Belt Series in the Region Around Snow Peak and Mallard Peak, Idaho

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By ANNA HIETANEN

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

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Petrologic and structural study of a part of the northwestern contact aureole of the Idaho batholith

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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

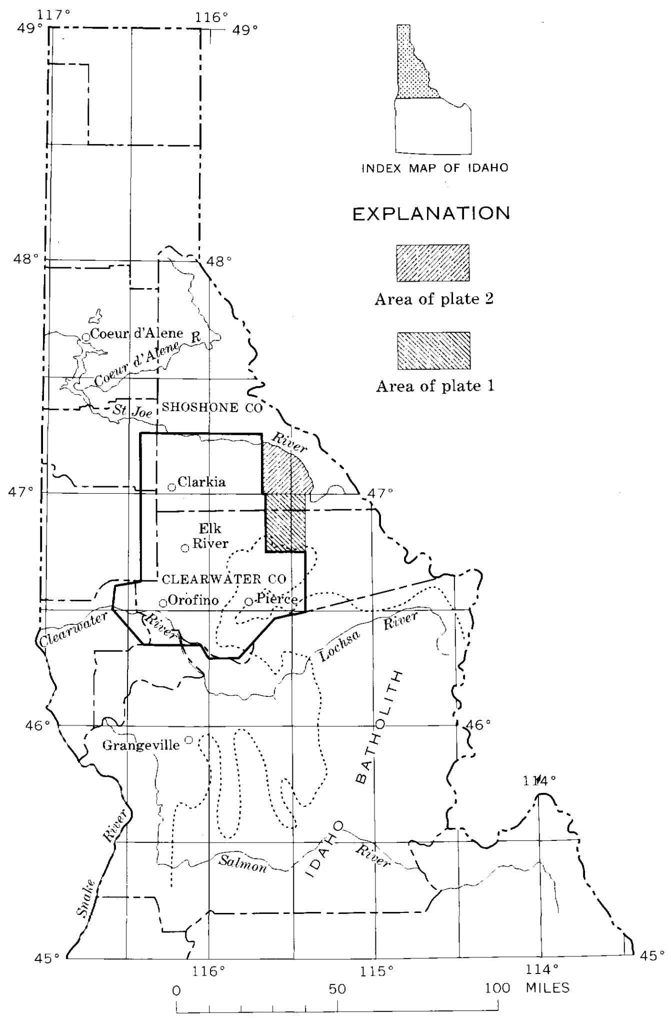

FIGURE 1.—Index map of northern Idaho. Outline of northern part of the Idaho batholith is shown by dotted line. Outline of the area described previously (Hietanen, 1962a, 1963a, b, c, 1967) is shown by heavy line.

(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.

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	500	
				Garnet-mica schist		
		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	

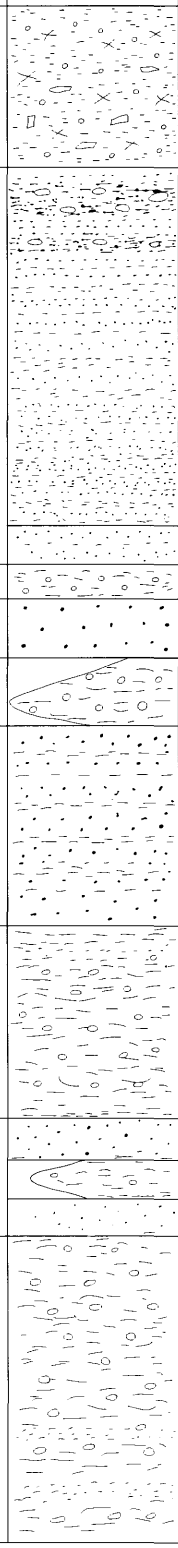

FIGURE 2.—Generalized stratigraphic section of the Belt Series near Snow Peak.

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.

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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).

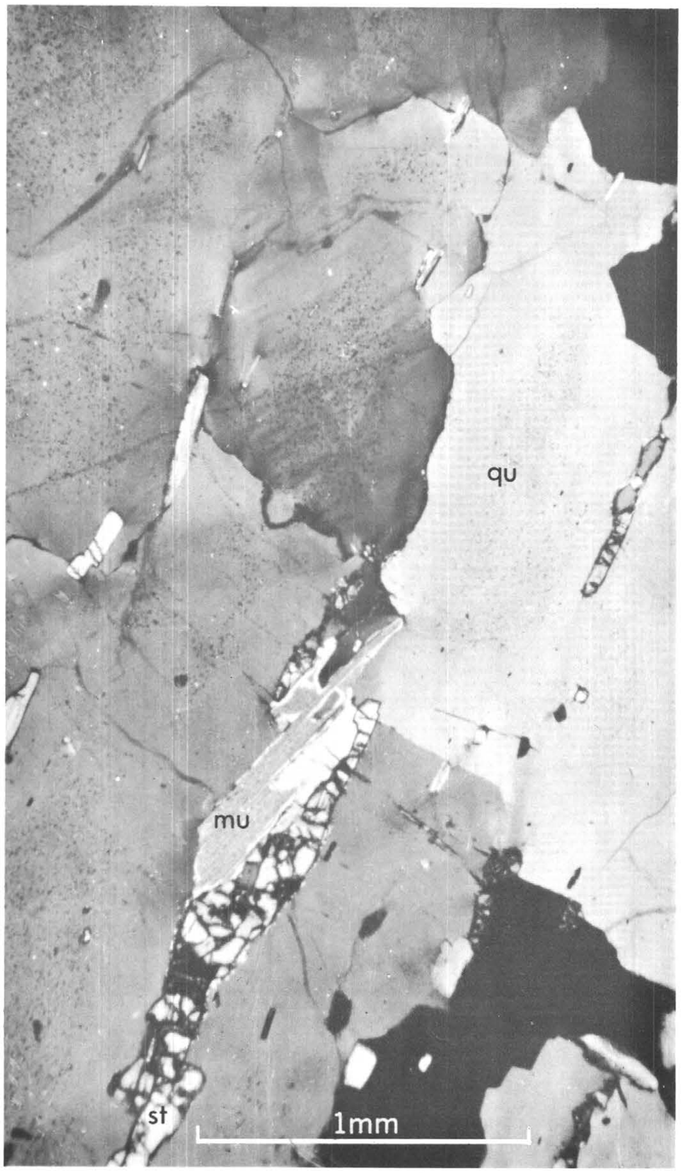

FIGURE 3.—Photomicrograph of quartzite of the Burke Formation. Small mica flakes (mu) and staurolite (st) crystals are parallel to the bedding and are included in large quartz grains (qu). Sawtooth Creek (loc. 1550).

The crystals of garnet, ranging from 2 to 6 mm in diameter, are more common in the lower part of the unit.

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.

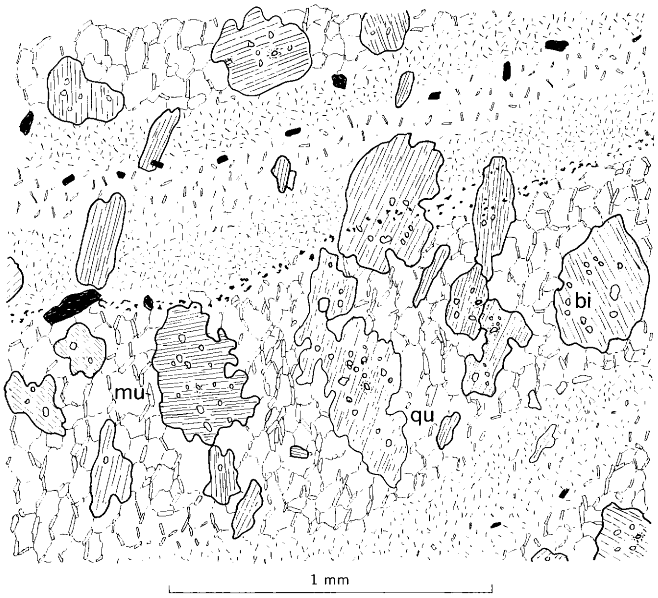

FIGURE 4.—Camera lucida drawing of biotite (bi) porphyroblasts in biotite zone; mu, muscovite; qu, quartz; black is ilmenite. Argillitic unit of the Wallace Formation, West Sister (loc. A–27). Micas are subparallel to cleavage that transects the bedding.

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 ($\gamma=1.644\pm0.002$) 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: $\alpha=1.633\pm0.001$, $\beta=1.642\pm0.001$ and $\gamma=1.651\pm0.001$.

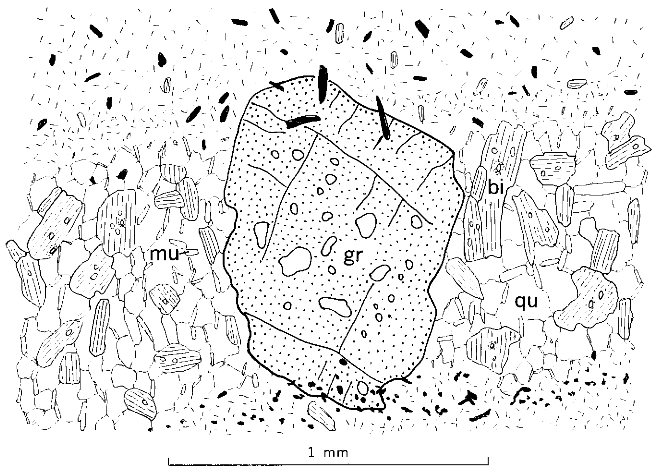

FIGURE 5.—Camera lucida drawing of garnet (gr) in garnet zone; bi, biotite; mu, muscovite; qu, quartz; black is ilmenite and magnetite. Inclusions in garnet are quartz, ilmenite, and magnetite, and their distribution reflects the distribution of these minerals in beds outside the garnet.

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 $+2V\approx30°$.

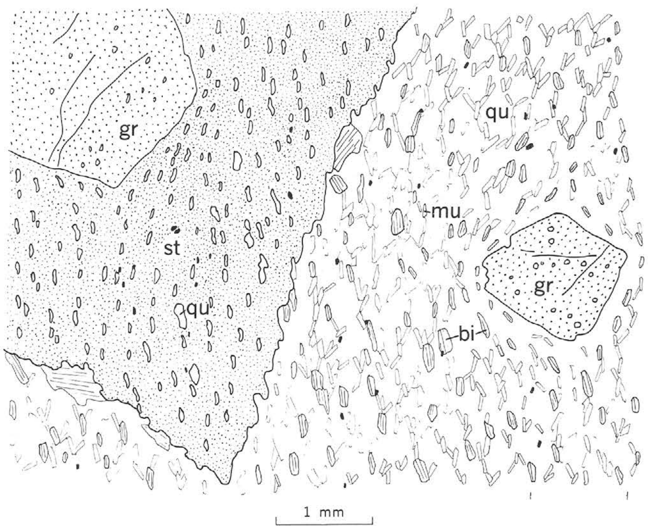

FIGURE 6.—Camera lucida drawing of a postkinematic staurolite (st) in garnet-mica schist of the staurolite zone north of Junction Peak (loc. A–143). Foliation bends around garnet (gr) but continues undisturbed through the late staurolite crystals as shown by the elongated quartz (qu) grains that are included in staurolite; mu, muscovite; bi, biotite.

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.

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

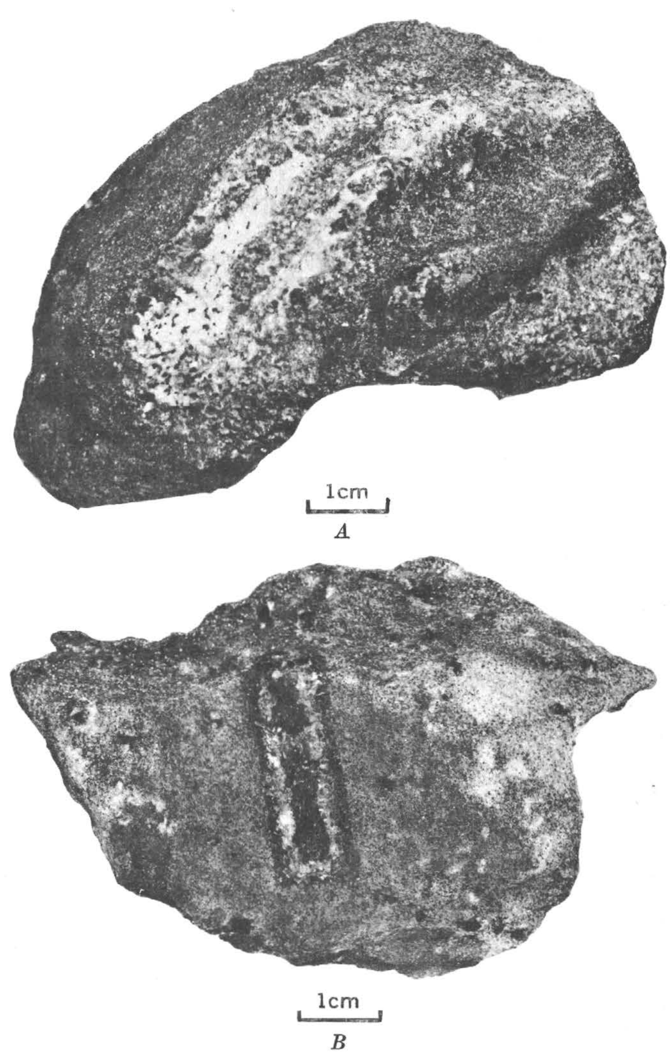

FIGURE 7.—Pseudomorphs after staurolite. *A*, Light-colored central part is a pseudomorph that consists of muscovite, kyanite, garnet, biotite, and some staurolite; Papoose Creek (loc. 2093). *B*, Staurolite crystal altered to an aggregate, the center of which consists of staurolite and garnet and the rim of muscovite and kyanite; west of Peggy Peak (loc. A–156).

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.

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 $\gamma = 1.625 \pm 0.001$; 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,

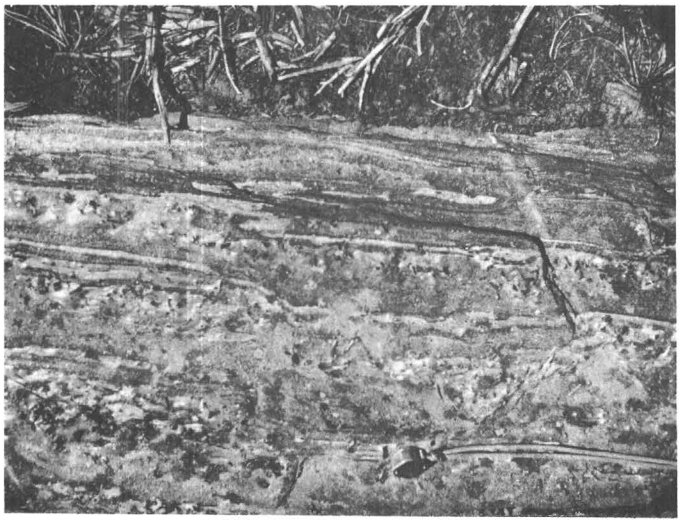

FIGURE 8.—Staurolite schist a mile southeast of Bathtub Mountain (loc. 2096). Distribution of the euhedral prisms of staurolite follows certain beds.

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

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				
		Total anions	187.44		
	100.15				

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.

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

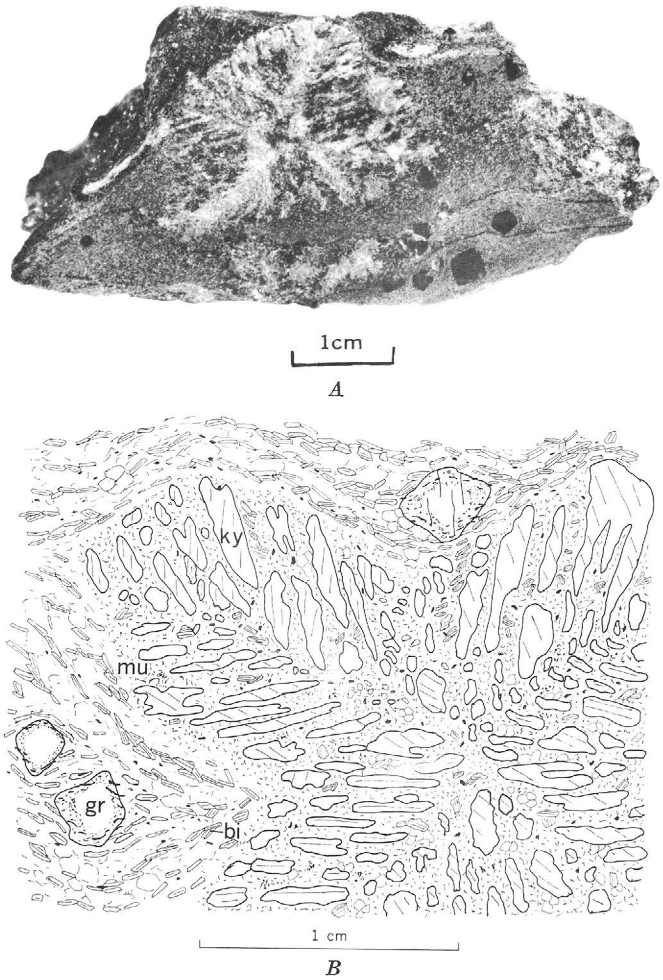

FIGURE 9.—Herringbone structure in a pseudomorph after a staurolite twin. *A*, Photograph of specimen from a mile east of Papoose Mountain (loc. 2104). *B*, Drawing made from thin section of specimen in figure 9*A* shows that the pseudomorph consists of kyanite (ky), muscovite (mu), biotite (bi), and 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:

$$Al_4(OH)_8Si_4O_{10} + Fe_2O_3 \rightarrow 2\ FeAl(OH_2)SiAlO_5 + 2\ SiO_2 + 2\ H_2O + O \quad (1)$$

Kaolinite + hematite → 2 chloritoid + 2 quartz + 2 water

$$9\ FeAl(OH)_2SiAlO_5 \rightarrow 2\ HFe_2Al_9O_8Si_4O_{16} + SiO_2 + 5\ FeO + 8\ H_2O \quad (2)$$

9 chloritoid → 2 staurolite + quartz + 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 + staurolite + biotite. (3)

This reaction can be expressed by three simpler equations:

$$3(OH)_4K_2Al_6Si_6O_{20} + 4\ FeO \rightarrow 2\ HFe_2Al_9Si_4O_{24} + 3\ K_2O + 10\ SiO_2 + 5\ H_2O \quad (3a)$$

3 muscovite + 4 iron oxide → 2 staurolite + 3 potassium oxide + 10 quartz + 5 water

$$(OH)_8(Mg,Fe)_6Si_4O_{10} + (OH)_4K_2Al_6Si_6O_{20} + 3\ FeO \rightarrow 3\ (Fe,Mg)_3Al_2Si_3O_{12} + SiO_2 + K_2O + 6\ H_2O \quad (3b)$$

chlorite + muscovite + 3 iron oxide → 3 almandite + quartz + potassium oxide + 6 water

The potassium freed in these reactions would concurrently react with chlorite to form biotite as follows:

$$(OH)_8(Mg,Fe)_4Al_4Si_2O_{10} + K_2O + FeO + 3\ SiO_2 \rightarrow (OH)_4K_2(Fe,Mg)_5Al_4Si_5O_{20} + 2\ H_2O \quad (3c)$$

chlorite + potassium oxide + iron oxide + 3 quartz → 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, ferriantigorite, 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:

$$10\ HFe_2Al_9Si_4O_{24} + 4\ (OH)_4K_2Al_6Si_6O_{20} + 5\ SiO_2 \rightarrow 4\ (OH)_4K_2Fe_5Al_4Si_5O_{20} + 49\ Al_2SiO_5 + 5\ H_2O \quad (4a)$$

10 staurolite + 4 muscovite + 5 quartz → 4 biotite + 49 kyanite + 5 water

$$6\ HFe_2Al_9Si_4O_{24} + 17\ SiO_2 + 2\ K_2O + H_2O \rightarrow 4\ Fe_3Al_2Si_3O_{12} + 2\ (OH)_4K_2Al_6Si_6O_{20} + 17\ Al_2SiO_5 \quad (4b)$$

6 staurolite + 17 quartz + 2 potassium oxide + water → 4 almandite + 2 muscovite + 17 kyanite

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:

$$\underset{\text{muscovite}}{(OH)_4K_2Al_6Si_6O_{20}} \rightarrow \underset{\text{3 kyanite}}{3\ Al_2SiO_5} + \underset{\text{3 quartz}}{3\ SiO_2} + \underset{\text{potassium oxide}}{K_2O} + \underset{\text{2 water}}{2\ H_2O} \qquad (5)$$

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:

$$\underset{\text{biotite}}{(OH)_4K_2(Fe,Mg)_6Al_2Si_6O_{20}} + 2O_2 \rightarrow \underset{\text{sillimanite}}{Al_2SiO_5} + 2Fe_3O_4 + 6SiO_2 + K_2O + 2H_2O. \qquad (6)$$

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.

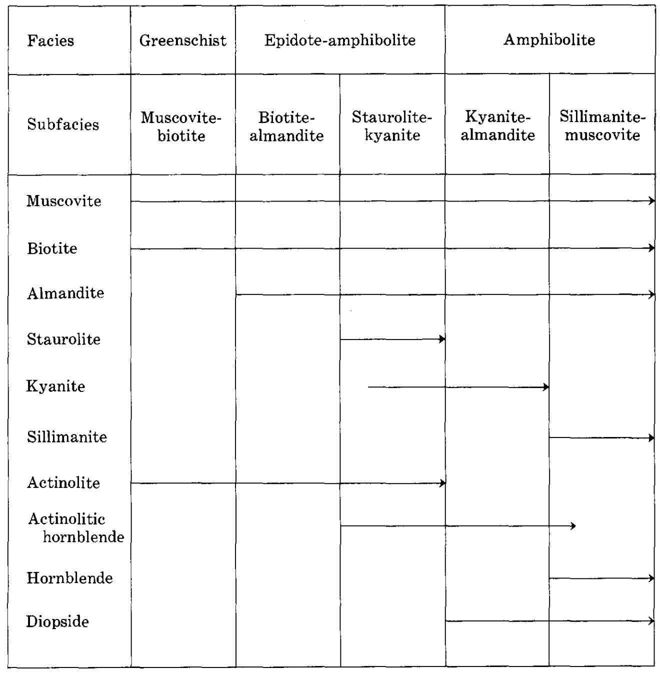

FIGURE 10.—Stability ranges of the index minerals in the area studied.

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 $300° \pm 50°C$ 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.

[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.

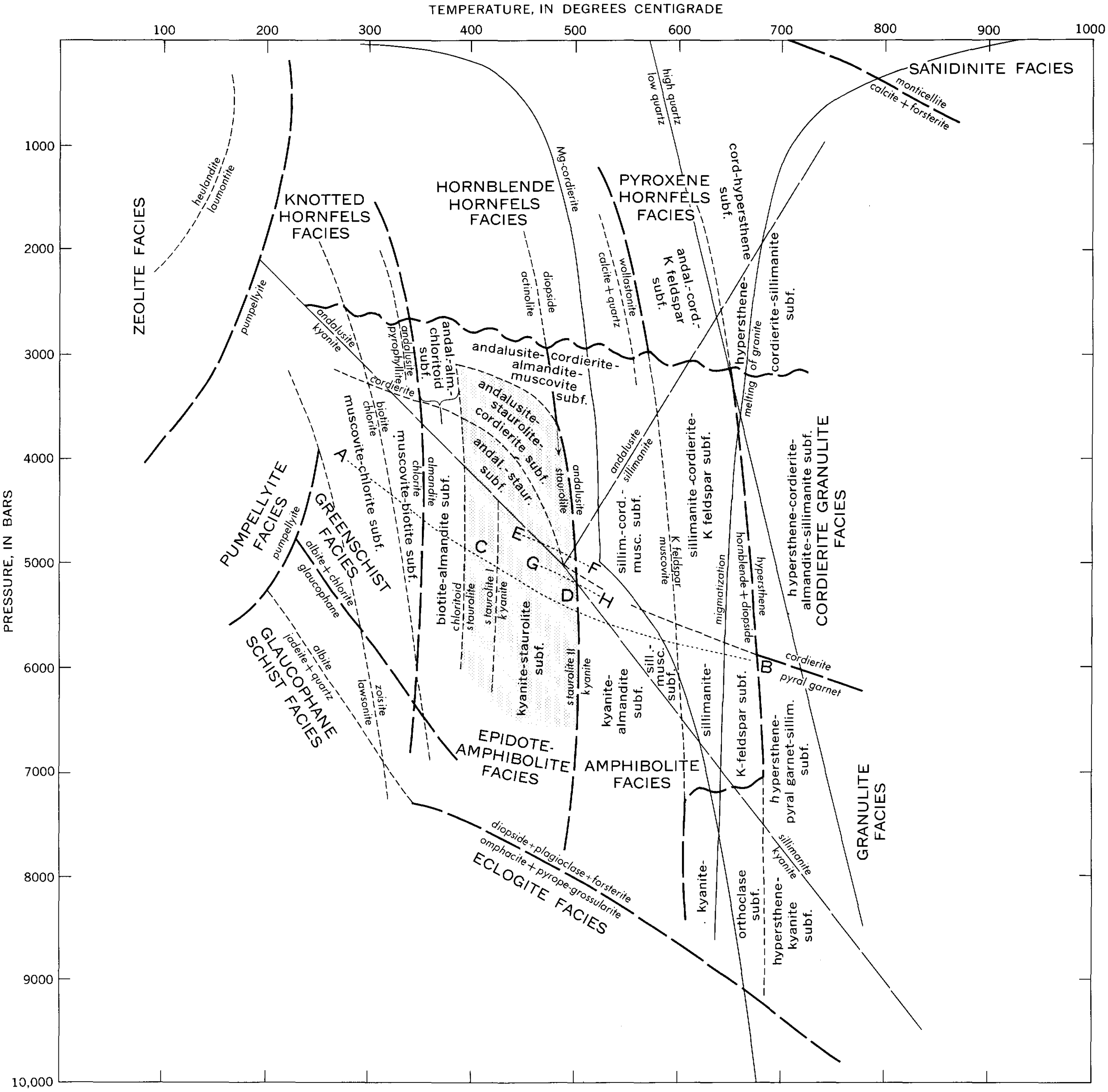

FIGURE 11.—A tentative pressure-temperature diagram showing a possible pressure-temperature gradient of recrystallization (dotted line A–B) and the stability fields of the subfacies north of the Idaho batholith. For comparison, pressure-temperature gradient for the rocks in the neighboring Boehls Butte 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.

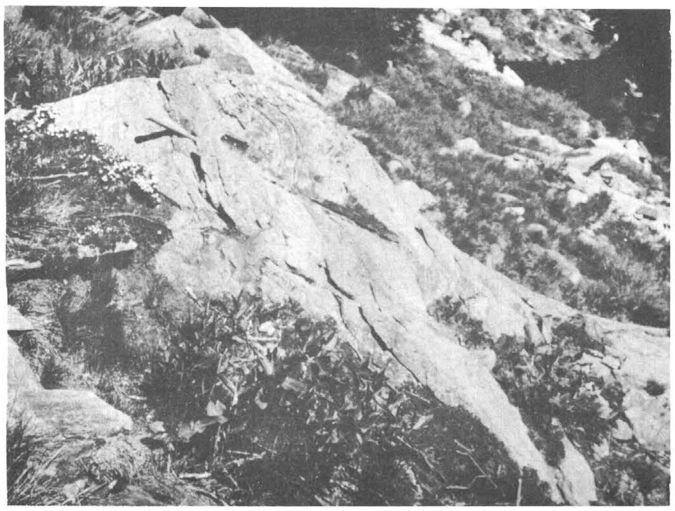

FIGURE 12.—Two sets of folds in garnet-mica schist of the Prichard Formation half a mile southwest of Black Mountain (loc. 2058). The handle of the hammer (upper left) is parallel to the axis of small folds plunging 30° S. In the photograph, the axis of the gentle large syncline plunges away from the 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.

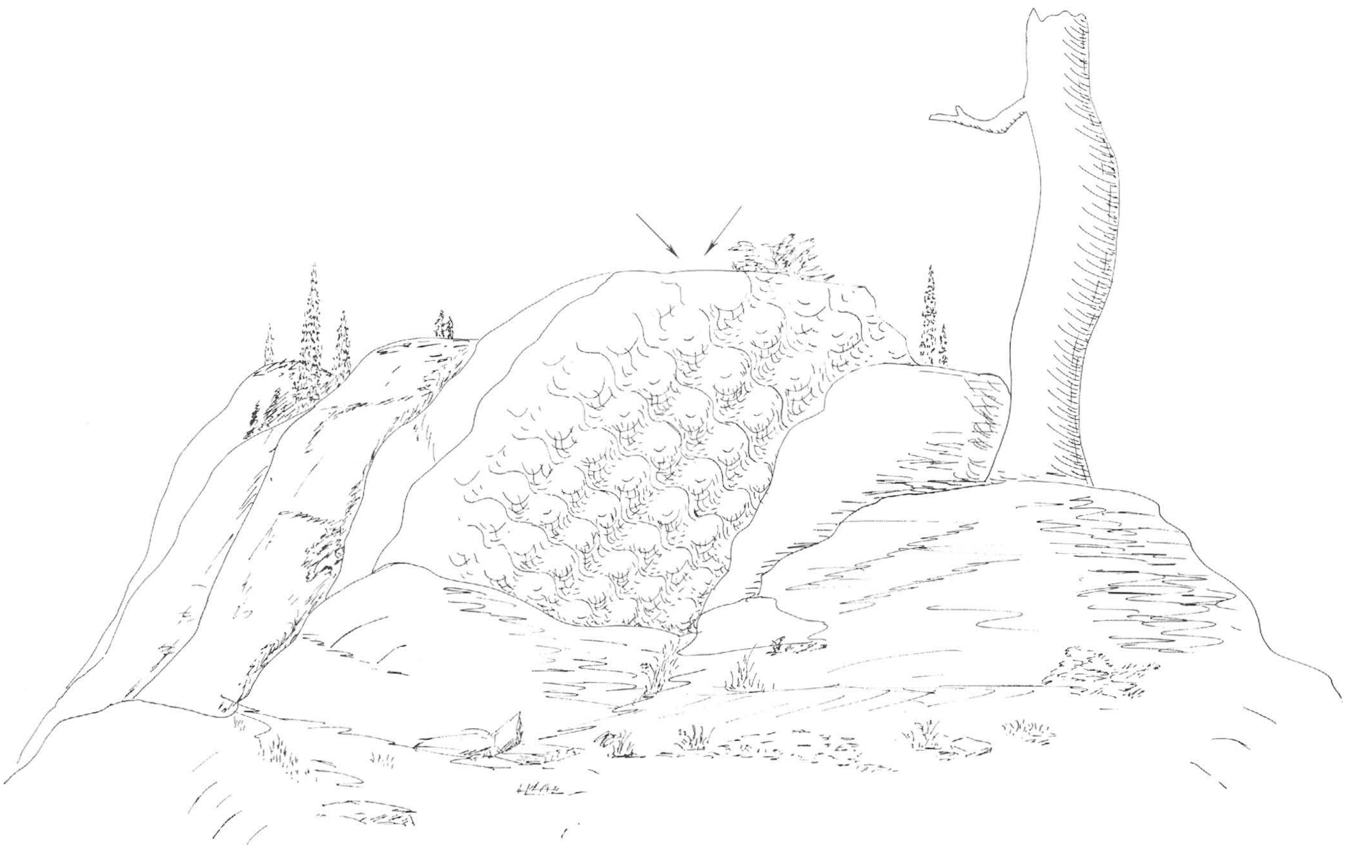

FIGURE 13.—Two sets of small folds in schist of the St. Regis Formation, 2 miles northeast of The Nub between Cliff Creek and Drift Creek. The arrows show the direction of fold axes. Folded surface faces westward and dips 70° away from the observer. The tree in the foreground is 30 cm thick.

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.

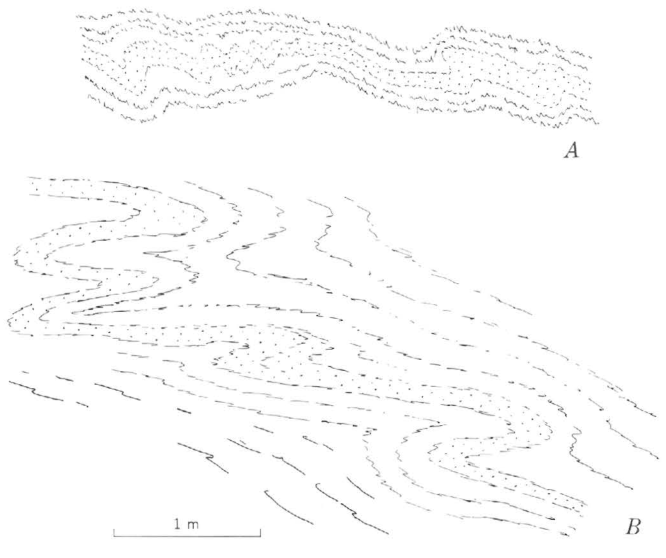

FIGURE 14.—Folding of quartzite layers (dotted) interbedded with schist that shows intense wrinkling instead of folding. *A*, Prichard Formation, southeast-facing vertical wall on the north slope of Black Mountain; *B*, St. Regis Formation, east-facing vertical wall on Canyon Creek, one-third of a mile 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,

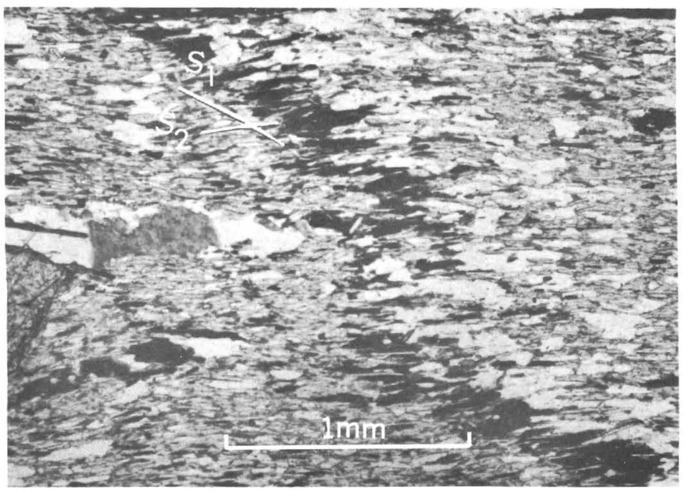

FIGURE 15.—Photomicrograph of garnet-mica schist showing bedding (s_1) and transection foliation (s_2). Micas are parallel to the foliation. Fly Creek, 1 mile east-southeast of Peggy Peak (loc. A–154).

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-

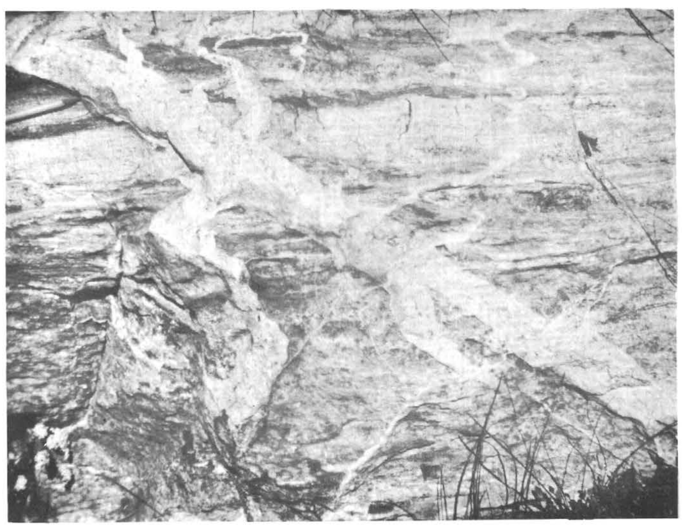

FIGURE 16.—Quartz vein in actinolite-diopside granofels 1.5 miles southwest of Dismal Lake. Note that there is no offset of the bedding planes. Pencil in the upper left corner is 10 cm.

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

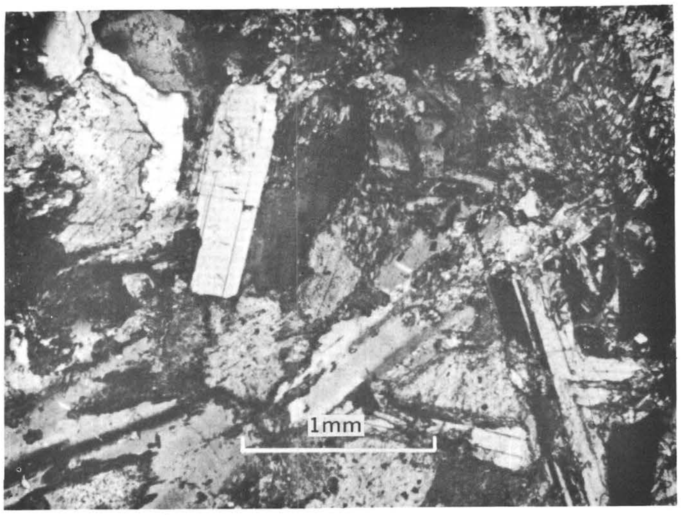

FIGURE 17.—Photomicrograph of gabbroic rocks in Wishards sill. Plagioclase is in twinned laths and the dark mineral is hornblende; black mineral is magnetite. Gabbroic rock is from St. Joe River, about a mile west of the mouth of Mosquito Creek (loc. A–298).

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).

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

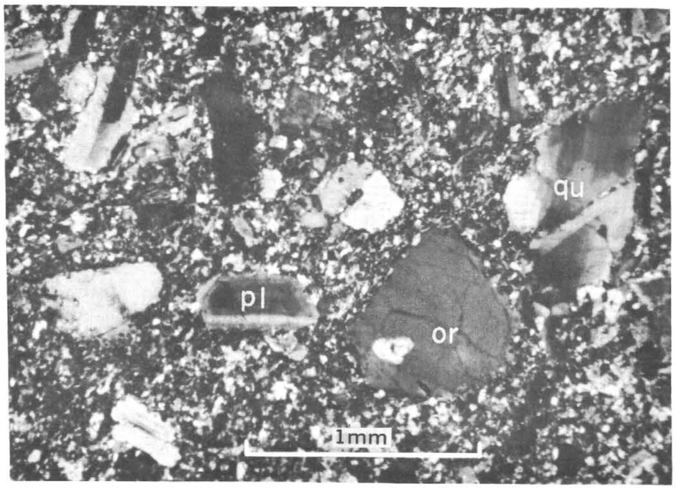

FIGURE 18.—Photomicrograph of a typical quartz monzonitic dike rock. Phenocrysts are quartz (qu), plagioclase (pl), and orthoclase (or); dark mineral is biotite. Dike rock is from Little Washington Creek (loc. 2050).

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.

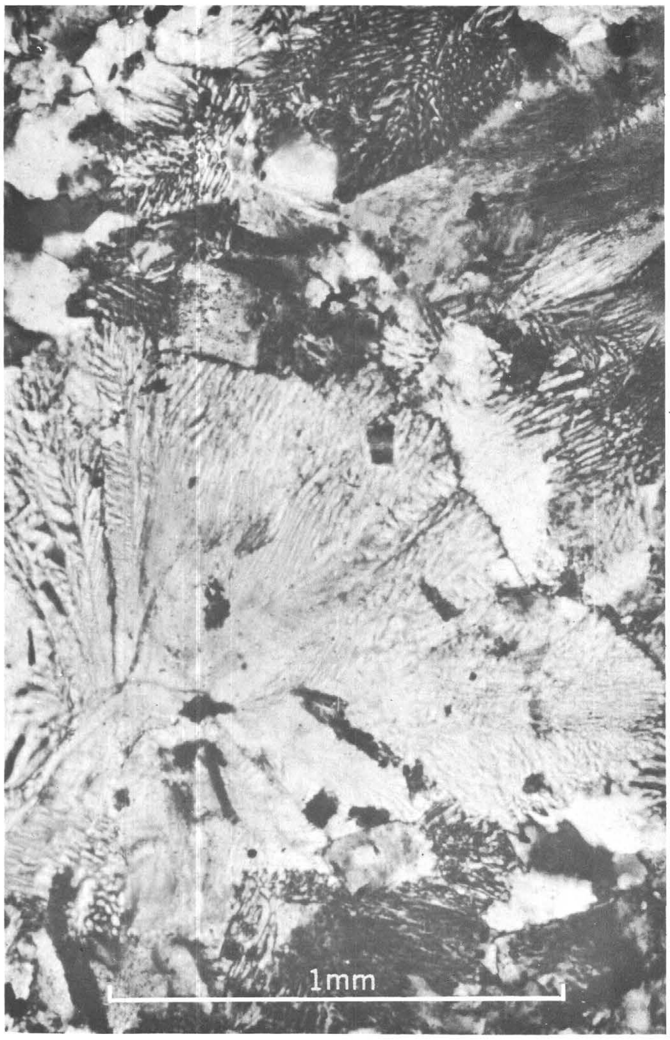

FIGURE 19.—Spherules consisting of granophyric intergrowth of quartz and feldspars in a granitic dike. South side of Eagle Creek (loc. 2155).

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	99.91	Total	100.00		100.00
Less O	.04	O	171.34		
		OH	.56		
	99.87	Cl	.02		
		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, Ti, 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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