U. S. Geological survey.
Reports. Open file

UNITED STATES DEPARTMENT OF THE INTERIOR

GEOLOGICAL SURVEY

THE GEOLOGY OF THE BISHOP 15-MINUTE QUADRANGLE, CALIFORNIA

By

Paul C. harles Bateman, 1910-

May 1957

U. S. Geological Survey
OPEN FILE REPORT
This report is preliminary and has not been edited or reviewed for conformity with Geological Survey standards or nomenclature.

Weld - 2548

UNITED STATES
DEPARTMENT OF THE INTERIOR
Geological Survey
Washington 25, D. C.

For Release JANUARY 10, 1957 8

GEOLOGICAL REPORTS AND MAPS RELEASED FOR PUBLIC INSPECTION

The Geological Survey is releasing in open files the following reports. Copies are available for consultation in the Geological Survey Library, Washington, D. C., and at other places as listed:

1. Geology and ore deposits of the East Shasta copper-zinc district, Shasta County, California, by John P. Albers. 337 p., 28 illus., 7 tbls.

2. Geology of the Basin quadrangle, Montana, by Edward T. Ruppel. 219 p., 15 illus., 7 tbls.

On file at the Geological Survey office, South 157 Howard St., Spokane, Washington.

3. The geology of the Bishop 15-minute quadrangle, California, by Paul C. Bateman. 155 p., 14 figs., 6 pl.

On file at the Geological Survey Library, 4 Homewood Pl., Menlo Park, and Office of the Chief, Calif. Div. of Mines, Ferry Bldg., San Francisco, Calif.

4. The application, technique, and theory of Gish-Rooney Instruments, methods, and interpretation in electrical resistivity measurements, by H. C. Spicer. 288 p., 117 figs.

On file at the Geological Survey Libraries, Denver Federal Center, Denver, Colo., and 4 Homewood Place, Menlo Park, Calif.; and Geological Survey offices at 468 New Custom House, Denver, Colo.; 504 Federal Bldg., Salt Lake City, Utah; 602 Thomas Bldg., 1314 Wood St., Dallas, Tex.; 724 Appraisers Bldg., San Francisco, Calif.; 1031 Bartlett Bldg., Los Angeles, Calif.; and 210 Glover Bldg., Anchorage, Alaska.

x x x

Table of Contents

Page

ILLUSTRATIONS

ABSTRACT

The purpose of the study of the Bishop 15-minute quadrangle in east-central California was to appraise the mineral resources and to delineate and interpret the stratigraphic and structural history of the area. A report on the mineral resources has been published and the emphasis here is on stratigraphy and structure. About 8 months were spent in field study, and several months were spent in the office and laboratory compiling and analyzing data and in microscopic study.

The quadrangle includes the northern end of Owens Valley, the southern end of the Volcanic Tableland, a salient of the Sierra Nevada, and a wedge-shaped strip along the western face of the White Mountains. Owens Valley is underlain by alluvial deposits and small masses of basalt of Cenozoic age; the Volcanic Tableland by rhyolite tuff of Pleistocene age; the Sierra Nevada by granitic rocks of Cretaceous(?) age, which enclose small masses of pre-Cretaceous metamorphic and mafic igneous rocks; and the White Mountains chiefly by strongly-deformed sedimentary strata of Early Cambrian and Late Precambrian(?) age.

The geologic structures are the products of 3 different tectonic episodes. The earliest structures -- folds, faults, and cleavage formed by compressional forces in Late Jurassic or Early Cretaceous time prior to the intrusion of granitic rocks -- are found only in the sedimentary rocks of the White Mountains and in metamorphic remnants in the Sierra Nevada. In the White Mountains the strata

were tightly folded and faulted along north- to northwest-trending lines. Faulting probably began after the rocks were so tightly folded that faulting required less energy than further folding. Slaty cleavage is present in slate and locally in limestone and sandstone.

Next are structures that are the result of the emplacement of granitic rocks in Cretaceous(?) time. Strata adjacent to the Poleta stock are believed to have been bent upward by the intrusion of the stock, and it is surmised that the distribution of metamorphic remnants in the Sierra Nevada was affected by the emplacement of the batholith.

The most recent structures--chiefly normal faults and warps of Cenozoic age--account for the larger features of the landscape and many of the smaller ones. The Sierra Nevada escarpment here is a warp rather than a fault scarp as it is to the north and to the south. An erosion surface on the warp slopes toward Owens Valley at about 10 degrees and continues beneath valley fill to the base of the White Mountains where it is cut off at a depth of 5,000 feet beneath the floor of Owens Valley along a system of faults that border the White Mountains.

The Cenozoic structures have involved countless small movements. Repeated movements along many faults is clear, and abundant fresh scarps along older faults, some formed as late as 1872, indicate that movements have continued to the present. Most structures indicate extension rather than compression, and a clockwise rotational couple seems to provide a satisfactory explanation for small horsts, graben, and tilted blocks in en echelon arrangement on the Volcanic Tableland. These structures are small-scale replicas of larger structures in the Great Basin region, and it is suggested that these larger structures were produced by similar forces.

INTRODUCTION

Purpose and scope of the report

The purpose of this report is to present the results of a geologic study of the Bishop 15-minute quadrangle in east-central California (plate 1). The mapping of the quadrangle was done as part of a study of the Bishop tungsten district for the U. S. Geological Survey in cooperation with the California Division of Mines. The entire study involved the geologic mapping of several 15-minute quadrangles, most of which lie chiefly in the Sierra Nevada where productive tungsten deposits are found. However, the Bishop quadrangle includes only a salient of the Sierra Nevada in its southwest corner, and it also contains segments of the White Mountains, Owens Valley, and a broad plateau of rhyolite tuff called the Volcanic Tableland--all features of exceptional geologic interest.

The report is in four parts. The first and second parts are short; they are an introduction and brief description of the geography. The third part is a description of the geologic formations. The fourth part is concerned with the geologic structures. A report on the mineral deposits of the Bishop district has been published (Bateman, 1956), and consequently the mineral deposits are not treated here.

Previous geologic work

One of the earliest accounts of the geology of the northern Owens Valley region is contained in the first report of the California Geological Survey, "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 bibliographies as having been published in 1873, actually was never published. Another early description of the geology of the area was written by W. A. Goodyear, who traveled through Owens Valley in 1870 and again in 1888 examining the geology and ore deposits. His interesting report is included in the 8th annual report of the state mineralogist (1888, p. 224-309). One of the earliest maps to show the geology of the region is a "Preliminary Mineralogical and Geological map of the State of California", published in 1891 by the California State Division of Mines as Map 1. On this map the Sierra Nevada opposite Bishop is shown to consist of Jurassic and Triassic rocks, which are cut by less abundant northerly trending lenses of granite, and the White Mountains are shown to consist of northwesterly trending Permo-Carboniferous strata along the range front and of granite and Jurassic and Triassic strata farther east. Several extensive volcanic areas are shown along the Sierra Nevada front in the span between Bishop and Big Pine. A map published in 1905 by J. E. Spurr, as part of a report on a reconnaissance of Nevada south of the 40th parallel and adjacent portions of California shows a threefold distinction in the northern Owens Valley region of granular or coarse porphyritic igneous rock in the Sierra Nevada, strata of Cambrian age in the White Mountains, and strata of Pleistocene age in Owens Valley. The following year a paper by W. T. Lee (1906) on the ground water resources of Owens Valley was issued in which are included 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 covered 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 resulting from these studies that deals most fully with the Bishop district is one called "Deformation in the interval Mt. Lyell-Mt. Whitney" (1941). Also during the 1930's C. M. Gilbert studied the volcanic region north of Bishop, including the Volcanic Tableland in the north-central part of the Bishop quadrangle (Gilbert, 1938 and 1941). Recently, Bateman and Merriam (1954) published a geologic map of the Owens Valley region.

Methods of investigation

Field work in the Bishop quadrangle was done chiefly during the summer and early fall of 1947, but parts of the map were revised in 1948 and 1949. In all, about 8 months were spent in the field. During part of the mapping the writer was aided by one or two temporary assistants, and part of the time he was by himself. The data collected in mapping were plotted in the field on prints of the aerial photographs from which the topographic map of the Bishop 15-minute quadrangle was made. These prints are on an approximate scale of 1/35,000. Observations were plotted with the use of a simple lens-type stereoscope with a 2X magnification. The use of a stereoscope permitted more accurate plotting of data than would otherwise have been possible, and was of considerable additional value in planning traverses and in evaluating possible alternative interpretations of the data. The data plotted on the photographs were transferred to the topographic base map of the Bishop quadrangle by the multiplex method in the Sacramento laboratories of the Topographic Division of the Geological Survey. Later a few observations were added to the maps and some modifications were made by simple inspection.

The topography within the Bishop quadrangle is gentle compared with adjoining areas and permitted ready access to all parts of the area. Nevertheless, the coverage of ground was most thorough in the White Mountains where extremely complicated structures require detailed observations to acquire the data that are necessary for their interpretation. The Sierra Nevada and the Volcanic Tableland were covered less thoroughly, and the floor of Owens Valley least thoroughly.

Modal analyses of granitic rocks were made with a point counter of the type described by Chayes (1949), and the counts were recorded on a blood counter. Determinations of plagioclase composition were made with a 4-axis universal stage. Lineations defined by the intersection of bedding planes with cleavage planes were computed with a Schmidt equal-area net.

All colors referred to in the report are in accord with a Rock Color Chart issued by the Geological Society of America (second printing, 1951). This chart is based on the Munsell system of color identification.

Acknowledgments

In mapping the quadrangle, the writer was aided during the summer of 1947 by J. W. Reid and M. W. Ellis, who served as field assistants. During the preparation of the report the writer benefited by discussions with his colleagues of the Geological Survey and with W. C. Putnam of the Department of Geology, University of California at Los Angeles, who has studied geomorphic problems in adjacent areas to the north. Many of the ideas expressed in this report had their inception in these discussions. The illustrations were drawn by Mrs. Esther McDermott, who also aided in their planning and provided counsel and assistance far beyond the mere mechanics of drafting.

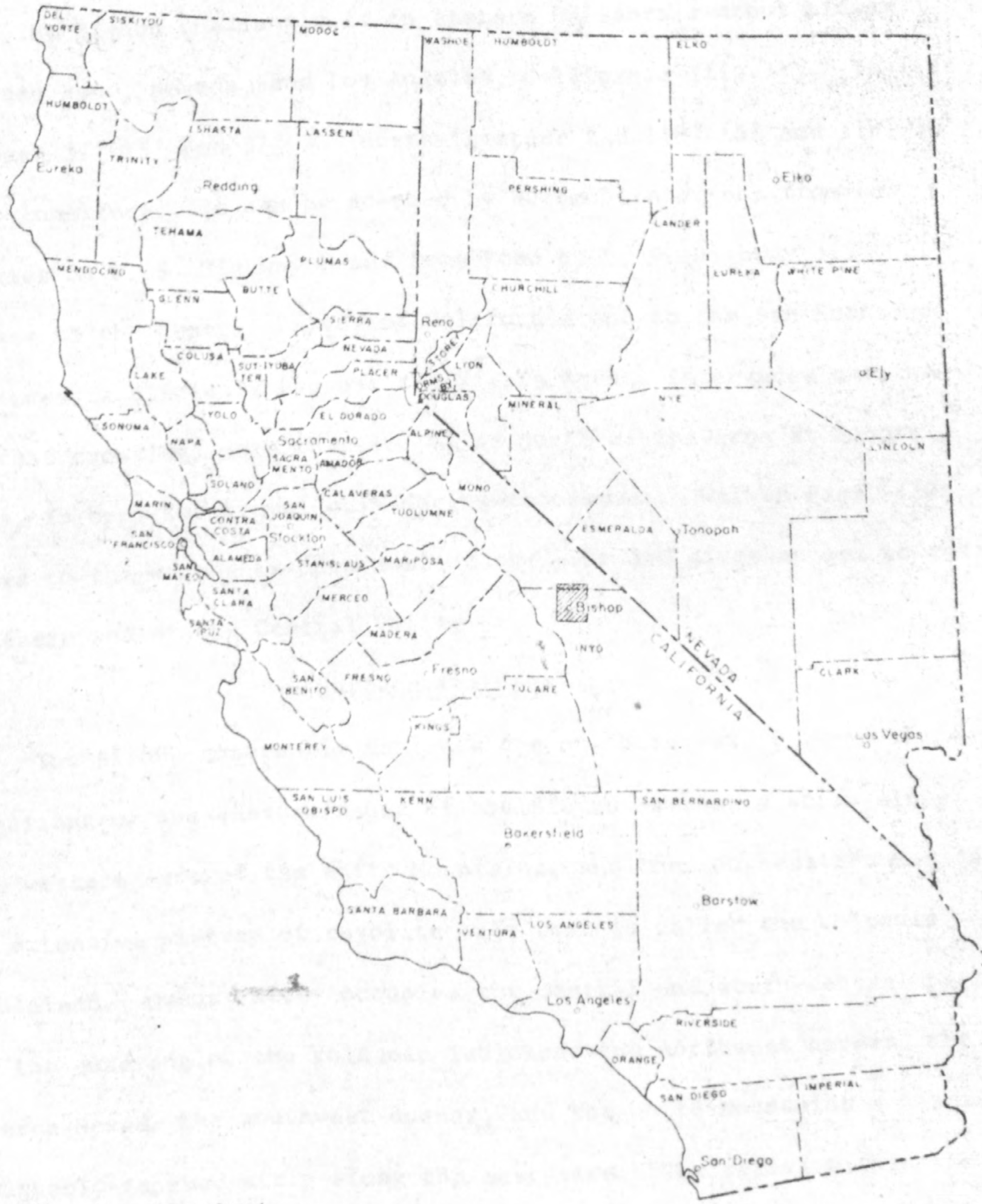


Figure 1. Map of California and Nevada showing the location of the Bishop 15-minute quadrangle.

GEOGRAPHY

Location and accessibility

The Bishop quadrangle is in eastern California about midway between Reno, Nevada, and Los Angeles, California (fig. 1). It lies between 37° 15' and 37° 30' north latitude and 118° 15' and 118° 30' west longitude. It can be reached by automobile or bus from Los Angeles on U. S. Highway 6 and from Reno by U. S. Highway 395. Access to the Central Valley of California and to the San Francisco Bay area is difficult because the Sierra Nevada intervenes and the nearest crossing, more than 100 miles north of the area at Sonora Pass, is open for travel only during the summer. Walker Pass, 130 miles to the south, is open most of the year and gives access to the southern end of the Central Valley.

Surface features

The Bishop quadrangle includes the northern end of Owens Valley, a salient of the eastern slope of the Sierra Nevada, a strip along the western face of the White Mountains, and the southeastern end of an extensive plateau of rhyolite tuff that is called the Volcanic Tableland. Owens Valley occupies the central and south-central part of the quadrangle, the Volcanic Tableland the northwest corner, the Sierra Nevada the southwest corner, and the White Mountains a narrow southerly-tapered strip along the east side. The relief within the quadrangle boundaries is more than 5,000 feet -- ranging from less than 4,000 feet on the floor of Owens Valley to about 9,600 feet in the White Mountains and 10,121 feet in the Sierra Nevada. Nevertheless, most of this relief is confined to the Sierra Nevada and White Mountains; the relief in Owens Valley and the Volcanic Tableland, which together comprise about two-thirds of the area of the quadrangle, is only about 1,000 feet.

Owens Valley proper terminates in the north part of the quadrangle, a few miles north of Bishop, along a steep, south-facing escarpment that bounds the Volcanic Tableland, although 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 are virtually a part of the Valley.

The floor of Owens Valley is a flat plain 3 to 5 miles wide that is flanked on the east and southwest sides by extensive alluvial fans formed from debris carried out of the White Mountains and the Sierra Nevada respectively. The Valley floor slopes gently southward -- a bench mark at Bishop is at 4,141 feet and the lowest contour that crosses the south boundary of the quadrangle, eight miles to the south, is at 3,960 feet.

The Volcanic Tableland slopes generally southeasterly at a rate of about 100 feet per mile, but it 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, extends entirely across the segment of the Tableland within the quadrangle and divides it into two parts. This scarp ranges from 200 to 400 feet high and faces westward; it is offset about half a mile to the west near the north boundary of the quadrangle. Most other scarps are a mile or less long and less than 100 feet high, but a few are longer and higher.

The Volcanic Tableland stands higher than Owens Valley, and is bounded on the south and east sides by escarpments. At the southeast corner, however, the Tableland slopes to the level of the valley floor. The escarpment on the south side of the Tableland is higher than the one on the east side. It averages about 200 feet high, but is higher toward the west and lower toward the east. The escarpment on the east side has a maximum height of about 200 feet.

Elevated, stream-cut terraces flank the Volcanic Tableland on the south and east sides. Three terrace levels can be readily identified in the south-side terraces, but only two were distinguished in the east-side terraces. The terraces are cut chiefly on the rhyolite tuff of the Volcanic Tableland. The south-side terraces are gravel covered, whereas gravels are present only locally on the east-side terraces.

The salient of the Sierra Nevada that lies within the quadrangle marks a local change in the trend of the Sierra Nevada escarpment from northerly to westerly. Remnants of an old upland surface are widely preserved in the salient. These remnants show that the old surface slopes north and east to the floor of Owens Valley at about 10 degrees. The escarpment is quite unlike the precipitous fault scarp that generally is considered to be typical of the eastern face of the Sierra Nevada. The salient is, in fact, the apex of a two-sided structural feature, here called the Coyote Warp. This warp stands out in front of the projected trend of the Sierra scarp both from the south and from the northwest, and thus constricts Owens Valley. The warped surface of the salient has been cut by several deep canyons, the largest of which is the canyon of Rawson Creek.

The White Mountain escarpment within the quadrangle in general is a trifle steeper than the Sierra Nevada escarpment, the slope angle ranging generally between 10 and 15 degrees, but it hardly can be characterized as precipitous. Slopes generally are steeper in the south part of the quadrangle than in the north part. In the north part, especially north of Silver Canyon, valleys parallel with the trend of the range front are present in the interfluves. The range front is cut into by several canyons, the 3 largest of which, Black, Silver, and Coldwater Canyons, contain perennial streams. These are steep-walled canyons, but with notably regular-sloping, gravel-covered stream beds.

The quadrangle is drained by the Owens River, which empties into Owens Lake at the south end of Owens Valley. From the western edge of the quadrangle the river flows due east along the base of the cliffs that mark the south side of the Volcanic Tableland to the lower ends of the fans from the White Mountains, then it flows southerly along the east side of Owens Valley. In the historic past the principal tributary within the quadrangle to the Owens River was

Bishop Creek, but the waters of this stream have been diverted into canals. Rawson Creek and the streams in Coldwater and Silver Canyons in the White Mountains also have been diverted, but these streams appear to have sunk into their fans before reaching the Owens River, except possibly when in flood.

Climate and vegetation

The 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 range 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 degrees F., but the nights are comfortably cool. Winter days generally are pleasant with the temperature well above freezing; nights though below freezing are rarely as low as zero degrees. Most of the precipitation is as winter snow, but summer thunder showers are common. The amount of snow on the ground at any one time rarely exceeds a foot, although as much as several feet has collected in severe winters.

At higher altitudes the temperatures are progressively lower and the precipitation higher. At altitudes of 10,000 feet snow collects to depths of many feet, and on north slopes banks of snow may persist through the summer. During July, heavy thunder showers, accompanied by spectacular displays of lightning, are common in the middle part of the day.

Climatic data supplied by the U. S. Weather Bureau for Bishop and for South Lake, a station several miles southwest of Bishop and outside the Bishop quadrangle, are summarized in the following table. Data from the weather station at Bishop, at about 4,100 feet, are representative of lower altitudes on the floor of Owens Valley; and those from the station at South Lake, at about 9,600 feet, are representative of higher altitudes in the Sierra Nevada.

Summary of Weather data at Bishop and at South Lake

	Bishop airport (Alt. 4108)			South Lake (Alt. 9620)		
	Jan.	July	Annual	Jan.	July	Annual
Ave. max. temp. (F.)	52.8	93.8	72.8	38.8	69.1	51.4
Ave. min. temp. (F.)	21.1	54.3	36.2	12.5	45.2	27.2
Ave. temp. (F.)	36.8	73.6	54.6	25.7	57.1	39.3
Ave. precipitation (inches)	1.5	0.1	5.8	2.2	.6	17.4
Ave. snowfall (inches)	5.5	0	17.4	33.9	Trace	174.0

Data are from the files of the U. S. Weather Bureau.

All data are averages of periods of at least 20 years.

Ave. precipitation data at Bishop are for a 40-year period.

Native vegetation is sparse on the lower slopes of the mountains and on the floor of Owens Valley, except along streams, where deciduous trees are found. Conifers are present locally in the Sierra Nevada above altitudes of 6,500 feet and in the White Mountains above 7,500 feet, but continuously forested areas like those on the west slope of the Sierra Nevada are absent.

GEOLOGIC FORMATIONS

The distribution of rocks within the quadrangle is correlative with the distribution of physiographic units. The White Mountains are underlain chiefly by strongly deformed sedimentary strata, the Sierra Nevada by granitic intrusive rocks in which small masses of metamorphic and mafic igneous and hybrid rocks are scattered, the Volcanic Tableland by rhyolite tuff, and Owens Valley 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 Early Cambrian and Late Precambrian(?) age; the scattered metamorphic remnants in the Sierra Nevada probably include rock of both Mesozoic and Paleozoic age; the granitic intrusives are of Cretaceous(?) age; and the rhyolite tuff of the Volcanic Tableland, the scattered masses of olivine basalt, and the various surficial alluvial deposits are of Cenozoic age. In the following sections, the rock descriptions are arranged generally in the order of decreasing age, but with some modification for the Cenozoic deposits in order to bring the volcanic rocks together into one group and the sedimentary deposits into another.

Lower Cambrian and Upper Precambrian(?) sedimentary rocks of the White Mountains

More than 9,000 feet of strongly deformed strata of Early Cambrian and Late Precambrian(?) age crop out in the wedge of about 40 square miles of the west slope of the White Mountains that is included within the quadrangle. The upper 1,600 feet of strata, belonging to the Silver Peak group, contain fossils that indicate an Early Cambrian age, but diagnostic fossils have not been found in the lower 7,500 feet of strata. Following is a summary of the exposed stratigraphic sequence with the approximate thicknesses of the mapped units:

Silver Peak group		
limestone unit	1,000 feet	Fossiliferous Lower Cambrian
lower slate unit	600 feet	
Campito sandstone	3,000 feet	No diagnostic fossils found
Deep Spring formation	1,500 feet	
Reed dolomite	2,000 feet	
Pre-Reed dolomite strata	1,000 feet	
	9,100 feet	

No unconformities were found in the area studied although previous workers have reported unconformities in the same stratigraphic section in the Blanco Mountain and Waucoba Mountain quadrangles, which adjoin the Bishop quadrangle on the east and southeast. The character of the stratigraphic sequence and the physical aspects of the contacts between formations seem entirely compatible with continuous unbroken deposition. However, the comparatively small size of the area mapped and the highly deformed character of the rocks make inadvisable a categorical statement that no unconformities exist within the exposed stratigraphic section.

With the exception of the Deep Spring formation no attempt has been made to make detailed measurements of stratigraphic sections or to describe the variations in the lithology; indeed, the complexity of the structure virtually precludes detailed measurements within the quadrangle. Contiguous areas to the east in the Blanco Mountain and Waucoba Mountain quadrangles are underlain by the same strata, generally less deformed and in better order.

Pre-Reed dolomite strata

Strata that underlie the Reed dolomite are 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. (plate 1). Both exposures are in the base of the range at the head of the alluvial apron.

At the southern locality about 1,000 feet of strata are exposed in a prominent ridge half a mile south of the mouth of Black Canyon. The exposed strata can be divided into four lithologically distinct units of about equal thickness. 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 with undolomitized residual masses of thin-bedded medium light gray limestone. Some of the limestone contains small oolitic nodules of concentric structure. The nodules range from an eighth of an inch or less to half an inch in diameter and may be algal.
4. Thin-bedded dark gray limestone and sandstone with 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 with some shale interbeds, then a massive medium-gray limestone bed about 30 feet thick. The top 110 feet consists of thinly-interbedded limestone and shale; medium gray limestone beds one to four inches thick commonly are 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 appears sharp.

The 280 feet of strata at the northern locality is lithologically similar, though not identical, with unit 4 at the southern locality. If these strata are stratigraphically equivalent, an unconformity at the base of the Reed dolomite is unlikely inasmuch as at both the northern and southern exposures the beds appear to be structurally conformable with the overlying formations.

In the Blanco Mountain quadrangle to the east J. V. Maxon reports a formation 3,700 feet thick unconformably beneath the Reed dolomite for which he has proposed the name Wyman formation (Maxon, 1935, p. 314). Maxon's brief description of the strata as spotted schist and phyllite with a few dolomite beds indicates a degree of similarity with the strata here described, but the unconformity he reports at the base of the Reed dolomite casts some doubt on direct correlation of the strata in the two localities. Kirk also has briefly described the strata beneath the Reed dolomite in the Blanco Mountain quadrangle as "oldest sandstones and dolomites." He writes, "In general, the series as seen at several points seem to consist of thin beds of arenaceous slate, some beds of impure dolomite, and thin beds of sandstone" (Knopf, 1918, p. 23).

Reed dolomite

The Reed dolomite is well-exposed along the lower slopes of the White Mountains between Redding Canyon and Black Canyon, and for about 2 miles south of Black Canyon. The most notable characteristic of the formation is its massiveness. From a distance, bedding is shown by thick members, but on close examination is obscure and difficult to identify. Well-developed joints cut the dolomite almost everywhere and the rock breaks along joint planes into angular blocks that form rough talus slopes of moderate steepness.

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

The lower contact of the formation with the pre-Reed dolomite strata is sharp and is marked by a conspicuous lithologic break, but the upper contact with the arenaceous dolomite of the Deep Spring formation is less well defined. The horizon chosen as the upper contact for representation in mapping is the base of a thick yellowish gray unit whose color contrasts with the homogeneous pale or medium gray dolomite beneath. 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. South of Black Canyon above the southern exposure of pre-Reed dolomite strata, however, the formation appears to be cut by only one fault that causes the apparent thickness of the Reed dolomite to appear greater than the true thickness. The apparent thickness of the formation at this locality is 2,200 feet, assuming no displacement on the fault. The throw on the fault is not known, but it is thought to be no more than 200-300 feet. In the construction of the accompanying structure sections (plate 3) and block diagram (plate 4) a thickness of 2,000 feet was assumed. Support for this assumption is derived from a measurement by Kirk (Knopf, 1918, p. 24) of 2,000$^{\pm}$ feet in the head of Wyman Canyon, a few miles to the northeast. Maxon (1935, p. 314), however, measured a thickness of 2,500$^{\pm}$ feet in the same general area.

The most probable correlative of the Reed dolomite is the Noonday dolomite of Hazzard, 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 Lower Cambrian age, and both 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 here designated as the Deep Spring formation are well-exposed along the White Mountain front southeast of Bishop, along the same distance as the Reed dolomite on which it rests. The formation consists of 1,500$^{\pm}$ feet of arenaceous dolomite with limestone in the upper 150 feet and two prominent quartzitic layers at 220 and 800 feet beneath the top. From the floor of Owens Valley between Bishop and Big Pine, the formation is visible for two miles north from Black Canyon (fig. 2a). In the late afternoon, the well-

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

Figure 2b. Exposure of older dissected alluvium in Poleta Canyon. Thin layers of shale, tuff, marl, and limestone are interbedded with coarse bouldery layers. Marl contains gastropods characteristic of sluggish water.

Figure 2a.

Figure 2b.

bedded alternately brown and gray weathering colors of the strata contrast with both the underlying massive yellowish gray Reed dolomite and with the overlying dusky brown to grayish black Campito sandstone. This section, measured and described by Walcott (1895, p. 142-143), is reproduced in plate 2 together with two other stratigraphic sections, one on the south side of Black Canyon measured by the writer, and one at the type locality of the Deep Spring formation on the north side of Deep Spring Valley measured by Kirk (Knopf, 1918, p. 25).

The section visible from Owens Valley and measured by Walcott lies just outside of the quadrangle, but the one measured by the writer, though not visible from the floor of Owens Valley, is equally satisfactory (plate 2). This section was measured along a ridge on the south side of Black Canyon in the SE¼ sec. 4, T. 8 S., R. 34 E.

The upper contact with the Campito sandstone is sharp and offers no problem in mapping. Beds in the basal part of the Campito sandstone and in the top of the Deep Spring formation are essentially conformable with each other and with the contact throughout the mapped area, giving no support for an unconformity as was suggested by Kirk (Knopf, 1918, p. 24-25). The lower transitional contact with the Reed dolomite offers more difficulties. The 500 feet of strata at the base of the formation are generally more massive than higher beds, and bedding is discernible only locally. The unit is truly transitional with the Reed dolomite and may almost as

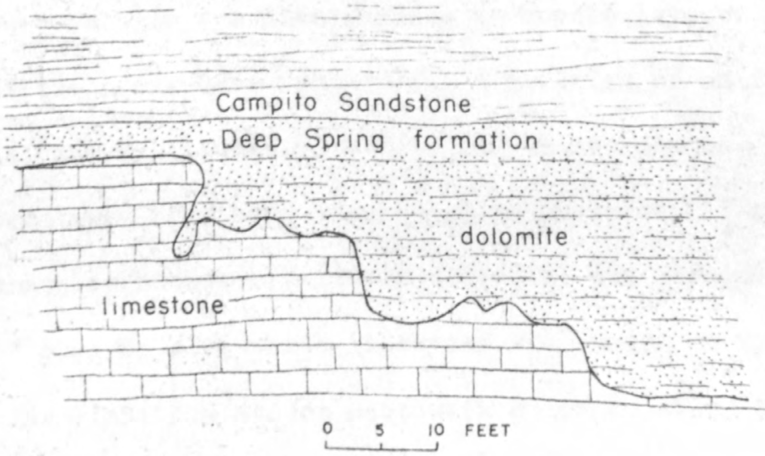


Figure 3 Cliff section on north side of Black Canyon showing irregular dolomitization of limestone in the Deep Spring formation. Undolomitized limestone residuals are most abundant at the top of the formation just beneath the Campito sandstone.

well be considered the uppermost member of that formation as the basal member of the Deep Spring formation.

One of the more interesting features of the formation is the presence locally of prominently developed cross-bedding 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 prominent cross-beds are in arenaceous layers, but in places cross-beds also are preserved in carbonate layers. The amplitude of the cross-beds ranges from a fraction of an inch to several feet, and amplitudes 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. The limestone masses meet with dolomite along the bedding with sharp transgressive contacts. Figure 3 is a sketch of an outcrop that illustrates the most common relationship between the limestone and dolomite within a single bed.

The Deep Spring formation at the type section along the northwest side of Deep Spring Valley comprises 1,600 feet of quartzitic and calcareous sandstone with a few layers of arenaceous limestone, according to Kirk (Knopf, 1918, p. 24-25). Maxon reports 3,100 feet (1935, p. 314). A brief examination of the strata beneath the Campito sandstone in the area between the type section and the west front of the range provides data that indicate that the dominantly arenaceous strata at the type locality of the Deep Spring formation interfingers laterally with the dominantly dolomitic strata along the west front of the White Mountains. In upper Black Canyon the strata underlying the Campito sandstone consist of interbedded arenaceous dolomite and quartzitic sandstone in about equal amounts, and both lithologic types locally are cross-bedded. Detailed mapping of the intervening area will be required to establish with certainty the equivalence of the strata along the range front with the Deep Spring formation at its type locality, but the readily visible relationships seem sufficiently convincing to make that tentative assumption. On the other hand, Kirk included the well-bedded strata in Black Canyon, here placed in the Deep Spring formation, in the upper part of the Reed dolomite, and concluded that at Marble Canyon (Black Canyon) the Campito sandstone 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 sandstone.

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 type locality of the Johnnie formation is too great to propose any correlation until intervening areas have been studied, but the two formations have certain lithologic similarities and appear to occupy approximately the same stratigraphic interval.

Campito sandstone

The dusky brown to grayish black Campito sandstone is widely exposed along the west front 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 can be seen to overlie the Deep Spring formation. North of Gunter Creek it is exposed in the cores of several anticlines. The most readily accessible extensive exposure of the formation is in Silver Canyon where it has an outcrop width of about 3 miles.

The top and bottom of the Campito sandstone are exposed in different areas, but a complete section was not found. Seventeen hundred feet of strata of the Campito sandstone crop out in sec. 17, T. 7 S., R. 34 E., south of Redding Canyon, between exposures of the underlying Deep Spring formation and overlying lower shale unit of the Silver Peak group, but the Campito sandstone is cut by two reverse faults that have telescoped the section. In the south half of secs. 8 and 9, T. 8 S., R. 34 E., about 2,700 feet of strata in very good order rest on Deep Spring formation and are truncated at the top by a fault. Walcott (1895, p. 143) measured a thickness of 2,000 feet in Silver Canyon, but the base of the formation is not exposed there, and Walcott's description suggests that the top of his section is a fault contact with limestone of the Silver Peak group. Kirk (Knopf, 1918, p. 27) gives a paced measurement of 3,200 feet for the thickness of the formation on the west side of Deep

Spring Valley, which he thought to be fairly accurate. Maxon, however (1935, p. 314), reports a thickness of only 2,000± feet, measured presumably in the same general area. In the construction of the structure sections and block diagram (plates 3 and 4) a thickness of 3,000 feet was assumed.

The formation consists chiefly of dark sandstone, siltstone, and silty shale. Under the microscope the rock can be seen to consist largely of quartz and plagioclase in a matrix of sericite or chlorite. Accessory minerals are magnetite, tourmaline, and zircon. The most common rock is grayish-black, silty sandstone in which bedding planes are very difficult to identify. In silty or sandy layers, intricate cross-bedding with amplitudes generally in the order of less than an inch are common. Locally, in quartzite layers, cross-beds several feet long and with amplitudes of almost a foot are found.

The formation is cut by numerous joints and locally exhibits an imperfect cleavage. The rock breaks along these surfaces and along bedding planes into irregular shaped slabs. The cleavage, ordinarily, is imperfect with wavy cleavage surfaces, although in some argillaceous layers it approaches in perfection that of the Silver Peak slates. Locally, the rock is composed of recemented fragments, the fragments and the matrix being of the same material. These fragmental zones coincide with axes of folding and are thought to be tectonic breccias.

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 thicknesses 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 ribbed with "fucoids". 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 the 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 cross-bedded. Interfaces between thicker sandy beds and thinner silty beds are exceedingly sharp, causing the stratification to be conspicuous. This part of the section crops out boldly because of the sandy beds and is in contrast 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 a larger percentage of 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 stratigraphically overlying shale of the Silver Peak group, but the highest exposed beds are siltier than the typical shale. Lenses and beds that contain carbonates and weather out cavernously, present throughout the formation, are increasingly common toward the top, and also suggest transition toward the overlying shale unit, which contains similar carbonate beds and lenses.

The upper limit of the Campito sandstone is placed by Kirk (Knopf, 1918, p. 27) at the lowest horizon at which fissile calcareous shales and fairly pure masses of limestone appear. In most places the contact with the overlying strata is sheared, which probably accounts for Kirk's mention of limestone as well as shale in the definition of the upper contact. In the NW¼ sec. 21, T. 7 S., R. 34 E., the beds in both the Campito sandstone and in the overlying shale dip gently and the contact appears undisturbed. There, the contact can be picked within a few feet on the basis of grain size. In other places, especially where cleavage is developed in the Campito sandstone, it is difficult to identify the contact with precision. Furthermore, it is not at all unlikely that within the areas represented on the maps as Campito sandstone are infolded masses of Silver Peak slate that more properly belong to the Silver Peak group.

No fossils were found except for branching, tubelike forms near the base of the formation that have been characterized as "fucoids". Kirk (Knopf, 1918, p. 28) reported finding only annelid trails and trilobite(?) tracks. The annelid trails possibly are identical with the "fucoids".

The Campito sandstone probably is correlative with the Stirling quartzite of the Death Valley region and with the Prospect Mountain quartzite of Nevada, as well as with many other sandstone formations in the western United States of Early Cambrian or Late Precambrian age (Wheeler, 1943, p. 1808-1811).

Silver Peak group

The name Silver Peak was applied by Turner (1902, p. 264-265) to strata of Lower Cambrian age in the Silver Peak quadrangle, northeast of the Bishop quadrangle. This usage was extended by Walcott (1908, p. 185-188) to include the fossiliferous series of Lower Cambrian strata in western Nevada and eastern California, and followed by Kirk (Knopf, 1918, p. 28-31) in his description of several thousand feet of limestone, shale, and sandstone in the Inyo Range and White Mountains that carry the genus Olenellus.

According to Kirk (Knopf, 1918, p. 28) the two most complete sections of Silver Peak are north of Waucoba Springs on the northwest side of Saline Valley, which was measured by Walcott (Walcott gives the locality as follows: East of Waucoba Springs, on the Saline Valley road, east of the Inyo Range, Inyo County, California), and along the northwest face of Andrews Mountain, northwest of

Waucoba Mountain. The section described by Walcott near Waucoba Springs has a total thickness of 5,670 feet with Olenellus through 4,900 feet, but apparently the section does not extend downward to the Campito sandstone; consequently an unknown thickness of strata is missing from the base of the group. The section along the northwest face of Andrews Mountain has a total thickness of 6,000 feet, according to Kirk (Knopf, 1918, p. 29), who states that it probably lacks more of the basal portion than that at Waucoba Springs. Kirk has described only the uppermost 557½ feet of the section at Andrews Mountain.

In the area mapped in this study the fossiliferous Lower Cambrian strata are deformed in an exceedingly complex manner owing to their incompetence as compared with the underlying strata. Nevertheless, it has been possible to separate and map a lower slate unit that rests directly on the Campito sandstone, and an overlying limestone unit. It is not unlikely that lithologically similar units higher in the section are infolded with these strata, but have not been recognized.

Lower slate unit

The sporadic distribution of the lower slate unit of the Silver Peak group as shown on the geologic map of the Bishop quadrangle is a good measure of the incompetence of the rock and of the deformation that is concentrated in it. Not only does the slate pinch and swell in an erratic fashion, but in places it is missing entirely, and the overlying limestone unit rests directly on the Campito sandstone. Commonly the bedding in the slate is at an angle to the bedding in the contiguous formations, even where other evidence of deformation is not obvious. North of Poleta Canyon the slate is too disturbed to measure its thickness, but south of Poleta Canyon, where the rocks are less tightly folded, it is about 600 feet thick. This thickness was measured in the SW¼ sec. 16, T. 7 S., R. 34 E., where the beds are nearly flat. At this locality the strata lack cleavage and are better termed shale.

Typically the slate 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 bands and by thicker brownish-gray bands that are both calcareous and arenaceous. Generally slaty cleavage is well-developed and the rock splits readily along cleavage planes. In many places both cleavage and bedding are distinguishable and the cleavage can be seen to lie at all angles to the bedding; only locally are the cleavage and bedding parallel.

Under the microscope the typical grayish-olive slate can be seen to consist of an extremely fine felt of sericite, generally partly altered to chlorite, with variable amounts of quartz in tiny grains. The calcareous beds consist chiefly of calcite and quartz in variable but usually nearly equal amounts, with a little plagioclase, sericite, and rare tourmaline.

Hornfelsed rock in igneous aureoles.--In the vicinity of the stock of porphyritic quartz monzonite in Poleta Canyon and the diorite stock north of Coldwater Canyon the slate was converted to hornfels, which locally is spotted. Unspotted varieties commonly are dark gray on fresh surfaces and weather grayish brown. In appearance they closely resemble finer grained parts of the Campito sandstone and were initially so mapped. They consist chiefly of sericite and quartz with streaks and spots of chlorite, generally a little coarser grained than slaty specimens.

The spotted rocks, fortunately, offer less difficulty in mapping than the unspotted ones. The groundmass is greenish gray to medium dark gray and the spots are grayish black. The spots range from megascopically barely decipherable specks that give the rock a mottled appearance to ovoids a millimeter or more across and 2 to 3 millimeters long. In specimens with larger spots the spots coalesce and commonly make up about half of the rock. The groundmass of these rocks consists of sericite and quartz with scattered grains of magnetite. The ovoids consist chiefly of chlorite developed from sericite and quartz in roundish grains much larger than any in the groundmass.

One specimen from strata on the north side of the Poleta stock consists of approximately equal amounts of quartz, andalusite, and muscovite, with a lesser amount of reddish brown biotite. The rock here initially was mapped as Campito sandstone but the mineralogic composition suggests it is a derivative of the slate. Presumably this rock is the higher temperature equivalent of the spotted hornfels; the specimen was collected from a locality closer to intrusive rock than any of the specimens of spotted hornfels.

A less common kind of hornfels, collected from the vicinity of the diorite stock north of Coldwater Canyon, consists chiefly of zoisite and tremolite in the ratio of 3 to 1, with a small amount of quartz. This rock apparently was developed from somewhat calcareous shale or slate.

Fossils.--Fossils were collected from the slate unit at 3 localities shown on the geologic map of the Bishop quadrangle.

B-1. 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. Identified by G. A. Cooper. Following is his description of the collection.

> 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.

B-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.

Archaeocyathus sp.
Kutorgina sp.

B-3. On the ridge on the south side of the south branch of Gunter Creek, ½ mile north of peak 7824. Identified by Edwin Kirk.

<u>Archaeocyathus</u> sp.
<u>Kutorgina</u> sp.

<u>Correlation</u>.--The stratigraphic position of the shale unit is similar to that of the Pioche shale which rests on lower Cambrian sandstone (the Prospect Mountain quartzite) and is overlain by limestone (the Lyndon limestone of Middle Cambrian age). Further work may well demonstrate that the two shale formations are correlative or partly correlative. Wheeler (1943, p. 1811-1815) has suggested correlation of the Pioche shale with a number of other shale units of the Great Basin and Colorado Plateau.

<u>Limestone unit</u>

The distribution of the limestone unit of the Silver Peak group in the mapped area is almost as erratic as that of the underlying shale. Masses of limestone of varying size and shape are irregularly distributed in the northern two-thirds of the mapped segment of the White Mountains. All of the limestone outcrops in this area, except those within the Deep Spring formation, belong to the limestone unit of the Silver Peak group. In most places, the limestone has been strongly deformed, as the outcrop pattern indicates. Commonly the direction of bedding is clearly recognizable, but in some outcrops, the limestone is massive, and in others cleavage is developed that can be confused with bedding. In mapping, only lithologically distinctive layers were taken to indicate the attitude of the bedding.

The lower contact of the limestone unit with the underlying slate unit is sharp everywhere that it was observed, although in many places the contact is probably sheared. Nevertheless, the contact also is sharp in the SW¼ sec. 16, T. 7 S., R. 34 E., where the beds in both the limestone and slate are little disturbed. The upper contact of the limestone unit is exposed in the adjoining Blanco Mountain quadrangle where the limestone is overlain by slate and sandstone, but is not exposed within the mapped area. In the construction of the accompanying structure sections (plate 3) and block diagram (plate 4) a thickness of 1,200 feet has been assumed, but this thickness may be in error by as much as several hundred feet.

The limestone unit is predominantly limestone, but does include interbeds of calcareous shale or slate similar in appearance to that in the underlying slate unit. Dolomite is present within limestone beds in the form of thin anastomosing layers that lie along bedding, cleavage, and fracture planes, and clearly have been formed by the replacement of limestone, chiefly subsequently to 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 calcite. Weathered surfaces of the dolomite commonly are yellowish orange to light brown, which makes it easy to distinguish dolomite from calcitic limestone. In places archeocyathids and flattened, almond-shaped ovoids (Girvanella?) consisting of calcitic limestone are embedded in a matrix of dolomite.

Fossil collections from the limestone unit were made at only 2 localities, which are shown on the geologic map of the Bishop quadrangle (plate 1).

B-4 North side of Silver Canyon, 1,000 feet east of contact between Campito sandstone on west and limestone unit of Silver Peak group on east. In green shale within limestone unit. Possibly the same as Kirk's locality 7 (Knopf, 1918, p. 31). Identified by Edwin Kirk.

Archaeocyathus sp.

B-5 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.

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

Metamorphic remnants in the Sierra Nevada

Remnants of metamorphic rocks in the salient of the Sierra Nevada that extends into the quadrangle are considered in three groups. One group includes several areally disconnected masses of metamorphic rock that are distributed along the lower slopes of the east-trending segment of the Sierra front. Inasmuch as most of these masses are bounded on one side (commonly the north side) by alluvium, their true relationships are not apparent. Conceivably most of them might be fingers of a large pendant or even of the main wall rock on the east side of the batholith, or they could be a series of isolated or semi-isolated inclusions surrounded on all sides by granitic rock of one kind. The second group includes several metamorphic septa along the contact between orthoclase-albite granite and quartz monzonite that extends southerly through the salient.

The third group consists of a single mass in the southwest corner of the quadrangle that is merely a corner of a large mass of metamorphic rock known as the Bishop Creek pendant.

Most of the metamorphic rocks were derived from sediments, but a septum four miles west of Keough Hot Springs consists chiefly of meta-andesite. The metasedimentary remnants are tentatively assigned to the Paleozoic and the predominantly metavolcanic remnant west of Keoughs Hot Springs to the Mesozoic on the basis of the lithologic relations of unmetamorphosed strata in the ranges to the east. In the Inyo and White Mountains, strata of Paleozoic and Lower Triassic age contain little or no volcanic material, whereas strata that overlie fossiliferous Middle Triassic metasedimentary strata are composed predominantly of volcanic material (Bateman and Merriam, 1954).

The metamorphic rocks were subjected to thermal metamorphism during the emplacement of the granitic rocks, and almost all of the resulting mineral assemblages are in the amphibolite facies of Eskola (1939; see also Turner, 1948; and Barth, 1952). Amphiboles and plagioclase are common in all of the rocks of appropriate composition. Although diopside is common in rocks with a high calcium content, the association diopside-hypersthene, diagnostic of the next higher pyroxene hornfels facies, was not found. Locally plagioclase has been saussuritized, but such regressive effects are not widespread. In terms of Bowen's 13 steps in the progressive metamorphism of siliceous dolomite (Bowen, 1940, p. 225, 274) the mineral assemblages indicate that the metamorphism was above step 5, everywhere below step 8, and in most places below step 6. Dolomite, stable below step 5 was not recognized, and the assemblage calcite plus diopside, stable below step 8, is common. The assemblage quartz plus calcite, stable below step 6, commonly occurs in apparent equilibrium, although locally, especially in the marginal parts of the metamorphic masses wollastonite has formed.

On the map the metasedimentary rocks are subdivided on the basis of the dominant rock within a unit of mappable extent into siliceous hornfels, calc-hornfels, and marble.

Siliceous hornfels

Under the designation siliceous hornfels are included all of the metasedimentary rocks with low lime content. Most of them are fine-grained quartz-bearing rocks that were derived from impure silt or sand. Much of the siliceous hornfels that extends westward along the range front from the Bishop antimony mine is dark-gray rock that consists of a fine granoblastic intergrowth of quartz and potassium feldspar with abundant disseminated carbonaceous material, accompanied by minor sphene, tremolite, and sericite. Locally this rock contains thin light-colored layers of similar texture and mineral content except that they contain poikiloblastic diopside crystals and are almost devoid of carbonaceous material. Obviously, these layers are richer in lime than the country rock. Most of the light-colored layers are parallel with obscure bedding planes in the country rock, but locally such layers cut across the bedding, indicating that lime was introduced into the rock, presumably along fractures, prior to the metamorphism.

Other somewhat different kinds of rocks included under siliceous hornfels occur in small scattered metamorphic remnants. The inclusion at the Rossi mine contains grayish red micaceous quartzite that consists almost entirely of roundish quartz grains and well-oriented plates of biotite. The inclusion at the Chipmunk mine contains a unit that consists of thinly interbedded very light gray wollastonite-rich layers and dark-gray quartz-rich layers. Small diopside grains generally are present in the wollastonite-rich layers in amounts that vary from layer to layer. Quartz-rich layers also contain plagioclase and potassium feldspar, plus minor sphene, calcite, and pyrite.

South of the dominantly metavolcanic septum, along the south boundary of the quadrangle, is the northern tip of another septum that consists chiefly of feldspathic and micaceous quartzite. The rock within the quadrangle is chiefly feldspathic quartzite, and consists of detrital quartz, biotite, sericite, potassium feldspar, and epidote in proportions that vary from layer to layer, plus minor garnet, sphene, amphibole, apatite, and ilmenite.

Calc-hornfels

The largest mass shown as calc-hornfels on the map is the corner of the Bishop Creek pendant in the southwest corner of the quadrangle. The rock here is extremely fine grained and platy, and ranges in color from pale yellowish gray to light olive gray. Under the microscope the rock is seen to consist of a fine-grained mat of tremolite, quartz, and potassium feldspar with scattered grains of sphene. The mineral content indicates that the rock very likely was formed from a calcareous shale. Near the center of sec. 17, T. 8 S., R. 32 E. worm borings stand out in bas-relief on weathered bedding plane surfaces. According to C. A. Merriam, who examined the borings, similar borings are abundant in the Great Basin in strata of Cambrian and Ordovician age (oral communication).

A second smaller mass is shown on the map in the septum four miles west of Keough Hot Springs. The strata there are dominantly hornfelsic, but include thin layers of marble and a single thicker layer of somewhat calcareous metaconglomerate. The calc-hornfels is dense, fine-grained, dirty white to greenish rock that consists chiefly of diopside and plagioclase. The metaconglomerate is composed of pebbles of felsic metavolcanic rock in a matrix of lime silicate minerals. The pebbles consist chiefly of a very fine granoblastic intergrowth of quartz and potassium feldspar in which broken crystals of quartz, potassium feldspar, and plagioclase are set. The matrix around the pebbles consists of quartz, potassium feldspar, epidote, tremolite, and diopside.

Marble

Most of the masses composed predominantly of marble lie along the north side of the salient of the Sierra Nevada. The marble commonly is coarsely crystalline and white or light gray; marble at the Rossi mine is light bluish gray. Commonly the marble contains intercalated layers of calc-hornfels derived from shaly or siliceous layers. The calc-hornfels generally consists of diopside, tremolite, plagioclase, and potassium feldspar in various combinations and proportions. Locally garnet is present, and idocrase, though not positively identified, is probably present.

Meta-andesite

Metamorphosed mafic volcanic rock of andesitic composition comprises the bulk of the septum four miles west of Keough Hot Springs. Most of the rock is massive and dark gray, but part of it is amygdaloidal and is mottled and veined with light yellowish-gray diopside-plagioclase rock. The massive dark-gray rock consists of unoriented zoned plagioclase (An_{30-65}) plates or laths as much as 2 mm long, embedded in a fine-grained groundmass of plagioclase, biotite, and hornblende, with minor sphene, magnetite, and microcline. Most of the groundmass is granoblastic, but locally, where biotite is abundant, the texture of the groundmass is decussate. The rock is cut by thin quartz veinlets, which, like the quartz cores of the amygdules, are bordered by pyroxene or hornblende.

The mottled appearance of the rock was caused by the local conversion of hornblende to almost white diopside, thus producing yellowish-gray rock. The pyroxene-bearing zones generally are marginal to quartz-diopside veinlets, but in some outcrops many square feet of the exposed rock, except for small residual areas, are converted to diopside-plagioclase rock (fig. 4a). Usually the original felty texture of the andesite is preserved in the diopside-

Figure 4a. Mottled meta-andesite partly altered to calc-hornfels in septum 4 miles west of Keough Hot Springs. Dark areas represent residuals of metavolcanic rock.

Figure 4b. Closely-spaced joints in granite in Rawson Canyon.

Figure 4a.

Figure 4b.

plagioclase rock, but locally the texture is granoblastic and the rock is true calc-hornfels. In the conversion to calc-hornfels the feldspar has lost its zonal structure.

The yellowish-gray diopside-plagioclase rock apparently has been formed in places where lime was introduced into the original igneous rock. In the amphibolite metamorphic facies any lime in excess of the amount needed to form tremolite will combine with tremolite to form diopside. The mol ratio CaO:MgO is 1:1 in diopside and 2:5 in tremolite. This relationship is shown graphically on various ACF diagrams of the amphibolite faces of metamorphism (for example Barth, 1952, figs. 132 and 133, and Turner, 1948, fig. 17). The association of pyroxene-bearing zones with the quartz-diopside amygdules and veins suggests that lime was introduced into the rock along with silica in the formation of the amygdules and veins, and that some lime permeated the contiguous rock. This permeation was not necessarily at the time of initial introduction--in the course of the thermal metamorphism of the rock by the intrusive granitic rocks the lime in excess of that needed for the formation of diopside may have migrated laterally from the veins and amygdules until it was used up in the conversion of hornblende to diopside.

<u>Quartz diorite, diorite, and hornblende gabbro</u>

The rocks grouped under the general heading of quartz diorite, diorite, and hornblende gabbro are mafic and darker hued than any of the granitic rocks. In the Sierra Nevada small masses are scattered through the dominantly granitic terrane; in the White Mountains a stock of altered diorite crops out on the north side of Coldwater Canyon, and diorite dikes are abundant north of Silver Canyon.

The diorite in the White Mountains is equigranular, and has an average grain size of about 1 mm. The rock is of much the same appearance in both the stock and in dikes. In thin section it can be seen to be highly altered; the alteration was essentially isochemical. The rock consists chiefly of unzoned albite (An_{10} approx.), chlorite, and epidote-clinozoisite, with albite making up about half the rock. 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, giving the rock a panidiomorphic-granular texture. Undoubtedly the original feldspar was more calcic than albite; rare unaltered parts of grains indicate it was in the andesine range. The form of much of the chlorite and epidote suggests that they were derived at least in part from original biotite and hornblende respectively, although some epidote-clinozoisite must have been formed from plagioclase as a byproduct of its alteration to albite. The intrusive relationships of the stock and dikes indicate the diorite crystallized from a magma.

The mafic rocks of the Sierra Nevada are extremely variable in texture, grain size, and color index, although most of the rocks are medium grained. The darker hued ones generally are hornblende gabbro and the lighter hued ones quartz diorite or mafic granodiorite. Masses of hornblende gabbro commonly consist of about equal amounts of hornblende and calcic plagioclase, together with several percent of magnetite, apatite, and sphene. Secondary minerals include epidote, chlorite, and sericite. The plagioclase in these rocks is in strongly and progressively zoned almost euhedral crystals that give the rock a panidiomorphic-granular texture. Commonly the plagioclase ranges from bytownite in cores to calcic oligoclase or sodic andesine in rims; discontinuities and reversals in the zoning are common. Hornblende generally is anhedral, and has a pleochroic formula X=grayish yellow, Y=light olive brown or pale olive, Z=dusky yellow green, locally grayish green. Augite cores are present in a few grains. In some exposures the rock is conspicuously layered; light-colored layers rich in plagioclase alternate with dark-colored layers rich in hornblende. The panidiomorphic-granular texture, deep zoning of the plagioclase, presence locally of layered facies, and occurrence of dikes cutting metamorphic rocks indicate that the rock crystallized from magma and is truly igneous.

Somewhat lighter-colored quartz diorite ranges in texture from fine equigranular to coarse inequigranular with poikilitic hornblende crystals as long as an inch. Most masses of quartz diorite, however, are equigranular and hypidiomorphic-granular. Plagioclase commonly makes up 40-65 percent of the rock, hornblende about 15 percent, biotite 15-20 percent, quartz 5-15 percent; accessories include apatite, sphene, magnetite; the usual secondary minerals are epidote, chlorite, and sericite. Plagioclase commonly is progressively zoned, but with oscillations and discontinuities or strong changes in composition through narrow zones. Many crystals have a broad central zone with 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 Z generally is grayish blue green.

In texture and in composition the quartz diorites are transitional between hornblende gabbro and the granitic rocks. Almost certainly most of them are igneous, but some appear to be granitized amphibolite that was derived either from mafic metavolcanic rocks or from calcic metasedimentary rocks. The metamorphic grade of the metamorphic rocks is in the amphibolite facies, so that any rock of appropriate composition could be converted to a stable assemblage of hornblende and plagioclase. A few miles south of the quadrangle is granitized amphibolite that was derived from andesite which is very similar in appearance to other diorites and quartz diorites. Less strongly granitized specimens of this rock contain euhedral to subhedral plagioclase crystals set in a finer grained groundmass that contains more sodic plagioclase. Locally this rock contains quartz amygdules that are rimmed with amphibole and these amygdules are preserved in even the most strongly granitized areas. In most places where amphibolite is associated with calc-hornfels the hornfels has been formed from the amphibolite, but in places, especially in the margins of inclusions in granite, calc-hornfels or marble has been altered to amphibolite. Alteration of calc-hornfels or marble to amphibolite must have involved metasomatic

exchange of material between the calcareous rock and the granitic magma. The variety of calc-hornfels associated with amphibolite is one that consists chiefly of diopsidic pyroxene and plagioclase. Consideration of the compositions of diopsidic pyroxene and common hornblende indicates that addition of water and subtraction of lime are the principal changes required in the conversion of the diopsidic pyroxene to hornblende. Formation of amphibolite from marble, of course, involves much greater transfer of materials. Amphibolite selvages between calcareous metamorphic and granitic rocks can be seen at the Brown prospect, which is a mile west of the Bishop Antimony mine.

Approximate contemporaneity of the diorite in the White Mountains with the mafic igneous rocks of the Sierra Nevada is suggested by the known intrusive relationships. The diorite stock and dikes of the White Mountains cut across folds and cleavage in the sedimentary rocks, which probably originated just prior to the intrusion of the granitic rocks, as also do hornblende gabbro dikes in a mass of metasedimentary rocks in the Sierra Nevada west of the Bishop quadrangle. The mafic igneous rocks of the Sierra Nevada are older than the granitic rocks, which intrude them. Mafic rocks derived from amphibolites also are intruded by the granitic rocks with which they are in contact, but their present appearance must date from the intrusion of the granitic rocks.

Calc-alkaline granitic rocks

Calc-alkaline granitic rocks constitute a large proportion of the salient of the Sierra Nevada that extends into the southwest corner of the quadrangle. In the White Mountains the only mass considered granitic in the sense in which the term is used in this report is a stock at the head of Poleta and Redding Canyons. Three different granitic formations have been distinguished, each of which is in sharp contact with the other two. These are called simply granodiorite, quartz monzonite, and orthoclase-albite granite, since only one formation of each compositional group was recognized. These compositional names are used in accordance with the nomenclature of Johannsen (1931, p. 141-161).

Each formation occurs in one or more bodies. Composition, texture, and intrusive relationships are the principal features that were used to distinguish one formation from another and to correlate separated masses of similar rock with one another. Correlation of separated masses is not without hazard, but erroneous correlations seem unlikely because the range in texture and composition in one formation generally is small compared with the range of these characteristics in the entire suite of granitic rocks. Within the quadrangle, contacts between the intrusives are remarkably plain and featureless and yielded no relationships that establish the sequence of emplacement. However, contact relationships outside the quadrangle between these same intrusives indicate that they were emplaced in the usual order--granodiorite first, quartz monzonite second, and orthoclase-albite granite last.

Foliation is conspicuous only in the granodiorite, presumably because it contains more abundant mafic minerals and small tabular mafic inclusions, which are rare in the quartz monzonite and granite. Hornblende is present in the quartz monzonite and granite only in contaminated parts.

All of the rocks are cut by a regional joint system that consists of two steeply-dipping sets--one trending northeast and one trending northwest. The joints are continuous across boundaries between intrusive masses and thus are of later origin than the granitic rocks. Nevertheless, because the joints are more closely spaced in finer grained rocks, the spacing of joints is useful in the recognition of different granitic rocks from a distance.

Granodiorite

Distribution and general character

Parts of two bodies of granodiorite lie within the quadrangle. The larger area of about 3 square miles in the southwest corner is part of a mass that has a total area of about 10 square miles. The smaller area of about half a square mile is near the western edge of the quadrangle one and a half miles north of Bishop Creek in secs. 7 and 8, T. 7 S., R. 32 E.; it constitutes about two-thirds of a much smaller mass. The rocks in the two masses are similar in appearance and in mineral content, except for larger average grain size in the larger mass.

Typical rock is light gray, finer grained than most other granitic rocks in the region, equigranular, and generally is either unfoliated or weakly foliated. The average color index is 13, and most specimens fall between 8 and 15. Biotite generally is in euhedral plates. The average grain size in the larger mass is about 2 mm, whereas it is only about 1 mm in the smaller mass. Foliation that is parallel to the margins of the mass is most conspicuous in marginal parts of the larger mass, chiefly because of an abundance of tabular mafic inclusions. Regional joints are more closely spaced in the granodiorite than in the coarser grained quartz monzonite or orthoclase-albite granite, and are less conspicuous because the rock splits along these joint surfaces and breaks down into rubble.

Petrography

The mineral content of the rock in the two masses is approximately the same, although differences are recognizable in texture and in the characteristics of certain minerals. A typical specimen contains about 50 percent of plagioclase, 5 percent of quartz, 15 percent of potassium feldspar, 7 percent of biotite and 3 percent of hornblende. Accessory minerals include apatite, sphene and zircon; alteration products are epidote, chlorite and sericite.

The rock in the smaller mass is hypidiomorphic granular, whereas that in the larger mass is poikilitic. In the larger mass large anhedral grains of quartz and potassium feldspar with quadrille structure form a mesostasis in which euhedral to subhedral grains of the other minerals are enclosed. In both masses biotite and hornblende are euhedral to subhedral, plagioclase is subhedral, and quartz and potassium feldspar are anhedral.

Plagioclase is conspicuously zoned in the andesine range, with an uncommonly large range of An_{20} to An_{50} in small euhedral crystals in the larger mass. The principal trend of the zoning is from calcic cores to sodic rims, but superimposed oscillations are common. The calcic cores of many grains contain sericite or hydromica that presumably formed from excess potassium in the calcic plagioclase. Discontinuous marginal rims of albite, locally present along segments of plagioclase grains that are in contact with potassium feldspar, presumably were derived from the potassium feldspar in accordance with the mechanism described by Tuttle (1952, p. 115-116 and plates 3 and 4).

Potassium feldspar in the smaller mass rarely exhibits quadrille structure and is not notably perthitic, although it does contain thin lamellae of albite; the poikilitic grains in the larger mass, on the contrary, exhibit conspicuous quadrille structure and are perthitic. Quartz generally extinguishes without undulation. Hornblende is of the ordinary variety--X=grayish green, Y=olive, and Z=grayish blue green. The Z direction of biotite is grayish red in the larger mass and grayish brown in the smaller one.

Quartz monzonite

Distribution and general character

About 11 square miles in the west part of the salient is underlain by quartz monzonite that extends beyond the quadrangle boundary. Commonly the rock is medium light gray, medium grained, and has a porphyritic texture. Phenocrysts of perthitic potassium feldspar as much as a centimeter across and 2 centimeters long are set in a groundmass with an average grain size of 2 to 3 millimeters. The average color index is about 6, and most rocks fall within a range of 4 to 10. The mafic minerals, chiefly biotite, are generally in clusters of anhedral grains that are sporadically distributed through the rock. Peripheral concentrations of mafic minerals not uncommonly provide frames for the potassium feldspar phenocrysts. More rapid weathering of the groundmass further accentuates the prominence of the phenocrysts.

Exposures of quartz monzonite within the quadrangle are virtually structureless. Mafic inclusions are rare, except locally adjacent to diorite and hybrid mafic rocks. The regional joints are widely spaced and weathering along them has produced bold rounded forms that characterize much of the area of quartz monzonite.

Petrography

The rock generally consists of roughly equal amounts of quartz, plagioclase and perthitic potassium feldspar. Biotite is present in amounts that generally range from 3 to 8 percent. Hornblende is absent from most specimens, and where present rarely exceeds 2 percent of the rock. Common accessories are magnetite, ilmenite, sphene, apatite, allanite, zircon, and thorite. Alteration products are epidote, chlorite, hematite and limonite, and sericite or hydromica.

Plagioclase generally is in subhedral zoned crystals with an average composition of about An_{30}. Both gradational zoning from a calcic core to a sodic rim and superimposed oscillatory zoning are common. Potassium feldspar is both in large phenocrysts that are twinned on the Carlsbad law and in smaller interstitial masses. Commonly it exhibits quadrille structure and contains about 10 percent of albite in perthitic lamellae. Phenocrysts generally enclose euhedral- to subhedral-crystals of zoned plagioclase, hornblende, and magnetite, whereas smaller masses of potassium feldspar are interstitial to these minerals. Both relationships indicate late crystallization of perthite. The zone of mafic and accessory minerals peripheral to many phenocrysts is interpreted as material that was expelled from the potassium feldspar during its growth.

Quartz generally is in large grains that either exhibit undulatory extinction or are composed of a granoblastic mosaic of differently oriented components. Both structures are thought to be the result of strain that was imposed after crystallization. Undulatory extinction and internal polygonal patterns that are visible when a grain is near the extinction position are thought to result from less intense strain than that which produced granoblastic mosaics. All gradations between large crystals with undulatory extinction and quartz grains that consist of granoblastic mosaics can be found.

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

Cataclastic or protoclastic effects are evident in some specimens but are by no means widespread. The most reliable evidence of cataclasis is fine granoblastic mortar between grains, but myrmekite and strained or granoblastic quartz are most abundant in specimens where granoblastic mortar is conspicuous, and the formation of both is thought to have been conditioned by strain. Myremekite is common along boundaries between plagioclase and potassium feldspar; where thin bands appear in thin section to be enclosed in potassium feldspar plagioclase is inferred in the third dimension.

Granite

Distribution and general character

Orthoclase-albite granite underlies about 15 square miles of the east side of the salient of the Sierra Nevada within the quadrangle boundaries and extends southward beyond the limits of the quadrangle. In the White Mountains rock that is somewhat similar to the Sierran granite comprises a stock at the head of Poleta and Redding Canyons. Only about half a square mile of the western side of this stock, called the Poleta stock, lies within the quadrangle.

The typical granite from the Sierra Nevada is megascopically equigranular with an average grain size of 3 to 4 millimeters. Much of the rock is deeply weathered to a loose iron-stained aggregate. Where the granite is least altered the potassium feldspar generally has a pinkish cast, but this may be a product of incipient weathering rather than a primary characteristic. The rock is notably felsic and generally contains less than 2 percent of biotite as the only dark mineral. Mafic inclusions are exceedingly rare. Joints commonly are widely spaced and consequently conspicuous though locally they are spaced as close as a few inches (fig. 4b).

The rock in the Poleta stock in the White Mountains differs in appearance from the Sierran granite in being porphyritic and in having a slightly higher content of biotite. Abundant subhedral phenocrysts of potassium feldspar 3 to 5 millimeters across are set in a finer grained matrix with an average grain size of about 2 millimeters.

Petrography

The granite of the Sierra Nevada contains approximately equal amounts of quartz, potassium feldspar and sodic plagioclase. The texture is hypidiomorphic granular, and although the rock appears megascopically equigranular, thin sections reveal it to be seriate.

Commonly the plagioclase is in subhedral grains that are normally zoned within the range of An_7 and An_{13}, with an average composition of about An_{10}. Knopf (1918, p. 67-69) originally determined the albitic composition of the plagioclase. His optical data also indicate an average composition of An_{10}, although a partial chemical analysis of a rock collected by him from Rawson Canyon indicates a normative composition of An_6. Knopf pointed out, however, that An_{10} may be closer to the true average composition than An_6 because the potassium feldspar contains lamellae of almost pure albite. The plagioclase in the Poleta stock is more calcic than in the Sierran granite, and has an average composition of about An_{18}. Thus, the rock in the stock is properly quartz monzonite rather than granite.

Potassium feldspar commonly is anhedral, perthitic, and exhibits quadrille structure. Much of it contains small grains of plagioclase and accessory minerals, but rarely of quartz. Quartz is in anhedral grains that extinguish sharply. Biotite is in subhedral plates that generally are evenly distributed through the rock. The color in Z commonly is moderate brown, but in a few specimens it is moderate yellowish brown or moderate olive. Common accessory minerals are zircon, apatite, magnetite, and sphene.

Volcanic rocks

Olivine basalt

Scattered masses of olivine basalt in the form of flows, necks, and cinder cones are present in the southwest part of the quadrangle. The typical rock in flows and shallow intrusives is dark gray to black on fresh surfaces, and consists of large olivine and more numerous but smaller and less conspicuous augite phenocrysts in a fine-grained matrix of plagioclase, augite, and magnetite. Cinders are grayish red to moderate red, which makes them easy to distinguish from other basaltic rock even at a distance.

All outcrops of basalt have been eroded, and are considered to be of Pleistocene age or older. The deep dissection of some flows and the exposure of basaltic necks required the removal of large volumes of rock after the basalt was extruded or emplaced. Basalt of several ages may be represented, and some may have been extruded prior to the inception of the structural movements by which Owens Valley was downdropped relative to the bordering ranges. However, the basalt that can be dated is younger than the earliest of these movements. Structural movements earlier than basalt are indicated by the localization of some feeders along Basin and Range faults and by lava that flowed down slopes toward Owens Valley.

Two rounded basalt hills crop out west of Bishop, one just west of Oteys Sierra Village and the other north of the lumber mill on the north side of U. S. Highway 395. Presumably each of these hills marks the locus of a feeder. In the NE¼ sec. 17, T. 7 S., R. 32 E., on the north side of the hard-surface road along Bishop Creek, is a basaltic neck that has become well known through descriptions of Knopf (1918, p. 74-75). Here, fingers of basalt intrude quartz monzonite and locally have caused refusion of the quartz monzonite and the production of glass. Both the basalt and the adjacent refused quartz monzonite have conspicuous columnar joints that are continuous from one rock into the other.

Farther south, along the north slope of the Sierra Nevada and west and southwest of the Chipmunk mine are several remnants of a once extensive flow that has been dissected by the headward erosion of Coyote Creek and an unnamed creek that flows past the Chipmunk mine. Locally, this basalt is capped by glacial till or outwash of pre-Tahoe age. The distribution of basalt remnants suggests that the flow was localized in a broad north-trending valley, probably the ancestral valley of Coyote Creek.

Other small flows and necks are exposed south of the Bishop Antimony mine. Most of the necks and the feeders for the flows are along faults. One particularly youthful-appearing flow in the SE¼ sec. 26, T. 7 S., R. 32 E. extends N. 80° E. 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 the cone and the flow, considerable antiquity is indicated by the fact that 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 flow about N. 45° E.

Bishop tuff

The Bishop tuff crops out in the northeast corner of the quadrangle and continues southward into Owens Valley where it is intercalated with valley fill at increasingly greater depths southward. Discontinuous segments of what is believed to be a thin basal member of the tuff also crop out locally in the alluvial deposits that lie along the west side of the White Mountains.

The Bishop tuff was named and described by Gilbert (1938, p. 1829-1862) who concluded that it 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. Gilbert's most detailed studies were in the Rock Creek and Owens River Gorges, several miles west of the outcrops in the Bishop quadrangle, where the exposed tuff differs in several ways from the tuff within the Bishop quadrangle. In the gorges, thick sections of tuff are progressively denser with depth, but the base of the tuff is not exposed. In the Bishop quadrangle tuff that is more or less equivalent to that exposed in the gorges is much thinner and none of it is as dense as tuff in the lower parts of the gorges; furthermore, two lower members of the tuff not described by Gilbert are present, and the base of the formation is exposed.

The tuff within the Bishop quadrangle was subdivided into 3 readily-distinguishable members--a thin basal member composed of tightly-packed, but unwelded or uncemented angular white pumice fragments, an overlying thicker unconsolidated tuff member, and an upper consolidated member (fig. 5a).

Pumice layer

The basal pumice layer is best exposed in the cliff along the south side of the Volcanic Tableland and in quarries along the east side. The pumice underlies the higher members of the tuff in a

Figure 5a. Abandoned pumice pit on east side of Volcanic Tableland. Lower half of face of pit is basal pumice layer of Bishop tuff; upper half is unconsolidated tuff.

Figure 5b. Basal pumice layer of Bishop tuff showing overall upward increase in size of fragments and superimposed minor variations.

Figure 5a.

Figure 5b.

continuous sheet that ranges from 15 to 20 feet in thickness. Borings into the floor of Owens Valley for water show that the pumice layer extends under the valley at increasingly greater depths southward. The most southerly boring to intersect the pumice layer was made in the north-central part of sec. 23, T. 7 S., R. 33 E., where the base of the pumice is at an altitude of 3,300 feet--more than 700 feet beneath the floor of the valley (plates 5 and 6). None of the borings farther to the south was deep enough to intersect the pumice layer, if present there. The distribution of pumice in borings suggests that the pumice layer is continuous beneath the central part of the valley and that it has a cuspate margin, being convexly indented at larger alluvial fans. Discontinuous masses of pumice exposed intermittently in the alluvial deposits of the White Mountains are intercalated in alluvial detritus. Locally, some pumice rests directly on bed rock rather than on alluvial detritus. Some of these masses of pumice crop out because of faulting, but the pumice masses appear to be present only in small protected basins; subsurface continuity is unlikely.

In the south and east sides of the Volcanic Tableland the pumice layer rests on sands and gravels. The pumice here, as in most outcrops in the White Mountains, consists almost exclusively of angular, unrounded fragments of pumice that range from sand-size to 2 inches in average diameter. The pumice fragments are unaltered, firm, and perfectly white. Angular fragments of foreign material are sparsely and sporadically distributed through the member. 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 well-exposed outcrops along the White Mountain front, is gradational increase upward in the average size of the pumice fragments (fig. 5b). Ordinarily, beds in which the size of the fragments increases upward are considered to be overturned, but regional relationships leave no doubt that the pumice layer is right-side-up. The complete angularity of the fragments precludes transportation of the pumice in any way except through the air, but within this limitation two hypotheses merit consideration as providing an explanation for the inverse size gradation of the fragments.

The first hypothesis is that the mechanics of intrusion produced more finely comminuted fragments at the beginning of an explosive spasm than toward the end, and that the ejected material fell on dry land. Wentworth and MacDonald (1953, fig. 38 and p. 73) have suggested this mechanism to explain the upward increase in the size of basaltic fragments in the flanks of a cinder cone. This hypothesis seems to explain all of the features of the pumice layer and is favored as the most likely one even though it requires special conditions at the source.

The alternate hypothesis and one that formerly was favored by the writer is that pumice of random size blown into the air fell into a body of standing water (Bateman, 1953, p. 1499-1500). All of the pumice fragments would have floated for a period, but the smaller pumice fragments would have become saturated first and progressively larger fragments would have sunk with time. To test the workability of 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, but it was several hours before the larger pumice fragments sank. The pumice collection in the beaker was rudely layered, and the fragments were graded in size from small at the bottom to large at the top. This mechanism has the advantage over the first hypothesis of not requiring special conditions at the source. Nevertheless, several objections to the hypothesis have been brought to the attention of the writer. Possibly the most significant objection is that accidental fragments sporadically distributed through the pumice also appear to be larger toward the top of the layer. The second objection is that along the White Mountain front pumice with inverse size gradation of fragments in places rests directly on Cambrian formations with no intervening layer of sediment. Completely clean bedrock surfaces in a body of standing water must certainly be regarded as rarities; where pumice rests on clean bedrock surfaces it seems most likely to have been deposited subaerially.

Unconsolidated tuff

Unconsolidated, pale-pink tuff overlies the pumice layer in the margins of the Volcanic Tableland and probably also beneath Owens Valley. Drillers' logs of borings into Owens Valley show that the part of the Bishop tuff above the pumice is thinner beneath the Valley, but do not permit distinction between unconsolidated tuff and consolidated tuff. Nevertheless, it is a reasonable supposition, based on relationships along the south edge of the Tableland and on exposures of unconsolidated tuff in the terraces on the south and east sides of the Tableland and in a quarry in the SE¼ sec. 14, T. 6 S., R. 33 E. at the base of the White Mountains, that all of the tuff that overlies the pumice layer beneath Owens Valley belongs to the unconsolidated tuff member.

In Fish Slough tuff in the stratigraphic position of the unconsolidated tuff is semi-consolidated and physically intermediate between typical consolidated and typical unconsolidated tuff. This tuff is included on the geologic map (plate 1) with consolidated tuff. The areal relationships suggest that the unconsolidated tuff grades to the north and toward the interior of the Volcanic Tableland to consolidated tuff. Thus the unconsolidated tuff represents the marginal and distal parts of a layer that elsewhere is consolidated.

In the margins of the Volcanic Tableland the unconsolidated tuff rests on the pumice layer with a very sharp and regular contact. At the top it grades through about 15 feet into overlying consolidated tuff. The unconsolidated tuff is about 200 feet thick at the west edge of the quadrangle, and it appears to thin toward the east.

Typically the unconsolidated tuff consists of rounded pumice fragments as much as several inches in diameter in an ashy matrix. In addition to ash and pumice fragments the matrix contains rounded pellets of obsidian and discrete crystals of quartz and sanidine, and less commonly accidental fragments of granitic and metamorphic rock, or basalt. Crystals of quartz and sanidine occur as phenocrysts in the pumice fragments. Locally, accumulations of larger pumice fragments in mounds as much as 10 feet high are present in the base of the formation. An interesting color variation occurs in the east margin of the Volcanic Tableland where the tuff adjacent to a tongue of consolidated tuff has a bright pink color caused by the oxidation of small amounts of iron.

In contrast to the pumice fragments in the underlying basal pumice layer the pumice fragments in the unconsolidated tuff member are physically much weaker, apparently because of larger less symmetrical vesicles with thinner walls, and are rounded rather than angular. The roundness of the pumice fragments indicates abrasion, but lack of sorting, the presence of only sparse exotic material, the widespread areal distribution, and the apparent lateral gradation of unconsolidated to consolidated tuff indicates that the transporting agent was unlikely to have been water. The general appearance of the member suggests that it was emplaced as a nuée ardente, which along its margins and at its distal end lacked sufficient heat to weld itself.

Consolidated tuff

The cap rock of the Volcanic plateau is hard, consolidated tuff that is extremely resistant to erosion. To the north and west exposures of consolidated tuff as much as 500 feet thick can be seen in the gorges of Rock Creek and the Owens River, but in exposures along the margins of the Volcanic Tableland within the Bishop quadrangle the cap rock is only about 50 feet thick.

Most of the cap rock is homogeneous and appears to consist of a single layer. The typical rock is light brown consolidated tuff that owes its coherence more to the formation of tiny crystals of tridymite or cristobalite across fragment boundaries than to welding through partial melting. Stratigraphically lower layers exposed in the Owens River and Rock Creek Gorges are truly "welded", but the tuff here cannot be termed a "welded" tuff without undesirable broadening of the term. Fenner (1948, p. 883) has proposed the term "sillar" for agglutinated tuffs of similar description and Marshall (1932, p. 198-202, and 1935, p. 323-366) suggested "ignimbrite" for tuffs formed by deposition from glowing clouds of intensely heated minute fragments (nuées ardentes). The term "ignimbrite" has been criticized because it is based on an interpretation of genesis, and because of growing doubt, expressed in recent publications, that all agglutinated tuffs, and especially those that are truly "welded", have originated from nuées ardentes.

The semi-consolidated tuff along Fish Slough is included with the consolidated tuff on the map, but appears to grade laterally, toward the south and east, into unconsolidated tuff.

Petrographically, the consolidated tuff is much like the unconsolidated tuff except for greater compaction and higher density. A specimen from an old quarry in the SE¼ sec. 14, T. 6 S., R. 32 E. has a specific gravity of 1.49. Pumice fragments have been only slightly flattened, and shards are readily visible.

<u>Age</u>

At several places north of the Bishop quadrangle the Bishop tuff is locally interstratified with tills and therefore is of Pleistocene age. These tills have been intensively studied by Putnam (1938, p. 68-82; 1949, p. 1281-1302; 1950, p. 115-122; 1952, p. 1291), and he has concluded that the Bishop tuff rests on till of the Sherwin stage, and is overlain by moraines of the Tahoe stage. In the past some confusion existed as to the relative ages of the Bishop tuff and the Sherwin till, but Putnam, in his most recent publication that deals with the subject, (Putnam, 1952, p. 1291) states, "the Sherwin glacial stage is older than the Bishop tuff rather than younger as evidence in the Mono Lake region suggested." This conclusion is based on examination of the logs of cores and of exposures in a tunnel driven by the Los Angeles Department of Water and Power in the west wall of the Owens River gorge and seems likely to stand.

Sedimentary deposits

Moraine and till

Glacial deposits are present at three localities in the southwestern part of the quadrangle. The lowest and the only one whose form is constructional rather than erosional is the end of an extensive moraine that follows the canyon of Bishop Creek; the lowest point on this moraine is at about 5,000 feet. The next lowest glacial deposits are remnants of bouldery gravels about 3 miles farther south, on the northeast side of Coyote Creek at altitudes of 7,000 to 8,000 feet. The highest is an extensive sheet of bouldery gravel in the southwest corner of the quadrangle that lies at 8,400 to 10,000 feet within the quadrangle and extends to higher altitudes beyond the quadrangle boundaries.

The deposit along Bishop Creek is the terminal part of a moraine that is believed to be correlative with the Tahoe stage of Blackwelder (1931, p. 881-884). The original form is readily discernable, although individual features are not as sharp as in younger moraines higher along Bishop Creek, which are thought to represent Blackwelder's Tioga stage. The moraine consists of relatively fresh, poorly-sorted material that ranges from silt-size particles to boulders.

The next higher glacial deposits, on the northeast side of Coyote Creek consist of 3 formless masses of unsorted bouldery gravel that rest on remnants of a once-continuous flow of olivine basalt. Boulders exposed at the surface appear moderately fresh, but those in a road cut are so deeply weathered that they can be cut through with a shovel as readily as their finer-grained, poorly sorted matrix. The distribution of the basalt remnants suggests that it flowed northerly down a smooth slope, prior to dissection of the slope by Coyote Creek and its tributaries, and it is reasonable to suppose that the glacial debris was carried down the same slope. Whether the glacial deposits are till or outwash was not determined. Coyote Creek has cut a canyon as much as 800 feet deep since the deposition of the basalt and glacial debris; most of this cutting probably was accomplished before the deposition of younger moraines that are considered to be correlative with Blackwelder's Tahoe and Tioga stages along Bishop Creek, to which Coyote Creek is tributary. The deep weathering of the boulders also indicate that the till or outwash is older than the Tahoe moraine in Bishop Creek.

Most of the glacial deposits in the southwest corner of the quadrangle are in a thin sheet with a hummocky surface that is part of a broad field of similar material that covers Coyote Flat to the southwest. The age of this material is difficult to establish. It apparently has been cut through by the upper part of Coyote Creek, which flows in a smoothly rounded glacial canyon. This relationship suggests that two ages of deposits may be present, younger ones along Coyote Creek and older ones at higher altitudes away from the creeks. It is very probable that during the youngest stage (Tioga of Blackwelder) ice occupied upper Coyote Creek, and the younger deposits may be of that age.

Dissected alluvial fan and lake bed deposits

Both the Sierra Nevada and the White Mountains within the boundaries of the quadrangle are flanked by partially-dissected alluvial deposits as well as by more recent fans with depositional upper surfaces (Knopf, 1918, p. 54-57). Some of the dissected deposits have remnants of their original upper surfaces that slope parallel with adjacent, more recent alluvial fans, but stand at higher altitudes. Other dissected deposits are in fault blocks that have been rotated toward Owens Valley. Raised but untilted deposits are present along the north side of the salient of the Sierra Nevada and along the White Mountains south of Poleta Canyon where they extend high along the range front. Some extend upward to altitudes of more than 6,000 feet within the quadrangle, and to higher altitudes farther to the east. Dissected deposits in rotated fault blocks are exposed along the east side of the salient in the Sierra Nevada, and 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.

<u>Dissected fanglomerate along the base of the Sierra Nevada</u>

The dissected deposits along the Sierra Nevada front have been 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 boulders set in a fine sandy and clayey matrix. The material well may represent outwash from glaciers that existed higher in the range during the Pleistocene epoch.

Dissected fanglomerate and lake bed deposits along the base of the White Mountains

In the White Mountains the thickest unbroken sections of dissected material are in Black and in Redding Canyons, where about 600 feet of coarse, crudely-bedded fanglomerate is exposed. 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 into by streams. White pumice that is considered to belong to the lower basal unit of the Bishop tuff locally is intercalated in the fanglomerate at several places. 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 feet 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 material 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 Taylor (written communication) 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, as well as finer-grained pumiceous sandy layers, are interstratified with coarse-grained layers identical with the fanglomerate of Black and Redding Canyons (fig. 2b).

The detailed stratigraphy of the dissected alluvial deposits along the base of the White Mountains was not determined; in particular, neither the lateral nor vertical extent of finer-grained sediments that appear at least in part lacustrine nor their relationship to coarse-grained fanglomerate was established by direct observation. The overall distribution of materials suggests that the finer-grained materials grade into the coarser--fanglomerate was deposited at the mouths of powerful streams that flowed in the main canyons simultaneously with finer-grained materials in interfluves and peripheral to the fans. Conceivably much of the finer-grained material could have been deposited sub-aerially and lacustrine deposits confined to small ephemeral ponds such as exist in low-lying places on the floor of Owens Valley today. Nevertheless, it is equally probable that all of the finer-grained sediments were deposited in a larger lake and deposition in the lower ends of the fans was, in fact, deltaic. Distinction between coarse fan material and similar material deposited in a delta would be difficult. Younger beds in the dissected series, especially those that lie stratigraphically above the pumice layer, may well have been deposited everywhere sub-aerially, though until the extent of the lacustrine deposits has been established, this cannot be ascertained.

Basis of separation of dissected alluvial deposits

The distinction between the dissected series and undissected fan deposits, based as it is on the operation of erosional rather than depositional processes at the surface, can hardly be made in the usual terms of stratigraphic relationships. The same beds presumably are present in deeper parts of the undissected fan deposits as are exposed in the dissected deposits, although the more recent beds of undissected fans are certainly younger than adjacent dissected beds. Further difficulties are that the dissected series cannot be defined in terms of top or bottom, or assigned an age span that is distinct from that of the lithologically similar fan deposits. Stratigraphically the distinction of the two series is almost valueless; nevertheless, structurally it serves a very useful purpose.

The dissection of the deposits was a consequence of structural depression of Owens Valley relative to the White Mountains and the Sierra Nevada. This process went on for a long time and included many movements, most of which probably involved only limited parts of the area of dissected deposits. Two periods of deformation can be identified in the White Mountains, and evidence for earlier periods of deformation may well have been obliterated by erosion.

Age

The strata in the dissected series are almost certainly of Pleistocene and probably late Pleistocene age. The preservation of original features of many glacial deposits in larger canyons of the Sierra Nevada, including easily eroded recessional ridges on the floors, indicates that little erosion of the ranges and consequently little deposition, has taken place since the end of the Pleistocene epoch. For this reason, 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 pumice layer, the basal member of the Bishop tuff, among the strata of the dissected series indicates a late Pleistocene age for beds adjacent to the pumice. The rate of deposition during the Pleistocene epoch must have been rapid, and it seems unlikely that strata older than Pleistocene are exposed, although beds of Late Tertiary age may well be present at depth.

Terrace gravels

Thin veneers of river gravel cap elevated terraces along the south and east sides of the Volcanic Tableland. The gravels consist of well-rounded particles that range in size 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 rock 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 north half of sec. 32, T. 6 S., R. 32 E. Most 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 4400-foot contour crosses both terraces at about equal distances from the 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, and the east-side terraces may also have been deformed.

The gravels in the terraces are much coarser than the gravels in the bed of the Owens River, indicating that the terrace gravels were deposited by faster-flowing and presumably larger streams capable of carrying larger-size 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, it is not impossible that at least the last terrace was cut more recently, during a pluvial period in which many ancient lakes in the Great Basin were filled with fresh water. Putnam (1950) has concluded that the most extensive terrace that borders Mono Lake was cut during this pluvial period.

Undissected alluvial fan deposits

Fans with constructional forms and with no greater dissection of their surfaces than is usual in the ordinary processes of fan formation border the Owens Valley and extend into canyons of the White Mountains and the Sierra Nevada. The material in the fans has been derived both from bedrock formations and from raised and dissected alluvial deposits; the relative proportion of material derived from each source varies from place to place. The fans have distinct forms; they are convex upward in cross section and concave upward in longitudinal section. The crests of the larger fans slope toward the Valley at an average rate of about 300 feet per mile, but the upper ends slope more steeply and the lower ends more gently. Each fan has its own distinct distributary stream pattern.

At the upper ends most of the fans pass without interruption into the beds of their parent streams. At the lower end, on the other hand, the fan of Bishop Creek is the only one within the quadrangle that merges with the alluvium of Owens Valley. Bishop Creek, unlike the parent streams of other fans within the quadrangle, has continued to contribute substantial amounts of material to Owens Valley. A distinct train of material that was carried by Bishop Creek can be traced into the Valley and southward beyond the southern boundary of the Bishop quadrangle for several miles before its identity is lost through admixture with material from other sources. Bishop Creek appears to have recently contributed larger volumes of material to Owens Valley than the Owens River, although the Owens

River may have supplied a larger proportion of material in the past. On the map the lower boundary of the Bishop Creek fan is an arbitrarily chosen line that passes through Bishop nearly parallel with contours. This line approximately separates fine, well-sorted sandy and silty alluvial fill downstream from pebbly and conglomeratic material up the fan.

The other fans in the quadrangle have been overlapped by alluvial fill that was carried into Owens Valley by Bishop Creek and other streams. The boundary between the fans of such streams and the alluvial fill in the Valley can be readily drawn because the slope of the Valley floor is southward, approximately at 90° to the average slope of the fans; thus the contours make a sharp bend where the fans meet the floor of the Valley.

The bulk of the material in fans must have been deposited in Pleistocene time when the rates of erosion and deposition were accelerated because of heavier precipitation. The fan of Bishop Creek almost certainly was formed from glacial outwash, and other fans along the Sierra front may have been built partly from glacial debris that was washed down from higher altitudes. Locally, the upper surfaces of the fans have been veneered by material deposited by floods, in Recent time.

Alluvial fill

Under alluvial fill is included the detrital material that has accumulated in the central part of Owens Valley between the bordering alluvial fans, the alluvial deposits in Fish Slough, and alluvium in small structural basins in the Sierra Nevada. In general, alluvial fill is finer-grained and better sorted than fan material or terrace gravel, but nevertheless exhibits considerable variation in the degree of sorting and in the average size and roundness of particles. The alluvial fill caught behind fault scarps in the Sierra Nevada is less well sorted than most other material called alluvial fill, and the fragments are more angular. Most of it is slope wash that has been transported only short distances.

The fill in Owens Valley was deposited chiefly on a flood plain by streams, and in ponds or lakes, and varies from place to place as is common in such deposits. Well-sorted granitic sands underlie the Bishop tuff in the west end of the cliffs that bound the Volcanic Tableland on the south, whereas, pebble beds, chiefly of metamorphic material, are common beneath the Bishop tuff in the east end of the same cliffs. The pebble beds contain obsidian pellets with shatter patterns caused by impact with other rocks during transportation, and are interstratified with sand and pumiceous beds. Cross-laminated beds are common, indicating deposition by streams. On the highway between Bishop and Laws pebble beds are exposed in a road cut into a low cliff on the west side of the Owens River.

The exposed 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.

Dune sand

Sand dunes that have been stabilized and are overgrown with vegetation occur on the highest terrace in the west-central part of the quadrangle and along the highway between Bishop and Laws. Accumulations of sand are also present in sec. 21, T. 7 S., R. 32 E., where the sand is in pockets on the northeast face of a rounded spur of the Sierra Nevada. Most of the dunes are formless, but several on the highest terrace are crescentic. The convex sides of these dunes face northeasterly, indicating the prevailing wind was from that direction at the time of deposition. The dunes are probably of Recent age, but are no longer growing or moving.

Talus

Talus cones are present locally in both the Sierra Nevada and the White Mountains, though they are not as abundant as in higher glaciated parts of the Sierra Nevada. The size and shape of the fragments in the talus depends largely on the nature of the source rock, and the rate at which it accumulated. Talus from Campito sandstone generally is in relatively large angular fragments, whereas the talus from the slate unit of the Silver Peak group is slabby and breaks down rapidly to clay. Older talus in the White Mountains commonly is cemented by carbonate, whereas no carbonate-cemented talus was observed in the Sierra Nevada.

Most talus cones can be readily distinguished from small alluvial fans by their steeper slopes and by the fact that talus material is coarser downslope whereas fan material is coarser upslope. In a few places this distinction is difficult or impossible because seasonal flow of water down talus-filled channels results in partial reworking of the talus by water into alluvial fans.

GEOLOGIC STRUCTURES

The geologic structures of the Bishop quadrangle are the products of 3 different tectonic episodes, and the diversity of structures is correspondingly great. The earliest structures--folds, faults, and cleavage that were formed by compressional forces prior to the intrusion of the granitic rocks--are found only in the sedimentary rocks of the White Mountains and in the metamorphic remnants in the Sierra Nevada. However, within the Bishop quadrangle the metamorphic remnants are too small to reveal any significant pattern. Most of these structures are believed to have formed in Jurassic or early Cretaceous time before the emplacement of the batholith, but the possibility of earlier diastrophism in the late Paleozoic or Triassic cannot be ruled out. Merriam (oral communication) reports angular unconformities within the Permian strata and between Permian and Triassic strata in the south end of the Inyo Mountains, and Ferguson and Muller (1949, p. 7-8) have mapped Permian strata that lie unconformably above Ordovician strata in the Candalaria Hills about 60 miles north of Bishop. In the White Mountains some intrusive dikes and sills that are considered to be the products of early Mesozoic vulcanism were involved in the diastrophism, but a large majority of them were intruded later.

Next are structures that are the direct results of the emplacement and solidification of the batholith in Cretaceous time.

Deformation of the strata adjacent to the Poleta stock is believed to have been caused by the emplacement of the stock, and it is surmised that the distribution of metamorphic remnants in the salient of the Sierra Nevada is in part the result of the emplacement of the batholith. In adjacent areas the dislocation of the metamorphic rocks by granitic magma is indicated by geometrical relationships.

The most recent structures are Basin and Range faults and warps of Cenozoic age, which account for the larger features as well as many of the smaller features of the landscape. They are chiefly normal faults and warps or broad open folds, but include a few reverse or thrust faults. A regional system of conjugate joints that is conspicuous in the batholithic rocks also is believed to have been formed during this time.

Older structures in the White Mountains

The Lower Cambrian and Upper Precambrian strata of the White Mountains have been strongly folded and faulted along north- to northwest-trending axes. The effects of deformation vary with different stratigraphic units because of differences in their structural competence. The most conspicuous results of deformation are in the slate, followed by the overlying limestone unit of the Silver Peak group, and are least conspicuous in the rocks beneath the Campito sandstone--the dolomite of the Deep Spring formation, the Reed dolomite, and the pre-Reed dolomite strata. Cleavage is present in almost all outcrops of the slate unit except adjacent to the quartz monzonite stock at the head of Poleta Canyon and the diorite stock north of Coldwater Canyon, and is present locally in the Campito sandstone and in the limestone unit of the Silver Peak group, but is lacking in the formations beneath the Campito sandstone. The discontinuous outcrop pattern, variable thickness, and discordant contacts of the slate indicate that during deformation it behaved almost as a paste. Probably every contact of the slate 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 lower part of the limestone unit of the Silver Peak group and the upper part of the Campito sandstone commonly are almost as strongly deformed as the slate itself. This deformation in the strata adjacent to the slate probably resulted from the relative incompetence of the slate.

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 sub-parallel east-trending faults, all of which are downthrown to the north.

The interpretation of the structure shown in the block diagram (plate 4) is far simpler than the true structure, for many minor structures could not be shown at the scale of the diagram. In some complex areas the beds can be seen to be 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 may as well 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. Most of the cleavage mapped is approximately parallel to the axial planes of folds and consequently is useful for determining the trace and dip of axial planes and the plunge of fold axes. The principal geometrical relationships between bedding, cleavage, lineation, the attitude of axial planes, and the plunge of fold axes are shown in figure 6. The dips of axial planes shown on the map and block diagram were derived by calculating the average dips of adjacent cleavage observations. Likewise, the plunges of fold axes were determined by averaging the plunges of adjacent lineations formed by the intersection of bedding and cleavage (plate 5).

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 farther to the east outside of the quadrangle.

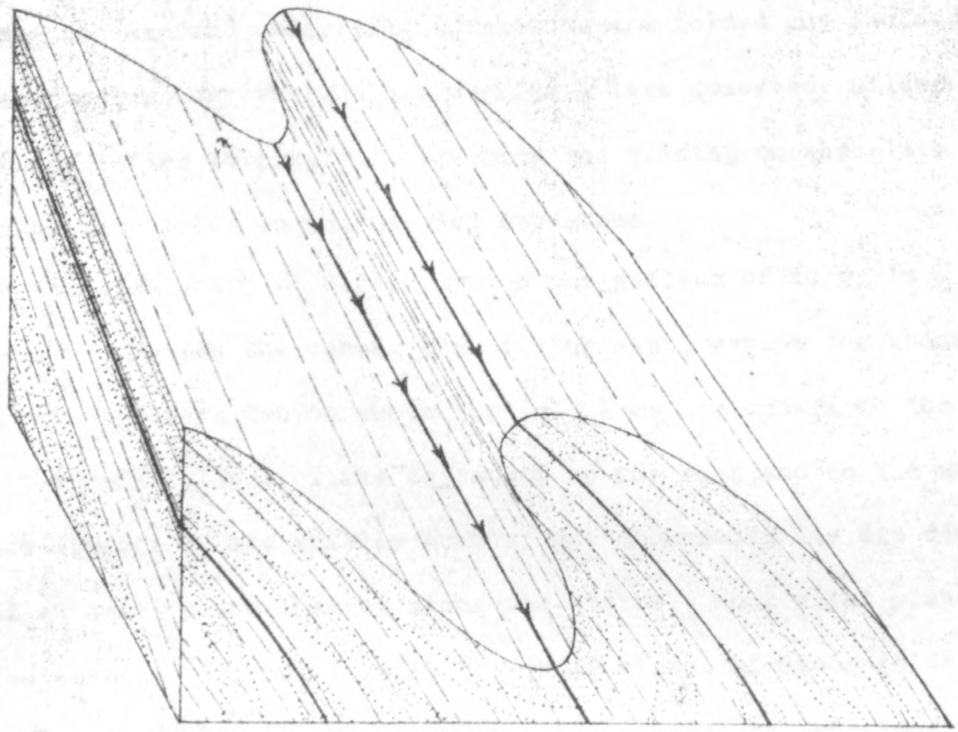


Figure 6. Idealized diagram showing the geometrical relationships of bedding, cleavage parallel to axial planes, and fold axes. The lineations formed by the intersection of beds and cleavage are parallel to fold axes. Top directions can be determined as follows: Beds that dip in the same direction as 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 axial plane is more than 90° from vertical or fold axes are more than 90° from horizontal the rules for tops of beds are reversed.

North of Poleta Canyon, where the width of the mapped span is greater, the folding is complex, and is complicated by faulting. The Campito sandstone and exposed underlying formations are folded and faulted in a competent manner whereas the overlying less competent Silver Peak strata failed complexly, collapsing and sliding on the slate between growing folds in the Campito sandstone.

In the area north of Poleta Canyon the pattern of folds is broadly arcuate with the convex side to the east, except for about a mile north of 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 directions are equally common; not uncommonly the dip direction of an axial plane changes along the strike. Most axial planes dip 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 local plunges of axes range from horizontal to vertical although plunges of more than 45° are uncommon.

Although the details of the folding are complex, the major fold structures are few. They consist of three anticlines marked chiefly by outcrops of Campito sandstone and three synclines marked chiefly by limestone of the Silver Peak group (plate 5). The anticlines are here 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 in a belt of limestone just north of Poleta Canyon, which marks the core of a transverse syncline. This syncline is believed to have been formed at a later period in connection with the emplacement of the stock in Poleta Canyon or with a post-intrusive fault along Poleta Canyon.

North anticline

The North anticline, present on the map for only a short distance, is marked by the outcrop of Campito sandstone in the northeast part of the quadrangle. This structure appears to be fairly simple. Its axial plane strikes N. 40° W. and dips, on the average, about 55° to 65° southwest. The axis is nearly horizontal at the north edge of the mapped area, but farther south plunges to the south 25° to 45°. The slate appears to be abnormally thin in the east limb and abnormally thick in the west limb, but otherwise it is uncomplicated.

East syncline

The East syncline lies along the east edge of the quadrangle from Poleta Canyon northward to the ridge between Gunter Creek and Coldwater Canyon, where it bends northwesterly between the Central and North anticlines. The crest line follows a sinuous course, and the axial plane dips alternately to the east and to the west. At several places the Campito sandstone in the Central anticline has been thrust laterally into the East syncline, and in the vicinity of Silver Canyon the slate between the Campito sandstone and limestone of the Silver Peak group has been concealed so that the Campito sandstone in the Central anticline overrides the limestone in the East syncline. In the span north of Coldwater Canyon the limestone in the core of the syncline locally is intricately interfingered with slate in the map pattern. These interfingers have been interpreted to be drag folds on the flanks of the syncline, but they could have been interpreted as fault slivers.

Central anticline

The Central anticline is marked by the belt of Campito sandstone that extends north from the first canyon north of Poleta Canyon. This structure has a broad undulating crest and many subsidiary superimposed folds which make it difficult to identify the crest of the major fold. In places the crestal area is bounded on the east and west by two sharp knee folds that together make the anticlinal structure (plate 4, sec. 3). The axis of the anticline trends a few degrees west of north for most of its length, but in the vicinity of Gunter Creek it swings more westerly and plunges northwesterly. The structure in the span between Gunter Creek and Coldwater Canyon is the least understood of any in the area studied, but it is possible that the Central anticline continues northwesterly beneath structurally unglued slate and limestone to two small outcrops of Campito sandstone near the mouth of Coldwater Canyon. At the south end the axis steepens and was not surely identified in the limestone that lies along the north side of Poleta Canyon although the pattern of strikes and dips in the limestone suggests southward continuation. Strikes and dips in the Campito sandstone south of the fault along Poleta Canyon suggest a north-plunging anticlinal structure, but the beds also strike parallel with the contact of the granite stock at the head of Poleta Canyon and dip outward from the stock. The anticlinal structure here is believed to be a result of forcible intrusion of the stock.

Central syncline

The Central syncline lies west of the Central anticline from Poleta Canyon to the ridge north of Silver Canyon. Probably the anticline extends farther north, but south of Gunter Canyon several reverse faults cut into it and make it unidentifiable. Some of the fault-bounded slivers of limestone and slate in this area may mark the core of the syncline. Throughout most of its length the syncline is open, and the axial plane is almost vertical. However, near the south end, in the slate and limestone of the Silver Peak group exposed north of Poleta Canyon, it appears to be overturned to the west. As compared with the other folds here described the amplitude of the Central syncline is small, probably ranging between 1,000 and 2,000 feet.

West anticline

The West anticline extends north from Poleta Canyon to the vicinity of Gunter Canyon between the Central syncline on the east and the West syncline on the west. The reverse faults south of Gunter Canyon cut the anticline and make it difficult to identify with certainty. The trace of the axial plane as shown on the maps (plates 1 and 5) is highly problematic in this area. Throughout most of its length the anticline is overturned to the west--the axial plane dips to the east at about 60°. At the south end this fold, like the Central syncline and Central anticline to the east is cut off by the fault along Poleta Canyon.

West syncline

The West syncline is identifiable along the range front from the first canyon north of Poleta Canyon northward to Gunter Creek. Near the south end the trough of the syncline is shown by the tongue of slate near the mouth of the first canyon north of Poleta Canyon, which projects northward into Campito sandstone. Northward to Silver Canyon the syncline can be traced within the Campito sandstone by means of bedding attitudes, and between Silver Canyon and Gunter Creek it is shown by limestone and shale along the range front. The syncline is by no means simple and in the span between Silver Canyon and Gunter Creek both minor anticlinal and synclinal folds can be seen. The axial plane dips steeply east and the plunge of the axis varies widely, plunging steeply to the north and to the south in different places.

At the south end, the West syncline is cut off by a fault that separated slate exposed in the first canyon north of Poleta Canyon from Campito sandstone along the range front. However, a faulted syncline on the southwest side of this fault, marked by thin slivers of slate and limestone within Campito sandstone, may represent an offset segment of the West syncline.

Faults

Most of the pre-batholithic faults strike northwesterly 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 are believed to have resulted from the same compressional movements that produced the folds. Probably the folding was well advanced when the first fault movements began. Faulting began only after the rocks were so tightly folded that the internal resistance to folding was so high that faulting required less energy than further folding. 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 later during a period of relaxation that followed the period of compression.

In mapping it was difficult to distinguish some pre-batholithic faults from some Basin and Range faults inasmuch as both 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	Pre-batholithic 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 are the result of erosion.

Longitudinal faults south of Poleta Canyon

South of Poleta Canyon the only faults that can be ascribed with certainty to pre-batholithic diastrophism are two reverse faults that strike about N. 30° W. across the first prominent ridge south of Redding Canyon (plate 4, 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 steepen in depth; consequently in cross section they appear convex upward and to the west. Because the western fault dips generally more steeply than the east fault the two faults are presumed to join at depth. The exposed segments of both faults are chiefly within the Campito sandstone, but along the east fault Campito sandstone locally has been carried eastward over the slate unit of the Silver Peak group, and the beds in the Deep Spring formation west of the western fault strike into inliers of Campito sandstone in the alluvium to the north, indicating that along this fault the Deep Spring formation locally must have been carried east over the Campito sandstone. The apparent stratigraphic thickness of the Campito sandstone has been reduced by about 1,000 feet on the two faults, and assuming approximately equal movement 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 the Campito sandstone and the slate and limestone units of the Silver Peak group, or between the slate and limestone units. 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 along the strike. In a general way the fault planes 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 slate unit plus unknown thicknesses of the underlying or overlying formations have been cut out.

The southernmost group of 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 slate and limestone 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 limestone unit are cut out against the over-riding slate. The steepening and overturning of the beds are pictured as early movements that culminated in the shearing.

Farther east slate overrides the east side of the limestone on a flat fault that can be traced into an unfaulted contact. The flat fault is cut by steep faults that are apparently downthrown to the west but which may have moved laterally like the fault along Poleta Canyon which they parallel.

From the south wall of Silver Canyon north to Coldwater Canyon is a belt of faults that separate slate, limestone, and the Campito sandstone. In the lower walls of Silver Canyon limestone on the east is in contact with Campito sandstone on the west along a fault that strikes north and dips west (plate 4, sec. 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 dips gently west (plate 4, sec. 2). Although the fault plane may be twisted along the strike, these relationships suggest that the fault plane is convex upward and toward the east, and bends in dip through more than 100 degrees (plate 4, secs. 5 and 6). Locally in this span, slate is present along the contact, but in most places limestone is in direct contact with the Campito sandstone.

Farther north, in the canyons of Gunter Creek, slate lies between the Campito sandstone and the limestone unit of the Silver Peak group, but in most places the bedding planes within the slate dip steeply and are discordant with the gently-dipping upper and lower contacts, indicating structural ungluing of the shale from the overlying and underlying units. Two parallel branches of the fault system that extend south from the south fork of Gunter Creek between slate on the east and Campito sandstone on the west steepen northward at topographically lower altitudes, giving the impression of either twisted fault planes or of planes that are convex to the east.

Just south of Coldwater Canyon, the limestone along the west side of the quadrangle is structurally discordant with both underlying and overlying slate, and hence both contacts were mapped as faults. The fault that follows the lower contact of the limestone steepens northward and is nearly vertical where it crosses Coldwater Canyon. North of Coldwater Canyon several steep faults may mark the continuation of this belt, but it is equally likely that these faults are related to the emplacement of the diorite stock in this area.

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 Campito sandstone along the front of the range is in fault contact on the east side with Silver Peak limestone and slate. The fault trends N. 30° W. and in the north part is vertical and in the south part dips steeply east. At the north end, in the first canyon north of Poleta Canyon, the faulted contact grades into an apparently unfaulted contact between Campito sandstone and the slate units of the Silver Peak group. The faulted segment of contact is clean and sharp, and cuts across folds in the Campito sandstone at a high angle and truncates bedding in the Silver Peak strata at a slightly smaller angle.

Within the Campito sandstone on the west side of the fault is a thin sliver of Silver Peak slate and limestone that is bounded on both sides by faults. This sliver trends a few degrees east of north, approximately parallel with fold axes in the Campito sandstone and appears to mark the approximate position of the southward continuation of the trace of the axial plane of the West syncline. The faults that bound the sliver of Silver Peak strata are poorly defined but appear to terminate against or bend into the fault that cuts off the Campito sandstone on the east.

Faults in the Central syncline and West anticline between Silver Canyon and Gunter Creek

The Campito sandstone exposed between Silver Canyon and Gunter Creek is cut by several east-dipping reverse faults. These faults are along northern projections of the Central syncline and West anticline, which they have almost destroyed. A thin belt of slate is believed to mark the trough of the Central syncline. Most of the faults dip steeply but some dip as low as 35°. The displacements were not determined, but the movement on most of the faults appears to have been small because the faults are either entirely within the Campito sandstone or between Campito sandstone and Silver Peak strata. They are approximately parallel both in strike and in dip with fold axes. Parallel faults are also exposed in the ridge of Campito sandstone 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 Campito sandstone is exposed in the walls of the faults.

Along the range front at Coldwater Canyon and also about one mile farther to the south are two slivers of Campito sandstone that are surrounded by slate and limestone. Both slivers may mark the cores of anticlines, but the northern sliver is bounded on both sides by steep faults and at the south end is in contact with limestone of the Silver Peak group.

Cleavage

Slaty cleavage is present in many outcrops of Campito sandstone and Silver Peak limestone, but is prevalent only in the Silver Peak slate (figs. 7a and 7b). Cleavage is present throughout the slate except adjacent to the Poleta quartz monzonite stock and the diorite stock north of Coldwater Canyon. The thermal metamorphism that accompanied the emplacement of the stocks is believed to have destroyed any cleavage that existed in the adjacent rocks. In the Silver Peak slate the cleavage is pervasive and the rock splits readily along smooth cleavage surfaces that generally are not parallel to the bedding. In the Campito sandstone the cleavage is approximately parallel with cleavage in adjacent masses of slate but the cleavage is not pervasive and surfaces generally are wavy and branching rather than smooth and parallel. 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 Silver Peak limestone, impure argillaceous limestone cleaves readily along smooth surfaces, but in cleaner carbonate rocks the cleavage is shown chiefly on weathered surfaces by thin parallel ridges and trough-like depressions. This sculpturing on weathered surfaces along cleavage planes is so conspicuous in places as to cause some confusion in identifying bedding. Such rocks tend to break along rough surfaces that are only sub-parallel 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, cleavage parallel with faults, which in the slate is pervasive adjacent to the fault and increasingly more widely spaced farther away, was also observed in a few places. Cleavage of this

Figure 7a. Slate unit of Silver Peak group showing divergent attitudes of beds and cleavage. Strike of cleavage is from left to right, and of bedding from upper left to lower right.

Figure 7b. Minor folds in slate shown by calcareous sandy layer.

Figure 7a.

Figure 7b.

kind was observed to transgress and in places obliterate cleavage parallel to axial planes and is clearly younger than the latter. This second cleavage is identical in appearance to the first kind 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 (plate 5) 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 of folds permits the existence of two cleavages 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.

Structures related to intrusion

Structures that resulted from the emplacement of the granitic rocks in Cretaceous(?) time are difficult to identify in the Bishop quadrangle, although they are evident in adjoining areas. Two kinds of structures are included, those within intrusive rocks, and those in the adjacent wall rock. Planar foliation and in places lineation are present in some granitic rocks, especially granodiorite, as a result of magmatic movements during their emplacement and crystallization.

The strata marginal to the Poleta stock on the north and northwest sides are bent upward into a northwesterly plunging anticline. The beds strike parallel to the contact of the stock and dip away from it. This geometrical arrangement is believed to be the result of forcible emplacement of the stock, although it is possible that it is the result of earlier folding. The distribution of remnants of metamorphic rock in the Sierra Nevada was undoubtedly affected by the emplacement of the granitic rocks, but evidence of the amount is lacking.

North of Poleta Canyon the limestone of the Silver Peak group is flanked on both the north and south sides by slate and must therefore have a synclinal structure; the rock distribution seems to suggest that the pre-intrusive East and West synclines join here to form one continuous syncline. Nevertheless, the attitudes of cleavages and bedding indicate that the north-trending folds bend easterly but continue across the limestone in the synclinal structure. Apparently the transverse syncline originated after the north-trending folds. A plausible explanation is that the intrusion of the Poleta stock, which is now offset to the east of its original position on the Poleta fault, forced up the strata on the south side of the limestone and brought the slate there to its present level. An alternate explanation is that the strata were bent upward during a later period in conjuction with movement on the fault along Poleta Canyon.

Block faults and warps of Cenozoic age

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 formed during the Cenozoic era. These structures account for most of the larger features and many of the smaller features of the landscape. 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.

Most of the structures mapped were recognized by their effect on the topography; and only rarely were internal layers of use in identifying structures. The escarpments along faults are, almost without exception, true fault scarps, though most of them have been modified varying amounts by erosion. The traces of faults across irregular topography and rare exposures of fault planes indicate that almost all of 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 forms that are very different 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 structures are warps, and some conspicuous faults are merely secondary features on warps or folds.

In the evolution of the Sierra Nevada and White Mountain 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 faults that are downfaulted on the valley side, here called valley-down (synthetic) faults, are faults with the mountain side downfaulted, or mountain-down (antithetic) 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 or antithetic faults in a similar gross setting (the Rhine graben) 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 Owens 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 that 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. As they were formed at the surface, the warping was 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 this arrangement are marked by escarpments that face toward the mountains. The fact that the faults are normal indicates tension and suggests that the valley subsided under the influence of gravity. If it had been folded down as a result of horizontal compression normal faults would be expected to be less abundant, and reverse faults might be present.

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 through the aggregate tilting of blocks.

Narrow graben a few hundred feet or less in width and as much as a mile in length are present at several places along the range fronts, both in alluvial fans and in older rock. Along some graben the fault with 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, more mountain-down than valley-down faults are likely to be identifiable by their scarps, thus giving an erroneous impression of their relative abundance.

The Sierra Nevada escarpment

The salient of the Sierra Nevada that extends into the Bishop quadrangle has little in common either physiographically or structurally with the precipitous fault scarp that ordinarily is envisaged when speaking of the Sierra Nevada escarpment. Steep fault scarps occur to the south and to the northwest, but in the salient broad interfluvial areas between steep-walled canyons slope gently north and northeasterly into Owens Valley at average slope angles of about 10 degrees. These interfluvial areas are part of an old erosional surface whose present configuration is the result of structural warping. The salient is in fact the apex of a 2-sided structure called the Coyote warp that extends into adjacent quadrangles. This warp stands out in front of projected trends of faulted segments of the Sierra Nevada escarpment to the south and to the northwest (fig. 8).

The ancient surface that records the warp is deeply weathered and in places has a cover of residual soil. Where it is best preserved it is gently rolling with relief normal to the average slope angle of a few hundred feet, part of which is assignable to recent dissection. With dissection the amount of relief increases to the point where the ol surface is no longer identifiable. Local deep dissection of the old surface by streams such as Rawson Creek

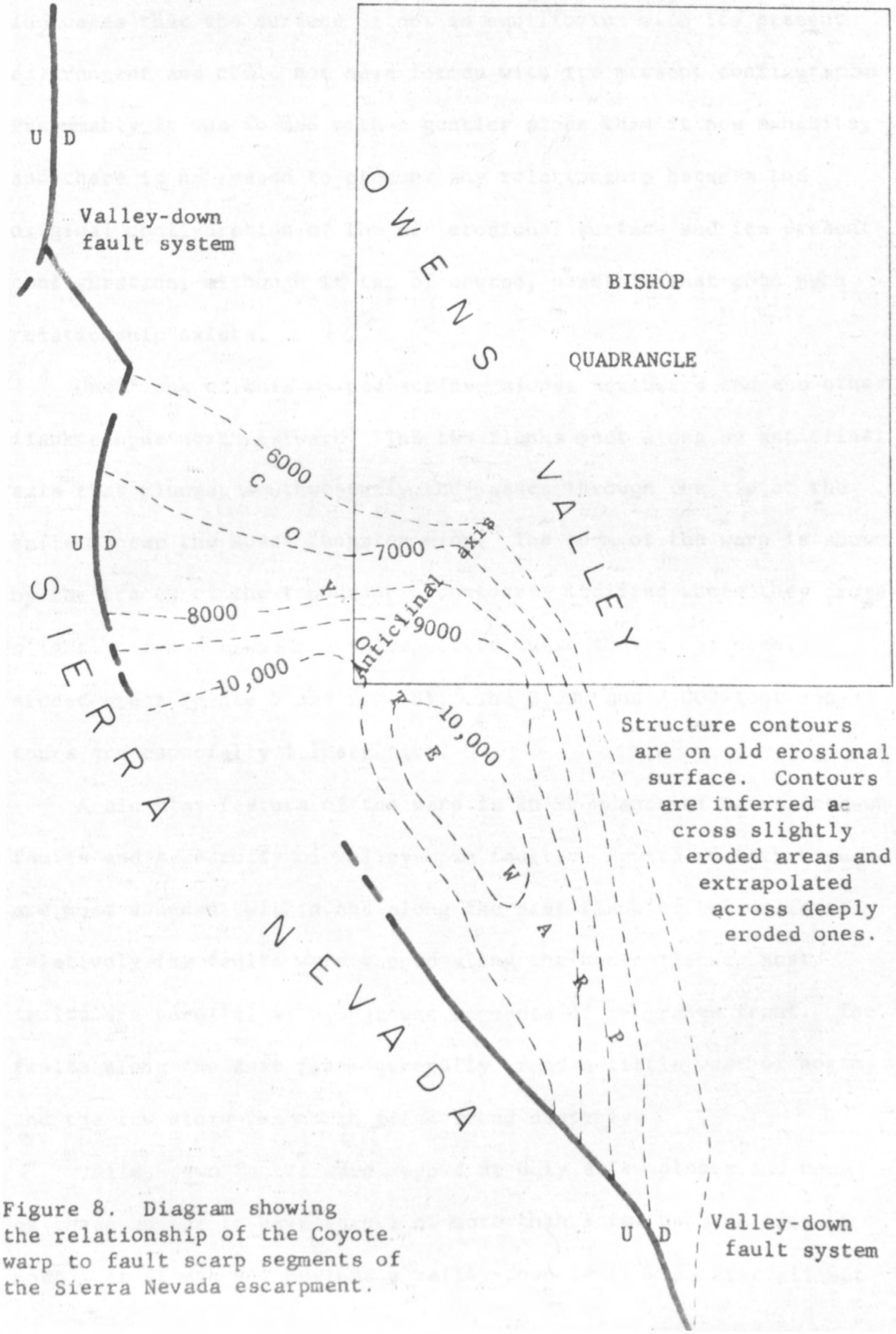


Figure 8. Diagram showing the relationship of the Coyote warp to fault scarp segments of the Sierra Nevada escarpment.

indicates that the surface is not in equilibrium with its present environment and could not have formed with its present configuration. Presumably it was formed with a gentler slope than it now exhibits, and there is no reason to presume any relationship between the original configuration of the old erosional surface and its present configuration, although it is, of course, possible that some such relationship exists.

One flank of this warped surface slopes northward and the other flank slopes northeastward. The two flanks meet along an anticlinal axis that plunges southwesterly and passes through the tip of the salient near the Rossi tungsten mine. The form of the warp is shown by the traces of the topographic contours, modified where they cross slightly eroded areas and extrapolated where they cross deeply eroded areas (plate 5 and fig. 8). The 6,000 and 7,000-foot contours are especially illustrative.

A singular feature of the warp is an abundance of mountain-down faults and a scarcity of valley-down faults. Faults of both kinds are most abundant within and along the east flank of the salient; relatively few faults were mapped along the north flank. Most faults are parallel with adjacent segments of the range front. The faults along the east flank generally trend a little west of north, and the few along the north flank trend easterly.

Valley-down faults were mapped at only a few places and none of these appear to have throws of more than a few hundred feet at most. At Keough Hot Springs a valley-down fault separates all but the very youngest fan deposits along the range front from the granitic rocks of the range, and also provides the conduit for the hot springs. The springs break out at the lowest point of intersection of the fault plane with the ground surface. This fault appears to terminate just north of Keough Hot Springs, but a second fault, offset about a quarter of a mile to the west, continues northward for about a mile. Throughout most of its exposed length the northern offset segment bounds the west side of a narrow graben and is paralleled by a mountain-down fault a few hundred feet to the east on the other side of the graben. The throw on the valley-down fault is difficult to estimate because at the south end where the land form suggests the largest throw the granite on the upthrown side of the fault appears to have been tilted toward Owens Valley an unknown amount on mountain-down faults that lie to the west higher along the range front.

The only other valley-down fault that was mapped in the salient is an east-trending fault in the vicinity of the Bishop Antimony Mine, which is paired with a sub-parallel mountain-down fault that lies just to the north. However, although no fault was observed along the east-trending segment of the range front west of the Bishop Antimony Mine or along the east-trending segment in the vicinity of the Yaney Mine, locally steep escarpments at these places suggest bounding valley-down faults.

The most conspicuous effect within granitic rock of mountain-down faulting is a prominent bench at half height in the range front south of Rawson Creek, which superficially resembles a stepped block bounded on the west by valley-down faults. This impression, however, is erroneous. Careful examination 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. Instead, the evidence points to a formation of the bench as a result of 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 (fig. 9a). Smaller parallel benches of similar origin 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, extending about N. 20° W. for about a mile from the triangulation station at 5,398 feet just west of Keough Hot Springs.

On the west side the graben along the range front north of Keough Hot Springs is bounded only by granitic rocks, but on the east side it is bounded by a small area of granitic rock that is flanked on the north, east and south sides by older dissected alluvial deposits. The alluvial deposits have been rotated toward the east several degrees and a steep contact between these deposits and the underlying granitic rock suggests the possibility that the granite has been rotated even more as a result of earlier movements. The dissected alluvial deposits at the south end slope to the east at about 7°, whereas adjacent unfaulted alluvial fans slope at only 3°. In the north end of the graben an orchard flourishes, subsisting on

Figure 9a. Bench parallel to range front south of Rawson Canyon formed by mountain-down faulting.

Figure 9b. Minor thrust fault in Bishop tuff. Basal pumice layer is thrust above stratigraphically overlying unconsolidated tuff.

Figure 9a.

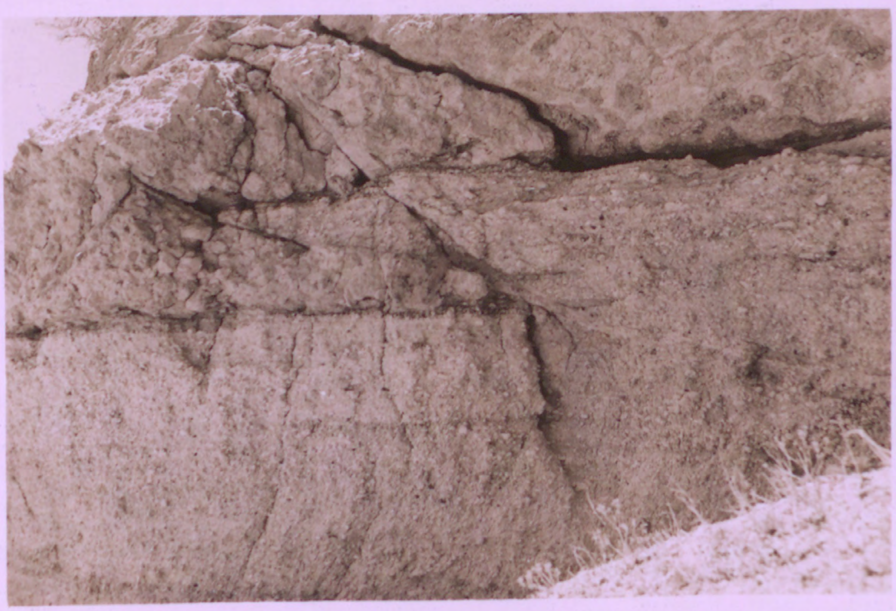


Figure 9b.

fine alluvial fill and water that were trapped behind the mountain-down fault on the east side of the graben. This situation is repeated in many places elsewhere along both sides of Owens Valley in connection with mountain-down faults that cut alluvium.

About one mile farther north a mountain-down fault bounds a block of bouldery fanglomerate that is tilted strongly toward the valley (fig. 10). This fault has diverted Rawson Creek, which flows south along the base of the scarp and around the end of the tilted block. Several wind gaps in the scarp together with fine alluvial fill that has collected west of the scarp as a result of the faulting indicate that the course of the stream has changed repeatedly, presumably as a result of repeated small movements along the fault.

Other parallel mountain-down faults occur along the east base of the salient to the east and to the north and others lie to the west within the range. A north-trending fault that extends southerly from the center of sec. 31, T. 7 S., R. 32 E. into the center of sec. 6, T. 6 S., R. 32 E., has relationships similar to the fault just described, except that its throw is less and the block of alluvium east of the fault has not been tilted so much. A marshy area of fine alluvium lies just to the west of this fault. This fault may be the southern extension of one that bounds the west side of the granite block at the point of the salient, which contains the Rossi tungsten mine. Other faults that bound tilted alluvial deposits lie east and northeast of this block. In this general area a multitude of minor faults and fractures are evident,

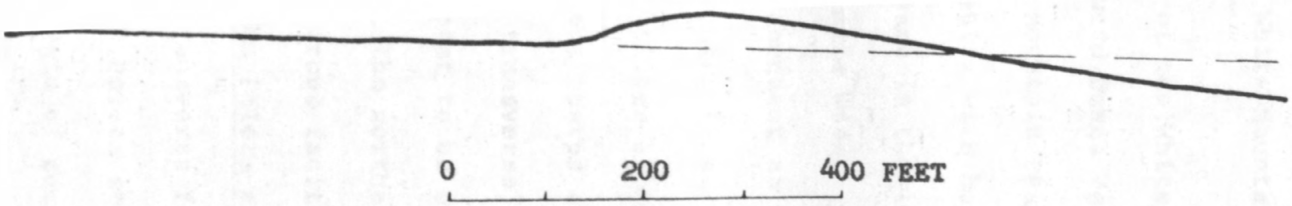


Figure 10. Profile across mountain-down fault scarp in alluvial fan of Rawson Creek.

both in the field and on aerial photographs (plate 5). Some of the minor faults are in such poorly consolidated material that they must be the products of very recent movements.

The White Mountain escarpment

The western side of the White Mountains within the mapped area also slopes gently toward Owens Valley, although farther north, at the latitude of White Mountain Peak, the range front is very precipitous. Numerous faults with 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 chiefly because they mark the boundaries of tilted blocks. Older than the range front faults and lacking recent scarps are three faults between Poleta and Black Canyons that are transverse to the range front. The middle and southern faults appear to be older than the range front faults, which cut them off, but the northern fault may bend near the range front and join a range front fault beneath alluvium.

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, called the Poleta fault, strikes northeasterly. All three faults are descriptively normal faults that dip toward 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° W., parallel with the fault along Poleta Canyon, and are downthrown on the northwest side.

These faults account for the exposure south of Poleta Canyon of the formations beneath the Campito sandstone. Along the southern two faults Reed dolomite on the south has been brought into contact with Campito sandstone on the north, indicating a stratigraphic displacement on each fault in excess of the thickness of the dolomite of the Deep Spring formation (1,500 feet). Along the Poleta fault Campito sandstone on the south is in contact with slate and limestone of the Silver Peak group on the north--relationships 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.

Left lateral movement on the order of about one mile is suggested on the Poleta fault by the fact that slate with 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 one mile west of the stock. These relationships also indicate that the principal movement on the fault was after the emplacement of the Poleta stock. Lateral movement along the southern 2 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. C. A. Nelson (oral communication) reports left lateral movement along several transverse faults in the Waucoba Mountain quadrangle to the southeast, which are nearly parallel with these faults. He states that one of these faults locally forms part of the escarpment along the range front.

Structures parallel with the range front

Near the south boundary of the 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. Just south of Coldwater Creek a valley-down fault passes southward into a mountain-down fault that bounds a tilted block on the mountain side.

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. 34 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 to the west at about 15° whereas those east of the fault dip to the west at only about 6°. The 15° dip is not exceptionally steep--beds exposed in a pumice pit in the SE¼ sec. 14, T. 6 S., R. 34 E. dip 45° to 55° toward the valley. The 15° angle of 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 toward the mountains and that at the south end faces the valley. The height of the scarps diminish toward each other and almost meet in a small area of no displacement. The block west of the fault was dropped at the south end on the valley-down segment of the fault and at the north end by a tilted block west of the mountain-down scarp.

Repeated movements are also shown in the part of the scarp that faces the mountains. Two periods of dislocation can be clearly identified in the scarp both to the north and to the south of the stream that flows from Coldwater Canyon (fig. 11). 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 that continues westward beneath the valley floor. The aggregate deformation of the bed rock surface can be presumed to be far greater than the deformation represented in the exposed rocks, many of which are fan or lake bed deposits and record only very recent movements. No single fault was recognized through geologic mapping which alone would account for a major part of the structural relief.

However, gravity studies made in northern Owens Valley by L. C. Pakiser and M. F. Kane of the 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 (written communication). The westernmost mapped valley-down fault between Silver and Black Canyons may mark the projection of this escarpment to the

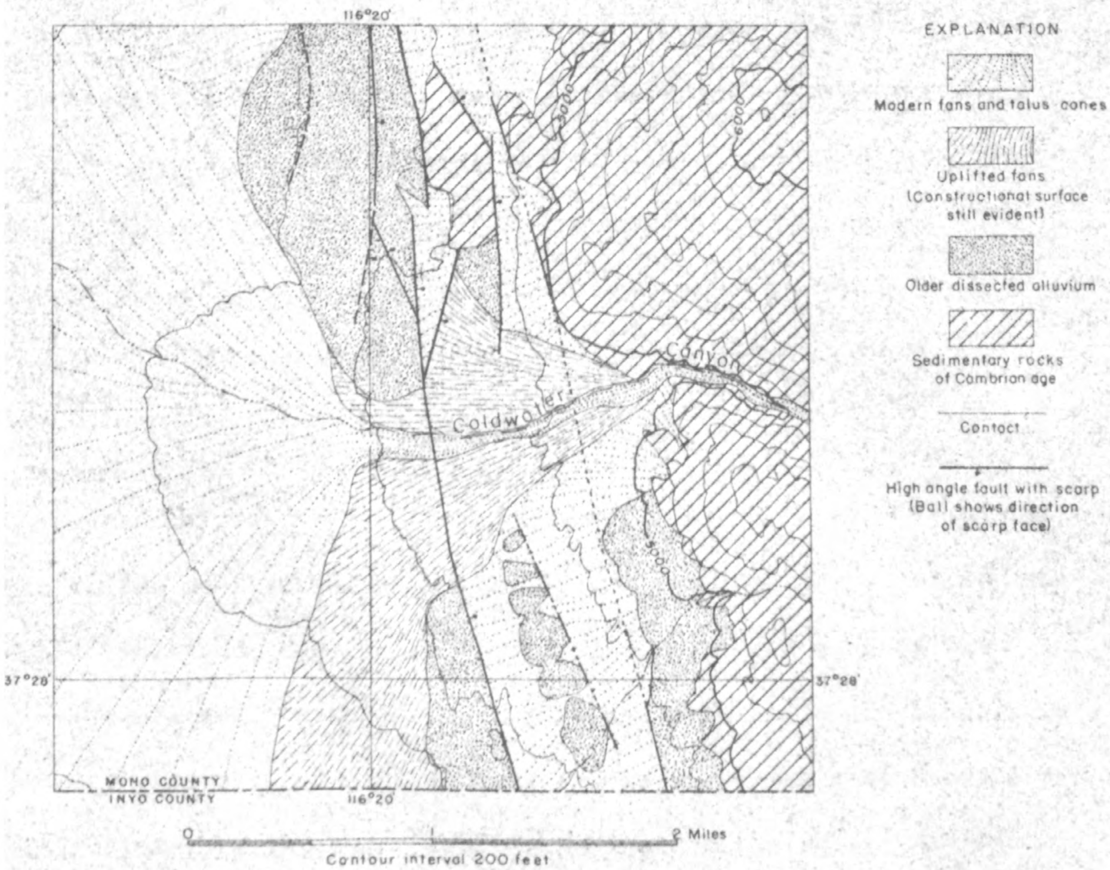


Figure 11. Map of area at the mouth of Coldwater Canyon. The uplifted fans are younger than the dissected alluvium. Both uplift and dissection result from fault movements. Two periods of faulting are inferred from the relationships.

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 part of that in the exposed escarpment is accounted for by movement on the master fault. Nevertheless, the mapped structures must account for a substantial part of the structural relief within the exposed part of the range front. The accompanying geologic cross sections (plate 3, secs. A-A' and B-B') were constructed in accordance with this interpretation.

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, 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.

Structures in the Volcanic Tableland

The surface of the Volcanic Tableland provides a reference plane for recording faults and warps that is both sensitive to strain and resistant to destruction by erosion. Within the Bishop quadrangle the surface approximately parallels internal layering in the Bishop tuff and is chiefly constructional, although locally less resistant layers appear to have been stripped away.

The general slope of the surface is to the southwest, but this slope is broken by many minor features. The most conspicuous features are steep, fresh appearing, north-trending fault scarps (fig. 12a). The largest scarp, along the east side of Fish Slough, is more than 5 miles long though it consists of two segments that are offset from each other near the north edge of the quadrangle. A few other faults are as much as 3 miles long but most of them 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 the height of most scarps is less than 50 feet. Most scarps are highest at the center and are progressively lower toward the ends. Alluviation has taken place locally in basin-like depressions at the bases of some scarps, but except in Fish Slough the amount of alluvium deposited generally is negligible. Thus the height of most scarps is a measure of the throw on the fault since the extrusion of the Bishop tuff. Scarps that face to the east and to the west are equally abundant, and are of the same average height. 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 with larger throws appear to dip vertically or nearly vertically, whereas the planes of some minor faults dip at angles as low as 55°. Many of the fault planes can be observed to pass through the Bishop tuff and into the underlying alluvial deposits without recognizable increase or decrease in the throw.

Figure 12a. Aerial view of south part of Volcanic Tableland showing systems of en echelon faults. Note segments of dislocated stream channel in lower left part of photograph. Fish Slough extends across lower right-hand corner of photo.

Figure 12b. View looking northeast of tilted fault blocks along Fish Slough. Surface is the same on ridge in middle distance, on tilted blocks in front of ridge, and in foreground.

Figure 12a.

Figure 12b.

The two 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 almost half a mile to the west. Although the scarp attains a maximum height of about 300 feet, the downthrown side is covered with alluvial fill and the maximum throw on the fault may be 400 feet or more. A gravity high is reported to exist in the part of the Volcanic Tableland east of Fish Slough (Pakiser, written communication) which, if caused by upfaulting of the basement suggests a throw of several thousand feet.

In the lower end of Fish Slough, in the W½ of sec. 7 and NW¼ sec. 18, T. 6 S., R. 33 E., three tilted blocks lie along the fault (fig. 12b). All of these blocks have been rotated so 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. The lower ends of all the blocks are covered by alluvial fill, so that the relationship 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 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 Volcanic Tableland. The fault surfaces reveal that most of the faults there are vertical or high-angle normal faults, but that a few are high-angle reverse faults, and that one is a low-angle thrust fault. Slickensides identified on about 25 percent of the fault planes 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 with no obvious difference 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° north.

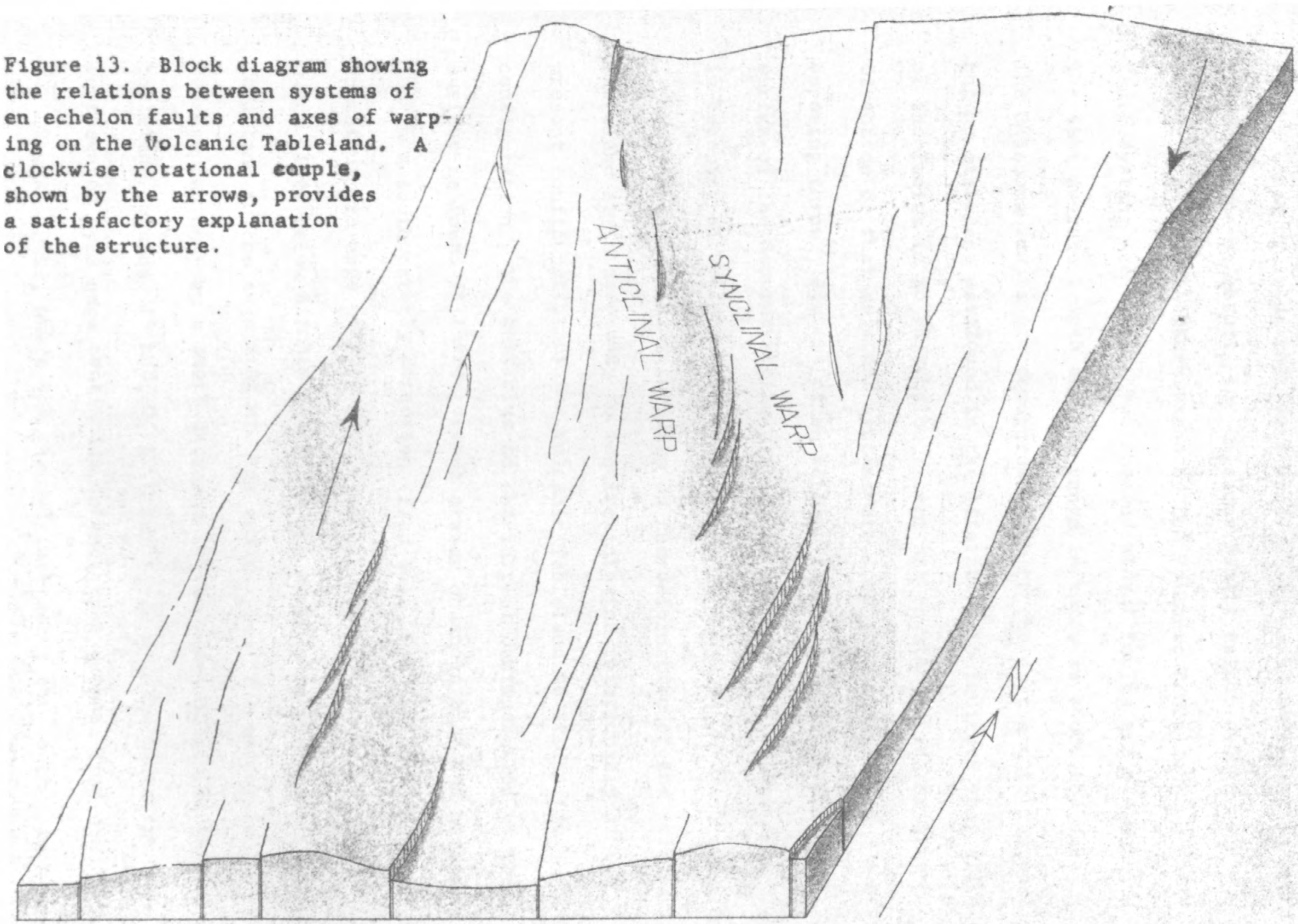
Most of the faults west of Fish Slough are arranged in northwest-trending en echelon systems. All of the faults in a system are downthrown in the same direction, and systems in which the faults are downthrown to the west alternate with systems in which the faults are downthrown to the east. Several of these systems can be readily identified on the maps (plates 1 and 5) but even where these 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 with opposing directions of throw are paired to bound graben or horsts, but most blocks are tilted and up-faulted on only one side. In a few places the throw on a fault diminishes 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.

It is evident that two systems of en echelon faults whose scarps face inward define a structural low and that two series 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 folds and the structural lows to mark the axes of synclinal folds (plate 4). These relationships lead to the tentative conclusion that the broad inconspicuous anticlinal and synclinal folds are really the primary structures here and that the fault systems are merely secondary features of contemporaneous origin that help define the flanks of the folds.

To better understand the strain pattern, the faults in part of the area were traced from a map 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 right (east) side of the sheet was moved south relative to the left (west) side, both folds and faults formed in a pattern that is in agreement with the field observations, indicating that a rotational couple was involved (fig. 13). The principal component of movement on the faults was dip-slip even though they are 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 structures 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.

A common explanation of systems of en echelon faults is a rotational couple, but so far as the writer has been able to determine 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 (1929) suggested regional torsion, augmented by slight uplift. Fath (1920) and Folley (1926) postulated horizontal shifts in the basement beneath each system. Link (1929) suggested tension

Figure 13. 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.



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 of the faults in an en echelon system are downthrown in the same direction or that adjacent systems are composed of faults of opposing throw. Systems of en echelon faults in Montana on the strike of the Osburne strike-slip fault include faults that face in opposing directions.

Before summarizing the data on the deformation of the Volcanic Tableland it is necessary to inquire into the relationship 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 southwesterly direction (fig. 14). 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 in the adjoining Mt. Tom quadrangle. 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 has been broken into segments by faults and some segments are lifted high above others (fig. 12a). It is most likely that these channels were formed by consequent streams that flowed down a southeasterly sloping surface; and in the absence of evidence to the contrary it is reasonable to assume that this slope was the initial one. This conclusion is in agreement with considerations as to the probable source of the Bishop tuff, and its probable direction of flow.

Both the northwest-trending folds and associated systems of en echelon faults and the tilted blocks along the Fish Slough fault 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. It should be emphasized that 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 at 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). A fault in the NW¼ sec. 9 bends easterly at its north end toward the upthrown block, presumably as a south-dipping thrust similar to the one exposed in the pumice pits along the east edge of the

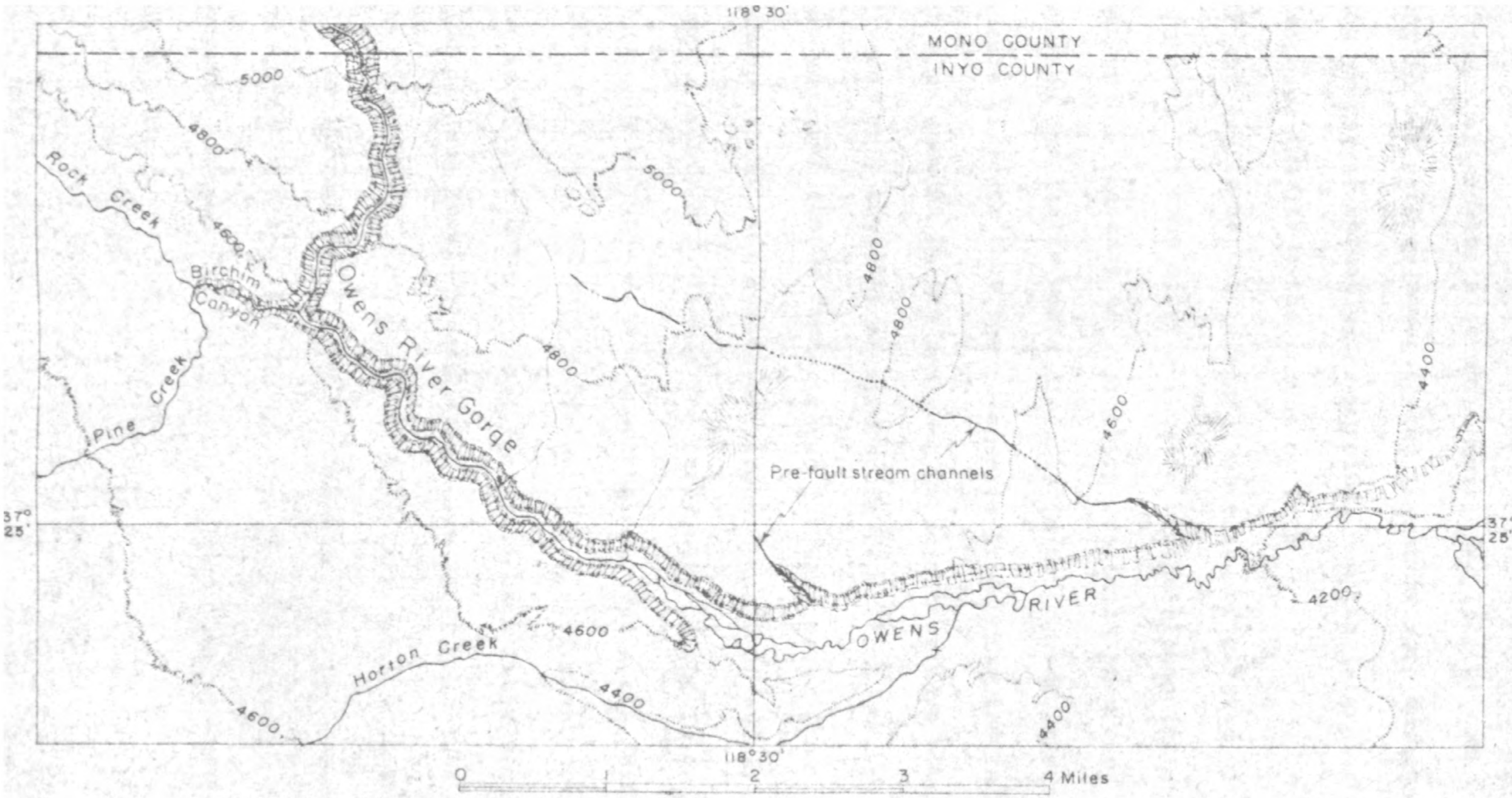


Figure 14. Map of the southwest part of the Volcanic Tableland. Shows the location of three acient stream channels and their relationship to modern topography. One channel, Birchim Canyon and the lower part of the Owens River gorge, is occupied by the Owens River. The other two have been disrupted by faulting and warping and are dry.

Volcanic Tableland. The north-trending segments of faults that bend into thrusts or terminate at warps must have lateral components of movement in the part adjacent to the thrust or warp commensurate with the amount of thrusting or warping, and the lateral component must decrease away from the thrust or warp.

If the faults in the Volcanic Tableland originated within a brief interval of time after extrusion of the Bishop tuff, it might be inferred that because of increasing confining pressures the displacement on the faults would diminish in depth, but several lines of evidence indicate that the reverse is true. The faults have greater displacement in the basement because they have moved repeatedly and only the last movements are recorded in the Bishop tuff. Repeated movement can be demonstrated for many range front faults, and inasmuch as all the faults are probably part of one system, repeated movements seem likely for the faults in the Volcanic Tableland. In the Casa Diablo quadrangle to the north, Rinehart and Ross (1957) have demonstrated that the displacement in the basement is greater than in the Bishop tuff on a fault along the west side of Casa Diablo Mountain and on a fault that is exposed in Owens River gorge. Furthermore, a gravity anomaly coincident with the Fish Slough fault (Pakiser and Kane, written communication) is interpreted by the writer to signify that the displacement of the concealed bedrock surface is much greater than of the Bishop tuff.

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 peripheral to the salient in the Sierra Nevada front and possibly also at the time of the faulting along the western base of the White Mountains that resulted in the dislocation of older fan deposits. 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 stand 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. Of the two groups of terraces those 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 west of the quadrangle. It is evident from inspection of the map (plate 1) 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 has cut into the northern margin of its predecessor, local benches and accumulations of gravels that hang on the cliff on the south side of the Volcanic Tableland indicate that some of the terraces, at least, extended north to the cliff.

These relationships suggest that at the inception of each period of terrace development the Owens River flowed easterly along the south edge of the Volcanic Tableland and that gradually it broadened its valley by cutting laterally toward the southeast. For this to have happened, the surface in the vicinity would have had to be tilted periodically to the east; each period of tilting would have resulted in restoration of the river to an easterly course and in the inception of a new period of terrace formation. After a period of structural stability, during which alluviation worked upstream, the river would have begun to cut laterally to the southeast until a new cycle was initiated.

A test for this hypothesis is found in the slope angle and amount of deformation of the terraces. If the hypothesis is correct, each higher terrace should slope eastward more steeply than its next higher predecessor; and this appears to be so. 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 all of the terraces 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 slope 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. West of the longitude of Bishop the Bishop Creek fan lies somewhat lower than either the terraces or the dissected fans and is bounded on the northwest by the southeasterly-facing escarpment on the southeast side of the terraces and on the south by an intricately indented escarpment on the north side of the fans. A reasonable explanation of the existing relationships is that 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 that not only were the terraces 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 relationship 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 side in the Mt. Tom quadrangle and in the Bishop quadrangle have been tilted toward the east. The east end of the middle terrace is broken by several faults with scarps of as much as 20 feet that face both to the east and to the 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. On examination, these lineaments appear to be low scarps generally 2 to 3 feet or less 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 the most recent movements on the same system of faults that cut the Volcanic Tableland.

The terraces along the east side of the Volcanic Tableland were cut by a strong tributary ancestral to the Owens River, which may have 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. Each step represents a terrace indicating that the escarpment is erosional. Data from well logs indicate that the escarpment coincides closely with the axis of a warp along which the Bishop tuff is bent down to the east. This warp in turn may coincide with a fault in the basement--it coincides with a strong gravity anomaly (Pakiser and Kane, written communication). 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.

<u>Subsurface structure of Owens Valley</u>

The ancient warped surface preserved in interfluves in the salient of the Sierra Nevada is believed to continue beneath the alluvial fill of Owens Valley and to dominate the bedrock configuration. If the eastern flank of the anticlinal warp preserved in the salient is projected eastward at the 10° average slope angle that exists in the salient, the bedrock surface will lie about 5,000 feet beneath the valley floor at the western base of the White Mountains. The north flank also is believed to continue northward beneath valley fill and the Volcanic Tableland, but is assumed to bend upward again along an easterly trending axis because granitic and metamorphic rocks crop out not far north of the quadrangle boundary and because evidence for large easterly trending faults in the valley block is lacking.

The postulated configuration of the bedrock has recently been substantiated by gravity studies made by L. C. Pakiser and M. F. Kane of the U. S. Geological Survey, the salient results of which have been made available through written and oral communications. The geölogical data that bear on the subsurface structure is summarized here.

The slope of the ancient erosional surface in the salient of the Sierra Nevada toward the valley block together with the absence of marginal valley-down faults of large throw constitutes a first point of evidence. A second point 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 conjuction 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 terraces indicate that the east side of the valley has been subsiding more rapidly for an extended period. The possibility that the progressive tilting of terraces and the cutting of new ones is correlative 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 in Owens Valley, but only about 40 were of use in interpreting the subsurface structure in the northern part of the valley. All of the logs used penetrated into the Bishop tuff and through the layer of white pumice at its base. This layer provides an excellent 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 chief means of identification of the pumice in the logs is that it is conspicuously white and a good aquifer--properties that usually were noted by the drillers. The structure derived from these logs is shown in a fence diagram (plate 6) 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, and in structure contours drawn on the base of the pumice which are shown on the structure map of the Bishop quadrangle (plate 5).

The well data show that in a general way the pumice layer slopes southeasterly 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 the faults in the Volcanic Tableland and in particular the fault along Fish Slough, and by a northwesterly trending depression whose deepest part is just northeast of Bishop. The depression beneath Bishop seems likely to be the result of a fault downthrown to the southwest that had a maximum throw in the pumice layer of about 350 feet. It is conceivable but unlikely that it resulted entirely from warping. The position of the depression, opposite the point of the salient in the Sierra Nevada front suggests that it may be related to the crest of the anticlinal Coyote warp.

The present differences between the configuration of the pumice layer and a horizontal surface cannot be taken to be entirely the result of deformation, for undoubtedly the pumice had an initial dip. The average slope of the surface is in the same direction as the initial slope deduced for the initial slope-direction of upper surface of the Bishop tuff in the Volcanic Tableland. This suggests that the initial slope of the pumice was in the same direction as at present. Nevertheless, the initial slope farther to the south in the valley seems likely to have been more southerly than the initial dip in the area of the Volcanic Tableland. Probably the existing average dip south of the Volcanic Tableland is the composite result of southerly, initial dip and eastward tilting since deposition.

Conjugate joint system in the Sierra Nevada

A regional system of 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 that have not undergone recent rapid erosion. In such areas weathering of the rock adjacent to joints has resulted in linear depressions that follow the joints. Joints locally exert considerable control over drainage pattern, and even in deeply dissected regions straight segments of streams coincide with joints.

The joints occur in all of the granitic rocks and cross boundaries between different intrusive masses. Generally two principal joint sets that dip steeply and are almost at right angles to each other can be identified. One of these strikes northwesterly and the other northeasterly; the precise direction of strike changes from place to place although the two sets maintain their right-angle relationship to one another.

In addition to these principal joint sets other steeply-dipping joints are present locally and a gently-dipping set may also be recognizable, especially at higher altitudes where frost action has split the rock along joint surfaces. The attitude of the gently-dipping joints varies from place to place, but everywhere the joints strike parallel with the prevailing direction of the topographic contours and dip a trifle more gently than the ground surface. These relationships indicate that these joints were formed in relation to the present ground surface, probably as a result of unloading through erosion.

Individual joints can be traced for distances of as much as several miles. Most joints are almost straight, but some are gently curved. The joints in a given set generally are only 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, the joints from one direction interfinger with joints of different strike from another direction, and rarely does a joint curve from one direction to another.

Close inspection of the joints indicates that any movement that has taken place was parallel with the joint surface rather than perpendicular to it. Displacement can be established for relatively few joints, but offset dikes record displacements of as much as several inches, although the average probably is half an inch or less.

The fact that the joints continue without interruption across boundaries between intrusive masses and that the pattern is regional indicates that it is younger than the batholith. It seems likely that the joints are the product of regional forces that may be related in origin to the Cenozoic blocking and warping.

Summary

The subsidence of the valley block relative to the bordering ranges is believed to have been accomplished by means of countless small increments of movement, each 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 movements along many individual faults is clear, and abundant fresh fault scarps indicate that movements continued to the present at a significant rate and no doubt they will continue into the future.

Apparent continuity in the long and complicated sequence of structural events that took place in connection with the progressive structural depression of the valley block strongly suggests that a continuous pattern of movement has dominated events throughout the period of block faulting and warping. The prevalence of normal faults indicates tension; the concept of Owens Valley as a keystone block progressively subsiding between the Sierra Nevada and the White Mountains, possibly in the apex of an arch, fits the known facts. Downwarping in place of faulting along some segments of the boundary of the depressed block fits well with such a hypothesis, since the formation of the valley in such a mechanism would be largely the result of collapse under the influence of gravity.

The small horsts, graben, and tilted blocks on the Volcanic Tableland can be explained satisfactorily as the result of a clockwise rotational couple. The evidence of rotation might be dismissed as a result of local effects that were contingent on the depression of Owens and Round Valleys on the two sides of the Tableland, except for two factors. The first is that the north-trending faults displayed on the Volcanic Tableland are identified chiefly because the Bishop tuff preserves scarps so well; it seems likely that similar faulting occurred in adjacent areas but that the fault scarps were not preserved because the surface material was easily eroded alluvium. Probably similar faults cut the bedrock and older alluvial strata beneath Owens Valley. North-trending faults that moved during the earthquake of 1872 showed right-lateral displacement and the gross pattern of deformation is similar to that on the Volcanic Tableland. The second factor is that the structures on the Tableland are literally miniatures of Basin and Range block fault structures, with which they are in parallel orientation. In view of these considerations, it seems not unreasonable to suggest that the larger fault block structures in the Great Basin are also the result of a rotational couple.

REFERENCES CITED

Barth, T. F. W., 1952, Theoretical petrology: John Wiley and Sons, Inc., New York.

Bateman, P. C., 1953, Up-side-down graded bedding in right-side-up lacustrine pumice: Geol. Soc. America, v. 64, abstract, p. 1499-1500.

Bateman, P. C., and Merriam, C. W., 1954, Geology of the Owens Valley region, Inyo County, California: in Geology of Southern California, Calif. Div. of Mines, Bull. 170, Map sheet No. 11.

Bateman, P. C., 1956, Economic geology of the Bishop district, California: Calif. Div. of Mines, Special Report 47.

Blackwelder, Eliot, 1931, Pleistocene glaciation in the Sierra Nevada and Basin Ranges: Geol. Soc. America Bull., v. 42, p. 865-922.

Bowen, Norman L., 1940, Progressive metamorphism of siliceous limestone and dolomite: Jour. Geology, v. 48, p. 225-274.

Chayes, Felix, 1949, A simple point counter for thin section analysis: Am. Min., v. 34, p. 1-11.

Cloos, Hans, 1939, Hebung, Spaltung, Vulkanismus: Geol. Rundschau, Band 30, p. 401-527.

Eskola, P., 1939, Die Entstehung der Gesteine (Barth, T. F. W., Correns, C. W., and Eskola, P.): Springer, Berlin.

Fath, A. E., 1920, The origin of the faults, anticlines and buried "granite ridge" of the northern part of the Mid-Continent oil and gas field: U. S. Geol. Survey Prof. Paper 128c, p. 75-84.

Fenner, C. N., 1948, Incandescent tuff flows in southern Peru: Geol. Soc. America Bull., v. 59, p. 879-893.

Ferguson, H. G., and Muller, S. W., 1949, Structural Geology of the Hawthorne and Tonopah Quadrangles, Nevada: U. S. Geol. Survey Prof. Paper 216.

Folley, L. L., 1926, The origin of the faults in Creek and Osage Counties, Oklahoma: Am. Assoc. Petroleum Geologists Bull., v. 10, p. 293-303.

Hazzard, John C., 1937, Paleozoic section in the Nopah and Resting Springs Mountains, Inyo County, California: Calif. Jour. of Mines and Geology, v. 33, p. 273-339.

Gilbert, C. M., 1938, Welded tuff in eastern California: Geol. Soc. America Bull., v. 49, p. 1829-1862.

Gilbert, C. M., 1941, Late Tertiary geology southeast of Mono Lake, California: Geol. Soc. America Bull., v. 52, p. 781-816.

Goodyear, W. A., 1888, Inyo County: Calif. Div. of Mines, 8th Rept. of the State Mineralogist, p. 224-309.

Johannsen, Albert, 1931, A descriptive petrography of the igneous rocks; Vol. 1, Introduction, textures, classifications, and glossary: Univ. Chicago Press.

Knopf, Adolph, 1918, A geologic reconnaissance of the Inyo Range and eastern slope of the Southern Sierra Nevada: U. S. Geol. Survey Prof. Paper 110.

Lee, W. T., 1906, Geology and water resources of Owens Valley, California: U. S. Geol. Survey Water-Supply Paper 181.

Link, T. A., 1929, En echelon tension fissures and faults: Am. Assoc. Petroleum Geologists Bull., v. 13, p. 627-637.

Marshall, P., 1932, Notes on some volcanic rocks of the North Island of New Zealand: Jour. Sci. and Tech., v. 13, p. 198-202.

Marshall, P., 1935, Acid rocks of the Taupor Rotorva volcanic district: Trans. and Proc. Roy. Soc. of New Zealand, v. 64, p. 323-366.

Maxon, John V., 1935, Pre-Cambric stratigraphy of the Inyo Range: Geol. Soc. America Proc., abstract, p. 314.

Mayo, Evans B., 1941, Deformation in the interval Mt. Lyell-Mt. Whitney, California: Geol. Soc. America Bull., v. 52, p. 1001-1084.

Putnam, W. C., 1938, The Mono Craters, California: Geol. Rev., v. 29, p. 68-82.

Putnam, W. C., 1949, Quaternary geology of the June Lake district, California: Geol. Soc. America Bull., v. 60, p. 1281-1302.

Putnam, W. C., 1950, Moraine and shoreline relationships at Mono Lake, California: Geol. Soc. America Bull., v. 61, p. 115-122.

Putnam, W. C., 1952, Origin of Rock Creek and Owens River Gorges, California: Geol. Soc. America Bull., v. 63, p. 1291-1292 (abstract).

Rinehart, C. D., and Ross, D. C., 1957, Geology of the Casa Diablo Mountain quad.: U. S. Geol. Survey, Map series report GQ 99.

Sherrill, R. E., 1929, Origin of the en echelon faults in north-central Oklahoma: Am. Assoc. Petroleum Geologists Bull., v. 13, p. 31-37, 1398-1399.

Spurr, J. E., 1905, Descriptive geology of Nevada south of the fortieth parallel and adjacent portions of California: U. S. Geol. Survey Bull. 208.

Turner, F. J., 1948, Mineralogical and structural evolution of the metamorphic rocks: Geol. Soc. America Memoir 30.

Tuttle, O. F., 1952, Origin of the contrasting mineralogy of extrusive and plutonic salic rocks: Jour. Geology, v. 60, p. 107-124.

Walcott, C. D., 1895, Lower Cambrian rocks in eastern California: Am. Jour. Sci., 3rd ser., v. 49, p. 141-144.

Walcott, C. D., 1908, Cambrian sections of the Cordilleran area: Smithsonian Misc. Coll., v. 53, p. 185-188.

Wentworth, C. K., and MacDonald, G. A., 1953, Structures and forms of basaltic rocks in Hawaii: U. S. Geol. Survey Bull. 994.

Wheeler, Harry E., 1943, Lower and Middle Cambrian stratigraphy in the Great Basin area: Geol. Soc. America Bull., v. 54, p. 1781-1822.

Since list does not agree paging

REFERENCES CITED

Barth, T. F. W., 1952, Theoretical petrology: John Wiley and Sons, Inc., New York.

Bateman, P. C., 1953, Up-side-down graded bedding in right-side-up lacustrine pumice: Geol. Soc. America, v. 64, abstract, p. 1499-1500.

Bateman, P. C., and Merriam, C. W., 1954, Geology of the Owens Valley region, Inyo County, California: in Geology of Southern California, Calif. Div. of Mines, Bull. 170, Map sheet No. 11.

Bateman, P. C., 1956, Economic geology of the Bishop district, California: Calif. Div. of Mines, Special Report 47.

Blackwelder, Eliot, 1931, Pleistocene glaciation in the Sierra Nevada and Basin Ranges: Geol. Soc. America Bull., v. 42, p. 865-922.

Bowen, Norman L., 1940, Progressive metamorphism of siliceous limestone and dolomite: Jour. Geology, v. 48, p. 225-274.

Chayes, Felix, 1949, A simple point counter for thin section analysis: Am. Min., v. 34, p. 1-11.

Cloos, Hans, 1939, Hebung, Spaltung, Vulkanismus: Geol. Rundschau, Band 30, p. 401-527.

Eskola, P., 1939, Die Entstehung der Gesteine (Barth, T. F. W., Correns, C. W., and Eskola, P.): Springer, Berlin.

Fath, A. E., 1920, The origin of the faults, anticlines and buried "granite ridge" of the northern part of the Mid-Continent oil and gas field: U. S. Geol. Survey Prof. Paper 128c, p. 75-84.

Fenner, C. N., 1948, Incandescent tuff flows in southern Peru: Geol. Soc. America Bull., v. 59, p. 879-893.

Ferguson, H. G., and Muller, S. W., 1949, Structural Geology of the Hawthorne and Tonopah Quadrangles, Nevada: U. S. Geol. Survey Prof. Paper 216.

Folley, L. L., 1926, The origin of the faults in Creek and Osage Counties, Oklahoma: Am. Assoc. Petroleum Geologists Bull., v. 10, p. 293-303.

Hazzard, John C., 1937, Paleozoic section in the Nopah and Resting Springs Mountains, Inyo County, California: Calif. Jour. of Mines and Geology, v. 33, p. 273-339.

Gilbert, C. M., 1938, Welded tuff in eastern California: Geol. Soc. America Bull., v. 49, p. 1829-1862.

Gilbert, C. M., 1941, Late Tertiary geology southeast of Mono Lake, California: Geol. Soc. America Bull., v. 52, p. 781-816.

Goodyear, W. A., 1888, Inyo County: Calif. Div. of Mines, 8th Rept. of the State Mineralogist, p. 224-309.

Johannsen, Albert, 1931, A descriptive petrography of the igneous rocks; Vol. 1, Introduction, textures, classifications, and glossary: Univ. Chicago Press.

Knopf, Adolph, 1918, A geologic reconnaissance of the Inyo Range and eastern slope of the Southern Sierra Nevada: U. S. Geol. Survey Prof. Paper 110.

Lee, W. T., 1906, Geology and water resources of Owens Valley, California: U. S. Geol. Survey Water-Supply Paper 181.

Link, T. A., 1929, En echelon tension fissures and faults: Am. Assoc. Petroleum Geologists Bull., v. 13, p. 627-637.

Marshall, P., 1932, Notes on some volcanic rocks of the North Island of New Zealand: Jour. Sci. and Tech., v. 13, p. 198-202.

Marshall, P., 1935, Acid rocks of the Taupor Rotorva volcanic district: Trans. and Proc. Roy. Soc. of New Zealand, v. 64, p. 323-366.

Maxon, John V., 1935, Pre-Cambric stratigraphy of the Inyo Range: Geol. Soc. America Proc., abstract, p. 314.

Mayo, Evans B., 1941, Deformation in the interval Mt. Lyell-Mt. Whitney, California: Geol. Soc. America Bull., v. 52, p. 1001-1084.

Putnam, W. C., 1938, The Mono Craters, California: Geol. Rev., v. 29, p. 68-82.

Putnam, W. C., 1949, Quaternary geology of the June Lake district, California: Geol. Soc. America Bull., v. 60, p. 1281-1302.

Putnam, W. C., 1950, Moraine and shoreline relationships at Mono Lake, California: Geol. Soc. America Bull., v. 61, p. 115-122.

Putnam, W. C., 1952, Origin of Rock Creek and Owens River Gorges, California: Geol. Soc. America Bull., v. 63, p. 1291-1292 (abstract).

Rinehart, C. D., and Ross, D. C., 1957, Geology of the Casa Diablo Mountain quad.: U. S. Geol. Survey, Map series report GQ 99.

Sherrill, R. E., 1929, Origin of the en echelon faults in north-central Oklahoma: Am. Assoc. Petroleum Geologists Bull., v. 13, p. 31-37, 1398-1399.

Spurr, J. E., 1905, Descriptive geology of Nevada south of the fortieth parallel and adjacent portions of California: U. S. Geol. Survey Bull. 208.

Turner, F. J., 1948, Mineralogical and structural evolution of the metamorphic rocks: Geol. Soc. America Memoir 30.

Tuttle, O. F., 1952, Origin of the contrasting mineralogy of extrusive and plutonic salic rocks: Jour. Geology, v. 60, p. 107-124.

Walcott, C. D., 1895, Lower Cambrian rocks in eastern California: Am. Jour. Sci., 3rd ser., v. 49, p. 141-144.

Walcott, C. D., 1908, Cambrian sections of the Cordilleran area: Smithsonian Misc. Coll., v. 53, p. 185-188.

Wentworth, C. K., and MacDonald, G. A., 1953, Structures and forms of basaltic rocks in Hawaii: U. S. Geol. Survey Bull. 994.

Wheeler, Harry E., 1943, Lower and Middle Cambrian stratigraphy in the Great Basin area: Geol. Soc. America Bull., v. 54, p. 1781-1822.

Pocket contains
6 pieces

(200)
R290
no. 436

(Dashed where app
dotted whe

Fault,
(Dashed where ap
dotted whe

High-angle f
(Ball shows dire

Faul

4294
list # 117
no 436

U. S

This map or
and has not
conformity w
standards or

(11)

PLE
IN