Geology and Tungsten Mineralization of the Bishop District California

GEOLOGICAL SURVEY PROFESSIONAL PAPER 470

Prepared in cooperation with the State of California, Department of Conservation, Division of Mines and Geology

Geology and Tungsten Mineralization of the Bishop District California

By PAUL C. BATEMAN

With a section on Gravity Study of Owens Valley

By L. C. PAKISER *and* M. F. KANE

And a section on Seismic Profile

By L. C. PAKISER

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CONTENTS

ILLUSTRATIONS

[Plates are in separate volume]

TABLES

GEOLOGY AND TUNGSTEN MINERALIZATION OF THE BISHOP DISTRICT, CALIFORNIA

By Paul C. Bateman

ABSTRACT

The Bishop district is in eastern California about midway between Reno, Nev., and Los Angeles, Calif. The area mapped comprises a little more than 800 square miles and includes the Mount Tom, Bishop, and Big Pine and the northeastern part of the Mount Goddard 15-minute quadrangles. The crest and eastern escarpment of the Sierra Nevada occupy the western half of the area, and the western foothills of the White Mountains lie along the eastern margin. Between the Sierra Nevada and White Mountains is a structural trough that contains Owens and Round Valleys and the Volcanic Tableland. Altitudes range from approximately 4,000 feet on the floor of Owens Valley to more than 14,000 feet in the highest peaks of the Sierra Nevada.

The White Mountains are underlain chiefly by strongly deformed sedimentary strata, the Sierra Nevada by granitic intrusive rocks in which remnants of metamorphic and mafic igneous rocks are scattered, the Volcanic Tableland by rhyolite tuff, and Owens and Round Valleys by alluvial deposits and a few masses of olivine basalt. A wide range of ages is represented, but much more of the record is missing than is present. The sedimentary strata of the White Mountains are of late Precambrian(?) and Early Cambrian ages; the scattered metamorphic remnants in the Sierra Nevada are of Paleozoic and early Mesozoic ages; the granitic intrusive rocks are of Cretaceous age; and the rhyolite tuff of the Volcanic Tableland, the scattered masses of olivine basalt, and the alluvial deposits are of Cenozoic age, chiefly Quaternary.

The rocks and structures are divisible into three groups: those formed before the Sierra Nevada batholith was emplaced, those formed at the time of emplacement of the batholith, and those formed after the emplacement. The rocks of the oldest group include the Precambrian(?) and Cambrian strata of the White Mountains, and the remnants of Paleozoic and Mesozoic strata in the Sierra Nevada. In the White Mountains more than 10,000 feet of sedimentary rocks are present in which *Olenellus* has been found in the upper 3,500+ feet. In the Sierra Nevada, remnants of sedimentary rocks probably range in age from Cambrian or Ordovician to Pennsylvania and Permian(?); stratigraphically overlying (probably unconformably) metamorphosed metavolcanic and intercalated metasedimentary rocks are probably of Triassic(?) and Jurassic age. In Late Jurassic or Early Cretaceous time these rocks were folded and faulted along north- to northwest-trending axes.

The Sierra Nevada batholith is a mosaic of discrete intrusive masses of plutonic rock, which are in sharp contact with one another or are separated by thin septa of metamorphic or mafic igneous rock. The intrusive rocks range from diorite, quartz diorite and hornblende gabbro to alaskite. Most of the intrusive rocks are grouped into formations on the basis of composition, texture, and intrusive relations, but a few are unassigned. In general the rocks were emplaced in order of increasing silica content, but with many exceptions. Some intrusives are zoned, and their interiors are more siliceous than their margins.

The following features indicate that most of the plutonic rocks are magmatic:

1. Contacts of individual plutons with each other and with older rocks are sharp, clean, and regular.
2. Finer grained rock is present in the marginal parts and apophyses of some plutons.
3. The geometry of the wall rocks suggests that some dislocations were caused by forcible emplacement of magma. In one place a separation of 3 miles seems clearly attributable to forcible intrusion, and in another a separation of 8 miles seems probable. Intrusive breccias are present locally.
4. Internal foliation in the margins of plutons parallels external contacts.
5. The walls of aschistic dikes marginal to some plutons are dilated.
6. Granitization and assimilation effects are confined to amphibolite and other wall rocks that consist chiefly of minerals earlier in Bowen's reaction series than those crystallized in the granitic rocks. The effects are in accord with theoretical expectations of reactions between granitic magma and wall rocks.
7. The metamorphic grade of the wall rocks and of inclusions is that of the amphibolite facies, which is in accord with the temperatures believed to exist in nonsuperheated granitic magmas.
8. Variations in the compositions of the intrusive rocks are in accord with variations predicted from experimental studies of melts.

During cooling and solidification, the granitic intrusives expelled heated solutions that reacted with lime-rich rocks in metamorphic remnants and wall rocks to form contact-metasomatic tungsten deposits. These deposits yielded 1.3 million units of WO_3 (tungsten trioxide) to the end of 1953 and were still at an early stage of exploitation. The Pine Creek mine of the Union Carbide Nuclear Co. contains the largest known deposit of tungsten ore in the district and has yielded more than a million units of WO_3 as well as substantial amounts of molybdenum and copper.

The solutions given off by the cooling magma contained silicon, aluminum, iron, manganese, titanium, tungsten, sulfur, and other metals. These solutions reacted with lime-rich rock to form tactite, a rock composed of dark silicate minerals such as garnet of the grossularite-andradite series, pyroxene of the diopside-hedenbergite series, epidote, idocrase, and amphibole.

Scheelite, the most important tungsten-bearing mineral, is present locally in tactite. The largest tungsten deposits are in tactite composed mainly of garnet and pyroxene, which is believed to have formed at the highest temperatures, but smaller and richer deposits have formed in rock composed of quartz and epidote, locally accompanied by pyrite or pyrrhotite, which formed at somewhat lower temperatures. As the temperature fell quartz rather than silicate minerals was deposited in veinlets; quartz also replaced fractured tactite and adjacent granitic rock. Valuable sulfides were deposited locally with the quartz, but silicification of garnet-pyroxene tactite generally involved removal of scheelite.

Most commercial ore bodies were formed in clean marble because it can react completely with magmatic substances. Deposits formed in impure marble are generally of lower grade because silicate minerals that were formed isochemically by earlier thermal metamorphism could react only very slowly with introduced magmatic substances.

Solutions expelled from the cooling intrusive mass collected along intrusive contacts and moved upward, especially along channelways that were kept open by fracturing caused by regional forces or by forces related to the emplacement and cooling of intrusives. Ore bodies were formed where these solutions were brought into intimate contact with lime-rich rock. Effective traps for solutions include irregularities in the intrusive contact, especially those where granitic rock is wrapped around or protrudes over marble, in fractured lime-rich rocks, and in favorable beds. Small inclusions that contain lime-rich rocks were especially favorable traps; tactite was formed in the tops of many inclusions beneath an already solidified crust of intrusive rock.

The contact-metasomatic tungsten deposits are preferentially related to the most silicic intrusives in the district. Of 53 known deposits, 38 are associated with either the Tungsten Hills quartz monzonite or with alaskite. Furthermore, all 21 deposits that have yielded appreciable amounts of scheelite are closely associated with these two intrusive rocks. Evidently the distribution of these rocks determined the gross distribution of deposits within the district, and the existence of the district in this particular place may be a result of their juxtaposition with Paleozoic carbonate rocks.

After a period of erosion that exposed the batholith and produced a surface of low relief, the Sierra Nevada was tilted westward, probably as part of a broad upwarp that affected areas to the east and to the west. Stratigraphic studies in the San Joaquin Valley, which is on the downslope part of the Sierra Nevada block, show that tilting began in Late Cretaceous to Eocene time, recurred repeatedly during the Tertiary, and culminated in an orogeny in middle Pleistocene time. The structural movements that resulted in the depression of Owens Valley began later than the tilting, probably in late Pliocene time. The subsidence was accomplished by means of countless small increments of movement, each probably of about the magnitude of those that occurred in connection with the Owens Valley earthquake of 1872, and with historical earthquakes elsewhere in the Great Basin. Evidence of repeated movement along many individual faults is clear, and abundant fresh fault scarps indicate that movements have continued to the present, and no doubt they will continue into the future, at a significant rate.

As the trough subsided, alluvial debris eroded from the bordering ranges poured into it. Basaltic cinder cones and associated flows broke out along bounding faults, and the rhyolitic Bishop tuff poured into Owens and Round Valleys from a source to the north.

The White Mountains escarpment and segments of the Sierra Nevada escarpment north of Bishop Creek west of the Tungsten Hills and south of Tinemaha Creek are fault scarps. The intervening span of the Sierra Nevada, however, is warped rather than faulted. This warp is recorded in an old erosion surface, and geological and geophysical evidence show that it continues to slope eastward beneath the alluvial fill of Owens Valley to the fault that bounds the White Mountains, and northward under the Volcanic Tableland, where it bends upward along an east-trending synclinal axis. The maximum depth of alluvial fill, at the base of the White Mountains, is about a mile; the fill thins toward the Sierra Nevada.

The east flank of the warped surface is broken by many antithetic normal faults, which indicate extension; the concept of Owens Valley as a keystone block that progressively subsided between the Sierra Nevada and White Mountains in the apex of an arch or in a zone of structural weakness along the east side of the batholith fits the known facts. En echelon fault systems in the Volcanic Tableland, movements on the faults that formed in 1872, and consideration of the structural pattern suggest compressional forces, even though almost all of the faults are normal.

Both the Sierra Nevada and the White Mountains escarpments have been deeply dissected by streams, and the higher parts of the Sierra Nevada have also been sculptured by glaciers. Since retreat of the last glaciers, erosion has been slight and many glacial phenomena are well preserved.

INTRODUCTION

PURPOSE, SCOPE, AND ORGANIZATION OF THE REPORT

The purpose of this report is to present the scientific and economic-scientific results of a geologic study of the Bishop tungsten district in east-central California (fig. 1). The area is about midway between Los Angeles, Calif., and Reno, Nev., and includes part of the steep eastern face of the Sierra Nevada. An earlier report (Bateman, 1956) was prepared for persons interested primarily in mining and prospecting; the present report is written especially for those interested in the many challenging geologic problems of the district, both economic and academic. Detailed descriptions of individual mines and prospects contained in the earlier report are not repeated here, but certain concepts regarding the tungsten mineralization have been amplified, and the emphasis has been shifted from a descriptive to a genetic standpoint.

The geology of this report is discussed as "Prebatholithic geology," "Geology of the batholith," and "Cenozoic geology." "Prebatholithic geology" contains descriptions of the stratigraphy and structure in upper Precambrian(?) and lower Cambrian strata in the White Mountains, and in Paleozoic and Mesozoic strata in pendants, septa, and inclusions in the Sierra Nevada. "Geology of the batholith" includes descriptions of the intrusive rocks, their composition, structure, and mode of emplacement. Contact-metasomatic tungsten deposits for which the district is well known were formed during the cooling and consolidation of the granitic rocks, and the process of tungsten mineralization is considered. "Cenozoic geology" contains descriptions of alluvial and volcanic deposits of Cenozoic age, and discussions of the Cenozoic structures and the evolution of the landscape.

Geophysical data are presented in two independently prepared sections. L. C. Pakiser and M. F. Kane prepared a section dealing with gravity studies, and Mr. Pakiser prepared a section on seismic studies. Some geologic interpretations made in these sections are not in agreement with interpretations made elsewhere in the report and reflect the views of the authors of these sections.

PREVIOUS GEOLOGIC WORK

One of the early descriptions of the geology of the northern Owens Valley region is contained in a report of the Geological Survey of California, "Geology," by J. D. Whitney, published in 1865; no geologic map accompanies the report. (Whitney's "Geologic Map of the State of California," listed in some bibliographies as having been published in 1873, was not actually ever published.) W. A. Goodyear, who traveled through Owens Valley in 1870 and again in 1888 examining the geology and ore deposits, gives an especially interesting early description of the geology of the area. His report is included in the 8th Annual Report of the State Mineralogist (1888, p. 224–309). A "Preliminary Mineralogical and Geological map of the State of California" published in 1891 by the California Mining Bureau as Map 1 is one of the earliest maps to show differences among the rocks of the region. On this map, the Sierra Nevada west of Bishop is shown to consist of Jurassic and Triassic rocks cut by northward-trending lenses of granite; the White Mountains are shown to consist of northwestward-trending Carboniferous and Permian strata along the range front and of Triassic and Jurassic strata and granite farther east. Several extensive volcanic areas are shown along the Sierra Nevada front in the span between Bishop and Big Pine. A map by J. E. Spurr, published in 1903 as part of a report on a reconnaissance of Nevada south of the 40th parallel and adjacent parts of California, shows a threefold distinction in the northern Owens Valley region of (a) granular or coarse porphyritic igneous rock in the Sierra Nevada, (b) strata of Cambrian age in the White Mountains, and (c) strata of Pleistocene age in Owens Valley. A paper by W. T. Lee (1906) on the ground-water resources of Owens Valley includes descriptions of the surficial deposits and an interpretation of the structure of Owens Valley. In 1912 and 1913 Adolph Knopf (1918) made a geologic reconnaissance of the Inyo Range and eastern slope of the southern Sierra Nevada, which extended into the south half of the area described in the present report. About 20 years later, during the 1930's, a series of structural studies of the metamorphic and intrusive rocks of the crest and eastern slope of the Sierra Nevada was made by Evans B. Mayo. The report (1941) resulting from these studies, called "Deformation in the interval Mt. Lyell-Mt. Whitney, California," deals most fully with the Bishop district. Also during the 1930's, C. M. Gilbert studied the volcanic region north of Bishop, including Volcanic Tableland in the north-central part of the Bishop region (1938; 1941). Bateman and Merriam's geologic map of the Owens Valley region was published in 1954.

The tungsten deposits within the Bishop region have been described in several papers, including my 1956 report on the economic geology of the region. The earliest geologic report on the tungsten deposits was by Adolph Knopf (1917), who visited the deposits of the Tungsten Hills in 1916 when the deposits discovered in the preceding 3 years were being brought into production. Shortly afterward, in the summer of 1918, Esper S. Larsen, Jr., also examined the deposits in the Tungsten Hills as well as the Pine Creek mine in Pine Creek Canyon. His report (Hess and Larsen, 1921) includes sketch maps of the Little Sister, Round Valley, and Pine Creek mines. The next geologic study of the tungsten deposits was not made until 1934, when Randolph Chapman studied the contact metamorphism along the north side of the Round Valley septum, at the Round Valley mine (Chapman, 1937).

After the outbreak of World War II in 1939, the Geological Survey intensified its investigations of the strategic minerals of the United States, and in 1941 published two preliminary papers by Dwight M. Lemmon on tungsten-bearing districts within the area covered by the present report. One of Lemmon's papers deals with the deposits in the Tungsten Hills (1941a), and the other with deposits in higher parts of the Sierra Nevada near Bishop (1941b).

Prior to publication of my report on the economic geology, two preliminary publications based on the present study were issued; one of these is on the Pine Creek and Adamson mines (Bateman, 1945), and the other is on the tungsten deposits in the Tungsten Hills (Bateman, Erickson, and Proctor, 1950).

Brief descriptions of many deposits of both metals and nonmetals are included in reports of county or commodity surveys by the California Division of Mines, in "Mineral Resources of the United States" published by the U.S. Geological Survey and the U.S. Bureau of

Mines, and in "Minerals Yearbook" published by the U.S. Bureau of Mines. Paul Kerr's memoir "Tungsten mineralization in the United States" contains not only brief descriptions of most of the tungsten deposits in the Bishop district but also a theoretical discussion of the means by which tungsten is introduced into masses of metamorphic rocks such at the Pine Creek pendant (1946, p. 17–18, 142–147).

METHODS OF INVESTIGATION

Many of the tungsten deposits in the district were mapped between 1939 and 1944 in connection with strategic minerals studies of the Geological Survey; the data derived from this work served as a nucleus for the study of the district. Fieldwork for the district study was carried out chiefly during the summers of 1946 to 1950, inclusive, although a few weeks were spent in the field in 1951 and 1952. In all, about 30 months were spent in the field. During most of the fieldwork I was aided by one or two temporary assistants. Laboratory and office studies were chiefly by me, but I received some assistance in making modal analyses and in making routine determinations of the composition of plagioclase in granitic rocks. Chemical analyses, and certain mineralogical determinations, were made by members of the Geological Survey. Gravity studies were conducted by L. C. Pakiser and M. F. Kane of the Geological Survey, and seismic studies were made by L. C. Pakiser.

The data collected in regional mapping were plotted in the field on the aerial photographs from which the 15-minute topographic maps covering the area were prepared. These photographs are on scales ranging from about 1:24,000 to 1:40,000—from 1 inch=2,000 feet to 1 inch=3,333 feet. Observations were plotted by use of a simple lens-type stereoscope having a × 2 magnification. The use of a stereoscope permitted more accurate plotting than would otherwise have been possible and aided in planning traverses and in evaluating possible alternative interpretations of data. The data on the photographs were transferred to the topographic quadrangle base maps by multiplex. Later a few observations were added to the maps and some modifications made by simple inspection.

In mapping, it obviously was not possible, because of the rugged terrane, to gain access to all parts of the area. Where the terrane is too rugged to be occupied, however, the exposures are generally excellent, and contacts can be drawn across inaccessible areas with considerable confidence—indeed, many contacts are visible on the aerial photographs, especially when viewed through a stereoscope. The coverage of ground was most thorough in the vicinity of ore deposits and in areas of metamorphic rock; granitic and alluviated areas generally were less thoroughly covered.

Over 1,000 specimens were collected, about half of granitic rocks and about half of metamorphic rocks. More than 600 of the specimens were studied microscopically in thin section. In addition, larger samples were collected for chemical analyses and for age determinations by radiometric methods.

The tungsten mines and prospects were mapped on scales of 1 inch=20 feet or 1 inch=40 feet. Plane table, alidade, and stadia rod were used in the preparation of surface maps, and plane table, open-sight alidade, and tape, or Brunton pocket compass and tape were used for underground maps. Base maps of mine workings were prepared only where none were available from the operator.

Modal analyses of granitic rocks were made by using thin sections of finer grained rocks and stained slabs of coarser grained rocks. Plagioclase determinations and measurements of the optic angles of pyroxenes were made by using a 4-axis universal stage. All colors referred to in the report are in accord with the Rock Color Chart issued by the Geological Society of America (Goddard and others, 1948 (1951)). This chart is based on the Munsell system of color identification.

ACKNOWLEDGMENTS

The study of the Bishop tungsten district is part of a cooperative program between the U.S. Geological Survey and the California State Division of Mines. Study of the mines was largely made possible through the cooperation extended to the writer both by mining companies and by individuals engaged in mining or prospecting. Among the company mine staffs that have extended assistance are those of the Union Carbide Nuclear Co., the Tungstar Corp., and Panaminas, Inc. I am especially indebted to Mr. Lawson Wright of the Union Carbide Nuclear Co., who supplied much of the data on the Pine Creek mine and participated in the integration of these data into a cohesive treatment of the geology of the mine. The following individuals aided materially by supplying information or assistance: D. R. Adamson, Morris Albertoli, "Cap" Aubrey, J. F. Brackett, George Brown, C. W. Churchill, A. L. Covington, C. W. Fletcher, Gale Green, Gerald B. Hartley, Jr., Charles M. Herron, Kenneth G. Irons, E. E. Ives, H. O. Johansen, Robert Kelso, Otis A. Kittle, Victor E. Kral, Stanley Lambert, Fred R. Lee, Mike Millovitch, J. E. Morhardt, F. L. Murphrey, Nick Pappas, A. H. "Salty" Petersen, Joseph Rossi, Bert Shiveley, Robert Simpson, Joseph Smith, Al Stephens, Howard Stephens, Robert Symons, and B. W. Van Voorhis. The logs of water wells bored in Owens Valley were

made available by the Los Angeles Department of Water and Power. Others too numerous to mention also contributed information that has been incorporated in this report.

I was aided in the field by the following members, or former members, of the Geological Survey, most of whom assisted me for one summer field season: M. P. Erickson, 1943; P. D. Proctor, 1946; M. W. Ellis, 1947; J. W. Reid, 1947; R. M. Campbell, 1948; M. F. Carman, 1948; E. D. Jackson, 1949; L. D. Clark, 1949; R. F. Johnson, 1950; R. L. Parker, 1950; H. S. Imholz, 1950; E. M. MacKevett, 1951; Dallas Peck, 1952. In the laboratory, many of the point counts of granitic rocks were made by S. H. Huddleson and A. C. Bettiga.

During the preparation of this report the writer profited from discussions with colleagues of the Geological Survey, and many of the ideas expressed are a result of these discussions. Edgar H. Bailey and Dwight M. Lemmon critically reviewed the manuscript, and vastly improved its accuracy and technical presentation. David B. Stewart served as consultant and critic on parts of the report dealing with the relations of the granitic rocks to experimental data. In the planning of the illustrations Esther McDermott provided council and assistance far beyond the mere mechanics of preparation.

GEOGRAPHY

LOCATION AND ACCESSIBILITY

The study area is in east-central California about midway between Reno, Nev., and Los Angeles, Calif. (fig. 1). It lies between lat 37°00′ and 37°30′ N. and long 118°15′ and 118°45′ W., and includes a little more than 800 square miles. The area is traversed by U.S. Highway 6 between Los Angeles and Tonopah, Nev., and by U.S. Highway 395 from Reno. Access to the Central Valley of California and to the San Francisco Bay area from the study area is difficult because the Sierra Nevada intervenes, and the nearest road crossing, 70 miles north of the area, State Highway 120 through Tioga Pass, is open for travel only during the summer months. U.S. Highways 50 and 40, through Echo and Donner Passes, respectively, 175 and 205 miles to the north, and State 178 and U.S. 466, through Walker and Tehachapi Passes, 130 and 170 miles to the south, generally are open all year.

SURFACE FEATURES

The area includes parts of two high mountain ranges, the Sierra Nevada in the west and the White Mountains in the east, separated by a deep trough that is occupied by Owens Valley, Round Valley, and the Volcanic Tableland. The area is one of considerable relief—altitudes range from less than 4,000 feet on the floor of Owens Valley to more than 14,000 feet in the Sierra Nevada. Nevertheless, great relief is confined to the two ranges; the relief within the trough generally is only about 1,000 feet, except in the Mount Tom quadrangle where fans that flank Round Valley on the west and the Volcanic Tableland north of Paradise Camp extend upward to altitudes of more than 6,000 feet.

The crest and eastern face of the Sierra Nevada occupy more than half of the mapped area. The Sierra Nevada divide culminates in North Palisade Peak at an altitude of 14,242 feet, and most of the named peaks along the crest attain altitudes of more than 13,000 feet. In most places the divide is a "knife-edged" ridge, passable on foot in only a few places. The upper slopes are largely steep-walled glacial cirques that are mantled with talus. Moraines commonly fringe the lower sides of cirque basins, and in the larger canyons extend downward to altitudes as low as 5,200 feet. Below the glaciated zone the slopes are less precipitous but, in most places, are still steep.

The Sierra Nevada escarpment is precipitous only in the northern and southern parts of the area where the average slope of the escarpment exceeds 30°. The middle span between the Tungsten Hills on the north and Fish Springs Hill on the south is a broad convex bulge having comparatively gentle slopes of about 10°. The summit surface of this bulge, a gently rolling, till-covered upland that includes Coyote Flat, Coyote Ridge, and Table Mountain, occupies many square miles between altitudes of 9,000 and 11,000 feet. Northward the surface slopes down to the Tungsten Hills, and eastward and southeastward it slopes into the foothills southwest of Big Pine, and thus forms reentrants with the precipitous escarpments to the north and south. Deep canyons cross these slopes, but broad surfaces between the canyons are composed of soil and weathered rock.

Among the many steep canyons that drain the Sierra escarpment, those of Rock Creek, Pine Creek, Bishop Creek, Baker Creek, and Big Pine Creek are the largest and drain the most extensive areas. The lower parts of all these canyons except that of Baker Creek are accessible by road; and in Rock Creek and Bishop Creek canyons, roads extend to altitudes of 9,500 feet or more.

The average slope of the White Mountains escarpment within the mapped area is generally between 10° and 15°, a trifle steeper than the gently sloping middle span of the Sierra escarpment. Slopes generally are steeper in the south part than in the north part. Inasmuch as the area extends only part way up the slope, altitudes do not exceed 9,600 feet, but summit altitudes

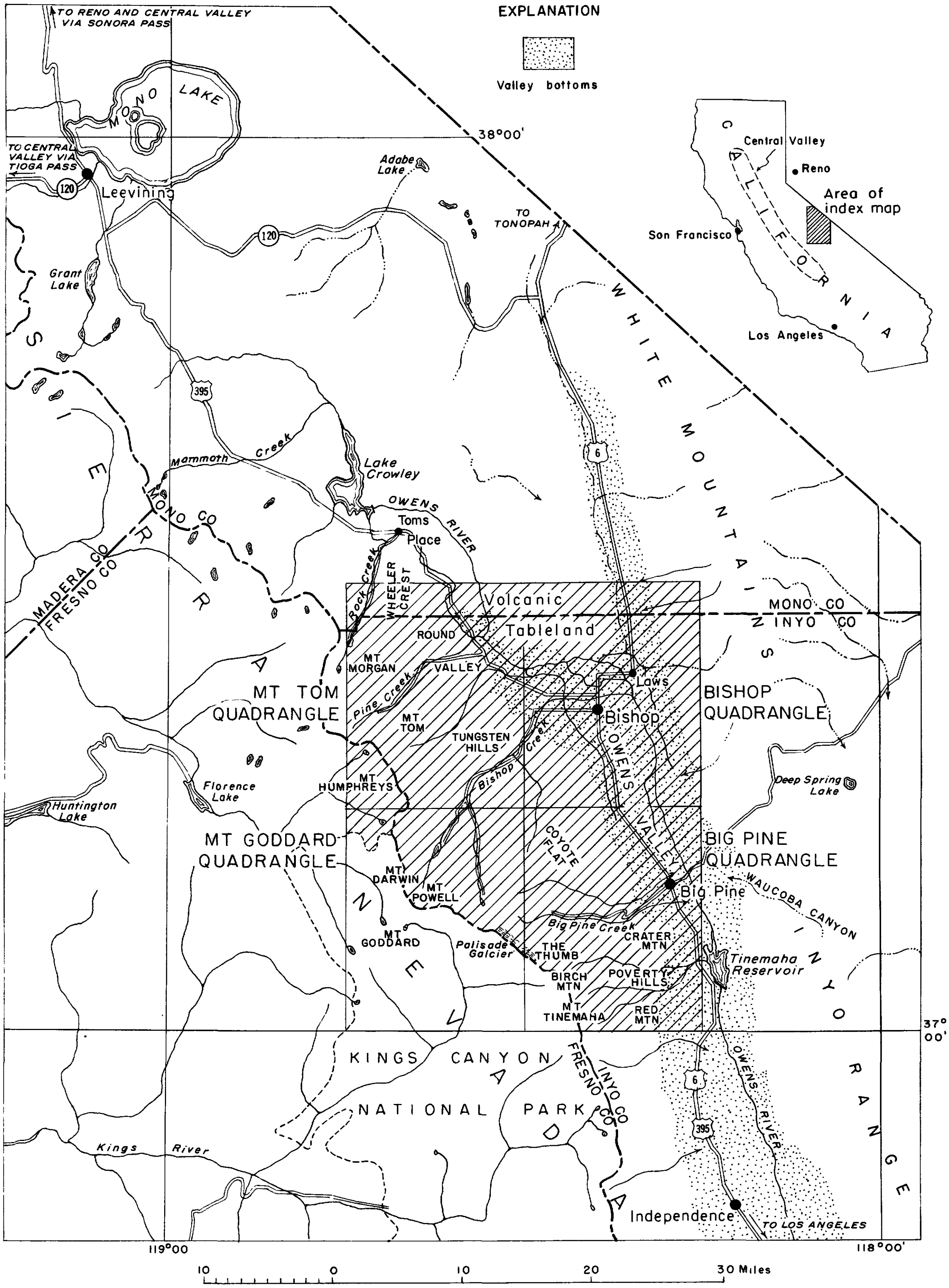

FIGURE 1.—Index map showing the area mapped for this report.

farther east exceed 11,000 feet, and White Mountain Peak to the north reaches 14,242 feet. In the north part, especially north of Silver Canyon, structurally controlled shallow valleys parallel with the trend of the range front are present in the interfluves. The range front is cut by several canyons, the three largest of which, Coldwater, Silver, and Black Canyons, contain perennial streams. The canyons are steep walled, and the streambeds are gravel covered and slope uniformly.

South of Bishop, Owens Valley occupies the entire trough area between the Sierra Nevada and the White Mountains, but a few miles north of Bishop, Owens Valley proper terminates along a steep south-facing escarpment that bounds the Volcanic Tableland. However, a narrow continuation of the valley extends northward beyond the quadrangle boundary between the east side of the Volcanic Tableland and the White Mountains. West of Bishop, Owens Valley is bounded by low river terraces that separate it from Round Valley. The floor of Owens Valley is a flat plain 3 to 5 miles wide, which is flanked on the east and southwest by extensive alluvial fans formed from debris carried out of the White Mountains and the Sierra Nevada. The valley floor slopes gently southward from about 4,100 feet at Bishop to 3,900 feet 16 miles to the south near Big Pine. South of Big Pine several hills rise above the alluvial apron from the Sierra Nevada. These include Crater Mountain, Red Mountain, and a low hill 2 miles northwest of Red Mountain, all of volcanic origin, and Fish Spring Hill and the Poverty Hills, which are upfaulted blocks of the crystalline basement.

Round Valley lies in a reentrant in the Sierra Nevada front which bounds it on the west and south; on the northeast it is bounded by the Volcanic Tableland. The valley is formed by the meeting of coalescent fans from the Sierra Nevada with the Volcanic Tableland. The lower, gentler sloping parts of the fans constitute the valley floor. The floor is a nearly flat area 6 miles long and as much as 2 miles wide, but the entire basin, including the upper parts of the fans, is at least twice as large.

The Volcanic Tableland is a large plateau of rhyolite tuff, which extends into the northern part of the area. It slopes generally southeastward at about 125 feet per mile, except in the southwestern part, where it slopes southwestward into Round Valley. In the western part of the tableland, small rounded hills rise above the general level, and deep gorges have been cut by Rock Creek and the Owens River. In the eastern part the fairly even surface of the tableland is broken by many north-trending fault scarps and contains several undrained depressions. The longest and highest scarp, along the east side of Fish Slough, averages about 300 feet high, but most other scarps are less than a mile long and less than 100 feet high. On the south and east sides the Volcanic Tableland is bounded by escarpments. The average height of the escarpment on the south side is about 200 feet, but it is higher toward the west and lower toward the east. The maximum height of the escarpment on the east side is about 200 feet, but most of it is much lower. Elevated stream-cut terraces flank the Volcanic Tableland on the south and east sides. They are cut chiefly on the rhyolite tuff that composes the Volcanic Tableland, but the south-side terraces and locally the east-side terraces are gravel covered.

The area is drained by the Owens River, which empties naturally into Owens Lake at the south end of Owens Valley, but much of the water in the river has been diverted to an aqueduct which carries about 330,000 acre-feet annually to the city of Los Angeles. The principal tributaries from the Sierra Nevada to the Owens River within the map area are, from north to south, Rock Creek, Pine Creek, Bishop Creek, Baker Creek, Big Pine Creek, and Tinemaha Creek. In the northeast part of the Mount Tom quadrangle the Owens River flows toward the south in the deep gorge that it has cut into the Volcanic Tableland. On leaving the tableland it flows due east along the base of the cliffs that mark the south side of the tableland to the base of the fans along the White Mountains; from there it flows southward along the east side of Owens Valley.

SETTLEMENTS AND CULTURE

The population of Inyo County is about 11,500, at least half of whom live within the mapped area. Bishop, with a population of about 3 000, is the largest town. Smaller settlements are Big Pine, 16 miles south of Bishop; Laws, 3 miles northeast of Bishop; and in Pine Creek Canyon. Rovanna, at the mouth of Pine Creek Canyon, provides housing for employees of the Pine Creek tungsten mine of the Union Carbide Nuclear Co.

The economy of Bishop and of most of the other settlements in Owens Valley is based primarily on tourist trade, but mining and cattle raising furnish additional income. Although the area is more than 200 miles from Los Angeles, the spectacular scenery in the Sierra Nevada and the many well-stocked lakes and streams attract many thousand tourists into the area every year.

The area is also visited for deer, dove, sagehen, duck, and pheasant hunting, for winter sports, especially skiing, and for aerial gliding. Air plunging over the Sierra escarpment rises from the valley floor in a huge vertical updraft that reaches upward thousands of feet (Barnett, 1955, p. 74–75), and many of the world's gliding records have been made above Owens Valley.

Although since World War II the economy of Owens Valley has been based chiefly on tourist trade, it has not always been so. Early discoveries of gold caused some excitement, but the economy of the first settlers in the valley in the early 1860's was founded on sheep and cattle raising and to a lesser extent on agriculture. Irrigation districts were organized, and a network of canals for irrigation was constructed. At the turn of the century a ready market for produce existed in the booming mining camps of Nevada, especially at Tonopah. The need for water in Los Angeles, 200 miles to the south, resulted in the building of an aqueduct to the lower end of the valley, and to acquisition by the city of Los Angeles of most of the arable land within the valley itself.

In 1939 and 1940, as economic conditions improved throughout the nation, tourists appeared in the valley in large numbers, and the tremendous growth of population in southern and central California after World War II resulted in further increases. Many of the tungsten mines, idle or operated only intermittently since World War I, were reopened in the late 1930's because of increased price and demand for tungsten, and have been operated intermittently since then. In 1959, the price and demand were low and only the Pine Creek mine of the Union Carbide Nuclear Co. was in operation.

CLIMATE AND VEGETATION

Except for a few square miles west of the Sierra Nevada divide, the study area is in the western part of the Great Basin, a semiarid region in the rain shadow of the Sierra Nevada. Both temperature and precipitation vary widely in different parts of the area, corresponding to the wide differences in altitude. As in most arid regions, the diurnal range of temperature is large. On the floor of Owens Valley, daytime temperatures in the summer often exceed 100° F, but the nights are comfortably cool. Winter days generally are pleasant with a temperature well above freezing; night temperatures, though below freezing, are rarely as low as zero. Most of the precipitation is during the winter, but summer thundershowers are common. At higher altitudes almost all of the winter precipitation is as snow, but at lower altitudes much of it is as rain. The amount of snow in the valley at any one time rarely exceeds a foot, although as much as several feet has collected.

At higher altitudes the temperatures are progressively lower and the precipitation higher—snow collects to depths of many feet, and on north slopes banks of snow may persist through the summer. During July, thundershowers, accompanied by spectacular displays of lightening, are common in the middle part of the day.

Climatic data supplied by the U.S. Weather Bureau for Bishop and for South Lake are summarized in table 1. Data from the weather station at Bishop, altitude about 4,100 feet, are representative of lower altitudes, and those from the station at South Lake, about 9,600 feet, are representative of higher altitudes. At low altitudes the winters generally are mild enough to permit mining operations all year without undue difficulty, whereas at high altitudes the more severe winter climate requires careful, and generally costly, preparation for successful winter operation.

TABLE 1.—*Summary of weather data at Bishop and at South Lake, Calif.*

[Data from U.S. Weather Bureau, San Francisco office. Average-precipitation data at Bishop are for a 40-year period; all other data are averages of periods of at least 20 years. Temperature data in degrees Farhrenheit, other in inches]

Average—	Bishop airport (alt 4,108 ft)			South Lake (alt 9,620 ft)		
	January	July	Annual	January	July	Annual
Maximum temperature	52.8	93.8	72.8	38.8	69.1	51.4
Minimum temperature	21.1	54.3	36.2	12.5	45.2	27.2
Temperature	36.8	73.6	54.6	25.7	57.1	39.3
Precipitation	1.5	0.1	5.8	2.2	.6	17.4
Snowfall	5.5	.0	17.4	33.9	Tr.	174.0

Native vegetation is sparse on the lower slopes of the mountains and also on the floor of Owens Valley, except along streams, where willow, alder, and other deciduous trees are found. Cottonwoods and Lombardy poplars planted by early settlers are common in settled parts of the valley, and in the towns a wide variety of shade trees has been introduced.

Locally at altitudes of about 8,000 feet manzanita or sage is present, and much of the rolling upland east of Coyote Flat, which lies between 9,500 and 11,000 feet, is sage covered. Conifers are present locally in the mountain regions between altitudes of 6,500 and 10,500 feet in the Sierra Nevada and above 7,500 feet in the White Mountains, but even though some of the larger canyons are heavily wooded, continuously forested areas like those on the west slope of the Sierra Nevada are absent. Singleleaf pine and juniper grow at the lowest altitudes and are the only conifers growing in the part of the White Mountains discussed here. In the Sierra Nevada, juniper and singleleaf pine merge with groves of Jeffrey pine at about 7,500 feet, and at somewhat higher altitudes they are replaced by stands of lodgepole pine. Limber pine and red and white fir are not common in this area. White bark pine is the timberline tree and occurs in low windblown forms. Aspen and willow are present along streams and in other well-watered areas. The trees are generally too small to support a lumbering operation except for small stands of

Jeffrey pine, which are mostly in areas set aside for recreation. Lumber cut farther north in the Mammoth Lakes area or trucked in from Los Angeles can be purchased locally, and very little timber has been cut in the area in recent years. Brown (1954) and Schumacher (1962) give additional information about the flora of the area.

PREBATHOLITHIC GEOLOGY

Rocks older than the Sierra Nevada batholith are found in the Bishop district only as remnants. Strongly folded and faulted strata of late Precambrian(?) and Early Cambrian ages are exposed in the White Mountains, and similarly deformed metamorphosed strata of Paleozoic and early Mesozoic ages are present in the Sierra Nevada in scattered roof pendants, septa, and inclusions (pl. 3).[1] In spite of their scarcity and scattered distribution, these remnants yield information which, when used with data obtained from nearby areas, provides a generalized picture of the events that took place between late Precambrian(?) and Cretaceous time, when the Sierra Nevada batholith was intruded.

During late Precambrian(?) and Paleozoic time many thousand feet of marine sedimentary deposits accumulated in the Bishop district, and in Mesozoic time they were followed by thick dominantly volcanic deposits. Unconformities have been observed in nearby areas and may be present in the Bishop district, within the Paleozoic and between the Paleozoic and Mesozoic. In the Candelaria Hills, 55 miles north-northeast of Bishop, Ferguson and Muller (1949, p. 7–8) report Permian strata resting with angular discordance on Ordovician rocks; they also report an unconformity between the Triassic and Permian. Kirk (in Knopf, 1918, p. 45–46) describes an unconformity between Triassic and Carboniferous rocks in the southern part of the Inyo Range.

The structural movements that caused the folding and faulting of the prebatholithic strata probably began in Triassic time, probably contemporaneously with the onset of Mesozoic vulcanism, and ended in Late Jurassic or Early Cretaceous time.

UPPER PRECAMBRIAN(?) AND LOWER CAMBRIAN SEDIMENTARY ROCKS OF THE WHITE MOUNTAINS

More than 10,000 feet of strongly deformed sedimentary strata of late Precambrian(?) and Early Cambrian age crop out in a southward-tapering area of about 40 square miles mapped along the west slope of the White Mountains. The upper 3,500 feet of strata contain Lower Cambrian fossils, but no diaguostic fossils have been found in the lower 6,000 feet of strata. The entire sequence appears conformable, but the area mapped is too small to rule out unconformities. A summary of the exposed stratigraphic sequence with the approximate thickness of the mapped units follows:

Formation	*Thickness (feet)*	*Remarks*
Harkless formation	1,000+	Fossiliferous; Lower Cambrian.
Poleta formation	1,000+	
Campito formation:		
Montenegro member	600±	
Andrews Mountain member	3,000±	
Deep Spring formation	1,500	No diagnostic fossils found; upper Precambrian or Lower Cambrian.
Reed dolomite	2,000±	
Wyman formation	1,000+	
	10,100±	

The stratigraphic nomenclature used in this report is in accord with that of Nelson (1962) who has studied generally better preserved sequences in the Blanco Mountain and Waucoba Mountain quadrangles, which adjoin the area on the east. No attempt has been made to measure and describe stratigraphic sections in detail, except for the Deep Spring formation; indeed, the complexity of the structure precludes making detailed measurements of most units.

Only formations that have yielded Olenellid trilobites are designated as unequivocal Lower Cambrian. The lowest Olenellid trilobites thus far reported were collected in the Blanco Mountain quadrangle by C. A. Nelson (written communication, 1959) from the Andrews Mountain member of the Campito formation about 900 feet below the top of the member. The Deep Spring and older formations are designated upper Precambrian or Lower Cambrian inasmuch as no basis exists for assigning them a more specific age.

Kirk (in Knopf, 1918, p. 25) placed the base of the Cambrian in this region at the base of the Campito formation, although diagnostic fossils had not then been found in it. He felt that the Deep Spring formation and rocks below it have little in common with the overlying Campito formation, and also believed a pronounced unconformity separated the two groups. Undoubtedly, he was influenced by the prevailing belief that systems are separated by worldwide unconformities, and further that the first sediments deposited would be sandstones.

The basis for Kirk's unconformity beneath the Campito formation was that the lithology of underlying but structurally conformable strata is different in the vicinity of Black Canyon from what it is along the west side of Deep Spring Valley 15 miles east in the

[1] In the explanation of plate 3, the Bishop agglutinated tuff is erroneously shown as green; it should be light tan as shown in the northwestern part of the map itself.

Blanco Mountain quadrangle. Kirk thought that at Black Canyon the Campito formation rested directly on the Reed dolomite; however, the upper part of what he identified as Reed dolomite is a facies of the Deep Spring formation.

Although no unconformities were identified, the comparatively small size of the area mapped and the highly deformed character of the strata make inadvisable a categorical statement that no unconformities exist within the section. Previous workers have reported unconformities in a similar stratigraphic section in the Blanco Mountain and Waucoba Mountain quadrangles, which adjoin the area on the east.

WYMAN FORMATION

The Wyman formation is exposed at two places about 3 miles apart. The smaller and more northerly exposure is in the north wall of a fault canyon in the NE¼ sec. 19, T. 7 S., R. 34 E.; the larger and more southerly exposure is in sec. 5, T. 8 S., R. 34 E. (pl. 3). Both exposures are in the base of the range at the head of the alluvial apron.

At the southern locality, where about 1,000 feet of strata are exposed in a prominent ridge south of the mouth of Black Canyon, the strata can be divided into four lithologically distinct units, each about 250 feet thick. The description of these units from the bottom up is as follows:

1. Platy thin-bedded dark-gray sandstone and siltstone.
2. About 20 feet of well-bedded dolomite at the base of the unit is overlain by thin-bedded medium-gray limestone and dark-gray sandstone.
3. Massive medium-gray, somewhat recrystallized dolomite and residual areas of thin-bedded medium light-gray limestone. Some of the limestone contains small oolitic nodules of concentric structure. The nodules range in diameter from an eighth of an inch or less to half an inch and may be algal.
4. Thin-bedded dark-gray limestone and sandstone and a few shaly layers. The upper contact with the Reed dolomite is sharp.

At the northern locality only about 280 feet of strata crop out. The lower 100 feet consists of thin-bedded medium-gray silty shale and limestone. Next overlying is 40 feet of thin-bedded limestone and some shale interbeds, then a massive medium-gray limestone bed about 30 feet thick. The upper 110 feet consists of thinly interbedded medium-gray limestone beds 1 to 4 inches thick separated by an inch or two of sandy shale. In two places, however, shaly beds 3 to 4 feet thick occur. As at the southern locality, the upper contact is sharp.

The 280 feet of strata at the northern locality is lithologically similar, though not identical, to unit 4 at the southern locality. If these strata are stratigraphically equivalent, an unconformity at the base of the Reed dolomite is unlikely in this area.

The name Wyman formation was applied by J. H. Maxson to 3,700 feet of strata lying unconformably beneath the Reed dolomite in the Blanco Mountain quadrangle to the east (Maxson, 1935, p. 314). He described these beds as spotted schist and phyllite and a few dolomite beds. Kirk had previously described the same strata as "oldest sandstones and dolomites." He wrote, "In general, the series as seen at several points seems to consist of thin beds of arenaceous slate, some beds of impure dolomite, and thin beds of sandstone" (Kirk, in Knopf, 1918, p. 23). These descriptions indicate that the strata beneath the Reed dolomite in the Blanco Mountain quadrangle are similar to the strata beneath the Reed dolomite in the Bishop district, although the presence of an unconformity beneath the Reed dolomite in the Blanco Mountain quadrangle casts some doubt on precise equivalence of the strata in the two areas. C. A. Nelson has restudied the Blanco Mountain quadrangle and also has inspected the strata on the south side of Black Canyon, and he regards the strata as belonging to the same general sequence (written communication, 1959).

REED DOLOMITE

The thick Reed dolomite is well exposed along the lower slopes of the White Mountains between Redding Canyon and a point 2 miles south of Black Canyon (pl. 3). The most notable characteristic of the formation is its massiveness. Bedding, shown by thick members from a distance, on close examination is obscure and difficult to identify. Conspicuous joints a few inches to a few feet apart cut the dolomite almost everywhere, and the rock breaks along joint planes into angular blocks that form rough talus slopes.

In most places the dolomite is microcrystalline, but locally it is recrystallized to a medium-grained dolomitic marble. Fresh surfaces of typical fine-grained rock are white or pale gray, but weathered surfaces generally are grayish yellow and are etched by the curving lines and pock marks that constitute the "elephant-hide" surface typical of many dolomite rocks exposed in desert areas. South of Black Canyon the dolomite is medium gray and has been recrystallized. The only organic remains found are poorly preserved forms that Kirk says strongly suggest calcareous algae of the type of *Girvanella* (in Knopf, 1918, p. 24).

The lower contact of the formation with the Wyman formation is sharp and is marked by a conspicuous lithologic break, but the upper contact with the Deep Spring formation is less well defined. The horizon chosen as the upper contact in mapping is the boundary between a thick yellowish-gray unit and a contrasting pale- or medium-gray dolomite below. In most places the outcrop of this horizon is recognizable on aerial photographs—a feature that proved to be an invaluable aid in mapping. The actual upper contact, based on color, is sharp; in a broad sense it follows a stratigraphic horizon, but it crosses beds locally and is stratigraphically a few feet higher in some places than in others.

Although the formation is well exposed, a complete and unfaulted section was not found. Above the southern exposure of the Wyman formation south of Black Canyon, however, the formation appears to be cut by only one fault that causes the apparent thickness of the Reed dolomite to be greater than the true thickness. The apparent thickness at this locality is 2,200 feet, assuming no displacement on the fault. The throw on the fault is probably no more than 200 to 300 feet. In the construction of the structure sections (pl. 5) and block diagram (pl. 6), a thickness of 2,000 feet was assumed. Support for this assumption is derived from a measurement by Kirk (in Knopf, 1918, p. 24) of 2,000± feet in the head of Wyman Canyon, a few miles to the northeast. Maxson (1935, p. 314), however, measured a thickness of 2,500± feet in the same general area.

A probable correlative of the Reed dolomite is the Noonday dolomite, which is widespread throughout the Death Valley region (Hazzard, 1937, p. 300–302). The two formations are of comparable thickness and lithology, occupy similar positions beneath fossiliferous strata of Early Cambrian age, and contain algae but no other fossils. Chemical analyses of specimens from the two formations reported by Hazzard (1937, p. 302) indicate that both are nearly pure dolomite and of almost identical composition.

DEEP SPRING FORMATION

The strata assigned to the Deep Spring formation are well exposed along the White Mountain front southeast of Bishop, in the same area as the Reed dolomite on which it rests. Lithologically these strata are quite different from the strata at the type locality of the formation 15 miles to the east, but the two lithologies interfinger in the intervening area and are clearly equivalent. According to C. A. Nelson (written communication, 1960), a few selected lithologies can be traced from the type locality to the exposures southeast of Bishop.

The formation along the White Mountain front consists of 1,500± feet of arenaceous dolomite containing limestone in the upper 150 feet and two conspicuous slaty quartzitic layers at 320 and 785 feet beneath the top. From the floor of Owens Valley between Bishop and Big Pine the formation can be seen to extend for 2 miles north from Black Canyon (fig. 2). In the late afternoon, the well-bedded, alternately brown and gray strata contrast with both the underlying massive gray-

FIGURE 2.—Unfossiliferous Upper Precambrian or Lower Cambrian strata along White Mountain front north of Black Canyon. Dark rock on summits is Andrews Mountain member of the Campito formation. Conspicuously banded rock beneath Andrews Mountain member is Deep Spring formation. Underlying massive light-colored rock is Reed dolomite. Middle foreground consists of older dissected alluvial fan and lakebed deposits of Cenozoic age, and foreground consists of undissected alluvial fan deposits.

ish-yellow Reed dolomite and with the overlying dusky-brown to grayish-black Andrews Mountain member of the Campito formation. This section, measured and described by Walcott (1895, p. 142–143), is reproduced in figure 3 together with two other sections, one on the south side of Black Canyon measured by me and one at the type locality on the north side of Deep Spring Valley measured by Kirk (in Knopf, 1918, p. 25).

The section visible from Owens Valley and measured by Walcott lies just east of the boundary of the mapped area, but the one measured by me, though not visible from the floor of Owens Valley is equally satisfactory. This section was measured along a ridge on the south side of Black Canyon in the SW¼ sec. 4, T.8 S., R. 34 E. The upper 1,200 feet of the formation was measured with telescopic alidade and stadia rod—a span that was thought, at the time it was measured, to include all the strata that might be assigned to the Deep Spring formation—and the lower 300 feet of strata was calculated approximately from the trace of the unit on the quadrangle map. Unfortunately for comparative purposes, the aggregate thickness of the lithologic units tabulated by Walcott is 135 feet less than the 1,525 feet he gives for the total thickness of unit 2, which comprises the strata here assigned to the Deep Spring formation. Comparison with the section south of Black Canyon suggests that most and possibly all 135 feet should be added at the top of the column, but perhaps a part of it may represent omissions lower in the section.

The upper contact with the Campito formation is sharp and offers no problem in mapping. The fact that beds in the basal part of the Campito formation and in the top of the Deep Spring formation are mainly conformable with each other and with the contact throughout the mapped area gives no support for the unconformity suggested by Kirk (in Knopf, 1918, p. 24–25). The lower contact with the Reed dolomite offers more difficulties, inasmuch as both formations are chiefly dolomite. The contact is placed at the base of a 500-foot-thick unit which contrasts in color with the underlying Reed dolomite.

An interesting feature of the formation is locally conspicuous crossbedding, which together with the arenaceous character of the dolomite indicates that much, and possibly all, of the formation is of clastic origin. Some of the most conspicuous crossbeds are in arenaceous layers, but in places crossbeds also are preserved in carbonate layers. The height of the crossbeds ranges from a fraction of an inch to several feet, and heights of 2 to 6 inches are common. Generally the false beds meet true beds sharply at an angle of 25° to 30°.

The common carbonate mineral is dolomite, but undolomitized residuals of blue or blue-white limestone are common in the upper 150 feet. In a single bed the limestone masses meet the dolomite with sharp transgressive contact. Figure 4 is a sketch of an outcrop that illustrates the most common relation between the limestone and dolomite within a single bed.

According to Kirk (in Knopf, 1918, p. 24–25), the Deep Spring formation at the type section 15 miles to the east along the northwest side of Deep Spring Valley comprises 1,600 feet of quartzitic and calcareous sandstone containing a few layers of arenaceous limestone. Maxson reports 3,100 feet (1935, p. 314). In the area between the type section and the west front of the White Mountains the dominantly arenaceous strata at the type locality interfingers laterally with the dominantly dolomitic strata along the west front of the White Mountains. In upper Black Canyon the Deep Spring formation consists of interbedded arenaceous dolomite and quartzitic sandstone in about equal amounts, and both the dolomite and sandstone are locally crossbedded. Kirk (in Knopf, 1918, p. 24–25) included the strata in Black Canyon, assigned here to the Deep Spring formation in the upper part of the Reed dolomite, and concluded that at Marble Canyon (Black Canyon) the Campito formation rests directly on the Reed dolomite. From this and similar observations in other areas he postulated an important unconformity at the base of the Campito formation.

The only known formation that seems a possible correlative of the Deep Spring formation is the Johnnie formation of southern Nevada and contiguous areas in the Death Valley region of California. The distance between the White Mountains and the closest outcrop of the Johnnie formation is too great to permit correlation at this time, but the two formations have many lithologic similarities and appear to occupy approximately the same stratigraphic interval.

CAMPITO FORMATION

The Campito formation consists of two members, a lower sandstone member designated the Andrews Mountain, and an upper shale and slate member designated the Montenegro. The Andrews Mountain sandstone member was included by Kirk (in Knopf, 1918, p. 27–28) in his Campito sandstone, but the Montenegro shale member may have been included, at least in part in his overlying Silver Peak group. The name Silver Peak was applied by Turner (1902, p. 264–265) to strata of Early Cambrian age near Silver Peak, about 45 miles northeast of Bishop. The term was extended by Walcott (1908, p. 185–188) to include the fossiliferous Lower Cambrian strata of western Nevada and eastern California. In this report the term Silver Peak group will not be used because its lower boundary in the

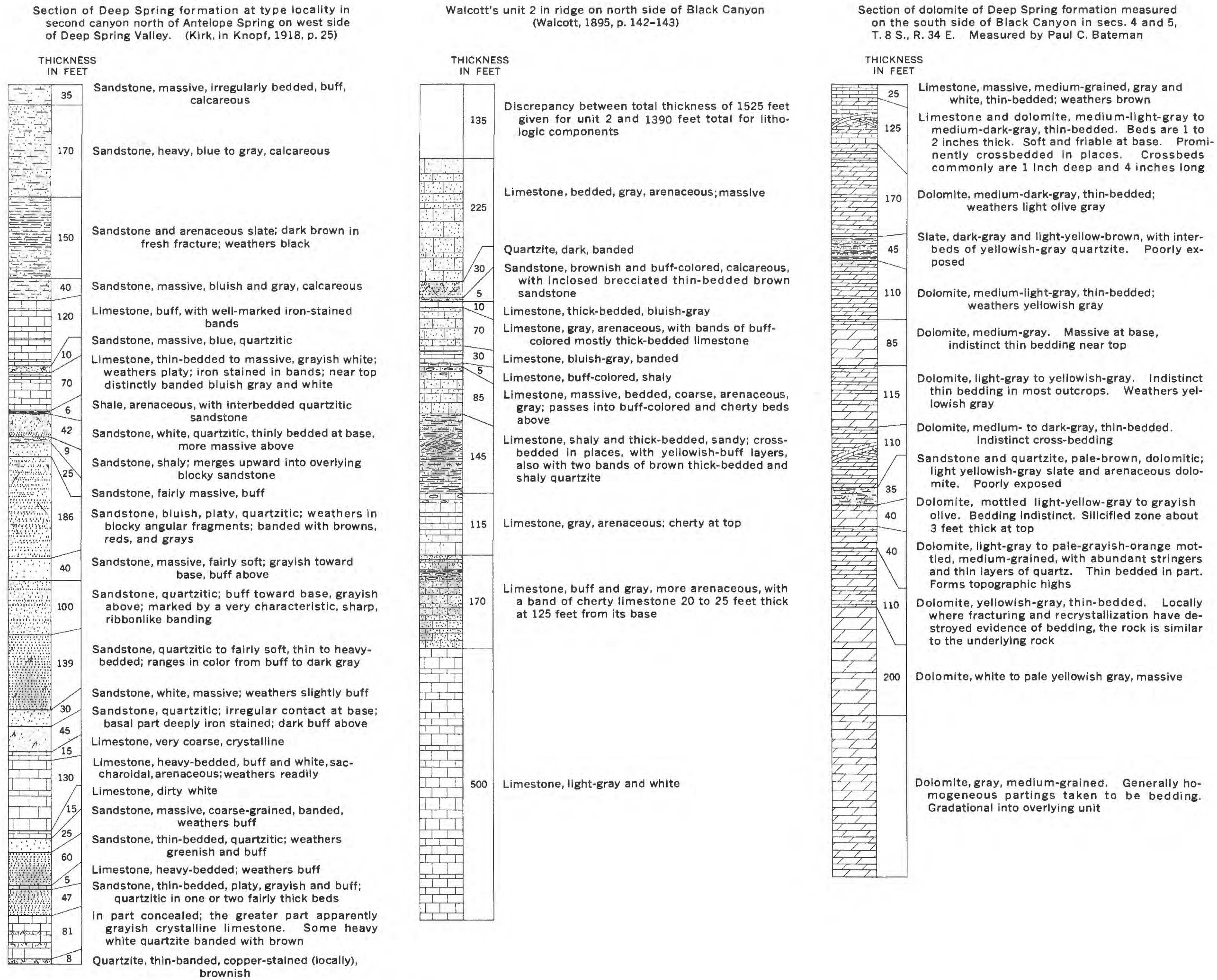

FIGURE 3.—Columnar sections of the Deep Spring formation.

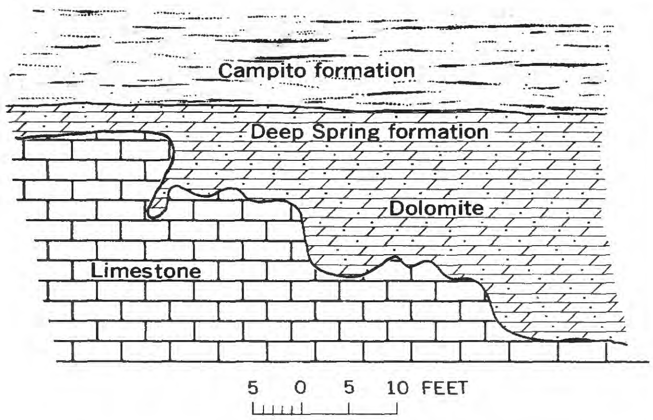

FIGURE 4.—Cliff section on north side of Black Canyon showing irregular dolomitization of limestone in the top of the Deep Spring formation.

Silver Peak region is in doubt and because the term is not necessary to the discussion.

ANDREWS MOUNTAIN MEMBER

The dusky-brown to grayish-black Andrews Mountain member of the Campito formation is named from exposures at Andrews Mountain, approximately 15 miles southeast of the mapped area in the southern part of the Waucoba Mountain quadrangle. Within the mapped area it is widely exposed along the west side of the White Mountains. Between Silver and Poleta Canyons and on the west flank of Black Mountain it crops out along the range front; between Poleta and Black Canyons and for 2 miles south of Black Canyon it crops out higher in the range and overlies the Deep Spring formation (fig. 2). North of Gunter Creek it is exposed in the cores of several anticlines. A readily accessible exposure of the member is in Silver Canyon where it has an outcrop width of about 3 miles.

The top and bottom of the Andrews Mountain member are exposed in places, but a complete and unfaulted section was not found. The thickest and most complete section is in the south half of secs. 8 and 9, T. 8 S., R. 34 E., where about 2,700 feet of strata in good order rest on the Deep Spring formation and are truncated at the top by a fault. Seventeen hundred feet of strata of the Andrews Mountain member crop out south of Redding Canyon in sec. 17, T. 7 S., R. 34 E., between exposures of the underlying Deep Spring formation and overlying Montenegro member, but the Andrews Mountain is cut by two reverse faults that have telescoped the section. Walcott (1895, p. 143) measured a thickness of 2,000 feet in Silver Canyon, but the base of the member is not exposed there, and his description suggests that the top of his section is the fault contact on the east side of the Andrews Mountain member with the Poleta formation. Kirk (in Knopf, 1918, p. 27) gives a paced measurement of 3,200 feet for the thickness of his Campito sandstone on the west side of Deep Spring Valley, which he thought to be fairly accurate. Whether the Montenegro shale member was included is not known. Maxson however (1935, p. 314), reports a thickness of only 2,000± feet, measured presumably in the same general area. In the construction of the block diagram (pl. 6) a thickness of 3,000 feet was assumed for the Andrews Mountain member.

The member consists chiefly of dark, very fine grained sandstone, but locally contains shaly or slaty layers and quartzite layers. Dense grayish-black rock in which bedding planes are difficult to identify in hand specimen is the most common type. In some layers intricate crossbeds having heights of less than an inch are marked by thin black streaks of magnetite (fig. 5). Locally, very light colored quartzitic rock contains crossbeds that are several feet long and have amplitudes of almost a foot.

FIGURE 5.—Specimen of quartzitic facies of the Andrews Mountain member of the Campito formation showing crossbedding.

Under the microscope, specimens of the common grayish-black rock are seen to consist of angular to subangular grains of quartz and a little plagioclase set in a matrix of sericite, greenish biotite (partly chloritized), and magnetite. Accessory minerals include tourmaline and zircon. Generally quartz and plagioclase grains are well graded; equidimensional grains are about 0.1 mm across and inequidimensional grains are about 0.1 mm in the shorter dimension and as much as 0.5 mm in the longer. Overgrowths on the quartz can be seen in some sections but were not recognized in most. The angularity of the grains probably reflects overgrowth rather than original shape of the grains. Some quartzitic layers are cemented with carbonate, and some light-brown rocks contain abundant limonite in the matrix.

The member is cut by numerous joints, and locally an imperfect cleavage is present. The rock breaks along these surfaces and along bedding planes into irregular-shaped slabs that ring when walked over. The cleavage ordinarily is wavy and imperfect, although it is almost as perfect in some slaty layers as in the Montenegro member. Locally, in the axial regions of folds, the rock has been brecciated; subrounded fragments of sandstone are contained in a matrix of sand and smaller fragments.

Following is a qualitative description of the unbroken 2,700-foot section in secs. 8 and 9, T. 8 S., R. 34 E.; no measurements were made in the field and the thickness and distances given are approximate:

Base to 200 feet: Thin bedded and platy; beds generally are less than 2 inches thick and many bedding surfaces are marked with branching tubelike fucoid forms. Crops out poorly.

200 to about 1,000 feet: Sandy dark-gray beds 1 to 2 feet thick are interbedded with thin silty layers similar to those in lower 200 feet. A few layers 1 to 4 feet thick of grayish-yellow quartzitic sandstone are also present. Both the dark-gray and grayish-yellow sandy layers commonly are conspicuously crossbedded. The fact that the interfaces between thicker sandy beds and thinner silty beds are exceedingly sharp causes the stratification to be conspicuous. This part of the section crops out boldly because of the sandy beds, and contrasts with the underlying and overlying strata which crop out poorly.

1,000 to about 1,500 feet: Silty and shaly strata that form poor outcrops. Interfaces between beds are not sharp and stratification is inconspicuous—in many outcrops, bedding is difficult to identify.

1,500 to about 2,500 feet: Much like the section between 200 and 1,000 feet, but contains more dark silty and shaly layers.

2,500 to about 2,700 feet (fault): Chiefly dark silty shale and shaly siltstone. Cleavage common and generally more conspicuous than bedding, which is obscure.

The progressively finer grain of the upper part of the section suggests transition to the overlying Montenegro member, but the highest exposed beds are siltier than the shale. Lenses and beds that contain carbonates and that weather out cavernously are present throughout the formation and are increasingly common toward the top; they suggest transition toward the overlying shale member, which contains similar carbonate beds and lenses.

In most exposures the contact between the Andrews Mountain member and the overlying Montenegro member is sheared or obscured by cleavage that extends from the shale into the sandstone. The least disturbed exposure of the contact is in the NW¼ sec. 21, T. 7 S., R. 34 E., where the beds dip gently and the contact is gradational through about 50 feet. It is not at all unlikely that within areas mapped as Andrews Mountain member there may be infolded slate that properly belongs to the Montenegro member.

No fossils were found other than the branching tubelike forms characterized as fucoids. Kirk (in Knopf, 1918, p. 28) reported finding only annelid trails and trilobite(?) tracks. The annelid trails possibly are the fucoids.

The Andrews Mountain member may be correlative with the Stirling quartzite of the Death Valley region.

MONTENEGRO MEMBER

The Montenegro member of the Campito formation is named from Montenegro Spring in the Blanco Mountain quadrangle where it has been studied by C. A. Nelson (written communication, 1959). The least disturbed section, however, is in the SW¼ sec. 16, T. 7 S., R. 34 E., 3 miles east of Bigelow Station, a siding on the Southern Pacific Railway. About 600 feet of flat-lying shale was measured along a line that extends northeastward and upslope from a point near the southwest corner of sec. 16. Near the section corner the Montenegro member rests on the Andrews Mountain member, and at the northeast end of the line of measurement it is overlain by the Poleta formation. The lower contact with the Andrews Mountain member is gradational through about 50 feet of strata, and the upper contact with the Poleta formation is abrupt.

In most places the Montenegro is pale olive to grayish olive, but locally, generally in the vicinity of faults.

it is grayish yellow, and in a few places it is medium to dark gray. Bedding commonly is shown by thin dark greenish-gray silty layers and by thicker brownish carbonate-cemented sandy layers. Slaty cleavage is conspicuous in most outcrops. Where both cleavage and bedding are distinguishable, the cleavage lies at all angles to the bedding; only locally are the cleavage and bedding parallel. Along the measured section the shale lacks cleavage, possibly because of less intense deformation than elsewhere, but more likely because of slight thermal metamorphism caused by intrusion of a stock of granitic rock in upper Poleta Canyon.

Under the microscope the common grayish-olive slate can be seen to consist of a fine felt of pale yellowish-brown mica and subordinate chlorite, tiny grains of quartz, and scattered sericite. Sandy beds consist chiefly of quartz grains in a carbonate matrix; a little plagioclase and mica are common in small amounts, and tourmaline is present locally.

The Montenegro member is incompetent, and as a result its distribution is highly irregular (pl. 3). Not only does it pinch and swell in an erratic fashion, but in places it is missing entirely and the overlying Poleta formation lies against the Andrews Mountain member. Not uncommonly the bedding is at an angle to the bedding in contiguous formations, even where other evidence of deformation is not obvious. Deformation within the shale is especially evident north of Poleta Canyon.

HORNFELS

Adjacent to the diorite stock north of Coldwater Canyon and extending away from it for about half a mile, slate of the Montenegro member has been converted to hornfels. Most of it is unspotted or very faintly spotted, but locally it is conspicuously spotted. Most unspotted hornfels is dark gray on fresh surfaces and grayish brown on weathered surfaces. In hand specimen it resembles finer grained parts of the Andrews Mountain sandstone member. The hornfels is a little coarser grained than the slate from which it was formed and consists chiefly of pale yellowish-brown mica and quartz that contains microscopic streaks and spots of chlorite. Some specimens also contain a little magnetite. One specimen from the south side of the stock, composed chiefly of zoisite and tremolite in the ratio of about 3 to 1, doubtless was derived from a calcareous shale or slate.

The groundmass of the spotted hornfels is greenish gray to medium dark gray and is composed of the same minerals as unspotted hornfels. The spots are grayish black ovoids that generally are a millimeter or less in longest dimension. They consist chiefly of chlorite, quartz, and pale yellowish-brown mica. Many spots exhibit a concentric arrangement—cores of fine-grained chlorite, quartz, and mica are enclosed by an inner zone of large rounded quartz grains and by an outer zone of coarser chlorite.

FOSSILS

Fossils were collected from the Montenegro member at three localities marked on the geologic map (pl. 3):

F_1 in SE¼ sec. 24, T. 5 S., R. 34 E., on north side of draw on northwest side of road to mines on bench between Piute and Coldwater Canyons. Written communication from G. A. Cooper, 1949:

"This collection consists of numerous specimens of an olenellid trilobite of Lower Cambrian age. Inasmuch as the specimens preserve the head and thorax only, I am unable to decide what the correct generic name for the trilobite is. Unfortunately the tail is needed for accurate generic identification but none is preserved. Furthermore I am unable to identify the specimens with any known species of olenellid in the National Museum collections. The Bishop quadrangle specimens are unusual in having the eyes originating almost at the anterior end of the glabella. No described species is like this and none like this occurs in the National Museum collections. I therefore conclude that the specimens submitted are a new species. Some resemblance to *Nevadella gracilis* (Walcott) can be detected but the two are not the same. I cannot therefore state from what part of the Lower Cambrian strata the specimens were taken."

F_2, a few hundred feet northeast of the S¼ corner of sec. 1, T. 6 S., R. 33 E., on the north side of the road along Gunter Creek just below the main fork. Identified by Edwin Kirk, 1947:

Archaeocyathus sp.
Kutorgina sp.

F_3, on the ridge on the south side of the south branch of Gunter Creek, half a mile north of peak 7824. Identified by Edwin Kirk, 1947:

Archaeocyathus sp.
Kutorgina sp.

POLETA FORMATION

The Poleta formation includes all the carbonate sequences in the northern two-thirds of the mapped segment of the White Mountains, except those of the Deep Spring formation. The Poleta formation is chiefly limestone, but contains much shale in its upper half. It is named after Poleta Canyon where it is well exposed. The thickness there is at least 1,000 feet and may be several hundred feet more. In most places, the formation has been strongly deformed, as the outcrop pattern indicates (pl. 3). Commonly the direction of bedding is clearly recognizable, but some limestone outcrops are massive, and others exhibit conspicuous cleavage that can be confused with bedding. In mapping, only the trace of lithologically distinctive units was taken to indicate bedding.

Both the lower contact of the Poleta formation with the underlying Montenegro shale member and the upper contact with the Harkless formation are sharp everywhere that they were observed. In many places these contacts are sheared, but they are equally abrupt where they are little disturbed.

Although the Poleta formation is predominantly limestone, it includes interbeds of calcareous shale or slate similar in appearance to that in the underlying Montenegro member and in the overlying Harkless formation. Dolomite is present within limestone beds in the form of thin anastomosing layers that lie along bedding, cleavage, and fracture planes, and clearly has replaced limestone, chiefly after the deformation of the strata.

Most of the limestone and dolomite is medium to dark gray on fresh surfaces, but some has been recrystallized to white marble. Weathered surfaces of the dolomite commonly are yellowish orange to light brown—a feature which makes it easy to distinguish dolomite from calcitic limestone. In places archeocyathids and almond-shaped ovoids (*Girvanella?*) consisting of calcite are embedded in a matrix of dolomite.

Fossil collections from the Poleta formation were made at only two localities, which are marked on the geologic map (pl. 3).

Fossils collected from the Poleta formation, Bishop quadrangle:

F_4, on north side of Silver Canyon, 1,000 feet east of contact between Andrews Mountain member of the Campito formation on west and the Poleta formation on east; in green shale within limestone unit; possibly the same as Kirk's locality 7 (in Knopf, 1918, p. 31). Identified by Edwin Kirk, 1947:

Archaeocyathus sp.

F_5, on south side of second ridge north of Silver Canyon at an altitude of about 7,500 feet; one-quarter mile westerly from peak 7824; in green shales above buff limestone. Identified by Edwin Kirk, 1947:

Archaeocyathus sp.
Kutorgina sp.
Olenellus sp. (fragments)

HARKLESS FORMATION

The Harkless formation, named after Harkless Flat in the Waucoba Mountain quadrangle, crops out discontinuously along the east side of the mapped area between Poleta Canyon and Silver Canyon. In addition, an area of slate between Gunter Canyon and Coldwater Canyon that forms a dip slope is shown on the map as Harkless formation, although the identification of the slate is uncertain. Some of the slaty shale along the west edge of the White Mountains north of Silver Canyon may also belong to the Harkless formation rather than to the Montenegro shale member of the Campito formation as shown on the map.

Within the mapped area, the formation consists largely of greenish to grayish slaty shale much like the Montenegro member of the Campito formation, but farther east in the Blanco Mountain and Waucoba Mountain quadrangle it contains beds of siltstone and lenses of moderate reddish-brown to grayish-yellow massive to thick-bedded vitreous quartzite. According to C. A. Nelson (written communication, 1960), the Harkless formation is about 2,000 feet thick, but only the lower part—perhaps a thousand feet—is present in the area mapped here.

HORNFELS

Adjacent to the granitic stock in Poleta Canyon, slate of the Harkless formation has been converted to spotted hornfels. The hornfels is much like the spotted hornfels formed from the Montenegro shale member of the Campito formation adjacent to the diorite stock north of Coldwater Canyon except that the spots are generally larger, many being 2 to 3 mm. long. In some specimens the spots coalesce and make up half or more of the rock.

A specimen collected within a few feet of the stock consists of approximately equal amounts of quartz, andalusite, and sericite, and a lesser amount of reddish-brown biotite, locally altered to chlorite. This hornfels represents a higher metamorphic grade than the spotted hornfels.

STRUCTURES IN THE WHITE MOUNTAINS

North-to northwest-trending folds and related faults in the sedimentary strata of the White Mountains and in remnants of metamorphosed sedimentary and volcanic rock in the Sierra Nevada were formed during Mesozoic time. These structures are generally well preserved in the White Mountains, and except in the vicinity of stocks have been modified only by Cenozoic faulting and warping. Tectonic movements probably began in Early Triassic time and culminated in Late Jurassic time. Movements no doubt took place earlier during the middle or late Paleozoic, but if structures were formed then in this area they were not recognized.

The effects of deformation vary with different stratigraphic units because of differences in their structural competence. The results of deformation are most conspicuous in the Montenegro member of the Campito formation and in the Harkless formation, are next most conspicuous in the Poleta formation, and are least conspicuous in the Deep Spring formation, the Reed dolomite, and the Wyman formation. Cleavage is present in almost all outcrops of the Montenegro member and Harkless formation except adjacent to the porphyritic quartz monzonite stock in Poleta Canyon and to the diorite stock north of Coldwater Canyon, and is present locally in the Andrews Mountain member and in the Poleta formation, but is lacking in the formations beneath the Andrews Mountain member. The discontinuous outcrop pattern, variable thickness, and discordant contacts of the Montenegro member indicate that during deformation it behaved almost as a paste. Probably every contact of the shale with other units is a plane of slippage, although on the maps and structure sections only obviously discordant contacts have been shown as faults. The upper part of the Andrews Mountain member and the lower part of the Poleta formation commonly are almost as strongly deformed as the shale itself. This deformation in the strata adjacent to the Montenegro probably resulted from the relative incompetence of the shale.

A fault transverse to the range front along Poleta Canyon marks the boundary between more complexly deformed younger strata on the north and less complexly deformed older strata on the south. This fault is the northernmost of three subparallel east-trending faults, all of which are downthrown to the north.

The interpretation of the structure as shown in the block diagram (pl. 6), although consistent with the data, undoubtedly is simpler than the true structure, for many minor structures could not be shown on the scale of the diagram. In some complex areas the beds are isoclinally folded on both large and small scales. Most faults that lie wholly within single stratigraphic units are not shown, and features represented as drag folds could have been shown as fault slivers; both structures were observed in the field.

In interpreting the structure, such common geologic features as order of superposition, known thickness of units, bedding attitudes, and outcrop pattern were utilized; in addition, both cleavage and lineation formed by the intersection of beds and cleavage planes were used. In many places the cleavage can be demonstrated to parallel approximately the axial planes of folds as delineated by bedding attitudes; consequently, the cleavage can be used to determine the trace and attitude of an axial plane where the position of the axial plane is not otherwise apparent. The orientation of fold axes is shown by the intersection of bedding and cleavage. Where bedding and cleavage are not parallel, top directions can be determined as follows: Beds that dip in the same direction as the cleavage but more gently and beds that dip oppositely to the cleavage are right-side-up. Beds that dip in the same direction as the cleavage but more steeply are overturned. If the axial plane is more than 90° from vertical or the axis more than 90° from horizontal, the rules for tops of beds are reversed. The dips of axial planes shown on the cross sections (pl. 5) and block diagram (pl. 6) were derived by averaging the dips observed in adjacent cleavage. Likewise, the plunges of fold axes were determined by averaging the plunges of adjacent lineations formed by the intersection of bedding and cleavage (pl. 7).

FOLDS

South of Poleta Canyon the beds dip eastward at moderate angles, apparently in the west limb of a syncline, but the trough of the fold lies further to the east, outside the mapped area. North of Poleta Canyon, where the width of the mapped span is greater, the folding is complex, and is complicated by faulting. The Andrews Mountain member and exposed underlying formations were folded and faulted in a competent manner, whereas the overlying less competent Montenegro member and Poleta and Harkless formations failed complexly, collapsing and sliding on the shale between growing folds in the Andrews Mountain.

In the area north of Poleta Canyon the pattern of folds is broadly arcuate with the convex side to the east, except for an area extending northward from Poleta Canyon where the fold axes are convex to the west. The axial planes of the folds dip to the east and to the west, and both dips are equally common; not uncommonly the dip of an axial plane changes along the strike. Most axial planes are steeper than 45°, but locally the axial planes of minor folds dip as low as 25°. The axes of most of the folds undulate and plunge locally from horizontal to vertical, although generally not more than 45°.

The major folds are few although the details of their structure are complex. They consist of three anticlines marked chiefly by outcrops of the Andrews Mountain member of the Campito formation and three synclines marked chiefly by the Poleta formation (pl. 3). The anticlines are called the North, Central, and West anticlines, and the synclines are called the East, Central, and West synclines. Several of these folds lose their identity just north of Poleta Canyon where the Poleta and Harkless formations bend westward across the trends of the fold axes. Intrusion of the granitic stock at the head of Poleta Canyon or movement along the fault that follows Poleta Canyon may have contributed to the complexities of this area.

FAULTS

Most of the prebatholithic faults strike northwestward and dip either to the east or west parallel with the axial planes of the folds and with the cleavage. Both normal and reverse faults are present, but nevertheless most faults probably resulted from the same compressional movements that produced the folds. Faulting probably began after the folding was well advanced, and it increased in magnitude as the folds became tighter and as internal resistance to further folding built up. Although many and perhaps most of the faults were deformed by contemporaneous folding, no evidence was found of faults that are clearly older than the folds. Some normal faults seem to be best explained as the result of local tension in a dominantly compressional field, but these faults could have been formed during a later period of relaxation.

In mapping it was difficult to distinguish some prebatholithic faults from some Basin and Range faults inasmuch as the two types are roughly parallel in strike. The following three criteria proved helpful in making distinctions as to the affiliations of questionable faults:

Basin and Range faults	*Prebatholithic faults*
1. Commonly normal faults with little or no evidence of strong contemporaneous compression.	1. Mostly compressional faults.
2. No cleavage of contemporaneous origin parallel with the fault plane.	2. Parallel cleavage in the walls of some faults, which diminishes in intensity with distance from the faults.
3. Scarps common as a direct result of fault movement. Many scarps wholly or partly in Cenozoic deposits.	3. Linear depressions along traces of faults, the result of erosion.

FAULTS SOUTH OF POLETA CANYON

South of Poleta Canyon the only faults that can be ascribed with certainty to prebatholithic diastrophism are two reverse faults that strike about N. 30° W. across the first prominent ridge south of Redding Canyon (pl. 6, sec. 7). The faults are cut off to the south by an east-trending normal fault and are overlapped to the north by alluvial deposits. Both faults dip to the west, and the more westerly one steepens in depth. Because the western fault dips generally more steeply than the eastern fault, the two faults are presumed to join in depth. The exposed segments of both faults are chiefly within the Andrews Mountain member, but along the eastern fault the Andrews Mountain locally has been carried eastward over the Montenegro member; along the western fault, beds in the Deep Spring formation on the southwest side strike into inliers of the Andrews Mountain in the alluvium on the northeast side. The apparent stratigraphic thickness of the Andrews Mountain member has been reduced by about 1,000 feet on the two faults; if movement was approximately equal, the stratigraphic displacement on each fault is about 500 feet.

FAULTS ON THE EAST LIMB AND CREST OF THE CENTRAL ANTICLINE

An interconnecting system of faults of highly variable strike and dip follows roughly the limb between the Central anticline and the East syncline. Most of the faults are contacts between sandstone of the Andrews Mountain member and shale of the Montenegro member of the Campito formation and the Poleta formation, or between the shale and the limestone. The traces of the faults commonly are curved, partly because of the effect of topography and partly because the fault planes curve both in dip and strike. In a general way the fault surfaces follow the folded structure of the beds. The movement on most faults appears to have been small, although in places the total thickness of the Montenegro member plus an unknown thickness of the underlying or overlying formations have been cut out; a stratigraphic displacement in excess of 600 feet is thus indicated.

The southernmost faults of this system are between Poleta and Silver Canyons. Most of the faults here dip steeply to the east, but a few are almost flat, and one dips to the west. An especially interesting one follows the contact between the Montenegro member and the Poleta formation near the head of the first drainage north of Poleta Canyon. On the ridge between the two branches of this drainage the beds dip to the east, and no evidence of significant faulting of the contact was found. Northward the contact steepens and then overturns to the east concomitantly with increasing magnitude of shearing along the contact. Just south of a northwest-trending cross fault that offsets the contact, several hundred feet of strata from the bottom of the Poleta formation are cut out against the overriding Montenegro member. The steepening and overturning of the beds are pictured as early movements that culminated in the shearing.

From the south wall of Silver Canyon north to Coldwater Canyon is a belt of faults that separate the Montenegro member, the Poleta formation, and the Andrews Mountain member. In the lower walls of Silver Canyon, the Poleta formation on the east is in contact with Andrews Mountain along a fault that strikes north and dips west (pl. 6, section 3). Northward and topographically higher this fault steepens, then overturns to the west, and on the ridge on the south side of the south fork of Gunter Creek it dips gently west (pl. 6, section 2). Although the fault may be twisted along the strike, the spacial relations suggest that it is convex upward and toward the west and bends in dip through more than 100° (pl. 6, sections 2–4). Locally in this span, the Montenegro member is present along the contact, but in most places the Poleta formation is in direct contact with the Andrews Mountain member.

Farther north, in the canyons of the two branches of Gunter Creek, the Montenegro member lies between the

Andrews Mountain member and the Poleta formation, but in most places the bedding planes within the shale dip steeply and are discordant with the gently dipping upper and lower contacts—a feature that indicates structural ungluing of the shale from the overlying and underlying formations.

FAULTS ASSOCIATED WITH THE WEST SYNCLINE

North of Poleta Canyon in sec. 6, and the NE¼ sec. 7, T. 7 S., R. 34 E., the Andrews Mountain member along the front of the range is in fault contact on the east side with the Montenegro member and the Poleta formation. The fault trends N. 30° W.; in the north part the dip is vertical and in the south part is steeply east. At the north end, in the first canyon north of Poleta Canyon, the faulted contact grades into an apparently unfaulted contact between the Andrews Mountain and the Montenegro. The faulted segment of contact is clean and sharp, cuts across folds in the Andrews Mountain at a high angle, and truncates bedding in the Montenegro member and Poleta formation at a slightly smaller angle.

Within the Andrews Mountain member on the west side of the fault is a thin sliver of shale and limestone bounded by faults. This sliver trends a few degrees east of north, approximately parallel with fold axes in the Andrews Mountain, and may mark the approximate position of the southward continuation of the West syncline. The faults that bound the sliver of shale and limestone are poorly defined, but appear to terminate against or bend into the fault that cuts off the Andrews Mountain member on the east.

FAULTS IN THE CENTRAL SYNCLINE AND WEST ANTICLINE BETWEEN SILVER CANYON AND GUNTER CREEK

The Andrews Mountain member exposed between Silver Canyon and Gunter Creek is cut by several east-dipping reverse faults. These faults are chiefly along northern projections of the Central syncline and the West anticline, which they have almost destroyed. A thin belt of slate probably marks the trough of the Central syncline. Most of the faults dip steeply, but some dip as low as 35°. No displacements were determined, but movement on most of the faults appears to have been small because the faults are either entirely within the Andrews Mountain member or between it and the Montenegro member and the Poleta formation. They are approximately parallel with axial planes of folds. Parallel faults are also exposed in the ridge of Andrews Mountain member on the north side of Silver Canyon near the range front, and the movements on these faults also appear to have been small inasmuch as only the Andrews Mountain is exposed in the walls of the faults.

Along the range front at Coldwater Canyon and also about 1 mile farther to the south are two slivers of Andrews Mountain member that are surrounded by shale and limestone. Both slivers may mark anticlines, but the northern sliver is bounded on both sides by steep faults and at the south end is in contact with the Poleta formation.

CLEAVAGE

Slaty cleavage is prevalent in the shales of the Montenegro member and the Harkless formation, and is also present in many outcrops of the Andrews Mountain member and the Poleta formation (fig. 6). Cleavage

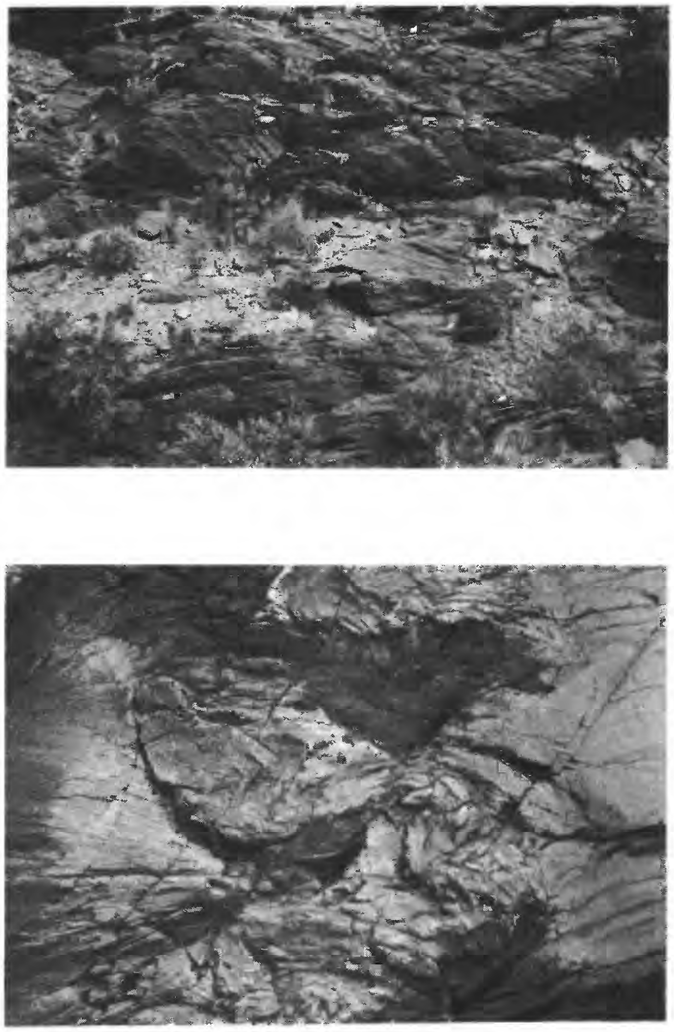

FIGURE 6.—Slaty cleavage in Montenegro member of Campito formation showing divergent attitudes of bedding and cleavage. Above, steeply dipping cleavage in steeply dipping homoclinal section. Cleavage strikes from left to right and bedding from upper left to lower right. Steep plunge of lineation formed by intersection of bedding and cleavage parallels locally steep axis of West syncline. Below, minor folds in slate shown by calcareous sandy bed. Cleavage is generally parallel to axial planes of folds. Late stages of folding caused competent sandy bed to bend cleavage.

is present throughout the shales, except adjacent to the stock of porphyritic quartz monzonite in Poleta Canyon and the diorite stock north of Coldwater Canyon, where thermal metamorphism destroyed the cleavage. The cleavage is pervasive and the shale splits readily along flat cleavage surfaces that generally are not parallel to the bedding. In the Andrews Mountain member the cleavage is approximately parallel with cleavage in adjacent shale, but the cleavage is not pervasive, and cleavage surfaces generally are wavy and branching rather than flat. The character of the cleavage surfaces appears to be chiefly a function of the grain size of the rock; the smoothest cleavage surfaces are in rocks composed of clay-size particles, wavy surfaces are indicative of silt-size particles, and rocks composed of sand-size particles generally exhibit no cleavage. In the Poleta formation, impure argillaceous limestone cleaves readily along flat surfaces, but in cleaner carbonate rocks the cleavage is shown chiefly on weathered surfaces by thin parallel ridges and troughlike depressions. This sculpturing along the trace of cleavage on weathered surfaces of limestone can be confused with the trace of bedding. Cleaved limestone tends to break along rough surfaces that are only subparallel to the cleavages.

Most of the cleavage is approximately parallel to the axial planes of folds and, therefore, is related in origin to the folding. However, a second cleavage parallel with faults was also observed in a few places in the Montenegro member. This second cleavage is pervasive adjacent to faults and increasingly more widely spaced farther away. Cleavage in this orientation was observed to transgress and in places obliterate cleavage parallel to axial planes and is clearly younger than the latter. Because this second cleavage is identical in appearance to the first and was not recognized until the mapping was nearly completed, no attempt was made to discriminate between the two cleavages. With a few exceptions, the cleavage attitudes plotted on the structure map (pl. 7) fall into a systematic pattern that appears to parallel the axial planes of folds. The fact that many of the faults associated with the folding are approximately parallel with the axial planes permits the existence of two kinds of cleavage in such a pattern. Nevertheless, cleavage can be demonstrated to parallel the axial planes of folds in so many places that it seems likely that most of the recorded cleavage is related to folding rather than to faulting.

METAMORPHIC ROCKS OF THE SIERRA NEVADA

Two large pendants (the Pine Creek and Bishop Creek pendants), several septa, and groups of small inclusions are distributed through the predominantly granitic terrane of the Sierra Nevada (fig. 7). The rocks in these masses include metasedimentary, metavolcanic, and mafic intrusive and hybrid rocks that are older than the granitic rocks that make up the bulk of the batholith. Broadly, the metamorphic rocks can be separated into two series—an older one, probably of Paleozoic age, composed entirely of metasedimentary rocks, and a younger one, probably of Mesozoic age in which metavolcanic rocks predominate over metasedimentary rocks.

The rocks of sedimentary derivation include micaceous quartzite, pelitic hornfels, metachert, marble, calc-hornfels, conglomerate, gneiss, and mica schist. The metavolcanic rocks are chiefly mafic flows, tuffs, and shallow intrusives of andesitic to dacitic composition, and the Pine Creek pendant also contains felsic metavolcanic dike rocks of quartz dacitic and tuff of rhyolitic composition. Intercalated with the metavolcanic rocks are minor marble, calc-hornfels, pelitic hornfels, micaceous quartzite, and conglomerate.

In the two pendants and in some of the larger septa, it has been possible to subdivide the metamorphic rocks into mappable units. However, the relatively small extent of the units, their strong deformation and high metamorphic grade, which makes detailed stratigraphic descriptions impractical, and the absence of fossils are strong deterrents to adding new formational names for these units to already overburdened lexicons. On the geologic map the units are designated in terms of the dominant lithologic assemblages. The rocks in smaller inclusions are classified simply, on the basis of the most abundant rock, into (1) marble, (2) calc-hornfels, (3) pelitic hornfels, and (4) mafic metavolcanic rocks.

The units that were distinguished in mapping far exceed in number the different kinds of rock that were recognized. All but a few of the mapped units consist of more than one kind of rock, and most kinds of rock are in more than one mapped unit. To avoid repetitious rock descriptions, the principal kinds of metamorphic rock that were recognized are described in a section that precedes the systematic descriptions of the metamorphic remnants.

Although no diagnostic fossils were found in the metamorphic rocks, a comparison with fossiliferous strata in the less highly metamorphosed Laural-Convict pendant in the Mount Morrison quadrangle adjoining the Bishop area on the northwest (Rinehart, Ross, and Huber, 1959) and in the Inyo Mountains (Knopf, 1918) indicates that the metasedimentary series is Paleozoic and that the dominantly metavolcanic series is Mesozoic.

The only metamorphic remnants large enough to incorporate more than a very simple structural pattern on

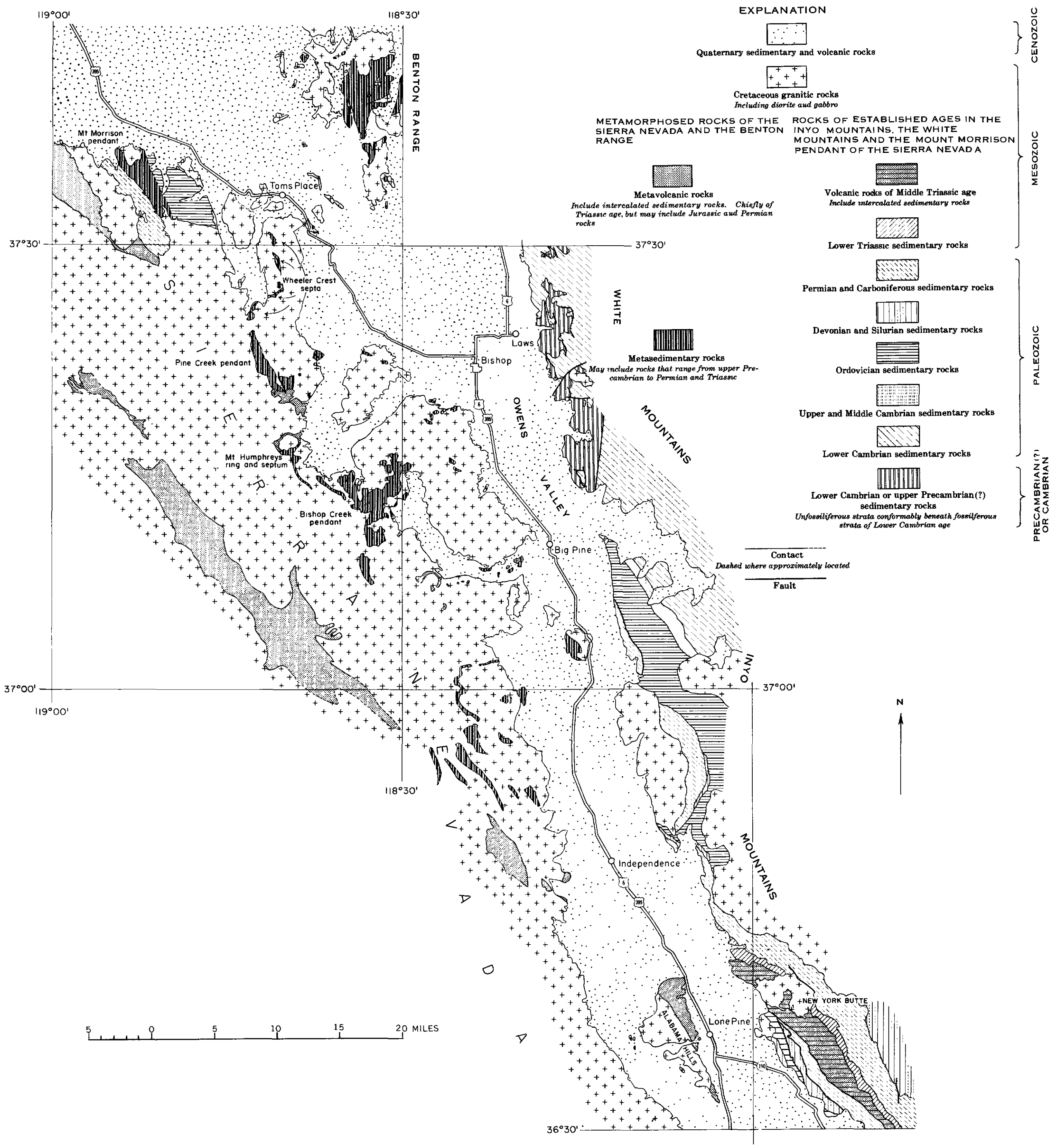

FIGURE 7.—Geologic map of the Owens Valley region showing the distribution of the pre-Cenozoic rocks.

the map scale are the Pine Creek and Bishop Creek pendants. The rocks in these pendants have been subjected not only to the regional deformation that resulted in the north- to northwest-trending folds and compressional faults in the White Mountains, but also to later deformation caused by the forcible intrusion of the batholithic rocks. Although folds produced during earlier regional deformation cannot always be distinguished from folds caused by later intrusion of a pluton, several criteria are generally applicable. The axes of the earlier regional folds are generally nearly horizontal and parallel with the north-to-northwest regional trend. These folds may extend for many miles and show no geometric relation to bordering intrusives. Folds caused by intrusion, on the other hand, generally are of more limited extent, their axes trend in directions that bear geometric relation to the configuration of the bordering intrusives, and the axes generally dip steeply rather than gently. The relation of intrusion to wall rock deformation is considered in a later section of this report dealing with evidence of mechanical emplacement of the intrusive rocks.

The rocks in the Sierra Nevada pendants, though folded and faulted, are not mashed to the same degree as some of the rocks in the White Mountains, presumably because of the greater structural competence of the strata. Cleavage was not recognized in the Bishop Creek pendant but is locally conspicuous in the Pine Creek pendant. If cleavage formerly was more conspicuous and widespread, it has been destroyed by subsequent thermal metamorphism.

METASEDIMENTARY ROCKS

The metasedimentary rocks were derived from miogeosynclinal sedimentary rocks—shale or slate, limestone, sandstone, siltstone, chert, and sediments of intermediate composition. Shale or slate has been metamorphosed to pelitic hornfels, marl and siliceous dolomite to calc-hornfels, limestone to marble, sandstone and siltstone to quartzite, and chert to metachert, and so forth.

MICACEOUS QUARTZITE DERIVED FROM ARGILLACEOUS SANDSTONE AND SILTSTONE

The rock here designated micaceous quartzite is a dense fine-grained rock, commonly dark gray to grayish red, which was derived from dirty sandstone or siltstone. The largest mass is in the Pine Creek pendant, but it is a common rock type and is present in many of the metamorphic remnants. In the Bishop Creek pendant, together with finer grained pelitic hornfels, it constitutes a unit that is almost as extensive as the one in the Pine Creek pendant.

Commonly the rock megascopically appears structureless and breaks into angular fragments on weathering, but locally it has a more or less conspicuous cleavage and breaks along cleavage planes into tabular slabs. Bedding generally is obscure, and to locate it requires careful searching for compositionally distinctive layers.

Quartz is the most abundant mineral and generally makes up 50 to 60 percent of the rock. Most specimens also contain 20 to 30 percent of biotite and 10 to 15 percent of sericite. Many rocks also contain 5 to 10 percent of K feldspar and a few contain as much as 25 percent. Rocks that contain abundant K feldspar generally contain little or no sericite and are light yellowish gray. Minor accessory minerals, which commonly are sporadically distributed through the rocks, include plagioclase, apatite, tourmaline, sphene, garnet, magnetite, and pyrite. In addition, small amounts of chlorite, epidote, hematite, and sericite are present. Fragments of fine-grained siliceous rock (chert?) also are common, though not abundant. Most of the quartz is in somewhat rounded grains, but the margins are rough in detail owing to overgrowths of new quartz. In some specimens, the quartz grains are all about the same size, but more commonly the quartz grains are not so well size sorted and consist of a wide variety of grain sizes. Pebbles, present locally in restricted stratigraphic zones, commonly consist of aggregates of quartz rather than of quartz in a single orientation. Generally, the quartz grains are seen under crossed-nicols to be unstrained and to extinguish uniformly.

K feldspar is interstitial to the quartz grains. Most of the biotite is in tiny flakes that are evenly distributed through the rock, but some is in patches or streaks in which the individual biotite flakes are much larger than the dispersed flakes. All the biotite is pleochroic from reddish brown to colorless and causes the grayish-red color of fresh surfaces of many specimens. The biotite flakes show some degree of preferred orientation in almost all the rocks, but the orientation is best in finer grained, more schistose ones. Sericite is present in tiny flakes in some rocks and in large poikiloblastic plates in others. Where the sericite is in tiny plates it is commonly distributed with the biotite and oriented parallel with it, but with a lower degree of preferred orientation than the biotite. Larger poikiloblastic plates, on the other hand, are oriented at random—a feature that suggests that they were formed late.

METACONGLOMERATE

Metaconglomerate was found only in the Deep Canyon area of the Tungsten Hills and in the northern segment of the septum 3 miles west of Keough Hot Springs. In the Tungsten Hills subrounded pebbles that generally are less than an inch across are set in a calcium-rich matrix that consists chiefly of diopside.

Many of the pebbles consist of granoblastic quartz and interstitial diopside, and a few consist of zoisite or other calc-silicate minerals. In the septum 3 miles west of Keough Hot Springs, pebbles of felsic volcanic rock are contained in a matrix of quartz, K feldspar, and calc-silicate minerals, including epidote, tremolite, and diopside. The pebbles consist chiefly of a very fine granoblastic intergrowth of quartz and K feldspar in which broken crystals of quartz, K feldspar, and plagioclase are set.

Pebbly beds are present locally within the micaceous quartzite unit of the Pine Creek pendant, but the pebbles generally make up less than half the bulk of the rock. The pebbles consist of granoblastic quartz probably derived from metachert and schist. The matrix is micaceous quartzite.

PELITIC HORNFELS DERIVED FROM SILTY SHALE, ARGILLACEOUS SILTSTONE, AND CLAY SHALE

The rock here classed as pelitic hornfels is one of the more ubiquitous of the metamorphic rocks. It commonly occurs with micaceous quartzite, but is also interlayered with other metasedimentary rocks. Megascopically it is a dense fine-grained structureless rock that on fresh surfaces ranges from light yellowish or olive-gray to dark gray or even black; the reddish color and cleavage found in some micaceous quartzite is not present in pelitic hornfels, but otherwise the two rocks appear megascopically similar. In terms of origin, pelitic hornfels is derived from finer grained, more argillaceous rock than micaceous quartzite. Parent sedimentary rocks probably include argillaceous siltstone, silty shale, and clay shale or slate, and consequently they are finer grained and contain less quartz and more feldspar and mica than the quartzite. The varieties having finest grain are carbonaceous and contain andalusite.

The mineral content is quite variable and reflects the wide compositional range of the rocks included. Most specimens that were examined under a microscope contain quartz, mica, and feldspar, but the mica may be either biotite or sericite or both, and the feldspar may be K feldspar or plagioclase; a few specimens that contain abundant plagioclase lack quartz. Plagioclase and actinolite are present in rocks having a high content of CaO and occur in assemblages that are transitional to calc-hornfels. The most common accessory minerals are apatite, magnetite, and sphene.

Andalusite-bearing pelitic hornfels is not common, but with metachert it forms an extensive mappable unit in the Bishop Creek pendant. Typically it is dark gray to black, except for conspicuous white prisms of andalusite as much as half an inch long. The groundmass is fine grained, dense, and structureless, and the andalusite prisms are randomly oriented. The content of andalusite generally falls between 25 and 35 percent, but in some specimens it constitutes as much as 50 percent of the rock. In all specimens that were examined under the microscope, the groundmass contains quartz, sericite, and carbonaceous material. In some the groundmass also contains biotite or albite. Apatite commonly is a minor accessory mineral. Most andalusite porphyroblasts are bordered by a dark rim of expelled carbonaceous material, and many exhibit the internal carbonaceous cross that is characteristic of the variety called chiastolite. Much of the andalusite has been replaced along the margins by sericite, and in some specimens it has been completely replaced by a sericite aggregate.

METACHERT

Metachert was identified only in the Bishop Creek pendant where it is interstratified with pelitic hornfels in an extensive mappable unit, but other rocks that resemble quartzite and consist chiefly of recrystallized quartz may also be derivatives of chert.

The metachert in the Bishop Creek pendant occurs in two principal varieties, the more common of which is a light-gray vitreous rock megascopically indistinguishable from quartzite. This rock grades locally into the less common second variety, a dark-gray fine-grained carbonaceous metachert.

The dark carbonaceous chert consists of a very fine grained granoblastic mosaic of tiny quartz grains through which is disseminated finely divided carbonaceous material. Some specimens contain a very small amount of interstitial K feldspar or sericite (estimated to be less than 2 percent), and others contain calcite or calc-silicate minerals such as garnet, epidote, or diopside. The calcite and calc-silicate minerals, though disseminated in the chert, are most abundant adjacent to thin veinlets that contain the same minerals. This association suggests that calcium was introduced into the rock prior to or during thermal metamorphism. In some specimens, nevertheless, microcline and diopside occur together in thin beds that probably mark original marly layers.

Although no fossils were identified in the metachert, small round masses of granoblastic quartz much coarser than the groundmass, and rimmed with carbonaceous material, probably are vestiges of radiolaria. Similar-appearing structures in the cherts of the Franciscan formation in the California Coast Ranges, as well as in cherts elsewhere, are fossil radiolaria.

Derivation of the more common coarser grained light-gray metachert from the dark-gray carbonaceous chert by recrystallization and expulsion of the carbonaceous material is readily demonstrable. Most specimens of dark carbonaceous metachert are cut by light-colored veinlike streaks of quartz having a relatively coarse-grained granoblastic texture. These streaks are most abundant in transitional zones between dark-gray metachert and light-gray metachert. In the direction of the light-gray metachert the veinlike streaks are increasingly common, and the light-gray metachert is simply the end stage of this veination. The relations in the transition zones leave no doubt that the veins as well as the more extensive masses of light-gray metachert resulted from recrystallization of the dark-gray metachert (fig. 8). In the recrystallization the carbonaceous material was expelled, and various stages of this process can be traced in thin sections. The carbonaceous material was aggregated into irregular veinlets before complete elimination. The light-gray rock is commonly mottled in various shades of gray, and the different shades reflect the degree of recrystallization and elimination of carbonaceous material. Other constituents, however, were not eliminated during recrystallization, and even the most coarsely recrystallized specimens contain a little K feldspar and sericite.

FIGURE 8.—Specimen of metachert showing light and dark varieties. Light-colored rock was derived from dark rock by recrystallization and elimination of carbonaceous dust.

MARBLE DERIVED FROM LIMESTONE

Marble includes all the rocks that were derived from limestone and that consist predominantly of coarsely crystalline calcite. Marble ranges from very light gray (almost white) to dark gray, depending upon the content of carbonaceous material, and in granularity from coarse to fine. In general the lighter colored varieties have been more intensively recrystallized than the darker varieties and are coarser grained. Carbonaceous material was driven out of the rock during recrystallization and almost completely eliminated in the coarsest grained rocks.

The carbonate mineral in the marble is everywhere calcite; dolomite or dolomitic marble was not identified in the Sierra Nevada, presumably because it is not stable under the condition of thermal metamorphism that existed. The presence of diopside or tremolite in many silicated marbles indicates the rock formerly was impure dolomitic limestone or dolomite, and scarce brucite indicates that some varieties were clean dolomitic limestone or dolomite. Most marble in the Sierra Nevada contains argillaceous and siliceous impurities, which have combined with calcite and dolomite to form calc-silicate minerals. These calc-silicate minerals are disseminated through the marble, but are most abundant in thin well-defined layers that mark impure beds. Metamorphism was sufficiently intense to eliminate any carbonaceous material. Silicate minerals in addition to those already mentioned include grossularite, idocrase, plagioclase, and scapolite. Locally, quartz-rich layers represent layers of quartz sand or of chert.

Most marble beds are at least tens of feet thick, and the more recrystallized ones are massive and structureless. The bedding is best shown by compositional differences among beds. The thickest and most extensive marble lies along the west side of the Pine Creek pendant, and less extensive marble is present in most other metamorphic remnants.

CALC-HORNFELS DERIVED FROM ARGILLACEOUS AND SILICEOUS LIMESTONE AND DOLOMITE

Calc-hornfels is widespread, and in the Bishop Creek pendant constitutes an extensive stratigraphic unit. The rocks here designated as calc-hornfels are dense fine-grained, predominantly light-colored rocks that are composed of several different assemblages. Common minerals include plagioclase, diopside, tremolite, grossularite, wollastonite, K feldspar, quartz, and calcite. Fresh surfaces commonly range from light greenish gray to pale greenish yellow, or to light gray, but lay-

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ers rich in diopsidic pyroxene are usually greenish gray to dark greenish gray, and layers rich in garnet generally are light brown. In many places intercalated with calc-hornfels are a few thin medium- to dark-gray quartzose or feldspathic layers that contain a cloud of carbonaceous material. Weathered surfaces are lightly stained with limonite, but the stain does not affect the overall light tone of the rocks in outcrop.

Bedding generally is shown by compositional layers that commonly are less than an inch thick. Secondary cleavage is lacking and the rock breaks on weathering into angular fragments, although some calc-hornfels splits along bedding planes into flat fragments. Calc-hornfels is one of the more resistant rocks to erosion, and commonly it crops out in ridges or forms cliffs.

Common mineralogic assemblages include plagioclase-diopside, quartz-diopside, quartz-diopside-tremolite, quartz-tremolite, quartz-diopside-K feldspar, diopside-tremolite-K feldspar, quartz-diopside-K feldspar. Assemblages that contain abundant tremolite and K feldspar are characterized as argillaceous calc-hornfels, and those that contain abundant quartz are called siliceous calc-hornfels. Plagioclase-diopside assemblages were derived from argillaceous dolomite or magnesian limestone, and quartz-diopside assemblages from siliceous dolomite or from dolomitic sandstone. Tremolite occurs in rocks that is relatively poor in calcium, where it takes the place of diopside. Quartz-tremolite rock was probably derived from dolomitic sandstone. K. feldspar is present only in rocks that originally contained significant amounts of argillaceous material. Calcite is present in calcium-rich, diopside-bearing assemblages, but does not occur with tremolite. Wollastonite is found most commonly near intrusive contacts, where it normally occurs with calcite and diopside. Locally, favorable beds and thin selvages between marble and granitic intrusive rock consist almost entirely of wollastonite.

Rock that contains large amounts of quartz might be more accurately characterized as sandy calc-hornfels or as silicated quartzite depending on whether quartz or silicate minerals predominate. Similarly, calc-hornfels grades to marble through calcitic calc-hornfels and silicated marble. As the content of argillaceous material increases calc-hornfels grades to argillaceous hornfels, but no names have been designated for intermediate assemblages.

TACTITE DERIVED FROM MARBLE AND CALC-HORNFELS

Tactite is a rock of more or less complex mineralogy formed from calcareous or dolomitic rocks by contact metasomatism. It is the common host rock for scheelite, the principal tungsten-bearing mineral in the Bishop district, and is more fully described in the part of this report that deals with tungsten mineralization. The term "tactite" is probably synonymous with the term "skarn." No single mineral is essential by definition to tactite, but most of the tactite in the Bishop district is composed chiefly of pale to moderate reddish-brown garnet of the andradite-grossularite series and grayish-green pyroxene of the diopside-hedenbergite series. Epidote, quartz, scheelite, and various sulfides are also common in tactite.

GNEISS

Gneiss was found at only two places within the mapped area. The largest mass is a lenticular septum about 3 miles long that is exposed in the canyon of the South Fork of Bishop Creek between Long Lake and Bishop Pass; and gneiss is present locally in several small metamorphic remnants that crop out in the lower part of the ridge between Red Mountain and Taboose Creek, in the vicinity of Stecker Flat. The gneiss in both places consists of thin lenticular alternating mafic and felsic layers that generally are an eighth of an inch or less thick. In the South Fork of Bishop Creek the mafic layers are generally thinner than the felsic layers, but in the vicinity of Stecker Flat the layers are about equal in thickness. In both places the felsic layers consist chiefly of a granoblastic intergrowth of quartz and feldspar. In the mass in the South Fork of Bishop Creek the mafic layers contain biotite plates and hornblende prisms oriented in the plane of foliation. The mafic layers in the gneiss in the vicinity of Stecker Flat contain biotite but no hornblende. Magnetite is a minor constituent in the mafic layers.

The gneiss in the South Fork of Bishop Creek provides little basis for speculation as to its origin; its appearance suggests highly sheared plutonic rock. The gneiss in the vicinity of Stecker Flat, however, grades to schist and to hornfels of the same mineral composition, and all three rocks were produced from the same parent rock, presumably slightly feldspathic siltstone. All three structural varients are composed of quartz, plagioclase, and biotite, and accessory magnetite and apatite. Some specimens also contain as much as 15 percent of sericite associated with biotite and in aggregates interstitial to quartz and plagioclase. All the minerals in the hornfels are nearly equant and unoriented, and the texture is granoblastic, whereas biotite in the schist and gneiss is in strongly oriented plates. In hornfels the quartz and plagioclase grains are of about the same size throughout the rock, whereas in schist and gneiss coarser grains of quartz and plagioclase are bordered by finer quartz and plagioclase, and conspicuously fine-grained streaks are common. In gneiss the dark layers are both richer in biotite and finer grained than in the felsic layers. The finer grain size in streaks and marginal to larger grains was probably produced by pervasive shearing at the time of metamorphism.

In figure 9, specimens of hornfels having recrystallized orbicular spots and gneiss containing relict hornfelsic areas are shown. The hornfels in the spotted rock is mineralogically and structurally identical with that in the gneiss. The orbicular spots were formed by recrystallization without obvious structural control; nevertheless, the light and dark minerals were segregated during recrystallization. The gneissic structure probably formed in the same way, except that biotite recrystallized preferentially along shears.

METAVOLCANIC ROCKS

Metavolcanic rocks were mapped simply as either felsic metavolcanic rocks or mafic metavolcanic rocks. Mafic metavolcanic rocks are more abundant than the felsic metavolcanic rocks, which are limited to a small area in the Pine Creek pendant on the southwest side of Mount Tom. The felsic metavolcanic rocks include rhyolite and quartz latite, and the mafic ones include andesite and dacite. The rocks in the two groups are distinguishable megascopically by color index and specific gravity. The specific gravities of hand specimens of felsic rock generally are between 2.60 and 2.65, and the specific gravities of mafic rocks are between 2.75 and 2.85.

FELSIC METAVOLCANIC ROCKS

Metarhyolite tuff and quartz latite, mostly intrusive but some possibly in flows, are present in the south part of the Pine Creek pendant between micaceous quartzite on the north and mafic metavolcanic rocks on the south. Metarhyolite tuff is massive and structureless except locally where it has a secondary foliation.

Unweathered surfaces are light gray, but weathered ones commonly are stained pinkish gray. Small phenocrysts of quartz 2 to 3 mm across and lens-shaped foreign rock fragments, the largest less than a centimeter in greatest dimension, can be readily identified in hand specimen.

Under the microscope the tuff can be seen to consist of about half mineral and foreign rock fragments and half a fine-grained granoblastic groundmass composed chiefly of quartz and feldspar. Angular fragments of quartz, microcline, and a little sodic plagioclase (oligoclase) are abundant. Presumably the microcline was derived from original sanidine during thermal metamorphism. The quartz, both in angular fragments and in crystals, has been converted to a granoblastic mo-

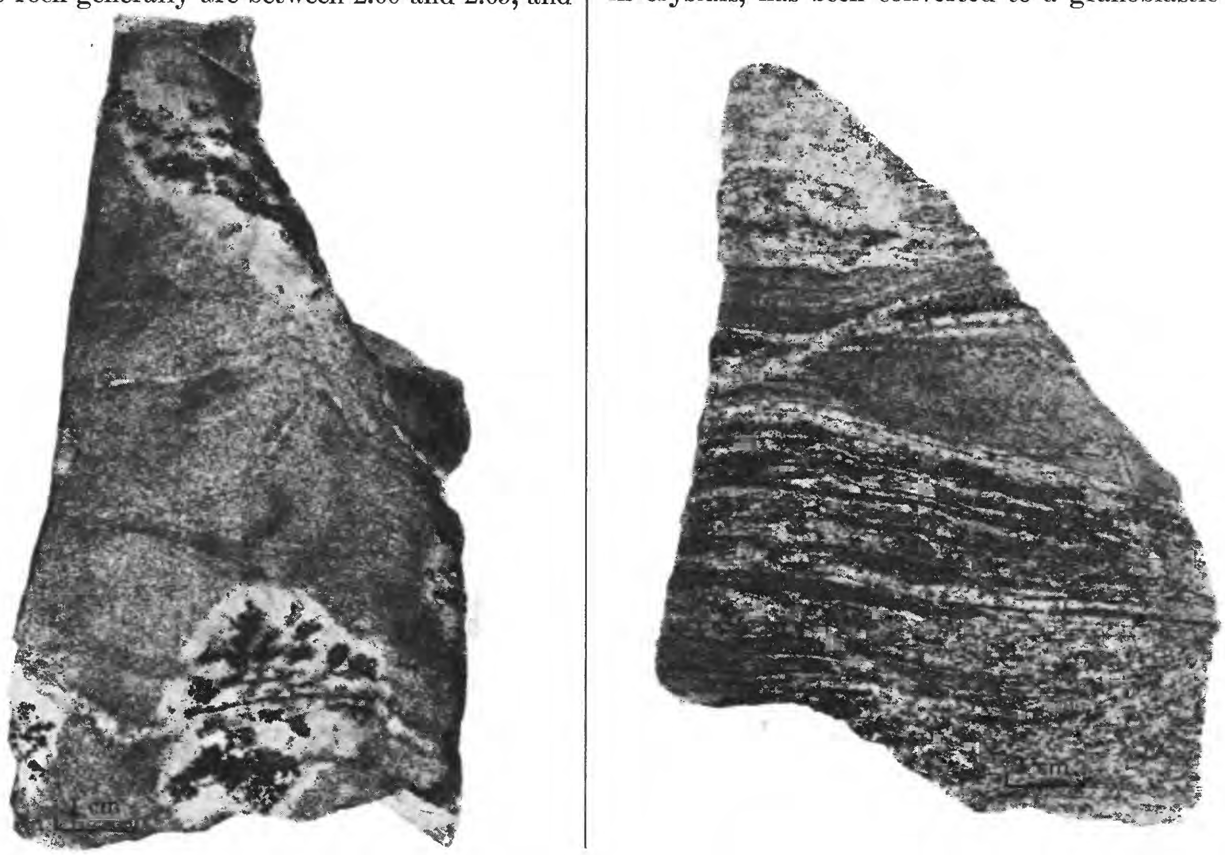

FIGURE 9.—Products of metamorphic differentiation in hornfels and in sheared rock of the same bulk composition. Left, recrystallization spots in quartz-biotite-plagioclase hornfels. Right, gneiss layers in foliated rock. Undifferentiated rock in the two specimens are mineralogically identical, and the minerals in the dark and light parts of the spots are the same as in the dark and light layers in the gneiss.

saic. The boundaries of the mineral fragments and crystals are still quite sharp against the groundmass and have been only slightly modified. Most of the foreign fragments are composed of plagioclase grains of intermediate composition a millimeter or so across and clusters of hornblende prisms intergrown in and bordering the plagioclase, but some are composed of biotite, quartz, and plagioclase. Magnetite and apatite are present in both kinds of fragments. The granoblastic groundmass of the tuff consists chiefly of quartz and microcline, but sodic plagioclase is also present, and some specimens examined contained as much as 10 percent of biotite.

The angular shapes of the crystal fragments, the slight amount of corrosion of the original borders of mineral fragments by the groundmass, the presence of foreign rock fragments, and the absence of flow structure are the chief lines of evidence that the rock is a tuff rather than a flow or dike rock. The absence of conspicuous metamorphic effects in the crystal fragments, except for the conversion of sanidine to microcline and the breakdown of quartz to mosaics, contrasts with the complete reconstitution of the groundmass. Finer grain size and the presence of glass may have made the groundmass more susceptible to recrystallization than the mineral fragments.

Quartz latite, chiefly and perhaps everywhere, in dikes and sills is present south of the Tungstar mine and along the contact between micaceous quartzite and metavolcanic rocks. Sills penetrate from the contact distances of several hundred feet along bedding planes in the micaceous quartzite. Fresh quartz latite commonly is medium light gray, slightly darker than the light-gray metarhyolite tuff. In most places the quartz latite is almost structureless, but in a few places it is strongly foliated. On weathering it breaks into polygonal blocks.

The quartz latite is porphyritic and contains abundant phenocrysts of feldspar as much as 3 mm across and somewhat smaller aggregates of biotite. Under the microscope, most of the large feldspar grains can be seen to be zoned plagioclase, but some are microcline. The feldspars and biotite aggregates are set in a granoblastic groundmass of quartz, microcline, and plagioclase, which has corroded the margins of the feldspars. Biotite in small plates also is present in some rocks, generally with a moderate to high degree of preferred orientation. Present in minor amounts are hornblende (locally in amounts of as much as 5 percent), sphene, magnetite, ilmenite, and apatite. Locally, biotite is altered to chlorite and plagioclase to epidote.

Following is a chemical analysis by the rapid method of Shapiro and Brannock (1956) of a typical specimen of dike rock from the west side of Mount Tom (lab. no. 138267; analysts, Harry F. Phillips, Paul L. D. Elmore, and Katrine E. White; locality, NE¼ sec. 22, T. 7 S., R. 30 E., 2,000 ft south of the Tungstar mine). Comparison of the analysis with the average chemical compositions of Nockolds (1954) indicates that the dike rock is very similar to his dellenite.

	Weight percent		*Weight percent*
SiO_2	70.5	K_2O	4.3
Al_2O_3	15.6	TiO_2	.32
Fe_2O_3	1.0	P_2O_5	.11
FeO	1.4	MnO	.06
MgO	.68	H_2O	.47
CaO	2.2	CO_2	.05
Na_2O	3.6		
		Total	100

MAFIC METAVOLCANIC ROCKS

The most extensive remnants of mafic metavolcanic rocks are in the southern one-third of the Pine Creek pendant; smaller masses are in a ring around Mount Humphreys, the western part of the Tungsten Hills, and 3 miles west of Keough Hot Springs. Most of the mafic metavolcanics are dark gray, slightly foliated nonporphyritic rocks of andesitic or dacitic composition, which lack distinctive characteristics. Locally, however, units are porphyritic, amygdaloidal, or a breccia.

The mafic metavolcanic rocks are closely associated with fine-grained rocks of panidiomorphic granular texture, which have been included with them on the map. These rocks are doubtless the hypabyssal equivalents of the mafic extrusives. The mafic metavolcanic rocks are also commonly associated with medium-grained quartz dorite, diorite, and hornblende gabbro of hypidiomorphic-granular texture. The medium-grained hypidiomorphic granular rocks are probably of diverse origins, and some may be genetically related to the mafic metavolcanic rocks. They have been mapped separately.

Under the microscope, pilotaxitic, trachitic, and panidiomorphic-granular textures can be seen among the metavolcanic rocks. A fine-grained granoblastic groundmass composed of plagioclase and hornblende generally is interstitial to plagioclase laths or plates, biotite plates, and hornblende prisms. Locally, the groundmass has encroached on the plagioclase crystals. Originally, the groundmass may have been glass or finely divided crystalline material, although no relicts of these materials were found.

The total plagioclase content generally is between 45 and 65 percent, and the combined biotite and hornblende content is between 30 and 40 percent. Commonly biotite and hornblende vary inversely with each other in relative abundance. Rocks of dacitic composition generally contain 5 to 15 percent quartz, and a few specimens also contain 5 to 10 percent K feldspar. Accessory minerals are apatite, magnetite and ilmenite, and sphene. Locally, biotite is altered to chlorite and hornblende to epidote.

Most of the euhedral or subhedral plagioclase crystals are strongly zoned and are undoubtedly relict primary igneous grains. In contrast, the plagioclase in granoblastic intergrowths generally is untwinned and zoned only slightly. In most zoned crystals of plagioclase the zoning is progressive from calcic core to sodic rim, but oscillating zoning is present in places, and may be superimposed on progressive zoning. Commonly, selected zones, usually the core, are strongly sericitized. Biotite is in thin pleochroic plates and generally exhibits moderate preferred orientation. Most of the hornblende is in small, elongate crystals, but locally larger and evidently later formed crystals poikiloblastically enclose small plagioclase crystals.

Amygdules were found in the septum 3 miles west of Keough Hot Springs, in meta-andesite, and in the south part of the Pine Creek pendant, in dacite. These amygdules commonly are ovoid with the greatest dimension as much as 1½ cm. The core consists of a mosaic of quartz, which is bordered by a rim of hornblende or diopside. On weathered surfaces the amygdules stand out and exhibit smoothly rounded external surfaces, which reflect the shapes of the vesicles. The presence of diopside marginal to some amygdules suggests that carbonate as well as quartz was present. In some amygdules a rim of antiperthitic plagioclase lies outside the hornblende rim, and in a few places hornblende in the rock marginal to the amygdules is altered to epidote.

Remnants of a layer of volcanic breccia are present in the south part of the Pine Creek pendant, in the floor of Horton Creek canyon; the best exposure is in a knob that projects through the talus a few hundred feet northwest of the west end of Horton Lake. A second good exposure is a few hundred feet east of the small lake below Horton Lake. This rock is a cemented aggregate of flattened volcanic rock fragments. The largest fragments are several feet long, but most fragments are a few inches or less in length and less than an inch across. Fragments include both mafic and felsic metavolcanic rocks that are identical with rocks in the surrounding terrane. The matrix is andesitic or dacitic in composition and presumably is tuffaceous.

The rocks containing amygdules are either flows or shallow intrusives, and the volcanic breccia layer is either of pyroclastic or sedimentary origin. Rocks having felty textures are presumed to be tuffs, flows, or shallow instrusives. Rocks of coarser panidiomorphic-granular texture are presumed to be intrusive. If the granoblastic texture was formed most readily in glassy or very finely crystalline material, the bulk of the mafic volcanic rocks seems likely to have been tuffs or flows rather than small intrusives, but the local presence of the granoblastic texture in rock having relict panidiomorphic-granular texture indicates that all rocks of granoblastic texture cannot be assumed to be tuffs or flows.

The following rapid chemical analysis (Shapiro and Brannock, 1956) is of a typical rock of andesitic composition from the southwest side of Mount Tom (lab. no. 138274; analysts, Harry F. Phillips, Paul L. D. Elmore, and Katrine E. White; locality, SW¼ sec. 22, T. 7 S., R. 30 E., one-half mile southeast of the Hanging Valley mine). In this rock, crystals of biotite, hornblende, and plagioclase are set in a granoblastic groundmass of quartz and plagioclase.

	Weight percent		*Weight percent*
SiO_2	55.5	K_2O	1.8
Al_2O_3	18.5	TiO_2	.74
Fe_2O_3	3.2	P_2O_5	.37
FeO	5.3	MnO	.16
MgO	3.9	H_2O	.85
CaO	7.0	CO_2	.05
Na_2O	3.0	Total	100

Norm

Quartz	9.00	Hypersthene	15.77
Orthoclase	10.58	Magnetite	4.64
Albite	25.18	Ilmenite	1.46
Anorthite	31.69	Apatite	.93
Diopside	.39		

CALC-HORNFELS DERIVED FROM MAFIC IGNEOUS ROCK

In several places, mafic metavolcanic rock and finer grained diorite have been converted to very light gray (almost white) plagioclase-diopside rock that is megascopically identical with plagioclase-diopside-hornfels derived from impure limestone. The altered rock differs from true calc-hornfels in two ways: (1) Much of its original felty texture is retained, and the hornfelsic texture, though present locally, is not dominant, and (2) plagioclase commonly retains its primary zoning, although locally the zonal structure has been obliterated. The altered rock, nevertheless, is here called calc-hornfels because without microscopic study it cannot be distinguished from true calc-hornfels.

Juxtaposition of mafic metavolcanic rock with felsic plagioclase-diopside rock indicates that the formation of hornblende in one and of diopside in the other cannot be explained in terms of different pressure-temperature environments. Both plagioclase-hornblende and plagioclase-diopside are stable assemblages in the amphibolite facies, and the factor that determines whether diopside or an amphibole forms is composition, chiefly the content of CaO—diopside is the stable mineral in assemblages rich in CaO (Barth, 1952, p. 328). CaO was introduced into the diopside-bearing rock either prior to or during the metamorphism, and the source is not likely to have been the granitic intrusive that caused the metamorphism. In some places CaO was obviously derived from associated calcareous metasedimentary rocks, and in others it was supplied by carbonate-bearing veins and amygdules that were late stage accompaniments to the intrusion of dikes and sills or the extrusion of lavas.

Some of the best examples of hornfelsed mafic igneous rocks are in the northeast part of the Bishop Creek pendant where dikes that cut marble are partly altered to calc-hornfels, and in the northern segment of the septa 3 miles west of Keough Hot Springs where amygdular lavas or shallow intrusives were partly hornfelsed. The mafic dikes in the Bishop Creek pendant are intricately penetrated by conspicuous light-colored calc-hornfels having geometrical relations that indicate the alteration involved no change in volume (fig. 10). Under the microscope the unaltered dike rock can be seen to consist chiefly of hornblende and plagioclase and the altered rock of diopside and plagioclase. In the altered rock, aggregates of tiny diopside grains have taken the place of the hornblende; in most specimens the plagioclase was not altered and retains its zonal structure. The boundaries between altered and unaltered rock generally are sharp, though in a few places they are gradational. The proportions of the altered and unaltered rock vary from place to place, from a few thin veinlets of calc-hornfels in unaltered dike rock to rock composed chiefly of calc-hornfels but containing a few residuals of unaltered rock.

FIGURE 10.—Mafic dike rock partly altered to plagioclase-diopside hornfels.

TABLE 2.—*Rapid chemical analyses of mafic dike rock and of hornfelsed rock derived from it*

[Gains and losses in the alteration are given in milligrams per cubic centimeter. Analysts: Paul D. L. Elmore, Katrine E. White]

	Unaltered dike rock (lab. No. 141351)		Hornfelsed rock (lab. No. 141352)			
					Mg per cc gained or lost in alteration	
	Weight percent	Mg per cc (sp gr 2.91)	Weight percent	Mg per cc (sp gr 3.03)	Oxides	Elements
SiO_2	48.6	1,414	49.4	1,497	+83	Si, +39.
Al_2O_3	16.1	469	15.0	455	−14	Al, −7.
Fe_2O_3	1.2	35	.6	18	−17	Fe, −61.
FeO	7.4	215	5.0	152	−63	
MgO	7.9	230	6.6	200	−30	Mg, −18.
CaO	11.0	320	18.9	573	+253	Ca, +187.
Na_2O	2.3	67	1.4	42	−25	Na, −18.
K_2O	1.4	41	.63	19	−22	K, −18.
TiO_2	.76	22	.66	20	−2	
P_2O_5	.20	6	.18	5	−1	
MnO	.19	6	.24	7	+1	
H_2O+	2.3	67	1.1	33	−34	
H_2O-	.14	4	.12	4	0	
CO_2	.08	2	.44	13	+11	
Total	100	----------	100	----------	----------	

The CaO required to produce diopside probably was derived from the calcareous strata into which the dikes were intruded and was introduced into the dike rock along cracks either before or during the metamorphism. To determine what chemical differences exist in the altered and unaltered rocks, samples of rock sorted from a single specimen having a patchy alteration pattern were analyzed chemically. Table 2 gives the analytical results and the calculated gains and losses of oxides and of elements.

The data show that Si, Ca, and CO_2 were gained in the altered rock, and that Fe, Mg, K, Na, Al, and H_2O+ were lost. The percentage changes in the amounts of other constituents are too small to be significant.

In the northern segment of the septa 3 miles west of Keough Hot Springs much of a mass of amygdular andesitic lava or dike rock is of mottled appearance because of splotchy alteration of the dark-gray rock to yellowish-gray calc-hornfels. The alteration is similar to the one that affected the mafic dikes in the Bishop Creek pendant, but the lime that was added to the hornfelsed rock was derived from amygdules and veinlets of quartz and carbonate.

The dark parts of the mottled rock consist of a microcrystalline felty mass of zoned plagioclase plates or laths (compositional range about An_{30-60}) and lesser amounts of hornblende and biotite, and minor K feldspar, sphene, apatite, and magnetite. In some rocks plagioclase also occurs in phenocrysts.

The yellowish-gray hornfelsic rock contains diopside rather than hornblende. The felty texture of the original flow rock is preserved in most parts of the altered rock, but locally the texture is granoblastic and the rock is true calc-hornfels. In rock that is granoblastic the plagioclase feldspar has lost its primary zoning.

The light-colored pyroxene-bearing rock generally is adjacent to amygdules that have cores of granoblastic quartz bordered by rims of diopside or green amphibole, or to veinlets composed of quartz and diopside. The excess CaO required to produce diopside rather than hornblende in the amphibolite facies doubtless was supplied by the amygdules and veinlets. Calcite is now a minor constituent in both, but diopside indicates an original higher content of carbonate.

KIND AND GRADE OF METAMORPHISM

The metamorphic rocks were subjected to weak regional metamorphism during the later stages of folding prior to the emplacement of the batholith, and to higher grade thermal metamorphism later, at the time of the emplacement of the batholith. The most conspicious products of the earlier regional metamorphism are the slates of the White Mountains and relict slaty cleavage in metamorphic remnants in the Sierra Nevada. In the Pine Creek pendant cleavage is identifiable in micaceous quartzite and in tectonically flattened fragments in metavolcanic breccia. The later thermal metamorphism resulted chiefly in the formation of hornfelses, but also, locally, in the formation of mica schist and gneiss.

The mineral assemblages in the metamorphic remnants are chiefly in the amphibolite facies of Eskola (1939; see also Turner, 1948, and Barth, 1952). In 1958, Turner (Fyfe, Turner, and Verhoogen, 1958, p. 199–239) separated facies of contact metamorphism from facies of regional metamorphism. In this scheme most of the hornfelses of the Bishop district fall into the hornblende hornfels facies, which formerly was called the cordierite-anthophyllite subfacies of the amphibolite facies. Brucite and wollastonite in calcareous assemblages indicate transition to the pyroxene hornfels facies, and garnet-bearing mica schist and assemblages that include both epidote and intermediate to calcic plagioclase indicate transition to the amphibolite facies of regional metamorphism, renamed, however, the almandine amphibolite facies.

In the hornfelses of the Bishop district, amphiboles and plagioclase are common in all rocks of appropriate composition. Diopside is common in hornfels having a high calcium content, but the association of diopside and hypersthene, considered diagnostic of the pyroxene hornfels facies, was not found. One criterion suggested by Turner (Fyfe, Turner, and Verhoogen, 1958, p. 206) for distinguishing the hornblende hornfels facies from the pyroxene hornfels facies, incompatibility of potassium feldspar with andalusite or cordierite, does not apply in the Bishop area because K feldspar and andalusite are found together in a terrane in which amphibole also is present. Rose (1958, p. 1703), who also recognized this problem in the metamorphic rocks of the Sierra Nevada, concluded that if the upper limit of the amphibolite (or hornblende hornfels) facies is defined by the transition of hornblende and calcic plagioclase to clinopyroxene, hypersthene, and plagioclase, the transition of the assemblage muscovite-biotite-quartz to andalusite-cordierite-microcline must occur well below the upper limit of the amphibolite facies. The presence of assemblages that include andalusite and potassium feldspar together indicates, however, that the associated rocks are in the upper rather than the lower part of the hornblende hornfels facies. No facies of lower grade in which the place of calcic plagioclase is taken by albite and epidote were found, although locally plagioclase has been saussuritized.

The mica schist and the gneiss pose a problem, since their formation requires both elevated temperatures and dynamic conditions. Their mineral grade is too high for them to have been formed during the earlier folding and regional metamorphism. It seems certain that they must have been formed approximately contemporaneously with the hornfelses, but under kinematic rather than static conditions. The composition of the rock also is significant, and pelitic rocks become schistose more readily than rocks of any other composition. Durrell (1940, p. 100–115) explained similar relations in the western Sierra Nevada by postulating regional orogenic stress continuing into the period of magmatic intrusion. He concluded that the granitic magmas were emplaced passively and exerted no deforming force on the wall rocks because he found no relation between the wall-rock structure and intrusive contacts, because of an absence of peripheral schistosity, and because of the presence of antistress minerals in contact-metamorphosed rocks (Durrell, 1940, p. 30). He wrote further (Durrell, 1940, p. 108):

> This high temperature in the wall rocks ahead of the magmas, combined with the stress existing there, would result in the formation, by recrystallization, of schists from such rocks as are capable of giving rise to minerals which can assume an orientation under such conditions. Close to contacts, in areas free of stress or where stress was very weak, where temperatures would be higher and recrystallization would be more intense, hornfelses would be expected to form.

Durrell's concept of regional stress contemporaneous with the intrusion of the granitic rocks cannot be tested within the Bishop district, but it does not afford a satisfactory explanation of the schist and gneiss there. Schist and gneiss are found in only two places, in septa along the South Fork of Bishop Creek and along Red Mountain Creek. The septum along Bishop Creek consists of hornblende-biotite-quartz-feldspar gneiss, and that along Red Mountain Creek of mica-quartz schist and associated garnetiferous gneiss. The fact that both the schistosity and gneissic foliation parallel the intrusive contact suggests a relation to the intrusions. The rock in both septa is pervasively sheared, and the gneissic layering appears to have been produced by the migration of mafic minerals to shear planes. According to Durrell's concept of regional stress, both septa should have been chiefly stress-free at the time of intrusion, since they are bounded by granitic rocks. On the whole the schistosity appears to be the "peripheral schistosity" which was absent in the area studied by Durrell. A freely flowing magma would hardly produce peripheral schistosity. Probably it could only be formed in rock of appropriate composition after the margins of the intrusive granite had achieved some strength through cooling and crystallization. Thus, peripheral schistosity is akin to marginal protoclasis, except that it takes place in the wall rock rather than in the marginal parts of the intrusive.

Forcible emplacement of certain other intrusives in the Bishop district is shown by wall-rock separations measurable in miles. Dislocations of wall rocks by intrusions are discussed on pages 115–116.

Bowen has distinguished thirteen steps in the progressive thermal metamorphism of siliceous dolomite (1940, p. 225–274), and Turner (1948, p. 80) includes the mineral assemblages formed above steps 3 to 6, inclusive, in his cordierite-anthophyllite subfacies of the amphibolite facies (hornblende hornfels facies). The steps are marked by the upper limits of stability of various mineral assemblages as follows:

Stable below step—	*Mineral assemblage*
1	Dolomite and quartz
2	Dolomite and tremolite
3	Calcite, tremolite, and quartz
4	Calcite and tremolite
5	Dolomite
6	Calcite and quartz
7	Calcite, forsterite, and diopside
8	Calcite and diopside
9	Calcite and forsterite
10	Calcite and wollastonite
11	Calcite and akermanite
12	Spurrite and wollastonite
13	Spurrite and akermanite

Harker and Tuttle (1956, p. 239–256) have shown that for pressures of less than 40,000 pounds per square inch (equivalent to the weight of about 35,000 feet of rock) steps 5 and 6 should be in reversed order; wollastonite will form from quartz and calcite at a lower temperature than calcite and periclase will form from clean dolomite.

Both diopside and tremolite are present in the metamorphic remnants of the Sierra Nevada, but whereas diopside occurs in assemblages that include calcite, tremolite was found only in assemblages free of calcite. Therefore, Bowen's step 4 was attained and step 8 was not. Dolomite was not identified in any of the metamorphic remnants, although its former existence is shown by diopside, tremolite, and in one place, brucite. The presence of brucite indicates clearly that step 5 (step 6 according to the data of Harker and Tuttle) was exceeded locally, but it cannot be assumed that it was attained everywhere because the paucity of brucite and the abundance of diopside and tremolite show that clean dolomite, necessary to record the step, was rare. Bowen's step 6 (step 5 according to Harker and Tuttle), reaction of calcite and quartz to form wollastonite, was attained in many smaller inclusions and in the marginal parts of larger ones, but calcite and quartz appear together in apparent equilibrium in many specimens from larger metamorphic remnants. Forsterite, included in the critical assemblage that becomes unstable above step 7, was not identified, and it could not be determined therefore whether step 7 was attained. In summary, it appears that step 4 was attained everywhere, that steps 5 and 6 were reached in most smaller inclusions and in the marginal parts of larger ones, and that step 8 was not attained anywhere.

Among the pelitic rocks, andalusite is the common aluminosilicate, although sillimanite was tentatively identified in a few places. The occurrences of sillimanite are too few to warrant speculation about the relation of sillimanite to andalusite.

The thermal metamorphism in most places appears to have been chiefly isochemical with introduction of material from the granitic magmas indicated only in connection with the formation of tactite, the host rock for the contact metasomatic tungsten deposits. The prevalence of sharp boundaries between unlike mineral assemblages, which coincide very closely with original boundaries between lithologically different sedimentary layers indicates that very little redistribution of material has taken place within the metamorphic remnants. The sharp boundaries between contiguous layers suggest further that the metamorphism was rather dry—that little water was present in the original sedimentary rocks and that little was introduced.

Nevertheless, locally some constituents were removed during metamorphism and others were redistributed. In the silication of limestone or marble to form silicated marble or calc-hornfels carbon dioxide was removed, and in the formation of tactite carbon dioxide was removed and calcium was removed in part. Carbonaceous material was driven out of some marble, chert, and calcareous and argillaceous hornfels. In places, very dark gray to black rocks that contain carbonaceous material grade along their strikes into white or very light gray rocks from which carbonaceous material has been removed.

Constituents that have been locally recirculated include calcium, iron, and magnesium. Beds of calc-hornfels not uncommonly encroach unevenly on bordering layers of fine siliceous rocks and on amphibolite, and in places veinlike layers of calc-hornfels anastomose through these same rocks. The encroachment of calc-hornfels layers on adjoining layers probably took place during metamorphism by progressive enlargement and modification of the chemical systems within the calcareous layers to incorporate the material in the transgressed layers. Dikelike stringers of calc-hornfels in siliceous hornfels or amphibolite may reflect premetamorphic calcite or dolomite veins. The source of the calcium generally was calcium-rich layers, but in mafic metavolcanic or dike rock a common source was calcitic amygdules and veinlets.

Iron and magnesium locally have been redistributed in mafic metavolcanic and plutonic rock to form veinlike layers and orbicular structures of hornblende and biotite.

The septum at the Blue Star Talc mine is of interest because it is the site of a commercial talc deposit which is believed to have been formed by the redistribution of constituents contained in the adjacent rocks. The outcrop is somewhat elongate, having a maximum dimension of about 1,000 feet and a minimum dimension of about 400 feet. Most of it consists of amphibolite, but its core is white and coarsely crystalline marble that contains brucite and small amounts of epidote. The mantle of amphibolite around the marble core suggests a large scale corona formed by the interaction of solutions from the enclosing granitic rocks with the marble. Petrographic study, however, indicates that at least part of the amphibolized rock was originally igneous. It is dense and very dark gray, and much of it consists chiefly of variable proportions of plagioclase and hornblende, although part of it consists almost entirely of amphibole with or without chlorite. Minor accessory minerals are biotite, sphene, magnetite (partly altered to hematite), and apatite. The amphibole commonly is pleochroic: X=pale brownish yellow, Y=brownish green, Z=olive green, but the pleochroism is variable, and some samples contain both a strongly colored amphibole and a very pale one. The habit of the amplibole also is variable. Some crystals are euhedral, and most are subhedral. Commonly the ends of larger prisms are frayed. In some specimens, the plagioclase also is of two compositions and of two habits. Zoned plates (An_{40-80}) appear lathlike in thin section and form a mat of unoriented crystals. In the interstices of these plates is oligoclase which has a lower index of refraction than that of the plates. The presence of plagioclase of two compositions is the chief evidence for an igneous parent rock. In other rocks in this area all the plagioclase is in irregular plates having the composition of oligoclase. Such rocks as well as rocks composed entirely of amphibole have been thoroughly reconstituted.

Talc is present locally in the amphibolite, adjacent to the marble. It is pseudomorphous after coarse radiating crystalline aggregates of amphibole as much as 4 inches across. Some of the adjacent amphibolite also contains small amounts of chlorite. The chemical change from amphibole to talc involves principally the addition of CO_2 and the loss of SiO_2. If the alteration took place with constant volume, the reaction can be expressed by the following equations (Turner, 1948, p. 133): (1)

$$\underset{\substack{\text{Tremolite}\\(810\text{ g; }270\text{ cc})}}{Ca_2Mg_5Si_8O_{22}(OH)_2} + 4CO_2 \rightarrow \underset{\substack{\text{Dolomite}\\(368\text{ g; }130\text{ cc})}}{2CaMg(CO_3)_2} + \underset{\substack{\text{Talc}\\(378\text{ g;}\\140\text{ cc})}}{H_2Mg_3Si_4O_{12}} + \underset{\substack{\text{(remov-}\\\text{ed in}\\\text{solution)}}}{4SiO_2}$$

and (2)

$$\underbrace{4H_4Mg_2Al_2SiO_9 + 6H_4Mg_3Si_2O_9}_{\text{Pennine (2768 g; 1025 cc)}} + 13.2SiO_2 \rightarrow$$

$$\underset{\substack{\text{Talc}\\(2759\text{ g; }1022\text{ cc})}}{7.3H_2Mg_3Si_4O_{12}} + \underbrace{4.1MgO + 4Al_2O_3}_{\text{(removed in solution)}} + 12.7H_2O$$

The CO_2 required for the conversion of amphibole probably was derived from the adjacent marble, possibly as a result of the formation of periclase (hydrated to brucite) and epidote.

Near the talc are veins as much as 1-foot thick that are composed chiefly of magnetite and specular hematite. The distribution of these veins with respect to the talc suggest that they are byproducts of the formation of talc from amphibolite. The iron oxides from which they were formed could have been those expelled from the amphibolite during its conversion to talc.

ARRANGEMENT OF METAMORPHIC REMNANTS IN RELATION TO THE REGIONAL DISTRIBUTION OF THE PREBATHOLITHIC ROCKS

The regional distribution of the prebatholithic rocks provides a framework that makes possible an interpretation of the stratigraphic and structural relations of the metamorphic rocks in pendants, septa, and inclusions in the Bishop district. The trends of the prebatholithic rocks that crop out in the ranges east of Owens Valley can be projected into the Sierra Nevada. Even though the batholithic rocks in the Sierra Nevada have bent and dislocated the rocks which they intrude, and have engulfed or pushed upward large segments of the rocks, a highly deformed skeletal pattern of these rocks still can be deciphered in the metamorphosed remnants.

The known distribution of pre-Cenozic rocks in the east-central Sierra Nevada and the adjacent White and Inyo Mountains is shown in figure 7. The general trend of the rocks is about N. 35°–40° W., a trend that cuts across Owens Valley and locally strikes obliquely into the Sierra Nevada batholith. Although the rocks in the White and Inyo Mountains are strongly folded and faulted, the rocks exposed along the western fronts of these ranges are, on the whole, progressively younger toward the south, ranging in age from Early Cambrian or older at the latitude of Bishop to Middle Triassic in the New York Butte quadrangle east of Owens Lake. This distribution suggests that the strata are progressively younger toward the west and constitute the west limb of an anticlinorium or the east limb of a synclinorium. Further evidence that the strata in this region are younger toward the west is found northwest of Bishop in the Mount Morrison and Ritter Range pendants, which contain a steeply dipping homoclinal sequence of Paleozoic and Mesozoic strata of more than 60,000 feet (Rinehart, Ross, and Huber, 1959, p. 941–945). These strata range in age from Ordovician in the east to Early Jurassic or younger in the west. In the western part of the Ritter Range pendant is a synclinal axis, west of which the section is reversed and the tops of beds are to the east.

The Paleozoic and Mesozoic strata in the western foothills of the Sierra Nevada, on the west side of the batholith, strike N. 35°–45° W. at the latitude of Bishop, parallel with the strata in the Owens Valley region. Studies by L. D. Clark (1960, p. 492–493) reveal that the tops of beds in most exposures are to the east but that the sequence is cut by several eastward-dipping reverse faults of large displacement. Very likely these strata constitute the west limb of a synclinorium whose east limb is in the Owens Valley region. The batholith occupies the axial region of this synclinorium.

The stratigraphic sequence in the White and Inyo Mountains can be divided into two lithologically distinguishable series; an older one composed of limestone, dolomite, shale, and quartzite, and little or no volcanic material, and a younger one that contains abundant volcanic material. The older series is chiefly Paleozoic and the younger series Mesozoic, but the nature and precise position of the contact between the two series are in some doubt. Near New York Butte (fig. 7) the volcanic series rests unconformably on Middle Triassic strata that contains little or no volcanic material (C. W. Merriam, oral communication, 1959). In the Mount Morrison and Ritter Range pendants, fossils of Permian(?) age have been found about 3,000 feet below the volcanic series, and fossils of Early Jurassic age have been found about 10,000 feet above the base of the volcanic series (Rinehart, Ross, and Huber, 1959); the contact therefore may be in about the same position as near New York Butte.

In a span of more than 100 miles, between the New York Butte area and the Mount Morrison pendant, most of the remnants of the volcanic series lie west of the nonvolcanic series, as they should if the strata are in the east limb of a synclinorium. The trace of the contact between the two series is a remarkably straight line trending N. 30°–35° W., although irregularities are present locally. The comparative straightness of the overall contact results from the prevailing steep or vertical dips of the beds.

Inspection of figure 7 shows that the metamorphic remnants in the Bishop district lie between strata whose age is known or can be inferred within broad limits. Both the volcanic series and the nonvolcanic series are represented in the district and are in contact in the Pine Creek pendant. The highly generalized conclusion that the volcanic rocks in the metamorphic remnants are chiefly of Mesozoic age is warranted. Furthermore, sedimentary strata that stratigraphically underlie metavolcanic strata are likely to be of late Paleozoic age.

DESCRIPTIONS OF METAMORPHIC REMNANTS

PINE CREEK PENDANT

The Pine Creek pendant (pl. 2; fig. 7) is a lens of metamorphic rocks almost 7 miles long and 1 mile wide. From the northeast face of Basin Mountain it extends N. 20° W. across Horton Creek, Mount Tom, and Pine Creek into the south end of Wheeler Crest. Outcrop altitudes range through more than 6,000 feet—from a low point of about 7,400 feet on the floor of Pine Creek to a high point of 13,652 feet on the summit of Mount Tom. Everywhere except at the north end where it is cut off by dark hornblende gabbro, the pendant is bounded by light-colored granitic rocks assigned to several different formations. Both north and south of the pendant, but separated from it by intrusive rocks, are thin septa and inclusions composed of metamorphic rocks like those in the pendant. The northern continuation, the Wheeler Crest septa, extends beyond the northern boundary of the Mount Tom quadrangle and separates Round Valley Peak granodiorite on the west from Wheeler Crest quartz monzonite on the east. A southern extension from the pendant forms a ring of metamorphic rock that circles Mount Humphreys and a septum that extends south from the ring to the North Fork of Bishop Creek, called the Mount Humphreys ring and septum on the map (fig. 7).

The Pine Creek pendant and its extensions include three mappable metasedimentary rock units of Paleozoic(?) age and two predominantly metavolcanic units of Mesozoic(?) age. The metasedimentary rocks comprise the northern two-thirds of the Pine Creek pendant, the Wheeler Crest septum, and the southern one-third of the ring around Mount Humphreys and the attendant septum that continues to the south; the metavolcanic rocks comprise the southern one-third of the pendant and the northern two-thirds of the ring around Mount Humphreys.

The oldest metasedimentary unit is composed of pelitic hornfels, micaceous quartzite, and vitreous quartzite. The next stratigraphic unit is marble, which is overlain, in turn, by micaceous quartzite. The metavolcanic rocks are divided, simply, into a unit of felsic metavolcanic rocks and a unit of mafic metavolcanic rocks. Probably the felsic metavolcanic rocks are the older, but the age relations are uncertain.

The main structure in the northern two-thirds of the Pine Creek pendant is a tight syncline whose axis lies within the micaceous quartzite. Except at the south end, the synclinal axis trends N. 20° W., and the axial plane is vertical. The beds in both limbs dip steeply and are generally parallel; gently dipping beds are found only within a few feet of the axial plane. Beds with gentle dips are visible on the south side of Pine Creek Canyon a few hundred feet west of the juncture of Pine and Gable Creeks, along the trail to the Tungstar mine. If the same general locality is viewed from the switchbacks on the road on the north side of Pine Creek leading to the Pine Creek mill, beds can be traced visually through a half circle, from one nearly vertical limb to the other. The synclinal axis can also be identified in the north wall of Pine Creek Canyon and 1½ miles to the south in the canyon cut across the pendant by Gable Creek. Because of the steepness of the terrane, the structure in the north end of the pendant was not worked out, but the syncline is presumed to continue into this area. Farther south on the northwest side of Mount Tom, the syncline is bent around to an easterly trend. At the bend it is cut by a wedge-shaped dike along which the pendant is offset laterally about half a mile.

The marble that flanks the micaceous quartzite on the west is absent from the east side of the pendant because it is cut out by quartz monzonite. However, a metamorphic remnant that consists chiefly of marble crops out in the ridge on the southeast side of the cirque at the head of Elderberry Creek. This remnant is separated from the pendant about half a mile by quartz monzonite. The presence of slivers of micaceous quartzite within the marble on the west side of the pendant suggests folding or faulting within the marble unit, but the slivers may be the result of lenticular sedimentation.

The metavolcanic and associated metasedimentary rocks in the south end of the pendant are in fault contact with the marble and micaceous quartzite in the central and north parts. An unconformity is unlikely because it would require an improbable series of events: folding of the micaceous quartzite and, rotation of the fold axes to near-vertical positions; then, after deposition of the overlying volcanic rocks, rotation of the fold axes back to their present near-horizontal positions. Nevertheless, in most places the angular discordance between the micaceous quartzite and the metavolcanic sequence is only a few degrees. Much of the contact is occupied by diorite and felsic dikes, some of which penetrate into the micaceous quartzite distances of a few hundred feet. Both the metavolcanic rocks and the south end of the micaceous quartzite have been bent into a broad S-fold which also affects, in part, the synclinal axis within the micaceous quartzite. The strata are bent from their N. 20° W. strike farther north to an easterly strike in the south side of Mount Tom for about 1½ miles west from Horton Lake, and then are bent to the southeast in the northeast side of Basin Mountain. In this part of the pendant, hornblende gabbro, quartz diorite, quartz monzonite, and thin granitic dikes have intricately penetrated, reacted with, and physically displaced the metamorphic rocks.

The syncline in the northern two-thirds of the pendant parallels the regional fold axes and is probably the result of prebatholithic deformation. The S-fold in the south end of the pendant, however, can be most readily explained as the result of deformation caused by the forcible emplacement of the quartz monzonite. In the formation of the S-fold, the earlier structures were deformed, and the pattern is not one that suggests continuation of the same stresses that produced the prebatholithic folds. On the contrary the S-fold is confined to a part of the pendant that is in contact with the quartz monzonite. Furthermore, the intricate penetration and shattering of the metamorphic rocks in the S-fold by dikes, sills, and irregular masses of granitic rocks that are satellitic to the quartz monzonite suggests that the S-fold was caused by the forcible emplacement of the quartz monzonite.

PELITIC HORNFELS, MICACEOUS QUARTZITE, AND VITREOUS QUARTZITE

A lens of metamorphosed clastic sedimentary rock, including pelitic hornfels, micaceous quartzite, and vitreous quartzite, is preserved along the west side of the pendant in the vicinity of Pine Creek, and a smaller mass of similar rocks forms a dislocated mass along the east side. The beds in the west side of the pendant dip steeply and strike parallel with the long axis of the pendant. They have a maximum stratigraphic thickness of about 1,000 feet, but are cut off on the west by quartz monzonite. The strata are mostly iron-stained pelitic hornfels and micaceous quartzite, and are only locally unstained vitreous quartzite. The unit has been penetrated by numerous aplitic and pegmatitic dikes that are offshoots from a quartz monzonite pluton that lies to the southwest. A few hundred feet west of the clastic strata and separated from it by quartz monzonite is a mass of marble and tactite that has a stratigraphic thickness of about 100 feet. The tactite in this mass has been exploited in the Brownstone tungsten mine. Whether this remnant of marble and tactite represents a thin interbed in the clastic strata or part of a stratigraphically lower carbonate unit is not known.

The small mass of pelitic hornfels, micaceous quartzite, and vitreous quartzite on the east side of the pendant is at the east end of a small mass of marble that contains the Lambert tungsten mine. This mass is in the opposite limb of the major syncline of the pendant, and the clastic strata lie on the east side of the marble. The presence of vitreous quartzite is considered good evidence of the stratigraphic equivalence of the strata here with those in the west side of the pendant, because it is such an uncommon rock in the region.

MARBLE

East of and stratigraphically overlying the lens of pelitic hornfels, micaceous quartzite, and vitreous quartzite in the west side of the Pine Creek pendant is an extensive belt of marble. This marble is more than 3 miles long, extending from the south wall of Pine Creek northward to the north end of the pendant, and about one-third of a mile wide at the widest place. The strata dip steeply or vertically, but the outcrop width probably does not represent the stratigraphic thickness because both isoclinal folds on a small scale and bedding plane shears are present. Enclosed in the marble are two lenses of micaceous quartzite, which may be either sedimentary lenses that were deposited with the marble or infolded segments of the stratigraphically overlying unit of micaceous quartzite.

The marble also crops out in the east side of the pendant in the dislocated mass that contains the Lambert tungsten mine. The apparent thickness of marble there, measured between the pelitic hornfels, micaceous quartzite, and vitreous quartize unit on the east and the overlying micaceous quartzite on the west, is about 800 feet.

The marble in the west flank of the pendant is light gray to medium light gray away from granitic contacts. Near granitic contacts it has been recrystallized and bleached to very light gray (almost white). In most places this bleached and recrystallized zone includes only a few feet adjacent to the granitic contact, but near the ore bodies of the Pine Creek mine it includes most of the width of the marble.

In hand specimen the marble appears to be fairly clean, but quartz and various calc-silicate minerals can be seen in thin section—grossularite, plagioclase, idocrase, diopside, and wollastonite. Samples from the Pine Creek mine that were analyzed chemically contained only 70 to 80 percent $CaCO_3$. Common impurities are 15 to 20 percent of SiO_2, 3 to 4 percent of Al_2O_3, and about 3 percent of MgO.

Because of its association with micaceous quartzite, the marble present in the Wheeler Crest septa is believed to be correlative with the marble unit of the pendant. Marble in the south part of the ring around Mount Humphreys and in the septum that extends south from it also is associated with micaceous quartzite and may be correlative with the marble in the west flank of the Pine Creek pendant, but locally it contains abundant thin hornfels layers that are not found in the marble unit of the pendant.

MICACEOUS QUARTZITE

The micaceous quartzite unit makes up the larger part of the northern two-thirds of the Pine Creek pendant and most of the Wheeler Crest septa, and is probably represented in the septum that extends south from the Mount Humphreys ring. This unit is almost all micaceous quartzite, but locally it includes a few interbeds of calc-hornfels, and at the Ridge ore body of the Adamson mine it contains tactite formed from a bed of marble.

All the micaceous quartzite is fine grained except for a few pebbly beds exposed in the glaciated floor of Pine Creek Canyon in the NW cor. sec. 9, T. 7 S., R. 30 E. The bedding is steep and strikes parallel with the external boundaries of the unit except close to the axial plane of the syncline where the beds dip gently or are horizontal. Bedding is generally obscure and can be determined only by careful searching for compositionally distinctive layers. In many outcrops the micaceous quartzite is cleaved parallel with the axial plane of the major syncline. Generally the cleavage is at only a slight angle to the bedding, but near the axial plane of the syncline it is at a large angle. The whole unit is stained various shades of brown by iron oxides derived from pyrite, but on fresh surfaces most rocks are dark gray to grayish red. Feldspathic varieties generally are pale yellowish brown, and calc-hornfels layers are medium light gray. The rock weathers into polygonal blocks a few inches on a side.

About 3,000 feet of steeply dipping strata are exposed in the north wall of Pine Creek Canyon between the axial plane of the syncline and the marble unit in the west side of the pendant. Inasmuch as the top of the micaceous quartzite unit is not present, the total thickness of the unit must be somewhat greater. Duplication by folding is possible but unlikely, hence 3,000 feet is a reasonable figure for the minimum thickness.

FELSIC METAVOLCANIC ROCKS

Felsic metavolcanic rocks crop out on the south and southwest sides of Mount Tom in a belt about 9,000 feet long and 2,000 feet wide. They are in an area that is structurally complex and penetrated by numerous dikes of aplite, quartz monzonite, diorite, quartz diorite, and hornblende gabbro. On the north the felsic metavolcanic rocks are in fault contact with micaceous quartzite, and on the south they are bordered by mafic metavolcanic rocks. They terminate on the west against metasedimentary rocks and diorite, and on the east against diorite, hornblende gabbro, and quartz monzonite.

The principal felsic metavolcanic rocks are metarhyolite tuff, and quartz latite sills, dikes, and probably flows. Metarhyolite tuff is well exposed in the western part of the area and quartz latite in the eastern part. Metarhyolite tuff is particularly well exposed along the ridge northwest of the Hanging Valley mine. The tuff is generally massive, but locally shows a foliation that probably is secondary. Euhedral quartz and microcline (originally sanidine) and small dark lithic fragments are visible in hand specimen. Quartz latite is well exposed south of the Tungstar mine. Many of these rocks are foliated. Rounded and corroded plagioclase phenocrysts are conspicuous in hand specimens. Along the contact of the felsic volcanic rocks with micaceous quartzite, sills of quartz latite penetrate the micaceous quartzite along bedding planes.

Associated with the felsic metavolcanic rocks are amygdular volcanic rocks of dacitic composition and such metasedimentary rocks as siliceous hornfels, marble, and schistose rocks that probably are tuffaceous. Siliceous hornfels is the most abundant metasedimentary rock and is present in the Tungstar mine area. Marble is also present at the Tungstar mine, and a string of lenses extends east from the mine toward the summit of Mount Tom.

The metasedimentary rocks east of the felsic metavolcanic rocks, in the vicinity of Gable Lakes, are chiefly feldspathic quartzite and pelitic hornfels, but include a thin belt of marble that can be traced northwest from the Lakeview mine for about half a mile. The Hanging Valley mine, about half a mile east of the Lakeview mine, may be in the same marble bed.

MAFIC METAVOLCANIC ROCKS

Mafic metavolcanic rocks lie southwest of the felsic ones and extend from the south side of Mount Tom, across Horton Creek, into the northeast side of Basin Mountain. Similar rocks farther south in a ring around Mount Humphreys are separated from those in Basin Mountain by Tungsten Hills quartz monzonite.

The most common mafic volcanic rock is dark-gray fine-grained andesite having a faint to conspicuous foliation that is probably secondary. The principal minerals are intermediate plagioclase, quartz, hornblende, and biotite. Accessory minerals are magnetite, apatite, and, less commonly, sphene. Most specimens have a granoblastic groundmass, which locally encloses larger crystals of zoned plagioclase and hornblende. Oriented biotite gives some specimens a lepidoblastic texture. Some rocks are obviously bedded, and most of them are probably tuffs. Lenses of marble, calc-hornfels, and micaceous quartzite are present locally, and several lenses include tungsten prospects.

The only marker bed recognized within the mafic metavolcanic rocks is a layer of volcanic breccia a few hundred feet thick, which extends from the northwest side of Horton Lake eastward along Horton Creek for more than a mile. This bed consists of angular flattened fragments of metavolcanic rock in a fine-grained tuffaceous matrix containing abundant quartz. The fragments range in length from less than an inch to more than 3 feet, and are fairly well sorted. Most of the volcanic rocks in the pendant are represented among the fragments.

PROBABLE CORRELATION WITH THE MOUNT MORRISON PENDANT

The metasedimentary strata in the Pine Creek pendant match closely those in the Bloody Mountain block of the Mount Morrison pendant 15 miles to the northwest (fig. 7), which contain fossils of Pennsylvanian and Permian(?) age (Rinehart, Ross, and Huber, 1959). The suggested correlation between the Pine Creek pendant and the Bloody Mountain block is shown in figure 11. Correlation of the strata in the

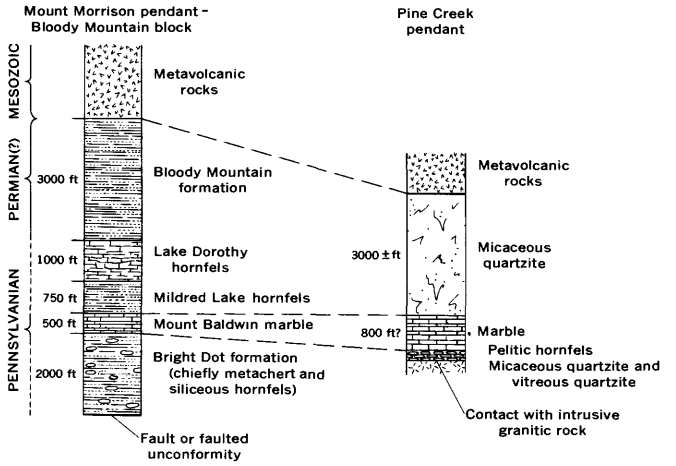

Figure 11.—Suggested stratigraphic correlation of the Pine Creek pendant with the Bloody Mountain block of the Mount Morrison pendant.

two pendants is based on similarity of lithologic sequence. In both pendants a thick marble unit stratigraphically overlies metachert or siliceous hornfels and is overlain by a thick section of fine-grained siliceous clastic rocks. The section of siliceous clastic rocks is overlain in turn by metavolcanic rocks, though in the Pine Creek pendant the contact is a fault contact. In the Mount Morrison pendant the strata above the marble include a calc-hornfels unit about 750 feet thick, which was not identified in the Pine Creek pendant. Two explanations for the apparent absence of the calc-hornfels unit can be offered: (1) The calc-hornfels unit is present in the Pine Creek pendant but was not mapped separately—beds of calc-hornfels are present within the micaceous quartzite unit, and more detailed study might result in recognition of a calc-hornfels unit. (2) The calc-hornfels unit in the Mount Morrison pendant may grade to micaceous quartzite or pelitic hornfels in the interval between the Mount Morrison and Pine Creek pendants. Even within the Bloody Mountain block the calc-hornfels unit grades along the strike to more siliceous rock (Rinehart and Ross, 1964).

BISHOP CREEK PENDANT

The Bishop Creek pendant occupies an irregularly shaped area of about 20 square miles. The strata have been deformed, as in the Pine Creek pendant, by regional compression that resulted in tight north- to northwest-trending folds, and somewhat later by emplacement of the intrusive rocks, which caused complex and irregular deformation of the regional structures. Apophyses from the bordering granitic plutons and dikes, sills, and stocks of diorite and hornblende gabbro penetrate the pendant extensively.

The heart and most cohesive part of the pendant lies along Coyote Ridge, on the east side of the South Fork of Bishop Creek. From this central core, lobes extend northwest across Table Mountain, northeast to Lookout Mountain and beyond, and south to the ridge southwest of Green Lake. From the northwest lobe a septum extends northwest across the Middle Fork of Bishop Creek. Many small inclusions are peripheral to the pendant and separated from it by granitic rock.

The pendant is composed of metasedimentary rocks, and in the central core six units were distinguished in mapping. These are, in order of decreasing age, (1) pelitic hornfels with interbeds of marble, (2) marble, (3) banded calc-hornfels and pelitic hornfels, (4) metachert and andalusite-bearing pelitic hornfels, (5) siliceous calc-hornfels, and (6) micaceous quartzite and pelitic hornfels.

Structural features that resulted from the earlier regional deformation are most readily identified in the central body of the pendant. The principal structures are a syncline on the east and an anticline on the west. Most of the beds are steep or vertical and trend north. The anticline is well exposed in the steep wall east of the South Fork of Bishop Creek about 1.6 miles north from Parchers Camp, but the syncline is only poorly exposed in Coyote Ridge, and its existence and position was established only by the duplication in inverse order of the stratigraphic sequence. Neither fold exhibits a geometric relation to any of the intrusive masses that indicates genetic relation, whereas the fold axes are approximately parallel with the regional fold axes. The anticline where exposed in the South Fork of Bishop Creek is generally symmetrical but pinched at the crest. As a result of this pinching the axis for about 1 mile southeast from the exposure of the anticlinal structure lies within a narrow band of calc-hornfels (pl. 1). The syncline shows little effect of the emplacement of the intrusives, except at the south end where strata have been pushed westward by a tongue of quartz monzonite of Cathedral Peak type. The west side of the anticline, on the other hand, has been deformed by a tongue of Tungsten Hills quartz monzonite along the South Fork of Bishop Creek, by a salient of Lamarck granodiorite that penetrates the calc-hornfels unit at the core of the anticline, and by the tongue of quartz monzonite that cuts off the syncline at the south end. Each intrusive pushed the strata in the syncline laterally and produced irregularities in their strike.

The pendant is cut off at the north end of the central body by Tungsten Hills quartz monzonite that protrudes southward and separates the two lobes of the pendant. The quartz monzonite pushed southward into the pendant and spread the strata in the lobes.

Distortion of the strata by the quartz monzonite is especially clear in the northeastern lobe where several stratigraphic units can be traced from the east flank of the syncline in the central body of the pendant through a broad curve concave to the northwest, which follows the margin of the quartz monzonite. In the south part of the lobe, the strata are bent smoothly and regularly, whereas in the north part, in the vicinity of Lookout Mountain, they are cut by faults and are penetrated by small apophyses from the quartz monzonite. These faults and intrusive apophyses have so dislocated the strata that it was not possible to match the strata of the blocks. In the very southeast corner of the Mount Tom quadrangle is a block that contains a synclinal axis, but the relationship of this structure to the syncline of the central body of the pendant was not determined. The fact that locally some faults have served as guides for apophyses of quartz monzonite suggests that the faults were formed during an earlier period of regional deformation.

The northwest lobe of the pendant has been almost separated from the central body by a tongue of quartz monzonite exposed along the South Fork of Bishop Creek. Isoclinal folds along the east side of the lobe could have been caused by the emplacement of the tongue of quartz monzonite; if they are regional folds they have been deformed by the quartz monzonite. Most of the beds in the north and northwest parts of the northwest lobe dip moderately toward the northwest. At the west edge of the lobe the beds were pushed upward by Tungsten Hills quartz monzonite, which may also have pressed the rocks in the septum against the earlier emplaced Lamarck granodiorite and caused the steep dips in the septum.

PELITIC HORNFELS WITH INTERBEDS OF MARBLE

The oldest strata that can be fitted into the stratigraphic sequence in the central part of the pendant are in a poorly exposed sequence that lies along the southeast side of the West Fork of Coyote Creek. The strata are chiefly dense medium-gray pelitic hornfels layers 100 feet or more thick separated by layers of marble 50 to 100 feet thick. The hornfels generally is composed of quartz and K feldspar in variable proportions. Coarser grained layers commonly contain more quartz than K feldspar, but finer grained layers generally contain more K feldspar. Diopside is a common constituent and is most abundant in coarser layers. The hornfels probably was derived from interlayered limy shale and very fine calcareous silt.

The thickness of the unit is difficult to determine because it has been irregularly intruded by granitic rocks. The beds dip steeply, and the maximum outcrop width is about 3,000 feet, which is the approximate thickness of the sequence if the strata are not duplicated.

The sequence of pelitic hornfels and interbeds of marble is bounded on the southeast by granitic and mafic intrusive rock, but beyond the igneous rocks are several isolated masses of calc-hornfels. The relation of these masses of calc-hornfels to the pendant is uncertain, but it seems reasonable to assume that they are remnants of a once continuous formation that lay stratigraphically just below the pelitic hornfels and interbeds of marble. The largest remnants, in secs. 29 and 31, T. 8 S., R. 32 E., and secs. 6 and 7, T. 9 S., R. 32 E., are shown on the geologic map (pl. 1) of the Mount Goddard quadrangle as calc-hornfels. The rock is dark greenish gray, dense, and structureless except for compositional layering that reflects bedding.

MARBLE

Crystalline white marble lies stratigraphically above the sequence of pelitic hornfels and marble interbeds along the West Fork of Coyote Creek. It is also present in the escarpment on the east side of Coyote Ridge. The width of outcrop of the marble is about 500 feet, which probably approximates its thickness. Near the middle of the marble is a distinctive quartzite layer about 50 feet thick. This layer is present in several outcrops along the east side of Coyote Ridge, but apparently is absent to the northeast.

BANDED CALC-HORNFELS AND PELITIC HORNFELS

Contiguous with the marble on the northwest and west is a hornfels unit that is characterized by conspicuously banded sequences of alternately light-gray calc-hornfels and dark-gray pelitic hornfels. The banded sequences are thin bedded—each bed is a quarter of an inch or less in thickness. The banded appearance is accentuated on weathered surfaces by staining of the light-gray beds to grayish orange. Although the banded sequences are the most distinctive assemblages, grayish-black to brownish-gray or light olive-gray to yellowish-gray hornfels that are thin bedded but which are not conspicuously banded are also present.

In the banded hornfels, the dark-gray pelitic beds are much finer grained than the lighter colored ones of calc-hornfels. The dark pelitic beds consist chiefly of very fine grained quartz, K feldspar, and carbonaceous material, accompanied in places by tremolite. The calc-hornfels beds contain abundant diopside, generally with quartz and K feldspar, although in some specimens diopside is the only mineral. Accessory minerals include apatite and pyrite. In beds where quartz is more abundant than diopside, the quartz is in grains that commonly are size graded across the beds. The color difference between the pelitic hornfels and calc-hornfels beds is caused by the presence of much fine carbonaceous material in the pelitic hornfels.

Light-colored calc-hornfels layers generally have wavy contacts with the contiguous darker pelitic layers; streaks from the light-colored layers penetrate into the dark ones and indicate that the light-colored layers have grown at the expense of the dark layers. This encroachment involved movement of calcium from the light-colored layers into the margins of adjacent pelitic layers and the removal of carbonaceous material. This replacement could have taken place either before or during thermal metamorphism. In the Lookout Mountain area light-colored hornfels locally cuts across dark pelitic layers, apparently along fractures (fig. 12).

The hornfels that is not conspicuously banded consists chiefly of approximately equal amounts of tiny grains of quartz and unoriented biotite flakes, and a small percentage of sericite and carbonaceous material. Layers of predominantly quartz alternate with layers in which sericite predominates, and apparently reflect original alternate layers of silt and clay.

METACHERT AND ANDALUSITE-BEARING PELITIC HORNFELS

Metachert and andalusite-bearing pelitic hornfels are stratigraphically above the banded hornfels and comprise the lowest unit that is exposed in both limbs of the syncline in the core of the pendant. In the eastern limb the unit lies along the east side of Coyote Ridge and the north side of the West Fork of Coyote Creek, and in the western limb it crops out along both sides of the South Fork of Bishop Creek. Disconnected masses having unknown structural relations are also present in the northeastern lobe of the pendant.

FIGURE 12.—Banded calc-hornfels (light-colored) and pelitic hornfels in which calc-hornfels encroaches on and cuts across pelitic hornfels. Crosscutting calc-hornfels was formed by introduction of calcium into pelitic hornfels along fractures and elimination of carbonaceous material. Narrow white streaks within crosscutting calc-hornfels are composed of quartz.

The most abundant rock in the unit is light-gray vitreous metachert that closely resembles quartzite. Locally this rock grades into dark-gray to black metachert from which it was derived by further recrystallization of quartz and expulsion of carbonaceous material. Interbedded with metachert of both colors is dark-gray to grayish-black, carbonaceous andalusite-bearing pelitic hornfels. Although the metachert is the more abundant rock in most places, pelitic hornfels predominates over metachert in the northeast lobe of the pendant. Commonly metachert and hornfels are interlayered in beds ranging from 1 to 6 inches thick.

Fine-grained micaceous or feldspathic quartzite or quartz-sericite hornfels is locally interlayered with the two rocks just described. The distribution of sericite in patches in some of these rocks suggests that the sericite was derived from andalusite. The rocks are readily distinguishable from metachert by the abundance of K feldspar or sericite, and in some specimens also by either the microscopically visible size-grading of quartz grains or by the angular outline of the quartz grains. Many of these rocks contain abundant carbonaceous material.

SILICEOUS CALC-HORNFELS

Siliceous calc-hornfels stratigraphically overlies the metachert and andalusite-bearing pelitic hornfels unit in the flanks of Table Mountain and along both sides of Coyote Ridge. The calc-hornfels on the east side of Coyote Ridge crops out in a band that extends northeastward toward Lookout Mountain where several disconnected masses with unknown structural relations occur. The unit consists predominantly of siliceous hornfels, but locally, especially along the lower stratigraphic boundary, it includes gray to white marble. Most beds of siliceous calc-hornfels are very light to yellowish gray, but some quartz-rich layers are darkened by carbonaceous material.

Individual beds are usually an inch or less thick, but some are several inches thick. The mineral content varies from layer to layer, but quartz is a principal constituent of most layers and is generally accompanied by tremolite or diopside or both. Commonly quartz forms a fine-grained granoblastic groundmass in which larger crystals of diopside or tremolite are enclosed. Microcline is present in the groundmass of some layers and in a few it predominates over quartz. Tremolite is in prisms that in some specimens have a random orientation and in others lie with their long axis in the plane of the bedding. Larger diopside crystals are poikiloblastic, and enclose many small grains of the groundmass minerals. Accessories include pyrite, magnetite, sphene, and apatite. Some quartz-rich rocks are darkened by carbonaceous dust.

These rocks were probably derived from dolomitic sandstone or siltstone, but beds derived from less siliceous carbonate rocks also are present. They contain such minerals as calcite, diopside, wollastonite, garnet, idocrase, scapolite, plagioclase, and zoisite.

MICACEOUS QUARTZITE AND PELITIC HORNFELS

A unit composed chiefly of micaceous quartzite and pelitic hornfels overlies siliceous hornfels in the north part of Table Mountain and in Coyote Ridge, and is the youngest unit in the pendant. Coarser grained rocks in which quartz is the most abundant mineral are included under micaceous quartzite, and finer grained rocks that contain abundant feldspar and biotite under pelitic hornfels, but the two kinds of rock are completely gradational. The only other rock in the unit is lenticular calc-hornfels. Megascopically, the micaceous quartzite and the pelitic hornfels are similar. Fresh surfaces commonly are light to dark gray—less commonly light olive gray, yellowish brown, or yellowish gray. Some surfaces are mottled gray, yellowish gray, and yellowish brown, and black lenticular streaks and equidimensional white spots are present locally. Weathered surfaces commonly are stained with iron oxides, though the staining is not as heavy or as widespread as in the micaceous quartzite unit of the Pine Creek pendant. Although the unit somewhat resembles the micaceous quartzite of the Pine Creek pendant, the rocks are generally finer grained, pelitic hornfels layers are more common, and individual layers are better sorted. The grayish red color on fresh surfaces, so common in the micaceous quartzite of the Pine Creek pendant, is lacking.

AGE OF THE STRATA IN THE BISHOP CREEK PENDANT

The sequence of strata in the Bishop Creek pendant is unlike the sequence in the Pine Creek pendant, and probably includes a different stratigraphic interval. The absence of metavolcanic rocks indicates that the strata are older than those in the Pine Creek pendant. The only fossil remains are worm castings that were found at a single locality in the siliceous calc-hornfels. Charles W. Merriam, who examined them, states that castings of this sort are abundant in the Cambrian, although they are found in rocks of a wide age span (oral communication, 1959). The stratigraphic sequence does not appear to correlate with either the sequence of Ordovician strata in the Mount Morrison pendant (Rinehart, Ross, and Huber, 1959) or the Precambrian(?) and Lower Cambrian sequence mapped in the White Mountains in connection with this report, but it may fall between these sequences. C. A. Nelson (1962) has mapped 12,000 feet of *Olenellus*-bearing strata in the Inyo Mountains, only the lower 1,500 feet of which is represented in the Bishop district. Strata of Middle and Late Cambrian and Early Ordovician age also are present in the Inyo Mountains, and the strata in the Bishop Creek pendant may be wholly or partly equivalent to the Ordovician strata of the Mount Morrison pendant, but of a different facies. The Mount Morrison Ordovician strata belong to the western clastic facies of the Great Basin, whereas the Ordovician strata of the Inyo Mountains belong to the eastern carbonate facies (Kirk, in Knopf, 1918, p. 32–33). The Bishop Creek pendant is in a proper position for strata transitional between the two facies.

GNEISS IN THE CANYON OF THE SOUTH FORK OF BISHOP CREEK

A lenticular mass of gneiss 3 miles long and as much as three-quarters of a mile wide crops out in the canyon of the South Fork of Bishop Creek between Long Lake and Bishop Pass. The trail from South Lake to Kings Canyon passes through the gneiss for several miles. At the north end the gneiss is bordered on the east side by the metasedimentary rocks of Chocolate Peak, but farther south it forms a septum between the Lamarck granodiorite on the west and the Inconsolable granodiorite on the east. Foliation strikes generally N. 30° W., parallel with the long axis of the mass; the dip is vertical or nearly so.

The appearance of the gneiss in the field and the mineral content and fabric suggest that it is a strongly sheared granitic rock. The gneiss is variable both in color and in composition. The most common rock is medium gray; less common rocks are medium light gray or light gray. The common dark-gray gneiss consists of 35 percent plagioclase (oligoclase), 30 percent microcline, 20 percent quartz, 7 percent biotite, 6 percent hornblende, and 2 percent accessory minerals (magnetite, ilmenite, sphene, and apatite). A light-gray felsic type contains 45 percent microcline, 40 percent quartz, only 10 percent plagioclase (oligoclase), 5 percent biotite, plus a little apatite, sphene, and pyrite.

The foliation of the gneiss results chiefly from alternating felsic and mafic lenses an eighth of an inch or less in thickness. The lenses are subparallel and discontinuous; the trace of a single lens is seldom identifiable for more than an inch. The mafic lenses consist chiefly of biotite, hornblende, and magnetite, and the felsic ones of quartz and the feldspars.

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Biotite and hornblende are strongly oriented in the foliation plane, whereas the felsic minerals form a granoblastic intergrowth. Most of the felsic minerals are nearly equidimensional, but some quartz grains are flattened in the foliation plane. The average grain size of the felsic minerals differs materially in adjacent lenses or streaks, ranging from a tenth to half a millimeter in finer lenses to half a millimeter to a millimeter in coarser lenses. The finer grained lenses are also the dark-colored lenses and contain most of the biotite and hornblende.

A concordant lens of very light gray rock composed chiefly of quartz and microcline in granoblastic intergrowth and more than a mile long and 500 feet wide is enclosed in the southern part of the gneiss. Faint streaks of biotite and sericite are also present, and coincide with finer grained streaks within the rock as do the mafic minerals in the darker gneiss.

MARBLE AND CALC-HORNFELS IN THE NORTHEAST BASE OF MOUNT TOM

Two remnants of marble and calc-hornfels at the east base of Mount Tom are bounded by quartz monzonite and, on the northeast, by alluvial fan deposits. The eastern part of the larger and more northerly remnant contains noncalcareous hornfels and schist as well as marble and calc-hornfels, and all diminish in abundance toward the west. If the rocks are not isoclinally folded, the maximum stratigraphic thickness is about 600 feet.

The smaller remnant, about half a mile south of the larger one, is largely calc-hornfels, locally altered to talc, and the south end has been prospected for talc.

METAMORPHIC ROCKS IN THE TUNGSTEN HILLS

In the Tungsten Hills, the largest masses of metamorphic rocks are the Round Valley septum and a somewhat smaller septum about 2 miles farther east. Both septa lie along a contact between alaskite of Cathedral Peak type and Tungsten Hills quartz monzonite. Smaller inclusions are present in the Deep Canyon area in the east-central part of the Tungsten Hills, and many small inclusions of meta-andesite are present in the southwest part of the Tungsten Hills.

ROUND VALLEY SEPTUM

The metamorphic rocks in the Round Valley septum are all metasedimentary; they include micaceous quartzite, impure argillaceous marble, calc-hornfels, quartz-sericite hornfels, and tactite. The strata in the main part of the septum strike northward, across the long axis, and dip 55° to 70° to the west. The rocks in the east end of the septum are in fault contact with the main part of the septum; they strike eastward and dip steeply south. The predominant rock in the eastern part of the septum is brown-weathering micaceous quartzite, but a thin layer of marble lies along the north side of the micaceous quartzite and extends westward along the contact between the two granitic rocks that bound the septum. Exposures in the Little Shot (Tungsten Hill) mine indicate that the width of the marble increases at depth.

In fault contact with the micaceous quartzite and marble on the west is coarse-grained garnetiferous calc-hornfels which is the predominant rock in the central part of the septum. This rock is mottled pale red and grayish green. Skeletal garnet crystals as much as 1 cm across and idocrase crystals as much as 4 mm across give the rock its coarse granularity. Plagioclase and clinozoisite form finer grained aggregates. All these minerals enclose abundant tiny grains of diopside.

In the vicinity of the Round Valley mine, the garnetiferous calc-hornfels encloses impure argillaceous marble, which at the Round Valley mine grades through a zone of light-colored hornfels into tactite. The hornfels is layered, and the different layers are composed of varying proportions of plagioclase, microcline, and diopsidic pyroxene, with or without garnet, clinozoisite, sphene, calcite, quartz, chlorite, altered magnetite, and muscovite or sericite. The field relations suggest that the argillaceous marble is the parent rock of both the calc-hornfels and the tactite.

West of the garnet-diopside-idocrase-plagioclase-clinozoisite rock is a mass of quartz-sericite hornfels that contains two layers of garnetiferous tactite, which are the loci of the ore bodies in the Western Tungsten mine. The tactite is a layered rock that consists of alternate layers of garnet-epidote rock and diopside-rich rock.

If the section west from the fault in the east end of the pendant is homoclinal, the thickness of the exposed section is about 2,500 feet, of which 1,500 feet is tactite and related argillaceous marble and calc-hornfels and 1,000 feet is quartz-sericite hornfels, including the intercalated tactite layers. Detailed mapping in the Western Tungsten mine area, however, suggests that the quartz-sericite hornfels may be isoclinally folded and that the true exposed thickness of strata in this unit may be no more than 500 to 700 feet.

SEPTUM 2 MILES EAST OF THE ROUND VALLEY SEPTUM

The septum 2 miles east of the Round Valley septum, at hill 5949, is composed chiefly of quartz-sericite hornfels and micaceous quartzite, but the septum also contains lesser amounts of marble and calc-hornfels. Because the rocks are poorly exposed and complexly deformed, their distribution on the geologic map (pl. 2) is generalized.

METAMORPHIC INCLUSIONS IN THE DEEP CANYON AREA

Small inclusions of metasedimentary rocks are abundant in the hornblende gabbro and quartz diorite of the Deep Canyon area and in the surrounding quartz monzonite. The most common rocks in these inclusions are vitreous quartzite, marble, calc-hornfels, and tactite. In the vicinity of the Aeroplane and Lucky Strike mines metaconglomerate accompanies the quartzite, and at the Aeroplane mine a unit composed of micaceous quartzite and biotite-plagioclase-microcline hornfels is found.

The vitreous quartzite generally is clean, but contains minor interstitial feldspar and sericite. The conglomerate which accompanies the quartzite is composed chiefly of pebbles of diopside-bearing quartzite and zoisite-clinozoisite rock in a matrix of tremolite, diopside, quartz, and clinozoisite. The common calc-hornfels is a dense massive fine-grained rock composed chiefly of calcic plagioclase (An_{80-90}) and lesser amounts of pyroxene and sphene. At the Aeroplane mine, coarse-grained calc-silicate rock is composed chiefly of equal amounts of grayish-pink garnet and light olive-gray idocrase, and contains minor amounts of pyroxene, clinozoisite, and plagioclase. This rock might be classified as a tactite, but would be an uncommon variety.

The micaceous quartzite and biotite-plagioclase-microline hornfels at the Aeroplane mine are medium dark-gray fine-grained rocks. A typical specimen of the micaceous quartzite is composed of 50 percent quartz in rounded grains, 25 percent muscovite, and 20 percent chlorite that appears to have been formed from biotite, and minor amounts of biotite, magnetite, and apatite. The biotite-plagioclase-microcline hornfels consists of approximately 40 percent microcline, 30 percent plagioclase, and 20 percent well-oriented biotite that gives the rock a faint cleavage, plus a little muscovite, pyrite, apatite, and sphene.

MAFIC METAVOLCANIC ROCK IN THE SOUTHWESTERN PART OF THE TUNGSTEN HILLS

A small mass of meta-andesite in the southwestern part of the Tungsten Hills, in the SW¼ sec. 15, T. 7 S., R. 31 E., is visible from the unpaved road to the Buttermilk Country. It consists in part of volcanic breccia and in part of massive dark-gray rock containing abundant corroded relict phenocrysts of plagioclase as much as 2 mm across. Both rocks contain about 65 percent plagioclase in relict phenocrysts and in a granoblastic groundmass, variable proportions of biotite and hornblende, plus minor quartz, magnetite, and apatite.

SEPTA 3 MILES WEST OF KEOUGH HOT SPRINGS

Three miles west of Keough Hot Springs are two disconnected metamorphic remnants. The northern remnant, or septum, lies along a contact between alaskite similar to the Cathedral Peak granite and Tungsten Hills quartz monzonite, and the southern remnant extends from the contact between the plutons into the alaskite. The northern remnant is more than 4,000 feet long and about 2,000 feet wide. It is composed chiefly of dark meta-andesite, but contains conglomerate, marble, and calc-hornfels in its northern part. The meta-andesite adjacent to the metasedimentary rocks is amygdaloidal and is mottled and veined with light yellowish-gray, diopside-plagioclase hornfels.

The southern remnant is also about 4,000 feet long in a northwesterly direction, but is only about 800 feet wide. It is composed chiefly of metasedimentary strata in the northern part and of metavolcanic strata in the southern part. Beds strike northward, diagonally across the remnant, and dip westward. Dips are to the west, 55°–70° in the southern part and 15°–20° in the northern part. If the structure is homoclinal, as it appears to be, the stratigraphic thickness is in excess of 2,000 feet.

The metasedimentary rocks in the northern part of the southern remnant are chiefly highly feldspathic quartzites. Light- and dark-colored beds an inch or two thick alternate. In thin sections the rock can be seen to consist chiefly of poorly sorted angular grains of microcline and quartz in a granoblastic groundmass of the same minerals. The light-colored layers contain a higher percentage of larger grains than the darker ones and also contain moderate amounts of epidote. The darker layers contain biotite, amphibole, and magnetite.

The volcanic rocks in the southern part of the remnant generally are felsic and may be tuffs. Plates and aggregates of biotite and hornblende and knots of epidote are set in a granoblastic matrix of microcline and quartz that locally includes biotite and magnetite. Scattered porphyroblasts of andalusite are generally present. Associated with these rocks are layers of micaceous hornfels, which may have been derived from pelitic sedimentary rocks.

REMNANTS ALONG THE RANGE FRONT SOUTHWEST OF BISHOP

Along the east-trending segment of range front southwest of Bishop and extending west into the range are several discontinuous remnants of metasedimentary rocks. The easternmost one is a small inclusion at the Rossi mine, and the westernmost is an inclusion at the Chipmunk mine. The metasedimentary rocks include marble, tactite, calc-hornfels, and carbonaceous feldspar-quartz hornfels. The masses are so discontinuous that the stratigraphic sequence is unknown.

Marble and calc-hornfels make up about half of the metamorphic rocks, and the remainder is dark carbonaceous hornfels. Commonly the marble is coarsely crystalline and white, but at the Rossi mine the marble is light bluish gray. Much of the marble is moderately clean, but in places it is thinly interlayered with calc-hornfels, and in places calc-hornfels predominates. The largest lens composed predominantly of calc-hornfels is half a mile west of the Bishop Antimony mine.

A specimen of dark carbonaceous feldspar-quartz hornfels from the vicinity of the Bishop Antimony mine, which is typical of the dark carbonaceous rock in this area, consists of a fine granoblastic intergrowth of quartz and K feldspar, and contains abundant disseminated carbonaceous material and minor sphene, tremolite, and sericite. Thin light-colored layers intercalated with the dark carbonaceous hornfels have a a similar texture and mineral content, but contain large skeletal or poikiloblastic crystals of diopside and are almost devoid of carbonaceous material. A dark rock, slightly different from the inclusion at the Rossi mine, contains abundant grayish-red biotite and megascopically appears very similar to the micaceous quartzite in the Pine Creek pendant.

A rock unit at the Chipmunk mine consists of alternating very light gray wollastonite-rich layers and dark-gray quartz-rich layers. The wollastonite-rich layers contain small grains of diopside, which vary in amount from layer to layer. Tiny quartz grains are abundant in the margins of the wollastonite layers, and larger quartz masses locally are associated with pyrite in the central parts of the layers. The dark layers contain, in addition to quartz, plagioclase and K feldspar, plus minor diopside, sphene, calcite, and pyrite.

BIG PINE SEPTUM

Several inliers of metasedimentary rocks crop out through the moraines that lie between Big Pine and Baker Creeks. These outcrops are parts of a large buried septum, here designated the Big Pine septum, which extends west from the range front continuously for about 4 miles and appears to have a maximum width of about 1 mile. The septum lies between quartz monzonite on the north and granodiorite on the south. The rocks in the inliers—marble, tactite, calc-hornfels, micaceous quartzite, and micaceous hornfels—are probably representative of the rocks in the pendant because the inliers are chiefly up-faulted blocks rather than erosional remnants.

The most abundant rock in the Big Pine septum is coarsely crystalline, locally silicated, white to medium-gray marble. The marble commonly contains a few percent quartz, biotite, actinolite, and wollastonite. In a few places the marble next to intrusive granitic rocks has been converted to tactite. Calc-hornfels is present in only two outcrops. A small inclusion in the Tinemaha granodiorite on the south side of the septum consists of rock that is composed of approximately 70 percent of talc, 20 percent fibrous actinolite, and 10 percent magnetite.

Dark-gray micaceous quartzite and quartz-mica hornfels are only slightly less abundant than marble. The micaceous quartzite generally consists of silt-size detrital quartz grains disseminated through a fine-grained matrix of sericite and a little quartz. Large poikiloblastic plates of biotite, locally altered to chlorite, are common, and a few percent magnetite generally is present. Spotted quartz-mica hornfels is a common rock in the westernmost inlier. The groundmass of this rock consists of a fine intergrowth of approximately equal amounts of quartz, biotite, and sericite, and a little magnetite. The spots are dark gray in hand specimen and consist of quartz-sericite intergrowths that are much finer than the groundmass of the rock. A thin marginal rim of biotite causes the dark color of the spots in hand specimen. Another rock in the westernmost inlier is composed of about 50 percent plagioclase (oligoclase) and 25 percent each biotite and sericite.

All these rocks appear to have been formed from miogeosynclinal sediments—limestone, argillaceous siltstone, calcareous shale. However, the fact that a specimen from the south side of Big Pine Creek consists of large corroded crystals of quartz and oligoclase in a fine sericitic matrix suggests derivation from an acid crystal tuff.

SEPTA AND INCLUSIONS IN UPPER BIG PINE CANYON

In the upper part of Big Pine canyon, west of the Blue Star talc mine, are many septa and inclusions that are composed chiefly of mafic metavolcanic or hypabyssal rocks. The more easterly of these also contain marble and calc-hornfels. The most interesting septum is the one that contains the Blue Star talc mine (see p. 33). Most of the inclusions west of Glacier Lodge, in the North Fork of Big Pine Creek, are predominantly of fine-grained metavolcanic or hypabyssal rock; in the South Fork of Big Pine Creek diorite predominates.

The metavolcanic or hypabyssal rocks have been recrystallized and corroded along their margins by the enclosing granitic rocks. Notable variation of the mineral content of metavolcanic rocks at contacts with granitic rocks suggests that they have also been altered metasomatically. The plagioclase content generally ranges between 40 and 60 percent, but some specimens contain no plagioclase and others contain as much as

70 percent. The hornblende content is exceedingly variable, but most rocks contain more than 45 percent. Specimens from inclusions of metadiorite in the North Fork of Big Pine Creek which have been corroded and embayed by granite, generally contain less than 15 percent hornblende, 10 to 25 percent biotite, 5 to 10 percent K feldspar, and 2 to 20 percent quartz. Locally, these highly altered rocks contain relict quartz xenocrysts with coronas of hornblende (Muir, 1953 b, p. 409–428). The common minor accessory minerals are magnetite, apatite, and sphene. Epidote or clinozoisite has been formed from hornblende and plagioclase, chlorite from biotite, and hematite from magnetite.

A large inclusion of mafic igneous rock at the east base of Mount Alice, at the junction of the North and South Forks of Big Pine Creek, consists of fine-grained mafic rock in association with coarser grained rock of variable texture, some of which is pegmatitic. The coarser grained rocks could conceivably have resulted from original crystallization as hypabyssal intrusives, but seem more likely to have resulted from the metamorphism of fine-grained rocks.

MIDDLE PALISADE SEPTUM

The Middle Palisade septum is a thin belt of calcareous rocks that extends south from the Middle Palisade Glacier for about 3 miles. The septum lies between the Tinemaha granodiorite on the east and the Inconsolable and Lamarck granodiorities on the west. About 2 miles farther south along the same trend, and half a mile southwest of the summit of Split Mountain, is an inclusion composed of similar calcareous rocks.

The calcareous rocks include marble, calc-hornfels, and tactite; all the rocks were formed from limestone and from impure limy strata. Most of the marble is white and coarsely crystalline, and contains scattered blebs of silicate minerals. Much of the calc-hornfels contains plagioclase and pale-green pyroxene as principal constituents. The mineral content of the tactite is variable; some consists predominantly of grayish-green pyroxene and quartz and a little light-olive epidote, and some consists chiefly of reddish-brown garnet. A little scheelite occurs locally in the garnetiferous varieties.

SPLIT MOUNTAIN SEPTUM

The Split Mountain septum is a thin lens of schist and calc-hornfels that extends from the range front westward along the north side of Red Mountain Creek to the east face of Split Mountain, then southward to the edge of the mapped area. The exposed length within the mapped area is about 5½ miles, and the outcrop width is from a few feet to more than a thousand feet. The septum separates quartz monzonite similar to the Cathedral Peak granite from Tinemaha granodiorite on the north and from Lamarck granodiorite on the west. The segment north of Red Mountain Creek dips steeply, whereas the south-trending segment steepens westward from almost flat to vertical (pl. 5, section *F–F′*). Several small pods of schist and gneiss that crop out in Stecker Flat are along a gently dipping contact between quartz monzonite similar to the Cathedral Peak and Tinemaha granodiorite, and are probably also part of the septum (pl. 4). The distribution and attitudes of the metamorphic rocks suggest that the septum is marginal to the quartz monzonite of Cathedral Peak type and that before erosion it discontinuously overlay the quartz monzonite along a gently east dipping upper contact (pl. 5, section *F–F′*).

The most common rocks in the septum are schist and calc-hornfels, but pelitic hornfels and gneiss of the same mineral composition as the schist are present in the remnants in Stecker Flat.

Locally, however, the septum contains calc-hornfels. In one place, fine-grained pyroxene-plagioclase hornfels contains irregularly shaped coarse-grained zones as much as 2 inches across composed of large hornblende and quartz crystals which have poikiloblastic margins. In hand specimen the coarse-grained rock superficially resembles diorite, but examination with the microscope reveals that plagioclase is present only in tiny grains that are enclosed in the sievelike margins of the large hornblende and quartz crystals.

MARBLE IN THE POVERTY HILLS

In the west side of the Poverty Hills, a mass of marble having an outcrop area of about 1½ square miles is embraced on the north and east sides by the Tinemaha granodiorite. On the west side the marble is bounded by alluvium and on the southwest side by basaltic lava. The marble is white and medium grained, and appears to be clean except in the south end where thin beds of calc-hornfels are interlayered with marble. No clastic beds were observed within the marble.

GEOLOGY OF THE BATHOLITH

The Sierra Nevada and the Sierra Nevada batholith are two different entities and are not to be confused. The Sierra Nevada is a physiographic feature—a mountain range—whereas the Sierra Nevada batholith is the plutonic terrane that makes up the central and eastern parts of the range. The metamorphic remnants between and enclosed in the plutonic rocks are not part of the batholith, although they are within it. The batholith is composite; it is a mosaic of discrete intrusives in sharp contact with one another or separated by thin septa of metamorphic rock or by late aplitic dikes that follow contacts. The individual intrusive masses are called plutons. The largest plutons are composed of felsic quartz-bearing granitic rocks ranging in composition from granodiorite to alaskite. Plutons of granitic rock comprise more than 90 percent of the batholith within the area mapped for this report (fig. 13). The rest of the batholith is composed of smaller scattered bodies of older and darker rocks ranging in composition from hornblende gabbro to granodiorite. For convenience and because the relation of the older and darker rocks to the granitic rocks is uncertain, the two groups of rocks are described separately.

DIORITE, QUARTZ DIORITE, AND HORNBLENDE GABBRO

The small bodies of diorite, quartz diorite, and hornblende gabbro have been aptly referred to by Mayo (1941, p. 1010) as "basic forerunners," or simply as "forerunners." Their distribution is reminiscent of the metamorphic rocks; they occur as inclusions or small pendants within individual plutons of more silicic rock, or as septa between plutons. Commonly they are associated with metamorphic rocks, and many are crowded with metamorphic inclusions. This relation is understandable because diorite, quartz diorite, and hornblende gabbro were the first of the plutonic rocks to be emplaced and they came into contact with metamorphic rocks on all sides. The original sizes and shapes of most masses were destroyed by later granitic intrusives, which tore apart the masses and recrystallized, granitized, and assimilated fragments. In this report the term "assimilation" is used to describe the incorporation of solid rock in magma and "granitization" to describe conversion, in essentially the solid state, of nongranitic rock to granitic rock.

Partly as a result of original differences and partly as a result of subsequent modification, the rocks grouped under diorite, quartz diorite, and hornblende gabbro are heterogeneous in composition and texture; very likely they include rocks of diverse origin. Although they are discussed separately they are not delineated separately on the maps (pls. 1–4). Quartz diorite is used here for quartz-bearing plutonic rock in which quartz comprises more than 10 percent of the total felsic constituents and K feldspar less than 10 percent (fig. 14). It passes into granodiorite where K feldspar comprises more than 10 percent of the total felsic constituents and into diorite or gabbro where quartz comprises less than 10 percent. The plagioclase of diorite contains less than 50 percent anorthite, and the plagioclase of gabbro more than 50 percent. Hornblende gabbro contains hornblende rather than augite, the usual principal mafic mineral of gabbro.

ALTERED DIORITE OF THE WHITE MOUNTAINS

The altered diorite of the White Mountains is equigranular and has an average grain size of about 1 mm. The rock is of much the same appearance in the stock and in dikes. In thin section it can be seen to be highly altered; it consists chiefly of unzoned albite (An_{10}, approximately), chlorite, and epidote-clinozoisite. Quartz in small amounts is interstitial to most of the other minerals; sericite is disseminated through the albite; and sparse magnetite and hematite are present. The albite crystals have irregular boundaries in detail, but are generally euhedral in gross form—a feature that gives the rock a panidiomorphic-granular texture. Undoubtedly the original feldspar was more calcic than albite; sparce unaltered parts of grains indicate it was in the andesine range. The form of much of the chlorite and epidote suggests that these minerals were derived, at least in part, from original biotite and hornblende, respectively, although some epidote-clinozoisite must have been formed from plagioclase as a by-product of its alteration to albite.

The intrusive relations of the stock and dikes indicate that the diorite crystallized from a magma. Nowhere was the diorite found in contact with other plutonic rocks; its age relation to the intrusive rocks of the Sierra Nevada is therefore not known. No evidence was found to indicate that the diorite was involved in the deformation that affected the enclosing rocks. Very likely it is of about the same age as the masses of quartz diorite and hornblende gabbro of the Sierra Nevada, but it may be genetically related to and of the same age as swarms of mafic dikes that are younger than some of the granitic plutons.

HORNBLENDE GABBRO

The quartz diorite and hornblende gabbro of the Sierra Nevada are extremely variable in color index, grain size, texture, and proportions of constituent minerals. The darker appearing rocks (color index 40–60) generally contain plagioclase having more than 50 percent anorthite and contain hornblende or uralitic amphibole rather than augite, and are therefore classed as hornblende gabbro. Most of the hornblende gabbro ranges from 1 to 5 mm in average grain size and thus is medium grained, but these limits are so wide as to permit very great differences in the appearance of different rocks.

In addition to hornblende and uralitic amphibole and calcic plagioclase, the hornblende gabbros generally contain magnetite, apatite, and sphene. Secondary

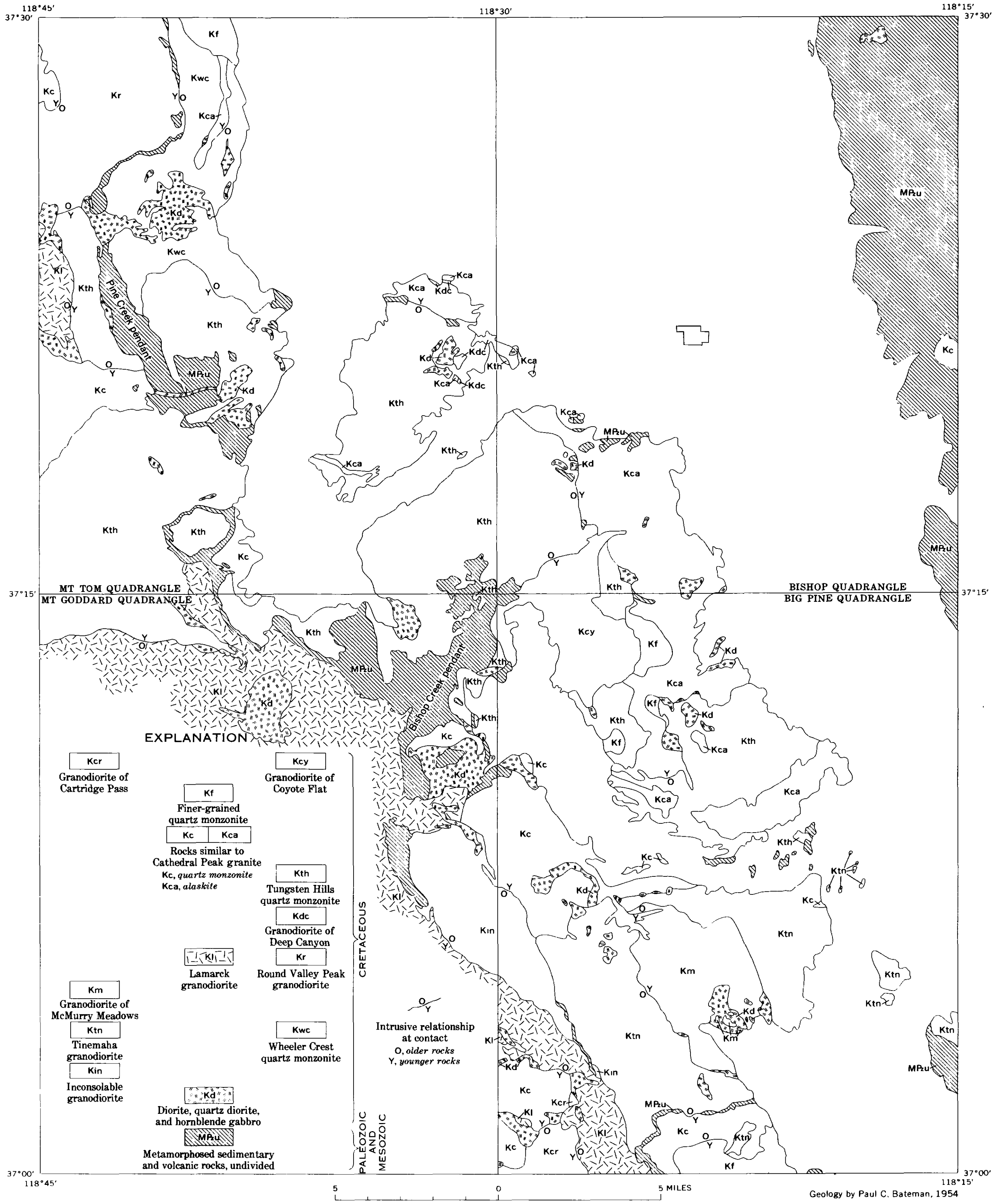


FIGURE 13.—Bedrock map of the Bishop area showing the distribution of the plutonic rocks.

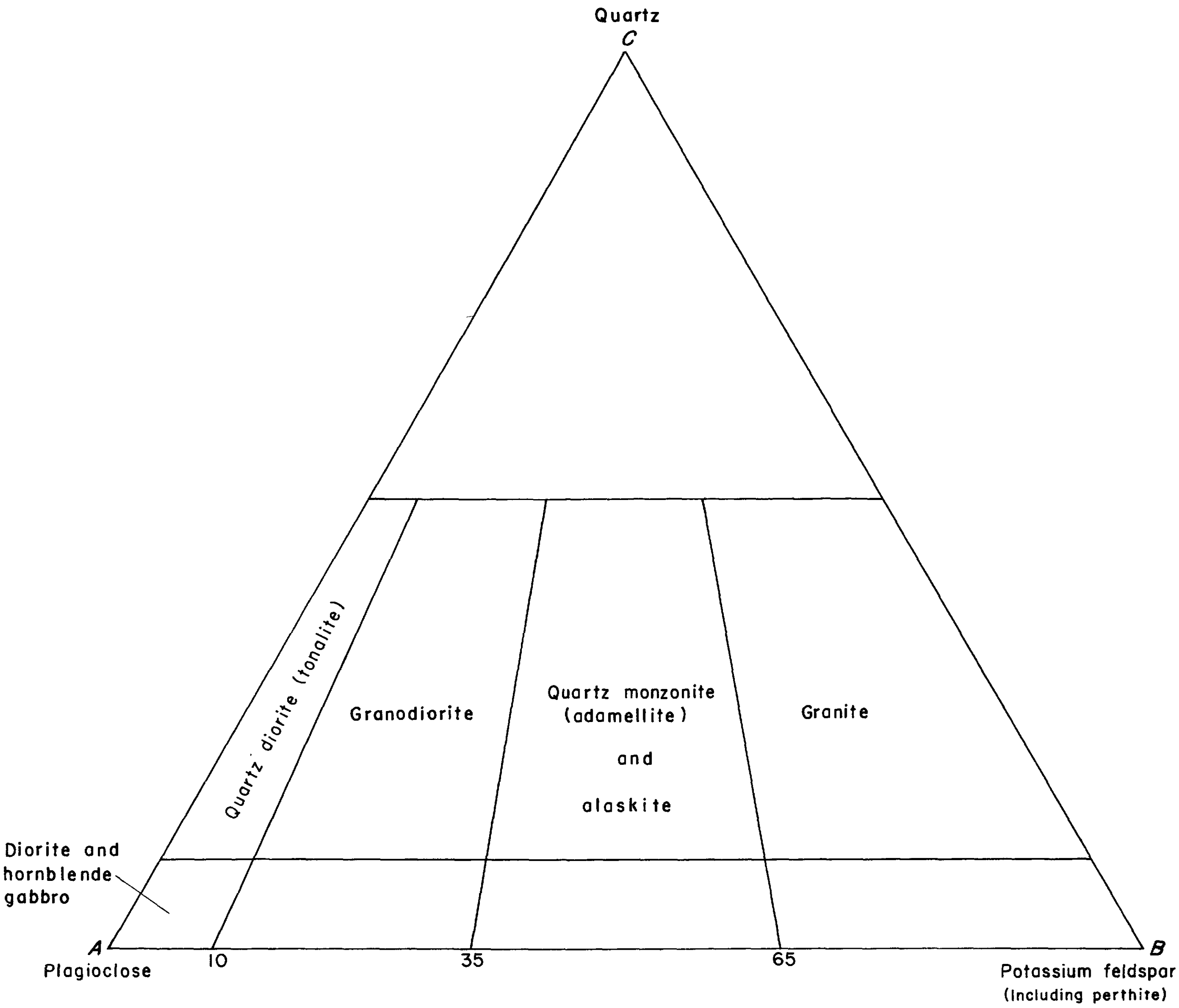


FIGURE 14.—Triangular diagram showing the classification system used in this report. Classification is modal. The plagioclase of quartz monzonite contains more than 10 percent anorthite, and the plagioclase of alaskite less than 10 percent.

minerals include epidote, chlorite, sericite, and scant serpentine group minerals. Plagioclase generally is in almost euhedral crystals that give the rock a panidiomorphic-granular texture. Commonly these crystals are strongly and progressively zoned from bytownite cores to calcic oligoclase or sodic andesine rims; discontinuities and reversals in zoning are common. Hornblende generally is anhedral, but in many rocks is in euhedral prisms having a wide range of proportions of length to thickness. Uralitic amphibole is colorless or mottled colorless and pale green and is rimmed by pale-greenish amphibole that was probably formed at the time of uralitization by reaction with feldspar. Blades of pale-greenish amphibole are also usually present scattered through the adjacent rock. Uralitic amphibole locally contains residual cores of augite, and some hornblende encloses augite or uralitic amphibole.

Many of the rocks exhibit unusual and interesting textures. Locally coarse-grained, almost pegmatitic rock is found, which contains euhedral prisms of hornblende 1 inch or more long and ¼ to ½ an inch across. In several places, notably west of McMurry Meadows in the Big Pine quadrangle, hornblende gabbro contains almost equidimensional crystals of plagioclase. An interesting fabric is shown by elongate hornblende crystals that lie in a well-defined foliation plane, but which are randomly oriented within that plane. Rock of this fabric generally occurs in thin tabular masses that are bordered by younger silicic granitic rock—a spacial arrangement that suggests a metamorphic rather than an igneous origin. However, the fact that many masses of hornblende gabbro intrude metamorphic rocks and exhibit panidiomorphic-granular texture, deep zoning of plagioclase, layered facies, and dikes that cut metamorphic rocks indicates that they crystallized from magma and are truly igneous.

LAYERED GABBRO IN THE TUNGSTEN HILLS

A mass of gabbro in the Tungsten Hills locally contains layered facies that merit special attention. The gabbro crops out over an area of about a quarter of a square mile and is entirely surrounded by later quartz monzonite. Patches of layered rock are exposed in a low knoll about 500 feet north of the Tungsten Blue (Shamrock) mine. The largest patch has an area of only about a hundred square feet, and most of the others have outcrop areas of only a few square feet. The layered patches exhibit "angular unconformities," "crossbedding," and penecontemporaneous faults, which are strikingly similar to structures found in sedimentary rocks. Figure 15 is a large-scale map of the largest patch of layered gabbro, and figure 16 consists of photographs of several interesting exposures. In figure 15, the largest mass shown is in place; the two outlying masses are slightly displaced from their original positions.

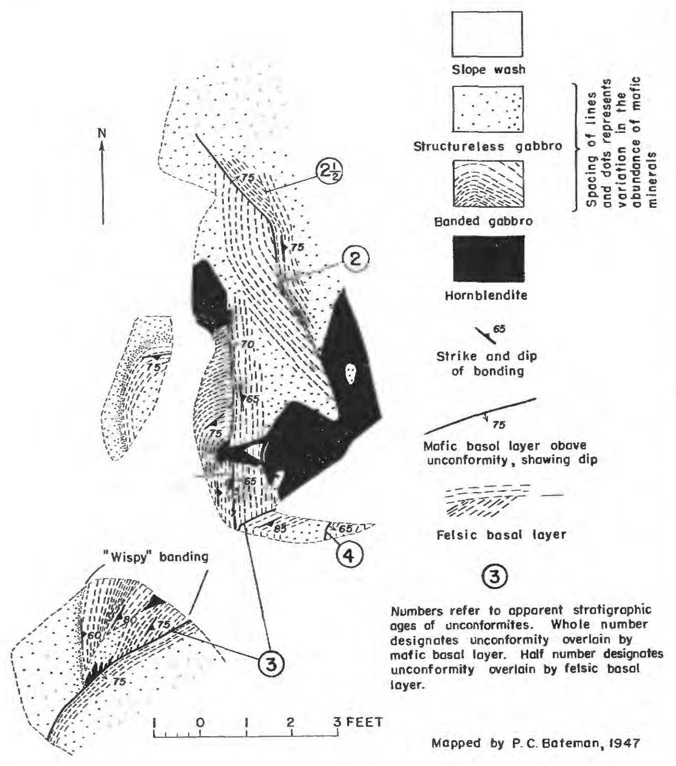


FIGURE 15.—Map of the main patch of layered gabbro in the Tungsten Hills.

In general, the more conspicuous "angular unconformities" are marked by dark basal layers. Other generally less distinct "unconformities" are overlain by felsic basal layers. In figure 15, four "unconformities" having dark basal layers and one having a felsic basal layer interrupt the layering. All the "unconformities" are numbered in the order of their apparent ages according to sedimentary criteria.

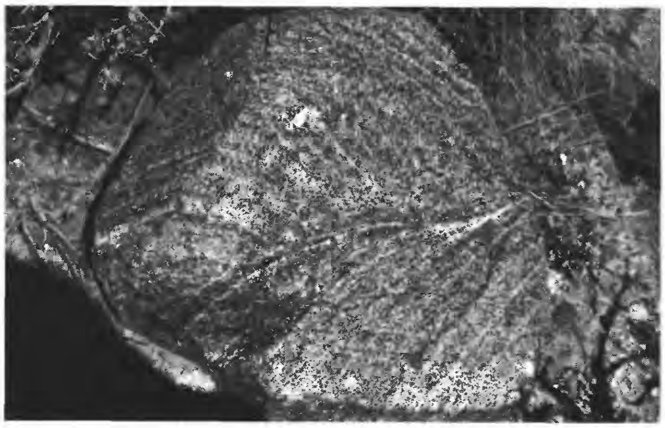
The layered sequence beneath some "unconformities," especially those overlain by mafic basal layers, are curved (see layers beneath "unconformities" 1 and 2 of figure 15). In one small mass, the ends of layers truncated by an "unconformity" bend sharply. In a few places, the layers beneath "unconformities" are not curved at all, as beneath "unconformity" 2½, figure 15. The southern part of the dark basal layer at "unconformity" 1 is broken into several segments, which are partly rotated and displaced as much as an inch from their original positions. The slip planes bounding these segments are not very evident, only the layers adjacent to the "unconformity" having been displaced.

The layering is caused by the rhythmic alternation of light- and dark-colored layers. Most layers are parallel, but a few are wispy or "crossbedded." The layers dip 65° to 75° to the east and southeast, away from the nearest contact of the gabbro with quartz monzonite, which is about 300 feet distant. At the borders of the layered areas the layers become less distinct, and the layered rock merges imperceptibly with structureless gabbro. In most places, individual layers can be traced for several feet with little change in thickness or in relative abundance of light- and dark-colored minerals. Discontinuous wispy layering, however, is displayed in some places, for example in the isolated boulder shown in the lower part of figure 15. In figure 16*C*, the mafic layers above the "unconformity" lens out where they overlap onto the "unconformity."

The combined thickness of a dark layer and of the overlying light-colored layer averages about half an inch, but pairs of layers range from ¼ to 1 inch in thickness. Immediately above the "unconformities" that are marked by mafic basal layers the layers are thin, but upward they gradually increase in width until they attain average thickness an inch or two above the basal layers. By contrast, the layers just above "unconformities" with felsic basal layers are of average width.

Faults offset the layers less than an inch, but they contain dark minerals that make them easily visible. The dark hornblendite shown in the central part of figure 15 dips to the southeast and cuts across the layers to the southwest. From near the southwest end of

A



C

B

FIGURE 16.—Photographs of several exposures of layered gabbro. *A*, Layering and "unconformities" 1 and 3 (fig. 15) from the south. *B*, Layering and "unconformities" 2 and 2½ from the west. *C*, Boulder showing layering and an "unconformity" with a felsic basal layer.

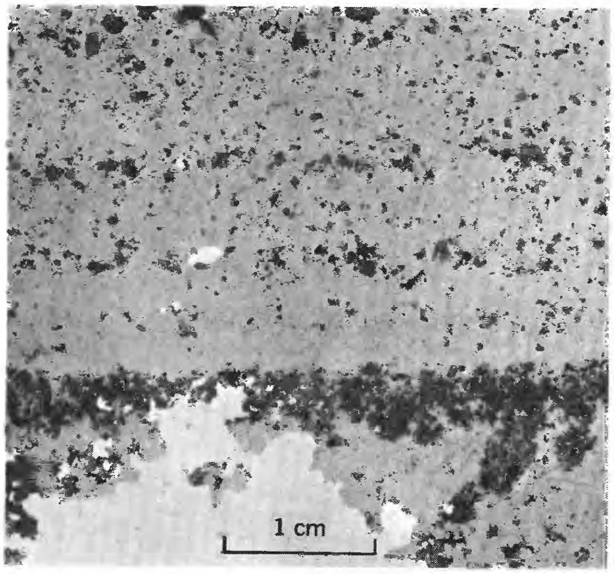
the hornblendite, a branch cuts "stratigraphically" downward and across "unconformity" 1 into the underlying layers. The hornblendite follows along the layering, and becomes less distinguishable and loses its identity a foot beneath the "unconformity."

Microscopic study included the examination of several extra-large (2 by 3 inches) oriented thin sections. These sections were studied both under a petrographic microscope and between polaroid filters either without magnification or under a low-power binocular microscope. The polaroid filters made it possible to see in a single field the structures and textures over the full area of a thin section.

The light- and dark-colored layers contain the same minerals, but in different proportions. The felsic layers consist mainly of calcic plagioclase (bytownite), whereas grains of uralitic amphibole, some having residual augite cores, are abundant in the dark-colored layers. Hornblende, magnetite, apatite, sphene, and locally biotite are common, especially in the dark-colored layers. Epidote and clinozoisite are uncommon derivatives of plagioclase, while chlorite sporadically replaces augite, the amphiboles, or biotite. Antigorite pseudomorphous after pyroxene is rare.

The bytownite generally is in zoned euhedral to subhedral crystals. Compositions range from An_{90} to An_{65}, but the compositions of many large areas of zoned crystals and of most unzoned crystals fall between An_{75} and An_{85}. Uralitic amphibole, in some places enclosing residual cores of augite, is ragged, but the gross outlines of many crystals resemble augite. The uralitic amphibole is weakly pleochroic and contains numerous small grains of exsolved magnetite. Commonly the pale uralite is bordered by more deeply colored rims of amphibole, pleochroic in blue green. Small needles having approximately the same optical properties as the amphibole of these rims occur within and between the surrounding plagioclase crystals. Hornblende is in anhedral grains that are pleochroic in dark olive green, brownish green, and brown. Locally, brownish-green hornblende encloses augite or uralite.

Plagioclase commonly is in euhedral to subhedral crystals tabular parallel to (010). Under the micro-



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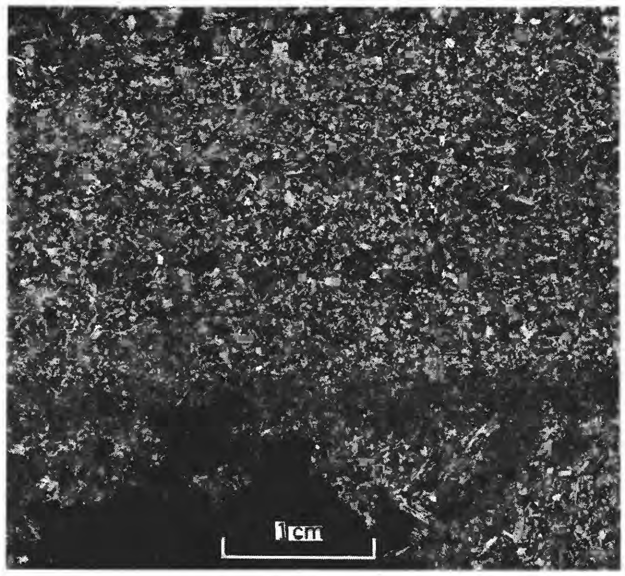

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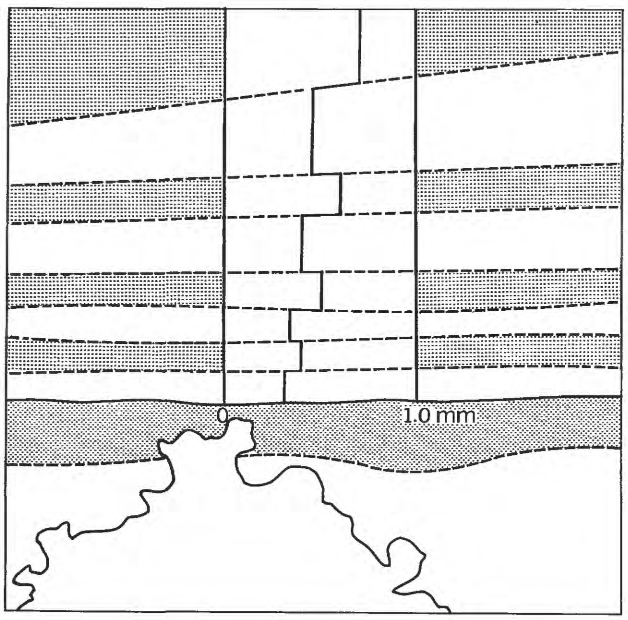


B

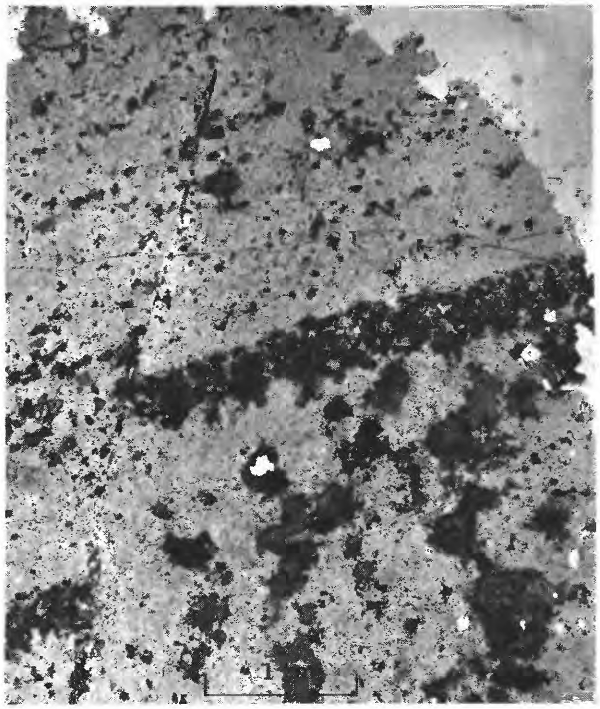
FIGURE 17.—Thin section cut across the layering and an "unconformity" in layered gabbro. *A*, Ordinary light. *B*, Crossed nicols. *C*, Sketch of thin section with superimposed graph showing (in millimeters) the average apparent length in thin section of plagioclase crystals in the layers above the mafic basal layer.

scope, consequently, the crystals showing albite twinning appear elongate, and the crystals not showing twinning appear equidimensional. Individual crystals are small, the average measurement on (010) being less than 1 mm. The other minerals in the gabbro are of comparable size.

Crystals are interlocking across contacts between felsic and mafic layers just as they are in the centers of layers. A few crystals are fractured, but evidence of widespread cataclasis or protoclasis is lacking. Locally, augite encloses plagioclase, but in general plagioclase, augite, and magnetite seem to have mutual interference boundaries with one another. Hornblende and biotite appear to be late in the sequence of magmatic crystallization. Uralitic amphibole probably formed from augite after crystallization of the gabbro, possibly at the time of intrusion of the adjacent quartz monzonite.

The (010) planes of the feldspar crystals, especially of the larger ones, have a marked tendency to lie subparallel to the layering, although crystals in all orientations are present (figs. 17, 18). Within the planes of layering the *a* and *c* crystallographic axes lie in any direction. Thin sections cut in the plane of the layering show the random orientation of crystals in this plane.

Two kinds of systematic variation in the size of the feldspar crystals were observed. The first, present throughout the layered rock, is between adjacent dark- and light-colored layers, the feldspar crystals in the dark layers being larger than those in contiguous light-colored layers (figs. 17, 18). The second kind of variation is limited to the thin layers immediately above "unconformities" overlain by dark basal layers. In these layers the average size of the feldspar crystals in both light- and dark-colored layers increases systematically with distance from the "unconformity." Graphs drawn to show the size variations in these layers with mafic basal layers above "unconformities" are saw-toothed because of the effect of the difference in size of the plagioclase crystals in the light- and dark-colored



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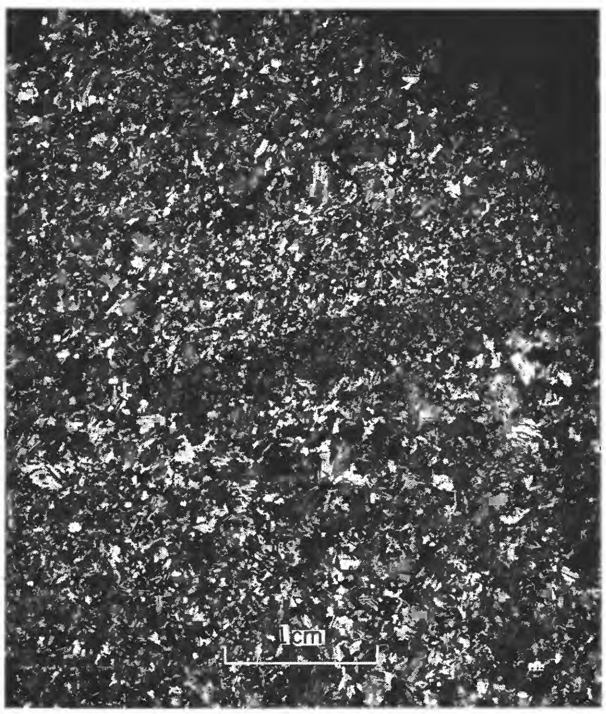

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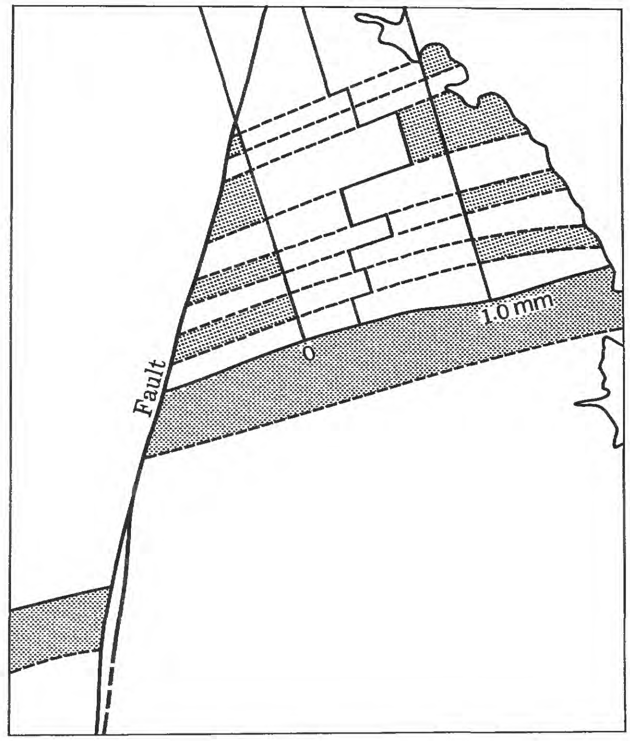

C

FIGURE 18.—Thin section cut across the layering in layered gabbro. *A*, Ordinary light; note fault offset and mafic minerals present along the fault. *B*, Crossed nicols. *C*, Sketch of part of thin section with superimposed graph showing (in millimeters) the average apparent length in thin section of plagioclase crystals in the layers above a mafic basal layer.

layers, but the graphs also show the overall increase in crystal size with distance from the "unconformity" (figs. 17, 18).

Many hypotheses have been advanced to explain the origin of layering in mafic, ultramafic, and alkaline igneous rocks. Grout (1918, p. 452–454), Coats (1936, p. 407–412), and Wager and Deer (1939, p. 284–289) have summarized the hypotheses advanced up to the times of their respective publications. Faults and folds and unconformities have been described from the Stillwater complex (Hess, 1938, p. 264–268; 1960, p. 129–131).

Any acceptable hypothesis of origin for the various features in the layered gabbro should explain the layering, the gradational relations between the layered patches and the unlayered gabbro, the faults, the cross-cutting mafic streaks, and the "unconformities," including those overlain by felsic basal layers as well as those overlain by mafic basal layers. The explanation of the layering should consider the larger size of the plagioclase crystals in any particular mafic layer as compared with those in contiguous felsic layers, the preferred orientation of the (010) faces of the plagioclase crystals in the plane of the layering, and the upward increase in the average size of the plagioclase crystals through the first few layers above dark-colored basal layers.

The mechanism that seems best to explain the layering is a process of gravity-controlled crystal settling in combination with magmatic flow. This is essentially the hypothesis advanced by Wager and Deer (1939, p. 271–275) to explain the layering in the Skaergaard intrusion. Hypotheses considered and rejected because they failed to explain one or more critical features of the gabbro are (1) other hypotheses based on sorting of continuously forming crystals in a magma chamber, (2) heterogeneous intrusion, (3) repeated sill-like intrusion, (4) rhythmic crystallization, (5) deformation during consolidation, (6) replacement or migmatitization, and (7) metamorphic differentiation.

The main feature that indicates gravity-controlled settling of crystals is the localization of the larger plagioclase crystals in the dark layers that contain heavier minerals. This size distribution of minerals is the kind found in clastic sedimentary rocks where the controlling factor is the relative settling velocities of particles from turbidity currents.

The density relations between plagioclase crystals and magma of intermediate to calcic composition have been a subject of considerable controversy; descriptions in the literature suggest that plagioclase is a little heavier than some magmas and a little lighter than others. The gabbro of the Tungsten Hills is not an extremely mafic rock; consequently, the magma may have been less dense than that of many gabbros. The fact that augite (largely altered to uralite) is interlayered with plagioclase indicates that if augite was heavier than the magma, plagioclase was also heavier. The association of larger plagioclase crystals with the heavier mafic constituents lends further strong support to this view.

Features most indicative of current action are (1) lateral variation in the content of mafic minerals within groups of layers, (2) the presence of wispy layering or "crossbedding," (3) the association of large plagioclase crystals with the mafic constituents, which indicates good sorting, and (4) overlaps such as are shown in figure 16*C*. Grout (1918, p. 453–454) has expressed the opinion that planar orientation of feldspars cannot be accomplished by direct settling of crystals alone and must be accompanied by some movement of the magma, but the validity of this conclusion seems doubtful unless almost every crystal is oriented in the plane of the layering; tabular and linear minerals in sediments are oriented preferentially parallel to bedding.

If continuous crystallization of all minerals is assumed, the layering can be explained by crystal sorting through rhythmic changes in the velocity of currents in the magma. During periods of relatively greater current velocity, heavier mafic minerals and larger feldspars were deposited. With lesser velocity, the smaller feldspars and a few small mafic crystals held in suspension were deposited.

The exceedingly small size of all the crystals in the dark basal layer present at most "unconformities" can be satisfactorily explained as an accumulation of newly formed crystals, the development of the "unconformity" having been accompanied by an influx of hotter magma that at first contained no suspended crystals. The first crystals to form and settle out were very small. Mafic minerals are more abundant in the basal layer because of their greater settling velocity. With time, larger percentages of the accumulating crystals came from greater distances and consequently had longer periods for growth. This hypothesis offers a reasonable explanation for the overall increase in crystal size through the first few layers above an "unconformity." The roughly uniform size of the crystals above these first few layers, except for the variation between light- and dark-colored layers, suggests that after the deposition of the first few layers their growth period approached constancy.

The felsic basal layers may not reflect cessation of crystallization such as is postulated to precede the deposition of mafic basal layers, for the average size of the plagioclase crystals does not decrease toward the "unconformity." Probably they result from "stratigraphic" overlap. "Stratigraphically" lower layers would have lain to the right in figure 16*C*, and it is conceivable that the very lowest is a fine-grained layer like those present above most "unconformities."

The crosscutting dark streaks (including those along obvious faults) consist mainly of hornblende and a little biotite. Probably all the crosscutting dark streaks (not including dark basal layers above "unconformities") are localized along fractures. The hornblende in these streaks is similar to hornblende in layers and may be magmatic; nevertheless, some of the crosscutting hornblendite shown in figure 15 may in fact be metamorphic in origin, possibly a result of intrusion of the surrounding quartz monzonite.

Probably most of the "unconformities" were produced by corrosion by magmatic currents, which picked up or dissolved crystals previously deposited, but some sharp bends or even broad curves in the layers beneath "unconformities" could have been caused by slumping and sliding within the magma chamber during deposition. One difficulty in evaluating such a process is that the original configuration of the layers is not known. Although the layers now dip steeply, their original inclination very likely was more gentle. Slumping would require a sloping surface; the slope could have been initial or it could have resulted from tectonic movements. The possibility of deltaic deposition cannot be ruled out.

QUARTZ DIORITE AND RELATED GRANODIORITE

Quartz diorite includes rocks that are somewhat lighter colored than the hornblende gabbro, but darker than the large masses of granodiorite and quartz monzonite. In texture and in composition the quartz diorite is transitional between hornblende gabbro and rock in the larger plutons of granodiorite. The color index ranges from 20 to 40, plagioclase contains less than 50 percent anorthite (and commonly more than 40 percent), and the rock generally contains 10 percent or more quartz. Associated with the quartz diorite are slightly lighter colored rocks of granodioritic composition. Many of these lighter colored rocks are hybrids between more silicic granitic rock and quartz diorite, hornblende gabbro, or mafic volcanic rock.

The grains generally range from 1 to 5 mm, but more quartz diorite is in the lower part of this range than in the hornblende gabbro size range. Most quartz diorite is equigranular and hypidiomorphic-granular, but some is coarse inequigranular and in places contains poikilitic hornblende crystals an inch long. In quartz diorite of hypidiomorphic-granular texture, plagioclase commonly makes up 40 to 65 percent of the rock, hornblende about 15 percent, biotite 15 to 20 percent, and quartz 1 to 15 percent. Accessories include apatite, sphene, and magnetite; the usual secondary minerals are epidote, chlorite, and sericite. Plagioclase commonly is progressively zoned, but exhibits oscillations and discontinuities or strong changes in composition through narrow zones. Many crystals have a broad central zone and a compositional range of An_{40} to An_{50}. This zone commonly contains small cores as calcic as An_{60}, and is discontinuously rimmed with more sodic plagioclase that ranges in composition from An_{20} to An_{30}. Hornblende is similar to that in the hornblende gabbro except that the color in the Z direction generally is grayish blue green.

Most quartz diorite of hypidiomorphic-granular texture appears to have crystallized directly from a magma. Rock of this kind is found in the Deep Canyon area of the Tungsten Hills in the vicinity of the Little Sister mine, along the west side of the Pine Creek pendant, on the east side of Wheeler Crest north of the mouth of Pine Creek, and in many other places. The masses of quartz diorite north of Pine Creek may originally have belonged to a single large east-trending mass that was broken up and partly assimilated and granitized by later more silicic intrusives.

Much of the quartz diorite and associated granodiorite of irregular fabric may well be hybrid rock produced by metamorphic or metasomatic processes or by the contamination of silicic magma with femic wall rock (Bowen, 1928, p. 175–223; Nockolds, 1933, p. 561–589). Some coarse inequigranular textures, such as are found in quartz diorite along Big Pine Creek in the vicinity of the junction of the two main forks of Big Pine Creek, could have been produced by recrystallization of hornblende gabbro, quartz diorite, or mafic volcanic rock as a result of the intrusion of larger quartz monzonite or granodiorite masses. Small amounts of felsic material could have been added during recrystallization.

In other places the indications of interchange of material between early more mafic rocks and larger more silicic intrusives, and consequent hybridization of one or both are much stronger. Along the North Fork of Big Pine Creek, progressive hybridization of mafic rocks by quartz monzonite similar to the Cathedral Peak granite is evident. This hybridization has been achieved in part by granitization of original mafic rocks and in part by contamination of the quartz monzonite magma. Weak panidiomorphic-granular texture in the least altered rocks suggests that the original rock may have been an equigranular gabbro. The least altered rock contains about 50 percent plagioclase (strongly zoned approximately An_{50}–An_{10} or less), 40 percent hornblende, 5 percent biotite, and about 1 percent each quartz, microcline, magnetite and ilmenite, and sphene. Locally, it contains rounded quartz grains ⅛ to ¼ inch across, which are mantled with hornblende. These grains, which appear to be xenocrysts, survived extreme changes in the rock and provide a key to its original nature.

The least altered rock appears to have been granitized different amounts in different places. In general the granitization resulted in coarser grained rock that contains larger amounts of quartz, K feldspar, and biotite, and smaller amounts of plagioclase and hornblende. In slightly granitized rock that still retains weak panidiomorphic-granular texture, the approximate mineral content is 50 percent plagioclase (An_{38-32}), 10 percent hornblende, 15 percent biotite, 10 percent quartz, 10 percent K feldspar, and about 1 percent each magnetite, ilmenite, and sphene. In more strongly granitized rock the texture is hypidiomorphic-granular, and the approximate mineral content is 40 percent plagioclase (An_{35-20}), 25 percent biotite, 20 percent quartz, 10 percent K feldspar, and about 1 percent each magnetite, apatite, sphene, and hornblende. Rocks of all stages of granitization have disintegrated marginally, and ragged fragments half an inch or less in the longest dimension are strewn through the adjacent quartz monzonite. As distance from the mafic rock increases, the fragments are progressively less distinguishable, and generally they are not identifiable at distances of more than a few tens of feet, although the resultant rock has a darker color than uncontaminated quartz monzonite, with which it is locally in sharp contact.

During the granitization of the mafic rocks along the North Fork of Big Pine Creek, metamorphic differentiation was in operation along or near fractures. Figure 19 shows a partly granitized specimen of mafic rock that contains a veinlet of hornblende-rich rock, apparently localized along a fracture. The fact that the rock within an inch or so of the veinlet is deficient in mafic minerals suggests that the hornblende in the vein was derived from the walls. The thickness of the

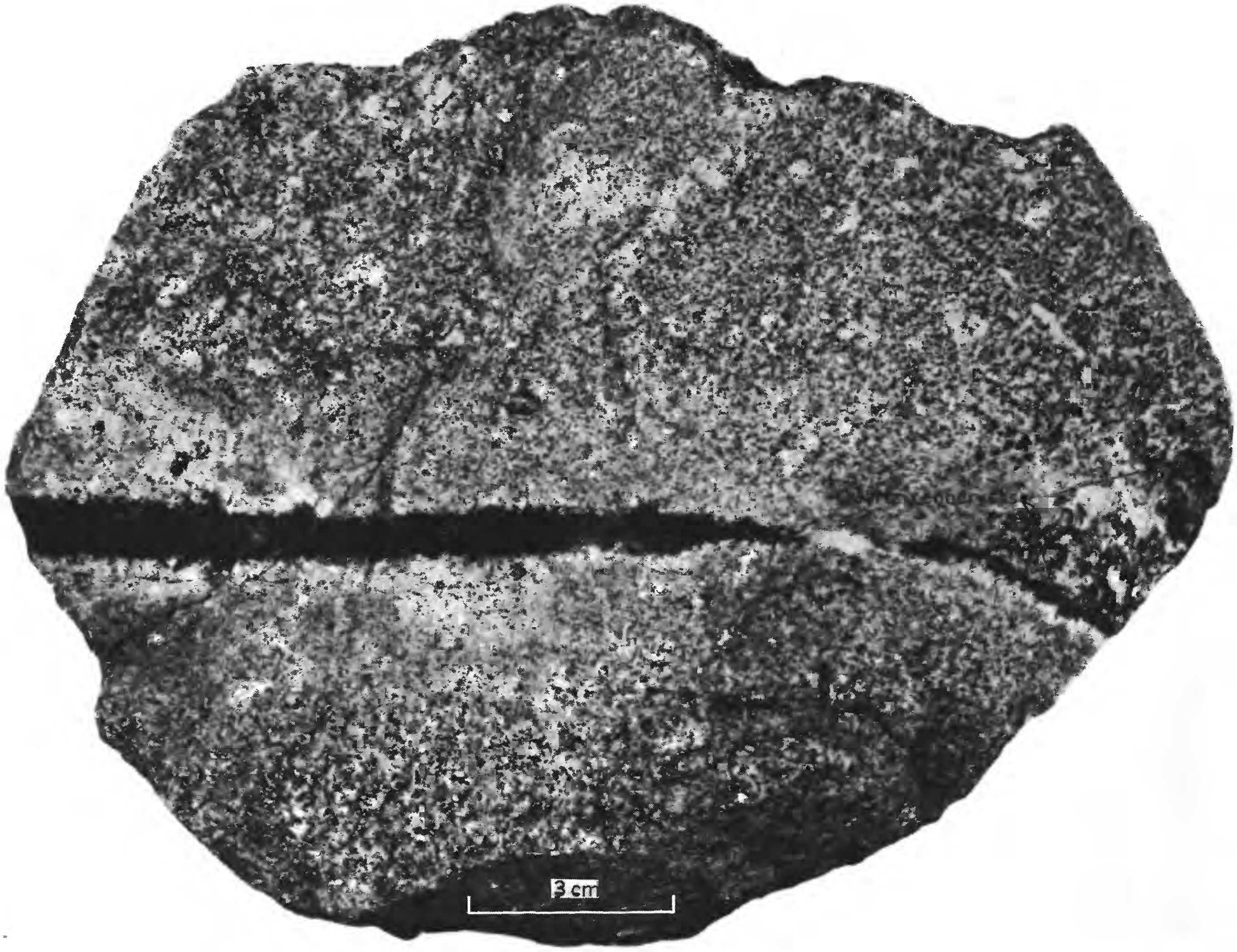

FIGURE 19.—Partly granitized mafic rock cut by a hornblende-rich vein. The hornblende in the vein appears to have been derived from the adjacent wall rock, which is deficient in mafic constituents. The thickness of the vein varies from place to place, generally in accord with variations in the thickness of wall rock deficient in mafic minerals.

vein changes from place to place, and the thickness of the zones in the walls deficient in mafic minerals varies accordingly. The vein consists of anhedral hornblende grains in a wide variety of shapes and sizes, accompanied by a little magnetite and quartz. The hornblende grains in the margins of the veinlet are highly poikiloblastic. The hornblende-poor wall rock consists of a ragged mosaic of plagioclase, quartz, and microcline. Large crystals of microcline are abundant close to the dikelet, whereas only small blebs enclosed in plagioclase occur farther away.

In the south end of the Pine Creek pendant, mafic metavolcanic rock and diorite, quartz diorite, and hornblende gabbro are all present. However, some of the rock mapped with quartz diorite may well be contaminated quartz monzonite. In Shannon Canyon and in the east side of Wheeler Crest just north of the mouth of Pine Creek canyon, darker colored rock having good hypidiomorphic-granular texture grades marginally through lighter colored hybrid rocks to quartz monzonite.

An example of the contamination of granodiorite magma where the contamination can be identified is found in the upper Pine Creek drainage, northwest of Pine Lake. There, hornblende gabbro has been intruded on the north by granodiorite, which is increasingly darker toward the hornblende gabbro. The contact, though difficult to locate because the contaminated granodiorite so closely resembles the hornblende gabbro, is sharp, and many details of the contact were mapped. The conclusion seems inescapable that here the granodiorite magma assimilated large amounts of hornblende gabbro. The contaminated gabbro was recognizable here ,but how many isolated masses of dark-colored rock of similar origin have been mapped as quartz diorite?

Scattered along the southwest side of the area from Piute Pass southeast to Mather Pass is dark fine-grained granodiorite having odd textures that suggest hybridization. Typical of these rocks is the mass in which the glacial basin occupied by Lake Sabrina was carved. The odd texture is largely the result of the distribution of K feldspar, which is scattered through the rock in patches $\frac{1}{4}$ to $\frac{1}{2}$ inch across. These patches are crowded with all the other minerals in the rock. Conceivably the K feldspar could have been of late magmatic origin, but it seems equally likely that it was derived from the enclosing granodiorite. If so, the parent rock was a more mafic plutonic rock, or possibly even a mafic volcanic rock.

One other possible origin for some masses of diorite as well as of hornblende gabbro is that they were derived from calcium-rich sedimentary rock. The metamorphic grade of the innermost wall rocks is in the amphibolite facies; therefore, any rock of appropriate composition could be converted to a stable assemblage of hornblende and plagioclase. In many places where amphibolite is associated with calc-hornfels, the hornfels has been formed from the amphibolite, but in a few places, especially in the margins of inclusions in granitic rock, calc-hornfels has been altered to amphibolite. Such alteration must have involved metasomatic exchange of material between the calcareous rock and the granitic magma. The variety of calc-hornfels commonly associated with amphibolite consists chiefly of diopsidic pyroxene and plagioclase. Conversion of diopside to hornblende would require subtraction of calcium and addition of water and femic constituents. In places, amphibolite is also present along contacts between marble and granitic rock, but the formation of amphibolite from marble requires much greater exchange of materials and is not common. Amphibolite margins in calcium-rich rocks adjacent to granitic rock can be seen at the Brown tungsten prospect, which is along the range front southwest of Bishop and a mile west of the Bishop Antimony mine, and at the Lakeview tungsten mine at the head of Gable Creek.

GRANITIC ROCKS

The formations of granitic rock in the Bishop district have been described in a preliminary paper (Bateman, 1961b). They range in composition from granodiorite to alaskite, but rocks in the compositional range of quartz monzonite are most abundant and are followed closely by granodiorite. The average granitic rock of the area is calcic quartz monzonite.

The plutons of granitic rock are of variable orientation and size. Most of the larger ones parallel the axis of the batholith, but some smaller ones are elongate in other directions or are not elongate at all. They generally range in outcrop area from 1 to 50 square miles. The largest continuous mass, of Lamarck granodiorite, underlies 54 square miles in the mapped area, and extends northwest, west, and southeast into bordering areas. Its total area may be several hundred square miles.

Many of the plutons are grouped as formations on the basis of composition, texture, and intrusive relations. Each mass assigned to a formation is given a name, and one has been selected as the type mass and contains the type locality. Thus the type mass of the Tungsten Hills quartz monzonite is the Tungsten Hills mass. Inclusion under one formational name of the rock in several discrete masses is not without hazard, but it is believed that most of the correlations that have been made are valid because the range in texture and composition within most individual masses is small compared with the differences between intrusives assigned to different formations. Nevertheless, most plutons are compositionally and texturally zoned, and in a few zoning is pronounced. The formally named granitic formations are the Inconsolable granodiorite, Tinemaha granodiorite, Wheeler Crest quartz monzonite, Round Valley Peak granodiorite, Lamarck granodiorite, and Tungsten Hills quartz monzonite.

Some plutons cannot, with confidence, be assigned to formations and are assigned to one of two informal rock groups, or are unassigned. The first informal rock group includes all plutons composed of rock similar in appearance and approximately correlative with the Cathedral Peak granite of Yosemite (Calkins, 1930, p. 126–127). These rocks are called "rocks similar to the Cathedral Peak granite," and include two lithologic facies, alaskite, and quartz monzonite. The second informal group includes all plutons composed of relatively fine-grained quartz monzonite and are designated "finer grained quartz monzonite." These intrusions are compositionally and texturally similar and seem to have been emplaced at about the same time. Some or perhaps all of them may be offshoots from the same parent magma and temporal equivalents, but the evidence is too weak to make any such assumptions. The rocks in four unassigned plutons are named informally in terms of locality and average composition. They are called "granodiorite of Coyote Flat," "granodiorite of Cartridge Pass," "granodiorite of Deep Canyon," and "quartz monozonite of McMurry Meadows."

The granitic rocks are composed dominantly of quartz and feldspar, and can be conveniently represented on a triangular diagram whose corners are the felsic constituents: quartz, K feldspar, and plagioclase. The classification used here differs only slightly from several other classifications that have been proposed. Quartz comprises at least 10 percent of the felsic constituents in all the granitic rocks. Boundaries between the fields of the different granitic rocks are in terms of the amount of K feldspar (including perthite) to total feldspar as follows: quartz diorite, 0 to 10 percent; granodiorite, 10 to 35 percent; quartz monzonite, 35 to 65 percent; granite, more than 65 percent (fig. 14).

Few individual specimens and no average compositions of plutons fall in the granite field as defined. Rocks of such composition do not seem to be common. The abundance of granite reported in the literature results partly from other usages of the term "granite": for all granitic rocks, for the most felsic rocks of a granitic suite regardless of composition, and for granitic rocks in which the average composition of the plagioclase is albite. The first two usages are loose and not acceptable for precise classification, but the third is part of several classifications and is incorporated in the classification used here with a modification—because the rocks in which the average composition of the modal plagioclase is albite contain only a few percent mafic minerals they are called alaskite rather than granite. Most of this alaskite plots in the quartz monzonite field on the diagram in figure 14.

MINERALOGY

The mineral content is similar in the granitic rocks and is discussed for the suite as a whole. Such differences in mineral composition as do exist are dealt with in discussions of the distinguishing characteristics of each rock. The essential minerals are quartz, plagioclase feldspar, and K feldspar. The varietal minerals include biotite, hornblende, and augite. Accessory minerals are magnetite and ilmenite, sphene, apatite, zircon, allanite, thorite, and monazite. Secondary minerals include epidote, sericite, and hematite.

ESSENTIAL MINERALS

QUARTZ

Quartz is in anhedral grains that have a wide range of size and shape. Commonly it contains numerous tiny liquid inclusions, but it contains few mineral inclusions. Most larger grains extinguish irregularly or consist of a mosaic of diversely oriented components, but in some rocks the larger grains extinguish regularly. The irregular extinction generally is either undulatory or by sharply defined polygonal areas that are visible near extinction, but in a few rocks linear twinning can be seen. Some of these conjugate twins are oriented symmetrically with respect to fractures that bisect the acute angle between the lineations.

POTASSIUM FELDSPAR (K FELDSPAR)

In hand specimen K feldspar is white or pinkish. Almost all of it is perthitic and most of it also exhibits the quadrille structure (grid twinning, gridiron structure, or grating structure) of microcline, but in places quadrille structure is inconspicuous or absent. Most of the albite in perthite is in thin, generally parallel but somewhat irregular lamellae of the sort commonly thought to be products of exsolution, but some is in irregular streaks and blebs. Except for subhedral to euhedral phenocrysts, which are conspicuous in some rocks of intermediate composition, K feldspar is anhedral and interstitial with respect to all the other minerals. Euhedral phenocrysts twinned according to the Carlsbad law are characteristically tabular parallel to (010) in some rocks and elongate or nearly equidimensional in others. Some phenocrysts are several inches in greatest dimension.

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Commonly the K feldspar contains inclusions of all the other minerals in the rock. Uncorroded prisms of plagioclase are the most common inclusions, and in phenocrysts generally are oriented parallel with the nearest crystal face. Inclusions of quartz are not abundant and are generally in rounded, possibly partly resorbed blebs. Phenocrysts of K feldspar commonly are accentuated by peripheral concentrations of mafic minerals, which are highly reminiscent of the dark carbonaceous material marginal to many andalusite porphyroblasts. The rough faces of the phenocrysts, caused by interference with the bordering grains during growth, can be seen in thin section and in phenocrysts that have weathered out of the enclosing rock.

PLAGIOCLASE

Plagioclase is present in relatively small white to light-gray subhedral grains that constitute phenocrysts only in finer grained rocks from dikes or small apophyses. The grains generally are smaller than most grains of quartz and K feldspar, and their size range is less. Many grains are twinned on the albite law, and some also are twinned on the pericline or Carlsbad laws. Lamellar albite twins generally are very closely spaced in albite and sodic oligoclase and more widely spaced in more calcic varieties. Inclusions are uncommon, but more calcic zones usually contain an abundance of secondary sericite.

The main part of almost all plagioclase grains is zoned continuously from a calcic core to a sodic rim, and many grains also exhibit thin oscillations that are superimposed on the broader zoning. Some grains also have unzoned, relatively albitic rims, or small exceptionally calcic core areas. The average compositions of the plagioclase commonly range from albite to andesine; albite is present in alaskite, oligoclase prevails in quartz monzonite, and andesine is the common plagioclase in granodiorite.

The composition of plagioclase was determined optically for most of the modally analyzed samples. The method followed was to record the extinction angle X′∧ (010) in sections that were oriented by means of a universal stage perpendicular to the albite twin plane and (010) cleavage and to the basal cleavage (001). The extinction angle was converted to percentage of An (fig. 20) with the use of a curve determined by F. C. Calkins (unpublished data, 1940) and modified by H. H. Hess (unpublished data, 1941). This method is not generally considered to yield results as accurate as other more cumbersome methods, but it has two advantages that outweigh the lack of precision: (1) it is rapid and thus permits making determinations of the composition of the plagioclase in a large number of samples, and

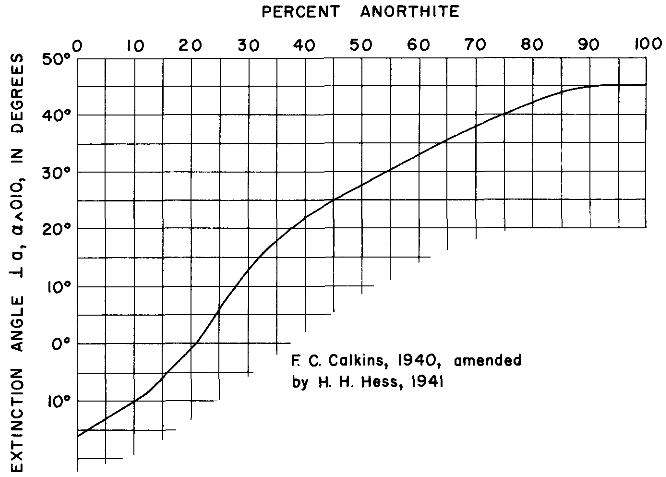


FIGURE 20.—Curve used to determine the anorthite content of plagioclase.

(2) it is ideally suited for determining the range of composition in zoned crystals. Checks provided by calculating the normative compositions of analyzed specimens do indicate, however, that the optical determinations are in the proper compositional ranges.

In tabulating the plagioclase compositions, three sets of figures are given. The first shows the composition of the rim where a discontinuity exists between the rim and the main body of the grain; the second shows the composition of the main body of the crystal; and the third shows the composition of any small exceptionally calcic cores. In addition to these recorded zones, many intergranular contacts between plagioclase and perthitic K feldspar are occupied by albite that was derived from the perthite in accordance with observations of Tuttle (1952, p. 115 and pls. 3, 4).

VARIETAL MINERALS

BIOTITE

Biotite is present in all the granitic rocks; in some it is characteristically in individual hexagonal plates and in others it is in clusters of smaller, irregularly shaped plates, associated with accessory minerals and with hornblende, where present. The pleochroic color in the X direction in almost all specimens is pale grayish yellow, but in the Z direction it ranges in different rocks through various shades of red, grayish red, brownish red, yellowish brown, olive brown, olive, grayish olive, and olive gray.

HORNBLENDE

Hornblende is present in all of the granodiorite and in calcic quartz monzonite but is absent in sodic quartz monzonite and in alaskite, except in hybrid rocks. It occurs both in well-formed prismatic crystals and in ragged anhedral grains, many of which poikilitically enclose grains of plagioclase. The typical prismatic habit of hornblende in some rocks is one of the distinguishing characteristics of those rocks.

AUGITE

Augite is present as cores in hornblende in the Tinemaha and Inconsolable granodiorites and in more mafic parts of the granodiorite of McMurry Meadows.

ACCESSORY MINERALS

The common accessory minerals include magnetite and ilmenite, sphene, apatite, zircon, and allanite. Thorite and monazite have also been identified; Larsen recovered sufficient thorite from the Lamarck granodiorite and sufficient monazite from the granodiorite of McMurry Meadows to make age determinations by the lead-alpha method (Larsen and others, 1958, p. 50).

Magnetite and ilmenite are present in small grains in all the granitic rocks, but are most abundant in those having the highest color index. Much of the magnetite is altered to hematite. Sphene occurs in discrete double wedges and in irregular masses associated with ilmenite. Most of the sphene is brownish or yellowish and only faintly pleochroic, but some of it is pleochroic in brilliant shades of red, and superficially resembles piedmontite. Apatite is in euhedral prismatic crystals. Commonly the prisms are short and stubby, but in some rocks they are long and very thin. Tiny euhedra of zircon are present in all the rocks, but are most abundant in quartz monzonite and alaskite. Allanite is a rather uncommon accessory, but where present it is in elongate conspicuous grains. It is somewhat variable in color and in pleochroism; colors range from reddish brown through grayish orange to yellowish orange.

ALTERATION PRODUCTS

Secondary minerals include sericite, epidote, and chlorite. Sericite is present chiefly in the cores and other calcic zones in plagioclase. Most chlorite is an alteration product of biotite, and most epidote is derived from hornblende or calcic plagioclase.

TEXTURES

Each of the granitic rocks represented on the map is characterized by a "typical appearance" which is largely determined by the color index, grain size, texture, and fabric of the rock (fig. 21). These features as much as any others are the bases for correlating and discriminating among the granitic rocks, and are useful criteria for field use. Except for uncommon local variants, the granitic rocks are medium grained, and hypidiomorphic-granular, and have color indices that fall between 2 and 20. Most rocks are equigranular, but some are porphyritic, and a few are seriate. The term "color index" is the content of dark-colored minerals—biotite, hornblende, and the opaque minerals—expressed in volume percent.

The average grain size of the common nonporphyritic rocks and the groundmasses of the common porphyritic ones ranges from 1 to 5 mm. Phenocrysts in porphyritic rocks range widely in size, and some phenocrysts of K feldspar in the Wheeler Crest quartz monzonite are as much as several inches long. Although correlation between kind of rock and common or "typical" grain size is apparent, the connection is probably largely indirect. Grain size appears to be primarily a function of the size and shape of the granitic mass; some rocks occur commonly in masses of larger size than others, and the average grain size of these rocks, therefore, is greater than that of rocks that commonly occur in smaller masses. The common grain size of the fine-grained quartz monzonite, which occurs in relatively small masses having no more than a few square miles of outcrop, is 1 to 2 mm., but the common grain size in the main mass of the Lamarck granodiorite, which has an outcrop area of at least 50 square miles, is 4 to 5 mm. Shape, too, is a factor that determines grain size, for thin masses commonly are finer grained than thicker ones of the same area.

The grain size within most intrusions is fairly uniform, but in some rocks is finer grained toward the margins and in small apophyses. For example, the Morgan Creek mass of the Tungsten Hills quartz monzonite, which lies along the west side of the Pine Creek pendant, is increasingly finer grained toward the western margin, where it is in contact with older granodiorite and hornblende gabbro. Constricted projections of the Tungsten Hills quartz monzonite into the northern margin of the Bishop Creek pendant likewise are finer grained than is common for that rock.

Equigranular rocks are more common than prophyritic rocks, but both are well represented. Generally one texture or the other prevails, but in some rocks one grades to the other. The Wheeler Crest quartz monzonite, for example, is porphyritic in most places, but locally grades to equigranular rock.

In equigranular hypidiomorphic-granular rock, plagioclase, hornblende, and biotite commonly are subhedral and quartz and K feldspar are anhedral. The Lamarck granodiorite is characterized by conspicuous euhedral or nearly euhedral crystals of hornblende and biotite, and most of the biotite in both the Tungsten Hills quartz monzonite and Wheeler Crest quartz monzonite is in streaky aggregates of subhedral grains. In some specimens, some or all of the hornblende is anhedral and poikilitically encloses smaller grains of pla-

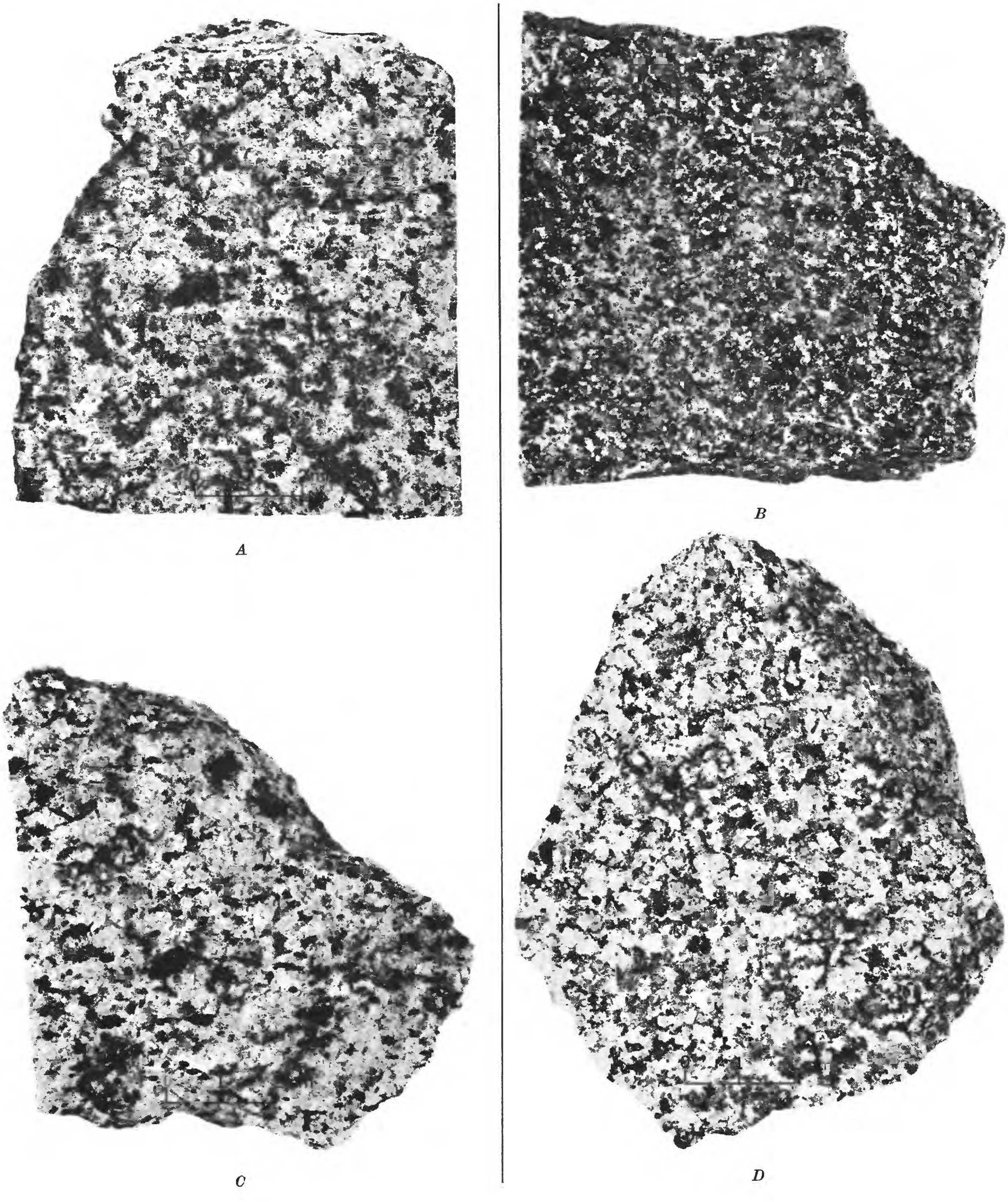


FIGURE 21.—Typical granitic rocks from the Bishop district. *A*, tinemaha granodiorite. *B*, Inconsolable granodiorite. *C*, Lamarck diorite. *D*, Round Valley Peak granodiorite.

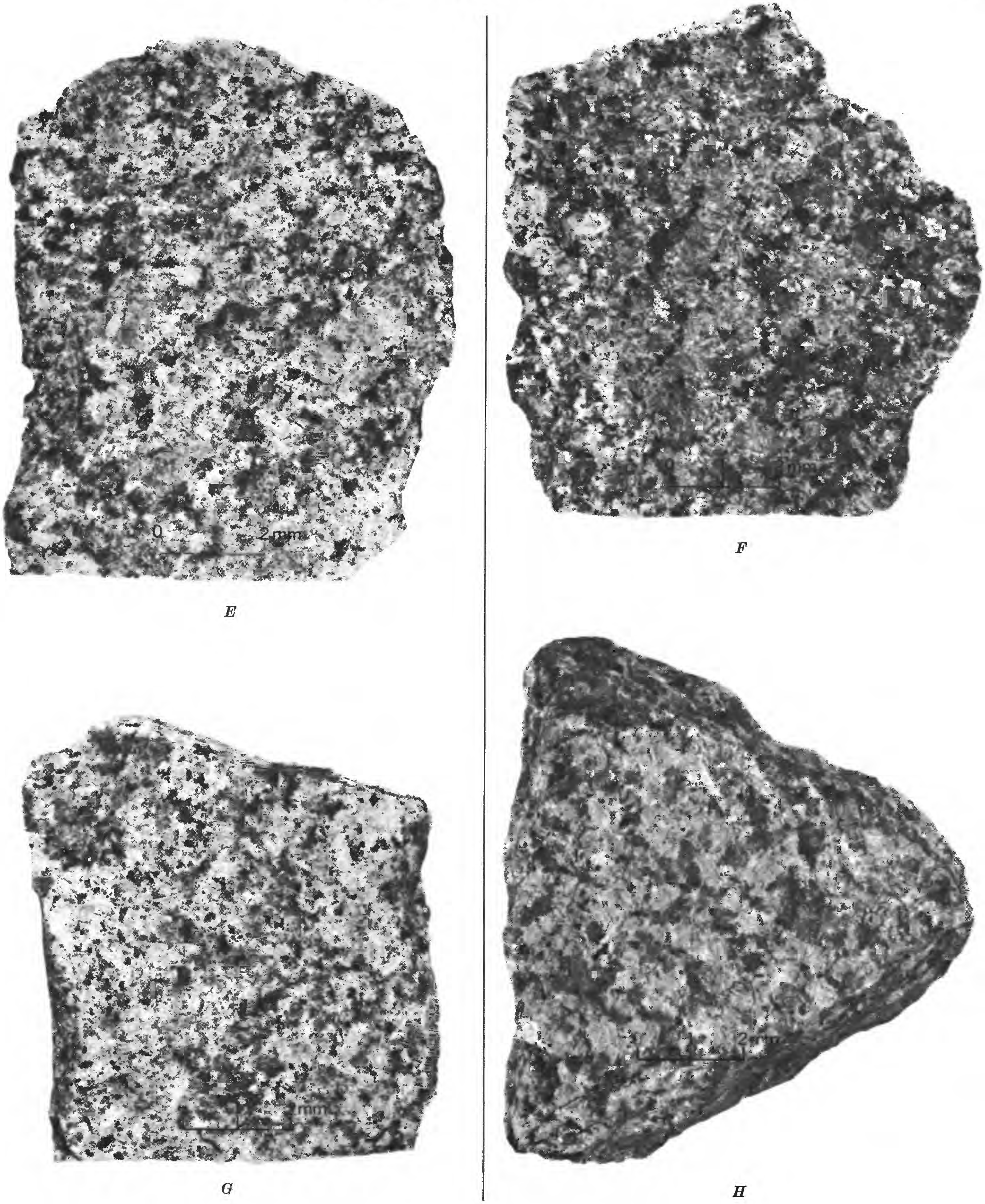


FIGURE 21—Continued—*E*, Wheeler Crest quartz monzonite. *F*, Tungsten Hills quartz monzonite. *G*, Quartz monzonite similar to the Cathedral Peak granite. *H*, Alaskite similar to the Cathedral Peak granite.

gioclase and the accessory minerals. Both quartz and K feldspar are interstitial to the other minerals, K feldspar being the more ramifying of the two. Much of the K feldspar is notably poikilitic and encloses grains of the accessory minerals, hornblende, biotite, plagioclase, and more rarely, rounded, apparently partly resorbed blebs of quartz. Most of the cores of K feldspar are clean, and included minerals are increasingly abundant toward the margins. The accessary minerals—apatite, magnetite, zircon, sphene, and allanite—commonly are closely associated with hornblende and biotite.

The porphyritic rocks can be conveniently placed into two groups, those with phenocrysts of K feldspar and those with phenocrysts of other minerals, usually plagioclase with or without hornblende, quartz, and K feldspar. The phenocrysts in plutons are generally K feldspar whereas the phenocrysts in finer grained dikes, marginal rock, and small apophyses are generally earlier formed minerals such as hornblende or plagioclase. The phenocrysts in finer grained rocks are of about the same habit and size as the same minerals are in coarser grained rocks of typical aspect, and not uncommonly are set in an allotriomorphic-granular groundmass. In some of these rocks plagioclase is the only mineral that forms phenocrysts, and in others all the essential and varietal minerals occur both as phenocrysts and in the groundmass. These relations can be readily explained by the mechanism that is usually assumed for porphyritic rocks; for example, the phenocrysts are early crystallized minerals that were suspended in the magma when it was moved to an environment that resulted in the more rapid crystallization of the groundmass.

The phenocrysts of K feldspar in more typical porphyritic rock, however, cannot be explained in the same way because they formed during the later rather than the earlier stages of crystallization. Knopf infers from the abundance of included minerals that the large tabular phenocrysts of K feldspar in the porphyritic granite of Mount Whitney began to crystallize late and suggests that the rock owes its porphyritic texture "to a superior velocity of crystallization and a superior power of attracting the crystallizing molecules to a few centers of crystallization" (Knopf, 1918, p. 66).

In some rocks, notably the Wheeler Crest quartz monzonite and, locally, the quartz monzonite similar to the Cathedral Peak granite, the phenocrysts are euhedral; in others, such as the Tinemaha granodiorite and the Tungsten Hills quartz monzonite, they are anhedral to subhedral. Regardless of external form, the phenocrysts enclose grains of the other minerals in the rock, including blebs of partly resorbed quartz, and are accentuated by thin peripheral concentrations of mafic minerals. The faces of even euhedral crystals are rough because of interference between the phenocrysts and the bordering grains. The inclusions and the marginal concentrations of mafic minerals support Knopf's hypothesis that the phenocrysts crystallized later than most of the other minerals. The peripheral zone of mafic minerals marginal to many crystals is highly reminiscent of the dark carbonaceous zones commonly associated with chiastolite, and suggests that the minerals in the dark zone were expelled or pushed aside by the growing phenocrysts at the time of their formation. It is true that in Becke's (1903) crystalloblastic series K feldspar occupies the lowest position, but the concept of simultaneous growth of all the constituents does not seem applicable here. If most of the growth of the phenocrysts was made after crystallization of most other minerals was well advanced, competition with other minerals for a euhedral form would be small.

Phenocrysts of K feldspar are restricted almost entirely to rocks of intermediate composition (sodic granodiorite and calcic quartz monzonite). Experimental studies of artificial melts show that plagioclase is ordinarly the first of the essential felsic minerals in granite to form, and that quartz and K feldspar do not crystallize until the temperature has fallen to a point where sodic plagioclase or albite is being formed. In more calcic rocks a large part of the rock may already be crystallized before K feldspar begins to form, and consequently the K feldspar may be relegated to interstices. On the other hand, in very silicic rocks such as alaskite, the felsic constituents may all crystallize at about the same time and inhibit the growth of euhedral crystals of any one kind. Only in rocks of intermediate composition, where some minerals are formed in advance of K feldspar but not so many that K feldspar cannot grow into euhedral crystals through its strength of crystallization, are conditions especially favorable for the formation of phenocrysts of K feldspar.

ANALYTICAL DATA

Two kinds of analytical data were obtained for the granitic rocks, (1) limited chemical data and (2) extensive modal data. Fourteen samples were analyzed chemically, nine by standard methods and five by the rapid method described by Shapiro and Brannock (1956). The nine specimens analyzed by standard chemical methods were also analyzed by spectrographic means for semiquantitative determination of the minor elements. The standard chemical analyses are all of granodiorite and quartz monzonite; four of the rapid analyses are of rocks similar to the Cathedral Peak granite (alaskite and quartz monzonite), and one is of quartz diorite. In addition to these analyses a partial analysis of a sample of alaskite similar to the Cathedral Peak granite published by Knopf (1918, p. 68) was utilized. The chemical and spectrographic analyses and norms are summarized in table 3.

The modes of about 300 samples of granitic rocks were determined. First, modes for all samples were made by using thin sections of standard size (about a square inch of rock on the average) and a technique similar to that described by Chayes (1956). Plots of the modes made from thin sections on quartz-K feldspar-plagioclase diagrams showed less scatter for the finer grained rocks than for the coarser grained rocks. This and other comparisons of the modes of coarser and finer grained rocks showed that ordinary size thin sections are too small to be representative samples of coarser grained rocks.

In early 1960, the coarser grained rocks were recounted by using rock slabs in which K feldspar was stained yellow and plagioclase light red to reddish brown by a method described by Bailey and Stevens (1960). In this procedure K feldspar is stained with sodium cobaltinitrate and plagioclase is stained with rhodozonic acid, which acts on previously introduced barium. The actual counting was done by projecting

TABLE 3.—*Summary of chemical and spectrographic analyses and norms of the granitic rocks*

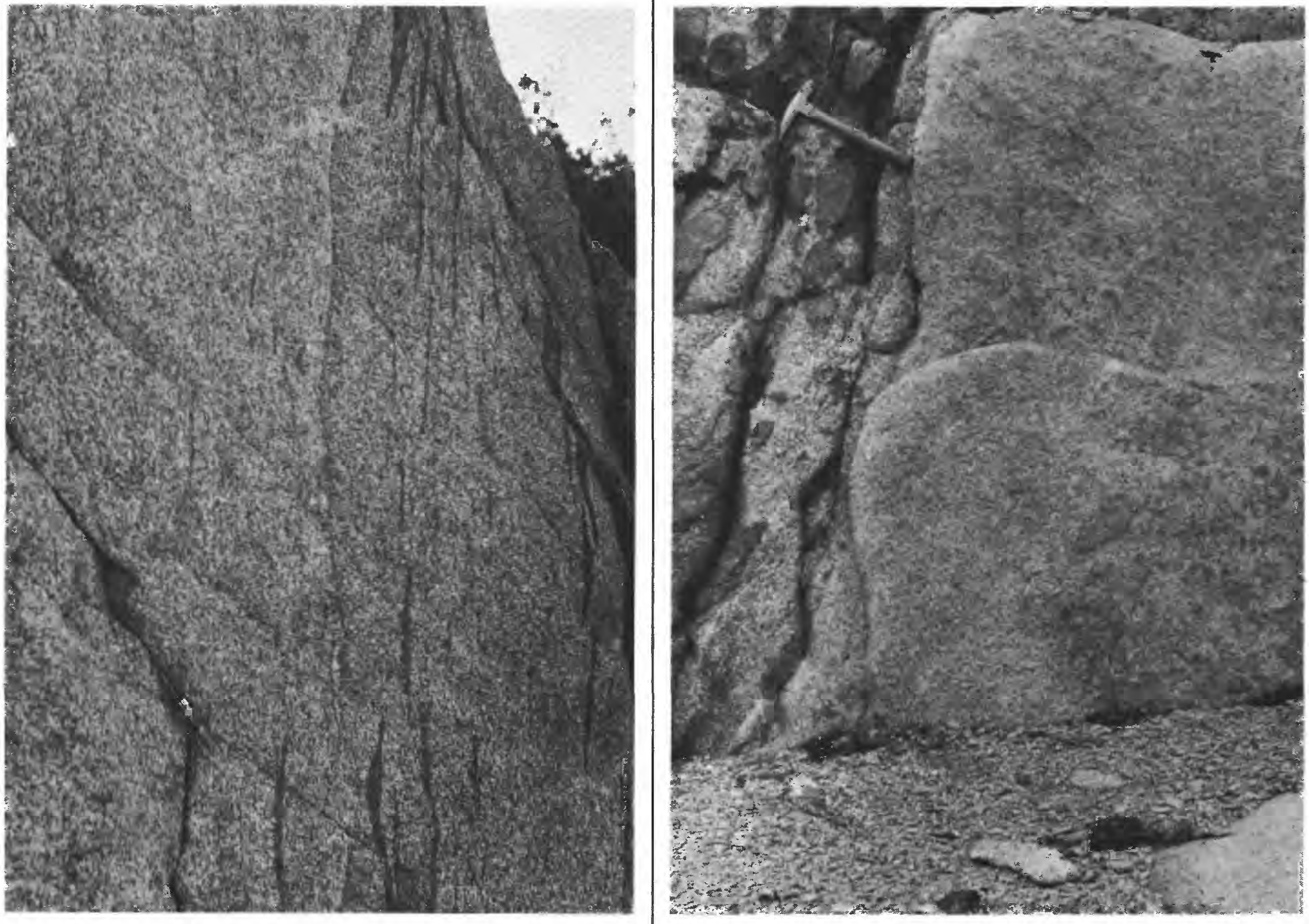
	Quartz diorite	Inconsolable granodiorite	Tinemaha granodiorite	Granodiorite of McMurry Meadows	Wheeler Crest quartz monzonite	Round Valley Peak granodiorite	Lamarck granodiorite	Tungsten Hills quartz monzonite, specimen—		Rocks similar to Cathedral Peak granite					
										Quartz monzonite, specimen—			Alaskite, specimen—		
								5	52	4	12	23	37	46	K
Chemical analyses															
[1, Rapid rock analysis; analysts: H. F. Phillips, P. L. D. Elmore, and K. E. White. 2, Standard rock analysis; analyst: L. M. Kehl. 3, Rapid rock analysis; analysts: P. L. D. Elmore, S. D. Botts, and M. D. Mack. 4, Partial analysis reported by Knopf (1918, p. 68) of alaskite from Rawson Creek canyon]															
Lab. No.	138275(1)	53-1303 SCD(2)	53-1299 SCD(2)	53-1301 SCD(2)	53-1295 SCD(2)	53-1302 SCD(2)	53-1300 SCD(2)	53-1298 SCD(2)	53-1297 SCD(2)	152445(3)	152446(3)	53-1296 SCD(2)	152443(3)	152444(3)	(4)
SiO_2	59.0	61.00	62.82	64.86	71.42	63.53	66.92	71.42	69.60	71.8	73.0	74.11	71.0	75.4	76.28
Al_2O_3	17.2	16.06	15.44	16.12	14.47	15.61	15.19	14.03	14.89	15.3	15.2	13.73	15.7	13.3	------
Fe_2O_3	2.0	1.86	2.59	1.90	1.03	2.35	1.45	.89	1.07	1.0	.4	.60	.9	.3	------
FeO	4.6	4.06	3.17	2.52	1.38	3.25	2.52	1.63	1.99	.65	.34	.88	.81	.74	------
MgO	2.8	3.10	2.35	1.55	.78	2.54	1.74	.70	.91	.34	.16	.32	.39	.12	------
CaO	6.2	5.46	5.04	3.80	2.86	4.58	3.79	1.91	2.70	1.8	1.1	1.29	1.6	.48	.47
Na_2O	3.3	3.45	3.15	3.44	3.44	3.31	3.16	2.86	3.18	3.8	3.8	3.44	3.8	4.1	4.72
K_2O	2.1	2.95	3.72	4.03	3.69	2.98	3.82	5.35	4.45	4.1	5.0	4.92	5.0	4.5	4.73
H_2O-	1.0	.05	.03	.06	.06	.04	.06	.08	.08	.48	.48	.12	.69	.46	------
H_2O+		.58	.62	.51	.21	.61	.48	.35	.31			.18			------
TiO_2	.82	.88	.64	.57	.25	.63	.47	.36	.42	.22	.09	.18	.22	.10	------
CO_2	<.05	.00	.01	.00	.03	.03	.02	.02	.01	.08	.10	.01	.07	.11	------
P_2O_5	.34	.25	.30	.23	.10	.23	.18	.09	.12	.05	.05	.06	.05	.01	------
MnO	.14	.10	.11	.09	.08	.12	.08	.05	.07	.06	.06	.05	.04	.08	------
BaO	------	------	------	.13	------	------	------	------	------	------	------	------	------	------	------
Total	100.0	99.80	99.99	99.81	99.80	99.81	99.88	99.74	99.80	100	100	99.89	100	100	------
Quantitative spectrographic analyses for minor element															
[Looked for but not found: Ag, As, Au, Bi, Cd, Ge, In, Mo, Pt, Sb, Sn, Ta, Th, Tl, U, W. A trace of Be (less than 0.00005) was found in all samples. Analyst: P. R. Barnett															
B	------	0.004	0.001	0.0	0.0	0.0	0.0	0.0	0.0	------	------	0.0	------	------	------
Ba	------	.1	.2	.2	.2	.2	.1	.1	.2	------	------	.1	------	------	------
Co	------	.002	.002	.0009	.0004	.001	.001	.0005	.0005	------	------	.0001	------	------	------
Cr	------	.004	.001	.001	.0003	.002	.001	.0006	.0005	------	------	.0003	------	------	------
Cu	------	.002	.001	.001	.0003	.0009	.0006	.0004	.0008	------	------	.0002	------	------	------
Ga	------	.002	.002	.001	.001	.002	.001	.001	.001	------	------	.001	------	------	------
La	------	.006	.01	.009	.01	.009	.009	.008	.01	------	------	.005	------	------	------
Nb	------	.002	.002	.002	.002	.002	.002	.002	.002	------	------	.001	------	------	------
Ni	------	.003	.0008	.0004	.0001	.0007	.0005	.0003	.0002	------	------	.0001	------	------	------
Pb	------	.002	.002	.002	.001	.002	.002	.002	.002	------	------	.002	------	------	------
Sc	------	.001	.001	.0007	.001	.001	.001	.001	.001	------	------	.0007	------	------	------
Sr	------	.1	.09	.06	.05	.09	.07	.03	.03	------	------	.04	------	------	------
V	------	.01	.01	.007	.005	.009	.006	.004	.005	------	------	.001	------	------	------
Y	------	.003	.005	.005	.003	.003	.004	.004	.005	------	------	.002	------	------	------
Yb	------	.0003	.0004	.0003	.0002	.0002	.0002	.0003	.0004	------	------	.0001	------	------	------
Zn	------	.01	.0	.0	.0	.01	.0	.0	.0	------	------	.0	------	------	------
Zr	------	.01	.02	.02	.01	.01	.01	.05	.02	------	------	.03	------	------	------
Norms															
Q	13.30	13.32	16.69	18.90	29.94	15.84	22.62	28.38	26.52	29.64	29.22	31.92	25.68	32.58	29.76
or	12.29	17.27	22.20	23.91	21.71	17.76	22.76	31.68	26.13	24.46	29.47	28.94	29.47	26.69	27.24
ab	27.84	29.32	26.71	28.82	28.85	32.42	26.72	24.11	26.72	31.96	31.96	28.85	31.96	34.58	29.82
an	25.95	19.46	16.68	16.68	13.35	16.13	15.84	9.46	12.51	8.90	5.56	6.40	8.06	2.50	2.50
di	2.51	4.73	5.16	.36	------	5.44	1.83	------	------	------	------	------	------	------	------
hy	11.67	10.00	6.18	5.91	3.43	6.55	6.18	3.55	4.53	.90	.66	1.72	1.26	1.36	------
mt	2.78	2.78	3.74	2.78	1.46	3.43	2.09	1.39	1.57	1.39	.70	.89	1.39	.46	------
il	1.52	1.67	1.22	1.07	.47	1.21	.90	.65	.78	.46	------	.32	.46	------	------
ap	.64	.64	.69	.62	.27	------	.35	------	.29	------	------	------	------	------	------
c	------	------	------	------	------	------	------	.10	.31	1.33	1.53	.51	1.12	.71	------
Total	98.50	99.19	99.27	99.05	99.48	98.78	99.29	99.32	99.36	99.04	99.10	99.55	99.40	98.88	------

colored photographic transparencies of the slabs onto a grid of about 2,000 dots.

Counting projections of transparencies permitted the use of larger and more representative samples than was possible with thin sections, and most plots of modes determined from stained slabs show less scatter than plots determined from thin sections cut from the same samples. The average slab counted had an area about five times that of a standard thin section. The stained slabs have the disadvantage that they permit counting only four constituents: quartz, K feldspar, plagioclase, and mafic minerals; biotite and hornblende cannot be counted separately. Modes of most samples of the Inconsolable granodiorite, Tinemaha granodiorite, Wheeler Crest quartz monzonite, Lamarck granodiorite, Round Valley Peak granodiorite, Tungsten Hills quartz monzonite, and rocks similar to the Cathedral Peak granite were counted on stained slabs, but a few modes of these rocks, which were of finer grain size or for which stained slabs could not be obtained, were counted in thin section. Modes of the other rocks were determined from thin sections.

ROCK DESCRIPTIONS

The plutonic rocks are described in order of their decreasing age as interpreted from intrusive relations and from the map pattern. Slight divergence from this plan is made to permit bringing together the rocks within each of two sequences not in contact; one of these, called informally the Tinemaha sequence, is confined to the area south of the septum along Big Pine Creek and the other, called informally the Bishop sequence, is con-



A *B*

FIGURE 22.—External contacts of the Inconsolable granodiorite in the drainage basin at the head of the South Fork of Big Pine Creek. *A*, Contact with Tinemaha granodiorite (left half of photograph) southeast of Elinore Lake about half a mile. Mafic inclusions that are present in both rocks parallel to contact and to foliation suggest proximity to the original intrusive contact of both masses; note greater abundance of inclusions in the Inconsolable. Relative ages are uncertain. *B*, Contact with quartz monzonite similar to the Cathedral Peak granite about half a mile south of Contact Pass. Note absence of flattened mafic inclusions in quartz monzonite, which is the younger rock.

fined to the area north of the septum. The Tinemaha sequence includes the Tinemaha granodiorite, the Inconsolable granodiorite, and the quartz monzonite of McMurry Meadows; the Bishop sequence includes the Wheeler Crest quartz monzonite, the Round Valley Peak granodiorite, the Tungsten Hills quartz monzonite, and the granodiorite of Coyote Flat. The order of emplacement is usually also an order of decreasing silica content; thus the first rocks described are rich in calcium, iron, and magnesium, and later rocks are increasingly richer in silicon and the alkalis. A notable exception is that the very last plutons emplaced are of granodioritic composition.

INCONSOLABLE GRANODIORITE

The Inconsolable granodiorite is represented by medium-grained, somber-hued rock that constitutes the Sierra Nevada divide from Middle Palisade to Mount Agassiz, and extends northward in the Inconsolable Range to the latitude of Chocolate Peak. The spectacular cirques at the heads of the main and the South Forks of Big Pine Creek, which are still occupied by glaciers, are carved in this rock. The main mass of rock is elongate in a northwesterly direction and has an outcrop area of a little more than 12½ square miles. A second very small mass, having an outcrop area of only about a tenth of a square mile, is about half a mile south of the main mass. The type locality is in the Inconsolable Range; characteristic rock is also exposed in the Palisade Crest and the cirques at the head of Big Pine Creek.

The Inconsolable granodiorite is megascopically equigranular and medium grained, and the average grain size is about 2 mm (fig. 21*B*). It is distinctly finer grained toward most margins. In overall aspect the rock is medium to medium-dark gray; this relatively dark hue results partly from a high average color index of about 18, and partly from the prevalent gray to grayish-red color of the feldspar. Many specimens, especially those from marginal parts, contain scattered small but conspicuous grains of moderate-red to reddish-brown plagioclase.

Primary foliation generally is recognizable and is especially conspicuous in the margins of the mass. The rock contains abundant mafic inclusions that are oriented parallel with the foliation and are increasingly flattened near external contacts (fig. 22*A*). Mafic dikes, abundant in the adjacent Tinemaha granodiorite and present farther north in the Lamarck granodiorite, were observed only in the extreme north end of the Inconsolable granodiorite.

Quartz and K feldspar are approximately equal in abundance, and plagioclase generally is more than twice as abundant as either. Biotite is the predominant mafic mineral, and hornblende and augite are also present. The accessories are the usual ones. The texture in thin section is seriate, and the largest grains are plagioclase.

Plagioclase is in subhedral zoned crystals of a wide size range. The average composition is about An_{40}, the most calcic plagioclase in any of the granodiorites. Commonly, it is zoned in the general range of An_{30} to An_{50}; locally, small calcic cores are present, and many crystals are rimmed with calcic plagioclase (An_{30}) (table 4). Quartz and K feldspar commonly are interstitial to the plagioclase. The quartz grains lack the irregular patterns of extinction prevalent in most other intrusives. In most specimens K feldspar does not exhibit the quadrille structure of microcline and is only weakly perthitic, although in places it contains a few laminae of albite.

Biotite is in conspicuous plates having a pleochroism X=grayish yellow, Y=Z=grayish red to redddish brown. Especially at the margins of grains, the grayish red to reddish brown color grades to light olive, and in a few grains Y=Z=olive gray. Large plates of biotite not uncommonly enclose hornblende, which in turn may enclose augite. Both augite and hornblende also are found in discrete grains, although most augite grains are irregularly bordered with hornblende. Augite appears to be more abundant than hornblende, a feature that is unique to this rock.

The locations of modally analyzed specimens are shown in figure 23. The plot of modes (fig. 24) which is elongate away from the plagioclase corner, indicates considerable range in the plagioclase content. Comparison of the range of modes with the positions of the specimens within the intrusion fails to reveal any systematic relation between composition and position.

The Inconsolable granodiorite is probably the oldest granitic rock in the area south of Big Pine Creek, but its relations to the Tinemaha granodiorite are uncertain. The fact that about half a mile southwest of Mount Bolton Brown mafic dikes in the Tinemaha terminate at the contact with the Inconsolable granodiorite suggests they were cut off, but half a mile southeast of Lake Elinore a swarm of inclusions in the Inconsolable granodiorite terminates at the contact with the Tinemaha granodiorite in such a way as to suggest it was cut off. The age of the mafic dikes has not been definitely determined, and they may be younger than either intrusive. Inasmuch as the Inconsolable granodiorite is more mafic than the Tinemaha granodiorite, it probably crystallized at higher temperatures and a little earlier.

The Inconsolable granodiorite is clearly intruded by the Lamarck granodiorite and by quartz monzonite similar to the Cathedral Peak granite. Both the Lamarck

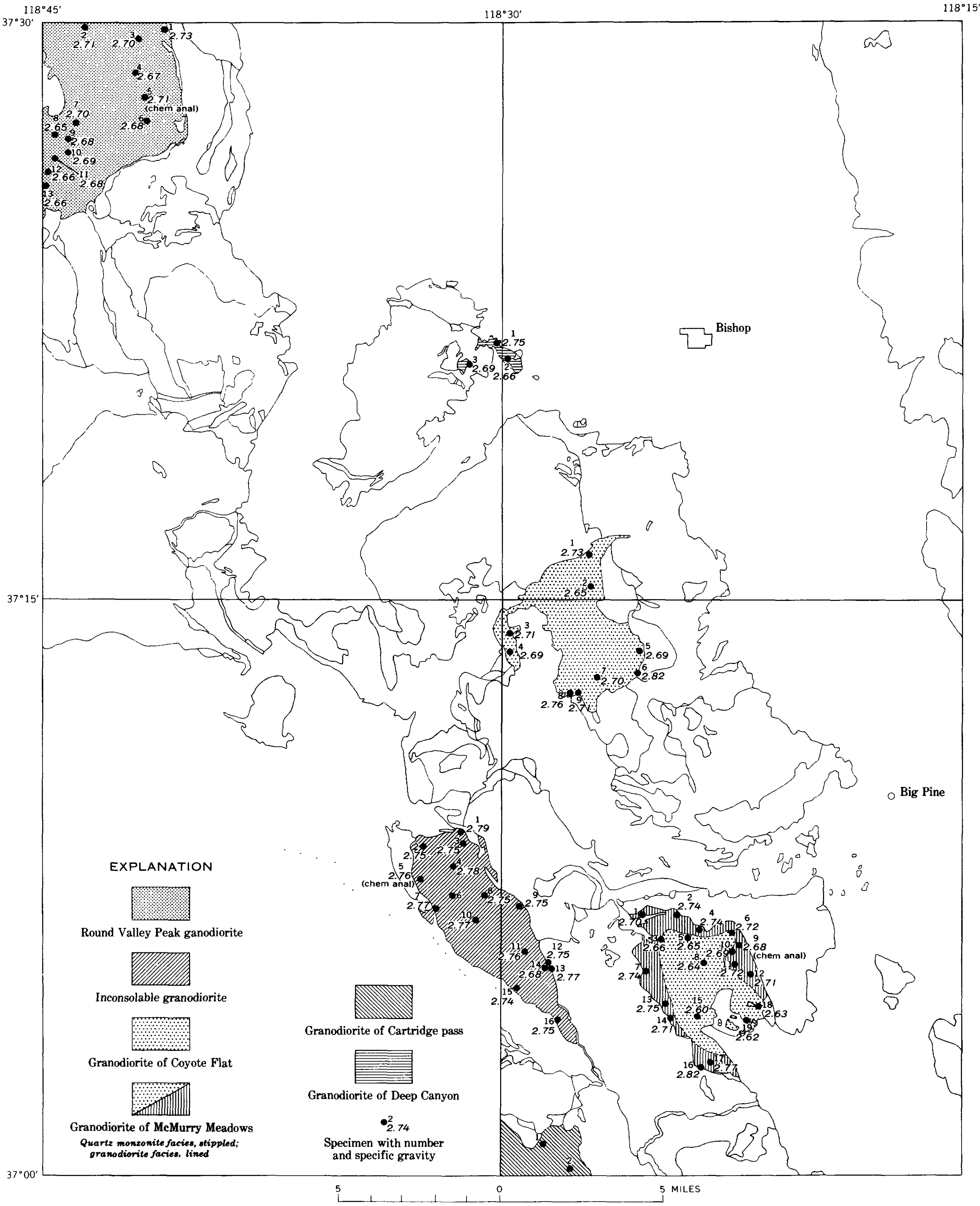


FIGURE 23.—Map showing the locations of modally analyzed specimens of the Round Valley Peak and Inconsolable granodiorites, the granodiorite of Coyote Flat, the quartz monzonite of McMurray Meadows, the granodiorite of Cartridge Pass, and the granodiorite of Deep Canyon.

granodiorite and the quartz monzonite similar to the Cathedral Peak granite contain inclusions of the Inconsolable granodiorite and penetrate it with dikes. North of Contact Pass the contact between the Inconsolable granodiorite and the Mount Alice mass of quartz monzonite similar to the Cathedral Peak granite makes a westerly bend of about 40°, and the granodiorite in the east side of Temple Crag is embraced by the quartz monzonite. The granodiorite there is conspicuously brecciated and cemented by fine-grained felsic igneous rock and by quartz (fig. 25), which almost certainly

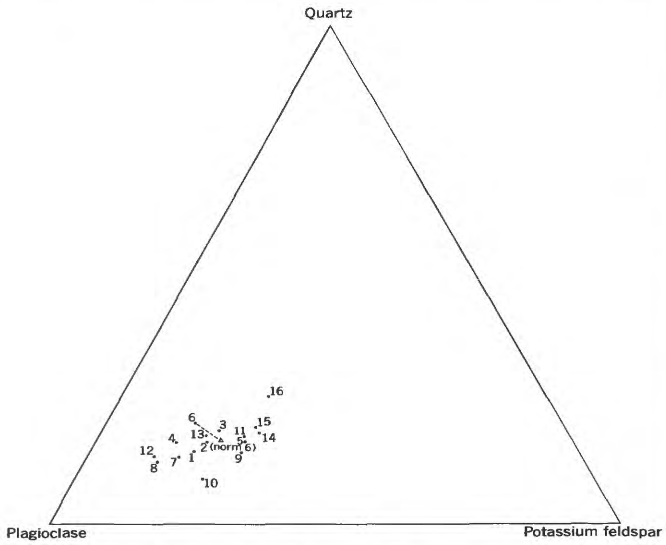


Figure 24.—Plot of modes of Inconsolable granodiorite on quartz-K feldspar-plagioclase diagram.

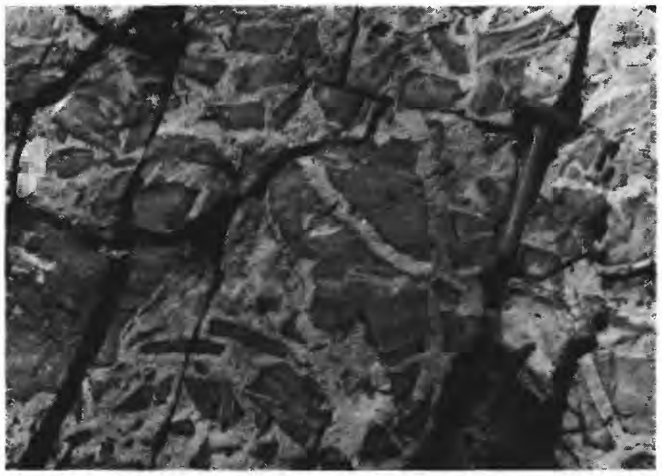


Figure 25.—Brecciated Inconsolable granodiorite in the east side of Temple Crag. Fragments are cemented by fine-grained felsic igneous rock and by quartz. The cementing materials almost certainly were derived from adjacent quartz monzonite similar to the Cathedral Peak granite, which caused the brecciation.

Table 4.—*Modal analyses of Inconsolable granodiorite, in volume percent*

[Where the content of biotite, hornblende, and accessory and secondary minerals is shown separately, the mode was determined from a thin section. Where only the total content of mafic minerals is shown, the mode was determined from a stained slab. Location of specimens is shown on figure 23]

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Biotite	Hornblende	Accessory and secondary minerals	Total mafics	Percent of anorthite in plagioclase			Remarks
									Rim	Body	Core	
1	2.79	10.5	13.5	49.9	--	--	--	25.7	--	36–43	55	Seriate; grayish-red feldspar; high color index.
2	2.75	11.7	14.2	46.2	--	--	--	27.9	--	32–38	--	Seriate; a few scattered reddish-brown feldspars; no augite.
3	2.75	13.6	14.6	43.1	--	--	--	28.7	30	34–53	--	Seriate; grayish-red plagioclase.
4	2.78	12.3	10.8	52.6	--	--	--	24.3	--	40, 36–53	72	Seriate; scattered moderate red feldspar.
5	2.76	12.8	20.0	44.4	--	--	--	22.8	29	33–47	64	Seriate.
6 [1]	--	16.4	12.5	51.8	11.0	6.1	2.2	19.3	--	38	--	Do.
7	2.77	9.5	11.0	48.5	--	--	--	31.0	--	36	51	Seriate; scattered reddish-brown feldspar.
8	2.75	9.7	10.3	59.0	--	--	--	21.0	31	28–39	--	Seriate; somewhat foliated and coarser grained than usual.
9	2.75	10.8	20.0	44.3	11.0	10.1	3.8	24.9	29	36	45	Seriate; scattered reddish-brown feldspar.
10	2.75	6.9	17.0	51.5	--	--	--	24.5	--	40	51?	Seriate.
11	2.76	13.6	20.1	44.5	--	--	--	21.8	31	37	52	Seriate; somewhat foliated; scattered moderate red feldspar.
12	2.75	10.7	9.7	60.9	--	--	--	18.7	29	31–45	--	Seriate; some moderate red feldspar.
13	2.77	15.2	16.2	54.4	--	--	--	14.2	--	34, 32–36	--	Seriate; some moderate red feldspar; specimen 6 inches from contact with Lamarck granodiorite.
14	2.68	16.0	24.3	47.4	--	--	--	12.3	26	32–37	--	Seriate; somewhat foliated.
15	2.74	18.0	24.3	49.9	--	--	--	18.8	28	31–36	47–55, 50	Do.
16	2.75	20.3	20.4	38.7	--	--	--	20.6	29	34	48	Seriate.
Average	2.75	13.1	15.6	49.3	--	--	--	22.0	--	--	--	

[1] *Standard chemical analysis, specimen 6*

[Lab. No. 53-1303scd. Analyst: L. M. Kehl]

	Weight percent		Weight percent
SiO_2	61.00	H_2O-	0.05
Al_2O_3	16.06	H_2O+	.58
Fe_2O_3	1.86	TiO_2	.88
FeO	4.06	CO_2	.00
MgO	3.10	P_2O_5	.25
CaO	5.46	MnO	.10
Na_2O	3.45	BaO	-----
K_2O	2.95		
		Total	99.80

Norm, specimen 6

	Weight percent	
Q	13.32	
or	17.27	
ab	29.32	(Plagioclase composition An_{40})
an	19.46	
di	4.73	
hy	10.00	
mt	2.72	
il	1.67	
ap	.64	
Total	99.13	

are differentiates of the quartz monzonite. Many fragments of granodiorite have been recrystallized to finer textured rock (Joplin, 1935b), and locally the mafic minerals have collected into knots and layers much like those in pelitic hornfels at Stecker Flat (fig. 9). The presence of differentiates of the quartz monzonite similar to the Cathedral Peak granite as cementing material places the time of brecciation as not later than intrusion of the quartz monzonite, and localization of the breccia zone in the bend in the intrusive contact suggests strongly that the brecciation was caused by intrusion of the quartz monzonite.

Some facies of the granodiorite are similar in composition and appearance to rock in the Deep Canyon area of the Tungsten Hills, which has been classified as quartz diorite. The granodiorite is also similar to some facies of the Sentinel granodiorite of the Yosemite region (Calkins, 1930, p. 125), but the Sentinel granodiorite also includes facies that show equal similarity to the Lamarck and Tinemaha granodiorites.

TINEMAHA GRANODIORITE

The Tinemaha granodiorite has an aggregate outcrop area in the south half of the Big Pine quadrangle of approximately 32 square miles (fig. 13). Most of the granodiorite is in a large oval mass containing inclusions of diorite, nearly bisected by the granodiorite of McMurry Meadows. The western part crops out continuously in about 18½ square miles, whereas the eastern part crops out discontinuously in about 13½ square miles. The largest exposure in the eastern part, an outcrop area of about 12 square miles, is separated from several small outcrops, that project through the basalt of Crater Mountain. A small outcrop having an area of half a square mile lies south of Red Mountain Creek in the vicinity of Stecker Flat.

The type locality is along Tinemaha Creek, and rock of typical appearance crops out in all the canyons between Tinemaha Creek and the South Fork of Big Pine Creek. Granodiorite of typical appearance also occurs in road cuts on the south side of Big Pine Creek near the crossing at Bench Mark 5066, but the rock is more weathered than that exposed higher in the range.

The appearance of the Tinemaha granodiorite is very nearly the same in all exposures throughout the mass; differences in texture and in color index do not materially affect the general appearance of the rock. Adjacent to the Inconsolable granodiorite, it is somewhat finer textured than in most other places, but the difference in grain size is not great.

Commonly the granodiorite is porphyritic and contains large subhedral to anhedral grains of perthitic microcline as much as 1½ cm across, although some specimens are equigranular or seriate (fig. 21*A*). The texture of the groundmass is hypidiomorphic-granular, and the grain size generally ranges from 2 to 4 mm. The color index of the rock averages about 14, and ranges from 6 to 25. A characteristic of the Tinemaha granodiorite that is unique among the granitic rocks of the Bishop district is that hornblende generally is in excess of biotite, the average ratio being 4:3. Much of the hornblende is in euhedral or subhedral prisms, whereas biotite, unlike the biotite in the Lamarck granodiorite, is rarely euhedral. Plagioclase in the Tinemaha granodiorite is variable in amount, but generally is more abundant than either K feldspar or quartz. Quartz is a little less abundant than K feldspar. Plagioclase commonly is zoned; the main part of many grains is about An_{40} near the center and about An_{30} near the margins, but some cores are as calcic as An_{50}, and some rims are An_{20} or less. Commonly the zoning is continuous except for minor but conspicuous oscillations. Most of the K feldspar is perthitic microcline. Several thin sections of specimens (locations on fig. 26) from marginal parts of the western limb of the intrusion exhibit augite cores in the hornblende.

The granodiorite contains numerous mafic inclusions, and in most places a foliation is defined by the inclusions and by planar orientation of biotite and hornblende. In the western part of the intrusion, steeply dipping foliation in the west side and gently dipping foliation in the east side adjacent to the granodiorite of McMurry Meadows define the western half of a foliation arch (pl. 5, sec. *E–E'*). Foliation was recorded in the eastern part at only a few places because it is obscured by poor exposure and deep weathering.

The western part of the main mass of Tinemaha granodiorite ends to the south against the septum along Red Mountain Creek. However, rocks similar to those in the septum occur near Stecker Flat and form small septa along a gently dipping contact between the small mass of Tinemaha granodiorite and underlying quartz monzonite similar to the Cathedral Peak granite. The quartz monzonite is underlain in turn by finer grained quartz monzonite. The relations suggest that originally the Tinemaha granodiorite extended farther to the south but has been cut out by younger intrusions or eroded away (pl. 5, sec. *F–F'*).

The Tinemaha granodiorite is cut by many mafic dikes (pl. 4), which generally dip steeply or are vertical and trend N. 70° W. These dikes do not continue in abundance into any of the bordering intrusives, and many terminate approximately at the contacts.

The Tinemaha granodiorite is distinguished from the Wheeler Crest quartz monzonite and the Lamarck granodiorite, by the preponderance of hornblende over

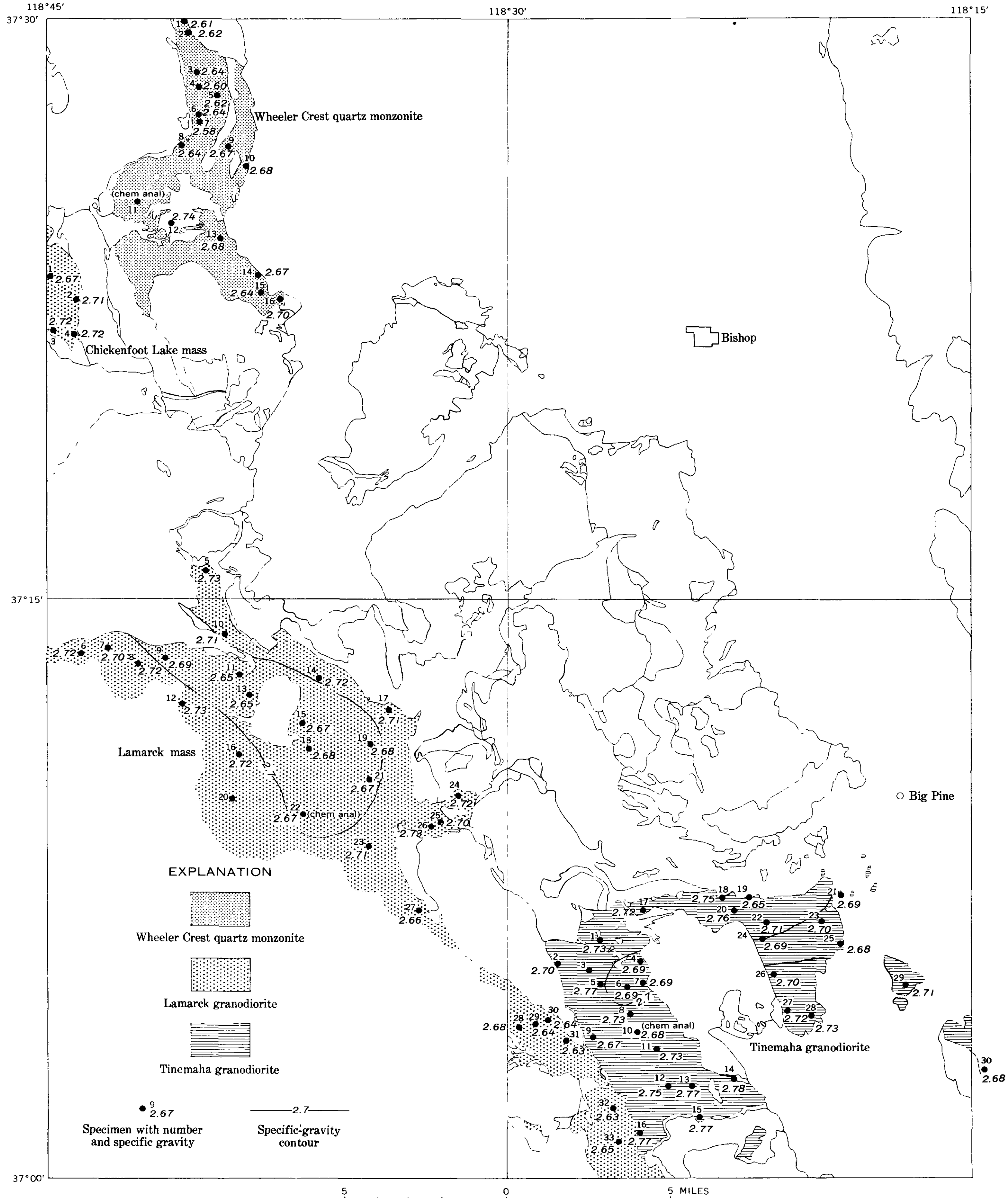


FIGURE 26.—Map showing the locations of modally analyzed specimens of the Wheeler Crest quartz monzonite and of the Lamarck and Tinemaha granodiorites.

biotite in the Tinemaha granodiorite, although the average grain size and color index are about the same in all three rocks (fig. 21). Actually no problem arises in the field in distinguishing the Lamarck granodiorite from the Tinemaha granodiorite because the south end of the Lamarck granodiorite, where it is in contact with the Tinemaha granodiorite, is exceptionally light colored. The Tinemaha granodiorite also can be readily distinguished from the Lamarck by the absence of euhedral biotite plates, which are characteristic of the Lamarck granodiorite, and by its generally porphyritic habit. The anhedral to subhedral habit of the phenocrysts distinguish it from porphyritic facies of the Wheeler Crest quartz monzonite.

The modes and specific gravities of the 29 specimens whose locations are shown in figure 26 are tabulated in table 5 and plotted on a quartz–K feldspar-plagioclase triangular diagram in figure 27. The plot of modes forms a large elongate field that extends away from the plagioclase corner toward the quartz-orthoclase sideline. The elongate shape of the field reflects range in the plagioclase content of more than 35 percent.

The Tinemaha granodiorite is not known to intrude any other plutonic rock except small masses of quartz

TABLE 5.—*Modal analyses of Tinemaha granodiorite, in volume percent*

[Where the content of biotite, hornblende, and accessory and secondary minerals is shown separately, the mode was determined from a thin section. Where only the total content of mafic minerals is shown, the mode was determined from a stained slab. Location of specimens is shown on fig. 26]

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Biotite	Hornblende	Accessory and secondary minerals	Total mafic	Percent of anorthite in plagioclase			Remarks
									Rim	Body	Core	
1	2.73	15.7	17.8	48.6				17.9		39–46		Weakly porphyritic; faint foliation.
2	2.70	13.6	24.7	49.1				12.6		32–36		Weakly porphyritic; some hornblende grains enclose augite cores.
3		15.4	22.7	43.6				18.3		31–46		Porphyritic.
4	2.69	32.4	30.6	22.7				14.3		34–48, 39		Weakly porphyritic.
5	2.77	17.9	18.6	42.8				20.6		39–48		Porphyritic.
6	2.69	22.7	28.6	38.5				10.2	10	21–40		Porphyritic; some intergranular granoblastic mortar and myrmekite.
7	2.69	21.2	27.6	37.2				14.0		23–34		Weakly porphyritic.
8	2.73	25.1	24.0	39.2				11.5		20–46		Porphyritic; some intergranular granoblastic mortar and myrmekite.
9	2.67	25.8	31.4	33.4				9.3	4	40–48	40	Equigranular to seriate; foliated; abundant granoblastic mortar and myrmekite; hornblende encloses augite.
10 [1]	2.68	22.8	17.4	45.8	5.6	7.2	1.3	14.1		38–44	48	Porphyritic; abundant granoblastic mortar and myrmekite.
11	2.73	18.8	25.0	41.0				15.2		20-32	38	Porphyritic; some granoblastic mortar and myrmekite.
12	2.75	17.4	16.3	47.5				18.8		38–44		Equigranular.
13	2.77	24.1	24.4	33.5				18.0		37–43, 38		Porphyritic; some hornblende grains enclose augite.
14	2.78	24.4	12.2	46.3				17.1				Porphyritic; some hornblende grains enclose augite. (Mode poor.)
15	2.77	17.0	23.8	51.7	4.1	2.5	.8	7.4		33–39, 36		Equigranular; some hornblende grains enclose augite.
16	2.77	11.3	11.6	49.4				27.7		40–44		Equigranular; very mafic; much granoblastic mortar.
17	2.72	18.9	11.3	49.4				20.4	12	21–52		Weakly porphyritic; some granoblastic mortar and myrmekite.
18	2.75	20.5	21.2	40.2				18.1	10, 27	34, 37		Seriate.
19	2.65	31.0	30.4	29.6				9.0		24–36		Porphyritic; some granoblastic mortar.
20	2.76	14.2	13.5	47.9				24.4		22–34		Seriate.
21	2.69	19.6	31.3	37.0				12.1		37–52		Porphyritic; both perthite and plagioclase phenocrysts.
22	2.71	24.3	27.4	30.4				17.9		22–37		Seriate; much epidote.
23	2.70	24.9	27.2	30.7				17.2	21	29–43		Porphyritic; both perthite and plagioclase phenocrysts.
24	2.69	18.3	18.6	42.4				20.1		34?		Porphyritic (Plagioclase determination poor).
25	2.68	27.5	28.5	26.7				17.2		21–46		Seriate.
26	2.70	34.8	22.1	27.8				15.3	21	29–38		Porphyritic.
27	2.72	13.7	34.4	39.7				12.4	22	30–43		Seriate.
28	2.73	27.5	22.5	32.4				17.5		34–46	52	Seriate; abundant quartz.
29	2.71	19.5	33.1	36.9				10.5		33–44		Porphyritic; plagioclase much sericitized; some epidote; mafic mineral content too low.
30	2.68	21.4	33.5	29.3	5.1	3.1	7.7	15.9		28, 33		Porphyritic; highly sericitized plagioclase phenocrysts in hypidiomorphic groundmass.
Average	2.72	21.3	23.6	39.0				16.1				

[1] *Standard chemical analysis, specimen 10*

[Lab. No. 53–1299scd. Analysts: L. M. Kehl]

	Weight percent		Weight percent
SiO_2	62.82	H_2O-	0.03
Al_2O_3	15.44	H_2O+	.62
Fe_2O_3	2.59	TiO_2	.64
FeO	3.17	CO_2	.01
MgO	2.35	P_2O_5	.30
CaO	5.04	MnO	.11
Na_2O	3.15	Total	99.99
K_2O	3.72		

Norm, specimen 10

	Weight percent	
Q	16.69	
or	22.20	
ab	26.71	(Plagioclase composition An_{38})
an	16.68	
di	5.16	
hy	6.18	
mt	3.74	
il	1.22	
ap	.69	
Total	99.27	

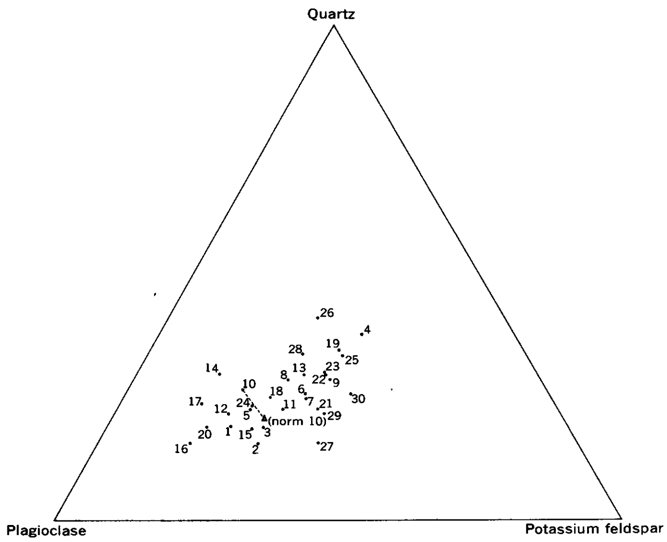


FIGURE 27.—Plot of modes of Tinemaha granodiorite on quartz-K feldspar-plagioclase diagram.

diorite and hornblende gabbro, but it probably is intrusive into the Inconsolable granodiorite. Inclusions of the Tinemaha granodiorite in the Lamarck granodiorite and dikes of Lamarck granodiorite and of quartz monzonite similar to the Cathedral Peak granite in the Tinemaha granodiorite indicate that the Tinemaha granodiorite is older than these intrusives. The Tinemaha granodiorite is also intruded by the granodiorite of McMurry Meadows. The position of the granodiorite of McMurry Meadows relative to the Tinemaha granodiorite and certain similarities of mineral content and texture suggest that the granodiorite of McMurry Meadows crystallized from the core magma of the Tinemaha granodiorite following crystallization of the margins and movement of the still-liquid core.

The Tinemaha granodiorite megascopically resembles the Wheeler Crest quartz monzonite, which like the Tinemaha granodiorite is apparently older than any of the plutonic intrusives with which it is in contact, except for masses of quartz diorite and hornblende gabbro. Although correlation of these two rocks is an attractive possibility, differences in the quartz-K feldspar ratio, in the size and shape of the perthitic microcline phenocrysts, and the preponderance of hornblende over biotite in the Tinemaha granodiorite suggest they are different rocks. Porphyritic granodiorite somewhat similar in appearance to the Tinemaha granodiorite also is found in the Benton Range, 30 miles north of Bishop, and at the Jumbo mine in the Inyo Range west of Independence. The granodiorite of the Benton Range contains more hornblende than biotite and is porphyritic, but the K feldspar phenocrysts are euhedral rather than subhedral as in the Tinemaha granodiorite.

GRANODIORITE OF McMURRY MEADOWS

The granodiorite of McMurry Meadows constitutes a single pluton that is intrusive into and enclosed by the Tinemaha granodiorite. It exhibits the greatest range in composition of any plutonic mass within the mapped area (fig. 28; table 6). The pluton is concentrically zoned; it grades from quartz monzonite in the core to granodiorite in the margins. The southwest margin is more mafic than other parts and approaches quartz diorite in composition. Specific gravities of modally analyzed specimens range from 2.62 to 2.69 in the quartz monzonite core and from 2.69 to 2.82 in the granodiorite rim. (See fig. 23 for specimen locations and table 6 for specific gravities.) Modal analyses show that the plagioclase content ranges from 24 to 64 percent (fig. 28; table 6).

The cause of the strong zonation has not been established. The presence of small masses of older mafic rocks around the margins suggests the possibility of contamination and hybridization by these rocks. Contamination of other intrusives, for example the southern part of the Chickenfoot Lake mass of the Lamarck granodiorite, by earlier mafic rock is demonstrable. However, this hypothesis seems incompatible with the presence of almost uncontaminated quartz monzonite in the southern part of the granodiorite of McMurry Meadows adjacent to large inclusions of hornblende gabbro. Many other plutons in the mapped area exhibit systematic compositional zoning, though none through as wide a range as the granodiorite of McMurry Meadows, and several lack mafic rocks around their margins. The best explanation for zoning of intrusive rocks where hybridization cannot be invoked is differentiation during cooling.

Hand specimens of rock of quartz monzonite composition from the pluton usually can be distinguished by their texture and mineral content from those of granodiorite composition. Commonly those of quartz monzonite composition are porphyritic, containing conspicuous phenocrysts of microcline perthite and not more than 1.5 percent of hornblende. Specimens of granodiorite composition, on the other hand, are equigranular and contain as much as 10 percent hornblende. Furthermore, much of the hornblende in the granodiorite contains augite cores, whereas no augite was found in the hornblende of the quartz monzonite. An exception is specimen 9 (the chemically analyzed specimen), which on the basis of the modal content of plagioclase and K feldspar would be classed as calcic quartz monzonite, but which has an equigranular texture identical with that of the granodiorite specimens and contains hornblende having augite cores.

Although the general appearance of the rocks suggests that the presence or absence of microcline perthite phenocrysts determines whether the rock is quartz monzonite or granodiorite, the plot of modes in figure 28 indicates quite clearly that difference in the amount of plagioclase is of greater significance. The plagioclase in both granodiorite and quartz monzonite is deeply zoned though about the same in compositional range. The central body generally is zoned within the limits of An_{30} to An_{40}, although the range in some specimens is as great as An_{20} to An_{50}. Rims as sodic as An_{20} and tiny cores as calcic as An_{50} to An_{70} are present locally.

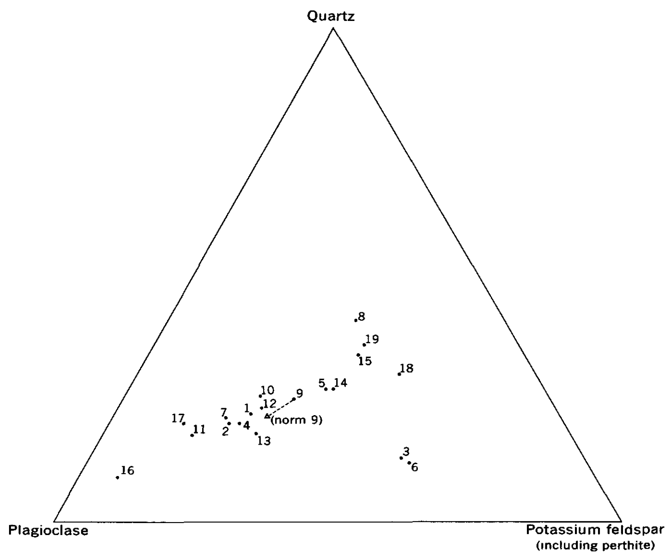


FIGURE 28.—Plot of modes of the granodiorite of McMurry Meadows on quartz-K feldspar-plagiocase diagram.

Two kinds of porphyritic texture are exhibited by the quartz monzonite, a common kind in which the phenocrysts are of perthitic microcline, and a less common kind in which quartz, plagioclase, and K feldspar phenocrysts occur in a granoblastic groundmass of the same minerals plus biotite and accessory minerals. As with the other granitic rocks, the K feldspar phenocrysts appear to have formed late, because they enclose plagioclase, hornblende, and magnetite and in a few places penetrate in thin veinlets into quartz. On the other hand, textural relations indicate that in rock that contains phenocrysts of plagioclase and quartz as well as K feldspar, the phenocrysts crystallized earlier than

TABLE 6.—*Modal analyses of the granodiorite of McMurray Meadows, in volume percent*

[Location of specimens is shown on fig. 23]

Specimen	Specific gravity	Quartz	K feldspar	Plagio-clase	Biotite	Horn-blende and Augite	Accessory and secondary minerals	Percent anorthite in plagioclase: Rim	Body	Core	Remarks
1	2.70	20.3	23.0	50.4	4.2	1.4	0.7	22, 23	32, 35	----------	Equigranular; augite cores in hornblende.
2	2.74	16.3	17.3	47.8	10.1	6.6	1.9	18, 25	30, 36	40, 44	Do.
3	2.66	12.0	49.6	29.3	5.5	----------	----------	20, 24	27, 30	----------	Porphyritic; large and abundant microcline perthite phenocrysts.
4	2.74	16.0	17.7	45.2	13.0	6.2	1.8	23, 29	21, 30	45	Equigranular; augite cores in hornblende.
5	2.65	24.9	31.9	34.2	5.9	1.3	1.8	4, 21	30, 36	----------	Porphyritic; large microcline perthite phenocrysts.
6	2.72	11.4	55.5	30.8	1.6	----------	0.7	----------	21	----------	Equigranular, but with abnormally high content of microcline perthite.
7	2.74	16.5	15.8	47.3	10.3	7.9	2.2	30, 33	33, 35	----------	Equigranular; Augite cores in hornblende.
8	2.64	38.1	30.0	23.5	4.9	1.5	1.8	15, 19	35, 37	----------	Porphyritic; large microcline perthite phenocrysts.
9 [1]	2.68	23.5	27.6	41.0	4.3	2.5	1.1	23, 27	32, 37	40	Equigranular; augite cores in hornblende.
10	2.69	21.3	19.9	42.0	11.9	2.9	1.9	21, 27	33, 37	42, 45	A few large perthite phenocrysts; hornblende contains augite cores.
11	2.72	15.1	13.4	57.3	9.4	4.7	0.1	17, 29	29, 40	----------	Equigranular; augite cores in hornblende.
12	2.71	20.1	21.7	44.6	7.8	3.9	2.0	17, 20	21, 25	----------	Do.
13	2.75	14.2	21.8	43.3	14.9	4.1	1.7	20, 23	35	----------	Do.
14	2.71	23.9	31.8	32.4	9.9	----------	2.0	9, 21	25, 45	60, 79	Porphyritic; large microcline perthite phenocrysts.
15	2.60	31.5	33.7	26.6	5.4	0.4	2.4	18, 29	35, 36	57	Porphyritic; plagioclase, perthite, and quartz in two generations; groundmass allotriomorphic.
16	2.82	6.9	5.1	64.4	9.4	10.0	4.2	33	42	----------	Equigranular; high color index; hornblende contains augite cores; high plagioclase content gives rock subpanidiomorphic texture.
17	2.77	17.3	10.7	56.3	8.0	5.9	1.8	30, 36	40, 56	----------	Equigranular; high color index; hornblende contains augite cores.
18	2.63	27.2	41.6	21.3	7.7	----------	2.1	13, 25	34	----------	Porphyritic, plagioclase quartz, and perthite in two generations; groundmass allotriomorphic.
19	2.62	32.3	33.9	24.9	7.2	0.5	1.1	20, 30	32, 36	----------	Do.
Average	2.70	20.5	26.4	40.1	8.0	3.1	1.6	----------	----------	----------	

[1] *Standard chemical analysis, specimen 9*

[Lab. No. 53-1301scd. Analyst: L. M. Kehl]

	Weight percent		*Weight percent*
SiO_2	64.86	H_2O-	0.06
Al_2O_3	16.12	H_2O+	.51
Fe_2O_3	1.90	TiO_2	.57
FeO	2.52	CO_2	.00
MgO	1.55	P_2O_5	.23
CaO	3.80	MnO	.09
Na_2O	3.44	BaO	.13
K_2O	4.03		
		Total	99.81

Norm, specimen 9

	Weight percent	
Q	18.90	
or	23.91	
ab	28.82	(Plagioclase composition An_{37})
an	16.68	
di	.36	
hy	5.91	
mt	2.78	
il	1.07	
ap	.62	
Total	99.05	

the accompanying allotriomorphic granular groundmass. Even in such rocks, however, the marginal parts of K feldspar phenocrysts not uncommonly enclose rounded quartz grains, a relation which indicates that the phenocrysts continued to grow during the crystallization of the granoblastic groundmass.

The granodiorite of McMurry Meadows is in contact with only two intrusives other than hornblende gabbro. It is intrusive into the Tinemaha granodiorite and is intruded by dikes of quartz monzonite similar to the Cathedral Peak granite. Its position relative to the Tinemaha granodiorite suggests that it crystallized from the core magma of the Tinemaha granodiorite following magmatic movement that caused the intrusive contact between the two formations.

WHEELER CREST QUARTZ MONZONITE

The Wheeler Crest quartz monzonite is in a single mass that crops out principally in the steep eastern face of Wheeler Crest. It underlies a little more than 17 square miles within the mapped area and extends from the north edge of the Mount Tom quadrangle south across Pine Creek into the lower northeastern slope of Mount Tom. Typical quartz monzonite is exposed on both sides of the entrance to Pine Creek Canyon, and this area can be considered the type locality.

Most of the quartz monzonite contains conspicuous phenocrysts of potassium feldspar, which are set in a medium-grained hypidiomorphic-granular groundmass (fig. 21E). The groundmass minerals commonly are 2 to 4 mm across, and the phenocrysts, many of which are euhedral tabular crystals, average about half an inch in thickness and range from 1 to several inches in maximum dimension. Fresh surfaces are light gray, and the color index averages about 12, but ranges from a little less than 5 to 18 (table 7). Porphyritic facies are distinctive in appearance and superficially resemble only porphyritic facies of the quartz monzonite similar to the Cathedral Peak granite, which, however, is a much more felsic rock, and the Tinemaha granodiorite, which is more mafic. By decrease in the abundance of phenocrysts, the porphyritic rock grades into equigranular rock which has a texture identical with that in the groundmass of porphyritic rock. The dark minerals, biotite and hornblende, are evenly scattered in small clusters through the rock, but in concentrations along the margins of phenocrysts of K feldspar. Individual grains are small, generally less than 1 mm across, and anhedral.

Locally, the quartz monzonite contains irregular, fine-grained, dark-colored aggregates as much as an inch in the greatest dimension, which consist chiefly of biotite plates and lesser amounts of accessory minerals. These aggregates are thought to be small inclusions of schist or pelitic hornfels.

In most places the quartz monzonite has a primary foliation that is marked by planar orientation of ovoid clots of mafic minerals and by lenticular mafic inclusions. In a few places a secondary gneissic foliation is shown by layers of hornblende and biotite that lie along closely spaced shears.

Modal analyses were made of 16 specimens, and one specimen was analyzed chemically. The plot of the modes on the triangular diagram (fig. 29) shows that the ratio of quartz to perthitic K feldspar is variable but averages about 1 to 1, and that the content of plagioclase is variable.

Potassium feldspar is present as anhedral grains in the groundmass as well as in phenocrysts. Albite is regularly distributed in wavy streaks of about the same size and density in all the crystals of K feldspar; this distribution indicates that the albite was exsolved from the host crystal. Both Carlsbad twinning and the quadrille structure characteristic of microcline are present in some crystals, but many exhibit neither. The plagioclase in most specimens is zoned andesine having an average composition of about An_{32}, but the plagioclase in specimens from the northernmost part is more sodic. Many individual grains are zoned through determinable ranges of as much as 12 percent An. The plagioclase generally is more sodic toward the margin, but in some crystals minor oscillations are superimposed on the general trend. In most specimens the plagioclase is subhedral, but in specimen 12 (table 7), which is from a dike, much of the plagioclase is euhedral and the texture of the rock is panidiomorphic granular.

Both hornblende and biotite are notably ragged as seen in thin section, and the hornblende commonly encloses many tiny rounded plagioclase grains. Minor accessory minerals are magnetite and ilmenite, apatite, sphene, allanite, and zircon. Common alteration products are chlorite and epidote.

Most of the specimens of the quartz monzonite that were studied contain abundant evidence of cataclasis and recrystallization. This evidence includes fractured and dislocated grains, fine granoblastic mortar between many grains and across some, extensive myrmekite along boundaries between perthite and plagioclase, granoblastic texture in quartz masses that appear to pseudomorph primary quartz crystals, and undulatory extinction and strain shadows in quartz that has not been reduced to granoblastic aggregates. Many boundaries between primary grains are occupied by a fine granoblastic mortar of quartz and feldspar together with small amounts of the mafic and accessory minerals.

735-925 O—65——6

Some zones of granoblastic mortar cut across individual crystals. The granoblastic mortar zones commonly coalesce in such a way as to show that the rock has been cut by through-going shears. In some sheared rocks the segregation of biotite and hornblende along the shears, presumably by a process of metamorphic differentiation, gives these rocks a secondary gneissic folation.

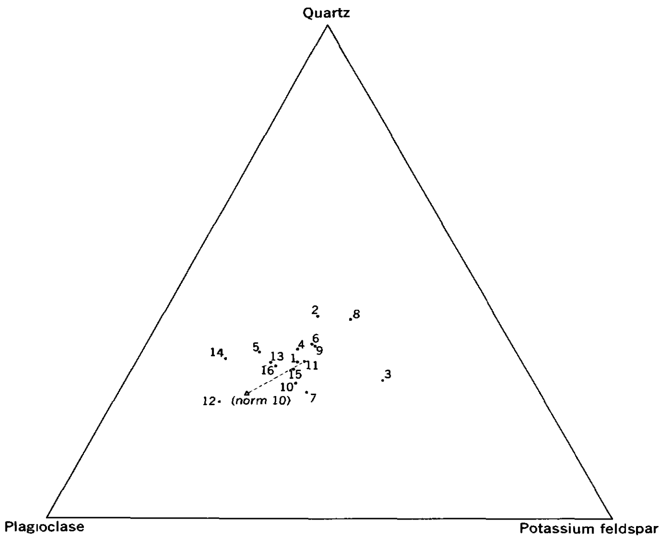


FIGURE 29.—Plot of modes of Wheeler Crest quartz monzonite on quartz-K feldspar-plagioclase diagram. Modes were determined from stained rock slabs.

Boundaries between grains of plagioclase and K feldspar are more commonly occupied by vermicular myrmekite than by granoblastic mortar, but the continuity of myrmekite with zones of granoblastic mortar indicates that cataclasis along grain boundaries was a factor in its origin. The presence of "cats paws" of myrmekite that embay the contiguous perthite indicate that the myrmekite formed, in accordance with classical theory, by replacing potassium feldspar (Tyrrell, 1929, p. 94).

The areas of granoblastic quartz probably are products of the recrystallization of fractured primary quartz crystals, but some masses of granoblastic quartz may have been precipitated from solutions. Strain shadows in quartz are common. In most specimens the strain shadows are in a polygonal pattern, but in some specimens they are in the form of wavy lines that are not only parallel within individual grains but which are also parallel throughout a thin section. A few quartz grains exhibit two sets of shadowy lines that intersect at a wide angle and which superficially resemble microcline twinning. Locally the acute angle between the two sets is bisected by fractures that are sealed with

TABLE 7.—*Modal analyses of Wheeler Crest quartz monzonite, in volume percent*

[All modes were determined from stained slabs. Location of specimens is shown on fig. 26]

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Total mafic	Percent of anorthite in plagioclase			Remarks
						Rim	Body	Core	
1	2.61	30.7	27.8	38.9	2.6				
2	2.62	39.4	26.4	30.2	4.0	4	14–20		Porphyritic; phenocrysts smaller than in typical rock; rock unusually felsic.
3	2.64	24.7	39.8	23.4	12.0				
4	2.60	30.7	24.6	34.7	10.0				
5	2.62	30.9	19.0	41.4	8.6	15	22–27		Equigranular; foliated; finer grained than usual and unusually felsic.
6	2.64	33.0	27.0	32.5	7.5		24–33		Somewhat porphyritic.
7	2.58	22.3	33.4	31.6	12.7		25–32		Porphyritic; quartz is in ovoids about ¼ inch across; notably cataclastic.
8	2.64	36.6	30.8	23.4	9.2	23	25–32		Porphyritic; some cataclasis, not severe; aggregates of biotite, sphene, and magnetite.
9	2.67	32.5	27.5	32.2	7.8		27–33		Porphyritic, but phenocrysts anhedral; notably cataclastic.
10	2.68	25.2	27.8	39.0	8.0		32–38		Equigranular; moderate cataclasis; finer grained than typical rock.
11 [1]		30.5	28.4	36.3	4.8	22	33, 38		Porphyritic; some granoblastic mortar, but not in abundance.
12	2.74	18.8	14.8	45.5	20.9	15	30–39		Panidiomorphic granular with finer grain than usual; a dike rock.
13	2.68	27.8	20.4	39.3	12.5	24, 25	32–40		Porphyritic; extensive cataclasis.
14	2.67	30.4	14.3	48.5	10.5	21, 23	28, 32, 35		Do.
15	2.64	26.9	25.2	36.4	12.2		22–40		Do.
16	2.70	26.2	21.2	37.2	15.4	24, 28	32		Equigranular; extensive cataclasis.
Average	2.65	29.3	25.3	35.3	10.1				

[1] *Standard chemical analysis, specimen 11*

[Lab. No. 53-1295scd. Analyst: L. M. Kehl]

	Weight percent		*Weight percent*
SiO_2	71.42	H_2O-	0.06
Al_2O_3	14.47	H_2O+	.21
Fe_2O_3	1.03	TiO_2	.25
FeO	1.38	CO_2	.03
MgO	.78	P_2O_5	.10
CaO	2.86	MnO	.08
Na_2O	3.44	BaO	-----
K_2O	3.69		
		Total	99.8

Norm, specimen 11

	Weight percent	
Q	29.94	
or	21.71	
ab	28.85	(Plagioclase composition $An_{31.6}$)
an	13.35	
hy	3.34	
mt	1.46	
il	.47	
ap	.27	
Total	99.39	

secondary quartz of a different orientation than that in the enclosing grain.

The Wheeler Crest quartz monzonite is intruded by all the plutonic rocks with which it is in contact, except quartz diorite and hornblende gabbro, and is the oldest major intrusive in the area north of Big Pine Creek. Finer grained quartz monzonite and the Round Valley Peak granodiorite send dikes into the Wheeler Crest. Marginal to the dike of alaskite similar to the Cathedral Peak granite, which extends along the east face of Wheeler Crest and clearly intrudes the Wheeler Crest quartz monzonite, are many thin fine-grained felsic dikes that dip gently toward the main dike. These satellitic dikes appear as thin white lines when seen from U.S. Highway 395 along Sherwin Grade in the forenoon, when the sun lights up the eastern face of Wheeler Crest.

Evidence that the Tungsten Hills quartz monzonite intrudes the Wheeler Crest quartz monzonite was found on the northwest side of Mount Tom, on the west side of Elderberry Canyon. The contact there is sharp but featureless; it lacks dikes and other conspicuous diagnostic features, although a few angular inclusions of Wheeler Crest quartz monzonite a few feet across were found close to the contact in Tungsten Hills quartz monzonite. Within 2 to 4 feet of the contact the Tungsten Hills quartz monzonite is finer grained than usual and has a mottled appearance caused by small rock fragments and phenocrysts of Wheeler Crest quartz monzonite. The Wheeler Crest quartz monzonite is cut by numerous shear zones that are generally about parallel with the contact, and along some of these the rock has been reduced to an augen gneiss. Shear zones were observed half a mile or more away from the contact, but they are increasingly abundant as the contact is approached. These relations suggest that the shearing resulted from the emplacement of the Tungsten Hills quartz monzonite.

The Wheeler Crest quartz monzonite resembles quartz monzonite similar to the Cathedral Peak granite where both rocks are porphyritic. However, the Wheeler Crest is darker colored and contains tabular phenocrysts of K feldspar, whereas phenocrysts in porphyritic quartz monzonite similar to the Cathedral Peak granite characteristically are more equant. The younger age of the quartz monzonite similar to the Cathedral Peak granite is clearly indicated by the fact that it intrudes the Round Valley Peak and Tungsten Hills quartz monzonites, both of which are younger than the Wheeler Crest quartz monzonite.

The Wheeler Crest quartz monzonite also resembles the Tinemaha granodiorite, but is more silicic and its phenocrysts are more nearly euhedral. Nevertheless, they may be temporal equivalents, for each rock is the oldest intrusive in its area except for quartz diorite and hornblende gabbro.

ROUND VALLEY PEAK GRANODIORITE

The Round Valley Peak granodiorite is represented within the mapped area by a single mass, which lies in the northwest corner of the Mount Tom quadrangle and extends into the adjoining three quadrangles. It underlies a little more than 18 square miles within the mapped area, and its total area is about 40 square miles. Good exposures can be examined along both sides of upper Rock Creek. The intrusive was mapped in the Casa Diablo quadrangle as granodiorite of Rock Creek by Rinehart and Ross (1957), and this usage was followed in the north half of the Mount Abbot quadrangle by Sherlock and Hamilton (1958). Adoption of a formal name became necessary when a correlative mass was found farther to the northwest in the Mount Morrison quadrangle by Rinehart and Ross (1964, p. 44–47). Because Rock Creek has long been preempted in formal stratigraphic nomenclature, the name Round Valley Peak granodiorite was given the rock (Bateman, 1961); Round Valley Peak is a high point along Wheeler Crest, which is composed of granodiorite of typical appearance.

The granodiorite is notably equigranular and medium grained (fig. 21*D*). Both biotite and hornblende are evenly distributed in discrete euhedral crystals that give the rock a distinctive "tidy" look. The average grain size is about 3 mm, but the grain size is about 2 mm in the eastern part, and 3 to 4 mm in the southwestern part. The feldspars are white and quartz light gray, and the rock has an overall light-gray hue. The average color index of 13 modally analyzed specimens is 12.7, but the finer grained rock in the eastern part is darker than coarser grained rock in the southwestern part (table 8). The specific gravities of the 13 specimens range from 2.65 to 2.73; samples from the darker and finer grained eastern part are heavier, mostly above 2.7, whereas those from the southwestern part are less than 2.7.

Foliation parallel to contacts with older rocks is pronounced near the eastern and southern contacts, and is progressively less conspicuous away from them. The foliation is shown best by mafic inclusions, which appear in outcrop to be from a few inches to 2 feet across and from less than an inch to several inches thick at the middle. The inclusions decrease in abundance and are progressively less flattened with distance from contacts with older rocks. Near contacts with older rocks the foliation is also shown by orientation of the mafic minerals, but granodiorite more than a few hundred feet from these contacts appears structurally isotropic.

The modes group rather closely on a quartz-K feldspar-plagioclase diagram (fig. 30). The average mode is about 27 percent quartz, 19 percent K feldspar, 44 percent plagioclase, and 10 percent mafic minerals. The small content of K feldspar probably is partly the result of the relatively large amount of biotite (about 7 percent) in the rock, which contains about two-thirds as much K_2O as K feldspar.

In thin section the texture of the granodiorite can be seen to be hypidiomorphic granular. No mortar structure or other evidence of cataclasis was observed. Biotite and hornblende are in euhedral to subhedral crystals; plagioclase is subhedral; and K feldspar and quartz are anhedral.

The body of most plagioclase crystals generally is unzoned or only slightly zoned in the compositional range of An_{36} to An_{38}. However, specimens 12 and 13 (table 8) are of composition An_{32} and An_{33}. Small cores of about An_{45-50} are present in some crystals, and most crystals are bordered by more sodic plagioclase that is zoned from An_{28} on the inside to An_{21} on the outside. Undulatory extinction in quartz is rare. Hornblende is of the ordinary kind.

The Round Valley Peak granodiorite is clearly younger than the Wheeler Crest quartz monozonite and older than quartz monzonite similar to the Cathedral Peak granite. Dikes of Round Valley Peak granodiorite intrude the Wheeler Crest quartz monzonite, and dikes of quartz monzonite similar to the Cathedral Peak granite intrude the Round Valley Peak granodiorite. Intrusive relations with the Tungsten Hills quartz monzonite are less certain, although the Tungsten Hills quartz monzonite is believed to be the younger. At the contact, older hornblende gabbro, Round Valley Peak granodiorite, and Tungsten Hills quartz monzonite are all involved in a mixed zone. In a few places, dikes of probable Tungsten Hills quartz monzonite intrude the Round Valley Peak granodiorite.

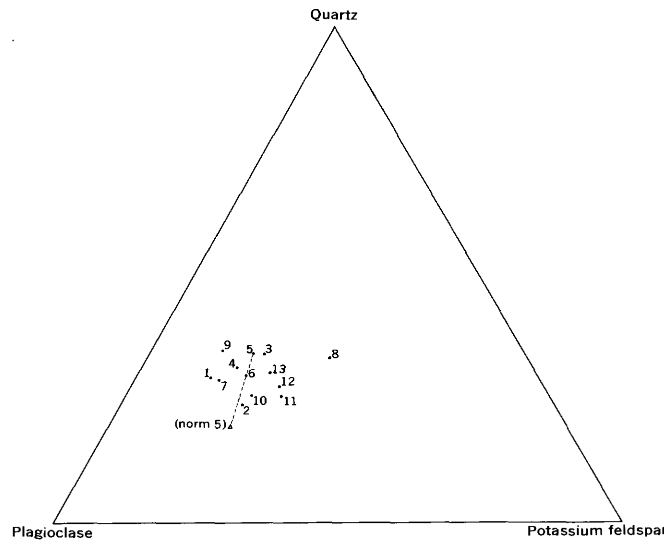


FIGURE 30.—Plot of modes of the Round Valley Peak granodiorite on a quartz-K feldspar-plagioclase diagram.

TABLE 8.—*Modal analyses of Round Valley Peak granodiorite, in volume percent*

[All modes were determined from stained slabs. Location of specimens is shown on fig. 23]

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Total mafic	Percent of anorthite in plagioclase			Remarks
						Rim	Body	Core	
1	2.73	25.1	11.2	49.0	14.7	21–30	36		Equigranular hypidiomorphic; typical of north and east part of mass.
2	2.71	21.2	19.1	48.6	11.1	21–28	37	45	Equigranular hypidiomorphic, typical.
3	2.70	31.2	18.4	42.0	8.4	21–27	38		Do.
4	2.67	28.4	15.4	47.4	8.8	22–30	38	49	Do.
5 [1]	2.71	28.3	16.1	39.9	15.7	22–30	43	47	Do.
6	2.68	26.2	16.7	44.9	12.2	21–28	36		Do.
7	2.70	26.8	13.4	50.1	10.7	21–26	34		Do.
8	2.65	30.3	29.5	32.0	8.2	30	38		Equigranular hypidiomorphic; slightly coarser and more felsic than above.
9	2.68	31.2	11.2	47.0	10.6	29	37		Do.
10	2.69	24.0	20.9	48.5	6.6	24–29	37	45	Do.
11	2.68	23.4	25.5	43.2	7.9	30	38		Do.
12	2.66	24.9	23.8	41.8	9.5	21–25	32	49	Do.
13	2.66	27.7	21.7	42.9	7.7	21–25	33		Do.
Average	2.69	26.8	18.7	44.4	10.1				

[1] *Standard chemical analysis, specimen 5*

[Lab. No. 53-1302scd. Analyst: L. M. Kehl]

	Weight percent		Weight percent
SiO_2	63.53	H_2O-	0.04
Al_2O_3	15.61	H_2O+	.61
Fe_2O_3	2.35	TiO_2	.63
FeO	3.25	CO_2	.03
MgO	2.54	P_2O_5	.23
CaO	4.58	MnO	.12
Na_2O	3.31	BaO	-----
K_2O	2.98	Total	99.81

Norm, specimen 5

	Weight percent	
Q	15.84	
or	17.76	
ab	32.42	(Plagioclase composition An_{33})
an	16.13	
di	5.44	
hy	6.55	
mt	3.43	
il	1.21	
Total	98.78	

In composition and intrusive relations, the Round Valley Peak granodiorite is similar to the Lamarck granodiorite, and the two possibly should be considered the same formation. The Lamarck granodiorite is generally somewhat coarser grained than the Round Valley Peak granodiorite and in some parts is subporphyritic to porphyritic. However, the Lamarck mass of the Lamarck granodiorite constitutes a much larger body than the Round Valley Peak granodiorite, and the difference in the sizes of the masses may explain the differences in grain size and texture.

LAMARCK GRANODIORITE

The Lamarck granodiorite is in two masses that lie along the west and southwest sides of the mapped area. The larger one, called the Lamarck mass, underlies about 54 square miles within the mapped area. The smaller one, called the Chickenfoot Lake mass following the usage of Sherlock and Hamilton (1958) in the adjoining Mount Abbot quadrangle, underlies only 2.7 square miles of the area discussed in this report. Both masses extend west or southwest beyond the limits of the mapped area. The type locality is in the cirques east of Mount Lamarck, but typical rock crops out in all the cirques at the head of Bishop Creek. The mapped part of the granodiorite is chiefly within the Mount Goddard quadrangle, and extends into the southwest corner of the Big Pine quadrangle. The Lamarck mass is elongate and trends northwestward. Even though the contact on the southwest side was mapped only near the south end, it is obvious that the mass is widest near the north end and that it thins southward. In the Bishop Creek drainage it is at least 6 miles wide whereas in the Big Pine quadrangle, where the contacts on both sides are exposed, it averages little more than a mile wide.

The Chickenfoot Lake mass is elliptical, the longer axis trending northwestward. The total outcrop area is about 7 square miles, most of which lies within the Mount Abbot quadrangle to the west of the mapped area. Although this mass is correlated with the Lamarck mass, the correlation is by no means certain, because the rock is generally finer grained and in the south part is darker colored than typical rock.

The rock in the Lamarck mass appears in the field to be remarkably homogeneous in both composition and texture, except in the thin southeastern part, which is more felsic and appears somewhat porphyritic. The change in color index and texture between the northern and southern parts takes place gradually in the vicinity of Hurd Peak; the rock at Bishop Pass is identical in appearance with rock in the vicinity of the Palisade Lakes. Modal analyses (table 9) show that the northern part (fig. 26) is compositionally zoned and that a broad core area has a lower color index and specific gravity than the margins, though not as low as the southeast part.

Rock typical of the main northern part of the mass can be collected along the roads into any of the forks of Bishop Creek, west of the Bishop Creek pendant. The rock is medium grained and generally seriate; less commonly, it is equigranular (fig. 21*C*). The average grain size is 3 to 4 mm., but K feldspar is in grains as much as a centimeter across in the longest dimension. The seriate texture is especially conspicuous in the thin southeastern part, where in hand specimen the rock appears porphyritic. Biotite and hornblende generally are evenly distributed both in clusters and in discrete crystals. Both minerals commonly are subhedral, but plates of biotite and prisms of hornblende are present. The color index of the whole mass averages 11.5, but ranges from 4.4 to 18.4. The average color index of specimens 27 to 33, from the southeastern part of the mass (table 9), is 6.5 and the average of the northern part is 12.2. Specific gravity varies in accordance with the color index (table 9). The measured range in specific gravity is from 2.63 to 2.73. In the margins of the main northern part of the mass the range is from 2.70 to 2.73; in the core it is from 2.65 to 2.69; and in the southeastern part it is from 2.63 to 2.65 except for specimen 28 which has a specific gravity of 2.68.

Commonly the rock has a conspicuous planar foliation, which is best shown by numerous lenticular mafic inclusions, but which also is shown by planar orientation of biotite and hornblende. The foliation is most conspicuous in the margins of the intrusive; the foliation becomes less conspicuous and the number of mafic inclusions diminishes toward the interior of the intrusion.

West of Chocolate Peak and adjacent to the north end of the Inconsolable granodiorite the Lamarck granodiorite contains clusters and individual plates of biotite half an inch or more across, and this rock is represented separately on the geologic map (pl. 1). The large biotite plates are probably the product of hybridization of the magma with sediment, a hypothesis that is supported by the presence of small unassimilated inclusions of calcareous and siliceous metamorphic rock.

The rock in the Chickenfoot Lake mass is finer grained and in the southern part is much darker than typical rock from the Lamarck mass. The finer grain size may be a function of the smaller size of the Chickenfoot Lake mass; the increasing abundance of mafic minerals in the southern part is almost certainly the result of assimilation of hornblends gabbro which lies along the south edge. A sharp contact can be found in most places between the Chickenfoot Lake mass and the hornblende gabbro but only after careful examination of the rocks. The Chickenfoot Lake mass is increasingly contaminated toward the south, and at the contact the rock is almost indistinguishable in hand specimen from the hornblende gabbro. The average color index of four modally analyzed specimens is 16.5, and specimens 3 and 4 from the strongly contaminated part exceed 20.

Thin sections show that the Lamarck granodiorite generally contains about equal amounts of quartz and K feldspar and variable amounts of plagioclase. Although the rock is designated granodiorite, the rock in the southeastern part and much of the core of the northern part of the Lamarck mass is quartz monzonite (table 9). Most thin sections show little evidence of cataclasis.

Quartz is in anhedral grains that have undulatory extinction, but only locally does it comprise a granoblastic mosaic. Commonly the K feldspar is perthitic and shows the quadrille structure of microcline, but some grains show no quadrille structure, and a few are not perthitic. The K feldspar is in interstitial or anhedral grains of small to medium size.

Plagioclase commonly is in subhedral grains that exhibit a wide range of sizes. Most grains are zoned, but some are not. The bodies of plagioclase grains from the Lamarck mass (not including the felsic southern section) are generally zoned in the range of An_{30} to An_{45}. Unzoned plagioclase from the same rock generally has an An content of 38 to 40 percent, figures that agree well with the normative composition of the plagioclase in specimen 9 (table 9) of 37 percent. In the felsic southern part of the Lamarck mass the plagio-

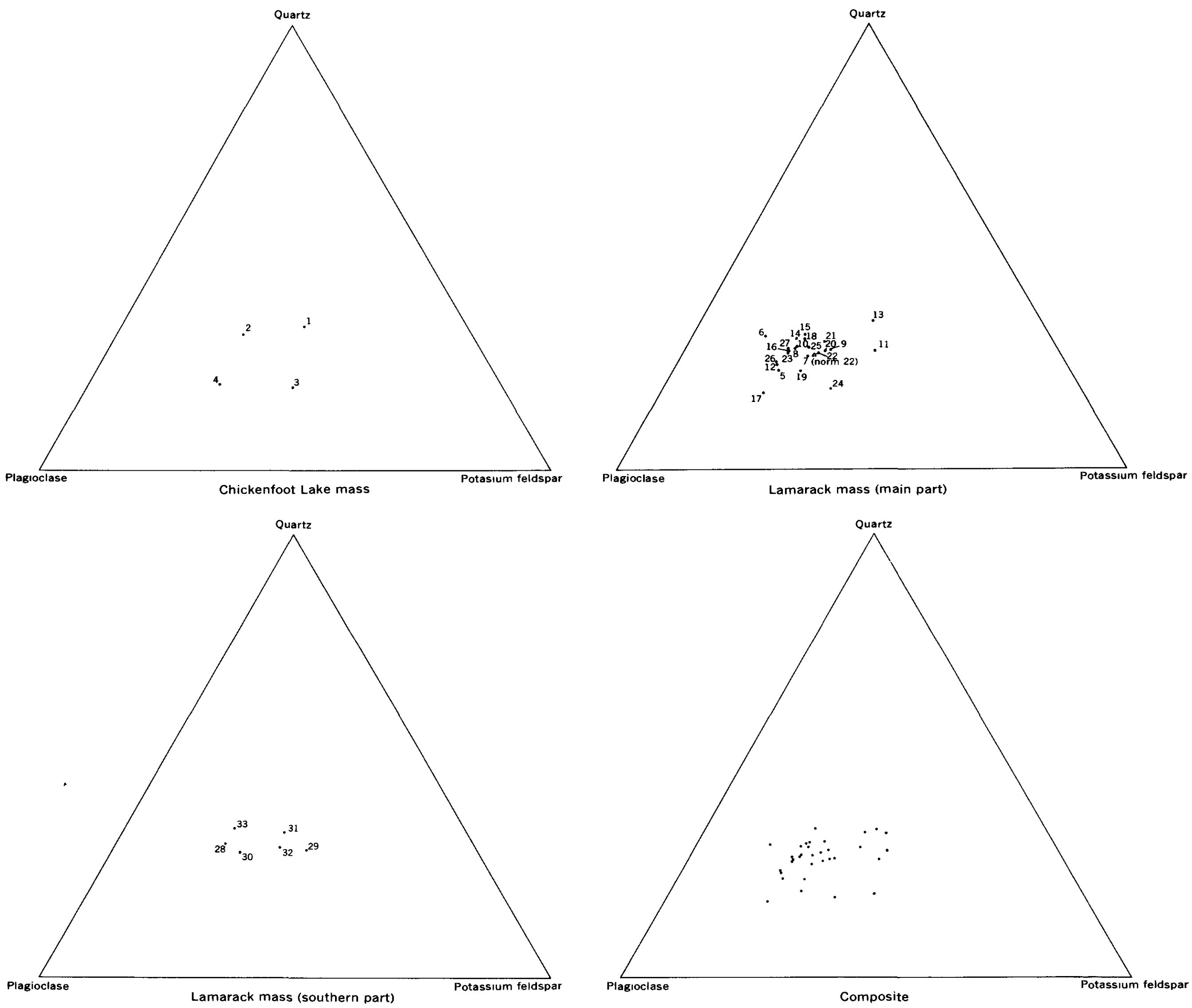


FIGURE 31.—Plots of modes of Lamarck granodiorite on quartz-K feldspar-plagioclase diagrams.

clase is progressively more sodic southward, and the average compositional range in the bodies of grains from the four most southerly specimens (specimens 30 to 33 of table 9) is $An_{17.2}$ to $An_{27.6}$. The plagioclase in specimens from the Chickenfoot Lake mass is erratically zoned, the cores as well as the rims varying widely in composition, but the average is close to that in the main part of the Lamarck mass.

The plot of modes (fig. 31) produces an elliptical field elongate in the direction of the plagioclase corner. The field shows about the same amount of scatter as the plots for the Inconsolable granodiorite and the

TABLE 9.—*Modal analyses of Lamarck granodiorite, in volume percent*

[Where the content of biotite, hornblende, and accessory and secondary minerals is shown separately, the mode was determined from a thin section. Where only the total content of mafic minerals is shown, the mode was determined from a stained slab. Location of specimens is shown on fig. 26]

Specimen	Specific Gravity	Quartz	K feld-spar	Plagio-clase	Biotite	Horn-blende	Acces-sory and second-ary min-erals	Total mafic	Percent of anorthite in plagioclase			Remarks
									Rim	Body	Core	
1	2.67	28.8	31.9	27.9	------	------	------	11.4	------	29–45, 38, 40	------	Porphyritic; a few potash feldspar phenocrysts; hornblende ragged and poikilitic.
2	2.71	25.5	20.9	37.3	------	------	------	16.3	33	39–53, 39	------	Fine grained and equigranular; foliated owing to partial parallel orientation of plagioclase and hornblende.
3	2.72	16.4	34.7	35.3	------	------	------	13.6	34	28–38, 38	------	High mafic mineral content; foliated; seriate; largest crystals are plagioclase.
4	2.72	15.8	20.9	44.3	------	------	------	19.0	------	34–49, 46	------	High mafic mineral content; porphyritic with plagioclase phenocrysts in allotriomorphic groundmass.
Subaverage	2.71	21.6	27.1	36.2	------	------	------	15.1	------	------	------	
5	2.73	19.8	18.3	50.2	------	------	------	11.7	22	38, 40	------	Seriate; large slightly strained quartz grains.
6	2.72	25.3	12.3	43.8	------	------	------	15.5	------	36–43, 38	------	Equigranular; foliated.
7	2.70	22.9	22.2	40.0	------	------	------	10.8	------	31–36, 33	------	Do.
8	2.72	23.0	18.3	43.1	------	------	------	15.6	------	36–44, 37	------	Seriate.
9	2.69	23.1	24.7	37.8	------	------	------	14.4	------	34–43, 38	------	Porphyritic; phenocrysts are plagioclase and potash feldspar; granoblastic mortar is quartz.
10	2.71	24.1	19.1	44.1	------	------	------	12.7	------	32–44	------	Equigranular; foliated; some granoblastic mortar; quartz in granoblastic mosaic.
11	2.65	25.1	35.1	33.8	------	------	------	6.5	------	30–44, 38	------	Equigranular; nonfoliated; some granoblastic mortar; quartz in granoblastic mosaic.
12	2.73	21.5	17.8	51.4	------	------	------	11.3	27	36–45, 40	------	Seriate; large perthite, plagioclase, and quartz grains.
13	2.65	28.9	29.5	28.8	------	------	------	12.8	------	29–46	------	Seriate; large strained quartz grains; some granoblastic mortar and myrmekite.
14	2.72	26.5	18.6	44.6	------	------	------	10.3	27	32–44	45	Seriate; large microcline perthite grains.
15	2.67	29.4	21.2	45.5	------	------	------	8.9	------	26–46, 40	------	Do.
16	2.72	23.5	17.7	45.5	------	------	------	13.7	23	34–40	------	Do.
17	2.71	15.4	17.9	55.9	------	------	------	15.8	------	37–44	------	Seriate.
18	2.68	25.6	19.0	41.4	------	------	------	14.0	21–24	32–45	------	Do.
19	2.68	18.5	20.9	44.4	8.0	6.2	1.8	16.0	7–24	32–44	------	Do.
20	------	22.9	23.9	39.0	------	------	------	14.2	------	37–45	------	Seriate, foliated.
21	2.67	24.9	22.6	38.2	------	------	------	14.3	7	26–45	------	Seriate; large quartz, microcline, and plagioclase grains.
22 [1]	2.67	23.6	24.0	42.5	6.2	3.0	0.5	9.7	20	27–38, 40–46	55	Seriate.
23	2.71	23.3	18.6	47.5	------	------	------	10.6	------	25–39, 33	------	Seriate; quartz grains strained in layers.
24	2.72	14.7	26.5	39.2	------	------	------	19.6	------	29–40	------	Porphyritic; finer grained than usual; plagioclase phenocrysts in allotriomorphic granular groundmass.
25	2.70	23.3	20.6	41.4	------	------	------	14.7	------	28–34, 40	------	Seriate; quartz grains strained in layers.
26	2.73	20.3	16.4	47.0	------	------	------	16.3	------	25–40	------	Porphyritic; finer grained than usual; plagioclase phenocrysts; microcline poikilitic.
27	2.66	25.2	18.8	49.0	------	------	------	7.0	------	21–31	------	Seriate.
Subaverage	2.70	23.1	21.0	43.2	------	------	------	12.9	------	------	------	
28	2.68	28.0	19.9	44.6	------	------	------	7.5	------	26–31	------	Seriate to porphyritic; large perthite crystals.
29	2.64	26.6	34.9	30.3	------	------	------	8.2	0	15, 18	------	Porphyritic; large pink microcline perthite phenocrysts.
30	2.64	27.0	24.0	44.4	------	------	------	4.6	------	12–30	------	Equigranular; abundant microcline perthite.
31	2.63	31.2	30.3	33.6	------	------	------	4.9	------	15–29	------	Equigranular.
32	2.63	27.6	30.3	35.8	------	------	------	6.3	2	16–26	------	Seriate.
33	2.65	31.0	19.6	40.7	------	------	------	8.7	------	17–22	------	Seriate; granoblastic mortar common between grains.
Subaverage	2.65	28.6	26.5	38.2	------	------	------	6.7	------	------	------	
Average	2.69	23.9	22.8	41.5	------	------	------	12.0	------	------	------	

[1] *Standard chemical analysis, specimen 22*

[Lab. No. 53-1300scd. Analyst: L. M. Kehl]

	Weight percent
SiO_2	66.92
Al_2O_3	15.19
Fe_2O_3	1.45
FeO	2.52
MgO	1.74
CaO	3.79
Na_2O	3.16
K_2O	3.82
H_2O-	0.06
H_2O+	.48
TiO_2	.47
CO_2	.02
P_2O_5	.18
MnO	.08
BaO	-----
Total	99.88

Norm, specimen 22

	Weight percent	
Q	22.62	
or	22.76	
ab	26.72	(Plagioclase composition An_{37})
an	15.84	
di	1.83	
hy	6.18	
mt	2.09	
il	.90	
ap	.35	
Total	99.29	

Wheeler Crest quartz monzonite. The rock there could as well be considered quartz monzonite as granodiorite, despite a color index that in some specimens is as high as 20.

The Lamarck granodiorite is younger than the Inconsolable and Tinemaha granodiorites, and older than the granodiorite of Cartridge Pass, the Palisade Creek mass of quartz monzonite similar to the Cathedral Peak granite (fig. 13), and the Basin Mountain mass of Tungsten Hills quartz monzonite. It contains inclusions of the Inconsolable and Tinemaha granodiorites and also sends dikes into both intrusives. The granodiorite of Cartridge Pass contains inclusions of Lamarck granodiorite; dikes from the Palisade Creek mass of quartz monzonite similar to the Cathedral Peak granite penetrate the Lamarck granodiorite southwest of the Palisade Lakes, and dikes of the Basin Mountain mass of Tungsten Hills quartz monzonite penetrate the Lamarck granodiorite along the North Fork of Bishop Creek in the vicinity of Loch Leven.

The Lamarck granodiorite resembles the Round Valley Peak granodiorite, which occupies a similar position in the intrusive sequence, but which is somewhat finer grained and is equigranular. The Lamarck granodiorite is much like the Half Dome quartz monzonite of the Yosemite region (Calkins, 1930, p. 120). The two rocks are of about the same grain size, though the Half Dome may be a trifle coarser textured in some places, have the same range in color index, and have euhedral crystals of biotite and hornblende distributed in the same ratio and in the same pattern.

GRANODIORITE OF DEEP CANYON

Two masses of dark-colored granodiorite having a combined area of about a square mile crop out in the eastern part of the Tungsten Hills, in the Deep Canyon area, and several small patches of similar rock lie along the north edge of the Tungsten Hills (pls. 2, 3; fig. 23). In hand specimen the granodiorite appears equigranular, the average grain size being about 1 mm. The color index is variable and ranges from about 10 to more than 20. The darkest grandiorite is in the easternmost part of the larger and more easterly of the masses along Deep Canyon. Within this mass the rock is progressively a lighter shade and coarser grained toward the west. The rock in the smaller masses resembles average rock in the largest mass.

In thin section the rock can be seen to be seriate or faintly porphyritic rather than equigranular; embayed and corroded crystals of plagioclase, hornblende, and biotite, 1 to 2 mm long, are set in a finer grained allotriomorphic-granular groundmass of quartz and perthitic microcline. In some thin sections, optically continuous masses of quartz several millimeters across can be seen to enclose plagioclase, hornblende, and biotite. Modal analyses of three thin sections suggest that plagioclase constitutes a little less than half of the rock, that quartz is more abundant than K feldspar, and that biotite is two to three times as abundant as hornblende (table 10 and fig. 32).

The central bodies of most plagioclase crystals are of composition An_{36-40}. They contain small cores as calcic as An_{52} and are rimmed with oligoclase of composition An_{21-30}. Potassium feldspar is somewhat perthitic and exhibits the quadrille structure of microcline. Quartz contains abundant tiny liquid or gaseous inclusions, and most grains extinguish irregularly. Biotite and hornblende have the usual properties for these minerals. Common accessories are magnetite, sphene, and apatite.

The granodiorite is older than alaskite similar to the Cathedral Peak granite, and it probably is also older than the Tungsten Hills quartz monzonite. Evidences of these relations are that the small patches of the granodiorite along the north edge of the Tungsten Hills are inclusions in alaskite similar to the Cathedral Peak granite and that the smaller and more westerly of the two masses along Deep Canyon appears to be an inclusion in Tungsten Hills quartz monzonite. However, relations along the contact between the mass of granodiorite at the east end of Deep Canyon and the Tungsten Hills quartz monzonite suggest that the age difference between these intrusives may be small and that the granodiorite may be, in fact, an early marginal phase of the quartz monzonite. The granodiorite grades westward toward the quartz monzonite from dark fine-grained calcic granodiorite to lighter colored, coarser grained, more silicic granodiorite, very similar to the adjacent quartz monzonite, which at the contact is finer grained than usual. The actual contact between the granodiorite and quartz monzonite is exposed at only a few places; in some it is sharp and in others it appears gradational.

TUNGSTEN HILLS QUARTZ MONZONITE

The Tungsten Hills quartz monzonite crops out discontinuously in a northwest-trending belt that passes through the central part of the mapped area (pls. 1–4; fig. 33). The hill just west of Longley Meadow in the southwestern part of the Tungsten Hills has been designated the type locality because it is readily accessible and the exposed rock is representative of most of the formation. Fresher rock of similar appearance is exposed in Grouse Mountain and east of Bishop Creek below the junction of the South and Middle Forks.

TABLE 10.—*Modal analyses of granodiorite of Deep Canyon, in volume percent*

[Location of specimens is shown on figure 23]

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Biotite	Hornblende	Accessory and secondary minerals	Percent anorthite in plagioclase			Remarks
								Rim	Body	Core	
1	2.75	20.2	9.0	47.7	12.3	6.7	4.1		40	52	Equigranular, hypidiomorphic.
2	2.66	26.3	19.2	41.5	7.7	2.0	3.3	21–26	44	48	Do.
3	2.69	24.5	16.2	47.0	8.0	2.3	2.0	21–30	36	51	Equigranular, hypidiomorphic, nearly panidiomorphic.
Average	2.70	23.6	14.8	45.4	9.3	3.7	3.1				

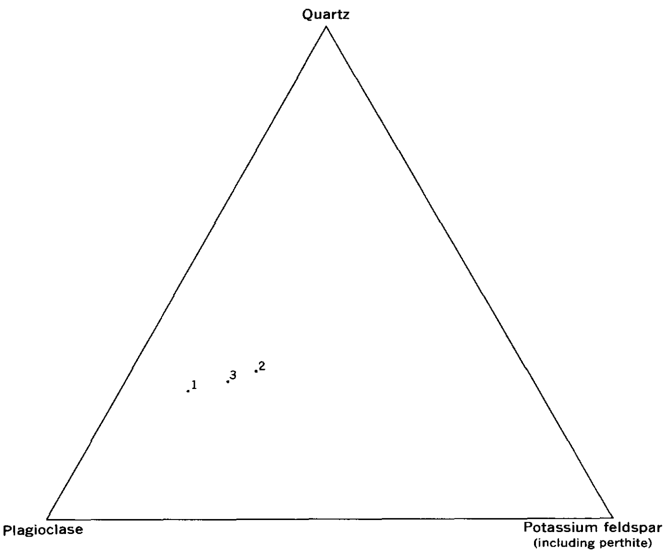


FIGURE 32.—Plot of modes of granodiorite of Deep Canyon on quartz-K feldspar-plagioclase diagram.

The Tungsten Hills quartz monzonite underlies about 120 square miles within the mapped area, and continues westward into the Mount Abbot and Blackcap Mountain quadrangles. It includes the following named masses: Morgan Creek, Tungsten Hills, Basin Mountain, Bishop Creek, and Shannon Canyon (fig. 33). The individual masses are irregular in shape, and exhibit no general parallelism either with one another or with the long axis of the belt. Originally there were probably only two plutons, one lying west and another east of the Pine Creek pendant and the metamorphic rocks around and south of Mount Humphreys. The western pluton was composed of the Morgan Creek and Basin Mountain masses, which are separated from each other only by younger quartz monzonite similar to the Cathedral Peak granite, and the eastern pluton was composed of the Pine Creek, Tungsten Hills, Bishop Creek, and Shannon Canyon masses. The Pine Creek, Tungsten Hills, and Bishop Creek masses are separated from one another only by alluvial deposits and doubtless form a single continuous mass in bedrock. The Shannon Canyon masses are separated from the Bishop Creek mass and from one another by younger intrusives—granodiorite of Coyote Flat, alaskite similar to the Cathedral Peak granite, and finer grained quartz monzonite.

With a few notable exceptions, the quartz monzonite is homogenous both in composition and in appearance. The exceptions are rock of finer grain size than is usual in the Morgan Creek mass and locally in the west side of the Pine Creek mass, light-colored albitized rock adjacent to the north and east margins of the Bishop Creek pendant, and calcic rock of granodiorite composition in the west half of the Basin Mountain mass. The typical quartz monzonite, away from the margins, is medium-grained and medium light-gray on fresh surfaces. Commonly it is porphyritic or subporphyritic, containing subhedral phenocrysts of perthitic K feldspar that are as much as 3 centimeters in the longest exposed direction, parallel with the composition plane of Carlsbad twins, and 1 cm across (fig. 21*F*). The mafic minerals, chiefly biotite but including a little hornblende, are in small irregularly shaped crystals and clusters of crystals that are distributed evenly through the rock. Peripheral concentrations of mafic minerals commonly provide frames for the K feldspar phenocrysts. More rapid weathering of the groundmass further accentuates the prominence of the phenocrysts. Rock in the margins of masses generally is finer grained than rock that is considered to be typical, and the K feldspar phenocrysts are absent, or smaller and less obvious.

About a mile southwest of the Aeroplane mine, rock of typical appearance was observed in sharp contact with rock having a slightly greater abundance of phenocrysts, but, though exposures are good, the contact could be followed along the strike for only a few hundred feet. This contact is thought to be a minor internal contact within the quartz monzonite, caused by the local movement of magma along an already crystallized part.

In most places the rock appears structureless, and planar foliation was mapped in only a few places. Mafic inclusions are scarce, and most of those that were observed are ovoid rather than flattened or spindle

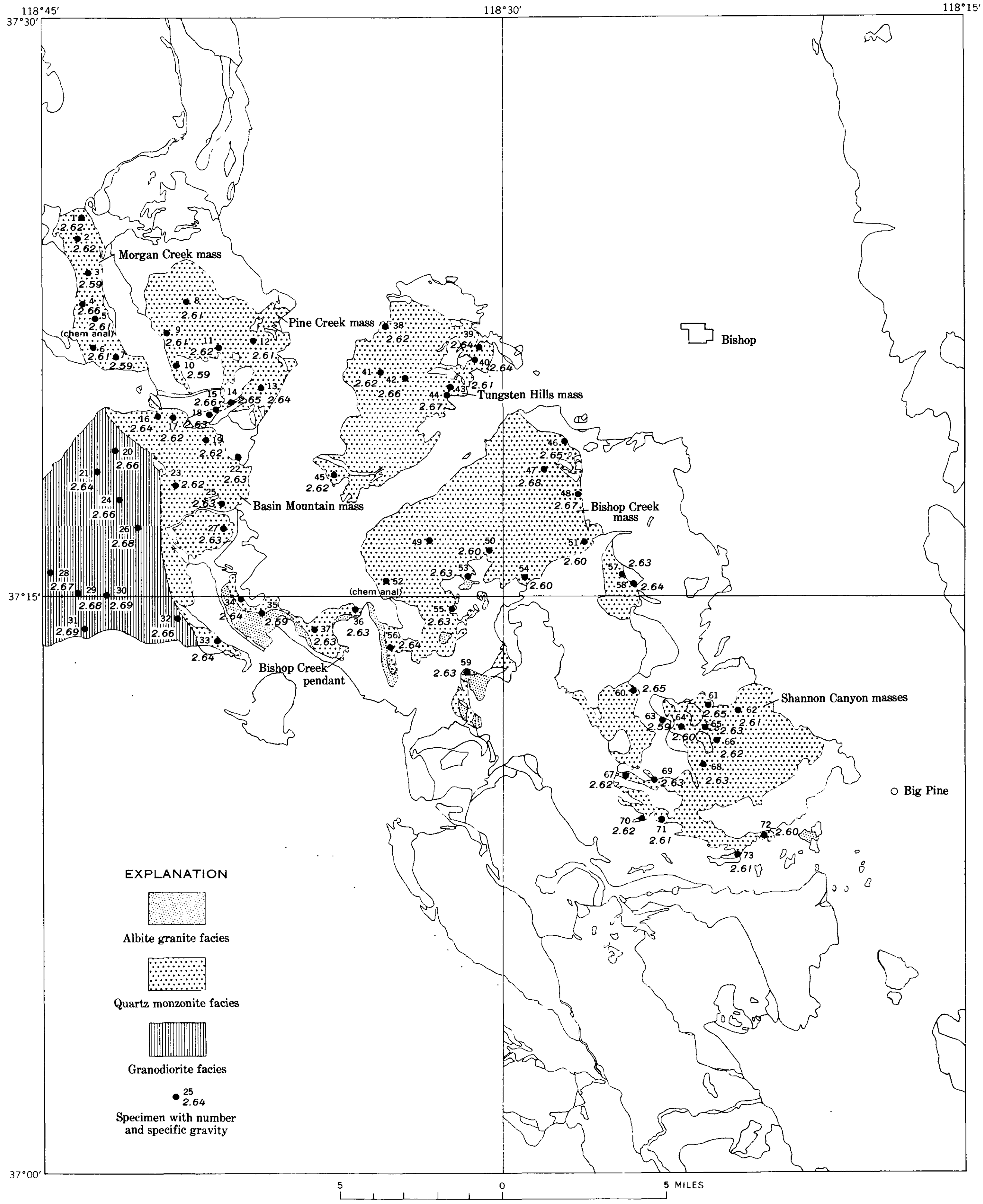


FIGURE 33.—Map showing the locations of modally analyzed specimens of Tungsten Hills quartz monzonite.

shaped. Joints, on the other hand, are conspicuous. Commonly they are widely and regularly spaced. In the southern part of the Tungsten Hills and especially at the type locality, the hill west of Longley Meadow, weathering along two sets of nearly vertical joints that intersect at very close to 90°, has produced a rectalinear pattern of deep slots. Bold rounded forms produced by weathering along the joints are characteristic of the quartz monzonite.

The finer grain of the rock in the northeastern end of the belt, in the Morgan Creek mass and in the southwestern part of the Pine Creek mass, seems to be related to dikelike forms which could have cooled especially rapidly. In the Morgan Creek mass the grain size of the rock decreases gradually toward the west. This gradation is observable along the trail to Pine Lake. At the contact with the Pine Creek pendant the grain size is about 2 mm; it decreases westward and at Pine Lake is less than 1 mm, although scattered perthite phenocrysts and clots of biotite and hornblende are as much as 4 mm across. A similar gradation exists in the western part of the Pine Creek mass, from finer grain in the tongue that extends south along Gable Creek to coarser toward the north and east. Much of the rock in the eastern and southeastern part of the mass is nearly as coarse as typical rock from the Tungsten Hills. Readily recognizable planar foliation in the Morgan Creek mass is evident chiefly in flattened small clots of mafic minerals.

Typical Tungsten Hills quartz monzonite consists of roughly equal amounts of quartz, plagioclase, and perthite (table 11). Biotite is always present, generally in the range 3 to 8 percent, although some rocks contain as little as 1.3 or as much as 11.0 percent. Hornblende, on the other hand, is absent in most specimens and where present exceeds 2 percent of the rock only in rock of granodiorite composition in the western half of the Basin Mountain mass. Common accessory minerals are magnetite and ilmenite, sphene, apatite, allanite, zircon, and zirconlike minerals (possibly monazite and thorite). Epidote is present locally; other common alterations are fine sericite in cores and selected zones of plagioclase, chlorite after biotite, and hematite or limonite after magnetite.

Plagioclase generally is in subhedral zoned crystals. Most of the crystals measured are zoned through some part of the range An_{20} to An_{35}; in some, a central zone is rimmed by albite or sodic oligoclase. Changes in composition are both progressive, from calcic toward the interior of the crystal to sodic toward the exterior, and oscillatory. Albite twins are common, and Carlsbad and pericline twins are less common.

Perthitic K feldspar is both in large twinned phenocrysts and in smaller interstitial masses. Much of it exhibits the quadrille structure diagnostic of microcline, but some shows no such structure. Commonly it contains about 10 percent albite in perthitic intergrowths. In some crystals the albite is in somewhat irregular but generally parallel lamellae, and in others it is in irregular patches. Phenocrysts of K feldspar commonly enclose euhedral to subhedral crystals of zoned plagioclase, hornblende, and magnetite, whereas smaller masses of K feldspar are interstitial to these minerals.

Quartz generally is in large grains that either exhibit conspicuous strain shadows or are composed of a granoblastic mosaic of differently oriented components. Grains that show strain shadows and extinguish irregularly may also show a polygonal internal pattern near extinction. All gradations can be found from grains having undulatory extinction and polygonal strain patterns to grains that consist of granoblastic mosaics.

Biotite generally is in groups of small plates associated with the minor accessory minerals and with hornblende if it is present. Commonly the biotite is pleochroic—X=grayish yellow, Y=Z=olive gray or moderate olive brown. Hornblende is usually in euhedral or subhedral prisms that are pleochroic—X=grayish yellow or moderate yellow, Y=dark yellowish green, Z=blue green.

The results of cataclasis are evident in thin sections of most specimens from the two northernmost masses (the Morgan Creek and Pine Creek masses) and from the north part of the Basin Mountain mass. Elsewhere, only a few scattered specimens yield evidence of cataclasis. The cataclasis is shown by fine granoblastic mortar between grains, by myrmekite that is most abundant in the areas where granoblastic mortar is conspicuous, and by strained or granoblastic quartz grains.

The modes of 73 thin sections (not including the albite granite facies) show that the bulk of the rock is of quartz monzonite composition, except locally in the Tungsten Hills and in the western part of the Basin Mountain mass where the rock is close to granodiorite in composition (fig. 34; table 11). Samples from the western part of the Basin Mountain mass, whose modes fall in the granodiorite field, contain appreciable amounts of hornblende. Hornblende is characteristic of granodiorite rather than quartz monzonite, and these rocks contain 1.5 to 3.7 percent as compared with a maximum of 1.5 percent among all the other modes from the intrusive. Recognition of the near-granodiorite composition of the west part of the Basin Mountain mass in conjunction with a nonporphyritic texture sug-

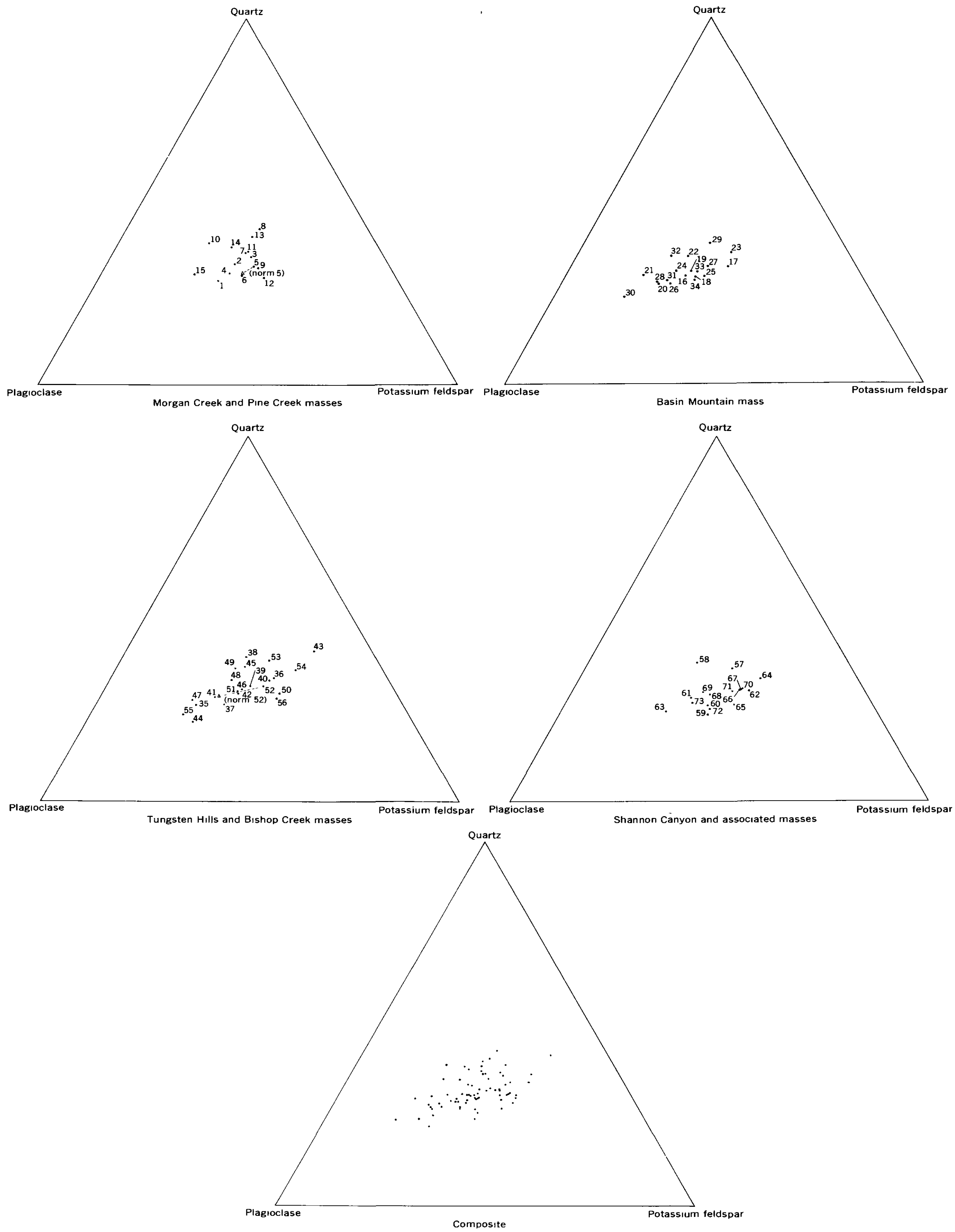


FIGURE 34.—Plots of modes of Tungsten Hills quartz monzonite on quartz-K feldspar-plagioclase diagrams.

Table 11.—*Modal analyses of Tungsten Hills quartz monzonite, in volume percent*

[Where the content of biotite, hornblende and accessory and secondary minerals is shown separately, the mode was determined from a thin section; where only the total content of mafic minerals is shown, the mode was determined from a stained slab. Location of specimens is shown on fig. 33]

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Biotite	Hornblende	Accessory and secondary minerals	Total mafic	Percent of anorthite in plagioclase			Remarks
									Rim	Body	Core	
1	2.62	26.4	27.0	40.0	5.0	1.5		6.5		18–34		Less than average grain size; granoblastic mortar.
2	2.62	30.0	27.8	33.0	5.7	1.2	2.3	9.2		21–34		Fine grained; groundmass either allotriomorphic granular or granoblastic.
3	2.59	33.4	32.0	30.0	3.6		1.0	4.6				Fine grained; porphyritic groundmass allotriomorphic granular.
4	2.66	27.2	27.4	34.9	8.7	1.2	.6	10.5				Less than average grain size; porphyritic with allotriomorphic granular groundmass.
5[1]	2.61	30.1	32.8	30.6	5.2		1.3	6.5		22–34		Fine grained; foliated; larger crystals in allotriomorphic granular or granoblastic groundmass.
6	2.61	28.0	31.6	33.7	5.6		1.1	6.7	0+	25–34		Fine grained; some granoblastic mortar and myrmekite
7	2.59	32.7	28.7	29.1	7.4	0.7	1.4	9.5		17–33		Less than average grain size; some granoblastic mortar and myrmekite.
8	2.61	40.3	30.2	24.5	1.3		3.7	5.0		22–33		Average grain size; nonporphyritic.
9	2.61	30.1	35.6	31.2				3.1				Fine grained; some granoblastic mortar and myrmekite.
10	2.59	36.6	20.1	37.4	5.9			5.9				Fine grained (in narrow tongue); porphyritic with allotriomorphic granular groundmass.
11	2.62	34.5	30.3	30.0				5.2		22–29		Average grain size; nonporphyritic; some granoblastic mortar, myrmekite, and quartz mosaics.
12	2.61	28.3	38.0	30.2				3.5		28–34		Average grain size; nonporphyritic; granoblastic mortar and myrmekite.
13	2.64	38.9	29.9	27.6	2.8		0.8	3.6		23–32		Average grain size; nonporphyritic; some granoblastic mortar and myrmekite.
14	2.65	34.0	24.9	31.7				9.4				Average grain size; nonporphyritic; much granoblastic mortar, quartz mosaics, and myrmekite.
15	2.66	27.8	20.4	44.0				7.8		21–44		Average grain size; nonporphyritic; a little granoblastic mortar and myrmekite.
Subaverage	2.62	31.9	29.1	32.5				6.5				
16	2.64	26.4	25.6	37.6				10.0				Typical porphyritic; local granoblastic texture, quartz mosaics, and myrmekite.
17	2.62	28.8	33.7	27.5				10.0				Typical porphyritic; myrmekite bands of quartz mosaics common.
18	2.63	27.4	29.3	37.2				6.1	21	32		Typical porphyritic; myrmekite, granoblastic mortar, and quartz mosaics abundant.
19	2.62	29.3	27.4	37.6				5.5				Typical porphyritic; some granoblastic mortar and myrmekite.
20	2.66	24.6	21.2	44.5				9.7	5, 15	23–35		Nonporphyritic; granoblastic bands that suggest mortar.
21	2.64	28.0	17.3	48.8				5.9	20, 25	31		Nonporphyritic.
22	2.63	32.8	25.1	36.8				5.2				Typical porphyritic; some granoblastic mortar and quartz mosaics.
23	2.62	34.4	34.8	26.9				3.9				Typical porphyritic; granoblastic mortar between grains.
24	2.66	27.8	23.3	39.4				9.5	24, 29	37		Nonporphyritic.
25	2.63	27.7	31.3	35.4				5.6				Typical porphyritic.
26	2.68	23.4	22.3	40.0				14.3	30	37		Nonporphyritic.
27	2.63	29.9	30.4	37.8				6.9				Typical porphyritic.
28	2.67	25.3	20.5	45.1				6.9	21, 23	35		Nonporphyritic.
29	2.68	32.6	25.5	26.6				1.3	26, 29	31–36		Nonporphyritic. (Mode questionable.)
30	2.69	20.8	14.8	51.9	8.1	2.2	2.1	12.4	20, 25	35, 39		Nonporphyritic.
31	2.69	25.1	22.1	41.6				11.2	23, 29	34, 39		Nonporphyritic; some myrmekite.
32	2.66	31.7	20.5	38.5				9.3		40		Nonporphyritic; some myrmekite and granoblastic mortar; quartz mosaics.
33	2.64	28.0	28.3	35.5				8.2	20	34–37		Nonporphyritic; abundant granoblastic mortar.
Subaverage	2.65	28.4	25.4	38.4				7.8				
34	2.64	26.9	29.3	37.7				6.0	22	27		Finer grained than typical and nonporphyritic.
35	2.59	25.6	23.3	47.7				3.4	12, 18	14, 15		Finer grained than typical and nonporphyritic and felsic.
36	2.63	32.0	36.6	25.6				5.8	8–22	23		Nonporphyritic.
37	2.63	25.3	28.5	39.7				6.4				Medium grained; moderately large plagioclase and larger perthite crystals.
38	2.62	38.6	29.1	30.2				2.1				Typical porphyritic; granoblastic mortar and quartz mosaics.
39	2.64	29.8	31.9	32.5				5.8				Fine grained; perthite phenocrysts small.
40	2.64	31.1	35.7	26.7				6.5				Typical porphyritic.
41	2.62	27.7	26.2	42.1				4.0				Do.
42	2.66	26.9	29.7	34.2				9.2				Do.
43	2.61	38.8	42.3	12.9				6.0				Typical porphyritic; much fine granoblastic rock.
44	2.67	21.2	24.6	50.3				3.9				Typical porphyritic.
45	2.62	35.4	29.4	30.8				4.4				Do.
46	2.65	27.6	29.3	32.4				10.7				Typical porphyritic; quartz in mosaics or in larger strained crystals.
47	2.68	26.4	21.3	46.6				5.7				Typical porphyritic; anomalous high color index.
48	2.67	30.6	26.0	34.9				8.5				Typical porphyritic.
49		35.7	27.6	33.6				3.1				Do.
50	2.60	28.4	40.6	26.5				8.4				Do.
51	2.60	25.9	27.6	32.1	11.0		3.4	14.4				Do.
52[1]		30.7	36.6	29.5	2.9		0.3	3.2				Do.
53	2.63	37.0	34.8	24.7				4.5	0–25	33		Typical porphyritic; a little myrmekite.
54	2.60	33.8	40.5	19.7				6.0				Typical porphyritic.
55	2.63	22.3	20.5	49.8				7.4	0–22	30, 31		Finer grained than usual; nonporphyritic.
56	2.64	26.9	40.5	27.7	3.0		1.9	4.9	20	13, 19		Finer grained than usual; larger crystals in granoblastic or allotriomorphic granular groundmass.
Subaverage	2.63	29.8	31.0	33.2				6.0				

See footnote at end of table.

TABLE 11.—*Modal analyses of Tungsten Hills quartz monzonite, in volume percent*—Continued

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Biotite	Hornblende	Accessory and secondary minerals	Total mafic	Percent of anorthite in plagioclase			Remarks
									Rim	Body	Core	
57	2.63	34.1	33.6	26.6				6.3				Typical porphyritic.
58	2.64	34.7	23.7	32.9				8.7				Typical porphyritic; a little granoblastic mortar.
59	2.63	23.1	34.4	39.5				3.0				Finer grained than typical; nonporphyritic.
60	2.65	24.6	32.3	36.7				6.4				Typical porphyritic.
61	2.65	26.0	27.2	39.0				7.8				Do.
62	2.61	29.2	41.0	26.8				3.0				Do.
63	2.59	23.1	23.4	46.6				6.9				Do.
64	2.60	33.1	42.6	22.6				1.7				Do.
65	2.63	24.8	38.9	31.4	4.3		0.6	4.9	20–29	31, 36		Felsic and shattered; much granoblastic mortar; some plagioclase with chessboard structure.
66	2.62	28.7	37.6	27.3				6.4	0	13–21	10	Typical porphyritic.
67	2.62	29.2	38.6	28.7				3.5				Do.
68	2.63	27.3	31.2	35.2				6.3				Do.
69	2.63	28.0	29.0	35.1				6.9	0, 21	9, 12		Do.
70	2.62	28.7	37.5	26.7				7.1	15, 20	8, 12		Do.
71	2.61	29.2	37.2	30.3				3.3				Typical porphyritic; quartz mosaics.
72	2.60	23.8	33.3	36.8				6.1				Typical porphyritic.
73	2.61	23.9	26.8	37.9	7.8		3.0	11.6				Do.
Subaverage	2.62	27.7	33.5	32.9				5.9				
Average	2.63	29.2	29.7	34.3				6.6				

[1] *Standard chemical analyses, Tungsten Hills quartz monzonite, in weight percent*

[Specimen 5: lab. No. 53–1298scd. Specimen 52: lab. No. 53–1297scd. Analyst: L. M. Kehl]

	Specimen 5	Specimen 52
SiO_2	71.42	69.60
Al_2O_3	14.03	14.89
Fe_2O_3	.89	1.07
FeO	1.63	1.99
MgO	.70	.91
CaO	1.91	2.70
Na_2O	2.86	3.18
K_2O	5.35	4.45
H_2O-	.08	.08
H_2O+	.35	.31
TiO_2	.36	.42
CO_2	.02	.01
P_2O_5	.09	.12
MnO	.05	.07
BaO	----------	----------
Total	99.74	99.80

Norms, Tungsten Hills quartz monzonite, in weight percent

	Specimen 5		Specimen 52	
Q	28.38		26.52	
or	31.68		26.13	
ab	24.11	(Plagioclase composition An_{28}.)	26.72	(Plagioclase composition An_{32}.)
an	9.46		12.51	
di	-----		-----	
hy	3.55		4.53	
il	.65		.78	
mt	1.39		1.57	
C	.10		.31	
ap	-----		.29	
Total	99.32		99.36	

gests that the rock belongs to a different intrusive, but the fieldwork failed to reveal an intrusive contact; on the contrary the appearance of the rock changes from east to west across the mass from typical Tungsten Hills quartz monzonite to rock typical of the granodioritic facies.

The Tungsten Hills quartz monzonite is younger than the Wheeler Crest quartz monzonite, the granodiorite of Deep Canyon, the Lamarck granodiorite, and the Round Valley Peak granodiorite, and older than quartz monzonite and alaskite similar to the Cathedral Peak granite and the granodiorite of Coyote Flat, and probably older than finer grained quartz monzonite. The intrusive relations with the older rocks have been discussed in connection with each of these rocks. Swarms of dikes satellitic to the Mono Recesses mass of quartz monzonite similar to the Cathedral Peak granite penetrate the Tungsten Hills quartz monzonite along Pine Creek on the west side of the Pine Creek pendant. Dikelike masses of alaskite similar to the Cathedral Peak granite penetrate the largest of the Shannon Creek masses of Tungsten Hills quartz monzonite a mile northwest of Piper Peak. Intrusive relations with finer grained quartz monzonite are not entirely certain, but the Sugarloaf mass of finer grained quartz monzonite appears to be a stock and to penetrate Tungsten Hills quartz monzonite.

ALBITE GRANITE FACIES

The albite granite adjacent to the north and east margins of the Bishop Creek pendant is a rusty-weathering light-colored rock that contains almost no mafic minerals but does contain pyrite. It may have been formed by the soda-metasomatism of slightly shattered quartz monzonite adjacent to the Bishop Creek pendant. Albite pseudomorphs the preexisting feldspars, and the resultant rock is not texturally very different from the adjacent quartz monzonite.

The albite granite crops out in a discontinuous zone having an aggregate area of about 3½ square miles, which ranges in width from a few inches to as much as 4,000 feet. The albitized rock grades imperceptibly into quartz monzonite through transitional zones that range in width from less than an inch to several feet, but broad zones of mixed rock as wide as several hundred feet are common. Although it has been possible to map the albite granite separately from the unaltered quartz monzonite, the gradational contacts between the two rocks leave no room for doubt that the albitized rock is a facies of the quartz monzonite.

Albitized rock can be readily examined along the road in the North Fork of Bishop Creek. The hill slope west of North Lake and a rock slide at the base of this slope, a quarter of a mile west of the lake, are composed almost exclusively of albitized rock. The albitized rock is hypidiomorphic-granular—and the average grain size about 2 mm. Although it is finer grained than typical Tungsten Hills quartz monzonite, it is of about the same grain size as the adjacent quartz monzonite. On fresh surfaces quartz is light gray and feldspar is mottled, commonly white to yellowish gray or grayish yellow.

The composition ranges from rock composed of approximately equal amounts of quartz, albite, and microcline, to rock composed of about 35 percent quartz, 65 percent albite, and almost no K feldspar. The rock commonly contains, in addition, a little biotite, traces of sphene and magnetite (or ilmenite), and rare grains of epidote. Very nearly pure albite is indicated by its low index and the extinction angles of 15° to 17° in the zone normal to (010) and (001). Albite with two kinds of structures can be distinguished, (1) albite with closely spaced, throughgoing lamellar twins, and (2) albite with chessboard structure. The chessboard structure results from intricate alternation of closely spaced polysynthetic albite twins and polysynthetic pericline twins. The lamellar twinned albite with throughgoing twins may contain finely disseminated flakes of sericite and biotite, whereas the chessboard albite is clean except for a little fine dust. This difference in the two kinds of albite causes the mottled coloration of the feldspar of some specimens; the mica-bearing lamellar twinned albite corresponds with yellowish gray to grayish yellow grains and the chessboard albite with light gray grains. Some of the chessboard albite is twinned according to the Carlsbad law.

Chessboard albite and microcline are inversely proportional to each other in abundance. Where both are present, they commonly occur with one enclosed in or intergrown with the other. Commonly the chessboard structure of the albite is in parallel (or nearly parallel) orientation with the quadrille structure of the microcline. In places, the chessboard albite contains tiny blebs of quartz, which locally are sufficiently abundant to produce myrmekite. Some of the lamellar twinned albite is enclosed in microcline, as it is in the adjacent quartz monzonite; in specimens lacking microcline some lamellar twinned albite may be enclosed in chessboard albite.

Quartz is in anhedral grains that have undulatory extinction and abundant tiny liquid inclusions, and appears to be identical with the quartz in the adjacent quartz monzonite. Granoblastic mosaic structure within the quartz is present locally, though it is not conspicuous, and most larger grains of quartz exhibit undulatory extinction. In albite-rich rock, sparse flakes of biotite generally are pale—X=colorless and Y=Z=moderate yellowish brown, whereas in rock with appreciable amounts of microcline the biotite is more strongly colored—X=grayish yellow and Y=Z=grayish olive. Some of the biotite is altered to green chlorite.

Several considerations indicate that the albite granite probably is not a product of direct crystallization from a magma. The most significant of these are the theoretical improbabilities first, that a melt of the composition of the thoroughly albitized rock could exist and, second, that albite more sodic than An_5 is ever pyrogenic (Gilluly, 1933, p. 73–74). The manner of transition of the albite granite to quartz monzonite precludes the possibility that the granite is a separate intrusion, and it seems unlikely that ordinary quartz monzonite and albite-rich granite could crystallize from the same magma. A final argument against an origin by crystallization from a magma is the chessboard albite, which is generally held to be of a replacement origin (Gilluly, 1933, p. 73).

The origin of the albite granite seems best explained by some postconsolidation process of metasomatism of the quartz monzonite by means of which Na was added and Ca, K, Fe, and Mg were removed. Unlike the albite granite at Sparta, Oreg. (Gilluly, 1933), the silica content in these rocks appears to have remained approximately constant; indeed no evidence is apparent in the thin sections that quartz was in any way affected during the alteration. Only two sources for the albitizing solutions seem possible: the metamorphic rocks in the Bishop Creek pendant, and some source deeper than the exposed level, possibly late magmatic solutions from the quartz monzonite or from younger intrusions. The fact that albitized rock is found only contiguous with the pendant strongly favors the pendant as a source for the required Na even though no sodium-rich metamorphic rocks have been found. Possibly the sodium came from a layer of salt, which was removed during metamorphism.

Cataclasis of sufficient intensity to have produced permeability is indicated by weak cataclastic structures. Some relict granoblastic mortar borders larger crystals, the twins in lamellar twinned plagioclase are commonly bent and offset on small fractures, and quartz is strained and locally reduced to a granoblastic mosaic.

The albitization probably took place chiefly by the pseudomorphic replacement of both the K feldspar and calcic plagioclase. Lamellar twinned albite, because it contains tiny mica flakes distributed similarly to the plagioclase in the quartz monzonite and because it is in grains of the same size and habit as the plagioclase of the quartz monzonite, is believed to have replaced plagioclase. However, in the thin sections that were studied none of the original plagioclase was observed. The chessboard albite, on the other hand, appears to have taken the place of microcline and to have acquired the chessboard structure thereby. Comparison of thin sections containing chessboard albite and very little microcline with thin sections containing abundant chessboard albite shows that texturally albite substitutes for microcline. The replacement of microcline by chessboard albite is strongly substantiated by the close association and parallel intergrowth of the two minerals where they are present together.

The fact that microcline is associated in some slides with lamellar twinned albite in which chessboard albite is lacking indicates that plagioclase was replaced before microcline. The mafic minerals also were eliminated early. The replacement of microcline by chessboard albite was the last effect of the metasomatism.

ROCKS SIMILAR TO THE CATHEDRAL PEAK GRANITE

Alaskite and quartz monzonite similar in appearance to the Cathedral Peak granite of Yosemite (Calkins, 1930, p. 126–127) occur in several discrete bodies within the part of the Sierra Nevada mapped in connection with this report. The total area of rocks similar to the Cathedral Peak is about 68 square miles, divided almost equally between quartz monzonite and alaskite (fig. 35). The largest mass that lies entirely within the mapped area is the Rawson mass of alaskite, but the North and South lobes of the Mono Recesses mass of quartz monzonite are both parts of a very large pluton, which extends entirely across the adjoining Mount Abbot quadrangle (Sherlock and Hamilton, 1958).

Masses of alaskite and masses of quartz monzonite occur in geographically different areas. Alaskite is confined to the range front from Big Pine Creek north to Wheeler Crest; quartz monzonite lies west of the alaskite in a belt that extends northwestward from near the southeast corner of the mapped area to near the northwest corner. At one stage of mapping, the alaskite and quartz monzonite were considered to be different formational units, but overlapping mineral and textural characteristics and the absence of any evidence that rock of one composition intrudes the other favor the view that the rocks are genetically and temporally very closely related. Nevertheless, it is unlikely that the two facies are precise temporal equivalents, inasmuch as their mineralogic differences probably reflect different temperatures of crystallization.

QUARTZ MONZONITE FACIES

Typical quartz monzonite similar to the Cathedral Peak granite is medium to coarse grained, has an average color index of 3.5, and commonly is equigranular to weakly seriate, but locally is porphyritic (fig. 21G). The average grain size is 3 to 4 mm, but locally the average grain size may be 5 mm or more, and in some marginal parts of masses it is less than 1 mm. Conspicuously porphyritic rock is present in the McGee Creek mass, in the stock between Poleta and Redding Canyons in the White Mountains, and locally in the South lobe of the Mono Recesses mass (fig. 35). Quartz is light gray, and both feldspars are white except in porphyritic rock where K feldspar phenocrysts may have a pinkish caste.

Inclusions are few except in the immediate vicinity of intrusive contacts with metamorphic or mafic igneous rocks. Foliation is rarely discernible, probably because of the paucity of mafic inclusions or minerals. Joints commonly are widely spaced and conspicious.

Quartz, K feldspar, and plagioclase, which constitute more than 95 percent of the average rock, are present in variable amounts (table 12), but almost all of the modes fall within the quartz monzonite field (fig. 36). The average biotite content of the specimens analyzed modally is 3 percent; hornblende was found in only two specimens, where it probably reflects contamination. K feldspar commonly exhibits quadrille structure and is perthitic; except for phenocrysts, which are generally euhedral, it is anhedral. Most larger grains contain smaller grains of the other minerals, and the outer surfaces of phenocrysts are rough because of the abundance of inclusions in their outer parts. Quartz commonly is in large clear anhedral grains that extinguish cleanly. Plagioclase is generally in conspicuously zoned subhedral crystals having a compositional range of An_{10} to An_{30} and an average composition of about An_{20}. The anorthite content varies considerably (table 12).

The Mount Alice mass locally has been contaminated by inclusions of mafic igneous rock. The quartz monzonite adjacent to the mafic rock is darker than the ordinary quartz monzonite, and near rafic rock it contains numerous tiny inclusions of recrystallized and partly granitized mafic rock. Sharp contacts occur between uncontaminated and contaminated quartz monzonite, and in several places contaminated quartz monzonite forms inclusions in uncontaminated quartz monzonite.

An especially interesting feature of the Mount Alice mass is the presence of compositionally layered rock.

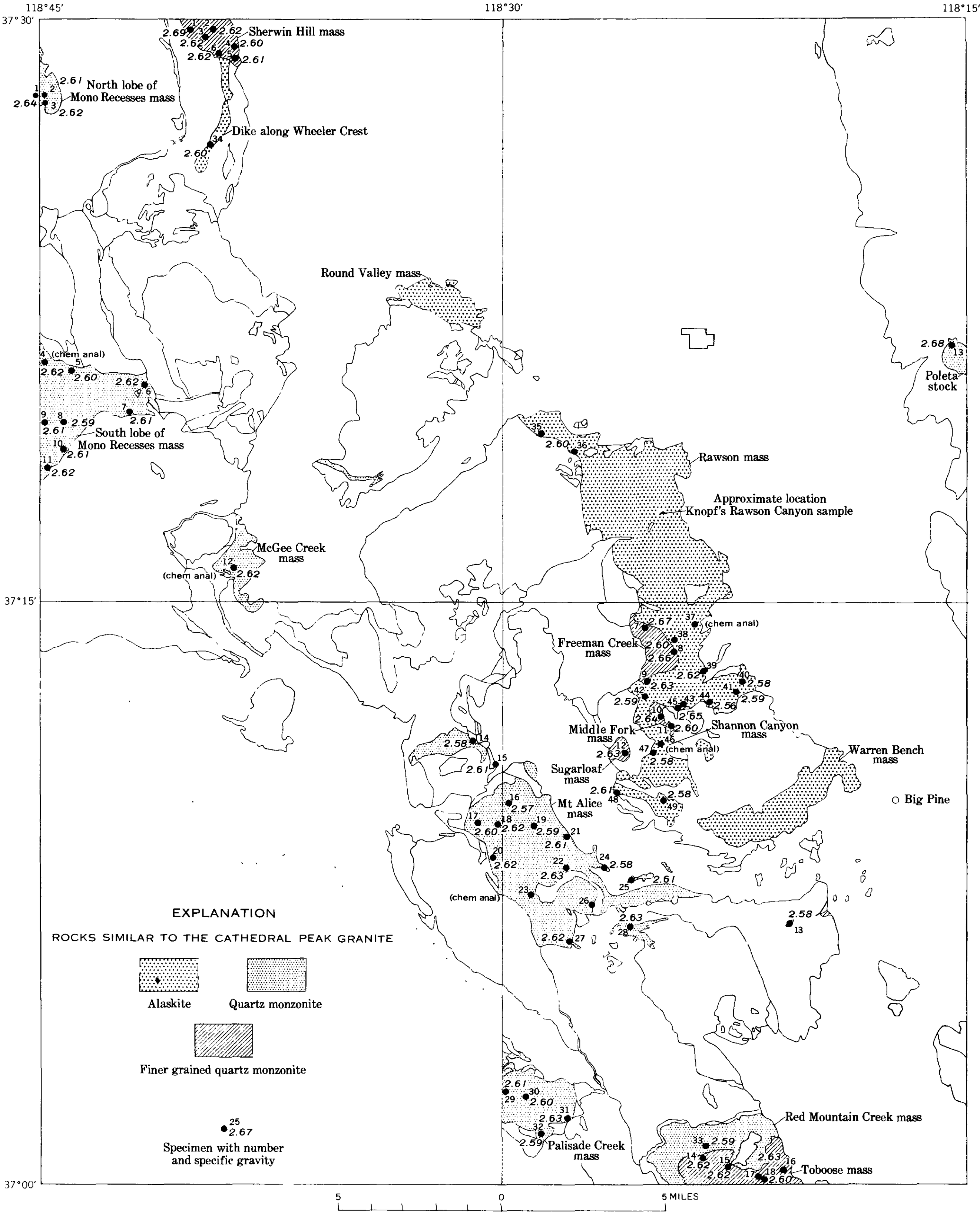


FIGURE 35.—Map showing the locations of modally analyzed specimens of rocks similar to the Cathedral Peak granite and of finer grained quartz monzonite.

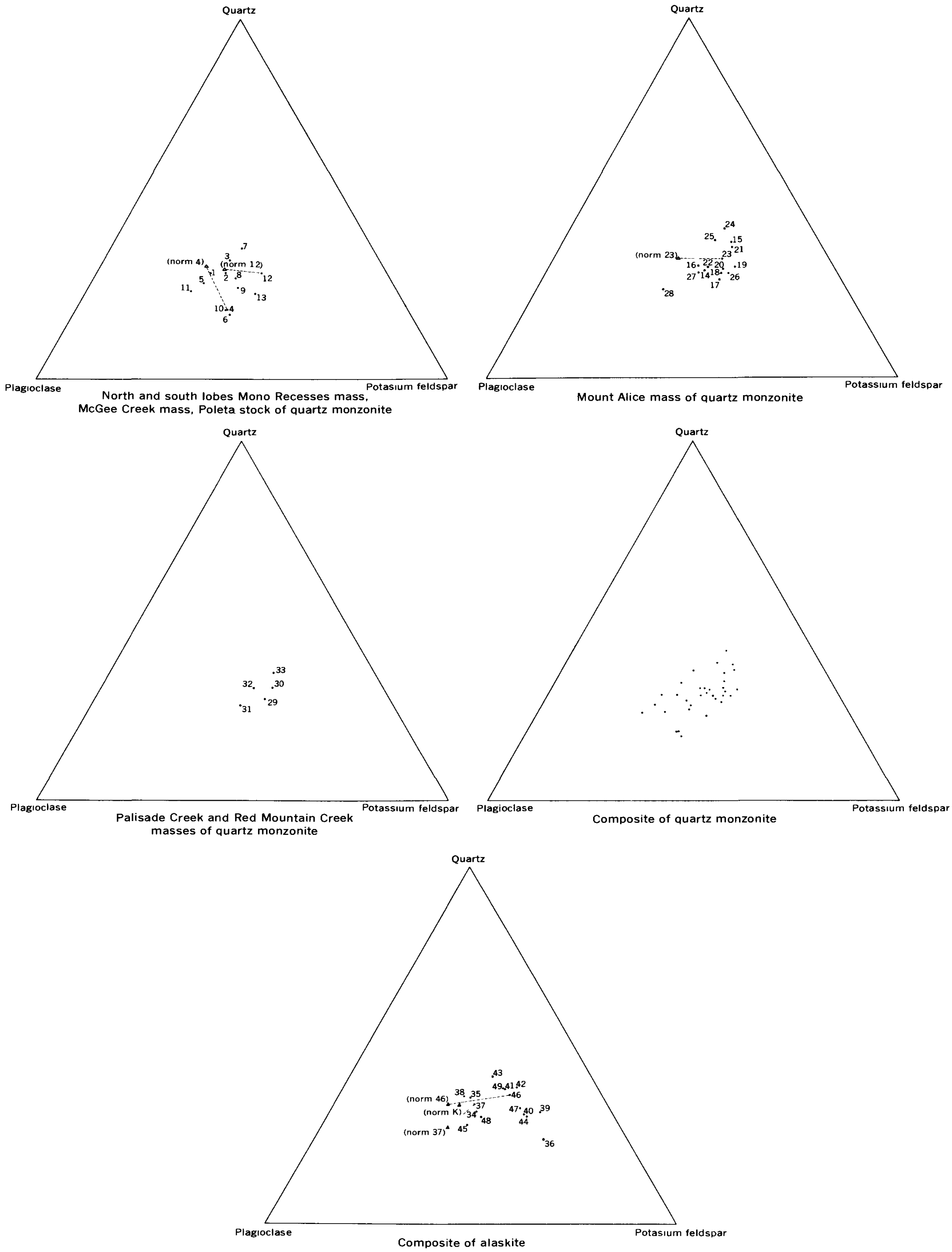


FIGURE 36.—Plots of modes of rocks similar to the Cathedral Peak granite on quartz-K feldspar-plagioclase diagrams.

TABLE 12.—*Modal analyses of rocks similar to the Cathedral Peak granite, in volume percent*

[Where the content of biotite, hornblende, and accessory and secondary minerals is shown separately, the mode was determined from a thin section; where only the total content of mafic minerals is shown, the mode was determined from a stained slab. Location of specimens is shown on fig. 35]

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Biotite	Accessory and secondary minerals	Total mafic	Percent of anorthite in plagioclase			Remarks
								Rim	Body	Core	
1	2.64	28.3	26.4	41.5			3.8		23–31		Equigranular; somewhat finer grained than usual.
2	2.61	28.9	30.6	38.1			2.4		20–31		Do.
3	2.62	32.0	29.8	34.8			3.4		21–30		Do.
4 [1]	2.62	18.6	35.4	42.1			3.8		21–25	22	Do.
5	2.60	26.1	26.2	43.4			4.3		12–26		Do.
6	2.62	17.4	36.8	42.4			3.4		21–29		Somewhat porphyritic; contains K feldspar phenocrysts.
7	2.61	30.8	34.8	30.7			3.6		15–25		Do.
8	2.59	27.0	33.1	36.1			3.8		17–24		Do.
9	2.61	24.6	35.6	37.1			2.5		17–28		Do.
10	2.61	18.6	35.1	42.6			3.6		14–22		Seriate but with most grains in two size groups; finer grained than usual.
11	2.62	23.8	24.9	48.3			3.0		16–22		Equigranular.
12 [1]	2.62	28.7	39.2	29.7			2.4		8–18	15	Porphyritic; K feldspar phenocrysts.
13	2.68	21.6	37.2	31.4			9.8			±20	Porphyritic; pale-red microcline phenocrysts in fine groundmass of biotite, quartz, hypersthene, plagioclase.
Subaverage	2.62	25.1	32.7	38.3			3.8				
14	2.58	29.6	37.7	31.3			1.4	10	26–30		Equigranular; fine grained.
15	2.61	37.4	40.6	21.3			0.7	10	24–29		Seriate; fine grained.
16	2.57	30.5	35.3	31.9			2.3	5	10–20		Equigranular typical.
17	2.60	27.1	42.6	29.1			1.2	7	10–13		Equigrannlar; granoblastic mortar beetween many grains.
18	2.62	28.2	41.2	26.9			4.2	6	22–29	45	Equigranular to porphyritic; a few large K feldspar grains.
19	2.59	30.5	44.4	23.5			1.6	5	12–18		Equigranular; some granoblastic mortar between grains.
20	2.62	29.4	41.5	26.1			3.0	5	17–18		Equigranular; conspicuous granoblastic mortar between grains.
21	2.61	35.4	40.7	21.4			2.5	5	8–17		Equigranular; some granoblastic mortar between grains.
22	2.63	29.6	37.3	29.2			3.9		9–15		Do.
23 [1]		31.8	39.3	24.9	3.3	0.7	4.0	6	15–20		Do.
24	2.58	40.2	35.9	20.3	0.8	2.8	3.6		16–17		Equigranular; typical granoblastic mortar between grains.
25	2.61	37.0	35.3	23.7			3.8		2–11		Equigranular; two grain size groups; may be result of cactaclasis; strongly altered to muscovite.
26		28.7	43.5	26.1			1.7	9	15		Equigranular; some granoblastic mortar between grains.
27	2.62	28.9	36.5	32.8			1.9				Equigranular; typical granoblastic mortar between grains.
28	2.63	24.0	29.8	42.6			3.6	20	26–31		Equigranular; some granoblastic mortar between grains; strongly altered to muscovite.
Subaverage	2.61	31.2	38.8	27.4			2.6				
29	2.61	26.8	39.3	28.6			5.3	10	25–28		Equigranular; typical granoblastic mortar between grains; strongly altered to muscovite.
30	2.60	30.6	40.6	26.2			2.6		9–28		Seriate; large poikilitic K feldspar grains.
31	2.63	25.9	35.2	36.0			2.9	10	18–34		Seriate; finer grained than usual; high biotite content.
32	2.59	31.2	37.0	31.0			0.9	5	11–17		Seriate; slightly finer grained than usual.
33	2.59	35.4	39.3	24.3			1.0				Equigranular; some granoblastic mortar between grains.
Subaverage	2.60	30.0	38.3	29.2			2.5				
34	2.60	31.5	35.3	32.4			0.8		4–14		Equigranular to seriate.
35	2.60	35.1	31.8	32.4			0.7		2–15		Equigranular; finer grained than usual.
36		22.6	54.6	20.9	0.2	1.7	1.9		4–10		Equigranular; average grain size.
37 [1]		33.1	33.4	31.6			1.9		5–17	20	Equigranular; typical.
38	2.60	35.0	29.8	32.6			2.6		22–31		Do.
39	2.62	30.8	49.8	16.7			2.7		10–20		Do.
40	2.58	38.9	45.0	20.4			6.0		5–12		Equigranular.
41	2.59	36.9	38.6	21.9			2.6		10–17		Do.
42	2.59	37.5	41.1	19.2			2.2		5–9		Do.
43		37.5	21.7	32.1	5.8	2.8	8.6		31–33		Mode poor; not plotted on fig. 34.
44	2.56	30.0	48.2	21.6			0.2		2–18		Equigranular.
45	2.65	25.8	32.2	33.8			8.0		27–33		Equigranular; fine grained; probably a hybrid.
46 [1]		34.9	40.5	21.8			2.7				Equigranular.
47	2.58	32.4	45.4	21.2			1.0				Do.
48	2.61	29.5	33.2	31.9			1.4				Equigranular; scattered conspicuous biotite.
49	2.58	36.6	37.5	22.3	0.9	2.6	3.5		5–10		Equigranular.
Subaverage	2.60	33.0	38.6	25.8			2.9				
Average	2.61	29.9	37.1	30.0			3.0				

See footnote on following page.

Layered rock crops out along the trail that follows the North Fork of Big Pine Creek between Second Lake and Third Lake. The layered rock is in tabular masses of diverse orientation, which appear to be included in the quartz monzonite. Superficially the layered rock appears to be part of the quartz monzonite, but careful examination reveals the presence of scattered grains of pink plagioclase, found only in the Inconsolable granodiorite. Most contacts between the quartz monzonite and the layered rock are concordant with the layering, but in a few places tongues of quartz monzonite penetrate across the layers. Presumably the layering was caused by migration of the mafic material in the inclusions to selected surfaces by a process of metamorphic differentiation. These surfaces probably are shear planes, which may have been formed as a result of forcible emplacement of the quartz monzonite. The granodiorite in nearby Temple Crag has been shattered and cemented with quartz and fine-grained felsic rock, very likely as a result of the emplacement of the quartz monzonite.

Footnote for table 12:

1 *Chemical analyses*

[a, Standard rock analysis. Analyst: L. M. Kehl. b, Rapid rock analysis. Analysts: H. F. Phillips, P. L. D. Elmore, and K. E. White. c, Analysis reported by Knopf of sample of alaskite from Rawson Canyon (Knopf, 1918, p. 68)]

	Quartz monzonite				Alaskite	
	Specimen (lab. No.)					
	4(b) (152445)	12(b) (152446)	23(a) (53-1296-scd)	37(b) (152443)	46(b) (152444)	K(c)
SiO_2	71.8	73.0	74.11	71.0	75.4	76.28
Al_2O_3	15.3	15.2	13.73	15.7	13.3	----------
Fe_2O_3	1.0	.4	.60	.9	.3	----------
FeO	.65	.34	.88	.81	.74	----------
MgO	.34	.16	.32	.39	.12	----------
CaO	1.8	1.1	1.29	1.6	.48	.47
Na_2O	3.8	3.8	3.44	3.8	4.1	4.72
K_2O	4.1	5.0	4.92	5.0	4.5	4.73
H_2O- / H_2O+	.48	.48	.12 / .18	.69	.46	----------
TiO_2	.22	.09	.18	.22	.10	----------
CO_2	.08	.10	.01	.07	.11	----------
P_2O_5	.05	.05	.06	.05	.01	----------
MnO	.06	.06	.05	.04	.08	----------
Total	100	100	99.89	100	100	----------

Norms

	Quartz monzonite				Alaskite	
	Specimen					
	4	12	23	37	46	K
Q	29.64	29.22	31.92	25.68	32.58	29.76
or	24.46	29.47	28.94	29.47	26.69	27.24
ab	31.96	31.96	28.85	31.96	34.58	29.82
an	8.90	5.56	6.40	8.06	2.50	2.50
hy	.90	.66	1.72	1.26	1.36	----------
mt	1.39	.70	.89	1.39	.46	----------
il	.46	----------	.32	.46	----------	----------
C	1.33	1.53	.51	1.12	.71	----------
Total	99.04	99.10	99.55	99.40	98.88	----------

ALASKITE FACIES

The alaskite was described by Knopf (1918, p. 67–69) as orthoclase-albite granite. Megascopically it closely resembles nonporphyritic facies of the quartz monzonite, although slight differences are generally recognizable (fig. 21H). The most obvious features of the alaskite that distinguish it from the quartz monzonite are a pinkish cast of the K feldspar, and a very low biotite content—about 1.6 percent on the average for uncontaminated rock, but many specimens contain less than 1.0 percent. However, the most distinctive feature of the alaskite (plagioclase in the albite compositional range) can be ascertained only through microscopic study or by chemical analysis.

The average grain size of the alaskite is 3 to 4 mm, but it ranges from less than 2 mm in the dike along Wheeler Crest to more than 5 mm locally in the Rawson mass. The rock is megascopically equigranular, but appears slightly seriate in thin section. The largest grains are of K feldspar. In the field, foliation is conspicuous only in the dike along Wheeler Crest. Inclusions are scarce. Joints commonly are widely spaced and consequently conspicuous. Much of the rock that crops out in the Tungsten Hills and in the north part of the Rawson mass and in the mass just north of Baker Creek is deeply weathered to a loose iron-stained grüs; consequently, few modes were made on rocks from these areas.

The average mineral composition of 15 modally analyzed specimens is 33 percent quartz, 39 percent K feldspar, 26 percent plagioclase, and less than 3 percent biotite. Hornblende was found in only one specimen, which probably is contaminated.

K feldspar generally is perthitic and exhibits conspicuous quadrille structure. Quartz commonly is in large clear grains that extinguish with little undulation. Biotite is in subhedral plates, which generally are evenly distributed through the rock. Under the microscope, Z commonly is moderate brown, but in some specimens is moderate yellowish brown or moderate olive.

Albite is in anhedral to subhedral grains that exhibit closely spaced lamellar twinning on the albite law. With the exception of contaminated rocks from the marginal zones, the plagioclase is in zoned crystals that generally have an average An content of less than 10 percent. Knopf (1918, p. 67–69) originally determined the albite composition of the plagioclase. His optical determinations indicate a composition of An_{10}; a partial chemical analysis of a rock collected by him from Rawson Canyon indicates a composition of An_6. Knopf pointed out that An_{10} may be closer to the true composition of the plagioclase than An_6 because the K feldspar contains lamellae of almost pure albite. The sodic nature of the plagioclase is supported by optical determinations of the composition of the plagioclase tabulated in table 12, and the calculations of two additional norms. It is evident, however, that the composition is variable and that the plagioclase of some specimens is more calcic than that in some specimens from the quartz monzonite facies. The maximum range of zoning in all the uncontaminated specimens is from An_2 to An_{18}, but most grains are zoned through only a small part of the total range.

MARGINAL DIKES OF APLITE, PEGMATITE, AND ALASKITE

A notable feature of both the quartz monzonite and alaskite facies is that some masses are parent to marginal swarms of felsic dikes that dip gently into and merge with the parent body. Dips of dikes generally are less than 40° and mostly between 20° and 30°. Dikes marginal to the south lobe of the Mono Recesses mass of quartz monzonite are well exposed and can be readily examined on the north side of Pine Creek between Pine Lake and the Pine Creek pendant (fig. 37). Marginal dikes are also abundant on the north side of the Palisade Creek mass, and marginal to the alaskite dike along Wheeler Crest.

The dikes consist chiefly of quartz and feldspar, but locally contain garnet, sphene, biotite, magnetite, and other minerals. Most of the plagioclase in the dikes along Pine Creek is conspicuously zoned in the oligoclase range, as is the plagioclase in the parent mass. The dikes range from aplitic to pegmatitic, and not uncommonly both textures are present in the same dike.

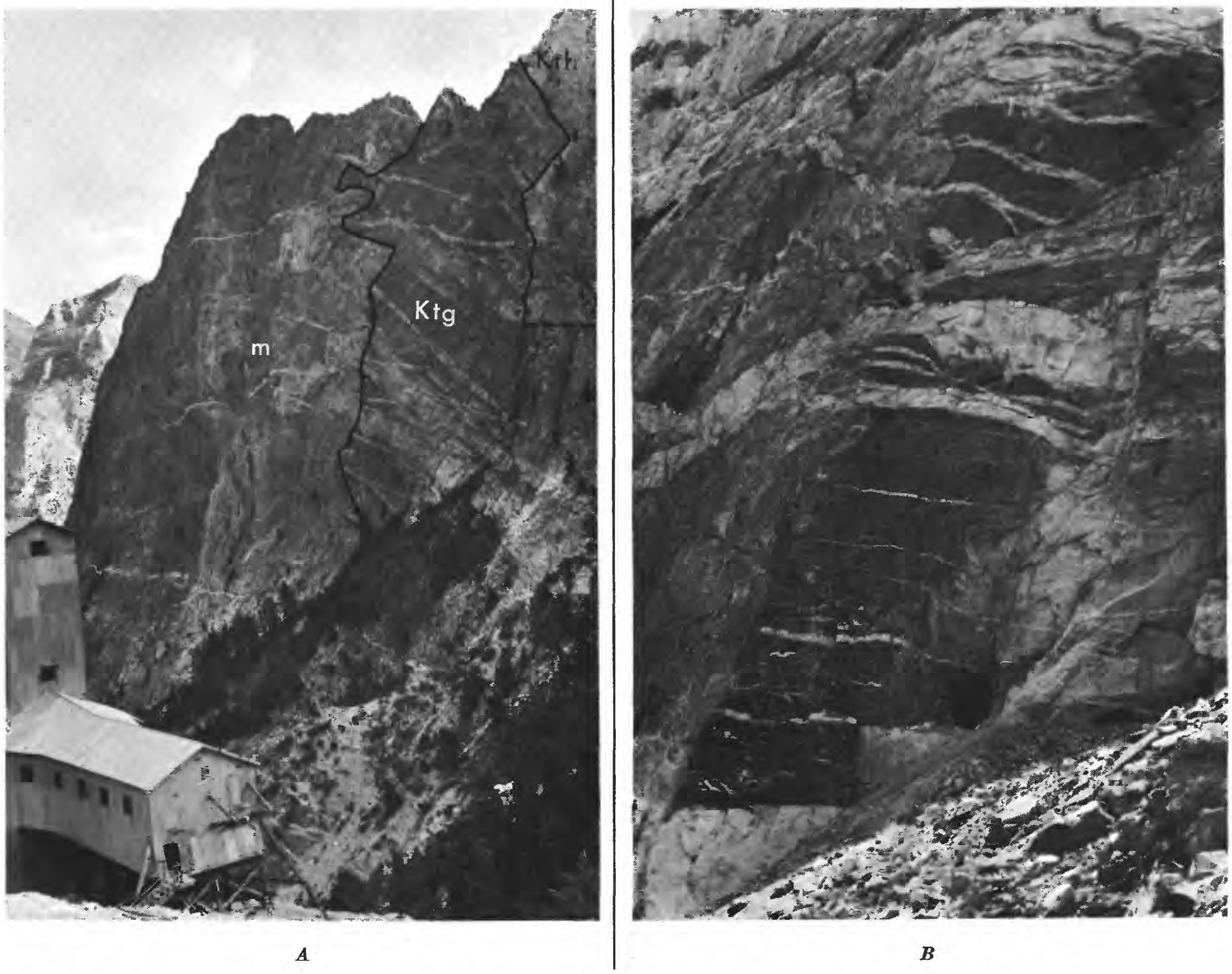


FIGURE 37.—Aplite, pegmatite, and alaskite dikes along Pine and Morgan Creeks. The dikes dip toward the parent mass of quartz monzonite similar to the Cathedral Peak granite, commonly at 20° to 45°. *A*, Along the west side of the Pine Creek pendant dikes cut Tungsten Hills quartz monzonite (Kth) and quartz diorite (Kdg) and finger out in marble (m). *B*, At the Brownstone tungsten mine dikes cut Tungsten Hills quartz monzonite (right), tactite (center), and marble (left).

Dikes having a predominant pegmatitic texture almost invariably cut dikes having a predominant aplitic texture, although no great age difference seems likely. Matching irregularities in the opposing walls of many dikes indicate that the dikes were emplaced by dilation of their walls, although a component movement parallel to the dike walls is required by the geometry of some matching walls. In a few places, thin marginal zones of fine-grained aplite appear to have accreted to the dikes without dilation, and presumably by replacement.

Dikes extend in abundance from the north side of the south lobe of the Mono Recesses mass for more than a mile, and a few dikes are found at distances of more than 2 miles. Along the south side of Pine Creek in the vicinity of the Brownstone mine, close to the parent mass, dikes comprise at least half of the rock exposed in cliff faces, whereas half a mile away, in cliffs on the north side of Pine Creek, dikes comprise only about 10 percent of the exposed rock. Few of these dikes continue into the Pine Creek pendant even where it is very close to the intrusive mass parent to the dikes; most dikes finger out in the marble along the west side of the pendant (fig. 37*A*).

Similar dike swarms have been described by Hans Cloos and his associates (in Balk, 1948, p. 101–106). They interpret the fractures occupied by the dikes as the result of tension caused by the upward movement of magma along a steep or vertical face. According to Balk (1948, p. 104), "The zones of marginal fissures are of utmost importance for the evaluation of the forces acting during the consolidation of many igneous masses. Even where the rock may be structureless, marginal dikes and upthrusts must be regarded as evidence of a strong upward motion of the plutonic mass." This concept of origin fits the known facts in the Bishop district. Matching dike walls indicate dilation, locally accompanied by slight movements parallel with the walls. The fact that the dikes die out away from the parent mass indicates progressively greater dilation toward the parent pluton, which, in turn, must signify upward drag of the wall rocks adjacent to the parent mass. The fact that the dikes maintain a constant range of dips indicates that the fractures they occupy must have formed contemporaneously with the bending, else they would not have been formed or would have been bent upward near the parent pluton from their original configuration. These considerations indicate the fractures must have been formed and the dikes emplaced generally in order from bottom to top as the parent pluton rose.

INTRUSIVE RELATIONS

Rocks similar to the Cathedral Peak granite are among the youngest of the plutonic rocks, and only the granodiorites of Coyote Flat and Cartridge Pass and masses of finer grained quartz monzonite could have been emplaced later. Intrusive relations with older rocks were discussed in connection with each of these rocks. The granodiorite of Coyote Flat was observed to intrude rock similar to Cathedral Peak granite, but evidence of the younger age of the other rocks rests primarily on the gross geometric relations of the rocks, as shown by the map pattern. The granodiorites of Coyote Flat and of Cartridge Pass occur as rather symmetrical bodies that appear to cut across the quartz monzonite similar to the Cathedral Peak granite. Likewise the Wheeler Crest mass of finer grained quartz monzonite appears on the map to cut off the north end of the alaskite dike along Wheeler Crest. Additional suggestive evidence is that thin flat-lying aplite dikes marginal to the alaskite dike, present in the older Wheeler Crest quartz monzonite, were not found within the finer grained quartz monzonite.

The Red Mountain Creek mass of quartz monzonite, unlike any other mass assigned tentatively to the Cathedral Peak, is cut by mafic dikes; this relation suggests the possibility that the Red Mountain Creek mass is older than other masses of rock similar to the Cathedral Peak granite and improperly assigned. Evaluation of this possibility can be made only after the time of emplacement of the mafic dikes, now in doubt, has been established. The Red Mountain Creek mass has a concentric pattern on the map and in section, which suggests that it is younger than the overlying Tinemaha granodiorite and older than the underlying Taboose mass of finer grained quartz monzonite (pl. 4; pl. 5, section *F–F′*).

FINER GRAINED QUARTZ MONZONITE

The finer grained quartz monzonite includes five small masses having an aggregate outcrop area of only a little more than 6 square miles (pls. 3–4; fig. 35). Although most or perhaps all of the masses may have been emplaced at about the same time, evidence bearing on their relative ages is too meager for them to be considered a formational unit. The two largest masses, the Sherwin Hill and Taboose Creek masses (fig. 35) continue beyond the boundaries of the mapped area. The three smaller masses are closely grouped in upper Freeman and Shannon Canyons and in Sugarloaf.

The rock in all the masses is generally similar in texture and mineral content, although minor differences are recognizable. Typical rock is medium grained, though finer grained than any other intrusive, and locally porphyritic. The average grain size is 1 to 2 mm. Phenocrysts of K feldspar ½ to 1½ cm across are present in some parts of the Sherwin Hill and Freeman

Canyon masses. Although the average grain size varies slightly from mass to mass, generally the grain size throughout a single mass is nearly constant.

The rock ranges from very light gray to yellowish gray. Biotite, ordinarily the only dark mineral, is present in amounts ranging from 1.0 to 6.5 percent; most rocks contain between 2.0 and 5.0 percent. Much of the rock appears structureless, but locally feldspar grains and biotite flakes are oriented, and aggregates of biotite flakes form thin layers that give the rock a planar foliation. Mafic inclusions are rare. Some of this foliation may be the result of cataclasis after the original consolidation of the rock.

Much of the mass along Taboose Creek is composed of two very similar appearing rocks of about equal abundance, one of which intrudes the other in narrow irregularly branching dikes. The rock in the dikes is lighter colored than the intruded rock, because the feldspars are white rather than light gray or yellowish gray. Under the microscope the two rocks are indistinguishable, and for this reason are thought to be very closely related to each other.

Average rock consists of about equal amounts of quartz, K feldspar, and plagioclase (table 13). Biotite, with Z=dark yellowish brown, is the only varietal mineral. Accessories are apatite, zircon, allanite, magnetite, and pyrite. The plot of modes (fig. 38) shows considerable scatter, which, in view of the relatively fine grain-size probably reflects real differences in composition. Perthitic K feldspar is commonly in large anhedral grains that enclose smaller subhedral to euhedral plagioclase grains. Most K feldspar exhibits conspicuous quadrille structure. Plagioclase commonly is finely twinned according to the albite law; it generally ranges from An_{25-30} in the cores of crystals to An_{20} in the rims, but cores more calcic than An_{30} and rims more sodic than An_{20} are not uncommon. Quartz is in anhedral crystals of moderate size, many of which consist of granoblastic aggregates which may have been produced from original homogeneous grains by strain.

The texture in nonporphyritic rocks is hypidiomorphic-granular. Granoblastic mortar is present along the boundaries of some grains, and is associated with myrmekite along boundaries between perthitic K feldspar and plagioclase. Myrmekite also is present locally in thin veins that lie along contacts between two K feldspar crystals or more rarely cut across perthite crystals.

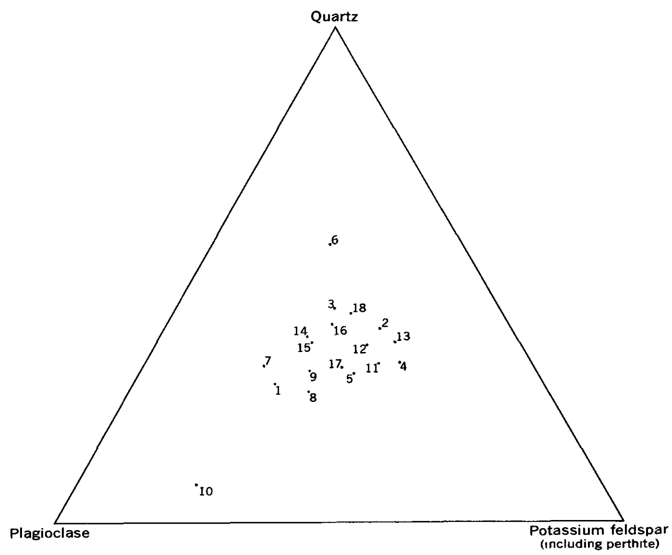


Figure 38.—Plot of modes of finer grained quartz monzonite on quartz-K feldspar-plagioclase diagram.

Table 13.—*Modal analyses of finer grained quartz monzonite, in volume percent*

[Location of specimens is shown on fig. 35]

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Biotite	Hornblende	Accessory and alteration minerals	Percent anorthite in plagioclase			Color index	Remarks
								Rim	Body	Core		
1	2.69	26.3	22.7	43.0	4.5	2.1	1.4	21	27–36	40	6.7	
2	2.62	37.9	36.6	21.9	2.4		1.1	2–12	23–24		3.5	
3	2.62	41.8	27.2	27.8	2.3		.9	8–17	22–25		2.3	
4	2.60	30.3	42.7	22.2	2.8		1.9	4, 17	23–27		4.7	
5	2.61	29.8	37.1	31.8	1.1		.2	8	12–18		1.3	
6	2.62	51.8	19.5	21.6	6.4		.7	4–5	16–17		7.3	
7	2.67	30.3	20.2	44.9	3.9		.7	5	24–38	38–42	4.6	
8	2.66	24.7	29.6	39.7	4.9		1.2		28–37		6.0	
9	2.63	29.1	28.4	37.8	3.6		1.1		23–31		4.7	
10	2.64	7.3	19.8	66.3	5.7		.8	8	21–33		6.5	A few large perthite crystals; plagioclase is in small subhedral zoned crystals.
11	2.60	30.2	39.2	25.5	4.6		.6	5, 16	26–32		5.2	
12	2.63	34.3	35.6	25.9	3.6		.6	21, 23	28–36		4.2	
13	2.58	34.3	40.0	20.8			4.9		2–17		2.1	
14	2.62	36.1	25.0	35.3	2.9		.6	8	17–38		3.5	
15	2.62	34.6	25.9	34.5	4.3		.6		21–39	40	4.9	
16		38.7	28.2	30.1	2.8.		.2	23	28–34			
17	2.63	29.2	32.7	31.8	3.9		2.4		30–31		5.5	
18	2.60	40.1	30.2	25.5	3.9		.4	5, 8	16, 20		4.3	
Average		32.5	31.5	31.2	3.4		1.4					

Few intrusive relations of masses of finer grained quartz monzonite to the rocks with which they are in contact were established by field observation. In part this lack results from the fact that the finer grained quartz monzonite is closely jointed and breaks down to rubble that obscures the intrusive contacts.

Among the older intrusives, only the Wheeler Crest quartz monzonite is in contact with finer grained quartz monzonite. Small dikes from the Sherwin Hill mass of finer grained quartz monzonite penetrate the Wheeler Crest quartz monzonite and clearly indicate the younger age of the finer grained quartz monzonite. The three small masses at the heads of Freeman Creek and Shannon Canyon are in contact with the Tungsten Hills quartz monzonite, alaskite similar to the Cathedral Peak granite, and the granodiorite of McMurry Meadows, and the Taboose mass is in contact with quartz monzonite similar to the Cathedral Peak granite. The Sugarloaf mass has the general appearance of a stock intrusive into the Tungsten Hills quartz monzonite, but all contacts are covered with rubble. The nearly circular shape on the map of the granodiorite of Coyote Flat gives the impression that it is younger than the Freeman Creek mass of finer grained quartz monzonite, but such a relation was not established. The Taboose mass of finer grained quartz monzonite underlies the alaskite of Red Mountain Creek and Tinemaha granodiorite in a concentric arrangement in plan and section and is probably younger (pl. 4; pl. 5, section *F–F'*).

Some or all of the masses of finer grained quartz monzonite may be correlative with the Johnson granite porphyry of the Yosemite region (Calkins, 1930, p. 127–128). The Johnson granite porphyry is similar in appearance to the finer grained quartz monzonite, and its intrusive position as the youngest formation of the Tuolumne intrusive series is compatible with the inferred position of the finer grained quartz monzonite in the intrusive sequence.

GRANODIORITE OF COYOTE FLAT

Granodiorite underlies about 13 square miles in the northeastern part of Coyote Flat and adjoining areas to the north and east. It constitutes a nearly equidimensional pluton in map pattern except for a tongue that extends northeastward along upper Rawson Creek. Most of the pluton is well exposed, but the west-central part is covered by alluvial fill in Coyote Flat.

Most of the rock is light gray and medium grained; the average grain size is 1 to 2 mm. A marginal zone a few hundred feet thick is generally present, which is darker and finer grained than the more abundant rock from the interior of the pluton. The marginal rock is conspicuously foliated parallel with the external contacts; it contains few mafic inclusions, and the foliation is shown by orientation of the constituent minerals.

In composition the rock is granodiorite. The average mineral content of nine modally analyzed specimens is 22 percent quartz, 13 percent K feldspar, 48 percent plagioclase, 9 percent biotite, 5 percent hornblende, and 4 percent accessories (table 14). Except for specimens 1, 6, and 8, from the margins of the pluton, the plot of modes shows little scatter (fig. 39). Samples 1, 6, and 8 are considerably richer in plagioclase than most other specimens. The average color index of all the specimens is 17, but ranges from 12 to 25.

In thin section the most striking feature of the granodiorite is that it is conspicuously poikilitic; large anhedral grains of quartz and K feldspar having quadrille structure comprise a groundmass, which encloses euhedral to subhedral grains of the other minerals in the rock. The range of zoning of the small euhedral plagioclase grains is exceptionally great. Large unzoned or slightly zoned central body areas generally fall in the compositional range of An_{36} to An_{44}. The body areas contain small cores as calcic as An_{70} and are enclosed by rims of sodic plagioclase that may be as calcic as An_{30} at the inner edge and as sodic as An_{20} at the outer edge. Biotite is somewhat unusual in that it is grayish red in the Z direction.

The granodiorite of Coyote Flat sends dikes into and contains inclusions of the Tungsten Hills quartz monzonite and alaskite similar to the Cathedral Peak granite, and is clearly younger. Intrusive relations with finer grained quartz monzonite were not established by field observations, but the roundish shape in plan of the granodiorite suggests that it intrudes all the rocks

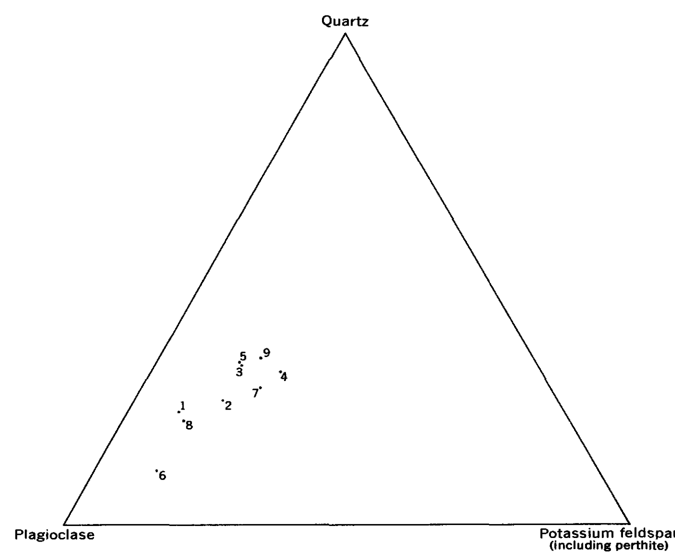


FIGURE 39.—Plot of modes of granodiorite of Coyote Flat on quartz-K feldspar-plagioclase diagram.

TABLE 14.—*Modal analyses of granodiorite of Coyote Flat, in volume percent*

[Location of specimens is shown on fig. 23]

Specimen	Specific gravity	Quartz	K feldspar	Plagioclase	Biotite	Hornblende	Accessory and secondary minerals	Remarks
1	2.73	18.5	7.6	56.2	10.5	4.0	3.2	Megascopically equigranular but notably poikilitic in thin section.
2	2.65	22.4	13.5	51.6	8.2	3.4	.8	Do.
3	2.70	27.2	12.6	43.6	8.6	4.6	3.4	Do.
4	2.69	25.7	19.2	38.8	7.3	3.6	5.4	Do.
5	2.69	28.5	12.4	44.7	5.7	3.7	5.0	Do.
6	2.82	8.5	8.1	58.4	11.5	8.5	5.0	Megascopically equigranular but notably poikilitic in thin section; finer grained and darker than typical.
7	2.70	22.3	17.4	41.3	9.0	5.7	4.3	Megascopically equigranular but notably poikilitic in thin section.
8	2.76	16.5	8.7	54.3	9.3	6.0	5.2	Megascopically equigranular but notably poikilitic in thin section; finer grained and darker than typical.
9	2.71	29.3	15.4	42.0	7.8	2.5	3.0	Megascopically equigranular but notably poikilitic in thin section.
Average	2.71	22.1	12.8	47.9	8.7	4.7	3.9	

with which it is in contact, including finer grained quartz monzonite. It may be correlative with the granodiorite of Cartridge Pass.

GRANODIORITE OF CARTRIDGE PASS

Light-gray equigranular granodiorite having an average grain size of 1 to 2 mm occupies a little less than 3½ square miles in the southwestern corner of the mapped area, and extends into adjoining areas to the south and west (pl. 4; fig. 23). James G. Moore has made intensive studies of the pluton in the Mount Pinchot quadrangle in the vicinity of Cartridge Pass and has found that it is concentrically zoned and that the average composition is granodiorite close to quartz monzonite (Moore, 1963, p. 60). Only two thin sections were studied; one of granodiorite, and the other of quartz diorite (table 15; fig. 40). Both specimens are from outer, more calcic and ferromagnesian zones. The texture of both specimens is hypidiomorphic-granular, but the plagioclase, especially in the specimen of quartz diorite, approaches euhedral, and the rock texture is close to panidiomorphic-granular. In both specimens the central parts of plagioclase crystals are of andesine composition (An_{37-38}), and they are rimmed with calcic oligoclase.

The pluton contains inclusions of Lamarck granodiorite, and is therefore younger. Along the north side it is in contact with quartz diorite and with quartz monzonite similar to the Cathedral Peak granite. It clearly intrudes the quartz diorite, but its intrusive relations with the quartz monzonite were not established. Whether the granodiorite of Cartridge Pass is correlative with either the quartz monzonite of McMurry Meadows or the granodiorite of Coyote Flat depends on whether it is younger or older than rocks similar to the Cathedral Peak granite.

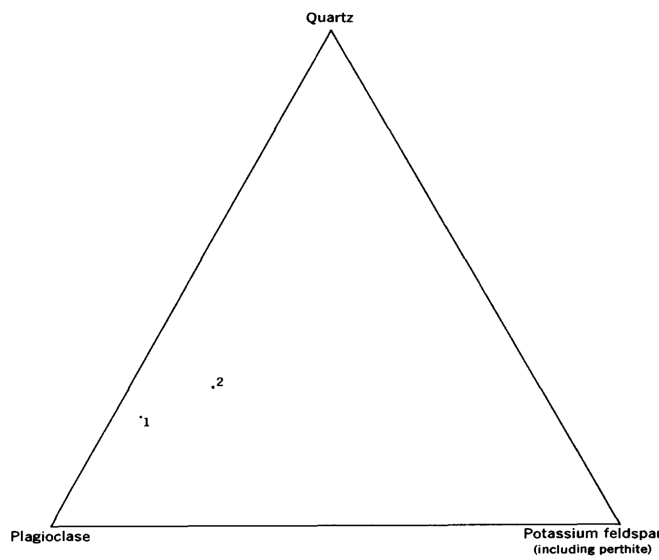


FIGURE 40.—Plot of modes of granodiorite of Cartridge Pass on quartz-K feldspar-plagioclase diagram.

TABLE 15.—*Modal analyses of granodiorite of Cartridge Pass, in volume percent*

[Location of specimens is shown on fig. 23]

Specimen	Quartz	K feldspar	Plagio-clase	Biotite	Horn-blende	Accessory and secondary minerals	Percent anorthite in plagioclase			Remarks
							Rim	Body	Core	
1	15.9	3.8	54.6	15.7	6.8	3.2	28	38	55	Dark, fine-grained and equigranular.
2	22.8	12.3	46.1	12.1	1.7	5.0	21.27	37		Lighter colored, and coarser than specimen 1; equigranular.

MAFIC DIKES

Swarms of dark-colored dikes of dioritic composition are present locally in the Bishop district. The rock in these dikes has been termed "malchite" by Gilbert (1941, p. 784). The most extensive swarm is in the south half of the Big Pine quadrangle and extends southward many miles into the Mount Pinchot quadrangle (Moore and Hopson, 1961). Most of the dikes dip steeply, and within the Big Pine quadrangle strike about N. 70° W., but to the south in the Mount Pinchot quadrangle the average strike is about N. 30° W. A second dike swarm extends westward through the central part of the Tungsten Hills and across the north side of Mount Tom to the Pine Creek pendant. Dikes of this swarm are well exposed in the walls of Pine Creek canyon above the village of Scheelite. These dikes dip steeply and strike westward, parallel with the long axis of the swarm, except just east of the Pine Creek pendant where the dikes swing to the southwest. A third swarm is in the northwest corner of the Mount Goddard quadrangle, along the Glacier Divide. Mafic dikes are also in the cirques at the head of the South Fork of Bishop Creek; dikes in the east face of Hurd Peak can be seen from the trail to Bishop Pass.

The dikes are generally a few inches to a few feet thick; the thickest observed, in the northwest side of Mount Tom and in the north-central part of the Tungsten Hills, are 15 to 20 feet thick. All the rock is fine grained, but differences in grain size and texture are evident. Most dikes are equigranular but some appear porphyritic. Many dikes have been strongly sheared parallel to their walls and appear schistose. Commonly the dark minerals in sheared dikes are in lenticular streaks.

Only a few specimens from the mafic dikes were examined in thin section, and in all these the primary constituents are plagioclase, hornblende, and biotite, accompanied by accessory magnetite, apatite, and sphene. A comparatively coarse grained and little-altered dike from the Tungsten Hills consists of about 70 percent plagioclase and 15 percent each of hornblende and biotite, plus the accessories. The plagioclase is in unoriented thick tabular euhedral to subhedral crystals that average about 2 mm across and 1½ mm thick. Most crystals are discontinuously zoned from cores of about An_{55} to rims of about An_{35}. Interstitial to the plagioclase are jumbled aggregates of the mafic minerals and accessories. The hornblende and biotite in these aggregates are in small anhedral grains.

A sample of dike rock from Fish Springs Hill in the Big Pine quadrangle is highly altered to chlorite and epidote, but scattered relicts of original plagioclase crystals indicate an original weak panidiomorphic-granular texture. Other specimens are all very fine grained and schistose. In these rocks some of the plagioclase is in larger (5 mm) rounded unzoned grains. Hornblende in subhedral to anhedral grains of about the same size range as the plagioclase generally is about as abundant as plagioclase, whereas biotite is much less abundant than either. Streaks of aggregates of the mafic minerals impart a foliation to the rock. The texture of these rocks may be metamorphic rather than igneous—granoblastic rather than allotriomorphic granular.

Apparently swarms of mafic dikes have been intruded at several times during the period of emplacement of the granitic rocks. The oldest swarm, in the Tinemaha granodiorite, does not extend into the adjacent Lamarck granodiorite or Mount Alice mass of quartz monzonite similar to the Cathedral Peak granite, although a few dikes extend into the Red Mountain Creek mass of quartz monzonite similar to the Cathedral Peak granite. Nevertheless, mafic dikes are present in the Lamarck granodiorite in and near Hurd Peak and along the Glacier Divide where they are clearly truncated by the Tungsten Hills quartz monzonite. The dikes that are probably youngest are in the Tungsten Hills quartz monzonite along Pine Creek and in the Tungsten Hills. These relations suggest three periods of dike formation if the Red Mountain Creek mass is assumed to be older than other masses of quartz monzonite similar to the Cathedral Peak granite. The possibility also exists that the Tungsten Hills quartz monzonite along the Glacier Divide is younger than the Tungsten Hills along Pine Creek and in the Tungsten Hills; if so, only two periods of dike emplacement are required.

The source of the dike magma is an unsolved problem. It is difficult to understand how usual processes of magmatic differentiation could produce magma of dioritic composition from a parent magma, which before, and afterward, yielded magma of granodioritic or quartz monzonitic composition. In view of this difficulty two other sources for the mafic dike magma merit consideration: (1) the magma for the dikes was derived from a deep earth layer, and (2) the dike magma was remobilized from masses of older diorite, gabbro, and mafic volcanic rock. The chief argument in favor of the first suggestion is that the dikes are in swarms that extend many miles along the strike. The arguments for the second suggestion are that many dikes are indistinguishable on the basis of their texture and mineral content from mafic inclusions and that in many places, as along the north side of the Glacier Divide and along the South Fork of Big Pine Creek, the dikes are closely associated with abundant inclusions of dark-colored plutonic or volcanic rocks. Further support for the second suggestion is that in gross distribution the dikes are associated with the string of Mesozoic metavolanic roof pendants that extend southward to the Alabama Hills (fig. 7).

BROAD PROBLEMS RELATING TO THE BATHOLITH

SEQUENCE OF EMPLACEMENT

The age relations of the granitic rocks were established wherever possible by observing the contacts between pairs of rocks in the field. The principal features that were used to determine the relative ages were aschistic dikes and apophyses, inclusions, truncated structures, and the presence of characteristic marginal features. Not all of the granitic rocks are in contact with one another, and the relative ages of some that are could not be determined. Contacts that provided no critical data include those where two intrusions meet along a featureless surface, those occupied by dikes or by septa of metamorphic rock, and those obscured by weathering and cover.

The intrusive relations of the granitic rocks, based on the available data and an interpretation of the sequence of intrusion, are given in fig. 41.

The granitic rocks can be conveniently divided on the basis of geographic distribution into three sequences, two of which are not in contact with each other. These are the Tinemaha sequence, which lies entirely south of Big Pine Creek; the Bishop sequence, which lies entirely north of Big Pine Creek; and a third group, which is present both north and south of Big Pine Creek.

In the Bishop sequence the oldest rock is the Wheeler Crest quartz monzonite, followed successively by the Round Valley Peak granodiorite, the Tungsten Hills quartz monzonite, and the granodiorite of Coyote Flat. In the Tinemaha sequence the oldest rock is probably the Inconsolable granodiorite, followed by the Tinemaha granodiorite, and finally by the quartz monzonite of McMurry Meadows. However, the intrusive relations between the Inconsolable and Tinemaha granodiorites are uncertain; in one place the Tinemaha granodiorite appears to intrude the Inconsolable granodiorite, and in another place the reverse is suggested. The quartz monzonite of McMurry Meadows intrudes the Tinemaha granodiorite and is not in contact with the Inconsolable granodiorite.

The relative ages of the rocks in these two sequences is fixed within broad limits by their intrusive relations with rocks of the third group, but their positions as shown in figure 41 also reflect textural and compositonal similarities and differences in the rocks of the two sequences. Thus, the Tinemaha granodiorite is

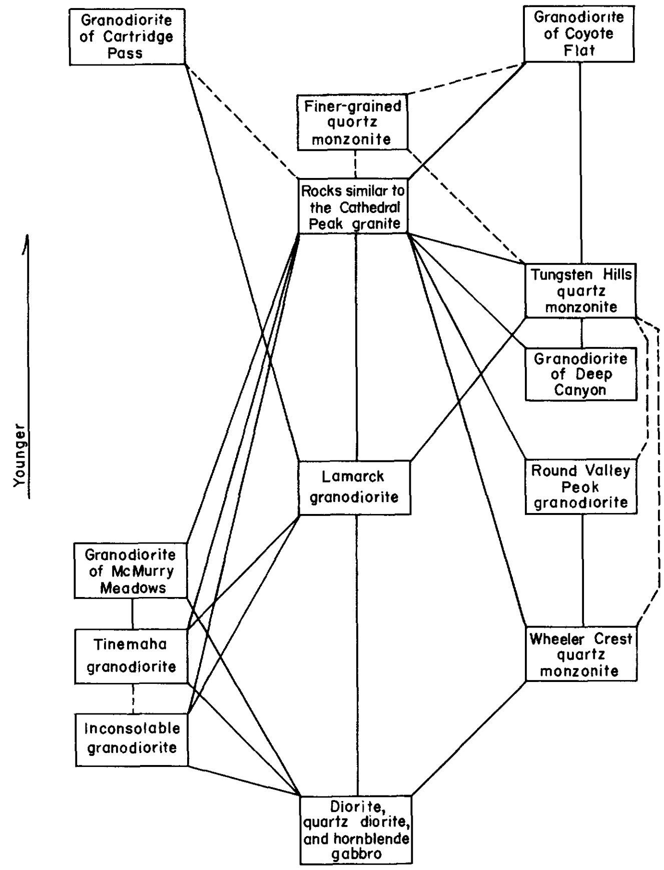


FIGURE 41.—Diagram showing the intrusive relations and probable age sequence of the granitic rocks. Solid lines indicate observed relations; dashed lines indicate probable relations inferred from the map patterns, or from inconclusive field observations.

somewhat similar in appearance and composition to the Wheeler Crest quartz monzonite; in lieu of any other age criteria these rocks are shown to be temporal equivalents.

The rocks of the third group, which occur both north and south of Big Pine Creek, include the oldest and some of the youngest rocks. Masses of mafic rock—diorite, quartz diorite, and hornblende gabbro—are intruded by all the other rocks. The Lamarck granodiorite intrudes both the Inconsolable and Tinemaha granodiorites of the Tinemaha sequence, and is intruded by the Tungsten Hills quartz monzonite of the Bishop sequence. It also is intruded by quartz monzonite similar to the Cathedral Peak granite. With the exception of the granodiorite of Coyote Flat and the possible exception of the granodiorite of Cartridge Pass and finer grained quartz monzonite, alaskite and quartz monzonite similar to the Cathedral Peak granite intrude all the other rocks. The rocks mapped as finer grained quartz monzonite are closely jointed and are weathered to rubble that drifts across and obscures most contacts, but the map pattern suggests that at least some masses intrude quartz monzonite similar to the Cathedral Peak granite. The map pattern also indicates that the granodiorites of Coyote Flat and Cartridge Pass may be intrusive into all the rocks with which they are in contact.

On the whole the rocks appear to have been emplaced generally in order of their increasing silica content. Notable exceptions are the granodiorites of Coyote Flat and Cartridge Pass, which may be early members of a younger sequence that is not otherwise represented.

AGE

The Sierra Nevada batholith has been variously referred to as of Jurassic or Cretaceous age; the geologic relations provide no basis for discriminaton. In the western foothills, granitic rocks intrude Upper Jurassic strata (Mariposa formation), and the metamorphic terrane which the granitic rocks intrude is unconformably overlain by Upper Cretaceous strata (Chico formation). In the Devils Postpile quadrangle, northwest of the Bishop area, granitic rocks intrude fossiliferous strata of Early Jurassic age (Rinehart, Ross, and Huber, 1959), and in the Inyo Mountains they intrude Middle Triassic strata.

Hinds (1934) demonstrated that the Shasta Bally batholith, a granitic mass near Redding, Calif., on the trend of the Sierra Nevada but separated from it many miles, is almost certainly of Late Jurassic age, and suggested that the Sierra Nevada batholith is of the same age. Although this suggestion was accepted by many, the reported presence in Lower California of fossiliferous Lower Cretaceous and lower Upper Cretaceous strata in a terrane intruded by granitic stocks and batholiths raised doubt in the minds of some (Woodford and Harriss, 1938, p. 1330; Böse and Wittich, 1913, p. 394).

The first radiometric determinations of Sierran rocks were of eight samples collected from the Bishop area. The determinations were made by E. S. Larsen, Jr., and associates by using the lead-alpha method (Larsen and others, 1958, table 9, p. 52–53). The results are summarized in table 16. The calculated ages range from 88 to 116 million years, and the average is 105 million years. The lead-alpha ages show little relation to observed intrusive relations. Larsen and others (1952) believe that the error inherent in the lead-alpha method is about 10 percent; this error probably is too large to permit comparing the age of one intrusive with another.

Later the ages of a suite of granitic rocks from Yosemite National Park were determined by the potassium-argon method (Evernden, Curtis, and Lipson, 1957, p. 2120–2125; Curtis, Evernden, and Lipson, 1958, p. 7–9). The calculated ages of these rocks range from 76.9 to 95.3 million years and show remarkable agreement with the order of intrusion as established in the field by Calkins (1930, p. 120–129). The error in the method is thought to be only about 1 to 2 percent. The Cathedral Peak granite was determined to have an age of 83.7 million years; if masses in the Bishop district are correctly assigned to the Cathedral Peak this figure provides an index to the ages of the Bishop district rocks by the potassium-argon method.

The Cretaceous period lasted from 135 to 70 million years ago according to Holmes' revision of his time scale (1960), and from 133 to 58 million years ago according to a calculation by Curtis, Evernden, and Lipson (1958, p. 11). On these scales both the lead-alpha determinations of Bishop district rocks and potassium-argon determinations of the Yosemite rocks fall close to the middle of the Cretaceous. Many difficulties are still to be worked out in integrating radiometric age with paleontologic age, but it seems reasonable to assume that the main part of the Sierra Nevada batholith is of Cretaceous age, although some plutons intrusive into the metamorphic terrane on the west side of the main

TABLE 16.—*Lead-alpha ages of granitic rocks from the Sierra Nevada near Bishop*

[Data from Larsen and others (1958)]

Intrusive	Mass	Location	Probable order of emplacement	Mineral	Activity (α per mg per hr)	Lead (parts per million)	Calculated age (10^6 years)
Quartz monzonite similar to Cathedral Peak granite.	Mount Alice	Big Pine quadrangle SW¼ sec. 26, T. 9 S., R. 32 E. At end of Big Pine Creek road.	6	Zircon	618	26	105
Tungsten Hills quartz monzonite.	Morgan Creek	Mt. Tom quadrangle West of surface workings Pine Creek mine.	5	do	796	37	116
Tungsten Hills quartz monzonite.	Bishop Creek	Mt. Goddard quadrangle NE¼ sec. 20, T. 8 S., R. 31 E. Along Bishop Creek road.	5	do	792	35	110
Lamarck granodiorite	Lamarck	Mt. Goddard quadrangle NW¼ sec. 14, T. 9 S., R. 31 E. NE side of South Lake.	4	Zircon Thorite	400 4,670	15 93 205 88	91
Round Valley Peak granodiorite.	Round Valley Peak.	Mt. Tom quadrangle ¼ mile NE of Rock Creek Lake.	4	Zircon	396	12, 13, 14, 16 (13.8)	88
Granodiorite of McMurry Meadows.		Big Pine quadrangle NE¼ sec. 9, T. 10 S., R. 33 E.	3	Monazite	4,897	234, 238 (236)	100
Tinemaha granodiorite		Big Pine quadrangle NW¼ sec. 14, T. 10 S., R. 33 E.	2	Zircon	331	15, 16 (15.5)	116
Inconsolable granodiorite		Big Pine quadrangle SE cor. sec. 33, T. 9 S., R. 33 E. ¼ mile S. of Third Lake.	1	do	221	10	112

batholith are older (Curtis, Evernden, and Lipson, 1958, p. 6).

BROAD CHEMICAL AND MINERALOGICAL VARIATIONS

A broad trend from calcic and ferromagnesian to silicic and alkalic is shown by the sequence of intrusion. The chemical and modal analyses provide a basis for examining variations in compositions and mineral content in a quantitative as well as a qualitative way. The curves in figure 42 are based on 15 chemically analyzed specimens; 14 of the specimens were collected during this study, and the 15th is a specimen of alaskite similar to the Cathedral Peak granite that was collected by Knopf (1918, p. 68). The chemical analyses are given in table 3. In figure 42 the principal oxides are plotted against SiO_2; the anorthite content of normative plagioclase also is shown.

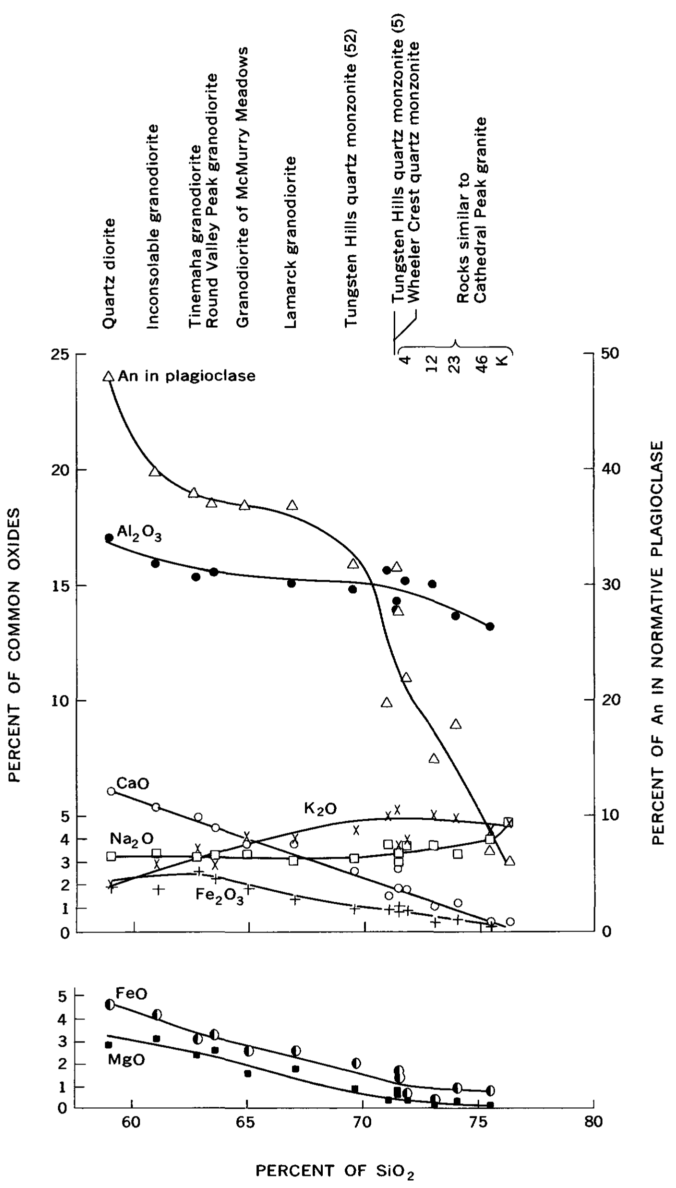


FIGURE 42.—Variation diagram of common oxides in granitic rocks of the Bishop district plotted against SiO_2.

The SiO_2 in the 15 specimens ranges from 59.0 percent in quartz diorite to 76.3 percent in Knopf's specimen of alaskite. Except for minor variations, the curves of the oxides appear regular. As SiO_2 increases, Al_2O_3, CaO, MgO, FeO, and Fe_2O_3 all decrease. Na_2O is nearly constant below 70 percent SiO_2, but increases slightly above 70 percent, and a single analysis suggests it rises sharply above 75 percent. K_2O increases regularly with SiO_2 to 70 percent SiO_2, and is nearly constant above 70 percent SiO_2. The curve for plagioclase composition shows that the normative content of anorthite diminishes with increasing SiO_2, and that the amount of decrease is greater in the lower and higher ranges of SiO_2, and smaller in the intermediate range (62–70 percent SiO_2).

The normative compositions of the 15 chemically analyzed specimens were calculated according to the CIPW method, and are reported in tables 4–9, 11, and 12. Quartz, orthoclase, and plagioclase (albite plus anorthite) were calculated to 100 percent and plotted on a triangular quartz-orthoclase-plagioclase diagram (fig. 43). On this diagram the norms fall in a narrow band that extends from the center of the diagram toward the plagioclase corner. This pattern shows that the ratio between normative quartz and orthoclase is virtually constant, very close to 1:1, and that quartz plus orthoclase is inversely proportional to plagioclase.

The arithmetic average of the modal analyses from different masses and plutons are given in table 17, and the figure 44 the modal averages of quartz, K feldspar, and plagioclase for different plutons and masses have been calculated to 100 percent and plotted on a triangular diagram. In a general way the plot of modes resembles the plot of norms (fig. 43). The chief differences are that the field of modes extends farther away from the plagioclase corner, past the center of the triangle, and that the long axis of the field of modes is slightly inclined to that of the field of norms. Inasmuch as modes are determined from real minerals and norms are theoretically pure molecules computed from chemical analyses, it is hardly expectable that they would plot in the same positions on a triangular diagram.

The differences in the fields of norms and modes probably result chiefly from two factors: (1) crystallization of increasingly larger amounts of biotite in the rocks toward the plagioclase corner, and (2) increasingly larger amounts of albite in K feldspar in rocks away from the plagioclase corner. Crystallization of biotite deletes K_2O from the magma, and results in

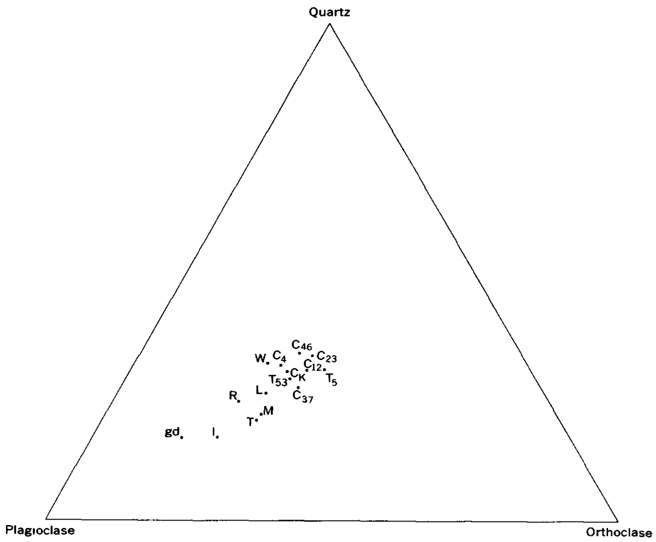
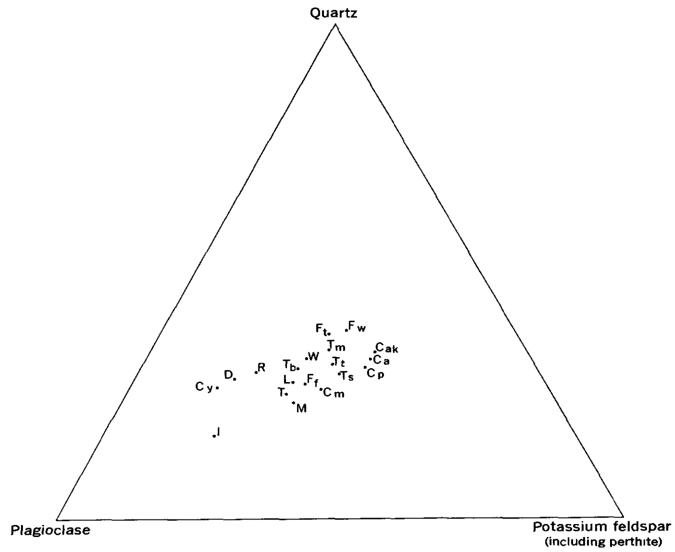


FIGURE 43.—Plot of norms of granitic rocks on quartz-orthoclase-plagioclase diagram; gd, quartz diorite; T, Tinemaha granodiorite; I, Inconsolable granodiorite; M, granodiorite of McMurry Meadows; W, Wheeler Crest quartz monzonite; R, Round Valley Peak granodiorite; L, Lamarck granodiorite; T, Tungsten Hills quartz monzonite; C, Rock similar to Cathedral Peak granite.



R, Round Valley Peak granodiorite
L, Lamarck granodiorite
M, Granodiorite of McMurry Meadows
W, Wheeler Crest quartz monzonite
T, Tinemaha granodiorite
I, Inconsolable granodiorite
C_y, Granodiorite of Coyote Flat
D, Granodiorite of Deep Canyon
Finer grained quartz monzonite:
- F_w, Sherwin Hill mass
- F_f, Freeman Creek, Shannon Canyon, and Sugarloaf masses
- F_t, Taboose mass

Rocks similar to the Cathedral Peak granite:
- C_m, Lobes of Mono Recesses mass
- C_a, Mount Alice mass
- C_p, Palisade Creek mass
- C_{ak}, Alaskite masses

Tungsten Hills quartz monzonite:
- T_m, Morgan Creek and Pine Creek masses
- T_b, Basin Mountain mass
- T_t, Tungsten Hills and Bishop Creek masses
- T_s, Shannon Canyon mass

FIGURE 44.—Plot of arithmetic modal averages of quartz, K feldspar, and plagioclase in different plutons.

less modal K feldspar than normative orthoclase; because biotite increases in abundance with plagioclase, toward the plagioclase corner the modal field is shifted progressively away from orthoclase as compared with the normative field. Solid solution of albite in K feldspar increases the amount of modal K feldspar over normative orthoclase. The amount of albite in K feldspar is greatest in rocks having the least plagioclase because these rocks contains the most K feldspar and because the K feldspar is conspicuously perthitic. The result is that the field of modes is expanded at the end most distant from plagioclase away from plagioclase and toward K feldspar.

In figure 45 the average percent of modal quartz, K feldspar, biotite, and hornblende, given in table 17, are plotted against average percent of modal plagioclase. The diagram shows that quartz and K feldspar are about equally abundant in all rocks and that they decrease systematically as plagioclase increases. Biotite and hornblende increase with plagioclase, and biotite is, on the average, about 4 percent more abundant than hornblende. A notable exception is the Tinemaha granodiorite, which contains more hornblende than biotite. Hornblende generally is present only in trace amounts in rocks that contain less than 37 percent of plagioclase, although a few masses of rock having less than this amount of plagioclase, such as the Morgan Creek and Pine Creek masses of Tungsten Hills quartz monzonite, contain several percent of hornblende. Thirty-seven percent of the modal plagioclase falls in the lower part of the quartz monzonite field near the boundary between the quartz monzonite and granodiorite fields, and the presence or absence of appreciable hornblende is a useful and generally reliable criteria for distinguishing between granodiorite and quartz monzonite among eastern Sierra granitic rocks. Calcic quartz monzonite generally contains a few percent hornblende, however.

SYSTEMATIC COMPOSITIONAL VARIATIONS WITHIN INDIVIDUAL MASSES

Some masses of granitic rock are obviously compositionally zoned, and many others that are superficially of uniform appearance and mineral composition are found on careful study to vary systematically in composition from place to place. Compositional variations are either concentric or lateral.

The granodiorite of McMurry Meadows is an excellent example of a concentrically zoned intrusion;

TABLE 17.—*Arithmetic averages of modal analyses for different plutons, in volume percent*

[Percentages of biotite, hornblende, and accessory and secondary minerals were determined from thin section, and their sums do not necessarily equal the percentages of total mafic, which were determined chiefly from stained slabs]

Formation and mass	Specific gravity	Quartz	K feldspar	Plagioclase	Biotite	Hornblende	Accessory and secondary minerals	Total mafic
Granodiorite of Coyote Flat	2.71	22.1	12.8	47.9	8.7	4.7	3.9	17.3
Finer grained quartz monzonite:								
Sherwin Hill mass	2.63	36.3	31.0	28.1	3.3	.4	1.0	4.7
Freeman Creek, Middle Fork, and Sugarloaf masses	2.64	26.0	28.8	40.0	4.4	0	.8	5.2
Taboose mass	2.62	35.7.	28.4	31.4	3.6	0	.8	4.4
Rocks similar to the Cathedral Peak granite:								
Lobes of Mono-Recesses mass	2.62	25.1	32.7	38.3	2.9	0	2.2	3.8
Mount Alice mass	2.61	31.2	38.8	27.4	2.7	0	2.8	2.6
Palisade Creek and Red Mountain Creek masses	2.60	30.0	38.3	29.2	2.7	.2	1.7	2.5
Alaskite	2.60	33.0	38.6	25.8	1.8	.2	2.2	2.9
Tungsten Hills quartz monzonite:								
Morgan Creek and Pine Creek masses	2.62	31.9	29.1	32.5	4.8	3.9	1.4	6.5
Basin Mountain mass	2.65	28.4	25.4.	38.4	5.9	1.4	2.7	8.2
Tungsten Hills and Bishop Creek masses	2.63	29.8	31.0	33.2	5.5	.2	2.1	6.0
Shannon Canyon masses	2.62	27.7	33.5	32.9	4.0	.1	2.9	5.9
Granodiorite of Deep Canyon	2.70	23.6	14.8	45.4	9.3	3.7	3.1	16.1
Lamarck granodiorite:								
Chickenfoot Lake mass	2.71	21.6	27.1	36.2	6 1	8.1	3.2	15.1
Lamarck mass (main part)	2.70	23.1	21.0	43.2	6.8	3.3	1.5	12.9
Lamarck mass (south part)	2.65	28.6	26.5	38.2	5.3	.6	1.0	6.7
Round Valley Peak granodiorite	2.69	26.8	18.7	44.4	7.2	3.6	1.9	10 1
Wheeler Crest quartz monzonite	2.65	29.3	25.3	35.3	5.8	1.6	1.6	10.1
Granodiorite of McMurry Meadows	2.70	20.5	26.4	40.1	8.0	3.1	1.6	12.7
Tinemaha granodiorite	2.72	21.3	23.6	39.0	4.6	6.0	3.4	16.1
Inconsolable granodiorite	2.75	13.1	15.6	49.3	10.4	5.7	2.9	22.0

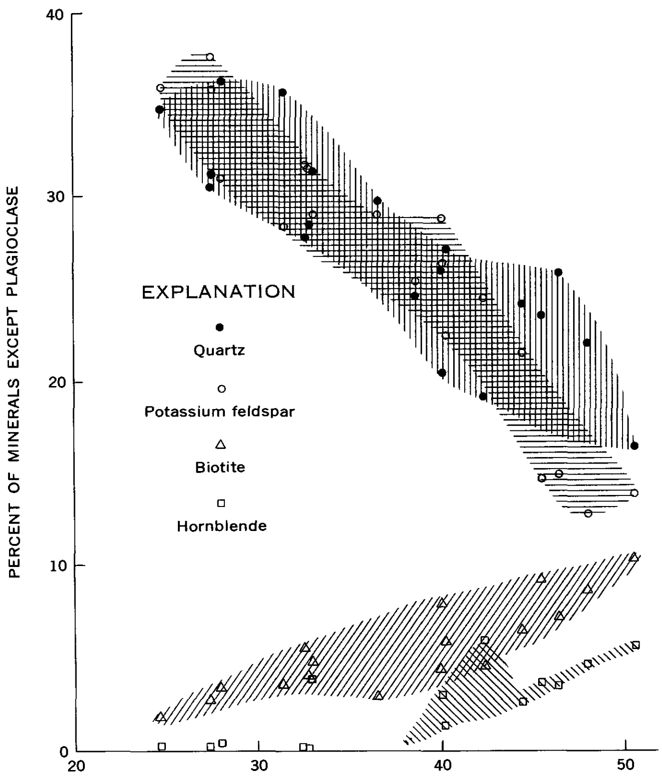


FIGURE 45.—Average percent of modal quartz, K feldspar, biotite, and hornblende in different plutons plotted against average percent of modal plagioclase.

the rim is granodiorite rich in plagioclase and ferromagnesian minerals, and the core is light-colored quartz monzonite. The specific gravities of specimens from the margin are greater than of specimens from the core. Concentric zoning is also shown by the granodiorite of Cartridge Pass, the Round Valley Peak granodiorite, and to a lesser degree, by the granodiorite of Coyote Flat. Even the Tinemaha granodiorite and the main northern part of the Lamarck granodiorite have core areas in which the specific gravity is less than in marginal areas.

Lateral variations are obvious chiefly in the largest intrusive masses, especially in the Lamarck granodiorite and the Tungsten Hills quartz monzonite. The narrow southern extension of the Lamarck granodiorite south of Hurd Peak and Bishop Pass is notably lighter in color, lower in specific gravity, and less calcic than the thicker northern part. The granodiorite part of the Basin Mountain mass of Tungsten Hills quartz monzonite that lies west of Desolation Lake is darker and heavier, and contains more plagioclase than most of the rest of the mass. The Tungsten Hills quartz monzonite also may grade in the Tungsten Hills into the largest mass of granodiorite of Deep Canyon, but this relation was not established.

In most intrusive masses, systematic changes in composition probably reflect differentiation during cooling, but in some masses contamination by wall rocks has been the cause of compositional differences. The usual basis for distinguishing between the effects of contamination and of differentiation is the presence or absence of rock of the proper composition and in the proper position to indicate the cause of compositional variations in an intrusive. In places where contamination has been effective in modifying the composition of a granitic intrusive, the contaminating rock generally is either early mafic intrusive rock or mafic volcanic rock. A good example of contamination of an intrusive by mafic igneous rock is in the Chickenfoot Lake mass of Lamarck granodiorite, which is progressively darker and more calcic toward its southern margin where it is in contact with hornblende gabbro. Calcium-rich wall rocks also seem likely to influence the composition of a magma.

Where no evidence is found of contamination by wall rocks, crystal fractionation is usually invoked to explain systematic variations in composition. In a granitic intrusive, early formed crystals of plagioclase, hornblende, and the accessory minerals are presumed to settle or to accrete to the walls. Convection could be effective in the marginal accretion of crystals by bringing crystals into contact with the walls.

A possible contributing process may be that of thermodiffusion. At one time the Soret effect was thought to be a principal cause of differentiation in magmas, but this explanation has been in disfavor for many years because of the presumed slow velocity of diffusion in viscous silicate melts. Wahl (1946) is one of the few advocates in recent years of diffusion in magmatic differentiation. My concept of the kind of diffusion that might operate is that of a gradual shift of calcium and the femic constituents outward from the core of the magma body in response to impoverishment of these constituents in the liquid phase near the margins as a result of crystallization. Even though the rate of diffusion may be very slow—perhaps only a few centimeters a year—over the period of crystallization of a pluton, which may extend over hundreds of thousands of years, diffusion could play a significant role in differentiation.

Both crystal fractionation and thermodiffusion permit formation of rock in the margins of an intrusion that is more salic than the magma, and can be responsible for a concentrically zoned intrusive. The marginal part of a zoned intrusion does not necessarily represent the original composition of the magma, even if it is fine grained as compared with the interior of the intrusion (unless it is glass). On the contrary, if it can be assumed that either or both convection and high viscosity inhibited the sinking of significant amounts of crystals, the initial composition of the magma will be represented more accurately by the average composition of the exposed rocks.

Lateral changes in composition from one side or one end of an intrusive to the other can be most easily explained if it is assumed that after partial differentiation and solidification of the margins, renewed movements of the still liquid magma in the central part took place. Abrupt contacts could result from movements of the magma after consolidation of the margins. A local internal contact in the Tungsten Hills between two different facies of the Tungsten Hills quartz monzonite could be a result of such movement. However, abrupt contacts would result only if parts of the intrusive were rather completely crystallized; otherwise gradational contacts should result. The Lamarck granodiorite may have been intruded initially as an elliptical body whose southern limit was near Bishop Pass. After partial crystallization in the margins and differentation of the core magma to quartz monzonite composition, the magma may have broken out of its chamber southeastward, probably along a fracture, to form the felsic southern extension. Inasmuch no sharp contacts have been found, it is not likely that the margins of the original elliptical body were entirely solidified. An excellent example of a concentrically zoned intrusive in which the residual core magma moved repeatedly is the White Creek batholith in British Columbia (Reesor, 1958).

CORRELATION OF NORMATIVE COMPOSITIONS OF THE GRANITIC ROCKS WITH EXPERIMENTAL DATA

Since granitic rocks consist chiefly of quartz and feldspar, they can be approximately experimentally by mixtures of SiO_2, $KAlSi_3O_8$, $NaAlSi_3O_8$, and $CaAl_2Si_2O_8$, hereafter referred to as Qz, Or, Ab, and An. In the east-central Sierra Nevada the content of normative feldspar and quartz ranges from 79.4 in a specimen of quartz diorite to 99.3 in a specimen of alaskite. A considerable amount of experimental work has been carried on which such artificial mixtures, particularly by N. L. Bowen, J. F. Schairer, and O. F. Tuttle at the Geophysical Laboratory of the Carnegie Institution of Washington, D.C. These three workers, together with R. R. Franco, have experimentally studied the liquidus relations at atmospheric pressure in all the binary and ternary combinations of this four-component system. The ternary system Qz-Or-Ab was studied by Bowen (1937), the system Qz-Or-An by Schairer and Bowen (1947), the system Or-Ab-An by Franco and Schairer (1951), and the system Qz-Ab-An by Schairer (written communication, 1957). Bowen and Tuttle have also determined certain liquidus relations in systems that include H_2O under pressure (Bowen, 1954; Tuttle and Bowen, 1958). Yoder, Stewart, and Smith (1957) have studied the system Or-Ab-An-H_2O at 5,000 bars, and Stewart (1958) has studied the system Qz-An-H_2O at various pressures.

The liquidus relations in the system Or-Ab-An-Qz-H_2O at 5,000 bars H_2O pressure are shown in figure 46. In figure 46*A* projections of norms of the granitic rocks from the east-central Sierra Nevada are shown on the faces of the tetrahedron. The positions of the norms can be misleading if due allowance is not made

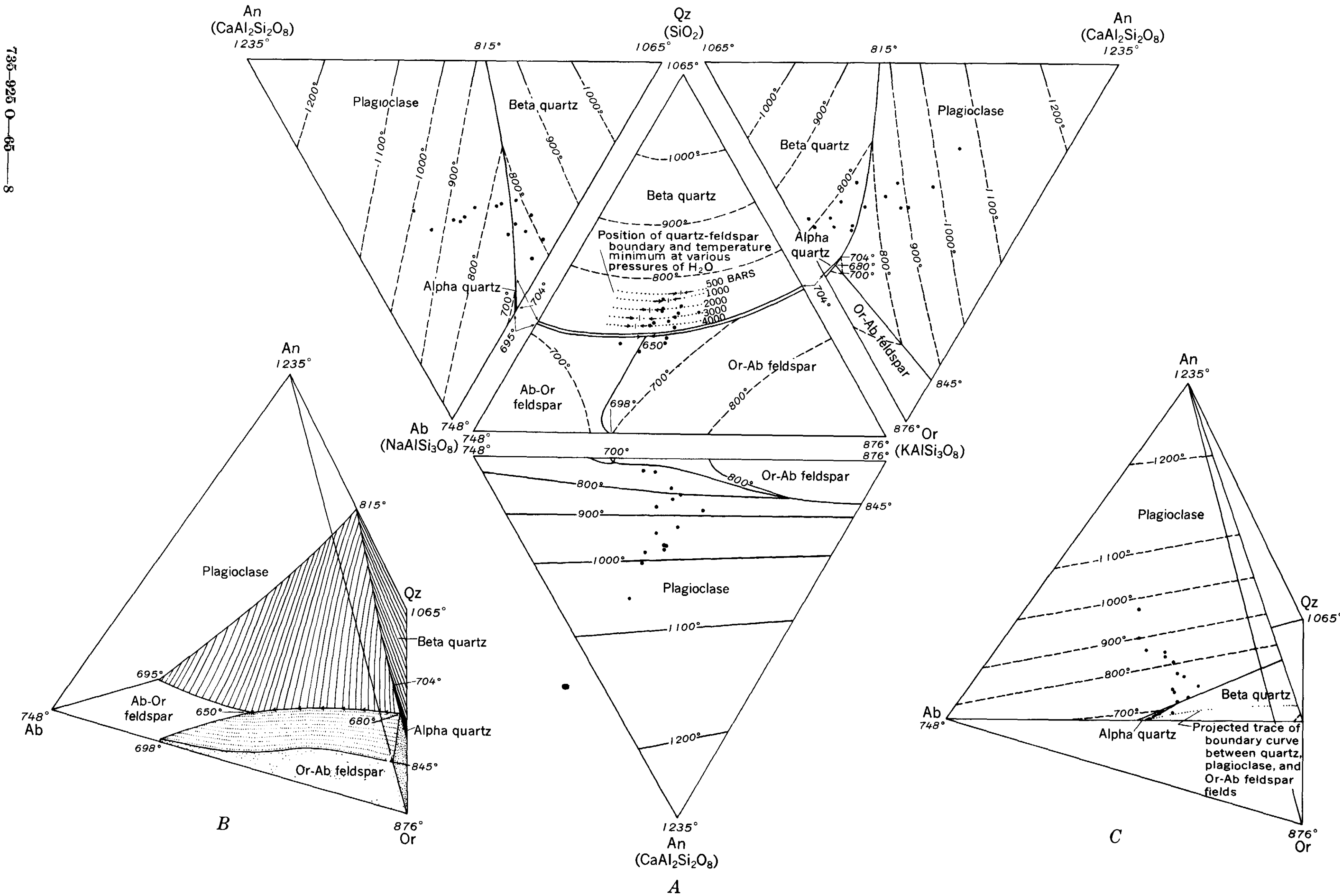


FIGURE 46.—Tetrahedron showing the liquidus relations in the system Or($KAlSi_3O_8$)-Ab($NaAlSi_3O_8$)-An($CaAl_2Si_2O_8$)-Qz(SiO_2)-H_2O at 5,000 bars H_2O pressure. All components are in weight percent. *A*, Faces of tetrahedron showing field boundaries (heavy lines), isotherms (light lines and dashes), and projections of norms of granitic rocks from the east-central Sierra Nevada (points). Or-Ab-An face is after Yoder, Stewart, and Smith (1957). An-Qz join is after Stewart (1958). Inversion temperature of quartz is after Yoder (1950). Temperature and position of Or-Ab-Qz minimum and positions and temperatures of other field boundaries were projected by D. B. Stewart (written communication, 1958) from data of Tuttle and Bowen (1958) and H. R. Shaw (Stewart, written communication, 1960). Quartz-feldspar boundary and temperature minimum at various H_2O pressures after Tuttle and Bowen (1958). Isotherms, except on Or-Ab-An face and along An-Qz join were constructed with reference to the above data and are approximate. *B*, Three-dimensional drawing of tetrahedron showing field boundaries. The minimum lies at 650°. *C*, Bisecting section through tetrahedron passing through Ab and An corners and intersecting the Or-Qz join at midpoint. On the section are plotted isotherms; field boundaries between alpha quartz, beta quartz, and plagioclase; projections of norms of rocks from the east-central Sierra Nevada; and projected trace of boundary curve between quartz, plagioclase, and Or-Ab feldspar fields.

for the fact that they are projections. Their positions were determined by calculating to 100 percent the three constituents represented in each face. From a visual standpoint, the points are plotted as they would appear within the tetrahedron if they were viewed by looking at each face with the eye at the opposite corner. As a consequence of the construction, the spread of norms on the tetrahedron faces is greater than the true field within the tetrahedron. The tetrahedron is shown in three-dimensional form in figure 46*B*. The plagioclase field occupies most of the tetrahedron, a quartz field occupies a part of the tetrahedron near the Qz corner, and a flattish Qz-Ab feldspar field extends outward from the Or corner toward Ab and Qz. Because most of the granitic rocks contain almost equal amounts of normative quartz and orthoclase, the plane within the tetrahedron that approximates most closely the true field of the norms is one which bisects the tetrahedron as shown in figure 46*C*. The norms are projected only short distances to this plane and appear in very nearly their true positions. The plane intersects the boundary between the plagioclase and quartz fields at a small angle, and does not intersect the Or-Ab feldspar field at all. However, the boundary line between quartz, Or-Ab feldspar, and plagioclase converges on the section at a small angle and very nearly intersects it at the minimum. Because this boundary line is so close, it is projected onto the section.

It is apparent from the section that the norms fall either within the plagioclase field or very close to the field boundary between quartz and plagioclase. In general, but not in detail, the oldest rocks are those whose norms plot highest in the diagram and nearest An, and the youngest plot closest to the temperature minimum. The pattern of norms is very close to that of a theoretical path for the composition of a differentiating liquid, and strongly supports the view that the granitic rocks in the east-central Sierra Nevada are fundamentally of magmatic origin (Barth, 1952, fig. 20, p. 101). Magma of the composition of the norm having the most An (quartz diorite) would first crystallize calcic plagioclase at the H_2O saturated liquidus (slightly below 1000°C). Crystallization of the calcic plagioclase would cause the composition of the remaining liquid to be displaced away from An and, to a lesser degree, Ab. With falling temperatures the amount of Ab in the crystallizing plagioclase would increase, which would cause the composition of the remaining liquid to change along a curved path, convex toward Ab. On reaching the quartz or Or-Ab feldspar field boundary, a second mineral would begin to crystallize. If the quartz boundary surface were intersected first, as appears to have happened, quartz would begin to crystallize in addition to plagioclase, and the composition of the melt would move away from the SiO_2 corner of the tetrahedron as well as from the An and Ab corners in a path that would be determined by the relative amounts of quartz and plagioclase crystallizing and by the composition of the plagioclase. This path would ultimately so change the composition of the remaining liquid that Or-Ab feldspar also would begin to crystallize. The liquid would then lie on the common boundary line of the quartz, plagioclase, and Or-Ab feldspar fields. Inasmuch as the path along this boundary curve is toward Ab, and because at the low temperature (about 700°C) the plagioclase crystallized would be Ab rich, much more quartz and Or-Ab feldspar would crystallize than plagioclase. At the temperature minimum the relative amount of albitic plagioclase crystallizing probably would increase; the theoretical composition of the liquid and of the crystallizing constituents would be Ab 44.5 percent, Or 28.5 percent, and Qz 27.0 percent. However, extreme fractionation would have to occur for the minimum to be reached, and separation of a gas phase may affect the ratio of constituents in the liquid.

If no crystals were subtracted from the melt during cooling and crystallization, the final rock would have the same composition as the initial melt. However, if during crystallization some crystals were removed, the bulk composition of melt plus the remaining crystals would keep changing. If parts of this residual magma were intruded at different times, the rocks formed from them would have different compositions.

The norms as plotted on the section (fig. 46*C*) seem to be in a plane parallel to, and nearly coinciding with, the position of the field boundary between quartz and Or-Ab feldspar at 5,000 bars of water-vapor pressure; however, it must be remembered that the norms are projected, even though for only short distances, and the field boundary is cut at an unfavorable angle for accurately depicting the position of the norms with respect to this field boundary. It is therefore possible that the position of the quartz Or-Ab feldspar field boundary at some other vapor pressure would fit the distribution of norms better. The position of the boundary shifts toward the SiO_2 corner at lower pressures of H_2O, as is shown in figure 46*A* (Bowen, 1954).

Goranson (1931) determined the solubility of water in glass made from granite from Stone Mountain, Ga., to be—at 800° C—about 8.9 percent at 3,000 bars and 9.3 percent at 4,000 bars. These figures are in close agreement with figures obtained by Tuttle and Bowen (1958, fig. 28, p. 58) for melts of minimum melting composition in the system Or-Ab-Qz-H_2O. The amount of H_2O present in an undifferentiated magma is probably less than the maximum amount that is soluble in the magma. In all probability an amount of H_2O equal to the amount soluble is attained only in the late stages of crystallization, after the crystallization of substantial amounts of anhydrous or nearly anhydrous minerals.

The diagrams indicate that plagioclase began to crystallize in the most calcic magma represented by a point on the diagrams at about 970° C and that quartz and alkali feldspars completed crystallization in the most silicic magma represented at about 650° C. However, these temperatures would be correct only if the magmas represented were saturated with H_2O throughout crystallization and only if the additional components in the magma not represented on the diagrams did not affect the crystallization temperatures. Experimental evidence suggests that additional components may lower the crystallization temperatures markedly (D. B. Stewart, written communication, 1960). If saturation of the melts with H_2O was achieved only in the later stages of crystallization, the first plagioclase may have crystallized at temperatures higher than 970° C.

Another possibility is that the most calcic rocks represent collections of early crystallized crystals in magma less calcic than the rock. This possibility is partly negated by the fact that the norms of these rocks plot along the theoretical path followed by a silicate melt during cooling.

If the pressure of H_2O in the closing stages of differentiation was about 5,000 bars, as the data suggest, and if the pressure of H_2O at that time was roughly equal to the rock pressure resulting from load, the depth of differentiation was at least 11 miles. The depth of final crystallization could have been less, for masses of magma from a parent body differentiating at depth could have moved higher into the crust with only local additional differentiation. Nevertheless, the zonation of many plutons suggests that significant differentiaton took place at the present level of exposure. I infer from these considerations that the present level of exposure was about 11 miles beneath the surface of exposure at the time of emplacement of the granitic rocks.

COMPARISON OF COMPOSITIONAL TRENDS WITH TRENDS OF GRANITIC SUITES FROM OTHER AREAS

For comparison with suites of granitic rocks from other areas, plots of norms on triangular Qz-Or-Pl (An+Ab) diagrams (fig. 47) are used because significant similarities and differences can be represented readily without resorting to oxide variation diagrams or four-component Qz-Or-Ab-An diagrams. Plots of modes would be useful for making comparisons of suites of rocks, but unfortunately few suites have been analyzed with sufficient accuracy and in sufficient detail.

Plots of norms of granitic rocks from various areas in the western United States and Canada are shown in figure 47. All the plots are elongate and extend away from the Pl corner to an area of convergence near the center of the diagram. Some plots extend nearly to the Pl corner, whereas others, including that of the eastern Sierra Nevada, terminate at considerable distances from the Pl corner. In part the failure to extend closer to the Pl corner results from a lack of chemical analyses rather than from a lack of rocks that would plot near the Pl corner. Nevertheless, the absence of analyses of plagioclase-rich rocks generally reflects a paucity of such rocks in the terrane. In the eastern Sierra Nevada, plagioclase-rich rocks are generally in small bodies of variable texture and percentage mineral content; consequently, analyses of these rocks are not of much significance for general purposes, and few have been made. On the other hand, in granitic areas such as the batholith of southern California, where plagioclase-rich rocks are in large bodies and constitute a significant part of the terrane, abundant analyses have been made.

All the plots except those of the Idaho and Boulder batholiths, which are irregular, converge at the center of the diagram, and the most elongate ones also converge at the Pl corner. Between these two areas of convergence the plots follow different trends. On the basis of these trends the plots can be categorized into three groups, although it is recognized that the groups are completely gradational. The plots of the batholith of southern California, the Bald Mountain batholith, and the Mount Garibaldi area lie along trends that extend away from Pl in a direction toward Qz; at about 30 to 40 percent of quartz they bend toward the center of the diagram. In contrast, the plots of the Laramide stocks of Colorado and New Mexico and of rocks from the Kuskokwim region, Alaska, lie along trends that extend from the Pl-Or sideline between 20 and 50 percent of Or toward the center of the diagram. The third group, which includes the eastern Sierra Nevada and the Cowichan Lake area of Vancouver Island, British Columbia, is intermediate to the other two; these plots of norms extend away from Pl almost directly toward the center of the diagram.

The patterns of the trends show close correlation with experimental data. In figure 48, the interpreted median lines of the various fields of norms are plotted on a triangular quartz-orthoclase-plagioclase diagram on which the experimentally determined boundary curve between quartz and feldspar is shown for P_{H_2O} 1,000 and 5,000 bars, since this boundary shifts with differences in water vapor pressure. The plagioclase of the rocks is, of course, not albite, but the field boundaries shown, nevertheless, are as good approximations

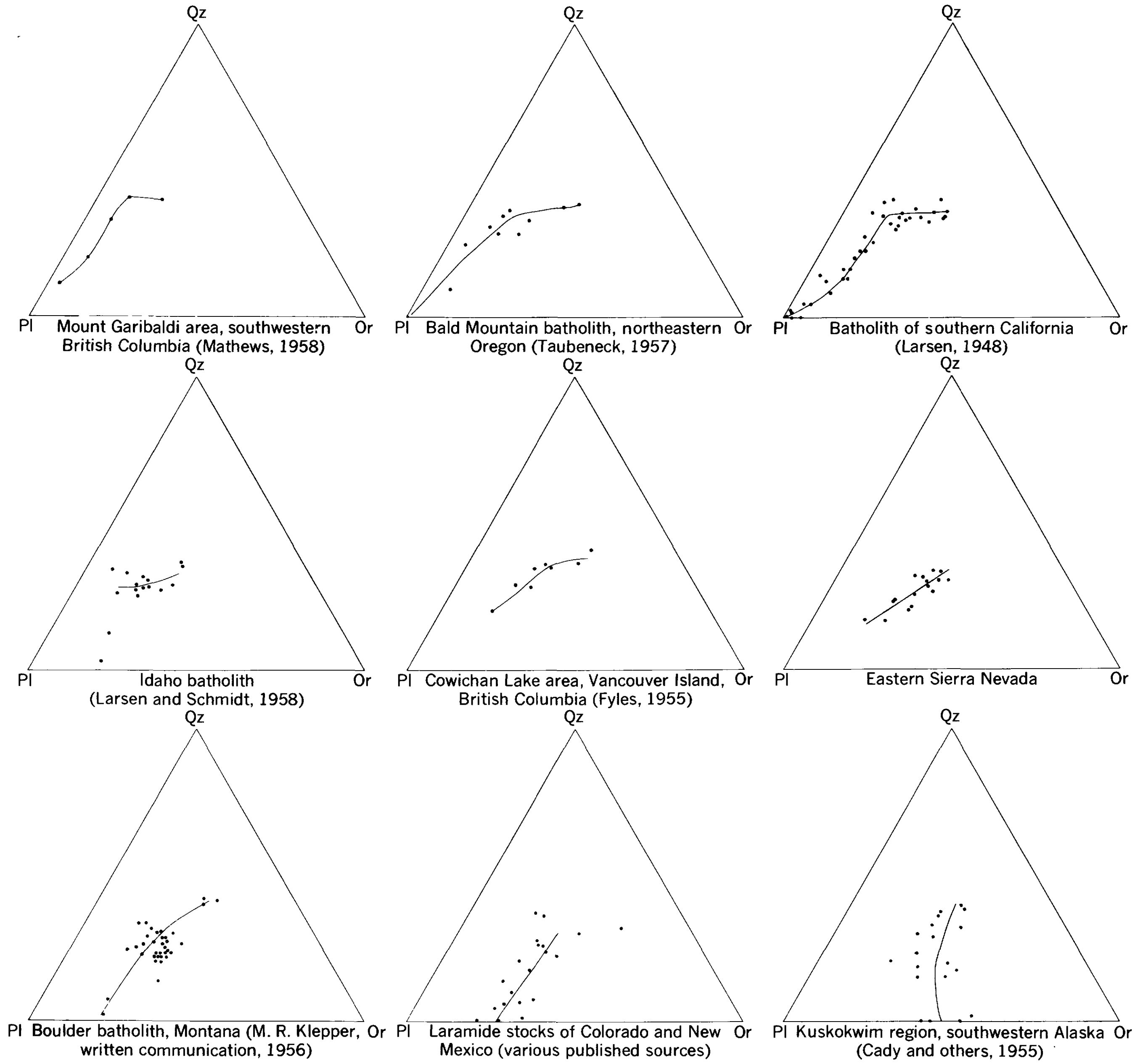


FIGURE 47.—Plots of norms for areas of granitic rocks in the western United States and Canada.

as can be made. By comparing figure 47 with figure 46, it can be seen that the norms of most suites of rocks lie either within or very close to the plagioclase (albite) field. The field boundaries of the field of norms approximate the positions of the field boundaries between quartz and feldspar and between plagioclase and Or-Ab feldspar.

It is known that in a general way the norms that plot closest to the Pl corner in each suite were the earliest rocks emplaced, and that those that plot near the center of the diagram were the latest, although exceptions to this generalization no doubt exist. The distributions of the norms and the sequence of emplacement of the rocks which they represent are strong arguments for considering all the different suites to be chiefly products of differentiation under conditions of general crystal-liquid equilibrium.

The different paths followed by the various plots can be explained as a result of differences in the proportion of normative quartz to orthoclase in the melt. These differences may have been original, resulting from differences in the bulk composition of the rock that was liquified, or they may have resulted from early crystallization of some such mineral as horn-

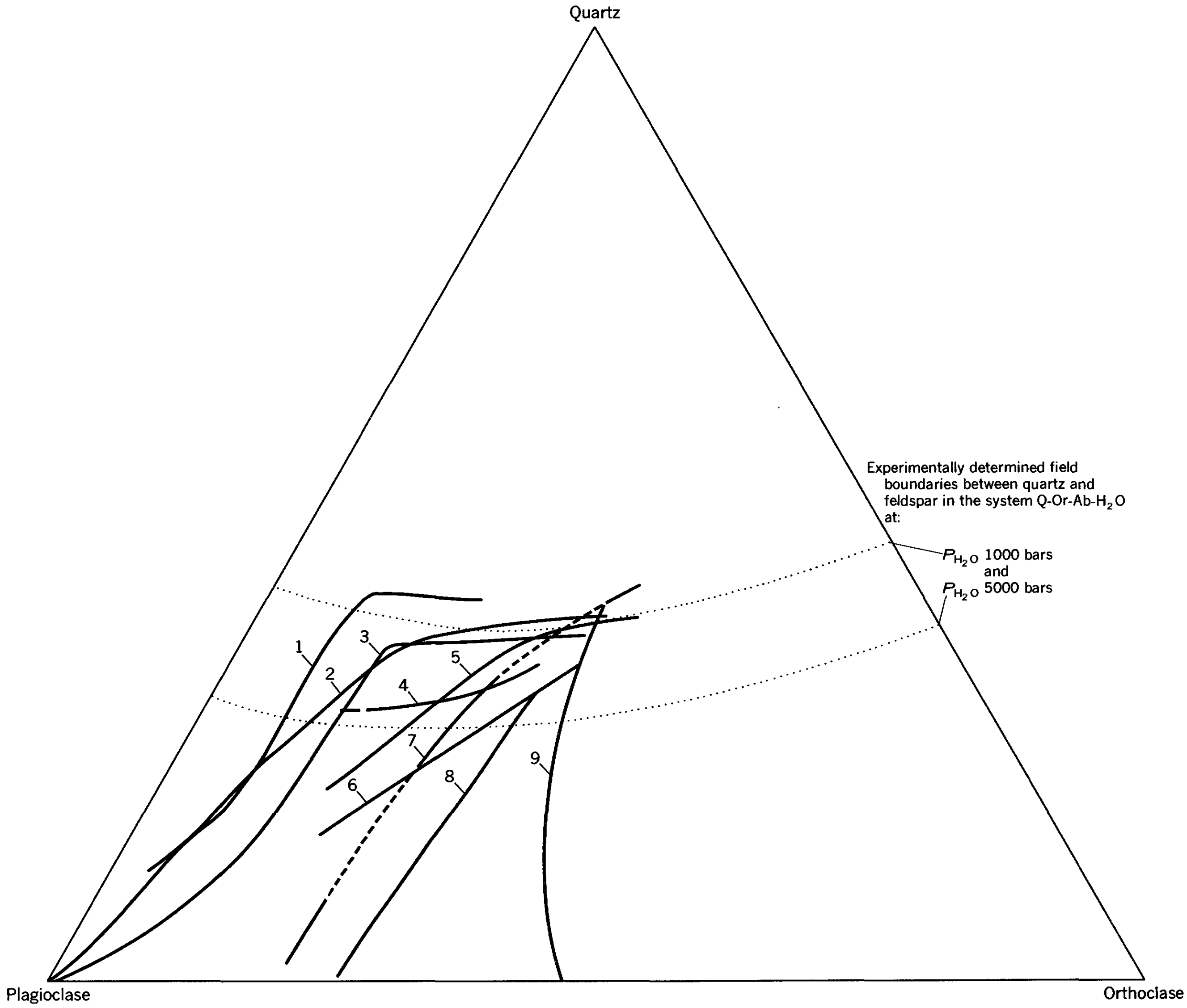


FIGURE 48.—Composite of median lines through fields of norms shown on figure 47. 1, Mount Garibaldi area, southwestern British Columbia. 2, Bald Mountain batholith, northeastern Oregon. 3, Batholith of southern California. 4, Idaho batholith. 5, Cowichan Lake area, Vancouver Island, British Columbia. 6, Sierra Nevada batholith. 7, Boulder batholith, Montana. 8, Laramide stocks of Colorado and New Mexico. 9, Kuskokwim region, Alaska.

blende, which could have altered the proportion of normative quartz to orthoclase in the liquid. A very slight change could have a significant effect on the path of differentiation. Moore (1959) has suggested that the different trends of crystallization in the granitic rocks of different parts of the western United States are related to the composition of the initial melt and that this initial composition is a product of the position of the magma chamber relative to the edge of the continent.

The median lines (fig. 48) of the batholith of southern California, the Bald Mountain batholith, the Mount Garibaldi area, and the Cowichan Lake area at the ends near the center of the diagram lie closer to the quartz-feldspar field boundary at 1,000 bars than at 5,000 bars. Smaller water-vapor pressure could reflect either smaller load pressure than pertained during differentiation of the eastern Sierra Nevada granitic rocks, or a deficiency of water at the same load pressure, or both.

CONTACTS BETWEEN DIFFERENT GRANITIC ROCKS

Most observed contacts between the different granitic rocks are sharp and nearly vertical; plutonic breccias occur locally along only a few contacts. Generally the traces of contacts between granitic rocks are flowing curves, but in places straight-line segments meet at sharply angled corners. Dikes of the later rock in the earlier one are present near contacts, but are rarely abundant. At many contacts adjacent granitic rocks are separated by thin discontinuous septa of metamorphic rock or by later aplitic dikes. A notable feature along all contacts between granitic rocks is the absence of any evidence of chemical reaction between the rocks in contact.

Typical sharp contacts are shown in figure 22. Both contacts shown were followed in the field for many miles. The contact between the Mount Alice mass of quartz monzonite similar to the Cathedral Peak granite and the Inconsolable granodiorite is the same everywhere that it was examined; the contact between the Inconsolable and Tinemaha granodiorites, on the other hand, is occupied intermittently by thin discontinuous metamorphic septa, only the largest of which are shown on the map (pl. 4). A typical septum that crosses Bishop Creek just below Camp Sabrina separates Tungsten Hills quartz monzonite from Lamarck granodiorite (pl. 1). Steep joints appear to be entirely later than the granitic rocks, and therefore do not influence the intrusive contacts; however, some gently dipping joints are early and in cliff sections offset contacts in places.

The most extensive zone of plutonic breccia, and the only one of sufficient size to be represented on the quadrangle maps, is between the Tungsten Hills quartz monzonite and the Lamarck granodiorite in the vicinity of Piute Pass. A good deal of metamorphic and dioritic or mafic hybrid rock is present along the contact, and the metamorphic and mafic rock and the Lamarck granodiorite are intricately penetrated in the breccia zone by anastomosing dikes and apophyses of Tungsten Hills quartz monzonite. Mixed rock also is present along Big Pine Creek on the hillside south of Big Pine Lodge, where scattered angular fragments of the Tinemaha granodiorite are present in the Mount Alice mass of quartz monzonite similar to the Cathedral Peak granite. Farther west, south of Second Lake along the North Fork of Big Pine Creek, the Inconsolable granodiorite adjacent to the Mount Alice mass of quartz monzonite similar to the Cathedral Peak granite is shattered, and the interstices between fragments are filled with felsic rock and with milky quartz (fig. 25). Local small-scale segregations of the light and dark minerals in the Inconsolable granodiorite have formed; some of the segregations are linear and follow lines of fracture and some are ovoid. The shattered zone coincides with a bend in the contact in which the quartz monzonite lies on the convex side.

Aschistic dikes, though not abundant, generally are common enough to provide a means of determining the relative ages of the granitic rocks. In general, masses of rock similar to the Cathedral Peak granite are accompanied by more dikes than the other intrusives, and marginal dike swarms are present along the northern margins of the Palisade Creek mass and the south lobe of the Mono Recesses mass.

CONTACTS BETWEEN GRANITIC ROCKS AND METAMORPHIC ROCKS OR DIORITE

Most contacts between granitic rocks and metamorphic rock or diorite are sharp and clean, but in places irregular contacts and broad zones of mixed rock are present. The sharpest, cleanest, and most regular contacts are between granitic rocks and metasedimentary or felsic metavolcanic rocks. Mixed zones involving these rocks are intrusive breccias that consist of angular fragments of metamorphic rock enclosed in granitic rock. The angular outlines of the fragments and the straightness of their sides indicate that little or no chemical reaction has taken place between the fragments and the invading granitic rock. Contacts that are concordant with the layering or bedding of the metamorphic rocks generally are more regular than contacts that are discordant. The generally concordant contacts along the sides of the northern two-thirds of the Pine Creek pendant are especially regular (fig. 49). Intrusive breccias generally coincide with zones in the metamorphic rocks where prebatholithic structures in the metamorphic rocks have been strongly disturbed and the rocks fragmented, generally by the invading granitic magma. For example, at the south end of the Pine Creek pendant the major fold axes have been bent sharply and the rocks fractured and intricately penetrated by dikes and apophyses of the Tungsten Hills quartz monzonite. Locally, in places where quartz-rich metasedimentary rocks predominate, the pendant rocks have been expanded into clusters of slightly displaced angular blocks that are separated from one another by thin branching dikes of quartz monzonite.

Most contacts between granitic rock and older mafic metavolcanic rock, diorite, or hornblende gabbro also are sharp, but many are highly migmatitic because of chemical reaction between the granitic rocks and the invaded rocks. Sharp contacts, nevertheless, are commonly cuspate because of reaction between granitic magma and mafic rock. The products of reaction are irregular, both in composition and in texture. They are hybrid rocks of convergent compositions, which are in part the products of contamination of the granitic rock with more or less assimilated mafic material, and in part the products of additions of quartz and feldspar to metavolcanic rock.

In the part of the south end of the Pine Creek pendant where meta-andesite predominates, rocks of all compo-

FIGURE 49.—Aerial view of Pine Creek pendant in north wall of Pine Creek Canyon. Contacts between the pendant rocks and the bordering Tungsten Hills quartz monzonite are generally clean and sharp. Mono Recesses mass of quartz monzonite similar to the Cathedral Peak granite is light-colored rock along skyline. This mass is parent to aplitic, pegmatitic, and alaskitic dikes, which penetrate the granitic rocks on the far side of the pendant. Few dikes penetrate the pendant itself. The dikes generally dip 20° to 45° into the parent mass. Photo by Symons Flying Service.

sitions between quartz monzonite and diorite and andesite are complexly scrambled. Most contacts between rocks of different composition or texture are obscure, but enough relations can be established to demonstrate that the rocks are hybrids and that the parent of some is quartz monzonite and of others meta-andesite. In general, later dikes of quartz monzonite have sharper contacts with the meta-andesite than earlier ones.

Northwest of Pine Lake a mass of hornblende gabbro appears to grade over a distance of about half a mile to granodiorite. Nevertheless, the gradational zone contains a sharp though highly irregular contact between gabbro and heavily contaminated granodiorite of only slightly different appearance. The compositional gradation here appears to have taken place largely within the granodiorite by means of progressively greater contamination in the direction of the hornblende gabbro. Reciprocal hybridization of the hornblende gabbro was not established.

Most contacts between granitic rocks and metamorphic rocks are steep. The nearly vertical sides of the Pine Creek pendant are exposed through a relief of more than 5,000 feet. Flat or gently dipping contacts, nevertheless, occur locally. In Table Mountain, quartz monzonite underlies the northwest end of the Bishop Creek pendant with an almost flat contact. In the Tungsten Hills, alaskite dips under the north side of the Round Valley septum at 60° or less in the Round Valley mine and at somewhat flatter angles in the Western mine.

MAFIC INCLUSIONS

Inclusions of fine-grained mafic material, here referred to as mafic inclusions, constitute the most abundant xenolithic material in the granitic rocks. These or similar inclusions have been called "basic segregations" (Knopf and Thelen, 1905, p. 239), "autoliths" (Holland, 1900; Pabst, 1928), "basic concretions" (Grubenmann, 1896), and "inclusions" (Hurlbut, 1933, p. 614). The mafic inclusions have regular shapes and megascopically sharp boundaries with the enclosing granitic rock; they cannot properly be described as schlieren, which generally are conceived as streaky masses of irregular form.

The inclusions are variable in size, shape, texture, and mineral content, but the range of variation is rather narrow. They range from less than an inch to several feet across, and from ovoid to lenticular, spindle shapes reported elsewhere in the Sierra Nevada (Balk, 1948, p. 12, pl. 2 (fig. 2), pl. 3 (fig. 1); Pabst, 1928, p. 334) were not recognized in the Bishop region. Some lenticular inclusions less than an inch in thickness are many feet in outcrop length.

The minerals in the mafic inclusions are the same as in the enclosing granite, but they are present in very different proportions. Plagioclase generally makes up 40 to 60 percent of the inclusion, biotite 5 to 20 percent, hornblende 20 to 50 percent, and quartz 5 to 15 percent. K feldspar is scarce, and apatite and magnetite are relatively abundant. The composition of the plagioclase appears to be about the same as in the enclosing granitic rock, but in some inclusions it may be a trifle more calcic.

The common texture is granoblastic or allotrimorphic-granular. Most inclusions are structureless, but some lenticular ones have a planar foliation that is caused by the planar orientation of platy and elongate minerals. Porphyroblasts of plagioclase, hornblende, and biotite are present in most inclusions, although in larger ones they are confined to the marginal parts. These porphyroblasts are of the same size and habit as in the surrounding granitic rock. Locally, especially near the margins of inclusions, aggregates of porphyroblasts exhibit a texture indistinguishable from that of the surrounding granitic rock, and make identification of the contact difficult or impossible.

Although amphibolite inclusions are in almost all of the granite rocks, they are common only in the granodiorites. The finer grained quartz monzonite, rocks similar to the Cathedral Peak granite, and Tungsten Hills quartz monzonite contain only a few inclusions, except in the immediate vicinity of mafic igneous or amphibolitic wall rock. The inclusions that have been observed in these rocks are ovoid or irregular in shape and not lenticular.

In the granodiorites, mafic inclusions are generally more abundant near the margins than in the cores of intrusive masses. They are most abundant within a mile of the contact, common between 1 and 2 miles, and uncommon beyond 2 miles. In the Inconsolable granodiorite, however, amphibolite inclusions are common throughout the mass, though more abundant near the margins.

The inclusions are also more flattened toward the margins of the granodiorite masses. At the contact, inclusions an inch or less thick and several feet in outcrop length are common (fig. 22). With distance from the contact the inclusions appear thicker and shorter in outcrop, and beyond a mile or two from the contact generally are ovoid or irregularly rounded.

Lenticular inclusions generally parallel the nearest contact of the enclosing intrusive rock, and because most of the intrusive contacts are steep, most lenticular inclusions also are steep. Recognition of elongation within the plane of flattening is difficult unless the elongation is extreme. Most extensive faces of granitic rock are along joint planes, and the trace of the inclusion on these faces is a function of the angle of intersection of joint and inclusion. In an effort to determine whether tabular inclusions are elongate in the plane of flattening, careful watch was kept during mapping for places where the inclusions are perfectly parallel to a joint set. Strict parallelism is rare, and the effort was not successful. It is unlikely, however, that the inclusion are very greatly elongate in one direction.

The origin of mafic inclusions has never been completely explained, but three alternative hypotheses merit consideration: (1) the inclusions are clots of early formed minerals; (2) the inclusions are refractory material that was not melted when the magma was formed; (3) the inclusions are fragments of wall rocks of appropriate composition. The first and second alternatives are difficult to evaluate. How early crystallized minerals would collect into clots of roughly the same size is not obvious. The second alternative is attractive because it explains what happens to material that is left over after selective fusion of part of the earth's crust, and because it provides an explanation for the dissemination of mafic inclusions of the same size range throughout an intrusive mass, especially where wall rock of appropriate composition to form the inclusions is not exposed. Unfortunately, it is an alternative that cannot be tested in the Bishop area. Deeper level migmatic terranes where magmas are presumed to have formed are suitable for testing this hypothesis.

The third alternative, that the mafic inclusions are truly inclusions of wall and roof rock, has been investigated, and it can be shown that at least some mafic inclusions are of this origin. Any rock of appropriate chemical composition can be converted into amphibolite like that in the mafic inclusions by recrystallization. The formation of mineral assemblages such as are in mafic inclusions is assured because the metamorphic grade is in the hornblende hornfels facies (Fyfe, Turner, and Verhoogen, 1958). Mafic volcanic rocks can be made over by recrystallization, and coarser grained mafic plutonic rocks can be converted by reduction of grain size, possibly followed or accompanied by recrystallization (Joplin, 1935b). Probably some exchange of material is always involved. Hurlbut (1933) has presented a strong case for the making over of gabbro fragments into the mafic inclusions in the batholith of southern California. The conversion involved both recrystallization and minor exchange of material with the enclosing magma. The most common parent rock for the formation of inclusions of known origin in the Bishop area was mafic volcanic rock of basaltic and andesitic composition, and in a few places hornblende gabbro also appears to have served as a parent.

Other rocks such as marble and calc-hornfels require metasomatic interchange of larger amounts of material between wall rock or fragments of wall rock and magma for the formation of amphibolite inclusions. The conversion of plagioclase-diopside hornfels to amphibolite requires the addition of small amounts of H_2O, Fe, Mg, and possibly Al, and the subtraction of Ca. Calc-hornfels that has been partly converted to amphibolite was observed at several places, for example, at the Lakeview mine at the head of Gable Creek, but it was not observed in any great volume. On the other hand, no examples were found on a hand-specimen scale of residual cores of plagioclase-diopside in amphibolite; such cores as were found are composed of quartz and diopside and have rims of quartz and hornblende. Apparently diopsidic pyroxene is more easily converted to hornblende if plagioclase is present than if quartz is present, presumably because the plagioclase supplies constituents necessary for the formation of hornblende.

Clean marble also was observed to have been converted to amphibolite by the metasomatic exchange of much larger amounts of material than is required to convert plagioclase-diopside hornfels to amphibolite. At the Brown tungsten prospect, southwest of Bishop and at the Scheelore mine in the Mount Morrison quadrangle to the north, an amphibolite selvage has been formed along contacts between marble and granitic rock. At the Scheelore mine the amphibolite in the selvage is identical with abundant inclusions in the adjacent granodiorite.

Several stages in the breaking up of amphibolite wall rock into inclusions can be seen. The amphibolite is intricately penetrated with anastomosing dikes of granitic rock, which separate or nearly separate individual masses of amphibolite. These dikes appear to be chiefly granitization effects and to have formed by reaction of magmatic substances with the amphibolite in accordance with principles laid down by Bowen (1928, p. 197–198). Fractures presumably served as avenues for the movement of the magmatic substances. In a few places along the margins of the amphibolite, individual rounded remnants of amphibolite appear to have been caught in the act of drifting away from the main mass. This relation suggests that the granitic rock formed by reaction with the amphibolite was not completely rigid, possibly because it was close to its melting point.

The presence of inclusions in granodiorite and their scarcity in quartz monzonite and granite can be explained in two ways. The first is that the mafic inclusions were all picked up during intrusion, and that granodiorite, having been generally emplaced earlier than quartz monzonite and granite, had greater opportunity to come in contact with wall rocks of suitable composition to form amphibolite. The alternative explanation, which seems to fit the observed relations better, is that mafic inclusions were semistable in magma of granodioritic composition, but were unstable in more salic magmas. Granitic or quartz monzonitic magma could be expected to react with mafic inclusions in accordance with Bowen's reaction principle more readily than granodioritic magma, which is closer to amphibolite in composition. It may be significant that mafic inclusions are present in the hornblende-bearing rocks and absent in the hornblende-free rocks. The Wheeler Crest quartz monzonite is the only mass of quartz monzonite that contains abundant mafic inclusions and also the only one that contains appreciable hornblende. With respect to the batholith of southern California, Larsen (1948, p. 162) states that the inclusions were in almost perfect equilibrium with the granitic magma and probably would persist in stagnant magma for a long time without mixing to give a homogeneous rock. The magma to which he chiefly refers is tonalitic.

Progressive decrease in the abundance of mafic inclusions away from the margins of many intrusives can be explained by a change of composition of the residual magma and a progressively longer period during which inclusions can be digested. Generally there is some evidence of compositional zoning in intrusives in which mafic inclusions become less abundant away from the margins. A likely explanation for the residual magma's becoming more felsic when it is digesting mafic inclusions is that the effect of differentiation outweighs that of contamination.

Reesor (1958, p. 47–48) has postulated that the inclusion-bearing marginal granodioritic shell of the strongly zoned White Creek batholith in British Columbia is simply contaminated quartz monzonite. Two objections can be made to Reesor's hypothesis as applied to Sierra Nevada plutons. First, demonstrable contamination of the Sierran granitic rocks is spacially related to specific rock types, notably mafic ones, and the contamination of a pluton is not concentric, most particularly where several different kinds of rocks are intruded by the pluton. Second, the hypothesis implies that granodiorite that contains mafic inclusions whether in zoned plutons or not is contaminated rock that originally was more felsic. The compositional range of plagioclase in granodiorite, however, is systematically more calcic than in quartz monzonite, and except in obviously contaminated zones contains no relict picked-up crystals of calcic plagioclase. These relations suggest strongly that granodiorite crystallized at higher temperatures than quartz monzonite and was not derived from it by contamination.

The progressively flatter shapes of inclusions toward the margins of many intrusives doubtless represents physical flattening of the inclusions, and must have been accomplished by plastic flow. During flattening the inclusions probably were not much stiffer than the enclosing granitic magma. The inclusions doubtless were softened by incipient or actual melting of the lowest melting constituents, and the magma was stiffened by crystallization and increased viscosity of the melt as a result of lowering temperature.

The nearly circular shapes of inclusions in their planes of flattening can be explained if it is assumed that during emplacement the plutons grew by expansion, much as a balloon grows as it is blown up. Note that not all balloons are spherical. Mackin (1947, p. 26–32) has made the very significant point that the direction of flow in a radially spreading horizontal tabular intrusive body is toward the contact and that the direction of elongation of an equidimensional clot would be normal to that direction. Thus the foliation in granitic rocks defined by lenticular mafic inclusions does not show the flow plane. On the contrary, it is normal to the direction of flow and shows a plane of stretching. If the surface of an intrusion were hemispherical, the inclusions would be stretched equally in all directions perpendicular to the direction of flow. However, any departure from the hemispherical form should be reflected in elongation of the inclusions in the plane of flattening. If magma is pictured as flowing toward the margins of a hemisphere as it expands, the area of the external surface and parallel phantom surfaces within the magma will be expanded proportionally to the squares of the radii of the hemispheres. The inclusions nearest the margins will be flattened most, and those farther away will be flattened progressively less.

In a cylindrical pluton the circular cross section of the surface also will stretch proportionally to the square of the radius. In the axial direction, however, stretching will be equal to the elongation. Thus whether an inclusion in the margins of a cylindrical pluton will be stretched more in one direction than the other depends on whether the extension in the axial direction is greater or less than the increase in the square of the radius.

EMPLACEMENT OF THE BATHOLITH

In much of the foregoing discussions, crystallization of the granitic rocks from magma that was intruded from greater depths has been tacitly assumed. The following lines of evidence taken together indicate that the major part of the granitic rocks crystallized from a melt.

1. Contacts of individual plutons with one another and with older rocks commonly are sharp, clean, and regular.
2. Finer grained rock is present in the marginal parts and apophyses of some plutons.
3. In areas of diverse wall rocks, most plutons are either homogeneous or are compositionally zoned in patterns that bear little or no relation to the wall rocks.
4. The geometry of some dislocations of the wall rocks suggests strongly that the dislocations were caused by the forcible emplacement of magma.
5. The internal foliation in the margins of plutons parallels external contacts and results from flow.
6. Mafic inclusions are oriented parallel to intrusive contacts rather than to features in the wall rocks, and in many plutons are progressively less flattened and less abundant inward—all facts that are compatible with a magmatic origin, but not with origin by granitization.
7. The walls of aschistic dikes marginal to some plutons are dilated.
8. Granitization and assimilation effects are confined to amphibolite and other mafic wall rocks that consist chiefly of minerals earlier in Bowen's reaction series than those crystallized in the granitic rocks. The effects are in accord with theoretical expectations of reactions between granitic magma and wall rocks.
9. The metamorphic grade of the wall rocks and of inclusions is that of Turner's hornblende hornfels facies and almandine amphibolite facies (in Fyfe, Turner, and Verhoogen, 1958, p. 199–239), which form at temperatures believed to exist in the wall rocks of nonsuperheated granitic magmas.
10. Variations in the compositions of the granitic rocks are in accord with variations predicted from experimental studies in melts.

Nevertheless, even though the evidence indicates overwhelmingly that the larger part of the granitic rocks were molten, the further problem of the relative importance of different processes of emplacement remains to be considered. Before this can be done, evidences of different processes that were involved in emplacement must be summarized. In a broad way, these evidences can be divided into mechanical processes and thermochemical processes.

EVIDENCE OF MECHANICAL EMPLACEMENT

The best evidence of mechanical emplacement of the granitic rocks is deformation in wall and roof rocks that can be attributed to intrusion of magma. Several other lines of evidence are also cited, but this line is the most convincing one and the only one that permits quantitative measurement of deformation. The following topics relating to mechanical emplacement will be discussed: (1) bends and dislocations in the metamorphic remnants resulting from the intrusion of granitic magma, (2) deformation caused by intrusion of swarms of marginal dikes, (3) protoclastic borders, intrusive breccia, and related marginal features, and (4) the significance of flattened mafic inclusions.

BENDS AND DISLOCATIONS IN METAMORPHIC REMNANTS RESULTING FROM THE INTRUSION OF GRANITIC MAGMA

In the study of deformation in the metamorphic remnants, the deformation caused by intrusion must be distinguished from the earlier regional deformation. The regional deformation resulted for the most part in folding and faulting along north- to northwest-trending lines. Although the axes of regional folds undulate they average about horizontal. Deformation caused by intrusion, on the other hand, only coincidentally follows north- to northwest-trending lines, and most folds caused by intrusion have steep axes. The pattern of deformation is directly related to the shape of the intrusion.

Plate 8 is a map showing the distribution of metamorphic remnants in the area mapped in connection with this report. The adjacent intrusive masses also are shown. On this map the metamorphic rocks are subdivided into a metasedimentary series, chiefly of Paleozoic age, and a metavolcanic series, chiefly of Mesozoic age. Principal fold axes and the traces of bedding planes are shown in the larger remnants.

Deformation caused by the intrusion of granitic masses is most clearly demonstrable in the Pine Creek and Bishop Creek pendants, and the resultant structures in these pendants are described in detail in an earlier section of this report. The south end of the Pine Creek pendant, including its major synclinal axis, was bent eastward into an S-shaped structure by the intrusion of Tungsten Hills quartz monzonite. In the Bishop Creek pendant the two lobes at the north end were spread apart by the intrusion of the Tungsten Hills quartz monzonite. Faults in the northeast lobe may also have been caused by the intrusion. The strata in the west side of the northwest lobe, along the west side of Table Mountain, probably have been bowed upward by Tungsten Hills quartz monzonite. In the south part of the pendant the beds have been bent westward around a protrusion of quartz monzonite similar to the Cathedral Peak granite, which appears to have penetrated from the east. The magnitude of separations represented in these structures is variable; the largest separations—the S-fold in the south end of the Pine Creek pendant and the spreading of the north end of the Bishop Creek pendant—are about 3 miles.

Less clearly demonstrable dislocations caused by forcible intrusion of the granitic masses can be inferred from the relations between two or more separated metamorphic masses. The magnitude of displacement inferred from these relations is somewhat greater than is shown by structures within single masses. One such structure is continuation of the S-fold in the south end of the Pine Creek pendant. The contact in the Mount Humphreys ring between metavolcanic rocks and metasedimentary rocks suggests that these rocks represent the opposite limb of the fold at the south end of the Pine Creek pendant. In this interpretation, the rocks in the span between Mount Tom and Mount Humphreys would have been pushed eastward and perhaps broken through by a lobe of Tungsten Hills quartz monzonite.

The largest dislocation inferred from the arrangement of metamorphic rocks is just north of the mapped area between the south end of the Mount Morrison pendant and the north end of the Wheeler Crest septum. Strata in the Pine Creek pendant strike roughly, though not precisely, in the direction of the Pennsylvanian and Permian(?) formations in the Mount Morrison pendant with which they are correlated. The strata in the Pine Creek pendant continue to the north in the discontinuous Wheeler Crest septum, and at the north end of the septum are offset 8 miles to the east of the probable correlative strata in the Mount Morrison pendant. Round Valley Peak granodiorite occupies the region between the north end of the septum and the south end of the Mount Morrison pendant (pl. 8). The overall pattern suggests that the correlated strata in the Pine Creek and Mount Morrison pendants were once connected, and that the strata in the Wheeler Crest septum have been bent eatward and broken off from the south end of the Mount Morrison pendant. The present distribution of strata is difficult to explain on the basis of regional deformation but could have been caused by the forcible emplacement of either the Round Valley Peak granodiorite or of the Mono Recesses mass of quartz monzonite similar to the Cathedral Peak granite. Inasmuch as the Wheeler Crest quartz monzonite was emplaced along the east side of the septum before intrusion of the Round Valley Peak granodiorite, it must also have been involved in the deformation. This may explain an abundance of cataclastic structures in the Wheeler Crest quartz monzonite.

Most of the cited examples of mechanical dislocation involve lateral separation, although the one on the west side of the northwest lobe of the Bishop Creek pendant involved upward bulging. Nevertheless, if all the metamorphic remnants were pushed together, large amounts of rock would still be missing; in other words, only part of the space for the granitic rock was made by pushing aside the wall rocks. Undoubtedly some of the missing rock was pushed upward, but some was also incorporated in the magma.

DEFORMATION CAUSED BY INTRUSION OF SWARMS OF MARGINAL DIKES

Several intrusives are bordered locally by swarms of aschistic felsic dikes. The most notable swarm is exposed in the walls of Pine Creek Canyon and is satellitic to quartz monzonite similar to the Cathedral Peak granite. These dikes dip into their parent masses at angles that commonly range from 20° to 45°. A nearly perfect match of the opposite dike walls indicates that they are simply spread apart. Inasmuch as the dikes thicken and are more abundant in the direction of the parent mass, their emplacement must have caused considerable deformation of the wall rocks. In a highly diagrammatic section through a marginal dike swarm (fig. 50), all the dikes dip 25° into the parent mass, and the external contact of the parent mass is vertical. Figure 50 shows the traces of hypothetical planes which, before emplacement of the underlying dikes, were parallel to medial planes through the dikes. These traces show the amount of deformation of the wall rocks that must have been caused solely by emplacement of the underlying dikes; that is, the amount of upward displacement of these traces at any place equals the aggregate thickness of the dikes below the planes of reference measured perpendicular to the medial planes of the dikes. The lifting is accumulative through only limited distances, and the amount of lifting at any place equals the thickness of underlying dike rock measured perpendicular to the dip of the dikes; it does not equal the aggregate thickness of all the underlying dikes.

The dikes as represented in the diagram (fig. 50) are parallel and unfolded and are emplaced in order from bottom to top. In nature the dikes are generally parallel, though intersecting ones are common, and along Pine Creek the dikes most distant from the parent intrusive dip more steeply than dikes closer to the parent intrusive. These relations suggest that the dikes may be convex upward and that the flatter dips closer to the parent intrusive may have been caused by upward drag by the parent intrusive or by the later emplacement of dikes below. A correct statement might be that the dikes are emplaced in the general order of bottom to top, but with many exceptions.

The fractures which the dikes occupy appear to be feather joints caused by upward movement of the magma mass parent to the dikes. Cloos (1932) has experimentally determined that the angle between the direction of movement and tension joints in soft clay

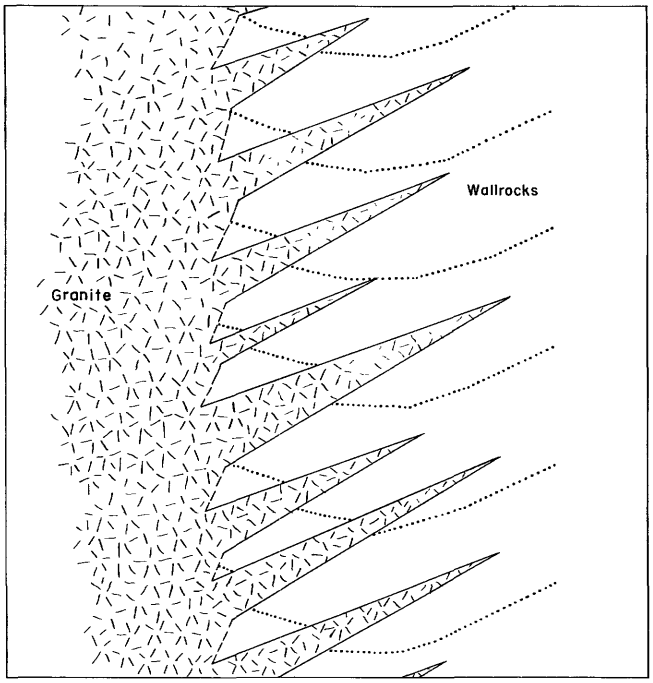


FIGURE 50.—Diagrammatic section through an intrusive and its marginal dikes to show the amount of plastic deformation in the wall rocks caused solely by the emplacement of the dikes. Dotted lines show traces of planes of reference originally parallel to the medial planes through dikes before the underlying dikes were emplaced. Inasmuch as dikes are parallel and dip into the parent intrusive, it is assumed they were emplaced in order from bottom to top and that space for the dikes was made by lifting the overlying rock in a direction normal to the dike.

is about 45°; he made field observations of about 30°. The angle between the dikes and principal contact of 45° to 70° is somewhat anomalous in the light of these data, but the higher angles could have been caused by drag from the parent intrusive or by late emplacement of dikes at depth.

PROTOCLASTIC BORDERS, INTRUSIVE BRECCIA, AND RELATED MARGINAL EFFECTS

No protoclastic borders have been recognized within the Bishop area, but Sherlock and Hamilton (1958, p. 1261) and Rinehart and Ross (1964, p. 48–49) report a protoclastic zone farther west in the north margin of the Mono Recesses mass of quartz monzonite similar to the Cathedral Peak granite. According to Sherlock and Hamilton (1958, p. 1261), the deformation occurred after the emplacement of marginal dikes and involved the marginal parts of both the pendant and the quartz monzonite.

Within the Bishop area two intrusives, the Wheeler Crest quartz monzonite and the Red Mountain Creek mass of quartz monzonite similar to the Cathedral Peak granite, exhibit shearing, granulation, and recrystallization which may have been caused by the forcible emplacement of neighboring plutons. However, the Wheeler Crest quartz monzonite is the oldest granitic rock in the northern part of the Bishop area and the Red Mountain Creek mass of quartz monzonite may be one of the oldest in the southern part; the structures in the two bodies of quartz monzonite could have resulted from regional deformation. Shearing and granulation in the south end of the Wheeler Crest quartz monzonite increase in intensity westward toward a contact with Tungsten Hills quartz monzonite—a feature that suggests that the deformation resulted from intrusion of the Tungsten Hills quartz monzonite. Shearing and granulation elsewhere in the Wheeler Crest quartz monzonite may have been caused by intrusion of the Round Valley Peak granodiorite and quartz monzonite similar to the Cathedral Peak granite west of the granodiorite. Likewise, cataclasis within the Red Mountain Creek mass of quartz monzonite similar to the Cathedral Peak granite may have been caused by the intrusion of a pluton of finer grained quartz monzonite that borders it on the south and east. Schist peripheral to the quartz monzonite on the north and east sides was formed from pelitic rocks during the emplacement of the intrusive rocks by shearing, granulation, and recrystallization. Although the schistosity could have been caused entirely by the emplacement of the quartz monzonite similar to the Cathedral Peak granite, it may have been caused in part by emplacement of the finer grained quartz monzonite; this emplacement could also have caused the shearing within the quartz monzonite similar to the Cathedral Peak granite. The origin of the schist was discussed on page 31.

Contact breccias were described briefly in descriptions of the granitic rocks. Most breccia zones are in places where the wall rocks also have been bent by the intrusive magma, and the shattering of the wall rock was caused by the same forces that caused the bending. Some wall rocks may have been shattered during preintrusive deformation, but generally the pattern of breaking indicates that the force causing the shattering and bending was supplied by adjacent intrusives. An excellent example of intrusive breccia occurs at the south end of the Pine Creek pendant, along Horton Creek. Shattering of the marginal part of the Inconsolable granodiorite by intrusion of quartz monzonite similar to the Cathedral Park granite is described on p. 6.

Most of the larger remnants of metamorphic rock surrounded by granitic rock have been so modified in outline that adjacent blocks of similar lithology cannot now be related to one another by the shapes of their walls. Nevertheless, lithology and shape of a few moderately large blocks indicate they were split off from adjacent masses. One such block is at the north end of the Pine Creek pendant, on the west side, and contains the principal workings of the Adamson tungsten mine. Another block is on the southeast side of the Pine Creek pendant and contains the Lambert tungsten mine. Two masses of diorite and hornblende gabbro in the Deep Canyon area of the Tungsten Hills are separated by a tongue of quartz monzonite that apparently split them apart.

FLATTENED MAFIC INCLUSIONS AS INDICATORS OF FORCIBLE EMPLACEMENT

The conclusion was reached on page 114 that mafic inclusions of random shape are flattened as a result of stretching of the outer, more viscous shell of a cooling pluton through increase in size. Growth of a pluton requires space, and unless large amounts of wall and roof rock were incorporated in the magma the walls must have been crowded upward and outward. The mafic inclusions themselves may represent wall-rock material, and some additional material probably has been digested by magma. Nevertheless, most intrusions appear to have incorporated far too little of the exposed wall rock to provide the needed space. The conclusion seems inescapable that during emplacement most plutons grew primarily by pushing their walls upward and outward. Deformation of the wall rocks is evident in many places, but not in others, especially where the walls are granitic. One possible explanation for the apparent near absence of deformation in granitic rocks intruded by later plutons is that the earlier rocks were still not entirely crystallized and so show no evidence of deformation; another is that the evidence of deformation has simply been overlooked and that the problem needs further investigation.

EVIDENCE OF THERMOCHEMICAL EMPLACEMENT

Thermochemical effects include all the effects that indicate chemical reaction between granitic rock or its magma and wall or roof rocks. These are the effects that have been variously attributed to granitization or to assimilation. Tendencies to broaden the term "granitization" to include all reactions between magma and solid rock give the term a double or uncertain meaning and lessen its value for precise description. In this report "assimilation" is used to describe the incorporation of solid rock in magma by partial reaction, solution, or melting, and "granitization" is reserved for the conversion in the solid state or nongranitic country rock to granitic rock. Undoubtedly these processes have an area of overlap, and discrimination between them commonly is difficult or impossible; nevertheless, some merit is attached to at least theoretical distinction.

Thermochemical effects in the Bishop district include two general kinds of features—mixed zones of granitic rock and wall rock in which sharp contacts predominate but in which wall rock dilation has been negligible, and hybridized granitic rocks that have been contaminated by wall or roof rock. Almost all of the conspicuous thermochemical effects involve mafic, generally fine-grained igneous rock; locally calcareous metasedimentary rocks have been converted to amphibolite which is chemically and mineralogically similar to the mafic igneous rocks and exhibits similar thermochemical effects. Extensive areas of sedimentary or volcanic country rock characterized by porphyroblasts or the local formation of igneous texture, such as are present in many Precambrian terranes, are absent here.

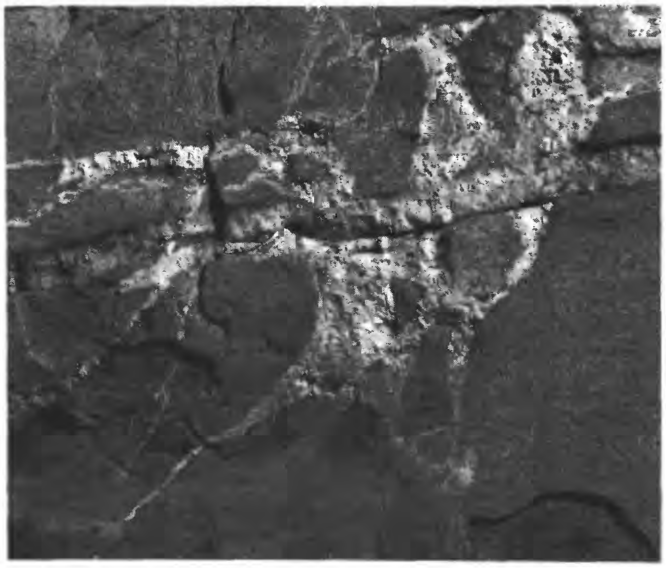
GRANITIZATION OF MAFIC ROCK

Examples of mixed zones that result from the granitization of mafic igneous rocks are shown in figures 51, 52, and 53. The geometric relations of many irregularly shaped masses of granitic rock that penetrate mafic igneous rock indicate that the granitic rock has taken the place of mafic igneous rock and has not been emplaced by simple dilation of the walls. Small inclusions of mafic igneous rock in granitic rock could be interpreted equally well as either undisplaced residuals of chemical attack by granitic rock or its magma, or as dislocated inclusions whose boundaries have been irregularly corroded chemically.

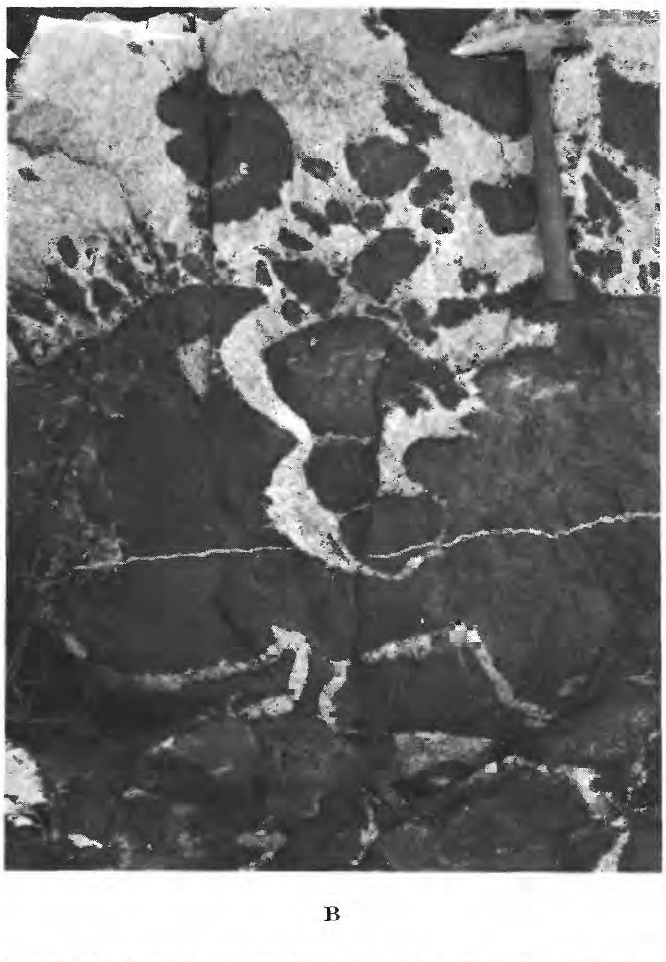
The geometry of the walls of the early aplite shown in figure 51*A* is such that emplacement of the aplite forcibly by spreading of the walls would have been very difficult unless the mafic rock was quite soft and plastic. Darker colored, somewhat rounded areas of mafic rock adjacent to or enveloped in aplitic material are areas in which the grain size has been reduced and in which new minerals have been formed. This process is accomplished by recrystallization in conjunction with exchange of substance. Common changes are the disappearance of hornblende and plagioclase, which are the most abundant minerals in most mafic rocks, and the appearance of a fine-grained granoblastic or schistose intergrowth of biotite, quartz, K feldspar, and, near the granite, epidote. These new minerals require addition of Si and K and subtraction of Ca, Mg, and Fe. Reduction of grain size is a common feature along contacts between earlier mafic rock and later granitic rock.

In figure 51*B*, contacts between mafic rock and granitic rock are embayed and cuspate, typical features of reaction contacts. The shapes of the dikelike masses of granitic rocks indicate they replaced the mafic rocks. Careful examination of many outcrops like that shown in figure 51*B* indicates no movement in the third dimension.

Mafic dikes younger than the enclosing granitic rock and mafic inclusions in the granitic rock have been at-



A



B

FIGURE 51.—Typical granitization effects in mafic rock. *A*, Replacement of mafic rock by aplitic material; later aplite dike, emplaced by dilation, cuts earlier replacement aplite; note areas of darker, finer grained mafic rock (L) present locally adjacent to replacement aplite. *B*, Replacement of mafic rock by medium-grained felsic granite rock. Note irregularly embayed contact.

tacked and selectively replaced by aplitic material. Commonly the first signs of such replacement of a mafic body are thin stringers of aplitic material in the margins. The contact of the aplitic stringers with the enclosing granitic rock is straight and sharp, and coincides with the original contact of the mafic rock, as is shown in figure 52. The contact of the aplitic material with the mafic rock, in contrast, is irregularly penetrating or conspicuously cuspate. Stringers may extend entirely across a dike or inclusion, and in places the mafic material is cut by many such stringers which may irregularly pinch and swell. Mafic dikes were observed that could be traced through segments containing progressively thicker marginal stringers of aplite into a dike entirely of aplite.

Reaction between an aplite dike and an older mafic dike is illustrated in figure 53. In the lower right-hand corner of the photograph, where the aplite dike cuts Tinemaha granodiorite, the positions and shapes of the dike walls indicate that they were spread to accommodate the dike. At the upper end where the thin dike intersects an older mafic dike, reaction has taken place within the mafic dike to produce the mixed and aplitic rock. The granodiorite was not involved in this reaction, and the conspicuous contact across the lower part of the photograph between granodiorite and aplitic rock marks the original wall of the mafic dike.

Progressive hybridization of mafic rocks by quartz monzonite similar to the Cathedral Peak granite along Big Pine Creek was discussed in connection with the description of quartz diorite and related granodiorite. The early stages of hybridization involved progressive granitization of the mafic rock through recrystallization, accompanied by interchange of substances. The later stages involved disintegration of the granitized rock at the margins and incorporation of the fragments in the granitic magma. At some distance from the mafic rock the granitic rock is even textured, but has a higher content of dark minerals than uncontaminated quartz monzonite with which it is locally in sharp contact.

The mechanism by which ganitic rock replaces mafic igneous rock can be explained most satisfactorily in terms of Bowen's reaction series. The fine-grained mafic rocks are composed of minerals that would crystallize from a silicate melt at higher temperatures than the minerals of the quartz-bearing granitic rocks. When a mafic rock is brought into contact with granitic magma, the mafic rock is so modified as to bring its mineral assemblage into equilibrium with the magma. The granitic magma does not dissolve the mafic rock because it does not contain sufficient heat; it works it over by reaction. Bowen states (1928, p. 198), "These

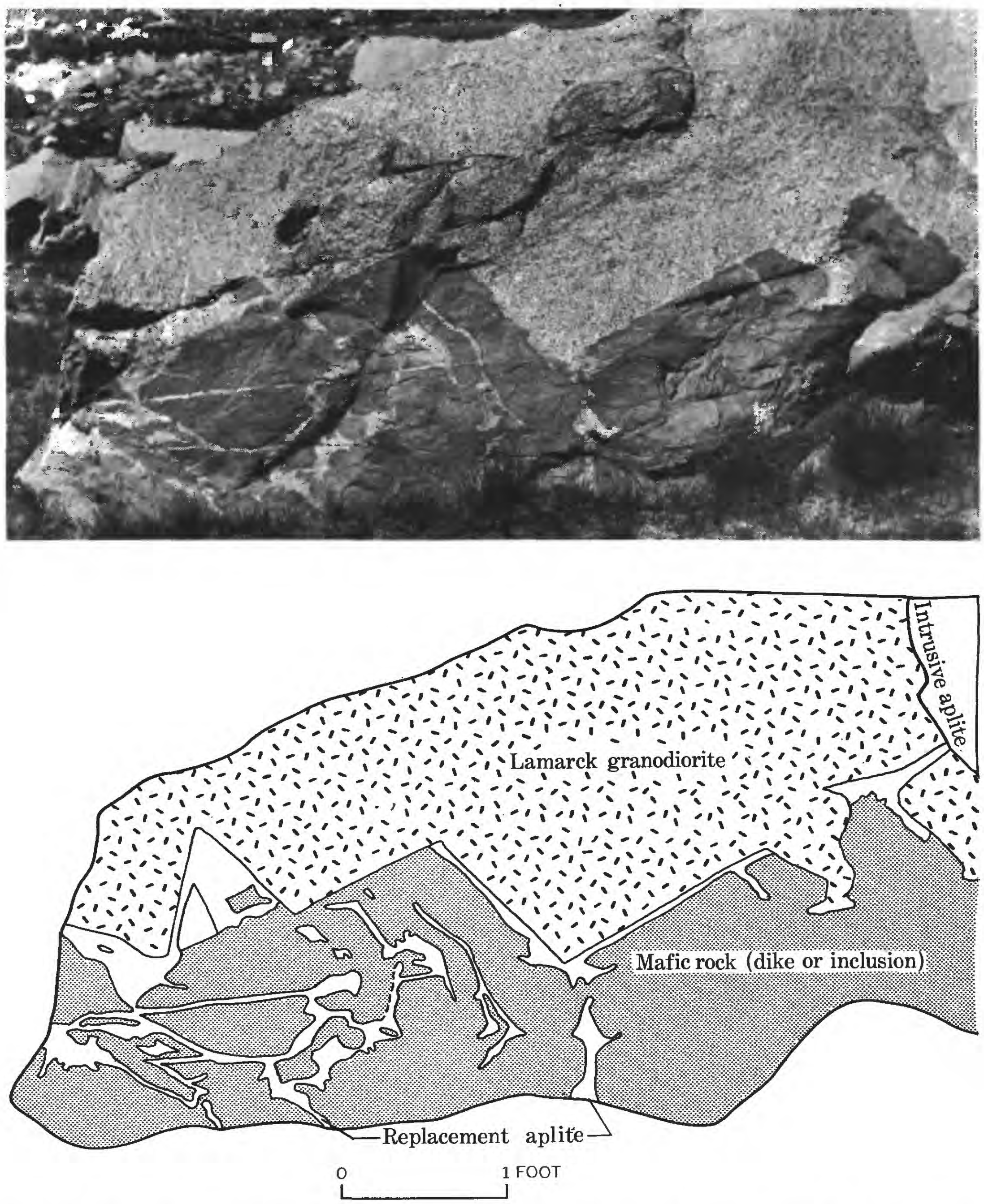


FIGURE 52.—Selective replacement of mafic rock (dike or inclusion) over Lamarck granodiorite by aplitic material. Late intrusive aplite dike is present in upper right-hand side of photograph.

remarks are tantamount to the statement that saturated granitic magma cannot dissolve inclusions of more basic rocks. The magma will, however, react with the inclusions and effect changes in them which give them a mineral constitution similar to that of the granite. These changes will often be accompanied by disintegration of the inclusions and the strewing about of the products which may be indistinguishable from the ordinary constituents of the granite."

Several problems can be envisaged in connection with the actual operation of the process. One is the mechanism by which material (chiefly Si, Na, and K) is introduced into the mafic rock and excess material (Ca, Mg, Fe, and Al) is removed. Another problem is the nature of the contact between granitic rock that crystallized from a magma and similar rock that formed by the replacement of mafic rock. The irregular penetration of granitic rock into mafic rock and the dikelike

FIGURE 53.—Selective reaction of aplite dike with mafic dike in Tinemaha granodiorite. Aplite was intruded into Tinemaha granodiorite, with which it shows no evidence of reaction. Clean straight walls and geometry of walls where aplite enters mafic dike indicate emplacement by dilation. Reaction with mafic dike is apparent. Probably both granitization and assimilation effects are represented. Note darker appearing, finer grained rounded masses of mafic rock adjacent to aplitic material.

form of some thin granitic layers suggest that the granitization was guided by fractures. That true magma entered along these fractures seems dubious. Nockolds (1933) has emphasized the role of volatiles which he suggests furnish a medium of low viscosity through which material can travel by diffusion from a magma both into and from a xenolith. This medium could be a hydrous melt of ternary minimum composition, for although a silicic magma cannot totally melt a ferric one, it can supply sufficient heat to melt the least refractory fraction. Garrels, Dreyer, and Howland (1949) have shown experimentally that solute diffusion through intergranular spaces in rocks can be an effective geologic process. Conceivably the contact between granitic rock that crystallized from a magma and similar rock that formed from granitization could remain in its initial position, but both Nockolds (1933) and Turner and Verhoogen (1951, p. 121–122) suggest that marginally the granitized rocks would tend to disintegrate, chiefly by mechanical means, and the minerals and rock fragments would be scattered through adjacent magma. If the granitized rock disintegrates marginally, the magma will encroach on the wall rock and thus reduce the distance through which diffusion must operate.

ASSIMILATION OF MAFIC ROCK

Complementary to granitization of mafic igneous and metamorphic rock is assimilation of the same rock by granitic magma. The product of assimilation is a dark-colored hybrid rock having a splotchy appearance that is caused by uneven distribution of the mafic minerals and by variations in texture. Most of the hybrid rock is granodiorite in composition, but some is quartz diorite. Generally the hybrid rock is darker than the rock in large intrusive masses of granodiorite composition. On a small scale, some dikes intrusive into mafic host rock are progressively more contaminated with greater distance into the host rock.

Contaminated granitic rock is extensive in the south end of the Pine Creek pendant, on the north side of Pine Creek northwest of Pine Lake, and in the ridge between the Middle and South Forks of Shannon Canyon. An illustrative area (described on p. 56) is northwest of Pine Lake where hornblende gabbro and diorite grade laterally into granodiorite through a distance of about half a mile. Nevertheless, the intrusive contact between the two rocks is sharp but highly irregular, although the rocks are very similar in appearance at the contact. The earlier hornblende gabbro does not appear to have been much affected by the granodiorite, but the gabbro obviously contaminated the intrusive granodiorite magma.

Assimilation of the mafic material and consequent contamination of the granitic magma are probably accomplished by diffusion through volatiles or a pervasive hydrous melt of ternary minimum composition in conjunction with the physical incorporation in the magma of fragments of the wall rock. Accompanying the introduction of material into wall rock in granitization is the reciprocal process of removal of material (principally Ca, Mg, Fe, and probably some Al). No evidence was found to indicate that the removed material was driven ahead of the granitization; presumably it moved in the reverse direction, back into the magma. A difficult problem in this connection is the absence in so many places of any evidence of contamination of the granitic rock adjacent to granitized mafic rocks. Whether the material that was expelled from the wall rock traveled only far enough after reaching the magma to produce a zone of hybrid rock or whether it was dispersed so widely as to leave no recognizable trace probably was determined by the rate of dispersal after reaching the magma. Very little is known about the rate of diffusion in siliceous magma except that the rate decreases as the magma cools, becomes more viscous, and crystallizes. Certainly, movements within the magma, which would aid in the dissemination of

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material and homogenization of the magma, would decrease with cooling.

Marginal disintegration of the granitized or partially granitized mafic wall rock and strewing of rock or crystals through the adjacent magma may also contribute to the formation of the hybrid rock. The texture of some areas in the hybrid rock is sufficiently similar to that of the wall rock to suggest a xenolith. Partially granitized xenoliths would, of course, be further granitized, and the displaced elements diffused into the surrounding magma. The splotchy texture of the hybrid rock may be due to the partly granitized fragments of wall rock distributed through contaminated granitic rock.

EVALUATION OF PROCESSES OF EMPLACEMENT OF THE BATHOLITH

The Sierra Nevada batholith at the latitude of Bishop is about 60 miles wide. It has been established that the batholith consists of a mosaic of separate intrusive masses and that magma was involved in the formation of these masses, although a minor amount of rock was formed by granitization. The chief problems to be considered here are the means by which space for the intrusives was provided and the relative effectiveness of forcible displacement of the country rock by magma, stoping, granitization, and assimilation in their emplacement. These two problems are interdependent and can conveniently be considered together.

One of the difficulties of evaluating the different processes related to the emplacement of the intrusives is that they are characteristically on different scales. Neither the eye nor the camera is equally receptive to features of all sizes. Outcrops a few inches or a few feet across can be observed in considerable detail; outcrops a few hundred feet across if viewed in their entirety can be seen in much less detail, and larger features can be viewed only in a broad way and only if conditions that affect visibility are exceptionally good. Effects associated with emplacement of the intrusive masses of about the right size for field inspection include replacement dikes and other granitization effects, zones of intrusive breccia, and dike swarms. Somewhat larger features that can be seen in the field in some places are hybridized zones in granitic rock and large foundered blocks of wall rock that are separated from an originally conjoined mass by granitic rock, but the understanding of the significance of these features commonly is aided by maps or diagrams. Large features that ordinarily escape the eye and which can be satisfactorily understood only after they have been represented on a map include large-scale bends or dislocations of the country rock that were caused by the forcible emplacement of intrusive magma. Thus the small-scale effects of granitization and brecciation can be readily observed and photographed in many places in the Sierra Nevada, whereas larger scale physical dislocations of the wall rocks by forcibly emplaced magma commonly can be recognized only after a map has been prepared.

Physical separations of as much as 3 miles, caused by the forcible emplacement of intrusives, have been demonstrated for structures within individual pendants, and a separation of about 8 miles has been inferred between the Wheeler Crest septum and the Mount Morrison pendant. Separations of these magnitudes are by far the most impressive effects associated with the emplacement of the intrusives. In some places the entire space for the granitic rocks apparently was made by pushing aside the wall rocks. All evidence indicates that dislocation by intrusion must certainly be assigned an important role in making space for intrusive masses.

The role of stoping is difficult to assess because the boundaries of most blocks of pregranite rock have been so modified that blocks of similar lithology cannot be related to one another by the shape of their walls. At only a few places can large blocks (blocks at least a quarter mile long) be shown to have been split off from adjacent blocks of similar lithology. Fragmentation and piecemeal stoping are locally demonstrable in zones of intrusive breccia, but appear to be secondary to either large-scale bending or to granitization and assimilation. Nevertheless, it seems reasonable to suppose that magma intruded with such force as to cause more than 3 miles of lateral separation of the wall rocks would have entered along lines of weakness and would have split off blocks. Once these blocks were separated they would be moved in accordance with their specific gravity and the currents and viscosity of the magma. Some of the exposed masses of pregranitic rock probably came from higher or lower levels than the present erosion level.

Visible traces of granitization and assimilation are quantitatively insignificant, though conspicuous because of their optimum size for observation. Nevertheless, the end stage of granitization and assimilation is granitic rock that is virtually indistinguishable from rock that crystallized directly from magma; hence these processes are self-effacing. Granitic magma does not appear to have reacted appreciably with most metasedimentary rocks—nor with earlier granitic rocks. Reaction is largely confined to the mafic igneous rocks and to amphibolite, and it is in these mafic rocks that granitization and assimilation are conspicuous. Time as well as composition is important to reaction. Dilation dikes having sharp walls invariably are later than granitization effects in the same area (figs. 51*A*, 52), and earlier dikes generally are more strongly hybridized and have less sharply defined walls than later dikes (fig. 54). These relations indicate that the chemical activity of granitic magma decreases with lowering temperature and that reaction is only important during the earlier stages when the magma is hottest.

Compton (1955) has studied an area in the western Sierra Nevada where plutons west of the main Sierra Nevada batholith intrude a terrane that is composed predominantly of mafic metavolcanic rock. For the Bald Rock batholith he estimates from approximate projections of the country rock units through the batholith that most of the space for the batholith was provided by forceful intrusion and outward forcing of the country rock, but that about a quarter of the required space was provided by stoping, assimilation, or granitization of the country rock. For the adjacent Swedes Flat pluton, which has a broad gradational contact zone with the metavolcanic country rocks, he expresses the opinion that most of the granitic rock is of replacement origin, but that the granitization was produced by fluids that emanated from a core of magma. His estimates, which appear to be well founded, suggest that in terranes composed of mafic igneous rock a very significant part of the space required for the emplacement of an intrusive can be provided by granitization,

FIGURE 54.—Mafic inclusion cut by dikes of slightly different ages, both of which are offshoots from the surrounding quartz monzonite similar to the Cathedral Peak granite. The earlier dikes extend from upper left to lower right. They are darker than the parent rock, apparently because of contamination. Although it seems clear that the dike walls are dilated, they are ragged and do not match perfectly across the dikes, presumably because of reaction. In contrast, later dikes, which extend from top to bottom, are uncontaminated, and the walls are smooth and match perfectly.

probably in conjunction with the complementary process of assimilation.

Physical dislocations caused by forcible intrusion of granitic magma are most conspicuous in metasedimentary terranes, where the rocks are of such a composition as to react very slowly with the magma, or in terranes composed of igneous rocks compositionally similar to the intrusive magma. Nevertheless, there is little reason to believe that intrusions were less forcible in terranes composed of mafic igneous rocks and other rocks reactive toward granitic magma; however, granitization and assimilation obscure the effects of forcible intrusion.

In summary, the most important role in the emplacement of the granitic rocks in the Bishop area must be assigned to forcible intrusion which thrust aside, and ultimately upward, the older rocks. These dislocations were accompanied locally by piecemeal stoping, and in places stoping may have provided substantial amounts of space now occupied by intrusive rocks. Thermochemical processes were most effective where the wall rocks were mafic igneous rocks or amphibolite, but even there were probably of less importance than mechanical processes. Where volcanic rocks predominate, as in certain parts of the western Sierra Nevada, thermochemical effects may be of greater importance in the emplacement of granitic rocks.

CONTACT-METASOMATIC TUNGSTEN MINERALIZATION

The contact-metasomatic, scheelite-bearing tungsten deposits contain the principal ores of the Bishop district. To the end of 1953 they had yielded about 1.3 millions short-ton units [1] of WO_3 (tungsten trioxide)—about 12 percent of the total domestic production of the United States. More than 50 tungsten deposits have been discovered in the district, including 21 that have yielded more than 100 units of WO_3 (table 18). The Pine Creek mine of the Union Carbide Nuclear Co. has supplied more than three-fourths of the district's past production of tungsten, and substantial amounts of copper and molybdenum as byproducts. To the end of 1953, it yielded 1,057,498 units of WO_3, 1,967.97 tons of copper, and 6,130,559 pounds of molybdenum.

In an earlier report (Bateman, 1956), individual tungsten deposits as well as deposits of other mineral commodities mined in the Bishop district—gold, silver, antimony, cobalt, barite, expansible rhyolite (perlite), marble, talc, clay and feldspar, and sand and gravel—

[1] The tungsten content of ore and of concentrate is commonly given in terms of units of WO_3, regardless of the actual form of the tungsten. A unit is one percent of a ton. Thus, a short-ton unit (the unit of this report and the one commonly used in the United States) is equivalent to 20 pounds of WO_3.

TABLE 18.—*Tungsten mines and prospects in the Bishop district showing the parent metamorphic and plutonic rocks*

[Closest silicic rock is shown by dashed arrow for deposits in inclusions in quartz diorite or hornblende gabbro. Question mark indicates probable parent rock]

Mines (production greater than 100 units WO_3) and prospects	Parent metamorphic rock		Associated plutonic rock			
	Marble	Calc-hornfels	Alaskite	Tungsten Hills, quartz monzonite	Quartz diorite and hornblende gabbro	Other
Pine Creek pendant:						
Pine Creek	X			X		
Adamson	X			X		
Brownstone	X			X		
Tungstar	X			X←	——X	
Hanging Valley	X			X		
Lakeview	X			X (or alaskite)		
Lambert	X			X		
Round Valley Peak	X					Wheeler Crest quartz monzonite.
2 miles N. 25° W. from Scheelite	?				X——→	Wheeler Crest quartz monzonite.
Blue Grouse	X			X (also felsic dikes)		
Moore (Sunnyboy)	X			X		
Mountain Basin	?			X		
Bishop Creek pendant:						
Schober	Unknown			X←	——X	
Merrill	do			X←	——X	
Lindner		X				Lamarck granodiorite.
Snow Queen	X			X		
Black Monster	X			X		
Munsinger		X		X		
Hilltop	X			X		
Little Egypt		?		X		
Coyote Creek	X					Granodiorite of Coyote Flat.
Lookout Mountain	X			X		
Waterfall	X	X		X		
Stevens		X		X		
North Lake		X		X		
Green Lake area	X	X	X←	——	——X	
Chocolate Peak (Bishop silver-cobalt prospect)	X	X				Lamarck or Inconsolable granodiorite.
Tungsten Hills:						
Round Valley	X		X			
Western		X	X			
Tungsten Hill (Little Shot)	X		X			
Little Sister	X			X←	——X	
Jackrabbit	Unknown			X←	——X	
Lucky Strike	X			X←	——X	
Tungsten Blue (Shamrock)	Unknown			X←	——X	
Aeroplane	?		Uncertain			
White Caps	X		X			
Tungsten Peak	X			X		
Lookout		?		X		
Hilltop	Unknown				X——→	Granodiorite of Deep Canyon.
Van Loon (Raindrop)		X		X		
Range Front south of Bishop:						
Rossi	X		X			
Chipmunk	X			X		
Yaney	X	X	X			
Marble	X			X		
Early-Morhardt	X		X			
West of Keough Hot Springs	Unknown		X			
Brown	X			X		
Buckshot	X		X←	——	——X	
Middle Fork Shannon Creek	X	X		X		
Rattlesnake	X		Uncertain			
McVan	X		X←	——	——X	
Bakock	X			X		
Taboose Creek	X					Finer grained quartz monzonite.
Totals			8	26	12	5
Adjusted totals (deposits enclosed in quartz diorite or hornblende gabbro reassigned to closest silicic intrusive)			11	33		6

were described in detail. The focus of this report is primarily on the process of tungsten mineralization as interpreted in the light of information acquired in the study of the Bishop district. To be considered are the mineral content and internal constitution of the deposits; factors that influenced the localization, size, and grade of the ore bodies and ore shoots; and regional relations that may explain the existence of the district. Among other mineral commodities, only byproducts of tungsten mining, which were deposited as part of the tungsten mineralization, will be dealt with.

MINING HISTORY

Tungsten was not discovered in the Bishop district until 1913 and not mined until 1916, but prospecting, chiefly for gold, was carried on by the earliest settlers. The Golden Wedge mine, later part of the Silver Belle mine, 4 miles north of Laws at the base of the White Mountains, was worked in the early 1860's. The Poleta mine, east of Bishop in the White Mountains, and the mines in Fish Spring Hill were worked in the 1880's and may have been worked earlier. The Cardinal mine, on Bishop Creek, by far the largest producer of gold in the Bishop district, was operated from 1911 to 1922 and from 1934 to 1940. During the depression years of the 1930's many deposits of gold were worked. Gold mining ceased at the beginning of World War II, and no gold mine in the Bishop district had been reopened by 1958.

Sand, gravel, granite, marble, and rhyolite tuff have been also mined intermittently since the time of the earliest white settlers. The Bishop tuff was used for building block in Owensville in 1863, and was locally a popular building material until the 1930's, when cement blocks with pumice aggregate proved to be cheaper and easier to use. Pumice for aggregate and for use in plaster has been mined more or less constantly since then. Barite was mined from the Gunter Canyon barite mine in 1928 and 1929, but the deposit was idle from 1930 to 1958. Talc was most recently mined at the Blue Star talc mine in 1945. The newest nonmetallic mineral commodity is expansible ryholite (perlite), which has been mined from a rhyolite hill south of Big Pine only since 1949.

Tungsten mining has been very sporadic because of tremendous fluctuations in the price of concentrate. During World War I and again in World War II the price rose to high levels, then at the end of the wars dropped abruptly to low levels. Tungsten ore was first discovered in August 1913 at the site of the Jackrabbit mine in the Tungsten Hills. Early in 1916, when the price of tungsten rose to unprecedented heights, mining was begun in the Deep Canyon area. The first mining was on the Aeroplane claims by the Standard Tungsten Co., and shortly thereafter the Little Sister, Jackrabbit, and Lucky Strike mines were brought into operation by the Tungsten Mines Co. The town site of "Tungsten City" was laid out in Deep Canyon, below the mines. In the following year, mining was begun at Nobles Camp (Round Valley mine), and exploratory work was carried out at the Chipmunk mine, the Mineral Dome Prospect (Rossi mine), the McVan claims, and the Buckshot prospect. High-grade molybdenum ore was handpicked from the outcrop at the Pine Creek mine in 1916, but the first mining for tungsten there was not until 1918, by the Pine Creek Tungsten Co.

The tungsten market collapsed at the end of World War I, and between 1920 and 1923 no production of tungsten was reported in the United States. In the following decade, however, most of the known mines in the district were in operation for brief periods at one time or another. In 1924, the Tungsten Products Co. reopened the Pine Creek mine, and it was in operation through 1925 and part of 1926, but was shut down between 1926 and 1936.

In the middle 1930's increased demand and higher prices for tungsten resulted in the reopening of many mines. A wave of prospecting followed the introduction of the ultraviolet light, and by 1941 all the known deposits in the Bishop district, except the Yaney mine, had been discovered. New discoveries were made during the period between 1936 and 1941 at the Tungstar, Brownstone, Hanging Valley, Lakeview, Schober, Lambert, Tungsten Blue, and Marble tungsten mines.

The main period of exploitation of the Pine Creek mine began in 1936 when it was purchased by the U.S. Vanadium Co. (Union Carbide Nuclear Co.). Between 1937 and 1939 the upper part of the mine, above level A, was developed for mining (pl. 9). All the ore mined prior to 1948 was from above level A; in 1948 the Zero adit, having its portal 7,000 feet south of the outcrops of the ore bodies and 1,500 feet lower than level A, was completed. By 1948 the ore reserves in the upper part of the mine were depleted, and most of the ore mined between 1949 and 1958 was from between the Zero adit and level A.

In 1937, ore was discovered at the Tungstar mine, second in total production to the Pine Creek mine, on the west side of Mount Tom at an altitude of 12,000 feet. Mining was begun in 1939 and carried on until October 1946, when the mine installations and upper tram terminal burned.

Peak production for the district was reached during World War II, when the price and sale of tungsten concentrates were fixed by Federal law; but at the end of the war, discontinuation of Government purchases of tungsten concentrates caused a price decline that once again forced curtailment or abandonment of most operations.

In 1951, following the outbreak of war in Korea, the Federal Government increased its purchases of tungsten concentrate and stimulated domestic production by offering to purchase 3 million units of WO_3 at $63.00 per unit. Tungsten mining flourished in the United States until mid-1956 when the Government stockpile was filled and purchasing was terminated. In early 1958 the price on the open market was only about $20.00 per unit, too low for profitable exploitation of any of the mines in the Bishop district. In 1958, only the Pine Creek mine was in operation and that on a limited scale.

GRADE OF ORES

The average grade of ore mined in the district is about 0.5 percent of WO_3, but a substantial amount of ore having several percent of WO_3 was mined from the Tungstar, Schober, and Yaney mines. The lowest grade ore that has been profitably exploited exclusively for tungsten, from the Shamrock and other deposits in the

Tungsten Hills, contained about 0.4 percent of WO_3. From time to time, ore containing even less tungsten but with recoverable copper and molybdenum has been profitably produced at the Pine Creek mine. Factors favorable to the profitable exploitation of lower grade ores are low altitude of the deposit, large volume of ore mined daily, reliable and inexpensive transportation, and an ore body of sufficient size and of such configuration as to favor cheap methods of mining.

In most deposits the ore shoots have fairly sharp walls and do not grade through a zone of submarginal ore into barren or nearly barren tactite. Common practice in most mines is to examine each face under ultraviolet light before blasting. In this way the ore mined is kept at a profitable grade, and low-grade or barren tactite is left in pillars or discarded as waste. In the Pine Creek mine, where mining is by means of blast hole stopes, the distribution of ore is determined by examination of cores taken from holes to be used for blasting.

The grade of the ore within most ore shoots is constant, but not in all ore shoots. In deposits containing both high-grade and low-grade ore, the ore is commonly mixed to maintain an average grade. Hand sorting under ultraviolet light has not been employed widely in the district.

OUTLOOK

Ore reserves are large, even though some formerly productive deposits are exhausted and others contain only submarginal ore. Recently discovered extensions of known ore bodies provide a basis for considerable optimism regarding the life of the district. New deposits undoubtedly will be discovered, but their discovery will require increased use of geologic and engineering skill. As in the past, most new discoveries will probably be relatively small.

SUMMARY OF GEOLOGIC RELATIONS

The tungsten deposits in the Bishop district, with one or two exceptions, are contact-metasomatic tactite deposits. Such deposits are generally conceived to have formed at high temperatures by the interaction of calcium-rich sedimentary or metamorphic rock with hot solutions that emanated from intrusive magma. The term "contact metasomatic" is used here in preference to the more commonly used term "contact metamorphic" in order to distinguish between two different though related processes: recrystallization under high temperatures without addition of substances, and recrystallization under high temperatures with addition of substances. As used here "contact (or thermal) metamorphism" is restricted to the high-temperature recrystallization of rocks without significant addition of substances from an intrusive magma. Barrell (1907) made this same distinction and pointed out that nonadditive contact metamorphism generally precedes metasomatism and takes place with rising temperatures, whereas contact metasomatism takes place later with falling temperatures, after partial recrystallization of the intrusive rock. He pointed out further that the effects of nonadditive contact metamorphism are most striking in impure limestone, whereas metasomatism takes place most readily in purer limestone.

The principal product of contact metasomatism is tactite, which locally contains scheelite ($CaWO_4$). Tactile generally is marginal to masses of calcium-rich rock adjacent to intrusive granitic rock. Some masses of tactite are bordered by a peripheral zone of bleached and silicated marble that may have been formed either contemporaneously with the tactite or earlier, at the peak of thermal metamorphism. In addition, in many deposits thin quartz veins and thicker silicified zones locally penetrate the tactite and the adjacent granitic rock. These quartz veins and silicified zones formed somewhat later than the tactite and at lower temperatures, but nevertheless as part of the same process. The quartz veins and silicified zones are accompanied in places by valuable sulfides. Generally they contain little or no scheelite, but where present it is abundant and may form rich, though small, ore bodies.

Although the contact metasomatic tungsten deposits commonly are called tactite deposits, all tactite does not contain commercial amounts of scheelite. On the contrary, many tactite masses contain no scheelite; where scheelite is present it usually is restricted to certain zones within the tactite. Tactite masses that contain scheelite-bearing zones of a grade that is commercially exploitable are called ore bodies; the exploitable scheelite-bearing zones themselves are called ore shoots.

DISTRIBUTION OF DEPOSITS

The tungsten deposits are confined to masses of calcium-rich metamorphic rock adjacent to intrusive granitic rock. Most of the deposits are in groups, but a few are isolated. The largest groups are in the marginal parts of the Pine Creek and Bishop Creek pendants and Round Valley septum, in a belt of metamorphic rocks along the Sierra Nevada front southwest of Bishop, and in groups of inclusions of metamorphic rock along Deep Canyon in the Tungsten Hills and in Shannon Canyon. Of 21 commercial deposits, 7 are in the Pine Creek pendant, 1 is in the Bishop Creek pendant, 9 are in the Tungsten Hills, 3 are along the Sierra Nevada front southwest of Bishop, and 1 is in Shannon Canyon. The mines in the Pine Creek pendant have produced by far the greatest amount of tungsten and they include the two deposits with the greatest individual yields; the mines in the Tungsten Hills are next.

THE CONTACT ZONE

The term "tactite" was introduced by F. L. Hess (1919) and defined by him as "a rock of more or less complex mineralogy formed by the contact metamorphism of limestone, dolomite, or other soluble rocks into which foreign matter from the intruding magma has been introduced by hot solutions or gases. It does not include the enclosing zone of tremolite, wollastonite, and calcite." The term "tactite" is now deeply entrenched in the United States, both in the literature and in local usage, in connection with tungsten deposits. An older synonymous term is "skarn," which was originally used to described similar assemblages of silicate minerals associated with Swedish iron ores. Garnetite is a variety of tactite in which garnet predominates.

Although no single mineral by definition is essential to tactite, most of the tactite in the Bishop district consists chiefly of various proportions of pale to moderate reddish-brown garnet of the andradite-grossularite series and grayish-green pyroxene of the diopside-hedenbergite series. Light-olive to grayish-green epidote is present in many tactite masses and in a few is a principal constituent, but it is not ubiquitous. Some of the richest tungsten ore consists chiefly of quartz and epidote, accompanied in places by pyrite or pyrrhotite and scheelite. Green to black amphibole is present locally in some tactite, but is not common or abundant. Quartz usually is present in only minor amounts in garnet-pyroxene tactite, except where the tactite has been partly silicified. Scheelite is sporadically distributed.

In addition to these more typical minerals, many tactite masses contain subordinate amounts of calcite, fluorite, idocrase, sphene, and such light-colored silicate minerals as wollastonite, diopside, grossularite, zoisite, and feldspars. However, if these minerals predominate, the rock is not, properly speaking, tactite, but rather a calc-silicate rock or hornfels. Metallic sulfides and oxides locally are disseminated in tactite, but generally were deposited with quartz after the silicate minerals were formed. In the Pine Creek and adjacent mines, molybdenite, chalcopyrite, and bornite are the most common sulfides; sphalerite, galena, and magnetite are found locally in other parts of the district.

The grain size ranges from less than a millimeter to more than a centimeter, but the average is probably closer to a millimeter than to a centimeter. Larger and more nearly euhedral grains of garnet, epidote, scheelite, or sulfides are locally associated with masses of quartz, which are present in many tactite bodies. The fabric of tactite ranges from isotropic to layered or veined. Veins are especially common in dark garnet-pyroxene tactite, and layering is more abundant in lighter colored varieties, especially those that contain light-colored silicate minerals.

Tactite bodies vary widely in size and shape—from thin layers a few inches thick to masses several hundred feet long and more than a hundred feet thick. The contact with granitic rock generally is sharp, but in places is obscured by silicification. Diorite or hornblende gabbro commonly is altered to epidote adjacent to tactite, and where the tactite also contains appreciable epidote the contact may be difficult to place within a few inches or even a few feet. The contact with marble generally is sharp or is gradational through a few feet, and may be irregular, especially where the granite cuts across bedding and tactite extends outward from the contact different distances along different beds. Along some contacts, discrete crystals of garnet or idocrase may trail off from the tactite into the adjacent marble through distances of several feet. In detail, these crystals may lie along beds or fractures.

CHARACTERISTIC MINERALS OF TACTITE

Following are brief descriptions of the more characteristic minerals of tactite. Calcite, quartz, and the light-colored silicates locally present in but not typical of tactite are not included.

Garnet.—Garnet generally is pale to moderate reddish brown, but may be pale pink in the marginal parts of tactite masses adjacent to or interfingered with marble. In contact with other crystals of garnet or of pyroxene, external crystal faces are rarely formed, although inner zones may be euhedral. Against quartz and calcite, garnet crystals generally are euhedral, commonly with dodecahedral habit. Most garnet is isotropic to weakly birefringent; birefringent crystals are twinned parallel to crystal faces and are in pyramidal sectors whose bases are crystal faces. Some garnet crystals are zoned in a narrow range of composition and color. Most determinations of index of refraction fall between 1.78 and 1.80; the highest determined index is 1.85 for moderate to dark reddish-brown garnet from the Lakeview mine. This garnet is color mottled with lighter colored garnet (n=1.81) that is increasingly abundant toward the crystal margins. The lowest determined index is 1.76, for pale pink garnet associated in many places with calcite, idocrase, and light-colored silicates in the marginal parts of tactite masses. The occurrence and properties show that the garnet belongs to the grossularite-andradite series. If small amounts of spessartite and possibly almandite are ignored, the indices indicate the following compositions according to Winchell and Winchell (1951, p. 492):

n	*Andradite (percent)*
1.78–1.80	27–41 (most common range)
1.85	72 (highest andradite)
1.76	10 (lowest andradite)

R. G. Coleman (written communication, 1958) found isotropic garnet from the Pine Creek mine to have the following properties:

$$a_o = 11.92 \text{ A}$$
$$n = 1.787 \pm .005$$
$$D_{25°C} = 3.72 \pm .02 \text{ (measured)}$$
$$D = 3.694 \text{ (calculated from } a_o\text{)}$$

According to charts given by Winchell and Winchell (1951, p. 485), the andradite-grossularite ratio is about 3:7, and the garnet also contains notable amounts of spessartite or almandite. An analysis of tactite from the Pine Creek mine indicates 2.9 percent MnO (table 19), which suggests 10 percent or more spessartite in the garnet.

Pyroxene.—The pyroxene generally is grayish green; it may be somewhat lighter colored in the marginal parts of tactite bodies adjacent to marble. Locally black pyroxene is present in the Pine Creek mine. In thin section both green and black pyroxene is pale green. Commonly it occurs in small stubby prisms, but in places is in elongate prisms and larger anhedral grains that enclose grains of calcite and silicate minerals. The optical properties vary, but common ones are as follows:

$$N_x = 1.70\text{–}1.71$$
$$N_y = 1.72$$
$$N_z = 1.735\text{–}1.74$$
$$2V = 60°\text{–}62°$$
$$Z \wedge c = 46°\text{–}47°$$

These properties indicate a pyroxene of the diopside-hedenbergite series that contains 60 to 70 percent of the hedenbergite molecule (Winchell and Winchell, 1951, fig. 302, p. 413). Another sample of grayish-green pyroxene from the Pine Creek mine having approximately these optical properties was studied by R. G. Coleman by X-ray methods; it has the following physical properties (written communication, 1958):

hkl	*dÅ*	
220	3.27	
221		
310	3.003	$D_{25°C} = 3.59 \pm .02$
131	2.91	
$22\bar{1}$	2.555	

According to Zwaan (1954), these physical properties indicate 93 to 95 percent of the hedenbergite molecule.

A black pyroxene from the Pine Creek mine has the following optical properties:

$$N_y = 1.71$$
$$2V = 65°$$
$$Z \wedge c = 43°\text{–}44°$$

The values of N_y and $Z \wedge c$ suggest a pyroxene of the diopside-hedenbergite series having about 55 percent of the hedenbergite molecule. Possibly a small amount of MnO present as the Johannsenite molecule ($CaMnSi_2O_6$) could cause the black color and large optic angle.

Epidote.—Epidote is not common in tactite, but is present locally, and in a few places is abundant. The variety present is iron rich and is megascopically light olive to grayish olive green. The pleochroism in thin section is colorless to greenish yellow. Epidote occurs in irregular-shaped masses and with quartz in elongate crystals that are conspicuously striated parallel with the long dimension of the crystals. Crystals an inch long are common, and some several inches long are found in places. The epidote of tactite always has a large (−) optic angle, and the fact that the value of N_y generally is about 1.75 indicates about 25 percent $HCa_2Fe_3Si_3O_{13}$ and 75 percent $HCa_2Al_3Si_3O_{13}$.

Idocrase.—Idocrase is not a common mineral in tactite. It is apparently confined to iron-poor varieties and has not been observed in iron-rich or scheelite-bearing varieties. In the Round Valley mine it is restricted to certain bedded zones, in the Aeroplane mine it occurs with grossularite in a widespread mass, and in the Pine Creek mine it is in the marginal parts of the tactite in a narrow zone transitional to silicated marble.

The idocrase at the Aeroplane mine is in elongate olive-gray prisms that commonly are arranged in conspicuous radial structures as much as 2 inches across. It is intergrown with grossularite and accompanied by clinozoisite, which appears to have partially replaced the idocrase. In thin section {110} cleavage is conspicuous, and under crossed nicols the mineral is birefringent and shows zonal structures. The optic angle is small and the optic sign negative. Both N_o and N_e fall between 1.70 and 1.71.

The idocrase in the marginal parts of the tactite at the Pine Creek mine is in dark yellowish-brown to moderate olive-brown, irregularly shaped masses as much as 6 inches across. Individual crystals are surrounded by numerous tiny crystals of diopside and wollastonite, some of which are also enclosed in the idocrase. The idocrase is cut by narrow veinlets of calcite, which follow fractures. These relations indicate that the idocrase was formed somewhat later than diopside or wollastonite and grew, partly at least, at their expense. Under crossed nicols, mottled and irregularly concentric zonal structure can be seen.

Sphene.—Sphene is a sporadically distributed minor constituent of many tactite masses, but locally it constitutes as much as several percent of the tactite. Commonly it is in tiny wedges that can be identified only in thin section, but in the Stevens ore body at the Tungstar mine it is in yellowish veinlike chains of crystals that are easily seen. In the Pine Creek mine, sphene is locally abundant in quartz-pyroxene tactite in small wedges having a bright violet color in the Z direction.

Scheelite.—Generally scheelite is very light gray, but some is pale olive or clear. The size and shape of grains is variable. In tactite that consists chiefly of garnet and pyroxene, scheelite commonly is scattered through the tactite somewhat sporadically in tiny equidimensional grains the size of a pinhead or smaller; in tactite that consists chiefly of quartz and epidote, crystals a quarter to half an inch across are common. Scheelite also occurs in irregular streaks and masses, and in thin sheets along fractures. Crystal faces generally are not present against garnet or pyroxene, but commonly are numerous against quartz and in places against epidote.

Under ultraviolet light scheelite composed of pure $CaWO_4$ fluoresces bluish white; with increase of powellite molecule ($CaMoO_4$) the fluorescence is increasingly yellowish. Most scheelite in the Bishop district fluoresces yellowish, and the intensity of yellow is extremely variable within short distances. Some single crystals exhibit several shades of yellow and bluish-white fluorescence. Most bluish-white fluorescent grains are associated with quartz.

BLEACHED AND SILICATED MARBLE

In many places the calcium-rich rock peripheral to tactite is bleached and silicated. In dominantly sedimentary terranes, where tactite bodies are formed adjacent to stocks or other smaller intrusive masses this zone can usually be identified. In dominantly granitic terranes such as the Bishop district, however, the masses of calcium-rich rocks are generally so small that they are bleached and silicated throughout. The only place in the Bishop district where the calcium-rich rock is extensive enough to show the spacial relations between tactite, bleached and silicated marble, and unbleached marble is in the Pine Creek mine. There, marble peripheral to the tactite is bleached and silicated across an outcrop width of several hundred feet.

The inner contact between the bleached and silicated rock and tactite is sharp or gradational through a few feet or less. In places, euhedral crystals of reddish-brown garnet or brownish or greenish idocrase are irregularly distributed in discrete crystals through the marble that lies within a few feet of the tactite. The outer contact of the bleached and silicated zone is highly irregular and is gradational in places.

The most common minerals in the bleached rock other than calcite are highly variable amounts of diopside, wollastonite, potassium feldspar, grossularite, and quartz. In places, zoisite accompanies diopside. All these minerals are in layers that appear to be parallel to bedding and are believed therefore to mark originally impure beds. This interpretation implies that little material has been added to the rock and that the chief process involved was recrystallization and loss of carbon dioxide. In most thin sections studied, wollastonite appears to be stable, and quartz and calcite are not in direct contact with each other where both are present. However, in one thin section wollastonite is present in one layer and blades of calcite are present in a contiguous quartzite layer. A similar anomalous situation has been explained by Watters (1958, p. 715–716) to result from the growth of a compact layer of wollastonite which hindered the escape of carbon dioxide and inhibited reaction between quartz and calcite. An equally acceptable though less likely explanation is that the pressure-temperature conditions were at the position where calcite and quartz as well as wollastonite are stable.

The geometrical pattern of the bleached and silicated zone with relation to the tactite seems to indicate that the bleaching and recrystallization of the carbonate took place in conjunction with the formation of tactite. Possibly carbon dioxide expelled during the formation of tactite contributed in some way to the bleaching and recrystallization. Most of the silicate minerals, however, seem likely to have been formed earlier during the period of thermal metamorphism when peak temperatures reached the range in which wollastonite is formed.

ZONES OF SILICIFIED ROCK AND QUARTZ VEINS

In many contact-metasomatic tungsten deposits, late quartz in veins and in silicified zones has been introduced into both tactite and the adjacent granitic rock. The zones of silicified rock and quartz veins are economically important because in places valuable sulfides accompany the quartz and because scheelite is usually removed from tactite during silicification. The main loci for quartz are the intrusive contacts, but quartz is not confined to these contacts; silicified rock and quartz veins occur in both the granitic rock and in tactite, in places at considerable distance from the intrusive contact. Some quartz-rich tactite may have been silicified. Obviously much, and perhaps all, of the quartz was introduced after the formation of tactite, doubtless as a continuation of the process that resulted in the formation of tactite.

The introduction of quartz has been guided by fractures. Zones of silicified rock can be traced into brecciated tactite or granitic rock, and thin sections of rock adjacent to silicified rock commonly reveal mortar between grains. In partly silicified rock, fractures are evident in the structure of the rock. Small veinlets occupy fractures in which the offset walls either match or very nearly match each other.

Mineral and structural relations suggest that quartz has been introduced over a long period and with falling temperatures. The earliest quartz introduced at any one place is in zones of silicified rock that were formed at least in part by replacement, and the latest is as fracture fillings in veins; however, transitions between these two kinds of occurrences can be found, and in ideal form they appear to represent end members of a continuous sequence.

Zones of silicified rock were formed earlier than veins, and presumably at higher temperatures. Evidence that they were formed, partly at least, by replacement is found in incompletely silicified rock present in many silicified zones, especially near the margins, which still retains the texture of the tactite or granitic rock. In many places, silicified rock contains unreplaced residual masses of the granitic rock or tactite whose orientation was not noticeably disturbed during silicification. In places where no relicts remain it could not be determined whether the quartz has replaced preexisting rock or simply filled open fractures.

Probably the earliest rock to form in the Pine Creek mine during silicification is composed of quartz and green pyroxene. Its presence along the margins of silicified zones in garnet-pyroxene tactite indicates that it was formed through the silicification of the tactite. In this process, scheelite was removed and garnet removed or converted to pyroxene. This rock probably formed at relatively high temperatures, possibly only slightly below those that prevailed at the time of formation of tactite. In other places in the Pine Creek mine where the garnet in the tactite has not been selectively removed, silicification probably took place at somewhat lower temperatures. Narrow zones of silicified rock are transitional to veins in which the walls have been so extensively replaced that no larger irregularities in the opposing walls can be matched. In the youngest veins no evidence of replacement of the walls can be found, and every irregularity in the vein walls matches—a feature that indicates that dilation was the sole means by which space was made available for the emplacement of the quartz.

In the Pine Creek mine, rich ore shoots of molybdenite and of chalcopyrite and bornite have yielded notable amounts of molybdenum and copper. Elsewhere in the district, sulfides other than pyrite or pyrrhotite have been found in tungsten deposits only in small amounts and have not been recovered. Quartz is accompanied in the Aeroplane mine by sphalerite, in the Round Valley mine by galena and ramdohrite ($Pb_3Ag_2Sb_8S_{18}$?), in the Tungstar mine by pyrite, in the Schober mine by pyrrhotite and scheelite, and in the Coyote Creek prospect by galena that contains inclusions of bismuth and schapbachite? ($AgBiS_2$). These sulfides and sulfo-salts are not likely to have been deposited at the same temperature. Molybdenite, chalcopyrite, bornite, and sphalerite probably were deposited in the highest temperature range, galena in a medium range, and the sulpho-salts in the lowest range.

No direct correlation was recognized between the sulfides and the habit of the associated quartz. In the Pine Creek mine, molybdenite, for example, occurs in quartz-rich (presumably silicified) tactite, in silicified granite, and in quartz veins having sharp walls. Molybdenite was locally very abundant in quartz-rich rocks in both the North and South ore bodies above level A, but extensive masses of similar appearing rocks in the Main ore body below level A contain only small amounts of molybdenite.

CHEMICAL GAINS AND LOSSES IN THE FORMATION OF TACTITE

Inasmuch as the constituents of tactite are very different from those of the host carbonate rock and its specific gravity is 25 to 30 percent greater, a large amount of material must have been introduced and a somewhat smaller amount removed in the formation of tactite. In order to calculate gains and losses of substances, the chemical compositions and specific gravities of the host carbonate rock and of tactite must be known, and an assumption must be made regarding the volume relation.

Since Lindgren's study of the Clifton-Morenci district (Lindgren, 1905), it has been customary to assume that the volume has remained almost unchanged during contact metasomatism. Nevertheless, uncertainties exist. It is well known that sediments of mixed calcareous nature can lose considerable volume if thermally metamorphosed without addition of material (Barrell, 1902 and 1907; Cooper, 1957). Barrell (1907, p. 148) states that no general basis exists for calculating volume changes where metasomatism accompanies metamorphism, for not only are some substances brought in but others are taken out. The problem is extremely difficult to deal with in the Bishop district because layers altered to tactite cannot be followed into unaltered rock and the thicknesses of altered and unaltered strata therefore cannot be compared. Most marble contains silicate minerals such as diopside, wollastonite, garnet, and feldspar, which are distributed along bedding planes with no apparent relation to fractures or to intrusive contacts, and apparently formed from impurities in the original limestone without addition of material. These minerals were formed as a result of thermal metamorphism, which is believed to have reached a maximum before contact metasomatism. It seems certain that CO_2 was lost as a result of the formation of the new minerals, and this loss, together with the greater density of the new minerals, implies some shrinkage of the rock. The amount of shrinkage would depend on the kinds and amounts of new minerals formed; impure silicated marble such as is present in the Round Valley septum in the vicinity of the Round Valley mine undoubtedly has shrunk more than lightly silicated marble such as is present in the west side of the Pine Creek pendant. Fortunately, shrinkage caused by nonadditive thermal metamorphism should not affect calculations of gains and losses during contact metasomatism because it can be assumed that both the tactite and the adjacent host rock, with which the tactite is compared, underwent the same changes.

The only evidence bearing on volume relations during contact metasomatism itself is in small irregularities in the contact between tactite and the host rock. These contacts do not indicate any change of volume, although the evidence is not conclusive; therefore, it is assumed in calculating gains and losses that any volume change was negligible.

To gain a quantitative concept of the gains and losses involved in the formation of tactite, five samples from the vicinity of the Main ore body of the Pine Creek mine were analyzed (table 19). Samples 1 and 2 are of partially bleached and silicated strata east of the Main ore body and were not used in the calculation. Samples 3 and 4, which were used, are from strata that, closer to the granitic rock, have been converted to tactite. Sample 5 is of typical tactite from the Main ore body.

The chemical analyses show that the marble is impure and contains significant amounts of SiO_2, Al_2O_3, Fe, MgO, and K_2O. In thin sections, the minerals containing these constituents are recognized to be diopside, orthoclase, grossularite, and pyrite. The calc-hornfels (sample 2) from strata east of the ore zone also contains quartz, grossularite, and wollastonite. The fact that silicate minerals and quartz are distributed along bedding planes suggests that they were formed by simple recrystallization of sedimentary layers without addition of material. If this assumption is true, then it can be assumed further that these impurities were also present in the limestone that has been converted to tactite, and that gains and losses can be calculated directly.

Gains and losses were calculated as follows: The milligrams per cubic centimeter of each oxide present in samples 3, 4, and 5 were calculated. The values obtained for samples 3 and 4, of gray marble continuous with the ore zone, were then averaged, and differences between these average values and the values for tactite were determined. These differences represent the gains and losses in terms of oxides. Gains and losses were also calculated in terms of the metallic elements. However, if, as seems highly probable, the principal loss was

TABLE 19.—*Chemical analyses of tactite and bleached and silicated marble from the main ore body, Pine Creek mine, and gains and losses in the formation of tactite*

[Rapid rock analyses (Shapiro and Brannock, 1956). Analysts: P. L. D. Elmore and K. E. White]

	Marble and calc-hornfels east of tactite zone (weight percent)		Marble continuous with tactite zone					Typical tactite		Gains and losses in transformation of marble to tactite (mg per cc)	
	1	2	3		4		3 and 4	5		Oxides	Elements
Lab. No.	138271	138272	138270		138273			138277			
			Weight percent	*Mg per cc*	*Weight percent*	*Mg per cc*	*Average mg per cc*	*Weight percent*	*Mg per cc*		
SiO_2	19.7	61.0	17.1	465.12	16.0	444.80	455.0	40.4	1,426.1	+971.1	Si +456.4
Al_2O_3	3.6	8.3	3.8	103.36	3.3	91.74	97.6	11.1	391.8	+294.2	Al +155.9
Fe_2O_3	1.4	2.4	1.2	32.64	1.2	33.36	33.0	7.4	261.2	+228.2	Fe +275.4
FeO								4.2	148.3	+148.3	
MgO	2.9	1.8	3.1	84.32	3.0	83.40	83.9	3.9	137.7	+53.8	Mg+ 32.3
CaO	44.7	19.0	43.6	1,185.92	45.1	1,253.78	1,219.9	29.1	1,027.2	−192.7	Ca −142.6
Na_2O	.24	.86	.20	5.44	.20	5.56	5.5	.15	5.3	−.2	
K_2O	.60	2.6	.76	20.67	.76	21.13	20.9	.06	2.1	−18.8	K − 15.6
TiO_2	.16	.41	.16	4.35	.14	3.89	4.1	.29	10.2	+6.1	Ti + 3.6
P_2O_5	.07	.10	.05	1.36	.06	1.67	1.5	.06	2.1	+.6	
MnO	.06	.14	.09	2.45	.12	3.34	2.9	2.9	102.4	+99.5	Mn+ 76.6
H_2O	.12	.46	.84	22.85	.46	12.79	22.8	.20	7.1	−15.7	
CO_2	26.7	2.0	29.0	788.80	30.1	836.78	812.8	.35	12.4	−800.4	

1. Gray marble (1100 level; coordinates 36,750N; 35,560E).
2. Calc-hornfels (1100 level; coordinates 36,450N; 35,730E).
3. Gray marble (1500 level; coordinates 36,265N; 35,680E); sp gr 2.72.
4. Gray marble (1300 level; coordinates 36,200N; 35,660E); sp gr 2.78.
5. Tactite (1500 level; DDH 408); sp gr 3.53.

of CO_2, approximately as much O must have been added as is represented by O in the gained oxides.

The chief calculated losses were of CO_2 and CaO; much smaller losses were of K_2O and H_2O. Total calculated losses in terms of oxides =1027.8 mg per cc, or more than 37 percent by weight of the marble. The amounts of K_2O and H_2O lost, though small, are significant because the losses represent very large parts of these constituents.

The principal calculated gains are of SiO_2, FeO, Fe_2O_3, Al_2O_3, MnO, and MgO, and a smaller gain of TiO_2. The calculated gains total 1801.7 mg per cc, or 51 percent by weight of the tactite. Not reported in the chemical analyses are small amounts of tungsten and fluorine known to be present in scheelite and fluorite. Although the apparent gain of MgO is substantial it is not large, and the MgO content of the gray marble could have been sufficiently variable to cause the apparent gain. On the other hand, an abundance of sphene in some thin sections of tactite indicates that TiO_2 was added. A very small gain of P_2O_5 is indicated, but the amount is so small that its reality is dubious. Na_2O appears to be the only oxide that is clearly unaffected.

During conversion of the impure limestone to silicated marble prior to the formation of tactite, the volume of the rock was reduced about 10 percent, if we assume no addition of substance and loss only of carbon dioxide. This shrinkage is calculated as follows: The analyses of samples 3 and 4 (table 19) show that the gray marble stratigraphically continuous with the tactite contains as impurities about 16.5 percent SiO_2, 3.5 percent Al_2O_3, 3.0 percent MgO, and 0.76 percent K_2O, plus 1.2 percent Fe expressed as Fe_2O_3. The Fe is contained in pyrite and can be neglected. The other oxides are distributed among the following minerals, known to be present in the gray marble: 16.2 percent diopside, 4.5 percent orthoclase, and 11.9 percent grossularite. The diopside contains 3.0 percent of the rock as MgO and 4.2 percent as CaO, and grossularite contains 4.4 percent of the rock as CaO. These quantities of CaO and MgO were derived from dolomitic limestone by expulsion of CO_2 equivalent in weight to 10.1 percent of the rock. This loss figure could be further refined, but in view of the crudity of the basic data and possibility of erroneous assumptions, it suffices to conclude that the volume loss has been about 10 percent.

STABILITY RELATIONS OF GARNET, PYROXENE, AND EPIDOTE

Comparison of chemical compositions (table 20) shows that garnet contains substantially more CaO than pyroxene or epidote, that pyroxene contains more SiO_2 than epidote or garnet, and that epidote is rich in Al_2O_3 and contains H_2O. MgO and FeO are confined to pyroxene, which contains no Fe_2O_3 or Al_2O_3. Thus, a high CaO content probably favors formation of garnet, a high silica content and the presence of MgO and FeO pyroxene, and a high Al_2O_3 content and the presence of H_2O epidote. Pyroxene is absent or present in small quantities in tactite adjacent to marble or carbonate veins, whereas garnet is common. Conversely the silicified wall rock of many quartz veins in tactite consists of quartz and pyroxene and lacks garnet. The state of oxidation of Fe may also be significant in determining whether garnet or pyroxene will form. Butler (1923, p. 398–404) suggested that the larger ratio of ferric to ferrous oxide in limestone contact zones than the ratio in the intrusive rock may result from the high concentration of CO_2 in the contact zone. The equation proposed by Butler is: $Fe_3O_4 + CO \rightleftharpoons 3\ FeO + CO_2$. According to Butler the temperature most favorable for the reaction to move to the left is 490° C. Field and petrographic relations of ordinary tactite unveined by quartz or calcite indicate that garnet and pyroxene formed approximately contemporaneously, and it seems reasonable to suppose that the concentrations of available constituents determined which of these two minerals formed.

The relation of epidote to the other two minerals, however, is uncertain. In some tactite, garnet and epidote appear to be in stable equilibrium with each other, but the textural relations in other masses of tactite and the common association of epidote with

TABLE 20.—*Chemical compositions of some common silicate minerals in tactite*

[Common pyroxene: approximately 25 percent diopside and 75 percent hedenbergite. Common garnet: approximately 10 percent spessartite; andradite: grossularite=3:7 Epidote: Al: Fe=3:1]

	Pyroxenes			Garnets				Epidote
	Diopside	Hedenbergite	Common pyroxene	Grossularite	Andradite	Spessartite	Common garnet	
SiO_2	55.6	48.4	50.2	40.0	35.5	36.4	38.4	37.87
CaO	25.9	22.2	23.2	37.3	33.0	----	32.4	23.51
MgO	18.5	----	4.6	----	----	----	----	----
FeO	----	29.4	22.0	----	----	----	----	----
Fe_2O_3	----	----	----	----	31.5	----	8.5	12.60
Al_2O_3	----	----	----	22.7	----	20.6	16.4	24.13
MnO	----	----	----	----	----	43.0	4.3	----
H_2O	----	----	----	----	----	----	----	1.89

quartz suggest that it formed a little later than pyroxene and garnet. Goldschmidt (1911) identified the assemblage diopside-grossularite among the quartz-bearing hornfels of the Oslo region, which Turner (1948) includes in his pyroxene hornfels facies. Both Barth (1952) and Turner (1948) also represent the diopside-grossularite assemblage as being stable in the amphibolite facies, and Turner (in Fyfe, Turner, and Verhoogen, 1958, p. 205–211) has also included it in his hornblende hornfels facies. Epidote is considered by both Turner (Fyfe, Turner, and Verhoogen, 1958, p. 229–230) and Barth (1952, p. 341) to be stable in the lower temperature range of the amphibolite facies. Inasmuch as there is no evidence in the hornfelsed sedimentary rocks that temperatures attained were as high as those required for the pyroxene hornfels facies, it seems likely that the assemblage pyroxene-garnet formed under conditions of Turner's hornblende hornfels facies. Epidote may have formed in part contemporaneously with pyroxene and garnet in this facies, but in part probably formed somewhat later at lower temperatures and at probably higher H_2O pressures.

LAYERING AND STREAKS IN TACTITE

Much of the tactite is structurally isotropic, and the minerals are evenly distributed, but layered structures and streaks composed of any of the constituent minerals are common. The layers and streaks are of two general types—those that coincide with and appear to be inherited from original bedding, and those that are unrelated or only accidentally related to bedding.

Layering inherited from original bedding generally is regular and exhibits considerable continuity. It is shown by alternating layers of different color and mineral content, and appears to reflect alternating pure and impure layers in the parent limestone. During thermal metamorphism impurities react with one another and with carbonate to form light-colored silicate minerals. Metasomatic solutions, introduced later from the granite, react readily with the remaining carbonate to form the usual iron-rich silicates of typical tactite, and more slowly with the relatively insoluble and inert light-colored silicate minerals already formed. The resultant rock has an overall lighter color than tactite formed from clean marble and consists of alternating darker tactite layers and lighter colored calc-silicate layers. In some tactite, layering is inconspicuous near the granitic rock but becomes progressively more pronounced in the rocks farther from the intrusive.

An excellent example of layered tactite formed from impure limestone is in the Western tungsten mine area where granitic rock transects bedding in what must have originally been thin-bedded shaly and sandy dolomite or magnesian limestone. Alternate layers are rich in garnet (containing some epidote) and in pyroxene. At the contact with granitic rock the tactite is quite dark and obviously layered, though not as conspicuously as it is farther away from the granitic rock where the tactite is lighter colored and where less substance has been introduced. During thermal metamorphism the original sediment must have been largely converted to silicate minerals—argillaceous layers to grossularite and epidote, and siliceous magnesian layers to diopside. Other light-colored minerals also were formed, and calcite probably was present in some layers. Substances introduced from the granitic magma worked away from the contact along bedding planes and reacted readily with the remaining calcite and more slowly with the already formed silicates.

Some parallel layers are of a more subtle nature, and their inheritance from bedding is difficult to prove. In parts of the Pine Creek mine the rock is layered in discontinuous parallel streaks of garnet and pyroxene. The most conspicuous layering is in tactite that is notably lighter colored than usual, but obscure layering is also present in tactite of typical darker color. Some of the layered tactite of lighter color is in marginal parts of the tactite, and the layering is parallel to bedding in adjacent marble. Lighter colored tactite generally contains less scheelite than darker tactite from the same mass, and where the tactite is layered, scheelite is in streaks parallel with the layers. The lighter colored layers having a lower scheelite content originally contained more impurities than the darker layers and received smaller amounts of substance from the magma. The observed parallelism of layering with bedding suggests that some and perhaps all of the layering in abnormally light-colored tactite is inherited.

Streaks of silicates independent of bedding control are believed to follow fractures that formed in tactite while it was still quite hot, and while solutions from the granitic magma were still circulating. Streaks and irregular masses of garnet and of pyroxene and quartz are common, and streaks and masses of epidote are present locally. Formation of garnet is probably conditioned by circulating calcium-bearing solutions, which reacted with pyroxene to form garnet, inasmuch as the common garnet of tactite contains considerably more calcium than diopside. Magnesia must have been subtracted and some alumina probably was added in the transformation. Small amounts of quartz present locally in garnet streaks and masses may actually have been released as a consequence of the conversion of pyroxene to garnet. Quartz-pyroxene streaks that may or may not contain sulfides appear, on the other hand, to result from the introduction of silica. Many early formed quartz veins in tactite are bordered by zones of quartz and pyroxene unaccompanied by garnet. The common pyroxene of tactite contains more silica than the common garnet (table 20). Alumina must have been removed and magnesia probably was introduced. Most quartz-pyroxene streaks contain no scheelite, although sulfides are common. Locally, however, scheelite is abundant in large euhedral to subhedral crystals. These crystals generally lie in the marginal parts of the quartz-pyroxene streaks adjacent to common tactite. In a few places scheelite lies in thin sheets along fractures in tactite, quartz, and even granite.

DISTRIBUTION OF SCHEELITE

The distribution of scheelite in tactite is often spoken of as "spotty," and this characterization is apt. Nevertheless, the spotty distribution cannot be dismissed as the result of capriciousness—it must result from definite processes and relations. A complete explanation of the distribution of scheelite in tactite will require extensive investigations, and at this time only a few simple relations can be described on the basis of data available.

In the Bishop area, scheelite-bearing ore shoots in tactite can be considered to be of two general kinds, those in ordinary garnet-pyroxene tactic composed of iron-rich silicates and poor in quartz, and those in rocks abnormally rich in quartz. Ore shoots in ordinary tactite are more abundant and generally more extensive than those in quartz-rich tactite, but those in quartz-rich tactite include the highest grade ore shoots. Most quartz-rich tactite contains very little sheelite, and the local presence of rich ore bodies in such rocks is decidedly anomalous.

In garnet-pyroxene tactite, scheelite usually is disseminated through the tactite in rounded grains that range from pin-point size to half an inch across. Generally the distribution of scheelite is not uniform, and the quantity present from place to place may range widely. If the tactite is structurally isotropic, the pattern of scheelite distribution also is structureless, whereas if the tactite is layered, the scheelite generally is distributed in streaks parallel to the layers.

In garnet-pyroxene tactite, dark-colored tactite is a commoner host for scheelite than lighter colored tactite. A reasonable explanation for this association is that such tactite usually is formed from relatively pure marble, and consequently its formation involved introduction of the maximum amounts of silica, alumina, iron, and other substances, including tungsten, from the granitic magma. Tungsten is, of course, simply one of the constituents introduced during metasomatism. Variations in the amount of scheelite within apparently homogeneous tactite are difficult to explain. However, even in the most homogeneous tactite some variation in the proportions of minerals is present, probably as a result of variations in the amount of available silica, lime, alumina, magnesia, iron, manganese, and other constituents at the time of formation. Fluorite is present in some of the richest tungsten ore in the Pine Creek, Adamson, and Brownstone mines, but was not identified in ore from any other mines in the district.

Scheelite is generally appreciably less abundant in light-colored tactite whose formation involved the introduction of a smaller total amount of substance. Light-colored tactite can be formed under at least two circumstances. Most commonly it is formed from impure limestone that was already partly altered to silicate minerals at the time of metasomatism. Such tactite commonly is layered, and scheelite is largely confined to darker layers to which the greatest amount of magmatic substance was introduced. Light-colored tactite may also constitute transitional zones between dark tactite and marble. Tactite in this setting generally is composed of grossularite, idocrase, and other iron-poor silicate minerals, and possibly calcite. Its formation from clean carbonate rock involves the addition of as much silica as does the formation of iron-rich tactite, but obviously smaller amounts of iron and tungsten were introduced. Watters (1958, p. 703–724) has shown that silica has traveled farther from the granite contact than alumina, iron, and titanium in skarn in the eastern Pyrenees. Possibly some light-colored tactite formed from clean marble results from a paucity of iron, titanium, manganese, and tungsten, and a preponderance of silica and alumina among the introduced substances.

Tactite containing more than a few percent quartz ordinarily is poor in scheelite, probably because scheelite was soluble in the solutions from which quartz was deposited. Quartz commonly is a late mineral in tactite, and its appearance may mark the beginning of the period of silicification that followed immediately after the period of tactite formation. Locally, scheelite is present in late quartz masses, generally in clusters or masses of large euhedral to subhedral crystals that provide an impressive display under fluorescent light. The presence of these sporadically distributed masses of scheelite in quartz is consistent with the hypothesis that tungsten was contained in the solutions that deposited the quartz but that special conditions were required to bring about the precipitation of scheelite. Possibly the essential condition for scheelite precipitation was the availability of calcium, for quartz has been deposited not in limestone or marble, but in tactite and granite where calcium was usually bound in silicate minerals and was not available to form scheelite.

Whatever the special conditions required may be, they have been met in the district at several deposits where the richest ores, containing several percent WO_3, have been mined. These deposits include the Schober mine, where the ore consisted chiefly of quartz and pyrrhotite, and lesser amounts of garnet, epidote, and calcite; and the Lakeview mine, where the ore consists of epidote and quartz, and some calcite. Equally rich ore from the Tungstar mine was composed of pyrite, oligoclase, and quartz, accompanied locally by garnet and epidote. Although quartz is not abundant, the presence of pyrite, which like the other sulfides commonly is associated with quartz, suggests that silica was present in the depositing solutions.

These quartz-rich deposits appear to have been deposited a little later, and presumably at lower temperatures, than deposits in quartz-poor but iron-rich tactite, and may constitute a transition to hydrothermal vein deposits. In a few places scheelite has been observed in fractures, where it obviously was formed later than the enclosing rock. In the Western tungsten mine, where the tactite is conspicuously layered because it was formed from thinly interlayered marble and calc-hornfels, much of the scheelite is along fracture planes in paper-thin crystals as much as an inch across. The tactite breaks preferentially along these fractures, yielding surfaces that fluoresce brilliantly under ultraviolet light and give an erroneous impression of high-grade ore. In the Pine Creek mine, thin veinlets of scheelite cut across quartz veins, and on the east side of Mount Tom, in the vicinity of the Lambert mine, scheelite is present in places along fractures or joints in the granitic rock adjacent to marble.

LEACHED OUTCROPS AND SECONDARY ENRICHMENT

Leaching of tungsten from the outcrops of ore bodies and its secondary enrichment at depth are controversial subjects among geologists and engineers who have worked with tungsten deposits. Gannett (1919) showed that preferential leaching of tungsten from outcrops by sulfate-bearing waters is theoretically possible, but almost no reliable field observations in support of either leaching or secondary enrichment of tungsten have been published, and the differences of opinion are not likely to be resolved until more observational data are available. In the Bishop district, geologically recent, deep dissection of most parts of the region precludes the kind of ground-water circulation that favors surface leaching or secondary enrichment at depth, and the tactite in most outcrops is little altered.

Nevertheless, abundant pyrite in the Tungstar mine and abundant pyrrhotite in the Schober mine were oxidized near the surface. Both deposits were notably high-grade at the surface, and the grade of the ore fell at depth. These deposits are in small inclusions, however, and a characteristic of such deposit is decreasing grade at depth as a result of the primary mineralization. The ore in the oxidized zone doubtless was enriched as a result of the removal of the sulfides. The scheelite in the surface outcrop of the Tungstar mine exhibited no evidence of leaching or enrichment. No record of the mineral content of the near-surface oxidized ore is available for the Schober mine, and it is not known whether the scheelite was altered.

At the Yaney mine the disposition of minerals provides a basis for postulating secondary enrichment (chiefly residual), but the mine is unique in the district. The chief tungsten-bearing mineral is ferberite, which occurs in pyramidal crystals that are pseudomorphs after euhedral scheelite. These crystals are embedded in a matrix of jarosite, opal, and quartz, and the matrix also contains dispersed tungsten in an unknown form. A small amount of scheelite also is present, and tungstite has been identified. The setting of the deposit, which is in the marginal part of a small calcareous inclusion adjacent to alaskite, and the presence of partly altered remnants of tactite and calc-hornfels layers suggest that the deposit is an altered tactite deposit. The position of the deposit, adjacent to a range-front fault, suggests that hot spring waters were the agents of alteration. The average grade of the ore is about 2.0 percent, which is high for an average tactite body, but not abnormally high if the change in density through loss of calcium, iron, and other substances is considered. The fact that some of the tungsten is dispersed indicates some movement of tungsten, though perhaps through only limited distances. The euhedral form of the pseudomorphs probably results from growth of the original scheelite crystal to obtain the euhedral form, for the scheelite in tactite is not generally in euhedral crystals.

FACTORS THAT INFLUENCE POSITION, SIZE, AND SHAPE OF ORE BODIES

Tactite bodies exhibit great diversity in their size, shape, internal structure, and scheelite content. Tactite does not occur along all contacts between marble and intrusive rock, and it is notably irregular in thickness and scheelite content where it does occur. Along some contacts between intrusive granitic rock and marble only a thin selvage of tactite a few inches thick is present, whereas along others great masses 100 feet or more thick occur. At the Marble tungsten mine a few inches of wollastonite is present along most contacts between quartz monzonite and marble, but two tactite masses having average thicknesses of about 10 feet are also present, and one of these has been followed underground more than 200 feet. Along the west side of the Pine Creek pendant, granitic rock is in contact with marble for more than 3 miles, yet all the thick masses of tactite that constitute known ore bodies are localized along a 3,000-foot segment at the north end. Within this span only about 1,000 feet is occupied by tactite ore bodies. Here, tactite is present everywhere along the intrusive contact, but is a few inches, or at most a few feet, thick between ore bodies; the ore bodies, in contrast, range from 50 to 200 feet thick. Much of the tactite between ore bodies is partially silicified and contains little or no scheelite.

Some tactite masses are tabular, some are tubelike or chimneylike, and many exhibit complex shapes that defy simple descriptive terms. The longer dimensions of some tactite masses lie along intrusive contacts, whereas the longer dimensions of others extend outward from the contact along beds or fractures in the metamorphic rocks. Tactite masses having regular shapes dip and plunge in various directions.

The distribution of scheelite in commercial amounts within tactite is almost as varied as the distribution of tactite itself, but in many places the shapes of ore shoots reflect in a modified way the configuration of the enclosing tactite masses.

In some places it is possible to relate the size, shape, or tungsten content of tactite to recognizable features of the geologic setting, and every irregularity of the tactite bodies no doubt has its explanation, even though no relation has been established. Three features seem to have been of special significance in determining the size and shape of tactite masses: (1) irregularities in the intrusive contact, (2) the stratigraphy and lithology of the metamorphic rock, and (3) fractures along the intrusive contact and in the calcium-rich host rock.

IRREGULARITIES IN THE INTRUSIVE CONTACT

Many tactite bodies that constitute ore bodies are localized in embayments, troughs, and other irregularities in intrusive contacts. Irregularities are most common where the beds in the metamorphic rocks meet the intrusive contact at an angle in dip, in strike, or in both. Along such contacts the granitic rock commonly alternately follows and cuts across the bedding—a feature that produces steps in the contact. At crosscutting segments of contact, sills and apophyses of granitic rock may penetrate along bedding planes in the metamorphic rock.

In the Bishop district, irregularities in which granitic rock wraps around and embraces a salient of marble are especially effective as traps for tactite. The largest number of ore bodies are associated with irregularities in which granitic rock overlies marble; fewer ore bodies are associated with irregularities having steeply dipping axes in which granitic rock wraps around tactite in plan. Irregularities formed by a salient of granite in marble are apparently much less effective traps. Only one small ore body, that at the Lambert mine, was associated with a bench in the intrusive contact where marble overlies granitic rock, and no tactite bodies of commercial size or grade have been found associated with steeply dipping salients of granitic rock in marble.

Large tactite bodies at irregularities in igneous contacts may be the result of extra heat supply and of movement of mineralizing solutions. Metamorphic rocks that projected into granitic magma received the maximum available heat because a large area per unit of volume was exposed to the magma. The metamorphic projections into the magma chamber were also exposed to magma for a longer period because the granitic intrusives cooled inward from their margins (see fig. 55A).

Control over the movement of mineralizing solutions is suggested by the shape of many ore bodies and ore shoots. Vertical elongation of many ore bodies and preferential localization of others beneath granitic dikes and apophyses in marble suggest that these solutions moved generally upward along the intrusive contact. Irregularities having steeply dipping axes very likely acted as channelways, whereas irregularities having horizontal or gently dipping axes probably forced the solutions to spread laterally and thus acted as traps. Control of the movement of mineralizing solutions was probably aided by fracturing, which provided greater permeability. Fracturing could have been caused by stresses related to the emplacement and cooling of an intrusive or from regional forces. Whatever the cause, fracturing was probably more intense where irregularities were present than where the intrusive contact was smooth and regular. In addition to providing increased permeability conducive to the movement of mineralizing solutions, fracturing would also provide

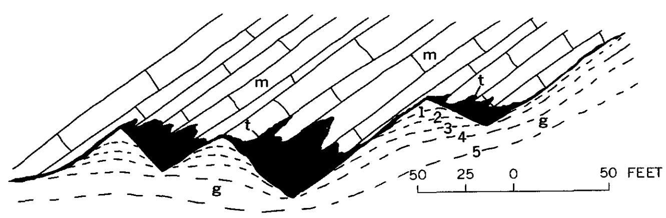

FIGURE 55.—Hypothetical sketch map showing common relations between marble (m), granitic rock (g), and tacite (t) along a discordant intrusive contact. Dashed lines in granite represent successive boundaries between magma and crystallized rock as the magma crystallized away from the contact.

greater surface area for the solutions to come into contact with large volumes of rock, including marble.

STEEPLY DIPPING SALIENTS OF MARBLE

The South ore body in the Pine Creek mine is the best example of a tactite body localized in a steeply plunging irregularity in an intrusive contact where granitic rock wraps around metamorphic rock in plan. Except for the Main ore body in the Pine Creek mine, the South ore body has yielded more tungsten, molybdenum, and copper than any other in the district. The ore body is at the nose and on the west side of a wedge-shaped marble salient in quartz monzonite along the west side of the Pine Creek pendant (fig. 56). At the surface, quartz monzonite envelopes the north end of the marble salient, and a tongue of quartz monzonite penetrates south into the east side of the marble for about 100 feet. The marble salient dips vertically, and the nose plunges southward at about 60°.

The tactite ore body dips vertically, rakes about 60° to the south, and has been followed downward from the surface for more than a thousand feet. The greatest thickness is near the surface. The ore body was more than 150 feet wide and 300 feet long in an open pit, but less than 100 feet wide on level C, about 90 feet deeper, and not more than 50 feet wide on level A, about 275 feet beneath the open pit; 600 feet beneath level A it is only about 15 feet thick. The ore body contained a single ore shoot, which became narrower and shorter at depth with diminishing size of the enclosing tactite. At levels A and C the tactite in the tip of the salient is barren, probably as a result of late silicification.

Molybdenum in commercial amounts was restricted to the upper part of the ore body, where it was associated with silicified tactite and quartz veinlets. Only insignificant amounts were present on levels A and C.

Exploration of the deeper part of the South ore body shows that its bottom coincides with the lower limit of the reentrant in the intrusive contact; the reentrant was obviously the dominant control in the localization of the ore body. Nevertheless, other types of controls are suggested by the shape of the ore body. The thin tail at the south and at depth may have resulted from a favorable bed that lay along the contact, or from premineral fracturing along the contact. The abrupt upward thickening of the ore body and the presence upward of rich molybdenum ore suggest a former bench or protrusion of granite above the ore body analogous to the one at the Main ore body.

Several smaller masses of tactite elsewhere in the district are also embraced in granitic rock in reentrants in the intrusive contact. Among these are the Little Egypt (fig. 57) and Coyote Creek prospects in the Bishop Creek pendant and the Tungsten Peak prospect in the Tungsten Hills. Each of these tactite masses occupies a small segment of an otherwise regular intrusive contact along which a thin selvage of tactite is present. They have not been explored underground because the scheelite content is low.

BENCHES OR APOPHYSES OF GRANITIC ROCK

Examples of ore bodies beneath benches or apophyses of granitic rock are numerous and include the Northwest ore body of the Adamson mine, the North ore bodies of the Pine Creek mine, the upper part of the Main ore body of the Pine Creek mine, and smaller bodies in the Marble tungsten mine. At all the deposits the marble just beneath intrusive rock has been converted to tactite. Downward the tactite passes into marble, which is usually silicated, except along intrusive contacts where tactite may persist downward with diminishing thickness. In some deposits the lower contact of tactite is with silicated marble at a sharp, though in places highly irregular, contact. In others, tactite passes downward through progressively lighter colored tactite to silicated marble. The mineralizing solutions probably worked upward along the intrusive contacts and reacted with the adjacent marble to form tactite; they were slowed down or stopped upward beneath benches of granitic rock and permeated into the adjacent marble.

The Northwest ore body of the Adamson mine was at the end of a nearly vertical slab of metamorphic rock, principally marble, which was split off from the west side of the Pine Creek pendant at the time of the intrusion of the quartz monzonite. The northwest end of the slab is enveloped by quartz monzonite. Tactite was present everywhere along the intrusive contact, but was much thicker in the upper part of the marble beneath quartz monzonite than in the sides or bottom of the marble.

The North ore body of the Pine Creek mine lay along a segment of intrusive contact where the contact and beds are discordant in dip. The general attitude of the contact is vertical, and the adjacent beds dip 75° east. Sills penetrate from the main contact downward along bedding planes. The ore body consisted in reality of several small ore bodies; an upper one cropped out at the surface, and lower ones lay beneath apices formed by the intersections of sills with the main body of quartz monzonite (fig. 58). The lower ore bodies widened downward for a distance, then split into two limbs separated by marble. The upper ore body widened upward to the surface where it was split longitudinally by a sill-like mass of quartz monzonite. The upper body, like the lower body, was probably localized beneath a cap of quartz monzonite, which has been eroded away.

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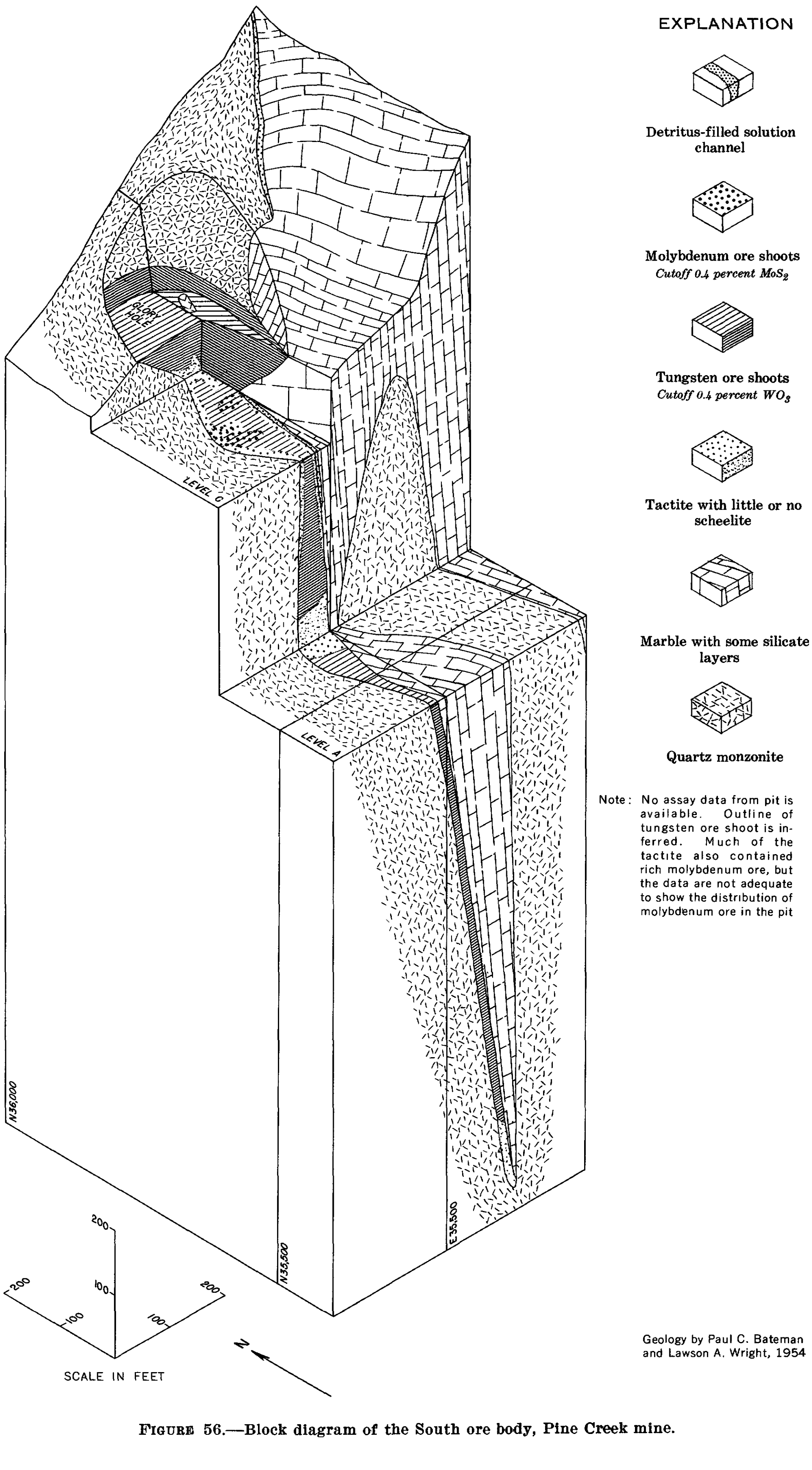


FIGURE 56.—Block diagram of the South ore body, Pine Creek mine.

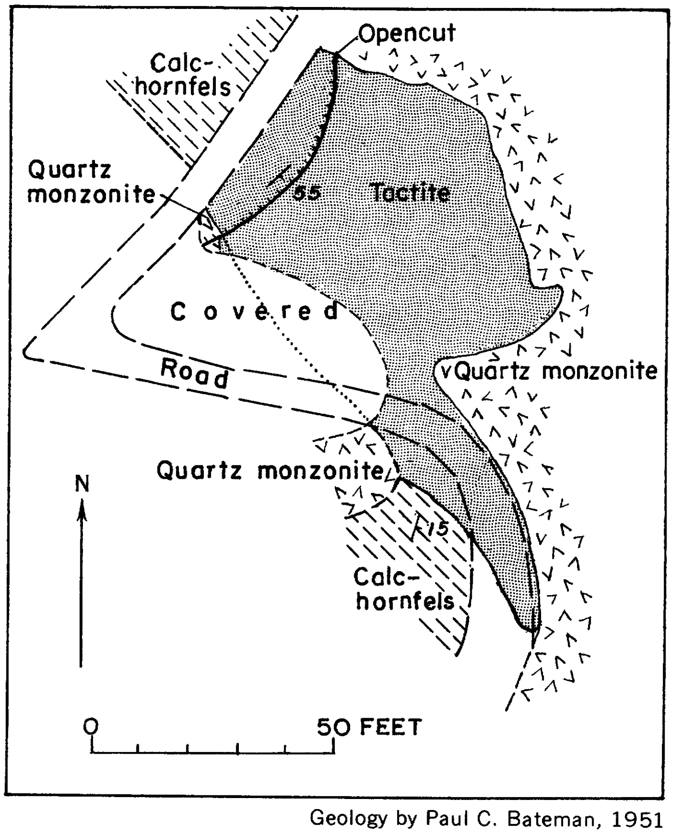


FIGURE 57.—Geologic sketch map of the Little Egypt prospect, sec. 7, T. 8 S., R. 32 E.

Although the principal control over localization of the Main ore body of the Pine Creek mine was not an irregularity in the intrusive contact, the ore body is locally thicker and contains rich molybdenum ore beneath two benches of granite. The upper bench is about 150 feet and the lower bench about 500 feet beneath the outcrop (fig. 59). At the lower bench a wedge-shaped mass of silicified granitic rock extends more than a hundred feet downward along the bedding in the marble. The thickness of the ore body averages about 60 feet above the upper bench, about a hundred feet between the benches, and more than 130 feet below the lower bench for about 150 feet. Between the benches the molybdenum ore lies along the intrusive contact, and below the lower bench it follows the wedge of silicified granitic rock. The tactite body probably was more fractured near the benches in the contact because of the irregularity in the contact. Such fracturing would have permitted late introduction of quartz and sulfides.

At the Marble tungsten mine in Shannon Canyon, tactite was formed locally along the main intrusive contact and also within the marble beneath a thin dike of quartz monzonite (fig. 60). This dike enters the marble flatly, but sends offshoots downward along the beds. Tactite bodies underlay the dike and extended downward along the main quartz monzonite contact and along contacts with the sill-like offshoots from the dike.

SMALL INCLUSIONS OF METAMORPHIC ROCK

Deposits in small inclusions of metamorphic rock are closely allied with deposits localized beneath benches or apophyses of granitic rock in terms of the factors that controlled the mineralization. Deposits in small inclusions have yielded disproportionately large amounts of tungsten ore, and it would be erroneous to equate small inclusions with small tungsten deposits. Most inclusions, though small in terms of the features that can be represented on the maps (pls. 1–4), are sufficiently large to contain extensive tungsten ore bodies. The yield from the Tungstar mine, which is in a small inclusion, is exceeded only by that from the Main and South ore bodies of the Pine Creek mine. Large yields have also come from the Schober, Rossi, Little Sister, Tungsten Blue, and Aeroplane mines, and lesser yields have come from the White Caps, Chipmunk, and Jackrabbit mines, all of which are in small inclusions.

The term "inclusion" implies that the body is enclosed in intrusive rock in outcrop, and is underlain and before erosion was overlain by intrusive rock. The bottoms of inclusions at the Jackrabbit and Tungsten Blue mines were determined in mining, and the inclusion at the White Caps mine is capped by quartz monzonite. Furthermore, at the Tungsten Blue, Schober, and Tungstar mines, the fact that the inclusions broadened downward from very small outcrops suggests that the inclusions were capped by igneous rock at a level not much higher than the present erosion surface.

Although an ore body can occupy an entire inclusion, and apparently does at the Tungsten Blue mine, most inclusions contain nonscheelite-bearing rocks. All the stratigraphic, lithologic, and structural controls that are effective in localizing ore bodies in the margins of larger pendants or septa are also effective in small inclusions; nevertheless, the capping of intrusive rock over inclusions overshadows all other controls. Where tactite is present it generally is in the tops of inclusions, and it passes downward into marble in a fashion similar to that of tactite beneath benches and apophyses along more extensive intrusive contacts. The amount of scheelite commonly also decreases with depth as the tactite becomes lighter colored or passes into other calcareous rocks in which the effects of additive metasomatism are less apparent. The distribution of the rocks suggests that mineralizing solutions worked upward along the sides of the inclusions to their tops where the solutions accumulated and reacted with the marble to form tactite.

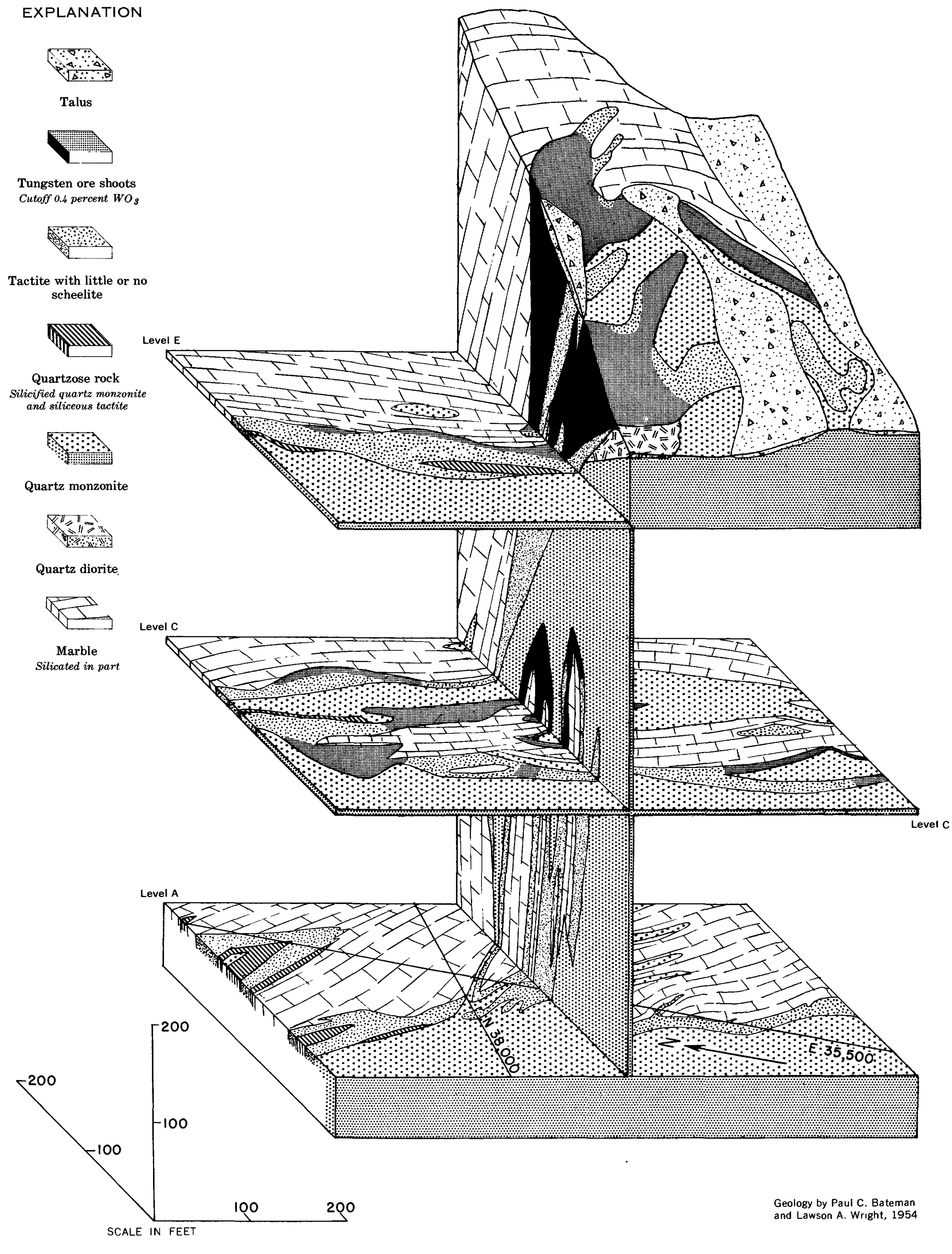


FIGURE 58.—Sectional diagram of the North ore body, Pine Creek mine, showing the tungsten ore shoots.

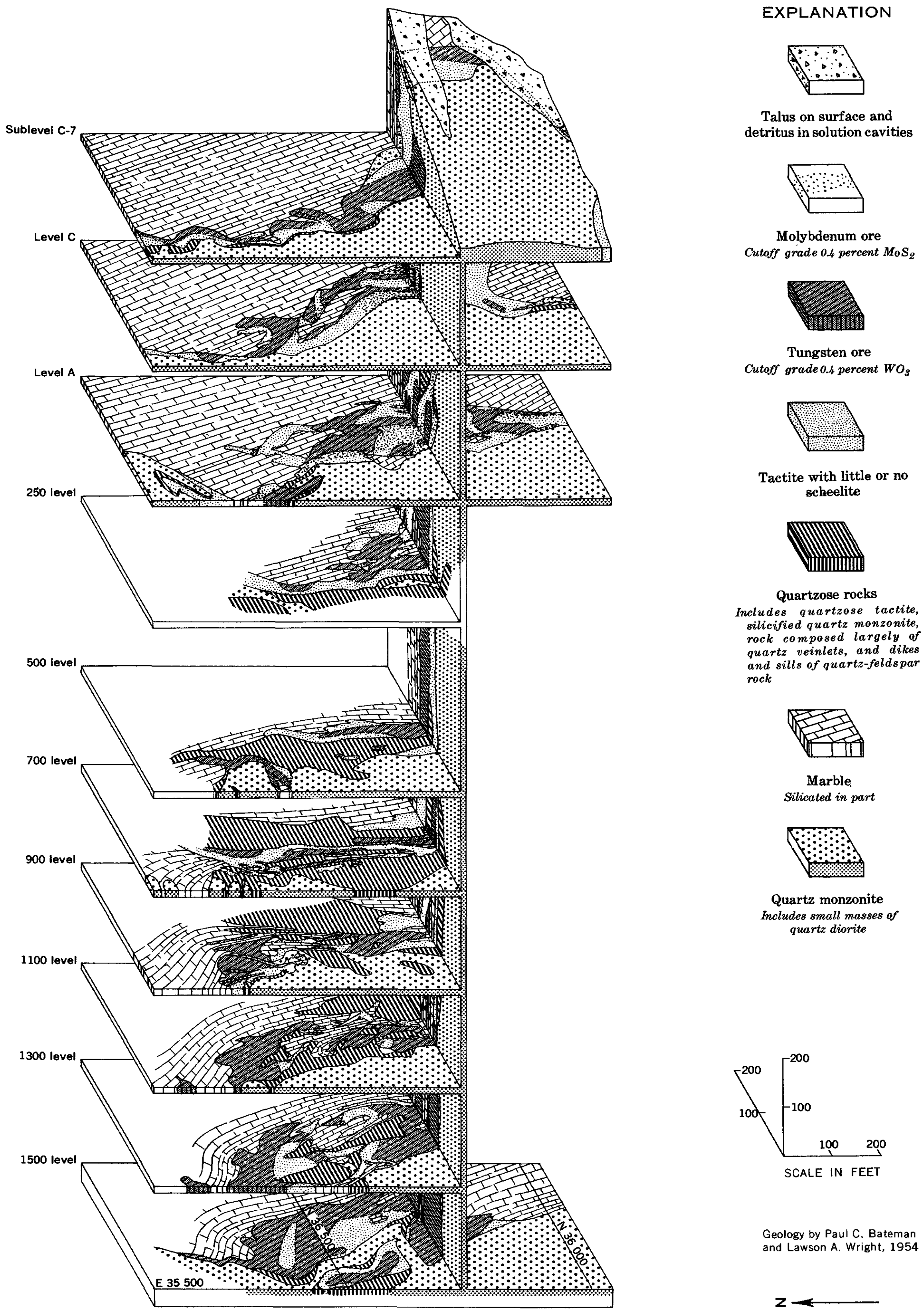


Figure 59.—Sectional diagram of the Main ore body, Pine Creek mine, showing the tungsten and molybdenum ore shoots.

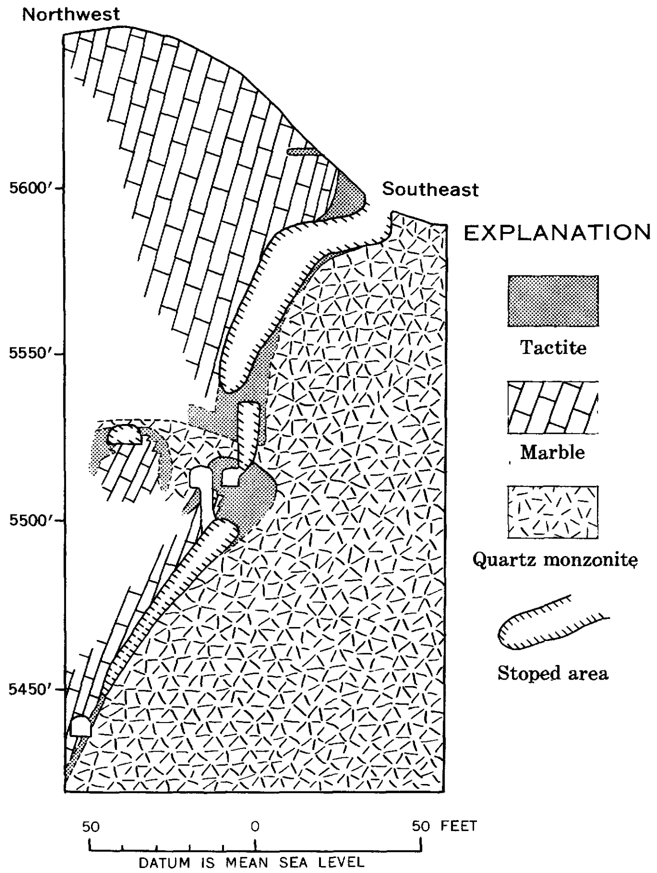


FIGURE 60.—Section through the Marble tungsten mine showing the relation of tactite ore bodies to the intrusive contact.

Where many small inclusions are present in a small area, as in the Deep Canyon (Tungsten City) area of the Tungsten Hills, inclusions containing tactite ore bodies are generally accompanied by more numerous inclusions composed of marble, light-colored silicated rock, or noncalcareous rock. Some of the calcareous inclusions exposed at the surface may be the roots of inclusions that contained tactite ore bodies in their eroded upper parts. Concealed inclusions must also be present in these areas and, if a practical method for locating them were found, they might provide a significant source for tungsten ore. An attempt to locate them with a magnetometer proved unsuccessful (Bateman, 1956, p. 21–22 and fig. 2, p. 50).

Following are brief descriptions of the geological relations in some of the deposits in small inclusions:

The Tungstar mine, on the west side of Mount Tom at an altitude of 12,000 feet, consists of two ore bodies in separate inclusions in quartz-diorite. Almost all of the production came from the Greene ore body, a steeply dipping tabular inclusion having a vertical extent of more than 400 feet, an average width of about 100 feet, and an average thickness of about 25 feet. Most of the inclusion was tactite, but the upper part contained some pelitic hornfels and the lower part had relicts of coarse white marble. The mineral content of the tactite is somewhat unusual: it consists chiefly of garnet, epidote, and pyrite (oxidized in the upper part of the ore body), but locally includes quartz, oligoclase, apatite, and sphene. The scheelite generally occurs in large crystals, a quarter to half an inch across. Decrease in the WO_3 content of the mill heads during the life of the mine indicates that the scheelite content of the ore decreased with depth. The grade diminished from 2.6 percent of WO_3 for the first 17,000 tons of ore mined in 1939 (Lenhart, 1941, p. 67–71) to about 2.0 percent in 1943, then to a little more than 1.0 percent in 1946 (Bateman, 1956, p. 37). Analyses of cores taken below the 280-foot level indicate further decrease in grade to 0.73 percent WO_3.

The Schober mine, in the Bishop Creek pendant, is in an inclusion composed of tactite and hornfels in the marginal part of a mass of hornblende gabbro. In a shallow glory hole the inclusion is elliptical in plan—about 100 feet across in the longer direction and about 60 feet across in the shorter. Tungsten ore of commercial grade was found only in the upper part of the inclusion; the floor of the glory hole is approximately the lower limit of ore. The ore was coarsely crystalline rock composed chiefly of pyrrhotite (oxidized to about 20 feet beneath the surface), quartz, and scheelite, and a little garnet, epidote, and calcite. The typical garnet-bearing tactite here contained very little scheelite. The contact between the inclusion and the enclosing hornblende gabbro is steep except on the west side where the inclusion clearly dips under the hornblende gabbro. Before erosion the inclusion was probably covered by hornblende gabbro only a short distance above the outcrop.

The grade of the ore in all the ore bodies in the Deep Canyon (Tungsten City) area of the Tungsten Hills decreased with depth, and the tactite passed downward into light-colored silicate rock and marble. Only at the Tungsten Blue and Jackrabbit mines. however, can a capping of intrusive rock be observed. At the Tungsten Blue mine the deposit was partly covered at the surface by epidotized gabbro.

The only outcrop of tactite at the White Caps mine is a nearly barren showing which is entirely surrounded by quartz monzonite. This unpromising outcrop is one end of an elongate, steeply dipping mass of tactite that extends laterally beneath quartz monzonite for more than 100 feet. The upper contact of the tactite with quartz monzonite is very nearly horizontal. An ore shoot within the tactite was stoped downward to the water table, which is only about 25 feet beneath the top of the ore body. The ore body has not been explored below the water table, but the presence of marble in the bottom of the stope and analogy with other deposits in inclusions suggest that the grade of ore will diminish with depth and that commercial ore may not continue to the bottom of the inclusion.

STRATIFICATION AND LITHOLOGY OF THE METAMORPHIC ROCKS

The limitation of tactite to calcium-rich rocks and especially to cleaner marble is an obvious example of lithologic control. Because the lithology can change abruptly from bed to bed, though it changes more gradually along beds, the lithologic control generally is also stratigraphic. This kind of control is common where beds of clean marble are interlayered with heavily silicated marble, with calc-hornfels, or with noncalcareous strata. Excellent examples are present at the Western and Round Valley mines in the Round Valley septum. At the Western tungsten mine thinly bedded calc-hornfels and silicated marble is interstratified in pelitic schist and hornfels. The calcareous rocks have been converted to tactite and light-colored silicate rock in alternating layers.

At the Round Valley mine the prevailing rock is somewhat argillaceous marble, which is bounded on the west by pelitic schist and hornfels and on the east by partly garnetized calc-silicate rock. The ore body apparently was formed in argillaceous marble in preference to other strata that contained little or no free calcite after thermal metamorphism. Within the ore body, the scheelite-bearing ore shoots are closely restricted to the cleaner beds (fig. 61).

Stratigraphic control is indicated in some places where no compositional or lithologic differences between the mineralized and unmineralized beds are recognizable. For example, at the Munsinger prospect in the Bishop Creek pendant, quartz monzonite cuts across the bedding in calc-hornfels at a large angle, and a few beds are mineralized outward from the intrusive contact (fig. 62). Beyond the mineralized zone all the beds in the calc-hornfels appear to be of about the same lithology. Almost identical relations exist in the nearby Lindner, Waterfall, and Stevens prospects. Conceivably some subtle differences of lithology exist between the mineralized and unmineralized beds, but it seems equally likely that the mineralization followed premineral bedding-plane fractures.

Another example of stratigraphic control where no obvious differences in the lithology of the mineralized and unmineralized beds were recognized is the Main ore body of the Pine Creek mine (fig. 59). The east boundary of this ore body coincides in most places with a stratigraphic horizon. The marble in extensions of the mineralized beds appears to be very similar to marble in beds just outside the ore zone. Here also, slight differences in lithology could have escaped notice, but a premineral bedding-plane fracture that has been obscured by the formation of tactite or by recrystallization in the marble is equally likely to have controlled the boundary of the ore body.

FRACTURES ALONG THE INTRUSIVE CONTACT AND IN THE CALCIUM-RICH HOST ROCK

Fractures open at the time of tactite formation along the intrusive contact and in the adjacent calcium-rich host rock probably contributed to the localization of virtually every tactite ore body in the district, including those localized in "favorable beds" and those associated with irregularities in the intrusive contact. Permeability was required for the movement and accumulation of mineralizing solutions along intrusive contacts; fractures seem to provide the most likely avenues for such movement. The lithology of the beds can govern the intensity of additive metasomatism once the solutions come in contact with the calcium-rich rocks, but the permeability of the beds must have been too small for the movement of ore solutions, else ore bodies would not be as closely limited to the intrusive contact as they are. Nor could irregularities in the contact have provided permeability in themselves. Contact irregularities are probably effective principally in that they are apt to be loci for fracturing.

Many detailed descriptions appear in British literature of reciprocal exchange of substance along contacts between granitic rock and marble (Joplin, 1935a; Muir, 1953a; Gindy, 1953; Watters, 1958.) In all these descriptions it is shown that silica, alumina, and iron, and in some places magnesia and alkalis have been introduced into the marble, and that calcium has been introduced into granitic rock. In all places where reciprocal exchange of substance has been demonstrated the contact zone is narrow, generally measurable in inches rather than in feet. The relations along such contacts seem entirely compatible with the concept of a frozen contact along which limited amounts of material moved from the cooling granitic magma into the adjacent marble and from the marble into the adjacent granitic rock or magma.

The accumulation of masses of tactite a hundred feet or more thick, however, such as form many tactite ore bodies, seems unlikely to have resulted from the direct flow of material across the contact from the contiguous granitic rock. The mere irregularity in distribution and thickness of tactite requires the accumulation of material bled to the contact from a large volume of igneous rock and also demands circulation along the contact. The tactite beneath granitic apophyses in the

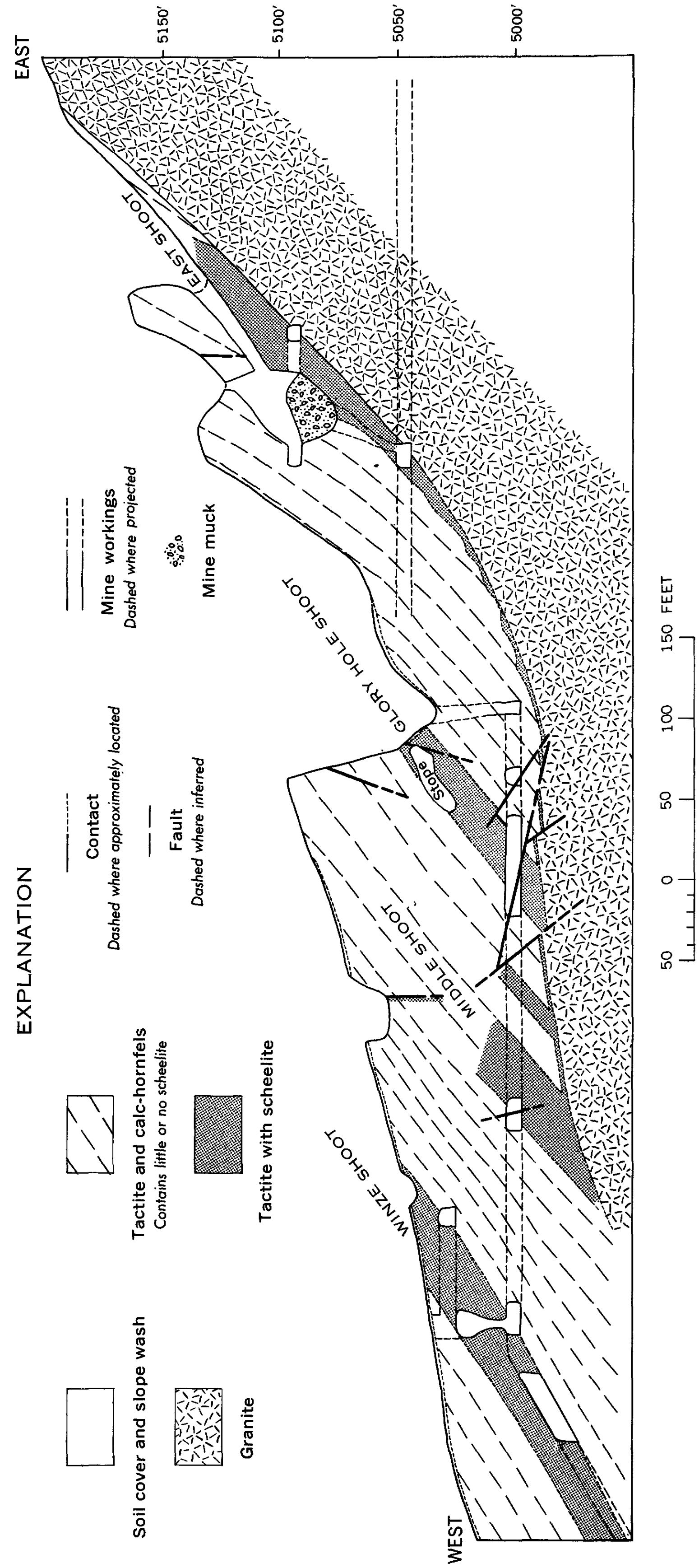


FIGURE 61.—Geologic section through the Round Valley mine.

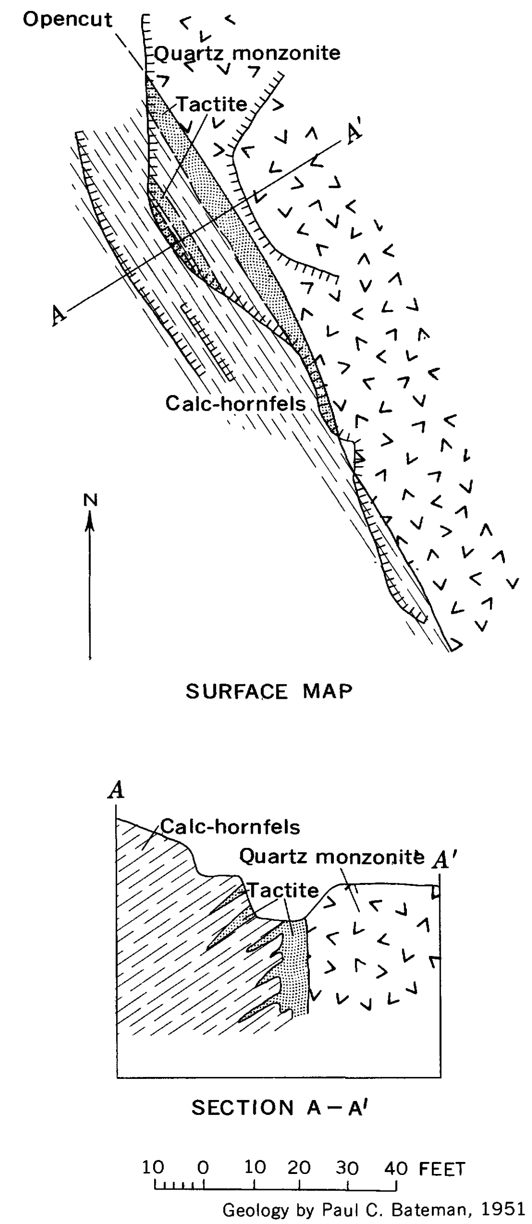


FIGURE 62.—Geologic sketch map and section of the Munsinger prospect, sec. 17, T. 8 S., R. 32 E.

Marble tungsten mine and North ore body of the Pine Creek mine is volumetrically almost as great as the apophyses and was surely not all derived from the magma that crystallized in these apophyses. Intrusive contacts at the Marble tungsten mine illustrate the probable difference between contacts that were frozen and contacts that were fractured at the time of tactite formation. There, apophyses of quartz monzonite penetrate complexly a small mass of coarsely crystalline white marble. These contacts have an aggregate length at the surface of at least 2,000 feet, yet tactite is confined to a 500-foot span. At other places along intrusive contacts the only metasomatic effect in the adjacent marble is a local selvage of wollastonite a few inches thick. The small amount of silica needed to form the wollastonite selvage could have been supplied by the adjacent granitic magma. In contrast, solutions from which the tactite was formed probably accumulated and moved laterally along fractured parts of the contact to its present position.

Either regional forces or stresses arising from the emplacement and consolidation of intrusive masses could cause fractures to form along intrusive contacts at the time of tactite formation. That fractures formed locally along some intrusive contacts during the general period of mineralization is shown by the localization of quartz veins and silicified zones in fractures. The evidence that quartz veins and silicified zones followed fractures can be viewed in many outcrops, because the introduction of quartz, though partly a replacement process, has not completely destroyed the fabric of the rock. On the other hand, all evidence of earlier structures can be destroyed during the formation of tactite. It seems especially possible that in ore bodies where fracturing has guided silicification earlier similar fractures were important in the localization of the tactite ore bodies. This possibility is particularly pentinent to the Main and South ore bodies of the Pine Creek mine, where silicification is prevalent.

Although fractures provided channelways for the movement of ore-forming solutions, proved examples of fractures that localized ore bodies are few. An excellent example is the Pinnacle ore body of the Pine Creek mine (fig. 55). This ore body extends outward into marble from the Main ore body, which lies along the intrusive contact. Although the outcrop of the ore body is too steep for careful examination of the wall rocks for evidenec of offset strata, localization along a fracture seems the only reasonable explanation for the shape and orientation of the ore body.

Fractures that restricted the flow of mineralizing solutions are found in the Hanging Valley and Western mines. In the Hanging Valley mine the ore is in vertically elongate shoots, most of which are bounded by steep east-trending beds and by north-trending fractures (fig. 63). In plan, the shoots are only a few feet on a side; one was followed upward 40 feet above the upper adit, and another downward 45 feet beneath the same level. Furthermore, ore intersected in diamond-drill borings as deep as 90 feet beneath the upper adit seems to represent extensions of the small prisms of ore.

In the Western tungsten mine much of the scheelite lies in fracture planes. Nevertheless, scheelite-bearing zones locally terminate abruptly against similar frac-

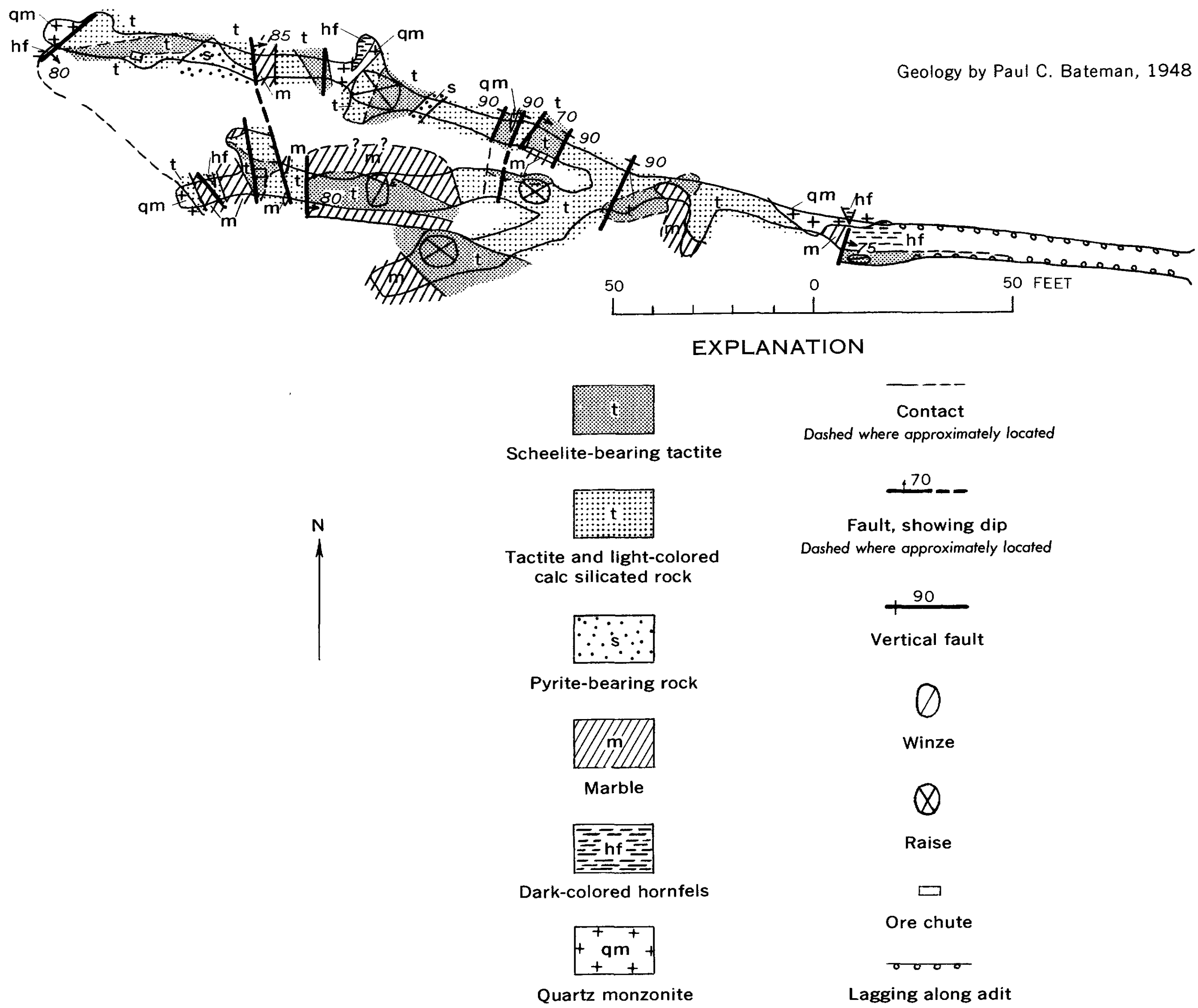


FIGURE 63.—Geologic map of the upper adit in the Hanging Valley mine.

tures. It seems apparent that some fractures provided openings for the movement of mineralizing solutions, whereas others acted as seals.

RELATION OF THE TUNGSTEN DEPOSITS TO THE GRANITIC ROCKS

The controls discussed thus far are of the kind that can influence the position, size, and shape of a single tungsten deposit, or even of several closely associated deposits, but they do not provide a satisfactory explanation of either the gross distribution of deposits within the Bishop district or of the existence of the district. The distribution of individual kinds of granitic rocks seems to have considerable bearing on the distribution of deposits within the district and may possibly have some bearing on the localization of the district. No satisfactory hypothesis can be proposed to explain the existence of the tungsten district until detailed information is obtained over a wider area than has been investigated.

The tungsten deposits appear to be preferentially associated with certain of the intrusive rocks. Of the 53 tungsten mines and prospects known in the district (table 18), 26 are in contact with the Tungsten Hills quartz monzonite, 8 are in contact with alaskite similar to the Cathedral Peak granite, 12 are in small metamorphic inclusions in mafic rocks that range from quartz diorite to hornblende gabbro, 2 are in contact with more than one intrusive and are therefore of uncertain association, and only 5 are in contact with other intrusive rocks. Furthermore, of the 21 deposits that have yielded more than 100 units of WO_3, 8 are in contact with the Tungsten Hills quartz monzonite, 6 are in contact with alaskite similar to the Cathedral Peak granite, 6 are in inclusions in quartz diorite and hornblende gabbro, and 1 is in contact with more than one intrusive and is of uncertain association.

Tungsten deposits commonly are associated with and assumed to be genetically related to granitic rocks in the range of granodiorite to granite (Buddington, 1933, p. 371). The distribution of tungsten in different igneous rocks and rock minerals bears upon the problem of the relation of ore bodies to igneous rocks, but the data available seem contradictory. Sandell (1946) concluded, on the basis of analyses of granitic rocks, that in general the amount of tungsten in igneous rocks containing 60 to 80 percent silica increases with the silica content. He found that the average silicic igneous rock contains 1.5 ppm (parts per million) of tungsten and that the range is from about 1.0 to 2.6 ppm. Jeffery (1959) arrived at 1.4 ppm for the abundance of tungsten in granitic rocks of Uganda, a figure in good agreement with Sandell's, but he found several times as much tungsten in amphibolite, dolerite, and the alkalic volcanic rocks of Uganda. He also found that the tungsten in the granite rocks is contained chiefly in the accessory minerals and biotite, and concluded that tungsten does not readily replace silicon in silicate structures. Obviously, more analyses for tungsten in igneous rocks and rock minerals are needed.

A figure that can be readily grasped is the amount of tungsten contained in a cubic mile of granitic rock having a specific gravity of 2.65 and a tungsten content of only 1 ppm. This amount is about 24 million pounds of tungsten or about 1.5 million units of WO_3, approximately the yield of the Bishop district to 1958.

The occurrence of so many tungsten deposits in metamorphic inclusions in quartz diorite and hornblende gabbro is decidedly anomalous, especially in view of the small total outcrop area of quartz diorite and hornblende gabbro. Masses of quartz diorite and hornblende gabbro are small as compared with the more silicic intrusives, and most of them are inclusions in or septa between more silicic intrusives. Tactite ore bodies within hornblende gabbro or quartz diorite are never far from a silicic intrusive. The only metamorphic inclusions known to contain tungsten ore bodies in the hornblende gabbro east of Shreves Camp (formerly Andrews Camp, as shown on the Mount Goddard quadrangle) on the South Fork of Bishop Creek are adjacent to Tungsten Hills quartz monzonite. The distribution of tungsten deposits in inclusions in quartz diorite and hornblende gabbro in the Deep Canyon (Tungsten City) area of the Tungsten Hills, on the other hand, is haphazard, and may be related chiefly to the inclusions that contained clean marble. The outcrop areas of quartz diorite and hornblende gabbro in the Deep Canyon area are quite small, and they may be underlain at a relatively shallow depth by quartz monozonite.

Almost all of the tactite-bearing inclusions in quartz diorite and hornblende gabbro are bordered by a zone in which the igneous rock has been epidotized irregularly for a thickness of a few inches to 20 feet or more. In places, the igneous rock adjacent to fractures is altered to epidote for distances of 50 feet or more from a tactite inclusion. Generally, rock close to an inclusion has been more or less completely altered to epidote, and the intensity of alteration diminishes with distance from the inclusion. In most places, the original texture of the epidotized rock is preserved well enough to show that it is identical with that of adjacent unaltered quartz diorite or hornblende gabbro. Thus the epidotization must have taken place after crystallization of the mafic intrusive rock adjacent to the inclusion. Because epidote contains a large amount of calcium, it would form more readily in quartz diorite and hornblende gabbro than in the granitic rocks. Any extra calcium needed to form epidote may have been expelled from the metamorphic inclusions during alteration to tactite. Similar zones of epidote and clinozoisite rock were attributed by Joplin (1935a), Gindy (1953), and Watters (1958), to reciprocal exchange of substances between an intrusive and calcium-rich sedimentary or metamorphic rock.

All tungsten-bearing metamorphic inclusions in quartz diorite or hornblende gabbro have as the closest granitic rock either the Tungsten Hills quartz monzonite or alaskite similar to the Cathedral Park granite, the same intrusives with which almost all other tactite ore bodies are associated. This presumed genetic relation is neither proved nor disproved by the solid shell of mafic rock which was epidotized when the tactite was formed, for still fluid parts of the mafic rock could have been the source for the tactite-forming solutions. The field relations strongly suggest but do not prove that tactite-bearing inclusions in quartz diorite and hornblende gabbro were formed from solutions derived from silicic magma.

If the deposits in inclusions are reassigned genetically to the nearest silicic intrusive mass, the association of the 53 deposits in the district would be retabulated as follows: 33 with the Tungsten Hills quartz monzonite, 11 with the alaskite similar to the Cathedral Peak granite, 6 with other silicic intrusives, and 3 uncertain. All 6 commercial tungsten deposits in quartz diorite or hornblende gabbro are associated with Tungsten Hills quartz monzonite, and their reassignment to that intrusive rock would give the following tabulation for the 21 commercial deposits in the district: 14 with the Tungsten Hills quartz monzonite, 6 with the alaskite similar to the Cathedral Peak granite, and 1 uncertain.

Before a close genetic relation between the tungsten deposits and the Tungsten Hills quartz monzonite and the alaskite can be accepted, the distribution of calcium-rich host rocks must be appraised with relation to the various intrusives. Several granitic intrusives are nowhere in contact with calcium-rich rocks; these intrusives include the granodiorites of McMurry Meadows and Cartridge Pass, the Inconsolable granodiorite, and several masses of quartz monzonite similar to the Cathedral Peak granite and of finer grained quartz monzonite. On the other hand, many intrusive rocks not associated with any significant tungsten deposits are in contact with calcium-rich rocks at one or more places; they include the Tinemaha granodiorite, the Round Valley Peak granodiorite, the Wheeler Crest quartz monzonite, the Lamarck granodiorite, and the granodiorite of Coyote Flat. The aggregate length of linear contact of calcium-rich rocks with the Tungsten Hills quartz monzonite and the alaskite similar to the Cathedral Peak granite is greater than that with all other granitic rocks together, but the number of tungsten deposits associated with the intrusives is, nevertheless, disproportionately large.

On the whole, it seems likely that the relation of the tungsten deposits to the Tungsten Hills quartz monzonite and alaskite similar to the Cathedral Peak granite is genetic. These two intrusives are among the most silicic in the district, and the ones which experimental data indicate would crystallize at the lowest temperatures. The alaskite in particular would crystallize close to the lowest temperatures expectable for granitic rocks. A possible explanation of the preferential association of tungsten deposits with these two intrusives is that tungsten was progressively enriched in the magma in the course of differentiation. No tungsten deposits have been found associated with the quartz monzonite similar to the Cathedral Peak granite, but marble is absent along the margins of this intrusive.

Kerr (1946, p. 19) has suggested that tungsten concentrated in the core of a partially crystallized intrusive mass "may follow the aplitic and pegmatitic juices along the fan-like fractures that radiate from the core of the typical massif." He suggested that the ore bodies in the Pine Creek pendant were formed by such a mechanism (Kerr, 1956, p. 18–19, and figs. 7, 8, and 9). However, the mechanism proposed cannot apply to the ore bodies in the Pine Creek pendant. Aplite and pegmatite dikes marginal to the Mono Recesses mass of quartz monzonite similar to the Cathedral Peak granite intersect the marble on the west side of the pendant, but are younger than the contact rocks, for the walls of the dikes were spread after the formation of the tactite and associated silicified rocks. A few grains of molybdenite found locally in the dikes were very likely picked up by the dike magma. Dikes cutting tactite are well displayed at the Brownstone mine (fig. 37*B*).

On the east side of the pendant a swarm of mafic dikes extends eastward from the pendant, but no ore bodies have been found along the east side of the pendant in the vicinity of the dikes. It is not likely that these dikes could have been carriers of tungsten ores.

SUMMARY OF GENESIS

During the cooling and crystallization of granitic magma, silicon, aluminum, iron, manganese, tungsten, titanium, and other metallic elements were expelled and became available to react with calcium-rich wall rocks and to form contact metasomatic deposits. Magmas of different compositions contained different amounts of these constituents; in the Bishop district the tungsten deposits are preferentially associated with the more silicic intrusives, which crystallized at the lowest temperatures. Tactite formed adjacent to granodiorite generally contains less scheelite than tactite adjacent to quartz monzonite or alaskite. Shortly after the emplacement of an intrusive mass, maximum temperatures were reached in the wall rocks and they brought about thermal metamorphism. The recrystallization generally was essentially isochemical, although such fugitive constituents as carbon dioxide were lost. Only small amounts of substance were discharged from the magma at this time, and it was later, when temperatures were lower and crystallization of the magma more advanced and after a crust of granitic rock had formed about the magma, that substantial amounts of material were expelled.

The form in which the substances given off by the magma traveled through the cooling intrusive is unknown; both pneumatolytic and hydrothermal processes have supporters. Barth (1952, p. 277) states, "The primary magmatic gases are acid and show in consequence high reactivity. If the contact rock is basic, especially limestone, the acid gases will react effectively with it. Limestone, therefore, acts as a filter capturing the escaping gases, with the formation of a great variety of reaction minerals; the corresponding rocks are known as skarns. Reaction rocks at the contact of limestone and composed of lime silicates form mainly garnet and pyroxene, often accompanied by fluorite and phlogopite, and with sulfides of iron, zinc, lead, or copper; in other occurrences magnetite is formed." Barth cites the following equation as an example of the kinds of reactions that can take place when gases rich in iron fluoride meet with limestone:

$$2FeF_3+3CaCO_3=Fe_2O_3 \text{ (hematite)}+3CaF_2 \text{ (fluorite)}+3Co_2$$

or, if silica is present in the limestone:

$$2FeF_3+3SiO_2+6CaCO_3=Ca_3Fe_2Si_3O_{12} \text{ (andradite)}+3CaF_2 \text{ (fluorite)}+6CO_2.$$

Geijer (1958, p. 211) has suggested the following equation to show how both iron and silicon could be transported as fluorides and precipitated by limestone to form andradite and fluorite:

$$2FeF_3+3SiF_4+12CaCO_3=Ca_3Fe_2Si_3O_{12}+9CaF_2+12CO_2.$$

On the other hand Turner (1948, p. 125) states, "In some cases metasomatism has been pneumatolytic, the iron having been introduced in some volatile form such as FeF_3 or $FeCl_3$; but the more generally hydrothermal reactions following upon the main phase of thermal metamorphism, and governed by somewhat lower temperatures, seem to have been responsible for the development of skarns."

The effectiveness of these two processes in the formation of the tactite deposits in the Bishop district is difficult to evaluate because so few critical data are available. One of the chief difficulties is that the processes themselves probably merge at the pressures and temperatures that pertained at the time of formation of tactite. The rock pressure was above 217.7 atmospheres and the temperature, except in the last stages, probably was above 374°C, the critical pressure and temperature of water. Fluorite is present in small amounts in the Pine Creek, Adamson, and Brownstone mines of the Pine Creek pendant, and phlogopite is present in the Round Valley mine and in the unproductive Stevens ore body of the Tungstar mine, but minerals considered indicative of pneumatolytic action appear to be lacking in most other deposits. Although fluorine may have been present in the mineralizing solutions that formed all the deposits, and escaped as a fugitive constituent from all but a few, such an assumption is unwarranted without some supporting evidence. On the other hand, few hydrous minerals are present in tactite to indicate that an appreciable amount of water was present in the mineralizing solutions. However, hydrous minerals are hardly to be expected in view of the high temperatures that existed at the time of formation of tactite.

Both pneumatolytic and hydrothermal processes (if they are to be distinguished) probably contributed to the formation of tactite, pneumatolytic processes being dominant in the early stages and hydrothermal processes in the later stages. The garnet-pyroxene-scheelite tactite at the Pine Creek mine (which also contains fluorite) may therefore be considered pneumatolytic and the later quartz and sulfides hydrothermal. Tactite containing abundant quartz and epidote, such as that at the Tungstar, Schober, and Lakeview mines, may also be considered hydrothermal.

The mineralizing substances expelled from a granitic magma collected along the intrusive contacts and migrated generally upward, especially along fractured segments of these contacts. Fractures along intrusive contacts could have resulted from regional stresses or from stresses related to the intrusion, crystallization, or cooling of the magma. Fracturing generally was more intense at irregularities in the intrusive contact, as would be expected. More irregularities are present along discordant than along concordant intrusive contacts because along discordant contacts the intrusive magma alternately followed and broke across the bedding in the metamorphic rocks; furthermore, at discordant contacts apophyses and sills of granitic rock penetrated into the metamorphic rock along beds. Reentrants in the intrusive contact provided channels for the accumulation and movement of mineralizing solutions, presumably because of the more pervasive fractures there. Benches or protrusions of granitic rock acted as traps and forced accumulating solutions to spread into adjacent metamorphic rock.

Where the solutions came into contact with calcium-rich rock in an inclusion or in the wall rock, they reacted with it to form tactite. At frozen contacts or at contacts where fracturing was slight, only thin selvages of tactite were formed. Thick masses were localized at irregularities in the intrusive contact where pervasive fracturing is believed to have taken place, and along fractures and favorable beds in the metamorphic rock. In the formation of tactite the greatest addition of substances, including tungsten, was to clean limestone or marble, because all the clean rock entered into reaction with the mineralizing solutions. Dirty limestone had already been altered to silicated marble or to calc-hornfels as a result of isochemical thermal metamorphism, and the relatively inert and insoluble silicate minerals reacted very slowly with the introduced substances. Thus a bed of silicated marble that consisted half of calcite and half of silicate minerals reacted with not much more than half as much of the introduced substances as an adjacent bed of clean marble. If the silicate minerals that were formed without addition of material lay along beds, a layered tactite could result.

In tactite that is formed from clean marble, half or more of the substances calculated as oxides (about 25 percent in terms of the metallic elements) in the rock were introduced. Carbon dioxide and calcium were the principal constituents that were released as a result of the formation of tactite. Some displaced calcium moved into the adjacent intrusive rock, and in calcic rock combined with plagioclase and hornblende to form epidote or clinozoisite.

The first minerals to form in clean marble were iron-rich pyroxene of the diopside-hedenbergite series and garnet of the grossularite-andradite series, together with fine-grained scheelite. The most extensive ore bodies were formed at this time. The proportion of garnet and pyroxene at each deposit was determined by the relative abundance of the principal constituents, chiefly silica, which favors formation of pyroxene, and calcium, which favors formation of garnet. The composition of the pyroxene probably was determined chiefly by the amount of magnesium in the marble, but some magnesium also may have been introduced.

As temperatures fell and as the calcium in the marble adjacent to the intrusive contact was used up, quartz was deposited in place of silicate minerals. Where fractures were formed between and within newly formed tactite and intrusive rock, quartz was deposited along the fractures and it replaced the silicates in the adjacent rock. In tactite, pyroxene usually survived in silicified rock after garnet and scheelite had disappeared, probably because pyroxene contains more silica. Tungsten was probably highly soluble in the silica-rich solutions because it was so readily removed from the tactite during its silicification, and because in a few places quartz locally contains small but rich masses of euhedral to subhedral scheelite crystals. Formation of the scheelite in these small masses probably was caused by the local presence of a precipitating agent, such as calcium. In the Bishop district, tactite rich in quartz usually contains little scheelite.

At temperatures a little below those at which pyroxene-garnet tactite was formed, quartz and epidote were deposited, in places accompanied by iron sulfides. Quartz-epidote rock at the Lakeview mine, quartz-epidote-pyrrhotite rock at the Schober mine, and feldspar-epidote-pyrite rock at the Tungstar mine contained coarse-grained scheelite and constituted the richest, though not the largest, ore bodies in the district. Whether the ore at the Tungstar and Schober mines was formed from earlier garnet-pyroxene tactite or directly from marble is not known; the scheelite-bearing quartz-epidote rock at the Lakeview mine was undoubtedly formed from earlier garnet-pyroxene rock.

At still lower temperatures, sulfides of valuable metals locally accompanied quartz. These sulfides include molybdenite, chalcopyrite, bornite, sphalerite, and galena. In places, the galena contains small inclusions of sulpho-salts. Presumably these sufides were deposited at different temperatures and probably represent a considerable temperature range.

CENOZOIC GEOLOGY

FORMATIONS OF CENOZOIC AGE

VOLCANIC ROCKS

BASALT DIKES, NECKS, AND DISSECTED FLOWS

Eroded dikes, necks, and dissected flows of early Pleistocene or late Tertiary age are present west and southwest of Bishop, and scattered outcrops of eroded basalt have been found as far west as the Glacier Divide and as far south as Big Pine Creek. Much of the basalt is massive and dark gray on fresh surfaces. The mineralogy is simple—olivine, augite, and, rarely, plagioclase constitute phenocrysts, and augite, plagioclase, and magnetite, with or without interstitial glass, constitute a felty groundmass. Olivine crystals as much as a millimeter across generally are readily visible in hand specimen, whereas smaller augite and plagioclase phenocrysts may be difficult to identify megascopically. The relative abundance and mode of occurrence of the minerals differ from mass to mass.

Deep dissection of some flows and exposure of basaltic dikes and necks, which requires the removal of large volumes of rock after the basalt was emplaced, show that the basalt is much older than well-preserved cinder cones and flows south of Big Pine. In the west side of the Tungsten Hills, basalt dikes and necks crop out in low ridges on an old erosion surface. Apparently the ridges are held up by the basalt, and at least the last stages of erosion of the old surface took place after the basalt was emplaced. Southwest of Bishop, west and southwest of the Chipmunk mine, remnants of flows that are capped by till of Sherwin or older glacial stages have been cut through by Coyote Creek to a depth of as much as 800 feet below the basalt. These flows apparently moved down an ancient surface that sloped toward Owens Valley. The only feeders upslope from the flows that were mapped crop out in the hilly country east of Coyote Flat, but it is possible that closer feeders are concealed beneath glacial till, or even beneath the upper ends of some of the flow remnants.

Patches of basalt exposed along the North Fork of Bishop Creek, on the slope north of North Lake, probably are remnants of once extensive flows; older glacial till on the north side of Bishop Creek contains abundant boulders of basalt that must have come from this area because basalt has not been found elsewhere in the Bishop Creek drainage basin. Inasmuch as adjacent moraines assigned to the Tahoe glacial stage contain no basalt boulders, the dissection of the basalt at North Lake must have been completed at an early time.

Although the basalt is deeply eroded, most of the masses appear to have been emplaced after the inception of the structural movements by which Owens Valley was downwarped relative to the bordering ranges.

Structural movements earlier than basalt are indicated by the localization of some feeders along faults that formed in connection with the subsidence of Owens Valley and by the lava flows above the Chipmunk mine, which moved downslope towards Owens Valley.

An unusually youthful appearing flow in the SE¼ sec. 26, T. 7 S., R. 32 E., extends east from a small cinder cone at its head. This flow is about 2,000 feet long and 200 feet wide. In spite of the fresh appearance of both cone and flow, the surface north of the flow has been steepened and dissected since the flow erupted; if a flow broke out from the same crater today, it would follow the same path on leaving the crater but would turn gradually to the north into a north-trending canyon.

An interesting compositional variant among these basalt masses is a holocrystalline trachybasalt in the west side of the Tungsten Hills, in the south half of sec. 3, T. 7 S., R. 31 E. This rock consists of plagioclase, augite, olivine, hypersthene, potassium feldspar, and magnetite. The most abundant mineral is plagioclase in tabular or elongate crystals that appear in thin section to average about half a millimeter long; rounded grains of augite and olivine are next most abundant. A few larger subhedral grains of plagioclase a millimeter across also are present. Potassium feldspar interstitial to these grains constitutes about 15 percent of the rock. Hamilton and Neuerburg (1956) attribute similar rocks in the Huntington Lake area of the Sierra Nevada to the assimilation of large quantities of wall rock in a basaltic magma.

Melting and assimilation of granitic rock by basaltic magma are shown in a basalt neck in the NE¼ sec. 17, T. 7 S., R. 32 E., on the north side of the hard-surface road along Bishop Creek. This locality is well known through the descriptions of Knopf (1918, p. 74–75; 1938). Here, conspicuous columnar joints are present in partially re-fused quartz monzonite adjacent to fingers of basalt. The fact that the basalt locally contains numerous ragged xenocrysts, chiefly of quartz, indicates that quartz survived longer than the feldspar or biotite. In thin section the quartz xenocrysts are surrounded and embayed by brown glass that contains wedge-shaped crystals of tridymite.

Glass is present along all contacts between minerals in granitic fragments included in the basalt. Adjacent to basalt the glass is brownish, and away from the basalt it is progressively lighter colored; at a distance of 1 to 2 mm into the quartz monzonite it is almost colorless. The colorless glass is largely re-fused quartz monzonite, whereas colored glass is derived, at least in part, from basalt. Vesicles in the brown glass commonly are lined with a green isotropic or near-isotropic substance (chlorite?). No biotite remains in the rock fragments, and the margins of feldspar grains adjacent to glass have a frosty appearance that is caused by the presence of extremely fine vermicular glass.

BISHOP TUFF [2]

The Bishop tuff, a name here adopted by the U.S. Geological Survey, crops out in the Volcanic Tableland in the north-central part of the mapped area and extends 60 miles to the north. It also underlies the alluvium in the northern part of Owens Valley where it has been intersected at increasing depths southward in borings made for water. Discontinuous segments of a thin basal pumice layer also crop out locally along the west side of the White Mountains in older dissected alluvial deposits (fig. 64).

The Bishop tuff was named and described by Gilbert (1938), and his report has been drawn on freely, but with no attempt to repeat his excellent petrographic descriptions. He carefully studied a section in the Owens River Gorge at the north boundary of the Mount Tom quadrangle, in sec. 22, T. 5 S., R. 31 E., which is considered the type locality. This section includes about 500 feet of tuff but does not extend to the base of the formation. At the base of the formation is a thin layer of airborne pumice, which is well exposed along the south and east sides of the Volcanic Tableland.

Gilbert concluded that the tuff (exclusive of the basal pumice layer) is a "welded tuff", and that it originated as nuées ardentes—flows of intensely hot, discrete but viscous, glassy fragments that are lubricated by gases emitted from the fragments. Heat contained within the mass itself caused the fragments to become welded after the mass came to rest.

The basal pumice layer was included in an earlier report (Bateman, 1956) with the alluvial deposits of Owens Valley; however, it seems more properly considered a part of the Bishop tuff. This windblown pumice was deposited a short time before the outpouring of the nuées ardentes. Quite likely it is the product of early explosive activity in the same vents that a short time later emitted the nuées ardentes.

Although the main part of the tuff described by Gilbert is treated here as a single member, it includes several layers of different color and texture. The member is, however, subdivided into two facies that are gradational to one another, an agglutinated facies, and a marginal unconsolidated facies.

BASAL PUMICE LAYER

The pumice layer at the base of the Bishop tuff is best exposed along the south and east sides of the Volcanic Tableland (fig. 65). Along the east side similar pumice crops out discontinuously at the base of the

[2] In the explanation of plate 3, the Bishop agglutinated tuff is erroneously shown as green; it should be light tan as shown in the northwestern part of the map itself.

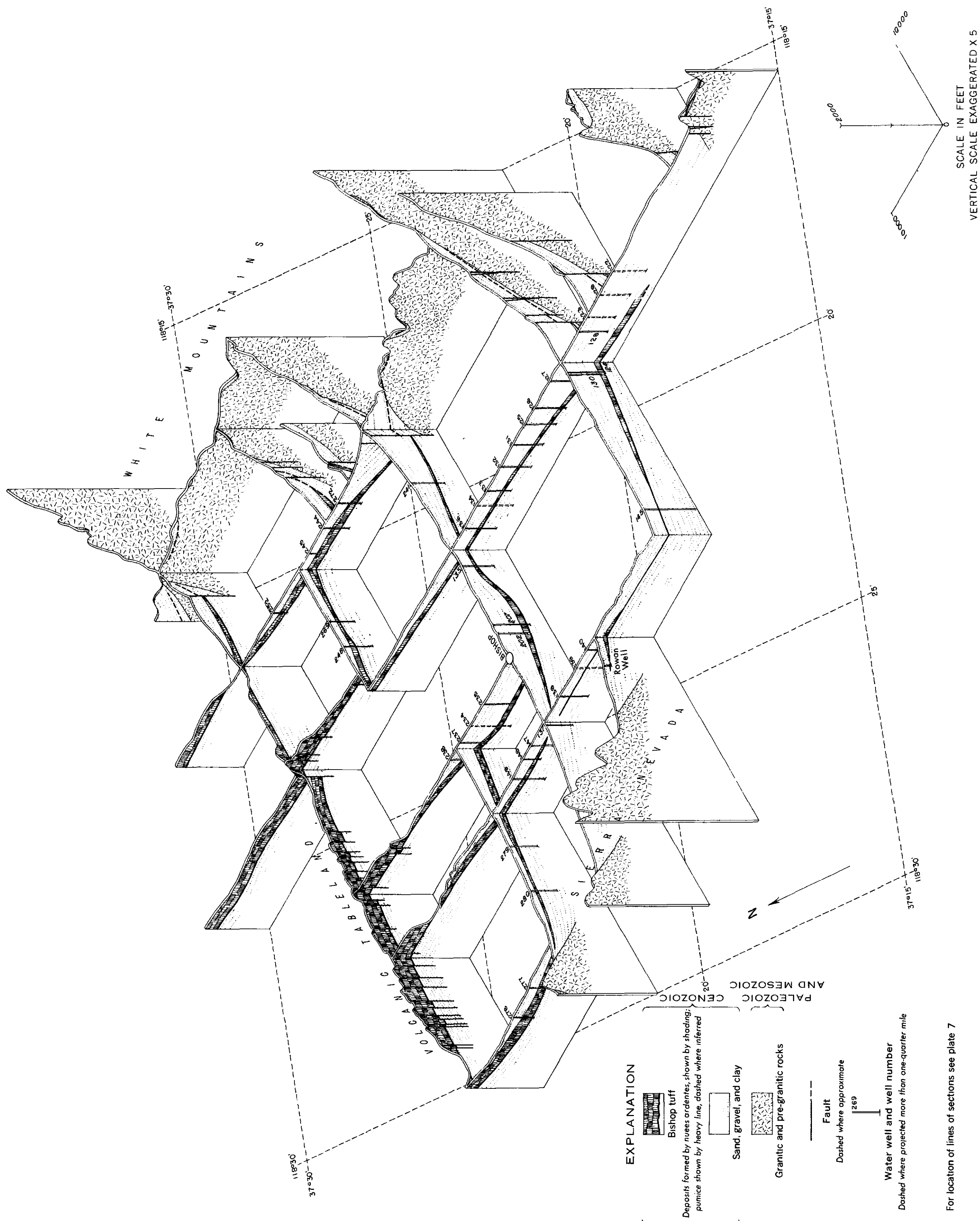
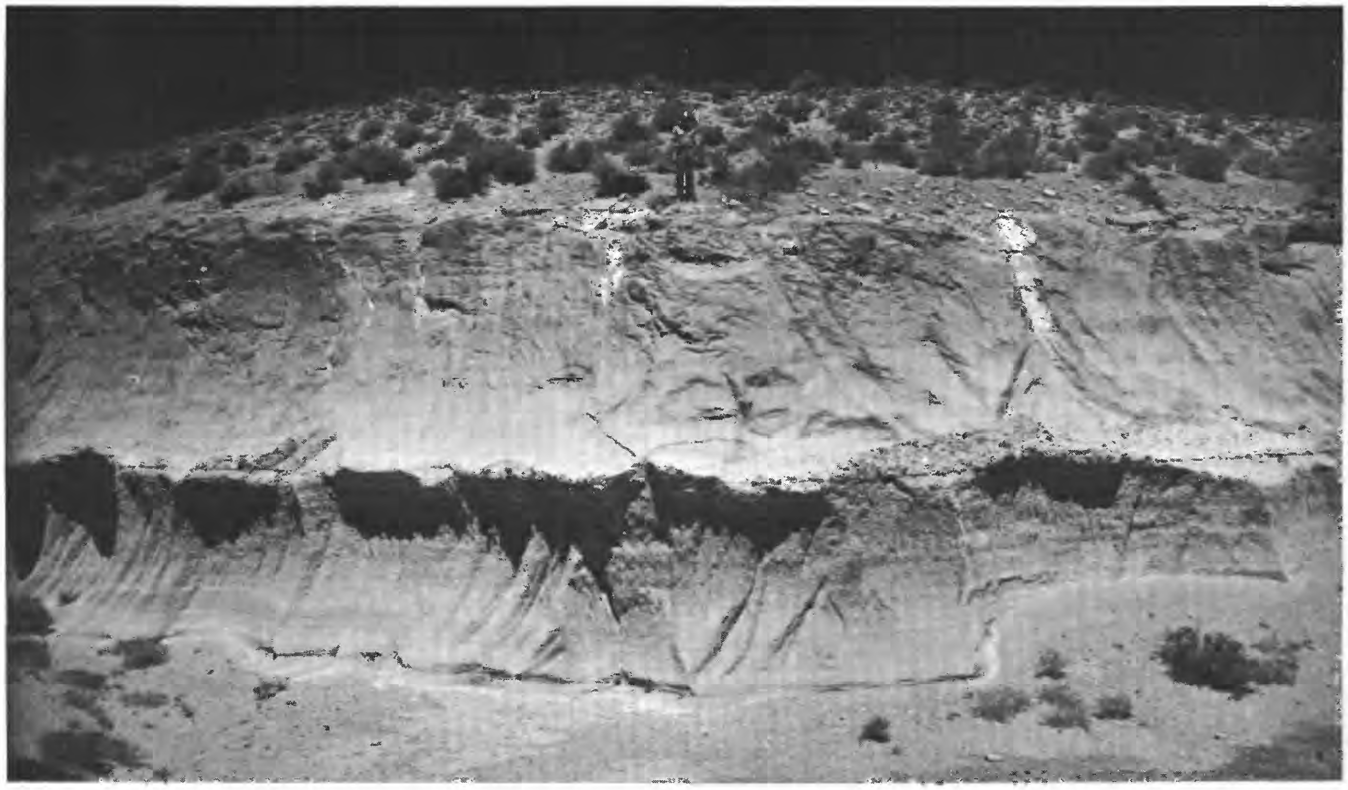


FIGURE 64.—Fence diagram showing the distribution of the Bishop tuff in the Bishop quadrangle.



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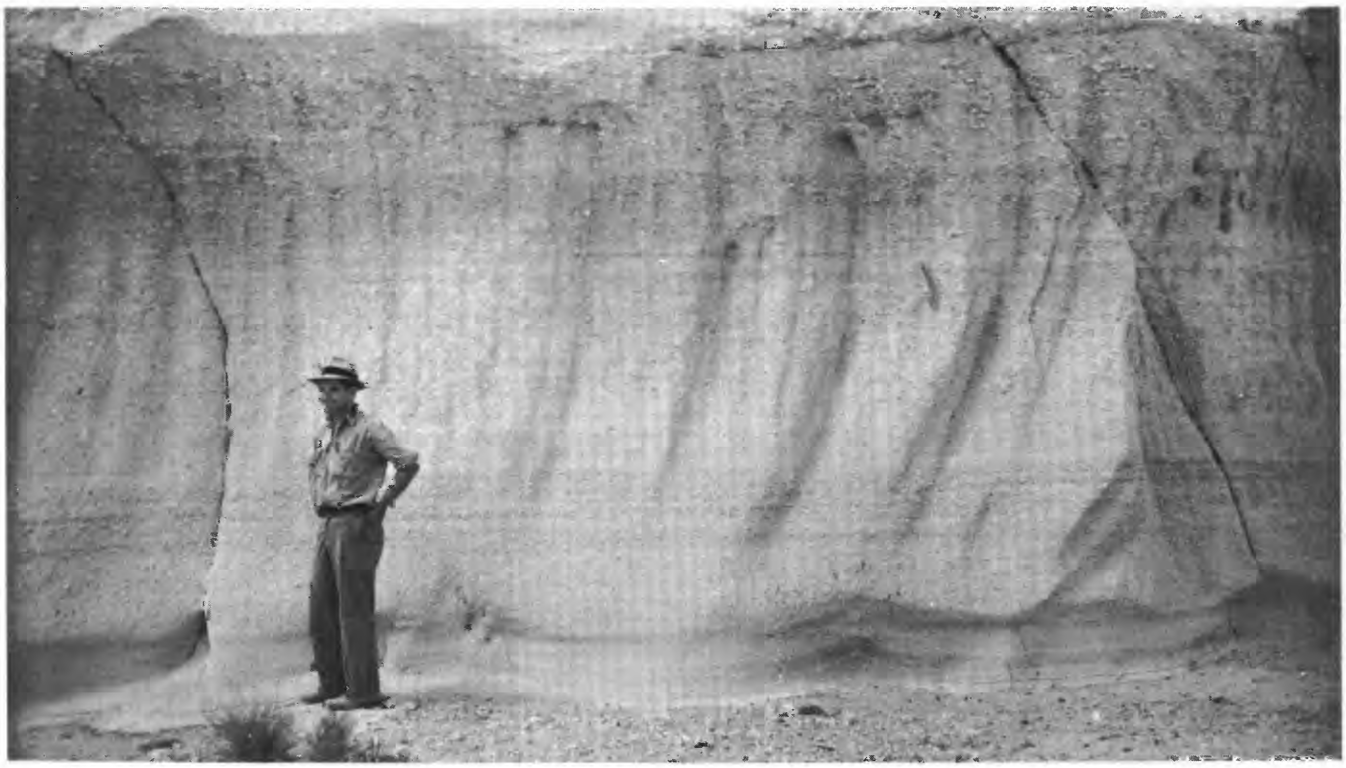

B

FIGURE 65.—Views of Bishop tuff in pit of Insulating Aggregates Co. in east side of Volcanic Tableland, sec. 32, T. 5 S., R. 33 E. *A*, Lower half of face is basal pumice layer; upper half is unconsolidated tuff. Mounds of rounded pumice fragments are present in base of unconsolidated tuff. *B*, Close view of basal pumice layer to show overall upward increase in size of fragments, three-ply structure, and laminations.

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Bishop tuff as far north as Blind Spring Hill, 30 miles north of Bishop. The pumice also extends below the surface south into Owens Valley where borings intersect it at increasing depths southward. The most southerly boring is in the north-central part of sec. 23, T. 7 S., R. 33 E., where the base is at an altitude of 3,300 feet—more than 700 feet beneath the floor of the valley (pl. 7; fig. 64). None of the borings farther south was deep enough to intersect the pumice layer. Discontinuous masses of correlative pumice are intercalated in alluvial detritus of the White Mountains. Most of these exposures are in upfaulted or tilted blocks. Outcrops along Black Canyon and farther south in the Bishop quadrangle indicate that the pumice was deposited south of Bishop for a distance of at least 8 miles. Undoubtedly the pumice in the south and east margins of the Volcanic Tableland extends under the tuff some distance to the northwest, but it is lacking where the Bishop tuff is exposed north of the mapped area along the west side of Rock Creek and in Tunnel 1 of the Los Angeles Department of Water and Power. The pumice crops out, however, in a road cut along U.S. Highway 395 in sec. 34, T. 4 S., R. 30 E., and at the Delta Placer, 1 mile south of Lake Crowley.

A continuous sheet of pumice exists beneath Owens Valley and the south and east sides of the Volcanic Tableland except where locally removed by erosion. Data from borings indicate that this sheet is cuspate along the margins of Owens Valley, being indented at the larger alluvial fans. Most outcrops of pumice along the White Mountain front were undoubtedly part of this sheet before faulting, but a few in the northeast corner of the Bishop quadrangle, 2,000 feet above the valley floor, may have been deposited in separate basins.

The pumice consists almost entirely of angular, unrounded fragments that range from sand size to 2 inches in maximum dimension. The pumice fragments are unaltered, firm, and perfectly white. They contain conspicuous phenocrysts of quartz and sanidine, which also occur loosely throughout the layer. Angular fragments of foreign material are sparsely and sporadically distributed through the layer. Although the pumice fragments are tightly packed, they are not cemented.

The most conspicuous feature of the pumice layer in exposures along the margins of the Volcanic Tableland and in most outcrops along the White Mountain front is a gradational increase upward in the average size of the pumice fragments. In most outcrops, the pumice layer also is three-ply; the size gradations of fragments within each division are superimposed on the overall gradation. In pits of the Insulating Aggregates Company, along the east side of the Volcanic Tableland, the lower division is less than 2 feet thick, the middle division 3 feet thick, and the upper division 9 feet thick (fig. 65). The layer rests conformably on sand and gravel, and the three divisions meet along sharp planes. In the lower division, fine pumiceous sand at the base grades upward to pumice in which the fragments average about one-quarter inch across. In the middle division, pumice at the base is composed of tiny fragments about one-eighth inch across, but fragments one-half inch across become increasingly abundant upward. In the upper division the pumice fragments at the base average about one-quarter inch across, and in the upper 3 feet pumice fragments having a maximum dimension of 2 inches are abundant. At the top, an inch or two of fine well-bedded ash and dust is present locally. The pumice has a rather distinct laminated appearance, which in part results from the presence of the three divisions within the layer and in part from thin layers throughout the pumice, some only one pumice fragment thick, composed of fragments of about the same size, which alternate with other thin layers composed of fragments of a different size. The laminated appearance is heightened by the orientation of the larger pumice fragments with their maximum dimensions parallel with the laminations, and by the presence of carbonaceous streaks in the upper 4 feet of the layer parallel with the laminations. In a few places, faint cross-laminations can be identified.

The internal structure of most pumice outcrops in the White Mountains is similar to that in the margins of the Volcanic Tableland. Even a small outcrop on the northeast corner of the Bishop quadrangle in the SE¼ sec. 24, T. 5 S., R. 33 E., at an altitude of 6,500 feet, which was trenched to its base and studied in detail, is structurally almost identical to outcrop in the margins of the Volcanic Tableland. In this outcrop the pumice rests on detrital material composed of angular cobbles of basement rock in a clay matrix, and is overlain by slope wash. Pumice along Black Canyon in the southeast corner of the Bishop quadrangle interstratified in coarse fanglomerate is only about 6 feet thick, apparently because of erosion immediately following deposition. At an exposure in Black Canyon examined in detail, about 3 feet of pumice crops out and another 3 feet at the top is concealed by slope wash. At the base, the pumice is sand size, and the size increases upward through 7 inches to fragments about one-quarter inch across. Overlying this pumice is 2 inches of well-bedded pumiceous dust. About 2 feet of pumice exposed above the pumiceous dust grades from fragments one-eighth to one-quarter inch across at the bottom to fragments one-quarter to one-half inch across at the top.

Ordinarily, beds in which the size of the fragments increases upward are considered to be overturned, but regional relations leave no doubt that the pumice layer is right-side-up. The angularity of the fragments precludes transportation of the pumice in any way except through the air. Within this limitation two hypotheses merit consideration as providing an explanation for the inverse size gradation and related features.

The first hypothesis and the one that I favor is that more finely comminuted fragments were produced at the beginning of an explosive spasm than toward the end and that the ejected material fell on dry land. The three-ply structure could have resulted from slight variations in the activity of the vent. Wentworth and Macdonald (1953, p. 73 and fig. 38) have suggested such a mechanism as one of two possible explanations for upward increase in the size of basaltic fragments in the flanks of a steep-sided cinder cone. The alternative explanation offered by Wentworth and Macdonald (1953) for the inverse size gradation of the fragments in the cinder cone—progressive accumulation by rolling and sliding, on the premise that coarse material will roll on finer material, but not the reverse—could not apply to the flat-lying basal pumice layer of the Bishop tuff.

The second hypothesis, which I originally favored, is that pumice of random size blown into the air fell into a body of standing water (Bateman, 1953). All the pumice fragments would have floated for a brief period, but the smaller ones would have become saturated first, and progressively larger ones would have sunk with time. The three-ply structure would result from three explosions, each throwing out more and larger fragments than the preceding one. The fine dust locally present at the tops of the internal divisions would result from dust settling slowly from the air after each explosion.

To test this hypothesis, some of the pumice was dumped into a glass beaker that was about two-thirds full of water. Loose quartz and sanidine crystals sank immediately, and within a few minutes the smaller pumice particles began to settle. The pumice collection in the bottom of the beaker was rudely layered, and the fragments were graded in size from small at the bottom to larger at the top. After most of the pumice had settled, more pumice was dumped into the beaker and a second inversely graded layer resulted. It was several hours before the largest pumice fragments sank (fig. 66).

FIGURE 66.—Two inversely graded layers of pumice resulting from dumping two batches of pumice into a beaker partly full of water.

This second mechanism has the advantage over the first of not requiring special conditions at the source, but fails to explain why accidental fragments are larger and more abundant in the upper half of the pumice than in the lower half and why no concentration of accidental fragments is present in the base of the pumice. Rafting of the accidental fragments by the pumice may have had some effect on distribution, but is difficult to evaluate. Another objection to the lacustrine hypothesis is an apparent absence of diatoms, ostracods, and other lacustrine organisms in the silt and fine sand at the base of the pumice. Samples collected at several places from just beneath the pumice contained no organisms that could be seen with the binocular microscope.

PRINCIPAL TUFF MEMBER

The bulk of the Volcanic Tableland is composed of agglutinated tuff that varies considerably in color, specific gravity, and texture. The term "agglutinated" is here used in preference to "welded" because much of the tuff owes its coherence more to the growth of tiny crystals of tridymite or cristobalite across fragment boundaries than to actual "welding" through fusion. Lower, but not basal, parts of thicker sections of the tuff, nevertheless, are truly "welded."

Porphyritic pumice fragments that range in size from lapilli to blocks as much as 2 feet in maximum dimension, but average about 2 inches in maximum dimension, are set in a matrix of fine vitric tuff. Conspicuous crystals of quartz and sanidine are present both as phenocrysts in the pumice and as discrete crystals in the matrix. Less abundant are crystals of sodic plagioclase and biotite. Accidental fragments are sparse and generally small; most are less than 2 inches in diameter.

Two samples from different layers exposed in the west wall of the Owens River Gorge were analyzed chemically (table 21). The analyses show that the rock is rhyolitic and that the compositions of the two layers are nearly identical.

TABLE 21.—*Analyses of Bishop tuff*

[Analyses by rapid method (Shapiro and Brannock, 1956). Both samples are from the west side of the Owens River Gorge at the north edge of the Mount Tom quadrangle near the west quarter corner sec. 22, T. 5 S., R. 31 E. Sample 1 is from a white member which crops out from between 10 and 80 feet below the rim of the gorge. Sample 2 is from a pale grayish-purple member exposed in the lower slopes and bottom of the gorge. Analysts: Paul L. D. Elmore, Samuel D. Botts, and Marvin D. Mack]

	1 (*Lab. No. 152441*)	2 (*Lab. No. 152442*)
SiO_2	75.6	76.7
Al_2O_3	13.1	13.0
Fe_2O_3	.8	.5
FeO	.13	.26
MgO	.18	.13
CaO	.56	.52
Na_2O	3.8	3.7
K_2O	4.8	4.5
TiO_2	.11	.06
P_2O_5	.01	.00
MnO	.04	.03
H_2O	.35	.34
CO_2	.16	.08
	99.64	99.82

According to Gilbert (1938, p. 1835–1836) the tuff consists of several layers of different color, each of which represents a separate eruption. These layers generally are readily distinguishable in the sides of gorges and in cliff faces, but are difficult to follow across the surface of the Volcanic Tableland. Each layer changes laterally, and some layers grade to loose unconsolidated tuff in the marginal parts of the Volcanic Tableland, epecially in the southeast part where the tuff is probably most distant from its source. Although unconsolidated tuff grades to agglutinated tuff, unconsolidated tuff can be represented separately on the geologic maps (pls. 2, 3), largely because most of the exposed contacts of unconsolidated tuff are with the stratigraphically underlying pumice layer and with an overlying agglutinated layer rather than with laterally equivalent agglutinated tuff.

Except for the gross layering resulting from the separate extrusive impulses, no internal layering is apparent. In some thick sections, however, systematic changes that involve several layers are present. The most marked changes are downward increase in specific gravity, coherence, and hardness, a decrease in porosity, and progressive flattening of pumice fragments. In sections where these relations exist, separate extrusive impulses must have followed one another so closely that the whole mass cooled together and compacted in accordance with the weight of the overlying tuff. In other sections tuff layers of higher specific gravity overlie layers of lower specific gravity—a feature that indicates that the two layers did not cool together, although their cooling periods doubtless overlapped.

Joints are locally conspicuous in the gorges of Rock Creek and the Owens River, where joint columns are oriented in several directions. In places the columns radiate through a full circle. If they were formed normal to a cooling surface (or along the direction of maximum temperature gradient), the cores of radiating columns probably represent "hot spots" within the cooling tuff.

In the eastern part of the Volcanic Tableland the surface is dotted with rounded mounds as much as a hundred feet high. Most of the mounds are distributed haphazardly, but some are alined. East and southeast of Birchim Canyon the mounds give place to straight but discontinuous ridges that lie in a polygonal pattern. The two commonest directions of the ridges and alined mounds are N. 25° W. and N. 20° E., but trends in other directions are also present. Southeast of Birchim Canyon, ribs in a polygonal pattern stand up above the average surface and superficially resemble rock walls built between fields.

The mounds are erosional remnants, for layers pass through them nearly horizontally, parallel with layers in the underlying tuff. They consist chiefly of white or nearly white agglutinated tuff, whereas the tuff between the mounds is pink and unconsolidated. Exposures in road cuts through the sides of mounds show that the white tuff is derived from the pink tuff by alteration. In thin section, crystals of tridymite are abundant across boundaries between fragments in the white tuff, whereas very few such crystals are present in the pink tuff. The mounds undoubtedly mark avenues of escaping gases. Gilbert (1938, p. 1856–1858) observed variations in crystallinity, which he suggested could be explained by differences in the amount of gas that passed through the rock.

Conspicuous steeply dipping elongate vesicles in the rock ribs southeast of Birchim Canyon indicate that gases also escaped along the lines of these ridges. The northeast- and northwest-trending lines along which the ridges are oriented may mark conjugate shear planes that formed locally as a result of late movement in the tuff after partial consolidation. Where such localized avenues were present, it is expectable that they would have served as loci for escaping gases; where such planes were not formed, the gases would have escaped in a haphazard pattern.

At the north boundary of the Mount Tom quadrangle five different colored layers can be seen in the walls of the Owens River Gorge. Samples were collected from the west wall of the gorge at vertical intervals of about 25 feet along an old road near the west side of sec. 22, T. 5 S., R. 31 E. The boundaries between the layers appear sharp from a distance, but are difficult to place precisely on close inspection. Changes in color and specific gravity are shown in figure 67. Except in the uppermost grayish orange-pink layer the specific gravity increases in depth. Pumice fragments are not conspicuously flattened in the top two layers, are progressively flattened downward through the next two layers, and are extremely scarce in the bottom layer. The rock in the top layer and in the two lower layers is very tough and cohesive, whereas the white layer is soft and the underlying pale red-purple layer crumbles easily. Both the white and pale red-purple layer contain abun-

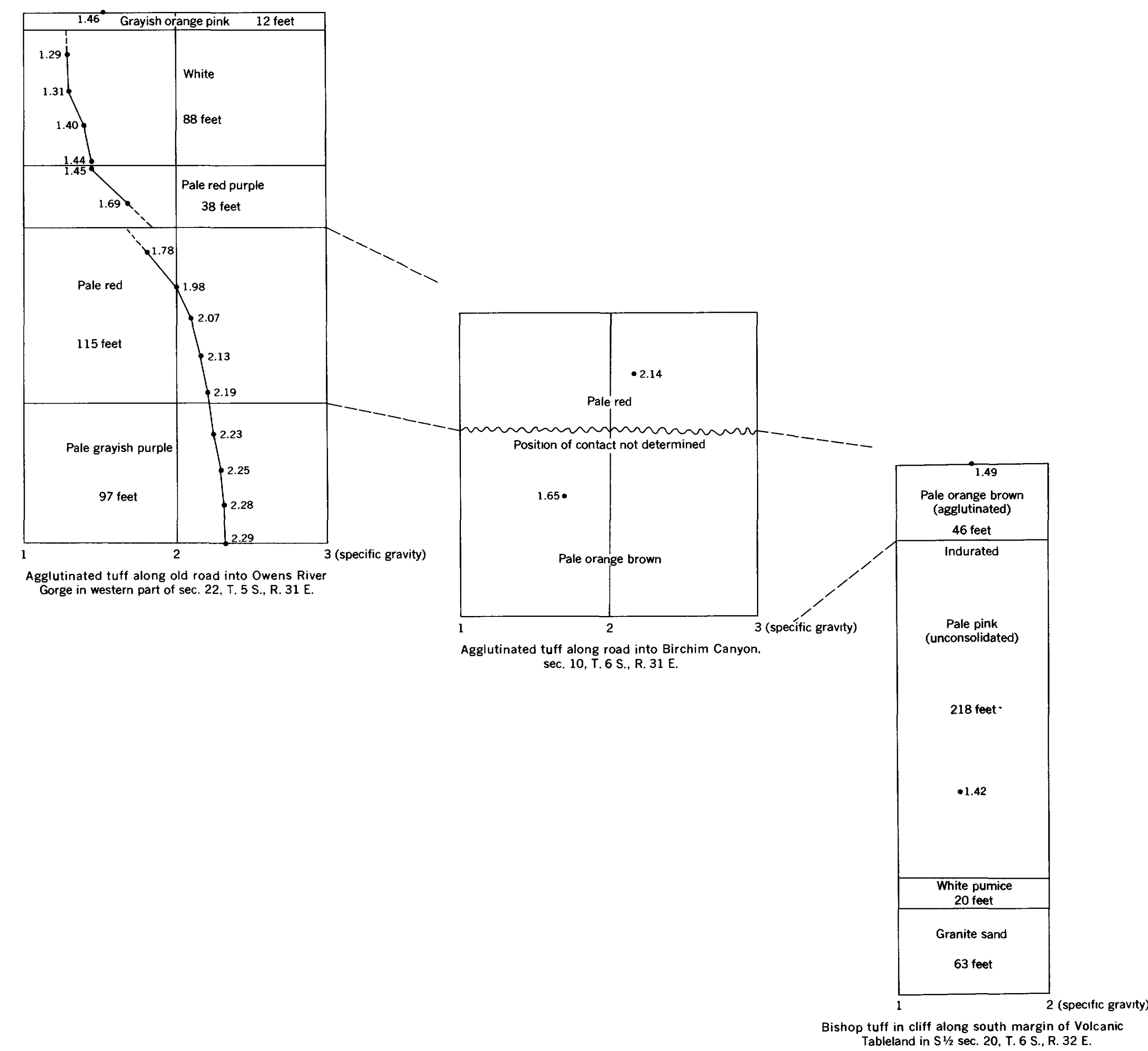

FIGURE 67.—Vertical sections through the Bishop tuff showing thickness, color, and specific gravity.

dant small devitrified pumice fragments that weather out readily. The variation in specific gravity suggests that the four lower layers were extruded closely enough to have been somewhat fluid during cooling and to have cooled together as a single cooling unit. The top layer, however, must have been extruded somewhat later, its high density as compared with the immediately underlying tuff may be the result of loss of volatiles without replacement from cooling tuff at depth. Furthermore, the devitrification of the pumice fragments in the white layer and in the upper part of the pale red-purple layer was almost certainly caused by escaping gases. The fresh pumice fragments in the top layer indicate that this layer was extruded after gas activity had greatly diminished.

The white layer can be followed in the walls of the Owens River Gorge downstream to about a mile above Birchim Canyon. The fact that in this distance the white layer is just below the rim indicates that the surface is a dip slope. For a mile above Birchim Canyon exposures are poor because of loose rubble on the walls of the gorge, and in Birchim Canyon the rocks are significantly different. Pale-red tuff in the canyon rim, identical in appearance and specific gravity to the pale-red tuff exposed halfway down the walls of the Owens River Gorge upstream, is underlain by less dense pale orange-brown tuff resembling the grayish orange-pink layer that forms the rim rock in the Owens River Gorge upstream (fig. 67). If the pale-red layer in Birchim Canyon is the same as the one upstream, and the underlying pale orange-brown layer is either the lateral equivalent of the pale grayish-purple layer upstream or, more likely, is a unit not exposed upstream. Very likely the pale orange-brown layer is a new layer not exposed in the section upstream, for even though layers do change laterally, a layer could hardly have been emplaced at one locality so close in time with an overlying layer as to have cooled with it, and in another to have preceded the same layer so far as to have solidified alone.

Downstream from Birchim Canyon the pale orange-brown tuff gradually rises along the walls of the gorge to the surface and forms the rim along the south margin of the Volcanic Tableland. In the south half of sec. 20, T. 6 S., R. 32 E., less than 50 feet of pale orange-brown tuff is present overlying more than 200 feet of pale-pink unconsolidated tuff, which rests on 20 feet of pumice of the basal member (fig. 67). Eastward the pale orange-brown tuff thins; it is missing over part of the Volcanic Tableland east of Fish Slough.

The unconsolidated pale-pink tuff is composed of the same constituents as agglutinated tuff, but is very different in appearance because of the absence of any agglutination. Typically, pumice fragments as much as several inches across are embedded in an ashy matrix. The pumice fragments are rounded and are physically much weaker than in the underlying basal pumice layer, probably because they contain larger, less symmetrical vesicles having thinner walls. The formation is generally unsorted, but mounds 5 feet high of pumice fragments as much as 8 inches in maximum dimension are present locally in the base of the tuff (fig. 65*A*). The color of the tuff changes slightly from place to place, and in the eastern margin of the Volcanic Tableland is locally light red adjacent to a tongue of agglutinated tuff.

Unconsolidated tuff is exposed in the south and east margins of the Volcanic Tableland where it rests on the basal pumice member along a very sharp and regular contact (fig. 65*A*). At the top it grades through about 15 feet into the overlying agglutinated tuff. However, along Fish Slough it is increasingly coherent northward and intertongues with and grades into agglutinated tuff. Much of the tuff in the northern part of Fish Slough shown on the geologic map (pl. 3) as agglutinated tuff is physically intermediate between typical agglutinated and unconsolidated tuff.

Unconsolidated tuff is much more widespread than its outcrop area. Driller's logs of borings show that the principal member of the Bishop tuff continues into the alluvial fill of Owens Valley and thins progressively southward. Although the driller's logs make no distinction between agglutinated and unconsolidated tuff, the tuff underlying the valley is almost certainly unconsolidated. Unconsolidated tuff is exposed in a few places in the terraces south and east of the Volcanic Tableland and also in a quarry in the SE¼ sec. 14, T. 6 S., R. 33 E., at the base of the White Mountains. Also the tuff beneath the valley rests directly on the basal pumice layer, and only unconsolidated tuff has been observed in this stratigraphic position.

Unconsolidated tuff also was intersected at the base of the formation in at least one tunnel driven for the Los Angeles Department of Water and Power in the west wall of the Owens River Gorge north of the mapped area. Here, and in a road cut along U.S. Highway 395 just east of Rock Creek, unconsolidated tuff rests on glacial till of Blackwelder's Sherwin stage. Elsewhere along Rock Creek unconsolidated tuff rests on granitoid and metamorphic rocks.

Rounding of the pumice fragments indicates abrasion, but lack of sorting, the presence of only sparse exotic materials, widespread areal distribution, and lateral gradation to agglutinated tuff indicate the transporting agent was something other than normal winds or water. The distribution of the unconsolidated tuff indicates that it is a marginal facies of the agglutinated tuff, and must therefore have been deposited as marginal, distal, and basal parts of nuées ardentes, which lacked sufficient heat to weld themselves. According to Gilbert (1938, p. 1853), the "ignimbrite" of the North Island of New Zealand, which is petrographically similar to the Bishop tuff, grades downward to unwelded tuff at its base. Water or ice at the base of tuff would, of course, have increased the cooling rate and favored the formation of unconsolidated rather than agglutinated tuff.

AGE

At several places north of the Bishop quadrangle the Bishop tuff is locally interstratified with tills, and therefore is of Pleistocene age. The tills have been intensively studied by Putnam (1938, 1949, 1950, 1952) who has concluded that the Bishop tuff rests on Sherwin till along Rock Creek and is overlain by Tahoe till west of the Mono Craters.

A potassium-argon age determination of 870,000 years for the Bishop tuff was published by Evernden, Curtis, and Kistler (1957); subsequent work by D.G. Dalrymple (oral commun. 1964) indicates that 0.7 million years is a better figure.

RHYOLITE SOUTH OF BIG PINE

Pumiceous rhyolite in the central part of a low rounded rhyolite hill which rises above the alluvial fan 2 miles west of the Poverty Hills was being exploited in 1958 for expansible rhyolite (perlite). The hill is a mile long in an easterly direction, has a maximum width near the middle of half a mile, and rises about 200 feet above alluvial fans from the Sierra Nevada. The geology of the hill was studied some years ago by Mayo (1944), and his report includes detailed descriptions of the rocks and structures.

The hill is probably an extrusive dome having short flows. The distribution of rocks and the attitudes of flow banding suggest that the dome is shaped like an asymmetric mushroom and that the vent is approximately beneath the high point in the east end of the hill. From the vent viscous rhyolite flowed in all directions, but more flowed down the alluvial slope than in other directions. In the central and north parts of the hill the flow layers generally are steep, whereas along the south and east sides they are flat or gently dipping. Buckles are present in the layers along the north side of the hill.

Three kinds of rocks were distinguished: (1) dense, gray vitrophyre, (2) thinly interlayered black glassy obsidian and gray vitrophyre, and (3) light-gray pumiceous rhyolite (fig. 68). In most places the rocks are in sharp contact and readily distinguishable from one another. Pumiceous rhyolite occupies the central part of the hill and, except at the west end where alluvial fan deposits overlap the pumiceous rhyolite, is flanked concentrically first by a discontinuous obsidian-bearing layer, then by gray vitrophyre. Locally on the northeast side of the hill, a second obsidian-bearing layer is intercalated in the gray vitrophyre. In the western part of the hill an obsidian-bearing layer is missing between the pumiceous rhyolite and gray vitrophyre. In the east and southeast sides of the hill, where the flow layers generally are flat, gray vitrophyre appears as the lowest layer and is overlain approximately conformably by an obsidian-bearing layer; the pumiceous rhyolite occurs at the top. It seems certain that here the gray vitrophyre was extruded first and the pumiceous rhyolite last. Dikes of pumiceous rhyolite that cut both the gray vitrophyre and the obsidian-bearing layers clearly indicate that it was the last extruded rock. Probably the time that elapsed between extrusious was slight and they were all part of a single period of volcanism.

At the surface the pumiceous rhyolite is highly brecciated, but in quarries at the two ends of the hill it is increasingly coherent at depth. The obsidian-bearing units crop out only locally, but their presence is clearly marked by dark-gray soil that contains abundant obsidian pellets. In most exposures, the obsidian bands have been stretched into thin discontinuous lenses, and, locally, where the rock is most strongly contorted, the obsidian is in rounded pellets that are enclosed in gray vitrophyre. At least part of the contortion in the rock may have been caused by partial remelting at the time of extrusion of the pumiceous rhyolite.

The rhyolite probably is of Pleistocene age, possibly of the same general age as the Bishop tuff. The gross shape of the dome may not be very different from the original shape, but it is considerably modified in detail. Certainly it is older than adjacent basaltic cones and flows of late Pleistocene age.

BASALT FLOWS AND CINDER CONES

Well-preserved cinder cones and associated basalt flows are present in the fan slopes south of Big Pine in the southeast corner of the Big Pine quadrangle. They compose the northern half of a volcanic field that extends to the east and south into the Waucoba Mountain and Mount Pinchot quadrangles. The cones consist chiefly of grayish-red cinders, but contain abundant bombs and large angular blocks of basalt. The mineral content of the basalt is similar to that of the older basalt west and southwest of Bishop. Conspicuous phenocrysts of olivine together with smaller crystals of augite are contained in a groundmass of augite, plagioclase, and magnetite, in places accompanied by glass.

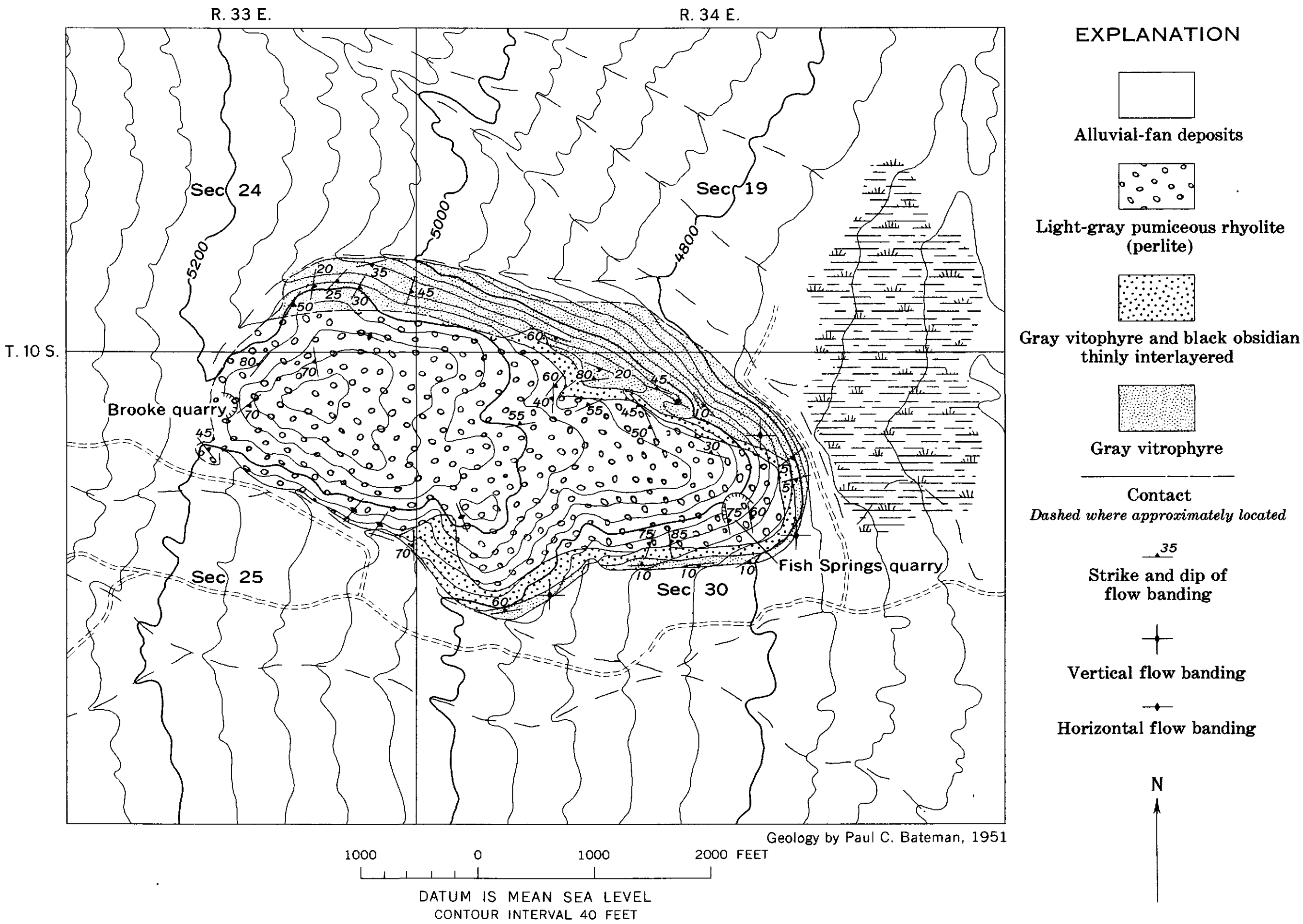

FIGURE 68.—Geologic map of the rhyolite hill south of Big Pine.

Although the cones and flows are exceedingly fresh, they are, nevertheless, probably of late Pleistocene age. A flow in a similar state of preservation a few miles to the south in Sawmill Canyon is, according to Knopf (1918, p. 77–78), overlain by a lateral moraine of his later glacial epoch, which is equivalent to Blackwelder's (1931) Tioga stage.

Red Mountain and Crater Mountain are the two main volcanic centers within the mapped area. Red Mountain is a very nearly symmetrical cinder cone that rises 700 feet above the alluvial slope on which it rests. A flow of dark lava has broken out from its eastern base and extends to beyond the eastern boundary of the mapped area. This flow is extremely rough and scoriaceous, and at the lower end exhibits row on row of arcuate pressure ridges.

Crater Mountain, on the other hand, consists chiefly of lava; cinders are present only adjacent to two vents in the crest of the mountain. Crater Mountain rises at least 1,500 feet above the alluvial slope, and more than 2,000 feet above the floor of Owens Valley. Because it stands well out into the valley, it is readily visible for many miles to the north and south. It is, however, built on an upfaulted horst of granite that crops out at the south end of the mountain in Fish Springs Hill and on the north side in several smaller areas. Fish Springs Hill rises 1,400 feet and the higher patch to the north 1,200 feet above the alluvial slope. The logs of water wells along U.S. Highway 395 show that along the east edge the basalt interfingers with alluvial fill to a depth of at least 200 feet beneath the valley floor.

All the volcanic feeders are localized along faults. Crater Mountain, Red Mountain, and an intervening mass of basalt lie along the same north-trending fault. A west-facing scarp in the alluvium between Red and Crater Mountains and extending into the Crater Mountain flows was formed at the time of the Owens Valley earthquake of 1872, according to Knopf (1918, p. 77). Earlier movements along the fault are indicated by the presence of the granite mass of Fish

Springs Hill, which required large vertical movement for its exposure.

A small cinder cone west of Fish Springs lies along a parallel fault marked by a scarp that breaks the alluvial fan as well as the cinder cone itself. The freshness of the scarp suggests that it too may have been formed at the time of the earthquake of 1872. Basaltic lava at Fish Springs on the upthrown side of a fault may be part of a flow, elsewhere buried, that broke out of the base of the cinder cone.

SEDIMENTARY DEPOSITS

GLACIAL DEPOSITS OF THE PLEISTOCENE EPOCH

Huge piles of unstratified glacial drift of Pleistocene age lie at the mouths of all the larger canyons in the eastern Sierra Nevada from Big Pine Creek north, and extend high into the range along some. Glaciation in the Sierra Nevada has been studied most thoroughly by Knopf (1918), Matthes (1930), Blackwelder (1931), Putnam (1949, 1950), and Birman (1954a, b). Blackwelder's paper is still, after more than 30 years, the standard reference to glaciation in the eastern Sierra Nevada.

I did not study the glacial deposits of the Bishop district in detail. The distinctions made on the geologic maps (pls. 1-4)) are those that can be made in a general geologic study without recourse to the time-consuming techniques of a glaciologist. The information gained, however, should provide a basis for further detailed studies. Wherever possible, the deposits are separated into different age groups chiefly on a physiographic basis with the aid of aerial photographs, which were intensively studied under a stereoscope. The principal criteria used are the spatial relations of deposits to one another and to other Cenozoic features, the degree of dissection, the relative abundance of faults, the differences in displacement along individual faults, the general abundance of boulders at the surface, and the general condition of weathering of boulders.

The tills, subdivided in accordance with the glacial stages of Blackwelder (1931), are from youngest to oldest, Tioga, Tahoe, Sherwin, and McGee. According to Blackwelder, the Tioga glaciation is represented by relatively small fresh undissected moraines, the Tahoe glaciation by much larger more dissected moraines, the Sherwin by generally formless piles of till, and the McGee by ancient accumulations of boulders, especially at high altitudes. In this report the Tioga and Tahoe tills are mapped separately, and all older tills are included under Sherwin and older tills. In addition to these main divisions, two subdivisions were recognized locally—a late subdivision of the Tioga, and an early subdivision of the Tahoe. The late Tioga till appears to represent a brief readvance of a retreating Tioga glacier. The early Tahoe till, on the other hand, includes morainal ridges more dissected and apparently older than the later Tahoe moraines, and may in fact represent an unnamed period of glaciation suggested by Blackwelder between his Tahoe and Sherwin glaciations. If detailed work indicates the existence of glacial till in this position, such till will include parts of the glacial deposits mapped here with both the Tahoe and the Sherwin and older tills.

The most complete representation of glacial deposits in the mapped area is along Bishop Creek, and in the descriptions of the different tills special attention is given these deposits. This place is the most favorable within the area for making more detailed glacial studies (pls. 1, 2; fig. 69). Differences in the degree of dissection and in the throws along two faults that cut across the different tills on the northwest side of Bishop Creek indicate that the glacial deposits are successively younger to the southeast. Each successive glacier was southeast of its predecessor, and all the morainal ridges on the northwest side of Bishop Creek are lateral moraines that were deposited along the northwest sides of these glaciers.

The glacial deposits along Big Pine Creek (pl. 4) are as extensive as those along Bishop Creek, and the deposits there also are successively younger toward the south. However, relations along Big Pine Creek are less clear than those along Bishop Creek because the older tills there have been cut by many normal faults, and the uplifted segments are deeply eroded.

SHERWIN AND OLDER TILLS

Sherwin and older tills include all the deeply eroded glacial deposits of pre-Tahoe age. Several glaciations may be represented. Surfaces of such tills generally are hummocky, and only the largest of original glacial features can be dubiously identified. Boulders are generally less abundant at the surface than on younger moraines, and many are deeply weathered and cavernous. In road cuts and other exposures of lower levels of the tills in place, the granitic boulders commonly are so deeply weathered that they can be sliced through easily with a shovel.

Much of the extensive till on Coyote Flat and adjacent areas shown on the maps (pls. 1–4) as undifferentiated is probably Sherwin. U-shaped valley forms indicate that ancient glaciers existed in the upper reaches of Coyote and Rawson Creeks and along upper Onion Creek. Early glaciers probably also mantled the east side of Coyote Ridge, where Tioga and probably Tahoe glaciers existed later, and may have extended into Coyote Flat. Small patches of older till or drift are present on the southeast side of Bishop

FIGURE 69.—Aerial view of the Sierra Nevada crest west of Bishop. Lateral moraines along Bishop Creek are in center and right foreground. The light-colored rock in the escarpment in the right central part of the photograph is porphyritic quartz monzonite similar to the Cathedral Peake granite. To the left and lower than the picture center is a probable landslide that resembles a stabilized rock glacier. Lake Sabrina is near the left edge of the photograph; in the right middle distance is Desolation Lake. Photo by Symons Flying Service.

Creek at higher altitudes than the crests of nearby Tioga and Tahoe lateral moraines. Two patches near Andrews (Shreves) Camp, one about a mile north and the other about a mile and a half east, are probably erosional remnants of a lateral moraine deposited by an early glacier along Bishop Creek. Three patches along Coyote Creek at altitudes of 6,600 to 8,000 feet, however, had their source in a glacier that headed in the Coyote Flat area. These patches rest on remnants of basalt flows. The antiquity of the till is indicated by the depth of Coyote Creek canyon, locally as much as 800 feet below the base of the till. This cutting must have taken place largely before the deposition of the Tahoe and Tioga moraines along Bishop Creek, else extensive destruction of the lower parts of the moraines below the junction of Coyote Creek with Bishop Creek would have occurred.

On the northwest side of the Bishop Creek moraine, several small disconnected mounds of bouldery gravel are assigned to the Sherwin and older glaciations. Some doubt exists as to whether these mounds are till from Bishop Creek or uplifted and dissected outwash, possibly from the part of the Sierra Nevada escarpment west of the Buttermilk Country. In gross appearance the material in them resembles the till on Coyote Flat. The largest mound is cut longitudinally by a stream channel tributary to Birch Creek. Flat-topped ridges on both sides of the channel parallel Tahoe and Tioga lateral moraines, and possibly are themselves terraced morainal ridges. West of Dutch Johns Meadow these ridges are cut by a northwestward-trending normal fault that is downthrown to the southwest. The escarpment along the fault is more than 200 feet high in the northwest ridge, but only 80 feet high in the southeast ridge. This difference results from unequal planation of the downthrown side of the fault by ancestral Birch Creek, and does not reflect different amounts of throw in the two ridges. However, progressively diminishing heights of escarpment toward the southeast in progressively younger lateral moraines is attributed to lesser total throw during the smaller time intervals presumed to be represented.

The Sherwin or older till in the north side of the Big Pine moraine and in the north side of the moraine along the Birch Creek in the Big Pine quadrangle is weathered and dissected as much as the old till on Coyote Flat and in the north side of the Bishop Creek moraine. However, the till along Big Pine Creek is cut by a group of parallel north-trending faults, which may have accelerated erosion and produced a high degree of dissection in a brief period. A lateral moraine in Bishop Creek assigned to an older subdivision of the Tahoe was not recognized along Big Pine Creek and may be included with the dissected till. Tills along McGee Creek, in the Mount Tom quadrangle, and along Tinemaha Creek, in the Big Pine quadrangle, mapped as Sherwin and older tills may also be correlative with the older subdivision of the Tahoe. This possibility is suggested by probable morainal ridges in these tills, by the abnormally small mass of debris assigned to the Tahoe, and by the apparent absence of the older Tahoe deposits.

Undoubtedly the Sherwin and older till formerly was much more widespread and has either been removed by erosion or concealed by younger deposits. Too, some older till may be included with Tahoe deposits on the geologic maps (pls. 1–4).

TAHOE TILL

According to Blackwelder (1931, p. 884), "The most conspicuous moraines in the Sierra Nevada are those of Tahoe age. They are well developed and easily studied in the Big Pine, Bishop, and Pine Creek districts of Owens Valley, and in nearly every important canyon northward to Truckee and beyond. In general, moraines of this age have been recognized in all the valleys that also contained glaciers of the Tioga age and also in some that did not." In the Bishop district, in addition to the moraines mentioned by Blackwelder along Pine, Bishop, and Big Pine Creeks, Tahoe moraines were mapped in the morainal piles of most other canyons along the Sierra Nevada front.

The original glacial form of Tahoe moraines generally is easily recognized, but the crests of glacial ridges have been rounded by erosion and finer textured details are obscured. Boulders in the till are fresher and more abundant at the surface than in older tills, but not as fresh and abundant as in Tioga till. Tahoe till is especially well represented along the northwest side of Bishop Creek where it lies between Sherwin and older till on the northwest and Tioga till on the southeast.

Two parallel morainal ridges of considerably different aspect are included within the Tahoe till mapped along Bishop Creek; a younger, higher, sharp-crested one to the southeast, and an older, lower, round-crested one to the northwest. Both ridges are believed to be lateral moraines. In the vicinity of Dutch Johns Meadow, the crest of the younger moraine stands 150 to 200 feet above both the Tioga moraine to the southeast and the older Tahoe moraine to the northwest. This younger moraine nearly parallels the Tioga lateral on the northwest side of Bishop Creek in most places, but is cut into by the Tioga moraine in secs. 34 and 35, T. 7 S., R. 31 E. The older lateral moraine is here assigned to an older subdivision of the Tahoe, although it might represent an independent glaciation. A substantial difference in the ages of the two lateral moraines is suggested by degree of erosion and by greater heights of fault escarpments in the older lateral moraine in the vicinity of Dutch Johns Meadow. A fault southwest of Dutch Johns Meadow has a 60-foot escarpment in the older lateral moraine and displaces minor ridges on the northwest side of the younger one, but does not cut the crest. Presumably the crest is younger than the last displacement along the fault. Northeast of Dutch Johns Meadows, a fault displaces both morainal ridges, but the scarp in the older ridge is greater and suggests that more increments of movement are represented in the older ridge than in the younger. Neither fault displaces the Tioga moraine.

The Tahoe terminal moraine is exceptionally well preserved in Bishop Creek canyon over an area of more than 2 square miles. Southeast of Sand Canyon, glacial ridges and depressions that probably were caused by lobes of ice are present. The excellent preservation of these features suggests that they are of the same age as the younger Tahoe lateral moraine, but some of the debris in the terminal moraine may be correlative with the older lateral moraine.

Along Big Pine Creek, Tahoe lateral moraines are present on both sides of the canyon below the Tioga terminal moraine at Sage Flat. Above this point, the Tahoe moraine is present only on the north side of the canyon westward to the fault in the east half of sec. 30, T. 9 S., R. 33 E. West of the fault the moraines are deeply dissected and were not mapped separately. The main Tahoe moraine is probably equivalent to the younger Tahoe moraine along Bishop Creek because of similar dissection. Northwest of Sage Flat two morainal ridges, south of the main Tahoe moraine and between it and the Tioga moraine, are younger than the main moraine and may represent a late subdivision of the Tahoe that was not recognized along Bishop Creek. The older Tahoe till of Bishop Creek was not recognized along Big Pine Creek and may have been mapped with the Sherwin and older tills.

The older subdivision of Tahoe till was recognized elsewhere only along Pine Creek and Taboose Creek where short well-preserved lateral spurs are truncated and overlapped by more extensive younger Tahoe lateral moraines. Undoubtedly, this till is also present in other piles of glacial debris where it may be included in undifferentiated Tahoe till or in Sherwin and older till.

Moraines assigned to the Tahoe glaciation are present along both sides of Rock Creek canyon. The highest and most distinct lateral moraine on the east side, which is considered correlative with the younger lateral moraine of Bishop Creek, extends from the northern margin of the Mount Tom quadrangle south about 3 miles, then southeast toward Round Valley Peak to the base of Wheeler Crest. On the west side of Rock Creek canyon all the glacial debris on the broad flat in the northeast corner of the Mount Tom quadrangle is shown on the map (pl. 2) as Tahoe till. This flat is flanked on both east and west by morainal ridges that are probably correlative with the younger Tahoe laterals of Bishop Creek. The central part of the flat is covered with angular fragments belonging to a deposit of unknown age.

TIOGA TILL

Moraines assigned to the Tioga glaciation are much fewer and smaller than the moraines of the Tahoe glaciation, but are better preserved. Most Tioga moraines are "nested" in Tahoe moraines, but Tioga glaciers did not occupy all canyons that held Tahoe glaciers; some smaller canyons that underwent glaciation in Tahoe time escaped glaciation during Tioga time. Tioga lateral moraines are sharp crested, and the details of terminal and recessional ridges are clearly visible. Knob-and-kettle topography, common at altitudes of 10,000 feet or more between lateral moraines, appears as fresh as if formed very recently.

In most canyons only one period of Tioga glaciation was recognized, but along Bishop Creek a late subdivision is represented. The Tioga end moraine of the Bishop Creek glacier is largely in sec. 25, T. 7 S., R. 31 E. It consists of two lobes, one north of Bishop Creek in the old Tahoe valley, and the other along the stream-cut canyon of Bishop Creek. An outcrop of quartz monzonite at the juncture of the lobes may have contributed to the split of the glacier. Upcanyon, numerous recessional moraines cross the canyon floor, and are cut through only by a narrow slot about 20 feet deep that contains Bishop Creek. A single lateral moraine flanks the canyon on the southeast side as far upcanyon as Egypt Creek, and its mate flanks the canyon on the northwest side to the south boundary of the Mount Tom quadrangle.

The late subdivision along Bishop Creek is represented only in the Middle Fork of Bishop Creek above the junction of the Middle and South Forks. One lateral moraine of the subdivision extends along the northwest side of the Middle Fork from the junction of the forks to a ridge of bedrock northeast of North Lake. The other on the opposite side of the canyon lies across the mouth of the canyon of the South Fork. These lateral moraines descend rapidly and are at the level of the canyon bottom at their lower ends. The features of the younger subdivision are not discernibly sharper than those of ~~the older Tioga glaciation.~~ Nevertheless, the configuration of the moraines indicates that the

Tioga glacier withdrew and then readvanced, and that the moraines are not simply parts of a recessional moraine. Till between the North and Middle Forks of Bishop Creek, in the vicinity of George Lake and below Green Lake, probably also belongs to the younger subdivision.

In Big Pine Creek the Tioga moraines terminate in and below Sage Flat. Two steeply descending laterals on the north side of the canyon curve at their lower ends across the canyon and may be recessional moraines, or one or the other may be correlative with the younger Tioga subdivision along Bishop Creek.

Extensive Tioga moraines are present along upper Baker Creek and along the east side of Coyote Ridge. The glacial deposits along Coyote Ridge are unusual in that they extend in a broad loop into Coyote Flat. Much of the loop consists of knob-and-kettle topography.

In Pine Creek the precise boundary between the Tahoe and the Tioga tills is not clear. On the geologic map (pl. 2) the boundary is shown to be well toward the lower end of the morainal pile, but the proper division may be at a morainal loop present where the canyon leaves bedrock.

Extensive Tioga glacial deposits are present in Rock Creek, but the relations there are obscure. The principal lateral moraine on the east side of the canyon is plastered on the side of the Tahoe lateral moraine. Its crest generally lies a hundred feet or so below the crest of the Tahoe moraine, but extensive slumping causes it to be very irregular. On the north side, several lateral moraines are present, most of which probably bend into recessional moraines.

Smaller Tioga moraines are present in the morainal piles of Horton, Birch (Mount Tom quadrangle), McGee, Birch (Big Pine quadrangle), and Tinemaha Creeks. All these piles consist of a series of looped end moraines that lie within older till just outside the mouths of bedrock canyons.

OLDER DISSECTED ALLUVIAL FAN AND LAKEBED DEPOSITS

The White Mountains in the mapped area and the Sierra Nevada locally between Bishop Creek and the Birch Creek in the Big Pine quadrangle are flanked by alluvial deposits, some of which have dissected surfaces whereas others have depositional upper surfaces (Knopf, 1918, p. 54–57). Remnants of alluvial fans also are present in the eastern part of the Volcanic Tableland. Some of the dissected deposits have remnants of their original upper surfaces, which slope parallel with adjacent, more recent alluvial fans but which stand at higher altitudes. Such untilted deposits are present along the westward trending segment of the Sierra Nevada escarpment southwest of Bishop and along the White Mountains south of Poleta Canyon, where they extend upward to altitudes of more than 6,000 feet within the mapped area and to higher altitudes farther to the east (fig. 2). Other dissected deposits in fault blocks that have been rotated toward Owens Valley are exposed along the Sierra Nevada front south of Rawson Canyon, along the White Mountains north of Poleta Canyon, and, locally, in a narrow belt adjacent to the alluvial fill of Owens Valley, south of Poleta Canyon.

The basis of distinction between the dissected series and undissected fan deposits—whether the surface in the aggregate is erosional or depositional—precludes many of the common stratigraphic relations between formations. The dissected series cannot be defined in terms of top and bottom, or assigned an age span that is distinct from that of the undissected fan deposits. The upper beds of some dissected blocks conceivably may be as young as the upper beds of undissected fans.

The dissection of the deposits is the result of structural movements related to progressive depression of Owens Valley relative to the bordering ranges. Two periods of faulting can be clearly identified in the White Mountains, and the probable structural history requires a great many periods, the evidence of which has been concealed or obliterated. Because the dissection results from movements of different magnitudes which occurred at different times, distinction between the dissected and undissected fan deposits is difficult in some places, although clear enough in most. A typical difficult place is along the Birch Creek drainage in the Big Pine quadrangle, south and southwest of Crater Mountain, where fans have been faulted so recently that erosion is confined to the scarps.

DISSECTED FANGLOMERATE AND LAKEBED DEPOSITS ALONG THE BASE OF THE WHITE MOUNTAINS

In the White Mountains the thickest unbroken exposed sections of dissected material are in Black and in Redding Canyons, where about 600 feet of coarse, crudely bedded fanglomerate is exposed. Virtually all of the material in the fanglomerate in these canyons was derived from older rocks that crop out higher in the same drainage basins. Much of the fanglomerate, except in the upper part, is cemented with carbonate, which produces a tough cohesive rock that stands in vertical or near vertical cliffs where it has been cut by streams. White pumice that is considered to belong to the lower basal unit of the Bishop tuff locally is intercalated in the fanglomerate. One of the more extensive and better exposed outcrops of pumice, in the south wall of Black Canyon, is at an altitude of 5,200 feet, about 150 beneath remnants of the original upper surface of the fan.

The material exposed in these larger canyons, however, is not typical of the entire formation. Much of the material that crops out lower in the alluvial slope, and between the larger canyons, contains finer grained deposits, including granitic sand and pebble beds whose source could not be found in adjacent parts of the White Mountains. Locally, thin layers of fresh-water limestone and calcareous shale that contain abundant remnants of the thin-shelled gastropod *Hydrobia* sp. are present. According to Dwight W. Taylor (written communication, 1956), this gastropod lives in quiet water—in lakes, ponds, or in a backwater along a stream. Such finer grained material is well exposed between Redding and Black Canyons and in dissected slopes north and south of the mouth of Silver Canyon. In the north wall of Poleta Canyon, in the northwest corner of sec. 18, T. 7 S., R. 34 E., limestone and gastropod-bearing marly shale and finer grained pumiceous sandy layers, are interstratified with coarse-grained layers identical with the fanglomerate of Black and Redding Canyons.

The detailed stratigraphy of the dissected alluvial deposits along the base of the White Mountains was not determined; in particular, neither the lateral nor the vertical extent of finer grained sediments that appear to be at least in part lacustrine nor their relation to coarse fanglomerate was established by direct observation. The overall distribution of materials suggests that the finer grained sediments grade laterally into the coarser—at the mouths of powerful streams that flowed in the main canyons, fanglomerate was deposited simultaneously with finer grained materials in interfluves and peripheral to the fans.

One interpretation of these relations is that the finer grained sediments were deposited in a single very extensive lake in Owens Valley. Fans would have flanked this lake, and their lower ends would have extended into the lake, where deposition would, in fact, have been deltaic. Intertonguing of very coarse material with lacustrine sediments could result from periodic torrential storms which caused coarse fanglomerate to be carried by rain-swollen streams into the lake, or from fluctuations in the lake level. If such a lake existed, it probably extended at least as far north as Benton, 30 miles north of Bishop where the northernmost exposures of the basal pumice layer of the Bishop tuff are found, and as far south as Zurich where extensive lakebeds lie in an embayment in the White Mountain front (Walcott, 1897). According to L. C. Pakiser and M. F. Kane (oral communication, 1958), gravity studies indicate very shallow bedrock in the floor of Owens Valley just south of the Poverty Hills, which could mark the southern limit of the lake.

A second possible interpretation of the relations is that no single large lake existed. All the demonstrable lacustrine deposits could have been deposited in small ephemeral ponds such as are present in low-lying places on the floor of Owens Valley today. Ponds may even have been formed high on the fans behind fault scarps that faced upslope. Such scarps are abundant today, and although none dams a pond or lake, many bound marshy areas.

DISSECTED FANGLOMERATE ALONG THE BASE OF THE SIERRA NEVADA

The dissected deposits along the Sierra Nevada front are less deeply incised, and consequently are less well exposed, than those along the base of the White Mountains. The material exposed is coarse fanglomerate, similar to material in undissected fans, and consists of well-rounded granitic pebbles, cobbles, and boulders set in a sandy and clayey matrix. The material probably represents outwash from glaciers that existed higher in the range during the Pleistocene epoch.

ALLUVIAL REMNANTS IN WEST PART OF VOLCANIC TABLELAND

Scattered remnants of fanglomerate are present along the west side of the Volcanic Tableland in the north-central part of the Mount Tom quadrangle. The remnants are thin and probably are nowhere more than 50 feet thick. Formerly, fans extended from the north into this area, but structural depression of Round Valley has caused dissection of these older fans and the building of new ones west of the Volcanic Tableland.

AGE

The strata in the dissected series are almost certainly of Pleistocene age. The preservation of original features in the youngest Pleistocene glacial deposits, including easily eroded recessional ridges in canyon bottoms, indicates that little erosion of the ranges, and consequently little deposition, has taken place since the end of the Pleistocene epoch. It seems probable that not only are the dissected alluvial deposits older than Recent, but that a large part of the undissected alluvial fans must also be older than Recent. The presence of the basal pumice member of the Pleistocene Bishop tuff among the strata of the dissected series along the base of the White Mountains indicates a middle Pleistocene age for the adjacent beds. The stratigraphically high position of this pumice layer in the section exposed in Black Canyon, only 150 feet beneath the original surface in an exposed section of 600 feet, suggests that the lower beds may be early Pleistocene or even late Tertiary in age.

LANDSLIDE

Landslide material on the south side of Bishop Creek at the lower end of the younger substage of the Tioga moraine includes both quartz monzonite and glacial till. The landslide probably resulted from oversteepening caused by glaciation and may have occurred during the Pleistocene.

TERRACE GRAVELS

Thin veneers of river gravels cap elevated terraces along the south and east sides of the Volcanic Tableland. The gravels consist of well-rounded particles that range from sand to cobbles 6 inches or more in diameter. The gravels are in crudely stratified, poorly sorted layers. The material in the south-side terraces is chiefly granitic and obviously had its source in the Sierra Nevada. The material in the east-side terraces, on the other hand, is largely metamorphic, similar to rock exposed in the north end of the White Mountains. Three terrace levels are distinguishable in the south-side terraces, but only two in the east-side terraces. The highest and oldest terraces are most distant from the modern stream channels, and the terraces closer to these channels are progressively lower and younger.

The terraces on the south side of the Volcanic Tableland were cut by the Owens River and a tributary from the west that joined the Owens River where it leaves the Volcanic Tableland. Gravels cover the terrace levels almost continuously to depths of several feet, and have slumped over the edges of the terraces. Outcrops of the underlying rock are mostly in the terrace edges or in road cuts, although a mound of olivine basalt projects through the middle terrace in the N½ sec. 32, T. 6 S., R. 32 E. Most of the outcrops are of Bishop tuff.

The terrace gravels on the east side of the Volcanic Tableland are much less continuous than those on the south side, and are in long north-trending strips that are separated from one another by gullies in which Bishop tuff is exposed. The east-side terraces were cut by an ancient tributary to the Owens River from the north, which at one time carried a large volume of water, but which at present is dry except immediately after storms. The metamorphic material in the gravels indicates that some tributaries to this stream headed in the western flank of the White Mountains. Some water also may have come from Mono Lake during periods of overflow in the Pleistocene epoch.

The middle terrace on the south side of the Volcanic Tableland and the upper terrace on the east side are considered correlative because they are more extensive than any other terraces and because the 4,400-foot contour crosses both terraces at about equal distances from the probable point of confluence of the streams that cut the terraces. Such correlation, nevertheless, is hazardous because the south-side terraces have been slightly deformed by faults and warps; the east-side terraces may also have been deformed.

The fact that gravels in the terraces are much coarser than the gravels in the bed of the Owens River indicates that the terrace gravels were deposited by faster flowing and presumably larger rivers capable of carrying larger particles. More water was, of course, available during certain parts of the Pleistocene epoch, and quite likely the terraces were cut and the gravels deposited in late Pleistocene time. Nevertheless, the lowest terrace at least may have been cut more recently, during a pluvial period in which many ancient lakes in the Great Basin were filled.

UNDISSECTED ALLUVIAL FAN DEPOSITS

Alluvial fans having constructional forms and no greater dissection of their surfaces than is usual in the ordinary processes of fan formation border Owens and Round Valleys and extend into canyons of the White Mountains and the Sierra Nevada. Much smaller fans have been deposited in broad glaciated canyons in the Sierra Nevada by streams flowing in from the sides. The material in the fans along Owens Valley has been derived both from bedrock formations and from older raised and dissected alluvial deposits. The proportion of material derived from each source varies from fan to fan; presumably larger proportions of White Mountain fans than Sierran fans have been derived from older alluvial deposits. Individual fans have distinct forms even though they merge to form piedmont alluvial slopes, and each fan has its own distinct distributary stream pattern. The crests of the larger fans slope toward the valleys at an average rate of about 300 feet per mile, but the upper part slopes more steeply and the lower more gently.

The upper boundaries of the fans with bedrock or with older sedimentary or volcanic deposits commonly are easily distinguished, but because all the fans are progressively finer grained and better sorted downslope and many grade imperceptibly into fine-grained alluvial fill that constitutes the valley bottoms, some difficulties arise in mapping the lower boundaries. Among the fans having gradational lower boundaries are those of Bishop Creek, Big Pine Creek, and all the fans that flank Round Valley. A distinct train of material from Bishop Creek can be traced on aerial photographs into Owens Valley as far south as Big Pine. Both Bishop and Big Pine Creeks appear to have contributed larger volumes of material to Owens Valley recently than the Owens River, although the Owens River may have supplied a larger proportion of material earlier. On the map the boundary between these fans and valley fill is a line that approximately separates fine well-sorted sandy, silty, and clayey alluvial fill in the valleys from pebbly and conglomeratic material in the fans. In Round Valley this line approximates the boundary between cultivated and barren cobble-covered fan slopes.

The fans along the White Mountain front and along the Sierra Nevada front between Bishop and Big Pine Creeks have been overlapped by alluvial fill carried into the valley by the larger, more powerful streams, and this overlapping provides a basis for the separation of the fans from alluvial fill. The general slope of the valley floor is southward, whereas the fans slope radially from the points where the parent streams emerge from their canyons. Thus the contours bend sharply where the fans meet the valley floor, and the boundary contact can be drawn through these bends.

The bulk of the material in fans must have been deposited during Pleistocene time when the rates of erosion and deposition were accelerated because of heavier precipitation, although some smaller fans in glacial canyons are Recent. The fans of Pine, Bishop and Big Pine Creeks almost certainly were formed by outwash from glaciers higher up, and other fans along the Sierra front may have been built partly from glacial debris washed down from higher altitudes.

Locally the upper surfaces of fans have been veneered by material deposited by floods in Recent time. Blackwelder (1928, p. 471) reports bouldery mudflows in the alluvial fans along the east side of the Sierra Nevada south of the Bishop district. In 1946, a cloudburst in the Gable Creek drainage caused a fan to be built in Pine Creek canyon at the mouth of Gable Creek within a few hours time. Buildings near the head of the fan belonging to the Tungstar Corporation were destroyed, and houses lower down on the fan occupied by employees of the Union Carbide Nuclear Co. were partly buried.

ALLUVIAL FILL

Alluvial fill includes the detrital material in the lower central parts of Owens and Round Valleys and in Fish Slough. In general, alluvial fill is finer grained and better sorted than fan material or terrace gravel, but nevertheless exhibits considerable range in the degrees of sorting and in the average size of particles. The fill was deposited chiefly on flood plains by streams and in shallow ephemeral ponds and lakes such as Warren and Klondike lakes north of Big Pine and the small ponds along Fish Slough.

At the surface the alluvial fill probably is chiefly Recent in age, although detritus of Pleistocene age must be present at shallow depths, and may be exposed locally at the surface. The Bishop tuff has been intersected in borings into Owens Valley as far south as Bigelow station, and the adjacent beds are certainly of Pleistocene age.

DUNE SAND

Sand dunes that have been stabilized and are overgrown with vegetation are present 4½ miles west of Bishop where U.S. Highway 395 crosses the highest terrace. Accumulations of sand also are present in the W½ sec. 21, T. 7 S., R. 32 E., where the sand is in pockets on the northeast side of a rounded spur of the Sierra Nevada. Most of the dunes are formless, but several are crescentic. The fact that the convex sides of the crescentic dunes face northeastward indicates that the prevailing wind was from that direction at the time of deposition. The dunes probably are of the same age as the highest terrace.

TALUS AND ROCK GLACIERS

Talus is present in both the Sierra Nevada and White Mountains, but is especially abundant along the lower slopes of the Sierra Nevada escarpment and in glacial cirques at higher altitudes in the Sierra Nevada where frost action is effective in fragmenting the well-jointed granitic rocks. In the White Mountains, talus generally occurs in areas too small to be mapped, whereas in higher glaciated parts of the Sierra Nevada individual areas of talus exceed a square mile, and extensive tracts are more than 25 percent covered with talus.

The size and shape of the fragments in the talus depends largely on the nature of the source rock. Talus from the Andrews Mountain standstone member of the Campito formation generally is in relatively large angular fragments, whereas talus from Montenegro shale is slabby and breaks down rapidly to clay. In granitic rocks, the size of the fragments is largely a function of the spacing of joints, which in a measure is correlative with grain size—finer grained granitic rocks generally yield smaller fragments than coarser grained ones. Granitic fragments several feet on a side are common near the bottoms of some talus slopes. Older talus in the White Mountains commonly is cemented by carbonate derived from limestone and dolomite in the bedrock, but no cemented talus was observed in the Sierra Nevada.

Most talus cones differ from small alluvial fans by their steeper slopes and by the fact that material is coarser downslope rather than upslope as in a fan. In a few places, however, distinction on this basis is difficult or impossible because partial reworking by seasonal flow of water down talus-filled channels results in a hybrid deposit.

At higher altitudes in the Sierra Nevada, talus in glacial cirques in many places grades downslope into rock glaciers and allied forms transitional from talus to rock glaciers. The rock glaciers appear to be formed from and fed by the talus. At several places along the Sierra Nevada crest and along the Glacier Divide, rock glaciers lie below active ice glaciers rather than below talus slopes. Possibly the debris required for the rock glacier is derived from the glacier, but it is also possible that the rock glaciers formed earlier than the glaciers and are now being pushed ahead of them. Most rock glaciers are at the base of north and northeast slopes, above altitudes of 10,400 feet, but a few are lower. Fully formed rock glaciers are tongue-shaped or spatulate and have steep fronts and sides on which the talus fragments lie at the maximum angle of repose (fig. 70). The upper surfaces slope gently toward the rock glacier fronts and, especially near their fronts, are marked by arcuate ridges that are convex downslope. Longitudinal ridges and furrows may extend almost the full length of the rock glaciers. The longitudinal profiles of active rock glaciers are convex upward, whereas the the profiles of inactive ones are concave upward. Melting of interstitial ice is accompanied by partial collapse and change of the surface configuration (Wahrhaftig and Cox, 1959, p. 427).

On the geologic maps (pls. 1, 2, and 4) rock glaciers and related forms are shown by semidiagrammatic representation of the principal arcuate and longitudinal ridge crests. The largest rock glaciers are a mile or more long and more than a quarter of a mile wide, but most are considerably smaller. Related forms range from isolated, generally arcuate ridges that cross talus slopes approximately horizontally to complexes of ridges and furrows called by Wahrhaftig and Cox (1959) rock-glacier aprons (fig. 71). In some cirques several rock glaciers and allied forms are present, fed by different talus slopes, and in the cirque at the head of the south fork of Red Mountain Creek rock glaciers from high in the cirque override some that originate lower in the cirque.

Convex-upward longitudinal profiles, absence of vegetation on fronts, and delicate balance of many large fragments on ridge crests indicate that many of the rock glaciers above about 11,000 feet are moving. Inactive rock glaciers are shown by growth of vegetation on their

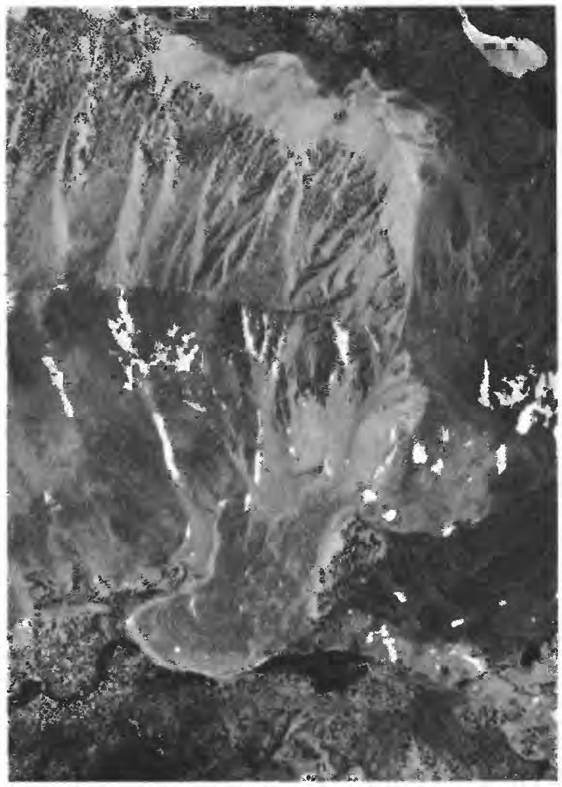

FIGURE 70.—Rock glacier between Second and Third Lakes, North Fork of Big Pine Creek.

FIGURE 71.—Vertical view of rock-glacier apron southeast of Rock Creek Lake.

fronts and by the presence of broad, concave depressions in their upper surfaces. These inactive rock glaciers must date from older, colder periods.

Just above North Lake the road into the North Fork of Bishop Creek passes along the base of the front of a physiographic form that has all of the characteristic features of inactive rock glaciers but which is probably a landslide (fig. 72). The altitude of only 9,400 feet and the fact that the form is at the base of a south slope are the principal lines of evidence that it is not an inactive rock glacier.

Wahrhaftig and Cox (1959) conclude that the rock glaciers of the Alaskan Range are mixtures of debris and ice to within a few feet of the surface, and that they move as a result of the flow of the interstitial ice. Steep cliffs, a near-glacial climate cold enough for the ground to be perennially frozen, and bedrock that breaks by frost action into coarse block debris having large interconnected voids are listed by Wahrhaftig and Cox as favorable to the formation of rock glaciers. These concepts of the composition and movement of rock glaciers and of the factors necessary for their formation are compatible with the limited observations made of Sierran rock glaciers. The Sierran rock glaciers are confined to steep-walled cirques where a near-glacial climate and frozen ground prevail. The granitic rocks are readily broken along joint planes by frost action and supply abundant talus. Some straight ridges in the lower parts of talus slopes conceivably may have been formed as protalus ramparts by the sliding of talus over snow banks, but the arcuate form of most ridges is more easily explained as the result of flow in interstitial ice.

CENOZOIC STRUCTURE AND EVOLUTION OF THE LANDSCAPE

Landscapes represent the resolution of two opposing groups of processes. On the one hand are the leveling processes of erosion and deposition, which work toward the reduction of the land to a featureless plain; on the other hand are the processes that interrupt the leveling processes and increase the relief.

In the Bishop area, situated in an active orogenic region, most of the larger features of the landscape, and many of the smaller ones, owe their existence to structural movements or to vulcanism. Of these two, structural movements have been far the more important. The Sierra Nevada, the White Mountains, and the broadest features of Owens and Round Valleys are the products of faulting and warping; the Volcanic Tableland, Crater Mountain, and Red Mountain are the prod-

FIGURE 72.—View of probable landslide in the North Fork of Bishop Creek above North Lake which resembles an inactive rock glacier.

ucts of vulcanism. None of these features exists because it is made of rock that is especially resistant to erosion.

Nevertheless, erosion has been effective in modifying almost all slopes, steepening some and flattening others, and in carving out canyons, cirques, and peaks, all notable features of the landscape. Sedimentation, likewise, has been effective, and has formed the floors of Owens and Round Valleys, the huge alluvial fans that flank these valleys, and moraines and talus cones.

CENOZOIC STRUCTURAL HISTORY OF THE SIERRA NEVADA AND ADJACENT REGIONS

During and after the emplacement of the Sierra Nevada batholith, thick sections of rocks were stripped away, and by the beginning of Cenozoic time the batholith was exposed and a broad region that extended many miles to the east and to the west was degraded. Probably beginning in Late Cretaceous or Eocene time and continuing into middle Tertiary time, the area of the present Sierra Nevada began to tilt westward. By middle Pliocene time these structural movements and concomitant erosion had produced a broad low topographic arch whose crest lay somewhat east of the present Sierra Nevada divide in about the position of Owens Valley. In late Pliocene time, stronger deformation began, which caused further arching accompanied by block faulting. During this time, east-central California and adjacent parts of Nevada were broken up into a series of warped and tilted, fault-bounded blocks, the largest of which is the Sierra Nevada. Structural movements have continued to the present, but were greatly accelerated in middle Pleistocene time, when movements of exceptional magnitude took place.

The most complete record of the progressive westward tilting of the Sierra Nevada block during Cenozoic time probably is that found in the geosynclinal sequence of the Central Valley of California. The record there has been interpreted by Hoots, Bear, and Kleinpell (1954). They state (p. 116): "The comparatively gentle eastern flank is in reality the down-dip western part of the Sierra Nevada block; its broad structural character has resulted principally from westerly tilt of this fault block throughout much of Cenozoic time." Sections drawn by them (Hoots, Bear, and Kleinpell, 1954, pl. 6) show considerable westward thickening of the strata above the base of the Miocene but only slight thickening of the strata between the base of the Eocene and the base of the Miocene. These relations suggest that tilting was slow during the Eocene and Oligocene, and greatly accelerated after the beginning of the Miocene. Unconformities in the stratigraphic section mark the following events: middle Eocene uplift, post-"Oligocene" uplift, middle early Miocene uplift, post-early Miocene uplift, orogeny near the end of Miocene time, and middle Pleistocene orogeny (p. 127–128). Regarding the middle Pleistocene orogeny, they state (p. 128): "At the close of Tulare deposition, in mid-Pleistocene time, there occurred a pronounced orogeny, the magnitude and intensity of which exceeded that of any disturbance since the Nevadan revolution of the Late Jurassic. It was this orogeny that expelled remaining small seas from several parts of California and produced through folding, faulting, and subsequent erosion, most of the present structure and topography."

Matthes (1930, 1947), Hudson (1955), and Axelrod (1957) have all outlined the Cenozoic history of the Sierra Nevada from different kinds of data found within the Sierra Nevada itself. Their interpretations of the structural history agree quite well on the whole; principal disagreements concern the magnitudes rather than the times of structural movements.

Matthes (1930, p. 27–30; 1947, p. 179–180) concluded from the physiographic record in the Yosemite region and along the eastern escarpment that upwarping of the Sierra Nevada and adjacent areas to the east began in the Eocene, and that strong movements in the Miocene raised the Sierra Nevada crest near Tioga Pass to about 7,000 feet. He thought that in late Pliocene or early Pleistocene time further arching raised the range to about its present height. After the McGee glaciation, the area east of the Sierra Nevada, including Owens Valley, was faulted down. Matthes considered faulting later in the Pleistocene and Recent along the east side of the Sierra to represent minor adjustments, and concluded that the Sierra Nevada and its neighboring blocks are in fairly stable adjustment.

Hudson (1955) concluded from studies of the Yuba River drainage that the altitude of Donner Pass in Miocene time was about 5,500 feet and that the Pliocene and Pleistocene uplift was only about 2,000 feet. He thought that the northern Sierra Nevada was not tilted as a rigid block, as has been rather generally assumed, but that at least five zones of deformation were internal to the Sierra Nevada. He also concluded that the aggregate throw of the faults on the east side of the Sierra amounted to more than the absolute uplift of the range since middle Eocene, and that the valleys east of the Sierra Nevada in the vicinity of Reno, Nev., were downfaulted in an absolute sense.

Axelrod (1957) concluded from studies of the Tertiary floras that the Sierra Nevada was quite low just before the inception of vulcanism in Miocene and Pliocene time. At that time the Sierra Nevada was a broad ridge having its summit near an altitude of 3,000 feet in the vicinity of Donner and Carson passes, and lowlands in central Nevada had an average altitude of 2,000 to 2,500 feet. Axelrod inferred that the major vertical displacement of the range took place during late Pliocene and early Pleistocene time, and that the uplift was partly the result of faulting and partly the result of general regional uplift. His figure 6 (p. 39) is illustrative of this concept. Axelrod's interpretation of the magnitude of the Pliocene and Pleistocene movement seems more compatible with the sedimentary record in the San Joaquin Valley than does Hudson's interpretation of smaller Pliocene and Pleistocene movement.

Matthes' belief that the valleys adjacent to the Sierra Nevada on the east were downfaulted only after they had been uplifted with the Sierra Nevada to nearly its present height and after the McGee glaciation is not supported by the evidence of late Tertiary block faulting close to the Sierra Nevada escarpment elsewhere (Thompson, 1956, p. 64–65). The evidence that persuaded Matthes of the middle Pleistocene downfaulting of Owens Valley is the presence of McGee till on flat-topped McGee Mountain several miles to the north of Owens Valley, bounded on the north by the Sierra Nevada escarpment and on the east and west by deep canyons. Matthes inferred that the crest of the range must have been at approximately its present height for the McGee glaciation to have taken place and that the range front escarpment could not have existed in McGee time. The McGee till has not been identified east of the escarpment, any eastward extension is presumably buried beneath fill in Long Valley, which lies at the east base of McGee Mountain. However, Long Valley has been shown by geophysical studies (Pakiser and Kane, 1956; Pakiser, Press, and Kane, 1960) to be filled with sediments and volcanics to a depth of 12,000 or more feet. Unless all this fill has accumulated since McGee time, which seems unlikely, a fault, though possibly not much of an escarpment, must have existed in McGee time. Even if Matthes' interpretation of the significance of the McGee till is correct, extrapolation of the age of faulting there to Owens Valley is hardly warranted. Probably the most reasonable picture for Owens Valley is that significant faulting began in late Pliocene or early Pleistocene time, that faulting movements were greatly accelerated in middle Pleistocene time after the McGee and before the Sherwin glaciations, and that movements at a reduced rate continue to the present.

A cross section (pl. 10) shows the structural relations of the Sierra Nevada to the San Joaquin Valley on the west and to the desert ranges on the east. The east part of the section is through the Darwin and Panamint Butte quadrangles mapped in detail by Wayne E. Hall, and the section was drawn with his help. The cross section is very similar to one drawn by Hopper (1947), with which it is in good agreement.

The major structure shown is a great faulted arch having the Sierra Nevada as its west limb and the desert ranges as its east limb. The southern extension of Owens Valley, and presumably Owens Valley itself, is a graben in the axis of the arch. The sedimentary history in the desert basins and ranges shows that the normal faults that bound the ranges formed after late Pliocene time simultaneously with arching. There is no evidence to suggest that the arch was completely formed before it was faulted, although the faulting may have been confined to the later stages of arching. In the Argus and Panamint Ranges, lower Pleistocene deposits dip generally about 10° to 15° E. parallel with an underlying erosion surface of low relief and subparallel with the Paleozoic strata except where the Paleozoic has been deformed by thrust faults or by intrusive rocks. Middle and upper Pleistocene deposits rest unconformably on the early deposits, and dip east less steeply. Faults that cut across Cenozoic deposits of different ages exhibit successively greater displacements in the older deposits, and show that faulting has been progressive (W. E. Hall, oral communication, 1959). The middle Pleistocene unconformity marks a period of intensified deformation and may coincide with the period of faulting along the Sierra Nevada escarpment that followed the McGee glaciation.

A late Pliocene or early Pleistocene erosion surface of low relief is preserved in many places in the desert ranges beneath sedimentary deposits of early Pleistocene age, and it is assumed that this surface was originally widespread. In the Sierra Nevada the Pliocene and Pleistocene surface has been modified and in many places entirely destroyed by erosion, but was probably only slightly above the present general upland surface.

CENOZOIC STRUCTURAL FEATURES OF THE BISHOP AREA

The structural features formed in late Cenozoic time, which include most of the larger and many of the smaller features of the landscape, are chiefly normal faults and broad warps and open folds related to the uplift of the Sierra Nevada and White Mountains and the correlative subsidence of Owens Valley. The Owens Valley is generally termed a graben, and this designation is correct if boundary slopes that result from warping as well as from faulting are included. Along both the Sierra Nevada and White Mountains escarpments various combinations of warping and faulting occur. The structural relief of the central part of the Sierra Nevada escarpment within the mapped area is principally the product of warping, but elsewhere along Owens Valley faulting predominates. Owens and Round Valleys have subsided as virtually coherent blocks, but nevertheless have suffered internal deformation. This deformation consists chiefly of broad warps and related faults.

Most of these late structural features were recognized by their effect on the topography, and only rarely were internal layers of use in identification. The escarpments along faults are, almost without exception, true fault scarps, though most of them have been modified to varying degrees by erosion. The traces of faults across irregular topography and scarce exposures of fault planes indicate that almost all the faults are normal, although on the Volcanic Tableland a few high-angle reverse faults and a low-angle thrust fault were identified. Warps are reflected in surfaces that have been bent or warped to very different forms from the original. To the casual observer the faults are far more obvious than warps or open folds where both are reflected in the topography, because abrupt fault scarps are more conspicuous than surfaces that have been bent a few degrees. Nevertheless, some of the largest features are warps, and some conspicuous faults are merely secondary features on warps or folds.

In the evolution of the Sierra Nevada and White Mountains escarpments, the two most significant structures are warps and faults in which the block on the side nearest Owens Valley has been relatively dropped. In contrast with the faults that are downfaulted on the valley side, here called valley-down faults, are faults having the mountain side downfaulted, or mountain-down faults. Mountain-down faults generally are subsidiary features that are closely related to warps. The association of mountain-down faults with warps is so general that in places where mountain-down faults are abundant warping might be inferred even in the absence of other evidence.

Mountain-down faults in a similar gross setting have been explained by Hans Cloos (1939, p. 416), on the basis of experimental models and observations in a tunnel that cuts the eastern border fault of the Rhine graben, as minor complementary faults that dip into a master fault on the downthrown side. This explanation does not fit the mountain-down faults of the Owen Valley because a master fault is absent in those areas where mountain-down faults are most abundant. A closer analogy is found in faults on the flanks of anticlinal folds, which strike parallel with the fold axes and are downthrown toward the anticlinal core.

If the warps were formed in less brittle rock or under high confining pressures such as pertain at great depths, the rocks would be expected to bend. Inasmuch as they were formed at the surface, the warping must have been accompanied by breaking along planes perpendicular to the stretching caused by warping. Thus, the warps are composed of a series of discrete blocks, each tilted toward the valley. The planes of separation between blocks in the arrangement are marked by escarpments that face toward the mountains. The fact that the faults are normal implies extension and indicates tension at the surface.

The observed throw on most mountain-down faults is relatively small, rarely exceeding 100 feet. Many of these faults, however, have moved repeatedly, and the throw observed in the dislocation of a surface may represent only a small component of the total throw. Nevertheless, neither the total displacement on a given fault nor the aggregate throw on a series of faults is of any great structural consequence as compared with the difference in structural elevation achieved by warping and accompanying tilting of groups of fault blocks in the same direction.

Narrow grabens a few hundred feet or less in width and a mile or more in length are present at several places along the range fronts, both in alluvial fans and in older rocks. Along some grabens the fault having the larger throw is a valley-down fault, and along others it is a mountain-down fault. In either setting the secondary fault is interpreted to be the result of local tension in the hanging wall of the primary fault.

The downthrown side of most faults is clearly evident in displaced surfaces, but the scarps of valley-down faults appear to erode more rapidly than those of mountain-down faults. Ordinarily, the height of the scarp of a mountain-down fault is diminished by alluviation behind the scarp as well as by erosion of the scarp, but in places where subsequent drainages follow mountain-down faults, the height of the escarpment may be temporarily increased. Among a group of faults of the same age, the fact that more mountain-down faults are likely to be identifiable by their scarps gives an erroneous impression of their relative abundance.

THE SIERRA NEVADA ESCARPMENT

Along the Sierra Nevada escarpment within the mapped area, the broad, gently sloping central span that extends from the Tungsten Hills on the north to the latitude of Crater Mountain on the south, a distance of more than 20 miles, is chiefly the product of warping, whereas the steep scarps to the north and south are chiefly the products of faulting. The central warped span is called the Coyote warp, the northern faulted span is called the Wheeler Crest scarp, and the southern faulted span is called the Tinemaha scarp. Because of the gentle slope of the warp, the range front in the warped span projects farther into the valleys than it does along the more precipitous faulted spans, and narrows the valley by several miles (pls. 1–4, 7). The faulted spans extend from the northern and southern boundaries of the mapped area into the warped span with progressively diminishing throw. The faulted spans are offset from each other and do not join.

THE COYOTE WARP

The Coyote warp has little in common either physiographically or structurally with the precipitous fault scarp that is ordinarily envisaged when thinking of the Sierra Nevada escarpment. Within the area of the warp, broad interfluvial areas between steep-walled canyons slope gently north, northeast, and east into Owens and Round Valleys. These interfluvial areas are parts of an old erosional surface—presumably the mid- to late-Pliocene surface—whose present configuration is the result of structural warping. The old surface is preserved in Coyote Flat, Coyote Ridge, and Table Mountain, and in the sloping interfluvial areas south of Shannon Canyon to Big Pine Creek, and north and west from Freeman Creek, across Bishop Creek and into the Tungsten Hills. The surface is missing over a broad area between Rawson Creek and Shannon Canyon because of deep dissection caused by headward erosion of Rawson, Freeman, and Shannon Creeks. However, the sloping surfaces south of Shannon Canyon and north of Freeman Cayon can be projected across this span without serious adjustments.

The ancient surface that records the warp is deeply weathered and in places has a cover of residual soil. Locally, as in Coyote Flat, it is covered with glacial moraine and outwash. Where the surface is best preserved, it is gently rolling and has a relief of a few hundred feet at right angles to the average slope. Part of this relief is assignable to recent dissection and to dislocations along faults. The local relief increases with dissection, to a point where the old surface is no longer identifiable. Local deep dissection of the old surface by such streams as Bishop Creek, Rawson Creek, Freeman Creek, Baker Creek, Big Pine Creek, and especially Shannon Creek indicates that the surface is not in equilibrium with its present environment and could not have been formed in its present configuration. Almost certainly it was formed with a gentler slope than it now exhibits, and there is no reason to presume any relation between the original configuration of the old erosional surface and its present configuration.

The warp is two sided—one flank slopes northward, and the other flank slopes eastward (pl. 7). The two flanks meet along an anticlinal axis that plunges northeastward and passes through the northeast-facing salient in the range front south of Bishop. The form of the surface is shown by the traces of topographic contours, modified where they cross slightly eroded areas and extrapolated where they cross deeply eroded areas (pl. 7). The 6,000- and 7,000-foot contours are especially illustrative. South of Shannon Canyon the old surface slopes eastward about 6,000 feet in 6 miles, or less than 11°—hardly a precipitous slope (pl. 5, section *D–D′*). On the north flank of the warp the upper slopes from Coyote Flat are at about this same angle, and the lower ones across Grouse Mountain and the Tungsten Hills are even gentler (pl. 5, section *G–G′*). The downwarped nature of the north flank was first recognized by Taylor (1934), who believed that it was bounded on the east by a fault that died out northward.

A significant and interesting feature of the Coyote warp is the abundance of mountain-down normal faults, although valley-down faults are also present. Almost all of these faults are north-trending, and they are mostly confined to the summit and east flank of the warp. The north flank is broken by relatively few faults.

Crest and east flank

A shallow graben bounded by north-trending normal faults of relatively small displacement occupies the topographically low eastern part of Coyote Flat. West of the graben are several nearly parallel faults along which the western part of Coyote Flat has been stepped up. Coyote Flat is believed to be separated from Coyote Ridge by two en echelon faults that are about a mile apart. If Coyote Flat and Coyote Ridge are faulted segments of the same surface, the throw along the northern fault is about 800 feet at the south end and diminishes northward to nothing. These faults are the north end of the fault system that lies along the base of the Tinemaha scarp and bounds the Coyote warp on the west.

East of the graben, subparallel north-trending faults are present eastward down the east flank of the warp to the base of the range and beyond into the alluvial fans. Almost all of these faults are downthrown to the west, but a few are accompanied by faults of opposite throw and define long narrow grabens. Topographically the mountain-down faults are expressed in bedrock by long narrow alluviated trenches parallel with the range front. These are most conspicuous in the broad sloping surface south of Shannon Canyon. Section *D–D′* of plate 5 illustrates the relation of the faults to the deformed surface in the summit and east flank of the warp.

The mountain-down faults between Baker Creek and Shannon Canyon continue southward across Big Pine Creek and into the area west of Crater Mountain. Bedrock is exposed through the moraines of Big Pine and Baker Creeks as a result of faulting. South of Big Pine Creek are several scarps a hundred feet or more in height. The largest of these bounds McMurry Meadows on the east and has acted as a trap for the alluvial detritus there. The height of the scarp indicates a throw of several hundred feet.

A prominent bench at half height in the range front south of Rawson Creek was formed by alluviation behind the composite scarp of a closely connected chain of mountain-down faults. Where the scarp has not been cut through by gullies, a narrow alluviated valley parallel with the range front is preserved. From the floor of Owens Valley the bench gives the erroneous impression of a stepped block bounded by valley-down faults, but careful search of the walls of the canyon of Rawson Creek, where the plane of such a fault would be exposed, proved fruitless even though exposures are excellent. Smaller parallel benches formed by mountain-down faulting are present about a mile to the west in sec. 13, T. 8 S., R. 32 E., and also about a mile to the east near the range front, where they extend about N. 20° W. for nearly a mile from the triangulation station at 5,398 feet just west of Keough Hot Springs.

Mountain-down faults displace alluvial fans along the base of the east flank of the Coyote warp at several places. Profiles across two faults, one cutting the Shannon Creek fan and the other cutting the Rawson Creek fan, are shown in figure 73. Fault scarps that cross the Shannon Creek fan are continuous with fault-controlled trenches in the old surface to the south. The fault represented by *A* and *B* in figure 73 extends south from sec. 20 into the east half of sec. 29, T. 8 S., R. 33 E. The fault was dug into where it crosses the north bank of Shannon Creek, and the dip of the fault plane was found to be 60° W. The block east of the fault has been tilted strongly to the east, and its lower end has obviously been lowered substantial amounts by the tilting. Alluviation has taken place upslope from the scarp, which is about 25 feet high. The faulting and tilting have caused the entrenchment of Shannon Creek into the fan.

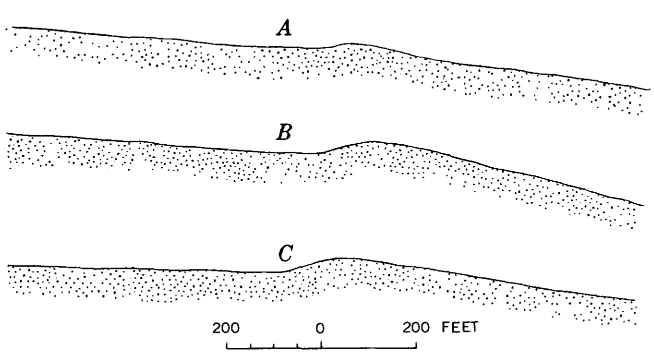

FIGURE 73.—Profiles across mountain-down fault scarps in alluvium along the base of the east flank of the Coyote warp. *A* and *B* are across a scarp in the fan at the mouth of the Shannon Canyon. *A* is in the SE¼ sec. 20, T. 8 S., R. 33 E., and *B* is one-half mile to the south in sec. 29. *C* is across a scarp in the fan of Rawson Creek which extends from the NW¼ sec. 5, T. 8 S., R. 33 E., northward into secs. 31 and 30.

The more easterly of two faults that cross the Rawson Creek fan also bounds a block of bouldery fanglomerate, which is tilted strongly toward the valley. This fault has diverted Rawson Creek, which flows along the face of the scarp and around the end of the tilted block. Several wind gaps in the scarp together with fine alluvial material that has collected west of the scarp as a result of the faulting indicate that the course of the stream has changed several times, presumably as a result of a succession of small movements along the fault. South of the scarp and on line with it, the alluvial slope steepens downslope, undoubtedly as a result of warping, inasmuch as the depositional surface of a fan normally flattens downslope.

Just north of Keough Hot Springs a prominent graben lies along the base of the range. On the west the graben is bounded only by granitic rock, but on the east side it is bounded by a small area of granitic rock that is covered on the north, east, and south sides by dissected alluvial deposits that are tilted several degrees toward the east. A steep contact between these deposits and the underlying granitic rock suggests the probability that, as a result of earlier movements, the granitic rock has been tilted still more than the alluvium. The dissected alluvial deposits at the south end slope about 7° E., whereas adjacent unfaulted alluvial fans slope only 3°. Alluvial fill and water were trapped behind the mountain-down fault and support a flourishing orchard in the north end of the graben. Similar situations exist in many places along both sides of Owens Valley in connection with mountain-down faults that cut alluvial fans. The graben may well continue northward in bedrock along the range front and west of the Rossi tungsten mine as suggested by the topography, but the valley-down fault on the west side of the graben was not identified along this span.

Faults of large displacement west of Big Pine

Although the north and east flanks of the Coyote warp north of Shannon Canyon are known to continue into the valley block beneath alluvial fill (pl. 5, sections *C–C′*, *G–G′*), south of Shannon Canyon the downdip part of the warp is cut by several faults of large displacement, including at least two valley-down faults. At Keough Hot Springs a valley-down fault displaces all but the very youngest fan deposits along the range front, and also provides a conduit for the water of the hot springs which break out at the lowest exposure of the fault. This fault ends just north of Keough Hot Springs about a quarter of a mile east of the valley-down fault of the graben.

The fault can be traced in alluvium only about half a mile south of Keough Hot Springs, but it probably continues southward in bedrock, and very likely is the valley-down fault east of Warren Bench; a valley-down fault along the intervening span is suggested by the steep escarpment. The fault at Warren Bench is marked by an eroded escarpment in granitic rock. The fault probably passes along the east side of the outcrops of metamorphic rock between Big Pine and Baker Creeks to an escarpment in granitic rock along the west side of the Crater Mountain basalt field. The rhyolite dome 3 miles south of Crater Mountain may mark this fault. The height of the scarp opposite Crater Mountain indicates a throw of about 800 feet. No scarps in alluvium are present along this fault south of Freeman Creek, but the moraines along Big Pine and Baker Creeks terminate along the line of the fault and have no doubt been downdropped along it.

East of this fault is a mountain-down fault along which the feeders for Red Mountain and Crater Mountain are localized. In bedrock this fault bounds the west side of a horst that projects through the basalt flows of Crater Mountain in Fish Springs Hill and in small inliers of Tinemaha granodiorite in the northern flank of Crater Mountain. A recent scarp is present in the alluvial fans between Red Mountain and the south flank of Crater Mountain, and an alined scarp in the alluvial fan of Big Pine Creek probably is along the same fault. Orchards are present in alluvium collected against these scarps along Big Pine Creek and north of Tinemaha Creek.

West of Big Pine the alluvial fan is displaced along a valley-down fault having a very large aggregate displacement (pl. 5, sections *D–D′*, *F–F′*). Gravity measurements and strong movements at the time of the earthquake of 1872 indicate that this fault extends southward for at least 100 miles and separates shallow bedrock in the west side of Owens Valley from deep bedrock in the east side. Undoubtedly this is one of the major Cenozoic faults of the region, and is a major step fault in the downdropping of Owens Valley. The fault can be followed south of Big Pine along a scarp that crosses the east side of the basalt field of Crater Mountain. At two places the throw along this scarp is reversed, but in each place companion faults to the west are downthrown to the east and the net throw on the pairs of faults is the same as elsewhere along the scarp. This anomalous situation probably reflects local near-surface deflection of the principal fault plane to the positions of the companion faults; the segments having reverse throw along the line of the main scarp are merely secondary fractures.

In the south margin of Crater Mountain the fault splits, one branch bends eastward toward the east side of the Poverty Hills, and the other branch bends westward to the west side of the Poverty Hills, where with a companion mountain-down fault it forms a narrow graben. A small cinder cone along the west branch between Crater Mountain and the Poverty Hills has been displaced by the most recent faulting, which here has downdropped the fan on the east side of the fault about 15 feet.

Movements in this area at the time of the earthquake of 1872 were described by Whitney (in Goodyear, 1888, p. 295).

> * * * nowhere are the effects of the earthquake in fissuring and depressing the surface so manifest as in the vicinity of Big Pine. A large body of water issues from the gorges of the Sierra west of this place, and this water spreads out after leaving the sagebrush slope, and runs in numerous channels through a low and swampy meadow, several hundred acres in extent. Here there is a series of extensive fissures, which may be traced uninterruptedly for several miles. In one place an area of ground, two or three hundred feet wide, has sunk to the depth of twenty or thirty feet in places, leaving vertical walls on each side, and these depressions have become partly filled with water, so that ponds have been formed of no inconsiderable size. One noticed was fully one-third of a mile in length, and would have been much larger had not the depression been so situated as to afford partial drainage of the area at one end, so that the basin could not be entirely filled.

North flank

The north flank of the Coyote warp, unlike the east flank, is broken by relatively few faults. Those that are present fall into the following three groups: (1) east-trending faults along the range front south and southwest of Bishop, (2) northeast- and northwest-trending faults in the vicinity of Lookout Mountain, and (3) northeast- and northwest-trending faults in Grouse Mountain and along the west side of the Tungsten Hills. Although additional unrecognized faults may be present, faults having significant recent throw could hardly have been overlooked.

East-trending faults were observed southwest of Bishop, and one of these faults contains the vein that has been worked in the Bishop Antimony mine. Faults downthrown either to the east or to the west are present, but the dips of most fault planes were not ascertained. The fault exposed in the Bishop Antimony mine is downthrown to the north and is nearly vertical. Minor slips within the fault zone dip north or south. The mapped faults are all short and fall within an area of only about a square mile, but the physiography of the range front between Bishop Creek and the Rossi mine suggests that the Coyote warp may be splintered by two or three east-trending northwesterly en echelon faults that die out toward the west. One fault probably lies along the edge of bedrock from the Bishop Antimony mine westward about 2 miles into the lower part of the canyon in which the Chipmunk mine is located. A second fault almost certainly lies along the north edge of the spur that contains the Yaney mine, particularly since the ore in the mine has been altered by hot spring action. And a third fault may lie along the northern edge of the next spur to the northwest.

In the vicinity of Lookout Mountain only two faults have physiographic expression, and some or all of the others may be relatively old features; some of those in metamorphic rock may possibly be of prebatholith age. The two faults having physiographic expression are northeast of Lookout Mountain. These faults trend northeast and each are downthrown to the southeast several feet.

The faults in Grouse Mountain and along the west side of the Tungsten Hills separate the Coyote warp from the southerly extension of Round Valley. The northwest-trending faults lie in and adjacent to Grouse Mountain. The most easterly of these faults appears to be downthrown to the northeast, and two other faults are downthrown to the southwest. Displacement on the easterly fault probably is minor. Repeated movements along the faults downthrown to the southwest are indicated by progressively greater displacement in older moraines and outwash, and obviously greater throw in bedrock than in the moraines. The Tioga moraines, however, are not cut by these faults.

The northeast-trending faults comprise two en echelon segments along the west side of the Tungsten Hills. The more northerly fault continues northeast of the Tungsten Hills, and recent movements along it have displaced the surface of the Volcanic Tableland and faulted off the terraces on the north side of the Tungsten Hills. A succession of movements is shown by the difference between the throw required to cause the escarpment along the west side of the Tungsten Hills and the throw in the Volcanic Tableland. Recurrent movements along the southern fault segment have caused both McGee Creek and Horton Creek to change their courses at least twice. Both streams have been forced from original courses directly across the Tungsten Hills—McGee Creek into a course farther to the north, and Horton Creek into a northward course into Round Valley.

WHEELER CREST SCARP

The Wheeler Crest scarp is one of the highest and most precipitous escarpments along the east face of the Sierra Nevada. The base of the scarp in the span north of Pine Creek is at altitudes of 5,200 to 6,200 feet at the head of the alluvial slope, and summit altitudes about 2 miles to the west range from a little less than 11,000 feet to a little less than 13,000 feet. The average slope angle thus is about 30°, although in places it is as much as 35°. Profiles across the slope, following ridge crests, are approximately straight lines; neither the upper part nor the lower part of the escarpment is systematically steeper than the other. South of Pine Creek, the escarpment is less precipitous, and because the base rises to the south in the Coyote warp, the height progressively diminishes southward. The progressively lesser height of the scarp reflects, of course, progressively smaller throw on the faults.

Along most parts of the escarpment, valley-down faults can be identified along the base. North of Pine Creek an almost continuous line of faults was mapped, but south of Pine Creek some segments lack mapped faults; thus the south part of the scarp may not have been as completely involved in recent faulting as the north part. High on the east side of Wheeler crest and in the lower eastern slopes of Mount Tom, mountain-down faults occur, but elsewhere the escarpment itself appears unbroken.

The fault zone along the base of the escarpment is made up of interconnecting straight or gently curved segments. The fact that fault segments follow every reentrant in the base of the scarp indicates that, even though the scarp is deeply eroded, the base has not retreated. Although actual fault planes were observed in only a few places, the positions of the fault segments can be established by other means. The most useful criterion was low scarps that offset alluvial fans, talus cones, or moraines. Almost as useful are lines of springs. Not uncommonly the bedrock adjacent to fault segments is sheeted parallel to the fault plane, and sheeted zones were taken to indicate faulting in a few places where other clear evidence of faulting was lacking.

The full width of a fault zone was observed at only one place, in the wall of an entrenched gulley a few hundred feet west of the west quarter corner, sec. 13, T. 6 S., R. 30 E. At this locality the fault lies within Wheeler Crest quartz monzonite and is a brecciated and mylonitized zone about 20 feet wide; sheeting in the quartz monzonite west of the fault dips 70° E. In other places along the base of Wheeler Crest, sheeting dips from 60° to 65° E.

A branch of the fault system along the base of the range north of Pine Creek probably follows into the range along Pine Creek, and joins a fault along the west side of the thick dike of quartz monzonite that penetrates southward from Pine Creek into the Pine

Creek pendant. This fault can be traced south from Pine Creek into Horton Lake by means of a low scarp so recent that it offsets talus cones. An exposure of the fault plane at the summit of the escarpment on the north side of Horton Creek dips 65° E.

At Wells Meadow a small alluvial-filled graben is present at the base of the scarp. On the east the graben is bounded by a horst of granitic rock. Apparently the graben and horst lie between overlapping segments of the principal valley-down fault.

Mountain-down faults, so abundant in the Coyote warp, were identified by their effect on the topography at only two places along the Wheeler Crest scarp—on the lower eastern slope of Mount Tom, and along the east side of the summit surface of Wheeler Crest. The mountain-down faults in the lower east slope of Mount Tom are most conspicuous in the broad ridge south of Elderberry Canyon, but they extend northwest onto the ridge just south of Pine Creek. South of Elderberry Canyon these faults are subparallel with the contours and cut metamorphic and granitic rocks, and some also cut fan surfaces. Each of the faults is marked by a small valley or bench that runs along the slope (fig. 74). These valleys and benches are caused by alluviation behind the scarps. The traces of the fault planes across the topography indicate that most of them dip west at angles close to 45°. These faults are in a salient in the range front that is bounded along the base by valley-down faults, and may result from the downward drag of the salient.

The mountain-down fault on the east side of Wheeler Crest furnishes a smooth bench which makes for relatively easy access for about 4 miles along the very crest of the Wheeler Crest scarp. The fault has resulted also in alluviation behind the fault scarp at the heads of several steep canyons. One alluviated canyon head is north of the Adamson Round Valley Peak tungsten deposit, a second is about 2 miles farther north just south of peak 11,498, and a third is about three quarters of a mile south of the north boundary of the Mount Tom quadrangle. The northernmost alluviated canyon head contains a pond or small lake, and the others probably also contained lakes in the recent past.

The position of the valley-down faults immediately at the base of the escarpment indicates that dislocation has continued to the present. Had it not, the base of the escarpment would have retreated from the position of the bounding faults, especially along larger streams. The dislocations of slope surfaces by recent movement on the faults also indicate that dislocation has continued, except that they might be dismissed as minor movements that followed the main movements after a long period of stability.

TINEMAHA SCARP

The span of the Sierra Nevada escarpment that is here designated the Tinemaha scarp is the north end of a scarp that continues to the south many miles, although in several places it is offset. In the mapped area, the scarp can be traced for about 8 miles, from Taboose

FIGURE 74.—Bench on northeast side of Mount Tom formed by mountain-down faulting parallel with the range front.

Creek at the south edge of the Big Pine quadrangle northwestward into the south side of Kid Mountain. The northwest half of this span, from Tinemaha Creek northwestward, lies west of the Coyote warp and progressively diminishes in height into the warp. The escarpment is about as steep as the Wheeler Crest scarp south of Pine Creek, which has a similar structural setting. Farther north, across Big Pine Creek, the more southerly and easterly of two east-facing escarpments that bound Coyote Flat on the west is alined with the Tinemaha scarp, and no doubt reflects northward continuation of the faulting that produced the Tinemaha scarp.

The fault pattern is similar to the pattern along the base of the Wheeler Crest scarp and consists of an interconnecting system of curved and straight-line segments. The height of the scarp is certainly chiefly the product of displacement along valley-down faults at the base of the scarp. However, south of Red Mountain Creek both valley-down and mountain-down faults were mapped along the face of the escarpment to altitudes as high as 10,000 feet. Some of these faults may continue farther to the northwest, but were not recognized because of the greater recent dissection of the scarp to the northwest.

The principal valley-down fault dips 70° E. where it is exposed along the east face of Birch Mountain, and 45° E. in the lower east slope of Mount Tinemaha as indicated by the trace across talus. Recent movement on most of the faults is shown by physiographic displacements. At Tinemaha and Birch Creeks, faults displace moraines of both Tahoe and Tioga age, and in Birch Creek all but the most recent stream deposits are displaced.

The faults in the escarpment south of Red Mountain Creek are shown by physiographic dislocations in the slope. The faults here fall into two groups; those between the base of the escarpment and the east side of Stecker Flat are all mountain-down faults, whereas those farther west and higher are valley-down faults. Inasmuch as the traces of the faults across irregular topography indicate that the mountain-down faults dip gently to the west and the valley-down faults dip moderately to the east, all the faults are normal. The displacement on most of them is believed small, but the highest and most western fault at the upper edge of Stecker Flat may have a throw of several thousand feet. The fault pattern here is similar to the pattern along much of the White Mountain escarpment within the mapped area in that a broad zone that includes both valley-down and mountain-down faults is present in the mountain block behind a principal valley-down fault. The presence of mountain-down faults suggests that warping as well as faulting was involved in the production of the escarpment. The warping may be caused by drag along the principal valley-down fault. The probable structure of this part of the range front is shown in section *F–F′*, plate 5.

CONJUGATE JOINT SYSTEM

A regional system of steeply dipping conjugate joints is present almost everywhere in the granitic rocks of the Sierra Nevada. The joints are most conspicuous in areas of low topographic relief where the rock adjacent to joints is subject to rapid weathering, chiefly by chemical action and hydration. Erosion of the weathered rock has resulted in linear depressions along the joints. At low altitudes, erosion has been chiefly by wind and water, and at high altitudes glaciation has also been effective (fig. 75). Joints locally exert considerable control over the drainage pattern, and even in deeply dissected regions straight segments of streams coincide with joints.

The steeply dipping joints occur in all the granitic rocks and cross boundaries between different intrusive masses without obvious deflection (pl. 7). Generally two principal joint sets almost at right angles can be identified. One of these strikes northwestward and the other northeastward; the precise direction of strike changes from place to place although the two sets maintain their right-angle relation to one another.

Individual steep joints can be traced for several miles. Most are straight or gently curved. The joints in a given set generally are subparallel, and not uncommonly one joint cuts across other joints only slightly different in strike. Where significant change takes place in the strike of the sets, joints of one strike generally interfinger with joints of slightly different strike, and rarely does a single joint curve from one direction to another (pl. 7).

The spacing of the steep joints ranges from an inch or less to many feet. Regional differences in the spacing undoubtedly reflect differences in the intensity of the regional stresses causing the joints. The paucity of joints in the Yosemite region, which makes possible the formation of magnificent topographic domes, probably reflects less intense stress than existed in the Bishop area. The spacing within a given broad area, however, appears to be primarily a function of the grain size of the rock in which the joints occur: The widest spacing is in the coarsest grained rock, as is illustrated in figure 76, which shows joints that cross an aplite dike within the Tungsten Hills quartz monzonite. The joints in the aplite are much more closely spaced than those in the quartz monzonite. The most conspicuous joints are the widest spaced; where the joints are closely spaced, the rock breaks into rubble and does not permit the

FIGURE 75.—Aerial view of Sierra crest southwest of Bishop; Blue Lake in center foreground. Grid pattern in right foreground results from the influence on erosion by a conjugate system of steeply dipping joints. Photo by Symons Flying Service.

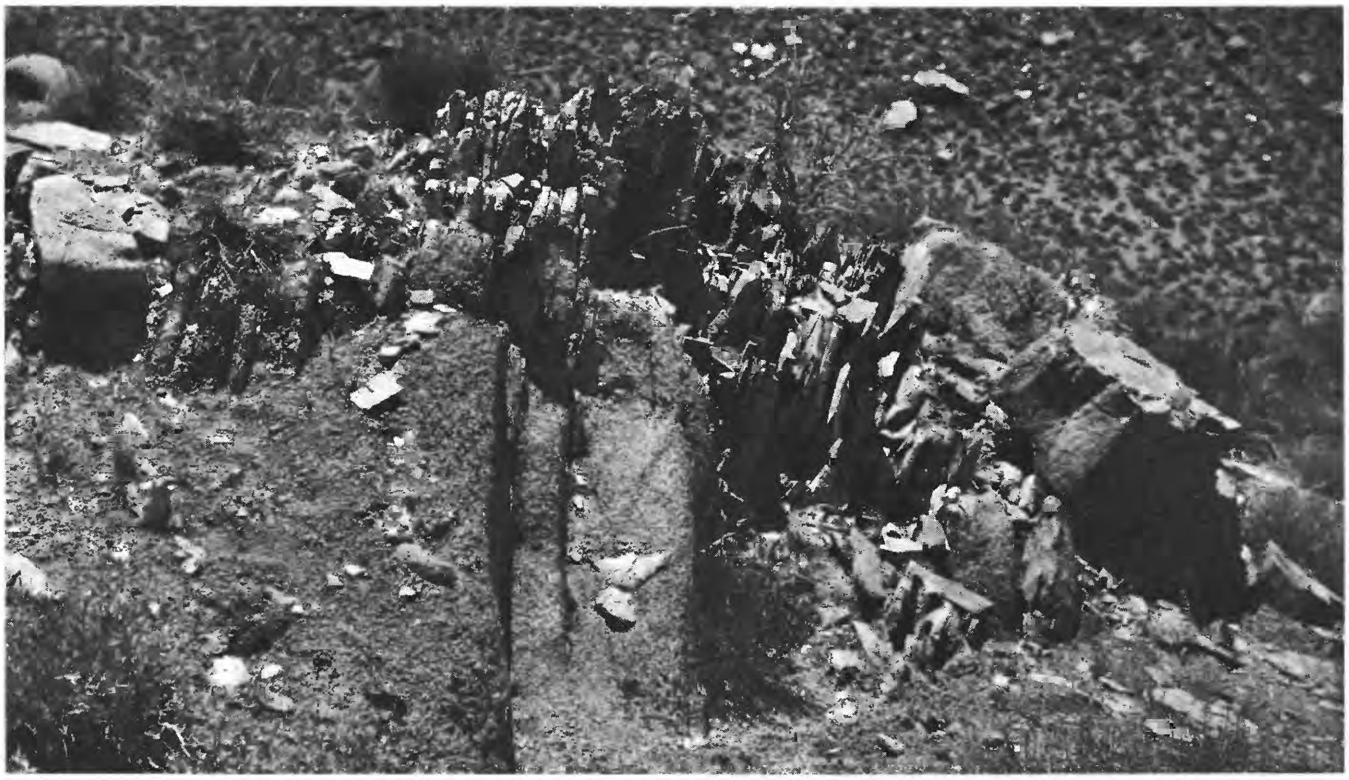

FIGURE 76.—Joints crossing aplite dike in Tungsten Hills quartz monzonite. Closer spacing of joints in aplite is a function of its finer grain size.

formation of deep and continuous linear depressions or trenches along individual joints.

Most steep joints exhibit no evidence of movement, but a few offset dikes as much as several inches. The movement along these joints was parallel to the joint. The fact that the joints continue without interruption across boundaries between intrusive masses and that the pattern is regional indicates that the joints are younger than the batholith. It seems likely that the joints are the product of regional forces. The principal stress was very likely horizontal, and the joints can therefore be regarded as incipient shears. Because displacements along the joints were not systematically recorded, the stress pattern cannot be analyzed.

Gently dipping joints are conspicuous at higher altitudes where frost action has opened them. In most places they strike parallel to the prevailing direction of the topographic contours and dip a trifle less steeply than the ground surface but in the same direction. This relation between the orientation of the gently dipping joints and the average local slope of the surface suggests genetic connection. The most likely explanation is that the joints were formed as a result of unloading through erosion. Any movement along the gently dipping joints would have been normal to the joint surfaces and not parallel with it as is inferred for the steep joints.

THE WHITE MOUNTAINS ESCARPMENT

The western side of the White Mountains within the mapped area slopes gently toward Owens Valley, although farther north at the latitude of White Mountain Peak the range front is very precipitous. Numerous faults having both valley-down and mountain-down displacements are present in the lower slopes of the range in a belt that averages about a mile wide. In this belt, mountain-down faults are at least twice as abundant as valley-down faults but have much smaller throws and are structurally significant because they indicate warping and mark the boundaries of tilted blocks. Older than the range-front faults and lacking recent scarps are three faults between Poleta and Black Canyons, which are transverse to the range front. The middle and southern faults are cut off by range-front faults, but the northern fault may bend near the range and join a range-front fault beneath alluvium.

STRUCTURES PARALLEL WITH THE RANGE FRONT

Near the south boundary of the Bishop quadrangle, in the vicinity of Black Canyon, almost all of the faults parallel with the range front are downthrown on the valley side, and most of the displacement appears to have been on a single fault. Elsewhere, valley-down faults and tilted blocks combine to give the visible structural relief along the range front.

Several valley-down faults that bound stepped blocks 3 to 6 miles long can be identified. One such fault extends in fan deposits just west of the base of the crystalline rocks, from Coldwater Canyon southward to about a mile north of Silver Canyon; a second extends from Silver Canyon south to Poleta Canyon, and a third lies along the west side of the older dissected fan deposits from Poleta Canyon to Black Canyon. Mountain-down faults generally are much shorter and most can be traced less than a mile. One of the most prominent mountain-down faults crosses the alluvial deposits at the mouth of Poleta Canyon in the north-central part of sec. 13, T. 7 S., R. 33 E. This fault lies on the east side of a tilted block and is marked by an east-facing scarp almost 100 feet high. The beds in the tilted block west of the fault dip about 15° W., whereas those east of the fault dip only about 6° W. The 15° dip is not exceptionally steep—beds exposed in a pumice pit in the SE¼ sec. 14, T. 6 S., R. 33 E., dip 45° to 55° toward the valley. The 15° angle of the dip in the tilted block carries the beds downward at a rate sufficiently great to offset the mountain-down displacement on the bounding fault within a distance of about 600 feet. Beyond that distance the beds are structurally depressed relative to their position before they were tilted.

Another illustrative fault, a hinge fault, extends southward from the mouth of Coldwater Canyon. The scarp at the north end of this fault faces the mountains and that at the south end faces the valley. The height of the scarps diminish toward the hinge, and they meet in a small area of no displacement. The area west of the fault was dropped at the south end on the valley-down segment of the fault and at the north end by the tilted block west of the mountain-down scarp.

Recurring movement is shown in the part of the scarp that faces the mountains. Two periods of dislocation can be identified in the scarp both to the north and to the south of the stream that flows from Coldwater Canyon. Near the stream the fan deposits are broken by a low, very slightly eroded scarp; farther away from the stream they are broken by a higher and strongly eroded scarp. The most recent fan deposits immediately adjacent to the stream are unfaulted. In other places, comparison of the dissection in different tilted blocks and different parts of the same blocks suggests two or more periods of movement.

The mapped structures taken alone lead to the supposition that the belt of faults and tilted blocks along the range front are but part of a system of faults of distributive deformation, which continues beneath the valley floor. The aggregate deformation of the bedrock surface can be presumed to be far greater than the deformation in the exposed rocks, many of which are fan or lake deposits and record only very recent movements. No single fault was recognized through geologic mapping which alone would explain a major part of the structural relief.

However, gravity studies by L. C. Pakiser and M. F. Kane of the U.S. Geological Survey indicate the presence of a steep buried escarpment about 5,000 feet high along the west side of the exposed belt of deformation. The westernmost mapped valley-down fault between Silver and Black Canyons probably marks the projection of this escarpment to the surface. The most reasonable interpretation of the combined gravity and geologic data is that the concealed escarpment is a steep master fault and that the exposed faults and tilted blocks are subsidiary structures. According to this interpretation, all or almost all of the structural relief of the concealed part of the range front and probably also part of that in the exposed escarpment is a result of movement on the master fault. Nevertheless, the mapped structures must explain a substantial part of the structural relief within the exposed part of the range front. The situation here is somewhat similar to that along the southern part of the Tinemaha scarp within the mapped area and also to that along the southern part of the east flank of the Coyote warp opposite Big Pine and southward. Sections *A–A′* and *C–C′* (pl. 5) illustrate the structure of the range front.

TRANSVERSE FAULTS BETWEEN POLETA AND BLACK CANYONS

The northernmost transverse fault follows Poleta Canyon and passes just north of the Poleta stock; the next fault to the south lies in the northern margin of secs. 19, 20, and 21, T. 7 S., R. 34 E., and the most southerly fault follows closely the south boundary of sec. 21. All three faults continue to the east beyond the limits of the quadrangle, and to the west are overlapped by alluvium. The two southern faults strike almost due east, but the fault along Poleta Canyon strikes northeastward. All the faults are normal faults and are downthrown to the north. Two other smaller faults that may belong to this group are north of Poleta Canyon in sec. 4, T. 7 S., R. 34 E. These faults strike about N. 35° E., and are downthrown on the northwest.

The transverse faults explain the exposure south of Poleta Canyon of the formations beneath the Campito formation. Along the southern two faults, Reed dolomite on the south has been brought into contact with the Andrews Mountain sandstone member of the Campito formation on the north; indicating a stratigraphic displacement on each fault in excess of the thickness of the Deep Spring formation (1,500 feet). Along the

Poleta fault the Andrews Mountain member on the south is in contact with the Montenegro member and the Poleta formation on the north—relations that provide little basis for calculating the stratigraphic displacement. Nevertheless, the gross distribution of formations in the vicinity of Poleta Canyon suggests that the movement along the Poleta fault has been at least as great as along the other two faults and that it may have been much greater.

Lateral movement of about 1 mile is suggested on the Poleta fault by the fact that slate having excellent cleavage occurs north of the fault adjacent to the Poleta stock, whereas spotted hornfels derived from similar slate by thermal metamorphism is exposed on the north side of the fault about 1 mile west of the stock. These relations also indicate that the principal movement on the fault was after the emplacement of the stock in Poleta Canyon. Lateral movement along the southern two faults is a distinct possibility, but no criteria were found to identify such movement except for a faint suggestion of drag in the Deep Spring formation on the south side of the middle fault.

STRUCTURES IN THE VALLEY BLOCK

The block referred to here as the valley block lies between the Sierra Nevada and the White Mountains, and includes Owens Valley, Round Valley, the Volcanic Tableland, and the river terraces south and east of the Volcanic Tableland. Each of these physiographic units contains structures of interest in themselves, and in the aggregate they provide a basis for interpreting the overall structure of the relatively depressed intermontane block. To augment the geologic data a geophysical study of the variation in gravity in the intermontane area was made by L. C. Pakiser and M. F. Kane of the U.S. Geological Survey, and is reported herein. A seismic profile across Owens Valley is also described by Mr. Pakiser.

STRUCTURES IN THE VOLCANIC TABLELAND

Within the mapped area the surface of the Volcanic Tableland almost everywhere parallels internal layering in the Bishop tuff. The central and eastern parts have undergone relatively little erosion, and the surface provides a reference plane that is sensitive to deformation and resistant to erosion; fault scarps and broad warps are remarkably preserved. The western part has been more deeply eroded, and all but the largest features of the original surface were destroyed by streams that headed in the Sierra Nevada. The most conspicuous features of the landscape are mounds and ridges of resistant tuff that has been indurated by gas action. The following discussion deals mainly with the structures preserved in the central and eastern parts of the Volcanic Tableland.

The largest structure on the Tableland is a broad asymmetric arch that plunges S. 30° E.; the crest lies 1 to 2 miles east of the Owens River Gorge. The east flank of this arch slopes southeastward toward Owens Valley and the southwest flank slopes southwestward toward Round Valley.

Superimposed on this broad arch are many minor warps and faults. In the central and eastern part of the Tableland the most conspicuous structures are steep, fresh-appearing, north-trending scarps (fig. 77). The largest scarp, along the east side of Fish Slough, is more than 5 miles long and consists of two en echelon segments. A few other faults are as much as 3 miles long, but most are less than a mile long. The scarp on the east side of Fish Slough is more than 300 feet high locally, and a few others are 100 to 200 feet high, but most scarps are less than 50 feet high at the center and progressively lower toward the ends. Alluviation has taken place locally in basinlike depressions at the bases of some of the scarps, but except in Fish Slough the amount of alluvium deposited is generally negligible. Thus, the height of most scarps is a measure of the throw on the fault since the extrusion of the Bishop tuff.

Scarps downdropped to the east and to the west are equally abundant, and are of the same average height and length. Most fault planes exposed in cliff faces, road cuts, and pumice quarries along the south and east margins of the Volcanic Tableland are vertical or dip steeply toward the downthrown side, but a few dip toward the upthrown side. Most of the faults having larger throws appear to be vertical or nearly vertical, whereas some minor faults dip at angles as low as 55°. Many of the faults can be observed to pass downward through the Bishop tuff and into the underlying alluvial deposits without recognizable change in throw.

The two en echelon segments of the fault along Fish Slough are offset about half a mile. As the southern segment dies out at the north end, the displacement increases on the northern segment, which is offset to the west. Although the scarp attains a maximum height of about 300 feet, the downthrown side is covered with alluvial fill, and the throw on the fault may be 400 feet or more. A gravity high in the part of the Volcanic Tableland east of Fish Slough was noted by Pakiser and Kane. If caused by upfaulting of the basement, throw of several thousand feet on both sides of a horst is indicated, and basement rock must lie very close to the surface.

In the lower end of Fish Slough, in the W½ sec. 7 and NW¼ sec. 18, T. 6 S., R. 33 E., three tilted blocks lie

FIGURE 77.—Aerial view of the southeast part of the Volcanic Tableland showing systems of en echelon faults. Fish Slough extends across the lower right corner of photograph. Photo by Roland von Huene.

along the fault (fig. 78). All these blocks have been so tilted that the south end of each block is lowest and the north end highest. The north ends of the two northernmost blocks are not quite so high as the upthrown side of the scarp, but the north end of the south block is even higher. Because the lower ends of all the blocks are covered by alluvial fill, the relation of the lower ends to the downthrown side was not established. The positions of these blocks can be most readily explained if oblique slip movement on the fault is assumed, with the east side moving relatively upward and to the south. The amount of lateral movement, at least since the deposition of the tuff, probably is commensurate with the amount that is represented in the warp that exists at the offset of the two segments of the main fault along Fish Slough.

Just east of the Fish Slough scarp the upper surface of the Tableland is broken by a closely spaced group of faults, some of which are downthrown to the east and some to the west. Farther east the surface has been beveled by stream action, which has effectively destroyed any scarps that may have existed. Nevertheless, the surfaces of many faults are exposed in pumice pits along the eastern escarpment of the Tableland.

Most of the faults are vertical or high-angle normal faults, but a few are high-angle reverse faults, and one is a low-angle thrust fault. Slickensides identified on about 25 percent of the faults suggest that the normal faults and most of the vertical ones were formed by dip-slip displacement but that the reverse faults and some vertical faults were formed by oblique-slip movement. Both the normal and reverse faults range in strike from a few degrees east of north to a few degrees west of north, but there are no obvious differences in the average strike of the two kinds.

The single thrust fault strikes N. 75° W., almost at right angles to the other faults, and dips 30° N. About half a mile to the northwest of the thrust fault, pumice identical in appearance with the pumice in the base of the Bishop tuff rests on unconsolidated tuff. The pumice outcrop is too small and exposures are too poor to arrive at any definite conclusion about this anomalous outcrop, but it is probably part of the basal pumice layer carried to its present position along a thrust. The stratigraphic throw required would be only about 50 feet, and the dip-slip displacement, assuming a 30° dip, would be about 100 feet.

Most of the faults west of Fish Slough are arranged in northwest-trending en echelon systems. All the faults in a system are downthrown in the same direction, and systems in which the faults are downthrown to the west alternate with systems in which the faults are downthrown to the east. Several of these systems can be readily identified on the maps (pls. 1–4, 7), but even where these systems are not readily apparent, as in the south-central part of the Tableland where the faults are very closely spaced, en echelon systems can be shown to exist by plotting the east-facing and west-facing scarps separately. Although the average trend of the faults is about due north, the faults range in strike from northwest to northeast. In places, faults having opposing directions of throw make grabens or horsts, but most blocks are tilted and upfaulted on only one side. In a few places the throw on a fault dimi-

FIGURE 78.—Tilted fault blocks along Fish Slough. Surface on ridge in middle distance, on tilted blocks in front of ridge, and in foreground, was once continuous and unbroken.

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nishes along the strike to no displacement, then increases with the throw in the opposite direction. In such faults the segments with opposing throw belong to adjacent systems.

Here it is evident that two systems of en echelon faults whose scarps face inward define a structural low, and that two systems that face outward define a structural high. Along the lines of the structural lows, several undrained basins have been formed in the surface, whereas the topographically highest areas are in the structural highs. The structural highs can be considered to mark the axes of anticlinal warps and the structural lows to mark the axes of synclinal warps (pl. 7). The broad inconspicuous anticlinal and synclinal warps are really the primary structures here and the fault systems are merely secondary features of contemporaneous origin that help define the warps.

To understand the deformation pattern better, the faults in part of the area were traced onto a sheet of paper and the lines showing the faults were slit. When the paper was pushed in a direction normal to the fold axes, folds formed but the slits remained closed. However, when the east side of the sheet was moved south relative to the west side, both folds and faults formed in a pattern that is in agreement with the field observations; the fact that a rotational couple was involved is thus indicated (fig 79). The movement on the faults was dip-slip even though it is the product of a couple. This experiment is concerned only with the duplication in a reference sheet of the geometry of the surface of the Bishop tuff; it is not a scale model, which would entail use of material of appropriate strength and reproduction of the structural features without the initial aid of slits for the faults. The geometrical approach is believed to be valid because the structures were formed at the surface in brittle rock without the confining pressures that would be required for plastic flow in the tuff.

En echelon faults are commonly ascribed to a rotational couple, but so far as I have been able to deter-

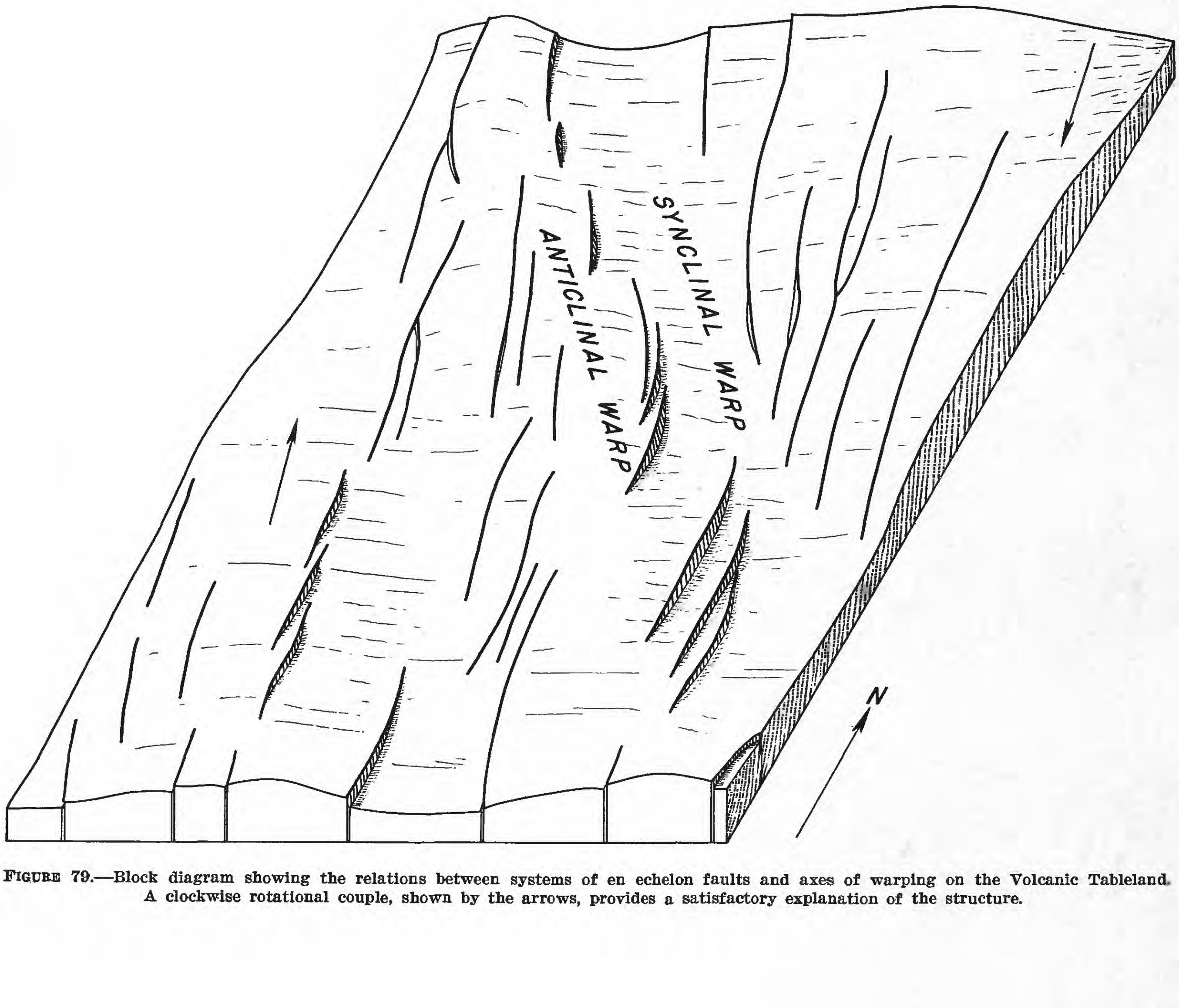

FIGURE 79.—Block diagram showing the relations between systems of en echelon faults and axes of warping on the Volcanic Tableland. A clockwise rotational couple, shown by the arrows, provides a satisfactory explanation of the structure.

mine from a cursory search of the literature the precise situation here has not previously been described. In the extended controversy during the 1920's over the cause of the en echelon faults in east-central Oklahoma several explanations were advanced, but only Sherrill's (1929) suggestion of regional torsion augmented by slight uplift is very much like the mechanism suggested for the en echelon fault systems in the Bishop tuff. Fath (1920) and Foley (1926) postulated horizontal shifts in the basement beneath each en echelon system. Link (1929) suggested tension in the crest of folds that were produced by compaction over buried ridges in the basement. The Oklahoma faults extend over a much larger area than is represented by the Volcanic Tableland, and the fault systems are longer, but the individual faults are of about the same average length and the range in throw is about the same. The Oklahoma faults are downthrown both to the east and to the west, but no statement was found in the literature to indicate that all the faults in an en echelon system are downthrown in the same direction, or that adjacent systems are composed of faults of opposing throw.

The few eroded fault scarps in the western part of the Volcanic Tableland trend either northeast or northwest. These directions parallel the two principal joint sets in adjacent parts of the Sierra Nevada, and joints in the basement may have controlled the pattern of faulting. The only fresh scarp among this group of faults is one along the northeast-trending fault that is alined with the west side of the Tungsten Hills.

Before summarizing the data on the deformation of the Volcanic Tableland, it is necessary to inquire into the relation of the present configuration of the Volcanic Tableland to its initial configuration. The best clue to the initial configuration of the surface is found in three ancient stream channels that run parallel and in a southeasterly direction (fig. 80). One includes the anomalous drainage of Rock Creek through Birchim Canyon and the Owens River below Birchim Canyon to the south edge of the Volcanic Tableland. The second is represented only by a short drainage in the south margin of the Tableland in secs. 19 and 20, T. 6 S., R. 32 E. The third is represented by a once continuous channel that extends from sec. 22, T. 6 S., R. 32 E., where the channel intersects the south margin of the Tableland, northwestward for about 5 miles to the SE¼ sec. 12, T. 6 S., R. 31 E. This channel, which probably was formerly occupied by the Owens River, has been broken into segments by faults, and some segments are lifted high above others (fig. 77). All three channels were probably formed by consequent streams that flowed down a southeastward sloping initial surface. This conclusion is in agreement with other considerations, such as the probable source of the Bishop tuff and its probable direction of flow. If the original slope of the Tableland was to the southeast, the southwesterly

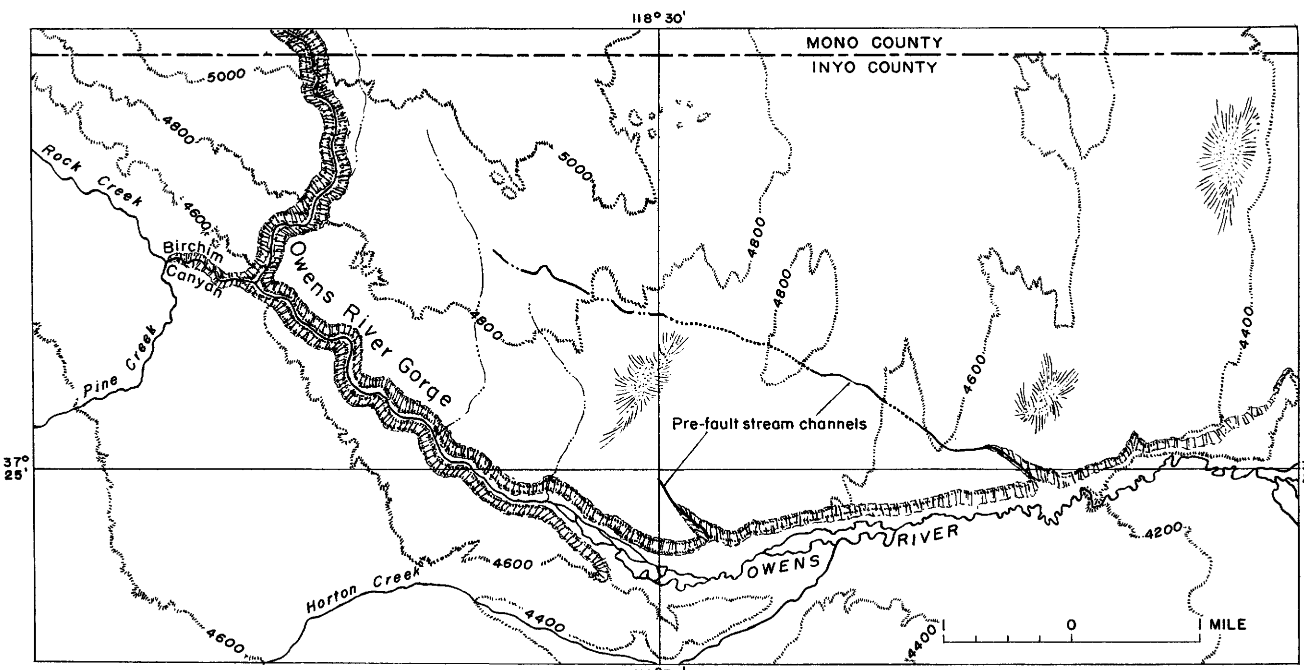

FIGURE 80.—Map of the southwestern part of the Volcanic Tableland showing the location of three ancient stream channels and their relation to modern topography. One channel, Birchim Canyon and the lower part of the Owens River Gorge, is occupied by the Owens River. The other two have been disrupted by faulting and warping and are dry.

slope in the vicinity of Round Valley must then be the result of later warping.

Both the northwest-trending folds and associated systems of en echelon faults and the tilted blocks along Fish Slough fit well into a rotational couple in which the east side has moved to the south with respect to the west side (right-lateral movement), and a rotational couple seems the best explanation for all the structures in the Tableland. In this mechanism, dip-slip normal or vertical movement on the faults in the en echelon systems is the rule. However, it is possible in the same movement pattern for both lateral and thrusting movements to have existed. Lateral movements could take place on faults that ordinarily would have dip-slip movement, if the faults bent at the ends into thrusts or ended in warps. The Fish Slough fault provides an example of a fault along which lateral movement was made possible by a warp (the warp at the offset in the scarp). Two northeast-trending faults, one in the NW¼ sec. 9 and the one alined with the west side of the Tungsten Hills, bend eastward at their north ends toward the upthrown block, presumably as south-dipping thrusts which are perhaps similar to the thrust exposed in the pumice pit along the east edge of the Volcanic Tableland. The north-trending segments of faults bend into thrusts or terminate at warps, and must have lateral components of movement in the part adjacent to the thrust or warp commensurate with the amount of thrusting or warping; the lateral component must decrease away from the thrust or warp.

The destruction of the scarps of faults in the west side of the Volcanic Tableland by erosion must have taken place largely before the Owens River and Rock Creek entrenched themselves in their gorges. Hence, the principal post-Bishop tuff movement along the faults must have taken place not long after the extrusion of the Bishop tuff. It might be inferred that the displacements on the faults would diminish in depth because of increasing confining pressures downward, but the regional pattern of faulting and a few specific observations indicate that the reverse is true. Successive movements can be demonstrated for many faults along range fronts and, inasmuch as all the faults probably are part of the same system, recurrent movements seem likely for the faults in the Volcanic Tableland. If the faults have moved repeatedly, only the later movements are recorded in the Bishop tuff. Within the mapped part of the Volcanic Tableland, the fault that continues into the Volcanic Tableland from the west side of the Tungsten Hills has clearly had earlier movements along it. If the gravity high east of Fish Slough (pl. 7) signifies a horst, pre-Bishop tuff movements of large magnitude must have taken place there also. In the Casa Diablo Mountain quadrangle to the north, Rinehart and Ross (1957) have demonstrated that the displacement in the basement is greater than in the Bishop tuff along a fault on the west side of Casa Diablo Mountain and along a fault exposed in the Owens River Gorge.

STRUCTURAL RELATIONS OF THE TERRACES

The river terraces on the south side of the Volcanic Tableland between Round Valley on the west and Owens Valley on the east and southeast are considered to be correlative with the terraces that are cut across the east side of the Volcanic Tableland. They probably were formed at the same general time as the dissected fans south of Bishop along the Sierra Nevada front and also at the time of the faulting along the western base of the White Mountains that resulted in the dislocation of older fan deposits. This faulting is believed to have been at the time of the Tahoe glaciation because it was then that Mono Lake overflowed into Adobe Valley, and from there flowed southward into Owens Valley (Putnam, 1949, p. 1296). Three terrace levels are present on the south side of the Tableland whereas only two are present on the east side. In general, the terraces are above the depositional surfaces of Owens and Round Valleys, although the east-side terraces plunge under valley fill at the south end and the south-side terraces plunge under valley fill at the east end. The terraces on the south side of the Tableland are structurally the more interesting, and most of the discussion concerns them.

The terraces on the south side of the Volcanic Tableland are the products of the Owens River and an ancestral tributary that flowed along the north side of the Tungsten Hills. The maps (pls. 2, 3) show that the southern margin of each succeeding lower and younger terrace lies north of the southern margin of its next higher predecessor. Although the original northern limits of the terraces cannot be determined precisely because each succeeding lower terrace was cut into the northern margin of its predecessor, local benches and accumulations of gravel hang on the cliff on the south side of the Volcanic Tableland and indicate that some of the terraces, at least, extended north to the cliff.

Each period of terrace formation started with eastward tilting, which caused the Owens River to flow eastward along the base of the Volcanic Tableland. The river would gradually broaden its valley by cutting laterally toward the southeast. Lateral cutting would continue until tilting restored the river to an easterly course and initiated a new period of terrace formation.

A test for this hypothesis is found in the slope angle and amount of deformation of the terraces. If the hypothesis is correct, the youngest and lowest terrace should slope more gently eastward than the older, higher terraces, and apparently it does. In secs. 21, 22, and 23, the present flood plain slopes east at an average rate of less than 30 feet per mile, whereas the lowest terrace slopes at 40 feet per mile, and the middle terrace slopes at 70 feet per mile—the area of the highest terrace is too small to determine the slope angle. As a consequence of their different slope angles, the terraces converge toward the east and plunge under valley alluvium at about the same place.

South of Bishop, in the vicinity of the Rossi mine, the salient in the base of the Sierra Nevada is flanked by dissected fans that probably are of the same general age as the terraces. These dissected fans, like the terraces, plunge eastward beneath recent alluvium. Between the terraces and the dissected fans lies the recent fan of Bishop Creek, which lies somewhat lower than either the terraces or the dissected fans. The fan occupies a trough that was cut by Bishop Creek as a consequence of eastward tilting that affected both the terraces and the dissected fans.

The absence of conspicuous scarps on most parts of the terraces indicates not only that the terraces were cut after the deposition of the Bishop tuff, but also that they were cut since most of the fault scarps on the Volcanic Tableland were formed. This relation also is indicated in a few places by faults that cut the strata exposed in the margins of both the south-side and east-side terraces, most of which are overlain by a thin unfaulted layer of gravel. Nevertheless, the south-side terraces have been faulted off on the west and tilted toward the east. The east end of the middle terrace is broken by several faults having scarps of as much as 20 feet that face either east or west. These faults may have formed in connection with the tilting or warping. The most easterly scarp faces west and bounds a block that plunges eastward beneath the alluvium. Furthermore, although the terraces contain few conspicuous scarps, north-trending lineaments that make a pattern similar to the fault pattern in the Volcanic Tableland are visible on aerial photographs of the south-side terraces. These lineaments are low scarps generally 2 to 3 feet in height. Because the gravels capping the terraces are not as resistant to erosion as the Bishop tuff, the scarps are rounded. Presumably these scarps reflect a late increment of movement on the same system of faults that cuts the Volcanic Tableland.

The terraces along the east side of the Volcanic Tableland were cut by a strong river ancestral to the Owens River, which during the Tahoe glaciation probably carried the overflow from ancestral Mono Lake. The escarpment on the east side is in a single step in some places and in two steps in others. The tread of each step is a terrace and the escarpment is erosional. The escarpment coincides closely with the axis of warp along which the Bishop tuff is bent down to the east. This warp may overlie a fault in the basement rocks—it coincides with a strong gravity anomaly. Probably the escarpment was produced by the lateral cutting of a stream that was first entrenched in more easily eroded alluvial material that lay to the east of the warp. Such lateral cutting could have been induced by fans from the White Mountains which pushed the stream westward. The southward plunge of the terraces beneath the fill of Owens Valley strongly suggests a southward component of tilting since the terraces were cut.

STRUCTURES OF ROUND VALLEY

Round Valley is a structurally depressed block that is bounded by faults on the west and southeast, and by a warp accompanied by relatively minor faults on the northeast. The fault on the west side follows the base of Wheeler Crest and the range front south of Pine Creek, and the fault on the southeast side follows the west side of the Tungsten Hills. The part of the Volcanic Tableland adjacent to the northeast side of Round Valley slopes southeastward toward the valley as a result of warping, and the downward projection of this surface carries the Bishop tuff beneath the alluvium in Round Valley. Nevertheless, this warp is modified by one and possibly two faults that are downthrown toward Round Valley and which locally control the contact between the Bishop tuff and alluvium.

One fault that trends N. 30° W. and is marked by an eroded scarp that faces toward Round Valley intersects the contact between the Bishop tuff and alluvium about half a mile north of Birchim Canyon at a point where the contact turns southward parallel with the direction of the scarp. The scarp is about 50 feet in average height and is somewhat rounded by erosion. The fault does not displace the alluvium, but straightness of the contact between the Bishop tuff and alluvium south of the point of intersection and alinement of this contact with the projected trace of the fault suggest that the alluvium may have been deposited against the eroded scarp of the fault.

A fault along the west side of the Volcanic Tableland was drawn at the base of a prominent escarpment. Although faulting seems the most likely explanation for the escarpment, it could have been caused by lateral erosion of streams flowing from the Wheeler Crest escarpment.

Gravity data indicate that the alluvial fill in Round Valley is relatively shallow. No bedrock outcrops are present in Round Valley proper, but bedrock is exposed at one place in the bottom of a stream channel that crosses the Buttermilk Country.

SUBSURFACE STRUCTURE OF THE VALLEY BLOCK

The ancient surface preserved in the Coyote warp is believed to continue beneath the Volcanic Tableland, Round Valley, and Owens Valley north of Big Pine, and to dominate the bedrock configuration. South of Big Pine, however, the warp is faulted off along the fault that displaces the basalt in the east side of Crater Mountain and along which strong movements took place at the time of the earthquake of 1872. If the eastern flank of the warp north of Big Pine is projected eastward at its 10° average slope, the bedrock surface will lie about 5,000 feet beneath the valley floor at the western base of the White Mountains (pl. 5, section *C–C′*). The north flank also can be projected northward beneath valley fill and the Volcanic Tableland, but it must turn upward along an easterly trending synclinal axis because granitic and metamorphic rock crop out not far north of the quadrangle boundary and because evidence for large easterly trending faults in the critical area is lacking.

The postulated configuration of the bedrock has been substantiated by the gravity studies of Pakiser and Kane and more recently by seismic data of Pakiser. Although this information is of the utmost significance, several lines of geologic evidence, including data from well logs, also bear on the bedrock configuration.

The slopes of the ancient erosional surface in the Coyote warp, together with the absence of marginal valley-down faults of large throw north of Big Pine, constitute a first point of evidence. A second is provided by the eastward course of the Owens River along the south edge of the Volcanic Tableland to the foot of the fans from the White Mountains and its course southward along the east side of Owens Valley, in conjunction with the greater slope toward the east of older river terraces along the southern margin of the Volcanic Tableland. The course of the river merely suggests that the most recent depression of the valley has been along the east side, but the progressively greater easterly slopes of the older and higher terraces indicate that the east side of the valley has been subsiding for an extended period and that tilting was involved in the subsidence. The possibility that the tilting of terraces and the cutting of new ones coincided in time with periods of deformation along the base of the White Mountains is a third point.

A fourth clue is provided by a study of well logs supplied by the Department of Water and Power of the City of Los Angeles. A great many borings for water have been made into Owens Valley, but only about 40 were of use in interpreting the subsurface structure in the northern part of the Valley. All the logs used recorded penetrations of the Bishop tuff and the layer of white pumice at its base. This layer provides an excellent reference marker because it is relatively easy to identify in the well logs which were kept by drillers, because it is a continuous sheet except near the range fronts, and because it was deposited in a brief interval of time. The pumice is conspicuously white and a good aquifer—properties that usually were noted by the drillers. All the wells that supplied useful data are at the north end of Owens Valley, in the Bishop quadrangle. The geologic structure is shown in a fence diagram (fig. 64) in which the sections were drawn in such directions as to make maximum use of the well data as well as of outcrops of the Bishop tuff and of the pumice layer. Structure contours drawn on the base of the pumice are shown on the structure map (pl. 7).

In a general way the pumice layer intersected in wells slopes southeastward at a rate of about 100 feet per mile beneath valley fill, from 4,300 feet in the south margin of the Volcanic Tableland in the western part of the Bishop quadrangle to 3,300 feet at the base of the White Mountains near Redding Canyon. This slope, however, is not smooth; it is interrupted by faults in the Volcanic Tableland and in particular the fault along Fish Slough, and by a northwestward-trending depression whose deepest part is just north of Bishop. The depression lies between the northeastward-plunging anticlinal axis of the Coyote warp and the southward-trending structural high east of Fish Slough, and very likely marks a structural trough between these features. This interpretation is supported by the line of gravity lows (pl. 7), which extends northward along Owens Valley and splits near the depression in the Bishop tuff; one branch continues northward along the base of the White Mountains, and the other extends along the depression northwestward into the south margin of the Volcanic Tableland, where it bends westward to Round Valley.

The differences between the present configuration of the pumice layer and a horizontal surface cannot be taken to be entirely the result of deformation, for undoubtedly the pumice layer had an initial dip. As interpreted from old stream channels the south part of the Volcanic Tableland originally sloped in a S. 60° E. direction and swung more southward along the south edge. Almost certainly the swing in slope direction continued southward into Owens Valley where the slope direction was probably about the same as the present surface. It also seems reasonable to assume that the magnitude of the slope of the valley floor was about the same then as now. A reasonable interpretation is that the present slope of the pumice layer is the result of southeasterly initial dip and easterly tilting.

GRAVITY STUDY OF OWENS VALLEY

By L. C. Pakiser and M. F. Kane

A gravity study of Owens Valley north of Poverty Hills was completed by a U.S. Geological Survey party headed by M. F. Kane during three weeks of fieldwork in February 1954. This survey is part of a large, regional geophysical study of the structural geology of the area included within the modern drainage basin of Owens River. Earlier, in the summers of 1952 and 1953, Howard Oliver (written communication) made some gravity measurements in the Bishop area in conjunction with a regional gravity survey of the Sierra Nevada, and these measurements revealed that the valley block is expressed by a pronounced gravity low. The existence of this gravity low was confirmed by the later work, which permits it to be described and interpreted in considerable detail. About 300 gravity stations were established in an area of about 300 square miles in 1954, and 40 gravity stations that had been previously established by Oliver have been incorporated into the present study.

Field methods and reduction of gravity readings

All gravity stations were run from two base stations, at Bishop and at Big Pine, for which the relative observed gravity values had been accurately determined by several readings taken between them. These bases in turn were tied to Howard Oliver's earlier survey and more recent work in the area to the east by D. R. Mabey (written communication). Two instruments were used to make the survey: a Worden gravity meter with a sensitivity of 0.5046 milligal per scale division, and a Frost gravity meter with a sensitivity of 0.0729 milligal per scale division . The more sensitive instrument was used mainly in the valley floor where greater accuracy was desired; the less sensitive one was used mainly in the adjacent mountain slopes where its larger reading range and portability were useful.

Gravity stations were run in single loops from an initial reading at a base station. Base readings were repeated at intervals of four hours or less to determine the instrument drift and tidal variations. In each loop a station from a previous loop was read to check on the accuracy of the gravity data. A maximum of 0.3 milligal difference between the two readings was permitted. The elevation and position control for about 40 percent of the stations was obtained from U.S. Geological Survey topographic maps and bench marks. Bench marks, section corners, and road intersections were used as stations where their elevations were known and where they could be identified in the field. The control for the remainder of the stations was surveyed by plane table and alidade. A maximum vertical error of closure of 3 feet was permitted in the valley; 10 feet was allowed on the mountain slopes. The horizontal error of closure did not exceed 500 feet. Thus the permissible errors of closure for the gravity readings and the equivalent gravity errors of closure for the survey control were about the same.

All gravity measurements were referred to the base at Bishop which was tied, through the surveys of Oliver and Mabey, to absolute gravity bases of the U.S. Coast and Geodetic Survey and the University of Wisconsin. Elevation and latitude corrections were computed by using standard methods (Nettleton, 1940, p. 51–56). An elevation correction factor of 0.060 milligal per foot was used, corresponding to a density of 2.67 g per cm^3, the average density of the crystalline rocks in which the greatest variations of elevation are found. Latitude corrections were determined from the tables of theoretical gravity on the International Ellipsoid computed from the International Gravity Formula (Nettleton, 1940, p. 137–143). Terrain corrections through zone "O" were determined by using the Bullard modification of the Hayford-Bowie method (Swick, 1942, p. 67–68). (The outer radius of zone "O" is 546,914 feet.)

Finally, the gravity field was contoured at a scale of 1:62,500 and a contour interval of 1 milligal. On plate 7 this gravity contour map is reproduced with a contour interval of 2 milligals. The gravity contours are complete Bouguer gravity plus 1,000 milligals to make all values positive.

Interpretation of the gravity data

The predominantly clastic deposits of Cenozoic age that fill the valley block have a much lower density than the pre-Cenozoic rocks that confine them and that form the surrounding mountain masses. Therefore the gravity field (corrected for elevation, latitude, and terrain) in the valley area should be lower than that in the surrounding mountain slopes. By a careful study of the corrected gravity field, at least qualitative conclusions can be reached on the subsurface configuration of the buried bedrock floor. If the density contrast between the Cenozoic and the older rocks is known, or if some depth control is available from drill hole or seismic information, the depth and attitude of the buried bedrock floor can be determined within narrow limits along selected profiles. From the subsurface information thus determined, it is possible to draw some conclusions on the results of the tectonic movements.

In the valley block the average density contrast between the valley fill and the buried pre-Cenozoic rocks is not known accurately. It is assumed to be between 0.3 and 0.7 g per cm^3, and 0.5 is taken as the most probable density contrast. All theoretical interpretations in this study are based on these limiting and most probable assumed density contrasts. (Seismic refraction data given on p. 195 of this report support a density contrast near 0.5 g per cm^3.)

In this study, the interpretation of the gravity data has been approached in the following order:

1. A qualitative study was made of the gravity contours (pl. 7) to reach broad, general conclusions on the configuration of the buried surface of the pre-Cenozoic rocks.
2. Selected profiles were analyzed by using a two-dimensional gravity graticule designed by D. C. Skeels (in Dobrin, 1952, p. 98–99) and an assumed density contrast of 0.5 g per cm^3. Then upper and lower limits of the greatest depth were computed for the assumed limiting density contrasts of 0.3 and 0.7 g per cm^3. Certain local gravity anomalies were analyzed in detail.
3. The subsurface structure of the valley block was redescribed on the basis of the quantitative information obtained from the detailed analyses above.
4. Some conclusions were reached on the nature of the tectonic movements that led to the erosion of the mountain masses and deposition of the thick valley fill.

The gravity field shown by the gravity contours is assumed to be a true representation of the gravity field on the surface of the ground, corrected for elevation, latitude, and terrain effects. This assumption is believed to be valid over most of the area, except on the Volcanic Tableland, which has considerable topographic expression in materials of lower density (the Bishop tuff) than the 2.67 g per cm^3 assumed for the elevation corrections. The gravity field on the Volcanic Tableland may, therefore, actually be a few milligals higher than that shown on the map, but the general shape and amplitude of the anomalies is reliable and the theoretical interpretation of the subsurface structure in this area is valid.

The assumption that the subsurface masses analyzed by using a graticule are two-dimensional (that is with infinite extent along their axes) is not literally true, but the errors that result from this assumption are not great. Because of the uncertainties concerning densities, no further refinements are considered to be worthwhile, but the error, if any, in this assumption of two-dimensional masses results in depths somewhat less than the true depths.

The gravity contours

Owens Valley is an area of great gravity relief; the maximum range is about 45 milligals from the low southeast of Bishop in the valley block to the high northeast of Big Pine in Waucoba Canyon (pl. 7). There is a broad regional gravity gradient, and the gravity field generally decreases in a westerly direction. This regional gradient has been removed from each of the profiles analyzed in detail.

The most striking feature shown by the gravity contours is the elongated gravity low in the valley block. This low has extremely steep gradients along the White and Inyo Mountains and relatively gentle gradients along the Sierra Nevada, except near Big Pine where the gradients are steep. The axis of this low lies near the base of the White and Inyo Mountains and very nearly coincides with the present course of Owens River. The greatest gravity relief of this low, along profile *X–X′*, plate 11, is 27 milligals (corrected for the regional gradient). The steep gradient along the White-Inyo mountain chain indicates that the interface between the pre-Cenozoic and Cenozoic rocks there dips very steeply to the west. This zone of steep gradients extends to the northern limit of the area covered. Along the Sierra Nevada front just north of Big Pine the steep gravity gradient suggests that there the bedrock floor dips steeply to the east. Elsewhere along the Sierra Nevada front the bedrock surface probably slopes rather gently to the east. The deepest part of the Owens Valley structure lies very close to the base of the White and Inyo Mountains.

The conspicuous gravity high north of Bishop and just east of Fish Slough is the expression of a mass of dense rocks of unknown origin that stands in great vertical relief above the bedrock floor.

The conspicuous gravity low swings sharply to the west north of Bishop and into Round Valley. In this area the gravity gradients are gentle in all directions—a feature that indicates that the surface of the buried bedrock slopes gently toward the center of the gravity low from the Sierra Nevada front to the south and west and from a large mountain block to the north. A gravity high trends northeastward through Bishop and divides the gravity low into two parts.

Analysis of gravity anomalies

Three gravity profiles (*W–W′*, *X–X′*, and *Y–Y′* of pl. 7) were analyzed in detail by using a two-dimensional graticule (pl. 11).

The total gravity relief along profile $W-W'$, after removal of the regional gradient, is about 19 milligals. The gravity gradients are rather gentle on both sides of this profile. The interpretation shown, which is reasonably consistent with the gravity data, indicates that the bedrock floor slopes off gently into a syncline from the Sierra Nevada front to the south and from a large mountain block to the north. The probable depth of the valley fill is about 4,000 feet, and the depths range from 3,000 to 7,000 feet. The field and calculated gravity profiles could have been fitted much more closely, but the uncertainty about the elevation correction over the Volcanic Tableland would make this refinement meaningless. There is no geophysical evidence of faulting along this profile.

Profile $X-X'$ (pl. 11) was drawn across the largest local gravity relief in the part of Owens Valley included in this report. A total gravity relief of 27 milligals remains after removal of the regional gradient. The gravity gradients along the front of the White Mountains are much steeper than those along the Sierra Nevada front. The assumed geologic cross section meets the requirements of the observed gravity data almost exactly, although variations of this interpretation are possible. The valley fill is assumed to be in fault contact with the older rocks along the steeply dipping (about 60°) wall of the east side of the valley. On the west side, the bedrock floor slopes off gently from the Sierra Nevada front to the deepest part of the valley near the base of the White Mountains. Along the west side of the valley, the gravity profile is convex upward. In order to explain this condition, part of the buried bedrock surface is assumed to be convex upward also. The prominent steepening of the bedrock floor about 3 miles east of the Sierra Nevada front may reflect a buried fault scarp, possibly a continuation of one farther south of the west side of Owens Valley shown on profile $Y-Y'$. Density changes within the valley fill, which may result from a gradational change eastward from dense coarse, poorly sorted alluvial fan materials to less dense finer well-sorted clastic rocks, may possibly cause the convex upward gravity feature. The deepest part of the valley along profile $X-X'$ is filled with 4,000 to 10,000 feet of Cenozoic rocks. The most probable depth is about 6,000 feet.

Along profile $Y-Y'$ through Big Pine the total gravity relief, after the regional gradient has been removed, is 24 milligals. The gravity gradients are similar on the two sides of the valley, and the gravity low lies very near the center of the valley. The computed theoretical gravity profile for the assumed subsurface configuration nearly coincides with the field gravity profile, and the assumed structure therefore meets the requirements of the gravity data, although other similar geologic cross sections would do so as well. The greatest depth of the valley fill, if a density contrast of 0.5 g per cm^3 is used, is about 6,000 feet. The maximum and minimum depths of the deepest part of the valley based on assumed limiting density contrasts of 0.3 and 0.7 g per cm^3 are, respectively, about 10,000 and 4,000 feet. The steep gradients indicate that the valley walls dip steeply (roughly 60°) toward the center of the valley from both sides; the pre-Cenozoic and Cenozoic rocks are assumed to be in fault contact.

A detailed study was made of the conspicuous positive anomaly north of Bishop and east of Fish Slough. This anomaly has a total relief of about 21 milligals after subtracting the assumed anomaly of the valley structure. If a density contrast of 0.5 g per cm^3 between the causative mass and the surrounding Cenozoic sediments is assumed, the gravity anomaly can be explained by a truncated rectangular pyramid rising above the bedrock floor of the valley (about 6,000 feet deep) to the height of about 5,000 feet. The base of the mass is a rectangle about 3½ miles wide and 7 miles long, and strikes in a northerly direction. The top of the mass is a rectangle about 1 mile wide and perhaps 4 miles long, rising to within about 1,000 feet of the surface. The gravity gradient on the east side of the anomaly is probably somewhat greater than that to the west, therefore the east slope of the pyramid may be significantly steeper than the west slope. Other hypothetical masses could be used to explain the gravity data, but that described above is a reasonably accurate representation of the actual subsurface body. The actual subsurface body, however, is probably much more irregular in shape. If the density contrast were greater than 0.5 g per cm^3 the vertical relief of the causative mass would be less.

The origin of the mass is unknown. The mass could be an igneous intrusive, a series of lava flows that were deposited concurrently with the sediments, or a fault block.

The gravity high trending northeastward through Bishop may be caused by a large dense mass within the pre-Cenozoic rocks. It may be caused by a bedrock high, although several thousand feet of relief in the bedrock floor would be required to explain the gravity anomaly. It is equally possible that a thick section of lava may be buried under the gravity high. The cause of the anomaly remains indeterminate.

Conclusions and discussion

Owens Valley is a deep structural trough filled with about 6,000 feet of predominantly clastic rocks of Cenozoic age. A narrow, nearly linear fault zone dipping from 50° to 70° to the west (pl. 7) extends along the front of the White-Inyo mountain chain, and along this fault zone the valley fill is in fault contact with the pre-Cenozoic rocks. Along most of the Sierra Nevada front, the buried surface of the pre-Cenozoic rocks slopes rather gently to the east and north into Owens Valley and Round Valley. There is a fault zone near Big Pine, however, where the interface between the older rocks and the valley fill dips steeply from the Sierra Nevada front into Owens Valley (pl. 7). Here the Cenozoic and pre-Cenozoic rocks are probably in fault contact. There is a suggestion of a possible buried fault scarp about 3 miles east of the Sierra Nevada front along profile X–X' (pl. 11). Under the Volcanic Tableland, the buried surface of the pre-Cenozoic rocks slopes gently to the south from a large mountain block to the north. North of Bishop, and east of Fish Slough, a large buried mass of dense rocks of unknown origin rises to within about 1,000 feet of the surface. There is also a dense buried mass trending northeastward through Bishop, but it is not known whether this mass lies within the pre-Cenozoic rocks or projects above the general level of the bedrock floor.

Northeast of Big Pine, near the juncture of the White Mountains and the Inyo Mountains, the gravity contours and the postulated fault zone sharply change trend to nearly due north. This offset in the otherwise nearly linear fault zone may be caused by a northeastward-trending system of faults that continues into the White-Inyo mountain chain toward Deep Spring Valley. The volcanic field at Crater Mountain lies to the southwest in this zone.

The most satisfactory explanation of the steeply dipping interface between the clastic rocks of the valley block and the older rocks of the White-Inyo mountain block is that these rocks are in fault contact along a narrow fault zone. The interface cannot be buried topography, because such steep slopes would be quickly reduced by erosion to slopes comparable to those of the present Sierra Nevada and White Mountain escarpments (30° or less). Although warping is prevalent in this area, the warped surfaces all dip rather gently (as much as 20°). A system of faults in which displacement has been distributed through a wider zone, but in which the greatest amount of movement took place within the narrow zone described may exist. The fact that the Cenozoic and pre-Cenozoic rocks along this narrow zone are in fault contact indicate that the younger rocks must have been displaced by faulting against the older rocks. It seems unlikely that as much as 6,000 feet of clastic rocks could have accumulated prior to this faulting, for it is presumably the faulting that created the steep escarpment on which vigorous stream action eroded the older rocks and transported them as clastic sediments into Owens Valley. Therefore, it is concluded that the subsidence of the valley block was continuous or repeated at frequent intervals and concurrent with the accumulation of the clastic rocks. Repeated movement of the fault zone along the White-Inyo mountain chain is supported by the geologic evidence. Subsidence of the valley block must have been prolonged and may have begun in Tertiary time and continued to the present.

The bedrock floor was tilted downward to the east of the Coyote salient as the valley block subsided, but in the Big Pine area the floor of Owens Valley subsided as a graben along faults on both sides of the valley. The gently sloping bedrock surface in the synclinal area northwest of Bishop was formed primarily by downwarping but perhaps in part by distributive faulting. It is also possible that the gravity high trending northeastward through Bishop coincides with a bedrock high.

The coincidence of the gravity low axis along the base of the White-Inyo mountain chain with the course of Owens River suggests that the Owens River course was in about the same position throughout much of the period of subsidence of the valley block. Periodic shifting of the river channel to the east as the Owens Valley floor was tilting to the east is substantiated by the geologic evidence. Clastic sediments pouring into Owens Valley from the White Mountains held the river away from the base of the range.

The mass of dense rock north of Bishop and east of Fish Slough can be interpreted either as a series of lava flows that were deposited concurrently with the clastic sediments derived from the White Mountains to the east, a volcanic plug, a combination of these, or a fault block. The close relation of this feature to well-defined fault trends tends to support the fault-block interpretation, but it is difficult to understand how such a small mass could have been elevated as much as a mile relative to the bedrock floor around it. The local stress distribution required for such fault deformation is particularly puzzling in view of the fact that the mass is completely separated from the White Mountain block. On the other hand, there is no geologic evidence to support the interpretation of this feature as a mass of volcanic rocks, although volcanism is more readily understood than block faulting in terms of locally concentrated forces. (Aeromagnetic data obtained in 1956 and 1957 show that the material composing the mass is more magnetic than the material that surrounds it, but either plutonic rock of, say, dioritic composition or volcanic rock corresponding roughly to basalt could explain adequately the magnetic anomaly.)

The total throw on the fault zone bounding the west side of the White-Inyo mountain chain may have been as much as 13,000 feet. While the bedrock floor of Owens Valley was subsiding to its present altitude of about 2,000 feet below sea level, the White Mountains to the east were rising to an elevation of over 11,000 feet. However, warping may have caused part of the deformation.

SEISMIC PROFILE

By L. C. Pakiser

Five seismic-refraction profiles were shot in Owens Valley by a party under the supervision of W. H. Jackson during the summer of 1958. One of these (fig. 81) is in the area discussed in this report. Measurements were made by using conventional seismic-refraction instruments and methods, and the interpretation was done by conventional intercept-time and delay-time computations (Nettleton, 1940, p. 250–251).

The profile shown in figure 81 was run along an east-trending road 2½ miles south of Bishop (pl. 7, section *Z–Z′*). The maximum shot-point offset of the 12-geophone, 5,400-foot spread was 13,000 feet. A weathered layer having a velocity of 1,000 fps (feet per second) and a maximum thickness of 10 feet was assumed in making the calculations. The 5,820-fps segments recorded from shot points 1 and 3, together with the 6,050-fps segments from shot points 2 and 8, yield a true velocity of 5,900 fps, which is assumed to be that of the younger Cenozoic deposits. The interface between younger and older Cenozoic deposits dips to the east beneath the spread. The 6,720-fps segments are taken as the apparent downdip velocity, and the 7,120-fps segments as the apparent updip velocity to give a true velocity for the older Cenozoic deposits of 6,900 fps. Velocity layering for the part of the profile between shot points 1 and 8 was determined by using formulas for dipping beds; at shot points 3, 4, 5, 6, and 7, one-way delay times were used.

A precise calculation of the pre-Tertiary velocity was not possible for this profile because of small irregularities in arrival times; however, a reasonable interpretation of the data is made by taking 21,600 fps as the apparent updip velocity of the shots to the east and 12,300 fps as the apparent downdip velocity of the shots to the west. This correlation yields a true velocity of approximately 15,700 fps. No quantitative interpretation was made of the travel-time curve recorded from a shot point midway between shot points 3 and 4 because of an error in the shot-point location; however, the data are included to supply additional evidence of the 12,300 fps apparent downdip velocity of pre-Tertiary rocks.

The greatest depth computed to the pre-Tertiary rocks along this profile is approximately 4,800 feet (shot point 5). This depth and those calculated at shot points 6, 7, and 8 are assumed to be minimum depths, and the true depths may be 500 to 600 feet greater. This conclusion is based on the probability that the velocity of the older Cenozoic deposits east of the spread will increase with depth. The configuration of the pre-Tertiary bedrock as found from seismic measurements corresponds closely with that deduced from gravity nearby (pl. 11).

INTERPRETATION OF THE DEFORMATION PATTERN

The type of deformation active in the region today, represented by block faulting and warping, has characterized the structural history since the late Pliocene or early Pleistocene, and warping movements, precursors to the modern deformation, probably began as early as Eocene.

The sedimentary record in the San Joquin Valley indicates that although the Sierra Nevada region probably began to tilt westward at least as early as the Eocene, the rate was very slow before the beginning of the Miocene (Hoots, Bear, and Kleinpell, 1954). By late Miocene and Pliocene time a broad low topographic arch had been formed across the batholith and the adjacent area to the east (Axelrod, 1957). Approximate coincidence of the axis of this arch with the eastern edge of the batholith and with the axis of subsequent arching in late Pliocene and Pleistocene time suggests that the Miocene and Pliocene arch was an early product of the modern deformation pattern.

Evidence of progressive movements in the same sense along many individual faults and warps indicates that the subsidence of the valley block relative to the bordering ranges was accomplished by countless small increments of movement, each probably of about the magnitude of those that occurred in connection with the Owens Valley earthquake of 1872 (Whitney, 1872; Gilbert, 1884; Hobbs, 1910) and with other earthquakes in the Great Basin. Abundant fresh scarps indicate that movements have continued to the present at a significant rate and no doubt will continue into the future. The apparent continuity in the long and complicated sequence of structural events strongly suggests that continuous movement and stress patterns dominated events throughout the period of block faulting and warping.

Owens Valley is doubtless a keystone block in the crest of a broad north-trending regional arch of which the Sierra Nevada forms the west side and the desert ranges at least as far east as the Panamint Range form the east side (pl. 10). North-trending normal faults along Owens Valley and bounding the desert ranges in-

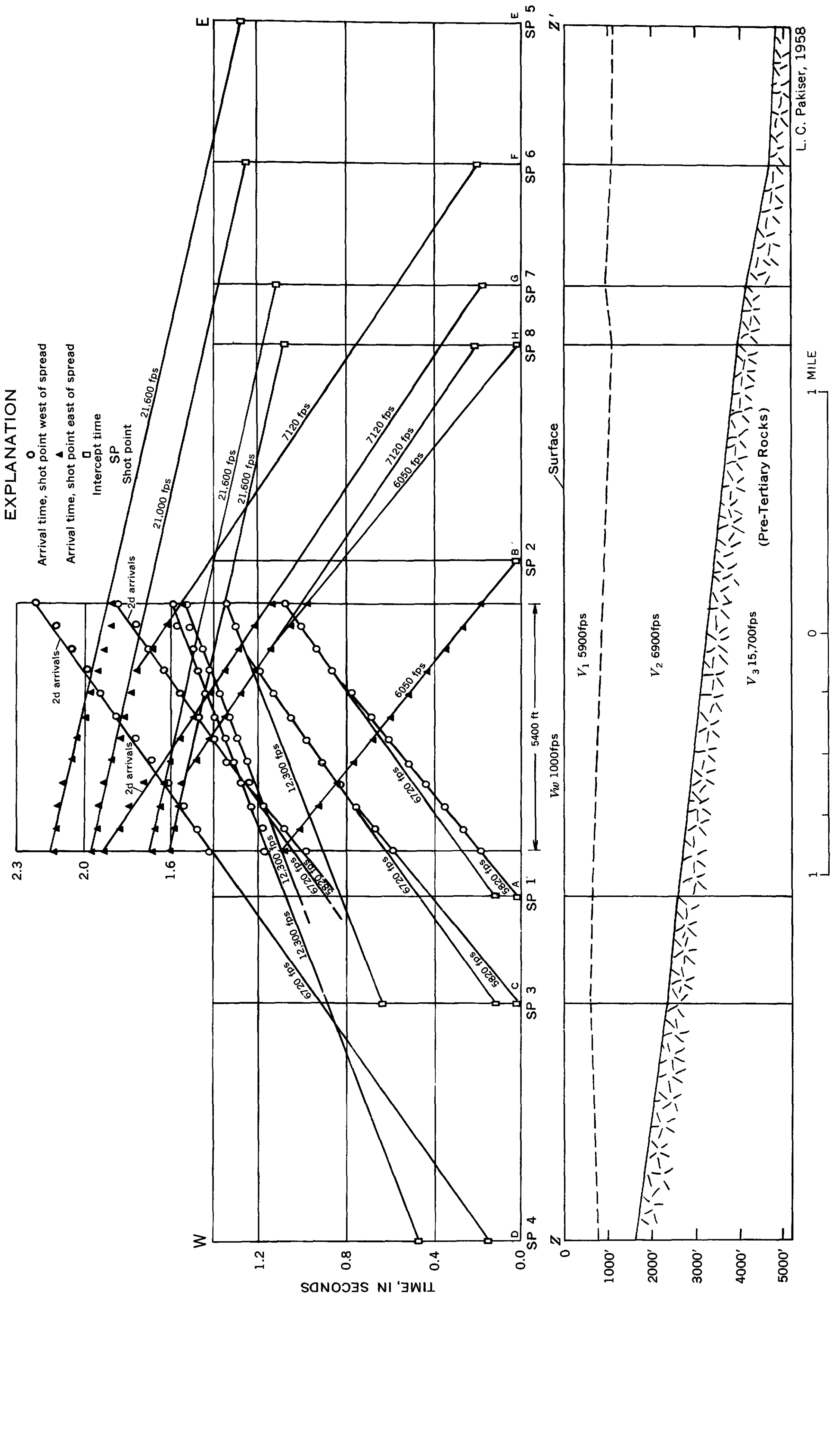

FIGURE 81.—Seismic profile across Owens Valley south of Bishop.

dicate east-west extension. The faults probably are merely near-surface secondary features that formed as a consequence of arching.

Similar relations between faults and warps or broad folds very likely exist elsewhere in the Great Basin, and mapping of the warps should lead to a better understanding of the nature of Basin and Range structure than we now have. The faults, marked as they are by conspicuous scarps, are usually given considerable attention. Warps, on the other hand, are seldom identified because they generally involve slopes of only a few degrees and must be delineated by mapping out reference surfaces.

Thompson (1960) has suggested that both the vertical uplift and lateral extension of the arch of which the Sierra Nevada forms the west limb were caused by internal expansion of the upper mantle, possibly as a result of abnormally high heat flow. Hafner (1951) has computed the probable deformation pattern in a block subjected to variable vertical and shearing stresses along its bottom. The similarity of Hafner's model (Hafner, 1951, pl. 1D) to the arch across the Sierra Nevada to the Panamint Range (pl. 10) is striking.

Within the Bishop district are structural features that indicate north-south shortening or a north-south clockwise couple. The absence of normal faults across north flank of the Coyote warp is in sharp contrast with the abundance of faults across the east flank and suggests north-south shortening. The small horsts, grabens, and tilted blocks on the Volcanic Tableland have been shown to be subsidiary features related to northwest-trending anticlinal and synclinal warps, and the deformation has been explained in terms of a north-south clockwise couple. This couple could have resulted from the buttressing effect of the Coyote warp in connection with southward-directed stresses.

A general north-south clockwise couple all along the edge of the batholith is suggested by the northwest-trending system of en echelon faults along the east side of the Sierra Nevada between Bishop and Lake Tahoe. Nevertheless, before lateral movement can be considered to have been systematic along the east side of the Sierra Nevada much more detailed geologic mapping is needed. Whitney (in Goodyear, 1888, p. 296) reported left-lateral movement east of Independence along a fault that moved in 1872, and Gianella (1959) has concluded on the basis of a review of the literature and undescribed field observations that the dominant horizontal movement in 1872 was left lateral. In the overall Cenozoic pattern, strike-slip movement undoubtedly was subordinate to dip-slip movement. Such strike-slip movements as occurred may have been caused by local inhomogeneities, and may have varied in sense from place to place and perhaps also from time to time.

The conjugate joints in the Sierra Nevada are interpreted to represent incipient shear planes but, lacking information as to the sense of movement on joints along which displacement took place, no conclusions can be made as to the orientation of the axis of greatest stress except that it probably was horizontal. If nonrotational stress is assumed, the axis of greatest compressional stress would have been north-south or east-west and would have bisected the average strikes of the joint sets.

ADJUSTMENT OF STREAMS TO THE STRUCTURAL MOVEMENTS

Because the structural movements have been progressive and in the same general sense, the chief adjustment of the streams within the mountain ranges has been simply to cut deeper. In certain areas, however, the stream drainages have been affected by structural movements strong enough to have changed the direction as well as the amount of slope. The adjustment of drainages to structural movements in two areas of exceptional interest will be described: that in the Volcanic Tableland and that in the Tungsten Hills. The diversion of Rawson Creek by a mountain-down fault in its fan has been described (p. 175) in connection with the discussion of the east flank of the Coyote warp.

ADJUSTMENT IN THE VOLCANIC TABLELAND

Three ancient consequent drainages that flowed southeastward across the Volcanic Tableland transverse to the present slopes can be identified (fig. 80). Of these, only one coinciding with Birchim Canyon and the Owens River Gorge below Birchim Canyon has maintained itself against the structural deformation that affected the Volcanic Tableland. A short drainage in secs. 19 and 20, T. 6 S., R. 32 E., seems never to have had any great length, and perhaps would never have maintained itself even without structural deformation. However, drainage farther northeast, now broken by faults into segments that stand at different altitudes, before deformation was a continuous drainage at least several miles long. Faulted segments of the old channel are identifiable from the south edge of the Volcanic Tableland in sec. 22, T. 6 S., R. 32 E., northwestward to the SE¼ sec. 12, T. 6 S., R. 31 E. No evidence of the channel was found farther to the northwest, probably because of increasingly intensive erosion in that direction.

If this stream were projected a little more than a mile and a half farther to the northeast along its course, it would intersect the Owens River Gorge in sec. 2, T. 6 S., R. 31 E., at an elbow. Below this elbow to Birchim Canyon, the gorge cuts directly across contours, whereas upstream it crosses contours obliquely. These relations probably indicate that the course of the stream above the elbow was established before the major deformation of the Volcanic Tableland and that the course below the elbow was established afterwards. The most probable explanation for this situation is that the Owens River below the elbow originally flowed downstream in the ancient fault-segmented drainage. This course makes a smooth sweeping curve from the north edge of the mapped area southeastward into Owens Valley, which would be a reasonable path for a consequent stream flowing across the original surface of the Volcanic Tableland.

The major structure formed in the Volcanic Tableland is a broad asymmetric arch, the crest of which plunges in a S. 30° E. direction. Warping of the southwest side of this arch toward Round Valley has changed the direction and amount of slope in the critical region from southeast to southwest. The stream in Birchim Canyon and the one that flowed through Birchim Canyon and the Owens River Gorge below Birchim Canyon was able to maintain itself across the new slope by downcutting.

However, it seems evident that the Owens River was not able to maintain its course against the warping and faulting movements and that it spilled out of its channel at the elbow in sec. 2, T. 6 S., R. 31 E., and flowed down the new slope to the stream flowing through Birchim Canyon. At that time the Owens River was not yet deeply entrenched and spillover could easily have taken place. The principal cause for doubt of this explanation is that the Owens River successfully maintained its course across a 100-foot-high scarp 2 miles upstream from the elbow while being defeated at the elbow where no evidence of a fault has been found. Very likely the relative rates of faulting and of warping in the two places were critical.

An unlikely alternative explanation of the changed course of the Owens River is that a capture was made at the elbow by a tributary from the stream that flowed through Birchim Canyon. The difficulty with this hypothesis is that the capture would have had to have been made at or before the time of principal warping and faulting; yet the adjustment of the segment of the Owens River between the elbow and Birchim Canyon to the present slope requires that the tributary worked its way headward after the present slope was established.

ADJUSTMENTS IN THE TUNGSTEN HILLS

Progressive movements along the west side of the Tungsten Hills have caused both McGee Creek and Horton Creek to change their courses at least twice. McGee Creek originally flowed across the Buttermilk Country in its present channel, through a wind gap on the south side of Grouse Mountain, and joined Birch Creek in the center of sec. 33, T. 7 S., R. 31 E. Differential movements between the Tungsten Hills and the graben between the Tungsten Hills and the main escarpment of the Sierra Nevada defeated the stream at Grouse Mountain and forced it to flow across a westerly spur of Grouse Mountain, in which it is now deeply entrenched as a result of later movements. For a time McGee Creek flowed eastward through a wind gap on the east side of Wells Upper Meadow, but further movements forced it into its present course.

Horton Creek at one time flowed across the Tungsten Hills through Deep Canyon; the stream that now occupies Deep Canyon is underfit. Horton Creek was defeated by structural movements and for a time flowed through a wind gap in the southwest corner of sec. 9, T. 7 S., R. 31 E., which it cut deeply, until further movements forced it into its present course. In the defeat of Horton Creek, warping rather than faulting may have been the effective structural agent inasmuch as the locale of the defeat is at the offset of the two fault segments that bound the west side of the Tungsten Hills.

An oddity in the drainage pattern is present in the east side of the Tungsten Hills, where McGee Creek and the stream that now occupies Deep Canyon leave the Tungsten Hills. There, McGee Creek flows through a notch that appears to have been originally occupied by the stream that carved Deep Canyon, and the stream in Deep Canyon spills over upstream from this notch through a deep slot in bedrock on the north side of its former course. A possible clue to solution of the drainage changes here is a terrace along the south side of the abandoned segment of Deep Canyon, opposite the notch through which the stream in Deep Canyon now flows. The terrace is level and lies close to the 4,880-foot contour. It may mark the shoreline of a small lake or pond that existed for a brief period. The pond very likely was caused by damming behind a fan built by outwash from the Bishop Creek glacier across the lower end of the valley. The height of the pond was regulated by a low point in the bedrock in its north side, through which the pond overflowed. Downcutting at this spillway eventually drained the pond and established the ancestral stream in Deep Canyon in its present course. It was probably subsequent to these events that McGee Creek was forced into its present course by the structural movements along the east side of the Tungsten Hills. Since then it has entrenched itself in the outwash from the Bishop Creek glacier, across several spurs of the Tungsten Hills, and through the notch formerly occupied by the stream that cut Deep Canyon.

SCULPTURING OF THE MOUNTAINS BY WATER AND ICE

While the valley block was being downdropped relative to the bordering ranges, valleys and canyons were carved in the ranges by stream and by ice, and the resulting sediment was deposited in the valleys. Many of the consequent streams along the range fronts are not related in any significant measure to the structures within the rocks, but flow directly down mountain fronts; however, Rock Creek, Pine Creek, Bishop Creek, the creek in Shannon Canyon, and many shorter drainages are adjusted along northeast-trending joints and faults. Bishop Creek and the creek in Shannon Canyon follow decidedly anomalous courses in the light of the present gross topography.

Headward erosion and entrenchment of streams was most rapid along the steep Wheeler Crest and Tinemaha scarps, and less rapid in the Coyote warp. Even now, Coyote Flat has not been reached, although headward eroding streams threaten it on all sides. Except for fault scarps, the local relief on Coyote Flat very likely has continued to be reduced during the deformation, and will continue to be reduced until dissected by headward eroding streams.

The streams in the White Mountains are smaller and less powerful than those in the Sierra Nevada because the White Mountains are in the rain shadow of the Sierra Nevada. Very likely the smaller amount of precipitation is the reason for the preservation of a gently rolling upland in the White Mountains. Broad rounded interfluvial areas between major stream canyons also reflect the slower rate of dissection by streams. Direct comparison of stream erosion in the Sierra Nevada and White Mountains is difficult because of the different kinds of rock of which these ranges are composed. Very likely the shale and slate, and possibly some sandstone, would erode more rapidly than granitic rocks under similar conditions, and carbonates would probably erode more slowly.

The contrasting landscapes of the two ranges are only partly the result of differences in the abundance and power of streams; during the Pleistocene the White Mountains were less severely glaciated than the Sierra Nevada and had only a few rather small glaciers (Blackwelder, 1934, p. 221). Inasmuch as both ranges have been uplifted to comparable heights, the less severe glaciation in the White Mountains cannot be attributed to a higher average temperature. The cause lies in the smaller precipitation in the White Mountains; annual snowfall was insufficient to form a cumulative snow pack.

This explanation of the scarcity of glaciers in the White Mountains during the Pleistocene carries with it two implications. The first is that in this area, increased precipitation could have been a more significant cause of glaciation than decreased temperature. After winters of exceptionally heavy snowfall in the Sierra Nevada, snow banks often survive the following summer. Present temperatures are low enough for glaciers to form if the winter precipitation were systematically increased over an extended period. Lower summer temperatures or summer cloudiness would, of course, favor the accretion of snow and the eventual formation of glaciers.

The second implication is that the glacial periods probably ended because the glaciers wasted away owing to decreased snowfall, rather than melted away owing to increased temperatures. Accordingly, the runoff was probably greatest and associated lakes at their highest levels at the peak of glaciation rather than during glacial retreat.

At the height of the Tahoe and Tioga glaciations the crest of the Sierra Nevada was a vast icefield through which the main divide and the principal ridges leading to the divide projected, and glaciers moved down the larger canyons to relatively low levels. These glaciers widened and deepened their canyons and deposited huge amounts of debris at lower altitudes in moraines and in outwash fans. When they retreated, the glaciers left behind them a legacy of glacial features, which add vastly to the beauty and interest of the landscape. Although the positions of the drainages were determined largely by streams, the modifying effects of glaciation were of very large magnitude, as has been emphasized by Matthes (1930, p. 84–103).

Chiefly because of the glacial features it illustrates the Mount Tom quadrangle map has been included in two sets of topographic maps, one of 100 maps and another more select group of 25 maps, that have been chosen to illustrate specified topographic features (Upton, 1955). In the southwest part of the Mount Tom quadrangle, along the Sierran divide and in the Humphreys Basin and French Canyon areas (fig. 69), features cited include Alpine topography, arête, basin, cirque, cirque headwall, cirque lake, col, cyclopean stair, glacier, Pater Noster lakes, and tarns. Pine Creek and Rock Creek canyons are given as examples of glacial troughs, and Pine Creek canyon is also cited for its U-shaped valley in cross section and for the lateral moraines at its mouth (fig. 82).

Since the last glaciation the rates of erosion and deposition have probably slowed down, especially at lower altitudes, because of the arid climate that prevails and because the modern landscape is almost completely fossil. Such easily destroyed constructional features as recessional moraines can be readily identified along most of the larger stream canyons, and they show that

FIGURE 82.—Aerial view looking southwest into Pine Creek Canyon. Note typical U-shape of glacial canyon, lateral moraines extending outward from canyon mouth, and loops of Tioga stage end moraine. Dark rock that crosses canyon at head is micaceous quartzite of the Pine Creek pendant. Photo by Symons Flying Service.

only minor erosion can have taken place since their deposition. Flash floods have locally added to the fill in canyon bottoms and along the range fronts, and talus continues to accumulate at higher altitudes and to form rock glaciers at the base of north slopes at altitudes above 11,000 feet. Otherwise, only rare periodic structural movements modify the landscape significantly.

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INDEX

[Italic page numbers indicate major references]