Geology of the Mount Pinchot Quadrangle Southern Sierra Nevada California

GEOLOGICAL SURVEY BULLETIN 1130

Prepared in cooperation with the California Department of Natural Resources, Division of Mines

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By JAMES G. MOORE

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CONTENTS

ILLUSTRATIONS

[Plates are in pocket]

TABLES

GEOLOGY OF THE MOUNT PINCHOT QUADRANGLE, SOUTHERN SIERRA NEVADA, CALIFORNIA

By James G. Moore

ABSTRACT

The Mount Pinchot quadrangle, California, includes the main crest and eastern scarp of a part of the southern Sierra Nevada and a segment of the west side of Owens Valley. The maximum relief is nearly 10,000 feet, and more than one-half of the quadrangle lies higher than 10,000 feet above sea level.

The Sierra Nevada here is underlain chiefly by a mosaic of grantic intrusives of Cretaceous age which encloses small masses of pre-Cretaceous metamorphic rocks and hybrid mafic rocks. The pre-Cretaceous rocks include two series; an older series of metasedimentary rocks of Paleozoic(?) age, and a younger series of predominantly metavolcanic rocks of Mesozoic age. The Paleozoic(?) metasedimentary rocks are present as nearly vertical tabular septa of schist and hornfels derived from shale, limestone, and sandstone. Jurassic and Triassic metavolcanic rocks are predominantly from rhyolite tuff, but meta-andesite, metadacite, and metabasalt are also present. The largest roof pendant of metavolcanic rocks represents a vertical homoclinal section more than 7,000 feet thick with the tops of the beds to the southwest. Both metasedimentary and metavolcanic rocks are of the hornblende hornfels grade of metamorphism.

Twenty-seven separate granitic intrusive masses were mapped. Of these, 9 are older than a large swarm of northwest-trending mafic dikes which completely crosses the quadrangle, and 18 are younger. In composition the plutons range from alaskitic quartz monzonite to dark-colored granodiorite; the average composition is close to the boundary between quartz monzonite and granodiorite. The plutons are believed to have made room for themselves chiefly by forcible intrusion, but stoping and assimilation of wallrock were locally operative.

Six of the intrusive masses contain porphyritic potassium feldspar. The composition of the porphyritic plutons is restricted to an intermediate composition, close to the average composition of all the granitic rocks. Phenocrysts apparently grew after initial emplacement of the mass, but before final intrusion of the remobilized core of zoned plutons.

Mafic inclusions are present in most of the granitic rocks except the alaskites. Mafic inclusions are more abundant and larger in the more mafic plutons and in the thinner plutons. At least five of the plutons are zoned from a silicic core to a calcic margin. In some plutons the silicic core has remobilized and intruded its margins.

The main uplift of the range occurred in late Tertiary(?) and Pleistocene time along a zigzag fault zone that follows closely the base of the range. Normal faults downdropped on the west are common in benches along the range front west of the principal fault zone and are related to warping.

Late Cenozoic basalt cinder cones and lava flows occur in the eastern part of the quadrangle at the foot of the Sierra Nevada. Olivine basalts were extruded

from several of the bounding faults. The mafic character of the basalt and inclusions of spinel-bearing dunite indicate that these faults penetrate to great depth. At least four periods of volcanism can be recognized; the most recent formed the volcanic field southwest of Big Pine during the interval between the last two stages of Pleistocene glaciation.

Scheelite, an ore mineral of tungsten, is present in tactite near the contact of metamorphosed limy rocks and granitic rocks. The known deposits are small. Gold-silver veins of the now idle Kearsarge district cut granodiorite and metavolcanic rocks; they are related to an alaskite pluton barely breached by erosion.

INTRODUCTION

LOCATION AND ACCESSIBILITY

The Mount Pinchot 15-minute quadrangle is in southeastern California and includes a part of the crest and eastern scarp of the Sierra Nevada (fig. 1). It is bounded by lat. 36°45′ and 37°00′ N. and long. 118°15′ and 118°30′ W. The quadrangle includes 240 square miles, of

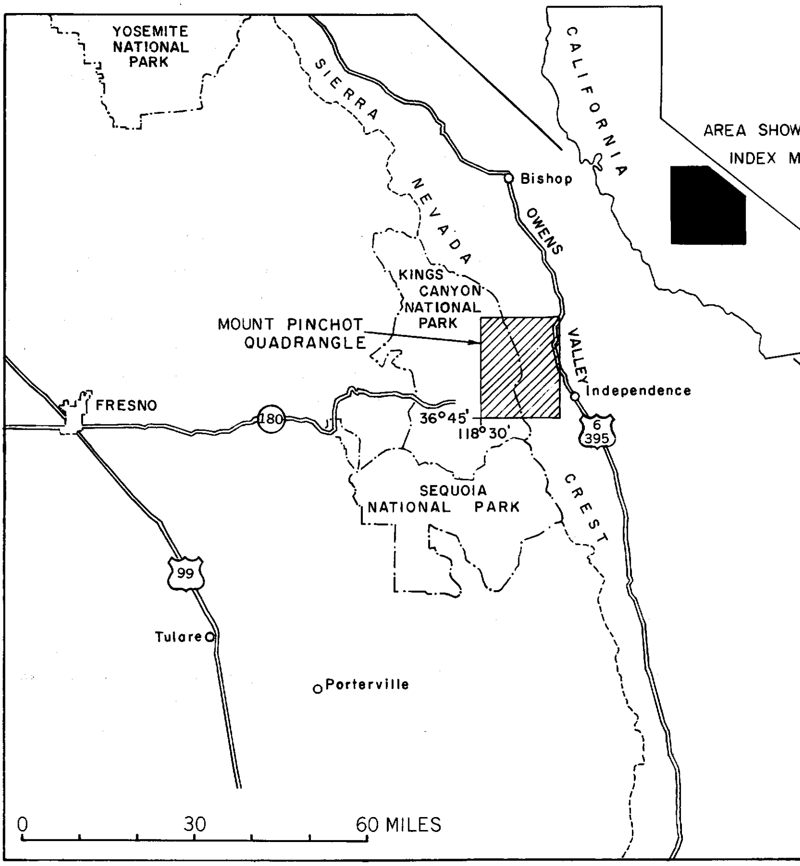

FIGURE 1.—Index map showing location of the Mount Pinchot quadrangle, Inyo and Fresno Counties, Calif.

which the major part is mountainous. The part east of the Sierra Nevada divide is in Inyo County, and that west of the divide is in Fresno County and within the Kings Canyon National Park.

U. S. Highway 395 (El Camino Sierra), the main route of travel to the area, runs near the east edge of the quadrangle in Owens Valley. From U. S. Highway 395 several small roads branch to the west and lead into the quadrangle and part way up the east scarp of the range. The most usable of these is the Onion Valley road in the southern part of the mapped area, which attains an elevation of 9,200 feet.

West of the Sierra crest there are no roads in the quadrangle, and the north-south John Muir Trail is the main route of travel. It can be reached from the west by trail from Cedar Grove at the terminus of State Route 180 in the canyon of the South Fork of the Kings River. Connecting the Muir trail with Owens Valley to the east are four high trails over the Sierra crest. These trails cross the divide at Taboose Pass (11,498 feet), Sawmill Pass (11,279 feet), Baxter Pass (12,320 feet), and Kearsarge Pass (11,160 feet), in order from north to south.

Permanent settlements in the quadrangle are at Aberdeen on Goodale Creek, Division Creek powerhouse on Division Creek, Parker Ranch on the South Fork of Oak Creek, and Seven Pines on Inpendence Creek. Pack stations are maintained in Onion Valley during the summer months. The nearest town is Independence, on Highway 395, a few miles east of the quadrangle boundary.

SURFACE FEATURES

The Mount Pinchot quadrangle is dominated by the main crest and eastern front of the Sierra Nevada (fig. 2). Within the boundaries of the quadrangle the trend of the Sierra is only slightly west of north; the entire range tends about N. 30° W.

The total relief within the quadrangle is nearly 10,000 feet, from 3,900 feet in Owens Valley to 13,495 feet on Mount Pinchot on the Sierra crest. More than one-half the area of the quadrangle lies above an altitude of 10,000 feet. The great eastern escarpment of the Sierra Nevada is the most striking topographic feature in the quadrangle (fig. 2). It rises 7,000 to 10,000 feet above the floor of Owens Valley and is incised by deep ravines, many of which were filled with glaciers during the Pleistocene. The range front is a single great escarpment, although it is irregular in plan and profile because of faulting.

The general slope on the west side of the crest is much less than that on the steep eastern scarp. However, glaciation on the western slope has produced an alpine topography of sharp ridges and peaks, broad cirques, and countless lakes.

FIGURE 2.—Aerial view looking west, showing approximately 10 miles of the Sierra Nevada escarpment. Dashed lines indicate principal east-dipping faults at foot of range, along which basalt cinder cones are alined. Small west-dipping faults occur on prominent range-front benches (A). The canyon of the South Fork of the Kings River is in the middle background, and the Great Valley of California lies beyond the most distant ridge crest (approximately 60 miles). Photograph by Symons Aerial Survey.

The mapped area west of the main divide lies in the watershed of the South Fork of the Kings River and its tributaries. It is drained by three major streams flowing west in deep U-shaped glacial canyons 3,000 to 4,000 feet deep. From north to south these master streams are the upper South Fork of the Kings River, Woods Creek, and Bubbs Creek. The three join west of the quadrangle and help compose the Kings River, which flows west into the Great Valley of California.

The eastern part of the quadrangle is occupied by an apron of coalescing alluvial fans sloping toward Owens Valley from the base of the Sierra Nevada. This apron forms a sloping plane 2 to 5 miles wide, inclined as much as 10°, and mantled by sand and gravel eroded from the mountains to the west.

Owens Valley is an arid basin without outlet, bounded by the fault scarps of the Sierra to the west and of the Inyo Range to the east. The valley is flat in its central part, where it contains unconsolidated sediments and Pleistocene lava flows. Precipitation falling east of the crest finds its way into Owens River. The water of the Owens River formerly flowed into Owens Lake, a typical desert sink without outlet, but now is carried by an aqueduct out of the valley to Los Angeles.

ACKNOWLEDGMENTS

Field work during 1953 and 1954 was done under the direction of the Department of Geology at The Johns Hopkins University, and a report on the eastern two-thirds of the quadrangle was prepared.[1] Laboratory work during this period was made possible by a National Science Foundation fellowship held from October 1953 to June 1954.

Work from June 1956 was done as part of a project of the U.S. Geological Survey in cooperation with the California Department of Natural Resources, Division of Mines. During this period, mapping of the quadrangle was completed and the earlier mapping was refined.

The writer was assisted in the field by J. G. Stone in 1956 and by C. A. Hopson in 1957.

PREVIOUS WORK

The Mount Pinchot quadrangle has received only general reconnaissance geologic study prior to this investigation. A reconnaissance study of the eastern Sierra Nevada by Knopf (1918) includes the eastern half of the quadrangle. Mayo (1941) made a general structural study of the entire east-central Sierra Nevada, which includes most of the Mount Pinchot quadrangle.

[1] Moore, J. G., 1954, Geology of the Sierra Nevada front near Mount Baxter, California: Baltimore, Johns Hopkins Univ. dissertation, 105 pp.

Geologic mapping in adjacent areas of the Sierra Nevada includes the Mineral King area (Knopf and Thelen, 1905), 20 miles to the south; the area around Three Rivers (Durrell, 1940), 30 miles to the southwest; an area northwest of Fresno (Krauskopf, 1953), 25 miles to the west; and the Bishop district, adjacent to the north (P. C. Bateman 1958; 1961).

METAMORPHIC ROCKS

Mayo (1941, p. 1032) has shown that two series of metamorphic rocks are present in the eastern Sierra Nevada: a Mesozoic metavolcanic series and a Paleozoic metasedimentary series. The metavolcanic rocks occur in a discontinuous septum at least 120 miles long, parallel to the trend of the range. A few miles east of this septum, and adjacent to it in places, is a discontinuous septum of metasedimentary rocks. This double septum, as Mayo calls it, is a dominant structural element of the eastern Sierra Nevada; its southernmost end is included in the Mount Pinchot quadrangle. South of the quadrangle the metavolcanic rocks occur again the Alabama Hills east of the Sierra front, and rocks equivalent to both metavolcanic and metasedimentary rocks occur on the west flank of the Inyo Range.

Within the quadrangle, metasedimentary and metavolcanic rocks occupy approximately equal areas. They occur in elongate bodies striking northwest and dipping nearly vertically. Most of the metasedimentary rocks are in the northeast half of the quadrangle and the metavolcanic rocks in the southwest half. Since the tops of the strata appear to be west, and the Mesozoic metavolcanic series is west of the Paleozoic metasedimentary series, the double septum is considered a homoclinal section with top to the west.

METASEDIMENTARY ROCKS

Micaceous schist and hornfels derived from argillaceous sediments predominate among the metasedimentary rocks. Calc-hornfels derived from limy and dolomitic argillite is next in abundance. Marble, quartzite, and metaconglomerate are less common. The age of the metasedimentary rocks is questionable since none contain recognizable fossils; individual masses are isolated from one another by granitic rocks, and metamorphism has obliterated most of the lithologic characteristics. The assumption that these metamorphic rocks are chiefly Paleozoic in age is held by most authors, though it has been suggested that some are of pre-Cambrian age (Webb, 1938, p. 312). Some workers have attempted to correlate metasedimentary rock of the roof pendants in the High Sierra with the Calaveras formation of the western Sierra foothills.

The actual fossil evidence pertinent to the age of the roof pendants of metasedimentary rock in southeastern Sierra Nevada may be listed as follows:

1. Walcott (1895) mentioned a Lower Cambrian sandstone west of Big Pine in the Bishop quadrangle, but the authenticity of this occurrence has been questioned (Mayo, 1941, p. 1008).
2. Mayo (1931) described a few poorly preserved fossils from crystalline limestone in Laurel Canyon, Mount Morrison quadrangle, tentatively referred to the Devonian.
3. Rinehart and others (1959) describe a weakly metamorphosed Paleozoic sedimentary section more than 30,000 feet thick near Mount Morrison approximately 50 miles north of the Mount Pinchot quadrangle. Ordovician graptolites, Pennsylvanian productid brachiopods and crinoid columnals, and Permian (?) bryozoans and brachiopods have been found. The Pennsylvanian fossils are from the formation Mayo tentatively referred to the Devonian.

From the available fossil evidence, it is assumed that the metasedimentary masses in the eastern Sierra include rocks belonging to nearly every period in the Paleozoic era.

In the relatively unmetamorphosed sedimenary rocks of the Inyo Range a few miles to the east, Knopf (1918) has described fossiliferous beds belonging to the Cambrian, Ordovician, Devonian, and Carboniferous periods. In all probability the metasedimentary rocks of the Sierra are some of these formations intensely folded and recrystallized.

PETROLOGY

In the field four broad groups of metasedimentary rocks were recognized and mapped: (1) marble, (2) calc-hornfels, (3) biotite schist, and (4) pelitic hornfels and quartzite. Because the metasedimentary rocks are commonly thinly bedded and diverse rocks are interbedded with one another, the designations on the map refer to the most common and abundant rock. An example of the mixed nature of the metasedimentary septa may be seen in the septum just west of Sawmill Pass, near the center of the quadrangle. The septum is about 200 feet thick at this point and contains all of the four rock types listed above, though marble predominates and the septum was mapped as marble (figure 3).

Marble is generally massive, coarsely crystalline, and white to light gray in color. Some of the marble contains thin layers of garnet- and diopside-rich layers which reflect original shaly partings. The marble, because of its susceptibility to recrystallization and flowage, is commonly intricately folded and contorted.

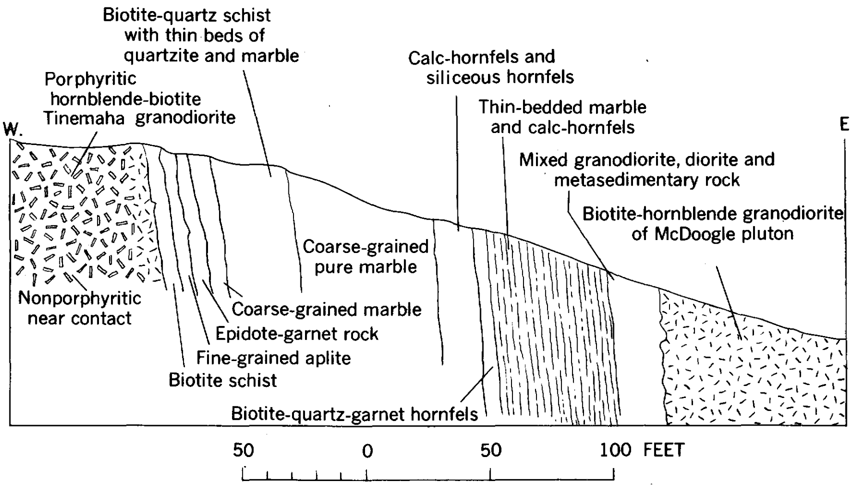

FIGURE 3.—Section across septum of metasedimentary rock 500 feet west of Sawmill Pass.

Calc-hornfels is composed mainly of diopside, quartz, and plagioclase. Orthoclase, hornblende, garnet (grossularite), and wollastonite are common (table 1). The calc-hornfels is generally light colored and massive. Diopside imparts a slight greenish color to the rock. Massive red-brown garnet-pyroxene rock (tactite) is commonly associated with calc-hornfels and marble and is prospected for tungsten.

Biotite schist comprises a large group of rocks with a wide range in mineral composition (table 1). All contain biotite and quartz, though biotite schist ranges in composition from that derived from dolomitic shale and igneous rock to that derived from alumina-rich shale. Most of the biotite schist is red brown to nearly black, contrasting markedly with the light-colored rocks. Of the four groups of metasedimentary rocks, the biotite schist has the best developed foliation.

Pelitic hornfels and quartzite are grouped together because they are commonly interbedded and because both contain abundant quartz. The quartzite was derived from the more sandy rather than shaly layers and contains more than 60 percent quartz. In addition to biotite and quartz, these rocks commonly contain plagioclase, cordierite, andalusite, sillimanite, muscovite, and almandine garnet.

The preservation of original bedding can be observed in the field and in thin section. Marble commonly contains thin layers of grossularite garnet which represent original beds richer in shaly material. In places, biotite schist contains intercalated beds of marble. Very fine bands in calc-hornfels (a few millimeters thick) are the result of original compositional differences in calcareous sediments. Microscope examination shows different mineral assemblages in each band; diopside-plagioclase-quartz, epidote-quartz, and quartz-grossularite-epidote assemblages.

The development of hornfelsic versus schistose texture is a result of both the mineral assemblage and the degree of deformation the rock has undergone. The calcareous and alumina-rich rocks are generally hornfelsic, since they contain relatively nonplanar and nonprismatic minerals such as diopside, calcite, garnet, epidote, cordierite, and andalusite. Rocks with a similar mineral makeup, however, locally develop into either strongly foliated schist or massive hornfels. The more foliated rocks are commonly in the thinner metamorphic septa. Presumably greater localization of penetrative deformation before or during metamorphism increases the schistosity.

Quartz is the most widespread and abundant mineral. It occurs in almost every rock that does not contain free calcite. Most of the metamorphism was at a temperature above that at which calcite and quartz reacted to form wollastonite.

Quartz generally occurs as irregular grains with sutured boundaries. Commonly it is sieve textured and contains abundant inclusions of biotite, apatite, muscovite, rutile, zircon, and magnetite. The volume percent of quartz in pelitic rocks (mica schist and pelitic hornfels), which are the most abundant metasedimentary rocks, averages about 50 percent.

K-feldspar (potassium feldspar) is widely distributed. It generally occurs as small grains which can be easily overlooked. Some orthoclase was recognized only after staining thin sections with sodium cobaltinitrite. Plagioclase is a common mineral in metasedimentary rocks ranging from calchornfels to biotite schist. Nearly all of the plagioclase is oligoclase or andesine. Progressive zoning (calci core) is common, although individual grains are rarely zoned through a range of more than 10 percent anorthite.

Sillimanite and andalusite have been identified in 12 specimens (fig. 4). Typical associated minerals are quartz, plagioclase, cordierite, muscovite, biotite, and garnet (almandite). Sillimanite and andalusite were found together in four specimens. In two of the specimens the sillimanite is of the fibrolite variety; it is clearly a late mineral concentrated along cracks and fissures, and locally developed at the expense of biotite. In the other two specimens, coarse sillimanite and andalusite appear to be in a stable, equilibrium assemblage.

Sillimanite occurs as prisms up to 1 mm long and 0.1 mm wide. Andalusite rarely shows crystal form but occurs either as anhedra or sieve-textured crystals; it commonly displays patchy red pleochroism. No chiastolite was recognized. In this respect the andalusite of the small metasedimentary bodies of this area differs from

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TABLE 1.—*Mineral assemblages of metasedimentary rock samples from the Mount Pinchot quadrangle*

[Minerals identified in thin section. x, common constituent; Tr, present trace amounts]

Sample No.	Quartz	Potassium feldspar	Plagioclase [1]	Al-silicates [2]	Cordierite [3]	Muscovite	Biotite	Hornblende [4]	Garnet	Diopside	Epidote [5]	Calcite	Other [6]	Original rock	Metamorphic rock
237								T				x		Limestone and dolomitic limestone.	Marble.
21a									x			x	H		
92		x								x		x			
90									x	x		x			
109									x	x		x	W		
21b	x		x						x	x		x		Limy and dolomitic shale and sandstone.	Calc-hornfels.
491	x								x	x		Tr.	W		
651	x	x	x							x		Tr.			
515	x								x	x					
534	x		x							x	C		Sp		
33	x		x							x					
176	x		37						x	x			Sp		
122	x								x	x					
671	x	Tr.	Tr.						x	x					
672	x	x	Tr.						x	x					
120	x		x					x		x					
127	x							x		x					
654	x		x				x	x	x	x				Dolomitic shale and intermediate igneous rock.	Biotite schist.
203		x					x			x					
244		x	x				x	T		x					
540	x	x	24				x	x							
531	x		32				x	x							
528	x	Tr.	32				x	x			x				
264a	x	Tr.	77				x	x							
116	x	x	37				x	x			x		Sp		
267	x	Tr.	x				x	x			x				
279	x	x	33				x	x			x		Sp		
531	x	Tr.	x				x	x							
16a	x	x	x				x	x			x		Sp		
580	x	x	x				x				x		Sp		
542	x		x				x				Tr.				
65	x	Tr.	29				x		x						
67	x		35				x		x						
264b	x	x	33				x								
179	x	x	30				x		x						
208	x	x	36				x								
210	x	Tr.	Tr.				x		x						
71	x	x	Tr.				x		x						
670	x		x				x								

676	x	--------	x	--------	--------	--------	x	--------	x	--------	--------	--------	--------	Shale	Pelitic hornfels.
706	x	x	x	--------	--------	--------	x	--------	--------	--------	--------	--------	T		
477	x	--------	25	--------	--------	--------	x	--------	--------	--------	--------	--------	--------		
479	x	--------	32	--------	--------	--------	x	--------	--------	--------	--------	--------	--------		
214b	x	x	34	--------	--------	x	x	--------	--------	--------	--------	--------	--------		
125	x	x	32	--------	--------	x	x	--------	--------	--------	--------	--------	--------		
236	x	Tr.	27	--------	xs	x	x	--------	x	--------	--------	--------	T		
16b	x	x	x	--------	s	x	x	--------	--------	--------	--------	--------	T		
115	x	x	40	--------	s	x	x	--------	x	--------	--------	--------	T		
213	x	--------	35	--------	s	x	x	--------	--------	--------	--------	--------	--------		
42	x	x	--------	--------	s	x	x	--------	x	--------	--------	--------	--------		
262	x	--------	25	--------	s	x	x	--------	--------	--------	--------	--------	--------		
214a	x	--------	36	--------	s	x	x	--------	--------	--------	--------	--------	--------		
263	x	--------	28	--------	xs	x	x	--------	--------	--------	--------	--------	--------		
281	x	Tr.	Tr. (15)	SA	s	x	x	--------	--------	--------	--------	--------	--------	Alumina-rich shale.	
235	x	--------	35	A	s	x	x	--------	x	--------	--------	--------	--------		
148	x	--------	45	SA	xs	x	x	--------	--------	--------	--------	--------	--------		
119	x	--------	--------	A	xs	x	x	--------	x	--------	--------	--------	T		
69	x	--------	15	A	xs	x	x	--------	--------	--------	--------	--------	--------		
278	x	--------	20	S	x	--------	x	--------	--------	--------	--------	--------	--------		
257	x	Tr.	--------	A	--------	x	x	--------	x	--------	--------	--------	T		
136	x	--------	x	FA	--------	--------	x	--------	--------	--------	--------	--------	--------		
274	x	--------	Tr.	FA	--------	--------	x	--------	x	--------	--------	--------	--------		
89	x	Tr.	x	S	--------	--------	x	--------	--------	--------	--------	--------	--------		
489	x	Tr.	Tr.	A	--------	x	x	--------	--------	--------	--------	--------	--------		
677	x	--------	Tr.	F	--------	x	x	--------	x	--------	--------	--------	--------		

[1] Numbers represent percent anorthite in plagioclase. [2] F, fibrolite sillimanite; S, prismatic sillimanite; A, andalusite. [3] s, shimmer aggregates pseudomorphous after cordierite. [4] Tc tremolite. [5] C, clinozoisite, [6] H, chondrodite; W, wollatonite; Sp, sphene; T, tourmaline.

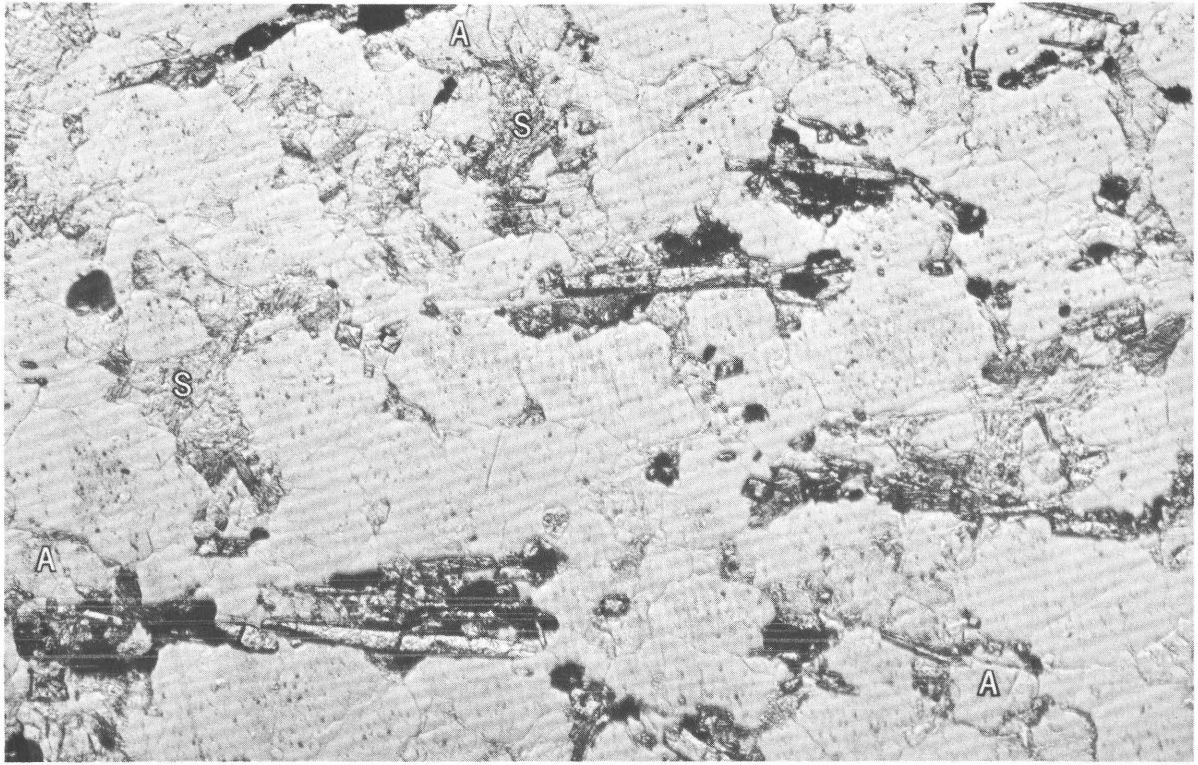

FIGURE 4.—Photomicrograph showing needles of sillimanite in quartzose pelitic hornfels derived from siliceous shale. The rock also contains andalusite (*A*), shimmer aggregates after cordierite (*S*), and plagioclase, red-brown biotite, and magnetite. ×30. Plain light.

andalusite of larger metasedimentary pendants, as for example the Mount Morrison pendant to the north (Rinehart and Ross, 1963).

The distribution of fibrolite sillimanite, prismatic sillimanite, and andalusite appears to be related to the distance from the plutonic contact. Fibrolite occurs only at and beyond about 500 feet from the contact. Sillimanite occurs adjacent to the contact and out to about 500 feet. Andalusite may occur close to the contact or as far distant as samples were collected, about 1,000 feet. It therefore appears that prismatic sillimanite is a mineral favored by the higher temperatures adjacent to the contact, whereas andalusite is stable down to a somewhat lower temperature which prevailed farther from the contact. The fibrolite sillimanite appears to be a lower temperature, retrogressive mineral which forms from biotite along zones of shear.

Biotite is a mineral of very wide occurrence and is present in all rocks containing muscovite. On the other hand, biotite is sparse in rocks derived from limy sediments, rare in rocks containing diopside, and unknown in association with wollastonite or calcite (table 1). The biotite has great variation in color. Its color, parallel to z, varies from dark reddish brown (color 10R ¾, Goddard, 1948) through light brown (5YR 5/6), dusky yellowish brown (10YR 2/2), to olive olive gray (5Y 3/2). The color does not appear to be related to the distance from the contact, or to the chemistry of the sediment from which the rock was derived.

Muscovite is common in rocks that contain alumina-silicate minerals or biotite; more than half of the rocks containing andalusite and sillimanite also contain muscovite. Muscovite never occurs with calcite, diopside, or hornblende.

Cordierite has been identified in 6 thin sections, and an additional 8 sections contain shimmer aggregates, the alteration product of cordierite. The mineral occurs in rocks poor in lime; it has not been identified in a rock containing calcite, wollastonite, diopside, hornblende, or epidote. Typical associated minerals are quartz, oligoclase-andesine, andalusite, sillimanite, muscovite, and biotite.

Generally cordierite occurs in rounded grains without twinning or crystal form. It is commonly the coarsest mineral in the hornfels and prefers the coarser areas in the rock. In one specimen spindle-shaped cordierite porphyroblasts attain 30 mm in length; however, in general, the mineral average 0.2 mm in size.

Cordierite alters very readily and is recognized primarily by its alteration products. Most commonly the mineral alters on its borders, and in a reticular pattern within, to a mixture of fine-grained muscovite (possibly some talc) and fine green chlorite. In a few specimens a yellow low-birefringent mineral (probably a chlorite) is an alteration product. Sillimanite was formed at the expense of cordierite porphyroblasts in one specimen. Pleochroic haloes surround small high-relief mineral grains in the cordierite.

Green hornblende is found in rocks derived from dolomitic shales and from igneous rocks, probably andesites. In a few specimens the igneous texture has not been entirely destroyed; the presence of sphene and epidote in some of these hornblende rocks strengthens the supposition that they were derived from igneous rocks. Hornblende is typically associated with quartz, andesine, biotite, and epidote. Tremolite is found in a few lime-rich rocks.

In one specimen, taken from the contact between quartz monzonite and lime-silicate hornfels, hornblende was found surrounding diopside and as discrete crystals, apparently as an alteration product of diopside. The amount of hornblende diminishes with distance from the plutonic contact. A hydrous environment near the contact apparently favored formation of hornblende.

Garnet is common throughout the sequence limestone–shale. The garnet in the limy rocks is presumably grossularite, but that in the lime-poor rocks, almandite.

Green tourmaline is found in many specimens as sparse, ragged crystals. The mineral is entirely confined to rocks derived from shales (table 1). The distribution of the tourmaline supports the findings of Goldschmidt (1954, p. 284) that boron is concentrated in marine argillaceous sediments. The tourmaline is presumably not of metasomatic origin (as suggested by its overall scarcity and confinement to one sediment type), and presumably it is a metamorphic mineral, that has utilized the boron in the shales for its growth.

METAMORPHISM

The original composition of the metasedimentary rocks ranged from limestone and dolomitic limestone to shale and siltstone. Sandstone and conglomerate were uncommon. A small amount of intermediate volcanic rock was tentatively identified in a few places. No chert, mafic volcanic rock, or graywacke has been identified in association with the metasedimentary rocks.

The metasedimentary rocks have been folded, intimately sheared, and recrystallized. Probably most of the metamorphism has been accomplished by the physical and thermal action of the incoming granitic masses. A large number of granitic intrusions of slightly different age are present in the area; hence the metamorphic rocks have been subjected to repeated metamorphism, though the closest intrusions no doubt exerted the greatest effect. If a preintrusion period of metamorphism occurred, all traces of it have been obliterated by the latest metamorphism.

The bedding and schistosity of the metasedimentary rocks are generally parallel and nearly vertical. Some beds appears to be isoclinally folded, but the folding is believed to be on a fairly small scale, since gross lithologic units can be traced almost the entire length of the metamorphic septa.

The mineral assemblages of the metasedimentary rocks are dominantly in the contact-metamorphic hornblende hornfels facies of Turner (Fyfe and others, 1958). However, the presence of almandine in some of the pelitic rocks suggest that they are transitional to the almandine amphibolite facies. Also the common presence of epidote in mafic assemblages indicates transition to the albite-epidote hornfels facies.

In general K-feldspar is absent or present only in trace amounts with andalusite or cordierite (table 1). But the occurrence of it with these minerals in places suggests that some of the rocks are in the upper part of the hornblende hornfels facies (Rose, 1958).

No metamorphic zoning was recognized within the septa of metasedimentary rocks, except for the localization of prismatic stillimanite within 500 feet of the plutonic contacts. However, the geometry of the septa is not suited to the study of progressive metamorphism because of their long and narrow shape and because the lithologic units are nearly always parallel to the plutonic contact rather than normal to it.

METAVOLCANIC ROCKS

The pregranitic rocks of the Mount Pinchot quadrangle consist in large part of metamorphosed volcanic rocks. The largest single body of metamorphic rock in the quadrangle is a pendant of metavolcanic rock more than 2 miles wide and 6 miles long in the upper drainage basin of Oak Creek. Smaller bodies of metavolcanic rocks occur on the range front north and south of the mouth of Independence Creek, and near the headwaters of Sawmill Creek. Others west of the Sierra crest are exposed on Mount Bago, in the drainage basin of the South Fork of Woods Creek, and on Cirque Crest in the northwest corner of the quadrangle. The Cirque Crest mass is adjacent to a large septum of metavolcanic rock which extends northwestward from the quadrangle for a distance of 40 miles.

No diagnostic fossils have been found in the metavolcanic section in the Mount Pinchot quadrangle. However, comparison of the rocks with fossil-bearing beds on strike to the north and south, and with similar beds to the southwest, suggests that this sequence may be Middle to Late Triassic and (or) Jurassic in age.

The rocks of volcanic origin in the Inyo Mountains east of the Sierra Nevada have been assigned to the Middle or Late Triassic by Kirk (*in* Knopf, 1918, p. 59):

The evidence on which this determination is based was found on the ridge on the south side of Union Wash, in the Inyo Range, where the basal part of the volcanic series, consisting of andesitic breccias, interleaves with the underlying limestones of Middle Triassic age. Moreover, angular fragments derived from these limestones are common inclusions in the breccia.

Because of the lithologic similarity of the volcanic rocks of the Inyo Range and the metavolcanic rocks of the Sierra Nevada, Knopf has assumed that the latter are also of Triassic age.

Merriam (written communication, 1951) suggests the possibility that an unconformity may exist between the marine Triassic and the basal conglomerate of the volcanic section as exposed on the Cerro Gordo road in the Inyo Mountains. If such is the case, the volcanic rocks may be younger than Late Triassic, perhaps Jurassic.

Extending northwestward from the northwest corner of the Mount Pinchot quadrangle through the Sierra Nevada batholith is a 40-mile-long pendant of metavolcanic rock (Mayo, 1941, pl. 1). Twenty-five miles north of this pendant Lower Jurassic fossils have been found in the Ritter Range metavolcanic pendant, approximately 10,000 feet stratigraphically above the top of a sequence of marine metasedimentary rocks of Paleozoic age (Rinehart and others, 1959).

In a roof pendant near the Mineral King region, 15 miles southwest of Mount Whitney, metavolcanic rock is interbedded with fossiliferous sediments. From the poorly preserved fossils, H. W. Turner (1894, p. 451) reports that the series is probably of Triassic age. A more complete study of these fossils from the Mineral King pendant by C. Durrell and S. W. Muller (Durrell, 1940, p. 17) establishes quite definitely "* * * that the Mineral King beds, or at least the known fossil-bearing members, are of Upper Triassic age."

LITHOLOGY AND METAMORPHISM

In general the metavolcanic rocks were mapped on the basis of their composition, determined in the field chiefly by color and percent of dark minerals. Three general groups have been distinguished in the field: (1) metabasalt, (2) metadacite and meta-andesite, and (3) metarhyolite. Petrographic examination and calculation of silica percent by specific gravity and refractive index of fused rock show that the rocks are slightly richer in silica than was supposed in the field (fig. 5).

Because of the poor bedding, rapid facies changes, and mixed nature of much of the metavolcanic sequence, the portrayal of rock types on the map is generalized. Mixed sequences are designated on the map by the name of the dominant or average compositional type.

The metavolcanic rocks, except some of the massive metabasalts, possess a foliation that is commonly parallel to the bedding, and to the regional elongation of the granitic bodies. Pyroclastic beds contain

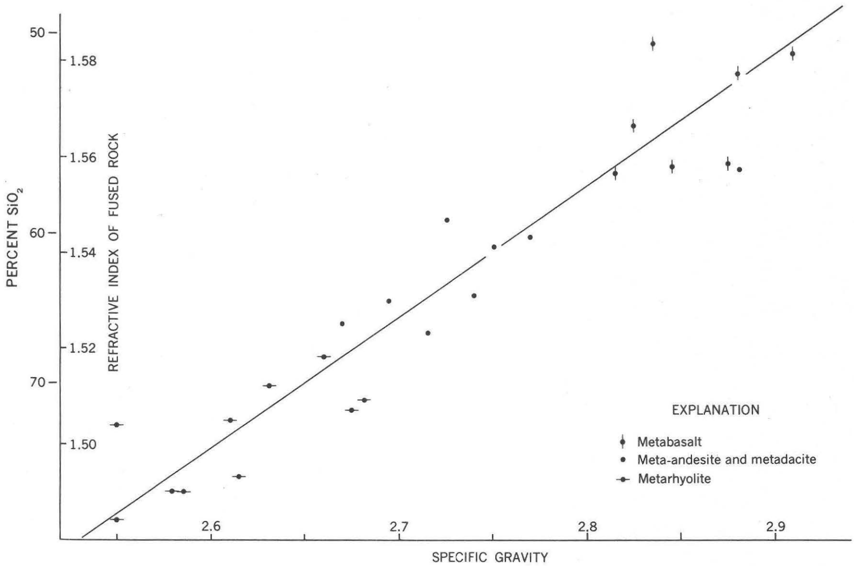

FIGURE 5.—Graph showing relation between hand specimen specific gravity and refractive index of fused rock in 25 specimens of metavolcanic rocks. The approximate percent SiO_2 corresponding to refractive index of fused rock is shown according to Mathews (1951). The metavolcanic rocks are divided into three groups as mapped in the field: (1) metabasalt, (2) meta-andesite and metadacite, and (3) metarhyolite.

rock fragments flattened and stretched parallel to the regional foliation. In some places these same rock fragments are stretched into spindle shapes in the plane of foliation. These spindles generally plunge steeply parallel to the lineation of the neighboring granitic intrusions. However, stretched rock fragments in a meta-andesite inclusion of the Dragon pluton are parallel to the gently dipping lineation of the Dragon pluton. These relations suggest that the lineations of the metamorphic rocks are caused by the same stresses that formed the lineation in the granitic rocks.

Mineral assemblages (fig. 7) indicate that the metavolcanic rocks, like the metasedimentary rocks, are dominantly of the contact-metamorphic hornblende hornfels facies (Fyfe and others, 1958). The presence of green hornblende and the absence of pyroxene is indicative of this facies.

METABASALT

Rocks derived from basalts constitute the smallest proportion of the metavolcanic sequence. However, they are generally the best preserved because of their original massive character. Nearly all of the basalt occurred as lava flows, whereas the more siliceous volcanic rocks were chiefly pyroclastics. The incompetent and porous pyroclastics underwent greater modification during diastrophism.

Some primary volcanic structures have been preserved in the metabasalt. Most notable is the preservation of original igneous subophitic textures. Flow alinement of the plagioclase crystals is common. Some flows show amygdaloidal tops and bottoms. Inclusions of an older granitic rock were seen in one flow. Textures and mineral content suggest that olivine was uncommon in the original basalts.

A metabasalt flow or group of flows east of Sardine Lake in the large metavolcanic pendant of Oak Creek can be traced nearly 2 miles although it is only a few hundred feet thick. The thinness of this flow, plus the absence of pillow structure, suggests that it was not of submarine origin.

The metabasalts are grey to black massive rocks, generally without cleavage or foliation; in part, they are interstratified in strongly foliated metarhyolite tuffs. Commonly they break with a subconchoidal fracture and form sharp-edged fragments. Microscopically the majority show excellent relict flow alinement of plagioclase laths, and subophitic textures (fig. 6). Most of the specimens have developed no schistosity.

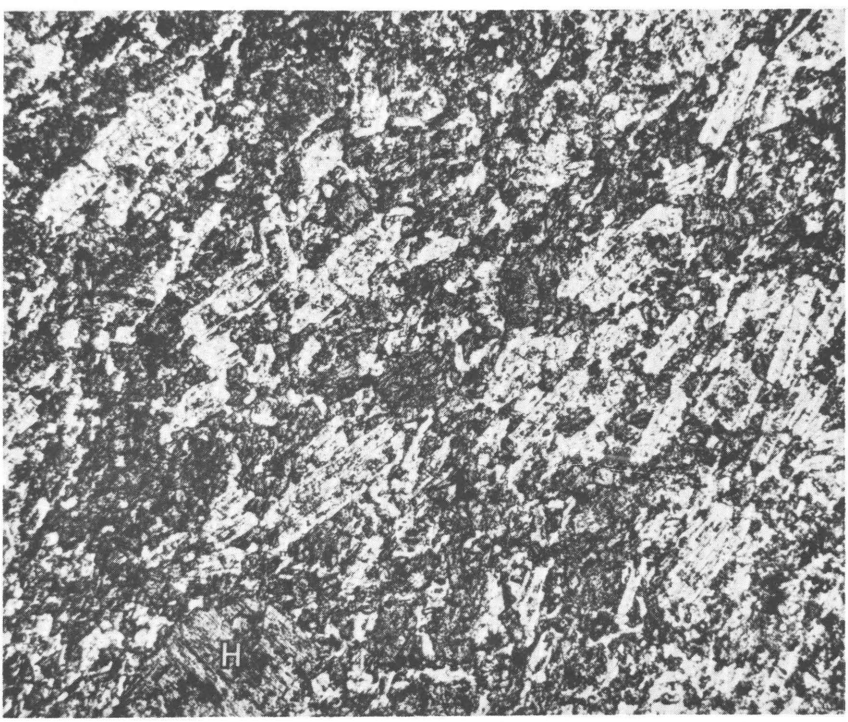

FIGURE 6.—Photomicrograph of metabasalt showing relict igneous texture and flow structure. Green hornblende forms a pseudomorph after a pyroxene phenocryst at lower left (H), and occurs throughout matrix. Calcic cores of zoned plagioclases are slightly saussuritized. ×30. Plain light.

Plagioclase is abundant as slightly clouded, lath-shaped crystals. It shows oscillatory zoning with a core of labradorite and a rim of andesine. The more calcic core is generally slightly saussuritized. The preservation of oscillatory zoning in a rock at this grade of metamorphism is notable. Presumably if the rock had gone through the greenschist facies, the plagioclase would have broken down into albite and clinozoisite, and upon higher temperatures, when a more calcic plagioclase crystallized, it would have been homogenized and shown no zoning.

Poldervaart (1953, p. 262) states ". . . . metabasaltic rocks throughout the world have been tardy in responding to plutonic conditions . . ." Perhaps the basalts never adjusted to a lower grade of metamorphism, but because of the quickly elevated temperatures possible adjacent to a plutonic intrusion, they did not have time to be broken down before the temperature was elevated enough to permit zoned labradorite to remain as a metastable mineral. Read (1949, p. 105) states that high-grade metamorphic rocks do not necessarily ever have to pass through a low-grade stage. Such a retardation in adjusting to the physical environment might have been favored by the relatively dry conditions under which much of the metamorphism took place.

Green hornblende is the chief mafic mineral in most of the metabasalts (fig. 7). It commonly pseudomorphoses the original pyroxene phenocrysts, and it also occurs as small grains in the groundmass. In some specimens the hornblende is ragged and partially replaced on the edges by biotite. Exsolved iron in the form of magnetite is abundant within and on the edges of the hornblende. Biotite is common in some of the metabasalts as a metamorphic mineral.

The mineral assemblages of the metabasalts indicate that metamorphism has been of the hornblende hornfels facies. The scarcity of any imposed schistosity and preservation of original textures, flow structures, and amygdules at the tops of some flows demonstrate that metamorphism was mainly thermal, with little shearing. Probably the massive, competent nature of the metabasalts caused them to be relatively immune to penetrative deformation.

Mineral makeup of the metabasalts, plus specific gravity determinations and refractive index measurements of the fused rocks, suggest that metamorphism was nearly isochemical. The presence of biotite suggests, however, that K_2O may have been added during metamorphism.

METADACITE AND META-ANDESITE

Metamorphosed dacite, andesite, and related rocks make up a large proportion of the Triassic and Jurassic metavolcanic sequence. Tuff breccias resembling the Tertiary andesites that cap the central Sierra

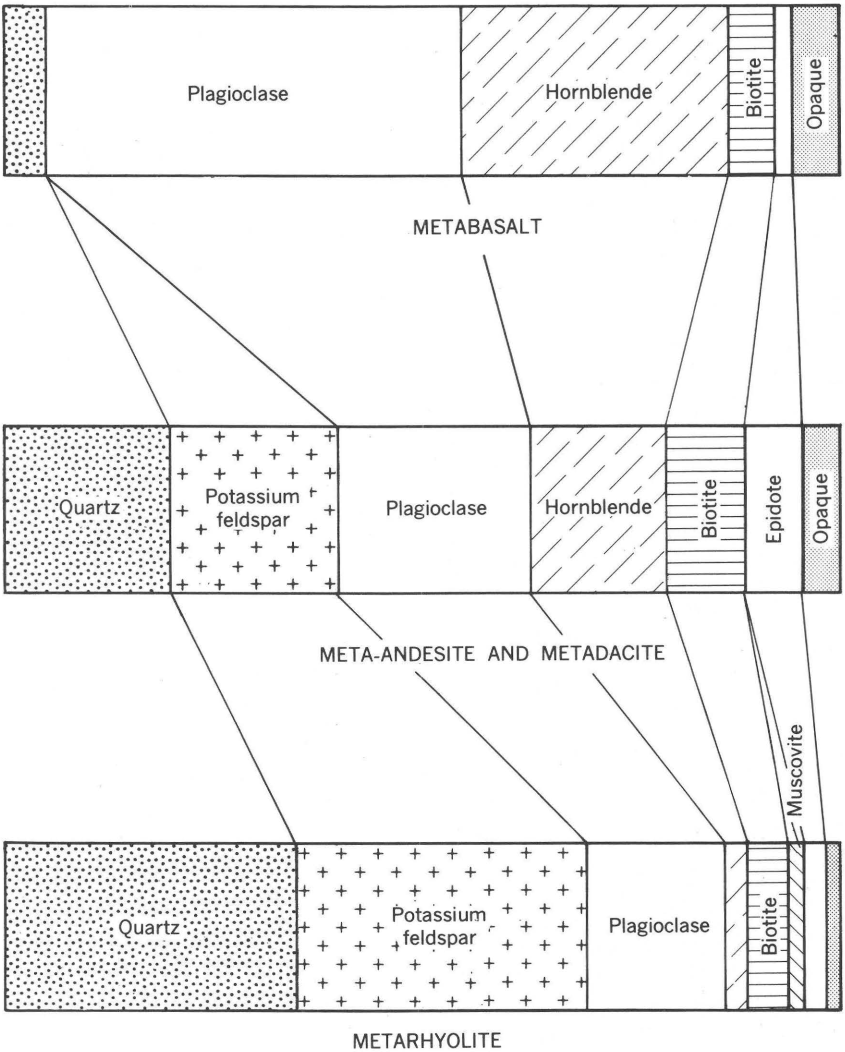

FIGURE 7.—Bar diagram showing the average mineral content of the three major groups of metavolcanic rocks.

Nevada are very common. They probably originated as agglomerate, volcanic avalanche debris, or perhaps mud flows. Some may have been deposited by submarine turbidity currents, as suggested by graded bedding in some intercalated tuffs. Lava flows (apparently subaerial), water-laid tuff, and volcanic conglomerate are also present.

The meta-andesites are well exposed in the easily accessible outlier east of the mouth of Independence Creek. The rock here has a distinct breccialike appearance with angular boulders exceeding 3 feet in diameter. The fragments are generally stretched and flattened and define a fairly consistent lineation or foliation. It is not uncommon for bands of finer fragments and shaly sediments to be interlayered with the dominant fragmental material. Where the bedding is thus defined, it is parallel with the foliation in most places, but in a few is inclined to the foliation. The meta-andesites are streaked and mottled with green; epidote is present in the groundmass as well as in veins and joints.

Microscopically most of the meta-andesites are fine grained (average grain size 0.05mm) and crudely foliated, and the crystals are sutured and ragged. Irregular spindle-shaped clots or masses of ferromagnesian minerals such as hornblende, biotite, epidote, and magnetite mark the site of former mafic rock fragments and ferromagnesian phenocrysts. Andesine and oligoclase are present as cloudy, commonly partly saussuritized crystals; the plagioclase rarely shows zoning. K-feldspar is present as small grains in the fine-grained matrix and as sparse relict phenocrysts.

Many of the rocks contain quartz grains, locally stretched out parallel with the foliation. Some quartz is probably detrital, originating in sediments associated with the pyroclastic rocks; quartz also is a devitrification product of the glassy matrix of the intermediate volcanic material. The chief mafic minerals, green hornblende, biotite, and epidote, generally occur together as scattered aggregates. Magnetite is common in some rocks.

METARHYOLITE

Metamorphosed rhyolites are light colored and occur in both phyllitic and massive form. Some have been so strongly recrystallized that they can be mistaken for orthogneiss in the field. The majority of these rocks were derived from structurally weak volcanic tuffs and have undergone severe deformation. In most of the metarhyolitic tuffs the bedding is poorly defined or absent. They contain abundant dark- and light-colored rock fragments as long as 4 inches. The light-colored fragments appear to be flattened and stretched pumice lapilli.

Under the microscope the fine-grained matrix (approx. 0.03 mm) presents a strong directional structure, which curves around rock fragments and quartz, feldspar, and mafic mineral porphyroclasts (fig. 8).

Quartz occurs abundantly through the matrix and as crystal fragments or original phenocrysts as large as 8 mm in length. The large quartz grains are stretched out parallel to the foliation and show undulatory extinction. As a rule the quartz grains are elongated more than plagioclase or orthoclase grains of the same size, which demonstrates that quartz responds to stress more quickly than feldspar. Many quartz crystals were observed which have corroded and

FIGURE 8.—Photomicrograph of metamorphosed rhyolite tuff with relicts of mafic minerals, probably biotite. Corroded and broken quartz grain (on right) probably inherited from original tuff. Matrix is fine quartz and feldspar. ×35. Plain light.

embayed edges; they are assumed to be relicts of original corroded quartz grains in rhyolite flows or tuffs.

The most abundant mineral of most of the metamorphosed rhyolite tuffs is orthoclase or microcline (fig. 7). K-feldspar is present both in the matrix of the rock and as large crystals. Many of these potassium feldspar crystals or relict phenocrysts are perthitic with intricate intergrowths and rims of K-feldspar and albite; the original volcanic sanidine evidently contained considerable soda, which has exsolved. Plagioclase, generally albite, is sparingly present in large crystals and common in the fine-grained matrix. The plagioclase is commonly unzoned and many grains are altered.

The mafic minerals are chiefly biotite and green hornblende, both of which occur in small amounts. Hornblende commonly is in radial or strung-out aggregates. Biotite crystals have exsolved iron in some places and remain as phlogopite with networks of magnetite pseudomorphing the original biotite cleavage.

Piedmontite is common along certain horizons in the metatuff; some of the more prominent purplish-colored, piedmontite-bearing zones have been mapped. The piedmontite is finely disseminated through the groundmass, in porous minerals such as altered biotite, and veins and pods with quartz. The piedmontite probably formed by contact metamorphism from a premotamorphic concentration of epithermal manganiferous minerals in the rhyolite tuffs.

Other minerals present in the metarhyolites are epidote, muscovite, chlorite (in small amounts in the groundmass), calcite, and apatite.

STRATIGRAPHY AND STRUCTURE

OAK CREEK PENDANT

The main mass of metavolcanic rocks in the Mount Pinchot quadrangle is a roof pendant elongate northwestward and measuring 6 miles in length by 2 miles in width. Most of the pendant, here named the Oak Creek pendant, lies east of the Sierra crest in the drainage area of the North and South Forks of Oak Creek and Sardine Creek. It extends southeastward across Kearsarge Peak into Independence Creek and northwestward across the main Sierra crest into the headwaters area of Baxter Creek.

To the northwest the pendant is cut off by the Baxter pluton, and to the southeast is is wedged out by the converging Tinemaha and Dragon intrusive masses. The concordant Sardine pluton has intruded parallel to the bedding into the center of the Oak Creek pendant, shouldering aside the metavolcanic rocks.

Field evidence suggests that the Oak Creek pendant represents an unrepeated section of volcanic rocks more than 7,000 feet thick. Graded beds and vesicular tops in basalt flows observed in five localities indicate that the top of the near-vertical beds are consistently to the southwest. The most decisive criteria were all found on the high ridge south of Summit Meadow and on the ridge north of Sardine Canyon.

The metavolcanic rocks of the Oak Creek pendant have been recrystallized and metamorphosed to the hornblende hornfels facies and stretched, sheared, and faulted, predominantly along strike faults. The degree to which these changes have affected the original thickness of the section is not known, but the present thickness is almost certainly less than the original thickness.

The Oak Creek metavolcanic section contains a great variety of volcanic rocks (fig. 9). The overall average rock type is rather silicic, perhaps a dacite or rhyolite. Forming the east edge of the main mass of the pendant and stratigraphically the lowest unit is a uniform siliceous metarhyolite layer, resembling a rhyolite flow though possibly a metamorphosed rhyolitic welded tuff. Above this basal metarhyolite is a series of metamorphosed andesitic lava flows from 100 to 300 feet thick. This rock is typically a porphyritic meta-andesite.

Above the meta-andesite flows is approximately 3,000 feet of poorly bedded metarhyolite tuffs. In general this unit is monotonous because

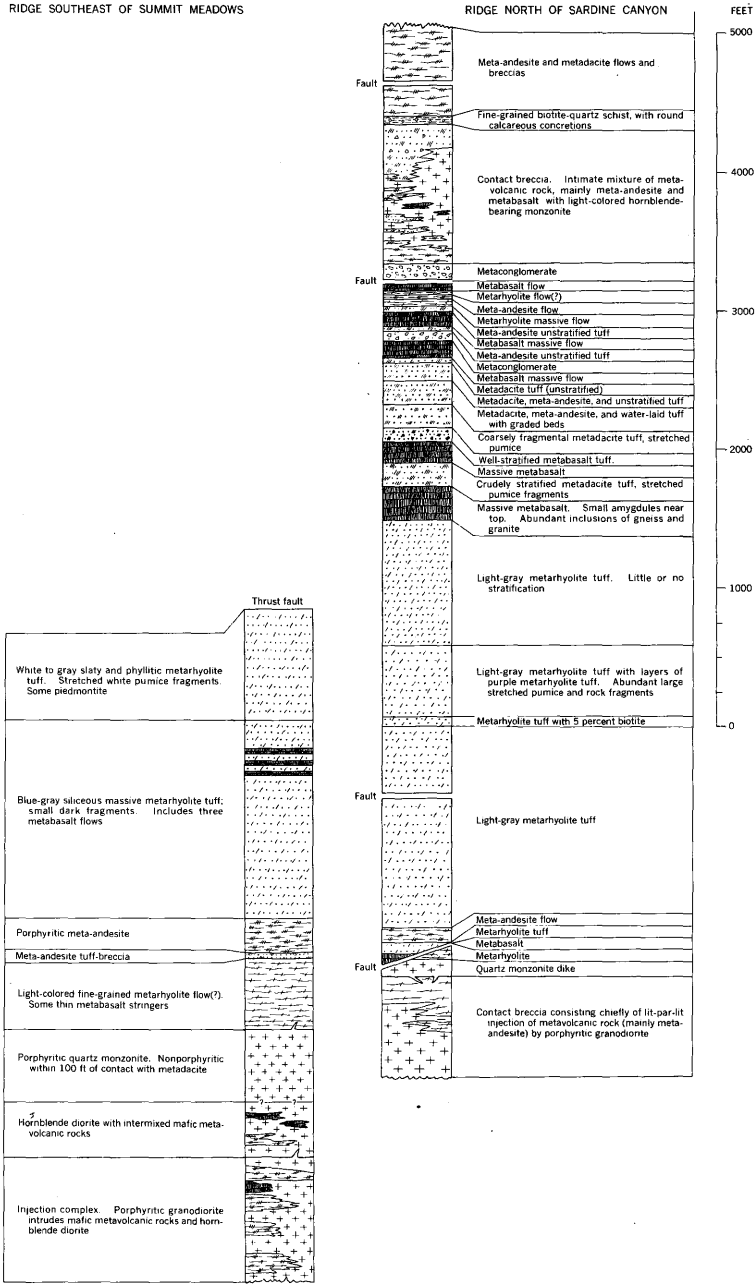

FIGURE 9.—Generalized section of metavolcanic rocks exposed in Oak Creek pendant.

of its uniformity. It varies in color from nearly white to light gray, blue gray, and dark gray. Commonly the average size of the original pyroclastic fragments is from 2 to 5 mm in greatest dimension.

Great thicknesses of the metatuff show little or no stratification. Some layers are characterized by dark rock fragments or pumice fragments several centimeters long. Commonly the fragments are stretched, so that their greatest dimension is three to six times their smallest dimension. Purplish horizons are present near the upper part of the thick metarhyolite section. Piedmontite, the manganese epidote, has been found to be the coloring agent in at least some of these purplish tuffs. Two of the more persistent of these piedmontite layers are shown on the map.

The lithologic break between the lower, relatively uniform metarhyolite tuff unit and an upper, mixed metavolcanic unit is important from the structural standpoint. The Sardine pluton was emplaced along this contact, making room for itself primarily by wedging apart the metavolcanic rocks along this one bedding plane.

The upper mixed unit of metarhyolite tuffs, meta-andesite and metadacite tuffs and flows, and metabasalt flows is probably greater than 3,000 feet thick. In this unit criteria were found which indicate that the top of the section is to the southwest. Some of the metabasalt flows show a zone of amygdules in the upper 10 or 15 feet. An amygdule zone is common at the base of some flows also, but it is generally much thinner. Some well-stratified water-laid andesitic and dacitic tuffs display graded beds from 1 to 5 cm thick. The grain size in a single bed may vary from pea size near the base to fine sand at the top.

SUMMIT MEADOW KLIPPE

Overlying the near-vertical flows and tuffs of the Oak Creek pendant south of Summit Meadow, near the headwaters of the North Fork of Oak Creek, is a mass of metavolcanic rocks dipping from 20° to 60° N. and consequently in strong angular discordance with the underlying rocks. These metavolanic rocks, which cover an area of one-half square mile, presumably represent the upper plate or klippe of a thrust fault which has carried them some distance to their present anomalous position.

The bedding of the metavolcanic rocks of the klippe is nearly parallel with the thrust plane; both bedding and thrust surface are nearly parallel with the canyon wall on which they are exposed. Consequently the klippe comprises a relatively thin metavolcanic section, about 600 feet thick.

In general the lithology of the klippe is most like the upper mixed unit of the main part of the Oak Creek pendant, though the rocks are metamorphosed to a lesser degree; the grain size is finer, and original structures such as bedding are better preserved. In order, from the bottom to the top of the klippe are exposed metarhyolite tuff, well stratified (water-laid) meta-andesite and metadacite

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tuffs, and dark meta-andesite and metabasalt flows. Interbedded with the water-laid tuffs is one silty bed, best exposed at 12,300 feet on the ridge crest three-fourths mile south of Summit Meadow, containing pisolitic structures from 0.1 to 0.5 inch in diameter. These spheroids have an outer concentric structure around an inner core of finely clastic material not unlike the matrix of the rock. All of the pisolites are tectonically stretched into spindle shapes such that the longer diameter is approximately three times the shorter; the long axis of these spindles lies in the plane of the near-vertical foliation and plunges steeply. The foliation cuts across the bedding and is parallel with the regional foliation of the rocks below the thrust fault. Many fragments of pisolites are also present in the bed. Despite the fact that these rocks have been sheared, stretched, and recrystallized, the fragments appear to be primary—that is, deposited as fragments in the original sediment.[2]

Except for the massive metabasalt, the rocks of the klippe possess a planar structure defined by a cleavage or phyllitic foliation. This foliation is nearly vertical and strikes parallel to the foliation of the rocks in the lower plate of the thrust. The parallelism of metamorphic structure both above and below the thrust shows that the metamorphism responsible for the foliation occurred after thrusting.

Mafic dikes, which are abundant in the lower plate of the thrust, were not found in the upper plate. Thrusting apparently took place after dike injection and before the last metamorphism, which probably occurred during the emplacement of the later granitic rocks. Hence the thrusting is most likely directly related to the period of plutonic activity. The place of origin of the klippe, and hence direction of thrusting, is not known. However, the lesser degree of metamorphism of the metavolcanic rocks of the klippe points to an eastern source.

OTHER MASSES OF METAVOLCANIC ROCK

Two masses of metavolcanic rock are present on each side of the Independence Creek road east of Grays Meadow, in the southeast corner of the map. Volcanic breccias and lava flows of andesitic composition appear to be the dominant rock types in these exposures. Metamorphosed rhyolites and some calcareous metasedimentary rocks have also been recognized.

West of the summit of Diamond Peak on the main Sierra Crest south of Baxter Creek is a septum of metavolcanic rock nearly 2 miles long. Dark metamorphosed lavas, probably andesites, make up the major part of the mass, an aggregate in thickness of about 600 feet.

[2] Additional study of these structures after this report was prepared shows that they are accretionary lapilli formed as the result of muddy rains accompanying pyroclastic eruptions (Moore and Peck, 1962).

A thin layer of calc-silicate hornfels is present on the east edge of the mass. The metavolcanic rocks exposed on Diamond Peak probably belong to the upper part of the Oak Creek pendant and have been wedged off by the intrusion of the early White Fork granitic body.

On the summit and west flank of Mount Bago in the southwest corner of the map are two septa of metavolcanic rock, chiefly metarhyolite tuff. Both extend out of the quadrangle to the south. The eastern septum, occupying the summit of Mount Bago, is 2 miles long; it separates normal alaskitic quartz monzonite of the Bullfrog pluton on the east from a fine-grained quartz monzonite on the west. The western septum, 4 miles long, separates the fine-grained quartz monzonite on the east from the granodiorite of the Paradise pluton on the west. Because of this septum, the relative ages of the Bullfrog and Paradise plutons are unknown. The eastern septum is interesting because, where exposed on the 3,000-foot south wall of Mount Bago, the metavolcanic rock grades downward into a migmatite composed of large masses of dark, partly mobilized metavolcanic rock intimately mixed with granitic material. Toward the bottom of the canyon, the migmatite grades into swirled and strung-out lenses of small (about 6 inches long) mafic inclusions, joined together and elongated parallel to the septum.

In addition to these larger masses of metavolcanic rocks, innumerable smaller bodies are included in many of the intrusive rocks. Only the larger or more conspicuous of these have been mapped. These smaller bodies are most common in the intrusive rocks flanking the Oak Creek pendant and were no doubt derived from that pendant by intrusion and wedging-off parallel to the foliation.

A small mass of metarhyolite west of the Cartridge pluton in the northwest corner of the map is associated with the large metavolcanic pendant of the Mount Goddard–Black Divide country northwest of the Mount Pinchot quadrangle. Some granitic masses have split off and incorporated vast quantities of their metavolcanic wall rocks, forming a plutonic breccia of lit-par-lit type. The Tinemaha, Dragon, and Independence intrusive masses are especially notable in this connection. Metavolcanic rock exists in some of the complexes of mafic plutonic rock, and perhaps much more is present and has lost its identity in these complexes as a result of metamorphism. Metabasaltic rocks are intimately mixed with dioritic rocks near the junction of the main and South Fork of Woods Creek.

MAFIC PLUTONIC ROCKS

Dark-colored mafic igneous and hybrid rocks have been combined and mapped as mafic plutonic rocks on the geologic map. This unit is extremely heterogeneous; it includes many different rock types of decidedly different origin. However, the rocks have one feature in common: they originated early in the plutonic cycle and have been subjected to repeated metamorphism by the younger intrusive bodies. In table 2 a few modal analyses of the mafic plutonic rocks are recorded. They include mafic granodiorite, quartz diorite, and hornblende gabbro. In addition, hornblendite, biotite rock, and peridotite are present.

TABLE 2.—*Volume percent of minerals of samples from mafic plutonic masses in Mount Pinchot quadrangle*

Sample No.[1]	Quartz	Potassium feldspar	Plagioclase	Biotite	Hornblende	Pyroxene	Opaque accessories	Nonopaque accessories	Color index
1	3.7	0.2	60.6	8.8	23.1	0.3	3.0	0.2	35.4
2	6.1	.4	55.3	13.6	23.1	----	1.2	.4	38.3
3	12.5	21.6	47.7	9.9	7.3	----	1.1	----	18.7
4	5.9	.5	61.8	10.9	19.6	1.1	.1	----	31.7
5	----	----	48.1	11.0	36.6	----	3.6	.8	52.0
6	9.4	----	50.7	22.4	14.8	----	1.1	1.5	39.8
7	7.6	11.4	50.6	12.9	3.8	11.3	1.8	.6	30.4
8	6.8	5.3	57.7	5.1	24.0	----	.7	.4	30.2
9	7.7	6.4	62.7	14.7	.5	5.6	2.4	----	23.2
10	9.9	14.4	48.9	12.1	5.7	6.1	2.3	.8	26.7
11	.9	1.1	91.9	----	4.7	.6	.8	----	6.1

[1] Samples 1–5 are from the dioritic mass south of Rae Lakes; No. 6, from the dioritic mass of Goodale Creek; Nos. 7–10, from the dioritic mass of Thibaut Creek; and No. 11, from the mass of anorthosite of Thibaut Creek.

The two largest masses of mafic plutonic rock are east of the Sierra crest, on the north and south sides of the Spook pluton. These two masses are lithologically similar and show other features in common. Both are cut by numerous mafic dikes, and both are closely associated with metasedimentary rock, predominantly marble. Furthermore, both are intruded on their east sides by the distinctive porphyritic granodiorite of the Mule Lake pluton. These facts suggest strongly that the mafic plutonic rock formerly was part of a single mass, which has been split and wedged apart about 3 miles by the younger Spook pluton.

In the southern part of the quadrangle is a nearly east trending mass of mafic plutonic rock that lies chiefly between the Dragon and Bullfrog plutons. This mass has been thoroughly injected and disrupted by a large number of dikes originating in the Bullfrog pluton.

Two other relatively large masses of mafic plutonic rock are in the drainage of Woods Creek. Both these masses are strongly sheared and intruded by a large number of mafic dikes of the Independence swarm. Many other mafic plutonic bodies of small size are scattered over the quadrangle; the larger of these are shown on the geologic map.

The dominant rock of the mafic plutonic bodies is mafic granodiorite or hornblende quartz diorite. These rocks vary notably in grain size and degree of foliation from place to place. They are locally strongly sheared. Irregular lenses and small bodies of hornblende gabbro, pyroxene gabbro, biotite rock, hornblendite, and perdiodite are present in the mafic plutonic masses. These lenses or inclusions are bordered by gradational feldspathized contacts, or by zones of contact breccia. Some of the rocks display typical igneous mineral assemblages and textures and are quite uniform over small areas. They are igneous mafic masses intruded early in the Sierra batholithic cycle.

In addition to mafic intrusive rocks, rocks of metamorphic origin are present in the mafic plutonic masses. Mafic metavolcanic rocks are common in the mass at the junction of the forks of Woods Creek along the John Muir Trail and on the east side of the mass of Armstrong Canyon. Many rocks have been so thoroughly recrystallized and feldspathized as to make it impossible to tell if they are of intrusive or volcanic origin.

The majority of the mafic plutonic rocks are similar mineralogically to the granitic rocks. They differ primarily in having a higher percentage of mafic minerals and a lower proportion of K-feldspar and quartz to plagioclase. K-feldspar is generally present in small amounts; even the gabbros generally contain a little interstitial K-feldspar. Quartz averages less than 20 percent of most of the mafic rocks. In many places it is strikingly interstitial and molded to the shape of the other minerals. Plagioclase is commonly strongly zoned. In composition it generally ranges from andesine to labradorite; and some of the gabbros and peridotites contain bytownite. In several thin sections, zoned plagioclase crystals were observed enclosing blebs of more calcic feldspar near the outer edge.

Most of the gabbros contain hornblende rather than pyroxene, but the stubby shape of the crystals, the presence of spongy cores of pyroxene surrounded by hornblende, and the locally interstitial character of the calcic plagioclase demonstrate the former presence of pyroxene. In some of the gabbros, pyroxene altered directly to a red-brown biotite; generally, however, it first altered to hornblende, which then commonly altered to biotite. The pyroxene is generally better preserved in the finer grained gabbros. Some gabbros contain two pyroxenes: hypersthene and a clinopyroxene.

Peridotite occurs as small, irregular masses in scattered localities in the mafic plutonic rock. One mass of peridotite on the east flank of Independence Peak consists chiefly of about equal amounts of olivine and hornblende. The olivine occurs in nearly euhedral crystals, 5 to 10 mm long, that are only slightly serpentinized. Brown hornblende, which encloses the olivine, contains some spongy cores of the pyroxene from which it was derived. Also present are pale biotite, interstitial labradorite, and abundant opaque minerals.

The mafic plutonic rock masses are commonly riddled by many kinds of dikes. A striking feature of some of the mafic plutonic bodies is the presence of swarms of nearly horizontal (dipping less than 30°) granitic dikes. The dikes occupy feather joints opened in the mafic plutonic body as a result of the emplacement of the parent granitic pluton. In some areas, as north of Glen Pass, the dikes are very thick and occupy a greater volume than the mafic country rock they intrude. These dikes are chiefly quartz monzonite and show the features of intrusive dikes. Many contain numerous inclusions of quartz diorite wallrock. The inclusions are in all stages of replacement and alteration. Some inclusions are sharp and angular, others are rounded or spindle shaped, drawn out in the direction of the dike. Even the angular inclusions are partly altered on their edges. One hornblende quartz diorite inclusion studied has a one-half-inch zone on its border in which the hornblende is pseudomorphosed by random-oriented coarse plates of biotite.

The subhorizontal dikes are probably best displayed in the mafic plutonic masses for three reasons: (1) the mafic bodies were generally early in the plutonic cycle and served as wallrock for many intrusions; (2) the mafic plutonic masses, unlike the metasedimentary or metavolcanic rock, do not have a strong, steeply dipping foliation which would restrain horizontal fractures; (3) the light color of the dikes contrasts with the darkness of the mafic plutonic rocks

Coarse hornblende pegmatites also cut the mafic plutonic masses. These pegmatites have the appearance of a replacement or recrystallization feature; they do not have matching walls and the borders are commonly gradational with the wallrock. One group of pegmatites is made up of large hornblende crystals up to 4 inches in length set in a coarse granitic matrix. Each hornblende crystal has a small core of plagioclase.

Many aplite and fine-grained granitic dikes which cut the dioritic areas have all the appearances of replacement dikes. Their walls are irregular and do not match, and some contacts are gradational. Shadows of replaced structures cross from the walls into the dike, and commonly transected structures are not offset. Many of the dikes cut inclusions without the other half of the inclusion appearing on the opposite side of the dike, indicating that the dike is either a replacement dike or was intruded along a fault for which there is no other evidence.

Large parts of the mafic plutonic rock areas are composed of plutonic breccia. Contact breccias are common along contacts of mafic rock with younger granitic plutons. In addition, large areas of breccia exist within the mafic bodies, particularly at internal contacts between hornblende gabbro or periodotite with quartz diorite or mafic granodiorite. The breccias are of at least two types: (1) intrusive breccias composed of intrusive plutonic rock enclosing angular fragments of more mafic rock (commonly dissimilar fragments), and (2) replacement breccias composed of coalescing replacement dikes enclosing rounded and corroded remnants of mafic rock (commonly similar fragments). In addition, a third type of breccia(?) is present. It is composed of a swarm of closely packed elliptical mafic inclusions of the type that is so common in the granitic rocks. The inclusions are rounded and generally fairly uniform in size (about 8 inches long and 3 inches wide) and compose over 50 percent of the rock. Such breccias are common in smaller areas of mapped mafic plutonic rock. They are believed to have formed by partial melting, flowage, and recrystallization of breccias formed by one of the two methods indicated above.

ANORTHOSITE

A single body of anorthosite was mapped in the quadrangle, near the headwaters of Thibaut Creek on the east scarp of the range. The mass is entirely enclosed in mafic plutonic rock and is generally in gradational contact with it. The rock to the east of the anorthosite is dominantly hornblende gabbro, and that to the west is diorite and mafic granodiorite. The age relation between the anorthosite and the mafic rock is not known.

The anorthosite is a light-gray, fairly uniform rock and can readily be mistaken for granodiorite from a distance. The grain size is about 4 mm. Over 90 percent of the rock is made up of euhedral plagioclase crystals of uniform size. The average composition is about An_{70}, but faint zoning, both progressive and oscillatory, is present in most crystals. The most abundant mafic mineral is clinopyroxene which has been altered to hornblende and chlorite. Opaque accessories are abundant. Quartz and K-feldspar are present in small amounts (table 2) in the interstices between the plagioclase crystals. Commonly several discrete interstitial grains of quartz show the same optical continuity.

The euhedral form and uniform size of the plagioclase crystals and the distinct interstitial and late crystallizing appearance of the quartz and K-feldspar suggest that the anorthosite is of igneous origin; probably it represents an accumulation of plagioclase crystals that were separated (perhaps by gravitational settling) from a mafic magma.

GRANITIC ROCKS

Granitic rocks crop out over a greater area in the Mount Pinchot quadrangle than any other rock type; they occupy about 160 of 190 square miles of exposed bedrock. Only scattered remnants of the metamorphic and mafic plutonic rocks into which the granitic rocks were intruded are present in the dominantly granitic terrane.

Twenty-seven separate granitic masses were mapped on the basis of texture, color index, internal structure, character of mafic inclusions, and other field criteria. Of these 27, 3 are correlated with 2 granitic formations named outside the quadrangle (Lamarck granodiorite and Tinemaha granodiorite, Bateman, 1961). Of the 24 remaining, 4 are believed to have been intruded as 2 separate masses (Pyramid pluton and Mule Lake pluton), each of which was split in half by a younger intrusion. Hence, on the map legend 24 separate granitic units have been shown, 2 formations and 22 plutons. Actually some of these may be truncated apices of intrusive masses that are continuous in depth, and others may be parts of masses that were originally continuous at the level of the present surface, which have been split apart by younger intrusives. However, a great many plutons appear to represent separate intrusive impulses. With respect to the great number of separate intrusions, the quadrangle is unlike adjacent areas to the north and northwest, and probably to the south, but the reason for its uniqueness is not known.

In composition, the granitic rocks range from granodiorite to alaskitic quartz monzonite according to the classification used in this report (fig. 10). Of the 27 granitic masses, 1 is quartz diorite, 15 are granodiorite, 5 are quartz monzonite, and 6 are alaskite (quartz monzonitic). Figure 11 shows the positions of the average mode for 24 of these masses on a triangular diagram whose corners are K-feldspar, plagioclase, and quartz. The greatest concentration of modes in this diagram, and the grand average, is close to the boundary line between the granodiorite and quartz monzonite fields.

In composition and texture, the rock within individual plutons is distinctive enough to be readily separated from the rock of neighboring bodies. Some plutons, however, are compositionally zoned from a felsic core to a mafic margin, and the appearance of the rock in the margins is very different from that in the cores. Laboratory studies suggest that zonation too subtle to be readily observable in the field may exist in most and perhaps all plutons.

In size the plutons range from less than 1 square mile to more than 20 square miles. Commonly they are elongate or elliptical in plan, with most long axes oriented about N. 30° W., subparallel to other elements of the bedrock structure. Where plutons meet, the boundary surface is nearly always sharp, and the intrusive relations can generally be determined readily. Nevertheless, a few sharp contacts are so featureless that the relative ages of the plutons could not be determined. Very commonly plutons are separated by thin septa of meta-

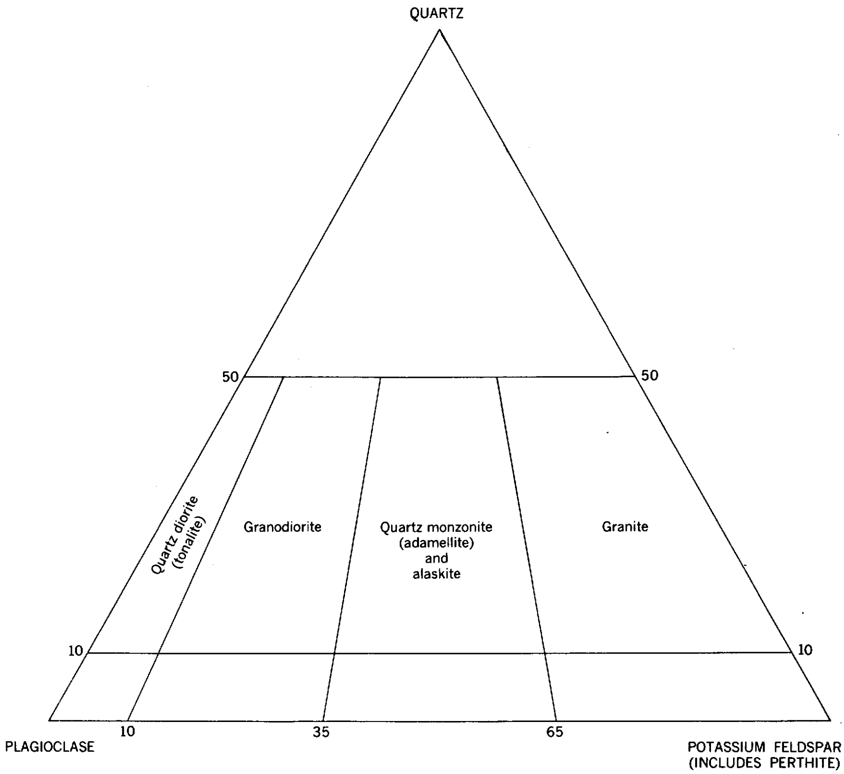
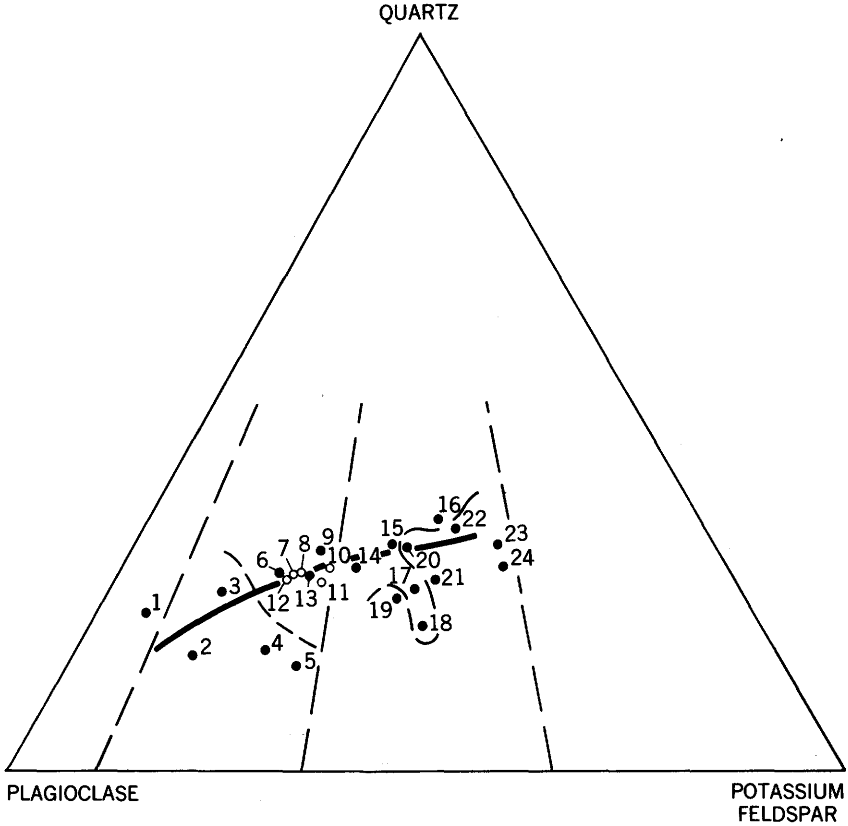


FIGURE 10.—Classification of granitic rocks used in this report. Diagram modified after Johannsen, 1931.

morphic rock. Most septa and contacts between plutons dip steeply or are vertical.

In the past the granitic rocks of the Sierra Nevada have been considered to be of Jurassic age on the basis of geologic relations in the vicinity of the Shasta Bally batholith in north-central California (Hinds, 1934). However, 7 granitic bodies in the Yosemite region have been determined by the potassium-argon method to range from 77 to 95 million years in age and the Shasta Bally batholith to be 134 million years old (Curtis, Evernden, and Lipson, 1958.) The age of the granitic rocks in the Bishop district, immediately to the north, has been determined by the lead-alpha method to be 91 to 116 million years (Larsen, Gottfried, Jaffe, and Waring, 1958). Inasmuch as the granitic rocks in the Mount Pinchot quadrangle are partly continuous with the granitic rocks in the Bishop district and probably equivalent in part to the granitic rocks of Yosemite, they are assumed to be of the same general age. All the ages cited fall within the Cretaceous period according to the Holmes "B" time scale (Holmes,



Calcic granodiorite:
1. Baxter pluton
2. Striped pluton
3. Pyramid pluton
4. McDoogle pluton
5. Dragon pluton

Silicic granodiorite:
6. Arrow pluton
7. Spook pluton
8. Lamarck mass (Lamarck granodiorite)
9. White Fork pluton
10. Woods Lake mass (Tinemaha granodiorite)
11. Paradise pluton
12. Cartridge Pass pluton
13. Cotter pluton

Quartz monzonite:
14. Colosseum mass (Lamarck granodiorite)
15. Siberian pluton
16. Taboose pluton
17. Goodale pluton
18. McGann pluton

Alaskite:
19. Twin Lakes pluton
20. Sardine pluton
21. Bullfrog pluton
22. Red Mountain Creek pluton
23. Independence pluton
24. Diamond pluton

FIGURE 11.—Averages of modes of 24 intrusive masses in the Mount Pinchot quadrangle. Open circles are plutons with porphyritic K-feldspar. Intrusive masses are divided into four compositional groups by dashed lines. Heavy line is apparent trend of differentiation.

1948) and other recently published charts that show probable correlations between radiometric and paleontologic time.

METHODS OF ANALYSIS

Micrometric modal analyses were made of over 230 thin sections of granitic rocks to determine the volume percent of the component minerals. The analyses were performed by the point-count method described by Chayes (1949) on a mechanical stage with an intercept of 0.3 mm. Full thin sections, 40 by 24 mm, were used. To facilitate the distinction of K-feldspar from quartz and plagioclase, all slides were stained with sodium cobaltinitrite after a hydrofluoric acid etch (Chayes, 1952).

On each slide between 1,100 and 1,500 points were identified and tabulated. The error arising from measurement alone should be less than 3 percent for the minerals composing more than 30 percent of the rock, and smaller for the minor constituents. The error arising from sampling, especially of coarser grained rocks, may be considerably larger. In order to determine the average composition, many samples of each rock unit were modally analyzed.

Modal analyses are reported in volume percent of the constituent minerals. Where five or more samples from the same pluton were analyzed, the standard deviation of percents of each mineral was calculated. The modes have been recalculated to 100 percent for the essential minerals (quartz, K-feldspar, plagioclase) and plotted on a triangular diagram for each pluton to portray graphically the range of composition within the mass.

Specific gravity determinations were made on a beam balance by balancing the specimen in air and in water, and measuring the change in length of the lever arm. Measurements are reproducible to 0.01.

Plagioclase compositions were determined by optical and X-ray methods. The powder X-ray method was used because an average value for the oscillatory zoned crystals can be determined. The sample is prepared by crushing the rock and staining the aggregate with sodium cobaltinitrite. The cleaved unstained white crystals (plagioclase) are hand picked, finely ground, and prepared with water on a glass mount. The $220\text{–}1\bar{3}1$ and $131\text{–}1\bar{3}1$ peak intervals are measured to determine mol percent anorthite; the curve for low-temperature plagioclase is employed (Smith and Yoder, 1956).

Field measurements of the abundance of mafic inclusions were made by pacing off an area, generally a glaciated surface, of 8 or 16 square yards and counting the number of mafic inclusions in this area.

GENERAL PETROLOGY

Although the grain size, texture, and mineral content of the granitic rocks are characteristic for each pluton, the general petrologic features are similar and will be discussed here. The distinctive and characteristic features of each pluton will be discussed in the individual pluton descriptions.

The average grain size of all of the granitic rocks is perhaps 2 to 4 mm, though this varies from specimen to specimen, from one mineral species to another, and within the same mineral species in the same specimen. Some plutons are porphyritic. The only mineral that occurs as phenocrysts is K-feldspar. The Paradise pluton has perhaps the largest K-feldspar phenocrysts in the quadrangle, and these attain a maximum length of 1½ inches. All the porphyritic granitic rocks occupy a somewhat restricted compositional range lying close to the quartz monzonite–granodiorite boundary (fig. 11).

The overall average texture can best be described as hypidiomorphic granular, except in the older plutons (such as the White Fork pluton) where gneissose and cataclastic textures are common. Plagioclase is generally fairly euhedral, with strikingly euhedral inner zones. Orthoclase (and microcline) is subhedral, with small-scale ragged and irregular borders, and poikilitically encloses plagioclase and ferromagnesian minerals. Quartz is generally interstitial, being molded between the feldspars; disconnected grains are commonly in optical continuity over large areas, and were doubtless connected above or below the plane of the section.

Nearly all the granitic rocks are believed to possess a preferred orientation of their mineral grains, but only in the more calcic granodiorites is it conspicuous. In most rocks the orientation is planar, but in a few it is linear; in some masses both foliation and lineation are visible (with the lineation in the plane of the foliation). The preferred orientation is marked by two elements: (1) mafic minerals, chiefly hornblende, and (2) elongate mafic inclusions. Both these elements are rare or absent in the lighter colored quartz monzonites and alaskites, and consequently the preferred orientation of the lighter colored granitic rocks is difficult to detect and measure.

In the lighter colored granitic rocks the mafic inclusions which mark the structure are small and widely spaced. In areas of horizontal outcrop, sections necessary to distinguish foliation from lineation from both foliation and lineation are not easy to find. In these places a bearing symbol, which records the strike of a foliation or the bearing of a lineation without regard to dip or plunge, has been recorded.

Under the microscope most granitic rocks present only a slight alinement of minerals. There is some indication of slight movement after crystallization; quartz commonly shows mosaic or undulant extinction, biotite flakes and plagioclase lamella are bent in some sections, and K-feldspar may show warped extinction and development of microcline grid twinning along zones of strain.

Plagioclase is the most abundant mineral in the average granitic rock. It occurs in well-formed euhedral to subhedral crystals which average 3 mm in greatest dimension. Virtually every plagioclase grain studied is zoned; oscillatory zoning superimposed on progressive zoning from a calcic core to a sodic rim is common. As many as 40 oscillations have been found in a single crystal. The average plagioclase in most plutons contains about 30 percent anorthite, but a single grain commonly varies 20 percent in anorthite content from core to rim. The plagioclase of the alaskite units contains 10 to 15 percent anorthite and is less strongly zoned.

Sericite is the chief alteration product of the plagioclase. The cores of the plagioclase crystals are much more susceptible to alteration, in part because of the lower percentage of silica in the more calcic cores and in part because of the less stable nature of anorthite in a deuteric environment of lower temperature; the cores are altered to sericite, epidote, and sometimes chlorite. In some specimens the plagioclase lamella are bent and faulted by postcrystallization movement.

Antiperthitic intergrowths of potassium feldspar in plagioclase crystals are common. The K-feldspar occurs in small, nearly equidimensional grains, and probably makes up less than 2 percent of the total volume of the antiperthite. Nearly all of the grains of K-feldspar in the antiperthite are in optical continuity.

Myrmekitic intergrowths of quartz are common in the plagioclase where it is in contact with K-feldspar. The border of the myrmekite and K-feldspar swells and projects slightly into the K-feldspar. Myrmekite is more common on the ends of plagioclase crystals than on the sides.

In all of the granitic rocks except alaskite K-feldspar is generally less abundant than plagioclase; commonly K-feldspar is in larger crystals (approx. 4 mm) than either quartz or plagioclase. The general larger size of the K-feldspar is reflected in the general higher standard deviation for the volume percents of K-feldspar than for the other essential minerals in the rock, especially since it is generally the least abundant of the three essential minerals. Perhaps this higher deviation from the mean for K-feldspar is also partly due to the somewhat irregular distribution of K-feldspar in the rock. The potassium feldspar grains are commonly poikilitic and contain abundant crystals of plagioclase and some biotite. Much of the K-feldspar shows twinned and untwinned areas within the same crystal. The twinned areas (microcline) are commonly in regions of strain within the crystal, for example, along small fractures and shears, or adjacent to corners of other large minerals impinging on the K-feldspar. The K-feldspar in most of the granodiorites contains little plagioclase, in contrast to the K-feldspar in the alaskites, which is characteristically perthitic.

Quartz occurs as fairly small (average 2 mm) grains molded in the interstices between the feldspar grains. The quartz has all the appearance of being a late crystallizing mineral which has had to take the space available and conform to the crystal outline of the earlier minerals. Many of the quartz grains are strained and show distorted extinction. In some cases the quartz displays bands or zones of slightly different orientation. In other specimens the quartz shows a mosaic extinction caused by shear in two nearly perpendicular directions breaking up the grain into many small rhombs. This mosaic extinction is very common in the quartz and serves as a reliable criterion for the separation of quartz and other transparent low-birefringent minerals, as in making a modal analysis.

Biotite is the most common varietal mineral in nearly all the granitic rocks. No granitic rock is without biotite. The biotite is dark brown and strongly pleochroic. In most specimens, about one-fifth of the mineral has been chloritized, with the resultant formation of magnetite, sphene, and (in some places) small blebs of orthoclase:

biotite + water———chlorite + K-feldspar + magnetite + sphene.

Where biotite is chloritized the chlorite has been tallied with the biotite in the modal analyses. In a few specimens biotite or chlorite is interlaminated with muscovite, though muscovite is rare and present only in the silicic granitic rocks. Epidote is also found interlaminated with biotite. Pleochoric haloes commonly surround small inclusions in biotite.

Hornblende is generally present in the granitic rock as small, somewhat irregular and spongy prisms. The hornblende is green; its pleochroic scheme is: X, light green; Y, yellowish green; Z, bluish green. As the color index of the rock increases the hornblende-biotite ratio increases and the hornblende becomes coarser. Many prisms of hornblende contain patches of biotite, and some prisms are surrounded by irregular rims of biotite. This evidence suggests that the hornblende is subject to alteration, probably late magmatic or deuteric, to biotite.

Only the more mafic granitic rocks contain pyroxene. Clinopyroxene, probably augite, is present in the rocks as cores in hornblende crystals, generally as highly irregular, spongelike remnants. Toward the pyroxene, the hornblende gradually changes to a colorless amphibole. Hornblende was clearly formed from the pyroxene.

Among accessory minerals, sphene is probably the most abundant. It occurs commonly in the typical rhomb shape and is probably largely magmatic; however, some sphene is a byproduct of the chloritization of biotite. Sphene is also present as irregular rims on some of the opaque grains and may represent alteration of ilmenite.

Opaque grains, probably mainly magnetite but including some ilmenite, are also common as accessory minerals. Apatite and zircon are nearly always present. Allanite, commonly showing an outer rim of epidote, is present in small amounts.

MODAL AND CHEMICAL TRENDS

The average mode of each pluton has been recalculated to 100 percent for the essential minerals (quartz, K-feldspar, plagioclase) and plotted on a triangular diagram (fig. 11). Most of the modal averages plot near a line extending from near the plagioclase corner to about the center of the triangle. This line is believed to represent the course of differentiation in magma that was parent to the individual plutons.

Several points fall below the apparent trend of differentiation; that is, they have less quartz than most of the other "normal" granitic masses of the same K-feldspar content. The McDoogle, Dragon, Goodale, and McGann plutons and a part of the Bullfrog pluton are poor in quartz. Each of these is adjacent to a major mass of one of the older dioritic complexes (pl. 1). Field evidence (discussed in the individual pluton descriptions) suggests that some of these plutons have reacted extensively with the dioritic complexes, and it is postulated that the unusually small amount of quartz in the plutons results from contamination by their mafic wall rocks. However, other intrusive masses which are adjacent to masses of mafic rock have a more normal composition and appear to be uncontaminated. Perhaps these relatively uncontaminated magmas were cooler and less reactive, or were in contact with mafic wall rock through a smaller vertical span.

The experimental data in the synthetic system albite–K-feldspar–quartz–water, determined by Bowen and Tuttle (Bowen, 1954), can be plotted on the same triangular diagram as figure 11. Such a comparison is approximate only, since several constituents present in the rock specimens are not present in the synthetic system. However, the specimens which plot near the center of the triangle are reasonably well represented, for they contain relatively minor amounts of anorthite and mafic minerals. The minimum melting trough of the system albite–K-feldspar–quartz, at 1,000 bars water pressure, falls near the center of the triangle, near the average modes of the Red Mountain Creek or Taboose plutons. The trend line of modes appears to head toward this trough and terminate within it, as would be expected if differentiation followed crystal–liquid equilibrium and tended toward producing a result with the lowest possible melting point.

Several parameters derived from modal analyses may be used as an index of the stage of differentiation of a granitic rock. However, as seen in the triangular diagram of figure 11, the main trend line of average modes is most nearly at right angles to the quartz–K-feldspar edge. Therefore the plagioclase content is a good indicator as to the position of a point on the trend line.

The plagioclase content of the granitic rocks varies from more than 55 volume percent to less than 25 volume percent (table 3). Clinopyroxene is commonly present in the cores of hornblende crystals in the plagioclase-rich rocks, but it disappears when plagioclase is less than approximately 40 percent. The percent anorthite in plagioclase decreases as the amount of plagioclase decreases; as the volume percent of plagioclase decreases from 55 to 25 percent, the mol percent of anorthite decreases from approximately 35 percent to 10 percent. As plagioclase decreases, both biotite and hornblende become less abundant (fig. 12). Biotite is always more abundant than hornblende by approximately 4 percent. Muscovite occurs in trace amounts when plagioclase is less than approximately 40 percent. Perthite is common only when the plagioclase carries less than 15 percent anorthite.

Chemical analyses of 10 representatives samples of granitic rocks are given in table 4, are plotted on a variation diagram, figure 13. These analysed samples were chosen on the basis of modal analyses as representing best the average composition of the mass from which they come. The average SiO_2 content of all the granitic rocks of the quadrangle, exclusive of mafic plutonic rocks, is 68.2 weight percent. This average has been calculated by considering the area and composition of each pluton and weighing the average in proportion to the area of the mass. The average rock would represent a quartz monzonite close to a granodiorite in composition, according to the chemical averages of Nockolds (1954, p. 1014), who uses a modal classification very similar to that used in this report (fig. 10).

DESCRIPTION OF PLUTONS

The compositions of the various granitic plutons of the quadrangle are fairly well known as a result of field and laboratory studies. However, the relative ages are only known in an incomplete fashion because many plutons are not in contact with one another. For this reason, the plutons will be discussed in compositional groups. The plutons of each compositional group are not necessarily of the same age, though any age correlations that can be made will probably be made among plutons of the same compositional group.

RELATIVE AGE RELATIONS

Several criteria have been used to determine the relative ages between two adjacent intrusive masses. Generally more than one critical feature is present at a single contact, and comparison between these features tests the reliability of any single criterion. Some of the

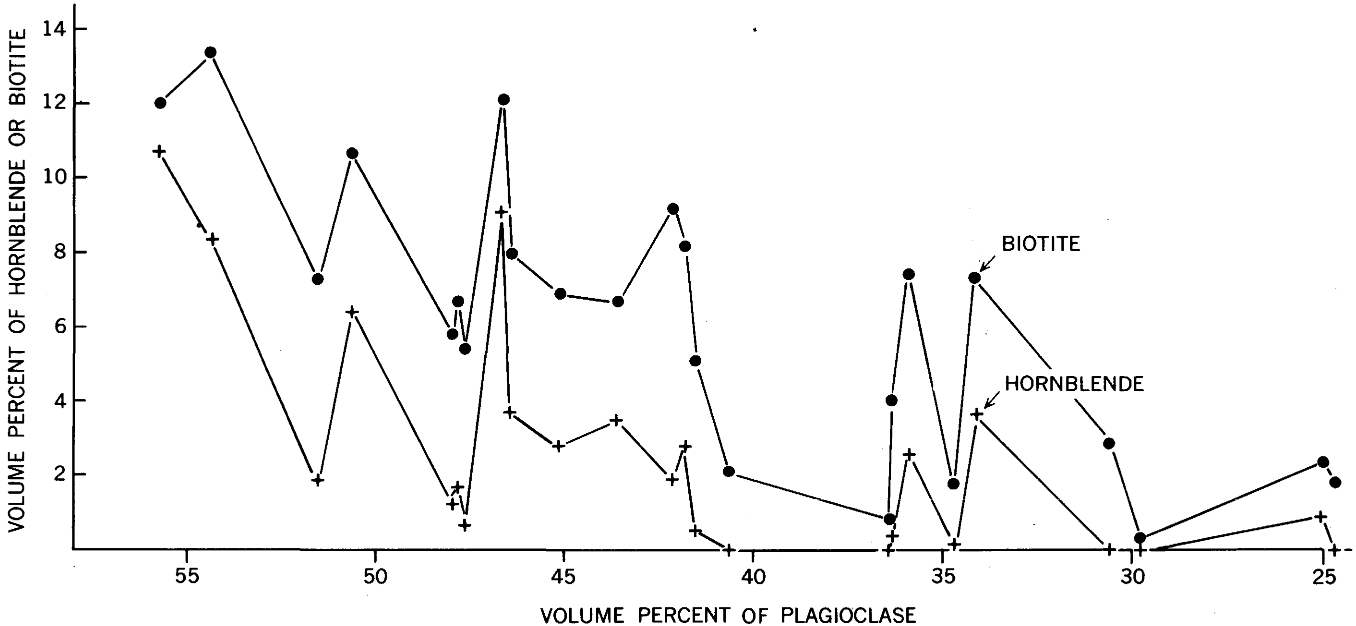


FIGURE 12.—Volume percent of biotite and hornblende plotted against volume percent of plagioclase. Each pair of points represents an average of several modal analyses from a single intrusive mass.

TABLE 3.—*Average modal composition of granitic intrusive masses in the Mount Pinchot quadrangle*

[Volume percent of minerals]

Intrusive mass	Area in sq miles	Quartz	K-feld-spar	Plagio-clase	Biotite	Horn-blende	Pyroxene	Opaque accesso-ries	Non-opaque accesso-ries	Musco-vite	Color index	Sp gr	Percent An in pla-gioclase
Baxter pluton	1.6	16.8	4.7	55.7	12.0	9.7	----	0.6	0.4	----	22.8	2.76	32
Striped pluton	.6	11.8	10.8	54.3	13.3	8.4	0.2	.5	.6	----	23.0	----	----
Pyramid pluton	[1] 3.2	19.7	10.9	50.6	10.7	6.4	.3	.9	.5	----	18.8	2.72	32.5
McDoogle pluton	5.3	12.4	17.5	46.6	12.1	9.1	.6	.9	.9	----	23.6	2.74	31
Dragon pluton	[1] 6.9	12.7	24.5	51.5	7.3	1.9	Tr.	1.2	.8	----	11.2	2.66	23
Lamarck granodiorite	[1] 9.6	----	----	----	----	----	----	----	----	----	----	----	----
Lamarck mass	[1] 7.0	24.8	20.3	47.6	5.4	.7	----	.8	.5	----	7.4	2.66	29
Colosseum mass	2.6	25.8	26.3	41.5	5.1	.5	----	.5	.3	----	6.4	2.62	30
Cartridge Pass pluton	14.1	23.4	18.3	47.8	6.7	1.7	----	.8	.5	----	9.7	----	----
Arrow pluton	[1] 14.4	23.3	17.0	46.4	8.0	3.7	Tr.	1.0	.6	----	13.3	2.69	33
Spook pluton	[1] 13.3	24.4	19.3	47.9	5.8	1.3	----	.7	.6	----	8.4	----	----
Inner facies	----	24.4	20.4	49.2	4.7	.1	----	.7	.5	Tr.	6.0	2.62	26
Outer facies	----	24.5	18.3	46.7	6.9	2.4	----	.7	.6	----	10.6	2.66	32
Tinemaha granodiorite													
Woods Lake mass	12.4	23.6	21.8	41.8	8.2	2.8	----	1.1	.8	----	12.9	2.65	30
Cotter pluton	6.9	23.5	20.3	45.1	6.9	2.8	----	.9	.5	----	11.1	2.67	31
Paradise pluton	[1] 11.6	22.5	22.0	43.6	6.7	3.5	----	.9	.9	----	12.0	2.67	30
Mule Lake pluton	2.2	----	----	----	----	----	----	----	----	----	----	----	----
White Forks pluton	7.9	26.4	19.5	42.1	9.2	1.9	----	.6	.3	----	12.0	----	----
Siberian pluton	2.5	29.5	29.6	36.3	4.0	----	----	.4	.2	----	4.6	2.60	26
Taboose pluton	[1] 3.5	32.9	33.2	30.6	2.9	----	----	----	.4	----	----	----	----
McGann pluton	[1] 8.3	17.2	35.5	35.9	7.4	2.6	----	.6	.9	----	11.5	2.64	28
Goodale pluton	2.3	21.9	32.1	34.2	7.3	3.7	----	.5	.3	----	11.8	2.65	----
Red Mountain Creek pluton	[1] 2.7	32.8	37.0	29.8	.3	----	----	.2	----	----	.4	2.60	16
Independance pluton	[1] 8.3	30.0	42.7	24.7	1.8	----	----	.6	.3	----	2.7	2.59	14
Twin Lakes pluton	1.2	22.7	34.1	40.6	2.1	----	----	.4	.1	----	2.6	----	----
Bullfrog pluton	[1] 22.9	25.0	37.3	34.7	1.8	.2	----	.6	.4	Tr.	3.0	2.60	10
Sardine pluton	.3	30.0	32.3	36.4	.8	----	----	.4	----	----	1.2	----	----
Diamond pluton	1.3	26 2	43.4	25.0	2.4	.9	----	1.7	.5	----	5.5	2.60	----

[1] Only that part of area within Mount Pinchot quadrangle is measured.

TABLE 4.—*Rapid chemical analyses and norms of granitic rocks from the Mount Pinchot quadrangle*

[Numbers at the head of columns refer to the following intrusive masses and samples: 1, McDoogle pluton, sample 1; 2, Dragon pluton, sample 8; 3, Pyramid pluton, sample 3; 4, Spook pluton, sample 10; 5, Arrow pluton, sample 15; 6, Woods Lake mass, Tinemaha granodiorite, sample 1; 7, Colosseum mass, Lamarck granodiorite, sample 15; 8, Cotter pluton, sample 4; 9, Bullfrog pluton, sample 6; 10, Siberian pluton, sample 3.
Analysts: Paul L. D. Elmore, Samuel D. Botts, Marvin D. Mack, and Herman H. Thomas, U.S. Geological Survey.
Samples analyzed are from localities shown on figures 18, 19, 17, 31, 27, 33, 20, 34, 47, and 49, respectively]

	1	2	3	4	5	6	7	8	9	10
Chemical analyses (weight percent)										
SiO_2	62.0	62.8	65.2	65.7	66.7	67.4	68.5	68.9	73.0	73.9
Al_2O_3	16.6	17.0	16.7	16.7	16.1	15.8	16.1	15.5	14.6	14.5
Fe_2O_3	2.1	2.4	1.9	1.8	2.1	1.7	1.6	1.6	.9	.7
FeO	3.4	2.8	2.2	2.1	2.2	2.2	1.6	1.4	1.0	.8
MgO	2.2	1.5	1.6	1.2	1.4	1.3	.88	.93	.59	.26
CaO	4.5	3.8	4.4	4.0	3.8	3.1	3.2	3.2	.71	1.2
Na_2O	3.7	4.8	3.7	4.0	3.6	3.5	3.9	3.8	4.5	3.3
K_2O	3.7	3.2	3.1	3.0	3.1	4.2	3.4	3.4	4.9	4.9
H_2O	.79	.62	.63	.92	.69	.48	.50	.37	.27	.41
TiO_2	.77	.76	.58	.49	.48	.42	.42	.43	.21	.15
P_2O_5	.21	.26	.18	.18	.18	.14	.14	.14	.02	.01
MnO	.10	.15	.09	.10	.10	.12	.10	.09	.08	.06
CO_2	.19	.12	.05	<.05	.08	.05	.03	<.05	.08	<.05
Sum	100.2	100.2	100.3	100.6	100.5	100.4	100.3	100.1	100.8	100.6
Norms (CIPW)										
Qu	13.1	12.6	19.9	20.2	23.3	21.8	24.5	25.9	25.2	32.9
Or	21.7	18.9	18.4	17.8	18.4	25.0	20.0	20.0	28.9	28.9
Ab	31.4	40.4	31.4	34.1	30.4	29.3	33.0	32.0	38.3	27.8
An	17.8	15.6	19.7	18.6	18.1	14.5	15.0	15.0	3.6	5.8
wo	1.5	.7	.6	.1	--------	--------	--------	--------	--------	--------
en	5.5	3.8	4.0	3.0	3.5	3.3	2.2	2.3	1.5	.7
fs	3.2	1.9	1.5	1.6	1.6	2.0	.9	.5	.7	.5
mt	3.0	3.5	2.8	2.6	1.6	2.6	2.3	2.3	1.4	.9
il	1.5	1.5	1.2	.9	.9	.8	.8	.8	.5	.5
ap	.3	.7	.3	.3	.3	.3	.3	.3	--------	--------
C	--------	--------	--------	--------	.2	.2	.5	.1	.5	1.6

criteria appear to be consistently reliable, others are only suggestive if used alone, and others are absent in all but a very few contacts and hence are of limited use.

One of the most useful criteria for determining relative age is the presence of a progressive change in one of the masses toward the contact while the other mass remains unchanged. The younger mass generally shows the greatest changes toward the contact, for the contact is the actual wall of the magma chamber and therefore had a strong physical discontinuity and thermal gradient across it at the time that the younger pluton was crystallizing. However, the older unit could have been split apart by the younger, and in that case the contact would have been an arbitrary line within the older mass. Commonly, the younger unit becomes finer grained or more mafic toward the contact. Mafic inclusions increase in abundance toward the contact. In places the size and number of K-feldspar crystals decreases toward the contact. In one contact studied, the younger pluton became progressively more pegmatitic toward the contact.

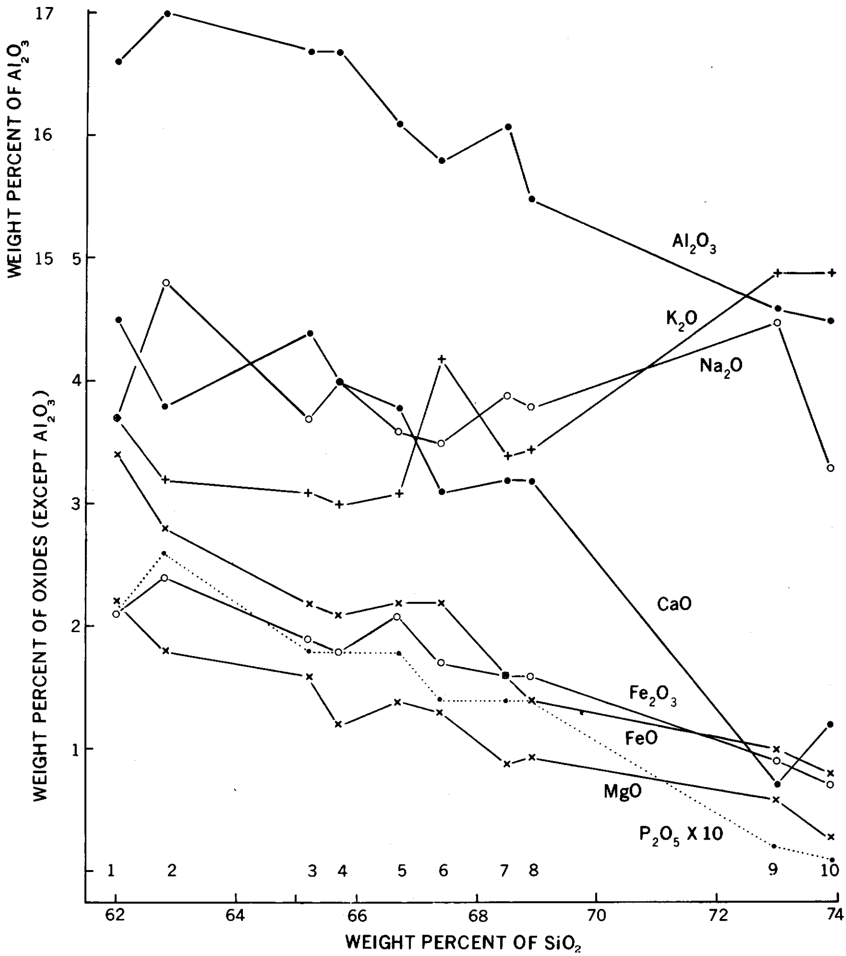


FIGURE 13.—Variation diagram of oxides plotted against SiO_2 for some representative granitic rocks of the Mount Pinchot quadrangle. Numbers on diagram refer to the following masses and samples: 1, McDoogle pluton, sample 1; 2, Dragon pluton, sample 8; 3, Pyramid pluton, sample 3; 4, Spook pluton, sample 10; 5, Arrow pluton, sample 15; 6, Woods Lake mass, Tinemaha granodiorite, sample 1; 7, Colosseum mass, Lamarck granodiorite, sample 15; 8, Cotter pluton, sample 4; 9, Bullfrog pluton, sample 6; 10, Siberian pluton, sample 3.

The internal structures of the younger intrusive mass are increasingly well developed toward the contact. Mafic inclusions become more elliptical and elongated and mafic minerals develop a more perfect preferred orientation. At some contacts the younger rock develops a layered concentration of mafic minerals. In other places mafic minerals or inclusions define a festoon or swirled structure parallel or tangential to the contact.

Where an intrusive contact transects a directional structure, that structure belongs to the older intrusive mass. This structure can be marked by a preferred orientation of mafic minerals, elongated mafic inclusions, or K-feldspar phenocrysts; or it can be a separate structure within the older rock, such as a vein, dike, shear, mafic inclusion, or a contact between two rocks types.

In some places a younger intrusive mass is intruded along the contact between two older plutons without cutting either of them. Where this occurs, the contact relations are commonly ambiguous and the two criteria described above (progressive change and cross-cutting relations) are useless.

Where apophyses or tongues of one rock are intrusive into another, the intrusive rock is younger. However, apophyses that are continuous with one mass and crosscut another are scarce. More commonly the apophyses are dikelike and cut both masses, even the younger mass to which they are satellitic.

Where inclusions of one rock are enclosed in another, the host rock is younger. Inclusions are not always positively recognizable, but commonly porphyritic or rather mafic or felsic inclusions can be correlated with a neighboring intrusive mass with reasonable accuracy.

The degree of shearing and recrystallization can in some places be used as a guide to age. If one mass is extensively sheared and recrystallized and a pluton in contact with it is not, the unaffected one is considered younger. The younger pluton is presumably younger than the period of metamorphism that affected the older pluton.

The presence or absence of a prevailing structural element is an indication of age, even if the element is not continuous enough to be actually observed as truncated at the contact. The unit with the prevailing structural element, such as a swarm of dikes or a set of shears or veins, is the older.

Some of the granitic intrusions are domical and concentric, appearing on the map as complete or nearly complete rings. Generally the inner pluton is the younger, and where there are more than two concentric masses, each successive inner mass is younger than the one that surrounds it. Likewise in domical intrusions, the overlying or roof rocks are older than the intrusion.

Because many plutons are not in contact with one another and also because some of the contacts are nondiagnostic, it is not possible to establish a complete intrusive sequence among all the mapped plutons. However, enough relations are known to deduce possible sequences and to establish some general relations. Figure 14 presents a possible intrusive sequence which is in agreement with known age relations. This is the same sequence as that shown on the map legend. The plutons are divided into a predike group and a postdike group. All pairs of plutons in the same group for which there is a relative age determination are connected with a line, the one higher on the chart being younger. One striking feature of this diagram is that at nearly every

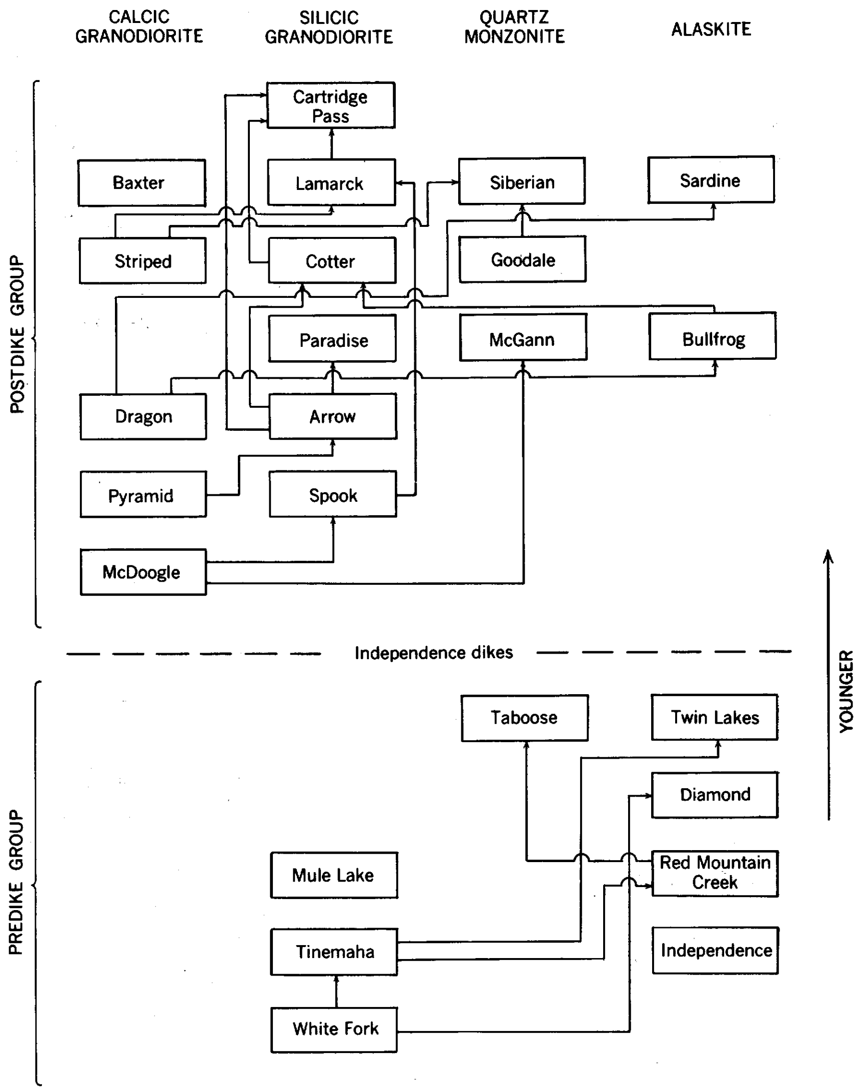


FIGURE 14.—Diagram showing a probable intrusive sequence among granitic masses, youngest toward the top. The intrusive masses are arranged in four compositional groups progressively more silicic to the right. Lines connecting boxes indicate known relative ages. Names on the same horizontal level are not necessarily of the same age.

contact the more silicic pluton is the younger. The only two exceptions are the contact between the Bullfrog and Cotter plutons and between the Taboose and Red Mountain Creek plutons (excluding contacts between masses of the same general compositional group). This fact raises some question of the validity of the intrusive sequence shown in figure 14. Perhaps each compositional group is of about the same general age and the next more silicic group is successively younger. On the other hand, perhaps each mafic mass regardless of its period of intrusion has differentiated to give rise to younger more silicic masses, and these more silicic masses, because of their proximity to the parent mass, are more likely to be in contact with the more mafic parent mass than with an unrelated pluton.

Regardless of which interpretation is correct, however, the known age relations suggest a rather pronounced break in the plutonic cycle at the time of intrusion of the Independence dikes (Moore and Hopson, 1961, p. 254–258). The data suggest that the dikes were preceded by the intrusion of rather silicic granitic masses and followed by the intrusion of rather mafic granitic masses.

CALCIC GRANODIORITE PLUTONS

Five plutons composed of granodiorite are present in a broad belt extending from the northwest corner to the southeast corner of the quadrangle. They are, from north to south, the Striped, Pyramid, McDoogle, Baxter, and Dragon plutons. All of these plutons appear to belong to the older part of the younger postdike series of intrusions. Known age relations do not exclude the possibility that all five of these separate masses are of the same age (fig. 14). However, they range quite broadly in composition and appearance from one mass to another.

BAXTER PLUTON

The Baxter pluton is a small (1.6 square miles) elliptical pluton just west of the Sierra crest in the middle of the quadrangle. The pluton is younger than all of its walls. It is situated squarely on strike with, and at the north end of, the Oak Creek metavolcanic pendant. The pluton is named after Baxter Creek, which traverses the southern part of the mass.

The Baxter pluton is composed of mafic, fine-grained granodiorite or quartz diorite (fig. 15 and table 5). The rock contains abundant small, dark, nearly round inclusions. Generally about 6 to 10 inclusions can be seen in each square yard of exposure.

The rock of the Baxter pluton is homogeneous over much of its extent. However, the southern part, west of the main Baxter Lake,

TABLE 5.—*Modes of Baxter pluton*

Sample No.	Quartz	K-feld-spar	Plagio-clase	Biotite	Horn-blende	Opaque acces-sories	Non-opaque acces-sories	Color index	Spgr	Percent An in plagio-clase
1	17.3	2.6	57.0	11.8	10.4	0.7	0.1	23.0	2.78	32
2	16.2	6.8	54.5	12.1	9.0	.6	.8	22.5	2.74	32
Av	16.8	4.7	55.7	12.0	9.7	.6	.4	22.8	2.76	32

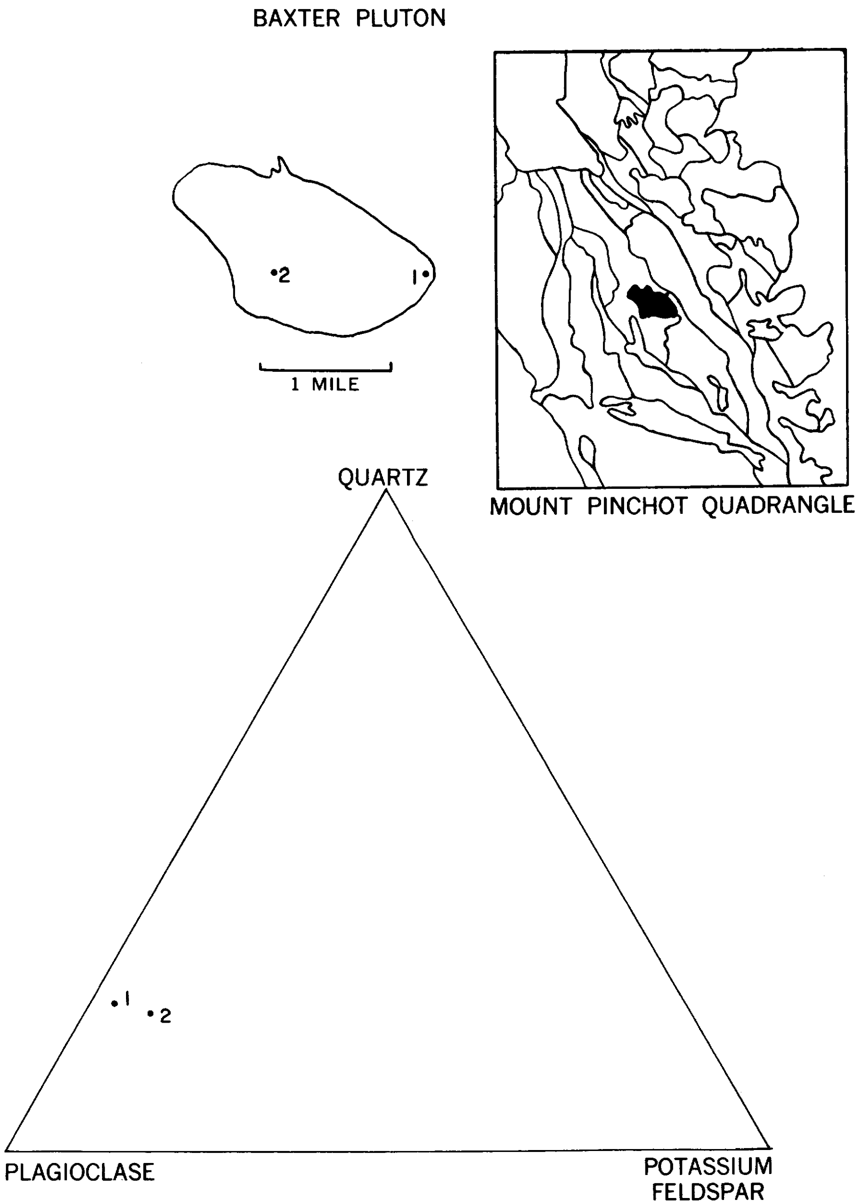


FIGURE 15.—Location and modal composition of samples of the Baxter pluton. Numbers refer to samples given in table 5.

is a migmatitic complex of mafic quartz diorite and lighter granitic material. Also much of the northern part is composed of injection zones and dikes of alaskitic rock that cannot be related to any neighboring pluton.

The Baxter pluton contains no mafic dikes of the Independence swarm, although it is surrounded on all sides by rock masses that are intruded by swarms of the dikes. The Baxter pluton belongs to the suite of plutons that were intruded after the mafic dikes, but the age relation of the Baxter pluton to other postdike plutons is not known.

STRIPED PLUTON

The Striped pluton is a small (0.6 square mile) mass in the north-central part of the quadrangle. It is confined to the slopes of Striped Mountain, from which it was named. Striped Mountain gets its name from the presence of a large number of light-colored granitic dikes which originate in the neighboring Lamarck granodiorite. The Striped pluton is composed of a fine-grained (1–3 mm) mafic granodiorite with a color index of 23 (fig. 16 and table 6).

On the west side, the rock of the Striped pluton grades into a migmatitic zone of mafic granodiorite and lighter colored quartz monzonite. On the south, the Striped pluton has sent out a large number of vertical dikes parallel to the foliation of the schistose wallrocks and formed an injection complex of biotite schist and mafic granodiorite.

The Striped pluton is not in contact with any older granitic rocks. However, lack of mafic dikes in it suggests that it is younger than the dikes. It is older than the Lamarck granodiorite (which has sent a swarm of gently dipping dikes into the Striped pluton) and the Siberian pluton (which cuts off the dikes associated with the Striped pluton).

PYRAMID PLUTON

The Pyramid pluton consists of two masses of a fairly dark colored granodiorite, in the northwest part of the map, included under a single pluton name because they are similar in composition and texture and because both are older than the pluton which separates them—the Arrow pluton. The pluton is named after Pyramid Peak in the center of the eastern mass. The East mass occupies 1.8 square miles and the West, 1.4 square miles within the quadrangle. The East mass lies between the Cotter and Arrow plutons, and the West mass lies west of the Cartridge Pass and Arrow plutons and is well exposed on Muro Blanco and Arrow Peak. It has not been traced far west of the quadrangle boundary.

Both masses are composed of a coarse-grained dark granodiorite (color index, 18.8) with large conspicuous hornblende crystals (fig. 17 and table 7). Mafic inclusions are abundant (4.6 per square yard) and rather large, commonly 6 inches to 1 foot in length. The northern part of the West mass is commonly somewhat migmatitic and fine-grained, and gneissose in places. At one spot a small septum of meta-

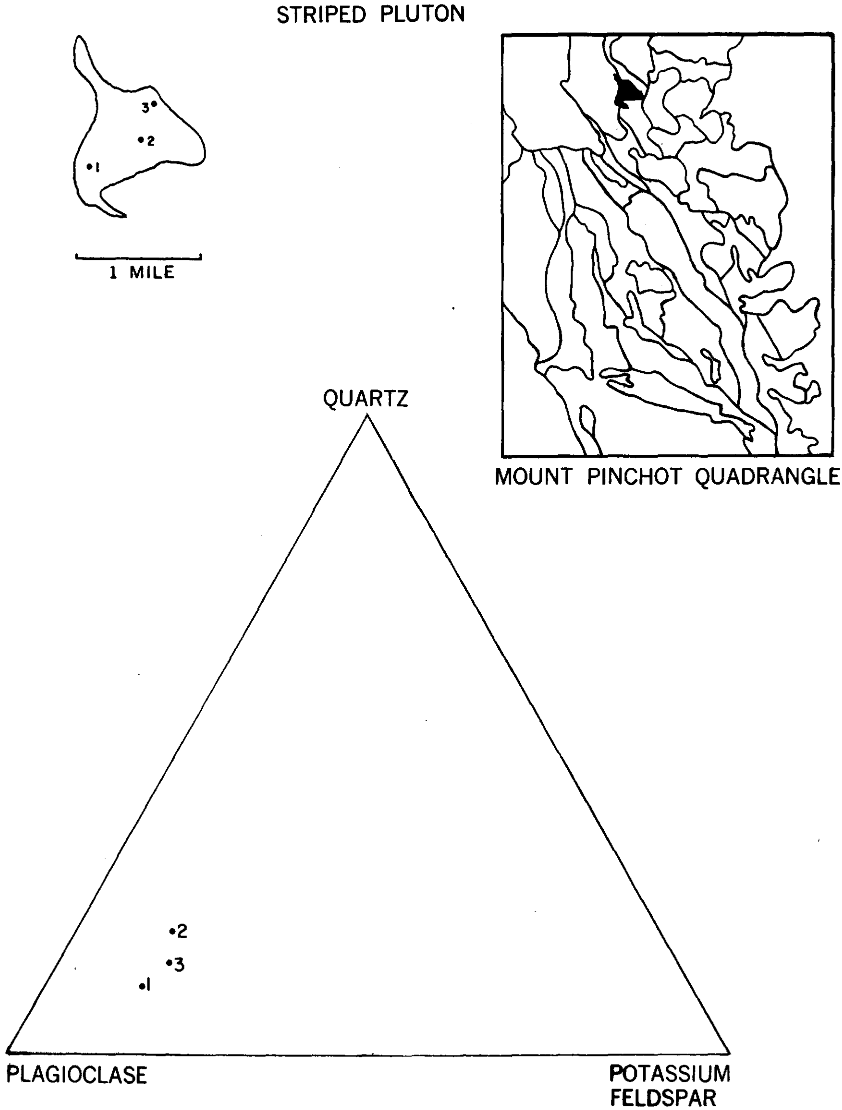


FIGURE 16.—Location and modal composition of samples of the Striped pluton. Numbers refer to samples given in table 6.

TABLE 6.—*Modes of Striped pluton*

Sample No.	Quartz	K-feld-spar	Plagio-clase	Biotite	Horn-blende	Pyroxene	Opaque accesso-ries	Non-opaque accesso-ries	Color index
1	9.4	10.7	60.7	11.6	5.7	Tr.	0.8	0.9	20.0
2	14.1	9.5	47.7	15.7	11.7	0.2	.4	.6	28.6
3	12.0	12.1	54.6	12.6	7.7	.3	.3	.4	21.3
Av	11.8	10.8	54.3	13.3	8.4	.2	.5	.6	23.0

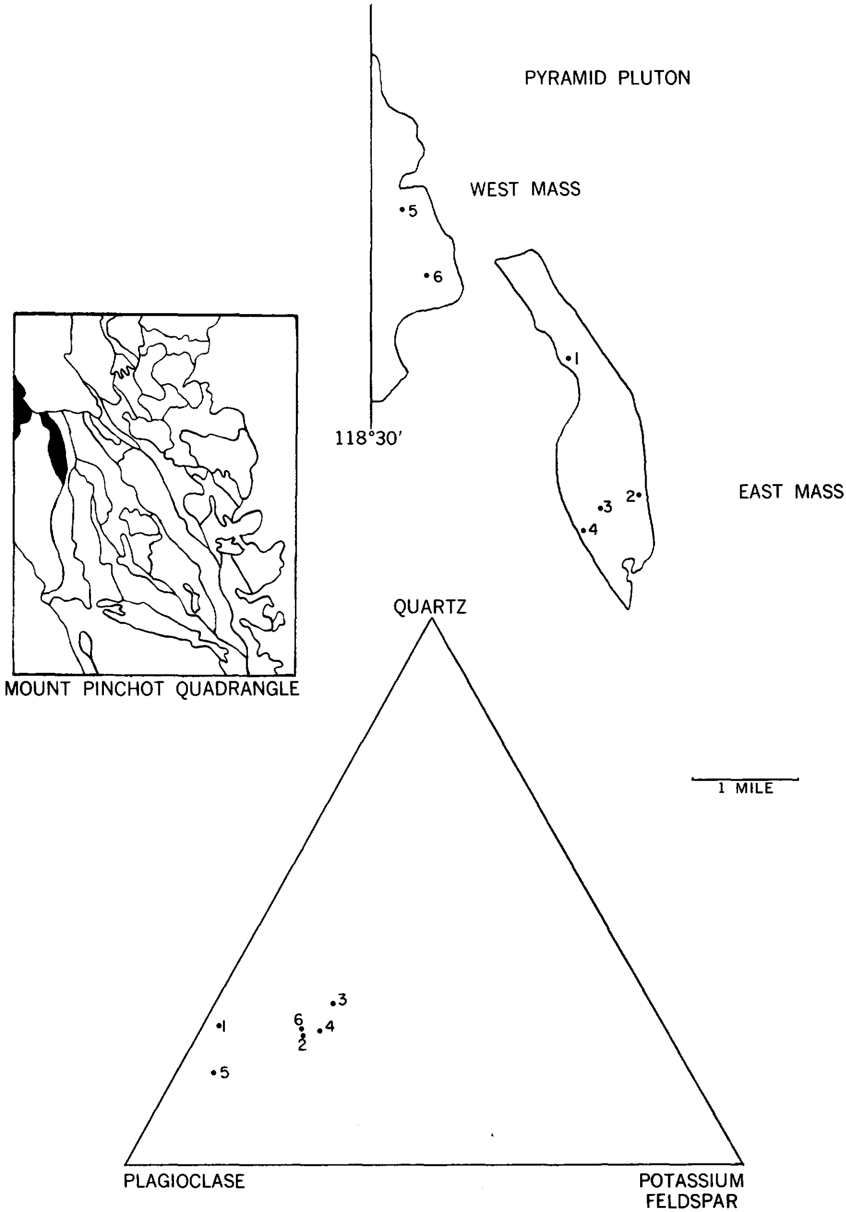


FIGURE 17.—Location and modal composition of samples from the Pyramid pluton. Numbers refer to samples given in table 7.

rhyolite was mapped. This is believed to be the southernmost extension of the large mass of metavolcanic rock to the northwest that centers in the Mount Goddard 15-minute quadrangle.

The Pyramid pluton is older than its walls. The East mass is intruded by the Arrow pluton on the west and the Cotter pluton on the

TABLE 7.—*Modes of Pyramid pluton*

Sample No.	Quartz	K-feld-spar	Plagio-clase	Biotite	Horn-blende	Pyrox-ene	Opaque Acces-sories	Non-opaque Acces-sories	Sp gr	Color index	Percent An in plagio-clase
1	20.6	1.3	57.5	12.6	6.4	----	0.9	0.7	2.77	20.6	----
2	20.3	14.2	51.2	5.4	7.7	----	.5	.7	----	14.3	----
3	24.2	15.7	43.7	7.2	7.8	----	.5	.3	----	15.8	32.5
4	21.0	16.2	47.9	10.1	4.2	----	.3	.2	2.69	14.8	----
5	11.6	3.9	53.2	18.9	8.4	1.5	2.1	.5	----	31.4	----
6	20.5	13.8	50.1	10.0	3.7	----	1.2	.5	2.71	15.4	----
Av	19.7	10.9	50.6	10.7	6.4	.3	.9	.5	----	18.8	----
SD [1]	4.2	6.5	4.7	4.7	2.0	.6	.7	.2	----	6.6	----

[1] Standard deviation.

east; it is truncated by the Cartridge Pass pluton to the north. The west mass is intruded by the Arrow and Cartridge Pass plutons.

McDOOGLE PLUTON

The McDoogle pluton is located in the central part of the quadrangle. The pluton is about 9 miles long and is relatively thin. It extends from Mount Wynne and Pinchot Pass on the northwest across Mount McDoogle (for which it was named) and the Sierra Nevada divide at Sawmill Pass to the North Fork of Oak Creek on the southeast. Mount McDoogle is a local name for the 12,372-foot peak north of Woods Lake.

The McDoogle pluton is composed of a dark granodiorite (color index, 23.6) which is commonly low in quartz (fig. 18 and table 8). Much of the pluton is composed of fairly homogeneous rock, yet areas of mixed and hybrid rocks are common. The northern part of the pluton from Pinchot Pass east to the summit of Mount Wynne is mixed with large amounts of more mafic plutonic rock. At Pinchot Pass the McDoogle pluton includes swarms of tabular dioritic inclusions, which in places are so abundant that they make up 50 percent of the rock.

The southern part of the pluton is intimately mixed with the mafic complex to the east. The mixed rock is represented separately on the map. South of Sawmill Lake, extensive reaction and mixing has

TABLE 8.—*Modes of McDoogle pluton*

Sample No.	Quartz	K-feld-spar	Plagio-clase	Biotite	Horn-blende	Pyrox-ene	Opaque acces-sories	Non-opaque acces-sories	Color index	Sp gr	Percent An in plagio-clase
1	13.6	20.6	43.5	11.3	10.4	0.1	0.1	0.8	22.7	2.72	31
2	7.1	13.2	56.8	11.3	9.8	.5	.7	.7	23.0	2.60	31
3	18.8	21.8	40.1	10.7	6.1	.6	1.1	.7	19.2	2.70	33
4	7.6	8.5	51.5	18.1	10.8	1.1	1.3	1.2	32.5	2.80	----
5	15.1	23.2	41.1	9.0	8.5	.9	1.1	1.2	20.7	2.69	29
Av	12.4	17.5	46.6	12.1	9.1	.6	.9	.9	23.6	2.74	31
SD	4.5	5.6	7.3	3.5	1.9	.6	.5	.3	5.2	----	----

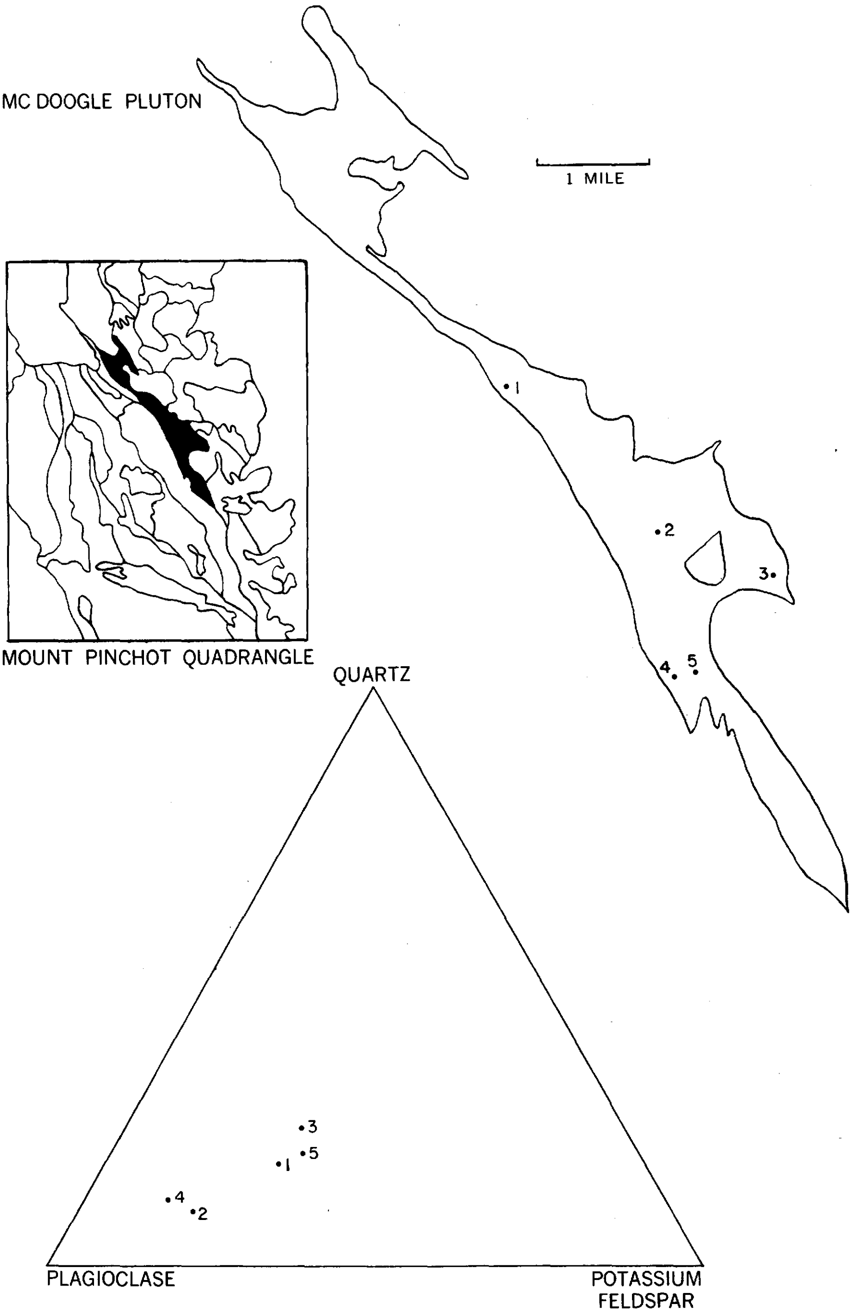


FIGURE 18.—Location and modal composition of samples from the McDoogle pluton. Numbers refer to samples given in table 8.

taken place and commonly the granitic rock of the pluton grades through a broad zone of mixed rock to diorite and gabbro.

The McDoogle pluton contains no mafic dikes and is younger than the dikes. Nevertheless, in many respects it resembles the predike plutons: it is commonly sheared and veined by epidote, it displays an irregular map pattern conforming to the shape of younger plutons; and it is commonly intimately mixed with wallrock. These characteristics suggest that the McDoogle pluton may be the oldest postdike mass. The McDoogle pluton is intruded by the Lamarck mass on the north and by the Colosseum mass, Spook pluton, and McGann plutons on the east.

DRAGON PLUTON

The Dragon pluton is in the south-central part of the quadrangle and extends southward beyond the map area. The pluton occupies 6.9 square miles within the quadrangle and is more than 11 miles long. The northern tip is near the junction of the North and South Forks of Woods Creek, and the pluton extends south across much of the drainage of the South Fork. The pluton is exposed at Dragon Lake (for which it was named), on Kearsarge Peak, and southeast of Onion Valley in Lime Canyon and Pinyon Creek.

The granitic rock of the Dragon pluton was not resistant to glacial action and it characteristically forms low cirques and lake basins. The Rae Lakes basins are excavated in the Dragon pluton.

The Dragon pluton wedges between the large Oak Creek metavolcanic pendant to the northeast and a dioritic complex to the southwest. It has consequently incorporated considerable mafic material from these wall rocks. The texture of the granodiorite of the Dragon pluton is characteristic: the mafic minerals appear to form a network or mesh structure, rather than discrete spots as is common with most of the other granitic rocks. Mafic inclusions are more shadowy and less distinct than in most other plutons.

Modal analyses show that the Dragon pluton is composed of a rock markedly different from most other plutons of similar appearance (fig. 19 and table 9). The average mode falls close to the quartz monzonite-granodiorite boundary. However, the rock is quite poor in quartz, averaging 12.7 percent. The low quartz content of the rock is likely a result of assimilation of mafic volcanic wallrock on the one side and dioritic rock on the other side. Evidence for wallrock contamination was seen in the broad mixed zone on the northeast side of the Dragon pluton. This complex is composed of large chunks of mafic granodiorite and diorite, commonly elongated in a horizontal direction, in a matrix of granodiorite of the Dragon pluton. Some of the individual mafic masses are several hundred feet in length.

Lineation is better developed in the Dragon pluton than in any other intrusive mass. Most of the mafic inclusions are rod shaped; they are nearly horizontal and bear northwest, parallel with the elon-

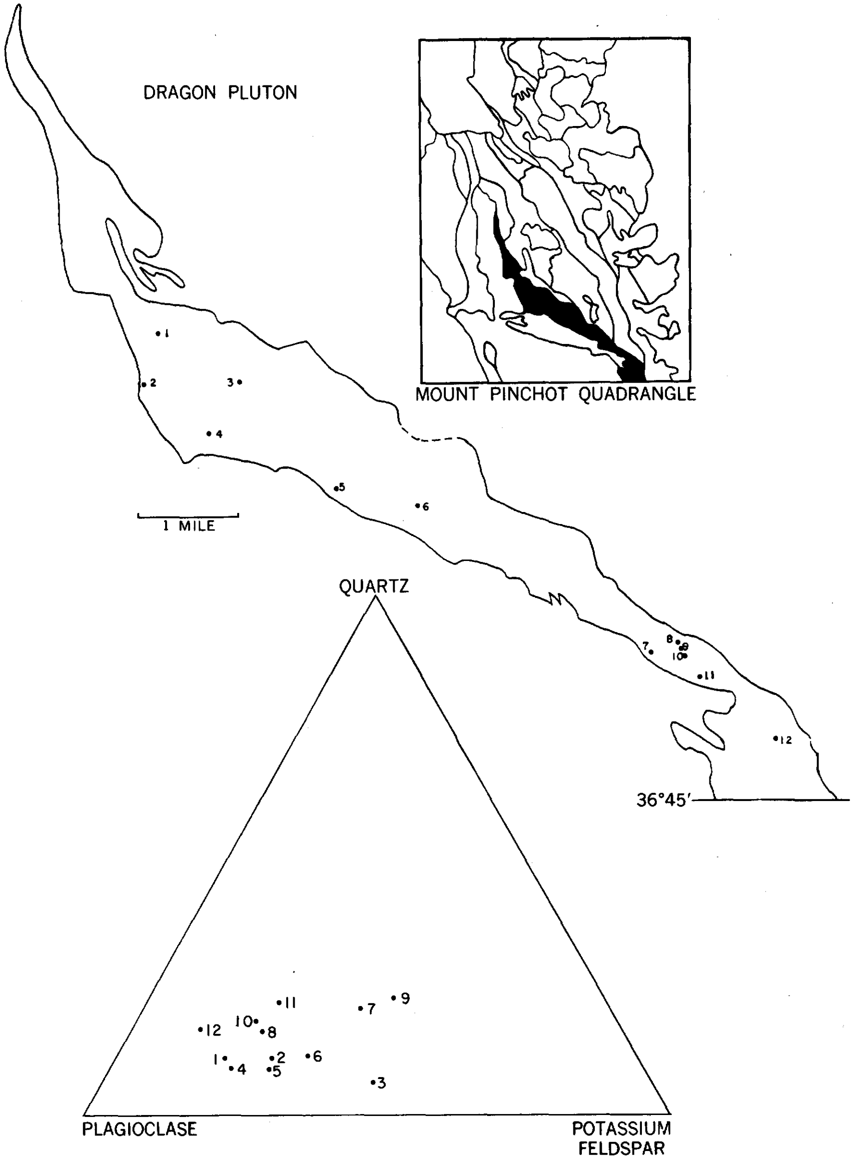


FIGURE 19.—Location and modal composition of samples from the Dragon pluton. Numbers refer to samples given in table 9.

gation of the pluton. In the center of the pluton a nearly horizontal planar structure is developed. This horizontal foliation is believed to be parallel with the roof of the intrusive mass and suggests that the pluton is not deeply eroded. The horizontal, northwest lineation suggests northwest-southeast arching and stretching over the top of the pluton.

TABLE 9.—*Modes of Dragon pluton*

Sample No.	Quartz	K-feldspar	Plagioclase	Biotite	Hornblende	Pyroxene	Opaque accessories	Nonopaque accessories	Color index	Sp gr	Percent An in plagioclase
1	9.9	16.5	62.6	7.2	1.7	-------	1.2	0.9	11.0	2.67	25
2	9.1	26.7	60.6	2.5	.1	-------	.7	.3	3.6	2.65	-------
3	6.4	43.0	44.5	4.5	.7	-------	.5	.4	6.1	2.65	-------
4	8.2	18.0	62.2	9.6	-------	-------	.8	.8	11.2	2.66	-------
5	8.4	24.8	58.5	5.8	.4	-------	1.5	.5	8.2	2.65	-------
6	9.8	28.2	49.8	7.7	.8	-------	2.4	1.2	12.1	2.65	21
7	19.4	34.8	39.7	4.7	-------	-------	.8	.5	6.0	2.63	15
8	14.3	18.5	52.3	7.2	5.4	0.2	1.5	.7	15.0	-------	-------
9	20.4	38.9	32.2	7.6	-------	-------	.7	.2	8.5	2.64	28
10	15.1	17.3	51.3	9.5	3.8	-------	1.3	1.7	16.3	2.66	-------
11	17.2	17.4	44.3	12.2	5.7	-------	1.8	1.4	21.1	2.70	28
12	14.6	9.6	59.8	9.0	4.7	.2	1.2	.9	16.0	2.73	-------
Av	12.7	24.5	51.5	7.3	1.9	Tr.	1.2	.8	11.2	2.66	23
SD	4.7	10.2	9.8	2.7	2.3	-------	.5	.5	5.1	-------	-------

The Bullfrog pluton is the only one of the intrusives in contact with the Dragon pluton that appears to be younger; it truncates the structure of the Dragon pluton. The White Fork and Diamond plutons on the northeast and the Tinemaha granodiorite on the southeast are all cut by mafic dikes of the Independence swarm and are presumed older than the Dragon pluton, which is not cut by mafic dikes. The contact of the Diamond and the Dragon plutons east of lower Rae Lakes further demonstrates that the Dragon pluton is the younger; (1) inclusions of alaskite of the Diamond pluton are present in the granodiorite of the Dragon pluton, (2) the Dragon pluton becomes finer grained and develops parallel schlieren toward the contact, (3) the Diamond pluton appears unaffected toward the contact, (4) in general, the rock of the Diamond pluton appears more sheared and altered, and contains abundant epidote-filled fissures.

SILICIC GRANODIORITE PLUTONS

Many of the plutons of the quadrangle are composed of a granodiorite that averages about twice as much plagioclase as K-feldspar. The rock therefore plots close to the boundary between the granodiorite and quartz monzonite fields on the classification used (fig. 10). Hence these rocks are designated silicic granodiorite for convenience in discussing them. Nine plutons are included in this composition group. Of these 9, 6 are porphyritic: Lamarck and Tinemaha granodiorites and Spook, Cartridge Pass, Paradise, and Mule Lake plutons. The composition of these rocks is believed to have favored growth of phenocrysts of K-feldspar. There are no porphyritic rocks in any of the other composition groups.

The silicic granodiorite masses cover a broad age span. Three of them are cut by mafic dikes: White Fork and Mule Lake plutons and Tinemaha granodiorite. Of these three, the White Fork pluton may be the oldest in the quadrangle. Six masses are younger than the mafic dikes: Cartridge Pass, Arrow, Spook, Cotter, and Paradise plutons and the Lamarck granodiorite.

LAMARCK GRANODIORITE

The main mass of the Lamarck granodiorite extends south from the north border of the quadrangle for 5 miles across Upper Basin and terminates within a few hundred feet of Pinchot Pass, on the drainage divide between Woods Creek and the South Fork of the Kings River. This portion occupies 7.0 square miles, and is here called the Lamarck mass of the Lamarck granodiorite to distinguish it from a correlated mass to the southeast. The Lamarck mass is slightly porphyritic over much of its extent within the Mount Pinchot quadrangle. North of the quadrangle, the mass becomes considerably darker and more mafic. It has been traced by Bateman (1961) continuously for 16 miles northwest to Mount Lamarck, for which it is named, in the Mount Goddard quadrangle.

One-half mile southeast of the southern end of the main Lamarck mass, in the cirque bottom on the southwest side of Mount Wynne, the top of a small pluton is exposed. This pluton extends southeast 3 miles beyond Colosseum Mountain and covers an area of 2.6 square miles. Because this pluton is directly on strike with the Lamarck granodiorite to the northwest, because it is separated by only one-half mile from the Lamarck mass and the tops of both masses dip toward one another, and because it is similar to the Lamarck granodiorite in quantitative petrology (fig. 20 and table 10), this pluton is tentatively correlated with the Lamarck granodiorite and is believed to be connected with it below the surface. This small pluton is called the Colosseum mass of the Lamarck granodiorite, after Colosseum Mountain on the main Sierra divide. The roof of the Lamarck granodiorite at Pinchot Pass dips southeast and must be very thin, probably not more than a few hundred feet thick.

The Lamarck mass is in contact with hybridized and mafic granodiorite of the McDoogle pluton at its southern end on the north side of Pinchot Pass. Here the Lamarck mass shows a foliation parallel to the contact and sends out many flat dikes into the McDoogle pluton, which are well exposed on Mount Wynne.

The Lamarck mass also sends out hundreds of near-horizontal granitic dikes into the older Striped pluton bordering it on the east. These dikes contrast vividly with the darker rock of Striped Mountain (fig. 21).

The Colosseum mass of the Lamarck granodiorite is intrusive into the McDoogle pluton on the west and the Spook pluton on the east. The north end of this mass, where it is exposed south of Pinchot Pass,

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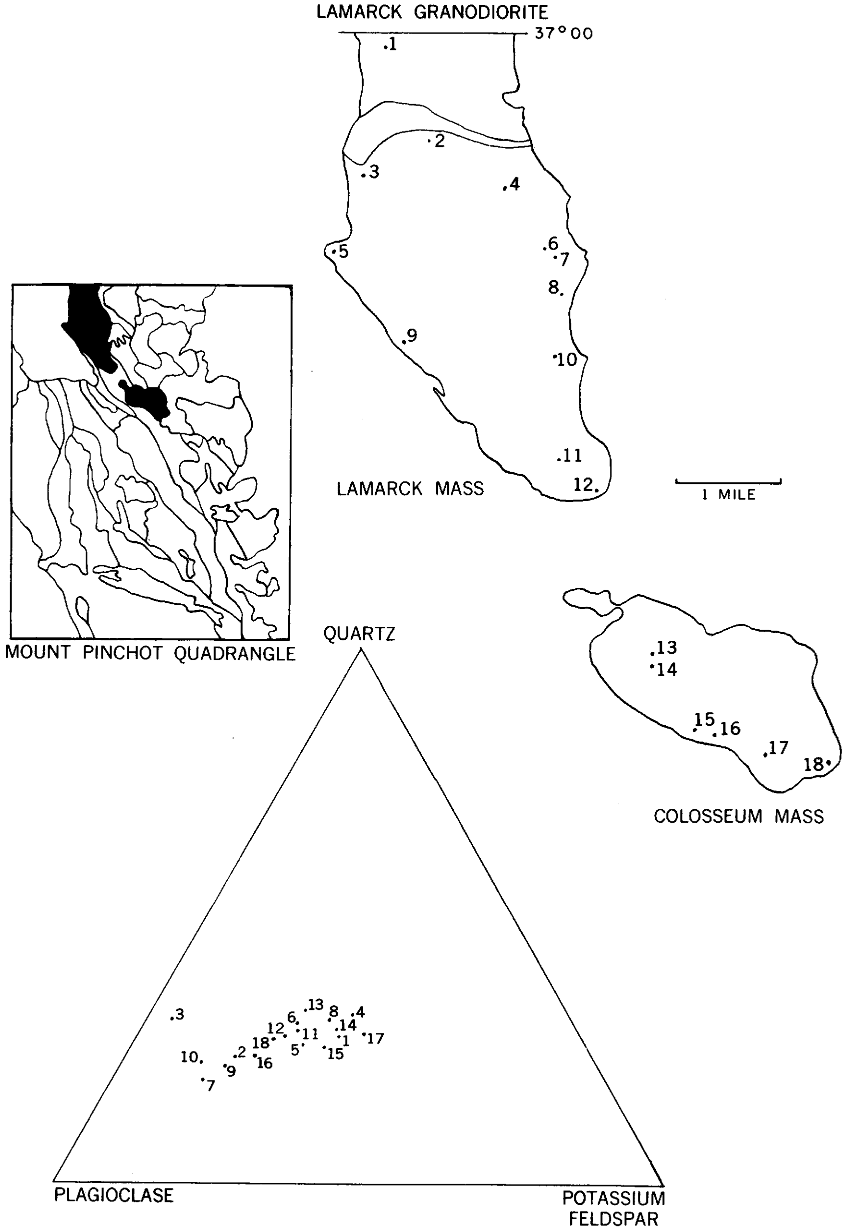

FIGURE 20.—Location and modal composition of samples from the Lamarck granodiorite. Numbers refer to samples given in table 10.

has a nearly horizontal top. It is finer grained than the rock of McDoogle pluton and has a horizontal layering parallel to the contact. Apophyses of Lamarck granodiorite cut the dark granodiorite of the McDoogle pluton (fig. 22).

TABLE 10.—*Modes of the Lamarck granodiorite*

Sample No.	Quartz	K-feld-spar	Plagio-clase	Biotite	Horn-blende	Opaque acces-sories	Non-opaque acces-sories	Color index	Sp gr	Percent An in plagio-clase
Lamarck mass										
1	25.3	30.3	37.1	5.3	1.0	0.7	0.4	7.4	2.63	
2	22.5	16.8	56.3	3.0	.5	.6	.3	4.4	2.66	
3	27.9	3.7	58.8	5.6	3.1	.7	.2	9.6	2.69	30
4	29.7	31.0	34.6	4.0		.6	.1	4.7	2.66	
5	24.7	25.0	42.6	5.1	.7	1.3	.5	7.6	2.68	
6	25.9	23.9	43.1	5.6		.5	.3	6.4	2.64	29
7	17.5	12.5	58.8	8.7	1.2	.5	.6	11.0	2.67	
8	29.0	27.9	37.8	3.6	.2	1.4	.2	5.4		28
9	20.0	14.7	55.3	7.1	.7	.8	1.5	10.1	2.67	29
10	20.7	10.9	58.4	7.0	.6	1.0	1.2	9.8	2.66	
11	27.1	23.6	43.0	5.5		.5	.3	6.3	2.60	
12	25.8	22.4	45.2	4.9	.1	.9	.8	6.7		
Av	24.8	20.3	47.6	5.4	.7	.8	.5	7.4	2.66	29
SD	3.8	8.6	9.3	1.6	.9	.3	.4	2.2		
Colosseum mass										
13	30.4	23.2	40.0	5.1	0.4	0.4	0.4	6.3	2.62	31
14	27.0	30.5	37.6	3.7	.3	.6	.3	4.9	2.69	29
15	24.7	29.8	40.9	3.6	1.0	.2		4.8	2.61	
16	21.2	18.3	49.3	8.2	1.4	.7	.8	11.1	2.62	
17	26.4	36.2	33.6	3.9		.6	.3	4.8	2.62	
18	25.1	20.7	47.4	5.9	.1	.6	.2	6.8	2.65	
Av	25.8	26.3	41.5	5.1	.5	.5	.3	6.4	2.62	30
SD	3.1	6.5	5.9	1.8	.6	.2	.3	2.4		

The contact between the Colosseum mass and the Spook pluton is exposed in the remote cirques south and east of Colosseum Mountain, but the contact is difficult to locate because the two rock types are similar in appearance. The older age of the Spook pluton is shown by the following relations: (1) the contact transects aplite dikes and mafic inclusions in the Spook pluton; (2) the Colosseum mass sends several tongues into the Spook pluton; (3) the Colosseum mass is very slightly banded and finer grained at the contact, whereas no appreciable difference was noted in the Spook pluton at the contact. A remarkable feature of this contact is that it dips from 40° to 65° N. under the younger Colosseum mass. In nearly every other contact observed, the younger pluton is on the footwall of the intrusive contact.

Specific gravity measurements made on samples from the Lamarck mass suggest that the mass is zoned, and that the central part has a lower density than the margins (fig. 23). The specific gravity is 2.66 or more in the margins and grades to less than 2.64 in the center.

CARTRIDGE PASS PLUTON

The Cartridge Pass pluton is located in the extreme northwestern corner of the map and extends some distance into the three adjacent quadrangles: Marion Peak, Mount Goddard, and Big Pine. All the

FIGURE 21.—Striped Mountain from the east. The dark granodiorite of the Striped pluton which makes up the bulk of the mountain, is intruded by a large number of subhorizontal dikes originating from the Lamarck granodiorite behind the mountain.

borders of the pluton have not been mapped outside the quadrangle, but it is believed that the greatest part of the pluton is within the Mount Pinchot quadrangle, where 14.1 square miles are exposed.

The pluton takes its name from Cartridge Pass (elevation 12,140 feet) on Cirque Crest, the drainage divide between the South Fork of the Kings River and Cartridge Creek on the Middle Fork of the Kings River. The pluton is excellently exposed over nearly all of its extent, except for the bottom of the canyon of the South Fork of the Kings River, which cuts across the middle. Exceptional exposures can be studied in upper Lake Basin, Upper Basin, and on the bench occupied by Bench Lake. The overall average rock of the Cartridge Pass pluton is a granodiorite comprising 23.4 percent quartz, 18.3 percent K-feldspar, and 47.8 percent plagioclase (table 11). However, many samples plot as quartz monzonite (fig. 24).

The Cartridge Pass pluton was recognized to be compositionally zoned in the field. The rock in the core of the pluton is light gray and has a color index of 6. Toward the margins it is progressively darker in color and richer in mafic minerals. In nearly every traverse made this change was observed to be completely gradational. However, in one area on the southeast flank of Mount Ruskin the inner, light-

TABLE 11.—*Modes of Cartridge Pass pluton*

Sample No.	Quartz	K-feld-spar	Plagio-clase	Biotite	Horn-blende	Opaque acces-sories	Non-opaque acces-sories	Color index	Sp gr	Pecent An in plagio-clase
1	20.5	32.9	40.2	4.4	0.4	0.9	0.8	6.5	2.65	30
2	20.4	19.6	48.7	8.7	1.3	.9	.5	11.4	2.66	32
3	26.9	20.3	47.5	4.2	.1	.4	.5	5.2	2.65	30
4	24.4	18.4	48.0	5.6	1.2	.8	.7	9.3	2.64	30.5
5	20.6	17.6	53.0	5.3	1.6	1.4	.6	8.9	2.68 9	32
6	25.4	16.3	47.1	8.2	2.1	.8	.2	11.3	2.67	31.5
7	17.5	19.4	47.3	9.2	3.8	1.6	1.1	15.7	2.6	31
8	----	----	----	----	----	----	----	----	----	32.5
9	22.9	23.2	42.6	7.6	2.5	.7	.4	11.2	2.68	32
10	23.8	10.8	54.1	8.2	2.0	.5	.6	11.3	2.70	35
11	26.7	20.2	46.7	5.3	.2	.7	.1	6.3	2.66	31
12	26.6	19.4	41.2	8.0	3.5	.4	.8	12.7	2.67	30.5
13	20.3	18.5	50.6	7.1	1.0	.7	1.0	10.7	2.69	31
14	11.9	7.2	56.5	12.7	9.8	1.4	.4	24.3	2.77	34.5
15	23.7	14.1	45.2	10.7	4.9	1.3	.2	17.1	2.68	33
16	27.5	12.0	51.7	7.1	.5	.8	.4	8.8	2.67	32
17	----	----	----	----	----	----	----	----	2.63	32.5
18	23.1	23.4	47.5	5.3	.1	.4	.1	5.9	2.65	31.5
19	23.8	19.4	47.3	6.7	2.0	.5	.2	9.4	2.67	31.5
20	21.0	14.1	47.6	11.5	4.3	1.2	.2	17.2	2.71	32.5
21	21.2	20.2	44.2	8.8	4.6	.4	.6	14.4	2.70	33
22	23.4	4.1	52.4	11.3	7.6	.6	.5	20.0	2.78	34
23	23.6	20.2	46.7	6.8	.9	1.2	.7	9.6	2.64	30
24	22.3	23.4	46.7	5.9	.2	.7	.7	7.5	2.64	32
25	24.6	22.7	46.9	4.6	.4	.6	.2	5.8	2.66	30
26	26.2	19.3	48.5	5.2	.1	.2	.5	6.0	2.64	29
27	25.9	22.0	46.7	4.6	----	.6	.2	5.4	2.64	29
28	21.5	23.1	50.0	4.6	.2	.6	.1	5.5	2.66	26.5
29	25.8	22.5	46.3	4.4	.1	.5	.5	5.5	2.63	25
30	22.1	21.9	50.3	4.7	.1	.7	.1	5.6	2.65 2	28
31	20.8	24.0	49.7	4.3	----	.7	.5	5.5	2.63	29
32	24.0	23.8	44.4	6.8	----	.8	.2	7.8	2.61	30
33	18.7	9.6	51.9	10.1	7.8	1.5	.4	19.8	2.72	34
34	19.2	10.2	54.8	8.9	5.2	1.3	.4	15.8	2.7	33.5
35	26.0	15.1	46.5	8.2	2.4	.9	1.0	12.5	2.66	30
36	21.7	14.2	48.4	11.6	2.6	1.0	.6	15.8	2.68	30.5
37	24.1	19.9	47.9	5.2	1.6	.7	.7	8.2	2.66	----
38	26.9	22.6	42.3	6.1	.7	.8	.8	8.4	2.67	31.5
39	27.5	22.9	43.7	4.5	.1	.9	.5	6.0	2.63	29.5
40	22.3	23.0	48.5	4.8	.3	.6	.5	6.2	2.62	29.5
41	20.6	23.8	49.1	5.0	.2	1.0	.4	6.6	2.64	27.5
42	28.1	30.3	38.8	2.3	.1	.1	.3	2.8	2.61	29
43	22.7	18.1	52.2	6.7	.3	.2	.8	7.0	2.65	27
44	26.4	18.5	47.4	5.8	.1	1.6	.2	7.7	2.66	28.5
45	23.4	22.3	48.3	4.4	.1	.8	.6	5.9	2.61	----
46	----	----	----	----	----	----	----	----	2.62	30
47	----	----	----	----	----	----	----	----	2.64	30
48	----	----	----	----	----	----	----	----	2.63	28.5
49	29.0	18.2	45.6	5.9	.1	.8	.4	7.2	2.66	----
50	29.1	16.9	47.8	4.7	----	.8	.6	6.1	2.62	29
Av	23.4	18.3	47.8	6.7	1.7	.8	.5	9.7	----	----
SD	3.3	6.8	3.7	2.3	2.4	.4	.2	4.7	----	----

colored granitic rock is intrusive into the outer, dark-colored border facies. The light-colored granitic rock contains schlieren of mafic material along and parallel to the contact, whereas the foliation of the dark-colored facies is truncated by the contact. However, to the north this intraplution contact becomes increasingly diffuse and completely disappears, and the outer and inner facies are gradational with one another. Such relations suggest that the central, more silicic, part

FIGURE 22.—Contact (*A–A'*) between the Colosseum mass of the Lamarck granodiorite (lower, light-colored) and the McDoogle pluton (upper, dark-colored). The Lamarck granodiorite has a near-horizontal layering parallel to the contact and sends apophyses into the older granodiorite of the McDoogle pluton. Hammer (arrow) is on the contact. Three-fourths mile southeast of Pinchot Pass on the John Muir Trail.

of the pluton was capable of local injection and reintrusion of its walls after its formation.

Excellent exposures and wide range in zoning afford an unparalleled opportunity to study a zoned pluton. Accordingly, samples were collected for study of variation in mineral content. Modal analyses (table 11) show that quartz and K-feldspar are increasingly abundant toward the center of the pluton, quartz increasing from less than 20 percent in the margins of the pluton to more than 26 percent in the center (fig. 25), and K-feldspar from less than 14 percent in the margins to more than 23 percent in the center. It was noticed in the field that the crystals of K-feldspar appear to become larger and porphyritic toward the core of the pluton. The porphyritic phase is best developed on the eastern tip of Bench Lake, where the K-feldspar phenocrysts average about 2 cm in length and are spaced approximately 15 cm apart. Over much of the center part of the pluton (roughly within the 9 color index contour, fig. 25) the phenocrysts average about 1 cm in length and are spaced roughly 25 cm apart.

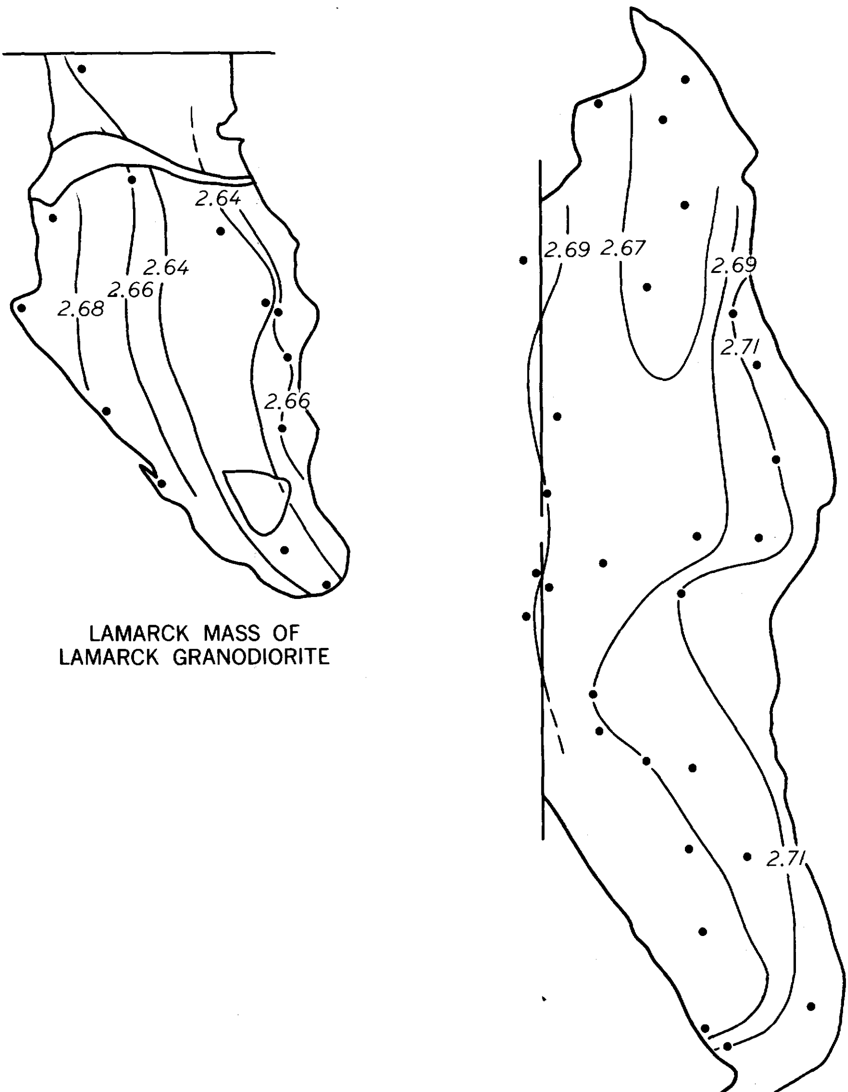

FIGURE 23.—Zonation as indicated by specific gravity measurements in the Lamarck mass and Arrow pluton. Dots indicate location of measured samples.

Along the margins of the pluton all traces of a porphyritic texture disappear.

Microscopic examination of the K-feldspar discloses that twinned and untwinned varieties are both present, commonly within the same crystal. The characteristic grid twinning is best developed along lines of shear or strain within the individual K-feldspar crystals.

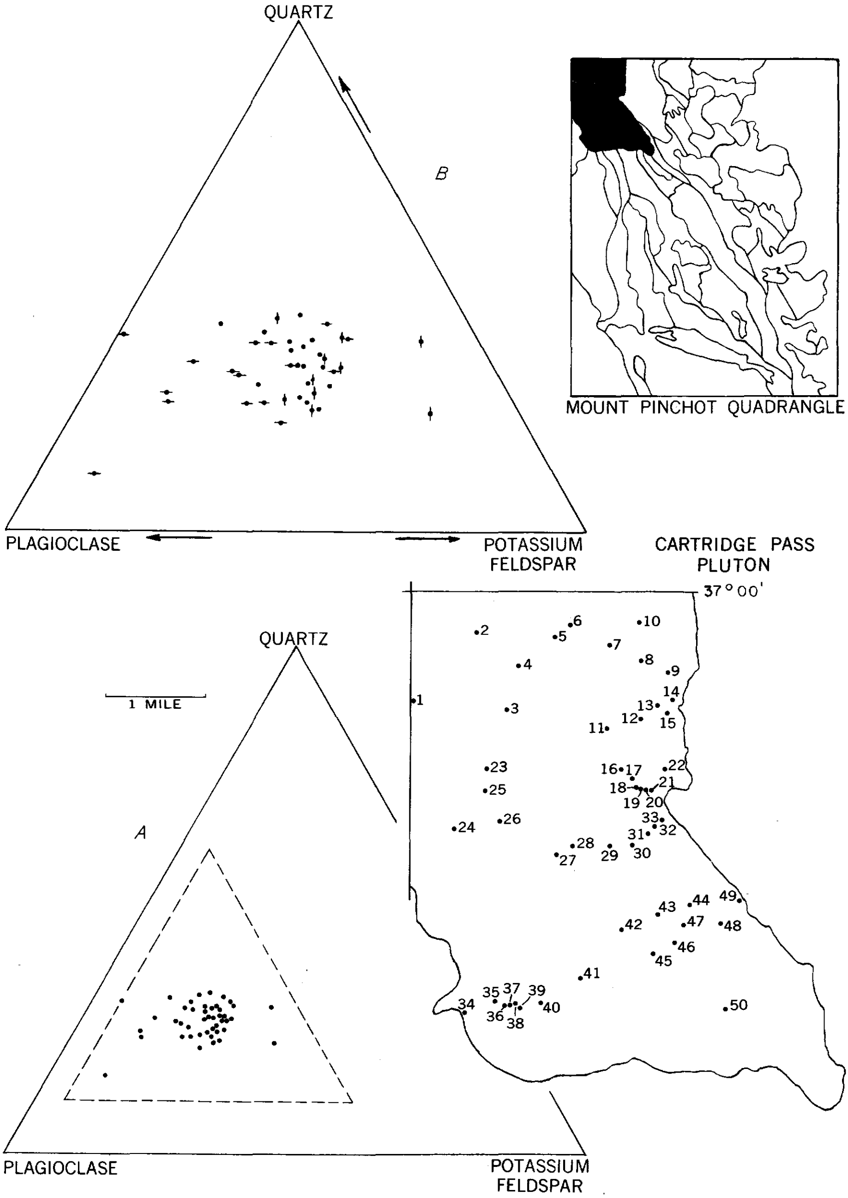

FIGURE 24.—Location and modal composition of samples of the Cartridge Pass pluton; *B* shows a magnified portion of the small triangle shown in *A*. Points with horizontal lines are those with specific gravity equal to or greater than 2.67, vertical lines, equal to or less than 2.63, all other points, between 2.67 and 2.63. Numbers refer to samples given in table 11.

Plagioclase decreases in abundance toward the core of the pluton, and becomes more sodic. The amount of plagioclase decreases from more than 53 percent at the margin to less than 44 percent in the

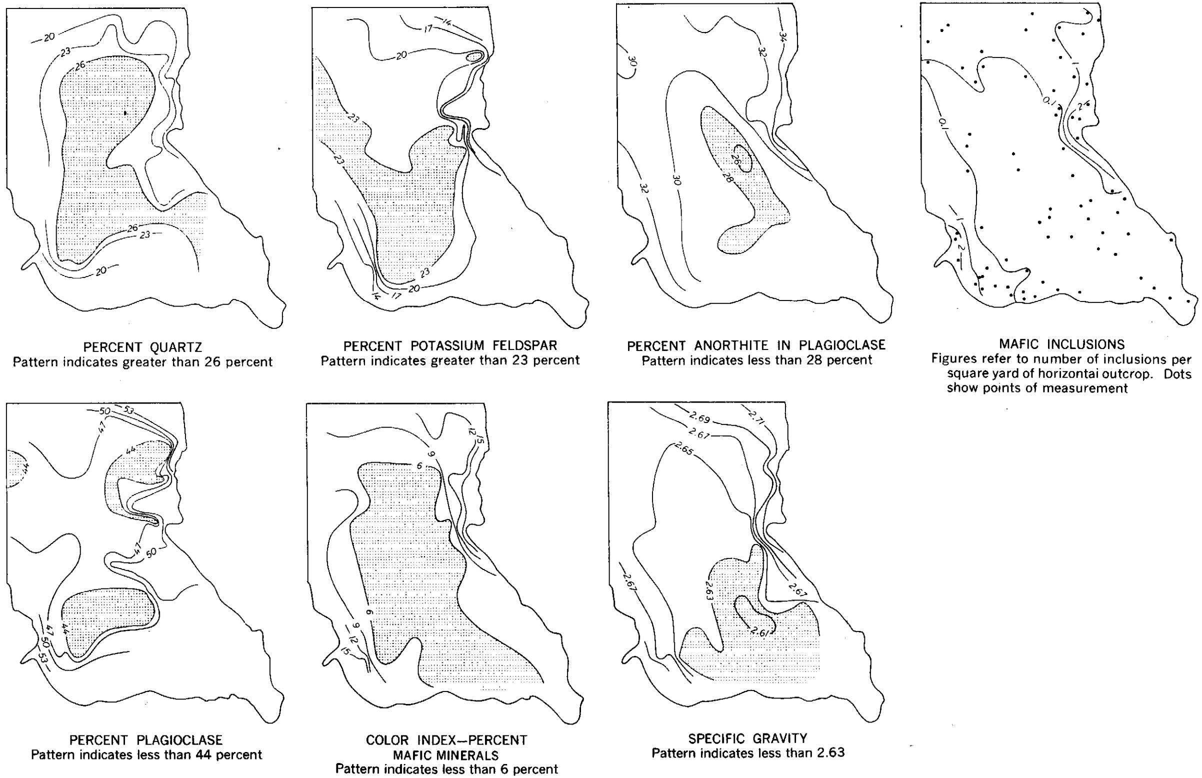

Figure 25.—Details in the zoning of the Cartridge Pass pluton.

core (fig. 25). The plagioclase shows strong oscillatory zoning, but average values of anorthite content determined by powder X-ray diffraction using the data of Smith and Yoder (1956) decrease from more than 34 percent anorthite in samples from the margin to less than 26 percent anorthite in samples from the center of the pluton.

The specific gravity measurements were made on large hand specimens and show the details of zoning more precisely than do the modes, which are each based on a single thin section. The area of lowest specific gravity is not in the center of the pluton, but lies south of the center (fig. 25).

Measurements were made in the field on the abundance of the typical ellipsoidal mafic inclusions that are so common in Sierran granitic rocks. The tabular inclusions in the Cartridge Pass pluton are 3 to 5 inches long and about half as wide. They are uniformly distributed over the area of an outcrop. Near the borders of the pluton they are more abundant than 2 per square yard of horizontal outcrop. Toward the center of the pluton they become smaller and less abundant, and in a broad central zone they are extremely scarce (fig. 25). Since the mafic inclusions are about the only element in most of the pluton that is capable of recording movement, nearly all structural readings were taken on their direction of elongation. In the center of the pluton, where inclusions are absent, no structure is evident.

The Cartridge Pass pluton may be the youngest intrusion in the quadrangle. It intrudes the granodiorite of the west mass of the Pyramid pluton, as shown by the following evidence: (1) In a marginal zone several hundred yards wide the Cartridge Pass pluton is progressively finer grained and richer in hornblende and in mafic inclusions toward the contact. (2) It incloses fragments of the distinctive coarse hornblende granodiorite of the Pyramid pluton. (3) The map pattern suggests that the west mass of the Pyramid pluton is transected by the Cartridge Pass pluton.

The contact between the Cartridge Pass pluton and Lamarck granodiorite is very well exposed in Upper Basin a few hundred feet east of the John Muir Trail, near the north boundary of the quadrangle (fig. 26). The younger age of the Cartridge Pass pluton is shown by the following relations: (1) The contact cuts across the foliation of the Lamarck granodiorite. (2) The contact is parallel to a planar alinement of mafic inclusions in the Cartridge Pass pluton. (3) The Cartridge Pass pluton has a dark fine-grained border facies several hundred feet thick parallel to the contact. (4) Small dikes and apophyses from the Cartridge Pass pluton cut the Lamarck granodiorite. (5) Large inclusions (as much as 10 feet long) of Lamarck granodiorite are present in the Cartridge Pass pluton, several hundred feet from the contact.

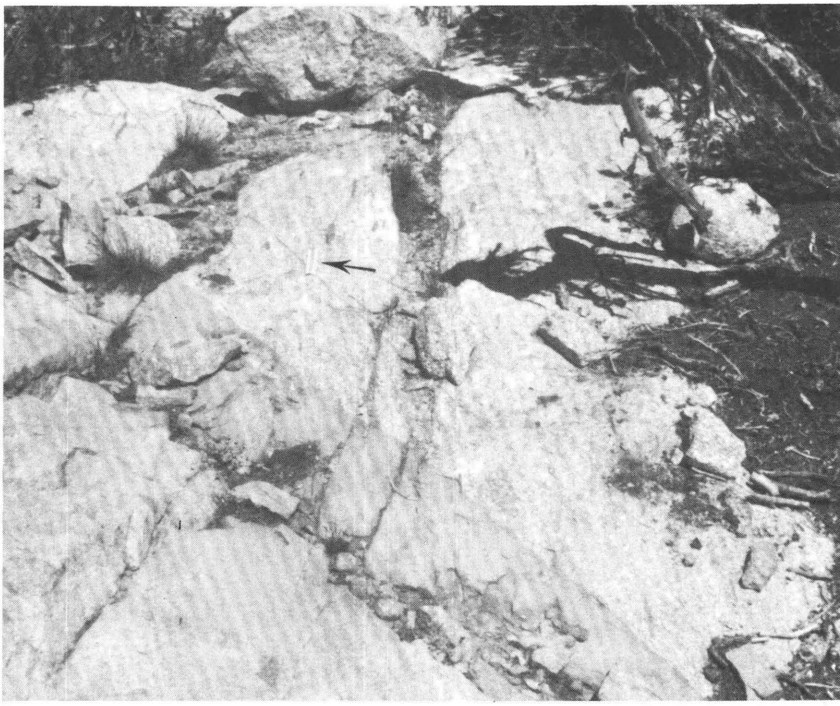

FIGURE 26.—Contact between dark border facies of Cartridge Pass pluton (left) and older main mass of the Lamarck granodiorite (right). The rock of the Cartridge Pass pluton has an indistinct foliation parallel to the contact, and the contact transects a foliation in the Lamarck granodiorite. Six-inch scale (arrow) is parallel to the trace of the foliation in the Lamarck granodiorite and lies on a small dike of granodiorite of the Cartridge Pass pluton cutting the Lamarck granodiorite. Few hundred feet east of John Muir Trail at north quadrangle boundary.

On the south, the Cartridge Pass pluton is in contact with the Arrow, Pyramid, Cotter, White Fork, and Twin Lake plutons. The map pattern suggests that the Cartridge Pass pluton cuts across all these plutons, and this concept is verified by field relations. A swarm of flat-dipping granitic dikes, satellitic to the Cartridge Pass pluton, intrude the Cotter pluton 1 mile south of Bench Lake.

The Cartridge Pass–Pyramid contact and the Cartridge Pass–Arrow contact were not studied in the field, but both the Pyramid and the Arrow plutons are known to be older than the Cotter pluton, and therefore older than the Cartridge Pass pluton.

The Cartridge Pass pluton cuts off mafic dikes of the Independence dike swarm. This relation can be seen along the Cartridge Pass–White Fork contact 1 mile south-southwest of Bench Lake, in a flat-floored glacial cirque. Along the contact, five mafic dikes in the White Fork pluton were observed to be truncated by the granodiorite of the Cartridge Pass pluton. Since the White Fork and Twin Lake plutons are both cut by mafic dikes, they must both be older than the Cartridge Pass pluton. Further support of this age relation are the truncation of the strong gneissic foliation of the White Fork pluton at the Cartridge Pass contact and the presence of near-horizontal graitic dikes satellitic to the Cartridge Pass in the White Fork and Twin Lake plutons. These granitic dikes cut the mafic dikes in the older plutons.

ARROW PLUTON

The Arrow pluton occupies 14.4 square miles in the west-central part of the quadrangle and extends into the neighboring Marion Peak quadrangle. The pluton is elongate in a northerly direction and extends for a distance of 9 miles, from near Arrow Peak on the north across Woods Creek to Gardiner Basin on the south. The pluton takes its name from Arrow Creek, which lies in the northern part of the pluton.

The average rock of this pluton is a granodiorite, yet many samples are quartz monzonite (fig. 27 and table 12). In the field the Arrow is distinguished from the neighboring Cotter pluton by slightly coarser grain and more prominent hornblende crystals, and by fewer mafic inclusions: Cotter, 5.4 per square yard; Arrow, 2.0 per square yard.

TABLE 12.—*Modes of Arrow pluton*

Sample No.	Quartz	K-feldspar	Plagioclase	Biotite	Hornblende	Pyroxene	Opaque accessories	Non-opaque accessories	Color index	Sp gr	Percent An in plagioclase
1	25.7	19.8	43.2	6.1	3.3	--------	0.6	0.4	10.4	2.66	33
2	19.0	26.2	38.5	8.8	4.8	--------	1.3	1.1	16.0	2.66	32
3	23.7	23.2	40.9	7.4	2.8	--------	1.4	.6	12.2	--------	33
4	26.6	11.0	48.2	8.9	2.5	--------	1.7	.9	14.0	2.71	36
5	19.9	5.9	55.3	11.2	6.3	0.5	.6	.3	18.6	2.73	--------
6	27.0	14.8	44.5	9.4	2.7	--------	.8	.7	13.6	2.69	--------
7	22.8	15.6	47.6	8.7	3.2	--------	1.3	.8	14.0	2.69	--------
8	27.9	16.6	42.8	8.8	3.1	--------	.4	.4	12.7	2.69	--------
9	25.6	14.3	49.9	6.6	2.1	--------	.6	.8	10.1	2.68	--------
10	24.2	26.8	37.1	7.6	2.6	Tr.	.7	.8	11.7	2.68	--------
11	22.7	16.9	43.5	11.7	4.3	.1	.5	.3	16.9	2.70	--------
12	23.7	14.1	48.1	8.1	3.9	--------	1.1	1.0	14.1	2.71	--------
13	27.8	17.5	44.4	6.4	2.0	--------	.9	1.2	10.5	2.68	--------
14	16.0	19.8	51.0	6.5	4.5	--------	1.9	.3	13.2	2.69	33
15	23.6	16.7	47.3	6.9	3.1	--------	1.3	1 0	12.3	2.67	32
16	22.2	21.8	46.8	5.6	2.3	--------	.7	.6	9.2	2.70	--------
17	22.8	19.4	43.2	7.4	5.7	--------	1.1	.3	14.5	2.68	--------
18	14.4	10.9	60.4	7.1	6.1	.1	.7	.4	14.4	2.75	--------
19	26.4	10.8	48.5	8.5	4.6	--------	.7	.4	14.2	2.71	--------
Av	23.3	17.0	46.4	8.0	3.7	Tr.	1.0	.6	13.3	2.69	33
SD [1]	3.7	5.3	5.5	1.6	1.4	--------	.4	.3	2.2	--------	--------

[1] Standard deviation.

Details of the internal structure of the Arrow pluton are defined by the orientation of mafic inclusions. In the southern part, south of Woods Creek, disc-shaped inclusions define a near-vertical foliation parallel to the walls. North of Woods Creek the foliation is less

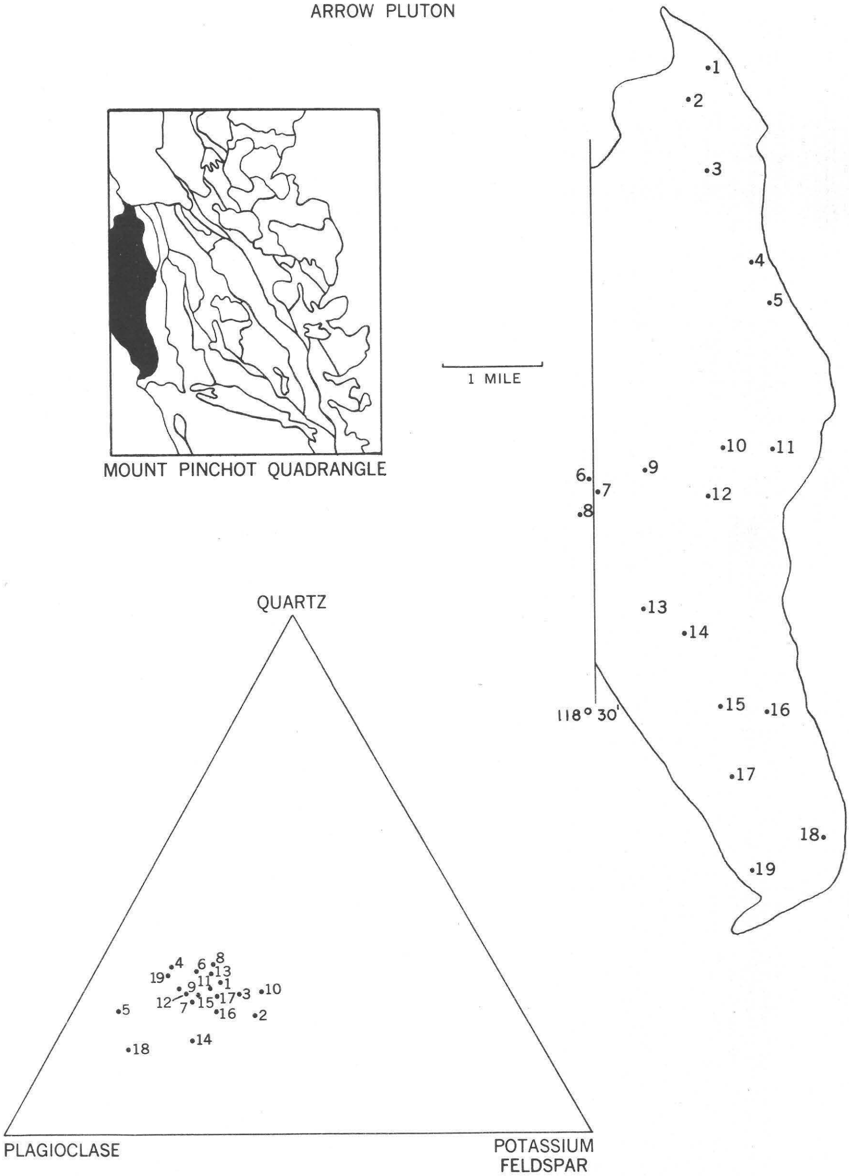

Figure 27.—Location and modal composition of samples from the Arrow pluton. Numbers refer to samples given in table 12.

steeply dipping and a near-horizontal lineation is present, bearing parallel with the long axis of the pluton. Hence the internal structure of the northern part of the pluton is domelike, probably reflecting the roof of the pluton; the Arrow pluton probably plunges south and is only slightly deroofed on its north end.

The Arrow pluton appears to be compositionally zoned in a manner similar to the Cartridge Pass pluton. A mafic border that is widest on the west and south sides of the pluton is indicated by 29 specific gravity determinations. The specific gravity ranges from 2.75 in this outer border to 2.66 in the core, which lies in the northern part of the pluton (fig. 23). The northern part of the pluton is believed to be closer to the roof than the southern; if so, the specific gravity of the center and the top of the pluton is less than that of the walls and the bottom.

The Arrow pluton is characterized by many small parallel left-lateral strike-slip faults that strike eastward and dip north at about 60° (fig. 28). The displacement on each fault is from an inch to a few feet. Figure 29 is a sketch map of a small part of the Arrow pluton, on the north side of Gardiner Creek at an elevation of 10,200 feet. In this region the faults are spaced about 10 feet apart. Indi-

FIGURE 28.—Small left-lateral strike-slip shears in the Arrow pluton on the north side of Gardiner Creek at an elevation of 10,200 feet.

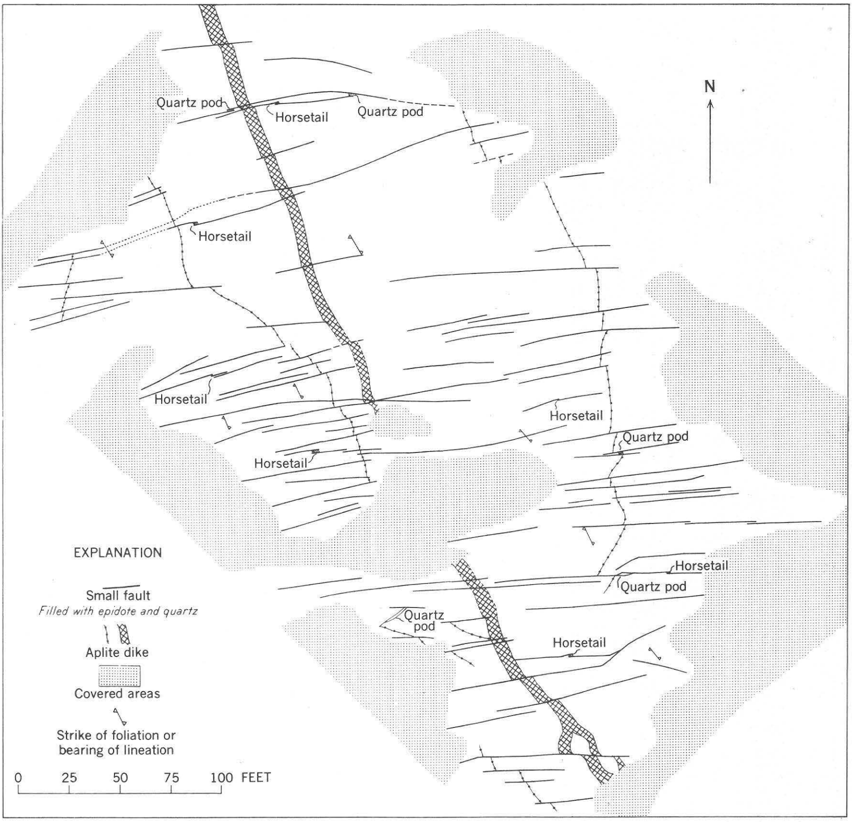

FIGURE 29.—Sketch map of glaciated surfaces of the Arrow pluton on the north side of Gardiner Creek at an elevation of 10,200 feet.

vidual faults can rarely be traced more than a hundred feet. All the faults have nearly horizontal slickensides. Such shearing is prevalent over nearly the entire pluton, though it may be slightly more intense in the area of the sketch map. The faults are almost completely confined to the Arrow pluton, though some cross the Arrow-Cotter contact and are present within the Cotter pluton.

Epidote and quartz are present in the shears, and no shears were observed with any coarse breccia or open spaces. Wallrock structures are commonly dragged around plastically to near parallelism with the shears. Large aplite dikes have been dragged plastically several inches. These relations suggest that shearing occurred at considerable depth and probably while the pluton was still in a semiplastic state.

Figure 30 A and B illustrates how different structures develop between the overlapping ends of two adjacent, nearly parallel, left-lateral faults depending upon the sense of the overlap. When one looks along the strike of one of these left-lateral faults, if the over-

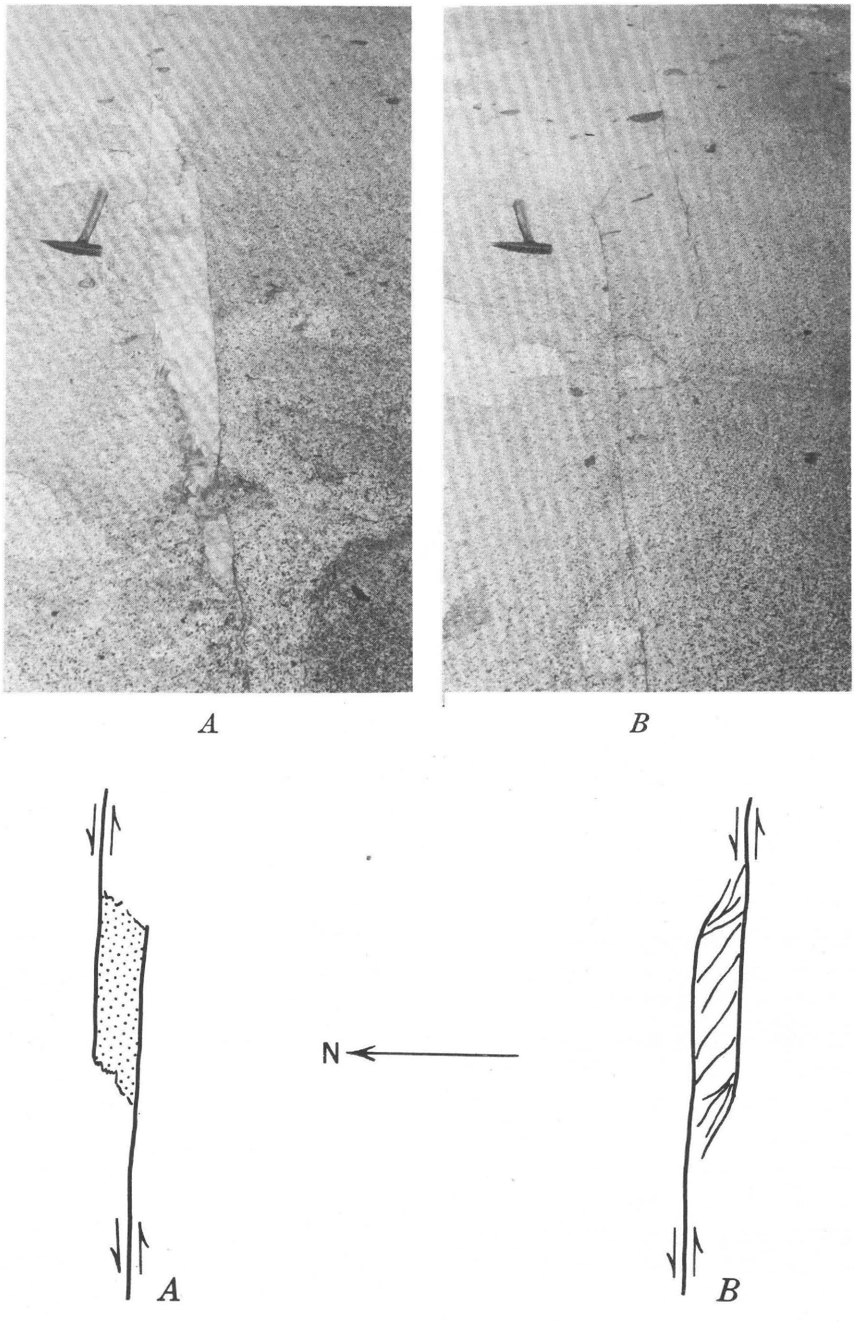

Figure 30.—Structures developed where one left-lateral, stripe-slip fault overlaps another. Where the northern fault overlaps the east end of another fault (*A*), a zone of tension is developed between the faults and a quartz pod results. Where the northern fault overlaps the west end of another fault (*B*), a zone of compression is developed between the faults and a horsetail shear structure results. Photographs from the south end of the Arrow pluton on the north side of Gardiner Creek at an elevation of 10,200 feet.

lapping fault is on the left (fig. 30, A) a quartz pod lies between the overlapping ends, but if the overlapping fault is on the right (fig. 30, B) a horsetail shear structure lies between the faults. The quartz pods occupy fractures caused by local tension between the faults, whereas the horsetail shears occupy a zone of local compression between the faults. The overlapping faults can be compared to a single fault whose trace is deflected by a double curve. Should any left-lateral fault be viewed along its trace, if the trace is deflected to the left, the deflection is a locus of tension, but if the trace is deflected to the right, the deflection is a locus of compression.

The extensive small-scale faulting of the Arrow plutton may be a result of the same stresses that formed large left-lateral strike-slip faults of the Rae Lake area, a few miles to the southeast. Here are at least three large east-trending left-lateral faults of the same strike and sense of displacement as the many small faults of the Arrow pluton. These faults cut and offset the Cotter, Bullfrog, Dragon, and Diamond plutons. The maximum displacement of the northern of these three is about one-half mile. The small-scale faulting of the Arrow pluton is related to the intrusive period and is probably Cretaceous in age. If the large parallel faults of the Rae Lakes are the result of the same stresses, they too are Cretaceous.

The Arrow pluton is clearly younger than the Pyramid pluton: (1) Adjacent to both the east and west masses of the Pyramid pluton, the Arrow pluton contains inclusions of the distinctive coarse-grained granodiorite of the Pyramid pluton, and (2) apophyses from the Arrow pluton penetrate the east mass of the Pyramid pluton in a cirque northwest of Window Peak.

Transgression of structures within the Arrow pluton at the contact with the Cartridge Pass pluton indicates that the Arrow pluton is older. The presence in the Arrow pluton of numerous small faults that are absent or sparingly present in the Cotter pluton suggests that the Arrow pluton is older than the Cotter pluton. The Arrow pluton is also considered older than the Paradise pluton on the basis of intrusive relations west of the quadrangle boundary.

SPOOK PLUTON

The Spook pluton is a zoned granitic intrusion in the northeast quarter of the quadrangle east of the Sierra crest. The pluton is named for Spook Canyon, a narrow canyon in its northern part. The exposed area of the Spook pluton is 13.3 square miles; basalt flows and alluvial deposits overlie the east part of the mass and may conceal a considerable part.

The Spook pluton is composed of two facies of slightly different age. An outer, mafic facies almost completely surrounds an inner,

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silicic facies. The two facies are separated by a sharp contact, along which there is ample evidence that the inner mass is the younger. (1) Schlieren in the inner mass parallel the contact. (2) The contact transects a foliation in the outer mass. (3) The contact is very sharp and distinct, but the two granodiorites are separated by a layer of pink pegmatite, 6 to 12 inches wide, that grades into the inner facies but is sharp against, and sends dikelets into, the outer facies. The contact dips out, under the older marginal facies.

The rocks of the two facies are actually quite similar and both plot close to the granodiorite-quartz monzonite field boundary (fig. 31 and table 13). The outer facies averages 2.4 percent hornblende and 0.5 mafic inclusions per square yard. In the inner facies both hornblende and mafic inclusions are rare to absent. Some of the specimens of the inner facies contain muscovite, a mineral rare in the granitic rocks of the quadrangle.

The main part of the inner facies of the Spook pluton is porphyritic; only that part east of about the 6,800-foot contour on the mountain

TABLE 13.—*Modes of Spook pluton*

Sample No.	Quartz	K-feldspar	Plagioclase	Biotite[1]	Hornblende	Opaque accessories	Nonopaque accessories	Color index	Sp gr	Percent An in plagioclase
Outer facies										
1	21.8	17.4	46.8	8.0	4.3	0.9	1.0	14.2	2.67	--------
2	21.6	18.8	47.2	7.2	3.4	1.1	.8	12.5	2.68	34
3	29.0	17.2	43.2	7.0	2.7	.3	.6	10.6	2.65	--------
4	24.4	18.8	47.1	6.0	3.1	.4	.1	9.6	2.65	--------
5	21.6	20.9	46.5	9.0	--------	1.4	.4	10.8	2.62	31
6	28.6	19.0	41.7	7.8	1.6	.7	.5	10.6	2.67	--------
7	21.5	21.0	48.0	7.2	1.3	.5	.6	9.6	2.67	--------
8	29.2	11.6	49.0	6.9	2.5	.4	.5	10.3	2.68	--------
9	22.2	22.6	44.3	6.3	2.8	1.2	.6	10.9	2.67	--------
10	19.2	15.8	55.5	6.2	2.4	.4	.7	9.7	--------	--------
11	30.1	16.2	46.1	4.6	2.0	.4	.5	7.5	2.67	32
12	24.8	20.8	44.7	6.3	2.2	.8	.4	9.7	2.66	--------
Av	24.5	18.3	46.7	6.9	2.4	.7	.6	10.6	2.66	32
Inner facies										
13	27.2	21.1	46.1	4.5	--------	0.8	0.3	5.6	2.57	--------
14	22.9	23.2	49.3	3.8	0.1	.6	.1	4.6	--------	--------
15	26.2	14.6	53.4	4.6	--------	.8	.4	5.8	2.64	--------
16	25.4	10.1	61.1	2.3	.5	.4	.2	3.4	2.64	27
17	26.2	19.0	51.1	2.9	.2	.4	.2	3.7	2.65	--------
18[1]	22.8	28.3	43.0	4.8	--------	.9	.3	6.0	2.59	--------
19	24.2	16.9	50.1	6.7	.6	.5	1.0	8.8	2.66	26
20	25.2	23.1	43.9	5.3	--------	1.0	1.4	7.7	2.62	19
21	22.3	27.1	45.2	4.5	--------	.7	.3	5.5	2.59	--------
22[1]	24.4	19.4	49.7	5.5	--------	.7	.5	7.3	2.62	28
23[1]	22.0	22.1	48.1	6.5	--------	.9	.4	7.8	2.61	31
Av	24.4	20.4	49.2	4.7	.1	.7	.5	6.0	2.62	26
Total Av	24.4	19.3	47.9	5.8	1.3	0.7	0.6	8.4	--------	--------
SD	2.9	4.3	4.4	1.6	1.4	.3	.3	2.8	--------	--------

[1] Samples 18, 22, and 23 contain a trace of muscovite.

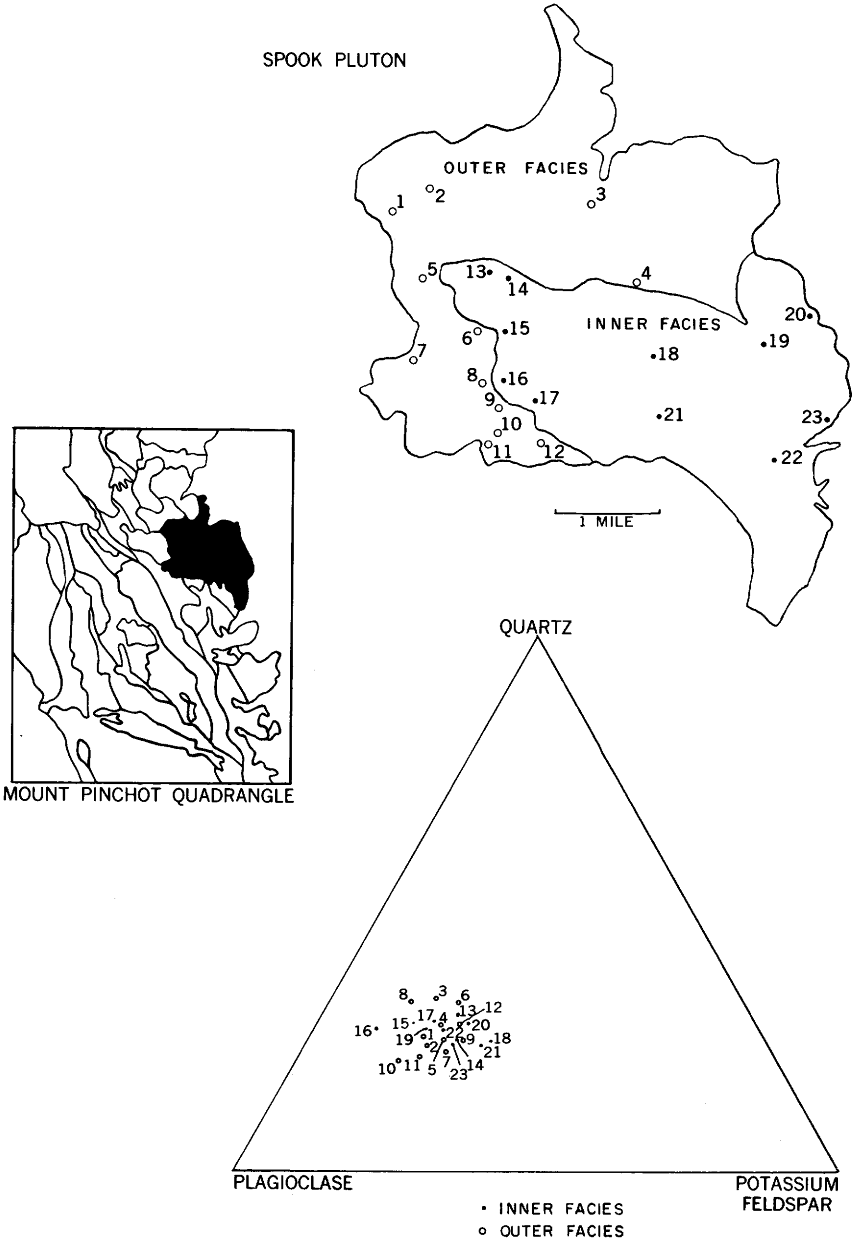

FIGURE 31.—Location and modal composition of samples from the Spook pluton. Numbers refer to samples given in table 13.

front does not contain porphyritic K-feldspar. The K-feldspar phenocrysts contain abundant zonally arranged inclusions of mafic minerals.

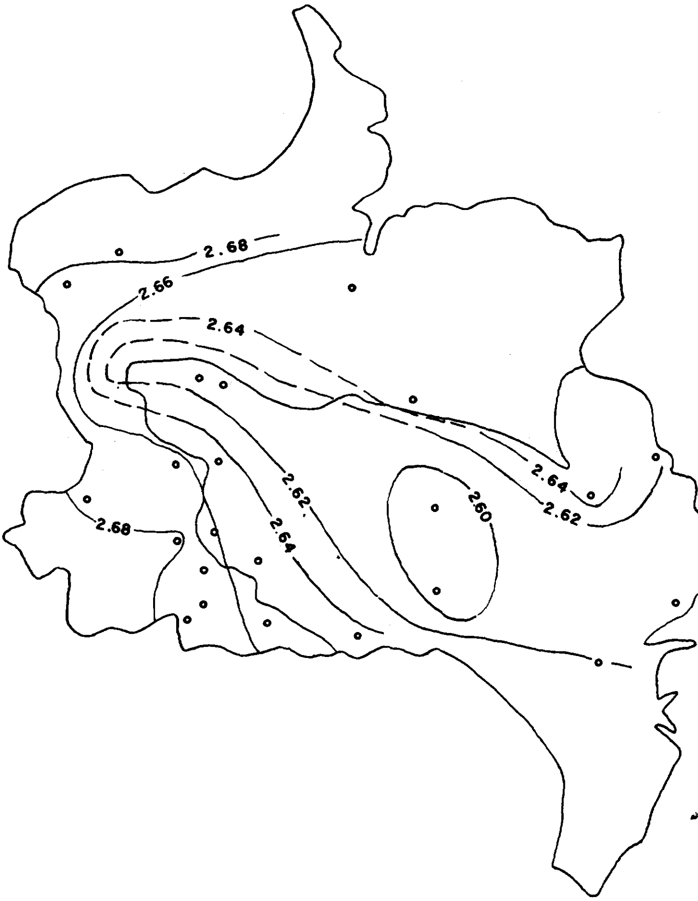

FIGURE 32.—Zonation as indicated by specific gravity measurements in the Spook pluton. Dots indicate location of measured samples.

Specific gravity measurements show that the Spook pluton is gradationally zoned and that the sharp internal contact is merely a minor hiatus within the pluton. The specific gravity is greater than 2.68 in the outer part of the more mafic margin and less than 2.60 in the center of the leucocratic core (fig. 32).

The Spook pluton is remarkably similar to the Cartridge Pass pluton. They are similar in size and shape, both possess a density zonation within similar limits, the composition and range in composition is remarkably similar (compare figs. 31 and 24), both possess a porphyritic core, and in both this inner, more silicic core is intrusive into the outer marginal portion. The main difference is that the core of the Spook is everywhere in intrusive contact with the margin, whereas in the Cartridge Pass the contact is gradational except for one small part, which is intrusive.

The Spook pluton is intruded by the younger Colosseum mass of Lamarck granodiorite, but is younger than all other masses with which it is in contact. Apophyses from the Spook pluton intrude the McDoogle pluton one-half mile north of Sawmill Lake. Both the north and south masses of the Mule Lake pluton are cut by mafic dikes that are clearly cut off by the Spook pluton. The map pattern strongly suggests that the two masses of the Mule Lake pluton were split apart and separated 3½ miles by the intrusion of the Spook pluton. An excellent example of the Spook pluton cutting off mafic dikes is in the small pendant of metasedimentary rock 1 mile east of Mount Perkins, which represents a part of the roof of the Spook pluton. Vertical mafic dikes parallel to the bedding of the metasedimentary rock are sharply cut off at the near-horizontal contact.

WOODS LAKE MASS OF TINEMAHA GRANODIORITE

Trending northwest across the center of the quadrangle is the elongate Woods Lake mass of the Tinemaha granodiorite. This mass is 11.5 miles long and occupies 12.4 square miles, all of which is within the Mount Pinchot quardangle. The northern end is in lower Woods Lake basin, for which the mass is named, very close to the junction of the John Muir Trail and the trail from Sawmill Pass. The mass extends southeast across Mount Baxter, Oak Creek, and Independence Creek to Pinyon Creek in the southeast corner of the map. Much of the eastern part has apparently been cut off by the Independence fault.

The Woods Lake mass has been tentatively correlated with the Tinemaha granodiorite mapped by Bateman (1961) in the Big Pine quadrangle to the north. The correlation is based on similarity of composition and texture and on an abundance of mafic dikes in both masses. The main mass of the Tinemaha granodiorite is about 7 miles north of the Woods Lake mass.

The average rock composing the Woods Lake mass is a slightly porphyritic granodiorite which lies very close to the quartz monzonite field boundary (fig. 33 and table 14). The K-feldspar phenocrysts are generally salmon to white in color and are as much as one-half inch in length.

The Woods Lake mass is extensively intruded by vertical mafic dikes of the Independence swarm. In a traverse 0.8 mile long perpendicular to the strike of the dikes in Woods Lake basin, 114 dikes were crossed. The mass is also somewhat sheared in a direction parallel to the strike of the dikes (northwest). Much shearing has taken place along the dikes, and in nearly every place observed it was of a left-lateral sense.

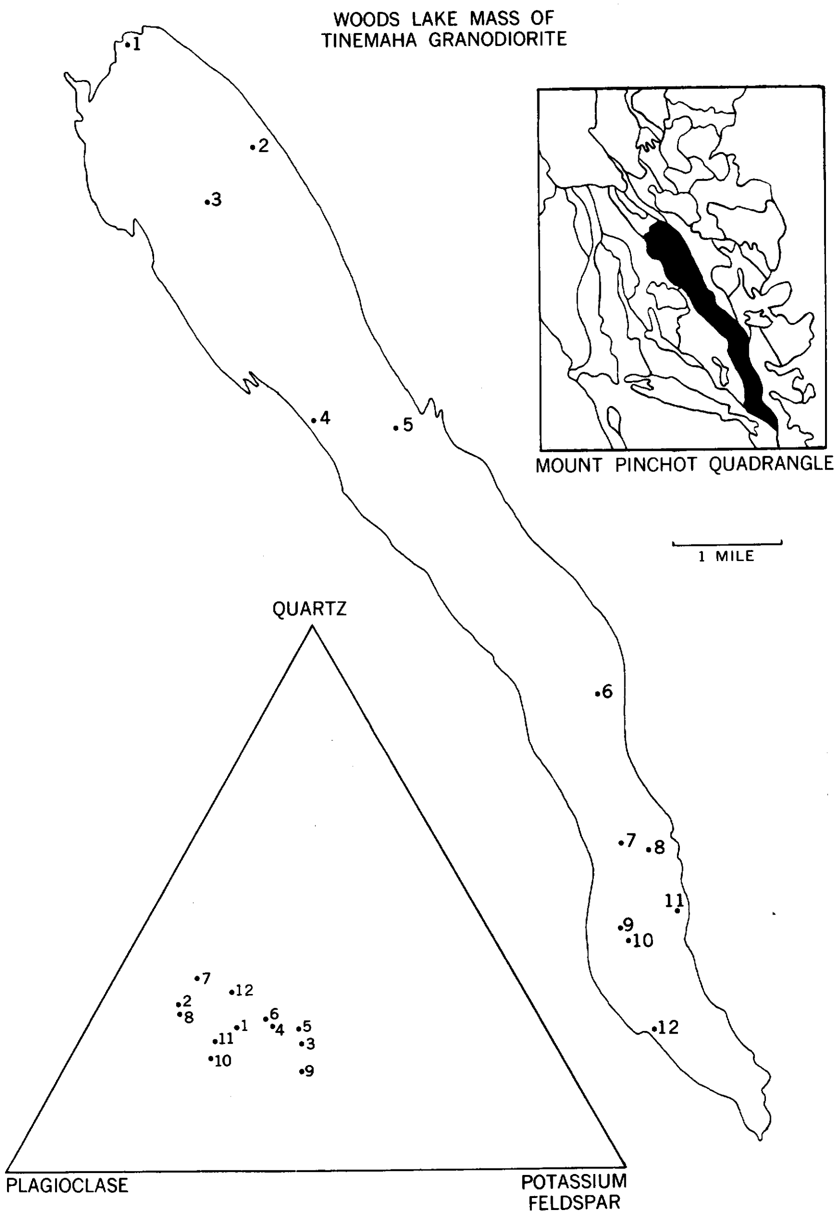

FIGURE 33.—Location and modal composition of samples from the Woods Lake mass of the Tinemaha granodiorite. Numbers refer to samples given in table 14.

The Woods Lake mass contains abundant mafic inclusion (3.3 per square yard) which define a foliation generally parallel to the walls. In the northern part of the mass, the inclusions are triaxial and mark a very steep lineation in addition to the steep foliation. Near-vertical

TABLE 14.—*Modes of Woods Lake mass of Tinemaha granodiorite*

Sample No.	Quartz	K-feld-spar	Plagio-clase	Biotite	Horn-blende	Opaqe accesso-ries	Non-opaque accesso-ries	Color index	Sp gr	Percent An in plagio-clase
1	23.9	22.0	45.2	4.9	2.7	0.4	0.9	8.9	2.61	--------
2	25.7	10.6	47.4	8.5	5.6	1.3	.9	16.3	2.66	33
3	21.2	33.1	36.2	5.5	2.6	.7	.7	9.5	2.65	--------
4	25.0	25.9	38.7	5.8	2.7	1.5	.3	10.3	2.67	--------
5	26.1	30.4	32.6	7.8	1.9	.7	.5	10.9	2.65	--------
6	23.2	24.0	36.7	11.5	3.4	.5	.8	16.2	2.67	--------
7	29.3	11.2	41.4	13.1	1.6	2.0	1.3	18.0	2.67	30
8	24.3	11.0	47.3	14.8	1.0	1.1	.5	17 4	2.68	--------
9	17.2	35.4	39.0	2.3	4.1	1.2	.7	8.3	2.62	26
10	18.3	20.2	49.2	9.5	1.2	.8	.7	12.2	2.69	30
11	20.1	18.7	45.9	8.0	4.8	1.5	.9	15.2	--------	--------
12	29.4	18.5	42.1	6.5	1.7	1.0	.8	10.0	2.63	--------
Av	23.6	21.8	41.8	8.2	2.8	1.1	.8	12.9	2.65	30
SD	3.9	8.5	5.3	3.6	1.5	.5	.2	3.6	--------	--------

lineation is not found in any of the other plutons and its significance is not understood.

Where the Woods Lake mass is in contact with the Oak Creek metavolcanic pendant on the west it forms an injection complex as wide as one-half mile. The granodiorite has split the metavolcanic rock apart into thin leaves approximately parallel to the bedding. Both metavolcanic and granitic rock are cut by mafic dikes, commonly at a slight angle to the lit-par-lit structure. The Woods Lake mass of the Tinemaha granodiorite is in contact with only one older mass, the White Fork pluton on the west side. Both masses contain mafic dikes. In some areas the White Fork pluton is quite gneissose and thin-section study shows that it is intimately sheared throughout most of its extent. For this reason it is considered older than the Woods Lake mass. The Twin Lakes alaskite on the north end of the Tinemaha pluton is instrusive into the Tinemaha granodiorite. It sends a large number of dikes and prongs into the Tinemaha granodiorite and also includes isolated masses of the distinctive porpryritic granodiorite.

COTTER PLUTON

The Cotter pluton is a slender, north-trending, body nearly 10 miles in length, which occupies 6.9 square miles in the western part of the quadrangle. The pluton is named after Mount Cotter on the southern end of King Spur between Sixty Lake Basin and Gardiner Basin. It crops out in the upper portion of Gardiner Basin and Mount Gardiner, and extends north across Woods Creek west of Castle Domes to the Cartridge Pass pluton. The silicic granodiorite of the Cotter pluton (fig. 34 and table 15) is characterized by its relatively fine grain, dark color (color index, 11.1), and an abundance of mafic inclusions (averaging about 5.4 per square yard). The apparent darker color of the Cotter pluton is due largely to its fine grain size.

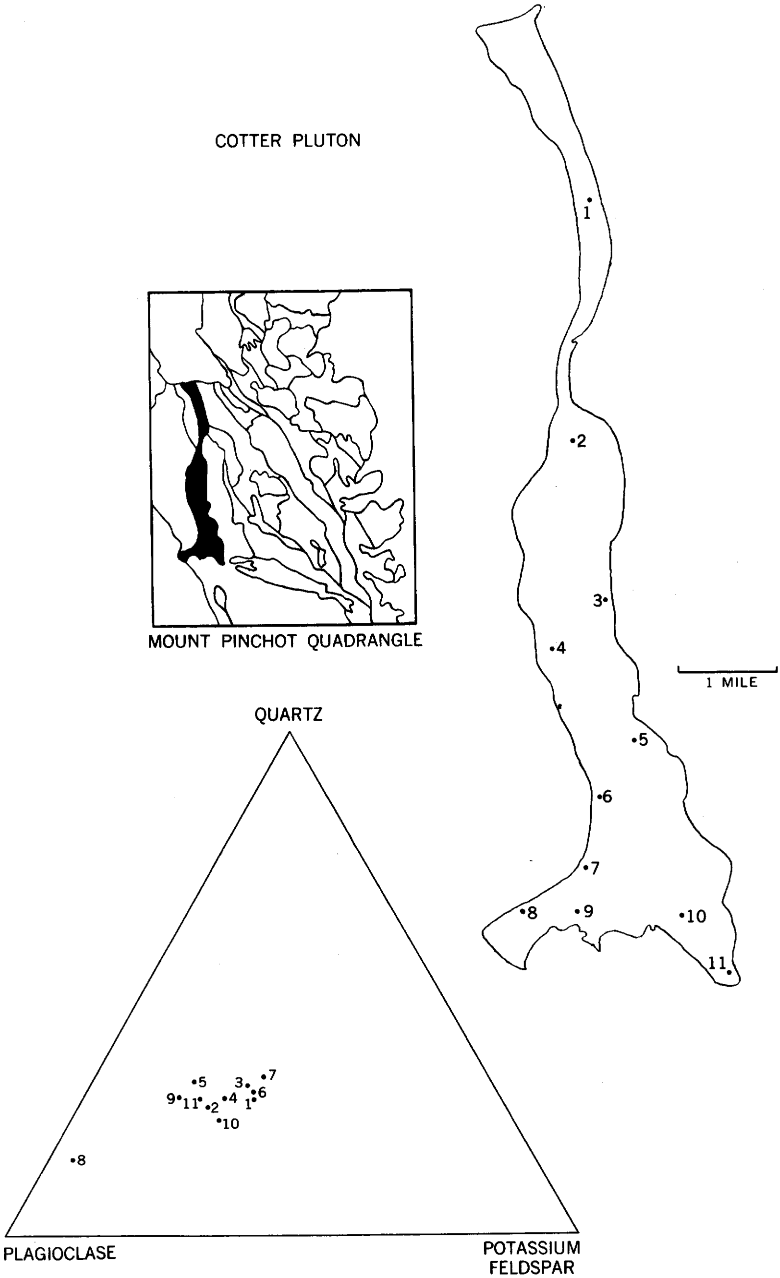

FIGURE 34.—Location and modal composition of samples from the Cotter pluton. Numbers refer to samples given in table 15.

On the south and east the Cotter pluton is in contact with alaskite of the Bullfrog pluton. This contact is well exposed on the west slope of Mount Gardiner. The Cotter pluton is younger as shown by the fol-

TABLE 15.—*Modes of the Cotter pluton*

Sample No.	Quartz	K-feld-spar	Plagi-oclase	Bio-tite	Horn-blende	Pyrox-ene	Opaque acces-sories	Non-opaque acces-sories	Color index	Sp gr	Per-cent An in plagi-oclase
1	24.8	27.2	39.2	6.3	1.2	-------	1.2	0.3	9.0	2.64	-------
2	21.4	19.6	44.4	7.6	5.5	0.1	.7	.6	14.5	2.67	-------
3	26.6	25.0	39.2	5.2	2.1	-------	1.0	.8	9.1	2.67	-------
4	25.1	22.7	44.9	5.1	1.2	-------	.4	.4	7.1	2.67	-------
5	26.4	15.9	46.0	6.5	1.8	-------	.9	.7	11.9	2.68	-------
6	25.6	26.6	39.0	7.3	.3	-------	.8	.5	8.9	2.64	31
7	28.9	26.9	37.0	6.0	.2	-------	1.0	.1	7.3	2.64	-------
8	10.8	3.7	64.6	10.2	8.3	-------	1.7	.6	20.8	2.79	-------
9	24.1	14.5	49.4	8.2	2.9	-------	.8	.2	12.1	2.68	30
10	20.6	23.1	46.2	6.3	2.5	-------	.7	.8	10.3	2.64	-------
11	23.9	18.0	46.2	5.4	4.8	-------	.7	.8	11.7	2.67	32
Av	23.5	20.3	45.1	6.9	2.8	-------	.9	.5	11.1	2.67	31
SD	4.8	7.1	7.5	1.6	2.5	-------	.3	.3	3.9	-------	-------

lowing evidence: (1) The foliation of the Cotter pluton as defined by parallel mafic inclusions parallels the contact. (2) A faint foliation (alinement of feldspar plates) of the Bullfrog pluton is cut off at the contact at an angle of approximately 60°. (3) The Cotter pluton is slightly finer grained toward the contact, whereas the Bullfrog pluton is unaffected adjacent to the contact. On the northeast the Cotter pluton intrudes an elongate septum of metasedimentary rock. Many mafic dikes belonging to the Independence dike swarm cut the metasedimentary rock but none intrude the Cotter pluton; the Cotter pluton is younger than the Independence dike swarm. On the northwest the Cotter pluton meets the coarse-grained granodiorite of the Pyramid pluton along a sharp contact. The foliation of each pluton parallels the contact, but dikes of the Cotter pluton in the Pyramid pluton show that the former is younger. South of the Pyramid pluton, the Arrow pluton is in contact with the Cotter pluton, and this contact though exposed almost continuously for many miles, shows few diagnostic age relations. Generally the two granitic bodies are separated by a zone of hybrid rocks up to several hundred feet in width. This mixed zone commonly is composed of closely packed mafic inclusions about 6 inches in average length, in a matrix of lighter granitic rock. On the west side of this septum, the matrix rock appears to be granodiorite of the Arrow pluton and on the east, granodiorite of the Cotter pluton; in the center the two merge. The internal structure of each pluton is parallel to the contact. Indirect evidence suggests that the Cotter pluton is younger than the Arrow pluton: (1) The Arrow pluton is characterized over most of its area by small, east-west, left-lateral shears filled with epidote and quartz (fig. 28). Though some of these shears are also present in the Cotter, they are far less abundant. A possible explanation is that shearing began before emplacement of the Cotter and continued somewhat after, but an equally probable explanation is that the two rocks simply responded differently to stress. (2) In several places abundant schlieren, dioritic stringers, and swarms of inclusions are present within the Cotter, near the contact and parallel with the contact. The absence of similar features in the Arrow pluton suggests the Cotter pluton is younger, but this relation is not conclusive.

PARADISE PLUTON

The Paradise pluton is in the southwest corner of the quadrangle, where 11.6 square miles are exposed. Reconnaissance mapping west of the quadrangle suggests that the pluton is quite large and that only a small part is in the Mount Pinchot quadrangle. It is known that the pluton does not extend far west of Zumwalt Meadows in Kings Canyon. This would place an upper width limit at the latitude of lower Bubbs Creek at 8 miles. Within the quadrangle, the pluton extends south from lower Gardiner Creek and Gardiner Pass, across Bubbs Creek toward the Kings-Kern divide. Reconnaissance mapping in the Marion Peak quadrangle west of the Mount Pinchot quadrangle shows that theParadise pluton is well exposed on both walls of Paradise Valley on the South Fork of the Kings River; the pluton was named for this locality.

The rock of the Paradise pluton is porphyritic and contains K-feldspar crystals as much as 4 cm in length. These phenocrysts are filled with inclusions, mainly hornblende and biotite, which are arranged in a crude zonal pattern. The abundant dark inclusions tend to camouflage the K-feldspar phenocrysts, though the K-feldspar becomes obvious by reflection of light from large cleavage surfaces. The average rock of the Paradise pluton is a granodiorite, which plots extremely close to the quartz monzonite boundary (fig. 35 and table 16). The Paradise pluton has relatively few mafic inclusions (0.2

TABLE 16.—*Modes of the Paradise pluton*

Sample No.	Quartz	K-feldspar	Plagioclase	Biotite	Hornblende	Pyroxene	Opaque accessories	Nonopaque accessories	Color index	Sp gr	Percent An in plagioclase
1	19.3	16.8	46.5	11.1	4.5	--------	1.2	0.6	17.4	2.68	--------
2	23.8	19.3	45.4	6.6	3.6	--------	.7	.5	11.4	2.70	--------
3	19.3	18.7	47.3	9.0	4.3	--------	.7	.7	14.7	2.69	--------
4	19.3	23.2	38.7	12.2	4.3	--------	1.7	.7	18.9	2.66	--------
5	24.7	29.0	35.2	5.6	3.6	--------	.6	1.2	11.0	2.65	31
6	22.2	30.2	32.4	6.1	5.7	--------	1.8	1.6	15.2	2.67	--------
7	22.6	19.7	47.5	5.1	2.7	0.2	1.2	1.0	10.2	2.67	31
8	21.5	13.9	52.8	6.7	2.4	--------	1.1	1.7	11.9	2.68	--------
9	22.0	20.7	48.8	5.4	1.4	--------	.4	1.3	8.5	2.65	31
10	24.2	25.7	39.2	6.4	3.4	--------	.8	.4	11.0	--------	--------
11	27.2	19.8	46.3	3.2	2.3	--------	.8	.5	6.8	2.64	30
12	26.0	19.9	45.2	4.8	2.9	--------	.7	.4	8.8	2.67	31
13	22.4	27.2	40.6	4.0	4.1	--------	.9	.9	9.9	2.65	31
14	19.8	23.9	43.9	7.8	3.3	--------	.5	.9	12.5	2.65	28
Av	22.5	22.0	43.6	6.7	3.5	--------	.9	.9	12.0	2.67	30
SD	2.5	4.7	5.6	2.6	1.1	--------	.4	.4	3.4	--------	--------

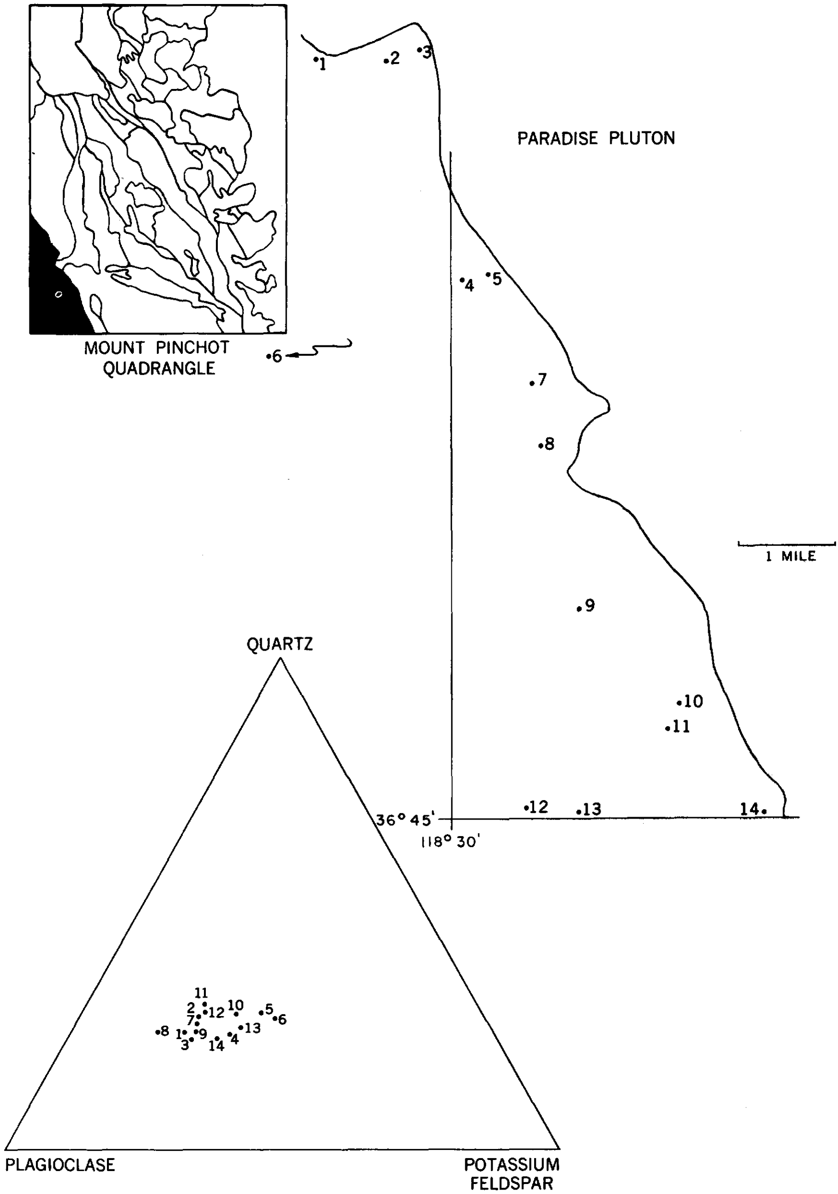

FIGURE 35.—Location and modal composition of samples from the Paradise pluton. Numbers refer to samples given in table 16.

per sq. yd.) in comparison with rocks of similar color index (12.0).

Specific gravity measurements of hand specimens of the Paradise pluton suggests that the mass is zoned in a manner similar to the Cartridge Pass pluton (fig. 36). Specific gravities on the east margin are generally greater than 2.67. These decrease westward toward the

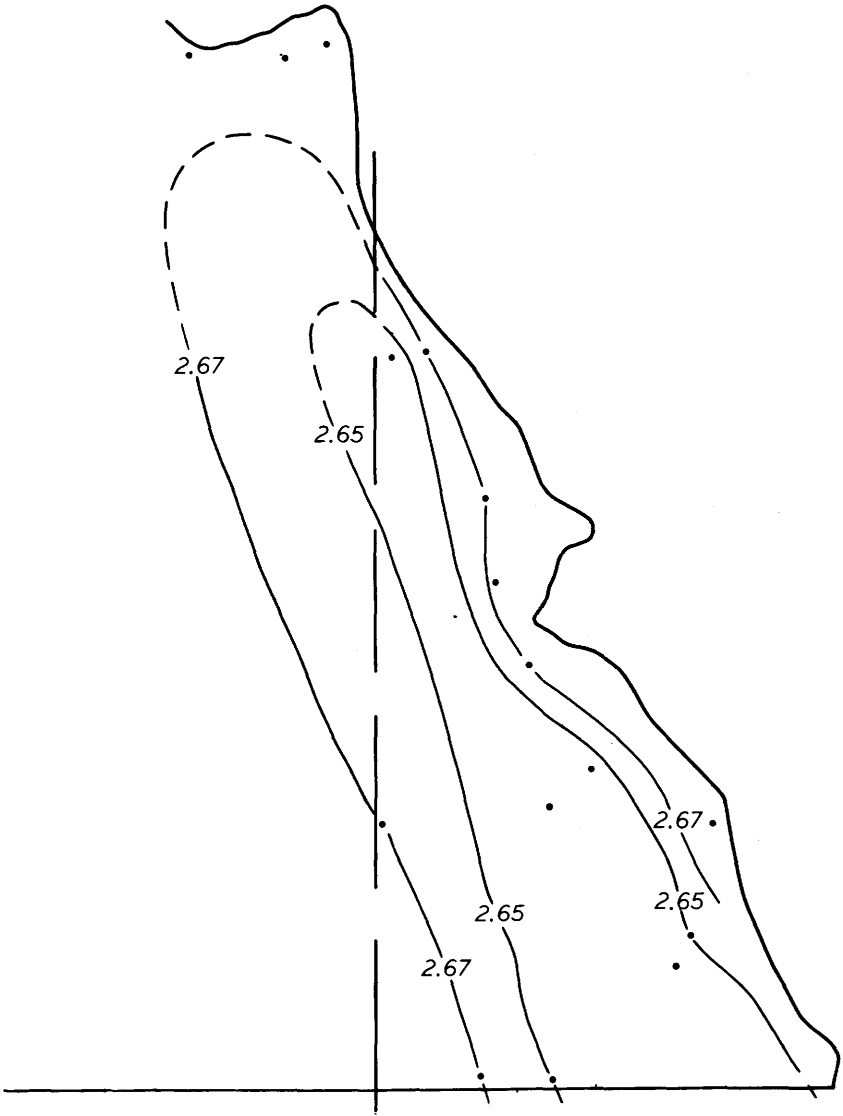

FIGURE 36.—Specific gravity zonation in the Paradise pluton. Dots indicate measured samples.

central part of the body to less than 2.65 and then increase again, presumably toward the unmapped west margin.

The Paradise pluton is in contact with only one other pluton in the Mount Pinchot quadrangle: the Arrow pluton to the northeast. On the east the Paradise pluton is separated from the Bullfrog pluton by a thin septum of metavolcanic rock. On the west side of Paradise Valley west of the Mount Pinchot quadrangle the Paradise pluton contains inclusions of rock from the Arrow pluton and therefore is younger.

MULE LAKE PLUTON

Two isolated masses make up the Mule Lake pluton which is in the northeast quarter of the quadrangle east of the Sierra crest (fig. 37).

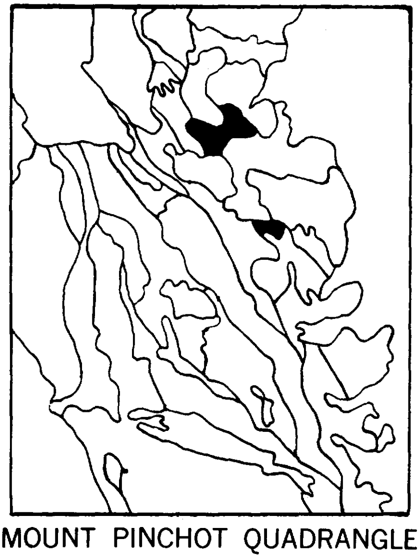

FIGURE 37.—North and South masses of the Mule Lake pluton.

Taken together the two masses comprise 2.2 square miles. The North mass is mainly in the headwaters region of Armstrong Canyon; the South mass underlies Mule Lake in the drainage of Sawmill Creek.

The Mule Lake pluton is composed of older granitic rock that has been intimately intruded by mafic dikes and has been somewhat sheared and epidotized. The rock is a silicic granodiorite (no modes were determined), and is porphyritic with small (about 1 cm) phenocrysts of K-feldspar. The Mule Lake pluton is older than all other granitic rocks adjacent to it. The Siberian pluton, Spook pluton, and Colosseum mass of the Lamarck granodiorite clearly cut off mafic dikes in the Mule Lake pluton. The Spook pluton has split the Mule Lake pluton and has presumably shouldered apart the two masses. The Mule Lake pluton is intrusive into the mafic complex to the north and east. Dikes of granodiorite of the Mule Lake pluton are abundant in the plutonic complex of lower Armstrong Canyon. The western part of the large mafic mass of Goodale Creek is cut by nearly horizontal granitic sills, which are probably offshoots from the Mule Lake pluton. These sills (not mapped) are clearly visible as a horizontal layering in the dark rock from U.S. Highway 395 in Owens Valley. Mafic dikes cut the sills and the enclosing dioritic rock.

WHITE FORK PLUTON

The White Fork pluton is a mass of silicic granodiorite which crops out west of the center of the map and extends from the Cartridge Pass pluton on the northwest to the Baxter Creek area on the southeast. The pluton is named for White Fork, a Woods Creek tributary that crosses the northern part of the mass. The pluton covers 7.9 square miles and is 8 miles long.

The rock of the pluton varies considerably in composition and texture from place to place (fig. 38 and table 17). Some of the rock is granodiorite and some is quartz monzonite. Much of the rock is sheared and some has been rendered gneissose or schistose by shearing. The rock of the northern part of the pluton is strongly foliated and contains small, strung-out dark spots which represent the former ferromagnesian minerals; it was called a spotted schist in the field, but microscopic study shows it to be an intimate sheared granodiorite.

TABLE 17.—*Modes of White Fork pluton*

Sample No.	Quartz	K-feldspar	Plagioclase	Biotite	Hornblende	Opaque accessories	Nonopaque accessories	Color index
1	22.8	20.4	41.1	11.1	3.5	0.8	0.4	15.8
2	27.4	28.5	33.7	9.6		.5	.3	10.4
3	31.7	13.4	48.8	5.6	.1	.2	.2	6.1
4	27.1	16.0	42.2	12.0	1.6	.7	.3	14.6
5	23.0	19.1	44.8	7.8	4.1	.8	.3	13.0
Av	26.4	19.5	42.1	9.2	1.9	.6	.3	12.0
SD	3.7	5.7	5.6	2.6	1.9	.3	.1	3.9

Under the microscope nearly every rock shows evidence of deformation and recrystallization of an original granodiorite or quartz monzonite. In most rocks biotite is present as aggregates of many small crystals rather than as large single crystals. Quartz shows mosaic extinction and occurs in elongated aggregates. K-feldspar is recrystallized and redistributed through the rock. The northern part of the White Fork mass is thoroughly impregnated with disseminated pyrite. The center of this area of pyritization is about 1 mile west of Crater Mountain. Pyritization affects the Twin Lakes pluton and the metamorphic septum between the Twin Lakes and White Fork plutons, but not the Cartridge Pass pluton to the north.

The White Fork pluton is older than the mafic dikes and is intimately cut by them. The pluton is a favorable host, because the dikes have intruded along the planes of the well-developed gneissic foliation,

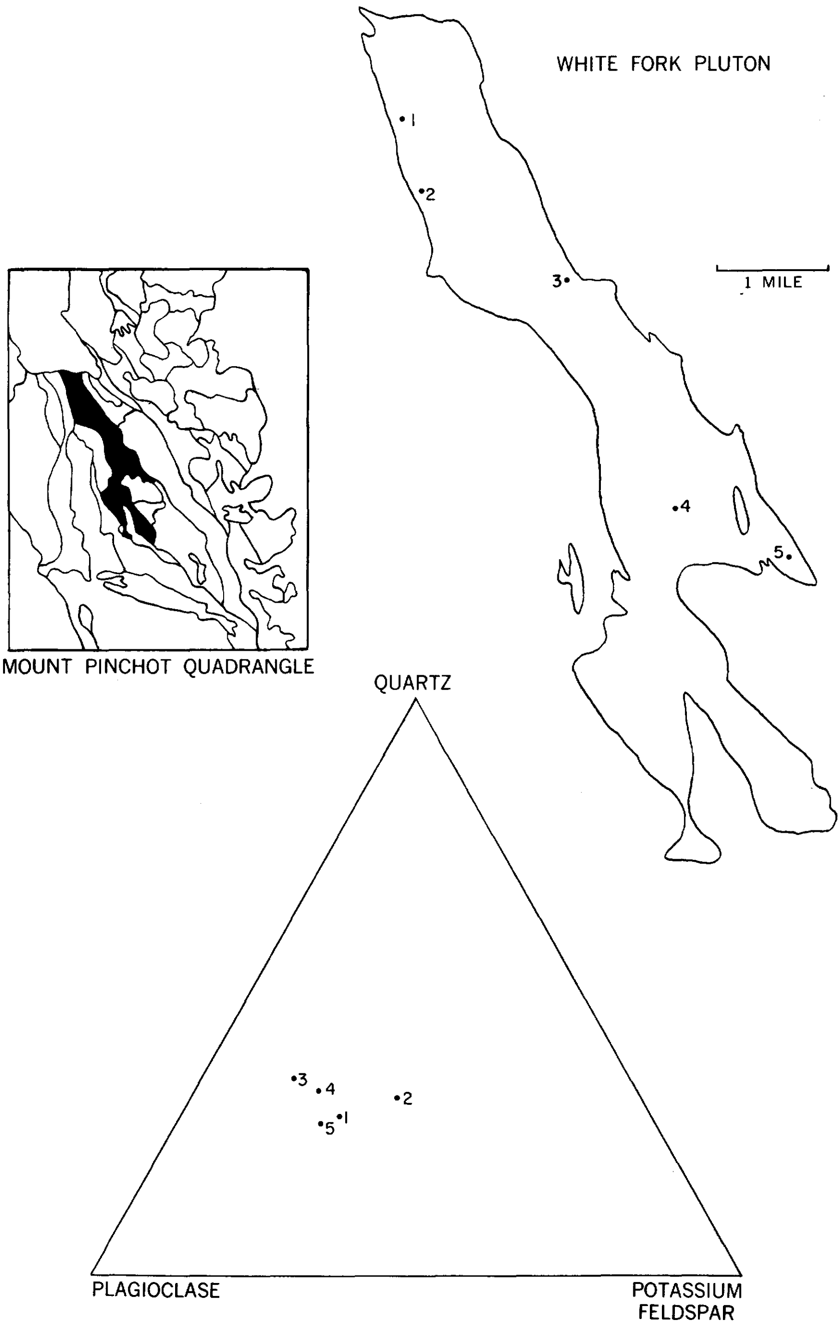

FIGURE 38.—Location and modal composition of samples from the White Fork pluton. Numbers refer to samples given in table 17.

which therefore largely predates the period of dike injection. The Cotter, Cartridge Pass, Bullfrog, Baxter, and Dragon plutons cut the mafic dikes and are clearly younger than the White Fork pluton. The contacts between the White Fork pluton and the Twin Lakes pluton and Tinemaha granodiorite are sheared and undiagnostic. However, the thorough deformation and recrystallization of the White Fork pluton suggests it is older, and is perhaps the oldest granitic pluton in the quadrangle.

QUARTZ MONZONITE PLUTONS

Four quartz monzonite plutons are present in the quadrangle: Siberian, Goodale, Taboose, and McGann plutons. These masses have about equal amounts of the two feldspars, although they differ from one another in the amount of quartz. All the quartz monzonites are younger than the mafic dikes except the Taboose pluton.

SIBERIAN PLUTON

East of the Sierra Nevada crest in the northern part of the quadrangle is a crescent-shaped pluton, convex to the west, 2½ miles long and nearly 1 mile wide (area, 2.5 sq. miles). This pluton has been called the Siberian pluton after the large lake east of Mount Pinchot, which is locally referred to as Siberian Lake. It is composed of a uniform, light-colored quartz monzonite (fig. 39 and table 18). Mafic inclusions are rare or absent.

TABLE 18.—*Modes of the Siberian pluton*

Sample No.	Quartz	K-feldspar	Plagioclase	Biotite	Opaque accessories	Non-opaque accessories	Color index	Sp gr	Percent An in plagioclase
1	26.1	33.6	36.4	3.4	0.3	0.2	3.9	2.64	31
2	25.9	23.2	43.6	6.5	.4	.4	7.3	2.61	26
3	32.8	30.0	33.2	3.5	.5	----------	4.0	2.59	26
4	29.1	33.3	34.2	3.0	.4	.1	3.5	2.59	----------
5	33.5	28.1	34.0	3.6	.5	.2	4.3	2.61	21
Av	29.5	29.6	36.3	4.0	.4	.2	4.6	2.60	26
SD	3.6	4.3	4.2	1.4	.1	.1	1.5	----------	----------

The Siberian pluton is younger than the Red Mountain Creek pluton to the north, the Mule Lake pluton to the south, and the dioritic complex of Goodale Creek to the east, because these plutons are cut by mafic dikes of the Independence swarm, whereas the Siberian pluton transects the dikes. The Siberian pluton appears to transect the swarm of dikes sent off by the Striped pluton and is believed to be younger. On its northern margin the Siberian pluton is in contact with the Goodale pluton. Inclusions of rock from the Goodale pluton in the Siberian pluton show that the Siberian pluton is younger. This

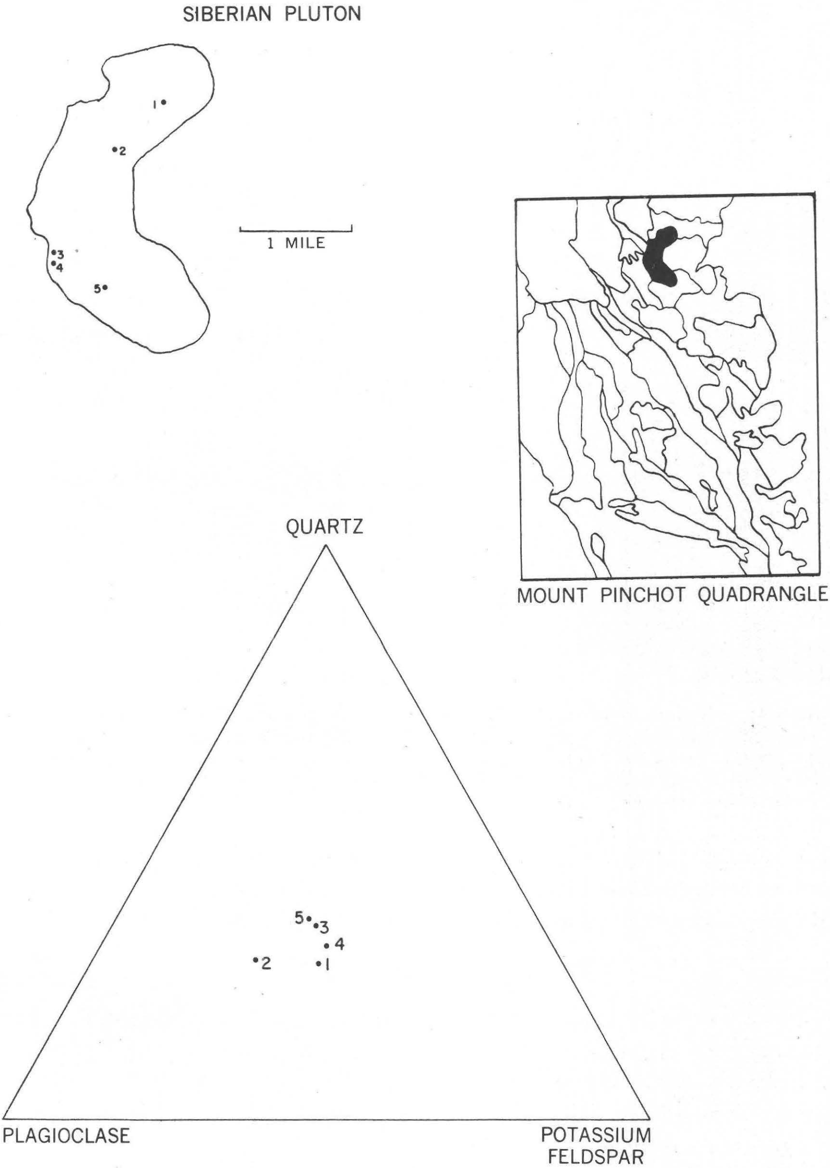

FIGURE 39.—Location and modal composition of samples from the Siberian pluton. Numbers refer to samples given in table 18.

contact is well exposed near the summit of Goodale Mountain and also on the precipitous south wall of Taboose Creek canyon. It is generally sharp, though in some places an intimate mixture of the two rock types is exposed.

On a cirque wall east of Striped Mountain, the contact between biotite schist and the Siberian pluton is well exposed (fig. 40). This

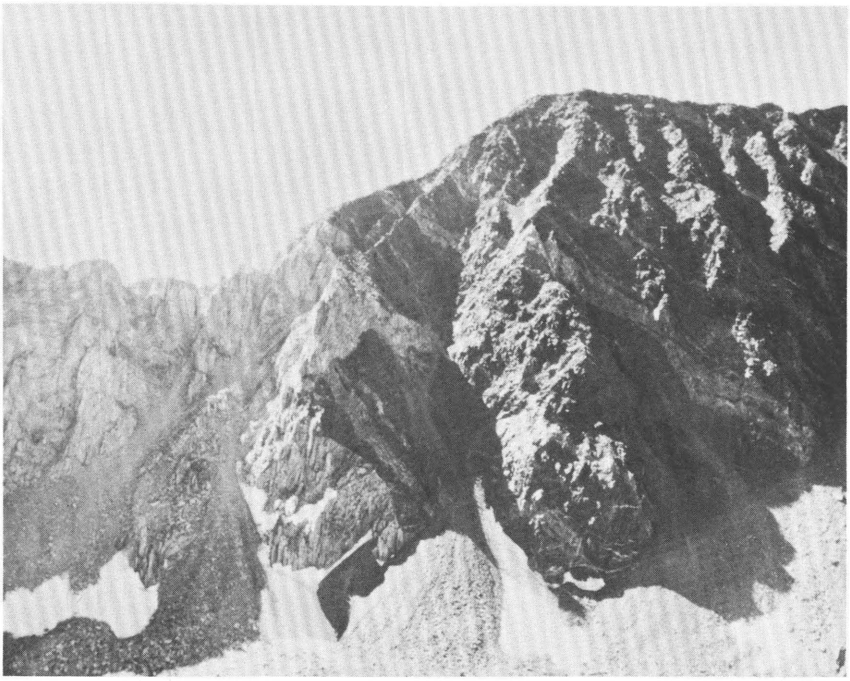

FIGURE 40.—Contact between biotite schist and quartz monzonite of the Siberian pluton on cirque wall east of Striped Mountain. The contact follows the schistosity in the metamorphic rock but jumps from one foliation plane to another half way up the cliff face. At the offset the pluton sends a tongue into the metamorphic rock. Cliff face is about 1,000 feet high.

is a typical sharp and broadly conformable intrusive contact. The contact follows the schistosity in the metamorphic rock but jumps from one foliation plane to another half way up the cliff face.

TABOOSE PLUTON

The Taboose pluton is on the northern border of the Mount Pinchot quadrangle. About 3.5 square miles is in the quadrangle and approximately an equal area is exposed in the Big Pine quadrangle to the north. The pluton is named after Taboose Creek, which drains most of the area covered by the pluton. It is grouped with the masses of finer grained quartz monzonite by Bateman (1961).

The rock of the Taboose pluton is generally a leucocratic quartz monzonite with about equal amounts of plagioclase, quartz, and K-feldspar (fig. 41). The eastern part of the pluton is slightly coarser grained and less uniform than the western part, and zones of mixed or layered, finer and coarser grained phases are present.

The Taboose pluton is in contact with the Red Mountain Creek pluton on the west, and is younger. The western part of the Taboose pluton is cut by a few mafic dikes continuous with those cutting the

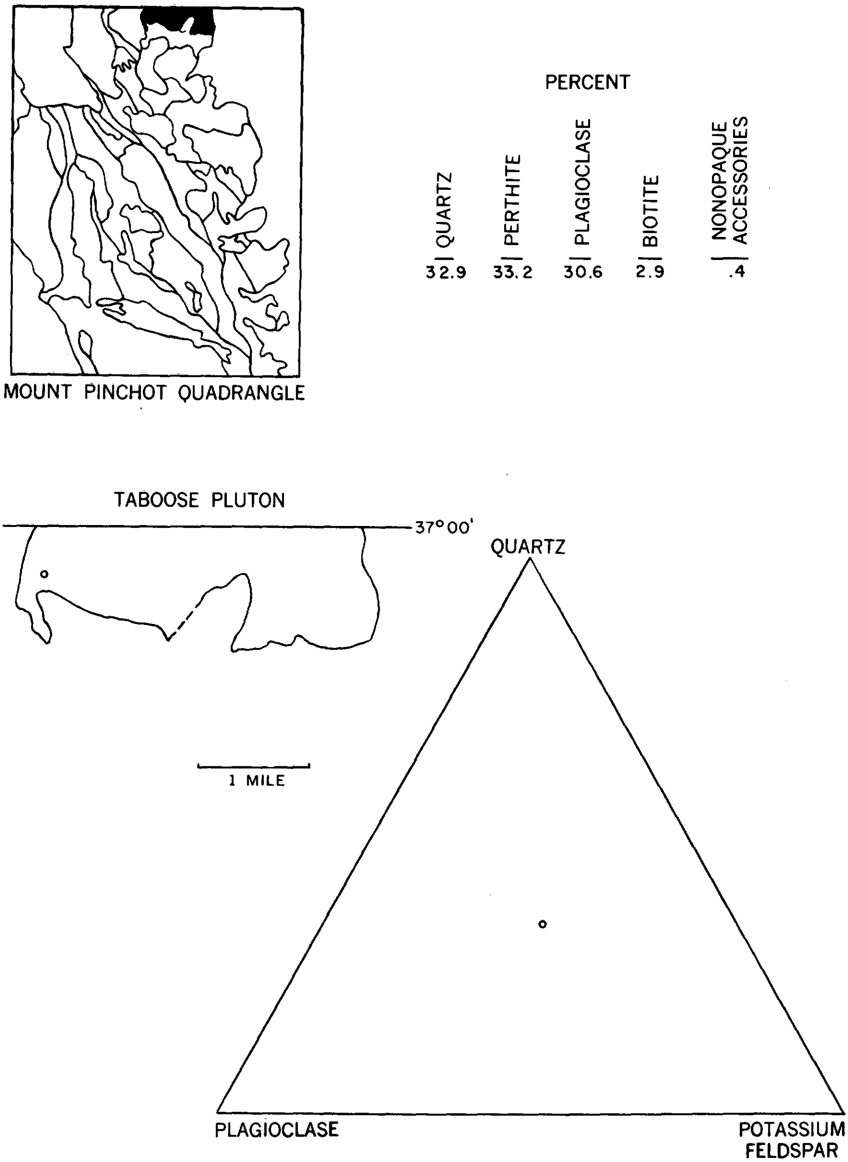

FIGURE 41.—Location and modal composition of one sample from the Taboose pluton.

alaskite of the Red Mountain Creek pluton, and is therefore considered older than the Independence dike swarm. On the south the quartz monzonite of the Taboose pluton is intruded by the Goodale pluton. Schlieren in the Goodale parallel the contact which transects a faint structure in the quartz monzonite of the Taboose pluton. Also the Goodale is younger than the mafic dikes which are believed to be the same as those which cut the Taboose pluton.

McGANN PLUTON

The McGann pluton is located east of the Sierra Nevada crest, mainly in an area drained by the North and South Forks of Oak Creek. The main mass of the pluton within the range is about 5 miles long, but outliers east of the range front show that the pluton may be twice as large as the 8.3 square miles of mapped area. The pluton is named after McGann Springs in its northeast part.

Most of the eastern part of the McGann pluton is exposed in rounded foothills along the range front and consequently the outcrops are poor. A large group of east-dipping aplite dikes which strike north to northeast cut across the eastern part of the pluton. In much of this country the aplite dikes are more resistant than the coarser granitic rock, and nearly all the outcrops on ridge crests are aplite. The McGann pluton is composed of a quartz-poor quartz monzonite (fig. 42 and table 19) which is unlike the rock in any other pluton. The average quartz content is 17.2 percent. Mafic inclusions are relatively common and average 2.4 per square yard. The McGann pluton is in contact to the north with a complex of mafic granodiorites, diorite, gabbro, and anorthosite, and its low content of quartz may be due to contamination.

TABLE 19.—*Modes of the McGann pluton*

Sample No.	Quartz	K-feldspar	Plagioclase	Biotite	Hornblende	Opaque accessories	Nonopaque accessories	Color index	Sp Gr	Percent An in plagioclase
1	20.9	33.8	34.8	6.3	3.3	0.8	0.1	10.5	2.65	--------
2	17.6	42.7	29.5	7.6	1.4	.5	.6	10.1	2.65	28
3	12.2	37.4	38.8	5.9	3.1	.6	2.0	11.6	2.65	31
4	16.3	35.6	32.5	10.0	3.4	1.1	1.2	15.7	2.60	--------
5	14.4	33.4	42.2	7.4	1.9	.3	.3	9.9	2.67	29
6	14.2	33.8	35.8	8.6	5.8	.3	1.4	16.1	2.67	28
7	15.1	37.2	39.2	6.6	1.2	.3	.4	8.5	2.66	--------
8	23.1	29.7	34.2	8.2	2.8	.7	1.4	13.1	2.65	27
9	21.4	36.1	35.8	5.7	.1	.4	.6	6.8	2.60	28
Av	17.2	35.5	35.9	7.4	2.6	.6	.9	11.5	2.64	28
SD	3.8	3.6	3.8	1.4	1.6	.3	.6	3.1	--------	--------

The McGann pluton contains no mafic dikes and is assumed to be younger than the dikes; it is in contact on all sides except the east with rocks which are cut by the dikes. It is therefore assumed to be younger than the mafic complex to the north and the Independence pluton to the southwest. On its western margin the McGann pluton is in contact with a mafic facies of the McDoogle pluton. Here the quartz monzonite of the McGann pluton is clearly younger than that of the McDoogle, since it sent dikes and tongues into the McDoogle pluton.

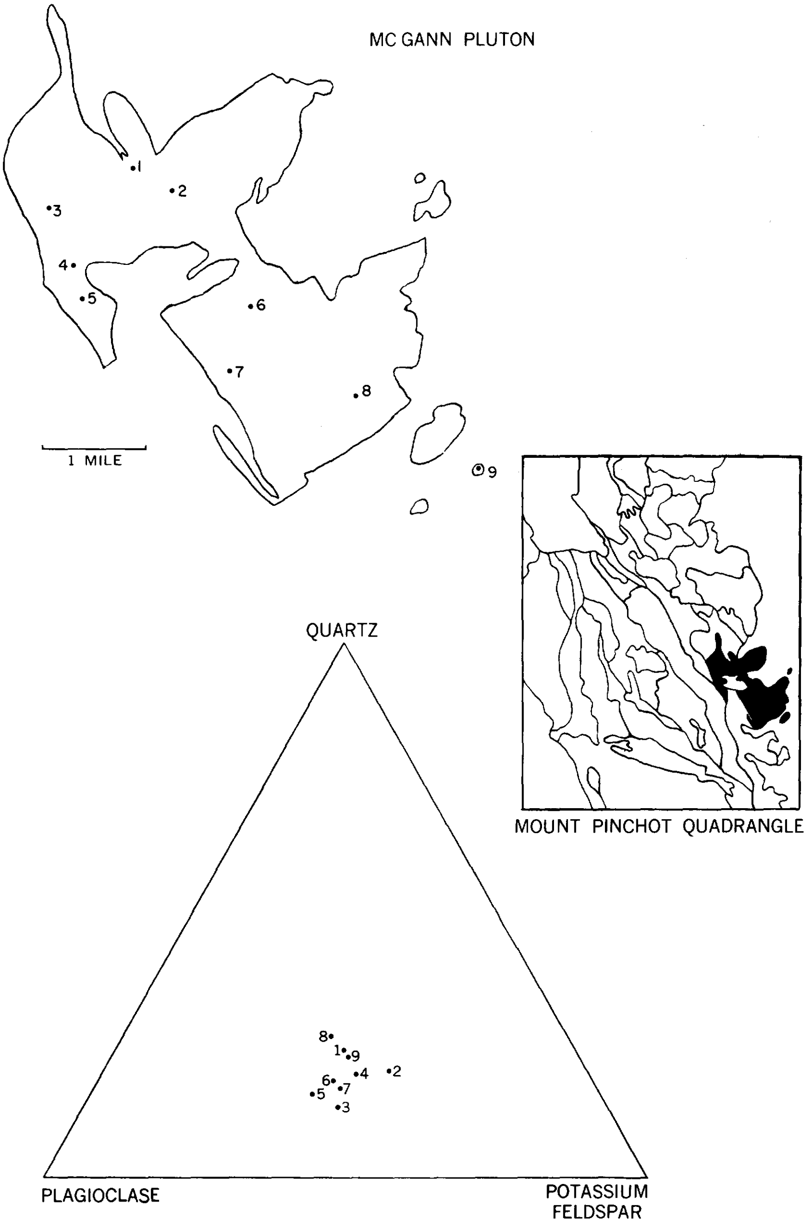

FIGURE 42.—Location and modal composition of samples from the McGann pluton. Numbers refer to samples given in table 19.

GOODALE PLUTON

The Goodale pluton is a small body (2.3 square miles) in the northern part of the quadrangle. The mass extends east from Goodale Mountain (from which it was named) to the foot of the range west of Aberdeen. On its southeast side the pluton is intimately mixed with the dioritic mass of Goodale Creek and is in fact an extensive contact breccia. Immense masses of dioritic rock (commonly hundreds of feet long) are present in the breccia.

The Goodale pluton contains slightly less than the usual amount of quartz (fig. 43 and table 20). The low quartz content has been interpreted in other plutons (Bullfrog, Dragon, McGann, and McDoogle plutons) as having been caused by contamination by (or assimilation of) mafic wall rock and this explanation seems also applicable to the Goodale pluton. The rock is characterized by abundant large mafic inclusions (4.4 per sq. yd.).

TABLE 20.—*Modes of the Goodale pluton*

Sample No.	Quartz	K-feldspar	Plagioclase	Biotite	Hornblende	Opaque accessories	Nonopaque accessories	Color index	Sp gr
1	23.4	32.2	32.5	6.8	3.9	0.5	0.7	11.9	2.66
2	22.9	32.1	33.9	7.0	3.6	.3	.2	11.1	2.66
3	19.5	32.0	36.1	8.0	3.7	.6	----------	12.3	2.63
Av	21.9	32.1	34.2	7.3	3.7	.5	.3	11.8	2.65

The Goodale pluton contains no mafic dikes and can be seen transecting the mafic dikes on its western contact with the older Red Mountain Creek pluton. Hence it is younger than the Red Mountain Creek and Taboose plutons—and a dioritic mass to the south—all of which contain mafic dikes. The Siberian pluton to the southwest is intrusive into the Goodale pluton and therefore younger.

ALASKITE PLUTONS

Six plutons have been called alaskite because of their low content of mafic minerals. The average of specimens from each mass contains less than 3 percent mafic minerals, except from the Diamond pluton. The Diamond pluton is inadequately sampled, but field appearance and high content of K-feldspar suggest that it is an alaskite. In the classification used, all the alaskites fall in the quartz monzonite field and close to the minimum melting trough of the system albite–K-feldspar–quartz at 1,000 bars H_2O pressure (Bowen, 1954). The alaskites contain about equal amounts of K-feldspar, quartz, and calcic albite or sodic oligoclase. Four of the masses—the Red Mountain Creek, Diamond, Twin Lakes, and Independence plutons—are cut by mafic dikes. The other two, the Bullfrog and Sardine plutons, are presumably younger than the mafic dikes for they contain no dikes.

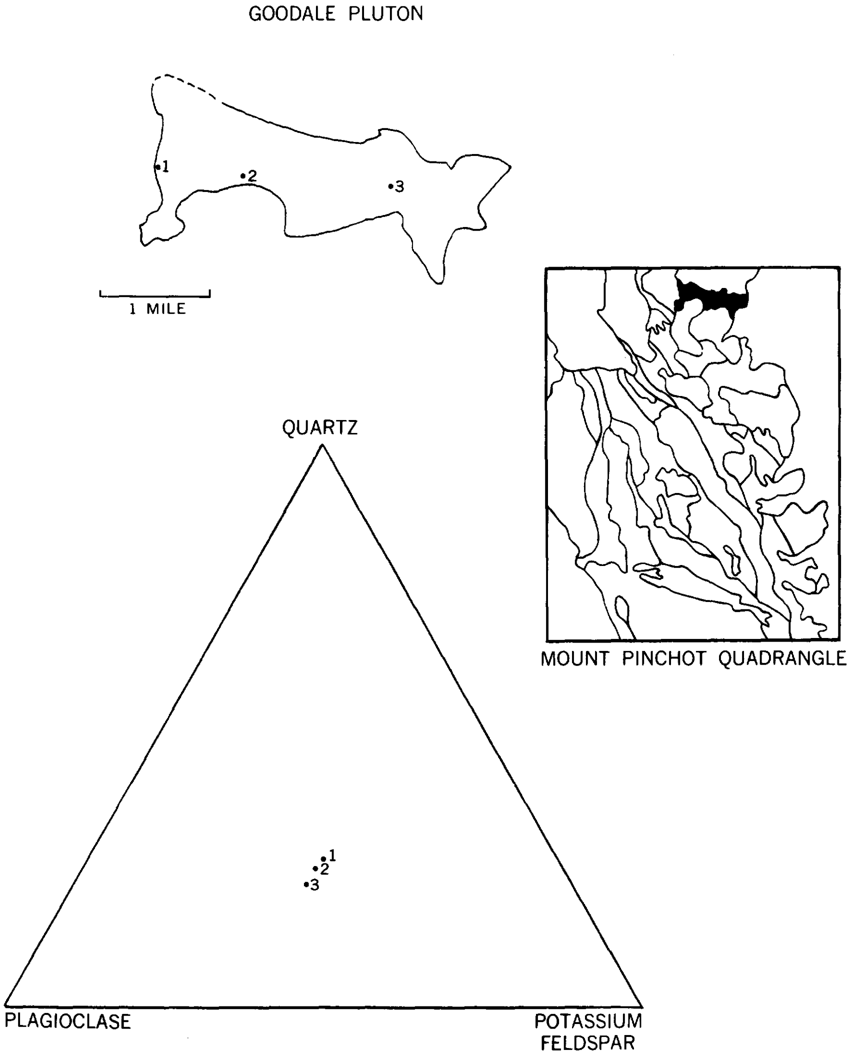

FIGURE 43.—Location and modal composition of samples from the Goodale pluton. Numbers refer to samples given in table 20.

RED MOUNTAIN CREEK PLUTON

The Red Mountain Creek pluton occupies 2.7 square miles in the north-central part of the quadrangle. The pluton lies about half in the Mount Pinchot quadrangle and half in the Big Pine quadrangle to the north. The pluton was named after Red Mountain Creek, the next large east-flowing stream north of Taboose Creek, and is grouped with the rocks similar to the Cathedral Peak granite by Bateman (1961).

The Red Mountain Creek pluton consists of quartz monzonitic alaskite (color index, 0.4) (fig. 44, table 21). It contains few mafic inclusions, and in some large areas it contains none. In the southern part of the pluton the alaskite has been locally albitized and contains no K-feldspar.

TABLE 21.—*Modes of Red Mountain Creek pluton*

Sample No.	Quartz	Perthite	Plagio-clase	Biotite	Opaque acces-sories	Color index	Sp gr	Percent An in plagio-clase
1	29.3	39.3	31.3	0.1	----------	0.1	2.60	16
2	36.2	34.8	28.4	.4	0.3	.7	2.60	15
Av	32.8	37.0	29.8	.3	.2	.4	2.60	16

Biotite schist forms a septum on the west side of the pluton and arches over the center of the pluton to form a nearly horizontal roof. This roof is well exposed on the summit of Cardinal Mountain, and it may be seen on the ridge north of Cardinal Mountain from Highway 395 in Owens Valley. The schist septum continues north of the quadrangle boundary for about 1 mile, but a little south of Split Mountain it turns abruptly east and continues to Owens Valley. The occurrence of similar biotite schist on Shingle Mill Bench and Stecker Bench, north of the quadrangle, suggests that the metamorphic septum completes its arch over the Red Mountain Creek pluton, which must be only slightly eroded. The main mass of the Tinemaha granodiorite overlies the schist septum that arches over the Red Mountain Creek pluton in the Big Pine quadrangle. Therefore P. C. Bateman (oral communication) has considered the Tinemaha granodiorite older than the Red Mountain Creek pluton. The Woods Lake mass of the Tinemaha granodiorite is also considered older than the Red Mountain Creek pluton.

The pluton is cut in its western part by many large mafic dikes of the Independence swarm. The dikes are well exposed at Taboose Pass (Knopf, 1918, p. 71), where they contrast vividly with the white alaskite.

The Red Mountain Creek pluton is older than all granitic rocks with which it is in contact. Mafic dikes that cut the Red Mountain Creek alaskite and the biotite schist to the west terminate at the contact with the Lamarck granodiorite. This relation can be seen on the mountain wall south of Cardinal Lake at the north border of the quadrangle. The Red Mountain Creek pluton is older than the Taboose pluton on its east. The Taboose is finer grained and contains schlieren near the contact, whereas the Red Mountain Creek alaskite is unchanged near the contact. Both are cut by mafic dikes. The

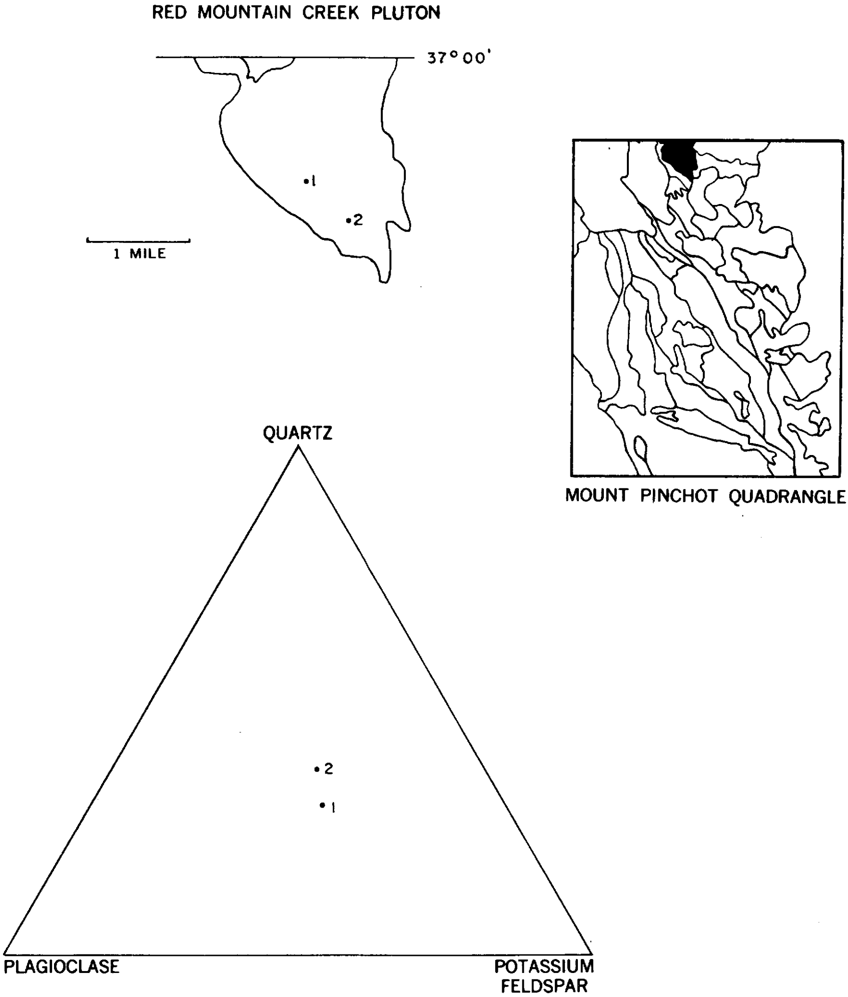

FIGURE 44.—Location and modal composition of samples from the Red Mountain Creek pluton. Numbers refer to samples given in table 21.

Red Mountain Creek pluton is older than either the Goodale or Siberian plutons, for both these plutons clearly cut off the mafic dikes that cut the alaskite of the Red Mountain Creek pluton.

INDEPENDENCE PLUTON

The Independence pluton, named for Independence Creek which flows through its center, is situated in the southeast corner of the quadrangle at the base of the east scarp of the Sierra Nevada. The pluton extends south from Charlie Canyon across the South Fork of Oak Creek, Independence Creek, and Pinyon Creek to the south boundary of the quadrangle. It is 6 miles long and occupies 8.3 square miles within the quadrangle; its extent to the south is not known. The pluton is generally poorly exposed over much of its area because of its location in the foothills.

The Independence pluton is composed of a fairly uniform alaskitic (color index, 2.7) rock, the average of which plots within the quartz monzonite field very close to the granite boundary (fig. 45, table 22). Mafic inclusions are very rare. The Independence pluton, like the Tinemaha granodiorite, contains abundant inclusions of identifiable wallrock, many of them sheetlike. Northeast of Little Onion Valley blocks and septa of meta-andesite and metarhyolite are present in the alaskite, and south of Seven Pines meta-andesite septa are abundant. Intimate mixtures of alaskite and hornblende diorite with metavolcanic rocks can be seen in the road cuts on the new road along Independence Creek above Seven Pines. The entire complex is cut by mafic dikes. Moreover, much of the Independence pluton is severely shattered, especially adjacent to the Independence fault.

TABLE 22.—*Modes of Independence pluton*

Sample No.	Quartz	K-feldspar and perthite	Plagioclase	Biotite	Opaque accessories	Nonopaque accessories	Color index	Sp gr	Percent An in plagioclase
1	23.5	40.3	29.8	5.1	0.9	0.5	6.5	2.64	----------
2	35.1	32.6	28.7	3.0	.6	.1	3.7	2.58	24
3	26.4	46.4	25.8	.9	.4	----------	1.3	2.56	11
4	31.1	46.2	19.9	.5	1.3	.9	2.7	2.58	10
5	36.2	43.6	20.1	.1	.1	----------	.2	2.60	----------
6	27.4	47.2	23.9	.9	.4	.2	1.5	2.59	11
Av	30.0	42.7	24.7	1.8	.6	.3	2.7	2.59	14
SD	5.0	5.6	4.2	1.9	.4	.4	2.3	----------	----------

The Independence pluton is in intrusive contact with only one other body, the McGann pluton to the north. Age relations on this contact were not observed, mainly because of the poor exposures. However, the Independence pluton is cut by mafic dikes, whereas the McGann pluton is not. On this evidence the Independence pluton is considered older than the McGann pluton. The Independence pluton is in fault contact with the Tinemaha and relative ages are uncertain inasmuch as both intrusions are cut by mafic dikes.

TWIN LAKES PLUTON

The Twin Lakes pluton is a small (1.2 square miles) mass of alaskite near the center of the quadrangle. The pluton extends from the John Muir Trail near Twin Lakes northwest to the Cartridge pluton. It is composed of a medium-grained quartz monzonite (fig. 46 and table 23) alaskite which is generally slightly altered. In thin section, the rock can generally be seen to be sheared and the biotite chloritized.

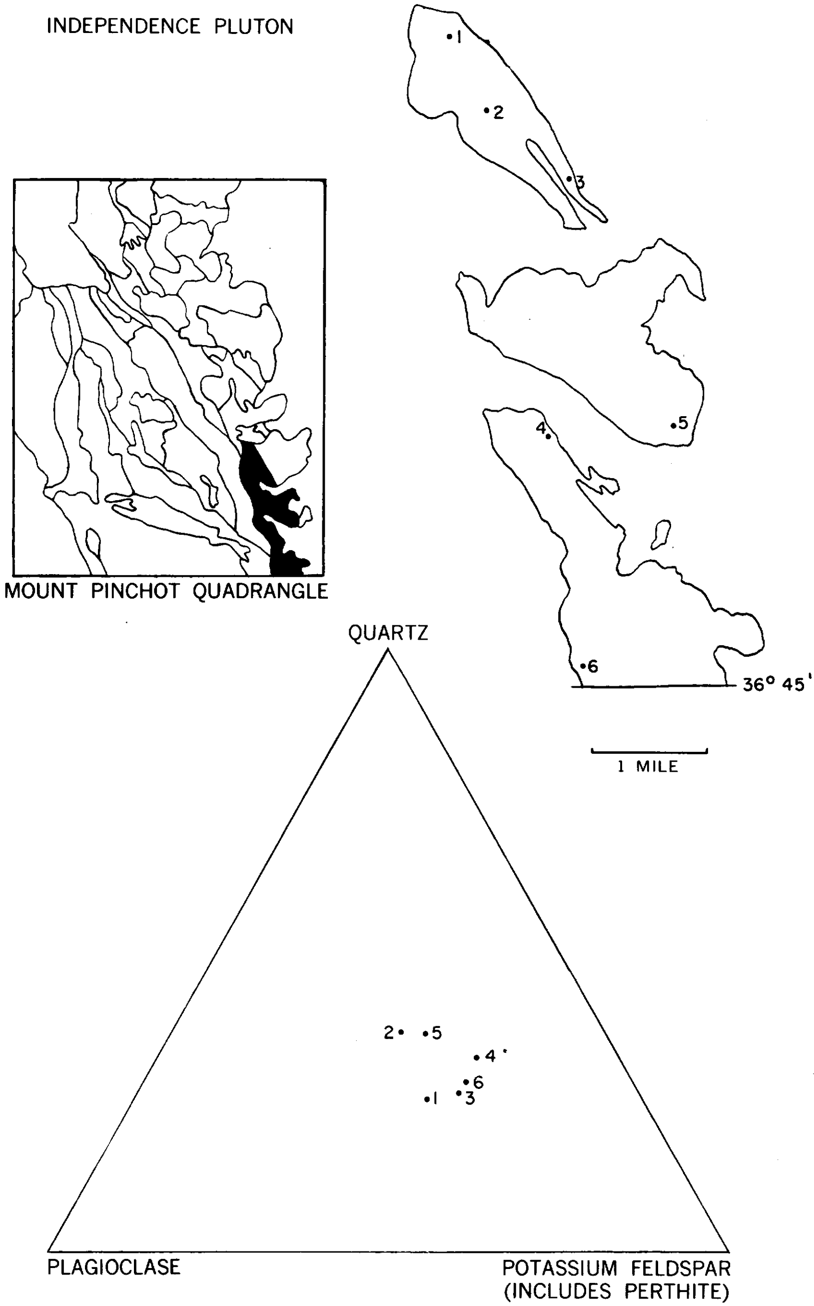

FIGURE 45.—Location and modal composition of samples from the Independence pluton. Numbers refer to samples given in table 22.

The Twin Lakes pluton is in contact with the Woods Lake mass of the Tinemaha granodiorite on the southeast and with sheared por-

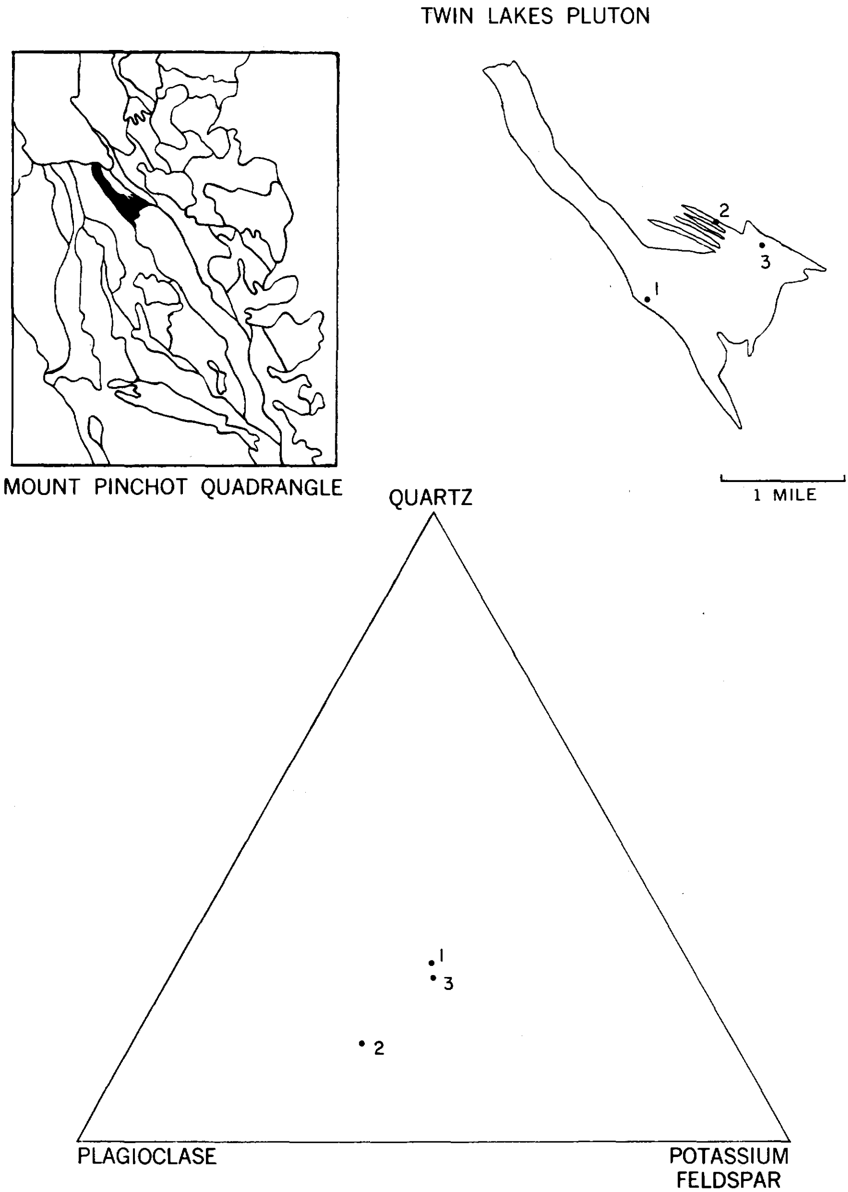

FIGURE 46.—Location and modal composition of samples from Twin Lakes pluton. Numbers refer to samples given in table 23.

phyritic mafic granodiorite or quartz diorite to the northeast. The alaskite is intrusive into both the Tinemaha granodiorite and the mafic granodiorite, and it forms intimate mixtures with both. These mixtures are best developed along the John Muir Trail in the southeastern part of the Twin Lakes pluton. Here the Twin Lakes pluton

TABLE 23.—*Modes of Twin Lakes pluton*

Sample No.	Quartz	Perthite	Plagioclase	Biotite	Opaque accessories	Nonopaque accessories	Color index
1	27.6	35.7	35.8	0.4	0.3	0.2	0.9
2	14.7	30.4	49.7	4.2	.8	.1	5.1
3	25.7	36.1	36.3	1.6	.2	.1	1.9
Av	22.7	34.1	40.6	2.1	.4	.1	2.6

is intimately mixed with northwest-trending sheets of foliated mafic granodiorite, with inclusions of Tinemaha granodiorite, and with irregular pods of metasedimentary rocks, mainly calc-hornfels. The Twin Lakes pluton is cut by mafic dikes and is older than the Cartridge Pass pluton, which cuts off both the pluton and the dikes.

BULLFROG PLUTON

The Bullfrog pluton is the largest in the quadrangle. It covers 22.9 square miles in the south-central part of the quadrangle. The major part of the pluton lies south of the east-trending dioritic complex on the divide between Bubbs Creek and Woods Creek, and extends south of the quadrangle boundary an unknown distance. One portion of the pluton breaks through the dioritic complex near Glen Pass and extends north 6 miles; it is exposed on King Spur and part of Castle Domes north of Woods Creek. The pluton is named after Bullfrog Lake, near the central part of the southern mass. The major part of the pluton is composed of a fairly uniform perthitic quartz monzonite, which is quite light-colored (color index, 3.0) and is called an alaskite (fig. 47 and table 24). The rock is composed of nearly equal amounts of perthite (with 30 volume percent of albite stringers), plagioclase (An_{10}), and quartz.

Two facies that are somewhat different from the dominant rock have been noticed: (1) In the southwestern part of the pluton, west of the metarhyolite mass of Mount Bago, the Bullfrog alaskite is notably fine grained and slightly poorer in K-feldspar. (2) In the southeastern part of the pluton, east of Kearsarge Pass, a facies of the alaskite of the Bullfrog pluton is very poor in quartz (table 24), averaging less than 10 percent. This rock is also somewhat higher in mafic minerals, especially hornblende. The presence of this quartz-poor rock near the dioritic complex bordering it to the north suggests that the low SiO_2 content was caused by some type of contamination by the dioritic complex.

The Bullfrog pluton is poor in mafic minerals and contains almost no mafic inclusions. The rare inclusions that occur are very small, generally less than 1 inch in length. Consequently movement in the pluton is not recorded, and internal structural measurements are diffi-

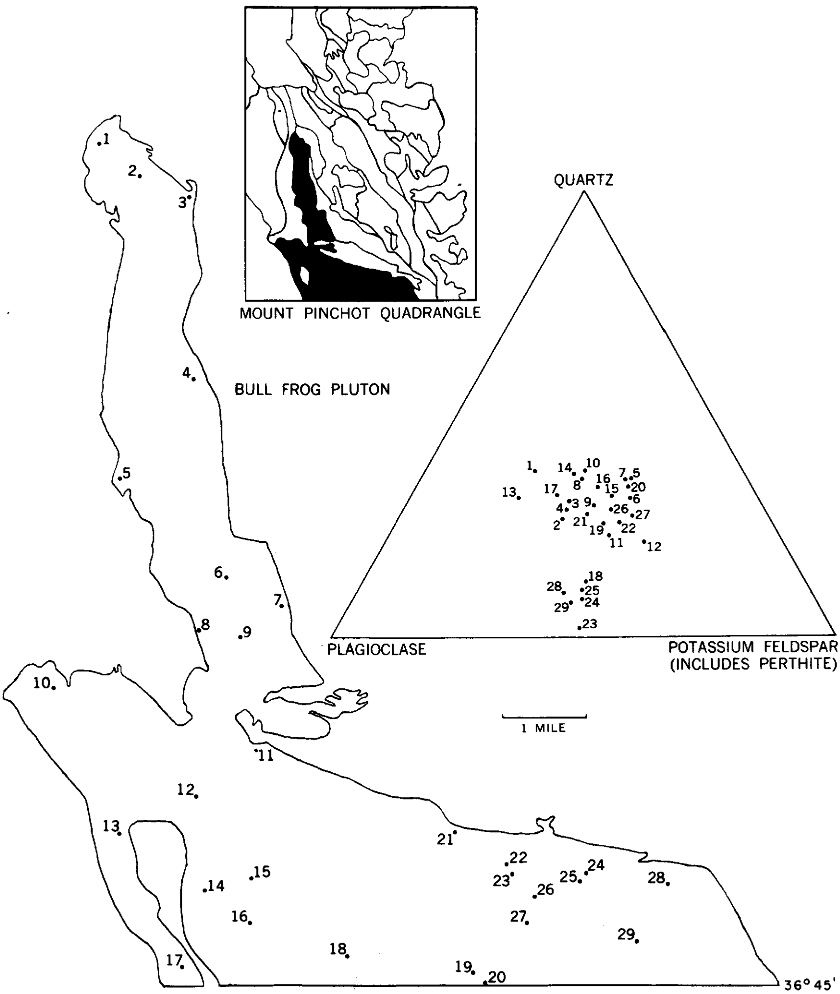

FIGURE 47.—Location and modal composition of samples from Bullfrog pluton. Numbers refer to samples given in table 24.

cult to make. In some places, weathering of the alaskite etches out a foliation which is apparently the result of a semiparallel arrangement of feldspar crystals.

The Bullfrog pluton characteristically shows very coarse jointing. The individual joint block may be up to 50 feet across. As a result, the rock is exceptionally resistant to glacial erosion by frost wedging and plucking and forms bold highlands and domes. Castle Domes, Fin Dome, King Spur, Mount Clarence King, Kearsarge Pinnacles, and Independence Peak all owe their elevation and form to the resistance of the rock in the Bullfrog pluton. University Peak (elevation 13,588 feet), just south of the southern quadrangle boundary,

TABLE 24.—*Modes of Bullfrog pluton*

Sample No.	Quartz	Perthite	Plagioclase	Muscovite	Biotite	Hornblende	Opaque accessories	Nonopaque accessories	Color index	Sp. gr	Percent An in plagioclase
1	35.3	21.4	40.7	--------	1.5	0.6	0.5	--------	2.6	2.60	--------
2	25.5	33.8	38.0	--------	2.6	--------	.7	0.1	3.4	2.61	--------
3	30.0	30.8	36.2	0.1	2.6	--------	.4	--------	3.1	2.60	16
4	27.1	29.0	41.2	--------	1.5	--------	.3	.8	2.6	2.63	--------
5	34.4	41.1	22.8	--------	.8	--------	.5	.3	1.6	2.60	--------
6	30.2	43.3	25.5	--------	.8	--------	.2	--------	1.0	2.61	--------
7	34.2	39.0	26.2	--------	.4	--------	.1	.1	.6	2.61	10
8	33.5	30.2	31.0	--------	2.7	--------	1.3	1.3	5.3	2.59	--------
9	26.5	36.0	33.5	--------	1.1	--------	.3	.6	2.0	2.59	--------
10	35.4	31.1	30.8	--------	1.4	.2	.8	.3	2.7	2.59	9
11	22.1	42.6	32.9	--------	1.9	--------	.3	.1	2.3	2.61	--------
12	20.3	51.0	26.8	--------	1.7	--------	.1	.1	1.9	2.58	--------
13	30.6	21.0	48.0	--------	1.6	--------	.4	.3	2.3	2.61	--------
14	35.8	31.3	31.0	--------	1.7	--------	.2	.1	2.0	2.58	--------
15	30.7	38.7	28.5	--------	1.2	--------	.7	.3	2.2	2.58	--------
16	32.4	35.3	30.1	--------	1.4	--------	.6	.3	2.3	2.61	6
17	30.4	27.9	38.4	--------	1.4	--------	1.4	.4	3.2	2.58	12
18	10.7	43.4	43.6	--------	.8	.1	.6	.5	2.3	2.58	6
19	24.8	39.6	32.1	.1	2.0	--------	.7	.7	3.5	2.59	--------
20	33.4	40.6	24.7	.1	.5	--------	.2	.4	1.2	2.61	8
21	27.1	36.0	34.6	--------	1.1	--------	1.0	.2	2.3	2.58	--------
22	25.0	44.2	29.6	--------	1.0	--------	.1	.3	1.4	2.57	--------
23	1.7	47.2	49.2	--------	1.2	--------	.6	.2	2.0	--------	--------
24	7.9	42.6	44.4	--------	3.4	.7	.9	.2	5.2	2.58	--------
25	8.8	42.1	43.2	--------	4.7	.3	.6	.3	6.9	--------	--------
26	27.4	40.5	30.4	--------	.5	.6	.6	.1	1.8	2.59	10
27	25.8	45.6	27.2	--------	.7	.2	.5	.1	1.5	2.57	13
28	8.9	38.2	44.8	--------	6.2	.2	.8	1.0	8.2	2.62	--------
29	7.3	39.6	44.9	--------	3.7	1.6	1.6	1.2	8.1	2.62	--------
Av	25.0	37.3	34.7	Tr.	1.8	.2	.6	.4	3.0	2.60	10
SD	10.0	7.2	7.5	--------	1.3	.3	.4	.3	2.0	--------	--------

is within the Bullfrog pluton and is higher than any point in the Mount Pinchot quadrangle.

The perthite grains in the rock of the Bullfrog pluton are 3 to 6 mm long and are generally the largest crystals. In gross outline the grains are euhedral to subhedral, though upon close inspection the margins are found to be crenulate and irregular (fig. 48). The K-feldspar host commonly shows microcline twinning over a considerable part of its area. X-ray measurements show it to be dominantly triclinic. Albite, other than perthitic lamellae, forms small grains and irregular veins on the borders of the perthites. It also occurs in large oscillatory zoned crystals comparable in size and abundance to the perthite crystals. The small discrete grains of albite are abundant along intergrain boundaries between perthite and large albite crystals and between two perthite crystals, but seldom appear between perthite and quartz. The albite stringers and patches in the cores of the perthite grains are generally thick and well developed, but there is a distinct narrowing and disappearance of the albite stringers toward the edge of the perthite crystal, giving an overall effect of zoning.

Eight specimens collected from localities scattered over the granitic body were studied (figs. 47 and 49). Thin sections were stained with sodium cobaltinitrite after a hydrofluoric acid etch, in the manner described by Chayes (1952). The resulting yellow stain on the K-

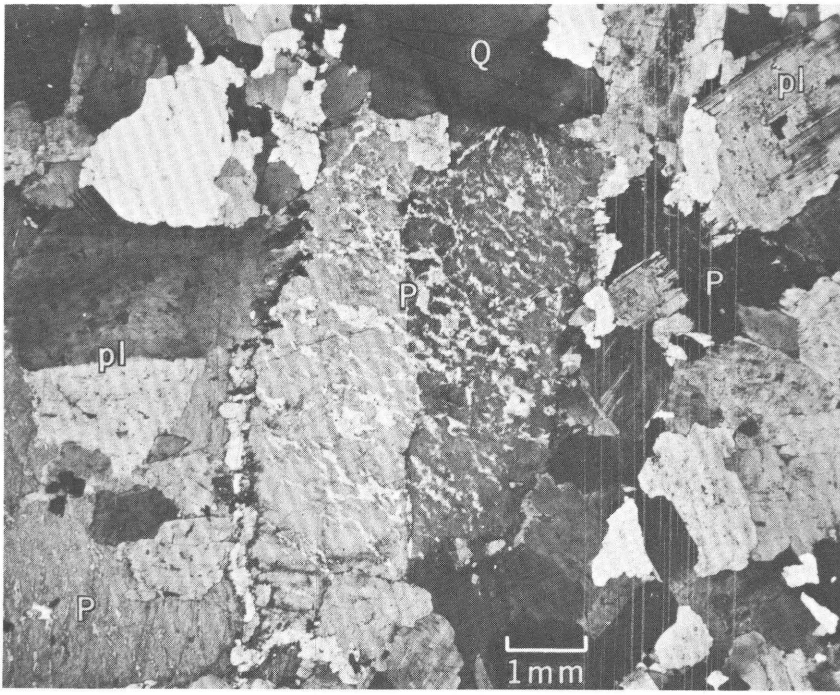

FIGURE 48.—Photomicrograph of alaskitic quartz monzonite from the Bullfrog pluton. P, perthite; Q, quartz; and pl, plagioclase. Thinning of albite stringers near edge of perthite crystals forms outer zone poor in albite. Small crystals on boundary of perthite are albite; none are present where perthite is in contact with quartz (top). Faint oscillatory zoning is visible in plagioclase crystal to left. Quartz (top) is cut by two sets of irregular shears into a mosaic of small rhombs. ×14. Crossed nicols.

feldspar is so selective that it leaves even the finest lamella of albite (less than 0.1 mm in width) unstained. A special point counter with an intercept of 0.05 mm was used to determine the relative amounts of albite and K-feldspar in individual perthite grains. In each slide about nine of the largest perthite grains were measured for their percent of albite. A total of 67 perthite grains were measured and an average of 450 points were counted and identified within each grain (table 25). The counting error is probably less than 3 percent in any grain.

Most of the albite lamellae in a single perthite grain are in optical continuity. Plagioclase grains not in optical continuity are assumed to be inclusions in the alkali feldspar and unrelated to the formation of the perthite. They have the same composition as the large plagioclase crystals (10 percent anorthite) and hence appear to be unrelated to the albite lamellae (5 percent anorthite). Such foreign plagioclase inclusions were omitted from the analyses.

The volume percent of albite ranges from about 20 to 40 percent and averages approximately 30 percent (table 25). Ninety percent of the

TABLE 25.—*Albite content of perthite grains*

[Samples obtained in localities shown in fig. 49]

Sample No.	Points per grain	Points in albite	Percent albite	Sample No.	Points per grain	Points in albite	Percent albite
15	163	48	29.3		535	164	30.7
	185	59	31.9		622	198	31.8
	215	54	25.1	24	177	64	36.1
	230	54	23.5		199	60	30.1
	247	79	31.9		226	86	38.0
	389	102	26.3		261	76	29.1
	672	216	32.2		490	130	26.6
19	89	22	24.8		491	133	27.1
	166	33	19.8		853	284	33.3
	186	72	38.7	26	142	40	28.1
	267	62	23.2		168	43	25.6
	296	90	30.4		307	97	31.5
	313	94	30.0		307	103	33.6
	349	100	28.7		310	116	37.4
	449	126	28.1		446	106	23.7
	458	105	22.9		474	132	27.9
	709	239	33.8		544	178	32.8
	740	233	31.5		568	214	37.7
20	161	37	22.9		665	245	36.9
	193	37	19.1	27	160	37	23.1
	268	73	27.2		264	77	29.1
	319	83	26.0		277	105	37.9
	360	108	30.1		300	88	29.3
	529	139	26.3		313	48	15.3
	554	131	23.7		373	120	32.2
	584	191	32.8		603	207	34.4
22	170	48	28.3		691	230	33.3
	226	76	33.6	29	356	107	30.1
	304	78	25.6		370	74	20.0
	382	143	37.4		470	132	28.1
	423	119	28.2		474	121	25.5
	490	176	36.0		711	243	34.2
	493	148	30.0		917	276	30.1
	504	200	39.7				

perthite grains have between 38 and 22 percent albite. Hence the percent of albite in perthite grains is reasonably constant over many square miles. The average percent of albite does not differ more than 5 percent from specimen to specimen. Figure 49 is a scatter diagram showing the percent albite in each perthite grain; the grains of each specimen are grouped together. The compositional range of perthite grains in each specimen is only slightly smaller than the range in composition of perthite grains over the entire area studied. The largest grain in each specimen in nearly every case has more albite than the smallest as shown in fig. 49. Also those specimens nearest the intrusive contact (samples 22, 24, 26) have the greatest amount of perthitic albite. The fairly constant albite: K-feldspar ratio in perthite over a large area indicates an exsolution origin for the perthite. The amount of albite exsolved from an originally homogeneous alkali feldspar would be fixed by the amount of soda contained in solid solution. Hence exsolution would free the same proportion of albite in every perthite grain provided they all have the same original composition and the same cooling history. The fairly constant amount of albite in the K-feldspar indicates that the perthite is not of replacement origin.

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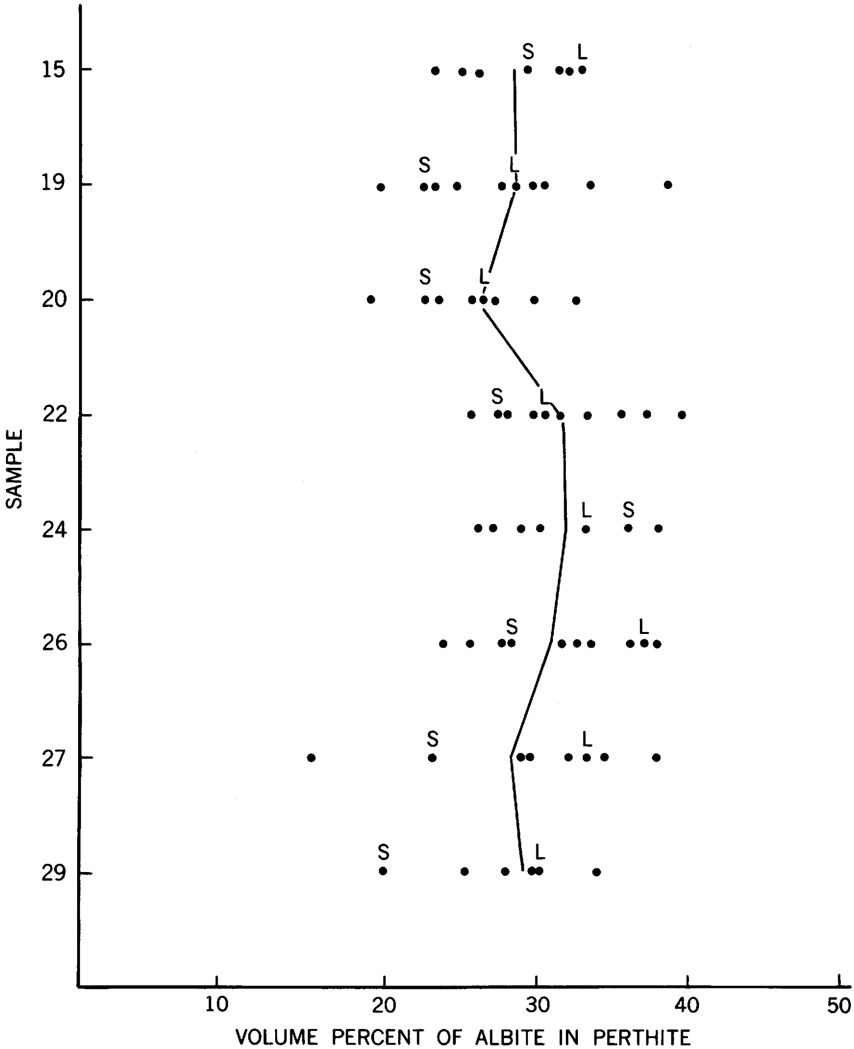

FIGURE 49.—Scatter diagram of perthitic albite (volume percent) in perthite crystals from eight specimens of alaskitic quartz monzonite of the Bullfrog pluton. Solid dots, individual grains; line connects specimen averages. The smallest and largest grains in each specimen are marked S and L, respectively. Sample numbers are the same as those given on figure 48 and table 24.

In addition to simple exsolution, however, other processes played a role in forming the perthites. The obvious decrease of albite near the edge of the perthite grains has been previously mentioned. This zoning is indirectly confirmed by the modal analyses, since large grains usually carry more albite than smaller ones. Assuming that all perthite grains in the rock are of about the same size, then the small grains in the thin sections are slices through the edges and corners of large grains. Such peripheral slices would contain less albite. In this way a part of the spread of figure 49 may be explained. If measurements were made through the center of every perthite grain, there would presumably be a smaller spread of values, and a slightly higher mean value of contained albite.

Does this perthitic zoning indicate unmixing of an alkali feldspar whose core was richer in albite? Laboratory investigations demonstrate (Bowen and Tuttle, 1950) that in a cooling environment, alkali feldspars (with approximately 30 percent albite and 70 percent K-feldspar) should contain cores richer in K-feldspar, not albite. Hence to explain the zoning of the perthites as inherited from an originally zoned alkali feldspar requires the postulation of an abnormal thermal history.

The hypothesis that best explains the zoning is that the perthites have reached an advanced stage of exsolution of an original homogeneous sanidine. Tuttle (1952, p. 113) shows that an original sanidine crystal undergoes several steps in the process of exsolution. The alkali feldspars of many salic rocks can be grouped into an "exsolution series" consisting of seven stages: (1) sanidine or anorthoclase (homogeneous crystal), (2) sub-X-ray perthite, (3) X-ray perthite, (4) cryptoperthite, (5) microperthite, (6) perthite, (7) albite and orthoclase (or microcline). In passing from stage 1 to stage 7, unmixing separates the two phases and concentrates each in larger and larger entities, until finally the albite migrates completely out of the original borders of the K-feldspar crystal. The perthites of the Bullfrog pluton are believed to represent the beginnnig of transition from stage 6 to stage 7. Albite from near the borders of the perthite crystals has migrated and consolidated along the intergranular boundaries, leaving an outer zone of albite-poor material. The albite migrates preferentially toward perthite-plagioclase or perthite-perthite boundaries (fig. 48). Why it avoids the quartz boundaries is not known. Much of the albite appears to replace the outer zone of large plagioclase crystals, thus forming an outer, albite-rich selvage. Those specimens collected near the intrusive contact of the Bullfrog pluton presumably contain more albite because the migration of albite out of the perthite crystals was partially arrested due to more rapid cooling at the margin of the pluton.

The Bullfrog pluton, though the largest in the quadrangle, is in contact with relatively few other plutons. It is in contact with the dioritic complex along the east-west drainage divide between Bubbs Creek and Woods Creek. Along this contact the Bullfrog pluton has sent flat-dipping dikes into the mafic complex on a large scale. In many places the bulk of the rock in the complex is actually Bullfrog alaskite belonging to these sheets. The hornblende diorite of the complex forms large sill-like slabs in the alaskite of the Bullfrog pluton near the south end of Sixty Lake Basin. Actually the alaskite is intrusive into the diorite, which has been so completely split apart by the alaskite that it occurs in isolated sheets. The diorite slabs dip gently south, 0° to 30°. There has been extensive feldspathization and replacement of the diorite by alaskite, and many intermediate products are present. The Bullfrog pluton is younger, as shown by these intrusive relations; in many places, however, the contact has been altered by postintrusion shears and may represent a fault zone similar to the east-west faults to the north.

The contact between the Bullfrog and Dragon plutons is well exposed on the divide between Rae Lakes and Sixty Lake Basin near Fin Dome. In most places where it was seen, this contact is gradational over 1 to 30 feet. Locally, dark schlieren separate the two plutons. South of Fin Dome the contact is gradational over 1 to 10 feet, but the Bullfrog appears to be intrusive into the Dragon because the north-south contact truncates the N. 40° E. trending internal foliation of the Dragon pluton; also, rare N. 7° E. trending schlieren in the Bullfrog near the contact are parallel to the contact. The relative ages of the Bullfrog and Paradise plutons is not known. The two bodies are separated along a 4-mile front by a metarhyolite septum a few hundred feet thick.

SARDINE PLUTON

The Sardine pluton intrudes the Oak Creek metavolcanic pendant in the southeastern part of the quadrangle. The pluton is exposed in Sardine Canyon, one-half mile east of Sardine Lake, whence it takes its name. The Sardine pluton is small (0.3 square mile) and is composed of fairly uniform alaskitic quartz monzonite (fig. 50). It is exposed in two masses, one in Sardine Canyon and the other in the South Fork of Oak Creek Canyon. The two exposed masses are separated in plan by a thin remnant of the roof, 300 feet wide and about 100 feet thick, on the high ridge separating the two canyons.

The Sardine pluton is conformable and has in general intruded along the contact between metarhyolite tuff and a metabasalt flow. It has made room for itself primarily by wedging apart the metavolanic wall rocks, but on its southern end the pluton feathers out and sends innumerable dikes into the metavolcanic wall rock. The regional steep west dip of the metavolcanic rocks is locally changed to a steep east dip on the east side of the pluton. The contact of the Sardine with the metavolcanic rocks is generally very sharp. The alaskite has an aphanitic chilled margin, and the metabasalt is recrystallized to a coarse spotted amphibolite in a zone several inches wide at the contact.

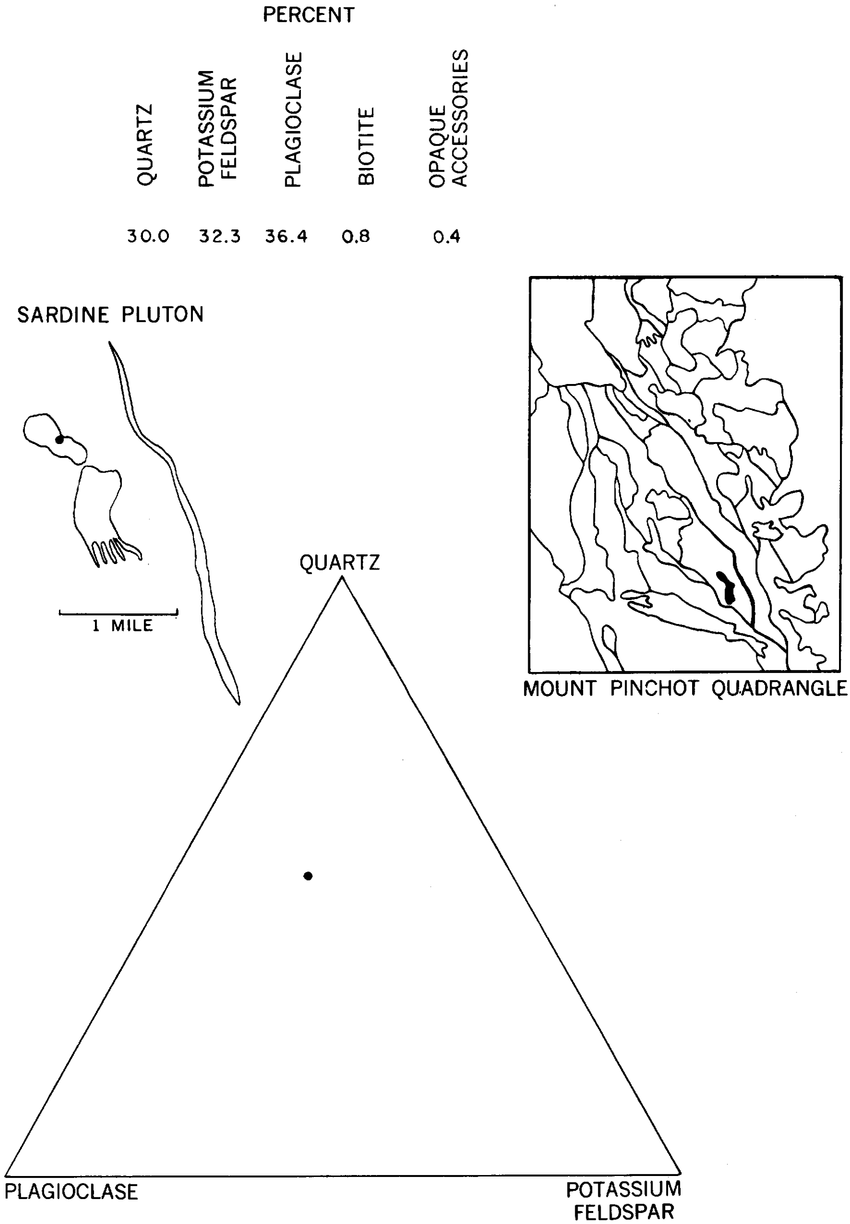

FIGURE 50.—Location and modal composition of one sample from the Sardine pluton. Color index is 1.2.

One-half mile east of the Sardine pluton is a dike of alaskite 3 miles long and up to 500 feet thick, which is also largely within the Oak Creek metavolcanic pendant. Because of the proximity of this dike to the Sardine pluton and the apparent lithologic similarity, it has been tentatively correlated with the Sardine pluton.

The small Sardine pluton is difficult to date because of its isolated position, entirely enclosed by metavolcanic rock. However, several relations bear on its age. The dike with which it has been tentatively correlated is intrusive into the Woods Lake mass of the Tinemaha granodiorite. Furthermore, neither dike nor pluton contain mafic dikes of the Independence swarm and hence are presumed younger than the dikes.

The Sardine pluton is considered the source of the sulfide and gold mineralization of the Kearsarge Peak area since the veins are localized around its southern end. Since some of these veins are in the Dragon pluton, the Sardine is younger than the Dragon. The upper age limit of the Sardine is not fixed. It could be the youngest intrusion in the quadrangle.

DIAMOND PLUTON

The Diamond pluton is a small (1.3 square miles), irregular mass of alaskite in the south-central part of the map. The Diamond pluton was first believed to be part of the Bullfrog pluton, which it greatly resembles petrologically, but the discovery of some mafic dikes cutting it and also the fact that it is older, rather than younger, than the Dragon pluton demonstrate that it belongs to an older, predike series of intrusions.

The Diamond pluton extends from several small disconnected patches near the lower (northern) Rae Lakes, across to the east slope of the Sierra at Diamond Peak (from which it was named), to Parker Lakes. The rock is generally light colored and alaskitic, though some darker facies are present (fig. 51 and table 26). Toward the southeast, the rock becomes darker and poorer in quartz and some facies are monzonitic in composition.

TABLE 26.—*Modes of Diamond pluton*

Sample No.	Quartz	K-feldspar	Plagioclase	Biotite	Hornblende	Opaque accessories	Nonopaque accessories	Color index	Sp gr
1	38.4	36.2	24.2	0.9		0.2		1.1	2.58
2	13.9	50.5	25.8	3.8	1.7	3.2	1.1	9.8	2.61
Av	26.2	43.4	25.0	2.4	.9	1.7	.5	5.5	2.60

The Diamond pluton is in contact with two other granitic masses: the Dragon pluton and the White Fork pluton. It is intruded by the Dragon pluton, as can be clearly seen in the glaciated exposures by the lower Rae Lakes on the John Muir Trail. The gneissic granodiorite of the White Fork pluton is considered older, as shown by its extensive shearing and recrystallization.

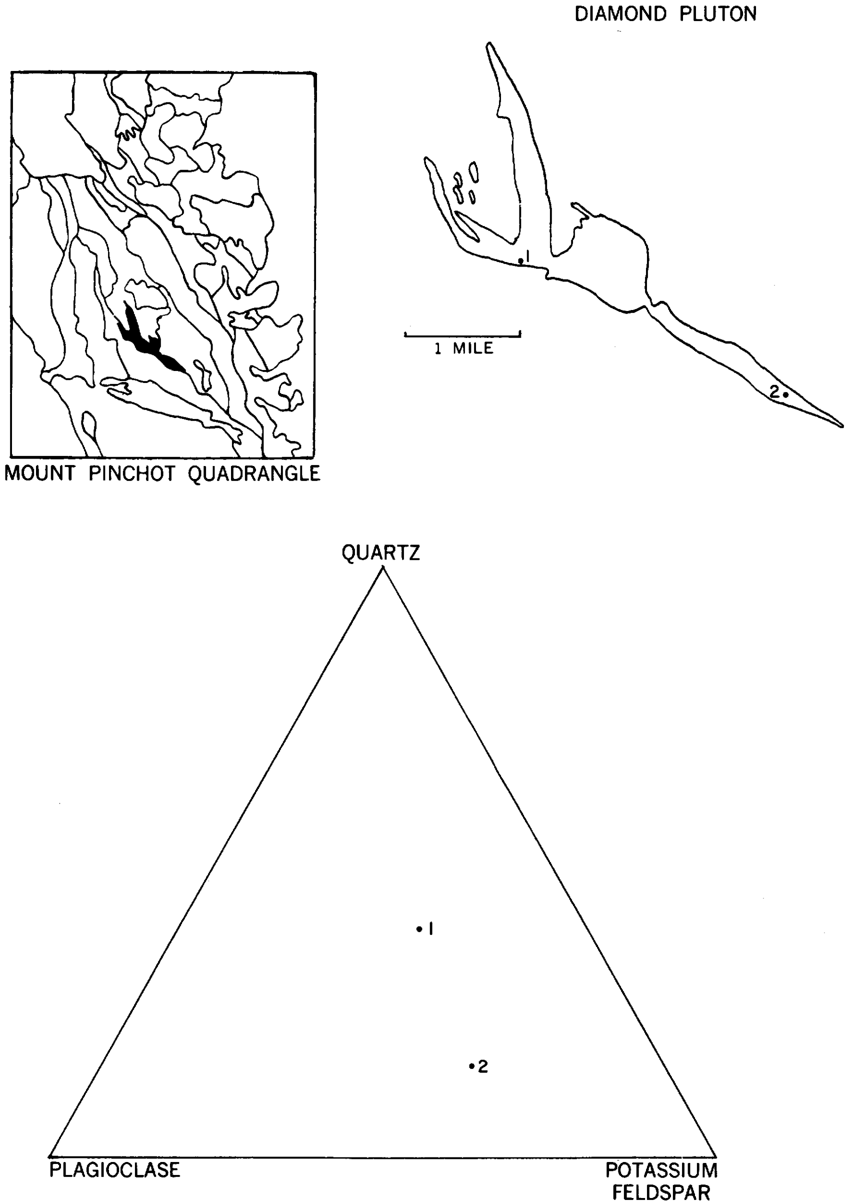

FIGURE 51.—Location and modal composition of samples from Diamond pluton. Numbers refer to samples given in table 26.

GRANITIC DIKES

A large number of dikes satellitic to the granitic plutons are in the quadrangle. They cut the metamorphic rocks, mafic hybrid rocks, and some of the granitic plutons. The dikes vary greatly in texture and composition. They include aplites and pegmatites in addition to fine-grained quartz monzonites.

On the geologic map only the more conspicious dikes or dike swarms have been mapped. One dike of interest is a strongly porphyritic granitic dike in the southeast corner of the quadrangle. The dike, which is nearly vertical, is cut by the Independence fault and is only slightly offset, demonstrating that this major fault has, if any, only a slight strike-slip component. North of Grays Meadow, a large number of north-trending near-vertical dikes cut the Independence pluton. These dikes are quartz poor and may best be called monzonite porphyry.

The most common granitic dikes are swarms of subhorizontal dikes sent off by the various plutons into tension fractures or marginal thrusts opened in the wallrock by the intrusive action of the parent pluton. On the geologic map such dikes are shown surrounding the Cartridge Pass pluton, Bullfrog pluton, and Lamarck granodiorite. There are also many unmapped satellitic dikes of this sort around the Spook, Cotter, and Mule Lake plutons.

PETROGENESIS

ORIGIN OF THE GRANITIC ROCKS

Most of the granitic rock composing the plutons of the Mount Pinchot quadrangle appears to be intrusive and to have crystalized from a magma. Some replacement granitic rocks were formed by recrystallization of other rocks, but they make a very small volume as compared with the exposed magmatic rocks.

The evidence concerning the origin of the granitic rocks is of three kinds: (1) field relations, (2) petrographic features, and (3) modal and chemical composition.

FIELD RELATIONS

One of the most striking features of the granitic rocks, not only in the Mount Pinchot quadrangle but throughout the eastern part of the Sierra Nevada batholith, is the sharp apparently intrusive character of the contacts between individual granitic masses and their walls, whether the walls consist of metamorphic rocks or of other granitic masses. The contact between two plutons many square miles in area can be located within a few inches (figs. 22 and 26). The contacts of the plutons with metamorphic rocks are equally sharp (fig. 40) and show many intrusive features such as tongues and apophyses of granitic rock extending into the wallrock. In most places intrusive features are so clearly shown that consistent relative age relations can be determined from them.

The internal structure of the plutons also suggests that each was emplaced as a unit and developed its own internal structures as a result of this movement. The foliation of the pluton, which is usually parallel to the walls, commonly cuts across the structures of the wallrock, as on the south side of the Cartridge Pass pluton.

Swarms of gently dipping granitic dikes are present in the wallrock around many of the granitic plutons, especially the more silicic ones. These dikes dip into the pluton and are probably akin to cone sheets, being formed by the upward motion of an intruding pluton, which opens tension fractures in its walls and fills them with the most mobile magma of the solidifying pluton. Gently dipping dikes are well exposed in the walls around the Cartridge Pass pluton and the Lamarck pluton. They are especially thick at the contact between the Bullfrog pluton and the adjacent dioritic mass.

The general rounded shapes of many of the plutons suggest that they were intruded as a unit and are not the result of crystallization of preexisting rocks. However, only the youngest plutons are equidimensional. In general the older plutons are more elongate and irregular in shape than the younger ones; apparently the younger plutons forced the older ones aside and distorted them.

Evidence for physical dislocations and forcing aside of wallrock structures is suggested by the map pattern of the Mount Pinchot quadrangle. The eastern offset of the Independence swarm of mafic dikes and of the metasedimentary septa near the northern part of the map may be an example of shouldering aside. The Cartridge Pass pluton and Lamarck granodiorite are in a position to effect this offset. Similarly, wallrock structures around the Baxter pluton appear to be disrupted and arched away from the regional trend.

Many of the plutons are fine grained near their margins, and the smaller and thinner plutons are fine grained throughout. For example, the Cartridge Pass pluton is distinctly finer grained on its border, and the relatively small Striped and Cotter plutons are entirely fine grained. Such relations suggest that the plutons are masses intrusive into relatively cool walls and developed a finer grained texture where rate of heat loss was greatest.

PETROGRAPHIC FEATURES

The chemical zonation of many of the plutons suggests that they were intruded as a unit and differentiated during cooling. Study of the minerals of the Cartridge Pass pluton shows that in the margins the minerals are higher temperature varieties, and that toward the core progressively lower temperature varieties are present. For example, plagioclase is progressively more sodic and biotite is more abundant relative to hornblende toward the center of the pluton. Mineralogic changes are related to the concentric zonation and not to changes in wall rock composition. Such changes are in accord with more rapid cooling of the margin of the pluton, which fixed the high-temperature forms. Slower cooling of the core allows adjustment to mineral varieties stable at lower temperature.

Textural evidence suggests that most of the granitic rocks have crystallized from a melt. Most of the granitic rocks have hypidiomorphic granular texture; progressive euhedralism of the essential minerals suggests crystallization in an order that is in agreement with Bowen's reaction series. Quartz, one of the last minerals to crystallize, is molded around the earlier crystallizing minerals.

Plagioclase crystals in most of the plutons show strong oscillatory zoning, with euhedral inner zones and as many as 30 to 40 recurrences in a single crystal. Such zoning demonstrates a temperature-pressure fluctuating environment, which is more probable in a moving magma than in rocks undergoing regional metamorphism.

MODAL AND CHEMICAL COMPOSITION

Modal analyses of the granitic rocks show that variation of the compositional averages of the plutons agree with differentiation following the laws of crystal-liquid equilibria as deduced from experimental studies. It has been shown that the trend of average modes of the plutons heads toward and terminates in the minimum melting trough of the system quartz–K-feldspar–albite. Such relations suggest that the parent rock of the plutons differentiated as a melt in equilibrium with solid material.

METHODS OF EMPLACEMENT

Several mechanisms of emplacement are operative for the plutons studied and the relative importance of each mechanism probably differs from one body to another. Much of the evidence suggests that the individual plutons were quite viscous, that both wallrock and intrusive rock behaved plastically, and that consequently stoping was not a dominating process. Forceful injection and shouldering sidewards and upwards was perhaps the most important single process in making room for the incoming granitic plutons. Nevertheless, some form of interaction with the wallrocks is believed to have been important in the emplacement of the Bullfrog, McDoogle, McGann, Dragon, and Goodale plutons—all of which are especially low in quartz.

The theory of stoping fails to explain several features of the overall map pattern. Most of the pluton contacts (either with other plutons or with metamorphic rock) are straight or gently curving over great distances. In those plutons in which the roof is exposed, such as the Colosseum mass, the contacts also appear fairly smooth and regular. If piecemeal stoping were operative, the contacts would be irregular, reflecting the size and shape of the individual blocks quarried out.

Likewise the long septa of fairly uniform width that separate individual plutons are not likely to have been preserved if the plutons were emplaced by stoping. The metavolcanic septum between the Bullfrog and Paradise plutons is 4 miles and only a few hundred feet wide, and that between the McDoogle and Tinemaha plutons is 5½ miles long and never wider than one-half mile. Why were these thin septa not breached if stoping was at all important in enlarging the incoming plutons?

Though cross-cutting relations are commonly seen on a small scale at the contacts (Noble, 1952, p. 54, points out that discordant contacts between intrusive bodies and wallrock are not uncommon in regions of forcible intrusion), the map pattern shows that the plutons are generally conformable with one another and with the metamorphic wallrock. In some areas the wallrock is shattered and large masses of it are included in the intrusion, as the north side of the Spook pluton, the east side of the Dragon pluton, and the dioritic complex adjacent to the Bullfrog pluton. Such features may record the operation of stoping, but they may also be expected if forceful intrusion were dominant.

The internal structure of the plutons is generally quite regular, with a steep foliation parallel to the walls. The last motion of the pluton, which these structures appear to record, is the emplacement of the pluton. They do not show the local turbulence and swirls one would expect to be associated with the stoping and foundering of individual blocks of wallrock.

ZONED PLUTONS

Specific gravity measurements show that at least five intrusive masses are zoned from a silicic core to a mafic border; these are the Cartridge Pass, Spook, Arrow, and Paradise plutons and Lamarck granodiorite. In addition, reconnaissance measurements indicate that the Cotter pluton and Tinemaha granodiorite are zoned, and that many other plutons may be zoned. Toward the center mafic minerals, plagioclase, and mafic inclusions are progressively less abundant, and K-feldspar and quartz are more abundant. Toward the center of some plutons the K-feldspar crystals increase in size.

Five possible mechanisms can be postulated to explain the zonation: (1) contamination of the margins of the pluton by the assimilation of mafic wallrock, (2) movement of early formed mafic crystals from the center to the edge by convection currents, (3) settling of early formed mafic crystals in the center downward by gravity while those at the cooler margin are frozen in place, (4) diffusion of mafic constituents toward the cooler margin of the pluton where they have been removed from the melt by crystallization, and of sialic constituents toward the warm core, and (5) multiple intrusion. Of the first four, mechanism 1 implies that the pluton is now more mafic than the original magma, mechanism 3 that it is more silicic than the original magma, and mechanisms 2 and 4 that it is about the same as the original magma.

Evidence is incomplete to establish which mechanism or combination of mechanisms was operative in the plutons. However, the lack of much mafic wallrock in contact with many of the plutons would suggest that mechanism 1 was not important at the level of exposure; many of the plutons are in contact chiefly with other granitic plutons. The lack of any structures indicating convection currents argue against mechanism 2. The gradational character of the zonation in most of the plutons, and the fact that the center is always more silicic, point away from multiple intrusion 5. The evidence is suggestive that mechanisms 3 and 4—crystal fractionation through gravity settling and diffusion—were important in producing zonation of the plutons. If settling occurred, however, it would have to have taken place before final movement of the pluton, since orientation of inclusions implies that there was no settling after they were oriented.

The cores of the Cartridge Pass and Spook plutons became partially mobilized and intruded the border of the pluton. Mobilization of the core of the Spook pluton progressed slightly further than in the Cartridge Pass pluton, in that the entire contact of the core with the margin is intrusive. Remobilization of the core proceeded until it breached the margin on the east and southeastern sides, and the core is now in contact with the original walls of the pluton. Such mobilization of the more silicic core of a zoned pluton would appear to be a logical consequence of renewed pressure from below, since the core would be more mobile for two reasons: (1) the core would be hotter, for heat loss from the pluton is greatest from the margins, and (2) the core has a lower melting point, for it is richer in silica and the alkalies.

The fact that the core of these plutons is slightly porphyritic, whereas the margin is not, suggests that the K-feldspar phenocrysts formed after emplacement of the pluton, during the period when redistribution of materials was forming a zonation in the pluton.

Similar concentric plutons are present in the Yosemite region (Calkins, 1930) where a rudely concentric arrangement of plutons is described in which each inner pluton is younger and more silicic. P. C. Bateman (1961) describes similar nested plutons in the Bishop area, and C. D. Rinehart (1963), in the Mount Morrison quadrangle.

PORPHYRITIC K-FELDSPAR

Six intrusive masses in the quadrangle contain phenocrysts of K-feldspar. They are the Paradise pluton, Spook pluton, Cartridge

Pass pluton, Lamarck granodiorite, Mule Lake pluton, and Tinemaha granodiorite, listed in order of decreasing size of their phenocrysts. The K-feldspar phenocrysts are small as compared with those in the Cathedral Peak granite (Calkins, *in* Matthes, 1930) in which 4-inch phenocrysts are common. The larger phenocrysts of the Paradise pluton are 1.5 inches long, and those of the Tinemaha granodiorite, 0.5 inches long. All the porphyritic plutons that were modally analyzed are silicic granodiorites and fall in a remarkably restricted range of mineralogical composition (fig. 11). There is no modal data on the Mule Lake pluton. The porphyritic granitic rocks lie about intermediate in the compositional range of all the granitic rocks (fig. 11) in the quadrangle. This intermediate composition (silicic granodiorite) is presumably necessary for the formation of the large K-feldspar crystals.

The K-feldspar phenocrysts are found in the center only of two of the zoned plutons: the Cartridge Pass and Spook plutons. The K-feldspar phenocrysts are larger and better developed in the larger plutons of silicic granodiorite. For these reasons it is assumed that the K-feldspar phenocrysts form after initial intrusion to about the present level. It is believed that the phenocrysts were not originally in the magma at the time of original intrusion; if they were, they should be present up to the margins of the zoned plutons. In the remobilized zoned plutons, the phenocrysts were present before the final intrusion of the core into its walls.

The intermediate composition of the melt (silicic granodiorite) is perhaps ideally suited to the production of large K-feldspar crystals. More potassic melts lie closer to the low-melting trough of the granite system (Bowen, 1954) and represent rocks in which the essential minerals crystallize more or less contemporaneously. Simultaneous crystallization and mutual interference would not favor the production of phenocrysts. On the other hand less potassic melts, simply because of the lower content of the constituents necessary for the K-feldspar phenocrysts, would not favor the growth of these large crystals, because a substantial part of the rock would already be crystallized at the time K-feldspar began to form and the K-feldspar (and quartz) would be relegated to interstices. Even though K-feldspar phenocrysts are confined to the silicic granodiorites, not all the silicic granodiorites contain phenocrysts. The phenocrysts are favored in the larger and thicker plutons; the three largest silicic granodiorite plutons (Cartridge Pass, Paradise, and Spook) have the best developed K-feldspar phenocrysts, whereas the thinner or smaller plutons like the Cotter, Colosseum mass, and White Fork contain no phenocrysts.

In summary, two factors are important in the production of K-feldspar phenocrysts: (1) an intermediate composition with approximately twice as much plagioclase as K-feldspar, and (2) a large-size pluton (roughly more than 1.5 miles in width). The phenocrysts grew at a time when the magma was still plastic enough to intrude its own walls after the growth of the crystals.

MAFIC INCLUSIONS

Dark elliptical or rounded inclusions are common in most of the granitic rocks of the quadrangle. They average 3 or 4 inches in greatest dimension, and are disc shaped, rod shaped, or almond shaped (triaxial ellipsoid). Preferred orientation of the inclusions produces planar, linear, or combined planar and linear structures. These inclusions are masses of slightly finer grained darker rock, richer in the mafic minerals of the enclosing rock. Mineralogically they represent a concentration of hornblende, biotite, and plagioclase, and a paucity of K-feldspar and quartz, relative to the surrounding granitic rock. Many have no K-feldspar and less than 5 percent quartz. Commonly the inclusions possess an internal structure formed by a planar or linear orientation of mineral grains, mainly hornblende. Papst (1928, p. 358) and Mayo (1941, p. 1022) point out that these dark inclusions appear to be separate from, and of a different origin than, plainly recognizable wallrock xenoliths.

In the Mount Pinchot quadrangle mafic inclusions commonly have a characteristic habit in each pluton. In the darker, K-feldspar-poor plutons, mafic inclusions tend to be more abundant and larger than in the leucocratic, K-feldspar-rich plutons. The Dragon pluton contains dark inclusions with especially gradational borders. The inclusions in the Goodale pluton are poorly sorted as to size, and in the Pyramid pluton inclusions are especially large, up to 12 inches in length. The texture, size, shape, and abundance of the inclusions are so characteristic of the pluton in which they occur, that they have been used as an aid in mapping the plutons.

Table 27 shows the average number of mafic inclusions per square yard of horizontal outcrop in 21 plutons as calculated from a simple average of a number of field measurements. Though the concentration of mafic inclusions is generally quite uniform in any small area, there are broad changes over the area of a single pluton. Hence the averages of measurements of the density of the inclusions is recorded only as a rough guide to their abundance.

On figure 52 the average number of inclusions per square yard is plotted against the color index of the pluton. Though the points scatter widely, in general the darker plutons contain more mafic inclusions. It is significant that the light-colored, silicic plutons contain few or no inclusions. However, the size or width of the pluton also has an effect on the number of mafic inclusions. Thin plutons (less

TABLE 27.—*Average number of mafic inclusions per square yard of horizontal outcrop in each of 21 granitic plutons*

Pluton	Number of field measurements	Average number of inclusions per square yard	Pluton	Number of field measurements	Average number of inclusions per square yard
Baxter	2	6	Tinemaha granodiorite,		
Pyramid:			Woods Creek mass	35	3.3
East mass	11	4.9	Cotter	55	5.4
West mass	7	4.4	Paradise	36	0.2
McDoogle	9	4.1	Mule Lake	1	1
Dragon	42	3.3	Siberian	10	0
Lamarck granodiorite:			McGann	34	2.4
Colosseum mass	15	0.5	Goodale	11	4.4
Lamarck mass	40	2.1	Red Mountain Creek	7	0
Cartridge	69	0.3	Bullfrog	15	0
Arrow	67	2.0	Independence	33	0
Spook:					
Margin	31	0.2			
Core	16	0			

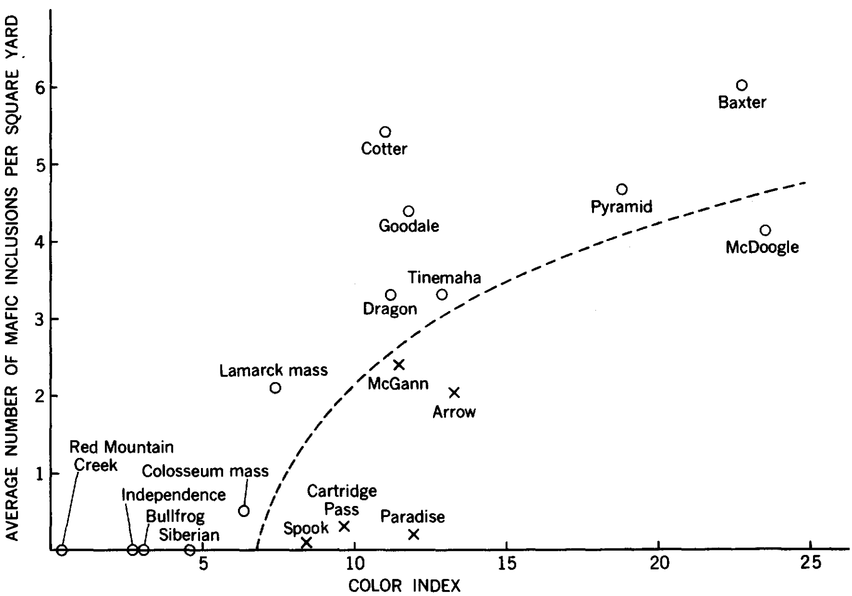

FIGURE 52.—Comparison of color index in some plutons with the average number of mafic inclusions per square yard. Plutons with an average width greater than 1.5 miles are shown by a cross, and those with an average width of less than 1.5 miles, by a circle. In general, thicker plutons contain fewer inclusions than do comparable thinner plutons.

than 1.5 miles in mean width) have a greater density of mafic inclusions than do the thicker plutons. The two extremes in this regard are illustrated by the Cotter and Paradise plutons. Both plutons are quite similar in composition, but the thin Cotter pluton (about one-half mile wide) contains 5.4 inclusions per square yard, and the thick Paradise pluton (greater than 3 miles wide) contains about 0.2 inclusions per square yard.

Within plutons there is also a variation in the number and character of the inclusions, but these changes are generally smaller than changes from one pluton to another. In most plutons the general picture is similar to that in the Cartridge Pass pluton: the inclusions are most abundant near the margins and decrease toward the center of the pluton, as the enclosing rock becomes more silicic.

The mode of origin of the majority of the mafic inclusions has not been established in this study. Three mechanisms of origin warrant consideration: (1) segregation of early formed minerals, (2) inclusion of wallrock, including the early crystallizing mafic border facies of the pluton itself, and (3) preservation of remnants of refractory material that were not melted when the magma was formed. The first process seems the least likely because of the absence of radial or concentric structures in the inclusions, which would be expected in segregations growing in a melt. Evidence bearing on the distinction between the second and third processes is somewhat ambiguous.

In some exposures every gradation between clearly recognizable wallrock xenoliths and the typical mafic inclusions can be seen. The greater abundance of inclusions in the margins of intrusives suggests the possibility that they have been derived from wallrock. However, even though some of the inclusions are clearly derived from wallrock, several facts raise doubt as to the importance of this process: Many of the plutons containing mafic inclusions are in contact (at the present level of exposure) for broad areas only with other granitic rocks, not with metamorphic or gabbroic rocks from which the inclusions could be made. Also, the fairly even distribution of inclusions (in spite of marginal zoning) would require that thorough mixing carried inclusions for great distances inward from the walls. Furthermore the uniformity of the inclusions in the plutons suggests that all were derived from similar types of rock (impossible for locally derived xenoliths) or that they have been in contact with the melt for a long period and have attained quasi-equilibrium with the melt.

Several additional facts suggest that the inclusions are an early feature in the history of the pluton and were present before intrusion to the present level. A distinct relation exists between composition of the pluton (as shown by color index, fig. 52) and the number of inclusions. The lighter colored plutons contain fewer inclusions, and the alaskitic plutons contain virtually no inclusions even in chilled margins. One explanation of this fact is that the inclusions were present in the parent magma *before* differentiation and were sorted out as to number and size by the same processes which effected the original differentiation (settling of heavy crystals, filter pressing, and so forth). Scarcity of inclusions in the center of plutons and in the large plutons (fig. 52) is perhaps a result of digestion of inclusions in the regions of slower cooling.

Hence, even though the evidence indicates that some of the inclusions are metamorphosed wallrock, the possibility of some or most of the inclusions originating as remnants of refractory premagma material is not eliminated. Furthermore, the data suggest that if the inclusions are wallrock xenoliths, they have been in contact with the melt for a relatively long period.

MAFIC DIKES

Hundreds of mafic dikes are present in the Mount Pinchot quadrangle. The average mafic dike has a specific gravity of 2.81, and the mineral makeup of a biotite-hornblende diorite. Several types of dikes may be distinguished, on the basis of both mineralogy and origin. However, field relations suggest that the great majority of the dikes belong to a swarm of northwest-trending dikes covering a large area. This swarm has been named the Independence dike swarm (Moore and Hopson, 1961), because the only easily accessible area in the quadrangle where the dikes are exposed is in the canyon of Independence Creek, a few miles west of the town of Independence. The mafic dikes not belonging to the Independence dike swarm will be discussed separately.

INDEPENDENCE DIKE SWARM

Only a part of a large swarm of northwest-trending, near-vertical mafic dikes is present in the Mount Pinchot quadrangle. The swarm trends north and northwest into the southern part of the Big Pine quadrangle. Southeastward the dikes are last seen in the Sierra Nevada near Independence Creek, in the southeastern corner of the Mount Pinchot quadrangle. However, similar mafic dikes reappear in large numbers in the Inyo Range in the New York Butte quadrangle (about 1 mile north of Swansea), more or less on strike with those of the Sierra Nevada. It is tentatively believed that these belong to the same swarm even though they are generally much more altered than the Sierra dike rocks. Similar dike rocks appear on strike to the southeast in the Ubehebe Peak quadrangle (McAllister, 1956) and the Darwin and Panamint Butte quadrangles (Hall, 1957, oral communication). Hence the dike swarm as described above may extend more than 80 miles parallel with the strike; on each end it is unmapped. Field evidence in the Mount Pinchot quadrangle indicates that the dikes were intruded in the Cretaceous plutonic cycle during a relatively short time. If these age relations hold true for the entire

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swarm, as the structural continuity of the swarm would suggest, then the dike swarm could serve as a time indicator of major importance.

FIELD RELATIONS

The mafic dikes cut almost every mass of metamorphic rock, most of the dioritic complexes, and some of the granitic rocks. About 40 square miles of granitic rock is older than the dikes, as compared with 120 square miles that is younger than the dikes. Since the granitic rocks appear to be middle Cretaceous in age, the dikes are considered to represent a much shorter span of middle Cretaceous time.

Mafic dikes that are cut off and included in a younger granite have been observed in several places, including the contact between the Red Mountain Creek and Goodale plutons on the north wall of Goodale Mountain, the contact between the Cartridge and White Fork plutons south of Bench Lake, and the contact between the Dragon and White Fork plutons east of Mount Clarence King. Generally, however, the dikes are nearly parallel to the interpluton contacts, and the presence or absence of dikes in each pluton has been assumed to indicate whether it is older or younger than the dikes. At some places where the dikes are cut off by a younger pluton, pieces of the dike rock may be seen included in the younger granitic rock.

The possibility of the mafic dikes being tabular inclusions, septa, or unreplaced remnants (skialiths) older than the host rock is disproven by obvious intrusive and dilational structures of the dikes. The dikes sharply cut the plutonic rocks, displaying knife-edge contacts and fine-grained chilled borders. They often contain angular inclusions of country rock and possess flow banding. Dilation with the offset of diagonal structures perpendicular to the walls is well shown; one dike was studied that cut and offset both foliation in the metavolcanic wallrock and a small aplite dike. The direction of offset is perpendicular to the dike walls a relation which eliminates the fortuitous possibility that the dike was formed along a fault by replacement (fig. 53). Irregularities on the opposite walls of the dikes match each other.

The mafic dikes range in thickness from less than 1 inch to more than 40 feet. However, the average thickness is between 1 and 2 feet (fig. 54). The thickness distribution closely approximates a log normal distribution.

The mafic dikes have been subjected to a complex series of changes after their intrusion and solidification. They have been (1) broken and torn apart by plastic movement of the granitic rocks; (2) recrystallized and metamorphosed by the emplacement of neighboring plutons; (3) intensively sheared; and (4) replaced partly or completely by granitic material.

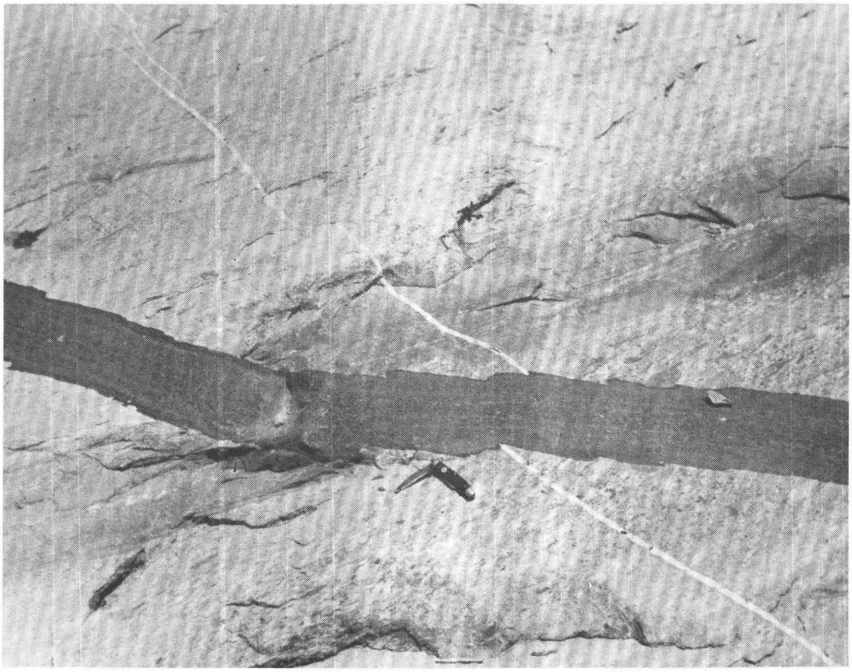

FIGURE 53.—Mafic dike south of Twin Lakes cutting layered metavolcanic rock. A thin aplite dike and dark layers in the host rock are offset perpendicular to the walls of the dike. Note matching walls, flow structure, and finer grained margins.

Many of the dikes are offset a few inches, yet the offset ends of the dike are nearly always connected by a thin septum of dike material (fig. 55). Also commonly small dikelets extend from the blunt ends of the truncated dikes. Such relations suggest that the dikes jump from one fracture to another and the small dikelets mark the continuation of the original fracture. Very likely, above or below the plane of exposure the fractures merge together.

In addition to the offset caused by dikes jumping from one fracture to another, some of the dikes have been offset and torn apart by later flowage of the host rock. This produces a structure similar to boudinage in disturbed sedimentary rocks. In some places a mafic dike may be traced into a train of isolated and somewhat rotated segments of the original dike. Such disruption of the dikes was probably caused by the physical and thermal effects of the intrusion of a neighboring granitic pluton.

Many of the dikes, especially the ones less than 3 inches in thickness, are more or less schistose. The fine-grained dikes, buttressed by their granitic wall rocks, were loci of considerable movement. Many of the thinner dikes present a diagonal pattern to their schistosity when viewed from above (fig. 56). The schistosity is most intense near

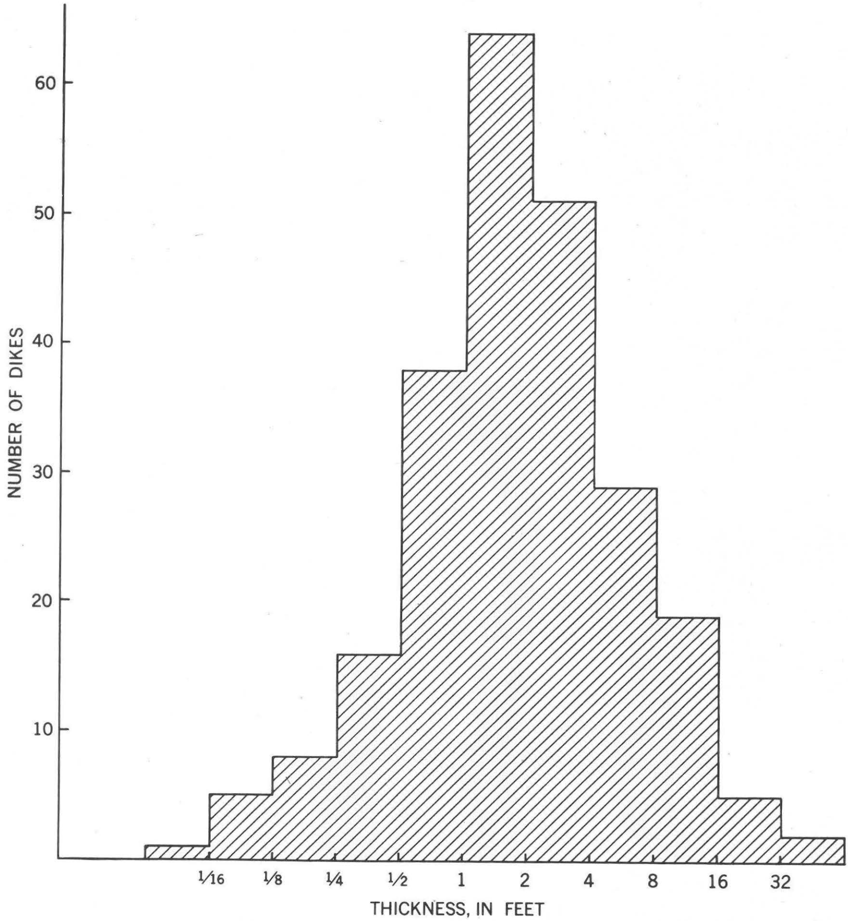

FIGURE 54.—Histogram showing frequency of thickness of mafic dikes in the Woods Lake region. The distribution of the 238 dike thicknesses plotted closely approximates a lognormal distribution.

the margin of the dike, where it swings around parallel to the walls. The orientation of the diagonal foliation demonstrates in nearly every case that movement was left lateral along the plane of the dike. The dike rocks show differing degrees of recrystallization; most now have crystalloblastic textures, but relict igneous textures are discernible in the mildly recrystallized dikes. The left-lateral movement described above was early, before complete recrystallization of the dike rocks.

The mafic dikes are subject to replacement by the residual fluids of the younger granitic instrusions. The contact of dike and granitic host rock commonly serves as a channelway for replacing solutions which convert the outer part of the dike into a lighter colored coarser

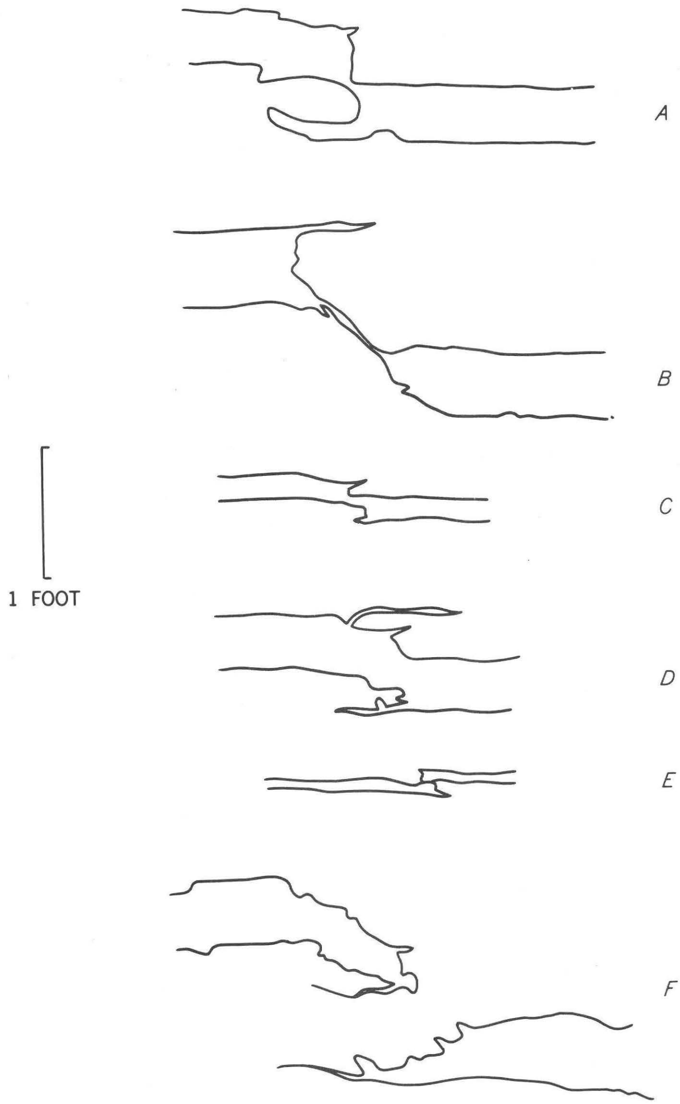

FIGURE 55.—Outlines of mafic dikes in the Woods Lake basin area. The dikes jump from one fracture to another and the small dikelets mark the continuation of the original fracture.

granitic-appearing rock. Rounded, unreplaced remants of the dark rock remain, giving the appearance of a breccia, whereas the structure is in fact a pseudobreccia of replacement origin.

Some mafic dikes are locally sheared apart by a set of conjugate shears into a train of angular fragments. The fragments are especially susceptible to replacement, and when the replacement has gone

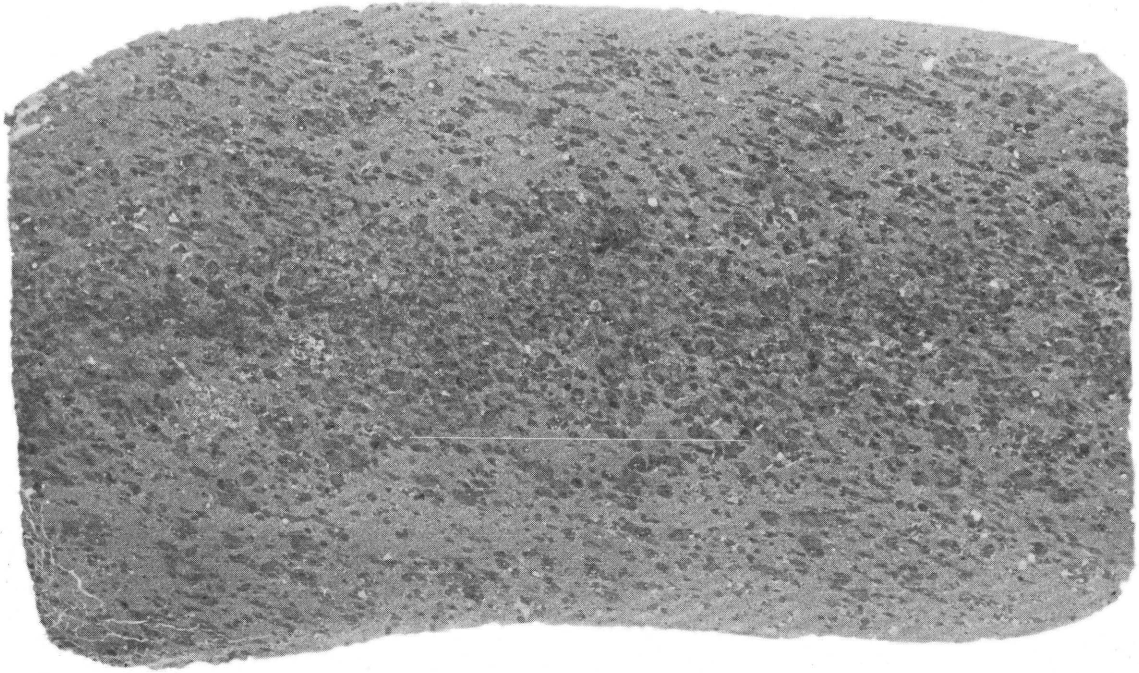

FIGURE 56.—Photograph of a thin section of a mafic dike 1 inch thick from Woods Lake basin. The thin section covers the entire width of the dike; the edges of the dike are at the top and bottom of the thin section. A foliation shown by the dimensional orientation of the mafic minerals crosses diagonally through the dike from upper left to lower right. The foliation is most conspicuous near the margins of the dike where large mafic crystals are least abundant. The diagonal foliation is believed to have resulted from left-lateral movement of the dike walls after intrusion of the dike but before final crystallization. The greater abundance of mafic minerals in the core of the dike probably reflects slower cooling of the core than of the margins, which were chilled.

nearly to completion the dike fragments are similar to the mafic inclusions of the granitic rocks.

In the replaced parts of the dikes, the plagioclase crystals are considerably larger than those in the unreplaced dike rock. The plagioclase in one dike was strongly zoned, with a core of An_{50} and a rim of An_{20}. In the replaced equivalent, the plagioclase was zoned in the same fashion, but the rims were considerably thicker and the crystals larger; material was apparently added to the outer rim of the original feldspars. There also is an addition of quartz and K-feldspar and a selective removal of hornblende in the replaced dike rock.

PETROGRAPHY

The mafic dikes are notably varied from place to place, both in texture and in composition. The swarm is, in fact, a dike series ranging from mafic to felsic members, with the more mafic greatly predominating. At the mafic end are melanocratic andesine-green hornblende rocks, and at the felsic end are leucocratic granodiorite to quartz monzonite porphyries.

Most common of the dike rocks are mesocratic green hornblende–andesine rocks with varied amounts of brown biotite and subordinate K-feldspar and quartz. Sphene, apatite, and magnetite are the common accessories, and epidote is commonly present. The specific gravity of the rock averages 2.81, and the color index about 45. Groundmass plagioclase is andesine; plagioclase phenocrysts, commonly not present, are progressively zoned from calcic andesine cores to calcic oligoclase rims.

Relict igneous textures are discernible in mildly recrystallized dikes, but most of the dike rocks have been completely transformed to a granoblastic mosaic. Many dike rocks are more or less schistose, presumably due to recrystallization during penetrative deformation. A rough correlation exists between intensity of dike recrystallization and the proximity of the dike to intrusive contacts of younger granitic masses; dikes near these contacts are strongly recrystallized, and the least recrystallized dikes are at least one-fourth mile away from such contacts. The metamorphism was not retrogressive, for it produced an andesine or oligoclase–green hornblende–biotite mineral assemblage similar to that of the original igneous dikes.

AGE OF DIKES

The dikes of the Independence swarm appear to be closely related in age and origin. Four lines of evidence together support a single period of mafic dike intrusion for the northwest-trending swarm:

1. The general petrographic similarity of mafic dike rocks throughout the quadrangle indicates close genetic relationship.

2. The structural relations of the mafic dikes are suggestive of a single stage of dike intrusion. The dikes were intruded into a set of vertically dipping northwest-trending fractures; throughout the quadrangle the mafic dikes occur in swarms having this orientation. Prominent sets of vertical fractures (both joints and faults) oriented in other directions have not been intruded by dikes. Evidently mafic dike magma was available for intrusion at the time regional tectonic forces created the northwest-trending vertical fractures, but was not available when other fracture sets formed. In view of the complexity of deformation in the region, and of the wide variety of variously oriented fracture sets, it would seem probable that, had there been more than one stage of dike intrusion, dike swarms of other orientations would also be present in the area.[3]
3. No evidence has been found for more than one period of mafic dike intrusion. For instance, there are no granitic bodies that truncate mafic dikes but that in turn are themselves cut by other mafic dikes. In view of the large number of granitic intrusions (at least 24) and the great abundance of mafic dikes throughout the quadrangle, there would be ample opportunity for relations of this sort to be present if there had been more than one stage of mafic dike intrusion.[4] Furthermore, in every place where a granitic pluton that is cut by dikes comes in contact with one that contains no dikes, intrusive relations (if diagnostic) show that the pluton not cut by dikes is the younger.
4. The mafic dike swarm is intermediate in age between rocks of the batholith which, by other criteria, form two fairly distinct groups. The early, predike granitic rocks commonly show evidence of shearing, recrystallization, and retrogressive alteration, whereas the younger, postdike granitic rocks show no evidence of deformation or recrystallization. This difference is quite evident in the field, as well as under the microscope.

Although each of the above four points is inconclusive in itself, the four together strongly support the interpretation that the Independence dike swarm was formed during a relatively short period of mafic dike intrusion.

In southwestern Finland, Sederholm (1926, p. 31–36) used the extensive swarms of mafic dikes to discriminate between the older and younger Archean granites; the dikes mark a major break in the history of the Finnish Precambrian. By comparison, the time interval be-

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[3] Near the South Fork of Woods Creek the mafic swarm swings around to an eastward trend. This appears to be partly a postdike shouldering aside of the dike swarm by the younger Baxter pluton.

[4] The local mafic dikes caused by mobilization of hybrid wallrock by a neighboring intrusion are not considered in this discussion.

tween predike and postdike granitic rocks in the eastern Sierra Nevada was a short one, for both series of granitic rocks are younger than (intrusive into) metavolcanic rocks believed to be of Jurassic and Triassic age, and both are older than early Tertiary. Beyond the fact that the chilled aphanitic dike margins indicate that the predike granites were completely solidified and at a relatively low temperature during dike intrusion, little more can be inferred about the time interval separating the predike and postdike granitic rocks.

MOBILIZED WALLROCK DIKES

Mafic dikes quite separate from the Independence dike swarm are present in localized areas in the quadrangle. These dikes are generally nearly horizontal. They are up to 1 mile in length and commonly are associated with the dioritic complexes or masses of hybrid wallrock.

These dikes are present along the contact between the Bullfrog and Cotter plutons on King Spur, and some of them are shown on the geologic map. The Bullfrog-Cotter contact is nearly vertical, and pods of hybrid mafic plutonic rock are present at the contact. The mafic dikes intrude only the Bullfrog pluton in nearly horizontal sheets, yet seem continuous with the mafic hybrid masses of the contact zone. They are up to several feet thick, are quite uniform in thickness, and show matching walls. In no place do these dikes cut the Cotter pluton. These relations suggest that the younger Cotter pluton opened flat tension fractures in the Bullfrog pluton during intrusion and mobilized the mafic hybrid rock of the contact zone which then filled these fractures. Other areas of these mobilized wallrock dikes are on Mount Bago associated with the migmatized metarhyolite pendant, and on the east shore of upper (southern) Rae Lake associated with the dioritic complex to the south.

CENOZOIC DEPOSITS

BASALT

A series of basalt cinder cones, lava flows, and dikes appear near the base of the main eastern escarpment of the range. This volcanic activity occurred in the Quaternary and perhaps late Tertiary periods. In general, the basalts are similar, regardless of which of the 20 or more vents conducted them to the surface. The rock is a mafic olivine basalt. Most of the basalts contain very calcic plagioclase phenocrysts, commonly bytownite or anorthite.

The spatial restriction of the volcanic vents near the range front and the alinement of cinder cones along straight or curved lineaments parallel to the range front strongly suggest that the basalts reached the surface along fault zones. In all probability these same fault zones are among those associated with the uplift of the range.

On the basis of geographical position and age the basalts have been divided into four main groups, which are also distinguishable by petrographic criteria. These groups will be described in order from oldest to youngest.

OLIVINE BASALT DIKES OF DIVISION CREEK

Near the mouth of Division Creek, at an elevation of about 6,500 feet, two basalt intrusions are exposed on the canyon wall. The larger of the two is a thick pod-shaped intrusion several hundred feet wide, the other is a dike about 3 feet wide. Petrographically the two masses are similar, except that the rock of the larger body is coarser grained, and its olivine is no longer fresh but has been deuterically altered to a brown mineral (iddingsite?).

The age of these dikes relative to the other basalts can only be guessed. They have suffered much more erosion than the cinder cones and related flows to the north and hence are older. Whether they are older than the olivine basalt of Oak Creek is not known.

The smaller dike has intruded parallel to a close-spaced jointing in the Spook pluton. The contact is very sharp, and thin chilled margins are developed on the dike. No contact effects on the granodiorite were seen. The basalt has good flow structure parallel to the contact and columnar jointing perpendicular to the contact.

This basalt is similar to the olivine basalt near Aberdeen except that phenocrysts of plagioclase (An_{80}) are considerably more abundant. In addition to phenocrysts of olivine, augite, and bytownite, a spinel (picotite) is found sparingly scattered throughout the rock. The spinel occurs as deep-brown isotropic, nearly euhedral crystals with squarish outline. The presence of picotite suggests that these basalts are exceptionally low in silica content.

OLIVINE BASALT OF OAK CREEK

Aside from the basalt dikes of Division Creek, the earliest evidence of Quaternary volcanism in the Mount Pinchot quadrangle is shown by a succession of lava flows in the canyon of the North Fork of Oak Creek. Remnants of lava flows extends from a vent area at about 8,400 feet, down the canyon to its mouth, and out on the alluvial slope at the foot of the range to an elevation of 4,400 feet.

Only a small part of the original basalt remains, as isolated erosional remnants in the upper part of the canyon and as islands protruding through the alluvium on the alluvial apron below the mouth of the canyon. The extent of the dissection and alluviation of the basalts is the best key to the age of the flows as compared to the other basalts of the area. None of the other basalt flows has been eroded so deeply.

A large remnant of the basalt occurs on the ridge between the North Fork of Oak Creek and Charlie Canyon. This mass is cut off at its upper end by a normal fault and throughout its length the base of the basalt is offset by faulting. At about 6,800 feet a well-defined glacial moraine of the early glaciation (Tahoe till) appears to partially cover one of the Oak Creek flows. At the mouth of the canyon the flows are covered by the bouldery alluvium which Knopf (1918, p. 74) ascribes to the earlier glaciation.

Despite the erosion and faulting which the Oak Creek basalt flows have undergone, the flows still are not older than the main uplift of the Sierra block, for they bear a very definite relation to the present topography. They were erupted high in Oak Creek Canyon and flowed east down the canyon, covering the alluvial slope at the mouth. Hence they postdate the major uplift of the range, which presumably occurred in latest Tertiary and early Pleistocene time (Gilbert, 1941 p. 802). They are tentatively assigned to the Pleistocene before the earlier glaciation (Tahoe till). A precise statement of the relative ages of the Cenozoic volcanic rocks and glacial stages along the east slope of the Sierra Nevada must await detailed geomorphological study.

The source of the Oak Creek flows appears to be a vent area at an elevation of 8,400 feet in the small northern tributary of the North Fork of Oak Creek. Here a mass of columnar basalt about 40 feet thick forms a circular outcrop on the granitic valley wall; this may be the feeder dike or neck.

Below the highest outcrop of basalt are two sizable masses of a fairly well indurated basaltic tuff breccia, believed to have been deposited by a mudflow at the time of activity of the basalt vent. The tuff breccia is composed of lapilli and fragments of scoriaceous basalt mixed with a large amount of granitic sand, pebbles, and boulders. Some granitic boulders up to 5 feet in diameter are included in the breccia. The deposit shows a crude bedding parallel to the slope of the valley wall and attains a thickness of 55 feet.

Under the microscope, the major part of this tuffaceous mudflow is seen to be composed of small, somewhat rounded fragments of vesicular olivine basalt averaging 1 mm or less in diameter. The reddish color of these basalt fragments is due to the complete oxidation of the magnetite grains to hematite. Mixed with the basalt ash, and composing nearly one-half of the rock, is fine granitic debris: crystal fragments of quartz, plagioclase, biotite, and hornblende. Set in this matrix are blocks and cobbles of granite and cinders of basalt.

In general there is little evidence of melting or reaction between the minerals of basaltic and granitic derivation. The granitic quartz is neither corroded nor rimmed by reaction products; biotite and hornblende are both fresh and clear, unaltered and free of exsolved iron; plagioclase is commonly fresh but shows corroded, saw-tooth edges where entirely included in a basalt lapillus. The rarity of reaction effects presumably indicates that the tuff was not subjected to any great temperature after the incorporation of granitic material. Probably some form of mudflow was the mode of origin of these bedded tuff breccias.

The basalt flows in the canyon and on the spur to the south are commonly associated with beds of basalt cinders and breccia. In one exposure on the south side of the North Fork of Oak Creek, the flow of massive basalt with wavy columnar jointing is about 30 feet thick. Beneath the flow is a pyroclastic zone, several feet thick, composed of blocks, lapilli, and stringers of scoriaceous basalt. Beneath this is a zone, 10 feet thick in the exposure studied, of an unconsolidated mixture of pyroclastic basalt and granitic boulders, some up to 3 feet in diameter. The matrix is basalt bombs, cinders, and reddish soil derived from volcanic ash. The emission of these basalts was preceded and accompanied by pyroclastic activity.

Most of the Oak Creek basalts possess an intergranular mesh of plagioclase (the larger grains contain 75 percent anorthite), augite, olivine, and magnetite. Phenocrysts are small and inconspicious; commonly they are less than 0.5 mm in diameter, and the rock is virtually nonporphyritic in contrast to all the other basalts in the area. The groundmass carries abundant small olivine crystals, in places entirely altered to iddingsite. The groundmass is comparatively coarse, ranging from 0.04 to 0.15 mm in grain size. In hand specimen, some of the basalt displays a spotted appearance, the spots being light colored and averaging 2 mm in diameter. Under the microscope the spots are almost indistinguishable from the rest of the rock. They are indefinite areas slightly richer in clinopyroxene and magnetite granules; the flow-alined feldspar laths pervade matrix and spots alike, precluding the possibility that the spots are partially assimilated inclusions. The spots probably represent some late-stage crystallization centers, controlled possibly by concentration of volatile materials.

A remarkable feature of the olivine basalts of Oak Creek is the presence of inclusions of dunite. These inclusions are fairly common; they average about 1 inch in diameter but attain 5 inches. Some of the inclusions are tabular and angular, but several with subrounded shapes were observed. Many have a faint gneissose structure. No reaction effects are noticed between inclusion and inclosing basalt.

The grain size of the inclusions is from 0.5 mm to several millimeters. Olivine is the most abundant mineral and composes the bulk of the inclusion; some clinopyroxene is present. Layers of dark yellow-brown spinel make up about 5 percent of the inclusions.

The mineralogy of these inclusions suggests that they are not locally derived. Ross, Foster, and Myers (1954), in studying many such dunite inclusions in basalt of very similar mineralogy, conclude that the inclusions are derived directly from the peridotite zone of the earth's crust.

In places, inclusions of granitic rocks and quartz are found in the basalt flows. These were either ripped off the walls of the volcanic conduit or picked up from the granitic terrain over which the lava flowed. Most of the quartz grains are resorbed, corroded, and surrounded by a reaction rim of small, radially arranged clinopyroxene needles.

OLIVINE BASALT OF SAWMILL CANYON

The basalt flow of Sawmill Canyon is presumably interglacial in age, for it appears to lie on top of a moraine of the earlier glaciation (Tahoe), and is covered by a large moraine of the later (Tioga) stage. If the material that covers the Oak Creek flows was deposited during the older glaciation, the Sawmill Canyon basalts are younger.

The basalt flows of Sawmill Canyon originate in a vent area on the north side of the canyon at an elevation of 7,600 feet. Here a large accumulation of basalt pyroclastic debris lies on and near the lower end of an earlier moraine. Many of the well-formed volcanic bombs associated with the tuffs and breccias are 1 and 2 feet in length, which indicates the close proximity of the volcanic orifice even though the shape of the cinder cone has been destroyed by erosion.

At the time the basalt flowed down the canyon of Sawmill Creek, the canyon was a steep-walled, V-shaped slot of much the form it has today. In the main part of the canyon there are at least two basalt flows, separated by several feet of frothy, cindery basalt. In several places in the canyon, the base of the basalt can be seen resting on granodiorite. A basal layer of basalt breccia about 2 feet thick, but locally 8 to 10 feet thick, is commonly present. This breccia generally consists of well-sorted subangular fragments of scoriaceous basalt (1–3 in. in diameter) which are unconsolidated or very poorly consolidated. The breccia grades upward into massive flows attaining 80 feet in thickness; the total thickness of the basalt is 150 feet in some places. Commonly the flows develop columnar jointing perpendicular to the steep canyon walls on which they rest.

After the lava had filled the V-shaped canyon to a depth of about 150 feet, stream erosion removed the major part of the basalt. The upper three-fourths mile of the flows was covered by the lower end of the Sawmill Canyon glacier during the last glacial advance (Tioga).

The flows in the lower canyon were cut entirely through by Sawmill Creek, which presumably carried more water during glacial times. As a result, only small remnants of the basalt were left clinging to the canyon walls above the present stream.

Modal analyses show that the Sawmill Canyon lavas are more porphyritic than the other Cenozoic basalts, carrying from 17 to 24 percent phenocrysts. These phenocrysts are olivine, augite, and plagioclase, in order of decreasing abundance (fig. 57). The groundmass is

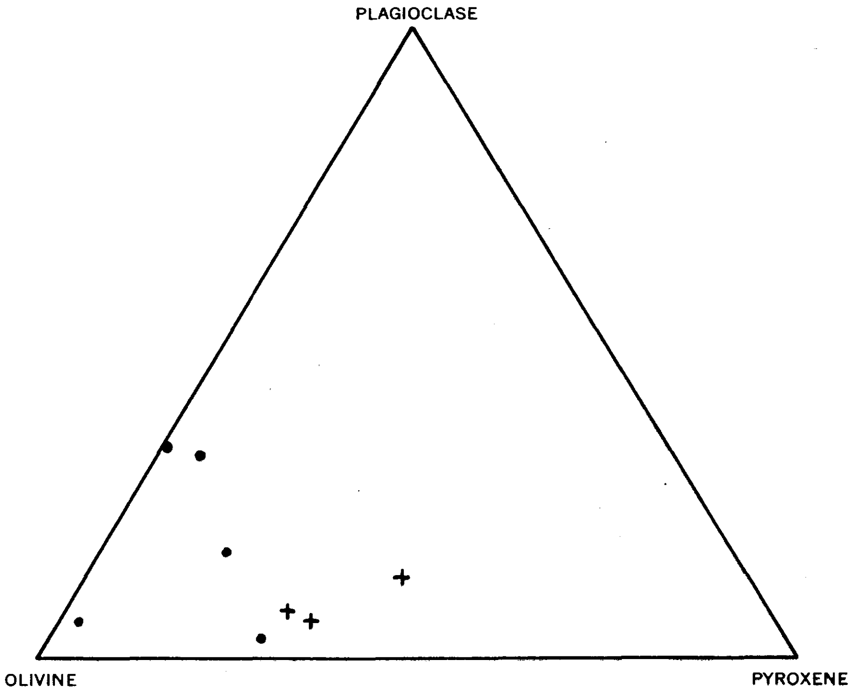

FIGURE 57.—Volume percent of phenocrysts in eight samples of basalt. Crosses, olivine basalt of Sawmill Canyon; dots, olivine basalt west of Aberdeen.

quite fine grained (0.02 mm to 0.07 mm) intergranular, speckled with magnetite or hematite dust.

The refractive index of the fused volcanic rock, which is an approximate measure of the SiO_2 content of the rock, shows that the basalt of Sawmill Canyon is the least silicic of all the basalts (fig. 58). The rock probably has about 47 percent SiO_2.

Olivine, both as phenocrysts and in smaller grains, forms euhedral crystals, commonly with a dark rim of exsolved magnetite and hematite. The olivine has $2V$ of 90°, and is hence rich in magnesia (about 15 percent fayalite).

Augite phenocrysts ($2V=59°$, extinction angle $= 45°$) are generally conspicuously zoned. The core has a smaller extinction angle

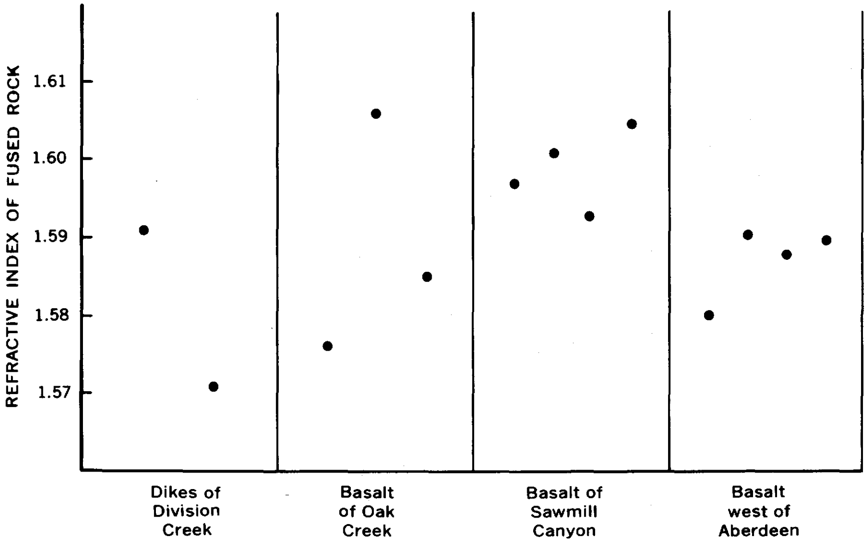

FIGURE 58.—Refractive indices of fused basalts. A refractive index of 1.58 corresponds to about 50 percent SiO_2, according to Mathews, 1951.

and is therefore poorer in iron or calcium. The core, and specific bands of the oscillatory zoned rim, commonly contain abundant fine inclusions (predominantly magnetite).

The larger, measurable plagioclase grains of the basalt are very calcic: 85 to 90 percent anorthite. The mineral rarely forms phenocrysts and is only faintly zoned. Pericline twinning is nearly as common as albite twinning.

OLIVINE BASALT WEST OF ABERDEEN

Scattered about in the large embayment of the range front near the mouths of Armstrong Canyon and Division Creek, west of Aberdeen, are 18 basalt cinder cones and associated lava flows. These basalts cover a much larger area than the other basalts. Because of the similarity and apparent contemporaneity of these volcanic rocks they will be considered together.

The Sawmill Canyon basalt flows that reach the mouth of Sawmill Canyon appear to be covered by basalt cinders from the large cinder cone north of the canyon mouth. Therefore this cinder cone and all the other prominent basalt cinder cones to the north are tentatively considered younger than the basalt flow of Sawmill Canyon. At the mouth of Armstrong Canyon, basalt cinders and lava are covered by moraines of the latest glaciation (Tioga). Hence the basalt cinder cones, like the flow of Sawmill Canyon, are also interglacial in age. On casual inspection these well-preserved cinder cones give the erroneous impression of having erupted very recently. The lower end of the basalt flows is commonly covered by alluvial wash from the mountains, and only isolated high points and pressure domes are exposed; at least 10 to 30 feet of valley fill have accumulated since eruption.

The position of the basalt cinder cones along straight or curved lines leaves little doubt that the basalts were erupted along fault planes. The string of 6 small cinder cones situated at the foot of the range between Division and Taboose Creeks are fairly evenly spaced along a remarkably smooth curve over 2 miles long; the line of cones is parallel to two fault scarps on the shoulder of Shingle Mill Bench to the west. Three larger cinder cones on the alluvial slope to the east of the above-mentioned cones also line up along a supposed fault. The northern continuation of this line passes through Red Mountain, a large cinder cone 2 miles north of the mapped area and presumably on the same fault. Nearly parallel with the line connecting these cones is a recent fault scarplet, upthrown on the west, which cuts the western sides of two of the cinder cones.

The relative ages of cinder cones and related lava flows are difficult to ascertain. However, the most northerly of the cinder cones on the range front provides some information on this problem. Lava flows on both the north and south of this cone, which appear to originate near the cone, swing around the cone. The evidence suggests that both these lava flows are younger than the cone. The large lava flow south of Goodale Creek appears to have originated from three simultaneously erupting vents located along a northwestward trending fault.

Along the range front north of Sawmill Creek a large basalt cinder area appears to have been erupted from at least two vents. South of the mouth of the canyon three circular areas of basalt cinders are aligned with the two northern vents. These small dark areas on the sandy slope of the range front are about 200 yards in diameter. The darker color is due to a scattered accumulation of basalt ash and lapilli less than three-fourths inch in diameter, which is scattered about on the thick slope-wash mantle of granitic sand and gravel. The areas must then represent the very initial breakthrough of basaltic material to the surface, along a fault plane or other linear zone of weakness. For some reason, after the first small amount of basalt was explosively spattered out in the form of basalt lapilli the activity stopped. Another small area of basalt cinders is present on the range front just north of Black Canyon.

The basalt cinder cones are composed of unconsolidated, extremely porous and scoriaceous pyroclastic basalt. Well-formed basalt bombs are common on the cinder-cone slopes, but much more abundant are angular blocks of basalt. Curved spines are rod-shaped masses with longitudinal fluting are abundant in the pyroclastic debris; they probably represent stringers and ends of volcanic bombs, modified by flight and broken up upon striking the ground.

Most of the basalt in cinders is redder in color than that of lava flows. The cinders are especially red around the vent itself. The red color is due to the oxidation of fine magnetite in the groundmass to hematite; the presence of volatile emanations at the vent would favor this oxidation.

Several of the cinder cones have well-developed craters. The large cone south of Taboose Creek has a crater with nearly 200 feet of closure. Generally, however, the cones are unsymmetrical and horseshoe shaped, one side of the crater wall being apparently blasted away by explosive activity. However, in the few perfect cones, it was noted that the south wall of the crater is generally the highest. This fact is true for two of the well-developed cones south of Taboose Creek and for Red Mountain, a large cone just north of the mapped area. The most reasonable explanation for this phenomenon may be that at the time of volcanic activity in the area, a prevailing north wind modified the shape of the cinder cones as they developed, causing more of the erupting cinders to fall south of the volcanic orifice.

The form of the lava flows which issue from the base of many of the cinder cones indicates that the lava was fluid at the time of eruption. The surface of the flows in places presents smooth contours and ropy structures. However, in general the surface of the flows is extraordinarily broken and rugged. Continued flow from beneath a crusted surface has formed lava tubes and jumbled mazes of breakdown material when the tubes collapsed. Tumuli and spatter cones developed as a result of hydrostatic head beneath crusted-over lava flows. Pressure ridges 40 and 50 feet high are both parallel to and perpendicular to the direction of flow. Columnar jointing is rare or poorly developed in these basalt flows; however, the blisters or tumuli commonly have a crude columnar jointing perpendicular to their domical layering.

On the surface of some of the larger lava flows are channelways which apparently continued to carry lava downhill after the major part of the flow was consolidated. These trenches are 150 to 250 feet wide and possess a flat floor some 10 to 20 feet lower than the general surface of the surrounding lava flow. A large channelway north of the Division Creek powerhouse can be traced for more than 1½ miles.

The basalt of the area west of Aberdeen is similar microscopically to the basalt of Sawmill Canyon. It possesses an intergranular texture, a somewhat fine grained groundmass (0.02 to 0.15 mm), and phenocrysts of olivine, pale-brown pyroxene, and plagioclase. It differs from the basalt of Sawmill Canyon in being less porphyritic (about 10 percent phenocrysts), and in possessing fewer augite phenocrysts relative to olivine. By means of modal analyses, as shown in

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figure 57, this basalt can be distinguished from the similar basalt of Sawmill Canyon. Olivine occurs as the largest and most abundant phenocrysts. Generally the olivine is remarkably fresh and unrimmed by its common alteration products. The olivine carries about 15 percent fayalite ($2V=90°$). Augite occurs both as phenocrysts and as an abundant constituent of the groundmass. The phenocrysts commonly show simple or oscillatory zoning. The plagioclase of the larger crystals has a composition of 80 percent anorthite.

SOURCE OF THE BASALTIC MAGMA

All of the Cenozoic olivine basalts are characteristically low in silica. The low content of silica is shown by a richness in olivine, by the calcic character of the measurable plagioclase (commonly bytownite), by the presence of spinel in the groundmass of the dikes of Division Creek, and by the presence of spinel-bearing dunite inclusions in the basalts of Oak Creek. These features suggest that the basalts came from great depth (perhaps near the Mohorovicic discontinuity) and reacted little with the granitic walls of the fissure which conducted them to the surface.

GLACIAL DEPOSITS

During Pleistocene time the area of the Mount Pinchot quadrangle underwent glaciation at least twice. The entire summit region of the Sierra was profoundly glaciated and large valley glaciers descended the major canyons on the western slope. The eastern escarpment of the range was less extensively affected, though glaciers occupied every major canyon and descended nearly to the upper limit of the alluvial slope above Owens Valley (between 6,000 and 8,000 feet).

At the present time small permanent snowfields persist the year around in north-facing cirques above 11,000 feet. North of the Mount Pinchot quadrangle small glaciers are active on the eastern flank of the Palisade group.

The glaciation of the east scarp of the Sierra Nevada in the Mount Pinchot quadrangle contrasts markedly with glaciation of few miles to the north, where moraines in the Big Pine and Bishop quadrangles extend onto the piedmont slope. The Pleistocene glaciers in the area mapped were small, terminating in the deep canyons of the east scarp rather than spreading out on the piedmont slope. Consequently their moraines are largely confined to canyon bottoms, where the form has been partly destroyed by stream erosion and talus creep.

The more southern latitude may account for part of the smaller size of glaciers from north to south. However, the relative sizes of the catchment basins is a factor of considerable significance. Farther north the eastern flank of the Sierra Nevada is broad and irregular, measuring up to 18 miles from crest to mountain front, whereas in the Mount Pinchot quadrangle this distance averages 8 miles.

A complete interpretation of the glacial history of this region would necessarily involve close comparison of the moraines of the mapped area with those better exposed and preserved to the north. The following discussion is based mainly on the interpretation of aerial photographs and is only tentative.

Workers studying the glacial history of the Sierra Nevada (Blackwelder, 1937; Putnam, 1949) have shown that there were at least four ice advances in the Pleistocene. From oldest to youngest they are: McGee, Sherwin, Tahoe, and Tioga. Constructional forms have been destroyed on the older moraines and are generally preserved only on the two younger ones. Since the moraines of the Mount Pinchot quadrangle still possess their original morainal form, they are tentatively referred to the last two glacial advances: Tahoe and Tioga.

MORAINES

Glacial moraines of two ages can be clearly seen in the valleys of Sawmill Creek and South Fork of Oak Creek. The moraines still show well-preserved topographic forms. Nested lateral moraines lie side by side along the north wall of Sawmill Canyon. The outer, older moraine is composed of debris which is more thoroughly decomposed and disintegrated. The older moraine lies below the olivine basalt flow of Sawmill Canyon and the younger moraine lies on top of the basalt. The older moraine is believed correlative with Tahoe glaciation and the younger moraine, called The Hogsback on the topographic map, with the Tioga glaciation. However, The Hogsback has been sharpened by the erosive action of the two forks of Sawmill Creek that flank it, hence its original morainal form has been altered.

In the South Fork of Oak Creek lateral moraines of two ages are preserved. The older and larger moraine has a more complete development of soil, and a predominance of dense, siliceous metamorphic rocks preserved on the surface. Granitic fragments, more subject to weathering, have been selectively decomposed. Actually, the form of the older moraine (Tahoe) suggests that it may represent two ice advances rather than a single one. The younger, inner moraine (Tioga) is better preserved and covered with many more boulders, of which the larger part are granitic.

The moraines of the other canyons on the eastern scarp have been compared with those of Sawmill and Oak Canyons as to topographic form and the weathering of the till and have been mapped accordingly. Earlier moraines are preserved in the mouths of Taboose Creek (just north of the mapped area), North Fork of Oak Creek, and Independence Creek.

ROCK GLACIERS

Rock glaciers occur in the high cirques on north- to northeast-facing slopes. The lower end of the rock glaciers is a steep front of large, angular blocks resting at about the angle of repose. The average elevation of the base of this front is about 11,500 feet. The rock glaciers are from one-half to one-fourth mile in length and are fed by talus cones at their upper end. On the upper surface, parallel to the lower end and convex downcanyon are parallel ridges 10 to 15 feet high and about five times as far from crest to crest. The upper end of the rock glacier is ill defined and merges with talus, ice, or bedrock at the base of the cirque headwall.

GEOLOGIC STRUCTURE

PRE-CENOZOIC STRUCTURE

In the Mount Pinchot quadrangle, the deformation that accompanied the emplacement of the composite batholith was so intense as to all but obscure any older structures that may have been present in the wallrock. The Sierra Nevada composite batholith was emplaced in a large number of small intrusive increments, or plutons. Most plutons are elongated to the northwest, and during emplacement each one is believed to have compressed its walls in a direction principally at right angles to its elongation. Judging by the span of ages of individual plutons in the Yosemite region (Curtis, Evernden, and Lipson, 1958), this period of granite intrusion lasted at least 18 million years, and possibly over 50 million years. Therefore, it is difficult to demonstrate that any structural feature, whether on the scale of the map or outcrop, predates the period of emplacement of the batholith and was formed by an earlier orogeny.

The pre-Cenozoic structural features will be discussed in two groups: those of the wallrocks, both older than and related to the emplacement of the batholith, and those of the granitic rocks themselves, formed during the period of emplacement.

STRUCTURES OF WALLROCKS

The older sedimentary and volcanic rocks that compose the walls of the plutons have been thrown into a series of vertical, highly metamorphosed septa. The septa are tabular masses of metamorphic rock that commonly separate two plutons of different ages.

Where original bedding can be recognized in these septa it is generally parallel to the schistosity. Steep-dipping lineation in the plane of the schistosity also is common. The lineation is marked by alinement of elongated minerals, trains of mineral aggregates, crumple axes, and stretched pebbles and fragments in volcanic breccia. The lineation is believed to be caused by squeezing induced on the metamorphic rocks by the emplacement of the granitic plutons. Apparently the squeezing produced upward extension and resultant stretching, as well as horizontal slip on foliation planes and crumpling around vertical axes.

Most septa extend downward to an unknown depth. The septum on Mount Bago, at the southern margin of the quadrangle, passes gradationally with depth into a migmatized mafic hybrid rock. The small septum on the Spook pluton, on the other hand, is cut off sharply at its base at right angles to its bedding schistosity.

On a map scale most of the wall rocks are unfolded, homoclinal masses. However, the calcareous pendant east of Sawmill Lake is synclinal, and the biotite schist septum between the Lamarck and Red Mountain Creek plutons swings over to the east in an anticlinal arch.

STRUCTURES OF GRANITIC ROCKS

The most conspicuous internal structure of the granitic rocks is the foliation marked by the orientation of flattened mafic inclusions. Of the foliation measurements in the granitic bodies 90 percent were made on these oriented mafic inclusions. The inclusions may be disc, rod, or almond shaped and define respectively planar foliation, lineation, or planar foliation and lineation. Preferred orientation of mafic minerals (especially hornblende), K-feldspar phenocrysts, and, rarely, small feldspar crystals also mark the foliation or lineation. Schlieren and swarms of mafic inclusions, where present, are commonly parallel to the foliation marked by the orientation of mafic inclusions.

Orientation of the mafic inclusions could have been accomplished by: (1) orientation of previously inequidimensional inclusions by lamellar flow, or (2) plastic stretching and distortion of roughly equidimensional inclusions as a result of being surrounded by a viscous flowing material. The fact that the inclusions are so uniform in size, orientation, and shape suggests that mechanism 2 is probably the more important.

Foliation is generally steeply dipping and follows the walls of the plutons. Where two plutons are in contact (the foliation of the younger is generally more nearly parallel to the contact than that of the older. Three intrusives have a well-developed lineation, the Dragon and Arrow plutons, and the Tinemaha granodiorite. In the Dragon and Arrow plutons, the lineation is nearly horizontal and parallel with the long axis of the intrusion; it is probably the result of arching and stretching near the top of a domical intrusion. The direction of the lineation parallel with the long axis of these plutons suggests that the intrusions grew not from a dikelike mass which expanded laterally, but from a more equidimensional mass which expanded longitudinally.

Gently dipping aplite and granitic dikes are commonly sent off by several of the more silicic granitic plutons. Wallrocks without a strong schistosity, such as mafic hybrid rock or other granitic plutons, are the most favorable hosts for these flat dikes. The dikes dip generally less than 30° into the pluton from which they originate. They may be relatively thin (less than 10 feet), as the dikes on Striped Mountain sent out by the Lamarck granodiorite, or quite thick (up to several hundred feet), as the dikes on Painted Lady (mountain) sent out by the Bullfrog pluton. The flat dikes swarms are believed to be injected into feather joints opened in the wallrock by the upward movement of the granitic pluton parent to the dikes.

The Arrow pluton is cut by a large number of small east-trending left-lateral strike-slip faults. The plastic drag and offset of aplite dikes and other structures in the granodiorite show that the faulting was deep seated and probably occurred not long after intrusion of the Arrow pluton.

Three large east-trending left-lateral strike-slip faults near Rae Lakes have the same trend and sense of displacement as the faults cutting the Arrow pluton. These faults cut and offset the Cotter, Bullfrog, Dragon, and Diamond plutons. Because of their similarity in strike and offset, and because they are on strike with many of the small faults of the Arrow pluton, they are believed to be related to the faults of the Arrow pluton and consequently to be earlier than the Cenozoic faults that occur on the east front of the range.

CENOZOIC STRUCTURE

According to Matthes (1930), the Sierra Nevada region was eroded to a surface of low relief after the late Mesozoic orogeny. During the Eocene epoch the area underwent upwarping, and intermittent uplift continued to early Pliocene. After a period of relative stability, vigorous tilting toward the west began in early Pleistocene time, when the faults bounding the block on the east seem to have originated. Mayo (1941, p. 1061) points out that these faults roughly follow the eastern margin of the granitic mass of the Sierra and appear to be localized along the zone where competent granite gives way to cleaved and bedded metamorphic rocks.

EAST-SIDE-DOWN FAULTS

The steep range-front escarpment in the eastern part of the quadrangle is part of the eastern border of the west-tilted Sierra Nevada block so commonly pictured in textbooks as a simple normal fault. The fault system is, however, far from simple. In detail the scarp possesses an irregular plan of salients and reentrants, and a steplike profile of shoulders and scarps. Evidence for Pleistocene (and possibly late Tertiary) faulting is well shown along the range front. The criteria for recognizing and mapping the faults are as follows:

Physiography.—Abrupt changes in slope at the foot of the range and on the east scarp indicate the position of faults. The presence of alined canyons and ridges and of actual fault scarps.

Fault trace.—In several places an actual fault trace can be seen in the field or on aerial photographs. The trace is a thin line a few feet wide along which there may or may not be a visible scarp. Probably most of the fault traces are caused by only the most recent movement along a fault surface. The visible trace appears to be the result of a slight concentration of vegetation along the fault. Where the Independence fault crosses Indepedence Creek and for about one-half mile north, aerial photos clearly show the actual trace of the fault traversing gorges and spurs.

Fault contacts.—The faults have had sufficient throw to offset rock units in some places an appreciable amount. The base of the lava flows of Oak Creek have been offset by several faults. The Independence fault separates the Independence pluton from the Tinemaha granodiorite along a 4½-mile fault contact.

Alinement of cinder cones.—In the volcanic field south of Taboose Creek, basalt cinder cones commonly occur along straight or curved lines, generally located at the foot of the range front. In all probability these lines of volcanoes represent faults up which basaltic magma has been conducted to the surface. The most striking of these concealed faults, at the east foot of Shingle Mill Bench, is 2½ miles long and has 6 basalt cinder cones fairly evenly spaced along its length.

Alinement of springs.—In several localities springs issue from faults revealed by one of the criteria listed above. South of the mouth of Sawmill Creek are 2 springs on the same fault along which 4 cinder cones are located. A fault has been mapped at Scotty Spring coinciding with a line of several springs at the base of the range.

A large fault, called the Independence fault, trends in a northerly direction across Independence Creek and is downthrown on the east. The fault extends 5 miles south of the quadrangle; where it passes south out of the range, near the mouth of Bairs Creek, there is an offset in the range front. Within the quadrangle the fault is over 8 miles long, tracing a zigzag map pattern as it cuts across valley and spur. The dip of the fault, calculated by the effect of topography on strike, is about 45° or 50° E.

The Independence fault cuts, but only slightly offsets, a large vertical granitic dike south of Pinyon Creek; hence a strike-slip component is only very slight or is lacking. The strong topographic effect of the fault and the fact that the Dragon and Independence plutons are in contact along the fault for 4½ miles suggests that considerable movement has taken place along the fault, probably a dip-slip component of several thousand feet. The fault cuts a moraine of the earlier glaciation in Independence Canyon, though this latest displacement is only a few feet.

The nature of the range-front faulting east of the Independence fault is not clear. The presence of bedrock knobs projecting through the alluvium near the east boundary of the quadrangle indicates that the surficial deposits are not thick in this area. Furthermore, measurements of gravity by L. C. Pakiser and M. F. Kane of the U.S. Geological Survey (oral communication) show no steep gravity gradients or anomalies east of the main Sierra Nevada front within the quadrangle. This suggests that there are no major buried faults in the eastern part of the quadrangle. Probably the bedrock surface east of the Independence fault is titled eastward and offset by relatively small faults.

The major east-side-down faults north of Lookout Point follow closely the base of the range and form a deep reentrant of the range front at Goodale Creek. Basaltic cinder cones are alined along these faults indicating that they are major fractures which extend to great depth. In addition to the faults at the base of the range, other major east-side-down faults half way up the range front form several prominent benches (fig. 2). These benches are warped and cut by west-side-down faults.

WARPING AND WEST-SIDE-DOWN FAULTS

North of Lookout Point many small normal faults have been mapped on range-front benches. These faults have a west-side-down displacement; they dip rather gently westward, approximately 45° or less, in the opposite direction of the principal normal faults which bound the range. These faults are present on Shingle Mill Bench, the bench south of Armstrong Canyon, and the bench south of Sawmill Canyon. Stecker Flat, north of Taboose Creek in the Big Pine quadrangle, is also traversed by many west-side-down faults, which appear to be related to those of Shingle Mill Bench.

The west-side-down faults are commonly better preserved than the dominant east-side-down faults. Erosion tends to emphasize the scarps of the antithetic faults, for streams cut down behind them and etch the scarps into greater relief. On the east-side-down fault scarps, streams erode the upthrown side, and alluviate the downthrown side, thus obliterating evidence of the fault rapidly.

The surfaces of the benches on which the west-side-down faults are present slope eastward and appear to be downwarped toward Owens Valley (fig. 2). However, the displacement of the west-side-down faults is such as would tend to make the surfaces of the benches slope westward. Hence tilting and rotation of fault blocks eastward is perhaps the dominant process.

West-side-down faulting is clearly related to warping in other areas along the Sierra Nevada front. Bateman (1958, p. 114) shows that southwest of Bishop the entire Sierra front is one large warp with a structural relief of 10,000 feet; west-side-down faults are common on this warp. A large number of west-side-down faults are believed related to a large warp on the east front of the Carson Range, southwest of Reno, Nev. (G. A. Thompson, oral communication).

JOINTS

The granitic rocks of the Sierra Nevada are cut by a large number of steeply-dipping joints. These joints appear to belong to a conjugate system of two sets approximately at right angles to one another. The joints cut across the contacts between the various granitic plutons; for this reason they are believed to be younger and unrelated to the emplacement of the plutons.

The joint system has not yet been thoroughly studied. The following discussion of the joints is largely based on a study of aerial photographs, which reveal the conjugate systems of steep-dipping joints. In addition to the steep-dipping joints, flat-dipping joints are rather common. These are commonly nearly parallel to the present rock surfaces and presumably resulted from the relief of pressure caused by erosion of overlying rock (Matthes, 1930, p. 115).

Most of the steep joints can be traced for at least several hundred feet, and some can be traced for a mile or more. The shorter ones are relatively straight, but the longer ones are apt to be gently curved. The steep joints are varied in the closeness of their spacing, their length, and their breadth. In a given area joints are generally more widely spaced in the coarser grained than in the finer grained granitic rocks. An example of this difference is seen where a fine-grained aplite dike traverses a coarse-grained quartz monzonite. The joints may be 2 feet apart in the quartz monzonite and 1 inch apart in the aplite, the strike and dip remaining approximately the same. The Bullfrog pluton, which is the largest in the quadrangle and is made up of one of the coarsest grained rocks, is cut by very widely spaced joints. The joints are from 10 to 50 feet apart. Because of this wide spacing of joints, the rock of this pluton is very resistant to frost wedging and glacial plucking and forms conspicuous ridges, domes, and highlands.

Plate 2 is a compilation of the master joints visible on aerial photographs. A comparison of this compilation with the geologic map shows that the joint systems cross interpluton contacts with little or no change in strike. The joints are not prominent in the finer grained and more heterogeneous metamorphic rocks.

The generally north-trending set shows a distinct regional curve convex to the west. In the northwest part of the quadrangle, the two sets of joints strike northwest and northeast. In the middle and southern parts of the quadrangle, the northeast set swings to the north, and the northwest set, to the west and even to the southwest.

The conjugate joint sets are younger than emplacement of the granitic rocks (Cretaceous), for they cut interpluton contacts. The fact that the joints are nearly vertical and about at right angles, and show only gradual changes in strike on the scale of a 15-minute quadrangle, suggest that they were produced by regional forces acting in a horizontal direction. There is some suggestion that the general zigzag pattern of the range-front faults may be in part controlled by faults following the joints (pl. 2).

MINERAL DEPOSITS

DIVISION CREEK MINES

The Division Creek mines actually lie in Armstrong Canyon north of Division Creek. They were presumably named after Division Creek because the road that leads to them follows Division Creek from Highway 395. The deposits lie in or near a metamorphic septum in a structurally complex area between the northern mass of the Mule Lake pluton on the west and a complex of mafic plutonic rocks on the east. The septum is intimately injected by numerous dikes of at least five intrusive periods (fig. 59).

The main part of the metamorphic septum is composed of metamorphosed limestone, much of it largely converted to tactite, some of which contains scheelite. However, other metasedimentary rocks are present, especially calc-hornfels. The metasedimentary rocks are intimately mixed with fine- and coarse-grained dioritic rocks. Some of the mafic rocks are dikes belonging to the Indepedence swarm, others are associated with the older dioritic complex to the north, and the possibility exists that some are metamorphosed volcanic rocks. Dioritic rocks that appear to be meta-andesites are exposed with limy rocks in the Valley View mine east of the main metasedimentary septum.

After mixing and intrusion of the metasedimentary rocks by the rocks of the dioritic complex, the complex was intruded and altered by subhorizontal dikes and sills that closely resembles the neighboring Mule Lake granodiorite. Next, nearly vertical thin dikes intruded the area; many of the dikes are microdiorites and others are intermediate in composition. These dikes, which presumably belong to the Independence swarm, are cut in turn by three types of granitic dikes: (1) large fine-grained quartz monzonite dikes apparently belonging to the Siberian pluton and related to some of the tactite, (2) large vertical

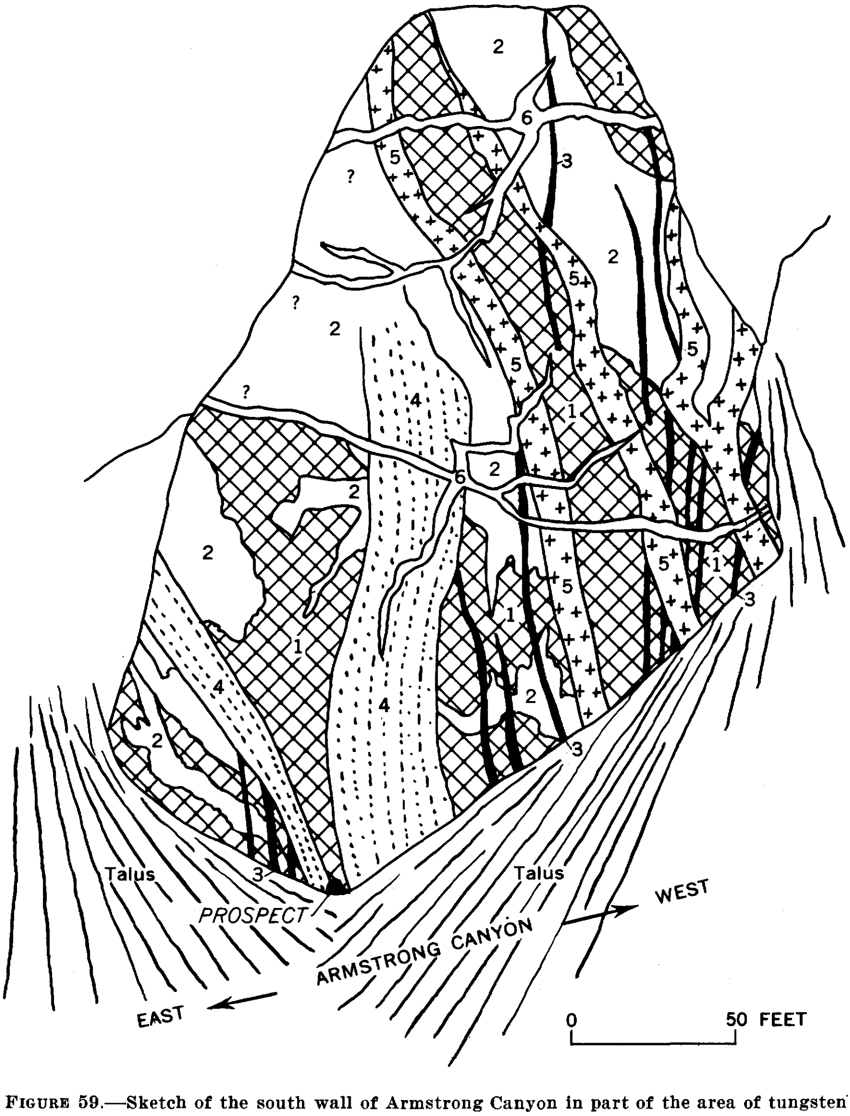

FIGURE 59.—Sketch of the south wall of Armstrong Canyon in part of the area of tungsten mineralization.

Rock units in order of decreasing age are:

1. Metasedimentary rocks (mainly marble and tactite), highly deformed.
2. Dikes from Mule Lake pluton (light-colored, fine-grained granodiorite).
3. Dikes (mainly mafic dikes, some light-colored dikes).
4. Dikes from the Siberian pluton.
5. Aplite and pegmatitic dikes.
6. Flat-dipping aplite dikes.

aplite and pegmatitic dikes, and (3) thin subhorizontal aplite dikes.

Tungsten production from these mines has been small. Thirty tons of scheelite-bearing tactite of approximately 1.75 percent WO_3 had been mined through 1941 (Tucker and Sampson, 1941). During the period 1952 to 1956 about 40 tons of ore was shipped averaging approximately 1 percent WO_3. The property is now idle.

KEARSARGE DISTRICT

The gold-silver mines on Kearsarge Peak in the southern part of the quadrangle were discovered in the fall of 1864. The discovery was made by Thomas W. Hill and others who prospected from a lumber camp at what is now Grays Meadow on Independence Creek (Chalfant, 1933, p. 235). The Kearsarge mining district was named after a Union battleship, in a counterstroke against Confederate sympathizers who shortly before had named the Alabama Hills west of Lone Pine after a Confederate privateer.

Work on the mines continued through 1865 and 1866. A ten-stamp mill was built and began operations in 1866. In the spring of 1867 a snow avalanche struck the mining camp on the slopes of Kearsarge Peak, destroyed 11 cabins, buried others, and killed 1 person. Activity continued in the district until about 1870, but the mines have been nearly idle since that time.

A series of northwest-dipping subparallel fissure veins a few inches to many feet wide strike northeastward across the Dragon pluton and into the south end of the Oak Creek metavolcanic pendant. They contain altered fault gouge and commonly quartz pods and stringers. Pyrite is common in all the veins. Gold and silver-bearing galena were apparently important ore minerals. In the early days of the district, some of these veins yielded considerable quantities of ore, rich in both gold and silver. One lot of 10 tons of ore is said to have yielded $900 per ton (Goodyear, 1888).

The mineralized veins of Kearsarge Peak lie south of the small Sardine pluton. The veins strike at about right angles to the northwest elongation of the pluton; most of them dip steeply northwest toward the pluton. These relations suggest that the fractures were opened by the force of intrusion of the Sardine pluton and mineralization was accomplished by late fluids from the crystallizing Sardine alaskite.

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INDEX

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