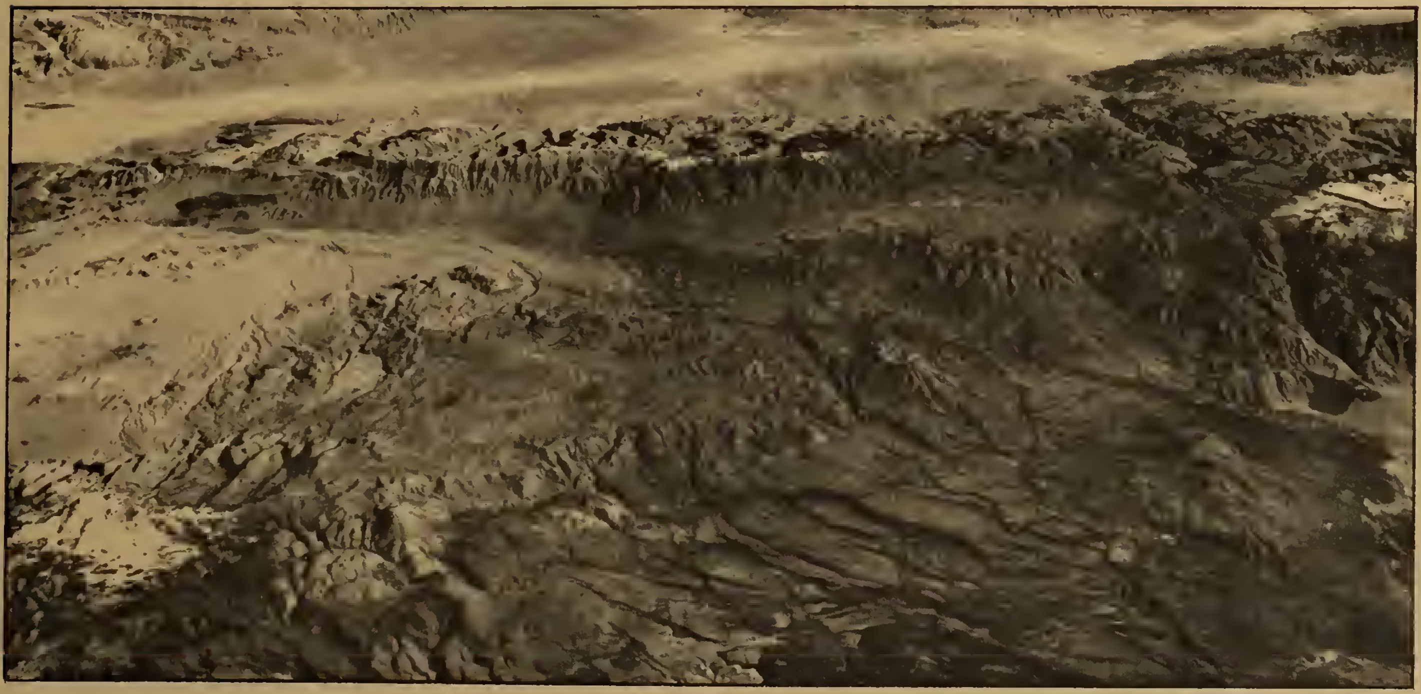
FIELD TRIP NO. 1

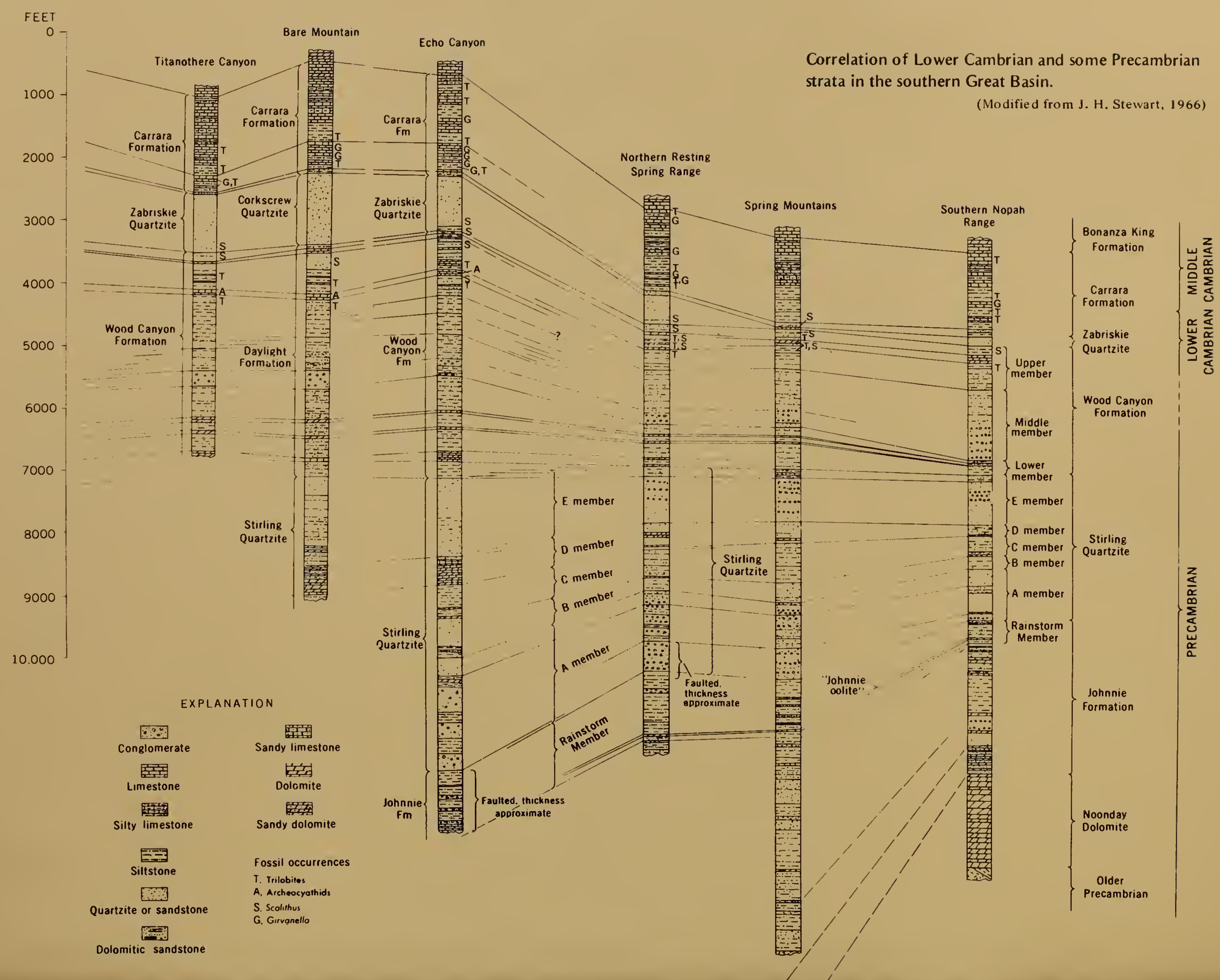
GUIDEBOOK:
DEATH VALLEY REGION, CALIFORNIA AND NEVADA



THE GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
70th Annual Meeting
March 29-31, 1974
Las Vegas, Nevada
·1888·

Correlation of Lower Cambrian and some Precambrian strata in the southern Great Basin.

(Modified from J. H. Stewart, 1966)



GUIDEBOOK:

DEATH VALLEY REGION, CALIFORNIA AND NEVADA

PREPARED FOR THE 70th ANNUAL MEETING OF THE CORDILLERAN SECTION
THE GEOLOGICAL SOCIETY OF AMERICA
FIELD TRIP NUMBER 1

CONTENTS

Published by
The Death Valley Publishing Company
Shoshone, California 92384

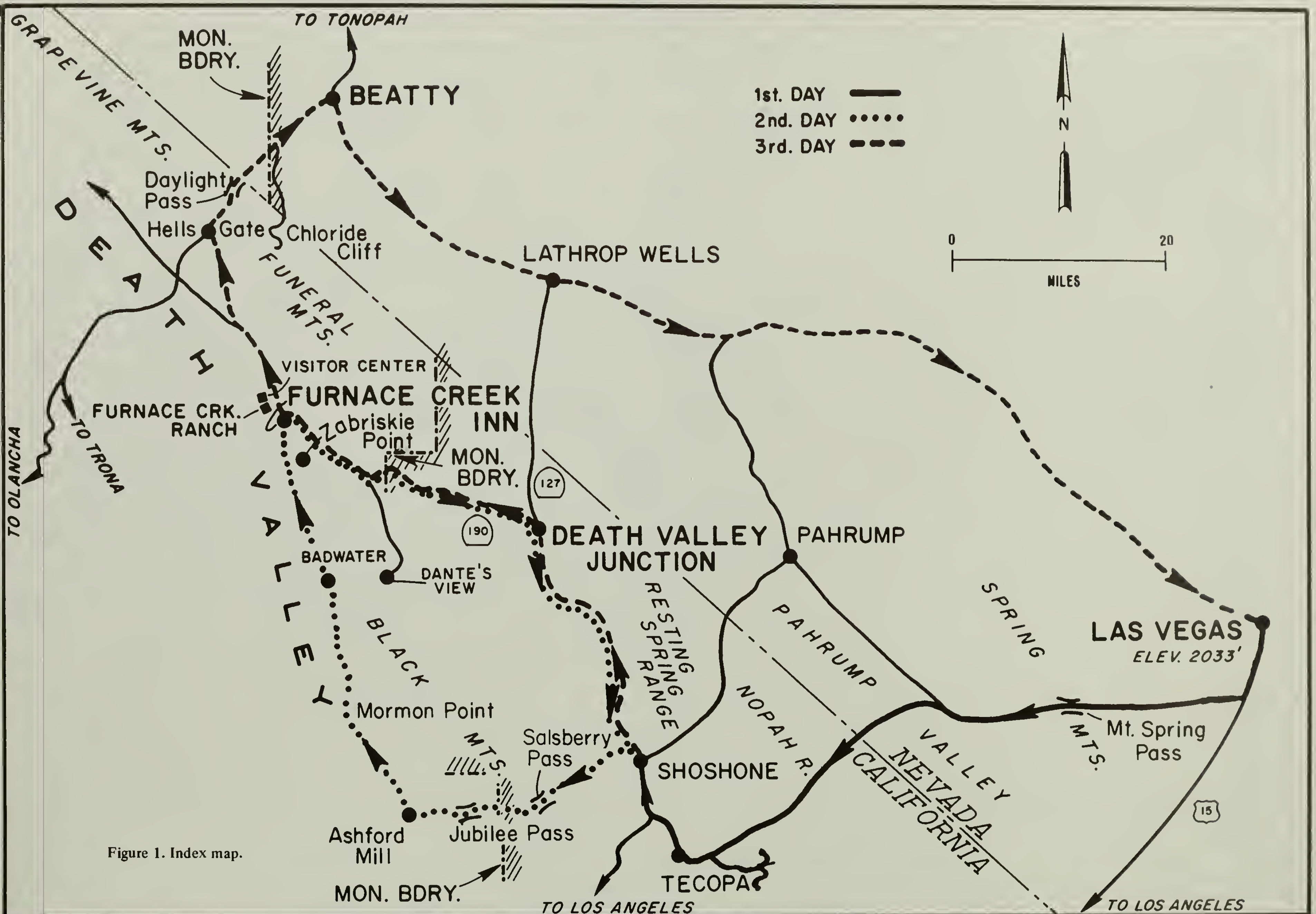


Figure 1. Index map.

Geologic Guide to the Death Valley Region, California and Nevada

Bennie W. Troxel[1]

INTRODUCTION

This guidebook was prepared for use on a 3-day bus trip of the Death Valley region, that starts and ends in Las Vegas, Nevada. The second and third day's trips begin in Shoshone, California. The guide may be useful to others who wish to follow selected segments of the trip or to review in more detail some of the geologic features of part of the Death Valley region which are highlighted in this guide. The guidebook can and should be supplemented by additional reports and maps. General papers and selected papers pertaining to areas along the route are included in the references.

The guidebook consists of a general guide (Part I) and preprints of several new papers, as well as reprints of others, that pertain to the Death Valley and Las Vegas regions (Part II).

Information for the general guide is derived from the published sources listed in the references, personal observations, and work in progress by the authors of the papers in Part II. Unfortunately, specific citations to individual works are lacking in many parts of the guidebook, especially to work underway.

Features most obvious from the roads are emphasized in this guide. Although no detailed road maps are provided, care has been taken to use reference points that are marked by signs on the road or are on common road maps.

On the trip, information will be provided along the route, as noted below, by:

NAME	AREA	SUBJECT
B. W. Troxel (general leader)		Geology along route
L. A. Wright (general leader)		Geology along route
B. C. Burchfiel	Las Vegas–Pahrump Valley	Geology along route
G. A. Davis	Las Vegas–Pahrump Valley	Geology along route
P. E. Diehl	Nopah Range	Wood Canyon Formation
L. A. Wright	Nopah Range	Noonday Dolomite
M. T. Roberts	Alexander Hills	Crystal Spring Formation
J. K. Otton	Black Mountains (vic. Mormon Point, Copper Canyon)	Structure and Precambrian rocks
J. F. McAllister	Furnace Creek	Structure; Paleozoic and Tertiary stratigraphy
M. W. Reynolds	Grapevine Mountains	Structure; Paleozoic and Tertiary stratigraphy
B. C. Burchfiel	Lathrop Wells–Las Vegas	Geology along route
G. A. Davis	Lathrop Wells–Las Vegas	Geology along route

[1] *Geological Society of America, Boulder, Colorado 80301.*

LAS VEGAS, NEVADA, TO NOPAH RANGE, CALIFORNIA

The rocks exposed in the steep east face of the Spring Mountains and visible from a distance of many miles consist of gray Cambrian and younger Paleozoic rocks in the crest of the range and bright red- and cream-colored Mesozoic sandstone in the cliffs.

The Cambrian rocks were thrust over the Mesozoic strata along the Keystone thrust fault, a major fault that can be traced easily along the top of the distinctively colored rocks in the east front of the range (see Burchfiel and others in Part II for the geology of the Spring Mountains).

West from the Las Vegas–Los Angeles highway (U.S. 15), the road to Pahrump passes through low hills made up of gently to moderately folded sedimentary rocks of late Paleozoic and Mesozoic age. Some of the rocks are an important source of gypsum, which is mined and made into various products at a large mine and plant situated north of the road to Pahrump.

Farther west, large masses of monolithologic breccia of gray carbonate rock crop out near the highway. Still farther west, the road crosses the Spring Mountains and extends into Pahrump Valley (Figs. 1 and 2).

The Keystone thrust fault is not exposed in roadcuts, but it can be observed in a small canyon north of the highway a few hundred yards east of Mountain Spring Pass, in the Spring Mountains.

It is easy to observe from points along the highway west of the pass that the rocks in the west part of the mountains are folded.

West of the Spring Mountains is Pahrump Valley, which has been developed in the last decade into farms whose main crop is cotton. The Pahrump Valley is bounded on the west by the Nopah Range.

NOPAH RANGE

The Nopah Range contains a very complete and well-exposed section of Precambrian and Paleozoic sedimentary rocks. The rocks in it dip eastward and are cut by many west-dipping faults. The oldest rocks are Precambrian gneiss at the southwest end of the range, whereas the youngest underlie a thrust fault at the northeast end of the range, where Cambrian and late Precambrian strata have been thrust over folded upper Paleozoic strata. Wright (1973) has described the geology and prepared a map of the southern part of the Nopah Range and the Alexander Hills.

At Emigrant Pass, where the county highway from Pahrump Valley to Tecopa crosses the Nopah Range, well-cemented coarse Cenozoic fanglomerate rests on Cambrian carbonate rocks of the Bonanza King Formation (Fig. 3).

At the first turnoff on the left (south) side of the road, a few hundred feet below the pass, there are several interesting features. The crest of the range north and south of the pass is underlain by east-dipping strata of the Bonanza King Formation. Underneath that is the Carrara Formation, which consists of an upper portion consisting of well-layered brown carbonate rocks, exposed at the bases of cliffs, and a lower silty and shaly portion downslope from the cliffs. Within the Carrara Formation are algal-rich girvanella carbonate beds that form small ridges and a green shale unit that yields good fragments of several trilobite species. They can be found in the rocks adjacent to the graded parking area at the turnoff.

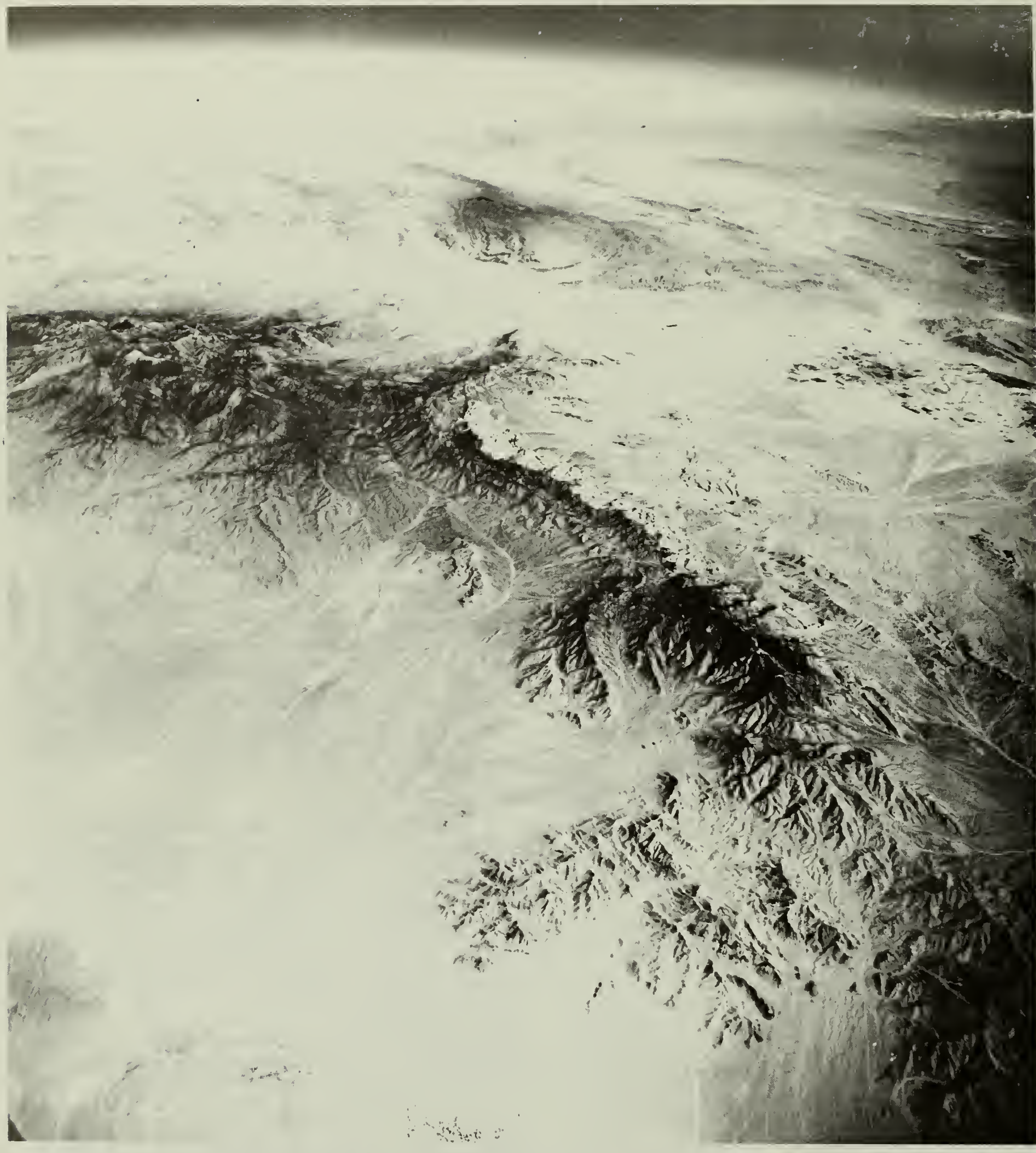


Figure 2. Oblique view northeast across Spring Mountains. Keystone thrust fault overlies pale-colored rocks on east side of mountains. Pahrump Valley in foreground. Las Vegas in valley on right side of photo. Route of trip crosses lower (near) end of pale rocks beneath thrust fault. U.S. Geological Survey photograph. U.S. Geological Survey – U.S. Air Force photograph.

To the north are several obvious low-angle normal faults that dip to the west, drop the overlying rocks west, and rotate the rocks downward to the east.

Below the Carrara Formation lies the Zabriskie Quartzite. It crops out in several places along or near the road west of the trilobite locality. The quartzite is recognized by its small, rugged outcrops, by its lavender-to-pink color, and locally, near its base, by the presence of locally abundant *Scolithus* tubes approximately 0.25 to 0.5 in. in diameter.

Below the Zabriskie is the Wood Canyon Formation, which is discussed in detail by Diehl in Part II.

West from Emigrant Pass, the road is parallel and north of the southwestern segment of the Nopah Range. In the afternoon sunlight one can easily recognize the following east-dipping Cambrian and older sedimentary units from east to west: Bonanza King Formation (dark to pale gray, striped; forms prominent outcrops in high parts of range), Carrara Formation (brown to gray and pale green; usually exposed in faces of cliffs), Zabriskie Quartzite (pale pinkish gray; forms ridges), Wood Canyon Formation (dark, reddish to lavender; mostly fine- to coarse-grained quartzite; moderately subdued topography with rounded slopes), Stirling Quartzite (upper part, pale pink; middle part, dark lavender; lower part, pale pink; forms regular slopes and ridges), Johnnie Formation (mostly siltstone and quartzite with thin beds of dolomite; forms subdued topography, generally brown colored from a distance), and Noonday Dolomite (prominent pale-tan, two-toned bold outcrops) in the southwest part of the range.

The Noonday Dolomite lies on various units of Precambrian rocks and is separated from them by a regional unconformity. In the southern Nopah Range and the Alexander Hills farther southwest, the Noonday Dolomite rests, from north to south, on successively younger rocks: Precambrian gneiss and schist, Crystal Spring Formation, Beck Spring Dolomite, and Kingston Peak Formation. At places a few tens of miles even farther south and southwest, the Noonday appears to be conformable upon the Kingston Peak Formation. Both tilting and downdropping appear to have occurred in Kingston Peak to Noonday time. (See the article by Wright and others in Part II.)

At the south end of the band of Noonday Dolomite, which crosses the Nopah Range, there are good exposures of the sedimentary features within the Noonday. A narrow paved road to a mine on the south side of the range crosses outcrops of the Noonday Dolomite. On the right (east) side of the road, just before the road crosses the Noonday, there is a small hill underlain by the lower part of the Noonday, which is an excellent site to see details of the Noonday Dolomite (see sketch by Wright and others and article by Williams and others in Part II). Farther east is the well-exposed section described by Hazzard (1937), well worth the review by those interested in Noonday, Johnnie, Stirling, Wood Canyon, and Zabriskie rocks. This section has become the standard reference section for the southern Death Valley region since Hazzard measured and described it over three decades ago.

Figure 3. Vertical air photograph of southern Nopah Range. Emigrant Pass near lower end of straight segment of road on right (east) side of Range. Alexander Hills along lower edge of photo. U.S. Geological Survey – U.S. Air Force photograph.

ALEXANDER HILLS

The Alexander Hills, southwest of the southern Nopah Range, contains a complete section of east-dipping sedimentary rocks below the Wood Canyon Formation. The west edge of the hills has small exposures of Precambrian gneiss, which is overlain by the Crystal Spring Formation. The Crystal Spring has been intruded by a large, thick sill of diabase, which, during its intrusion, caused the formation of talc bodies where it came into contact with dolomite and limestone. One of the larger talc mines of the Death Valley region is the Western talc mine in the Alexander Hills. Above the Crystal Spring Formation is the Beck Spring Dolomite, which forms a prominent gray band across the Alexander Hills. On the north side of the Alexander Hills it is overlain by the Noonday Dolomite. Only about 1 mi farther south, however, a complete section of the Kingston Peak Formation is preserved between the Beck Spring and Noonday Dolomite. Successively above the Noonday (eastward) are the Johnnie Formation, the Stirling Quartzite, and the lower part of the Wood Canyon Formation. These are more easily accessible but not as well exposed as the sections in the southern Nopah Range.

West of the Western talc mine in the Alexander Hills are good exposures of the lower units of the Crystal Spring

Formation. The formation is the subject of an article by Roberts in Part II.

NOPAH RANGE TO SHOSHONE

Precambrian metamorphic rocks crop out in an area a few square miles in extent in the southern Nopah Range. They are easily accessible from the road along the south side of the range. The rocks have been divided into several types by Wright (1974).

West from the Nopah Range, in the intersection of several valleys, there are very young lake beds that were deposited in an area several tens of square miles in extent, which were later dissected. The beds have yielded pumicite, clay, and borates, and contain other minerals of economic interest.

From Shoshone, the Charles Brown State Highway leads eastward, and about 4 mi east it crosses the Resting Springs Range. On the west side of the range, a few hundred feet northwest of the pass, there is a remarkable roadcut in Tertiary rocks, where many interesting geologic features are exposed. The parking area on the right side of the road is underlain by well-cemented breccia composed of carbonate clasts. It is also exposed in small patches at the base of the roadcut. The breccia is overlain conformably by volcanic-rich sedimentary rocks, ash flows, and altered ash. An unconformable extrusive flow with a black glassy center forms the north half of the roadcut exposure. All of the rocks are cut by small normal faults.

SHOSHONE TO SALSBERRY PASS

From an intersection 1 mi north of Shoshone, U.S. Highway 178 leads west through the Greenwater Range and Dublin Hills, Greenwater Valley, and the Black Mountains via Salsberry Pass and Jubilee Pass to Death Valley.

The road first passes along the north side of the Dublin Hills, which contains Cambrian and late Precambrian strata overlain by a succession of Tertiary rhyolite in flows. The rocks are cut by low-angle normal faults (Chesterman, 1973). North of the road is the Greenwater Range, the southern part of which contains Tertiary granite, Tertiary rhyolite, and Tertiary to Quaternary basalt. (See the article by Haefner in Part II.) Some of the west-dipping faults that cut the rocks can be seen from the highway.

Beyond Greenwater Valley, and well exposed at Salsberry Pass, are rocks of a Tertiary volcanic and sedimentary sequence that extends for many miles northward in the Black Mountains. Nearly everywhere these rocks have been cut by numerous normal faults, which causes them to have a patchwork pattern. Thus they were given the informal name of Calico volcanic rocks by Levi Noble (1941).

The southern limit of these and other volcanic rocks in the Black Mountains is formed by the Sheepshead fault, a major fault that extends from the vicinity of the southern Nopah Range to a point near the west edge of the Black Mountains several miles northwest of Salsberry Pass. Segments of the fault have been inactive since the time of volcanism as evidenced by rhyolite that has intruded parts of the fault zone. The road crosses a poorly exposed segment of the fault about 1 mi southwest of the pass.

SALSBERRY PASS TO ASHFORD MILL SITE

West from Salsberry Pass, the road follows the upper part of a broad drainage channel that drains into Death Valley. The drainage channel was the path of a flash flood caused by an intense rainstorm of a few hours' duration in October 1973, which washed out several miles of road.

Near the Death Valley National Monument boundary west of Salsberry Pass is a prominent small peak on the right side of the road, which is underlain by Precambrian gneiss. Deep-red east-dipping beds of Tertiary conglomerate crop out on each side of the road at the monument boundary. The conglomeratic rocks are seemingly older than the volcanic rocks at Salsberry Pass, since they are devoid of any clasts of the volcanic rock. An early to middle Tertiary age is most likely for them but is unconfirmed. The nearby outcrops south of the road at this point are deeply weathered and red-stained Precambrian crystalline rocks that are overlain by red-stained conglomerate that contains interbedded monolithologic breccia and andesite. The andesite has a trachytic texture and occurs throughout most of the southern Death Valley region but seldom in large masses.

West of the monument boundary, the road traverses the area where Levi Noble identified highly broken rock units that he named "chaos," in his classic paper published in 1941. Noble defined chaos as a product of brecciation by thrust faulting on the Amargosa thrust fault, and he recognized three episodes of faulting, each of which produced a phase of the chaos. He termed them the Jubilee phase, the Virgin Spring phase, and the Calico phase. Work begun with him by L. A. Wright and me in the early 1950s led the three of us to propose an origin for these rocks of gravity faulting rather than thrust faulting. (During his last few years of field work, Noble had accepted many aspects of this concept of gravity faulting to produce the chaos.)

Thus, the Amargosa fault is no longer identified as a continuous plane that developed as a thrust fault; rather, it is a plane or a series of planes that commonly occurs between Precambrian metamorphic rocks and the overlying younger rocks, which range in age from late Precambrian to Tertiary. Noble's Virgin Spring and Calico phases are products of normal faulting, but the Jubilee phase is composed mainly of megabreccia and is of sedimentary origin.

Between approximately 1 and 3 mi west of the monument boundary, many small prominent hills are evident near the highway. Most of these are isolated remnants of late Precambrian rocks and lie above faults. The prominent steep hill just south of the road and at the point where the road turns abruptly south about 3 mi from the monument boundary has an excellent north-facing exposure. The features exposed in the hill slope are typical of the features observed in the chaos. In the lower part of the exposure are thoroughly shattered Precambrian crystalline rocks. These are overlain in fault contact first by sedimentary rocks of the Crystal Spring Formation and then by the Noonday Dolomite and the Johnnie Formation. Not all units of the three formations are present, but those that are present are in their proper stratigraphic order and are greatly thinned down, brecciated, and bounded by faults on all sides. Some units are more steeply dipping than others (even overturned). Throughout this general area, the broken units of pre-Tertiary rocks in fault contact with the Precambrian metamorphic rocks are in similar condition—highly brecciated, greatly thinned down, but always in proper stratigraphic order. They commonly dip eastward.

About 3 mi farther west, the road rises out of the wash to a small pass (Jubilee Pass) after it has swung south and then northwest around a large mass of the Funeral Formation. The Funeral Formation here is composed primarily of coarse conglomerate and has a prominent basalt flow in it. At Jubilee Pass, the Funeral Formation contains megabreccia units composed of Precambrian gneiss derived from Jubilee Peak, the prominent peak south of the pass.

From the vicinity of Jubilee Pass there is a good view of the structural complexity in rocks to the northwest, where Levi Noble coined the term "chaos." The high dark-gray peak about 3 mi distant is Ashford Peak, which is along the axis of a feature named the Desert Hound anticline by Noble. The gray rocks in the peak are Precambrian gneiss and schist, locally containing abundant Precambrian pegmatite dikes. Left (southwest) of the peak in the main part of the range are bands and patches of rocks of contrasting colors, but they are in moderately small units of outcrop; farther southwest, each patch of rock is larger. The boundary between the crystalline rocks and the patches of differently colored rocks is the plane Noble named the Amargosa thrust fault, which here dips southwest. The rocks above (southwest) the fault are sedimentary strata of Crystal Spring Formation, Beck Spring Dolomite, Kingston Peak Formation, Noonday Dolomite, Johnnie Formation, Stirling Quartzite, Wood Canyon Formation, and Zabriskie Quartzite. These rocks are in their proper stratigraphic order, though not everywhere present nor continuous along strike. The blocks of rocks become progressively larger southwest and away (upward) from the fault zone.

The low, rugged isolated hills in the foreground beyond the outcrops of the Funeral Formation are underlain by a succession of Tertiary sedimentary rocks composed in large part of monolithologic breccia masses. The hills near the north side of the road about 1 mi west of Jubilee Pass are made up of a wide variety of rock types. One can distinguish in them many individual sheets, each of which is composed of a single type of brecciated rock. Noble originally proposed a tectonic origin for the brecciated rocks, but he later recognized their sedimentary origin. A common weathering phenomenon is the development of a cavernous surface on the brecciated rocks.

The prominent hills near the left side of the road below Jubilee Pass contain excellent exposures of east-dipping Tertiary conglomerate. Pale-colored tuff exposed in the south part of the hills is interbedded with the conglomerate. The conglomerate appears to be older than the volcanic rocks exposed at Salsberry Pass, since no clasts of that type of volcanic rocks have been noted in the conglomerate.

Rocks on the south side of the road at the foot of the range are east-dipping multicolored Crystal Spring and pale-gray Beck Spring Dolomite. North of the road are, from west to east, east-dipping Beck Spring Dolomite (gray), Kingston Peak Formation (red), Noonday Dolomite, and the Johnnie Formation. A west-trending fault must be projected approximately parallel to the road to explain the offset in strike projection of rocks on both sides of the road.

As the road emerges from the west front of the Black Mountains, the Owlshead Mountains loom ahead on the opposite side of Death Valley. The Owlshead is composed mainly of Mesozoic granite rocks that are overlain in many areas by Tertiary and Quaternary volcanic rocks. The northeast end of the Owlshead contains a nearly complete section of north-tilted Crystal Spring Formation intruded and locally metamorphosed by the Mesozoic rocks.

ASHFORD MILL SITE TO MORMON POINT

The floor of Death Valley in this area is interrupted by a long, low, barren ridge called the Confidence Hills, which extends about 8 mi southeast from Shoreline Butte (Fig. 4). Shoreline Butte, the prominent dark mass at the northwest end of the Confidence Hills, contains many remnants of shoreline cut by Lake Manly, which occupied much of Death Valley. The Confidence Hills contains late Tertiary and Quaternary lake beds, conglomerate, and basalt that were folded and uplifted during movement along branches of the Death Valley fault zone, which flank and traverse the Confidence Hills. The southeastern end of the fault zone merges with the Garlock fault a few tens of miles to the southeast, on the northeast side of the Avawatz Mountains. The northwesternmost trace of the Death Valley fault zone is in a small cinder cone about 1 mi north of Shoreline Butte (Fig. 5).

The Death Valley fault zone and its northern en echelon counterpart, the northern Death Valley–Furnace Creek fault zone, have been the subject of considerable discussion regarding the amount of right-lateral slip that has occurred along it. Stewart (1967) and Stewart and others (1968) presented arguments for right-lateral slip of 40 mi or more, whereas Wright and Troxel (1967) argued that no more than 6 to 10 mi can be demonstrated. In either case, small-scale right-lateral slip can easily be demonstrated at many points along the faults.

The small cone in the floor of the valley, for example, is cut by a branch of the Death Valley fault zone and is offset in a right-lateral sense and in very recent time. The cinder cone and related tephra provide evidence for very recent lateral slip and normal fault movement. The cone is andesitic in composition, about 600 ft long, 500 ft wide, and 100 ft high; it is only slightly dissected. The fault branch that cuts it trends northwest and divides the cone into two segments, the western segment of which comprises about two-thirds of the cone. The eastern segment has been relatively offset approximately 300 ft to the southwest. Tephra that has a chemical similarity to tephra on the flanks of the cone is preserved in gullies in at least 12 places within 5 or 6 mi of the cone. In each locality the tephra is protected from erosion by younger gravel and the younger gravel is cut by normal faults.

There is much evidence of normal faulting in the gravel along the east side of central Death Valley. This central segment of the valley trends more northerly, and the valley floor is asymmetrical, compared with the northerly and southerly segments. It is this segment of Death Valley that is interpreted (Burchfiel and Stewart, 1966) as being an opening that has developed between two en echelon right-lateral faults.

The fans along the east side of Death Valley between Shoreline Butte and Furnace Creek Wash are cut by hundreds of faults, many of which are very young. Some of the faults have scarps of only a few inches or feet, but others have scarps of 50 ft or more. Some are single traces, others occur in swarms; most are relatively near the front of the Black Mountains.

The west-facing scarp along the east side of the road, extending north from the vicinity of the turnoff marked West Side Road north of Shoreline Butte, is an example of young faulting along the east side of Death Valley. The prominent outcrop of Quaternary basalt forms a linear trace of a fault. Many small and younger faults cut the very young fans at the base of the basalt scarp.

Smith Mountain, the prominent mass rising along the east side of Death Valley north of Shoreline Butte, has a steep west front that also is an expression of moderately young faulting (Fig. 4). The slope is as steep as 30° in places and has been only slightly modified by erosion. Canyons emerging from it have narrow slots at the mountain fronts that are the result of downcutting in elevated stream channels.

Tertiary and Quaternary gravel and conglomerate in the small hills at the base of Smith Mountain indicate the magnitude of some of the normal faulting. One of the normal faults that extends along the edge of Smith Mountain dips west beneath the gravel hills at a moderately low angle. An older reddish conglomerate unit in the hills rests on the hanging wall of the fault and contains clasts that have a source only on the east side of the range. The clasts do not include Precambrian gneiss, diorite, or Tertiary quartz monzonite, which are exposed in the present mountain front. Thus, normal slip amounts to a minimum equal to the height of the mountains (at least 6,000 ft). The younger (tan) gravel contains clasts that were derived in part from the rocks in the crest of the range and the west front. This indicates that the major movement occurred after the red gravel was deposited but before the younger rocks were deposited. Subsequent faulting has cut the younger gravel as well.

Figure 4. Oblique view southwest over southern Death Valley. Black Mountains in foreground, Panamint Range (right) and Owlshead Mountains (left) beyond Death Valley. Confidence Hills, Shoreline Butte, and offset cinder cone in floor of the valley. Note sharp boundary of Black Mountains and Death Valley. Mormon Point situated where valley abruptly widens toward bottom of photo. U.S. Geological Survey – U.S. Air Force photograph.

Figure 6. Oblique view of west side of Copper Canyon turtleback. Tertiary rocks overlie turtleback surface plunging to northwest (left). Photograph by L. A. Wright. U.S. Geological Survey - U.S. Air Force photograph.

The south half of the front of Smith Mountain is not an expression of a single fault plane, but rather it contains a series of normal faults that strike obliquely into the mountain a few degrees southeast of the trend of the range front. A short distance in the range, these faults turn abruptly east, then farther in the range they turn northeast, where they appear to die out. The main zone of faults along the southwest side of Ashford Peak (Noble's Amargosa fault) can be traced to the range front south of Smith Mountain, where they, too, turn north and follow the west flank of the range.

From near Mormon Point, one can see the change in the trend of the valley (Fig. 4) from northwest between the Avawatz Mountains and Mormon Point to a nearly north trend between Mormon Point and Furnace Creek Wash. At Furnace Creek Wash the valley again assumes a northwest trend parallel to the northern Death Valley–Furnace Creek fault zone and more or less in line with Furnace Creek Wash.

Mormon Point offers an excellent view of the Copper Canyon turtleback, which lies northeast across the playa from Mormon Point. Its plunging northwest-trending axis is near the somber-colored skyline crest nearly east of Mormon Point. The axis of the Mormon Point turtleback is the same in trend and plunge as the Copper Canyon turtleback and is near the crest of the ridge that terminates at Mormon Point (Fig. 6). (See the articles by Otton and by Wright and others, 1974, in Part II.)

MORMON POINT TO BADWATER

Gravel of the Funeral Formation lies east of Mormon Point on the south side of the road. Faults that trend northeast cut and bound the gravel and shorelines of Lake Manly that are preserved on the gravel slopes. Younger faults that also trend northeast cut the modern fans in this area.

The steep fans dipping off the west slope of the Copper Canyon turtleback also show various hues of desert varnish.

You can easily hike to the mountain front near Copper Canyon from the road and thus observe detailed features of the turtleback surfaces and the rocks above and below it. The rocks exposed along the mountain front for a few miles north of Copper Canyon are similar to the Tertiary rocks exposed along the road between Salsberry Pass and Death Valley. Here, too, coarse, dark-red, well-cemented conglomerate is associated with andesite that has a trachytic texture.

BADWATER TO FURNACE CREEK WASH

Badwater, at the lowest point of elevation of the United States (282 ft below sea level) lies along the flank of the third and probably most picturesque turtleback. The Badwater turtleback plunges beneath Tertiary rocks at the canyon containing the natural bridge. Foot trails in the canyon follow the arcuate trace of the plunging turtleback surface. The west front of the range south of the canyon has remnants of Tertiary rocks still preserved in fault contact with the Precambrian rocks.

Dante's View, an excellent viewpoint from the crest of the Black Mountains, lies above Badwater, 25 mi by road from Badwater. The view from this location is especially good in bright morning sunlight.

Rocks exposed in the range front north from the Badwater turtleback are multicolored Tertiary sedimentary and volcanic rocks that are incompletely understood. Near Furnace Creek Wash, the rocks are subdivided into the older Artist Drive Formation, the Furnace Creek Formation, and the Funeral Formation. Each has coarse and fine facies.

The northern part of the Black Mountains does not appear to have been uplifted as much as the southern part. For example, north of the Badwater turtleback, Tertiary rocks are exposed on both sides of faults at the range front, no pre-Tertiary rocks are exposed in the mountains, the fans are much larger, and the range gradually dies out northward at Furnace Creek.

One mile north from the turnoff to Desolation Canyon, the paved road crosses the channel of a stream that now drains most of Furnace Creek Wash. The drainage from Furnace Creek Wash was diverted into this channel (Gower Gulch) at a point a few hundred yards east of Zabriskie Point. The diversion was created in order to protect man-made features in Furnace Creek Wash and at Furnace Creek Inn from damage from flash floods. Before the diversion, flood waters and debris that flowed through this small channel originally were derived from a drainage basin of less than 2 sq mi carved in soft sedimentary rocks. The original fan at Gower Gulch is modest in size and consists mostly of gravel rarely larger than pea size, except for occasional boulders a few feet in diameter that were deposited from flash floods.

Since the diversion, however, the channel has had to carry a much larger volume of flowing water and debris draining from a much larger basin. (See the article by Troxel in Part II.)

FURNACE CREEK INN TO ZABRISKIE POINT

A brief description and detailed map of the lower part of Furnace Creek Wash by McAllister (1970) should be used by those who want specific information on this region. Portions of his report are included in Part II of this guidebook.

Furnace Creek Wash, which separates the Funeral and Black Mountains, is underlain by the Artist Drive Formation (Oligocene(?) to Pliocene), Furnace Creek Formation (Pliocene), and Funeral Formation (Pliocene and Pleistocene(?)). These three formations each consist of sedimentary and volcanic rocks and are folded into a syncline whose axis is parallel to and near the center of Furnace Creek Wash. The folded rocks are cut by numerous faults and are overlain by Pleistocene gravel.

State Highway 190 follows along the southwest limb of the syncline between Furnace Creek Inn and the Zabriskie Point turnoff. The rocks near the road are beds of coarse gravel of the Furnace Creek Formation. They grade laterally and vertically into fine-grained sediments, most of which are lacustrine in origin.

At many places along the road in Furnace Creek Wash are good exposures of geologic features such as sedimentary

Figure 5. Cinder cone in floor of Death Valley. Rare coating of snow on cone and valley floor. U.S. Geological Survey - U.S. Air Force photograph.

features in the silt and gravel and unconformities between Quaternary gravel and the Tertiary to Quaternary sediments.

ZABRISKIE POINT

Zabriskie Point affords excellent views of the Tertiary rocks. South from Zabriskie Point, the principal rocks are dark-colored sedimentary and volcanic units of the Artist Drive Formation on the skyline ridge and pale-colored fine-grained sediments of the Furnace Creek Formation in the low hills between the skyline ridge and Zabriskie Point. In the dark hill northwest of Zabriskie Point, the interfingering relations of lacustrine sediments and gravel are well exposed.

A few tens of feet from the turnoff to Zabriskie Point is a gravel barrier across Furnace Creek Wash constructed diagonally from the foot of Zabriskie Point to the highway. On the upstream side of the barrier is a deeply incised drainage network cut into the floor of the wash as a result of the drainage now diverted into the channel that flows around the east and south sides of Zabriskie Point. (See the article by Troxel in Part II.)

ZABRISKIE POINT TO DEATH VALLEY JUNCTION AND SHOSHONE

Locally, on the southwest side of the road east from Zabriskie Point, dark-colored basalt dikes and sills can be seen in the pale-colored sediments.

About 1 mi southwest from the intersection with the road to Dante's View is an open pit mine operated by the Tenneco Company. It is the only borate mine that has been in operation in the Death Valley region for the past several years. About 2 mi east of the Tenneco mine are the gray metal buildings at the Ryan mine. The Ryan was a principal source of borax for many years but has not been in operation for many years.

The narrows in Highway 190, 1 mi or so northeast from the turnoff to Dante's View, is cut through coarse gravel of the Funeral Formation at a point where a basalt flow pinches out northward in the gravel.

Farther east, the gravel gives way down-section to fine-grained sediments deposited in a local basin. The east edge of the basin is marked by bold outcrops of pink-colored well-cemented sediments that have been the source for building stone. Continuing farther east, the road climbs in the section of coarse gravel of the Funeral Formation and follows along and over the north margin of basalt flows. Note that the drainage divide between Death and Amargosa Valleys lies east of the main crest of the Black Mountains, a phenomenon common to many parts of the ranges on the east side of Death Valley. Note also that the divide is in alluvial material.

The Furnace Creek fault zone lies along the base of the mountains on the northeast side of Furnace Creek Wash. It separates rocks primarily of late Precambrian and Paleozoic age in the Funeral Mountains from Tertiary and Quaternary sedimentary and volcanic rocks in the Greenwater Range and Furnace Creek Wash.

The trace of the Furnace Creek fault zone along the edge of Furnace Creek Wash projects southeast along a line that extends along the north margin of Eagle Mountain, a prominent point in the floor of the Amargosa Valley south of Death Valley Junction. Evidence is lacking for a continuation of the Furnace Creek fault zone through the Resting Spring Range farther east.

A major part of the Black Mountains, Greenwater Valley, and Greenwater Range is bounded by the north-trending segment of central Death Valley, by the north-trending Amargosa Valley between Death Valley Junction and Shoshone, by the northwest-trending Sheepshead fault between Amargosa Valley and Death Valley at Salsberry Pass, and by the northwest-trending Furnace Creek fault between Death Valley and Death Valley Junction.

Volcanic rocks that are widespread within this area are not known beyond the limits of it. This area is roughly shaped like a parallelogram that appears to be a large tectonic-volcanic feature that has formed as a result of spreading between the two northwest-trending faults. The parallelogram probably has a much more complex history than the simple extension and emission of volcanic rocks, because the unmetamorphosed Precambrian to pre-Tertiary rocks that are found on all sides of the parallelogram are lacking within it.

The trip for this day, which started and ended in Shoshone, has followed very closely the boundaries of this parallelogram.

SHOSHONE TO DEATH VALLEY JUNCTION

The Resting Spring Range east of State Highway 127 contains a thick succession of sedimentary rocks that are mainly of Cambrian and older age. The prominent small hill about 2 mi to the right of the highway from a point about 1 mi north of town contains Tertiary volcanic and sedimentary rocks. At the north end of the hill is an intermittently active borate mine in Tertiary sedimentary rocks. The rocks in the Resting Spring Range are cut by many low-angle west-dipping normal faults. The various rock units are moderately easily distinguished, the high point of the range being underlain by dark-gray, prominently banded carbonate rocks of the Bonanza King Formation. The underlying rocks are the same as those exposed in the Nopah Range, which are, in a downward succession, the Carrara Formation (pale reddish brown); Zabriskie Quartzite (a prominent thin pale pinkish-gray band); the dark-colored Wood Canyon Formation; and the uppermost unit of the Stirling Quartzite (moderately pale).

The highway between Shoshone and Death Valley Junction follows along the course of the Amargosa River, which has its headwaters in southwest Nevada. It flows south from Nevada to southern Death Valley, then west for several miles and finally north to Badwater.

The rock units in the Resting Spring Range north from a point about 7 mi north of Shoshone are noticeably much more faulted than they are farther south. Some of the faults are nearly flat and have considerable displacements. Note, for instance, the dark-colored Bonanza King Formation that forms the crest of most of the range. Near the north end of the visible point of the range, it also occurs low on the west front of the range where it has been downdropped along a large, low-angle, west-dipping normal fault (Fig. 7).

Eagle Mountain, about 4 mi west of the Resting Spring Range, contains the same rock units as those in the Resting Spring Range; they may have been transported to their present site in the hanging-wall block of a low-angle normal fault. The small patch of pale pinkish-buff rocks on the southwest side of Eagle Mountain is Tertiary in age. The strata are underlain by a unit composed mainly of angular blocks of gray carbonate rocks and poorly bedded gravel, similar to rocks at the base of the Titus Canyon Formation in other areas.

Along the left side of the road, southwest of Eagle Mountain, gravel—probably of the Funeral Formation—is overlain by basalt.

DEATH VALLEY JUNCTION TO FURNACE CREEK RANCH

The southern part of the Funeral Mountains lies along the right (north) side of Highway 190 between Death Valley Junction and Furnace Creek in Death Valley. The mountains contain a thick section of southeast-dipping late Precambrian and Paleozoic sedimentary rocks that locally are overlain by Tertiary rocks (Fig. 8).

The southeast tip of the range contains prominently exposed pale-brown conglomerate and sandstone of the Titus Canyon Formation. Beneath them are pale-tan lake bed sediments, also of the Titus Canyon Formation. The mouth of a small canyon cutting through the Titus Canyon Formation is choked by a landslide made up of the fine-grained sediments of the Titus Canyon. The stratigraphy and structure of the segment of the Funeral Mountains along Furnace Creek Wash have been investigated by J. F. McAllister. (See the maps, sections, and article by McAllister in Part II.) The range is transected by many northeast-trending, northwest-dipping normal faults, which result in duplication of many of the stratigraphic units by downdropping on the faults. Even without recognizing the specific stratigraphic units, the repetition of similar rocks by faults can be seen from the highway.

Two of the most easily recognized units are exposed in the high peak to the right (north) of the divide that lies 11 mi west of Death Valley Junction. The nearly white-colored unit is the Ordovician Eureka Quartzite, which is repeated by faults several times farther west. The quartzite is overlain by the very dark-colored Ely Springs Dolomite and is underlain by the Pogonip Group, which has a brownish color on weathered outcrops. The quartzite is also exposed low in the saddle east of the high peak.

The prominent, vertical white veins exposed in the cliffs of the Funeral Formation north of the highway near the monument boundary are calcite. Some are offset by bedding-plane faults.

At the point where the road turns southwest below the monument boundary, there is a good view across Furnace Creek Wash, where the variety of colors in the Tertiary rocks stand out in bold contrast. The two small hills visible in Furnace Creek Wash about 2 mi below the turnoff to Dante's View are made up of brown gravel, sandstone, and pale-colored lake beds.

FURNACE CREEK RANCH TO HELL'S GATE

The mountains east (right) of State Highway 190 leading north from Furnace Creek Ranch consist of Cambrian strata broken by many faults. Winters Peak is the highest (but slightly rounded) point in this part of the Funeral Mountains. The low, prominent dark-colored peak at the west flank and north edge of this segment of the range is Nevares Peak. It contains the Bonanza King and other Cambrian formations. A branch of the northern Death Valley–Furnace Creek fault zone extends along the base of the peak. After crossing the low hills 3 mi north of Furnace Creek, the northern and more distant part of the Funeral Mountains is visible. Rocks in this part of the mountains consist of the Johnnie through Wood Canyon Formations. The rocks become more highly metamorphosed and successive older rocks are locally exposed from south to north. Rocks that are siltstone or phyllite in the southern part of the range near Winters Peak are metamorphosed to such rock types as staurolite schist, garnet schist, and muscovite schist; some contain kyanite. Rocks in these mountains, in addition to having been folded into several large anticlines, have been cut by many west-dipping low-angle normal faults. The west face of this segment of the range is bounded by the Keane Wonder fault, which dips about 25° west where it is exposed. The fault extends southeast into the range about 2 mi along the east side of Winters Peak, where it turns and extends east and then northeast. The northern trace of the Keane Wonder fault becomes difficult to follow after it crosses the highway in Boundary Canyon. The rocks beneath the Keane Wonder fault plane are mildly to strongly metamorphosed. Those above the fault plane are unmetamorphosed.

A major fold in the rocks beneath the Keane Wonder fault has a doubly plunging axis, which is oriented approximately parallel to the mountain range. This fold lies along the west side of the range for several miles south from Boundary Canyon.

The major rock units in the Funeral Mountains can be identified with little difficulty from the highway, especially with afternoon sunlight, but this can best be done at a place such as the turnoff to Beatty, Nevada. Most of the rocks in the area between Winters Peak and the high peak in the north part of the Funeral Range opposite the Beatty turnoff are gray quartzite units of the Stirling Quartzite. The high peak, marked by a double white stripe on its south side, is Chloride Cliff. The smooth crest of the second peak south of Chloride Cliff contains the lower part of the Stirling Quartzite. The less obvious thin brownish bands are carbonate layers within the Stirling Quartzite. The western slope of the second peak south of Chloride Cliff contains fine-grained greenish-gray sedimentary rocks of the Johnnie Formation. These and all underlying rocks

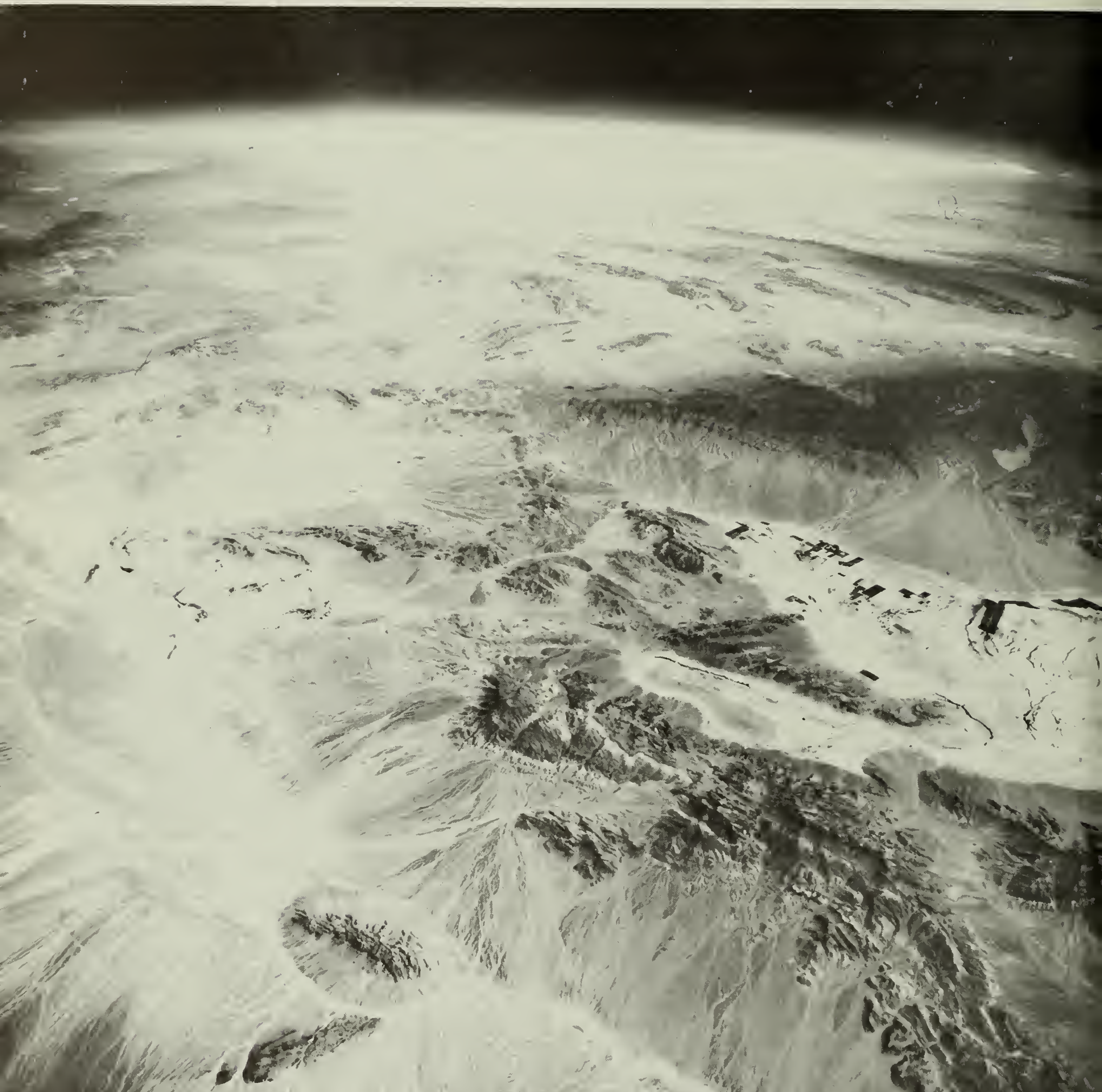


Figure 7. Oblique view to northeast from a point about 10 miles north of Shoshone. Eagle Mountain in left foreground. Resting Spring Range extends diagonally from lower right. Spring Mountains and Las Vegas Valley in upper right. U.S. Geological Survey – U.S. Air Force photograph.

Figure 8. Oblique view to northeast from upper part of Furnace Creek Wash. Near side of Funeral Mountains in middle foreground is bounded by northern Death Valley–Furnace Creek fault zone. U.S. Geological Survey - U.S. Air Force photograph.

are highly metamorphosed, except in the vicinity of Winters Peak. Beneath the Johnnie Formation, in the first peak south of Chloride Cliff, is the brown-colored rough weathering upper part of what appears to be the Kingston Peak Formation. The lower part of the Kingston Peak Formation is greenish in color, weathered to smooth slopes in a band that appears to project beneath Choride Cliff. The rocks at Chloride Cliff are most likely lithologically equivalent units to the Beck Spring Dolomite. All the rest of the multicolored rocks downslope (west) from Chloride Cliff are metamorphosed rocks that can be approximately correlated with units of the Crystal Spring Formation. Low in the canyon that drains the south side of Chloride Cliff is a prominent horizontal black band that overlies a very pale-colored rock. The prominent black band is composed of amphibolite, probably a metamorphosed diabase sill in the Crystal Spring Formation.

These rocks are exposed in the core of the anticline mentioned earlier. Not easily discernible from here, but given the proper vantage point, the southwest continuation of each unit can be seen, from the Stirling Quartzite downward, to turn north around the south end of the anticline at the flank of the range. The best unit to follow is the brown rough weathering unit of the Kingston Peak Formation, which is dipping southeastward in the first peak south of Chloride Cliff. It is greatly thinned and dips southwest, where it crops out in the range front south of the amphibolite band. The Johnnie Formation and Stirling Quartzite above the Kingston Peak Formation likewise turn and thin along the range front; the Crystal Spring and Beck Spring beneath the Kingston Peak are more complexly folded; and folding becomes much more intense and complex in the northern plunging axis of the anticline near Boundary Canyon.

Between Winters Peak and the road in Boundary Canyon, the only rocks exposed above the Keane Wonder fault are Tertiary sedimentary rocks that crop out in many places in the low foothills.

The large white patch of rock at the base of Chloride Cliff is travertine material formed from spring water seeping along the Keane Wonder fault.

The gravel in the fan crossed by the road that leads to Beatty from State Highway 190 is rich in metamorphic rocks derived from the northern part of the Funeral Mountains. It is worthwhile stopping along here to look at the variety of the metamorphic rocks.

Barely less than 2 mi north of the Beatty turnoff, the road is cut through a well-preserved gravel bar that was formed when Lake Manly occupied Death Valley. The roadcuts show the stratification within the bar; erosion has done little to alter its form. A smaller gravel bar lies on the right side of the road barely 0.1 mi farther north.

The low brownish-colored hills north of the gravel bars are composed of gravel and fine-grained sediments of the Funeral Formation. These hills are uplifted and folded along the northeast side of the northern Death Valley fault zone. The darker brown-colored hills farther north and east of the highway contain gravel, tuff, and fine-grained sediments that are probably part of the Titus Canyon Formation. A fault appears to separate the Titus Canyon from the Funeral Formation.

The isolated hill on the right side of the highway, which has an obvious pale-lavender tint, is composed of coarse quartzite debris derived solely from the Wood Canyon Formation.

Corkscrew Peak, the prominent peak in the Grapevine Mountains, is approximately in line with the north-trending highway. Corkscrew Peak, which lies in the axis of a large recumbent fold, is marked by a gray cliff-forming band between weaker rock that weathers pale reddish-brown. The structure and stratigraphy of the Grapevine Mountains are described by Reynolds in an article in Part II. The crest of Corkscrew Peak contains rocks of the Carrara Formation; the rocks on the slope beneath it are units of the Wood Canyon Formation.

Post-Titus Canyon sedimentary rocks that lie along the west front of the Funeral Mountains contain sediments derived from two directions. Coarse material transported from the east contains clasts derived primarily from the Wood Canyon Formation and Stirling quartzite. The coarse material grades westward into pink sand, then to pink silt; coarse material derived from the west contains clasts of rocks that are common to the Panamint Range. The gravel from the west is brown and grades laterally eastward into tan sand and farther eastward into cream-colored silt.

Tertiary rocks between Boundary Canyon and Nevares Peak are folded into a broad, shallow, gently plunging syncline, whose axis is parallel to the trace of the Keane Wonder fault. The rocks are repeated many times along northeast-trending northwest-dipping faults, which do not appear to cut the Keane Wonder fault.

Just beyond the point where the road swings eastward around brown rocks, there are, on the left (west), some layers of green and pink tuff within the conglomerate of the Titus Canyon Formation.

From a point about 1 mi farther north, the highly distorted beds of Precambrian rocks in the north crest of the plunging anticline low in the Funeral Mountains become obvious. The irregular black streaks are formed by amphibolite of the Crystal Spring Formation, and the irregular thin white streaks are some of the many coarse pegmatite dikes that intrude this part of the anticline.

HELL'S GATE TO DAYLIGHT PASS

The ridge on the right side of the road at Hell's Gate is underlain by quartzite and dolomite of the Stirling Quartzite. To the right (west) of the ridge, the basal conglomerate or breccia unit of the Titus Canyon Formation rests upon red-stained Stirling Quartzite. Most of the clasts are angular to subrounded boulders of quartzite and carbonate rock. On the north side of the road at Hell's Gate, the axis of the overturned syncline that goes through Corkscrew Peak to the north can be seen to project through rocks of the Wood Canyon Formation exposed high in the nearby small peaks east of the wash that lies east of Corkscrew Peak. Note that the second small peak to the right of the wash contains a dark layer of rock which can be seen to turn from a horizontal position to vertical and overturn in just a few tens of feet in the axis of the fold. One mile east of Hell's Gate, red rocks on the left side of the road and 0.25 mi distant mark the trace of the west-dipping Keane Wonder fault or an extension of it. Beyond the gravel bank on the right side of the road, the Keane Wonder fault separates the pale-gray Stirling Quartzite on the hanging wall (west side) of the Keane Wonder fault from the darker, highly metamorphosed rocks beneath the fault.

Where the bedrock is near the north side of the road, tan carbonate rocks can be seen to rest on grayish-green rocks along a very gently inclined plane that is parallel to the road and a few tens of feet north of it. This plane is a fault that is either the Keane Wonder fault or a low-angle fault that branches from it. The fault separates the unmetamorphosed Stirling Quartzite and Wood Canyon Formation from the underlying highly metamorphosed rocks, which here are gently undulating schist, gray marble, and stretched pebble conglomerate beds of the Johnnie Formation.

At 4.2 mi from Hell's Gate, the fault swings south across the road and continues along a remarkably smooth and well-exposed plane for 7 mi farther to the southeast.

From this point, the road continues up Boundary Canyon through folded and faulted Stirling Quartzite and Wood Canyon Formation. At 5.2 mi from Hell's Gate are two prominent craggy peaks composed of gray carbonate rocks on the left side of the road. These peaks consist of chunks of Bonanza King Formation, which are giant clasts in the Titus Canyon Formation, now mostly removed by erosion. Tan carbonate rocks on the right side of the road are part of the Stirling Quartzite and contain numerous small folds. The prominent ridge farther up the road on the right (east) is composed of the Wood Canyon Formation that strikes across the road another mile or so ahead.

DAYLIGHT PASS TO BEATTY, NEVADA

Daylight Pass, elevation 4,317 ft, marks the contact of the Titus Canyon Formation, the type locality of which is only a few miles to the west. This formation has yielded skeletons of Titanotheres, which give it an Oligocene age. Two features that are common to the lower part of the Titus Canyon Formation are the giant sedimentary clasts that lie at its base and a conglomerate unit that is noted for its well-rounded and exceptionally highly polished pebbles. In general, the Titus Canyon Formation occupies the low areas to the north and is overlain by younger Tertiary rocks largely volcanic in origin. The craggy peaks and slopes on both sides of the road at Daylight Pass are outcrops of the Zabriskie Quartzite, which rest on the Wood Canyon Formation.

From the east edge of the hills east of the Grapevine Mountains, the Bullfrog Hills, mainly north of the highway, and Bare Mountain in the eastern skyline come into view. These two features have been mapped and described by Cornwall and Kleinhampl (1964). The Bullfrog Hills consists nearly entirely of Tertiary volcanic rocks. A small mining community by the name of Rhyolite once existed at the end of a railroad spur that was extended into the Bullfrog Hills from Beatty. At the monument boundary, a gravel road extends southward to Chloride Cliff—a ghost mining town on the flanks of Chloride Cliff. From the townsite there is a spectacular view of central and northern Death Valley and outcrops of some of the metamorphosed Precambrian rocks.

From the valley east of the Grapevine Mountains, the east side of the Funeral Mountains is seen as much more subdued in appearance than the west side, as are most of the mountain ranges along the east side of Death Valley.

Near Beatty, the highway crosses some of the volcanic rocks of the Bullfrog Hills. Bare Mountain contains late Precambrian and Paleozoic rocks that are highly deformed by faulting and folding; in addition, the rocks in the northern part of Bare Mountain are metamorphosed. Some of the units that are easily identified from a distance are the nearly vertical striped gray rocks of the Bonanza King Formation, overlain by the Ordovician rocks and underlain by the Carrara Formation that has been bleached during slight metamorphism.

BEATTY TO LATHROP WELLS

Tertiary volcanic rocks at Beatty give way southward to metamorphosed Wood Canyon Formation that crops out on the right side of the road south of Beatty. Farther south and on the left side of the road are outcrops of other metamorphosed Cambrian rocks that include Zabriskie, Carrara, and probably Bonanza King.

The south boundary of the metamorphosed rocks appears to be a north-dipping fault that trends east across Bare Mountain. Most of the rocks just south of that fault are various units of the Stirling Quartzite. Still farther south, in the crest of the range, Cambrian and Ordovician (banded gray to white) rocks are overlain by more brownish-colored rocks that have been thrust over them. Southward and stratigraphically beneath the carbonate rocks are the Wood Canyon Formation and the upper part of the Stirling Quartzite. The pale-gray and tan rocks at the head of the straight gravel road that leads from the abandoned buildings near the highway are bleached units of the Carrara Formation. This is the type locality for the Carrara Formation, named after the abandoned site of Carrara. Still farther south in downward succession are the Bonanza King Formation (gray), Carrara Formation (pale brown), Zabriskie Quartzite (pale reddish brown), Wood Canyon Formation (dark gray), and Stirling Quartzite (tan). The isolated peak at the south end of Bare Mountain is of Bonanza King Formation. The low outcrops south and southeast of the Bonanza King Formation outcrops contain brecciated units along the east side of the fault that bounds the east side of Bare Mountain. Zabriskie Quartzite along the fault plane appears to have been offset right laterally from outcrops farther north in Bare Mountain. The rocks along the north side of the highway for many miles eastward from the south tip of Bare Mountain are Tertiary volcanic rocks of the giant Timber Mountain caldera, which has been intensively studied by U.S. Geological Survey personnel because the Nevada Test Site is situated within this caldera.

About 2 mi east of Bare Mountain, where a patch of very pale lake sediments crop out a few hundred feet north of the road, the ridge top is covered by a sheet of sedimentary breccia composed wholly of gray carbonate rocks. The sheet is several tens of feet thick and covers several square miles. A small cinder cone and basalt flow lie less than 1 mi north of the road and a few miles east of Bare Mountain.

LATHROP WELLS TO LAS VEGAS

From Lathrop Wells, the main road extends east around the north end of the Spring Mountains, then southeast to Las Vegas. Between Lathrop Wells and the Spring Mountains the road traverses small hills that are underlain predominantly by Cambrian and Ordovician rocks that are moderately deformed. The route is approximately along the trace of the Las Vegas shear zone.

SELECTED REFERENCES

Burchfiel, B. C., and Stewart, J. H., 1966, "Pull-apart" origin of the central segment of Death Valley, California: Geol. Soc. America Bull., v. 77, p. 439–442.

Chesterman, C. W., 1973, Geology of the northeast quarter of Shoshone quadrangle, Inyo County, California: California Div. Mines and Geology Map Sheet 18.

Cornwall, H. R., and Kleinhampl, F. J., 1964, Geology of Bullfrog quadrangle, and ore deposits related to Bullfrog Hills caldera, Nye County, Nevada, and Inyo County, California: U.S. Geol. Survey Prof. Paper 454–J, p. J1–J125.

Drewes, Harald, 1963, Geology of the Funeral Peak quadrangle, California, on the east flank of Death Valley: U.S. Geol. Survey Prof. Paper 413, 78 p.

Hazzard, J. C., 1937, Paleozoic section in the Nopah and Resting Springs Mountains, Inyo County, California: California Div. Mines, Jour. Mines and Geology, v. 33, no. 4, p. 273–339.

Hewett, D. F., 1940, New formations to be used in the Kingston Range, Ivanpah quadrangle, California: Washington Acad. Sci. Jour., v. 30, p. 239–240.

——— 1956, Geology and mineral resources of the Ivanpah quadrangle, California and Nevada: U.S. Geol. Survey Prof. Paper 275, 172 p.

Hunt, C. B., and Mabey, D. R., 1966, General geology of Death Valley, California—Stratigraphy and structure: U.S. Geol. Survey Prof. Paper 494–A, 162 p.

McAllister, J. F., 1970, Geology of the Furnace Creek borate area, Death Valley, Inyo County, California: California Div. Mines and Geology Map Sheet 14.

Noble, L. F., 1934, Rock formations of Death Valley, California: Science, v. 80, no. 2069, p. 173–178.

——— 1941, Structural features of the Virgin Spring area, Death Valley, California: Geol. Soc. America Bull., v. 52, p. 941–1000.

Noble, L. F., and Wright, L. A., 1954, Geology of the central and southern Death Valley region, California: California Div. Mines Bull. 170, chap. II, p. 143–160.

Stewart, J. H., 1966, Correlation of Lower Cambrian and some Precambrian strata in the southern Great Basin, California and Nevada: U.S. Geol. Survey Prof. Paper 550–C, p. C66–C72.

——— 1967, Possible large right-lateral displacement along fault and shear zones in the Death Valley–Las Vegas areas, California and Nevada: Geol. Soc. America Bull., v. 78, p. 131–142.

Stewart, J. H., Albers, J. P., and Poole, F. G., 1968, Summary of regional evidence for right-lateral displacement in the western Great Basin: Geol. Soc. America Bull., v. 79, p. 1407–1414.

Wasserburg, G. J., Wetherill, G. W., and Wright, L. A., 1959, Ages in the Precambrian terrane of Death Valley, California: Jour. Geology, v. 67, no. 6, p. 702–708.

Wright, L. A., 1968, Talc deposits of the southern Death Valley–Kingston Range region, California: California Div. Mines and Geology Spec. Rept. 95, 79 p.

——— 1973, Geology of the SE¼ Tecopa 15-minute quadrangle, San Bernardino and Inyo Counties, California: California Div. Mines and Geology Map Sheet 20.

Wright, L. A., and Troxel, B. W., 1967, Limitations on right-lateral, strike-slip displacement, Death Valley and Furnace Creek fault zone, California: Geol. Soc. America Bull., v. 78, p. 933–950.

——— 1970, Summary of regional evidence for right-lateral displacement in the western Great Basin: Discussion: Geol. Soc. America Bull., v. 81, p. 2167–2174.

Oblique view southwestward across southern Death Valley. Owlshead Mountains in center of photo. Death Valley extends at slight angle across photo beyond mountains in foreground. Shoreline Butte, cinder cone, and Confidence Hills in floor of Death Valley (on right). Normal faults in Owlshead Mountains and mountains in foreground trend approximately at right angles to Death Valley fault zone. Traces of Death Valley fault zone in floor of valley. Garlock fault zone extends obliquely away from observer front center of left edge of photo. U.S. Geological Survey – U.S. Air Force photograph.

Geology of the Spring Mountains, Nevada

B. C. Burchfiel,[1] R. J. Fleck,[2] D. T. Secor,[3] R. R. Vincelette,[4] G. A. Davis[5]

ABSTRACT

The northwest-trending Spring Mountains, Nevada, contain a 45-mi-wide (75-km) cross section of the eastern part of the North American Cordilleran orogenic belt and geosyncline. This cross section is probably the most southerly exposed section which exhibits structure and stratigraphy "typical" of the eastern part of the Cordillera.

Stratigraphically, the transition from Paleozoic craton to miogeosyncline is present from east to west across the Spring Mountains. The sedimentary succession throught the middle Permian thickens from 8,800 ft (2,660 m) east of the Spring Mountains to approximately 30,000 ft (9,000 m) in the west. Thickening of individual formations accounts for 6,800 ft (2,070 m) of added section, addition of formations at unconformities accounts for 4,600 ft (1,400 m) of added section, and addition of a thick terrigenous late Precambrian sequence accounts for 9,800 ft (3,000 m) of added section.

Three major thrust plates are exposed in the Spring Mountains, each structurally higher plate containing a thicker sequence of Paleozoic rocks. The easternmost thrust is the Keystone thrust, except where the earlier Red Spring thrust plate is present below the Keystone as isolated remnants. The Keystone thrust appears to be a décollement thrust, but complications at depth suggest that additional thrust slices may be present below the thrust, or several thousand feet of late Precambrian terrigenous rocks may be present above the thrust. The structurally higher Lee Canyon thrust plate probably contains at least 4,000 ft (1,200 m) of these terrigenous rocks at its base, and the Wheeler Pass thrust plate contains at least 11,000 ft (3,300 m) of these rocks. Pregeosynclinal basement could be involved in some of the higher thrust plates, particularly the Wheeler Pass plate, but depths of exposure are inadequate to determine its role.

Thrust faulting has produced a shortening of from 22 to 45 mi (36.6 to 75 km) in the geosynclinal rocks based on geometric constructions of cross sections at depth. This range probably represents a minimum figure. Some folding and thrusting occurred during the early Late Cretaceous, but data within the Spring Mountains only establish a much wider time bracket, post-Early Jurassic to pre-late Cenozoic for the easternmost thrust faults and post-Early Permian to pre-late Cenozoic for the westernmost thrusts.

INTRODUCTION

The Spring Mountains are located in southeastern Nevada, 10 mi west of Las Vegas (Fig. 1). Trending northwest more than 45 mi, they form a southern boundary for the general north- or northeast-trending ranges farther north in Nevada. Because the northwest topographic trend is transverse to the north- or northeast-trending regional structural strike, the Spring Mountains offer a unique opportunity to study a wide cross section of continuously exposed pre-Tertiary rocks. Stratigraphically, the Spring Mountains contain Paleozoic rocks which show the transition from craton to miogeosyncline. Structurally, they contain rocks of the undeformed craton and of three major thrust plates belonging to the easternmost part of the North American Cordilleran orogenic belt.

The earliest geologic work in the Spring Mountains was by G. K. Gilbert (1875), who served as a geologic assistant for the Wheeler expeditions of 1871–1872. R. B. Rowe of the U.S. Geological Survey did extensive work in the central part of the Spring Mountains during the period 1900–1901, but died before his results could be published. His field data were incorporated in the regional report of J. E. Spurr (1903). Rowe clearly recognized the fault along the east side of the Spring Mountains now referred to as the Keystone thrust fault.

In 1919, C. R. Longwell began systematic mapping in the region of southeastern Nevada. His work, together with that of Nolan (1929) and Glock (1929), led to the first regional geologic map of southeastern Nevada and of the Spring Mountains (Bowyer and others, 1958). Longwell's work forms the basis for most of our present understanding of the tectonics and stratigraphy in this part of the Cordilleran thrust belt. This report summarizes the geology of most of the Spring Mountains compiled from work completed at various times from 1961 to the present (Fig. 1). The authors of this paper owe a great debt to Chester Longwell for assistance given, ranging from selection of map areas, field excursions, and discussions to personal inspiration.

GEOLOGY OF THE SPRING MOUNTAINS

The Spring Mountains contain geology typical of the eastern part of the Cordilleran orogenic belt which can be traced continuously from Canada to southern Nevada. Late Precambrian and Paleozoic rocks are characterized by a thin (approx. 8,800 ft or 2.6 km) cratonal sequence to the east which thickens northwestward (to approx. 30,000 ft or 9.0 km) in the Spring Mountains into the miogeosynclinal part of the Cordilleran geosyncline (Fig. 2*). Mesozoic rocks are present only in the eastern part of the Spring Mountains, and their relations to Mesozoic geosynclinal development are not clear (Stanley and others, 1971; Burchfiel and Davis, 1972). Structures in the Spring Mountains are dominated by east-directed thrust faults along which thicker sequences of Paleozoic rocks have moved over thinner ones. At present levels of exposure, thrust plates carry only sedimentary rocks. These structural and stratigraphic characteristics are similar to those described from areas farther north along the eastern part of the Cordilleran orogenic belt.

Southward, however, the structural and stratigraphic relations change, such that the next exposed southeast cross section in the Mesquite and Clark Mountains, 30 mi farther south, contains Paleozoic sedimentary rocks characteristic of the craton and transitional between craton and miogeosyncline, and thrust plates that involve Precambrian cyrstalline basement rocks (Burchfiel and Davis, 1971, 1972). Regional relations

[1] *Rice University, Houston, Texas 77001.*
[2] *U.S. Geological Survey, Menlo Park, California 94025.*
[3] *University of South Carolina, Columbia, South Carolina 29210.*
[4] *Trend Exploration Limited, Denver, Colorado 80203.*
[5] *University of Southern California, Los Angeles, California 90007.*

Figures 2, 3, and 5 not included with this edition.

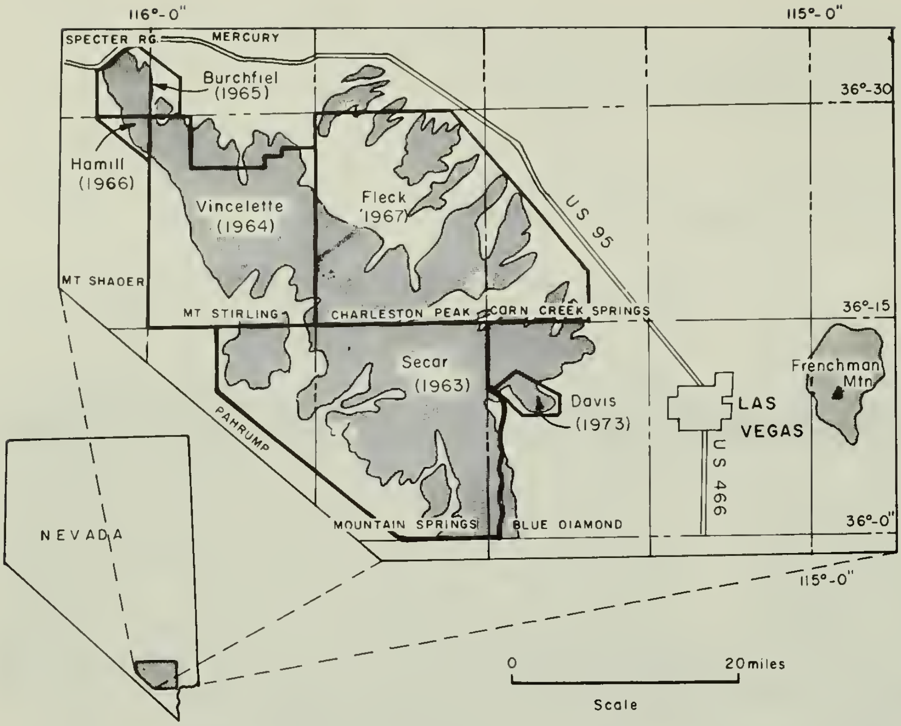


Figure 1. Location map of Spring Mountains, Nevada, showing areas of mapping responsibility. Quadrangle names are shown for Spring Mountains area.

suggest a divergence of structural and geosynclinal units (Burchfiel and Davis, 1972); thus, the Spring Mountains represent the southernmost example of "typical" Cordilleran tectonics.

Stratigraphy

No attempt is made here to describe all the stratigraphic units present in the Spring Mountains. Lithology, thickness, and facies changes are presented in Figure 2, and details of these sequences can be found in Burchfiel (1964), Secor (1963), Vincellete (1964), Fleck (1967, 1970, 1974), Gans (1970), and Longwell and others (1965). The four representative stratigraphic columns presented in Figure 2 demonstrate the change from a cratonal sequence at Frenchman Mountain (25 mi east of the Spring Mountains; Fig. 1) to a geosynclinal sequence in the northwest Spring Mountains.

Stratigraphically, the oldest exposed parts of the most easterly and westerly sections, and by inference the two central sections, begin with a nonmarine and shallow marine sequence of terrigenous rocks. At Frenchman Mountain, these rocks are Early and early Middle Cambrian in age, approximately 1,000 ft (300 m) thick, and rest unconformably on Precambrian crystalline rocks. In the northwestern part of the Spring Mountains, the terrigenous rocks are more than 11,000 ft (3.3 km) thick and range from late Precambrian to early Middle Cambrian in age. South and southwest of the Spring Mountains, the thick late Precambrian terrigenous sequence rests unconformably on either crystalline basement or on an older sequence of unmetamorphosed to weakly metamorphosed late Precambrian sedimentary rocks (Pahrump Group) that rests unconformably on crystalline basement. Regional correlations by Stewart (1970) suggest that the upper part of the Tapeats Sandstone in some cratonal sequences correlates with the Zabriskie Quartzite of the geosynclinal sequence (for example, northwestern Spring Mountains section) and that the lower part of the Tapeats correlates with the uppermost part of the Wood Canyon Formation. The transiton from the thin cratonal sequence to the thick geosynclinal sequence must occur beneath the Spring Mountains; the thick geosynclinal terrigenous sequence is first exposed in the Wheeler Pass thrust plate where it is already fully developed.

The upper part of the terrigenous sequence becomes calcareous and contains carbonate beds in lower Middle Cambrian rocks of the Bright Angel Shale, or its geosynclinal equivalent, the Carrara Formation, and grades upward into a thick sequence of shallow-water carbonate rocks that range in age from Middle Cambrian to Early Permian. Thin terrigenous units are present at several stratigraphic levels, and three can be followed from the cratonal into geosynclinal sequences: (1) the thin calcareous siltstone at the base of the Banded Mountain Member of the Bonanza King Formation; (2) the Dunderberg Shale Member of the Nopah Formation; and (3) the basal terrigenous rocks of the Bird Spring Formation. Three other prominent but thin terrigenous units are present in the geosynclinal sequence of the northern Spring Mountains, but are not present in the eastern Spring Mountains. Those three units are (1) Eureka Quartzite; (2) Ninemile Formation; and (3) a sandstone unit at the base of the Nevada Formation.

The carbonate formations can be followed through the Spring Mountains with little lithologic change: some are cut out by unconformities. Facies changes occur between the Spring Mountains and Frenchman Mountain, but the general characteristics of the formations are maintained. All the carbonate formations that have been studied suggest deposition in shallow marine or marginal marine environments (Gans, 1970).

Thickening of the carbonate sequence across the Spring Mountains takes place in two ways: (1) by thickening of individual formations, and (2) by addition of formations at unconformities. Thickening is marked in the Bird Spring Formation and equivalents of Late Mississippian to middle Permian age which thicken from 1,600 ft (485 m) at Frenchman Mountain (Callville Limestone and Permian red beds) to more than 6,700 ft (2,000 m) in the central Spring Mountains. Some formations thicken by 50 to 60 percent (for example, Bonanza King Formation), whereas others show little if any thickening (for example, Monte Cristo Limestone and Sultan Limestone).

Addition of formations at unconformities accounts for approximately 4,600 ft (1,400 m) of thickening across the Spring Mountains. In the cratonal sequence at Frenchman

Mountain, only one major unconformity is present in the Paleozoic carbonate sequence; the Middle Devonian Muddy Peak Limestone rests unconformably on the Upper Cambrian Nopah Formation. In the eastern part of the Spring Mountains above Red Rock Canyon, two and possibly three unconformities are present. One occurs between Lower and Upper Ordovician rocks (Goodwin Limestone and Mountain Springs Formation of Gans, 1970) which are here present below the Middle Devonian unconformity. A second unconformity is present below Middle Devonian rocks (Sultan Limestone) and is of regional extent. A third unconformity within unfossiliferous dolomites at the top of the Mountain Springs Formation is suspected but not proven (Gans, 1970). South along the Spring Mountains, the Ordovician and suspected Devonian rocks are cut out by the unconformities beneath Middle Devonian rocks, and the Middle Devonian Sultan Limestone rests on the Upper Cambrian Nopah Formation at Mountain Springs Pass. The trend of this wedge out of Ordovician and Devonian(?) rocks must be northeast across the structural trend (that is, west of Frenchman Mountain and north of Mountain Springs Pass).

Farther west, the Middle Ordovician Ninemile Formation, Antelope Valley Limestone, and Eureka Quartzite occur below the Upper Ordovician rocks, so that in the central part of the Spring Mountains, an unconformity is no longer recognized within Ordovician strata, and the Ordovician sequence is complete. Farther west, but still in the central Spring Mountains, the Silurian Laketown Dolomite occurs beneath the Middle Devonian Sultan Limestone and thickens westward. Finally in the northwest Spring Mountains, Lower Devonian dolomite beds of the Nevada Formation are present beneath Middle Devonian rocks and thicken rapidly; farther west the unconformity within Devonian rocks is no longer recognized. Thus, in the northwest Spring Mountains, all formations seem conformable and range from late Precambrian to Mississippian and probably Late Permian in age.

Pre–Late Permian strata (that is, pre–Coconino Sandstone) on the craton are 7,700 ft (2,300 m) thick. Assuming that a thick section of Bird Spring Formation once overlay the rocks exposed in the northwest Spring Mountains, the total stratigraphic thickness of pre–Late Permian geosynclinal rocks may have been approximately 30,000 ft (9.0 km). Estimates of the contributions to westward thickening of the section are the following: (1) thickening of individual formations that carry through to the craton: 6,800 ft (2,070 m) or 22 percent; (2) addition of formations at unconformities: 4,600 ft (1,400 m) or 32 percent; and (3) addition of thick terrigenous late Precambrian sequence: 9,800 ft (3,000 m) or 46 percent. The Bird Spring Formation alone accounts for one-half of the total effect of formational thickening. Thus, excluding this formation, the main cause of geosynclinal thickening is the addition of the late Precambrian terrigenous wedge at the base.

Formations that overlie the Paleozoic carbonate sequence are largely terrigenous with marine carbonate rocks present only in Late Permian and Lower Triassic units. These terrigenous formations range in age from Lower Permian to Cretaceous(?) and are exposed only in the cratonal sequence at Frenchman Mountain and in the eastern part of the Spring Mountains (Figs. 2 and 3). The oldest of these terrigenous units is the Lower Permian red beds which lie below the Kaibab and Toroweap Formations of Leonardian age. In the eastern Spring Mountains, they conformably overlie Lower Permian rocks of the Bird Spring Formation. Because the Bird Spring Formation in the central Spring Mountains contains fusulinids of Leonardian age, part or all of the red beds in the east may be equivalent to marine limestone units farther west. The red beds grade upward into the marine limestone and evaporite beds of the Toroweap and Kaibab Formations. The Kaibab and Toroweap Formations wedge out to the northwest and are not present in the central Spring Mountains where they are cut out by pre-Moenkopi erosion.

The Moenkopi Formation of Early Triassic age was deposited unconformably on older formations. Unlike the unconformity at the base of the Sultan Limestone, which from west to east truncates progressively older formations, the pre-Moenkopi unconformity truncates older formations from east to west. A basal detrital and conglomeratic unit overlies the unconformity and is overlain in turn by shallow-water marine limestone correlative with the Virgin Limestone member of the Moenkopi Formation. These rocks represent the last identified marine strata in the Spring Mountains area. A thin sequence of red beds forms the uppermost beds of the Moenkopi; they are the youngest rocks present in the allochthonous terrane of the Spring Mountains and are present only in the Keystone thrust plate. Rocks younger than Bird Spring Formation are absent above the Lee Canyon thrust fault. Detrital rocks of the Moenkopi in the cratonal sequence on Frenchman Mountain and below the Keystone thrust in the eastern Spring Mountains continue unbroken into the Lower Jurassic Aztec Sandstone. The uppermost part of the Moenkopi Formation and the Aztec Sandstone are not present west of the Keystone thrust fault.

Locally, channels filled by conglomerate with red sandstone matrix are present at the top of the Aztec and immediately below the Keystone and Red Spring thrusts. Secor (1963) and Davis (1973) have interpreted these channel deposits of post-Aztec age and suggest that they may be as young as Cretaceous. In any case, these conglomerate units are the youngest rocks in the area and indicate that the Keystone and Red Spring thrust plates may have moved across erosional surfaces (see below).

Structure

Structure of the Spring Mountains north of Mountain Springs Summit is dominated by three thrust faults which trend north to northeast through the northwest-trending range (Fig. 4). Passing through the eastern part of the range is the Keystone thrust which forms the easternmost thrust of the Cordilleran orogen throughout most of the Spring Mountains as well as farther south. The Lee Canyon thrust cuts through the central part of the mountains, and the Wheeler Pass thrust cuts through the western part. Four thrusts of probably smaller magnitude are also present: (1) the Red Spring thrust which lies below the Keystone thrust northeast of Red Rock Canyon; (2) the Kyle Canyon thrust which crops out over a small area in the central Spring Mountains; (3) the Deer Creek thrust; and (4) the Macks Canyon thrust. The latter two faults crop out only in the north half of the range (Fig. 4).

Folds occur either as minor intrafolial or intraformational structures of limited lateral extent or as very large amplitude, long wave-length structures, involving most or all of the stratigraphic section in the area. The latter are commonly associated with thrust faults, which appear to postdate the folds (Fleck, 1970; Vincelette, 1964; Nolan, 1929). High-angle faults cut and are cut by thrust faults suggesting at least two periods of high-angle faulting. The high-angle faults most commonly trend northwest, although north-trending faults are numerous locally.

The structural grain of the northeastern half of the Spring Mountains has been rotated by late Tertiary movement on the Las Vegas Valley shear zone (Longwell, 1960; Fleck, 1967) which lies in Las Vegas Valley immediately northeast of the Spring Mountains. Thrust faults, folds, and bedding that strike north in the southwestern part of the range arc northeast adjacent to Las Vegas Valley. This bending was interpreted by Longwell (1960) to be right-lateral drag along the south side of the shear zone.

Figure 4. Generalized tectonic map, Spring Mountains, Nevada.

Structure of the Autochthon. Autochthonous rocks crop out only on the eastern slope of the Spring Mountains where they form a homocline of Mesozoic strata that dips gently westward. In Red Rock Canyon, the Mesozoic rocks are overturned eastward below the Keystone thrust. This fold plunges below the thrust and is not exposed north or south of the Red Rock Canyon area. East of Mountain Springs Summit, near the southern boundary of Figure 3, a northwest-striking high-angle fault, the Cottonwood fault (Hewett, 1931), juxtaposes Devonian and Mississippian rocks on the south and Jurassic Aztec Sandstone on the north. Rocks south of the fault are part of the Contact thrust plate (Hewett, 1931; Davis, 1973) which overrides Aztec Sandstone south of the area shown on Figure 3. Thus, rocks south of the Cottonwood fault are displaced downward several thousand feet relative to rocks north of the fault. The fault displaces the thrust vertically by only 200 ft (65 m), with the south side down, suggesting that most displacement on this fault was pre-Keystone in age. Other faults that cut both the autochthon and the Red Spring thrust farther north suggest similar age relations.

Keystone Thrust Plate. The Keystone thrust fault can be followed along the entire eastern face of the Spring Mountains although its continuation east of sec. 14, T. 20 S., R. 58 E. has not been mapped since the work of Longwell and others (1965). Along the fault, the Cambrian Bonanza King Formation was thrust eastward over Jurassic Aztec Sandstone and remnants of the Red Spring and Contact plates. Along its trace, the thrust dips west approximately parallel to bedding in the Bonanza King Formation above and the Aztec Sandstone below. In Red Rock Canyon, the thrust is exposed at a deeper level and dips approximately 30° west, still approximately parallel to bedding in the Bonanza King Formation of the upper plate but cutting across bedding in overturned Mesozoic formations of the lower plate. Farther east, the thrust cuts across rocks of the Red Spring thrust plate which are locally thrown into an overturned syncline (secs. 22 and 23, T. 20 S., R. 58 E.); dips along the thrust contact vary from 16° north to near vertical.

At Red Rock Canyon, the thrust lies at the base of the Banded Mountain Member of the Bonanza King Formation, but southward, rocks at the base of the plate are folded, faulted, and brecciated on a small scale; how closely the thrust follows stratigraphic horizons in the Bonanza King is obscure. On a regional basis, the thrust fault is located near the boundary between the two members of the Bonanza King Formation (Fig. 2) for a distance of more than 35 mi (60 km), strongly suggesting stratigraphic control of the thrust.

Structure within the Keystone thrust plate consists of folds, high-angle faults, and thrust faults, which are variably developed in three blocks bounded by two northwest-striking faults.

AGE OF DEFORMATION

The age or ages of deformation in the Spring Mountains is not closely bracketed by stratigraphic units. The Keystone and Red Spring thrusts can be placed only in the interval of Early Jurassic (post-Aztec) to pre-late Cenozoic (alluvial deposits). The age of the channel-fill deposits below the Keystone and Red Spring thrusts unfortunately is unknown.

Closer dating of the deformation comes from the Muddy Mountains 40 mi (67 km) to the northeast and from the Clark Mountains 30 mi (50 km) to the south. In the Muddy Mountains, the Willow Tank Formation and Baseline Sandstone overlie the Aztec Sandstone with slight angular unconformity. Fossils from the upper part of the Willow Tank yield early Late Cretaceous ages (Longwell, 1949), and ash beds in the lower part of the Willow Tank have yielded K/Ar ages of 98.4 and 96.4 m.y. or early Late Cretaceous (Fleck, 1970). Regional studies show that pre-Silurian strata were not exposed in southeastern Nevada until early Late Cretaceous time, since the earliest occurrence of detrital material from these rocks is in the Baseline Sandstone (Longwell, 1952; Armstrong, 1968). The younger Overton Fanglomerate appears to be syn- or post-tectonic, but its age is uncertain and is Cretaceous(?) or Tertiary(?). Potassium-argon ages of approximately 23 m.y. from ash beds lying conformably above the Overton were determined by Armstrong (1963); thus the upper limit of deformation in this area can only be dated as Miocene.

In the Clark Mountain area to the south, thrust faults, including the continuation of the Keystone thrust, are intruded by plutons that yield radiometric dates between 84 and 94 m.y. (Adams and others, 1968). These relations are consistent with regional data presented by Armstrong (1968) that thrusting in the eastern part of the Cordilleran orogen from central Utah to southern Nevada was concluded before the end of the Cretaceous.

Recent work by Burchfiel and Davis (1971) has demonstrated that in the Clark Mountains, thrusting began as early as Late Triassic or Early Jurassic time in some of the thrust plates above the Keystone plate. Structures from this older episode (or episodes) project northward and perhaps continue into the central part of the western Spring Mountains. The time of inception of deformation in the central and western Spring Mountains thus remains uncertain.

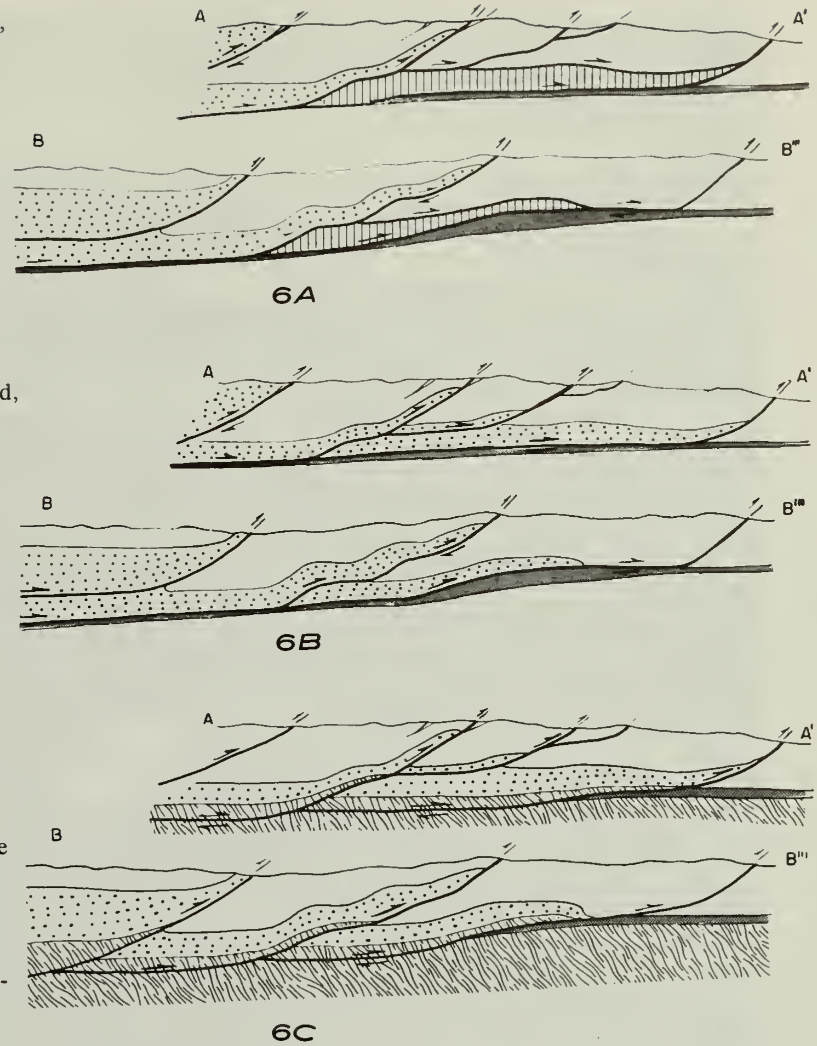


Figure 6. Interpretative cross sections from Figure 5. A. Space problem is solved by adding a thrust slice. B. Space problem is solved by adding late Precambrian terrigenous rocks to plates. C. Space problem is solved by adding Precambrian crystalline basement rocks to plates. Inclusion of crystalline basement in the Lee Canyon and Keystone thrust plates is to emphasize the nondécollement geometry which is possible. Note: high-angle faults have been removed, and topography is hypothetical. Dot pattern is late Precambrian terrigenous sequence, vertical line pattern is added thrust slice, gray pattern is autochthonous terrigenous sequence. and curved lines are Precambrian crystalline rocks.

SUMMARY

The Spring Mountains, Nevada, contain probably the most southern example of "typical" eastern Cordillera structure. Three major east-directed thrust plates move geosynclinal Paleozoic and late Precambrian sedimentary rocks eastward over cratonal equivalents. Too much uncertainty exists in the projection of thrust faults to depth to determine the involvement of older basement rocks, although the reconstructions suggest that these rocks may be present in the Wheeler Pass thrust plate. Thrusting in the Spring Mountains produced a minimum shortening estimated from the décollement model to be between 22 and 45 mi (36.6 and 75 km). Some deformation occurred during Late Cretaceous time, but recent work suggests that part of the deformation could be early or middle Mesozoic in age.

Stratigraphy across the Spring Mountains shows a transition from a Paleozoic cratonal sequence in the east to a geosynclinal sequence in the west. Addition of a thick section of late Precambrian terrigenous rocks at the base of the geosynclinal sequence accounts for nearly one-half of the thickening in the geosynclinal sequence. Where this section of terrigenous rocks first occurs beneath the Spring Mountains is unknown, but geometry of the thrust plates at depth suggests they may occur in or below the Keystone thrust.

REFERENCES CITED

Adams, J.A.S., Burchfiel, B. C., Davis, G. A., and Sutter, J. F., 1968, Absolute geochronology of Mesozoic orogenies, southeastern California: Geol. Soc. America, Abs for 1966, Spec. Paper 101, p. 1.

Armstrong, R. L., 1963, Geochronology and geology of the eastern Great Basin in Nevada and Utah [Ph.D. thesis]: New Haven, Conn., Yale Univ., 202 p.

——— 1968, Sevier orogenic belt in Nevada and Utah: Geol. Soc. America Bull., v. 79, p. 429–458.

Bowyer, B., Pampeyan, E. H., and Longwell, C. R., 1958, Geologic map of Clark County, Nevada: U.S. Geol. Survey Mineral Inv. Field Studies Map MF 138.

Burchfiel, B. C., 1964, Precambrian and Paleozoic stratigraphy of the Specter Range quadrangle, Nye County, Nevada: Am. Assoc. Petroleum Geologists Bull., v. 48, p. 40–56.

Burchfiel, B. C., and Davis, G. A., 1971, Clark Mountain thrust complex in the Cordillera of southeastern California: Geologic summary and field trip guide: California Univ., Riverside, Campus Mus. Contr., no. 1, p. 1–28.

——— 1972, Structural framework and evolution of the southern part of the Cordilleran orogen, western United States: Am. Jour. Sci., v. 272, p. 97–118.

Davis, G. A., 1973, Relations between the Keystone and Red Spring thrust faults, eastern Spring Mountains, Nevada: Geol. Soc. America Bull., v. 84, p. 3709–3716.

Fleck, R. J., 1967, The magnitude, sequence, and style of deformation in southern Nevada and eastern California [Ph.D. thesis]: Berkeley, Univ. California, 92 p.

——— 1970a, Tectonic style, magnitude, and age of deformation in the Sevier orogenic belt in southern Nevada and eastern California: Geol. Soc. America Bull., v. 81, p. 1705–1720.

——— 1970b, Age and possible origin of the Las Vegas Valley shear zone, Clark and Nye Counties, Nevada: Geol. Soc. America, Abs. with Programs (Rocky Mountain Sec.), v. 2, no. 5, p 333.

——— 1974, Geology of the Charleston Peak and part of the Corn Creek Springs quadrangles, Clark Co., Nevada: Nevada Bur. Mines Bull (in press).

Gans, W. T., 1970, The detailed stratigraphy of the Goodsprings Dolomite, southeastern Nevada–California [Ph.D. thesis]: Houston, Texas, Rice Univ., 87 p.

Gilbert, G. K., 1875, Report on the geology of portions of Nevada, Utah, California, and Arizona, p. 17–187, Pt. I, *in* United States Geographical and geological surveys west of the 100th meridian, Vol. 3: 681 p.

Glock, W. S., 1929, Geology of the east-central part of the Spring Mountain Range, Nevada: Am. Jour. Sci., v. 217, p. 326–341.

Hewett, D. F., 1931, Geology and ore deposits of the Goodsprings quadrangle, Nevada: U.S. Geol. Survey Prof. Paper 162, 172 p.

Longwell, C. R., 1949, Structure in the Arizona-Nevada boundary area [abs.]: Geol. Soc. America Bull., v. 60, p. 1905–1906.

——— 1952, Structure of the Muddy Mountains, Nevada: Utah Geol. and Mineralog. Survey Guidebook no. 7, p. 109–114.

——— 1960, Possible explanation of diverse structural patterns in southern Nevada: Am. Jour. Sci., v. 258-A, p. 192–203.

Longwell, C. R., Pampeyan, E. H., Bowyer, B., and Roberts, R. J., 1965, Geology and mineral deposits of Clark County, Nevada: Nevada Bur. Mines Bull. 62, 218 p.

Nolan, T. B., 1929, Notes on the stratigraphy and structure of the northwest portion of the Spring Mountains, Nevada: Am. Jour. Sci., v. 217, p. 461–472.

Secor, D. T., Jr., 1963, Geology of the central Spring Mountains, Nevada [Ph.D. thesis]: Stanford, Calif., Stanford Univ., 152 p.

Spurr, J. E., 1903, Descriptive geology of Nevada south of the fortieth parallel and adjacent portions of California: U.S. Geol. Survey Bull. 208, 229 p.

Stanley, K. O., Jordan, W. M., and Dott, R. H., Jr., 1971, Early Jurassic paleogeography, western United States: Am. Assoc. Petroleum Geologists Bull., v. 55, p. 10–19.

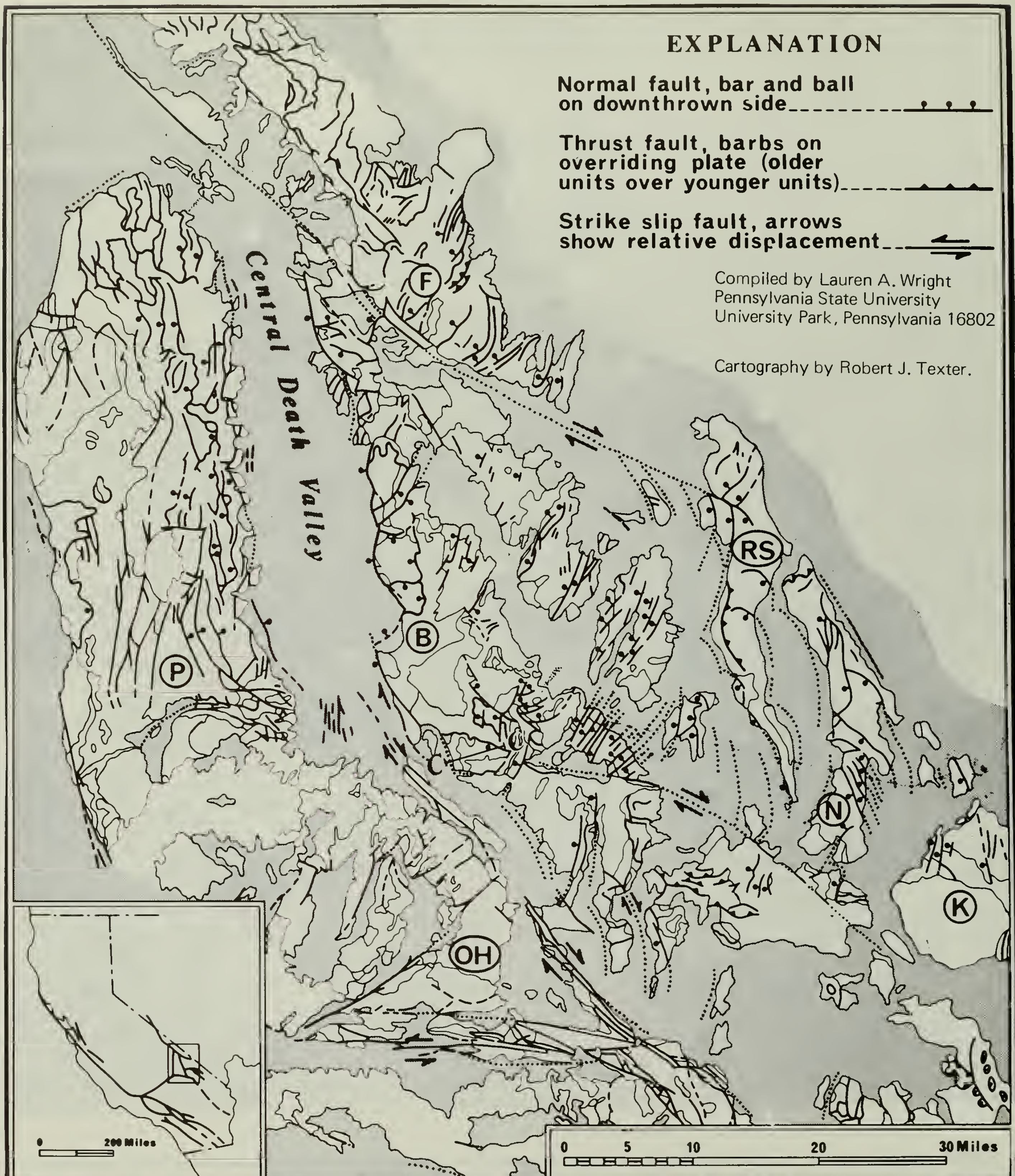
Stewart, J. H., 1970, Upper Precambrian and Lower Cambrian strata in the southern Great Basin, California and Nevada: U.S. Geol. Survey Prof. Paper 620, 206 p.

Vincelette, R. R., 1964, Structural geology of the Mt. Stirling quadrangle, Nevada, and related scale-model experiments [Ph.D. thesis]: Stanford, Calif., Stanford Univ., 141 p.

Preprinted (in part) by permission from Geological Society of America *Bulletin*, July, 1974. Subject to revision before final printing.

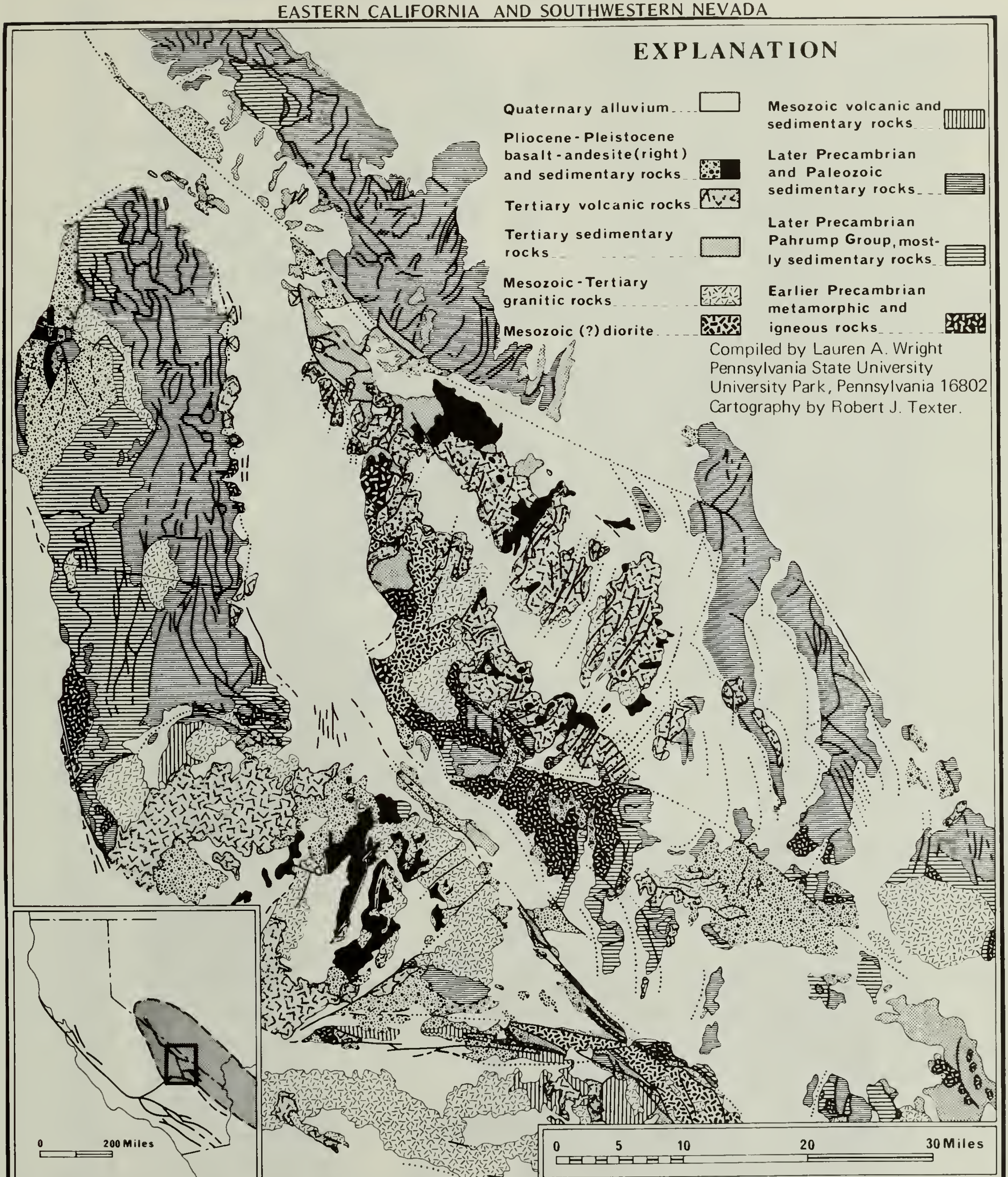
Oblique aerial view westward across central part of Death Valley. From lower to upper part of photograph are Greenwater Range, upper Furnace Creek Wash, Black Mountains, Death Valley, Panamint Range, Panamint Valley, Slate and Argus Ranges. Smaller ranges and valleys beyond not readily distinguished. Sierra Nevada forms skyline range. Badwater lies at east edge of Black Mountains at near edges of pale playa surface that lies athwart Death Valley. U.S. Geological Survey-U.S. Air Force photograph.

FAULT MAP OF THE REGION OF CENTRAL AND SOUTHERN DEATH VALLEY, EASTERN CALIFORNIA AND WESTERN NEVADA



Shaded pattern denotes Quaternary alluvium; blank areas denote pre-Quaternary rocks. B = Black Mountains; F = Funeral Mountains; K = Kingston Range; N = Nopah Range; OH = Owlshead Mountains; P = Panamint Range; RS = Resting Spring Range. Reproduced by permission and modified from *Gravity and Tectonics*, edited by Kees A. De Jong and Robert Scholtn (New York: John Wiley & Sons), 1973.

GEOLOGIC MAP OF THE REGION OF CENTRAL AND SOUTHERN DEATH VALLEY, EASTERN CALIFORNIA AND SOUTHWESTERN NEVADA



See fault map for identification of principal mountain ranges. Data from various sources acknowledged on Death Valley Sheet (1958; revision in preparation, 1973), and Trona Sheet, Geologic Map of California, California Division of Mines and Geology. Reproduced by permission and modified from *Gravity and Tectonics*, edited by Kees A. De Jong and Robert Scholtn (New York: John Wiley & Sons), 1973.

Low-altitude oblique aerial photograph to northeast of west side of southern Panamint Range. Mouth of Goler Canyon is in lower center of photograph. Pale rocks left of Goler Canyon are patch of the platform facies of the Noonday Dolomite. Basin facies crop out south of Goler Canyon (see preceding article). Crystal Spring and Kingston Peak Formations form most of the dark outcrops in foreground (see article by Roberts). Pale rocks (right of center) are Tertiary volcanic rocks. Dark rocks in foreground are intruded by granitic rocks (center and left center). Death Valley is barely discernable beyond the Panamint Range. *Photo 105 by John H. Maxson; courtesy of the National Park Service.*

Precambrian Sedimentary Environments of the Death Valley Region, Eastern California

L. A. Wright,[1] B. W. Troxel,[2] E. G. Williams,[1] M. T. Roberts,[1] P. E. Diehl[1]

INTRODUCTION

Noteworthy among the geologic features of the Death Valley region of eastern California is an extensively exposed accumulation of sedimentary rock and sill-forming diabase, Precambrian in age, as much as 5,500 m thick, and essentially unmetamorphosed. Although these rocks have yet to yield a reliable radiometric date, the oldest probably were deposited about 1.4 b.y. ago, as they rest with profound unconformity upon a crystalline complex from which 1.7-b.y. dates have been obtained, and they have been intruded by diabase sills and dikes that seem best correlated with similarly disposed diabase bodies, of central and southern Arizona, that are about 1.2 b.y. old. The youngest Precambrian strata of the Death Valley region conformably underlie strata that contain Early Cambrian fossils.

This Precambrian section (Fig. 1) has long been recognized as composed of (1) a lower succession, named the Pahrump Group and referred to by some as "Beltian" in age, and (2) an upper succession divided into several formations (Fig. 1) comparable in stratigraphic position with the Windermere Group of the more northerly parts of the North American Cordillera.

The two successions were originally defined as separated by an angular unconformity beneath the Noonday Dolomite (Noble, 1934; Hazzard, 1937; Hewett, 1956). The unconformity causes the Noonday to rest variously upon each of three formations of the Pahrump Group and upon the older crystalline complex.

Most workers in the region have assumed that somewhere in this impressive accumulation of sedimentary units and sills is recorded a change in provenance from one that was initiated with the beginning of Pahrump sedimentation and has remained little understood, to another that constitutes the provenance of the Cordilleran miogeosyncline. The sub-Noonday unconformity has frequently been inferred to be the principal indicator of this change, so that published stratigraphic cross sections of the southern part of the Cordilleran miogeosyncline consistently show the Noonday as the lowest formation in it.

Not everyone has accepted this interpretation, however, partly because carbonate rock rarely forms the initial deposit in a developing miogeosynclinal environment. Noble (1941, p. 592) observed that the entire Johnnie Formation, which overlies the Noonday, is lithologically more similar to parts of the Pahrump Group than to any of the younger formations, and on many occasions, he informally expressed the belief that the deposition of the enormous volume of detrital silica preserved in the Stirling Quartzite marked the initiation of the miogeosyncline. Stewart (1972), on the other hand, has suggested that the change in provenance is recorded in the pre-Noonday conglomeratic units, commonly diamictite, of the Kingston Peak Formation.

We will briefly summarize data, much of it recently acquired and described more completely elsewhere in this volume, that relates to the nature of sedimentary environments that existed before the beginning of the Cordilleran miogeosyncline. These data indicate that the occurrences of the Pahrump Group in the southern Death Valley region were deposited within a long-continuing west-northwest-trending trough or basin (Figs. 2 and 3) about 50 km wide and at least 140 km long, with bordering platform areas that at times were inundated by shallow seas and at other times stood well above sea level to become source areas of basinal clastic sediments. The structural and stratigraphic characteristics of this feature suggest that it is an aulacogen precisely in the sense that aulacogens are defined and characterized in a recent review by Hoffman and others (1974). As indicated by them, aulacogens were recognized by the Russian geologist Shatiski and consist of "long-lived, deeply subsiding troughs, at times fault-bounded. that extend at high angles from geosynclines far into adjacent foreland platforms." The trough that controlled Pahrump and subsequent sedimentation is named by us the Amargosa aulacogen. Its bordering, generally high areas to the north and south we designate as the Nopah (Wright and Troxel, 1967) and the Mojave uplands, respectively (Fig. 3B).

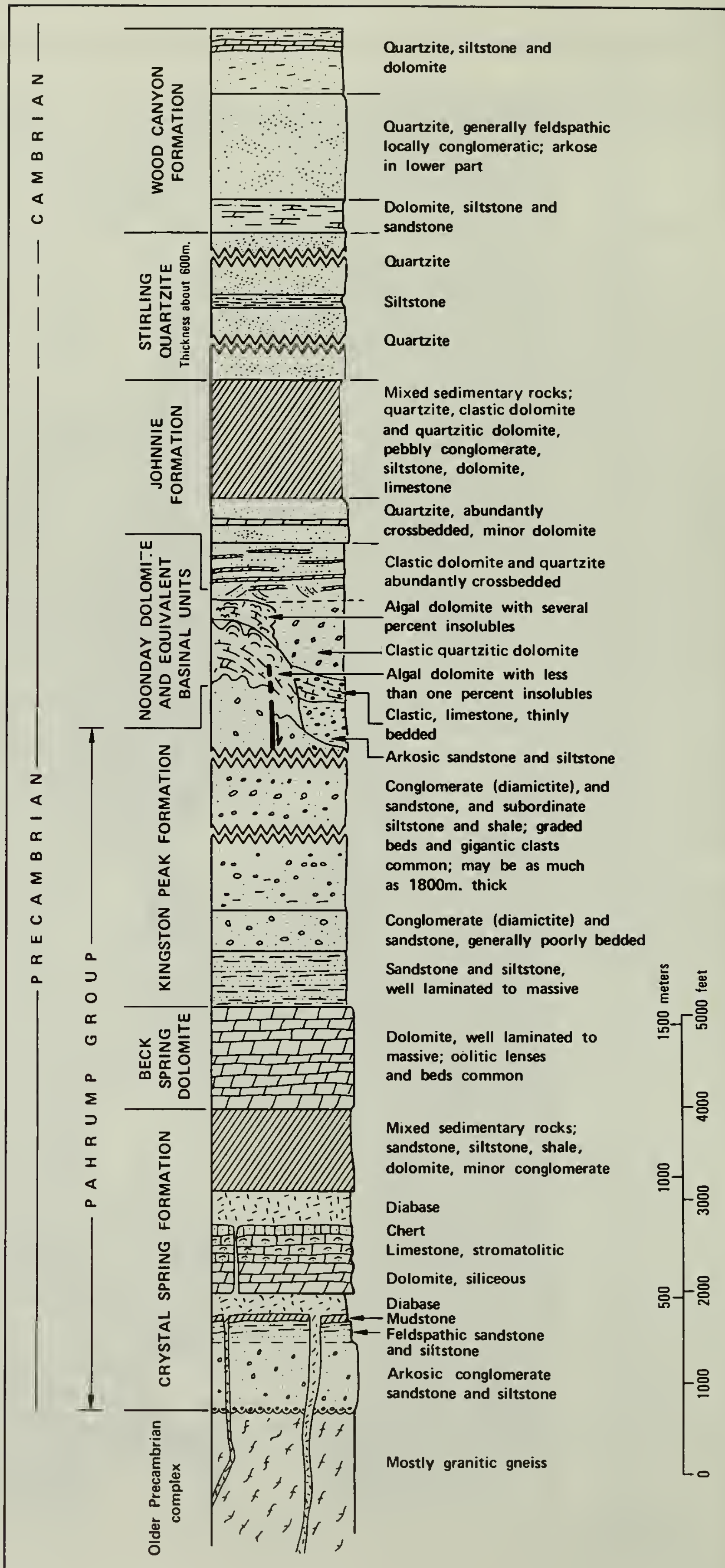
As the Noonday Dolomite passes abruptly southward into time-equivalent basinal deposits that are approximately coextensive with basinal units of the underlying Pahrump Group, we view the Noonday and its equivalent units as deposited within the provenance of the Amargosa aulacogen rather than within the Cordilleran miogeosyncline. The sub-Noonday unconformity we attribute to strong post-Pahrump emergence along the northern margin of the aulacogen. Thus interpreted, the Noonday Dolomite marks a carbonate platform developed on the deeply eroded edge of a trough.

We also observe evidence that the overlying Johnnie Formation and Stirling Quartzite were deposited on a south to south-southwesterly paleoslope, tectonically perpetuated since the beginning of Pahrump time, and that the northwesterly paleoslope, which seems to have characterized the part of the miogeosyncline that occupied the site of the southern Death Valley region, became firmly established with the deposition of the Wood Canyon Formation. The uplift of the Nopah upland apparently recurred briefly and finally to produce a southwesterly paleoslope upon which were deposited arkosic strata in the lower part of the middle members of the Wood Canyon (Diehl, this volume).

The existence of the depositional trough and the west-northwest orientation of its axis is indicated in numerous stratigraphic and sedimentologic features of the three Pahrump formations and the Noonday. Isopachous lines drawn on these formations collectively, on each formation, and on most individual members display a troughlike configuration. With respect to this configuration, deep-water facies, where present in the section, occupy the center of the trough; some carbonate units intertongue with siliceous clastic strata near the margins; clastic units of various types fine troughward or from one side of the trough to the other; and various current direction features indicate paleoslope directions consistent with a west-northwest-trending trough. Changes in facies and thickness are characteristically abrupt and commonly occur on either side of mappable faults or monoclines. The formation of the trough is thus largely attributable to vertical

[1]*Pennsylvania State University, University Park, Pennsylvania 16802.*
[2]*Geological Society of America, Boulder, Colorado 80301.*

Figure 1. Generalized columnar section of Precambrian to Lower Cambrian strata, Death Valley region.



movements on fault-bounded blocks. The lower part of the Kingston Peak Formation, on the other hand, probably was deposited while the earlier underlying strata were being folded into a broad downwarp.

NOPAH UPLAND

The northerly margin of the trough was first detected in the reconstruction of paleogeologic contacts on the sub-Noonday surface (Wright and Troxel, 1967; Fig. 3C, this paper). Their configuration indicates that, prior to the deposition of the Noonday Dolomite, an uplifted positive area of crystalline rocks extended from the present Black Mountains for at least 100 km eastward and was flanked southward by progressively younger units of the Pahrump Group. Additional evidence for the uplift and for the conclusion that the two lower formations of the Pahrump Group, namely the Crystal Spring Formation and the Beck Spring Dolomite, once extended northward into the area of crystalline rocks on the paleogeologic map and were eroded away in pre-Noonday time (Wright and Troxel, 1967; Troxel, 1967), is contained in the observations that Crystal Spring and Beck Spring debris is abundant in the conglomeratic units of the Kingston Peak Formation and that the debris was transported southward.

Other observations, made subsequently, indicate that this northerly positive area and the south to south-southwesterly paleoslope were in existence when the earliest of the Pahrump strata were deposited and that they persisted through all of Pahrump time. Roberts (this volume) has shown, for example, that the Crystal Spring Formation displays (1) a southward fining in average grain size and in maximum size of clasts in the sandstone-conglomerate arkosic strata that compose its lower part, and (2) a general south-southwest orientation of cross-bedding in conglomeratic strata in that unit (Fig. 3D). The Beck Spring Dolomite, which in the central part of the trough is free of siliceous detrital beds, contains two tongues of fine-grained quartzite in exposures along the indicated site of the basin margin immediately south of Ashford Canyon on the west face of the Black Mountains.

The Kingston Peak Formation, which remains to be studied in detail, shows evidence of the Nopah upland by a marked northward thinning (Fig. 4A), as well as by its content of debris derived from the Crystal Spring Formation and Beck Spring Dolomite. The thinning is well displayed in exposures (1) in the Black Mountains north of the highway between Jubilee Pass and the Ashford Mill site (L. A. Wright and B. W. Troxel, 1974, unpub. data), (2) in the Alexander Hills at the south end of Nopah Range (Wright, 1973), and (3) in the vicinity of the Excelsior talc mine on the northeastern side of the Kingston Range (Wright, 1968). It is caused in part by erosion that preceded the deposition of the Noonday Dolomite, in part by the fact that some units that were deposited in the central part of the basin were not received by the topographically higher fault blocks along the margins of the basin, and in part by a stratigraphic thinning of individual units (Fig. 2).

The southward change in facies of the Noonday Dolomite,

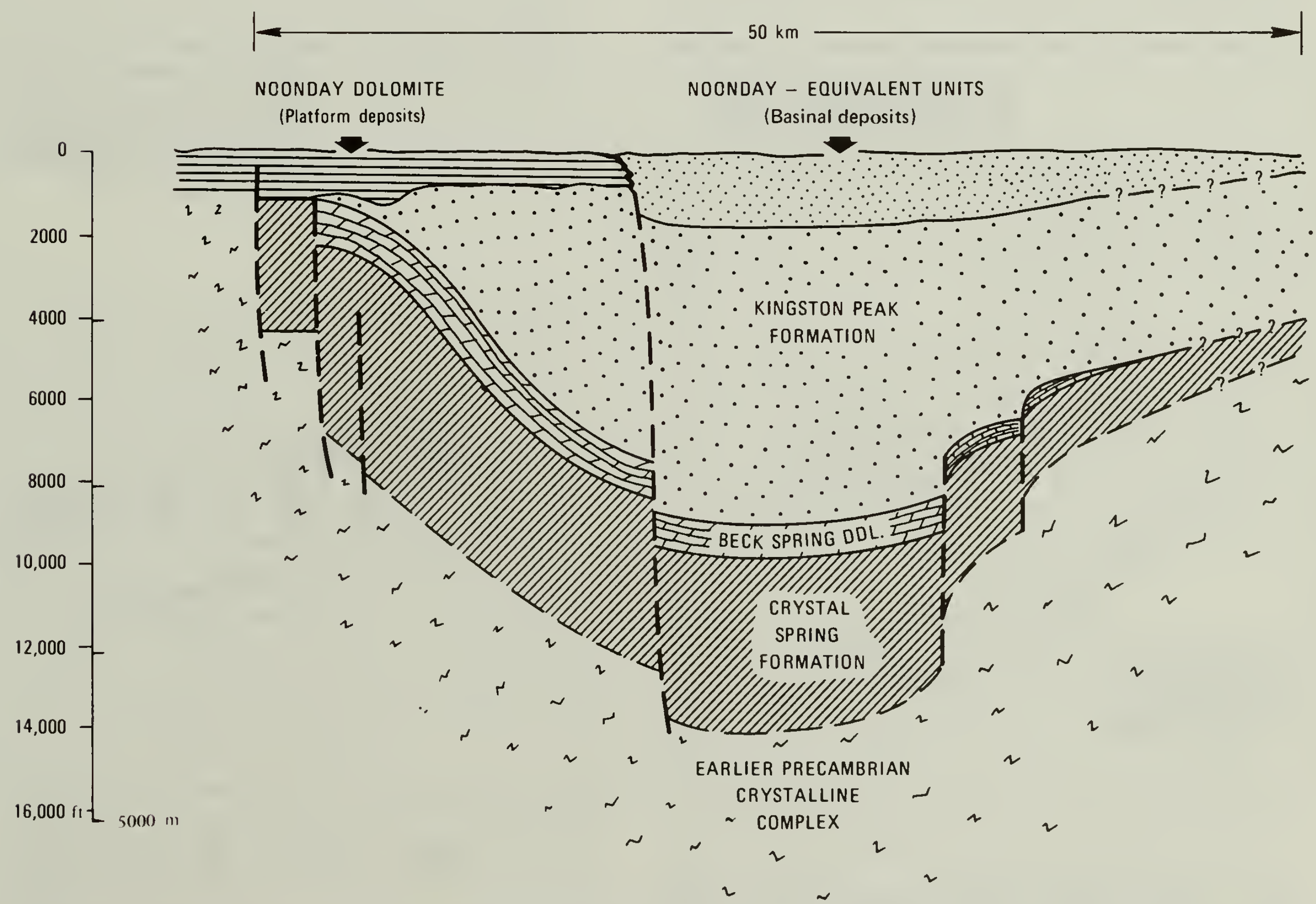


Figure 2. Diagrammatic cross section of Amargosa aulacogen at end of deposition of Noonday Dolomite and equivalent basinal sediments.

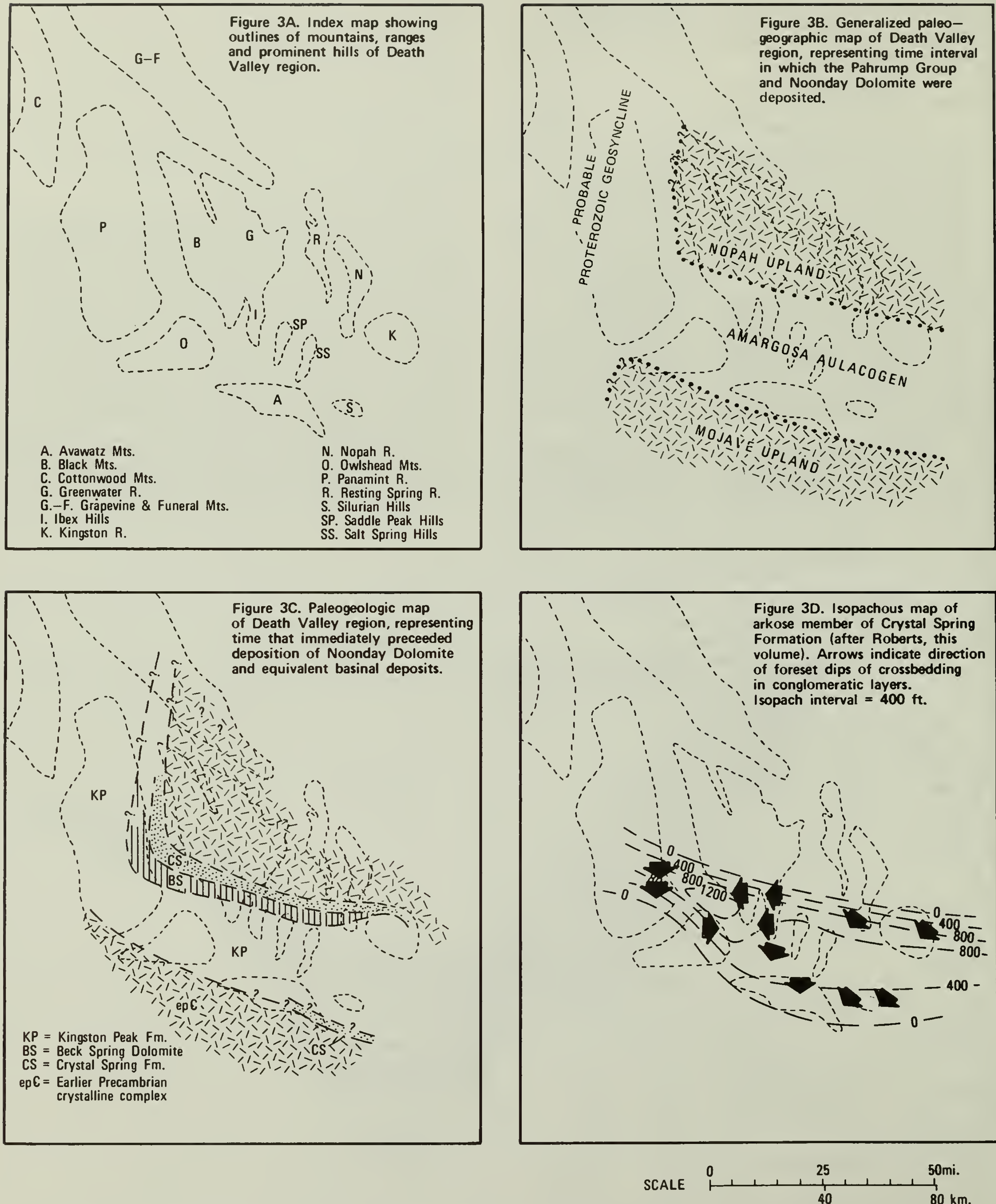


Figure 3. Paleogeographic maps of times when Pahramp Group and Noonday Dolomite were deposited, and isopachous map of Crystal Spring Formation.

from the platform carbonate to basinal strata (Williams and others, this volume), records a still later phase in the stratigraphic-tectonic development of the Amargosa aulacogen and its northern margin. The exposures of the Noonday that extend from the southwestern part of the Panamint Range for about 140 km south-southeastward (Figs. 2 and 4B) consist almost entirely of dolomite that has formed in the presence of algal mats. At various localities these platform deposits, which ordinarily are 200 m or less thick, can be traced abruptly into basin deposits two to three times at thick. In most places the latter are composed, in upward succession, of arkosic sandstone and siltstone in which graded bedding is common, thinly bedded clastic limestone and dolomite, and an almost structureless detrital unit of dolomite and quartz. Clastic dolomite and quartz also compose a strongly cross-bedded sandstone unit that overlies both the dolomite of the platform and the structureless unit of the basin and marks the transition from the Noonday Dolomite and equivalent units into the overlying Johnnie Formation.

A northerly to northeasterly source for the dolomite and quartz grains of the uppermost of the basinal units and the cross-bedded unit must be invoked to explain the intimate association of the two minerals, the quartz grains originating beyond the limits of the carbonate cover of the platform, passing over it, and mixing with clastic dolomite where the cover was being destroyed by wave or stream action. Measurements of current directions in the cross-bedded strata indicate a generally east-southeast strike for the paleoslope, essentially unchanged from the strike of the slope upon which the lowermost of the Pahrump strata were deposited. Indeed, measurements made to date of cross-bedding orientation in sandstone throughout the Johnnie Formation and Stirling Quartzite suggest that this strike persisted during the deposition of both (Fig. 4D). As these formations are exposed at various localities for many kilometers to the northeast of the basin, the source areas of the detrital material must have been much more distant from the basin site than the source areas of the detrital strata of the Pahrump units.

MOJAVE UPLAND

Evidence for exact location of the southern margin of the Amargosa aulacogen is less clear than that for the site of the northern margin, as Mesozoic and Cenozoic deformational and intrusive events have obscured the Precambrian record there. In the Silurian Hills, however, a conformable section of Pahrump strata, at least 2,220 m thick, has been identified (Kupfer, 1960; Wright and Troxel, 1966). The lower 300 m of this section is lithologically correlative with the lower and middle units of the Crystal Spring Formation as recognized elsewhere (Roberts, this volume). An apparently complete and thicker occurrence of the Crystal Spring also is exposed near the mouth of Sheep Creek Canyon on the north slope of the Avawatz Mountains. The distribution of a complex of post-Pahrump sedimentary units and Mesozoic(?) diorite, on either side of the valley that separates the Avawatz Mountains from the Silurian Hills, suggests right-lateral displacement of about 8 mi (13 km) and thus a strong possibility that these two Pahrump localities were once approximately opposite each other (Fig. 5D) on either side of the southern extension of Death Valley.

Exposures and intact sections of Pahrump- and Noonday-equivalent units in the southern part of the study area, although relatively few, indicate the existence of a southerly positive from which much detritus was fed into the Amargosa aulacogen (Fig. 2).

The earliest evidence of a southerly tectonic high is the southward thinning of the shallow-water arkosic lower part of the Crystal Spring (Roberts, this volume; Fig. 2), although Roberts has concluded that most or all of this detritus was derived from the northern upland. Quartzite layers in the middle part of the Crystal Spring Formation in the Silurian Hills (Kupfer, 1960) are interlayered with subordinate carbonate strata, the lower of which are dolomite and the upper of which are limestone. The clastic material is attributable to a southerly source. To the north, this part of the formation generally consists of two carbonate members, a lower member of siliceous dolomite and an upper limestone member containing stromatolites

The higher part of the section in the Silurian Hills, as Kupfer (1960) observed, consists mostly of sandstone and conglomerate. Although he assigned the entire section to the Pahrump Group, he hesitated to make firm correlations with the established formations of the Pahrump. He noted, however, that a limestone unit, about 30 m thick and about 780 m above the base of the section, may be correlative with the Beck Spring Dolomite. He also observed that, in general, the section of the Silurian Hills contains coarser clastic rocks and a smaller proportion of carbonate material than those farther north. He thus proposed a southerly source for the clastic material.

Subsequent observations of the more northerly of the Pahrump occurrences support Kupfer's observations and interpretations to the extent that they apply to the upper part of the Crystal Spring Formation and the Beck Spring Dolomite. In the Ibex-Saratoga Hills the Beck Spring Dolomite shows a marked southward increase in insoluble material in the form of shaly and sandy layers. The upper units of the Crystal Spring in the central and northern parts of the basin are distinctly finer grained and contain a higher proportion of carbonate material than do the strata beneath the aforementioned limestone unit in the Silurian Hills.

Although most of the clastic material in the Kingston Peak Formation is traceable to the northerly source area, Troxel (1967) has recognized a southern Kingston Peak facies which contains clasts of metamorphic rocks unlike those in the rest of the formation and apparently derived from the south (Fig. 4A). Troxel has since observed that the two facies intertongue in a more complex fashion than his published description implies.

A southern source area also is indicated for the unit of arkosic sandstone and siltstone in the succession of strata that forms the basinal equivalent of the Noonday Dolomite (Fig. 4B). As observed by Williams and others (this volume), this unit contains bottom markings that evidence northerly transport, displays a northerly fining, and, in its northernmost exposures, shows an on-lapping relationship with the lower dolomite of the platform. The Nopah upland, at that time, was a low area largely or entirely covered by carbonate rock and thus could not have provided detritus of arkosic composition.

Stratigraphic-Tectonic Environments of the Amargosa Aulacogen

In brief, the various stratigraphic and structural features of the Amargosa aulacogen record a history divisible into two stages. Each is characterized by (1) an early period of erosion of the northern (Nopah) upland concurrent with clastic sedimentation in the subsiding trough, and (2) a later period of stability when that upland became a surface of low relief to serve as a platform and receive a cover of carbonate rocks. The southern (Mojave) upland apparently also was elevated with respect to sea level in two stages, one preceding and the other following the deposition of the Beck Spring–equivalent(?) unit of limestone in the Silurian Hills. These were approximately

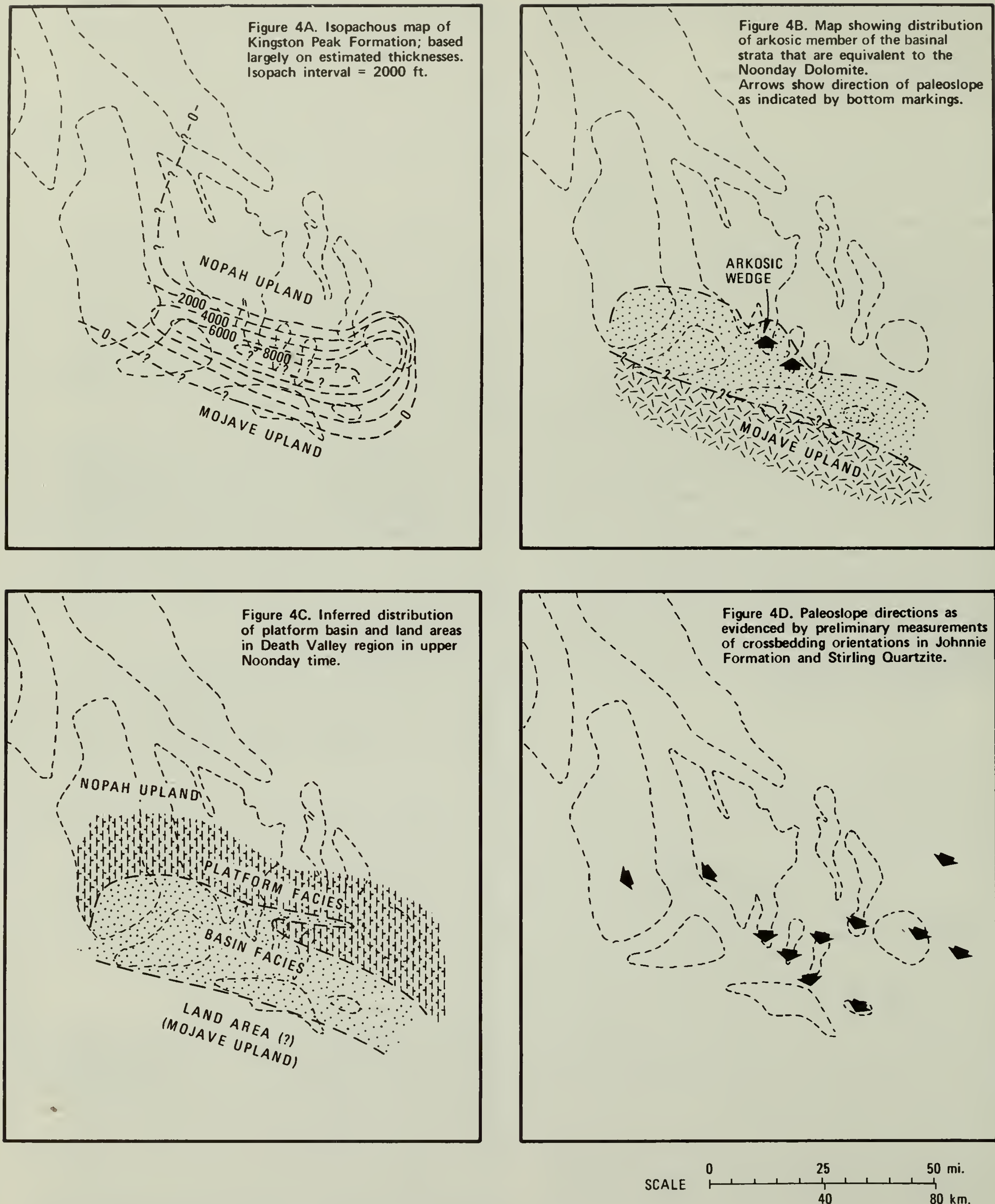


Figure 4A. Isopachous map of Kingston Peak Formation; based largely on estimated thicknesses. Isopach interval = 2000 ft.

Figure 4B. Map showing distribution of arkosic member of the basinal strata that are equivalent to the Noonday Dolomite. Arrows show direction of paleoslope as indicated by bottom markings.

Figure 4C. Inferred distribution of platform basin and land areas in Death Valley region in upper Noonday time.

Figure 4D. Paleoslope directions as evidenced by preliminary measurements of crossbedding orientations in Johnnie Formation and Stirling Quartzite.

Figure 4. Isopachous map of Kingston Peak Formation, distribution map of Noonday Dolomite and equivalent basinal facies, and paleoslope directions of Johnnie and Stirling, Death Valley region. See Figure 3a for identification of mountains.

concurrent with the two erosional periods of the Nopah upland. But the trough received detritus from the northerly source considerably earlier than it did from the southerly source. The latter, however, may then have remained continually above sea level and, almost continually, a source area for clastic sediment. If it did, at times, receive carbonate platform deposits, a record of the covering remains unobserved either in situ or in the basinal detritus.

The two stages, although grossly similar, differed from each other in several major respects. The first stage, as recorded in the sediments of the Crystal Spring Formation and Beck Spring Dolomite, was featured by relatively slow subsidence compared with the subsidence rate during the second stage and shallow-water environments. Shallow water is indicated by abundant cross-bedding, ripple marks, and mud cracks in the units of clastic strata in the Crystal Spring, by algal stromatolites in the carbonate members of the Crystal Spring, and by algal stromatolites and layers of oolite in the Beck Spring. A progressive slowing of subsidence, with respect to sea level, and a concurrent lowering of the Nopah upland are indicated by the overall upward fining displayed by the clastic units, and by the deposition, in the later part of the first stage, of carbonate units across the trough site and onto the Nopah upland, which had been eroded to a platform. The emplacement of bodies of basic igneous rock, represented by the sill-forming diabase of the Crystal Spring Formation and diabase dikes in the crystalline complex, was confined almost entirely to the first stage, although basalt flows have been observed locally in the Kingston Peak Formation (A. L. Albee, personal communs.), as have thin layers of pale, fine-grained material that may be altered volcanic ash.

The second stage, which is recorded in the Kingston Peak Formation, the Noonday Dolomite, and the Noonday-equivalent basinal strata, on the other hand, was marked by higher marginal relief than is evidenced in the first stage, and by rapid subsidence of the trough. The central part received deep water sediment throughout the stage.

The Kingston Peak and Noonday-equivalent strata of the central trough consistently show evidence of deposition below wave base. Most display graded bedding, some are thinly and continuously laminated, others are massive and essentially structureless, and all are devoid of the shallow-water features that characterize the sedimentary rocks of the first stage. The conglomeratic bodies of the Kingston Peak Formation that lie along the trough margins have long been believed to represent alluvial fans (Hewett, 1956; Kupfer, 1960).

In contrast to the modest marginal uplift and basin sinking of the first stage, which diminished with time, the basinal sinking and marginal relief related to the first half of the second stage apparently increased, as evidenced in a general upward coarsening of the Kingston Peak clastic sediments and an increase in the proportion of graded beds of sand. Conglomerate layers high in that formation contain clasts as much as 0.3 km long. Boulders associated with the deep-water turbidite and diamictite layers of the Kingston Peak, some of which are striated, are believed by some to constitute evidence of glacial rafting, but their presence in deep-water sediments might also be explained by transportation in dense submarine debris flows.

In the later part of the second stage, the Nopah upland was again reduced to a surface of low relief, but it was featured along its southern margin by ridges, as much as 300 m high, of resistant conglomerate of the Kingston Peak Formation. On this platform was deposited the algal carbonate of the Noonday Dolomite, while the trough was receiving the deep-water basinal arkosic deposits derived from the Mojave upland. Still later, the basin was filled and the platform was largely covered by the dolomite-quartz sandstone and breccia transported southward.

That the Amargosa aulacogen is open and joins on the west to a geosyncline that was simultaneously receiving sediments like those of the aulacogen is suggested by the distribution of the most westerly and northerly exposures of units correlative with the Pahrump Group. These units occur in a belt that lies at a high angle to the aulacogen axis and extends along the west side of the Panamint Range as far north as Tucki Mountain and thence to the northern part of the Funeral Mountains on the northeast side of Death Valley (Fig. 3B). They remain to be examined in detail, but are known to comprise a lower part composed of siliceous clastic and carbonate rocks intruded by basic sills and an upper part composed of conglomerate with carbonate clasts. Farther to the west and north only younger rocks are exposed.

In summary, the various structural, stratigraphic, and igneous features that compose the Pahrump-Noonday terrane and its two-stage evolution correspond in general to the distinguishing features attributed by Hoffman and others (1974) to a typical aulacogen. The trough that received the basinal Pahrump and basinal Noonday-equivalent units was long-lived, being evidenced in stratigraphic units that cover the approximate time span of 1.4 to 0.7 b.y. If we have correctly interpreted the distribution of the most westerly and northerly of the Pahrump exposures, the trough joins with a geosyncline and lies at a high angle to the geosynclinal margin. The structural evolution of the trough and the nature of the sediments it received was controlled primarily by vertical movements on faults that bound the blocks on the trough margin, moderate in the earlier stage and strong during most of the later stage of trough development. Igneous activity, represented entirely by basic rock, occurred mostly in the earlier stage when dikes and sills of diabase were emplaced.

Aulacogen-Miogeosyncline Transition

The problem of identifying the record of transition from an aulacogenic to a miogeosynclinal environment in the Precambrian strata east of the Death Valley region remains unresolved, awaiting the acquisition of further data pertinent to the depositional environments of the Johnnie, Stirling, and Wood Canyon. When the quartzite member in the middle of the Johnnie (Fig. 1) was deposited, the platform-trough boundaries of the proposed aulacogen were apparently unexpressed topographically as the unit can be traced through the site of the trough and the immediately bordering uplands with no obvious variations except a thinning southward or southwestward. But the south to southwest paleoslope that is strongly suggested by paleocurrent directional features throughout the Johnnie is oriented similarly to the paleoslope upon which the earliest of the Pahrump strata were deposited. Measurements of limited numbers of directional features in the Stirling Quartzite suggest continuation of the south to southwest paleoslope for that formation in the Death Valley region, although the source or sources of the detrital quartz probably lay well beyond the sites of the aulacogen margins.

As Diehl (this volume) suggests, the principal indication of the change in provenance may well be the change to a northwesterly orientation in paleoslope evidenced in the cross-bedding and carbonate units thickening through most of the Wood Canyon Formation in the sections that he has studied (Fig. 5, A and B). As only the arkose and arkosic conglomerate beds low in the middle member of the Wood Canyon display cross-bedding that is consistently oriented southwestward, these possibly constitute the final record of uplift related to the pre-miogeosynclinal provenance.

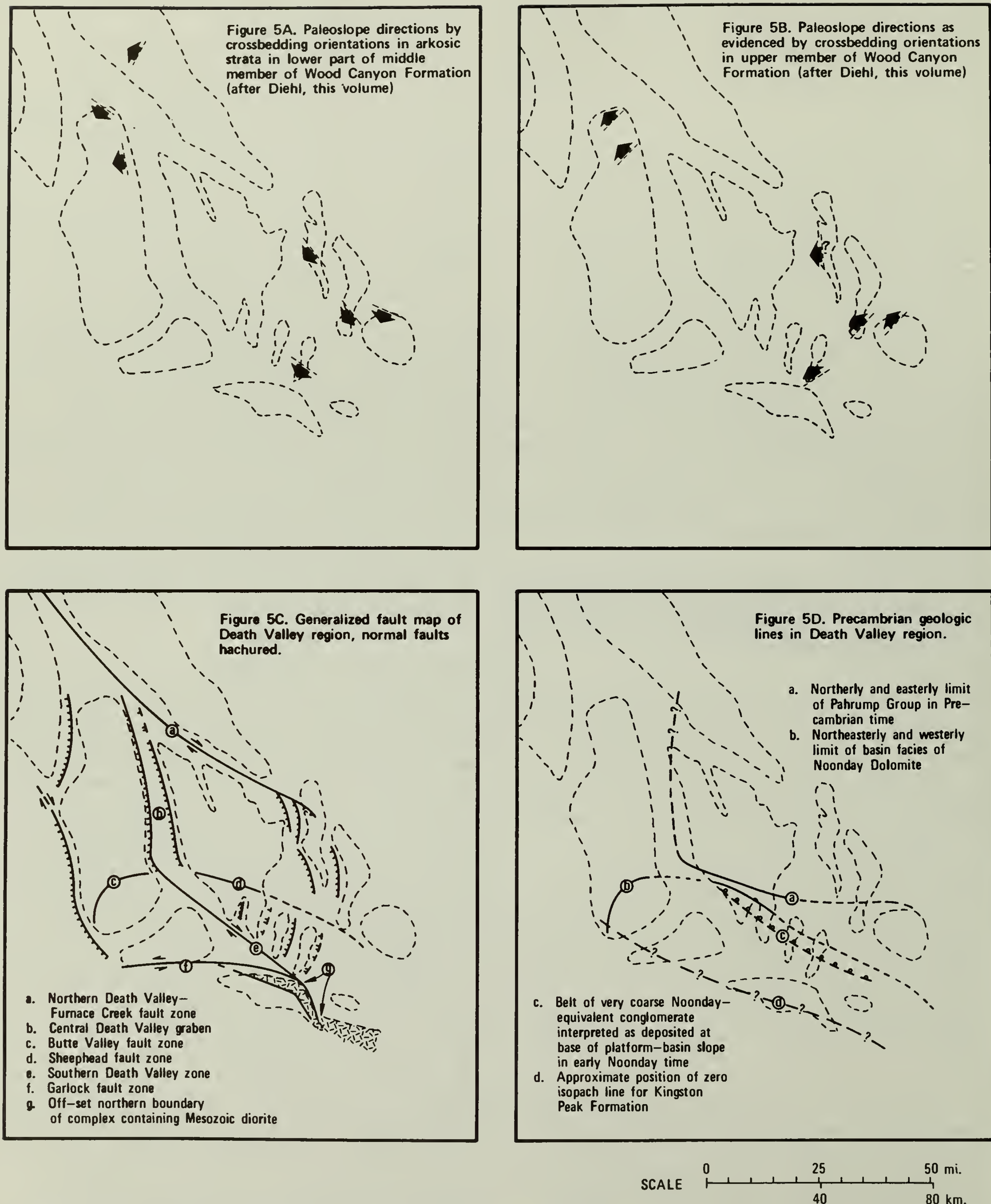


Figure 5. Paleoslope directions of Wood Canyon Formation, fault map, and Precambrian geologic lines, Death Valley region. See Figure 3a for identification of mountains.

REFERENCES CITED

Hazzard, J. C., 1937, Paleozoic section in the Nopah and Resting Springs Mountain, Inyo County, California: California Div. Mines, Jour. Mines and Geology, v. 33, no. 4, p. 273–339.

Hewett, D. F., 1956, Geology and mineral resources of the Ivanpah quadrangle, California and Nevada: U.S. Geol. Survey Prof. Paper 275, 172 p.

Hoffman, P., Dewey, J. F., and Burke, K., 1974, Aulacogens and their genetic relation to geosynclines with a Proterozoic example from the Great Slave Lake, Canada, *in* Dott, R. H., Jr., and Shaver, R. H., eds., Modern and ancient geosynclinal sedimentation: Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 19, p. 38–55.

Kupfer, D. H., 1960, Thrust faulting and chaos structure, Silurian Hills, San Bernardino County, California: Geol. Soc. America Bull., v. 71, no. 1, p. 181–214.

Noble, L. F., 1934, Rock formations of Death Valley, California: Science, v. 80, no. 2069, p. 173–178.

——— 1941, Structural features of the Virgin Spring area, Death Valley, California: Geol. Soc. America Bull., v. 52, p. 941–1000.

Stewart, J. H., 1972, Initial deposits in the Cordilleran geosyncline: Evidence of a late Precambrian (<850 m.y.) continental separation: Geol. Soc. America Bull., v. 83, p. 1345–1360.

Troxel, Bennie W., 1967, Sedimentary rocks of late Precambrian and Cambrian age in the southern Salt Spring Hills, southeastern Death Valley, California: California Div. Mines and Geology Spec. Rept. 92, p. 33–41.

Wright, L. A., 1968, Talc deposits of the southern Death Valley–Kingston Range region, California: California Div. Mines and Geology Spec. Rept. 95, 79 p.

——— 1973, Geology of the SE¼ of Tecopa 15-minute quadrangle, San Bernardino and Inyo Counties, California: California Div. Mines and Geology, Map Sheet 20.

Wright, Lauren A., and Troxel, Bennie W., 1966, Strata of late Precambrian–Cambrian age, Death Valley region, California-Nevada: Am. Assoc. Petroleum Geologists Bull., v. 50, no. 5, p. 846–857.

——— 1967, Limitations on right lateral strike-slip displacement, Death Valley and Furnace Creek fault zones, California: Geol. Soc. America Bull., v. 78, no. 8, p. 933–949.

This research was supported by NSF Grant GA-16119.

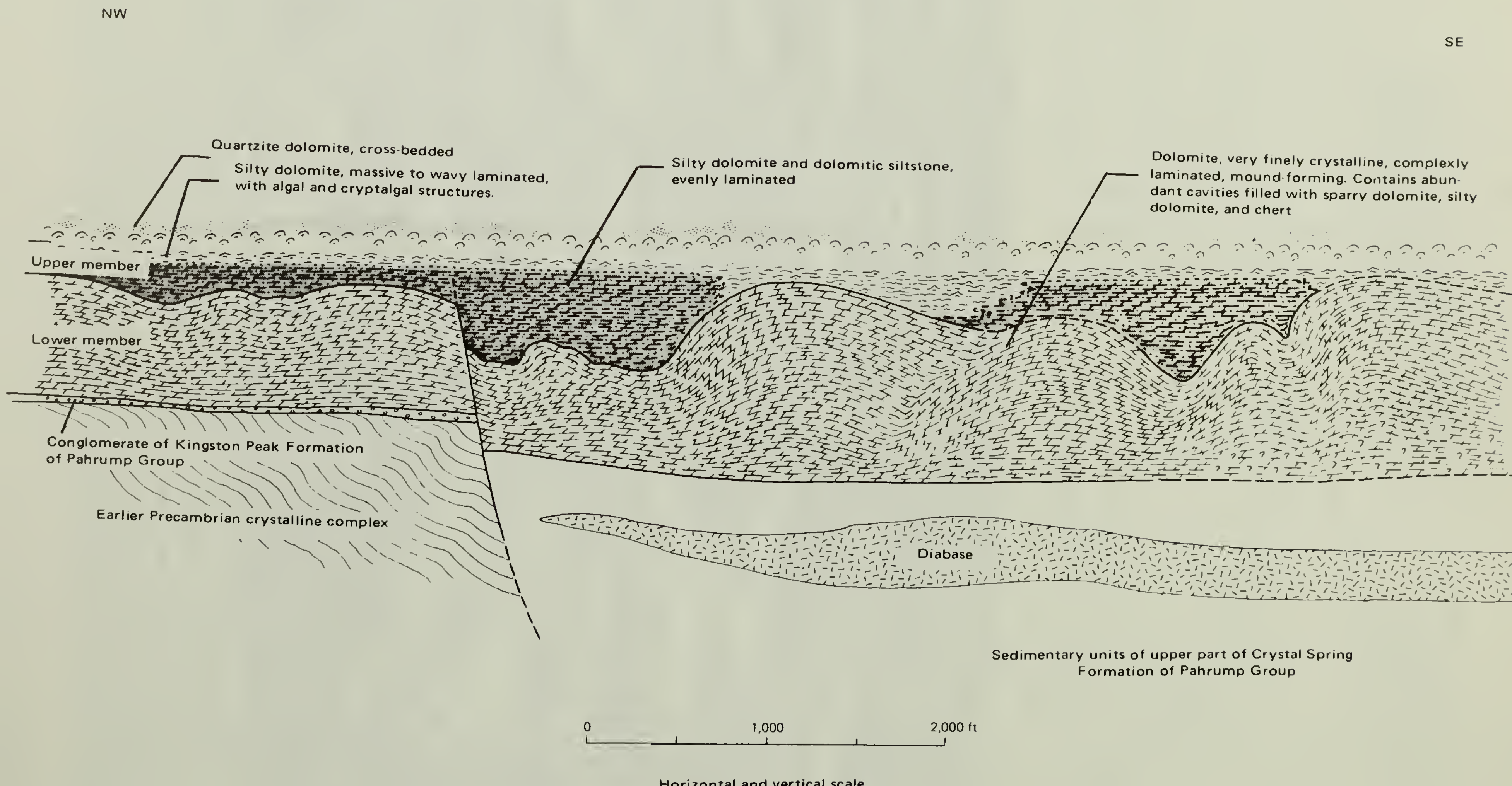
High-altitude oblique aerial photograph to south-southwest across the southern end of Death Valley. Avawatz Mountains barely above central part of photograph. South end of Nopah Range lies in lower center; Alexander Hills (with thin white streak) lie slightly farther south. Amargosa River flows south from lower right of photograph through patch of deformed pale Tertiary lake beds (just right of center foreground), then through deformed Tertiary-Quaternary gravel into the floor of southern Death Valley. The river then flows westward at the north edge of the fan extending from the Avawatz Mountains and flows northward near right edge of photograph. The east-trending Garlock fault and northwest-trending Southern Death Valley fault intersect along the north side of the Avawatz Mountains.

The Mojave River flows from the San Bernardino Mountains (dark ridge in upper center) across the Mojave Desert to Baker (on left side of dark hills beyond and left of Avawatz Mountains). From Baker an ancient river channel continued northward to the dark hills (Salt Spring Hills) in center of photograph. *Photo U.S. Air Force 374R 199, 6 September 1968; courtesy of the U.S. Geological Survey.*

Stratigraphic Cross Section of Proterozoic Noonday Dolomite, War Eagle Mine Area, Southern Nopah Range, Eastern California

L. A. Wright,[1] E. G. Williams,[1] and Preston Cloud[2]

[1] *Department of Geosciences, Pennsylvania State University, University Park, Pennsylvania 16802.*
[2] *Department of Geological Sciences, University of California, Santa Barbara, Santa Barbara, California 93106.*



Stratigraphy and Sedimentology of the Wood Canyon Formation, Death Valley Area, California

Paul Diehl[1]

INTRODUCTION

Nolan (1924) named the Wood Canyon Formation and described it from exposures in the northwest part of the Spring Mountains, Nevada. Hazzard (1937) extended the use of the name to the southern Nopah Range (Fig. 1), where he measured a section of Precambrian to Cambrian strata including those of the Wood Canyon Formation. Later, Stewart (1966, p. C71) divided the formation into lower, middle, and upper informal members.

Still more recently, Stewart (1970), in a systematic regional stratigraphic study of the Wood Canyon and associated formations of the southern Great Basin, California and Nevada, presented measurements of cross-strata dip directions that he interpreted as indicative of a west- to northwest-dipping paleoslope. The depositional environments of the Wood Canyon, however, remain relatively unstudied.

PURPOSE

Although the general physical stratigraphy of the Wood Canyon Formation has been well established, the sedimentary and tectonic environments in which it was deposited are still little understood, and are the subject of the present investigation. Included in this study are observations of small-scale primary sedimentary features within the formation.

The Wood Canyon Formation is of particular interest because it has diverse lithologies, occupies a stratigraphic position athwart or near the Proterozoic-Paleozoic boundary, and its areal distribution is well documented. The section that contains the Wood Canyon and the underlying Stirling Quartzite, Johnnie Formation, and Noonday Dolomite in the southern part of the Nopah Range has been suspected as a possible stratotype for the Proterozoic-Phanerozoic boundary (Cloud, 1973). In addition, environmental information from this formation, combined with environmental data being gathered by other workers in the area, is useful in attempts to reconstruct the late Precambrian–Cambrian evolution of the Death Valley region.

WOOD CANYON FORMATION

General Features

The Wood Canyon Formation has been divided previously into three informal members that are recognizable over a large area of eastern California and western Nevada. The lower member is composed of interbedded siltstone (36 percent), thinly laminated to platy bedded, fine- to medium-grained feldspathic and micaceous sandstone (50 percent), and laminated, siliceous dolomite (14 percent). The latter occurs primarily as three sub-units in the member. The middle member contains a lower sub-unit composed of arkosic conglomerate and grades up-section into cyclically bedded subarkosic sandstone and maroon siltstone. The interbedded light olive-gray to tan siltstone (23 percent) and dark brown to pale red weathering, fine- to very fine grained quartzitic sandstone (49 percent) of the upper member are similar to those in the lower member with respect to distribution and primary sedimentary features. Sandstone beds are 1 to 5 ft thick, well laminated to massive with occasional cross-laminations. The dolomite (12 percent) in the upper member is contained mostly in one sub-unit, which is characteristically cyclically bedded. Distinctive millimeter-sized platelets of echinodermal debris are found in the dolomite beds.

In addition to the gross subdivision of the Wood Canyon into three members, a further subdivision into at least 18 sub-units is recognizable. (Detailed descriptions of sub-units recognizable in the southern Nopah Range measured section are shown in Fig. 2).

Lower Member

The lowermost 66 ft of the measured section in the Nopah Range (Fig. 2) are composed of predominantly medium- to fine-grained feldspathic sandstone (76 percent) and siltstone (20 percent) that weathers drab red-brown. In sub-unit SN-3 (Fig. 2), the sandstone content decreases up-section, and sandstone beds form discontinuous wedges and lenses. This decrease in sandstone is accompanied by an increase in siltstone. One of the three dolomite sub-units (SN-2) occurs at 66 ft. The other two are near the top of the member. Sub-unit SN-1 is featured by cyclic bedding, although the scale of Figure 2 is too small to record it. A typical cycle consists of a lower layer of sandstone that is internally massively bedded at its base and contains little mica. It is 1 to 2 ft thick and grades upward into very evenly and thinly laminated, micaceous, very fine grained sandstone. The latter forms the top one-fourth to one-half of each cycle and gives a "shaly look" to the rock that separates the massive beds.

Mudcracks, trace fossils, interference ripples, loading features, scoured surfaces, and discontinuous, convex sandstone bodies that are cross-laminated suggest shallow water deposition. Bimodal and polymodal current rose diagrams are compatible with shallow water marine, probably tidal environments for the lower member.

Middle Member

The basal, ubiquitously cross-laminated arkosic conglomerate (Fig. 2) composes 99 percent of sub-unit SN-10. The overlying subarkosic and feldspathic sandstone and siltstone are cyclically bedded in fining-upward sequences. These cycles range from 1 to 20 ft thick. Most of them are from 3 to 6 ft thick. The base of a typical cycle is marked by a layer of coarse- to very coarse grained quartzitic sandstone or pebbly conglomerate that contains clasts of reddish siltstone. The siltstone clasts apparently were derived from the underlying siltstone bed with which the quartzitic sandstone or conglomerate is in sharp erosional contact (Fig. 3b). The basal layer grades upward into coarse-grained sandstone that is either planar or trough cross-laminated and featured by sets or cosets 2 to 5 in. thick. Planar cross-laminations with straight foresets are as much as 9 in. thick. The next higher layer consists of medium-grained feldspathic sandstone exhibiting complexely festooned cross-laminations and cosets of 2 to 4 in. This sandstone, in turn, grades upward into a massive to well-laminated, fine-grained feldspathic sandstone. Topping the cycle is an alternation of massive and fissile maroon siltstone. The beds within the cycle are 1 to 4 ft thick. The sandstone beds are generally thicker than the siltstone beds, which are as much as 6 in. thick.

[1] *Pennsylvania State University, University Park, Pennsylvania 16802.*

Beds of the middle member are from 1 to 4 ft thick, but change thickness as well as color and lithology laterally. Beds of cross-laminated sandstone commonly grade laterally into siltstone, but also interfinger with siltstone beds. Some of the beds that display the lateral wedging and gradation persist for only 3 to 4 ft, whereas others persist for 100 ft or more.

Upper Member

The upper member is similar to the lower member in thickness and in the proportion of sandstone, siltstone, and dolomite that it contains. The sandstone becomes more quartzitic upward and also shows an upward transition from even-lamination to cross-lamination. The siltstone content increase

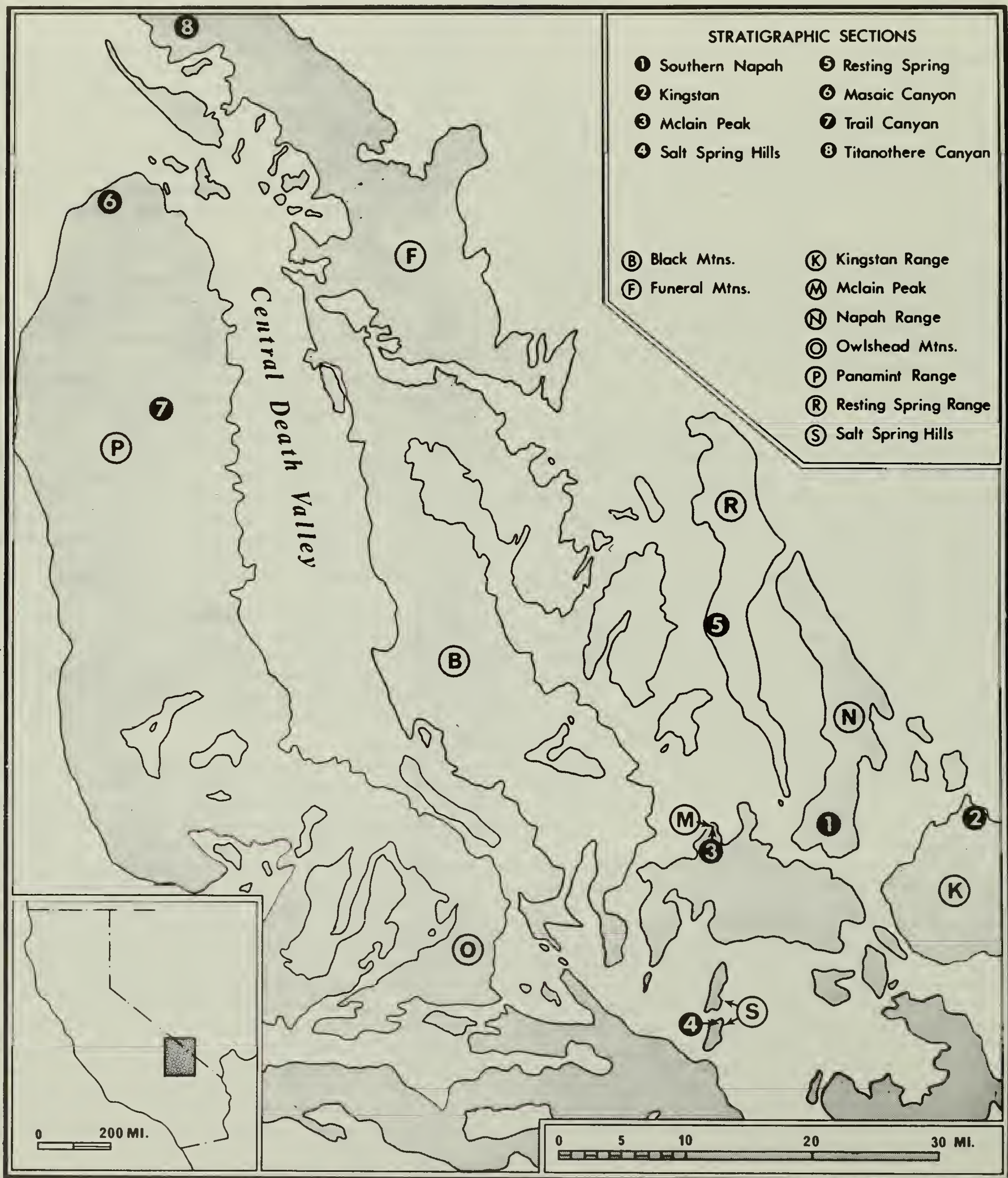
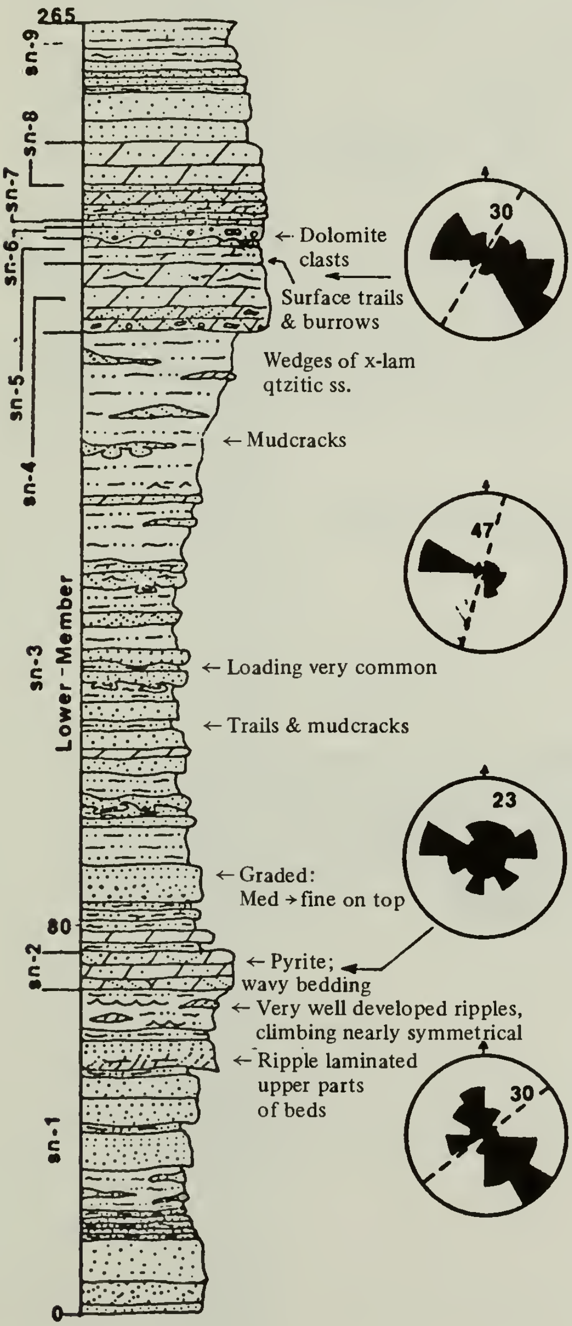


Figure 1. Location of measured sections, Wood Canyon Formation.

up-section, reaching a maximum in sub-unit 18 underlying the dolomite sub-unit. Details of the sub-units, such as first occurrence of fossil hard parts and sand casts of trilobite fragments, are illustrated in Figure 2.

The dolomite sub-unit within the upper member is cyclic in nature. An ideal cycle is illustrated in Figure 3a. This sequence is variable, and one or more lithologies are commonly absent. The cycle, where fully developed, consists of a thin siltstone bed followed vertically by an evenly laminated, well-sorted sandstone that characteristically contains calcareous cement, and locally is faintly cross-laminated. This sandstone is in turn overlain by persistently cross-laminated, dolomitic sandstone in which fossils are common and that grades upward into a cross-laminated to wavy-laminated dolomite displaying abundant quartz laminations and lenses. The sequence is capped by a massive, "elephant hide"-weathering, fine-grained dolomite. Cross-laminations in the dolomite are commonly trough shaped, but planar cross-laminations are also abundant. The sequence is commonly 8 ft thick, and the dolomite beds are as thick as 5 ft. The sandstone and siltstone beds, however, are usually less than 1 ft thick. The siltstone disappears from the sequence upward through the dolomite unit (Fig. 2). The dolomite also contains distinctive platelets of echinodermal debris, which serve to distinguish the upper member dolomite unit from the



sn-9
Intbbb fg, lam ss & qtzt w/ lt. br. to orange tan sltst & slty sh. Few lensatic, 2–6″ bdd mod br (5YR3/4) wthrd, lam dol ss; some xlam & ripples. 3′ massive vf to fg, pale mod reddish br wthrs pale yelsh br (10YR6/2) in middle. Ss, qtzt & sltst pale red (10R6/2) wthr patchy mod yelsh br (10YR5/4), mica on surfaces. Ss lens and wedge along strike. *Scolithus*?

sn-8
Intbdd dol ss & sandy dol; unit appears striped orange br. becoming mod yel br (10YR5/4) at top. Beds 4 in. at base becoming thicker and more dol up, capped w/ 4 ft vfg med dk gry (n4) dol in 1 ft beds. Ss mg-cg, dk yel br(10YR4/2), xlam in more sandy middle of the unit.

sn-7
Ss, vfg to fg, mod br(5YR4/4) wthr mod br (5YR3/4), 1/16 in. horiz lam, dol peb casts & chips to 2¼ in.; surface trails. Pods of granular ang qtz grains.

sn-6
Granular & vcg, poorly sorted, flat ped cgl ss; chips of dol to ¾ in., pods of gran qtz pebs in 3 in. layers, scoured base. Intbdd fg ss, thin lam qtzt ss w/ ripple lam; some 2 in. beds maroon sltst w/ sand-filled mudcracks or burrows?

sn-5
Intbdd platy, vfg, calc ss to sltst; lt. pinkish to grnsh gry wthrs lt. gry grn to yelsh br. 1–1½ ft beds fg sandy dol, mod br(5YR3/4) wthrs med by (5YR4/4), gritty surface w/ mica, grades lat. to dol ss.

sn-4
Dol, vfg, sandy, med dk gry wthrs dk yelsh br(10YR4/2). Qtz grains raised on dol surface in lower 5 ft. Wavy & xlam wthrs v. dusky br(10R2/2) at base. Pyrite.

sn-3
Intbdd ss, sltst, some slty sh. Ss mica rich, usually platy but beds to 1 ft, mg to fg, tan wthrs dk tan br. Sltst buff tan near base to olive drab w/ red streaks at top. Deformed bdd, loading, xlam, mudcrack, ripples common. Top 35–40 ft slty sh & fissile sltst, ss here 2–6 in. thick lenses of 10–15 ft lat. extent, xlam. Burrows & trails on bdg. Scattered discontinuous dol beds.

sn-2
Dol, mg, med dk gry wthrs lt. med br, sandy lam wthr in relief dusky red br, irregular w/ some xlam. Pyrite rosettes.

sn-1
Intbdd platy, mg to fg, feldspathic ss. Shaly bdd calc ss, v. micaceous. Silica cemented ss beds to 2 ft, med to dk gry wthr drab red br. Lower 10 ft xlam small scale planar & trough foresets. Symmetric ripples well developed in slty top 7 ½ ft. Beds massive base to well lam top. Parting lineations, trails & burrows on bdg surf.

(continued on page following)

Figure 2. Measured section of Wood Canyon Formation, Nopah Range. Section measured on southern slope of hill VABM 4238, approximately 1½ miles north of the Noonday mine, sec. 11, T20N, R8E, Tecopa, Calif. 15-minute quadrangle. Dashed lines on current rose diagrams indicate possible strike of paleoslope. Scale 1 in. = 40 feet.

other dolomite beds both within the Wood Canyon Formation and in the overlying and underlying formations.

TRENDS IN THE WOOD CANYON FORMATION

Cross-lamination measurements were recorded with close reference to their stratigraphic positions so that vertical variations in their orientation, discussed below, could be detected within each stratigraphic section. The lateral variation of preferred orientation of paleocurrent indicators could be investigated by comparing their trends from section to section within the Death Valley area.

Vertical Trends

Analysis of the orientation data gathered so far reveals a consistent vertical variation of paleocurrent orientation within each stratigraphic section. This variation can be correlated with the succession of sub-units defined above (Fig. 2). The bimodal and polymodal frequency diagrams of cross-strata measurements taken in the lower member are consistent in each section measured (Fig. 4a) and suggest a northwest-southeast paleocurrent orientation.

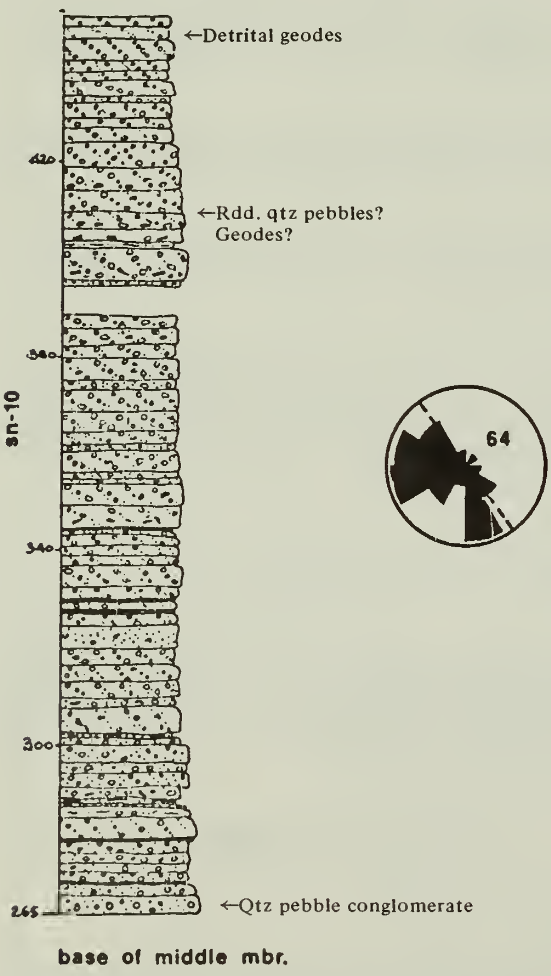
In marked contrast to the lower member, the paleocurrent indicated by current rose diagrams taken from the arkosic conglomerate at the base of the middle member, is to the southwest (Fig. 2, SN-10; Fig. 5a). This nearly 90° change in paleocurrent occurs abruptly at the contact between the lower and middle members. Up-section in the middle member, cross-lamination orientations show a re-establishment of a northwest-southeast paleocurrent. This change is gradual; the frequency modes of paleocurrent direction measurements from sub-units in the ramainder of the middle member show an up-section clockwise rotation.

Polymodal and bimodal cross-lamination frequency diagrams plotted from data taken in the upper member again reflect a northwest-southeast paleocurrent as defined in the lower member.

Regional Trends

Sub-units recognized within the southern Nopah Range can be recognized in sections throughout the Death Valley area, particularly if sub-units five, six, and seven of the lower member are grouped into one, thus leaving seven regionally recognizable sub-units of that member. In general, these sub-units occur with little lithologic variation throughout the Death Valley area, so that distinct lateral facies are not generally recognizable. The current-direction indicators that are measured from correlative sub-units in various stratigraphic sections (Fig. 1) yield consistently oriented paleoslope indicators except where tectonically disturbed (Figs. 4a, 5a). The same vertical pattern of paleoslope variation described above can generally be detected in sections thus far measured.

Although the nearly 90° rotation of current direction indicators up-section is clearly demonstrated in the Mclain Peaks section, the rose diagrams there are anomalous, as they suggest a northward paleoslope for the lower and upper members and a northwest paleoslope for the arkosic conglomerate at the base of the middle member. The differences between these orientations and the others already described seem best explained

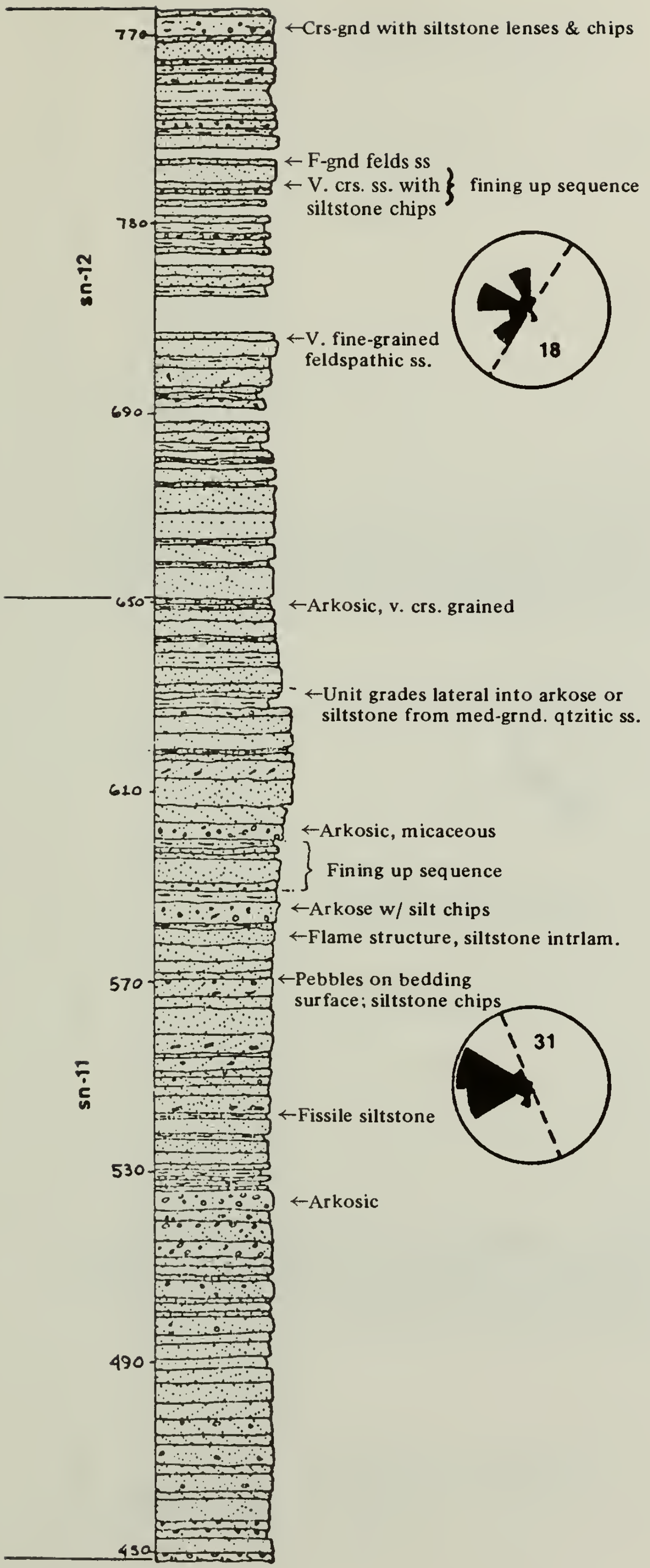


sn-10

First 4′ granular to vcg qtz peb cgl, poorly sorted vitreous white & transparent ang qtz fragments; contains specks of dk yelsh orange(10YR6/6) and lt. grn clay material, bed wthrs pale grn(10G6/2). Basal ctc sharp and uneven but top transitional w/ overlying dk grnsh gry(5G4/1) arkosic cgl, vcg & granular, planar xlam v. abundant w/ some trough foresets. Rare grysh-red purple (5RP4/2) sltst in beds 6″ or less. Felds abundance patchy. Beds 2-5′ w/ xlam sets up to 1′.

(continued on page following)

Figure 2 (continued).



sn-12

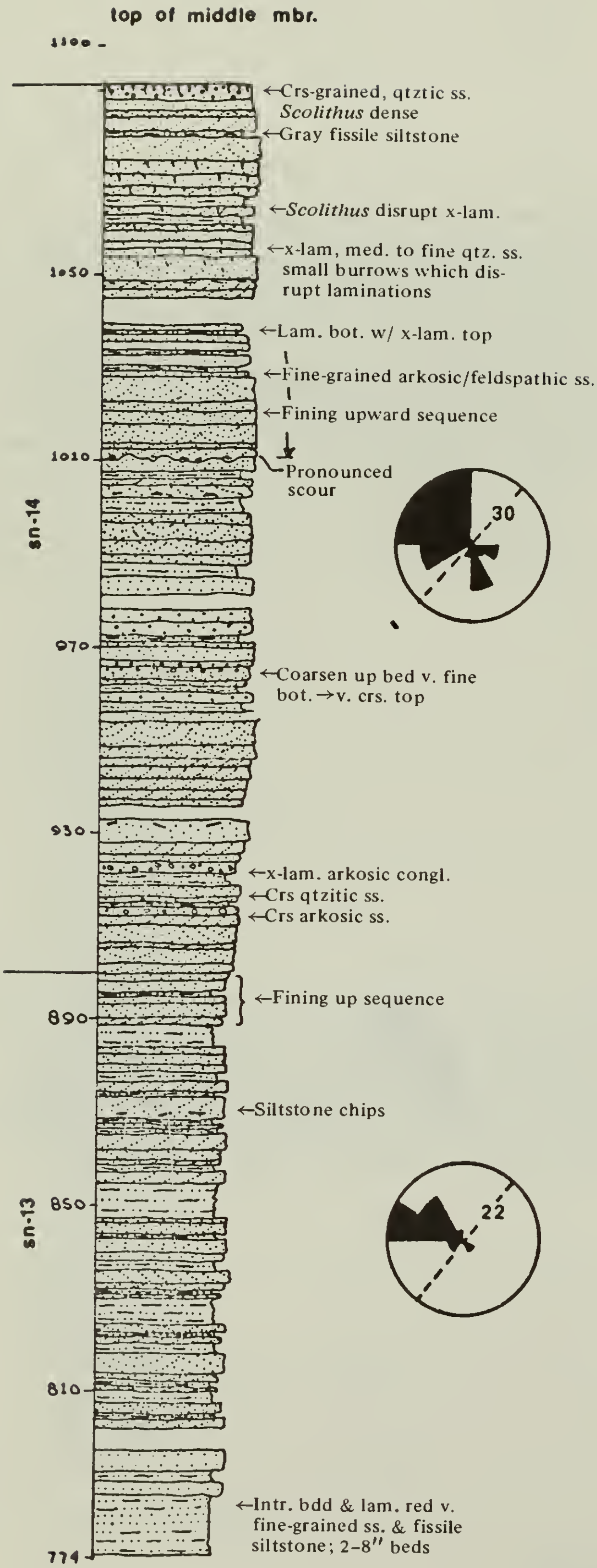
Cyclic intbdd ss & sltst. Ss are (1) fg, felds, grysh red purple (5RP4/2) wthr grysh red(5R4/2), beds 15″–4′, more abundant & thicker upward, horiz lam to massive w/ few planar xlam; (2) m–vcg ss, pale red purple(5RP6/2) wthrs streaky pale pink & grysh red, complex trough xlam, ang & rdd sltst flat peb chips to 2″ abundant, not well cemented, beds 1′ & less; (3) well cemented, m–cg, grysh purple(5P4/2) w/ some white to v. lt gry(N9–N8) wthrs mod yelsh br(10YR5/4), ang to subrdd qtztic ss, well defined complex trough xlam w/ long tangent foresets, also planar straight foresets truncated top & bottom, cosets 2–3″ but up to 1′ when planar, beds 2–4′ pinch & swell w/ lithologies grading laterally. Fissile sltst, much mica on bdg planes, beds to 2′, dusky red(5R3/4) to grysh red purple(5RP4/2).

sn-11

Intbdd qtz peb cgl, arkose cgl, qtzt & felds ss. Felds ss pale red(10R6/2) to grysh red(10R4/2), mg, very well developed complex trough xlam w/ cosets 2–8″. Planar xlam w/ larger sets up to 1′. Qtz peb cgl w/ pebs of qtz, red chert, & detrital geodes, beds med yelsh br(10YR5/4) & lt gry(N7). Arkose cgl becoming less frequent. Sltst becomes more abundant than in SN–10 w/ some beds to 1′. Lithologie form fining upward sequences of 7–14′ thick. Beds not laterally continuous, pinch & swell, wedge out, & grade laterally

(continued on page following)

Figure 2 (continued).



sn–14

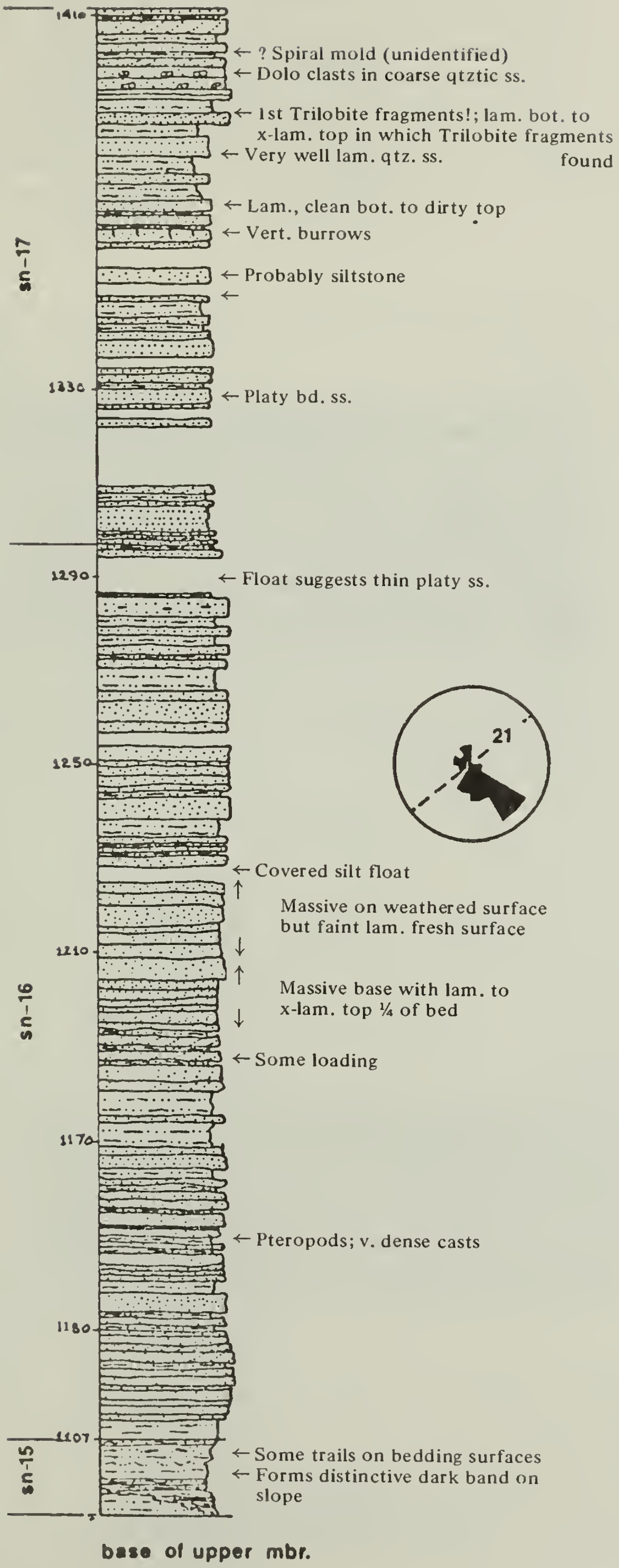
Qtztic & gwke ss, well defined complex xlam, troughs & complex v. tangential foresets, beds 2–3′ w/ cosets 3″ to 1′ but 3–8″ most common, m–vcg w/ cg predominating. Vcg & gran arkosic beds comprise 12′ of section 52′ from base of unit. Top of unit cg to vcg, well cemented qtzt ss w/ v. densely spaced *Scolithus*. Fissile sltst & massive fg felds ss rare.

sn–13

Dominated by fissile maroon sltst w/ beds up to 3′ thick, and grysh red purple felds to gwke ss, mg w/ parts of section composed of as much as 12′ continuous ss in 1–2′ beds. Chips of sltst v. common on bdg surface and w/in beds. Other lithologies of SN–11 & 12 also occur. Fining upward sequences.

(continued on page following)

Figure 2 (continued).



sn-17

Qtztic ss, fg & vfg, well cemented, thin horiz lam, ss dk grnsh gry(5G4/1) wthrs vary colored dk yelsh br(10YR4/2) & dusky yelsh br(10YR2/2). Covered section may be vfg platy bdd ss & sandy grysh grn sltst-seen in float. Beds 9″ to 5′ but most 1-2′. Some massive & platy bdd "sugar-like" qtzt, lt gry(N7) wthrs mod orange pink(5YR8/4) & yelsh gry(5Y7/2). Xlam ss 72′ from base & becomes more abundant up-section. Top of unit has fg qtztic ss w/ lens of vcg qtz & flat rdd dol cobbles 1″ to 3½″ long, xlam. First trilobite casts occur in place 92′ from base of unit. Trilobite frags. associated w/ lt to med gry(N7-N5) qtztic ss, horiz lam base & xlam top. *Scolithus*, bioturb. lam also found. Unidentified spiral mold found in xlam calc cemented qtztic ss at top of unit.

sn-16

Slope forming intbdd vfg-fg ss & sltst (60/40). Sltsts dk yelsh br (10YR4/2) shaly & lt olive gry(5Y5/2) massive & fissile. Sltst beds sandy & grade transitionally w/ ss. Ss are pale yelsh br (10YR6/2) to med dk gry (N4) to olive gry(5GY4/1) w/ some pale red wthrng to dk redsh br(10R3/4), grysh red(10R4/2) & dusky yelsh br(10YR2/2), but distinctive pinkish or redsh br wthrd surface. Beds massive, 9″ to 2 1/2′. Lower 8-12′ of unit gwke ss becoming fg & qtztic up w/ decrease in shaly sltst. Xlam rare in lower 1/4 of unit, some scour w/ sltst chips. Pteropod bed 3″ thick 9′ from base. Trilobite frags. in float in upper part of unit.

sn-15

Gwke ss, m-fg, grysh br wthrs blksh red(5R2/2), poorly cemented, lam w/ thin shaly wavy lam approaching xlam, platy bdd but some beds to 15″. Cg patches, poorly sorted, ang qtz grains & blk sltst/sh chips. Intercalated dk maroon sltst. Upper ctc transitional. Forms distinctive dk wthrng band. 6′ olive gry(5Y4/1) sltst occurs NW along strike.

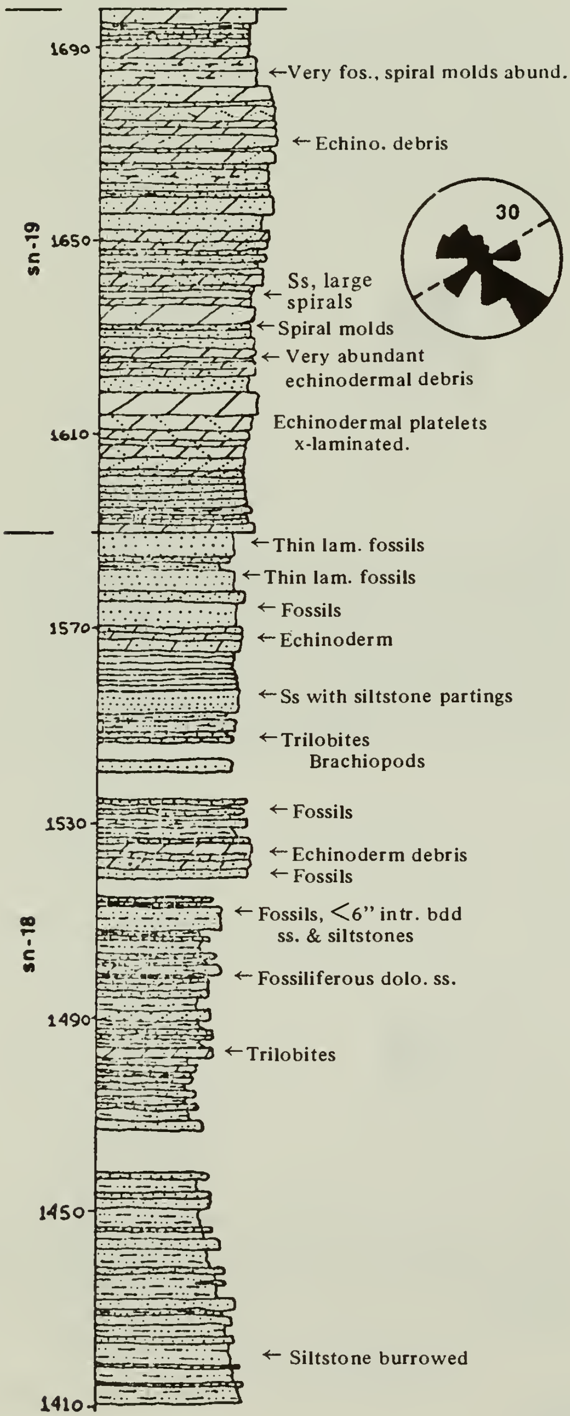
(continued on page following)

Figure 2 (continued).

by postdepositional tectonic rotation that occurred about a vertical axis and was related to the strike-slip movement of the Sheephead fault, which apparently passes immediately north of the peak.

The dolomite sub-units of the lower and upper members show gradual thickening to the northwest. The dolomite sub-unit of the upper member in the south Salt Spring Hills is 20 ft thick, whereas in Mosaic Canyon, at the north end of the Panamint Range, the correlative sub-unit is 340 ft thick. The three dolomite sub-units of the lower member are recognizable in each of the sections measured (except in the Salt Spring Hills, where the lower member is absent). They show a progressive thickening from southeast to northwest. These thickness trends are demonstrated in Figure 4a and b where thickness of the dolomite sub-units is plotted against distance from a line placed approximately parallel to the strike of the paleocurrent indicated by current rose diagrams for the two members. This line is positioned south of the Salt Spring Hills, where the lower member of the Wood Canyon Formation is absent.

Both the author's data and those of Stewart (1970) indicate that the lower member of the Wood Canyon Formation gradually thins to the southeast and disappears in the area of



sn-19

Intbdd dol, sandy dol, & qtztic ss. Dol m- cg, med dk gry(N4) wthrs olive gry (5Y4/1) w/ elephant hide texture, contains calcitic platelets which wthr in relief often defining crude xlams, oolitic. Dol f - mg, med dk gry(N4) wthrs dk yelsh br(10YR4/2), sandy dol w/ ribbons & stringers as well as xlam of brnsh blk(5YR2/1) f -mg silica cemented ss. Dol v. commonly xlam 32-60' from base of unit, xlam small scale w/ cosets 6" & less, foresets v. tangential bottoms w/ truncated tops, some mod red(5R4/6) slty wavy lam 1/8". Intbds of 1-3', lt gry, horiz lam, calc cemented qtzt, often xlam in middle part of unit. Ss becomes more common and more dolomitic upward w/ trilobite sand cast fragments and spiral molds abundant. Cyclic bdd is present but not as well displayed as in the sections to the NW.

sn-18

Gentle slope forming predominantly yelsh tan & lt olive gry fissile sltst in 6'-thick units w/ some v. thin beds of f - vfg horiz lam qtzt similar to that in SN-17. Toward top of unit cg, micaceous, highly fossiliferous ss is more common. Burrows & trails, load casts, flute casts, & ripples(interference) in phyllitic looking ss. Top ½ to 1/3 of unit dol beds appear and ss dominates w/ lt gry fossiliferous qtztic ss intbdd w/ dol & dol ss. Dol contains calcitic platelets - echinodermal debris? Fossils associated w/ m - cg lt gray qtztic ss beds w/ xlam darker less well cemented dol ss tops. Trilobite fragments, brach molds, spiral molds, vertical tubes abundant.

(continued on page following)

Figure 2 (continued).

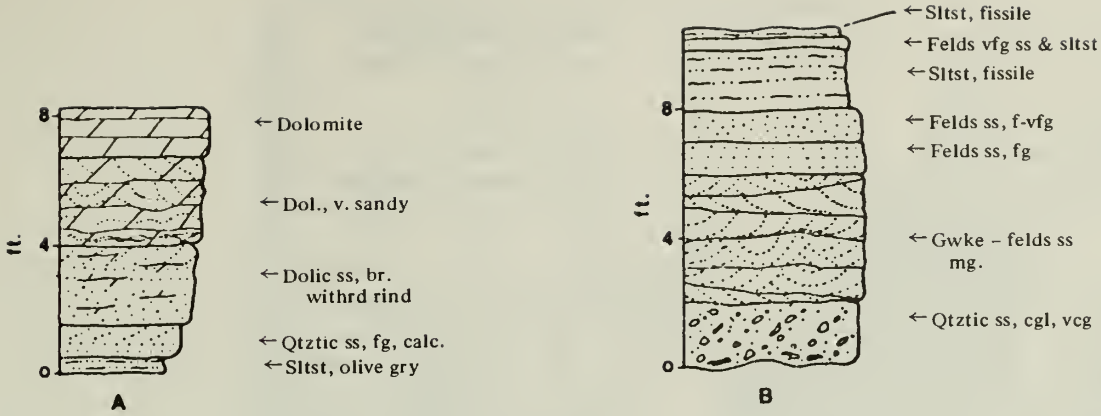


Figure 3. Cyclic bedding in: A. Upper member dolomite unit (sn–19); B. middle member cross-laminated sandstone and siltstones (sn–11 to 14). Scale 1 in. = 60 ft.

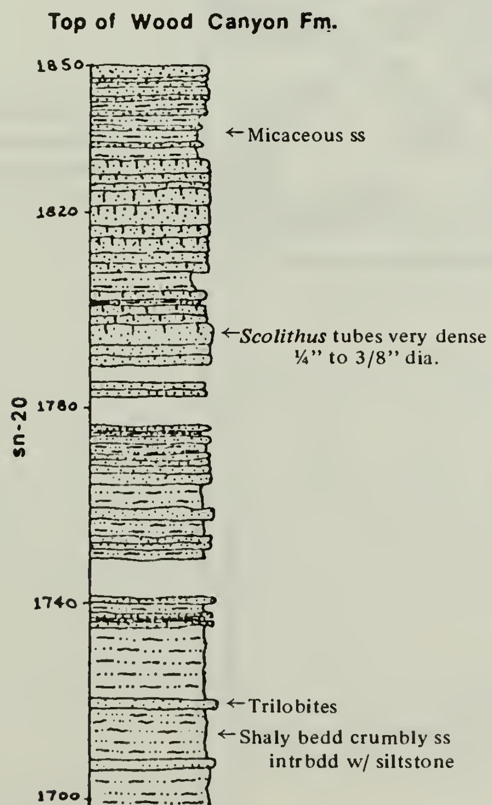
the Silurian Hills (Fig. 4b). In that area and to the southeast, conglomeratic arkose and arkosic sandstone of the middle member rest upon the Stirling Quartzite. Farther southeast, the Stirling disappears and the Wood Canyon correlative, the Tapeats Sandstone, lies directly upon older Precambrian gneiss and schist (Stewart, 1970, p. 39). This suggests a high to the southeast that is consistent with the paleocurrent indicators for the upper and lower members and is also compatible with the northwest thickening of the dolomite sub-units.

In contrast to the northwestward thickening of dolomite in the lower and upper members, no thickness trend is apparent in the middle member when it is considered in total. However, when the thickness of the arkosic conglomerate unit is plotted against distance southwest from an arbitrarily defined line which is parallel to the average strike of the paleocurrents indicated by the arkosic conglomerate rose diagrams, a marked thinning of sub-units to the southwest is obvious (Fig. 5b). Maximum diameters of pebbles measured in the arkosic conglomerate sub-unit at each section also decrease to the southwest (Fig. 5c).

PRELIMINARY INTERPRETATIONS

Bedding features and trace-fossil content, as well as current rose patterns mentioned above (Figs. 2, 3, 4a, 5a) suggest a tidal marine environment for the Wood Canyon Formation.

The vertical changes in paleocurrent directions demonstrable within single stratigraphic sections are consistent



Sn–20

Base of unit tan sltst intercalated w/ "crmbly" wthrng lt yelsh tan mg, v. micaceous ss in 2–5" beds. Also 9" & less beds of massive, f – mg, well cemented, qtztic ss, dk yelsh br(10YR4/2) wthrs to grysh br(5YR3/2); contains trilobite fragments. A less micaceous, mod red br wthrng, *Scolithus* bearing ss appears 1/3 the way up-section from base. Grnsh gry fissile sltst occurs nearer top of unit along w/ the tan sltst. Near top is 40', v. well cemented, dk yelsh orange(10YR6/6) qtzt wthrs mottled lt br(5YR5/6), v. dense *Scolithus*, beds 2–4' thick.
Top of Wood Canyon drawn at base of purplish red & red purple banded well cemented, vitreous qtzt.

Figure 2 (continued).

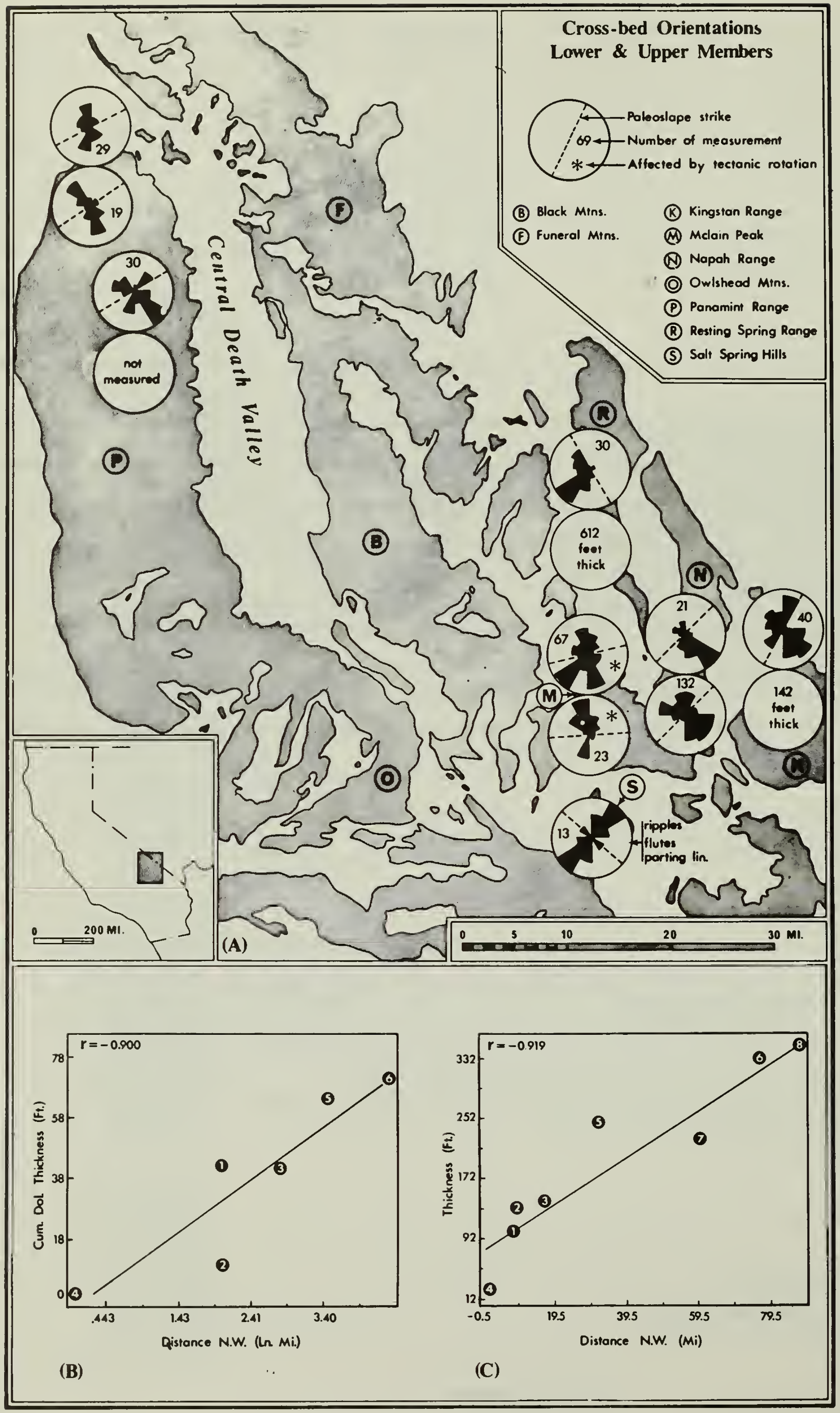


Figure 4. Cross-bed orientations of lower and upper members, Wood Canyon Formation.

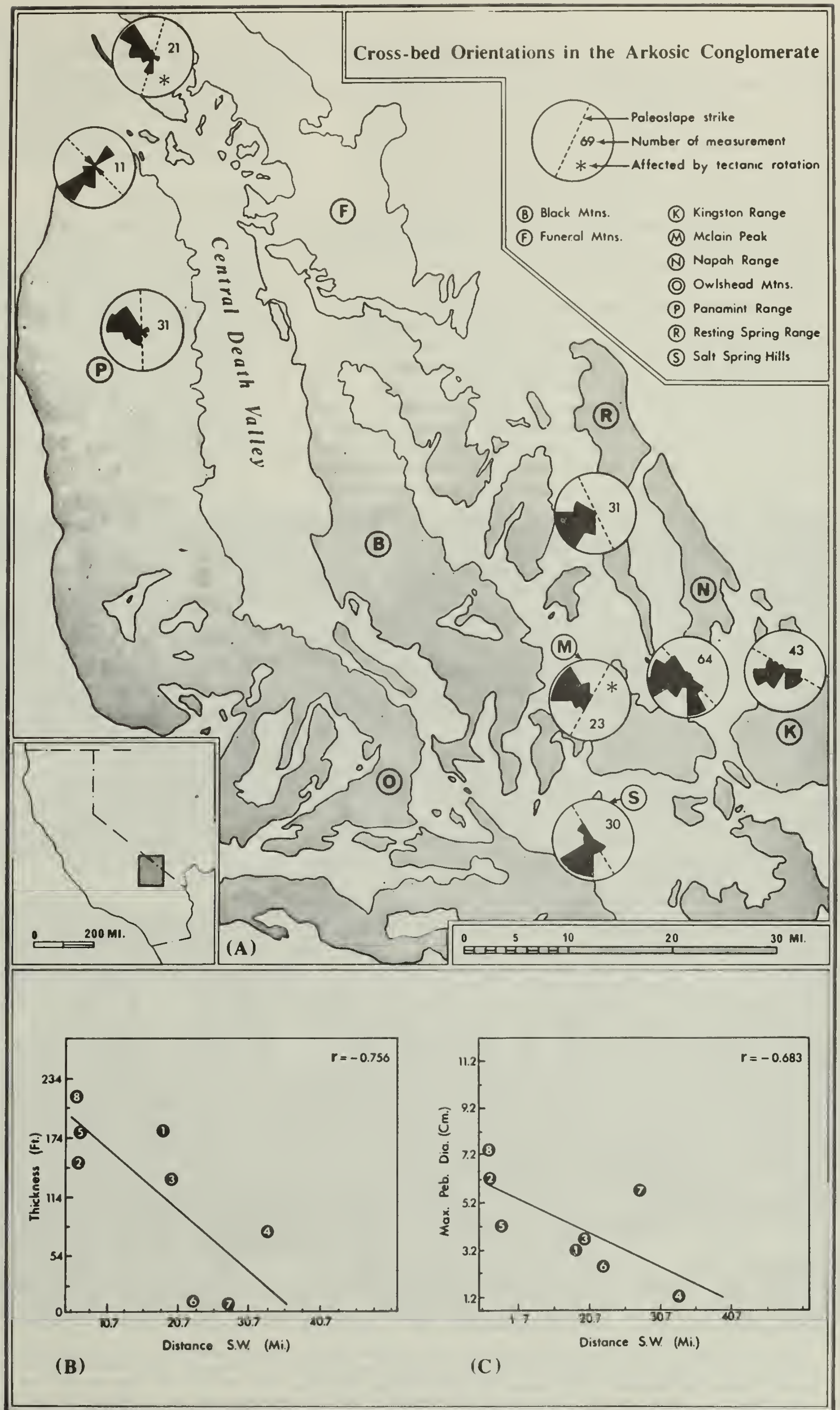


Figure 5. Cross-bed orientations in arkosic conglomerate, Wood Canyon Formation.

regionally and indicate a change in paleoslope during deposition of Wood Canyon strata. That the paleoslope was northwest during deposition of most of the formation is supported both by cross-lamination frequency diagrams and by the northwest increase of dolomite thickness and content in the upper and lower members. The northwest thickening of dolomite units, which is interpreted as a basinward increase in carbonate content, the existence of a southeastern high indicated by the gradual southeasterly disappearance of the lower member, and the paleocurrent data are consistent with a northwest paleoslope during the time of deposition of lower and upper Wood Canyon strata.

On the other hand, the southwestward thinning of the arkosic conglomerate, the decrease of maximum pebble diameter, also to the southwest, and predominantly unimodal paleocurrent frequency diagrams point to a southwest paleoslope during deposition of that coarse sediment sub-unit.

Thus, the data indicate a sharp change in transport direction from northwest to southwest at the base of the middle members with a gradual re-establishment up-section of the northwest trends shown in the lower member. This change in slope is intimately associated with the deposition of the arkosic conglomerate at the base of the middle member and is interpreted as indicating tectonic activity at that time. The arkosic sediments were probably derived from a granitic terrain to the east-northeast. The overlying and underlying beds may then well represent early stages of the Cordilleran miogeosynclinal development with paleoslope here to the northwest.

REFERENCES CITED

Cloud, P. E., 1973, Possible stratotype sequences for the local Paleozoic in North America: Am. Jour. Sci., v. 273, no. 3, p. 193–206.

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

Nolan, T. B., 1924, Geology of the northwest portion of the Spring Mountains, Nevada [Ph.D. thesis]: New Haven, Conn., Yale Univ.

Stewart, J. H., 1966, Correlations of Lower Cambrian and some Precambrian strata in the southern Great Basin, California and Nevada: U.S. Geol. Survey Prof. Paper 550–C, p. C66–C72.

——— 1970, Upper Precambrian and Lower Cambrian strata in the southern Great Basin, California and Nevada: U.S. Geol. Survey Prof. Paper 620.

This research was supported by NSF grant GA–16119.

High-altitude oblique aerial photograph to northeast across southern Funeral Mountains taken from a point barely south of Ryan (lowermost center). Trace of Northern Death Valley-Furnace Creek fault zone extends across lower one-third of photograph. Rocks mostly Tertiary in age lie on near side of fault zone, late Precambrian and Paleozoic rocks lie beyond the fault, in Funeral Mountains. Rocks on left in the range are older and are moderately to strongly folded; rocks on right are younger and deformed mainly by eastward tilting. Tertiary sedimentary rocks lie at extreme right end of range. All rocks are repeated by abundant west-dipping normal faults. A major part of the Basin and Range province occupies rest of photograph. Ranges are dark-colored; basins contain pale-colored playa sediment. *Photo U.S. Air Force 374L 193, 6 September 1968; courtesy of the U.S. Geological Survey.*

Stratigraphy and Depositional Environments of the Crystal Spring Formation, Southern Death Valley Region, California

Michael T. Roberts[1]

INTRODUCTION

This report is a summary of a study of the lower and middle part of the Crystal Spring Formation in the southern Death Valley region begun in 1972. The lower three members of the formation have been studied throughout the region; the dolomite and algal members were investigated at selected localities. The upper two members and the diabase sills in the formation were not studied, and data pertaining to them in this report are primarily from Wright (1968).

The late Precambrian Crystal Spring Formation is exposed within a 75 by 25 mi (121 by 40 km) belt that trends approximately east and extends from the Kingston Range to the southern Panamint Range. The Crystal Spring is the lowest of three formations of the Pahrump Group. It rests with profound unconformity on older Precambrian metamorphic and igneous rocks, and it is successively overlain by the Beck Spring Dolomite and the Kingston Peak Formation. The older Precambrian complex has been dated at 1,700 m.y. (Wasserburg and others, 1959). Diabase sills within the Crystal Spring have been correlated with sills 1,200 m.y. old in Arizona on the basis of their similarity in composition (Wrucke, 1972). The Crystal Spring Formation contains *Baicalia* stromatolites (Howell, 1971); this suggests a middle Riphean age (1,350 to 950 m.y.) for the algal member of the formation. The algal member also contains *Conophyton* stromatolites. Raaben (1969) suggests that the *Baicalia-Conophyton* association is typically middle Riphean. The age may be, then, between 1,350 and 1,200 m.y.

The Crystal Spring Formation is divisible into seven sedimentary members (Fig. 1) and contains diabase sills that range in thickness from a few tens of feet to about 1,500 ft (457 m). The sills are not shown in Figure 1 in order to better show the relations of the sedimentary members. The lower four members, redefined below from those of Wright (1968), are each remarkably persistent throughout the region, varying in thickness but not markedly in lithology. The thickest diabase sill generally occurs within the basal part of the dolomite member, but sills also occur higher in the section in several areas, and feeder dikes are exposed locally. Contact metamorphism has altered the rocks for a distance, generally, of 100 to 150 ft (30 to 46 m) on both sides of the sill. Below the sill, in most exposures, are hornfelsic calc-silicate rocks (the "upper quartzite member" of Wright, 1968, here included with the dolomite member) which were originally impure carbonate rocks rich in quartzose sand and shaly strata. In most areas above the sill, dolomite beds have been altered to talc-tremolite rocks, which are extensively mined in the area (Wright, 1968).

The upper part of the dolomite member and the upper three members (Fig. 1) are persistent in northern and central exposures but apparently show marked facies changes to the south. The southern facies are restricted in outcrop to the Silurian Hills and northern Avawatz Mountains. The exact relation of the northern to southern facies is unknown, because they are separated by a 5- to 15-mi (8- to 24-km) interval in which the Crystal Spring is not exposed, and, also, because of a lack of detailed data on the upper members.

[1] *Pennsylvania State University, University Park, Pennsylvania 16802.*

ARKOSE MEMBER

The arkose member of the formation is here defined as the arkosic sandstone and conglomerate between the basal unconformity and, generally, the top of the uppermost layer of conglomerate. The upper boundary is generally marked by the upward change in color from mainly gray and green to the red or purple of sandstone of the overlying member. In some areas—for example, the Owlshead Mountains—the uppermost conglomerate is absent and the top is defined on the basis of bedforms and color change. In other areas, such as the southern Panamint Range, the conglomerate is present, but the color change occurs within rather than above it. Preliminary petrologic study indicates that the boundary also separates more feldspathic strata from overlying less feldspathic strata, although the change may be gradational.

The member is composed of about 60 percent sandstone and 40 percent conglomerate. Thin beds of shale, siltstone, chert, and carbonate are very subordinate and occur within the finer-grained parts of the member. Thin sections of the member (36 thin sections, 500 points counted per thin section) indicate an average composition of 35 percent quartz grains (mainly undulatory extinction), 22 percent matrix (mainly illitic material), 20 percent rock fragments (15 percent polycrystalline quartz, 5 percent granitic), 17 percent feldspar (albite, microcline, orthoclase), and 6 percent accessory minerals (calcite, opaque minerals, mica, heavy minerals). Much of the matrix appears to be authogenic and derived from the feldspars. "Ghost" grains of feldspar are abundant in the matrix, and the feldspar displays all stages of alteration to illitic material. Consequently, the original feldspar content of the member may have exceeded 30 percent. The rock fragments, heavy minerals, and abundance of alkali feldspar indicate a source containing granitic igneous and metamorphic rocks. The basal and uppermost beds of the member tend to be more quartzose than the rest of the member, trending toward feldspathic sandstone in composition. Variations in quartz/feldspar ratios in the member seem to be related to differences in degree of alteration of feldspars in the sediment and to changes in weathering or composition in the source area, rather than to differences in depositional environment.

Thickness trends of the arkose member and the initial geometry of the depositional basin of the formation are shown in Figure 2. The arkose member contains a basal conglomerate unit as much as 27 ft (8 m) thick, which locally pinches and swells, but which appears to increase in thickness from about 4 ft (1.2 m) to more than 20 ft (6 m) east-southeastward. It consists of pebbles and cobbles composed mainly of metaquartzite and white quartz in a matrix of very coarse feldspathic sandstone. Exposures of this conglomerate in the central part of the region display a clast imbrication. Measurements of clast orientation indicate southerly flowing currents. The basal conglomerate is overlain in many areas by conglomeratic strata (described below) with cross strata that show southerly current directions; it is probably fluvial in origin. In the eastern Kingston Range, the basal conglomerate appears to be part of a fining-upward fluvial sequence. The basal conglomerate contains scour-and-fill structures and potholes at its base. However, in some areas—for example, the southern Ibex Hills—the conglom-

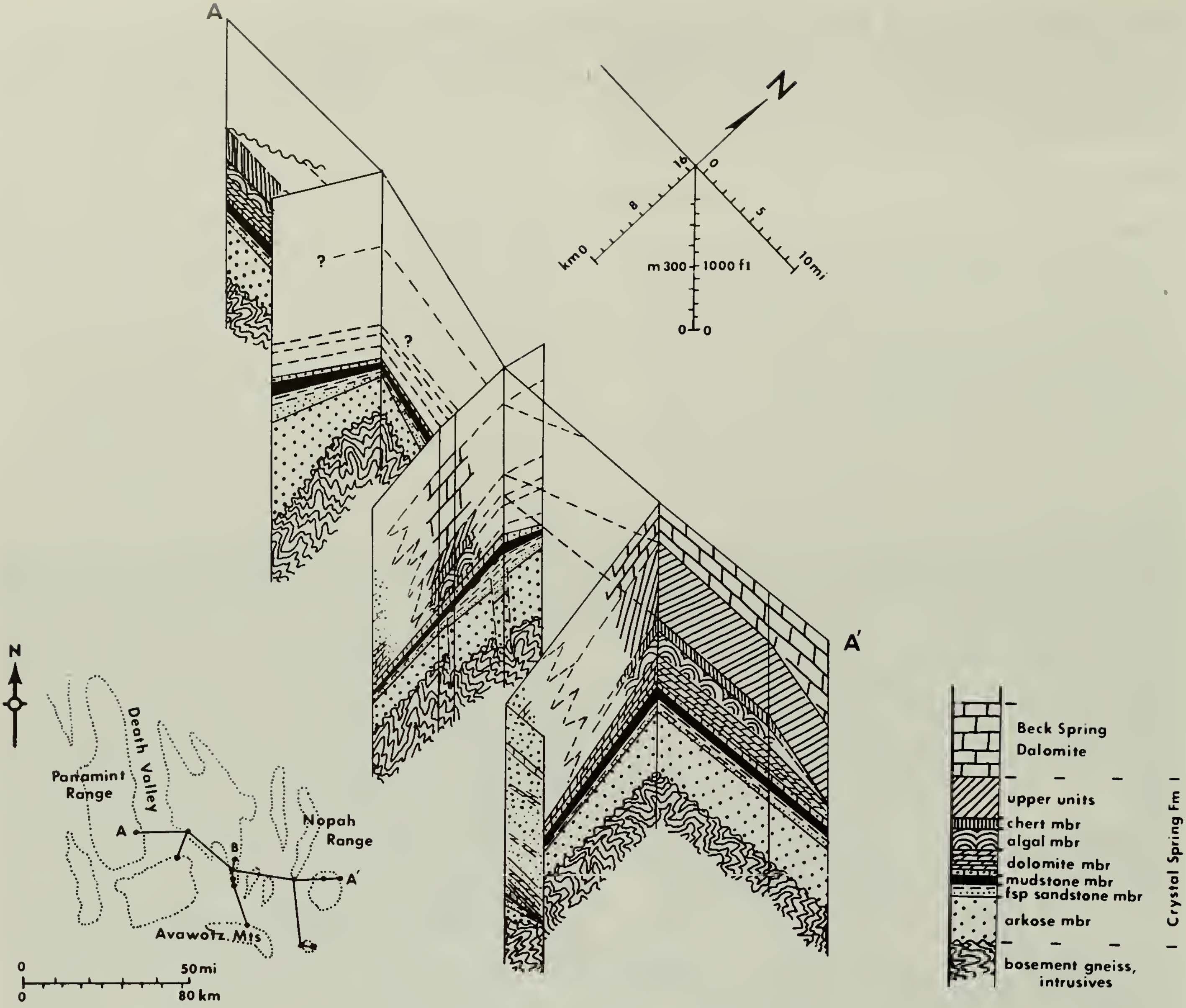


Figure 1. Fence diagram of Crystal Spring Formation. Vertical exaggeration x 26. Location of measured sections (vertical lines of diagram) shown as dots on inset map. Generalized stratigraphic column inset represents northern lithologies. Lithologic changes to south are shown diagrammatically on three southernmost sections. Data for chert, upper units, and Beck Spring Dolomite from Wright (1968). Data for Silurian Hills section in part from Kupfer (1960).

erate is overlain by cross-stratified tidal sandstone beds, which show northerly current directions (described below). The basal conglomerate thus may be the product of complex nearshore processes during the initial transgression over the area. It may, in part, represent a transgressive, cobble beach deposit. The conglomerate appears to have been derived primarily from the erosion of underlying rocks. In the southern Panamint Range, for example, it contains clasts of schist and vein quartz identical with the underlying rocks. The metaquartzite of the clasts is unlike any known older rocks exposed in the region. They may represent redistributed clasts from gravel that formerly overlay the basement or were transported directly from a distant source.

The rest of the member is a cyclical sequence of conglomerate and sandstone (Fig. 3) beds that contain very abundant cross strata. Generally, a cycle consists of a lower conglomeratic zone 30 to 100 ft (9 to 30 m) thick overlain by a sandstone zone 20 to 100 ft (6 to 30 m) thick. The conglomeratic zones are composed of beds of poorly sorted pebble to cobble conglomerate and very coarse sandstone. These zones are characterized by lenticular beds 1 to 6 ft (0.3 to 1.8 m) thick, commonly separated from one another by lenses of micaceous sandstone less than 6 in. (15 cm) thick. Internally, the large lenses are horizontally laminated, cross stratified, or both. The cross strata are chiefly trough types and commonly are about 6 in. (15 cm) to 1 ft (30 cm) thick, although sets of cross strata as much as 6 ft (1.8 m) thick have been observed. Preliminary determination of maximum pebble sizes in the conglomerate, excluding the basal conglomerate, indicates a decrease in size southward (Fig. 3).

Measurements of cross strata in the conglomerate revealed southerly or westerly transport directions (Fig. 2), except in the Owlshead Mountains section, where easterly directions were obtained. The directions in the Owlshead may reflect a local western source; and the western edge of the basin (Fig. 2) was drawn with a northerly trend close to this section, based

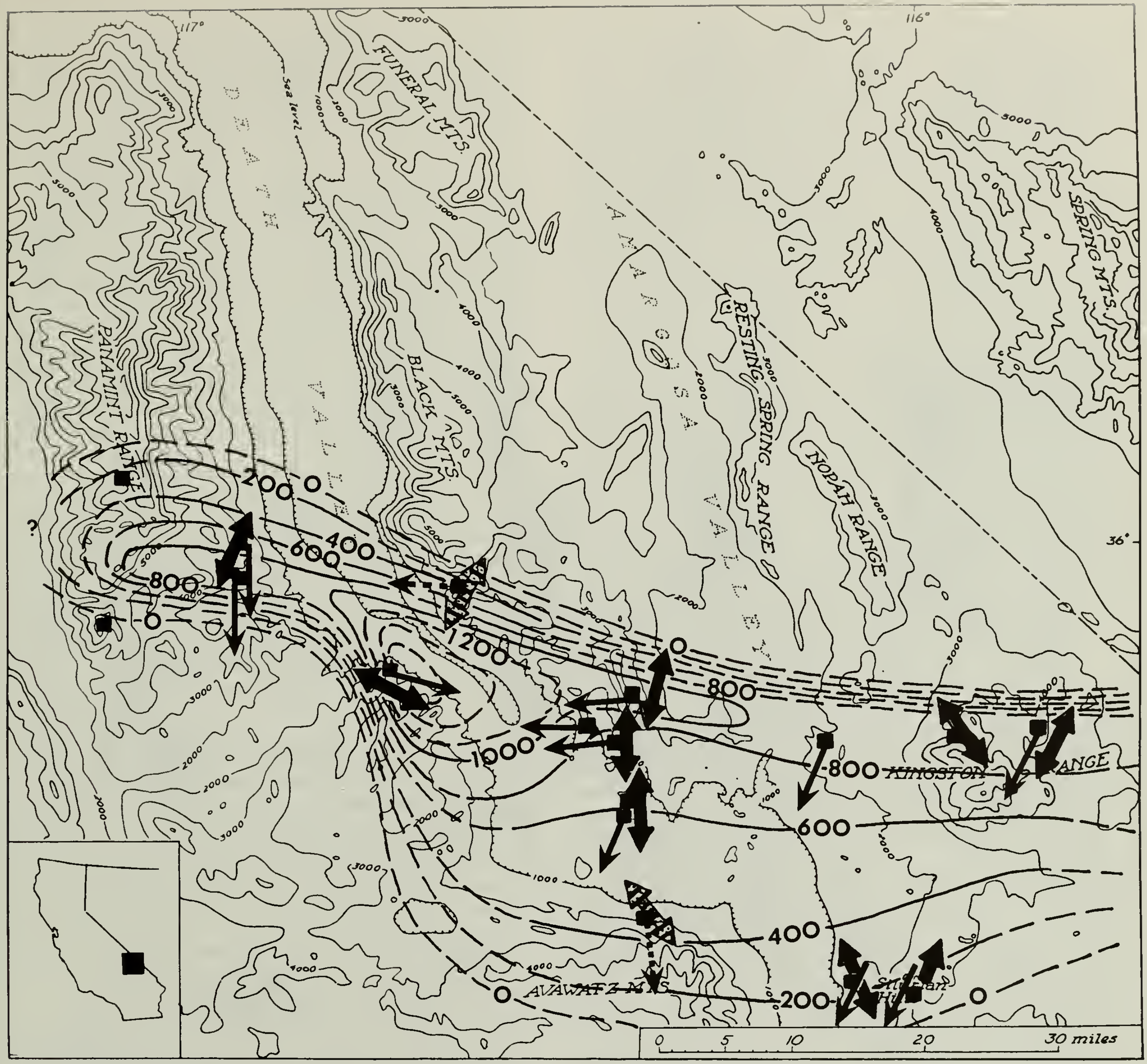


Figure 2. Isopach and paleocurrent map of arkose member. Small arrows are mean directions of foreset dips in conglomeratic units of member. Broad arrows are directions of foreset dips in sandstone units of member. Dashed arrows are probable directions in structurally complex areas. Data represent about 1,200 foreset measurements, 13 measured sections, and observations from additional sections. Isopach interval is 200 ft.

on these data. The cross strata rosettes show scatter, but are unimodal or bimodal with modes generally about 90° apart. The conglomeratic zones are interpreted to be the deposits of braided streams debouching into the basin.

The sandstone zones are composed of fine- to coarse-grained sandstone strata. In some zones, very subordinate, discontinuous beds of chert, shale, and carbonate, a few inches thick, are interbedded with the sandstone. The sandstone is typified by tabular beds that range in thickness from several inches to 3 ft (up to 0.9 m); internally, the beds are horizontally laminated, planar cross stratified, or both. The planar cross strata range in thickness from a few inches to 2 ft (0.6 m). Herringbone cross-strata sets are common in the sandstone zones. Foreset orientations of the cross strata show little scatter, compared to the conglomeratic units, and show unimodal northerly directions in some zones and bimodal northerly and southerly directions in others. The bimodal rosettes have 180° mode separation typical of tidal currents (Fig. 2). The unimodal rosettes that show directions opposite those of the conglomeratic zones represent flood-tide-dominated environments. In the Owlshead Mountains section, the sandstone zones, like the conglomeratic zones, show anomalous directions. The east-west tidal direction of the Owlshead section supports the north-trending border of the basin shown in Figure 2. The sandstone zones are interpreted as deposits of nearshore tidal environments such as tidal deltas.

The cycles are complicated by the presence of zones of interbedded sandstone and conglomerate, which appear either to be due to small-scale interfingering of the fluvial and tidal environments or to be tidal dominated. In addition, the

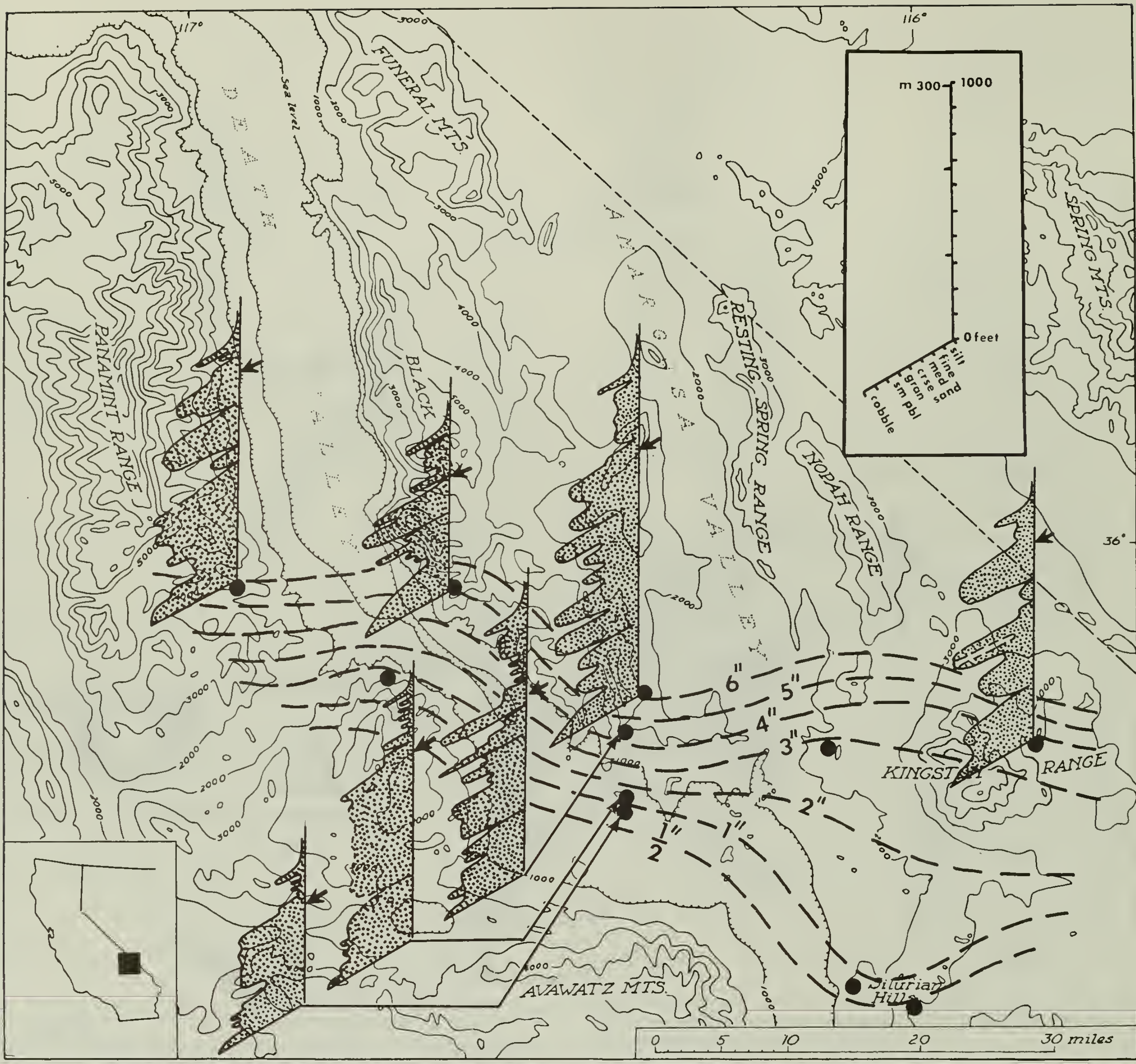


Figure 3. Variations of maximum grain size in lower members. Dashed contours are trend of maximum pebble sizes in arkose member, excluding basal conglomerate. Vertical sections are plots of maximum grain size versus stratigraphic position for selected sections on east-west and north-south lines. Arkose member is from base to top of last conglomerate "pulse" (arrow); feldspathic sandstone member is between arrow and last sandstone "pulse."

sandstone-conglomerate cycles are interrupted high in the member by about 100 ft (30 m) of horizontally laminated sandstone strata. This laminated zone contains scattered, contorted planar cross strata and resembles deposits of modern floods such as those described by McKee and others (1967). On the other hand, this zone may represent a tidal environment with a high regime flow.

The initial site of deposition of the Crystal Spring Formation was a generally east-trending shoreline zone at least 75 mi (121 km) long. Paleocurrent data (Fig. 2) and pebble-size data (Fig. 3) indicate a northern upland source for the sediments. If one applies Sternberg's Law to the pebble-size trends from the central part of the basin (excluding the basal conglomerate), the source is estimated to have been 2 to 4 mi (3.2 to 6.4 km) north of the Eclipse mine area in the central Ibex Hills (point B on inset map of Fig. 1).[2] A close source is also indicated by the angular and arkosic nature of the sediments and the near-

[2] *Sternberg's Law relates pebble size (sometimes stated as weight) to distance of transport.* $Y = Y_0 e^{-ax}$ *where Y is the diameter of the largest pebble a given distance from some reference point;* Y_0 *is the diameter of the largest pebble at the reference point; a was originally stated as a "wear coefficient" but is a constant probably related more to slope factors; and x is the distance between the point at which Y was measured and the reference point.*

Four sections on a nearly north-south line (central panel of Fig. 1) were used for the computation. The constant a was determined using known pebble sizes and distances for this north-south line and the shore character of the bedforms. Figure 3 also shows that the cyclicity of grain size is best defined in northern exposures and that the "pulses" of conglomerate are more subdued to the south. The cyclicity is interpreted as due to periodic uplift of the northern upland, possibly along a bounding fault zone. The contrast in detectable cyclicity is due to the distance from the bounding zone of the upland.

The bedforms, directional features, and grain-size trends indicate that the sediments were transported generally southward into the basin, where the finer fraction was reworked by generally north-south tidal currents. The cyclicity is attributable to an interfingering of fluvial and tidal environments controlled by periodic uplift in the source area. The member is thus interpreted as deposited in a complex system of nearshore, high-energy environments shifting laterally and vertically through time. Fluvial and tidal sediments are essentially the same in composition and roundness and, therefore, probably resided in their respective environments long enough to acquire distinctive bedforms, but not long enough to become differentially altered before burial.

Abrupt thickness changes within the member probably reflect faulting within the basin during sedimentation, as illustrated in the central panel of Figure 1 and in Figure 6. As the member records high-energy shoreline loci of deposition with a northern upland, finer offshore facies, such as siltstone and shale, might be expected to have been present to the south of the present exposures of the member. However, in later Crystal Spring time (detailed below) the southern region was uplifted and any record of such sediment was probably lost by erosion. An alternate explanation for the lack of this facies would be that little or no fine-grained material was available in the source.

FELDSPATHIC SANDSTONE MEMBER

The feldspathic sandstone member is here defined as the red to purple sandstone, siltstone, and shale that lie above the arkose member and below massive, purple mudstone. The top of the member is everywhere marked by a layer of coarse sandstone, generally only a few feet thick, which is represented by the last "pulse" in Figure 3. The isopach map of the member (Fig. 4) shows similar trends to that of the arkose member, indicating little change in the geometry of the basin. The feldspathic sandstone, however, may have extended farther northward beyond the northern limits of the arkose member.

The basal part of the member consists of tabular beds of fine- to coarse-grained sandstone generally less than 1 ft (0.3 m) thick. Internally, these beds are horizontally laminated, planar cross stratified, or ripple laminated, or they contain some combination of these features. The upper surfaces of the beds are commonly rippled. All types of ripples were observed, including interference ripples, oscillation ripples, lunate/linguoid ripples, and asymmetric ripples. The beds are commonly separated from one another by shaly partings exhibiting mudcracks. The planar cross strata are, in places, arranged in herringbone sets or show evidence of reactivation by opposing currents, as in tidal environments. This basal zone is 100 ft (30 m) or more thick in the central part of the basin, but elsewhere ranges from 30 to 70 ft (9 to 21 m) thick. Orientations of cross strata in the sandstone reveal unimodal northerly current directions at some localities. At other localities, bimodal, generally north-south currents are indicated (Fig. 4). Ripple directions, on the other hand, show multimodal directions typical of intertidal sand bodies. As in the arkose member, the Owlshead Mountains section shows western directions. The current directions are interpreted as due to tidal currents, with the unimodal northerly directions forming in areas dominated by flood tides.

The sandstone zone comprises most of the member at some localities. Ordinarily, however, the zone grades upward to shaly strata which form as much as 60 percent of the member. The sandstone of the lower part of the member generally grades upward into interbedded sandstone, siltstone, and shale which form cycles composed of fining-upward sequences typical of prograding tidal-flat deposition. The sandstone layers contain planar cross strata and ripples like those of the lower part of the member. The silty and shaly layers are thinly laminated and contain abundant mudcracks and shale-chip conglomerates formed from the breakup of mudcrack polygons. These features support an intertidal origin for these rocks. The fining-upward sequences range in thickness from about one to several meters. The thickness of fining-upward tidal sequences approximates the paleotidal range (Klein, 1971); however, the thicker sequences probably reflect subsidence more than the vertical distance between high and low tides. An approximate water depth of 0.5 m in one area was calculated from the grain size and wavelength of oscillation ripples in a fining-upward sequence.[3] The zone composed of interbedded sandstone, siltstone, and shale generally grades upward into shaly strata. Thus, the whole member forms a fining-upward sequence on a large scale, grading upward from sandstone to sandstone, siltstone, and shale and, finally, to shale. Locally, for example, in the Ibex Hills, where the member shows abrupt increases in thickness, it consists of repetitions of the large-scale fining-upward sequence. These thickness changes seem to be related to faulting in the basin during deposition of the sediments, as illustrated in Figures 1 and 6.

The member is interpreted as representing nearshore tidal sand bodies (lower part) and tidal flat (middle and upper part) deposits. The tidal current directions, like those of the arkose member, indicate northerly flood tides and southerly ebb tides. The fining-upward sequences suggest that the tidal flats prograded southward over the basin. The lack of sediment coarser than coarse sand and the abundance, as well as the lowermost occurrence, of shale indicates that the relief of the northern source, which was relatively rugged during deposition of the arkose member, became subdued.

A preliminary study of thin sections obtained from the feldspathic sandstone indicates a greater percentage of quartz than in the arkose member. The quartz content of the sandstone ranges from about 35 percent to about 70 percent. The more quartzose samples contain well-rounded grains and show better sorting than samples from the arkose. The red color of the member is due to abundant hematite in the matrix of the sediments. The relatively high quartz content may be due to removal of feldspar by weathering when the source area was reduced to low relief. It also may reflect a relatively long residence time in the tidal environment. Deeper weathering at the source is

---

Sternberg equation. In order to determine the distance from the northernmost point (B) to the source of the sediment, only Y_o must be known—that is, the probable maximum pebble size available in the source area. The critical assumption is that the transgressive basal conglomerate contains sizes representative of those available in the source. The maximum size observed in the basal conglomerate was 10 in. (254 mm). The range of 2 to 4 mi (3.2 to 6.4 km) was obtained by using maximum and minimum a values, based on the data, and the assumed 10-in. size for Y_o. Because of these assumptions, the distance to source should be taken as only a "ballpark" estimate. For comparison, even if the source contained boulders 1 m in diameter. it would be only 24 km away.

---

[3] *Harms (1969) has shown that wave period (T) can be calculated from $T = \pi(E_{max}/U_{w\,max})$ where E_{max} = average ripple crest spacing and U_w (current velocity) must exceed about 19 cm per sec for fine sand. Using crest spacing of 5 cm for one Crystal Spring sample of rippled fine sandstone, T = 0.826 sec. If the maximum depth of wave-affected rippling is about $4/9\lambda$, and $\lambda = (g/2\pi)T^2$, then $4/9\lambda$ (maximum depth for the oscillation ripple above) would be 47 cm.*

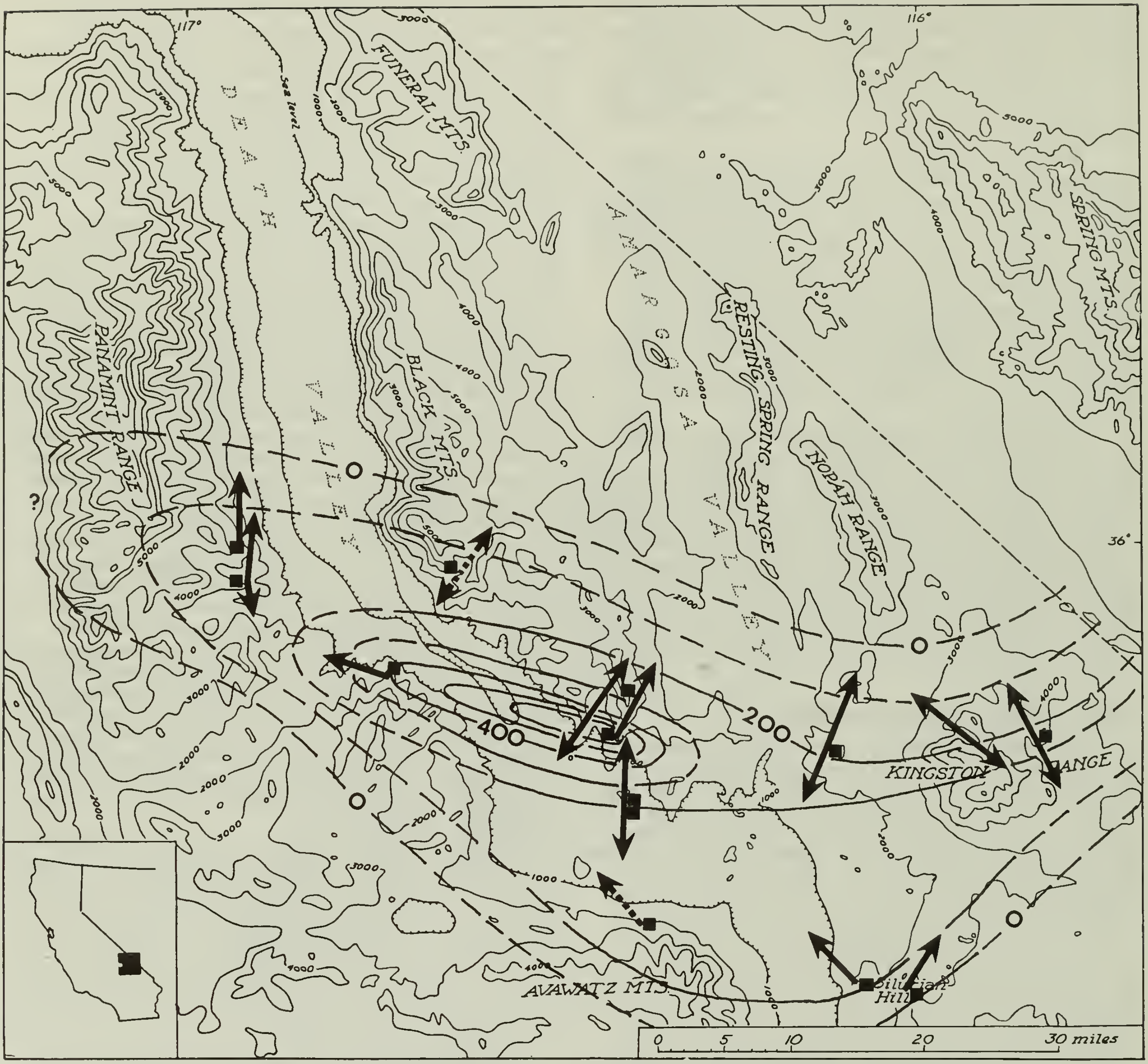


Figure 4. Isopach and paleocurrent map of feldspathic sandstone member. Arrows represent mean directions of foreset dips. Dashed arrows are probable directions in structurally complex areas. Data represent 356 foreset measurements and 13 measured sections. Isopach interval is 100 ft.

indicated by the abundant hematitic material in the matrix of the sediments.

PURPLE MUDSTONE MEMBER

The mudstone member is massive to poorly bedded, purple to red, silty to sandy mudstone that overlies the uppermost sandstone bed of the underlying member. It is overlain by the sandy carbonate zone at the base of the dolomite member. The mudstone member contains scattered grains of sand in its lower part and consists of laminated siltstone and shale in its upper part. The upper part in some areas is metamorphosed below the diabase sill. The member ranges from 30 to 160 ft (9 to 49 m) thick but is commonly about 125 ft (38 m) thick. It is distributed similarly to the lower members, being thickest through the central part of the basin and thinnest in the Silurian Hills. The member is termed a mudstone because it generally lacks regular parting planes, laminations, or bedding; it is poorly sorted and is clayey to silty in composition. A preliminary study of its petrology indicates that 75 to 80 percent of the mudstone is an illite-hematite mixture. The remainder consists of silt and sand grains of quartz (15 to 20 percent) and feldspar (less than 5 percent). The mudstone contains a few thin, discontinuous carbonate beds and very fine sandstone beds and is typified by bluish to greenish reduction spots.

Although the member lies between strata with bedforms that indicate an intertidal environment, it is lacking in distinctive bedforms. The member is apparently the last unit of predominantly clastic material in Crystal Spring time to be derived from the northern upland, although the lack of detailed data from the upper units makes this conclusion tentative. During the deposition of the mudstone, the northern upland was very low in relief. The mudstone may represent low-gradient stream, estuary, or tidal-marsh deposits.

DOLOMITE MEMBER

The dolomite member is here defined as the chiefly dolomitic strata that include the sandy carbonate zone ("upper quartzite member" of Wright, 1968) and the lower part of the "carbonate member" of Wright (1968). The sandy carbonate strata are about 100 ft (30 m) thick across the northern and central parts of the basin and thin southward to less than 50 ft (15 m) in the Silurian Hills. This part of the member does not show a north-south facies change.

In the southern occurrences of the dolomite member, the beds that overlie the sandy zone consist of a zone of dolomite 50 to 100 ft thick (15 to 30 m), in turn overlain by several hundreds of feet of interbedded quartzite, dolomite, shale, and limestone. In the northern exposures, the basal sandy zone is overlain by dolomitic strata as much as 300 ft (91 m) thick. The dolomitic strata display a cyclicity expressed by alternating layers of two types of dolomitic rock. One type consists of light-gray to light-brown dolomite that weathers to various shades of orange. It is very fine grained and contains scattered quartz sand grains, zones of dolomite intraclasts, and subordinate chert beds. It is characterized by wavy to irregular laminations and locally shows ripple cross laminations. The irregular laminae resemble those formed by algal mats. Layers of this type of dolomitic rock range from 1 to 30 ft (0.3 to 9 m) thick and are thickest low in the member (Fig. 5).

The other type of dolomitic rock is very thinly bedded, medium- to dark-gray siliceous dolomite. This type contains thin, wavy laminated beds of dolomite that alternate with thin layers of microcrystalline quartz. On weathered surfaces the latter stand out as brown ribs. The thin layers are typically contorted into folds apparently caused by soft-sediment deformation. The wavy laminae of the dolomite are archlike in places and resemble those produced by algal mats. Layers of this siliceous dolomite range from 1 to 15 ft (0.3 to 4.6 m) thick and increase in thickness and abundance upward (Fig. 5). The top part of the dolomite member in its northerly occurrences consists of a unit of clastic limestone as much as 30 ft (9 m) thick. This unit is composed of sand- to silt-size limestone clasts and contains quartz sand lenses. It is commonly rippled and shows small tidal cross laminations.

Although the dolomite member is about 400 ft (122 m) thick in the Alexander Hills, only 100 ft (30 m) of similar strata can be recognized in the Silurian Hills, 15 mi (24 km) to the south. Either the member simply thins to the south as do lower members, or in the south the upper part is replaced by chiefly quartzitic strata.

The alternation of the two types of dolomitic rock in the northern exposures of the dolomite member is interpreted as due to interfingering of rocks of two depositional environments. The light-gray to light-brown dolomite unit is interpreted as deposited in the higher-energy environment, possibly in a low intertidal or subtidal zone. The siliceous dolomite unit is interpreted as deposited in a lower-energy environment, possibly in a high intertidal or supratidal zone. As the strata that possibly indicate a high intertidal or supratidal environment become more common upward, the sequence can be attributed to prograding (regressive) carbonate tidal-flat deposition. The clastic beds at the top of the member in the northern exposures may represent transgression or the return of high-energy tidal environments, possibly related to uplift in the south, as indicated by the quartzose sand in the upper part of the dolomite member there.

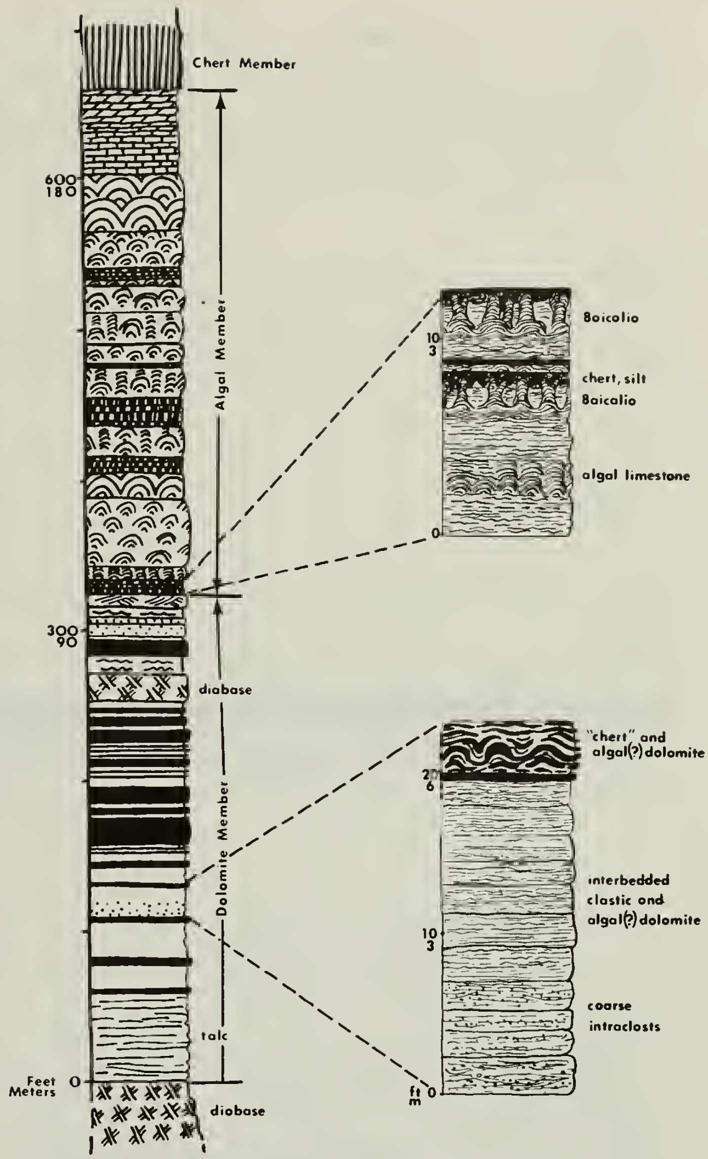


Figure 5. Stratigraphic column of carbonate members in Alexander Hills area. Lower inset details lithologies of single cycle of dolomite member (solid black units of column correspond to upper, ribbed strata of inset; solid white units of column correspond to lower strata of inset). Upper inset details stromatolite cycles within basal zone of larger stromatolitic cycles of algal member. At top of dolomite member is zone of rippled and cross-stratified clastic limestone.

ALGAL MEMBER

In the northern exposures of the dolomite member, it is overlain by a carbonate unit, as much as 300 ft (91 m) thick, and characterized by abundant stromatolitic structures. It is here designated as the algal member. It consists of limestone, except in the Saratoga Hills where it is dolomitic. The member contains cyclical repetitions of morphologically different stromatolite forms and includes about 50 ft (15 m) of non-stromatolitic limestone and dolomite at the top. The nature of the cycles varies from place to place. The cycles in the

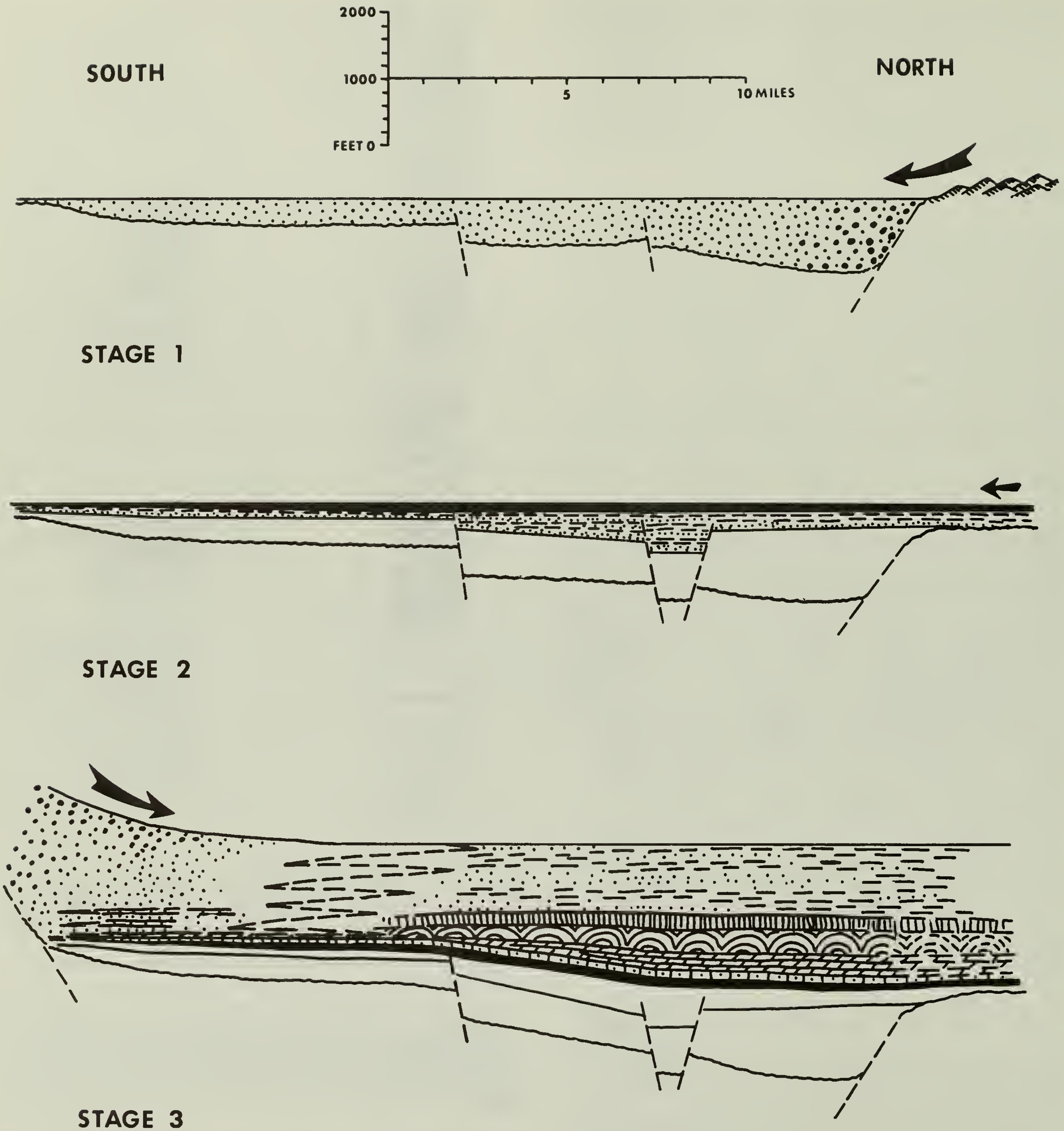


Figure 6. Summary diagram of general development of Crystal Spring basin. Section is composite north-south cross section through center of basin. Exact position of southern upland not known. See text for details of stages.

southern Panamint Range and the Alexander Hills are similar. In the Kingston Range and Saratoga Hills, different sequences of stromatolites were observed.

In the Alexander Hills (Fig. 5), a typical cycle consists of a well-defined basal zone, itself cyclical, overlain by massive zones containing as many as three stromatolite forms. The lowermost cycle in the Alexander Hills is considered to be complete. It consists of a basal zone, 12 ft (3.7 m) thick, composed of cyclical repetitions of the sequence: wavy laminae–algal heads–*Baicalia* columns–silt and chert (Fig. 5, upper inset). This sequence is repeated as many as five times in the basal zone. If intertidal models of Holocene stromatolites apply, this zone may represent transgressive intertidal sequences.

Overlying the basal zone is a bed, 8 ft (2.4 m) thick, consisting entirely of vertically stacked algal heads tending toward columnar development. This is followed by a massive limestone bed, 45 ft (13.7 m) thick, which appears to be composed entirely of algal heads, typically laterally linked, 3 to 4 in. (7 to 10 cm) in diameter. The upper zone of the lowermost cycle is 17 ft (5 m) thick and consists entirely of algal heads about 1 ft (30 cm) in diameter. These upper zones may represent slight regression, followed by relative stability of environment during subsidence prior to the next transgressive basal zone.

Each of the cycles that overlies the lowermost cycle in the Alexander Hills contains the basal *Baicalia*-bearing zone, but the successive zones may differ slightly from those of the lowermost cycle. Commonly the zone characterized by stacked algal heads is missing, and stacked algal heads occur within the zone of laterally linked small heads. The large (1 ft) heads are absent in some cycles. As independent evidence of an intertidal environment for the stromatolites, such as dessiccation features, has not been found, they may have inhabited subtidal areas. The elliptical shape, in plan view, of many of the heads indicates that they grew in current-influenced areas.

In the southern Panamint Range and Saratoga Hills, cylindrical *Conophyton* forms are interbedded with forms possibly of intertidal origin. The relations are not everywhere clear, but *Conophyton* may have grown in low intertidal environments, as in one area they lie between possibly low intertidal and mid-intertidal stromatolites in a vertical sequence. The *Conophyton* cylinders, generally 2 to 3 in. (5 to 8 cm) in diameter and less than 2 ft (0.6 m) long, are inclined and parallel. They apparently grew in that position and may have been aligned with tidal currents.

The algal member of the northern localities probably grades southward into strata composed chiefly of quartzite interlayered with subordinate limestone. The cycles in the algal member thus may be related to periodic uplift of the southern area, which was shedding quartz sand into the basin. The dolomite and algal members record persistent carbonate shelf environments over the northern part of the basin. The northern upland, which was emergent during deposition of the lower members, is not evident in the two carbonate members. The carbonate shelf thus probably extended northward beyond the limits of the lower members.

The large volume of carbonate in these members may have originated from carbonate-secreting algae or from precipitation of carbonate as a result of photosynthetic activity of the algae; however, many of the stromatolite forms are analogous to those created by the trapping of carbonate debris in Holocene stromatolites.

CHERT AND UPPER UNITS

The upper two members of the formation have not been studied in detail. (Their distribution is shown in Fig. 1.) The chert is dark, dense, and fine-grained massive rock, 100 to 500 ft (30 to 152 m) thick (Wright, 1968). The upper sedimentary units are composed of shale, sandstone, and dolomite 200 to 1,200 ft (61 to 366 m) thick (Wright, 1968). Both chert and upper sedimentary units occur only in northern exposures and may be lateral equivalents of quartzite and conglomeratic strata to the south (Fig. 1). These members, then, may be northern basin facies deposited from the rising southern landmass that began to form earlier in Crystal Spring time.

SUMMARY

The Crystal Spring Formation consists of sediments deposited during the initiation of a late Precambrian intracratonic trough. Figure 6 outlines the main stages in the history of the basin in Crystal Spring time. Apparently after an initial marine transgression, a trough formed and was filled with coarse arkosic debris derived from a rugged northern upland that was probably uplifted along a fault (Fig. 6, stage 1). When uplift ceased, the upland was worn down, and, first, tidal-flat sand and mud, then mud possibly of fluvial origin prograded over the basin (Fig. 6, stage 2). The rest of Crystal Spring time was dominated by deposition of quartzose detritus in the southern part of the basin, while, first, algal carbonate shelf sediments, then shale-rich clastic sediments were being deposited over the central and northern parts of the basin (Fig. 6, stage 3). The quartzite and conglomerate of the southern part of the basin were probably derived from a rising southern upland. These major tectonic elements—the trough and the northern and southern uplands—first active in Crystal Spring time, were persistent features, controlling sedimentation throughout the rest of late Precambrian time (Wright and Troxel, 1967; Wright and others, 1974).

REFERENCES CITED

Harms, J. C., 1969, Hydraulic significance of some sand ripples: Geol. Soc. America Bull., v. 80, no. 3, p. 363–396.

Howell, D. G., 1971, A stromatolite from the Proterozoic Pahrump Group, eastern California: Jour. Paleontology, v. 45, no. 1, p. 48–51.

Klein, G. deV., 1971, A sedimentary model for determining paleotidal range: Geol. Soc. America Bull., v. 82, no. 9, p. 2585–2592.

Kupfer, D. H., 1960, Thrust faulting and chaos structure, Silurian Hills, San Bernardino County, California: Geol. Soc. America Bull., v. 71, no. 1, p. 181–214.

McKee, E. D., Crosby, E. J., and Berryhill, H. L., 1967, Flood deposits, Bijou Creek, Colorado, June 1965: Jour. Sed. Petrology, v. 37, no. 3, p. 829–851.

Raaben, M. E., 1969, Columnar stromatolites and late Precambrian stratigraphy: Am. Jour. Sci., v. 267, p. 1–18.

Wasserburg, G. J., Wetherill, G. W., and Wright, L. A., 1959, Ages in the Precambrian terrane of Death Valley, California: Jour. Geology, v. 67, p. 702–708.

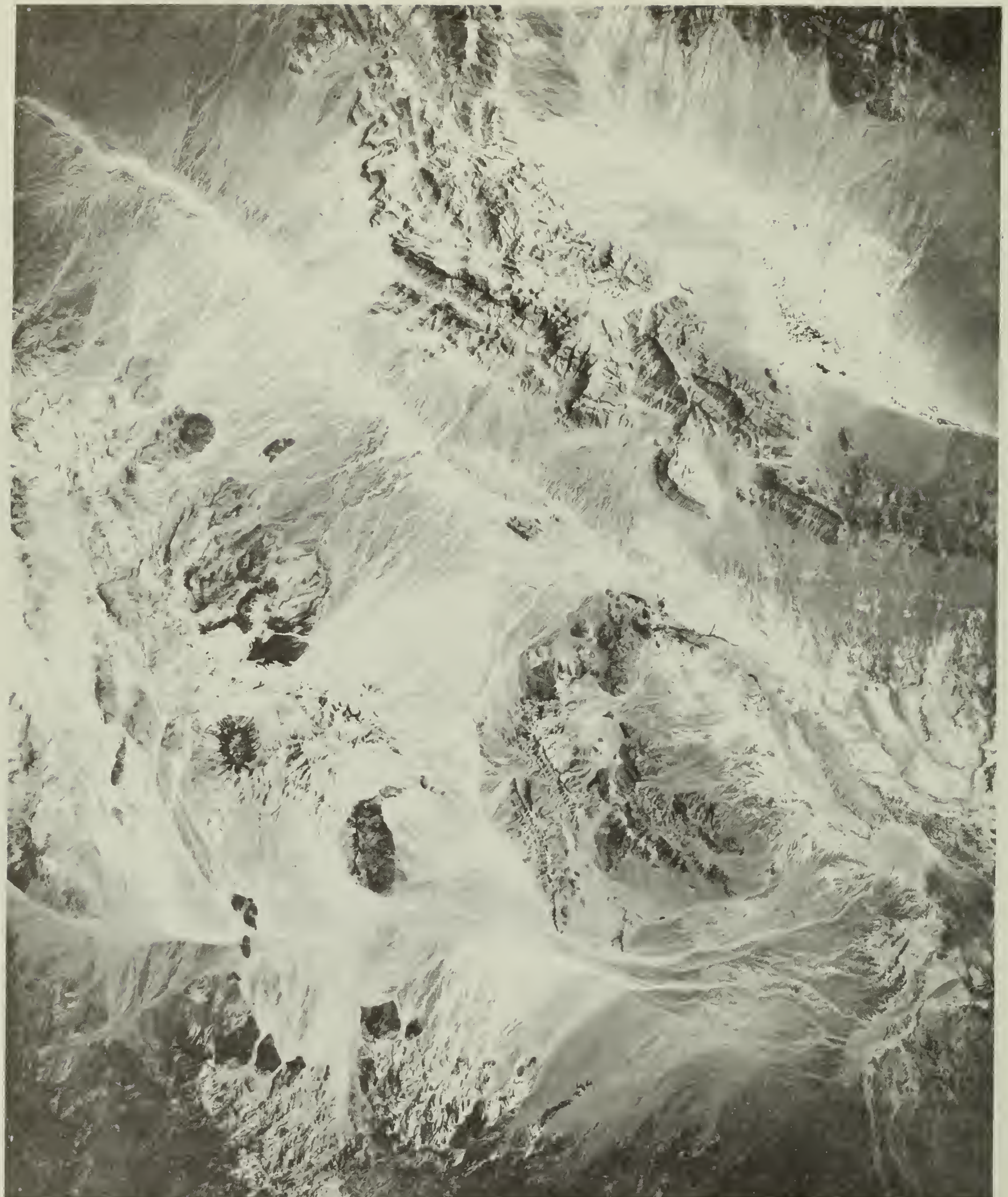
Wright, L. A., and Troxel, B. W., 1967, Limitations on right-lateral, strike-slip displacement, Death Valley and Furnace Creek fault zones, California: Geol. Soc. America Bull., v. 78, p. 933–950.

Wright, L. A., 1968, Talc deposits of the southern Death Valley–Kingston Range region, California: California Div. Mines and Geology Spec. Rept. 95, 79 p.

Wright, L. A., Williams, E. G., and Troxel, B. W., 1974, Late Precambrian sedimentation, Death Valley, evidence of prolonged and varied tectonic control of a basin and shelf environment: Geol. Soc. America, Abs. with Programs (Cordilleran Sec.), v. 6, no. 3, p. 278.

Wrucke, C. T., 1972, Correlation of Precambrian diabase in Arizona and California: Geol. Soc. America, Abs. with Programs (Cordilleran Sec.), v. 4, no. 3, p. 265.

This research was supported by NSF Grant GA-16119.



Vertical air photograph, vicinity of Shoshone (upper right side of Dublin Hills in right center of photo). Amargosa River drains southward from upper left corner of photo. Resting Spring Range on east side of river and part of Greenwater Range on west. Greenwater Valley crosses lower part of photo. Salsberry Pass is barely on lower part of photo. Normal, west-dipping faults offset many of the rocks. U.S. Geological Survey - U.S. Air Force photograph.

Geology of the Shoshone Volcanics, Death Valley Region, Eastern California

Richard Haefner[1]

GEOLOGIC SETTING

Within the Black Mountains block of the Death Valley region of eastern California is a terrane composed primarily of Cenozoic intrusive and extrusive rocks, which covers an area of about 1,300 sq km. The volcanic rocks of this terrane are predominantly rhyolitic but also include dacitic, andesitic, and basaltic units. They appear to be genetically related to the Cenozoic deformational features of the block (Noble and Wright, 1954; Wright and Troxel, 1971a, 1971b). The extrusive rocks that have been mapped thus far are mostly lava flows and air-fall tuffs, but ash-flow tuffs form much of the lower part of the volcanic sequence in the southern part of the field (L. A. Wright, personal commun.). Included within the rhyolitic rocks is an accumulation of lava flows and tuffs, as much as 900 m thick, to which the name "Shoshone Volcanics" is here applied and which is the subject of this summary report.

Although the field is incompletely explored, it appears to consist dominantly of acid intrusive and extrusive rocks; rocks of intermediate to basic composition are also present but subordinate.

Clastic sediments derived from the igneous rocks, together with evaporites, were deposited in local basins (Drewes, 1963). Some of these rocks contain commercial borate deposits. Two major sediment accumulations have been named Copper Canyon Formation (Curry, 1941; Drewes, 1963, p. 32) and Furnace Creek Formation (Noble, 1941, p. 956).

DESCRIPTION

The Shoshone Volcanics are exposed almost continuously in a belt about 26 km long, which extends northward (Fig. 1). They also are exposed in more westerly parts of the Greenwater Range and in the Black Mountains.

In the area of Figure 1, the Shoshone Volcanics overlie Cambrian sedimentary rocks, Cenozoic quartz monzonite plutons, and older acid volcanic rocks with an erosional unconformity. The contact between the two accumulations of volcanic rock shows marked angular discordance. Southwest of the area shown in Figure 1, the Shoshone Volcanics also overlie ash-flow tuffs and andesite lava flows. The eroded surface of the Shoshone Volcanics is in turn overlain by air-fall tuffs and lava flows of the Greenwater Volcanics. This contact commonly also shows angular discordance. As the dip direction of all three volcanic accumulations is generally eastward, the angular unconformities suggest continued eastward tilting of fault blocks during emplacement of the volcanic rocks. Dark colored lava flows of basaltic and andesitic composition overlie the Shoshone Volcanics in some places.

The Shoshone Volcanics are prominently layered, the layers producing a striped landscape colored in various shades of pink, yellow, and gray. A regular sequence of layers is obvious, even from a distance. A yellow layer is generally overlain successively by a gray layer, a pink layer, locally a second gray layer, and another yellow layer (Table 1). At many localities, three or four of these sequences are exposed in layer-cake fashion on a single slope, individual sequences being 45 to 120 m thick. The sequences on steeper slopes produce cliffs (gray and pink layers), alternating with benches (yellow layers).

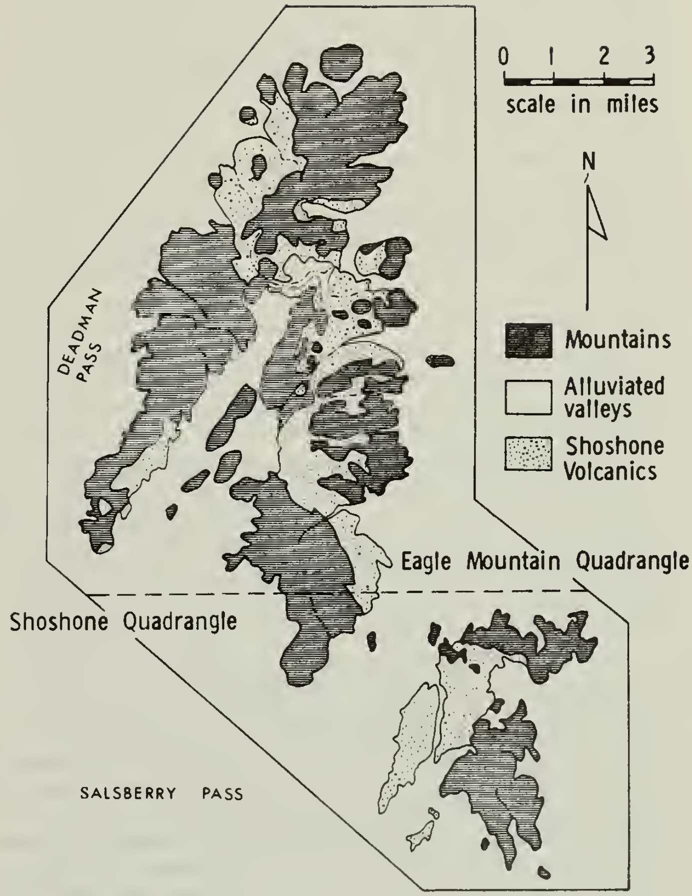


Figure 1. Distribution of the Shoshone Volcanics in southeast Greenwater Range (Eagle Mountain quadrangle) and Dublin Hills (Shoshone quadrangle).

As the volcanic rocks dip eastward, the best views of the layer sequences are from the west side of the mountains. The sequences in the Dublin Hills are among the best exposed (Chesterman, 1973) and are also easily accessible. These dominate the view eastward as one travels toward Salsberry Pass from Greenwater Valley.

CORRELATION AND AGE

The Shoshone Volcanics are equivalent to a part of Drewes's "older volcanics" in the Funeral Peak quadrangle to the west of the area shown in Figure 1 (Drewes, 1963). This unit is lithologically similar to the Shoshone Volcanics even to the zones within individual lava flows, and it also overlies eroded quartz monzonite plutons and underlies Greenwater Volcanics.

A tentative age for the Shoshone Volcanics may be inferred from isotope ages of rock units elsewhere in the volcanic field. Fleck (1970, p. 2810) reported seven K-Ar ages for

[1]*Department of Geological Sciences, State University College of New York at New Paltz, New Paltz, New York 12561.*

"older volcanics" from the Funeral Peak quadrangle: three from Dante's View and four from Hidden Springs. The dates from Dante's View (6.32 ± 0.13, 6.34 ± 0.13, 6.49 ± 0.13) are clearly younger than those from Hidden Spring (8.02 ± 0.16, 7.60 ± 0.30, 7.77 ± 0.15, 8.02 ± 0.16). These yield an early Pliocene age for Hidden Spring and a middle Pliocene age for Dante's View, based on the time scale of Evernden and others (1964). Although the "older volcanics" of Drewes include more than the equivalent of the Shoshone Volcanics, the early to middle Pliocene dates fit the permissible interval allowed the Shoshone Volcanics by radiometric dating of the quartz monzonite plutons and the Greenwater Volcanics. The former have yielded middle to late Miocene K-Ar dates (Stern and others, 1966; Armstrong, 1966). The Greenwater Volcanics, on the other hand, have yielded late Pliocene K-Ar dates (Fleck, 1970). The early to middle Pliocene age for the Shoshone Volcanics thus indicated may be refined when more detailed correlations of volcanic units are made, and when the age of the volcanic rocks underlying the Shoshone Volcanics is better known.

PETROGRAPHY

The Shoshone Volcanics consist primarily of four lithotypes: (1) gray perlitic vitrophyre, (2) red-brown to pinkish-gray felsophyre, (3) yellow felsophyre, and (4) yellow devitrified tuff (Table 1).

Phenocrysts of the Vitrophyre and Felsophyre. Phenocrysts comprise 3 to 20 percent of the total rock volume (Table 2); the remainder is glass or cryptocrystalline groundmass Some of the layer sequences are crystal rich, others are crystal poor (Haefner, 1972). The Dublin Hills sequences are crystal poor. Crystal-rich sequences overlie crystal-poor sequences in the vicinity of Brown Peak.

Plagioclase, the only feldspar, forms the largest (4 mm) and most abundant phenocrysts. Most crystals are andesine, but the composition ranges from oligoclase to labradorite.

Hornblende forms euhedral prisms to 1.5 mm long. It occurs as ordinary hornblende (pleochroic in green and brown), and as the varieties variously known as oxyhornblende, lamprobolite, or basaltic hornblende (pleochroic in red, green, and brown). Crystals of the latter variety are commonly rimmed with opaque minerals. Oxyhornblende occurs only in the pink felsophyre; ordinary hornblende occurs in the vitrophyre and yellow felsophyre.

TABLE 1. TERMINOLOGY AND INTERPRETATION OF LAYERING IN SHOSHONE VOLCANICS

Color bands	Topographic expression (in steep areas)	Zones*	Limits of eruptive unit	
Yellow	Slope	Lower altered vitrophyre zone		
		Tuff zone		
		Upper altered vitrophyre zone	Single lava flow	Eruptive unit
Gray	Cliff	Upper vitrophyre zone (not always present)		
Pink	Cliff	Red felsophyre zone		
Gray	Cliff	Lower vitrophyre zone		
Yellow	Slope	Lower altered vitrophyre zone		
		Tuff zone	Air-fall tuff	
		Upper altered vitrophyre zone		

** Zones are mappable units, drawn in this column in their average proportional thicknesses. Altered vitrophyre zones consist predominantly of yellow felsophyre, which at a distance may be difficult to distinguish from the tuff zone.*

Biotite occurs as euhedra to 1.5 mm in maximum dimension. It, too, occurs as two varieties: ordinary biotite (pleochroic in green and brown), and a type which, by analogy with hornblende, may be called oxybiotite (pleochroic in red, green, and brown). Oxybiotite is commonly rimmed with opaques, and it occurs exclusively with oxyhornblende in the pink felsophyre.

Opaque minerals, largely magnetite, occur as small anhedral to euhedral grains from dust size to 0.3 mm, and as larger masses partly to completely replacing oxyhornblende and oxybiotite. Orthopyroxene and clinopyroxene form euhedral phenocrysts that may be as large as the hornblende and biotite but are much less common.

A xenocrystic character is inferred for some grains that are rounded or embayed. Quartz grains as much as 1 mm in diameter occur in a few thin sections and are consistently rounded or embayed. Species that occur as phenocrysts also commonly show such outlines.

Groundmass of the Vitrophyre and Felsophyre. The glass groundmass of the vitrophyre commonly is gray to black. Brown glass occurs as a local variation that is mixed in swirls and patches with gray glass. The glass contains abundant pyroxene crystallites of several habits. Perlitic structure is nearly ubiquitous. In thin section, concentric fractures outline spheres with centers spaced at intervals of 0.5 to 1.0 mm and associated with a boxwork of planar fractures. Spherulites, 0.1 to 30 cm, occur in the glass but are confined to the several meters of vitrophyre along the margins of the red felsophyre (Fig. 2). Some spherulites are hollow and contain quartz crystals, agate, or opal.

The groundmass of the yellow felsophyre is cryptocrystalline. The presence of the alteration mineral jarosite is indicated by x-ray diffraction analysis. Magnetite commonly is absent. Relict perlitic fractures are locally present, indicating that the yellow felsophyre groundmass was originally glass, and that it acquired its present crystallinity through devitrification.

The groundmass of the red felsophyre is cryptocrystalline. The red to pink color, which is particularly conspicuous around magnetite microphenocrysts and around the opacitized rims of oxyhornblende and oxybiotite, is produced by abundant patches of hematite. In contrast to the vitrophyre and yellow felsophyre, the red felsophyre is vesicular; small vesicles to 2 mm are nearly ubiquitous, but large vesicles to 1 cm also occur, most of them adjacent to the vitrophyre (Fig. 3).

Devitrified Yellow Tuff. The same phenocryst species as found in the vitrophyre and felsophyre are found in the tuff matrix. The matrix has been devitrified and is now cryptocrystalline, with only rare traces of the original glass shards. Magnetite appears to be absent in the matrix, whereas jarosite is present. Essential pumice lapilli (devitrified) and accidental lapilli and blocks of volcanic rocks are the most common clasts; clasts of sedimentary and metamorphic rocks are rare. Vitrophyre clasts are absent nearly everywhere, but yellow and red felsophyre clasts are common.

Two localities have been found where the tuff is not devitrified. The most accessible of these is the lowest tuff

TABLE 2. APPROXIMATE MODAL COMPOSITIONS, BY CONVENTIONAL POINT COUNTING

	Crystal-poor eruptive unit* (4,200 points) (%)	Crystal-rich eruptive unit† (2,000 points) (%)
Groundmass (glass)	97.75	81.8
Plagioclase	2.02	11.7
Biotite	0.12	0.6
Hornblende	0.07	4.7
Opaques	0.02	1.2
Augite/hypersthene	0.02	tr.
Total	100.00	100.0

** Dublin Hills, average of 21 vitrophyre thin sections from one lava flow.*
† West of Brown Peak, average of 10 vitrophyre thin sections from one lava flow.

exposed on the west side of the Dublin Hills where vitrophyre lapilli and glass shards are abundant, and jarosite does not occur in the matrix.

Paleomagnetic Properties of the Rocks. Ninety-six determinations of paleomagnetic direction were made on vitrophyre and red felsophyre of the layer sequences only because yellow felsophyre and tuff commonly are not magnetically susceptible. In each case, strike and plunge of the remanant magnetization was determined. As the rocks are relatively young, the orientation of the remanant magnetization should approximately parallel the magnetic direction of the Earth's present field, after correcting for postcooling deformation.

All the layer sequences examined exhibit normal polarity, except for specimens from exposures of one layer sequence, which are reversed (NW¼ sec. 3, T. 22 N., R. 5 E.).

All the vitrophyre samples, including those from the reversed sequence, have directions of remanant magnetization which parallel that predicted from the present field (Fig. 4). However, among specimens from the red felsophyre zone, fewer than half have remanant directions that parallel the predicted direction; the remainder have remanant directions with widely varying strikes and plunges. It appears that only the vitrophyre can be relied upon to be a faithful recorder of the Earth's magnetic field at the time of emplacement.

INTERPRETATION OF THE LAYER SEQUENCES

Types of Deposits. Within each yellow layer is a deposit, about 9 m thick, of well-stratified tuff. The remainder of the layer is yellow felsophyre (Table 1). The tuffs, which consist of interbedded lapilli-rich and lapilli-poor strata, are interpreted as air-fall tuffs. The tuff deposits divide the yellow color bands into additional mappable units, or zones (Table 1).

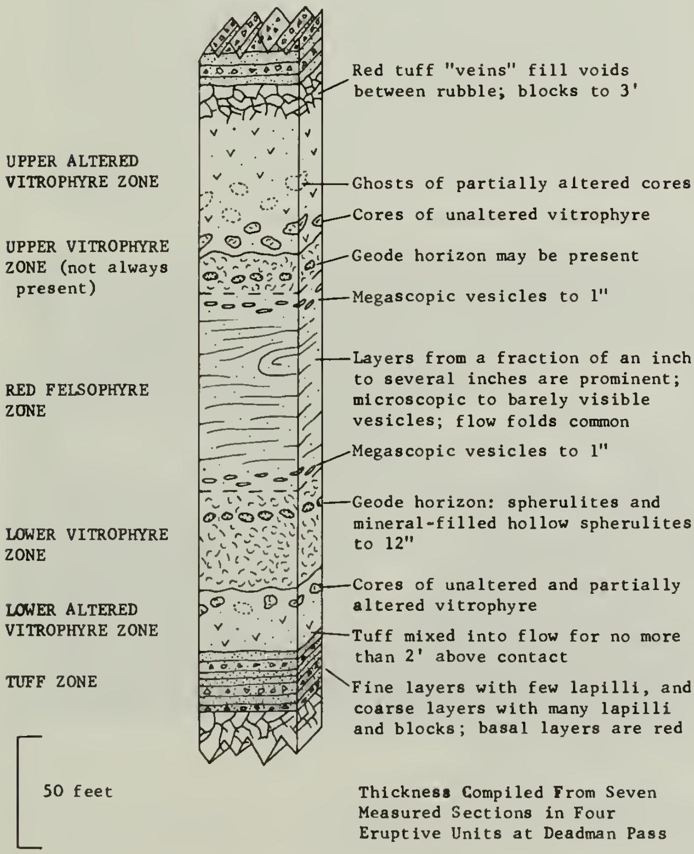


Figure 2. Features of typical eruptive unit in Shoshone Volcanics.

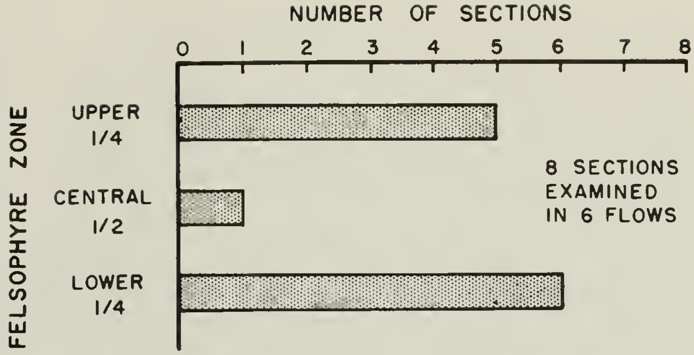


Figure 3. Distribution of megascopic vesicles in red felsophyre.

The sequence of zones between tuff deposits is interpreted as a single rhyolite lava flow (Fig. 2; Haefner, 1969). Glass shards, fiamme, and other characteristics of ash-flow deposits are absent in these zones. In addition, laminae in the red felsophyre zone are locally warped into folds, indicating that the magma flowed as a coherent mass. That this sequence of zones constitutes only one lava flow is evidenced by exposures of flow margins. Four flow margins are exposed in the Dublin Hills; at such localities, the red felsophyre constitutes a core, around which are wrapped the upper vitrophyre and yellow felsophyre zones to become the lower vitrophyre and yellow felsophyre zones (Fig. 5).

Alteration. The yellow felsophyre is interpreted as altered vitrophyre (Haefner, 1969, 1973). Evidence for this inference is the presence of relict perlitic fractures and the observation that yellow felsophyre merges with unaltered vitrophyre inward from both the upper and lower flow surfaces. Thus, a lava flow appears to consist essentially of a red felsophyre core encased in a vitrophyre sheath; the outer portions of the sheath have been altered, as have adjacent tuff deposits. Alteration is expressed as devitrification, the growth of jarosite, and the disappearance of magnetite.

Altered acid volcanic rocks are usually distributed about centers of fumarolic or hydrothermal activity. The Shoshone lava flows are unusual in that the altered rock occurs as a sheet-like body at the glassy top and at the glassy base of each lava flow. However, such alteration may not be unique to Death Valley; there are at least two localities in the Soviety Union from which apparently similar altered lava flows of acid composition have been described (Nasedkin, 1963).

A hypothesis for the origin of this alteration in the Shoshone Volcanics, which involves the interaction of magmatic with meteoric volatiles during the cooling of the flows, has been presented elsewhere (Haefner, 1969, 1973).

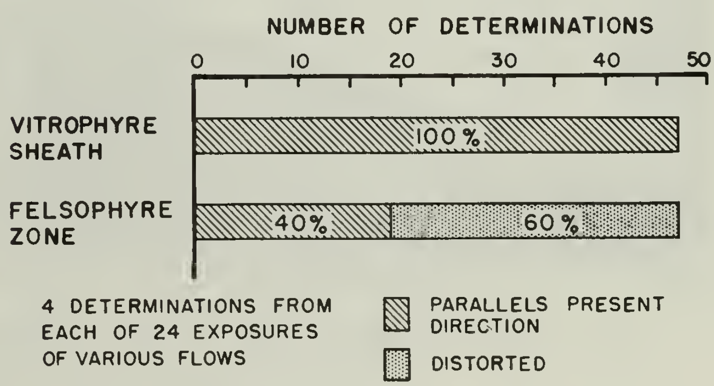


Figure 4. Determination of paleomagnetic direction.

EMPLACEMENT OF A TYPICAL ERUPTIVE UNIT

Crust of the Active Lava Flow. At the top of each lava flow that is overlain by a tuff deposit, veinlets of tuff extend downward into the flow, outlining individual blocks of yellow felsophyre (altered vitrophyre); locally, the blocks are pumiceous. These blocks apparently constituted a rubble crust formed by the chilling and breaking of lava on the upper surface of the active flow. The crust, about 4.5 m thick, composes only 3 to 9 percent of the total flow thickness; furthermore, crustal thickness does not depend on the thickness of the lava flow (Fig. 6).

Evidence of a rubble crust at the base of the lava flow would be generally difficult to detect, because tuff veinlets do not outline the rubble blocks. Nevertheless, ghostly remnants of angular blocks in the yellow felsophyre (altered vitrophyre) are observable at some localities and apparently originated as rubble. The lower crust is relatively thin, and massive, unaltered vitrophyre composes most of the part of the flow that underlies the red felsophyre zone (Fig. 2). If the lower crust was about as thick as the upper crust, then 80 to 95 percent of the lava flow remained liquid while the flow was active.

This large proportion of liquid magma contrasts with that inferred for many other acid lava flows, which are block lavas. These contained a much higher proportion of rubble during their active stage than do the flows of the Shoshone Volcanics. The overlying flows of the Greenwater Volcanics, exposed near Brown Peak, are block lavas. One measured section shows a flow to have been 70 percent rubble and only 30 percent liquid during its active stage.

Formation of Massive Vitrophyre. Adjacent to the rubble crusts are thick layers of massive gray vitrophyre and massive yellow felsophyre (altered vitrophyre). Together they form a sheath that encases the red felsophyre core of the flow. Patches of gray vitrophyre within the massive altered vitrophyre resemble rubble crust (Fig. 2), but they are not, because flow laminae or layers of spherulites can be traced across the patches into the adjacent vitrophyre, thus demonstrating that the patches are not rotated blocks.

The massive vitrophyre, fresh and altered, represents lava which must have chilled to a glassy rock only after movement of the lava ceased. Otherwise, the chilled glassy rock would have been broken and become part of the rubble crust. This chilling took place at least in part after the flow margin stopped advancing, because the massive vitrophyre sheath extends to the flow margin (Fig. 5). The chilling to form vitrophyre could

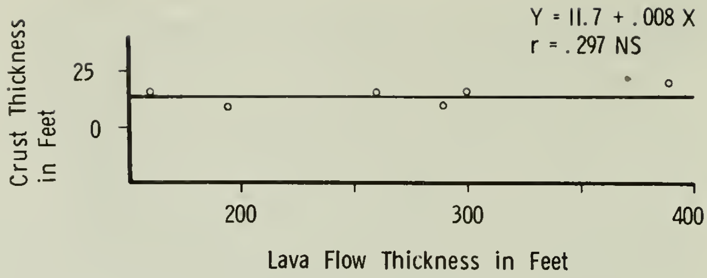


Figure 6. Thickness of upper crust of rubble in six lava flows of Shoshone Volcanics.

have begun in less mobile parts of the flow during the active stage, particularly in portions of the flow near the vent (Fig. 7).

Formation of the Felsophyre Core. An emplaced and cooling lava flow consisted of a molten core surrounded by a vitrophyre sheath (Fig. 7). The layers of massive vitrophyre must have continued to thicken as cooling proceeded, until conditions favored creation of features that characterize the lava flow core, namely, vesicularity, hematite stain, oxyhornblende and oxybiotite, and primary cryptocrystallinity, as contrasted with crystallinity caused by devitrification.

Except for local pumiceous rock in the rubble crust, only the felsophyre core is vesicular. This indicates that the lava, instead of degassing completely when it reached the Earth's surface, retained an appreciable quantity of gas in solution until after the flow was emplaced and had cooled for some time.

Lava flows of basic composition are generally devoid of evidence of delayed vesiculation, probably because they have a lower melt viscosity. The proportion of acid lava flows that display evidence of delayed vesiculation, is unknown, because detailed descriptions of the interiors of acid lava flows are rare. The delayed vesiculation probably is responsible for the other characteristic features of the red felsophyre, based on the following evidence: On heating in air to about 750°C, ordinary hornblende changes to oxyhornblende (Belovsky, 1891; Kozu and others, 1927; Barnes, 1930). This process is one of oxidation and is accompanied by resorption around the crystal margins, the products generally including iron oxides and pyroxene (Deer and others, 1962).

The conversion of ordinary hornblende to oxyhornblende and of biotite to oxybiotite in the flow interior probably occurred with vesiculation, the exsolved gases providing the

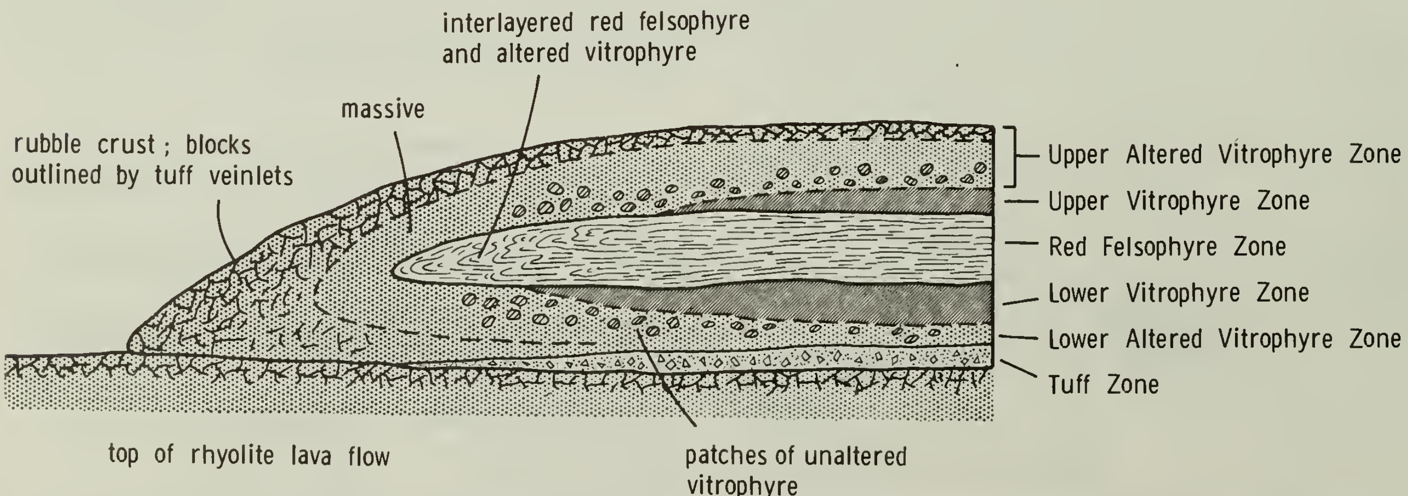


Figure 5. Cross-section sketch of lava flow margin in exposure of Shoshone Volcanics near Miller Spring. Flow margin overlain by another eruptive unit of Shoshone Volcanics. Not drawn to scale; length of section approximately 250 ft.

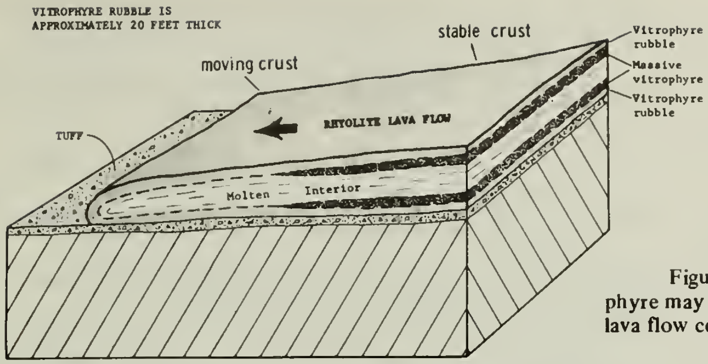


Figure 7. Interpretation of emplacement of an eruptive unit. Massive vitrophyre may have formed in stable crust areas of active lava flow but also formed after lava flow ceased advancing. Not drawn to scale.

necessary oxidizing environment. Because hematite, the colorant of the lava flow core, is concentrated around oxyhornblende and oxybiotite crystals, it probably formed as a resorption product.

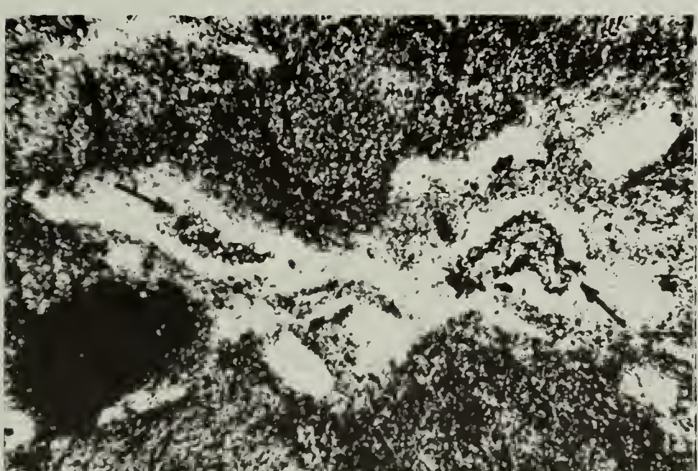
Because oxidation is an exothermic process, additional heat probably was generated in the molten lava flow core upon vesiculation. The oxidation of hornblende, experimentally, suggests that the temperature of the molten core was at least 750°C when the oxidation began. The additional heat provided by the oxidation apparently was sufficient to maintain the vesiculating molten core in a liquid state long enough for ions to migrate through the miscous melt and organize themselves as minute crystals. Thus the cryptocrystallinity of the red felsophyre groundmass also appears to have been favored by the delayed vesiculation.

Vesicle Collapse in the Felsophyre Core. The walls of vesicles are more coarsely crystalline than the rest of the groundmass. Streaks of similar material, observed in thin section, demonstrate the partial to complete collapse of vesicles (Fig. 8). Collapsing may have occurred when exsolving gases became depleted in the melt, or when the falling temperature caused a reduction in gas pressure to the point where it could no longer support the vesicle walls. The relatively large vesicles, which occur at the top and base of the red felsophyre zone, may have become "frozen" in highly viscous lava close to the chilled vitrophyre.

The peculiar paleomagnetic properties of the red felsophyre may be related to vesicle collapse. About 40 percent of determinations on red felsophyre yield magnetic directions that parallel the predicted directions of the rocks. The red felsophyre, even though it has been oxidized by exsolved gases, evidently can be a faithful recorder of the Earth's magnetic field, just as the vitrophyre can. Thus, the remaining determinations that do not yield the predicted magnetic direction seem best explained by mechanical rotation of mineral grains perhaps related to vesicle collapse. No other rotation is apparent as the paleomagnetism of the massive vitrophyre, particularly in the flow margins, demonstrates that the Curie point in the outer part of the lava flow was crossed only after the lava flow had stopped advancing. If this interpretation is correct, then the collapse of vesicles in the red felsophyre took place below 580°C, the measured Curie point of the rock.

Mechanism of Lava Flow Advance. Since freshly fallen tuff is notoriously susceptible to erosion but is conformable with the overlying flows, the lava flows must have moved over the tuff immediately after the tuff was deposited. Therefore, tuff is taken to represent deposits of explosions that cleared the vent just prior to a flow effusion.

The tuff deposits, which were unconsolidated lapilli and ash at the times the lava flows moved over them, are virtually undisturbed. Typically, the tuff fragments that are mixed into the overlying lava flow rock occur within 10 cm of the contact. In addition, the surfaces of the tuff layers apparently are unscoured by the overlying flows. Individual layers at the tops



A

B

Figure 8. A. Photomicrograph of collapsed vesicles in red felsophyre (crossed nichols). Coarsely crystalline walls of vesicles show as light streaks. Arrows point to tridymite, which filled in partially collapsed vesicles. Light streak above tridymite patch on right is completely collapsed vesicle, without tridymite filling. Length of photo, 3 mm. B. Large vesicles in red felsophyre near bottom of zone. Vesicles are lined with druses of silica minerals.

of tuff deposits can be traced for nearly 1 km. Therefore, a typical lava flow advanced without disturbing the underlying loose tuff. The nature of the flow margins and the existence of rubble crust at the base of the typical flow indicate that it advanced with a simple rolling tractor-tread motion, the crust that formed at the upper surface being dragged beneath the flow as it advanced (Fig. 7).

Although the areal extent of individual lava flows is obscured by abundant faults, individual flows have been traced continuously for as much as 5.5 km. A lava flow in the Dublin Hills that contains distinctive fragments of basaltic rock may be correlative with a similar lava flow exposed as far as 19.3 km north of the Dublin Hills.

Considered as a group, the lava flows of the Shoshone Volcanics appear to have filled the valleys. A map of the area near Brown Peak in the Eagle Mountain quadrangle, for example, shows flows accumulated which filled an ancient valley that possessed a relief of about 300 m (Haefner, 1972, Pl. 2).

The tractor-tread motion, the large proportion of liquid lava in the active flows, and the apparent extensiveness of some lava flows suggest a high fluidity for the Shoshone lava flows. This might be an effect of high initial temperature (Gibbon, 1964), but the presence of phenocrysts of the lava suggests a limited temperature range comparable to the range of the relatively viscous lava assumed for the porphyritic flows of block rhyolite. Thus, apparently high fluidity of the Shoshone rhyolite lava flows is more likely an effect of a high content of dissolved volatiles than of a high initial temperature, the volatiles being largely retained in solution during the advance of the flows.

REFERENCES CITED

Armstrong, R. L., 1966, K-Ar dating using neutron activation for Ar analysis: Granitic plutons of the eastern Great Basin, Nevada and Utah: Geochim. et Cosmochim. Acta, v. 30, no. 6, p. 565–600.

Barnes, V. E., 1930, Changes in hornblende at about 800°C: Am. Mineralogist, v. 15, p. 393.

Belovsky, Max, 1891, Über die Änderungen, welche die Optischen Verhältnisse der gemeinen Hornblende beim Glühen erfahren: Neues Jahrb. Mineralogie, Bd. 1, p. 291–292.

Chesterman, Charles W., 1973, Geology of the northeast quarter of Shoshone quadrangle, Inyo County, California: California Div. Mines and Geology, Map Sheet 18.

Curry, H. D., 1941, Mammalian and avian ichnites in Death Valley [abs.]: Geol. Soc. America Bull., v. 52, no. 12, p. 1979.

Deer, W. A., Howie, R. A., and Zussman, J., 1962, Rock-forming minerals, Vol. 2: New York, John Wiley & Sons.

Drewes, Harald, 1963, Geology of the Funeral Peak quadrangle, California, on the east flank of Death Valley: U.S. Geol. Survey Prof. Paper 413, 78 p.

Evernden, J. F., Savage, D. E., Curtis, G. H., and James, G. T., 1964, Potassium-argon dates and the Cenozoic mammalian chronology of North America: Am. Jour. Sci., v. 262, p. 145–198.

Fleck, R. J., 1970, Age and tectonic significance of volcanic rocks, Death Valley area, California: Geol. Soc. America Bull., v. 81, p. 2807–2816.

Gibbon, D. L., 1964, Origin and development of the Star Mountain rhyolite [Ph.D. dissert.]: Houston, Rice Univ., 119 p.

Haefner, Richard, 1969, Emplacement and cooling history of a rhyolite lava flow and related tuff at Deadman Pass, near Death Valley, California [M.S. thesis]: University Park, Pennsylvania State Univ., 82 p.

——— 1972, Igneous history of a rhyolite lava flow series near Death Valley, California [Ph.D. dissert.]: University Park, Pennsylvania State Univ., 281 p.

——— 1973, Alteration of rhyolite near Death Valley, California: A newly recognized phenomenon: Geol. Soc. America, Abs. with Programs (Ann. Mtg.), v. 5, no. 7, p. 647.

Kozu, S., Yoshiki, B., and Kani, K., 1927, Notes on the study of the transformation of common hornblende into basaltic hornblende at 750°C: Tohoku Univ. Sci. Repts., 3d ser., v. 3, no. 2, p. 143–159.

Nasedkin, V. V., 1963, Water-bearing volcanic glasses of acid composition–Their genesis and alteration: Acad. Sci. USSR (text not translated).

Noble, L. F., 1941, Structural features of the Virgin Spring area, Death Valley, California: Geol. Soc. America Bull., v. 52, no. 7, p. 941–1000.

Noble, L. F., and Wright, L. A., 1954, Geology of the central and southern Death Valley region, California, *in* Jahns, R. H., ed., Geology of southern California: California Div. Mines and Geology Bull. 170, p. 143–160.

Stern, T. W., Newell, M. F., and Hunt, C. B., 1966, Uranium-lead and potassium-argon ages of parts of the Amargosa thrust complex, Death Valley, California: U.S. Geol. Survey Prof. Paper 550–B, p. 142–147.

Wright, L. A., and Troxel, B. W., 1971a, Thin-skinned megaslump model of Basin Range structure as applicable to the southwestern Great Basin: Geol. Soc. America, Abs. with Programs (Ann. Mtg.), v. 3, no. 7, p. 758.

——— 1971b, Evidence for tectonic control of volcanism, Death Valley: Geol. Soc. America, Abs. with Programs (Cordilleran Sec.), v. 3, no. 2, p. 221.

High-altitude vertical aerial photograph near southeastern end of Greenwater Range. Pale-colored rocks in left of center of photograph are marine Tertiary volcanic rocks described by Haefner (see preceding article). The rocks also crop out on northwest (left) side of Dublin Hills (right of center). Amargosa River flows south across photograph. Pale-colored area on right is underlain by flat-lying lake sediments deposited during an interval when the flow of the Amargosa River was blocked. *Part of photo U.S. Air Force 374V 197, 6 September 1968; courtesy of the U.S. Geological Survey.*

Geologic Features of the Central Black Mountains, Death Valley, California

James K. Otton[1]

INTRODUCTION

The Black Mountains, which form the eastern margin of the central part of Death Valley, California, contain complex lithologic and structural features that have been difficult to integrate with the geologic framework of the surrounding region. They are underlain mostly by a crystalline complex composed of folded pre-Cenozoic metasedimentary units, in part marble bearing, and of intrusive igneous rocks of Precambrian, Mesozoic(?), and Cenozoic age. Especially abundant in the complex is diorite that postdates the sediments, now metamorphosed, and predates other intrusive bodies that are more acidic and largely to wholly Cenozoic in age. Overlying the complex, generally with fault contact, are structural units composed variously of slightly metamorphosed Precambrian and Paleozoic sedimentary rocks and Cenozoic sedimentary and volcanic rocks. These constitute a relatively thin, discontinuous, and faulted cover of the complex.

The following is a report on the progress of a mapping project in the southwestern Black Mountains. The area (Fig. 1) includes the southwestern quarter of the Funeral Peak and the southeastern part of the Bennetts Well 15′ quadrangles. In this area, the igneous and metamorphic complex—a dislocated cover of Cenozoic rocks—and two of the peculiar domical features known as turtlebacks are well exposed. Although many questions concerning the geologic history of the Black Mountains remain unanswered, the mapping to date has documented several features relevant to this history and the evolution of the region.

The general geologic features of the Black Mountains have been shown on several published maps, namely those of Curry (1954), Noble and Wright (1954—generalization of Curry's map), Drewes (1963), the Death Valley sheet of the State Geologic Map of California (scale 1:250,000), and Wright and Troxel (1973). Curry (1954) outlined the rock units and structural features along the western margin of the north and central Black Mountains and was the first to describe the turtleback surfaces. He concluded that they were parts of a folded and eroded thrust fault, and he correlated the fault with the Amargosa thrust fault that Noble (1941) had identified in the southern Black Mountains.

Drewes, working in the Funeral Peak quadrangle and in an adjacent part of the Bennetts Well quadrangle (Drewes, 1963), also described the rock units and structural features; he concluded that the turtleback surfaces and the Amargosa thrust were separate features. He favored the hypothesis that the turtlebacks were produced by differential erosion during Tertiary time of a terrane of folded metamorphic rocks, producing an undulating surface on which late Tertiary rocks were deposited and from which they slid along normal faults.

The geologic features of the area in Figure 1 present numerous avenues of investigation, including the following: (1) the stratigraphy and age of the sedimentary rocks now metamorphosed; (2) their metamorphic history and the role of the various intrusive events in the metamorphism; (3) the nature and time of folding of the metamorphic rocks; (4) the age and mode of emplacement of the diorite and the bodies of younger igneous rock; and (5) the evolution of the fault pattern, including the faulted surfaces between the complex and the bodies of rock that overlie it.

LITHOLOGIC AND STRUCTURAL FEATURES OF THE METASEDIMENTARY ROCKS

The metasedimentary rocks of the crystalline complex can be subdivided into two gross units. The stratigraphically lower unit is composed of augen gneiss, biotite schist, amphibolite, biotite-hornblende gneiss, quartz-feldspar gneiss, and gneissic granite. This unit lithologically resembles the older Precambrian complex extensively exposed in the more southerly parts of the Black Mountains and is here correlated with it. The stratigraphically higher unit is carbonate bearing and consists variously of massive light-brown to gray dolomitic marble; well-bedded tan stromatolite-bearing dolomitic marble (Fig. 3) with cherty lenses and nodules; tan dolomitic marble containing thin layers of hornfels; chert; quartzite; and light-gray limestone. It displays a maximum thickness of about 600 ft.

Wherever the two units are in contact, the carbonate-bearing unit overlies the generally gneissic unit. In addition to their lithologic differences and stratigraphic relations, differences in style of deformation and metamorphism are distinctive. The gneissic unit is characterized by variously oriented mesoscopic, isoclinal, locally recumbent folds and by textures attributable to shearing. The carbonate unit, on the other hand, is characterized by upright slightly overturned macroscopic disharmonic folds that consistently plunge northwestward. The older unit is pervasively metamorphosed to lower amphibolite or upper greenschist facies. The metamorphism of the carbonate-bearing unit ranges widely in intensity and apparently is related to distance from contacts with the diorite. At localities farthest from diorite contacts, chert bodies are devoid of prograde metamorphic minerals and of evidence of recrystallization. Chert nodules in exposures of the carbonate-bearing unit close to contacts with diorite are altered to talc, epidote, and (or) garnet and have been recrystallized. Finally, the contact between the carbonate-bearing unit and the older complex is discordant with foliation in the complex.

Between Copper Canyon and Sheep Canyon (Fig. 1), the older Precambrian schist and gneiss form an elongate dome that is mantled by a thin, discontinuous sheath of the carbonate-bearing unit. The morphology of this dome is displayed in a structure contour map of the contact between the gneissic unit and the carbonate-bearing unit (Fig. 2). The dome is steepest on its southwestern flank, which is marked by a broad fault zone containing abundant large fragments of carbonate rock. Diorite lies on the southwestern downdropped side of this fault. The type turtleback (Curry, 1954) is located on the northwestern nose of this dome. Geometrically, this structure resembles a mantled gneiss dome (Eskola, 1949), in that it consists of a core of gneiss that underlies a cover of younger, structurally discordant metamorphosed rocks.

The older Precambrian rocks mapped in the Mormon Point area are also mantled by a sheath of carbonate-bearing rocks. The sequence preserved there is thicker than on the other dome and obscures most of the older Precambrian units. The exposures of carbonate-bearing rocks terminate, however, along

---

[1] *Pennsylvania State University, University Park, Pennsylvania 16802.*

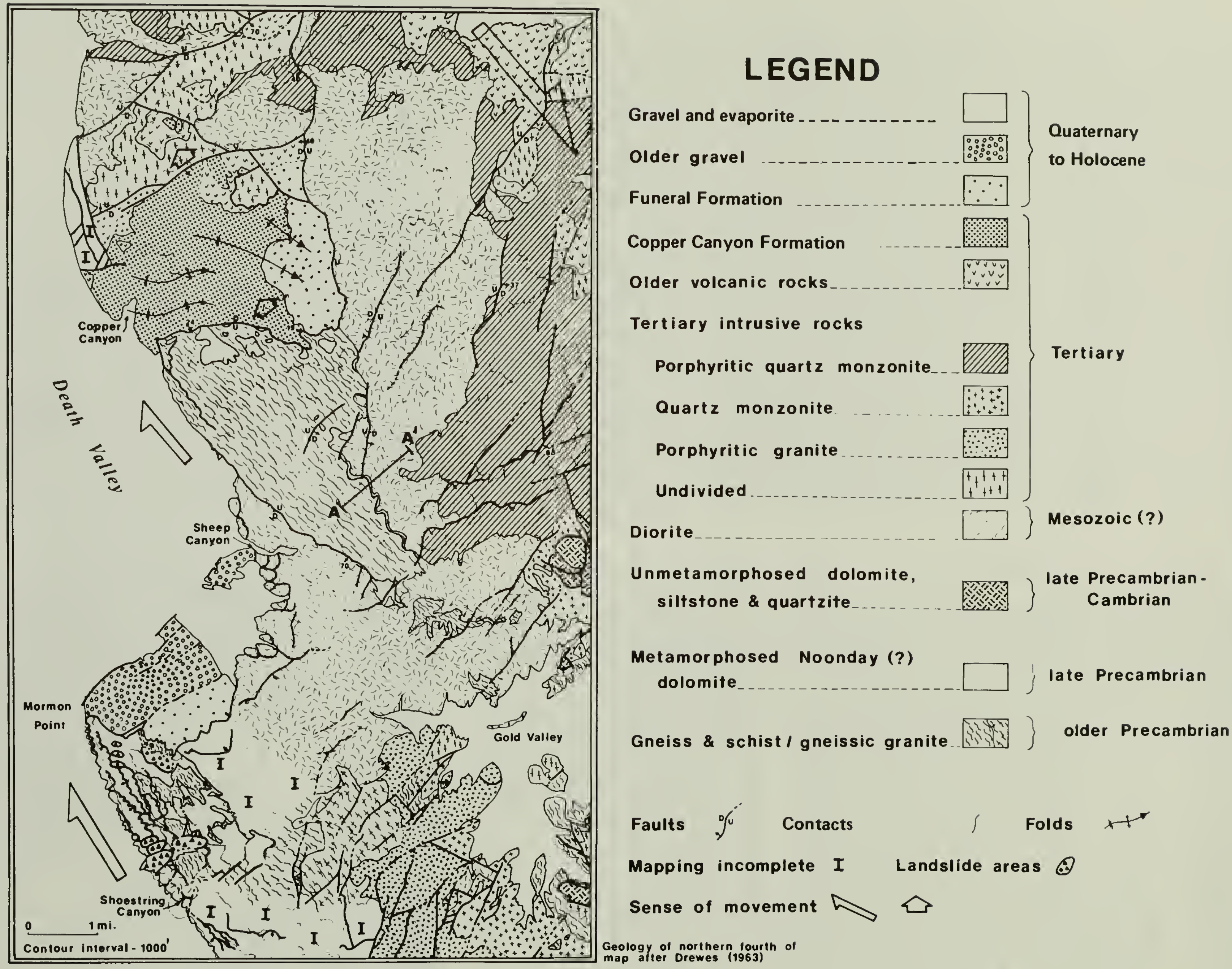


Figure 1. Generalized geologic map, central Black Mountains, Death Valley, California.

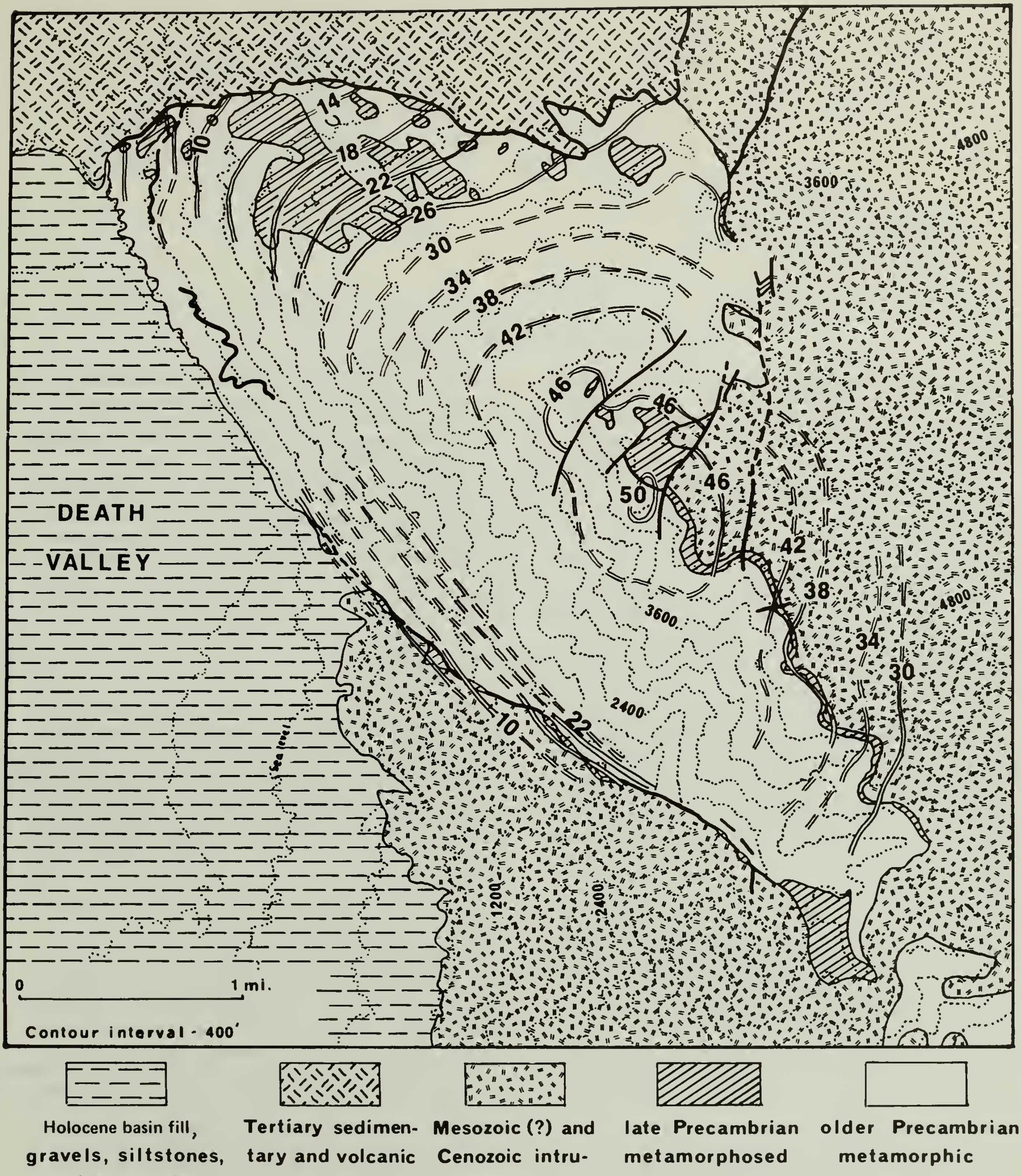


Figure 2. Generalized geologic map of a part of the Black Mountains showing structural contours on contact between older Precambrian and late Precambrian units.

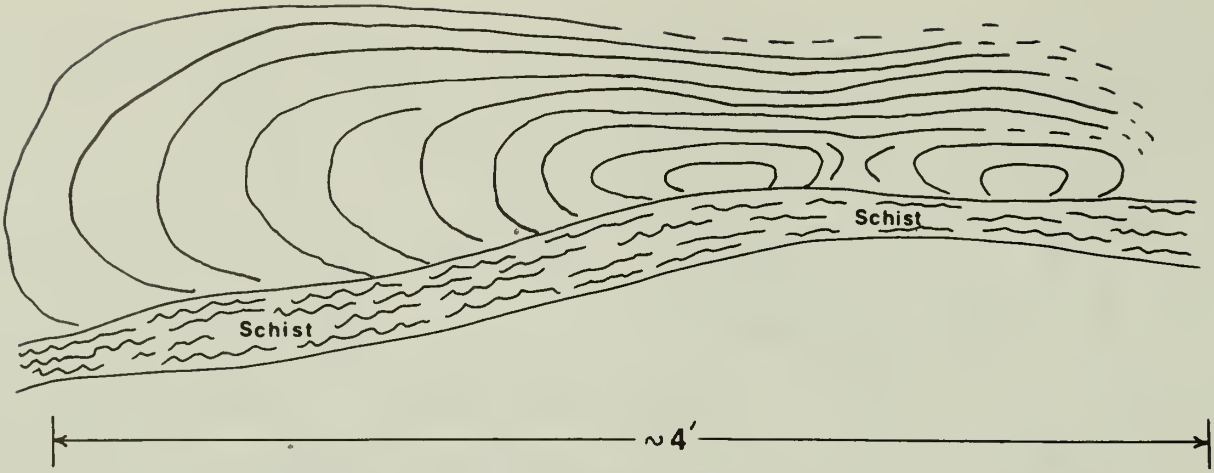


Figure 3. Sketch of stromatolites in carbonate, Mormon Point.

a reverse fault on the south side of Shoestring Canyon (Fig. 1). Although there are extensive exposures of older Precambrian schist and gneiss south of this fault, no mantling carbonate-bearing sequence is exposed. The unconformity beneath the carbonate-bearing rocks is folded in the Mormon Point area, but the shape of that surface cannot be completely reconstructed. Minimum relief on the surface is 1,200 ft.

CORRELATION OF THE MARBLE WITH OTHER CARBONATE UNITS OF THE REGION

If the diorite is Precambrian in age and is older than later Precambrian Pahrump Group and younger rocks that are extensively exposed in the southern part of the Black Mountains (Drewes, 1963), then the marble must be older than the Pahrump Group and composes a unit unexposed in the pre-Pahrump terrane of the Death Valley region.

If the diorite is Mesozoic in age, the marble may be correlated: (1) with carbonate units of the Pahrump, (2) with the Noonday Dolomite and (or) the Johnnie Formation, (3) with other carbonate units of later Precambrian or Paleozoic age, or (4) with carbonate units of Triassic age.

Correlation with the Noonday Dolomite or the Johnnie Formation is compatible with the paleogeographic reconstructions of Wright and Troxel (1966, 1967), with the lithologic descriptions of the Noonday by Williams and others (1974) and Williams (1974, personal communication) and with the lithologic descriptions of the Johnnie Formation by Benmore (1974). In wright and Troxel's reconstructions, the Mormon Point and Copper Canyon–Sheep Canyon areas are part of the Nopah Upland, an area from which Pahrump Group sedimentary rocks were eroded in late Pahrump and pre-Noonday time. It is thought that parts of the Noonday Dolomite and the Johnnie Formation then transgressed and possibly blanketed the post-Pahrump terrane (Fig. 5). If so, then these carbonate units could be rocks of the Noonday Dolomite or the Johnnie Formation metamorphosed near the diorite body, if it is younger than the Noonday.

In general aspects, much of the carbonate sequence of Mormon Point resembles the shelf facies of the Noonday Dolomite described by Williams and others (1974). The massive carbonate rocks and the thinly-bedded carbonate rock with hornfels lenses may be the metamorphosed equivalent of massive domal carbonate units and thinly-bedded carbonate units with cryptalgal structures in the Noonday shelf facies. Moreover, sharp lateral thickness variations noted by Williams and others (1974) also occur in carbonate units at Mormon Point.

Parts of the sequence contain rocks similar to parts of the Johnnie Formation. This includes a ferroan dolomite and cyclic units consisting of clastic carbonate and fine-grained siliceous material. The clastic carbonate is cross-bedded (cross-beds 6 in. to 18 in. high) and exhibits a bimodal rosette that suggests a west-dipping paleoslope and a tidal environment.

Correlation with carbonate units in the Pahrump Group (possibly preserved as in Fig. 4), namely a carbonate member in the Crystal Spring Formation (Roberts, this volume) or with the overlying Beck Spring Dolomite also is a possibility, however, the marble of the Black Mountains is unlike the carbonate units of the Pahrump or any of the younger units in the late Precambrian, Paleozoic, and Mesozoic terranes of the Death Valley region.

DIORITE BODIES

The pluton or plutons of diorite that intrude the metamorphic rocks of the Copper Canyon–Sheep Canyon and Mormon Point–Smith Mountain areas extend discontinuously from the southwestern flank of Smith Mountain to Badwater, 6 mi north of the present map area, a distance of 15 mi (Drewes, 1963, Pl. 1). It comprises more than half the mapped crystalline rocks of the central Black Mountains. The original shape of the diorite pluton(s) is difficult to document because many of its contacts are faulted or are intrusive contacts with later granitic bodies. Near the Copper Canyon turtleback, however, the diorite occurs as a sill that follows closely the contact between the two units of metamorphic rocks (Fig. 6). In the area of Mormon Point, the intrusive relations are less simple.

Diorite contacts in the Copper Canyon–Sheep Canyon area are sharp and can be located within a few feet. Near Mormon Point the contacts are gradational over a few hundred to several hundred feet. Foliation in the older Precambrian gneiss becomes indistinct, and the gneiss commonly develops feldspar porphyroblasts near the diorite contact.

Xenoliths of metamorphic rocks in the diorite are common particularly in a zone that extends southeast from the central part of Sheep Canyon into Gold Valley. The xenoliths consist mostly of the older Precambrian gneiss, but carbonate quartzite and schist form xenoliths on the north side of Gold Valley.

The petrology of the diorite, as described by Drewes (1963, p. 8–9), consists mostly of hornblende and plagioclase with minor or accessory proportions of biotite, clinopyroxene, sphene, titaniferous magnetite, and apatite. The texture is hypidiomorphic granular in most occurrences. In others, hornblende exhibits ophitic or poikilitic textures. Plagioclase is commonly poikilitic. Drewes stated that some of the hornblende may represent secondary growth from pyroxene and that some of the plagioclase may also be secondary. He added (1963, p. 9–10):

> . . . This replacement is the chief metamorphic process that has affected the original diorite or gabbro; the metamorphism could have occurred at the same temperature and pressure as the metamorphism of the (older Precambrian) metasedimentary rocks.
>
> The metadiorite is younger than the metasedimentary rocks, and it is possibly also younger than the folding of the metasedimentary rocks, for it is virtually unfoliated and it intrudes and appears to truncate the foliated rocks. In degree of metamorphism it has a far greater resemblance to the older Precambrian rocks of the Funeral Peak quadrangle than to the Pahrump series of late Precambrian age; therefore the metadiorite is tentatively assigned to early Precambrian.

The textures that Drewes attributed to metamorphism could also be ascribed to deuteric alteration of the intrusive body during cooling. Reliable radiometric dates have yet to be obtained for the diorite. However, it resembles diorite that is clearly Mesozoic in age in other bodies in the Death Valley region. These occur in the Avawatz Mountains (B. W. Troxel and L. A. Wright, personal commun.) and the southern part of the Panamint Range (Johnson, 1957). A post-Precambrian age for the diorite is also suggested by the observation that it extends south into an area where the Precambrian gneiss is cut by numerous diabase dikes that were emplaced in Pahrump time. The diorite there is not intruded by diabase, although none of the diorite contacts has been observed to actually show crosscutting relationships with the diabase. Moreover, unlike the metasedimentary units, the diorite is devoid of pervasive foliation. That it postdates the folding of the metasedimentary rocks is indicated by mafic dikes that extend into the folded rocks from the pluton(s) of diorite, without evidence of folding. The diorite crosscuts faults which, in turn, cut exposures of the folded carbonate sequence. Thus, the diorite is believed to represent a much later event related to the Mesozoic plutonism so common in the region west and south of Death Valley.

SILICIC PLUTONS AND DIKES

The igneous bodies that intrude the diorite and the metasedimentary units range in composition from granite to monzonite. The mappable igneous bodies comprise a pink porphyritic granite, two types of quartz monzonite, three types of porphyritic quartz monzonite or latite, red porphyritic rhyolite, and andesite dikes.

To completely reconstruct the sequence of intrusive events through cross-cutting relationships is difficult. The pink porphyritic granite and the porphyritic quartz monzonite intrude the diorite but not each other. A gray phase of the quartz monzonite clearly intrudes the pink porphyritic granite, as demonstrated by xenoliths of the latter in the monzonite A white phase of the quartz monzonite intrudes the diorite, but its contact with the porphyritic quartz monzonite north of Sheep Canyon is unexposed.

The red porphyritic rhyolite that forms plugs and the andesite that occurs as dikes probably represent the latest events. The andesite dikes are abundant on the south side of Gold Valley and Smith Mountain where they intrude the gray phase of the quartz monzonite. One andesitic plug intrudes the porphyritic quartz monzonite near the northeastern corner of the area. The red porphyritic rhyolite intrudes both the porphyritic quartz monzonite and the quartz monzonite at various localities.

The intrusive contacts between these various igneous bodies and their host rocks are commonly gradational over distances of hundreds of feet or more (Fig. 7).

The larger plutons commonly are bordered by an outer zone where only a few dikes penetrate the host rock, an intermediate zone where the number of dikes increases to form an anastomosing network in which the host rock "floats" as xenoliths, and finally an inner zone where the number, size, and volume of xenoliths approaches zero. The anastomosing dikes are largely horizontal, and xenoliths commonly maintain a tabular, subhorizontal dimensional orientation. On the other hand, many contacts are sharp. The pink porphyritic granite exhibits a sharp subhorizontal contact with the older Precambrian rocks on the southwestern flank of Smith Mountain.

Dike swarms and dike systems that are compositionally related to the silicic plutons extend beyond the margins of those intrusive bodies. Many of the dike systems show no systematic orientation, while others show a strong preferred orientation. One system of dikes occupies the crest of a ridge that lies between Sheep Canyon and Copper Canyon. This dike system consists of gray porphyritic quartz monzonite similar to one of the porphyritic intrusive phases a few miles to the east. The dikes are remarkably planar; they are all subvertical, and all but two strike approximately N. 30° E. They are spaced 200 to 800 yd apart, and their thicknesses reach 75 ft and

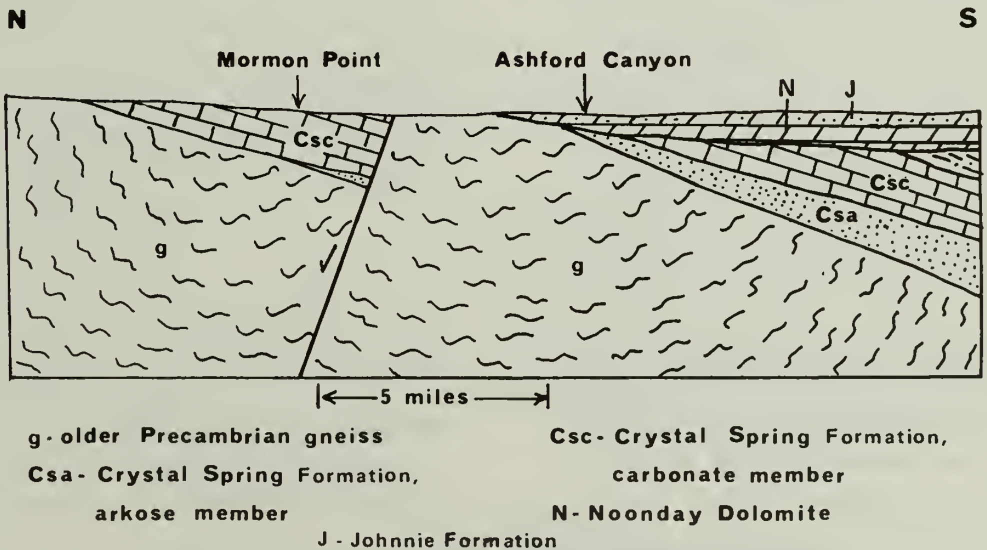


Figure 4. Cross section of hypothetical late Johnnie Formation Black Mountain terrane showing preservation of carbonate member of Crystal Spring Formation adjacent to a fault.

engths over 2 mi. This planarly oriented system of dikes is restricted to the older Precambrian rocks of the ridge. The dikes terminate to the southwest against the range front fault and to the northeast against the turtleback fault. Some of the dikes cross the intrusive contact into the diorite, but where they do, they change orientation.

A second system of dikes occupies the northerly and northwesterly faces of a ridge of diorite just to the south of the mouth of Sheep Canyon. Dikes in this system are essentially planar but interfinger more than the dikes of the previously described system. They strike N. 75° E. and dip 35° to 45° to the southeast. These dikes also consist of porphyritic quartz monzonite. They are more closely spaced than the other system—as close as 100 ft or less.

Most of the silicic dikes and all of the silicic plutons of the region lack a visible foliation. One swarm of dikes, composed of gray aphanitic to slightly porphyritic rock, shows well-developed layering. Layers are distinguished by varying shades of gray, or gray and reddish gray. Layer thickness varies from a few millimeters to a few centimeters. The layering is interpreted as flow layering because isoclinal folds symmetric with dike margins are common in the layers. In larger dikes, the central portions commonly are unlayered.

The age of these igneous bodies is uncertain. Drewes (1963) reported two lead-alpha age determinations on zircon in the monzonitic rocks, 45 ± 10 and 30 ± 10 m.y. These determinations, made in 1959, are probably too old by a large factor (Rose and Stern, 1960). Fleck (1970) reported K-Ar whole-rock age dates on the older volcanic rocks of the region of 6 to 8 m.y. Because these units were deposited on or were involved in faulting that cuts the monzonitic rocks, these dates probably can be considered a minimum age for the monzonites. The time of intrusion for these rocks seems best placed between 10 and 15 m.y. ago.

TERTIARY AND QUATERNARY SEDIMENTARY AND VOLCANIC ROCKS

Drewes (1963) described in detail the lithology of the Tertiary and Quaternary rocks of the region, the older volcanic rocks, the Copper Canyon Formation, the Funeral Formation, and gravel of various ages. The older volcanic rocks include rhyolite and rhyodacite flows and tuff with interbedded tuff-breccia and agglomerate and minor volcanigenic sediments. The Copper Canyon Formation comprises over 10,000 ft of reddish conglomerate, yellowish-gray siltstone, and evaporites and intercalated basalt. Based on mammalian fossils, it is thought to be Pliocene in age (Curry, 1941). The Funeral Formation consists of tan conglomerate and diorite megabreccia.

STRUCTURAL FEATURES

The structural features of the central Black Mountains appear to have developed in two or more stages. The latest stage has produced faults which are compatible with a northwesterly crustal extension of the Black Mountains–central Death Valley terrane. The earliest stages involved the development of the major structural features of the crystalline terrane before extension began. This earlier deformation and intrusion resulted in a tectonic fabric that controlled the faulting during the northwest crustal extension of the later stage. The silicic plutons may mark the beginning of the extensional stage.

The features of the latest stage include many northeast-trending intermediate to high-angle normal faults with displacements generally within the range of a few tens of feet to a few hundred feet. The features of the latest stage mainly consist of faults that are either clearly normal or seem best interpreted as normal in sense of movement. The most abundant of the well-defined normal faults strike northeast and ordinarily show displacements measurable in tens or hundreds of feet; some of the displacements are 1,000 ft or more. One of the larger of the northeast-trending faults bounds the western side of the depression that contains the Copper Canyon Formation (Fig. 1). The faults that derive their configuration from earlier structural features and that are also interpreted here as being normal include the turtleback faults at Mormon Point and Copper Canyon and extensions of the turtleback faults. Linear features, including slickensides, fault mullions, and drag folds, on the surfaces of these faults indicate that the downward movement of the hanging wall has been in a northwesterly direction.

Apparently related to the normal faults are two high-angle faults that strike northwest and can be traced for 5 mi or more (Fig. 1). The linear features of movement on the planes of these faults indicate movements in a normal sense and also in a right-lateral oblique-slip sense. One of these faults terminates along strike in the lower part of Sheep Canyon.

The nature and distribution of the faults of the Black Mountains escarpment support the long-held view that the Black Mountains front is essentially an irregular fault scarp developed during the latest stage of deformation. It consists of

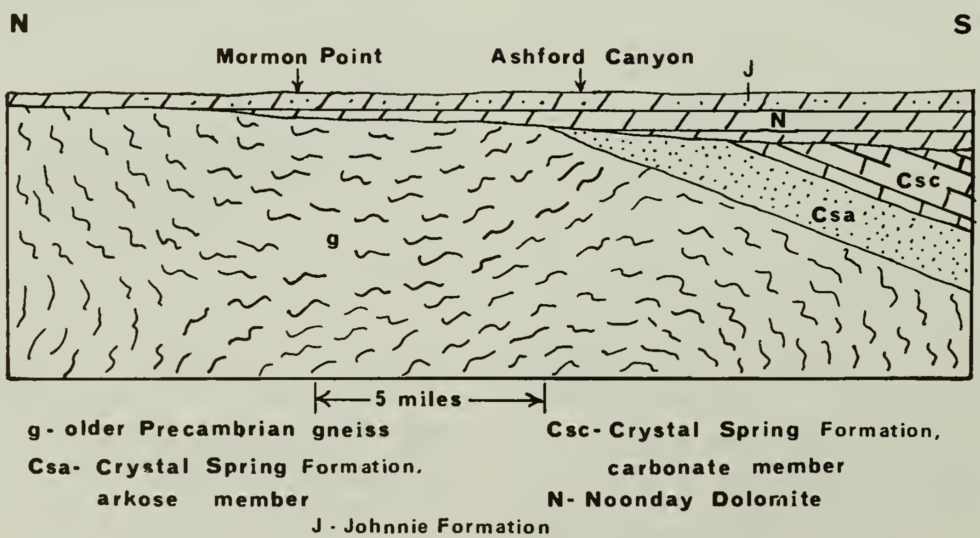


Figure 5. Cross section of hypothetical late Johnnie Formation Black Mountain terrane showing onlap of Noonday Dolomite, in turn overlain by Johnnie Formation.

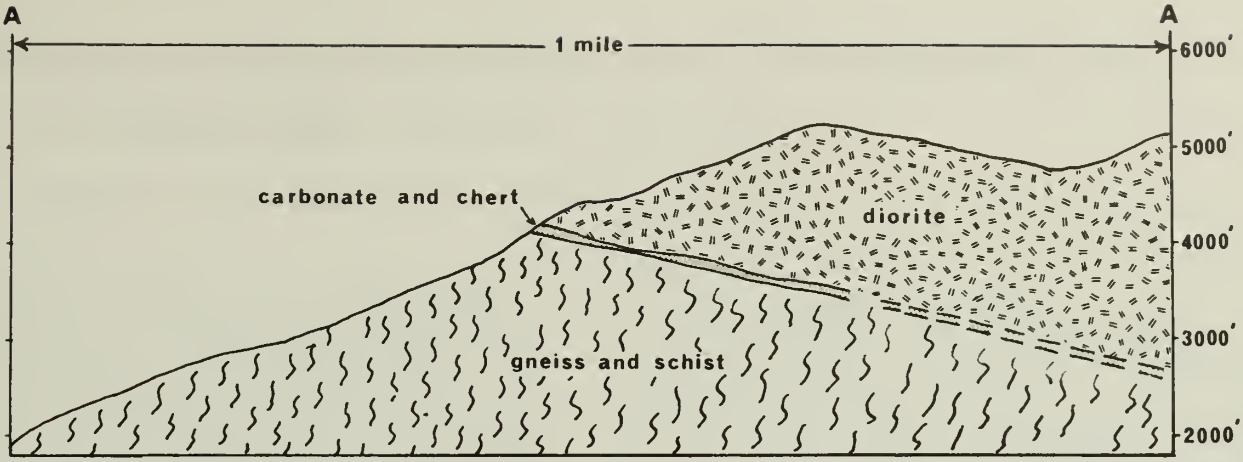


Figure 6. Cross section of ridge between Sheep Canyon and Copper Canyon showing diorite contact.

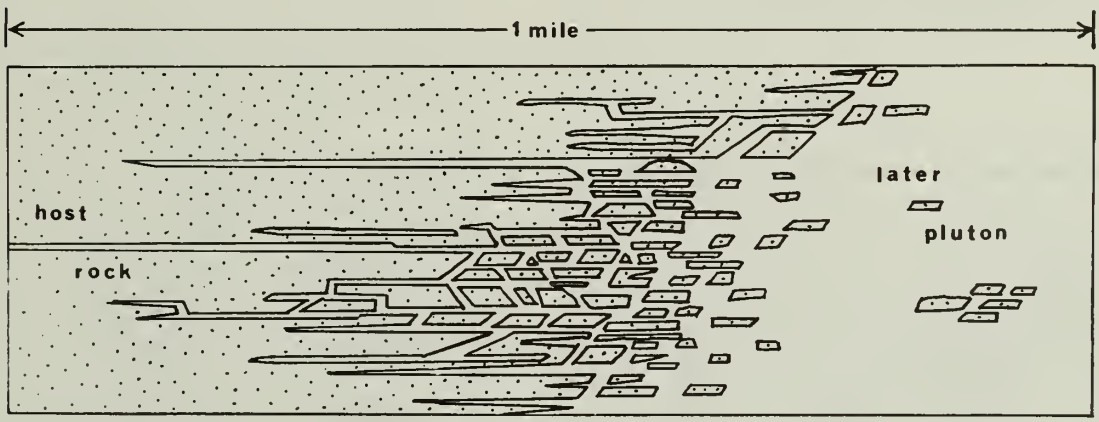


Figure 7. Schematic cross section of gradational contact between silicic plutons and host rock.

northwest-trending en echelon faults with linking northeasterly irregular zones. The irregular zones, however, contrary to previously expressed interpretations, are also related to crustal extension. They coincide geometrically with weak horizons (the carbonate-bearing units) in the early fabric of the crystalline rocks. This suggests that the orientation and position of these zones was controlled by zones of weakness in the inherited fabric.

This view of the Black Mountains front has recently been expressed by Wright and others (1974, reprinted in this volume) and is compatible with the pull-apart model (Fig. 8A) for the origin of the central segment of Death Valley (Burchfiel and Stewart, 1966). Figure 8B is a modification of that model and shows the structural features of the east wall of the central part of Death Valley in greater detail.

REFERENCES CITED

Burchfiel, B. C., and Stewart, J. H., 1966, "Pull-apart" origin of the central segment of Death Valley, California: Geol. Soc. America Bull., v. 77, p. 439–442.

Curry, H. D., 1941, Mammalian and avian ichnites in Death Valley [abs.]: Geol. Soc. America Bull., v. 52, no. 12, p. 1979.

——— 1954, Turtlebacks in the central Black Mountains, Death Valley, California, *in* Jahns, R. H., ed., Geology of southern California: California Div. Mines and Geology Bull. 170, chap. IV, contr. 7.

Drewes, H., 1959, Turtleback faults of Death Valley, California: A reinterpretation: Geol. Soc. America Bull., v. 70, p. 1497–1508.

——— 1963, Geology of the Funeral Peak quadrangle, California, on the east flank of Death Valley: U.S. Geol. Survey Prof. Paper 413, 78 p.

Eskola, P. E., 1949, The problem of mantled gneiss domes: Geol. Soc. London Quart. Jour., v. 104, p. 461–476.

Fleck, R. J., 1970, Age and tectonic significance of volcanic rocks, Death Valley area, California: Geol. Soc. America Bull., v. 81, p. 2807–2816.

Johnson, B., 1957, Geology of a part of the Manly Peak quadrangle, southern Panamint Range, California: Calif. Univ. Pubs. Geol. Sci., v. 30, no. 5, p. iv, 353–423.

Noble, L. F., 1941, Structural features of the Virgin Spring area, Death Valley, California: Geol. Soc. America Bull., v. 52, p. 941–1000.

Noble, L. F., and Wright, L. A., 1954, Geology of the central and southern Death Valley region, California, *in* Jahns, R. H., ed., Geology of southern California: California Div. Mines and Geology Bull. 170, chap. II, contr. 4.

Rose, H. J., Jr., and Stern, T. W., 1960, Spectrochemical determination of lead in zircon for lead-alpha age measurements: Am. Mineralogist, v. 45, p. 1243–1256.

Williams, E. G., Wright, L. A., and Troxel, B. W., 1974, Depositional environments of the late Precambrian Noonday Dolomite, southern Death Valley region, California: Geol. Soc. America, Abs. with Programs (Cordilleran Sec.), v. 6, no. 3, p. 276.

Wright, L. A., and Troxel, B. W., 1966, Strata of late Precambrian-Cambrian age, Death Valley region, California-Nevada: Am. Assoc. Petroleum Geologists Bull., v. 50, no. 5, p. 846–857.

——— 1967, Limitations on right-lateral, strike-slip displacement, Death Valley and Furnace Creek fault zones, California: Geol. Soc. America Bull., v. 78, p. 733–750.

——— 1973, Shallow-fault interpretation of Basin and Range structure, southwestern Great Basin, *in* De Joong, R., and Scholten, R., eds., Gravity and tectonics: Amsterdam, Elsevier Pub. Co.

Wright, L. A., Otton, J. K., and Troxel, B. W., 1974, Turtleback surfaces of Death Valley viewed as phenomena of extensional tectonics: Geology, v. 2, no. 2, p. 53–54.

This research was supported by NSF Grant GA-16119, The Geological Society of America, the P. D. Krynine Memorial Fund at Pennsylvania State University, and the Death Valley Natural History Association.

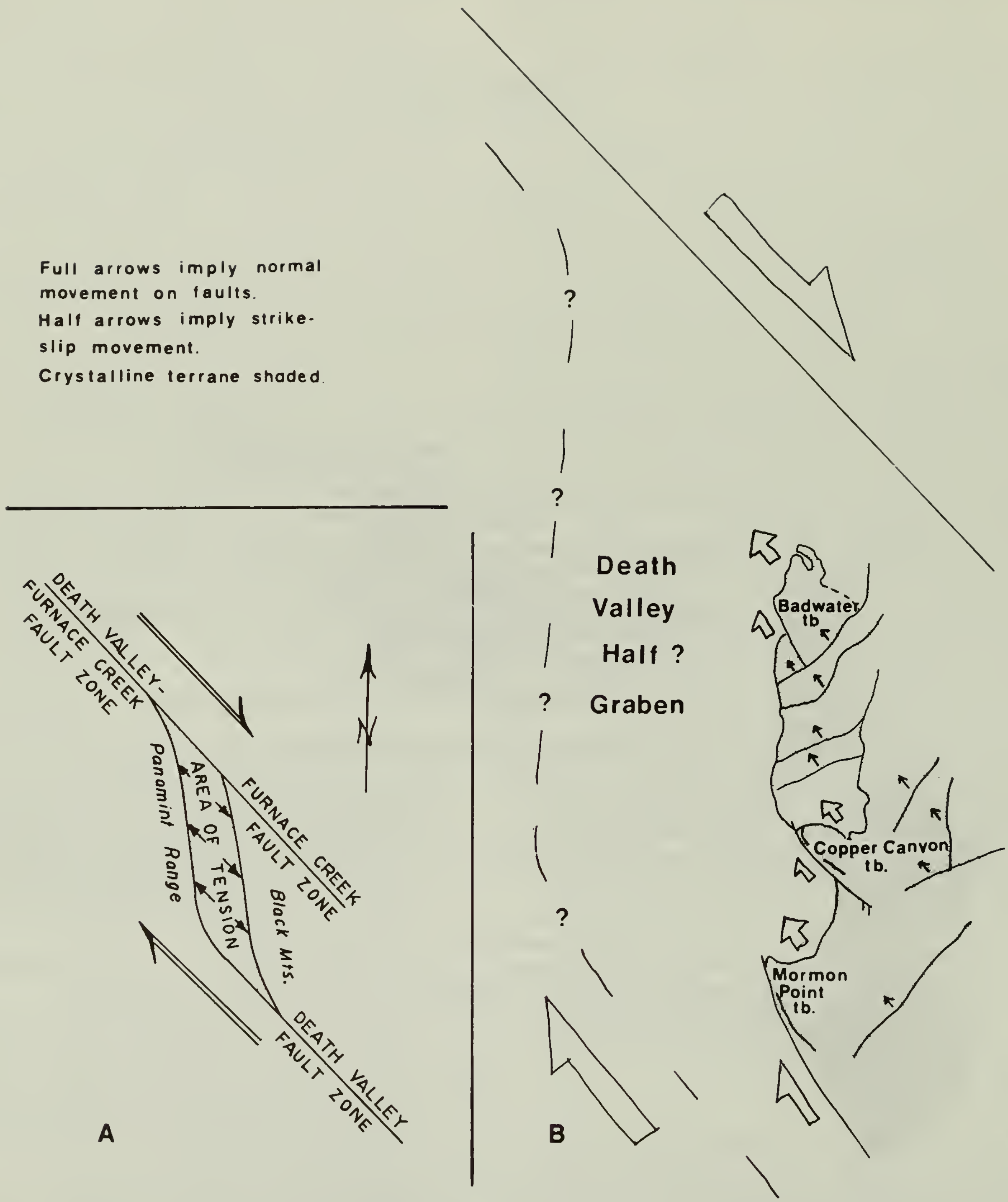


Figure 8. A. Diagrammatic map showing interpretation of strike-slip movement and area of tension (from Burchfiel and Stewart, 1966). B. Detailed map of east wall of area of tension (generalized from Death Valley sheet, Geologic Map of California).

The Noonday Dolomite and Equivalent Stratigraphic Units, Southern Death Valley Region, California

Eugene G. Williams,[1] Lauren A. Wright,[1] Bennie W. Troxel[2]

INTRODUCTION

The Noonday Dolomite of the Death Valley region of eastern California is generally assigned a Precambrian age because it lies as much as 1,200 m stratigraphically beneath the lowest beds that have yielded Lower Cambrian fossils. It was defined by Hazzard (1937) from exposures near the Noonday mine at the southern end of the Nopah Range. It rests upon various units of the late Precambrian Pahrump Group and also upon a pre-Pahrump complex of crystalline rocks that has yielded radiometric dates as old as 1.7 b.y. Most of the exposures of the Noonday, including those described herein, lie within an elongate area, about 50 km wide, that extends from the southern part of the Panamint Range east-southeastward for about 130 km. The Noonday also has been traced discontinuously northward in the Panamint Range to the vicinity of Tucki Mountain (Hunt and Mabey, 1966), but these more northerly occurrences remain relatively uninvestigated.

In the northern and eastern part of the east-southeast–trending area of exposure, the Noonday consists almost entirely of dolomite. As it contains abundant structures of algal and cryptalgal origin, it qualifies as a shallow-water platform deposit. It passes southward and very abruptly, however, into a deeper water facies of clastic carbonate, breccia, siliceous sandstone, and shale. Because these lithologies are totally unlike those of the Noonday Dolomite as originally defined, this facies warrants and will eventually be given a new formational name. But for the purpose of this report, it will simply be referred to as "the basin facies" and will be divided informally into four members.

In the east-southeast–trending belt of exposure, and for an undetermined distance northward in the Panamint Range, the Noonday Dolomite or "platform facies" comprises two well-defined members. Each has been previously noted and mapped (Wright and Troxel, 1966, 1967; Wright, 1973) but has remained unstudied in detail. These two members will be referred to as the lower and upper dolomite members or as the platform dolomite members. In most of the present exposures of the Noonday Dolomite, the two members display a combined thickness within the range of 60 to 400 m. They terminate southward, generally within a distance of several tens of meters, against units of the basin facies in the manner shown in Figures 2, 3, and 5.

LOWER DOLOMITE MEMBER

The lower algal dolomite is very finely crystalline and laminated. It is pinkish gray to grayish orange on fresh surfaces and weathers to very light gray or grayish orange cliff-forming exposures. The laminations, which commonly are obscured by minute fractures, consist of alternating light and dark layers. Discontinuous accumulations of banded cherty dolomite are common in the upper part of the member. Elsewhere the lower dolomite contains less than 1 percent of insoluble material.

Specimens of the lower dolomite member, as observed in thin section, consist of dark cryptocrystalline dolomite and dolomite spar in varying proportions, and exhibit a spongelike texture. The dark material commonly comprises spherical bodies distributed in a cement of dolomite spar. Variations in the percentages of these two materials account for the laminations.

At many localities, and especially those within 2 km of the platform-basin facies change, the lower dolomite member forms concentrically banded mounds as much as 200 m high and 600 m long (see sketch by Wright and others, 1974b, this volume). These can be interpreted only as algal growth structures and will be described in detail in later publications. The observation that the mound-forming dolomite is identical in physical and chemical characteristics with the more evenly structured occurrences of the member, together with the presence in the latter of smaller scale structures that are apparently of algal origin, has led us to view the entire member as having formed in the presence of algae.

UPPER DOLOMITE MEMBER

The between-the-mound low areas in the paleotopographic depressions outlined by the top of the lower algal dolomite are filled with onlapping younger strata in sharp contact with the underlying dolomite. These compose the lowest part of the upper member and ordinarily consist of dolomite with wavy laminations. Some of the larger low areas, however, are filled with deposits of very evenly laminated rock, composed of interbedded finely crystalline clastic dolomite and dolomitic siltstone, which grades upward into the wavy, laminated dolomite. The clastic dolomite and siltstone are generally dark gray on fresh surfaces, but weather to various shades of red, orange, and yellow. Even where the mounds are absent or only a few feet in relief, such strata commonly directly overlie the lower algal dolomite.

The upper dolomite member, apart from the discontinuous bodies of evenly laminated rock, is generally from 45 to 90 m thick. It is a fine-to-medium crystalline pale gray dolomite, which weathers to light brownish gray and persistently contains from 2 to 10 percent of insoluble residues, consisting of chert, clay, and silt-size quartz. The upper dolomite is less resistant than the underlying member. The lower half consists of internally laminated beds, 2 to 4 cm thick, and commonly displays wavy structures that are generally 10 to 20 cm in amplitude and cuspate to undulating in cross section. Some are clearly cross-laminations and others are slump folds, but most seem best interpreted as laterally linked algal stromatolites.

The upper half of the upper dolomite member consists mostly of interbeds, 20 to 40 cm thick, of dolomite of two types; (1) massive, medium crystalline, and (2) laminated and banded finely crystalline. Distributed through this unit, although more abundant near the top, are moundlike structures with amplitudes of 20 cm to 5 m. Some of these are clearly algal stromatolites of hemispherical form. Other structures are less regular and commonly are surrounded by dolomite clasts. We view these as cryptalgal, yet recognize the possibility that they may be caused by slumping or differential loading.

This mound-forming dolomite high in the upper member is overlain by and intertongued with an intensely cross-laminated dolomitic quartzite unit, which marks a transition into the overlying Johnnie Formation. In the 15 to 30 m of sandstone in which this gradational contact is ordinarily recorded, and in the sandstone that intertongues with the upper dolomite, the quartz grains vary in size from medium to very coarse, and are well rounded, although poorly sorted. The sandstone

[1] *Pennsylvania State University, University Park, Pennsylvania, 16802.*
[2] *Geological Society of America, Boulder, Colorado 80301.*

contains both igneous and metamorphic types of quartz; feldspar in proportions of 5 percent or less also is present and consists of albite, microcline, and orthoclase.

BASIN FACIES

The generally southward change from the Noonday Dolomite to the basin facies is so abrupt that the transition is represented in Figure 1 by single lines rather than by gradational symbols. South of these lines the two members of the Noonday are replaced by generally thicker clastic units of the basin facies which generally thicken abruptly. The abruptness of these facies and thickness changes is strongly indicative of vertical movements on fault-bounded blocks. Indeed, through-going faults that were active in Cenozoic and earlier time—the Butte Valley fault in the Panamint Range and the Sheephead fault in the Black Mountains—lie parallel with and close to the lines of facies change. In addition, abrupt variations in the thickness of the platform facies are controlled by vertical movements on contemporaneous faults, some of the faults being detectable in present exposures (Wright and others, 1974b, this volume).

The basin facies, which generally exceeds 300 m in thickness and has a maximum measured thickness of 488 m, is divisable into four members. In upward succession they are here named (1) the basal breccia member, (2) the arkose member, (3) the clastic carbonate member, and (4) the quartzitic clastic dolomite member.

The basal breccia member, which ranges from 3 m to 10 m or more thick in its observed occurrences, is characterized by a polymict breccia featured by gigantic clasts derived from the lower dolomite member of the platform facies and from conglomerate of the Kingston Peak Formation.

Well and evenly laminated dolomite, similar in lithology to the lower dolomite member, commonly forms a thin layer, no more than a meter or two thick, beneath the breccia. It marks a much thinned and basinward extension of the members. Well laminated dolomite also forms masses within the breccia. There the laminations are very irregular and appear to have been deformed contemporaneously with the transportation and deposition of the other components of the breccia. The basal breccia member has been observed only in a northwest-trending belt that lies with 1 km of the southward termination on the zone of marked thinning of the lower dolomite member (Figs. 1 and 5), and is especially well exposed in the southern part of the Ibex Hills (loc. 10, Fig. 1). Elsewhere the arkose member forms the lowest unit of the basin facies. We interpret the breccia as deposited at the base of the platform-basin slope and composed mostly of material dislodged from the slope.

The most complete, easily accessible, and continuous exposures of the other three members of the basin facies occur in the southern part of the Black Mountains immediately north of the highway between Jubilee Pass and the Ashford Mill site (loc. 11 in Fig. 1; Fig. 2), but they also are well exposed at numerous other localities extending from the southwestern part of the Panamint Range south-southeastward to the Silurian Hills.

The arkose member generally consists of arkose and feldspathic siltstone and shale. It forms a north-tapering clastic wedge with a maximum observed thickness of about 250 m. Its disappearance northward ordinarily occurs slightly north of the southern termination of the lower platform dolomite, the arkose thinning as the underlying dolomite thickens. Thus, when traced to its most northerly exposures, it generally can be observed to rest successively upon the breccia and the lower algal dolomite. In most places, however, the arkose member rests upon the Kingston Peak Formation.

In the thickest occurrences of the arkose member, the lower part consists of purple siltstone, shale, and limestone. These pass upward into a sequence of arkose and shale beds that are characterized by graded bedding and by bottom markings that indicate current directions from south to north. The arkose ranges from medium to very coarse grained and is very

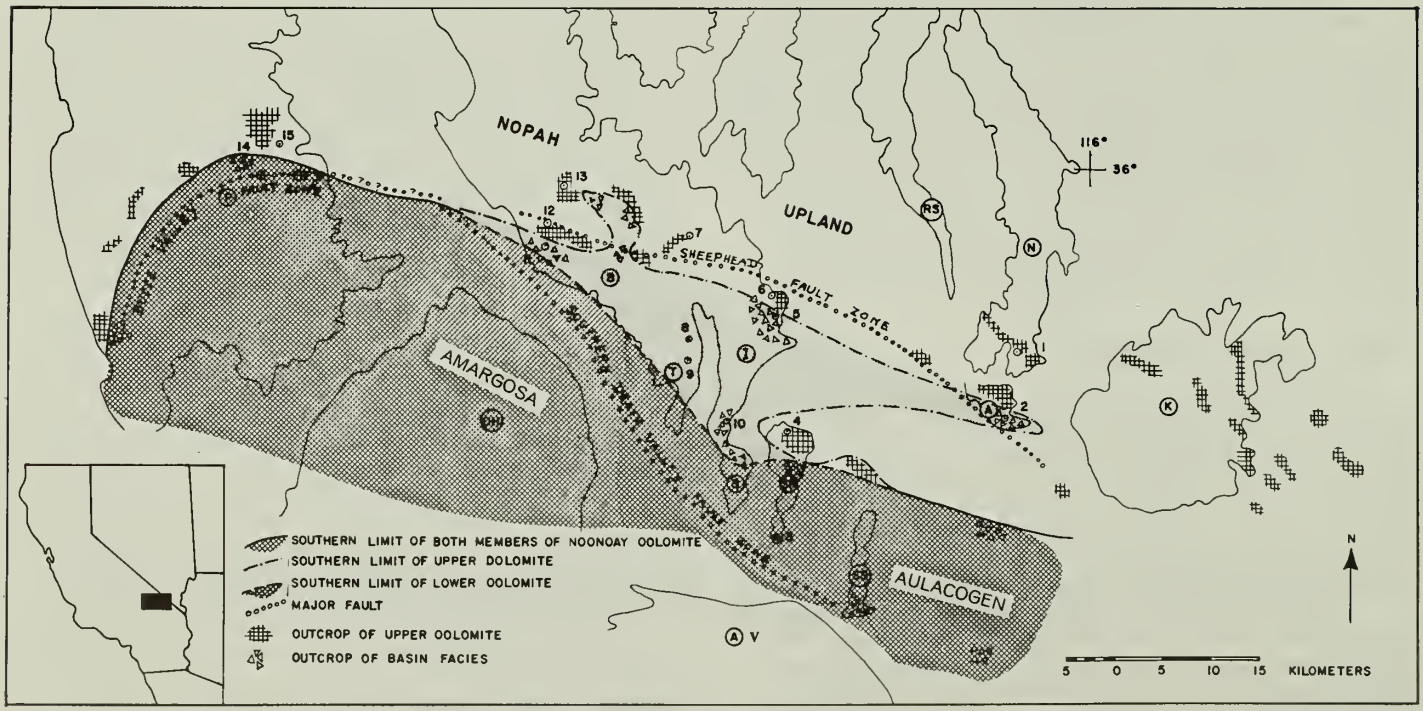


Figure 1 Sketch diagram showing distribution of Noonday Dolomite and equivalent basin facies in southern Death Valley region. Numbers refer to localities where sections were measured. A = Alexander Hills, Av = Avawatz Mts., B = Black Mts., I = Ibex Hills, K = Kingston Range, N = Nopah Range, OH = Owlshead Mts., P = Panamint Range, RS = Resting Spring Range, S = Saratoga Hills, SP = Saddle Pink Hills, SS = Salt Spring Hills, T = Talc Hills.

poorly sorted. It consists of approximately equal proportions of quartz and feldspar, the quartz grains being angular and the feldspar grains highly altered. The thinner, more northerly occurrences of the member consist of purple arkosic siltstone interbedded with argillaceous dolomite beds that weather yellowish brown.

The clastic carbonate member is 190 m in maximum thickness, and consists of varicolored interbeds of dolomite, limestone, and shale that are generally 2 to 30 cm thick and laminated. Close to the platform margin, however, the carbonate unit contains conglomeratic layers with angular fragments of limestone, limy dolomite, and dolomite. Some of the clasts are as much as 15 m long.

Both the bedded carbonate and the clasts of the carbonate-bearing conglomeratic strata weather reddish brown, yellowish brown, and lavender. The content of insoluble residues in the bedded carbonates ranges from 10 to 60 percent and consists of silt-size quartz and chert, illite, and minor kaolinite. A very small proportion of the beds contains well-rounded medium-grained quartz. Some of the beds exhibit contorted laminations that suggest contemporaneous, soft sediment deformation.

The uppermost member of the basin facies, which is as much as 460 m thick, consists of clastic dolomite mixed with 10 to 50 percent of coarse, well-rounded floating grains of quartz and very minor proportions of feldspar. Most of this member qualifies as quartzitic clastic dolomite. Occurrences that lie close to the line of facies change commonly consist of coarse breccia with clasts apparently derived, in part, from the upper platform dolomite. Some of the clasts are intraformational, consisting of the quartzitic clastic dolomite. Beds are generally within the range of 30 cm to 1 m thick. Most of them are devoid of internal structure but cross-bedding is locally present. In many outcrops, bedding is difficult or impossible to detect. The quartzitic clastic dolomite is gradational with the overlying transitional member of the Johnnie Formation, in which cross-bedding is abundant and well preserved.

REGIONAL STRATIGRAPHIC RELATIONS AND ENVIRONMENTS

The regional stratigraphic relations between the two memoers of the Noonday Dolomite and equivalent strata of

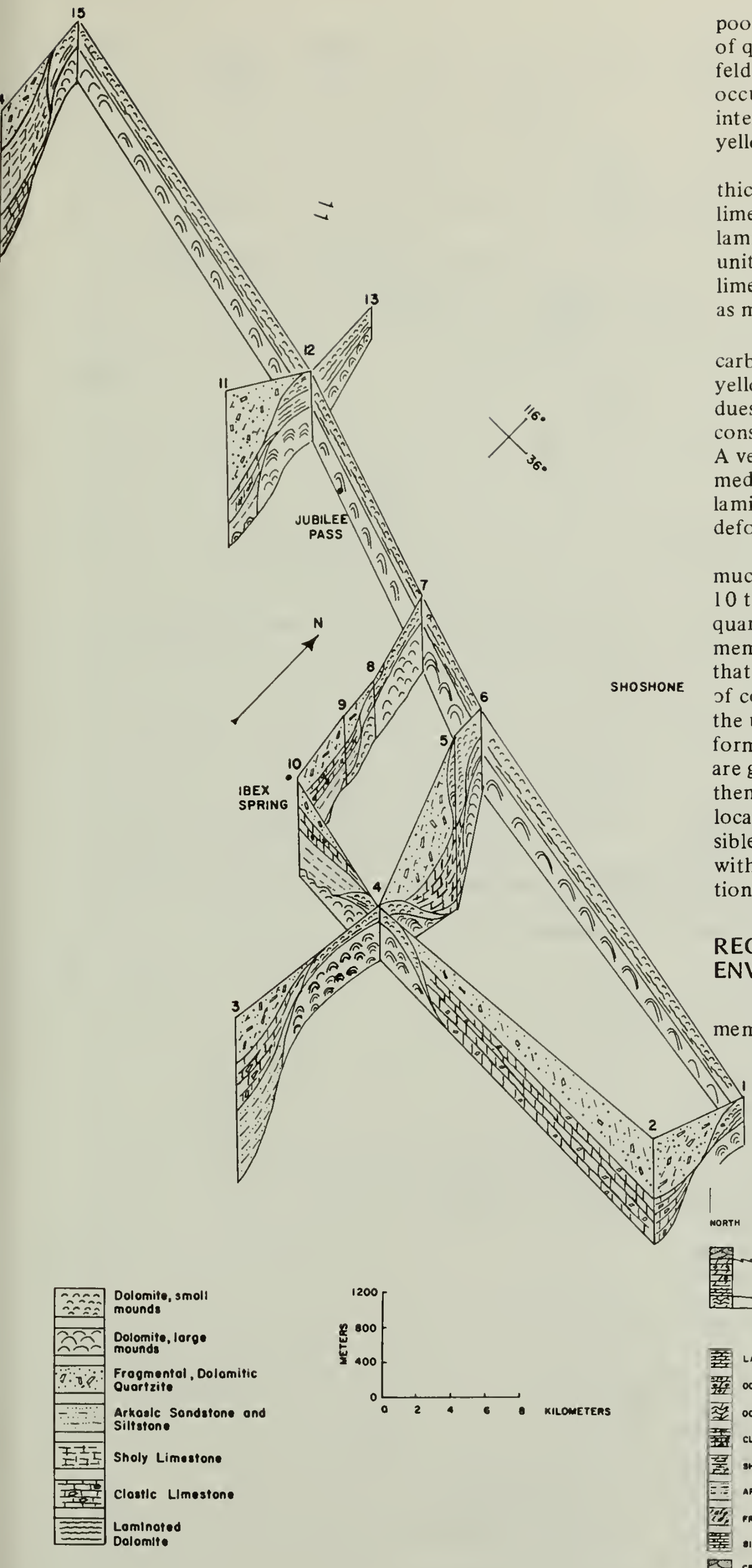


Figure 2. Fence diagram showing relations between members of Noonday Dolomite and equivalent basinal facies. Numbers refer to localities where sections were measured (see Fig. 1).

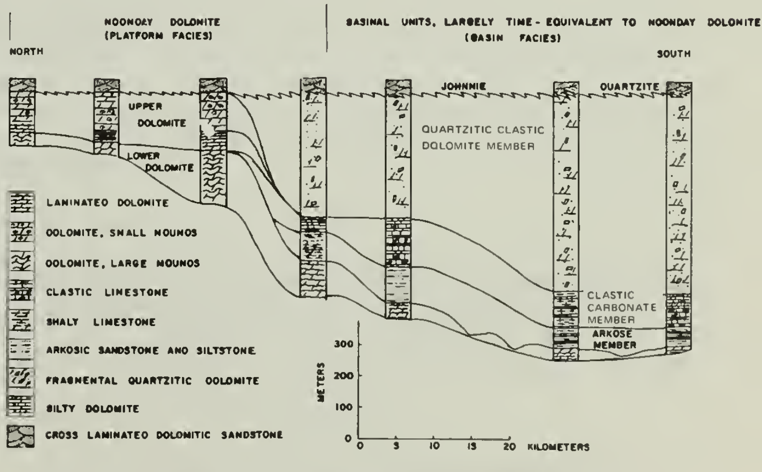


Figure 3. Stratigraphic cross section showing relationships between members of Noonday Dolomite and equivalent basinal facies. Data taken from measured sections near Jubilee mine in west-central part of the Black Mountains and projected to a common north-south line of section. Basal breccia member is missing at this locality.

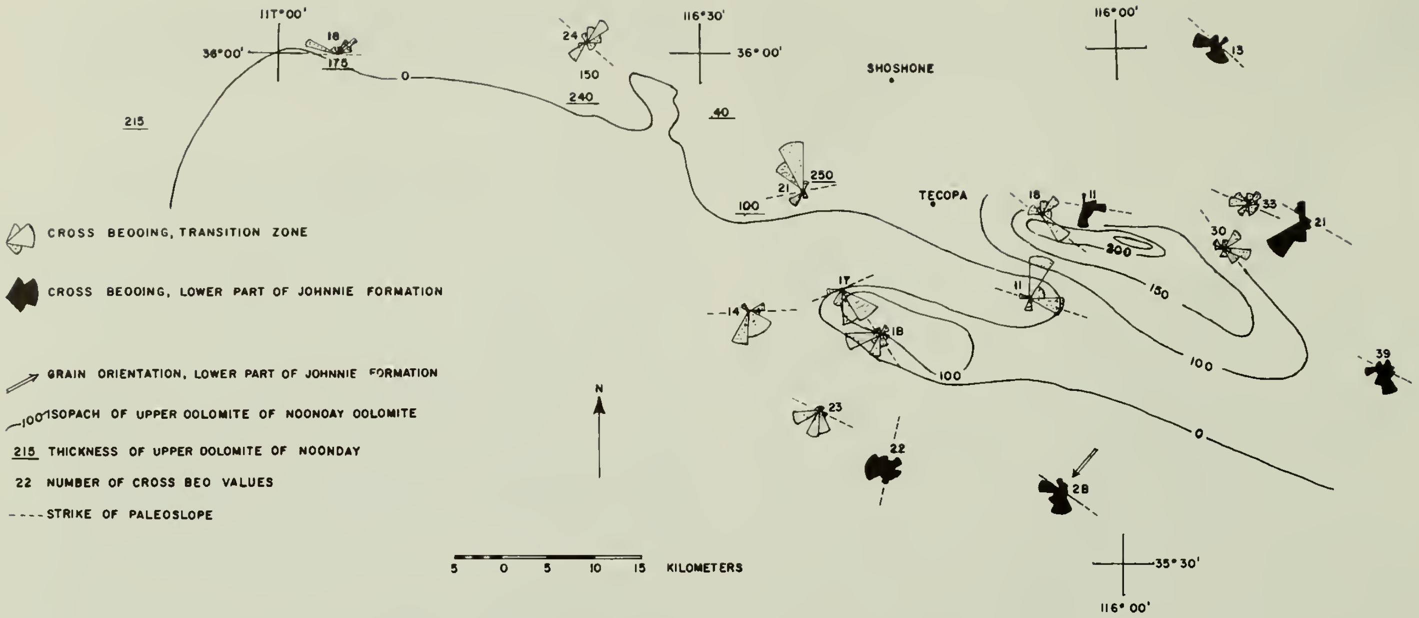


Figure 4. Cross bedding, grain orientation, and thickness information of upper and lower members of Noonday Dolomite, and lower part of Johnnie Formation.

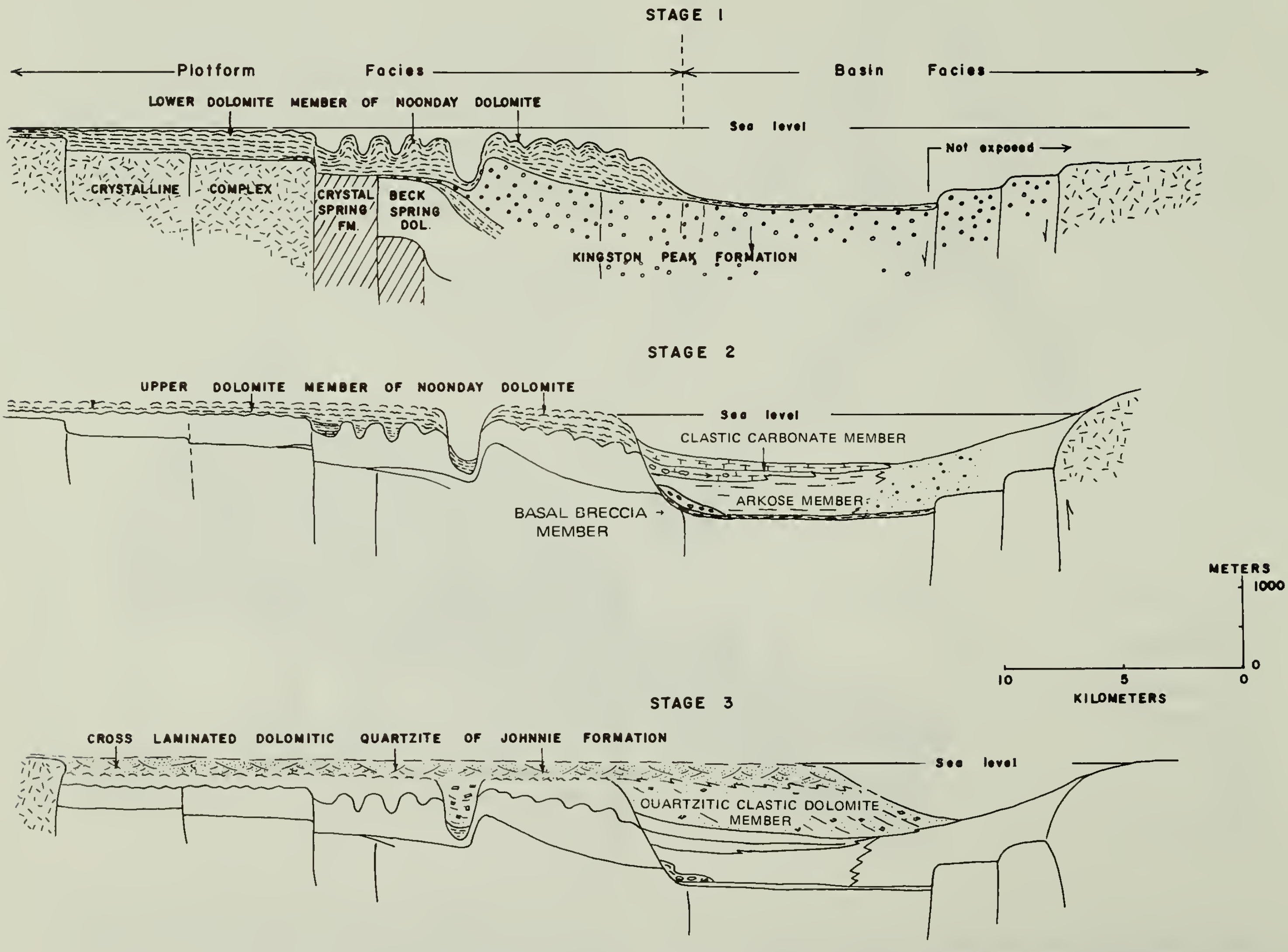


Figure 5. Composite sequential diagram showing development of Noonday Dolomite and equivalent basinal facies in these stages.

the basin facies are illustrated in Figure 3; the boundaries between the two are shown as lines in Figure 1. Observe that the southern limits of the lower and upper dolomite members of the Noonday are divergent in the eastern two-thirds of the region and coincident in the western third. These lines are interpreted as being controlled primarily by vertical movements along faults and contemporaneous with sedimentation. The faults, thus determining the trend of the basin and the adjacent platform.

Verification of the configuration of the northern edge of the basin boundary is independently obtainable through analysis of cross-bedding in the dolomitic quartzite of the overlying transitional member of the Johnnie Formation. A total of 360 measurements were made at the 17 localities shown in Figure 4. The south zero isopach of the upper dolomite marks the platform edge in upper Noonday time. The strike of the paleoslope during the deposition of the lower Johnnie strata, estimated from cross-bedding rosettes, is closely parallel to the trends of the isopachs. The tendency for modal directions in the lower part of the transition member to trend both platformward and basinward probably reflects tidal influences. The generally southwest directions shown by cross-bedding in the higher parts of the transition member may reflect influence of fluvial processes.

The various observations cited thus far are integrated into the composite sequential diagram in Figure 5. In stage one, laminated algal limestone was deposited during transgression of sediment-free marine waters across the beveled fault blocks upon which the Pahrump Group is preserved. Vertical movements on these faults in Pahrump time influenced the thickness and lithologies of the various Pahrump units (Wright and others, this volume). Relief on the surface was generally low, but hills as high as 300 m marked the position of a resistant unit of conglomerate in the upper part of the Kingston Peak Formation. During this first stage of Noonday deposition, the more northerly blocks were depressed slowly, with respect to sea level, and the algal mats grew evenly. We interpret the algal mounds as characterizing blocks that were being depressed less slowly, and that algal growth was significantly favored on highs of the basement topography and retarded on lows. The differential relief of highs and lows was increased by different rates of algal growth as depression proceeded. Blocks that were depressed to still greater depths contain very thin occurrences of the lower dolomite member, or none at all—the growth of the algal mats apparently being inhibited or prevented by insufficient sunlight.

Stage two began with the abrupt ending of the deposition of lower algal strata throughout the platform area, the paleotopography of the surface of that unit being preserved in the sharp contact between it and the overlying upper dolomite. Because this contact is equally sharp throughout the platform area, regardless of the thickness of the lower dolomite member or the presence or absence of the mounds, it seems attributable to a regional event, tectonic or eustatic, that caused a change in sea level. It probably represents a deepening of the ocean floor that killed the mat-forming algae, as no evidence of erosion has been observed even on the highest parts of the largest mounds, and the very evenly laminated silty dolomite and siltstone that fill some of the intermound areas are devoid of algal structures and of features attributable to waves or currents.

The coarse breccia at the base of the basin facies formed after much, if not all, of the lower dolomite member was deposited, as it contains large clasts of that unit as well as much debris from conglomerate of the Kingston Peak Formation, which, immediately north of the basin margin, underlies the lower dolomite member. The breccia probably records a further tectonic deepening of the basin, somewhat before or at the beginning of stage two, steepening the basin margin and favoring the accumulation of the breccia, through submarine sliding, on slopes at the base of the escarpment. The debris from the Kingston Peak could have been derived only from the escarpment, because to the north of the basin margin all older units were sealed beneath the platform cover of dolomite.

The arkose member rests upon basal breccia and lower algal dolomite near the northern margin of the basin and upon the Kingston Peak Formation farther south. Its superjacent relation with the breccia and lower dolomite along the basin margin, however, indicates that some and perhaps most of the arkosic unit was deposited during stage two.

The arkosic sediment provides a record of the elevation and erosion of a granitic land area to the south. Although the granitic terrane has been obliterated by later deformational and igneous events, its southerly location can be inferred from the following observations: (1) bottom markings on arkose beds indicate sediment transport was from south to north; (2) the crystalline basement to the north was sealed by the Noonday carbonate rocks; and (3) the Noonday carbonate and detrital rocks on the shelf contain very little feldspar.

Later, argillaceous limestone and shale were deposited in stage two in the basin, their formation being interrupted by debris flows and turbidity currents that carried broken material from the platform margin and platform-basin slope and deposited it near the base of the slope. The occurrence in the basin deposits of fragments and blocks of limestone, limy dolomite, and dolomite suggests that the algal reef had been partly dolomitized prior to the breaking away of the fragments.

The return, during stage two, of an environment favoring the growth of algae on the platform is recorded in the upper dolomite, but the original limestone of this unit was deposited during an influx of clay and quartz silt. We interpret the wavy structures as forming in the presence of sediment-entrapping algae. The final event, illustrated in stage 3 in Figure 5, is marked by the influx of large volumes of quartz sand bringing the destruction of the reef-forming algal mats. By the beginning of this influx, the reef had been completely dolomitized, because only dolomite fragments are found in the resulting basin deposits of quartzitic clastic dolomite and breccia.

On the shelf, tidal currents reworked quartz and fragmental dolomite into bars and beaches, whereas in the basin, submarine currents carried the quartz and reef debris below the wave base to be deposited as submarine fans. This process continued until the water was sufficiently shallow to permit action of tidal currents. At this point the paleotopography produced by faulting was obliterated. Roberts (this volume) and Diehl (this volume) have presented evidence that the southwesterly paleoslope of the Noonday platform existed as early as the beginning of Pahrump time and continued well after the Noonday Dolomite was deposited. The close parallelism between major faults that have been active in Cenozoic and earlier time and the inferred Precambrian faults that mark the margins of the ancient basin or trough suggest periodic reactivation of ancient zones of weakness over an immense segment of geologic time.

REFERENCES CITED

Hazzard, J. C., 1937, Paleozoic section in the Nopah and Resting Springs Mountain, Inyo County, California: California Div. Mines Jour. Mines and Geology, v. 33, no. 4, p. 273–339.

Hunt, C. B., and Mabey, D. R., 1966, Stratigraphy and structure Death Valley, California: U.S. Geol. Survey Prof. Paper 494–A, 162 p.

Wright, L. A., 1973, Geology of the SE¼ Tecopa 15-minute quadrangle, San Bernardino and Inyo Counties, California: California Div. Mines and Geology, Map Sheet 20.

Wright, L. A., and Troxel, B. W., 1966, Strata of late Precambrian–Cambrian age, Death Valley region, California-Nevada: Am. Assoc. Petroleum Geologists Bull., v. 50, no. 5, p. 846–857.

——— 1967, Limitations on right-lateral, strike-slip displacement, Death Valley and Furnace Creek fault zone, California: Geol. Soc. America Bull., v. 78, p. 933–950.

This research was supported by NSF Grant GA–16119.

Oblique aerial view southwestward across central Death Valley. Garlock fault extending diagonally across upper part of photo is marked by linear depression along southern edge of Owlshead Mountains (left side of photo) and Slate Range (range with bulge at far end). Wingate Wash forms broad topographic low between Owlshead Mountains (on left) and southern Panamint Range. Panamint Valley is near valley beyond Panamint Range. Sharp boundary between Black Mountains (left foreground) and Death Valley is caused by uplift along normal faults. Note small fans on east side of Death Valley as compared to broad fans from Panamint Range. Mormon Point is at tip of abrupt change in trend of edge of Black Mountains. Smith Mountain lies immediately to east (left). Copper Canyon is toward observer from Mormon Point. U.S. Geological Survey – U.S. Air Force photograph.

Turtleback Surfaces of Death Valley Viewed.as Phenomena of Extensional Tectonics

Lauren A. Wright and James K. Otton
Department of Geosciences, Pennsylvania State University,
University Park, Pennsylvania 16802

Bennie W. Troxel
Geological Society of America,
Boulder, Colorado 80301

ABSTRACT

The controversial turtleback surfaces of Death Valley may be colossal fault mullions resulting from severe crustal extension which were localized along undulating and northwest-plunging zones of weakness that were in existence prior to this deformation. Supporting evidence includes (1) a coincidence between the surfaces and carbonate layers in folded metamorphic rocks beneath the surfaces, and (2) striations, slickensides, and extensional fractures with orientations compatible with northwest extension.

The three peculiar domical fault surfaces that feature the Black Mountains along the east side of Death Valley have puzzled students of Basin and Range geology since these features were described and given the name "turtlebacks" by Curry (1938). Although the three surfaces occupy relatively small areas (Fig. 1), they are more than geologic curiosities; various interpretations of their origin support widely different views of the deformational and erosional history of the southwestern Great Basin in Cenozoic time.

As indicated by Curry, each turtleback surface is little eroded and broadly curved to resemble the carapace of a turtle. Each plunges from a height of 865 to 1,350 m in the Black Mountains northwestward to the floor of Death Valley. Each surface is coincident with an anticlinal fold in metasedimentary units of Precambrian age. Beyond the limits of the turtleback surfaces, the mountain front is underlain by other rock units, mainly by bodies of Mesozoic(?) and Tertiary igneous rocks. Resting with fault contact upon each turtleback surface are remnants of a cover of Cenozoic sedimentary and volcanic rocks which is more deformed than the underlying, broadly folded units.

The carapace-like form of the turtleback surfaces has been attributed to

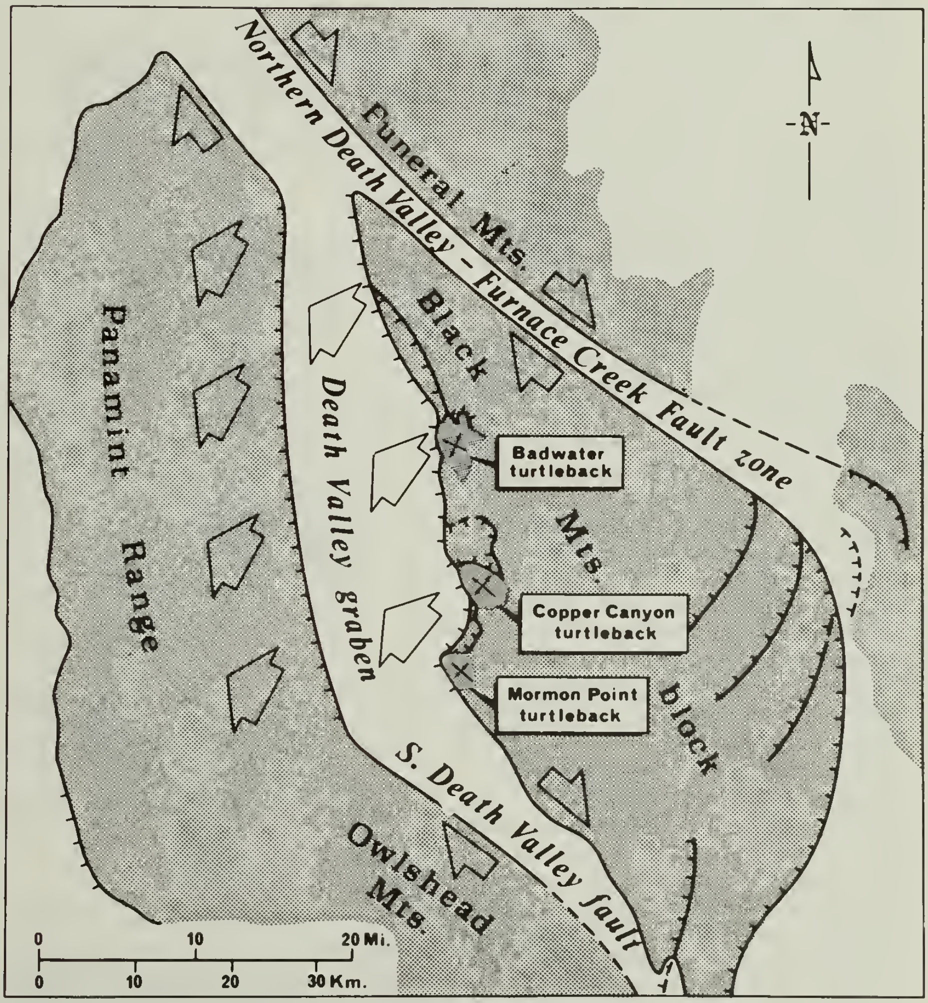


Figure 1. Generalized structural map of Death Valley region, showing position of three turtleback surfaces of Black Mountains. Hachured lines mark positions of major normal faults; full arrows show inferred direction of crustal extension; half arrows show relative displacement on strike-slip fault zones.

(1) compressional folding of a regional thrust fault (Curry, 1954; Hunt and Mabey, 1966; Noble, 1941); (2) differential erosion to produce an undulating topographic surface upon which the Cenozoic rocks were deposited and from which they later slid, propelled by gravity (Drewes, 1959); (3) uparching related to the intrusion of shallow plutons (Sears, 1953); and (4) compressional folding of both the basement rocks and the cover, the cover being essentially autochthonous (Hill and Troxel, 1966).

We propose yet another hypothesis, simpler than those already mentioned, but, we believe, more compatible with the available field data and with the generally held view that most or all of the Great Basin has been undergoing severe crustal extension since mid-Tertiary time. This hypothesis, which is subject to testing through detailed geologic mapping, holds that the turtleback surfaces are gigantic fault mullions developed along planes of weakness in a zone of normal faulting that penetrates the crust (Fig. 2). The cover of Cenozoic rocks is visualized as having been deposited along the fault zone following the inception of faulting. The cover then moved downward, generally parallel with the axes of the turtleback surfaces, as the hanging-wall side of the fault moved downward and northwestward to deepen Death Valley in the manner shown in Figure 2.

We formulated this pull-apart hypothesis in recent years during numerous visits to the Mormon Point and Copper Canyon turtlebacks, where we observed that these surfaces exhibit abundant linear features related to movement along the fault surface. They range in scale from minute slickensides to fault mullions tens or hundreds of meters in amplitude. All are similarly oriented and tend to lie parallel to the axes of the turtleback surfaces. Consequently, at the nose of each of the two turtlebacks, the linear features plunge moderately valleyward; along the crest and flanks, the linear features are horizontal to gently plunging. We also noted that the turtleback surfaces are well developed only on the folded metasedimentary units and tend to coincide with carbonate-rich layers, which would probably yield to stress more readily than the schist and gneiss units with which the carbonate layers are associated. In addition, the metasedimentary rocks are thoroughly broken by fractures that tend to lie normal to the turtleback axes, many of the fractures being occupied by dikes of Cenozoic age.

These features seem more compatible with the pull-apart hypothesis for turtleback formation than with the origin

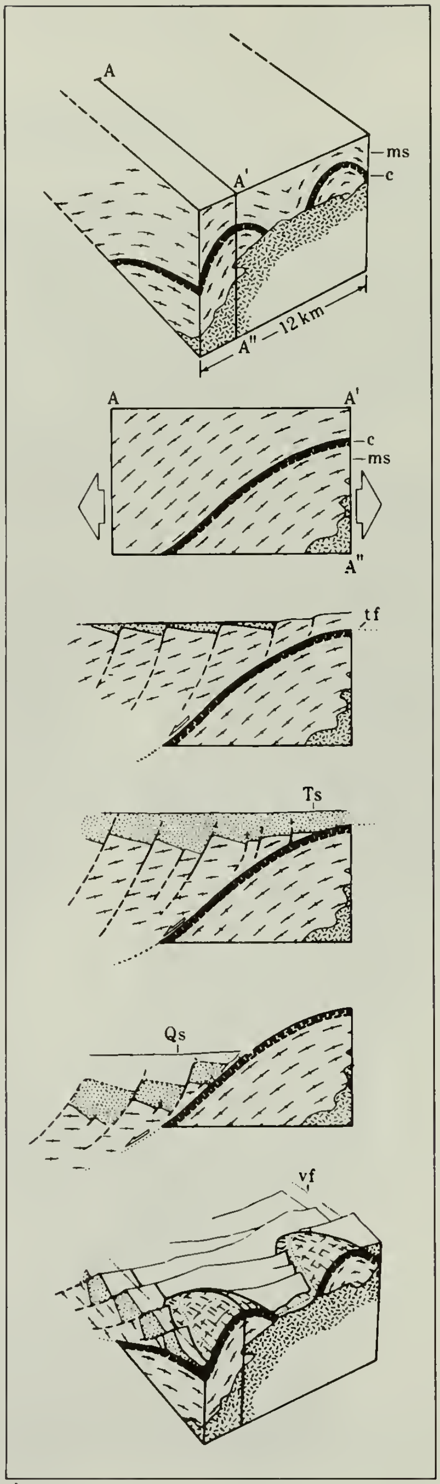


Figure 2. Idealized block diagrams and cross sections, illustrating pull-apart concept of turtleback formation; based on observations of Copper Canyon and Mormon Point turtlebacks, Death Valley. c = Carbonate layers; ms = mixed metasedimentary rock; Qs = Quaternary sediments; tf = turtleback fault; Ts = Tertiary sedimentary rock; vf = valley floor.

suggested by others. This interpretation is, in turn, compatible with the observations that the entire Death Valley region is characterized by innumerable normal faults that trend northward to northeastward; that central Death Valley is essentially a graben, formed in late Cenozoic time, between the en echelon ends of two northwest-trending, high-angle faults of right-lateral displacement as shown in Figure 1 (Burchfiel and Stewart, 1966); and that the graben is part of a rhombochasm that has been forming through a much lengthier segment of geologic time (Wright and Troxel, 1967, 1971) and also in an environment of northwesterly crustal extension.

If the turtlebacks were formed by a pull-apart mechanism, these surfaces can no longer be cited as evidence for large-scale Cenozoic thrusting in the Death Valley region, for northeast compression during or after turtleback formation, or for structurally controlled relief on an eroded surface that received the Tertiary sedimentary and volcanic units.

REFERENCES CITED

Burchfiel, B. C., and Stewart, J. H., 1966, "Pull-apart" origin of the central segment of Death Valley, California: Geol. Soc. America Bull., v. 77, p. 439-442.

Curry, H. D., 1938, "Turtleback" fault surfaces in Death Valley, California [abs.]: Geol. Soc. America Bull., v. 49, p. 1875.

—— 1954, Turtlebacks in the central Black Mountains, Death Valley, California: California Div. Mines Bull., v. 170, p. 53-59.

Drewes, H., 1959, Turtleback faults of Death Valley, California: A reinterpretation: Geol. Soc. America Bull., v. 70, p. 1497-1508.

Hill, M. L., and Troxel, B. W., 1966, Tectonics of Death Valley region, California. Geol. Soc. America Bull., v. 77, p. 435-438.

Hunt, C. B., and Mabey, D. R., 1966, Stratigraphy and structure, Death Valley, California: U.S. Geol. Survey Prof. Paper 494-A, p. A137.

Noble, L. F., 1941, Structural features of the Virgin Spring area, Death Valley, California: Geol. Soc. America Bull., v. 52, p. 994.

Sears, D. H., 1953, Origin of Amargosa chaos, Virgin Spring area, Death Valley, California: Jour. Geology, v. 61, p. 182-186.

Wright, L. A., and Troxel, B. W., 1967, Limitations on right-lateral strike-slip displacement, Death Valley and Furnace Creek fault zones, California: Geol. Soc. America Bull., v. 78, p. 947.

—— 1971, Evidence for tectonic control of volcanism, Death Valley: Geol. Soc. America, Abs. with Programs (Cordilleran Sec.), v. 3, no. 2, p. 221.

ACKNOWLEDGMENTS

Reviewed by B. C. Burchfiel.

MANUSCRIPT RECEIVED NOVEMBER 26, 1973

MANUSCRIPT ACCEPTED DECEMBER 14, 1973

Reprinted by permission from *Geology*, v. 2, p. 53-54, Geological Society of America, February 1974.

Geologic Maps and Sections of a Strip from Pyramid Peak to the Southeast End of the Funeral Mountains, Ryan Quadrangle, California

James F. McAllister[1]

Work done partly in cooperation with the California Division of Mines and Geology.

Table 1. Explanation of map units. Maps are contiguous, aligned along A–A″. (Thickness is rounded to nearest 100 in maximum section.)

Qal	ALLUVIUM (Quaternary)
Qao	OLDER ALLUVIUM (Quaternary)
Qlds	LANDSLIDE MATERIAL (Quaternary)
QTf	FUNERAL FORMATION (Pleistocene? and Pliocene)
Tcl	CONTINENTAL CLASTIC ROCKS (Tertiary)
Tls	LACUSTRINE LIMESTONE (Tertiary)
Mp	PERDIDO FORMATION (Upper and Lower(?) Mississippian) 500 feet (top eroded)
Mt	TIN MOUNTAIN LIMESTONE (Lower Mississippian) 300 feet
Dl	LOST BURRO FORMATION (Upper and Middle Devonian) 2,500 feet
DSh	HIDDEN VALLEY DOLOMITE (Lower Devonian and Silurian) 1,400 feet
Oes	ELY SPRINGS DOLOMITE (Upper and Middle Ordovician) 500 feet
Oe	EUREKA QUARTZITE (Middle Ordovician) 400 feet
Op	POGONIP GROUP (Middle and Lower Ordovician) 2,200 feet
Ꞓn	NOPAH FORMATION (Upper Cambrian) 1,700 feet
	BONANZA KING FORMATION
Ꞓbb	BANDED MOUNTAIN MEMBER (Upper and Middle Cambrian) 2,400 feet
Ꞓbp	PAPOOSE LAKE MEMBER (Middle Cambrian) 1,200 feet
Ꞓc	CARRARA FORMATION (Middle and Lower Cambrian) 1,600 feet

Lower formations exposed in the Funeral Mountains, Ryan quadrangle

ZABRISKIE QUARTZITE (Lower Cambrian) 800 feet

WOOD CANYON FORMATION (Lower Cambrian and Precambrian) 4,000 feet

STIRLING QUARTZITE (Precambrian) 4,800 feet

JOHNNIE FORMATION (Precambrian) 1,000 feet to concealed part

REFERENCES

McAllister, James F., 1970, Geology of the Furnace Creek borate area, Death Valley, Inyo County, California: California Div. Mines and Geology Map Sheet 14.

——— 1973, Geologic map and sections of the Amargosa Valley borate area—southeast continuation of the Furnace Creek area—Inyo County, California: U.S. Geol. Survey Misc. Geol. Inv. Map I-782.

——— 1974, Silurian, Devonian, and Mississippian formations of the Funeral Mountains in the Ryan quadrangle, Death Valley region, California: U.S. Geol. Survey Bull. 1386, 35 p. (in press).

[1] *U.S. Geological Survey, Menlo Park, California 94025.*

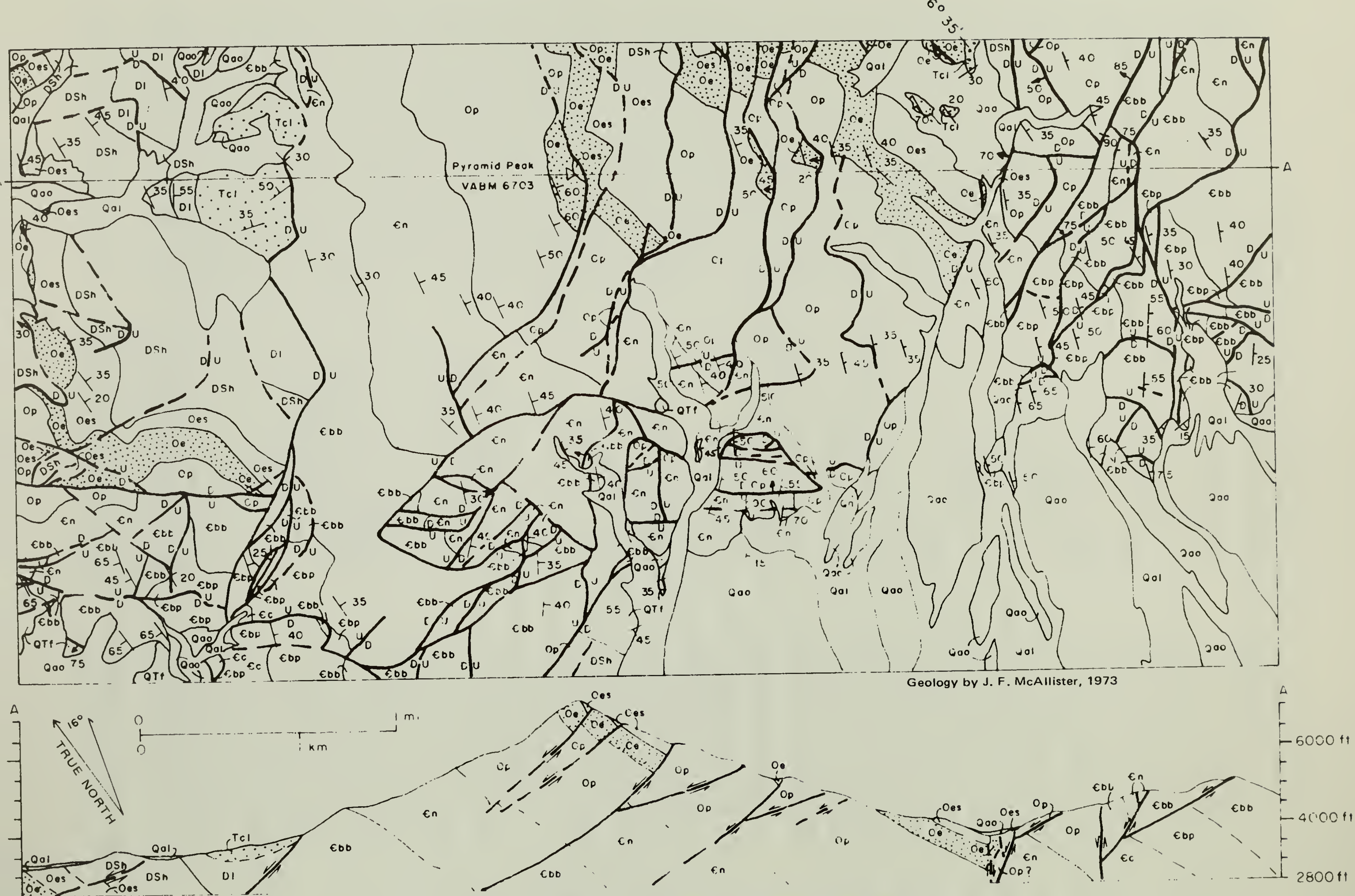


Figure 1. Geology at Pyramid Peak, Funeral Mountains. Explanation of map units in Table 1.

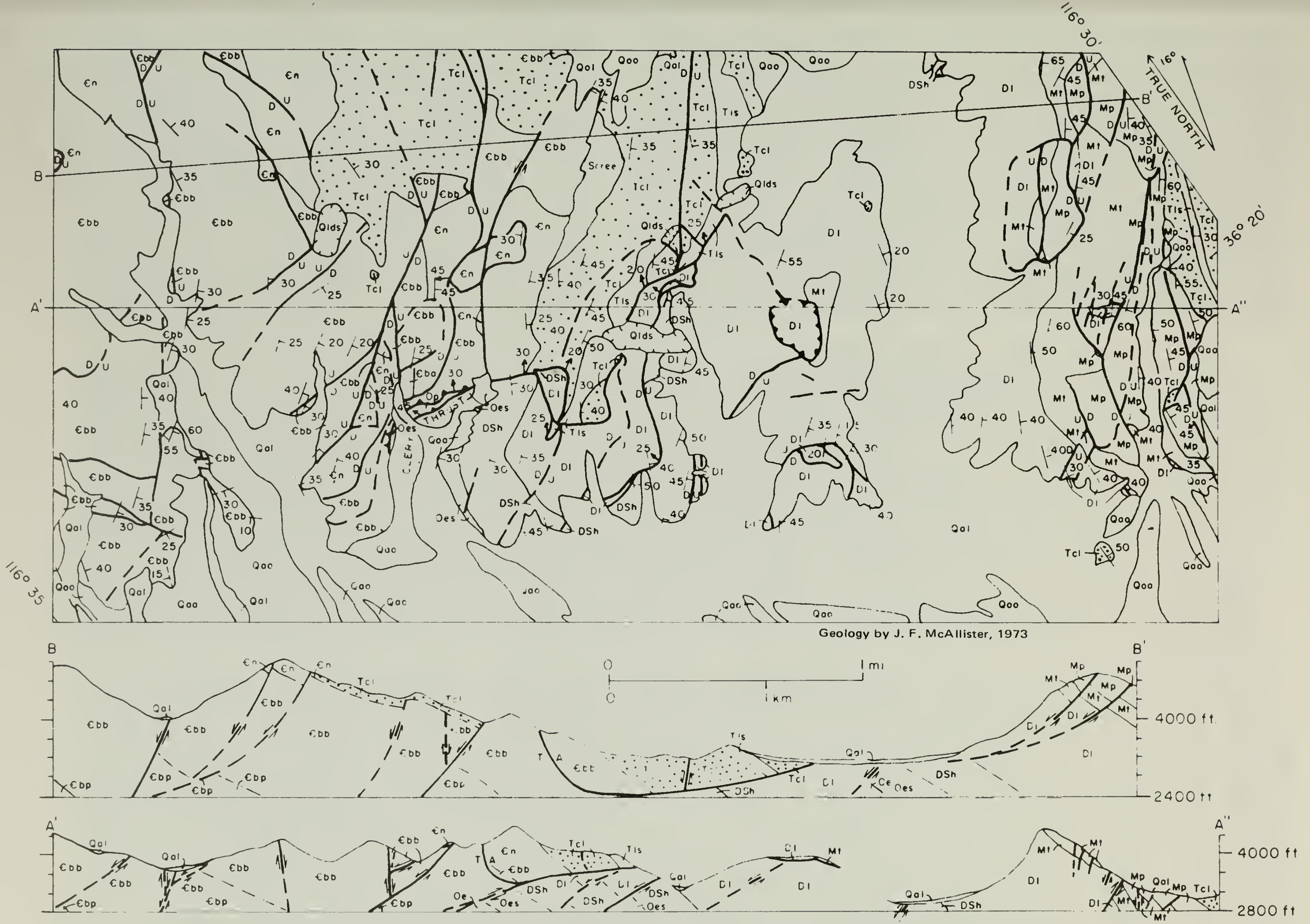


Figure 2. Geology at southeast end of Funeral Mountains. Explanation of map units in Table 1.

GEOLOGY OF THE FURNACE CREEK BORATE AREA, DEATH VALLEY, INYO COUNTY, CALIFORNIA

By JAMES F. McALLISTER

U.S. Geological Survey, Menlo Park, California

PRECAMBRIAN AND PALEOZOIC ROCKS

The oldest rocks in the area are in the Funeral Mountains. The sedimentary sequence extends concordantly from the youngest Precambrian into the oldest Paleozoic rocks and continues in geologic age through the Cambrian and Ordovician. A few structurally isolated blocks represent the Silurian and Early Devonian. In general terms of years, these sediments accumulated within the interval from somewhat more than 600 million to perhaps 400 million years ago.

The sediments were deposited in seas that spread over the Death Valley region and far across Nevada. At first predominantly siliceous clastic sediments accumulated as sand and mud, but soon after the Early Cambrian Epoch carbonate constituents of limestone and dolomite became dominant. Some distinctive beds of contrasting rocks—limestone or dolomite in the lower part, and quartzite or sandstone and shale in the upper part—relieve the lithic monotony. The marker beds and marine fossils diagnostic of the geologic age are most useful guides in mapping parts of the rock sequence. Their persistence through many a mountain range greatly aids in correlating the parts beyond each desert basin, and their wide distribution indicates the vastness of the ancient seas.

The Wood Canyon Formation, of Precambrian and Early Cambrian age, in the northeastern segment of the map area, forms the least conspicuous olive-gray slopes on the flank of the Funeral Mountains. The Wood Canyon consists of generally dark interlayered shale, siltstone, and quartzite, in which a few beds of lighter brownish-weathering dolomite or limestone distinguished a lower member (p€wl) and an upper one (€wu) from a more quartzitic middle member (€p€wm). The oldest Cambrian trilobites first appear in the regional correlative of the middle member, and Early Cambrian archaeocyathids as well as olenellids are widespread in the upper member, so that by convention the ages of the three members in ascending order are designated Precambrian, Precambrian and Early Cambrian, and Early Cambrian. The top of the Wood Canyon is well outlined by the Zabriskie Quartzite's contrastingly lighter color.

The Zabriskie Quartzite (€z), of Early Cambrian age, is generally shattered here by faulting and forms rough steep slopes, where it weathers from very pale orange to brown. The fresh quartzite is commonly pale red, ranging from vitrually white to grayish red. Although diagnostic fossils have not been found anywhere in the Zabriskie, those above and below limit the age to Early Cambrian.

The Carrara Formation (€c), of Early and Middle Cambrian age, displays the transition from siliceous clastic to carbonate rocks, apparent even from a distance, by a diversity of colors and varied relief of different layers. In the lowest part of the Carrara, light-colored quartzite and dark shale resemble rocks in the underlying Zabriskie and Wood Canyon Formations, whereas silty limestone in the highest part resembles limestone in overlying formations. Three prominent sets of limestone beds, separated by predominantly shaly and silty beds, stand out as ridges or as ribs on steep slopes. The two lower sets are conspicuously gray, but the highest is partly subdued by brown-weathering clastic material. Much of the limestone, either silty or purer, contains concentrically structured ovoids as much as two inches long, generally attributed to the alga *Girvanella*. Diagnostic trilobites in particular beds well spaced through the Carrara make it one of the most significantly fossiliferous formations in the section. They demonstrate that the lower part is Early Cambrian and the upper part is Middle Cambrian (A. R. Palmer, written communications, 1957–1965).

Extracted and reproduced with permission from California Division of Mines and Geology Map Sheet 14, 1970. (See original publication for full text, map, and references.)

The Bonanza King Formation, Middle and Late Cambrian in age, is distinguished on mountainsides by broad bands and narrow stripes of dark- and light-gray dolomite and limestone. The lower part, generally dark, makes up the Papoose Lake Member (€bp). The upper part, of contrastingly light and dark layers, constitutes the Banded Mountain Member (€bb). The pattern of the layering in the Banded Mountain Member is uniform for many miles, and some parts of it are traceable across the region. The base of the Banded Mountain Member is defined by conspicuous brownish-weathering clastic beds, which are a regional stratigraphic key, lying somewhat below the middle of the Bonanza King. In the Funeral Mountains, Middle Cambrian trilobites have been collected from these beds and from similar but much less conspicuous beds in the lower part of the Papoose Lake Member. No younger fossils have yet been found here in the Bonanza King Formation.

The Nopah Formation (€n), of Late Cambrian age, resembles the underlying Banded Mountain Member in that it consists of dark- and light-gray layers of dolomite above brownish-weathering clastic beds at the base. But the combination of light and dark gray is different from that of the Banded Mountain and may vary from place to place. The basal shale and silty limestone beds are persistently fossiliferous throughout the region and beyond, containing species of trilobites and acrotretid brachiopods diagnostic of a Late Cambrian age. The main, lower part of the basal unit is equivalent to the Dunderberg Shale, a formation of long standing in east-central Nevada. Dark dolomite above the middle of the Nopah Formation contains small silicified gastropods and another mollusk, the singular *Matthevia* Walcott, that are also Late Cambrian (Yochelson and others, 1965).

The Pogonip Group (Op), here of Early and Middle Ordovician age, includes limestone or dolomite, shale, siltstone, minor sandstone or quartzite, and some cherty beds in distinctive combinations that at other places are designated formations, and even members of a formation. The lowest dolomite or limestone, although somewhat gradational with the underlying Nopah, contains enough iron-bearing silt or other clastic material to add a light-olive or brownish tinge to the gray of the Pogonip. Within the lower part, shaly and silty beds are concentrated in a thin but readily discernible light-brown or grayish-orange zone in the carbonate rock. The middle third of the Pogonip Group consists of thicker similar zones interlayered with darker brown siliceous zones and gray carbonate rock. The shaliest layers, lightest in color, make notches on spurs and benches on mountainsides, supported by brown or gray ledges and crags. The upper third of the Pogonip consists mainly of cliff-forming dark- to medium-gray limestone or dolomite, tinged brown or pale red, but the topmost beds are varied and more colorful. The Early and Middle Ordovician age of the entire Pogonip Group in the southern Funeral Mountains is well documented by R. J. Ross, Jr. (written communications, 1962–67; 1967, p. D32–D34), who shows that the few trilobites collected from the lowest part are clearly Early Ordovician, although not of the earliest, and that abundant brachiopods and trilobites from near the middle of the Pogonip mark the change from Early to Middle Ordovician. The most conspicuous fossils are large gastropods, which occur widely in the upper, cliff-forming limestone or dolomite.

Three younger formations of Paleozoic rocks, although remaining closely associated, are broken up in faulted masses and ancient landslides widely spaced along the range front from Travertine Point to about two miles northwest of Echo Canyon. The fault boundaries of the formation blocks are much too generalized on the map to express the true disorder. The Eureka Quartzite (Oe), of Middle Ordovician age, is characterized by nearly white, pure quartzite in the upper part and brown-weathering quartzite and thinly interbedded less coherent, finer grained clastic rocks in the lower part. The Ely Springs Dolomite (Oes), of Middle and Late Ordovician age, consists of somewhat cherty dark-gray dolomite that contrasts sharply with the underlying Eureka and contains at the top some lighter gray dolomite, which outlines the boundary with the overlying dark part of the Hidden Valley Dolomite (DSh). Light-gray dolomite in the Hidden Valley is conspicuously thick above the dark cherty dolomite in the lower part of the formation. No Paleozoic formations younger than the Hidden Valley Dolomite, which is Silurian and Early Devonian in age, extend into the map area, although Middle and Late Devonian and Early Mississippian formations remain farther east in the Funeral Mountains. The younger formations contributed to the Cenozoic sediments.

CENOZOIC ROCKS

Long after the Paleozoic seas withdrew from the region and after the record of Mesozoic events here was obliterated, Cenozoic rock materials accumulated in a continental environment of basins among rugged mountains. Torrential streams carried muddy gravel from the mountains to gentler slopes in dry basins or into the margins of lakes. The finer sediments were distributed more evenly in lakes that fluctuated for long intervals over the length of the area and farther southeastward. Lakes in the area during the Tertiary Period occasionally became saline enough to precipitate calcium carbonate in widespread but irregular beds, limited dolomite, some gypsum also in beds, and easily soluble salts dispersed in the fine sediments. The lakes left no beds of rock salt or more soluble evaporites here although some accumulated in lower basins beyond. While Paleozoic rocks of the Funeral Mountains were providing sediments from the north, volcanoes erupting in the Black Mountains and the Greenwater Range supplied much more material. This material was washed down or settled as volcanic ash directly into the lakes or remained around the volcanoes as tuff-breccia, solidified lava flows, and intrusions.

Artist Drive Formation

The Artist Drive Formation is the oldest Tertiary formation in the Furnace Creek area, its age extending back from early in the Pliocene Epoch to perhaps the Oligocene. The base of the Artist Drive remains covered in the area, but southward in the Black Mountains equivalent volcanic rocks rest on exposed Precambrian metamorphic rocks. The typical Artist Drive Formation, as designated originally by T. P. Thayer (*in* Noble, 1941, p. 955; oral communication in the field, 1954), stands out imposingly along the mountain front north of Artists Drive and surmounts the crest of the Black Mountains. The thickness of the formation, according to sections derived from the geologic map, exceeds 4,000 feet. The Artist Drive Formation is divided into five successive parts, or informal members. Two massive pyroclastic members of vivid colors separate three well-stratified sedimentary members.

The lower sedimentary member (Tal), which is nearly half the thickness of the Artist Drive Formation, consists of brown-weathering mudstone, sandstone, and conglomerate. The brown color generally ranges from light brown to grayish orange, but some is grayer and much is redder, as seen in the hills north of the side road to Ryan. Most of the clastic constituents are the sizes of sand, silt, and clay and are cemented by calcite. Calcareous sandstone at some places grades into sandy limestone. Pebbles and cobbles are scattered through some of the finer grained sedimentary rocks or concentrated into conglomerate lenses or beds that are rarely more than a few inches thick in the Black Mountains but as much as 400 feet thick north of Ryan. The pebbles and cobbles consist of Paleozoic limestone, dolomite, and quartzite; granitic rocks like those in the region's intrusions of quartz monzonite, syenite, and gabbro; and coarse-grained marble along with a few other contact-metamorphic rocks. No source for these is known in the Black Mountains or the Greenwater Range, where only Precambrian and Cenozoic rocks are visibly in place.

The lower pyroclastic member of the Artist Drive

Formation (Tapl) is a massive unit of tuff-breccia about 500 feet thick. The fragmental volcanic rock is well lithified, at least partly by alteration to the zeolite clinoptilolite. The most distinctive aspect of the member is green and buff coloration. The green is pale blue green approaching very pale green in the isolated hills south of the Dantes View road junction or grayish yellow green near the northwestern end of the Black Mountains. The buff part, lying above the green or distributed irregularly through it, is very pale orange or grayish orange.

The middle sedimentary member (Tam), about 600 feet thick near the northwestern end of the Black Mountains, is generally an olive color or a light brown; the olive ranges from pale olive and greenish gray to yellowish gray and light greenish gray. The unit consists of mudstone, sandstone, and conglomerate, containing at some places basalt as fragments in a calcareous mudstone or as thin flows or sills. The finer grained sedimentary rocks contain much volcanic tuff, whereas the conglomerate has the same kinds of nonvolcanic constituents as the conglomerate in the lower sedimentary member.

The upper pyroclastic member (Tapu) consists of massive tuff-breccia about 350 feet thick. The lowest part is pale blue green, resembling the lower pyroclastic member, but most of it is a distinctive grayish pink. It grades upward into very pale orange and is somewhat browner at the top. The pink and green pyroclastics, contrasting with dark basalt in jumbled masses of the member, compose the Artists Palette, a scenic point on Artists Drive.

The upper sedimentary member (Tau) comprises well-stratified rocks, about 800 feet thick, that are the most varied in color and lithology of the Artist Drive. Brown, in many variations, is the commonest color of the weathered rocks and is uniformly darker in the lower part, which contains conglomerate and basalt. But greenish and reddish colors of volcanic detritus in freshly exposed lacustrine sandstone and mudstone are increasingly conspicuous upward. The volcanic constituents of some of the lake beds grade laterally southeastward into coarse pyroclastics south of Corkscrew Canyon.

Felsite and basaltic rocks recur as flows, sills, dikes, and volcanic necks in different parts of the Artist Drive Formation. The felsite (Taf) is generally light brownish gray and has flow laminae. The basaltic rocks (Tab) are darker gray, commonly tinged green or olive. Some of the basalt is conspicuously porphyritic from large phenocrysts of calcic plagioclase, but much of it is finer grained and indistinguishable from basalt in the Furnace Creek Formation.

The age of most of the Artist Drive Formation has not yet been clarified by fossils. The only fossils known from the lower part of the Artist Drive are fish (Noble, 1941, p. 955) that remain undiagnostic of the age within the Tertiary. Previous assignment of an Oligocene age by the tentative correlation (Noble, 1941, p. 956) with the Titus Canyon Formation of Stock and Bode (1935), which 25 miles north of the Black Mountains contains Oligocene mammal fossils, is tenuous. Nevertheless, an Oligocene age of the lower part of the Artist Drive is still possible.

The terms Oligocene, Miocene, and Pliocene in the text and explanation of the map follow the usage with the North American mammalian stages and ages. The Pliocene Epoch by this usage arbitrarily contains the Clarendonian, Hemphillian, and Blancan Ages respectively in the early, middle, and late parts.

The first direct clue to the age of any part of the Artist Drive Formation, from fossils collected in the area, is contributed by Kenneth E. Lohman. Along with Thomas P. Thayer in 1938, he collected samples 150 feet below the top of the Artist Drive Formation, at an outcrop 5,000 feet in a straight line S. 4° W. from the west junction of the Twenty Mule Team Canyon road. On examining the samples to assist the borate project, he found diatoms, the first from the Death Valley area. All the identified forms in the scrappy assemblage of highly altered diatoms, according to Lohman's written communications (1954, 1961, 1967, on USGS diatom loc. 3967), are still represented in living assemblages elsewhere, with one exception, a species known in late Miocene beds in Humboldt County, Nevada, but dominant in an early Pliocene (Clarendonian) formation in Nye County, Nevada. Two of the genera, which he has not found in strata older than early Pliocene, have a known geologic range of early Pliocene to Holocene. Therefore, according to Lohman, the best age assignment for the assemblage of 11 diatom genera from the uppermost part of the Artist Drive Formation is early Pliocene (Clarendonian).

Furnace Creek Formation

The base of the Furnace Creek Formation is marked by a conglomerate member (Tfc) composed of detritus from the Artist Drive Formation in the southwestern part of the area and from Paleozoic rocks in the Funreal Mountains. The basal conglomerate lies concordantly on the uppermost member of the Artist Drive in the Black Mountains, but a few exposures in the valley west of Ryan show the conglomerate on the middle member or the lower pyroclastic member of the Artist Drive. In the Funeral Mountains, the basal conglomerate rises stratigraphically as the Furnace Creek Formation laps higher over the Paleozoic rocks of the basin's side. The conglomerate unit in a measured section across Twenty Mule Team Canyon is nearly 400 feet thick in 3,400 feet of incomplete Furnace Creek Formation. The total thickness of sedimentary rocks in the formation, as derived from structure section B-B′, is about 7,000 feet.

The most characteristic, borate-bearing part of the Furnace Creek Formation, as exposed along Furnace Creek Wash from the floor of Death Valley to the Greenwater Range, is conspicuously made up of light-colored fine-grained rocks derived from lake sediments and darker, more variegated, coarser rocks derived from stream sediments. The sedimentary rocks end southeastward along the flank of the Black Mountains, where they interfinger abruptly with volcanic rocks. These pyroclastic rocks and flows of vitrophyric and basaltic rocks are retained, in accordance with early definitions (Curry *in* Axelrod, 1940, p. 527-528; Thayer *in* Noble, 1941, p. 956), as a major part of the Furnace Creek Formation. Three kinds of volcanic rocks, which recur in the sequence, are distinguished as map units.

Pale volcanic rocks in one of the map units (Tfp) are mostly pyroclastic but include some vitrophyre and vitrophyre breccia. Thick, well-lithified pyroclastic masses of pumice, vitrophyre, and felsite show no internal layering of the tuff-breccia, unlike some of the thinnest sheets that contain stratified volcanic conglomerate and sandstone deposited by streams between basalt flows. The color, although variegated in detail, effectively is greenish gray or pale orange and grades into lighter tints. The vitrophyre is gray, except where altered to pale red or virtually white, and is much less conspicuous than the other volcanic rocks on the northeastern flank of the Black Mountains.

Basaltic rocks (Tfb) in flows and intrusions make up the darkest masses in the Black Mountains and extend as layers into the light-colored sedimentary part of the Furnace Creek Formation. Close inspection shows that the gray color is commonly tinged green or olive from some alteration of the dark minerals, and many vesicles are lined if not filled with zeolites, calcite, or borate minerals at some places. The basalt contains generally calcic plagioclase and varied proportions of clinopyroxene, generally much altered olivine accompanied at places by extraneous quartz, and magnetite. Some of the basalt in contact with lacustrine mudstone is altered and veined by analcime. The map unit (Tfb) includes minor reddish scoria and gray pyroclastic basalt closely related to the ancient vents.

Another map unit (Tfa) consists of greatly altered fragmented basalt in large masses transgressing the mudstone sequence of the Furnace Creek Formation or in layers extending concordantly into it. Internal stratification, where discernible, is indistinct, and the unsorted or poorly sorted fragments range commonly from sand to pebble sizes. The rock in outcrops is distinctively colored pale olive to light olive gray or dusky yellow and is poorly coherent. It now consists of analcime and montmorillonite, according to X-ray diffractometer charts, but a few pieces of fresher basalt remain scattered through some of the masses. The fragmented basalt, probably from explosive eruptions through a saline lake, filled small vents and spread over the surrounding lake bottom. It contains basalt dikes, some of which are not distinguished on the map, and local concentrations of gypsum (including very coarse selenite) or borate veins.

Lacustrine mudstone and sandstone are prevalent in the main part of the Furnace Creek Formation (Tf). The light colors of these well-stratified rocks, ordinarily yellowish gray to light greenish gray or very pale orange in outcrops and shallow mine workings, have changed from darker colors, ranging from dark greenish gray to dark gray, seen in some deeper mines and drill cores. The constituents are clay minerals (montmorillonite and illite, in samples examined by John B. Droste, written communication, 1960) and predominantly volcanic debris cemented by calcite. Salines dispersed through mudstone are apparent to the taste and as efflorescent accumulations under the mud crusts produced by weathering.

A few beds of tuff, limestone, and minor dolomite, along with some conglomeratic or gypsiferous beds that are not distinguished from the main unit, are interstratified with the mudstone and sandstone. The tuff, in extensive beds as much as two feet thick but mostly a few inches thick, lacks calcite cement and is generally altered to clinoptilolite. A conspicuous color of the tuff is very pale blue green, but most of the tuff is very light gray tinged pinkish or yellowish and nearly white. Limestone in discontinuous beds grades from the cleanest limestone, showing algal structure in laminated beds and knobby columns, to marlstone and calcareous sandstone. Some of the thinly stratified limestone was broken into flat-pebble conglomerate. The rare dolomite is very fine grained, clay-bearing, and nearly white. No beds of salines are exposed. The principal borate deposits are widely distributed in the stratigraphically lowest part of the main unit and in some of the underlying muddy or calcareous conglomerate.

Gypsiferous beds are mapped as a member (Tfg), generally 100 to 200 feet thick, within the main part of the Furnace Creek Formation. The gypsiferous member forms the crest of the ridge between Twenty Mule Team Canyon and Furnace Creek Wash and continues westward, although interrupted by faults, along the highest ridge between Zabriskie Point and Gower Gulch to the floor of Death Valley. The member as mapped in widely separate places, such as in the East Coleman Hills, west and north of Navel Spring, and along Furnace Creek Wash northwest of Ryan, is perhaps not at the same stratigraphic horizon. Rough or knobby limestone at some places is closely associated with the top of the gypsiferous beds. Thin beds of granular gypsum, recrystallized to many sizes of grains and retaining some granular anhydrite, are interlayered with little or much mudstone in the member. Veins of fibrous gypsum are abundant. The gypsiferous member contains small quantities of borates widely distributed in the area.

The upper member of the Furnace Creek Formation (Tfu) is marked by distinctive pumiceous tuff beds overlain by light-colored mudstone. The member is generally about 350 feet thick, but thickens to as much as 800 feet. The white or pale orange-tinted tuff is mostly fine grained (sand and finer sizes) but at many places contains pebble-size fragments of pumice. Despite varied amounts of calcite cement, it is generally softer than the tuff beds in the lower part of the formation. Some thin beds of tuff are irregularly opalized at the eastern end of the area, where a thermal-spring environment during sedimentation is indicated by certain diatoms, according to K. E. Lohman (written communications, 1967), and by traces of certain elements in the locally manganiferous beds, according to D. F. Hewett (oral communications, 1967). The mudstone, which contains some swelling clay and a few thin beds of pale limestone and dolomite, grades into beds of sandstone and conglomerate.

Upper conglomerate members of the Furnace Creek Formation (Tfcu) intertongue with the finer grained rocks through much of the sedimentary sequence of the formation above the basal conglomerate, but most of the large masses are in the upper part. The occurrence of conglomerate is not as clear-cut as shown by the map; for example, gravelly mudstone is abundant between the conglomerate members mapped on the Death Valley side of the Black Mountains, and fine-grained sediments are abundant in the conglomerate member southeast of Zabriskie Point. Nevertheless, the conglomerate member as outlined from Texas Spring Camp Ground southeastward past Zabriskie Point depicts a tilted alluvial fan extending northeastward into lake sediments that tongue into the fan. The conglomerate is a poorly sorted mixture of fragments that range in size from boulders to clay. The predominance of the coarser sizes commonly decreases gradually both vertically in the sequence and laterally. An outstanding exception is exposed in the Hole in the Wall, where coarse fluvial conglomerate abruptly intertongues with lacustrine mudstone. The conglomerate here continues upward to the top of the Furnace Creek, where it takes the place of the upper finer grained member northeast of Navel Spring and again southwest of Travertine Point. The conglomerate along the Funeral Mountains was derived from pre-Tertiary sedimentary rocks, whereas the conglomerate along the Black Mountains was derived from Tertiary volcanic and sedimentary rocks, including the great variety of pre-Tertiary rocks reworked from conglomerate in the Artist Drive Formation.

The age of the Furnace Creek Formation, according to K. E. Lohman's diagnosis (written communications, 1961, 1967) of his diatom collections in terms of the North American mammalian chronology, is clearly middle Pliocene (Hemphillian) for the uppermost part of the formation and less definitely early Pliocene for the lower part. His lower collection (USGS diatom loc. 4159, foot of the Black Mountains 2.56 miles in a straight line S. 36° W. from Navel Spring) contains one extinct species having a known range of late middle Miocene to early Pliocene and two genera having ranges in nonmarine sediments from early Pliocene to Holocene; the combination thus suggests for the lower part an early Pliocene age (K. E. Lohman, written communication, 1967). The only significant fossil previously reported from the Furnace Creek Formation was found in about the same part of the formation at a locality for leaf fragments discovered by H. D. Curry (Axelrod, 1940, p. 526), almost 3 miles north-northwest of the diatom locality. From the impression of one diagnostic leaf fragment, Axelrod (1940, p. 531) derives an age probably not older than late Miocene or younger than early Pliocene. Mammal tracks in beds correlated with the Furnace Creek Formation, but isolated in Copper Canyon 10 miles south of the area, were found by Curry (1941), who states, ". . . preliminary study of the size of the horses as well as other diagnostic features of the fauna [as indicated by the footprints] suggests that the age of the Copper Canyon beds is middle Pliocene." Lohman's diatom collections from the uppermost part of the Furnace Creek (locs. 4070 and 4357, in the hills along the south side of California Highway 190 between Travertine Point and the east boundary of Death Valley National Monument) contain abundant and well-preserved assemblages extraordinarily similar to a diatom assemblage from the San Pedro Valley, Arizona, beds that have yielded a vertebrate fauna of middle Pliocene (Hemphillian) age (John Lance according to Lohman, written communications, 1967). Lohman concludes: "Based upon the known geologic ranges of many of the diatom species in the assemblages from localities 4070 and 4357, plus the very close similarity to the well-dated San Pedro Valley material, an age assignment of middle Pliocene can be made with considerable confidence."

Funeral Formation

The Funeral Formation of conglomeratic and basaltic rocks lies on the most conspicuous angular unconformity within the Tertiary sequence in the area. Basalt flows capping the Greenwater Range from Travertine Point southward past Ryan lie across the deformed and eroded Furnace Creek and Artist Drive Formations. The angularity at the contact disappears northwestward where conglomerate and pebbly or sandy mudstone of the Funeral were deposited along a trough in the Furnace Creek Formation. The thickness of the sedimentary part of the formation in structure sections near Death Valley is between 1,000 and 1,500 feet, whereas between Navel Spring and Travertine Point it is about 700 feet. The thickest exposure of volcanic flows, a mile east of Ryan, is about 500 feet.

The characteristic sedimentary part of the Funeral Formation (QTf), as exposed between Travertine Point and Navel Spring, consists of poorly sorted, indistinctly stratified conglomerate of pebbles, cobbles, and boulders in a mud and sand matrix. It is lithified sufficiently to stand in cliffs, particularly where supported by spring travertine and vertical calcite veins. The constituents here are from the Furnace Creek Formation and Paleozoic rocks in the Funeral Mountains, whereas along the road to Dantes View they are entirely of volcanic rocks—basalt from the Funeral Formation in the Greenwater Range and silicic volcanics from the Black Mountains.

A lower member (QTfl) consists mainly of gravelly sandstone and mudstone that grades into conglomerate or sandy mudstone in the northwestern part of the area near Death Valley and attains a thickness of a thousand feet, as shown in structure sections. It contains some isolated, very light colored limestone deposited by springs. The member intertongues into conglomerate of the upper Funeral, as readily seen around the lookout hill northwest of Texas Spring. Sandy mudstone in the lower member is distinguished from underlying Furnace Creek Formation by the Funeral's pinkish, brownish, and grayer tinges. At the eastern end of the area (north and east of Travertine Point), well-lithified sedimentary breccia (QTfx) and conglomerate, partly interlayered between giant landslide blocks of Paleozoic formations, are assigned somewhat questionably to the lower member of the Funeral. Similar jumbles of blocks, or megabreccias, extend into the Furnace Creek Formation around the Red Amphitheater and across the deformed Furnace Creek Formation farther northwest along the foot of the Funeral Mountains. The main conglomerate of the Funeral, in which basalt flows end, laps over the breccias near Travertine Point.

Lava flows in the Funeral Formation consist of olivine-phenocryst basalt (QTfb), which is fresh gray compared with the greenish-tinged gray of most of the basalt in older formations and lacks the conspicuous vesicle fillings. An exception exposed at the highway 1½ miles west of Travertine Point is greenish-gray altered basalt in a thin layer that extends into conglomerate of the Funeral. The usual color is medium to dark gray or brownish gray, made darker by a film of desert varnish after long exposure. A characteristic sample of the fresh basalt, collected about a mile southeast of Travertine Point, is microcrystalline except for about 2 percent olivine phenocrysts, commonly 0.5 to 1 mm in diameter, ten times larger than the olivine and clinopyroxene grains in the groundmass. The mineral composition is 63 percent plagioclase (labradorite), 30 percent total olivine and clinopyroxene, and 7 percent magnetite.

Basaltic agglomerate (QTfa) centered at explosive vents of the basalt consists of red scoria, some volcanic bombs, and breccia. The agglomerate, where it was solidified by basalt permeating a volcanic throat or reinforced by basalt dikes, has resisted erosion and stands out in relief against the surrounding agglomerate and flows. Shapes that resemble small volcanoes are merely resistant cores. Thin layers of scoria and breccia, as well as basalt dikes in agglomerate, are not distinguished on the map.

Surficial Deposits

Deposits that conceal the Funeral or older formations and express to some degree their original topography are grouped here as surficial deposits. Boundaries between the units of the group, as well as between them and the older rocks, are derived from aerial photographs. Differences between some of the units are gradational and difficult to apply uniformly. Talus rubble is not shown separately, but where it effectively conceals geology, the rubble is shown as the alluvium with which it merges. Remnants of old alluvium are not distinguished from the underlying Funeral Formation midway between the Funeral Mountains and Death Valley at the northern end of the area, and none of the basalt alluvium and rubble veneering much of the basalt in the Greenwater Range is shown at the eastern end. Most of the mapped alluvium is a thin cover (a few tens of feet or less) on pediments under present stream channels and on pediments that were left as alluvium-covered terraces by faults along the edge of the Death Valley floor and by intrenchment.

Alluvium is mapped in two generalized units. The alluvium in each unit was deposited by streams during immensely different lengths of time. The older alluvium represents the vastly greater span. The older alluvium (Qao) upstream is generally much dissected in a series of terraces, the highest of which is as much as 500 feet above the nearby valley floor, yet some parts of the older alluvium downstream coincide with or are covered by the younger alluvium. The surface of the older alluvium has been smoothed by the breaking down of coarse material and the development of desert pavement; rock fragments on it, if susceptible to desert varnish, are dark. Near the surface in favorable places, calcite is concentrated into caliche. Younger alluvium (Qal) along recent channels consists of stream sediments having rough gravel at most of the surface, from which the finer material has been sieved by the waning flow of the infrequent torrents. The finer sizes are thus concentrated in some major channels and at the lower margins of the alluvial fans. The surficial gravel is fresh gray in the youngest channels and brown from desert varnish along abandoned courses.

Other surficial deposits distinguished on the map are the salines and fine-grained clastic sedimentary deposits (Qsl) on the floor of Death Valley; travertine (Qtr) crust deposited by former springs; and undifferentiated constituents of landslides (Qlds) that are related to the present topography.

STRUCTURE

Most of the Furnace Creek area is part of the Black Mountains tectonic block, a northwest-trending wedge between the Death Valley and Furnace Creek fault zones (Noble and Wright, 1954, p. 151). The northwestern corner of the area is in the Death Valley tectonic depression, where structural features are concealed by young sedimentary deposits of the valley floor. Northeast of the Furnace Creek fault zone, a marginal strip of the area lies in the Funeral Mountains. The structural features of the Funeral Mountains that are truncated by the fault zone have not been recognized in the Black Mountains block.

Displacements on the Furnace Creek and Death Valley fault zones since the Furnace Creek Formation was deposited have been downward on the southwest and Death Valley sides. A horizontal component of relative movement toward the right by the block on the far side of each fault zone was proposed by Curry (1938, p. 1874–1875) and advocated by later writers in syntheses of the regional structure. Features shown on the present Furnace Creek map are ambiguous concerning the amount of lateral displacement.

A sequence of fault movements on the Death Valley side of the Black Mountains continued after the last movement in the Furnace Creek fault zone. Stream terraces of older alluvium are truncated along the front of the Black Mountains, and even the alluvial fans at the foot of the mountains south of Furnace Creek retain segments of a fault scarp, which Hunt and Mabey (1966, p. A100) consider to be about 2,000 years old. In contrast, the surface of the oldest alluvium that caps the mesa above Navel Spring extends across the entire Furnace Creek fault zone but is not displaced by any of the faults.

Faults within the Black Mountains block generally increase in number and displacement on the Death Valley side, strike generally eastward in the part south of Golden Canyon, and show relative displacement downward on the north side of some and the south side of others. In the eastern part of the area, other faults strike north and have the major downthrow on the west. The only exposed fault that crosses the Furnace Creek valley, in the middle of the area, strikes northeast and displaces a stream terrace of older alluvium a few feet down on the west side. Previously it displaced the Funeral Formation the same way, perhaps a thousand feet. Northeastward the fault does not go beyond the first fracture in the Furnace Creek fault zone but probably curves northwestward into the fault zone. The valley-crossing fault in the opposite direction joins the fault that continues, with the same downthrow, southeastward along the foot of the mountains. Farther southeastward beyond concealing alluvium in front of the Black Mountains, the Grand View fault continues the alinement and the west downthrow. No other significant fault in the southeastern part of the area displaces the Funeral Formation. Near Ryan, those that conspicuously displace the Artist Drive and Furnace Creek Formations end at the angular unconformity beneath basalt of the Funeral Formation.

Folds are subordinate features of the structure. Even the broad syncline in the Furnace Creek valley is shaped by steepening of beds faulted against the Funeral Mountains block. The Hole in the Wall, accessible by poor road from California Highway 190, exposes the abrupt steepening from open minor folds in the axial zone of the syncline to vertical conglomerate strata dragged on the Wall fault. The open folds plunge about 15° southeast up the valley from the East Coleman Hills to the mesa at Navel Spring, where a syncline in the Funeral and the underlying Furnace Creek Formations is visible from the highway on the west. Between the Texas Spring Camp Ground and the Funeral Mountains, an asymmetric anticline shares its steeper limb with the complementary syncline on the southwestern side. The limb steepens along the strike northwestward to the East Coleman Hills and becomes vertical or locally overturned under a reverse fault. At the other end of the area, in the narrows of Furnace Creek Wash on the road to Dantes View, the steep limb of an asymmetric syncline in the Funeral Formation is the result of drag on the downthrown southwestern side of the Grand View fault. The extensive basalt of the Funeral Formation in the Greenwater Range is younger than the small folds, as well as the other faults, in the underlying formations. Sets of small, tight folds in the Furnace Creek Formation at widely separate places show local adjustment of incompetent beds to an abrupt increase in competence near a fault. The older alluvium is folded slightly into the syncline at Texas Spring, into an open syncline from drag on the valley-crossing fault, and into a low anticline in the hills at Mustard Canyon.

Significance of a Man-made Diversion of Furnace Creek Wash at Zabriskie Point, Death Valley, California

Bennie W. Troxel[1]

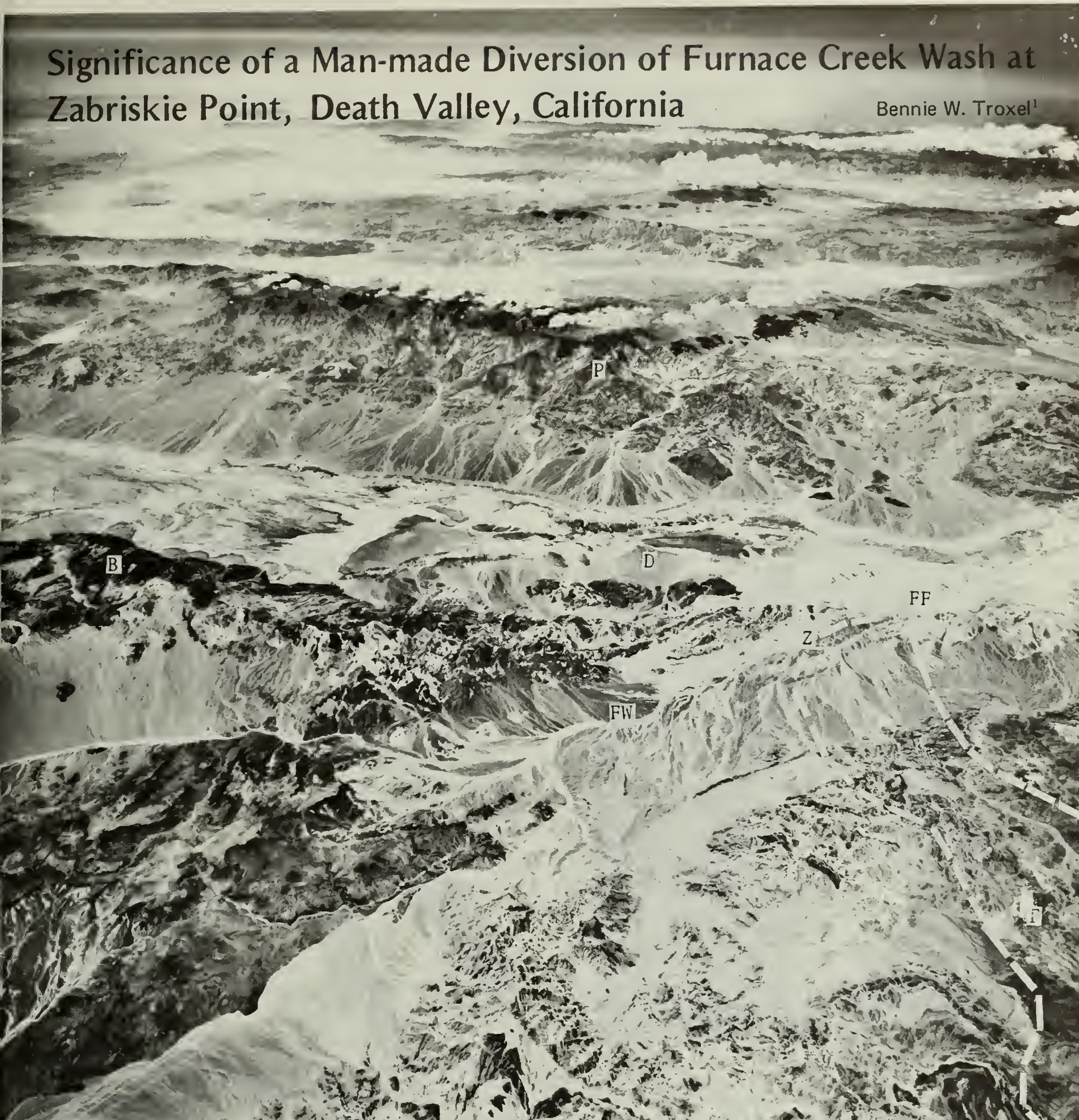


Figure 1. Oblique aerial photograph oriented west-southwest, showing most of drainage area of Furnace Creek Wash (foreground). Dashed line bounds approximate area of drainage onto Furnace Creek fan (FF) since the diversion. Zabriskie Point (Z), Black Mountains (B), Funeral Mountains (F), Death Valley (D), Panamint Range (P), Furnace Creek Wash (FW). Gower Gulch, too small to indicate, extends between (Z) and a very small fan on south (left) edge of Furnace Creek fan. Photo courtesy of U.S. Geological Survey - U.S. Air Force.

ABSTRACT

A man-made diversion of the flow of water and flood debris from Furnace Creek Wash into Gower Gulch at a point near Zabriskie Point has resulted in significant changes in the environment.

---

[1] *Geological Society of America, Boulder, Colorado 80301.*

Figure 2. Incised channel of Gower Gulch at the diversion point on southwest side of Furnace Creek Wash. Channel is about 15 ft deep at edge of wash in foreground.

Figure 3. A. Incised channel at mouth of Gower Gulch (view downstream). B. Steep-walled channel in Gower Gulch a few feet downstream from diversion point (view downstream). Rocks are steeply inclined fine-grained sediments of Furnace Creek Formation.

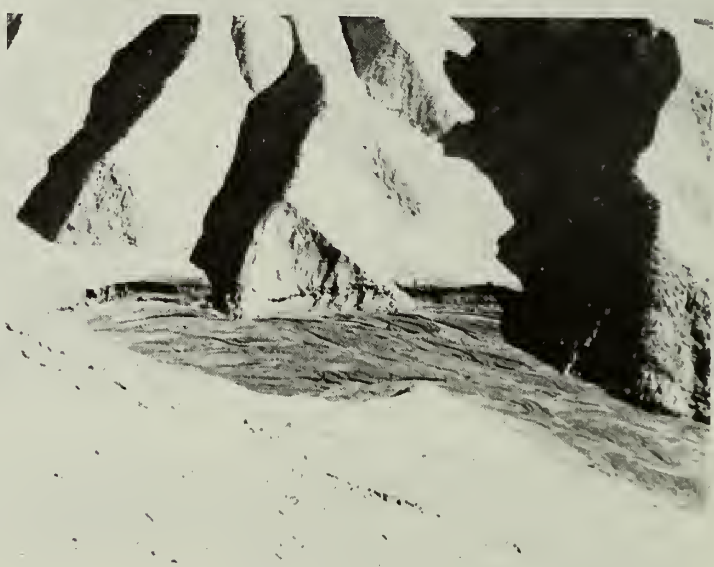


Figure 4. Hanging tributary channels in Gower Gulch at southeast base of Zabriskie Point.

INTRODUCTION

Furnace Creek Wash enters the floor of Death Valley at the head of a large alluvial fan near Furnace Creek Inn, a resort at the north tip of the Black Mountains. Until a major portion of the surface flow from the normally dry creek was diverted into Gower Gulch three decades ago, it drained some 200 sq mi of land (Fig. 1) that ranges in elevation from sea level, at Furnace Creek Inn, to 6,700 ft in the Funeral Mountains. Gower Gulch originally drained about 2 sq mi. Annual precipitation ranges from probably more than 10 in. in the higher area to about 1.8 to 2 in. at Furnace Creek Inn. Much of the precipitation falls during short periods of heavy rainfall.

Most of the mountainous part of the drainage area contains barren rock that is moderately impermeable and readily produces runoff during storms. The channels in the canyons and washes contain coarse gravel that is usually several feet to a few tens of feet thick and nearly everywhere is dry and loose. Slopes of many of the mountains contain appreciable amounts of debris that is easily eroded.

The combination of occasional rapid, heavy runoff, abundant loose material, and moderately steep gradients in the stream beds affords a high potential for erosion and debris transport during the heavy flows of runoff events.

Diversion of Furnace Creek into Gower Gulch reportedly was made in 1941 as a flood-control measure. It was accomplished by making a small cut into the soft, fine-grained sediments of the Pliocene Furnace Creek Formation, which crop out along the low, narrow divide along the southwest side of Furnace Creek Wash at the upstream base of Zabriskie Point. A gravel barrier was placed diagonally across Furnace Creek Wash at the downstream side of the cut, and waters from the wash were thus diverted into a tributary channel of Gower Gulch.

Before the diversion, Furnace Creek Wash carried intermittent flood waters to the floor of Death Valley in a stream system that appeared to be in balance. By gradual growth, the fan at the mouth of the wash accommodated the material that was being carried along the stream, and the channel system was well integrated so that no areas of unusual erosion or deposition were obvious in the drainage system. Tributary streams entered the wash at grade throughout the system except in local areas where recent flash floods had downcut a local segment of the main channel, thus leaving the channels of tributary streams temporarily at a higher level than the main channel.

Gower Gulch, too, in spite of its much smaller size, was apparently in balance throughout its channel, which flows onto a very small alluvial fan in Death Valley.

Gower Gulch and Furnace Creek Wash have the same ultimate base level (−282 ft at Badwater), but because of the elevation of the head of the fans, Gower Gulch debouches onto the apex of its fan at about −100 ft elevation, and Furnace Creek drains onto its fan at about sea level. The channel of Gower Gulch is a few tens of feet lower in elevation than Furnace Creek Wash at Zabriskie Point. The large, through-going Furnace Creek channel was at grade at a higher elevation than the small Gower Gulch. The result of a diversion from a higher and larger channel into the lower channel resulted essentially in the development of the equivalent of a waterfall that concentrated strong erosive force on the point of diversion (Fig. 2).

Thus the diversion at Zabriskie Point was done by man in a setting where stream capture would occur within a geologically short time, and the system was ready to respond immediately and strongly to the change. The magnitude of the changes in stream regimen is a function of the size of floods and their frequency. Several floods have occurred since the diversion, some with peak flows of several thousand cubic feet per second.

EFFECTS

By abruptly increasing its drainage area from about 2 sq mi at low altitude to about 170 sq mi at altitudes as great as 6,700 ft, the diversion has placed an extraordinary burden on Gower Gulch. The effect of this sudden influx of large volumes of water and debris during flood periods has, in only the 30 yr or so since the diversion, resulted in significant erosion and deposition. The erosion appears to be accelerating, but this may be illusionary because of the occurrence of some floods of larger-than-usual magnitude during the past few years. For example, during July 1968, a local storm centered in the southern part of the Furnace Creek drainage produced a peak flow of 7,000 to 10,000 cu ft per sec, which deepened the channel at the diversion several feet and caused appreciable scour (Fig. 3) in the channel.

The floor of Gower Gulch has been lowered (Fig. 4) and widened by the accelerated erosion; the fan at the mouth of Gower Gulch has been incised 10 to 15 ft (Fig. 5) in what appears to be an extension of the stream channel across the fan (none of the adjacent fans shows such deep channeling). gravel from Furnace Creek Wash now coats the channel of Gower Gulch (Fig. 4); tributary stream channels in Gower Gulch are hanging 1 or more ft above the level of the main channel (Fig. 4); lateral erosion of stream walls in Gower Gulch is accelerating by collapse after they have been undercut; and the lower reach of the Gower Gulch fan is growing more rapidly because of the increased volume of debris now being contributed to it.

Furnace Creek Wash. Furnace Creek Wash has become deeply incised at the point of diversion (Fig. 6) and headward erosion has migrated 1.7 mi upstream, as evidenced by hanging tributary stream channels; headward erosion is cutting a new drainage system into the floor of Furnace Creek Wash (Fig. 6); headward erosion is vigorously attacking State Highway 190 at a point across Furnace Creek Wash from the diversion (Fig. 6A), and at several points for some distance upstream.

The segment of Furnace Creek Wash below Zabriskie Point is receiving much less flood water, and as a result, the channel and fan will receive less debris.

Ground Water. The diversion is conceivably causing adjustments in ground-water storage and movement but the changes are apt to be slow and certainly cannot be detected without careful monitoring.

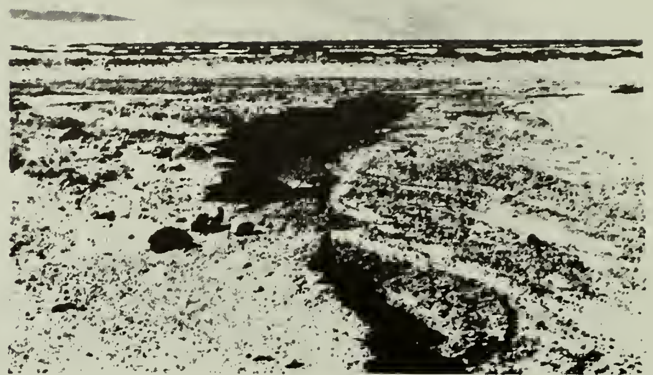
One obvious change that probably is occurring because of the loss of 80 percent of the drainage area is the reduction of recharge to the Furnace Creek fan by seepage losses during floods and by postflood underground flow in the lower part of Furnace Creek Wash. If the recharge is reduced significantly, the water table beneath the fan will be lowered. This could reduce the water available for plants in the wash and near the toe of the fan and some plants would die. Without recharge, salt water beneath Death Valley may intrude the fresh ground water that underlies the Furnace Creek fan.

More surface water will be lost to evaporation on the saline floor of Death Valley. The shorter channel, moderately impervious rocks in Gower Gulch, and lack of a large fan in which to spread the water near the mountain front are other significant deterrents to ground-water recharge and storage.

SUMMARY

The diversion of Furnace Creek Wash into Gower Gulch at Zabriskie Point has made significant changes in the environment. However, the magnitude of these changes can be assessed only by careful measurement of the changes to date and a program of monitoring future changes. The principal benefit of the diversion has been to lessen flood damage to (1) the high-

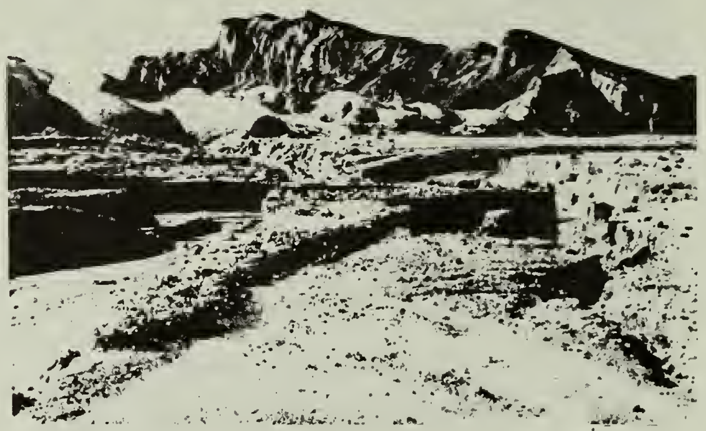
A.



B.

Figure 5. Incised channel on fan below mouth of Gower Gulch. Undercut and vertical walls collapse into channel during intervals between floods. A. View upstream (east). B. View downstream (west).

A.



B.

C.

Figure 6. Gullies cut into gravel of Furnace Creek Wash. A. View from highway toward diversion point. Edge of roadbed is actively eroded each time running water flows through. B. View downstream from center of wash toward diversion point (left of center). C. View upstream from northeast side of wash. Wash is incised 1.7 mi upstream from diversion point (to right of photo edge).

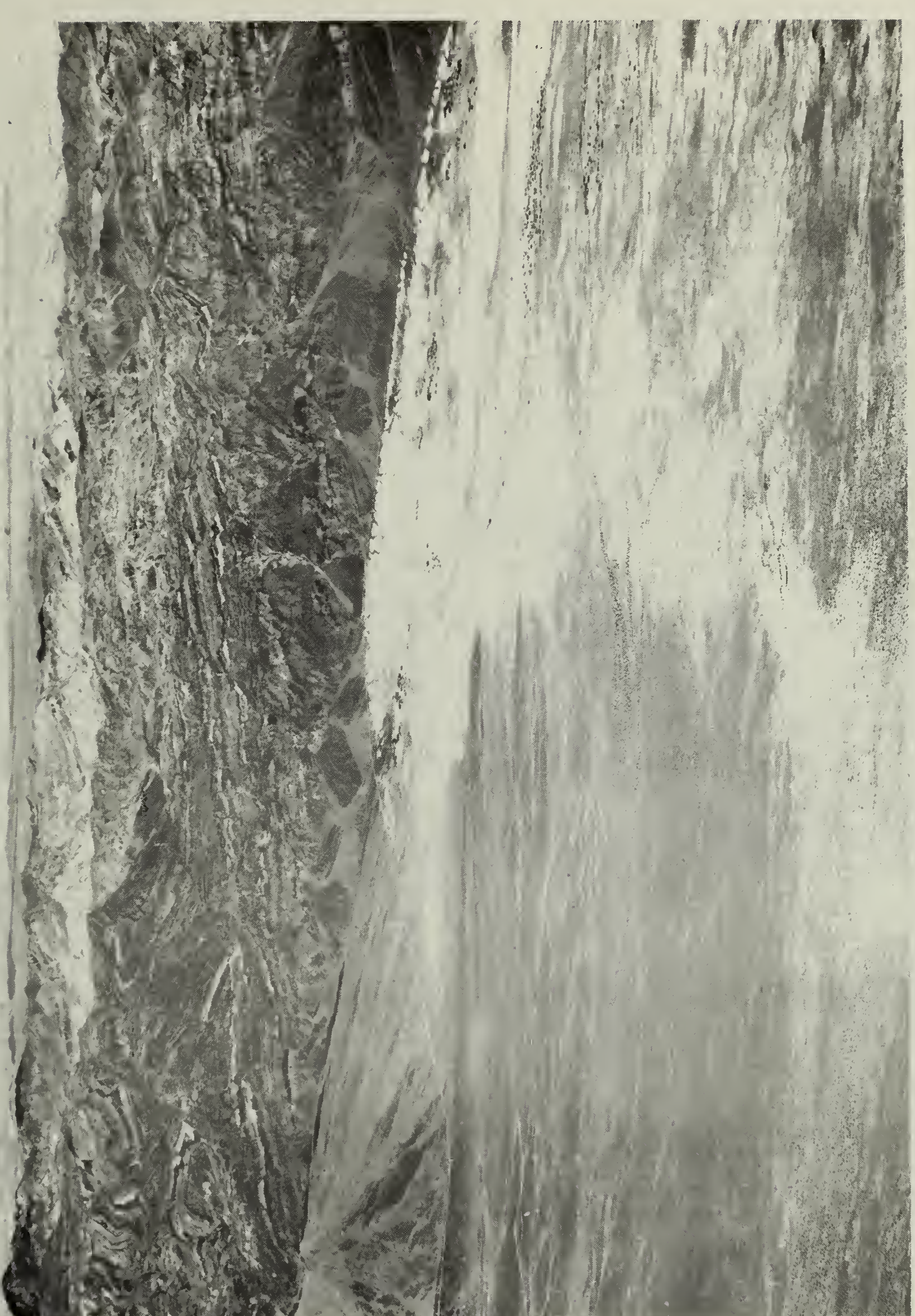
way in Furnace Creek Wash downstream from Zabriskie Point, (2) the water-supply collection and conveyance system from springs in the wash to Furnace Creek Inn and to users on Furnace Creek fan, and (3) to structures at the Inn and on the fan. To some degree the benefit is offset by jeopardizing a long segment of the highway upstream from the diversion point and a short segment of the paved highway from Furnace Creek to Badwater where it crosses Gower Gulch in Death Valley. The diminished ground-water recharge, intrusion of salt water, and transfer of flood damage from the Furnace Creek fan to the Gower Gulch fan are factors that are less easy to predict, measure, or evaluate.

What, in hindsight, appears to have been an ill-conceived action to remedy a local problem may cost much to rectify. The cost of plugging the diversion point (Fig. 2) so that the flow is returned to the original channel and thus to a more balanced condition would seem to be a small price to pay if it proved to be the best corrective measure. Upper Furnace Creek Wash probably would return to near its pre-diversion condition within several years or a few decades. The incised channel and secondary fan on the Gower Gulch fan will retain evidence of the diversion for a much longer time. If the diversion remains permanent, should the name Gower Gulch give way to Furnace Creek Wash or vice versa? Should lower Furnace Creek Wash be renamed Echo Wash?

ACKNOWLEDGMENT

I thank Glen Miller and Peter Sanchez for occasional interesting discussions over a period of years and the late Charles Hansen for corresponding with me about the matter.

Preprinted by permission from paper submitted for publication in *California Geology*, California Division of Mines and Geology. Subject to revision before final printing.



Low-altitude oblique aerial photograph across Death Valley northeast of Grapevine Mountains between mouth of Titus Canyon (left) and Titanothere Canyon (right). Pale rocks just below skyline are Tertiary rocks, mostly volcanic. Stratified rocks in Grapevine Mountains range in age from early Cambrian (right) to late Paleozoic (left) and are highly deformed. See article that follows. *Photo W16 by John H. Maxson; courtesy of the National Park Service.*

Geology of the Grapevine Mountains, Death Valley, California: A Summary

Mitchell W. Reynolds[1]

INTRODUCTION

The Grapevine Mountains, limited on the south by Boundary Canyon and on the north by Grapevine Canyon, form the eastern wall of northern Death Valley (Fig. 1). The mountains are rugged and have poor accessibility, but their high relief and barrenness provide excellent exposures of the complicated geology characteristic of the Death Valley region. Upper Precambrian and Paleozoic rocks exposed in the core of the mountains are the most northerly complete section of miogeosynclinal facies strata on the east side of the northern Death Valley–Furnace Creek fault zone. In the core of the mountains, a thrust fault of early or medial Mesozoic age juxtaposes upper Precambrian, Cambrian, and Ordovician rocks over middle and upper Paleozoic strata. Cenozoic rocks flank the core and include both the oldest dated Tertiary strata in the southwestern part of the Great Basin and an extensive succession of younger sedimentary and volcanic rocks. These and older strata are broken by late Cenozoic high-angle normal faults that flatten with depth east and west off the core of the range. In the southern part of the mountain core, folding and faulting of late Tertiary age produced a north-trending S-shaped fold system composed of a recumbent syncline and anticline. A low-angle fault, along which rocks are rotated and extended, replaces the upright limb of the anticline, and the entire fold system is separated from underlying metamorphosed Precambrian rocks by a low-angle fault that also displaces Tertiary strata at shallow depths. Strike-slip faults of late Cenozoic age displace rocks and structures in the range, as well as Quaternary deposits in adjacent Death Valley.

In this summary of the geology of the Grapevine Mountains, emphasis is on the southern part of the mountains, since this area is accessible by road through Boundary and Titus Canyons. The geology of the Grapevine Mountains has been mapped largely by me; mapping by Cornwall and Kleinhampl (1964) in the southeasternmost part of the mountains and by L. A. Wright and B. W. Troxel (1971, written commun.) along the southern edge has provided a base for more detailed mapping and structural interpretations of those parts by me.

STRATIGRAPHY

Nearly 9,100 m (30,000 ft) of pre-Mesozoic and Cenozoic sedimentary and volcanic rocks are exposed in the Grapevine Mountains (Fig. 2). For about 9.6 km (6 mi) southeast of Boundary Canyon to Chloride Cliff, the oldest rocks crop out in the anticlinal core of the Funeral Mountains (Wright and Troxel, 1970; 1971, written commun.). These rocks belong to the Pahrump Group of Precambrian age. Northwestward, a nearly continuous succession of pre-Mesozoic rocks is preserved down structural plunge from Chloride Cliff to the north end of the Grapevine Mountains, where strata of Pennsylvanian age crop out. The lowest quarter of the succession, widely exposed in Boundary Canyon, is a thick sequence of terrigenous clastic rocks, whereas much of the remainder of the section is composed of carbonate rock with only thin, interbedded terrigenous units. Some of the Precambrian, Cambrian, and Mississippian rocks may have accumulated in moderately deep marine waters, but the carbonate and terrigenous rocks were generally deposited in shallow-marine and intertidal environments on a slowly subsiding shelf. Siltstone, sandstone, and conglomerate beds among upper Paleozoic rocks were derived from uplands presumably to the northwest and north resulting from the Antler orogeny.

Terrigenous strata in the pre-Mesozoic sequence have generally been the loci of failure during deformation. Preferred units of failure are designated by asterisks in Figure 2. The Stirling and Zabriskie Quartzites, for example, controlled positions of faulting during both Mesozoic and Cenozoic deformation: these formations form the sole of the allochthon of the major thrust fault of Mesozoic age in the northern part of the Grapevine Mountains, and form the soles of flat faults of Cenozoic age in Boundary Canyon. Recumbent folding, probably late Tertiary in age, in the southern Grapevine Mountains involved rocks only as old as the Stirling Quartzite, which acted as a sole for detachment above metamorphosed and deformed older rocks. High-angle faults of Cenozoic age are commonly flattened with depth to follow siltstone and quartzite beds in Paleozoic strata.

Cenozoic rocks crop out widely along the east flank and the northwest front of the Grapevine Mountains and locally near Death Valley Buttes, west of the mouth of Boundary Canyon. Rocks on the northwest front differ in origin and age from those on the east and south. At the latter localities, strata of fluvial and lacustrine origin belonging to the Titus Canyon Formation of Oligocene and Miocene(?) age (Reynolds, 1969, 1974) rest unconformably on Paleozoic rocks and are, in turn, overlain unconformably by ash-flow tuff, lava flows, and sedimentary rocks of lacustrine and fluvial origin. This succession of volcanic and sedimentary rocks is about 22 to 20 m.y. old, and sources for most of the volcanic rocks lay within the Nevada Test Site east and northeast of the Grapevine Mountains.

Cenozoic rocks are exposed along the mountain front in the northern part of the Grapevine Mountains northwest of Fall Canyon and at the southern end of the mountains in the Kit Fox Hills (Fig. 1). Rocks northwest of Fall Canyon are divided into three parts—the lowest exposed part and the upper part are dominantly of fluvial origin, and rocks of the middle part are of lacustrine origin. The lower fluvial and lacustrine rocks are conformable, but the upper fluvial sequence lies unconformably on the lacustrine rocks. Thin, coarse basaltic intrusions occur in the lowest two units. Clasts in the lower fluvial sequence were derived from granitic intrusive rocks exposed at the north end and west of Death Valley; clasts of welded ash-flow tuff were eroded from units northeast and east in the Grapevine Mountains and beyond, but clasts of locally derived pre-Mesozoic rocks are not abundant. By contrast, the upper fluvial rocks are composed mainly of fragments eroded from pre-Mesozoic rocks in the Grapevine Mountains, onto which the fluvial strata lap or are faulted. The lower two units are here correlated with the Furnace Creek Formation as defined by McAllister (1970) and the upper unit with the Funeral Formation, both in the central part of Death Valley. Cenozoic lacustrine and fluvial rocks of the Kit Fox Hills west of Death Valley Buttes (Fig. 1) contain clasts derived from ash-flow tuff units of Miocene and early Pliocene age exposed east of Death Valley (Fig. 2), and hence are considered to be

[1] *11780 Swadley Drive, Lakewood, Colorado 80215.*

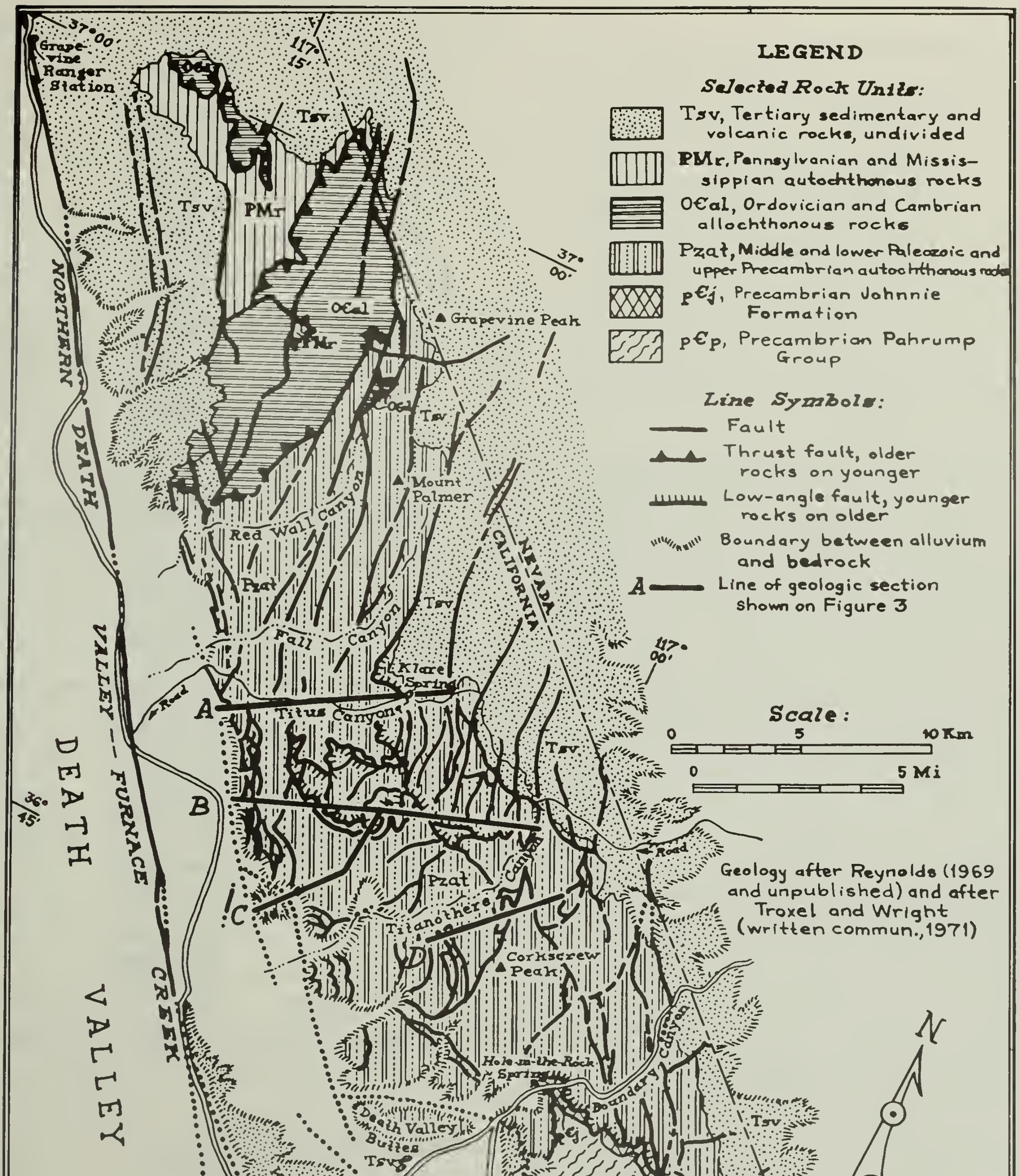


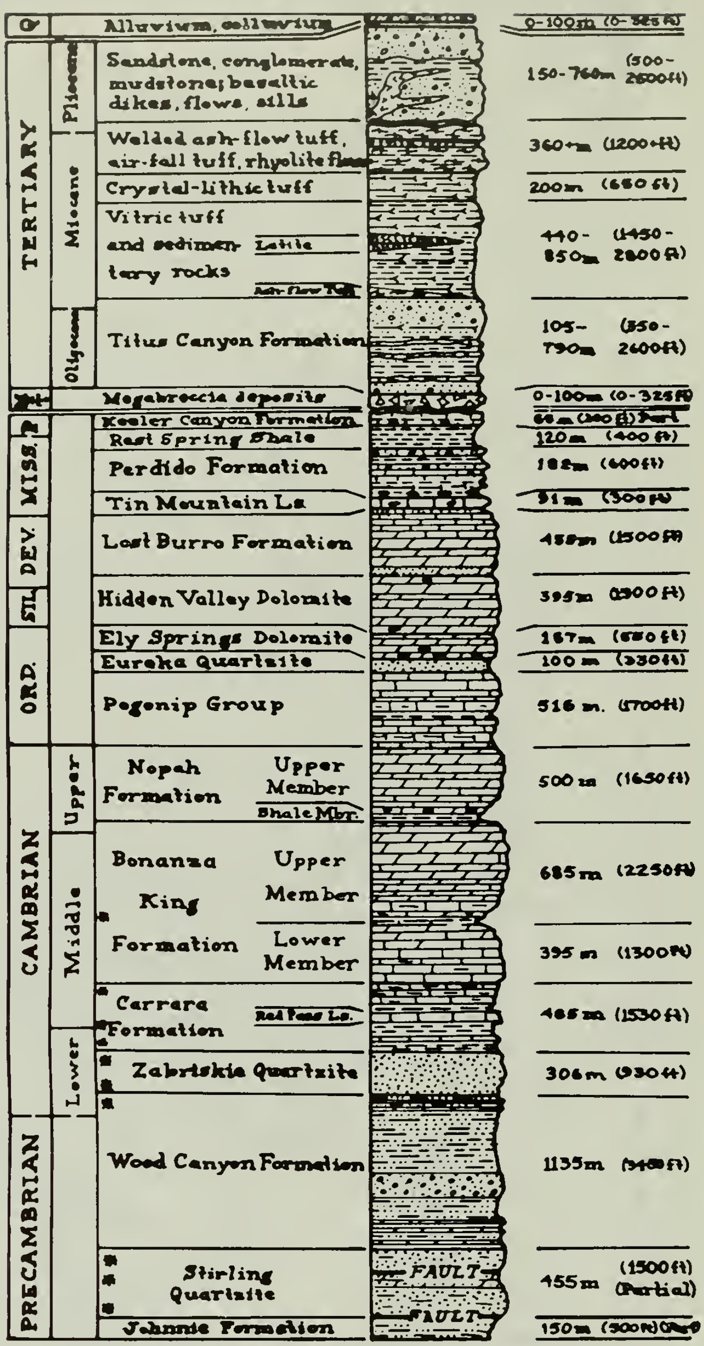
Figure 1. Generalized geologic map of the Grapevine and northern part of the Funeral Mountains, Death Valley, California, showing localities referred to in the text.

of middle or late Pliocene age, equivalent to the Furnace Creek Formation (compare with Hunt and Mabey, 1966, p. A57).

Alluvium eroded from the mountains spreads into Death Valley from major canyons and rills. Hunt and Mabey (1966) and Reynolds (1969) recognized four ages of alluvium northward from Boundary Canyon. The alluvium is both faulted against the mountain front from Titanothere to Red Wall Canyon (Fig. 1) and banked depositionally against several fault scarps, demonstrating intermittent movement along different segments of the mountain front. Alluvium adjacent to the mountains is unfaulted south of Titanothere Canyon to beyond Boundary Canyon, although locally warped. Clay, silt, and evaporite minerals are accumulating on a playa adjacent to the steep front of the Grapevine Mountains between Titus and Titanothere Canyons. The eastward displacement of the playa with respect to the axis of Death Valley toward the frontal scarp of the mountains suggests that that segment of the valley is tilting downward to the east. Both alluvium and playa sediments are displaced by faults along the northern Death Valley–Furnace Creek fault zone.

The general geologic structure of the Funeral and Grapevine Mountains is a broad anticlinorium with the deepest structural level exposed in the culmination near Chloride Cliff in the northern part of the Funeral Mountains. Progressively shallower structural levels are exposed northward across the Grapevine Mountains and southeastward across the southern part of the Funeral Mountains. Shallower structural levels are also exposed down the dip of the east flanks of the mountains and locally on the west flanks. The exposed structures developed during two principal times of deformation: the older during early Mesozoic time (probably Late Triassic or Early Jurassic [Burchfiel and others, 1970]), and the younger encompassing middle and late Cenozoic time. Different styles of deformation characterized by thrust faulting in which upper Precambrian and lower Paleozoic rocks were thrust over middle and upper Paleozoic rocks. The later deformation, however, is primarily distinguished by normal faulting and doming associated with regional tectonic extension and igneous activity. The following summary describes structures of early Mesozoic age and emphasizes structures of Cenozoic age.

Structures of Mesozoic Age. Across the northern part of the Grapevine Mountains, rocks of Precambrian, Cambrian, and Ordovician age are thrust over rocks ranging in age from Ordovician into Pennsylvanian. The fault is called the Grapevine thrust fault (Reynolds, 1971). At the base of the allochthonous plate about 8 km (5 mi) west-northwest of Mount Palmer, strongly contorted beds of the Stirling Quartzite are tectonically interleaved with deformed beds of the Wood Canyon Formation. There these rocks rest on the autochthonous Perdido Formation and Rest Spring Shale of Mississippian and Pennsylvanian(?) age. North and west-southwest of Mount Palmer, the Wood Canyon Formation is the lowest exposed unit of the allochthon. Rocks of the upper plate generally dip north across the north end of the Grapevine Mountains, so that the Cambrian Zabriskie Quartzite, Carrara, Bonanza King, and Nopah Formations, and the Ordovician Pogonip Group successively form the sole of the upper plate. Rocks of the upper plate are folded in a tight north-trending anticline 3.2 km (2 mi) southwest of Grapevine Peak, in the core of which the Wood Canyon Formation is exposed, but otherwise the plate was not strongly deformed during thrusting.



*Denotes interval of common failure during folding and faulting. Thicknesses are approximate. MzT denotes Mesozoic(?) and Tertiary age. Data of Mitchell W. Reynolds.

Figure 2. Generalized columnar section of rocks exposed in the Grapevine Mountains, Death Valley, California.

The authochthon includes the entire succession of Paleozoic and Precambrian rocks exposed from outcrops of the Grapevine thrust fault southward up the regional plunge to the major culmination of the anticlinorium near Chloride Cliff. Immediately beneath the fault, middle and upper Paleozoic rocks are strongly deformed: 9.7 km (6 mi) west of Grapevine Peak, the Mississippian Perdido Formation, the Rest Spring Shale, and the Pennsylvanian Keeler Canyon Formation are locally overturned west or northwest beneath the fault. This overturning suggests that the allochthon moved southeast and east. West of Mount Palmer the Devonian Lost Burro Formation is locally contorted in disharmonic folds beneath the thrust fault. Farther south the autochthon seems to have been little deformed during thrusting.

Stewart and others (1966) suggested that the Grapevine thrust fault may be equivalent to the Last Chance thrust fault that they described from extensive exposures west of Death

Valley. The Last Chance thrust fault is structurally the highest and youngest thrust fault and seems to have the greatest displacement in a succession of thrust faults west of Death Valley, including the Lemoigne, Gap Hills, Big Horn, Quartz Springs, and Racetrack thrust faults (McAllister, 1952; Hall and Stephens, 1962). I (1971) tentatively accepted the correlation of the Last Chance and Grapevine thrust faults and noted that the Grapevine thrust fault continues east from the Grapevine Mountains through exposures in the Bullfrog Hills to Bare Mountain, where Cornwall and Kleinhampl (1961) described a thrust fault at Meiklejohn Peak, which shows similar structural and stratigraphic relations. Although the specific surface of thrusting is not likely the same across the region, the exposed fault surfaces are probably part of the Last Chance thrust system. Thrust faults equivalent to the older described faults on the west side of Death Valley do not seem to be present on the east side in either the Grapevine Mountains or the northern part of the Funeral Mountains. Plutons of quartz monzonite that intrude the thrust faults west of Death Valley have approximate radiometric ages of 156 and 165 m.y., and Burchfiel and others (1970) consider that thrusting occurred during latest Paleozoic and early Mesozoic time, most likely during Middle Triassic to Early Jurassic time. That age of thrusting is also accepted for the Grapevine thrust fault.

No stratigraphic record of tectonic events younger than thrusting but older than Oligocene age is preserved in the Grapevine Mountains. Deformed Precambrian and Paleozoic rocks were tilted northward prior to erosion of the surface on which rocks of the Titus Canyon Formation (Oligocene and Miocene(?)) accumulated. That formation rests on Precambrian–Cambrian Wood Canyon Formation east of Chloride Cliff (L. A. Wright and B. W. Troxel, 1971, written commun.) and on progressively younger strata to the north where, near Grapevine Peak, it rests unconformably on allochthonous rocks of the Grapevine thrust fault. The relation suggests that the area of the Funeral Mountains was a structural high prior to the erosion and that the present anticlinorium represents a middle and late Cenozoic rejuvenation of that earlier structure.

Structures of Cenozoic Age. The most conspicuous geologic structures in the Grapevine Mountains and northern Funeral Mountains developed during Cenozoic time. Deformation was episodic in pulses between 20 and 16, ~14 and 13, and 11 and 7 m.y., and has been nearly continuous since (Reynolds, 1974). The first two episodes were characterized by uplift in the southern part of the area and by faulting along northeast and north trends. The episode between 11 and 7 m. y. seems to have been particularly intense, for, during that time, doming, recumbent folding, and extensive faulting along a north trend occurred. Intermittent younger deformation has produced much of the relief evident along the mountain fronts and right-lateral faulting that displaces Quaternary units in northern Death Valley.

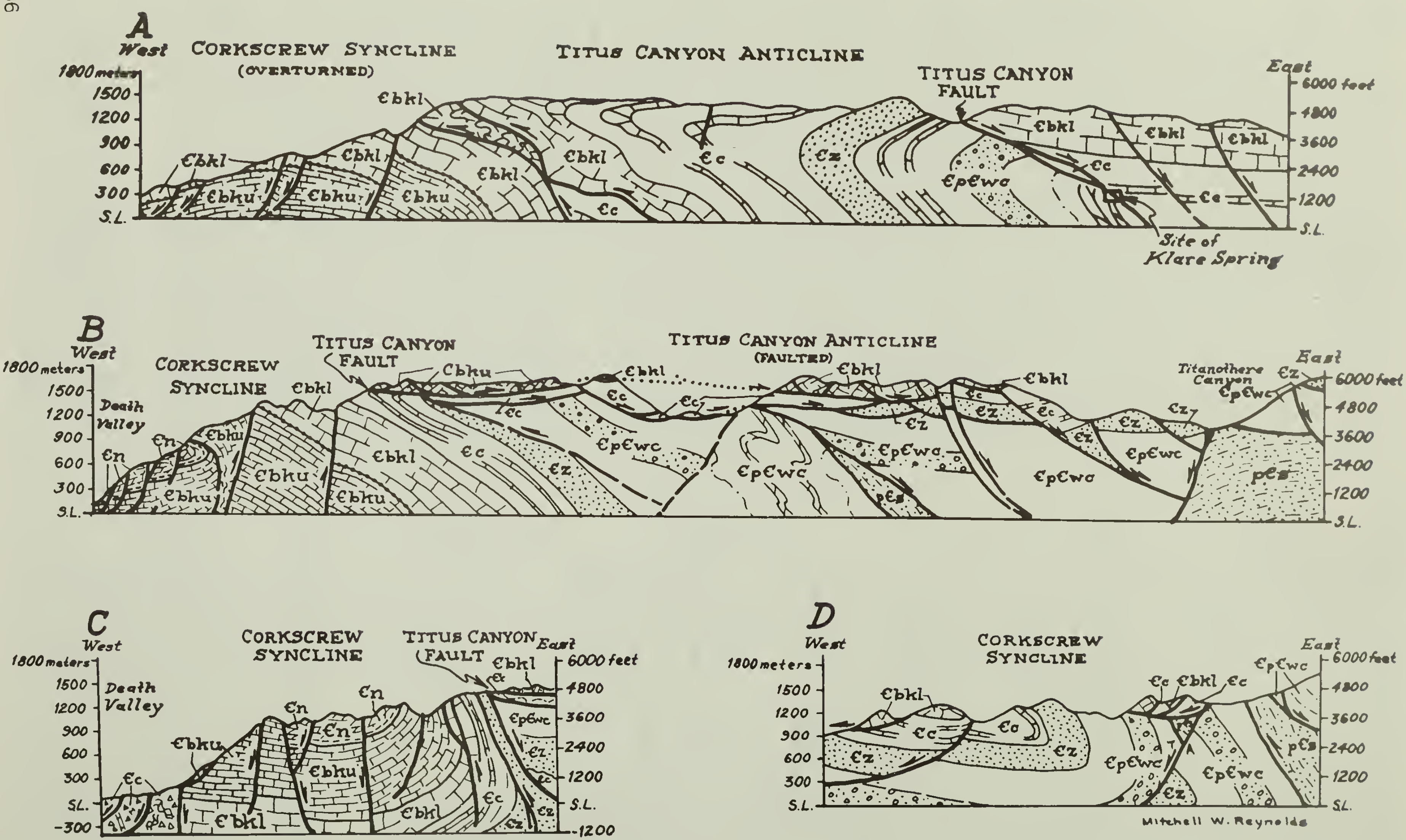
In section view from north-northwest to south-southeast along the Grapevine Mountains and northern Funeral Mountains, the general structure that developed during Cenozoic time is an anticlinorium with a major culmination in the vicinity of Chloride Cliff and a second culmination between Mount Palmer and Grapevine Peak (Fig. 1). From east to west the ranges are anticlinal with tilted, faulted Tertiary rocks on the east and west flanks. Minimum structural relief between the culmination at Chloride Cliff and the northern end of the Grapevine Mountains is 10.6 km (6.6 mi). Southeast down plunge from Chloride Cliff the apparent structural relief exceeds 6 km (3.7 mi). Structural relief between the crest of the second culmination near Mount Palmer and the equivalent stratigraphic horizon in the adjacent depression that lies between Titus and Titanothere Canyons is about 3 km (1.9 mi). Part of the relief had developed prior to Oligocene time, but most developed in Pliocene time, as demonstrated by strata of early Pliocene age tilted to angles of 45° on the flanks of the folds.

About 8 km (5 mi) east-southeast of Chloride Cliff, high-angle faults that displace rocks as young as Pliocene flatten westward and northward with depth, merging to form a single low-angle fault (L. A. Wright and B. W. Troxel, 1971, written commun.). Progressively north along the east flank of the range, north-trending faults also flatten with depth to join the master low-angle fault. In its southernmost exposures that fault is within the Stirling Quartzite, but from about 5 km (3 mi) east of Chloride Cliff to Boundary Canyon the fault forms a surface of décollement between the Precambrian Johnnie Formation below and the Stirling Quartzite and younger strata above. Across that area, Tertiary, Cambrian, and Precambrian strata, in blocks bounded by the high-angle faults that flatten to join the master fault, are rotated downward to the east or northeast to angles as steep as 75°. The low-angle master fault is well exposed between the Stirling Quartzite and Johnnie Formation on the north side of Boundary Canyon from south of Hole-in-the-Rock Spring east for about 1 km (0.7 mi).

From Boundary Canyon north as far as Titus Canyon in the Grapevine Mountains, the Stirling Quartzite, Wood Canyon Formation, and Cambrian beds are folded in an S-shaped fold system that consists of a recumbent syncline on the west and a structurally higher upright-to-recumbent anticline on the east (Fig. 3). The low-angle fault visible beneath the Stirling in Boundary Canyon is the surface of décollement between the recumbently folded strata and metamorphosed deformed older rocks. A limestone unit in the lower part of the Carrara Formation outlines the tight core of the recumbent syncline on Corkscrew Peak, and the Stirling and Wood Canyon beds east for 2 km (1.3 mi) from Hold-in-the-Rock Spring outline the broader, lower part of the syncline. The anticline is defined by outwardly opposing dips in the Wood Canyon Formation exposed on the northwest side of the highway through Boundary Canyon about 6.5 km (4 mi) above the spring. In Boundary Canyon, the anticline is faulted downward approximately 1.6 km (1 mi) against the overturned east limb of the syncline.

Traces of the fold axes trend about N. 60° W. toward a salient in the S-shaped fold system near Titus Canyon; north of the salient they trend north. The folds plunge northwest off the culmination of the Funeral Mountains anticlinorium. The trace of the axial surface is exposed in progressively younger beds northwest from Boundary Canyon in the recumbent syncline, so that between Titanothere and Titus Canyons the Upper Cambrian Nopah Formation forms the core of the fold (B and C in Fig. 3). Beds of the Cambrian Bonanza King Formation are the youngest exposed on the upright limb of the anticline near Titus Canyon, but farther north progressively younger Paleozoic rocks are continuous above the Bonanza King. Otherwise, only older strata are exposed in the anticline, and generally the overturned limb on the west has been faulted out. The axial surface of the syncline dips 20° to 60° to the east; for about 5 km (3 mi) southeast from the mouth of Titus Canyon, the recumbent syncline has been rotated downward toward Death Valley so that the axial surface dips about 15° to the west (A and B in Fig. 3). At Boundary Canyon the axial surface of the anticline is nearly vertical, but it dips northeast at progressively lower angles toward Titanothere Canyon. The recumbent part of the anticline is fully preserved only north of Titus Canyon where the axial surface dips from 5° to 10° to the east (A in Fig. 3).

Between Titus and Titanothere Canyons, where the west-directed salient of the fold system coincides with the structural depression in the range, beds of the upright limb of the anticline are broken by numerous normal faults (B in Fig. 3). Along these faults the strata have been rotated downward to the west and effectively extended over a wider area. The



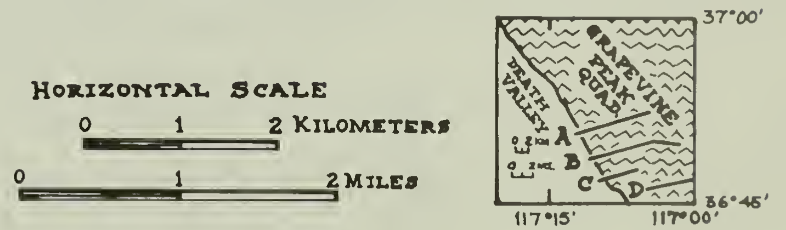


Figure 3. Generalized geologic sections of part of the southern Grapevine Mountains, Death Valley, California.

€n, Nopah Formation; €bku, Bonanza King Formation, upper part; €bkl, Bonanza King Formation, lower part; €c, Carrara Formation; €z, Zabriski Quartzite; €p€wc, Wood Canyon Formation; p€s, Stirling Quartzite.

faults flatten eastward to join a single, nearly flat fault called the Titus Canyon fault; it separates the upright but faulted beds from the overturned limb of the recumbent syncline beneath (Fig. 1; B in Fig. 3). That low-angle fault replaces the axial surface of the anticline south of Titus Canyon, whereas north from Klare Spring in the canyon, the fault steepens to become nearly vertical and parallel to the trace of the upper part of Fall Canyon (Fig. 1; A in Fig. 3). Displacement along the vertical segment of the fault is down on the east, and the fault offsets strata as young as early Pliocene.

I have (1969, 1970, 1971) interpreted the flat fault as a lag fault along which the upright limb of the recumbent anticline broke from and lagged behind the developing recumbent fold system. The fault steepens where the recumbent anticline remains intact (A in Fig. 3). Tertiary rocks are absent from the western part of the faulted limb, although east near Leadfield the Titus Canyon Formation is locally preserved in blocks rotated downward toward the principal low-angle fault. These relations, together with the displacement of lower Pliocene rocks along the steep segment of the fault and the style of faulting, all suggest that the recumbent folding and faulting occurred at shallow depths in the crust probably during middle Pliocene time. Evidence further corroborating the young age of the recumbent folding is derived from dating the décollement fault at the base of the fold system as probable middle Pliocene, because that fault offsets strata of early Pliocene age east of Chloride Cliff. The folding was accomplished before late Pliocene and Quaternary time, because beds of fanglomerate of those ages southwest and west of Death Valley Buttes and north of Fall Canyon contain clasts derived from rocks not exposed in the mountains prior to the folding and erosion.

North-trending normal faults on the east flanks of the Grapevine Mountains and northern Funeral Mountains displace the sequence of Tertiary rocks progressively down toward the east. These faults flattened eastward, or locally westward, with depth to follow horizons nearly parallel to bedding in the Cambrian Zabriskie Quartzite or in the upper part of the Wood Canyon Formation (Fig. 2; B in Fig. 3). Displacement on separate faults is as much as 1,000 m (3,200 ft). The offsets suggest that the Cenozoic rocks were extended in a west-northwest-east-southeast direction as they tore from and moved against the subjacent Paleozoic and Precambrian rocks that were being arched in the mountain core.

The abrupt west-southwest and south fronts of the Grapevine Mountains mark zones of normal faulting and warping along which the mountains have risen with respect to Death Valley. Gravity data of Mabey (1963) and the distribution of Tertiary rocks in the southern part of the Grapevine Mountains suggest that vertical separation on the pre-Tertiary surface between the valley and the mountains may be at least 4.3 km (2.7 mi). Fault scarps dip 44° to 75° toward the valley, and antithetic faults dip into the mountain front to break it into a mosaic of strongly fractured small blocks. Warping toward Death Valley accounts for an unknown but substantial part of the structural relief on the range front, for the axial surface of the recumbent syncline of probable middle Pliocene age has been warped 15° down toward Death Valley (A and B in Fig. 3). As a result of both the warping and recumbent folding, beds of the Cambrian Bonanza King Formation at the mouth of Titus Canyon have been rotated through as much as 235° from original horizontality. North of Fall Canyon the apparent vertical separation between Death Valley and the mountain front diminishes and right-lateral strike-slip faulting becomes important within the range. Movement seems to have recurred through latest Tertiary and Quaternary time.

From Death Valley Buttes to Titus Canyon along the steep mountain front, strata in blocks as much as 1.9 km (1.2 mi) across are faulted downward toward the valley along faults that flatten with depth to become nearly horizontal (Fig. 1; C and D in Fig. 3). The flat toes of such faults are commonly exposed at elevations higher than the valley floor. Clearly, these are block-glide landslide structures that formed during or after development of the high relief on the mountain front.

Evidence of Quaternary movement along the northern Death Valley–Furnace Creek fault zone is present almost continuously for 3.2 to 6.5 km (2 to 4 mi) west of the Grapevine Mountains in Death Valley. The linear west front of the Kit Fox Hills, west of Death Valley Buttes (Fig. 1), passes north-northwest into a linear fault trace along which alluvium and playa sediments are furrowed, and springs emerge. The northern Death Valley highway just south of its junction with the Titus Canyon road rises onto a scarplet in alluvium that varies from 0.6 to 2 m (2 to 7 ft) high; the west side is up relative to the east side. Northwest of Red Wall Canyon an old alluvial fan is displaced right-laterally about 46 m (150 ft) with no significant vertical offset, yet 12 km (7.5 mi) farther northwest, displacement on the fault is about 4.5 m (15 ft) west side up. Near Grapevine Ranger Station, at the north end of the Grapevine Mountains, the east side of the fault is elevated with respect to the west. The linear trace of the fault, patterns of offset of Quaternary deposits, and physiographic features characteristic of strike-slip faulting demonstrate that the fault is nearly vertical and that movement has been in a right-lateral direction. Right-lateral movement on the fault zone, probably occurring largely since middle Cenozoic time, has been variously estimated as being a few miles (Wright and Troxel, 1967, 1970) to scores of miles (Stewart, 1967; Stewart and others, 1968; Burchfiel and others, 1970).

REFERENCES CITED

Burchfiel, B. C., Pelton, P. J., and Sutter, J., 1970, An early Mesozoic deformation belt in south-central Nevada–southeastern California: Geol. Soc. America Bull., v. 81, p. 211–215.

Cornwall, H. R., and Kleinhampl, F. J., 1961, Geology of the Bare Mountain quadrangle, Nevada: U.S. Geol. Survey Geol. Quad., Map GQ–157.

——— 1964, Geology of the Bullfrog Hills quadrangle and ore deposits related to Bullfrog Hills caldera, Nye County, Nevada, and Inyo County, California: U.S. Geol. Survey Prof. Paper 454–J, 25 p.

Hall, W. E., and Stephens, H. G., 1962, Preliminary geologic map of the Panamint Butte quadrangle, Inyo County, California: U.S. Geol. Survey Mineral Inv. Field Studies Map MF–251.

Hunt, C. B., and Mabey, D. R., 1966, General geology of Death Valley, California—Stratigraphy and structure: U.S. Geol. Survey Prof. Paper 494–A, 162 p.

Mabey, D. R., 1963, Complete Bouguer anomaly map of Death Valley region, California: U.S. Geol. Survey Geophys. Inv. Map GP–305.

McAllister, J. F., 1952, Rocks and structure of the Quartz Spring area, northern Panamint Range, California: California Div. Mines and Geol. Spec. Rept. 25, 38 p.

——— 1970, Geology of the Furnace Creek borate area, Death Valley, Inyo County, California: California Div. Mines and Geology Map Sheet 14.

Reynolds, M. W., 1969, Stratigraphy and structural geology of the Titus and Titanothere Canyons area, Death Valley, California [Ph.D. dissert.]: Berkeley, Univ. California, Berkeley, 310 p.

——— 1970, Low-angle faults of lag and gravity origin, northeastern Death Valley, California: Geol. Soc. America, Abs. with Programs (Cordilleran Sec.), v. 2, no. 2, p. 134–135.

——— 1971, The Grapevine thrust and its significance to right-lateral displacement in the northern Death Valley area, California: Geol. Soc. America, Abs. with Programs (Cordilleran Sec.), v. 3, no. 2, p. 182–183.

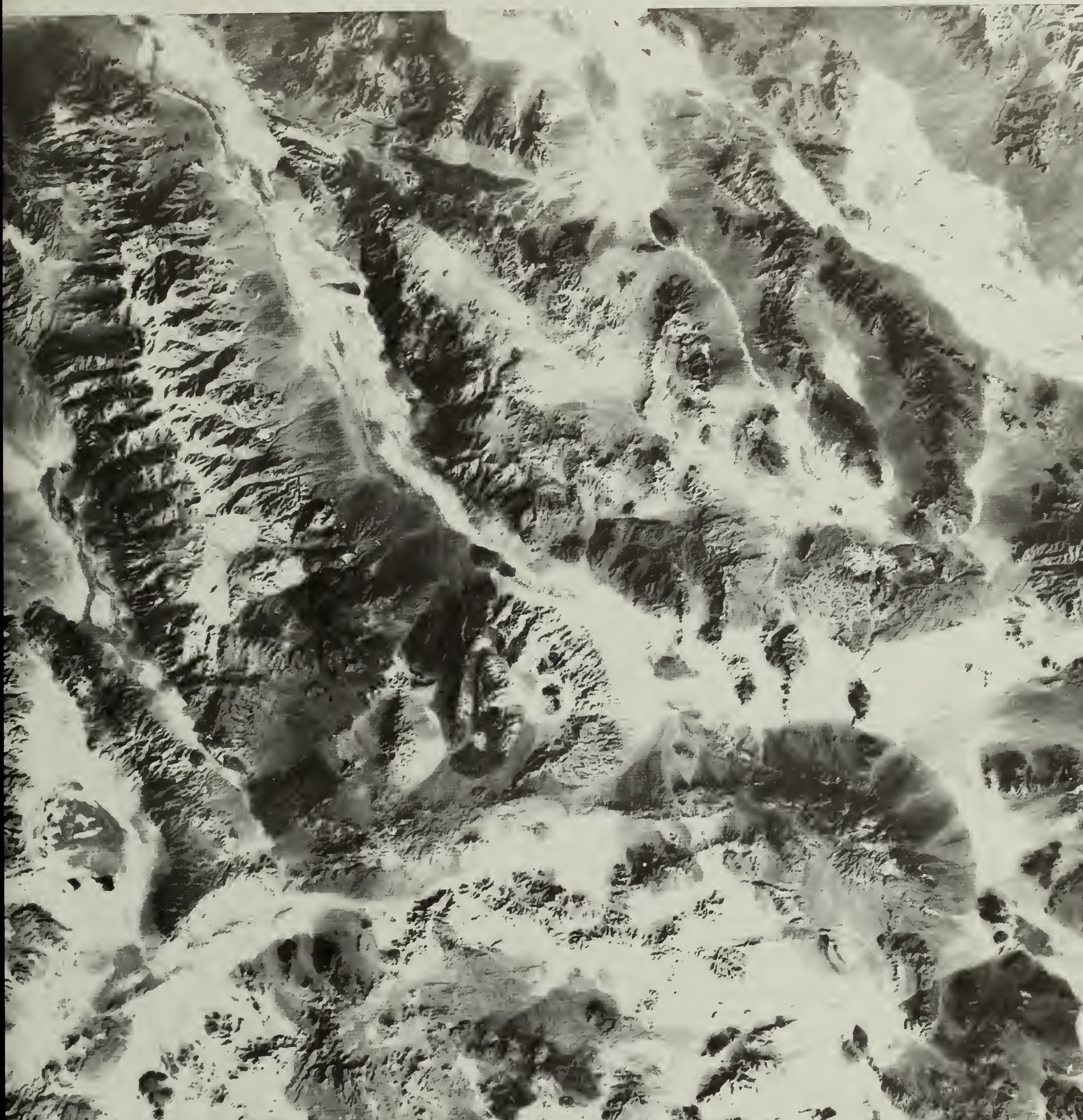
——— 1974, Recurrent middle and late Cenozoic deformation, northeastern Death Valley, California-Nevada: Geol. Soc. America, Abs. with Programs (Cordilleran Sec.), v. 6, no. 3, p. 241–242.

Stewart, J. H., 1967, Possible large right-lateral displacement along fault and shear zones in the Death Valley–Las Vegas areas, California and Nevada: Geol. Soc. America Bull., v. 78, p. 131–142.

Stewart, J. H., Ross, D. C., Nelson, C. A., and Burchfiel, B. C., 1966, Last Chance thrust—A major fault in the eastern part of Inyo County, California, *in* Geological Survey research, 1966: U.S. Geol. Survey Prof. Paper 550–D, p. D23–34.

Stewart, J. H., Albers, J. P., and Poole, F. G., 1968, Summary of regional evidence for right-lateral displacement in the western Great Basin: Geol. Soc. America Bull., v. 79, p. 1407-1414.
Wright, L. A., and Troxel, B. W., 1967, Limitations on right-lateral, strike-slip displacement, Death Valley and Furnace Creek fault zone, California: Geol. Soc. America Bull., v. 78, p. 933-950.
1970, Summary of regional evidence for right-lateral displacement in the western Great Basin: Discussion: Geol. Soc. America Bull., v. 81, p. 2167-2174.

High-altitude oblique aerial photograph to north-northeast from a point approximately above Skidoo, Panamint Range. Death Valley extends obliquely across center of photograph, Funeral Mountains (right) and Grapevine Mountains (left) lie beyond. The Amargosa River flows southward through the Amargosa Desert from its headwaters area in upper left-center of photograph. The Northern Death Valley–Furnace Creek fault zone forms a modest trace near the base of the fans extending into Death Valley from the Funeral and Grapevine Mountains. Tucki Mountain occupies center foreground. *Photo U.S. Air Force 374L 190, 6 September 1968; courtesy of the U.S. Geological Survey.*

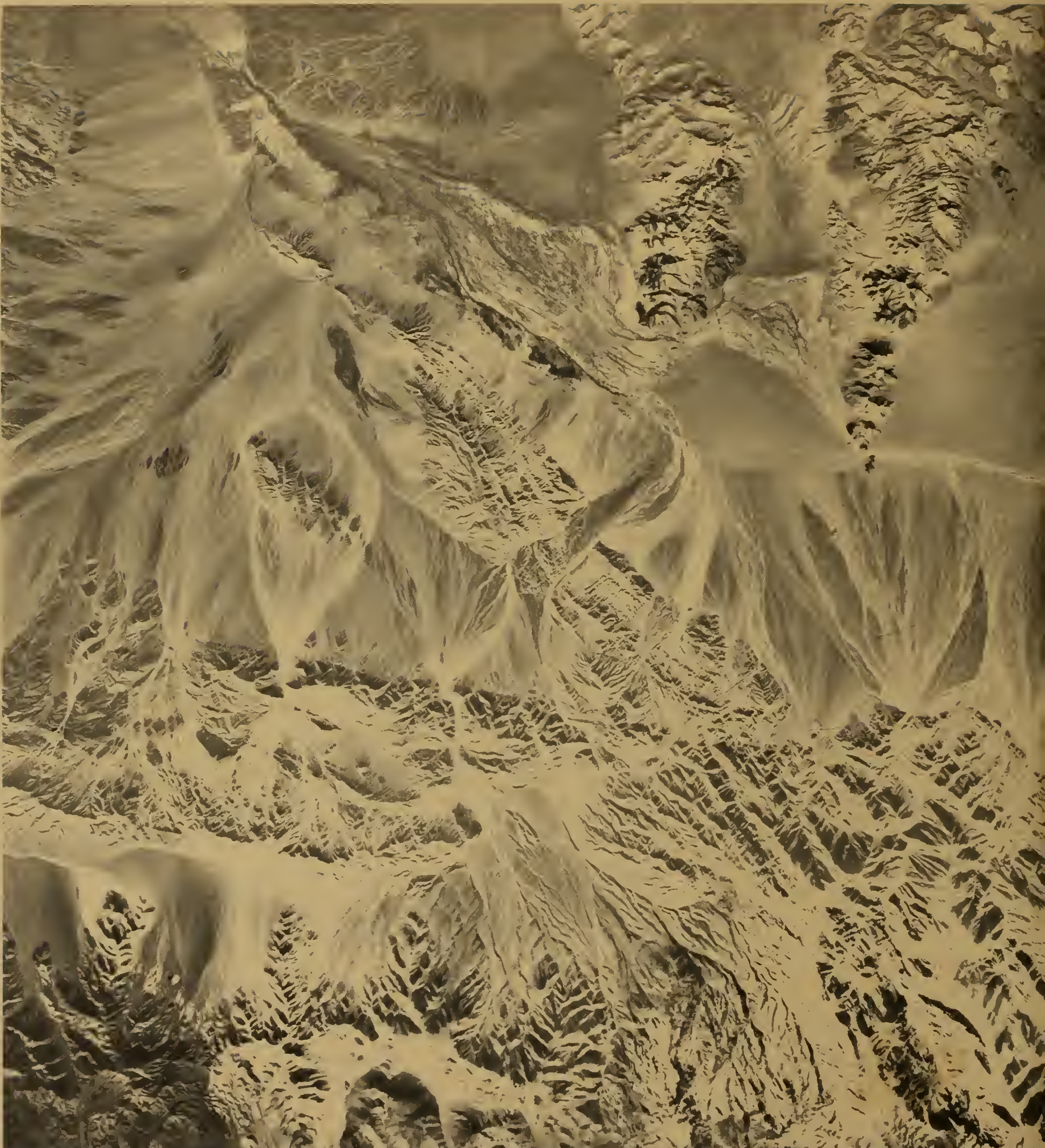


High-altitude imagery of southern and central Death Valley. *Image NASA ERTS 1125-17554 band 7, 25 November 1972; copy produced by Drew P. Smith, California Division of Mines and Geology.*

Oblique aerial view southwestward from point above vicinity of Tecopa. Avawatz Mountains, which form dark range in center of photo, lie south of intersection of Death Valley and Garlock fault zones. Salt Spring, Saddle Peak, Saratoga, and Ibex Hills occupy foreground. Mojave Desert beyond Avawatz Mountains is bounded on the south by San Bernardino and San Gabriel Mountains (skyline area). U.S. Geological Survey—U.S. Air Force Photograph.

FRONT COVER PHOTOGRAPH: Oblique aerial view eastward toward Las Vegas region from point above southern Owens Valley. Ranges and valleys cross photograph from lower to upper part in the following order: Argus Range (and bifurcating Slate Range to right), Panamint Valley, Panamint Range (capped with snow in center, covered by clouds to south), Death Valley, Black Mountains (Funeral Creek Wash and Funeral Mountains farther north—to the left). U.S. Geological Survey—U.S. Air Force photograph.

The preparation of this guidebook would not have been possible without the diligent efforts of Scott Brunk, Claire Hogle, Faith Ingalls, Beverly Kies, Renee Munoz, Harrison Shaffer, Dan Teaford, June Thomas, Betty Troxel, and Ann Wellck. Fourth printing, 1983, by Fairfield Printing Company, Fairfield, California.



Bock cover. High-oltitude verticol photogroph neor south end of Deoth Volley. North is toword top of photogroph. Moin moss of Avowotz Mountoins occupies lower right of photogroph, portly covered by snow. Troce of north bronch of Gorlock foult forms obvious stroight north boundory of port of Avowotz Mountoins ond hills to west. Southern Deoth Volley foult zone extends northwest to upper lett corner ot pnotogroph trom Avowotz Mountoins ond Gorlock foult in righi center of photogroph. Smoll unnomed northwest-trending hills contoin severol bronches of the Southern Deoth Volley foult zone; youngest bronch bounds northeost side of hills ond continues forther northwest in floor of volley.

Broided chonnels ore port of Amorgoso River which flows west oround southern end of Soddle Peok Hills (upper right) ond Sorotogo Hills (upper center) then north into centrol Deoth Volley. *Photo U.S. Air Force 744V 073, 29 November 1967; courtesy of the U.S. Geologicol Survey.*