek, the lower schist unit is underlain by another unit of coarse- to medium-grained strongly foliated very light gray to white micaceous quartzite, similar to the unit along Canyon Creek. This unit most likely is a structural repetition of the quartzite along Canyon Creek because it conformably overlies a thick middle schist unit that in turn overlies, in the Boehls Butte quadrangle, a still lower quartzite unit in the Prichard Formation (Heitanen, 1963a). Gabbro and amphibolite that are exposed on a ridge about 2 miles west of Devils Lake, just east of this quartzite, probably conceal a fault (Hietanen, 1967).

RAVALLI GROUP

The sequence of coarse-grained foliated quartzite and garnet-mica schist that overlies the upper schist unit of the Prichard Formation and that underlies thin-bedded granular quartzite of the Wallace Formation is divisible into stratigraphic units that, although metamorphosed, can be correlated on the basis of lithology with the Burke, Revett, and St. Regis Formation. The lower part of the Burke Formation consists principally of coarse-grained foliated fairly thin bedded micaceous quartzite containing some thick pure quartzite beds. The upper part, where present, is a garnet-mica schist. The Revett Formation is mainly coarse-grained slightly foliated thick-bedded quartzite containing minor micaceous layers. The St. Regis Formation is mainly schist and has some micaceous quartzite. A very light gray medium-grained slightly micaceous weakly granular layer of quartzite is interbedded near the middle of the formation. The thicknesses of many units vary consid-



GROUP	FORMATION	UNIT (MAP SYMBOL)	COLUMNAR SECTION	ROCK TYPE	THICKNESS (IN METERS)	
					CARIBOU RIDGE	ELSEWHERE
	Wallace	Lower schist (ws)		Kyanite-staurolite schist		
		Lower quartzite (wq)		Scapolite-bearing layers in biotite granofels Thin-bedded biotite quartzite	>300	1100, on southwest slope of Papoose Mountain 1000, south of Buck Point
				White granular quartzite	100	150, on Canyon Creek
Ravalli	St. Regis	(sr)		Garnet-mica schist	100	300, on Canyon Creek
	Revett	(rq)		Coarse-grained white thick-bedded quartzite	160	1500, near Mallard Peak 700, near The Nub
	Burke	Schist (bs)		Garnet-mica schist	180	1400, on Surveyors Ridge
		Quartzite (bq)		Coarse-grained light-gray to white micaceous quartzite, containing schist layers	500	
	Prichard	Upper schist (ps)		Biotite quartzite interbedded with garnet-mica schist Garnet-mica schist	500	
		Quartzite (pq)		Micaceous foliated quartzite	100	
				Mica schist	100	
				Micaceous foliated quartzite	100	
		Middle schist (ps)		Coarse-grained garnet-mica schist and some medium-grained biotite quartzite layers	>1500	

Figure 2.—Generalized stratigraphic section of the Belt Series near Snow Peak.

erably. It is noteworthy that the degree of foliation of the quartzite units increase steadily from the Wallace Formation downward to the Prichard Formation. The grain size, however, is coarsest in the Revett Formation.

The same sequences are well exposed both in the southwestern part of the Snow Peak area (pl. 2) and in the central part of the Mallard Peak area (pl. 1). The best continuous sections of the Ravalli Group are exposed on Caribou Ridge (just west of Caribou Creek, pl. 2), near Sawtooth Peak, and in several localities south of Mallard Peak and near The Nub. On Surveyors Ridge and on the south side of Sawtooth Creek, an overthrust and several high-angle faults interrupt the normal sequence.

In the section exposed on Caribou Ridge and on the steep slope at the southern end of this ridge (fig. 2), the upper schist of the Prichard Formation is overlain by a 500-m unit of coarse-grained micaceous quartzite, in which layers of white to light-gray quartzite from 1 to 20 m thick are separated by schist layers 5 cm to 1 m thick. In the quartzite layers, the individual beds, 10–20 cm thick, are separated by micaceous laminae. Some thicker beds of coarse-grained pure quartzite occur in the lowest part. This quartzite was mapped as the lower part of the Burke Formation; the upper part of the formation is garnet-mica schist, about 180 m thick. The total thickness of the Burke Formation is about 680 m. The schist of the Burke Formation is overlain by thick-bedded white to light-bluish-gray coarse-grained quartzite, the lithology and stratigraphic position of which are equivalent to those of the Revett Formation. The thickness of the Revett Formation on Caribou Ridge is about 160 m, which is less than elsewhere in the mapped area. The Revett Formation is overlain by about 100 m of garnet-mica schist that is mapped as the St. Regis Formation. Most of this schist on Caribou Ridge is medium-grained gray muscovite-biotite schist in which micas are well oriented parallel to the foliation.

West of Caribou Ridge the Ravalli Group thins and its lithology changes somewhat. Along the ridge northwest of Spotted Louis Creek, the St. Regis Formation consists of schist similar to, and about as thick as, that on Caribou Ridge. The Revett Formation consists of about 200 m of white to light-gray coarse-grained quartzite. The quartzite of the Revett Formation is separated from the garnet-mica schist of the upper schist unit of the underlying Prichard Formation by about 200 m of medium-grained gray very micaceous quartzite. Thus it seems that the Burke Formation is much thinner on the west side of the large fault that extends from Spotted Louis Creek southward. The schist of the Burke Formation that separates the quartzite of the Revett and Burke Formations on Caribou Ridge thins and disappears on the east slope of Snow Peak.

The belt of Ravalli Group rocks is cut by several faults southeast of Snow Peak. Only parts of the normal sequence are exposed in many fault blocks, but a complete section occurs on the slopes of Sawtooth Peak. In this section the quartzite unit of the Burke Formation is excellently exposed on the south slope of Sawtooth Peak (pl. 2) and along Sawtooth Creek (pl. 1). It is lithologically similar to quartzite of the Burke Formation on Caribou Ridge, but its thickness is about three times greater. Two thick schist layers (30–100 m) and several thinner ones (only the thickest is shown on maps) are interbedded with micaceous quartzite layers that range from 10 to 150 m in thickness and are distinctly bedded and well foliated.

This quartzite unit is overlain by coarse-grained garnet-mica schist that contains layers of thin-bedded biotite quartzite. Most of the north slope of Sawtooth Peak is covered by thick soil; a few outcrops and talus of thick-bedded white to very light gray coarse-grained quartzite, mapped as the Revett Formation, occur along streams that are tributary to Canyon Creek. Schist exposed along Canyon Creek to the north is stratigraphically above the Revett Formation and was mapped as the St. Regis Formation. Thin-bedded biotite quartzite and diopside-actinolite quartzite exposed on the north side of Canyon Creek between Buck and Caribou Creeks overlie the schist of the St. Regis Formation and are lithologically similar to the rocks of the quartzite unit of the Wallace Formation.

The white to light-gray coarse-grained quartzite forming the rugged peaks in the vicinity of Heart Lake (pl. 1) and Wasset Peak (cross section *B–B′*, pl. 1) has been mapped as the Revett Formation. These rocks are in part micaceous and contain a few thin layers of schist, but they also include much massive pure quartzite typical of the Revett Formation. A layer of schist separates this quartzite unit from underlying more micaceous quartzite exposed south and west of Heart Lake. These two units are considered to make up the Burke Formation. They are underlain by schist of the Prichard Formation.

A thick-bedded white to light-gray quartzite on the ridge west of Skyland Lake and along Northbound Creek (pl. 1) also was mapped as the Revett Formation. This quartzite is underlain by layers of schist and foliated light-gray micaceous quartzite of the Burke Formation which are exposed on the ridge west of Mallard Peak. At Martin Peak and along Sawtooth Creek, the Revett Formation is overlain by coarse-grained garnet-mica schist of the St. Regis Formation.

A similar sequence of coarse-grained quartzite and schist is exposed near Fawn Lake (pl. 1) and to the south. Near Fawn Lake most of the quartzite of the Revett Formation is very pure, coarse grained, thick bedded, and only slightly foliated. Similar beds continue on ridges north and east of Mallard Peak and on the ridge between Heather and Avalanche Creeks where the thickness of the Revett Formation is about 1,500 m. Toward the south, on Avalanche Ridge and in the vicinity of The Nub, about 700 m of the Revett Formation is exposed, but the structural features suggest that this section is thickened by folding and that the true thickness is about the same as that on Caribou Ridge.

The Revett Formation south of Mallard Peak and near The Nub is underlain by schist and quartzite units of the Burke Formation; each unit is about 250–300 m thick (pl. 1, cross section *A–A′*) and is overlain by schist of the St. Regis Formation, which is 600–700 m thick near Collins Peak. Much of the quartzite of the Burke Formation south of Mallard Peak resembles the Revett Formation except for a somewhat stronger foliation and locally more interbedded micaceous material. Several thin-bedded biotite gneiss layers are interbedded in the schist of the Burke Formation on the northern slope of Mallard Peak.

The schist of the St. Regis Formation east of The Nub is thickened by folding and faulting. Several large folds, one on top of another and all strongly overturned to the east, mark a fault on the steep eastern slope of The Nub. A layer of white to light-gray quartzite, about 20 m thick, is interbedded with the schist east of this fault. Toward the north, south of Collins Peak, the formation seems thinner, and the dips are consistently to the east. A layer of fairly coarse pure quartzite is also interbedded with the schist here as shown by exposures on the ridge southwest of Collins Peak and on a hill north of Heather Creek. Along Sawtooth Creek the total thickness of the schist indicates repetition of the strata by faulting.

The rocks of the Ravalli Group are also exposed along Quartz Creek in the southeastern part of the area (pl. 1) and along the contact of the granite in the southernmost part of the area. The exposed thickness of the Revett Formation on the north side of Quartz Creek is only about 50 m.

WALLACE FORMATION

The rocks of the Wallace Formation are exposed in the northern, eastern, and southern parts of the mapped area. Those in the northern part lie in a normal sequence above the Ravalli Group, but in the eastern and southern parts, faults have disturbed the original sequences. The rocks of the Wallace Formation in the southern part of the area, as in the vicinity of Eagle Point, are highly metamorphosed and isolated from the other occurrences either by faults or by intrusive bodies. The rocks consist mainly of diopside gneiss and have some biotite-plagioclase gneiss and are similar to the gneiss of the Wallace Formation in the Headquarters and Boehls Butte quadrangles (Hietanen, 1962a, 1963a) and in the southern part of the Elk River-Clarkia area (Hietanen, 1963b).

Several sections through the lower quartzitic part of the Wallace Formation are well exposed in the northern part of the area along Canyon Creek south and southwest of Papoose Mountain and south of Buck Point. The section southwest of Papoose Mountain consists mainly of thin-bedded biotite quartzite in which gray fine-grained biotite-bearing beds, 1–10 cm thick, alternate with very light gray to white granular quartzite beds in which the amount of biotite and muscovite is less. The lowest part of the section consists of white granular quartzite in which beds 5–30 cm thick are separated by thin micaceous layers. Some calcareous layers and some hornblende- or actinolite-bearing layers are interbedded. The thickness of the quartzite unit south of Buck Point is about 1,000 m and south of Papoose Mountain, 1,100 m. The amount of biotite and muscovite increases toward the north; scapolite is abundant in dark biotite-rich layers of the upper part of the unit which are exposed along Bluff Creek, its tributaries (Hietanen, 1967), and in several localities south and southeast of the St. Joe River. Generally the scapolite-rich layers are just under the argillitic unit.

In the eastern part of the area, the quartzite unit of the Wallace Formation is well exposed along Collins Creek (pl. 1). Around the headwaters of this creek, the unit consists mainly of thin-bedded biotite quartzite underlain by white granular quartzite. Layers rich in scapolite and biotite, or in scapolite and diopside, are common south of the mouth of Spud Creek, where strong folding and faulting have obscured the stratigraphic sequence. Scapolite-bearing layers also occur east of Perry Creek and southeast of the mouth of Collins Creek. Carbonate layers are interbedded with the quartzite near Granite Peak and at the mouth of Collins Creek.

In the north near the St. Joe River, thin-bedded argillite containing some interbedded quartzitic layers overlies the quartzite unit of the Wallace Formation; toward the south, this unit grades into coarse-grained schist through progressive zones of regional metamorphism. The thickness of the schist unit cannot be determined because the top has been removed by erosion. The exposed thickness is at least 400 m.

In the southeastern part of the area, between Quartz Creek and Bald Knob, the quartzitic unit of the Wallace Formation is an interbedded sequence of thin-bedded biotite quartzite, biotite gneiss, and diopside gneiss which conformably overlies schist of the St. Regis Formation. The white granular beds, which form a considerable part of this lower unit along Canyon Creek, are missing in this section; probably more clayey material was originally deposited with the quartz sand in the lower part of the formation near Quartz Creek. Toward the north, beds of white or very light gray quartzite, 5–20 cm thick and separated by thinner micaceous beds, are interbedded with thin-bedded gray quartzite; they are a conspicuous feature of the lower part of the quartzite unit exposed along Collins Creek north of the mouth of Spud Creek.

Quartzite is exposed on the west side of Collins Creek between Cliff Creek and Collins Peak and extends northwest to Surveyors Ridge. This thin-bedded gray granular quartzite shows stronger deformation, is coarser grained, and contains layers that seem more micaceous than the rest of the Wallace Formation in this area. This quartzite includes a few layers of white quartzite, 2–5 m thick, in which the quartz grains are strongly deformed and which contains less muscovite and biotite than the gray quartzite beds. A few thin calcareous beds are intercalated with thin-bedded white to light-gray beds west of Surveyors Peak. Toward the south this unit contains abundant thin-bedded biotite-rich layers and is definitely more thoroughly recrystallized than the rocks of the Wallace Formation on the east side of the Collins Creek fault. The alternating sequence of thin-bedded granular quartzite with micaceous layers and of quartz-rich schist with quartzite layers is lithologically similar to parts of the Wallace Formation near Elk River (Hietanen, 1963b). Between Surveyors Ridge and Surveyors Peak this unit lies under the thick-bedded white to light-gray coarse-grained quartzite that has been mapped as the Revett Formation. Overturned folds near Surveyors Peak and overturned beds near Collins Peak (see p. E25) support the view that the Revett and St. Regis Formations are brought up along an overthrust fault that is just under the Revett Formation (pl. 1, cross section *B–B′*). The gray granular quartzite can thus be a part of the Wallace Formation; its higher degree of recrystallization and deformation is due to its structural position in the lower plate of an overthrust.

PETROGRAPHIC DESCRIPTION

PRICHARD FORMATION

SCHIST

Three metamorphic zones—kyanite, kyanite + sillimanite, and sillimanite—are recognized in schist of the Prichard Formation on the basis of the occurrence of these aluminum silicates in the aluminum-rich layers. Because these index minerals are rare in most of the schist and because the common type of garnet-mica schist is very similar in all zones, all schist is described under the same heading. Most of the schist is a coarse-grained garnet-mica schist containing layers rich either in quartz or in quartz and plagioclase. The color of the surface of a fresh break of a common type of garnet-mica schist is gray, but slabs from weathered outcrops have bronze-colored glittering surfaces due to the combined effect of abundant large flakes of muscovite and weathered biotite. In the quartz-rich layers, micaceous laminae separate the individual thin beds. Quartz and plagioclase have segregated in thin laminae or in thin sheetlike masses that vary greatly in their dimensions and that are especially common in the southern part of the area. In many places such segregation has given rise to the formation of migmatitic rocks.

The major constituents of the schist are quartz, plagioclase (An_{23-27}), muscovite, biotite, garnet, kyanite, and sillimanite. In the micaceous layers, the amount of muscovite is much larger (20–35 percent) than that of biotite (5–20 percent), whereas in the quartz-rich layers (50–65 percent quartz), biotite is the predominant mica (10–20 percent) and the amount of muscovite is less (5–10 percent). Both micas are parallel to the foliation. Plagioclase is common throughout the schist, but most of it occurs in the medium-grained layers that are rich in quartz and biotite.

Garnet is ubiquitous but rarely abundant. The size of garnets generally ranges from 5 to 10 mm, but near Larkins Peak and in the central part of the area, between Mallard Peak and The Nub, garnet crystals 2–4 cm in diameter are common. Abundant dodecahedral garnet crystals 2–3 cm in diameter occur along the transitional contact zone between the quartzite and the underlying schist northwest of Larkins Peak. Many layers in the schist also are rich in garnet. In the southern part of the area, along the North Fork of the Clearwater River between Rock and Quartz Creeks and to the west, many layers of schist are peppered with small euhedral crystals of garnet. These small crystals rarely contain inclusions, whereas the large euhedral to subhedral grains that are common elsewhere include small grains of quartz and some magnetite.

Kyanite abounds locally at the northern end of the exposures of the Prichard Formation. For instance, outcrops of coarse-grained garnet-biotite schist about a mile northwest of Larkins Peak (loc. 2218) are studded with bright-blue euhedral to subhedral crystals of kyanite 2–4 cm long and 4–10 mm wide. In many layers, kyanite amounts to about 35 percent, biotite to about 25 percent, and garnet to about 10 percent. The garnet occurs in subhedral brownish-red crystals 1–2 cm in diameter. The light minerals are quartz and plagioclase in about equal amounts.

Kyanite and sillimanite occur together in garnet-mica schist that is exposed in the vicinity of Black Mountain and west of The Nub. Here small colorless crystals of kyanite are either scattered among muscovite and biotite or are clustered and accompanied by tiny needles of sillimanite. The two aluminium silicates crystallized during the same episode of recrystallization, and there are no signs of disequilibrium of either mineral.

In the southernmost part of the area along the North Fork of the Clearwater River, sillimanite is commonly the only aluminum silicate in rocks of the Prichard Formation; kyanite was found in only one specimen studied (pl. 1, loc. 876), and its occurrence in small elongate grains surrounded by muscovite give the impression of an armored relict. Sillimanite occurs in brown nodules, in colorless needles, and in large clusters of needles and prisms. The nodules consist of a fine mesh of fibrous sillimanite surrounded by larger needles. The nodules are resistant to weathering and form pebbles that are found in the river gravel. Many of these pebbles are green, brown, and yellow and are cut and polished for ornamental purposes by local mineral collectors. Clusters of needles and prisms as much as 20 cm in diameter occur in the schist exposed on the ridge due west from Sneak Point. Sillimanite needles in these clusters are oriented parallel to the lineation. A small amount of quartz, muscovite, and garnet are included among the sillimanite needles.

QUARTZITE

The quartzite of the Prichard Formation is coarse to medium grained, is well foliated, and contains muscovite and biotite. Most occurrences are coarsely granular and very light gray to white or pale brownish gray. Some layers contain large crystals of garnet.

Thin sections show that muscovite-bearing layers may contain as much as 95 percent quartz in strongly deformed grains that are elongate or irregular in shape and 1–7 mm in length. Muscovite is in flakes 0.5–1 mm long included in or interstitial to quartz grains. Ilmenite-magnetite, pyrite, zircon, and brown grains of rutile occur as accessory minerals. The muscovite-bearing layers are separated by thin biotite-bearing layers in which biotite either takes the place of muscovite or occurs with it.

BURKE FORMATION

QUARTZITE UNIT

In the vicinity of Snow Peak and Sawtooth Peak, the quartzite unit of the Burke Formation consists of light-gray distinctly bedded coarse-grained well-foliated micaceous quartzite having some interbedded garnet-mica schist. In the quartz-rich layers (1–20 m thick), many individual beds that range from 3 to 20 cm in thickness consist of 90–99 percent quartz and some muscovite and biotite. The interbedded micaceous layers are generally thinner (1–10 cm) and contain 10–25 percent muscovite and biotite. In some layers, as at locality 1550 (fig. 3), tiny strongly elongated crystals of staurolite and kyanite occur with the muscovite. Small brown grains of rutile, elongate and rounded grains of zircon, and some ilmenite and magnetite occur as accessory minerals. Near The Nub the quartzite of the Burke Formation contains more thick quartzite beds and less mica-rich layers than it does in the northern part, and it thus resembles quartzite of the Revett Formation.

Thin sections show that in all quartzite the quartz grains are strongly deformed—elongated and flattened parallel to the bedding. The grains are 3–10 mm in their longest dimension and 2–3 mm in their shortest. Many grains show Boehm lamellae, and all have strong strain shadows and well-developed orientation. The mica flakes, which are 0.5–2 mm long and less than 0.1 mm thick and thus very small in comparison to the size of the quartz grains, are included in quartz or transect its grain boundaries (fig. 3). In many beds, muscovite and biotite occur in separate thin layers; in some thicker layers the muscovite may be scattered throughout. Commonly the biotite forms paper-thin laminae that separate the thin layers of muscovite quartzite. The scarcity or absence of biotite in the staurolite-bearing layer (pl. 1, loc. 1550) indicates about the same iron content in each layer but less potassium and magnesium in the staurolite-bearing layer.

The schistose mica-rich layers (30 cm–10 m thick) that are interbedded with the quartz-rich layers contain 5–25 percent muscovite and biotite, 40–70 percent quartz, and 20–50 percent plagioclase (An_{20-27}). The layers rich in plagioclase are mineralogically similar to the gneissic layers in schist of the Prichard Formation. The layers rich in micas and plagioclase are finer grained than those consisting mainly of quartz. The flakes of muscovite are commonly larger (1–2 mm long) than those of biotite (0.5–1 mm long). Plagioclase and quartz occur in equant or slightly elongate grains with irregular or sutured borders; the grains range from 0.2 to 1 mm in size.

SCHIST UNIT

The schist unit of the Burke Formation consists mostly of coarse-grained garnet-mica schist that in the upper part contains some interbedded layers of medium-grained micaceous and plagioclase-bearing quartzite. This schist is similar to the garnet-mica schist of the Prichard Formation; it consists of quartz (40–60 percent), some plagioclase An_{20-27} (2–5 percent), muscovite (15–20 percent), biotite (10–15 percent), and garnet (0–20 percent). Muscovite generally occurs in larger flakes (2–3 mm long) than the biotite (1–2 mm long). The crystals of garnet, ranging from 2 to 6 mm in diameter, are more common in the lower part of the unit.

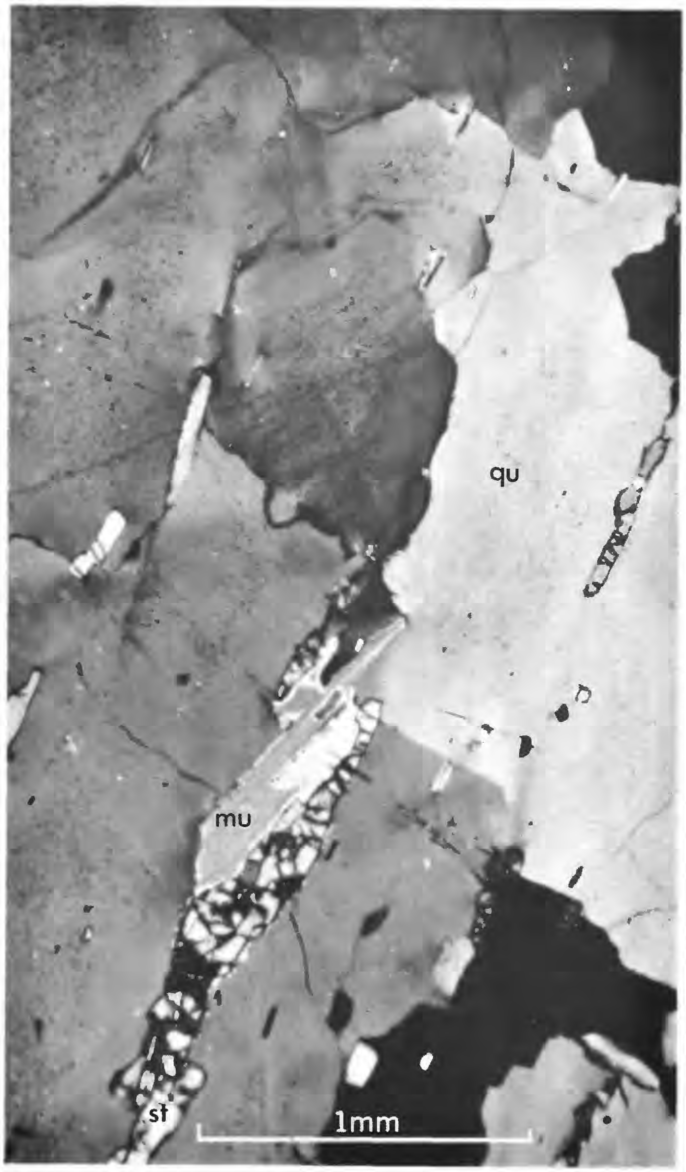


FIGURE 3.—Photomicrograph of quartzite of the Burke Formation. Small mica flakes (mu) and staurolite (st) crystals are parallel to the bedding and are included in large quartz grains (qu). Sawtooth Creek (loc. 1550).

Abundant garnet and some kyanite and sillimanite are contained in the schist that is exposed on the ridge south-southwest of The Nub. This schist is coarse grained and very rich in muscovite and biotite. The garnet crystals are 1–4 cm in diameter, brownish red, and have $n=1.801\pm0.002$. The crystals of kyanite are pale blue to colorless and rather small (2–5 mm long). Needles of sillimanite occur in small clusters or are included in garnet and micas. Ilmenite-magnetite and brown rutile are accessory minerals.

REVETT FORMATION

Much of the quartzite of the Revett Formation is thick bedded, white to very light gray, and coarse grained. It contains a total of only 1–10 percent muscovite, biotite, and plagioclase, mostly in paper-thin laminae, between the beds that consist of large grains of quartz. Locally, layers similar to those in the Burke Formation are interbedded; these layers contain 5–15 percent muscovite and biotite combined. The quartz grains in typical layers in the Revett Formation are 5–8 mm long and show undulatory extinction. Plagioclase (An_{10}) is in discontinuous laminae of small elongate grains (0.2–0.3 mm long) between the large quartz grains. Small flakes of muscovite, 0.1–0.3 mm long and well oriented parallel to the foliation, are sparsely scattered through the rock. They are either interstitial or included in the quartz grains; many transect the grain boundaries, some form thin laminae with small flakes of biotite. A chemical analysis of quartzite of the Revett Formation near Surveyors Peak has been published earlier (Hietanen, 1961a, table 1, no. 1558).

ST. REGIS FORMATION

The St. Regis Formation consists of medium- to coarse-grained mica schist and fine- to medium-grained gray micaceous quartzite. Garnet crystals, 0.5–2 cm in diameter, are exceptionally numerous in the lower part of the formation on Surveyors Ridge and along Sawtooth Creek. Muscovite and biotite, mostly in flakes 1–3 mm long, constitute 35–40 percent of many beds. In some layers, as in those 0.5 mile east of Surveyors Peak and 0.3 mile east of The Nub, round thick single crystals of muscovite (1–2 cm in diameter) are embedded in medium- to coarse-grained schist. The large single crystals of muscovite may be pseudomorphs after staurolite. The schist near The Nub contains about 10 percent plagioclase (An_{27}) and that near Surveyors Peak about 30 percent. In the schist just east of The Nub, many large crystals of garnet include long, slender needles of sillimanite. Some of these garnets are partly altered to chlorite. Small crystals of staurolite, most of them shelled by kyanite, are included in muscovite that envelops the garnet crystals. Sillimanite is abundant in the schist exposed along the North Fork of the Clearwater River north of the mouth of Quartz Creek.

A layer of light-gray medium- to coarse-grained quartzite having a low mica content is interbedded with the schist east of The Nub. Some layers are granular, resembling quartzite of the Wallace Formation; most, however, are coarser grained and more strongly foliated than any of the quartzite of the Wallace Formation. The textural features are intermediate between those of the quartzite typical of the Burke and Wallace Formations.

WALLACE FORMATION

QUARTZITE UNIT

The quartzite unit in the northern part of the mapped area is composed of four distinctive rock types: white granular quartzite, gray thin-bedded biotite quartzite, biotite granofels, and carbonate granofels. The lowest part of the unit consists mainly of white granular quartzite interbedded with thinner light-gray granular layers. The white layers consist of about 95 percent quartz, 3 percent muscovite, and 2 percent albitic plagioclase. Small grains of zircon and brown rutile occur as accessory minerals. The quartz grains are equidimensional or elongated parallel to the foliation and are 1–2 mm in diameter. The borders are sutured or the grains are angular and have round corners. Most grains show undulatory extinction and include tiny (0.2–0.5 mm long) flakes of muscovite. The gray layers interbedded with the white granular layers contain more biotite (3–5 percent) and albite (15–30 percent). Laminae consisting of large flakes of muscovite and biotite are common. Thus the grain size in the white granular quartzite of the Wallace Formation is considerably smaller than that in the thick quartzite beds of the Revett Formation.

The thin-bedded gray quartzite consists of alternating layers (1–10 cm thick) of light-gray quartz-rich rock and darker gray micaceous rock. The amount of minerals in the quartz-rich layers varies within the following limits: 80–90 percent quartz, 5–10 percent albitic plagioclase, 3–10 percent muscovite, 1–5 percent biotite, and 0–5 percent potassium feldspar. Small rounded grains of magnetite, zircon, and brown rutile occur as accessory minerals. The grains of quartz are round to polygonal and 0.2–1 mm in diameter. Feldspar occurs in grains of the same size or is interstitial. Mica flakes are 0.5–1 mm long. Muscovite and biotite in equal amounts make up about 30 percent of the micaceous layers, and the flakes are larger (1–1.5 mm long).

Some thick layers of biotite granofels that contain scapolite-bearing beds (Hietanen, 1967) are interbedded with thin-bedded gray quartzite east of Bathtub Mountain, along Bluff Creek and its tributaries, and along the St. Joe River. All these layers are fine grained; the quartz grains and biotite flakes are 0.1–0.2 mm in diameter. Scapolite, however, is in large round to subhedral grains that range from 1 to 15 mm in diameter and include numerous tiny round quartz grains and some mica flakes. The amount of biotite in some layers is as much as 60 percent, but in most it ranges from 10 to 40 percent. The amount of scapolite, where present, is 20–30 percent. Epidote minerals are more common in the biotite granofels than in the quartzite.

A few discontinuous carbonate-rich layers that also contain scapolite are interbedded in the upper part of the quartzite unit of the Wallace Formation south of East Sister and on the ridge between the forks of Bluff Creek (pl. 2). The proportions of the major constituents—calcite, dolomite, quartz, scapolite, and biotite or phlogopite—vary considerably. Most scapolite is either in the biotite-rich layers or in the carbonate-rich layers in amounts that range from 10 to 20 percent. The quartz content ranges from 10 to 40 percent, that of phlogopite or biotite from 5 to 40 percent, and that of carbonate from 10 to 70 percent.

In the contact aureole of the quartz monzonite that is exposed about a mile west of Dismal Lake, carbonates in feldspathic quartzite have reacted with quartz to form dark-green amphibole and light-green diopside. The indices of refraction of the amphibole measured in specimen A–272 are $\alpha=1.622\pm0.001$, $\beta=1.634\pm0.001$, $\gamma=1.644\pm0.001$ and $-2V\approx70°$, which values indicate it to be actinolitic hornblende. The diopside shows $\alpha=1.674\pm0.001$, $\beta=1.681\pm0.001$, $\gamma=1.705\pm0.001$. These optical properties indicate that both the actinolitic hornblende and diopside are magnesium-rich varieties. Biotite that has $\gamma=1.622\pm0.001$ is a common constituent of these layers. Calcareous layers in the same contact zone farther south contain abundant diopside and some tremolite (Hietanen, 1961a, table 1, no. 1541). Layers rich in scapolite are interbedded. Away from the contact, diopside is absent and only hornblende, tremolite, or actinolite occur in the calcareous layers as in several localities along the tributaries of Bluff Creek. For example, dolomite-bearing layers 2 miles east of Dismal Lake contain light-green actinolite prisms that show $\alpha=1.614\pm0.001$ and $\gamma=1.640\pm0.001$ (loc. A–290). Monomineralic discontinuous thin layers consisting of large tremolite prisms having $\alpha=1.609\pm0.001$ and $\gamma=1.633\pm0.001$ are interbedded with white granular quartzite exposed half a mile farther to the south (loc. A–291).

Zoisite in small grains 0.3–0.5 mm long is common in many carbonate- and actinolite-bearing layers that are interbedded with the biotite quartzite east of Pineapple Peak and north of Badger Mountain. In sec. 29, T. 43 N., R. 8 E., slender prisms of zoisite constitute 10–15 percent of several layers of thin-bedded biotite-rich quartzite. Indices of refraction of this zoisite measured in specimen 2221 are $\alpha=1.696\pm0.001$ and $\gamma=1.701 \pm0.001$. Sphene is abundant in paper-thin laminae that consist mainly of zoisite and some quartz, that are interbedded with biotite quartzite layers, 2–3 mm thick. A few layers of this biotite-zoisite quartzite contain scapolite (mizzonite with $\epsilon=1.549\pm0.001$, $\omega=1.576 \pm0.001$) in addition to quartz, biotite, and zoisite. Biotite shows $\gamma=1.611\pm0.001$. Magnetite, brown rutile, zircon, and apatite occur as accessory minerals in all layers. Small amounts of plagioclase are common in the actinolite-bearing layers.

Light-green actinolite ($\gamma=1.640$) and dark-green actinolitic hornblende ($\alpha=1.626\pm0.001$, $\beta=1.638 \pm0.001$, $\gamma=1.647\pm0.001$ measured in specimen 2085) are typical ferromagnesian minerals in the calcareous layers of the quartzite unit of the Wallace Formation near Buck Point and about 2 miles south of Pole Mountain.

Near Elk Prairie and Granite Peak and to the west, calcareous layers in the quartzite contain abundant dark-green prisms of actinolitic hornblende that are very pale green and show only a weak pleochroism under the microscope. In places this rock contains about 60 percent plagioclase (An_{27}) in large grains that include numerous tiny round grains of quartz. Epidote in small grains makes up about 1 percent of this rock.

Discontinuous layers and lenses of dolomite shelled by light- and dark-green amphiboles are interbedded with thin-bedded gray quartzite north of Granite Peak. The dolomite is coarse grained (2–3 mm in diameter), reddish beige and weathers to dark-brownish-red soil. The index of refraction ($\omega=1.700\pm0.002$) indicates that it is parankerite. The light-green actinolite, which together with calcite forms the inner shell around the parankerite rock, has indices of refraction of $\alpha=1.614 \pm0.001$, $\gamma=1.640\pm0.001$, and $Z\wedge c=20°$ (measured in speciment 2140). The dark-green outer shell consists of amphibole that shows a weak pleochroism (X=colorless, Y=pale green, Z=pale bluish green) and has indices of refraction of $\alpha=1.622\pm0.001$, $\gamma=1.645\pm0.001$, and it is thus probably an actinolitic hornblende.

Toward the south, at the mouth of Collins Creek, the reaction rims separating carbonate layers and thin-bedded quartzite consist of diopside in place of actinolite, shelled by dark-green actinolitic hornblende that is pale to bluish green under the microscope. The indices of refraction of the diopside are $\alpha=1.674\pm0.001$ and $\gamma=1.703\pm0.001$. These indices suggest it is rich in magnesium. The actinolitic hornblende has indices of refraction $\alpha=1.622\pm0.001$ and $\gamma=1.644\pm0.001$ and this refraction is similar to that in specimen 2140. About 2 miles to the north along Collins Creek, diopside showing indices of refraction of $\alpha=1.673\pm0.001$ and $\gamma=1.702 \pm0.001$ occurs as small grains together with actinolite ($\gamma=1.640\pm0.001$), zoisite ($\alpha=1.705\pm0.001$), biotite ($\gamma=1.605\pm0.001$), and scapolite ($\omega=1.582\pm0.001$) in thin-bedded quartzite.

Diopside and actinolitic hornblende are common constituents of calcareous layers in thin-bedded quartzite at the mouth of Buck Creek, at the mouth of Spud Creek, to the south along the lower drainage of Collins Creek, and also southeast of the mouth of this creek. Only diopside has crystallized in the chemically similar layers in the southernmost part of the area where the temperature of recrystallization was higher. Comparison of the distribution of diopside with that of staurolite and kyanite in the schist unit as shown by isograds on plates 1 and 2 suggests that the lower stability limit of diopside is in the kyanite zone near the higher stability limit of staurolite.

In the southern part of the area, rocks mapped as the quartzite unit of the Wallace Formation consist of thin-bedded biotite quartzite, biotite gneiss, and diopside-plagioclase gneiss, all interbedded. These rocks are similar to those mapped as equivalents of the Wallace Formation in the southern part of the Boehls Butte quadrangle (Hietanen, 1963a) and to the west near Elk River (Hietanen, 1963b). Actinolitic hornblende occurs with diopside in some layers. In many localities near the contacts of the intrusive rocks, common hornblende having pleochroism Z=green, Y=light green, X=pale green occurs with or instead of diopside. Sphene instead of rutile is a common titanium mineral in all diopside and hornblende gneiss.

SCHIST UNITS

Argillaceous units of the Wallace Formation show a steady increase in grade of metamorphism toward the south. Five distinct mineralogic zones can be mapped in the argillaceous unit exposed in the northern part of the area: biotite zone, garnet zone, staurolite zone, kyanite-staurolite zone, and kyanite zone. In the northernmost part of the mapped area, near Allen Point and Mount Chenoweth, this unit is argillite in which dark fine-grained layers, 0.2–2 cm thick, alternate with light-gray layers that are a little coarser grained and contain more quartz. Graded bedding is common but not ubiquitous. In the schist exposed in the southernmost part of the area, sillimanite occurs with muscovite, biotite, and garnet.

BIOTITE ZONE

The fine-grained layers consist mainly of quartz and sericite and a few scales of ilmenite and very little biotite. Dark dustlike inclusions, probably carbon and iron oxides, are common in quartz and micas. The sericite is interstitial to quartz and is randomly oriented. Recrystallization has advanced much further in the coarser layers. Flakes of sericite are larger, and biotite forms numerous elongated porphyroblasts 0.2 mm long that are almost perpendicular to the bedding.

Near West Sister and Little Sister, the dark fine-grained layers consist of about 50 percent muscovite, 40 percent quartz, and 5 percent biotite. The muscovite is recrystallized parallel or subparallel to the cleavage that transects the bedding. In the fine-grained layers, grains are 0.01–0.03 mm in diameter, and in the coarser layers are 0.03–0.05 mm. Coarser layers contain about 60 percent quartz and about 30 percent muscovite. Biotite porphyroblasts about 0.1–0.5 mm in diameter occur in both types of layers, but they are more numerous in the coarser grained layers (about 10 percent). The biotite porphyroblasts are either subparallel to the cleavage or oriented at random and contain minute round inclusions of quartz (fig. 4). Thus in this zone, biotite crystallized later than most of the muscovite. Small scales of ilmenite, tiny prisms of tourmaline, a few grains of zircon, and dustlike inclusions of iron oxide are the accessory minerals.

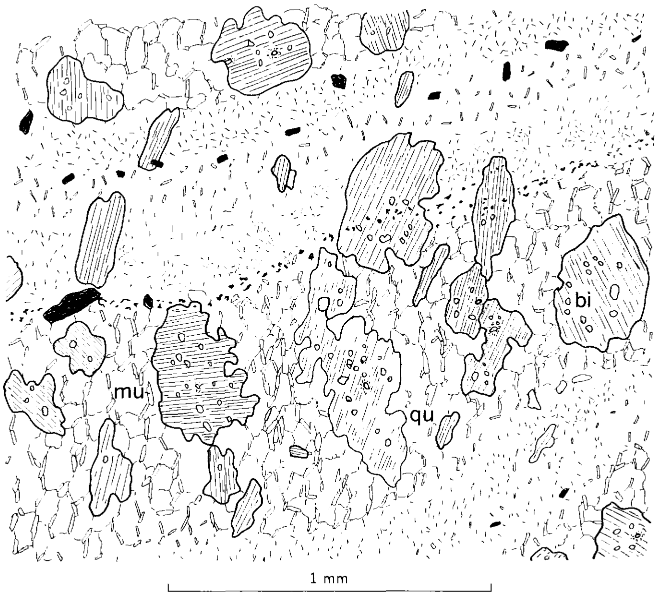


FIGURE 4.—Camera lucida drawing of biotite (bi) porphyroblasts in biotite zone; mu, muscovite; qu, quartz; black is ilmenite. Argillitic unit of the Wallace Formation, West Sister (loc. A–27). Micas are subparallel to cleavage that transects the bedding.

GARNET ZONE

Near Middle Sister all the layers are somewhat coarser grained than the equivalent layers near Little Sister, and tiny euhedral crystals of red almandite and more biotite have crystallized in the argillaceous layers. The flakes of muscovite are 0.02–0.1 mm long, the biotite porphyroblasts ($\gamma = 1.644 \pm 0.002$) are 0.1–0.5 mm in diameter, and the crystals of almandite 0.3–2 mm (fig. 5). Thin sections show that some muscovite flakes are subparallel to the cleavage and some are parallel to the bedding. Porphyroblasts of biotite contain tiny quartz inclusions and are subparallel to the transecting cleavage, which is discernible only in the coarser grained layers with the aid of a hand lens or under the microscope. The garnet has crystallized late, and has replaced all other minerals. The orientation of micas near it is not disturbed; this fact suggests postkinematic crystallization. In the outcrops, cleavage is poorly developed, and the name granofels (Goldsmith, 1959) is best suited for these rocks.

Layers rich in quartz are interbedded within the argillaceous units. Thin beds and lenticular masses (concretions) in some of these layers contain abundant dark-green amphibole. In many localities, as near Middle Sister, the amphibole crystals are much larger (1–2 cm long) than the other mineral grains. They are either perpendicular to the walls of the concretions or oriented at random. This orientation indicates that they crystallized after the deformation. Under the microscope this amphibole shows a fairly strong pleochroism: X = colorless, Y = pale green, and Z = light-bluish green. It has somewhat higher indices of refraction than the typical actinolite: $\alpha = 1.633 \pm 0.001$, $\beta = 1.642 \pm 0.001$ and $\gamma = 1.651 \pm 0.001$.

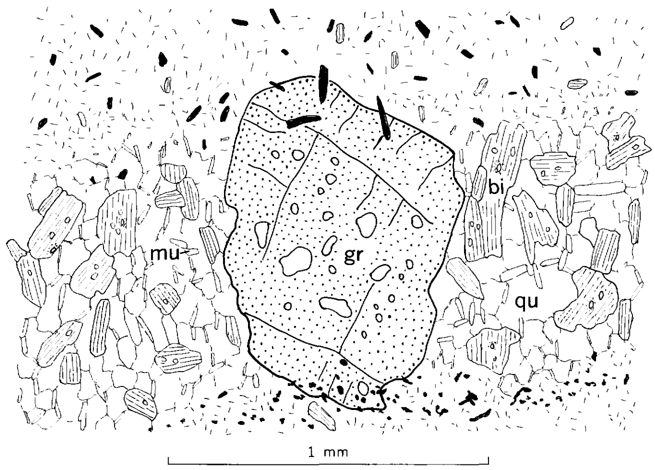


FIGURE 5.—Camera lucida drawing of garnet (gr) in garnet zone; bi, biotite; mu, muscovite; qu, quartz; black is ilmenite and magnetite. Inclusions in garnet are quartz, ilmenite, and magnetite, and their distribution reflects the distribution of these minerals in beds outside the garnet.

The schist unit near Conrad Peak is distinctly foliated, and the grain size is larger than that near Middle Sister. The muscovite flakes are 0.2–0.5 mm in diameter, the biotite flakes 1–2 mm, and the garnets 2–3 mm. Biotite is reddish brown and shows $\gamma=1.640\pm0.002$ and $-2V=0$ (specimen A–157).

In the southeastern part of the area of plate 2 near Red Ives and along the lower drainages of Beaver, Copper, and Timber Creeks, numerous large porphyroblasts and clusters of chlorite, 1–2 cm long and having the shape of a monoclinic mineral, are embedded in many layers of fine-grained garnet-mica schist interbedded with micaceous quartzite. The outlines of these clusters suggest that they are pseudomorphs after chloritoid. Two tiny crystals of chloritoid, however, was found only in one of the thin sections studied. The clusters consist of large flakes of dark-green chlorite that has indices of refraction of $\beta=1.632\pm0.002$ and $\gamma=1.637\pm0.002$ and $+2V\approx30°$.

STAUROLITE ZONE

Staurolite occurs in many layers of the fine-grained garnet-mica schist about 1.3 miles southwest of Conrad Peak and to the south as approximately shown by isograds on plate 2. This zone has been discussed in an earlier publication (Hietanen, 1962b). Most of the biotite in the staurolite zone is intermingled with muscovite; the flakes, about 1 mm long, are parallel to the foliation that in many places transects the bedding. Some of the biotite occurs as early-crystallized, elongate porphyroblasts, 2–4 mm long, that include a few tiny muscovite flakes and are enveloped by muscovite and biotite of medium size. Garnet is red almandite; euhedral crystals, 1–2 mm in size, contain only a few tiny inclusions of quartz and magnetite. Commonly the quartz is concentrated in a zone halfway between the core and the rim. Well-oriented flakes of muscovite and biotite distinctly mark the foliation, which curves around the garnet crystals. This suggests that the deformation continued after the crystallization of garnet.

A different sequence of deformation and recrystallization is indicated by the mineral inclusions in staurolite poikiloblasts. These poikiloblasts are 1–6 cm long and have well-developed m(110), b(010), c(001), and r(201) faces. Many staurolite crystals are twinned, the 60° crosses being most common. They include numerous small grains of quartz, some flakes of muscovite and ilmenite, and a few small crystals of garnet. The included quartz may amount to as much as 30 percent, and the grains are elongated parallel to the foliation. The arrangement of the inclusions reflects the position of bedding and foliation, both of which continue undisturbed through the crystals (fig. 6). In many outcrops east of Peggy Peak, micaceous laminae parallel to bedding are included in staurolite. The orientation of the mica flakes in the enclosed part of the laminae is the same as in the laminae outside the crystals, that is, parallel to the foliation (s_2). These relations show that staurolite crystallized after the deformation and included the preexisting *s* planes.

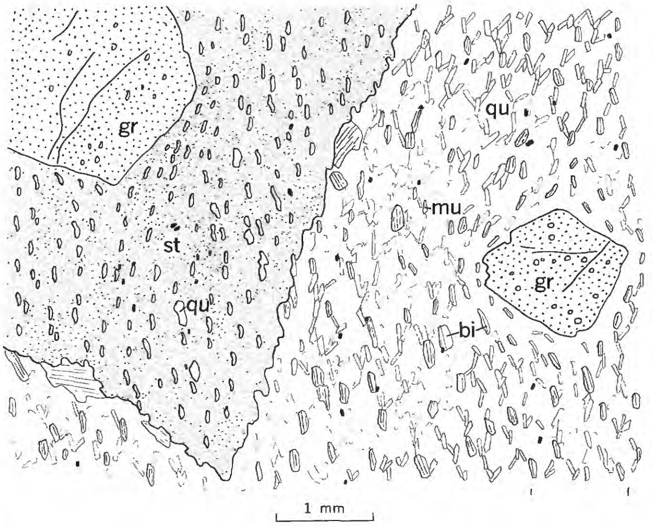


FIGURE 6.—Camera lucida drawing of a postkinematic staurolite (st) in garnet-mica schist of the staurolite zone north of Junction Peak (loc. A–143). Foliation bends around garnet (gr) but continues undisturbed through the late staurolite crystals as shown by the elongated quartz (qu) grains that are included in staurolite; mu, muscovite; bi, biotite.

Near Peggy Peak most of the staurolite has altered to chlorite or to a mixture of chlorite and muscovite. These altered crystals have preserved their euhedral shapes and weather out from the rocks just as their fresh counterparts. In places, all three ferromagnesian minerals present—garnet, staurolite, and biotite—are altered to chlorite. In some outcrops, a few pseudomorphs after staurolite crosses are scattered in the schist that contains pseudomorphs after chloritoid.

Individual outcrops containing kyanite occur at lower altitudes in the staurolite zone as, for example, along the east Fork of Bluff Creek (loc. A–348). The rock in this locality is a fine-grained quartz-biotite-muscovite rock with poikiloblastic kyanite and staurolite, small crystals of garnet, scales of ilmenite, and some tiny grains of brown rutile and green tourmaline.

KYANITE-STAUROLITE ZONE

In the kyanite-staurolite zone two generations of staurolite are evident: synkinematic and postkinematic. In the synkinematic crystals the relict *s* planes in the crystals (*si*) are rotated in relation to the *s* planes of the enclosing rock (*se*). The amount of rotation varies and some show S-shaped curvature of *si*. The *se* bends around these crystals and suggests that deformation outlasted the recrystallization. In all studied sections, foliation of this type is parallel to the bedding. In many specimens taken from the crests of folds with a transecting cleavage, the relict bedding enclosed in the poikiloblastic staurolite seems to form an integral part of the wrinkled s_1 but the relict cleavage in the crystal (si_2) makes an angle with the transecting cleavage in the rock (se_2). This angle indicates that these crystals also are synkinematic and not postkinematic and that the conformity of si_1 and wrinkled se_2 is purely coincidental. In some large staurolite crystals the inclusions are arranged parallel to certain crystallographic faces forming a herringbone structure.

Pseudomorphs after large twinned staurolite crystals are common in a zone southwest of Peggy Peak; this zone continues southward near Papoose Mountain and Badger Mountain. The pseudomorphs consist of a mixture of kyanite, muscovite, small staurolite, and garnet (fig. 7*A*). Some have a zonal structure (fig. 7*B*); the center consists of staurolite or muscovite, and the rims of small radial crystals consist of kyanite and muscovite, which are in turn shelled by a mixture of kyanite, biotite, garnet, and small staurolite crystals. Thus a second generation of staurolite was formed as small crystals with kyanite during a second episode of recrystallization. The textures show that this second episode was postkinematic. Crystallization of kyanite in addition to staurolite indicates that the temperature during the second (postkinematic) episode was higher than that during the first (synkinematic) episode.

At some localities, such as three-fourths of a mile southeast of Bathtub Mountain, two types of staurolite crystals occur in separate beds. Here most of the staurolite occurs as dull large stubby euhedral to subhedral crystals, 3–5 cm long; these are either twinned or clustered and contain abundant inclusions of quartz. However, in some rather thin-bedded layers, numerous small (1–2 cm long) crystals (fig. 8) have well-developed shiny prism faces and resemble the second generation staurolite.

Another striking mineralogic feature of this zone is a steady increase of the amount of kyanite and a decrease of the amount of staurolite toward the south. Near Bathtub Mountain many beds contain as much as 25 percent staurolite and only a few small crystals of kyanite, whereas about a mile southward only a few small staurolite crystals are evident and kyanite is the main aluminum silicate. Here it constitutes 15–20 percent of many beds and occurs in large crystals 3–8 cm long.

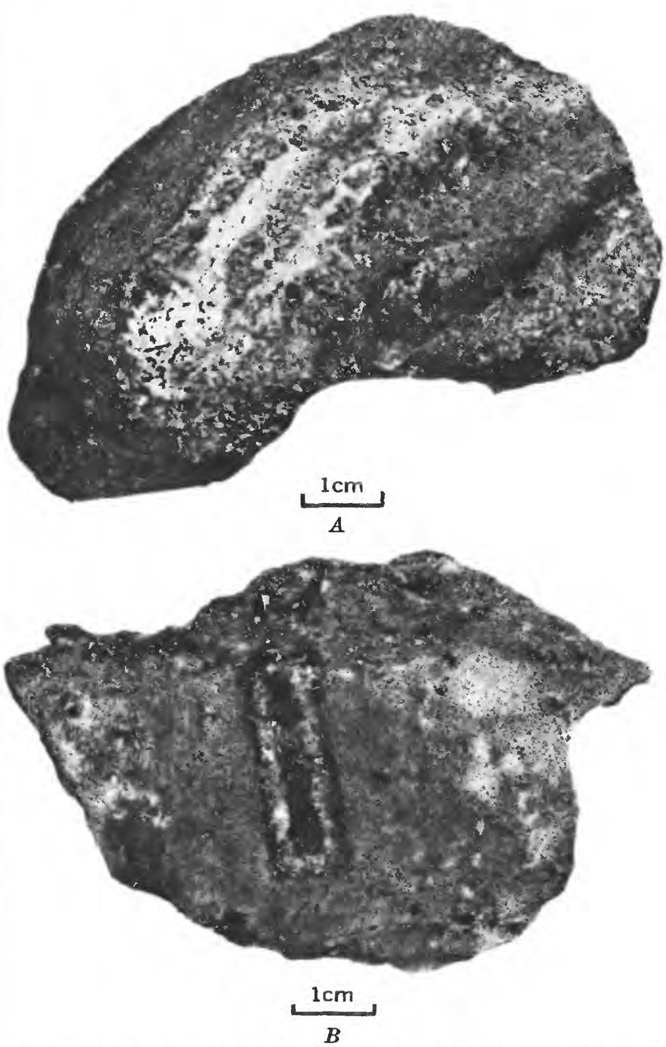


FIGURE 7.—Pseudomorphs after staurolite. *A*, Light-colored central part is a pseudomorph that consists of muscovite, kyanite, garnet, biotite, and some staurolite; Papoose Creek (loc. 2093). *B*, Staurolite crystal altered to an aggregate, the center of which consists of staurolite and garnet and the rim of muscovite and kyanite; west of Peggy Peak (loc. A–156).

Chemical analysis of staurolite schist (from loc. 2096) near Bathtub Mountain is shown in table 1. This schist is from a layer that contains an exceptionally large amount of staurolite in clusters of large crystals. In regrouping the cation percentages to form minerals present in the rock (the molecular mode), all the sodium was used to form albite, the potassium to form biotite and muscovite, and the magnesium to form biotite and garnet; the method of calculation is mainly the same as used by Barth (1959), except that the formulas of various minerals are modified to be in close agreement with their optical properties and with analyzed minerals from similar rocks in the neighboring quadrangles (Hietanen 1961a, p. 90–93). Staurolite was assumed to be an Fe-Al silicate. Biotite has $\gamma=1.625\pm0.001$; no chemical analysis is available as yet, but comparison with analyzed biotites and compositions calculated from rock analyses together with the optical properties suggests an Fe:Mg ratio of 2:3 and Al:Si ratio of about 2.5:5.5 for this biotite. After using this ratio for biotite,

FIGURE 8.—**Staurolite schist a mile southeast of Bathtub Mountain (loc. 2096). Distribution of the euhedral prisms of staurolite follows certain beds.**

TABLE 1.—*Chemical composition in weight and ionic percentage, molecular norm, and mode of staurolite schist from locality 2096*

[Analyst, V. C. Smith; spectrographic determinations, P. R. Barnett]

Weight percent		*Ionic percent*		*Molecular norm*	
SiO_2	64. 55	SiO_2	62. 52	Q	43. 54
TiO_2	. 71	TiO_2	. 52	Ab	5. 95
Al_2O_3	21. 33	$AlO_{3/2}$	24. 35	An	1. 40
Fe_2O_3	3. 30	$FeO_{3/2}$ [1]	. 36	Or	14. 65
FeO	2. 75	FeO [2]	4. 27	C	19. 67
MnO	. 07	MnO	. 06	Fs	6. 90
MgO	1. 81	MgO	2. 61	En	5. 22
CaO	. 38	CaO	. 40	Ap	. 19
Na_2O	. 63	$NaO_{1/2}$	1. 19	Il	1. 04
K_2O	2. 37	$KO_{1/2}$	2. 93	Mt	. 72
P_2O_5	. 08	$PO_{5/2}$	. 07	Crb	. 72
CO_2	. 54	C	. 72		
Cl	. 02			Total	100. 00
F	. 07	Total	100. 00		
H_2O^+	1. 49	O	177. 56		
H_2O^-	. 08	OH	9. 62		
		F	. 23		
Total	100. 18	Cl	. 03		
Less O	. 03				
	100. 15	Total anions	187. 44		

Molecular mode

Quartz	41. 58
Plagioclase An_{19}	7. 35
Muscovite	16. 22
Biotite	15. 65
Garnet	2. 66
Staurolite	23. 96
Apatite	. 19
Ilmenite	. 04
Rutile	. 34
Carbon	. 72
Total	108. 71
$+H_2O$	1. 17
	109. 88

Trace elements, in parts per million

Looked for but not detected: Ag, As, Au, Bi, Cd, Ge, In, Mo, Nb, Pt, Sb, Sn, Ta, Th, Tl, U, W, and Zn]

B	<20	Cr	70	Ni	17	V	80
Ba	560	Cu	14	Pb	20	Y	30
Be	8	Ga	39	Sc	14	Yb	4
Co	15	La	160	Sr	60	Zr	200

[1] Corrected for Fe_2O_3.
[2] Corrected for FeO.

the rest of the iron and magnesium was assumed to go to form garnet; this calculation gave a composition of about 80 percent almandite and 20 percent pyrope for this mineral.

Comparison of the molecular norm of the staurolite-garnet schist with that of the sillimanite-garnet schist of the Wallace Formation in the Headquarters quadrangle (Hietanen, 1962a, table 2, nos. 246 and 247) shows that the crystallization of a large amount of staurolite (23 molecular percent) in rock sample 2096 is not due to an exceptionally high iron-magnesium ratio of the rock. There is very little difference in the amount of biotite between the staurolite- and the sillimanite-bearing rocks, but the amount of garnet is higher in the sillimanite-bearing rocks. Thus it seems that much of the iron that is included in staurolite at the lower grades will form garnet at higher grades.

In the southwestern part of the kyanite-staurolite zone, kyanite occurs as large poikiloblasts that include mainly quartz and some biotite, ilmenite, and magnetite. In many crystals only about 60 percent of the material is kyanite, and the large crystal actually consists of an aggregate of smaller crystals that have a parallel orientation. South of Bathtub Mountain and along Papoose Creek, aggregates of the kyanite crystals form pseudomorphs after large twinned staurolite crystals. Occasionally the rows of inclusions in these aggregates form a herringbone structure that was inherited from the parent staurolite (fig. 9 *A*, *B*). Garnet crystals in this rock are euhedral, 2–4 mm in diameter, and have rims, 0.5–1 mm thick, that contain inclusions of magnetite, ilmenite, and quartz. The orientation of the scales of ilmenite and rows of small grains of quartz in the rims show that they are relicts of the plane of foliation that curved around the early small garnet crystals that are now centers of larger crystals (fig. 9*B*). Flakes of biotite and muscovite butt against the boundaries of these large garnet crystals and are not included in them. These relations show that the clear core of the garnet crystals was formed during the deformation that outlasted the crystallization of the centers and that the rims are postkinematic. Thus, there were two episodes of recrystallization separated from each other by the latest phase in deformation.

KYANITE ZONE

About 1.5 miles southwest of Bathtub Mountain, abundant kyanite without staurolite was formed. This indicates a slightly higher pressure-temperature field toward the south. Kyanite crystals also abound in some layers of garnet-mica schist 1.5 miles south of Pole

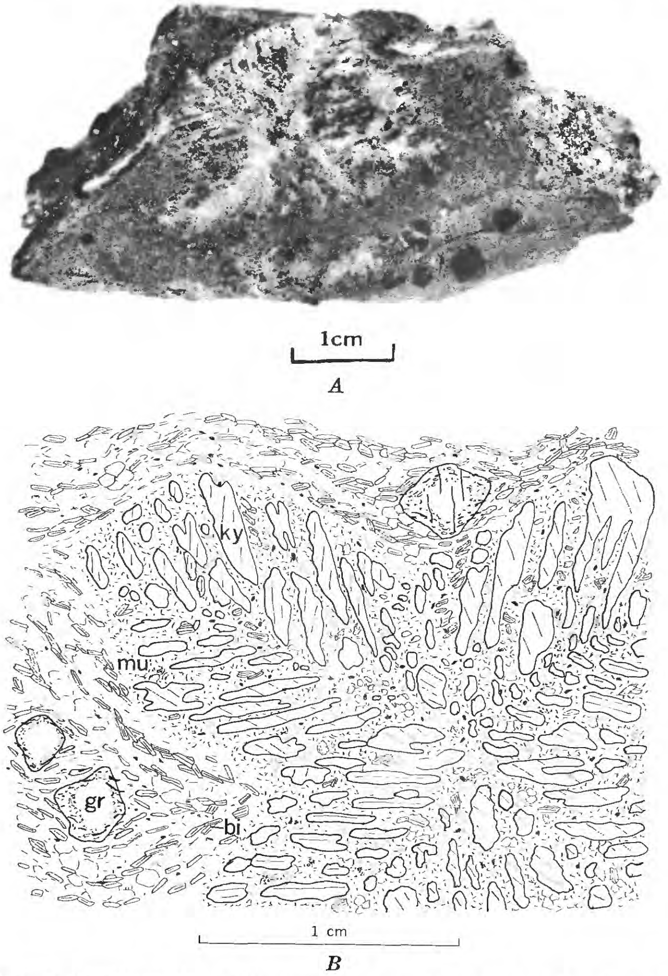


FIGURE 9.—Herringbone structure in a pseudomorph after a staurolite twin. *A*, Photograph of specimen from a mile east of Papoose Mountain (loc. 2104). *B*, Drawing made from thin section of specimen in figure 9*A* shows that the pseudomorph consists of kyanite (ky), muscovite (mu), biotite (bi), and magnetite (black). Garnet crystals (gr) show two episodes of recrystallization: clear cores are surrounded by rims that include magnetite, ilmenite, and quartz. The foliation bends around the cores but is included in the rims.

Mountain. The schist in these localities is fine grained and in it the kyanite forms aggregates of rather small crystals. These aggregates are 1–5 cm long and include abundant quartz but no micas. The orientation of some of the crystals is subparallel to the foliation. This orientation suggests late kinematic crystallization. However, most of the crystals cut across the foliation and are thus postkinematic. Garnet crystals in this rock are euhedral or subhedral and 3–7 mm in diameter.

SILLIMANITE-KYANITE ZONE

No schist of the Wallace Formation is exposed in this zone.

SILLIMANITE ZONE

Schist of the Wallace Formation is interbedded with diopside gneiss, biotite gneiss, and quartzite in the southern part of the area. There the schist is coarse grained and consists of quartz, biotite, muscovite, and garnet with some sillimanite, magnetite, and zircon. This schist is mineralogically similar to schist of the Wallace Formation in the Headquarters quadrangle (Hietanen, 1962a) and in the southern part of Boehls Butte quadrangle (Hietanen, 1963a) and shows the same grade of metamorphism.

METAMORPHIC REACTIONS AND METAMORPHIC FACIES

The mapped area is a representative cross section through the metamorphic aureole around the Idaho batholith. The northern part of the quartz dioritic border zone of the batholith and parts of satellitic intrusive bodies are exposed in the southernmost part of the area. Rocks near these igneous bodies were metamorphosed to the sillimanite-muscovite subfacies of the amphibolite facies. The grade of metamorphism decreases northward; along the northern edge of the mapped area, the pressure-temperature conditions during metamorphism were near the lower limit of the epidote-amphibolite facies (the biotite-muscovite subfacies). Farther north the rocks of the Belt Series were metamorphosed to the chlorite-muscovite subfacies of the greenschist facies. The distribution of aluminum silicates and staurolite in much of the schist and crystallization of minerals such as tremolite, actinolite, hornblende, and diopside in interbedded quartzite and gneiss provide adequate information for mapping the metamorphic zones in a general way (pls. 1, 2).

The isograds separating these metamorphic zones on the map indicate the first appearance of the index minerals in the pelitic rocks. The appearance of diopside in the calcareous layers coincides with the disappearance of staurolite in the interbedded pelitic layers and was used to trace this metamorphic zone in the quartzite units. The scarcity of aluminum-rich layers in the central part of the area made it impossible to show in detail the isograds for the kyanite and kyanite+sillimanite. Lack of suitable compositions and adequate outcrops also make it impossible to show the offset of isograds near faults. The long history of the faulting (see p. E22) seems to indicate that only a part of the total displacements occurred after the recrystallization. Most faults started to form early during the deformation that was accompanied by the first episode of recrystallization. The second episode, which gave rise to the isograds shown on the maps, was postkinematic and thus the offsets shown by the isograds would be only a part of the total offsets along the faults.

Each of the isograds is a result of a definite reaction which reestablished the equilibrium after a rise in temperature and pressure during the latest episode of metamorphism. This rise in temperature and pressure was connected with the formation and emplacement of the Idaho batholith in much the same way as described for the Elk River-Clarkia area (Hietanen, 1963b). As described above, two definite episodes of recrystallization can be identified on the basis of occurrence of pseudomorphs and orientation of mineral inclusions in poikiloblasts. The second-generation minerals within large staurolite pseudomorphs are kyanite, muscovite, small staurolite, and almandite—an indication of higher temperature and pressure during the second episode of metamorphism. The reactions and changes leading to the present mineral assemblages are complicated in the northeastern part of the area by hydrothermal alteration that affected some of the rocks. Near Peggy Peak and Red Ives the staurolite is hydrothermally altered to chlorite or to a mixture of chlorite, muscovite, and magnetite; this alteration indicates an addition of hydrous molecules and some potassium, and it must have occurred after the second episode of metamorphism because the staurolite elsewhere in this zone is postkinematic.

In the following discussion the most important reactions indicated by the field evidence and by thin-section study is presented in order as they occur from north to south in the field, that is, from the lower to higher grade of metamorphism. The formation of the present mineral assemblages involves at least two sets of reactions in each zone. For example, chloritoid that formed in the southeastern part of the area shown on plate 2 was early; it probably crystallized from kaolinite and iron oxides as suggested by Harker (1939, p. 214) and others. The pseudomorphs after chloritoid that occur near Red Ives are in muscovite-biotite-almandite schist that also contains a few scattered pseudomorphs after staurolite, but none of these occurs to the southwest in the unaltered staurolite schist. Staurolite commonly takes the place of chloritoid in the zone of higher grade metamorphism, and we can express these mineralogic relations by the following equations:

$Al_4(OH)_8Si_4O_{10}+Fe_2O_3 \rightarrow 2\ FeAl(OH_2)SiAlO_5$ (1)
Kaolinite +hematite→2 chloritoid
$+2\ SiO_2 +2\ H_2O+O$
+2 quartz+2 water

$9\ FeAl(OH)_2SiAlO_5 \rightarrow 2\ HFe_2Al_9O_8Si_4O_{16}+SiO_2$ (2)
9 chloritoid →2 staurolite +quartz
$+5\ FeO +8\ H_2O$
+5 iron oxide+8 water.

More staurolite may have been formed as shown by equation (3) from muscovite, chlorite, and iron oxides, the iron oxides being freed in reaction 2.

muscovite+chlorite+FeO⇆almandite (3)
+staurolite+biotite.

This reaction can be expressed by three simpler equations:

$3(OH)_4K_2Al_6Si_6O_{20}+4\ FeO \rightarrow 2\ HFe_2Al_9Si_4O_{24}$ (3a)
3 muscovite +4 iron oxide→2 staurolite
$+3\ K_2O +10\ SiO_2 +5\ H_2O$
+3 potassium oxide+10 quartz+5 water

$(OH)_8(Mg,Fe)_6Si_4O_{10}+(OH)_4K_2Al_6Si_6O_{20}$ (3b)
chlorite + muscovite

$+\ 3\ FeO \rightarrow 3\ (Fe,Mg)_3Al_2Si_3O_{12}+\ SiO_2$
+3 iron oxide→ 3 almandite +quartz

$+\ K_2O +\ 6\ H_2O$
+potassium oxide+6 water

The potassium freed in these reactions would concurrently react with chlorite to form biotite as follows:

$(OH)_8(Mg,Fe)_4Al_4Si_2O_{10}+\ K_2O$ (3c)
chlorite +potassium oxide

$+\ FeO +\ 3\ SiO_2$
+iron oxide+3 quartz

$\rightarrow (OH)_4K_2(Fe,Mg)_5Al_4Si_5O_{20}+\ 2\ H_2O$
→ biotite +2 water

Different compositions of chlorite are purposely used in reactions (3b) and (3c) above because most chlorite is a mixture of all four end members—antigorite, ferri-antigorite, amesite, and daphnite.

In the next higher metamorphic zone, postkinematic crystallization of kyanite, almandite, muscovite, biotite, and second-generation staurolite from the first generation (synkinematic) staurolite occurred as expressed by the following equations:

$10\ HFe_2Al_9Si_4O_{24}+4\ (OH)_4K_2Al_6Si_6O_{20}$ (4a)
10 staurolite + 4 muscovite

$+\ 5\ SiO_2 \rightarrow 4\ (OH)_4K_2Fe_5Al_4Si_5O_{20}$
+5 quartz→ 4 biotite

$+49\ Al_2SiO_5+\ 5\ H_2O$
+49 kyanite+5 water

$6\ HFe_2Al_9Si_4O_{24}+\ 17\ SiO_2\ +\ 2\ K_2O$ (4b)
6 staurolite +17 quartz+2 potassium oxide

$+\ H_2O \rightarrow 4\ Fe_3Al_2Si_3O_{12}$
+water→ 4 almandite

$+2\ (OH)_4K_2Al_6Si_6O_{20}+17\ Al_2SiO_5$
+ 2 muscovite +17 kyanite

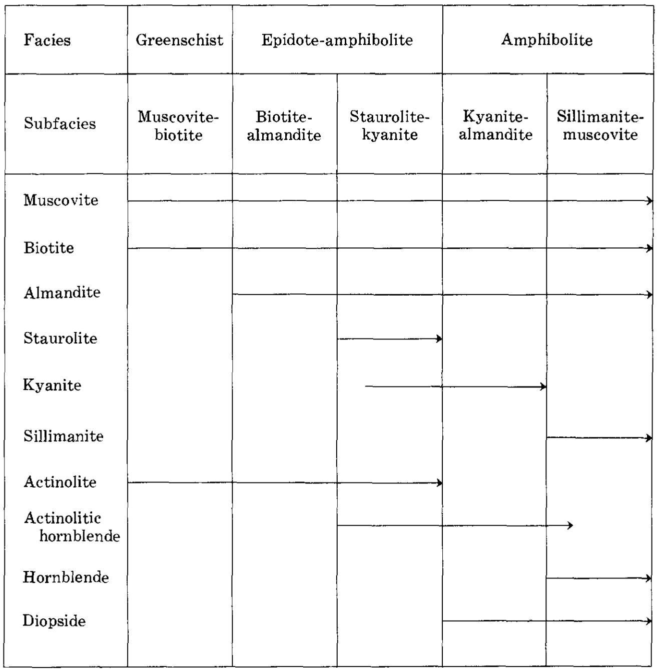


FIGURE 10.—*Stability ranges of the index minerals in the area studied.*

The reaction (4a) involves a decrease in the amount of muscovite and an adjustment in Mg/Fe ratios of staurolite, biotite, and garnet. Potassium needed for the reaction (4b) may have been released elsewhere in the same rock as follows:

$$(OH)_4K_2Al_6Si_6O_{20} \rightarrow 3\ Al_2SiO_5 + 3\ SiO_2 \qquad (5)$$
$$\text{muscovite} \rightarrow 3\ \text{kyanite} + 3\ \text{quartz}$$
$$+\ K_2O\ + 2\ H_2O$$
$$+\text{potassium oxide} + 2\ \text{water}$$

Toward the south, kyanite occurs without staurolite (kyanite zone) and farther south, sillimanite crystallized first with the kyanite (kyanite-sillimanite zone) and then alone (sillimanite zone). Crystallization of sillimanite from biotite was observed in schist of the Prichard Formation in the Boehls Butte quadrangle (Hietanen, 1963a). This reaction can be expressed as follows:

$$(OH)_4K_2(Fe,Mg)_6Al_2Si_6O_{20} + 2O_2 \qquad (6)$$
$$\text{biotite}$$
$$\rightarrow Al_2SiO_5 + 2Fe_3O_4 + 6SiO_2 + K_2O + 2H_2O.$$
$$\rightarrow \text{sillimanite}$$

Most of the iron oxide freed in this reaction would be precipitated as magnetite; magnesium would migrate to the country rock, where it would react with muscovite to form more biotite. An excess of potassium probably migrated to the granitic melt that was forming at depth in the place where the Idaho batholith was emplaced as has been suggested earlier (Hietanen 1961b, p. 164).

Water is freed in all these reactions, and most of it escaped during the metamorphism. However, the hydrothermal alteration of chloritoid to chlorite near Red Ives and alteration of staurolite to chlorite and muscovite near Peggy Peak indicate that some of the water and potassium freed in the inner zone during the second episode of metamorphism entered silicate structures in the outer zone.

Equations (2) and (3) mark the staurolite isograd in the field and equation (4) that of kyanite. The zone in which staurolite is stable overlaps into the kyanite zone for about 5 miles within the mapped area. The zone in which staurolite occurs alone is only about 2 miles wide. The width of the zone of kyanite-almandite schist is about 5 miles and that of kyanite-sillimanite schist is about 6 miles.

The approximate stability ranges of some of the major minerals in relation to the facies boundaries are shown schematically by lines in figure 10. Muscovite and biotite are stable over the whole area. Staurolite appears at a lower grade of metamorphism than kyanite, but most of its stability field overlaps that of kyanite. Sillimanite crystallized as the stable polymorph of Al_2SiO_5 in the zone of high-grade metamorphism next to the batholith. The stability ranges of the minerals in the calcareous layers are also shown. Actinolite and actinolitic hornblende occur together in the field of staurolite-kyanite subfacies. In the kyanite-almandite subfacies, diopside instead of actinolite occurs with actinolitic hornblende. In the sillimanite-muscovite subfacies, strongly pleochroic green hornblende generally occurs with diopside, although in some layers actinolite crystallized because of low content of aluminum and sodium. Ilmenite is the most common titanium mineral in all rocks of low-grade metamorphism, and it occurs also in the sillimanite schist with magnetite. Grains of brown rutile are typical accessory minerals in the rocks of the staurolite-kyanite and kyanite-almandite subfacies. Sphene instead of rutile occurs in the diopside gneiss in the zone of high-grade metamorphism.

Andalusite and cordierite occur in places near the quartz monzonite stock in the southwestern corner of the area. This occurrence indicates relatively lower pressures during the recrystallization there. Cordierite is one of the major constituents in many outcrops of this igneous body. All three polymorphs of Al_2SiO_5 and cordierite occur west of the quartz monzonite stock. This occurrence indicates pressures and temperatures close to those at the triple point (Hietanen, 1956, 1961a, and 1963a) and near the lower stability boundary of cordierite.

The next step after mapping the isograds in the field and establishing the relations between the facies boundaries and the stability ranges of index minerals (as shown in fig. 10) is to determine the temperatures along these facies boundaries. The geologic thermometers such as those quoted by Engel and Engel (1958) and by Ingerson (1955) and developed by Kullerud (1953), Schreyer, Kullerud, and Ramdohr (1964), and Clayton and Epstein (1961) give much useful information about the temperature at which similar assemblages as found in the present area will crystallize. The oxygen isotope ratios in coexisting pairs of quartz-ilmenite, quartz-biotite, quartz-garnet, and quartz-muscovite were determined by Garlick (1964) in several samples from the country rocks of the anorthosite in the neighboring Boehls Butte quadrangle and in staurolite schist specimen 2096. These determinations indicate that the temperature of recrystallization near the anorthosite bodies was above 650°C and in the staurolite-kyanite zone 525°–600°C. Comparison with the other geologic thermometers available (Hietanen, 1961a, p. 94) suggests that the metamorphism in the southern part of the present area took place at temperatures of 500°C–600°C. This estimate is in accordance with the lower stability limit of magnesium cordierite as determined experimentally by Schreyer and Yoder (1964).

The occurrence of all three polymorphs of Al_2SiO_5 and cordierite in the country rocks of anorthosite in the Boehls Butte quadrangle (Hietanen, 1963a) suggests that the triple point in this area is close to the lower stability limit of cordierite. It is possible that in the southern part of this quadrangle kyanite crystallized earlier and that andalusite, sillimanite, and cordierite are later. However, transformations kyanite→andalusite, sillimanite→kyanite, sillimanite→andalusite, and andalusite→sillimanite are common in the northern part and suggest that temperature and pressure fluctuated around the triple point during the metamorphism. Kyanite probably was formed during the deformation and was partly altered to andalusite when the tectonic pressures were diminishing but the temperatures continued to rise, or kyanite altered partly to andalusite during the second postkinematic episode when the pressures were lower. Kyanite, however, was formed in the central part of the present area during this second episode. All these relations together suggest that in the southwestern part of the area the pressures and temperatures during the recrystallization were never very far from the triple point.

Bell (1963) determined experimentally the triple point and the equilibrium boundaries of kyanite, andalusite, and sillimanite, using a modified Bridgman opposed-anvil pres. Synthetic gels of $Al_2SiO_5 \cdot XH_2O$ were used to reverse the reactions. The temperature-pressure plane of the triple point was determined at 300°±50°C and 8±0.5 kilobars, thus at much lower temperature and at higher pressure than geologic thermometry would indicate for the recrystallization of the rocks of a grade similar to those in which the three aluminum silicates were found together. Khitarov and others (1963) determined the triple point at 390°C and 9,000 atm (atmosphere), thus at higher temperature and pressure.

To satisfy the geologic relations, especially the closeness of the melting temperature of granite, and the limitation of the geologic thermometers available, the triple point was placed at 500°C and 5,000 atm pressure for the illustration (fig. 11).[1] Hypothetical lower boundaries for the stability fields of other index minerals—biotite, garnet, staurolite, K feldspar, and hypersthene—were superimposed on the stability fields of the three aluminum-silicates in the order that illustrates their entry as mapped in the field. It should be noted that the isograds mark the lower stability boundary of the index minerals and that each index mineral will appear only in layers that have suitable bulk composition. Thus, the presence of all common natural bulk compositions is prerequisite for correct determination of the isograds. To make the pressure-temperature diagram quantitatively valid, the absolute values for all stability boundaries should be determined experimentally and checked by using geologic thermometers. Comparison of the fields of stability of the index minerals in calcareous layers (such as epidote, actinolite, hornblende, and diopside, fig. 10) interbedded with the schist provides adequate basis for the facies classification of these layers.

Together the boundaries of stability fields of the index minerals and the equilibrium boundaries of the Al_2SiO_5 polymorphs divide the pressure-temperature field into subfields, each of which illustrates a stability field of a certain mineral assemblage. Because we do not know the absolute temperatures and pressures of these stability fields and because these absolute values may be functions of bulk composition, type of pressure, and activity of mobile components such as H_2O, CO_2, and O_2 (Thompson, 1955; Yoder, 1955; Eugster and Wones, 1962), this diagram shows relative temperatures and pressures only approximately. Still it is fruitful to base the classifications and comparisons of the petrologically similar metamorphic complexes on typical mineral assemblages and on their mutual relations as found in the field.

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[1] This is in good agreement with the experimental work published after preparation of this report. Newton (1966) determined the triple point of 500°C and 4.2 kilobars using water-vapor pressure.

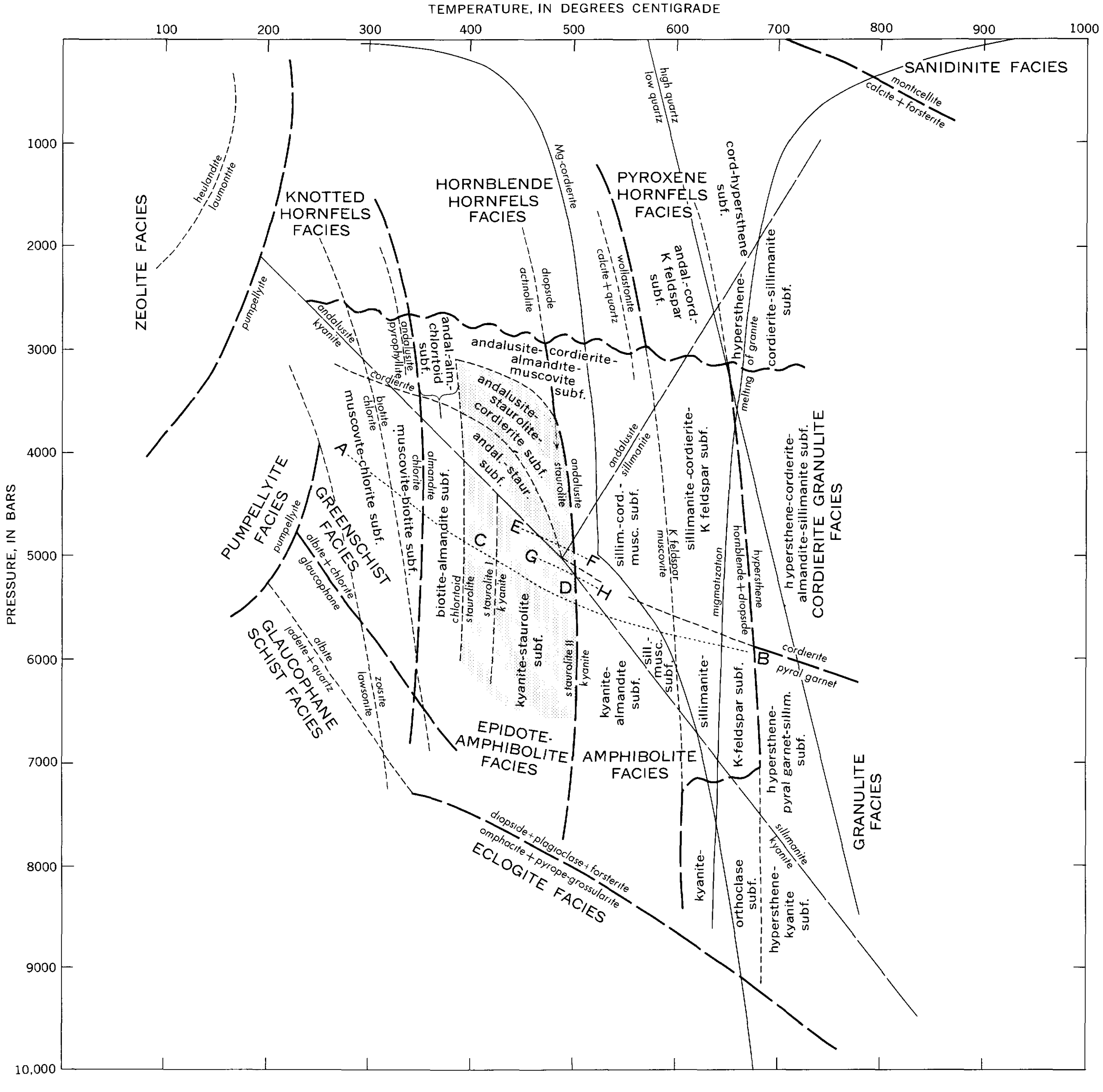

FIGURE 11.—A tentative pressure-temperature diagram showing a possible pressure-temperature gradient of recrystallization (dotted line A–B) and the stability fields of the subfacies north of the Idaho batholith. For comparison, pressure-temperature gradient for the rocks in the neighboring Boehls Butte quadrangle and to the west is shown (lines E–F and G–H). The melting curve of granite is after Tuttle and Bowen (1958), the curve for low-high quartz is after Yoder (1950), and the stability curve for magnesium cordierite is after Schreyer and Yoder (1964). Hypothetical lower boundaries for the stability fields of biotite, almandite, chloritoid, staurolite, K feldspar, and hypersthene are superimposed on the equilibrium diagram of kyanite, andalusite, and sillimanite. The temperature of the triple point was chosen to satisfy field relations and geologic thermometry. Pressures are estimated from geologic relations. Possible pressure-temperature fields of hornfelses, granulites, eclogites, glaucophane schist, and some low metamorphic facies are shown for comparison.

The mutual relations of the stability fields of the subfacies as they appear in schist of the Wallace Formation northwest of the Idaho batholith are illustrated along curve A–B in figure 11. The chemical composition of the aluminum-rich schist remains constant within rather narrow limits, but different minerals have crystallized because of an increase in temperature and pressure toward the batholith. Curve A–B illustrates the path of recrystallization during this increase. As the stability field of a certain mineral (for example staurolite, point C) is reached when moving from left to right along curve A–B, this mineral will start to crystallize in layers rich in aluminum and iron and will form a stable component of the rock until it becomes unstable at a higher temperature and pressure (point D). The temperature-pressure gradient during the recrystallization in the Boehls Butte quadrangle follows curve E–F and that in the Elk River-Clarkia area (Hietanen 1963b) curve G–H. The muscovite→K feldspar curve illustrates the lower limit of the stability field of K feldspar in natural occurrences, and it was placed at temperatures lower than the curve for the upper limit of the stability of muscovite in hydrous conditions as determined by Yoder and Eugster (1955). Transitional zones in the field are generally wide and indicate the sluggishness of the reactions.

The boundaries of the mineral facies and subfacies are conveniently defined by the stability ranges of certain minerals (index minerals) or mineral assemblages (Fyfe and others, 1958). Rarely do two minerals have the same field boundary, rather the stability fields overlap and there is no general agreement as yet as to which reaction surface should be defined as a facies boundary. This lack of agreement has introduced some differences in the facies classification (compare Turner, 1948; Fyfe and others, 1958; and Barth, 1962). For example, the upper limit of the greenschist facies is generally identified with the appearance of biotite and almandite and disappearance of chlorite. As the appearance of biotite and garnet rarely coincide and as either one can be earlier than the other, this upper limit can be placed at the beginning of crystallization of either mineral. Staurolite-bearing assemblages were originally included in the amphibolite facies by Eskola (1915). This division, however, would make the temperature range of the epidote-amphibolite facies very narrow in comparison to that of the amphibolite facies. Moreover, the lower limit of the amphibolite facies is defined by many writers as the appearance of sillimanite-bearing assemblages in the pelitic layers. To avoid the ambiguities and to more equally group the subfacies, the border between the epidote amphibolite and amphibolite facies is here placed at the upper stability limit for staurolite. Thus, such assemblages as staurolite-kyanite-garnet-biotite-muscovite are included in the epidote amphibolite facies, but those without staurolite are in the amphibolite facies (fig. 11). The facies boundary, if defined in this way, closely coincides with the lower stability limit of diopside. Actinolite instead of diopside occurs in the calcareous layers interbedded with the staurolite-bearing schist of the epidote amphibolite facies.

STRUCTURE

The structure of the Belt Series rocks in the northern part of the mapped area is fairly simple but becomes increasingly complex toward the south. Two to three sets of folds on axes that intersect at an angle of 60°–80° may be found within a small area. Two sets of folds, or two intersecting lineations, generally occur in a single outcrop. The intensity and style of folding changes from north to south; gentle, open folds are common in the northern part, whereas isoclinal folds are more characteristic in the southern part of the area. Transecting cleavage is pronounced in the schist in the northern and central parts; locally this cleavage is folded. Several major faults, and others branching off from these, break the continuity of the rock units. Piled-up recumbent folds indicating minor overthrusting occur locally.

FAULTS

Two major groups of faults can be recognized; the earlier ones trend northwest, and the later ones trend north or nearly north. Most of the faults are poorly exposed. They are shown where mapping revealed interruptions in the continuity of the stratigraphic sequence and structural trend. In many individual localities, however, rusty-brown weathered brecciated rock or yellow-brown clayey material is exposed along the faults. In some other places, as for example along Buck Creek and near Thor Mountain, strongly sheared quartzite is exposed along the fault. Hypabyssal and plutonic rocks are common along the faults; most were emplaced parallel to these structural zones of weakness, but some are strongly sheared. The shearing indicates that the movements continued after the emplacement of these intrusive rocks.

Such a relation was observed in the diabase sill along Periwinkle and Nugget Creeks and was reported earlier for the southwest border zone of the quartz monzonite along Beaver Creek (Hietanen 1961d, 1963c). Foliated gabbro is exposed along the west-trending fault that crosses the North Fork of the Clearwater River half a mile north of the mouth of Rock Creek.

These relations indicate that most of the faults were active already during the major period of folding and recrystallization which preceded the emplacement of the early Late Cretaceous batholith. (See p. E27.) Emplacement of most hypabyssal rocks was later than that of the batholith and later than much of the faulting. Some of the faults, however, were active after the emplacement of their igneous fill.

NORTHWEST-TRENDING FAULTS

The northwest-trending fault that separates the quartzite and gneiss unit of the Wallace Formation from schist of the Prichard Formation north of Canyon Ranger Station (pl. 1) is part of a major northwest-trending fault system north of the Idaho batholith. It is referred to as the Canyon fault in this report. About 2 miles east of Canyon Ranger Station a north-trending fault cuts the Canyon fault near the canyon of the North Fork of the Clearwater River. The eastern segment continues to the east-southeast and crosses the North Fork of the Clearwater River north of the mouth of Rock Creek. The western segment has been mapped in the southern part of the Boehls Butte quadrangle (Hietanen, 1963a) and in the Elk River-Clarkia area (Hietanen, 1963b). In these areas it forms the southern and western borders of an uplifted block in which anorthosite, its country rocks, and rocks of the Prichard Formation are exposed. Rocks of the Prichard Formation have been brought into contact with those of the Wallace Formation along this fault; a vertical displacement of about 1,000m is thus indicated.

Another northwest-trending fault, the Foehl Creek fault, extends from the canyon of Foehl Creek on the eastern border of the Boehls Butte quadrangle (Hietanen, 1963a) southeastward to Larkins Peak. This fault interrupts the quartzite units of the Prichard Formation northwest of Larkins Peak, where it apparently ends in coarse-grained biotite-plagioclase gneiss and amphibolite. Both of these rock types contain abundant igneous material that may conceal another fault.

Along a third northwest-trending fault, rocks of the Revett Formation rest on the quartzite unit of the Wallace Formation near Surveyors Peak. The fact that both formations dip to the north suggests an overthrust to the south (pl. 1, cross section *B–B′*). This structural analysis is supported by the occurrence of folds overturned to the south on slopes of Surveyors Peak.

NORTH-TRENDING FAULTS

Two major north-trending fault zones, the Buck Creek and the Collins Creek, extend across most of the area; considerable displacement has occurred along them. Other faults with northerly trends and displacements of smaller magnitude seem to branch off from these two.

The more conspicuous of the two, the Buck Creek fault, extends from the vicinity of Thor Mountain in the north to the vicinity of Eagle Point in the south. It passes to the east of Bathtub Mountain, follows the canyon of Buck Creek for 2 miles, passes over Black Mountain, crosses the North Fork of the Clearwater River 2 miles east of the Canyon Ranger Station, and probably continues southward for 3 more miles. The part of this fault near Buck Creek was shown on the map of Shoshone County by Umpleby and Jones (1923). Stratigraphic and structural relations east of the Bathtub Mountain show that the east side is upthrown and that the vertical displacement amounts to about 400 m in this vicinity. Overturned folds near Bathtub Mountain suggest that this fault may also have a considerable strike-slip component southward on the west side. The axes of these folds plunge gently to the east-northeast and make an angle of about 75° with the trend of the nearby fault. The overturning, which is to the south, indicates movement southward on the west side.

Several other north-trending faults branch away from the Buck Creek fault. The longest, the Skyland fault, branches away near the mouth of Buck Creek, passes to the east of the Skyland Lake, and continues south to the vicinity of the North Fork of the Clearwater River. North of Isabella Creek, the east side of Skyland fault is upthrown, and there appears to have been considerable vertical displacement of this type near Mallard Peak. The west side of Skyland fault is upthrown south of Isabella Creek, where quartz diorite conceals a northwest-trending fault. A fault near Snow Peak and another along Caribou Creek seem to join 0.5 mile east of Canyon Peak, where a small body of gabbro conceals their junction. From this body a fault continues south-southeast and ends against the Buck Creek fault. These faults, together with the Buck Creek fault and its longer branching faults, have broken the rocks of the Ravalli Group into five large blocks which show mainly vertical displacements.

The trace of the other north-trending major fault zone follows along the west side of Collins Creek and is referred to as the Collins Creek fault. Along this fault, quartzite of the Wallace Formation has been brought into contact with the Revett and St. Regis Formations southeast of Surveyors Peak and with schist of the St. Regis Formation along Skull Creek. Small slivers of coarse-grained mica schist (not shown on the map but presumably of the St. Regis Formation) are exposed in some localities along this fault, as for example east of Collins Peak and just south of Drift Creek, where the lowermost beds of the Wallace Formation are exposed just east of the fault and dip eastward. An east-trending fault along Cliff Creek is cut by the Collins Creek fault and in turn cuts off a north-branching fault from the Collins Creek fault. Along this branch, west-dipping beds of the upper part of the quartzite unit of the Wallace Formation are in fault contact with coarse-grained garnet-mica schist of the St. Regis Formation.

A north-trending fault whose trace passes over the east slope of The Nub and parallels the Collins Creek fault has doubled the thickness of the St. Regis Formation. High outcrops of schist on the steep eastern slope of The Nub show many large folds one above another and all strongly overturned to the east. It is the front of these piled-up folds (see also p. E25) that is indicated with a fault on plate 1.

FOLDING AND LINEATION

A characteristic structural feature of the mapped area is the occurrence of two sets of folds in many single outcrops. The axes of the major folds trend generally in a northwesterly direction and plunge either to the southeast or to the northwest. A second set of more minor fold axes trends mainly in a north-northeasterly direction and plunges northeastward or, less commonly, southwestward. Deviations from these trends occur near many faults in a manner that suggests a close relation between some folds and the conspicuous faults. Moreover, several of the large fault blocks seem to form structural units, each independent and slightly different from its neighbors. The style of the folding becomes more intense with an increase in the grade of metamorphism; the style also is partly dependent on the type of material folded. Each of these factors that contribute to the complexity of structural features is described separately.

TWO SETS OF FOLDS

In many outcrops, two sets of folds can be observed and their axes measured (Hietanen, 1961d); in others, one set of folds can be either observed or projected from the attitude of the bedding planes and the other set appears as a wrinkling of the bedding planes. In still other outcrops, especially in schist beds, two sets of wrinkling can be measured. On plates 1 and 2, symbols for fold axes are indicated where folds could be observed in the field and their axes measured. Lineation symbols indicate measured axes of wrinkling; because axes of folding and wrinkling are parallel, these symbols actually give directions of fold axes. In some rare places, two sets of folds and a wrinkling can be measured in a single outcrop. This wrinkling may appear as minor folding in neighboring outcrops.

The type of deformation resulting from two sets of folds can be studied in many outcrops near Black Mountain and near The Nub (pl. 1). The rock around Black Mountain is a coarse-grained garnet-mica schist of the Prichard Formation that contains discontinuous thin layers of white quartzite. On the ridge north of Black Mountain, where the schist is weathered out and leaves the bedding planes of quartzite exposed, saddle-shaped configurations of these planes are apparent. Saddle-shaped outcrops of schist 0.5 mile southwest of Black Mountain have small folds on south-plunging axes and larger ones on east-southeast-plunging axes (fig. 12). Where two vertical walls at right angles are exposed, each reveals a set of folds. A mile southwest of Black Mountain, larger folds strongly overturned to the southwest were formed on the southeast-plunging axis; and the smaller folds, overturned to the east are on south-southwest-plunging axes. The larger folds have an exposed amplitude of 1 m and the smaller ones a wave length of 15 cm and an amplitude of 6 cm. Biotite is parallel to the axial planes of smaller folds, which are therefore later than the larger folds.

FIGURE 12.—Two sets of folds in garnet-mica schist of the Prichard Formation half a mile southwest of Black Mountain (loc. 2058). The handle of the hammer (upper left) is parallel to the axis of small folds plunging 30° S. In the photograph, the axis of the gentle large syncline plunges away from the observer 5°–10° ESE.

The bottoms of beds that were deformed by two episodes of folding are exposed on the rocky ridge between Cliff Creek and Drift Creek 1.5 miles north of The Nub. The bedding planes in this locality strike generally N. 40° W. and dip 70° NE. The axes of the two sets of small folds trend N. 60° W. and N. 40° W. and plunge 20° ESE. and 40° NW., respectively. These directions coincide with axes of wrinkling and folding in the nearby outcrops; here the wrinkling is earlier. Where erosion has removed softer layers and exposed bottom sides of more resistant ones, the combination of the two transecting sets of small folds have twisted the bedding surface in such a fashion that it resembles the bottom of a huge egg carton (fig. 13). In addition to these two sets of folds, there is an earlier lineation (wrinkling) that trends N. 40° W. and plunges 5° SE.

The time relation between the observed sets of folds differs from place to place. The large overturned folds are commonly earlier because their axial planes are refolded. These overturned folds were formed mainly either on an east- or southeast-plunging axis and the second smaller folds on the northeast- or south-plunging axis. In the northeastern part, however, the axis of each large fold plunges southward and the cleavage is gently folded. Near Bathtub Mountain the large folds overturned to the southeast are on northeast-plunging axes. The major episode of recrystallization coincided with the peak of deformation, as indicated by the orientation of micas parallel to the axial planes of large overturned folds of any set (see Hietanen 1961d, fig. 9).

INFLUENCE OF THE GRADE OF METAMORPHISM

Variation in the intensity and style of folding is notable in different metamorphic zones northwest of the Idaho batholith (see Hietanen, 1961a). As shown on plate 1, the rocks in the southern part of the area near the igneous bodies are intensely folded, dips of 60°–90° being common. This intense folding is well exemplified by the isoclinal folds with steeply dipping flanks that are exposed along the North Fork of the Clearwater River south of the mouth of Rock Creek. In the central part of the mapped area, between the North Fork of Clearwater River and Canyon Creek, dips of 30°–45° prevail and folds have smaller amplitudes and larger wave lengths than those in the southernmost part. In the northern part, where dips of 10°–30° are widespread, folding is still gentler. Dips steeper than these occur only in a few localities where occasional small asymmetric or overturned folds are exposed. This rather simple change in the style of folding is complicated by two factors: (1) the differences caused by the type of material folded and (2) local deformation near some of the faults, as discussed below.

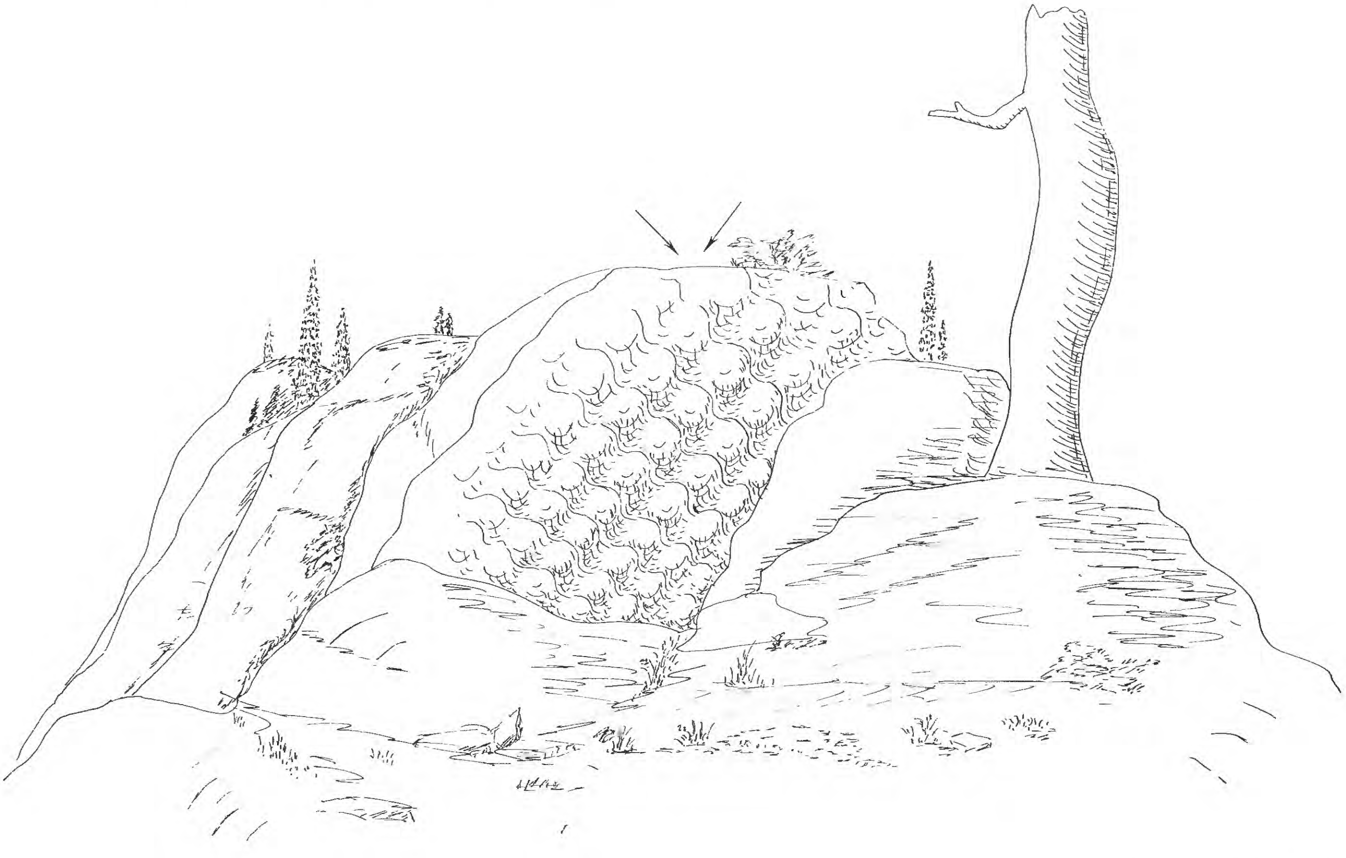

FIGURE 13.—Two sets of small folds in schist of the St. Regis Formation, 2 miles northeast of The Nub between Cliff Creek and Drift Creek. The arrows show the direction of fold axes. Folded surface faces westward and dips 70° away from the observer. The tree in the foreground is 30 cm thick.

INFLUENCE OF THE TYPE OF MATERIAL FOLDED

Schist layers are generally considered to yield more readily during deformation than more competent quartzite layers. Where the opposite is apparently true, as exhibited in the structures shown in figures 14*A* and 14*B*, it is somewhat startling. These structures are on outcrops of interbedded quartzite and schist, one within the Prichard Formation on Black Mountain (fig. 14*A*) and the other within the St. Regis Formation along Canyon Creek a third of a mile east of the mouth of Buck Creek. In these localities, layers of white medium-grained quartzite, 10–30 cm thick, are interbedded in coarse-grained garnet-mica schist. The quartzite layers are intensely folded, whereas the bedding planes of the enclosing schist farther from the immediate contact zone are fairly straight or appear only gently folded. A closer study, however, shows that the schist is intensely wrinkled. The shortening of the strata due to deformation is obviously of equal magnitude in both rock types, but difference occurs in the type of deformation. The more competent quartzite layers are thrown into round small folds, whereas the less competent schist is wrinkled.

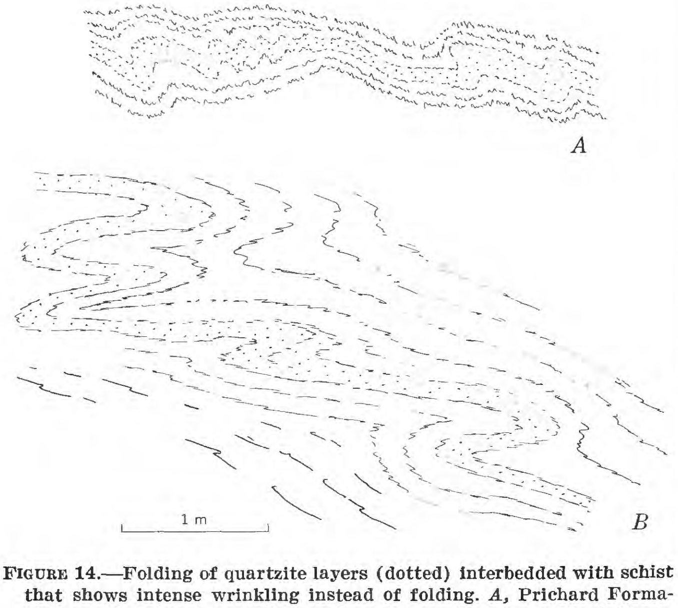

FIGURE 14.—Folding of quartzite layers (dotted) interbedded with schist that shows intense wrinkling instead of folding. *A*, Prichard Formation, southeast-facing vertical wall on the north slope of Black Mountain; *B*, St. Regis Formation, east-facing vertical wall on Canyon Creek, one-third of a mile east of the mouth of Buck Creek.

These differences in the type of deformation between quartzite and schist were observed only in the highly metamorphosed rocks. Quartzite and schist in the Wallace Formation in the zones of low-grade metamorphism show a regular relation: the schist is more intensely folded than the quartzite. For example, the lower quartzite unit of the Wallace Formation south of Papoose Mountain dips 20°–40° north-northeast and is obviously gently folded, whereas the schist on Papoose Mountain, above this quartzite, is strongly folded and overturned to the south-southwest. Here the quartzite formed a thick competent relatively unyielding unit and the less competent schist yielded more readily to the deformation. Thus it seems that two factors will determine the type of deformation in the interbedded schist and quartzite, the grade of metamorphism and the relative thickness of the interbedded layers. Relatively thin layers of quartzite in the schist will be folded, whereas the schist will be wrinkled, but thick layers of quartzite are less readily folded than the schist. The former type of deformation was observed only in the rocks subject to high-grade metamorphism. Cross section *A–A′* and the map pattern in the vicinity of Black Mountain and The Nub show such a difference in the style of deformation on a large scale.

FOLDS RELATED TO FAULTS

Strongly overturned folds occur near certain fault zones. Most of the folds observed are on the east slope of The Nub, just west of a north-trending fault; others were seen in the upper plate of the overthrust near Surveyors Peak and near Bathtub Mountain west of the Buck Creek fault. The steep cliffs on the east slope of The Nub reveal many large folds in schist of the St. Regis Formation strongly overturned to the east. Some outcrops, 20 m high, have six or more such folds, one on top of another. This arrangement indicates a considerable piling up of strata. The axial planes of these folds dip 10°–30° W. This structural complexity extends only about a mile to the south of The Nub, where it is replaced by a high-angle fault that extends several miles farther south. Toward the north, overturned beds were seen for a distance of 2 miles.

Near Surveyors Peak, folds on northwest-trending axes are overturned to the southwest; their axial planes are parallel to the northwest striking overthrust fault on the south side of this mountain. On Papoose Mountain, folds with corresponding axes, now plunging east-southeast, are also strongly overturned to the southwest. No faults were observed in the immediate vicinity of these folds, but their character is similar to those just described.

The folds near Bathtub Mountain (loc. 1579) are in the schist unit of the Wallace Formation exposed on the west side of the Buck Creek fault. They are overturned to the southeast and have a well-developed axial plane foliation, which dips to the northwest. The axes of the folds strike northeast, at an angle of about 60° to the nearby Buck Creek fault, and plunge gently northwest.

These folds indicate a crustal shortening of the second unit of the Wallace Formation on the west side of the fault, whereas the quartzite beds of the lowest unit on the east side are straight and dip gently to the east. These relations indicate that Buck Creek fault in this vicinity has a strike-slip component south on the west side (left-lateral displacement) in addition to the vertical component up on the east side.

DIFFERENCES IN TRENDS BETWEEN INDIVIDUAL FAULT BLOCKS

Several of the larger blocks bordered by faults seem to form structural units in which trends of fold axes differ from the trends in the neighboring block. Differences also occur between the role of the major and the secondary fold axes and between the fold axes and the lineation. Larger folds were formed on the major axes, which generally trend northwest. Secondary folds range in size from minute wrinkles to folds with an amplitude of 1–3 m.

In the southernmost part of the mapped area, the major fold axes trend west-northwest and the lineation is obscure or lacking. Another set of folds whose axes trend northeast becomes apparent near and north of the Canyon fault, which separates the Wallace Formation in the south from the Prichard Formation to the north. In the Prichard Formation, these folds on northeast-trending axes are as prominent as those with northwestern trends and southeastern plunges.

Along Collins Creek and south of Pole Mountain, the prominent folds have axes that trend north. A lineation which appears as a strong coarse wrinkling of the micaceous layers plunges north-northeast. The southeastern trends on Surveyors Peak are separated from these northerly trends by the Collins Creek fault.

North of Surveyors Peak, in the northeastern part of the mapped area, this same north-northeast-trending lineation is a common and characteristic feature of the structural pattern. It not only appears as a wrinkling on the bedding surfaces but also locally as axes of small folds. Another linear element, an east-plunging axis, was observed in only a few localities east of the Buck Creek fault, but it becomes prominent west of this fault.

In the vicinity of Snow Peak and Mallard Peak, three linear elements may occur in a single outcrop, a fold axis and two sets of wrinklings, one coarser and later than the other. Folds are either on northwesterly or easterly axes—more rarely on northeasterly ones; the structure in this area thus contrasts with the structure in the northeastern part.

FOLIATION, CLEAVAGE, AND FRACTURE CLEAVAGE

In the southwestern half of the mapped area, foliation is parallel to the bedding (bedding foliation of Fairbairn, 1949). Segregations in quartzo-feldspathic veins parallel to the same planes have accentuated the compositional differences and made this plane an excellent slip surface. In this part of the area, axial-plane foliation is rare; it occurs only in some thicker homogeneous schist layers.

In contrast, axial-plane cleavage is well formed in the northeastern half of the area. It may be parallel to axial planes of any set of folds, whichever is best formed. It is conspicuous in the schist of the staurolite and kyanite-staurolite zones and also occurs in some layers in the garnet zone. It is weak or lacking in the granofels and quartzite.

Axial-plane cleavage is due to the orientation of mica flakes parallel to the axial planes. In the biotite and garnet zones, only the newly crystallized biotite porphyroblasts (fig. 4) are parallel to the cleavage, which is therefore barely discernible in the field. In the staurolite zone both muscovite and biotite flakes are parallel to the axial plane cleavage. Thus they are at an angle to the compositional layering (fig. 15). The axial plane cleavage was not formed in the thin-bedded quartzite of the staurolite zone; in this rock the slip was parallel to the thin micaceous layers that separate the competent quartzite layers. Thus the distribution of axial plane cleavage follows a pattern similar to that described in earlier reports (Hietanen, 1961a, d)

An apparent folding of cleavage is an interesting feature of the northeastern corner of the mapped area. On the four ridges between the Forks of Bluff Creek, Mosquito Creek, Fly Creek, and Beaver Creek, the attitudes of the cleavage in the schist indicate folding on axes that have northerly trends and parallel the lineation shown on plate 2. Southwest of Conrad Peak and southwest of Angle Point these axes plunge southward,

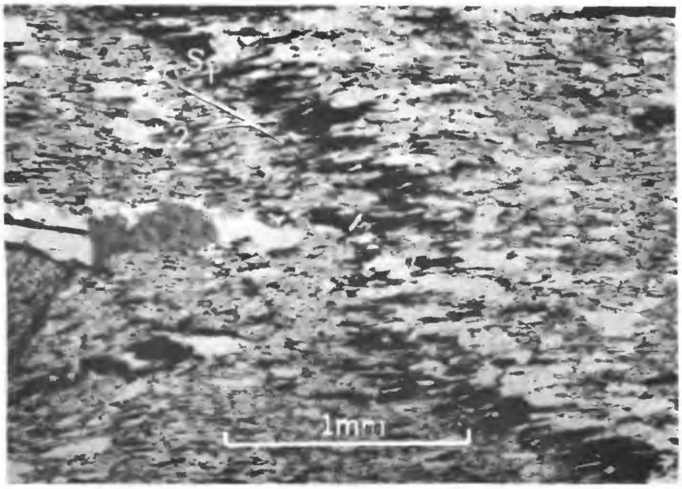

FIGURE 15.—Photomicrograph of garnet-mica schist showing bedding (s_1) and transection foliation (s_2). Micas are parallel to the foliation. Fly Creek, 1 mile east-southeast of Peggy Peak (loc. A–154).

but between Peggy Peak and Attention Point they plunge northward. It seems that this linear element is of later origin than the other structural features in the area.

In accordance with this conclusion, fracture cleavage, striking north-northeast and dipping 40°–55° W., is well formed in the garnet-mica schist of the staurolite-kyanite zone south of Pole Mountain. This fracture cleavage is parallel to the axial planes of microfolds, whose axes plunge north-northeast. The fracture cleavage makes an angle of 60°–90° with the bedding planes that dip 35°–65° E. and is not parallel to the axial planes of large folds. Apparently the fracture cleavage was formed after the major period of folding, as did also the folding of the cleavage north and east of Peggy Peak.

Fracture cleavage is a rare local feature elsewhere in the mapped area and has no significance other than as the occasional continuance of the second wrinkling to the brittle stage of deformation, or as a late feature connected with the faulting.

IGNEOUS ROCKS

There are two groups of igneous rocks: plutonic rocks of Cretaceous and perhaps of younger age and hypabyssal rocks of younger age. The plutonic rocks in the southern part of the area are associated with the Idaho batholith, which, according to Larsen and others (1958), is early Late Cretaceous. The same petrographic varieties as described previously for the areas to the south and west (Hietanen, 1962a; 1963a, b, c) are represented. Among the hypabyssal rocks, quartz monzonitic and granitic compositions are common in the southern and western parts, whereas gabbroic dikes and sills occur along the St. Joe River in the northeast. These rocks also are similar to the corresponding varieties described previously and therefore only some additional information is given here on all igneous rocks.

PLUTONIC ROCKS

The composition of the plutonic rocks ranges from olivine gabbro and hornblende gabbro through quartz diorite to quartz monzonite and granite. It is not clear whether the largest quartz diorite body is associated with the quartz monzonite or with the quartz dioritic border zone of the Idaho batholith that is exposed just south of the present mapped area. The large body of quartz monzonite along Beaver Creek and the included small bodies of olivine gabbro, pyroxene gabbro, and quartz diorite are an igneous complex that has been described in an earlier report (Hietanen, 1963c). The granite exposed along the south border of the area of plate 1 is the northern margin of the granite pluton near Bungalow described in the same report. The amphibolite and garnet amphibolite are similar to these rock types in the areas farther west (Hietanen 1962, 1963a).

QUARTZ DIORITE

A body of quartz diorite, 6 miles long and averaging 1 mile in width, is exposed in the southernmost part of the area. It is bordered by granite on the south and separated by a long body of hypabyssal quartz monzonitic rock from the quartz monzonite on the northwest. The quartz diorite is older than the granite, but its relation to the quartz dioritic border zone exposed south of the area shown on plate 1 remains uncertain. The quartz diorite is finer grained and less homogeneous than the quartz diorite exposed west of the Bungalow granite pluton. Most of it is medium grained and well foliated. Minerals are plagioclase (An_{33}), quartz, hornblende, and biotite, with some sphene, magnetite, zircon, apatite, and allanite. In some thin sections, chlorite occurs as an alteration product after biotite and hornblende. The plagioclase grains are rounded and larger (0.5–3 mm) than the small interstitial grains of quartz. Biotite and hornblende are also interstitial to plagioclase, or occur in long shreds parallel to the plane of foliation. Foliation strikes parallel to the length of the quartz diorite body and dips mainly to the south in the eastern part of the body and to the north in the western part.

A small body of rock mapped as quartz diorite is exposed on a ridge about a mile southwest of Mallard Peak. It consists of strongly foliated biotite-plagioclase-quartz gneiss and hornblende-biotite-plagioclase gneiss. The biotite-plagioclase-quartz gneiss is coarse grained and the biotite is segregated into irregular laminae. Plagioclase (An_{25}) occurring as round to subhedral grains, 1–2 mm long, constitutes about 55 percent of this rock. Quartz amounts to 20–25 percent and occurs in small (0.1–1 mm long) round grains that show strongly undulatory extinction. Biotite (about 20 percent) has a strong pleochroism X=pale brown, $Y=Z$= brown and contains inclusions of zircon, apatite, and allanite. The darker parts of this rock contain various proportions of hornblende (5–15 percent). The amount of quartz in these dark parts is less than in the light-colored parts and the plagioclase is richer in anorthite (An_{35}). Sphene, apatite, and zircon occur as accessory minerals. This body is mineralogically and structurally similar to the small bodies of gneissic quartz diorite in the northeastern part of the Elk River-Clarkia area (Hietanen, 1963b) and most likely is a synkinematic intrusive body that differentiated and became foliated during the deformation of the country rock. The country rock is strongly migamatized near the contact, which is concordant in single outcrops. Abundant pegmatite is common along the contact zones.

QUARTZ MONZONITE AND GRANITE

Parts of the two large plutons, one of quartz monzonite (Beaver Creek pluton) and the other of granite (Bungalow pluton), crop out within the southwestern and southern margin of the mapped area (Hietanen, 1963). Parts of two smaller bodies lie within the western margin; one is an elongate granitic mass along the Little North Fork of the Clearwater River and the other is a quartz monzonite body just west of Alpine Creek that forms the east end of the Roundtop pluton. Other similar rocks within the mapped area include a granitic sill-like body exposed in the east wall of the Little North Fork of the Clearwater River and an elongate granitic body along Perry Creek near the east edge of the mapped area. The Roundtop pluton was mapped as granodiorite by Umpleby and Jones (1923, pl. 1). Most of the rock, however, is mineralogically and texturally similar to the plutonic rocks of quartz monzonitic and monzotonalitic composition farther south (Hietanen, 1961c; 1963c, p. 20–23). The Roundtop pluton consists of about 20 percent quartz, 50 percent plagioclase (An_{32}), 15–25 percent orthoclase, and 5–15 percent hornblende and biotite combined. Zircon, apatite, epidote, and magnetite occur as accessory minerals. This quartz monzonite is a medium-grained rock in which clusters of dark hornblende prisms and biotite plates contrast sharply against the very light gray to white groundmass. Plagioclase is euhedral and zoned; many small grains of it are included in large quartz and orthoclase grains. Some of the orthoclase is interstitial. There is no apparent orientation of minerals in this eastern end of the pluton. A variant along the southeastern margin of the pluton is a very light gray medium-grained porphyritic quartz monzonite. Phenocrysts in this rock are euhedral zoned plagioclase, quartz, and biotite. Abundant orthoclase is interstitial and forms granophyric intergrowths with quartz.

Most of the granite exposed along the Little North Fork of Clearwater River is inhomogeneous and foliated. A few outcrops along the river consist of quartz monzonite similar to that in the eastern part of the Roundtop pluton. The gneissic parts of the pluton contain remnants of schist and amphibolite and are structurally similar to the gneissic tonalite in western Clearwater County (Hietanen, 1962a). The contacts are concordant and small lens-shaped bodies of igneous rock occur in the schist near the contact. A larger layerlike body of fine- to medium-grained granitic gneiss is exposed at a higher altitude along the eastern canyon wall. This rock consists mainly of quartz, plagioclase (An_{18}), biotite, and only a small amount of potassium feldspar. The foliation in the granitic gneiss is parallel to the bedding of the country rock. This and other similar small gneissic bodies have probably resulted from the introduction of elements that mainly formed plagioclase in the country rock. The source of these elements was most likely the quartz monzonite magma from which the largest body crystallized.

The granite stock that is exposed along Perry Creek and on the ridge to the west for 2.5 miles is very light gray medium-grained rock. It consists of plagioclase (An_{23-26}, about 35 percent), undulatory quartz (35 percent), microcline (28 percent), biotite (1 percent), and muscovite (1 percent). The microcline has a typical grid texture and occurs in grains smaller than those of the quartz and plagioclase. Three small bodies near this Perry Creek stock consist of similar granite and also contain abundant pegmatitic material.

PEGMATITE AND QUARTZ VEINS

Pegmatite and quartz veins are far more common in the southern than in the northern part of the area. They are also more common near the plutonic rocks than elsewhere. Veins in the schist and gneiss are pegmatitic and consist of quartz, feldspars, muscovite, and biotite. In the quartzites, most veins consist almost entirely of quartz with only a minor amount of feldspars. Most veins are parallel to the bedding and probably are a result of "sweating out" of quartz and feldspars parallel to the shear surfaces. Some of the veins cut the bedding, and small replacement veinlets branch off from the large ones. Figure 16 shows a vein of this type in tremo-

FIGURE 16.—Quartz vein in actinolite-diopside granofels 1.5 miles southwest of Dismal Lake. Note that there is no offset of the bedding planes. Pencil in the upper left corner is 10 cm.

lite-diopside granofels in the Wallace Formation southeast of the Roundtop quartz monzonite pluton. The bedding is not disturbed by this vein; it was formed either by replacement or by filling of an oblique joint that opened during stretching parallel to the beds.

HYPABYSSAL ROCKS

The composition of hypabyssal rock ranges from gabbroic and dioritic to quartz monzonitic and granitic; the quartz monzonitic dikes and sills are most common. No age determinations are available on the hypabyssal rocks. The geologic evidence shows that most are younger than the faults which are later than the plutonic rocks of the early Late Cretaceous batholith (Larsen and others, 1958), but are older than basalt of the Columbia River Group of middle Miocene through early Pliocene age (Heitanen, 1963c). Many dikes were emplaced along the north trending faults that are parallel to Laramide structural features. In Montana the age determinations made on the Boulder batholith, which has been considered to be of Laramide age, indicate an age of 60–80 million years (Knopf, 1964). Inasmuch as the hypabyssal rocks are the latest group of igneous rocks, an early Tertiary age seems probable.

GABBROIC SILLS AND DIKES

The gabbroic sill exposed along the St. Joe River between the mouths of Mosquito Creek and Bluff Creek is a part of the so-called Wishards sill which was considered older than the quartz monzonite by Calkins and Jones (1913, p. 174) and Umpleby and Jones (1923, p. 119). This sill and the dikes near it consist of greenish-black fine- to medium-grained hornblende-augite-plagioclase rock in which the amount of dark minerals is about 60 percent. Much of the augite is altered to light-green weakly pleochroic hornblende that is rimmed by blue-green hornblende (fig. 17). Radiating clusters of actinolitic hornblende and aggregates of chlorite also occur as alteration products. Numerous fairly large grains of iron-ore minerals occur with the hornblende and chlorite. Plagioclase (An_{46}) in stocky, subhedral crystals that include abundant epidote as an alteration product make up 30–40 percent of this rock. Some quartz, a very little of it in granophyric relation, is interstitial. Apatite, sphene, and leucoxene are the common accessory minerals.

In most of the smaller gabbroic dikes, diabasic texture is common. For example, in those dikes a mile east of Pineapple Peak (loc. 2082), 1.5 miles south-southeast of Mallard Peak (loc. 2213), and 1.5 miles south of Goat Ridge (loc. 1721), plagioclase (An_{50-52}) occurs in long lath-shaped crystals, and hornblende or hornblende and augite fill the interstices. In the dikes near Bathtub Mountain, all of the augite is altered to aggregates of small prisms of light-green hornblende rimmed by bluish-green hornblende. In those near Goat Ridge and Mallard Peak, only a part of the augite crystals are altered. A large amount of magnetite accompanied by some biotite and sphene is common in all these dikes. These gabbroic dike rocks are similar to the sill-like bodies of pyroxene gabbro in the Boehls Butte quadrangle (Hietanen, 1963a).

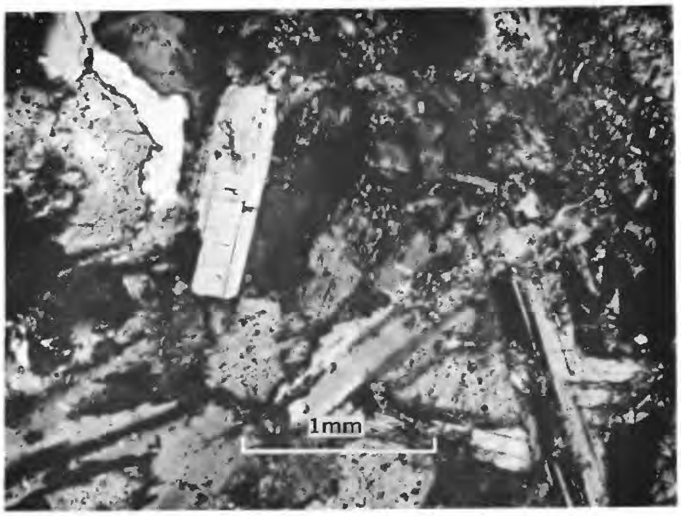

FIGURE 17.—Photomicrograph of gabbroic rocks in Wishards sill. Plagioclase is in twinned laths and the dark mineral is hornblende; black mineral is magnetite. Gabbroic rock is from St. Joe River, about a mile west of the mouth of Mosquito Creek (loc. A–298).

In contrast to the dikes in the southern part of the area, the specimens collected from the Wishards sill along the St. Joe River are highly altered and therefore seem different (older?) from the pyroxene gabbro in the Boehls Butte quadrangle. In most of the pyroxene gabbro there, only narrow rims of hornblende surround the augite crystals; exceptions are the chilled border zones of the large masses and most of the rock in the small bodies, which are more altered and in which secondary hornblende is a major dark constituent. It thus seems that the degree of alteration may depend on the proximity to the contact with the enclosing metasedimentary rocks and on the size of the body. The sill along the St. Joe River is thin in comparison with the larger masses in the Boehls Butte quadrangle. It is very likely that the metamorphic grade of the country rock also had a considerable influence—the Wishards sill was emplaced into a lower metamorphic grade rock that contained more water than higher grade equivalents. Moreover, some of the sills and dikes near the St. Joe River have been emplaced along fault zones near which alterations are extremely common elsewhere in the area.

It is not possible to tell on geologic grounds alone whether or not all pyroxene gabbro is of the same age. Correlation of the Wishards sill with the masses of pyroxene gabbro farther south would suggest an age younger than the early Late Cretaceous age of the Idaho batholith. On the geologic map, the age is tentatively given as Tertiary. A possibility of a late age for the northern part of the Wishards sill exposed in Mineral County, Mont., has been expressed by Wallace and Hosterman (1956).

The medium-grained gabbroic dikes on Goat Ridge contain more plagioclase and are lighter in color than the Wishards sill. The major constituents are plagioclase, augite, and two types of amphibole. Plagioclase (An_{16}) occurs in long lath-shaped crystals that include abundant alteration products—epidote, sericite, and chlorite. Most of the augite is fresh, but parts of some crystals are altered to hornblende or to chlorite with some leucoxene and magnetite. A few grains of green hornblende show $\gamma=1.680\pm0.002$, but most have $\gamma=1.652\pm0.002$. The latter occurs in aggregates and radiating groups of long slender prisms. The aggregates also include chlorite, epidote, magnetite, and leucoxene and are apparently pseudomorphs after some other ferromagnesian minerals, probably olivine and pyroxene. The accessory minerals are sphene and apatite.

DIORITIC DIKES

Most of the dioritic dikes are fine-grained hornblende-plagioclase rocks with some quartz, biotite, chlorite, and magnetite. Plagioclase and hornblende occur as long subhedral crystals having random orientation; the other minerals are interstitial. Some of the dikes contain scattered or clustered small phenocrysts of plagioclase; others are equigranular. Grains of sphene and long prisms of apatite occur as accessory minerals.

In many dikes, a part of biotite and hornblende are altered to chlorite. Chlorite also forms aggregates that have shapes of augite and include calcite. Some dikes, such as those near Granite Peak (loc. 2141), contain abundant epidote as an alteration product in plagioclase. This dike resembles the gabbroic dike on Goat Ridge except for the lack of augite.

A medium-gray porphyritic dike on the south side of Craig Lake consists of about 60 percent plagioclase (An_{32}), 15 percent augite, 10–15 percent chlorite, 5 percent epidote, and some quartz, calcite, hornblende, ilmenite-magnetite, and apatite. Epidote occurs as grains among the other minerals, as inclusions in plagioclase, and as round to oval amygdules that are sparsely scattered through the rock. The amygdules are 5–8 mm long and pistachio green in hand specimen. Under the microscope, epidote is pleochroic in yellow and shows $\alpha=1.732\pm0.002$, $\beta=1.752\pm0.002$, and $\gamma=1.768\pm0.002$, and $-2V\approx80°$—properties that would according to Winchell and Winchell (1951), indicate a pistacite with 28 percent of the iron end member. Many grains of an iron-ore mineral show lamellar intergrowths of magnetite and ilmenite, the latter partly altered to leucoxene. Apatite occurs in long slender prisms. This and other similar augite-bearing dioritic dikes form a group intermediate between the gabbroic and quartz monzonitic dikes.

QUARTZ MONZONITIC DIKES AND SILLS

Quartz monzonitic dikes and sills are most common near the two large quartz monzonite plutons exposed along the southwestern and northwestern borders of the mapped area. The largest mass borders the Beaver Creek pluton in the south and other fairly large dike-like bodies are along the northern border. Many small dikes (not shown on pl. 2) occur around the eastern end of the Roundtop pluton in the northwest.

Most of the dikes and sills consist of porphyritic fine- to medium-grained light-gray rock in which zoned plagioclase, orthoclase, quartz, and biotite occur as small phenocrysts (fig. 18). The groundmass is fine grained or very fine grained and consists of plagioclase, orthoclase, quartz, and biotite. Zircon, apatite, and magnetite are the accessory minerals. In a common type, the phenocrysts of quartz and plagoclase are 0.5–2 mm in diameter, whereas those of biotite are much smaller (0.4 mm). Some dikes are equigranular, the plagioclase occurring in euhedral to subhedral blocky laths that are strongly zoned. Such dikes also contain hornblende and less orthoclase than the normal porphyritic dikes. In their composition, these dikes are close

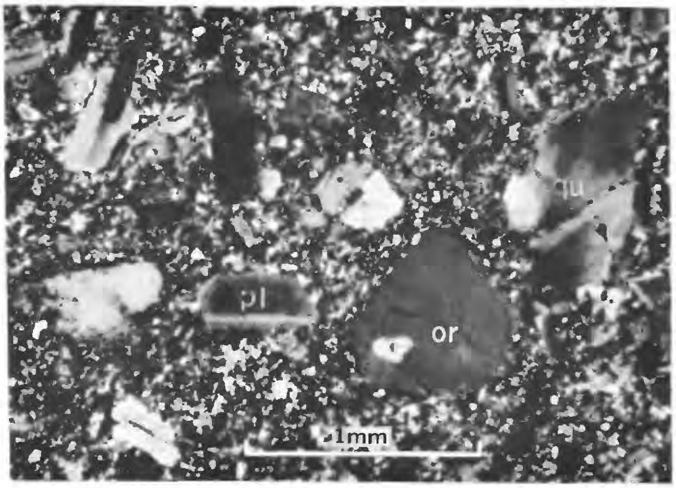

FIGURE 18.—Photomicrograph of a typical quartz monzonitic dike rock. Phenocrysts are quartz (qu), plagioclase (pl), and orthoclase (or); dark mineral is biotite. Dike rock is from Little Washington Creek (loc. 2050).

to the monzotonalite that occurs along the northern border zone of the Idaho batholith (Hietanen, 1963c).

GRANITIC DIKES

Most granitic dikes in this area are light brownish gray and contain abundant large phenocrysts of quartz, orthoclase, and plagioclase. These dikes can be easily distinguished in the field from the fine-grained gray quartz monzonitic dikes, but there are also fine-grained dikes whose composition can be identified only by examining their constituent minerals under the microscope. Phenocrysts in the common type of granitic dikes are 0.5–1 cm in diameter and grain size in the groundmass, which consists of quartz, plagioclase, orthoclase, and biotite, is 0.05–0.2 mm. The accessory minerals are magnetite, sphene, epidote, and apatite.

Granophyric and sperulitic textures (fig. 19) occur in a spotty way in the groundmass of many porphyritic dikes, and parts of some fine-grained dikes consist entirely of spherulitic intergrowths of feldspars and quartz. Needles of hornblende and biotite transect the spherules.

Figure 19.—Spherules consisting of granophyric intergrowth of quartz and feldspars in a granitic dike. South side of Eagle Creek (loc. 2155).

Granophyric texture indicates eutectic crystallization, and the chemical composition of the dike rock analyzed falls within the "ternary" minimum (table 2) if plotted on a Q-Or-Ab diagram. Comparison with the composition of the plutonic and hypabyssal rocks from the northwestern part of the Idaho batholith (Hietanen 1963c, table 1) shows that the granophyric dikes (table 2) form an end member of the quartz monzonite series. Plagioclase that constitutes a little more than a third of the rock is albite (An_3) and the amount of orthoclase

Table 2.—*Chemical composition in weight and ionic percentages, molecular norm, and mode of granitic dike No. 2155*

[Analyst, C. I. Parker; spectrographic determinations, Harriet Neiman]

Weight percent		*Ionic percent*		*Molecular norm*	
SiO_2	75.85	SiO_2	70.98	Q	29.85
TiO_2	.11	TiO_2	.08	Ab	37.75
Al_2O_3	12.55	$AlO_{3/2}$	13.85	An	1.17
Fe_2O_3	.75	$FeO_{3/2}$	.53	Or	29.15
FeO	.88	FeO	.69	Wo	.28
MnO	.04	MnO	.03	En	.02
MgO	.01	MgO	.01	Fs	.74
CaO	.42	CaO	.42	Ap	.04
Na_2O	4.16	$NaO_{1/2}$	7.55	Il	.16
K_2O	4.88	$KO_{1/2}$	5.83	Mt	.80
P_2O_5	.01	$PO_{5/2}$	.01	Cc	.04
CO_2	.02	CO_2	.02		
Cl	.01	Cl	(.02)	Total	100.00
F	.10	F	(.30)	Q=30.85	
H_2O^+	.09	OH	(.56)	Or=30.13	
H_2O^-	.03			Ab=39.02	
		Total	100.00		
Total	99.91	O	171.34		100.00
Less O	.04	OH	.56		
		Cl	.02		
	99.87	F	.30		
		Total anions	172.22		

Molecular mode

Quartz	30.19
Albite (An_3)	38.63
Orthoclase	28.65
Biotite	1.00
Hornblende	1.00
Magnetite	.75
Calcite	.04
Apatite	.04
Fluorite	.08
Total	100.38
+Cl	.02
+F	.18
$+H_2O$	.15

Composition used in calculating the molecular modes:

Biotite: $(H,F)_4K_2Fe_6Al_2Si_6O_{24}$

Hornblende: $NaCa_2Fe_5AlSi_7O_{22}(OH)_2$

Trace elements, in parts per million

[Looked for but not detected: Ag, As, Au, Bi, Cd, Ce, Co, Cr, Ge, Hf, Hg, In, La, Li, Mo, Ni, Pd, Pt, Re, Sb, Sc, Sn, Ta, Te, Tb, Tl, U, V, W, Y, Yb, Zn]

B	30	Cu	5	Pb	30
Ba	1,000	Ga	30	Sr	150
Be	3	Nb	10	Zr	50

is higher than that in the granite of the batholith. Because the amount of magnesium in the analysis is negligible, the small amounts of biotite and hornblende present must therefore be rich in iron. Only a few trace elements were detected (table 2), and of these only barium occurs in large amounts. Its amount is of the same order as in the quartz monzonite exposed along Beaver Creek just west of the area shown on plate 1 (Hietanen 1963c, table 3, no. 1064); strontium and zirconium, which follow in the order of abundance, are much less abundant than in the quartz monzonite.

CONCLUSIONS

The metasedimentary rocks of the area are divisible into formations that can be correlated with the five lowest formations of the Precambrian Belt Series—Prichard, Burke, Revett, St. Regis, and Wallace. The Prichard, Burke, and Wallace Formations were further subdivided into two or more mappable units on the basis of lithology. The schist units consist mainly of argillaceous material with some quartzite and are mineralogically much alike in every formation. The quartzite units show characteristic variation in structures, textures, and mineralogy which allow the identification of the formations where faulting has interrupted the stratigraphic sequence.

The grade of metamorphism increases from the muscovite-biotite subfacies of the greenschist facies near the St. Joe River in the north through epidote-amphibolite facies to muscovite-sillimanite subfacies of the amphibolite facies near the North Fork of the Clearwater River in the south. Within this 32-mile distance the following metamorphic zones were mapped: biotite, garnet, staurolite, staurolite-kyanite, kyanite, kyanite-sillimanite, and sillimanite. The isograds follow the periphery of the mapped batholith and crosscut the structural trends. The grade of metamorphism in any particular place depends mainly on three factors: distance from the batholith, stratigraphic unit, and structural position. The influence of these three factors is strikingly different and identifiable across late faults with vertical displacement.

The orientation of micas parallel to the axial planes indicates that the recrystallization was contemporaneous with the deformation. It is noteworthy that this well-developed orientation occurs also in the overturned folds that are most likely related to faults, such as those northeast of Bathtub Mountain. This orientation indicates that most faults have a long history; they were active through the period of major deformation and recrystallization, and many continued to be active after the emplacement of igneous rocks, as is shown by the occurrence of crushed and sheared rock along them. The isograds show no notable offset, however, partly because isograds can be shown only approximately, but also because there were no late displacements of great magnitude. Such displacements are striking farther west where rocks of the Prichard Formation were metamorphosed to a kyanite-staurolite subfacies and were brought into fault contact with rocks of the Wallace Formation of the biotite grade (Hietanen, 1967).

In general, the grade of metamorphism increases toward the batholith, but it also increases toward the lower stratigraphic units. A general explanation is that the increase of metamorphism toward the batholith is due mainly to an increase in temperature and that toward the lower stratigraphic units is caused by the combined effect of higher pressure and temperature. An example of the influence of increase due mainly to an increase in temperature is seen in the gently folded argillaceous unit of the Wallace Formation south of the St. Joe River. The grain size of the "groundmass" changes only little, but new index minerals—such as biotite, garnet, staurolite, and kyanite—appear as porphyroblasts in the successive zones toward the southwest. In contrast, grain size in the lower stratigraphic units (the Prichard and the Burke) within the same mineralogic zone (for example, in the kyanite zone) is much larger than the common grain size in the higher units such as the Wallace.

A marked difference was observed in the texture of the quartzite units. Quartz grains in the upper stratigraphic units, such as the Wallace Formation, are fairly equidimensional but become increasingly flattened toward the lower units. In the outcrops this appears as better developed bedding-plane foliation in the lower units. This difference in texture is partially preserved in the zone next to the batholith (Hietanen, 1962a).

Grain size in all stratigraphic units increases southward with the grade of metamorphism. Increase in the intensity of folding and deformation toward the south suggests that directed pressures were larger in the south and probably contributed, together with the higher temperature, to the larger grain size. This view is supported by the fact that in the central part of the area the places of strong deformation, such as vicinities of overthrusts, are places of larger grain size than common elsewhere in the same metamorphic zone.

The order of crystallization, as indicated by pseudomorphs after staurolite in the kyanite-staurolite zone and by textures, such as orientation of inclusions elsewhere, proves that there were at least two episodes of metamorphism, the earlier synkinematic and the later postkinematic. The isograds were moved 1–2 miles northward from their earlier position during the second episode. The grade of both episodes and the intensity of deformation increase toward the Idaho batholith. During the first episode of recrystallization (synkinematic), the temperature near the batholith may have reached a maximum of about 600°C and the pressure about 6,000 atm. Under these conditions, grantitic magma started to form at lower levels during the synkinematic phase. When this magma was emplaced to a higher level, it brought more heat to the level new exposed. This heat caused the second episode of metamorphism (postkinematic) of the country rock. Thus the two episodes of metamorphism described in this paper are considered to be intimately associated with these two phases in the formation of the batholith.

This conclusion is in accordance with those arrived at in earlier reports (Hietanen 1961b, 1962a, 1963c). The emplacement of the large bodies of quartz diorite, quartz monzonite, and granite of the Idaho batholith was found to be later than the major period of deformation and recrystallization. Formation of second-generation minerals in the zone next to the batholith was determined to be postkinematic and considered to be a result of an introduction of certain elements during the emplacement of the rocks of the batholith. No elements were introduced in areas farther from the batholith, but the later episode of recrystallization can be identified because of relict textures.

A similar relation between the deformation, metamorphism, and intrusion is probably common in other areas where large masses of granitoid rocks are formed through partial melting of a part of the crust at depth. The partial melting is an ultimate result of elevated temperatures during the deformation. In the cooler outer zones, the rocks would respond through recrystallization, the metamorphic grade decreasing away from the heat center. This synkinematic recrystallization would be the first. When the magmas start to move and are emplaced into cooler outer zones, they will bring more heat there and cause a second (late kinematic or postkinematic) recrystallization. The relict textures indicating the sequence of events of this more or less continuing process are not always preserved, especially in the zones of high-grade metamorphism where either the second recrystallization obliterated the earlier textures or where no readjustments were necessary because of similar pressure and temperature conditions during the first and second episode.

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