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Stratigraphy and Structure Death Valley, California

By CHARLES B. HUNT *and* DON R. MABEY

GENERAL GEOLOGY OF DEATH VALLEY, CALIFORNIA

GEOLOGICAL SURVEY PROFESSIONAL PAPER 494-A

Stratigraphy and structural geology, both of the surficial deposits and bedrock. Two companion reports describe the hydrology, saltpan, and plant ecology.

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GENERAL GEOLOGY OF DEATH VALLEY, CALIFORNIA

STRATIGRAPHY AND STRUCTURE

By Charles B. Hunt and Don R. Mabey

ABSTRACT

Death Valley is in southeastern California at the south edge of the Great Basin. The valley and the mountains around it have been the site of three major geosynclines—one formed during late Precambrian time, another during the Paleozoic, and a third during the early part of the Mesozoic Era.

During late Mesozoic and early Cenozoic time the southern Great Basin was part of a geanticline that was folded, thrust faulted, and invaded by granitic intrusions, and that shed sediments to surrounding regions. Later in Cenozoic time the southern Great Basin, including Death Valley, became fragmented, mostly by block faulting, into basins and ranges, and during this time sediments that were eroded from the ranges collected in the basins.

The rocks exposed in Death Valley and the adjoining mountains aggregate more than 60,000 feet in thickness, which is only a little more than half the aggregate thickness formed during the geosynclinal and other episodes in this part of the Great Basin. The rocks that are missing probably once were present, but they have been removed by erosion (represented by unconformities) or are concealed by structural discontinuities.

Precambrian rocks are in three major groups. The oldest, which are metamorphic rocks representing the crystalline basement complex, have a structural and topographic relief of more than 3,000 feet.

Overlying the basement complex is the Pahrump Series, which comprises much younger and only slightly metamorphosed formations that are mostly clastic sedimentary rocks, but they include some limestone and dolomite and some diabase. These formations total at least 10,000 feet in thickness. They are not exposed one above the other in this area, and their stratigraphy is inferred from known relations elsewhere. Westward across the Panamint Range, each of the three formations of the series rests in turn on the metamorphic basement complex, probably as a result of thrust faulting rather than stratigraphic changes.

The third and youngest group of rocks included in the Precambrian are sedimentary formations, mostly clastic, but they include considerable dolomite and some limestone. These rocks are slightly less metamorphosed than those of the Pahrump Series; the metamorphism is about the same as that of the Cambrian rocks. However, these formations lie below the *Olenellus* fauna, which is taken to mark the base of the Cambrian. Their thickness aggregates 7,000 feet.

An unusually complete section of Paleozoic formation is exposed in Tucki Mountain where rocks ranging in age from Early Cambrian to Permian, and representing all the intervening systems, are more than 20,000 feet thick. The Lower Cambrian formations are mostly clastic sedimentary rocks, but the rest of the Paleozoic formations are chiefly limestone and dolomite. The Permian rocks include much conglomerate or breccia derived from Paleozoic formations at least as old as Early Devonian and as young as Late Pennsylvanian. Evidently there was considerable deformation during Permian time; it may have begun in Pennsylvanian time.

At the south end of the Panamint Range, only a mile outside the area being reported upon, Triassic formations total 8,000 feet thick. These formations are composed of volcanic and clastic sedimentary rocks, and represent a return to conditions like those of the Precambrian. Moreover, the thick remnants of the Triassic, like the thick remnants of the Precambrian Pahrump Series, are restricted to a northwest-trending belt approximately coinciding with the edge of the Sierra Nevada batholith.

Two granitic intrusions that seem to be eastern satellites of the Sierra Nevada batholith lie within the mapped area. They are referred to as the granites at Hanaupah Canyon and Skidoo, and probably are floored intrusions that spread laterally along thrust faults and made the space they occupy by doming the rocks of the upper plates of the thrusts.

The intrusions of the batholith in the Sierra Nevada are Late Jurassic and Cretaceous in age. The granitic intrusions in the Death Valley area are closely related to the volcanism, which is of middle Tertiary age and these granites are younger than the main part of the batholith. This raises a philosophical question as to how widely apart in time and space individual plutons can be and still be part of a composite batholith.

Evidence for a close relationship between the granitic intrusions and the volcanics in the Death Valley area is found along the east foot of the Panamint Range. There a complex of may kinds of igneous, metamorphic, and sedimentary rocks occurs along a thrust fault believed to be the westward extension of the Amargosa thrust. Precambrian augen gneiss cut by a granitic intrusion and a swarm of still younger dikes, underlies the thrust fault. A similar augen gneiss and similar granitic intrusion underlie the Amargosa thrust at the Virgin Springs district 20 miles to the southeast across Death Valley. Zircons in the Precambrian rocks differ from those in the dikes; the granite contains both kinds.

Overlap of lavas and associated eruptives onto Paleozoic rocks of the Panamint Range shows that the eastward tilting of the Range occurred half before, and half after, the eruptives were deposited.

That the volcanic rocks along the belt of the Amargosa thrust complex are Tertiary is indicated by the stratigraphy of the very similar volcanic rocks in the Tertiary formations along the east and north sides of Death Valley. Tertiary formations in the Black Mountains east of Death Valley are at least 12,000 feet thick. The older deposits, volcanics 5,000 feet thick in the Artists Drive area, are quite like those in the Amargosa thrust complex. They are faulted onto the Precambrian core of the mountains. Northward these volcanics grade laterally into

playa and other sedimentary deposits. They dip northward and thin under a syncline separating the Black and Funeral Mountains. Where the older formations rise again on the north flank of the syncline, at the base of the Funeral Mountains, they are very similar to the Titus Canyon Formation (Oligocene) of Stock and Bode (1935), and are tentatively correlated with it.

In the trough of the syncline is the Furnace Creek Formation of Pliocene age, which is capped by and intertongues with the late Pliocene and early Pleistocene(?) Funeral Formation. Between the outcrops of the Furnace Creek and Titus Canyon(?) Formations is a faulted belt of different-looking sedimentary deposits which, on the basis of structural position, are assumed to be of an intermediate age and accordingly designated Miocene(?).

The oldest deposits in Death Valley classed as Quaternary are cemented fan gravels included with the Funeral Formation. In places the Funeral Formation is conformable on and intertongues with the playa deposits of the Furnace Creek Formation of Pliocene age, but more commonly the Funeral rests with angular unconformity on the older rocks. The Funeral Formation has been displaced thousands of feet by faulting and tilting during the late stages in the structural development of Death Valley and the bordering mountains.

Subsequent to most of that deformation huge gravel fans were built from the mountains to the floor of the valley. Some of these are 6 miles long and more than a thousand feet high. The oldest of these fan gravel deposits, referred to as the No. 2 gravel, still has a distinct fan form which the older Funeral Formation has lost because of deformation and erosion. Both the No. 2 gravel and Funeral Formation have smooth surfaces of desert pavement. Boulders and cobbles on these surfaces are deeply weathered and have disintegrated to produce a new mantle of angular rock fragments. The No. 2 gravel is surely late Pleistocene in age, but it may be pre-Wisconsin.

Other deposits of late Pleistocene age include a debris avalanche at the front of the Black Mountains and some isolated poorly developed beach deposits of a late Pleistocene lake, which had a maximum depth of about 600 feet. The lake, though, was of brief duration and evidently its level fluctuated rapidly, so that beach deposits and other shore features are poorly developed as compared with those around other Pleistocene lake basins in the Great Basin.

Younger gravels on the fans, referred to as No. 3, may include some late Pleistocene deposits and certainly include some Recent deposits.

Other deposits that may be of approximately this age are mounds of travertine at springs on the gravel fans. Some travertine, of course, is being deposited at present, but the occurrence on these mounds of stone artifacts representing the earliest human occupation of the area indicates that the main parts of the mounds have considerable antiquity.

The youngest gravel on the fans, the No. 4 gravel, is along the washes. These deposits are loose gravel composed of firm rocks without desert varnish.

During the Recent, but probably during the 3 millenia preceding the Christian Era, lakes flooded the floor of Death Valley to a maximum depth of 30 feet. The salt deposits comprising the saltpan were formed as a result of this lake.

The principal structural features of Quaternary age are (1) the north-south trough that is Death Valley and the bordering upfaulted mountain blocks; (2) the northwest-trending Furance Creek fault zone and the downwarp that extends along Furnace Creek and northwestward across the northern part of Death Valley; (3) the northwest-trending Confidence Hills fault zone that extends into the south end of Death Valley; and (4) some features of Copper Canyon and Butte Valley. Deformation is going on at present, as indicated by measurable tilt at seven tiltmeter stations that have been established in the valley.

The composition and extent of the Furnace Creek Formation of Pliocene age indicate that it was deposited in a narrow trough that extended southeastward from Mesquite Flat across the Salt Creek Hills and Cottonball Basin and along the Texas Spring syncline and north end of the Black Mountains. The playa in which the formation was deposited existed long enough to accumulate 5,000 feet of beds.

Much or most of the uplift of the Black Mountains occurred after the Furnace Creek Formation was deposited because the formation dips 45° or more off the north end of the mountains. Gravity data, however, indicate that the formation probably thins northward under the Texas Spring syncline, and presumably the thinning is by offlap from the mountains. If so, part of the uplift of the Black Mountains occurred while the Furnace Creek Formation was being deposited. It is inferred that roughly 4,000 feet of uplift at the Black Mountains occurred while the Furnace Creek Formation was being deposited, that another 3,500 feet of uplift occurred during early Pleistocene time, and the last 2,500 feet of uplift occurred in late Pleistocene and Recent time.

The Miocene(?) and older Tertiary formations exposed in fault blocks between the Funeral Mountains and the trough in which the Furnace Creek Formation was deposited are mostly coarse clastics that were derived from the Funeral Mountains. The mountains and the adjoining basin therefore were in existence in mid-Tertiary time.

The basins and ranges in this part of the Great Basin are at least as old as the Titus Canyon and Artist Drive Formations, although the structural limits of those basins and ranges probably were different from the present ones.

The structural history of the region during the earlier geanticlinal stage is obscure. The principal features are the westward-directed Amargosa thrust, the chaos that accompanies it, the smoothly exhumed surfaces of the thrust faults locally known as turtlebacks, and the granitic intrusions that seem to have spread along the thrust faults.

A short segment of the Amargosa thrust is exposed along the east foot of the Panamint Range. The lower plate there, composed of Precambrian metamorphic rocks, is cut by a granitic intrusion. The metamorphic rocks include an augen gneiss; locally the augen are collected into small pegmatitic masses that grade into the dikes of Tertiary age that cut all the rocks in the lower plate. Part of the metamorphism of the lower plate of Precambrian rocks may have occurred at the time the granitic intrusion was emplaced.

The Paleozoic and late Precambrian sedimentary rocks in the mountains bordering Death Valley occur in a series of thrust plates of the Amargosa thrust system. The thrusting moved younger rocks westward onto older ones. Within a thrust plate the rocks have uniform homoclinal dips, almost invariably to the east. The major structural units are grouped into four klippen and three fensters.

The most completely exposed klippe is at Tucki Mountain where Paleozoic formations ranging in age from Early Cambrian to Permian have been thrust westward onto the Kingston Peak(?) Formation of late Precambrian age. The klippe is divided into four plates by thrust faults that, towards the east, branch upward from the main one at the base. Along these branch faults the displacement is 4 miles westward;

along the main fault the displacement must be very much more than that.

The Panamint Range south of Tucki Mountain also is a klippe of east-dipping Paleozoic rocks thrust westward onto the Precambrian. The thrust fault at the base, the Amargosa thrust, is exposed at the east foot of the range. There Paleozoic formations in the upper plate dip east and rest on Precambrian metamorphic rocks in the lower plate. Other thrust faults within the Paleozoic formations also seem to be branches extending upward from the Amargosa thrust. Some of these branch faults are intruded by sills from the granite at Hanaupah Canyon.

The south end of the Funeral Mountains and the southern part of the Grapevine Mountains comprise klippen of Paleozoic formations thrust westward onto the Precambrian. Between these two thrust plates is a fenster of Precambrian formations forming the northern part of the Funeral Mountains. The two klippen may join under the Amargosa desert east of the Funeral Mountains.

The fenster in the northern part of the Funeral Mountains is formed by anticlinally domed Precambrian formations. So also is the west side of Tucki Mountain, another fenster, and the Black Mountains. The uplift at the Black Mountains is divided into three fensters, each a smooth-surfaced dome or turtleback.

It is suggested that the thrust plates of the Amargosa thrust system in part are detachment blocks, and that the turtleback fault surfaces were denuded tectonically.

Over the main part of the Panamint Range, Bouguer gravity-anomaly values are lower than over the mountains east of Death Valley and lower than those over the Slate Range to the southwest, suggesting that the Panamint Range is underlain by a granitic mass. In terms of deep crustal structure the geologic and gravity data suggest two possibilities. One is that deep under the Panamint Range is a large granitic intrusion that connects westward with the Sierra Nevada batholith and forms a bulbous thickened edge of the batholith. A second possibility is that the edge of the batholith is in the area that is seismically active west of the Panamint Range and that the deeply buried granite is mostly Precambrian. By the latter interpretation the granites at Skidoo and Hanaupah Canyon, and other granitic intrusions, could be attributed to palingenesis of the Precambrian granitic rocks.

INTRODUCTION

By Charles B. Hunt

LOCATION AND DESCRIPTION OF THE VALLEY

Death Valley, one of the valleys of the Basin and Range province, is at the south edge of the Great Basin, about midway between the Colorado River and the Sierra Nevada. Just to the south is the Mojave Desert (fig. 1).

The valley trends north-south between block faulted mountains (fig. 2). In the main part of the valley, the floor is a flat playa crusted with salts, one of the great natural saltpans of the world. The saltpan covers more than 200 square miles and is 250–280 feet below sea level. It is the sink for drainage from a hydrologic basin covering 8,700 square miles. An arm of Death Valley extends 60 miles north-northwestward from the saltpan and rises to about 3,000 feet in altitude. Discharging into the south end of the saltpan is the Amargosa River which rises northeast of Death Valley, flows southward along a valley 25 miles to the east, and then makes a great U-turn to discharge into the south end of Death Valley.

The mountains bordering Death Valley are north-trending fault blocks. Largest and highest of these blocks is the Panamint Range along the west side (fig. 3). Its highest peak, which is more than 11,000 feet above sea level, is only 12 miles from the edge of the saltpan. Both in terms of local relief and local ruggedness, this is one of the roughest terrains in the United States.

Along the east side of the saltpan are the Black Mountains which rise precipitously from the edge of the saltpan to a summit at about 6,000 feet in altitude (fig. 4). Northeast of the saltpan and set back from it by long gravel fans are the Funeral Mountains (fig 5). To the north are the Grapevine Mountains which lie along the east side of the northwest arm of Death Valley.

Death Valley is a desert area. The floor of the valley in fact is the hottest and driest part of the United States. Annual rainfall is only 1.65 inches; the evaporation rate is 150 inches annually! On the floor of the valley a maximum summer temperature of 134°F has been recorded. Minimum temperatures in winter almost never reach as low as freezing. The mountains, of course, have more moderate temperatures and considerably more precipitation; the summit of the Panamint Range is covered with snow most of each winter. Additional climatic data are given by Hunt and Robinson (in Hunt and others, 1965).

There is little surface water in Death Valley, but springs with potable water are surprisingly numerous.

Death Valley below an altitude of about 5,000 in part of the Lower Sonoran zone, characterized by the creosotebush. Vegetation is scanty, and there are almost no trees except for some mesquite around the edge of the saltpan where there is a zone of springs. The saltpan, covering more than 200 square miles, is without vegetation. The Panamint Range above 7,500 feet is wooded with piñon pine and juniper; limber and bristlecone pines grow at the summit.

The archeologic record in Death Valley reveals a long series of prehistoric occupations, or seasonal visitors, going back to early Recent and perhaps to late Pleistocene time. The first recorded entrance of white people into the valley was in 1849 when a party of emigrants, heading for the gold fields in California, left their caravan, sought a short cut, and became lost. One of the groups gave the name to the valley. During the next

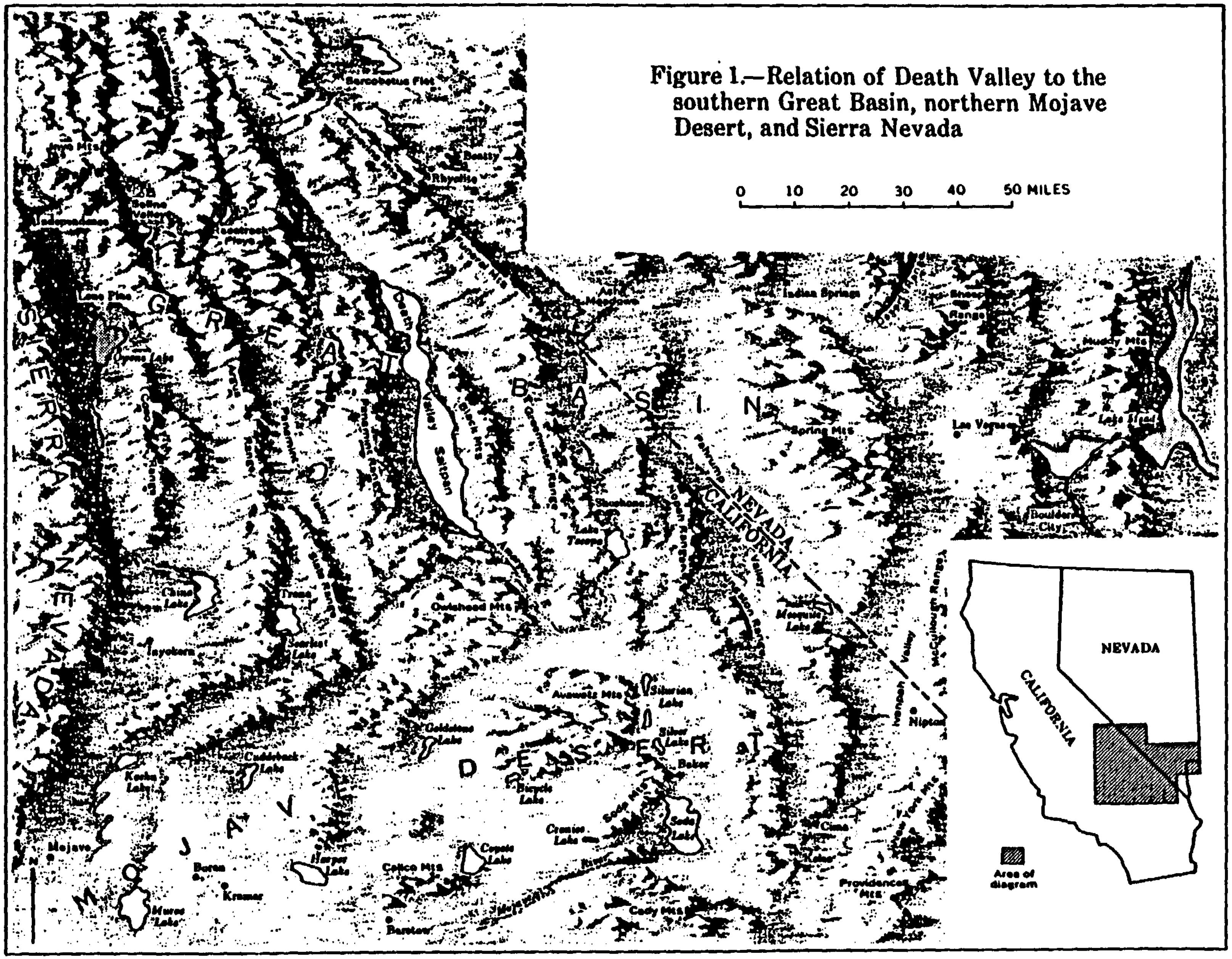

Figure 1.—Relation of Death Valley to the southern Great Basin, northern Mojave Desert, and Sierra Nevada

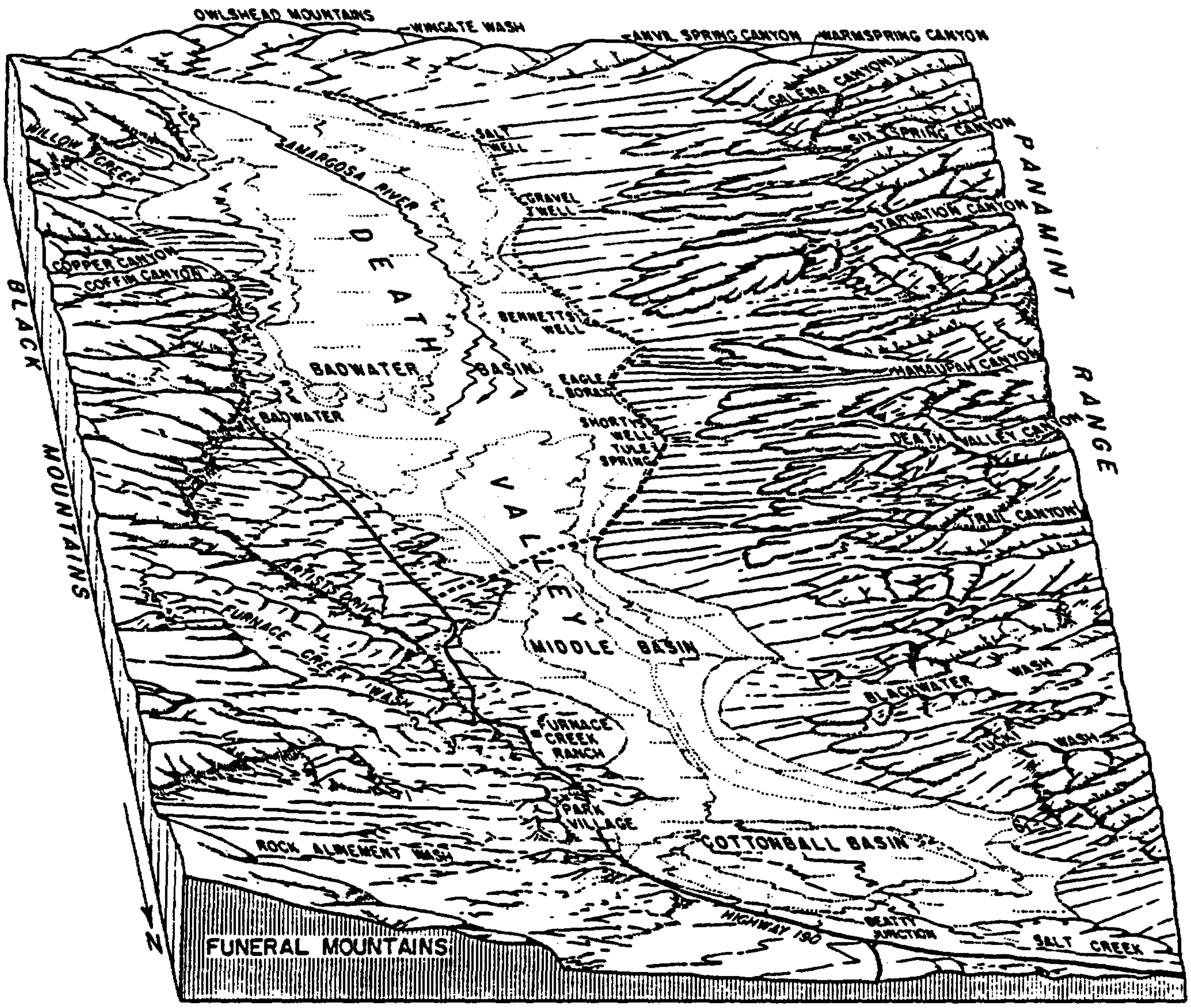

FIGURE 2.—Block diagram of Death Valley, Calif., looking south.

1. Nevares Spring.
2. Travertine Spring.
3. Texas Spring.
4. Corkscrew Canyon.
5. Coyote Hole.
6. West Side Borax Camp (Shoveltown).

FIGURE 3.—View west across Death Valley to the Panamint Range. Photograph by H. E. Maldé.

FIGURE 4.—View southeast across Death Valley to the south end of the saltpan and mouth of the Amargosa River. Photograph by W. B. Hamilton.

30 years prospectors searching for metal mines visited the mountains bordering Death Valley, and many of the local place names were given by them or commemorate their excursions.

Early in the 1880's borax was discovered on the Death Valley saltpan and the discovery led to production. In the next year or two richer deposits were found in the northern part of the Black Mountains and production shifted there. Produce from these mines was hauled to Mojave by the well-advertised 20-mule teams. In the first decade of this century the Skidoo mining district produced a little gold, but the Death Valley area has produced very little metal. Since the midtwenties mining activity has been slight, except for the production of talc in the mountains bordering the south end of Death Valley.

Land surveys in Death Valley were begun in 1857. The fact that the valley floor is below sea level seems to have been discovered about 1861 as a result of barometric observations by one of the several Nevada-California Boundary Commissions. This was confirmed in the midseventies by the Wheeler Surveys when satisfactory maps of the area first became available. During 1905–06 the first topographic map of the area was prepared and the altitude of the valley floor determined instrumentally.

Death Valley was made a National Monument in 1933. It has become a popular winter resort, and good highways lead to it from all directions.

Two major scientific contributions have come from Death Valley, one in the field of plant ecology, the other in geology. The ecology contribution was made in 1893 by Frederick V. Coville, a member of the Death Valley expedition. His report on the botany and its relationship to the environment is a classic.

The major contribution in geology was by Levi F. Noble (1941) who was the first to demonstrate clearly the existence of westward-directed thrust faults in the southern part of the Great Basin.

MAPPING OF THE VALLEY

Different parts of the geologic map (pl. 1) were prepared in very different ways and they are of quite different quality.

The saltpan was mapped by traverses across the valley floor at 1-mile intervals. The ground changes were plotted on aerial photographs, and the boundaries traced along the valley floor to tie into the preceding day's

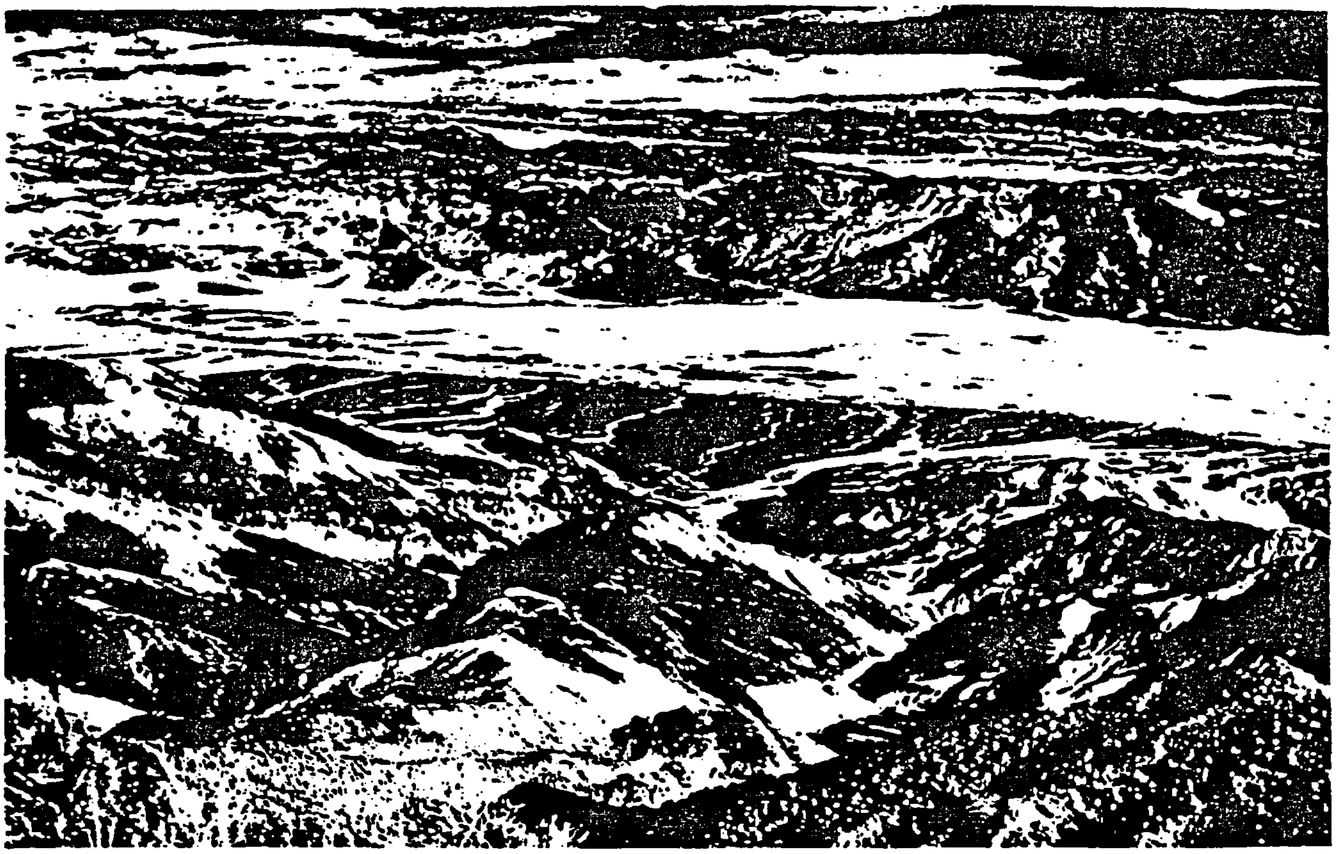

FIGURE 5.—View northeast across Death Valley to the Black Mountains and Funeral Mountains. Photograph by W. B. Hamilton.

traverse. The boundaries then were projected directly from the photographs onto the topographic base maps.

Mapping of the gravel fans involved sufficient traverses on each fan to identify the gravel deposits according to their distinctive color and tone on the aerial photographs. The boundaries of the several formations then were sketched with much generalization on the photographs, and these lines were transferred by projection and inspection to the topographic base. In general, the boundaries of the older of the late Pleistocene gravel (No. 2 gravel) are shown by the contours; the younger gravels commonly are in such small areas that a great deal of generalization was necessary, so much so that accurate projection onto the topographic base seemed an overrefinement.

The lower parts of the mountains were mapped on the ground. For the most part the traverses were along the valleys and along the ridges, although this is at right angles to the contacts. The higher parts of the mountains were mapped by helicopter. Within limits, this mapping from the air could be checked by tracing the formations to the lower parts of the mountains where mapping had been done on the ground and where fossil collections had helped control it. Later, some of the most questionable areas were checked by traverses down the ridges and canyons from the summit of the Panamint Range to the floor of Death Valley.

In the course of mapping the Paleozoic formations, about 100 collections of fossils were obtained. Their locations are shown on the geologic map, and the contents of most of the collections are described with the formations. It is expected that future work will show that the formations in some fault blocks have been incorrectly identified, but the fossil collections show the limits to which such corrections may be extended.

Fieldwork was interrupted while parts of the study still were incomplete. In particular, the Tertiary formations, the granitic intrusions, and the older of the Precambrian sedimentary formations need more study.

ACKNOWLEDGMENTS

Numerous specialists have contributed to various parts of this general report on the geology of Death Valley, and their contributions are acknowledged in the sections dealing with the particulars of the geology. Acknowledgment here is made to James Gilluly who was largely responsible for getting the project launched and persuading me to undertake it. T. S. Lovering visited the project each of the first 5 years and contributed ideas and other assistance on many phases of the

geology; his encouragement and guidance were a very real benefit to the project.

My wife, Alice P. Hunt, studied the archeology while I was working on the geology. We found that working together in the field was of mutual benefit, and each of us was able to assist the other.

We were headquartered in the field with the staff of the National Park Service. Their many favors and hospitality greatly assisted the project and made the field seasons most pleasurable.

STRATIGRAPHY

By Charles B. Hunt

Precambrian rocks exposed in the mountains bordering Death Valley include at least 3,000 feet of rocks belonging to the crystalline basement and a sequence of much younger and but slightly metamorphosed sedimentary rocks, mostly clastic, totaling roughly 10,000 feet thick, referred to as the Pahrump Series. Overlying the Pahrump, and also included in the Precambrian are three formations—the Noonday Dolomite, Johnnie Formation, and Stirling Quartzite, totaling about 7,000 feet thick.

Paleozoic rocks, mostly carbonate rocks and representing all the periods from Cambrian to Permian, aggregate about 20,000 feet thick. Triassic formations 8,000 feet thick are exposed just outside the mapped area; they include carbonate rocks, fine-grained clastic rocks, and volcanics. Table 1 summarizes the formations recognized in the area and nearby.

The mapping in the Panamint Range and Funeral Mountains on which this report is based was done by Hunt, partly by conventional ground methods and partly by helicopter (Hunt, 1960). The mapping of the Precambrian rocks in the Black Mountains was done by Harald Drewes (1963). The distribution of the formations is shown on the geologic map (pl. 1). Most of the identifications of Cambrian fossils are by Palmer; most of the identifications of Ordovician fossils are by Ross. Other palentologists who identified and reported on the fossils include Ellis L. Yochelson,

Table 1.—*Precambrian, Paleozoic, and Triassic formations exposed in the Death Valley area*

System	Series	Formation	Lithology and thickness	Characteristic fossils
Triassic		Butte Valley formation of Johnson (1957).	Exposed in Butte Valley 1 mile south of this area, 8,000 ft of metasediments and volcanics.	Ammonites, smooth-shelled brachiopods, belemnites, and hexacorals.
Carboniferous	Pennsylvanian and Permian	Formations at east foot of Tucki Mountain.	Conglomerate, limestone, and some shale. Conglomerate contains cobbles of limestone of Mississippian, Pennsylvanian, and Permian age. Limestone and shale contain spherical chert nodules. Abundant fusulinids. Thickness uncertain on account of faulting; estimate 3,000 ft +; top eroded.	Beds with fusulinids, especially *Fusulinella*.
	Mississipian and Pennsylvanian(?)	Rest Spring Shale	Mostly shale, some limestone; abundant spherical chert nodules. Thickness uncertain because of faulting; estimate 750 ft.	None.
	Mississipian	Tin Mountain Limestone and younger limestone.	Mapped as 1 unit. Tin Mountain Limestone, 1,000 ft thick, is black with thin-bedded lower member and thick-bedded upper member. Unnamed limestone formation, 725 ft thick, consists of interbedded chert and limestone in thin beds and in about equal proportions.	Mixed brachiopods, corals, and crinoid stems. *Syringopora* (open-spaced colonies), *Caninia* cf. *C. cornicula*.
Devonian	Middle and Upper Devonian.	Lost Burro formation.	Limestone in light and dark beds 1–10 ft thick give striped effect on mountainsides. Two quartzite beds, each about 3 ft thick, near base; numerous sandstone beds 800–1,000 ft above base. Top 200 ft is well-bedded limestone and quartzite. Total thickness uncertain because of faulting; estimated 2,000 ft.	Brachiopods abundant, especially *Spirifer*, *Cyrtospirifer*, *Productilla*, *Carmarotoechia*, *Atrypa*. Stromatoporoids. *Syringopora* (closely spaced colonies).
Silurian and Devonian	Silurian and Lower Devonian.	Hidden Valley Dolomite.	Thick-bedded, fine-grained, and even-grained dolomite; mostly light color. Thickness 300–1,400 ft.	Crinoid stems abundant, including large types. *Favosites*.

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TABLE 1.—*Precambrian, Paleozoic, and Triassic formations exposed in the Death Valley area*—Continued

System	Series	Formation	Lithology and thickness	Characteristic fossils
Ordovician	Upper Ordovician	Ely Springs Dolomite.	Massive black dolomite; 400–800 ft thick.	Streptelasmatid corals: *Grewingkia, Bighornia.* Brachiopods.
	Middle and Upper(?) Ordovician.	Eureka Quartzite	Massive quartzite, with thin-bedded quartzite at base and top; 350 ft thick.	None.
	Lower and Middle Ordovician.	Pogonip Group	Dolomite, with some limestone, unit at base; shale unit in middle; massive dolomite unit at top. Thickness, 1,500 ft.	Abundant large gastropods in massive dolomite at top: *Palliseria* and *Maclurites,* associated with *Receptaculites.* In lower beds: *Protopliomerops, Kirkella,* Orthid brachiopods.
Cambrian	Upper Cambrian	Nopah Formation	Highly fossiliferous shale member 100 ft thick at base; upper 1,200 ft is dolomite in thick alternating black and light bands about 100 ft thick. Total thickness of formation 1,200–1,500 ft.	In upper part, gastropods. In basal 100 ft, trilobite trash beds containing: *Elburgia, Pseudagnostus, Homagnostus, Elvinia, Apsotreta.*
	Middle and Upper Cambrian.	Bonanza King Formation.	Mostly thick-bedded and massive dark-colored dolomite; a thin-bedded limestone member 500 ft thick 1,000 ft below top of the formation; 2 brown-weathering shaly units, the upper one fossiliferous, about 200 and 500 ft, respectively, below the thin-bedded member. Total thickness uncertain because of faulting; estimated about 3,000 ft in Panamint Range; 2,000 ft in Funeral Mountains.	The only fossiliferous bed is the shale below the limestone member that occurs near the middle of the formation. This shale contains linguloid brachiopods and trilobite trash beds with fragments of "*Ehmaniella.*"
	Lower and Middle Cambrian.	Carrara Formation.	An alternation of shaly and silty members with limestone members; transitional between the underlying clastic formations and the overlying carbonate ones. Thickness about 1,000 ft but variable because of shearing.	Numerous trilobite trash beds in lower part yield fragments of olenellid trilobites.
	Lower Cambrian	Zabriskie Quartzite.	Quartzite, mostly massive and granulated due to shearing; locally in beds 6 in. to 2 ft thick; not much crossbedded. Thickness more than 150 ft; variable because of shearing.	No fossils.
Cambrian and Cambrian(?)	Lower Cambrian and Lower Cambrian(?).	Wood Canyon Formation.	Basal unit is well-bedded quartzite about 1,650 ft thick; shaly unit above this 520 ft thick contains lowest olenellids in the section; top unit of dolomite and quartzite 400 ft thick.	A few scattered olenellid trilobites and archaeocyathids in the upper part of the formation. *Scolithus*? tubes.
Precambrian		Stirling Quartzite	Well-bedded quartzite in beds 1–5 ft thick comprising thick members of quartzite 700–800 ft thick separated by 500 ft of purple shale; crossbedding conspicuous in quartzite. Maximum thickness about 2,000 ft.	None.
		Johnnie Formation	Mostly shale, in part olive brown, in part purple. Basal member 400 ft thick is interbedded dolomite and quartzite with pebble conglomerate. Locally, tan dolomite near the middle and at the top. Thickness more than 4,000 ft.	None.
		Noonday Dolomite	In southern Panamint Range, dolomite in indistinct beds; lower part cream colored, upper part gray. Thickness 800 ft. Farther north where mapped as Noonday(?) Dolomite, contains much limestone, tan and white, and some limestone conglomerate. Thickness about 1,000 ft.	*Scolithus*? tubes.
		Unconformity		

TABLE 1.—*Precambrian, Paleozoic, and Triassic formations exposed in the Death Valley area*—Continued

System	Series	Formation	Lithology and thickness	Characteristic fossils
Precambrian	Pahrump Series	Kingston Peak(?) Formation	Mostly conglomerate, quartzite, and shale; some limestone and dolomite near middle. At least 3,000 ft thick. Although tentatively assigned to the Kingston Peak Formation, similar rocks along the west side of the Panamint Range have been identified as Kingston Peak.	None.
		Beck Spring Dolomite	Not mapped; outcrops are to the west. Blue-gray cherty dolomite; thickness estimated about 500 ft. Identification uncertain.	None.
		Crystal Spring Formation	Recognized only in Galena Canyon and south. Total thickness about 2,000 ft. Consists of a basal conglomerate overlain by quartzite that grades upward into purple shale and thinly bedded dolomite; upper part, thick-bedded dolomite, diabase, and chert. Talc deposits where diabase intrudes dolomite.	None.
		—Unconformity—		
		Rocks of the crystalline basement	Metasedimentary rocks with granitic intrusions.	None.

P. E. Cloud, Jr., W. A. Oliver, Jr., Chas. W. Merriam, Mackenzie Gordon, Jr., and Richard Rezak.

I have illustrated a number of the formations with pictures of slabs of fossiliferous rock, which some may find helpful as lithologic guides to the formations, as pointed out 30 years ago by Noble (1934, p. 175). Trilobites are the most abundant fossils in the Cambrian beds; gastropods in the Lower Ordovician beds; corals in the Upper Ordovician, Silurian, and Devonian; and crinoids in the Mississippian. Such generalized observations are very helpful in identifying rock formations that have been severely faulted and crumpled.

PRECAMBRIAN SYSTEM

ROCKS OF THE CRYSTALLINE BASEMENT

Precambrian rocks of the crystalline basement in this area are most extensive in the steep front of the Black Mountains. Some smaller outcrops are at the head of Galena Canyon and along the east foot of the Panamint Range north of Hanaupah Canyon.

The outcrop of the Precambrian in the head of Galena Canyon is mostly schist. The foliation in the schist, which is about vertical, is cut off discordantly by conglomerate at the base of the Crystal Spring Formation, the lowest formation in the Pahrump Series. The contact is thought to be depositional.

The outcrop of Precambrian metamorphic rock along the east foot of the Panamint Range north of Hanaupah Canyon marks the lower plate of a thrust fault, probably the Amargosa thrust; this lower plate has been the site of much igneous activity. The rocks are referred to as the Amargosa thrust complex and are described more fully on page A 129.

The Precambrian in the Amargosa complex is mostly gneiss, much of which is highly distinctive augen gneiss like that underlying the Amargosa thrust southeast of Mormon Point (Noble, 1941). The augen are feldspar, ¼–1 inch long, and constitute from 10 to 40 percent of the rock. The matrix is quartz and coarse-grained mica, mostly biotite. Enclosed in the augen gneiss are crushed lenses of biotite schist and some muscovite schist.

Above the augen gneiss is a crush zone consisting of mylonite and breccia along a thrust fault, probably the Amargosa thrust. The upper plate here is Stirling Quartzite.

The gneiss is cut by a swarm of northward-trending felsite dikes; their strike parallels the foliation in the gneiss.

Northward the gneiss extends under interlayered volcanic and Paleozoic rocks suggestive of the "chaos" as described by Noble (1941). Underlying the gneiss is a granitic intrusion.

Much interest attaches to the age of the metamorphism of this gneiss, because part of the metamorphism may have occurred during the early or middle Tertiary when the thrust faulting occurred and the granitic intrusion became emplaced.

The rocks of the Black Mountains have been described by Drewes (1963) who has mapped the Funeral Peak quadrangle. The following descriptions are summarized from his report.

The rocks comprise metadiorite and smaller bodies of metasedimentary schist, gneiss, and marble. The schist, gneiss, and marble underlie the lower parts of northwest-trending mountain spurs at Mormon Point and the mountain southeast of Copper Canyon. The schist and gneiss are in about equal proportions. Near Mormon Point, marble is equally abundant, but elsewhere it is less so. The thickness of the metasedimentary rocks is unknown, but is of the order of many hundreds of feet.

The schist generally consists of quartz (10–25 percent), plagioclase (15–25 percent), chlorite (30–60 percent), biotite (10–20 percent), and sericite; it has minor amounts of magnetite, sphene, and possibly potassium feldspar.

The gneiss consists largely of quartz (20–40 percent), plagioclase feldspar (15–40 percent), and potassium feldspar (as much as 40 percent) in light-colored layers, mostly less than 1 inch thick, alternating with thin dark layers of biotite. Some facies of the gneiss have layers of muscovite instead of biotite. The layers are mostly even and distinct. In places the gneiss contains feldspar augen ½–1 inch long. Veins and irregular masses of epidote are common.

The marble is white to light olive gray and weathers yellowish gray to pale yellowish brown. It occurs in lenticular beds ranging from a few feet to a few tens of feet thick interbedded with the schist or gneiss. Most of the marble is coarsely crystalline calcite, but some at Mormon Point is dolomitic.

Most of the front of the Black Mountains is Precambrian metadiorite. This rock, which is not foliated, and consists largely of plagioclase feldspar (40–65 percent), hornblende (20–40 percent), and biotite (3–15 percent). It appears to be intrusive into the metasedimentary rocks.

In the course of my survey, samples were collected from all the formations for spectrographic scanning for trace elements. Table 2 lists the trace elements found in a piece of schist from Galena Canyon and in a piece of gneiss from the Black Mountains above Badwater. Analyses of the augen gneiss and biotite gneiss in the Amargosa thrust complex at the east foot of the Panamint Range north of Hanaupah Canyon are given in table 26.

The trace elements in the schist and gneiss are remarkably alike, considering the very different lithologies and locations of the rocks. They differ from the gneisses in the Amargosa thrust complex in containing more lead, manganese, cobalt, vanadium, titanium, boron, and gallium.

TABLE 2.—*Trace elements in Precambrian metamorphic rocks*

[Analyses of the metamorphic rocks in the Amargosa thrust complex, where the rocks may have been subject to Tertiary alteration, are given in table 26. Semi-quantitative spectrographic analyses by Uteana Oda, U.S. Geol. Survey. Values are in parts per million, except Mg, which is given in percent]

Element	Schist in Galena Canyon	Gneiss above Badwater
Pb	30	30
Mn	500	700
Cu	50	30
Zr	150	100
Ni	50	30
Co	20	70
V	150	300
Y	20	20
Be	3	1
Ti	5, 000	10, 000
B	70	70
Ga	70	30
La	70	50
Cr	100	30
Ba	700	2, 000
Sr	20	700
Mg	1. 5	5

NOTE.—Also found: W, <20; Nb, <10; Zn, <200; La, <50; Sc, <10; Mo, <2 or less; Ag, <1; Bi, <5; Sn, <10; As, <500; Sb, <50.

Metamorphism of schist under the Pahrump Series in the southern Panamint Range is dated as 1,700 million years by potassium-argon methods and 1,480 million years by strontium-rubidium methods (Wasserburg and others, 1959). A comparable age has been determined for the augen gneiss zircon in the Amargosa thrust complex (T. W. Stern, written commun., 1965).

PAHRUMP SERIES

The Pahrump Series, named for Pahrump Valley 50 miles east of Death Valley (see index map, fig. 1; Hewett, 1940, 1956), in its type area consists of three formations. At the base is the Crystal Spring Formation, about 2,000 feet thick, consisting of a basal quartzitic member overlain by thick-bedded dolomite, diabase, and chert. Above the Crystal Spring Formation is the Beck Spring Dolomite, 1,100 feet thick, which is made up mostly of beds of light-blue-gray dolomitte 2–4 feet thick. At the top, in the type area, is the Kingston Peak Formation, 1,000–2,000 feet thick, which consists of sandstone and shaly sandstone at both the top and base, separated by a middle unit of conglomerate with cobbles as much as 10 inches in diameter (Hewett, 1956).

In the Silurian Hills, southeast of Death Valley, the rocks assigned to the Pahrump Series are more than 11,000 feet thick, but correlation with the type section is uncertain (Kupfer, 1960, p. 188). In the Alexander Hills of Pahrump Valley, rocks assigned to the Pahrump total 8,000 feet thick and appear to represent all three of the formations recognized in the type section (Wright, 1954).

A section of the Pahrump Series very similar to that in the type area is present also in the Ibex Hills in the southern part of Death Valley (Wright, 1952), an area about midway between the type locality and the southern part of the Panamint Range.

In the southern part of the Panamint Range, both in Galena Canyon and in the section of Warm Springs Canyon south of the area I studied, only the Crystal Spring Formation is present. It rests with angular unconformity on schist of the crystalline basement and is overlain by Noonday Dolomite. This contact between the Crystal Spring Formation and the Noonday Dolomite probably is a thrust fault and not a depositional contact. In any case, younger formations of the Pahrump Series are not present in the southeastern part of the Panamint Range, but they are present along the summit and west slope of the range.

On the west side of the Panamint Range the Kingston Peak Formation rests on the Precambrian crystalline basement; the Crystal Spring Formation is absent there (Johnson, 1957, p. 360). Locally, a dolomitic limestone that is much crushed occurs along the contact between the Kingston Peak Formation and underlying crystalline rocks (Murphy, 1932, p. 343); this dolomitic limestone may correlate with the Beck Spring Dolomite.

The spatial relationships of the three formations and their possible structural relations are illustrated on figure 6.

CRYSTAL SPRING FORMATION

The Crystal Spring Formation in Galena Canyon is about 2,000 feet thick. It consists of basal conglomerate overlain by quartzite that grades upward into purple shale and thinly bedded dolomite. The upper part of the formation consists of thick-bedded dolomite, diabase, and locally, massive chert (fig. 7). The following is a section of the Crystal Spring Formation measured in Galena Canyon.

Section of Crystal Spring Formation in Galena Canyon

Top of section is base of Noonday Dolomite; the contact is concealed and may be a thrust fault (see fig. 6).

	Feet
Diabase	450
Chert, dense, greenish	150
Dolomite, weathers light brown; lower 50 ft in beds 6–20 ft thick, upper 125 ft massive. Talcose at base	300
Diabase, much-weathered, calcareous; red iron oxide along fissures	300
Dolomite, well-bedded and thin-bedded; beds 1 in to 2 ft thick; finely granular, grains less than 0.1 mm with vugs and veinlets of coarse dolomite (grains 0.5 mm): quartz grains (about 0.5 mm in diameter) at center of vugs (fig. 8*B*); chert lenses; weathers light brown; talcose where intruded by diabase sills	250
Purple argillite or shale, some brown-weathering dolomitic shale, some quartzite with sericitic groundmass	150
Quartzite, crossbedded, beds 1 in to 3 ft thick; in part conglomeratic with pebbles of quartzite, red chert, and gneiss as much as 6 in. in diameter. Well-rounded grains of quartz and microcline and clastic biotite in dolomitic groundmass (fig. 8*A*). Some beds weather white; others are black, especially those with carbonate in matrix	150
Conglomerate, light gray to white but weathers brown; quartz pebbles as much as 1 in. in diameter, but most of the unit is grit rather than conglomerate. Beds 6 in. to 2 ft thick	100
Total thickness of Crystal Spring Formation including 750 ft of diabase	1,850

Base. Angular unconformity. The basal conglomerate overlies steep to vertical foliation in Precambrian schist.

Table 3 shows the trace elements found in several different rock types in the Crystal Spring Formation. These samples are from Galena Canyon and are to be compared with the analysis of the schist from there (table 2). The quartzite and shale are most like the schist, but they contain substantially less of most trace elements, especially nickel, cobalt, vanadium, titanium, gallium, and chromium.

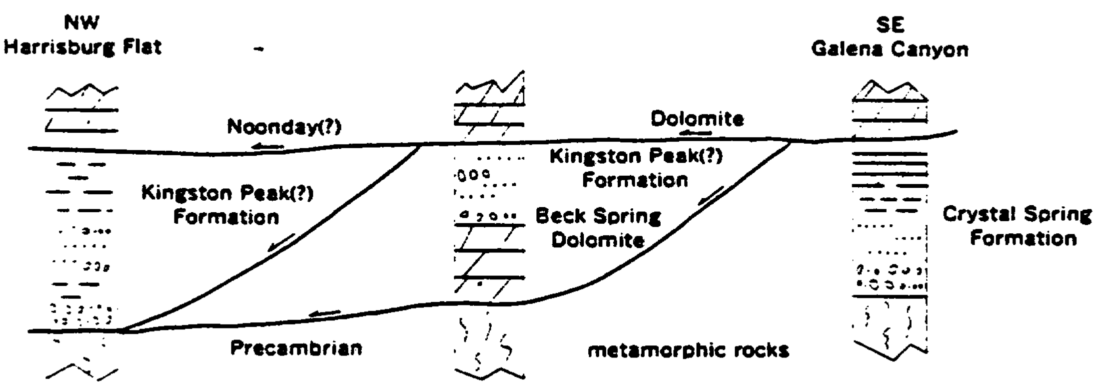


FIGURE 6.—Idealized diagram illustrating distribution and possible structural relationship of the three formations of the Pahrump Series in the Panamint Range.

FIGURE 7.—Crystal Spring Formation in Galena Canyon, view northwest. Q, quartzite member; sd, purple shale and thin-bedded dolomite; di, diabase sill with talcose beds (T) where the sill is in contact with dolomite; do, massive dolomite at top of the formation. In distance is Noonday Dolomite (pCn) capped by Johnnie Formation (pCj).

The Crystal Spring Formation is of much interest economically because it contains highly productive deposits of talc. The commercial product, which consists of the mineral talc [$H_2Mg_3(SiO_3)_4$] and tremolite [$CaMg_3(SiO_3)_4$] with accessory serpentine and calcite (Wright and others, 1954), occurs as an alteration product of dolomite where the dolomite is intruded by diabase (fig. 7).

The fact that diabase sills do not occur in formations younger than the Crystal Spring, even where younger formations of the Pahrump Series are present, has led to the interpretation that the sills and the alteration associated with them predate deposition of the Beck Spring Dolomite (Wright, 1952, p. 15).

BECK SPRING DOLOMITE

Rocks probably equivalent to the Beck Spring Dolomite, locally referred to as the Marvel Dolomitic Limestone by Murphy (1930), crop out at a dozen places along the west side of the Panamint Range west of the mapped area between the Kingston Peak Formation and underlying basement rocks (Murphy, 1932, p. 343).

It is a bluish-gray cherty rock containing about 30 percent CaO and 20 percent MgO, it also contains tremolite and muscovite (Murphy, 1932, p. 343). The formation is very much sheared and contorted—too much so to determine the thickness. The Crystal Spring Formation is absent in that area. An interpretation of the structural relationships is illustrated diagrammatically on figure 6.

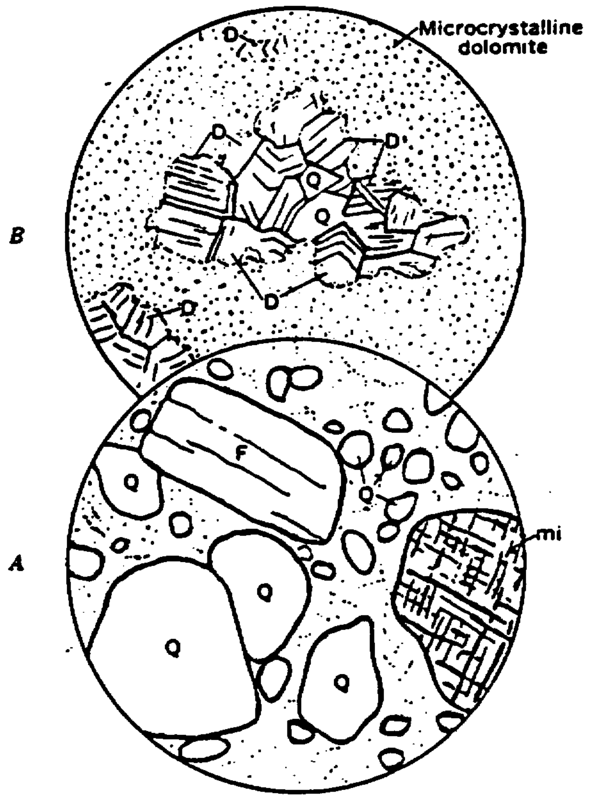


FIGURE 5.—Micrographs of quartzite (A) and dolomite (B) in Crystal Spring Formation. The quartzite consists of well-rounded grains of quartz (Q) and microcline (M) and other feldspar (F) in a dolomitic and sericitic matrix. The dolomite is finely granular with vugs of coarse dolomite (D) at the center of which are grains of quartz (Q). Diameter of field, 2.5 mm.

TABLE 3.—*Trace elements in the Crystal Spring Formation*

[Semiquantitative spectrographic analyses by Uteana Oda, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Quartzite	Dolomite	Diabase	Dark chert in diabase	Red-Brown chert in diabase	Talc	Shale
Pb	30	20	70	15	1,000	10	10
Mn	500	500	1,000	300	20	50	500
Cu	5	5	150	15	70	1	2
Zr	100	20	100	100	150	10	30
Ni	5	<5	50	30	20	<5	5
Co	<10	<10	50	10	15	<10	<10
V	10	15	200	200	150	20	20
Y	15	10	15	15	30	<10	10
Be	1	<1	<1	<1	<1	<1	2
Ti	300	100	20,000	5,000	7,000	200	700
B	20	<10	50	10	10	10	30
Ga	10	<10	50	50	50	<5	5
Cr	10	5	70	100	100	5	<5
Ba	700	15	500	3,000	3,000	15	300
Sr	30	70	200	200	500	20	700
Mg	0.7	>5	3	3	2	>5	>5

NOTE.—Also found: La, 30 or less; Zn, 200; Sc, 15 in diabase, 10 or less in others; Mo, <2; Bi, <1; Sn, <10; As, <500; Sb, <30; W, <20; Nb, <10.

KINGSTON PEAK(?) FORMATION

The youngest formation of the Pahrump Series, the Kingston Peak Formation, crops out extensively along the west side of the Panamint Range. Similar rocks tentatively assigned to this formation occur at Harrisburg Flat, Tucki Mountain, and the northern part of the Funeral Mountains.

Along the west side of the Panamint Range, the Kingston Peak Formation has been divided into three members (Johnson, 1957, p. 360). The lower member, 370–1,600 feet thick, is conglomeratic graywacke; above this is limestone 30–170 feet thick, and at the top is conglomerate, sandstone, and shale 260–1,000 feet thick (Johnson, 1957, p. 360–361).

In the northern part of the Panamint Range, in the vicinity of Harrisburg Flat, the rocks mapped as Kingston Peak(?) Formation are intruded by the granite at Skidoo in addition to being much faulted and folded. The stratigraphy there has not been determined satisfactorily. The most distinctive rocks are the stretched-pebble conglomerates. Some of these have a quartzite matrix, others have a sandy dolomitic matrix. The clasts are of quartzite and limestone. Much of the formation is platy quartzite, and there is some limy dolomite and limestone. In places the upper part is dark shale. Overlying these dominantly clastic beds is a thick section of carbonate rocks mapped as Noonday(?) Dolomite, and probably separated from the Kingston Peak(?) Formation by a flat fault. On Tucki Mountain some carbonate rocks below the thrust fault are doubtfully included in the Kingston Peak(?) Formation. Beds assigned to the Kingston Peak(?) Formation are at least 3,000 feet thick.

Six samples of the Kingston Peak(?) Formation were collected for spectrographic analysis. Three were obtained from the vicinity of Harrisburg Flat in the Panamint Range and three from the base of the Funeral Mountains. The analyses are given in table 4. The trace elements in the various rock types are quite different in the two areas. The limestone in the Panamint Range samples contains a greater concentration of trace elements than does the limestone from the Funeral Mountains. The quartzites differ less and their differences are more spotty. The differences, despite the small number of samples, cast further doubt on the correlation of the Kingston Peak(?) Formation between the Panamint Range and the Funeral Mountains.

TABLE 4.—*Trace elements in the Kingston Peak(?) Formation.*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Funeral Mountains			Panamint Range		
	Dolomite	Limestone	Quartzite	Limestone	Shale	Quartzite
Pb	15	20	300	10	30	50
Mn	1,500	200	500	1,000	2,000	700
Cu	2	10	150	100	150	200
Zr	30	10	200	500	700	700
Ni	5	7	50	70	100	10
Co	<10	<10	20	15	30	10
V	10	20	100	200	200	70
Y	10	20	20	50	50	70
Be	<1	<1	1	1	1	1
Ti	300	300	10,000	10,000	>10,000	7,000
B	10	<10	20	15	200	15
Sc	<10	<10	<10	20	50	<10
Ga	<5	<20	20	<20	<20	<20
La	<50	<50	<50	50	50	50
Cr	7	10	50	100	500	30
Ba	100	50	1,000	1,000	1,500	3,000
Sr	1,000	200	100	700	200	150
Mg	>5	>5	1.5	5	5	1

NOTE.—Also found: Mo 10 in shale in Panamint Range; <5; in other samples; Sn, <10; Ag, <1; Ge, <20; As, <1,000; Sb, <200, In, <50; Cd, <50; Tl, <100; Ta, <30; W, <50.

NOONDAY DOLOMITE

Overlying the Pahrump Series is the Noonday Dolomite. At its type locality at the south end of the Nopah Range, about 50 miles east of the Panamint Range, the formation consists of about 1,500 feet of light-cream-colored dolomite with sandy beds near the top (Hazzard, 1937b, p. 300).

In the southern part of the Panamint Range, in Galena and Six Spring Canyons, the Noonday Dolomite rests on the Crystal Spring Formation of the Pahrump Series. The contact between the 2 formations, which is concealed by rubble from the dolomite, probably is a flat fault that removed the upper 3,500 feet of the Pahrump Series and that cuts across at least 600 feet of the upper part of the Crystal Spring Formation (fig. 6). The Noonday Dolomite is conformably overlain by the Johnnie Formation.

The Noonday Dolomite in Galena and Six Spring Canyons consists of a light-cream-colored lower member about 500 feet thick and a gray upper member about 300 feet thick. Indistinct bedding in both members is characteristic of the formation throughout the region. Structures suggestive of *Scolithus* tubes were found in the dolomite in the fault block forming the foot of the mountain at the east tip of the spur south of Galena Canyon (loc. F-89, NE¼NE¼ sec. 17, T. 22 N., R. 1 E.; fig. 9). Similar structures are present in the Noonday Dolomite in the southern Nopah Range.

The lower member of the Noonday Dolomite is a granular recrystallized fine-grained dolomite mottled with spots of coarse-grained dolomite in euhedral grains (fig. 10*A*). The spots of coarse-grained dolomite have quartz, evidently secondary, at their centers. Petrographically this rock resembles the thick-bedded dolomite in the underlying Crystal Spring Formation (fig. 8).

The upper gray member of the Noonday Dolomite is coarser grained than the lower member; the grains are uniform in size, about 0.3 mm in diameter (fig. 10*B*). Through the rock are numerous grains of euhedral magnetite partially altered to hematite.

In this area, as elsewhere in the region, the Noonday Dolomite contains small irregular deposits of lead minerals; these have been prospected at numerous places along the mountaintop south and west of Galena Canyon (table 5). So far as known, none of the deposits has been productive.

Farther north in the Panamint Range, from Johnson Canyon northward to Tucki Mountain, a series of carbonate rocks 1,000 feet or more thick, lying below the Johnnie Formation, is assigned questionably to the Noonday Dolomite. The formation seems to be at the stratigraphic position of the Noonday, but it is enough different lithologically to warrant further study to confirm or correct the identification. The formation differs from the Noonday farther south in containing a great deal of limestone as well as dolomite and in having the dominant colors tan and white. In addition, there are beds of limestone conglomerate in which clasts of limestone 1–3 inches in diameter are contained in sandy dolomitic matrix.

A highly generalized section on the west side of Rogers Peak showed the following:

Feet	
Top	Highly contorted thin-bedded limestone.
10	White quartzite.
100	White limestone.
250	Banded light and dark limestone.
250	Light-brown limestone.
250	Limy sandstone.
200	Dark quartzite.
250	Banded light and dark limestone.
Base.	

The Noonday(?) Dolomite seems to be equivalent to the series of dolomitic limestone on the west side of the range that Murphy (1932, p. 349) referred to as the Sentinel Dolomite (base), Radcliff Formation, and Redlands Dolomitic Limestone (top). Johnson (1957, p. 370) correlated these beds with the Noonday.

Table 5 lists the trace elements in eight samples from the Noonday Dolomite. In the southern part of the

FIGURE 9.—Noonday Dolomite showing structures suggestive of *Scolithus* tubes. Location is at east foot of the mountain at the spur south of Galena Canyon.

Panamint Range the dolomite is highly mineralized, notably with lead and zinc. The sample of light facies differs little from the dark one. A much larger number of samples would be needed to suggest whether the Noonday(?) Dolomite should be correlated with the Noonday. The trace element content of the Noonday(?) Dolomite is about the same as that of the limestone in the Kingston Peak(?) Formation in the northern part of the Panamint Range.

JOHNNIE FORMATION

The Johnnie Formation at its type locality, near the town of Johnnie about 50 miles east of Death Valley, is mostly shale and is 4,500 feet thick (Nolan, 1929, p. 461).

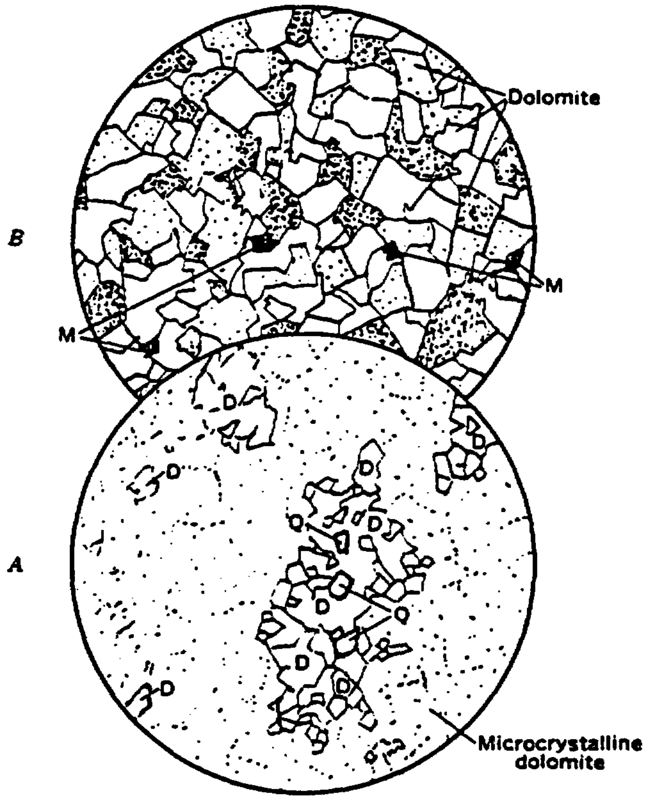


FIGURE 10.—Micrographs of Noonday Dolomite. *A*, Lower member, a fine-grained dolomite mottled with coarse dolomite (D). Quartz grains (Q) occur at the center of some of the areas of coarse dolomite. *B*, Upper member, a medium-grained dolomite in which the grains are of uniform size and contain scattered grains of magnetite (M). Diameter of field, 2.5 mm.

In the Panamint Range, where it has been referred to as the Hanaupah Formation (Murphy, 1932, p. 349), it is mostly shale, and its thickness is more than 4,000 feet.

The Johnnie Formation is conformable on the Noonday Dolomite and is gradational with it, for the lower part of the formation consists of interbedded dolomite and quartzite. The contact is taken at the lowest quartzite, as has been done elsewhere (Hazzard, 1937b, p. 303; Johnson, 1957, p. 373).

In the southern part of the Panamint Range the lower third of the formation consists of interbedded dolomite and sandstone or quartzite. The dolomite is thin bedded and ripple marked. Some of the quartzite is conglomeratic with pebbles as much as 1 inch in diameter.

The middle third of the formation is light-colored shale capped by dolomite; the upper third is purple shale with interbedded quartzite (fig. 11). A composite section of the Johnnie Formation in the southern part of the Panamint Range is as follows:

TABLE 5.—*Trace elements in the Noonday Dolomite*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Southern part of Panamint Range						Northern part of Panamint Range	
	Noonday Dolomite						Noonday(?) Dolomite	
	Light dolomite	Dark dolomite	Prospects					
	A	B	C	D	E	F	G[1]	H[2]
Pb	200	15	15	10,000	3,000	50	20	50
Zn	<200	<200	5,000	3,000	300	700	nd	nd
Mn	150	100	500	200	3,000	100	200	2,000
Cu	7	10	7	7	7	7	10	200
Zr	20	<10	<10	<10	20	<10	10	500
Ni	<5	<3	7	7	7	7	7	100
Co	<10	<10	<10	<10	<10	<10	<10	30
V	10	10	<10	<10	<20	<10	20	200
Y	<10	<10	<10	<10	10	<10	20	50
Be	<1	<1	<1	<1	<1	<1	<1	<1
Ti	100	50	10	15	200	30	300	10,000
B	<10	<10	70	30	10	70	<10	70
Mo	<2	<2	<2	5	7	<2	<3	<5
Ga	<5	<5	<5	<5	<5	<5	<20	<20
Cr	<5	<5	<5	<5	<5	<3	20	500
Ba	70	10	30	20	100	70	100	3,000
Sr	300	50	50	300	<20	<20	3,000	3,000
Mg	>5	>5	5	3	0.2	0.5	5	>5

[1] This sample also contained Ag, 70; Bi, 50; As, 1,500; Sb, >10,000.
[2] This sample also contained Sc, 50.

NOTE.—Also found: Sn, <10; Ag, <1 (except sample G); Ge, <20; As <1,000 (except sample G); Sb, <200 (except sample G); In, <50; Cd, <50; Tl, <100; Ta <50; W, <50; La 50 or less; Sc, <10 (except sample H).

Composite section of Johnnie Formation in Six Spring Canyon and Johnson Canyon

	Feet
Top. Base of Stirling Quartzite.	
Purple shale member. Shale, mostly purple, some red, some green; fissile. Upper 100 ft includes interbedded quartzite and shale; beds transitional with overlying Stirling Quartzite. The purple shale contains fine sand grains 0.1 mm in diameter in sericitic matrix (fig. 12*D*); the quartzite interbedded with the shale near the top of the member is fine grained (grains about 0.2 mm in diameter) with sericitic laminae and in a sericitic clay matrix	500
Yellow member. Olive-brown shale capped by yellow silicified dolomite about 25 ft thick in 1 or 2 beds; dolomite contains thin beds of quartzite. The dolomite is very fine grained (0.01 mm) with vugs of coarser dolomite (0.1 mm) (fig. 12*C*). The shale is sandy or silty with quartz grains in sericitic laminae (fig. 12*B*)	500
Dolomite member. Basal 200 ft consists of thin-bedded dolomite interbedded with ripple-marked sandstone, overlain by 35 ft of brown-weathering dolomite; upper 150 ft is well-bedded and thin-bedded quartzite and sandstone; at top is 5 ft of pebble conglomerate with pebbles as much as 1 in. in diameter. The dolomite is like that in the overlying yellow member but contains scattered quartz grains about 0.1 mm in diameter. The quartzite has rounded quartz grains with irregular sides due to recrystallization, and there is considerable associated microcline; the qartz shows strain shadows; there is very little matrix around the grains (fig. 12*A*)	400
Total thickness of Johnnie Formation	1,400

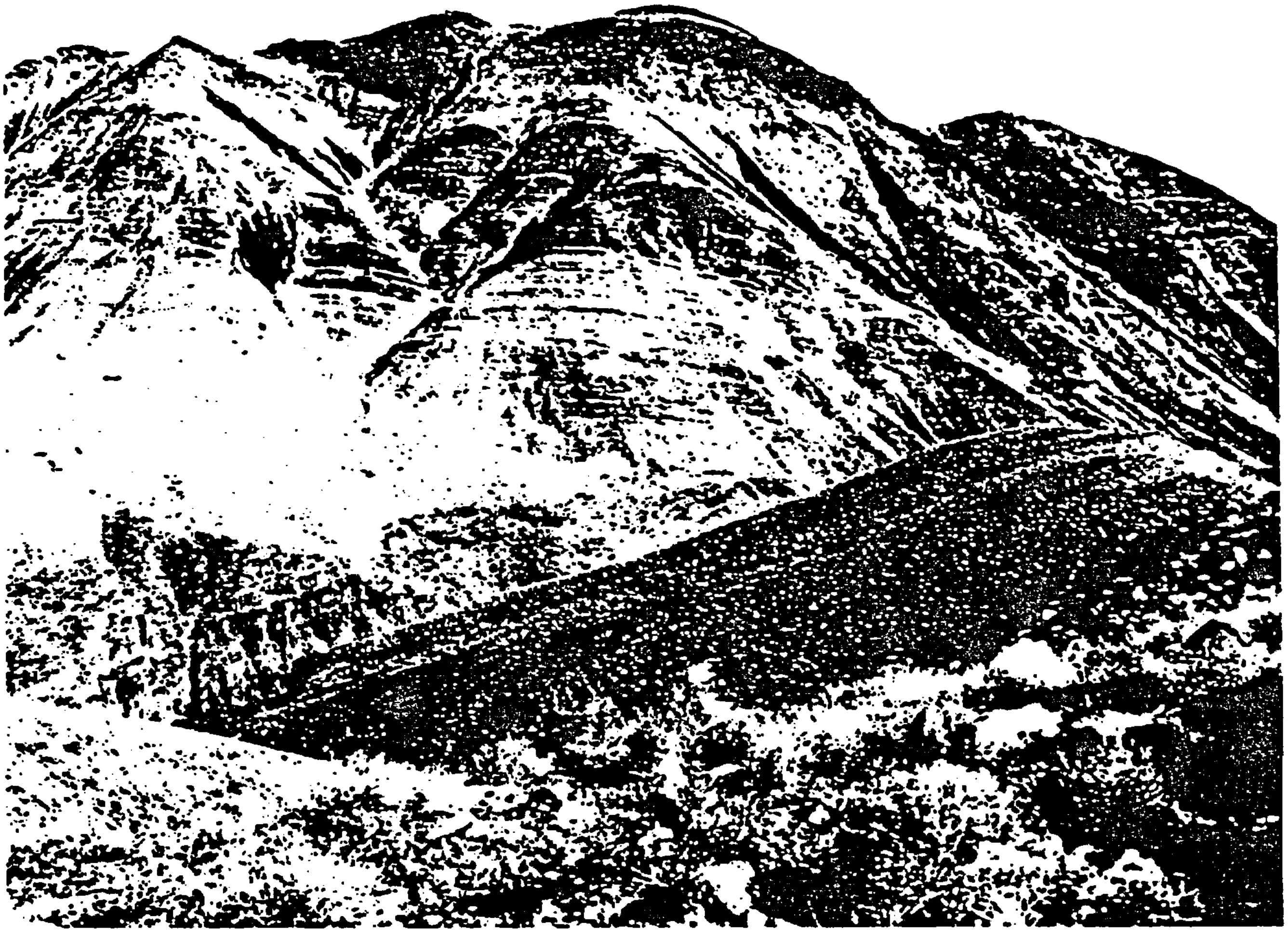


FIGURE 11.—Johnnie Formation on the north side of Six Spring Canyon. The hilltops are capped by Stirling Quartzite. Dark beds forming the upper half of the hillside are the purple shale member; light beds in the middle and lower half are the shale member capped by dolomite.

Northward along the Panamint Range the Johnnie Formation thickens and the lithology changes. In Hanaupah Canyon and farther north the Johnnie Formation is mostly shale. A section measured along the main (south) fork of Hanaupah Canyon is 2,300 feet thick, as follows:

Section of Johnnie Formation in Hanaupah Canyon, from Narrows at west edge of Bennetts Well quadrangle to mine workings at end of road

	Feet
Top. Base of Stirling Quartzite.	
1. Quartzite, thin-bedded; some argillite	100
2. Limestone, interbedded with banded purple argillite	750
3. Argillite, purple; upper part banded	550
4. Argillite, weathers tan, greenish on fresh surfaces; contorted bedding	175
5. Interbedded argillite, sandstone, and a few beds of thin-bedded, laminated, and highly micaceous dolomite; beds as much as 1 ft thick	700

Section of Johnnie Formation in Hanaupah Canyon, from Narrows at west edge of Bennetts Well quadrangle to mine workings at end of road—Continued

	Feet
6. Interbedded dolomite and argillite; argillite greenish, weathers light tan; dolomite beds 2 in to 1 ft thick; this is a transition zone between units 5 and the Noonday(?) Dolomite	50
Total thickness of Johnnie Formation	2,325
Base. Top of Noonday(?) Dolomite, white or light-cream color; much altered; no original structures left; 50 ft+ exposed.	

In the north fork of Hanaupah Canyon (Chuckwalla Canyon on the maps), there is a light-brown dolomite, about 200 feet thick, forming the top of the Johnnie. Below this is 50 feet of ripple-marked purple shale, and below this, 250 feet of fissile shale. The main part of the formation is argillite, as it is farther south.

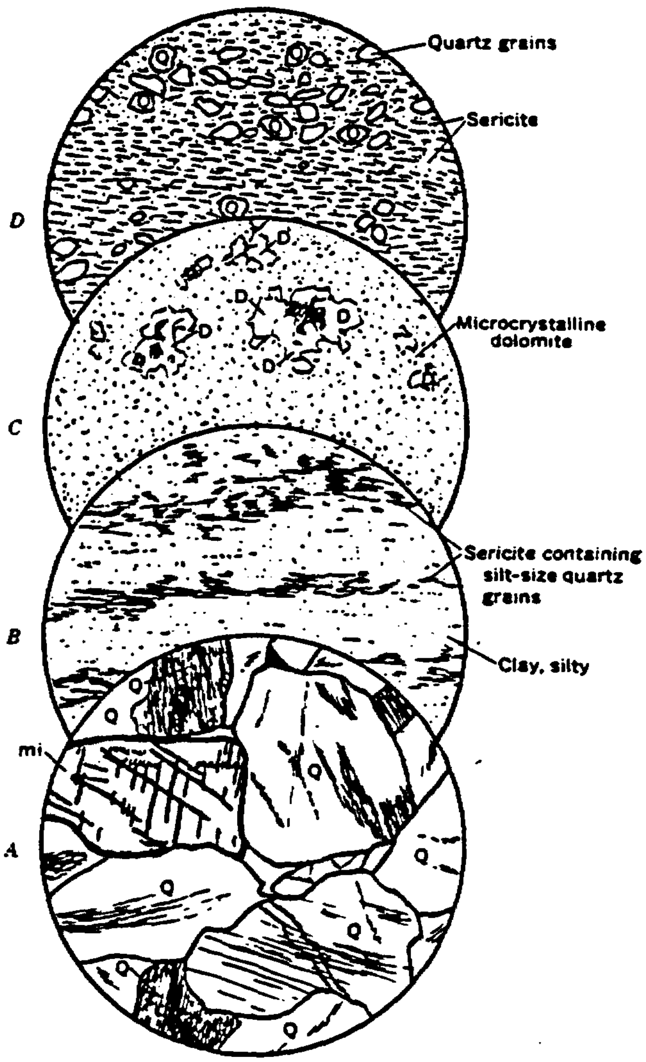


FIGURE 12.—Micrographs of rock types from the Johnnie Formation. *A*, Quartzite from basal dolomitic member (Q, quartz with strain shadows; mi, microcline). Very little matrix between the grains. *B*, Shale from yellow shale member. *C*, Dolomite from top of the yellow shale member, mostly very fine grained dolomite but mottled with vugs filled with coarser dolomite (D). *D*, Sandy sericitic purple shale. Diameter of field, 2.5 mm.

In Death Valley Canyon, dolomite at the top of the Johnnie Formation is white, but probably this color change is due to hydrothermal alteration. Light-colored dolomite was not seen elsewhere in the Johnnie Formation.

At the head of Trail Canyon the Johnnie Formation is about 4,000 feet thick. The beds are in a much-faulted and steeply dipping monocline, and the stratigraphy there is uncertain.

The following is an approximate section:

Feet
Top____Stirling Quartzite overlain by Wood Canyon Formation.
900____Dark shale.
350____Brown quartzite like the Stirling, and perhaps this is the Stirling repeated by faulting.
900____Dark shale.
900____Light-colored shale, green and tan.
350____Dark shale.
500____Brown shale.
Base__Noonday(?) Dolomite.

In the northern part of the Panamint Range the formation is mostly shale without evident marker beds of dolomite or coarse clastics. The thickness is uncertain on account of faulting and other deformation, but probably it is 4,000 feet or more.

Table 6 gives the results of analyses of trace elements in 18 samples from various rock types in the Johnnie Formation. Dolomite in the Johnnie Formation has about the same trace elements as does the Noonday Dolomite, but it averages more zirconium. The shale has about the same trace elements as does the shale in the Kingston Peak(?) Formation in the northern part of the Panamint Range (table 4).

STIRLING QUARTZITE

At the mouth of Johnson Canyon and along both sides of Starvation Canyon, the Stirling Quartzite consists of three members. At the base is 700 feet of reddish-brown-weathering quartzite that is in part conglomeratic. About this is 500 feet of purple shale with thin beds of quartzite; at the top is brown-weathering quartzite 800 feet thick. This section is very similar to that in the Nopah Range (Hazzard, 1937b, p. 306–307).

The quartzites are vitreous; most of the beds are coarse grained. Pebbles in the conglomeratic layers are as much as 1 inch in diameter, but most are smaller, ¼–½ inch in diameter. The pebbles are mostly white quartz; a few are red jasper. There has been much recrystallization and quartz veining, and the beds are firmly indurated. Smooth surfaces show fine layering due to size sorting of the sediments; individual beds are cross-bedded (fig. 13) and many are ripple marked. Figure 14 is a micrograph of a thin section of the quartzite.

Individual grains are subround and encased in secondary quartz. Both the original grain and the quartz deposited around it show strain shadows. Associated with the quartz is coarsely twinned microcline and some sericite. The rock is a true quartzite and breaks across the quartz grains.

The purple shale separating the upper and lower quartzite members of the Stirling consists of fine-grained well-rounded quartz in laminae of sericite; it is similar to that in the Johnnie Formation (fig. 12).

TABLE 6.—*Trace elements in the Johnnie Formation*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Shale or schist								Chaos	Conglomerate		Dolomitic sand	Dolomitic shale		Dolomite			
	Dark					Green	Yellow											
Pb	50	10	20	30	30	200	1,500	30	20	30	300	20	30	500	10	20	150	70
Mn	200	2,000	700	200	700	300	100	70	700	100	50	7,000	5,000	200	2,000	100	3,000	150
Cu	150	100	100	30	30	500	100	70	100	300	20	100	150	150	150	5	30	7
Zr	100	1,500	700	500	200	100	200	200	700	150	700	10	700	300	100	10	70	10
Ni	50	100	200	70	20	50	50	30	20	5	5	5	150	50	10	5	<5	<5
Co	10	30	100	10	<10	15	10	<10	<10	<10	<10	10	20	15	<10	<10	<10	<10
V	100	500	300	150	100	50	70	100	500	20	30	20	200	100	20	10	10	10
Y	70	150	150	30	20	20	20	20	150	<10	10	10	100	20	10	<10	<10	<10
Be	1	1	2	2	3	2	3	2	1	<1	<1	<1	<1	<3	<1	<1	<1	<1
Ti	10,000	>10,000	>10,000	7,000	7,000	2,000	5,000	5,000	>10,000	2,000	10,000	200	10,000	3,000	300	70	3,000	200
B	300	150	200	70	70	30	70	70	30	20	15	<10	150	20	<10	<10	<10	<10
Sc	30	50	70	20	15	10	15	15	50	<10	<10	<10	50	10	<10	<10	<10	<10
Ga	<20	<20	<20	50	50	20	50	50	<20	5	<5	<20	<20	20	<20	<5	<5	<5
Cr	200	300	700	150	70	50	70	100	700	20	30	20	500	70	20	5	5	5
Ba	700	2,000	5,000	700	500	300	500	500	7,000	300	500	50	2,000	500	100	15	50	70
Sr	150	100	70	20	30	20	20	30	1,000	20	<20	200	2,000	<20	100	500	50	70
Mg	1.5	>5	>5	1	2	1.5	1	1	3	0.2	0.5	>5	>5	0.5	>5	0.5	5	>5

NOTE.—Also found: Zn, <200; La, <50; Mo, <2; Ag, <1; Bi, <5; Sn, <10; As, <500; Sb, <50; W, <20; Nb, <10.

In places, there are some thin beds of shale and dolomite about 30–50 feet below the top of the formation. These beds are like those in the lower part of the Wood Canyon Formation.

FIGURE 13.—Detail of bedding in Stirling Quartzite at mouth of Johnson Canyon. Beds are 2–12 inches thick.

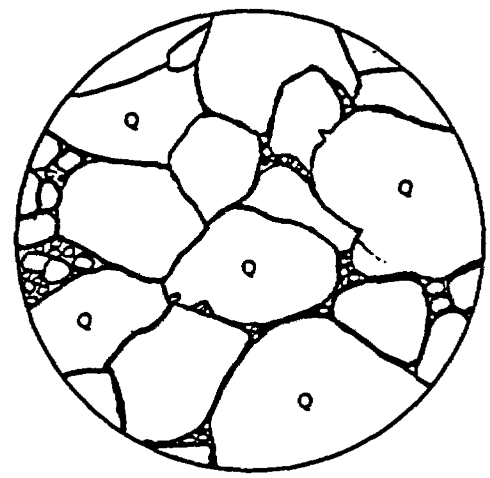


FIGURE 14.—Micrograph of Stirling Quartzite. The grains of quartz (Q) are rounded but also irregularly intergrown. Strain shadows conspicuous under cross nicols. Between the large grains is secondary quartz. In most beds sericitized feldspar and some muscovite occurs with the quartz grains. Diameter of field, 2.5 mm.

Trace elements in 9 samples of the Stirling Quartzite are given in table 7. Trace elements in the quartzite and shale facies are like those in the conglomerate and shale, respectively, of the Johnnie Formation (table 6). The quartzite differs from that of the Kingston Peak(?) Formation in the Panamint Range (table 4) in having less copper, barium, and strontium. The quartz veins in the Stirling contain trace elements only in small amounts compared to the parent quartzite.

In the Death Valley area the Stirling Quartzite has a maximum thickness of about 2,000 feet. At the type locality in the Spring Mountains (Nolan, 1929, p. 463) it is 3,700 feet thick. The boundaries of the formation, as shown on the geologic map of the Panamint Range (pl. 1), imply considerable variation in thickness of the

TABLE 7.—*Trace elements in the Stirling Quartzite*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Quartzite						Shale bed	Quartz vein	
								Clear	White
	A	B	C	D	E[1]	F	G	H	I
Pb	300	20	100	10	10	10	10	20	10
Mn	80	20	300	700	80	70	100	15	50
Cu	2	3	30	150	100	50	100	2	5
Zr	500	100	150	200	2,000	300	1,500	10	100
Ni	15	6	10	20	10	20	100	5	5
Co	<10	<10	<10	10	<10	<10	15	<10	<10
V	15	10	15	20	500	20	200	<10	10
Y	20	<10	10	10	150	50	100	<10	<10
Be	1	<1	<1	<1	1	<1	1.5	<1	<1
Ti	1,300	300	500	1,500	>10,000	7,000	>10,000	30	500
B	20	10	20	15	200	150	200	10	20
Sc	<10	<10	<10	<10	70	<10	50	<10	<10
Ga	10	<5	<5	<20	<20	<20	<20	<5	<5
Cr	30	5	15	10	300	30	200	<5	5
Ba	700	100	800	300	1,500	1,000	2,000	30	100
Sr	70	20	50	20	100	50	50	<20	<20
Mg	0.5	0.2	0.2	0.15	1	1	2	0.02	0.1

[1] La, 100; Mo, 15; sample E is from the base of the upper plate of the Amargosa thrust fault and has been hydrothermally altered.

NOTE.—Also found La,<50 (except sample E); Mo <2 (except sample E) Ag,<1; Bi,<5; Sn,<10; As,<500; Sb,<50; W,<20; Nb,<10.

quartzite; but this variation is attributable to lack of consistency in picking the boundary between the Stirling and overlying Wood Canyon Formation, the basal part of which also is quartzitic. It is difficult to distinguish the Stirling Quartzite from the thin-bedded quartzites that are included in the lower part of the Wood Canyon Formation, especially where there is granulation of the quartzite along faults.

The Stirling Quartzite coincides with a surface of flat faulting at its type locality (Nolan, 1929, p. 463, 470), and in the Funeral Mountains and Panamint Range. In the Funeral Mountains the flat fault is exposed at Echo Mountain and from Hells Gate to Daylight Pass. In the Panamint Range the Stirling Quartzite forms the base of the upper plate of the fault at the mouth of Mosaic Canyon and the top of the lower plate in Tucki Wash along the south and east sides of the Tucki Mountain thrust. As a result of this and other faulting, the Stirling Quartzite locally is absent, but such absence is attributable to deformation and not to stratigraphic thinning. Similarly, no stratigraphic significance can be placed on the variations in thickness that can be observed in short distances within this area. The thinning from the type area to the Nopah Range and westward to the Panamint Range may be real, because this thinning is accompanied by an increase in shale and thinner bedding westward in the formation.

CAMBRIAN SYSTEM

About 8,500 feet of beds representing all parts of the Cambrian System are present in the Death Valley region. These include the Wood Canyon Formation of Early Cambrian and Early Cambrian(?) age, the Zabriskie Quartzite of Early Cambrian age, the Carrara Formation of Early and Middle Cambrian age, the Bonanza King Formation of Middle and Late Cambrian age, and the Nopah Formation of Late Cambrian age.

The oldest beds in the Death Valley region containing animal remains are in the Wood Canyon Formation. The top of the Cambrian System lies somewhere near the top of the Nopah Formation, but it cannot be located more precisely because of the lack of fossils in this part of the section. For mapping, the upper and lower boundaries of the Cambrian System are placed at the formation boundaries.

WOOD CANYON FORMATION

The Wood Canyon Formation of Early Cambrian and Early Cambrian(?) age conformably overlies the Stirling Quartzite of Precambrian age, both at the type locality in the Spring Mountains (Nolan, 1929, p. 463) and in the mountains adjoining Death Valley. The top of the Wood Canyon in Death Valley is taken at the base of the Zabriskie Quartzite. At the type locality the top of the Wood Canyon Formation is at the base of a 20-foot bed of white quartzite which probably is equivalent to the Zabriskie (Hazzard, 1937b, p. 313). In the Nopah Range the Zabriskie Quartzite has been included as a member in the Wood Canyon Formation together with 630 feet of overlying beds (Hazzard, 1937b, p. 310). In this report the beds overlying the Zabriskie Quartzite are treated separately as part of the Carrara Formation.

The basal unit of the Wood Canyon Formation, about 465 feet thick, consists mostly of thin-bedded quartzite, but contains considerable shale and dolomite. Above this is 1,200 feet of quartzite in thicker beds. No fossils other than possible *Scolithus* tubes and possible algal structures have been found in the lower part of the formation in Death Valley or in the surrounding regions.

The upper member of the Wood Canyon Formation, about 900 feet thick, consists of shaly and dolomitic beds as well as thin beds of siltstone and quartzite (fig. 15). Fossils found in this member include fragmentary trilobites representing *Nevadella gracile* (Walcott), indeterminate molds of brachiopods, and molds of cystid plates in a unit of thin sandstones that generally lies just below a zone of oolitic and pelmatazoan dolomites and limestones. Thin sections of some of these pelmatazoan limestones in Death Valley and at Daylight Pass show the presence of fragmentary archaeocyathids. The association of sandstones with fragmentary trilobites and brachiopods and oolitic and pelmatazoan carbonates characterizes the upper part of the Wood Canyon Formation throughout the Death Valley region and as far north as the Groom district in Nevada (Palmer, oral commun., 1961).

FIGURE 15.—Detail of interbedded shale, quartzite, and dolomite in the upper part of the Wood Canyon Formation in the ridge along the north side of Blackwater Wash. The thick bed in the upper right is dolomite; below this is quartzite and shale.

Collections from these beds in Death Valley were studied by A. R. Palmer, who has reported on them as follows (locations are indicated on the geologic map):

F-31. Above Quartzite Spring, north side of Starvation Canyon, Bennetts Well quad. (NW¼NW¼ sec. 12, T. 21 S., 46 E.) "Olenellid scraps; certainly Early Cambrian age, but species not determinable."

F-64 (3104-CO). 1.5 miles south of Panamint Burro Spring; same general fault block as F-31, Bennetts Well quad. (2,000 ft north of NW cor. sec. 1, T. 21 S., R. 46 E.), estimated 500 ft below the Zabriskie Quartzite. "*Nevadella gracile*? (Walcott)."

F-36. Southeast side Hanaupah Canyon, alt 3,100 ft, 2 miles above mouth of Canyon. Bennetts Well quad. "Olenellid scraps, certainly Early Cambrian in age, but species not identifiable."

F-29 (2453-CO). East base of west butte of the Death Valley Buttes, Stovepipe Wells quad. (2,500 ft northeast of NE cor. sec. 36, T. 14 S., R. 45 E.). Pelmatazoan calcarenite. Indeterminate archaeocyathid.

F-50. Base of Zabriskie Quartzite in Blackwater Wash, top of hill 951, west side Furnace Creek quad. "*Kutorgina*? sp."

A section of the Wood Canyon Formation, measured along the north side of Blackwater Wash (Furnace Creek quad.) follows:

Section of Wood Canyon Formation along north side of Blackwater Wash

[Measured by Charles B. Hunt, A. R. Palmer, and R. J. Ross, Jr.]

	Feet
Top. Base of Zabriskie Quartzite.	
1. Brown-weathering dolomite and quartzite, some greenish shale. Dolomite and quartzite beds 1–10 ft thick; shale beds less than 1 ft thick (fig 15). Dolomite crossbedded, in part strikingly oolitic. Tubes suggestive of *Scolithus* tubes about 200 ft above the base	400
2. Shaly member. Lower 70 ft mostly green siltstone interbedded with dark-weathering quartzite in beds 5 ft thick. Overlying this is 210 ft of siltstone that is reddish along shear zones but greenish away from them. Above this is 105 ft of dark-weathering quartzite; 35 ft of green shale and siltstone, and, at the top, 100 ft of greenish micaceous fine-grained quartzite and siltstone. Some tubes suggestive of *Scolithus* in the uppermost unit	520
3. Quartzite member. Beds 1–3 ft thick; light gray on fresh fracture but weathers dark. Micaceous. Numerous gritty beds; some conglomeratic with pebbles as much as ¾ in. in diameter; most of these are milky quartz; some are red jasper. Lower 200 ft includes much grit and numerous shale beds about 1 ft thick; purple and green; increasing amount of shale downward. Two hundred feet above base is 50-ft bed of grit and conglomerate. Upper 700 ft is mostly fine-grained quartzite. This unit of section crossed by some faults and thickness is uncertain, estimate	1,200
4. Quartzite with interbedded shale, fine-grained, thin-bedded, transitional between units 3 and 5	120
5. Quartzite with interbedded siltstone and some fine-grained thin-bedded shale; a few thin beds of brown-weathering dolomite, light brown on fresh surfaces. Eighty feet above base is 10-ft bed of gray dolomite overlain by 15 ft of light-tan thin-bedded dolomite. Quartzite is light brown, weathers dark brown; micaceous. Twenty feet below top is a thin bed of dolomite having floating sand grains and pebbles of carbonate rock; intraformational conglomerate	150
6. Shale and quartzite. Quartzite beds are 1–12 in. thick, finely laminated, micaceous, light brown on fresh surfaces, reddish brown on weathered surfaces. Shale is olive green on weathered surface; also micaceous	130
7. Shale, sandstone, and dolomite. Shale and sandstone green and brown; dolomite dark blue on fresh surface; weathers brown. Dolomite beds 1–2 ft thick	65
Total thickness of Wood Canyon Formation	2,585
Base. Top of Stirling Quartzite. Contact gradational and taken at top of highest massive light-colored quartzite. There are thin beds of shale and dolomite 30–50 ft below the contact.	

Thin sections of the quartzite and shale beds of the Wood Canyon Formation are illustrated on figure 16. The shale contains muscovite (or sericite) and magnetite in addition to minute quartz grains that occur both scattered and in layers. The quartzite has interlocking

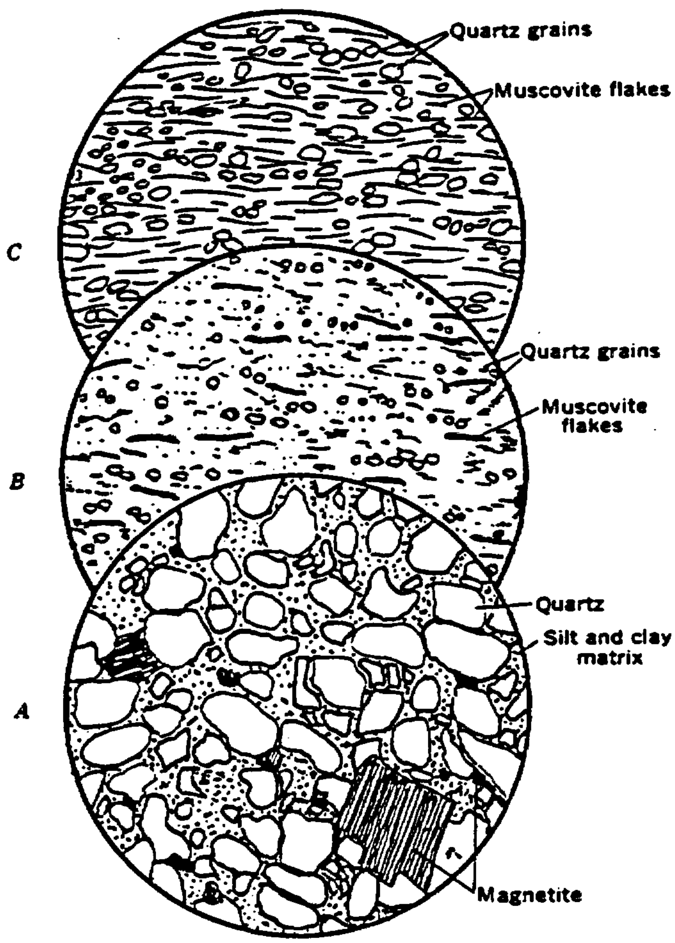


FIGURE 16.—Micrographs of rock types in the Wood Canyon Formation. *A*, Quartzite from basal member. Other quartzite has much less interstitial material and resembles the Stirling Quartzite. *B*, Shale composed largely of silt and clay. *C*, Shale composed largely of muscovite and sand grains. Diameter of field, 2.5 mm.

TABLE 8.—*Trace elements in the Wood Canyon Formation.*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U. S. Geol. Survey. Values in parts per million, except Mg, which is given in percent.]

Element	Wood Canyon Formation										Dolomitic quartzite
	Limestone	Shale			Sandy shale			Quartzite			
Pb	20	150	20	20	10	15	100	70	50	10	10
Mn	10,000	2,000	15	7,000	2,000	150	1,000	1,000	150	1,500	<10,000
Cu	3	20	3	30	30	200	30	5	5	20	50
Zr	30	2,000	70	100	30	200	150	300	100	200	15
Ni	7	30	<5	50	20	70	7	5	7	5	<5
Co	<10	10	<10	10	10	20	<10	<10	<10	<10	<10
V	15	30	<10	10	10	100	30	20	50	15	10
Y	10	70	<10	30	20	30	10	30	10	50	20
Be	<1	1	3	1	1	1	2	1	<1	<1	<1
Ti	1,500	7,000	300	1,500	1,000	7,000	1,500	1,500	1,500	2,000	700
B	10	50	20	15	15	30	70	20	20	20	<10
Sc	<10	15	10	<10	10	<10	<10	<10	<10	10	<10
Ga	7	10	10	5	10	30	20	5	10	5	<20
Cr	30	70	5	10	15	100	20	20	10	20	10
Ba	150	150	500	200	100	500	1,000	700	500	700	100
Sr	100	100	700	50	200	50	1,500	70	20	50	500
Mg	5	3	0.2	0.7	0.7	2	0.5	0.2	0.5	0.3	5

NOTE.—Also found: La, 50; Mo, 2; Ag, 1; Bi, 5; Sn, 10; As, 500; Sb, 50; W, 20; Nb, 10.

grains of quartz with associated magnetite and traces of mica.

Trace element concentrations in quartzite in the Wood Canyon Formation (table 8) are about the same as in the Stirling Quartzite (table 7), except that the Wood Canyon quartzite averages higher in manganese. A comparison of shales in the Wood Canyon Formation and in the Johnnie shows that the Wood Canyon contains more manganese and less vanadium, boron, gallium, and barium than does the Johnnie. The proportions of trace elements in the limestone are quite different from the proportions in dolomite in the Johnnie Formation (table 6).

ZABRISKIE QUARTZITE

The Zabriskie Quartzite, originally named and described as a member of the Wood Canyon Formation by Hazzard (1937b, p. 309), consists of white quartzite in laminated beds about 6 inches to 2 feet thick interbedded with micaceous purple shale, sandstone, and siltstone. The quartzite beds show little crossbedding; mostly they are evenly laminated. The quartzite contains few impurities; the rock consists of closely interlocked grains of quartz with little other foreign matter (fig. 17).

The Zabriskie Quartzite, like the Stirling Quartzite, has been subject to major deformation due to shearing along flat faults that approximately parallel the bedding. As a consequence the formation varies greatly in thickness, but the variation is attributable to tectonic deformation—not to stratigraphic changes.

Along the north side of Blackwater Wash, the Zabriskie Quartzite is 160 feet thick and is mostly massive

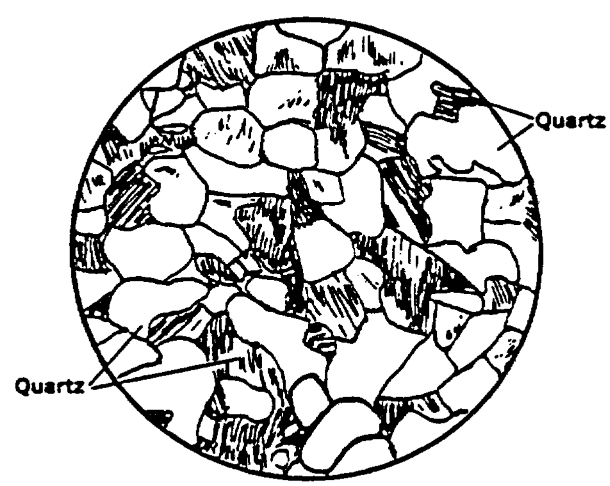


FIGURE 17.—Micrograph of the Zabriskie Quartzite. The rock consists of closely interlocked grains of quartz with almost no interstitial material. Quartz grains are rounded but show signs of irregular intergrowth. Diameter of field, 2.5 mm.

brecciated quartzite stained lavender. Fifty-five feet below the top is a 2-foot layer of mustard-colored shale and siltstone.

A thickness of 70 feet has been reported for the Zabriskie Quartzite at Aguereberry Point; gray quartzite just below the base of the Zabriskie contains rodlike structures several inches long oriented perpendicular to the bedding, possibly *Scolithus* tubes (Hopper, 1947, p. 406).

In the Funeral Mountains, north of Echo Canyon and at the east edge of the Furnace Creek quadrangle, the Zabriskie Quartzite is several hundred feet thick, mostly quartzite breccia. About a hundred feet of undisturbed beds at the top showed the following section:

Section of Zabriskie Quartzite north side of Echo Canyon center north side NW¼ sec. 15, T. 27 N., R. 2 E.

	Feet
Top. Base of Carrara Formation.	
1. Quartzite; interbedded white, black, and reddish beds 6 in to 2 ft thick, evenly laminated, not much cross-bedding	45
2. Shale, siltstone, and sandstone, purple, micaceous	6
3. Quartzite, white, vitreous; in beds 1 ft thick; grains as much as 1 mm	42
4. Tectonically crushed quartzite	200
Total thickness uncertain because of brecciation and faulting.	
Base. Wood Canyon Formation.	

Five samples of Zabriskie Quartzite analyzed for trace elements contain similar amounts of the several elements (table 9). The quantities are very much less than in the Stirling Quartzite (table 7), but the proportions appear not to be greatly different.

A block of sheared granulated quartzite surrounded by volcanic rocks near the north end of the Artists Drive fault blocks is represented by the two samples F and G in table 9. The quartzite probably is the Zabriskie, but it could be the Eureka (table 13). Its low content of trace elements makes it unlike any of the known Precambrian quartzites.

The Zabriskie Quartzite is considered Early Cambrian in age.

CARRARA FORMATION

The Carrara Formation was named by Cornwall and Kleinhampl (1962) for exposures at Bare Mountain, Nev., just north of Death Valley. The Carrara Formation represents a sequence of beds transitional between the underlying clastic formations (Zabriskie Quartzite and Wood Canyon Formation) and the overlying carbonate ones (Bonanza King and younger formations). The Carrara is widespread in the Death Valley region where it is characterized by an alternation of shaly or

TABLE 9.—*Trace elements in the Zabriskie Quartzite*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Panamint Range				Funeral Mountains	Fault block in volcanics at north end of Artists Drive	
	A	B	C	D	E	F	G
Pb	10	<10	<10	<10	<10	30	<10
Mn	100	20	10	10	20	10	10
Cu	3	100	5	3	5	3	20
Zr	200	200	70	50	70	50	20
Ni	5	5	5	5	5	3	<5
Co	<10	<10	<10	<10	<10	<10	<10
V	<10	10	<10	<10	<10	10	<10
Y	<10	<10	<10	<10	<10	<10	<10
Be	<1	<1	<1	<1	<1	<1	<1
Ti	300	500	300	100	150	100	150
B	15	10	10	10	10	10	<10
Sc	<10	<10	<10	<10	<10	<10	10
Ga	<5	<5	<5	<5	<5	<5	<20
Cr	<5	<5	<5	<5	<5	5	10
Ba	50	150	30	20	200	20	30
Sr	<20	<20	<20	<20	20	<20	<10
Mg	0.05	0.02	0.05	0.015	0.02	0.05	0.05

NOTE.—Also found: La, <30; Mo, <2; Ag, <1; Bi, <5; Sn, <10; As, <500; Sb, <50 W, <20; Nb, <10.

silty members and limestone members. Thin generally yellowish limestone beds composed mostly of fragmentary olenellid trilobites representing species of *Bristolia*, *Fremontia*, and *Peachella* are found throughout the region near the bottom of the formation. In the Funeral Mountains and at Bare Mountain, a prominent blue-gray limestone member with many "*Girvanella*" beds is found near the top of the Lower Cambrian part of the formation. This member is separated from an upper limestone member by an interval of shales and siltstones. The upper member includes one or two massive limestone units overlain by a yellow- or white-weathering thin-bedded limestone unit with some thin shale interbeds. Although no fossils were obtained from this thin-bedded unit in Death Valley, trilobites of the Middle Cambrian *Glossopleura* zone were collected from it at Eagle Mountain to the southeast and Lathrop Wells to the east (A. R. Palmer, oral commun., 1961).

In the Funeral Mountains and Panamint Range, the Carrara Formation varies greatly in thickness, probably due to crushing and shingling as a result of deformation. The best section, given below, is in Echo Canyon in the Funeral Mountains where the formation is more than 1,200 feet thick.

In the Manly Peak quadrangle, southwest of the area covered in this report, the carbonate rocks above the Zabriskie Quartzite aggregate 4,800 feet thick and are referred to the Lotus Formation by Johnson (1947, p. 380).

Section of Carrara Formation in Funeral Mountains, in Echo Canyon at east edge of the Furnace Creek quadrangle, sec. 15, T. 27 N., R. 2 E.

[Measured by Charles B. Hunt, A. R. Palmer, and R. J. Ross, Jr.]

	Feet
Top. Base of Bonanza King Formation; dark dolomite forms cliff.	
1. Probably limestone; inaccessible cliff; blue beds as much as 20 ft thick interbedded with well-bedded tan beds 30 ft thick; estimated thickness	200
2. Variegated shaly beds, pink, tan, and yellow; not accessible; estimated thickness	200
3. Limestone, massive, thick-bedded; dark blue below, top 25 ft is white; forms cliff. Thickness estimated	100
4. Sandstone, limy, and sandy limestone; some shaly beds near base; limestone thick bedded toward top; tan weathering	100
5. Shale, olive-green	175
Colln. F-56 (3097-CO), from base: *Olenellus* sp.	
6. Limestone, cliff former; lower part thin bedded and grades downward to unit 7; upper part consists of 2 thick-bedded units separated by a thin-bedded one; "*Girvanella*" common	175
7. Limestone, well-bedded; silty, mottled; abundant "*Girvanella*"	20
8. Limestone, blue-gray; cliff former; well bedded; beds about 3 in. thick	60
9. Covered	50
10. Shale, like unit 11; with thin interbeds of bioclastic limestone that weathers light tan; the number of limestone beds increase upward	70
Colln. F-55 (3101-CO), from this unit: *Bristolia* cf. *B. insolens* (Resser) *Peachella* sp. *Paedeumias* sp.	
11. Shale	90
Colln. F-83 (3148-CO), from lower 15 ft: *Paedeumias nevadensis* Resser *Fremontia* sp.	
12. Transition beds; mostly thin-bedded quartzite, 1 ft of brown-weathering limestone at base; other beds of limestone less than 6 in. thick	20
Total thickness	1,260
Base. Zabriskie Quartzite.	

Figure 18 is a view of the Carrara Formation at the mouth of Death Valley Canyon. Figure 19 shows one of the bioclastic "trilobite-trash" beds characteristic of the lower part of the Carrara Formation.

Limestone beds in the Carrara Formation are fine grained, commonly 0.01–0.05 mm in diameter. With the calcite are scattered grains of quartz and some muscovite (fig. 20*A*). Some of the carbonate grains are brownish and may be siderite.

Eleven collections of fossils have been made from the Carrara Formation in the Death Valley area. They were studied by A. R. Palmer who has reported on them as follows (locations of collections are shown on the geologic map):

F-2 (2454-CO). Dark-blue limestone about 100 ft above Za briski Quartzite, north foot of Tucki Mountain between Mo saic and Grotto Canyons, Stovepipe Wells quad. (NE¼NE¼ sec. 7, T. 16 S., R. 45 E.) "Scraps of olenellid trilobites ar definitely of Early Cambrian age. The scraps are not identif able as to genus, but the olenellid is a long-eyed form an most probably is late Early Cambrian in age." A collectio of fossils from this location but from the 100 ft of interbedde shale and sandstone overlying the Zabriski Quartzite include *Olenellus gilberti* Meek (Hopper, 1947, p. 406).

F-30 (2454-CO). About 150 ft above top of Zabriski Quartzit on hill 4213, east edge of Furnace Creek quad., ½ mile sout of the northeast corner of the quadrangle. "At least tw species of olenellid trilobites, one definitely referable to *Olen ellus*, and the other to one of the small-eyed olenellids. Th age is probably late, but not latest Early Cambrian."

F-34 (3100-CO). Carrara Formation, 150 ft above base; ! mile northeast of Nevares Spring, Chloride Cliff quad. N cor. sec. 36, T. 28 N., R. 1 E. "*Bristolia* sp.; *Peachella* sp."

F-38 (3102-CO). Hill above Chuckawalla Spring, 2.4 miles we of NW cor. sec. 6, T. 20 S., R. 47 E., Bennetts Well qua "*Peachella* sp., *Dictyonina* sp., silicified baby olenellids."

F-39 (3905-CO). North side of Hanaupah Canyon; hill 378 3.75 miles west of the SW cor. sec. 6, T. 20 S., R. 47 E., Be netts Well quad. "*Fremontia* sp."

F-52 (3091-CO). About 15 ft above the Zabriskie Quartzi in ridge north of Blackwater Wash (see section following Furnace Creek quad. "*Dictyonina* sp; silicified baby olene lids."

F-55 (3101-CO). Bioclastic beds 110–180 ft above base of Ca rara Formation. (See also F-83.) North side of Echo Ca yon, about 500 ft upstream from the narrows at the edge the Furnace Creek quad. (NE¼NW¼ sec. 15, T. 27 N., 2 E.). See measured section.

F-56 (3097-CO). Shale at top of cliff-forming limestone; 4 ft above base of Carrara Formation; see measured sectio Echo Canyon, just below narrows at east edge of the Furna Creek quad. Center NW¼ sec. 15, T. 27 N., R. 2 E.

F-57. Lowermost Carrara Formation, probably same hori as F-55; in Echo Canyon ½ mile downstream from the n rows at the east edge of the Furnace Creek quad. (Cen west side NW¼ sec. 15, T. 27 N., R. 2 E.). "Olenellid scra certainly of early Cambrian age, but the scraps are not id tifiable."

F-60 (3105-CO). At flat fault in Funeral Mountains, so side of the mouth of the northernmost canyon (unnamed) the Furnace Creek quad.; estimated 100 ft above Zabri Quartzite but position uncertain because of faulting. (ft northeast of SW cor. sec. 5, T. 27 N., R. 2 E.). "*Brist* sp., unidentifiable pytchoparoid."

F-83 (3148-CO). Shale 20–35 ft above top of Zabriskie Qu zite (see measured section) in Echo Canyon, north side, a 750 ft upstream from the narrows at the east edge of Furnace Creek quad. (North edge, NW¼ sec. 15, T. 27 N 2 E.).

Trace elements in 8 samples of the limestone and s beds of the Carrara Formation are given in tabl Samples of limestone from the Funeral Mountain much like those from the Panamint Range. The portions and amounts of the trace elements are s what different from those in a single specimen of l stone from the Wood Canyon Formation (table 8)

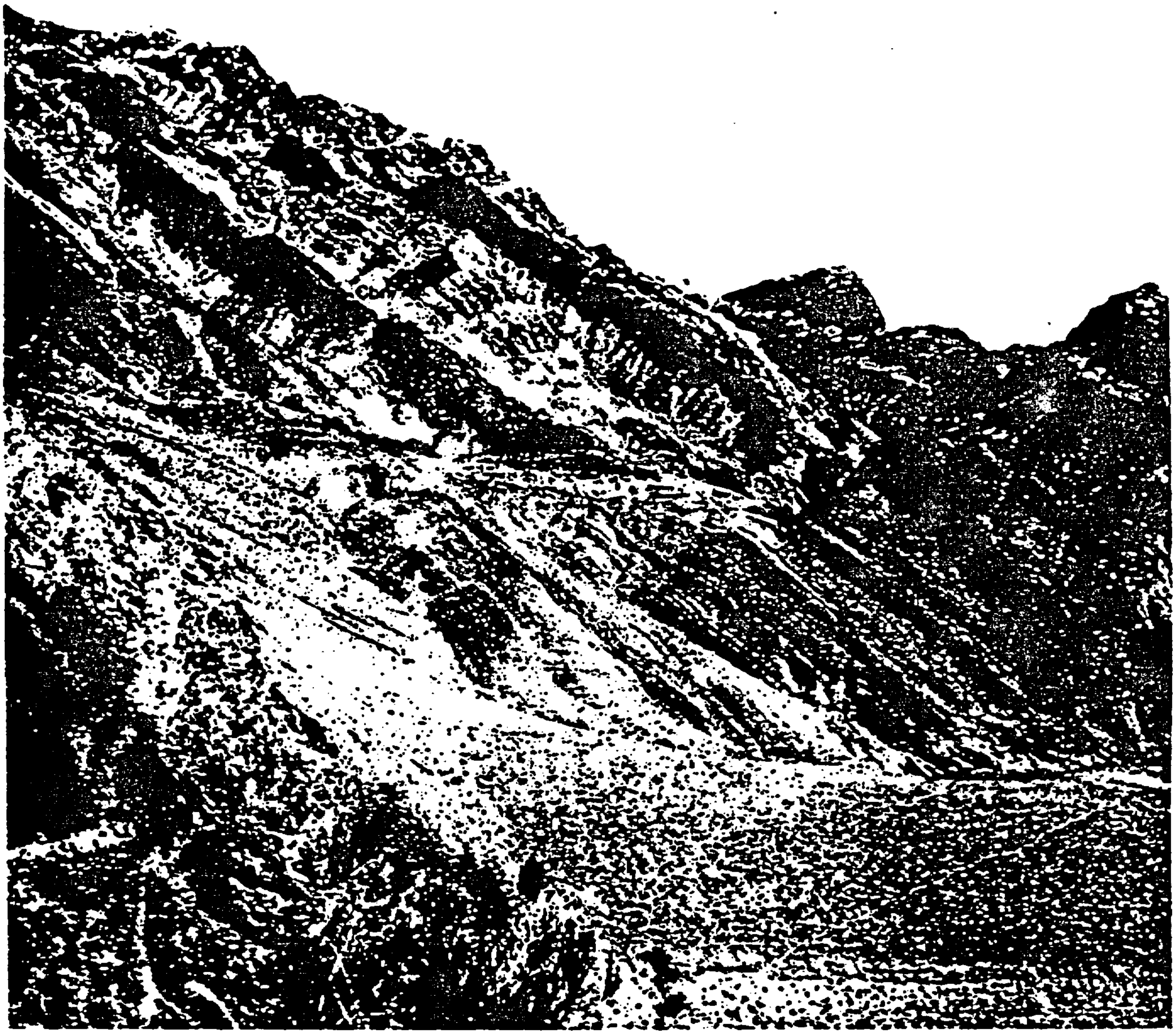


FIGURE 18.—Cambrian formations at the mouth of Death Valley Canyon, view north. In left foreground is Zabriskie Quartzite (Cz). To right of this and below the flat fault is 1,000 feet of shale and thin-bedded sandy shale and limestone of the Carrara Formation (Cc). The upper plate of the fault is mostly thick-bedded dolomite belonging to the Bonanza King and Nopah Formations (Cbn).

they are more like those in the Precambrian dolomite (tables 4–6).

BONANZA KING FORMATION

The Bonanza King Formation at the type locality in the Providence Mountains is about 2,000 feet thick (Hazzard and Mason, 1936, p. 234) and is of Middle and Late Cambrian age (Palmer and Hazzard, 1956, p. 2498). In the Nopah Range, 1,500 feet of rocks were originally assigned to the Bonanza King Formation (Hazzard, 1937b, p. 277), but the section there has recently been revised to include in the Bonanza King overlying rocks that had been assigned to the Cornfield Springs Formation (Palmer and Hazzard, 1956, p. 2495), giving a total thickness of almost 4,500 feet for the Bonanza King Formation in the Nopah Range.

In the Quartz Spring area in the northern Panamint Range equivalent beds, designated the Racetrack Dolomite, are more than 1,900 feet thick, but the base there is not exposed (McAllister, 1952, p. 9).

In the mountains adjoining Death Valley the Bonanza King Formation is about 3,000 feet thick. In Trail Canyon the computed thickness is 3,500 feet, but this

FIGURE 19.—Fragments of bioclastic "trilobite-trash" bed typical of the lower part of the Carrara Formation. The trilobites are olenellids.

thickness includes some beds that appear to be duplicated by faulting. The lower half of the formation in Trail Canyon is thick-bedded dark dolomite. Near the middle of the formation there are 2 light-tan shaly and sandy zones, each less than 50 feet thick and about 200 feet apart stratigraphically. The upper of these zones

TABLE 10.—*Trace elements in the Carrara Formation*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Panamint Range					Funeral Mountains		
	Limestone					Limestone		Shale
	A	B	C	D[1]	E	F	G	H
Pb	20	10	15	15	20	30	20	100
Mn	1,500	70	100	2,000	3,000	300	500	300
Cu	3	2	2	20	50	2	2	20
Zr	20	<10	10	30	200	<10	20	70
Ni	7	<5	<5	20	<5	5	10	20
Co	<10	<10	<10	10	<10	<10	<10	15
V	<10	<10	<10	15	50	10	10	50
Y	20	<10	<10	20	50	<10	10	30
Be	<1	<1	<1	1	<1	<1	<1	5
Ti	300	100	100	500	1,500	150	300	3,000
B	20	<10	<10	15	<10	10	20	70
Sc	<10	<10	<10	10	<10	<10	<10	15
Ga	5	<5	<5	10	<20	<5	5	20
Cr	5	<5	5	15	20	10	10	70
Ba	70	15	30	100	100	30	30	300
Sr	500	500	500	200	5,000	300	500+	50
Mg	0.5	0.7	0.5	1	5	0.5	0.7	0.7

[1] Sample D is from the Burro Trail fault at Chuckwalla Spring.

NOTE.—Also found: Sn, <10; Ag, <1; Ge, <20; As, <1,000; Sb, <200; In, <50; Cd, <50; Tl, <100; Ta, <50; W, <50; Mo, <2.

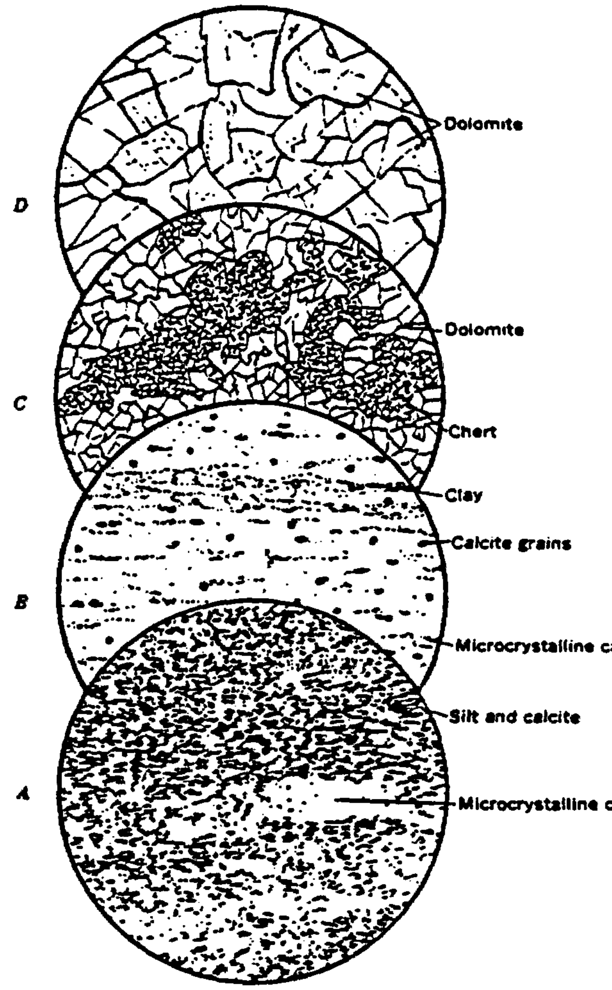


FIGURE 20.—Micrographs of carbonate rocks from Cambrian for tions. *A*, Silty limestone, Carrara Formation. *B*, Limestone fi middle member of Bonanza King Formation; this limestone dense and contains little silt or clay. *C*, Dark-colored dolo with chert from Nopah Formation. *D*, Light-colored dolo Nopah Formation. Diameter of field, 2.5 mm.

is fossiliferous and contains, in addition to indete nate linguloids, the Middle Cambrian trilo "*Ehmaniella*" sp. (colln. F-41, 3099-CO; fig. 21). the Nopah Range, shaly beds at about this same st graphic position have yielded the trilobite *Ehm* (Hazzard, 1937b, p. 319). The overlying beds are included in the Bonanza King Formation, alth originally referred to the Cornfield Springs Form (Palmer and Hazzard, 1956, p. 2498).

Overlying this fossiliferous zone in Trail Cany about 100 feet of dark thick-bedded dolomite, w is overlain by a distinctive unit, almost 600 feet t consisting of well-bedded and comparatively bedded limestone with only a few beds of dolomite 22). An intensive search of these beds failed to

FIGURE 21.—Linguloid brachiopods and trilobites ("*Ehmaniella*" sp.) from shaly zone near the middle of the Bonanza King Formation, Trail Canyon.

cover any fossils other than stromatolites. Figure 20*B* shows a micrograph of this limestone.

The top 1,000 feet of the Bonanza King Formation in Trail Canyon consists of drab-colored massive dolomite in four beds, two of which are dark gray and two lighter gray. The dark beds are poorly bedded; the lighter gray beds are well bedded in beds 1–3 feet thick and striped with a few dark beds. The top of the Bonanza King Formation is well marked here and elsewhere in the Death Valley area by the fossiliferous shale at the base of the Nopah Formation.

At the north foot of Tucki Mountain, between Mosaic and Grotto Canyon, the thickness of the Bonanza King Formation is computed as 3,000 feet. The well-bedded limestone unit about 1,000 feet below the top of the formation crops out at the mouth of Grotto Canyon, but the fossiliferous shaly and sandy layers that occur below it in Trail Canyon were not found.

An attempt to measure the thickness of the Bonanza King Formation in Echo Canyon in the Funeral Mountains gave a figure of only 2,050 feet. This figure is of doubtful stratigraphic significance, however, because the section along the canyon crosses several faults having displacements that are not well known. The fossiliferous zones were not found in that section.

More careful work in the Panamint Range probably will make it possible to subdivide the Bonanza King Formation into upper and lower thick-bedded dolomitic members separated by a middle thin-bedded limestone and shale member.

Table 11 gives the trace elements in samples from the Bonanza King Formation. The analyses suggest that the limestone has about the same trace elements as the more prevalent dolomite. The trace elements differ but slightly from those in the Carrara Formation (table 10).

TABLE 11.—*Trace elements in the Bonanza King Formation, Panamint Range*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U.S Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Sandy limestone	Limestone	Dolomite	
Pb	30	15	15	10
Mn	700	100	300	20
Cu	7	30	3	50
Zr	150	<10	<10	<10
Ni	20	<5	5	<5
Co	10	<10	10	<10
V	50	20	<10	20
Y	30	10	<10	15
Be	2	<1	<1	<1
Ti	2,000	150	20	30
B	50	<10	10	10
Sc	15	<10	10	<10
Ga	20	<20	<10	<20
Cr	50	10	5	10
Ba	700	15	10	20
Sr	200	500	100	500
Mg	0.7	>5	>5	>5

NOTE.—Also found: Sn, <10; Ag, < 1; Ge, <20; As, <1,000; Sb, <200; In, <50; Cd, <50; Tl, <100; Ta, <80; W, <50.

NOPAH FORMATION

The Nopah Formation at the type locality in the Nopah Range is 1,740 feet thick (Hazzard, 1937b, p. 276, 320) and consists of a basal shaly member about 100 feet thick overlain by alternating light- and dark-gray dolomites. In the northern Panamint Range the sequence of lithologies is similar to that in the Nopah Range, and the thickness is about 1,600 feet (McAllister, 1952, p. 9; 1955, p. 10; 1956). At both locations fossils indicative of Late Cambrian age occur in shaly beds at the base of the formation. Indeterminate gastropods of possible latest Cambrian age have been found in the upper 700 feet of the Nopah Formation in the Amargosa Range and in the northern Panamint Range (McAllister, 1952, p. 10). The Nopah Formation is correlated with the Cornfield Springs Formation in the Providence Mountains (Palmer, 1956, p. 673).

The Nopah Formation in the mountains bordering Death Valley is very similar lithologically to that at

FIGURE 22.—Thin-bedded middle member of the Bonanza King Formation on the north side of Trail Canyon, view north. This member, about 600 feet thick, forms a distinctive unit separating massive thick-bedded dolomite comprising the upper and lower members of the formation.

the type section and in the northern Panamint Range, and the characteristic fossils occur in the shaly beds at the base (fig. 23). The formation is about 1,500 feet thick in the Funeral Mountains, but it appears to be somewhat thinner in Tucki Mountain and in the southern part of the Panamint Range. Several computed thicknesses average about 1,200 feet. Sections through the whole formation and through the characteristic basal shale member are given below.

Figure 24 is a view of the banded light- and dark-colored dolomite forming most of the Nopah Formation. Much of this dolomite is [illegible]y, the chert occurring as nodules distributed along bedding planes and as irregular lumps that seem to have little or no relation to the bedding. Figure 20 shows some micrographs of the light dolomite and of the dark dolomite and chert. A section of the Nopah Formation measured in Trail Canyon follows.

FIGURE 23.—Bioclastic bed with fragments of trilobites and brachiopods from shale unit at base of Nopah Formation (colln. F-54-3098-CO).

Section of Nopah Formation, north side of Trail Canyon 2 miles above the canyon mouth

[Section measured by Charles B. Hunt and A. R. Palmer]

	Feet
Top. Base of Pogonip Group.	
1. Dolomite, black, thick-bedded	75
2. Dolomite, gray, in thick beds	275
3. Dolomite, black, massive	50
4. Dolomite, light-colored	120
5. Dolomite, black, massive	75
6. Dolomite, light-colored	110
7. Dolomite, banded black and light-colored	85
8. Dolomite, light-colored, thin-bedded	60
9. Dolomite, black, massive	60

Section of Nopah Formation, north side of Trail Canyon 2 miles above the canyon mouth—Continued

	Feet
10. Dolomite, light-colored, thin-bedded, forms slope	110
11. Shale and limestone. Shale greenish or brown in beds 1–4 ft thick; some shale has nodules of limestone. Limestones brown and in beds 6 in to 2 ft thick. Many are trilobite breccias; others are echinoderm breccias. Linguloid brachiopods in the shale	75

Colln. F-68, (3143-CO), top of unit:

Cheilocephalus sp.

Undetermined dokimocephalinid

Section of Nopah Formation, north side of Trail Canyon 2 miles above the canyon mouth—Continued

	Feet
11. Shale and limestone—Continued	
Colln. F-67, (3142-CO), 20 ft below top of unit:	
Apachia sp.	
Undetermined pterocephalinid	
Colln. F-66, (3141-CO), at base of unit:	
Homagnostus obesus (Belt)	
Strigambitus utahensis (Resser)	
Dunderbergia variagranula Palmer	
Apsotreta sp.	
Dysoristus sp.	
12. Concealed	25
Total thickness of Nopah Formation	1,120

Base. Thick-bedded dolomite, top of Bonanza King Formation.

Section of shale member at base of Nopah Formation, east side of mouth of Grotto Canyon, NE¼NE¼ sec. 8, T. 16 S., R. 45 E., alt. 1,175 ft

Top. Thick-bedded dolomite of Nopah Formation. Feet

Thin-bedded limestone and tan shale. Shale mostly in laminae separating thin beds of limestone, but some shale beds are 10 ft thick. Basal 15 ft is brown limestone; higher ones blue gray, in beds 1 in to 2 ft thick. Nodular limestone and shale 40 ft above base 115

Colln. F-69 (3144-CO) top of unit:
- *Elburgia quinnensis* (Resser)
- *Sigmocheilus* sp.
- *Cheilocephalus brachyops*? Palmer
- *Apachia* sp.

Colln F-1 (2433-CO), near middle of unit:
- *Elburgia quinnensis* (Resser)
- *Cheilocephalus* sp.
- *Strigambitus*? *blepharina* Palmer
- *Homagnostus* sp.
- *Morosa brevispina* Palmer

Colln. F-71 (3146-CO), 15 ft above base of unit;
- *Dunderbergia variagranula* Palmer
- *strigambitus utahensis* (Resser)
- *Homagnostus* sp.

Colln. F-70 (3145-CO), basal limestone of unit:
- *Minupeltis conservator* Palmer
- *Cernuolimbus granulosus* Palmer
- *Pseudagnostus* sp.

Base. Massive dolomite at top of Bonanza King Formation.

Section of shale member at base of Nopah Formation, south side Echo Canyon one-half mile above the canyon mouth

[Measured by Charles B. Hunt, A. R. Palmer, and R. J. Ross, Jr.]

Top. Base of lowest cliff-forming dolomite in Nopah Formation.

	Feet
1. Limestone, dark-brown to black; in beds 1 ft thick; cherty	15
Colln. F-82, (3147-CO), from top of unit:	
Pterocephalia? *punctata* Palmer	
Pseudagnostus sp.	
"*Acrotreta*" *spinosa* Walcott	

Section of shale member at base of Nopah Formation, south side Echo Canyon one-half mile above the canyon mouth—Con.

	Feet
2. Shale, tan, limy; in part sandy	45
Colln. F-54, (3098-CO), 15-25 ft above base of unit:	
Elburgia *quinnensis*? (Resser)	
Strigambitus? *blepharina* Palmer	
Apsotreta sp.	
3. Limestone pebble conglomerate and coarse bioclastic beds; some limestone beds ½ in thick; some silt beds ¼ in thick; chert	50
4. Covered	50
Total thickness	160

Base. Massive dolomite, top of Bonanza King Formation.

Two other collections of fossils from the Nopah Formation were reported upon by Palmer as follows:

F-53 (3103-CO). North side of Echo Canyon, ¼ mile above mouth, Furnace Creek quad. (SW¼SW¼ sec. 16, T. 27 N., R. 2 E.). "*Apsotreta* sp.; abundant siliceous sponge spicules a part of this collection is essentially a spiculite."

F-58 (3039-CO). Ridgetop south of Echo Canyon about 1 mile above the mouth of the canyon, alt 2,950 ft, Furnace Creek quad., probably near middle of the shale unit (SW¼SW¼ sec. 15, T. 27 N., R. 2 E.). "*Elburgia quinnensis* (Resser); *Pseudagnostus communis* (Hall and Whitfield); *Homagnostus tumidosus* (Hall and Whitfield); *Morosa brevispina* Palmer; *Strigambitus*? *blepharina* Palmer; *Apsotreta* sp.; '*Acrotreta spinosa* Walcott; conodont."

Trace elements in the Nopah Formation are listed in table 12. The samples from the Funeral Mountains are limestone from the base of the formation; those from

TABLE 12.—*Trace elements in the Nopah Formation*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Panamint Range			Funeral Mountains	
		Dolomite		Limestone	
Pb	<10	70	20	10	1
Mn	10	50	70	700	1,00
Cu	2	5	50	5	
Zr	<10	70	<10	<10	2
Ni	5	5	<5	5	<
Co	<10	<10	<10	<10	<1
V	10	<10	20	<10	<1
Y	10	<10	20	<10	<1
Be	<1	<1	<1	<1	<
Ti	<10	50	30	200	30
B	<10	<10	<10	10	1
Sc	<10	<10	<10	<10	<1
Ga	<5	<5	<20	<5	
Cr	5	5	10	10	
Ba	10	15	20	70	
Sr	150	200	300	700	1,00
Mg	>5	5	>5	0.7	0.

NOTE.—Also found: Sn, <10; Ag, <1; Ge, <20; As, <1,000; Sb, <200; In, <30; C <30; Tl, <100; Ta, <30; W, <80; Nb, <10.

the Panamint Range are dolomite from the upper part of the formation. Whereas limestone and dolomite in the Bonanza King Formation have about the same content of trace elements (table 11), the limestone and dolomite of the Nopah Formation have quite different proportions of some constituents; notably, manganese, titanium, barium, and strontium are very much more abundant in the limestone than in the dolomite. Dolomite in the Nopah Formation has about the same amount and proportions of trace elements as does the dolomite in the Bonanza King Formation.

ORDOVICIAN SYSTEM

POGONIP GROUP

The name Pogonip originally was applied to the considerable thickness of carbonate rocks lying above quartzite of Cambrian age and extending up to the Eureka Quartzite of Ordovician age (King, 1878, p. 188). The name has been redefined several times and now is restricted to rocks of Ordovician age; underlying rocks of Cambrian age now are separated from the Pogonip (Hazzard, 1937b; Hintze, 1949, 1951; Easton and others, 1953; McAllister, 1952; Nolan, Merriam, and Williams, 1956). In the Death Valley area the Pogonip Group overlies the Nopah Formation and is overlain by the Eureka Quartzite. According to Ross (oral commun., 1961) the Pogonip Group of this area probably is roughly equivalent to the Yellow Hill and Tank Hill Limestones of the Pioche district (Westgate and Knopf, 1932, p. 14).

In the Death Valley area the Pogonip Group is about 1,500 feet thick. In Trail Canyon it is composed of three distinct members. The lower member consists of thin-bedded dolomite, the upper of thick-bedded dolomite; the middle member is shaly (fig. 24). Very little limestone is found in this section, and there seems to be evidence of a considerable amount of secondary dolomitization, which makes comparison with measured sections in other areas difficult. In the northern part of Tucki Mountain the Pogonip may be represented in a limestone facies, but outcrops are in disjointed fault slices which make stratigraphic placement almost impossible.

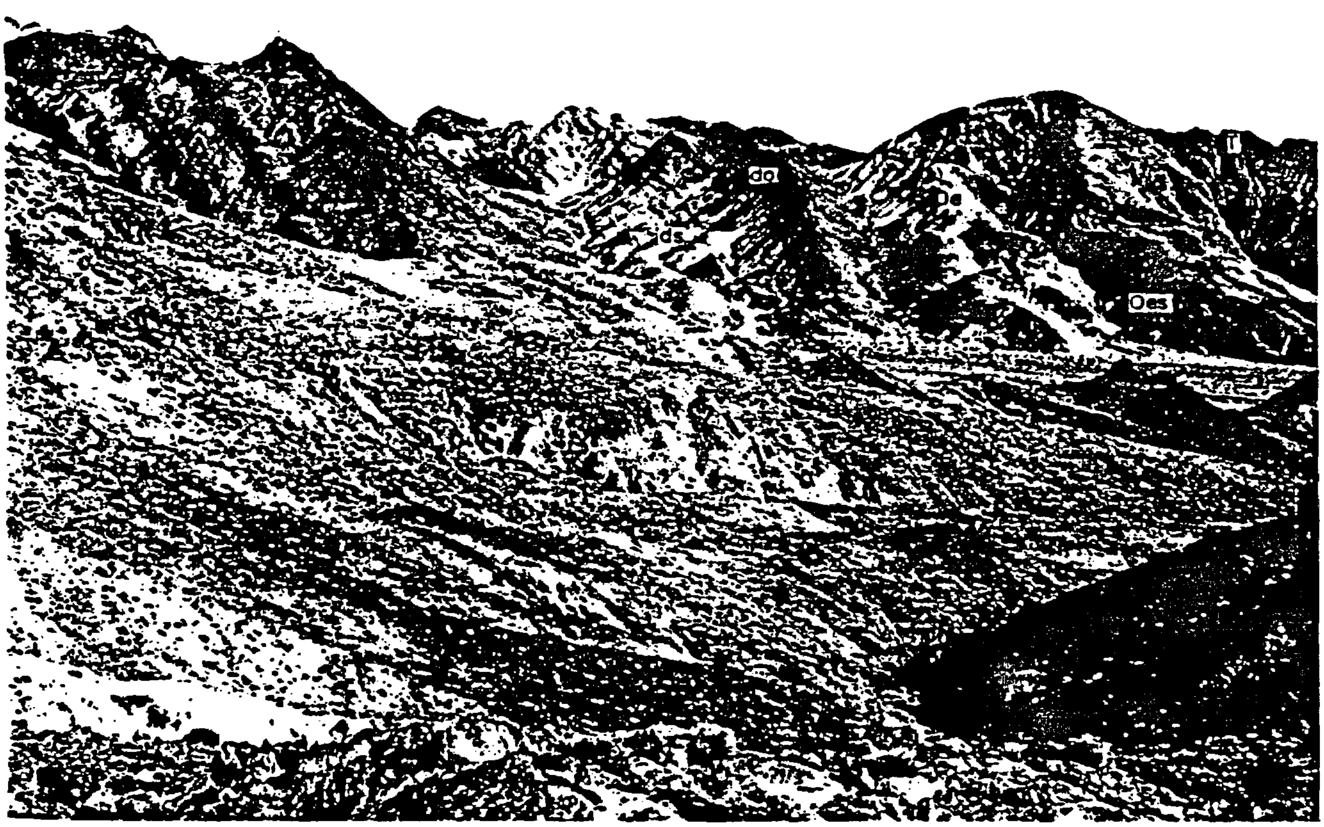


FIGURE 24.—View of Pogonip Group in Trail Canyon. View is north. At left is light- and dark-colored dolomite of the Nopah Formation (Cn). Thin-bedded dolomite and shale (ds) in the lower and middle part of the Pogonip Group form the saddle; thick-bedded dolomite (do) in the upper part of the Pogonip forms the dark ridge dipping under the light-colored Eureka Quartzite (Oe) at the right. Hill at extreme right is capped with Tertiary lavas (T); at the base of the hill is dark Ely Springs Dolomite (Oes).

A threefold division of the Pogonip is possible in several nearby areas; the three subdivisions according to Ross (oral commun., 1961), are roughly equivalent to the shaly limestones of the Goodwin Formation, overlain by the limy shales of the Ninemile Formation, which, in turn, are overlain by the more massive limestones of the Antelope Valley Limestone, all of the central Nevada Eureka district (Nolan and others, 1956, p. 24–25). Hazzard (1937b, p. 276) has recognized a similar tripartite division of the Pogonip Group in the Nopah Range. Similar subdivisions have been reported in the northern part of the Panamint Range by McAllister (1952, p. 11), and at Bare Mountain to the north of Death Valley (Cornwall and Kleinhampl, 1962), as well as in the general area of the Nevada Test Site farther to the east.

In the Death Valley area the basal unit of the Pogonip Group is mostly dolomite; but this may be due to metamorphism, because this member elsewhere includes considerable limestone. In adjacent areas it is mostly limestone. The middle unit of the Pogonip Group in the Death Valley area is reddish; this probably also is attributable to metamorphism, because relict sedimentary structures, such as intraformational conglomerate, coarse bioclastic beds, and occasional crossbedding in calcarenites, can be found. Also, numerous dikes cut the Pogonip Group from south of Trail Canyon to Blackwater Wash. Figure 25*A* is a micrograph of dolomite from the Pogonip Group.

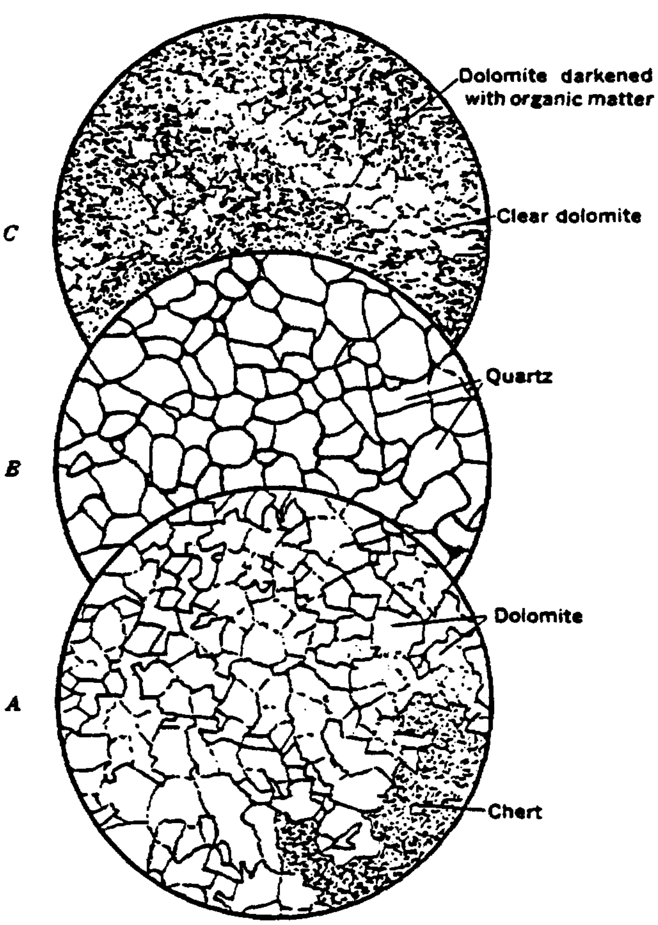


FIGURE 25.—Micrographs of rocks from Ordovician formations. *A*, Dolomite, with chert, from upper unit of Pogonip Group. *B*, Eureka Quartzite. *C*, Ely Springs Dolomite. Diameter of fields, 2.5 mm.

No fossils were found in the lower and middle parts of the Pogonip Group, but the cliff-forming dolomite comprising the upper third of the formation at many places contains large gastropods in such abundance as to be a lithologic guide to the dolomite (fig. 26).

Fossils from the Pogonip Group were collected at 10 localities in the Death Valley area, as follows:

F–3 (*D643–CO*). Dolomite in upper part of Pogonip Group; south side of canyon north of Trail Canyon, Furnace Creek quad.; 2.1 miles west and 0.3 mile north of SW¼ sec. 31, T. 18 S., R. 47 E., alt 1,280 ft. Identifications by R. J. Ross, Jr., *Receptaculites*? sp.: *Palliseria*? sp. Probably high Pogonip and equivalent to the Antelope Valley Limestone of the Eureka area, Nevada."

F–5 (not cataloged). Dolomite, upper part of Pogonip Group, north base of Tucki Mountain below mouth of Trellis Canyon, Stovepipe Wells quad. (SE¼SE¼ sec. 23, T. 16 S., R. 45 E., alt 1,280 ft). Identification by E. L. Yochelson, "The material consists of two pieces of dark-gray dolomite showing poor cross sections of three gastropods; one saw cut to determine the third dimensions shows a profile suggestive of *Palliseria*, a guide to the Antelope Valley Limestone of central Nevada."

F–13 (*D645–CO*). Dolomite in upper part of Pogonip Group, north side of Trail Canyon, alt 1,600 ft, Furnace Creek quad. Identification by R. J. Ross, Jr., "*Palliseria* sp."

F–27 (*D641–CO*). Dolomite in upper part of Pogonip Group, north side of second ridge south of the mouth of Trail Canyon, alt 1,600 ft, Furnace Creek quad. Identification by R. J. Ross, Jr., "Probably *Palliseria*."

F–28 (*D642–CO*). Same as F–27, lower in gulch, alt about 1,500 ft. Identification by R. J. Ross, Jr., "*Receptaculites* sp., *Palliseria* sp. Unquestionably high Pogonip."

F–40 (*3626–CO*). Pogonip Group at flat fault 1¼ miles south of Trail Canyon, 1.2 miles east of hill 4889, Furnace Creek quad. Identification by E. L. Yochelson, "Cross section of sponge?; *Maclurites* sp. indet.; *Palliseria robusta* Wilson. *Palliseria Robusta*, confined to the second oldest faunal zone in the Antelope Valley Limestone of central Nevada, is a guide to early Middle Ordovician age."

F–42 (*D587–CO*). Fault block under Eureka Quartzite, 1½ miles south of Trail Canyon, alt 1,000 ft, Furnace Creek quad. (7,500 ft south of west of SW cor. sec. 7, T. 19 S., R. 47 E.). Identification by R. J. Ross, Jr., "Very poorly preserved gastropods and trilobite fragments. None can be identified. Small brachiopod species suggests *Diparelasma*; it and the lithology suggest Pogonip."

F–43 (*D588–CO*). 1,000 ft northeast of F–42 and apparently overlying it. Identification by R. J. Ross, Jr., Unidentifiable gastropods; abundant *Girvanella*? Age indeterminate."

F–47 (*3625–CO*). 1½ miles southwest of "Dinosaur," 2½ miles southwest of SW cor. sec. 6, T. 18 S., R. 47 E. At south base of hill 430, Furnace Creek quad. Identification by E. L. Yoch-

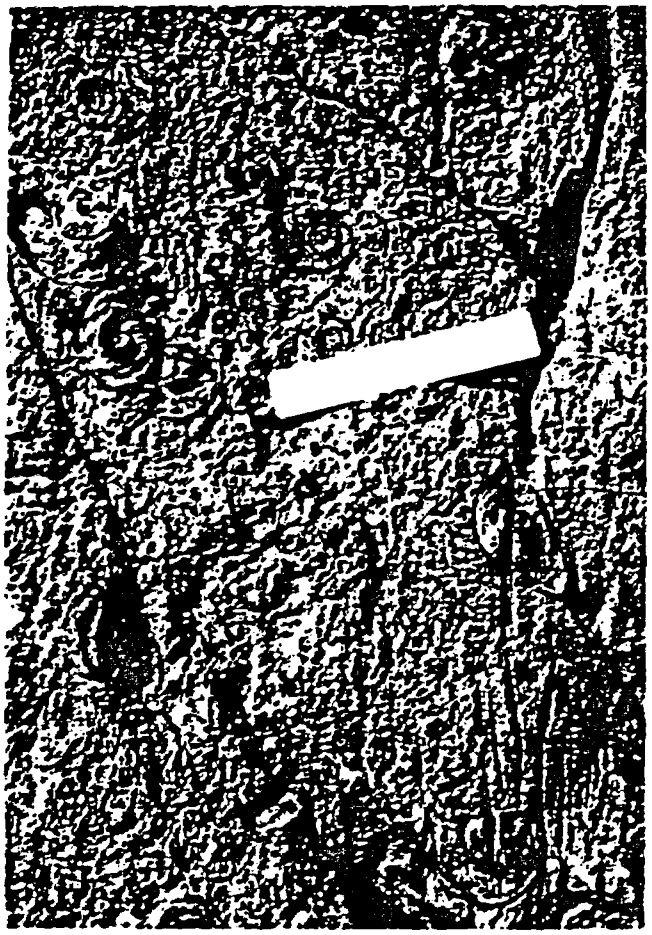


FIGURE 26.—Large gastropods (*Palliseria* sp.) in dolomite in upper part of the Pogonip Group.

elson, "*Receptaculites* sp., *Maclurites*, incomplete but suggestive of *M. magnus; Palliseria robusta* Wilson. The *Palliseria* is guide to early Middle Ordovician (see comment for F–40)." Also in this collection, according to R. J. Ross, Jr., is *Syntrophopsis*? sp.

Colln. F–61 (*D589–CO*). From fault block, probably Pogonip Group, in Red Amphitheater breccia at mouth of second canyon south of Echo Canyon, Furnace Creek quad. (Center, east side, SE¼NE¼ sec. 21, T. 27 N., R. 2 E.). According A. R. Palmer, "This collection contains orthoid brachiopods, gastropod cross sections, and an asaphid trilobite pygidium which collectively indicate an Ordovician age." According to R. J. Ross, Jr., "The trilobite segments are very characteristic of Ordovician proparian types, and a few brachiopod outlines suggest *Anomalorthis*."

Fossils collected in the Funeral Mountains (C. A. Richards [1]), in the Panamint Range (McAllister, 1952, p. 11; 1956), in the Nopah Range (Hazzard, 1937b, p. 276) and on the Nevada Test Site (Johnson and Hibbard, 1957, p. 346–347) indicate that the upper dolomitic unit in the Death Valley area correlates with the Antelope Valley Limestone.

The Pogonip Group is considered to be Early and Middle Ordovician in age.

In the Death Valley area the contact between the Pogonip Group and Eureka Quartzite appears to be gradational for it is marked by a series of interbedded quartzites and dolomites.

The following section of the Pogonip Group was measured in Trail Canyon (fig. 24).

Section of Pogonip Group, south side of Trail Canyon

[Measured by Charles B. Hunt, R. J. Ross, Jr., and A. R. Palmer]

	Feet
Top. Base of Eureka Quartzite; contact gradational. Contact taken at base of first quartzite; above this is 120 ft of interbedded thin-bedded quartzite and sandy dolomite transitional to overlying massive quartzite.	
1. Upper dolomite unit: mostly thick-bedded dolomite; bottom 75 ft is thin bedded, but the dolomite above this is massive with a few thin lenses of friable sandstone; abundant "*Girvanella*"; top 100 ft thinner bedded; several intraformational conglomerates; abundant *Palliseria*. Colln. F–3, F–13, F–27, and F–28 from this unit	385
2. Middle shaly unit; interbedded shale, siltstone, and dolomite; the clastics weather red and brown, probably because of metamorphism. Black limestone, 25–50 ft above base, with silt partings containing unidentifiable trilobites, brachiopods, and gastropods. Bed with cystid plates is 200–225 ft above base	285
3. Basal dolomite unit. Top 320 ft is thin-bedded dolomite interbedded with siltstone and shale; increasing shale upward gradational to unit 2. Lower 460 ft is thin-bedded dolomite, mostly weathering rustry brown; beds 2–4 in. thick; much black chert in lenses and in nodules elongated parallel to bedding; some blocky chert; a striking bed of thin-bedded blocky chert occurs 270 ft above the base. Intraformational conglomerate in beds 1–2 ft thick. Much of the dolomite is fine-grained calcarenite	780
Total thickness	1,450
Base. Top of Nopah Formation; contact taken at base of the thin-bedded dolomites.	

Trace elements in four specimens from Ordovician units are given in table 13. The single sample of dolomite from the Pogonip Group is similar to those of the Ely Springs Dolomite and to those of the dolomite rather than the limestone of the Nopah Formation (table 12).

EUREKA QUARTZITE

The name Eureka Quartzite was first used in central Nevada (Hague, 1883, p. 262; 1892, p. 54–57; see also, Kirk, 1933), and the formation has been widely recognized in the Great Basin southward to the Death Valley region. It is a massive vitreous quartzite that serves as a valuable easily recognized marker bed in the midst

[1] Richards, C. A., 1959, Geology of part of the Funeral Mountains: unpub. manuscript on file at Death Valley Natl. Monument and Univ. Southern California.

TABLE 13.—*Trace elements in Ordovician units, Panamint Range*

[Semiquantitative spectrographic analyses by Uteana Oda and E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Pogonip Group Dolomite	Eureka Quartzite	Ely Springs Dolomite	
Pb	15	<10	10	10
Mn	70	<10	30	50
Cu	3	2	2	50
Zr	<10	30	10	10
Ni	5	<5	<5	<5
Co	<10	<10	<10	<10
V	10	<10	<10	20
Y	10	<10	<10	20
Be	<1	<1	<1	<1
Ti	20	200	20	20
B	<10	15	10	<10
Sc	<10	<10	<10	<10
Ga	<5	<5	<5	<20
Cr	5	<5	5	10
Ba	10	15	20	20
Sr	150	<20	200	200
Mg	>5	0.05	>5	>5

NOTE.—Also found: Sn, <10; Ag, <1; Ge, <20; As, <1,000; Sb, <200; In, <50; Cd, <50; Tl, <100; Ta, <50; W, <50; Nb, <10.

of the thick section of carbonate formations (figs. 24, 27).

In Tucki Mountain the Eureka Quartzite is very much crushed and granulated, so that sections there cannot be regarded as meaningful for stratigraphic study. In this respect the Eureka Quartzite in that part of the area resembles the Stirling and Zabriskie Quartzites. In Trail Canyon, however, the formation seems to be less deformed. The quartzite there is not severely granulated (fig. 25*B*), and on the ridge south of Trail Canyon a measured section, which follows, indicates that the formation there is 350 feet thick.

Section of Eureka Quartzite, ridge south of the mouth of Trail Canyon

[Measured by Charles B. Hunt, R. J. Ross, Jr., and A. R. Palmer]

Top. Base of Ely Springs Dolomite; contact concealed by rubble.

	Feet
1. Quartzite, well-bedded in beds 2-5 ft thick	60
2. Quartzite, massive; weathers brown	110
3. Quartzite, mostly thin-bedded but with 2 ledges each about 15 ft thick; thin beds between the ledges fucoidal and mottled red	60
4. Quartzite and sandy dolomite, interbedded; gradational downward to dolomite unit at top of Pogonip Group; colors variegated	120
Total thickness Eureka Quartzite	350

Base. Top of Pogonip Group; Contact taken at base of lowest bed of vitreous quartzite.

Another section was measured across the crushed quartzite in Little Bridge Canyon (fig. 27). There the contact with the Ely Springs Dolomite is sharp but seems to have been sheared. The massive vitreous quartzite is 140 feet thick, and under this unit is 35 feet of thin-bedded very fine grained quartzite with a few thin beds of dolomite. The beds are 6 inches to 1 foot thick, the colors are variegated; the weathered surfaces are mottled red and green. The quartzite at this location undoubtedly is thinned by shearing.

In the northern part of the Panamint Range the Eureka Quartzite attains a thickness of 400 feet (McAllister, 1952, p. 12), in the Beatty area it is 350 feet thick (H. R. Cornwall, written commun., 1960), and in the Funeral Mountains it is about 360 feet thick (C. A. Richards [2]). In the Nopah Range the thickness is 265 feet (Hazzard, 1937b, p. 276), and at the Nevada Proving Grounds it is 285 feet (Johnson and Hibbard, 1957, p. 349–350).

No fossils have been found in the Eureka Quartzite in this area, but the age is restricted to Middle or early Late(?) Ordovician by the fossils in the underlying Pogonip Group and overlying Ely Springs Dolomite.

A single specimen of Eureka Quartzite, analyzed for trace elements (table 13), contains even less trace elements than does the Zabriskie Quartzite (table 9), which it most resembles.

ELY SPRINGS DOLOMITE

The name Ely Springs Dolomite was first applied to a formation of dark dolomite about 600 feet thick in the Ely Springs Range about 125 miles northeast of Death Valley (Westgate and Knopf, 1932, p. 15). The formation has since been widely recognized in the southern Great Basin. Its thickness ranges from 400 to 940 feet: in the Nopah Range, 800 feet (Hazzard, 1937b, p. 276); in the Beatty area, 400 feet (H. R. Cornwall, written commun., 1960); in the northern Panamint Range 940 feet (McAllister, 1952); and in the Darwin area 920 feet (Hall and MacKevett, 1958, p. 7). At most of these places and in the Death Valley area, the Ely Springs Dolomite contains Late Ordovician fossils; it overlies the Eureka Quartzite and is overlain by dolomite containing Middle Silurian fossils.

In the Death Valley area the thickness of beds assigned to the Ely Springs Dolomite is substantially less than in the surrounding region. In Trail Canyon the thickness is 425 feet; in the Funeral Mountains 403 feet of beds was assigned to the formation (C. A. Richards [3]). The formation is comparably thin at the north base of Tucki Mountain. At the widest place in the outcrop belt of the formation, at the south side

[2] See footnote, p. A35.
[3] See footnote, p. A35.

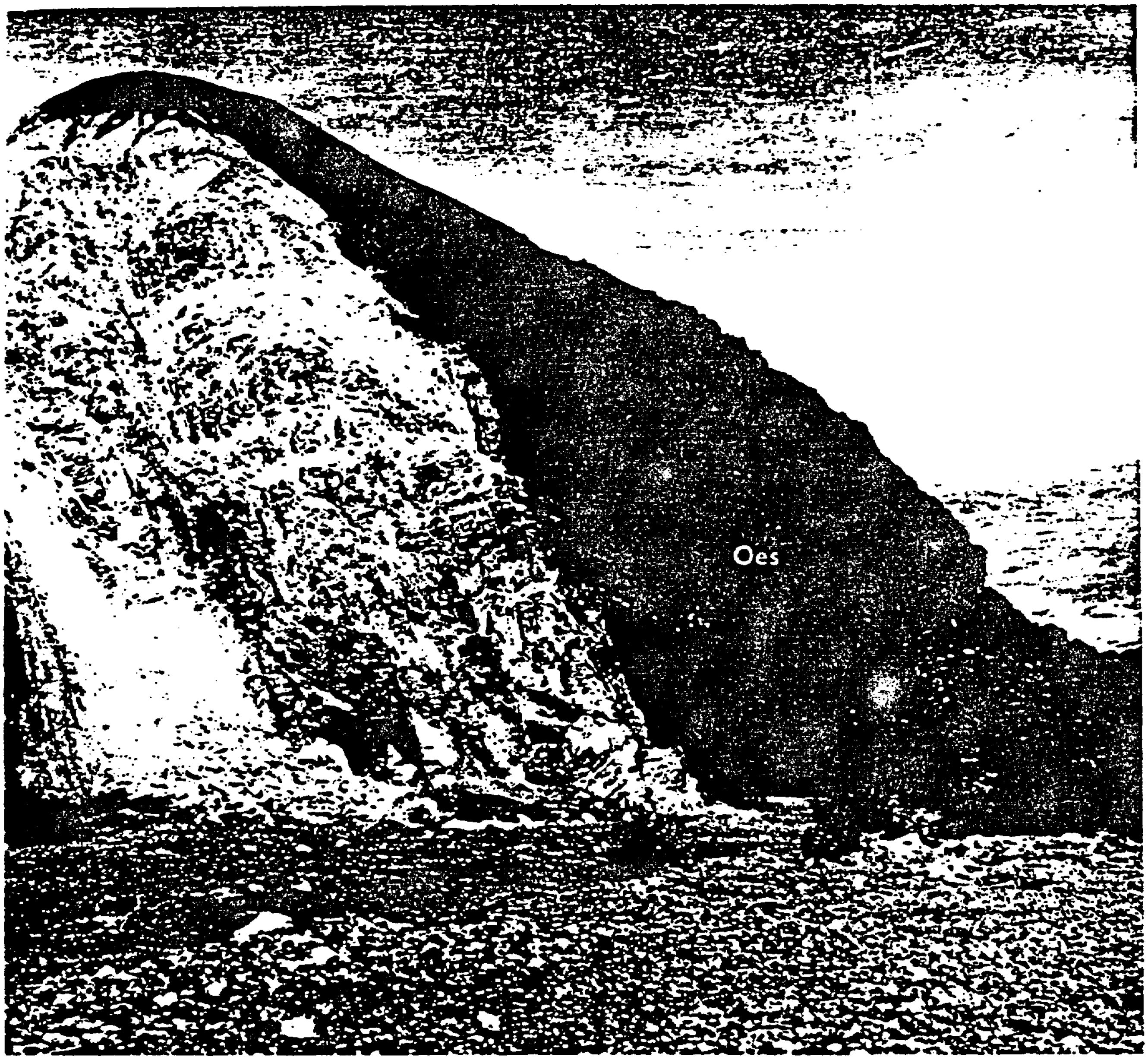


FIGURE 27.—View of Eureka Quartzite (Oe) and overlying Ely Springs Dolomite (Oes) at mouth of Little Bridge Canyon. The quartzite is much more crushed and granulated than is the dolomite and as a result has been eroded to form the valley in the foreground; carbonate formations form the ridges on either side.

of Tucki Mountain overlooking Tucki Wash, the computed thickness is 825 feet. More detailed work will be required to determine whether the differences in thickness are due to stratigraphic changes or to cutting out of beds by shearing along the bedding. Only a small part of the differences in thickness can be attributed to differences in boundaries selected for the formation. The basal contact with the Eureka Quartzite, though generally covered, can be located within a few feet (fig. 27). The boundary with the overlying light-colored dolomites, some of which contain Silurian fossils, is gradational through a zone of perhaps 100 feet.

The Ely Springs Dolomite in the Death Valley area is dark, thick bedded, and forms conspicuous cliffs above the light-colored Eureka Quartzite. The formation contains considerable dark-brown to black chert that occurs as nodules and as irregular lenses. The dark dolomite is streaked with curving light-colored lines 2–5 cm long and 1–5 mm wide, suggestive of scattered strands of spaghetti. Figure 25C is a micrograph of Ely Springs Dolomite.

The Ely Springs Dolomite has yielded a considerable fauna indicative of Late Ordovician (Richmond) age. Seven collections were made, as follows:

F-8 (*D458-CO*). Ely Springs Dolomite, 10 ft above base of the formation, in Little Bridge Canyon, Stovepipe Wells quad. (NW¼NW¼ sec. 15, T. 16 S., R. 45 E.). Identifications by R. J. Ross, Jr., "*Lepidocyclus* cf *L. laddi* Wang; *Austinella* sp.; strophomenid brachiopod; dalmanellid brachiopod; unidentified coral. The age is probably Late Ordovician, but it might be late Middle Ordovician."

F-12 (*D644-CO*). Lower part of Ely Springs Dolomite, south side of Trail Canyon, alt 1,340 ft at north base of butte with peak at 1,680 ft.

F-17 (*D151-SD*). Middle of the formation; 1 mile east of the mouth of Trellis Canyon, alt 1,200 ft, north foot of Tucki Mtn., Stovepipe Wells quad. (NE¼NE¼ sec. 26, T. 16 S., R. 45 E.).

Collections F-12 and F-17-58 were examined by W. A. Oliver, Jr., who states "they consist, respectively, of 7 and 6 fragments of small simple horn corals, mostly streptelasmatoids. These are very poorly preserved and cannot be identified. They could be of either Ordovician or Silurian age but not pre-Middle Ordovician."

F-44 (not catalogued). 1½ miles north of Trail Canyon, fault block at east end of the ridge dividing the valley; alt 1,800 ft. This collection, examined by E. L. Yochelson, yielded only isolated crinoidal columns.

F-45 (*3622-CO*). Upper part of Ely Springs Dolomite, north side of canyon next north of Trail Canyon and 2¾ miles west of SW cor. sec. 30, T. 18 S., R. 47 E., Furnace Creek quadrangle. Identifications by W. A. Oliver, Jr.,

"*Tollina* [*manipora*] sp.; streptelasmatical horn corals. The genus *Tollina* is known only from rocks of Late Ordovician age. One of the two specimens is very well preserved and is specifically distinct from representatives of the genera that I have previously seen or seen illustrated. The genus is known from the Montoya Dolomite in Texas and the Red River Formation of Manitoba as well as from the U.S.S.R. Streptelasmatid horn corals of this type range from the Middle Ordovician to the Silurian and Devonian."

F-49 (*3623-CO*). Corals from black dolomite believed to be Ely Springs near the middle of the formations; Captured Canyon at hill 800 ft in altitude near the mouth of the canyon; 2¾ miles west of SW cor. sec. 18, T. 25 N., R. 1 E., Furnace Creek quad. Identification by W. A. Oliver, Jr., "*Streptelasma*? sp., one specimen; streptelasmatid horn corals, three specimens. This collection may well be Upper Ordovician since streptelasmatids are common in rocks of this age. They are not diagnostic, however, and the age will have to be based on other criteria."

F-74 (*3624-CO*). Middle of Ely Springs Dolomite, ¾ mile south of Trail Canyon; saddle 600 ft west of hill 1932. Identification by W. A. Oliver, Jr.,

"*Bighornia* sp., two specimens; *Grewingkia* sp., one specimen; angulate streptelasmatid, one specimen; other streptelasmatoids, four specimens; small branching bryozoans. This assemblage is certainly Upper Ordovician as the genera *Bighornia* and *Grewingkia* are so limited. These corals are characteristic of the Ely Springs, Bighorn, and Red River Formations in western North America."

At F-75, an isolated hill at the mouth of Trellis Canyon (SE-NE-sec. 23, T. 16 S., R. 45 E.) dolomite thought to be Ely Springs, or possibly Silurian, contains biconvex cross sections.

Receptaculites has been reported from Ely Springs Dolomite at the east foot of the Panamint Range along Trail Canyon (Hopper, 1947, p. 407). The Ely Springs Dolomite was correctly identified by Hopper; but if the *Receptaculites* came from that formation, it is the only recorded occurrence of the genus in the formation in this entire region. Perhaps the fossil came from a fault block of dolomite belonging to the upper part of the Pogonip Group, which contains abundant *Receptaculites*.

The Ely Springs Dolomite is considered to be Late Ordovician in age.

Trace elements in two samples of the Ely Springs Dolomite are given in table 13. The amounts and proportions are about the same as in dolomite from the Pogonip Group.

SILURIAN AND DEVONIAN SYSTEMS—HIDDEN VALLEY DOLOMITE

The Hidden Valley Dolomite, named for exposures in the northern Panamint Range (McAllister, 1952, p. 15) where it is 1,365 feet thick, is a light-colored formation that contrasts strikingly with the dark underlying Ely Springs Dolomite (fig. 28). Throughout the region the Hidden Valley Dolomite is conformable on the Ely Springs Dolomite. At the type locality the upper contact is conformable (McAllister, 1952, p. 15), but in some areas the top of these beds is an unconformity (Hazzard, 1937b, p. 327).

In the Panamint Range, south from Tucki Mountain, the Hidden Valley Dolomite is of variable thickness. It has a computed thickness of 750 feet at the north base of Tucki Mountain between Little Bridge Canyon and Trellis Canyon. It has a computed thickness of slightly less than 600 feet at the south side of Tucki Mountain. In the ridge south of Trail Canyon it has a measured thickness of only 300 feet. In the Funeral Mountains east of Death Valley a thickness of 1,473 feet is indicated (C. A. Richards[4]). I have not determined whether these differences in thickness are due to stratigraphic changes or to crushing and shingling of the beds because of the intense structural deformation.

In this area the formation is light colored, thick bedded (fig. 28), fine grained, and even grained. Many beds contain crinoid stems; fragments of some large

[4] See footnote, p. A35.

FIGURE 23.—View of Ordovician, Silurian, and Devonian formations on the south side of Tucki Mountain. At left is Nopah Formation (Cn) overlain by Pogonip Group (Op). Eureka Quartzite (Oe) forms the light band under the dark Ely Springs Dolomite (Oes). The gray slope above the Ely Springs Dolomite is Hidden Valley Dolomite (DSh); the striped slope is formed by the Lost Burro Formation (Dl). Photograph courtesy of John H. Maxson.

ones are as much as half an inch in diameter. The top of the formation is taken at the base of the first quartzite beds that characterize the lower part of the Lost Burro Formation in this area.

No fossils, except crinoid stems, were found in the Hidden Valley Dolomite, but search was restricted to the Trail Canyon area and the north base of Tucki Mountain.

At the type locality and other nearby places in the northern Panamint Range, fossils in the lower part of the formation include (McAllister, 1952, p. 16–17):

Halysites catenularia (Linnaeus)
microporus (Whitfield)
Favosites cf. *F. niagarensis* Hall
Porpites porpita (Linnaeus)

Fragments of bryozoa and a few brachiopods of Silurian affinities.

Fossils from beds 15–65 feet below the top of the formation in the northern Panamint Range, indicating an Early Devonian age, include (McAllister, 1952, p. 17):

Favosites sp.	Branching *Cladopora*
Papiliophyllum elegantulus (Stumm)	*Heliolites* sp.
Breviphyllum lonensis (Stumm)	*Acrospirifer kobehana* (Merriam)
Unidentifiable cup corals	*Meristella robertsensis* Merriam
	Platyceras sp.

In the Funeral Mountains, on the east side of Death Valley, the lower 200 feet of the formation yielded (C. A. Richards [6]):

Syringopora sp.	Crinoid stems
Plectatrypa sp.	*Rhynchonella* sp.
Heliolites sp.	*Cladopora* sp.
Favosites sp.	*Stromatopora* sp.
Rugose corals	

Richards also reports the following from beds 200–550 feet above the base of the Hidden Valley Dolomite:

Halysites sp.	*Meristella*? sp.
Syringopora sp.	*Eospirifer* sp.
Plectatrypa sp.	Pentameroid brachiopod, *Virgiana*? sp.
Heliolites sp.	Brachiopod fragments
Favosites sp.	Gastropods
Rhynchonella sp.	Rugose corals
Cladopora sp.	Crinoid stems
Stromatopora sp.	
Cornulites sp. on *Syntrophina*? sp.	

From a 10-foot fossiliferous dolomite about 200 feet below the top of the Hidden Valley Dolomite, Richards obtained:

Halysites cf *H. labyrinthica* (Goldfuss)	*Heliolites* sp.
Favosites sp.	Zaphrentid-type corals

The fossils obtained from the formation in areas near Death Valley indicate a Silurian and Early Devonian age.

No samples of the Hidden Valley Dolomite were collected for trace-element analysis.

DEVONIAN SYSTEM—LOST BURRO FORMATION

The Lost Burro Formation at the type locality in the northern Panamint Range is 1,525 feet thick and consists chiefly of light-gray dolomite striped with nearly black dolomite, and limestone with some thin quartzite beds (McAllister, 1952, p. 18). In the Darwin area the thickness of the formation is more than 1,700 feet and may be as much as 2,400 feet (Hall and MacKevett, 1958, p. 8). Only the lower 750 feet is present in the Funeral Mountains, the upper part having eroded (C. A. Richards [6]). In the Nopah Range beds of Devonian age, referred to as the Sultan Limestone, are 1,720 feet thick (Hazzard, 1951, p. 1503).

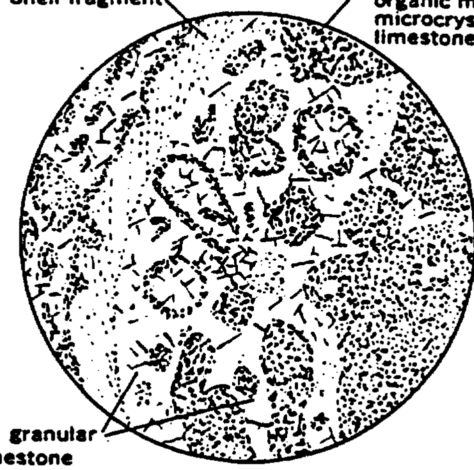


FIGURE 29.—Micrograph of thin section of limestone from Lost Burro Formation. The limestone is mottled with dusty microcrystalline limestone masses separated by more coarsely crystalline clear limestone. Many of the microcrystalline masses have structures suggesting an organic origin. Diameter of field, 2.5 mm.

In the Death Valley area, on Tucki Mountain, the formation has the striped appearance (fig. 28) so characteristic of the type locality. The formation is mostly limestone (fig. 29) with only minor amounts of dolomite and thin beds of sandstone and quartzite. Many of the dark beds are mottled by whitish bodies resembling chopped spaghetti.

An attempt was made to measure a section of the Lost Burro Formation along the north foot of Tucki Mountain eastward from the mouth of Trellis Canyon, but the attempt was only partly successful because of faults. The indicated thickness is about 2,000 feet.

The base, which seems to be of Middle Devonian age, was taken just below a pair of quartzite beds, each about 3 feet thick and separated by 20 feet of carbonate rocks. These quartzites are overlain by 800 feet of alternating light and dark limestone in beds 1–10 feet thick, the striped beds. Above this striped unit is 200 feet of limestone and dolomite with numerous thin beds of sandstone or quartzite. The quartzite, mostly medium grained, occurs in pods and in beds 2 inches to 3 feet thick. Fossil collection F–16 is from the top of this unit.

Above this unit are massive dolomitic beds, but the section appears to be duplicated by faulting. The

[6] See footnote, p. A35.

[6] See footnote, p. A35.

upper 200 feet of the formation consists of well-bedded and even-bedded limestone and quartzite in beds 5 and 6 feet thick. An unknown thickness of beds, but probably not over 500 feet, lies below this unit and the bed represented by fossil collection F-16. The top of the Lost Burro Formation is taken at a quartzite immediately underlying limestone of Mississippian age represented by collection F-93 (p. A44). The top of the Lost Burro Formation seems to be of early Late Devonian age.

Fossils from the base of the Lost Burro Formation were obtained at two locations.

F-18. Base of Lost Burro Formation, light-colored dolomite overlying the lowest quartzite on ridgetop (alt 1,600 ft) 1 mile east of Trellis Canyon, Stovepipe Wells quad. (NE¼NE¼ sec. 26, T. 16 S., R. 45 E.). Report by Jean M. Berdan,

"This collection contains brachiopods referable to *Emanuella* and small specimens of *Atrypa*. Although the genus *Atrypa* has a long range, occurring from the Silurian through the early Upper Devonian, according to Cooper (in Shimer and Shrock, 1944, p. 329) *Emanuella* is indicative of Middle or Upper Devonian. This collection, therefore, is probably Middle or early Late Devonian in age."

F-46. Base of Lost Burro Formation on south slope of ridge 2½ miles north of Trail Canyon and 2¾ miles west of SW cor. sec. 30, T. 18 S., R. 47 E., Furnace Creek quad. Report by C. W. Merriam,

"cyathophyllid rugose coral, deep calyx; *Stringocephalus* sp. cf. *S. burtini* Defrance; indeterminate gastropods, at least two genera. One large individual of *Stringocephalus* is fairly well preserved, showing the characteristic rodlike cardinal process of the dorsal valve. The rock contains abundant fragmentary silicified shell fragments of *Stringocephalus* and other smooth-shelled brachiopods, some of which could be the terebratuloid *Rensselandia*. These shells come out with acid but are not complete enough for positive identification. This collection represents the late Middle Devonian '*Stringocephalus*' zone now recognized widely in the Great Basin."

Other collections of fossils, obtained from the middle or upper part of the Lost Burro Formation, include:

F-4. Probably near middle of the Lost Burro Formation. Limestone butte below the mouth of Trellis Canyon. Stovepipe Wells quad. (SW¼ sec. 13, T. 16 S., R. 45 E.). Report by C. W. Merriam, "stromatoporoids; *Atrypa* cf. *A. missouriensis* fine-ribbed form, abundant; *Tabulophyllum* sp. early Late Devonian."

F-6. Near middle of Lost Burro Formation; ½ mile northeast of mouth of Little Bridge Canyon, Stovepipe Wells quad. (north side SE¼ sec. 10, T. 16 S., R. 45 E.). Report by C. W. Merriam, "stromatoporoids; *Amphipora* sp., *Atrypa* sp. Age: late Middle or early Late Devonian."

F-7. Near F-6. Report by C. W. Merriam, "stromatoporoids; abundant small indeterminate pelecypods resembling the genus *Edmondia; Spirifer* cf. *S. utahensis* Meek; abundant small indeterminate rugose corals with deep calyx. Age: Early Late Devonian."

F-16. Lost Burro Formation, about 1,000 ft above the base; 1½ miles east of mouth of Trellis Canyon, Stovepipe Wells quad. (SE¼NW¼ sec. 24, T. 16 S., R. 45 E.). Report by C. W. Merriam, "stromatoporoids; *Syringopora* sp.; ?*Orecopia mccoyi* (Walcott): indeterminate rugose coral. Age: Early Late Devonian."

Merriam goes on to report, "Rocks represented by coll. F-4, 6, 7, and 16 are seemingly correlative with middle and upper parts of the Devils Gate Limestone of central Nevada."

F-65. Brachiopods and corals from near the top of the Lost Burro Formation; overlies red limy shale and siltstone. In butte isolated from Tucki Mtn., NE¼ sec. 14, T. 16 S., R. 45 E. Mackenzie Gordon has reported as follows:

"Horn corals, genus and species indet.
Stromatoporoid cf. *Stachyodes* or *Idiostroma* sp. indet.
Cyrtospirifer or *Cyrtiopsis* sp."

"W. A. Oliver, Jr., says that the corals are simple types that are known to range through Silurian and Devonian rocks but not diagnostic of any one particular zone. Helen Duncan says that the corals are not Carboniferous types and that the small stromatoporoid is of a type characteristic of Devonian rocks. Jean Berdan confirms my belief that the silicified brachiopods are Late Devonian types and belong in one of the two mentioned genera, though they are not complete enough to be sure which."

F-90. North base, Tucki Mountain. Stromatoporoid reef about 750 ft above the base of the Lost Burro.

Stromatoporoid- and *Amphipora*(?)-bearing beds are particularly abundant near the middle of the Lost Burro Formation (fig. 30). Beds containing numerous brachiopods, including the diagnostic *Cyrtospirifer* (fig. 31), mark the top of the Lost Burro Formation. Syringoporoid corals are present in both Devonian and Mississippian limestones. The Devonian forms can be distinguished from the Mississippian forms on gross morphology (fig. 32) and can be useful field guides for distinguishing formations of these ages.

At the type locality in the northern Panamint Range the uppermost 35 feet of the Lost Burro Formation contains the following Late Devonian fossils (McAllister, 1952, p. 19):

Cyrtospirifer cf. *C. monticola* (Haynes)
disjunctus (Sowerby)
Tylothyris? cf. *T.*? *raymondi* Haynes
"*Camarotoechia*" aff. "*C.*" *duplicata* (Hall)
Cleiothyridina cf. *C. devonica* Raymond
Productella sp.

The Lost Burro Formation as mapped in this area is considered to be Middle and Late Devonian in age.

Trace elements in some samples from Devonian and younger formations are given in table 14. Most of the formations are represented by only a single sample. The samples suggest that the younger rocks have the greater concentrations of elements.

FIGURE 30.—Stromatoporoid (upper) and *Amphipora* (?) (lower) beds are abundant in the middle of the Lost Burro Formation.

FIGURE 31.—Limestone containing *Cyrtospirifer*, which is diagnostic of the uppermost Devonian. A somewhat similar appearing spirifer is present in the lower part of the Tin Mountain Limestone.

TABLE 14.—*Trace elements in Devonian and younger Paleozoic formations*

[Semiquantitative spectographic analyses by Uteana Oda and E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Lost Burro Formation Limestone	Tin Mountain Limestone	Limestone from upper part of the Mississippian	Rest Spring Shale: Limestone	Rest Spring Shale: Shale	Pennsylvanian and Permian formations on Tucki Mountain: Limestone		Pennsylvanian and Permian formations on Tucki Mountain: Altered rock at thrust fault	
Pb	<10	10	<10	<10	10	<10	800	>10,000	>10,000
Mn	50	100	30	100	7,000	20	15	50	10
Cu	7	7	<5	50	10	10	5	>500	100
Zr	<10	<10	<10	20	150	10	10	30	20
Ni	<5	<5	<5	10	100	7	<5	10	<5
Co	<10	<10	<10	<10	30	<10	<10	<10	<10
V	20	20	10	20	70	30	10	10	<10
Y	<10	10	10	15	100	10	<10	10	<10
Be	<1	<1	<1	<1	<1	<1	<1	1	<1
Ti	100	150	10	500	1,500	200	20	300	200
B	<10	<10	10	30	300	<10	10	<150	<100
Sc	<10	<10	<10	<10	10	<10	<10	<10	<10
Ga	<20	<20	<20	<20	<20	<20	<5	10	<5
Cr	<10	10	<10	70	50	100	10	15	10
Ba	10	10	10	30	300	500	15	200	50
Sr	300	1,000	100	700	500	2,000	300	200	300
Mg	0.2	1	0.1	5	1.5	3	0.2	0.1	0.03

NOTE.—Also found: In Sn, <10; Ag, <1; Ge, <20; As, <1,000; Sb, <200; In, <50; Cd, <50; Tl, <100; Ta, <50; W, <50; Nb, <10.

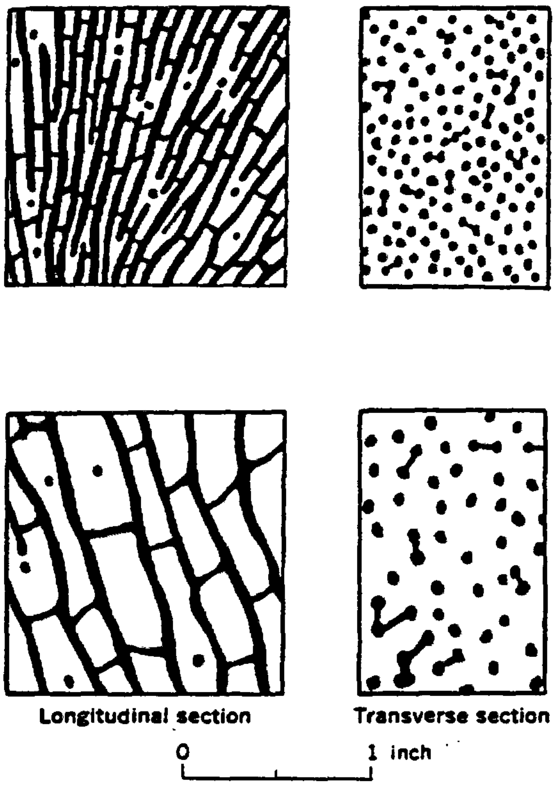


FIGURE 32.—Diagrammatic sections of Devonian (upper) and Mississippian (lower) syringoporoid corals. In the Devonian corals the corallites and connecting tubes are closely spaced; in the Mississippian corals they are widely spaced.

MISSISSIPPIAN SYSTEM—TIN MOUNTAIN LIMESTONE AND YOUNGER LIMESTONE

Mississippian formations in the northern part of the Panamint Range north of the area shown on plate 1, include the Tin Mountain Limestone, 475 feet thick, of Early Mississippian age and the Perdido Formation, about 600 feet thick, of Early and Late Mississippian age. The Tin Mountain Limestone consists largely of thin-bedded dark limestone with thin interbeds of shale. The Perdido Formation contains more shale, siltstone, sandstone, and chert and lithologically is transitional to the overlying Rest Spring Shale (McAllister, 1952, p. 20–24; 1956). Equivalent beds in the Darwin area have about the same thickness (Hall and MacKevett, 1958, p. 8–9). To the east, however, the thickness is much greater, for it is about 1,600 feet in the Nopah Range (Hazzard, 1951, p. 1503).

On Tucki Mountain, in the Death Valley area, about 1,725 feet of limestone ranging from Early to Late Mississippian in age overlies the Lost Burro Formation and is overlain by the Rest Spring Shale. The limestone of Late Mississippian age has little shale, and is sharply separable from, rather than transitional to, the overlying Rest Spring Shale. It was not mapped separately from the older Mississippian limestone, although its base can readily be determined by an absence of chert in the limestone below. The scarcity of shale in the younger limestone formation could be due either to facies changes or to squeezing out of the incompetent beds on account of the structural deformation. The lower limestone assuredly is equivalent to the Tin Mountain, but the younger limestone formation has not yet been correlated with formations in other parts of the region. Accordingly, these limestone beds on Tucki Mountain together are referred to as the Tin Mountain Limestone and younger limestone.

Beds regarded as equivalent to the Tin Mountain Limestone include a lower member, 600 feet thick, consisting of thin-bedded gray limestone with reddish beds of sandstone, and an upper member, 400 feet thick, of thick-bedded dark limestone. The overlying beds, in part at least of Late Mississippian age, consist of a lower cherty limestone, 475 feet thick, overlain by 250 feet of light-gray limestone interbedded with some sandstone and shale.

Several collections of fossils were obtained from various horizons in the formation. They have been reported on by Mackenzie Gordon, Jr., as follows:

In the following collections, for the most part only generic identifications have been made. This is sufficient to give the necessary age determinations and is practically mandatory as so little work has been done on the Carboniferous and Permian paleontology of this region.

In the faunal lists, Helen Duncan has identified the corals and bryozoans, Raymond Douglass the fusulinids, Ellis Yochelson the gastropods, and Mackenzie Gordon, Jr., the rest of the fauna.

F-15 (*17293-PC*). Dark-gray limestone with brown cherty lenses, about 500 ft thick, underlying thinly bedded limestone represented by F-14-58. SE¼SE¼ sec. 24, T. 16 S., R. 45 E., Death Valley, Calif. Collector, C. B. Hunt, 1958.

Homalophyllites sp.	*Spirifer* sp.
Vesiculophyllum? sp.	Spiriferoid brachiopod indet. (juvenile)
Cyathaxonia sp.	*Cyrtina* sp.
Aulopora sp.	*Crurithyris* sp.
Syringopora cf. *S. aculeata* Girty	*Torynifer* sp.
Crinoid columnals	*Hustedia* sp.
Rhipidomella sp.	*Cleiothyridina* 2 sp.
Camarotoechia sp.	*Composita?* sp.
Spirifer cf. *S. centronatus* Winchell	*Platyceras* sp.

This collection contains the typical Early Mississippian fauna found in the Tin Mountain Limestone in this region.

F-21. Crinoidal limestone, 400 ft thick. Underlies red beds unit represented by F-20. Center sec. 31, T. 16 S., R. 46 E., alt 400 ft, Stovepipe Wells quad., California. Collector, C. B. Hunt, 1958.

Crinoid columnals.

No diagnostic fossils found.

F-76 (19135-PC). Tin Mountain Limestone, about 500 ft above the base; north side Tucki Mtn. east of mouth of Trellis Canyon, NE¼SE¼SE¼ sec. 24, T. 16 S., R. 45 E., Stovepipe Wells quad., California.

Cyathaxonia? sp.	*Spirifer* sp.
Rylstonia sp.	*Cleiothyridina* sp.
Aulopora sp.	Compositoid brachiopod indet.
Syringopora sp.	
Crinoid columnals	*Platyceras* (*Orthonychia*) sp.
Rhipidomella sp.	
Camarotoechia sp.	

This is the typical fauna of the Tin Mountain Limestone, as represented also by field colls. F-15 and F-91.

F-91 (19990-PC). About 75 ft above wash on north basal edge of Tucki Mtn., 1½ miles east of Trellis Canyon, in the SE¼SE¼ sec. 24, T. 16 S., R. 45 E., Stovepipe Wells quad., California. Fossils through about 15 ft of gray limestone and dark-brown nodular chert, 60 ft stratigraphically below base of thick reddish-brown cherty limestone unit. Collectors, C. B. Hunt and M. Gordon, Jan. 2, 1961.

Homalophyllites sp.	*Rhipidomella* sp.
Cyathoxonia sp.	*Cranaena?* sp.
Rylstonia? sp.	*Camarotoechia* sp.
Aulopora sp.	Spiriferinid brachiopod, genus and species indet.
Syringopora cf. *S. surcularia* Girty.	*Cyrtina* sp.
Beaumontia sp.	*Spirifer* cf. *S. centronatus* Winchell
Fistuliporoid bryozoan, genus and species indet.	*Hustedia* sp.
Ramiporalia sp.	*Cleiothyridina* sp.
Cystodictya sp.	*Platyceras* sp.
Crinoid columnals	*Gattendorfia?* sp.
Chonetes sp.	Trilobite pygidium
Marginatia? sp.	
Small spinose productid, genus and species indet.	

This and the following collection are Lower Mississippian typical of the fauna found in the Tin Mountain Limestone.

F-92 (19991-PC). *Syringopora* in place about 3 ft above base of outcrop, 150 ft southwest of F-91 and about 100 ft stratigraphically lower. Collectors, M. Gordon and C. B. Hunt, Jan. 2, 1961.

Syringopora cf. *S. aculeata* Girty

F-93 (19992-PC). 500 ft west of F-91 at north foot of Tucki Mtn., 1½ miles east of the mouth of Trellis Canyon, SE¼SE¼ sec. 24, T. 16 S., R. 45 E., Stovepipe Wells quad., California. 10-foot zone of gray limestone about 10 ft stratigraphically above brown quartzite at top of Lost Burro Formation. Collectors, C. B. Hunt and M. Gordon, Jan. 1, 1961.

Rylstonia? sp.	*Avonia?* sp.
Dissepimented horn coral, indet.	*Camarotoechia?* sp. indet.
Syringopora sp.	*Spirifer* aff. *S. latior* Swallow
Syringopora cf. *S. aculeata* Girty	*platynotus* Weller
Crinoid columnals	*Cleiothyridina* sp.
Chonetes sp.	*Platyceras* sp.

This collection is Early Mississippian in age. The rocks from which it was collected represent the basal part of the Tin Mountain Limestone.

F-96 (19995-PC). Limestone bed 50 ft below top of the beds mapped as Tin Mountain Limestone and younger limestone. Collectors, M. Gordon and C. B. Hunt, Jan. 3, 1961.

Faberophyllum sp.	Fish dentition
Syringopora sp.	

Faberophyllum in the Rocky Mountain region, according to Helen Duncan, characterizes a zone that occurs approximately in the middle of the Upper Mississippian. It is not known to occur in lower Meramec or upper Chester equivalents. It is found, for example, in the upper part of the Humbug Formation and lower part of the Great Blue Limestone in central Utah. This would indicate that the upper beds of the Mississippian limestone sequence on Tucki Mountain are middle Late Mississippian in age.

The following is a section of the Mississippian formations, measured along the north foot of Tucki Mountain 3 miles northwest of Shoveltown.

Section of Tin Mountain Limestone and younger limestone 3 miles northwest of Shoveltown

	Feet
Top. Rest Spring Shale; contact much deformed by bedding faults. Limestone, younger than the Tin Mountain in part of Late Mississippian age.	
Light-gray partly recrystallized encrinital limestone interbedded with sandstone, shale, and thin-bedded limestone. F-96 (19995-PC) from 50 ft below top	250
Mostly dark-gray fine-grained limestone and brown weathering chert in beds 2-6 in. thick; at base is 30-40 ft of banded and bedded chert that weathers in gray blocks	475
Tin Mountain Limestone:	
Upper member: mostly thick-bedded limestone; at base is 40 ft of dark fossiliferous limestone, F-76 (19135-PC)	365
Lower member: upper 100 ft is thin-bedded gray limestone interbedded with reddish sandstone; lower 100 ft includes some black limestone in thick beds but containing chert partings and lenses parallel to the bedding F-15 (17293PPC) and F-91 (19990-PC), 300 ft above base; F-92 (19991-PC) is 250 ft above base; F-93 (19992-PC) is from the base	500
Total thickness	1,590
Base. Top of Lost Burro Formation.	

Micrographs of the limestone from the Tin Mountain Limestone and younger limestone are illustrated on figure 33.

Compared to other late Paleozoic formations, trace elements in a specimen of Tin Mountain Limestone were intermediate in amount, and least in a specimen from the limestone above the Tin Mountain (table 14).

MISSISSIPPIAN AND PENNSYLVANIAN(?) SYSTEMS—REST SPRING SHALE

The type locality of the Rest Spring Shale is in the northern Panamint Range, where the formation is largely shale and siltstone but includes some sandstone

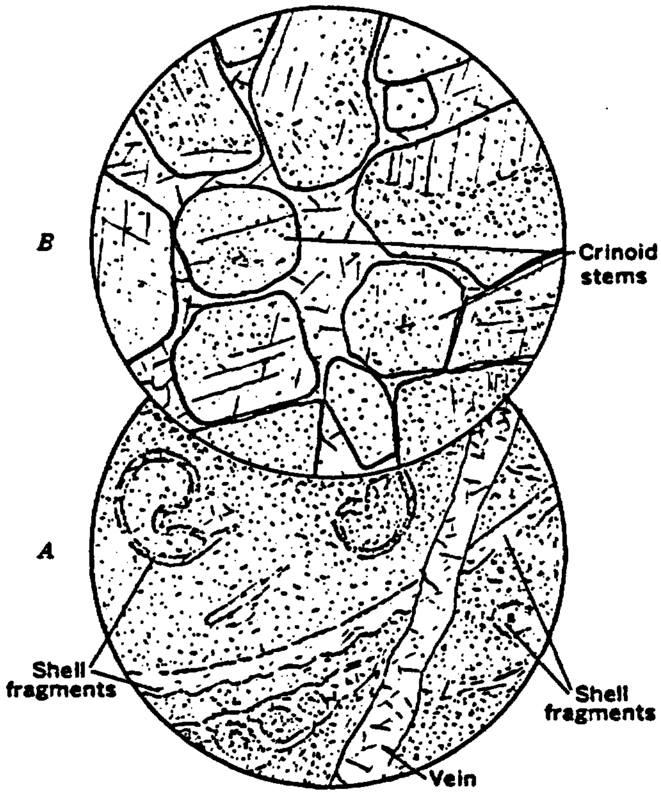


FIGURE 33.—Micrographs of limestone from Tin Mountain Limestone (*A*) and unnamed younger Mississippian formation (*B*). The limestone from the Tin Mountain is a bioclastic bed with many shell fragments set in a matrix of dusty (organic rich?) and more finely microcrystalline limestone. The limestone from the unnamed younger Mississippian formation is coarsely granular. Organic structures, probably large crinoid stems, are obscured as a result of the recrystallization of the limestone. Diameter of field, 2.5 mm.

and vitreous quartzite (McAllister, 1952, p. 25). The thickness at the type locality is uncertain (perhaps 400 ft) because the formation is incompetent and is greatly pinched and swollen in zones of folds and faults (McAllister, 1952, p. 26).

In the area covered by this report the beds referred to the Rest Spring Shale are restricted to a narrow strike valley half a mile from the east base of Tucki Mountain. The width of the outcrop belt is about 750 feet; the beds are practically vertical. This indicates a thickness for the formation of 750 feet, but as at the type locality, this thickness is uncertain because the formation is greatly crushed between the more competent underlying and overlying formations.

The formation is mostly light-gray to black shale and siltstone interrupted with beds of black limestone 1–5 feet thick. The beds contain rounded nodules, like golf balls, of dark chert (fig. 34). At the base is 35 feet of black shale; the beds above this are mostly shale and siltstone. Some of the siltstone is ripple marked. Poorly preserved brachiopods are common in the shale, particularly toward the base. Near the middle of the formation is 50 feet of dolomitic and dark cherty limestone (fig. 35). The upper half of the formation contains increasing limestone and calcareous cement in siltstone. The siltstone and limestone weather reddish. At the top is 15–30 feet of shale, crushed under coarse conglomerate of the Pennsylvanian and Permian formations.

Trace elements in the Rest Spring Shale are given in table 14.

Fossils collected at several places across the formation where it is exposed in the previously described strike valley, have been reported on by Mackenzie Gordon, Jr., and Helen Duncan, as follows:

F–20 (*17294–PC*) Center sec. 31, T. 16 S., R. 46 E., alt. 400 ft. Top of ridge on north side of first canyon north of Shoveltown. Brachiopods, etc., from crumbly limestone and red-bed unit, about 800 ft thick; overlies light-gray crinoidal limestone. Collector, C. B. Hunt, 1958.

Flexaria sp.
Inflatia sp.
Rhipidomella nevadensis (Meek)
Cranaena? sp.
Schizophoria? sp. indet.
Reticulariina campestris (White)?
Spirifer sp.

FIGURE 34.—Golfball-like nodules of dark chert are a characteristic lithologic feature of the Rest Spring Shale and basal part of the overlying limestone of Pennsylvanian age.

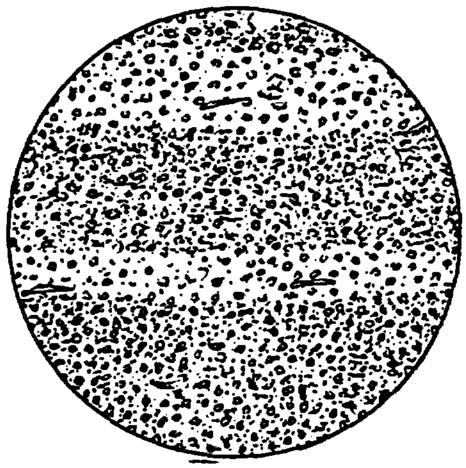


FIGURE 35.—Micrograph of limestone from Rest Spring Shale. The limestone is very fine grained, grain size 25 microns. The grains are mostly of calcite, many of them euhedral. About 10 percent are quartz, and 1 or 2 percent are iron oxide. The grains are set in a calcitic paste darkened with organic matter and (or) clay. Shell fragments are few; there are some circular structures, apparently composed of silica. Diameter of field, 2.5 mm.

In the Carboniferous rocks of the Great Basin, *Rhipidomella nevadensis* (Meek) is found in the uppermost Upper Mississippian and the lowermost Lower Pennsylvanian. The assemblage here does not contain any undoubted Pennsylvanian forms but has, rather, a Mississippian aspect. The containing beds are probably equivalent roughly to some part of the upper two or three hundred feet of the Chainman Shale as typically developed in east-central Nevada and western Utah.

F-77. Rest Spring Shale, limestone beds near middle of unit. 2 miles northwest of West Side Borax Camp (Shoveltown), center sec. 31, T. 16 S., R. 46 E., Stovepipe Wells quad., California.

Chonetid brachiopod indet.	*Spirifer* cf. *S. arkansanus* Girty?

F-78. Same location as F-77; from basal shale of Rest Spring Shale; 0-35 ft above the base.

Lissochonetes? sp.	Fish scale
Spinose productid brachiopod	

F-80. Same location as F-77. Shaly beds, 700 ft above the base of the Rest Spring Shale.

Lissochonetes? sp.	Compositoid brachiopod indet.
Spiriferoid brachiopod cf. *Torynifer*	

The collections from the Rest Spring Shale, above, are poorly preserved and do not permit determination as to possible Mississippian or Pennsylvanian age. The smooth chonetids suggest a Pennsylvanian age for the rocks, but some of the other brachiopods suggest Mississippian forms.

F-81 (17296). Same location as F-77, cherty limestone.

Rhomboporoid bryozoan indet.	Small spinose productid brachiopods
Lissochonetes? sp.	*Conocardium* sp.
Small Linoproductid brachiopod	*Hyolithes* sp.

This is the best preserved collection from the Rest Spring Shale. Although the forms are not strictly diagnostic of either Pennsylvanian or Mississippian age, the collection is notable for the presence of a number of well-preserved specimens of *Hyolithes*, a long-ranging early Paleozoic genus that is rare in the late Paleozoic.

F-97 (19996). Bottom of same gulch as F-77, possibly from the same bed, about 750 ft to south. Rest Spring Shale, near middle.

Button coral?	Pleurotomariid gastropod indet., with spiral ornament
Chonetes? sp.	
Echinocoelia sp.	

This unusual assemblage is a facies faunule expressing the special conditions of deposition of the shale sequence. These forms are not found commonly enough to be clearly indicative of either Mississippian or Pennsylvanian age. The brachiopod *Echinocoelia* is known elsewhere in Devonian and Mississippian rocks. To the best of our knowledge it has not been reported from rocks of Pennsylvanian age.

In a comment on these collections, Gordon states,

The Rest Spring Shale on Tucki Mountain appears to be the same as that in the Quartz Spring area. Diagnostic fossils, however, are few. In the type area the Rest Spring has *Cravenoceras merriami* Youngquist in the basal few feet. This shows it to be Late Mississippian in age, at least in the lower part. The one datable fauna in the Rest Spring on Tucki Mountain (F-20) is very late Mississippian in age; the fauna with *Rhipidomella nevadensis* (Meek) that elsewhere in the Great Basin is found a short distance above the beds with *Cravenoceras merriami*. The fossils came from the middle of the shale. Whether or not the upper part of the Rest Spring on Tucki Mountain is of Pennsylvanian age is not presently known.

PENNSYLVANIAN AND PERMIAN SYSTEMS—FORMATIONS AT EAST FOOT OF TUCKI MOUNTAIN

Formations at the east foot of Tucki Mountain include thin-bedded limestone, at least 2,500 feet thick, largely, if not wholly, of Pennsylvanian age, and limestone conglomerate, at least 3,000 feet thick, that is largely, if not wholly, of Permian age.

The formations crop out in a belt about 3,000 feet wide along the east foot of the mountain. They stratigraphically overlie the Rest Spring Shale; but the contact must be a fault, because towards the south the shale is overlain by the limestone of Pennsylvanian age, whereas towards the north it is overlain by the conglomerate which is Permian.

These formations are at the lower edge of the upper plate of the Tucki Mountain thrust fault, and they are turned up steeply and even overturned. They rest on Precambrian quartzite in the lower plate of the thrust fault. Because of the deformation, the thicknesses cited are little better than guesses. Intensive detailed work will be required to resolve the many uncertainties in the stratigraphy of the formations, and until that

FIGURE 36.—Thin-bedded limestone and dolomite near the base of the unnamed formation of Pennsylvanian age at the east foot of Tucki Mountain. These beds are overturned; their base is to the right.

work is done, it seems best to adopt an informal nomenclature.

At the south end of the outcrop of these formations the lowest exposed beds are thin-bedded limestone with some dolomite (fig. 36), which are estimated to be about 1,000 feet thick. These beds, which are vertical or overturned and lie against crushed, faulted, and contorted Rest Spring Shale, probably are Pennsylvanian in age (see colln. F–85, 87). Northward this belt of limestone narrows and in its place occurs coarse limestone conglomerate. The northward thinning of the limestone probably is due to faulting. Mountainward from Shoveltown (at Salt Springs) and farther north the formations are very largely conglomerate with comparatively thin lenses of limestone (fig. 37) and are mostly of Permian age (see colln. F–95 (19994–

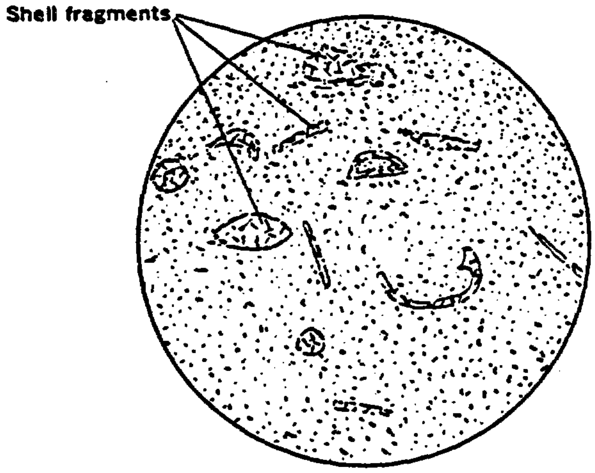


FIGURE 37.—Micrograph of limestone from the unnamed formation of Pennsylvanian age that overlies the Rest Spring Shale. The limestone is uniformly microcrystalline (grain size, 25 microns) and consists of interlocking anhedral grains of calcite and almost no quartz or iron oxide. Fragments of shells in the paste are more coarsely crystalline. Diameter of field, 2.5 mm.

PC)). The deformation, though, is too intense to be certain which lenses of limestone are interbedded with the conglomerate and which are faulted in with it.

The conglomerate consists of cobbles and small boulders as much as 1 foot in diameter. They are mostly of limestone and dolomite, but they include some quartzite. The beds of conglomerate are 5–10 feet thick; the bedding is indistinct except for differences in size of the cobbles. In most beds the cobbles, which are well rounded, touch one another, and in many beds there is almost no matrix. The conglomerate weathers reddish, perhaps because of alteration along closely spaced zones of shearing and crushing.

Trace elements in samples from these formations are given in table 14. The two samples of "altered rock at thrust fault" are from a prospect at the southeast corner of Tucki Mountain where the late Paleozoic formations form the upper plate of a thrust and rest on Stirling(?) Quartzite.

Several collections of fossils were obtained from the limestone of Pennsylvanian age south of Shoveltown.

F-84. Formation ¼ mile southwest of Shoveltown; limestone bed in a shaly unit; estimated 2,500 ft above base of the formation. Alt 200 ft, 0.9 mile south-southwest of SW cor. sec. 33, T. 16 S., R. 46 E., Chloride Cliff quad., California. Reported on by Mackenzie Gordon, Jr., as follows:

"Horn coral fragments, indet.

"This collection is too poor to be of value in dating the containing bed."

F-85. Fusulinids and silicified brachiopods from base of the shale and limestone represented by F-84. According to R. C. Douglass, this sample contains *Fusulinella sp.* and represents Middle Pennsylvanian age.

F-87 (19133-PC). Bryozoa, from limestone estimated 1,000 ft thick in the canyon 1 mile south of Shoveltown; fossils are from the middle of the unit, probably 750 ft above the base of the limestone. Examined by Helen Duncan who reports as follows:

"The sample collected contains a good deal of echinoderm and bryozoan debris. Preservation of the bryozoan material is poor, owing to partial silicification and recrystallization. Most of the fragments that can be identified as to genus are *Fenestella*. So far as one can tell from the types of meshwork noted, several species of the genus are probably represented, but structural details are not sufficiently well preserved to furnish data for specific identification. A few examples referable to *Polypora* and *Cystodictya* were found as well as an indeterminate rhomboporoid and a laminar fistuliporoid.

"The bryozoan material does not furnish objective evidence for precise dating. The assemblage of genera might occur anywhere from Devonian to Permian. However, bryozoans are not known to be prevalent in the Devonian and Lower Mississippian in this part of the Great Basin; and the generic association is one that commonly occurs in the later Mississippian and Pennsylvanian. The absence of genera (*Rhombotrypella*, *Rhomboporella*, *Ascopora*) that are diagnostic of and very common in the Middle and Upper Pennsylvanian of the Great Basin suggests that the interval may be older—Early Pennsylvanian or Late Mississippian. With the exception of one genus, which is not present in your material, I have not yet discovered criteria that can be used to distinguish Late Mississippian from Early Pennsylvanian bryozoan faunas. On the evidence available, a Pennsylvanian(?) assignment seems to be a reasonable one."

Collections of fossils from the conglomerate in the northern part of the outcrop belt indicate that it is mostly of Permian age and that it contains cobbles ranging in age from Early Mississippian to Late Pennsylvanian. The collections were studied by Mackenzie Gordon and Helen Duncan, who reported on them as follows:

F-23 (17295-PC). Corals, brachiopods, etc., from limestone breccia, probably high Carboniferous. The hill back of Shoveltown ½ mile southwest of SW cor. sec. 33, T. 16 S., R. 46 E., Chloride Cliff quad., California. Collector, C. B. Hunt, 1958.

Horn corals	Spirifer sp. indet.
Cyathaxonia sp.	Syringothyroid indet.
Syringopora cf. *S. surcularia* Girty	*Hustedia* sp.
	Platyceras sp.
Crinoid columnals	Fish dentition

"This collection no doubt represents a block or blocks of Ti[illegible] Mountain Limestone of Early Mississippian age incorporate[illegible] in the conglomerate sequence."

F-86 (19136-PC). Same location as F-84; float on hillsid[illegible] derived from weathering of 500-ft-thick conglomerate wes[illegible] of, and probably faulted against, the block of Pennsylvania[illegible] rocks represented by F-84 and F-85. This float represent[illegible] cobbles from the conglomerate, and the cobbles may hav[illegible] a wide range in age. Age of cobbles should be considere[illegible] separately.

Bradyphyllum sp.	*Spirifer* fragments indet.
Neokoninckophyllum? sp.	*Crurithyris* sp.
Zaphrentoid coral indet.	*Peruvispira* sp.
Multithecopora? sp.	*Platyceras* sp.
Goniocladia sp.	Gastropod fragments indet.
Crinoid columnals	Trilobite fragments indet.
Productid fragments indet.	

Another block of crinoidal limestone contains the following:

Wedekindellina? sp.

"The last block containing small fusulinids that either represent an advance form of *Fusulinella* or an early form of *Wedekindellina* indicate, according to Raymond Douglass, a Middle Pennsylvanian (Des Moines) age for the limestone block.

"The major part of the list includes forms that are found both in Pennsylvanian and Permian rocks. According to Helen Duncan, the bryozoan genus *Goniocladia* generally indicates Permian age for the rock. *Peruvispira* is, according to Ellis Yochelson, limited to rocks of Permian age and particularly to the Lower Permian of present Geological Survey usage. Part of the fauna is the same as that found in coll. F-95, and a Lower Permian (Wolfcamp) age for it is therefore indicated.

F-94 (19993-PC). Saddle between Shoveltown Ridge and Tucki Mtn., southwest corner Chloride Cliff quad., California. Scattered loose blocks. Collectors, M. Gordon and C. B. Hunt, Jan. 3, 1961.

Fusulinella? sp.	Crinoid columnals
Caninia sp. indet.	*Derbya* sp.
Fistuliporoid bryozoan, genus and species indet.	*Linoproductus* sp.
	Juresania sp.
Fenestella? sp.	*Rhynchopora* sp.
Rhabdomeson? sp.	*Spirifer rockymontanus* Marcou
Prismopora sp.	

Another loose block contains—

Triticites sp.

"As these are loose blocks from a conglomerate, they may be, and in fact are, in part, of different ages. For this reason the one block with *Triticites* is recorded separately. Raymond Douglass says that this species is definitely of Late Pennsylvanian age and probably of Virgil age. The fossils in the list above are Middle Pennsylvanian in age, possibly Atoka. The *Fusulinella*? is associated with the poorly preserved caninoid corals. The *Prismopora*, *Linoproductus*, and *Spirifer* are associated in a single block.

"Of the bryozoans, Miss Duncan reports as follows: 'The only potentially significant bryozoan is *Prismopora*. This genus ranges from the Devonian to the Permian, but in the West its occurrence seems to be confined to the Pennsylvanian and largely to the Middle Pennsylvanian. *Prismopora* is a common fossil in the Middle Pennsylvanian of the Rocky Mountain region, but it is apparently rare in Pennsylvanian assemblages from the Great Basin. In fact, I believe that this is the first time that I have seen the genus in a collection made west of the Wasatch and Oquirrh Mountains in Utah, and none are recorded in Girty's reports on Carboniferous faunas from Nevada and California.'

"This collection is of particular significance because it indicates that rocks as young as probable Virgil age are incorporated as boulders in this conglomerate."

F-95 (19994-PC). Limestone and dolomite beds making up a unit 350 ft thick the east foot of Tucki Mtn., on spur about 1,000 ft southeast of SW. cor. sec 31, T. 16 S., R. 46 E., Stovepipe Wells quad., California. Fusulinids are from zone 50–100 ft below top of 350-ft section. Collectors, M. Gordon and C. B. Hunt, Jan. 3, 1961.

Triticites sp.	Echinoconchoid productid fragments
Pseudofusulinella sp.	*Kochiproductus*? sp. (fragments)
Caninia sp.	*Horridonia*? sp.
Multithecopora? sp.	*Rhynchopora* sp.
Trepostomatous bryozoan, genus and species indet.	*Composita*? sp.
Fistuliporoid bryozoan, genus and species indet.	*Peruvispira* sp.
Crinoid columnals	*Omphalotrochus* sp.
Small strophomenoid brachiopod, genus and species indet.	

"Although the Rest Spring Shale on Tucki Mountain appears to grade upward into the overlying conglomerate beds, somewhere there must be a considerable hiatus. This is because the present evidence indicates that the conglomerate was laid down largely in Early Permian time, although deposition may have begun during Late Pennsylvanian. The fossils collected in place in nonconglomerate beds are of probable Wolfcamp age. Loose blocks from the conglomerate or in float on its surface contain fossils as late as probable Virgil (very Late Pennsylvanian) age. The conglomerate, therefore, appears to be related to the Permian Owens Valley Formation of the Inyo Range, which likewise contains conglomerate beds, although temporally it may be more directly equivalent with the upper beds of the Keeler Canyon Formation, which underlies the Owens Valley.

"Rocks of Bird Spring age, a formation which is roughly equivalent to the Keeler Canyon, are incorporated as blocks in the conglomerate on the east base of Tucki Mountain. The uppermost Bird Spring may be in part temporally equivalent to the conglomerate. However, it would be misleading to use the name Bird Spring for the conglomerate as the type Bird Spring is a well-bedded limestone sequence. The conglomerate on the other hand appears to be tied in with an orogenic belt that lay to the west and north.

"The fusulinids *Triticites* and *Pseudofusulinella* are forms, according to Raymond Douglass, that occur in rocks of late Virgil to Wolfcamp age. The occurrence of the snails *Omphalotrochus* and *Peruvispira* in association is, according to Ellis Yochelson, indicative of Wolfcamp age. *Omphalotrochus* has been found also in the *Uddenites* zone of Texas, considered by some geologists to be late Virgil in age. *Peruvispira*, however, has not yet been found as low as the *Uddenites* zone but occurs also in post-Wolfcamp rocks. On these grounds an Early Permian (Wolfcamp) age for the containing rocks is indicated by this collection. However, the possibility of a very Late Pennsylvanian age cannot be definitely excluded."

Mackenzie Gordon (written commun., 1961) has added the following comment about the significance of these collections:

The collections from the great carbonate rock conglomerate sequence cast considerable light upon its possible age. The collection of loose blocks from the saddle north of Shoveltown Ridge (F-94) include those of Middle Pennsylvanian and one of Late Pennsylvanian, probably Virgil age. The collection (F-95) from the 350-foot section of bedded limestone and dolomite about 1,500 feet above the base of the formation is probably Lower Permian (Wolfcamp) in age. These collections indicate that the conglomerate did not begin to be deposited until very Late Pennsylvanian or perhaps not until Early Permian time. The nearest contemporaneous equivalent is probably the upper part of the Keeler Canyon Formation, although the conglomeratic aspect is more like that of the overlying Owens Valley Formation.

As noted by Gordon, above, conglomerate like that at the east foot of Tucki Mountain occurs west of Death Valley, but to the east the Pennsylvanian and Permian formations are limestone with shale. In the Goodsprings quadrangle, Nevada, the Bird Spring Formation, comprising a sequence of well-bedded limestone 2,500 feet thick (Hewett, 1931, p. 21–30), is Pennsylvanian in age but is thicker in the Las Vegas quadrangle, where it includes beds of Permian age (Helen Duncan, written commun., 1964). The name was extended to an incomplete section of similar rocks of Pennsylvanian age in the Nopah Range (Hazzard, 1937b, p. 337) and to beds of Pennsylvanian and Permian age aggregating at least 5,000 feet thick in the northern part of the Panamint Range (McAllister, 1956). Merriam and Hall (1957) further revised the age of the Bird Spring to Late Mississippian to Permian.

The beds of Pennsylvanian and Permian age in the northern Panamint Range, according to McAllister (1956), consist largely of limestone and shale, but locally there is a lenticular conglomerate several hundred feet thick. In the southern part of the Panamint Range, in Butte Valley 35 miles south of Tucki Mountain, Permian rocks, 4,100 feet thick, also consist of limestone and shale, but they are apparently without conglomerate (Johnson, 1957, p. 382–383). The dominance of limestone and scarcity of shale in the formations at the east foot of Tucki Mountain may be due to squeezing out of the incompetent beds because of the severe deformation there.

Westward from Death Valley the formations equivalent to those at the east foot of Tucki Mountain are increasingly conglomeratic (Merriam and Hall, 1957, p. 4; Hall and MacKevett, 1958, p. 9–10; Bowen, 1954, p. 36–42).

The relationships of the conglomerates northward and southward along the Panamint Range and eastward and westward from there need study, not only for dating the onset of the deformation they record but for determining whether some of the mountain blocks have been moved long distances from their original position. As brought out in the section on "Structural geology" (p. A142), the Pennsylvanian and Permian formations at the east foot of Tucki Mountain have been thrust some miles westward along the Tucki Mountain thrust fault. Perhaps the position from which the fault moved them was not their original position either.

These conglomerates of Permian age are intermediate in type between those of the older Paleozoics and those of the Triassic in this region.

TRIASSIC SYSTEM

Rocks of Triassic age are not represented in the area covered by this report, but they are exposed in Warm Spring Canyon and Butte Valley only 1 mile to the south. There carbonate rocks of Permian age are overlain by metasediments and volcanic rocks aggregating 8,000 feet thick and containing fossils suggestive of an Early Triassic age (Johnson, 1957, p. 384–388). Triassic sedimentary rocks largely of volcanic origin also are known farther west, in the Inyo Mountains (Kirk, in Knopf, 1918, p. 47–48), to the south in Soda Mountains (Grose, 1959, p. 1523), and to the southeast in the Providence Mountains (Hazzard, 1937a, p. 329).

The Triassic rocks record a striking change in kind of sediments being deposited and in the location and trend of the troughs in which the sediments were deposited. The Triassic troughs apparently trended northwest at about the position of the Precambrian trough, and these may have become the site of the Sierra Nevada batholith. The troughs containing the Paleozoic carbonate formations trend north into the Great Basin.

CRETACEOUS OR TERTIARY SYSTEMS

GRANITIC INTRUSIONS

Death Valley is at the northeast edge of a belt of isolated granitic intrusions like the granitic plutons composing the Sierra Nevada batholith and are evidently eastern satellites of it. Three granitic intrusions in the Death Valley area, referred to as the granites at Skidoo and at Hanaupah Canyon, and another in the Amargosa thrust complex, seem to be floored intrusions that spread laterally, probably along thrust faults, and made the space they occupy by doming the upper plates of the thrust. These intrusions are mostly quartz monzonite and represent one of the more potassic types of the batholith. They are younger than the batholith and are closely related to, and only slightly earlier than, the volcanism in Death Valley, which is Tertiary.[7]

These intrusions are described on page 120ff.

[7] Since this manuscript was prepared, lead-alpha age determinations on zircon from some of these intrusions, and potassium-argon determinations on their feldspars, indicate that these intrusions probably are Middle Tertiary (T. W. Stern, written commun., 1965).

CHAOTIC COMPLEX ALONG THE AMARGOSA THRUST FAULT

The relation of the granites to the thrust faults and to the volcanics is best shown along the east foot of the Panamint Range between Starvation Canyon and Blackwater Wash, where a complex of many kinds of igneous, metamorphic, and sedimentary rocks occurs along the outcrop belt of a thrust fault. The fault is probably the westward extension of the Amargosa thrust, and the rocks along it are referred to informally as the chaotic complex along the Amargosa thrust fault, or, for short, as the Amargosa thrust complex; they are described on page 129 ff.

TERTIARY SYSTEM

The volcanic rocks in the Amargosa thrust complex closely resemble some of the Tertiary formations to the east and north of Death Valley and are probably part of the same volcanic sequence. The Tertiary formations east of Death Valley include the "chaos" and related formations along the Amargosa thrust fault at the south end of the Black Mountains (Noble, 1941) and a thick sequence of volcanics that grade northward into playa deposits at the north end of the mountains. The playa deposits and interbedded volcanics dip northward off the Black Mountains and under a syncline separating the Black and Funeral Mountains. Where the formations rise again on the flank of the Funeral Mountains, they resemble the Titus Canyon Formation (Oligocene) of Stock and Bode (1935) and are tentatively correlated with it. The youngest of the Tertiary deposits are mostly playa deposits and are of Pliocene age.

FORMATIONS IN THE BLACK MOUNTAINS

Tertiary formations in the Black Mountains have been studied and reported on by Noble (1941), Curry (1939, 1954), and Drewes (1963), and by J. F. McAllister (written commun., 1960). These formations were given little attention by me, and the descriptions that follow are summarized from the work of the others.

At the south end of the Black Mountains, in the Virgin Springs district (Noble, 1941), the Tertiary rocks include conglomerate, fanglomerate, shale, freshwater limestone, rhyolite, quartz latite, andesite, and tuff. These rocks originally were part of a volcanic and sedimentary sequence like that at the north end of the Black Mountains, but they now are masses in a breccia, the "chaos," along the Amargosa thrust fault. With these Tertiary rocks are blocks of late Precambrian sedimentary formations and breccia of Tertiary(?) granite, and the whole has been thrust onto Precambrian gneiss and other metamorphic rocks.

Individual blocks in the breccia may be as long as 1,000 feet and as thick as 300 feet. They are thrust one on the other, giving a shingled structure that dips east. The thrust was towards the west—towards the Panamint Range.

Tertiary formations at the north end of the Black Mountains have an aggregate thickness of at least 13,000 feet. They overlap and are faulted against Precambrian gneiss at a turtleback above Badwater (fig. 38). The Tertiary formations consist of volcanic rocks and sedimentary rocks. The percentage of sedimentary rocks increases northward.

The Precambrian gneiss under the turtleback surface is cut by a swarm of Tertiary dikes (Curry, 1954) trending N. 30° E. In places their volume exceeds that of the gneiss. These dikes do not extend into the upper plate. They are restricted to the gneiss underlying the turtleback surface and seem to be sheared-off feeders to volcanic rocks that once reached higher levels (Curry, 1954, p. 57).

The Tertiary formations north of the Badwater turtleback can be divided into an older set of deposits, the Artist Drive Formation[a] of Noble and Wright (1954) (Oligocene(?) to early Pliocene) and a younger set, the Furnace Creek Formation (Pliocene). For about 2 miles north from the turtlebacks, the Artist Drive of Noble and Wright is largely volcanic rocks, mostly felsitic flows, lapilli tuffs, and some explosion breccias (fig. 39). The lower 4,000 feet is dark-purple felsite; the upper 1,000 feet is variegated felsite and white tuffs. Northward the volcanic members thin and the sedimentary members thicken. A partial section across the upper 1,000 feet of the formations, where they consist of both volcanic and sedimentary units, shows the following:

Partial section, upper part of formations in the Artists Drive area at divide above Artists Palette and above the head of 20 Mule Team Canyon

Top. Tan playa beds.	
Conglomerate; includes cobbles of granite, upper Paleozoic rocks, and volcanics	50
Basalt	125
Purple and green, "calico" beds; mostly volcanic-derived playa deposits	100
Tan playa beds	125
Red felsite	250
Basalt	75
Red felsite and greenish; probably volcanic-derived playa beds	300
Tan playa beds	250
Base. Basalt and other dark layers interbedded with tan playa beds.	

[a] The correct place name, as approved on official maps of the U.S. Geological Survey, is Artists Drive. Preferred local usage (the guiding rule in adopting place names) is in the possessive, Artist's Drive. As used originally by Noble (1941), "Artist" is in the singular; consequently, the formal formation name retains this spelling.

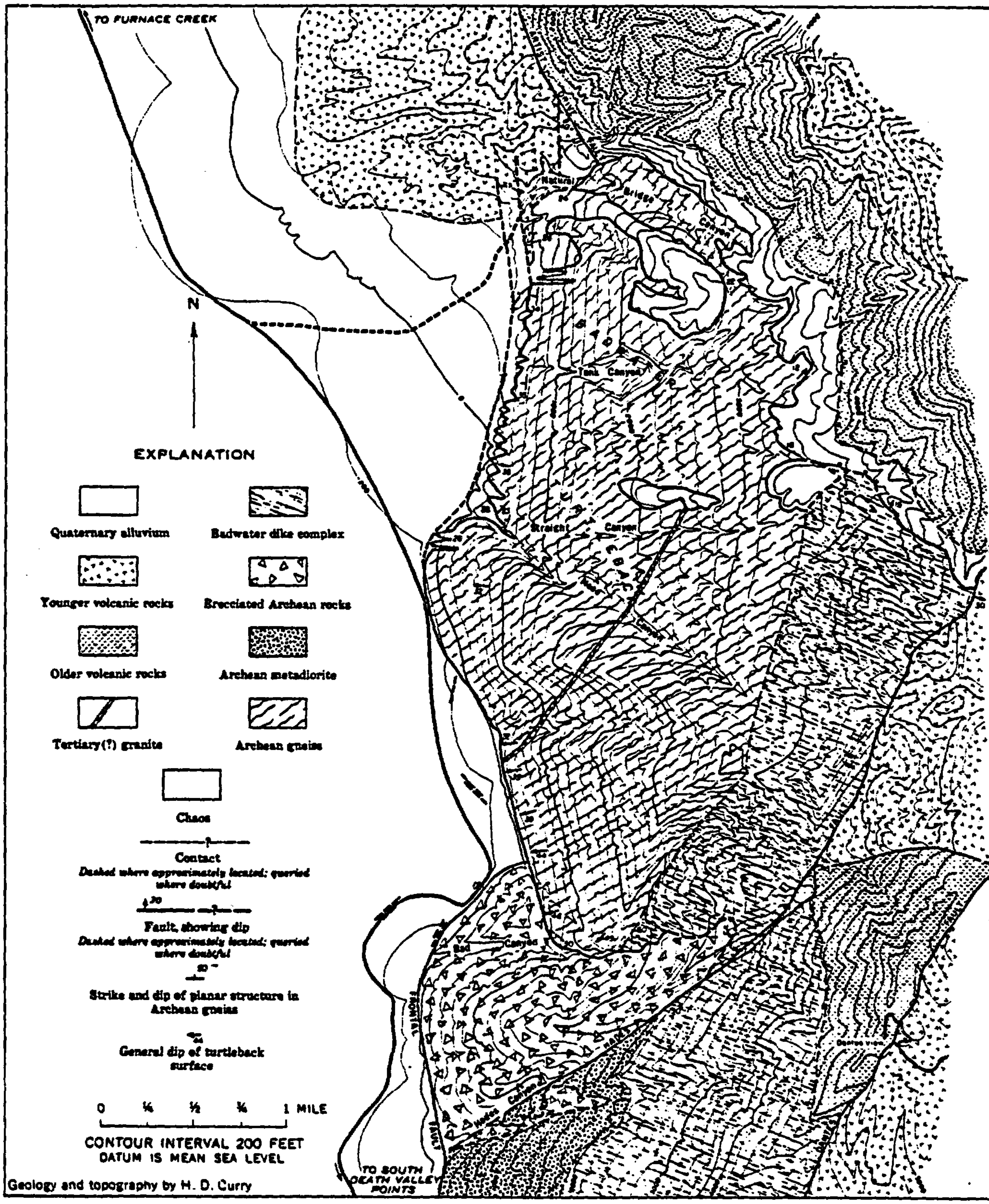


FIGURE 35.—Geologic map of Badwater turtleback and adjacent areas. From Curry, 1954.

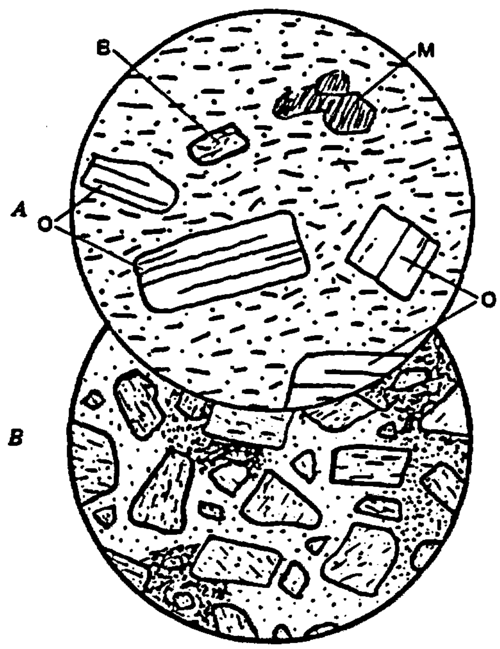


FIGURE 39.—Micrographs of thin sections of felsite (*A*) and explosion breccia (*B*) from the Artist Drive Formation. The felsite has phenocrysts of oligoclase (O), magnetite (M), and a much-altered mineral, probably originally biotite (S) in a glassy groundmass containing microlites of oligoclase and tiny specks of finely disseminated iron oxide which colors this felsite red. The explosion breccia consists of sharply outlined randomly oriented fragments of glassy rocks in a glassy matrix spotted and streaked bright red and yellow. Diameter of field, 2.5 mm.

The sedimentary deposits are mostly fine-grained clastics, probably playa deposits. The conglomerates, as much as 100 feet thick, contain cobbles of granite and some fossiliferous late Paleozoic formations as well as volcanics.

The nearest area having late Paleozoic formations and granite is to the northwest, across Death Valley, in the northern part of the Panamint Range. Moreover, because some conglomerates thin and become finer grained southeastward, as they rise onto the Black Mountains, it seems likely that the late Paleozoic and granite cobbles were brought from the northwest. The volcanic cobbles came probably from the Black Mountains because the sedimentary beds intertongue southeastward with volcanics which are quite like those occurring as cobbles. Probably, therefore, at the time the formations in the Artists Drive area were accumulating, the site of the north end of the Black Mountains was a low area. To the southeast was higher ground where volcanics were accumulating and perhaps being raised structurally as well. To the northwest, across Death Valley, there already were mountains at the site of the northern part of the Panamint Range.

Trace elements in volcanic rocks from the Artists Drive area are given in table 15. The amounts and proportions are similar to those in the felsitic rocks along the foot of the Panamint Range northward from Death Valley Canyon to Tucki Wash (table 26), part of which is dated by the potassium-argon method as late Miocene and by lead-alpha age determinations on the zircon in these rocks as middle Tertiary (T. W. Stern, written commun., 1965). They also resemble the tuffs in the Titus Canyon (?) Formation of Stock and Bode (1935) (table 16).

TABLE 15.—*Trace elements in volcanic rocks in the Artists Drive area*

[Semiquantitative spectrographic analysis by E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Rhyolite	Perlite in agglomerate	Purple felsite
Pb	20	20	20
Mn	200	200	300
Cu	30	10	15
Zr	200	150	300
Ni	20	10	30
Co	5	5	5
V	20	15	50
Y	15	10	15
Be	<1	<1	<1
Ti	3, 000	2, 000	2, 000
B	100	50	70
La	50	50	70
Sc	<10	<10	<10
Cr	50	20	70
Ba	1, 000	1, 000	1, 000
Sr	300	300	300
Mg	0. 5	0. 7	0. 5

NOTE.—All samples showed: As <1,000; Zn<200; Sn<10; Ge<20; Cd<50; Bi<10; In<10; Sb<200; Tl<100; Nb<30; Ta<50; W<100.

At the north end of the Black Mountains the formations in the Artists Drive area are overlain unconformably by the Furnace Creek Formation of Pliocene age.

The age of the greater part of the formations in the Artists Drive area and to the north is in doubt, but the uppermost tongue of playa beds in the Artist Drive Formation has yielded diatoms indicating an early Pliocene age. The fossils were obtained by K. E. Lohman (written commun., 1961), of the U.S. Geological Survey, who has reported as follows:

Loc. 3967. Hard thin limestone from center of W½NE¼ sec. 7, T. 26 N., R. 2 E., Furnace Creek 15-minute quad. Assigned to early Pliocene on basis of incomplete, partly altered diatom assemblage.

The formations in the Artists Drive area also are represented in several large downfaulted blocks in a belt 10 miles long and 2 miles wide at the foot of the Black Mountains north of the Badwater turtleback. The exposed beds, like those in the Black Mountains, are volcanic towards the south, and interbedded volcanic and sedimentary rocks at the north. They are from the upper part of the formations, and the displacement must be about 5,000 feet down towards Death Valley. If the Tertiary formations were as resistant to erosion as the Precambrian gneiss, this part of the front of the Black Mountains would also be a turtleback surface.

These downfaulted blocks in the Artists Drive area are overlapped unconformably by the Pliocene and early Pleistocene(?) Funeral Formation and basalts interbedded with it.

FORMATIONS AROUND COTTONBALL BASIN

Tertiary formations around Cottonball Basin at the north end of the Death Valley saltpan range probably from Oligocene to Pliocene. They occur in fault blocks protruding through and largely concealed by the Quaternary fan gravels. No fossils were found, and the outcrops are too isolated for satisfactory reconstruction of a stratigraphic succession. The stratigraphy of these deposits is uncertain; the structural geology necessarily even more so.

The deposits are in three principal belts (fig. 40). At the north, between the Kit Fox Hills and the Funeral Mountains and extending southeast in a narrow belt along the fault at the foot of the Funeral Mountains, are clastic deposits believed to correlate with the Titus Canyon Formation of Stock and Bode (1935) and accordingly thought to be Oligocene. Deposits believed to be somewhat younger, perhaps Miocene, form an intermediate belt at the Kit Fox Hills and southeastward along the northeast side of Cottonball Basin. The third belt extends northwest and southeast from Cottonball Basin and is represented by the Furnace Creek Formation of Pliocene age.

OLIGOCENE(?) FORMATIONS

Shale, sandstone, grit, and conglomerate thought to correlate with the Titus Canyon Formation of Stock and Bode (1935) from isolated hills protruding through the Quaternary fan gravels between the Kit Fox Hills and Funeral Mountains. These beds also occur in a narrow belt along the fault at the foot of the Funeral Mountains and on a turtleback fault surface on Paleozoic formations in the southern part of the Funeral Mountains. Most of the formation is dark red or dark brown; many clayey beds are light green. The bedding is highly lenticular. The total thickness of the Titus Canyon(?) Formation in these areas can only be guessed, because the outcrops are not continuous and the formation is very much faulted. At least 1,500 feet of beds is exposed, and the total thickness may greatly exceed this.

Of the several kinds of lithologies that are included in the formation, three are distinctive: conglomerates in which the cobbles are fractured; green clayey and silty beds that contrast strikingly with the enclosing thicker and more coarsely clastic beds; and a bright-red conglomerate along the faults at the foot of the Funeral Mountains.

The conglomerate having fractured cobbles occurs in beds 10–25 feet thick at many horizons through the middle 1,000 feet of the formation. Cobbles are as much as 6 inches in diameter and are displaced as much as one-fourth inch by fractures oriented normal to the bedding. Many of the fractures are healed, and the fractured cobbles can be removed intact. The fractured cobbles may touch one another, or they may be isolated in a matrix of gritty silt (fig. 41).

The distinctive greenish clayey beds, mostly less than 10 fet thick, are tuffaceous (fig. 42). They are conspicuous because most of the formation is dark red or dark brown.

The red conglomerate and sandstone is at least 500 feet thick and occurs along the faults at the foot of the Funeral Mountains. The conglomerate contains well-rounded cobbles and pebbles of quartzite along with the more angular boulders of limestone and dolomite. It is faulted against the Cambrian and Precambrian formations that form the foot of the mountain. The reddening may be due to hydrothermal alteration along the fault zone.

The following is a typical but partial section of the Titus Canyon(?) Formation:

Partial section, Titus Canyon(?) Formation, measured about midway between the Kit Fox Hills and Funeral Mountains

	Feet
Top covered. About 750 ft of beds like units 1 and 2 up to top of this member.	
1. Sandstone, gritty, with scattered cobbles; some cobbles as much as 3 in. in diameter; weathers dark brown	200
2. Interbedded tan siltstone and dark-brown grit and cobble conglomerate. Cobbles as much as 3 in. in diameter; fractured. Some tuff	30
3. Sandstone, gritty, with scattered cobbles; some cobbles as much as 3 in. in diameter; weathers dark brown	60
4. Tan beds like unit 2	60
5. Grit, sandstone, and conglomerate with pebbles as much as 2 in. in diameter of quartz, quartzite, carbonate rocks, black chert; no pebbles of volcanic rocks; dark brown. Possible tuff layers. Thin beds; apparently some interbeds of greenish shale	110
6. Greenish shale in beds 5–10 ft thick with 1- to 6-in. beds of tan limestone and silt and a bed of reddish cemented clay or silt	50
7. Like unit 5; weathers dark brown but is gray on fresh fracture. Across wash to southeast the top of this units is cut off discordantly by unit 6; this may be a flat fault and not an unconformity	40

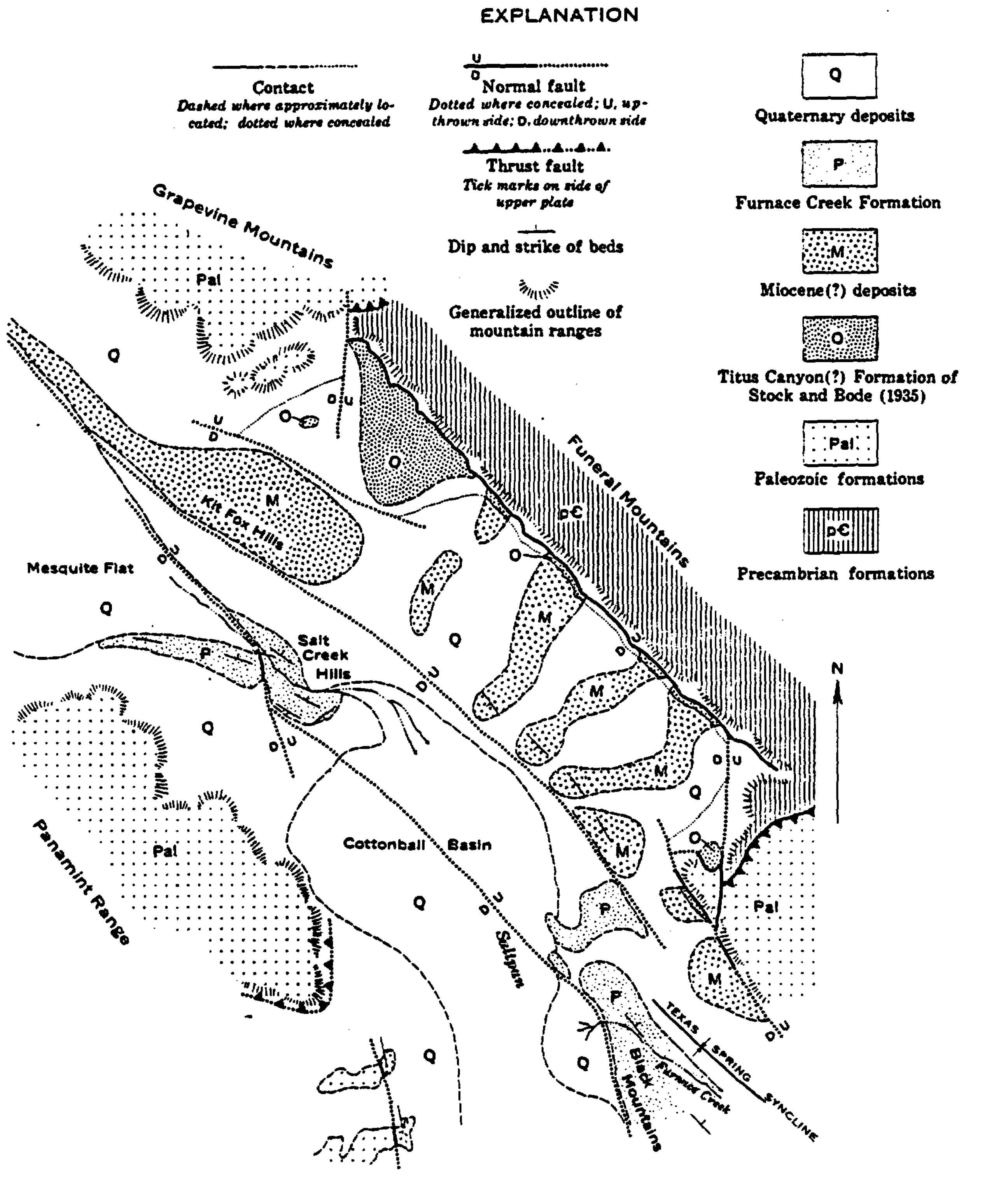


FIGURE 40.—Sketch map of Tertiary formations around Cottonball Basin.

FIGURE 41.—View of fractured cobble conglomerate in the Titus Canyon(?) Formation of Stock and Bode (1935). The fractures extend through the cobbles and some cross from one cobble to another. They are oriented at right angles to the bedding, which approximately parallels the elongation of the cobbles. Displacements along the fractures are mostly less than a quarter of an inch.

Partial section, Titus Canyon(?) Formation, measured about midway between the Kit Fox Hills and Funeral Mountains—Continued

	Feet
8. Interbedded sandstone, grit, shale, and fine conglomerate. Pebbles of Precambrian rocks as much as ¾ in. in diameter. Gray unit with salt-and-pepper appearance	25
9. Conglomerate, with fractured cobbles as large as 6 in. Brown sandstone above and below	15

Base concealed.

The uppermost beds in the Titus Canyon(?) Formation, estimated to be about 500 feet thick, are light colored and fine grained. The beds are mostly silt and sand, in part limy, and very well bedded—even laminated. Dominant colors are yellow and light green. Other beds are platy limestone, mostly brownish, but some are light gray. All these beds are cut by veinlets of gypsum and anhydrite.

Along the northeast side of the Kit Fox Hills the texture of these beds ranges from clay to medium sand. The sandy beds are thickest, some are 8 inches thick; the clay stains blue when tested with benzidine, and apparently is a montmorillonite. Colors are mostly pastel shades of lavender, brown, and red. Overlying these beds are red and buff sandy beds at the base of the buff conglomerate forming the Kit Fox Hills and mapped as Miocene(?).

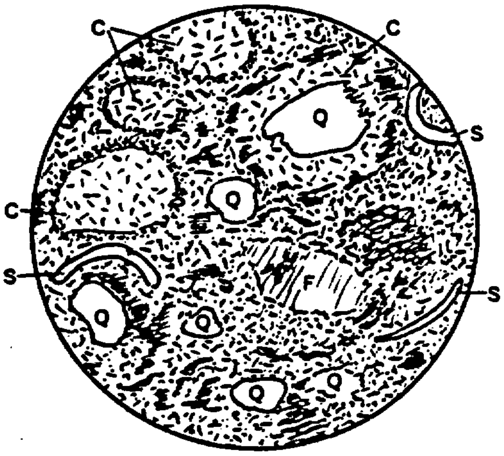


FIGURE 42.—Micrograph of thin section of greenish clay from the Titus Canyon(?) Formation of Stock and Bode (1935). Diameter of field, 2.5 mm. Well-rounded microcrystalline grains, probably chalcedony (C) have cryptocrystalline (or possibly argillized glass?) rims. Rounded quartz grains (Q) and some feldspar (F) occur in the groundmass; some, like the example in the northeast quadrant, are surrounded by microcrystalline material that is surrounded by an irregular discontinuous rim of highly birefrigent material, probably a clay mineral (black areas). Scattered through the rock are isotropic shards (S); many of these now are holes through the slide as if their glass had been removed by solution. Some rounded grains (not shown) are greenish and might be pollen, but more likely they are stained rims on grains of chalcedony.

In brief, the lowest part of the Titus Canyon(?) Formation appears to be fanglomerate. The middle part contains conglomerate interbedded with fine-grained beds. The upper part seems to be largely a playa deposit.

Some analyses of trace elements in beds of the Titus Canyon(?) Formation are given in table 16. The tuff is similar to the volcanics in the Artists Drive Formation (table 15).

As noted by Noble and Wright (1954, p. 149), these beds are similar to the Titus Canyon Formation which, in the Grapevine Mountains north of this area, has yielded Oligocene vertebrate fossils (Stock and Bode, 1935). The beds probably also correlate with the lower part of the formations in the Artists Drive area (Noble and Wright, 1954, p. 149) in the northern part of the Black Mountains.

The faults along the foot of the Funeral Mountains north of Echo Mountain dip 25°–30° towards Death Valley. The Titus Canyon(?) Formation was deposited against the fault surface and subsequently faulted down against it.

MIOCENE(?) FORMATIONS

Tertiary deposits in the Kit Fox Hills and in the fault blocks southeastward to the foot of the Funeral

776-623 O-66—5

TABLE 16.—*Trace elements in beds correlated with the Titus Canyon(?) Formation of Stock and Bode (1935)*

[Semiquantitative spectrographic analyses, by E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Greenish beds		Tuff	Gritty sandstone, brown
Pb	50	70	10	20
Mn	150	500	200	700
Cu	70	100	5	100
Zr	500	300	300	100
Ni	<5	50	5	10
Co	<10	10	<10	10
V	30	200	10	100
Y	100	50	20	20
Be	1	1	1	<1
Ti	3,000	7,000	2,000	3,000
B	200	700	200	100
La	50	50	50	<50
Sc	<10	20	<10	<10
Cr	20	150	10	70
Ba	700	700	500	2,000
Sr	300	300	70	1,000
Mg	1	5	0.3	5

Note—All samples showed: As <1,000; Sn <10; Ge <20; Ga <20; Cd <50; In <50; Sb <200; Tl <100; Ta <50; W <50; Ag <1.

Mountains constitute an intermediate belt between the outcrops of the Titus Canyon(?) Formation of Stock and Bode (1935) and those of the Furnace Creek Formation (fig. 40). The deposits, estimated to aggregate about 4,000 feet thick, are thought to overlie the Titus Canyon(?) Formation and to underlie the Furnace Creek Formation, but because of the faulting, this sequential relationship has not been established, and the stratigraphic position is assumed.

The color of these deposits tends towards buff and light red; it is suggestive of coarse facies of the Muddy Creek Formation—which is considered to be of Pliocene(?) age and which is extensive east of Death Valley. The Titus Canyon(?) Formation tends to be much darker, and the Furnace Creek Formation tends to be much lighter.

In the Kit Fox Hills the deposits are mostly reddish conglomerate containing quartzite and chert cobbles. Limestone, dolomite, and volcanic cobbles make up a minor part. A pebble count near the north side of the hills showed the following percentages: Red and purple quartzite and chert, 40; sandstone, 40; shale, 10; schist, 5; other, 5. This suggests a source in the northern part of the Funeral Mountains.

Along the south side of the Kit Fox Hills the felsite cobbles increase to about 15 percent, and limestone and dolomite to about 10 percent. Southeastward along the Kit Fox Hills the volcanic rocks in the cobbles increase to 35 percent; at one place a count showed 60 percent volcanic rocks. The other materials are chiefly quartzite and chert rather than carbonate rocks.

Northwestward the lower part of the conglomerate grades into fine-grained sedimentary deposits, and this, too, suggests a source to the northeast. Probably the conglomerate was thin, if ever present, northeastward across the belt of the Titus Canyon(?) Formation. Very possibly that area was a pediment that ended along a southward-facing fault scarp at about the position of the present northeast-facing edge of the Kit Fox Hills; if so that area could have been bypassed by the Miocene(?) deposits.

Along the northeast side of Cottonball Basin are three bare hills of red conglomerate similar to that in the Kit Fox Hills. The conglomerate is about 1,400 feet thick at the southeasternmost hill, 1,000 feet thick at the middle one, and about 900 feet thick at the northwest one. Three-quarters of the cobbles and pebbles are quartzite and chert; the remainder are volcanics and carbonate rocks. This conglomerate, like that in the Kit Fox Hills, probably was derived from the northern part of the Funeral Mountains. Fossiliferous limestone and dolomite from the Paleozoic formations, which compose a large fraction of the conglomerates in the Furnace Creek Formation, are notably lacking in these Miocene(?) deposits.

This conglomerate is overlain by about 350 feet of white and gray tuffaceous beds containing thin reddish beds with pebbles, perhaps reworked from the conglomerate and marking an unconformity. Overlying these beds and marking the top of the deposits mapped as Miocene(?) is a brown sandstone unit about 1,500 feet thick.

The brown sandstone unit consists of brown limy grit, limy sand, friable sandstone, thin-bedded limestone, and a little greenish silt and shale. A few layers are pebbly. Limy beds are a few inches thick and are strikingly ripple marked. Sandy beds are 1–8 feet thick. Figure 43*D* is a micrograph of a thin section of a limy sandstone bed. Unidentifiable stem fragments of monocotyledonous plants are common, and animal tracks have been reported in these brown beds (H. D. Curry, oral commun., 1960), but no identifiable fossils have been found.

The light-colored tuffaceous beds and the brown sandy beds that overlie the conglomerates and that have been taken as the top of the deposits mapped as Miocene(?) are transitional between the Miocene(?) and Furnace Creek Formation. They could as well have been included with the Furnace Creek Formation.

Underlying the conglomerate that forms the bare hills northeast of Cottonball Basin, and between those hills and the Funeral Mountains, is a series of sandy tuffaceous beds aggregating about 3,000 feet thick. A section northeastward from Cottonball Basin shows the following:

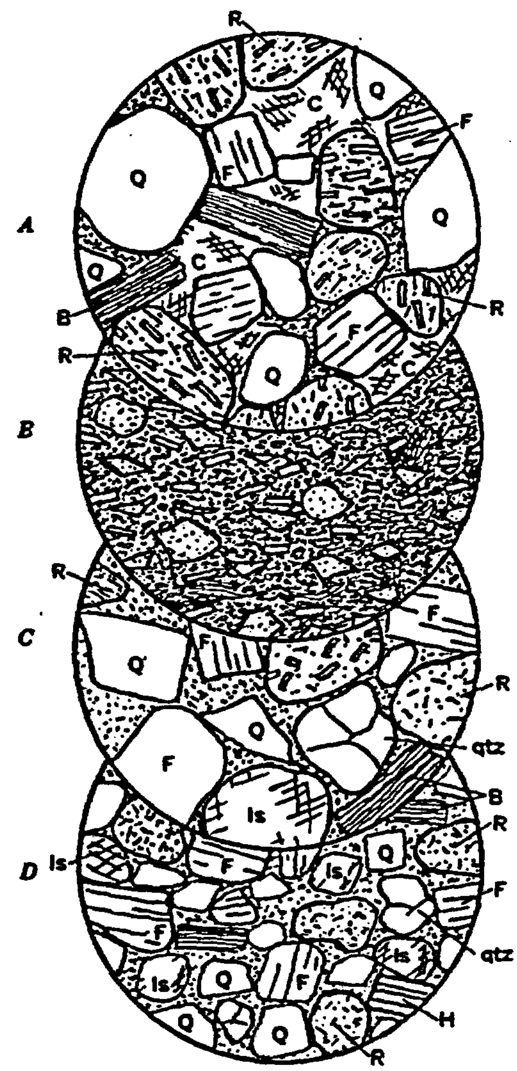


FIGURE 43.—Micrographs of thin sections of sedimentary rocks from Miocene(?) formations northeast of Cottonball Basin. Q, quartz; F, feldspars; H, hornblende; B, biotite; C, calcite; R, volcanic rock; qtz, quartzite; ls, limestone or dolomite. Diameter of fields, 2.5 mm. *A*, Crystal tuff from lower part of the Miocene(?) beds. The matrix is calcite, fine glass fragments, and clay. Crystals are quartz, feldspars, biotite, and fragments of glassy volcanic rocks. *B*, Clayey tuff. Rectangular crystals are mostly feldspar, some mica. Shards of glass are angular. Matrix clay. *C*, Arkose from near the middle of the Miocene(?) formations. Differs from *A* in having greater variety of volcanic rock fragments, quartzite, limestone or dolomite, both argillized orthoclase and plagioclase feldspar; also has quartz and biotite. Matrix silty clay. *D*, Brown calcareous sandstone at top of the Miocene(?) formations. Grains are quartz, argillized orthoclase, microcline, and plagioclase feldspar, hornblende, biotite, limestone or dolomite, quartzite, volcanic rocks of several kinds. Matrix is calcareous clay.

Section northeastward from Cottonball Basin to the Funeral Mountains

	Feet
Top. Red conglomerate forming bare hills at northeast edge of the basin	900–1,400
1. White and light-gray tuffaceous sandstone, some grit, very little conglomerate; pebbles isolated from one another. Beds 1–2 ft thick and well laminated	450
2. Brown limy sandstone, grit, conglomerate, and interbedded tuffaceous sandstone. Brown limy beds comprise about half the unit; they are resistant and form the most conspicuous outcrops. Conglomerate beds are pebbly, with only a few cobbles. Pebbles are well rounded, partly polished; mostly various kinds of felsite and not many are of Precambrian or Paleozoic rocks	800
3. Buff and yellow arkosic sand and silt (fig. 43*C*), some thin beds of gray tuffaceous sandstone. Well bedded and finely laminated	600
4. Conglomerate, reddish; subround and subangular pebbles, cobbles, and boulders as much as 5 ft in diameter; boulders larger than 2 ft not common; most large boulders are a banded purple quartzite containing quartz veins as wide as ½ inch; also present are cobbles of bull quartz. Source evidently was the Chloride Cliff district in the Funeral Mountains. Other materials include red and green banded quartzite, sandy dolomite. Sorting poor; beds are 1–3 ft thick and contain all sizes of materials. Many boulders and cobbles are slabby; their long axes parallel the bedding. Estimated size proportions: 15 percent larger than 6 in.; 45 percent 1–6 in.; 40 percent less than 1 in., including matrix	1,000

The conglomerate at the base of the section is a resistant unit that forms hills along the foot of the Funeral Mountains and in places lies against the frontal fault. Away from the fault the conglomerate seems to grade into and be intertongued with red sandstone that is in part tuffaceous (fig. 43*A*, *B*). The tuffaceous layers are as much as 5 inches thick, and are separated by beds of fine-grained sandstone ½–2 inches thick. The sandstone contains shale laminae and is silty; it grades downward into light-colored beds thought to be the upper unit of the Titus Canyon(?) Formation.

The composition of the tuffaceous rocks varies widely, depending on the proportion of volcanic debris to the clastics from the Precambrian. In table 17 the contrast is illustrated by analyses of three random samples from this part of the section.

In the Artists Drive area and in the Amargosa thrust complex the tuff differs in composition from the felsites, being notably higher in strontium and lower in titanium. Systematic chemical analyses of these volcanic rocks might help determine whether they are more closely related to those in the Amargosa thrust complex in the Black Mountains or to those in the rhyolite district northeast of the Funeral Mountains.

TABLE 17.—*Trace elements in the Miocene(?) deposits*

[Semiquantitative spectrographic analyses by E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	Tuffaceous sandstone	Tuff	Tan limy sandstone at top of formation
Pb	15	10	20
Mn	1,000	50	1,000
Cu	30	10	70
Zr	200	50	300
Ni	20	<5	10
Co	7	<5	10
V	70	10	100
Y	20	10	50
Be	<1	1	<1
Ti	5,000	700	5,000
B	100	70	100
La	50	<50	50
Sc	10	<10	10
Cr	70	15	20
Ba	500	300	1,500
Sr	300	1,000	700
Mg	1.5	0.5	3

NOTE.—All samples showed: As<1,000; Zn<200; Sn<10; Ge<20; Ga<20; Cd<80; Bi<10; In<10; Sb<200; Tl<100; Nb<50; Ta<80; W<100.

Despite uncertainties, the stratigraphy of these Miocene(?) deposits around Cottonball Basin does help in interpreting the structural history. The northwest-trending Furnace Creek fault zone is suspected to have had considerable lateral displacement (Noble and Wright, 1954, p. 153), but these Miocene(?) deposits are not offset laterally from their probable source in the Funeral Mountains. However, there may have been lateral displacement along faults farther to the southwest.

PLIOCENE FORMATIONS

Pliocene deposits are represented by the Furnace Creek Formation, which outcrops at the north end of the Black Mountains and northward from there to the fault along the southwest side of the Miocene(?) deposits (fig. 40). The formation probably underlies Cottonball Basin, for it reappears northwest of there in the anticlinal uplift at the Salt Creek Hills.

At the north end of the Black Mountains the formation is more than 5,000 feet thick and consists in large part of fine-grained light-colored playa deposits (fig. 44). Interbedded with the playa deposits are conglomerates and some basalts.

The formation consists of a basal conglomerate of variable thickness, but averaging perhaps 200 feet, overlain by 2,500 feet of light-colored fine-grained playa beds. Some of these beds are highly tuffaceous; others are clastic sand or silt (fig. 45). Interbedded with these are basalts and conglomerate beds. Some of the latter thicken westward toward the west front of the Black

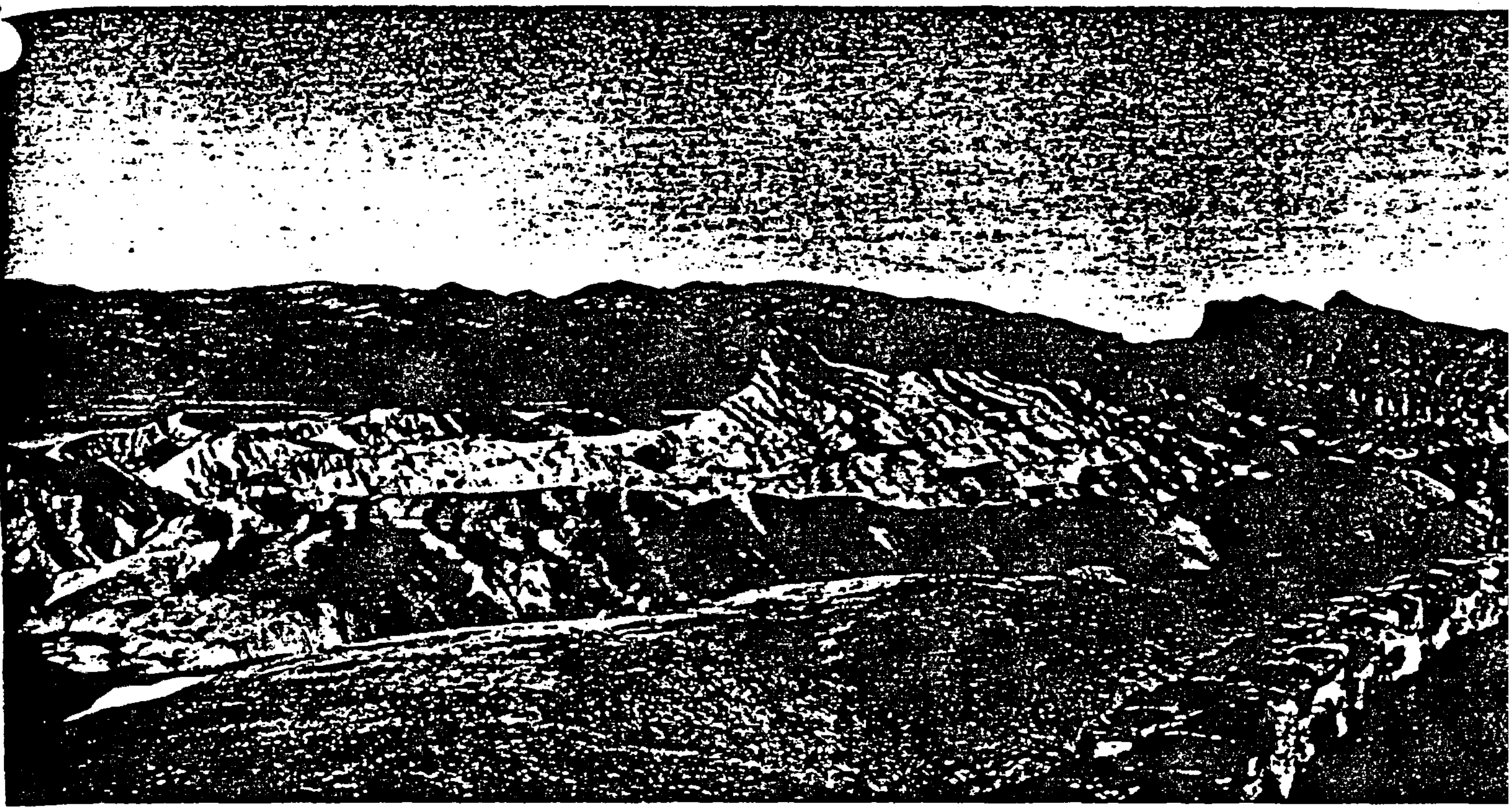
FIGURE 44.—View of Furnace Creek Formation at north end of t[illegible] Black Mountains. View is southwest and west from Zabriskie Poin[illegible] an overlook by Highway 190 about 3 miles up Furnace Creek Wa[illegible] from Furnace Creek Inn. The base of the Furnace Creek Form[illegible]tion is at the topographic break between the badlands and t[illegible] rougher and higher ground in the distance at the left. Ligh[illegible]

Mountains. The playa beds are overlain by the conglomerate seen at the right side of figure 44. This conglomerate reaches a maximum thickness of about 2,000 feet at the mouth of Furnace Creek. It is overlain by another series of playa beds about 1,300 feet thick. Younger units of the formation, if any, are concealed under the Pliocene and Pleistocene(?) Funeral Formation in the trough of the Texas Spring syncline (fig. 40).

The basal conglomerate is composed mostly of cobbles of Paleozoic limestone and dolomite, and there are boulders as much as a foot in diameter. Among these cobbles are fossiliferous rocks from upper Paleozoic formations—fusulinid limestone from Pennsylvanian or Permian formations; granular limestone containing large crinoid fragments, almost certainly from the Tin Mountain Limestone or younger limestone; black dolomite like that of the Tin Mountain; and limestone containing *Cyrtospirifer* from the Lost Burro Formation. These fossiliferous cobbles may have been brought from the northwest, possibly from Tucki Mountain but more likely from farther north in the Panamint Range. About 65 percent of the cobbles are Paleozoic carbonate rocks; 10 percent are quartzite, 20 percent are volcanics, and a few percent are granite and miscellaneous other types. The granite does not look like the granites at Skidoo or at Hanaupah Canyon, and it is different i[illegible] its trace elements. It contains a tenth as much lead an[illegible] five times as much copper as do the Skidoo and H[illegible]naupah granites. Its content of zirconium is only [illegible] fifth as great, and it contains 10 times as much boro[illegible] 10 times as much strontium, 4 times as much vanadiu[illegible] and cobalt, and twice as much nickel as do the oth[illegible] granites.

The percentage of volcanic rocks among the cobbl[illegible] is low, considering that this basal conglomerate unco[illegible]formably overlaps volcanic and other deposits of f[illegible]mations in the Artists Drive area. The proportion [illegible] volcanic rocks may increase southeastward, but su[illegible] change was not determined.

Whereas the basal conglomerate of the Furnace Cre[illegible] Formation seems to have been derived in large p[illegible] from the northwest, the conglomerate that outcrops [illegible] the mouth of Furnace Creek Wash and caps the low[illegible] playa beds member of the formation, seems to have be[illegible] derived mostly from the Black and Funeral Mountai[illegible] This conglomerate thins southeastward. At the mo[illegible]



colored playa beds about 2,500 feet thick extend to the base of a conglomerate which forms the dark cliff at the right. The beds are dipping to the right (north) into the Texas Spring syncline. Center of the picture looks west across Death Valley to the Panamint Range at Aguereberry Point; Tucki Mountain is at the right. Panorama by John Stacy.

of Furnace Creek it is 1,800 feet thick. Two miles southeast from there it is 1,600 feet thick, and in another 2 miles it is only 1,000 feet thick.

On the other hand, the proportion of volcanic rocks in the conglomerate increases southeastward as the conglomerate thins, from 10 percent at the mouth of Furnace Creek to 20 percent where the formation is 1,600 feet thick and to 40 percent where the thickness is down to 1,000 feet. The proportion of quartzite and carbonate rocks changes irregularly. At the mouth of Furnace Creek there is 75 percent carbonate and 15 percent quartzite and other clastics; 2 miles southeastward the percentages are 30 and 45, respectively; and where the formation is 1,000 feet thick, the percentages are 50 and 10. These rocks look like early Paleozoic types and probably came from the southern part of the Funeral Mountains.

Another conglomerate, perhaps 200 feet thick, is found along the west front of the Black Mountains in the playa beds about half way between the basalt conglomerate and the one that crops out at the mouth of Furnace Creek. This conglomerate is composed very largely of volcanic debris and thins southeastward, as if from a source at the site of Death Valley, or a source west of the valley, perhaps, in the Amargosa thrust complex.

The contrast in sources of the different materials also shows well in some of the fine-grained playa beds (fig. 45). Some of these beds are largely of volcanic materials; others are largely from Precambrian or Paleozoic formations.

Probably there was a playa elongated northwestward at the present site of the north end of the Black Mountains, formerly the north base of a pile of felsitic volcanics (formations at Artists Drive) that had accumulated on the Precambrian rocks farther south in the Black Mountains. Most of the fine-grained sediments deposited in the playa were derived from the volcanic pile to the south, but deposition was interrupted by influxes of coarse gravels from the northwest, presumably in response to structural movements in that area. During the second half of the time represented by the Furnace Creek Formation an increasing amount of sediment was brought from the south end of the Funeral Mountains.

Lateral changes in thickness and geochemistry of the playa beds suggest that the central part of the Pliocene playa was a short distance east of the present edge of

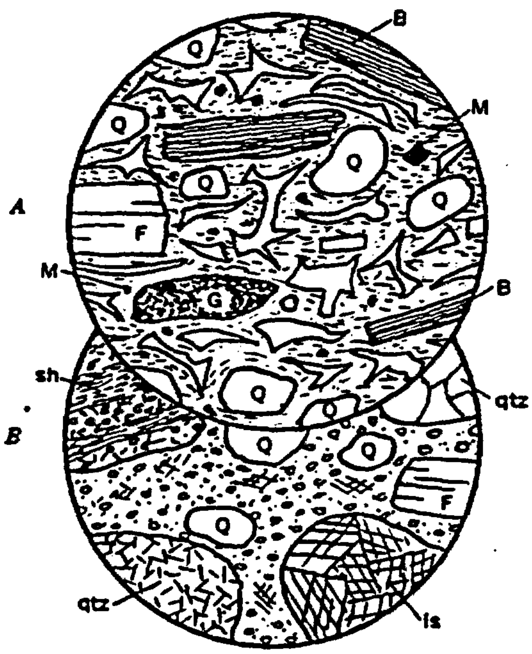


FIGURE 45.—Micrographs of thin sections of playa deposits in the Furnace Creek Formation. Q, quartz; F, feldspar; B, biotite; M, magnetite; G, volcanic glass; qtz, quartzite; ls, dolomite or limestone; sh, shale or schist. Diameter of fields, 2.5 mm. A, tuff layer, an ash fall. Angular shards of volcanic glass mixed with grains of quartz, biotite, feldspar, magnetite, and rounded grains of volcanic glass are in a matrix of sericite, quartz, and feldspar. B, poorly sorted sandy silt of clastic sediments, probably derived chiefly from the Funeral Mountains. The large grains are quartzite, both fine grained and coarse grained, dolomite or limestone, shale or schist, and scattered small grains of quartz and feldspar in a calcareous clay with minute grains of quartz.

the saltpan in Cottonball Basin. Playa beds there are highly saline. At the East Coleman Hills, which are at the north tip of the outcrop area at the north end of the Black Mountains (fig. 40), the upper playa beds are at least 2,400 feet thick and contain sulfates and borates. About a mile southeast the upper playa beds are about 1,300 feet thick, and some contain abundant veins of gypsum. Farther to the southeast, along the flank of the Texas Spring syncline, these beds continue to thin to about 650 feet and contain less sulfate, almost no chlorides, and more carbonate and granular tuff.

The lower playa beds of the formation contain thick deposits of gypsum and of borates that were productive during the days when the 20-mule teams operated. These deposits are southeast of those in the upper playa beds as if the sulfate-borate zone shifted northwestward towards Cottonball Basin. Today it is located at the edge of the saltpan.

At the Salt Creek Hills, about 2,500 feet of light-colored playa beds of the Furnace Creek Formation are

TABLE 18.—*Trace elements in the Furnace Creek Formation*

[Semiquantitative spectrographic analyses by E. F. Cooley, U.S. Geol. Surve
Values in parts per million, except Mg, which is given in percent]

Element	Tuff	Basalt	Element	Tuff	Basal
Pb	70	20	Ti	500	10,00
Mn	500	1,000	B	100	10
Cu	50	70	La	<50	10
Zr	50	150	Sc	<10	2
Ni	10	200	Cr	20	30
Co	<5	70	Ba	30	1,50
V	10	150	Sr	70	3,00
Y	15	20	Mg	0.2	
Be	3	<1			

NOTE.—Both samples showed: As<1,000; Zn<200; Sn<10; Ge<20; Ga<20; C
<50; Bi <10; In <10; Sb <200; Tl <100; Nb <50; Ta <50; W <100.

exposed in an asymmetrical anticline. Westward alon Salt Creek, basaltic (or andesitic) lavas and conglon erate are inter bedded with the playa beds. The co glomerate contains boulders of granite that may be pa of a boulder train extending southeastward to the nort end of the Black Mountains.

Analyses of trace elements in a tuff and a basalt fro the Furnace Creek Formation are given in table 1 The tuff has more lead, manganese, and copper, a less barium and strontium than does the tuff near t base of the Miocene(?) deposits (compare with tab 17). The basalt differs from those in the Amargo thrust complex (compare with table 26) in its high content of nickel, boron, lanthanum, chromium, a strontium.

Diatoms collected from the top and base of the Fu nace Creek Formation indicate that it spans much Pliocene time. The diatoms were collected by K. Lohman (written commun., 1961), of the U.S. Geolo ical Survey. A collection near the base of the form tion is described as follows:

Loc. 4159. Hard calcareous tufa immediately overlying trusive basalt, SW¼NE¼ sec. 27, T. 26 N., R. 2 E., R 15-minute quad. Assigned to early Pliocene on basis of complete, partly altered diatom assemblage.

The collection from near the top of the formatio described as follows (written commun., 1961):

Loc. 4070. 1.15 miles northwest o ertine Point on H way 190, roadcut in south side. 15-minute quad. E gray limestone containing an abundant and well-prese diatom assemblage extraordinarily similar to a diatom as blage from a diatomite in Sand Pedro Valley, Ariz., w has yielded a vertebrate fauna of middle Pliocene age. 4070 has therefore been assigned to the middle Pliocene.

Other plants from the Furnace Creek Format reported by Axelrod (1940), also indicate a Plio age for this formation.

PLIOCENE AND PLEISTOCENE(?) DEPOSITS—FUNERAL FORMATION

The Funeral Formation crops out extensively in the fault blocks that extend northwest from Furnace Creek along the east side of Cottonball Basin, in small areas in the Artists Drive fault blocks and near Mormon Point, and in an extensive area along Emigrant Wash and about as far as its head.

In Furnace Creek Wash and around Cottonball Basin, fanglomerate of the Funeral Formation rests conformably on the light-colored playa deposits of the Furnace Creek Formation. On the Artists Drive fault blocks the fanglomerate overlies volcanic and other rocks of the Artist Drive Formation with an angular unconformity. At Mormon Point the fanglomerate overlaps Precambrian rocks and is faulted against them. Along Emigrant Wash the fanglomerate overlies and is faulted against the turtleback surface on the west foot of Tucki Mountain (p. A143), and it overlies and is faulted against the west flank of the granite at Skidoo.

In the absence of fossils, it is doubtful that these widely separated deposits are of the same age. On the basis of diatoms in the upper part of the Furnace Creek Formation, it seems likely that the overlying Funeral Formation there is Pliocene and early Pleistocene(?). A Miocene age originally was assumed for the fanglomerate at the head of Emigrant Wash (Hopper, 1947), but more recently a Pliocene age has been inferred for those deposits on the basis of pollen obtained from them (Axelrod and Ting, 1960).

Included in the Funeral Formation at the north end of the Artists Drive fault blocks is a mass of quartzite breccia believed to be a chaotic breccia derived from the Zabriskie Quartzite (p. A25). How it became emplaced remains a mystery. Other blocks of Paleozoic rocks in that area are found along faults and are far removed from masses of the Paleozoic formations that might have supplied them. Perhaps they are blocks of chaos associated with low-angle faulting. The beds mapped as the Funeral Formation in the Salt Creek Hills have been regarded by Curry (1939) as Pleistocene on the basis of tracks found in them.

In Emigrant Wash the Funeral Formation is at least 3,000 feet thick; it was originally referred to as the Nova Formation (Hopper, 1947). It consists in large part of cobbles and boulders of Precambrian rocks like those on the high part of the Panamint Range; consequently, the formation has been thought to have been derived from that direction. However, Axelrod and Ting (1960) report that the formation becomes coarser grained westward and indicate a western source from the direction of Panamint Valley. This interpretation involves major topographic changes like those indicated by the southeastward-thinning gravels in the Furnace Creek Formation in the Black Mountains (p. A60). The Funeral Formation at the head of Emigrant Wash may record similar changes, but I gave the deposit there only cursory examination.

On the Artists Drive fault blocks the Funeral Formation is composed largely of debris from the volcanic formations. The Funeral unconformably overlaps the much-faulted and tilted rocks of the Artists Drive formation and includes flows of basaltic lava and some beds of volcanic ash as much as 4 feet thick. It dips westward under the playa where a drill hole on the floor of the saltpan opposite Artists Drive encountered 400 feet of fanglomerate with basaltic cobbles believed to be the Funeral. This hole, the log of which is given on page A74, did not reach the base of the formation. The Black Mountains must already have been high ground, shedding debris westward, but the vulcanism was continuing while these fanglomerate beds of the Funeral Formation were being deposited.

Along Furnace Creek only about 150 feet of fanglomerate of the Funeral Formation is exposed in the trough of the Texas Spring syncline and in the fault blocks farther north. A mile east of Zabriskie Point the fanglomerate is composed largely of volcanic rocks, commonly as much as 2 feet in diameter. Northward from Furnace Creek Wash the gravel contains increasing proportions of Paleozoic carbonate rocks and lower Paleozoic clastic rocks reflecting the composition of the source rocks in the Funeral Mountains. Near Echo Canyon the gravel contains about 70 percent carbonate rocks and 30 percent quartzite; from Echo Canyon north to Nevares Spring the proportions are reversed, carbonate rocks 30 percent, quartzite 70 percent. At Rock Alinement Wash and farther north the gravels are very largely quartzite, only 5–10 percent is carbonate rock. Clearly the Funeral Mountains were in existence and shedding debris into Death Valley when fanglomerate of the Funeral Formation in this area was being deposited, but the faulting and the downfolding into the Texas Spring syncline shows that much of the uplift of the mountains, especially the Black Mountains, was later. I assume that the fanglomerate was being deposited while the mountains were being elevated.

In the East Coleman and Salt Creek Hills the fanglomerate contains boulders of coarse-grained granitic rocks as large as 5 feet in diameter. Their source probably was to the northwest, in the northern Panamint Range, like that of the granite and upper Paleozoic cobbles in the conglomerates of the Furnace Creek Formation (p. A60).

Except for these granitic boulders, the fanglomerate is not very bouldery. North of Echo Canyon, for example, the common large size is 1 foot in diameter. At one outcrop 1½ miles from the mountain front (in SE cor. sec. 13, T. 27 N., R. 1 E.) a cliff 30 feet high and 30 feet long contains only 30 boulders as large as 1 foot in diameter; that is, only 1 per 30 square feet of outcrop. The common large size of the gravel is 6 inches in diameter, and even these cobbles are few. An outcrop 10 by 10 feet exposed 50 such cobbles; that is, 1 per 2 square feet. Three-quarters of a mile downstream the gravel is even less coarse and contains only a third as many small boulders. Similar proportions were found along Echo Canyon Wash.

At most places the fanglomerate is firmly cemented with calcium carbonate. The cemented layers are several feet thick and extend for hundreds of feet along the outcrops, especially where the gravel overlies fine-grained sediments so that ground water can be perched on top of the impermeable beds. The cemented layers are particularly thick and extensive in an area extending 2 miles southeastward from Park Village, an area that has many springs. The gravel also is cemented in the axis of the Texas Spring syncline down dip from Travertine Spring and Texas Spring (fig. 2, locs. 2, 3).

Where the fanglomerate is cemented with calcium carbonate it is cut by numerous veins of banded calcite, locally referred to as Mexican onyx. These veins, which generally parallel the principal faults, are a few inches to a few feet wide and may be several hundred feet long. Trace elements in the veins are the same and in about the same amounts as in the travertine mounds at springs. Analyses are given in table 20.

Caliche in the Funeral Formation in the Artists Drive area is mostly gypsum rather than calcium carbonate.

The surface of the fanglomerate has developed smooth desert pavement (fig. 50), a surface described more fully in connection with the No. 2 gravel where it is best developed.

The fanglomerate is so faulted, folded, and dissected that its remnants no longers retain their fan form. In the Park Village area, for example, the fanglomerate is in a fault block that forms a ridge about 350 feet high and trends north roughly along the contour of the fans that rise eastward to the Funeral Mountains (figs. 53, 62). Drainage down the fans has become incised across the ridge, and younger gravel deposits lie on both sides of it. The incised gorges are older than Lake Manly (p. A69). At the Salt Creek Hills the Funeral Formation is raised in a faulted structural dome having at least 250 feet of structural relief. Salt Creek is incised across this dome.

In both these areas fanglomerates of the Funeral u conformably overlie deposits of late Pliocene age th are uplifted more than the fanglomerate. The defo mation clearly began in Pliocene time and then co tinued. Very likely the deformation progressed small increments over a long period of time, and t drainage probably is antecedent—or anteposed (Hu 1956, p. 65)—across the uplifts. That is to say, t drainage incised across the uplifts probably w dammed repeatedly as uplift progressed, but the pond streams overflowed along their old channels. The sult is aggradation upstream from the dam, giving so suggestion of superposition in that direction; but dow stream the drainage would be antecedent or conseque

QUATERNARY SYSTEM

PLEISTOCENE DEPOSITS

NO. 2 GRAVEL

The No. 2 gravel unconformably overlaps fanglom ate of the Funeral Formation along Furnace Cr Wash and differs from it in being less well consolidat more bouldery, and less faulted, tilted, and dissect so that it retains its fan form. The No. 2 gravel la calcite veins; but it does have layers cemented w calcium carbonate, although these are thinner and l extensive than in the Funeral Formation. The No resembles the Funeral in having surfaces mantled w smooth desert pavement composed of disintegrat blocks, slabs, and flakes.

The No. 2 gravel can be distinguished from younger ones, Nos. 3 and 4, in at least six ways:

1. The No. 2 gravel forms the highest benches above present drainage.
2. The low parts of the No. 2 gravel are overlapped the younger gravels.
3. The No. 2 gravel is more bouldery than the your ones.
4. The No. 2 gravel is more cemented.
5. The surface of the No. 2 gravel is smooth desert p ment, whereas the younger gravels have r surfaces.
6. The streamworn cobbles and boulders on the surf of the No. 2 gravel have disintegrated to pro a new crop of angular rock fragments.

Along the foot of the Panamint Range and in fro the Funeral Mountains the No. 2 gravel forms be 100 feet above the present washes (pl. 2). Towar saltpan these benches commonly slope more steeply the average slope of the gravel fans, and their l edges extend under and are overlaped by the No. No. 4 gravels (fig. 64).

The No. 2 gravel contains considerable sand, but mixed with it are boulders many feet in diameter, as well as pebbles and cobbles of intermediate size. The largest granitic boulders are on the fans of Hanaupah and Starvation Canyons where many are more than 10 feet in diameter and some are as large as 30 feet. They are distributed along the entire length of the fan, down to 250 feet below sea level.

These abundant large boulders may indicate exterior drainage at the time they were deposited. No matter what mechanism is considered, vast amounts of water would be required to deposit such coarse debris in sufficient volume to build fans 6 miles long and 3 miles wide. With so much water, there should have been a lake, and had there been a lake, the bouldery deposits should have formed deltas, not fans. These coarse deposits in such large volume, though, pose no problem if Death Valley had exterior drainage to the south at the time they were deposited.

On the fans of Hanaupah and Starvation Canyons granitic rock comprises about 20 percent of the gravel. Sixty percent in quartzite, and about 10 percent is carbonate rocks and argillite.

On Trail Canyon fan, where the gravels contain about equal proportions of quartzite and carbonate rocks and only minor amounts of igneous and metamorphic rocks, the common large size of boulders is 2 feet in diameter, although some are 6 feet. On Johnson Canyon fan, where the gravel consists of about 80 percent quartzite, 10 percent monzonite, and 10 percent carbonate rocks and argillite, the boulders are small like those on Trail Canyon fan.

Northeast of Cottonball Basin the No. 2 gravel contains considerable salt, more salt than occurs in the gravel around the other sides of the saltpan. This is attributed to the winds transporting salt northeastward from the saltpan.

A gravel-filled former channel of Furnace Creek is well exposed along Furnace Creek Wash (fig. 46). Opposite Zabriskie Point this fill contains almost 90 percent carbonate rocks. This differs from fanglomerate of the Funeral in the area, which here contains a high proportion of volcanic rocks; it also differs from the fan gravels derived from the Funeral Mountains north of here, which contain a high proportion of quartzite. The principal source of this channel fill apparently was in the southern part of the Funeral Mountains; the principal source of fanglomerate of the Funeral in the area apparently was the northern part of the Black Mountains.

The downcutting that followed deposition of this channel gravel probably was caused by structural uplift of the wash relative to the present Furnace Creek fan, because the surface of the gravel fill, if projected, would extend about 50 feet above the fan. This uplift probably is the same deformation that raised the small gravel terraces along the west foot of the hills just north and just south of the mouth of Furnace Creek (fig. 47) and that produced hanging valleys along the foot of the Black Mountains farther south (fig. 77).

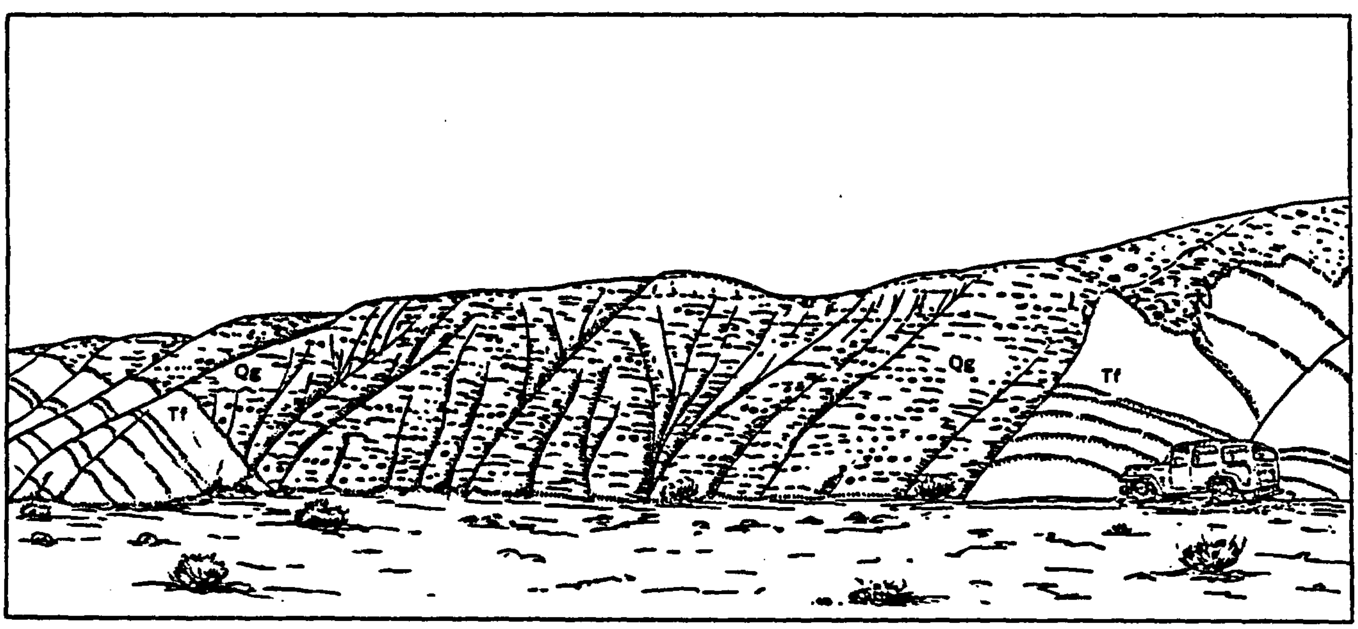


FIGURE 46.—Gravel fill in former channel of Furnace Creek Wash. The channel is eroded in Furnace Creek Formation (Tf), which dips steeply northeast (right), and the bottom is below the level of the present wash. The fill (Qg) is about 50 feet thick. View is north in tributary to Furnace Creek Wash opposite Zabriskie Point.

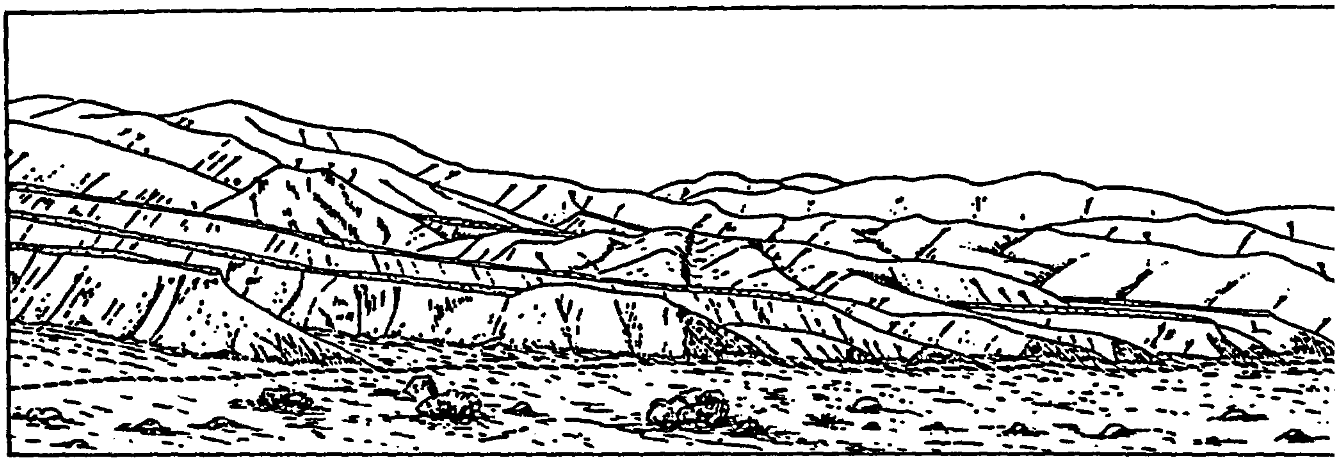


FIGURE 47.—Escarpment along the front of the north end of the Black Mountains a quarter of a mile south of Furnace Creek. The terraces, capped with No. 2 gravel, have been faulted upward about 75 feet.

Other deposits of No. 2 gravel occur in the canyons in the Panamint Range, considerably upstream from the gravel fans. There are sizable remants in most of the canyons, and these remnants are 50–100 feet higher than the present canyon bottoms. Evidently the canyons had been eroded to their present depth before the gravel was deposited, then buried to a depth of 100 feet or more with this gravel and subsequently re-excavated. Interbedded with the gravel in the south fork of Six Spring Canyon is a bed of volcanic ash as much as 4 feet thick.

Drainage on the No. 2 gravels reflects the fan form of the remants of that gravel. Washes are parallel and consequent. Moreover, washes that are a few hundred feet wide and more than about 10 feet deep commonly have low terraces of No. 3 gravel along them.

Karstlike solution features are common where Tertiary rocks are overlain by the No. 2 gravel, such as near Furnace Creek Ranch. On the west side of Death Valley, along the north edge of the wash a mile north of Hanaupah Canyon, there is a depression 15 feet deep, locally referred to as "the crater" in the No. 2 gravel. The depression has a floor of silt about 50 feet in diameter and 60 feet higher than the wash which is on the south side of the depression. This depression probably is due to water seeping from the gravel bench to the wash, dissolving calcium carbonate caliche from the gravel, and allowing the gravel above to collapse.

The fan of No. 2 gravel at Starvation Canyon has three tremendous ridges radiating down the fan and evidently marking old mudflows (fig. 48). The ridges, large enough to show on the topographic contours, are 2–3 miles long, 500–1,000 feet wide, and 50–75 feet high. Their volumes are 8–25 million cubic yards. Each ridge has a narrow crest with a wash along it; the sides are strewn with huge boulders and slope evenly to t[h]
adjoining fan surfaces.

The surface of the No. 2 gravel is smooth desert pav[e]ment. Boulders and cobbles on these surfaces have di[s]integrated to produce an entirely new crop of angul[ar] rock fragments—the kind that no longer are proper[ly] classified as water worn. They are better described [as] blocks, the equivalent of boulders; as slabs, the equiv[a]lent of cobbles; and as flakes, the equivalent of pebbl[es] (Woodford, 1925, p. 183; see also Pettijohn, 1949, [p.] 12–15). These desert pavements composed of slabs a[nd] flakes are the smoothest in the valley.

On a typical surface on the No. 2 gravel, and also [on] fanglomerate of the Funeral Formation, 75 percent [or] more of the boulders and cobbles have lost their origin[al] roundness. On some surfaces practically every bould[er] is fractured or crumbled. Although the kind and [de]gree of weathering varies considerably, depending [on] the composition and texture of the rock, no rock h[as] been spared, whether coarse or fine grained (fig. 49)[.]

The disintegration of these gravels is most advan[ced] where the gravels extend into the zone of abund[ant] salts. Striking examples of the effectiveness of salt[s in] accelerating disintegration are provided by the conc[rete] bases of bench marks along the highway crossing [the] saltpan and extending along its west side. Concret[e at] locations that are frequently wetted with saline w[ater] is badly disintegrated. The disintegration is less [ad]vanced at equally saline locations where the wett[ing] and drying is less frequent, and the concrete stil[l is] sound at locations that are dry and not notably sal[ine.]

But disintegration of stones on the No. 2 surfac[e is] general and not confined to the toes of the fans w[hich] are impregnated with salts. The disintegration oc[curs] all the way to the mountains. It is a near-surface [

nomenon because boulders and cobbles more than 2 or 3 feet deep in these deposits are sound.

The desert pavement consist of a single layer of closely spaced blocks, slabs, and flakes as illustrated on figure 50. Beneath it is a layer of vesicular sand and silt, 1-6 inches thick, containing as much as a tenth of a percent of salts. Gravel under this layer is cemented with salt and iron oxide. Stones forming the pavement creep down the slope, as is indicated by terracettes (fig. 51) and by trains of slabs extending downslope from blocks that are disintegrating.

An individual pebble on the desert pavements is subjected to three very different microclimates. The upper surface, exposed to maximum temperature and maximum temperature change, in general is being eroded, as shown by partial removal of desert varnish. Around the side of the pebble is a narrow band where the temperatures probably are less extreme and where there is maximum wetting, by dew as well as by other surface water. This narrow zone has a dense population of microorganism, and even some megascopic ones—algae. The underside of the pebble has moderate temperatures and soil moisture condenses on it. This surface is red with iron oxide. In an environment like Death Valley these differences in microclimate are extreme, and probably are an important factor in the continued weathering and disintegration of the No. 2 gravels.

Desert pavement may develop in a very short time. Where the ground consists of loose sand or silt containing pebbles, only a few windstorms are needed to blow away the fine materials and collect the coarse as a pavement of pebbles. Such very young pavements do not have a silt layer under the pebbles. On some of the archeological sites, however, a silt layer one-fourth inch thick occurs beneath the layer of pebbles. The thickest silt layer that I found on pavement developed on No. 3 gravel is about an inch, but no systematic search for thicker layers was made. The silt layer under the pebbles on the No. 2 gravel commonly is a few inches thick. The evidence is pretty good that the thickness of the

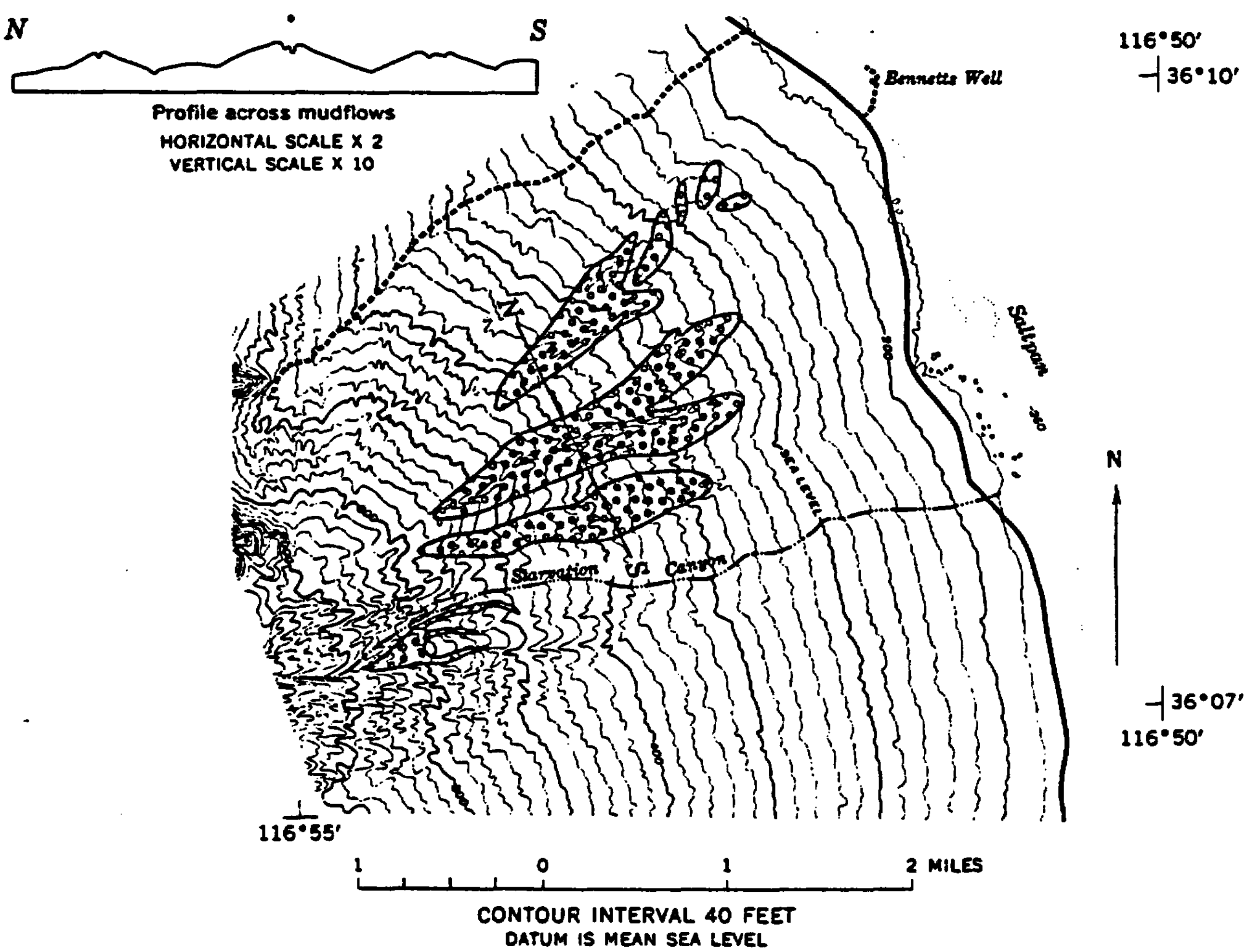


FIGURE 48.—Map and profile across mudflows on Starvation Canyon fan. Topography from U.S. Geological Survey topographic quadrangle; Bennetts Well, 1952.

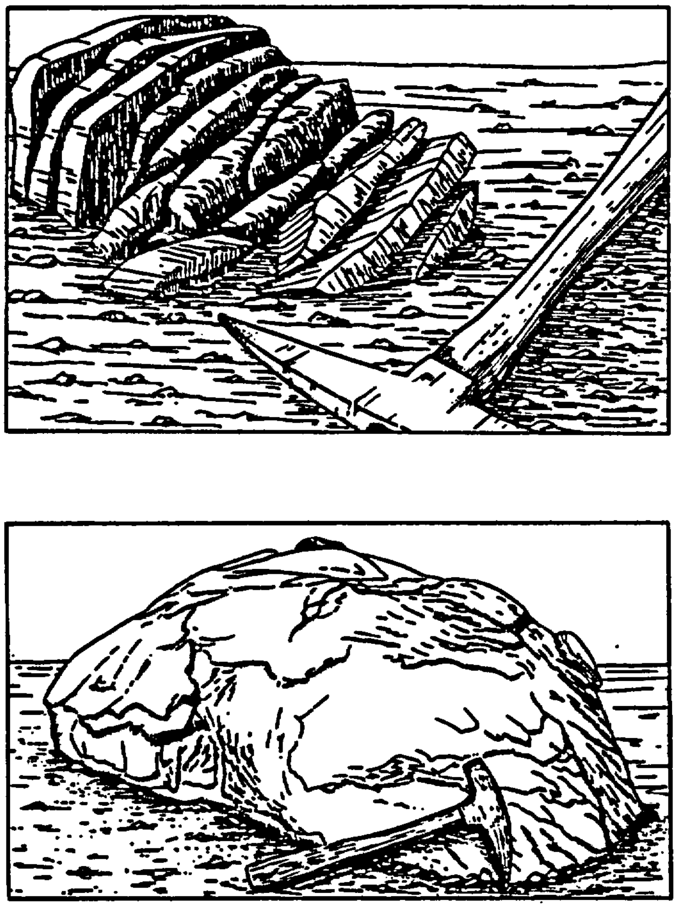


FIGURE 49.—Boulders disintegrating to slabs and flakes on the oldest gravel deposits (Funeral Formation and No. 2 gravel). Upper, Quartzite boulders commonly break into slabs along transverse fractures. Lower, Massive rocks like the porphyry boulders on the fans of the Hanaupah and Starvation Canyons exfoliate and crumble.

silt layer on old surfaces is greater than on young ones.

The terracettes on the No. 2 gravel commonly have treads 1–5 feet wide and risers 1–6 inches high (fig. 51). The surface inch or two on the treads commonly contains 1 percent or more of water-soluble salts whereas the adjacent stable surface without terraces contains as little as 500 parts per million of water-soluble salts. These ground patterns are described more fully by Hunt and Washburn (in Hunt and others, 1965).

Only once during the 6 years of the field study did I witness a rain that thoroughly soaked into the gravel. On February 16, 1959, 1 inch of rain fell in 24 hours, and the silt layer under gravel pavement on the Hanaupah Canyon fan became soaked. Walking on the pavement involved walking ankle deep in mud, because footsteps sank into the mud underlying the gravel of the pavement. Frequent soaking like this would accelerate mass-wasting processes, but there is evidence that these processes operate very slowly under the present

FIGURE 50.—Desert pavement, foreground. View west from Park Village fault block. Weathering of boulders and cobbles at the surface has produced a new mantle of blocks, slabs, and flakes, forming a smooth desert pavement in which the stones are closely spaced but barely or not at all shingled. Photograph by John R. Stacy.

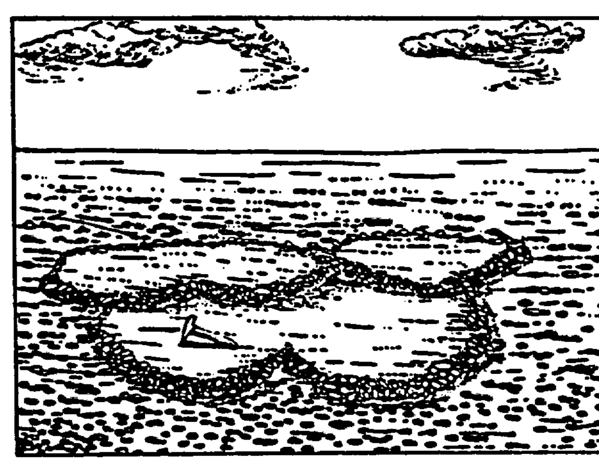


FIGURE 51.—No. 2 gravel with desert pavement interrupted by terracettes.

climate (Hunt and others, 1965). Most of the terracettes and other patterned ground probably are relicts from a wetter period.

DEBRIS AVALANCHE

A large avalanche of blocks and rubble of Precambrian rocks is at the foot of the Black Mountains midway between Badwater and Copper Canyon. At the source, above the debris avalanche, is a huge scar between 2,000 and 2,500 feet in altitude; the volume of the avalanche must be greater than 5,000,000 cubic yards. No lake features were recognized across the front of the avalanche, but its lower part has been displaced by faults that seem to antedate the Pleistocene lake deposits. It is composed of blocks of Precambrian rocks tens of feet in diameter in a matrix of rubble of similar rocks.

LAKE DEPOSITS

Late Pleistocene lake features in Death Valley are few, small, and not at all distinct. That Death Valley had contained a Pleistocene lake was stated widely long before positive evidence of its existence had been found. Before 1900 both Russell (1885, 1889) and Gilbert (1890) had referred to a former lake in Death Valley, and Bailey (1902) named it Death Valley Lake. Yet, as late as 1914, Gale, who was a student of the Quaternary basins, wrote (1914, p. 401):

> In spite of the immense drainage territory tributary to Death Valley there is no evidence that the waters from these streams ever accumulated in it to sufficient extent to form more than a shallow inconstant lake. A search for traces of any upper lines around the slopes leading into Death Valley has failed to reveal evidence that any considerable lake has ever existed there.

Not until 1926 was clear evidence found that a late Pleistocene lake had flooded Death Valley. Levi Noble identified the strand lines on the basalt hill, later known as Shoreline Butte, at the south end of Death Valley, and discovered other strand lines in the cove northeast of Mormon Point (Noble, 1926a, p. 69). The lake or lakes that produced these features have since been referred to as Lake Manly (Means, 1932; Blackwelder, 1933, 1954).

Small embankments of shingled gravel, evidently beach deposits or near-shore bar deposits of late Pleistocene lakes, are numerous but widely scattered along the north and east sides of Death Valley at altitudes as high as 380 feet above sea level; small horizontal terraces that may be wavecut features occur several hundred feet higher. Similar deposits or beach scars are curiously lacking along most of the west side of the valley; in fact, they are known at only 2 localities 40 miles apart—on the basaltic hill between Tucki Wash and Blackwater Wash and on Shoreline Butte at the south end of Death Valley and south of the area mapped.

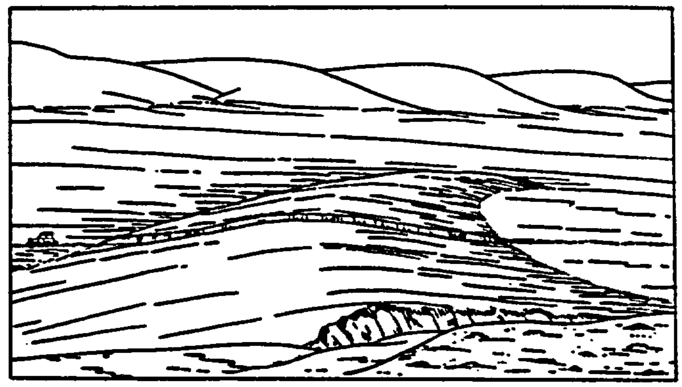


FIGURE 52.—Gravel bar of late Pleistocene Lake Manly resting on older fan gravel 2 miles north of Beatty Junction. Sketch from photograph.

The most accessible and best developed gravel bar is exposed along the highway 2 miles north of Beatty Junction. This bar (fig. 52) extends nearly a quarter of a mile east from a hill of Miocene(?) rocks which formed an island at the time the bar was built. The bar, 500 feet wide and 20 feet high, is composed of well-sorted, shingled, and crossbedded gravel, most of it an inch in diameter or less, and not at all like the poorly sorted fan gravels. The top of the bar is nearly level; it is 150 feet above sea level. The deposit narrows and then eastward. Other less well-developed lake gravels crop out below sea level a mile south of the bar. These and all the other gravel deposits of the late Pleistocene lakes are composed of firm pebbles showing no sign of disintegration. The pebbles commonly have a weathering rind and are stained with desert varnish.

Three miles southeast of this bar is another well-developed one forming an arcuate deposit half a mile long and 500 feet wide, resting on a bench of No. 2 gravel. The bar curves through an arc of 90°. The gravels are shingled, crossbedded, and usually about an inch in diameter, like those in the bar above Beatty Junction. The foreset beds in the gravel dip 10° NW. The top of this bar is nearly level and is less than 100 feet above sea level.

No other shoreline features were found between this bar and the one near Beatty Junction. The temptation is strong to assume that the 2 bars, which are similar and highly exceptional features in this area, were formed at the same time and that the difference in level is attributable to 50 feet of postlake faulting or tilting between the 2 localities.

Other shoreline features are exposed along the west face and top of the ridge of Funeral Formation in the fault blocks north of Park Village. Small deposits of shingled gravel are associated with long narrow terraces that, in part at least, are scars of old strand lines

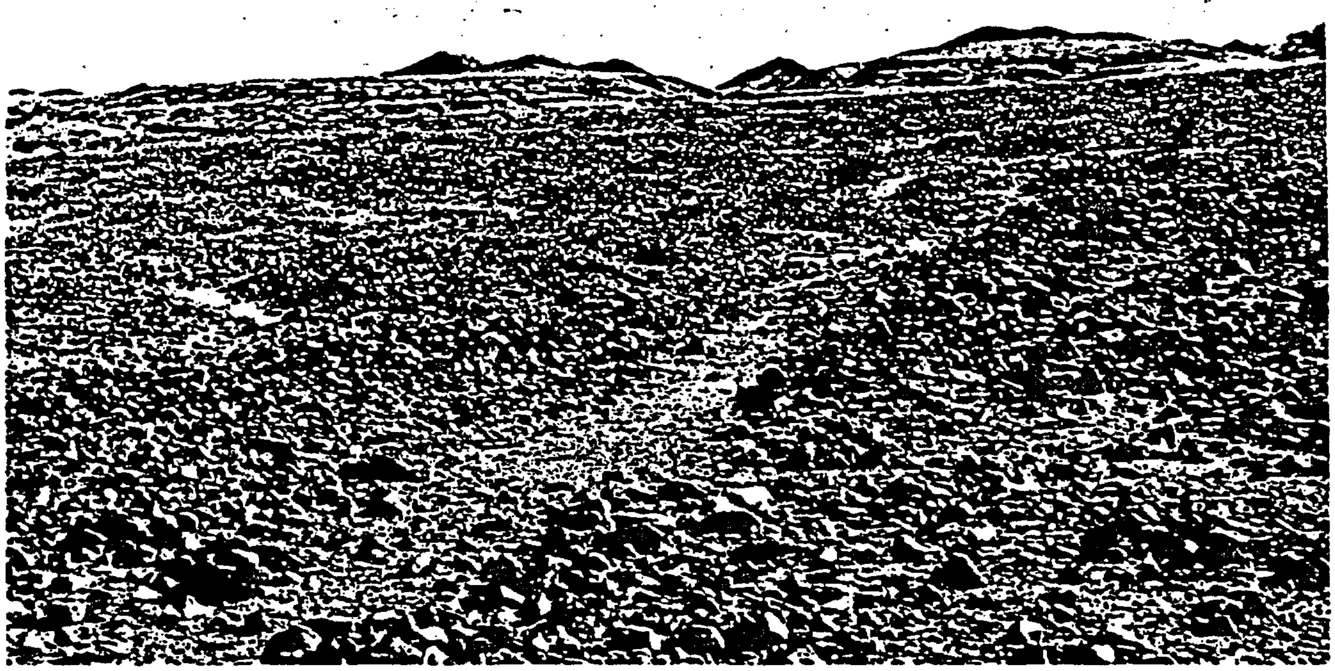


FIGURE 53.—View east to Park Village fault block of Funeral Formation showing long narrow terraces that are interpreted to be scars of strand lines of late Pleistocene Lake Manly. In foreground is No. 3 gravel.

(fig. 53); some similar terracettes, however, are attributable to mass washing. Most of the strand lines and associated deposits along the west face of the fault blocks are between sea level and 75 feet above sea level. Strand lines and associated deposits also occur on top of the fault blocks at 150 feet above sea level. Some of these lake deposits extend into the gorges across the fault block, showing that these gorges antedate the lake.

Another deposit of lacustrine gravel is at the north end of the Artists Drive fault blocks, and at the same level northward and southward from this deposit are narrow terraces, evidently wave cut, impressed on tilted strata. The deposit of gravel and the terraces are at sea level. Stone artifacts at this location have been interpreted to indicate human occupation in Death Valley at the time of the lake (Clements and Clements, 1953), but this interpretation is doubtful. No unequivocal artifacts have been found within the gravel deposit; the unequivocal artifacts are part of the desert pavement on top of the gravel and are therefore younger. Moreover, these artifacts are typologically quite like those characteristic of Recent occupations (A. P. Hunt, 1960).

Along the steep front of the Black Mountains from Badwater south to Mormon Point are numerous discontinuous horizontal embankments of gravel cemented with calcium carbonate. Most of these embankments are between sea level and 200 feet above sea level, but some are even higher. How many of them are truly lake deposits is problematical.

Well-sorted shingled lake gravel is exposed overlying a fault block of the Funeral Formation, at the north end of the steep part of the mountain front, about midway between Badwater and Bridge Canyon. The lake gravels are composed mostly of Precambrian rocks, whereas the fanglomerate contains, in addition many volcanic rocks. The lake gravels are better sorted and less well cemented than the fanglomerate. Foreset beds in the lake gravels dip north-northwest as if there had been northward shore drift at this location. The lake gravel intertongues with the lower part of a colluvial deposit that overlaps the lake beds. This colluvium is cemented with gypsum rather than calcium carbonate.

Other lake gravels are exposed at Mormon Point and extend 1½ miles eastward. These deposits, the most extensive lake deposits of gravel exposed in the valley, overlie the No. 2 gravel. They are at sea level and as much as 200 feet above sea level.

The only lake deposits and geomorphic features attributable to lake processes found thus far on the west side of Death Valley are at Shoreline Butte (Noble, 1926a; Blackwelder, 1933, 1954) and at the basaltic hill between Tucki Wash and Blackwater Wash. At Shoreline Butte are numerous shorelines between the foot of the butte at 150 feet below sea level nearly to the top, at 400 feet above sea level. The hill between Tucki Wash and Blackwater Wash has two shorelines. The lower one is an embankment of gravel at an altitude of 160 feet. Its gravel consists of basalt and of Paleozoic rocks derived from the fans in Blackwater Wash. The embankment thins northward, and the gravels, which are 2–3 inches in diameter at the south end of the hill, become finer northward (1 in. in diameter). This embankment is cemented by deposits of calcium carbonate that forms spotty masses of travertine. Both the gravel and the travertine are distributed irregularly through a vertical range of about 20 feet, but they can be followed discontinuously from the south to the north end of the hill.

On top of the hill, in the saddle between the peaks, at an altitude of 380 feet, is another small patch of shingle gravel derived from Paleozoic rocks.

Although the gravels from Paleozoic formations were drifted northward by shore currents across the face of this hill, the much lighter scoriaceous basalt from this hill does not occur as shore drift extending northward across the fans of No. 2 gravel in Tucki Wash. It would appear that the lake deposit is older than the No. 2 gravel, but this probably is not so. The surface of the No. 2 gravel in Tucki Wash may be younger than the lake, and if so, embankment deposits of basaltic scoriae that may have extended northward across the No. 2 gravel could have been destroyed.

The relationships at Tucki Wash illustrate the highly uncertain age of these lake deposits with respect to the No. 2 gravel. At 2 locations, Mormon Point and 2 miles north of North Side Borax Camp, the lake deposits rest on and must be younger than the No. 2 gravel. Nowhere has the reverse relationship been found. Moreover, the gravels at the surface of the No. 2 are much more weathered and disintegrated than those at the surface of the lake deposits. The difference in weathering is the kind that elsewhere in the West has been successfully used to distinguish pre-Wisconsin deposits from Wisconsin and younger ones. But why, then, are there no lake deposits or other shore features impressed on the many miles of No. 2 gravel exposed along the west side of Death Valley?

The west side of the valley was the lee side of the lake, and deposits there could have been thin and discontinuous. Even so, it is difficult to believe that all trace of them would be destroyed. Yet the evidence at the two localities where the stratigraphic relationships are certain, and the more general evidence about the difference in weathering, suggest that the No. 2 gravel everywhere is older than the Lake Manly deposits.

Lake Manly has been correlated with the Wisconsin (Tioga and Tahoe) stages of glaciation in the Sierra Nevada (Blackwelder, 1954; Clements and Clements, 1953), which correlate with stages of Lake Bonneville and Lake Lahontan. This correlation is supported by the fact that the pebbles on the surface are not disintegrated but are firm—suggesting an age no older than Wisconsin—yet many have developed a weathering rind that suggests an early Wisconsin (Tahoe) age.

The slight erosion and sedimentation record of Lake Manly may mean that the lake was of brief duration, and its level may have fluctuated rapidly. Whatever the cause, this California lake left one of the least distinct and most incomplete records of any Pleistocene lake in the Great Basin—another California superlative!

The water that accumulated in Death Valley to form Lake Manly has been attributed to overflow from a lake that formed in Panamint Valley when there was overflow from Searles Lake and the other lakes headward along the Owens River valley (Gale, 1914, p. 402; Blackwelder, 1954, p. 57). The overflow into Death Valley would have been by way of Wingate Pass and down Wingate Wash, but no trace remains of the floods that must have descended the wash to form Lake Manly. Perhaps much or most of the water came from the south, by way of the Mojave River and Soda Lake. This hypothesis has some support in the distribution of species of desert fish in the several drainage basins.

Owens Valley has two genera of desert fish, *Siphateles* and *Catostomus*, that are said to have come from the Lake Lahontan area; *Siphateles* also occurs in the Mojave River (Miller, 1948). Neither of these genera has been reported in the Death Valley—Amargosa River area. Further, a *Cyprinodon* that occurs in the Owens River, *C. radiosus*, is said to be more closely related to the Colorado River cyprinodonts than are any of the three species living in the Death Valley—Amargosa River area (Miller, 1948). This distribution of species suggests that the drainage system from Owens Valley to the Mojave River bypassed Death Valley.

Flooding from the direction of Soda Lake also is suggested by considering the possible tilt of the Lake Manly deposits. The principal deposits are at sea level in Mesquite Flat and along the north and east sides of

Death Valley as far south as Artists Drive, but they are 200 feet above sea level on the west side opposite Furnace Creek. They are 200 feet above sea level at Mormon Point, and there are large deposits as much as 300 feet above sea level at Shoreline Butte. At all these places there are higher shoreline features; the altitudes given refer to the principal deposits. They suggest an eastward tilting of 200 feet and a northward tilting of 300 feet. If such tilt is real, the lake probably extended to Soda Lake, which is where Russell (1885, pl. 1; 1889, pl. XVI), Gilbert (1890, pl. 2) and Bailey (1902) originally thought it went, and which was still considered a possibility by other later workers (Blackwelder and Ellsworth, 1936, p. 462).

VALLEY FILL

Gravity and magnetic surveys indicate that the fill in Death Valley has a maximum depth of about 9,000 feet near the west side of the valley a short distance south of Bennetts Well (p. A108). Drill holes a thousand feet deep near Badwater and in Cottonball Basin show rather uniform alternations of mud and salt to the bottom of the holes (table 19), and assuredly the uppe thousand feet is Quaternary. I assume that about third of the fill is Quaternary and that the rest i Tertiary.

The fill in Badwater Basin thins southward an northward. Opposite Artists Drive the fill is onl about 4,000 feet thick. A drill hole in this area en countered only 50 feet of mud and salt and then wen into basaltic conglomerate to a depth of 500 feet befor the hole was abandoned. This conglomerate is corre lated with the Funeral Formation that rises eastwar onto the fault blocks at Artists Drive and there uncon formably overlaps the older volcanic rocks (p. A63)

The fill thickens again northward under Cottonbal Basin, thins under the Salt Creek Hills, and thicken again under Mesquite Flat.

Logs of the three deep holes are given in table 1 Logs of some shallow holes drilled by the U.S. Geolog ical Survey in connection with the search for mor potash deposits during World War I are given in Hun and others (1965).

TABLE 19.—*Logs of wells drilled by Pacific Coast Borax Co.*

[Drillers' logs revised from mud samples]

Description	Depth (feet)
Well 1, 2½ miles northwest of Badwater, E½ sec. 30, T. 25 N., R. 2 E. (projected)	
Surface salt	0 - 4
Alternating salt and mud	4 - 20
Hard salt	20 - 28
Hard salt; occasional streaks of black mud	28 - 38
Hard salt	38 - 41
Soft mud	41 - 41½
Hard salt	41½ - 48
Thin streak soft black mud	48 - ..
Hard salt	48 - 50
Soft black mud, containing salt crystals	50 - 53
Hard salt	53 - 55
Very hard salt	55 - 59½
Soft black mud	59½ - 60½
Very hard salt; streaks of black mud and brown clay	60½ - 62
Very hard salt; one small streak black mud	62 - 65
Hard salt	65 - 67
Soft black mud	67 - 68
Hard salt	68 - 72
Soft black mud	72 - 72½
Hard salt	72½ - 75½
No record	75½ - 81
Hard salt	81 - 89
Mud and salt crystals	89 - 92
Hard salt	92 - 94
Mud and salt	94 - 100
Hard salt	100 - 103
Soft black mud	103 - 106½
Hard salt	106½ - 107
Very soft black mud; salt crystals	107 - 109
Very hard salt	109 - 114
Softer salt with some black mud	114 - 115½
Very hard salt; occasional streaks of soft black mud	115½ - 118
Salt mixed with black mud and a little red clay	118 - 122
Hard salt	122 - 125
Soft black mud	125 - 126½
Hard salt	126½ - 132
Black mud	132 - 134
Medium-hard black mud and salt	134 - 136
Very hard salt	136 - 138
Black mud	138 - 140
Medium-hard salt	140 - 142
Soft black mud	142 - 143
Hard salt	143 - 154
Soft black mud	154 - 158
Very hard salt	158 - 160
Hard and soft streaks; particles of reddish clay	160 - 163
Hard salt and soft streaks of clay	163 - 166
Soft gray clay and salt crystals	166 - 168
Soft gray clay and salt, mostly salt	168 - 170
Very hard salt	170 - 171
Softer material; clay and salt; gray and reddish clay	171 - 173½
Hard salt	173½ - 175
Softer material; some gray clay	175 - 176
Hard layers of salt and small streaks gray clay	176 - 190
Hard streaks of gray mud and salt	190 - 210
Hard salt; a little gray mud	210 - 215
No record	215 - 218
Hard salt; small streaks of gray clay	218 - 225
Hard salt	225 - 226
Hard salt and streaks of gray mud and clay	226 - 239
Same material; clay on the increase	239 - 241
Thin strata; alternating hard salt and dark muds	241 - 246
Grayish mud; very little salt	246 - 250
Salt and mud, principally salt	250 - 255
Soft clay; very little salt	255 - 260
Alternating salt and dark gray clay; some black clay	260 - 265
Dark-gray clay; thin (2-in.) streaks of salt	265 - 269
Dark-gray clay; very little salt	269 - 272

TABLE 19.—*Logs of wells drilled by Pacific Coast Borax Co.*—Continued

Description	Depth (feet)
Well 1, 2¼ miles northwest of Badwater, E½ sec. 30, T. 25 N., R. 2 E. (projected)—Continued	
Soft gray clay and salt crystals	272 - 273
Hard salt	273 - 274
Gray clay and dark mud, almost black	274 - 277
Clay and salt	277 - 278
Hard salt	278 - 280
Layers of salt and gray mud	280 - 284
Gray clay with small streaks of salt	284 - 285
Gray and black mud	285 - 290
Black mud	290 - 291
Very hard salt	291 - 292
Gray clay; few salt crystals	292 - 294
Gray clay; very tough in places	294 - 301
Very tough black clay	301 - 301½
Brown clay with salt crystals	301½- 305
Very hard salt	305 - 308
Gray clay with fine salt crystals	308 - 312
Hard salt	312 - 317
Gray clay and fine salt	317 - 318
Hard salt with "some coarse brown and gray clay"	318 - 328
Gray clay; some salt	328 - 329
Hard salt; some gray clay	329 - 336
Gray mud and clay; very little salt	336- 339
Gray mud and clay; a little salt and black clay	339 - 342
Yellow clay	342 - 343
Hard salt; streaks of gray mud	343 - 346
Hard salt and yellow clay	346 - 349
Hard salt; little gray mud and clay	349 - 350
Hard salt; gray mud and clay increasing	350 - 359
Gray and black mud; some salt	359 - 360
Gray mud and salt	360 - 365
Gray mud, salt, and clay	365 - 370
Gray mud and clay	370 - 375
Gray mud and clay, and salt	375 - 385
Tough brown salty clay	385 - 390
Same material, with streaks of black mud	390 - 394
Salty brown clay	394 - 400
Same material, but softer	400 - 410
Salty brown clay, streaks of gray clay	410 - 415
Brown mud, some particles of black mud	415 - 428
Same material, with a few salt crystals	428 - 433
Hard salt	433 - 436
Soft brown mud and salt crystals; a little glauberite	436 - 437
Salt; some very hard streaks; a little glauberite	437 - 440
Hard salt; some gray clay; a little glauberite	440 - 442
Hard salt; gray and black clay; notable amount of glauberite	442 - 443
Hard salt; gray clay	443 - 444
No record	444 - 446
Brown clay and hard salt	446 - 447
Same material, but softer; some gray clay	447 - 448
Hard salt; gray, brown, and black clay. Hardness varying	448 - 465
Same material, but very hard	465 - 468
Soft material: brown, gray, and black clays; black increasing	468 - 481
Hard salt and clays, brown, gray, and black	481 - 484
Hard salt and gray clay	484 - 495
Salt with gray, brown, and black clays; hardness varying	495 - 519
Clays, gray, brown, and black; some salt	519 - 523
Hard salt: gray, brown, and black clays	523 - 526
Hard salt: black and brown muds	526 - 528
Same material, black muds increasing	528 - 530
Same material, but softer	530 - 536
Salty black clay; streaks of brown clay	536 - 545
Same material, but brown clay decreasing	545 - 556
Salty black mud; some little streaks of clay	556 - 625
Stiffer clay, and black mud	625 - 631
Same material; clay alternating gray and brown	631 - 636
Black salt mud with gray and brown clay; clay increasing	636 - 650
Gray clay with streaks of black mud and brown clay	650 - 659
Gray clay with more black mud; streaks of brown clay	659 - 665
Clay, alternating gray and black, with some brown	665 - 680
Same material; probably some anhydrite ($CaSO_4$)	680 - 684
Soft black mud	684 - 685
Harder material, black mud and clays	685 - 695
Sticky and soft black mud and clays	695 - 697
Harder material, black mud and clays	697 - 700
Soft black clay	700 - 706
Mixed clays; no salt	706 - 710
Brown black clays; varying soft to tough	710 - 730
Gray salty clay; some streaks black and brown	730 - 742
Gray salty clay, gradually growing harder. (Probably more salt—RES)	742 - 748
Hard salt; layers gray clay and black mud	748 - 751
Hard salt with clay; slow drilling	751 - 764
Hard salty clays; gray and black, tough	764 - 766½
Same material, with some brown clay	766½- 767½
Black mud and gray clay; softer; mud increasing	767½- 785
Black mud; very little clay	785 - 800
Hard salt with clay and mud; a few fragments of gypsum	800 - 813
Hard gray clay, salty; thin strata of black and brown clays, showing a few fragments of gypsum	813 - 815
Same material, but hardness varying	815 - 830
Hard salt; with clays and mud	830 - 833
Black mud	833 - 843
Black mud with some gray clay and salt	843 - 847
Black mud	847 - 848
Black mud and salt	848 - 849
Hard salt; gray, brown, and black clay	849 - 854
Hard gray clay with layers of other clays and muds; a little crystal salt	854 - 863½
Softer material	863½- ...
Hard gray clay with some black and brown	863½- 865
Same material; shows white fragments of gypsum	865 - 878
Hard salt; gray-brown clay and black mud	878 - 887½
Hard gray clay; streaks of brown clay and crystals of salt	887½- 890
Same material, with black streaks	890 - 900
Same material. At 904 ft, about 1 in. very hard	900 - 904
Hard gray clay; streaks brown and black; crystal salt	904 - 909
Salt; gray mud; black clay	909 - ...
Softer material; gray clay	909 - 911
Black, brown, and gray clays	911 - 921
Salt; tough gray, brown, and black clays	921 - 922
Gray and black clay	922 - 928
Harder material, a little salt	928 - 929
Black and gray clay	929 - 937
Black mud	937 - 943
Black and gray mud; salt	943 - 945
Do	945 - 958
Gray, black, and brown clay	958 - 960
Gray and black clay	960 - 973
Gray and black clay and salt; a few fragments of gypsum	973 - 978
Gray and black clay; a very little salt; a few fragments of gypsum	978 -1,000
Gray clay	1,000 -1,000
Depth of well	1,000

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TABLE 19.—*Logs of wells drilled by Pacific Coast Borax Co.*—Continued

Well 2, by highway across the valley, at middle of Devils Golf Course NE¼ sec. [illegible], T. [illegible] N., R. 1 E.

Description	Depth (feet)
Salt	0 – 0½
Brown clay and salt and anhydrite	0½ – 5
Brown and gray clay and salt and anhydrite	5 – 6
Black bituminous clay, salt and anhydrite; very hard	6 – 13
Do	13 – 14½
Same material, with a little brown clay	14½ – 23
Softer brown clay, salt and anhydrite	23 – 26
Hard material, otherwise apparently same	26 – 28
Soft material, same	28 – 29
Hard material, same	29 – 30
Salt, with a little brown clay and anhydrite	30 – 31
A little salt and anhydrite and black and brown clays	31 – 40
Do	40 – 65
Same material, but more salt	65 – 70
Same material with less salt, and consequently softer	70 – 72
Black clay and very few crystals	72 – 77
Black and brown clay, with a little salt and anhydrite	77 – 90
Black clay, with very little salt and anhydrite, and a few tufts ulexite (cotton ball)	90 – 100
Black and gray clay; crystal strain	100 – 120
Harder material; a little calcium carbonate appears as a cement	120 – 121
Partly cemented black clay	121 – 129
Hard material; igneous breccia, principally basalt, with a little clay and limestone fragments, all more or less cemented with calcium carbonate	129 – 130
Same material, but softer	130 – 133
Same material, but hard	133 – 150
Black clay and breccia, partly cemented	150 – 155
Breccia of basalt, with a little granite, quartz and limestone; cemented in streaks; mostly the size of coarse sand; absorbs much water from drill hole	155 – 211
Boulders and calcium carbonate cement	211 – 213½
A little clay, and softer	213½ – 216
Angular gravel, principally basalt	216 – 217
Cemented gray clay; a few basalt fragments	217 – 219
Hard cemented material	219 – 220
Softer material; considerable clay	220 – 223
Hard material; gravel and boulders; cement gradually diminishing	223 – 232
Gravel, as above	232 – 235
Brown clay and sand and gravel	235 – 238
Gravel and small boulders	238 – 240
Same material; cemented	240 – 243½
Hard cemented gravel	243½ – 247
Softer material; contains some clay	247 – 248
Hard material; no clay	248 – 249
Softer material; some brown and light-gray clay	249 – 251
Do	251 – 253
Hard cemented sand and gravel	253 – 255
Brown clay	255 – 258
Hard cement; a little clay	258 – 260
Soft material; brown clay and gravel	260 – 263
Basalt boulder	263 – 265
Brown clays and gravel	265 – 268
Brown clay and gravel	268 – 270
Basalt, apparently two boulders	270 – 271
Brown clay; occasionally small boulders	271 – 273
Brown clay and gravel	273 – 275
Basalt boulder	275 – 276
Brown clay and a little siliceous sand	276 – 301
Streaks brown to gray clay	301 – 307
No record	307 – 311
Brown clay	311 – 421
Yellowish-brown clay	421 – 432
Same material with a little fine sand	432 – 438
Brownish-yellow clay	438 – 459
Red clay	459 – 462
Red clay; a little yellow clay	462 – 465
Brown, red, and yellow clays	465 – 467
Brown and gray clay, alternated	467 – 471
Mostly gray clay	471 – 481
Dark-brown clay and particles of quartz	481 – 483
Do	483 – 493
Dark-brown clay; a little sand; possibly slightly cemented	493 – 496
Hard material; brown clay; no evidence of calcium carbonate, but a little gypsum appears, which may possibly act as a cementing material	496 – 499
Gray and brown clay and sand	499 – 503
Same material; driller reports cement, but none shows in the sample	503 – 505
Gray and brown clay and sand	505 – 506
Gray clay; very little sand	506 – 508
No record	508 – 512[illegible]
Brown clay and slightly cemented gravel	512½ – 514[illegible]
Basalt boulder	514½ – 514
Gray and black clay; a little fine gravel; soft and caving	514 – 517
Gravel, principally basalt, a few particles of quartz, limestone and gypsum; driller reports hard cement, but sample gives no evidence of this	517 – 524
Depth of well	524

Well [illegible], Cottonball Basin, 2 miles northwest of Harmony Borax, SW¼ sec. [illegible], T. [illegible] N., R. 1 E.

Description	Depth (feet)
Salt, containing small amount of thenardite (sodium sulfate), borax, and a little yellowish-brown clay	0 – 2½
Yellow clay	2½ – 5
Soft salt and yellow clay	5 – 5½
Soft yellow clay	5½ – 18
Soft yellow clay; a few crystals anhydrite	18 – 19
Soft yellow clay	19 – 31
Soft yellow clay; a little fine salt and anhydrite	31 – 33
Black and green clay, with a little salt and anhydrite	33 – 37
Brown clay and salt	37 – 37½
Black, green and brown clay; a few crystals of salt and anhydrite	37½ – 38
Salt and a little pale-blue clay	38 – 43
Salt; a little anhydrite; blue and brown clays	43 – 44
Salt and clays, changing from brown to gray	44 – 46
Salt and gray clay; a few streaks of black clay	46 – 49
Gray and black clay; a little salt	49 – 53
Salt; a little anhydrite; some gray clay	53½ – 58
Black clay and salt	58 – 60
Gray clay and salt	60 – 68
Salt and a very little gray clay	68 – 73
Salt and a little gray, black, and brown clay	73 – 74
Gray clay and very little salt	74 – 7[illegible]
Salt and a little gray and brown clay; a few tufts of ulexite	75 – 7[illegible]
Same material, except clay principally brown	79 – 8[illegible]
Salt, with a very little clay; a few tufts of ulexite	83 – 8[illegible]
Black clay and salt	85 – 8[illegible]
Salt and a little black clay and ulexite	86 – 9[illegible]
Salt and a little ulexite	90 – 9[illegible]
Salt; gray and black clay; a little ulexite	93 – 10[illegible]
Salt; a little ulexite; a very little gray clay	108 – 11[illegible]

TABLE 19.—*Logs of wells drilled by Pacific Coast Borax Co.*—Continued

Description	Depth (feet)
Well 2, Cottonball Basin, 2 miles northwest of Harmony Borax. SW¼ sec. 32, T. 28 N., R. 1 E.—Continued	
Salt, and a little gray, black, and brown clay	110 - 125
Hard salt and a little ulexite	125 - 127
Salt and some gray, brown, and black clay	127 - 134
Salt; black clay; a little ulexite	134 - 137
Hard salt and a little gray clay	137 - 141
Salt; black clay; a little ulexite	141 - 142
Salt, and a little gray, brown, and black clay	142 - 146
Salt and black clay	146 - 154
Salt and very little clay (black and gray)	154 - 162
A little brown clay, otherwise same	162 - 186
Black clay and salt, varying in proportions	186 - 241
Salt and very little black clay	241 - 246
Salt; a little brown clay; a few crystals ulexite	246 - 255
Hard salt and a little ulexite	255 - 260
Hard salt and a little black clay	260 - 263
Hard salt and a little brown clay	263 - 264
Light-blue clay	264 - 265
Brown clay and salt	265 - 267
Brown clay and salt, and a little ulexite	267 - 270
Brown clay and salt	270 - 275
Salt and a little gray clay, and considerable ulexite	275 - 279
Salt and a little gray and black clay	279 - 284
Softer material, less salt	284 - 284½
Hard salt; a little black, brown, and gray clay	284½ - 288½
Salt; considerable brown clay; a little ulexite	288½ - 290
Salt and a little brown clay	290 - 292
Salt and brown and gray clays	292 - 298
Salt and brown clay, and a little ulexite	298 - 301½
Brown clay, containing fine crystals of salt and anhydrite	301½ - 320
Same material, except more anhydrite	320 - 323
Clay, with a little salt, anhydrite and ulexite	323 - 333
Brown clay and fine sand, containing a little salt and anhydrite	333 - 345
Same material (sand negligible)	345 - 360
Brown and gray clay, and a little salt and anhydrite	360 - 369
Bluish-gray clay and salt	369 - 373
Salt, with a little anhydrite and brown and gray clay	373 - 375
Brown clay and salt	375 - 376
Hard salt, and a little anhydrite and brown clay	376 - 385½
Pale-greenish-gray clay, and salt	385½ - 392
Hard salt, with very little gray clay and ulexite	392 - 398
Hard salt, and a little gray clay	398 - 400
Salt, and gray and brown clay in thin strata	400 - 405
A little gray clay, and small crystals of salt	405 - 410
Small salt crystals, and a little gray and brown clay	410 - 415
Fairly soft salt; gray and a little brown clay; occasionally a few tufts of ulexite	415 - 465
Hard salt and a little gray clay	465 - 495
Softer material, sticky and probably wet	495 - 514
Salt and a little anhydrite	514 - 519
Tough black clay, with very little salt	519 - 520
Salt and black and brown clay and a little sand	520 - 525
Same material, and sand decreasing in amount	525 - 538
Salt and tough black clay	538 - 540
Hard salt and a little gray-brown clay and ulexite	540 - 545
Very hard material (probably a salt stratum)	545 - 545½
Salt crystals, with a little gray and brown clay and sand	545½ - 570
Harder material. No sand	570 - 581
Gray-brown clay, very sticky	581 - 592
Coarse salt, and large crystals of anhydrite; a little brown and gray clay and traces of calcium-carbonate cement	592 - 595
Coarse salt, and some anhydrite, a little black and gray clay, and a few fragments thenardite	595 - 597
Same material, with a little blue clay	597 - 599
Same material, with a little brown clay	599 - 606
Blue-gray clay, about 50 percent; salt, anhydrite, and thenardite in about equal proportions (see note on thenardite below)	606 - 612
Blue and soft brown clays, otherwise same material	612 - 615
Brown clay; salt, anhydrite, and thenardite; a little ulexite and few borate fragments, apparently colemanite; some traces of calcium carbonate cement	615 - 617
Tough blue clay and salt	617 - 620
About 50 percent clay, blue, black and brown; crystals chiefly of thenardite, with a little salt and anhydrite	620 - 625
Same material; also a few nodules of clay, showing traces of cement	625 - 628½
Chiefly salt and clay; a little anhydrite and thenardite	628½ - 630
Salt and clay, and notable sand; no other crystals	630 - 635
Gray, brown, and blue clay; very few crystals, of salt only	635 - 640
Tough, dry clay, as above	640 - 646
Brown clay and sand, with a little salt and anhydrite	646 - 665
Brown clay, and a few crystals of salt and anhydrite; very little blue clay and sand	665 - 670
Bluish-green clay, and sand and salt	670 - 671
Very little sand, otherwise same	671 - 677½
90 percent clay, brown, blue, and gray; crystals of salt only	677½ - 680½
Salt stratum, hard	680½ - 681
Brown and gray clay, and a little salt	681 - 690
Same material, with some blue-green clay	690 - 695
Tough clays, gray, green and black; about 10 percent sand	695 - 704
Tough light-blue clay	704 - 706
Clays, brown, blue, black, and gray; about 50 percent salt	706 - 720
No record	720 - 721½
About 25 percent salt; remainder blue clay	721½ - 724
About 10 percent salt; blue and black clay	724 - 731
Very little salt, and no other crystals; brown clay, with sand increasing from 0 to 50 percent	731 - 750
Sand decreases from 50 to 5 percent	750 - 765
Brown, gray, and blue clay, and a little salt and anhydrite	765 - 770
Gray clay, and a little salt and anhydrite	770 - 775
About 10 percent salt, and very little anhydrite; remainder, gray, blue, and black clay	775 - 777
Gray, black, and brown clay, and a little salt	777 - 790
Brown clay and a little sand	790 - 795
Brown clay, and sand increasing from trace to 25 percent	795 - 808
Very tough blue clay and sand	808 - 810
Tough brown clay, and a little sand	810 - 812
About 80 percent gray and brown clay, 15 percent coarse sand, 5 percent crystals consisting of salt and very little anhydrite	812 - 815
Dark-brown sandy clay and a little salt; a few traces of ulexite	815 - 819
Sandy gray and brown clay, otherwise same	819 - 820

TABLE 19.—*Logs of wells drilled by Pacific Coast Borax Co.*—Continued

Description	Depth (feet)
Well 2, Cottonball Basin, 2 miles northwest of Harmony Borax, SW¼ sec. 32, T. 28 N., R. 1 E.—Continued	
Same material, except no ulexite	820 - 845
Same material, with a little black and blue clay in addition	845 - 850
Sandy gray and brown clay	850 - 865
Sandy gray clay	865 - 880
Very tough gray and black clay, with a few crystals of salt and anhydrite	880 - 890
Tough light-gray clay and salt	890 - 896½
Tough light-blue clay and salt, and a little thenardite	896½ - 900
Gray, black, and brown clay, and a little salt and thenardite	900 - 904
Mostly brown clay, about 5 percent sand, and a little salt and thenardite	904 - 909
Very hard blue and black clay, with a little salt and thenardite	909 - 911
Tough light-blue clay and salt; a little brown clay and ulexite	911 - 915
Light-gray clay and salt, and a little thenardite and ulexite	915 - 916
Same material, except no ulexite	916 - 920
Light-blue clay and salt; a little brown and black clay	920 - 925
Dark-brown sandy clay and a little salt	925 - 927
Very hard, gray, black, and brown clay, with a little salt and thenardite	927 - 9
Do	930 - 9
Gray clay and considerable salt (soft)	935 - 9
Gray and brown clays, otherwise same	945 - 9
Dark-brown sandy clay, with a little salt and anhydrite	951 - 9
Tough grayish-blue clay, with about 25 percent salt and a little thenardite	959 - 9
Gray clay; salt and considerable thenardite	965 - 9
Light-gray clay and sand	974 - 9
Soft gray sandy clay; salt and thenardite	975 - 9
Light-brown-gray clay, with some sand and about 10 percent salt and thenardite	984 - 9
Clay, mostly gray and brown, with a little sand, thenardite, and about 25 percent salt	993 - 9
Gray-brown clay and sand; about 10 percent salt; traces of thenardite	996 - 9
Same material, but harder	998 -1, 0
Gray clay and some sand, about 10 percent salt, with a little anhydrite	1, 000 -1, 0
Depth of well	1, 009½

PLEISTOCENE(?) AND RECENT(?) DEPOSITS

SAND AND SILT IN THE PLAYA

Sandy playa (and lake?) deposits crop out at the edge of the saltpan and foot of the gravel fans. The sand consists of very fine grained to medium-grained brown sand, most of which is rounded or subrounded quartz. Feldspar is abundant; there is some mica and hornblende. Depending on the source, there may be considerable amounts of volcanic glass or other volcanic rocks and of clastic grains of dolomite or limestone. A calichelike layer of salts occurs about a foot below the surface.

The sand is 3–10 feet thick and rests on gravel. Presumably there is more sand below the bed of gravel, for the position is where the facies would intertongue.

Originally, the sand must have graded into gravel on the fans; but the transition beds have been removed by erosion, and the sand now forms a low cuesta, 2–4 feet high, facing the gravel fans. Panward the sand grades into silt, which becomes increasingly clayey toward the center of the saltpan. Borings in the middle of the saltpan indicate that the sand deposit is 35–50 feet thick. There this deposit is overlain by the crust of salts forming the present saltpan; it overlies a layer of rock salt a few feet thick.

No fossils were found in the sand or silt. The sand is older than the sand dunes that can be equated with the earliest bow-and-arrow occupations in this area. It is older than the calichelike layer of salts contained in it, which is attributed to evaporation of ground water at the time of a Recent but pre-Christian lake (p. A79). The sand is considerably dissected. I crossed by numerous small washes draining from gravel fans to the saltpan, and it is being overlappe the No. 4 gravel that is being moved panward at pres

NO. 3 GRAVEL

The No. 3 gravel (pl. 2) differs from the No. several ways. The deposits contain less caliche generally are less well indurated. The cobbles and bles on the surface are firm and show little sign of integrating; the rocks are not angular but are round. Although not disintegrated, cobbles and bles on some of these deposits have thin weathe rinds; other deposits lack even this. The gravels a dark stain of desert varnish. Over a broad sur the stain may be darker than it is on the older gra because the No. 3 gravels are firm, whereas fang erates of the Funeral Formation and No. 2 gravel crumbly and the varnish there is partly destroye

The No. 3 gravel is much better stratified in co and fine-grained layers than is the No. 2 gravel. range in grain size is substantially less and the pr tion of gravel to sand is higher, although few bou are more than a foot in diameter. Where there nearby source of large boulders in erosion remnan the No. 2 gravel, some of these are reworked int No. 3 gravel, but such reworked boulders are few.

The surfaces on the No. 3 gravel are rough (fig The cobbles and small boulders are in ridges—na levees—1 or 2 fet high and as much as 10 feet wide. Washes between the levees are about the same width as the levees. The gravels also occur in small fanlike mounds that choke washes and disrupt the drainage. Nowhere is there desert pavement on these deposits like that on the No. 2 and older gravel. The range in size of the gravels on the surface of the No. 3 is the same as within the deposit.

Three kinds of surfaces have formed on the No. 3 gravel. Surfaces that have not been subject to flooding or washing are only a little less smooth than the desert pavements on the adjacent older gravels. Such surfaces are rough only because the ill-sorted small boulders, cobbles, and pebbles stand at different heights and are distributed irregularly on the surface. The stones are darkly stained with desert varnish.

Surfaces that have been subject to flooding, but not recently, are composed of levees of small boulders along the sides of washes floored with pebbles, and both the levees and washes are darkly stained with desert varnish (fig. 53). Desert varnish is thicker and darker on stones on these first two types of surfaces than on any other gravel deposits in this part of Death Valley.

The third type of surface is like the second, except for recent washing. On these surfaces the levees are stained with desert varnish, but the pebble floor of the wash is not. This third kind of surface grades into that of the No. 4 gravel.

The surfaces on the No. 3 gravel similarly grade into those that have formed on the No. 2 gravel. Where surfaces on the No. 2 have been overriden by flash floods, a layer of firm cobbles and pebbles overlies the pavement of partly disintegrated slabs and flakes. In these places the firm cobbles and pebbles form low ridges on the pavement, and the old desert pavement forms the beds of the little washes between the natural levees. Other surfaces on the No. 2 are dissected by shallow washes which have become mantled with firm cobbles and pebbles, leaving narrow interstream areas capped with the old desert pavement. A third kind of gradation is where the No. 3 gravel has been derived by erosion of old disintegrated gravel; depending upon how far such gravel was transported, there may be enough angular stones to form a surface like that on an older deposit. Such surfaces, however, lack the silt layer that is characteristic of older desert pavement.

Much of the ground shown as No. 3 gravel actually is only a thin veneer of this gravel on an eroded surface of the No. 2. The No. 3 gravel is neither as thick nor as extensive as the No. 2.

In general, the surface of the No. 3 gravel is lower than that of the No. 2 and generally less than 10 feet above the No. 4. But the No. 3 gravel overlaps the lower edges of the No. 2, and at such places has accumulated in small fans on top of it. Conversely, the No. 3 gravel is overlapped by the No. 4 (figs. 55, 64).

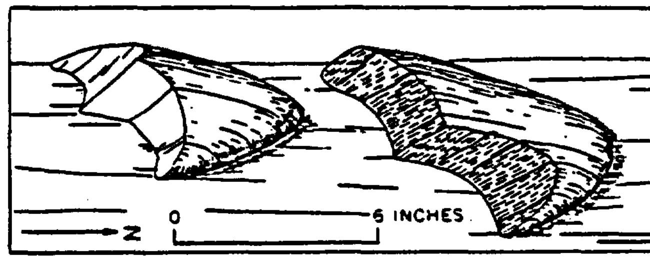

FIGURE 54.—Wind-faceted cobbles. Argillite (left) is smoothly faceted; limestone (right) has rillen on the facets. Both specimens oriented as in the field.

The Funeral Formation and the No. 2 gravel generally are without vegetation, but the No. 3 gravel generally has a sparse growth of shrubs along the shallow washes between the natural levees of cobbles and small boulders. This reflects the difference in permeability and runoff on the two surfaces. Runoff is greater on smooth desert pavement than it is on the rougher surfaces of the No. 3 gravel, and the ground is accordingly more xeric and less suitable for plant growth (see Hunt, 1965).

Pebbles and cobbles on the surface of the No. 3 gravel are wind faceted (fig. 54) at several localities, for example, along the south side of the Hanaupah Canyon fan 1–1½ miles due west of Eagle Borax, on a bench at the mouth of the wash at the north end of the Artists Drive fault blocks (NE¼NE¼ sec. 15, T. 26 N., R. 1 E.), and on the Salt Creek Hills. In the latter area some stone artifacts are clearly etched by sandblasting. Glass bottles that have been exposed are frosted and etched. The wind-facted pebbles may have been developing their facets over a long period of time, but certainly some of the shaping is Recent.

At several places the No. 3 gravel has been displaced by small faults. At the Hanaupah escarpment, 1 mile west of Shortys Well, the No. 3 gravel is displaced 6 feet along a fault that displaces the No. 2 gravel 75 feet (fig. 78). At most places, though, the No. 3 gravel overlaps faults without being displaced. Good examples are beside the highway 2 miles south of Bennetts Well (fig. 55) and at the south edge of the Trail Canyon fan 1½ miles southwest of the junction of the Trail Canyon road and West Side highway.

The No. 3 gravel is old enough to have been eroded into low benches and to have developed extensive desert varnish on the surface. Numerous archeologic sites on the gravel indicate that the bow-and-arrow and pottery occupations at those places are later than the No. 3 gravel. Further, the No. 3 gravel everywhere is darkly stained with desert varnish, but archeologic sites of the bow-and-arrow occupations are not. There

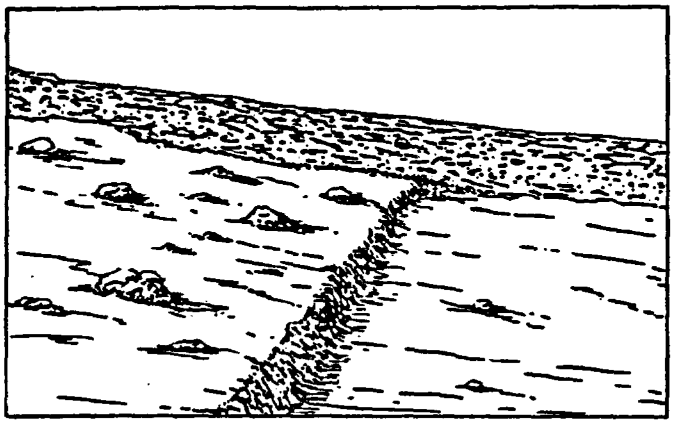

FIGURE 55.—No. 2 gravel (foreground) displaced 6 feet by a fault that is overlapped and buried by No. 3 gravel (distance). Locality is by West Side highway 2 miles south of Bennetts Well. View north.

is little reason to doubt that the gravel everywhere is older than these archeologic sites and antedates the Christian era. Probably the No. 3 gravel includes deposits that are early Recent in age and other deposits as old as late Pleistocene.

DEPOSITS OF TRAVERTINE AND CALICHE CEMENT IN GRAVEL

Travertine has been deposited in mounds at and near each of the large springs issuing along faults west of the Funeral Mountains, and a small mound has been built on the upper part of the Trail Canyon fan. The deposits are nearly pure calcium carbonate.

The largest deposit is at Nevares Spring at the foot of the mountains 2 miles east of Park Village.

Travertine has been deposited at Travertine and Texas Springs, and between them are some mounds of travertine that have become isolated by erosion. One of these deposits drapes over the side of Furnace Creek Wash and extends to the bed of the wash (fig. 56),

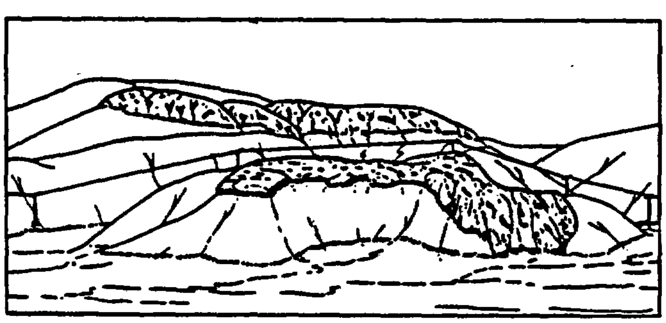

FIGURE 56.—Travertine deposit overlapping the bank of Furnace Creek Wash half a mile above the mouth of the wash. The old spring, now dry, that deposited the travertine was at the high mass of travertine that can be seen beyond the telephone line.

clearly dating the travertine as younger than this par of the gorge of Furnace Creek Wash.

On the gravel fan at Trail Canyon, a mile below th mouth of the canyon, is a mound of travertine 5 fee high and 30–50 feet in diameter. Another travertin deposit on the west side of Death Valley is at the sout base of the hill of basalt south of the mouth of Black water Wash. The deposit is at an altitude of 200 fe in a cove at the toe of a field of basalt boulders over lying tuff. The travertine probably was derived fro the carbonate caliche in the basaltic boulders up the hil side and probably dates from a time when there was spring here.

Travertine obviously is being deposited at Nevare Texas, Travertine, and similar springs at present, y much, probably most, of the travertine is an old deposi probably dating back to late Pleistocene time. Son of the deposits are at locations where springs have drie up. Other deposits are old enough to have been isolate by erosion from the spring areas. Projectile points types characteristic of the early occupations have bee found on the surface of some mounds.

A few fossils were found, but they are not meanin ful. At Nevares Spring, 4 feet below the surface, teeth were found and identified as mountain she (*Ovis canadenis* Shaw) by G. E. Lewis, of the U. Geological Survey, and C. B. Schultz and L. G. Tanne of the University of Nebraska State Museum. Son mollusk shells from the same layers of travertine Nevares Spring (U.S.G.S. Cenozoic loc. 21675) we identified by D. W. Taylor, of the U.S. Geological Su vey as Hydrobiidae indeterminate, a fresh-water sna

Some shells from travertine at Triangle Sprin (U.S.G.S. Cenozoic loc. 21575) on the northwest si of Mesquite Flat also were identified by Taylor follows:

- *Pisidium* sp., a fresh-water clam
- Hydrobiidae, 2 indeterminate species probably representi 2 genera of fresh-water snails
- *Physa*, a fresh-water snail
- *Vertigo*, a land snail
- *cf. Succinea*, a land snail

Taylor (written commun., 1961) offers the followi ecologic interpretation of these species:

> The two terrestrial species are inhabitants of moist situatio such as vegetation along streams, beside ponds, or in mar places. The fresh-water species do not inhabit wide range salinity; the water certainly was fresh rather than brack The water temperatures may have been warm, but hot—possibly as high as 80° F.
>
> The living molluscan fauna of the Death Valley area is sentially unknown. For this reason the significance of mollusks cannot be evaluated satisfactorily. Perhaps all species represented by the fossils are living; perhaps only so

Fresh-water snails in another collection from the irrigation ditch at Furnace Creek Ranch were identified by Taylor as:

> *Helisoma duryi seminole* Pilsbry, a Floridian species, probably introduced through aquaria
> *Physa*

The calcium carbonate caliche that cements layers of gravel on the fans is well developed where the gravels overlap the fine-grained Tertiary playa deposits, places favoring perched ground water. Such cemented ledges are extensive where the gravels overlap the Tertiary rocks below Nevares Spring and along the west edge of the Texas Spring syncline, where ground water comes to the surface. However, for reasons that are not obvious, the caliche also is well developed on the fans of Galena and Six Spring Canyons.

Although most of the caliche is calcium carbonate, there is considerable calcium sulfate caliche locally, especially along the foot of the Black Mountains, as had been noted in the description of the Funeral Formation on Artists Drive (p. A64). Certainly the greater part, and perhaps all, of the caliche in these gravel deposits has been deposited by ground water, or more likely, by water in the capillary fringe above the water table. The best evidence for this is the common occurrence of well-developed caliche where there is a perched ephermeral water table.

That much of the caliche is old, perhaps late Pleistocene in age, is indicated by the occurrence of earliest archeological sites (Death Valley I and Death Valley II; Hunt A. P., 1960) at shelters or ledges formed by the caliche.

Trace elements in the travertine are given in table 20. They are much the same as in the calcite veins cutting the Funeral Formation.

TABLE 20.—*Trace elements in spring-deposited travertine and in calcite vein in Funeral Formation*

[Spectrographic analyses by E. F. Cooley, U.S. Geol. Survey, values in parts per million, except Mg, which is given in percent]

	B	Ba	Be	Bi	Cd	Co	Cr	Cu	Ga	Ge
Travertine [1]	<10	150	<1	10	50	<5	<10	5	20	20
Calcite vein [2]	<10	10	<1	10	50	<5	<10	5	20	20

	In	La	Mg	Mn	Mo	Nb	Ni	Pb	Sb	Sc
Travertine [1]	10	<50	1.5	10	<5	<50	<5	10	200	<10
Calcite vein [2]	10	<50	1	10	<5	<50	<5	<10	200	<10

	Sn	Sr	Ta	Ti	Tl	V	W	Y	Zr
Travertine [1]	10	2,000	50	30	100	10	100	<10	<10
Calcite vein [2]	10	500	50	10	100	10	100	<10	<10

[1] Travertine from Texas Spring.
[2] Calcite vein from near foot of Funeral Mountains.

RECENT DEPOSITS

SHORELINE FEATURES OF THE RECENT LAKES

Near the base of the gravel fans on the west side of Death Valley, at about 240 feet below sea level, the desert varnish on the fan gravels abruptly ends. The color change, from darkly varnished gravels above to light unvarnished gravels below, follows the contour and occurs within a vertical range of about 5 feet. This contour also marks the upper limit of highly saliferous ground and is interpreted to be the high watermark of a lake (fig. 57).

The shoreline shows especially well in the cove north of West Side Borax Camp on the west side of Cottonball Basin (fig. 58). The salt-impregnated ground extends to a uniform level on the fans around the head of the cove and on a little butte of limestone that was an island in the cove. Above the salt-impregnated ground is comparatively salt-free gravel on the fans and salt-free colluvium on the butte.

At the east foot of the Salt Creek Hills the shoreline is impressed on the alluvial bank of an arroyo (fig. 59).

The shoreline is distinct also where the highway crosses Death Valley at the west foot of the fans on the Artists Drive fault block. There it is marked by a striking change in salinity of the ground which coincides with a small terrace along a contour 260 feet below sea level (fig. 60). The hill of the Funeral Formation just north of the highway has a strong caliche of gypsum above the terrace, and all the ground on the hillside is heavily impregnated with gypsum. Around the foot of the hill, at and below the terrace, the ground is impregnated with rock salt. A layer of rock salt, evidently marking the capillary fringe above the old water table, extends into the gypsiferous gravel. This layer of rock salt is broken by polygonal cracks that are manifested at the surface by shallow troughs in which pebbles have collected—a type of patterned ground distinctive of this shoreline (Hunt and Washburn, in Hunt and others, 1965).

Similar changes in salinity at about this same level appear at many places along the east side of Badwater Basin (fig. 61), and in the cove south of Copper Canyon, varnished gravel extends down to this level.

Sandy or other permeable beds along the edge of the saltpan generally have a calichelike layer of salts 1–4 inches thick and 3–15 inches below the surface. The composition of the salts varies from one part of the saltpan to another. Where little ground water moves into the saltpan, the salts in the caliche are mostly rock salt. Where much ground water moves into the saltpan, sodium chloride is flushed to the interior of the saltpan, and the caliche layer around the edge is mostly composed of sulfates. Along the north and east sides of

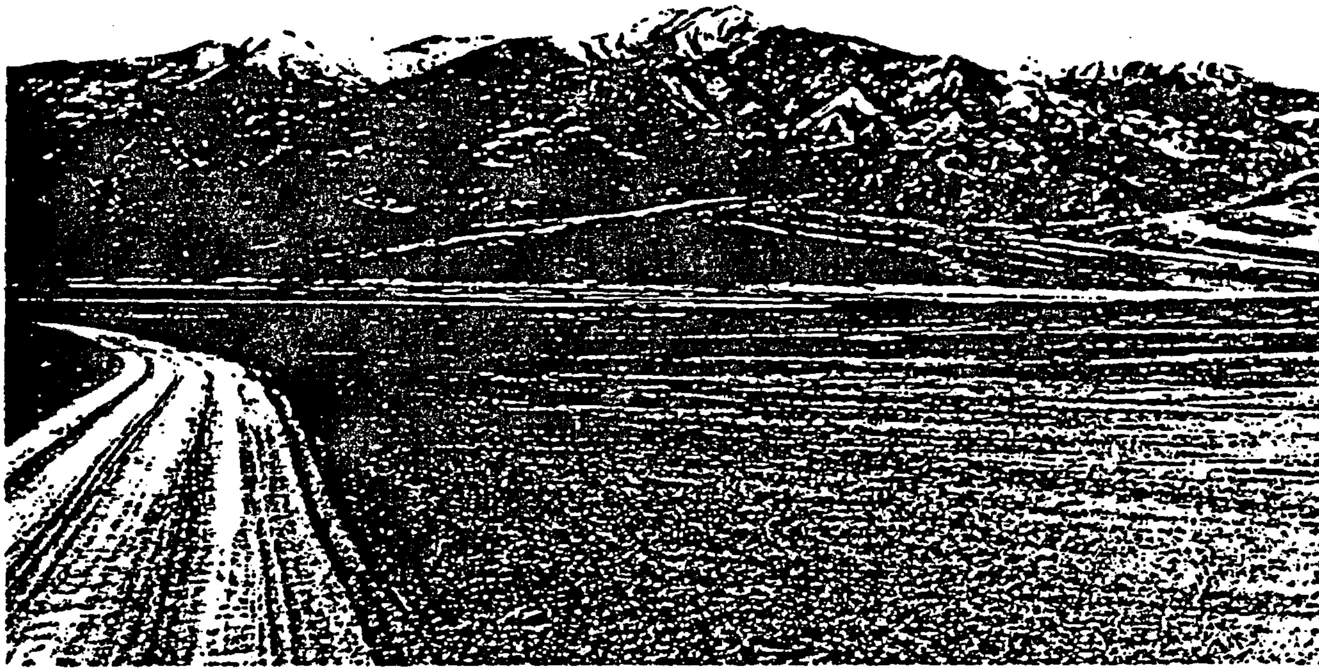

FIGURE 57.—View west across Death Valley to Trail Canyon fan showing how the lower limit of the desert varnish follows the contour. The light ground below the darkly varnished gravel not only is without desert varnish, it is highly saliferous.

FIGURE 58.—Shoreline of the Recent lake marked by change in ground. Highly saliferous ground, marked by pickleweed mounds in the foreground, forms a salt flat that ends along a contour near the foot of the gravel fans and around the base of the little hill in the center of the cove. View northwest across the cove north of the West Side Borax Camp.

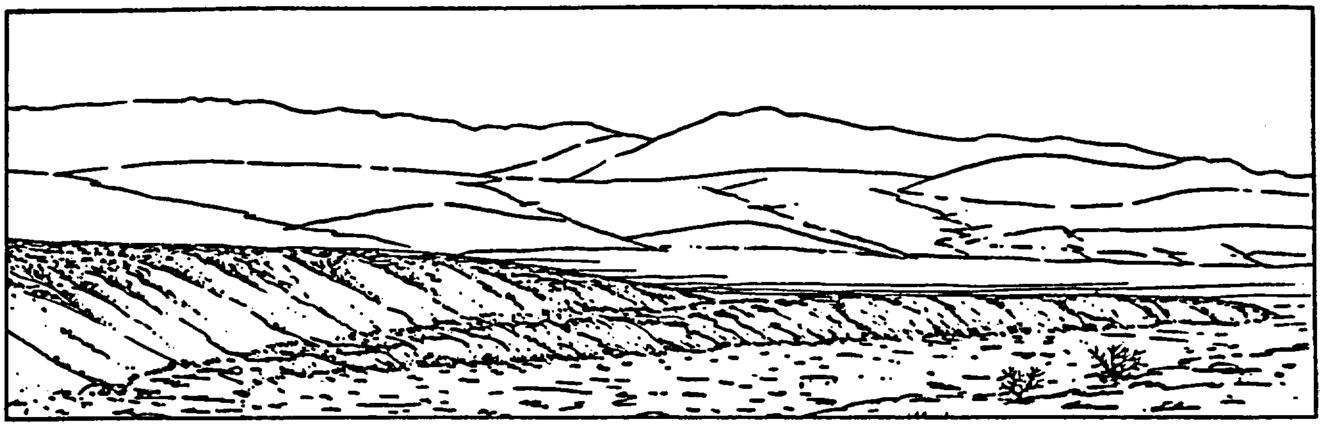

FIGURE 59.—Shoreline cut in alluvial bank of wash at east foot of the Salt Creek Hills (altitude about −240 ft; SW¼SE¼ sec. 10, T. 16 S., R. 46 E.). View east.

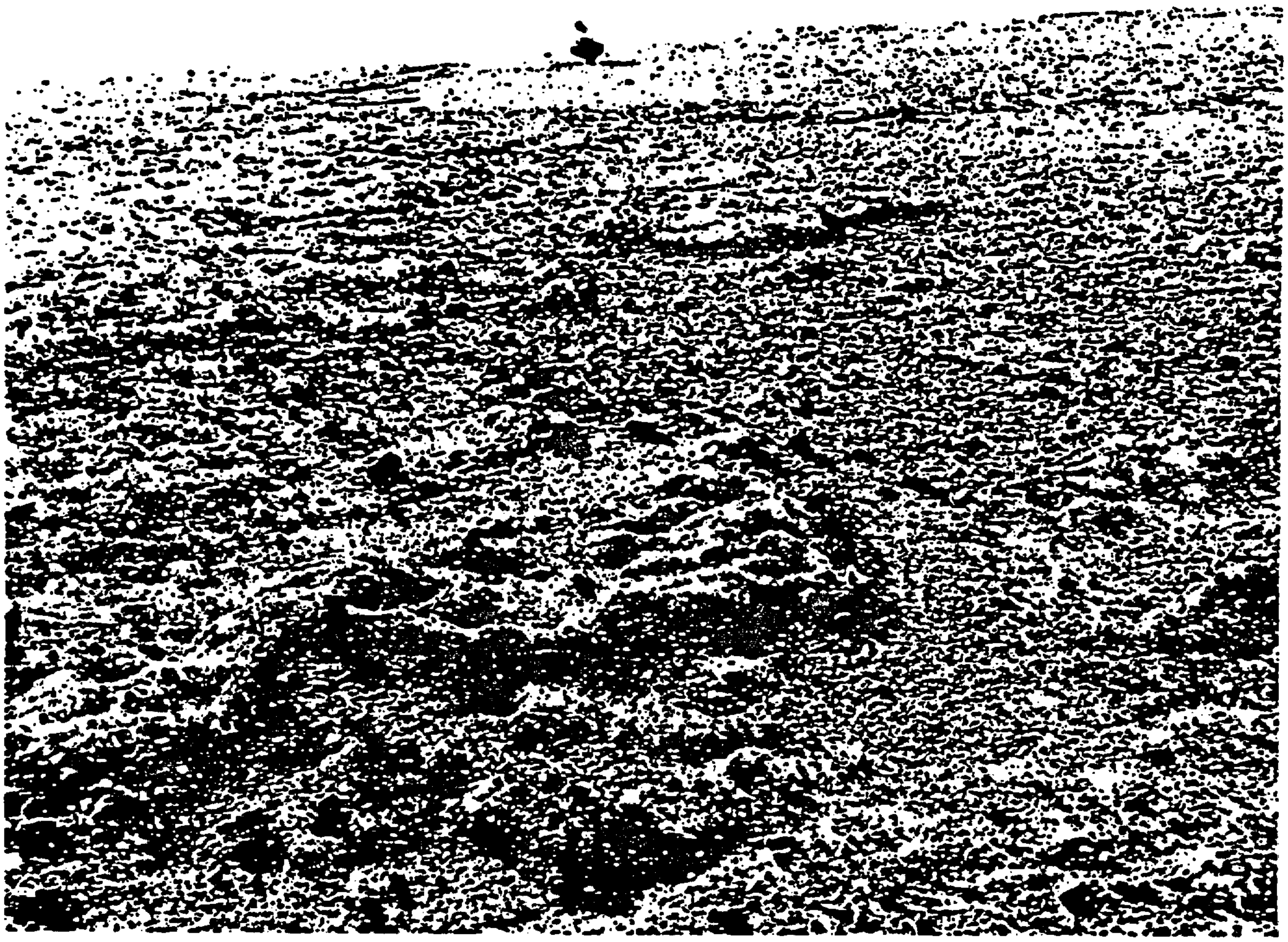

FIGURE 60.—View of narrow terrace around the foot of the Artists Drive fault block, 260 feet below sea level, which is interpreted to be the shoreline of a Recent lake. The ground below the terrace is roughened by heaving of rock salt; the ground above the terrace is smooth and impregnated with gypsum. View northwest, from half a mile north of the highway across the Devils Golf Course.

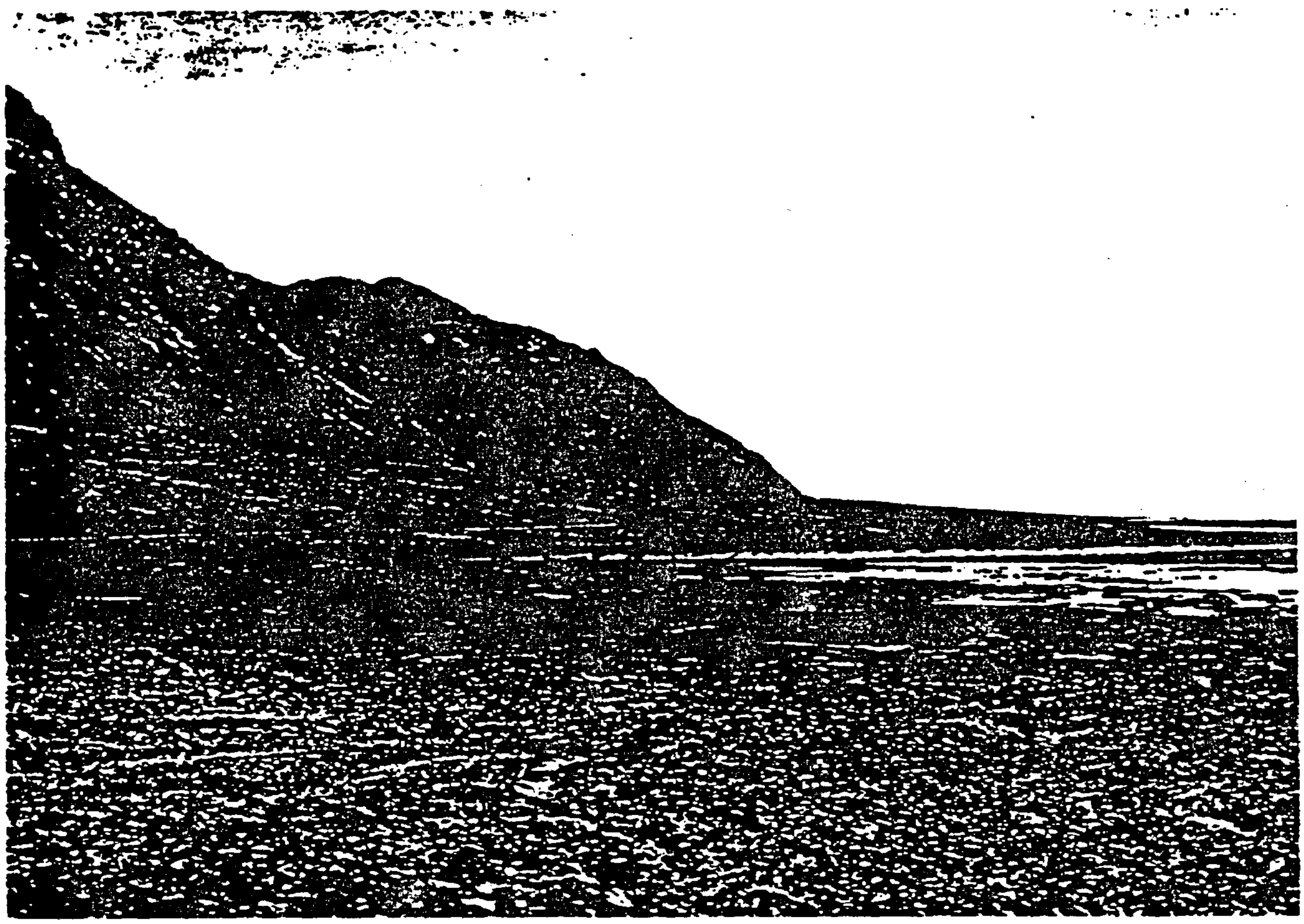

FIGURE 61.—View of shoreline at foot of fan north of Coffin Canyon. The greater salt content of the ground below the shoreline helps retain moisture and appears dark.

Cottonball Basin, where the ratio of calcium to sodium is low, the sulfates are mostly sodium sulfate. Along the west side of Badwater Basin, where the calcium-sodium ratio is high, the sulfates are mostly calcium sulfate.

At most places these caliche layers are 8–10 feet above the present water table and are 4–5 feet above the present capillary fringe. There is little doubt that they formed by evaporation of ground water at a time when the water table was higher than it is now, and very probably this occurred at the time the Recent lake filled the basin. There was more moisture at that time, and the caliche deposits largely coincide with the shoreline of that lake.

The lake that produced these features antedates sand dunes (p. A86) that have formed on the old lake floor along the west side of Badwater Basin. These sand dunes must have formed during the last 2,000 years, for they contain artifacts representing late archeological sites—specifically the Death Valley IV (pottery) occupation (about A.D. 1000) and the Death Valley III (prepottery but bow-and-arrow) occupation (about A.D. 1) (p. A87). The absence in the sand dunes of artifacts representing the pre-bow-and-arrow occupations is consistent with the idea that the dunes are younger than these occupations.

The lake must be older than the deposits of massive gypsum, for these deposits are not impregnated with rock salt, though they would have been flooded by the lake. The lake is younger than the sandy and silty beds that are exposed around the edge of the saltpan at the foot of the gravel fans, because these beds contain calichelike layers of salts and other salt impregnations that evidently were deposited when they were flooded by the young lake.

The lake probably is a feature of the Recent pluvial period that is widely represented by Recent but pre-pottery and pre-bow-and-arrow alluvial deposits in other parts of the Southwest (Hunt, 1953, p. 3).

Since the time of this lake the valley floor in Badwater Basin has been tilted eastward; the shoreline is 20 feet lower along the east side of that basin than it is along the west side.

The deposits that formed in an around the edge of the lake are mostly salines and constitute the Death Valley saltpan. The pan covers more than 200 square miles, all of it below sea level and most of it between 270 and 282 feet below sea level. The salt crust on the saltpan ranges from a few inches to a few feet thick; it is underlain by silt and clay. At the center of the pan the salts in the crust are mostly chlorides. These chlorides are surrounded by a narrow discontinuous zone in which the salts are chiefly sulfates, and these, in turn, are surrounded by a sandy zone containing carbonate salts.

The deposits are a few inches to a few feet higher than the areas that are subject to flooding at present, but although the surfaces of the deposits are elevated and protected against flooding, the edges of the deposits are being eroded where subject to washing by present-day floods. The zoning of the salts in the crust reflects the differences in solubility of the salts. These deposits, described fully in Hunt and others (1965), are summarized here.

OLDER SALINES

MASSIVE ROCK SALT

The massive rock salt is at least 3 feet thick and overlies silt and clay. It covers about 8 square miles in the lowest part of the saltpan (−280 ft), along the east side between the salt pools and Badwater (fig. 2). The deposit probably averages 95 percent or more of sodium chloride; the remaining 5 percent is mostly chlorides of calcium, magnesium, and potassium and sulfates of magnesium and sodium.

The deposit has an exceedingly rough surface of jagged pinnacles 6–10 inches wide and 1–2 feet high. The depressions between the pinnacles are 1–2 feet wide and are marked by cracks that divide the salt into polygonal slabs 4–6 feet wide. The composition and purity of the deposit, combined with the fact that it is located in the lowest part of the saltpan, suggest that it is the residue from the evaporation of a lake. This lake would have to be the youngest that has flooded Death Valley, the Recent lake.

ROUGH SILTY ROCK SALT

Peripheral to the massive rock salt and grading into it is a belt of equally rough rock salt that is silty. This rough silty rock salt is 1–3 feet thick. It extends onto ground that is 5–10 feet higher than the massive rock salt, about −275 to −270 feet. This deposit, covering about 25 square miles, contains 20–40 percent silt admixed with the salt. The deposit has a rough surface very much like that of the massive rock salt, and is also divided into polygonal slabs by cracks 4–6 feet apart.

Similar deposits are forming at present where perennial ground water is shallow enough for the capillary fringe to reach the surface. The resulting evaporation of water in the wet muds causes salts to precipitate in the mud, heaving it upward and producing a deposit that is mixed salt and mud. The distribution of the rough silty rock salt, peripheral to the massive rock salt, suggests that it formed in shallow parts of the lake where seasonal fluctuations of level would produce mud flats with ground water virtually at the surface.

SMOOTH SILTY ROCK SALT

Peripheral to the rough silty rock salt and gradational to it is a form of salt crust referred to as smooth silty rock salt. The smooth silty rock salt forms extensive smooth plains at the mouths of the principal streams discharging into Death Valley—the Amargosa River, Salt Creek, and Furnance Creek. The three areas where the deposit occurs are about 265 feet below sea level and aggregate about 50 square miles. The deposit consists of a surface layer of brown silt, 1–6 inches thick, resting on a layer of silty rock salt about 1 foot thick. This rock salt rests on clastic sediments. Panward the layer of silt thins, whereas the layer of rock salt thickens.

The salt is cracked into polygonal slabs 3–6 feet in diameter; the overlying silt is similarly cracked but also is divided by closely spaced desiccation cracks that end downward at the salt. The junctions of the polygonal cracks in the salt commonly are reflected in solution pits or depressions in the surface of the silt.

The smooth rock salt layer is interpreted as having been formed by evaporation of ground water, like the rough silty salt. This surface was smoothed, and the silt on it probably was deposited by floods from the main streams discharging onto the salt.

MASSIVE GYPSUM

Surrounding the chloride zone and slightly higher than the rough silty rock salt (at an average of about −265 ft on the west side of the saltpan and −270 ft on the east side) is a discontinuous belt of massive gypsum in deposits 1–5 feet thick. The gypsum overlies damp or wet silt and is capped by a layer of anhydrite or bassanite 1–6 inches thick.

All the present-day gypsum deposits are located near marshes, and presumably the massive gypsum was deposited in marshes at a time when the discharge of the springs was greater than it is today. At the marshes, the total of the dissolved solids is less in wet years than in dry years, chiefly because the amount and proportion of sodium chloride is less. In time of high discharge, gypsum continues to be precipitated, but the more soluble sodium chloride is flushed from this part of the system and transported in solution to the chloride zone.

The massive deposits of gypsum are interpreted as having formed during the period of the Recent lake when discharge from the marshes and springs would have been greater than now and great enough to keep the sodium chloride in solution and flushed out of the system. Under this interpretation the gypsum must have formed after the level of the lake had fallen below the level of those deposits, or they would have become impregnated with sodium chloride introduced by the lake water. The difference in level of the deposits on the two sides of the saltpan may be due to eastward tilting of the saltpan during the last 2,000 years (p. A100).

SALINE DEPOSITS FORMING AT PRESENT

FLOOD-PLAIN DEPOSITS

Salts and saliferous muds are being deposited at present on those parts of the saltpan that are subject to seasonal flooding, altogether about a third of the saltpan. A crust of salt is forming on the lowest parts of the flood plain where surface water collects and can escape only by evaporation. One such area is in Badwater Basin about midway between Badwater and Tule Spring. Another area is in Middle Basin, the low part of which is 1.5 feet lower than the channel that discharges from there to Badwater Basin.

The parts of the flood plain that are tributary to these low places are frequently washed by surface water, and they include extensive areas of bare mud flats. Salts that accumulate on the surface after one wetting are removed by later floods.

Much ground water, though, moves laterally from the channels to nearby areas that are flooded infrequently; the evaporation of this ground water leaves deposits of salts in the upper layers of the mud, forming a crust of silty rock salt.

MARSH DEPOSITS

Marsh deposits are forming at present at many places around the edge of the saltpan where ground water is moving laterally into the pan. The marshes are located where the sand facies grades laterally to silt; the movement of ground water is slowed by the silt, and the ground water level is held up in the adjoining sand (Hunt and Robinson, in Hunt and others, 1965).

All the marshes are depositing sulfate salts. At some marshes along the east side of Cottonball Basin, sodium carbonate is being deposited in addition to sodium sulfate. Elsewhere, the deposits are largely or wholly sulfates with some chlorides. In Badwater Basin where the calcium-sodium ratio is high, the sulfate being deposited is calcium sulfate. In Cottonball Basin where the calcium-sodium ratio is low, the sulfates that are being deposited are mostly sodium sulfate and sodium-calcium sulfate.

The deposits consist of califlowerlike lumps of granular and porous sulfate salts having the texture of wet bread crumbs and coated by a firm layer of salts that includes much sodium chloride. In dry seasons the proportion of sodium chloride, on the lumps and in brines, may be high; in wet seasons it is low, evidently because the discharge is sufficient to transport the more soluble chloride away from the marsh to the chloride zone.

NO. 4 GRAVEL

The No. 4 gravel is along the washes on the gravel fans. Where these washes are no more than a foot or two deep and the area between is frequently flooded, the interstream gravels are included with the No. 4. The No. 4 gravels are composed of firm rocks that are without both weathering rinds and desert varnish. The deposits are loose gravel containing few large boulders and without much sand (fig. 62).

The No. 4 gravel is thin, probably nowhere more than about 10 or 15 feet thick, and the volume of this gravel is correspondingly small compared to the older ones.

Not only is the volume of the No. 4 gravel small, most of the gravel has been derived by eroding the older deposits on the fans. Very little of it seems to have been new gravel from the mountains. Evidence for this is twofold. First, the volume of the No. 4 gravel approximately equals the volume of the channels that have been eroded into the older gravels. Second, large boulders that occur locally in the No. 4 gravel are as abundant along tributary washes that rise in those gravels as they are along the main washes that extend into the mountains. The boulders in the tributary washes must have been derived from the older deposits and probably most of those along the main washes were too.

That cloudbursts can produce floods and mudflows in the washes capable of transporting the largest boulders is clear enough. A striking example is along the main wash draining from Starvation Canyon, where a Recent flood was capable of lifting boulders 10 feet in diameter onto the bank which is 50 feet higher than the bottom of the wash. As noted above, there is adequate source for these boulders nearby in the older gravel; they need not have been moved far. They could be due entirely to reworking from the older gravel deposits along the channel.

On fans composed largely of fine-grained materials like those on Artists Drive, flash floods and resulting mudflows are frequent. At one place, about midway between the exit from Artists Drive and the junction with the West Side highway, an old highway pavement i

FIGURE 62.—Contrast between No. 4 gravel in present wash (left) and No. 3 gravel (right). Gravel in the wash is not stained with desert varnish, and it is loose. In middle distance is the ridge at Park Village, composed of the Funeral Formation. Photograph by John R. Stacy.

buried under about 4 feet of mudflow on top of which is the present pavement. It was cheaper to build a new road than to excavate the old!

Such examples of recently formed mudflows are impressive and provide a yardstick for visualizing the vastly larger ones represented by the older deposits, like those on figure 48.

The No. 4 gravel does not extend onto the floor of the valley, except as short narrow stringers of fine pebbles along rills at the foot of the fans. Few of the rills are as wide as 6 feet; most are only a foot or two wide and only a few inches deep. Although an occasional large pebble may be found along a rill a few hundred feet into the saltpan, for all practical purposes the gravels have not been moved onto the pan more than a few tens of feet, and this abrupt lower limit to the extent of the gravel coincides with a break in slope between the foot of the fans and the saltpan. The gravelly lower edges of the fans slope 2–6 percent; the stone-free edge of the saltpan slopes less than 1 percent.

On the high parts of the fans, the gravel along the present washes is lower than the older gravels, but on the lower parts of the fans the No. 4 overlaps the No. 3 and older deposits and forms fans on top of them (fig. 63). Just as the No. 3 gravel overlaps the lower edges of the No. 2, so also the No. 4 gravel overlaps the lower edges of the No. 3. Clear examples of these overlaps can be seen on practically every fan, and the position of overlap has shifted toward the foot of the fans (fig. 64). This shift could be attributed to downcutting on the high

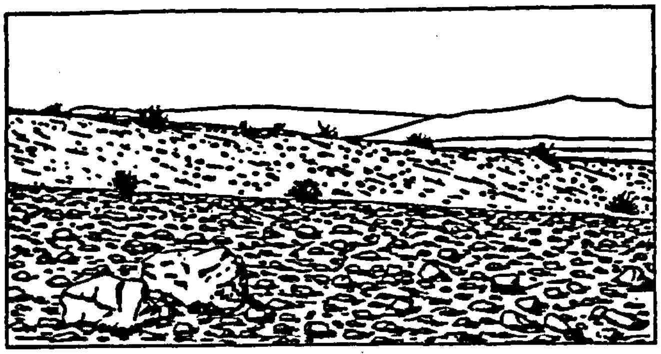

FIGURE 63.—Desert pavement with angular blocks and slabs on No. 2 gravel (foreground) is overlapped by a fan of No. 4 gravel (with creosote bush) which is being built higher than the old gravel. View north from road to Hanaupah Canyon about 1 mile west of Shortys Well.

parts of the fans and the building up of base level at the foot of the fans, but the process has been complicated by eastward tilting that steepened the fans while they were being built.

ALLUVIUM ALONG AMARGOSA RIVER AND SALT CREEK

Flood-plain deposits of alluvium occur along the Amargosa River and Salt Creek. The deposits are mostly silty sand or sandy silt with a little gravel. About 10 feet of alluvium is the maximum thickness exposed, but the maximum thickness of the deposits may be very much greater than this figure.

These alluvial deposits are overlain by sand dunes that date back to prepottery times—that is, the Death Valley III occupation (p. A87). Such dunes are widespread on the alluvium in Mesquite Flat and along Salt Creek in the vicinity of McLean Springs. Others are located on the alluvium along the Amargosa River 5 miles southeast of Coyote Hole and in the Amargosa Desert 35 miles east of Death Valley. Very likely, therefore, the alluvium was laid down about the time of the Recent lake in Death Valley.

In the 2,000 years since that time, the surface of the alluvium has been modified only slightly. The main streams have become trenched as much as 10 feet into the fill. Winds locally have excavated deflation hollows on the surface and have built, and still are building, dunes on the alluvium. Washing from the side hills locally has deposited a foot or two of younger alluvium on the old.

DUNE SAND

Dune sand is of very limited extent around the Death Valley saltpan for the reason that most of the sand there is firmly cemented with salt. Dunes are moderately extensive on Furnace Creek fan and along the west side of Badwater Basin opposite the mouth of Hanaupah and Starvation Canyons. At both localities the sand facies is somewhat wider than elsewhere, because of the source rocks; also, substantial quantities of fresh water are being discharged there to the saltpan, so that the ground contains less soil salts than does the sand around the rest of the saltpan. Too, these are the places where the fan gravels contain most calcium carbonate cement, indicating that ground water discharge has been greatest at these places in past as well as present times A78. The same is true but on a much smaller scale at isolated groups of small dunes at the mouths of Cow and Salt Creeks and on the west side of Badwater Basin opposite the mouths of Johnson and Galena Canyons.

Honey mesquite grows on the dunes and helps hold them in place. The occurrence of the honey mesquite is further indication of good quality water because this phreatophyte, in Death Valley, does not grow where the salinity of the ground water exceeds about 0.5 percent (see Hunt, 1965). Where mesquite plants have died, the dunes become destroyed by wind carrying the sand away from the locality.

Dunes on Furnace Creek fan average about 75 feet in diameter and 6 feet high. They overlie alluvial sandy silt and intertongue with the top 18 inches of that silt. Dunes along the west side of Badwater Basin average twice as wide and twice as high as those on Furnace Creek fan; some are as much as 20 feet high.

The dunes are not migratory. Rather, they are heaps of sand close to the parent formation. This is indicated partly by their distribution and partly, too, by their mineralogy. The mineral composition of the dune sand and of the underlying sand formation is alike at a given location, but it changes greatly from one part of the saltpan to another. For example, from Salt Well south to the foot of the Wingate Wash fan, about

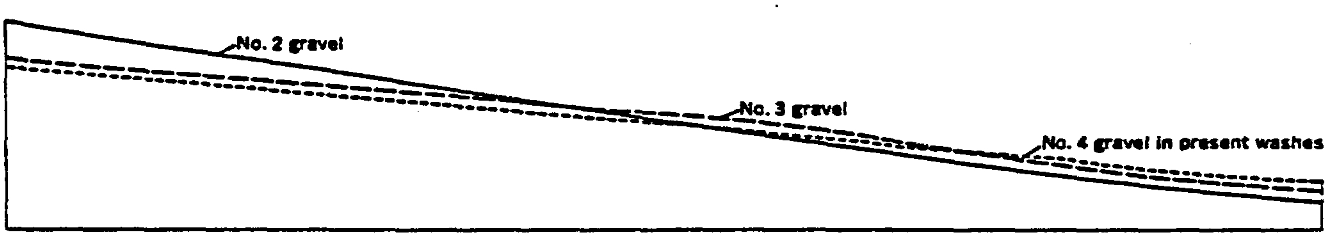

FIGURE 64.—Diagrammatic profiles of fans along the west side of Death Valley, illustrating the downfan shift in position and overlap of younger gravels on older gravels. No. 4 gravel overlaps the No. 3 gravel below where the No. 3 gravel overlaps the No. 2 gravel.

50 percent of the sand is volcanic rock, and a very little is carbonate rock. Northward from Salt Well the percentage of volcanic material decreases to 10 percent or less. At Gravel Well the sand contains numerous flakes of shale from the Johnnie Formation. On Furnace Creek fan, about 50 percent of the dune sand is volcanic rock and about 25 percent is carbonate rock. In all the dunes the grain size is about the same, commonly about 0.25 mm in diameter; about 65 percent is coarser than 0.15 mm and 35 percent is finer.

Sand dunes on Mesquite Flat are much more extensive, much larger and higher, and average finer in grain size than those along the edges of the saltpan. They rest on alluvium. Although the Mesquite Flat dunes are not stationary, they probably have not moved far, because the dunes are largest and most extensive along the sandy belt between the foot of the gravel fans and the silty flood plain that constitutes most of Mesquite Flat.

Other dunes occur along the Amargosa River and overlie the alluvium in the flood plain.

The dunes around the saltpan have developed on ground that was flooded by the Recent lake and have formed since that flooding. The dunes at Mesquite Flat and along the Amargosa River overlie alluvium. All these contain archeological remains representing the Death Valley IV (pottery) occupation and the Death Valley III (prepottery but bow-and-arrow) occupation (see below). These remains occur at all levels in the dunes and on the underlying salt-impregnated sand from which the dunes were derived. Accordingly, the dunes must have been forming throughout the last 2,000 years, and the Recent lake was before that time.

In other areas it has been possible to show that extensive dunes formed during the early part of the Recent. (See for example, Hack, 1941, 1942.) This probably also was true in Death Valley, but early Recent dunes that may have formed around the edge of the saltpan would have been destroyed by the rise and fall of the Recent lake. Relicts of the early Recent dunes would be expected in Mesquite Flat, but search there has not revealed satisfactory evidence of them. The flat was not submerged by the Recent lake, but it may have been flooded by stream wash sufficiently to level the supposed early Recent dunes.

ARCHEOLOGY OF THE DEPOSITS

Occupations of four different ages have been recognized around the Death Valley saltpan. The two early ones, called Death Valley I and II, predate the Christian era. They can be equated with the Lake Mojave, Pinto Basin, and early Amargosa cultures found elsewhere in the southern Great Basin and Mojave Desert. These occupations were chiefly hunting cultures and antedate the bow and arrow. Death Valley III is marked by the introduction of the bow and arrow and dates from about A.D. 500. Death Valley IV is marked by the introduction of pottery into Death Valley, probably about the 11th century A.D. Death Valley III and IV were gathering rather than hunting cultures. The archeology is described in a comprehensive report by A. P. Hunt (1960).

PHYSIOGRAPHY OF THE FANS

The gravel fans that are tributary to the Death Valley saltpan can be considered in four groups. The largest, both in area and volume, are those along the east foot of the Panamint Range. These fans are 5–6 miles long; their surfaces rise from below sea level, at the edge of the saltpan, to more than 1,000 feet above sea level. The summit of the Panamint Range is 7,000–11,000 feet in altitude, and the area of the mountains draining to the fans is almost twice as great as the area of the fans. The rocks are mostly of Precambrian and Paleozoic age. Geophysical surveys indicate that these fans and the fill under them attain a miximum thickness of about 6,000 feet. The thickness is greatest at the foot of the fans. As indicated below (p. A108), probably only about a third of this fill is Quaternary gravel; the remainder is Tertiary and most of volcanic origin.

A second group of gravel fans lies in front of the northwest-trending Funeral Mountains, which extend diagonally across the north end of the part of Death Valley that contains the saltpan. These fans are as long and as high as those along the foot of the Panamint Range, but the fan form is not distinct because the fans are interrupted by numerous hills and ridges of older rocks protruding through the gravel. Also these gravel deposits average very much thinner and their volume very much less than those along the foot of the Panamint Range, no doubt because the drainage basins from which these fans were derived are small compared to those in the Panamint Range. The summit of the Funeral Mountains is only 6,000 feet in altitude, and the area of the mountains draining to Death Valley is no greater than the area of the fans. The rocks are not unlike those in the Panamint Range.

A third group of gravel fans comprise those along the foot of the Black Mountains, along the east side of the saltpan south of Badwater. In contrast to the other two groups of fans, these are small, evidently because the floor of Death Valley has been tilted eastward during Pleistocene and Recent time. These fans have been sinking and are mostly buried by overlap of the playa sediments.

The fourth group of fans are those along the foot of the Black Mountains north of Badwater. These fans contain a high percentage of fine-grained sediments, because they were derived in large part from fine-grained Tertiary playa sediments and volcanics. The area of these fans is about equal to the area of the mountains drained to them, but the mountain summit is only 2,000–4,000 feet in altitude. The fans on Artists Drive, like those in front of the Funeral Mountains, are interrupted by hills of older rocks protruding through the fans, and the gravel deposits are equally thin.

On the fans the gravels of different ages form different kinds of ground, each having a distinctive drainage pattern. The differences are best illustrated on the fans along the foot of the Panamint Range (fig. 65).

Differences in the patterns of the fans in different parts of Death Valley reflect differences in their structural history. In terms of Davis' nomenclature (1925) the foot of the Black Mountains south of Badwater is partly fan based, as at Coffin Canyon (fig. 66*A*), and partly fan free, as in the coves just north and south of the Coffin Canyon fan. Clearly this reflects the eastward tilting of the floor of Death Valley and the overlap of the fine-grained sediments onto the fans and even onto the bedrock front of the mountain.

Differences in fan patterns northward along the foot of the Panamint Range very possibly reflect northward tilt of the Panamint Range. South of Johnson Canyon, at Six Spring and Galena Canyons, the foot of the range is fan bayed (fig. 66*B*). There the spurs are alined, and the fan gravels extend a short distance into the canyons. From Starvation Canyon northward to Trail Canyon the spurs are irregular, and the front of the range is fan frayed (fig. 66*C*). At Blackwater and Tucki Washes, hills of bedrock are surrounded by fan gravel, and the front of the range is fan wrapped (fig. 66*D*).

There is a difference, too, in the degree of dissection of the old gravels northward along the fans (pl. 2). In general, the extent and depth of dissection of the old gravel increases southward as if there has been northward tilting of this stretch of Death Valley since the No. 2 gravel was deposited.

On many of the fans the main washes, which are transporting coarse material from the mountains, have steeper gradients than tributaries. The main washes with their coarse debris rise in areas of relatively high rainfall; they are discharging into an area of low rainfall where the ground is permeable and where water is lost by seepage and evaporation. The courses of these streams are being aggraded. Their tributaries though, rising at the foot of the mountains, are cutting down, especially those eroding in fine-grained rocks; such tributaries may have flatter gradients than the main streams.

A very striking small-scale example of such difference in gradient can be seen on the east side of the highway at the south edge of Cow Creek. A small tributary draining the fine-grained rocks of the Furnace Creek

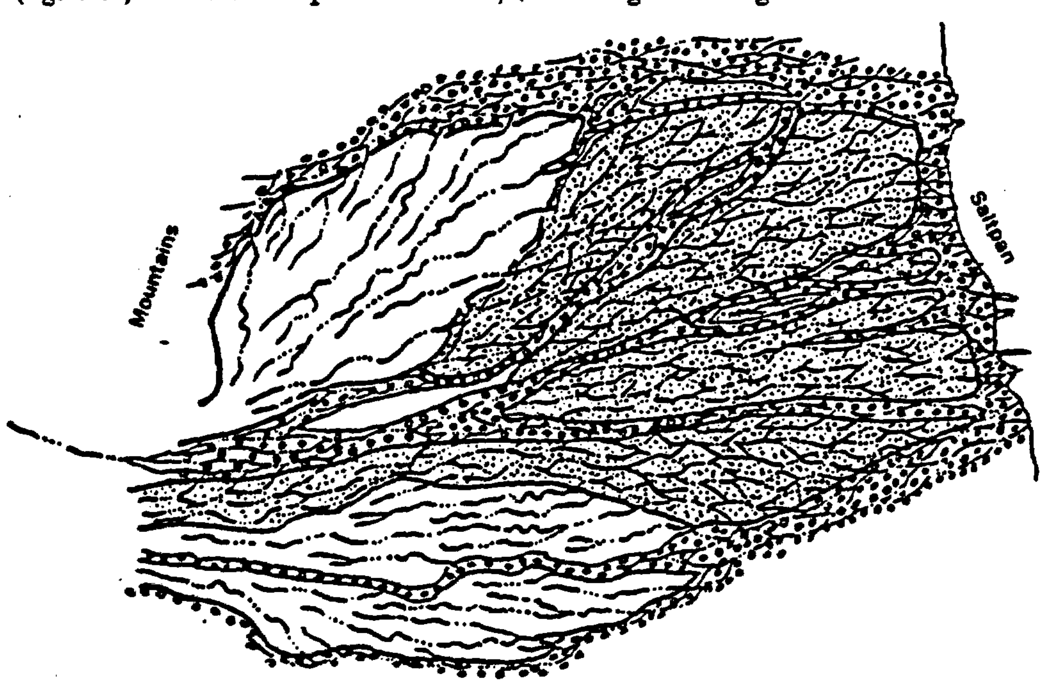

FIGURE 65.—Map illustrating differences in drainage pattern on the older and younger gravels. On the No. 2 gravel (white areas) the drainage is parallel tending towards dendritic. On the younger gravels, No. 3 (stippled areas) and No. 4 (circle pattern), the drainage is braided.

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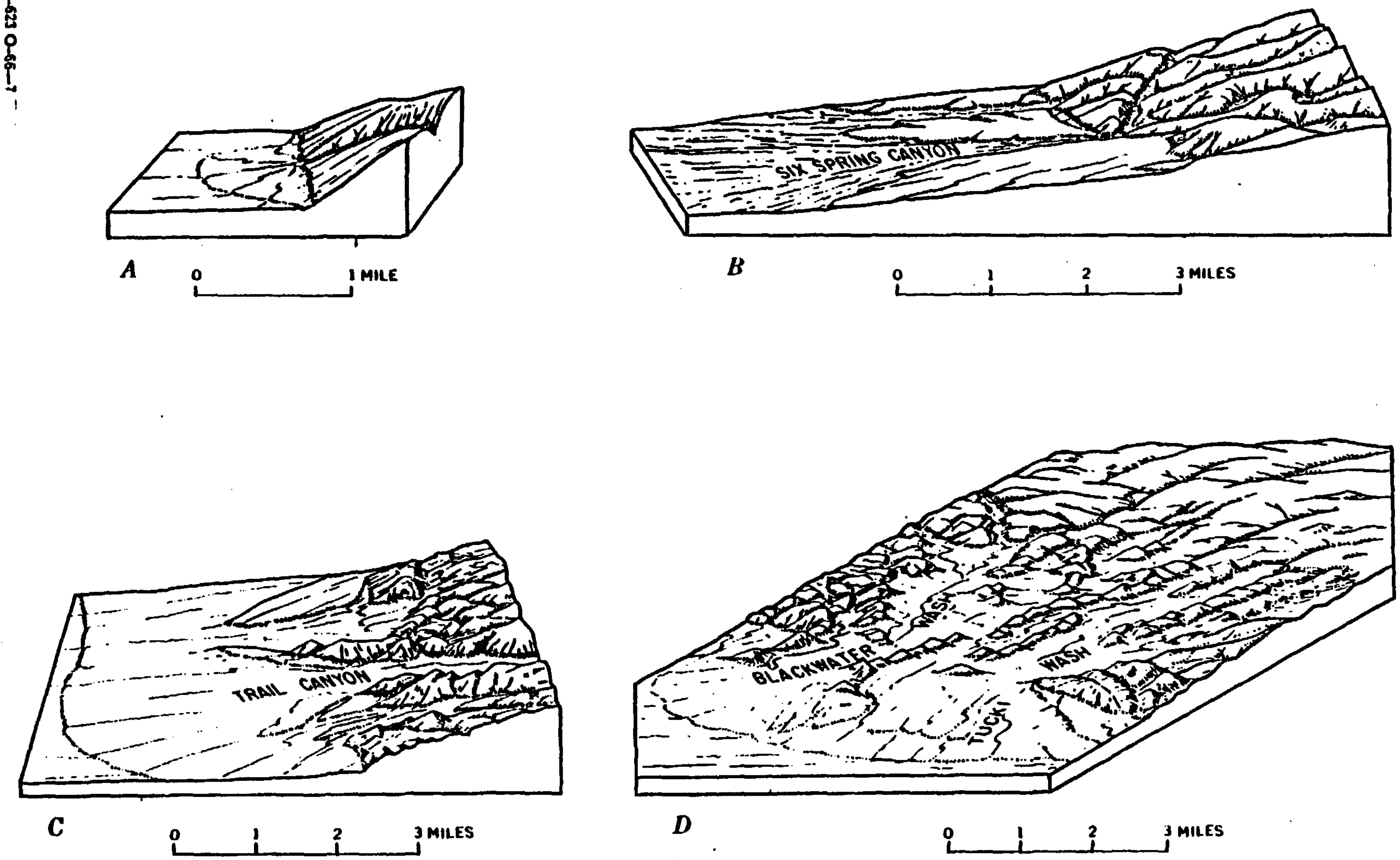

FIGURE 66.—Fan patterns at the foot of the Black Mountains and the Panamint Range. *A*, Fan-based front of Black Mountains at Coffin Canyon. View north. *B*, Fan-bayed east foot of the Panamint Range at Six Spring Canyon. View south. *C*, Fan-frayed east foot of the Panamint Range at Trail Canyon. View south. *D*, Fan-wrapped east foot of the Panamint Range at Blackwater and Tucki Washes. View southwest.

Formation has a much flatter gradient than Cow Creek, which it joins at the highway.

These relationships are similar to those better developed along the Book Cliffs (Rich, 1935) and around the Henry Mountains, Utah (Hunt and others, 1953, p. 204). A companion study of fan gravels in the Death Valley region, by C. S. Denny (written commun., 1965), has shown that gradients along the washes are proportional to the average grain size of the material being moved. Washes transporting coarse material have steeper gradients than those transporting fine material. Lawson (1915) noted that the angle of the rock slope in the desert commonly is determined by the maximum size of rock fragments shed from its surface.

Another feature of the small washes on the surfaces of the No. 2 gravel concerns differences in distribution of coarse and fine materials. Some washes are floored with coarse blocks, evidently a lag concentrate, and the banks are covered with finer gravel or flakes. A neighboring wash may have this distribution reversed. The coarse blocks may line the banks as a natural levee, and the bottom of the wash may be mostly finer gravel or flakes. Such differences may be related to the regimen of the last storm.

DESERT VARNISH

Desert varnish, a stain of iron and manganese oxides coating rock surfaces, is a conspicuous feature of the gravel fans in Death Valley. As seen from a distance, the different ages of gravel can be distinguished by the different degrees of stain (p. A76, 84).

Individual stones on the No. 2 gravel that are firm rock are stained as darkly as those on the surface of the No. 3 gravel; but a high proportion of stones on the No. 2 are crumbly, and these have light-colored surfaces. The varnish has flaked off. As seen from a distance or on an aerial photograph, therefore, these surfaces may appear less dark than do those on the younger No. 3 gravel.

As the name implies, desert varnish is best developed, or at least most conspicuous, in desert regions, but the stain is by no means restricted to such areas; iron and manganese oxides stain rock surfaces in humid regions too. The stain occurs on every type of rock, although it is less common on limestone and dolomite than on the less calcareous rocks. The surfaces stained may be the top or sides of isolated individual stones; they may be vertical or overhanging cliffs, or other surfaces splashed by rivers or wetted by seeps. The stained surfaces may be exposed to direct sunlight or surfaces never reached by the sun, such as joint planes or tunnel walls.

Engel and Sharp (1958), in an important contribution to the chemistry of desert varnish, studied the trace elements occurring with the iron and manganese stain and concluded that (1) varnish on stones in soil or colluvium is derived largely from that material, (2) varnish on large bedrock exposures comes from weathered parts of the rock, and (3) airborne material probably contributes little to the varnish.

Additional chemical studies of 15 samples or varnish removed by an ultrasonic separator from different kinds of rocks in different environments in Death Valley are presented in table 21. These analyses indicate that the proportion of iron to manganese ranges from 1:1 (sample 1B) to 10:1 (samples 3A, 4, 7). This range is comparable to that reported by Engel and Sharp (1958, p. 500).

As seen in thin section, though, the varnish occurs in layers, at least in some places. The bottom layer is brownish or reddish, and the surface layer is bluish black. Some cobbles that are losing their varnish develop a brownish or reddish band where the varnish is thin between the bluish-black surface, still coated with varnish, and the light-colored surface where the varnish has been removed. Hubert Lakin (oral commun., 1960) of the U.S. Geological Survey, experimenting with hydroxylamine hydrochloride solution to remove desert varnish, found that the color of bathed surfaces changed from bluish black to red and finally to that of the parent rock when all the varnish had been removed. These observations suggest not only that the varnish is in layers, but that the underlying layers may have more iron and less manganese than does the surface layer. The range in ratio of iron to manganese indicated by the analyses, therefore, may be due in large part to the range in composition of the layers.

The analyses on table 21 illustrate the similarity in composition of the desert varnish on different kinds of stones in single environments (compare 1A with 1B and 1C; 2A with 2B; 3A with 3B) and the differences in composition of the varnish in different environments (compare group 1 with groups 2 and 3). For example beryllium and molybdenum were found in the varnish on all three specimens from station 1, but they are absent at stations 2 and 3. Cobalt and gallium were found in the varnish on stones at stations 1 and 2, but they are absent at station 3.

Desert varnish is mostly a deposit of considerab[le] antiquity, a relic of past environments. It is being d[e]posited today or has formed in the very recent past on[ly] at exceptional isolated locations where conditions, st[ill] not understood, have been unusual and optimum for i[ts] deposition. Engel and Sharp (1958, p. 515–516) ove[r]emphasize a locality where they infer varnish has be[en] deposited in 25 years. If such deposition of the sta[in] were anything but highly exceptional, buildings a[nd]

TABLE 21.—*Semiquantitative spectrographic analyses of desert varnish*

Symbols:—not looked for: O, looked for, but not found: d, barely detected and concentration uncertain.
Elements looked for, but not found: As, Au, Bi, Cd, Ge, Hf, Hg, In, Ir, Os, P, Pd, Pb, Re, Rh, Ru, Sb, Ta, Te, Tn, Tl, U, W, Dy, Er, Ev, Ho, Lu, Pr, Sm, Tb, Tm.

[Analyst, John C. Hamilton]

	1A	1B	1C	2A	2B	3A	3B	4	5	6	7	8	9	10	11
Fe	3	1.5	1.5	1.5	7	0.7	1.5	3	0.3	3	3	3	1.5	3.0	3
Mn	1.5	1.5	.7	.3	3	.07	.7	.3	.3	.7	.3	.7	.7	.7	1.5
Ti	0	.3	.3	.3	.3	.15	.15	.7	.07	.3	.7	.7	.7	.3	.3
Ca	1.5	1.5	1.5	1.5	.7	1.5	1.5	.3	.3	1.5	3	.7	1.5	1.5	1.5
Mg	1.5	1.5	1.5	1.5	1.5	.7	.7	.7	.07	1.5	3	1.5	.7	1.5	1.5
Na	3	1.5	1.5	.7	—	.7	—	.7	.15	.7	1.5	1.5	3	—	1.5
K	3	7	7	3	3	1.5	0	3	0	1.5	1.5	3	3	7	3
Ag	.0003	.0003	.0007	.0003	.0003	.0003	0	.0003	d	.003	d	d	.03	.0007	.0003
B	.03	.03	.007	.015	.03	.015	0	.007	.007	.015	.007	.007	.015	.03	.015
Ba	.3	.15	.15	.07	.3	.03	.07	.03	.03	.15	.15	.07	.15	.15	.15
Be	.00015	.00015	.00015	0	0	0	0	.00015	0	0	0	.00015	0	0	.0003
Co	.007	.007	.007	.0015	.015	0	0	.003	.0015	.007	.003	.003	.003	.003	.015
Cr	.007	.003	.003	.003	.003	.0007	.0015	.007	.0007	.0003	.007	.03	.015	.007	.003
Cu	.03	.03	.03	.015	.15	.007	.015	.015	.007	.015	.007	.007	.03	.07	.03
Ga	.0003	.0007	.0003	.0003	.0003	0	0	.0007	0	.0003	.0003	.0007	.0003	.0007	.0003
Li	.15	.07	.03	.03	.07	0	0	.03	0	.03	0	0	0	0	.07
Mo	.0007	.0015	.0007	d	0	0	0	.0007	0	0	.0007	.0007	.0007	.007	.0007
Nb	.0015	.003	.0015	0	.003	0	0	.003	0	.0015	.0015	0	.0015	.003	.0015
Ni	.007	.007	.007	.003	.3	.0015	.003	.007	.0015	.007	.007	.007	.007	.007	.007
Pb	.015	.03	.015	.007	.07	.003	.015	.007	.007	.015	.003	.003	.15	.03	.03
Sc	.003	.003	.0015	.0015	.003	0	0	.0015	0	.0015	.0015	.003	.0015	.0015	.0015
Sn	0	0	.0015	0	.003	0	0	0	0	0	0	0	.007	.003	.0015
Sr	.07	.03	.015	.03	.03	.007	.007	.007	.007	.015	.07	.03	.03	.03	.03
V	.007	.007	.003	.003	.007	.0015	0	.003	.0015	.003	.007	.015	.003	.007	.007
Y	.007	.007	.007	.003	.015	.0015	0	.003	.003	.003	.003	.003	.007	.007	.007
Zn	0	0	.03	0	.07	0	0	0	0	0	0	0	.07	0	0
Zr	.03	.03	.015	.015	.03	.007	.015	.03	.015	.015	.015	.007	.03	.015	.015
Rare earths:															
Ce	.07	.07	.07	d	.15	0	0	.03	0	.03	0	0	.03	0	.07
La	.03	.03	.015	.007	.07	0	0	.015	.003	.015	.007	.003	.015	0	.015
Nd	.03	.03	.015	0	.07	0	0	.015	0	.015	0	0	.015	0	.015
Yb	.0007	.0015	.0007	.003	.015	.00015	0	.0007	.0003	.0003	.0003	.0003	.0007	.0007	.007

Figures are reported to the nearest number in the series 7, 3, 1.5, 0.7, 0.3, 0.15, etc., in percent. These numbers represent midpoints of group data on a geometric scale. Comparisons of this type of semiquantitative results with data obtained by quantitative methods, either chemical or spectrographic, show that the assigned group includes the quantitative value about 60 percent of the time.

Description of samples

1A. Dense gray chert from desert pavement on Death Valley I site by Furnace Creek at mouth of Corkscrew Canyon.
1B. Similar to A but more granular; same locality.
1C. Dense brown chert; same locality.
2A. Bull quartz or chalcedony, stained red; from desert pavement on benches between Travertine and Texas Springs.
2B. Quartzite stained black; from same locality as 2A.
3A. Small light-colored chert flake from Death Valley I site on desert pavement south of Nevares Spring.
3B. Large light-colored chert flake; same locality as 3A.
4. Cambrian quartzite in cliff wall above Schaub.
5. Pebble from face of Tertiary conglomerate at fault in NW¼ NE¼ sec. 7, T. 27 N., R. 2 E.
6. Quarry area at megabreccia of quartzite 2 miles southeast of Echo Canyon.
7. Basaltic cobble, with weathering rind, from Funeral Formation under Manly terrace at north end of Artists Drive fault blocks.
8. Granitic rock from debris avalanche, 3 miles south of Badwater.
9. Quartzite flakes from burial mound area at Tule Spring.
10. Upper Pleistocene gravel, Death Valley Canyon.
11. Choppers of volcanic rocks from quarry area at base of Dinosaur.

other surfaces, artificial and natural, that are as old as 25 years should generally be darkly stained, but such is not true.

There is, on the other hand, abundant and good evidence in Death Valley, in the Southwest generally, and in other parts of the world, indicating that very little desert varnish has been deposited during the last 2,000 years. In Death Valley, stone artifacts of the Death Valley III and IV occupations are not stained with desert varnish. Stone artifacts of the late Death Valley II occupation commonly are lightly stained. Stone artifacts of the early Death Valley II and Death Valley I occupations commonly are darkly stained. These generalizations are based on studies of hundreds of archeological sites.

The relationships observed in Death Valley are true generally through the Southwest. Pottery was introduced earlier elsewhere in the Southwest than in Death Valley—roughly 2,000 years ago. Stone artifacts and masonry dwellings of the pottery-making peoples rarely are stained with desert varnish, whereas stonework of older occupations commonly are stained. This evidence is regional and based on studies of thousands of sites.

Moreover, the conspicuous deposits of desert varnish today are being eroded, as on the surfaces of the No. 2 gravel in Death Valley. Whatever the rate of deposition of desert varnish, erosion is faster. The same is true on the Colorado Plateau where the varnish is preserved on the parts of cliff faces that are protected against erosion but is removed from exposed parts, such as the rounded edges of joint blocks and the upward-facing parts of cliffs or buttes. Recent rockfalls in the canyons there leave bright scars on surfaces that otherwise are darkened with varnish. The protected undersides of isolated boulders are still coated with varnish, but it has been removed from the weathered tops and rounded edges.

The same appears to be true in arid regions in other parts of the world. Blackwelder (1948) cites evidence from Egypt indicating practically no deposition of desert varnish in 2,000 years, slight deposition in 5,000 years, and dark strain on older stonework.

The origin of desert varnish is still uncertain; but a reasonable hypothesis can be offered that is based on conditions at places where similar varnish is being deposited today.

Oxides of iron and manganese are being deposited at present at many springs, seeps, and other damp places in the arid Southwest. Water obviously is needed to transport the metals to the surface where they are deposited, but the restricted occurrence of newly deposited iron and manganese oxide at wet places suggests that the moisture requirements may be considerable. Little or no varnish is forming on surfaces that are infrequently wet.

The deposition occurring at wet places probably is not due solely to physical-chemical processes. The quality of the waters varies greatly from place to place, yet the oxides of iron and manganese invariably are selectively precipitated with only minor contaminants. This selective precipitation of the metals without mixing with large amounts of other salts could readily be brought about by oxidation caused by bacteria, algae, or other micro-organisms, but it seems difficult to achieve in such varied environments by physical-chemical processes alone.

The importance of micro-organisms for hastening the oxidation of pyrite in mine waters has been studied extensively and seems well established (see bibliography in Temple and Koehler, 1954). Investigators of mine waters have concluded that oxidation by physical-chemical processes is too slow to account for the quantity of ferric iron transported and deposited by the water, and they attribute the oxidation and precipitation to specific micro-organisms.

Micro-organisms in water in a diversion tunnel of the Tennessee Valley Authority also have caused deposition of manganese oxides (Pollard and Smith, 1951).

I have attempted to apply some of what has been learned about mine waters to the desert-varnish problem. Samples of water were collected at seeps and springs in the Colorado Plateau and immediately added to various kinds of solutions. One solution consisted of a medium for growing bacteria, nutrient broth, to which had been added a trace of 10 percent solution of nonsterile ferrous sulfate. Seep waters added to this solution invariably produced abundant growths of new colonies of micro-organisms, and the iron became oxidized in a matter of hours. A control solution of the same but without the seep water did not change color for days; a second solution of seep water without the broth remained clear. This experiment was duplicated at many seeps.

In another experiment sterile mineral solutions without nutrient broth were prepared with the following composition (see Leathen, McIntyre, and Braley, 1951): In 1,000 ml of distilled water:

	Gram
$(NH_4)_2SO_4$	0.15
KCl	.05
$MgSO_4 \cdot 7H_2O$	.05
K_2HPO_4	.05
$Ca(NO_3)_2$	.01

The solution was autoclaved and to it added 10 ml of a 10 percent solution of nonsterile $FeSO_4 \cdot 7H_2O$. The pH was then adjusted to 3.5 by adding HCl or NaOH as necessary.

In this experiment, control solutions of the seep water remained clear. Control solutions of the basic medium remained clear for days. In a few cases, solutions of the basic medium to which seep water was added became discolored in about a day. No new colonies of micro-organisms were observed.

The experiments leave much to be desired, but despite obvious shortcomings the tests do indicate that micro-organisms of one kind or another abound at the seeps and springs where iron and manganese oxides are being deposited, and that those organisms hasten the oxidation of the metals. It still remains to be learned whether the micro-organisms are an essential part of the process or whether physical-chemical processes alone could deposit iron and manganese oxides containing only traces of contaminants at seeps and springs.

A study of some thin sections of desert varnish by Estella Leopold and Richard A. Scott (written commun., 1961), of the U.S. Geological Survey, did not reveal evidence favoring deposition attributable to micro-organisms. They suggest that the varnish layer may have been deposited as a silica gel around foreign particles serving as nuclei. The suggestion is favored by the fact that no extensive increment growth lines show in radial sections, but rings like liesegang rings show in tangential sections. Moreover, they found little or no penetration of the varnish into the rock in a way that would suggest surface weathering of the rock attributable to micro-organisms. A single hypha strand bearing two branches and looking very much like fungal hyphae was found in one radial section embedded in the varnish. What appear to be cells in the strand are 7–10 microns wide.

The age of the desert varnish and the evident need for moisture lead to the conclusion that the varnish in Death Valley and elsewhere in the Southwest is largely the product of a pluvial period. The last pluvial period was the time of the Recent lake in Death Valley. At that time there may have been sufficient moisture in the form of dew to maintain microfloras that could grow and hasten the oxidation and precipitation of iron and manganese oxide whenever occasional rains soaked the soil.

EROSION AND SEDIMENTATION

The huge gravel fans sloping from the mountains to the saltpan and the fill 1½ miles deep under the saltpan bear mute testimony to the vast amount of erosion in the mountains bordering Death Valley. Cross sections illustrating the depth of fill in the valley as estimated from gravity and magnetic surveys, and the depth and gradients of the canyons in the mountains, are shown on figure 67. The proportion of fill that is Tertiary and the proportion that is Quaternary must be inferred (p. A108), but it seems likely that half or more of the fill in Death Valley is Tertiary. The Quaternary is more than 1,000 feet thick at the localities that were drilled in Cottonball and Badwater Basins (table 19) where the total fill, as estimated from geophysical surveys, is about 4,000 feet. In the sections on figure 67, it is assumed that the thickness of the Quaternary fill averages about half that of the Tertiary.

Based on this assumption, the volume of the Quaternary fill in Death Valley aggregates about 80 cubic miles. The correct figure can hardly be less than half nor more than twice this estimate.

This estimated volume of the Quaternary fill is about equal to the estimated volume (90 cubic miles) of the canyons draining directly to the saltpan from the Panamint Range and Black and Funeral Mountains. Although neither estimate can be firm, the two quantities seem to be of the same order of magnitude.

Moreover, Tertiary eruptives overlap the east flank of the Panamint Range and rise high onto it along the divides between the canyons. It seems likely therefore that this surface of overlap represents the limit of depth of dissection on this side of the Panamint Range at about the beginning of Quaternary time, and that the canyons were cut below that surface during the Quaternary.

In attempting to equate the volume of fill in Death Valley with the volumes of the canyons, however, there is need to consider the sediments brought into the valley from other parts of the hydrologic basin, the volume of volcanics, and the possible losses of early Pleistocene sediments to the extent that Death Valley may have had exterior drainage.

The Quaternary fill in Death Valley includes sediment brought there by the Amargosa River, by Salt Creek, and by Furnace Creek Wash from its head in Greenwater Valley. Except at the mouth of Furnace Creek, these sources did not contribute to the gravel fans, and it is doubtful that they contributed as much as half of the silt facies of the fill (fig. 67). Even this proportion would amount to only a quarter of the total fill, and probably the correct figure is nearer a tenth—a quantity that is not significant considering other uncertainties in the estimates.

The Tertiary deposits are composed very largely of volcanic debris and contain no more than a tenth, and perhaps very much less, of sediments derived from the Precambrian and Paleozoic rocks. By contrast, more than 90 percent of the Quaternary fill is derived by erosion of the older rocks, including Tertiary as well as Precambrian and Paleozoic ones, and much less than a tenth is new volcanic debris.

Nevertheless, additions brought from other parts of the hydrologic basin and other additions due to vulcanism may aggregate a fifth, or perhaps a fourth, of the whole volume. This volume would be offset, and perhaps more than offset, by losses to the degree that Death Valley had exterior drainage during the Quaternary.

There is reason to believe that most of the erosion occurred at times when more water was coming into the valley than now, because under the present climatic regimen the processes have greatly slowed. The present rate of erosion can be estimated by noting changes along datable features, such as—

A. Extent of destruction of—
   1. Roads and flood-control ditches and other features that were constructed about 20–25 years ago.
   2. Old trails that were abandoned about 50 years ago.
   3. Archeological features.
   4. The Recent fault scarp, believed to be about 2,000 years old, along the foot of the Black Mountains.

B. Extent of weathering, erosion, and sedimentation on Quaternary deposits, such as the various formations on the saltpan and the younger gravel formations.

DAMAGE TO ROADS AND FLOOD-CONTROL DITCHES AND OTHER FEATURES IN 25 YEARS

The present highway from Mormon Point to Badwater, Furnace Creek Ranch, and Beatty Junction—45 miles—was paved between 20 and 25 years ago. Although no formal record has been kept of mudflows that have buried the highway, they are known to have occurred at half a dozen places. A flood in 1954 deposited 4 feet of mud on the highway half a mile south of the exit from Artists Drive; the present highway is on top of that fill. Other destructive mudflows have covered the highway at the crossing of the wash from Indian Pass, Furnace Creek Wash, at Desolation Canyon, Artists Drive between the exit and the junction with the road to the west side of the valley, and at the

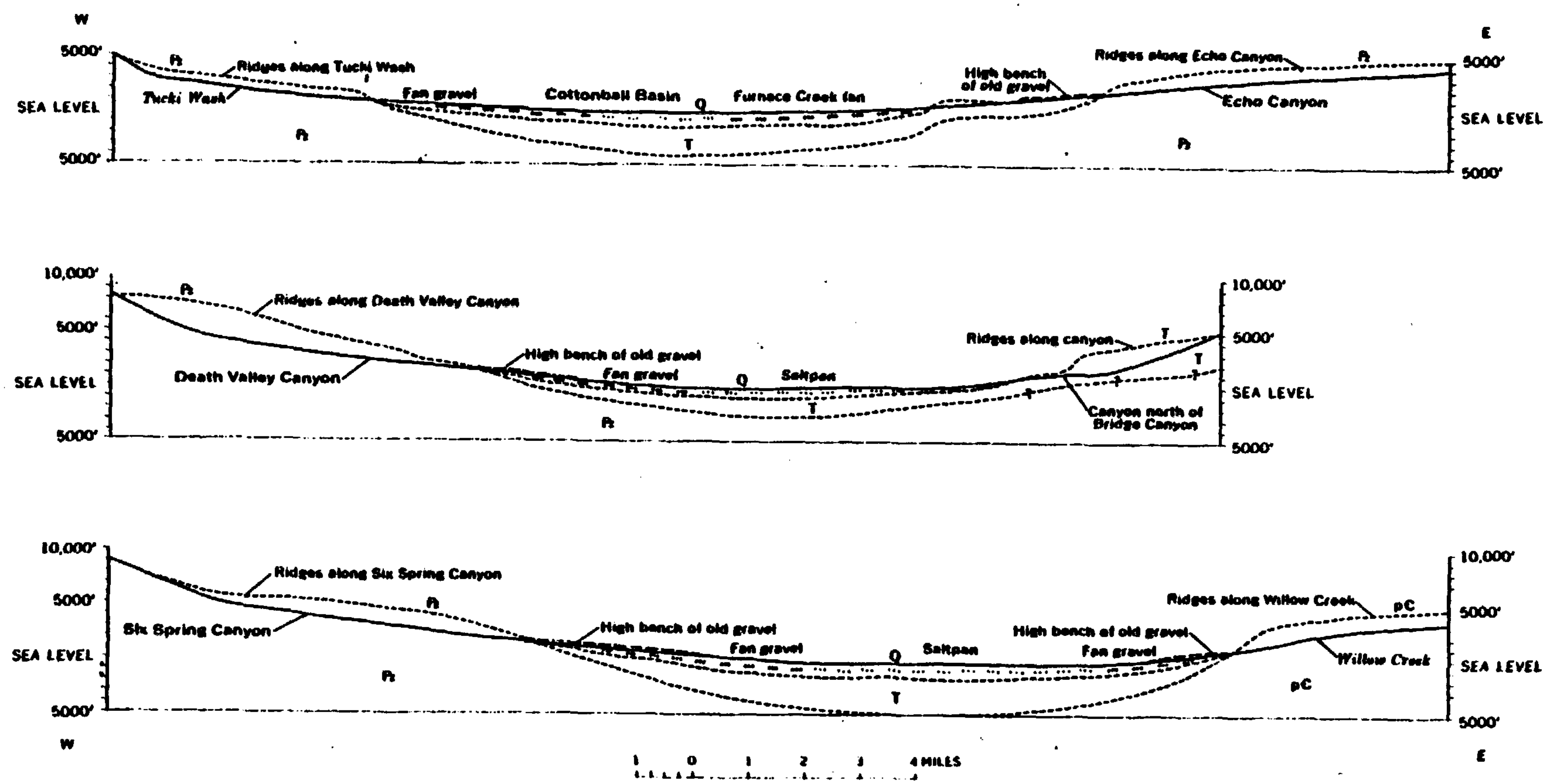

FIGURE 67.—Sections across Death Valley illustrating depth of fill under the saltpan, slope and thickness of fan gravels, and the depth and gradient of canyons in the adjoining mountains. Q, Quaternary fill (silt under the saltpan and gravel between there and the mountains); T, Tertiary fill (in large part volcanics); Pz, Paleozoic rocks; pC, Precambrian crystalline rocks.

junction of the road to the Salt Pools. No doubt there have been others, but only 1 percent or so of this 45-mile stretch of highway has been washed by destructive floods in the last 20 years.

The original highway along the east side of Death Valley was along the foot of the gravel fans; it was last bladed when the present highway was paved about 20 years ago. A survey was made of damage along the stretches of this old bladed road where it has not been disturbed by later construction; the shoulder on the uphill side of the road averages about 6 inches high and perhaps a foot wide. Floods across this road in the last 20 years have destroyed about 20 percent of this little feature, as indicated in table 22.

TABLE 22.—*Erosion, in 20 years, of road shoulder 6 inches high along old bladed road from Beatty Junction south along the foot of the gravel fans*

	Location of traverses along old road	Length of road	Road shoulder destroyed	
			Length	Percent
1	South from BM−233, 1 mile west of Beatty Junction:			
	BM −233 to BM −241 .... miles	1	....	25
	BM −241 to BM −245 .... do	1	....	25
	BM −245 to BM −250 .... do	1	....	25
	BM −250 to BM −249 .... do	1	....	12
	BM −249 to BM −255 .... do	1	....	17
	BM −255 to BM −254 .... do	1	....	10
	BM −254 to BM −255 .... do	1	....	5
	Total or average	7	....	17
2	½ mile north of Golden Canyon, south to Desolation Canyon .... feet	12,600	3,400	27
3	Near Artists Drive exit, south nearly to the highway across the valley .... do	13,600	2,000	15
4	Near highway across Death Valley, south to Badwater .... miles	10	3	30
5	Around foot of second fan south of Badwater .... feet	7,300	2,200	30

Fifteen miles of the traverses indicated in table 22 are on the fans between Badwater and Golden Canyon. These fans are composed of fine-grained sediments that are impermeable and favor runoff—far more so than the permeable gravel fans. In the last 25 years, floods on this impermeable ground have destroyed about 20 percent of a ridge of loose earth 6 inches high and 1 foot wide. The runoff, of course, has been channeled in washes, and in some washes there have been repeated floods. Nevertheless, 80 percent of the little ridge still stands.

When the new road was built, 15 flood-control ditches totaling about 5 miles in length were constructed in the northern part of the valley. The ditches were constructed at the most exposed locations. They extend diagonally across washes and are oriented about 45° to the highway. Originally the ditches were about 5 feet wide and had a maximum depth of about a foot. An earth embankment about as wide as the ditch and about a foot high was constructed along the lower side of each ditch. Floods in these washes in about 25 years have

TABLE 23.—*Erosion of flood-control embankments in 25 years*

	Locality	Approximate length of embankment (feet)	Destruction (percent)
1	2.5 miles northwest of Beatty Junction; embankment oriented slightly west of north	2,500	25
2	Same location; embankment oriented slightly north of east	4,000	30
3	0.9 mile southeast of Beatty Junction	2,500	30
4	1.5 miles southeast of Beatty Junction	1,200	10
5	1.8 miles southeast of Beatty Junction; embankment oriented slightly west of north	2,000	25
6	Same location; embankment oriented about east	2,000	10
7	2.3 miles southeast of Beatty Junction	2,200	30
8	3.3 miles southeast of Beatty Junction	2,000	15
9	3.6 miles southeast of Beatty Junction	1,000	20
10	4.6 miles southeast of Beatty Junction; embankment oriented northeast	2,000	40
11	Same location; embankment oriented southeast	1,000	25
12	East side of highway opposite exit from Mustard Canyon	1,200	30
13	1.3 miles north of Furnace Creek Ranch	800	30
14	1 mile north of Furnace Creek Ranch	1,000	30
15	0.3 mile northeast of Furnace Creek Ranch	2,000	30

destroyed about 25 percent of the embankments (table 23).

Even in washes where runoff is concentrated, after 25 years of floods about 75 percent of an earth ridge 1 foot high still stands.

On the flood plain of the saltpan at the Devils Speedway at the foot of Trail Canyon fan, some racetrack runways were scraped about 25 years ago. The area is flooded seasonally. A circular track half a mile in diameter, with a smaller one inside, was scraped north of the road across the valley; another nearly a mile long and more oval, was scraped south of the highway. At one curve an embankment about a foot high was built, but elsewhere the scraped track was bordered by only a low ridge of earth about 6 inches high. After 25 years, enough of the ridges still remain to outline plainly the position of the old tracks; there has not been enough silting or erosion on this part of the flood plain to destroy these minor artificial features.

DAMAGE TO TRAILS ABOUT 50 YEARS OLD

Old trails cross the fans diagonally or follow the contour and lead from one spring to another and to routes into and away from the valley. These trails are the original freeways, perhaps originally made by

Pleistocene animals, then used by the Indians who came into Death Valley, and finally by the pioneers. The trails, though, have been little used in the last half century; they were abandoned when vehicular traffic became heavy enough to require roads other than those that could be followed on foot or horseback. The trails are preserved because there has been no livestock in the valley.

That the trails were used by the Indians is indicated by the concentrations of stone artifacts and other archeological signs along them. That the pioneers and early prospectors used the trails is indicated by the common occurrence of pre-1900 relics along the trails. That the trails have been little used during the last 50 years is indicated by the scarcity of litter younger than about 1900 and by the occurrence of narrow coyote trails meandering within the wider, older trails. Results of some surveys along stretches of the old trails are given in table 24.

In general, the old trails are for the most part intact where they cross high benches on the No. 2 gravel, suggesting that in the present regimen erosion of these benches must proceed by retreat of the sides rather than by lowering of their tops.

Where the old trails cross No. 3 gravel, 10–20 percent of the alinement may be destroyed, and the destroyed stretches invariably are in the swales where runoff is concentrated. Part of this runoff originates on the No. 3 gravel, but most of it represents overflow from nearby washes.

Even where the trails cross washes with No. 4 gravel, as much as 25 percent of the alinement may be preserved. Destruction is more complete where main washes are crossed on the upper parts of the fans rather than on the lower. This may be due to the runoff being concentrated in a few channels on the upper parts of the fans, or it may be due to greater total runoff, or perhaps to both.

Reference has been made to the steps (terracettes, fig. 51) and related features resulting from mass wasting on the gravel fans, especially on the No. 2 gravel. (See p. A68; see also Hunt and Washburn, in Hunt and others, 1965.) The trails cross such features without showing signs of creep downhill.

TABLE 24.—*Erosion along old trails in the last half century.*

	Location of traverses along old trails	Segment of trail (feet)	Segment of trail destroyed	
			Feet	Percent
1	1¼ miles across Artists Drive fan, across secs. 15 and 22, T. 26 N., R. 1 E.	7,800	4,800	6
2	West of road 2 miles south of Bennetts Well	1,250	800	6
3	3 miles across northeast corner of Cottonball Basin. Begin at north end at BM —245, trail—			
	Across sand facies of carbonate zone	3,900	975	[1] 2
	To Natl. Park Service Monument, 2934, 8296, on gravel fan	2,000	2,000	10
	On fan southward to old road	2,000	675	3
	Southward from old road across saltpan	3,800	800	2
	To BM —255 along foot of gravel fan	3,300	3,000	9
4	Near Salt Springs, south to near Cow Creek	10,000	2,000	2
5	3.7 miles, from Nevares Spring to Texas Spring	1,300	0	Non
		1,050	1,050	10
		450	0	Non
		300	300	10
		550	0	Non
		4,525	750	1
		2,100	850	4
		900	900	10
		550	50	1
		1,250	625	5
		1,200	200	1
		350	175	5
		2,000	1,800	9
		925	90	1
		750	750	10
6	½ mile, from Furnace Creek Ranch to Nevares Spring. Begin at road into East Coleman Hills	2,600	750	2
7	North from Nevares Spring to Echo Mountain wash. Begin 1¼ miles north of Nevares Spring	7,400	1,400	1
		4,000	0	Nor
		1,400	975	7
	Total or average	67,650	25,715	3

[1] Half by salt heave and about half by erosion.

DAMAGE TO PREHISTORIC ARCHEOLOGICAL FEATURES

More than a dozen rock alinements dating from the Death Valley III occupation are preserved at various locations on the No. 2 gravel around Death Valley (Hunt, A. P., 1960). Some of these alinements were made by laying pebbles or small cobbles in a row; others were made by scraping the gravel into a ridge 2–4 inches high. These slight features still stand despite approximately 1,000 years' exposure to sheetfloods c the gravel. Perhaps there were other alinements th have been destroyed by sheetfloods, but probably r more than a few judging by the preservation of trai on the No. 2 gravel.

In addition to the alinements, there are approximat ly 1,500 rock circles. These are on the No. 3 as well on the No. 2 gravel. They are not well dated; some a historic. Others are prehistoric Death Valley IV, a some probably date from Death Valley III occupatic Most of these rock circles consist of a single layer cobbles arranged in a circle 5–7 feet in diameter. Sor circles located along the rims of washes have been partly destroyed by undercutting of the rim, but circles back from the rim show very little sign of washing.

DAMAGE TO RECENT FAULT SCARP ALONG FOOT OF BLACK MOUNTAINS

The Recent fault that extends from the mouth of Furnace Creek 30 miles south to Mormon Point (p. A 100) forms a discontinuous escarpment 5–10 feet high. The faulting is believed to have occurred about 2,000 years ago. Since that time about 90 percent of the escarpment has been destroyed, partly by dissection of the upthrown block and partly by burial under younger alluvial-fan deposits (fig. 72).

WEATHERING, EROSION, AND SEDIMENTATION ON QUATERNARY DEPOSITS

In the present climatic regimen disintegration of stones is slow enough not to be noticeable in the No. 3 or No. 4 gravels, yet in the No. 2 and older gravel deposits, disintegration is continuing, as indicated by the accumulation of fine debris around the foot of boulders (fig. 49). This difference between the No. 2 and the younger gravels could be attributed to difference in length of time that the gravels have been exposed at the surface, but more likely the difference is due chiefly to some changes in the processes causing disintegration—changes attributable to the climatic changes that occurred during late Pleistocene time.

Erosion and sedimentation under the present climate are slow too. Areas on the saltpan that are subject to flooding are restricted to the flood-plain areas which constitute about 30 percent of the valley floor. The salt crusts on the other 70 percent were deposited about 2,000 years ago, and the only part of this crust that has been flooded is the smooth silty rock salt, constituting about 30 percent of the crust. The smooth salt has been flooded often enough in 2,000 years to deposit 1–6 inches of silt on the salt. Only fine sand and silt is transported onto the flat valley floor; gravel is deposited at the foot of the fans. The only exception to this is provided by a few cobbles of light highly scoriaceous lava washed onto the salt flat at the foot of Furnace Creek fan.

The No. 4 gravel, covering about a third of the gravel fans, represents the part of the fans that is subject to much washing at present. The No. 3 and older gravels, darkly stained with desert varnish, have not been washed sufficiently in 2,000 years to destroy that stain, which is pre-Death Valley III occupation (p. A91). Erosion and sedimentation on the gravel fans in 2,000 years, therefore, have been restricted to about a third the area of the fans.

The volume of the No. 4 gravel is small compared to the older gravel formations and seems to be little, if any, more than the volume of the washes and channels that are eroded into the older gravels. Apparently not much gravel has been brought from the mountains to the gravel fans in 2,000 years; the No. 4 gravel seems to be derived chiefly from erosion of the older gravel formations.

Further evidence that little gravel is being transported from the mountains under the present climate is found where the No. 2 or No. 3 gravels overlap the foot of the mountains. Alluvial fans of No. 4 gravel have been built on the older gravels only at the mouths of large washes. Elsewhere, little new material has been moved from the mountainsides onto the surfaces of the old gravels.

The deposits of the last 2,000 years—the No. 4 gravel and floodplain deposits—if spread over the whole valley, would average considerably less than a foot thick, and much, perhaps most, of this is simply reworked older valley fill. Only a fraction can represent new sediment brought into the valley from the mountains. Exact figures cannot be had, but no matter what figure is assumed within the known limits, clearly the thick Quaternary fill in Death Valley could not have been deposited if the rate of erosion in the mountains had always been as slow as it has been during the last 2,000 years.

The history of erosion and sedimentation in Death Valley evidently is one of changing rates, presumably due to changing processes. Huntington (1907) offered the hypothesis that in arid regions the pluvial periods accelerate weathering and growth of vegetation. During such periods, waste is stored in the mountains. According to him, the transition to an ensuing arid period is the time when most of the waste is transported to the alluvial fans; when the supply of waste becomes exhausted, there is little new material brought from the mountains and the older fill deposits become dissected. The evidence in Death Valley fits well with Huntington's hypothesis.

STRUCTURAL GEOLOGY

By Charles B. Hunt and Don R. Mabey

The structural geology of the Great Basin, of which Death Valley is a part, has been a century-long subject of discussions and differences of opinion. One of the principal conclusions arrived at from our studies of the structural geology in Death Valley is that the discussions and differences of opinion will continue for a long time to come.

The mountains bordering Death Valley are uplifted fault blocks of the general kind first recognized in the Great Basin by G. K. Gilbert (1875), and later, with modifications, by King (1678) and Dutton (1880).

(For a review of early theories about Great Basin structure, see Davis, 1926, and Nolan, 1943.) But the mountains have had a complex history of early deformation and erosion that is not reflected in the shapes of the uplifted fault blocks, a complexity first clearly demonstrated in the Great Basin by Louderback (1904), although alluded to by the earlier investigators. Few today will seriously entertain the theory (Keyes, 1909) that the basins like Death Valley are chiefly erosional in origin, but many will ponder the degree to which the structural framework involves folding and thrust faulting as well as block faulting, an extreme view of which, advanced by Spurr (1901), discounts block faulting and attributes the basins and ranges to an Appalachian Mountain type structure in a desert climate. Too, our lack of knowledge allows for differences of opinion about the significance of the granitic intrusions and vulcanism to the deformation; indeed, this subject has received little attention.

Although Spurr's main thesis needs to be greatly modified, we think he was more right than wrong in his view that deformation has gone on "steadily though spasmodically from the close of the Mesozoic to the present." In the course of that long period of time, usually taken as 60 million years, the way in which the crust has yielded to stress has changed, and the earli structures therefore are obscured by the later ones. Tl record admittedly is not complete enough to demonstra continuity of the deformation.

Probably the most basic uncertainty to understandir the structural framework of Death Valley and the re of the Great Basin is uncertainty as to whether th part of the crust has been shortened by compressic or distended by tension. The evidence is conflicting in Death Valley and elsewhere in the Great Basin.

Whatever the origin and complete sequence of stru tural events have been, the result is a complex of stra effects superimposed on one another. We attempt unravel these by examining the younger first, and l eliminating their effects we attempt to restore seriati earlier structures to the way they first appeared; tl effort is only partly successful, for reasons that wi become apparent as we progress.

STRUCTURAL SETTING OF DEATH VALLEY

Death Valley is about centered in an area that is subsection consisting of the southernmost tenth of tl Great Basin (fig. 68). The Death Valley subsection characterized by block mountains that are of Precar

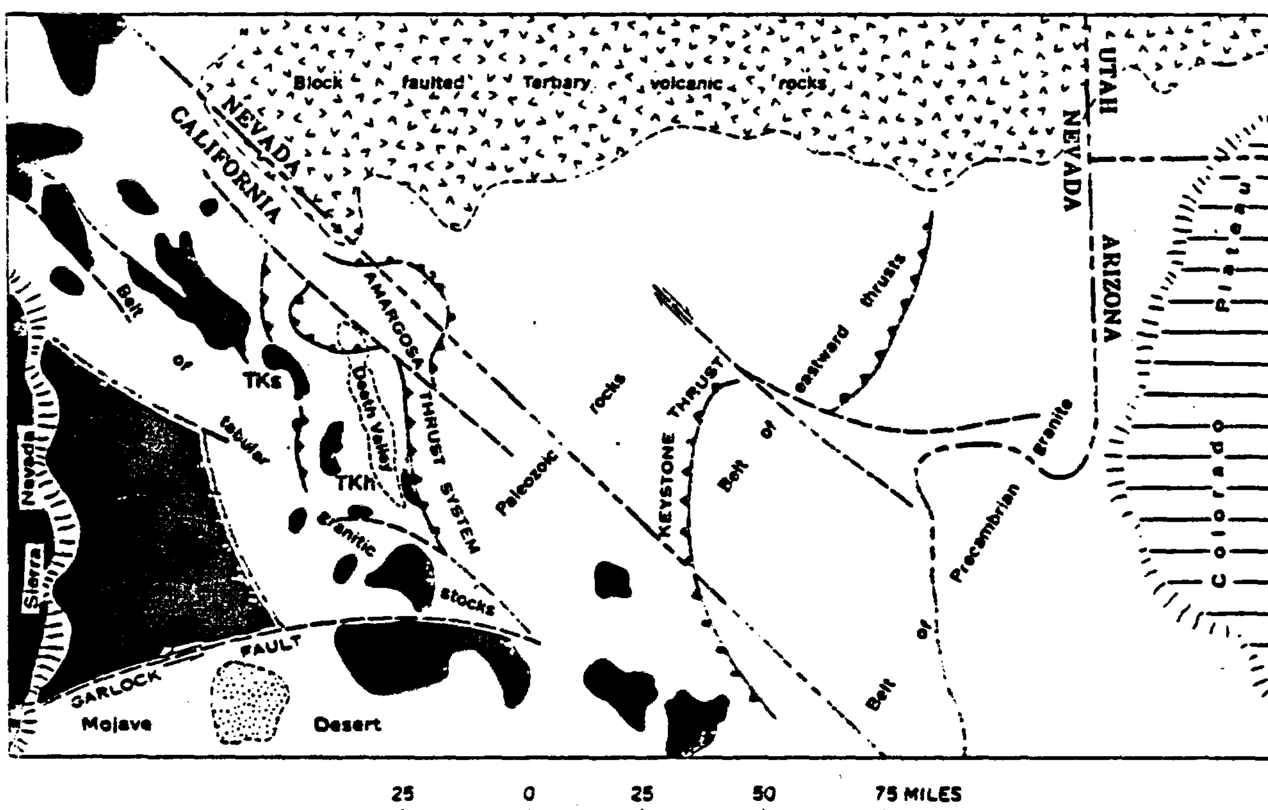

FIGURE 68.—Map showing the Death Valley subsection of the Great Basin. Mesozoic and Tertiary granitic intrusions shown in black. TKs, granite at Skidoo; TKh, granite at Hanaupah Canyon.

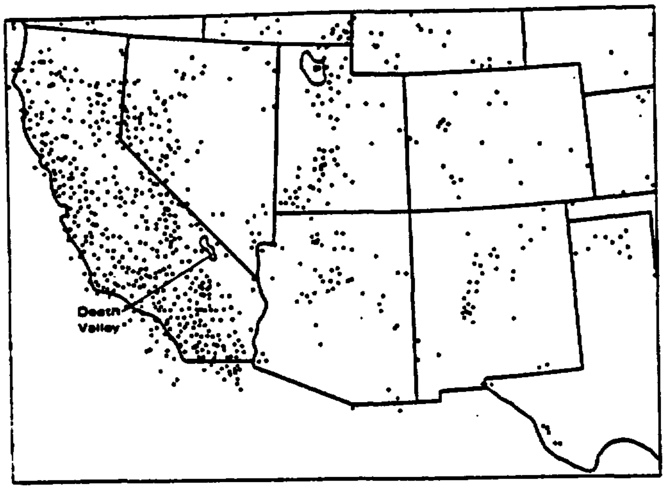

FIGURE 69.—Map showing seismic epicenters in southwestern United States. From Woollard, 1958.

brian and Paleozoic formations intruded by granitic rocks and locally capped by Tertiary volcanic rocks.

To the north, the next subsection of the Great Basin is characterized by block mountains that are almost entirely of Tertiary volcanic rocks. West of the Death Valley subsection is the Sierra Nevada batholith; to the east, along the Colorado River, is a small area that is largely Precambrian granite and assigned to the Mexican Highland; and beyond that is the Colorado Plateau. Southward the Death Valley subsection ends at the Mojave Desert (Hewett, 1955), a subsection of the Sonoran Desert.

Death Valley extends along the eastern edge of a cluster of granitic intrusions that are satellites of the Sierra Nevada batholith (fig. 68); it is at the eastern edge of the seismically active areas that extend westward to the coast (fig. 69). Along this border of the granitic intrusions and seismically active areas is a series of curious overthrust faults, first recognized and described by Noble (1941). These faults are curious because younger rocks have been thrust westward over older ones. Because the horizontal movement is measurable in miles, we follow Noble in referring to the faults as thrust faults, although the displacement on them in fact is that of a normal fault (Longwell, 1945). Noble, and later Curry (1954), also recognized that the thrust faults have had a complex history including later folding of the faults and later block faulting. Also, Noble recognized that granitic intrusions had spread laterally at some thrust faults as if controlled by them.

The thrust fault first described by Noble was named the Amargosa thrust. Because this type of thrust faulting now is recognized widely in the Death Valley subsection (Longwell, 1945; Mason, 1948; Kupfer, 1960), we extend the term and refer to this system of faults as the Amargosa thrust system, including the faults that branch upward from the main thrust.

About 60 miles east of Death Valley, and near the eastern edge of the Death Valley subsection, is a belt of conventional overthrusts along which the thrusting has been directed eastward—opposite to that of the Amargosa thrust system. At these overthrust faults, Paleozoic formations have been thrust eastward onto Mesozoic ones and the faulting has been interpreted as Late Jurassic, Cretaceous, or early Tertiary (Longwell, 1926, 1928, 1949; Hewett, 1931, 1956).

East of this belt of conventional thrust faults is an area exposing extensive Precambrian granite—part of the Mexican Highland—and east of that is the Colorado Plateau.

The dominant structural and topographic grain of the Death Valley subsection is northerly and northwesterly. The structures however rise southward from under the Tertiary volcanics in the subsection to the north. Within the Death Valley subsection the southward rise is further reflected by the dominance of Paleozoic formations across the northern part of the subsection and the dominance of Precambrian formations across the southern part. Structurally, the Mojave Desert appears to be as high or higher than the south edge of the Death Valley subsection.

Bouguer gravity-anomaly values in the Death Valley area are irregular but high compared to those under the Sierra Nevada to the west and under the Great Basin to the northeast (fig. 70). The gravity values suggest that the crust is relatively thin under the central part of the Death Valley subsection, and that it thickens westward and northeastward.

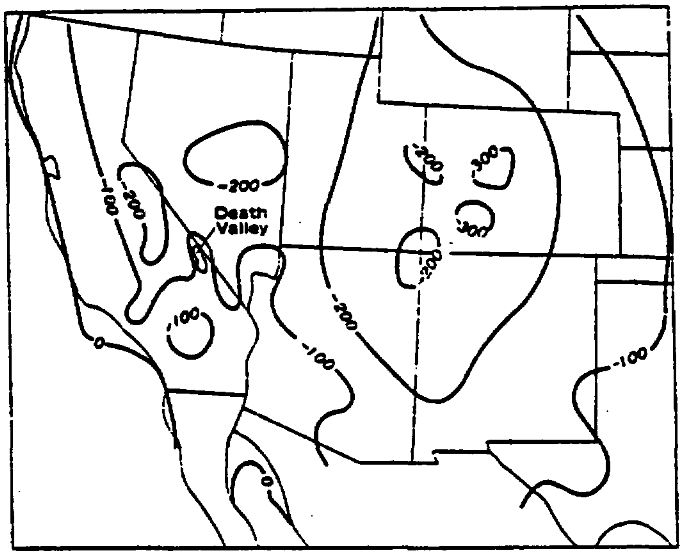

FIGURE 70.—Bouguer gravity-anomaly map of southwestern United States. Contour interval, 100 milligals. Based on G. P. Woollard (unpublished data) and Mabey (1960).

The major structural features of Death Valley and the mountains adjoining it are illustrated on plate 3.

RECENT AND LATE PLEISTOCENE STRUCTURAL GEOLOGY

STRUCTURAL FEATURES OF THE SALTPAN AND GRAVEL FANS

Recent and late Pleistocene structural features of the saltpan and gravel fans include (1) Recent eastward tilting of the saltpan and associated faulting; (2) five small Recent anticlines and five small Recent faults affecting the saltpan; (3) numerous faults including some graben structures and small folds of late Pleistocene age on the gravel fans (fig. 71).

RECENT TILTING OF THE SALTPAN

The saltpan is the product of a Recent lake about 30 feet deep, and dated archeologically about 2,000 years ago (p. A79). The eastern shoreline of this Recent lake is 20 feet lower than the western shoreline. About half the tilting may be accounted for by a 10-foot-high Recent fault scarp along the foot of the Black Mountains (fig. 72).

The eastward tilting of the saltpan is reflected also in differences in the drainage along the two sides. Along the east side of Badwater Basin the flood plain is smooth and is being aggraded; but along the west side the Amargosa River and its tributaries are entrenched in deep channels, and in places the flood plain is being dissected.

That the tilting occurred suddenly about 2,000 years ago rather than progressing gradually during that time is indicated by the arrangement of the concentric rings of salt in the saltpan, for they are crowded against the east side. That the fault scarp developed at the time of tilting is probable. The scarp is well enough preserved to leave no doubt that it is Recent, yet it has been partly destroyed by erosion and partly buried by younger gravel. That it is prehistoric is shown by the fact that Indians constructed mesquite storage pits in the colluvium that overlaps the scarp (figs. 72, 73).

The tilting, together with the faulting along the side towards which the tilting occurred, duplicates in many details the features that accompanied the earthquake at Hebgen Lake in the West Yellowstone area in 1959 (U.S. Geol. Survey, 1959), and very likely the tilting and faulting occurred equally suddenly.

RECENT ANTICLINES AND FAULTS AFFECTING THE SALTPAN

Other recognizable structures that affect the salt crust and that evidently are no more than 2,000 years old include 5 anticlines and 5 small faults. All these represent renewed movement along preexisting structures.

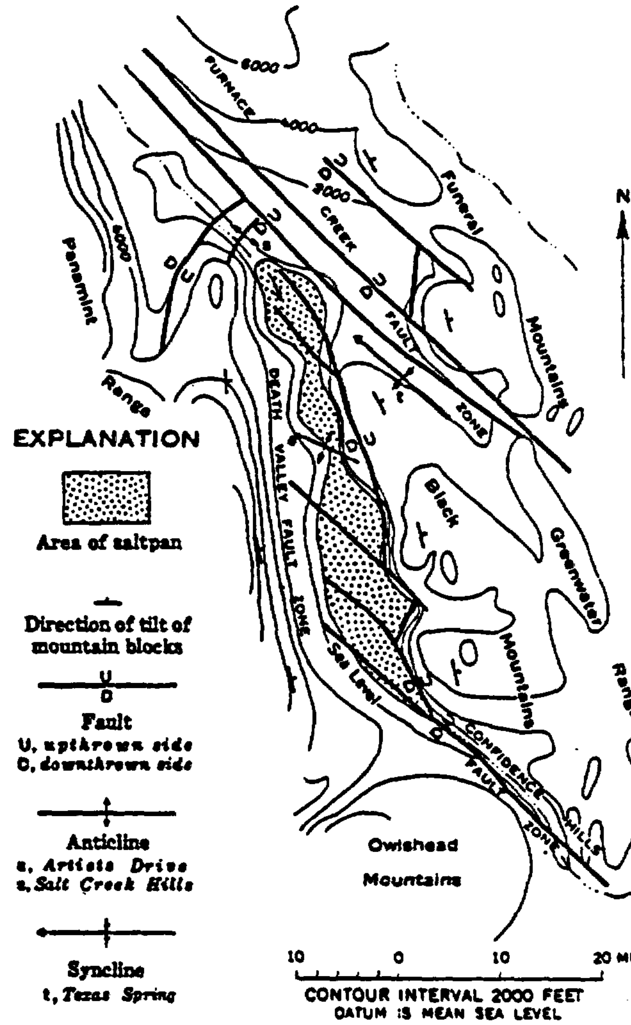

FIGURE 71.—Principal Quaternary structural features in Death Vall[ey]. The Death Valley fault zone is cut off at the north by the Furna[ce] Creek fault zone and at the south by the Confidence Hills fa[ult] zone.

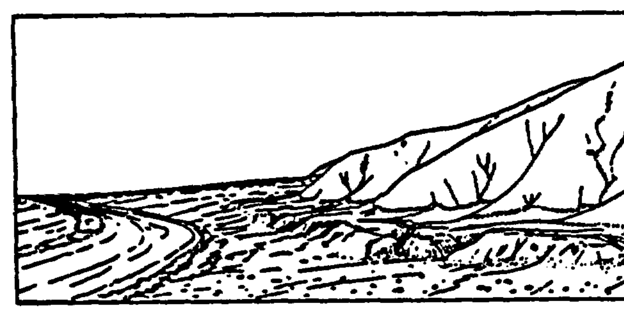

FIGURE 72.—Recent fault scarp at foot of Black Mountains. The and of the escarpment is buried by debris washed from large [can]yons and is dissected by small washes from small gullies on right. View is north on south side of Furnace Creek fan.

Two of the anticlines are in the southern part of Badwater Basin (pl. 3), and both trend northwestw[ard] from anticlines in the older rocks (Precambri[an]) marked by turtleback ridges (Curry, 1954), in the B

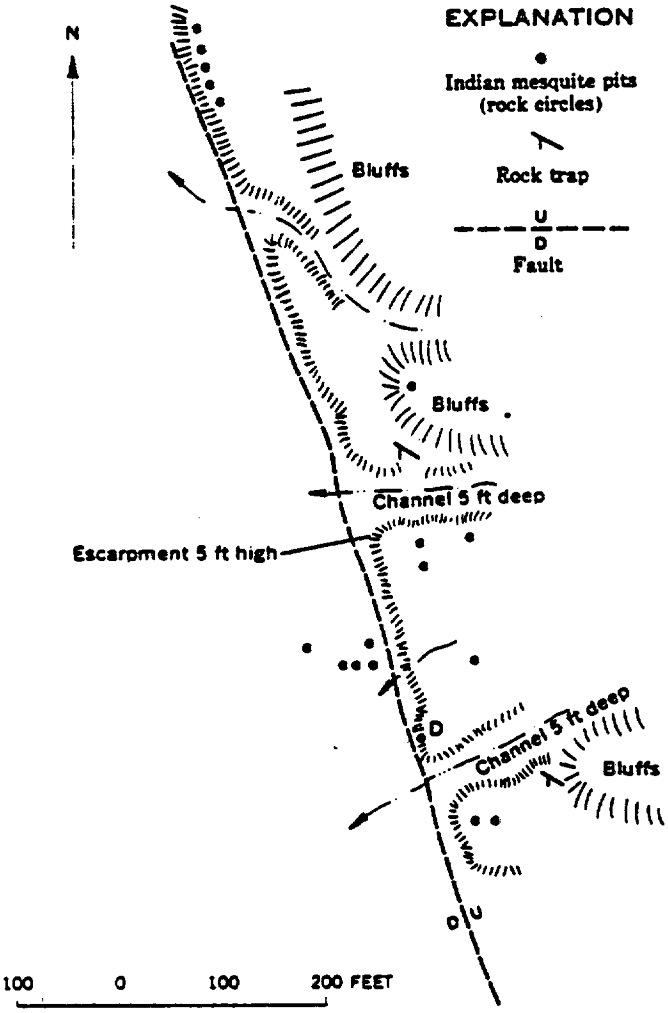

FIGURE 73.—Map of Indian sites along the escarpment of a Recent fault at the foot of the Black Mountains 3 miles south of Furnace Creek Wash (modified from A. P. Hunt, 1960). The fault and escarpment are older than the Indian mesquite storage pit at D, which is built in colluvium overlapping the scarp.

Mountains. They are expressed by sweeping curves in the drainage that lies between the foot of the mountains and the edge of the rock salt deposited by the Recent lake in the interior of the saltpan (fig. 74). The anticlines are manifested by northwest-plunging noses in the gravity contours (pl. 3); this indicates that the anticlines are underlain by highs on the surface of the bedrock, which probably is Precambrian.

A third anticline extends across the saltpan where the highway crosses it at the Devils Golf Course, west of Artists Drive. The trend of this anticline and the location of its highest part are uncertain. Nearby fault blocks of the Funeral Formation along the east edge of the saltpan trend southwest, but the anticlinal axis may or may not coincide with them. Because of Recent uplift on this fold, drainage that formerly was south to Badwater Basin now is ponded in Middle

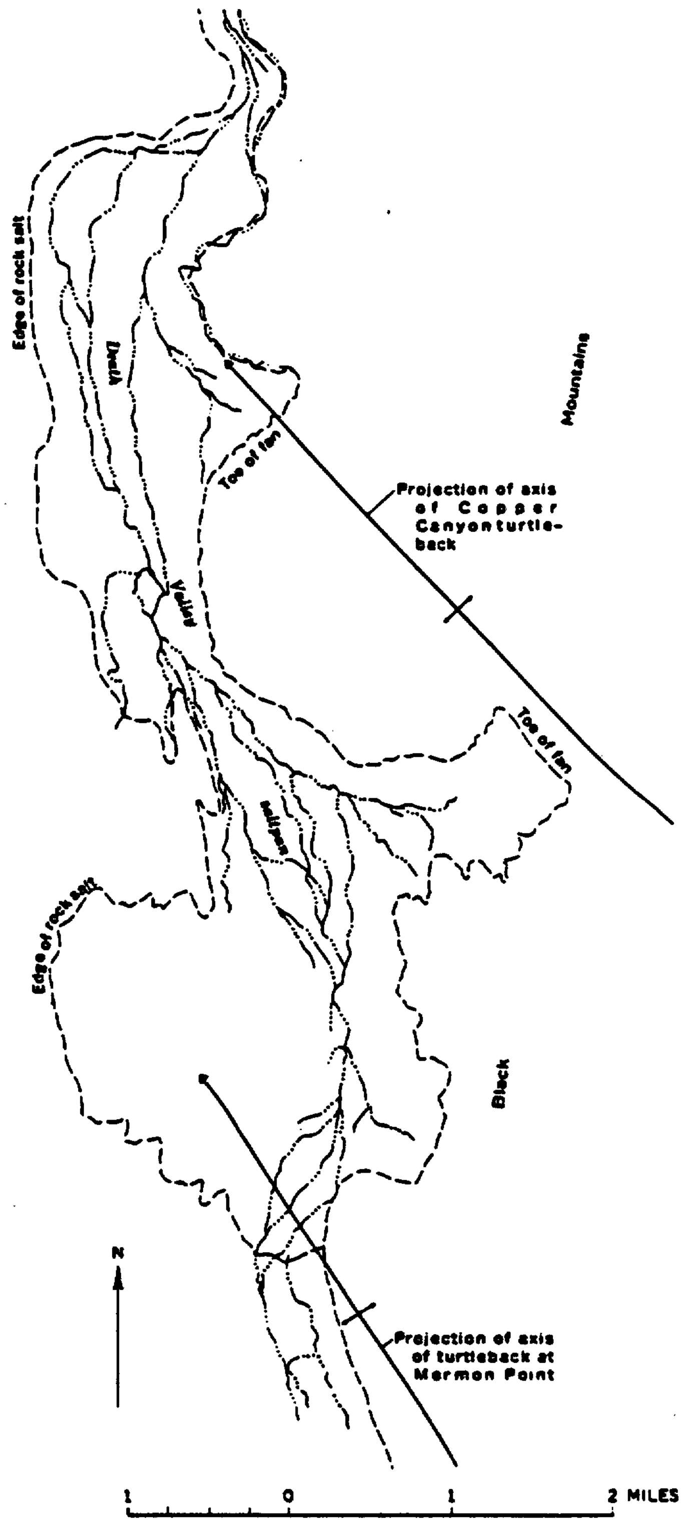

FIGURE 74.—Map showing drainage in the Death Valley saltpan deflected at the projected axes of turtlebacks in the Black Mountains.

Basin north of the fold. Leveling along the channel of Salt Creek, which extends from Middle Basin to

Badwater Basin, indicates that the stream course has been arched 18 inches and that a pond that deep must form in Middle Basin before it will overflow again into Badwater Basin. There was no such overflow during the 6 years 1955–61. Litter along the channel is mostly pre-1900 type, that is, planks with square nails and old bottles with hand-finished necks. There is almost no modern litter; thus, it would appear that there has been no flood of consequence along this stretch of channel during the last half century.

The buried anticline separating Badwater Basin from Middle Basin is marked by a gravity high and is confirmed by drilling that encountered the Funeral Formation 120 feet below the surface of the saltpan. The drill hole is near the middle of the saltpan beside the road across it; a log of the hole is given in table 19, well 2. The drill reached a depth of about 500 feet, still in the Funeral Formation so far as known. Another hole about 7 miles southeast, toward Badwater, was drilled to a depth of 1,000 feet, and did not reach the Funeral Formation so far as can be determined from the log. It may be inferred that the Pleistocene deposits thicken southward from 120 feet above the buried anticline to more than 1,000 feet west of Badwater.

A fourth anticline is in Cottonball Basin. It lies along the projected position of the faulted anticlines of Furnace Creek Formation where the Furnace Creek fault zone (pl. 3) extends under the saltpan. In the center of Cottonball Basin the anticline is marked by a broad arch, 10 inches high in 1958, that partly ponds drainage in the northeast corner of Cottonball Basin

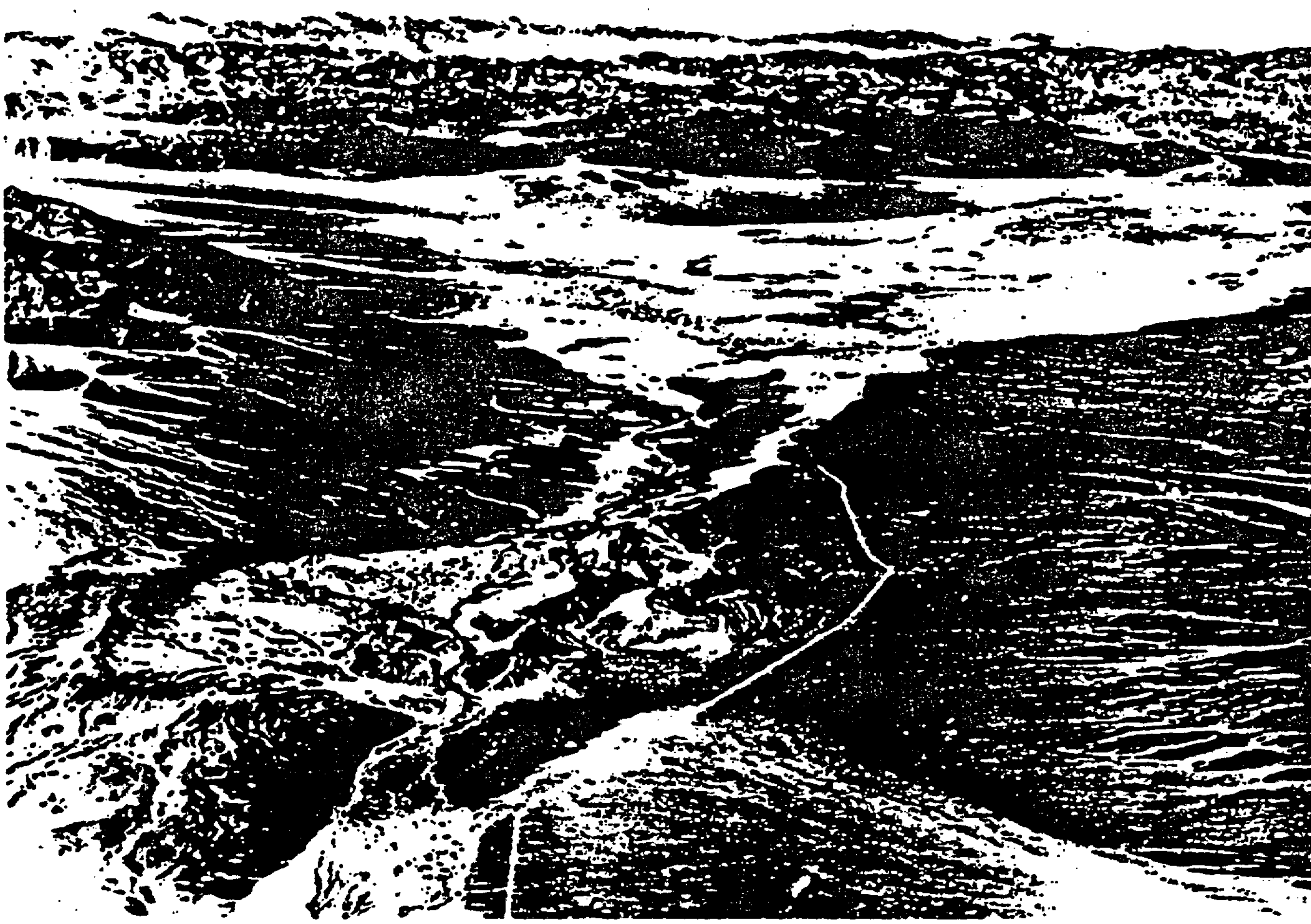

FIGURE 75.—Oblique aerial view of the Salt Creek Hills, an anticline of Pliocene and Pleistocene(?) beds dividing the saltpan (foreground) from the basin at Mesquite Flat (distance). View is northwest. Light-colored beds in the center of the anticline are Furnace Creek Formation. Dark gravel on the south flank of the anticline and gray gravels on the north flank are early Pleistocene and are uplifted less than the Furnace Creek Formation. Upper Pleistocene gravel forms terraces along the stream breaching the anticline from the southwest and is arched less than the lower Pleistocene gravel. Photograph courtesy of John H. Maxson.

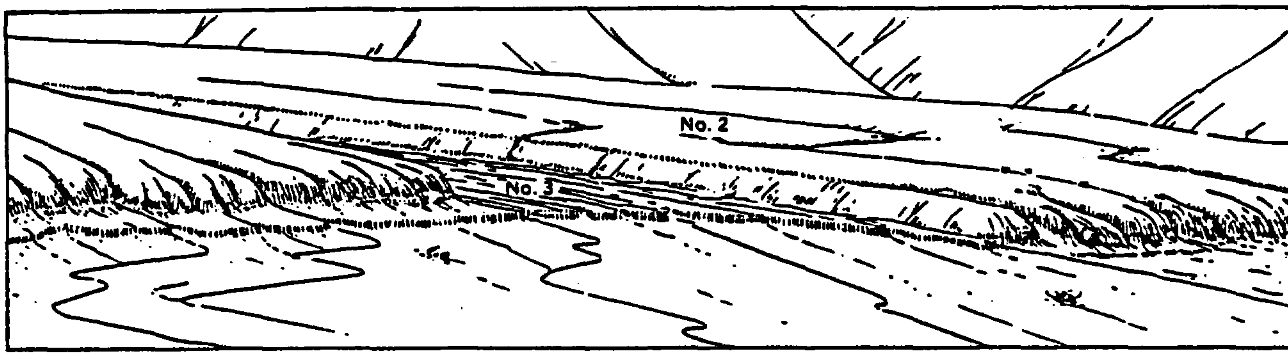

FIGURE 76.—Escarpment of a Recent fault that displaces No. 3 gravel 6 feet at the foot of a late Pleistocene escarpment that displaces No. 2 gravel 50 feet. View westward from near Shortys Well.

and separates it from the main drainage southward on the playa. Only at times of major floods does the ponded water discharge across the arch, and in protracted dry periods dry salt crust forms completely across the drainage line adding to the height of the arch. The tributary drainage in this part of Cottonball Basin is northwesterly, and on the smooth silty rock salt there are some northwest-trending ridges, one of them 5 feet high. The anticline coincides with a northwest-trending gravity low, indicating that the anticline is not marked at depth by a major high on the bedrock.

That the fill in Cottonball Basin is deep is confirmed by a drill hole 1,000 feet deep near the center of Cottonball Basin and on the arch extending northwest into the basin (table 19). From surface to bottom the hole encountered alternating layers of mud and salts that, so far as can be judged from the lithologic description, are entirely Pleistocene in age.

The fifth anticline affecting the saltpan is at the Salt Creek Hills. It marks the north end of the saltpan and divides it from Mesquite Flat (fig. 75). This faulted anticline records a succession of uplifts. The latest, amounting to 10–25 feet, has raised late Pleistocene gravel which now forms terraces along the lower part of Salt Creek and along a tributary from the southwest. An earlier stage of uplift is recorded by the dome of the Pliocene and lower Pleistocene(?) Funeral Formation on which the structural relief is not less than 250 feet and probably is much more. Structural relief on the Furnance Creek Formation is at least a thousand feet, and probably more. The gravity contours indicate that relief on the bedrock basement is still greater.

The gravity data also indicate another deep basin under Mesquite Flat. The anomaly here is larger than in the saltpan, and perhaps the fill of low density rocks is thicker.

Four small faults on the saltpan disrupt the rock salt in the northern part of Badwater Basin (pl. 3). These faults are marked at the surface by linear breaks in the salt crust, but the displacement is small enough to be obscured by irregularities in the salt growths. The position of the fractures is of interest because they are on the projection of the Mont Blanco fault, a southwest-trending rift in the Black Mountains. Whatever buried structure is manifested by these features may extend southwestward to the Hanaupah escarpment on the west side of the saltpan.

FAULTS AND FOLDS ON THE GRAVEL FANS

Many small faults displace upper Pleistocene deposits on the gravel fans, and most trend roughly parallel to the contour of the fans. Many show two stages of movement. The older of the upper Pleistocene deposits, the No. 2 gravel, commonly is displaced 50–75 feet by faults that displace the latest Pleistocene gravels 5–10 feet (fig. 76). There are good examples of faults showing two stages of displacement at the Hanaupah escarpment and at many places along the foot of the Black Mountains, especially at Mormon Point and near the mouth of Furnace Creek (fig. 47). Other faulted gravel fans are at Tucki Wash and at Trail, Starvation, and Johnson Canyons.

Parallel faults in the gravels commonly bound grabens. Some, 10–25 feet deep and 500–1,000 feet wide, break the older (No. 2) of the upper Pleistocene gravels on the fans at Death Valley Canyon and Tucki Wash (pl. 3). Several small grabens only 25–50 feet wide are along the northwest side of the fan south of Badwater. The paved highway follows one of these and appears to be in a roadcut.

In connection with the eastward tilting of the saltpan, reference already has been made to the Recent fault that extends along the foot of the Black Mountains to Mormon Point (p. A100; see also Noble, 1926b). The most recent displacement along this fault is 5–10 feet, but at many places along it, the older, No. 2, gravel is displaced 50–75 feet, as it is at the Hanaupah escarpment.

Late Pleistocene and Recent faulting along the front of the Black Mountains also is recorded by a series of

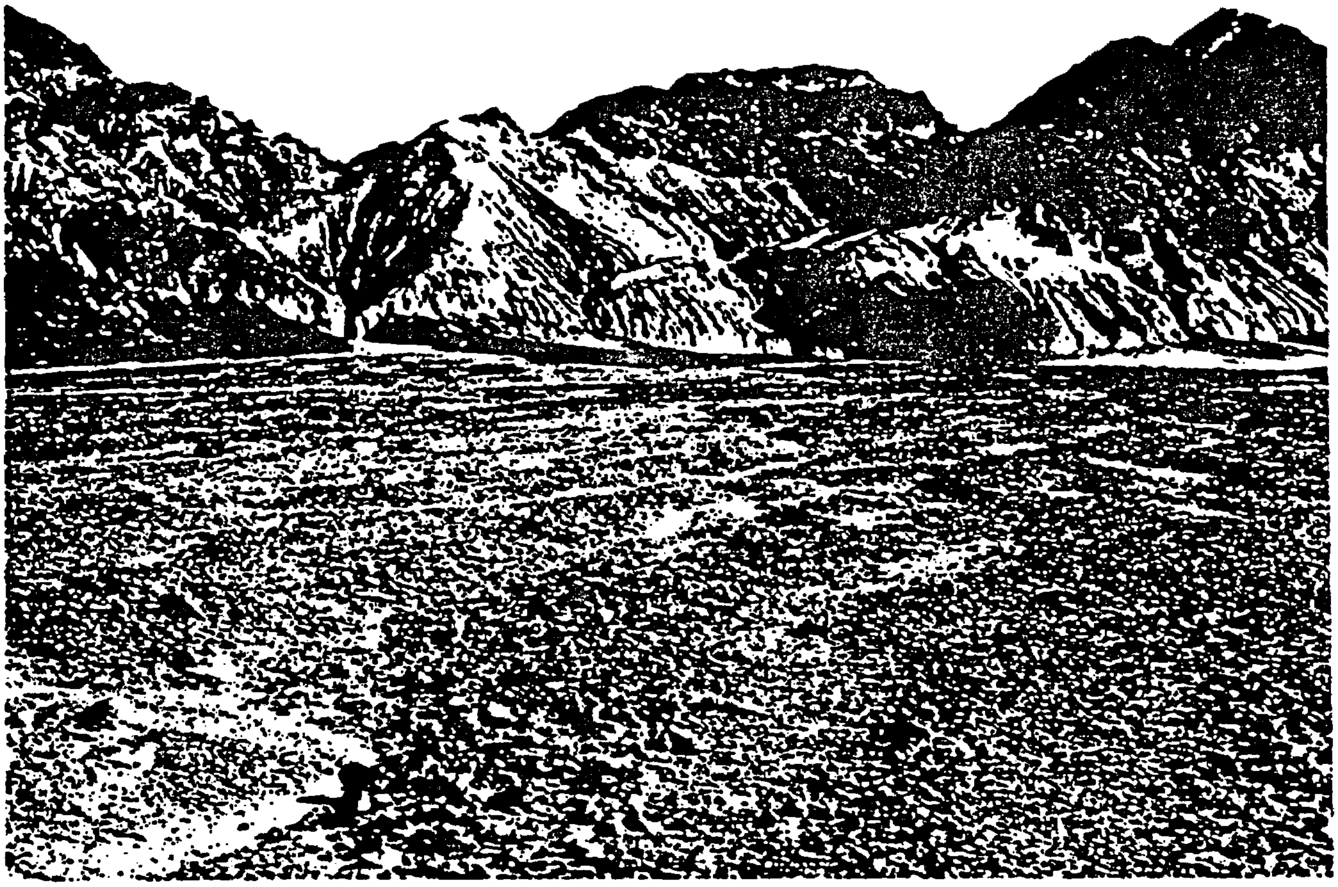

FIGURE 77.—Hanging valleys at the front of the Black Mountains, near the north end of the mountains at Desolation Canyon. Old valleys with U-shaped cross sections have been uplifted about 100 feet along this part of the mountain front, and the new valleys are narrow gorges incised into the bottom of the older, more open valleys. The bedrock consists of interbedded volcanic and sedimentary rocks of Miocene or Pliocene age.

hanging valleys in the mountain block (figs. 77, 78). In the vicinity of Furnace Creek fan the hanging valleys are about level with upfaulted benches of No. 2 gravel and evidently are no younger than that gravel; they may be somewhat older.

The height of the hanging valleys increases southward, which accords with other evidence that the mountains have been tilted northward as well as being faulted upward. It is difficult, however, to obtain very meaningful figures about the heights of the hanging valleys because of differences in the kinds of rock and ease of downcutting at the valleys.

For example, at one canyon, Golden Canyon, in the northern part of the Black Mountains the hanging valley is at an altitude of −40 feet. A mile south, at Gower Gulch, the hanging valley is at sea level. About 1¼ miles farther south, at Desolation Canyon, the hanging valley is 100 feet above sea level. But the northern canyon is cut in soft beds of the Furnace Creek Formation; the middle canyon is cut into moderately resistant conglomerate at the base of the Furnace Creek Formation; and the southern canyon is cut into a resistant block of dolomite. Part of the northward decrease in height may be due to northward tilting, but part probably is due to greater erodability of the rocks northward.

Another comparison that involves less variable channels is provided by 2 gulches 1 mile apart on the side of the Badwater turtleback of Precambrian rocks. Each widens upward into a more open gulch, the northern one at an altitude of 800 feet; the southern one at an altitude of 1,200 feet. This suggested tilt of 400 feet northward is about the same as the northward tilt of the upper surface of the turtleback.

The hanging valleys do not necessarily indicate northward tilt of the Black Mountains, but at least their heights do not conflict with such tilt.

The hanging valleys are approximately the age of the No. 2 gravel; this is suggested by the occurrence of benches of No. 2 gravel near the mouth of Furnace Creek that are displaced by about the same amount as nearby hanging valleys (fig. 47).

The No. 2 gravel is broken by many faults having displacements of 50–100 feet, specifically at Mormon Point, Furnace Creek fan (fig. 47), East Coleman Hills, Mustard Canyon, Salt Creek Hills, Tucki Wash (pl. 2), Blackwater Wash, Death Valley Canyon (pl. 2), Hanaupah escarpment (fig. 76 and pl. 2), and the fans at Six Spring Canyon (pl. 2) and south.

Upper Pleistocene gravels are downfolded into the northwest-plunging Texas Spring syncline separating the Black and Funeral Mountains (pl. 3). In this syncline Pleistocene deposits are less folded than the Pliocene and lower Pleistocene(?) Funeral Formation, and the fanglomerate, in turn, is less folded than the Furnace Creek Formation (fig. 79)—further evidence that the deformation occured in multiple stages.

Upper Pleistocene gravel is folded and faulted at many places along the Furnace Creek fault zone where the zone plunges under the saltpan at the edge of Cottonball Basin. At Mustard Canyon Hills, for example, upper Pleistocene gravel unconformably overlies and is anticlinally folded above upper Pliocene deposits (fig. 80). Strike faults along the flank of this dome have displacements 5–10 feet down in the direction of the dip. This late Pleistocene doming is 100 feet or more.

Reference has already been made to the doming of the upper Pleistocene gravel at Salt Creek Hills across Cottonball Basin from the Mustard Canyon Hills.

Evidence about tilting of shorelines of the upper Pleistocene lake, Lake Manly, is not at all satisfactory. Two hundred feet of eastward tilt and 300 feet of northward tilt since Lake Manly time is indicated—

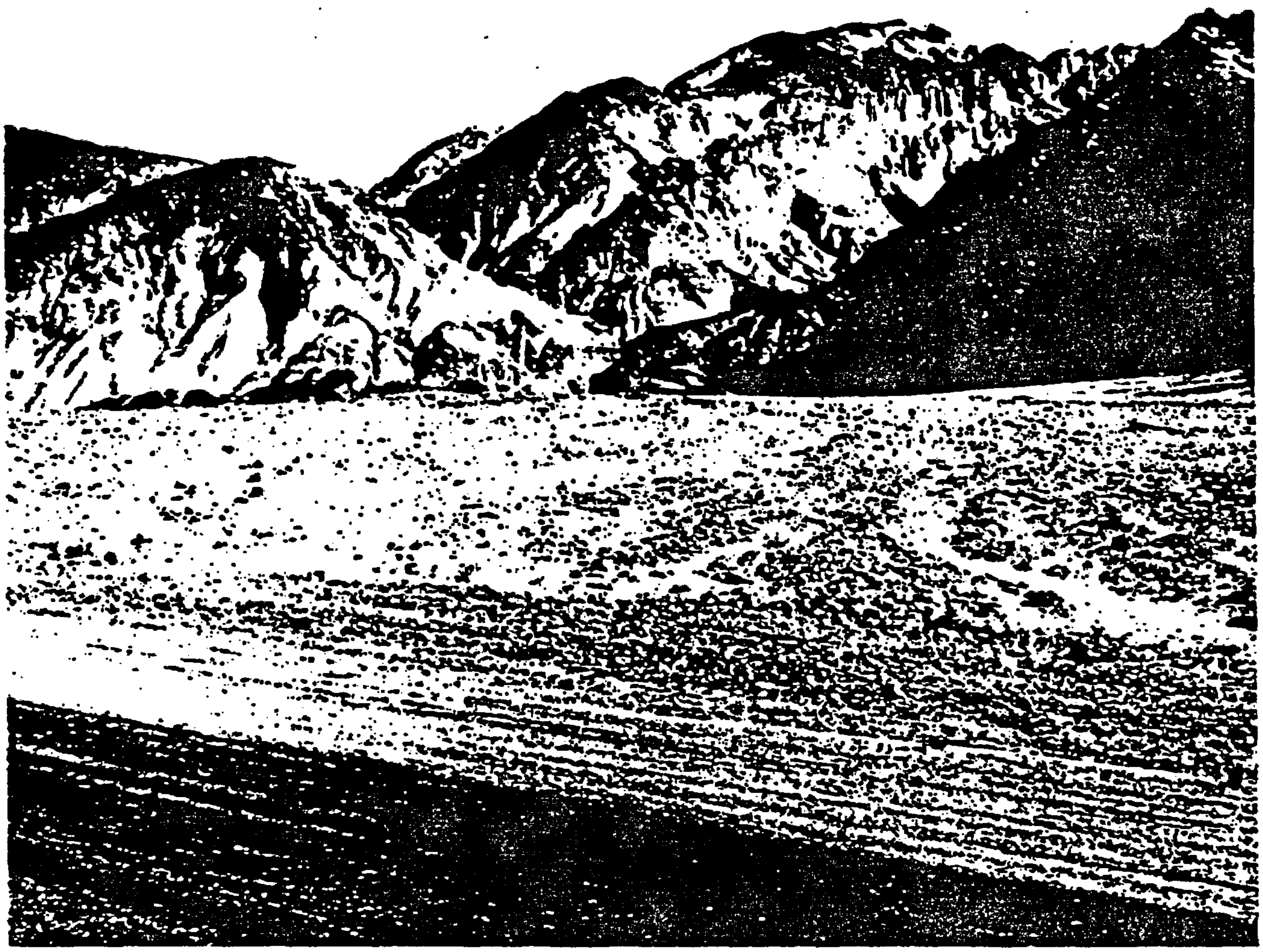

FIGURE 78.—Hanging valley at Gower Gulch at the front of the Black Mountains. The floor of the old valley has been raised 50–75 feet above the apex of the fan. Furnace Creek Formation (extreme left) overlies Artist Drive Formation.

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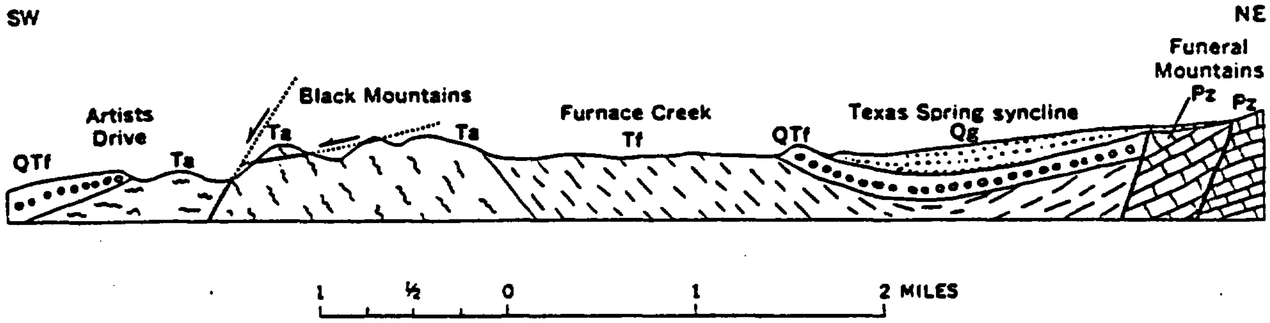

FIGURE 79.—Section from the Funeral Mountains southwestward across the Texas Spring syncline and Black Mountains to the fault blocks at Artists Drive. Pz, Paleozoic formations; Ta, Artist Drive Formation; Tf, Furnace Creek Formation; QTf, Funeral Formation; Qg, upper Pleistocene fan gravel. The Artist Drive Formation is more deformed than is the Furnace Creek Formation, and the latter is more deeply folded under the Texas Spring syncline than is the Funeral Formation. Upper Pleistocene gravel is slightly folded in the syncline. Vertical scale not exaggerated.

(p. A71) a reasonable supposition. But the deposits are discontinuous, and the correlations along and between the two sides of Death Valley are uncertain. About all that can be concluded from the evidence provided by the upper Pleistocene lake deposits is that this evidence does not conflict with what is known about the structural history. The same seems to be true of Panamint Valley (Maxson, 1950, p. 107).

Evidence of late Pleistocene deformation also is provided by the geomorphology of the gravel fans. The fans on the two sides of Death Valley are strikingly different; those at the foot of the Black Mountains are small, whereas those sloping from the Panamint Range are long and high. There is similar difference between the two sides of the Panamint Range (Murphy, 1932, p. 353). The differences assuredly reflect the eastward tilting, probably not of the whole area but of individual fault blocks.

Also, differences in the fans along the foot of the Panamint Range suggest there is a northward component in the tilting of that range. The fans at the north extend far into the mountains whereas those at the south extend only to the mountain front (fig. 66 and pl. 2).

FIGURE 80.—View along the southwest flank of the anticline at the Mustard Canyon Hills. Upper Pleistocene gravel (dark beds) unconformably overlaps the Furnace Creek Formation (light-colored beds) and is turned up 20° along the flank of the anticline. Photograph by John R. Stacy.

VALLEY FILL

GRAVITY FEATURES

The gravity contours shown on plate 3 are based on the complete Bouguer anomaly values. The data were reduced with an assumed density of 2.67 g per cm³ (grams per cubic centimeter). The terrain corrections are large over all the area and locally extreme (more than 70 mgals at Telescope Peak); however, the terrain corrected data are considered accurate enough to justify the 5- and 10-milligal intervals used in contouring.

The large local gravity lows in the valley are produced primarily by the relatively low-density Cenozoic fill in the troughs. The more extensive variations are produced by density variations in the older rocks and perhaps by effects deep within the crust or in the upper mantle. The quantitative interpretation of the gravity data is limited by uncertainties in the density of the rocks producing the anomalies. Horizontal and vertical changes in the density of basin fill produce gravity variations that cannot be distinguished from effects of relief on the bedrock-fill interface. In parts of Death Valley the problem of interpretation is further complicated by large gravity anomalies due to variation in the bedrock. Although the bedrock anomalies are more extensive than the basin anomalies, the two cannot always be separated satisfactorily. Despite these limitations, the gravity data can be used to estimate the approximate thickness of the basin fills and to indicate structures that produce relief on the interface between the Cenozoic rock and the more dense older rock. Some of the gravity effects produced by deeper mass anomalies are also useful in studying the structural geology.

In a basin several times as wide as it is deep, the amplitude of the gravity anomaly between Cenozoic fill and the dense older rocks primarily depend upon the density contrast and thickness of the fill. Computations of the thickness of fill are no more accurate than the assumed density contrast. In Death Valley there are no drill hole or seismic depths to bedrock to serve as control for interpreting the gravity data, and densities are inferred from surface samples. Gravity studies elsewhere in the Great Basin indicate that the average density contrast between the Cenozoic fill and the older rocks is about 0.4 g per cm³. The contrast may range from near zero to about 0.7 g per cm³. If the contrast is 0.4 g per cm³, a gravity anomaly of about 5 milligals will be produced by each 1,000 feet of fill. If the density contract is as great as 0.7 g per cm³, the actual thickness will only be about 60 percent as great as computed assuming a contrast of 0.4 g per cm³. However, if the density contrast approaches zero, the actual thickness may be several times greater than the depth computed, using a contrast of 0.4 g per cm³ (fig. 81). The depths

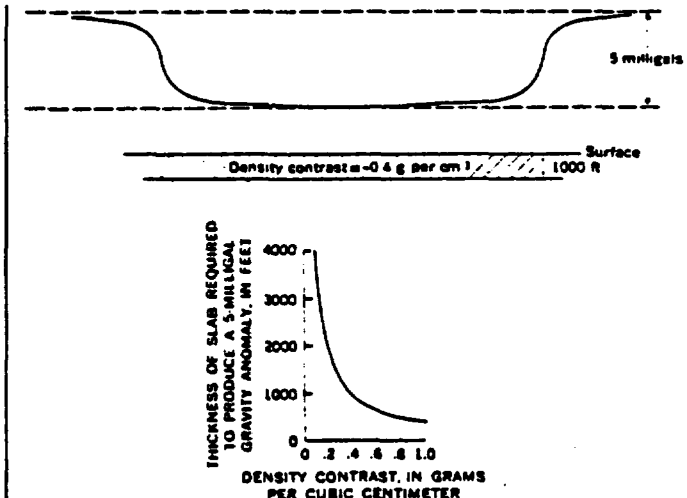

FIGURE 81.—Diagram illustrating the gravity anomaly produced by two-dimensional prism and the relationship between the amplitude of the anomaly and the density contrast between prism and the enclosing material.

we infer from the gravity data are based on an assumed density contrast of about 0.4 g per cm³.

A gravity low extends the entire length of Death Valley. This low is divided into three principal areas of low gravity closure separated by the two high trends across the valley. The low areas are in Mesquite Flat, Cottonball Basin, and Badwater Basin; the two high trends coincide with the anticlines at the Salt Creek Hills and opposite Artists Drive.

The gravity low in Mesquite Flat has the greatest relief of any of the lows in Death Valley. A part of this negative anomaly is probably related to the low anomaly values in the Cottonwood Range, but a local anomaly of 40–50 milligals is produced by rocks underlying the basin. This is the largest residual anomaly in the region, and it indicates a subsurface basin containing about 2 miles of Cenozoic fill with the thickest section near the center of the basin.

Near the south and east edges of Mesquite Flat are steep gravity gradients, which are interpreted as indicating faults with large vertical displacement. The fault indicated by the gravity data along the northeast side of the basin is part of the Furnace Creek fault zone. The steep gradient north of the front of Tucki Mountain probably indicates a fault trending a few degrees north of east. Not enough gravity data were obtained along the west side of the basin to define adequately the gravity gradient; however, a steep gradient, probably related to faulting, is indicated about 3 miles north of the mouth of Marble Canyon.

The minimum anomaly values in Cottonball Basin are near the center of the valley along the east side of the playa. The maximum gravity-anomaly relief between the floor of the valley and stations on bedrock in

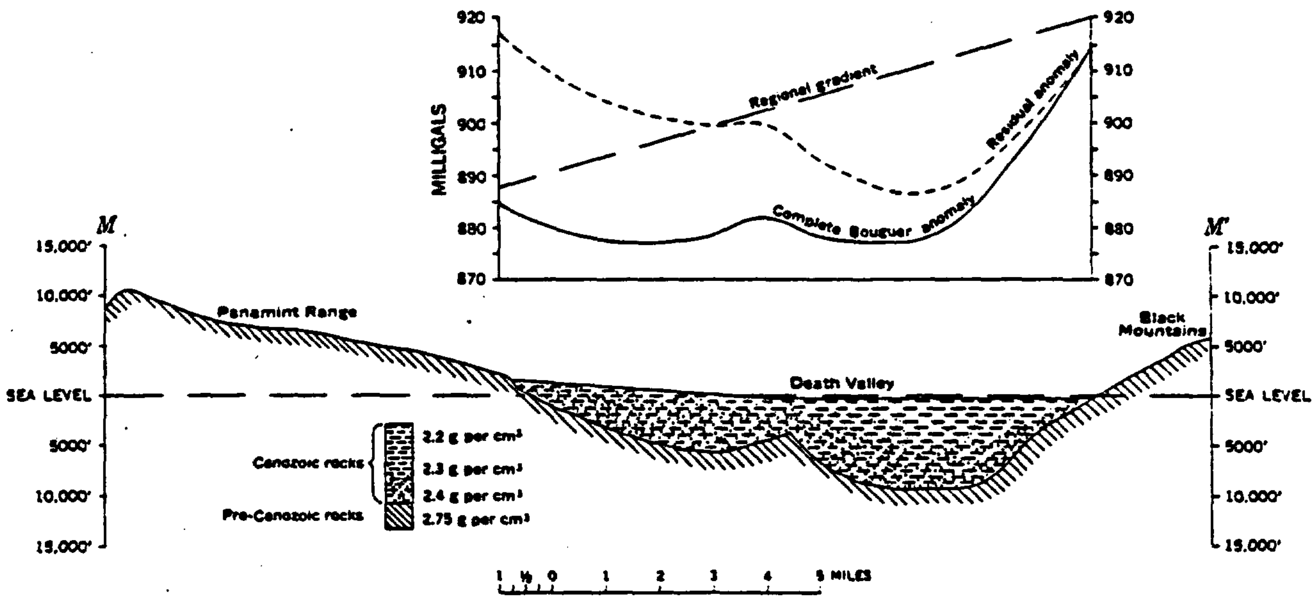

FIGURE 82.—Gravity and inferred bedrock profile across Badwater Basin.

the Funeral Mountains is 43 milligals. Along the northeast side of the Furnace Creek basin the gravity data indicate vertical displacement along more than one fault of the Furnace Creek fault zone.

To the south the gravity anomaly bifurcates with one branch trending south down Death Valley and the other trending southeast along the Furnace Creek fault zone under the Texas Spring syncline. The gravity data indicate a considerable thickness of Cenozoic rock; under the Texas Spring syncline most of it must be Tertiary. This Cenozoic rock thins southeastward. The gravity data do not indicate faulting along the northeast side of the Black Moutains.

The main branch of the gravity low in the Cottonball Basin continues down Death Valley to the Devils Golf Course opposite Artists Drive where a gravity saddle separates the low anomaly in the north from the low in Badwater Basin. This coincides with the buried anticline that has been confirmed by drilling (p. A102).

The low gravity anomaly in Badwater Basin consists of two areas of low closures separated by a north-trending high down the west side of the saltpan, but the significance of these gravity data is obscured by the large regional gravity variation across the valley (fig. 82). We infer a structural high along the west side of the saltpan separating two areas of deep fill. The fill under the saltpan is computed to be 9,000 feet deep, of which two-thirds is estimated to be Tertiary. The interpretation of the western half of the profile is very doubtful because a major change from the assumed regional gradient would substantially alter the residual anomaly.

South of Mormon Point and south of the area show on plate 3 the axis of the gravity anomaly trends a litt east of south and extends down the center of the narro part of the valley between the Owlshead Mountains ar the Black Mountains. In the Confidence Hills the min mum anomaly values occur directly over the hills, i dicating that the greatest thickness of Cenozoic roc occurs under the hills and that the surface relief is n reflected on the bedrock surface. Gravity evidence i dicates faulting on both sides of the valley in this are The low anomaly diminishes to the south to a small lo gravity closure north of the Avawatz Mountains.

MAGNETIC FEATURES

Aeromagnetic data were obtained along profiles Death Valley flown at an altitude of 3,500 feet abo sea level. One profile is along the axis of the valley ar six are at widely spaced intervals across the vall (fig. 83).

The local magnetic relief at the south end of profi *A–A'* is produced by Tertiary volcanic rocks. The ge eral rise in the total magnetic intensity toward t north is about equal to the normal regional magne gradient. The gentle undulations several miles in e tent with relief of less than 100 gammas probably r flect deep features, either relief on the top of baseme complex or, more probably, compositional chan within the basement rock. The local anomaly betw Badwater Basin and Cottonball Basin coincides w the buried anticline there. The anomaly is produ by a relatively shallow feature, probably volca material within the basin fill.

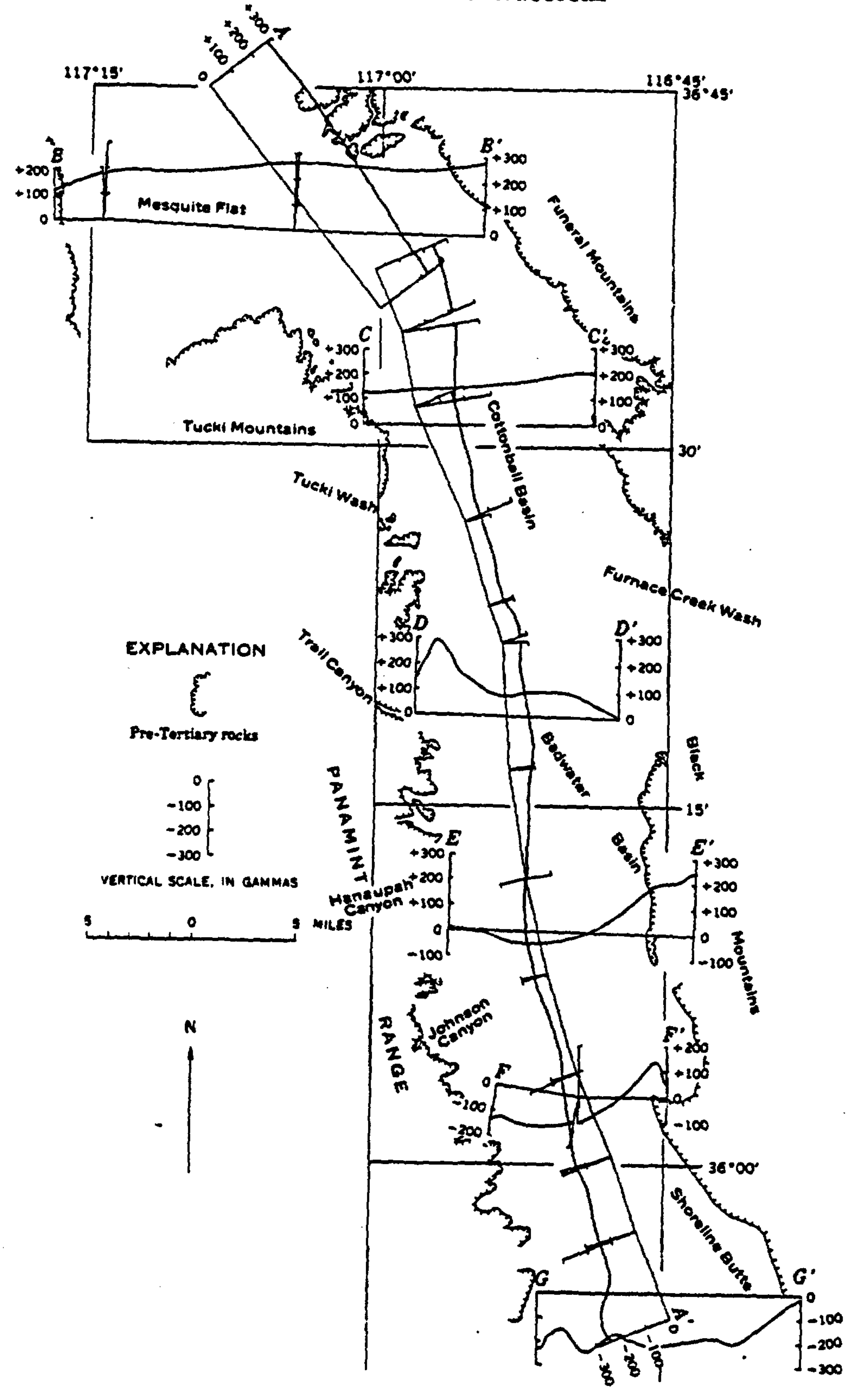

FIGURE 83.—Total intensity aeromagnetic profiles across Death Valley. Flight level is 3,500 feet above sea level.

Profiles *B–B′* and *C–C′* show the normal regional gradient across the valley without any large local anomalies. On profile *C–C′* there is an indication of the south end of the positive anomaly along the northeast side of Mesquite Flat, probably produced by the basement rocks.

Profiles *D–D′*, *E–E′*, and *F–F′* across the Badwater Basin show considerable local magnetic relief. Profiles *D–D′* and *E–E′* show a westward increase in the magnetic field along the west side of the valley, and at the west end of profile *D–D′*, the anomaly has a peak. Depth estimates on this anomaly indicate that the feature producing the anomaly is within 1,000 feet of the surface. These anomalies coincide with the volcanics and intrusions in the Amargosa thrust complex along the east foot of the Panamint Range.

On profile *D–D′* there is a decrease in the total intensity as the Black Mountains are approached; on profiles *E–E′* and *F–F′* the total intensity increases toward the Black Mountains. The latter two profiles are in the area where the older Precambrian complex makes up the core of the Black Mountains, and it seems probable that a major part of the rise in the total magnetic intensity is related to the elevation of the Precambrian rock. The local reversal near the east end of profile *G–G′* has a near-surface cause. The magnetic maximum is where the older Precambrian rock of the Mormon Point turtleback extends across the profile. To the east of this outcrop is Cenozoic rock. The magnetic reversal may be related to a high over the older Precambrian rock or to a low over buried volcanic rocks to the east.

The southernmost profile *G–G′* has considerable magnetic relief west of the center of the valley. One anomaly is a local high over Shoreline Butte, which contains a considerable volume of basalt; probably the other anomalies to the west are also produced by volcanic rocks in the Cenozoic fill. From Shoreline Butte the magnetic field increases eastward over the older Precambrian complex in the Black Mountains.

CHANGES IN THE ALTITUDES OF BENCH MARKS

Some evidence of continuing deformation in Death Valley is provided by the change in altitudes determined for bench marks between the original level surveys in the valley in 1907 and releveling in 1933–35 and 1942–43. In 1907 the U.S. Geological Survey surveyed level lines into Death Valley to provide vertical control for the 1:250,000 topographic maps of the region. Bench marks were set along a line which followed the road along the west side of Badwater Basin, crossed the valley floor at the Devils Golf Course and went around the north and east sides of Cottonball Basin and Mesquite Flat. Another line crossed the valley from Daylight Pass to Emigrant Canyon. This early leveling although providing adequate control for the topographic mapping, was not sufficiently accurate to provide a measure of the regional deformation when compared with later leveling. However, local deformation within the valley is indicated by comparing the original altitude differences between bench marks in the valley with altitude differences determined on the later surveys (fig. 84). Unfortunately only a few of the original bench marks were recovered in the later surveys, and these provide an incomplete picture.

The indicated general decrease in altitude relative to the reference bench mark near Wingate Pass may be produced by systematic errors in the original data, but in general the indicated changes within the valley are probably real. The largest subsidence is 0.73 foot for a bench mark along the southwest edge of Badwater Basin. At a bench mark about 5 miles to the southeast near the end of the saltpan the indicated subsidence is 0.16 foot. This is consistent with continued subsidence of the Badwater Basin.

On the east side of Cottonball Basin 3 bench marks over an interval of about 12 miles show changes of −0.07, +0.01, −0.65, and −0.03 foot. The bench marks with the −0.07 and +0.01 change are just southwest of the Kit Fox Hills fault with +0.01 bench mark closest to the fault. The −0.65 bench mark lies between the Kit Fox Hills fault and the anticline in the Cottonball Basin. The −0.03 bench mark is south of the anticline and over the gravity nose extending northwest from the Black Mountains.

The data for 3 bench marks southeast of Mesquite Flat indicate subsidence ranging from 0.28 to 0.68 foot.

These indicated subsidences may be due partly to compaction of sediments in the valley fill and partly to structural deformation.

SEISMIC ACTIVITY

Death Valley is in a seismically stable area at the edge of a highly active area (fig. 69). Fifty miles southwest of Death Valley, epicenters are closely spaced, and a rather sharp northwest-trending line marks the northeast edge of the seismically active area. Only a dozen earthquake epicenters have been located in the valley or in the mountains adjoining it, and none of these was sufficiently intense to be felt by persons in the area. There is no historic record of earthquake damage. Apparently the last major earthquake originating in Death Valley occurred 2,000 years ago when the floor of the valley was tilted 20 feet eastward and the 10-foot escarpment formed along the foot of the Black Mountains (p. A100). The geologic effects of that

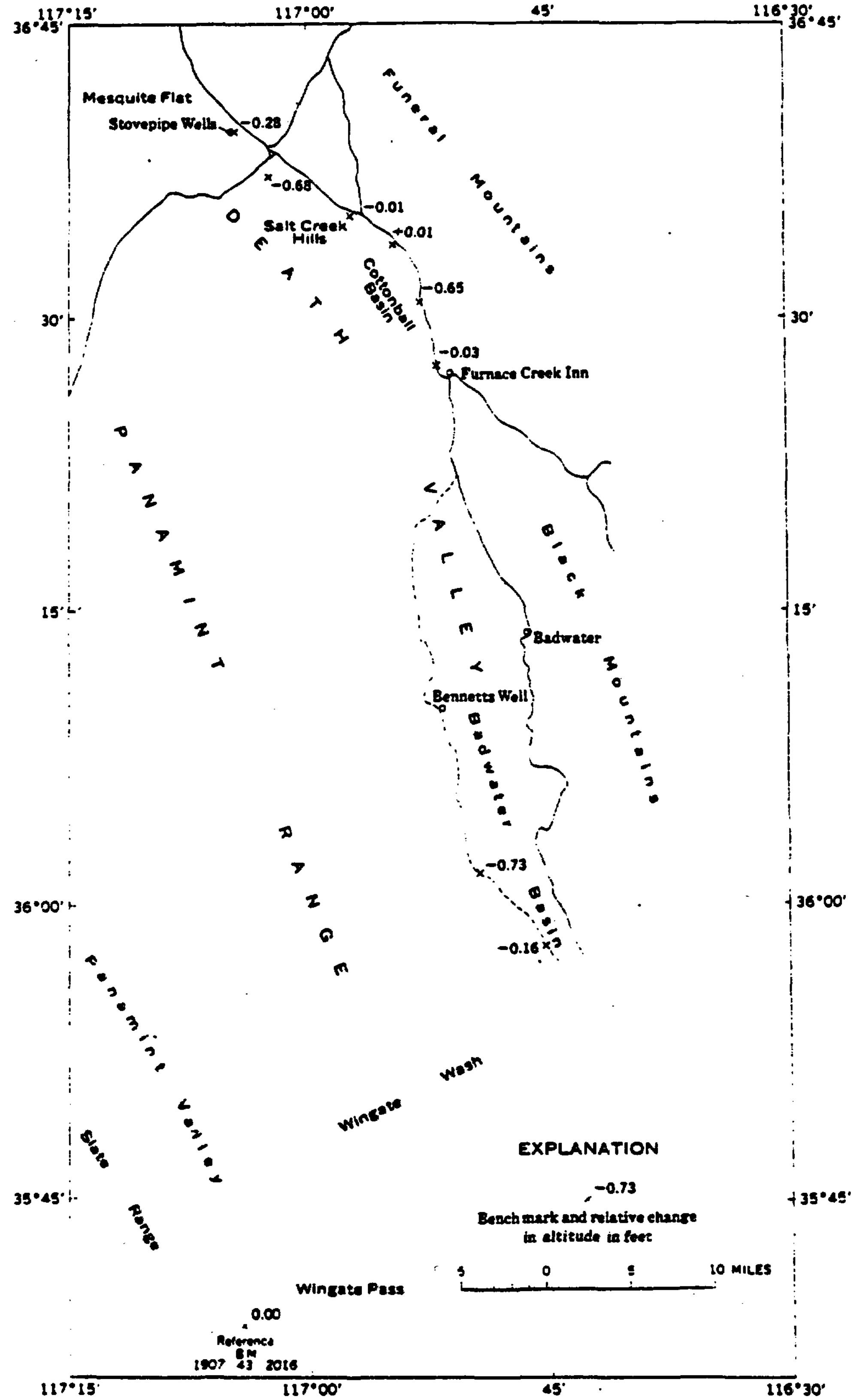

FIGURE 84.—Relative changes in altitude of bench marks in Death Valley between unadjusted 1907 U.S. Geological Survey data and adjusted 1933 and 1942–43 U.S. Coast and Geodetic Survey data. The changes are relative to bench mark BM 1907 43 2016 near Wingate Pass.

earthquake, and perhaps its intensity, approximately duplicated those of the 1959 earthquake at Hebgen Lake in the West Yellowstone area.

In the other direction, northeast of Death Valley, a series of epicenters alined northwesterly roughly along the California-Nevada boundary separates the Death Valley area from a more extensive stable area covering most of the eastern part of the Great Basin. This latter stable area coincides with a regional gravity low (fig. 70). Finally, another series of epicenters is located along the boundary between the Basin and Range province and the Colorado Plateau.

Data on earthquakes in the Death Valley region during the period from 1934 through 1958 were compiled and made available to us by C. R. Allen, of the California Institute of Technology. These data show 12 epicenters in the main part of Death Valley or on the adjoining mountain slopes. These quakes range in magnitude from 2.5 to 4.0 on the Richter scale. The uncertainty in the location of the epicenters is too great to justify any attempts to correlate them with known faults.

Seismic-refraction data indicate that the crust under much of the Great Basin consists of two layers and that it thickens northeastward (Press, 1960; Berg and others, 1960; Diment and others, 1961). The upper layer or layers have a velocity of 6.3 kmps (kilometers per second) or less and are underlain at depths between 20 and 25 km by a layer having a velocity of 7.6–7.8 kmps. The base of this layer slopes northeastward from a depth of about 50 km in southern Nevada to about 74 km in northwest Utah. The underlying material has a normal mantle velocity of about 8 kmps.

Such layering within the crust is not indicated westward from Death Valley to the Sierra Nevada (Carder and Bailey, 1958). Under the Sierra Nevada, velocities in the crust are uniform. The crust appears to be homogeneous and at least 40 and perhaps more than 50 km thick. Gravity data (fig. 70) and seismic data indicate that the upper crust in the Death Valley area is little more than half that thick, say 25 km. Assuming these thicknesses, the base of the upper crust must slope about 15° westward from Death Valley. The slope to the northeast would be about half that. Death Valley appears to be over a ridge on the mantle.

TILTMETER MEASUREMENTS

By Gordon W. Greene

METHODS

Seven tiltmeter stations were established in the Death Valley area in 1958 and 1959 to determine if measureable tilt was occurring. The location of the tiltmeter stations is shown on plate 3. A detailed description of the portable liquid tiltmeters used in this study is given by Eaton (1959).

Each tiltmeter station, except those in adits, was laid out in a nearly equilateral triangle with sides 30–50 meters long. At each apex a machined brass hub was set in concrete upon rock outcrops, and hub tops were established within ±0.5 cm of a level plane. Variations in the size and shape of the triangles were made because of local terrain. One station, Trail Canyon fan, was established by using large boulders embedded in gravel instead of bedrock outcrops.

Temperature gradients between hubs and rapid changes of air temperature cause erratic readings. To avoid heating of the system by solar radiation, the tiltmeter was used at night, or when the sky was heavily overcast. Heat radiated from the ground at night, especially when the sky is clear, also makes observations difficult. A gentle wind can assist in maintaining a constant temperature in the system.

Under ideal conditions, altitude differences between hubs can be measured with an error of less than three microns. Thus, if the hubs are 30 meters apart, a sensitivity of 1 part in 10 million can be realized. The precision of the measurements is checked by closing the measured altitude differences around the three sides of the triangle.

In practice, closure errors as much as 50 microns were encountered because of the difficulty in maintaining the entire system at a constant temperature. Readings were considered valid if the closure error was less than 10 percent of the total change in hub altitudes.

MEASUREMENTS

Tiltmeter observations show that the ground surfaces in the Death Valley area are being tilted at present. The direction and magnitude of tilting varies from one station to another, and at any given station both direction and magnitude vary from time to time. Figure 85 shows the tilting observed at five stations in the Death Valley area.

The direction of tilting is consistent with the structural geology and tends to follow the present dip of strata at each station. Tilting is to the northeast at Aguereberry Point, Trail Canyon, and Dantes View. These stations are located on blocks which have been tilted to the east or northeast. At Artists Drive, tilting is predominantly toward the southwest, although there is a distinct tendency at times to tilt southeastward. Here the Tertiary formations dip southeastward, and the overlapping early Pleistocene fanglomerate dips westward. Within the valley, tilting at Trail Canyon fan is chiefly toward the south, which accords with its position on the flank of the anticline north of Badwater Basin.

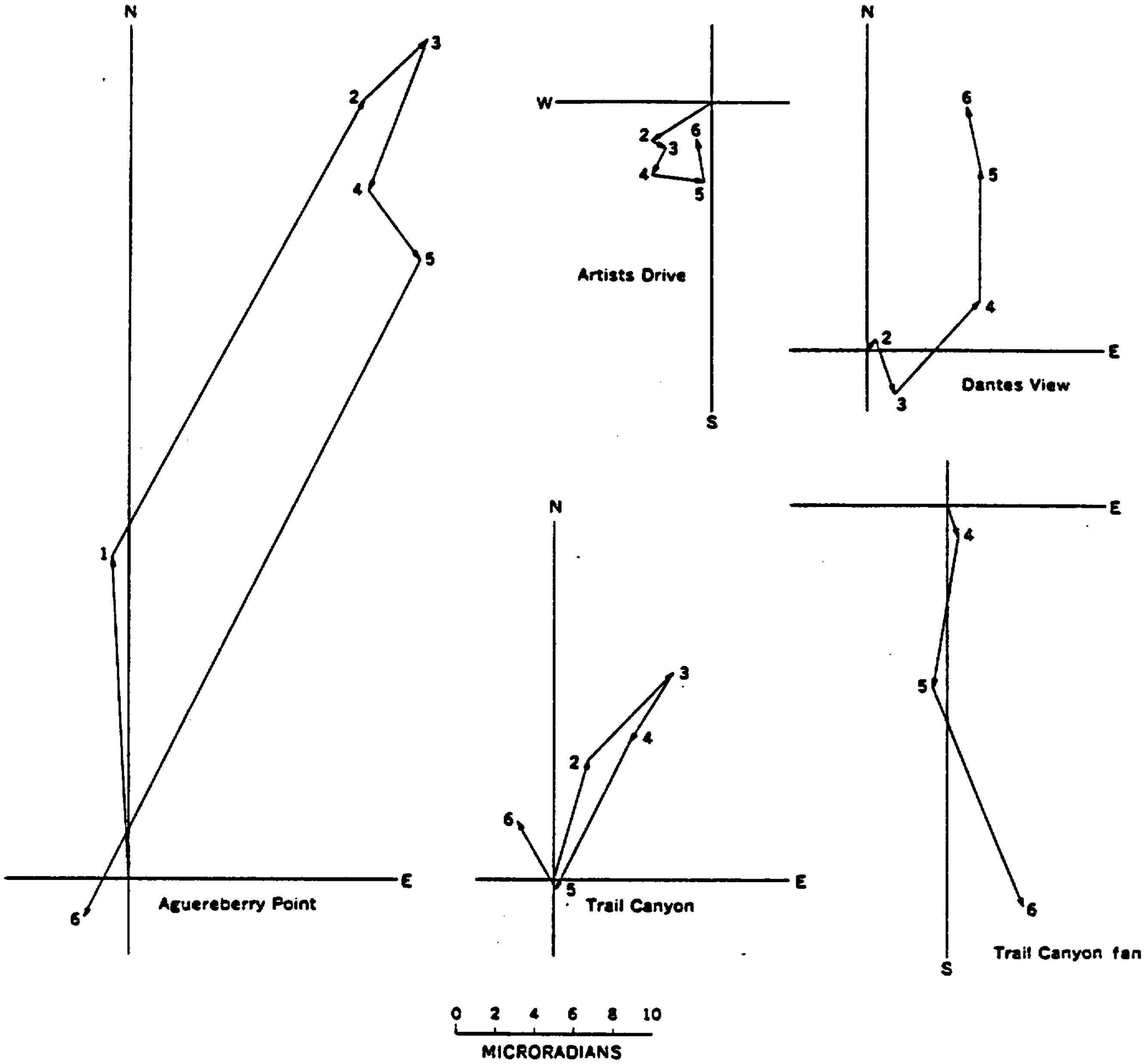

FIGURE 85.—Diagrams showing tilting observed in the Death Valley area. Vectors are used to indicate direction and magnitude of tilting between successive observations. Dates of observations: 1, December 1958; 2, April 1959; 3, October 1959; 4, April 1960; 5, October 1960; 6, February 1961.

The rate at which tilting occurs varies widely. Observations have been made at intervals of from 3 to 8 months, averaging about 6 months, and the total change of attitude, or tilting, that has occurred in the interval is measured. There is no evidence to indicate that the tilting has been at a uniform rate during the interval, nor is there any evidence to show that tilting occurred as a result of a single event or a series of events. It is convenient, however, to use "average" tilting rates when comparing tilting at different stations and intervals. An average rate as low as 0.15 microradian per month at Artists Drive and as high as 12 microradians per month at Aguereberry Point have been observed. Tilting at an average rate of 0.9 microradian per month (11 microradians per year) seems to be common in the Death Valley area.

Tilting at the rate of 11 microradians per year would produce an uplift of 581 feet over a distance of 5 miles in 2,000 years, an amount that is many times too large to be realistic.

During the years 1958–61, both the direction and magnitude of tilting have varied. At Aguereberry Point, Trail Canyon, and Artists Drive, reversals of tilting have occurred and, at the time this was written, the net tilt at some stations after 3 years was less than it was after 1 or 2 years. Observations over a much longer time are needed to establish the trends.

RELATION BETWEEN TILTING MOVEMENTS AND SEISMIC ACTIVITY

Because the Death Valley area is relatively free from earthquakes, it has not been possible to relate tilting to seismic activity in the vicinity of the tiltmeter stations. However, since the first tiltmeter measurements were made in May 1958, there have been five earthquakes of moderate intensity in the eastern Sierra Nevada and Owens Valley areas. These earthquakes, which are listed below, were not felt in Death Valley, although the quake of January 1961 was felt at the Defense mine in the Panamint Range where, according to the caretaker, several large rocks were rolled down the mountain.

Date	*Epicenter*	*Richter magnitude*
Jan. 5, 1959	Southern Owens Valley	4.7
Aug. 4, 1959	Northern Owens Valley	5.5
Jan. 28, 1961	Walker Pass	5.3
Feb. 2, 1961	Sierra Nevada, east of Big Pine	5.0
Oct. 18, 1961	Walker Pass	5.2

There may be some relationship between tilting movements in the Death Valley area and earthquakes about 50 miles away. The average rates of tilting at Aguereberry Point and Trail Canyon fan were greatly increased during the period between October 1960 and February 1961, but there was little change at the other stations. The direction of tilting at two stations, Trail Canyon and Artists Drive, was anomalous during this period, but most of the stations have shown some anomalous tilting in the past.

EARLY PLEISTOCENE STRUCTURAL FEATURES

Deposits of the Funeral Formation are exposed in fault blocks at the foot of the Black Mountains at Mormon Point and Artists Drive, in the Texas Spring syncline and in some of the anticlines northwest of it, in the Park Village fault blocks, Salt Creek Hills, and along Emigrant Wash (pl. 3). At all these places the deposits are mostly fanglomerate, but they include minor amounts of interbedded basaltic lava and volcanic ash. The fanglomerate is sufficiently faulted, folded, and eroded, so that its original fan form is no longer apparent. The structure of the lower Pleistocene fanglomerates is accordant with that of the upper Pleistocene deposits, but the faulting and folding are much greater.

In discussing structural features involving the lower Pleistocene deposits, it is necessary to recall the uncertainties in correlating the deposits mapped as early Pleistocene. Only in two localities is there paleontologic evidence for dating the deposits mapped as Funeral Formation (p. A63).

At Mormon Point the Funeral Formation overlaps and is faulted against the steeply dipping surface on the Precambrian metamorphic rocks (Drewes, 1959). The overlapped surface is a turtleback fault surface, very likely the Amargosa thrust, from which the overlying rocks have been stripped (Noble, 1941; Curry, 1954). Subsequent uplift of the turtleback has raised the fanglomerate a few hundred feet above the valley floor, and the fanglomerate has been faulted valleyward along the contract of overlap. A high-angle fault at the north edge of the Funeral Formation separates it from the younger fill in the saltpan. Displacement on this fault is at least 200 feet (Drewes, 1959). There is difference of opinion about the continuity of the turtleback fault surface with the Amargosa thrust. Noble (1941) and Curry (1954) connect them; Drewes (1959) suggests that the turtleback fault surface is a younger normal fault.

The structural relations at Mormon Point are duplicated along Emigrant Wash where the Funeral Formation overlaps the west-sloping turtleback surface marking the west side of the Tucki Mountain fenster, and the fanglomerate subsequently has been faulted downward along the old fault surface. The dip of the fault is about 25° to the west. Likewise, as at Mormon Point, a second fault, probably a high-angle one, marks the valley edge of the fanglomerate, and the displacement on this fault must also be down towards the valley, that is, to the west. The displacement on this high-angle fault is at least 500 feet.

The faulting that involves the lower Pleistocene deposits at Mormon Point and along Emigrant Wash raises problems about nomenclature of the faults. The autochthonous block at Mormon Point is regarded as that of the Amargosa thrust; the autochthonous block along the turtleback at Emigrant Wash is that of the Tucki Mountain thrust fault (fig. 86). At both locations two generations of movement are recorded. The latest movement, which involves the Funeral Formation, is sufficiently later than the earlier one for the fault to have been folded anticlinally and the upper plate stripped from it, probably by detachment faulting.

On Artists Drive the Funeral Formation overlaps middle and older Tertiary volcanic rocks. At the time this was mapped, in 1957, it was assumed that the contact was simply a depositional one of overlap. Question now arises whether faulting along that overlap contact

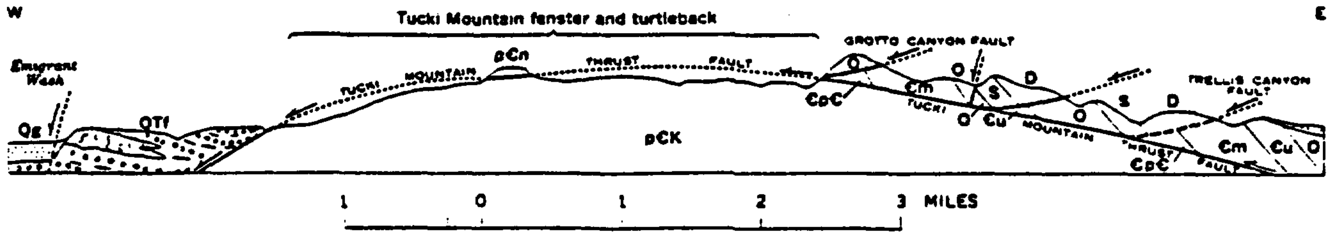

FIGURE 86.—Section of Tucki Mountain showing the Tucki Mountain thrust fault and its branches on the east side of the mountain, the turtleback on the west side, and the Funeral Formation in Emigrant Wash that overlapped the turtleback and that was later faulted down against it. pCk, Kingston Peak(?) Formation; pCn, Noonday(?) Dolomite; CpC, Sterling Quartzite and Lower Cambrian; Cm, Middle Cambrian; Cu, Upper Cambrian; O, Ordovician; S, Silurian; D, Devonian; QTf, Funeral Formation; Qg, upper Pleistocene fan gravel. Vertical scale not exaggerated.

was overlooked. Faulting is suggested by the occurrence in the fanglomerate of granulated Paleozoic quartzite, like the chaotic blocks associated with the Amargosa thrust (Noble, 1941).

Along the west side of the Artists Drive area, at the edge of the saltpan, are four southwest-trending fault blocks of Funeral Formation. The geomorphology of the ridges suggests that they are separated by as many southwest-trending faults and that each fault is downthrown on the northwest side. The two northerly hills are opposite the anticline buried in the saltpan that separates Badwater Basin from the basins to the north (p. A102). The two southerly hills lie just north of the projection of the Mont Blanco fault and between its outcrop on the higher part of Artists Drive and the faults thought to represent its extension in the saltpan. The Mont Blanco fault and its projection in the Artists Drive area marks the southern limit of Funeral Formation exposed in those fault blocks. South of the fault the Funeral Formation either is absent or has been faulted downward and buried under younger valley fill.

North of the Black Mountains the Funeral Formation is downfolded in the Texas Spring sycline (fig. 79) and anticlinally folded over the East Coleman Hills (pl. 3). The structural relief of the Funeral Formation across the Texas Spring syncline and adjoining anticline is at least 750 feet and probably more nearly a thousand. This structural relief is four of five times that of the upper Pleistocene gravels, but probably less than half that of the underlying Furnace Creek Formation (Pliocene).

In the Texas Spring syncline a block of the Funeral Formation ends southeastward at the Mont Blanco fault, which trends northeastward. Displacement on this fault, where it cuts off the fanglomerate, is down to the northwest. This direction of displacement is opposite to the apparent displacement farther southwest along the Mont Blanco fault where it enters the Tertiary formations and seems to have dropped the base of the Furnace Creek Formation downward a thousand feet on the southeast side against the Artists Drive Formation. Perhaps the displacement in the Tertiary formations is that of a tear fault with right-lateral displacement, and the horizontal displacement was taken up in part by downfolding of the Texas Spring syncline.

In the Park Village block, between Cottonball Basin and the Funeral Mountains, the Funeral Formation forms a northwest-trending ridge bounded by faults. This probably is not a horst, however, because the fault along the northeast side of the ridge very probably is a faultline scarp. The displacement there probably is down to the west, that is, towards the valley, and the Funeral Formation probably has been stripped from the structurally higher fault blocks between the Park Village block and the Funeral Mountains. This interpretation is favored because the Tertiary formations along this fault and others paralleling it in the Furnace Creek fault zone are dropped down towards the valley.

There may have been lateral displacement along the Furnace Creek fault zone (Curry, 1938a). There is some evidence of right-lateral displacement along several faults between the Park Village block and the Funeral Mountains, where the beds on the southwest side of the faults, the downthrown side, show southward drag.

Right-lateral displacement also is indicated along faults at the west foot of the Panamint Range (Maxson, 1950, p. 107) and in the Confidence Hills (Noble and Wright, 1954, p. 157).

In the Salt Creek Hills the Funeral Formation is involved in both anticlinal folding and faulting (fig. 75) and has a structural relief measurable in hundreds of feet (p. A103).

Now we attempt to visualize how the physical geography and structure of the Death Valley area might appear if we undo the folding and faulting attributable to middle and late Quaternary time—that is, post-Funeral Formation. The distribution of the lower Pleistocene deposits shows that the Funeral Mountains, Black Mountains, and Tucki Mountain were in existance, but structurally they were at least 1,000 feet lower, and probably more nearly 2,500 feet lower.

If we raise the trough of the Texas Spring syncline (fig. 79) by 2,500 feet—that is, unfold it—dips in the Tertiary rocks in the Black Mountains are reduced almost one half, to an average in the neighborhood of 20°–25°.

Similarly, if the top of Tucki Mountain (fig. 86) were flattened structurally by 2,500 feet, the eastward dip of the Tucki Mountain thrust fault and the westward slope of the turtleback surface would be reduced to about 5°. The westward dip of the faults branching upward from the Tucki Mountain fault would be approximately doubled to an average of about 25°; the average east dip of the faulted Paleozoic formations would be reduced to roughly 25°.

No basis was found for estimating the amount of middle and late Quaternary uplift along most of the Black Mountains, but in view of the regional structural rise southward (p. A71, 88, 99) the probabilities are that the uplift towards the south was as great or greater than it was at the north. We assume the same amount—2,500 feet.

To continue the attempt to visualize conditions as they were at the beginning of the Quaternary, it is necessary to remove something like 3,000 feet of sediment from the structurally deepest part of Badwater Basin and perhaps 2,000 feet from Cottonball Basin.

It seems doubtful that Death Valley ever was 3,000 feet below sea level. On the contrary, it may even have had exterior drainage during part of the late Pleistocene time. Some of the late Pleistocene lakes may have connected with those at Soda Lake (p. A72) and possible exterior drainage from Death Valley is suggested by the abundant large boulders in certain upper Pleistocene (No. 2) gravel deposits (p. A65). Accordingly, it is reasoned that the floor of Death Valley was no lower than sea level, and probably was higher, at the beginning of the Quaternary. The Black Mountains probably were about half as high above the floor of the valley as they are now.

There is reason to infer that, during the Quaternary, the main part of the Panamint Range was raised much more than has been assumed for Tucki Mountain or for the Black Mountains. The Funeral Formation at the west foot of Tucki Mountain (fig. 86) rises southward in 10 miles to an altitude of 7,400 feet. The fanglomerate may have had a source in or beyond what is now Panamint Valley west of the range (Axelrod and Ting, 1960, p. 22). This suggests the possibility that at the beginning of the Quaternary and during early Quaternary time a structural valley and trough extended southward from Mesquite Flat along the site of Emigrant Wash and connected with the north end of Panamint Valley.

Regardless of the continuity of the trough, its altitude probably was not more than half the maximum now reached by the fanglomerate. Whatever that altitude was, the difference between it and the present maximum altitude of the deposit (7,400 ft) is the amount by which the Panamint Range has been raised during middle and late Quaternary time. If this is assumed to be about 3,500 feet, the structures in the Panamint Range should be rotated westward 5°–10° to restore dips as they were when the Funeral Formation was being deposited.

An upland surface of lower relief is represented on the Panamint Range by broad open valleys about 5,000–6,000 feet in altitude (Maxson, 1950, p. 102). The age of the open valleys is uncertain. One of them, Harrisburg Flat, is eroded partly in Funeral Formation; but the open valley could be as old or even older than the fanglomerate, and the subsequent erosion attributable to the fact that the valley provided a local base level for erosion of the fanglomerate. The open valleys contain upper Pleistocene gravels; therefore they must be middle Pleistocene or older erosional features. Most of the differential uplift of the Panamint Range above Death Valley and Panamint Valley has occurred since the broad open valleys were formed.

LATE TERTIARY STRUCTURAL FEATURES

The Furnace Creek Formation records the existence of a playa that extended from Mesquite Flat southeastward across the site of the Salt Creek Hills, Cottonball Basin, Furnace Creek fault zone, and north end of the Black Mountains. Facies changes in the Furnace Creek Formation suggests that the playa did not extend southward along the trough that later became the Death Valley saltpan, because conglomerates in the formation in the northern part of the Black Mountains appear to have had a source far to the northwest in the northern part of the Panamint Range (p. A60). Had there been a playa extending southward along the site of the Death Valley saltpan, the gravels should have moved southward and not across the valley to the east side. The playa lasted long enough to accumulate more than 5,000 feet of fine-grained sediments, mostly derived from erosion of volcanic rocks.

Facies changes within the Furnace Creek Formation are not known well enough to reconstruct closely the limits of the trough that contained the playa. The northeast edge probably was at or close to the Kit Fox Hills fault (pl. 3). Major displacement on that fault occurred after the Miocene(?) formations were deposited in the Kit Fox Hills. It is assumed that the displacement of this fault, down on the southwest side, progressed while the Furnace Creek Formation was being deposited and that the scarp along the fault formed a northeast edge of the playa in which the Furnace Creek Formation was deposited.

The southwest edge of the trough evidently coincided with the flank of Tucki Mountain. On the Black Mountains there was a pile of older volcanic rocks, the Artist Drive Formation, that shed debris northward into the playa, but those mountains were not high enough or dissected deeply enough to contribute Precambrian debris to the Furnace Creek Formation. To what extent the mountains also had started to be raised as a major fault block is uncertain.

Much of the uplift of the Black Mountains occurred after the Furnace Creek Formation was deposited. At the front of the Black Mountains the Furnace Creek Formation is cut off by the fault along that front, one of the Basin and Range type of block faults, and dropped at least 1,000 feet, and perhaps more, into Death Valley.

Structural relief on the Furnace Creek Formation dipping off the north end of the Black Mountains could be as great as 7,500 feet (fig. 79), but this figure probably is excessive. It seems unlikely that the Furnace Creek Formation under the Texas Spring syncline is so thick because the gravity data (pl. 3) indicate that the bedrock floor there is about 5,000 feet deep, and part of this is Pleistocene.

The conflicting data provided by the gravity measurements and by the dips observed in the Furnace Creek Formation can be resolved by assuming that the playa shifted northeastward while the formation was being deposited. Such shift also is suggested by the fact that the sulfate and chloride zones in the upper members of the formation are north of their positions in the lower members (p. A62). If this shift is assumed, beds in the Furnace Creek Formation would offlap northeastward and would not be so thick under the Texas Spring syncline as would be assumed from the thick steeply dipping section exposed on the flank of the Black Mountains.

The structural significance of this interpretation is that much of the uplift of the Black Mountains may have occurred while the Furnace Creek Formation was being deposited; perhaps no more than half, and possibly much less, of the uplift of the Black Mountains need be attributed to early Pleistocene time after the Furnace Creek Formation had been deposited. We favor this interpretation because it resolves an apparent conflict between the structural and gravity data, and because it accords with an inferred northward shift of the salt zones of the Furnace Creek Formation.

At the Salt Creek Hills the Furnace Creek Formation is exposed in a sharply folded and faulted dome. The fold is asymmetrical. Its southwest flank is almost vertical where it is overlapped by the Pliocene and lower Pleistocene(?) Funeral Formation; the northeast flank dips 20°–45° northeastward. About 3,500 feet of beds assigned to the formation is exposed in the dome, and its indicated structural relief would be at least that much.

Outcrops of upper Tertiary deposits northwest of the Salt Creek Hills (pl. 3) suggest that a major fault trends northwestward for at least 6 miles. Southeastward the fault probably underlies the ridge of lower Pleistocene gravel that extends into the saltpan south of Salt Creek. Projected across Cottonball Basin the fault would join with the frontal fault of the Black Mountains where that fault turns northwestward into Cottonball Basin. Very likely this is one of the faults through which ground water is discharged from Mesquite Flat to maintain the extensive marsh on the west side of Cottonball Basin (Hunt and Robinson, 1960; see also Hunt and others, 1965).

At the Salt Creek Hills, as elsewhere, the Furnace Creek Formation is more strongly folded than the lower Pleistocene fanglomerate, which, in turn, is more folded than the upper Pleistocene gravels.

Reference has already been made to the curious Mont Blanco fault that trends northeast across the north end of the Black Mountains and Texas Spring syncline (p. A103). In the Black Mountains the base of the Furnace Creek Formation is offset to the right, as if the displacement were that of a normal fault down on the southeast side. Yet displacement of the Pleistocene beds in the Texas Spring syncline is down on the northwest.

Many assumptions have had to be made as to conditions at the time the Furnace Creek Formation was deposited, but it is inferred that the playa represented by it occupied a long, narrow, southeast-trending structural trough—at least 40 miles long and no more than 5 miles wide. The trough may have branched southward along the site of the Death Valley saltpan, but this branch was higher than the playa; its subsidence to a level below that of the southeast-trending playa could have occurred late in Pliocene time, but it more probably occurred early in Pleistocene time.

It has been inferred that the Black Mountains were raised structurally about 4,000 feet while the Furnace Creek Formation was being deposited, and that another 3,500 feet of uplift occurred during early Pleistocene time.

To visualize conditions at the beginning of the Pliocene, it is necessary to imagine the valley fill without the Furnace Creek Formation and younger sediments. The thickness of these, as suggested by the gravity data, is less than the apparent thickness across the steeply dipping beds, and is assumed to be about 3,500–4,000 feet. To the degree that the Furnace Creek Formation in the valley is underlain by older Tertiary deposits, the thickness assumed for the formation there would have to be even less. Because the amount of uplift inferred at the Black Mountains is about equal to the thickness of sediments under the Texas Spring syncline, the relief across that part of the area probably was little different from what it is at present.

There is no stratigraphic or structural evidence to suggest how the Panamint Range changed during this period. At least the northern part of the range was in existence because it provided some of the conglomerates in the Furnace Creek Formation. The part south of Tucki Mountain may have been much lower. It was certainly partly buried, and may have been largely buried, under the lower Tertiary volcanics that overlap the east side of the range.

MIOCENE(?) AND EARLY TERTIARY STRUCTURAL FEATURES

Deposits mapped as Miocene(?) are extensive in the Kit Fox Hills, east of Mesquite Flat, and they extend from there southeastward to the foot of the Funeral Mountains. The Kit Fox Hills fault, a straight high-angle fault and part of the northwest-trending Furnace Creek fault zone, extends for at least 30 miles along the southwest edge of the deposits. Displacement is down on the southwest side, and the gravity data indicate that displacement totals several thousand feet. Probably the Pliocene deposits southwest of the fault never extended northward across it (p. A117), but if they did, they have been stripped from the upthrown block. Part of the displacement on this fault is Quaternary in age; much of the displacement is Pliocene, and most of it may antedate the Furnace Creek Formation.

Along the northeast side of the Park Village block the Kit Fox Hills fault may be marked by a faultline scarp, as already noted (p. A115).

Between the Kit Fox Hills and the Funeral Mountains are hills of Titus Canyon(?) Formation of Stock and Bode (1935) (Oligocene) that protrude through the fan gravels. The northeast side of the Kit Fox Hills probably is a faultline scarp like that along the east side of the Park Village block. The scarp along the northeast side of the Kit Fox Hills faces northeast and overlooks older formations; it seems too straight to be simply a cuesta at the updip side of Miocene(?) fanglomerate. This fault may be older than the Kit Fox Hills fault.

Along the Keane Wonder fault at the foot of the Funeral Mountains, the Titus Canyon(?) Formation is faulted down against the Precambrian in the mountains. The fault dips 25°–40° towards the valley, about the same as the dip of the Precambrian beds in that part part of the Funeral Mountains, which form a dissected turtleback (fig. 118).

The occurrence of the Titus Canyon(?) Formation along this fault and turtleback surface is like that of the Funeral Formation at the Tucki Mountain turtleback in Emigrant Wash and at the Amargosa thrust fault and turtleback surface at Mormon Point (p. A114). These relations also are duplicated at the Copper Canyon turtleback (Drewes, 1959) (fig. 87). The Keane Wonder fault and the turtleback surface are continuous with the thrust faults at Boundary Canyon and Echo Mountain, along which Cambrian formations are thrust westward onto the Precambrian. The Titus Canyon(?) Formation is faulted against the turtleback surface of the Funeral Mountains; but the main thrust appears to be pre-Titus Canyon in age, because to the north that formation overlies the upper plate. The similarity of this structure to those where the Funeral Formation has overlapped and has been faulted against turtleback surfaces suggests similar histories. If so, the main thrust is old enough to have had the upper plate stripped from the turtleback surface before the Titus Canyon(?) Formation was deposited and faulted against it.

At the north foot of Nevares Peak, beds mapped as Titus Canyon(?) Formation are dragged upward along a smooth fault surface dipping northward off the Cambrian rocks that form that mountain. This feature

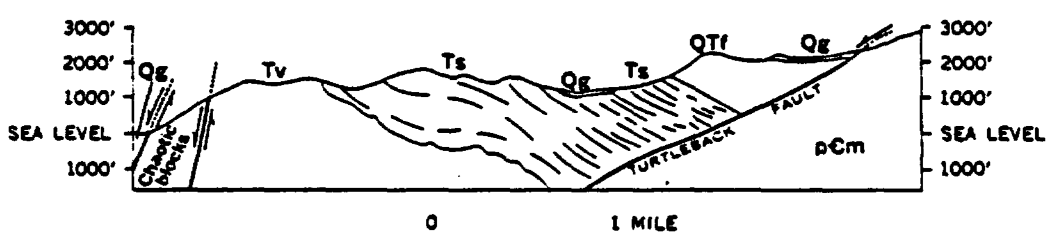

FIGURE 87.—Section of the Copper Canyon turtleback fault (generalized from Drewes, 1959). pCm, Precambrian metamorphic rocks; Tv, Tertiary volcanic rocks; Ts, interbedded Tertiary sedimentary rocks and basalt; QTf, Funeral Formation; Qg, upper Pleistocene gravel.

shows well on the State geologic map of Death Valley (Jennings, 1958).

Mapping of the Miocene(?) formations between Cottonball Basin and the Funeral Mountains suggests that they aggregate more than 4,000 feet thick. But considering that there must be some Titus Canyon(?) Formation beneath these beds, the thickness seems to be excessive. The gravity data suggest that bedrock is no more than about 4,000 feet deep in that area. The conflicting data can be reconciled by assuming either that the Titus Canyon(?) Formation is denser than was assumed in compiling the gravity data (p. A107) or by assuming an offlap relation in the Tertiary formations valleyward from the Funeral Mountains, as has been inferred for the Furnace Creek Formation northward from the Black Mountains. The latter interpretation seems reasonable because the outcrops of Titus Canyon(?) Formation north of the Kit Fox Hills probably never were buried under Miocene(?) fanglomerate as thick as that exposed in the Kit Fox Hills.

In the northern part of the Black Mountains, the formation at Artists Drive dips northeastward under the Furnace Creek Formation. About 6,000 feet of beds are exposed, mostly volcanics, but these grade northward to and intertongue with playa deposits. The formation must thin northward under the Furnace Creek Formation, for the same reason that the Furnace Creek Formation must thin northward under the Texas Spring syncline. Gravity data indicate a total of about 5,000 feet of Quaternary and Tertiary deposits down to the bedrock under the syncline. Figure 88 is a diagrammatic section illustrating the probable thinning northeastward of the Tertiary and Quaternary formations on the northeast flank of the Black Mountains.

Very likely the bedrock floor under the Tertiary formations is a series of fault blocks, but these cannot be satisfactorily reconstructed. While the formation at Artists Drive was being deposited, the axis of the trough was near A, B and C (fig. 88). At the site of the axis of the Texas Spring syncline, dips were towards the Black Mountains.

At the front of the Black Mountains the formation at Artists Drive is faulted down about 5,000 feet. The blocks faulted down on the Death Valley side are referred to as the Artists Drive fault blocks. The fault is the frontal fault of the Black Mountains, and it continues northward and cuts off the Furnace Creek Formation at the front of the north end of the Black Mountains. The Furnace Creek Formation is displaced at least 1,000 feet by the fault. If there was any displacement on this fault before the Furnace Creek Formation was deposited, the fault remained inactive while the formation was being deposited, because the facies

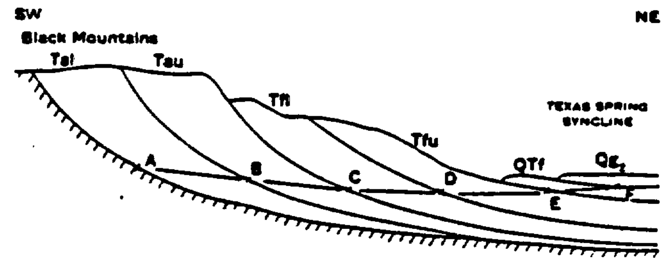

FIGURE 88.—Diagrammatic section illustrating probable thinning and offlap of Tertiary and Quaternary formations from the Black Mountains northeastward to the Texas Spring syncline. Length of section about 6 miles. Axes of the troughs in which the formations were deposited shifted progressively northeastward from A to F. Tal, Tau, lower and upper members of the Artist Drive Formation; Tfl, Tfu, lower and upper members of Furnace Creek Formation; QTf, Funeral Formation; Qg, No. 2 gravel.

changes in the formation are cut off by the fault and are not related to it. The probabilities are that the frontal fault along the Black Mountains is a Quaternary structure. Block faulting that outlined the present basins and range elsewhere in the region is regarded as largely of Quaternary age (Gilbert, 1941; Axelrod and Ting, 1960).

On the other hand, there was earlier faulting along the northwest-trending Furnace Creek fault zone and along other northwest-trending faults in the region. Near Leadfield, in the Grapevine Mountains north of this area, the Titus Canyon Formation overlaps an earlier northwest-trending high-angle fault (James Gilluly, oral commun. 1960). The block faulting therefore appears to have started as early as Oligocene time. The faulting has continued to the present, but since late Pliocene time the dominant trend of the faults has changed from northwesterly to northerly.

If the north-south faults are mostly Quaternary, the uplift of the Black Mountains that caused the northward offlap and thinning of the formation at Artists Drive and the Furnace Creek Formation is attributed to folding rather than faulting.

The formation at Artists Drive above the Artists Drive fault blocks also is broken by some flat faults that dip only 20°–40° towards Death Valley (fig. 79). Displacements are at least as great as 2,500 feet and in places may be very much greater. These structures were examined only in reconnaissance; they could be dismissed as megabreccia, except for the fact that in that area there are large chaotic blocks of Paleozoic rocks, hundreds of feet in diameter, and the ensemble suggests the possibility that major thrusting, like that of the Amargosa thrust farther south (Noble, 1941), has been overlooked in this part of the Black Mountains.

One of the blocks of Paleozoic rock is dolomite wedged into the Tertiary formations at the frontal fault. Another is granulated quartzite, a monolithologic breccia, under (or in?) the Funeral Formation at the northwest end of the Artists Drive fault blocks. The chaotic blocks probably are Cambrian—from the Zabriskie Quartzite and Bonanza King Formation. The blocks are suggestive of the Paleozoic blocks that are mixed with Tertiary ones and associated with the Amargosa thrust. There is no nearby source, and they must have been brought a long distance. The flat faults in the formation at Artists Drive may be branches from a larger fault that has not yet been identified.

Volcanic rocks along the east foot of the Panamint Range in part at least are Miocene (p. A120), and are similar to and probably correlate with parts of the formation at Artists Drive. The eruptives are tilted 20° towards the east and overlap more steeply dipping Paleozoic formations (fig. 89), showing that about half the eastward tilt of the Panamint Range has occurred since those eruptions.

GRANITIC INTRUSIONS

Two large granitic intrusions, probably Cretaceous or early Tertiary, are well exposed at the west edge of the Death Valley area and are shown on the geologic map (pl. 1). One of these, the granite at Skidoo, crops out in an area of about 12 square miles along the north and east sides of Harrisburg Flat and extends southeastward into the head of Trial Canyon. The other, the granite at Hanaupah Canyon, occupies an equal area at the head of Hanaupah and Starvation Canyons. Smaller outcrops of granitic intrusions in the Panamint Range are in Wildrose Canyon about midway between the granites at Skidoo and at Hanaupah Canyon, in Warm Spring Canyon about 15 miles south of Hanaupah Canyon, and along the east foot of the range.

Although labeled on the map as granitic intrusions, and loosely called granite or granitic rocks, most of the intrusions would more nearly be classed as quartz monzonite porphyry. However, the field study was interrupted before chemical analyses were made, and not enough thin sections were examined to be certain of the range in composition of the facies of the intrusions. Under the circumstances, a more precise nomenclature for the few rocks studies would be misleading, and we use the term "granitic intrusion" throughout.

The granites at Skidoo and at Hanaupah Canyon differ texturally and, to a minor degree, mineralogically. The granite at Hanaupah Canyon is porphyritic with distinct fluidal structures; the dark phenocrysts are biotite and hornblende. The granite at Skidoo is mostly gneissic, but a part of it is porphyritic and has fluidal structures; the dark phenocrysts are biotite. The kind and amounts of trace elements in the two intrusions and the trace elements that are in quantities too small to determine are similar, as brought out in table 25. The apparent greater concentration of lead and zinc in the granite at Hanaupah Canyon and the sill east of it, as compared to the granite at Skidoo, may be real, but additional analyses of the rocks are needed.

The three-dimensional form of intrusions probably is the most important criterion for determining the relationship between the intrusions and the structural geology of the region; this relationship, in turn, is basic for understanding the origin of the intrusions. Yet the form must remain a matter of considerable conjecture. We have attempted to apply some of the principles that have been learned about intrusive forms on the Colorado Plateau, where complexities are at a minimum and where exposures are far more complete. As G. K. Gilbert wrote (1876) in citing the Colorado Plateau as a field for geologic study, "with the facts of structure conspicuous and beyond question, the mind is left free to search for causes."

Before attempting to compare these intrusions with those on the Colorado Plateau, which is remote and structurally very different, some similarities may be noted between the intrusions of the Death Valley area and the individual plutons of the Sierra Nevada batho-

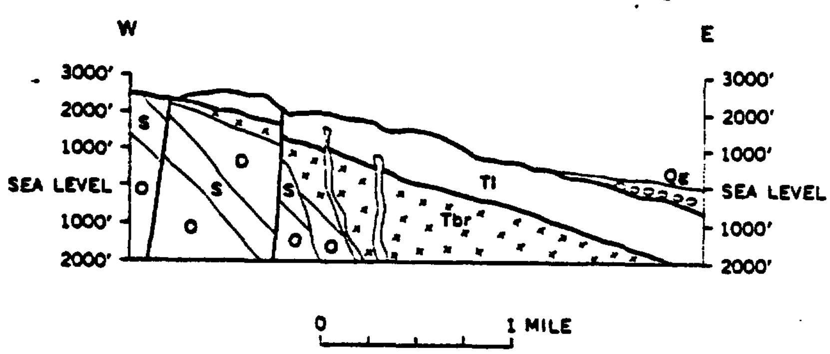

FIGURE 89.—Section along the ridge 3 miles north of Trail Canyon. Felsitic lavas dipping 20° east overlap more steeply dipping Paleozoic formations. About half the eastward tilt of the Panamint Range has occurred since the lavas were erupted. O, Ordovician; S, Silurian; D, Devonian; Tbr, Tertiary volcanic breccias, in part intrusive; Tl, felsitic lavas of Tertiary age; Qg, Quaternary gravel.

TABLE 25.—*Trace elements in the granites at Skidoo and at Hanaupah Canyon*

[Cu, Pb, and Zn determined by colorimetric methods by H. L. Neiman; other determinations are semiquantitative spectrographic analyses by E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

Element	At Skidoo, gneissic facies	At Hanaupah Canyon	
		Boulder from central intrusion	Sill along fault east of the canyon
Cu	4	4	6
Pb	26	80	280
Zn	50	90	80
Mn	300	1,000	700
Zr	100	300	500
Ni	15	15	10
Co	5	<5	7
V	2	20	70
Mo	<5	7	5
Y	15	30	30
Be	1.5	1.5	<1
Ti	2,000	2,000	5,000
B	20	10	30
La	50	100	150
Sc	<10	<10	10
Cr	30	30	20
Ba	500	500	1,500
Sr	300	200	1,000
Ag	<1	<1	<1
Mg	0.7	0.7	1

NOTE.—All samples contained: As >1,000; Sn >10; Ge >20; Ga >20; Cd >50; Bi >10; In >10; Sb >200; Tl >100; Nb >50; Ta >50; W 100.

lith. In the batholith the earliest intrusions were concordant and followed stratigraphic or tectonic boundaries (Cloos, 1932). Where the intrusions are crowded together, this concordance is masked because the later intrusions became guided by the walls or other structures of the earlier ones. There is abundant evidence that the intrusions emplaced themselves more by physical injection than by replacement, assimilation, or stoping (Knopf, 1929; Calkins in Matthes, 1930; Cloos, 1932, 1933, 1935, 1936; Bateman, 1958). The batholith plunges northward, and the intrusions at the north end are separated from one another like those to the east in the Death Valley area. The structural settings of the Sierra and Death Valley areas are quite different. Seismic and gravity data indicate that dense rock is shallow under the intrusions just west of Death Valley, but the main part of the batholith, including its north end, is underlain by a considerable thickness of light rock. But even where the intrusions of the batholith are crowded, there has been structural doming.

The earliest intrusions of the Sierra Nevada batholith are along its west edge and are thought to be Late Jurassic or Early Cretaceous (Knopf, 1929, p. 9). Eastward across the batholith the individual plutons are younger and generally less mafic (Calkins in Matthes, 1930; Cloos, 1936, p. 431–434; compare with Bateman, 1961, fig. 5). The intrusions in the Death Valley area may extend this pattern to the east. These age relationships have long been recognized, and the problems they pose have been well stated by Ferguson (1929, p. 118):

It may be that the locus of intrusion moved gradually eastward from the Sierra Nevada to the Rocky Mountain region and that the areas of granitic rocks, intermediate in position between the Sierra Nevada batholiths and the Tertiary batholiths to the east are also intermediate in time (Lindgren, 1915, p. 260), or there may have been two distinct and sharply separated episodes of granitic intrusion.

Whether the Death Valley intrusions should be regarded as part of the composite Sierra Nevada batholith depends on one's definition of the batholith and on assumptions about the form of the buried parts of the intrusions. But regardless of whether the definition, which perforce must be arbitrary, includes or excludes them, clearly these intrusions are genetically related to each other and to the batholith. In the Death Valley area the granitic intrusions may have reached almost to the surface and may have developed into volcanic rocks at the surface. These volcanic rocks are interbedded with playa and related deposits ranging from Oligocene to Pliocene.

The interpretation presented in this report is that the granitic intrusions in the Death Valley area are related to, but younger than, the batholith to the west and that they are related to, but older than, the volcanism. The granitic intrusions in the Panamint Range are probably Miocene (p. A50) and are interpreted as having immediately preceded the volcanism which continued long thereafter. The intrusions seem to have spread laterally along the Amargosa thrust fault and its branches and to have domed the overlying rocks (figs. 90, 91, 94, 108) because—

1. The intrusions have concordant roofs. There is discordance to be sure, but the degree of discordance is no greater than the discordance of laccoliths at their type area, the Henry Mountains, Utah, where structural relationships are much simpler than in the Death Valley area.
2. The intrusions spread in incompetent formations where they are overlain by competent ones, a favored mechanism of intrusions that are demonstrably floored.
3. Because there is little evidence of reaction along the contacts and little other evidence of assimilation or replacement, the intrusions must have emplaced themselves by physical injection. Despite the large size of the intrusions, hydrothermal effects are little, if any, greater than around intrusions in the Colorado Plateau.
4. The volumes of the structural domes over the intrusons are too small to accommodate physically in-

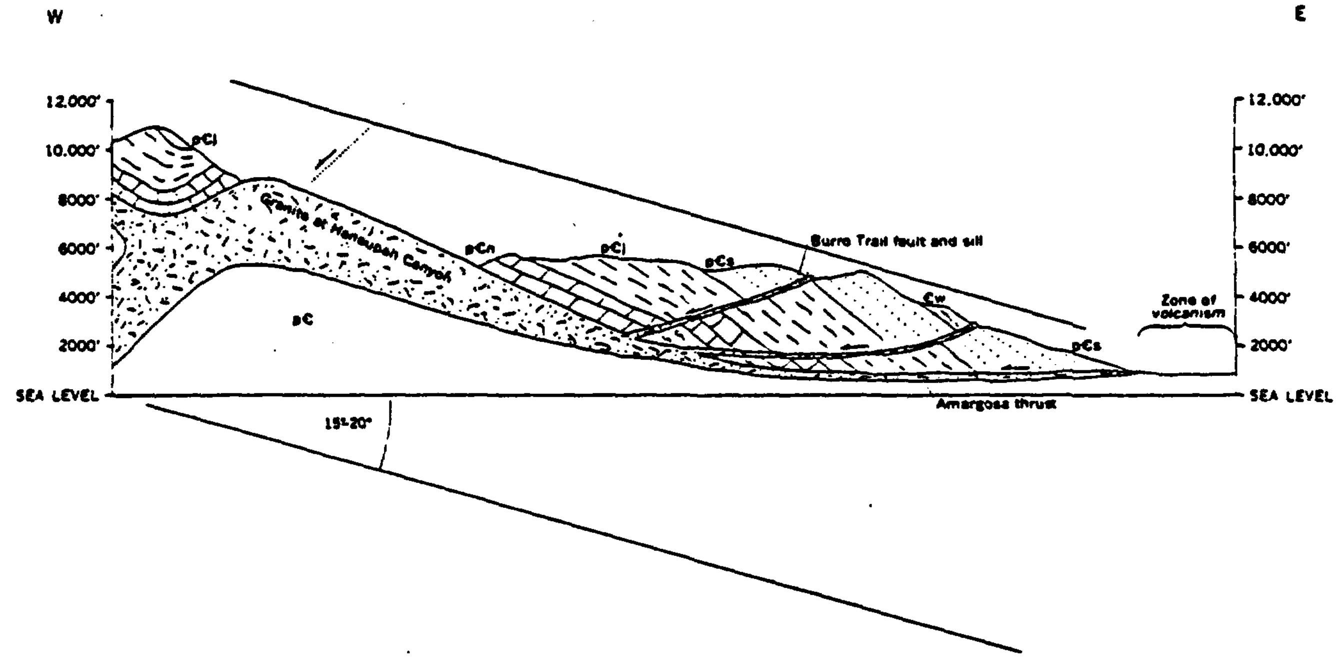

FIGURE 90.—Cross section of the granite at Hanaupah Canyon. Length of section 8½ miles; vertical scale not exaggerated. pC, Precambrian metamorphic rocks; pCn, Noonday(?) Dolomite; pCj, Johnnie Formation; pCs, Stirling Quartzite; Cw, Wood Canyon Formation. Since the faulting and intrusion of the granite, the Panamint Range has been tilted 15°–20° east.

jected crosscutting intrusions as wide as these (see Hunt and others, 1953, p. 139).

5. Sills that extend from the granite at Hanaupah Canyon along thrust faults, like the Burro Trail fault and others north of the intrusion (pl. 3), show that the granite is later than the thrust faulting and that part of the granite demonstrably spread laterally along the faults.

Except for their large size, the intrusions are much like those on the Colorado Plateau. They are not like the individual intrusions in the Sierra Nevada batholith where, although the plutons seemingly were forcibly injected, there is abundant evidence of steep contacts (Bateman, 1958; Sherlock and Hamilton, 1958). Nor is there evidence of much assimilation or stoping like that described at some intrusions in the Mojave Desert (McCulloh, 1954, p. 21).

Intrusions in the Great Basin probably are floored to a far greater degree than has generally been assumed. Few workers have considered this problem. Noble (1941, p. 963) inferred that the granite under the Amargosa thrust fault in the Virgin Spring area was controlled by the fault because the roof of the granite parallels the fault (fig. 108). Hewett (1956, p. 63) interpreted the Teutonia Quartz Monzonite in the Ivanpah quadrangle, 75 miles southeast of Death Valley, as having been intruded along one of the eastward directed thrust faults. He states (p. 63), "There is nothing about the relations of the monzonite to the limiting rocks in any part of the region to indicate that any large part of it is a crosscutting stocklike mass."

Granitic intrusions in the Mina quadrangle in western Nevada have been interpreted as having spread laterally at the thrust faults (Ferguson and others, 1954), and one at the south end of the Panamint Range, in Warm Spring Canyon, is intrusive along a fault (Wasserburg and others, 1959). Granitic intrusions in the southern part of the Panamint Range have emplaced themselves along faults (Johnson, 1957, fig. 2), and some in the Silurian Hills were first localized by the thrusting and later displaced by it (Kupfer, 1960). Intrusions in the Darwin area also have the flat form (W. E. Hall, written commun., 1964).

Mackin (1947, 1954) has shown that doming in the Iron Springs district in southwestern Utah is quite like that on the Colorado Plateau. He has been the principal advocate (1960) of the idea that igneous intrusions have been underestimated as a cause of many of the structural features in the Great Basin.

But such interpretations are exceptional for the Great Basin. In most reports the form of the intrusion either is ignored by drawing cross sections away from them, or the cross sections are drawn to indicate that the intrusions widen downward. A case can be made for this interpretation by assuming that the Great Basin is underlain by a granitic batholith and that the stocks are minor apophyses rising upward from it. In favor of such interpretation is seismic refraction evidence suggesting that the crust under the Great Basin consists of 2 layers, the upper one having a velocity of 6.3 kmps or less, and the lower one having a velocity of 7.6–7.8 kmps (Press, 1960; Berg and others, 1960; Diment and others, 1961). Such interpretation, however, is not presented, and one gains the impression that the stocks were drawn in cross sections to widen downward chiefly to conceal and dispose of complex structures at depth. A bad effect of this practice, however, is that there has been too little thought given to the three-dimensional forms of the intrusions, the mechanics of how they became emplaced, and their part in the structural history of this complex region.

The granites at Skidoo and at Hanaupah Canyon have domed roofs of Precambrian sedimentary formations that are roughly concordant with the upper surfaces of the granites. The degree of discordance is little greater than over laccoliths in the Henry Mountains where, despite the simplicity of the structure of the host rocks, even the most orderly laccoliths cut across several hundred feet of beds in a mile. Extensive concordant roofs, like those on the granites at Skidoo and at Hanaupah Canyon, imply equally concordant floors.

Moreover, the area domed by the granites at Skidoo and at Hanaupah Canyon is too small and has too little structural relief to be attributable to doming by steep-walled crosscutting intrusions having cross sections as wide as the exposed granites. On the Colorado Plateau, stocks 1 mile in diameter produce domes having a base 6 miles in diameter and a structural height of almost 1½ miles (Hunt and others, 1953, p. 139). The intrusions in the Panamint Range are much wider than those in the Colorado Plateau, but the doming is not correspondingly greater. This fact also suggests that the granitic masses have spread laterally and that the steep-walled stocklike source for them is much smaller than the area of granite at and near the surface.

Finally, the roof and side contacts of the intrusions are sharp and contact metamorphism is slight, both of which indicate that there was no great amount of reaction that would cause replacement or assimilation of the country rock by the granite. Indeed, the contact metamorphism is more like that around the floored intrusions than around the stocks on the Colorado Plateau.

The occurrence of a pyritic and an epidotic zone of alteration over the west side of the granite at Hanaupah Canyon is the basis for inferring that the source of that intrusion is under its west side (fig. 90). Pyritic alteration and considerable discordance along the west side of the granite at Skidoo suggest that its source is under its west side. The eastern edge of the granite at Skidoo is inferred to be along the north-trending monocline that forms the head of Tucki and Blackwater Washes. The intrusions therefore are interpreted to be wedge shaped, the form to which the names sphenolith (Burckhardt, 1906) and harpolith (Cloos, 1921) have been applied.

GRANITE AT SKIDOO

The granite at Skidoo (an abandoned mining camp) crops out in an area about 12 miles long and 1–5 miles wide, elongated north-northwestward parallel to the general strike of the enclosing formations. The age is presumably Cretaceous or Tertiary. The southwest contact is steep and crosscutting; but the roof is concordant, east dipping, and mostly in the Kingston Peak(?) Formation although cutting upward discordantly to the Noonday(?) Dolomite. The structural form (fig. 91) is broadly wedge shaped thinning eastward.

The contact along which the intrusion has spread almost certainly is a thrust fault of the Amargosa fault system. The contact between the Kingston Peak(?) Formation and Noonday(?) Dolomite is a fault; so also is the contact between the Noonday(?) Dolomite and Johnnie Formation. The spreading of the granite has obscured evidence for faulting at the plane of intrusion, and not enough is known about the local stratigraphy of the Kingston Peak(?) Formation and Noonday(?) Dolomite to estimate the thickness of beds cut out by faulting at any particular place—although locally more than a thousand feet of beds have been cut out. The amount of lateral displacement along the fault is uncertain.

The granite at Skidoo has spread in the Kingston Peak(?) Formation, which is an incompetent unit, and under the Noonday(?) Dolomite, which is a competent unit. This relationship duplicates that on the Colorado Plateau where the favored horizons for spreading of floored intrusions are in the upper part of incompetent formations that are overlain by competent ones (Hunt and others, 1953, p. 142).

The roof of the granite at Skidoo is well exposed almost continuously from the head of Trail Canyon to a little beyond the abandoned mining camp at Skidoo. In Trail Canyon the roof is Noonday(?) Dolomite. This roof rises steeply westward to the summit where it flattens. Along the summit from Trail Canyon northward along the east side of Harrisburg Flat, the roof of the granite and the overlying dolomite and limestone dip gently 5°–10° E. The steep dip noted in Trail Canyon, however, extends northward as a monocline 1½ miles east of the summit, and this fold probably marks the eastern limit of the granite. This interpretation is based partly on the structure (fig. 91) and partly on the occurrence of small intrusions of alaskite along the monocline. The granite, therefore,

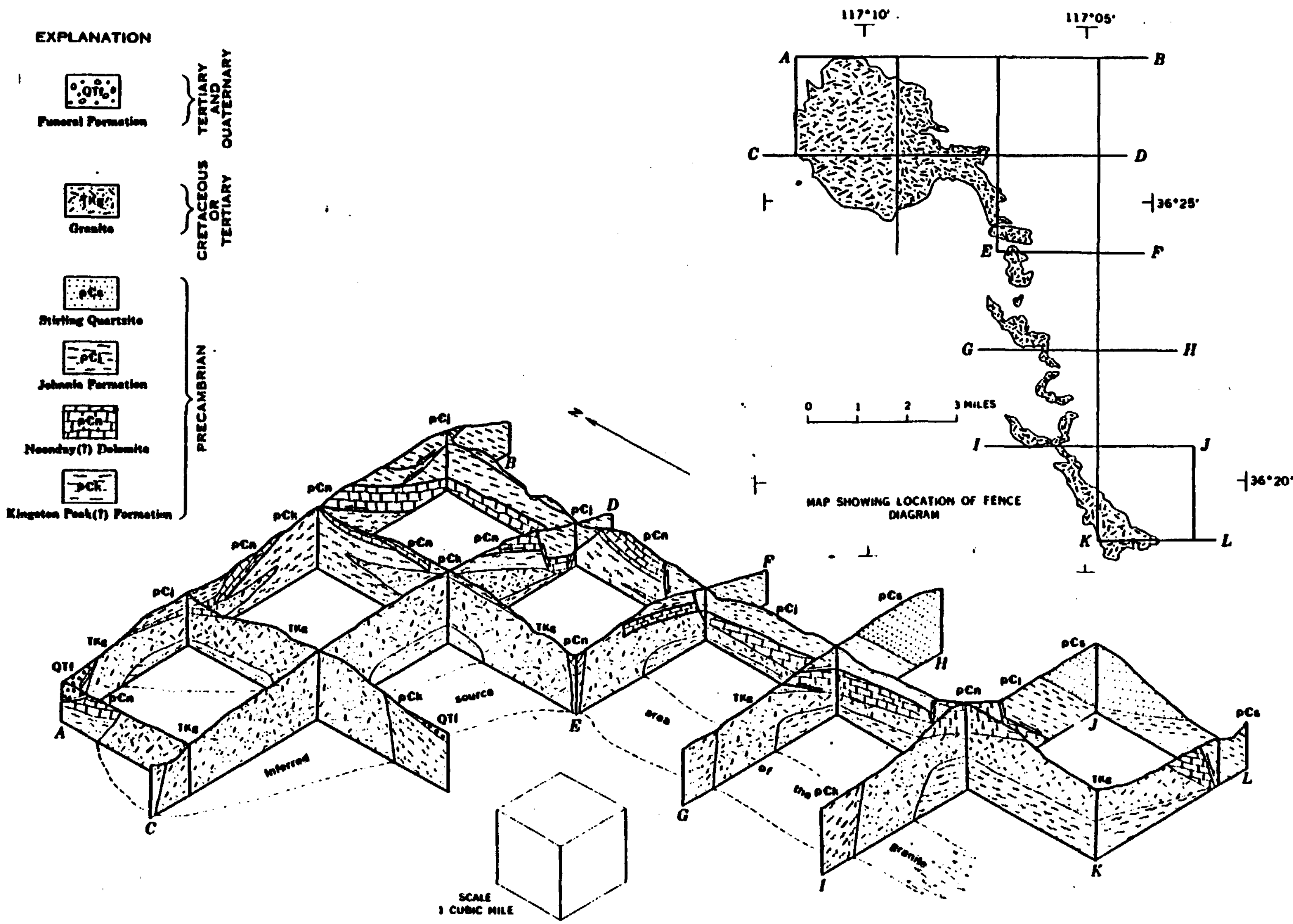

FIGURE 91.—Fence diagram, isometric projection, illustrating inferred shape of the granite at Skidoo.

is inferred to have a maximum thickness about equal to the structural relief along the monocline, that is, about 3,500 feet.

At the head of Trail Canyon the Noonday(?) Dolomite above the granite is about 800 feet thick. Within a mile and a half to the north, however, the dolomite has thinned to about 200 feet, probably by faulting, and about 200 feet of shaly beds of the Kingston Peak(?) Formation lies between the dolomite and the top of the granite. Farther north the dolomite again thickens to about 1,000 feet; it overlies the granite at 2 places, but 2 miles southeast of Skidoo, 500 feet of shale, quartzite, and stretched-pebble conglomerate belonging to the Kingston Peak(?) Formation lies under the dolomite and above the granite.

The steep west wall of the granite cuts across faulted and folded beds, mostly belonging to the Kingston Peak(?) Formation. At 2 places about 3 miles southeast of Skidoo the west wall is dolomite, but whether this dolomite is part of the Noonday(?) Dolomite or a part of the Kingston Peak(?) Formation was not determined.

In the vicinity of Skidoo the granite is in the Kingston Peak(?) Formation, and the roof plunges northwestward.

Westward from the north end of Harrisburg Flat and northward along the west edge of the intrusion, Pliocene and lower Pleistocene(?) Funeral Formation has been faulted against the granite. This roof of the granite is sheeted roughly parallel to the faulted contact along Emigrant Canyon; the sheeting and the roof contact strike north and dip about 25° W. Steeply dipping fissures in the granite also strike north. Dikes of basalt intrude both the sheeted joints and the steeply dipping fissures. Along the canyon there are at least 5 dikes in the upper 250 feet of the granite; their thicknesses range from 6 inches to 10 feet. These dikes are not related to the granite. Probably they are early Pleistocene in age and related to the basaltic lavas and minor intrusions that occur nearby in the Funeral Formation.

The southern third of the granite at Skidoo is porphyritic and clearly an intrusive igneous rock; but the northern two-thirds of the intrusion is banded gneiss, and many outcrops there look like metamorphic Precambrian rocks. Along the monocline east of the outcrop of the granite are small intrusions of alaskite (fig. 92).

The contacts along both the roof and west wall of the granite at Skidoo are sharp. In the porphyritic facies dike-like and sill-like apophyses of granite extend upward into the fractured roof rocks. The contact zone commonly is a few inches wide (fig. 93), about twice as wide but otherwise similar to roof contacts over laccoliths in the Henry Mountains on the Colorado Plateau. Four or five inches from the contact is normal porphyry with phenocrysts as much as 1 inch in diameter. Nearer the contact is a zone in which the crystals

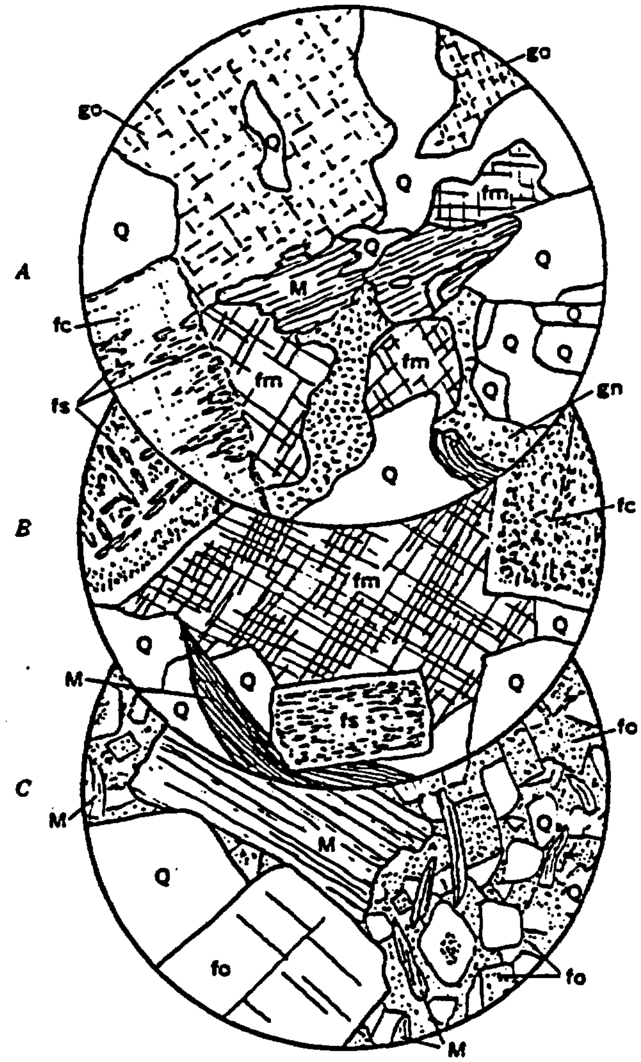

FIGURE 92.—Micrographs of thin sections of the granite at Skidoo. q, quartz; fo, orthoclase; fm microcline; fs, sericitized feldspar; fc, argillized feldspar; go, groundmass oriented; gn, groundmass not oriented. Diameter of field, 2.5 mm. A, Gneissic facies. Quartz with strain shadows (30 percent); euhedral feldspars (20 percent) altered to sericite and to clay; anhedral microcline (20 percent); muscovite (10 percent) with ragged sides and ends and embayed with quartz; and a feldspathic groundmass (20 percent) with sericite, some of it oriented and some not. B, Porphyritic facies. Euhedral plagioclase and orthoclase (30 percent) altered to sericite and to clay. These have clear rims of low index feldspar and are set in microcline (about 30 percent) which occurs in irregular growths. Other common minerals are quartz (about 30 percent) and biotite (about 10 percent). There is also a trace of augite (not shown). C, Alaskite facies. Consists of quartz (30 percent); clouded feldspar, probably mostly orthoclase (60 percent); sericite (10 percent) and occasional phenocryst of muscovite (M). Both the quartz and feldspar occur as euhedral crystals in a paste of anhedral quartz and feldspar. A few aggregates of phenocrysts of the same minerals are rounded and may have been floated in from porphyry facies from which this is believed to have been derived.

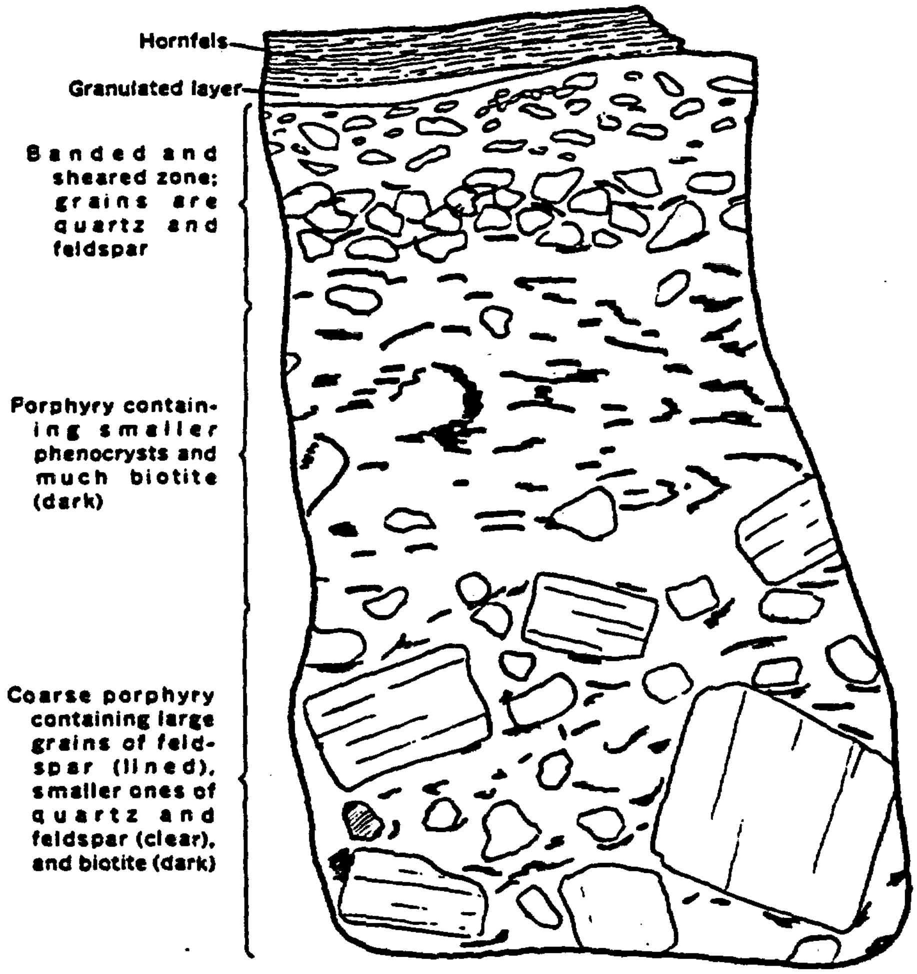

FIGURE 93.—Hand specimen of roof contact of the porphyry facies of the granite at Skidoo. Natural size.

are small and not very distinct in hand specimens. A selvage 1 or 2 inches wide at the contact is sheared and banded. At the contact is a granulated layer 2 or 3 mm wide.

Contact-metamorphic effects are surprisingly slight. Shale and schist are baked to hornfels, and carbonate rocks are bleached—effects that are only a little more intensive and extensive than the alteration zones above the much smaller floored intrusions on the Colorado Plateau.

The gneissic facies may be an ancient granite. There is much brecciation, and cataclastic structures are strongly developed along the roof contact. Quartz veins in the gneissic facies are numerous, and some, like those at Skidoo, have produced gold.

GRANITE AT HANAUPAH CANYON

A granitic intrusion at the head of Hanaupah Canyon crops out in an oval area about 6 miles long and 3 miles wide extending from Hanaupah Canyon to Starvation Canyon. The roof is highly domed, much more sharply domed than that of the granite at Skidoo. Across th north end of the intrusion and along the northwest sid of the roof is Noonday(?) Dolomite. Along the south east side, which, however, was examined only from th air (p. A8), the granite may cut discordantly upwar to the Johnnie Formation.

The structural relief across the top of this intrusio is at least 6,000 feet, but half of this can be attribute to the homoclinal eastward dip of the country rock The doming, superimposed on the homoclinal eastwar dip, amounts to about 3,000 feet and could be produce by a partly floored intrusion about that thick (fig. 94

The only part of the granite at Hanaupah Canyc that was examined on the ground is the northern ar northwestern contact from 4,000 feet altitude in Ha aupah Canyon to the summit at 10,000 feet. Along th contact the Noonday(?) Dolomite is bleached and h been dragged steeply upward; but the drag is not steep as the side wall of the granite, which in plac cuts upward to the Johnnie Formation.

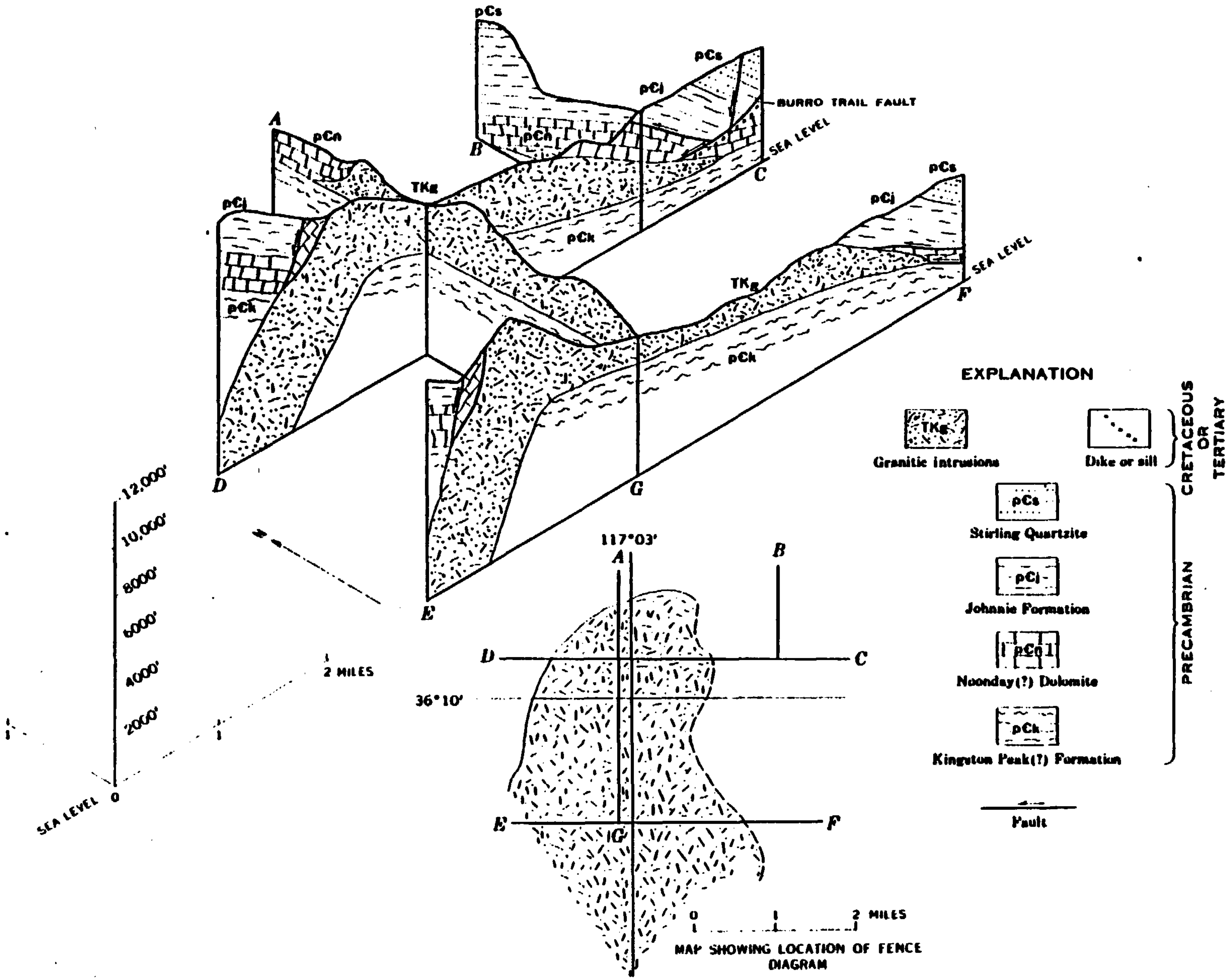

FIGURE 94.—Fence diagram, isometric projection, illustrating inferred shape of the granitic intrusion at Hanaupah Canyon.

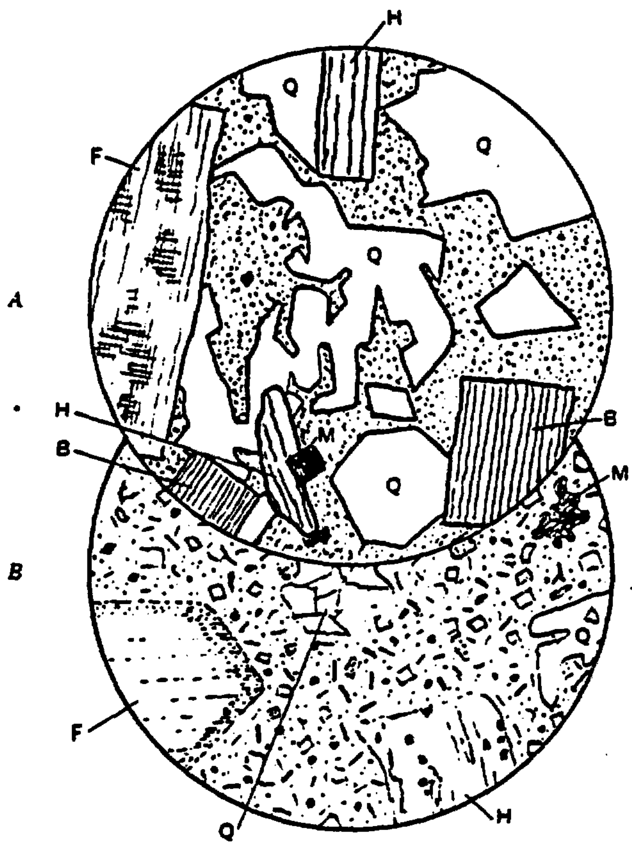

FIGURE 95.—Micrographs of thin sections of the granitic intrusion at Hanaupah Canyon and associated sill. Q, quartz; F, feldspar; H, hornblende; B, biotite; M, magnetite. Diameter of field, 2.5 mm. *A*, Quartz monzonite from the intrusion. Phenocrysts of feldspar are mostly oligoclase mottled with potash feldspar. Most of the rock is a graphic intergrowth of quartz and clouded feldspar stippled areas). Dark minerals are biotite, hornblende and magnetite. *B*, Monzonite porphyry from a sill along the Burro Trail fault. Phenocrysts of oligoclase have strongly argillized borders. Hornblende is altered to calcite, magnetite, and epidote(?). Quartz occurs as deeply embayed phenocrysts and as secondary nests. Magnetite occurs in small phenocrysts, some of which are embayed like the quartz. The groundmass, strongly argillized, consists of lathlike and rectangular feldspars and tiny specks of magnetite.

The granite is a homogeneous porphyry as far as could be judged by examining many boulders from it along Hanaupah and Starvation Canyons and by examining the north edge of the intrusion. The rock contains large phenocrysts of K-feldspar with rims of oligoclase associated with quartz, biotite, hornblende, and some magnetite (fig. 95).

Trace elements in the granite at Hanaupah Canyon (table 25), as already noted, are similar to those in the granite at Skidoo.

The contact at the edge of the granite is sharp like the roof and sides of the porphyry facies of the granite at Skidoo. Porphyry containing large phenocrysts occurs within a few inches of the contact, but at the contact the porphyry is fine grained and granulated (fig. 93).

Contact-metamorphic effects consist of bleaching of the Noonday(?) Dolomite and developemnt of nests of tremolite and calcite. Minor quantities of sulfides have been deposited along faults and fissures extending north from the intrusion. In addition, under the summit ridge for 2 miles north from Telescope Peak, the lower part of the Johnnie Formation and upper part of the Noonday(?) Dolomite are stained brown, presumably because of oxidation of disseminated iron sulfide. Northward from this belt along the summit ridge for another 2 miles the joints and fissures in the Johnnie Formation are coated with epidote.

On the Colorado Plateau, alteration zones like these are found only around the stocks, not around the laccoliths. The position of these alteration zones, therefore, suggests that the source for the granite at Hanaupah Canyon is under the west side of the intrusion, and this is a basis for the interpretation given on figure 94.

The alteration zones extend 4 miles north-northwestward from the granite to within 2 miles of a granitic body in Wildrose Canyon. The granite in Wildrose Canyon, which is gneissic and contains biotite, more closely resembles the granite at Skidoo than the granite at Hanaupah Canyon.

Dikes and sills extend 2 miles northward from the granite of Hanaupah Canyon, and others occur along the Burro Trail and other faults east of the granite (figs. 94, 96). The dikes and sills are 5–10 feet thick and provide a measure of the fluidity of the magma that was intruding. By contrast, at the laccolithic mountains on the Colorado Plateau, only the latest intrusions were sufficiently fluid to form thin dikes and sills; the earlier ones were highly viscous.

The rocks comprising the dikes and sills tend to be finer grained than the porphyry in the main intrusion, but many closely resemble it (fig. 95). Moreover, their content of trace elements is similar (table 25). Almost certainly the sills along the thrust faults east of the granite are connected with it like those that extend to the north.

The sills along the thrust faults locally have chilled contacts against the faulted surfaces and clearly are later than the faulting. However, there was renewed movement on some thrust faults later in Tertiary time; possibly later movement on the Burro Trail and neighboring faults followed shaly layers above or below the sills without severely fracturing those intrusions.

These sills along the faults are much altered, but it was not determined to what extent the alteration was caused by deuteric action, hydrothermal activity, or to weathering due to ground water percolating along the faults. Throughout the Panamint Range, the thrust faults have served as aquifers.

CHAOTIC COMPLEX ALONG THE AMARGOSA THRUST

The chaotic complex along the Amargosa thrust, along the east foot of the Panamint Range (p. A51), is structurally arched. Its high part is between Hanaupah and Death Valley Canyons where the thrust fault and underlying Precambrian metamorphic rocks are exposed (fig. 96). The Precambrian rocks are intruded by a granitic mass and a still younger swarm of felsite dikes (fig. 97). At Hanaupah Canyon the upper plate of the thrust is Precambrian Stirling Quartzite. Farther north the upper plate consists of progressively younger Cambrian formations.

At Death Valley Canyon these arched rocks of the complex plunge northward under a mass of volcanic rocks having interlayered slabs of Paleozoic dolomite (fig. 98), a mixture highly suggestive of the chaos which was described by Noble (1941) in the Virgin Spring district 20 miles southeast of here. The swarm of felsitic dikes that intrude the granitic mass and the metamorphic rocks south of Death Valley Canyon also intrude the chaos north of the canyon. The chaos also is cut by felsite plugs.

The thrust and the underlying metamorphic rocks and granitic intrusion are not exposed farther north or south, but in both directions the complex is represented by lavas, dikes, and areas of hydrothermal alteration. Very possibly the granitic mass that underlies the highest part of the structural arch may thin northward and southward and be a main cause of the arching. The thrust fault is well exposed where it crosses the divide a mile north of the mouth of Hanaupah Canyon. There the upper plate is Stirling Quartzite dipping about 45° E; the fault dips 15° W. (fig. 96). The lower part of the quartzite is thoroughly granulated, in part mylonitized, in a zone about 50 feet thick. Below this, the mylonitized quartzite is mixed with crushed Precambrian metamorphic rocks from the lower plate.

The metamorphic rocks include augen gneiss, feldspar-biotite gneiss, biotite schist, and quartz gneiss. The most conspicuous and most abundant rock is the augen gneiss (figs. 99, 100), which consists of augen of feldspar as much as 1 inch long in a matrix of feldspar and quartz cut by stringers of biotite. The augen make up perhaps 15 percent of the rock. Just below the thrust the augen gneiss is finely layered, and augen are few. The rock there grades into feldspar-quartz-biotite gneiss.

The augen gneiss is about 50 percent quartz. The augen, which constitute about 20 percent of the rock, are altered (silicified or argillized?) feldspar, probably microcline or albite. Some augen contain irregular grains of quartz, plagioclase, and biotite. The rest of the rock is about half plagioclase and half biotite, the latter occurring mostly as bands or layers in the gneissic structure. Zircons from several specimens of augen gneiss are rounded and colorless (Ralph L. Erickson, written common., 1961) (fig. 103). Their lead-uranium and lead-thorium ages are greater than a billion years (T. W. Stern, written commun., 1965).

Some facies of the biotite gneiss contain considerable potassium feldspar, as much as 25 percent. Other constituents are quartz, 40 percent; plagioclase, 20 percent; and biotite, 15 percent. Other facies are without potassium feldspar and contain 60 percent quartz, 20 percent plagioclase, and 15 percent biotite. In some specimens of these rocks Erickson (written Commun., 1961) found two kinds of zircon (fig. 103), mostly colorless and rounded like those in the augen gneiss, but some euhedral and pale pink with rodlike inclusions. More work is needed to determine how these types are distributed in the different facies of the biotite gneiss. Trace elements in the augen gneiss and biotite gneiss are given in table 26 (p. A140).

The gneiss is strikingly foliated and the dips are to the west. There are belts 50–250 feet wide in each of which the dip of the foliation increases westward from about 20°–70° (fig. 101). The parts of the belts having the most steeply dipping folia generally are occupied by dikes.

The dikes cutting the augen gneiss range in width from stringers less than 1 inch to dikes about 6 feet wide. Contacts are sharp and most of them show distinct chilled edges, although some contacts are obscured by subsequent shearing.

The age of the metamorphism and the age of the metamorphosed rock are not clear. The rock is Precambrian (see above), but some of the metamorphism may be very much later, because biotites from these same rocks give a late Miocene potassium-argon age (T. W. Stern, written commu., 1965). This could account for the close spatial association of the augen gneiss with so much volcanic activity and with the Amargosa thrust fault, not only here but also in the Virgin Spring area (Noble, 1941). A close relationship between the fissuring that localized the dikes and the structure of the metamorphism is indicated by the steepened dips in the folia adjacent to dikes. Also, numerous thin stringers of aplite in the augen gneiss are more like the dike rocks than the metamorphic rocks. The aplitic rocks are mostly potassium feldspar (as much as 60 percent) and quartz with only a little plagioclase and biotite.

Part of the metamorphism seems to be later than the thrust faulting. Locally, there has been reaction between the gneiss and the granitic rock, suggesting local assimilation. Augen locally are collected in pegmatitic

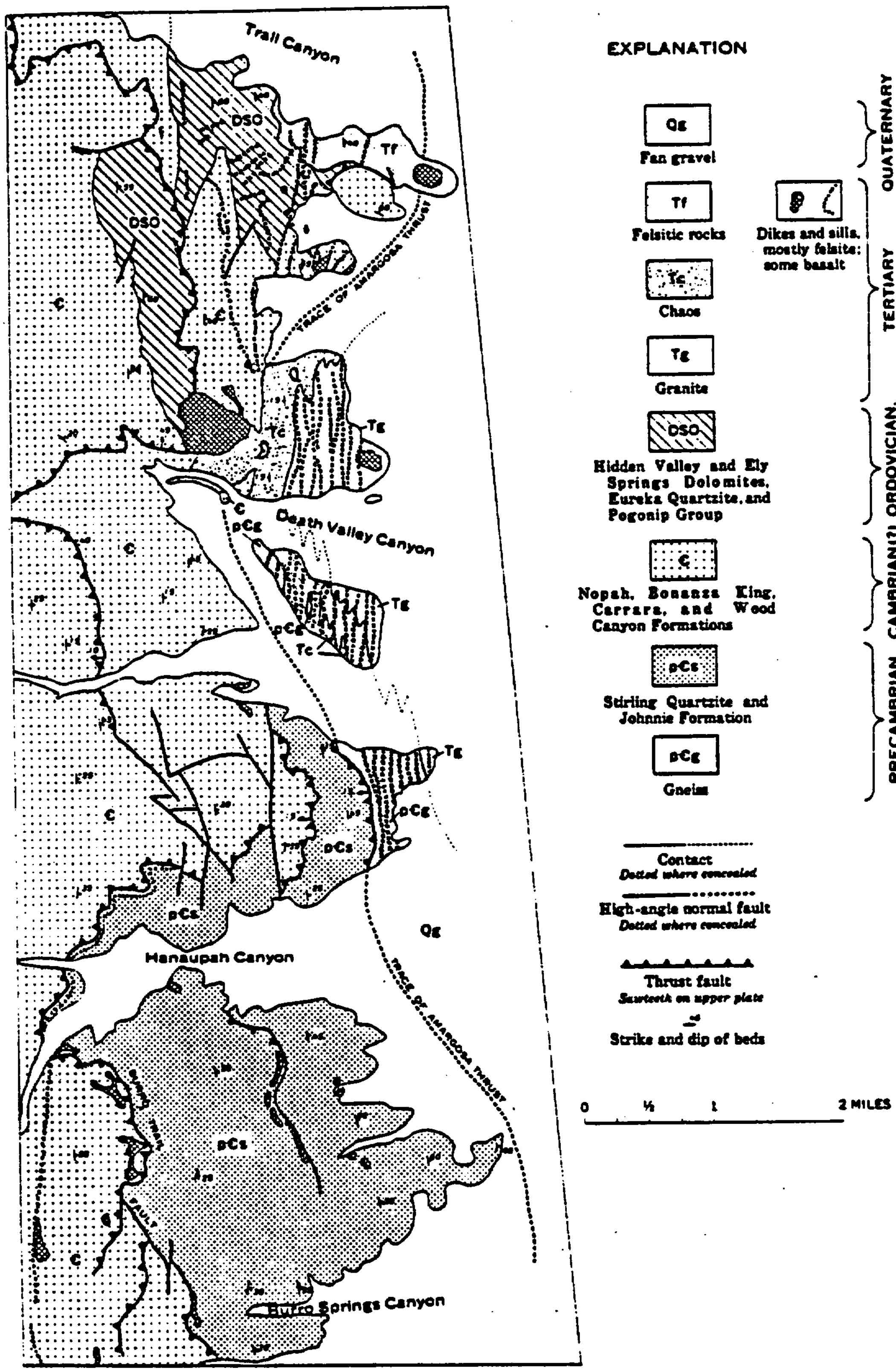

FIGURE 96.—Map of Amargosa thrust complex along the east foot of the Panamint Range.

FIGURE 97.—View of dike swarm in Amargosa thrust complex. At this location, the mouth of Death Valley Canyon, the host rock is a granitic intrusion.

or aplitic masses that seem to provide a gradation between the gneiss and some of the felsitic dikes that cut the gneiss. Also, the trace elements in these rocks are similar (table 26).

Other evidence indicating metamorphism attributable to the igneous activity is the occurrence of incompletely replaced metasediments in the granitic rocks (fig. 102). Some of the granite and some of the other igneous rocks may have been similarly generated from older rocks along the thrust fault. In short, the concentration of augen gneiss, granite, felsite dikes, and volcanics, together with various kinds of metasediments and partly granitized rocks along the Amargosa thrust, suggests that the fault zone is an old structure which repeatedly

FIGURE 98.—View of chaoslike formation in Amargosa thrust complex. The formation consists mostly of volcanic rocks (light colored) having interlayered slabs of Paleozoic dolomite (dark rock in center and foreground). Location is north side of Death Valley Canyon.

has been the site of mylonitization, recrystalliaztion, igneous invasion, metamorphism and granitization, and perhaps of magma generation.

Brecciation of the metamorphic rocks at the Amargosa thrust shows that the metamorphism in part at least antedates the thrust, yet this brecciation could partly be due to renewed movement on the thrust. Probably much of the metamorphism is Precambrian, but part of it may be much younger.

Ralph L. Erickson (written commun., 1961) found two kinds of zircons in the Amargosa complex (fig. 103).

Zircons from the augen gneiss are colorless, rounded, and frosted; zircons from the granite and volcanic rocks are brownish, pink, euhedral, and somewhat lar[illegible]r. Zircons from the biotite gneiss are mixed, but the colo[illegible] riety predominates strongly. Zircons in the aplite are th[illegible] ral pink variety.

The occurrence of these two varieties suggests that the rounded colorless zircons are detrital and much older than the euhedral pink variety, and further, that the augen gneiss and biotite gneiss are metasediments intruded and further metamorphosed by younger granitic rocks.

Along the east side of the arched metamorphic rocks is a granitic intrusion. Its exposed part widens north-

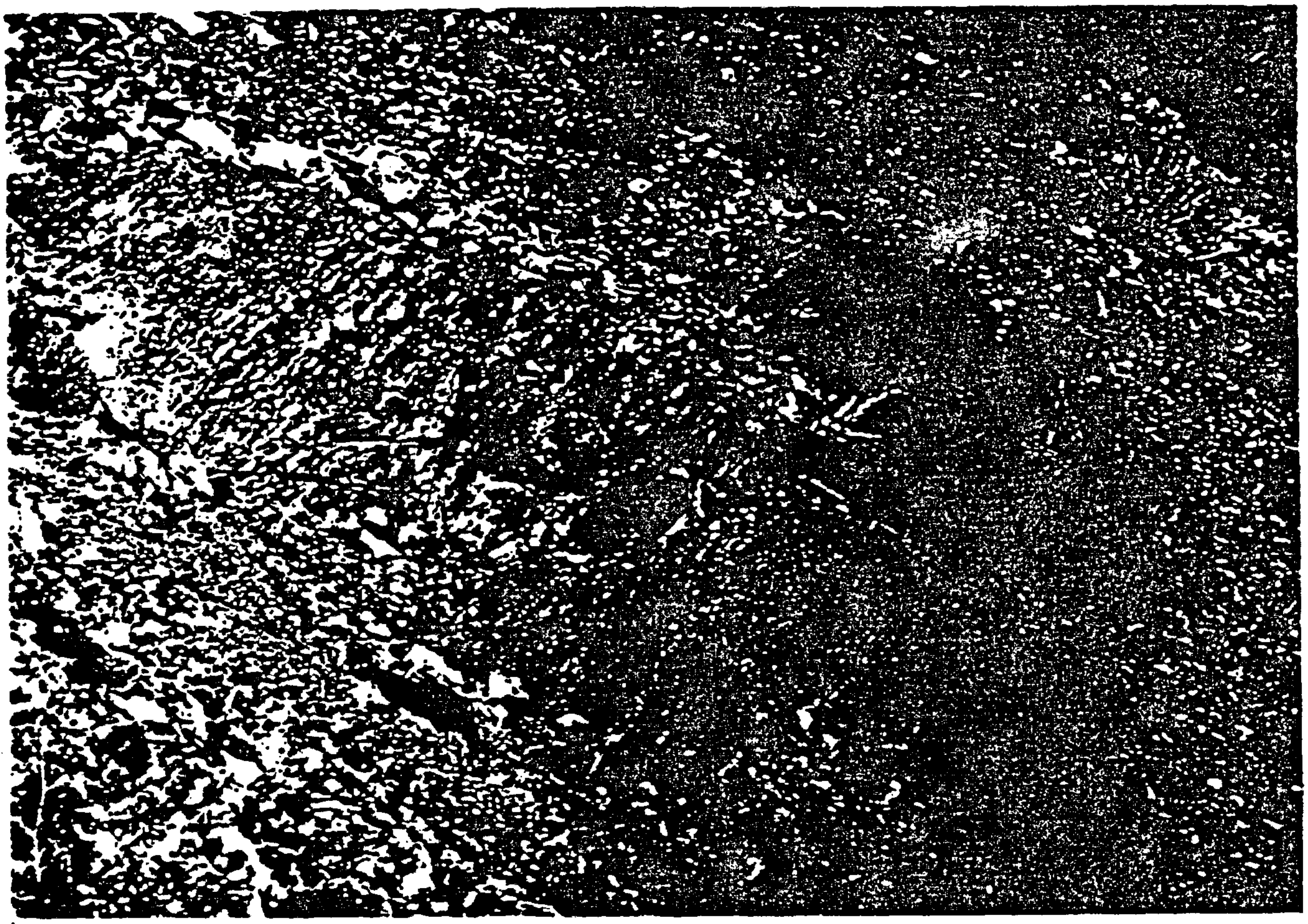

FIGURE 99.—View of augen gneiss in Amargosa thrust complex north of Hanaupah Canyon at the east foot of the Panamint Range.

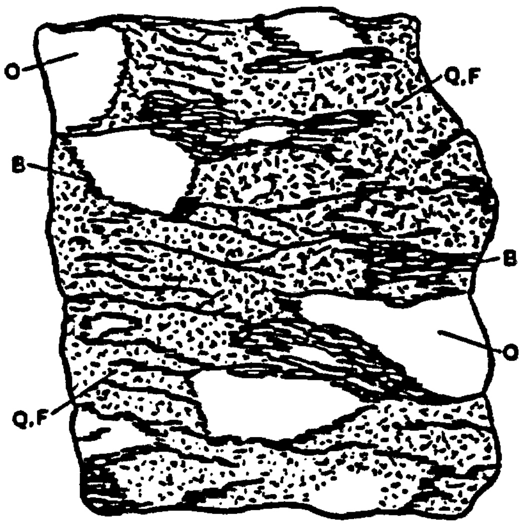

FIGURE 100.—Specimen of augen gneiss from Amargosa thrust complex 1 mile north of the mouth of Hanaupah Canyon. The rock consists of dark biotite-rich stringers (B) between lighter colored augenlike masses that are mostly fine-grained quartz and feldspar (Q, F) and large augen of orthoclase (O). Natural size.

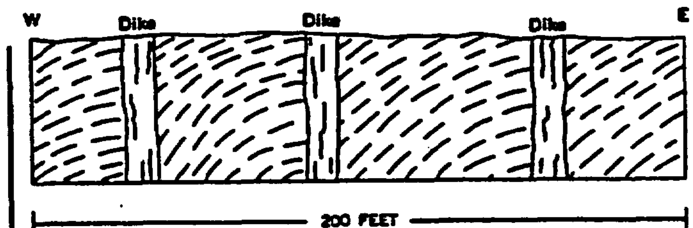

FIGURE 101.—Diagrammatic section illustrating flattening of foliation of augen gneiss eastward from dikes. The dikes are 1–3 feet wide and are 6–15 feet apart. Along the east wall of many dikes the foliation dips steeply west into the dike, but within a few feet eastward the dip flattens.

ward from the tip of the spur north of Hanaupah Canyon to the north side of the Death Valley Canyon, but not enough of the granite is exposed to reveal its structural form. Probably it is tabular and approximately parallel to the thrust like the similar intrusion below the augen gneiss under the Amargosa thrust fault in the Virgin Spring area (Noble, 1941). The swarm of dikes, mostly felsite, crossing the metamorphic rocks also crosses this granite (figs. 96, 97).

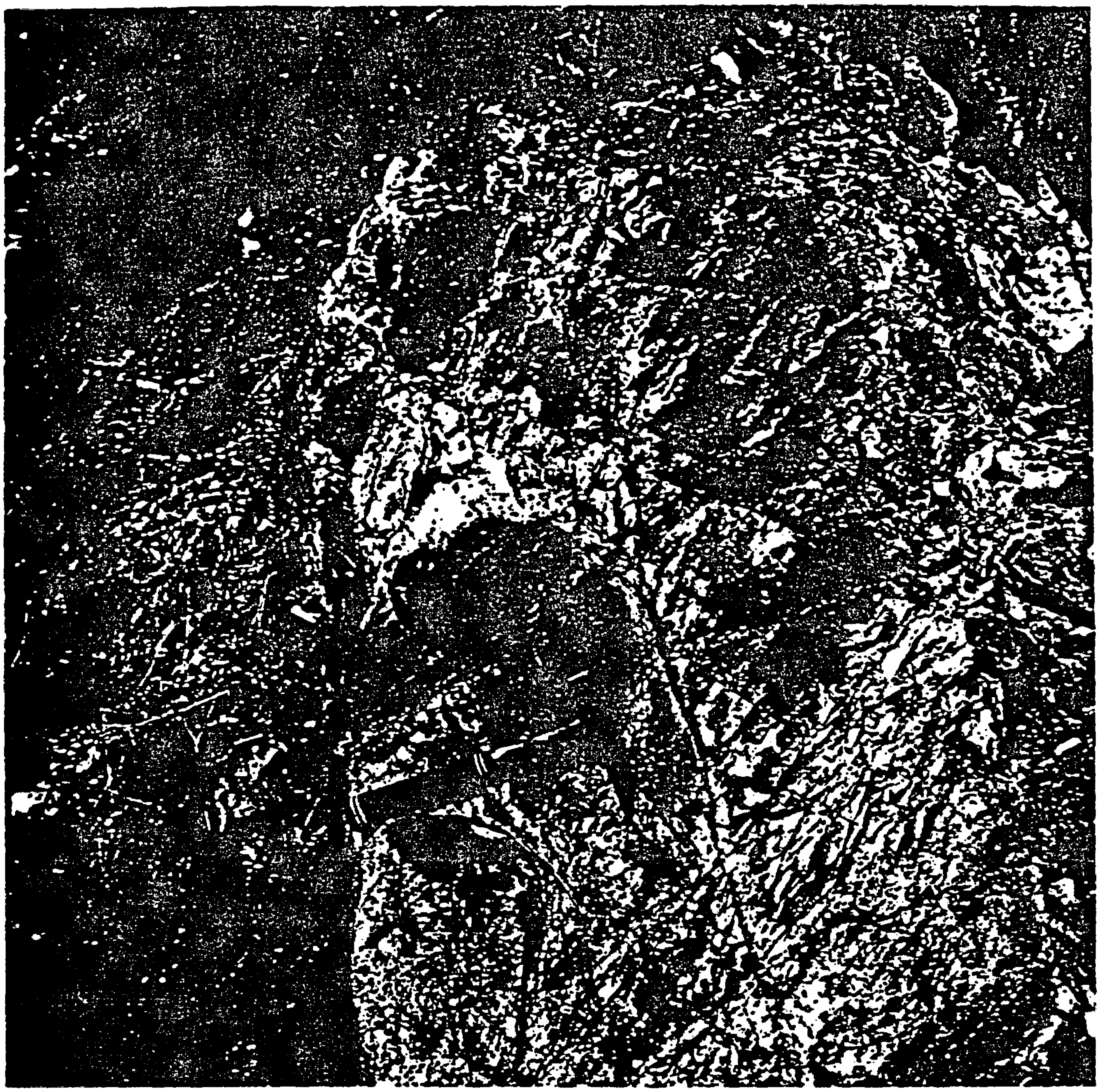

FIGURE 102.—Granitized metasediment (light rock) and incompletely replaced metasediment (dark rock above the hammer) in Amargosa thrust complex. At left is a felsite dike. A layer of green argillite containing lenses of mica cuts through the dark rock above the hammer and extends upward to the right as veins of the greenish mineral in the porphyry. Locality is about midway between Hanaupah and Death Valley Canyons.

The rock is finer grained than the granites at Skidoo or at Hanaupah Canyon. It contains roughly 40 percent potassium feldspar, 30 percent plagioclase, 20 percent quartz, and 10 percent biotite. The quartz and orthoclase are in graphic intergrowths (fig. 104). Some facies might be referred to as alaskite. The texture, which is more that of a hypabyssal rock than a deep-seated pluntonic one, is intermediate between the textures of the other "granites" and the volcanic rocks. The zircons are the euhedral pink variety (fig. 103).

Above the granite is 400–500 feet of Precambrian augen gneiss, and above the augen gneiss is the Amar-

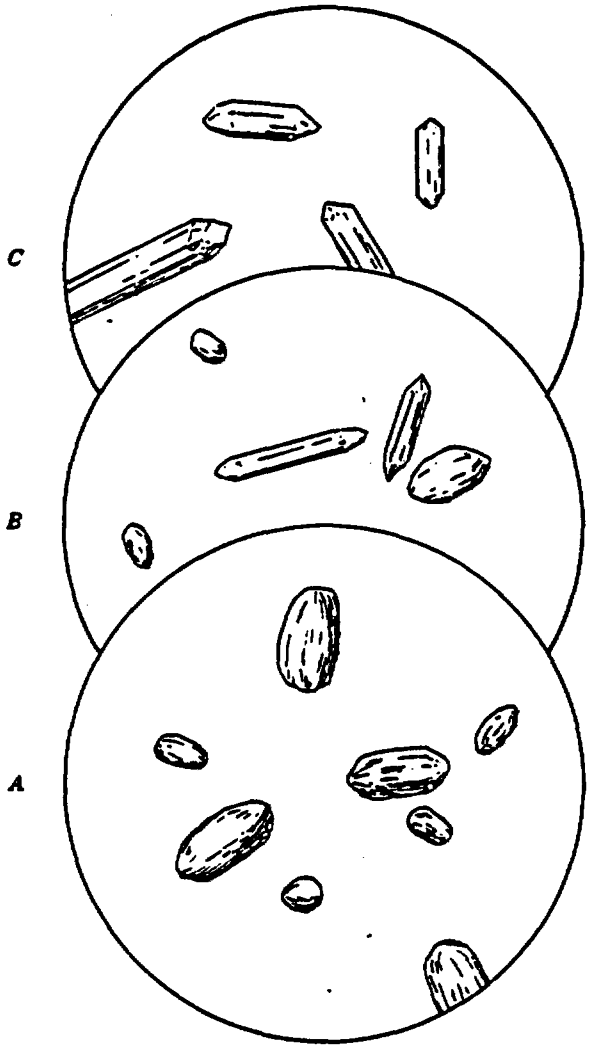

FIGURE 103.—Zircons from Amargosa thrust complex between Hanaupah and Death Valley Canyons. The Precambrian augen gneiss contains colorless round zircons (*A*). The granitic rock in the complex that may have been derived in part by melting such gneiss contains similar round zircons and sharply terminated euhedral pink zircons (*B*). Dikes cutting the granite, and seemingly in part derived from it, contain only the pink euhedral zircons (*C*). Lead-uranium and lead-thorium ages of zircon from the augen gneiss indicate Precambrian age; lead-alpha ages of zircon from the other rocks indicate middle Tertiary age (T. W. Stern, written commun., 1965). Width of fields, 1.35 mm.

gosa thrust fault, the upper plate of which is Stirling Quartzite. The sequence is very similar to that of the Amargosa thrust in the Virgin Springs district (Noble, 1941) where augen gneiss, identical to that at the east foot of the Panamint Range, lies above a granitic intrusion and below the Amargosa thrust. In the Virgin Springs district the upper plate of the thrust consists of a chaotic mixture of upper Precambrian sedimentary rocks and Tertiary volcanic rocks, a mixture referred to as "chaos" (Noble, 1941).

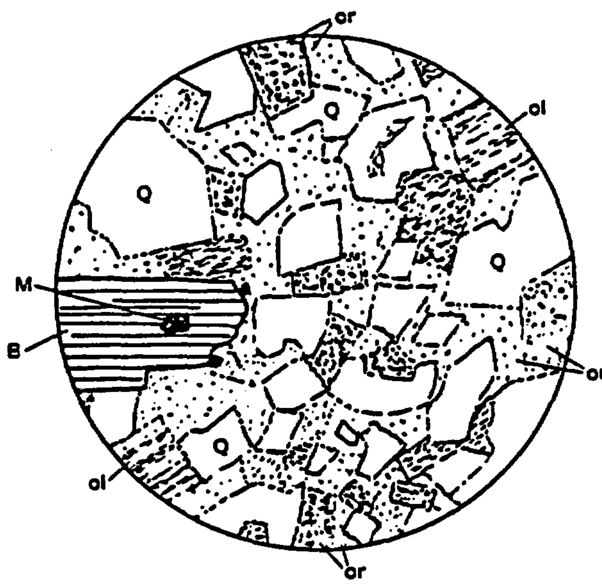

FIGURE 104.—Micrograph of thin section of granitic rock in Amargosa thrust complex at Death Valley Canyon. The rock is mostly quartz (Q) and orthoclase (or) in graphic intergrowths. Oligoclase (ol) and biotite (B) comprise the rest of the rock, and there is a trace of magnetite (M). Diameter of field, 2.5 mm.

The gneiss is brecciated at the crush zone along the thrust fault as if the gneiss antedates the fault. In part it probably does, but the relationships may have been complicated by later movements along the fault, and perhaps by some metamorphism dating from the time of the thrusting and the granitic intrusion.

The chaoslike formation exposed at the mouth of Death Valley Canyon is composed mostly of volcanic rocks (fig. 98). Intercalated with these rocks are slabs of Paleozoic dolomite and some slabs of Precambrian metamorphic rock including augen gneiss. Individual slabs are 10–50 feet thick and as much as 500 feet long; they dip at about 25° W. Some dolomite slabs are bordered by skarn. The dike swarm also cuts this chaos, but the dikes seem to be fewer than in the augen gneiss and granite. In addition to dikes, there are some plugs of felsite (fig. 105). One on the north side of Death Valley Canyon is 1,500 feet in diameter.

The chaos lies above the granitic intrusion and between it and the Cambrian formations in the upper plate of the Amargosa thrust. The contact between the intrusion and the chaos is obscured by shearing, but it seems to dip west about parallel to the sheeting in the chaos.

The structural relationship between this chaos and the Cambrian formations in the upper plate is not at

FIGURE 105.—Felsite plug (dark conical hill) in granite in Amargosa thrust complex. The plug is associated with a swarm of dikes trending across the direction of view. Locality is north side of Death Valley Canyon.

all clear, because the contact is obscured by a felsite plug north of the canyon mouth and by faulting.

Between Hanaupah and Death Valley Canyons the dikes, plugs, and other intrusions which are nearly vertical, are restricted to the lower plate. They do not cut the Stirling Quartzite and younger formations in the upper plate.

Moreover, the belt of volcanic activity extends as a narrow linear belt northward and southward from the arched lower plate, indicating that the lower plate was very near the surface at the time of the volcanism.

One of the most extensive eruptive masses is north of Trail Canyon where felsite flows and lapilli tuffs 500 feet thick overlap the Paleozoic formations in the upper plate. The flows dip about 25° E.; the Paleozoic formations dip about 50° E. Parallel layering in the eruptives indicates that their eastward dip is not original and depositional; evidently half the eastward tilting of the Panamint Range occurred before and half after these eruptives were formed.

The volcanism north of Trail Canyon also is marked by dikes of felsite and basalt cutting the upper plate, by a felsite plug 2 miles south of Blackwater Wash, by tuffs and basalt flows on each side of the mouth of Blackwater Wash, and by hydrothermal alteration of the Paleozoic formations. The Pogonip Group especially is reddened northward and southward from Trail Canyon.

Southward from Hanaupah Canyon the belt of volcanism similarly is marked by eruptives, dikes, sills, plugs and extensive areas of hydrothermal alteration. The eruptives, mostly tuffs, dip east and overlap westward onto the more steeply dipping rocks of the upper plate.

At Perlite Hill, porphyritic rhyolite is overlain by perlite that is capped by rhyolitic tuff and agglomerate, the whole about 800 feet thick.

A deposit of lapilli tuff high on the mountain west of Perlite Hill and north of Starvation Canyon (fig. 96) is broken by small thrust faults on which the movement has been westward. The tuff deposit south of the mouth of Starvation Canyon is broken by a series of high-angle normal faults. On all these faults the displacements are small, less on the average than the displacements of the neighboring Cambrian and Precambrian formations along similar faults. It seems likely that both the thrusting and the high-angle normal faulting were renewed after the volcanism.

Sills of porphyry intruded along the Burro Trail fault are like the porphyry in the granite at Hanaupah Canyon (fig. 95; also table 25). Moreover, these sills occur only along that part of the fault that is nearest the granite—the part from Starvation Canyon to the north side of Hanaupah Canyon. The sills very likely are connected with the granite at Hanaupah Canyon (p. A128), and perhaps the granitic intrusion in the lower plate of the Amargosa thrust north of Hanaupah Canyon is too.

Despite the considerable variation in texture and composition of phenocrysts in the volcanic rocks of the Amargosa thrust complex, the rocks are more alike than different (fig. 106). The dike rocks generally contain considerable potassium feldspar. Plagioclase tends to occur as phenocrysts, whereas the potassium feldspar is mostly in the groundmass. Zircons in these dikes, are the euhedral pink variety (fig. 103). The rocks occurring in the volcanic plugs are like the dike rocks. The distribution and concentration of minor elements in the different rocks are shown in table 26.

The inferred relation of the thrust to the granites and volcanics in the Panamint Range and eastward to the Virgin Spring area is presented in figure 108.

At the time the volcanism occurred the east foot of the Panamint Range had been eroded to within one or two hundred feet of the present surface, because from Trail Canyon north and at Starvation Canyon and farther south, lavas and other eruptives lie at the foot of the Panamint Range and rise part way onto it (fig. 89). Volcanic rocks probably underlie the gravel fans in those areas too, as indicated by magnetic anomalies on the Trail Canyon fan and on the gravel fans south of Hanaupah Canyon. This distribution of the volcanic rocks coupled with the fact that sills from the granite at Hanaupah Canyon spread eastward along branches of the Amargosa thrust, like the Burro Trail fault, suggests that the igneous activity progressed upward and eastward under the Amargosa thrust and erupted when the surface was reached at the east foot of the range. This is the interpretation illustrated on figure 90.

The suggestion is strong that the chaos in some way is related to the volcanism as well as to the thrust faulting, a relationship also favored by L. F. Noble (written commun.); but what the relationship is, remains to be demonstrated. There was thrust faulting before the volcanism, but in some places there was later thrusting too, both in the Panamint Range (see above) and in the Virgin Spring district (Noble, 1941).

To restore dips in the Panamint Range as they were at the time of the volcanism, the range must be rotated almost 20° back to the west (fig. 89). This places the highest part of the granite in Hanaupah Canyon (fig. 90) about a mile lower than the surface on which the eruptives spread. In addition, the domes produced by the granitic intrusions should be flattened. These changes practically eliminate the Panamint Range as a topographic feature. The Paleozoic and Precambrian formations composing the Panamint Range would have had only moderate dips to the east, mostly less than 25°, and the Amargosa thrust and its branches would have dipped west 10°–45°.

Folding of the Amargosa thrust and removal of upper plate rocks from above the thrust have exposed the gneissic cores of anticlines at three places in the Black Mountains. One of these is at Copper Canyon (fig. 107), another at Mormon Point (fig. 108), and a third at Badwater (fig. 38). The surface of the gneiss at these anticlines is smooth, and the domed surfaces are the so-called turtlebacks (Curry, 1938b, 1954).

Figure 107 is a view of the turtleback at Copper Canyon. Tertiary formations overlap the smooth surface of the gneiss and have been faulted against it. The surface of the Precambrian gneiss is the domed surface of the Amargosa thrust. An interpretation of the relations is that the thrusting, doming, and removal of the chaos from this turtleback occurred before deposition of the Tertiary formations seen in this view. Subsequently these formations were downfaulted against the turtleback surface. Drewes (1963) found boulders of volcanic rocks like those in the chaos in the Tertiary formations in Copper Canyon.

Figure 108 is an interpretation of the relations between the Amargosa thrust, turtleback surfaces, and granitic intrusions in the Black Mountains and their counterparts in the Panamint Range. This interpretation, although arrived at independently, was first suggested by Noble in 1941. He wrote (p. 995) that "the dominant structure of the Panamint Range may be an anticline in a thrust plate like the Amargosa 'chaos,' in the core of which the Precambrian gneiss of the autochthonous block is exposed."

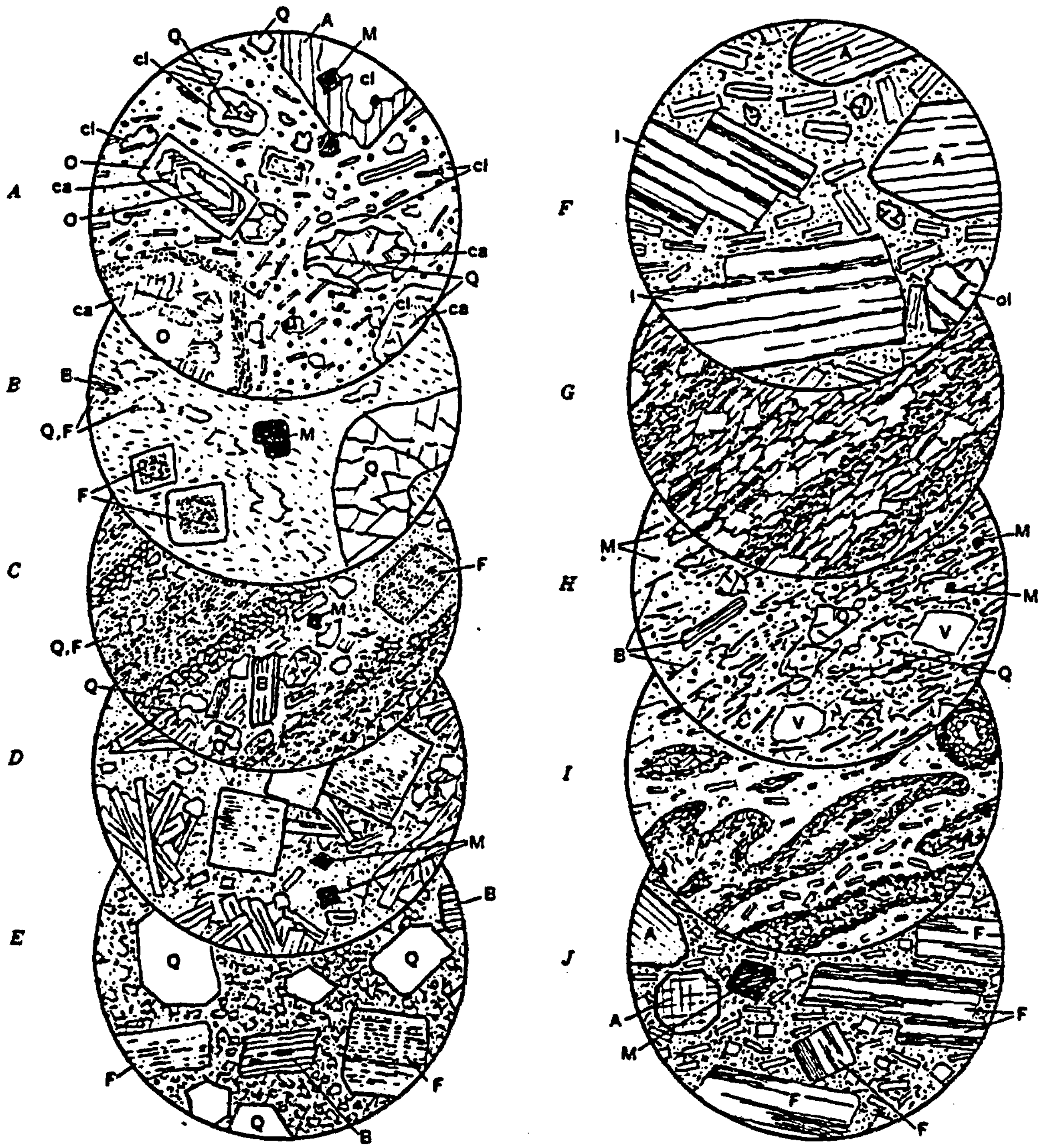

FIGURE 106.—Micrographs of thin sections of volcanic rocks in Amargosa thrust complex at base of the Panamint Range. Diameter of fields, 2.5 mm.

Q, quartz
F, feldspar
(andesine in E and J)
O, oligoclase

L, Labradorite
A, augite
B, biotite
cl, chlorite

ol, olivine
M, magnetite
ca, calcite
V, vug

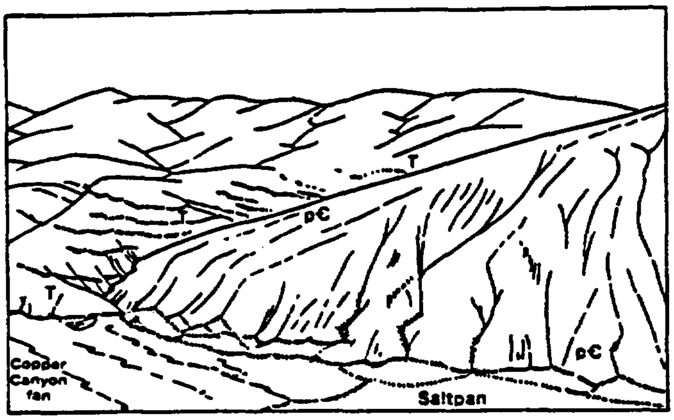

FIGURE 107.—Sketch of Copper Canyon turtleback, from oblique aerial photograph. View is east. The turtleback, composed of Precambian gneiss (pC) has a smooth top and forms a smooth but steep front of the mountain rising more than 2,000 feet above the saltpan. Tertiary formations (T) overlap the north end of the turtleback and are faulted down against it.

INDICATED GRANITIC ROCK UNDER THE PANAMINT RANGE

The highest Bouguer gravity-anomaly values in the Death Valley region are over the Precambrian metamorphic rocks in the Black Mountains, and anomalies almost that high were found over the Funeral Mountains and Tucki Mountain. Over the main part of the Panamint Range the anomaly values are low, even on the west side where Precambrian metamorphic rocks are exposed.

The gravity-anomaly values decrease westward towards the Sierra Nevada, but the low values under the Panamint Range are, in large part, a local feature related directly to that range.

The steepness of the gravity gradients on the southeast side of Tucki Mountain suggests that the density contrast producing the anomaly is at least, in part, in the upper few kilometers of the crust. The extent of the anomaly requires that a relatively low density mass underlies most of the Panamint Range south and west of Tucki Mountain. Too, the anomaly values over the metamorphic rocks on the west side of the Panamint Range are lower than those in the metamorphic rocks in the Slate Range, a few miles to the southwest.

If the low density mass underlying the Panamint Range is granitic rock in contact with Precambrian metamorphic rock, the granitic mass must be about 2 miles thick.

In terms of deep crustal structure the combined geological, seismic, and gravity data (p. A107) suggest three quite different possibilities. One is that deep in the crust under the Panamint Range is a large granitic intrusion that is continuous westward with the Sierra Nevada batholith. Its surface would rise irregularly

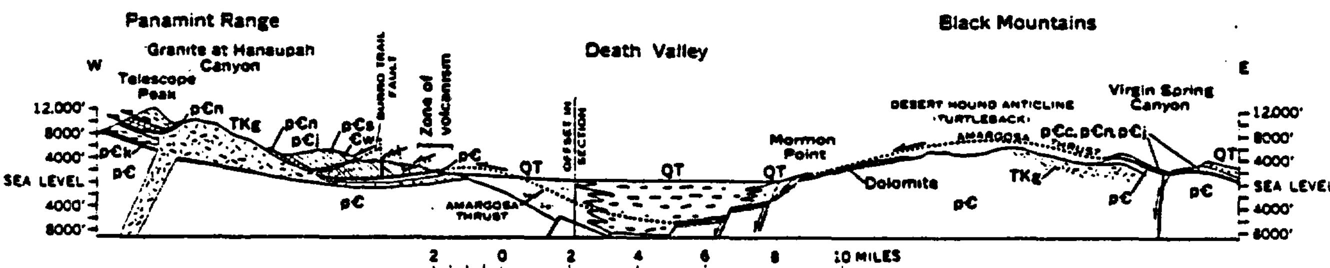

FIGURE 108.—Section of Black Mountains, Death Valley, and Panamint Range showing supposed extent of Amargosa thrust and relations of granitic intrusions and volcanism to it. Section across the Black Mountains from Noble (1941). QT, Quaternary and Tertiary valley fill; TKg, granitic intrusions; Cw, Wood Canyon Formation; pCs, Stirling Quartzite; pCj, Johnnie Formation; pCn, Noonday Dolomite; pCc, Crystal Spring Formation; pCk, Kingston Peak Formation; pC, Precambrian metamorphic rocks.

FIGURE 106—EXPLANATION

A. Dark-greenish augite andesite dike from the dike swarm 1 mile north of Hanaupah Canyon. Phenocrysts of augite are altered to chlorite and magnetite. Phenocrysts of oligoclase are altered to calcite and many have an argillized border. The groundmass consists of laths of oligoclase and cubes of magnetite. Interspersed with these are nests of chlorite, quartz, and calcite.

B. Pink felsite dike, near *A*. Phenocrysts of feldspar are mostly altered to a cryptocrystalline material resembling the groundmass. There are a few phenocrysts of magnetite and biotite. Vugs contain secondary quartz. The groundmass consists of microlites of oligoclase in a cryptocrystalline paste and is interrupted by irregular growths of feldspar and quartz.

C. Banded pink feldsite dike, near *A*. The bands differ in texture and composition. Some contain phenocrysts of biotite, argillized feldspar, probably oligoclase, and magnetite. Other layers are mostly quartz, which also occurs as irregular growth in the other layers and as veins.

D. Dacite dike, near *A*. Phenocrysts of andesine and laths of oligoclase with occasional magnetite are in a cryptocrystalline argillized groundmass.

E. Quartz dacite dike, near *A*. Euhedral quartz, andesine, and biotite phenocrysts in granular groundmass of feldspar (oligoclase?). In the biotite are irregular growths of magnetite.

F. Basalt dike on north side of Trail Canyon. Phenocrysts of augite and labradorite and a few small ones of olivine. Groundmass has small laths of andesine and small olivine crystals.

G. Felsite, interior of plug 2 miles south of Blackwater Wash. The rock has thin dark bands of glassy material separating wider and lighter feldspathic bands.

H. Contact zone at *G*, dark perlite. Biotite and magnetite occur as small phenocrysts and in the groundmass. The groundmass is largely feldspar, quartz, and glass. Many vugs are open; others contain quartz.

I. White banded tuff, underlies basalt at The Dinosaur. Feldspathic layers with microlites of oligoclase and tiny flattened blebs of glass are separated by glassy layers that are folded. Some vugs are lined with quartz and feldspar and are surrounded by glass.

J. Augite andesite lava at The Dinosaur. Phenocrysts are andesine, augite, and magnetite. The groundmass contains microlites of andesine and some olivine(?) in a glassy paste.

TABLE 26.—*Trace elements in the Amargosa thrust complex*

[Semiquantitative spectrographic analyses by E. F. Cooley, U.S. Geol. Survey. Values in parts per million, except Mg, which is given in percent]

ROCKS FROM AMARGOSA THRUST COMPLEX BETWEEN HANAUPAH AND DEATH VALLEY CANYONS

Pb	Mn	Cu	Zr	Ni	Co	V	Y	Be	Ti	La	B	Sc	Cr	Ba	Sr	Mg
Augen gneiss																
10	300	20	150	5	<10	50	50	1	3,000	70	10	10	20	1,500	150	1
<10	200	20	150	5	<10	20	20	1	3,000	50	20	<10	10	1,500	100	.7
10	200	10	150	5	<10	50	20	<1	3,000	70	10	10	10	1,000	100	1
70	200	30	300	20	7	50	30	<1	3,000	70	20	15	50	500	70	1
Biotite gneiss																
<10	150	5	300	50	15	50	30	1	5,000	50	20	10	50	1,000	50	1.5
10	150	70	300	10	10	50	20	<1	5,000	50	20	20	20	700	150	1
<10	100	30	200	5	<10	20	20	1.5	3,000	50	15	<10	10	300	200	.7
<10	200	50	200	7	10	50	20	1	3,000	70	10	10	15	1,000	70	1
20	700	50	150	20	10	70	20	<1	3,000	70	10	10	50	700	100	1.5
<10	200	15	150	7	<10	50	20	1	3,000	50	10	<10	20	100	70	1
Mylonite																
<10	300	3	200	10	10	50	20	2	2,000	50	10	10	20	70	70	2
Granitoid rocks																
20	500	30	100	<5	<10	15	15	1	1,500	50	10	<10	<10	700	150	0.2
10	500	20	200	5	<10	20	20	1	3,000	70	10	<10	<10	1,000	150	1
30	500	30	200	5	<10	20	10	1	2,000	70	10	<10	<10	1,000	200	.5
20	500	20	100	20	<5	10	20	1	1,500	<50	20	<10	30	500	70	.5
Felsite dikes																
<10	100	30	500	5	<10	50	20	1	7,000	70	20	10	10	2,000	150	1
30	700	70	200	<5	<10	<10	20	2	2,000	50	15	<10	<10	2,000	200	.7
<10	50	50	150	<5	<10	<10	15	1	1,500	<50	10	10	<10	100	100	.1
<10	100	5	150	<5	<10	10	20	1	1,000	50	20	<10	<10	500	50	.1
10	50	20	200	<5	<10	<10	20	1.5	1,000	50	10	<10	<10	300	50	.1[illegible]
20	300	30	150	<5	<10	10	20	1	1,000	70	10	<10	<10	300	70	.2
20	700	30	150	<5	<10	20	20	1	3,000	50	10	<10	<10	1,500	150	1
10	200	10	200	<5	<10	10	20	1	1,000	70	20	<10	<10	200	70	.5
10	500	15	150	<5	<10	10	15	1	2,000	50	15	<10	<10	1,000	200	.3
Basaltic dikes																
20	1,000	100	200	7	20	300	50	<1	10,000	50	15	20	10	1,000	700	>5
20	1,000	70	200	5	15	150	30	1	7,000	70	15	10	10	700	300	1.5
<10	1,000	30	150	10	20	100	30	1	7,000	70	15	15	20	1,000	500	2
20	1,500	5	150	<5	10	150	20	<1	10,000	50	15	10	<10	1,500	700	2
30	1,000	70	700	300	30	150	15	5	20,000	150	<10	15	300	10,000	1,500	>5
DEATH VALLEY CANYON TO TUCKI WASH																
Felsitic rocks																
30	100	5	700	10	<5	10	30	1	2,000	150	50	10	20	1,500	300	0.5
20	700	15	300	20	5	20	30	1	3,000	70	30	<10	50	1,000	300	.7
20	700	15	300	20	7	50	20	<1	3,000	70	10	10	70	1,500	500	1
100	300	10	100	20	<5	10	20	1	1,000	<50	50	<10	50	500	70	.3
TRAIL CANYON TO TUCKI WASH																
Basaltic rocks																
10	1,000	100	150	50	20	150	20	<1	7,000	50	50	30	100	700	1,000	2
15	1,000	20	150	15	20	200	20	<1	7,000	50	50	20	50	1,000	1,500	5

NOTE.—All samples showed: As <1,000; Zn <200; Sn <10; Ge <20; Ga <20; Cd <50; Bi <10; In <10; Sb <200; Tl <100; Nb <30; Ta <50; W <100.

westward, emerge at the Argus Range 20 miles west of the Panamint Range, and, with more interruptions, continue its rise westward to the summit of the Sierra. The floor of the batholith dips westward as the surface rises, greatly increasing the thickness of the granitic rock.

By this interpretation the granitic mass under the Panamint Range would be a bulbous edge of the Sierra Nevada batholith, which thickens northeastward, and would be the source for the injected granitic intrusions represented by the granites at Skidoo and at Hanaupah Canyon.

A second possibility is that the edge of the batholith is west of the Panamint Range and that the range is underlain by several isolated granitic masses. These masses need not be connected with the batholith. If by palingenesis of Precambrian granitic and metamorphic rocks a magma was produced at moderate depth within the crust, the magma could have been the source for the physically injected granitic masses now exposed at the surface of the Panamint Range.

The third possibility is that the low-density rocks underlying the Panamint Range may be Precambrian granite, like that to the east under the Colorado Plateau. Cretaceous (or Tertiary?) intrusions could have been derived from the old granite. According to this interpretation, the boundary between the Precambrian and Sierran granites lies west of the Panamint Range, perhaps where the local reversal of the gravity field coincides with the northeast edge of the seismically active area that extends westward to the coast.

The great height of the Panamint Range combined with the occurrence of a negative gravity anomaly under it suggests that part of the height and uplift may be attributable to local isostatic adjustment. If so, the adjustment began in middle or late Tertiary time, after deposition of the volcanics that overlap the east side of the Panamint Range (p. A120; fig. 89). This was later, and presumably much later, than the time the Skidoo and Hanaupah granites were injected.

STRUCTURE OF PRECAMBRIAN AND PALEOZOIC ROCKS

The Paleozoic and upper Precambrian sedimentary rocks in the mountains bordering Death Valley occur in a series of folded thrust plates of the Amargosa thrust system. The thrusting moved younger rocks westward onto older ones.

Within a thrust plate the formations have rather uniform homoclinal dips, that, with very few exceptions, are eastward.

In terms of competency and density the formations involved in these structures can be considered in three major groups. At the base is the competent and dense Precambrian metamorphic complex, exposed in the Black Mountains and the west face of the Panamint Range.

Overlying the metamorphic complex is a section of upper Precambrian and lower Paleozoic formations aggregating about 30,000 feet thick and consisting of very thick incompetent shaly units separated by moderately thick competent units. At the base is the Pahrump Series (Crystal Spring Formation, Beck Spring Dolomite, and Kingston Peak Formation), the upper part of which is composed largely of incompetent beds. Above this series is the competent Noonday Dolomite, a thousand feet or more thick. The Noonday is overlain by the highly incompetent Johnnie Formation, a few thousand feet thick; above the Johnnie is the competent quartzite of the Stirling and lower part of the Wood Canyon Formation. At the top is shale of the Wood Canyon Formation, which is Early Cambrian.

The rest of the Paleozoic formations comprise a competent unit of limestone and dolomite aggregating about 25,00 feet thick.

Much of the thrusting here as in other parts of the region (see for example, Kupfer, 1960) has been along bedding planes in the incompetent upper Precambrian formations; this introduces difficulties in evaluating unconformities and determining to what extent evident cutting out of beds is attributable to the angular unconformities or to deformation. Also, it introduces difficulties in estimating displacement along the thrust faults.

At the base of the Noonday Dolomite and top of the Pahrump Series probably is a considerable angular unconformity (Hazzard, 1937b; Hewett, 1956; Kupfer, 1960), but in places this contact also may be a thrust fault. It is, to say the least, curious that in the eastern part of the Panamint Range the Noonday Dolomite rests on the Crystal Spring Formation of the Pahrump Series and the Kingston Peak Formation is missing, wheras on the west side of the range a carbonate formation thought to be the Noonday rests on the Kingston Peak(?) Formation which, in turn, rests directly on the Precambrian metamorphic rocks—the Crystal Spring Formation being cut out (fig. 6).

There are no major unconformities in the rest of the Paleozoic section up to the Permian, which contains thick breccias and conglomerates.

Deformation in late Paleozoic or early Mesozoic time in this region is recorded by an unconformity between Permian and overlying Triassic rocks and by the major change in lithology between them (Johnson, 1957). The Triassic formations, consisting of metasedimentary and volcanic rocks, 8,000 feet thick, are exposed just south of the area shown on plate 3, in Warm Spring Canyon and Butte Valley.

The thrust faulting and folding of the Amargosa system began before the granitic intrusions were emplaced in this area and continued long into the Tertiary. The faulting may have begun when the granitic intrusions west of here were being emplaced, perhaps during the Cretaceous. The igneous intrusions in this area invade the earlier faults.

TUCKI MOUNTAIN KLIPPE

In the Virgin Springs area, the type area for the Amargosa thrust, the upper plates are chaos, which in large part consists of Tertiary volcanic rocks. Tucki Mountain provides the prototype of faults in the Amargosa system that involve great plates of Paleozoic formations thrust onto Precambrian formations. The Tucki Mountain klippe, covering more than 35 square miles, consists of a slab of Paleozoic formations dipping steeply to the east and thrust westward onto less steeply dipping Precambrian formations. That the direction of thrusting was westward is shown by the way the thrust slices are shingled on one another (figs. 86, 109) and by drag along the faults. The general structure has long been recognized and is shown quite satisfactorily on the Death Valley sheet of the State geologic map (Jennings, 1958).

The Tucki Mountain thrust fault is exposed almost continuously for 15 miles—along Mosaic Canyon, across the top of Tucki Mountain, and along the north side of Tucki Wash. Rocks of the upper plate range in age from late Precambrian (Stirling Quartzite) to Permian.

A series of secondary thrusts branch upward from the main thrust and divide the upper plate into a series of shingled thrust plates. On each of these thrusts the direction of movement was westward and carried younger rocks in the upper plates over older rocks in the lower plates. The mountainside along the north side of Tucki Wash provides an excellent exposure of a natural cross section showing the structure; a good view of this can be had from Aguereberry Point.

The Tucki Mountain thrust fault dips northeastward about 20°. It crosses the top of Tucki Mountain at an altitude of 5,600 feet, and it is buried under the northeast foot of the mountain, parts of which are below sea level.

The strike and dip of the thrust fault is oblique to the strike and dip of the formations in the upper plate, which strike almost north and everywhere dip east more steeply than 45°. At the east foot of the mountain the Pennsylvanian and Permian formations are overturned (pl. 3). The secondary faults that divide the upper plate into separate shingled thrust plates in general

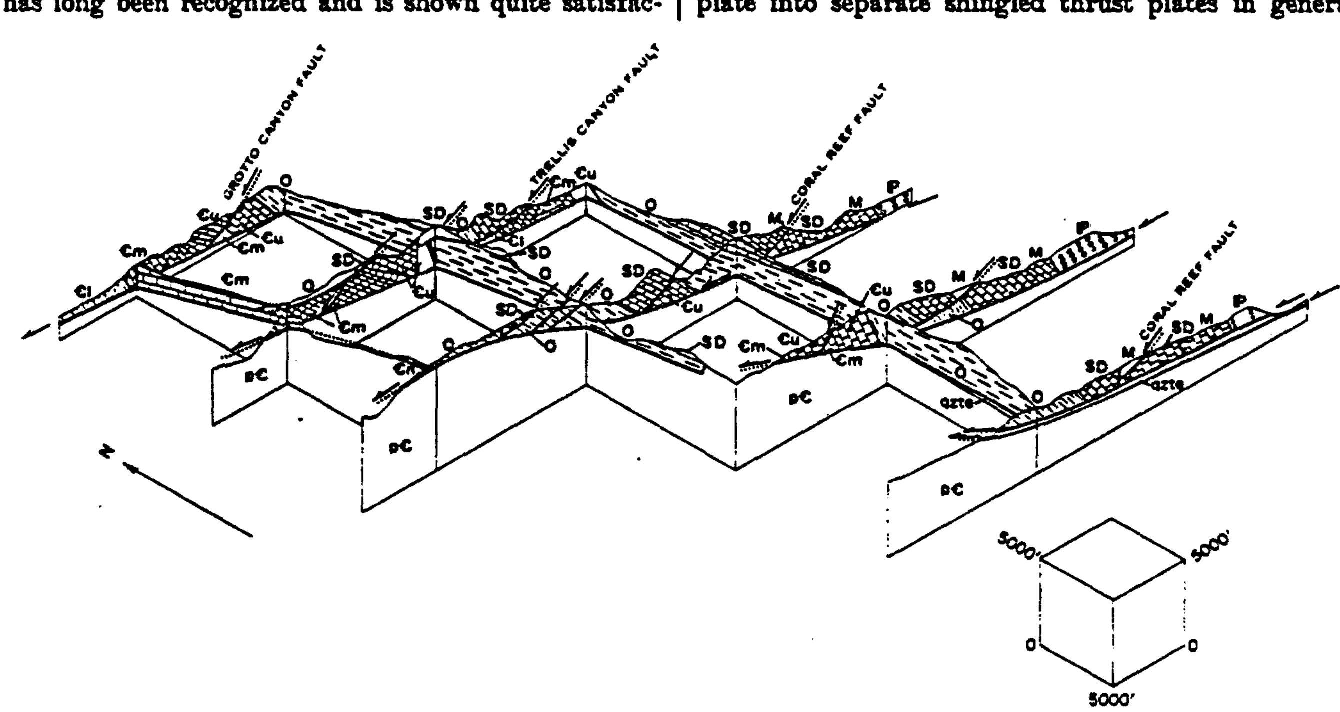

FIGURE 109.—Isometric fence diagram of the Tucki Mountain klippe. The Tucki Mountain thrust fault (heavy black line) has moved Paleozoic formations westward onto the Precambrian. pC, Precambrian formations (Kingston Peak(?) Formation and Johnnie Formation; qzte, a brecciated quartzite at the thrust at the south edge of the klippe (probably Stirling Quartzite); Cl, Lower Cambrian; Cm, Middle Cambrian; Cu, Upper Cambrian; O, Ordovician; SD, Silurian and Devonian; M, Mississippian; P, Pennsylvanian.

strike north-northwest, a direction that is intermediate between the strike of the main fault and that of the formations in the upper plates.

The principal secondary faults that divide the upper plate are the Grotto Canyon, Trellis Canyon, and Coral Reef faults. The Grotto Canyon fault, the most westerly of the three, has moved Ordovician, Silurian, and Devonian formations westward onto Middle Cambrian. The displacement is nearly 2 miles to the west.

The Trellis Canyon fault also carries Ordovician, Silurian, and Devonian formations about 2 miles westward onto the Middle Cambrian. Part of this fault is represented by an exhumed surface on the lower plate along the north side of Tucki Wash, 2 miles south of the trace of the fault. This exhumed surface covers about 1 square mile and, when viewed from Aguereberry Point, appears as a level skyline. The topographic contours show it well (fig. 110).

Displacement on these branch faults totals at least 4 miles to the west; displacement on the main fault must be very much greater than that.

The most easterly of the secondary faults, the Coral Reef fault, carries Mississippian rocks about a mile westward onto the Devonian. This fault is branched southward.

At the east foot of the mountain the dips are greatly steepened and locally overturned. The stratigraphic

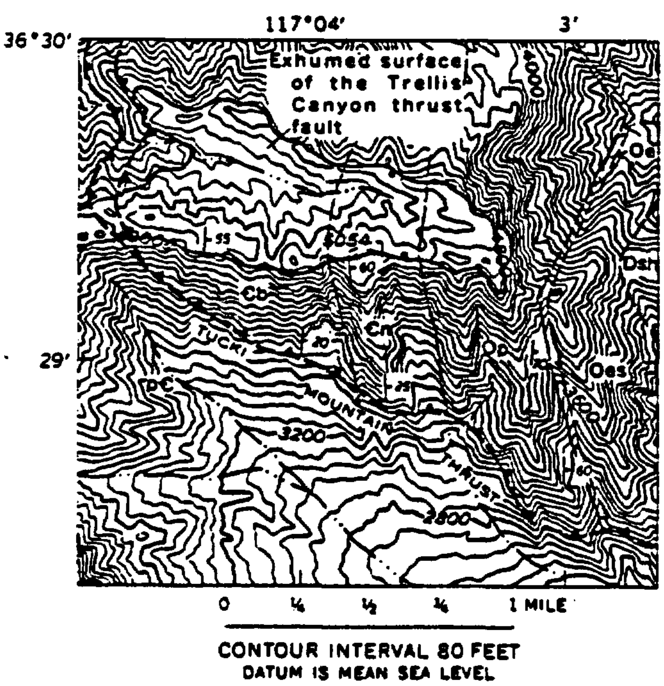

FIGURE 110.—Map of the exhumed surface of the Trellis Canyon fault at the south side of Tucki Mountain. pC, Precambrian under the Tucki Mountain thrust fault; Cb, Bonanza King Formation; Cn, Nopah Formation; Op, Pogonip Group; Oe, Eureka Quartzite; Oes, Ely Springs Dolomite; DSh, Hidden Valley Dolomite. Topography from Emigrant Canyon quadrangle map.

section in this part of the upper plate is confused, because there has been considerable faulting along the bedding.

TUCKI MOUNTAIN FENSTER

The west and south sides of Tucki Mountain are composed of the late Precambrian Kingston Peak(?) Formation folded into a dome of which the high part is approximately at the mountain top. From the top the beds dip northeastward under the Tucki Mountain thrust fault, westward towards Emigrant Wash, and southward towards the granite at Skidoo (pl. 3). The fenster is a dissected turtleback. Structural relief on the dome is greater than the topographic relief, that is, something more than 7,000 feet.

Capping many of the divides on the west slope of the turtleback are small klippes of carbonate rocks, probably from the Noonday(?) Dolomite but possibly from the Paleozoic. The thrust faults under these klippes are approximately concordant with the dip of the underlying Precambrian strata.

The turtleback ends westward at the Emigrant Spring fault where the Funeral Formation has overlapped the turtleback surface and later has been downfaulted along the overlap contact (p. A114; fig. 86).

Extending southeastward from the top of Tucki Mountain is a steep east-dipping monocline of Noonday(?) Dolomite and overlying Johnnie Formation. The monocline emerges from beneath the Tucki Mountain thrust fault where that fault is joined by the Trellis Canyon fault. The monocline turns southward at Tucki Wash and extends across the head of Blackwater Wash. It approximately parallels the granite at Skidoo and probably marks its eastern limit.

Evidence is conflicting about the relative age of the monocline, the Skidoo granite that is thought to have produced it, and the Tucki Mountain thrust fault. The monocline extends northward under the thrust, but it merges there with the east flank of the dome of the autochthonous rocks. The fact that the monocline is partly cut off by the thrust does not necessarily mean the thrust is later, because incompetent beds may have been folded under competent ones. Further, the westward offset of the Tucki Mountain thrust across the monocline suggests that the monoclinee is later. Detailed mapping would determine the relations, but this mapping is still to be done.

SOUTHERN PART OF THE PANAMINT RANGE

Fault blocks on the east side of the Panamint Range south of Tucki Mountain can be divided into two major units. From Starvation Canyon southward to Warm Spring Canyon, 1½ miles south of the area mapped (pl. 3), the formations are mostly late Precambrian. They dip east and are much broken by high-angle faults. This area is the upper plate of a thrust fault, probably one that is exposed in Warm Spring Canyon. From Starvation Canyon northward to Blackwater Wash the rocks are Paleozoic. They, too, dip east but are broken by west-dipping faults, all of which involve thrusting towards the west and are part of the Amargosa thrust system.

At the head of Galena Canyon, Precambrian schist is overlain by the Crystal Spring Formation of the Pahrump Series. The contact appears to be depositional. Above the Crystal Spring Formation is the Noonday Dolomite; but the contact at its base, which is poorly exposed, may be a thrust fault. Overlying the Noonday is the Johnnie Formation and that contact appears to be depositional, but there are thrust faults both within the Johnnie Formation and at the base of the Stirling Quartzite which is on top of it.

Where rock formations are very much sheared and faulted, it is difficult to represent the dips on a general map. For example, on figure 111, a series of individual fault blocks may each dip 45°, but because of repeated small faults, too small to show, the effective dip is only 30°. On the map (pl. 3), strikes and dips are recorded as measured, but cross sections accompanying this report show the effective dips.

Most of the faults in this part of the area are high-angle faults and are closely spaced. The principal ones trend northwest or north. North of Six Spring Canyon, dips are towards the east; south of the canyon the formations are domed. This doming, which shows well in Warm Spring Canyon too (C. T. Wrucke, oral commun., 1961), may be due to doming over a granitic intrusion in Warm Spring Canyon.

Shearing in these formations is reflected in fracture

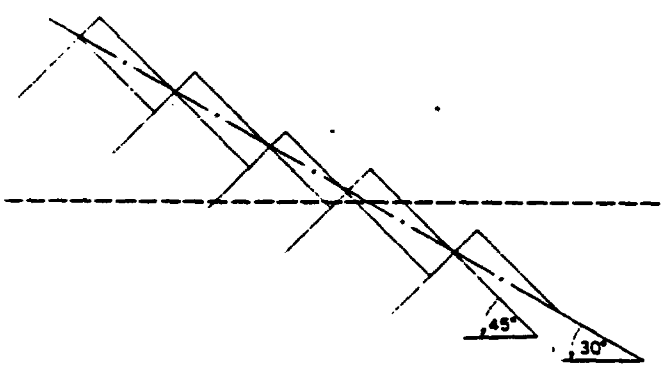

FIGURE 111.—Diagram illustrating difference between actual dip and effective dip in a series of fault blocks. The individual blocks dip 45°, but because of repeated small faults, the effective dip is only 30°.

cleavage and discordances along the contacts (fig. 112). The fracture cleavage is nearly horizontal across beds that dip east. The cleavage is about parallel to the faults of the Amargosa thrust system. The relationships are suggestive of those at overturned folds, but the fracture cleavage here evidently is related to the thrust faulting. The beds are not overturned.

North of Starvation Canyon the formations in the mountains are mostly Paleozoic and dip 45°–60° E. High on the mountain, structurally and stratigraphically below these formations, are the Stirling Quartzite, Johnnie Formation, and Noonday(?) Dolomite also dipping east. These east-dipping formations are broken by numerous west-dipping faults along which the upper plates have moved westward. Maximum displacement on these faults is about a mile. They appear to be branches of the Amargosa thrust fault, which is exposed along the east foot of the Panamint Range (p. A129).

Best known of the branch faults is the Burro Trail fault (figs. 113–116). It first appears at the south at Starvation Canyon where a tear fault (fig. 117) that dips steeply northward has been traced 3 miles westward to the west edge of the Bennetts Well quadrangle. This fault may extend along the south edge of the granite at Hanaupah Canyon, which is 3 miles farther west.

North of the mouth of Starvation Canyon the tear fault divides. An upper branch turns northward and flattens into the Burro Trail fault. The lower branch continues northeast and probably connects with the Amargosa thrust where that fault is buried by fan gravel. Eruptives and volcanics, with some admixed Paleozoic or Precambrian rocks suggestive of chaos, occur along the southeast side of this branch of the tear fault. Also, volcanic rocks have been faulted onto the Wood Canyon Formation in the upper plate north of the tear fault (fig. 117).

The Burro Trail fault is well exposed and intruded by monzonite porphyry sills for 6 miles to the north (pl. 3 and fig. 114). The sills extend to the divide north of Hanaupah Canyon, but they are not found north of there. The composition of these sills, together with their distribution in a belt directly east of the granite at Hanaupah Canyon, leaves little doubt that they are connected with that granitic intrusion (fig. 90).

North of Hanaupah Canyon the Burro Trail fault is broken by a series of north-trending high-angle faults that are downthrown to the east (pl. 3). North of this disturbed area the Burro Trail fault separates into several branches, but these join again north of Death Valley Canyon (fig. 113–115). North of Trail Canyon

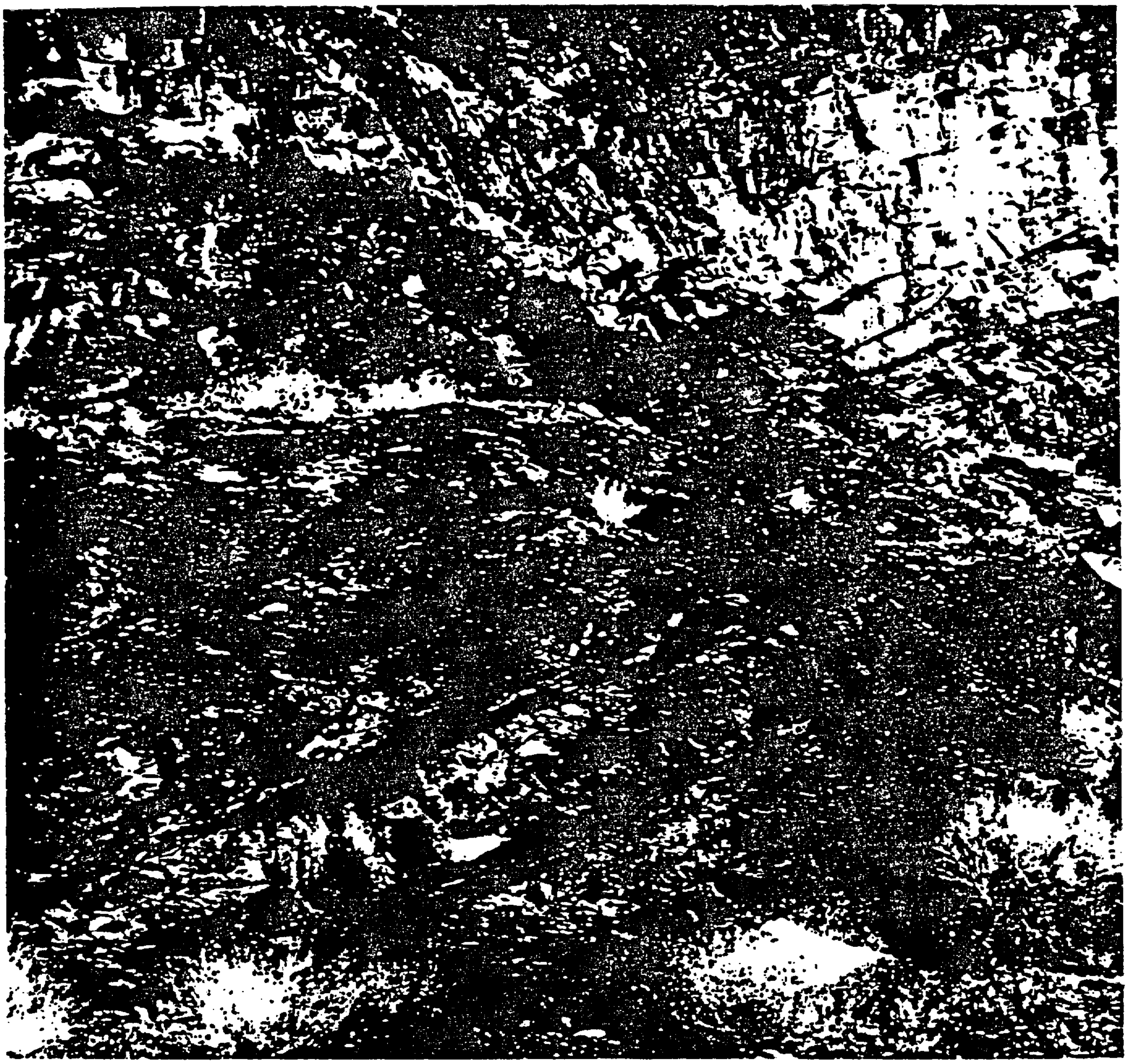

FIGURE 112.—Horizontal fracture cleavage in east-dipping shale member of the Stirling Quartzite between Hanaupah and Starvation Canyons. Looking south. The cleavage approximately parallels the faults of the Amargosa thrust system; the beds are not overturned.

the fault again divides into branches along which the displacement commonly is no more than a thousand feet.

On all these faults the hanging-wall rocks are the younger, and the displacement therefore is that of normal faults. Possibly these faults were originally high-angle normal faults that later were rotated to dip gently west. But although it is true that the Panamint Range has been rotated east (p. A137), and the faults originally must have dipped moderately steeply to the west, the structures along these faults are not those of high-angle normal faults. Crushing along them is much more intense than along the high-angle faults in this region, and locally the upper-plate rocks are almost concordant on the fault, as in Trail Canyon (fig. 116).

The Burro Trail fault, although a striking feature, probably is only a minor fault in the Amargosa system. It probably connects at depth with the Amargosa thrust under the Panamint Range and branches upward from

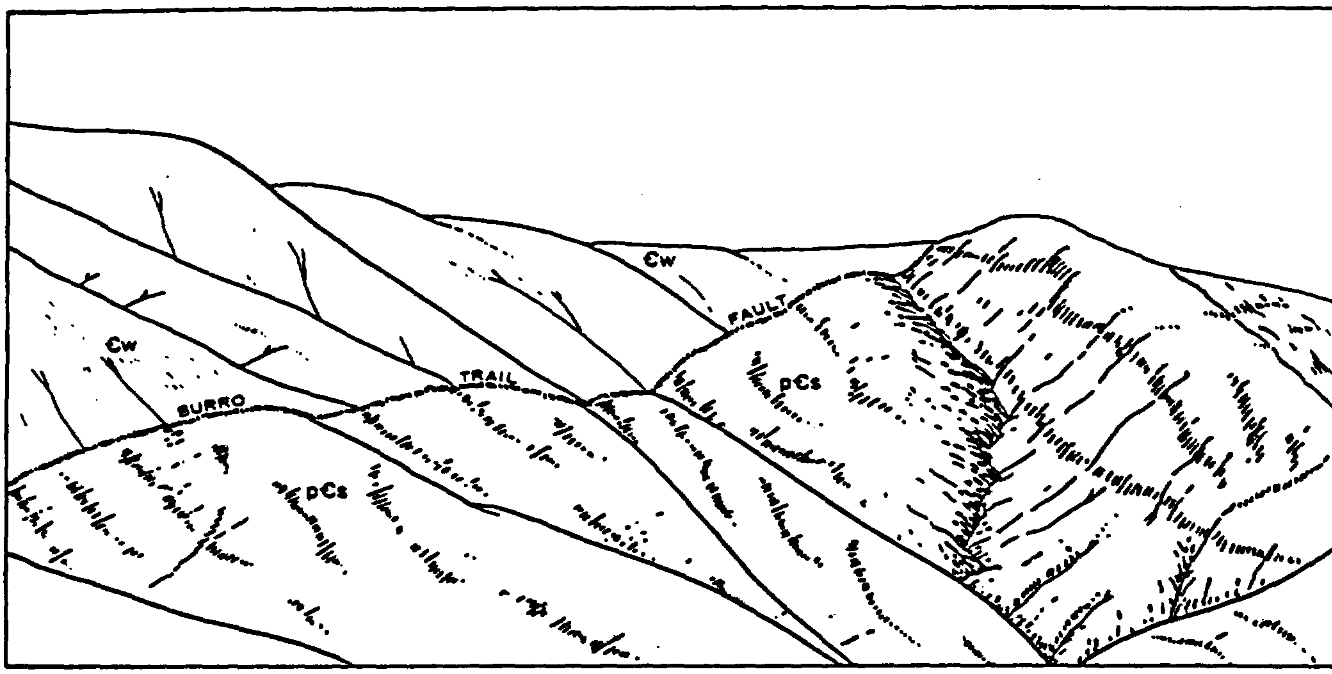

FIGURE 113.—Burro Trail fault. View north from the divide north of Starvation Canyon. The Wood Canyon Formation (Cw) has been faulted westward onto Stirling Quartzite (pCs). Both formations dip about 25° E.; the fault dips about 30° W.

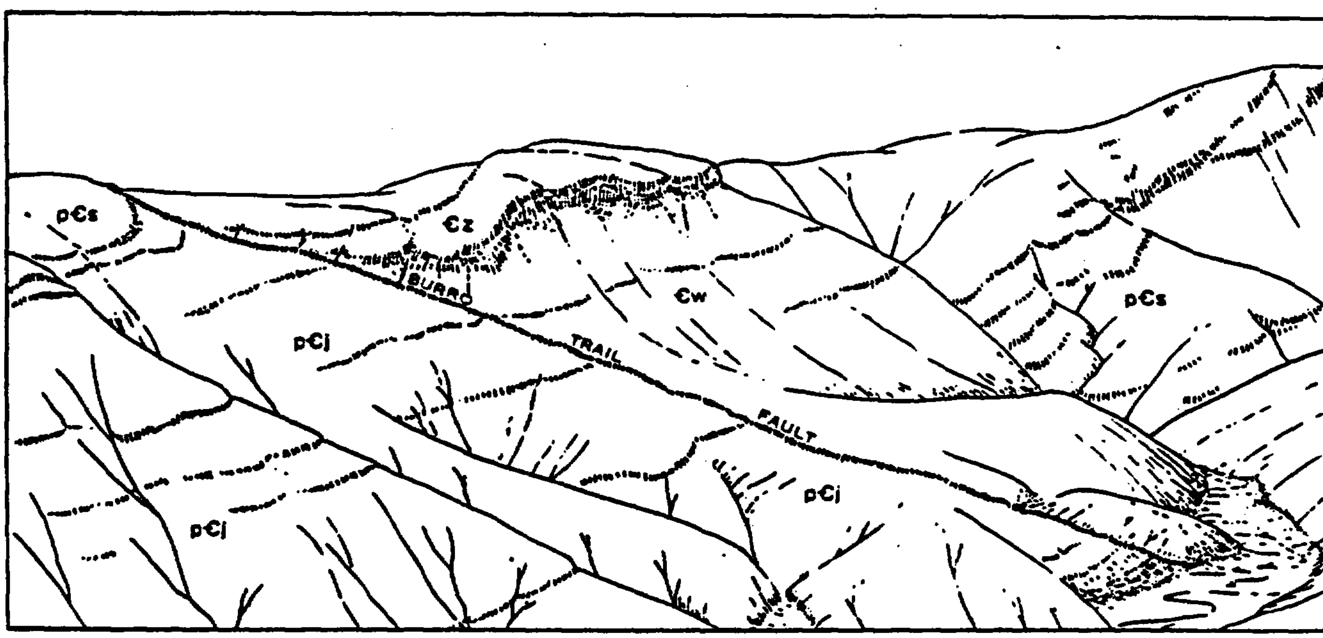

FIGURE 114.—Burro Trail fault. View south across Hanaupah Canyon. Stirling Quartzite (pCs) at the right and overlying Wood Canyon Formation (Cw) and Zabriskie Quartzite (Cz) are faulted westward onto Johnnie Formation (pCj) and Stirling Quartzite. The formations dip about 25° E.; the fault dips 20° W. Small sills of monzonite porphyry are intruded along the fault.

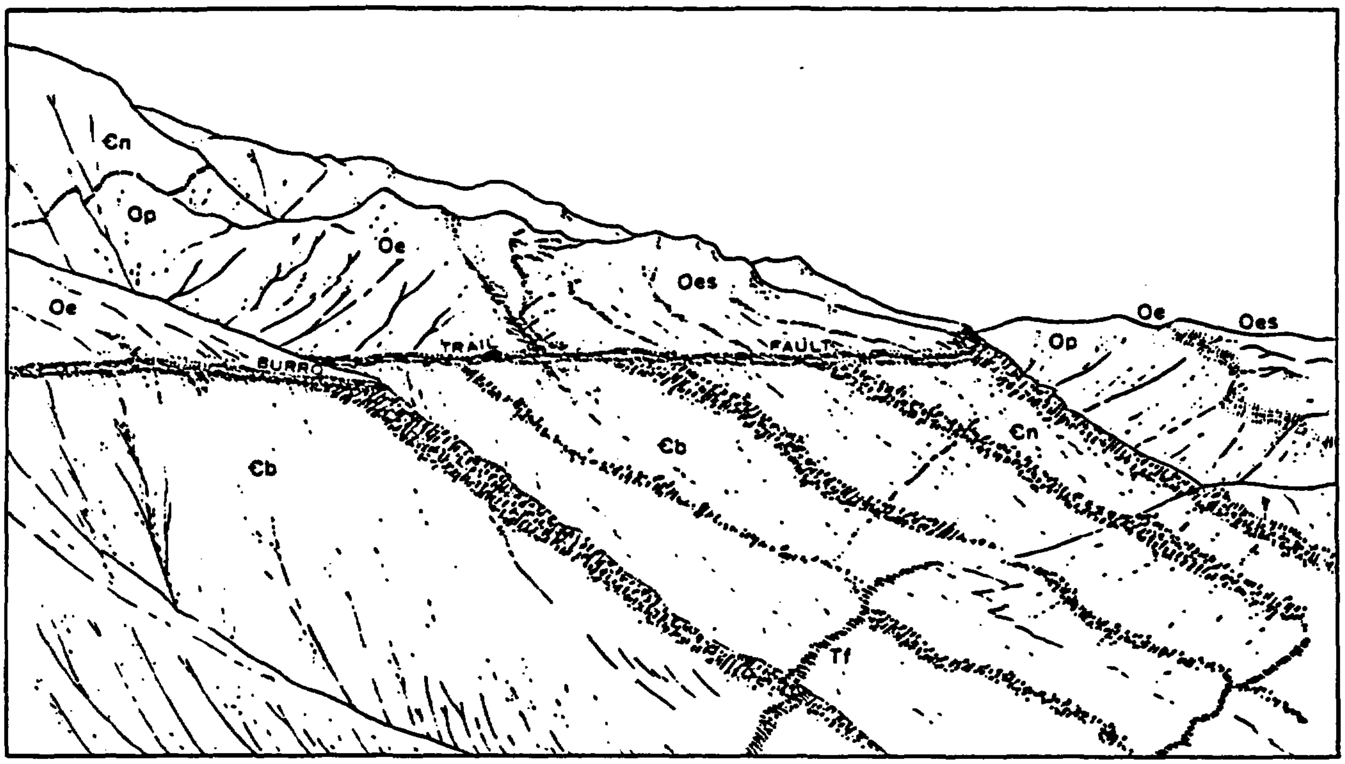

FIGURE 115.—Burro Trail fault. View north between Death Valley and Trail Canyons. The upper plate in this view is mostly Ordovician; it has moved westward on top of Middle Cambrian. Єb, Bonanza King Formation; Єn, Nopah Formation; Op, Pogonip Group; Oe, Eureka Quartzite; Oes, Ely Springs Dolomite. Felsitic dikes (Tf) intrude the lower plate.

it, as the Grotto Canyon, Trellis Canyon, and Coral Reef faults branch upward from the Tucki Mountain thrust fault (fig. 109).

BLACK MOUNTAINS FENSTER

The Black Mountains expose the domed lower plate of the Amargosa thrust. Precambrian metamorphic rocks which form the lower plate are exposed for 20 miles northward from Mormon Point to the north end of the Badwater turtleback (pl. 3). The Amargosa complex exposed along the east foot of the Panamint Range north of Hanaupah Canyon (p. A129) represents the domed lower plate of the thrust on the west side of Death Valley, and probably is the westward continuation of the faulted anticlines represented by the Black Mountains.

The Amargosa thrust fault is exposed along the east side of the Black Mountains. The fault was first recognized and named in the Virgin Springs district a few miles southeast of Mormon Point (Noble, 1941).

At the north the formations at Artists Drive overlap and are faulted against the Precambrian metamorphic rocks that comprise the Badwater turtleback. The Black Mountains between the Virgin Springs district and the Badwater turtleback have been mapped by Drewes (1963).

The folding and faulting of the Black Mountains has produced three turtlebacks that have been described by Curry (1938b, 1954). Axes of the turtlebacks plunge northwestward into Death Valley at Mormon Point, Copper Canyon, and Badwater. These structures are oblique to the northward-trending block mountain, but they are reflected in Quaternary structures on the floor of Death Valley (p. A100). The folds seem to have continued to develop through the stage of block faulting.

In brief, the anticlinal uplift of the Black Mountains evidently began during middle Tertiary time by folding along the northwest-trending axes represented by the turtlebacks. Uplift continued into Qaternary time when block faulting along the faults trending north-south further raised the mountain block (p. A116).

FUNERAL MOUNTAINS FENSTER AND FUNERAL MOUNTAINS KLIPPE

The Funeral Mountains comprise two very different structural units. At the north is Precambrian Kingston Peak(?) Formation in an anticlinal uplift having a northwest-trending axis and representing a fenster between thrust plates in the Grapevine Mountains to the north and in the southern part of the Funeral Mountains (pl. 3). The fenster is a dissected turtleback

FIGURE 116.—View of Burro Trail fault on the south side of Trail Canyon. Nopah Formation in the upper plate lies almost horizontally on the almost horizontal thrust fault. The lower plate is Bonanza King Formation.

much like the west side of Tucki Mountain. The turtleback surface approximately conforms to the dip of the Precambrian formations and to the dip of the thrust faults.

The northern end of the fenster is at Boundary Canyon, a mile north of the area mapped, where a thrust fault carries Lower Cambrian formations westward onto the Precambrian. The southwest edge of the fenster is at the Keane Wonder fault which trends northwest along the foot of the mountain and dips 25°–40° toward Death Valley. At Hells Gate the hanging wall is Lower Cambrian overlain by beds correlated with the Titus Canyon(?) Formation of Stock and Bode (1935), but from there southeast to the edge of the quadrangle the hanging wall consists of crushed and contorted beds of the Titus Canyon(?) Formation (fig. 118).

The age of this faulting is equivocal. At least part of the displacement is younger than the Titus Canyon(?) Formation, which is faulted against the Precambrian. But the structural geology is very similar to that along the west side of the Tucki Mountain turtleback where the Funeral Formation has overlapped and has been faulted against that turtleback surface (p. A114). Along the southwest side of the Funeral Mountains fenster the Titus Canyon(?) Formation is similarly faulted against both the upper and lower plates, and probably the thrust faulting occurred before the Titus Canyon(?) Formation was deposited. The Titus

FIGURE 117.—View north across tear fault in Starvation Canyon. pCs, Stirling Quartzite; Cw, Wood Canyon Formation; T, Tertiary volcanic rocks faulted onto the Wood Canyon. This tear fault marks the southern end of the Burro Trail fault.

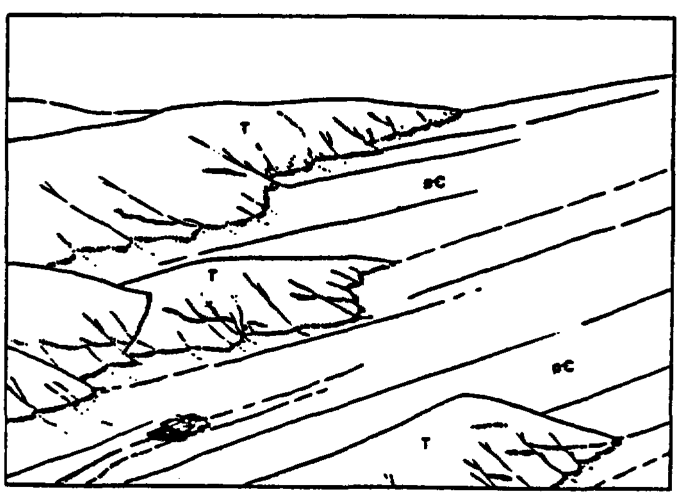

FIGURE 118.—View northwest along the turtleback fault surface marking the Keane Wonder fault. Crushed and contorted beds correlated with the Titus Canyon(?) Formation of Stock and Bode (1935), (Tt), are faulted against the smooth turtleback surface on the Precambrian formations (pC).

Canyon(?) Formation may have overlapped both the upper plate and the turtleback. By this interpretation the faulting of the Titus Canyon(?) Formation against the turtleback along the Keane Wonder fault is attributable to renewed anticlinal folding of the fenster.

Volcanic rocks have been reported in thrust breccia along what is presumed to be the Keane Wonder fault east of this area (James Gilluly, oral commun., 1960). Some possible relationships between the thrust faulting and early volcanism have already been discussed (p. A137). Whether the occurrence along the Keane Wonder fault is evidence of early volcanism or of late movement along the fault remains unresolved.

Along the southeast side of the Funeral Mountains klippe the overthrust plates are folded steeply downward toward Death Valley. East of Nevares Peak is a north-trending high-angle fault that drops Middle Cambrian formations down on the west against the Lower Cambrian and Precambrian (fig. 119).

At the north base of Nevares Peak, the Titus Canyon(?) Formation is downfaulted against a smooth

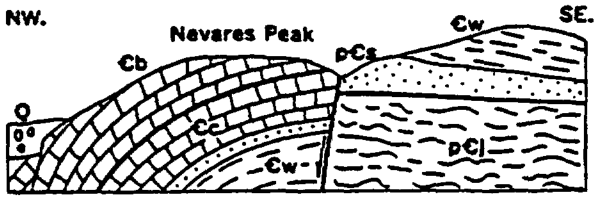

FIGURE 119.—Section of west end of Funeral Mountains klippe. Middle Cambrian formations at Nevares Peak are downfolded and downfaulted into Death Valley and downfaulted against Lower Cambrian and Precambrian formations on the east. pCj, Johnnie Formation; pCs, Stirling Quartzite; Cw, Wood Canyon Formation; Cc, Carrara Formation; Cb, Bonanza King Formation; Q, Quaternary gravel. Length of section, 3 miles; vertical scale not exaggerated.

fault surface on the Cambrian formations (p. A118). The downfolding and faulting of these formations into Death Valley seem to have antedated the Titus Canyon(?) Formation, but there was later movement, too, as there was along the Keane Wonder fault (fig. 118). Beds tentatively correlated with the Titus Canyon(?) Formation, on the southeast spur of the Funeral Mountains klippe, have the same dip as the underlying Paleozoic formations (C.S. Denny, oral commun., 1961); the dips in that part of the klippe are post-Titus Canyon(?), indicating eastward rotation of the upper plate like that in the Panamint Range.

The relationships at Nevares Peak can be interpreted as due to vertical movements with little or no horizontal displacement along the high-angle fault, but another interpretation could be that the Cambrian formations under Nevares Peak have moved laterally 2 miles northwestward from where these formations form the upper plate of the thrust and extend to the mountain front.

The Mont Blanco fault, on which there has been Pleistocene movement (p. A115), may extend northeastward through the Funeral Mountains klippe, for the Ryan quadrangle map, just to the east, show a conspicuous trough extending northeastward through that part of the Funeral Mountains to the Amargosa Desert.

SOUTHERN GRAPEVINE MOUNTAINS

The southern part of the Grapevine Mountains (pl. 3) is composed of Cambrian formations in one or more thrust plates resting on Kingston Peak(?) Formation and Noonday(?) Dolomite at the north end of the Funeral Mountains fenster. The thrust fault is arched anticlinally across the north end of the Funeral Mountains; the anticlinial axis trends about northwest.

At Hells Gate the thrust fault exposed in Boundary Canyon either is cut off by a high-angle fault, down to the west, or the thrust there dips steeply west. The structure is buried under Quaternary gravel, but it probably is like that at Nevares Peak (fig. 119) where the Funeral Mountains klippe is dragged steeply downward into Death Valley.

ORIGIN OF THE STRUCTURAL FEATURES

It still does not appear possible to offer a satisfactory explanation for the structural features in the Death Valley area. The structure of the thrust slices in the Amargosa thrust system is very similar to that of a group of landslid blocks (Mason, 1948). The similarities are so great that one feels the thrust must have moved by gravity (Sears, 1953). But none of these thrust slices has yet been found overriding surficial deposits attributable to an old land surface, so that the parts of the faults of the Amargosa system that are now exposed are presumed to be subsurface features.

We do not refer to these thrust slices as megabreccia (Longwell, 1951), but prefer to restrict that term to fractured blocks that can be related to an existing or a preexisting land surface like the clearly landslid megabreccia at the east foot of the Funeral Mountains (C. S. Denny, written commun., 1965). A similar distinction has been noted by Jahns and Engel (1950) and by Kupfer (1960). It is not possible, with present information, to relate the thrusts in the Panamint Range to a preexisting surface. Etymologically, of course, the range is a megabreccia but so, too, is any other broken-up part of the earth's crust.

The parent block from which the Tucki Mountain thrust plates were torn must have contained a virtually complete section of Paleozoic formations, one above the other in an orderly plateaulike structure, because 30,000 feet of Paleozoic formations are represented in the thrust blocks, and their dips are homoclinal to the east. The angle between the dipping beds and basal faults (such as the Tucki Mountain and Amargosa faults) is about 45°. The angle between the beds and the faults that branch upward from the basal ones approaches 90°. The homoclinal eastward dip of the beds and their angular relation to the faults could not have been obtained from a source that was complexly folded or faulted.

It is difficult to believe that there ever was a 30,000 foot escarpment facing west with a plateau that high east of it. It seems more likely that the thrust faults that are exposed represent segments of faults that extended downward into the crust.

Since the thrust faulting occurred, the Panamint Range has been rotated at least 20° to the east, because this is the dip of the Tertiary eruptives that overlap thrust blocks along the east side of the range. It is supposed, then, that when the thrust faults formed they had an average dip westward of about 45°. The faults are concave upward, so that the dip would be flatter at depth and to the west and steeper upward and to the east. The dip of the formations in the upper plates need have been no more than a third as great as the present dip, somewhere around 5°–10° E. in the southern part of the range and 10°–20° E. in the northern part.

The rocks in the upper plates are spread over more area today than they were before they were faulted to their present position. That this is so can be determined by measuring the length of individual beds in individual fault slices (see fig. 119 or pl. 1). These show 4 miles of displacement within the upper plate, between the westernmost and easternmost thrust slices, and with less than 1 mile length of displaced beds. The deficiency can only be made up by imagining it all in a block to the west that has since been destroyed by erosion. The length of individual beds averages only about 20 percent of the length of the area across which they are spread.

Faults like those in the Amargosa system are numerous across the Death Valley subsection (Longwell, 1945), and they appear very similar to some reported at the edge of the Colorado Plateau (fig. 120) that apparently were derived from a plateaulike block.

There is considerable evidence suggesting that the thrust faulting was in some way related to the intrusion of the granites and to the volcanism, because igneous activity accompanied the faulting. This relationship is reported in the Silurian Hills too (Kupfer, 1960).

Many features of the Amargosa fault system are suggestive of the detachment thrust faults described by Pierce (1957) in Wyoming. The similarities are that the thrust blocks have moved onto older rocks along thrust faults that have no roots, and the geometry strongly suggests that gravity was important as a propelling force. Also, there is a suggestion that both may have been triggered by volcanic or other igneous activity. A major difference is that the detachment thrusts in Wyoming must have moved onto a ground surface, but the faults of the Amargosa system seemingly extended downward. Despite this, the temptation is strong to attribute the turtleback fault surfaces to detachment thrusting and tectonic denudation.

Perhaps detachment thrusts developed from the upper plate of the Amargosa and other thrusts. The detached blocks could have moved off the areas now represented as fensters and be buried in the valleys under the fill. The block off the Funeral Mountains fenster would be composed of Paleozoic rocks and would have had to become detached before the Titus Canyon(?) Formation was deposited; a similar block of Paleozoic rocks off the west side of Tucki Mountain would have had to be detached before the Funeral Formation was deposited in Emigrant Wash. The blocks off the turtlebacks on the Black Mountains would be composed of chaotic blocks of volcanic and Paleozoic rocks.

The suggestion requires that all the detached blocks slid westward, but this direction of displacement would be favored in any case because the thrust faults converge westward. Slight movement eastward and valleyward of the Tucki Mountain klippe may be indicated by the increasingly steeper dips in the thrust plate eastward toward the base of the mountain. At the east foot of Tucki Mountain the Permian beds are overturned.

Such might be the explanation for the turtlebacks; we still need to try to explain the main Amargosa thrust.

In Late Cretaceous time the Death Valley subsection must have been high land contributing its share of sedi-

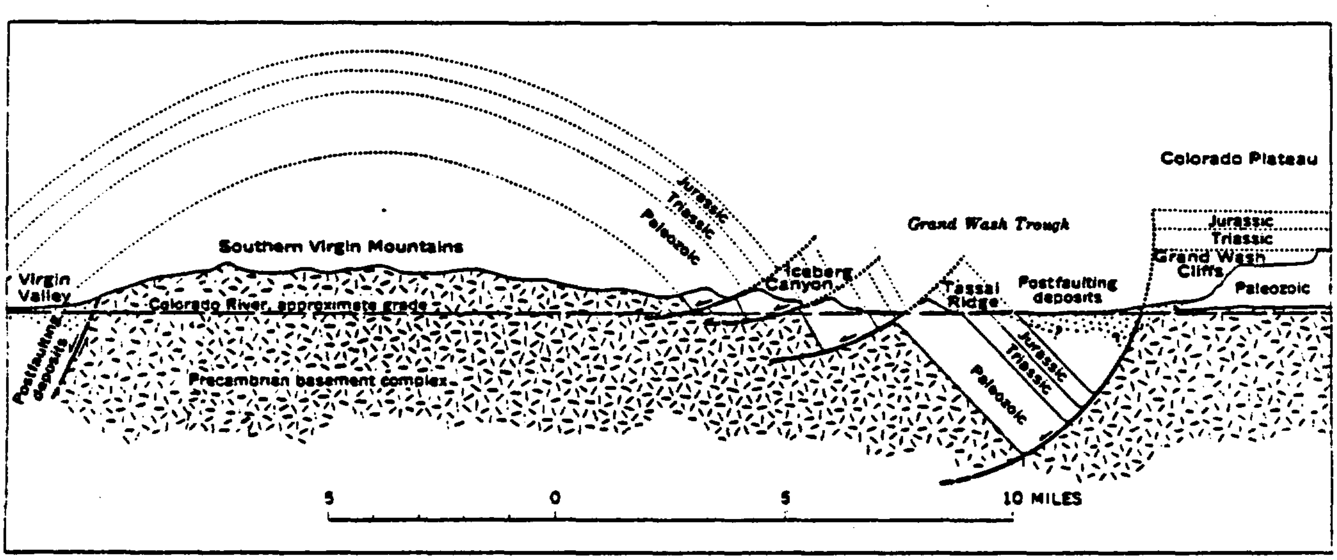

FIGURE 120.—Low-angle faults at west edge of the Colorado Plateau have displacements like those of the Amargosa thrust. From Longwell (1945).

ments to the geosyncline that extended eastward a thousand miles from somewhere along the east edge of what is now the Great Basin. There may have been a narrow belt of folding and thrusting at the position of the Keystone and other thrust faults 60 miles east of Death Valley, but from there westward must have been a simple plateau-type structure that could serve as source for the homoclinal dips in the upper plates of the Amargosa thrust faults. Perhaps the Keystone thrust had not yet developed; it may be younger than has been assumed.

The high part of this ancient plateau may have been across the section that lies between Death Valley and the folded belt to the east. In Late Cretaceous time, while part of the granitic batholith was being formed to the west, the site of Death Valley was on the west flank of this broad arched highly elevated plateau. The imagined plateau had to be high enough to contribute sediments to the geosyncline to the east, and its structure had to be simple enough to provide homoclinal dips in the fault blocks moved westward from it.

Longwell suggested (1945) that the flat faults in the region were due to collapse of an anticline (fig. 120), but there do not seem to be adequate west flanks for such folds. The dips are homoclinal to the east and are largely, if not wholly, those attributable only to an eastern limb. We visualize, instead of an anticline, a series of plateaulike blocks all tilted towards the east. In effect, these would comprise a series of west-facing and east-dipping blocks, perhaps continuous with the Colorado Plateau. The blocks would be normal faults, downthrown to the west. The westernmost part of the Colorado Plateau, between the Grand Wash Cliffs and the Hurricane fault, may be such a block still attached to the plateau.

Assuming this approximates the structural setting from which the Death Valley subsection evolved, we visualize a high plateau surfaced with flat-lying Triassic and Jurassic sediments. But all the formations thicken greatly westward. Along the belt of the Keystone thrust the Triassic and Jurassic rocks are about 4,000 feet thick, the Paleozoic rocks are about 9,000 feet thick, and the late Precambrian sedimentary formations are thin or absent (Hewett, 1931, p. 9). In the Death Valley area the Triassic is 8,000 feet thick (p. A50), and there was an unknown thickness of additional Triassic and Jurassic above this. The Paleozoic formations are 30,000 feet thick, and the late Precambrian are at least 13,000 feet thick. In the 60 miles from the Keystone thrust to Death Valley the section thickens westward about threefold. The base of the sedimentary formations would have dipped west almost 10°.

If this ancestral plateau be divided into major blocks by fractures that extended to the base of the crust, the high block of the series may have been the one at the present area of high gravity values extending from Death Valley to the folded belt 60 miles to the east. The Keystone thrust would be one of the fractures. Its extension into the lower part of the crust could very well cause remobilization of the Precambrian granite which would become forcibly intruded upward along the fault and along the secondary fractures branching from it in the shallower levels of the crust. Such could well have been the history of the extensive Teutonia Quartz Monzonite which spread along thrust faults in the Ivanpah quadrangle (Hewett, 1956, p. 61–63).

A second major fracture extending to the base of the crust may have been at the west edge of a block where the present seismic and gravity data indicate that the base of the crust begins sloping steeply westward under the Sierra Nevada (p. A139). This would be along a line trending northwest about at the Panamint Range.

This fracture, too, could have been the cause or effect of remobilization of Precambrian granite in the deep levels of the crust and upward movement of that granite to form the intrusions that are clustered along that belt and to the west.

From this fracture west to the Sierra Nevada, the base of the crust, or layers deep within the crust, slope westward perhaps as steeply as 15°. We already have noted a 10° westward dip at the base of the sedimentary part of the crust. With such slopes, the structural setting would seem to be favorable for westward sliding on fractures as they became lubricated by igneous activity.

SUMMARY OF THE STRUCTURAL FEATURES

The principal structural features of Death Valley were developed during late Mesozoic and Cenozoic time. There had been structural deformation during late Precambrian, Permian, and Permian and Triassic time, but little is known about the kind and extent of the ancient structures that were developed. The late Mesozoic and Cenozoic structures are of four principal kinds and developed at four principal stages:

1. Stage 1 was in Cretaceous, or possibly Jurassic time. The area is visualized as being a high plateau that became fragmented by northwest-trending rifts extending to, or nearly to, the base of the crust where granitic intrusions rose along the rifts. The base of the crust slopes west. As the lower parts of the rifts, downslope, became lubricated, the blocks began sliding westward. This began the Amargosa thrust system; later the thrusting was accompanied by folding.
2. Stage 2 was in early Tertiary time. During this stage the granitic intrusions reached the upper

levels of the crust, spread along the thrust faults, and the thrust plates above the intrusions became domed.

3. Stage 3 was in middle Tertiary time. During this stage folding continued, the igneous activity developed into volcanism. Blocks of the upper plates of the thrust faults became detached, slid off the uplifts, and left the lower plates exposed.
4. Stage 4 is the period of Basin and Range type block faulting; this began in middle Tertiary time and is continuing.

Thrust faults of the first stage strike northerly or northwesterly and involve the thrusting (sliding) of younger rocks westward onto older ones. The folds that developed during the latter part of this stage and during stages 2 and 3 are asymmetrical with their steep flanks facing west. Fold axes trend northwest; all the lesser ones plunge northwest. The axes of these lesser folds are arranged en echelon in northward-trending belts.

The granitic intrusions that were injected during stage 2 spread along the thrust faults. The long axes of the intrusions and of the anticlines over them parallel the strike of the thrust faults.

During stage 3 folding continued, and probably there was continued thrust faulting. During this stage, though, blocks of the upper plates became detached from the steep west flanks of the folds, slid into the synclinal valleys, and left the lower plate rocks exposed as westward-sloping turtleback surfaces.

The earliest block faults, illustrated by those of the Furnace Creek fault zone, trend northwest and are downthrown on the southwest side. These faults further accentuate the asymmetry of the northwest-trending folds as if the folding of the steep flanks progressed to the point of fracturing. This faulting probably began before the deposition of the Titus Canyon(?) Formation of Stock and Bode (1935), but we cannot be sure how early these synclinal basins of deposition became faulted. However, the faulting and the downfolding of the northwest-trending synclines has continued to the present.

Block faulting during Quaternary time developed the north-south trough represented by Death Valley, but some of the individual faults outlining the trough may trend northwest and be arranged en echelon like the steep flanks of the folds to which they seem related. The probabilities are that Death Valley is as low today as it has ever been, and that the valley floor has been sinking in altitude as well as in relation to the adjoining mountains. There is a little evidence, not at all satisfactory, that Death Valley drained southward in middle or late Pleistocene time (p. A65).

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General Geology of Death Valley California

GEOLOGICAL SURVEY PROFESSIONAL PAPER 494

This volume was published in separate chapters A and B

Geology of Northern Nellis Air Force Base Bombing and Gunnery Range, Nye County, Nevada

By E. B. EKREN, R. E. ANDERSON, C. L. ROGERS, *and* D. C. NOBLE

GEOLOGICAL SURVEY PROFESSIONAL PAPER 651

Prepared on behalf of the U.S. Atomic Energy Commission

Stratigraphy and structure of 2,400-square-mile area of dominantly Tertiary volcanic rocks in the Great Basin, with brief descriptions of small mines and prospects

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1971

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GEOLOGY OF NORTHERN NELLIS AIR FORCE BASE BOMBING AND GUNNERY RANGE, NYE COUNTY, NEVADA

By E. B. Ekren, R. E. Anderson, C. L. Rogers, and D. C. Noble

ABSTRACT

The area of study, covering about 2,400 square miles in Nye County, southwestern Nevada, lies east of Goldfield. Elevations range from 4,700 feet in the westernmost alluvial valley, Stonewall Flat, to more than 8,900 feet in the Reveille Range in the northeastern part of the area. The climate is arid; rainfall ranges from about 4 inches in the valleys to about 6 inches in the higher ranges and mesas. All the streams are intermittent and, with the exception of Thirsty Canyon and its tributaries, all end in closed basins, which make up about half of the total area. Igneous rocks of Tertiary age form at least 90 percent of the outcrops. The remainder consists of sedimentary rocks of late Precambrian and Paleozoic age and a single small horst of crystalline basement.

Rocks of late Precambrian age have an aggregate thickness of about 8,400 feet and include the Stirling Quartzite and the lower five-sixths of the Wood Canyon Formation; they consist of quartzite, siltstone, phyllite, and rather minor carbonate rock. The upper part of the Wood Canyon Formation is Early Cambrian in age and consists largely of micaceous siltstone and shale, with subordinate quartzite and a few carbonate layers.

The Wood Canyon Formation is overlain by the thin Zabriskie Quartzite of Early Cambrian age, which in turn is overlain by Lower to Middle Cambrian rocks that are at least 4,500 feet thick and are transitional between the older dominantly clastic rocks and younger dominantly carbonate rocks.

Younger rocks are dominantly of miogeosynclinal origin and belong to the eastern carbonate assemblage. At the base they include an incomplete section of the Middle and Upper Cambrian Bonanza King Formation (largely dolomite) and the Upper Cambrian Nopah Formation, which is about 3,000 feet thick and consists of the Dunderberg Shale and the Halfpint and Smoky Members.

Most of the Ordovician rocks occur in the Pogonip Group, which is also about 3,000 feet thick and consists of the Goodwin Limestone, Ninemile Formation, and Antelope Valley Limestone. The Pogonip is overlain by the Middle Ordovician Eureka Quartzite, about 315 feet thick, and this in turn is overlain by the Middle and Upper Ordovician Ely Springs Dolomite, which is about 340 feet thick.

The Ordovician rocks are succeeded by the dolomite of the Spotted Range, of Early Silurian to Early Devonian age. The dolomite is about 1,400 feet thick and is overlain by an incomplete section of the Nevada Formation. The Nevada is Early and Middle Devonian in age and at least 1,000 feet thick.

Most of the younger Paleozoic rocks appear to lie in the upper plate of a major thrust fault. They include a limestone and dolomite unit of Middle Devonian age, which has an exposed thickness of almost 1,300 feet, and the overlying Eleana Formation, which is Late Devonian to Mississippian in age and more than 4,000 feet thick.

Small, exposures of granite of Mesozoic age occur in the Cactus Range and southern Kawich Range. The granite is nearly void of mafic minerals and closely resembles the alaskite at Goldfield.

Rocks of Tertiary age form a composite section more than 20,000 feet thick. They include minor fanglomerate at base, numerous widespread ash-flow tuff sheets that range in age from about 27 to 7 m.y. (million years), thick piles of variegated lavas, and several sequences of interbedded ash-flow tuff and sedimentary rocks. The oldest volcanic rock is an ash-flow tuff of late Oligocene age named herein Monotony Tuff. The rock is rhyodacitic and contains abundant phenocrysts of plagioclase, quartz, and biotite. It is overlain by the tuffs of Antelope Springs in the western part of the mapped area and the Shingle Pass Tuff in the eastern part. Both units are dominantly rhyolitic and quartz latitic. The next higher unit of regional significance is the rhyolitic tuff of White Blotch Spring. This unit includes ash flows from more than one center, but all are characterized by abundant large crystals of quartz and sparse mafic minerals. In most areas the White Blotch Spring is overlain by minor sedimentary rocks and ash-flow tuff and then by widespread lavas of intermediate composition. The lavas were extruded from many vents that are widely scattered in the mapped area and beyond. They form the bulk of the outcrops in many parts of the area and are the principal host rocks for gold and silver ore at Goldfield and Tonopah adjacent to the area of study.

The Fraction Tuff, which overlies the lavas of intermediate composition, is more than 7,000 feet thick at the exhumed Cathedral Ridge caldera in the southern extension of the Kawich Range. The Fraction is a lithic-fragment-rich, crystal-rich, and generally quartz-rich ash-flow sheet of rhyolitic and quartz latitic composition. During a pause in volcanic activity after the eruption of Fraction Tuff, local areas were deeply eroded.

The period of relative quiescence and erosion was followed by extrusion of rhyolite lavas that overrode wide areas, and these eruptions in turn were followed by the extrusion of the Belted Range Tuff and related sodic rhyolites. The Belted Range is overlain by strata of the Paintbrush Tuff and by massive strata of the Timber Mountain Tuff of which two members are present: the Rainier Mesa Member and the Ammonia Tanks Member. In early Pliocene time, after the region had acquired a topography similar to that of the present, the Thirsty Canyon Tuff was extruded. This tuff, whose source was the Black Mountain caldera, forms the surface strata over broad areas in the western and southwestern parts of the area.

The project area contains eight structural blocks or units: the Belted Range, Kawich Range, Mellan Hills, Cactus Range, Trappman Hills, Mount Helen, Black Mountain, and Pahute Mesa. The Belted Range block exposes a major thrust fault having a displacement of several tens of miles. This fault places lower Paleozoic rocks over upper Paleozoic rocks and correlates with the C P thrust in the Yucca Flat area at the Nevada Test Site. Two normal fault systems are present throughout the mapped area, exclusive of Pahute Mesa and Black Mountain. The earlier system consists of two sets that strike northeast and northwest; the later system, a single set, strikes north. Both systems postdate the Tertiary volcanic rocks, but the older system is confined to rocks older than about 17 m.y.

Deposits of gold and silver occur in several localities, and small mines and prospect pits are abundant in parts of the Cactus Range and the southern Kawich Range. The deposits are small, and the combined gold and silver production to date is between 10,000 and 100,000 ounces.

Several localities within the area may provide favorable environments for underground nuclear testing.

INTRODUCTION

LOCATION AND GEOGRAPHY

The mapped area (fig. 1) lies within the Basin and Range physiographic province and consists primarily of alluvium-covered valleys separated by northerly trending mountain ranges. Pahute Mesa lies in the southern part of the mapped area and forms an east-west terminus to the north-trending mountains and valleys. Thick coalescing alluvial fans flank most of the mountain ranges. Relief in the area generally ranges from 1,000 to 3,000 feet; the lowest elevation, at Stonewall Flat on the west, is 4,729 feet, and the highest elevation, at Reveille Peak on the northeast, is 8,910 feet. All the streams are intermittent, and, with exception of Thirsty Canyon and its tributaries, all end in the closed basins of Cactus Flat, Reveille and Kawich Valleys, and

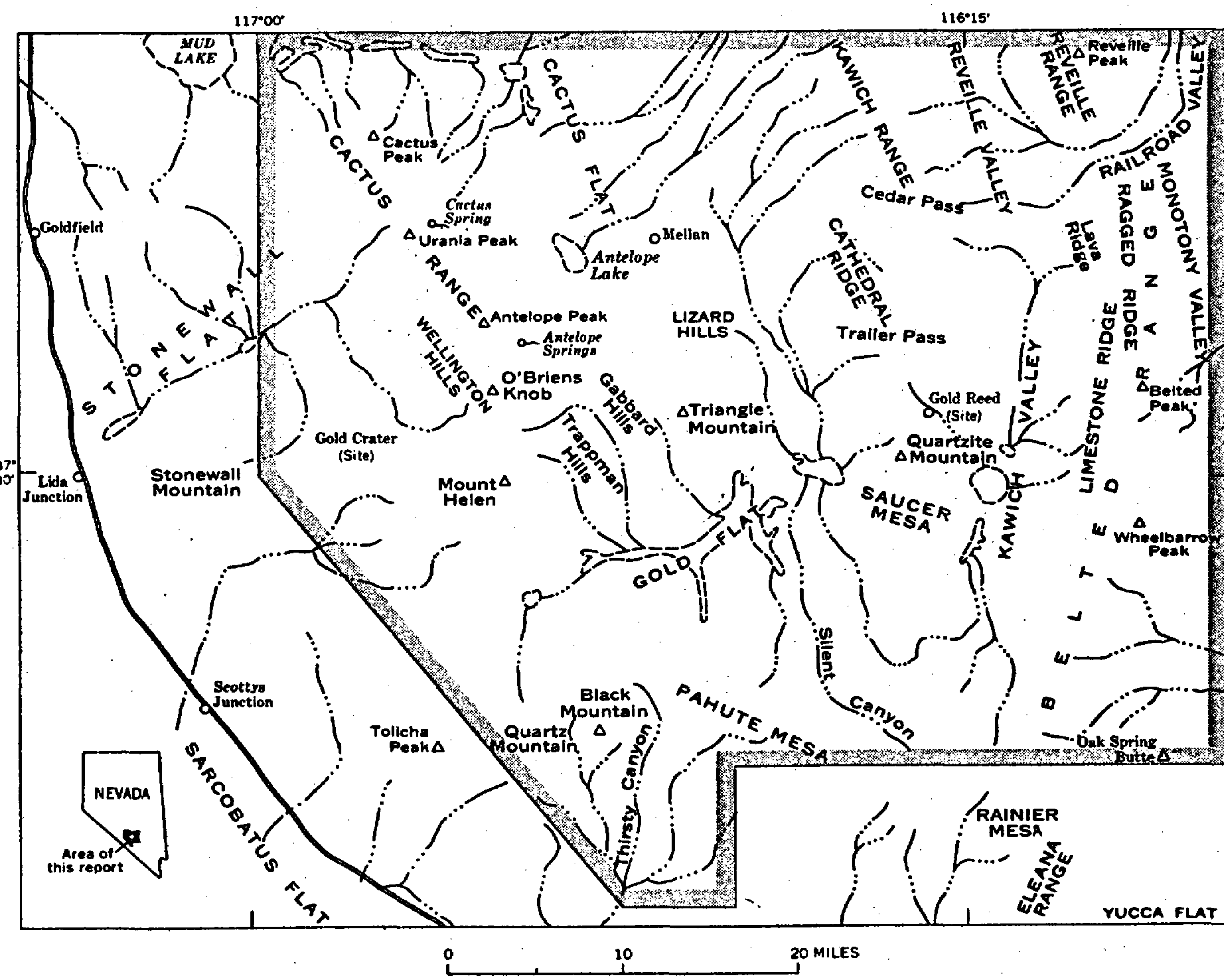

FIGURE 1.—Location of studied area (shaded outline) and major physiographic features.

Stonewall Flat. The basins are almost wholly enclosed by surrounding ranges and by Pahute Mesa.

Playa lakes occur in Gold Flat, Kawich Valley, and Cactus Flat. During the Pleistocene Epoch the lakes contained water several tens of feet deep as shown by former shorelines still outlined by thin zones of fine sand. The playas are used by the U.S. Air Force as bombing sites.

PURPOSE AND SCOPE OF THE INVESTIGATION

The northern Nellis Air Force Base Bombing and Gunnery Range was mapped as part of the long-range geologic investigations undertaken by the U.S. Geological Survey on behalf of the U.S. Atomic Energy Commission. The immediate objective was to provide geologic information necessary to predict whether parts of the area might be suitable for underground testing. The study included a detailed gravity survey to determine the subsurface configuration of the major basins in the area and the thicknesses of fill in the basins. The results of the gravity survey were described by C. H. Miller and D. L. Healey (written commun., 1964), and the gravity contours are shown on plate 1.

After Pahute Mesa was developed as an underground test area, consideration of the remainder of the Nellis Air Force Base Bombing and Gunnery Range for a test area was abandoned. As a result, this report has been written to provide general geologic information without specific emphasis on the evaluation of the area for nuclear testing.

CLIMATE, VEGETATION, AND WILDLIFE

The climate of the mapped area is arid (table 1) and the summer months are extremely hot. The most pleasant season is early fall, when the daytime temperatures range from low sixties to high eighties and high winds are rare. In the spring, the weather is extremely variable and few days are without strong winds.

Vegetation is sparse in the lower elevations but is fairly abundant on Pahute Mesa (table 2) and in the higher ranges. Grasses are abundant at elevations above 6,000 feet and support several herds of wild horses. Kawich and Reveille Valleys and Gold Flat are used by local ranchers as winter ranges for cattle, but the valleys constitute marginal grazing land at best. A herd of six or seven antelope inhabits the Gold Flat area, and a herd of about 20 antelope inhabits the Kawich Valley area. Deer are common in the higher ranges and on Pahute Mesa, and the mapped area is the home of many golden eagles. These can be seen nearly every day, and eagle nests are common in higher crags. Coveys of chukar partridge are abundant on Pahute Mesa, in the Belted Range, and in the vicinity of all flowing springs.

TABLE 1.—*Rainfall and temperature at Goldfield and Tonopah, Nev.*

[From U.S. Weather Bureau]

Year	Goldfield				Tonopah			
	Rainfall (inches)	Temperature (° F)			Rainfall (inches)	Temperature (° F)		
		Average	Extremes			Average	Extremes	
			Low	High			Low	High
1958					3.38	52.6	6	100
1959	2.87	53.4	6	100	2.37	52.7	0	103
1960	6.55	51.4	−2	99	3.69	52.6	0	104
1961	4.47	50.8	0	97	2.90	51.3	0	102
1962	6.23	51.9	−10	96	5.84	50.7	−15	97

TABLE 2.—*Plants in the northern Nellis Air Force Base Bombing and Gunnery Range*

[Identified and compiled by Helen Cannon, U.S. Geol. Survey]

Kawich Valley

For about 5 miles north and south from the playa the plant assemblage includes these xerophytic halophytes, which are very tolerant of salts:
- *Atriplex confertifolia*—shadscale
- *Kochia americana*—green molly
- *Artemisia spinescens*—budsage

For the next several miles both north and south, in soils that contain more alluvial material and less salts:
- *Eurotia lanata*—winterfat
- *Atriplex linearis*—narrowleafed saltbush
- *Tetradymia glabrata*—littleleaf horsebrush
- *Artemisia spinescens*—budsage
- *Atriplex confertifolia*—shadscale

In both ends of the valley and in washes along the sides of the valley, where water is available and the salt content is much lower:
- *Chrysothamnus* sp.—rabbitbrush
- *Artemisia arbuscula*—sagebrush

Higher in the drier parts of the alluvial fans:
- *Ephedra nevadensis*—Mormon tea
- *Grayia spinosa*—Hopsage

Gold Flat

The plants of Gold Flat are generally similar to those in Kawich Valley but there is more evidence of water. An assemblage of phreatophytes was noted along the south edge of the playa. They included:
- *Sarcobatus vermiculatus*—greasewood
- *Chrysothamnus viscidiflorus*—rabbitbrush
- *Atriplex canescens*—four-wing saltbush

A large area of selenium-indicator plants occurs in the center of the valley; however, these species do not indicate a very high concentration of selenium:
- *Astragalus lentiginosus*—Specklepod loco
- *Stanleya pinnata*—Princesplume
- *Aster abatus*—Mohave aster

Pahute Mesa

Trees:
- *Pinus monophylla*—pinyon
- *Juniperus osteosperma*—juniper
- *Quercus gambelli*—Gambels oak

Shrubs:
- *Ephedra* spp.—Mormon tea
- *Chrysothamnus* spp.—rabbitbrush
- *Tetradymia* spp.—horsebrush
- *Ribes montegenum*—gooseberry
- *Artemisia* spp.—sagebrush

TABLE 2.—*Plants in the northern Nellis Air Force Base Bombing and Gunnery Range*—Continued

Pahute Mesa
Purshia tridentata—antelope brush
Cowania stansburiana—Stansburys cliffrose
Symphoricarpos parishii—snowberry
Atriplex canescens—four-wing saltbush
Rhus trilobata—skunkbush
Cercocarpus ledifolius—mountain mahogany
Herbs—Many species of the following genera are common:
Lupinus—lupine
Cryptantha—cryptanth
Phacelia—heliotrope
Penstemon—penstemon
Castilleja—paintbrush
Arenaria—sandwort
Senecio—groundsel
Arabis—rockcress
Oenothera—primrose
Gilia—gilia
Eriogonum—wild buckwheat

FIELDWORK AND ACKNOWLEDGMENTS

The base map for this study was prepared by enlarging the Army Map Service Goldfield, 1° by 2° topographic sheet from its original scale of 1:250,000 to a scale of 1:125,000. Geologic mapping was done on aerial photographs, and the data were transferred to the topographic base by use of a Kelsh stereoplotter.

The following U.S. Geological Survey geologists participated in the mapping: Theodore Botinelly and J. E. Weir mapped with the authors during parts of 1962–63; in November 1962, E. J. McKay, J. T. O'Connor, and E. N. Hinrichs mapped parts of Triangle Mountain and Cathedral Ridge, and H. R. Cornwall mapped the Reveille Range and Reveille Valley. Credit for mapping quadrangles in the southern part of the area is given on plate 1.

E. G. Halligan (deceased) of the U.S. Army coordinated the ground surveying by the U.S. Geological Survey during 1962–64 with target schedules of the U.S. Air Force. The necessity of coordinating fieldwork with bombing schedules required that all project personnel work in limited areas for scheduled periods of time. As a result, none of the ranges or intervening areas was mapped entirely by one person; however, certain areas were designated as the responsibility of one or more geologists.

R. E. Anderson mapped most of the Cactus Range and wrote the descriptions of the stratigraphy and structure pertinent to that range. C. L. Rogers and J. E. Weir mapped most of the younger volcanic rocks exposed southwest of the Cactus Range, and Rogers also mapped nearly all the upper Precambrian and Paleozoic rocks in the project area. He is the sole author of the descriptions of these rocks in this report. E. B. Ekren mapped the geology of the Trappman Hills, and the volcanic rocks in the Belted Range, Mount Helen, and the Mellan Hills. Anderson, Ekren, and Rogers shared equally in the mapping of the Kawich Range and Cathedral Ridge area from the north boundary of the project area to the north rim of Pahute Mesa. Ekren and Anderson wrote the descriptions of the older volcanic rocks, and D. C. Noble, who participated in the mapping of most of the 7½-minute quadrangles along the south boundary (pl. 1), wrote the descriptions of the younger volcanic rocks. The sections on structure and on mines and mining were written by Ekren, Anderson, and Rogers.

We are indebted to R. L. Smith for his enthusiastic interest in the project and for his continued availability for consultation on problems pertaining to the volcanology of the area.

PREVIOUS WORK

The area between lat 36°30′ and 38°00′ and long 116° and 117°30′, which includes the present project area, was mapped in reconnaissance by Ball (1907). Ball delimited, with considerable accuracy, Paleozoic strata from volcanic strata, and the present authors marvel at the amount of geology he and his companions mapped in the very short field period of June–December 1905. Ball's study was made during the period of active mining in the area, and he described in considerable detail several of the mining districts that are now abandoned.

STRATIGRAPHY

Sedimentary, igneous, and metamorphic rocks that range in age from Precambrian to Holocene are exposed in the area. Tertiary lavas, ash-flow and ash-fall tuff, and intrusive rocks form at least 90 percent of the outcrops (table 3).

The pre-Tertiary rocks, exclusive of a small area of crystalline basement, are about 29,000–30,000 feet thick and include three distinct sequences. The lower part of the section—which is 10,000–11,000 feet thick and consists of quartzite, siltstone, shale, and minor carbonate—is late Precambrian to Middle Cambrian in age. It is overlain by a section—which is more than 14,000 feet thick and consists of limestone, dolomite, and minor siltstone, quartz and chert—that is Middle Cambrian to Devonian in age. This part of the section is of miogeosynclinal origin and is part of the eastern or "carbonate" assemblage as defined by Silberling and Roberts (1962, p. 5).

The Antler orogeny greatly modified the Cordilleran geosyncline in middle Paleozoic time, and during the remainder of the Paleozoic, deposition took place only locally in shallow restricted waters. Within the mapped

TABLE 3.—*Major geologic rock units in the northern Nellis Air Force Base Bombing and Gunnery Range*

Age	Rock unit	Stratigraphic thickness (feet)	Lithologic character
Quaternary and late Tertiary	Alluvium and colluvium	0–3, 000+	Valley and stream alluvium, terrace and pediment gravels, talus and landslide debris.
Pliocene	Basalt	0–100	Lava flows; a few dikes and one small cinder cone.
	Basalt	0–100	Lava flows; many dikes.
	Thirsty Canyon Tuff	0–500	Trachyte, trachytic sodic rhyolites, comendite, and pantellerite.
	Timber Mountain Tuff: Ammonia Tanks Member	0–350	Rhyolitic welded tuff.
	Timber Mountain Tuff: Rainier Mesa Member	0–600	Do.
Miocene	Paintbrush Tuff	0–500	Rhyolitic ash-fall tuff and intercalated rhyolite lavas.
	Rhyolite lavas and tuffs	0–1, 000	Sodic rhyolite lava flows, welded tuff, ash-fall tuff.
	Belted Range Tuff and associated lavas	0–1, 000	Comendite, trachytic sodic rhyolite, trachyte.
	Tuff of Tolicha Peak	0–400	Rhyolitic welded tuff.
	Rhyolite	0–1, 600	Lava flows and numerous dikes and plugs of rhyolite and rhyodacite.
	Sedimentary rocks	0–800	Ash-fall tuff, tuffaceous sediment, and thin-bedded lake sediment.
	Fraction Tuff	0–7, 200	Rhyolitic welded tuff, composite ash-flow sheet.
	Lavas of intermediate composition	0–3, 000	Lava flows, dikes, plugs, and small stocks.
	Tuff of Wilsons Camp and bedded tuff	0–500+	Rhyolitic welded tuff, ash-fall tuff, and tuffaceous sedimentary rock.
	Tuff of White Blotch Spring	0–3, 000	Rhyolitic welded tuff.
	Shingle Pass Tuff	0–1, 000	Rhyolitic welded tuff.
	Tuffs of Antelope Springs	0–4, 000+	Rhyodacitic and rhyolitic welded tuff; several cooling units.
Oligocene	Monotony Tuff	0–2, 300	Rhyodacitic welded tuff.
	Unconformity		
Mississippian and Late Devonian	Eleana Formation	5, 000±	Argillite, quartzite, and conglomerite; some limestone and limestone conglomerate at base.
Middle Devonian	Limestone and dolomite	1, 285+	Limestone, silty lim[illegible]tone, and dolomite.
Middle and Early Devonian	Nevada Formation	1, 000+	Dolomite, sandy d[illegible]lomitic sandstone, with subordina[illegible] and chert.
Early Devonian and Late and Middle Silurian	Dolomite of the Spotted Range	1, 415	Dolomite; sandy [illegible]
Late and Middle Ordovician	Ely Springs Dolomite	340	Dolomite, wit[illegible]
Middle Ordovician	Eureka Quartzite	315	Quartzite; g[illegible] lying unit[illegible]
Middle and Early Ordovician	Pogonip Group	3, 010±	Limestone [illegible] calcareo[illegible]
Late Cambrian	Nopah Formation: Smoky Member	950±	Limestor[illegible]
	Nopah Formation: Halfpint Member	1, 900±	Limest[illegible] and [illegible]
	Nopah Formation: Dunderberg Shale Member	200±	Shale [illegible]

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TABLE 3.—*Major geologic rock units in the northern Nellis Air Force Base Bombing and Gunnery Range*—Continued

Age	Rock unit		Stratigraphic thickness (feet)	Lithologic character
Late and Middle Cambrian	Bonanza King Formation		3, 300+	Dolomite and limestone; silty in part; minor chert.
Middle Cambrian	Limestone, silty limestone, and shale		1, 323+	Limestone and silty limestone, local chert; shale sequence at top.
Middle and Early Cambrian	Carrara(?) Formation		1, 577	Shale, and subordinate limestone, silty limestone, and siltstone.
Early Cambrian	Zabriskie Quartzite		150	Quartzite and subordinate siltstone.
Early Cambrian and Precambrian	Wood Canyon Formation		3, 750	Quartzite and siltstone, partly micaceous; carbonate beds in upper and lower parts.
Precambrian	Stirling Quartzite	Upper part	2, 300	Quartzite and siltstone, partly micaceous, and silty phyllite; subordinate carbonate material.
		Lower part	2, 970+	Quartzite, subordinate micaceous siltstone and quartzite and minor phyllite.
	Gneiss and schist of Trappman Hills area			Gneissic quartz monozonite and biotite-amphibole schist.

area this period is represented solely by the Eleana Formation of Late Devonian and Mississippian age, which is largely clastic, consisting dominantly of argillite, siltstone, quartzite, and conglomerite. This formation is at least 4,000 feet thick.

Although the Paleozoic strata have been telescoped by a thrust fault system of Mesozoic age, the total displacement was insufficient to juxtapose the western or detrital-volcanic assemblage. Some of the Devonian rocks in the Belted Range, however, may represent a transitional facies introduced by the large-scale thrusting.

PRECAMBRIAN

LOWER PRECAMBRIAN—GNEISS AND SCHIST OF TRAPPMAN HILLS AREA

Gneissic quartz monzonite and biotite schist of probable early Precambrian age crop out in the Trappman Hills (pl. 1) in an area about 3 miles long and 1 mile wide between Mount Helen and Gold Flat. The rocks are poorly exposed and form low rounded hills.

The gneissic quartz monzonite, which forms at least 70 percent of the outcrops, is light gray to brownish gray, fine to medium grained, locally pegmatitic, and generally moderately to well foliated. The rock contains 33–46 percent quartz, 26–30 percent orthoclase, about 25 percent plagioclase, and 1–15 percent muscovite and biotite. The potassium feldspar is generally fresh and clear, and a few grains are perthitic. The plagioclase is cloudy, and in zones of hydrothermal alteration the rock contains sericite, calcite, clay, abundant limonite, and cubes of pyrite. Biotite is altered to chlorite, sericite, and iron oxide. Biotite, muscovite, and chlorite occur as tiny isolated flakes and as lenticular masses along foliation planes. Quartz grains have been intensely strained, and undulatory extinction is pronounced in eight thin sections from widely separated outcrops.

The gneissic quartz monzonite contains xenoliths of schist, and in many outcrops it appears to have intruded the schist lit-par-lit. Contacts between schist and gneiss are very sharp in most areas; however, in places the contacts are gradational between mica-poor gneissic quartz monzonite and mica-rich schist.

The schist is of variable composition. Some contains the same minerals as the quartz monzonite but has more biotite and actinolite or tremolite. Other schist is very calcareous and contains abundant clinozoisite. Float of argillite and quartzite of probable Precambrian age is common on some of the hills, but these rocks were not observed in place.

The quartz monzonite and schist are cut by a few thin aplite dikes and by many quartz veins. The quartz veins range in width from a fraction of an inch to more than 50 feet and in places are several hundred feet long. They contain some pyrite and gold and were mined in the early 1900's. Ball (1907, p. 138) visited the Trappman Hills in 1905 when the quartz veins were being mined and concluded that there were three distinct periods of vein formation. He considered the first to be of pegmatitic origin related to the quartz monzonite, which he considered to be Mesozoic in age, and the second and third to be of Tertiary age.

UPPER PRECAMBRIAN—STIRLING QUARTZITE

DEFINITION AND DISTRIBUTION

The Stirling Quartzite was named by Nolan (1929) for exposures on Mount Stirling, about 5 miles east of the Johnnie mine, in the northwestern part of the Spring Mountains, Nye County, Nev. In the type locality it is 3,700 feet thick and is composed mainly of crossbedded quartzite, with minor amounts of siltstone and shale.

In the Groom district, east of the mapped area, the Stirling is 3,400–3,500 feet thick, as estimated by Barnes and Christiansen (1967), and consists dominantly of nonmicaceous quartzite, with less abundant siltstone and micaceous quartzite and a few thin beds of limestone. In the mapped area the Stirling Quartzite is exposed mainly in the southern extension of the Kawich Range and in one very small area on the west flank of the Belted Range. Quartzite Mountain, which has a total relief of about 2,500 feet and is a major landmark in this region, is composed entirely of Stirling Quartzite.

A nearly complete section was measured by J. H. Stewart and C. L. Rogers in the southern extension of the Kawich Range, in secs. 13, 24, 26, 27, and 35 (all unsurveyed), T. 4 S., R. 50 E. This was a composite section measured in six parts extending from the west flank of Quartzite Mountain to a point about 5.8 miles north-northwest of Quartzite Mountain. The section has a total thickness of almost 5,300 feet and, according to Stewart (oral commun., 1964), who has made a regional stratigraphic study of the formation (Stewart, 1970), is the thickest known section of the Stirling Quartzite. The lithology is intermediate between predominantly quartzite lithology to the southeast and predominantly siltstone lithology with conspicuous carbonates to the northwest.

LITHOLOGY

The base of the Stirling is not exposed in the Quartzite Mountain area, but the lowest beds exposed resemble the basal beds of the formation in other localities and probably lie only a short distance above the contact with the underlying Johnnie Formation. The remainder of the formation is complete, and at the north end of the outcrop the uppermost beds of the Stirling are overlain by beds of the Wood Canyon Formation. In this report the formation is divided informally into lower and upper parts. This bipartite division is a natural one, for although the lower part is dominantly quartzite, the upper part is dominantly siltstone and silty phyllite, with subordinate quartzite and an appreciable amount of carbonate material. The formation is further divided into five informal members, A through E, which are correlative with members A through E described by Stewart (1966) in the Spring Mountains-Death Valley area, and elsewhere (1970). They are summarized as follows:

	Thickness (feet)
Stirling Quartzite	5,290+
Upper part	2,320
Member E: largely siltstone and quartzite	1,195
Member D: dolomite and limestone	285
Member C: largely silty phyllite	840
Lower part	2,970+
Member B: quartzite, silty sandstone and siltstone; partly micaceous	1,130
Member A: largely quartzite	1,840+

Member A is about 1,840 feet thick and is almost wholly quartzite except for thin partings of greenish-gray micaceous siltstone or phyllite, which become more abundant from the base upward. The lowest beds of the formation in this area, which are well exposed on the west flank of Quartzite Mountain, are composed of gray to dusky-grayish-purple laminated to thick-bedded quartzite that weathers reddish brown. The quartzite forms somber-looking cliffs and steep rubble-covered slopes. The quartzite is fine to coarse grained and contains rather abundant thin conglomeratic lenses that are generally composed of granules and pebbles less than 5 millimeters in diameter but may include a few fragments as much as 1 centimeter in diameter. Many beds are laminated to cross laminated, and a few show rather poorly developed graded bedding. Ripple marks occur at some horizons but are nowhere abundant. Thin sections reveal that the quartzite consists mainly of subrounded to subangular quartz, 5–10 percent altered feldspar, a few quartzite fragments, minor red jasper, a variable amount of sericite that occurs interstitially and as an alteration product within the feldspar grains, finely disseminated limonite or hematite, and a small amount of heavy minerals such as magnetite, sphene, zircon, and tourmaline.

The upper part of member A contains a zone 35 feet thick that is predominantly medium-light-gray to olive-gray phyllite, with subordinate grayish-purple micaceous fine-grained quartzite. Above this zone is a sequence 375 feet thick, in which dark quartzite grades upward into lighter quartzite and in which micaceous siltstone and silty sandstone laminae are fairly common. This sequence is medium dark purplish gray to pinkish and yellowish gray on fresh surfaces and weathers to light reddish brown.

Member B, which overlies the lighter quartzite of member A, is about 1,130 feet thick. Its lower part, which is less resistant than the underlying quartzite and forms slopes and saddles, is composed of nonmicaceous to highly sericitic silty sandstone, quartzite, and siltstone, and it weathers to platy or flaggy fragments. The member is partly laminated to thick bedded, but it is predominantly thin bedded. The rock types are all gradational, and they characteristically alternate from purple to green. These rocks become gradually coarser and less micaceous toward the top and grade into a pinkish-gray to grayish-purple quartzite that is about 210 feet thick. The quartzite is mostly fine to medium grained, laminated to thin bedded, and contains a small amount of silty sandstone. The upper 110 feet of member B, which is composed of phyllitic, platy-splitting siltstone with subordinate pinkish-gray quartzite, represents a transition to the upper part of the formation.

The upper part of the Stirling Quartzite, which is about 2,320 feet thick, is a heterogeneous sequence that originated largely as fine-grained clastic sediments with subordinate carbonate material. Individual members are distinctive and could be easily mapped.

Member C is a relatively nonresistant sequence of rocks that is about 840 feet thick and is estimated to be 90 percent silty phyllite or phyllitic siltstone and about 10 percent quartzite. Quartzite is most abundant in the basal beds, which consist of micaceous siltstone and quartzite interbedded with silty to sandy phyllite. These rocks are dark steel gray to almost black, with occasional purplish hues, and are laminated to medium bedded. In thin section the quartzose rocks consist largely of subangular quartz grains in a fine-grained sericitic matrix. The quartz grains range in size from silt to sand, having an average diameter of about 0.05 mm and a maxium diameter of about 0.2 mm. The thin sections also contain a few feldspar grains, minor chlorite, calcite, and dark opaque minerals, minor tourmaline, and rather abundant small carbon flecks that occur predominantly along the micaceous laminae. The phyllitic interbeds are characterized locally by a conspicuous slaty cleavage.

Above the basal beds, member C consists largely of laminated to thin-bedded silty phyllite or phyllitic siltstone. To a point about 600 feet above the base the member is predominantly medium gray to grayish purple, with numerous greenish-gray zones, but toward the top it is almost wholly green. This green sequence exhibits some variation laterally from a silty phyllite to a silty argillite, and it is not clear whether this simply reflects some difference in the degree of metamorphism, or whether it reflects a change in lithology and the presence of a minor unconformity. The first explanation seems more likely. The phyllitic rocks are light greenish gray to greenish gray and may have a silky sheen or a distinctly schistose look, depending on the coarseness of the mica. Two specimens that were examined in thin section contain 30–50 percent silt-sized quartz and a little feldspar in subrounded to subangular grains, embedded in a very fine grained matrix of chlorite and sericite. These specimens also contain a considerable amount of iron oxide, both disseminated and filling small fractures. The argillite is a friable olive-drab rock that in thin section contains scattered small quartz grains in a very fine grained sericitic groundmass. The quartz grains average 0.02–0.03 mm in diameter and probably form 10–20 percent of the rock. The rock also contains a few feldspar grains, common disseminated limonite, and minor tourmaline. The limonite occurs as a stain and in euhedral and anhedral grains that are probably pseudomorphous after pyrite.

Near the base, member C generally contains several zones of dolomite 1–3 feet thick. The dolomite is light gray on fresh surfaces but weathers to grayish orange. Several thin zones of yellowish-brown dolomite also occur near the top of the member, and the highest beds are gradational into the overlying carbonate unit, member D. The basal dolomite beds may be lenticular, because they apparently are absent in some areas.

Member D is a carbonate unit about 285 feet thick. It is light to medium gray and aphanitic to fine grained; it ranges in composition from limestone to dolomite. It is predominantly laminated to thin bedded, and locally crosslaminated. The upper one-third is somewhat silty and includes some limestone that has weathered to brown. Some beds contain small platy to rodlike structures suggestive of poorly preserved algae and other fossils. According to J. H. Stewart (oral commun., 1964), this carbonate unit is probably correlative with part of the Reed Dolomite in the region to the west. The nearest exposure of Reed Dolomite lies only a short distance beyond the west boundary of the project area, near the northwest corner of Stonewall Mountain.

Member E, which overlies the carbonate unit, member D, is a heterogeneous sequence that is almost 1,200 feet thick; it is composed predominantly of siltstone and quartzite and contains very minor sandstone and carbonate. The lower 440 feet consists of silty to sandy argillite, micaceous siltstone, rusty limonitic sandstone, quartzite, and carbonate beds. The argillite and siltstone range in color from gray to greenish gray, olive drab, buff, and various shades of rusty brown. They are laminated to thin and medium bedded and weather to small platy fragments. In thin section the rock was found to contain abundant subangular quartz grains in a very fine grained sericitic and chloritic groundmass. The quartz grains are about 0.02–0.15 mm in diameter and form 40–50 percent or more of the rock, which ranges in composition from a silty argillite to a phyllitic siltstone and which contains rather abundant finely disseminated limonite, minor feldspar, and sparse small grains of tourmaline. The carbonate layers weather to reddish gray and yellowish gray and are similar to the rock in the underlying member. Carbonates occur throughout this zone but are somewhat less abundant in the upper part. The sandstone is greenish gray to yellowish gray and pale yellowish brown, fine to medium grained, laminated to thin bedded, locally crosslaminated, and commonly calcareous. The sandstone locally grades into quartzite. The quartzite in the upper two-thirds of this zone is predominantly pinkish gray, laminated to thick bedded, and relatively clean; it contains little mica.

The middle part of member E, 355 feet thick, is composed largely of interbedded siltstone, quartzite, and micaceous siltstone. The quartzite is increasingly abundant from the base upward and occurs in layers 1–3 feet thick.

The upper part of member E is a cliff-forming quartzite 400 feet thick; it has a few thin layers of micaceous siltstone in the upper part. The quartzite is predominantly a pinkish-gray clean-looking rock consisting largely of fine- to medium-sized quartz grains, with a small percentage of feldspar and rather abundant iron oxide specks. Thin lenticular pebble beds are common, and the rock is laminated to thin bedded with common cross-laminations.

AGE

No identifiable fossils have been found in the Stirling Quartzite. Its stratigraphic position indicates a probable late Precambrian age.

PRECAMBRIAN AND CAMBRIAN—WOOD CANYON FORMATION

DEFINITION AND DISTRIBUTION

The Wood Canyon Formation was named by Nolan (1929) for exposures in Wood Canyon in the northwestern part of the Spring Mountains, Nye County, Nev. In the type locality it comprises 2,100 feet of thin-bedded quartzitic sandstone and sandy shale with a few carbonate beds near the top.

The formation is exposed only in the two major belts of sedimentary rocks in the Belted and Kawich Ranges. In the Belted Range the unit occurs beneath a major gravity slide or thrust fault, and all but the basal part is well exposed. In the Kawich Range only the lower part is exposed.

LITHOLOGY

A section of the Wood Canyon Formation in the Belted Range, in an unsurveyed area about 2.5 miles west-southwest of Cliff Spring and 5.5 miles southwest of Belted Peak, was measured by J. H. Stewart and C. L. Rogers. Because the area contains numerous small faults and several rhyolite dikes, the thicknesses obtained may be slightly in error. The measured thickness is about 3,750 feet, and this represents an increase of about 1,465 feet over the nearest measured section, which lies approximately 20 miles to the east-southeast in the Groom district (Barnes and Christiansen, 1967).

In the Belted Range the contact of the Wood Canyon with the underlying Stirling Quartzite is masked by alluvium, and small differences in attitudes between the highest exposed Stirling beds and the lowest Wood Canyon beds suggest the possibility of a concealed fault. In the Kawich Range the contact with the Stirling Quartzite is transitional. Because the lower contact in these areas cannot be precisely located, the thickness of the lower unit could only be determined approximately at 1,320 feet.

Three informal units of the Wood Canyon are recognized in the Groom Range, outside the project area, and these have also been recognized within the project area. They are described separately as lower, middle, and upper units but are not shown separately on the geologic map.

The lower unit is composed largely of siltstone in the lower half and quartzite in the upper half. The siltstone is olive or olive gray to grayish red, platy, and variably micaceous. The quartzite is yellowish to greenish gray and pale yellowish brown, predominantly fine grained, evenly laminated to thin bedded, and locally micaceous. The unit also contains three carbonate zones which, according to J. H. Stewart (oral commun., 1964), are very persistent and have been recognized in widely separated areas. In the Kawich Range the lowest carbonate zone, which is 25 feet thick and lies about 360 feet above the base of the unit, is a grayish-orange, brownish-weathering, finely crystalline, sandy dolomite that grades locally to dolomitic sandstone and calcitic dolomite. In the Belted Range the lowest carbonate zone is 33 feet thick, and the overlying carbonate zones, which are similar in lithology, are 3 and 23 feet thick, respectively. Stewart believes that the basal carbonate zones in the two areas are correlative, and if this is true, it would indicate that the underlying sequence is considerably thicker in the Belted Range, because the lowest carbonate zone in the Belted Range apparently lies about 640 feet above the Stirling Quartzite. Much of this basal sequence in the Belted Range, however, is covered with alluvium, and the contact with the Stirling Quartzite can only be approximately located. Also, concealed faults in this covered interval may cause some duplication of beds.

The middle unit is about 1,115 feet thick and, because it is predominantly quartzite with subordinate siltstone, is more resistant than the neighboring units. The quartzite ranges in color from pale red to grayish red and yellowish gray and is fine to coarse grained, but predominantly fine grained. The beds in the unit are 1–6 inches thick and some are crosslaminated. The basal beds, 10–20 feet thick, are partly conglomeratic and contain granules and small pebbles of quartz and quartzite. The siltstone, which may constitute as much as 30 percent of the unit, is grayish purple to grayish olive, platy, and micaceous in varying degrees.

The upper unit is about 1,315 feet thick and is predominantly micaceous siltstone with smaller amounts of quartzite and carbonate rock. The siltstone is grayish olive to dusky yellow and platy. The quartzite is pale yellowish brown to yellowish gray, fine grained, laminated, and micaceous in part. Quartzite is rather sparse in the lower part of the unit but is increasingly abundant upward and may be dominant in the upper part. The carbonate rock is mainly concentrated in a zone 100 feet thick that lies about 850 feet above the base of the unit. This zone is estimated to contain about 60 percent carbonate rock, 30 percent quartzite, and 10 percent siltstone. The carbonate rock is largely dolomite or calcitic dolomite, except for a few feet of limestone near the base, and is pale yellowish brown and pale red on fresh surfaces, predominantly weathering to a dark reddish brown. The rock is finely crystalline or oolitic and in some parts contains disseminated sand grains. It is generally laminated to thin bedded and is locally cross-laminated. A few thin dolomite layers occur above the main zone. In the uppermost part of the unit, which is somewhat transitional in character, the siltstone grades into a finer grained rock that is more shaly in appearance and is similar in lithology to some of the shaly rocks in the overlying Zabriskie Quartzite and Carrara(?) Formation.

AGE

The lower and middle units and the basal part of the upper unit contain only indeterminate worm trails and borings and are considered to be Precambrian in age. Rocks in the upper 635 feet of the Belted Range section contain olenellid trilobites, vertical worm borings known as *Scolithus*, pelmatozoan debris, and some poorly preserved brachiopods. These rocks are considered to be Early Cambrian.

CAMBRIAN

ZABRISKIE QUARTZITE

DEFINITION AND DISTRIBUTION

The Zabriskie Quartzite was described by Hazzard (1937) in the Nopah-Resting Springs area of Inyo County, Calif., and was considered by him to be a member of the Wood Canyon Formation. The unit was redefined by Wheeler (1948, p. 26) as a formation between the underlying Wood Canyon Formation and overlying Carrara Formation.

The Zabriskie crops out only in the southern part of the Belted Range, where it can be traced for about 4 miles and is cut by numerous small transverse faults. The quartzite has been mapped with the Wood Canyon Formation on plate 1 because it is too thin to be shown separately. Its thickness is about 150 feet in the Belted Range, 220 feet in the Groom district (Barnes and Christiansen, 1967) 20 miles to the east-southeast, and 1,150 feet in the Bare Mountain area more than 50 miles to the southwest (Cornwall and Kleinhampl, 1961). The Belted Range section was measured by J. H. Stewart and C. L. Rogers.

LITHOLOGY

The lower one-third of the Zabriskie is a poorly exposed zone that is transitional into the underlying Wood Canyon Formation and is composed of interbedded quartzite and siltstone. This sequence has been arbitrarily included with the Zabriskie Quartzite, but a study of better exposures, if they could be found, might change this assignment to the underlying formation.

The upper two-thirds of the Zabriskie is a conspicuous ridge former and is composed of yellowish-gray to pale-yellowish-brown quartzite that is characteristically massive in appearance, although it is laminated and cross laminated in places. It contains a few thin lenticular conglomeratic layers containing pebbles and small cobbles of quartzite and red chert that are generally less than 1 cm long. The Zabriskie Quartzite is different from the quartzite in the older formations in that it is completely free of feldspar. A zone, 40 feet thick, at the top of the formation contains subordinate fissile olive-drab shale, and it apparently represents a transition into the overlying Carrara(?) Formation.

AGE

The only fossil evidence in the Zabriskie consists of vertical *Scolithus*-like worm borings. However, the stratigraphic position of the formation is sufficient for dating purposes, and it is clearly Early Cambrian in age.

CARRARA(?) FORMATION

DEFINITION AND DISTRIBUTION

The Carrara Formation was named by Cornwall and Kleinhampl (1961) for an abandoned mining camp located 8 miles east-southeast of Beatty, Nev. It lies between the Zabriskie Quartzite and the Bonanza King Formation and is transitional between the older dominantly clastic rocks and the younger dominantly carbonate rocks. In the type locality the Carrara consists of 1,785 feet of shale and limestone with relatively small amounts of quartzite, sandstone, and siltstone. The lower half consists mainly of shale with very subordinate limestone, but the upper half consists almost wholly of carbonate rock.

In the Belted Range the rocks lying above the Zabriskie Quartzite are intermediate in lithology between the typical Carrara Formation in the area to the southeast and the typical Emigrant Formation in Esmeralda County to the west. Rather than introduce a new stratigraphic name for these transitional rocks, the authors have resorted to a compromise. The lower part, which is about 1,577 feet thick and resembles the lower part of the Carrara, is called Carrara(?) Formation, and the upper part, which is about 3,000 feet thick and more closely resembles the Emigrant Formation, is identified simply as "limestone, silty limestone, and shale of Middle Cambrian age."

The Carrara(?) Formation is exposed over an area of several square miles on the west side of the Belted Range (pl. 1) and in several small isolated outcrops along the west base of a ridge 5 miles west-southwest of Mount Helen. These isolated outcrops are too small to be shown on the geologic map, but they are near the exposures of Eureka Quartzite in that area. In the Belted Range, in an area 2 miles southwest of Cliff Spring (unsurveyed secs. 21 and 22, T. 5 S., R. 52 E.), the formation was measured by J. H. Stewart and C. L. Rogers.

LITHOLOGY

For descriptive purposes, the Carrara(?) Formation is divided into four units, which correlate only in part with the units described by Barnes, Christiansen, and Byers (1962, 1965), and by Barnes and Christiansen (1967) in the Halfpint Range and the Groom district southeast of the project area. The four units are as follows:

	Thickness (feet)
Carrara(?) Formation	1,577
Unit D: limestone, silty limestone, calcareous siltstone, and shale	237
Unit C: papery shale	400
Unit B: shale, with rather minor siltstone, sandstone, and limestone	710
Unit A: shale and limestone in roughly equal proportions	230

Unit A is about 230 feet thick and consists of about equal parts of shale and limestone. The shale generally is grayish olive to green, and it contains sparse to rather abundant fine-grained mica. Some zones are less fissile and more resistant and are more accurately described as siltstone. The limestone is generally medium gray, finely crystalline, and thin to medium bedded, but it contains some buff silty limestone in discontinuous irregular laminae and a few thin layers that are silty throughout. The major limestone sequence, which is about 80 feet thick, occurs at the top of the unit. The lithology and fossil evidence indicate that unit A, though thinner, is the approximate equivalent of members A and B of Barnes and Christiansen (1967) in the Groom district, about 20 miles east-southeast of the Belted Range.

Unit B is about 710 feet thick and is composed largely of shale with relatively small amounts of silstone, sandstone, and limestone. The shale is generally grayish olive, though locally it is greenish gray to medium dark gray, and in the upper part of the unit is characterized in part by relatively coarse grained mica, giving it a schistose appearance. The upper part of the sequence also contains more abundant resistant siltstone layers and a few thin beds of greenish-gray micaceous sandstone. The limestone forms several thin zones near the base of the unit and a sequence about 20 feet thick at the top of the unit.

Unit C, about 400 feet thick, is a homogeneous sequence of papery pale-olive shale that weathers light yellowish gray or gray. The shale contains abundant fine-grained sericite, which imparts a silky sheen to the cleavage surfaces.

Units B and C, although more than twice as thick as member C, may be roughly equivalent to it in the Groom district. According to A. R. Palmer (written commun., 1964), the limestone zone at the top of unit B is comparable stratigraphically and in faunal content to a similar zone in the Groom section that forms the only prominent ledge between members B and D of that area. Trilobites collected from this limestone zone in the Belted Range and from the shales immediately above it are of earliest Middle Cambrian age.

Unit D, about 237 feet thick, is transitional in character. It is composed of alternating zones of limestone and clastic to semiclastic rocks ranging in type from yellowish-weathering micaceous shale near the base to thin-bedded buff- to brownish-weathering calcareous siltstone and silty limestone in the upper part. The limestone becomes more abundant upward and is generally medium to dark gray, aphanitic to finely crystalline, and laminated to medium bedded. Some beds are rather coarsely oolitic, and some limestone contains thin laminae of brown dolomite.

No fossils have been collected from unit D, but the general lithology and the fossil evidence obtained from neighboring units suggest that it represents member D of the Groom District.

AGE

An Early and Middle Cambrian age is assigned to fossils collected from the Carrara(?) Formation in the Belted Range and identified by A. R. Palmer (written commun., 1964).

Limestones in unit A yielded abundant fragments of the olenellid trilobites *Olenellus* and *Paedumias*, a smaller number of specimens of a nonolenellid trilobite that are probably referable to *Antagmus*, *Bristolia*, *Sombrerella*, several poorly preserved orthoid brachiopods, and numerous examples of *Girvanella*, an algal form, which may occur as scattered individuals or as clusters in biostromelike masses. This assemblage indicates an Early Cambrian age.

The limestone layer at the top of unit B yielded *Poliella?* sp. and a kochaspid trilobite, *Dictyonina* sp. and orthoid brachiopods, and *Chancelloria* sp.; shales in the basal part of unit C yielded *Pagetia* sp., *Oryctocephalus* sp., and undetermined ptychoparoid trilobites. The trilobites from these two horizons are of earliest Middle Cambrian age. *Girvanella* was collected from the top of unit C.

Collections from the small isolated exposures of possible Carrara beds in the western part of the project area, which were also studied by A. R. Palmer (written commun., 1964) contained abundant specimens of *Girvanella*, an indeterminate olenellid trilobite, trilobites referable to *Crassifimbra*, and a capuliform mollusk referable to *Scenella*. Palmer stated that it is difficult to make a formational assignment because, although these rocks are clearly Early Cambrian in age and equivalent to the lower part of the Carrara Formation, the locality lies between the Nye County and Esmeralda County Lower Cambrian areas, which have almost totally different formational sequences. The exposures that were found consist of thin- to medium-bedded gray to yellowish-gray silty limestone and thin interbeds of brown-weathering silty dolomite.

LIMESTONE, SILTY LIMESTONE, AND SHALE OF MIDDLE CAMBRIAN AGE

In three localities in the Belted Range the Carrara(?) Formation is overlain by a sequence that is referred to simply as limestone, silty limestone, and shale of Middle Cambrian age. The unit, which is incomplete and on the east is faulted against Tertiary volcanic rocks, can be divided into three units:

	Thickness (feet)
Limestone, silty limestone, and shale of Middle Cambrian age	2,970±
Upper unit: fissile shale	700±
Middle unit: limestone with minor silty limestone and chert	2,000±
Lower unit: limestone and silty limestone	270

The lower unit is about 270 feet thick and is composed of alternating zones of medium- to dark-gray aphanitic to fine-grained limestone and a buff- to brown-weathering silty limestone that grades locally into a calcareous siltstone.

Only the lower 450 feet of the middle unit has been measured, owing to an abundance of small-scale folds in its upper part. However, it is estimated that the total thickness of the unit may be as much as 2,000 feet. The middle unit is composed largely of medium- to dark-gray aphanitic to finely crystalline laminated to medium-bedded limestone, with scattered partings or thin interbeds of silty limestone or calcareous siltstone. Locally the beds have a wavy appearance, and, in the upper half of the unit, they contain scattered thin lenses of black chert. Chert increases in amount gradually upward, and near the top it is locally abundant.

The lower and middle units are lithologically similar to the Emigrant Formation (J. H. Stewart, oral commun., 1964) and, as shown by fossils, are partly correlative with it. The Emigrant Formation was named by Turner (1902) for exposures lying to the south of Emigrant Pass in the Silver Peak Range of Esmeralda County, Nev., and its age was revised by Albers and Stewart (1962) to Middle and Late Cambrian. It is equivalent in part to the upper part of the Carrara Formation and in part to the overlying Bonanza King Formation in the Groom and Jangle Ridge areas.

The upper unit, which is about 700 feet thick, is composed of gray to slightly greenish-gray fissile shale that weathers to various shades of brown and breaks into small platy to elongate fragments. It occurs only in the northernmost exposure of the sequence being described and is in normal-fault contact with volcanic rocks on the east. It seems probable that this shale occurs in approximately normal stratigraphic sequence and that it corresponds to a shaly unit in the Emigrant Formation. The contact between the shale and the underlying limestone cannot be determined because of poor exposures, however, and the shale may represent a sequence in the lower part of the Carrara(?) Formation that has been repeated by faulting.

AGE

In the lower unit, fossil material was collected by A. R. Palmer, J. H. Stewart, and C. L. Rogers, and was identified by Palmer (written commun., 1964) as *Ogygopsis typicalis* (Resser), *Pagetia* sp., *P. clytia* Walcott, and *P. maladensis* Resser, *Oryctocephalus* sp., *Tonkinella?* sp. and *?Tonkinella idahoensis* Resser, "*Agnostus*" *lautus* Resser and cf. "*Agnostus*" *lautus* Resser, *Alokistocare* sp., undetermined ptychoparioid trilobites, *Hyolithes*, and *Girvanella*. According to Palmer, this assemblage is approximately equivalent to the *Albertella* zone, which lies immediately below the Jangle Limestone Member of the Carrara Formation in the area east of Yucca Flat and which is Middle Cambrian in age.

The middle and upper units are probably Middle Cambrian in age, but no determinable fossils were found in them.

BONANZA KING FORMATION

DEFINITION AND DISTRIBUTION

The Bonanza King Formation was named by Hazzard and Mason (1936) for the Bonanza King mine on the east side of the Providence Mountains, San Bernardino County, Calif., where the formation consists of about 2,000 feet of dolomite and minor siltstone, silty limestone, and chert.

To the south and east of the project area the formation crops out widely in the C P Hills and in the Halfpint and Groom Ranges. A virtually complete stratigraphic section of the formation, 4,600 feet of limestone and dolomite, was measured across Jangle Ridge and Banded Mountain in the northwestern part of the Halfpint Range by Barnes, Christiansen, and Byers (1962, 1965), who divided the formation into Papoose Lake and Banded Mountain Members. In the Groom Range a complete section of the formation, which is 4,355 feet thick, is also largely limestone and dolomite with minor siltstone and silty limestone (Barnes and Christiansen, 1967). Within the project area the formation occurs mainly in the extreme southeast corner, where a thick, though incomplete, section is exposed. These rocks are tentatively assigned to the Papoose Lake and Banded Mountain Members, but the members are not separately mapped.

LITHOLOGY

In the main outcrop area an incomplete section of the Papoose Lake Member is about 1,450 feet thick and is composed almost wholly of dolomite. The dolomite is light to medium and dark gray, locally laminated but generally medium to thick bedded, and fine grained and is characterized in some places by a sugary texture. The lowest beds contain a small amount of buff platy silty dolomite; small lenses of gray chert were noted at one horizon. This part of the formation contains a large amount of light-colored secondary silica, some of which lies along fractures but most of which occurs in abundant very small, thin irregular lenses that on weathered surfaces bear a superficial resemblance to chain corals. The rocks have been dolomitized and extensively silicified, and as a result their original lithologic character has been largely obliterated. The Papoose Lake Member has been divided into three informal units by Barnes and Christiansen (1967), but these have not been recognized in the project area, perhaps owing to the extensive alteration of the rocks.

The upper part of the Bonanza King Formation, which has an exposed thickness of about 1,850 feet, is tentatively assigned to the Banded Mountain Member. The top of the formation and the overlying Dunderberg Shale Member of the Nopah Formation may lie beneath the covered area to the north, for this interval is succeeded on the north by a large exposure of the Halfpint Member of the Nopah Formation, which is structurally concordant with the Bonanza King beds.

The Banded Mountain Member has been divided into four informal units by Barnes and Christiansen (1967), but these units cannot be recognized with assurance in the project area, owing to alteration and poor exposures.

The basal 35 feet consists of buff to brownish-buff weathering laminated to thin-bedded silty dolomite which, except for being dolomitized, is similar to part of a zone marking the base of the Banded Mountain Member in the Halfpint and Groom Ranges.

The next 915 feet is composed of light- to dark-gray dolomite. The lower part consists of thick-bedded dark-gray dolomite, which is overlain by a thick sequence, incompletely exposed, of light- to dark-gray laminated to thick-bedded fine-grained dolomite that closely resembles the rocks assigned to the Papoose Lake Member. The upper part is light-gray laminated to thin-bedded dolomite and silty dolomite that weathers to brown. The thin-bedded dolomite contains a minor amount of lenticular gray chert. Locally these rocks contain rather abundant secondary silica similar to that occurring in the Papoose Lake Member.

The next 550 feet is mainly medium- to thick-bedded light-gray fine-grained locally calcitic dolomite in its lower part and thick-bedded light-gray to very light-gray aphanitic to fine-grained limestone in its upper part. The limestone weathers to a smooth rounded surface that contrasts with the rough pitted surface of the dolomite, and some beds appear to have been marmorized.

The uppermost 350 feet is medium- to thick bedded light- to medium-gray fine-grained dolomite having a rather saccharoidal texture. In the darker beds, which occur in the lower part of the unit, are a few fossiliferous layers that contain poorly preserved brachiopods and numerous small ovoid forms with a concentric structure that resemble *Girvanella*.

Two small exposures tentatively identified as Bonanza King occur on the west side of the Kawich Valley west of the north end of the Belted Range. They are composed of light- to dark-gray aphanitic to fine-grained medium- to thick-bedded dolomite. The color banding is partly stratigraphic, but it has been modified by irregular bleaching. One exposure is adjacent to a granite porphyry stock, and the rocks in both areas exhibit some fracturing and brecciation, as well as bleaching and minor silicification. The general lithology and the presence of small ovoid structures resembling *Girvanella* suggest a correlation with the Bonanza King Formation. The rocks also contain some linguloid brachiopod fragments, but these are nondiagnostic for the purpose of identifying the Bonanza King Formation.

AGE

Most of the Bonanza King Formation is Middle Cambrian in age, but the uppermost part is of Late Cambrian age (Barnes and Palmer, 1961).

NOPAH FORMATION

DEFINITION AND DISTRIBUTION

The Nopah Formation was named by Hazzard (1937) for the Nopah Range in southeastern Inyo County, Calif., where it consists dominantly of varicolored dolomite.

The formation is widely exposed between the project area and the Nopah Range. Barnes and Christiansen (1967) have described and redefined the Nopah in the Groom district and have correlated its three members southwestward to Bare Mountain and the Nopah Range. It is about 2,000 feet thick in the Groom district and is divided into the Dunderberg Shale, Halfpint, and Smoky Members. This area is only a few miles east of the southeast corner of the project area (pl. 1), and owing to its proximity and to lithologic similarity between the Groom and Belted Range rocks, the authors have adopted the nomenclature employed by Barnes and Christiansen.

In the Groom district the Dunderberg Shale Member is about 310 feet thick and consists of highly fissile pale-reddish-brown to olive-gray shale, with thin interbeds of platy to wavy and nodular limestone. The upper part grades into the overlying Halfpint Member, and the top is placed at the point where shale gives way to limestone as the dominant rock type. The Halfpint Member is 1,055 feet thick and is composed of platy-to flaggy-splitting very thin bedded limestone, with intercalated laminae of clayey and silty limestone and very thin beds of chert. The Smoky Member is 670 feet thick and consists mainly of blocky- to massive-splitting limestone, with little or no chert and silty limestone.

In the project area the Nopah Formation is partially exposed for a length of about 9 miles on the west flank of the Belted Range, in several relatively small areas on the east flank of the Belted Range, and in the southeast corner in the vicinity of Oak Spring Butte. It may have a total thickness of at least 3,000 feet.

LITHOLOGY

The Dunderberg Shale Member, which forms the basal member of the Nopah Formation within the project area, crops out in a single locality on the west flank of the Belted Range, where an incomplete section occurs within a small fault block. The exposed part may be about 200 feet thick. The contact with the underlying Bonanza King Formation is not exposed, and there appears to have been at least minor gravity slide or thrust movement along the contact with the overlying Halfpint Member. The lower part of the section exposed consists of very fissile reddish-brown shale and scattered thin beds of medium-gray limestone that weather to brownish gray. The limestone becomes relatively abundant in the upper part of the section and occurs in thin nodular to wavy beds resembling the basal part of the Halfpint Member.

The Halfpint Member may be about 1,900 feet thick in the Belted Range, but the member cannot be measured accurately because of severe crumpling of the beds and some fault displacement.

The Halfpint Member is composed largely of thin-bedded medium-gray limestone and dolomite and intercalated laminae of silty limestone and dolomite which weather to reddish gray and brownish gray. The rock is mostly laminated and wavy bedded, and it breaks characteristically into fairly large platy or flaggy fragments. Thin lenses and nodules of medium- to dark-gray chert are rather abundant at many horizons, and locally the rock contains a large amount of secondary carbonate material which weathers to brownish gray along bedding planes and fractures and in small irregular masses. In the southeast corner of the project area and locally in the Belted Range, the member consists largely of limestone. In several small gravity slide or thrust plates in the Belted Range, the Halfpint Member is extensively broken, deformed, and dolomitized. The dolomite is secondary and may be structurally controlled, as shown by the dominantly limy character of equivalent rocks in the lower plates. The contact with the overlying Smoky Member is gradational and is placed in the zone where there is a change upward to thicker splitting beds and less chert. Locally along this contact, where some movement apparently occurred, the change in lithology is abrupt. In these areas the Halfpint Member has been altered to a sugary-textured dolomite that is buff to brown on weathered surfaces.

The Smoky Member is about 900 feet thick and is composed of thin- to thick-bedded light- to dark-gray limestone and dolomite; in many places it forms rather smooth-weathering massive-looking outcrops. In many exposures the member is characterized by conspicuous alternating bands of light- and dark-gray carbonate rock, the bands ranging from several feet to tens of feet in thickness. Laterally along strike, dolomite and limestone occur in highly variable proportions, although dolomite seems to predominate. The dolomite is fine to medium grained, sugary textured, and locally laminated to cross laminated. Some is obviously clastic in origin. In the Belted Range the Smoky Member is brecciated, fractured, and altered in many places and locally contains abundant light-gray to buff or brownish-gray relatively coarse-grained secondary dolomite that occurs along bedding planes and fractures and in small irregular masses. The member contains scattered small chert nodules, and some dolomite beds are characterized by light-gray to reddish-gray mottling. Well-developed stromatolites and *Girvanella* have been noted at several horizons in the Smoky Member in the southeast corner of the project area. They were not observed in the Belted Range, however, perhaps because they have been largely obliterated there by the extensive alteration that has taken place.

AGE

Most of the Nopah Formation in the project area, in the Groom district, and in the Halfpint Range (Barnes and Christiansen, 1967), is Late Cambrian in age. However, no fossils have been collected from the uppermost part of the Nopah Formation or the basal part of the Goodwin Limestone in this region, and the Cambrian-Ordovician boundary has not been accurately located.

ORDOVICIAN

POGONIP GROUP

DEFINITION AND DISTRIBUTION

The name Pogonip, which is taken from Pogonip Ridge in the Hamilton district 30 miles southeast of Eureka, Nev., has survived many changes in usage since it was introduced in 1878. In the Eureka area, Nolan, Merriam, and Williams (1956) used the term Pogonip Group for the rocks lying between the Cambrian Windfall Formation and the Eureka Quartzite of Middle Ordovician age, and it is in this sense that the name is now being used in the Nevada Test Site and adjacent areas. Nolan and his associates divided the group into three formations, the Goodwin Limestone, Ninemile Formation, and Antelope Valley Limestone, and these have been recognized and described in the project area. They are shown separately only in cross section.

The Pogonip Group (fig. 2) is exposed for a length of about 8 miles on the west flank of the Belted Range; it crops out in two small areas on the east flank of the Belted Range and is exposed for a length of about 2.5 miles in the trough of a syncline in the Carbonate Wash area. The Pogonip occurs also in one small exposure near the Oswald mine on the east boundary of the project area. Complete, well-exposed sections occur in both the Belted Range and Carbonate Wash areas; the section on the west flank of the Belted Range was measured by R. J. Ross, Jr., and L. A. Wilson. The area in which the section was measured is not surveyed, but it lies about 3 miles west-southwest of Belted Peak and 3.5 miles northwest of Cliff Spring. The Pogonip has a total thickness of about 3,000 feet.

GOODWIN LIMESTONE

The Goodwin Limestone is about 1,010 feet thick in the project area and can be divided into three units. The lower unit is about 400 feet thick and grades into the underlying Smoky Member of the Nopah Formation. This unit is composed of light- to medium-gray fine-grained limestone, dolomitic limestone, and dolomite, which are generally thin to thick bedded, but locally laminated to crosslaminated. It is silty in part and commonly weathers to a buff to orange hue. The limestone contains rather abundant lenticular gray chert and a small amount of intraformational conglomerate. The rock is most readily distinguished from the Smoky Member by its weathered color, but it is also characterized by a higher proportion of limestone, more abundant chert, and thinner splitting beds that form more distinctly ledgy outcrops. Generally it is a better ridge and cliff former.

The middle unit, about 230 feet thick, is a relatively soft sequence which forms topographic saddles and strike valleys. It is composed of silty limestone and calcareous siltstone, which weather to yellowish gray, and subordinate laminated to thin-bedded medium-gray limestone. Much of the limestone is lenticular to somewhat nodular. This unit has been widely recognized and described in southern Nevada but has not been named.

The upper unit, about 380 feet thick, is composed predominantly of laminated to thick-bedded medium-gray aphanitic to medium-grained limestone. The upper half contains rather abundant laminae of silty limestone that weathers to shades of yellow, brown, and red. Some silty limestone occurs in an irregular network resembling the "chicken-wire" pattern in the Antelope Valley Limestone. The thicker limestone beds

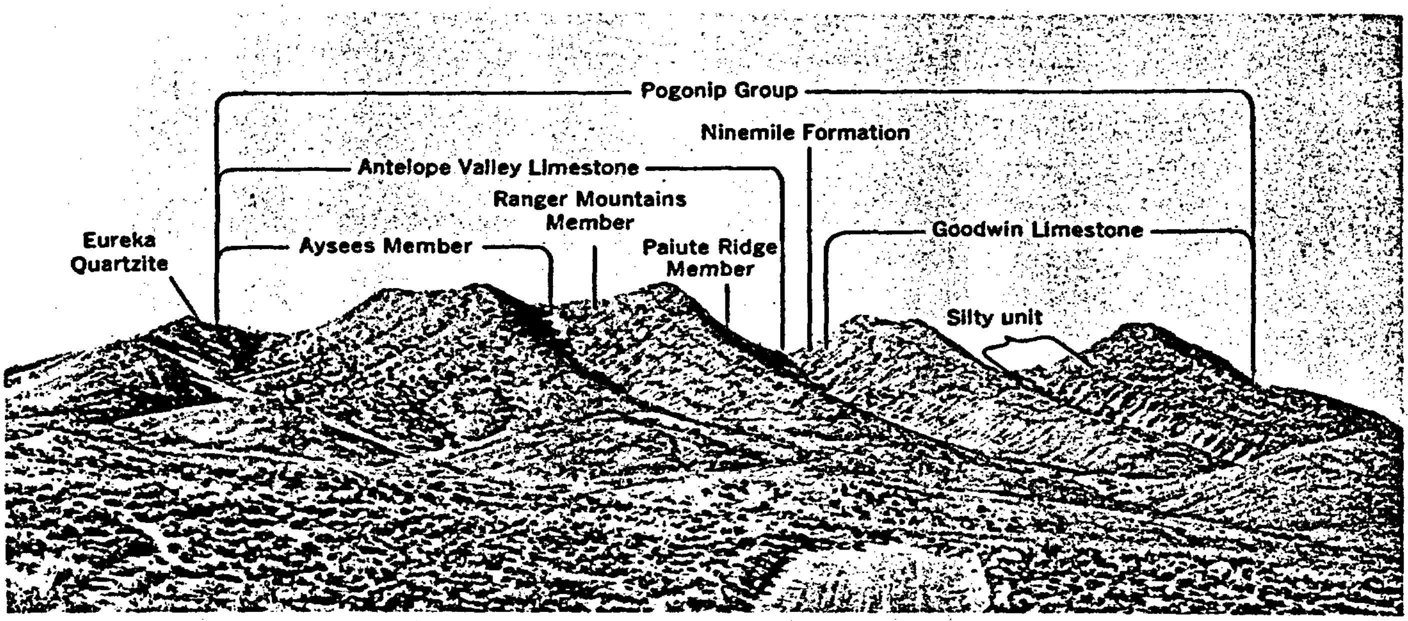

FIGURE 2.—Widest part of Limestone Ridge on the west flank of the Belted Range. This part of the ridge is composed of rocks of the Pogonip Group and Eureka Quartzite of Ordovician age. The Paleozoic rocks are in normal-fault contact with younger volcanic rocks in the foreground. View to the south.

in the upper part of the unit contain some intraformational conglomerate. Lenticular gray chert is common in the lower part of the section but sparse in the upper part.

The Goodwin Limestone is moderately fossiliferous and has been proved to be of Early Ordovician age. Fossils collected in the Belted Range, which were identified by R. J. Ross, Jr. (written commun., 1968), include two trilobites, *Hystricurus* sp. and *Symphysurina* sp., and a brachiopod, *Apheoorthis* sp.

NINEMILE FORMATION

In the project area the Ninemile Formation is a thin, easily eroded unit on which saddles and strike valleys have been formed, and exposures are generally poor. Its average thickness is about 300 feet, and it generally weathers to yellowish-gray. The Ninemile is composed of laminated to thin-bedded silty limestone and calcareous siltstone with intercalated fissile claystone. The unit includes a subordinate but rather variable amount of medium-gray fine-grained limestone that is commonly nodular to lenticular but may occur in persistent beds.

The Ninemile Formation is abundantly fossiliferous and is Early Ordovician in age, like the underlying Goodwin Limestone. Fossils from the Belted Range were identified by R. J. Ross, Jr. (written commun., May 9, 1968), and include two brachiopods, *Elliptoglossa* sp. and *Conotreta*? sp., collected from the southern part of Limestone Ridge on the west side of the range (USGS colln. D1409 CO), and two trilobites, *Ptyocephalus* sp. and *Lachnostoma*? sp., collected from an isolated exposure on the east side of the range (USGS colln. D1418 CO).

ANTELOPE VALLEY LIMESTONE

In the Belted Range and in the area north of Oak Spring Butte, the Antelope Valley Limestone contains three recognizable units that are correlated with the Paiute Ridge, Ranger Mountains, and Aysees Members. These members were defined by Byers, Barnes, Poole, and Ross (1961) and were named by them for features lying within and just east of the Nevada Test Site. They are well exposed in the Ranger Mountains, where they were mapped and described by Poole (1965), and have an aggregate thickness of about 1,530 feet. In the Belted Range the three members have an aggregate thickness of at least 1,600 feet.

The Paiute Ridge Member, about 400 feet thick, forms a bold ridge lying between two less resistant units. It consists of thin- to thick-bedded dominantly medium-gray fine-grained limestone with abundant silty limestone in regular laminae and beds and in the etched irregular networks which form a "chicken-wire" pattern and which are characteristic of this member (Byers and others, 1961). The silty limestone weathers to shades of brown and orange. Sparse small chert nodules occur in the member.

The Ranger Mountains Member is about 300 feet thick. It is a soft sequence characterized by poor exposures and consists of laminated to thin-bedded flaggy-splitting medium-gray fine-grained limestone interbedded with yellowish- to brownish- and reddish-gray siltstone and calcareous siltstone. The limestone may be wavy bedded to lenticular and nodular, and locally it contains numerous thin lenticular beds of black chert.

The Aysees Member, about 910 feet thick, can be divided into lower and upper parts. In the Belted Range the lower part is about 540 feet thick and is a ridge-forming sequence composed of medium-gray thin- to thick-bedded dolomite that is locally laminated and slightly silty. Weathered surfaces are pale gray to grayish and brownish orange. The upper part is about 370 feet thick and is a less resistant sequence composed of laminated to medium-bedded pale-gray- to yellowish-gray- and brownish-gray-weathering dolomite, dolomitic limestone, and limestone, which are slightly to strongly silty and which contain scattered small chert nodules. In places the silty rocks are gradational into siltstone and fine-grained quartzite. In the southeast corner of the project area the same division can be recognized, but in both lower and upper parts the rocks tend to be siltier and thinner bedded, and the yellowish-gray to brownish-gray weathering is more pronounced.

All members of the Antelope Valley Limestone are fossiliferous, and the formation is Early and Middle Ordovician in age.

Fossil material was collected from the formation in the Belted Range by R. J. Ross, Jr., and the authors and was studied by R. J. Ross, Jr., J. M. Berdan, J. W. Huddle, and O. L. Karklins.

Two trilobites were collected from a locality 165 feet above the base of the Paiute Ridge Member (USGS colln. D1501 CO) and were identified by R. J. Ross, Jr. (written commun., May 9, 1968), as *Ptyocephalus* sp. and *Nileus*? sp. This member also contains a few *Girvanella*.

A locality 460 feet above the base of the formation (USGS colln. D1500 CO), and in the lower part of the Ranger Mountains Member, yielded two trilobites, which were identified by R. J. Ross, Jr. (written commun., May 9, 1968), as *Ectenonotus* sp. and aff. *Miracybele* sp., and a number of conodonts, which were identified by J. W. Huddle (R. J. Ross, Jr., written commun., May 9, 1968) as:

Belodina aff. *B. inclinatus* Branson and Mehl
Ligonodina tortilis Sweet and Bergström
Oistodus sp.
Periodon sp.
Prioniodina sp.

According to Ross, this combination of trilobites and conodonts is characteristic of the *Orthidiella* zone in the Toquima Range of central Nevada, where it is found in the lower member of the Antelope Valley Limestone.

A collection from the Aysees Member of the Antelope Valley (USGS colln. D1089 CO) yielded three trilobites and three ostracodes. The trilobites were identified by R. J. Ross, Jr., and included *Illaenus* sp., *Pseudomera* sp., and *Isotelus*? sp.; the ostracodes were identified by J. M. Berdan and included *Schmidtella* sp., *Leperditella*? sp., and ?*Leperditella* sp. cf. *L. bulbosa* Harris. This unit also contains numerous examples of the gastropods *Maclurites* and *Palliseria*, but they were not collected.

Silty to sandy limestone lying 30–60 feet below the Eureka Quartzite yielded numerous fossils (USGS colln. D1499 CO) and might be considered as uppermost Antelope Valley or as equivalent to the lower part of the Copenhagen Formation.

This collection included pachydictyd(?), trepostome, and monticuliporid bryozoans, which were identified by O. L. Karklins (written commun., 1965) and indicate an age probably younger than the Oil Creek Formation and older than the Bromide Formation of Oklahoma. From the same collection R. J. Ross, Jr., identified the brachiopods *Syndielasma*? sp., *Valcourea* sp., and *Leptellina*? sp., as well as an indeterminate large fine-ribbed orthid and one trilobite, *Isotelus*? cf. *I. spurius* Phleger. From the same collection J. W. Huddle (written commun., March 9, 1965) identified the following conodonts:

Belodina cf. *B. ornata* (Branson and Mehl)
Cordylodus sp.
Dichognathus n. sp.
Distacodus sp.
Drepanodus sp.
Phragmodus undatus Branson and Mehl
Prioniodina? sp.
Trichonodella sp.
Zygognathus sp.

This conodont fauna is probably Middle Ordovician in age. The overall aspect of the fauna from the collection being described indicates, according to R. J. Ross, Jr. (written commun., May 9, 1968), a correlation with the upper part of the Antelope Valley Limestone exposed at Ikes Canyon in the Toquima Range of central Nevada; this interval may also be correlative in part with the lower beds of the Copenhagen of the northern Monitor Range in central Nevada.

EUREKA QUARTZITE

DEFINITION AND DISTRIBUTION

The Eureka Quartzite was named by Hague (1883, 1892) for outcrops near Eureka, Nev., but the type area was later transferred to the western base of Lone Mountain (Kirk, 1933) about 18 miles west-northwest of

Eureka, where the unit is better exposed and where it is 350 feet thick (Nolan and others, 1956).

The Eureka Quartzite and related quartzitic units of Ordovician age are remarkably persistent, as Ketner (1968) has recently pointed out, and extend from the Peace River in British Columbia to the Owens River in southern California. Ketner believes that the sand forming these deposits may have been derived fom Cambrian sandstone covering the Peace River-Athabaska arch in northern Alberta.

Small incomplete exposures of the Eureka Quartzite occur in several places around the margins of Yucca Valley. A complete section was measured and described by Byers, Barnes, Poole, and Ross (1961) and Poole (1965) in the Ranger Mountains, about 15 miles southeast of Yucca Flat. Poole reported that the section has a total thickness of 380 feet and consists of a basal quartzite unit 55 feet thick, which is dolomitic at the base; a carbonate unit 35 feet thick; a cross-laminated sandstone-quartzite unit 20 feet thick; a varicolored quartzite unit 150 feet thick; and an upper white quartzite unit 120 feet thick.

In the project area the Eureka Quartzite occurs mainly in the southeast corner, where it continues a short distance south into the Oak Spring quadrangle (Barnes and others, 1963); on the west flank of the Belted Range, where it occurs in four small exposures, the largest of which is about 1.3 miles long; at the Oswald mine in the northeast corner; and along the west base of a ridge 5 miles west-southwest of Mount Helen.

LITHOLOGY

A complete section of the Eureka Quartzite was measured by F. G. Poole, C. L. Rogers, and A. R. Niem about 1 mile northeast of Oak Spring Butte, where it is 315 feet thick and is readily divided into three major lithologic units that probably correlate with the five units recognized in the Ranger Mountains.

The basal 55 feet is poorly exposed and consists largely of laminated to thin-bedded silty limestone with smaller amounts of dolomite and dolomitic sandstone. It is correlated with the lower three units of the Eureka Quartzite in the Ranger Mountains and may be stratigraphically equivalent to the Copenhagen Formation in the Eureka district (F. G. Poole, oral commun., 1964). In the Belted Range the basal unit is largely covered by talus and may have been included in large part with the Antelope Valley Limestone.

The middle unit consists of 180 feet of varicolored quartzite, which weathers to shades of brown and gray, and, in the middle and upper parts, contains minor amounts of olive-gray argillite. The quartzite is generally laminated to thin bedded, but massive in appearance; many layers are cross laminated. The rock is composed of fine- to medium-sized well-rounded quartz grains, and it contains no feldspathic material.

The upper unit consists of 80 feet of homogeneous quartzite that is white to tan on weathered surfaces and only locally brown. It is characterized by poorly defined bedding, and only a few layers are laminated.

AGE

Fossils were found in the carbonate rocks of the basal unit, but no collection was made because material that was obtained previously from equivalent beds in the Ranger Mountains indicates a Middle Ordovician age for the formation. Vertical and horizontal animal borings are fairly abundant in the middle dark-weathering varicolored quartzite unit in the project area but are sparse in the upper light-colored quartzite unit.

ELY SPRINGS DOLOMITE

DEFINITION AND DISTRIBUTION

The Ely Springs Dolomite was named by Westgate and Knopf (1932) for exposures in the Ely Springs Range of the Pioche district, Lincoln County, Nev., where two measured sections total 525 feet and 770 feet, respectively. F. G. Poole (oral commun., 1964) has studied and remeasured the type section and has arrived at a total thickness of 685 feet. The formation is predominantly dark-gray dolomite and contains a considerable amount of chert.

Poole (1965) has measured and described the bipartite formation in the Ranger Mountains, a short distance east of the Nevada Test Site, where it is about 280 feet thick. The lower part (±130 ft) is dark- to medium-gray dolomite and the upper part (±150 ft) is medium-gray, light-gray-weathering dolomite. The lower part is unusually thin in the Ranger Mountains (F. G. Poole, oral commun., 1964).

Within the project area several small exposures of the Ely Springs Dolomite are present on the west flank of the Belted Range, but the least altered outcrops occur in the southeast corner of the area, north and northeast of Oak Spring Butte.

LITHOLOGY

A composite section of Ely Springs Dolomite totaling 340 feet was measured by F. G. Poole, C. L. Rogers, and A. R. Niem about 0.3 mile east-northeast of Oak Spring Butte. The formation may be comparably thick in the Belted Range to the north, but it is thinner in the Ranger Mountains and thicker in the Specter Range immediately south of the Nevada Test Site.

Near Oak Spring Butte the dolomite consists of two distinct parts. The lower part is 270 feet thick and is dominantly thin- to thick-bedded fetid medium- to dark-gray dolomite, which contains abundant poorly preserved small fragments of pelmatozoans and vaguely defined small round to ovoid forms that may be *Girvanella*. A brown and gray sequence about 30 feet thick composed of dolomitic sandstone and sandy dolomite crops out at the base immediately above the light-colored Eureka Quartzite. The lower part contains a persistent zone of layered chert about 20 feet from the top and numerous scattered blebs, lenses, and discontinuous layers of chert. In places part of the chert has been partly or wholly replaced by dolomite, and white secondary dolomite is rather common in small irregular masses as fracture fillings and as replacements of fossil debris.

The upper part is 70 feet thick and is composed of light- to medium-gray partly color-banded dolomite that is generally finer grained and more distinctly bedded than the dolomite in the lower part. It is laminated to thin bedded and may be slightly silty and clayey, as suggested by some yellow-weathering zones. It contains comparatively little chert. The contact between this part and the overlying Silurian and Devonian dolomite was difficult to determine because the quartz sand and oolitic or pelletal dolomite zones that characterize the top of the Ely Springs in many areas (F. G. Poole, oral commun., 1964) are not present. However, the Ordovician-Silurian boundary was approximately located by a persistent zone containing chain corals, which apparently mark the basal part of the Silurian in the Oak Spring Butte area.

AGE

Numerous fossils have been found in the Ely Springs Dolomite in other areas; these have established a Middle and Late Ordovician age for the formation (F. G. Poole, written commun., 1964). Within the project area it contains solitary and colonial corals, brachiopods, pelmatozoan debris, and *Girvanella*.

ORDOVICIAN AND SILURIAN DOLOMITE

Ordovician and Silurian dolomite crops out on Limestone Ridge about 2 miles southwest of Belted Peak in a gravity slide block or overthrust sheet that rests on Cambrian strata. (See "Structure.")

The lowermost strata consist of very cherty dark-gray dolomite, and they correlate with the Ely Springs Dolomite. The dark-gray dolomite grades upward to massive fine- to coarse-grained dolomite that is generally light to medium gray but locally is buff or reddish gray. It contains only minor chert. These strata are altered and recrystallized in most exposures, and bedding is indistinct or wholly missing. The uppermost beds must be of Silurian age, but because of the alteration they could not be separated from the Ely Springs.

Dolomite of Ordovician and Silurian age has also been mapped in the vicinity of the Oswald mine, on the northeast edge of the project area. The rock has been hydrothermally altered and is largely medium- to thick-bedded fine-grained light-gray dolomite with only minor chert and with no recognizable fossils.

SILURIAN AND DEVONIAN—DOLOMITE OF THE SPOTTED RANGE

The thick dolomite unit of Silurian and Early Devonian age that overlies the Ely Springs Dolomite is referred to informally as dolomite of the Spotted Range. The Spotted Range, where the dolomite is well exposed, lies immediately east of Mercury in the southeast corner of the Nevada Test Site.

LITHOLOGY

The dolomite of the Spotted Range is largely limited to the southeast corner of the project area, and a complete section, 1,415 feet thick, was measured by F. G. Poole, C. L. Rogers, and A. R. Niem about 1.5 miles northeast of Oak Spring Butte. It can be divided into two principal units, but these have not been shown on the geologic map.

The lower unit is about 465 feet thick and is probably correlative with unit C in the Ranger Mountains (Poole, 1965). It is composed of laminated to thick-bedded dolomite that is generally medium to dark gray on fresh surfaces but weathers to somewhat lighter shades of gray. In much of the unit bedding is rather vague and indistinct. The dolomite is locally aphanitic but generally fine to medium grained, and it contains scattered chert in blebs and lenses. Much of the chert occurs in two zones. One zone lies only 15–20 feet above the base of the section and contains rather regular thin lenticular beds of dark-gray to black chert that weathers to brown. The second zone lies 80–90 feet above the first and contains chert and dolomite in about equal proportions. The chert is gray but weathers to brownish gray; it occurs in rather irregular lenses and nodules that may coalesce to form a reticulate pattern. Where the section was measured the lower chert zone is about 25 feet thick and the upper zone about 60 feet thick, but the zones exhibit considerable variation in thickness owing apparently to partial replacement of the chert by dolomite in many places.

The lower unit is succeeded rather abruptly by a very thick sequence of dominantly light-gray dolomite of the upper unit. However, the boundary between light- and dark-gray dolomite is not everywhere stratigraphic, because it has been blurred by secondary bleaching, which may extend locally throughout the lower unit to its contact with the underlying Ely Springs Dolomite. The color change is very useful as a structural guide in areas where there has been no extensive alteration or bleaching and can be used to determine the relative movement and approximate magnitude of faults.

The upper unit is about 950 feet thick and contains three fairly distinct lithologic zones that may be equivalent to units, D, E, and F in the Ranger Mountains (Poole, 1965). To a point 215 feet above the base, the rock is a light-gray dolomite with some yellowish-gray and light-olive mottling. It is generally fine to medium grained, though coarsely crystalline in places, and massive in appearance, with rather indistinct bedding. The poorly bedded character is probably the result of extensive fracturing and recrystallization, which has left the bedding planes rather vaguely defined. Above this zone is a color-banded sequence about 500 feet thick consisting of light- to medium-gray and olive-gray dolomite, which is predominantly fine to medium grained and indistinctly bedded. This sequence is well preserved in places but locally is difficult to trace, owing to bleaching and alteration, and it is not a mappable unit. One zone only a few feet thick is medium-dark-gray laminated to thin-bedded dolomite containing poorly preserved spaghettilike forms that resemble *Amphipora*. The uppermost sequence, 235 feet thick, is a highly fractured light- to olive-gray dolomite that is finely crystalline and exhibits rather blocky splitting. The highest part of this sequence contains a zone about 35 feet thick that partly consists of sandy (quartz), color-banded, light- to dark-gray dolomite, which is locally fossiliferous.

AGE

The dolomite of the Spotted Range contains little good fossil material in the vicinity of Oak Spring Butte, but it can be dated by comparison with the equivalent rocks in and near the Nevada Test Site. Units A and B of the Ranger Mountains, which are Early Silurian in age, seem to be missing in the area north of Oak Spring Butte, and this absence indicates the presence of a major unconformity (F. G. Poole, oral commun., 1964). The remainder of the dolomite may be represented, however, and is probably late Early Silurian to Late Silurian and Early Devonian in age.

DEVONIAN

NEVADA FORMATION

DEFINITION AND DISTRIBUTION

The name Nevada Limestone was introduced by Hague (1892) and defined to include all the Devonian rocks in the Eureka district of Nevada. Merriam (1940) later divided the sequence on a faunal basis into the Nevada Formation and the Devils Gate Limestone. In the type locality the restricted Nevada Formation comprises almost 2,500 feet of limestone and dolomite.

In the project area the formation is exposed in the vicinity of Oak Spring Butte and to the northeast of Wheelbarrow Peak. North of Oak Spring Butte the Nevada forms a sharp ridge about 2.6 miles long and as much as 0.4 mile wide, where the base is well exposed but the upper part is missing. On the west the Nevada is faulted against rocks of the Pogonip Group. This incomplete section of the Nevada was measured by F. G. Poole, C. L. Rogers, and A. R. Niem about 1.5 miles north-northeast of Oak Spring Butte, where it is about 1,000 feet thick and can be divided into two informal units designated simply as lower and upper units. These have not been separately mapped, however.

LITHOLOGY

The lower unit is 430 feet thick. Its basal part, which is 180 feet thick, consists largely of light- to medium-gray and olive-gray finely crystalline laminated to thin-bedded sandy dolomite. About 80 feet above the base this sequence contains a zone 30 feet thick that is partly limestone and contains common brachiopods and cup corals. *Papiliophyllum elegantulum* Stumm, near the top of this zone, probably represents the upper part of the *Acrospirifer kobehana* and (or) the lower part of the *Eurekaspirifer pinyonensis* brachiopod zone; it indicates an Early Devonian age.

The upper part of the lower unit is 250 feet thick. At the base is a soft 40-foot-thick sequence which is poorly exposed in slopes and saddles and which consists largely of light-reddish-brown to brownish-red fissile siltstone. Interbedded with the siltstone are smaller amounts of calcitic partly sandy dolomite and silty pinkish-gray limestone. The siltstone is overlain by 210 feet of cherty limestone and dolomite, with minor laminae of silty limestone in the lower part. The limestone and dolomite are laminated to thick bedded, medium dark gray, and slightly sandy. The brownish-weathering chert is black on fresh surfaces and occurs in nodules and lenticular beds as much as 3 inches thick.

The incomplete section of the upper unit that was measured is 570 feet thick. The lower 300 feet is composed of alternating dolomite and quartz-sandy dolomite. The dolomite is light to medium gray, finely crystalline, and rather poorly bedded. The sandy dolomite weathers to brown and contrasts rather sharply with the gray dolomite. The quartz grains in the sandy dolomite are rather well sorted, well rounded, and fine to medium; they vary greatly in abundance. Locally the sandy dolomite and dolomitic sandstone are virtually quartzites. The lowest sandy beds contain some brown tubular structures that may be algae.

The upper 270 feet of the upper unit consists of light-, medium-, and dark-gray fine- to coarse-grained dolomite. The dark-gray dolomite contains coarsely recrystallized white dolomite in blebs and veinlets and as a replacement of fossil material; several layers contain abundant rodlike fossils that may be the stromatoporoid *Amphipora.*

AGE

The Nevada Formation is indicated by fossil evidence and stratigraphic position to be Early and Middle Devonian in age. In the Spotted Range, lying immediately southeast of the Nevada Test Site, it overlies dolomite of Early Devonian age and underlies the Devils Gate Limestone of Middle and Late Devonian age.

LIMESTONE AND DOLOMITE

An incomplete limestone and dolomite unit which is of Middle Devonian age, and which is at least partly equivalent to the Nevada Formation but represents a slightly more western facies, is exposed in the vicinity of Carbonate Wash in the southeast corner of the project area and occurs in one small area on the east flank of the Belted Range. These rocks appear to lie in the upper plate of a major thrust. (See "Structure.")

LITHOLOGY

The limestone and dolomite in the Carbonate Wash area, which was measured by F. G. Poole, C. L. Rogers, and Reginald Hammond, is 1,285 feet thick. It can be divided into three lithologic units.

The lower unit has an exposed thickness of 275 feet and is dominantly limestone and limestone conglomerate, which are commonly biohermal or biostromal and contain minor silty to clayey limestone in the matrix and in thin beds and partings. The limestone is aphanitic to coarsely crystalline, laminated to thick bedded, and medium to dark gray; it generally weathers to medium gray. It exhibits mostly flaggy to slabby splitting and locally contains a few black chert nodules. The conglomerate contains rounded to subangular fragments, which range in size from pebbles to boulders about 2 feet long and are lithologically similar to the flaggy limestone. The basal 20 feet of the unit contains the coral *Hexagonaria* sp., which may represent the lower part of the *Warrenella kirki* brachiopod zone.

The middle unit, about 445 feet thick, consists of relatively weak rock that weathers to a gentle slope. It is composed largely of interbedded limestone and silty limestone, but near the base it contains one layer of limestone conglomerate. The limestone is la to thin bedded, platy to flaggy splitting, aphanitic, fine grained, and medium gray to pinkish gray. The silty limestone is laminated, aphanitic to finely crystalline, and pale red to yellowish gray on weathered surfaces. A few small lenses and nodules of black chert occur locally in the unit. Corals and brachiopods from the upper 275 feet are representative of the *Warrenella kirki* faunal zone. Rocks near the base of the unit yielded styliolinids and a few specimens of *Nowakia.*

The upper unit, about 565 feet thick, is composed of relatively resistant rocks that form a ridge. The lower 230 feet consists of light-olive-gray to medium-gray limestone and, near the top, includes some thin dolomitized zones and minor cherty zones. The limestone is fine to coarse grained, laminated to thin bedded and commonly wavy bedded, and partly biostromal. The upper 335 feet consists almost wholly of dolomite, but near the base it contains local irregular zones of calcitic dolomite and unaltered limestone. The dolomite is medium to light gray, generally fine to medium grained, laminated to thin bedded, and locally cross laminated. On weathered surfaces it is yellow gray and light olive to medium gray and is commonly mottled or color banded. It contains common white coarse-grained dolomite in wisps and veinlets, and locally it is highly fractured and brecciated, with indistinct bedding.

About 350 feet above the base of the upper unit, F. G. Poole (written commun., 1967) collected silicified specimens of *Stringocephalus* sp., a large brachiopod that marks the upper part of the Nevada Formation in the Eureka district.

About 5 miles southeast of Belted Peak there is a small exposure of light-gray laminated to thin-bedded limestone that contains abundant rod-shaped fossils suggesting *Amphipora* of Middle to Late Devonian age (F. G. Poole, oral commun., 1963). It may be correlative with the formation being described.

DEVONIAN AND MISSISSIPPIAN

ELEANA FORMATION

DEFINITION AND DISTRIBUTION

The Eleana Formation was named by Johnson and Hibbard (1957) for incomplete exposures in the Eleana Range on the west margin of Yucca Flat. The formation has a probable minimum thickness of 7,700 feet and is known only by a composite of partial sections (Poole and others, 1961). Owing to structural complexities and partial cover by younger rocks, correlation of these sections is difficult and is considered tentative.

The Eleana Formation is exposed in the southeast corner of the project area and in several small areas on the east flank of the Belted Range. Rocks that probably correlate with the Eleana occur on the west flank of the Cactus Range, and these are described separately.

LITHOLOGY

The Eleana Formation contains six distinctive units in the southeast part of the mapped area. These correspond to units A–E and G of Poole, Houser, and Orkild (1961). Units A, B, C, and the lower part of D have been studied and remeasured by F. G. Poole, C. L. Rogers, and Reginald Hammond.

Unit A is about 112 feet thick and rather heterogeneous. It consists of limestone conglomerate, limestone, silty to sandy limestone, calcareous sandstone, sandstone, and quartzite. A 60-foot-thick zone near the base resembles a reef complex and consists of fossiliferous limestone conglomerate, limestone, and sandy limestone. The rock is medium to light gray, commonly mottled, and aphanitic to coarsely crystalline; the conglomerate contains subrounded boulders as much as 4 feet in diameter. Another major zone consists of platy- to flaggy-splitting beds of limestone and silty limestone, which are medium gray to pale red or brown, laminated to thin bedded, and aphanitic to finely crystalline. The quartzite is light gray to olive gray, medium grained, and laminated to thin bedded, and it weathers to brown and yellowish-to light gray; it commonly contains small cavities that may have formed by the leaching of carbonate material.

Unit B, about 1,230 feet thick, is composed largely of argillite and quartzite. The basal 240 feet, which appears to be gradational into the underlying unit A, consists of flaggy- to platy-splitting laminated to thin-bedded limestone, silty to sandy limestone, limy sandstone, laminated limy argillite, subordinate quartzite and limestone conglomerate, and, in the lower part, minor chert. Dominantly orange-pink to grayish-orange shaly partly calcareous argillite and sparse thin beds and lenses of quartzite form the upper part. These basal beds are overlain by about 470 feet of yellowish-brown to medium-gray and light-olive-gray thinly laminated shaly argillite that weathers to various shades of pale yellowish brown, gray, and green. The lowermost beds contain argillite, which locally grades to fine-grained quartzite, and in some places they contain sparse fine- to medium-grained quartzite. The argillite contains many cubic iron oxide pseudomorphs after pyrite and, in many places, small sinuous markings on bedding surfaces that are probably worm trails. The shaly argillite sequence is overlain by about 520 feet of dominantly platy- to slabby-splitting very fine to fine-grained quartzite with subordinate coarser grained quartzite and many shaly argillite partings. A few plant stem imprints occur throughout the sequence, and abundant convolute laminae occur in the fine-grained quartzite. These upper beds contain features characteristic of turbidites (F. G. Poole, oral commun., 1965).

Unit C, 430 feet thick, consists of quartzite, subordinate conglomerate, and minor argillite. The quartzite is olive gray but weathers to brown. It is fine to medium grained, laminated to very thin bedded, and characterized locally by convolute laminae. The rock exhibits partly flaggy to slabby splitting, and it contains scattered granules and pebbles of chert and sparse plant stem imprints. The conglomerate weathers to olive gray and contains rounded to subrounded pebbles of chert, quartzite, and argillite as much as 2 inches long, which are set in a vitreous quartzitic matrix.

The lithologic descriptions of units D, E, and G are based on a report by Poole, Houser, and Orkild (1961).

Unit D is about 400 feet thick at Carbonate Wash and is predominantly grayish-orange to yellowish-brown laminated argillite; it also contains numerous beds of pale-brown to grayish-brown quartzite that is fine to coarse grained, thin bedded, and characterized by abundant convolute laminae and small-scale cross strata. Many of these beds contain features characteristic of turbidites (F. G. Poole, oral commun., 1965). These rocks generally weather into sharp elongate fragments and form steep rubble-covered slopes. According to F. G. Poole (oral commun., 1965), the original measured thickness of 520 feet (Poole and others, 1961) for this unit may be excessive, inasmuch as the unit was originally measured in an area where faulting has probably resulted in some duplication of beds. A nearby section that may be nearly complete was later measured by Poole and C. L. Rogers and found to have a thickness of about 375 feet.

Unit E, about 2,400 feet thick, consists largely of argillite with minor interbedded quartzite. The argillite is yellowish brown to pale red and greenish gray to dark gray and is laminated to thin bedded. The quartzite is similar to that in unit D.

Unit G is 1,400 feet thick on Quartzite Ridge northwest of Yucca Flat, but only the lower part extends into the mapped area. The unit consists of quartzite, conglomerate, and argillite. The quartzite is brown to yellowish brown and gray, thin to thick bedded, and commonly cross-laminated. The conglomerate is brown to reddish brown and is composed of rounded fragments, as much as 2 feet long, of quartzite, chert, argillite, and limestone. The argillite is light brown to reddish, sandy, and laminated to thin bedded.

The exposure of the Eleana Formation located 5 miles southeast of Belted Peak (cross section *F–F′* of pl. 1) was visited by F. G. Poole (oral commun., 1963), who identified four major lithologic units that he tentatively correlates with the upper part of the Eleana Formation of Yucca Flat. If the correlation is valid, these units are much thinner than equivalent beds in the Yucca Flat area. The lowest beds consist of 100–200 feet of argillite and quartzite with minor conglomerite, and they may represent the top of unit E. The argillite exhibits numerous worm trails on bedding surfaces. These beds are overlain by 200–300 feet of quartzite and conglomerite that may correlate with unit G. The gravels are rounded to subangular and consist largely of greenish-gray and light- to dark-gray chert with some gray quartzite. Pebbles and cobbles are as much as 4 inches in diameter, but most are less than 1 inch. Unit H is probably represented by about 200 feet of light-gray to very light gray argillite that contains some plant stems. The highest beds exposed, which have an incomplete thickness of about 100 feet and which consist of gray cherty fossiliferous limestone, may represent the basal part of unit I.

Along the east side of the exposure southeast of Belted Peak the Eleana beds appear to be in fault contact with limestone that is probably Devonian in age (F. G. Poole, oral commun., 1963). The two units are separated by a limestone breccia with angular blocks as much as 2 feet long, but the rocks are poorly exposed in this area and their relations are obscure. It could not be determined whether the breccia is a sedimentary breccia in the Eleana Formation or a tectonic breccia related to faulting. However, it is more likely a tectonic breccia related to faulting, and if this is so, then a fault separates the two units.

The Eleana is also exposed in the area of Paleozoic rocks 3 miles northeast of Wheelbarrow Peak. The exposure is in a structural window and is very small, but because of its structural importance, it is slightly exaggerated on the geologic map (pl. 1). It consists largely of argillite and some conglomerate.

AGE

At its type locality in Carbonate Wash, located in the southeast corner of the project area, the Eleana Formation overlies rocks equivalent to the Nevada Formation of Early and Middle Devonian age and in the Yucca Flat area, lying a few miles to the south, underlies the Tippipah Limestone of Pennsylvania age. Its stratigraphic position and the available fossil evidence indicate an Early to Late Mississippian age for most of the formation; the basal part is Late Devonian in age (Poole and others, 1965). Initial fossil evidence suggested that the highest beds were of Pennsylvanian age (Poole and others, 1961), but recent fossil evidence indicates that these beds are Mississippian in age (Poole and others, 1965).

Numerous fossils have been collected from the Eleana Formation within the project area, and some of these have been of great importance in the solution of the age problem. Only fossil evidence not previously published will be presented here.

In the Carbonate Wash area F. G. Poole collected an impression of a bony plate from a quartzite layer in the upper part of unit A that was described by Edward Lewis (written commun., 1965) as being referable to an undetermined genus of archaic fish of the Class Placodermi, Order Arthrodira, and Infraorder Brachythoraci. It is probably Late Devonian in age.

According to F. G. Poole (written commun., 1968), a Late Devonian conodont assemblage, collected about 50 feet above the base of unit B and studied by J. W. Huddle, represents the *Palmatolepis crepida* zone of earliest Famennian age.

At the exposure of the Eleana Formation located 5 miles southeast of Belted Peak F. G. Poole (oral commun., 1963) found some silicified corals, crinoids, brachiopods, and other forms in the limestone of the highest beds, which may be correlative with unit I of the Yucca Flat area. The fossil collections from this locality were studied by Helen Duncan, E. L. Yochelson, I. G. Sohn, and J. W. Huddle and yielded some tiny gastropods, some ostracodes, a conodont, and three small horn corals. One of the horn corals might be a *Rylstonia*, and the other two appear to be zaphrentoids. However, neither the gastropods nor the corals are sufficiently well preserved to indicate whether they are Mississippian or Pennsylvanian in age.

ELEANA(?) FORMATION IN THE CACTUS RANGE

Most of the Paleozoic rocks exposed in the Cactus Range resemble the Eleana Formation and may correlate with it, although this has not been proved. A few hundred feet of strata is exposed and has been briefly examined, but the rocks have not been measured or studied in detail. They are uniform structurally—the beds strike north to north-northwest and dip 10°–20° W.

The rocks are composed chiefly of rusty-weathering thick-bedded to rather indistinctly bedded conglomerate that contains gray, brown, black, and green chert and quartzite fragments. Locally, the fragments also include micaceous quartzite and siltstone, phyllite, schist, quartz, pegmatite, and granite or related rocks of plutonic origin. Many of the metamorphic rocks resemble the Precambrian Stirling Quartzite in lithology. The fragments are well rounded to subangular and range in size from pebbles to boulders 3–4 feet in diameter. Some of the layers are strongly cemented with silica and can be described as conglomerite, but most of the rock contains little matrix and is friable, and as a result it disintegrates readily to form a rubble-covered slope. The conglomerate contains a few thin layers of argillite and a few beds of gray medium- to coarse-grained quartzite. The argillite contains plant stem imprints in a few places.

In the northernmost outcrop, about 3 miles south-southwest of Cactus Spring, the uppermost rocks contain two zones of predominantly limestone breccia whose aggregate thickness is about 250 feet. The carbonate fragments, which range in size from pebbles to boulders as much as 4 feet in diameter, are distinctly angular in contrast to the noncarbonate fragments. The fragments are largely monolithologic and consist of laminated to thin-bedded medium- to dark-gray silty limestone, with minor granular dolomite. The silty limestone contains sparse fossil material that includes linguloid brachiopod fragments and rodlike fossils resembling *Amphipora*, the latter suggesting a Middle to Late Devonian age. Immediately west of the area being described, the rocks of the Eleana(?) Formation are in fault contact with laminated to medium-bedded silty dolomite, which closely resembles the carbonate clasts in the limestone breccia and which has been tentatively assigned to the Halfpint Member of the Nopah Formation. This dolomite crops out adjacent to a small granite stock to the west and exhibits some contact metamorphic effects, such as recrystallization to dolomite and a rather fine color banding in shades of light to medium and dark gray. The outcrop has been exaggerated on the geologic map.

The Eleana(?) rocks in the Cactus Range are coarser than any Eleana within or near the Nevada Test Site that has been studied (F. G. Poole, oral commun., 1965), and this feature, together with the resemblance of the rock to fanglomerate and the angularity of the limestone fragments, suggests that the Cactus Range was near the source of the Eleana Formation. However, there is a marked difference between the limestone breccia and the cherty conglomerite, and the latter must have formed under somewhat different conditions. It contains abundant well-rounded clasts, which are widely diverse in lithology and which must have been transported a considerable distance.

MESOZOIC(?) GRANITE

Leucogranite of probable Mesozoic age intrudes Paleozoic limestone 2 miles south-southwest of Urania Peak (not shown on pl. 1) in the Cactus Range and Precambrian quartzite about 3 miles southeast of Quartzite Mountain in the Kawich Range. The granite masses in both areas are almost completely covered by alluvium, and their dimensions are unknown. The Precambrian rocks near Quartzite Mountain have been metamorphosed for a distance of nearly a mile adjacent to the granite outcrops, indicating that the intrusive mass in that area may be rather large.

The granite in both areas is pink to salmon colored, hypidiomorphic granular, and medium to coarse grained. The rocks are devoid of mafic minerals and contain only a fraction of a percent muscovite. Quartz makes up 30–40 percent of the volume, and the remainder is almost wholly perthite. Plagioclase averages less than 1 percent in the rock from the Kawich Range and less than 5 percent in the rock from the Cactus Range.

Alaskitic granite at Goldfield intrudes shale of Cambrian age and is probably of Cretaceous age (Ransome, 1909). The granite rocks in the Cactus Range and near Quartzite Mountain are probably also of Cretaceous age inasmuch as they are not known to intrude Tertiary strata and, in contrast with all known Tertiary intrusive rocks in the area, they are equigranular and coarsely crystalline. A postthrusting and postfolding age is suggested by the lack of features characteristic of dynamometamorphism. The feldspars are fresh and clear, and quartz grains show no strain effects.

TERTIARY

The mapped area includes some of the thickest and best exposed Tertiary volcanic sections in the Great Basin. These volcanic rocks are chiefly ash-flow tuffs but they include thick piles of silicic lavas and several sequences of interbedded ash-fall tuff and tuffaceous sedimentary rocks. The volcanic rocks form a composite section over 20,000 feet thick and range in age from about 27 to 7 m.y. with the oldest rocks exposed in the northern part of the area and the youngest (the Thirsty Canyon Tuff) exposed in the southern part. The area contains five major volcanic centers and parts of two others (pl. 1) that gave rise to thick sections of welded tuff and large volumes of silicic lava.

In order to keep cartographic units at a reasonable minimum for the small-scale map, many lavas and tuffs are combined into single map units. These composite map units include (1) lavas that show marked chemical similarities throughout the area, and that appear to be closely related in time, (2) ash-flow tuffs and ash-fall tuffs that form a mappable interval between easily recognizable marker beds, and (3) widespread sedimentary rocks of late Miocene age which are interbedded with volcanic rocks of limited areal extent and which reflect a comparative lull in volcanic activity.

The terminology used in this paper to describe ash-flow tuffs is that of Smith (1960) and Ross and Smith (1961); the size classification of pyroclastic fragments is that of Fisher (1960); and the classification for quartz-rich igneous rocks (tuffs and lavas) is that of O'Connor (1965). The textural terms are those of Williams, Turner, and Gilbert (1954).

FANGLOMERATE

Weakly cemented fanglomerate composed entirely of pre-Tertiary debris crops out in several fault blocks west of Mount Helen. The fanglomerate consists principally of angular cobbles and boulders of quartzite derived from the Stirling Quartzite, and it locally contains a few fragments of fossiliferous carbonates from Wood Canyon Formation. Contacts with adjacent strata are poorly exposed, and only in one locality was it determined conclusively that the strata rest directly on pre-Tertiary rocks. This occurrence, the total lack of volcanic detritus in the rock, and the juxtaposition of the rock with downdropped volcanic rocks dated elsewhere at 25 m.y. of age indicate conclusively that the rock predates the volcanic activity in the mapped area.

The source areas for the detritus probably were very close to the present outcrops; however, these have been downfaulted and blanketed by volcanic strata. The direction from which the detritus was derived can generally be determined from the wedge shape of most of the deposits.

MONOTONY TUFF

The oldest volcanic rock is an ash-flow tuff of late Oligocene age herein named Monotony Tuff for Monotony Valley in the extreme eastern part of the mapped area. The best exposures and the type locality are along the northeastern flank of Monotony Valley about 2.5 miles south-southeast of the Oswald mine, in secs. 2 and 3, T. 3 S., R. 53 E., where the tuff is about 2,300 feet thick, rests on Ordovician strata, and is overlain by the Miocene Shingle Pass Tuff. In the mapped area, the Monotony is perhaps the most widespread of the ash-flow sheets exposed; it occurs throughout the northern two-thirds of the area, where it has an average thickness of at least 1,000 feet. Beyond the mapped area, it was traced 50 miles northeastward to the Pancake Range (fig. 3) and 35 miles eastward to the northern Pahranagat Range. The west edge of the sheet is inferred to be near the west boundary of the mapped area. The northern edge is very indefinite. The south boundary, though vague because the contact between Paleozoic and Tertiary rocks is deeply buried in the region of Pahute Mesa, is inferred to be near the latitude of Quartzite Mountain and Gold Flat. At least one lobe of the sheet extended as far south as Groom Pass southeast of Quartzite Mountain within the Nevada Test Site, at lat 37° 12′ and long 116°.

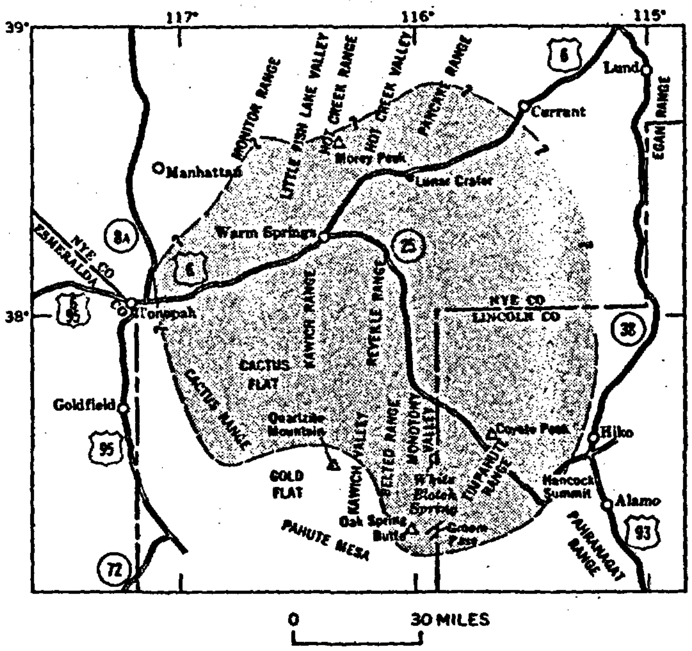

FIGURE 3.—Distribution of Monotony Tuff; edge of shaded area has queries where extension is doubtful.

Despite the indefinite northern edge, the Monotony Tuff has been correlated with reasonable certainty in an area about 100 miles in diameter. On the basis of an average minimal thickness of 500 feet, the unit had an astounding volume of more than 700 cubic miles.

The location of the vent area of the Monotony is uncertain, but it may have been the southern Pancake Range, as suggested by the following: (1) the southern Pancake Range lies near the center of the distribution area, and (2) the thickest sections known in the area—probably 2,000–4,000 feet thick—are exposed at the south end of the range and along the east flank of the Reveille Range. Whether these thick deposits lie within an area of collapse is unknown, but several arcuate post-Monotony Tuff collapse features are present farther north in the Pancake Range near Lunar Crater. The features include "scallops" similar to those around the edges of the Timber Mountain caldera (P. P. Orkild, oral commun., 1966), which probably are related to volcano collapse. Detailed mapping is necessary to determine the boundaries of the collapsed area and to define the volcanic history, but with the data now available, the Pancake Range seems the best choice for the source of the Monotony Tuff.

In outcrop the tuff characteristically weathers to rusty brown or green brown and forms gentle slopes and valleys between underlying and overlying more resistant rocks. The soft weathering habit results in a general paucity of good outcrops, although a few do occur as in Monotony Valley and in the Cactus Range just east of Wellington Hills where the tuff forms steep slopes along a receded fault scarp. The rock is densely welded in most exposures, but eutaxitic structure is vague, and the rock weathers to massive joint-controlled spheroidal outcrops and hoodoos very similar to weathered granite. Fresh rock is extremely scarce, but shards are visible in nearly all thin sections. Within the mapped area the Monotony appears to comprise a single compound cooling unit that is completely devitrified, mostly as a result of cooling-history crystallization (Smith, 1960, p. 152). In many other areas, however, part of the devitrification resulted from hydrothermal alteration which occurred after cooling and which commonly affected the mafic minerals. East of the mapped area, for example at Coyote Peake in the Timpahute Range (fig. 3), the rock is fresh and only partly devitrified, and two cooling units occur. These units are each 200–250 feet thick and are reddish gray to brown in their lower parts and light gray to very pale brown in their upper weakly welded tops.

Within the bombing and gunnery range the rock has a nearly constant phenocryst assemblage that shows only slight variation laterally and vertically. The rock contains 45–60 percent phenocrysts consisting of 15–20 percent clear to slightly smoky quartz (as much as 7 mm in diameter, average 2–3 mm), 15 percent biotite (commonly as much as 5 mm across), 5–15 percent alkali feldspar, 50–60 percent plagioclase, and less than 5 percent pseudomorphs after pyroxene and hornblende. The high ratio of plagioclase to alkali feldspar, the abundance of biotite, and the large quartz and biotite grains are distinguishing mineralogical features. The Monotony Tuff is very similar to the Hiko Tuff of Dolgoff (1963) exposed near Hiko, Nev., and at Hancock Summit in the Pahranagat Range, where the Hiko overlies the Shingle Pass Tuff. Petrographic study is commonly necessary to distinguish the two rocks; the Hiko generally contains abundant sphene, which is extremely rare in the Monotony; it contains much more hornblende, lacks clinopyroxene, and has a lower ratio of plagioclase to alkali feldspar. Both rocks have about the same amount of phenocrysts and both have similar "granitic" weathering habits.

Hydrothermal alteration of the Monotony in the type locality and also in the Belted Range has resulted in the partial replacement of mafic minerals by chlorite, calcite, and iron oxide. Locally, in the southern Belted Range due west of White Blotch Spring, the rock is weakly mineralized and was heavily prospected for gold and silver during the 1930's. A small mine was operated there during the late 1930's or very early 1940's. In the Kawich Range, the rock is similarly altered and is locally very weakly mineralized; the plagioclase phenocrysts are commonly albitized. In the Cactus Range, where the Monotony was subjected to various types of hydrothermal alteration, no fresh rock is known. The megascopic appearance of most of the altered rock, however, does not differ greatly from that of relatively fresh rock exposed in the other ranges. Thin sections show that few crystals have entirely escaped alteration. Alkali feldspar phenocrysts are commonly albitized and partly replaced by epidote; plagioclase phenocrysts are extensively sericitized, albitized, and partly replaced by calcite. Mafic phenocrysts are altered to chlorite, iron oxide, and locally calcite; groundmass constituents are altered to epidote, chlorite, sericite, calcite, and iron oxide.

Chemical analyses of five samples of relatively unaltered Monotony Tuff from widely separated areas are given in table 4. According to the classification for quartz-rich igneous rocks proposed by O'Connor (1965, p. B79–B84), the rock plots in the fields of rhyodacite and quartz latite (fig. 4).

Potassium-argon ages of 26.1 ± 0.71 and 27.6 ± 0.8 m.y. (average of two splits) were obtained from samples of Monotony Tuff from outcrops in the Belted and Timpahute Ranges respectively (table 5). These determinations indicate a late Oligocene age based on the time scale of Kulp (1961, p. 1105–1114) and the Geological Society of London (1964), or an early Miocene age based on the scale of Evernden, Savage, Curtis, and James (1964, p. 167). It is here considered late Oligocene in age.

PROBLEM OF CORRELATION

Recent reconnaissance mapping and stratigraphic studies by the authors and others north and east of the Nellis Air Force Base Bombing and Gunnery Range indicate that the lithology of the Monotony Tuff is far from unique. Tuffs that are megascopically and petrographically nearly identical with the Monotony occur both above and below the Monotony Tuff. The oldest of these probably exceeds 30 m.y. in age (F. J. Kleinhampl, written commun., 1966), and the youngest, the Hiko Tuff (Dolgoff, 1963, p. 885–888), is probably about 25 m.y. In most areas a definite correlation can be made on the basis of stratigraphic succession; however, in isolated fault blocks and in areas where units are missing, the problem of correlation is a major one. The Needles Range Formation (Mackin, 1960) in eastern and central Nevada (Cook, 1965) can be distinguished from the Monotony only by detailed study of many thin

TABLE 4.—*Chemical analyses and norms of Monotony Tuff*

[Analyses by P. L. D. Elmore, S. D. Botts, G.W. Chloe, Lowell Artis, and H. Smith, by rapid method described by Shapiro and Brannock (1962). Norms determined from chemical analyses recalculated to 100 percent minus H_2O+CO_2]

Sample	1	2	3	4	5
Laboratory No.	160722	160980	160808	160969	159664
Field No.	BP-28	QM-71	R-16E	M-220	PL-2374
Chemical analyses					
SiO_2	66.4	67.3	67.2	65.3	66.2
Al_2O_3	15.2	15.4	16.1	15.6	15.4
Fe_2O_3	2.5	2.4	2.8	2.3	2.0
FeO	.87	1.5	.64	2.0	1.1
MgO	.70	1.6	.73	1.5	1.6
CaO	4.0	2.9	2.9	3.3	3.1
Na_2O	2.5	3.0	2.6	2.8	2.8
K_2O	3.6	3.4	4.4	3.5	3.4
H_2O-	.87	.12	.72	.19	1.2
H_2O+	1.4	1.6	1.3	1.8	2.4
TiO_2	.43	.50	.43	.55	.46
P_2O_5	.13	.13	.15	.15	.14
MnO	.04	.08	.03	.13	.07
CO_2	1.2	.20	<.05	1.2	<.05
Sum	100	100	100	100	100
Norms					
Q	29.9	29.1	29.0	27.0	29.4
or	22.1	20.5	26.5	21.3	20.9
ab	21.9	25.8	22.4	24.4	24.6
an	19.7	13.8	13.7	15.8	15.0
C	.2	1.9	2.2	1.6	1.9
en	1.8	4.1	1.9	3.8	4.1
fs	------	.1	------	1.1	------
mt	1.8	3.5	.9	3.4	2.5
hm	------	------	2.2	------	.3
il	.8	1.0	.8	1.1	.9
ap	.3	.3	.4	.4	.3

Sample, locality and description

[Numbers following mineral names are percent of total phenocrysts]

1. In Belted Range 2 miles southwest of Belted Peak. 52 percent phenocrysts: quartz 33, plagioclase 53, alkali feldspar 2, biotite 8, calcite 2. Crystals are mostly fragments, quartz shows some resorption.
2. In southern extension of Kawich Range 2½ miles northwest of Gold Reed. About 50 percent phenocrysts with about same proportions of minerals as sample 1. Rock is very fragmental—most of the crystals are broken.
3. East flank of Kawich Range, 1 mile north of Cedar Pass. Same crystal content as sample 1.
4. Southwest flank of Cactus Range, 1½ miles north-northeast of O'Briens Knob. 56 percent phenocrysts: quartz 22.3, alkali feldspar 3.4, plagioclase 59.6, biotite 14.7.
5. East side of Jangle Ridge 7½-minute quadrangle, south of project area at lat 37°9.5', long 115°64.7'. About 50 percent phenocrysts in about same proportions of minerals as sample 1.

sections. The Needles Range Formation probably is at least 28 or 29 m.y. old (Armstrong, 1963) and may be as old as 31 m.y. (D. C. Noble and H. H. Mehnert, written commun., 1966). The Needles Range Formation generally contains more hornblende and pyroxene than the Monotony Tuff and locally contains abundant sphene which is extremely sparse in the Monotony.

The possibility seems good that several major ash-flow sheets of Monotony lithology were erupted from centers in central and eastern Nevada in late Oligocene and early Miocene time. Two of these are the Monotony

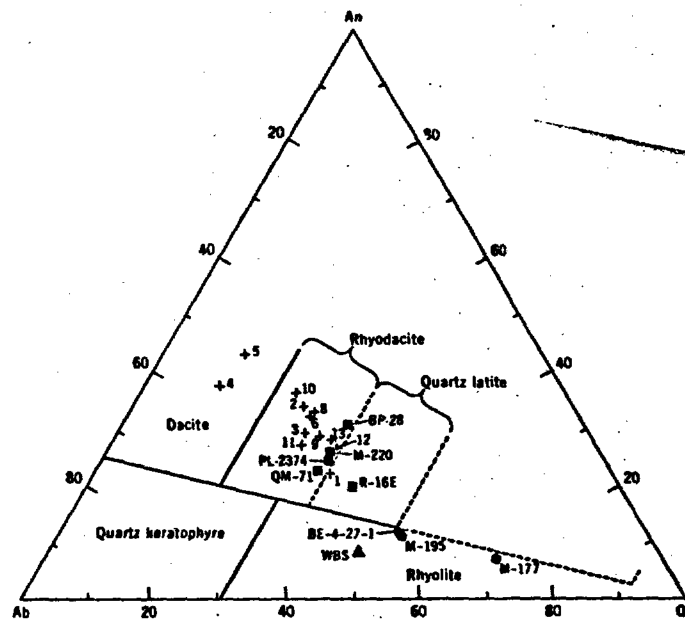

FIGURE 4.—Plot of normative albite-anorthite-orthoclase ratios. Squares, Monotony Tuff (from table 4); triangle, tuff of White Blotch Spring (average of four norms of samples 2, 3, 5, and 6 from table 6); crosses, lavas of intermediate composition exclusive of rocks from Mount Helen (from table 7); dots, rocks from Mount Helen (from table 8). Only rocks with more than 10 percent normative quartz are plotted. Fields are from O'Connor (1965).

Tuff and Needles Range Formation. Additional units will be found, with detailed mapping, that at present are correlated with either the Monotony Tuff or the Needles Range Formation largely on the basis of gross megascopic and petrographic features. Detailed mapping of the Tertiary strata in all the ranges in central and eastern Nevada is necessary before this stratigraphic problem can be resolved.

SEDIMENTARY ROCKS OF CEDAR PASS AREA

Clastic and tuffaceous sedimentary rocks more than 400 feet thick are exposed near Cedar Pass on the east side of the Kawich Range. The rocks are light to dark brown or light to dark gray; where hydrothermally altered they are white, pale green, or pale yellow. They consist of thin-bedded mudstone and sandstone with minor stratified shard tuff and pumice tuff.

The base of the unit is exposed 1.3 miles west of Cedar Spring. Here the clastic rocks rest on an erosion surface developed on Monotony Tuff, and they are overlain by highly altered welded tuff that correlates with either the tuffs of Antelope Springs or the Shingle Pass Tuff of early Miocene age. The sedimentary rocks are inferred to be also early Miocene in age.

TABLE 5.—*Summary of potassium-argon ages for volcanic strata in and near Nellis Air Force Base Bombing and Gunnery Range*

[Analysts: R. W. Kistler, H. H. Mehnert, R. F. Marvin, J. D. Obradovich, and Violet Merritt]

Sample	Field No.	Unit	Age (m.y.)	Analyzed material	Locality
		Thirsty Canyon Tuff:			
1	N80A	Labyrinth Canyon Member.	6.2±0.17	Alkali feldspar	4 miles northwest of Black Mountain; lat 37°18′ N., long 116°35′ W.
2	Age-11	Spearhead Member	7.5±0.20	do	3 miles north-northwest of Scrugham Peak, lat 37°38.5′ N., long 116°26′ W.
		Timber Mountain Tuff:			
3	Age-17	Ammonia Tanks Member	10.9±0.35	Biotite	Piapi Canyon; lat 36°59′ W., long 116°15′ W.
4	Age-16	do	10.8±0.40	Sanidine	Do.
5	Age-12	do	11.2±0.49	Biotite	6 miles west of south Timber Peak, lat 37°2′ N., long 116°35.5′ W.
6	Age-20	do	11.4±0.50	do	6 miles northwest of Scrugham Peak, lat 37°12′ N., long 116°28.5′ W.
7	Age-5	do	12.1±0.45	do	Massachusetts Mountain, lat 36°54.5′ N., long 115°58′ W.
8	GM-1-B	Rainier Mesa Member	11.3±0.3	Sanidine	Southeast corner of Pahute Mesa, lat 37°13.5′ N., long 116°16.7′ W.
		Paintbrush Tuff:			
9	Age-4	Tiva Canyon Member	12 4±0.46	Biotite	Piapi Canyon; lat 36°58.6′ N., long 116°14.5′ W.
10	BC-308	do	12.4±0.40	do	3 miles west of Rhyolite, Nev.; lat 36°54.5′ N., long 116°53′ W.
11	Age-2A	Topopah Spring Member	13.2±0.42	do	1 mile southeast of Topopah Spring; lat 36°54.5′ N. long 116°17′ W.
12	REA-62-SC	Rhyolite of Saucer Mesa	13.1±0.5	Nonhydrated glass.	Apache Tear Canyon; T. 6 S., R. 50 E. lat 37°24′ N., long 116°21′ W.
		Belted Range Tuff:			
13	WPN-23A	Grouse Canyon Member	13.8±0.6	do	Belted Range, Oak Spring Butte quadrangle; lat 37°20′30″ N., long 116°01′ W.
14	WPN-500	Rhyolite of Kawich Valley	14.8±0.6	Sanidine	Southern Belted Range; lat 37°21′10″ N., long 116°01′ W.
15	Age-24	Fraction Tuff	15.0±0.55	Biotite	Test well 8, 4,384 ft; lat 36°55.5′ N., long 116°16′ W.
16	Age-25	do	17.8±0.48	do	Trailer Pass, Kawich Range; lat 37°37′ N., long 116°19.5 W.
17	CS-817	do	15.7±0.5	Sanidine	Cactus Range; lat 37°45′ N., long 116°58′ W.
		do	16.4±0.5	do	Do.
18	EA-17-6	Dacite lava	18.7±0.7	Biotite	Gabbard Hills; T. 4 S., R. 48 E., lat 37°32′40″ N., long 116°32′ W.
		do	17.2±0.7	do	Do.
19 [1]	644-3	"Dacite vitrophyre" [2]	21.1±0.6	Biotite	South flank of Goldfield Hills.
20	OM-WBL	Tuff of White Blotch Spring	24.4±0.7	Sanidine	East flank of Monotony Valley; lat 37°45′ N., long 116°00′ W.
		do	25.2±0.8	Biotite	Do.
21	66-E-10	do	23.4±0.7	Sanidine	West flank of Kawich Range; lat 38°00′ N., long 116°34′ W.
		do	23.8±0.7	Biotite	Do.
22	438-6	do	21.5±0.5	do	Central Kawich Range; lat 37°56′ N., long 116°25′ W.
23	CS-796	do	22.9±0.7	Sanidine	Northern Cactus Range; T. 2 S., R. 46 E., lat 37°45′ N., long 116°52′ W.
		do	21.8±0.7	do	Do.
24 [1]	163-3	Shingle Pass Tuff	25.3±0.68	do	6 miles northeast of Belted Peak; lat 37°38′ N., long 115°59′ W.
25 [3]	15195-20	do	25.4±0.8	do	Southern Pancake Range; NE¼ sec. 6, T. 6 N., R. 54 E.
26	CS-869	Tuff of Antelope Springs	27.7±0.8	Alkali feldspar	Northern Cactus Range; lat 37°46′ N., long 116°53′ W.
		do	26.2±0.8	Biotite	Do.
27	MT-1	Monotony Tuff	27.4±0.8	do	Coyote Peak, Timpahute Range; T. 4 S., R. 56 E., lat 37°34′ N., long 115°40′ W.
		do	27.8±0.8	do	Do.
28 [4]	Age-26	do	26.1±0.71	do	West flank of Belted Range; lat 37°34′30″ N., long 116°06′ W.

[1] Samples collected by H. R. Cornwall.
[2] Named by Ransome (1909, p. 61); rock is a quartz latite welded tuff.
[3] Sample collected by Frank Kleinhampl.
[4] Hornblende and pyroxene are completely altered in this sample; biotite appears fresh but probably incipiently altered.

TUFFS OF ANTELOPE SPRINGS

A sequence of rhyolitic to rhyodacitic ash-flow tuffs crops out above the Monotony Tuff and beneath the tuff of White Blotch Spring in the western part of the project area. These rocks are here informally called tuffs of Antelope Springs after the excellent exposures near Antelope Springs in the Cactus Range, where the tuffs have a composite thickness of about 5,600 feet.

In almost all areas the tuffs of Antelope Springs are displaced and tilted by many normal faults, and nowhere is a continuous and unfaulted stratigraphic section available. Most of the rocks have been moderately to intensely altered by hydrothermal solutions. Rock textures normally used to interpret the cooling history of ash-flow tuffs (Smith, 1960) have been partially destroyed by this alteration. Also, the primary minerals (except quartz, apatite, and zircon) have been modified or replaced in the altered rocks. The rocks are generally drab, and many have greenish casts resulting from abundant secondary sericite, chlorite, and epidote in both phenocrysts and matrix. Low-silica varieties are yellowish brown where fresh but tend to be purplish gray where altered. The rocks are bleached to light gray, pink, or pale yellow adjacent to faults and intrusive masses where silicic and (or) argillic hydrothermal alteration has been intense.

Because these tuffs are generally faulted and altered, individual cooling units are very difficult to recognize and cannot be unambiguously correlated between separate areas. A broad tripartite division (lower, middle, and upper), based on the abundance of quartz and alkali feldspar and on color, was made in parts of the Cactus Range, but elsewhere the unit is undivided. Contacts between the three parts are depositional horizons not marked by bedded tuff or sedimentary rocks.

LOWER ASH-FLOW TUFFS

The lower ash-flow tuffs are 800+ feet thick in the southern Cactus Range. At the base the rock is pale green, weakly welded, and typically slabby weathering; it grades upward to greenish-gray densely welded columnar-jointed rock that contains about 15–20 percent phenocrysts. Phenocrysts of clear quartz and alkali feldspar averaging about 1.5 mm in diameter are conspicuous, but plagioclase and biotite, which are mostly replaced by white mica, are inconspicuous or unrecognizable megascopically. The greenish-gray rock is overlain by densely welded maroon to light-purple tuff characterized by conspicuous pink euhedral alkali feldspar phenocrysts. In six thin sections of this rock the penocryst content was 15–30 percent. The average phenocryst percentages are quartz, 16; alkali feldspar, 42; and plagioclase, 38. Altered locally.

All the lower welded tuffs merous greatly flattened sm to be green in the greenish-g light-purple and maroon ro generally less than 1 inch l cate wisps normal to the plane thin films in it. The rocks are notably poo fragments in contrast to some of the overlying upper tuffs of Antelope Springs that are similar in general appearance.

MIDDLE ASH-FLOW TUFFS

A decrease in abundance of alkali feldspar and quartz marks the lower boundary of the middle ash-flow tuffs, a sequence of rhyolitic to rhyodacitic ash flows that probably exceeds 2,300 feet in thickness, although nowhere is a complete section available. The rocks are well indurated, but it is not known whether the induration is everywhere due to original dense welding or to later postemplacement alteration. In an incomplete section that is well exposed on Antelope Peak 2 miles west-northwest of Antelope Springs, the lower 100–150 feet is dark gray to dark purple and locally brown, and it contains abundant dark-gray to black greatly collapsed pumice lapilli. The interval is generally very resistant and forms a dark cliff. Secondary epidote clusters as much as 1 cm in diameter are conspicuous. Phenocrysts make up about 20 percent of the rock and consist of 80 percent plagioclase, 5 percent quartz, and 12 percent alkali feldspar. Secondary epidote, chlorite, sericite, calcite, and clay occur in both the phenocrysts and groundmass of the rock.

The overlying rock, of which about 700 feet is exposed, is buff to light gray and weathers red to dark red. It is monolithologic, moderately altered, and pumice poor; it contains less than 10 percent phenocrysts. Tuff structures are very inconspicuous megascopically, but abundant shards and scattered tiny pumice fragments are clearly visible in thin section. Phenocrysts consist of plagioclase and minor iron-depleted biotite. Quartz is sparse to absent. The rock is massive in outcrop and weathers to steep slopes covered with angular blocky scree. Gradational contact relations observed near Antelope Peak indicate that this tuff and all the underlying tuffs of Antelope Springs may be part of a compound ash-flow tuff cooling unit.

A thick section of crystal-poor brown, tan, and gray densely welded rhyolitic(?) tuff is exposed between Roller Coaster Knob and Antelope Springs. These rocks, which are included with the middle tuffs on the basis of low quartz content, dip vertically to 25° SE.

It is not known to what extent these strata are repeated by faulting, and neither the base nor the top is exposed, but 1,900 feet probably is a conservative estimate of their total thickness. Phenocrysts, which make up about 5 percent of the rock, have been largely replaced by secondary minerals. The rocks are fractured and flow layered. Other exposures of this unit occur in Sleeping Column Canyon and on the northeast flank of Urania Peak, but they are not shown separately on plate 1.

A greenish-gray, greenish-brown, and locally brown slabby-weathering rhyodacitic welded tuff occurs locally above the typical crystal-poor tuffs just described. The rock contains 25–35 percent crystals, 82 percent of which are plagioclase and 12 percent, biotite. Biotite phenocrysts are preferentially oriented parallel to the plane of compaction. Pumice lapilli are either lacking or indistinct, and the rock is easily mistaken for a lava. The best exposures are about 2 miles northwest of Roller Coaster Knob, where an incomplete east-dipping section has an estimated thickness of 400 feet.

UPPER ASH-FLOW TUFFS

The upper ash-flow tuffs of Antelope Springs are about 2,000 feet thick at Antelope Springs. They appear to be much thicker in the northern Cactus Range, but they are intensely deformed there and thus cannot be measured accurately. The upper tuffs rest conformably on the middle tuffs in the southern part of the range and also in a single exposure on the northeast flank of Urania Peak (not shown on pl. 1). Between these two areas, about 3 miles southeast of Urania Peak, they apparently rest unconformably on the Monotony Tuff and contain lithic fragments and blocks of the middle tuffs of Antelope Springs. Only the upper tuffs are recognized in the northern Cactus Range. In contrast to the pervasive alteration of equivalent tuffs in the south and central parts of the range, the upper tuffs in the north are sufficiently unaltered to permit reliable modal analyses of phenocryst contents. In the descriptions that follow, the modal data given for the rocks in the north are inferred to apply to stratigraphically equivalent rocks in the central and southern part of the range.

At Antelope Springs, the upper tuffs comprise three distinctive zones, each of which probably represents an ash-flow cooling unit. The three zones are separated from each other by a few inches to several feet of bedded ash. The lower zone is about 1,000 feet thick and, in the lower half, consists of greenish-gray partially welded to densely welded tuff rich in lithic fragments of older welded tuffs, argillite, and quartzite; the upper half contains brown welded tuff rich in biotite, large red-stained quartz, and large pumice lapilli. The greenish-gray tuff also contains biotite, quartz, and pumice lapilli, but they are smaller and less abundant than in the overlying brown rock. The greenish-gray tuff contains less plagioclase and more alkali feldspar than the brown tuff.

The middle zone of the upper tuffs is about 300 feet thick and consists of slabby- to massive-weathered steel-gray to pale-purplish-gray densely welded tuff rich in lithic fragments of rust-colored carbonate and dacite lava. The rock contains conspicuous quartz and less biotite than the underlying brown tuff.

The upper zone is at least 700 feet thick and consists of a basal slabby-weathering pastel-green poorly welded tuff that grades upward to purple to light-brown densely welded tuff. This zone contains abundant small embayed quartz, alkali feldspar, and white pumice lapilli, and minor altered biotite. Plagioclase phenocrysts are completely altered to sericite and clay and are removed easily during weathering, leaving numerous small euhedral holes on weathered surfaces. Rocks in this zone form the prominent hogback ridges that mark the east edge of the Cactus Range north and south of Antelope Springs.

The stratigraphy of the upper tuffs of Antelope Springs in the northern Cactus Range is poorly understood owing to the structural complexity of that area. Most of the rock exposed probably correlates with the lower zone at Antelope Springs. The rocks equivalent to the lithic-rich lower part of that zone are generally brown to reddish brown, locally gray, and are composed of as much as 50 percent lithic fragments, some of which are 10 feet in diameter. The predominant lithic fragments are Monotony Tuff, argillite and quartzite of pre-Tertiary age, pink granite of Mesozoic(?) age, lower and middle tuffs of Antelope Springs, and porphyritic pilotaxitic dacite. Rocks equivalent to the brown tuff are widely distributed. They are generally brown with abundant yellowish-brown pumice lapilli and blocks but are locally purplish gray with light-gray pumice. They contain about 40 percent phenocrysts consisting of 15–28 percent quartz as much as 3 mm in diameter, 8–20 percent alkali feldspar, 50–56 percent zoned plagioclase, and 8–12 percent mafic minerals consisting of biotite and altered hornblende(?). Magnetite, zircon, and apatite are the principal accessory minerals. The rock generally contains 1–2 percent of small lithic fragments of welded tuff, dacite(?), and quartzite. Other welded tuffs of unknown stratigraphic position are poor in mafic minerals and contain as much as 43 percent alkali feldspar; still others are quartz latitic and contain as little as 7 percent quartz.

UNDIFFERENTIATED TUFFS

In the area between Stonewall Mountain and Wilsons Camp, numerous scattered exposures of quartz-bearing calc-alkaline welded tuffs are mapped as tuffs of Antelope Springs, undivided. Most of these rocks probably correlate with the upper tuffs just described. At Mount Helen, however, the tuffs exposed at the top of the sequence include two units, each about 750 feet thick, which are separated by at least a partial cooling break and which are not recognized with certainty at Antelope Springs. The lower unit is a densely welded purple tuff that contains abundant small grains of quartz, chatoyant alkali feldspar, minor biotite, argillized plagioclase, and indistinct pumice lapilli; it crops out on both sides of the mountain. This rock is overlain by a unit several hundred feet thick that is identical with the underlying rock except that it contains larger phenocrysts of quartz and alkali feldspar. The tuff with the smaller quartz grains closely resembles the tuff that forms the prominent hogback ridges at Antelope Springs, but it is different in that it lacks conspicuous white pumice lapilli and contains catoyant alkali feldspar. The tuff with the larger quartz grains apparently is not present at Antelope Springs or in the central core of the Cactus Range; it is thickest in the vicinity of Mount Helen. These tuffs, although they resemble some of the upper tuffs of Antelope Springs, are inferred to have been extruded from a different volcanic center, presumably the Mount Helen volcano.

SHINGLE PASS TUFF

The Shingle Pass Tuff was named by Cook (1965, p. 20) for a highly welded dark-red to pale-purple vitric ignimbrite exposed at Shingle Springs "just west of a dirt road that leads through Shingle Pass in the Egan Range, in sec. 8, T. 8 N., R. 63 E., Lincoln County." Scott (1965) later correlated the Shingle Pass with two to 10 chemically similar ignimbrites above the Needles Range Formation in the Grant Range, about 50 miles northeast of the Nellis Air Force Base Bombing and Gunnery Range. The name is used herein for several closely similar ignimbrites or ash-flow tuff cooling units exposed in the eastern part of the mapped area, some of which are probably the direct equivalent of units in the Grant Range. Whether any of the units, however, correlate with the single ignimbrite at the type locality remains to be proved by detailed mapping.

The rocks included in the Shingle Pass Tuff in the area of study lie between the Monotony Tuff and the tuff of White Blotch Spring. They are correlated with certainty only within the Belted Range, Monotony Valley, and as far south as the Jangle Ridge quadrangle in Nevada Test Site, where a s· ...sh called informally the "red ...d Christiansen, and Byers (1965). northward, northwestward, and east ward extension within the bombing and is extremely vague. Two hydrothermally and orange cooling units east of Quartzite M in the southern extension of the Kawich Range tentatively correlated with the Shingle Pass Tuff, and a unit north of Cedar Pass in the Kawich Range may be an equivalent, but because of intense hydrothermal alteration there, this correlation is uncertain. The rocks in both areas are shown on plate 1 as tuffs of Antelope Springs and Shingle Pass Tuff undivided. In Cactus and Gold Flats the interval between the Monotony Tuff and the tuff of White Blotch Spring is poorly exposed, and the few rocks that are exposed are also hydrothermally altered. They most closely resemble the tuffs of Antelope Springs that were extruded from the Cactus Range. Thus, the western limit of the Shingle Pass Tuff appears to lie somewhere between the Kawich and Cactus Ranges, and the southern limit is just south of the mapped area at about lat. 37°10′ N.

The tuff in the Belted Range and Monotony Valley occurs in a mosaic of fault blocks. It consists of four to possibly as many as seven separate cooling units. Most of the units average less than 100 feet in thickness and the combined sequence averages about 700 feet. Most of the flows are nonpersistent, a feature which probably reflects the great distance from the source area—presumably at least several tens of miles to the northeast. Probably only two or three of the thickest flows persist throughout the Belted Range. Everywhere the rocks are slightly to intensely hydrothermally altered and are more altered and fractured than the overlying tuff of White Blotch Spring. All the flows are dominantly red, orange, or grayish purple except for their chilled nonwelded bases which commonly are yellow, white, or green; all are densely welded, and most have conspicuous black to dark-greenish-black basal vitrophyres 5–20 feet thick. Nearly all units contain conspicuous lithophysal or gas-bubble zones that start below the contact between the basal glassy vitrophyres and devitrified rock and grade upward well into the devitrified densely welded interiors. Many of the lithophysae apparently formed in collapsed pumice fragments, but many others formed in shard tuff almost wholly devoid of pumice. In all the cooling units some pumice occurs, and eutaxitic structure is generally well defined. Several units have a vague planar structure that formed as a result of slight flowage after the units were emplaced.

With the exception of a red welded tuff about 50 feet thick that occurs locally near the base and contains fairly abundant small phenocrysts of quartz, the cooling units are quartz poor and contain 8–20 percent phenocrysts consisting of plagioclase and alkali feldspar—in ratios that range from 6:1 to about 1:4—biotite, minor quartz, sparse corroded pseudomorphs after olivine or clinopyroxene, and sparse hornblende. One cooling unit exposed along the east flank of Monotony Valley east of the mapped area contains biotite as the only mafic mineral in a crystal-poor base, but in a more crystal-rich upper part it contains both biotite and clinopyroxene. This unit is as much as 450 feet thick in Monotony Valley and at least 200 feet thick throughout the exposures in the Belted Range. It is the only unit in which pyroxene was an appreciable constituent in thin section. Allanite was noted in several thin sections from cooling units at the base of the sequence and in one thin section from an upper cooling unit exposed in Monotony Valley.

A potassium-argon date of 25.3 m.y. was obtained from an outcrop of Shingle Pass Tuff on the east flank of Monotony Valley, sampled by H. R. Cornwall, and 25.4 m.y. from an outcrop in the Pancake Range about 40 miles north of the Belted Range, sampled by F. J. Kleinhampl. These dates (table 5) indicate an early Miocene age.

The contact between the Shingle Pass and the underlying Monotony Tuff appears to be conformable throughout the Belted Range area. In places, however, the upper weakly welded top of the Monotony Tuff was completely removed by erosion prior to the deposition of the Shingle Pass. The gentle basal contact contrasts with the upper contact, which in most places is a surface of considerable relief marked by rubble zones containing boulders and cobbles of Shingle Pass Tuff as well as fragments derived from the Monotony Tuff.

LACUSTRINE SEDIMENTARY ROCKS OF THE CACTUS RANGE

Sedimentary rocks that include abundant siltstone and shale rest unconformably on a surface of considerable relief developed on tuffs of Antelope Springs and Monotony Tuff in the central Cactus Range. In this area the rocks are at least 800 feet thick, they are everywhere hydrothermally altered, and they are locally baked to dense dark-brown and black hornfels adjacent to intrusive masses. Where intensely silicified or argillized, the rocks are light gray to white, and where propylitized, they are brown to greenish brown.

The strata consist of siltstone, shale, sandstone, and ash-fall tuff; tuff is most abundant in the lower 300 feet. The upper strata, which are thin bedded, consist of approximately 20 percent coarse arkose and volcanic conglomeratic sandstone and 80 percent dark-gray silicified shale and siltstone. Most siltstone and shale beds are only a few inches thick, and most sandstone beds are less than 10 feet thick. The persistent thin and even bedding indicates deposition in water without vigorous currents, and a lacustrine environment is inferred. No fossils have been found in these strata. Although tuff of White Blotch Spring is not seen resting on them, these sedimentary rocks are inferred to be pre-White Blotch Spring in age. They probably accumulated in a lake that developed after collapse related to the withdrawal of magma to form the upper ash-flow tuffs of Antelope Springs. These sedimentary rocks probably were deposited over a broader area than that in which they now occur. They may have been removed from the northern part of the range and other areas during and after a stage of postcollapse doming.

TUFF AND RHYOLITE OF GOLD FLAT

Several isolated exposures of welded tuff occur in Gold Flat south of Coyote Cuesta along the northern flank of Pahute Mesa. The rocks crop out beneath rhyolite lavas that underlie the tuff of Wilsons Camp (p. 42). The stratigraphic position of the rock with respect to older volcanic strata in the project area is unknown.

The base of the tuff has not been observed, although a vitrophyre, probably basal, is present in several exposures. If the vitrophyre is actually at or near the base, the tuff is about 100 feet thick. The rock is pastel pink, red, and green on fresh fractured surfaces and weathers pinkish gray. It is densely welded, poor in pumice and crystals, and fairly rich in small rhyolite and andesite lithic fragments. Crystals make up 10–20 percent of the rock and include zoned plagioclase, alkali feldspar, quartz, and very sparse clinopyroxene, hornblende, and biotite.

Rhyolite crops out in a few isolated areas beneath the tuff of Wilsons Camp and above the welded tuff just described. The rhyolite has abundant tiny spherulites and is light gray to red, highly flow layered and laminated, brecciated at the base, and void of crystals. It weathers to small flat angular fragments.

TUFF OF WHITE BLOTCH SPRING

The name tuff of White Blotch Spring is applied to a sequence of quartz-rich welded tuff that crops out throughout the bombing and gunnery range. The sequence forms a readily mappable unit of strikingly similar strata from range to range, but slight differences in the type and abundance of phenocrysts are evident, and marked differences in nonbasal accumulations of lithic fragments indicate that the unit as mapped contains ash flows from different centers. Each center seemingly stamped its identity on its ash flows by means of lithic fragments that are representative of the crust through which the magma was erupted.

The Cactus Range and the northern Kawich Range (beyond the mapped area) have been identified as two of the centers of ash-flow eruptions. An unidentified third center is inferred near the east boundary of the mapped area. Potassium-argon dates indicate a slight progressive decrease in age of volcanic activity from east to west, but the similarity of strata strongly suggests a common substratum source.

The ash-flow tuffs exposed in the eastern, central, and western parts of the mapped area probably are juxtaposed at depth in the intervening valleys. They may also be juxtaposed in the ranges, but this cannot be demonstrated conclusively with the available data.

EASTERN PART OF MAPPED AREA

Rocks mapped as tuff of White Blotch Spring in the eastern part of the mapped area are distributed widely in the ranges on both sides of Monotony Valley, where they are 800–900 feet thick and rest disconformably on the Shingle Pass Tuff in most exposures. They are as much as 2,000 feet thick in the southern Reveille Range, where three or more densely welded cooling units occur that weather to massive reddish-gray and brown cliffs and steep slopes. (The Reveille Range was mapped by Cornwall (1967), and it was examined only briefly by the authors.) As far as is known, the base of the tuff of White Blotch Spring is not exposed in the southern Reveille Range. Reconnaissance mapping by Ekren in 1966 in the northern part of the range where older tuffs are exposed failed to disclose the Shingle Pass Tuff. The possibility should not be overlooked, therefore, that the tuff of White Blotch Spring in the Reveille Range predates the Shingle Pass.

Exclusive of the Reveille Range, the strata consist of two cooling units, each about 400 feet thick, that are separated locally by bedded tuff. Where bedded tuff is absent, the break between cooling units is extremely subtle, as in the exposures at White Blotch Spring in the locale of the measured section. (See p. 82.) Starting about 1 mile north of the spring, however, the break is easily recognized and can be traced northward to the Oswald mine, where the bedded tuff between cooling units is as much as 450 feet thick. In addition, a welded ash flow, 20–50 feet thick, is present in several exposures a few feet below the upper cooling unit. This rock contains 14 percent phenocrysts of plagioclase and alkali feldspar, in a ratio of about 2 : 1, and 0.5 percent biotite. Quartz is absent. Despite its occurrence between ash flows of the tuff of White Blotch Spring, this welded tuff obviously comprises a genetically unrelated cooling unit. The "alien" tuff was observed only in the eastern part of the area, and it is presumed, therefore, that its source lies east of the project area.

In Monotony Valley and the Belted Range the two cooling units of tuff of White Blotch Spring form a series of cliffs and slopes (fig. 5) produced by differential erosion of zones that differ slightly in welding and devitrification. Several of the cliff-forming zones display columnar jointing. The rocks range in color from light grayish tan and pinkish tan at the base, through alternating light brown and medium brown, to light reddish brown in an upper cliff-forming zone (upper cooling unit). Weakly welded rock above the highest cliff (not visible in fig. 5) is commonly light blue gray to white. Pumice fragments are generally small and indistinct in the lower cooling unit but are large (as much as 6 inches) and conspicuous in the upper unit. Lithic fragments, which are sparse and are confined to the lower one-third of the section, are chiefly vitrophyric cobbles and boulders of Shingle Pass Tuff but include lavas of intermediate composition and well-rounded boulders of massive Paleozoic quartzite. These fragments are inferred to be mostly basal accumulations—rocks picked up by the tuff as it rode over an irregular surface.

The two cooling units are petrographically nearly identical. Both contain 30–35 percent phenocrysts, of which quartz is the most conspicuous and commonly the most abundant mineral. The quartz occurs as euhedral bipyramids as much as 5 mm in diameter. The grains show slight embayment in thin section, and very few are "worm eaten" in hand specimen. Plagioclase and biotite are generally more abundant in the lower unit than in the upper, but some modal analyses show no differences. The plagioclase is commonly resorbed, "worm eaten," and charged with glass in the lower unit; it contrasts with the euhedral clear plagioclase in the upper unit. Allanite and zircon are conspicuous accessory minerals in the lower unit and are exceedingly sparse in the upper.

CENTRAL PART OF MAPPED AREA

Rocks mapped as tuff of White Blotch Spring in the central part of the mapped area crop out only in the Kawich Range, where two cooling units are present. Both units are characterized by abundant large crystals of quartz. These rocks were called "tuff of the Kawich Range" by Rogers, Anderson, Ekren, and O'Connor (1967). The lower cooling unit is poor in fragments and pumice; it could, on the basis of phenocryst content,

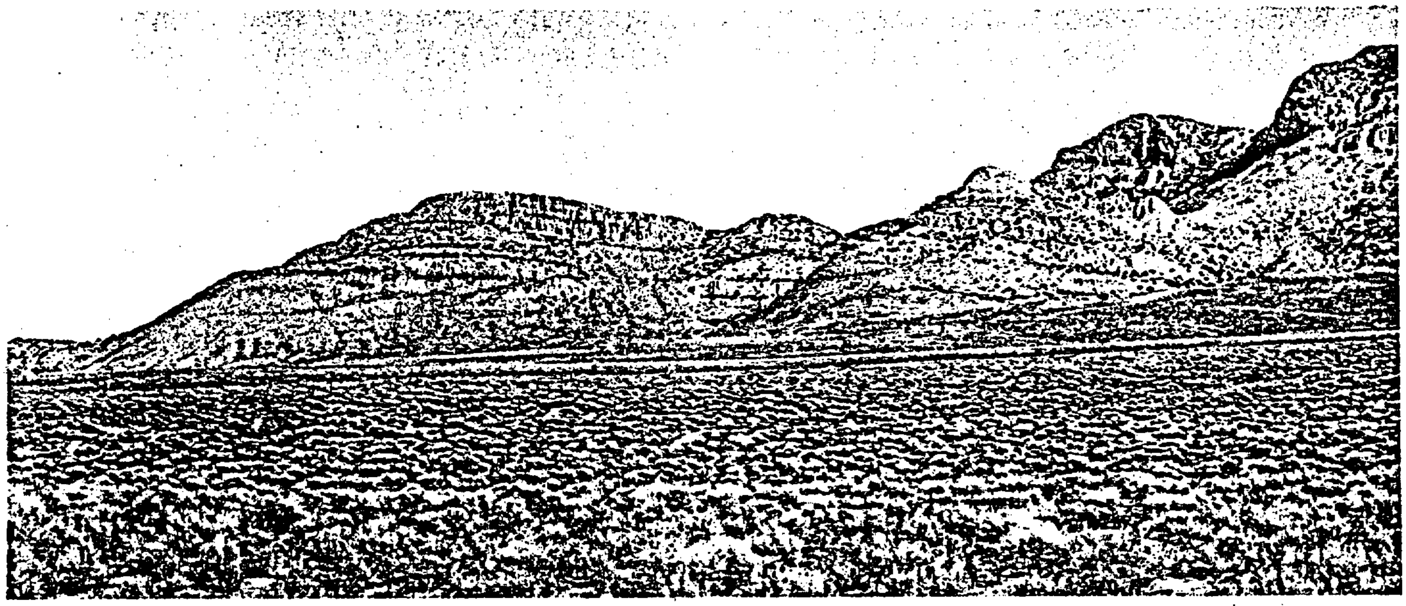

FIGURE 5.—Outcrop of tuff at White Blotch Spring, which is located just to the right of the area shown. In the high bluffs in the right background, strata include ash-fall tuff (forms slope) and quartz latite lavas. Pediment surface in right middleground is cut on Shingle Pass Tuff and also on downfaulted tuff of White Blotch Spring.

correlate with either of the cooling units exposed in the Belted Range and Monotony Valley. This unit, however, contains sparse but ubiquitous foliated quartzite lithics, fragments which were not observed in the eastern units. The foliation in the quartzite is defined by sparse flakes of biotite. This type of quartzite was observed also in all the cooling units exposed in the extreme northern part of the Kawich Range beyond the mapped area. The rocks there form the central dome of the resurged Kawich caldera and were undoubtedly extruded from that center. The unit is about 400 feet thick; it is densely welded, reddish gray, and mostly devitrified. Quartz makes up 25–50 percent of the total phenocrysts. Alkali feldspar and plagioclase occur in proportions that range from about 2:1 to 1:2; biotite is the sole identifiable mafic mineral, but pseudomorphs occur that appear to be after hornblende.

The upper unit, also about 400 feet thick, is rich in lithic fragments and pumice throughout its exposure. The lithic assemblage, in order of decreasing abundance, consists of red-brown Monotony Tuff, dark-gray argillite, dacite lava, quartzite, and sparse gneissic granite. This assemblage indicates that the unit does not correlate with either of the two cooling units exposed in the eastern part of the mapped area. The phenocryst assemblage, in contrast, matches very well. The rock contains 35 percent phenocrysts of which quartz is 30 percent, alkali feldspar is 37 percent, plagioclase is 27 percent, and biotite is 3–4 percent.

WESTERN PART OF MAPPED AREA

Western facies rocks of the tuff of White Blotch Spring have been mapped only in the northern Cactus Range, where they rest with steep angular unconformity on the upper tuffs of Antelope Springs. No continuous section is exposed in the Cactus Range, but the estimated composite thickness exceeds 3,000 feet. The rocks are intensely faulted and, except locally, are moderately to intensely hydrothermally altered. Cooling breaks possibly are present, although none were observed. The tuff of White Blotch Spring has not been identified in the Goldfield area to the west, the Tonopah area to the northwest, or the Monitor Hills to the north of the Cactus Range. Rocks younger and older than the tuff of White Blotch Spring are exposed in all those areas. Much of the rock mapped as tuff of White Blotch Spring and tuffs of Antelope Springs undivided in the area east of Mellan probably correlates with the tuff of White Blotch Spring in the Cactus Range. These rocks probably are widely distributed in the shallow subsurface south and west of Mellan. The tuff of White Blotch Spring is much more heterolithic in the Cactus Range than in the eastern area and somewhat more heterolithic in the Cactus Range than in the Kawich Range.

The tuff is characterized by a pale-orange-brown 100- to 200-foot-thick basal zone, which is rich in lithic fragments and which grades upward to drab brown-gray rock rich in yellowish-gray pumice lapilli and blocks (fig. 6) as much as 1 foot in diameter. The pumice-rich

FIGURE 6.—Specimen of pumice-rich tuff taken a few feet above a basal lithic-rich zone in the tuff of White Blotch Spring, Cactus Range. The nonflattened inclusions are fragments of cognate tuff and not pumice blocks. Quartz forms the most conspicuous crystals, but feldspar grains are more abundant and are of equal size. Note that the largest phenocrysts are in the pumice and cognate inclusions.

zone averages about 700 feet in thickness. It is overlain by massive medium-gray densely welded tuff which is virtually free of lithic fragments and pumice lapilli. This tuff is well jointed and locally columnar jointed; it forms steep rugged slopes partially covered with dark-purplish-gray or dark-red-gray blocky talus and scree. The uppermost unit is light red brown to salmon, slabby to hackly weathering, and massive and densely welded. It is indistinguishable from the rocks in the Kawich Range that are poor in lithic fragments. The compaction foliation is extremely difficult to detect in both this and the underlying zone.

Lithic fragments locally make up 50 percent of the lower part of the unit in the Cactus Range and are commonly several feet in diameter. The most abundant fragments, in order of decreasing abundance, are Monotony Tuff, tuffs of the Antelope Springs area, and dacite. Pink Mesozoic(?) granite and sedimentary rocks of probable Paleozoic age are locally common. The fragments are also moderately abundant in outcrops near Mellan. The tuff is similar in general appearance and lithology to the upper tuffs of Antelope Springs exposed in the northern Cactus Range. Both units are characterized by abundant grains of quartz and locally by abundant large pumice lapilli and blocks; both contain similar assemblages of lithic fragments. The tuffs of Antelope Springs are richer in biotite. The two units can generally be distinguished by differences in color; exposures of the tuff of White Blotch Spring tend to be dominantly pale orange, brown, or buff, and exposures of tuffs of Antelope Springs are commonly lavender or purple and locally mottled in purplish gray and yellowish brown.

CHEMISTRY

Chemical analyses of the tuff of White Blotch Spring from White Blotch Spring and the Kawich and Cactus Ranges are shown in table 6. The abnormally high normative orthoclase and quartz and the low normative anorthite of sample CS-228 reflect not only the partial replacement of plagioclase by alkali feldspar but also the weak silicification of this rock. Weak silicification is also apparent in sample R-16D. The average normative albite, anorthite, and orthoclase for the other four samples plots within the rhyolite field (fig. 4, point WBS).

An early Miocene age is indicated for the tuff of White Blotch Spring on the basis of four potassium-argon dates from rocks exposed in the eastern, central, and western areas (table 5).

ROCKS BETWEEN THE TUFF OF WHITE BLOTCH SPRING AND FRACTION TUFF

The rocks between the tuff of White Blotch Spring and Fraction Tuff consist of tuffs and lavas without distinctive marker zones. The strata are interbedded and interfingered, and, as a consequence, beds that are lowest in one locality occur in the middle or near the top of the section in another. The chief units are (1) bedded tuffs, commonly zeolitized, that underlie and are interbedded with lavas of intermediate composition, (2) lavas of intermediate composition, (3) older rocks of Mount Helen, (4) the tuff of Wilsons Camp, which is probably interbedded with the lavas, (5) intrusive rocks of intermediate and rhyolitic compositions which occur in the raised central core of the Cactus Range and to a lesser extent in the Kawich and Reveille Ranges, and (6) rhyolite and interbedded ash-fall tuff and sedimentary rocks that overlie the lavas (chiefly at White Ridge in the Kawich Range).

ZEOLITIZED BEDDED TUFF AND SEDIMENTARY ROCKS

In most of the ranges the tuff of White Blotch Spring is overlain by bedded ash-fall tuff and sedimentary rocks that form a conspicuous slope between the relatively resistant welded tuff below and the hard lavas of intermediate composition above. Most of the bedded tuffs are zeolitized, but some are vitric; in the eastern part of the mapped area the vitric tuffs predominate.

In the Belted Range the strata consist of 60–100 feet of dazzling-white well-stratified vitric ash-fall tuff that contains minor sandstone and conglomerate at the base. The tuff consists of fine shards and small pumice lapilli and contains phenocrysts of plagioclase, biotite, and minor quartz, an assemblage that persists throughout

TABLE 6.—*Chemical analyses and norms of tuff of White Blotch Spring*

[Analyses by P. L. D. Elmore, S. W. Botts, G. W. Chloe, Lowell Artis, and H. Smith, by rapid method (Shapiro and Brannock, 1962). Norms determined from chemical analyses recalculated to 100 percent minus H_2O+CO_2]

Sample	1	2	3	4	5	6
Laboratory No.	162061	162062	162063	160807	162064	162065
Field No.	CS-228	BE-10-15-3	BE-10-15-7	R-16D	WBS-4	WBS-5
Chemical analyses						
SiO_2	74.4	75.9	74.7	75.5	73.8	72.1
Al_2O_3	13.6	13.0	12.8	13.0	13.5	14.3
Fe_2O_3	1.3	1.1	1.4	.94	1.0	1.5
FeO	.40	.16	.12	.24	.24	.52
MgO	.46	.43	.05	.27	.48	.70
CaO	.22	.59	1.1	.24	.86	1.6
Na_2O	1.4	3.0	3.4	2.8	2.9	2.4
K_2O	5.7	4.5	4.7	4.7	4.9	4.3
H_2O-	.30	.41	.70	.42	1.0	1.0
H_2O+	2.0	.85	.75	1.2	1.1	1.0
TiO_2	.28	.20	.19	.17	.17	.30
P_2O_5	.09	.05	.04	.07	.09	.07
MnO	.03	.04	.04	.02	.03	.04
CO_2	<.05	<.05	<.05	<.05	<.05	<.05
Sum	100	100	100	100	100	100
Norms						
Q	44.5	39.9	35.2	41.4	36.6	38.3
or	34.4	26.9	28.2	28.6	29.6	26.0
ab	12.1	25.6	29.2	24.2	25.0	20.8
an	.5	2.6	5.3	.8	3.8	7.6
C	5.1	2.3	.2	3.1	2.1	3.0
en	1.2	1.1	.1	.7	1.2	1.8
mt	.6	.1	--------	.4	.4	1.0
hm	.9	1.1	1.4	.7	.8	.9
il	.5	.4	.3	.3	.3	.6
ap	.2	.1	.1	.2	.2	.2

Sample, locality and description

[Numbers following mineral names are percent of total phenocrysts]

1. About 1.5 miles west-southwest of Cactus Spring in Cactus Range. Rock is rich in white pumice lapilli; 30 percent phenocysts: quartz 36, iron oxide=1, altered biotite 2.8, altered sanidine and plagioclase 58, epidote 2. Biotite is chloritized, plagioclase is partly replaced by albite, epidote, and clay.
2. Summer Spring on east flank of Kawich Range. 30 percent phenocrysts: quartz 52, plagioclase 13.5, sanidine 32, biotite 3.
3. About 1 mile northwest of Summer Spring, east flank of Kawich Range. Rock is richer in pumice than sample 2; 40 percent phenocrysts: quartz 37, sanidine 39, plagioclase 23, biotite 2.
4. Kawich Range just north of Cedar Pass. Rock has been slightly silicified; about the same mode as sample 2.
5. White Blotch Spring; upper part of slope-forming zone above lowest cliff. 35 percent phenocrysts: quartz 42, sanidine 36, plagioclase 21, biotite 1.7.
6. White Blotch Spring; base of highest cliff-forming zone. 33 percent phenocrysts: quartz 40, sanidine 20, plagioclase 31, biotite 8. This rock is fairly rich in devitrified pumice lapilli which contain more plagioclase than the matrix; the lapilli were counted in the mode.

the project area. The conglomerate contains pebbles, cobbles, and boulders of probable Cambrian or Precambrian age and densely welded tuff derived from the underlying White Blotch Spring.

In the vicinity of Gray Top Mountain and Lava Ridge in the northern Belted Range, 50–200 feet of tuff and sedimentary rocks crops out above a black andesitic basalt in the approximate middle of the lava pile. At Gray Top Mountain these rocks consist of gray well-bedded tuffaceous sandstone, gravel, conglomerate, and, at the base, minor massive-weathering tuff that is rich in lithic fragments, principally of intermediate lavas. On Lava Ridge west of Gray Top Mountain, the rocks also consist of massive lithic tuff at the base, but there the overlying sandstone and conglomerate is dominantly yellow brown, red brown, and dark gray. At the top, a red conglomeratic sandstone, rich in iron oxide, contains large boulders of rhyodacite lava. All the conglomerate beds are cross-stratified and most contain well-rounded pebbles and cobbles. This sedimentary sequence is not shown separately on plate 1.

In the Kawich Range, bedded tuffs beneath the lavas rest on an eroded surface of considerable relief. Near Gold Reed, for example, 200–500 feet of vitric ash-fall tuff rich in small pumice lapilli overlies altered tuffs mapped as Shingle Pass Tuff and tuffs of Antelope Springs undivided. South of Quartzite Mountain the bedded tuffs rest directly on Precambrian strata. The tuff of White Blotch Spring is absent in both areas. North of Gold Reed and south of Cedar Pass, in Tps. 2 and 3 S., R. 51 E., the bedded tuffs are very thin or absent, and in most of this area lavas of intermediate composition rest directly on the tuff of White Blotch Spring or on older sedimentary rocks of the Cedar Wells area (older than tuffs of Antelope Springs).

In the Lizard Hills, Wilsons Camp, and Mellan areas, the bedded tuff is zeolitized in nearly every exposure; it is fresh only in sporadic outcrops. In some outcrops the zeolitized tuff is massive and contains no visible bedding, and thus it is not easily distinguished from zeolitized tuff of Wilsons Camp (a weakly welded ash-flow tuff). In other outcrops the tuff is thin bedded and locally cross-stratified. It is composed chiefly of small pumice lapilli and fine shards, but locally it contains beds rich in coarse pumice fragments as much as 3 inches long. Where fresh, the tuff is dominantly white or light gray; where zeolitized, it is mostly yellow brown.

Near Mount Helen the bedded rocks are more diverse than in the Lizard Hills and Mellan areas. They are as much as 100 feet thick and contain coarse conglomeratic sandstone, thin-bedded siltstone, and ash-fall tuff rich in pumice lapilli. These strata lie beneath the oldest lavas and above a densely welded ash-flow tuff tentatively correlated with tuffs of Antelope Springs. In addition, bedded tuff, siltstone, and poorly stratified tuff crop out locally on the west and south sides, possibly between the oldest lava (andesitic basalt) and overlying lavas of quartz latite. Some of the poorly stratified rock on the west side, which includes ash-flow tuff, is rich in metamorphic lithic fragments that are identical with the gneiss and schist exposed east of Mount Helen in the Trappman Hills. The fragments possibly were picked up as the tuff rode over the gneiss and schist of Trappman Hills, but it seems more likely that they were derived from the crust at the Mount Helen volcano during extrusion, because the Trappman Hills rocks were probably completely covered by volcanic strata at that time.

LAVAS AND INTRUSIVE ROCKS OF INTERMEDIATE COMPOSITION

The rocks treated herein are exclusive of the rocks of Mount Helen and the intrusive rocks in the central cores of the Cactus and Kawich Ranges.

Lavas of intermediate composition are exposed in all the ranges within the mapped area. Beyond the mapped area they are widely distributed to the north, west, and east. The rocks were extruded from many widely dispersed vents, but their stratigraphic position indicates that they are of approximately the same age in all localities. They apparently were extruded principally from fissure-type feeders, several of which are exposed. The intrusive and extrusive rocks are petrographically and chemically very similar.

The intermediate lavas are light gray to black and commonly weather to somber shades of brown; in most outcrops they are flow layered and folded. They are generally more resistant to erosion than adjacent strata and tend to form steep ridges and rugged hills. Where the lavas have been weakly or moderately hydrothermally altered, they tend to form distinctive lavender or purplish-gray outcrops and soils. Where the lavas are intensely altered, as at Gold Reed, they are commonly bleached to pale gray, yellow, or pink. The rocks are described by area in a general progression from east to west.

Belted Range and Reveille Valley

Intermediate lavas nearly 1,000 feet thick form the bulk of the outcrops in the northern Belted Range. They crop out as far south in the range as Wheelbarrow Peak and also form a broad area of hills in Reveille Valley. The rocks are well exposed and show considerable variation in composition and texture. All are porphyritic and most contain about 30 percent phenocrysts. The groundmasses range from glassy to completely crystalline and consist predominantly of plagioclase and alkali feldspar microlites, iron oxides, and fine biotite. The principal phenocryst in all the rocks is plagioclase, whose sizes and zoning are varied and whose average composition is andesine-labradorite. Biotite is the chief mafic mineral, but hornblende occurs in nearly all the rocks, and augite and hypersthene occur in some. In general, the least silicic rocks and those richest in mafic minerals occur at or near the base of the lava pile; however, in most exposures within the Belted Range thin andesitic basalt lavas crop out near the middle of the sequence. These rocks are black where glassy and dark greenish gray where stony. They contain about 20 percent small phenocrysts of labradorite, augite and, hypersthene in a pilotaxitic groundmass of andesine microlites and iron oxide. The basalt is a major rock unit southeast of the Oswald mine and east of the mapped area.

Chemical analyses of two samples from the Belted Range are shown in table 7 (samples 1 and 2). Sample 1 is from the top of the pile at Lava Ridge, and sample 2 is from the approximate base; both contain abundant normative quartz, but only sample 1 contains quartz phenocrysts. Both samples, as shown on the plot of normative albite, anorthite, and orthoclase (fig. 4), fall in the rhyodacite and quartz latite fields.

Kawich Range

Intermediate lavas form two outcrop belts in the Kawich Range. The largest and best exposed belt strikes north-northwest and extends from the east side of Saucer Mesa to the west flank of the Kawich Range at Trailer Pass. The other belt lies entirely along the east flank of the range and extends roughly from the south boundary of T. 3 S., R. 51 E., to Cedar Pass in T. 2 S. The rocks in both belts are the same age, but they have been displaced from each other by several northwest-trending faults that are inferred to bound the Cathedral Ridge caldera.

East of Saucer Mesa the lavas are as much as 500 feet thick and crop out in a series of isolated knobs and hills. The oldest flows are rhyodacites. A modal analysis of one of these is given in table 7 (sample 3). The plagioclase phenocrysts are as much as 10 mm long and commonly are aggregates of several grains. The hornblende is brown and has a very small extinction angle. The groundmass, which is hyalopilitic, contains microlites of andesine and tiny grains of iron ore in brown glass. The rhyodacite is overlain by andesite lava which contains about 44 percent small phenocrysts of plagioclase, clinopyroxene, hypersthene, and iron ore. Except for a basal vitrophyre, the andesite has a completely devitrified groundmass composed of a dense felt of plagioclase microlites and interstitial iron ore. Apatite is a common accessory in both the rhyodacite and andesite.

North of Quartzite Mountain, where the lava belt is shifted westward by high-angle normal faults, an estimated 2,000–3,000 feet of lava is exposed. At Gold Reed these rocks are moderately to intensely hydrothermally altered and are weakly mineralized. The rocks are principally dacites and rhyodacites and they contain 30–40 percent phenocrysts of which plagioclase is always dominant. Mafic minerals are altered partly or wholly to chlorite, calcite, and iron ore. The chief pseudomorph appears to be after biotite, but several thin sections show relicts that are after hornblende and pyroxene.

TABLE 7.—*Chemical analyses and norms of lavas of intermediate composition exclusive of Mount Helen rocks*

[Analyses by P. L. D. Elmore, S. W. Botts, G. W. Chloe, Lowell Artis, and H. Smith, by rapid method (Shapiro and Brannock, 1962). Norms determined from chemical analyses recalculated to 100 percent minus $H_2O + CO_2$]

Sample	1	2	3	4	5	6	7	8	9	10	11	12	13
Field No.	BP-57	BP-58	QM-88	QM-70	QM-75	QM-94	R-14C	R-16A	M-13	M-16	CS-326	CS-253	CS-392
Laboratory No.	160725	160726	160982	160979	160981	160983	160804	160805	160963	160964	162860	162058	162859
Chemical analyses													
SiO_2	67.7	60.5	63.0	60.9	59.6	59.3	53.7	59.8	63.3	57.2	62.5	59.5	62.6
Al_2O_3	15.2	16.6	16.0	17.1	16.2	16.0	15.4	16.9	16.8	17.6	15.0	16.1	14.3
Fe_2O_3	1.9	4.2	4.6	3.4	4.3	5.0	2.5	4.5	2.9	3.1	4.9	6.1	3.8
FeO	.97	1.9	.76	2.0	1.2	.56	5.1	1.2	1.4	3.7	.60	.40	1.1
MgO	1.0	2.0	2.1	3.3	3.1	.90	7.9	2.4	1.7	3.6	1.7	2.5	2.0
CaO	3.1	5.2	4.6	4.3	5.6	6.7	8.3	6.4	4.4	6.4	4.2	4.3	4.4
Na_2O	3.0	3.1	3.2	3.1	3.1	3.0	2.6	3.1	3.2	3.2	3.2	3.1	2.7
K_2O	3.8	3.0	3.3	1.0	1.4	3.1	2.0	3.3	3.6	2.8	3.2	4.1	3.4
H_2O-	.32	1.1	.70	.47	1.6	.77	.47	.64	.75	.31	.96	.84	.90
H_2O+	2.5	.94	1.0	2.9	2.8	1.3	.90	.87	.97	.84	1.7	1.4	2.8
TiO_2	.30	1.1	.77	.78	.79	.74	1.0	.86	.54	.96	.70	.81	.96
P_2O_5	.18	.37	.22	.32	.27	.32	.40	.36	.34	.40	.28	.38	.32
MnO	.04	.10	.07	.12	.05	.10	.12	.06	.08	.14	.07	.15	.07
CO_2	<.05	<.05	<.05	.07	.06	2.3	<.05	<.05	<.05	<.05	<.05	<.05	<.05
Total	100	100	100	100	100	100	100	100	100	100	99	100	99
Norms													
Q	28.9	18.5	20.1	27.5	21.3	18.2	2.9	14.6	20.3	9.5	22.1	14.5	24.1
C	1.0			4.2	.1				.5				
or	23.1	18.1	19.8	6.1	8.7	19.1	11.9	20.0	21.7	16.7	19.6	25.0	21.0
ab	26.1	26.7	27.4	26.4	27.4	26.5	22.2	26.5	27.5	27.3	28.1	27.0	24.0
an	14.6	23.0	19.8	20.0	27.2	22.0	24.7	22.7	20.0	26.0	18.0	18.4	18.0
wo		.4	.8			4.4	6.0	3.0		1.6	.8	.1	1.3
en	2.6	5.1	5.3	8.5	8.1	2.3	20.0	6.0	4.3	9.0	4.4	6.4	5.2
fs							5.9			2.9			
mt	2.5	3.3	.5	4.8	1.8		3.7	1.6	3.3	4.5	.1		1.1
hm	.3	2.0	4.4	.3	3.2	5.2		3.5	.7		5.0	6.3	3.2
il	.6	2.1	1.5	1.5	1.6	1.5	1.9	1.7	1.0	1.8	1.4	1.2	1.9
tn												.5	
ap	.4	.9	.5	.7	.67	.8	1.0	.9	.8	1.0	1.0	.9	.8

Sample, locality and description

[Numbers following mineral names are percent of total phenocrysts]

1. Highest flow in northern part of Lava Ridge, Belted Range. Medium-gray vitrophyric rhyodacite; 28 percent phenocrysts: calcic andesine 62, biotite 17, hornblende 17, quartz 3, and minor sphene, magnetite and apatite.
2. Northern part of Lava Ridge, Belted Range. Dark-gray to black, dense rhyodacite; 17 percent phenocrysts: plagioclase 53, clinopyroxene and orthopyroxene 16, and magnetite 14; groundmass is trachytic to felty.
3. About 2.2 miles south-southeast of Gold Reed, Kawich Range. Reddish-brown rhyodacite; about 30 percent phenocrysts: plagioclase (as much as 10 mm) 70, brown hornblende 20, and minor clinopyroxene, biotite, magnetite and quartz; groundmass is hyalopilitic.
4. About 1.5 miles south of Trailer Pass near western limit of exposed dacite lavas, Kawich Range. Greenish-gray dacite; about 50 percent phenocrysts consisting predominantly of zoned plagioclase (sodic labradorite); mafic minerals include partly altered clinopyroxene and biotite and completely altered and unidentifiable grains that may have been hornblende. Note high corundum in norm; possibly indicates loss of alkalis through alteration. Also, plot of partial molecular norm (fig. 4) indicates possible loss of potassium.
5. About 3.5 miles northwest of Gold Reed, Kawich Range; just below base of Fraction Tuff. Reddish-brown dacite with phenocryst assemblage similar to sample 3 but with clinopyroxene as principal mafic phenocryst.
6. About 1.5 miles southwest of Gold Reed on north flank of Quartzite Mountain, Kawich Range. Red weakly altered rhyocacite; about 20 percent phenocrysts of plagioclase, altered mafic minerals, magnetite, and quartz in a pilotaxitic groundmass; calcite, hematite, and quartz are common secondary minerals.
7. About 4.5 miles south-southeast of Mellan. Black andesite; 30 percent phenocrysts: plagioclase (an_{52}) 39, diopsidic augite 38, magnesian olivine 20, and minor magnetite and hornblende, in a pilotaxitic groundmass of plagioclase, pyroxene, and iron oxide.
8. About 2.5 miles southeast of Mellan. Dark-gray to black rhyodacite; 32 percent phenocrysts: plagioclase (calcic andesine) 70, augite 20, biotite 5, and magnetite 5, in a dense trachytic groundmass that contains sparse small plagioclase laths.
9. Eastern part of Lizard Hills. Black rhyodacite dike; 33 percent phenocrysts: plagioclase (an_{36}) grains (as much as 5 mm) 72, biotite 14, green hornblende 7, and minor clinopyroxene and magnetite in a cryptocrystalline felty groundmass.
10. Northeastern part of Lizard Hills. Black rhyodacite; 35 percent phenocrysts: plagioclase 64, augite 15, hypersthene 9, magnetite 10, brown hornblende 2, and a trace of quartz in a cryptocrystalline felty groundmass.
11. About 2.5 miles east of Cactus Spring. Reddish-gray rhyodacite; similar to sample 13 but with sparse quartz phenocrysts and slightly more augite.
12. About 5 miles southwest of Cactus Peak. Olive-green rhyodacite; 20 percent phenocrysts: plagioclase (calcic andesine) 60, brown hornblende 22, biotite 10, and quartz 4, augite 2, and magnetite 2, in a hyalopilitic groundmass composed predominantly of plagioclase and magnetite.
13. East part of Goldfield Hills west of project area. Gray rhyodacite; 23 percent phenocrysts: andesine 75, biotite 11.5, augite 8.5, hornblende 2.3, and magnetite 2 in a hyalopilitic groundmass.

The plagioclase is extensively replaced by more sodic plagioclase and locally by calcite. Quartz phenocrysts are visible in nearly all the rocks, but they are subordinate to the mafic phenocrysts and plagioclase.

Northeast of Trailer Pass, in the second belt of lavas outlined above, most of the rocks are nearly identical with those already described; however, several flows in this area contain phenocrysts of alkali feldspar, as much as 2 cm in diameter, and moderate amounts of quartz. These flows have not been analyzed, but they are petrographically very similar to analyzed rocks that are quartz latites in composition. In this area the lavas locally rest unconformably on sedimentary rocks that predate the tuffs of Antelope Springs; this fact suggests extensive prelava erosion. North of Cedar Pass, on the east flank of the range, lavas of intermediate composition occur in isolated exposures. These rocks are similar megascopically to the lavas to the south and are presumed to be equivalent stratigraphically. They form part of the downfaulted margins adjacent to the raised central mass of the range.

Analyses of four rocks from the Quartzite Mountain (southern Kawich Range) area are shown in table 7 (samples 3, 4, 5, 6).

Mellan Hills Area

For simplicity the term "Mellan Hills area" is used for the area of ridges and hills lying between the Cactus and Kawich Ranges. It comprises the hills near the old townsite of Mellan, the Lizard Hills, Gabbard Hills, and Triangle Mountain.

The intermediate rocks exposed in the Mellan Hills area display a wide variety of colors and textures and range in composition from andesite to quartz latite. Most are lavas that form a group of fault-controlled northwest-trending ridges, and some are dikes that trend mostly northeast. In the area between Mellan and the old road between Antelope Springs and Trailer Pass, the lavas overlie the tuff of Wilsons Camp, which in turn rests on the tuff of White Blotch Spring or older rocks. The lavas, which are overlain by the Fraction Tuff or younger rhyolite lavas, are thin, rarely exceeding 50 feet. The thinness is probably due principally to shortening of the section by low-angle faults, but may be partly due to deep pre-Fraction Tuff and prerhyolite erosion. The rock still preserved in outcrop is dark-gray rhyodacite (sample 8, table 7). The rocks exposed between the Antelope Springs-Trailer Pass road and the Antelope Springs-Gold Reed road are mostly black and glassy with compositions ranging from andesite to quartz latite. The black latites and andesites are not easily distinguished from basalt. They contain about 30 percent small phenocrysts consisting of about equal amounts of plagioclase and pyroxene and minor hornblende, biotite, fayalite, and locally a trace of quartz. Magnetite is abundant as small phenocrysts and as tiny grains in a felty to hyalopilitic groundmass. The phenocrysts of hornblende and biotite consistently display thick reaction rims of magnetite. The andesites and latites are intruded by and locally overlain by rhyodacite and quartz latite which form the bulk of the outcrops southward to Triangle Mountain.

At Triangle Mountain three rock types occur in a steep west-dipping sequence. The easternmost and oldest rock is an andesite or dacite. The rock is black, glassy, and flow brecciated at the base and top and is brown and gray in the thin flow-layered stony interior. Bedded tuff at the base of the lava is fused for a distance of 2–5 feet from the contact, and the beds dip vertically or 70° W. The rock contains nearly 50 percent small phenocrysts of plagioclase, hornblende, biotite, augite, and hypersthene in a pilotaxitic to hyalopilitic groundmass composed of plagioclase microlites, specks of iron ore, and glass. The middle flow in the lava pile is flow layered light- and brownish-gray mostly devitrified dacite or latite that contains 30 percent phenocrysts of plagioclase, biotite, and hornblende in a dense groundmass of plagioclase microlites, "cryptofelsite," and glass. On the west flank, this rock is overlain by quartz latite that contains 30 percent phenocrysts of plagioclase, quartz, hornblende, biotite, augite, and fayalite. The plagioclase grains are commonly as much as 10 mm long, and the rock, therefore, is conspicuously porphyritic. This rock is in fault contact with the tuff of Wilsons Camp which is locally silicified for a distance of about 2 feet from the fault. Locally, the tuff of Wilsons Camp has been forced into steep dips along the fault zone, and older bedded tuffs are in contact with the quartz latite. On the southeast flank of the mountain the lava flows change strike from north-northwest to east and here the dip in the flow layering decreases from vertical and 70° W. to about 15°–25° S. In this area the conspicuously porphyritic quartz latite rests on latite or dacite, which in turn rests on zeolitized tuff. The andesite, which is the oldest flow in the exposures to the northwest, is not present.

The quartz latite at Triangle Mountain contains abundant inclusions of hornblende-rich rock with a peculiar pseudo-ophitic texture. Crystals in the inclusions consist of randomly oriented tabular plagioclase, hornblende, and minor clinopyroxene, and interstices are filled with brown glass. Hornblende is mostly in the form of very slender prisms or "needles" which are intergrown with the plagioclase; individual crystals commonly penetrate as many as five grains of plagioclase. The hornblende is pleochroic from olive to deep brown and has a maximum extinction angle of about 10°. The plagioclase is labradorite or calcic andesine and is weakly zoned. The clinopyroxene is commonly altered along crystal edges to hornblende. The inclusions, which range in size from a few millimeters to several feet, probably constitute early magmatic segregates that were brought to the surface in the form of semisolid masses of crystal mush.

Wilsons Camp

Intermediate lavas form several rounded hills east and south of Wilsons Camp. The oldest lava at Wilsons Camp is a dark-gray to dark-brownish-gray quartz latite. It contains about 30 percent phenocrysts consisting of glomeroporphyritic plagioclase (clots as large as 10 mm), minor quartz (as large as 3 mm), sparse alkali feldspar (as large as 1½ cm), and abundant mafic minerals consisting of about equal amounts of clinopyroxene and hornblende and minor altered biotite. The hornblende and biotite are largely altered to iron oxide, chlorite, and calcite; the clinopyroxene, in contrast, is fresh and clear. The groundmass is pilotaxitic with tiny plagioclase and alkali feldspar microlites, abundant dust and specks of iron oxide, and interstitial "cryptofelsite" and glass.

The youngest lava is a dacite or rhyodacite that is finer grained than the older rock and contains about 20 percent phenocrysts comprising, in order of decreasing abundance, plagioclase (as much as 5 mm), clinopyroxene, hornblende, hypersthene, and biotite (all less than 5 mm). The hornblende and biotite are largely altered to iron oxide. The groundmass consists of a dense felt of tiny plagioclase and alkali feldspar microlites. Apatite, in prisms nearly 1 mm long, is a conspicuous accessory mineral.

Cactus Range

In the Cactus Range, lavas of intermediate composition occur only on the flanks of the range, where they form isolated outcrops surrounded by alluvium. The basal contacts have not been observed.

In the northwestern part of the range, the lavas are about 600 feet thick. They are unconformably overlain by sedimentary rocks of pre-Fraction Tuff age. Both the lavas and sedimentary rocks are intruded by broad rhyolite dikes that were emplaced in faults along which the strata were tilted westward to moderate and steep angles. The strata are repeated several times by these faults and form discontinuous bands parallel to the range. The lavas are mostly reddish brown to dark gray and tend to weather to subdued dark ridges that are locally paralleled by ridges of contrasting lighter colored rhyolite. The lower part of the unit is commonly olive green as a result of the partial alteration of the groundmass to celadonite.

On the eastern flank of the range, east of Cactus Spring, rhyodacite lavas are reddish brown to reddish gray. These colors contrast sharply with the dark-greenish-gray to greenish-brown colors characteristic of the propylitized intrusive rhyodacites that lie southwest of these exposures. A small isolated outcrop of reddish-gray rhyodacite similar to the lavas near Cactus Spring is located in Civet Cat Canyon southwest of the Cactus Range, and small exposures of altered rhyodacite lava occur also in the extreme southern part of the Cactus Range near the Wellington Hills.

In the northern Cactus Range the reddish-brown lavas contain 20–30 percent phenocrysts consisting of 60–75 percent conspicuous plagioclase (calcic andesine) and varying proportions of biotite, hornblende, clinopyroxene, and magnetite. Clinopyroxene is sparse or absent in some rocks, and occasional large phenocrysts of quartz and alkali feldspar (as large as 1.5 cm) occur in some flows. Much of the reddish-brown color of these rocks is caused by pseudomorphs of hematite after biotite and hornblende. Clinopyroxene was generally unaffected by the alteration that gave rise to hematite in the biotite and hornblende. Some plagioclase grains show deep magmatic corrosion but are otherwise unaltered. In most rocks the groundmass is hyalopilitic to pilotaxitic and is composed predominantly of plagioclase, magnetite, biotite, and glass. Phenocrysts range in size from about 1 to 10 mm.

Chemical analyses of lavas from the east and west flanks of the Cactus Range are given in table 7 (samples 11 and 12, respectively). In thin section both rocks show some evidence of deuteric alteration, and sample 12 contains appreciable celadonite in the groundmass. For comparison, an analysis of unaltered lava from the east flank of the Goldfield Hills is also given (sample 13). All analyses plot in the rhyodacite field of figure 4.

Lavas of intermediate composition, similar to those in the Cactus Range, make up most of the Tertiary exposures in the Goldfield Hills, a volcanic center adjoining the west margin of the mapped area. These lavas rest on and are overlain by welded tuffs and tuffaceous sedimentary rocks that can be correlated unequivocally with rocks in similar stratigraphic positions in the Cactus Range. The lavas in the Goldfield Hills, therefore, are interpreted as belonging to the same period of eruptive activity as the lavas in the Cactus Range. A potassium-argon age of 21.1 m.y. (sample 19, table 5) was obtained from biotite from a quartz latitic welded tuff that overlies the lavas on the south flank of the Goldfield Hills (H. R. Cornwall, written commun., 1964). Thus an early Miocene or older age is indicated for the underlying lavas in the western part of the mapped area.

OLDER ROCKS OF MOUNT HELEN

Mount Helen was a source area for lavas of intermediate to basic composition. Intrusive rocks and lavas derived from this source make up most of the mountain and form many of the adjacent hills. These rocks form a mappable group that is related spatially and chemically. The lavas rest locally on thin-bedded sandstone, siltstone, and tuff, which in turn rest on welded tuff that probably was extruded from the Mount Helen area. The lavas are overlain by sedimentary rocks and the tuff of

Tolicha Peak. Three distinctly different rocks occur; from oldest to youngest they are andesitic basalt, rhyodacite, and quartz latite. The rhyodacite is described herein but is not shown separately on the geologic map.

In addition to these rocks, much younger basalt was extruded from the central feeder at Mount Helen. This basalt is described on page 65.

Andesitic Basalt

The andesitic basalt underlies Mount Helen and crops out sporadically around the perimeter of the mountain. It also forms several broad outcrops as far as 3 miles west of Mount Helen. The rock is largely a lava, but in places near the mountain it occurs as intrusive dikes and irregular apophyses. A small dike can be seen in the saddle south of the mountain on the old road to Wellington.

The andesitic basalt is dark gray to black where fresh and unaltered; it is brown and massive weathering without visible flow layering. The presence of highly vesicular zones near the top and of small pebbles of basalt in the overlying quartz latite lavas indicates that the rock is a lava.

The rock is porphyritic, containing conspicuous dark-green to black prisms of augite as much as 5 mm in length and 3 mm in diameter. The dense groundmass is composed of randomly oriented microlites of calcic andesine, tabular crystals of labradorite as much as 1.5 mm in length, a few small crystals of olivine, abundant magnetite, and very sparse interstitial alkali feldspar. A few thin sections contain sparse corroded microphenocrysts of biotite or hornblende as much as half a millimeter in diameter that are rimmed with iron oxide. Commonly the rock is altered and the augite has been replaced by iron oxide and chlorite, and the larger crystals of plagioclase have been replaced by calcite and yellowish-green clay or chlorite. The altered augite phenocrysts weather to cavities that closely resemble vesicles in outcrop.

A chemical analysis of the andesitic basalt is given in table 8 (sample 1).

Rhyodacite

Rhyodacite makes up a very small part of the rock at Mount Helen and is exposed only in small patches on the east and west sides, where it may be a single lava flow. The rock is dark greenish gray and, in contrast with the andesitic basalt, is conspicuously flow layered. It is porphyritic but the phenocrysts are small and inconspicuous, rarely exceeding 3 mm in length. Plagioclase (sodic andesine) is the most abundant crystal, followed by augite, hypersthene, and magnetite. The total phenocryst content is about 15–20 percent. The groundmass is glassy to cryptocrystalline.

Table 8.—*Chemical analyses and norms of rocks from Mount Helen*

[Analyses by P. L. D. Elmore, S. D. Botts, G. W. Chloe, Lowell Artis, and H. Smith, by rapid method (Shapiro and Brannock, 1962). Norms determined from chemical analyses recalculated to 100 percent minus H_2O+CO_2]

Sample	1	2	3	4	5
Laboratory No.	164094	165058	162066	160965	162067
Field No.	E8-6-1	E11-20-3	BE-4-27-1	M-195	M-177
Chemical analyses					
SiO_2	51.1	65.3	62.8	63.5	65.2
Al_2O_3	17.4	16.0	17.1	16.7	15.5
Fe_2O_3	6.4	1.3	5.0	4.8	3.3
FeO	1.3	1.4	.16	.32	.12
MgO	3.3	1.1	.40	.33	.47
CaO	9.1	3.6	2.4	2.1	2.1
Na_2O	3.2	4.0	3.2	3.4	2.2
K_2O	2.4	2.4	6.8	6.8	9.2
H_2O-	1.5	1.3	.37	.26	.17
H_2O+	1.1	2.7	.57	.68	.45
TiO_2	1.4	.49	.74	.78	.44
P_2O_5	.71	.23	.31	.26	.23
MnO	.07	.07	.06	.02	.05
CO_2	.54	.09	.08	.11	.67
Sum	100	100	100	100	100
Norms					
Q	4.2	24.8	13.5	15.4	12.2
or	14.7	14.8	40.6	40.7	55.0
ab	28.1	35.3	27.4	26.9	18.8
an	27.0	17.1	10.0	8.3	8.2
C		.9	.9	1.2	
wo	6.3				
en		2.9	1.0	.8	1.2
fs		.9			
mt	.4	2.0			
hm	6.4		5.1	4.9	3.3
il	2.8	1.0	.5	.7	.4
ap	1.7	.6	.7	.6	.6
ru			.5	.4	
tn					.6

Sample, locality, and description

[Numbers following mineral names are percent of total phenocrysts]

1. South flank of Mount Helen. Andesitic basalt lava; 20 percent phenocrysts consisting of labradorite, augite, hornblende rimmed with iron oxide, and magnetite in a pilotaxitic groundmass.
2. East side of Mount Helen. Rhyodacite lava; 18 percent phenocrysts: plagioclase 74, clinopyroxene 13.5, hypersthene 10, magnetite 2.5, and a trace of biotite in a glassy groundmass.
3. South flank of Mount Helen. Biotite-hornblende quartz latite outcrop. Rock is blue gray; about 30 percent phenocrysts of andesine, quartz, altered biotite, and hornblende; mafics are almost completely altered to iron oxide, chlorite and clay(?). Plagioclase is slightly altered to calcite and potassium feldspar; groundmass is aphanitic and contains much iron oxide dust. Rock (in outcrop) contains scattered potassium feldspar phenocrysts as much as 2 cm long (none of these believed to be in rock analyzed).
4. South flank of Mount Helen. Biotite-hornblende quartz latite. Rock is dark purplish gray and probably part of same flow as sample 3. Same phenocryst content; a few potassium felsdpar phenocrysts visible in thin section but andesine is dominant feldspar; mafic minerals are almost completely altered to iron oxide and chlorite; plagioclase partly altered to calcite and potassium feldspar.
5. South flank of Mount Helen. Biotite quartz latite; rock is light gray; 46 percent phenocrysts: quartz 19; alkali feldspar 29; andesine 28; biotite 19; hornblende 2. Plagioclase is partly replaced by alkali feldspar and calcite; biotite and hornblende are almost completely altered to iron oxide and calcite.

A chemical analysis of the rhyodacite is shown in table 8 (sample 2).

Quartz Latite

Rocks mapped as quartz latite consist of lavas, a few thin sills that were intruded at the base of the pile adjacent to Mount Helen, and a large feeder plug at the south end of Mount Helen. Two types of quartz latite are recognized. The older is dark green gray or dark blue gray, conspicuously flow laminated and layered, and characterized by sparse potassium feldspar phenocrysts that commonly are 2 cm long, rarely as much as 5 cm. This rock contains about 30 percent phenocrysts consisting of plagioclase that range in size from about 1 mm to about 1 cm, considerably less but more conspicuous potassium feldspar, biotite, hornblende, and minor quartz. Iron oxide is fairly abundant as small phenocrysts, as fine dust and specks in the groundmass, and as a replacement product of hornblende and biotite; in many thin sections it has entirely replaced the mafic minerals. The groundmasses of all rocks examined in thin section are too dense for point counting with the microscope. All contain tiny grains of quartz and microlites of alkali feldspar. A conspicuous feature is the partial replacement of plagioclase phenocrysts by alkali feldspar. A lack of twinning and the unusually high K_2O content of the rock (see samples 3 and 4, table 8) strongly suggest that the replacement feldspar is potassium feldspar rather than albite.

The younger quartz latite is characterized by abundant phenocrysts of quartz and biotite. It forms lavas in and adjacent to Mount Helen and occupies the central part of the feeder plug. This rock is light gray to medium gray and is also conspicuously flow layered. It contains as much as 50 percent phenocrysts that include oligoclase and potassium feldspar in ratios of about 2:1, abundant quartz and biotite, and sparse hornblende. The mafic crystals are largely replaced by iron oxide in most thin sections and entirely replaced in some.

The two types of quartz latite from Mount Helen are extremely rich in K_2O (table 8); the older quartz latite (samples 3 and 4) contains 6.8 percent K_2O. Because of the possibility that single very large potassium feldspar crystals were present in these analyzed samples, an additional analysis was made of several small fragments in order to preclude the occurrence of the large potassium feldspar grains. This analysis showed 7.7 percent K_2O and 2.3 percent Na_2O (W. M. Mountjoy, written commun., 1965). The younger quartz latite (sample 5) contains 9.2 percent K_2O and 2.2 percent Na_2O. The high K_2O contents indicate that all the alkali feldspar in the groundmasses of the quartz latite is probably potassium feldspar. The fact that plagioclase phenocrysts have been largely replaced by alkali feldspar suggests that much of the groundmass feldspar is probably similarly altered. Whether this alteration was hydrothermal or deuteric is not known. It appears to be confined to rocks in the immediate vicinity of Mount Helen. Prior to alteration the rocks may have been normal calc-alkalic types similar chemically to the lavas of intermediate composition that crop out throughout the project area.

TUFF OF WILSONS CAMP

The name tuff of Wilsons Camp is here informally applied to an ash-flow tuff sheet that crops out in Tps. 3, 4, and 5 S., Rs. 47, 48, and 49 E., and locally along the southwest flank of the Cactus Range. The fresh rock is best exposed between Coyote Cuesta and Triangle Mountain in T. 5 S., R. 48 E., about 5 miles southeast of Wilsons Camp.

The tuff comprises two cooling units, each of which is 50–300 feet thick. Both units are predominantly weakly welded and vitric, and both locally have thin vitrophyric zones at their bases. The tuff is characterized by abundant nonflattened to slightly flattened pumice lapilli and by blocks of light-gray to white silky pumice and greenish-gray and brown cindery pumice. The pumice fragments are as much as 4 feet in length but average less than 4 inches. They make up as much as 50 percent of the total rock. The tuff also contains abundant fragments of porphyritic lava of intermediate composition, rhyolite, basalt, and a few variegated metamorphic rocks—mostly argillite and quartzite. The fragments range in size from about 1 inch to 2 feet and are largest and most abundant at the base of the lowest cooling unit. The tuff matrix is light gray and weathers pale brown to buff. It contains 20–33 percent phenocrysts consisting of 4–16 percent quartz, 40–70 percent plagioclase, 10–30 percent alkali feldspar, and 8–15 percent biotite. Other mafic minerals are hornblende, clinopyroxene, and hypersthene; these are most abundant in the lowest unit. Magnetite, sphene (as much as 1 mm), and allanite are common accessory minerals.

Tuff that is mapped as tuff of Wilsons Camp in the valley east of Gabbard Hills and in the valleys northeast of Triangle Mountain is intensely zeolitized. The zeolitized rock is yellow, orange, and brown, and the abundant pumice lapilli and blocks which characterize the vitric tuff of Wilsons Camp are rare and obscure. Biotite is the only identifiable mafic mineral in thin section. Map unit identification, therefore, is tentative and the zeolitized strata may include cooling units that are older than the tuff of Wilsons Camp. If this is the case, the rocks are probably at least genetically related to the tuff of Wilsons Camp as shown by the same lithic fragment assemblage and nearly the same abundance of fragments.

At the old townsite of Mellan the tuff has been intensely silicified by hydrothermal solutions. The two cooling units are separated by 20–100 feet of siltstone, ashfall tuff, and conglomeratic sandstone, all of which are intensely silicified. Both cooling units are characterized by abundant cavities that originally contained pumice lapilli. The two cooling units and the sedimentary rocks that separate them are best seen in the hanging-wall block of a northwest-trending fault that displaces the tuff of Wilsons Camp down on the northeast against tuff of White Blotch Spring on the southwest. An inclined shaft follows this fault and penetrates about 300 feet of the lower cooling unit.

The tuff of Wilsons Camp is similar in general megascopic appearance to the Fraction Tuff (p. 50), and the two can easily be confused. The Fraction also contains a great abundance of fragments of intermediate rocks and has about the same percentage of phenocrysts; it is more densely welded, contains more metamorphic lithic fragments and less pumice and plagioclase, and, apparently, contains no pyroxene. It does contain both sphene and allanite.

INTRUSIVE ROCKS OF THE CENTRAL CORE OF THE CACTUS RANGE

The intrusive masses within the central core of the Cactus Range form a group of intergrading hypabyssal calc-alkalic rocks that range in composition from melanodiorite to rhyolite. This assemblage persists throughout, but high-silica varieties dominate in the northern part of the Cactus Range, and intermediate and low-silica varieties predominate in the central and southern parts. There is good evidence that the rocks ranged from the generally low-silica varieties to coarse-grained porphyritic rhyolite as intrusion proceeded, and they appear to be genetically related. All the rocks are porphyritic but the texture of many is medium-grained granitoid; such a texture suggests that they may have been intruded at moderate depth.

The rocks form dikes, sills, plugs, laccoliths, and stocks. These bodies range in size from apophyses a few feet in diameter to a large complex laccolithic mass exposed between Cactus Spring and Antelope Springs, herein named the Roller Coaster laccolith for exposures on Roller Coaster Knob. The laccolith is 6, and possibly 10, miles long and more than 3 miles wide. Only the larger intrusive masses are shown on plate 1. Probably more than 100 additional small isolated masses occur.

Most, if not all, of the larger masses are composite, and it is within them that the relative age of the various rock types has been established. North of the Wellington Hills, for example, the sequence of intrusion is low-silica porphyritic granodiorite followed by melanodiorite, biotite lamproyphyre, a high-silica variety of porphyritic granodiorite, and finally coarse-grained porphyritic rhyolite. In the Roller Coaster pluton the oldest rocks are aphanitic low-silica rhyodacites similar in composition to the low-silica variety of porphyritic granodiorite. These are intruded in turn by high-silica porphyritic granodiorite, quartz latite porphyry, rhyolte, and coarse-grained porphyritic rhyolite. Other rock types, gradational between these, may also occur among the aphanitic and fine-grained phases which cannot be accurately identified. Small isolated dikes and apophyses crop out around the peripheries of the larger masses and are presumed to be the same age as rocks of similar composition in the composite masses.

The intrusive rocks postdate the tuff of White Blotch Spring, and they probably predate the Fraction Tuff. The intrusive masses very possibly are equivalent in age to the lavas of nearly identical composition that flank the range and occur in adjacent areas.

Most of the intrusive masses are extensively propylitized and large areas of some are highly altered; unaltered rocks are exceptional. This alteration tends to obscure intrusive contacts between petrographically similar rocks, and it renders the recognition of subtle differences in texture and mineralogy very difficult. In certain bodies, where alteration is sufficiently mild to permit distinction of separate phases, it is apparent that too many phases occur to be shown on the geologic map. Similarly, other more altered masses, even if they could be differentiated by detailed field study, could not be shown at the scale of plate 1. For this reason many intrusive masses are shown as single cartographic units although several rock types are present. The rocks are divided into three broad units: (1) rocks of basic and intermediate composition, (2) rhyolite, and (3) coarse-grained porphyritic rhyolite.

Similar intrusive rocks occur in isolated masses in the Kawich and Reville Ranges. They are approximately the same age as the rocks in the Cactus Range and are included in the same cartographic unit.

Rocks of Basic and Intermediate Composition

Rocks in the basic and intermediate composition group include melanodiorite, biotite lamprophyre, low-silica granodiorite, rhyodacite, high-silica granodiorite, and quartz latite porphyry. Of these, only the quartz latite porphyry is mapped separately; the remaining rocks are mapped together and subdivided on the basis of texture into aphanitic and phaneritic varieties.

Melanodiorite.—The melanodiorite is dark gray to black and porphyritic; it occurs as dikes and plugs in the composite stock north of Wellington Hills and as irregular-shaped small apophyses and dikes which intrude the tuff of White Blotch Spring in the northern part of the range. The total area of all exposures probably does not exceed 1 square mile. The rock contains sparse euhedral phenocrysts of clinopyroxene and plagioclase as much as 33 mm in length and glomeroporphyritic clots composed mostly of clinopyroxene as much as 7 mm in diameter. The phenocrysts are set in a diabasic groundmass in which interstices between abundant plagioclase laths are filled with alkali feldspar, clinopyroxene, biotite, and iron oxide. The groundmass plagioclase laths are about 0.15–1 mm long. The finer grained varieties closely resemble basalt.

The typical melanodiorite contains strongly zoned labradorite (50 percent), augite (25 percent), and biotite, alkali feldspar, and iron oxide (about 6 percent each). Some rock is conspicuously porphyritic with euhedral labradorite and augite in a fine-grained granular groundmass. Alkali feldspar is more abundant in the groundmasses of the conspicuously porphyritic rocks than in the more equigranular diabasic varieties. The composition of zones in plagioclase ranges from calcic labradorite in the cores to albite in thin outer rims. Although chemical data are not available, the granular, porphyritic varieties are probably gradational between melanodiorite and granodiorite; the diabasic textured varieties are gabbroic in composition. Thus, a considerable range in composition is indicated for the melanodiorite.

Biotite lamprophyre.—The biotite lamprophyre is medium gray and has conspicuous abundant plates of biotite and subhedral to anhedral alkali feldspar in a medium-grained (about 1.5 mm) matrix composed predominantly of andesine, partly uralitized clinopyroxene, and magnetite. The large biotite and alkali feldspar grains are poikilitic; the biotite grains enclose euhedral andesine crystals, and the feldspar grains enclose euhedral to subhedral grains of andesine, unaltered clinopyroxene, biotite, and magnetite. The rock occurs only in the composite stock north of Wellington Hills.

Low-silica granodiorite.—The low-silica granodiorite is dark gray to black, porphyritic, and not easily distinguished from melanodiorite. It is similar to melanodiorite in abundance and areal distribution, and it also makes up a minor border phase of the large mass located northwest of the Antelope Springs (patterned area, pl. 1). In addition, it is similar in composition and is gradational in texture with the rhyodacite intrusive rock described below. Most of the rock has seriate texture with grains ranging in size from small microlites to crystals about 1 mm long. The larger crystals consist of andesine, augite, hornblende, and magnetite. Mafic phenocrysts are partly uralitized and chloritized. The groundmass is a granular aggregate composed of sodic plagioclase, alkali feldspar, magnetite, quartz, and secondary uralite and chlorite.

Rhyodacite.—Rhyodacite forms plugs, dikes, sills, and most of the larger Roller Coaster laccolith. These masses, which together make up about 15 square miles of outcrop, are distributed throughout the range but are most common in the south and central parts. Rhyodacite is the oldest rock in the Roller Coaster laccolith, which probably is a single large floored and roofed mass that was divided into northwest and southeast segments by intrusion of younger discordant masses of porphyritic granodiorite.

All the rhyodacite, except a few small isolated plugs and dikes, has been propylitically altered. The alteration ranges from incipient albitization of plagioclase and (or) chloritization of biotite and hornblende to complete replacement of phenocrysts and groundmass by chlorite, calcite, albite, epidote, and iron oxide in varying combinations and proportions.

The color of propylitically altered rhyodacite ranges from olive green through light greenish gray to dark greenish gray. Less altered rhyodacites are dark gray, black, or dark purple, and they commonly contain small flecks or spots of secondary green chlorite and (or) epidote. Most rocks weather to greenish brown. Flow alinement of crystals is common and upon weathering gives rise to an accentuated planar structure.

Thirty thin sections of rocks, collected from the Roller Coaster and satellite masses, show considerable petrographic variations. In most rocks plagioclase is the dominant phenocryst; quartz phenocrysts are not sparse, but generally only a few grains occur per thin section; alkali feldspar phenocrysts are sparse or absent. The mafic minerals are mostly altered, but the morphology of replaced grains commonly permits an estimate of original phenocryst assemblages. Some prealteration mafic mineral assemblages are (1) biotite-hornblende (either may dominate), (2) biotite-hornblende-pyroxene (any one may dominate), (3) clinopyroxene-hornblende, and (4) clinopyroxene. Common accessory minerals are magnetite, apatite, zircon, and allanite. Groundmass textures are hyalopilitic, pilotaxitic, trachytic, or granular. Plagioclase is the principal groundmass constituent and is accompanied by biotite, hornblende, and iron oxide. In general, the isolated small intrusive masses tend to be less silicic than the large masses.

High-silica granodiorite.—The high-silica granodiorite is medium gray to greenish gray and porphyritic, and it forms very irregular, mostly discordant masses. These occur in a band extending from 1 mile north of Roller Coaster Knob southwestward to Wellington Hills (patterned area, pl. 1), where it makes up the youngest intermediate intrusive rock. Most of the rock has been weakly propylitized and deuterically altered, but it is distinctly less altered than the rhyodacite just described. It is massive, nonfoliate, and fine to medium grained. Where the rock is fine grained, it is conspicuously porphyritic, carrying phenocrysts of plagioclase, hornblende, and pyroxene. Where it is coarse grained, it is only slightly porphyritic. The rock contains 40–50 percent plagioclase (andesine), about 10 percent hornblende, 10–16 percent quartz, 15–20 percent alkali feldspar, minor biotite, and magnetite. Alkali feldspar and quartz are commonly micrographically intergrown and are generally restricted to interstices between plagioclase and hornblende or pyroxene. The rock is gradational with low-silica granodiorite and is distinguished from it by its more conspicuous quartz. Some rocks gradational between low-silica and high-silica granodiorite contain as much as 20 percent augite. These rocks also contain less alkali feldspar than is typical for the high-silica phase.

Quartz latite porphyry.—Dikes, plugs, and laccoliths of quartz latite porphyry are exposed in a broad northwest-trending band in the north half of the Cactus Range. The bulk of the quartz latite occurs in a large intrusive mass near Sleeping Column Canyon. This mass has a flat-lying to gently dipping roof (fig. 7) from which many dikes and small apophyses extend upward (not visible in fig. 7). The mass is inferred to be a laccolith, although its true shape is unknown because of extensive faulting. Like other large intrusive bodies in the Cactus Range, this laccolith is composite. Part is composed of plagioclase-rich quartz latite, in which alkali feldspar makes up less than 5 percent, and part is composed of coarse-grained porphyritic rhyolite, in which alkali feldspar is the dominant phenocryst. These rocks are not shown separately on plate 1.

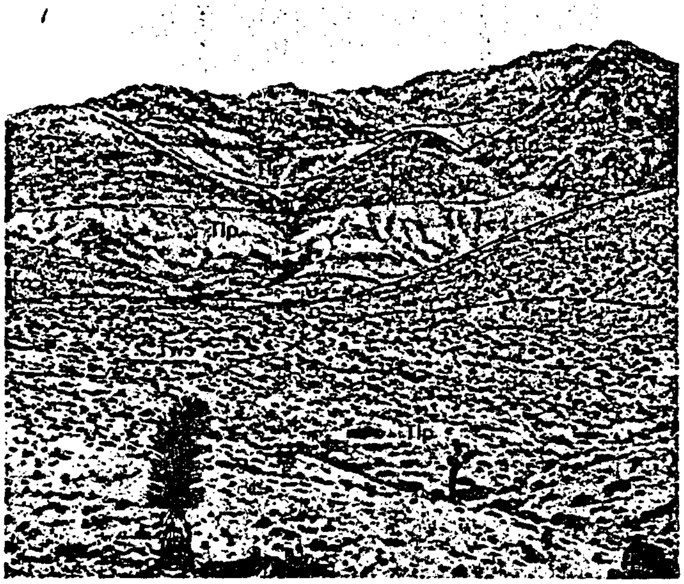

FIGURE 7.—Outcrops of quartz-latite porphyry laccolith (Tlp) and tuff of White Blotch Spring (Tws). The strata are repeated by normal faults that displace the rocks down toward the viewer. Both rocks are highly altered at the contact. Light dashed lines indicate formation contacts. Solid lines outline topography. Heavy dashed lines indicate faults. Fault traces are dotted where concealed behind hills; U, upthrown side; D, downthrown side. View to the northeast into the Cactus Range from a point 2 miles southwest of Urania Peak.

In the northern part of the range the quartz latite porphyry is unaltered to weakly altered. Where unaltered, the rock is gray; where altered, it is green or greenish gray as a result of secondary chlorite and epidote. The freshest rocks contain about 40 percent plagioclase (commonly glomeroporphyritic clots of andesine), 8–10 percent biotite, 4–9 percent altered pyribole, 2–5 percent conspicuous large alkali feldspar (as large as 2 cm), 10 percent quartz, and 2.5 percent magnetite. The groundmass is mostly fine grained and felsic. The quartz latite porphyry is distinguished from other intrusive rocks by the sparse large and conspicuous alkali feldspar and quartz phenocrysts. Porphyritic rhyolite phases included in this unit are similar to those described below.

RHYOLITE

The rhyolite is white to light gray, crystal poor, massive to flow layered, and generally silicified. It forms several large plugs and numerous small dikes and plugs. Phenocrysts comprise less than 5 percent of the rock and are predominantly quartz and alkali feldspar with minor plagioclase and biotite. The groundmass is felsic and is cryptocrystalline to fine-grained saccharoidal.

COARSE-GRAINED PORPHYRITIC RHYOLITE

Coarse-grained porphyritic rhyolite crops out discontinuously over about 2 square miles in the northern Cactus Range. Most of the rock is part of a gently dipping body that has a flat-lying roof. Contacts with enclosing tuffs are marked in many outcrops by altered zones several feet thick. The tuff of White Blotch Spring is the youngest rock intruded by the porphyry in this area.

The coarse-grained porphyritic rhyolite is light brown to light orange brown and weathers to light-brown grus-covered rounded knobs and hills that are not easily distinguished from outcrops of tuff of White Blotch Spring. Phenocrysts makeup 30–75 percent of the rock and consist of quartz (20–30 percent) and plagioclase (30–40 percent) that range in size from 1 to 5 mm, and alkali feldspar (40–50 percent) as much as 2 cm long. Biotite and magnetite are the only mafic minerals, and they constitute less than 1 percent of the phenocrysts. The groundmass is aphanitic and is composed predominantly of equigranular quartz and alkali feldspar. The abundance of quartz and the occurrence of pink to pale-orange alkali feldspar phenochysts are distinguishing mineralogical characteristics.

An impressive system of east- and northwest-trending dikes, some of which are as much as 6 miles long, transects the Cactus Range south of Cactus Spring. The dikes are composed predominantly of light-gray to light brown highly porphyritic rhyolite that is similar in general appearance and mineralogy to the rock just described. Many of the porphyritic dikes have thin selvages of white fine-grained flow-laminated rhyolite that grade within short distance (a few inches to 2 ft) into highly porphyritic rock. Where the dikes are narrow, they are composed entirely of white, crystal-poor rhyolite.

Chemistry

Chemical analyses and norms of 18 intrusive rocks in the Cactus Range are given in table 9. Samples 1–2 and 11–13 are of phaneritic rocks, ranging from quartz latite porphyry to melanodiorite, which are only incipiently or mildly altered in contrast to samples 3–7, 9, and 14–15, which are highly propylitized. This contrast in degree of alteration is reflected by the presence of less than 0.25 percent CO_2 in the first group and 0.9–3.2 percent CO_2 in the second group.

Chemical variation among these rocks is shown in figures 8 and 9. The trend lines are drawn on the figures as best visual fits to the points representing rocks with negligible CO_2 (dots). Tielines extend from the open circles to positions corresponding to ratios calculated after the reported CaO was reduced by an amount sufficient to form calcite from all the reported CO_2. Although the magnitude of shift between each pair of points is large for some rocks with high CO_2 contents, the direction of shift so nearly parallels the trend line that it is impossible to unambiguously determine whether CaO should be corrected for CO_2. There is, nevertheless, a slightly better fit of the trend line to the open circles than to the positions at the ends of the tielines; this fit suggests that calcium was not introduced along with CO_2 during the alteration process. On the basis of this relationship, the normative percentages for the altered rocks reported in table 9 were calculated assuming no addition of CaO. Water and CO_2 were removed and the analyses were recalculated to 100 percent before the norm was calculated. When the norm is calculated in this way, corundum appears in only three rocks (samples 9, 17, 18, table 9), whereas corundum will appear in the norms of all the altered rocks if the percentage of CaO is reduced by an amount sufficient to form calcite from all Co_2.

The general absence of corundum in the reported norms is assumed to indicate that only minor amounts of alkalis were lost during the alteration process. This conclusion is also supported by comparison of partial molecular norms (fig. 10) for unaltered and altered pairs of rocks. For example, sample 16 is of quartz latite porphyry that shows partial alteration of biotite phenocrysts to chlorite, hornblende to chlorite and epidote, and plagioclase to sericite. The rock contains no secondary calcite or albite and is considered to have undergone only incipient propylitic alteration. In contrast, sample 14 is of quartz latite porphyry that is completely altered to an assemblage of albite, quartz, sericite, calcite, epidote, and iron oxide. The partial molecular norms of the two rocks plot very close together when CO_2 is considered a mobile component (fig. 10). A similar comparison can be made between samples 5 and 6, which are rhyodacite from the Roller Coaster laccolith southeast of Cactus Springs. Both rocks are altered and both contain CO_2, but sample 5 is much more intensely altered. If the partial molecular norms for these two rocks had been calculated after correcting the CaO contents for CO_2, the points would be widely separated in figure 10. These comparisons

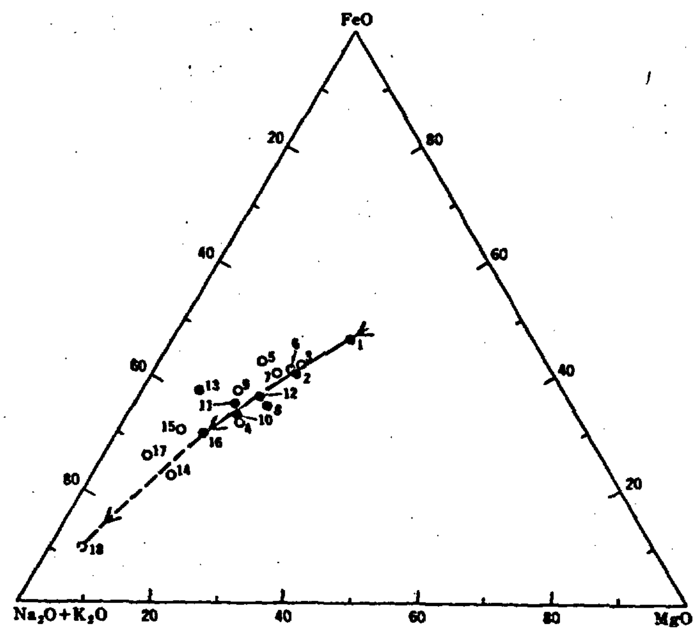

FIGURE 8.—Plot of $FeO:Na_2O+K_2O:MgO$ ratios of intrusive rocks from the core of the Cactus Range. Open circles correspond to rocks having appreciable CO_2; solid circles, to rocks having little or no CO_2. Numbers refer to sample numbers in table 9. Arrows point in direction of more siliceous and, in general, younger rocks. The dashed extension of the trend line is based on a single analysis of weakly altered porphyritic rhyolite (sample 18, table 9).

TABLE 9.—*Chemical analyses and norms of fresh and propylitically altered intrusive rocks in the core of the Cactus Range*

[Analyses by P. L. D. Elmore, S. W. Botts, G. W. Chloe, Lowell Artis, and H. Smith, by rapid method (Shapiro and Brannock, 1962). Norms determined from chemical analyses recalculated to 100 percent minus H_2O+CO_2]

Sample	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
Field No.	CS-57	CS-61	CS-299	CS-227A	CS-83	CS-93	M-198	CS-291	CS-91	CS-383	CS-105	M-214	M-205	CS-379	CS-244	CS-371	CS-353	CS-76
Laboratory No.	162051	160949	162054	162057	164507	160951	160966	162052	164506	162861	160952	160968	160967	162858	162055	162857	162866	160950
Chemical analyses																		
SiO_2	50.2	58.1	53.8	56.1	57.1	59.1	59.3	60.4	61.5	63.5	61.4	61.7	62.9	62.8	65.3	68.1	69.4	67.2
Al_2O_3	17.7	16.3	16.7	17.8	15.9	16.1	16.2	16.9	16.5	14.1	16.1	16.2	16.1	15.7	15.5	13.3	15.6	16.1
Fe_2O_3	4.9	3.8	4.5	4.3	3.5	2.9	2.7	3.0	4.7	3.7	3.4	3.1	3.0	2.0	2.7	2.9	2.6	1.2
FeO	4.9	2.8	2.6	2.8	2.6	2.8	3.5	2.0	1.4	1.5	2.2	2.6	2.2	1.1	1.1	1.0	.40	.93
MgO	5.3	2.3	3.4	2.1	2.2	2.7	2.7	2.7	1.6	2.4	2.3	2.7	1.1	1.5	1.0	1.6	.67	.55
CaO	8.2	6.1	7.2	6.5	6.2	6.0	5.4	5.2	4.6	3.6	4.0	4.8	4.4	3.6	3.9	2.4	1.9	1.4
Na_2O	3.2	2.8	3.4	3.0	2.7	2.6	2.8	3.7	2.4	2.9	3.4	3.2	3.3	3.8	3.1	3.1	2.8	3.5
K_2O	2.3	3.1	2.5	3.2	2.9	2.7	3.2	2.4	3.1	4.4	4.2	3.7	3.8	4.6	3.8	4.0	4.4	6.0
H_2O-	.20	.32	.30	.18	.23	.29	.22	.88	.54	.33	.28	.17	.17	.17	.32	.44	.53	.26
H_2O+	.92	1.6	1.30	1.4	2.1	1.9	1.8	1.7	2.1	1.8	1.9	.94	1.4	1.6	1.0	1.6	1.3	1.1
TiO_2	1.3	.78	.95	.86	.74	.66	.73	.73	.79	.69	.60	.84	.69	.48	.48	.45	.42	.31
P_2O_5	.58	.39	.52	.32	.39	.27	.20	.35	.42	.36	.41	.33	.31	.22	.21	.20	.00	.12
MnO	.15	.12	.08	.08	.11	.10	.10	.09	.11	.00	.00	.10	.09	.08	.10	.04	.02	.06
CO_2	.07	.21	2.10	1.6	3.2	1.8	1.6	<.05	.12	.16	.07	<.05	<.05	1.2	1.2	<.05	<.05	.90
Total	100	100	99	100	100	100	101	100	100	100	100	100	100	100	100	99	101	100
Norms																		
Q	....	14.4	7.8	11.6	16.6	16.5	15.5	16.1	25.4	20.8	15.2	15.8	19.6	18.3	24.4	28.7	31.8	23.6
C	....	....	....	....	....	....	....	....	1.9	....	....	....	....	....	....	....	2.8	3.8
or	13.6	18.8	15.4	19.5	18.2	16.2	19.6	14.6	18.9	26.7	25.3	22.0	22.9	28.3	23.1	24.3	26.5	36.8
ab	27.1	24.3	30.1	26.2	24.2	22.9	24.5	32.1	20.9	25.2	29.3	27.3	28.5	33.5	27.0	27.0	24.1	30.8
an	27.2	23.9	24.0	25.8	24.1	29.1	22.9	23.0	20.7	12.8	16.6	19.1	18.3	12.7	17.7	10.9	9.6	.5
wo	6.2	1.5	4.1	2.2	2.4	....	1.4	.5	....	1.3	.2	1.2	.8	1.8	.4	....	....	....
en	9.3	5.4	5.9	5.4	5.8	7.0	6.9	6.9	4.1	6.1	5.5	6.8	2.8	3.9	2.6	4.1	1.7	1.4
fs	2.2	1.0	....	.3	.9	1.9	3.3	.2	....	....	.3	1.0	.6	....	....	....	....	.4
fo	2.8	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....
fa	.7	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....
mt	7.1	5.7	6.2	6.4	5.4	4.4	4.0	4.5	2.7	3.2	5.0	4.5	4.4	2.5	2.6	2.1	.1	1.8
hm	....	....	.5	....	....	....	....	....	3.0	1.6	....	....	....	.3	1.0	1.5	2.6	....
il	2.5	1.5	1.9	1.7	1.5	1.3	1.4	1.4	1.5	1.3	1.3	1.6	1.3	1.0	.9	.9	.8	.6
ap	1.4	1.0	1.3	.8	1.0	.7	.5	.9	1.0	.9	1.0	.8	.8	.5	.5	.5	....	.3

Sample, locality, and description

[Numbers following mineral names are percent of total phenocrysts]

1. About 2.5 miles southeast of Antelope Peak, Cactus Range. Medium-grained melanodiorite composed of strongly zoned plagioclase (sodic labradorite) 56, clinopyroxene 24, magnetite 5; 5 percent each of interstitial alkali feldspar and biotite and 4 percent secondary chlorite and sericite.
2. About 2.5 miles southwest of Antelope Peak, Cactus Range. Low-silica granodiorite; rock contains phenocrysts of plagioclase, clinopyroxene, hornblende, and magnetite in a very fine-grained groundmass composed predominantly of alkali feldspar, quartz, plagioclase, and magnetite.
3. Incipiently altered rhyodacite; plagioclase phenocrysts are fresh; secondary calcite, chlorite, and magnetite have completely replaced pyroxene and partly replaced hornblende; groundmass is a delicate felt of feldspar and iron oxide with small patches and veinlets of secondary calcite.
4. Small plug 1 mile southwest of Cactus Spring. Weakly propylitized low-silica granodiorite; mafic minerals are completely replaced by calcite, chlorite, and iron oxide; plagioclase phenocrysts (sodic labradorite) are fresh and commonly glomeroporphyritic; groundmass is pilotaxitic with andesine microlites and interstitial alkali feldspar.
5. Small satellitic intrusive mass west of Roller Coaster laccolith 1.5 miles southeast of Urania Peak. Intensely propylitized rhyodacite; principal secondary minerals are chlorite, magnetite, and abundant calcite (shown by high CO_2 content); minor secondary quartz; no epidote and only minor albite; mafic phenocrysts are completely replaced and plagioclase phenocrysts partly altered to calcite and sericite; groundmass is completely altered.
6. Roller Coaster laccolith 3.5 miles southeast of Cactus Spring. Moderately propylitized rhyodacite; principal secondary minerals are calcite, chlorite, epidote, magnetite, and clay; biotite, pyroxene, and hornblende are completely replaced, plagioclase phenocrysts and groundmass laths are mostly fresh; no apparent albitization.
7. Roller Coaster laccolith at Antelope Springs. Moderately propylitized rhyodacite; principal secondary minerals are chlorite, calcite, and clay; mafic minerals are completely replaced; no secondary epidote or albite.
8. About 1.5 miles south of O'Briens Knob, Cactus Range. Black vitric dacite lava containing 30 percent phenocrysts consisting of 70 percent plagioclase (an$_{48}$) and 30 percent augite, brown hornblende, and magnetite in a ratio of 2:2:1 and a trace of biotite set in a glassy groundmass containing tiny blebs of magnetite and small laths of plagioclase.
9. Thin sill 3 miles southeast of Urania Peak. Intensely propylitized and weakly pyritized rhyodacite; entire rock is altered to secondary albite, epidote, sericite, chlorite, quartz, iron oxide, and pyrite; notable lack of secondary carbonate.
10. About 2.8 miles southeast of Urania Peak. Weakly altered quartz latite porphyry.
11. About 1.2 miles west of Antelope Peak, Cactus Range. Granodiorite similar to 12.
12. About 1.2 miles southwest of Roller Coaster Knob, Cactus Range. Granodiorite similar to 13 with 40 percent plagioclase (an$_{32}$), 28 percent alkali feldspar, 10 percent hornblende, 16 percent quartz, 3 percent partially chloritized biotite and minor clinopyroxene, magnetite, sphene, and apatite.
13. About 1.5 miles northwest of Antelope Springs, Cactus Range. Greenish-gray porphyritic granodiorite with phenocrysts of plagioclase, hornblende, clinopyroxene, biotite, and magnetite in a medium-grained matrix composed predominantly of quartz, alkali feldspar, and plagioclase; biotite grains are wholly or partially chloritized.
14. About 2 miles south-southeast of Urania Peak. Highly propylitized quartz latite porphyry; rock is altered equivalent(?) of sample 16 but contains less quartz; principal secondary minerals are albite, epidote, sericite, calcite, chlorite, quartz, and magnetite; quartz phenocrysts are unaltered; alkali feldspar phenocrysts are partly replaced by sericite, epidote, and calcite; all other primary minerals are completely pseudomorphed.
15. About 2 miles south-southeast of Cactus Peak. Light-gray to light-greenish-gray quartz latite porphyry; 40 percent phenocrysts consisting of plagioclase 70, quartz 9, biotite 8, altered hornblende(?) 9, and minor alkali feldspar and magnetite in a fine-grained groundmass composed predominantly of quartz. alkali feldspar, and plagioclase.
16. About 1.2 miles southwest of Urania Peak. Quartz latite porphyry; 38 percent phenocrysts consisting of andesine 70, biotite 11, quartz 11, and minor alkali feldspar, pyrobole, and magnetite in a felsic fine-grained granular groundmass. Note anomalously high silica content—some of granular quartz in groundmass may be secondary; silicification is locally intense in adjacent terrain.
17. South side Sleeping Column Canyon 3 miles west of Cactus Spring. Quartz latite porphyry; about 30 percent phenocrysts which are, in order of decreasing abundance, plagioclase, alkali feldspar, biotite, and quartz in a granular groundmass of the same minerals plus magnetite. Rock is weakly argillized and may be weakly silicified.
18. About 2.5 miles south-southwest of Cactus Peak. Coarse-grained porphyritic rhyolite.

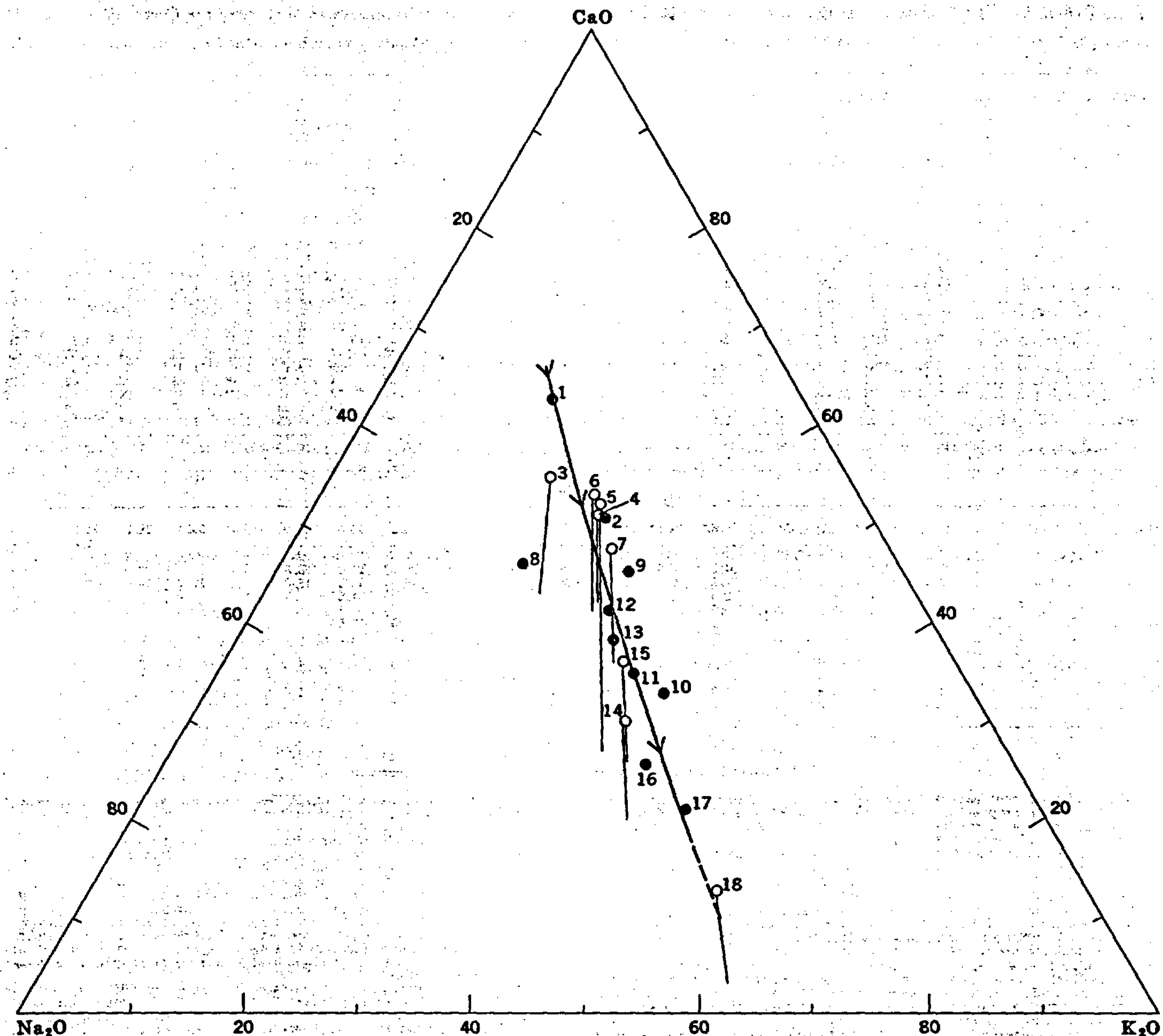

FIGURE 9.—Plot of $CaO:Na_2O:K_2O$ ratios of intrusive rocks from the core of the Cactus Range. Open circles correspond to rocks having appreciable CO_2; solid circles, to rocks having little or no CO_2. Numbers refer to sample numbers in table 9. Arrows point in direction of more siliceous and, in general, younger rocks. The dashed extension of the trend line is based on a single analysis of weakly altered porphyritic rhyolite (sample 18, table 9). Tielines extend from the open circles to positions corresponding to ratios calculated after the reported CaO was reduced by an amount sufficient to form calcite from all the reported CO_2.

indicate that only minor gains or losses of alkalis and calcium occurred during propylitic alteration. This stability is in sharp contrast with a rather high mobility of sodium and calcium recognized in the silicic and argillic or "vein-forming" type of alteration based on unpublished analyses of rocks from the Cactus Range.

The arrows shown on the trend lines in figures 8 and 9 point in the direction of the more high-silica rocks. The positions of points along these lines correspond to a general decrease in age in the direction of the arrows. The rocks lowest in silica are from the early intrusive rocks in the Wellington Hills area and from the Roller Coaster laccolith. The progression passes through the low-silica variety of granodiorite, then through quartz latite porphyries, and finally to the porphyritic rhyolite, which is the youngest intrusive

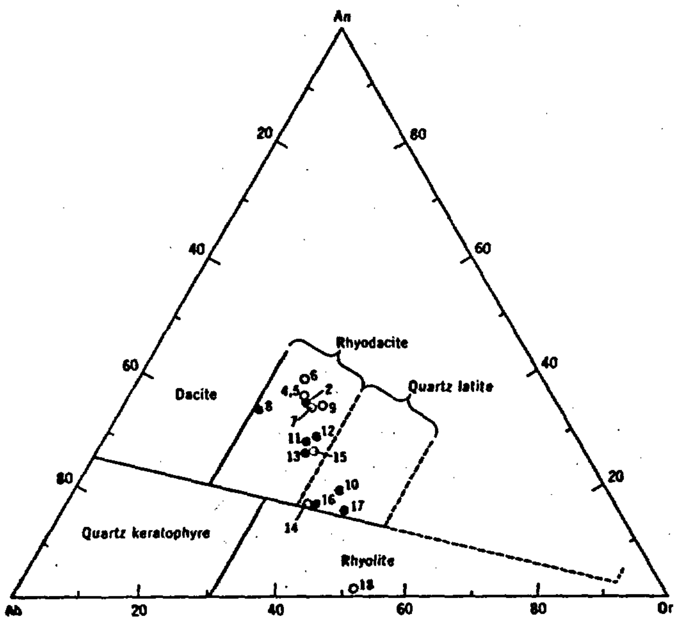

FIGURE 10.—Plot of normative albite-anorthite-orthoclase ratios of intrusive rocks from the core of the Cactus Range. Solid circles are plots of relatively unaltered rocks with little or no CO_2; open circles correspond to rocks with appreciable CO_2 and the norms of these rocks were calculated considering CO_2 as a mobile component. Numbers refer to sample numbers in table 9. Fields are from O'Connor (1965).

rock in the central core of the range. The concordance of this chemical trend with age is consistent with the interpretation that these rocks have a common parental magma.

RHYOLITE OF WHITE RIDGE AND TUFFACEOUS ROCKS

A small volcano from which ash-fall tuff and rhyolite lava were erupted is centered on White Ridge about 2.5 miles southeast of Cedar Pass in the Kawich Range. Stratified ash-fall tuff, which accumulated around the center to distances of a few miles, represents the earliest record of volcanic activity at the White Ridge volcano. The tuff was deposited on an erosional surface of considerable relief developed on dacite and andesite lava. The tuff is white to light gray and even bedded, and it contains sparse small phenocrysts of alkali feldspar, quartz, and plagioclase. Locally, the tuff is fused to greenish-gray and dark-gray vitrophyre adjacent to rhyolite intrusive masses that crop out in the center of the volcano and in several radial dikes. These fused zones are no wider than 50 feet. Other fused or welded bedded tuffs occur at distances of as much as 1 mile from the nearest visible intrusive mass. These tuffs are red brown to orange brown and contain collapsed vitrophyric pumice lapilli. They apparently accumulated very rapidly near the vent and retained sufficient initial heat to permit welding. Similar welded bedded tuffs around volcanic vents have been recognized by R. L. Smith (oral commun., 1964), who has suggested the term "agglutinate" for such tuffs.

The intrusive rhyolite is white to pale yellow, crystal poor, flow laminated, and locally brecciated. The only rocks that can be identified as lavas are exposed west of Cedar Pass, 5 miles northwest of the volcano. It is not known whether these were extruded from the volcano or from other unrecognized vents. The lavas are slightly richer in crystals than is the intrusive rhyolite. Alkali feldspar and quartz are the predominant phenocrysts; plagioclase and biotite occur in lesser amounts.

Rocks occupying the same position as the rhyolite of White Ridge crop out in both the Belted and Cactus Ranges. Although no genetic or absolute age equivalence is inferred for these widely separated rocks, they are mapped as the same cartographic unit as the rhyolite of White Ridge and associated tuffaceous sedimentary rocks. The rhyolite in the Belted Range is 50–100 feet thick and overlies about 30 feet of well-bedded ash-fall tuff which, in turn, overlies quartz latite lava. The rhyolite is crystal poor, pink to gray, and flow laminated, and at the base it contains a thick vitric flow breccia. About 4 miles southeast of these exposures ash-fall tuff and tuffaceous sedimentary rocks, having a combined thickness of about 500 feet, crop out. These strata rest on lavas of intermediate composition and are overlain conformably by Fraction Tuff. The strata are best exposed near White Saddle Pass, where the bulk of the rock consists of vitric to slightly devitrified ash-fall tuff. Tuffaceous sedimentary rocks containing fragments of dacite, rhyolite, quartzite, and chert crop out near the base. The rocks are commonly cross-stratified and range in color from pastel shades of brown and green to white.

About 400 feet of well-stratified mudstone, sandstone, gravel, conglomerate, and minor ash-fall and ash-flow tuff is exposed in the northwestern Cactus Range. These strata rest on dacite lavas and are overlain by Fraction Tuff and post-Fraction rhyolites. They dip about 50° W. and are repeated several times by faulting. Locally they are separated from the underlying lavas by remnants of quartz-bearing, plagioclase-rich, biotite-rich, quartz latitic welded tuff. This tuff probably correlates with the dacite vitrophyre of Ransome (1909), which is nearly identical with it and which occupies the same stratigraphic position. Ransome (1909, p. 64) noted the chemical similarity between the tuff (dacite vitrophyre) and the underlying dacite lavas and suggested that the two rocks are of nearly the same age. The present authors share this opinion and interpret the tuff as a late eruptive phase of igneous activity of dominantly intermediate composition. The sedimentary rocks are brown, greenish brown, reddish gray, and pastel shades of yellow and pink. The lighter colors are characteristic of zeolitized, silicified, or argillized rock. Pink pumice-rich massive air-fall(?) tuff is common in the upper part subjacent to the Fraction Tuff. Where the rocks rest on dacite lavas, they contain boulders and cobbles derived from the lavas. Much of the finer grained clastic rock is rich in plagioclase and biotite; this richness indicates that these clastic rocks were also derived in part from rocks of intermediate composition. The deposition of the clastic rocks is probably related to erosion of the central core of the Cactus Range as it was elevated following the extrusion of lavas and tuff of intermediate composition (p. 72).

A small exposure of brown sandstone rich in biotite and plagioclase is located 1 mile northeast of Antelope Springs. The sandstone dips about 15° E. and is overlain by minor zeolitized tuff and at least two flows of rhyolite of O'Briens Knob (p. 56). Biotite and plagioclase in the sandstone are unaltered. These fresh grains could not have been derived from any rocks now exposed in the central core of the Cactus Range. They are presumed to have been derived from tuffs or lavas that once covered the range and were stripped off when the range rose.

FRACTION TUFF AND RELATED ROCKS

FRACTION TUFF

The Fraction Tuff was first described by Spurr (1905), who named the unit the Fraction Dacite Breccia. The type area was given as the south half of the Tonopah mining district, where the rock overlies lavas of intermediate composition and underlies bedded tuffs and lake beds. Nolan (1930), realizing that the rock was not a dacite in composition, changed the name to Fraction Breccia. The Fraction Tuff is a major map unit in the project area, and in nearly all exposures it overlies intermediate lavas and underlies bedded tuffs or sedimentary rocks.

Distribution and Thickness

The known area of distribution of the Fraction Tuff extends from the Belted Range, where it is as much as 1,000 feet thick, to Tonopah, Nev., where it is at least 745 feet thick (Spurr, 1905). The thickest and best exposures in the project area are at Trailer Pass in the Cathedral Ridge caldera, where a composite measured section [1] indicates a thickness of about 7,200 feet. The tuff between Quartzite Mountain and Pahute Mesa has been faulted out or covered by younger strata, and its extent is therefore unknown. Thicknesses of 1,000 and 1,150 feet have been penetrated in two drill holes located about 30 miles south of Cathedral Ridge on the south flank of Pahute Mesa, and several hundred feet has been penetrated in a drill hole in northern Yucca Flat (W. J. Carr, oral commun., 1966). The tuff is absent from the vicinity of Coyote Cuesta and Triangle Mountain, where older lavas of intermediate composition are exposed, but several hundred feet is exposed north of Coyote Cuesta near Mellan. The rocks at Mellan are identical with those at Cathedral Ridge.

Ash-flow tuffs that are petrographically similar to the Fraction Tuff crop out on the northwest flank of the Cactus Range. These strata are mapped as Fraction Tuff (pl. 1), but the correlation is questionable. Common rock colors are light chocolate brown, charcoal gray, or salmon pink, none of which is characteristic of the Fraction elsewhere in the mapped area. Also, lithic fragments are sparse and those of metamorphic rocks are very rare.

In all the areas of outcrop there is evidence that Fraction Tuff was deposited on an erosional surface of considerable relief. The thinning of the Fraction toward Quartzite Mountain and at White Ridge probably indicates that these areas were topographically high during the period of Fraction Tuff eruption. In the Belted and Cactus Ranges the Fraction is locally separated from the underlying dacite lavas by 50–300 feet of tuffaceous clastic sedimentary rocks including coarse conglomerate containing well-rounded cobbles of pre-Tertiary quartzite and volcanic rocks; these rocks record an erosional interval of unknown duration. Subsurface investigations in the western part of the Tonopah mining district (Bastin and Laney, 1918; Nolan, 1935) have shown that the Fraction rests unconformably on lavas of intermediate composition, and ore-bearing veins are locally truncated by the tuff.

Lithology and Petrology

At Cathedral Ridge the Fraction Tuff is pinkish gray, pale brown, and light red; it weathers to pale brown, brown, and dark brownish gray. The tuff forms a prominent sequence of steep cliffs and slopes on Cathedral Ridge. The differences in hardness and weathering characteristics result from variations in degree of devitrification and argillic alteration rather than pronounced differences in the degree of welding. Pods and lenses of dark-gray to black vitrophyre, indicative of partial cooling breaks, occur at several intervals, and at least one vitrophyre can be traced the length of Cathedral Ridge. Thin zones of welded ash-fall tuff

[1] The lower approximately one-third part of the measured section is in a single fault block about 1 mile south of Trailer Pass; the upper two-thirds is in another block located about 2,000 feet south of the lower line of section. Covered intervals in the upper two-thirds may conceal small faults, and the total thickness, therefore, might be slightly in error. A gradual upward decrease in welding indicates that repetitions, if they do exist, are minor.

have also been recognized within the sequence in this area. Except for a nonwelded to poorly welded light-yellow zeolitized basal zone of variable thickness, this sequence of ash flows is densely welded and is interpreted as a compound cooling unit. South of Trailer Pass, however, at about 4,200 feet above the base, the tuff is very weakly welded and has a low bulk density. (See field No. Ttp–17, table 10.) In this locality a partial cooling break is inferred. The interval is poorly exposed and it is not known whether the poorly welded rock constitutes the upper or lower part of a cooling zone.

North of Wild Horse Draw in the Kawich Range, the formation loses its stratified appearance and weathers to massive, brown, devitrified outcrops without visible cooling breaks.

The assemblage of lithic fragments is a diagnostic feature of the Fraction Tuff throughout Cathedral Ridge, the Kawich Range, and at Tonopah. Indeed, it was the abundance of these fragments at Tonopah that led Spurr (1905) and others to consider the rock to be a breccia. The fragments, in order of decreasing abundance, are intermediate lavas, gneiss, schist, granite, and sedimentary rocks of probable Paleozoic age. At Cathedral Ridge the fragments increase in abundance and size from the base to the approximate middle of the section and remain fairly constant from the middle to the top. At the base of the section, fragments average less than 2 inches in diameter and make up less than 10 percent of the rock; near the middle they are commonly 10 inches long, rarely as much as 3 feet, and make up 20–30 percent of the rock. The increase in abundance of fragments upward in the section and the occurrence of metamorphic fragments indicate that the fragments were derived from the crust during the ascent of the Fraction magma and were not merely picked up by the tuff as it flowed over the ground surface. Pumice lapilli are common. They are flattened, black, and vitrophyric where the tuff is densely welded and white to light gray where it is weakly welded.

In the Belted Range the Fraction is 500–1,000 feet thick, and two cooling units are inferred to be present. The lowest unit is about 300 feet thick and has as much as 200 feet of nonwelded to partially welded light-gray tuff at base that grades upward to a weakly welded gray-tan top. The overlying unit is pale brown to gray brown and contains abundant dark-gray to black vitric pumice lapilli. This unit is partially welded at base, densely welded in the middle, and nonwelded at the top. The nonwelded top was deeply eroded prior to the deposition of overlying sedimentary rocks and is preserved in only a few localities. The entire sequence contains abundant lithic fragments of rhyolite and dacite and sparse fragments of gneiss and schist.

In the northern Cactus Range the rocks that are tentatively assigned to the Fraction Tuff include three cooling units totaling 450 feet in thickness. The lower cooling unit, about 100 feet thick, is pumice rich and light chocolate brown. The middle cooling unit consists of a basal white to pink nonwelded zone 50 feet thick, a pink shard-rich vitric densely welded zone 75 feet thick, and an upper partly welded vapor-phase

TABLE 10.—*Physical properties of Fraction Tuff and underlying lavas at Trailer Pass, Kawich Range, Nye County, Nev.*

[Analysts: E. F. Monk, W. H. Lee, and W. R. Eberle. All samples were thin rectangular slabs]

Lab. No.	Field No.	Rock type	Porosity (percent)	Dry bulk density (g/cc)	Grain density (g/cc)	Saturated bulk density (g/cc)	Magnetic susceptibility (10^{6} cgs)	v_p longitudinal velocity (fps)	v_s Shear velocity (fps)	σ Poisson's ratio	E Young's modulus (10^6 psi)	G Shear modulus (10^6 psi)	K Bulk modulus (10^6 psi)
D700708	Ttp- 1	Dacite lava	6.7	2.50	2.68	2.57	3,517	15,070	8,381	0.276	6.03	2.37	4.49
709	2	Latite lava	4.3	2.52	2.64	2.57	(1)	(1)	(1)	(1)	(1)	(1)	(1)
710	4	Welded tuff	6.3	2.46	2.63	2.53	254	13,497	7,840	.245	5.07	2.04	3.32
711	5	do	1.9	2.38	2.42	2.40	214	14,641	8,470	.249	5.74	2.30	3.81
712 [2]	6	Welded tuff, vitrophyre	1.7	2.38	2.42	2.39	223	12,502					
713	7	Welded tuff	9.0	2.36	2.59	2.45	146	14,155	8,142	.253	5.28	2.11	3.56
714	8	Welded tuff, vitrophyre	2.4	2.38	2.44	2.40	176	8,552	5,189	.209	2.09	0.86	1.19
715	9	Welded tuff	7.5	2.40	2.60	2.48	118	13,937	8,803	.168	5.85	2.51	2.94
716	10	do	10.1	2.34	2.60	2.44	22	12,891	7,258	.268	4.21	1.66	3.02
717	11	Welded tuff, vitrophyre	1.0	2.42	2.44	2.43	410	17,646	10,238	.246	8.52	3.42	5.59
718	12	Welded tuff	7.7	2.40	2.60	2.48	578	13,570	8,142	.219	5.22	2.14	3.10
719 [2]	13	do	6.5	2.45	2.62	2.51	304	12,262					
720 [2]	14	do	1.7	2.39	2.43	2.41	401	14,949					
721 [2]	15	do	7.7	2.37	2.57	2.45	916	13,254					
722 [2]	16	do	10.5	2.30	2.57	2.40	974	12,974					
723	17	do	27.7	1.77	2.45	2.05	331	12,285	6,898	.270	2.88	1.13	2.09
724	18	do	6.4	2.39	2.55	2.45	851	12,579	6,986	.277	4.01	1.57	3.00
725 [2]	19	Welded tuff, vitrophyre	3.0	2.38	2.46	2.41	596	11,562					
726	20	Welded tuff	12.5	2.25	2.57	2.37	609	10,827	6,723	.186	3.25	1.37	1.73
727	21	do	12.2	2.21	2.52	2.33	469	11,732	6,463	.282	3.19	1.24	2.44
728 [2]	22	do	17.5	2.08	2.53	2.26	548	9,293					
729 [2]	23	do	19.1	2.06	2.54	2.25	388	11,023					

[1] No sample received.
[2] S-wave indeterminable.

zone 25 feet thick. The upper cooling unit, about 150 feet thick, consists of completely devitrified partially welded charcoal-gray to pink tuff which is characterized by a distinctive clinkery aspect. The rocks contain 25–35 percent phenocrysts, which consist of subequal plagioclase, alkali feldspar, quartz, and accessory amounts of biotite, hornblende, allanite, and sphene. On the west flank of the range, 1 mile north of Kawich Road, these tuffs are at least 1,200 feet thick. An unknown thickness of similar tuffs is exposed on the south flank of the Goldfield mining district just west of the project area.

Petrographic studies of 16 thin sections from the measured section on Cathedral Ridge reveal mineralogical variation from plagioclase-rich, relatively mafic-rich quartz latite to alkali feldspar-rich, mafic-poor rhyolite. Modal data obtained from eight thin sections which represent the extremes of variation throughout the section together with modal estimates of the remaining eight thin sections are plotted in figure 11. The lowest specimen (sample 2) is quartz latite lava that crops out beneath the Fraction Tuff. This lava is very similar mineralogically to the tuffs directly above it. The tuffs are rich in zoned plagioclase and mafic minerals and poor in alkali feldspar. In the interval between samples 5 and 10, the rocks are poor in mafic minerals and contain approximately equal amounts of plagioclase and alkali feldspar. They are notably rich in quartz, which is the dominant phenocryst. Phenocryst assemblages in the rocks between samples 10 and 19 show wide variations, but in most of the rocks plagioclase is the dominant phenocryst and quartz averages about 30 percent. In these rocks, biotite is more abundant than in the underlying rocks, and hornblende is rare. Between samples 19 and 22 the rocks are rich in plagioclase and contain about 20 percent each of quartz and alkali feldspar. These rocks also contain less hornblende than biotite. In general, throughout the section plagioclase varies antithetically with alkali feldspar and sympathetically with biotite.

The Fraction Tuff exposed on the east flank of the Belted Range is similar petrographically to the alkali feldspar-rich, mafic-poor tuff that occurs in the interval between samples 5 and 10 in the section measured south of Trailer Pass on Cathedral Ridge (fig. 11).

Despite its varied phenocryst content, the Fraction Tuff can be distinguished from other tuffs in the region by (1) the lack of pyroxene even in mafic-rich varieties, (2) the presence of trace amounts of allanite, which is common also in the tuff of Wilsons Camp (see earlier pages), (3) the presence of plagioclase phenocrysts that generally show strong normal zoning and are commonly aggregates of several grains, and (4) by a distinctive assemblage of lithic fragments.

Chemical analyses and norms of six samples of Fraction Tuff are given in table 11. The tuffs are typically calc-alkalic. Corundum appears in the norm of all these rocks as an indication of their peraluminous character or of the leaching of alkalis. Petrographic studies showed that the amount of normative anorthite varies with the amount of modal plagioclase and that the amount of TiO_2 varies with the amount of modal sphene. A comparison of normative and model data is justified by a general parallelism of variations among analyzed specimens. Analyses 1 and 2 represent the approximate maximum mineralogical variation in the unit at the type area, and they are, therefore, assumed to approximate the maximum chemical variation. The tuff that crops out on the east flank of the Belted Range (sample 4) is chemically similar to the higher silica variety from the Cathedral Ridge (sample 1), but the latter rock has a slightly lower ratio of K_2O to Na_2O. These two rocks, plus a weakly altered sample of Fraction Tuff from the Belted Range and a sample of vitrophyre from the Cactus Range, plot in the rhyolite field; samples 2 and 3 plot in the quartz latite field (fig. 12).

Age

A Miocene age is indicated for the Fraction Tuff from a sample collected at Trailer Pass (table 5, sample 16). This age is compatible with that deduced by Spurr (1905) for the volcanic rocks that include the Fraction Breccia at Tonopah.

TUFFACEOUS CONGLOMERATE AND DEBRIS FLOWS

East of Cedar Pass and White Ridge in the Kawich Range, the Fraction Tuff is separated from older lavas and welded tuffs by a chaotic zone of very coarse clastic material, some of which is tuffaceous. In most areas the zone is thin, but just west of Cedar Pass it is as much as 400 feet thick. Because only the upper ash flows of the Fraction Tuff are known to overlie them, the clastic deposits are inferred to be interstratified with the Fraction Tuff (cross section *C–C′*, pl. 1). The deposits probably accumulated adjacent to the wall of the Cathedral Ridge caldera. The chaotic zone is composed of very crudely stratified beds of nontuffaceous debris flows, tuffaceous conglomerate, and massive tuffs rich in lithic fragments. Locally, the debris flows are composed of brecciated dacite or rhyodacite and elsewhere they are predominantly rhyolite. The tuffaceous conglomerates contain blocks of dacite or rhyolite as much as 100 feet long completely surrounded by tuff. Much of the tuff matrix is rich in crystals of plagioclase, quartz, alkali feldspar, hornblende, and biotite, an assemblage that is very similar to that of

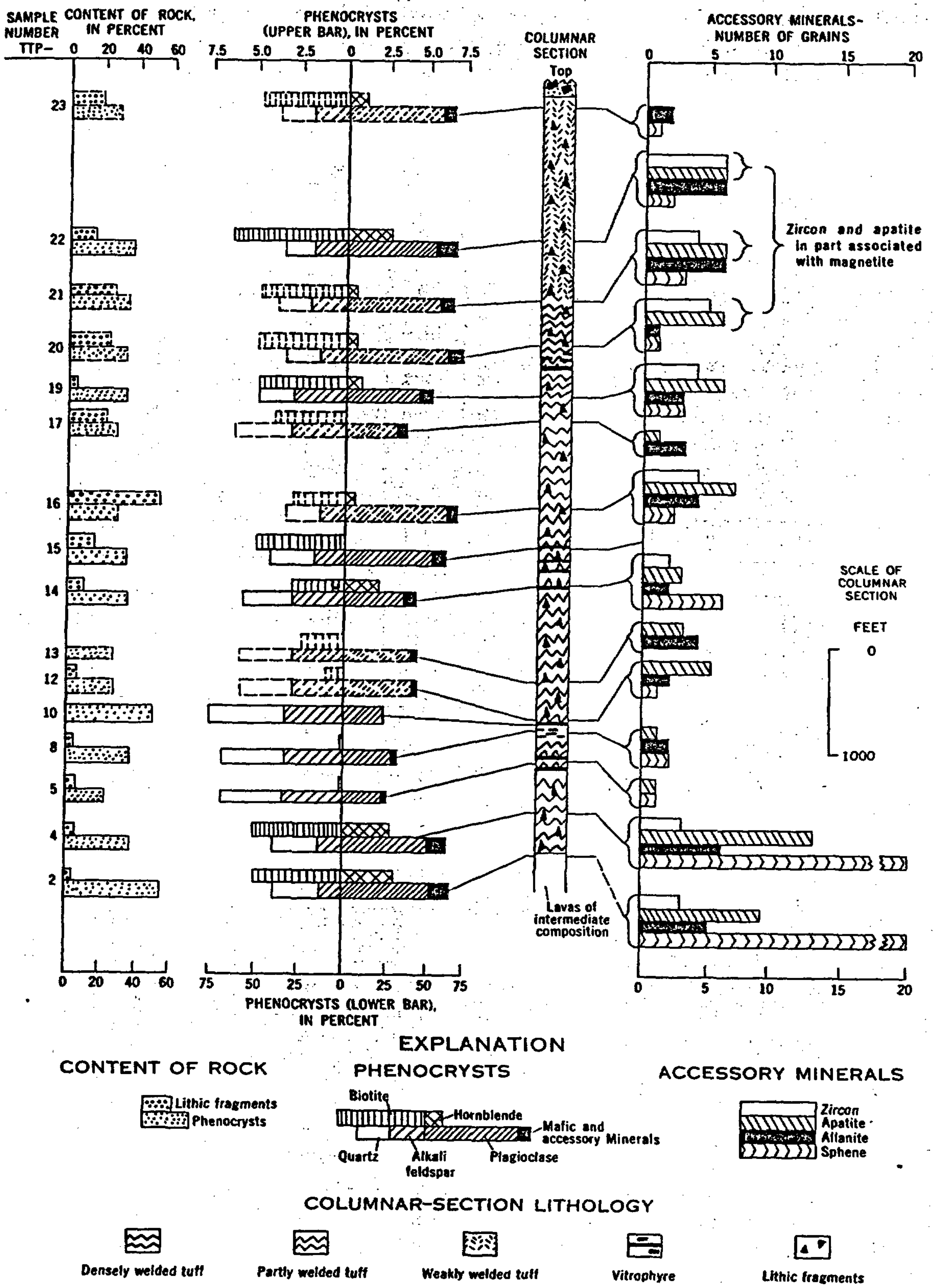

FIGURE 11.—Columnar section and thin-section petrography of Fraction Tuff near Trailer Pass in southern part of Kawich Range. Dashed bars and patterns indicate estimates. Petrography and compilation by F. M. Byers, Jr.

TABLE 11.—*Chemical analyses and norms of Fraction Tuff, rhyolite of O'Briens Knob, and andesite of Stonewall Flat*

[Analyses by P. L. D. Elmore, S. W. Botts, G. W. Chloe, Lowell Artis, and H. Smith, by rapid method (Shapiro and Brannock, 1962). Norms determined from chemical analyses recalculated to 100 percent minus H_2O+CO_2]

Unit	Fraction Tuff						Rhyolite of O'Briens Knob					Andesite of Stonewall Flat
Sample	1	2	3	4	5	6	7	8	9	10	11	12
Laboratory No.	160974	160975	160976	160720	160724	160955	160723	160977	160954	162053	160956	162056
Field No.	QM-3	QM-13	QM-17	BP-23	BP-33	CP-62	BP-32	QM-19	CS-24	CS-292	CP-8	CS-259
Chemical analyses												
SiO_2	73.8	70.2	71.1	75.1	72.7	73.7	68.7	70.2	72.2	70.2	73.6	53.0
Al_2O_3	12.6	14.8	14.5	12.7	12.7	13.9	15.0	15.6	13.4	14.1	13.5	17.4
Fe_2O_3	.51	1.8	1.5	.64	.85	1.2	1.3	1.5	.80	2.6	1.6	5.7
FeO	.18	.49	.37	.18	.10	.48	.97	.67	.37	.2	.21	2.6
MgO	.27	.55	.40	.31	.31	.40	.80	.61	.36	.71	.49	4.5
CaO	.66	2.1	2.2	.68	1.1	1.7	2.7	2.3	1.2	2.2	1.8	8.2
Na_2O	3.0	3.2	3.2	2.7	2.2	3.2	3.1	3.6	2.8	3.0	3.0	3.1
K_2O	4.6	4.4	4.2	5.3	5.6	4.4	3.5	4.0	4.9	3.9	4.5	2.0
H_2O-	.71	.45	.31	.43	2.2	.22	.71	.14	.26	1.2	.31	.75
H_2O+	3.3	1.1	1.8	2.0	2.0	.91	2.9	.86	3.1	1.4	.91	1.20
TiO_2	.11	.32	.28	.07	.08	.21	.29	.32	.16	.36	.24	1.0
P_2O_5	.00	.08	.07	.03	.02	.08	.13	.10	.05	.18	.08	.45
MnO	.06	.02	.05	.05	.03	.04	.06	.05	.05	.08	.04	.14
CO_2	<.05	<.05	<.05	<.05	<.05	<.05	<.05	<.05	<.05	.06	<.05	<.05
Sum	100	100	100	100	100	101	100	100	100	100	100	100
Norms												
Q	38.6	30.3	32.0	38.2	37.3	34.5	31.6	28.7	35.5	33.4	34.9	6.6
or	28.4	26.5	25.4	32.0	34.6	26.2	21.4	23.9	30.1	23.6	26.8	12.1
ab	26.5	27.6	27.7	23.4	19.4	27.3	27.2	30.8	24.6	26.0	25.6	26.7
an	3.4	10.1	10.7	3.3	5.6	8.0	13.0	10.9	5.8	10.0	8.5	28.2
C	1.6	1.2	.9	1.4	1.1	1.0	1.6	1.4	1.5	1.4	.6	------
wo												4.3
en	.7	1.4	1.0	.8	.8	1.0	2.1	1.5	.9	1.8	1.2	11.4
fs							.4					
mt	.5	.7	.6	.6	.2	1.1	2.0	1.4	.9	------	.1	6.5
hm	.2	1.3	1.2	.3	.8	.5	------	.5	.2	2.7	1.5	1.6
il	.2	.6	.5	.1	.2	.4	.5	.6	.3	.6	.5	1.9
ap		.2	.2	.1	.1	.2	.3	.2	.1	.4	.2	1.1

Sample, locality, and description

[Numbers following mineral names are percent of total phenocrysts]

1. About 350 ft above base of unit and directly above the lower plagioclase-rich zone at the reference area at Trailer Pass; dark-gray to black vitrophyric ash-flow tuff; stratigraphic position and mineralogy is about equivalent to that of specimen 6 in figure 12.
2. From dipslope on northeast flank of Cathedral Ridge. Densely welded vitric plagioclase-rich biotite-hornblende ash-flow tuff; stratigraphic position and mineralogy is about equivalent to that of specimen 22 in figure 12.
3. From base of unit 2.3 miles east of Wild Horse Ranch, Kawich Range. Gray vitrophyric crystal-rich plagioclase-rich biotite-hornblende ash-flow tuff; stratigraphic position and mineralogy is about equivalent to that of specimen 4 in figure 12.
4. From east flank of Belted Range at long 116°3′ W., lat 37°29.5′ N. Gray vitric weakly welded pumice-rich quartz-rich ash-flow tuff; ratio of quartz:alkali feldspar:plagioclase is approximately 5:3:2.
5. From east flank of Belted Range 2 miles northeast of Belted Peak. Light-tan partly devitrified welded ash-flow tuff; mineralogy is same as sample 4; some sodium probably leached by weak hydrothermal alteration.
6. From 2 miles southwest of Cactus Peak, Cactus Range. Partly devitrified vitrophyre, contains 32 percent phenocrysts consisting of plagioclase 35, alkali feldspar 29, quartz 28, biotite 6, and accessory hornblende, magnetite, allanite, sphene, and zircon.
7. From 2 miles north-northwest of Belted Peak, Belted Range. Gray vitrophyric rhyodacite flow containing 34 percent phenocrysts consisting of strongly zoned calcic andesine (?) 60, quartz 27, biotite 9, hornblende 3, and accessory magnetite and allanite.
8. From 3.5 miles east-southeast of Wild Horse Ranch, Kawich Range. Gray vitrophyric crystal-rich intrusive rhyolite, mineralogy is similar to sample 4.
9. From west flank of O'Briens Knob, southern Cactus Range. Dark-gray intrusive vitrophyric rhyolite.
10. From ridge 1 mile west of Sulfide Well, southern Cactus Range. Rock is a reddish-gray autobrecciated quartz latite intrusive containing 20 percent phenocrysts of andesine 69, biotite 19, quartz 3, magnetite 7, and minor zircon and apatite. Mineral assemblage is not typical of unit.
11. From 2 miles southwest of Cactus Peak, Cactus Range. Vitrophyric rhyolite dike containing 33 percent phenocrysts consisting of andesine 46, alkali feldspar 20, quartz 24, biotite 7, and accessory magnetite, hornblende, sphene, allanite, and apatite.
12. From 6.5 miles southwest of Cactus Peak, Cactus Range. Black, dense andesite dike rock, contains about 28 percent phenocrysts as much as 2 mm in diameter consisting of partly corroded sodic labradorite 63, augite 20 (commonly glomeroporphyritic), hornblende rimmed with iron oxide 8, iddingsite 5, and magnetite 3, in a hyalopilitic groundmass.

the Fraction Tuff. Locally, the tuffaceous conglomerate and tuffs rich in lithic fragments contain cobbles and blocks of welded tuff, some of which appear to be Fraction Tuff. The occurrence of these cobbles and blocks substantiates the inference that these deposits are interstratified with the Fraction.

North of Gold Reed the nonwelded upper part of the Fraction is overlain with apparent low-angle unconformity by a debris flow, tuffaceous conglomerates, and tuffs that together are at least 2,000 feet thick and possible as much as 3,000 feet. These rocks are interpreted as post-Fraction caldera-fill deposits. The lowest deposit consists of 50–300 feet of nearly monolithologic debris flow composed of boulders and smaller fragments of porphyritic dacite. The lack of a basal vitrophyre and the lack of baking in the underlying soft vitric tuffs indicate that the rock was cold when deposited and that it may have originated as a landslide. Locally, the flow contains a sandy tuff matrix and large fragments of Fraction Tuff, and in many places the matrix is moderately well cemented.

The debris flow is overlain by a thick sequence of conglomerate, a bed of limestone about 20 feet thick, ash-fall tuff, and minor ash-flow tuff. The strata consist of about 60 percent conglomerate and 40 percent interbedded tuff. A broad pediment surface bevels the area of outcrop, and pediment gravels and alluvium blanket the bedrock except where recent streams have cut into the pediment surface.

The conglomerate consists of large blocks, boulders, and smaller fragments of Fraction Tuff and dacite lavas in a tuffaceous sand matrix. The rock is weakly cemented and nonresistant, and it is not easily distinguished from the overlying unconsolidated pediment gravels. Crossbedding is visible locally, especially where the rock is moderately indurated. In places near the base the conglomerate contains abundant fragments of silicified wood, and locally logs as much as 2 feet in diameter and 100 feet in length may be found.

The interbedded tuffs are white, vitric, and rich in biotite, quartz, and plagioclase. Ash-flow tuffs occur near the middle of the exposed sequence and carry abundant pyroxene in addition to the minerals listed above. With the exception of a densely welded tuff at Gold Reed (not mapped separately), these tuffs are nonwelded to weakly welded. Some of the tuff was quarried for building stone in the early 1900's; the Kawich Post Office at Gold Reed was built with this stone. The densely welded tuff at Gold Reed is medium gray to dark brownish gray and vitrophyric, and contains 40–50 percent phenocrysts of plagioclase, quartz, biotite, and clinopyroxene. No alkali feldspar was observed.

The thick largely sedimentary sequence, which includes boulders of Fraction Tuff as much as 30 feet long, was unquestionably derived from a large nearby block that came into high relief following the formation of the Cathedral Ridge caldera. This block must have existed east and northeast of the area that received the deposits and has since been either downfaulted by basin-range faults or eroded and covered by younger deposits.

SEDIMENTARY ROCKS AND BEDDED TUFF

Exclusive of the tuffaceous conglomerate and debris flows in the Kawich Range just described, a large variety of sedimentary rocks and bedded tuff accumulated in adjacent areas after the extrusion of the Fraction Tuff and preceding the eruption of the rhyolite of O'Briens Knob. The evidence is good that this interval represents a long period of erosion and relative paucity of volcanic activity. The rocks were mapped as a single cartographic unit, although they were not deposited contemporaneously in all areas.

BELTED RANGE

Bedded strata above the Fraction Tuff in the Belted Range include thin beds of crossbedded conglomeratic sandstone at base, ash-fall tuff in the lower half, and interbedded sandstone and ash-fall tuff in the upper half. The strata are at least several hundred feet thick where exposed in Monotony Valley near Juniper Pass.

The tuff in the lower half is poorly bedded and, in places, probably includes nonwelded ash-flow tuff. The rock is light gray to white and rich in small pumice lapilli, and it contains quartz, biotite, and feldspar. The rock is vitric and unaltered except for local occurrences of abundant siliceous concretions as much as 2 inches in diameter. This rock is overlain by well-bedded ash-fall tuff and tuffaceous sandstone; the bedded tuff is generally poorer in crystals than the underlying tuff and is nearly void of biotite and other mafic minerals. The beds of sandstone are white to light gray, cross-stratified, rich in quartz, and commonly conglomeratic.

The interval just described is overlain by the Grouse Canyon Member of the Belted Range Tuff, a very distinctive and easily mapped ash-flow tuff. Bedded strata of the Belted Range Tuff probably underlie the Grouse Canyon Member. The lowest beds of the sedimentary rocks and bedded tuffs, however, may be nearly as old as the Fraction Tuff. Thus, a range in age of several million years is inferred for the beds in this interval in the Belted Range area.

CACTUS RANGE

Thick- to thin-bedded fine-grained to conglomeratic tuffaceous sedimentary rocks crop out on the north and east flanks of the Cactus Range and are downfaulted against the older volcanic and intrusive rocks that make up the central core of the range. These strata dip 10°–25° away from the range; they rest unconformably on rocks inferred to be tuffs of Antelope Springs, and they are interstratified with basalt lavas that are not mapped separately (pl. 1). The beds are overlain by the rhyolite of Cactus Peak (p. 57) and locally contain abundant inclusions of that rhyolite; these inclusions indicate that the deposition of the sediments coincided in part with the intrusion and extrusion of post-Fraction Tuff rhyolites. The sedimentary rocks are white, light gray, and pale yellow; they are locally vitric but mostly highly zeolitized. Similar tuffaceous sedimentary rocks and ash-fall(?) tuff are exposed in gullies west of Wellington Hills. These strata, which are mostly white and vitric, dip toward the range 15°–20°. Their age is known only as pre-Thirsty Canyon (p. 65) but they are inferred to be post-Fraction Tuff in age.

MOUNT HELEN

The sedimentary rock near Mount Helen consists of cross-stratified fluvial conglomerate and fine- to coarse-grained sandstone composed predominantly of poorly sorted volcanic detritus. Finely laminated siltstone and shale and thick- to thin-bedded ostracode-bearing limestone, all of lacustrine origin, are also common at Mount Helen. Thick beds of ash-fall tuff and three cooling units of the tuff of Tolicha Peak (p. 59) are interbedded with these strata. With the exception of medium- to dark-gray limestone, the rocks are yellowish brown to buff, pink, and white. Nearly everywhere they have been modified by hydrothermal or ground-water alteration, and locally they are intensely silicified.

Fossil wood fragments, ranging in size from less than an inch to a few large trunks 2 feet in diameter and as much as 100 feet long, are fairly abundant in the section, but they are so completely silicified that they can be identified only as conifer (R. A. Scott, oral commun., 1963).

Ostracode samples were examined by I. G. Sohn, who reported (written commun., 1964):

> The preservation is too poor for study of shell morphology, but one group, represented by steinkerns, has a diagnostic node on the surface of the valves. These were compared with described species of *Tuberocypris* Swain, 1947, described from the Salt Lake Formation of Utah that have a similar node. * * * The specimens on hand differ from the described species of *Tuberocypris*. The Russian workers have described a similar form as *Herpetocyprella* Daday, 1909, that ranges from the Miocene, Pliocene, and Holocene. The specimens on hand do not resemble any of the Russian species.

Sohn concluded that the ostracodes are probably of Pliocene age.

Vertebrate bone fragments collected from sandstone were examined by Edward Lewis (written commun., 1963), who reported that all 51 fragments are seemingly from a single individual camelid artiodactyl, but 32 are morphologically indeterminate. Morphologically determinate fragments are:

one carpal—right pyramidal
one tarsal—left cuboid
one tarsal—left ecocuneiform
four of a right metacarpal
four of the shaft of a metapedial
six of proximal phalanges of a manus
two of a proximal phalanx of a pes

Lewis concluded:

> Camelids of this size occur commonly in lower Miocene (Arikaree) to upper Pleistocene (Wisconsin) rocks of North America. These nondiagnostic fragments most closely approach the morphology and size of camelids such as those of the genus *Procamelus* of the upper Miocene and lower Pliocene.

Fishes from calcareous siltstone were identified as killifishes of the family Cyprinodontidae by D. M. Dunkle (written commun., 1963). Dunkle pointed out that the fossils were too poorly preserved for precise determination but that they can be tentatively referred to the genus *Fundulus*. According to Dunkle, fish of this type are known from several localities in California and Nevada and are considered by Miller (1945) to be either Pliocene or Pleistocene in age.

The age of the strata at Mount Helen is not clearly indicated by the above paleontological data. The strata are interbedded with the tuff of Tolicha Peak for which a late Miocene age is inferred on the basis of potassium-argon dates of higher and lower strata. (See later pages.) The strata are here considered to be of late Miocene age.

RHYOLITE OF O'BRIENS KNOB

After the deposition of a large variety of sedimentary rocks and ash-fall tuff in the project area, silicic lavas ranging in composition from quartz latite to rhyolite were erupted from many widespread vents. These rocks are generally rich in crystals and have a distinctive phenocryst assemblage that enables correlation from one range to another. Hence, they are mapped herein as a single unit named informally for exposures at O'Briens Knob in the Cactus Range. Rhyolite and quartz latite that are very similar to the rhyolite of O'Briens Knob occur in ranges east, north, and northwest of the project area, and stratigraphic relations in these areas indicate that they are virtually the same age as the rhyolite of O'Briens Knob. These rocks include rocks equivalent to the Oddie Rhyolite and Brougher Dacite (quartz latite) mapped by Spurr (1905) at Tonopah and the Oddie Rhyolite mapped by Ferguson and Cathcart (1954) in the western Manhattan district 42 miles northeast of Tonopah.

Characteristically the rhyolite of O'Briens Knob is light to medium gray or reddish gray where devitrified and dark gray where vitrophyric. It weathers to form reddish-gray or reddish-brown rounded hills and slopes, but locally it forms rugged craggy outcrops, especially in areas where the rock occurs as flow breccia or in intrusive masses. The rhyolite is generally conspicuously flow layered with the layers averaging several inches in thickness, but locally it is massive without visible layering. The rock at the borders of intrusive masses is commonly vitrophyric and autobrecciated, containing abundant shattered and broken phenocrysts.

Phenocryst contents range from 5 to 45 percent, with the higher values greatly predominating. Crystal-rich varieties are distinguished by abundant conspicuous plagioclase phenocrysts (mostly andesine as much as 5 mm in diameter) and conspicuous quartz and biotite. Plagioclase, the dominant phenocryst, typically makes up 50–60 percent of the phenocrysts. The plagioclase generally shows oscillatory and strong normal zoning and complex twinning, and commonly it occurs as glomeroporphyritic clots. Other phenocrysts include grains of quartz (25–30 percent), sanidine (0–25 percent), biotite (5–10 percent), and minor hornblende, sphene, magnetite, and allanite. This mineral assemblage is also characteristic of the tuff of Wilsons Camp and Fraction Tuff.

Chemical analyses and norms of the rhyolite of O'Briens Knob from each of the ranges are given in table 11 (samples 7–11). These rocks fall in the rhyolite and quartz latite fields of O'Connor (fig. 12) with the exception of sample 7, a rock from the Belted Range which falls in the rhyodacite field close to the boundary of quartz latite.

The rhyolite of O'Briens Knob and the Fraction Tuff are very similar chemically, petrographically, and in the range of variation among the normative minerals. The only apparent difference between the two rocks,

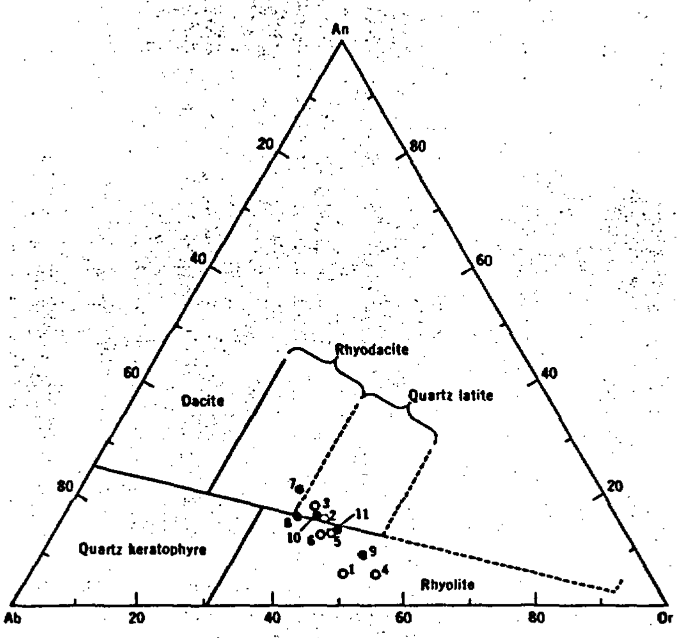

FIGURE 12.—Plot of normative albite-anorthite-orthoclase ratios for Fraction Tuff (open circles) and rhyolite of O'Briens Knob (solid circles). Numbers refer to samples in table 11. Fields from O'Connor (1965).

aside from one being a tuff and the other a lava, is the high lithic content characteristic of the tuff. These rocks possibly are genetically related. The fact that the rhyolite of O'Briens Knob occurs throughout such a broad area indicates that it, like the lavas of intermediate composition previously described, was withdrawn from a substratum that was broader than the mapped area.

STRATIGRAPHIC RELATIONS AND AGE

In the Belted Range, the lavas rest unconformably on or intrude faulted eroded blocks of tuff of White Blotch Spring, lavas of intermediate composition, Fraction Tuff, and, locally, sedimentary rocks and bedded tuff. In the Kawich and Cactus Ranges, the rhyolite of O'Briens Knob occurs mostly as broad dikes and plugs, some of which intrude the Fraction Tuff. In the southern Belted Range, the rhyolite is overlain by the rhyolites of Belted Peak and Ocher Ridge, which in turn are overlain by the Belted Range Tuff dated at 13.8 m.y. (table 5). It seems reasonable to conclude that the rhyolite of O'Briens Knob is late Miocene in age.

In the west half of the mapped area, especially along the margins of Cactus and Gold Flats, rhyolites that are not typically of O'Briens Knob lithology were included in the same cartographic unit to keep rhyolite units at a minimum. Most of these rhyolites contain phenocrysts of plagioclase, and stratigraphic relations indicate that they are closely related in time to the rhyolite of O'Briens Knob.

RHYOLITE OF CACTUS PEAK

Several plugs, broad dikes, and flows of white to gray crystal-poor plagioclase-free rhyolite occur in the northern part of the Cactus Range. Cactus Peak, one of the most prominent landmarks in the area, is composed of this rhyolite. At several localities the rhyolite intrudes rocks mapped as rhyolite of O'Briens Knob. The rhyolite of Cactus Peak is massive to intensely flow layered and locally columnar jointed; it weathers to reddish gray where unaltered and light gray to pale yellow where silicified. It contains 1–10 percent phenocrysts of alkali feldspar, small resorbed quartz, and sparse biotite in a microcrystalline granular, felty, or spherulitic groundmass that generally contains some secondary zeolite and sericite. It is vitrophyric locally at the base of flows or at the border of intrusive masses. A thick pile of similar rhyolite crops out east of the Goldfield mining district and extends into the extreme northwestern part of the project area (pl. 1). Although the age equivalence is not established, these rocks are mapped with the rhyolite of Cactus Peak.

RHYOLITE OF BELTED PEAK AND OCHER RIDGE

Thick rhyolite lavas and several large rhyolite intrusive masses form the bulk of the southern Belted Range. Aside from the relatively young rocks that are genetically related to the sodic rhyolitic Belted Range Tuff, the rhyolites are calc-alkalic in composition and occur as two types. One type generally contains abundant phenocrysts (rhyolite of Belted Peak); the other contains few phenocrysts (rhyolite of Ocher Ridge). The rhyolite of Belted Peak characteristically contains moderately abundant and conspicuous flakes of biotite and abundant large phenocrysts of quartz. In contrast, no mafic minerals of any kind have been observed in the rhyolite of Ocher Ridge either in outcrop or in five thin sections taken from widely separated outcrops. Quartz phenocrysts are small. Both rocks contain alkali feldspar and minor oligoclase.

The rhyolites of Belted Peak and Ocher Ridge appear to have been extruded nearly simultaneously from several feeder necks and plugs that are alined north-south along the west flank of the Belted Range. Near Wheelbarrow Peak the rhyolite of Ocher Ridge is younger; to the southwest, however, rhyolite of Ocher Ridge is interbedded with rhyolite of Belted Peak. Several of the plugs are well exposed, and in these the discordant relations with country rock are visible (fig. 13). The rocks are definitely younger than the rhyolite of O'Briens Knob and older than the Belted Range Tuff.

RHYOLITE AND COARSE PORPHYRITIC RHYOLITE OF UNCERTAIN AGE

Rhyolite lavas and rhyolite porphyry intrusive masses that crop out (1) on the east flank of Stonewall Mountain, (2) near Mount Helen, (3) in the Kawich Range, and (4) in the Belted Range and northern Kawich Valley can be dated only as Tertiary in age. These rocks are included in a single cartographic unit.

On the east flank of Stonewall Mountain, the lavas are about 1,300 feet thick and are in fault contact with the Thirsty Canyon Tuff. They are light gray to pink gray and conspicuously flow layered. Most are moderately rich in phenocrysts of quartz, sanidine, plagioclase, biotite, and magnetite. Some of the sanidine crystals are as much as 10 mm long. The groundmass is glassy to finely crystalline.

Near Mount Helen the rhyolite is so intensely silicified that nothing is known of its original petrography except that the rock was probably poor in phenocrysts.

In the Kawich Range, rhyolite exposed in small patches north of Quartzite Mountain and as large masses north of Cedar Pass is petrographically dissimilar to the rhyolite at Stonewall Mountain. These rocks are very poor in crystals. They contain sparse

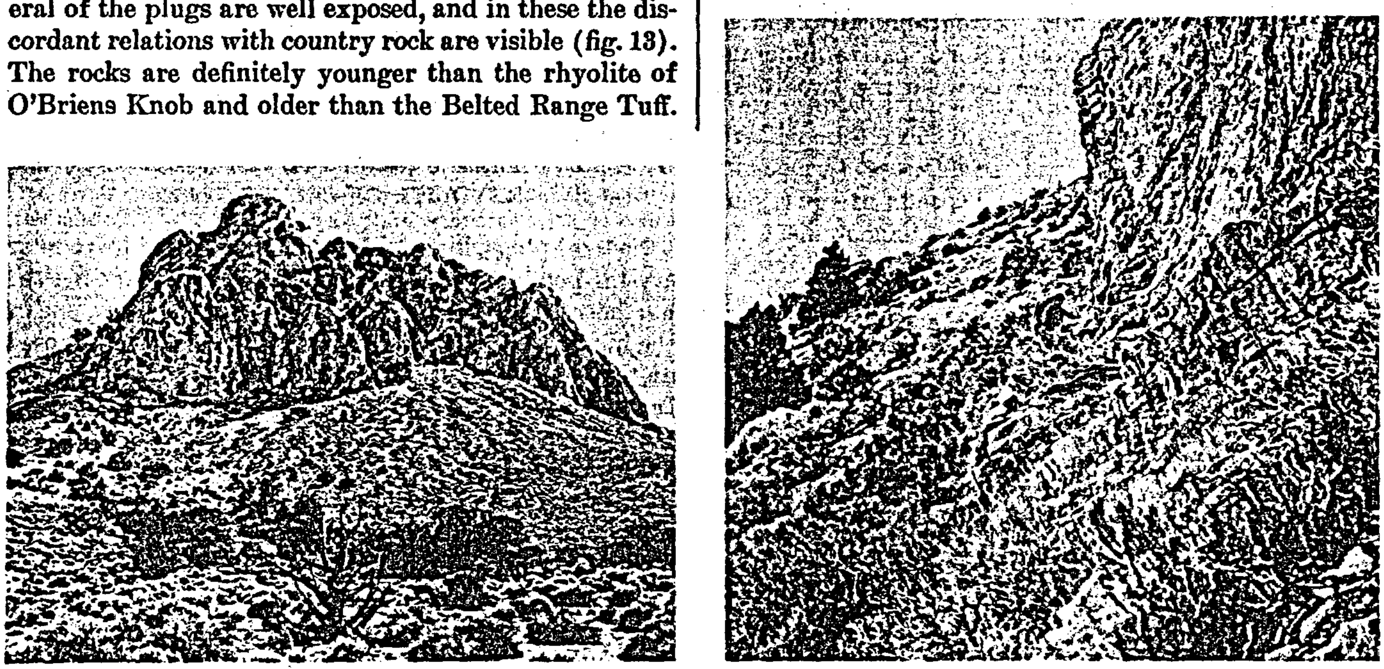

FIGURE 13.—Feeder plug about 2½ miles south-southwest of Belted Peak. Rock in the plug is gradational between rhyolite of Belted Peak and rhyolite of Ocher Ridge. It intrudes gently dipping Monotony Tuff and Shingle Pass Tuff. *A* (left), View to the east. *B* (right), View to the north along discordant contact with Shingle Pass Tuff on west side. Note that flow layering in the rhyolite parallels contact. (Man is standing in contact zone.) Rock in front of tree on left is nearly flat-lying altered tuff.

phenocrysts of alkali feldspar, quartz, and biotite. They are similar to the rhyolite of Cactus Peak but are not mapped with that rhyolite because of the uncertainty in age equivalence and because of the great distance separating the exposures.

A rhyolite porphyry plug located in the northern part of Kawich Valley and a rhyolite dike located southeast of Belted Peak intrude strata of Paleozoic age and the Monotony Tuff, respectively. The rock in Kawich Valley is massive weathering with only weak flow layering and is rich in crystals including alkali feldspar as much as 10 mm in length and quartz as much as 8 mm in diameter. Chloritized biotite is the only mafic mineral. The interior of the mass has a much coarser grained groundmass than the outer margins. The dike rock southeast of Belted Peak is petrographically similar, containing alkali feldspar crystals as much as 10 mm in length and abundant crystals of quartz and small crystals of biotite.

ANDESITE OF STONEWALL FLAT

Andesite dikes and plugs intrude Fraction Tuff along the northwest margin of Stonewall Flat. The rock is dark gray to black and produces a distinctive black pattern on aerial photographs. The rock weathers to dark-brown slopes composed of angular joint-faceted blocks. It is massively to very faintly flow layered and contains about 30 percent phenocrysts consisting of plagioclase, pyroxene, hornblende, magnetite, and altered olivine. A chemical analysis is given in table 11 (sample 12). The andesite predates Thirsty Canyon Tuff, but its age relative to other post-Fraction Tuff strata is not known.

TUFF OF TOLICHA PEAK

The tuff of Tolicha Peak is well exposed at Tolicha Peak, which is outside the project area about 2 miles west of Quartz Mountain. The best exposures within the area are 12 miles north-northeast of Tolicha Peak at triangulation station Pahute located on a large knoll or hill about half a mile east of Road D and 6 miles south-southwest of Mount Helen in T. 6 S., R. 46 E., Nye County, Nev. The rock is discontinuously exposed over a considerable area in this vicinity and adjacent to Mount Helen, and it has been correlated as far eastward as the southern Belted Range. It has been tentatively identified in drill cores from wells on Pahute Mesa. Its northern extent is unknown because vast areas to the north that were probably topographically low during the tuff eruptions are covered by Thirsty Canyon Tuff and alluvium. The Cactus Range was probably a topographic high during Tolicha Peak time, and the absence of the tuff there does not preclude the possibility that it originally extended that far north. The extent of the tuff west and southwest of the mapped area is unknown.

The tuff is about 300 feet thick at the triangulation station, where it forms a compound cooling unit with a conspicuous vitrophyre zone 30–50 feet thick at the base. The rock is densely welded there and weathers to hard clinkstone. It is gray, buff, and reddish brown where fresh and unaltered, and yellow or buff where zeolitized and silicified. In most exposures the tuff forms valleys or low hills. It contains less than 1 percent phenocrysts consisting of plagioclase, quartz, alkali feldspar, and sparse crystals of biotite and hornblende. The rock is composed dominantly of shards and pumice fragments that average less than half an inch in length and diameter. The pumice fragments commonly weather to elongated pits in exposures north of the triangulation station, especially where the tuff has been silicified. The pits together with incipient desert varnish, give the altered rock a distinctive rough and dingy aspect. The rock contains many small spherulites that average 1 mm or less in diameter. The spherulites are visible even where the enclosing matrix is highly zeolitized or silicified; they are a useful guide for recognizing the rock.

North of the triangulation station, at the south end of Mount Helen, three cooling units are present in the tuff of Tolicha Peak. These are separated from each other by fluvial and lacustrine sedimentary rocks and ash-fall tuff. The lowest of the three units crops out adjacent to the mountain, where it rests directly on eroded lavas of Mount Helen or on thin-bedded fluvial sandstone. This unit is massive weathering, contains abundant fragments of lavas of Mount Helen at the base, and is moderately rich in phenocrysts. It grades upward to crystal-poor tuff identical with that at the triangulation station. The tuff is about 200 feet thick and is overlain by 500–800 feet of crystal-poor ash-fall tuff, thin-bedded siltstone and limestone, and fluvial crossbedded sandstone. The middle cooling unit, which is thought to correspondent to the unit at the triangulation station, contains abundant rhyolite lithic fragments and has a thick altered vitrophyric zone at the base. It is overlain by ash-fall tuff and sedimentary rocks that include zones of sandy limestone as much as 15 feet thick and several zones of thinly laminated siltstone and shale. The bedded interval is at least 200 feet thick and possibly as much as 400 feet. It is overlain by weakly welded ashflow tuff that is nearly identical with the tuff of Tolicha Peak below but is different in that it contains virtually no lithic fragments. It is the youngest bedrock in the syncline valley in the SW. cor. T. 5 S., R. 47 E. All three units are zeolitized in the Mount Helen exposures, and locally they are intensely silicified. Because the three units are so strikingly similar in outcrop, no attempt was made to differentiate them. In fact, in many areas the tuff of Tolicha Peak includes not only the three units but the interbedded ash-fall tuffs and sedimentary rocks as well.

A single chemical analysis (table 12) of vitrophyre from the triangulation station shows the rock to be a salic rhyolite extremely low in femic constituents.

TABLE 12.—*Chemical analysis and norm of tuff of Tolicha Peak from 12 miles north-northeast of Tolicha Peak*

[Analysis of sample 1, laboratory No. 165058, field No. E 65-4-19-4, by P. L. D. Elmore, S. W. Botts, G. W. Chloe, Lowell Artis, and H. Smith, by rapid method]

Chemical analysis		*Norm*	
SiO_2	73.6	Q	40.8
Al_2O_3	12.2	C	1.8
Fe_2O_3	.48	or	26.2
FeO	.28	ab	26.8
MgO	.16	an	3.0
CaO	.57	en	.4
Na_2O	3	fs	.07
K_2O	4.2	mt	.74
H_2O-	.66	il	.22
H_2O+	3.9		
TiO_2	.11		
P_2O_5	0		
MnO	.07		
CO_2	.05		
Sum	99		

The source for the tuff of Tolicha Peak is not known with absolute certainty. The fact that several cooling units of the tuff are intercalated with thick sequences of lacustrine sedimentary rocks in and adjacent to Mount Helen suggests that collapse occurred there concomitantly with the extrusion of the tuff. The caldera may be centered south of Mount Helen on a small gravity low (pl. 1).

The age of the tuff of Tolicha Peak, as indicated by available data, is probably late Miocene. It is younger than the rhyolite of Ocher Ridge, which underlies the tuff along the southeast flank of Kawich Valley, and is older than the Belted Range and Paintbrush Tuffs, which overlie the tuff of Tolicha Peak in the southern Belted Range and near Tolicha Peak. The Belted Range Tuff (Grouse Canyon Member) has been dated at 13.8±0.6 m.y. (table 5, sample 13).

RHYOLITE OF QUARTZ MOUNTAIN

Calc-alkalic rhyolite lavas with minor basal ash-fall tuffs locally are at least 400 feet thick in the general vicinity of Quartz Mountain in the extreme southwest part of the project area. The rhyolite overlies the tuff of Tolicha Peak and underlies the Grouse Canyon Member of the Belted Range Tuff. In outcrop the rock ranges from grayish white to gray and buff. It contains 20–25 percent phenocrysts, which consist of subequal amounts of quartz, sanidine, and sodic plagioclase, 1–2 percent biotite, and accessory iron ore and sphene. The groundmass is typically vitric, although locally it is devitrified. Locally the rhyolite has been silicified and argillized by hydrothermal solutions.

Chemical analyses of two samples of the rhyolite are given in table 13.

TABLE 13.—*Chemical analyses of rhyolite of Quartz Mountain*

[Analyses by P. L. D. Elmore, S.W. Botts, G.W. Chloe, Lowell Artis, and H. Smith, by rapid method (Shapiro and Brannock, 1962)]

Sample	TP-12	BM-20
SiO_2	71.3	71.1
Al_2O_3	13.0	14.2
Fe_2O_3	1.2	1.4
FeO	.37	.46
MgO	.35	.40
CaO	1.3	1.1
Na_2O	3.4	3.6
K_2O	3.8	4.8
H_2O-	.43	.26
H_2O+	4.9	2.3
TiO_2	.20	.24
P_2O_5	.05	.04
MnO	.05	.08
CO_2	.05	.05
Sum	100	100
Powder density g/cc	2.38	2.38

Sample, locality, and description

TP-12. Vitrophyre from lava exposed 3 miles west-northwest of Quartz Mountain.
BM-20. Vitric rhyolite lava from 1.5 miles south of Quartz Mountain.

YOUNGER VOLCANIC ROCKS

In the southern part of the mapped area, the older Tertiary volcanic rocks described herein are overlain by relatively undeformed and virtually unaltered volcanic rocks of late Miocene and Pliocene age. These rocks consist of extensive sheets of silicic and intermediate ash-flow tuff with intercalated ash-fall and reworked tuff, and of local silicic to mafic lavas.

Most of the younger volcanic rocks belong to one of five major genetic groups, which, from oldest to youngest, are: Belted Range Tuff and associated lavas and tuffs, Paintbrush Tuff, Timber Mountain Tuff, Thirsty Canyon Tuff and associated lavas, and basalt of Basalt Ridge. The rocks of a genetic group are closely related chemically, petrographically, and with respect to their source areas.

The rocks of these five genetic groups are most widespread, thickest, and best exposed in the southern part of and south of the mapped area, where they have been mapped at a scale of 1:24,000 under the U.S. Geological Survey's long-range geologic studies project of the Nevada Test Site. A voluminous body of data on these rocks has been accumulated in conjunction with this 1:24,000-scale mapping, and a series of reports, both published and in preparation, describes their stratigraphy, structure, and petrology. For this reason, only short summary descriptions of these rocks are given here.

In addition, several genetically unrelated units, mainly lavas, interfinger with the rocks belonging to the five major groups. These rocks will not be described elsewhere and thus are described here in somewhat greater detail.

BELTED RANGE TUFF AND ASSOCIATED LAVAS AND TUFFS

Rocks of the Belted Range Tuff are widespread in the southern part of the mapped area (pl. 1) and are extensive south of the mapped area. The formation is composed of the Tub Spring and the overlying Grouse Canyon Members (Sargent and others, 1965; Hinrichs and Orkild, 1961), both of which are compound cooling units of ash-flow tuff. Both members were erupted from a vent area in eastern Pahute Mesa (pl. 1), which is the site of a caldera complex 7–10 miles in diameter (P. P. Orkild and K. A. Sargent, written commun., 1968; Orkild and others, 1968; Noble and others, 1968). In the general vicinity of the caldera, the two members are overlain, underlain, and locally separated by areally restricted units of lava, ash flows, and nonwelded and welded ash-fall tuff, which, as shown by field relationships, chemical composition, and petrography, are genetically related to the Belted Range Tuff. These lavas and tuffs are here subdivided into four informal units: the rhyolite of Kawich Valley, the rhyolite of Quartet Dome, the trachyte of Saucer Mesa, and the rhyolite of Saucer Mesa.

The composition of the rocks of the Belted Range Tuff and the associated lavas and tuffs, except the trachyte and some of the rhyolite of Saucer Mesa, is comenditic (table 14). Most of the rocks are peralkalic and contain a molecular excess of alkalis over alumina, as shown both by chemical analyses and by the presence of sodic amphibole and pyroxene in the groundmass of rocks that have crystallized in a nonoxidizing environment. The rocks are also characterized by relatively high contents of such trace elements as zirconium, niobium, beryllium, gallium, and the rare earths.

TABLE 14.—*Average major-element chemical compositions of rocks of younger volcanic units*

[These compositions are averages for densely welded devitrified tuffs and dense devitrified lavas. Some of the compositions are significantly different from the inferred original compositions of the unit prior to crystallization and prolonged contact with ground water. Averages are based on approximately 250 rapid and standard rock analyses made in the U.S. Geological Survey laboratories in Washington, D.C., and Denver, Colo.]

	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO_2	48	74	69.5	66	72.5	66	73.5	69	60.5	68.5	64	63	68.5
Al_2O_3	16.5	13	11.5	16	14	15	12.5	14.5	17	15	16.5	17.5	15.5
Fe_2O_3	4.5	1.9	4	2.1	1.8	3	1.9	3	3	2.8	2.7	2.2	3
FeO	7	.1	1.2	1.3	.2	.5	.3	.3	2.6	.7	1.7	1	.2
MgO	5.2	.4	.3	.4	.2	.5	.2	.4	1.3	.25	.9	.9	.4
CaO	8.5	.6	.5	1.5	.9	1.	.7	.8	2.8	.55	2.2	2.2	.6
Na_2O	4	4.2	5.2	5.3	4.2	5.1	4.5	4.8	5.2	5.1	4.8	5.1	4.8
K_2O	1.8	4.8	4.8	5.8	5.2	5.3	4.8	5.4	4.8	5.5	5	5.5	5.3
H_2O+	.7	.4	.6	.6	.5	.6	.2	.5	.6	.4	.6	.7	.4
TiO_2	2.2	.12	.3	.45	.15	.35	.15	.35	.9	.35	.85	.7	.45
MnO	.19	.08	.16	.16	.08	.08	.08	.15	.15	.15	.13	.15	.15
P_2O_5	.9	.06	.07	.10	.04	.16	.03	.07	.5	.04	.38	.35	.1
CO_2	<.05	.5	.05	.3	.2	.3	.1	.1	.05	<.5	<.05	<.05	.05

	14	15	16	17	18	19	20	21	22	23	24	25
SiO_2	61.5	76.5	68.5	72	74	76	74.5	69.5	69.5	76	65.5	75
Al_2O_3	17.5	13	16	13	11.5	11	12.5	14.5	14	11.5	15.5	11.5
Fe_2O_3	3.3	.5	1.5	2.6	2.5	2.1	1.4	2.2	2.6	2.3	4	2.1
FeO	1.7	.2	.7	.5	1	.4	.9	1.3	.7	.2	.7	.4
MgO	.85	.15	.8	.4	.1	.35	.1	.1	.3	.1	.35	.2
CaO	3	.6	1.7	1	.3	.5	.3	.8	.5	.2	1.1	.4
Na_2O	5.1	3.6	4.5	4.4	4.6	4.3	4.6	5.2	4.6	4.2	5.5	4.2
K_2O	4.6	4.8	5.2	5	4.6	4.7	4.7	5.2	5.2	4.5	5.6	4.5
H_2O+	.5	.5	.5	.9	.4	.5	.4	.2	.6	.3	.3	.3
TiO_2	.85	.15	.5	.35	.22	.15	.15	.30	.30	.16	.6	.15
MnO	.12	.06	.08	.15	.17	.07	.09	.20	.17	.11	.23	.07
P_2O_5	.34	.05	.2	.10	.02	.02	.02	.03	.04	.02	.16	.01
CO_2	<.05	.1	.1	.5	.05	.1	<.05	.1	.1	.05	<.05	.3

1. Basalt of Basalt Ridge.
2. Labyrinth Canyon Member, Thirsty Canyon Tuff.
3. Gold Flat Member, Thirsty Canyon Tuff.
4. Trail Ridge Member, Thirsty Canyon Tuff.
5. Trail Ridge Member, Thirsty Canyon Tuff, crystal-poor ash flows.
6. Trail Ridge Member, Thirsty Canyon Tuff, lower crystal-rich ash flows.
7. Spearhead Member, Thirsty Canyon Tuff, comendite ash flows.
8. Spearhead Member, Thirsty Canyon Tuff, trachytic sodic rhyolitic ash flows.
9. Trachyte of Hidden Cliff.
10. Rhyolite of Pillar Spring, middle trachytic sodic rhyolite lavas.
11. Rhyolite of Pillar Spring, lower trachyte lavas.
12. Trachyte of Yellow Cleft, trachyte lava.
13. Rhyolite of Ribbon Cliff, upper trachytic sodic rhyolite lavas.
14. Rhyolite of Ribbon Cliff, lower trachyte lavas.
15. Timber Mountain Tuff, rhyolitic ash-flow tuffs.
16. Timber Mountain Tuff, quartz latitic ash-flow tuffs.
17. Grouse Canyon Member, Belted Range Tuff, upper crystal-rich ash-flow sheet.
18. Grouse Canyon Member, Belted Range Tuff, lower crystal-poor ash-flow sheet.
19. Tub Spring Member, Belted Range Tuff.
20. Rhyolite of Saucer Mesa.
21. Trachytic sodic rhyolite of Saucer Mesa.
22. Tuff of Basket Valley.
23. Rhyolite of Quartet Dome.
24. Trachyte of Saucer Mesa.
25. Rhyolite of Kawich Valley.

Sodium-rich sanidine is the dominant phenocryst mineral; sodium-and iron-rich clinopyroxene, fayalite, zircon, and apatite are ubiquitous phenocryst minerals, but form less than 0.1 percent of the rock. Quartz phenocrysts are abundant in the rhyolite of Quartet Dome, the Tub Spring Member, and the rhyolite of Kawich Valley, and less abundant in rocks of the Grouse Canyon Member. Crystals of sodic amphibole of vapor-phase origin are common, particularly in the ash-flow tuffs.

Rhyolite of Kawich Valley

Numerous discontinuous bodies of petrographically similar commendite lavas, which were erupted in the general vicinity of Kawich Valley and in the southern Belted Range prior to the eruption of the Tub Spring Member, are here included in the rhyolite of Kawich Valley. The form and areal distribution of these lavas show that they were erupted from numerous vents.

Rocks of the unit are typically flow layered and highly contorted. The lavas are crystallized in most outcrops although vitrophyres are present locally. Spherulitic devitrification is common and zeolitic alteration is fairly common. Colors range from gray, bluish gray, and green to various shades of red and yellow. Phenocrysts, consisting predominantly of sanidine and subordinate quartz with minor fayalite and clinopyroxene, make up about 1–10 percent of the rock. The rhyolite has been dated at 14.8±0.6 m.y. (table 5, sample 14).

Trachyte of Saucer Mesa

Peralkalic trachyte lavas crop out below the rim of Saucer Mesa on the southeast flank of Gold Flat and also in a strike valley southeast of Saucer Mesa. The rock is reddish gray or green and weathers to dark red or brown. Phenocrysts—which include sodium-rich sanidine-anorthoclase (10–30 percent), clinopyroxene (<5 percent), and sparse crystals of iron-rich olivine—are set in either a glassy groundmass or a trachytic groundmass of alkali feldspar and sodic amphibole.

Rhyolite of Quartet Dome

The rhyolite of Quartet Dome comprises several thick, small bodies of crystal-rich comendite lava which, as shown by their shape and structure, were emplaced at various times as relatively viscous bodies which erupted from a different volcanic vent. Some were emplaced before the eruption of the Tub Spring Member, others were emplaced after the eruption of the Tub Spring Member but before the eruption of the Grouse Canyon Member, and one was emplaced after the eruption of the Grouse Canyon Member. It is possible that some bodies of the rhyolite of Kawich Valley postdate some of the lavas of Quartet Dome in areas where stratigraphic relations are doubtful. When stratigraphic relations are straightforward, however, the rhyolite of Quartet Dome overlies the rhyolite of Kawich Valley.

Rocks of the unit are typically gray, crystallized, and well foliated on a large scale. Rocks of the various domes are petrographically nearly identical. Phenocrysts of sanidine and quartz, 1–4 mm in diameter, together compose 25–30 percent of the rock, and clinopyroxene and fayalite compose less than 1 percent. Vitrophyre and zones of spherulitic and zeolitic alteration are relatively uncommon.

In Grass Spring Canyon the rhyolite of Quartet Dome is underlain by a unit composed of poorly welded ash-flow or welded ash-fall tuff and nonwelded ash-fall tuff. This unit has the same phenocryst content as the overlying lavas, and locally it contains lithic fragments that are lithologically identical with those lavas. This tuff was probably erupted from the same vent immediately prior to the eruption of the rhyolite.

Tub Spring Member

The Tub Spring Member crops out on the east and south flanks of the Belted Range, east of Yucca Flat south of the project area, and locally in Kawich Valley and vicinity. It is as much as 300 feet thick.

Rocks of the member are typically buff or bluish gray, but locally they are brick red. The member is mostly devitrified, but at the base and top it contains poorly welded glassy tuff that is partly or completely zeolitized in many outcrops. Vitrophyre is locally present in the central part of the cooling unit, where the member is thick and densely welded. The member contains about 20–25 percent phenocrysts of sanidine and quartz, minor amounts of clinopyroxene and fayalite, and locally fragments of pumice, rhyolite, welded tuff, and Paleozoic sedimentary rocks.

Grouse Canyon Member

The Grouse Canyon Member has a much wider distribution than the Tub Spring Member. As inferred from its present known distribution, the Grouse Canyon Member originally covered at least 1,500 square miles.

The Grouse Canyon differs from the Tub Spring in being more densely welded and in having a much smaller percentage of phenocrysts. The various ash flows that make up the member contain about 0.01–25 percent phenocrysts, with an average of about 2–3 percent. In most places the member is densely welded (fig. 14) and, with the exception of a thin basal vitrophyre, is almost everywhere devitrified. The rocks are greenish gray to bluish gray to grayish buff and brown and locally are brick red.

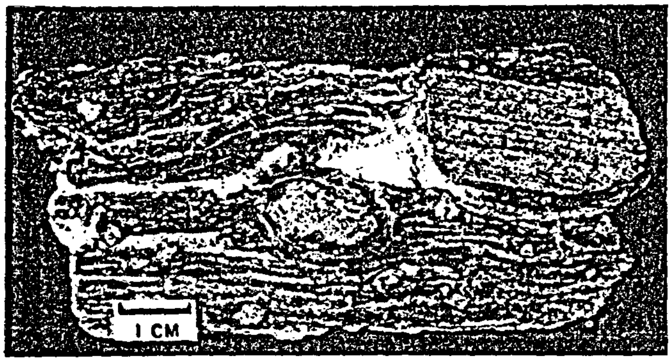

FIGURE 14.—Hand specimen of Grouse Canyon Member of Belted Range Tuff cut normal to prominent compaction foliation. Note that foliation parallels the boundaries of lithic inclusions of flow-banded rhyolite.

RHYOLITE OF SAUCER MESA

The rhyolite of Saucer Mesa consists of generally crystal-poor and well-flow-banded lavas that overlie the Grouse Canyon Member at Saucer Mesa and in Kawich Valley. The lavas and associated tuffs are locally as much as 1,000 feet thick.

The rocks range in color from bluish gray, greenish gray, and buff to reddish brown and red. Phenocrysts, consisting almost entirely of sodium-rich sanidine, compose about 1–20 percent of the lavas. The rocks of the upper part are almost entirely crystal poor, and they are bluish gray and greenish gray, whereas rocks of the lower part contain both crystal-poor and moderately crystal-rich flows. Although most of the rocks are comendites, trachytic soda rhyolite lavas (table 13) occur locally at the base of the unit.

The trachytic sodic rhyolite of Saucer Mesa and the tuff of Basket Valley, which are mapped separately on 1:24,000- and 1:62,500-scale maps, respectively, are included with the rhyolite of Saucer Mesa on plate 1.

ASH-FALL AND REWORKED TUFF

In most places the units of welded tuff and lava are underlain and separated by variable thicknesses of nonwelded ash-fall tuff, reworked tuff, and epiclastic tuffaceous material. Generally these rocks are bedded and partly or completely zeolitized. Crystal content is about 1–25 percent. Some tuffs, particularly those containing abundant crystals, contain numerous lithic fragments. Where too thin to be shown separately on the geologic map, ash-fall and reworked tuffs are, by convention, included with the overlying map unit.

Ash-fall and reworked tuff occurs stratigraphically above and below the Tub Spring Member (pl. 1). Above the Tub Spring the tuffs are almost wholly genetically related to the Belted Range Tuff and associated lavas, but in many places below the Tub Spring the tuffs appear to be related to older volcanic units.

RHYOLITE OF AREA 20, PAHUTE MESA

Calc-alkalic rhyolite lavas locally crop out beneath the Timber Mountain Tuff and between members of the Timber Mountain Tuff in the northern and southwestern parts of Pahute Mesa. The oldest, which is called the rhyolite of Area 20 by Orkild, Sargent, and Snyder (1969), is flow brecciated, vesiculated, and glassy at the top and base, and flow layered and devitrified in the interior. The rock contains phenocrysts of quartz, alkali feldspar, plagioclase, biotite, hornblende, and, locally, sphene. Drill-hole data on Pahute Mesa (P. P. Orkild, oral commun., 1966) show that this rhyolite, although it crops out directly below the Rainier Mesa Member of the Timber Mountain Tuff in Silent Canyon and tributaries, is actually older than the Paintbrush Tuff.

The youngest lava, which is called rhyolite lava of Timber Mountain Caldera Moat by Orkild, Sargent, and Snyder (1969), crops out in tributaries of Thirsty Canyon. The rock is light gray to purplish gray, massive to flow layered, dense to vesicular, and generally crystallized except for locally glassy tops and bases. It contains 5–10 percent phenocrysts—mainly quartz and alkali feldspar with minor plagioclase, biotite, and iron ore. In places the rock is quartz free and contains alkali feldspar and plagioclase in a ratio of about 2:1. Drill-hole data indicate that this rhyolite lies between the Ammonia Tanks Member and the Rainier Mesa Member of the Timber Mountain Tuff.

PAINTBRUSH TUFF

Two members of the Paintbrush Tuff (Orkild, 1965), the Stockade Wash Member and an unnamed unit of ash-fall and reworked tuff, are present in the southeastern part of the map area.

STOCKADE WASH MEMBER

The Stockade Wash Member is a simple cooling unit of buff poorly welded rhyolitic ash-flow tuff which locally is as much as 300 feet thick; in most places it is much thinner. Sparse phenocrysts include quartz, plagioclase, alkali feldspar, biotite, hornblende, and iron ore.

Ash-Fall and Reworked Tuff

The unit of ash-fall and reworked tuffs is approximately equivalent to the Survey Butte Member of former usage exposed on Rainier Mesa (Gibbons and others, 1963), but probably only the middle part of the map unit is genetically related to the Paintbrush. Tuffs immediately underlying the Rainier Mesa Member are undoubtedly genetically related to the Timber Mountain Tuff. Where the Stockade Wash Member is absent, ash-fall and reworked tuffs genetically related to the Belted Range Tuff and associated lavas and tuffs are locally mapped with the bedded tuffs of the Paintbrush Tuff. In the southern part of Kawich Valley a local unit of tuffaceous conglomerate, which in places is as much as 200 feet thick, is included in the unnamed unit.

LATITE OF SOUTH KAWICH VALLEY

Black latite lavas that closely resemble basalt in outcrop are here termed the latite of south Kawich Valley. The latite interfingers with the ash-fall and reworked tuffs in south Kawich Valley and at several localities north of Pahute Mesa on the southeast edge of Gold Flat. In the area north of Pahute Mesa the rock was emplaced as a series of thin flows generally no more than 30 feet thick, whereas in south Kawich Valley the unit is as much as 250 feet thick and covers an area of more than 1 square mile.

The rock is dark gray to black and sparsely porphyritic, containing a few phenocrysts of labradorite, clinopyroxene, olivine, and iron ore in a pilotaxitic groundmass. The plagioclase microlites in the groundmass, which appear to be oligoclase or andesine in composition, surround tiny grains of pyroxene, iron ore, and iddingsite. A few crystals of greenish- to dark-brown hornblende mantled with opaque iron oxide are visible in two thin sections from exposures near Gold Flat. The rocks contain an abnormally large amount of apatite, which occurs in the groundmass and as euhedral prisms as much as 0.3 mm long.

TIMBER MOUNTAIN TUFF

The Timber Mountain Tuff (Orkild, 1965) is composed of a sequence of silicic ash-flow tuffs that were erupted from sources in the immediate vicinity of Timber Mountain just south of the mapped area. The two most extensive members of the formation, the Rainier Mesa and the Ammonia Tanks, are present within the mapped area.

Devitrified rocks of both members range in color from gray and maroon to buff; glassy bases and tops of the cooling units are pink, buff, or dark gray. Degree of welding ranges from nonwelded and poorly welded to densely welded.

Within the mapped area the Rainier Mesa Member ranges in thickness from 0 to 600 feet, and the Ammonia Tanks Member, from 0 to 350 feet. Both members are thickest in the eastern part of Pahute Mesa. Numerous partial cooling breaks, reflected by alternating zones of dense and partial welding are visible in many places in thick sections of both the Rainier Mesa and Ammonia Tanks Members; no complete breaks have been observed.

Rocks of the Timber Mountain Tuff range in composition from rhyolite to quartz latite. In both members the ash flows that compose the lower parts are more silicic and less mafic than those that compose the upper parts. Average chemical composition for the rhyolite and quartz latitic phases is given in columns 15 and 16, table 14. Phenocrysts of quartz, sanidine, plagioclase, biotite, hornblende, and clinopyroxene compose 15–25 percent of both members. The Ammonia Tanks Member has an appreciably higher ratio of sanidine to total quartz and plagioclase than does the Rainier Mesa Member (F. M. Byers, Jr., oral commun., 1964).

In isolated outcrops it may be difficult to distinguish the two members. In many thin sections, however, the Ammonia Tanks Member may be recognized by the presence of abundant accessory sphene. In addition, in southeastern Pahute Mesa, the Ammonia Tanks Member may be distinguished by the presence of numerous fragments of red densely welded tuff of an older unit of the Timber Mountain Tuff, and by a thick and distinctive sequence of poorly welded pumice-rich ash-flow tuffs that makes up the lower and middle parts of the member in that area. The difference in the quartz: sanidine-plagioclase ratios and in the direction of remanent magnetization may also be used to identify rocks of the two members; studies by Gordon Bath (unpub. data, 1966) of the U.S. Geological Survey indicate that the Rainier Mesa Member is reversely polarized and the Ammonia Tanks Member is normally polarized.

BASALT OF STONEWALL MOUNTAIN AREA

Porphyritic basalt crops out between the Thirsty Canyon Tuff and the Rainier Mesa Member of the Timber Mountain Tuff in several localities in the southwestern part of the project area near Stonewall Mountain and at the south end of Coyote Cuesta. None of this basalt was examined in thin section or chemically analyzed. The rock contains small phenocrysts of labradorite and a few crystals of olivine largely altered to iddingsite. None of the basalt lavas are more than a few tens of feet thick. South of Stonewall Mountain, west of the project area, several basalt flows are separated by bedded tuff. The basalt and interbedded tuffs at Stonewall Mountain are several hundred feet thick (F. M. Byers, Jr., written commun., 1964).

YOUNGER ROCKS OF MOUNT HELEN

Local poorly exposed units of fanglomerate, alluvium, and basalt underlie the Thirsty Canyon Tuff in the general vicinity of Mount Helen. Although the Thirsty Canyon Tuff provides an upper limit on the age of these rocks, it is difficult to place definite lower limits on their age.

Alluvium

Weakly lithified alluvium, colluvium, and fanglomerate crop out beneath the Thirsty Canyon Tuff in several areas around the perimeter of Mount Helen, notably near Gold Crater. The unit is heterolithic; in places it consists of boulders, cobbles, and pebbles of Paleozoic rocks and fragments of volcanic rocks older than the Thirsty Canyon Tuff in a locally lithified matrix of sand or silt. Material that is clearly of alluvial-fan origin and that consists wholly of volcanic detritus crops out along the west and north flanks of Mount Helen. In the vicinity of Gold Crater, the alluvium apparently is dominantly the valley-fill type, although some may be of alluvial-fan origin. The alluvium is locally at least 50 feet thick, but it may be considerably thicker under Stonewall Playa, which appears to have been topographically low prior to the extrusion of the Thirsty Canyon Tuff.

Porphyritic Basalt

Basalt caps the south end of Mount Helen and the tops of several hills or buttes flanking Mount Helen. The lavas, which apparently were originally only a few feet thick, were erupted onto a fairly extensive erosional surface that sloped away from Mount Helen in all directions.

The rock is characterized by sparse to abundant phenocrysts of plagioclase that average about 1 cm in length but locally are as much as 10 cm. In thin section the groundmass is glassy to extremely dense. At the summit of Mount Helen, however, at or near the feeder vent for the basalt, the rock grades to porphyritic leucodiabase containing a few large plagioclase phenocrysts in a subophitic groundmass of calcic labradorite, 65 percent; olivine (fa_{35}), 11 percent; clinopyroxene, 15 percent; and iron ore, 3 percent. Less than 1 percent alkali feldspar is present interstitially and in the rims of plagioclase. The remainder of the rock is composed of 6 percent calcite, which occurs as vesicles and locally as an alteration product of plagioclase.

The presence of basalt boulders in alluvium beneath the Thirsty Canyon Tuff shows that the basalt is older than the Thirsty Canyon. The basalt itself rests on an erosional surface developed on a variety of older volcanic rocks. This surface was not extensively dissected until after Thirsty Canyon time, as shown by erosional remnants of Thirsty Canyon Tuff on or very near the same surface as the basalt. It seems, therefore, that the rock is not appreciably older than the Thirsty Canyon Tuff and is herein considered tentatively to be of early Pliocene age.

FANGLOMERATE OF TRAPPMAN HILLS

Weakly cemented fanglomerate composed dominantly of fragments of gneissic quartz monzonite and biotite schist forms gently rounded hills west and southwest of Trappman Hills. The fragments are about 1–12 inches in diameter, and in most exposures they contain less than 10 percent of Tertiary volcanic rocks. The percentage of volcanic fragments increases southward; in the southernmost outcrops volcanic material comprises about 50 percent of the fanglomerate.

Although the source of the metamorphic fragments is clearly the Trappman Hills, where the bedrock consists of identical rock, the age of the fanglomerate is not so simply deduced. Locally, the fanglomerate rests on tuff of Tolicha Peak; this fact suggests a Miocene or Pliocene age. However, the occurrence of boulders or fragments of gneissic quartz monzonite in rhyodacitic breccia (rocks of Mount Helen, pl. 1) suggests that the fanglomerate may have started to form prior to the deposition of the tuff of Tolicha Peak. The evidence here, however, is not straightforward inasmuch as the fragments could have been derived directly from Precambrian bedrock.

THIRSTY CANYON TUFF AND ASSOCIATED LAVAS

A complex sequence of genetically related tuffs and lavas had its source in the volcanic center of Black Mountain (Christiansen and Noble, 1965).

The ash-flow and closely associated ash-fall tuffs that erupted from the Black Mountain volcanic center are named the Thirsty Canyon Tuff (Noble and others, 1964). Five formal members are recognized within the Thirsty Canyon. From oldest to youngest, they are: the Rocket Wash, Spearhead, Trail Ridge, Gold Flat (Noble, 1965), and Labyrinth Canyon Members. On plate 1 the tuffs of the Rocket Wash, Spearhead, and Trail Ridge Members are combined. Each member consists primarily of ash-flow tuff that had completely cooled before the deposition of succeeding units of ash-fall or ash-flow tuff. A relatively thin unit of ash-fall tuff occurs at the base of the members in most outcrops.

The rock and cooling unit types of the members and their relative original volume are given in table 15, and average chemical compositions for the Spearhead, Trial Ridge, Gold Flat, and Labyrinth Canyon Members are given in table 14.

TABLE 15.—*Rock and cooling unit types and relative original volume of members of the Thirsty Canyon Tuff*

Member	Rock type	Cooling unit type	Relative original volume
Labyrinth Canyon.	Comendite	Single-flow (?) simple cooling unit.	1
Gold Flat	Pantellerite	Compound cooling unit.	2
Trail Ridge	Nonperalkaline rhyolite.	Multiple-flow simple cooling unit.	4
Spearhead	Comendite and trachytic sodic rhyolite.	Compound cooling unit.	5
Rocket Wash	Trachytic sodic rhyolite and comendite.	do	3(?)

With the exception of the Gold Flat Member, which is typically buff, ocher, or green, the rocks of the Thirsty Canyon Tuff are maroon, gray, buff, and locally pink. Phenocrysts of sodium-rich sanidine compose about 5–25 percent of the rock. Phenocrysts of fayalite and sodic clinopyroxene are ubiquitous, but each composes less than 0.1 percent of the rock. Sparse phenocrysts of an iron- and sodium-rich amphibole are present in the Spearhead, Gold Flat, and Labyrinth Canyon Members. Sparse phenocrysts of quartz and plagioclase are present in the Gold Flat Member, and flakes of biotite occur very sparsely in the Spearhead, Trail Ridge, and Gold Flat Members. Lithic fragments, mostly genetically related material derived from the Black Mountain volcano, and large blocks of pumice are common in many ash flows.

Trachyte and trachytic sodic rhyolite lavas underlie and interfinger with the Thirsty Canyon Tuff in the immediate vicinity of Black Mountain. These lavas are here divided into four informal units, which are, from oldest to youngest, the rhyolite of Ribbon Cliff, the trachyte of Yellow Cleft, the rhyolite of Pillar Spring, and the trachyte of Hidden Cliff. The rocks are typically gray or blue-gray holocrystalline rocks, which in many localities exhibit large-scale flow layering. Feldspar phenocrysts compose 10–30 percent of the lavas. The trachytic sodic rhyolites contain phenocrysts of sodium-rich sanidine and anorthoclase averaging almost a centimeter in diameter, whereas the trachytes contain somewhat smaller phenocrysts of plagioclase, thickly rimmed by anorthoclase or sodium-rich sanidine, in addition to phenocrysts of alkali feldspar. All the rocks contain phenocrysts of iron-rich olivine and clinopyroxene. Biotite phenocrysts are also sparsely present in a few of the lavas of Ribbon Cliff. The groundmass of the lavas consists principally of alkali feldspar, with minor amounts of quartz, iron ore, and aegirite or sodic amphibole. Average chemical compositions of the various units are given in table 14.

The rhyolite of Ribbon Cliff underlies the Rocket Wash and Spearhead Members east, south, and north of Black Mountain (pl. 1). The best exposures are at Ribbon Cliff, 5 miles east of the summit of Black Mountain, where the unit is more than 400 feet thick. Although most of the lava flows that make up the rhyolite of Ribbon Cliff are of trachytic sodic rhyolite composition (table 14), flows of trachyte are locally present at the base of the unit.

A collapse caldera 7 miles in diameter centered on the summit of Black Mountain formed after the deposition of the rhyolite of Ribbon Cliff and the extrusion of the Rocket Wash and Spearhead Members. After collapse, the trachyte of Yellow Cleft, a complex sequence of lavas, tuffs, breccias, and hypabyssal intrusive bodies of trachytic and trachytic sodic rhyolite composition, was erupted within the depression.

The rhyolite of Pillar Spring was erupted during the interval between the deposition of the Trail Ridge and Gold Flat Members. The locality is at an informally named spring (pl. 1), 3 miles south-southwest of the summit of Black Mountain. Three distinct rock types can be recognized in the formation: a basal sequence of trachyte flows, and middle and upper units of trachytic soda rhyolite. With the exception of the upper unit, which has overflowed the northwest edge of the post-Spearhead depression, the rhyolite of Pillar Spring is presently, and probably was originally, restricted to the area of post-Spearhead subsidence.

The trachyte of Hidden Cliff, which is named for Hidden Cliff, located 1 mile east of the summit of Black Mountain, was erupted between the deposition of the Gold Flat and Labyrinth Canyon Members. These lavas, erupted from a vent near the present summit of Black Mountain, probably originally did not extend much beyond their present distribution.

BASALT OF BASALT RIDGE

Basalts younger than the Thirsty Canyon Tuff crop out at Basalt Ridge, at several localities west of Gold Flat, and in Thirsty Canyon and vicinity. Although the various outcrops are similar in lithology, they may vary considerably in age. No isotopic dates are available for the basalt of Basalt Ridge, but the topographic expression and degree of erosion of the various outcrops make a Quaternary age improbable. The basalt is here assigned a Pliocene age.

The basalts are porphyritic in most exposures, but on Basalt Ridge they are conspicuously porphyritic with about a third of the rock containing phenocrysts of plagioclase and clinopyroxene as much as 3 cm in length. In most thin sections the rock is seriate with grains of sodic labradorite, clinopyroxene, olivine (fa_{15}), and iron ore, which range in diameter from several millimeters to less than 0.1 mm. Groundmass texture is typically intergranular, but in some thin sections is subophitic or intersertal. The cores of the larger plagioclase phenocrysts are sodic labradorite and the smaller phenocrysts and microlites range from andesine to calcic oligoclase.

Chemically the basalt of Basalt Ridge is characterized by relatively high iron and alkalis. The average chemical composition (table 14) is similar to that of the average hawaiite described by MacDonald (1960).

TERTIARY AND QUATERNARY

BASALT

Basalt of Tertiary and Quaternary age crops out in Reveille Valley, along the west flank of Kawich Valley near Gold Reed, northeast of Mellan in Cactus Flat, and north of Cactus Peak. The rocks are dense black olivine basalts with sparse phenocrysts of plagioclase; none were examined in thin section.

The youngest basalt appears to be that in Reveille Valley, which, according to H. R. Cornwall (oral commun., 1964), was erupted from fissures in the Reveille Range and flowed out onto the valley where it rests on valley-fill alluvium (pl. 1). The basalt along the west flank of Kawich Valley near Gold Reed is probably older than basalt of Reveille Valley because it is largely covered by valley-fill alluvium. Dikes and highly pumiceous basalt rubble occur in an isolated knob, 4 miles north-northeast of Gold Reed, that probably marks the site of a vent zone or small cone that fed the flows in the Gold Reed area. The basalt northeast of Mellan, shown at the north boundary of the map, rests on an erosional surface developed on the Fraction Tuff.

ALLUVIUM AND COLLUVIUM

Approximately half of the mapped area is blanketed by alluvium and colluvium of Quaternary and Tertiary age. This category includes fan alluvium (locally fanglomerate) deposited on pediment surfaces that slope radially away from the mountain ranges, valley-fill alluvium, lake and shoreline deposits, and, locally, landslide deposits and talus. That some of this material is of Tertiary age is indicated by interbedding with volcanic rocks. For example, in the vicinity of Mount Helen, alluvial material that crops out beneath the Thirsty Canyon Tuff was mapped as older alluvium, but where the Thirsty Canyon is not present because of erosion or nondeposition, alluvium that is probably of pre-Thirsty Canyon age is not distinguished from more recent alluvium.

The alluvial material is thickest in the major basins, but the actual thicknesses are unknown. The basins in the project area are similar in size, areal extent, and geologic setting to the Yucca Flat intermontane basin in which the alluvial fill ranges in thickness from 0 to as much as 2,200 feet (W. P. Williams, W. L. Emerick, R. E. Davis, and R. P. Snyder, written commun., 1963). Gravity data (Healey and Miller, 1962) suggest that the alluvial fill and volcanic rocks may be as much as 4,500 feet thick in Kawich Valley and Gold Flat. Drill-hole data at Yucca Flat (W. P. Williams and others, written commun., 1963) show that the composition, texture, and physical properties of the alluvium vary according to the distance from the source, the types of rock at the source, the carbonate content and resultant degree of cementation, and the amount of compaction of the alluvium with depth. The bulk of the material cut by the drill within the project area is sand and clay. (See p. 84.)

Several of the canyons cut into Thirsty Canyon Tuff adjacent to playas have been partly filled with alluvium. For example, in Civet Cat Canyon south of Stonewall Flat, a well dug 75 feet into the floor of the Canyon bottomed in alluvium of post-Thirsty Canyon origin. The mechanism of this occurrence of aggradation is not perfectly understood, but it seems likely that the canyon was cut primarily during a pluvial cycle of the Pleistocene Epoch when increased rainfall probably gave rise to a perennial stream. With diminished rainfall after the Pleistocene Epoch, rock waste was moved into the canyon by constantly overloaded "gully washing" streams. This aggradation continued until the canyon was filled to a level slightly higher than its present depth, as indicated by several alluvial terraces flanking the streambed a few feet above the present stream level. The shift to downgrading may reflect a slight change to relatively heavier precipitation in fairly recent times or relative settling of the Stonewall Flat basin which effectively increased the stream gradient.

STRUCTURE

GENERAL SETTING

The mapped area is in the western part of the Basin and Range province, about 50 miles east of the average east border of the Cordilleran eugeosyncline as defined by Gilluly (1965). The eugeosynclinal border also approximately marks the east border of the region that is characterized by large granitic plutons of Mesozoic age that are related to the intrusion of the Sierra Nevada batholith. Granite plutons are present east of the line, but they are relatively small and widely scattered. The area lies east of a zone of transcurrent faulting first defined by Gianella and Callaghan (1934) and named the Walker Lane by Locke, Billingsley, and Mayo (1940). This zone divides southern Nevada into an area of north-trending ranges (includes most of mapped area) and an area of northwest-trending ranges (fig. 15). The eastern part of the Walker Lane as defined by Locke, Billingsley, and Mayo (1940) was named the Las Vegas Valley shear zone by Longwell (1960), who concluded that the zone was a right-lateral fault with at least 25 miles of displacement. Stratigraphic studies by U.S. Geological Survey personnel working at Neveda Test Site (F. G. Poole, oral commun., 1966) support Longwell's conclusion. The exact location and the nature of the shear zone in and adjacent to the area of study are problematical. Burchfiel (1965) concluded that the Las Vegas Valley shear zone and the Walker Lane do not form a single continuous fracture zone. Nevertheless, there is general agreement that these two faults together with others (for example, the Furnace Creek fault zone in Death Valley) define a zone of structural weakness in the crust along which there has been considerable right-lateral movement. Shawe (1965) prefers to extend one of the major transcurrent faults through Tonopah and southeastward from there through the Cactus Range. Although the present authors have found no unequivocal evidence of strike-slip movement in the area southeast of Tonopah, the prominent northwest-trending grain of the Cactus Range and the occurrence of numerous volcanic centers along a line [2] extending southeastward from the Cactus Range into Nevada Test Site (fig. 15) tend to suggest that a major crustal rift is present in the area along which magmas, generated at great depth, moved upward. That this zone is a major transcurrent fault may never be proved, but it certainly must be considered a reasonable possibility.

The structure of the adjacent Goldfield area, a volcanotectonic feature west of the mapped area (fig. 1), has been described by Ransome (1909) and recently by Albers and Cornwall (1967); the Tonopah area to the northwest, by Spurr (1905), Nolan (1935), and Ferguson and Muller (1949); and the structure of the Bare Mountain area to the southwest, by Cornwall and Kleinhampl (1961).

[2] P. P. Orkild (oral commun., 1962) of the U.S. Geological Survey was the first to note the alinement of volcanic centers, and he coined the expression "line of fire" to designate the concentration of centers.

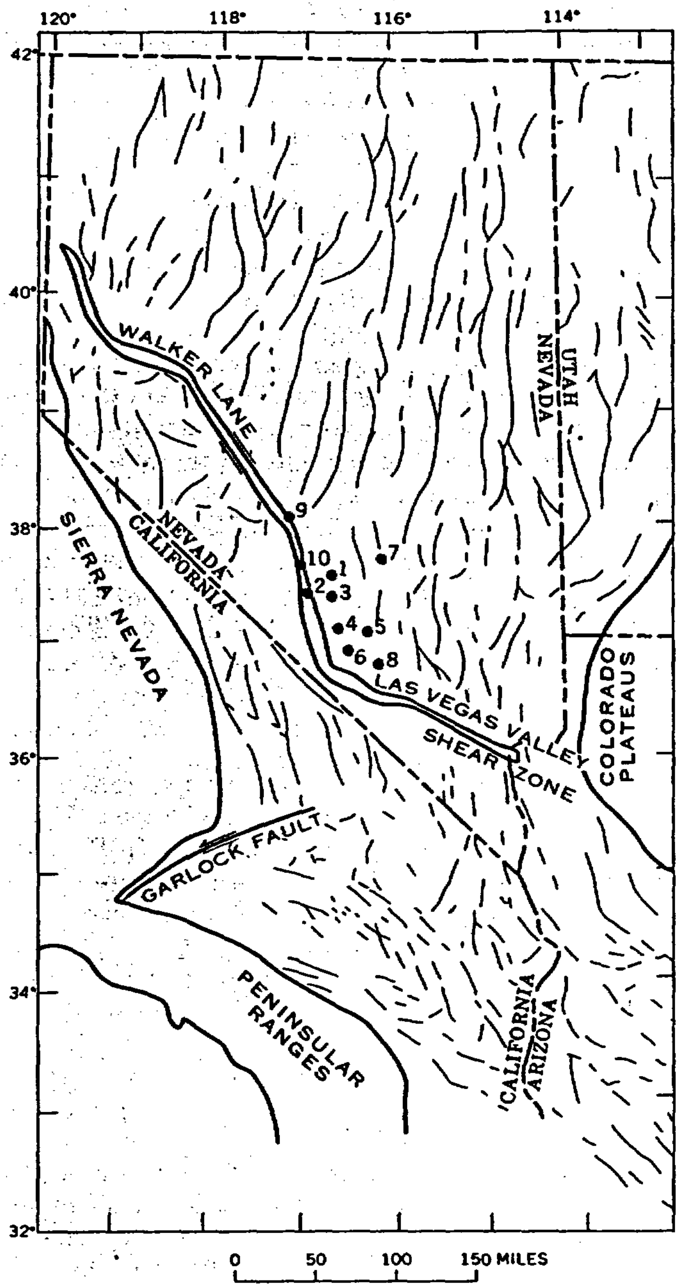

FIGURE 15.—The Walker Lane–Las Vegas shear zone and trends of the ranges in the Basin and Range province. Dots indicate location of major volcanic centers: 1, Cactus Range; 2, Stonewall Mountain; 3, Mount Helen; 4, Black Mountain; 5, Pahute Mesa; 6, Timber Mountain; 7, Cathedral Ridge; 8, Wahmonie; 9, Tonopah; 10, Goldfield. (Modified from Burchfiel, 1965.)

The mapped area contains eight structural blocks or units that lend themselves to separate descriptions, although the relations between units are not entirely clear. The units are separated from each other by structural basins and (or) faults. The units are the Belted Range, the Kawich Range-Quartzite Mountain Block, the Mellan Hills, the Cactus Range, Mount Helen, the

Trappman Hills, Black Mountain, and Pahute Mesa. Of these, only the Belted and Kawich Ranges appear to be predominantly basin-and-range type structural blocks; the others are entirely or in large part volcano-tectonic structural features.

BELTED RANGE BLOCK

The Belted Range block lies along the east border of the mapped area and includes all the topographically high ground between Railroad Valley on the north and Oak Spring Butte on the south. It is bounded on the west by Kawich and Reveille Valleys and on the east by Monotony Valley and an unnamed valley. The block, as a separate structural entity, probably extends beyond the mapped area as far south as the Eleana Range, which lies directly along the projection of the Belted Range and exposes strata of Paleozoic age.

Tertiary strata exposed in the block decrease in age from north to south and dip 20°–60° E. in the northern part of the range and about 5°–10° E. in the southern part. They are cut by numerous normal faults with displacements that range from a few feet to at least 1,000 feet. Paleozoic and Precambrian strata dip at steeper angles both eastward and westward and are cut by numerous normal faults and several low-angle faults, most of which are interpreted herein as gravity-slide faults.

PRE-TERTIARY FOLDS AND THRUST FAULTS

Pre-Tertiary strata in the Belted Range are very steeply dipping, and they contrast, for example, with the strata in the Quartzite Mountain-Cathedral Ridge area where dips on pre-Tertiary strata are nearly conformable to dips in the overlying volcanic rocks. Along Limestone Ridge vertical beds and overturned beds are a common feature. These steep dips are the result of the composite effects of folding, thrust faulting, gravity sliding, and tilting related to basin-and-range faulting.

The oldest beds (Precambrian Stirling Quartzite) crop out on the west flank of Limestone Ridge adjacent to Kawich Valley near Cliff Spring. The youngest beds (Eleana Formation of Mississippian and Devonian age) crop out in three localities on the east flank of the Belted Range between lat 37°16′ and 37°33′.

A thrust-fault system that occurs in the Yucca Flat area of the Nevada Test Site is regional in extent, and it appears to continue northward through the Belted Range and on into central Nevada (F. G. Poole, oral commun., 1966). Thrusting was from the west and northwest. In the part known as the C P thrust (Barnes and Poole, 1968), rocks of late to middle Paleozoic age are overridden by Cambrian and Precambrian rocks. A lower, subsidiary, part of the system, known as the Mine Mountain thrust, has moved Devonian and Silurian carbonate rocks over rocks of Mississippian and Pennsylvanian age.

Both thrust plates are present in the ridge that lies 2.5 miles northeast of Wheelbarrow Peak, at about lat 37°28′. The C P thrust can be observed in the central part of the ridge, where Eleana rocks are overlain by the Halfpint Member of the Nopah Formation and the fault is clearly flat to gently dipping. The upper plate in this area also includes younger rocks from the Smoky Member of the Nopah and from the lower part of the Pogonip Group. A continuation of the structure is the buried thrust shown in section F–F′ (pl. 1) which lies several miles to the north. The Mine Mountain thrust, which occurs in the south half of the ridge, consists mainly of the Nevada Formation with subordinate Silurian dolomite (not shown on pl. 1), and it forms a thin, highly deformed thrust remnant that rests on clastic rocks of the Eleana Formation.

A still lower thrust, known as the Tippinip, may extend from Yucca Flat into the Belted Range. It is covered by volcanic rocks in the Carbonate Wash area (southeast corner of pl. 1), but is believed by F. G. Poole and the authors to be present at depth and to form a structural contact between dolomite of the Spotted Range (pl. 1) on the east and limestone and dolomite of Middle Devonian age (pl. 1) on the west. The latter unit is, at least in part, stratigraphically equivalent to the Nevada Formation in this area, but is dominantly limestone rather than dolomite and represents a facies that is slightly more western and may be described as transitional (F. G. Poole, oral commun., 1967). The Tippinip thrust may also be present in the small Paleozoic exposure that lies on the east margin of the mapped area and 5 miles southeast of Belted Peak, where Eleana rocks may overlie a Devonian limestone.

Several additional thrust faults are present to the east of the Tippinip fault in the Carbonate Wash area, and one of these may have been the mechanism that overturned the Carbonate Wash syncline. A sizable thrust fault may also be present at depth in the vicinity of the Butte fault, and this may have caused the Carbonate Wash syncline to override the adjacent gently north-plunging anticline lying to the east.

The absence of deposits of Mesozoic age in the mapped area precludes accurately dating the period of thrust faulting. The thrust faults do not cut Tertiary strata and it seems reasonable to assume that the faults formed during the same orogeny that gave rise to thrust faults in adjacent areas. Thrust faults northwest of the mapped area, in the Hawthorne and Tonopah quadrangles, have been dated as Jurassic by Ferguson and Muller (1949). The earliest thrusting was from the north followed by thrusts from the northwest. Longwell (1949) dated intense thrusting in the Muddy Mountains near Las Vegas as Middle to Late Cretaceous, and Nolan (1962) dated the thrust faults at Eureka, Nev., as Early Cretaceous. The thrusts in the Belted Range are inferred to be Jurassic or Cretaceous in age.

GRAVITY SLIDE BLOCKS

The strata on Limestone Ridge north of the Kawich Valley-Cliff Spring road are part of a large gravity-slide block that is inferred to have moved from east to west (section *F–F′*, pl. 1). A higher block in the northern part of the ridge (section *E–E′*) and other blocks not shown in cross section are inferred to have the same origin.

The conclusion that these blocks slid into their present positions by gravity movement rather than by thrust faulting from the west is based entirely on the consistent west shift of stratigraphic marker zones in the successive overlying plates. For example, the beds on Limestone Ridge (section *F–F′*) decrease in age from west to east and can be restored to normal stratigraphic position with respect to the underlying beds by moving the overlying block about 4 miles to the east. Thrust faulting from the west would require shifting the east limb of one anticline over the east limb of another. The same relations apply to the higher block shown in section *E–E′*. The beds can be restored to their normal position with respect to the underlying beds by moving them approximately 2,000 feet to the east. Parts of a third and still higher block are found about 1 mile south of section *E–E′*. In this area beds of the Smoky Member of the Nopah Formation have moved about 1,500 feet westward with respect to the underlying strata. To thrust these beds from the west would again entail moving part of an east limb of one anticline over the east limb of another. That this is possible is not questioned; however, the possibility seems remote that each thrust movement could result in younger strata being moved over older strata, and that each thrust would superpose only the east limbs of anticlinal folds in the Limestone Ridge area.

Brecciation at the base of the gravity slide blocks is intense, and this feature largely masks any drag folds or other features that might be used to determine movement directions of the blocks. Thicknesses of breccia range from a few feet to as much as 100 feet, even in blocks that show only a few tens of feet of apparent lateral shift. Locally, the zones are intensely silicified and (or) dolomitized. In the northern part of Limestone Ridge, for example, a small block composed of the Nopah Formation and the Goodwin Limestone (section *E–E′*, pl. 1; and fig. 16) probably moved no more than 200 feet westward, but at the base this block has a breccia zone that is about 50 feet thick on the south and east sides. Brecciated rock is sparse on the west side of the block, but the lack of breccia there may be caused by a normal fault that drops the brecciated rock a few feet down to the east. The breccia fragments and the beds in the overlying block are coarsely recrystallized to dolomite, whereas the underlying strata, especially the Goodwin, are mostly fine- to medium-grained limestone. This occurrence indicates that some of the dolomitization in the Belted Range is related to groundwater or hydrothermal alteration that postdated the major folding and thrust faulting. The gentle dip of the fault plane (section *E–E′*, pl. 1; and fig. 16) resulted in part from eastward tilting, related to basin-and-range faulting, which eliminated part of the original west dip on the fault plane. Direction and degree of dips of the adjacent Monotony Tuff indicate that the fault plane could have been rotated as much as 30° during late Tertiary time.

The age of the gravity sliding apparently is pre-late Oligocene, because nowhere do the volcanic rocks appear to be involved in the sliding. The most likely time was immediately after the period of thrust faulting.

NORMAL FAULTS

Two ages of normal faults are recognized in the Belted Range and throughout the project area (Ekren and others, 1968). The oldest faults, which strike east, northeast, north-northeast, and northwest, are entirely pre-Belted Range Tuff in age. These faults have displacements of as much as several thousand feet. They are as abundant in Tertiary strata as in pre-Tertiary and it is presumed, therefore, that they are post-late Oligocene in age. The youngest faults, which strike north, displace all the strata and give rise to the north strike of the Belted and southern Kawich Ranges. Most of the north-trending faults within the ranges have displacements of only a few tens of feet; however, the range-front faults are inferred to have displacements that locally exceed 1,000 feet. For example, gravity data indicate that the fault that bounds the Belted Range on the west side near Cliff Spring has a displacement of at least 2,000 feet (D. L. Healey, oral commun., 1964). This fault is marked by a feeble lineament on aerial photographs and by small offsets of older alluvium along the east flank of Kawich Valley. The northward fracturing probably began just before the eruption of the rhyolites of Belted Peak and Ocher Ridge (late Miocene), as shown by the fact that these rhyolites were intruded along many north-trending fractures in sev-

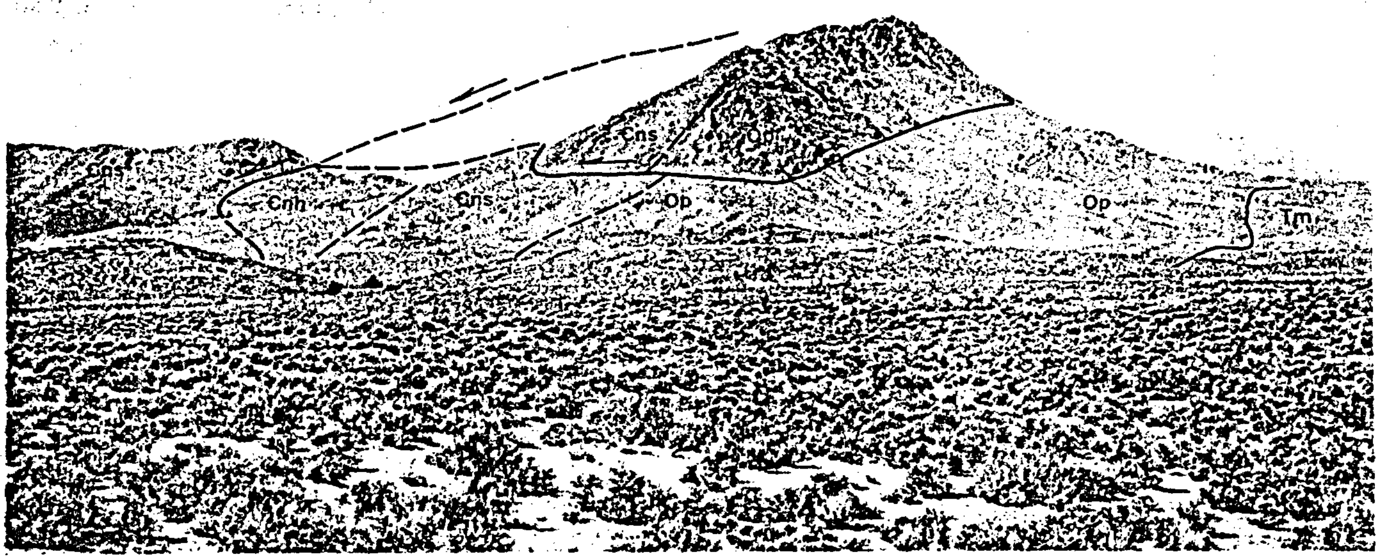

FIGURE 16.—Gravity-slide blocks in northern part of Limestone Ridge, Belted Range. The planes of movement are inferred to have dipped steeply to the west prior to late Tertiary faulting and subsequent eastward tilting. See also section *E–E′* (pl. 1). Tm, Monotony Tuff (Tertiary); Op, Pogonip Group (Ordovician); Ꞓns, Smoky Member, and Ꞓnh, Halfpint Member of Nopah Formation (Cambrian). View to the north.

eral areas, principally in the vicinity of Indian Spring. Evidence is good, however, that the basins and ranges controlled by the north-trending faults did not acquire their present configuration and relief until after the Timber Mountain Tuff was extruded. Drilling in Yucca Flat, for example, discloses that the tuffs are no thicker in the basin than on the flanks, an impossible situation if the basin had formed prior to the eruption of the tuffs from the nearby Timber Mountain volcanic center. By Thirsty Canyon time the area had virtually the same topographic grain but lacked the relief that it has today. This conclusion is deduced from the general distribution of the tuff, evidence of its lapping up against some of the ranges and hills, and its occurrence in old valleys and draws, especially in the vicinity of Lizard Hills in the central mapped area.

KAWICH RANGE-QUARTZITE MOUNTAIN BLOCK

The area of positive relief extending northward from Saucer Mesa and flanked on the east by Kawich and Reveille Valleys and on the west by Gold Flat is here designated the Kawich Range-Quartzite Mountain block. The part of this block that lies within the mapped area averages about 7 miles in width, and trends northward in the south two-thirds of the area and N. 35° W. in the north one-third. The block can be divided into three northwest-trending en echelon structural segments—the southern, central, and northern segments.

The southern segment, which consists predominantly of pre-Tertiary sedimentary rocks, extends from east of Saucer Mesa northwestward to Cathedral Ridge. The sedimentary rocks are nearly all of Precambrian age and are well exposed in a belt almost 12 miles long and as much as 3.5 miles wide. North of Quartzite Mountain the strata have an average strike of north-northwest and low to moderate easterly dips. On Quartzite Mountain and in the outcrops to the southeast, the sedimentary strata strike north, northeast, and northwest and dip about 20°–40° E. East of Saucer Mesa the Precambrian rocks are cut by numerous faults that trend east to northeast and northwest. The contacts with the adjacent volcanic rocks to the east, west, and south are almost wholly fault contacts, and some of these faults have displacements of several hundred feet. Quartzite Mountain itself is an upthrown block that is bounded on all sides by normal faults and is broken into a mosaic of small blocks by variously oriented faults, only a few of which are shown on plate 1. The faults bounding Quartzite Mountain on the west (section *D–D′*, pl. 1.) are inferred to be part of a major north-trending range-front fault system that truncates both the southern and central segments of the structural block. This fault or fault zone is concealed throughout most of its length, but at Quartzite Mountain a small exposure of Stirling Quartzite in the downthrown block indicates a stratigraphic throw of at least 4,000 feet. To the south the range front fault is interpreted as splitting into two faults of small displacement that cut Saucer Mesa; however, several other faults cutting the western part of the mesa may also converge northward beneath the alluvium to join the range-front fault zone.

The central segment, which lies en echelon to the southern segment and extends from Gold Reed to Cedar Pass, consists entirely of Tertiary igneous rocks, and it contains a structural depression named herein the Cathedral Ridge caldera. This collapse structure contains the thickest cogenetic sequence of ash-flow tuffs known to the authors—the Fraction Tuff that is more than 7,000 feet thick. This tremendous thickness suggests that this area subsided concomitantly with the extrusion of the tuff and is, therefore, part of the original caldera. The possibility that a large depression existed there prior to the Fraction Tuff eruptions seems most remote because where the tuff is thickest it rests directly on intermediate lavas without intervening thick deposits of basin fill. The presence of a volcanotectonic collapse structure or caldera in the vicinity of Cathedral Ridge is further suggested by geologic relationships in the area west of Cedar Pass and White Ridge, where some of the youngest ash flows in the Fraction Tuff are deposited on a thick sequence of tuffaceous conglomerate, rubble, and debris flows that apparently accumulated along the wall of the caldera subsequent to the first phases of subsidence (section *C–C′*, pl. 1). This zone of very coarse clastic material thus marks the location of the north and northeast margins of collapse. Rocks within the zone are mapped with similar debris flows and coarse tuffaceous sedimentary rocks north of Gold Reed, but it is not known whether these presumably younger strata at Gold Reed also mark the margin of the caldera. Much of the coarse clastic material north of Gold Reed consists of pre-Fraction debris that must have been derived from the east, presumably from a caldera wall that is now either eroded away or downfaulted into Kawich Valley. The south margin of the caldera is inferred to lie along the trace of two major normal faults that strike N. 65° W. northwest of Gold Reed. These faults have combined throws that total about 7,000 feet, almost equal to the thickness of Fraction Tuff exposed at Trailer Pass. The west margin is inferred to be downfaulted into Gold Flat and to be buried by basin-fill deposits.

The Fraction Tuff dips east and northeast at an average angle of about 25° throughout most of the caldera; this dip suggests that the entire structure was rotated eastward by basin-range faulting. Locally along the collapse zone west of White Ridge, the Fraction dips westward into the caldera at 10°–45°. The inferred area of collapse is elongate northwest and seemingly reflects control by the same forces that controlled the northwest-trending northern and southern segments of the Kawich Range-Quartzite Mountain structural block.

The northern segment, which extends from near White Ridge northwestward to beyond the mapped area (pl. 1), consists predominantly of Tertiary igneous rocks that are older than those of the central segment. This segment is a horst that averages about 4 miles in width and trends N. 35° W. Except for three small masses of Paleozoic sedimentary rocks that might be landslipped blocks totaling less than 1 square mile, it consists of Tertiary tuff, lava, and intrusive masses of pre-Fraction Tuff age. The horst is poorly defined south of Cedar Pass, where it consists mostly of lavas of intermediate composition and where, at White Ridge, it contains a well-defined rhyolite volcano. Lavas and bedded tuffs in that area dip gently southward and are cut by numerous normal faults that trend northwest and west-northwest. The east boundary of the horst south of Cedar Pass is probably buried in Kawich Valley, and the west boundary is nearly coincident with the collapse zone of the Cathedral Ridge caldera.

North of Cedar Pass the horst is composed predominantly of welded tuff of White Blotch Spring. The rocks in this sequence strike northwest and dip moderately to steeply near the north edge of the mapped area (pl. 1), but to the south they have more variable strikes with dips generally less than 20°. They are intruded by a large rhyolite mass, similar in composition to the rhyolite at White Ridge, and by other smaller less silicic intrusive masses. The possibility seems good that intrusive masses are widespread in the shallow subsurface and that their emplacement was a major factor in the uplift of the horst—as much as 3,000 feet on the southwest side. Aside from the range-front faults, numerous normal faults cut the horst with displacements ranging from a few feet to as much as 500 feet. These give rise to a mosaic of tilted blocks with random dips, a feature that contrasts with other ranges that constitute tilted east-dipping blocks rather than horsts; for example, the Belted Range. Although the rocks are well exposed throughout the range, the faults are generally concealed beneath talus and stream gravels, and their dips are rarely observable.

CACTUS RANGE

The Cactus Range is a northwest-trending raised structural block bounded on the east by Cactus Flat and on the west by Stonewall Flat. It terminates abruptly to the northwest a few miles beyond Cactus Peak; to the south and southeast it passes into numerous low hills that extend nearly to Mount Helen. The block is at least 18 miles long and its average width is about 5 miles; it strikes N. 40° W.

The Cactus Range is one of at least five separate mountain masses that lie along the Las Vegas valley-Walker Lane lineaments. In these mountain masses volcanic rocks of Tertiary age are steeply tilted, highly faulted, and invaded by numerous hypabyssal intrusive masses (R. E. Anderson and E. B. Ekren, unpub., data). In each area the steeply tilted strata are cut by wrench and low-angle faults. The similar structural patterns recognized in each area apparently result from superposition of structures of probable volcanotectonic origin on those resulting from wrench and related styles of regional tectonism. The structures are extremely complex and difficult to interpret. They are the subject of present field studies designed to evaluate the interrelationships between the volcanotectonics and regional tectonics. The present report includes only a summary of the major structural features in the range and brief tentative interpretations of the sequence of events that produced them.

The main mass or central core of the Cactus Range is composed of minor upper Paleozoic sedimentary rocks, one small exposure of Mesozoic granite, a thick sequence of widespread Tertiary extrusive and sedimentary rocks ranging stratigraphically from Monotony Tuff to tuff of White Blotch Spring, and a wide variety of intrusive rocks believed to postdate the White Blotch Spring and known to predate the Thirsty Canyon Tuff. The central core rocks are flanked locally by post-White Blotch Spring volcanic and sedimentary rocks that are downdropped along known or inferred faults that tend to gird the range. These younger strata include the Fraction Tuff, and they predate the Thirsty Canyon Tuff which laps onto the range.

The fault density throughout most of the range is much greater than shown on plate 1, where only faults that in most places juxtapose rocks of contrasting lithology are shown. Northwest, east, and northeast fault trends prevail, but locally the fault pattern is a random mosaic. The principal faults trend northwest. The pre-Tertiary rocks, for example, are restricted to a northwest-trending horst about 1 mile wide and 7 miles long along the west margin of the southern range. Also, major northwest-trending faults and related structures that may be of volcanotectonic origin occur in the northern range.

The Monotony Tuff was deposited on a surface of moderate relief developed on gently dipping upper Paleozoic rocks. Its deposition was followed by emplacement of the several ash-flow tuffs that make up the lower and middle parts of the tuffs of Antelope Springs. The first major event of structural significance in the Cactus Range appears to have been the eruption of the upper tuffs of Antelope Springs. That these tuffs were erupted from within or near the range is indicated by their great thickness, especially in the northern range, and by the local occurrence of many large lithic inclusions. The lithic assemblage in these tuffs includes all rock types known to have been present in the area prior to emplacement of the tuff and suggests derivation from disrupted roof rocks in or near the range. The assemblage would be anomalous in most other areas. Evidence for postextrusion collapse within the range is twofold.

1. Along the northern range axis the tuffs of Antelope Springs are steeply tilted, contorted, and locally sheared. In some places they are overlain by gently dipping to flat-lying tuff of White Blotch Spring; the deformation would thus be pre-White Blotch Spring in age. The intensity of this pre-White Blotch Spring deformation far exceeds that which could reasonably have been produced by normal block faulting. It is inferred to have been produced by postextrusion collapse along a northwest-trending rift or graben. The tuffs of Antelope Springs along the northwest margin of the range also dip steeply and are overturned locally, but similar attitudes prevail in the adjacent younger tuff of White Blotch Spring, and this fact precludes unambiguous assignment of a pre-White Blotch Spring age to the deformation there.
2. As much as 800 feet of thin-bedded lacustrine sedimentary rocks accumulated on relatively undeformed tuffs of Antelope Springs in the middle part of the range, and subsidence of the underlying rock is indicated. The sedimentary rocks are reasonably inferred to predate the White Blotch Spring. They may have been deposited in a less deformed southeastward extension of the postulated postcollapse rift or graben.

Partial resurgence within the inferred Antelope Springs collapse structure is suggested by deep pre-White Blotch Spring erosion of intensely deformed tuffs of Antelope Springs in the northern part of the range.

The second major event was eruption of the tuff of White Blotch Spring. Evidence for eruption of these tuffs from a source within or adjacent to the Cactus Range and for subsequent collapse is essentially the same as for the earlier tuffs. After eruption of the White Blotch Spring, subsidence may have occurred along a northwest zone as extensive as the present range. Steep northeast tilting and local overturning to the southwest of tuff of White Blotch Spring and older rocks occurred in a zone as much as 1 mile wide and at least 8 miles long flanking the northern range on the west. Some areas of complete structural chaos occur within the zone. The White Blotch Spring dips more gently or is locally flat lying east of this zone of deformation. Along the eastern range flank the White Blotch Spring dips moderately west toward the axis of the inferred rift. The fact that no zone of steep tilting is recognized there suggests that if the steep dips to the west are related to postextrusion collapse of a northwest rift or graben then the structure is asymmetric with maximum downthrow along the western edge. The pattern of deformation is similar to that produced in the upper tuffs of Antelope Springs along the range axis, and only locally is it possible to distinguish between these two periods of deformation. Subsidence of the central range is supported by intrusion of stocks, sills, and laccoliths of post-White Blotch Spring age into the post-Antelope Springs lacustrine rocks. The considerable thickness of overburden younger than the lacustrine rocks required for intrusion of these masses probably consisted of tuff of White Blotch Spring that accumulated to a much greater thickness in the subsiding block than outside it. This inferred thickness of White Blotch Spring is assumed to have been removed by erosion during later uplift of the range.

The third major event consisted of prolonged intense intrusive activity following the postulated subsidence related to extrusion of tuff of White Blotch Spring. An estimated 200 exposed intrusive masses were emplaced. Lavas and tuffs flanking the northern part of the range probably were erupted concomitantly with emplacement of the intrusive masses. These strata may have covered most of the area now occupied by the Cactus Range. During the latter part of this period of igneous activity, the range was uplifted to its present structural level. Uplift occurred along a system of faults that tend to gird the range. They are well delineated in the northern part of the range but are mostly inferred in the southern part. Little or no additional drag folding, tilting, or brecciation occurred along these fractures during uplift. The contrast between lack of deformation during uplift and intense deformation during subsidence indicates that the fractures along which postextrusion subsidence occurred dip inward. The volume constrictions required by collapse along inward-dipping fractures are assumed to have caused the intense tilting of the strata at the margins of the downdropped blocks or rifts. The volume increase required for the abundant dikes and plugs within the blocks may be related to the volume increase accompanying uplift along inward-dipping faults.

Emplacement of the stocks was accomplished mainly by upward displacement of large blocks of roof rock along preintrusive faults. The stocks are composed mostly of relatively low-silica granitoid-textured rocks and are among the oldest intrusive rocks in the range. They were presumably emplaced when the range was greatly depressed. Emplacement of laccoliths and sills was primarily along zones of sedimentary rocks, but locally these masses are markedly discordant. The dikes were intruded mainly along west- and northwest-trending faults. They are mostly aphanitic rhyolite and are the youngest intrusive masses in the range. As noted previously, their emplacement probably marks a period of volume inflation attending uplift of a volcanotectonic rift or graben.

Although the interpretation of the structural evolution of the Cactus Range is based largely on inference, there is little doubt that large volumes of rock were extruded from the area and that collapse occurred. The range is unique among reported deeply eroded volcanotectonic structures in that postcollapse uplift was of sufficient magnitude to elevate much of the floor of the structure. The range offers an unusual opportunity to observe the fracture zone along which collapse occurred, a feature normally buried beneath the "moat fill" of less eroded calderas.

THE MELLAN HILLS

The Mellan Hills are a series of north-northwest-trending topographically subdued lava ridges separated by valleys formed of soft tuffs. The geologic structure is so completely dominated by northwest- and northeast-striking faults that the area has a distinctly rectilinear, almost diagrammatic-appearing structural and topographic grain. The block is about 10 miles wide (east to west) and 20 miles long (north to south). It is separated from the Cactus Range by Cactus Flat and from the Kawich Range by the northern extension of Gold Flat. On the southwest, the hills merge topographically with the Trappman Hills, from which they are separated structurally by a major fault with large displacement that drops Tertiary strata in the Mellan Hills against Precambrian crystalline rocks in the Trappman Hills.

Despite low elevations and a consequent lack of vegetation, the hills have the poorest exposures of any bedrock area in the Bombing and Gunnery Range. This is due to the abundance of rubble-weathering flow breccias in the lava piles and the tectonic breccias that formed as the lava ridges were faulted, titled, and shifted eastward by northwest-striking very low angle faults (section *A–A′*, pl. 1). The rubble obscures most of the faults that are the key to understanding the area, and were it not for the recent experience gained by R. E. Anderson mapping in the Boulder City area in extreme southeastern Nevada, where remarkably similar geology is well exposed a sensible interpretation of the structure of the Mellan Hills probably would not have been possible.

Very few fault planes were actually observed. All that were observed show eastward dips of 25°–35° with slickensides alined directly downdip. Beds in the hanging walls (mostly lavas) display steep reverse dips of 50° to nearly vertical, averaging 70° into the fault plane. The footwalls are formed of zeolitized tuff of Wilsons Camp which is intensely silicified in a zone about 6 inches wide adjacent to the fault plane. The silicified rock weathers to hard rubble and bears a striking resemblance to flow-laminated crystal-poor rhyolite.

Some of the faults that are inferred in the valley areas where exposures are especially poor were located principally on the occurrence of the silicified "rhyolite" rubble. In other areas, especially along the western ridges where partially welded tuff of Wilsons Camp is mostly fresh and vitric, the fault zones are marked by float of fine-grained brown gouge that formed by the grinding of the tuff of Wilsons Camp in preference to the more competent lavas in the hanging-wall block. This type of float occurs in all instances within a few feet of the west margins of the west-dipping lava ridges, and it is inferred, therefore, that the ridges are all bounded very closely on their west flanks by low-angle faults.

The structural relations of a faulted lava ridge that is thought to be a typical example are well displayed northeast of Triangle Mountain near the east township line of T. 4 S., R. 48 E. Here one can see the controlling low-angle fault and also see beneath the lava ridge. The base of the topographic ridge clearly is also essentially the base of the west-dipping lava mass. It is apparent, however, that in addition to the eastward and downward movement along the main controlling fault, which dips 25° E. there has been some gliding of the lava mass at the base of the ridge along planes developed above the main fault in soft bedded tuffs. These planes dip at angles less than 10° E. Total movement along these gently dipping planes is obviously very minor for this particular ridge; however, this type of movement could have given rise to considerable displacement for some of the larger masses, especially the eastern ridge shown at the end of section *A–A′* (pl. 1). This mass is inferred to have moved eastward principally along a plane that probably dips about 5°. The lavas in the eastern ridge display flow layering that dips generally east, whereas flow layering in nearly all the other ridges dips about 70° W. The layering, however, is in marked discordance with dips in the subjacent bedded tuffs. This discordance and the lack of reverse dip into the fault plane suggest a different mode of faulting and transport, a mode consistent with movement along gentle glide planes similar to those visible in T. 4 S., R. 48 E.

A feature of the hills that is nearly as conspicuous on aerial photographs as the northwest-trending ridges is a series of lineaments and dikes that strike east-northeast, nearly at right angles to the ridges. The lineaments that are not occupied by dikes are extremely straight. There is an apparent horizontal offset of the lava ridges across some of the northeast-trending lineaments. The lineaments are inferred to mark the traces of lateral faults that have horizontal displacements of several hundred feet. The sense of movement is right lateral across some of the faults and left lateral across others; the total offset could be either right or left lateral. The occurrence of quartz latite dikes in the northeast set of faults indicates that faulting started before the intermediate rock volcanism had completely ended.

Thus far this description has been concerned with those rocks lying south of the Antelope Springs-Trailer Pass Road. A tectonic fabric that appears to be similar to that of the southern hills prevails north of the road. There the younger rocks, exclusive of the rhyolite, dip consistently 20°–30° E. and the older tuffs of Antelope Springs and White Blotch Spring dip as much as 70° W. Two interpretations are possible: (1) the Fraction Tuff and dacite lavas rest on the older tuffs with a nearly 90° unconformity, or (2) they have moved over the underlying rocks in a manner similar to the gliding transport inferred for the eastern lava ridge south of the road. We favor the gliding interpretation because of (1) the intensely broken nature of the dacite lava and the Fraction Tuff, and (2) the conflict that exists between a nearby flat outcrop pattern and an internal dip of 20°–30° E.

Either of the two possible explanations indicates, however, that the northern Mellan Hills was the site of profound tectonic activity that either postdated the Fraction Tuff (gliding hypothesis) or postdates the tuffs of Antelope Springs and White Blotch Spring and predates the Fraction Tuff (unconformity hypothesis). The rhyolite masses in the northern hills do not seem to be involved in the deformation that affected the Fraction and older rocks. The rhyolite contains rubble of dacite at its base and is clearly younger than the dacite. Poor exposures suggest that the rhyolite rests unconformably on the Fraction Tuff.

The rhyolite has not been dated isotopically. The lavas of intermediate composition that form the faulted ridges south of the road have been dated at about 18 m.y. (table 5), and they are overlain unconformably by the Grouse Canyon Member of the Belted Range Tuff dated at about 14 m.y. (table 5). If one ignores the unconformity hypothesis that is applicable only to the northern hills, the period of the intense structural activity seems to be bracketed by these dates.

The type of faulting that affected the Tertiary rocks of the Mellan Hills is not unique in the Basin and Range province. Longwell (1945) described strikingly similar faults in southeastern Nevada, and we have observed them in the Cactus Range. Anderson, studying such faults in a broad area in the vicinity of Hoover Dam, thinks that none of the several possible explanations currently in vogue to explain this type of faulting—for example, simple gravity sliding—fully satisfies all the data. At Mellan, the occurrence of northeast-trending dikes and tear faults that formed during a period of intermediate rock volcanism and virtually simultaneously with the development of northwest-trending low-angle glide faults indicates deeper seated control than mere gravity sliding would require. More data are needed on a regional scale before this type of deformation can be satisfactorily explained.

TRAPPMAN HILLS

The Trappman Hills form a north-northwest-trending horst bounded on the north by Cactus Flat, on the east by a syncline that is inferred to be the southward extension of Cactus Flat, on the west by a structural low that flanks Mount Helen, and on the south by Gold Flat. The hills are formed of Precambrian gneiss and schist, the only known occurrence of crystalline basement rocks in the mapped area. High-angle northeast- and northwest-trending faults bound the horst on the north, west, and east. Gravity data suggest that the horst extends southward for several miles beyond the outcrop area. The Precambrian rocks there are probably at shallow depth beneath alluvium and Thirsty Canyon Tuff. See Rogers, Ekren, Noble, and Weir (1968).

The Precambrian rocks are cut by several northwest-trending rhyolite dikes that nearly parallel northwest-striking foliation and by one dike of dark-gray rhyodacite. This dike follows a northeast-trending fault that drops upper tuffs of Antelope Springs against gneiss and schist.

The lack of upper Precambrian and Paleozoic rocks in fault blocks in and adjacent to the Trappman Hills suggests that these rocks are absent or very thin in and near the hills and that Tertiary rocks may directly overlie the basement strata adjacent to the horst block. This is suggested also at Mount Helen, about 5 miles west, where a tuff rich in lithic fragments crops out that is found in no other area and that is presumed, therefore, to be locally derived. The chief lithic fragments in this rock are gneiss and schist of the Trappman type. This fact indicates, perhaps, that little sedimentary rock of Paleozoic age was encountered by this tuff during its ascent through the crust. The chief lithic fragments in dikes of rhyodacite, exposed near Triangle Mountain about 7 miles northeast of Trappman Hills, are also gneiss and schist of the Trappman type.

The thinness or absence of Paleozoic strata suggested by these data indicates either that prior to Tertiary volcanic eruptions the area stood as a topographic high for a long period during which the Paleozoic rocks were stripped away or that the crystalline rocks at Trappman Hills have been thrust over the Paleozoic rocks. Data on this point are sparse. The gravity high over the hills seemingly is low for basement rocks, because higher anomalies were obtained over Paleozoic and Tertiary rocks in the mapped area. This relatively low anomaly, however, probably reflects only the low density of the quartz monzonite gneiss.

MOUNT HELEN

Mount Helen is a volcano and lava pile that lies at the center of a structural dome which, in turn, lies near the center of a collapsed area or caldera that is 9 miles wide measured east to west (sections *A–A′* and *B–B′*, pl. 1). The structure on the north and south is covered by younger volcanic rocks and alluvium. In the center of the 9-mile-wide zone of collapse (section *A–A′*, pl. 1), an additional collapse zone or graben, which is about 4 miles wide, occurs. This zone contains rocks that are younger than the lava pile; these include the tuff of Tolicha Peak, and lacustrine and fluvial sedimentary rocks totaling more than 2,000 feet in thickness. The graben is inferred to be part of a caldera that formed during the eruption of the tuff of Tolicha Peak.

The lavas at Mount Helen are radially distributed around a large feeder neck that is exposed at the south end of the mountain. The neck is a composite mass composed of an older quartz latite that forms a peripheral zone in the neck and a younger quartz latite that fills the central part. The neck is at least three-fourths of a mile in diameter, but the exact dimensions are unknown because on the northwest and north the rocks grade laterally to lava flows. The feeder may actually be elongate north to south and could underlie the entire length of the mountain (this possibility is not inferred, however, in section *A–A′*, pl. 1).

The andesitic basalt that underlies the quartz latite is not found in the feeder neck, but was undoubtedly fed from the same locality as indicated by several radial dikes that are exposed south and west of the mountain.

SEQUENCE OF EVENTS

The volcanic and structural history of Mount Helen is long and varied, and deciphering the early history is hampered by the occurrence of alluvium and young volcanic rocks that not only mask large parts of the 9-mile-wide structure but also cover vast areas beyond. Little is known, therefore, of the distribution of the older tuffs that can reasonably be inferred to have been extruded from the Mount Helen volcanic center and that caused the initial collapse. Mount Helen possibly lies in a large rift zone which is bounded on the east by crystalline basement rocks of Precambrian age and on the west by sedimentary rocks of early Paleozoic age. The rift may have controlled the location of the Cactus Range, Black Mountain, and Timber Mountain centers as well as Mount Helen.

Whether or not such a rift zone exists, evidence is good that tuffs were extruded from the Mount Helen center prior to the lava eruptions and that these extrusions caused collapse across a broad area. Two ash-flow–tuff cooling units, which have not been positively identified elsewhere crop out on both the east and west flanks of the graben. They probably are radially distributed around the mountain at depth. These ash-flow tuffs were called tuff of Mount Helen and are shown separately on a map covering the north half of the Black Mountain quadrangle by Rogers, Ekren, Noble, and Weir (1968). On plate 1 the two units are included with the tuffs of Antelope Springs. (See p. 29.) They probably are at least 1,500 feet thick. That a topographic and presumably a structural depression existed at Mount Helen after the tuffs were extruded is indicated by the occurrence of coarse fluvial conglomerate, sandstone, and lacustrine siltstone and shale above the tuffs and beneath the lava pile. These rocks are poorly exposed and, although their average thickness is only 100–200 feet, they are widespread, occurring both inside and outside the inner graben.

After the deposition of the sedimentary rocks, lavas ranging in composition from andesitic basalt to quartz latite were erupted from Mount Helen; a long period of erosion then ensued. The lavas from Mount Helen were deeply eroded, at least on the south and southeast flanks, and the volcanic neck was exhumed. In places some of the underlying bedded sedimentary rocks and older welded tuffs were exposed at the surface before the basal member of the tuff of Tolicha Peak was deposited. This tuff locally rests on prelava rocks and locally fills deep gullies cut into the lavas. It seems certain that the tuff lapped up against the feeder neck itself. At the base the tuff contains abundant boulders and cobbles of quartz latite and basalt, but 20 feet above the base and upward it is virtually free of lithic fragments. It is overlain by thin beds of ash-fall tuff, cross-bedded sandstone, lacustrine siltstone, and thin-bedded limestone. The occurrence of lacustrine strata almost directly above the ash-flow tuff indicates that a peripheral low area, which might have existed prior to the ash-flow eruption, was not filled with tuff or that collapse immediately followed the extrusion of the basal member.

After the deposition of these sedimentary rocks, the middle member of the tuff of Tolicha Peak was deposited, and collapse certainly occurred. This member south of the mountain is overlain directly by thin beds of limestone and by several tens of feet of finely laminated siltstone, some of which contains fossil plant debris and fish. The collapse associated with the eruption of the second member may have resulted in the burial of the entire mountain, as suggested by the occurrence of finely laminated siltstone near the top of the mountain at the south end. This material is identical with siltstone that crops out around the flanks. The third member of the tuff of Tolicha Peak was then deposited, but whether this unit also gave rise to collapse is uncertain inasmuch as it is almost completely covered by alluvium.

It seems likely that the vent for the tuff of Tolicha Peak was located some distance south of the mountain and that the greatest collapse occurred there, possibly centered on a small gravity low in T. 6 S., R. 47 E. (pl. 1).

After the deposition of the tuff of Tolicha Peak, a dome formed at Mount Helen. Faulting undoubtedly accompanied the doming, and reverse movement probably occurred along faults that formed the earlier graben. Rhyolite dikes that now crop out on the north and northeast flanks probably were intruded during this episode. During a subsequent period of erosion, pediment surfaces were formed radially around the mountain, and the top of the lava pile was exhumed. Porphyritic basalt was extruded from the central feeder zone and also from one arcuate fault on the east side. The eruption of basalt was followed by extensive erosion that removed much of the basalt before the deposition of the Thirsty Canyon Tuff, which lapped onto the flanks of Mount Helen and rests on fanglomerate or other alluvial material containing many large boulders of the Mount Helen porphyritic basalt. Since Thirsty Canyon time, the area has been extensively dissected and most of the Thirsty Canyon adjacent to the mountain has been stripped away.

BLACK MOUNTAIN

The Black Mountain volcano is an excellent example of multiple caldera subsidence uncomplicated by either penecontemporaneous radial faulting or subsequent basin-and-range normal faulting.

The main collapse, which formed an elliptical depression 7 miles in average diameter centered on the summit of Black Mountain, took place after the deposition of the Spearhead Member of the Thirsty Canyon Tuff. A smaller collapse caldera formed within the post-Spearhead caldera immediately west of Black Mountain after the deposition of the Gold Flat Member. Another caldera probably formed within the area of post-Spearhead collapse after the eruption of the Trail Ridge Member. These calderas, however, if they existed, have been completely obliterated by the rhyolite of Pillar Spring.

The cycle of (1) eruption of ash-flow tuff, (2) caldera subsidence, and (3) partial or complete filling of the caldera by lavas has been repeated several times in the Black Mountain center. Thus the trachytes of Yellow Cleft and the trachytes of Hidden Cliff partially filled the post-Spearhead and post-Gold Flat calderas, respectively. The lower and middle units of the rhyolite of Pillar Spring, emplaced after the deposition of the Trail Ridge Members, are probably closely related to the inferred post-Trail Ridge caldera.

Almost all the lavas that erupted after the deposition of the Spearhead Member seem to have been derived from from central vents located within collapsed calderas. In only one place were lavas erupted from the marginal fractures of the post-Spearhead, and those lavas were in very small amounts.

PAHUTE MESA

Pahute Mesa is a broad elevated plateau of relatively gentle relief measuring approximately 20 miles east to west and about 10 miles north to south. On the south, adjacent to the moat area of the Timber Mountain caldera, the mesa is bounded by steep slopes with total maximum relief of about 1,700 feet; on the west it abuts Black Mountain; on the northwest it slopes gently into Gold Flat; and on the north and east it joins the southern extensions of the Kawich and Belted Ranges. The mesa is unique in the Nevada Test Site area in that it is a broad area of gently dipping strata not broken by large fault-controlled basins. Gravity data (pl. 1) indicate that the mesa overlies a deep structural basin, and recent drill-hole data indicate that the volcanic strata are at least 13,000 feet thick in the basin. This structural basin, with its great thicknesses of many different volcanic strata, has been interpreted by geologists working in the Nevada Test Site (Orkild and others, 1968; Nöble and others, 1968) to be a large caldera that gave rise to the Belted Range Tuff and related lavas and tuffs.

Most of the faults on Pahute Mesa strike dominantly north, and the strata are downthrown mostly on the west. Vertical displacements average less than 100 feet but range from a few feet to as much as 600 feet. Most of the faults are of post-Thirsty Canyon (early Pliocene) age; a few are older, and some are recurrent, showing greater displacements in pre-Thirsty Canyon strata than in Thirsty Canyon. The faults are inferred to be basin-range structures wholly unrelated to the volcanotectonic basin that underlies the mesa.

MINES AND MINING

The principal mining areas in and adjacent to the project area are at Gold Reed and Silverbow in the Kawich Range; Antelope Springs, Cactus Spring, and Wellington Hills in the Cactus Range; Gold Crater northwest of Mount Helen; at Wilsons Camp; and in the Trappman Hills. These mining areas and others in Nye County have been briefly described by Kral (1951). Prospect pits and exploratory shafts also are abundant in other areas but only in the above listed areas is it apparent that ore was actually shipped. All areas are abandoned, but the Silverbow mining area, just north of the mapped area, enjoyed limited exploration activity during the present field study.

ANTELOPE SPRINGS DISTRICT

The Antelope Springs mining district covers a rectangular area measuring 12 miles northward and 10½ miles eastward, with Antelope Springs in the Cactus Range at the approximate center (Schrader, 1913). It includes scattered mines and prospects at Antelope Springs and Wilsons Camp and in the Wellington Hills.

At Antelope Springs several shafts were sunk along north-trending faults that drop the strata down to the west. The mineralized fault zone along which most of the mines are located dips about 30° W. and displaces the upper part of the tuff of Antelope Springs at least 1,000 feet down to the west. This fault is cut by several west- and northwest-trending faults of small displacement. Drifts in the mines follow both the main fault zone and the west-trending minor faults. The tuffs are propylitically altered throughout the area, and adjacent to ore-bearing veins they are either intensely silicified or argillized and are bleached to light greenish gray and light gray. Schrader (1913), who studied the district in 1912 when the mines were in full operation, noted that sericite, calcite, chlorite, alunite, and hydrous iron oxide, in addition to secondary clay and silica, are common gangue minerals. Development at that time was restricted to the oxidized zone, which yielded cerargyrite and argentite as the chief ore minerals with a silver-gold ratio of 4:1. At the old sulfide prospect

2½ miles south of Antelope Springs (Schrader, 1913, p. 14), the lower part of a 120-foot-deep shaft penetrated the sulfide zone. According to Horton, Bonham, and Longwill (1962a, b), the Antelope Springs mines produced at least 10,000 ounces of silver. The mines are filled with water to very shallow depths, and the good pay zones apparently were below the present water level.

At Wilsons Camp the mines are along a fault zone trending N. 60° E. that drops the strata down on the northwest side. Most of the fractures in the fault zone are filled with quartz, but many are open and contain cavities as much as 6 inches wide that are lined with dogtooth crystals of quartz and minor calcite. Pyrite and hydrous iron oxides are locally abundant. The rocks in the mine are dacitic lavas and welded tuffs of Antelope Springs that have been silicified and argillized. The welded tuffs occur principally in the upthrown block and the lavas, in the downthrown. According to Ball (1907, p. 139), the ore averaged 1 ounce of gold to 5 or 6 ounces of silver. These mines were worked also in the late 1930's and very early in the 1940's. Total production is unknown.

In the Wellington Hills and in the low areas east of them, altered Monotony Tuff, tuffs of Antelope Springs, and intrusive masses of intermediate to rhyolitic composition form most of the weakly mineralized outcrops. Hydrothermal alteration was locally intense along numerous faults in this area. One mine shaft follows a high-angle fault in fine-grained melanodiorite, and numerous shafts and prospect pits are located in country rock adjacent to intrusive masses and along faults some distance from the intrusives. There is, however, no apparent genetic relation between mineralized rock and the intrusion of the various masses; the period of hydrothermal alteration and mineralization postdates intrusive activity.

Several small mines that were being worked during 1905 (Ball, 1907, p. 95) are located south of Wellington Hills and north of Gold Crater in a small patch of brecciated and altered tuff or rhyolite.

Other mines, located 3 miles northwest of Mount Helen near the old townsite of Jamestown (Schrader, 1913, pl. 1), are in altered dacite and andesite that intrudes(?) tuffaceous sediment. At the surface the rocks are brecciated and highly silicified or argillized, and along fault zones they contain barren vein quartz. Hydrothermal alteration occurred in the rocks throughout the area, but most intensely in rocks along faults. Silicification is most intense near the surface and gives way at depth to predominantly argillic alteration; this distribution suggests silica enrichment during weathering. The mine dumps at three of the four main shafts in this area contain both oxidized and unoxidized minerals. The occurrence of oxidized minerals indicates that the workings penetrated the unoxidized zone where primary pyrite and chalcopyrite occur in veinlets a few millimeters wide and as disseminated grains. The rock was mined for silver and gold associated with the sulfide minerals.

TRAPPMAN HILLS

The workings in the Trappman Hills consist of two shafts sunk into Precambrian strata near the south end of the hills. One shaft is sunk in a north-trending quartz vein about 60 feet thick; the other is in a pyritized fault zone that trends north-northeast and dips 60°–70° W. Ball (1907) visited these workings shortly after they were opened in June 1905 and recognized three distinct periods of quartz vein formation. The earliest is pegmatitic, and the associated veins are apparently all barren except where they are brecciated and stained by limonite. In the two youngser sets of veins, which carried the principal ores, Ball (1907, p. 139) recognized minor amounts of native silver, galena, and horn silver. Ball reported that the ratio of gold to silver was 1:4.

GOLD CRATER

Gold Crater consists of a few small hills and knobs of pre-Thirsty Canyon rock surrounded by Thirsty Canyon Tuff, the edge of which forms a discontinuous subcircular rim resembling a crater rim. Here, as in other mining areas in the region, the hydrothermal alteration predates the Thirsty Canyon Tuff. The alteration and associated mineralization were most intense in the north half of the "crater," where most of the mines and prospects are located. Several different lithologies can be recognized in the northern "crater" area, but the chief rock is quartz latite lava of the Mount Helen type. Alteration is predominantly argillic and is most intense along faults. Most of the mining was restricted to the oxidized zone, where the rocks show no visible minerals other than limonite and hematite. Rocks in the mine dump at one of the deeper shafts contain pyrite, indicating penetration below the oxidized zone. Altered rocks, possibly mineralized, probably extend beneath the Thirsty Canyon Tuff adjacent to the mining area.

GOLD REED (KAWICH)

The Gold Reed (originally Gold Reef or Kawich) mining area lies 6 miles northwest of Kawich playa on the east side of the Kawich Range. The first locations were made in December 1904, and 10 miners were at work in August 1905 after a rush of several hundred men in the spring of 1905 (Ball, 1907, p. 111).

The principal mines are located along a northwest-trending silicified horst along which the strata have been dropped both to the northeast and southwest. The silicified zone forms a reeflike ridge; hence, the original name Gold Reef. It is not known when or how the original name was corrupted to Gold Reed.

None of the major mines are accessible at present; however, all the deep shafts are sunk in porphyritic dacite which seems to be the principal ore bearer. The dacite is bleached to light gray and pastel shades of yellow and pink. The gold is not visible to the eye but apparently is associated with iron oxide and pyrite. According to Ball (1907, p. 111–112), some of the limonite-stained casts of phenocrysts contained visible plates of hackly gold, but these relatively rich oxidized zones apparently were restricted to shallow depth and have been mined out.

The amount of gold ore produced at Gold Reed is unknown. Data compiled by Horton, Bonham, and Longwill (1962b) indicate only that the Kawich (Gold Reed) area produced an estimated total of between 0 and 1,000 ounces of silver.

SILVERBOW (TICKABO) MINES

The Silverbow (Tickabo) mines are located half a mile north of the mapped area, in T. 1 N., R. 49 E., along the west flank of the Kawich Range northeast of Mellan; only a few minor prospects are in the project area. The mines were operated intermittently during 1904–40, and they produced at least 10,000 ounces of silver and 1,000 ounces of gold (Horton and others, 1962a, b).

The mines are along several northwest- and west-trending faults that drop the Fraction Tuff and dacite lavas down on the west and south against the tuff of White Blotch Spring and older tuffs; mines are also in and adjacent to rhyolite plugs. Horn silver is concentrated locally in thin veinlets, and in places it is disseminated in the country rock. Secondary copper minerals occur locally; Ball (1907, p. 109) reported that specks of stephanite and "ruby silver" occur and that gold occurs free in some of the veins but that silver was the predominant metal.

The authors visited the Silverbow mines in April 1964 and were guided through the area by Mr. Dan Sheek of Tickabo Mining and Milling Co. At that time the company was in the process of reopening many of the mines and was building a small concentration mill. (The reactivation was short lived; by 1967 all activity had ceased.) Several of the prospects controlled by the Tickabo Mining and Milling Co. are in Fraction Tuff. Inasmuch as the lavas of intermediate composition are the principal ore bearers in adjacent areas, especially at Tonopah where they also underlie the Fraction Tuff, those lavas possibly are mineralized at depth along the west flank of the range in the Silverbow area. The Fraction has been pervasively altered in much of the Silverbow area, but in most places the alteration and associated mineralization were restricted to the lower part of the tuff except where faults or open fractures cut the tuff. Where the rock is unfractured, the hydrothermal solutions apparently were unable to penetrate above a densely welded impermeable zone in the tuff. The altered rock beneath the impermeable zone is moderately to densely welded in most exposures, but it weathers easily and closely resembles nonwelded or partially welded tuff.

SUGGESTIONS FOR PROSPECTING

Future mining exploration in and adjacent to the project area, especially in the Silverbow area, should include sampling the lavas of intermediate composition wherever practicable. These rocks are generally the most intensely altered rocks in areas of hydrothermal alteration and, as stated previously, are the chief ore bearers at Goldfield and Tonopah. In the Antelope Springs area the possibility seems good that weakly mineralized welded tuffs at the surface are intruded by dacite and andesite at shallow depths. If the intrusive rocks are present, they may contain ore.

In mineralized areas adjacent to the Thirsty Canyon Tuff or alluvium-filled valleys, the possibility of buried deposits beneath the tuff or alluvium should be considered, inasmuch as these strata postdate most periods of hydrothermal activity (Anderson and others, 1965). The most likely areas for buried deposits appear to be (1) the area north of Stonewall Mountain bordering Stonewall Flat, (2) the area between Stonewall Mountain and Mount Helen, and (3) Stonewall Flat adjacent to the Cactus Range.

POTENTIAL SITES FOR UNDERGROUND NUCLEAR TESTING

The project area, exclusive of Pahute Mesa which is currently being used as a testing area, offers various sites that may be suitable for underground nuclear testing. The sites and the thicknesses and physical properties of available media are discussed below.

ALLUVIUM-FILLED BASINS

Most of the underground nuclear testing at Nevada Test Site has been in alluvium-filled basins. The basins in the area of study are very similar geologically to those in Nevada Test Site and should, therefore, provide suitable environments for testing. Of the four principal basins–Kawich Valley, Gold Flat, Cactus Flat, and Stonewall Flat–the first two appear to be most suitable because they are far removed from population centers and have the thickest deposits of alluvium. Gravity data (Healey and Miller, 1962, p. 20–22; C. H. Miller and D. L. Healey, oral commun., 1968) suggest 4,500 feet of alluvial fill and volcanic rocks in the deepest parts of both valleys and indicate sizable areas where the alluvium and volcanic rocks can be expected to be at least 2,000 feet thick.

As of June 1966 no drill holes had penetrated sufficiently deep to determine the thickness of alluvium and volcanic rocks in any of the valleys. The logs of several wells that have been drilled for water (pl. 1) very briefly describe the material penetrated (p. 84). The driller's log of Gold Flat well 1 indicates bedrock at from 400 feet to 486 feet, but this log may be erroneous inasmuch as it conflicts with gravity data and the geologic setting. It seems more likely that the driller reached well-cemented gravel rather than bedrock, but possibly the alluvium in this part of Gold Flat is thin. In general, the greatest thicknesses of alluvium within the valley areas probably occur within or adjacent to the lowest gravity contours along the east flanks of the valleys, but drill-hole data are needed for accurate thickness determination.

NONWELDED AND PARTIALLY WELDED ZEOLITIZED TUFF

Thick zones of nonwelded and partially welded zeolitized tuff have been used for contained explosions for many years at the Nevada Test Site. As much as 800 feet of such tuff (bedded) crops out along the west flank of the Belted Range east of Pahute Mesa in the vicinity of Quartet Dome (Sargent and others, 1966). Here the tuff is capped by 80 feet of the densely welded ash-flow tuff of the Grouse Canyon Member of the Belted Range Tuff. The bedded tuff dips gently to the east, and it may thicken eastward. As much as 1,000 feet of cover could be obtained by driving tunnels 3,000 feet eastward from an elevation of 6,200 feet approximately 2 miles south-southwest of Quartet Dome (K. A. Sargent, written commun., 1963).

Thick sections of zeolitized nonwelded and partially welded tuff have been penetrated by the drill beneath the Timber Mountain Tuff in local areas on Pahute Mesa. The possibility seems good that similar tuff in thinner beds is also present beneath the Timber Mountain Tuff along the south flank of Gold Flat north and northwest of Silent Canyon. In this area, the alluvium probably ranges in thickness from 0 to at least 1,000 feet and, excluding the western part of the flat, the first bedrock is probably either basalt, Thirsty Canyon Tuff, or Timber Mountain Tuff. Along the west and southwest flanks of Gold Flat, lavas and tuffs that are older than the Timber Mountain Tuff crop out. These outcrops indicate that the maximum width of ground that can be inferred to be underlain by Timber Mountain Tuff and thick alluvium in Gold Flat is less than 5 miles measured east to west. Most of the alluvium above the volcanic rocks is probably poorly lithified and rich in boulders; therefore, drill holes will undoubtedly meet with lost circulation zones in the alluvial section. The volcanic rocks beneath the alluvium are probably no more fractured and faulted than those on Pahute Mesa; they should drill easily and provide tight impermeable zones suitable for test media. Drill holes would be required for accurate thicknesses of the total alluvium-volcanic rock section in this area.

WELDED TUFF

Thick sections of densely welded tuff are probably present at various depths in all the basin areas, and in a few areas they may be sufficiently homogeneous and impermeable to be potentially valuable as test media. The most favorable areas for such occurrences appear to be the west flank of Kawich Valley east of Trailer Pass and north of Gold Reed, in Tps. 3 and 4 S., R. 51 E., and in Gold Flat adjacent to the Cathedral Ridge caldera. In both areas several thousand feet of Fraction Tuff probably underlies the sedimentary rocks of Tertiary age and the alluvium of Cenozoic age. Depending on the distance outward from the range, the tuff could be expected to occur at depths from zero to several thousand feet. Drilling would be required to determine the depth to ground water and the occurrence of impermeable zones suitable for test media. Physical properties of Fraction Tuff and of two samples of subjacent lava of intermediate composition are given in table 10.

BASEMENT ROCKS AND GRANITE

The Trappman Hills are a structural horst formed of gneiss and schist of Precambrian age. The gneiss (gneissic quartz monzonite) forms the bulk of the outcrops in the hills and intrudes the schist. Both rock types are well foliated and intensely fractured; locally they are hydrothermally altered. The rocks are cut by numerous quartz, aplite, and pegmatite veins as well as by several rhyolite dikes of Tertiary age. The schist is characterized by abundant femic minerals and relatively high dry bulk density (2.8 grams per cubic centimeter, E. F. Monk, written commun., 1963); the gneissic quartz monzonite is characterized by relatively sparse femic minerals and lower dry bulk density (2.6 g/cc). In several areas gneissic quartz monzonite forms the only visible outcrops and probably forms a homogeneous medium at depth suitable for some types of nuclear tests.

Granite of probable Mesozoic age in T. 3 S., R 46 E., in the Cactus Range along the northeast flank of Stonewall Flat, and southeast of Quartzite Mountain in the southern Kawich Range about 4½ miles west-northwest of the Floyd Lamb water well provides relatively unfractured media similar to the quartz monzonite in the Climax stock at Nevada Test Site.

MEASURED SECTIONS

Tuff of White Blotch Spring

[Outside mapped area, about 1 mile north of White Blotch Spring in NE¼ T. 5 S., R. 53 E., White Blotch Spring quadrangle, Nye County, Nev. Measured by E. B. Ekren, C. L. Rogers, Theodore Botinelly, and J. E. Weir, April 1963]

	Unit thickness (feet)	Accumulated thickness (feet)
Top of section is fault contact with younger ash-flow tuff.		
Tuff of White Blotch Spring (incomplete):		
Ash-flow tuff, weathers white; moderately welded. Contains abundant white fiamme as much as 2 in. long, sparse biotite as sole mafite, 30 percent phenocrysts consisting of 40 percent quartz, 40 percent sanidine, and 20 percent andesine. (A comparison of this zone with outcrops where top of White Blotch Spring is preserved indicates no more than a few tens of feet are cut out here by fault.) Forms bench	50	895
Ash-flow tuff, light-gray, densely welded, crumbly weathering; vague flow layering with layers about 3 ft thick. Contains fairly abundant lithic fragments of shard-rich gray welded tuff about 1 in. in diameter, abundant white fiamme, same phenocrysts as above, sparse biotite. Forms bench above cliff. Thickness of unit, approximate	120	845
Ash-flow tuff, medium-gray, weathers brownish gray; otherwise similar to above but fiamme are very sparse and groundmass is mostly vitric. Contains 24 percent phenocrysts consisting of 33 percent quartz, 42 percent sanidine, 24 percent plagioclase, and 1 percent biotite. Forms upper part of cliff	30	725

Tuff of White Blotch Spring—Continued

	Unit thickness (feet)	Accumulated thickness (feet)
Ash-flow tuff, light-gray, weathers light reddish brown, densely welded; groundmass partly vitric. Contains small white fiamme and 32 percent phenocrysts consisting of 40 percent quartz, 43 percent sanidine, and 17 percent plagioclase. Forms steep jointed cliff	160	695
Ash-flow tuff, light-gray, weathers light reddish brown. Basal less-welded slope-forming part of same zone as above, but is richer in biotite and plagioclase; no apparent cooling break. Contains sparse small andesite lithic fragments, 34 percent phenocrysts consisting of 36 percent quartz, 20 percent sanidine, 33 percent plagioclase, and 11 percent biotite	30	535
Ash-flow tuff, very light gray, densely welded, fiamme-poor, shard-rich. Groundmass vitric to weakly devitrified. Unit contains 35 percent phenocrysts consisting of 41 percent quartz, 36 percent anidine, 21 percent plagioclase, 2 percent biotite. Forms slope	100	505
Ash-flow tuff, light-gray, weathers to alternating light-, medium-, and dark-brown zones. Pumice lapilli very sparse, eutaxitic structure vague, groundmass vitric to weakly devitrified. Contains 30 percent phenocrysts consisting of 45 percent quartz, 30 percent sanidine, 18 percent plagioclase, 7 percent biotite. Forms cliff that is columnar jointed at base	70	405
Ash-flow tuff, tan, pale-pink, green, and gray, biotite-rich, weakly to moderately welded. Contains abundant light-greenish rhyolite fragments and flattened pumice lapilli and blocks as much as 4 in. long; at about 150 ft above base lithic fragments diminish. Unit forms slope. Mostly covered, thickness approximate	335	335
Total incomplete tuff of White Blotch Spring	895	
Shingle Pass Tuff (incomplete):		
Ash-flow tuff, vitrophyre, black. Contains 23 percent phenocrysts of plagioclase, hornblende, biotite, and magnetite. Unit appears to be vitrophyre at top of flow that is very poorly exposed but may be basal part of flow extending into covered zone above	20	

Fraction Tuff

About 1½ miles south of Trailer Pass in N½ T. 4 S., Rs. 50, 51 E., Quartzite Mountain quardrangle, Nye County, Nev. Measured by E. B. Ekren and C. L. Rogers, May 23, 1963]

	Unit thickness (feet)	*Accumulated thickness (feet)*
Top of section represented by a poorly exposed contact with overlying dacitic debris flow.		
Fraction Tuff:		
Ash-flow tuff, white, weathers white; vitric, weakly welded. Contains phenocrysts of plagioclase, alkali feldspar, quartz, biotite, hornblende, sphene, and magnetite. Gently sloping surface. Generally poorly exposed	1, 300	7, 220
Ash-flow tuff, light-pinkish-gray, weathers brownish gray, vitric, moderately welded. Contains fairly abundant white flattened pumice lapilli, plagioclase, quartz, alkali feldspar, biotite, and abundant hornblende. Weathers to rubbly slope, forms gentle hills. Contact gradational with units above and below	600	5, 920
Ash-flow tuff, dark-brownish-gray at base grading upward to light-pinkish-gray at top; white flattened pumice lapilli, vitric, moderately welded. Plagioclase is dominant phenocryst, quartz and alkali feldspar also present; abundant biotite, very little hornblende, no sphene	700	5, 320
Vitrophyre of ash-flow tuff, brown-black; same mineralogy as unit above	30	4, 620
Ash-flow tuff, brown-gray; same mineralogy as unit above, very rich in lithic fragments, including dacite, schist, and quartzite, the latter in boulders as much as 3 ft in diameter; unit is moderately welded, vitric	300	4, 590
Ash-flow tuff, light-brownish-gray, lithic-rich, weakly welded; approximately equal amounts of plagioclase, alkali feldspar, and quartz, with minor biotite and no hornblende or sphene. Zone of hydrothermal alteration may indicate fault contact with unit above	100	4, 290
Ash-flow tuff, light-brownish-gray, weathers dark brownish gray, moderately welded, lithic-rich; same mineralogy as unit above	800	4, 190
Ash-flow tuff, weathers brown, densely welded. Phenocrysts are quartz, alkali feldspar, plagioclase, abundant biotite, very sparse hornblende. Forms rugged hills above two vitrophyre zones that show as two dark conspicuous bands on arerial photographs	700	3, 390
Ash-flow tuff. Interval includes two vitrophyres about 30 ft thick with about 120 ft of intervening devitrified welded tuff; upper vitrophyre apparently is base of of ash flow described above; lower vitrophyre is base of 150-ft-thick welded tuff. Unit contains abundant biotite; phenocrysts consist of plagioclase (dominant), quartz, and alkali feldspar	180	2, 690

Fraction Tuff—Continued

	Unit thickness (feet)	*Accumulated thickness (feet)*
Ash-flow tuff, light-buff, weathers brown, densely welded, partly devitrified. Contains small amount of biotite, no hornblende or sphene; quartz dominant phenocryst, alkali feldspar exceeds plagioclase; contains a few dacite lithic fragments. Forms steep jointed cliffs that weather to spires; columnar jointed at base. Thickness, approximate	1, 100	2, 510
Ash-flow tuff, discontinuous. Basal 35 ft is black vitrophyre; above the vitrophyre the unit is light gray and weathers gray brown. Unit is partly devitrified, forms stepped cliff near base grading upward to slabby- and crumbly-weathering slope former. Same mineralogy as unit above	200	1, 410
Ash-flow tuff, some lenses of vitrophyre at base, light-gray at base, light-violet-gray at top. Crystal rich with same mineralogy as unit above	70	1, 210
Ash-flow tuff, densely welded. At base, black vitrophyre pods, as much as 10 ft thick; above base, unit is gray green and devitrified. Phenocrysts: quartz (dominant), alkali feldspar and plagioclase about equal, small amount of biotite	40	1, 140
Ash-flow tuff. At base, discontinuous dark-brownish-gray to black vitrophyre as much as 40 ft thick; above base, unit is largely devitrified. Unit contains sparse hornblende and biotite, very minor sphene; quartz, plagioclase, and alkali feldspar in about equal amounts, abundant dacite lithic fragments	230	1, 100
Ash-flow tuff, olive-gray, hackly-weathering, densely welded at base of exposure; grades upward within about 20 ft to black vitrophyre as much as 50 ft thick; rock above vitrophyre same lithology as rock below; some biotite, no hornblende, minor sphene; quartz, plagioclase, alkali feldspar; no visible ash fall at base of unit in vicinity of section, but in exposures to north about 50 ft of stratified yellow tuff crops out at base	120	870
Ash-flow tuff, light-red to light-reddish-gray, densely welded, devitrified, irregular slabby weathering. Contains abundant flattened light-gray to yellow-gray pumice; biotite abundant, some hornblende and large sphene; plagioclase, quartz, alkali feldspar. Unit forms gentle slopes and hummocky hills in line of section; to north weathers to small rounded "haystack" hills; lower 5 ft crystal poor and locally silicified	750	75
Total Fraction Tuff	7, 220	
Lavas of intermediate composition (incomplete):		
Latite or rhyodacite, light-gray to yellowish-green, flow-layered, crystal-rich; plagioclase, quartz, biotite, and hornblende. Base not exposed	40	

DRILLERS' LOGS OF WATER WELLS IN MAPPED AREA

Drillers' logs throughout this section are quoted verbatim.

Gold Flat 1

[Well drilled by E. C. Ferguson, Tonopah, Nev. Casing 6 in. in diameter to 390-ft (?) depth]

Material	Depth (feet)	Thickness (feet)
Gravelly sand	40	40
Whitish soil	200	160
Reddish soil	255	55
Gravelly sand	260	5
Whitish soil	390	130
(Not logged)	400	10
Bedrock	486	86

Gold Flat 2

[Well drilled by E. C. Ferguson, Tonopah, Nev. Casing 8¼ in. in diameter to 290-ft depth, perforated 225-290 ft. Could not lower water level with 5-in.-diameter bailer 20 ft long]

Material	Depth (feet)	Thickness (feet)
Soft volcanic gravel	90	90
Slightly harder volcanic gravel	200	110
Reddish volcanic gravel, water	250	50
Blackish volcanic gravel	290	40

Sandia 1

[Well drilled by S. R. McKinney and Son, Las Vegas, Nev. Uncased test hole. Bailed 330 gal of water from zone at 45 ft, remainder of well dry]

Material	Depth (feet)	Thickness (feet)
Loose sand and gravel	3	3
Cement gravel and boulders	28	25
Black, pink, and white cemented gravel	35	7
Granite clay	45	10
Cement gravel with clay	50	5
Cement gravel	55	5
Cement gravel and clay	60	5
Brown cemented gravel	75	15
Black, red, and white cement gravel	100	25
Brown cemented gravel	110	10
Pink, white, and brown cemented gravel	130	20
Gray and white cemented gravel	160	30
Brown cemented gravel and sand	175	15
Gray cemented gravel and sand	220	45
Gray clay and gravel	260	40
Gray clay and sand	325	65
Gray clay	435	110
Gray and green clay	440	5
Green clay	465	25
Gray and green and gray and black shale	470	5
Blue shale	475	5
Blue and green shale	480	5
Gray limestone	485	5
Dark-gray shale	535	50
Brown shale	541	6
Brown and gray shale	548	7
Blue shale	588	40
Blue and brown shale with limestone shelves	593	5
Blue shale	600	7

Sandia 2

[Well drilled by S. R. McKinney and Son, Las Vegas, Nev. Casing 8 in. in diameter to 525-ft depth; perforated 325-485 ft. Well yielded 15 gpm with 142-ft drawdown during a test of unspecified length of time]

Material	Depth (feet)	Thickness (feet)
Gravel and clay	4	4
Brown cement gravel	20	16
Red, brown, and white cement gravel	35	15
Brown cement gravel	45	10
Brown and white cement gravel	70	25
Brown, red, and white cement gravel	99	29
Brown and gray gravel, hard	110	11
Brown and gray cement gravel	125	15
Brown, gray, and red cement gravel	170	45
Brown, gray, and red cement and sand	215	45
Brown, gray, and white cement gravel	280	65
Brown and white cement gravel	327	47
Brown, white, and pink cement gravel	346	19
Brown, white, and pink cement gravel and sand	357	11
Brown, white, and pink cement gravel and clay	362	5
Cement gravel, water	367	5
Brown, red, and white cement gravel	387	20
Brown, red, and white cement gravel, water	392	5
Brown, red, and white cement gravel	397	5
Cement gravel and clay	407	10
Brown and white cement gravel	417	10
Brown and white cement gravel and clay	437	20
Yellow clay	442	5
Yellow clay and gravel	452	10
Brown cement gravel, hard	455	3
Muddy clay	467	12
Clay and gravel, water	477	10
Gray sand and clay	525	48

Sandia 3

No log available.

Sandia 4

[Well drilled by S. R. McKinney and Son, Las Vegas, Nev. Casing 8 in. in diameter to 380-ft depth, perforated 351-466 ft. Well yielded 6 gpm with 67-ft drawdown during a test of unspecified length of time]

Material	Depth (feet)	Thickness (feet)
Soil and gravel	5	5
Gravel	15	10
Cement gravel and boulders	65	50
Gravel	74	9
Cement gravel and boulders	81	7
Cement gravel	95	14
Cement gravel and boulders	130	35
Gravel and little clay	134	4
Cement gravel	196	62
Gravel	201	5
Cement gravel	225	24
Gravel and sand	230	5
Cement gravel and sand	260	30
Cemented sand	347	87
Cemented sand and gravel	362	15

Sandia 4—Continued

Material	Depth (feet)	Thickness (feet)
Sand, gravel, and clay; lost mud	367	5
Cemented sand	397	30
Cemented sand, little clay, little water	402	5
Cemented sand, little water	412	10
Cemented sand	422	10
Cemented sand and clay	437	15
Sandy clay, little gravel	442	5
Sandy clay	477	35
Gray sandy clay	497	20
Gray clay, muddy	546	49
Pink clay	580	34

Sandia 5

[Well drilled by Perry Bros., Flagstaff, Ariz.]

Material	Depth (feet)
Sandy clay	0–18
Sand and gravel	19–30
Brown clay	30–36
Fine sand	36–51
Sandy clay and gravel	51–82
Sand and gravel	82–125
Brown sandy clay	125–132
Sandy clay with streaks of semicemented gravel	132–180
Sand and gravel	180–280
Black volcanic rock	280–300

Test pumping results: Pump setting 210 ft to end of suction. At start 11 gpm, water level 210 ft. Surging and backwashing increased yield to 15 gpm. At completion of test, yield had increased to 16 gpm. After setting overnight, well would pump 20 gpm for 10 hours then drop off to 16 gpm. Maximum yield was from 210-ft level.

Sandia 6

[Well drilled by Perry Bros., Flagstaff, Ariz.]

Material	Depth (feet)
Soil	0–2
Boulders	2–34
Large gravel and boulders	34–60
Large loose gravel	60–140
Gravel and brown sandy clay	140–245
Cemented gravel (small)	245–305
Sand and gravel	305–415
Sandy clay with streaks of sand	415–510
Gray clay	510–525
White soapstone	525–568
White and blue shale with limestone streaks	568–620
Green shale with limestone streaks	620–705
White sandy clay	705–755
Dark-gray and black clay	755–790

Test pumping results: Pump setting at 510 ft; water level 351 ft. Ninety gpm resulted in 75-ft drawdown to 426 ft; 70 gpm resulted in 74-ft drawdown to 425 ft. Pumped 48 hours at 90 gpm.

Sandia 7

[Well drilled by Perry Bros., Flagstaff, Ariz.]

Material	Depth (feet)
Blow sand	0–4
Sand and sandy clay	4–110
Sand and large gravel	110–120
Sand and small gravel (water level 137 ft)	120–280
Gravel, sand, and clay	280–305

Test pumping results: Static water level 137 ft; 125 gpm with 23-ft drawdown; 150 gpm with 38-ft drawdown.

Sandia 8

[Well drilled by Perry Bros., Flagstaff, Ariz. Casing 8 in. in diameter to 793-ft depth, perforated 368–773 ft. Well yielded 122 gpm with 100-ft drawdown during test of unspecified length of time]

Material	Depth (feet)	Thickness (feet)
Brown clay with gravel and sand	250	250
Basalt sill or cemented gravel	268	18
Clay, gravel, and sand with basalt stringers	328	60
Sandy clay	483	155
Clay, gravel, and sand	710	227
Sand with clay	793	83

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