GUIDEBOOK

GEOLOGIC EXCURSIONS IN THE CALIFORNIA DESERT

Prepared for the 78th Annual Meeting of the Cordilleran Section of the Geological Society of America, Anaheim, California, April 19 - 21, 1982

FIELD TRIP NUMBERS 2, 7, 13

Contents

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LATE CENOZOIC TECTONIC AND MAGMATIC EVOLUTION OF THE CENTRAL MOJAVE DESERT, CALIFORNIA

FIELD TRIP NUMBER 2

LEADERS: R. K. Dokka and A. F. Glazner

Contents

FIELD TRIP ROADLOG: LATE CENOZOIC TECTONIC AND MAGMATIC EVOLUTION OF THE CENTRAL MOJAVE DESERT, CALIFORNIA

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INTRODUCTION

Welcome to the Mojave Desert! Over the course of the next two days we hope to expose you to many of the geologic wonders and natural beauty of this magnificent land, emphasizing aspects of its late Cenozoic history. The central Mojave has been a long neglected entity of California geology, perhaps due to the hostile environment or rugged topography. Detailed work on Cenozoic problems of the Mojave Desert has increased dramatically in the last 10 years (e.g. see Howard and others, 1981). This guidebook and its associated descriptive papers represents the product of that activity and presents the results of some of the recent research into Cenozoic problems.

This guidebook is arranged into two parts, a roadlog and a collection of papers. The roadlog is divided into two legs, Day 1 and Day 2, and includes descriptions of specific stops that are to be made. In order to increase readability of the roadlog, we have kept explicit mileage directions to a minimum. Therefore, when such directions occur (e.g. "note mileage in Kelso"), they should be noted, because they will help locate an important turn or feature that might otherwise be missed.

We are indeed fortunate to have had the cooperation of several active Mojave researchers in contributing papers that deal with specific aspects of this trip. These articles provide a more complete discussion of the topics raised in the roadlog and will greatly enhance your "outcrop experience" if you read them beforehand. For your convenience the papers are arranged in an order that reflects the trip itinerary.

ACKNOWLEDGMENTS

We would like to take this opportunity to thank the many people who made this field trip and volume possible. John D. Cooper, Field Trip Chairman, 1982 GSA/SSA Annual Meeting handled most of the administrative and logistical tasks required for this excursion. A. Boettcher, Susan Miller, Steve Semken, Robert P. Sharp, Ray Weldon, II, and Mike Woodburne made many valuable suggestions regarding the sights to see at the various stops. Clifford Duplechin, Mary Lee Eggart, and James Kennedy, Jr. drafted or reproduced many of the figures. Dan Pease printed many of the photographs in this guidebook. The LSU College of Arts and Sciences Word Processing group produced the final camera-ready product. Ms. M. M. Kolb and Ms. P. Philips assisted in editing the final draft of this guidebook.

Special thanks go out to A. L. Boettcher, K. Howard, B. John, D. Miller, S. Miller, R. P. Sharp, R. H. Tedford, R. Weldon, II, and M. Woodburne for their valuable contributions and cooperation.

This work has been supported by the National Science Foundation (EAR-8107524 awarded to R. K. Dokka), Sigma Xi (A. F. Glazner, R. K. Dokka) and the Geological Society of America (A. F. Glazner).

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Begin trip at Barstow Rd.- Interstate 15 intersection. Proceed along Barstow Rd. 1.2 miles and park. DAY 1--REFER TO FIGURE 1.

STOP #1 - GEOLOGIC OVERVIEW AT BARSTOW

Objectives

1. Discuss the Cenozoic geologic evolution of the central Mojave Desert.
2. Discuss controls on present-day drainage and physiography.

The central Mojave Desert owes much of its present-day physiography and structural grain to late Cenozoic tectonic and magmatic events. These events were part of regional deformations that affected western North America since the Oligocene, namely, intraplate extension and distributed transform shear. Contrary to earlier syntheses of the tectonic evolution of western North America, the Mojave has not behaved as a stable block during the last 25 m.y.

At this stop we can see the effects of most of the Cenozoic tectonic regimes imposed on the Mojave Desert. To the southeast lie the Newberry Mountains, whose sawtooth skyline is the result of early Miocene detachment faulting. Tilting and faulting of these rocks represents some of the earliest evidence for distension of western North America during late Cenozoic time. These rocks and structures originated along the southwestern margin of a west-northwest-trending volcano-tectonic depression referred to as the Barstow-Bristol trough (Glazner, 1981; Dokka and Glazner, this volume; Gardner, 1980). This extensional belt was the site of thick accumulations of Tertiary volcanic and sedimentary rocks (Fig. 2).

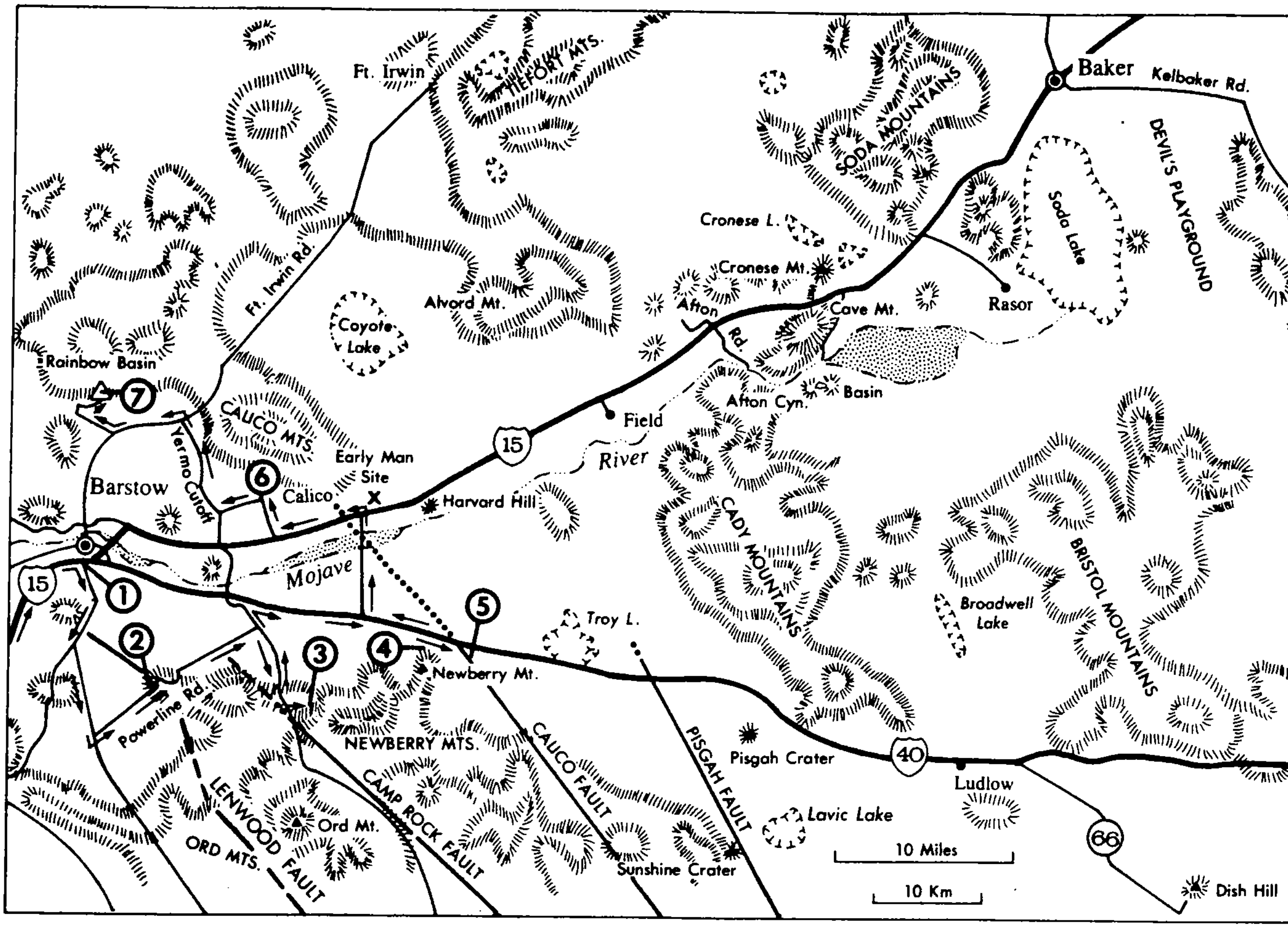

Figure 1. Index map of the central Mojave Desert, California. Numbers refer to field trip stops that are described in the text.

To the north are the Mud Hills, the type area for the middle Miocene Barstow Formation. These continental basinal facies and alluvial fan deposits contain the type Barstovian faunal assemblage described by Merriam(1919), Wood (1941), and Lewis (1964). These rocks were deposited in a single basin or a series of basins that were coextensive with the earlier extensional trough (see the article by Woodburne and others (1982) in this volume for a more complete discussion of the Barstow Formation). The abrupt margin of the Mud Hills is the result of truncation by the Calico fault, a right, strike-slip fault.

The Calico fault belongs to a family of northwest-striking wrench faults that traverse the central Mojave Desert. Local geometric irregularities along these faults have given rise to zones of shortening and extension that are manifested in topographic uplifts and depressions, respectively. We will examine one of these fault-controlled physiographic features at Stop #2. Although many of these wrench faults are long and seismicly active, displacements on individual faults are small. Cumulative distributed right lateral shear across the Mojave Desert from Lenwood to Ludlow is only 20-30km (Dokka, in prep.). Field data suggest that initiation of strike-slip faulting in the central Mojave Desert is fairly recent (Pliocene?).

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Return to California Highway 247 (Barstow Road) and proceed south for approximately 8.1 miles. Turn left at the Southern California Edison powerline road (three major high tension lines) and pass through a cattle gate (please resecure the gate). This road is periodically graded and will generally support most vehicles.

To the right (south) along the skyline are the Ord Mountains (highest point, 6270 ft.), named for General Edward O. C. Ord, leader of the first official survey of Los Angeles. This range is composed largely of andesite, dacite, and tuff of the Mesozoic Sidewinder Volcanic Series and younger Mesozoic granitic intrusive rocks. Copper, along with minor amounts of gold and silver, has been mined there since 1876. Straight ahead lie the Newberry Mountains (Daggett Ridge), our next stop.

Proceed east along the powerline road for approximately 3.4 miles and park at the bottom of the playa.

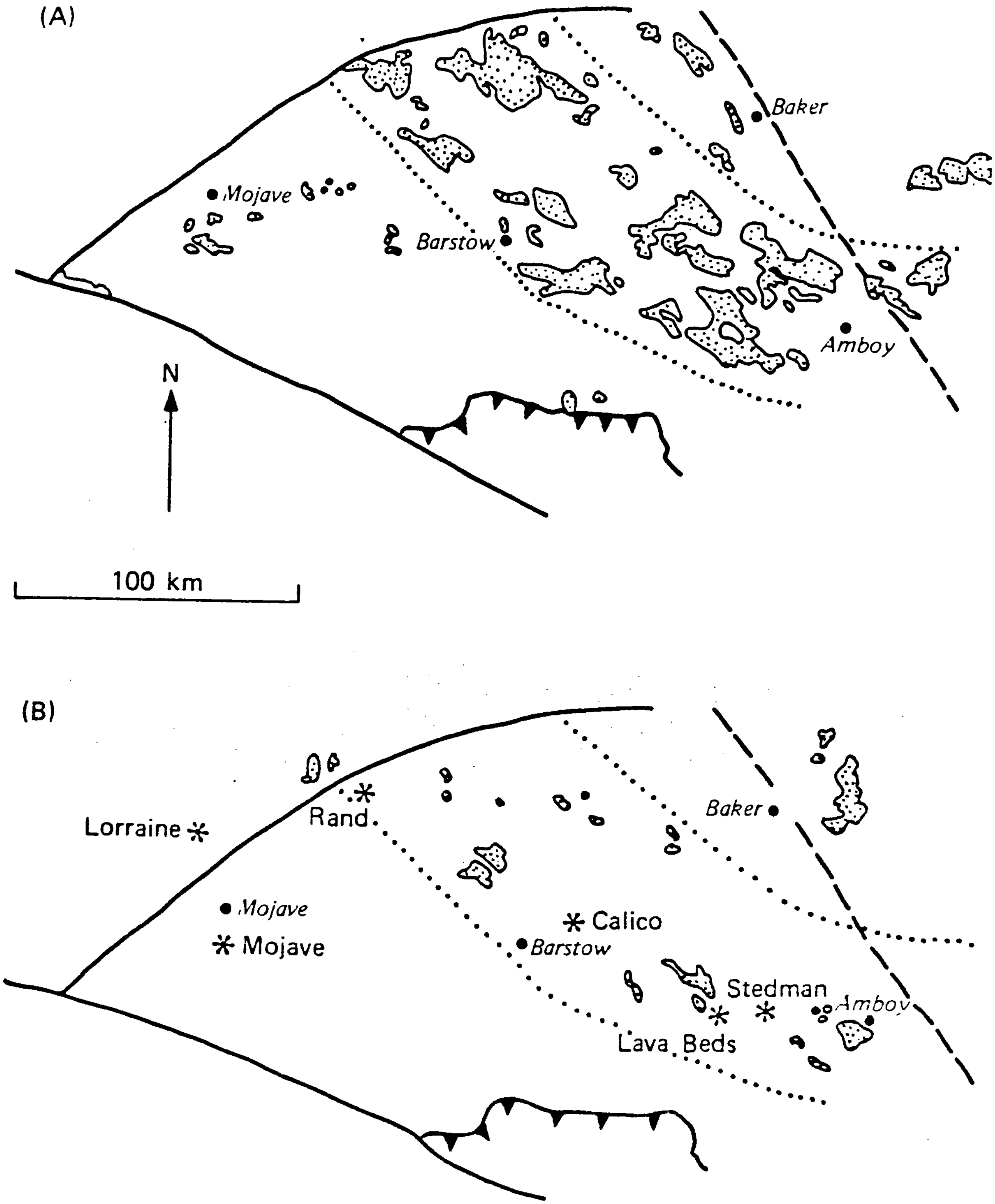

Figure 2. The Barstow-Bristol Trough (Glazner, 1981).

a) distribution of Tertiary rocks

b) distribution of Quaternary basalts and major districts

STOP #2 - PULL-APART BASIN ALONG THE LENWOOD FAULT

Objectives

1. Inspect a newly formed pull-apart basin along an active northwest-striking strike-slip fault.

2. Observe lower Miocene sedimentary breccias and conglomerates and discuss their depositional environment.

3. Consider evidence for crustal extension that commenced at the beginning of the Miocene in the central Mojave Desert.

As shown on figure 3, the areal geometry of this depression suggests that it is a rhombochasm or pull-apart basin. The valley lies between two strands of the Lenwood fault, a northwest-striking right-slip wrench fault. The geology of the area covered by figure 3 is presented in figure 4. The valley formed (and is presumably still evolving) as a result of oblique extension, produced as the fault transferred motion from one strand to another. The southwest margin of the basin is unique because it displays the highest relief and is adjacent to the lowest portion of the basin. This edge is marked by a steep scarp composed of the highly indurated sedimentary breccias of the lower Miocene formation of Slash X Ranch. Oblique-slip along the portion of the fault is suggested by the occurrence of low-angle slicken-side striations. Cross-cutting relationships on this fault and other northwest-striking wrench faults of the central Mojave Desert suggest that this style of faulting was imposed no earlier than middle Miocene (Dokka, ms). The Lenwood fault displays only 1-2 km of strike separation.

Sedimentary breccias in the road cut and the hogback to the east (Daggett Ridge) consist mainly of large angular clasts of Mesozoic-age Sidewinder

Figure 3. Vertical airphoto of a pull-apart basin along the Lenwood fault, western Newberry Mountains.

Volcanic Series (andesite-dacite). Large masses of granitoid-bearing breccia and conglomerate also occur in the unit. Sidewinder-bearing breccias are extremely resistant to weathering, forming the resistant unit that caps the southwest dipping homoclinal ridge of Daggett Ridge. The lower portion of the formation of Slash X Ranch was derived from the southwest whereas rocks of the formation of Daggett Ridge were transported from evolving volcanic fields to the northeast (Dokka, 1980). Figure 5 illustrates the stratigraphic relations of these two units.

These coarse, angular sediments were deposited in a rapidly evolving, narrow basin that lay adjacent to an active high-angle normal fault (Kane Springs fault). This fault, although not seen at the surface, is suggested to lie beneath the younger sediments of Stoddard Valley. The Kane Springs fault is kinematically related to a family of northwest-striking, northeast-dipping normal faults that are responsible for distending and tilting lower Miocene and older rocks of the central Mojave Desert(Dokka, 1980). Because sedimentation was occurring in concert with the faulting and tilting, sediment packages such as the formation of Slash X Ranch provide a key to understanding aspects of the tectonic evolution of the area.

The formation of Slash X Ranch most probably represents the proximal facies of an alluvial fan system that developed adjacent to a rising mountain front that lay to the west (Stoddard Mountain-Stoddard Ridge area). Geometry, lithology, and internal structure of the breccia units are similar to those described as landslide deposits (Heim, 1932; Longwell, 1945; Drewes, 1963; Burchfiel, 1966; Shreve, 1968; Hsu, 1975; Krieger, 1977). These coarse, angular, jumbled deposits are thought of as originating as huge rock slivers that became detached from a mountainside, broke up during their downhill descent, and were deposited as either air-cushioned slides (Shreve, 1968) or as flows (sturzstrom; Heim, 1932; Hsu, 1975). Conglomeratic facies are considered to be debris flow deposits on the basis of their matrix-supported fabric and are similar to the sediments found on the upper and middle portions of active fans (Bull, 1972).

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Return to vehicles and continue east along the powerline road. From the lake floor to the

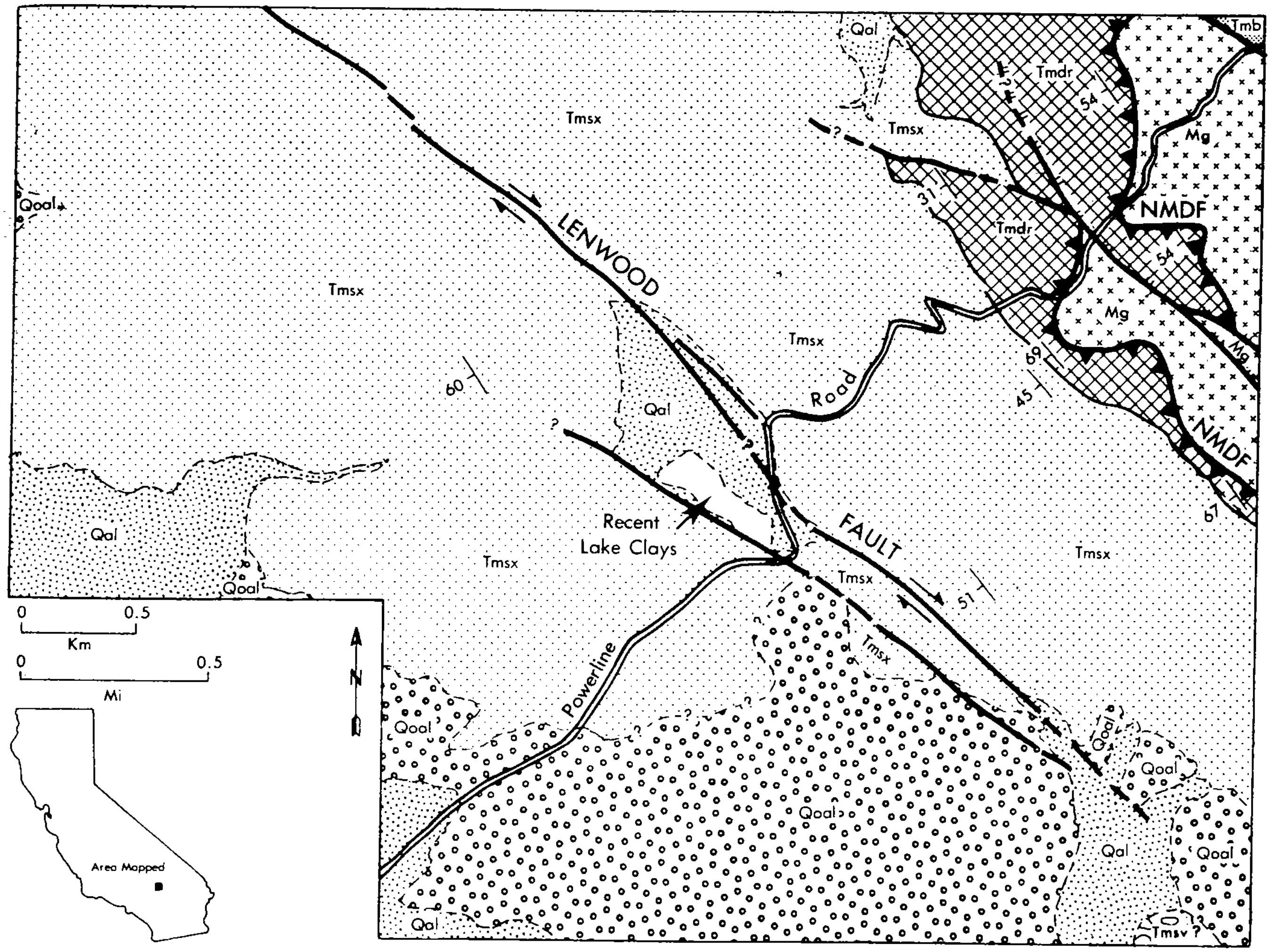

Figure 4. Geologic map of the area covered by figure 3. Symbols are standard. NMDF = Newberry Mountains detachment fault; Mg = Mesozoic crystalline rocks; Tmsx = Slash X Ranch breccia; Tmdr = Daggett Ridge fm; Tmb = Barstow Formation.

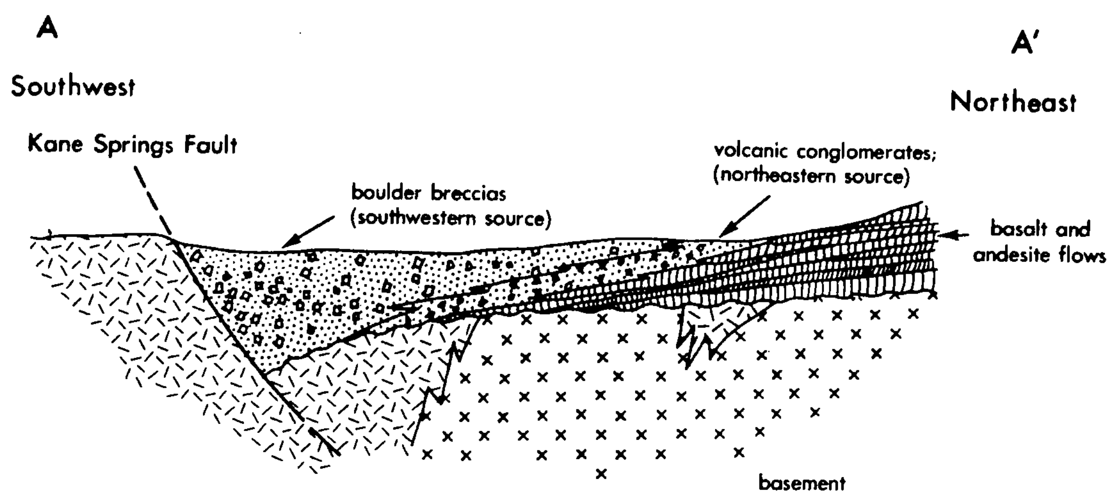

Figure 5. Paleogeologic cross section Stoddard Valley and Daggett Ridge (from Dokka, 1980).

top of the grade you will be traveling down section through the upper part of formation of Slash X Ranch. Ledge-forming out-crops are composed of Sidewinder Volcanic Series whereas the slopes are underlain by granite-volcanic-bearing conglomerates and breccias. From the top of the grade east, one encounters rolling hills composed of the formation of Slash X Ranch and distinctive green to gray weathering volcaniclastic alluvial fan deposits of the formation of Daggett Ridge. Approximately 2 miles from Stop #2 the road intersects the Newberry Mountains detachment fault (fig. 4). Moderately dipping sedimentary rocks dip directly sharply into the contact. Lower plate rocks are locally intensely sheared. Emerging from the rolling hills, the road cuts through several exposures of middle Miocene Barstow Formation. These outcrops consist of limestone, marl, mudstone, siltstone, and sandstone. Continuing to the east, the road traverses several generations of Quaternary alluvial fan and stream deposits. Turn right at the Camp Rock Road (it is a well-graded gravel road). Reset your odometer and continue south on the Camp Rock Road approximately 5.1 miles. (Note: at 1.9 miles, the road forks: take the left fork) and turn left onto an unimproved jeep trail. Bearing right at all forks, proceed 1.2 miles east and park at the crest of the ridge (Do not continue past the crest unless you have 4-wheel drive!).

STOP #3 - AZUCAR MINE WINDOW

Objectives

1. Observe the Newberry Mountains detachment fault.
2. Discuss interpretations of this contact.
3. Observe a thick section of tectonic micro-breccia developed beneath the Newberry Mountains detachment fault.

A generalized structure map of the Newberry Mountains is shown on figure 6. Climb up the nearby slope to the north and study the contact exposed there between the granite-quartz monzonite and the lower Miocene fanglomerate (point A on figure 7). Two interpretations have been proposed for this contact. Dibblee (1970, 1971) considered this surface to be the "great Middle Tertiary Buttress Unconformity" of the Mojave Desert along which 25,000 ft. of sedimentary and volcanic rocks were deposited. Dokka (1976, 1979, 1980) interpreted this surface as a regional low-angle fault.

Evidence for a dislocation origin of this surface include: 1) extreme shearing along the contact that is expressed by coherent and incoherent breccias; 2) occurrence of fault-related kinematic indicators along the fault; 3) truncation at a high angle of upper plate strata along the fault; 4) sedimentary and volcanic rocks juxtaposed across this zone do not contain any clasts of the adjacent basement rocks.

This tectonized zone has had at least two separate stages of development. Formation of the thick (a few meters to greater than 100 meters) microbreccia apparently occurred first and was later cross-cut by centimeters thin, subparallel zones of cataclasite and microbreccia. In thin section, these narrow zones are composed of phacoidal shaped porphyroclasts that rest in a more comminuted matrix (fig. 8). New minerals that have grown in this rock are mainly quartz and calcite, with very minor amounts of chlorite. Shears within the thick unit have three common orientations. One group is subparallel to the overall orientation of the unit whereas the others, a conjugate set, strike northeast and dip at low angles to the other set.

Follow the contact down the canyon until it intersects the jeep trail. Proceed to Point B (fig. 7).

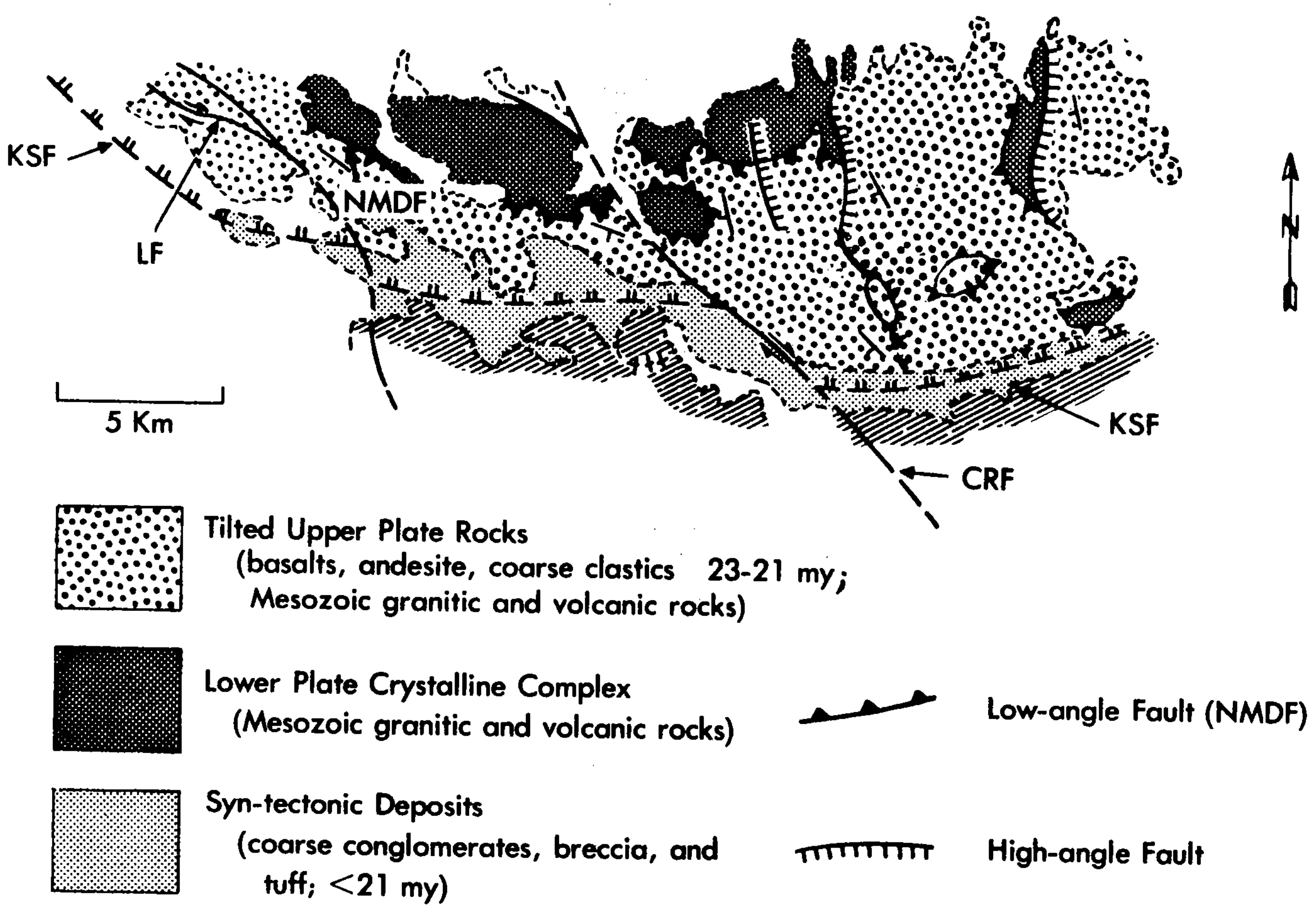

Figure 6. Generalized geologic map of the Newberry Mountains. NMDF = Newberry Mountains detachment fault; KSF = Kane Springs fault; LF = Lenwood fault; CRF = Camp Rock fault

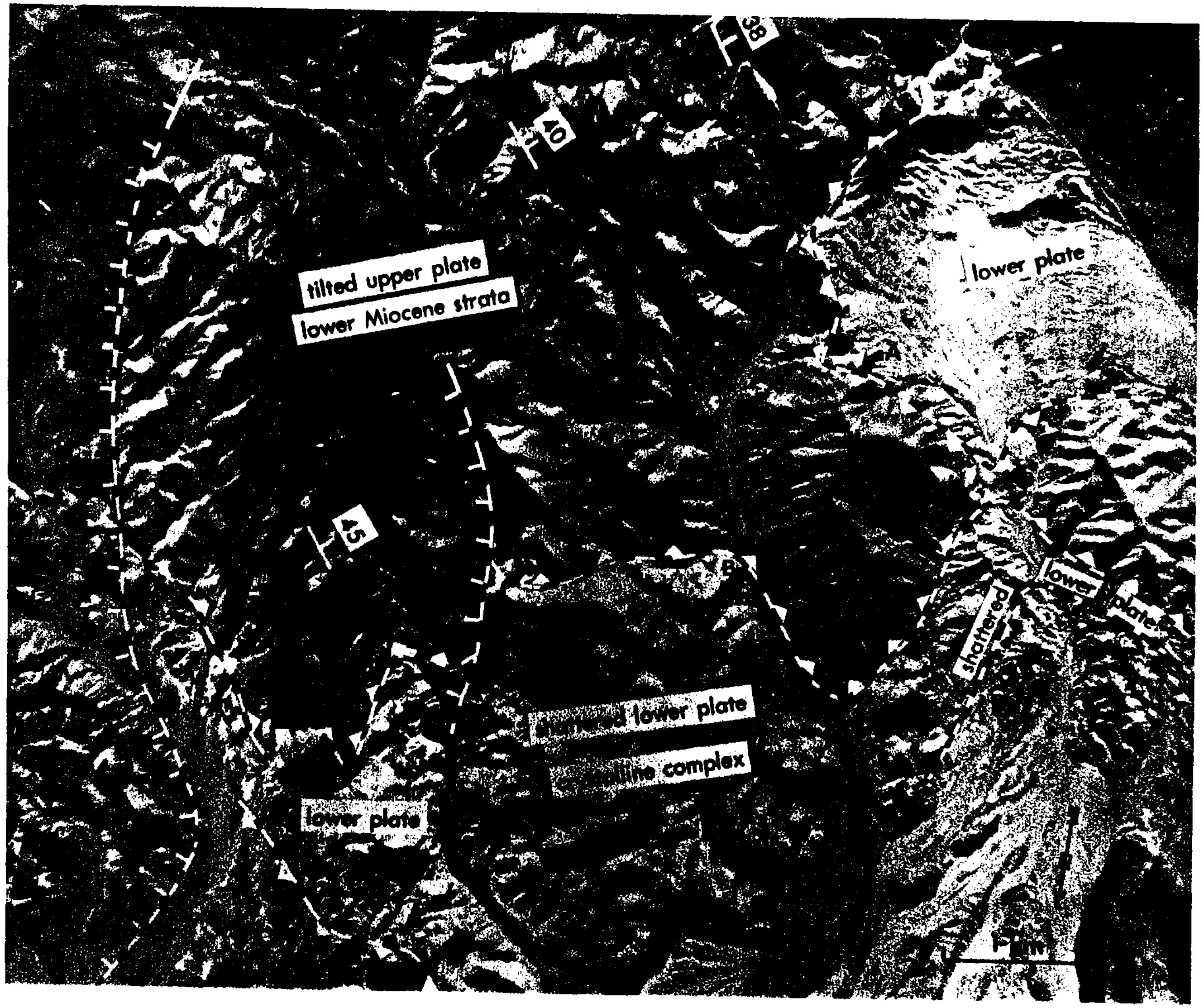

Figure 7. Airphoto of a portion of the northwestern Newberry Mountains (east of the Camp Rock Road). The Newberry Mountains detachment fault (NMDF) and associated upper plate high-angle faults are highlighted. Note the curviplanar character of the NMDF and the uniformly tilted package of lower Miocene volcanic and sedimentary strata that lie above it.

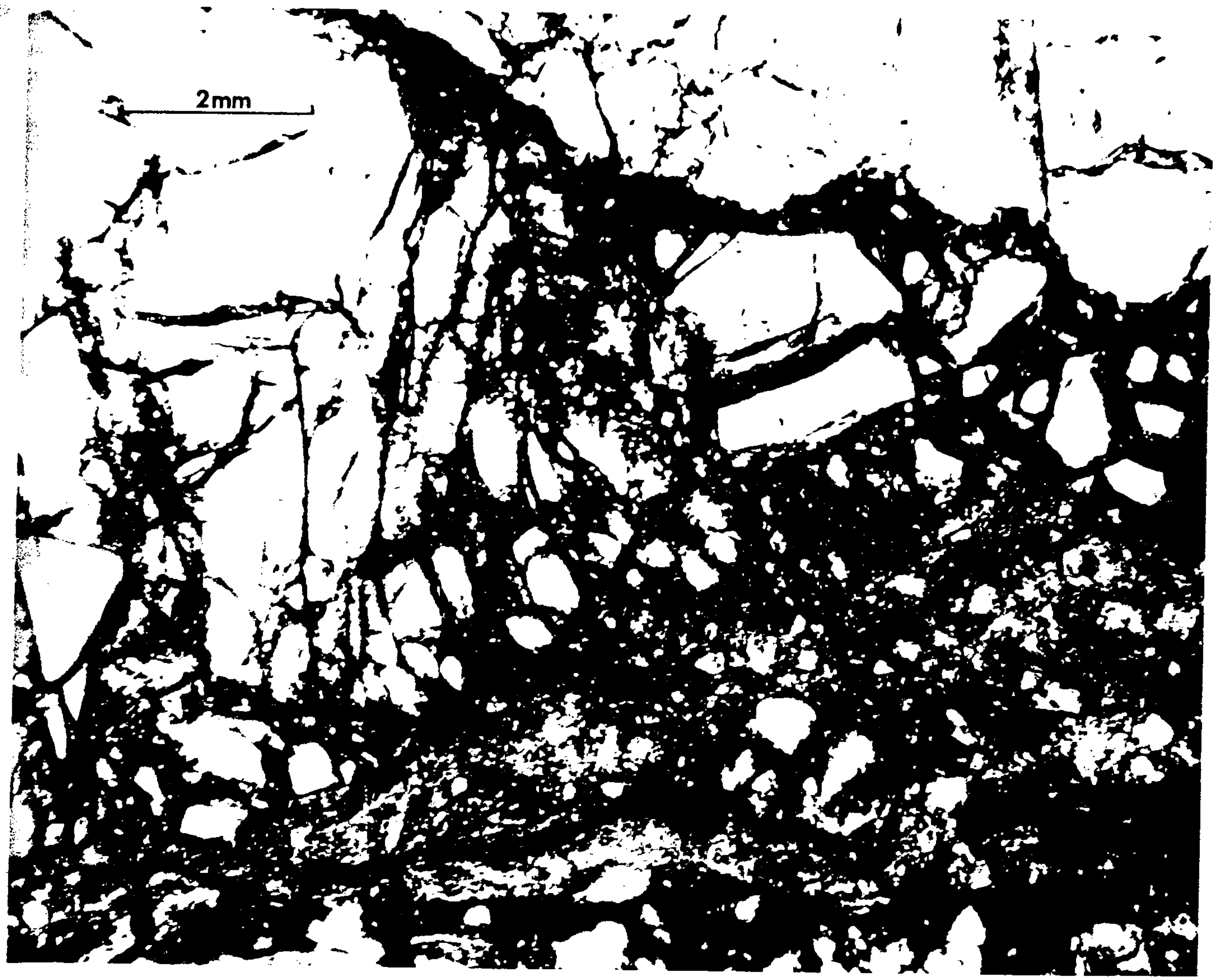

Figure 8. Photomicrograph of the tectonic microbreccia. Sample was taken 1 m below the NMDF.

At Point B, observe the thick (several hundred feet) section of tectonic microbreccia lying below the NMDF. As can be seen on figure 7, the NMDF is not flat but an undulatory surface. Here the fault is dipping moderately to the south separating cataclasized granite-quartz monzonite from undeformed lower Miocene andesite, basalt, and conglomerate. Kinematic indicators are sparse along the NMDF but consistently suggest lateral transport along a NE-SW line (N.51°E. $\pm$ 13°) (Dokka, 1980).

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Return to the Camp Road, turn right, and proceed back to Interstate 40. Reset your odometer at the interstate. Get on I-40 heading east (toward Needles), continue approximately 11.2 miles and exit at Newberry Springs. There are several important features that can be seen from the highway along this leg of the journey. The northern Newberry Mountains (two miles from the onramp) are fringed by an impressive family of Quaternary (and older?) alluvial fans. These fans project deeply into the interior of the range and contain rock fragments that could have only been derived from the Ord Mountains (located south of the Newberry Mountains (figure 1). These rocks call out for someone to study them!

The uniformly tilted pattern of lower Miocene volcanic rocks of the Newberry Mountains are well-exposed on the southern skyline. The orange-red weathering, rugged low hills are composed of the tectonic microbreccia we just saw at Stop #3. The man-made structure to the left (@ 2.5 miles) that looks like a drive-in movie for Titans is actually a solar power generating station, owned by Southern California Edison. "Solar One" as it is called, is scheduled to be on line by late 1982. Upon exiting at the Newberry Springs offramp, turn right and then turn right again on Quarry Road (0.05 miles past offramp). Reset your odometer and drive approximately 0.3 miles and turn left onto a dirt road. Follow the well-traveled path 1.5 miles into the range until the road forks. Take the left road that leads into the small canyon then proceed 0.2 miles and park.

STOP #4 - NEWBERRY SPRINGS KLIPPE

Objectives

1. Observe a low-angle detachment fault (presumed to be the NMDF).

2. Observe evidence for post-detachment fault extension by normal faulting.

3. Observe a section of lower Miocene dacites and tuffs.

The geology of this area is depicted on figure 9. The low-angle fault that is so spectacularly exposed on the north wall of this canyon separates an upper plate consisting of lower Miocene volcanic rocks (vent deposits, viscous flows, ash-flows, and feeder dikes) from a sheared and shattered lower plate (granite-quartz monzonite, Sidewinder Volcanic Series) (fig. 10). These upper plate rocks have been informally designated as the formation of Newberry Springs (Dokka, 1980). The exposed section of these rocks is about 2.1 km thick, but can only be considered as a minimum because the lower contact is not exposed.

A high-angle, east-dipping normal fault passes through the saddle at point A (figure 9). This major north-striking fault traverses the range and has juxtaposed the upper member of the formation of Newberry Springs against the crystalline basement. Kinematic indicators observed along the fault consistently suggest down-dip movement. Because the truncated low-angle fault is not seen again in the hanging wall to the east, a minimum slip of 1 km must be considered on this fault.

Three members of unit can be seen at this locality. The lowest unit crops out at point A (fig. 9) and consists of tuff breccia and lapilli tuff. These rocks contain granite xenoliths that were probably ripped out of the volcanic conduit. The middle unit, a sequence of pale purple hornblende-biotite dacite flows, constitutes much of the hill above the low-angle fault. An irregular pattern of flow foliations suggests the unit experienced locally intense churning during viscous flow. Lying on top of this unit is a sequence of massive dacites and tuffs that were deposited as viscous flows and ash-flows, respectively. These rocks are volumetrically the most important unit in the eastern Newberry Mountains and make up most of the rugged mountain-side to the east.

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Return to the vehicles and proceed to the intersection of Quarry Road and National Trails Highway. Reset your odometer. Turn right on National Trails Highway and proceed 2.8 miles to Poniente Rd. Turn right (south) and proceed 1.0 miles to an exposure of the Calico fault. We will be passing through the town of Newberry Springs, named for an important watering hole that served the local residents in the early 1900's.

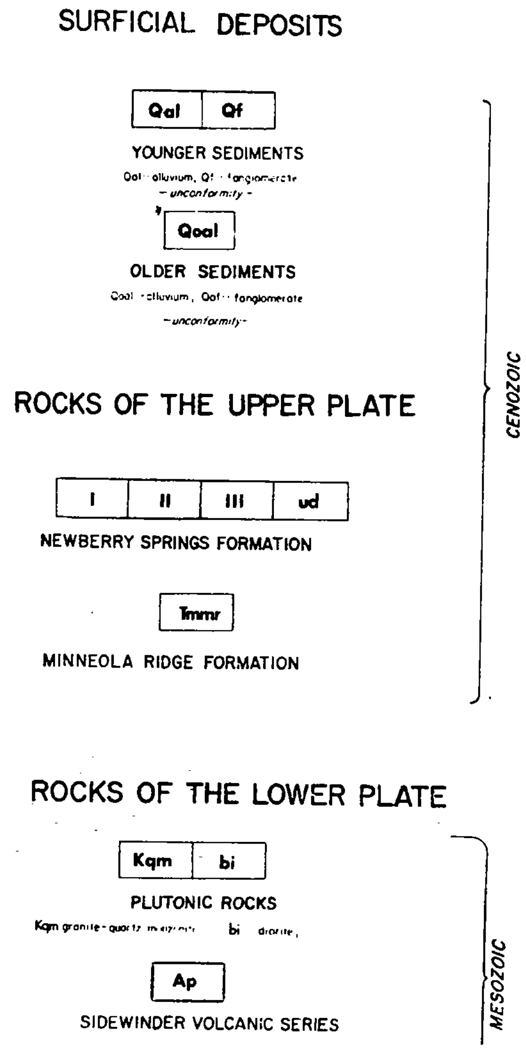
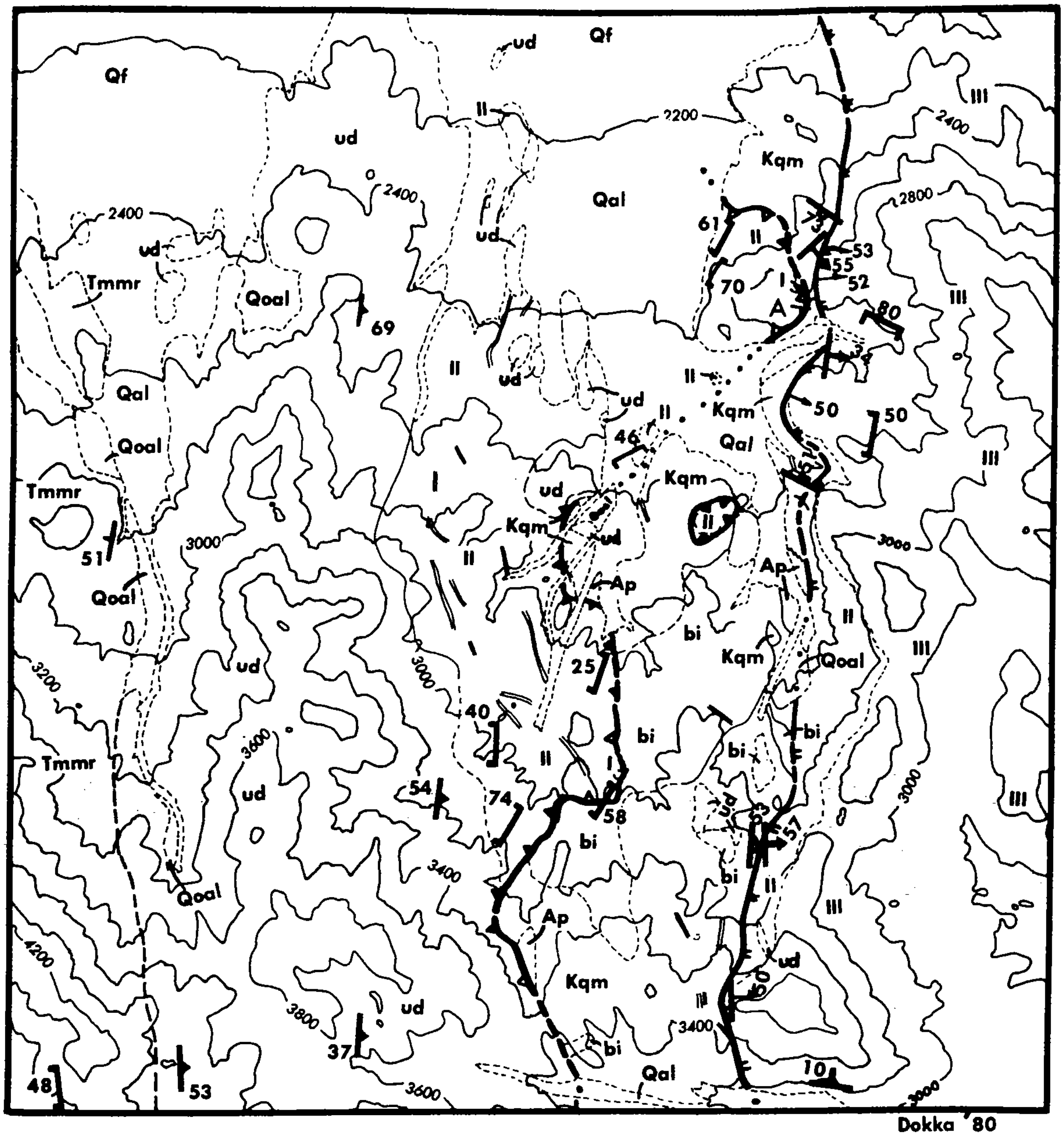


Figure 9. Geologic map of the Newberry Springs klippe area (from Dokka, 1980). Formation names are informal.



STRIKE AND DIP OF STRATA

CONTACT

dashed where approximate, questioned where queried

STRIKE AND DIP OF FOLIATION

HIGH-ANGLE FAULT

dashed where indefinite, dotted where concealed; queried where doubtful U, upthrown side; D, downthrown side

STRIKE AND DIP OF JOINTING

LOW-ANGLE FAULT

barbs on upper plate; lower plate generally cataclasized

DIKES

1.5 km

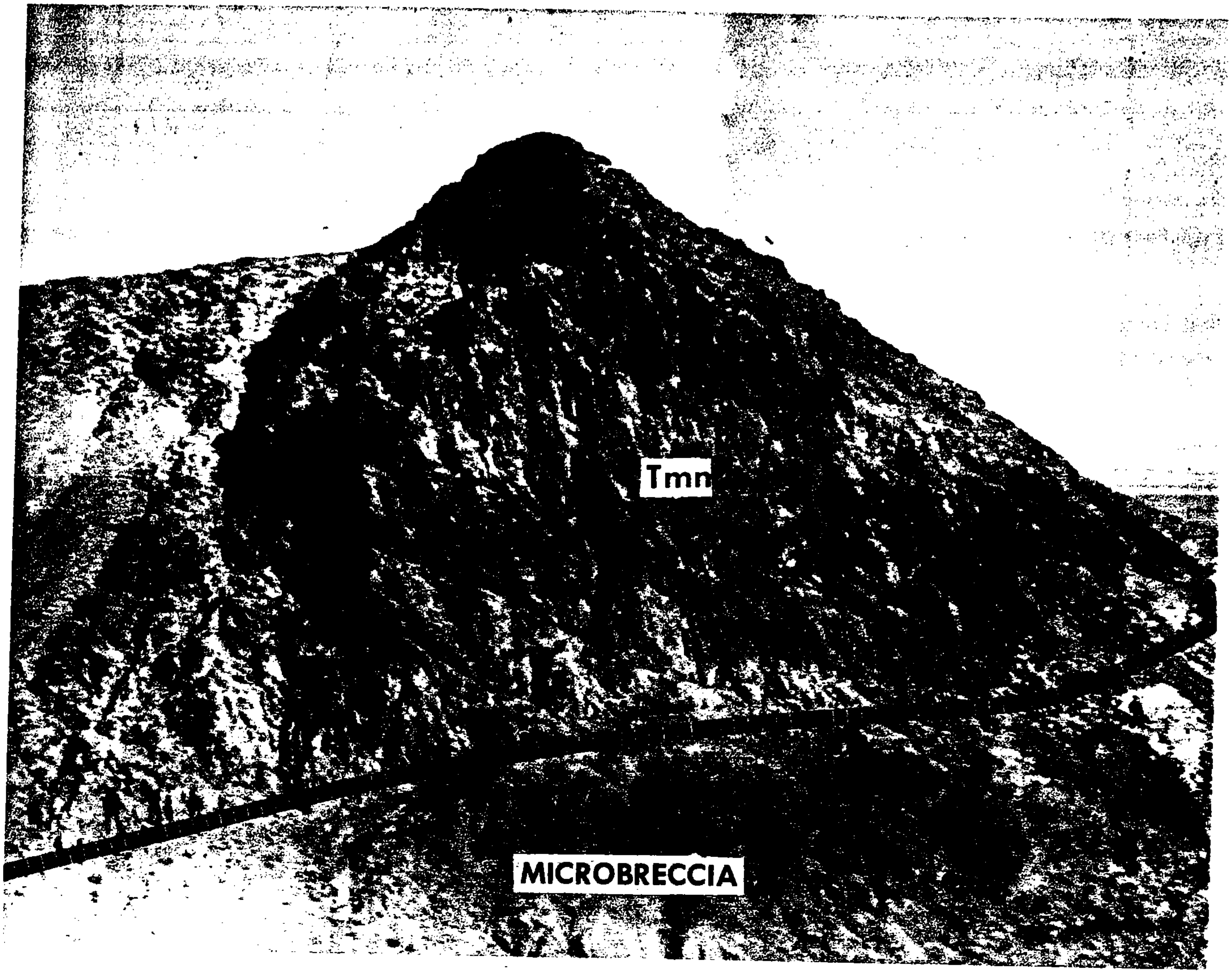


Figure 10. Low-angle fault located near Newberry Springs. View is to the north. Upper plate rocks are dacite flows and tuffs. Lower plate rocks are Mesozoic granite and volcanic rocks that have been cataclasized.

STOP #5 - CALICO FAULT

Objective

1. To observe the geomorphic effects of the Calico fault.

The Calico fault, like the Lenwood and Camp Rock faults to the west, is part of a family of northwest-striking, right-slip wrench faults that traverse the central Mojave Desert. These faults, along with their associated shortening and extensional structures, are responsible for giving this region much of its present-day structural grain and topography.

Look to the northwest and note the following fault-related features: 1) a sharp break in vegetation; 2) sharp, straight boundary of the playa; and 3) slightly stepped topography. Strike separation along the Calico fault is approximately 8.2 km. If strain effects are added, then the figure increases to 9.6 km (Dokka, ms).

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Return to the vehicles and trace our route back to Interstate 40. Proceed 6.2 miles west on I-40 back to the Daggett exit, turn right, and head north on Daggett Rd. to the intersection with Interstate 15. Follow the signs to the Calico Ghost Town parking lot (approximately 3.4 miles).

STOP #6 - CALICO MOUNTAINS

Objectives

1. Observe middle Miocene Barstow Formation.

2. Observe post-middle Miocene gravity-induced folds.

The Calico district was an active silver producing area from 1881 until the middle 1890's. Silver ore containing as much as 200 oz./ton were extracted during this interval from the variegated upper Tertiary volcanic rocks that comprise the Calico Mountains. Silver occurs in steeply-dipping, northwest trending veins. Ore minerals of silver consist mainly of cerargyrite (AgCl), argentite (Ag_2S), and embolite (Ag(Cl,Br)) that reside in a gangue of jasper and barite. Lesser amounts of gold, barite, lead, and copper have also been mined as well as borate minerals (colemanite, probertite) and, most recently, decorative stone. Billingsly (1929) reported that as much as 50 million dollars worth of silver was mined during the early years. In 1975, ASARCO reported that an ore body of about 30 million tons had been located just west of the ghost town. For a reported average of 3 oz./ton, this discovery (at late 1981 silver prices) is worth over one billion dollars!

The Barstow Formation at and near the ghost town consists mainly of thinly bedded shales, mudstones, sandstones, and conglomerates (see the article by Woodburne and others, this volume) for a more complete discussion of the Barstow Formation). Link (1979) considers these rocks to represent continental lacustrine, nearshore and shoreline depositional environments. Ledge-forming limestone, pastel-colored tuff and colemanite also occur higher in the section. In the Calico area, the Barstow Formation overlies lower Miocene Pickhandle Formation, a thick (1500m; McCulloh, 1952) accumulation of intermediate composition flows, tuffs, breccias, and volcaniclastic sedimentary rocks along with granitic breccias, conglomerates and sandstones. The Barstow Formation, in turn, is overlain by unnamed andesitic flows and volcanic breccias (DeLeen, 1949; McCulloh, 1952; Mayo, 1972). Recent K-Ar studies performed on rocks lying above and below the Barstow Formation bracket the age of the unit as 16 to 13 m.y. old (Susan Miller, U.S.G.S., personal communication, 1980).

The spectacular south-vergent folds in Barstow Formation rocks exposed at the ghost town

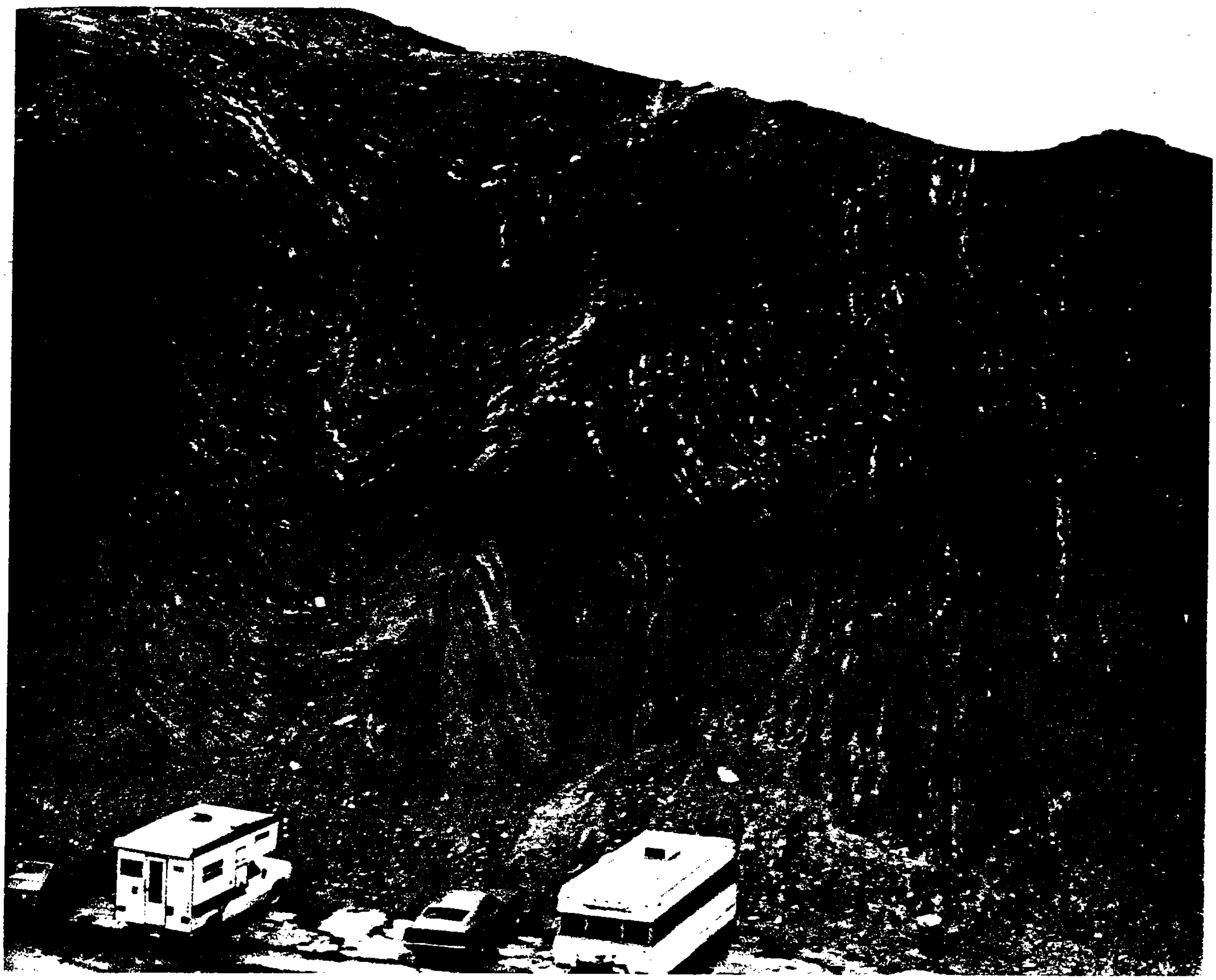

Figure 11. Chevron and rounded folds at Calico Ghost town parking lot. The rocks are middle Miocene Barstow Formation.

being the result of downslope movement (Mayo, 1972; and Weber, 1976). Relations in the range (especially in Mule Canyon, east of Calico) show abundant evidence for additional gravity-induced structures such as decollement faults (Fig., 12). Kinematic indicators suggest that detached sheets and cascade folds moved to the south off a rising highland located to the north (Mayo, 1972; Weber, 1976). The highland may have formed as a result of doming associated with igneous intrusion or perhaps to block uplift along a restraining bend of the Calico fault.

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Return to the vehicles and proceed back to the Calico Loop Road. Turn right (west) and drive 1.0 mile to Yermo Cutoff. The brown-red mountain located to the west-southwest, just to the left of Interstate 15, is Elephant Mountain. It is a shallow intrusive plug composed of middle Miocene (?) andesite. Dibblee (1971) suggested that the plug intruded Barstow Formation at the time of its (Barstow) deposition.

Turn right onto Yermo Cutoff and proceed 5.7 miles to Fort Irwin Road. As you travel along this road, look back to the south across the valley at the sawtooth skyline of the Newberry Mountains. To the right, the abrupt front of the Calico Mountains is the result of strike-slip movements along the Calico fault. Turn left at Fort Irwin Road and drive 3.2 miles to Owl Canyon Campground-Rainbow Basin turnoff (watch for the sign). In the opposite direction, Fort Irwin Road leads to a major army base of the same name and to the Goldstone Tracking Station. Goldstone is a major communications link with deep space probes such as Pioneer, Viking, and Voyagers 1 and 2. Also located here is the major antenna of ARIES (Astronomical Radio Interometric Earth Surveying) network, a system that measures distances between points on Earth by precise measurement of pulses from extragalactic radio sources. The distance between Goldstone and a radio telescope in Spain can be measured to within a few centimeters!

Turn right and drive 2.9 miles on the dirt road and turn left at the "Y" intersection to the Rainbow Basin Natural Area.

Figure 12. Décollement in middle Miocene Barstow Formation sandstones and shales; location - Mule Canyon, Calico Mountains, California.

STOP #7 - MUD HILLS - RAINBOW BASIN

Objectives

1. Observe stratigraphic relations of the Barstow Formation.

2. Examine fossils of the type Barstovian North American land mammal stage.

3. Observe folds and faults associated with late Cenozoic strike-slip faults.

The Mud Hills area (Solomon Canyon, 2 miles east) is the type area for the middle Miocene Barstow Formation, a package of continental lake and basin-edge deposits (Fig. 13). Maps and cross-sections of the Mud Hills are given on figures 14 and 15. In this area, the unit is approximately 650m thick (Dibblee, 1967) and is

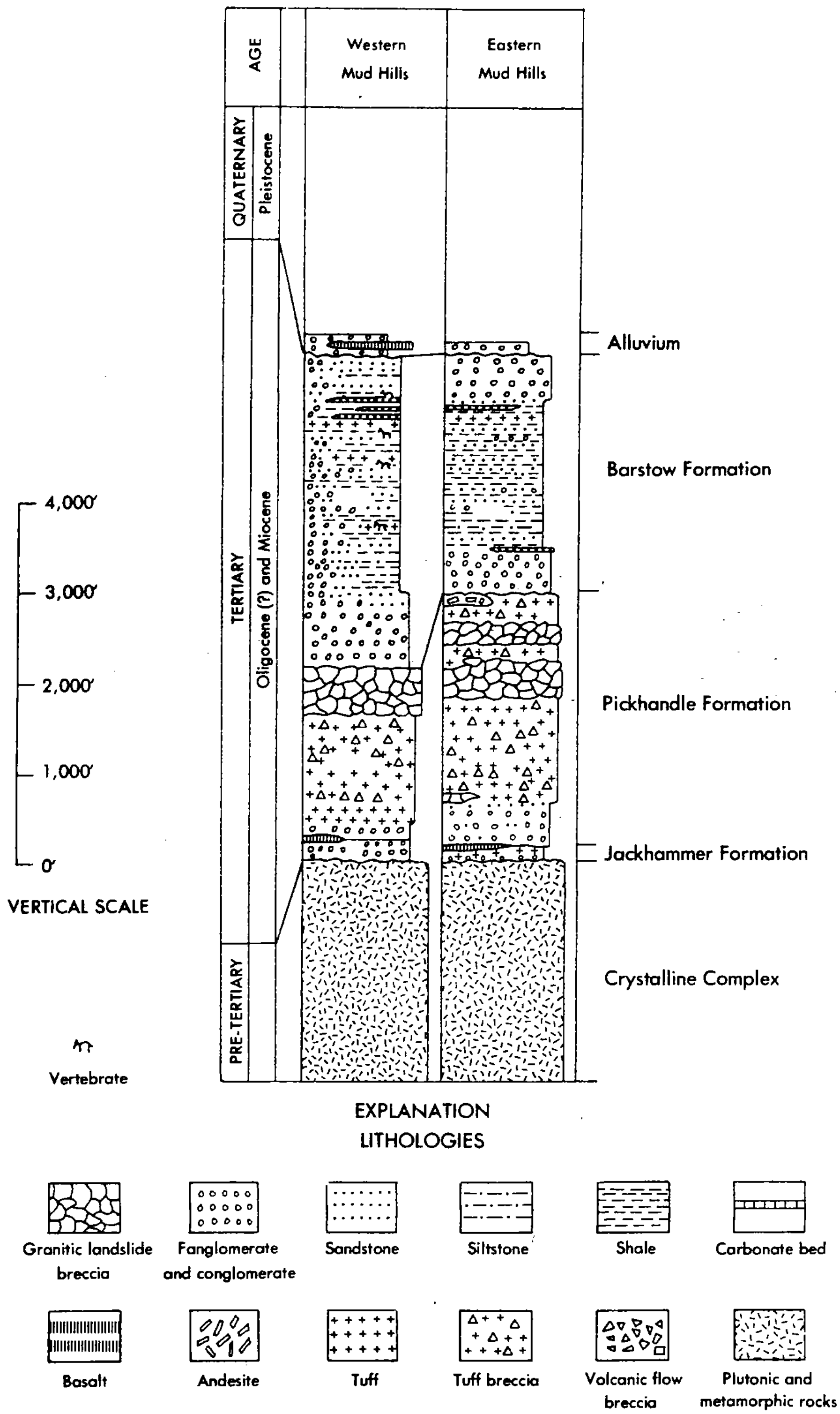

Figure 13. Stratigraphy of the Mud Hills, California (from Dibblee, 1967).

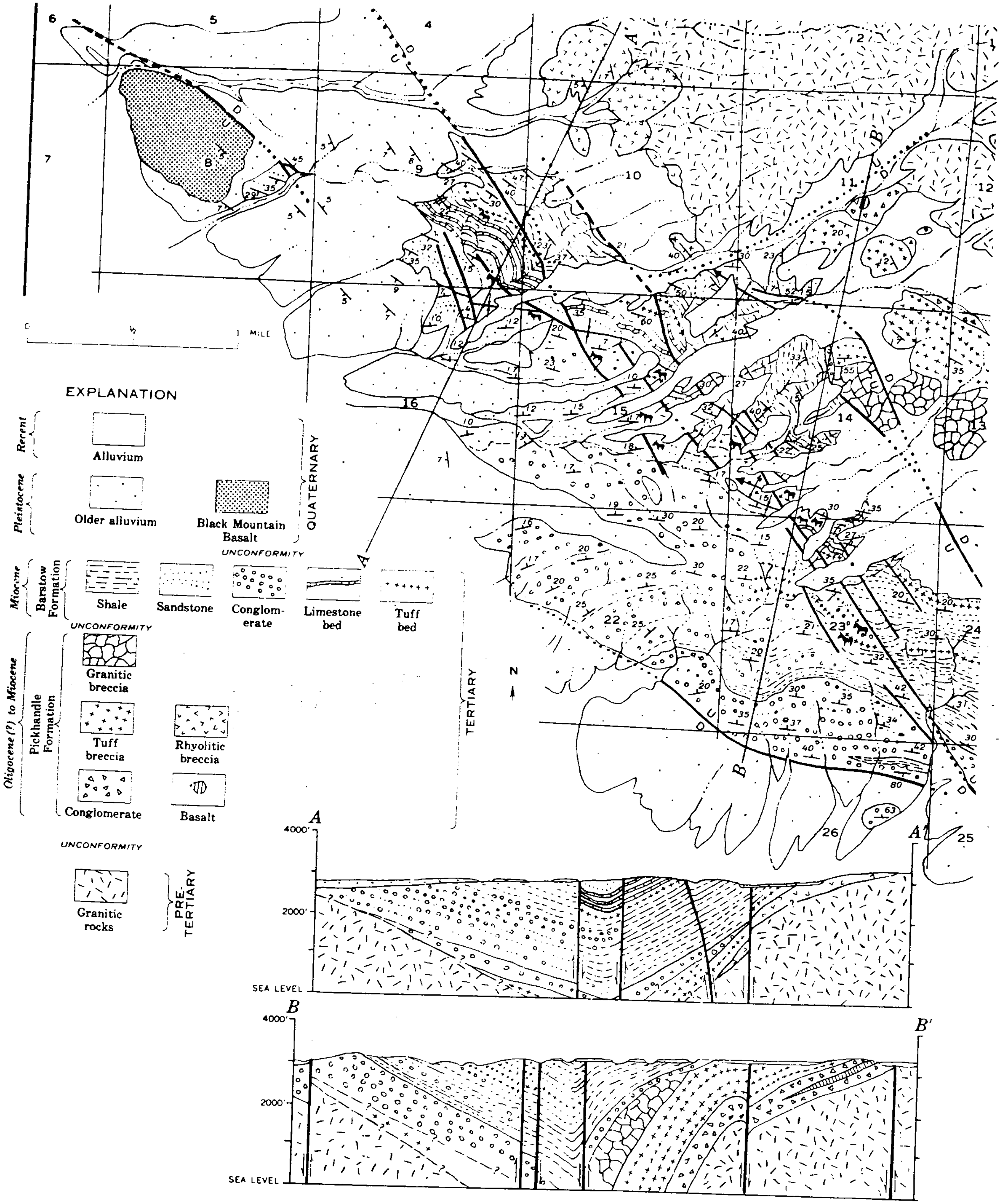

Figure 14. Geologic map of the western Mud Hills, California (from Dibblee, 1967).

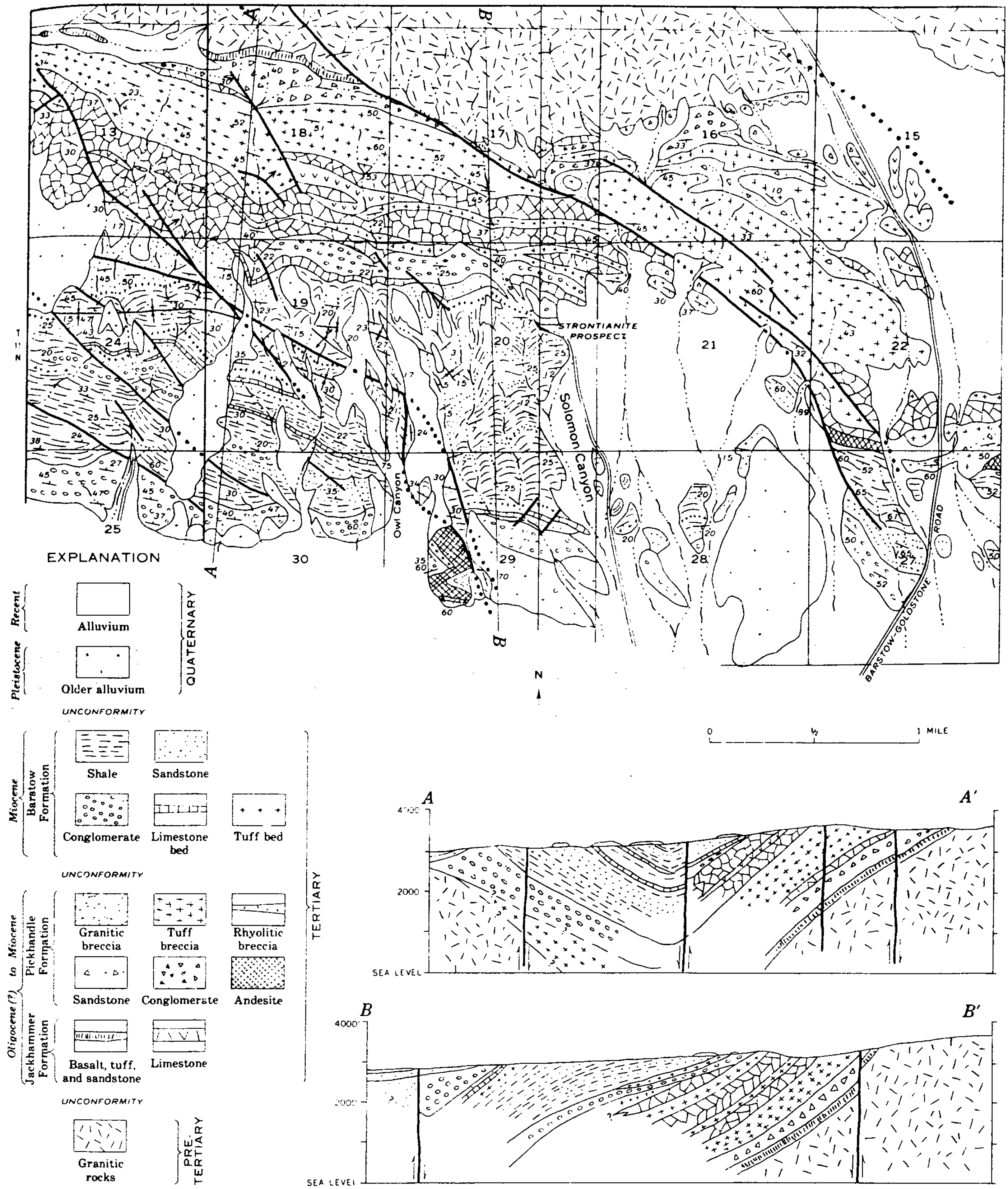

Figure 15. Geologic map of the eastern Mud Hills
(from Dibblee, 1967).

composed of shale, sandstone, and conglomerate along with minor amounts of limestone and tuff. Rocks here have been folded into broad, open, west-trending folds in association with a family of short, northwest-striking, right strike-slip faults.

The Barstow Formation contains a wealth of middle Miocene mammalian and molluscan fossils. J. C. Merriam (1919) first described the assemblage in detail and found the bones and teeth of small, three-toed horses (Merychippus intermontanus), antelopes (Merycondus), and camels. Merriam inferred that the Barstow area was a warm, grassy plain during Barstovian time. The best fossil localities occur in the western part of the range.

PLEASE NOTE: Collecting of any kind (fossils, rocks, flora, fauna) is strictly forbidden within the park.

* * * * * * * *

Return to the vehicles and proceed to motels in Barstow or to Owl Canyon Campground.

END OF DAY 1

DAY 2

After meeting in Barstow for breakfast, gas, and supplies, get onto Interstate 15 north (to Las Vegas). A map of todays stops is given on figure 16.

The road out of Barstow passes through hills of Miocene sediments and dark red Miocene dacite (?) plugs. After passing the southern tip of the Calico Mountains, we pass by the Calico Early Man Site. The site is a key element in the history of Man in North America. Evidence at Calico suggests that the site was occupied at least 50,000 years ago, and perhaps much earlier. This is far older than previous estimates for the antiquity of Man in North America.

Artifacts at Calico are found in a dissected alluvial fan (the Yermo Formation) which crops out on the southeastern slopes of the Calico Mountains. Artifacts occur in an unsorted mudflow facies of the fan complex, and are composed mostly of chalcedony derived from the Barstow Formation in the Calicos. The fan apparently contains no material suitable for either isotopic or paleomagnetic dating, so estimates of its age are based on geomorphology. Clements (1979) believes that the fan formed about 75,000 years ago, before the start of the Wisconsin glaciation. Haynes (1973) estimated the fan's age at about 500,000 years, based on somewhat looser arguments than Clements'. Haynes used this extreme age as part of his argument that the artifacts are not artifacts at all, but "geofacts" produced by normal erosional processes in mudflows. Lee (1979) countered Haynes' argument by noting that striking two boulders together in a mudflow (matrix-supported conglomerate) is like putting two boulders together in a bag of dough and trying to strike them together.

The interpretive program at the Early Man Site is excellent, so visitors have a chance to decide for themselves which argument holds up best. For the geologist, the pit tour offers a unique chance to view the inside of an alluvial fan. A self-guided tour around the pits explains the history of the Site and affords a close look at a well-developed desert pavement.

About 5 miles past the Early Man Site, Harvard Hill, 736 m in elevation, is on the right. This hill, composed of Miocene sediments an volcanic rocks, was an island in pluvial Lake Manix.

Soon after Harvard Hill, the Alvord Mountains come into view on the left. This end is composed of Mesozoic granitoids enclosing a belt of pre-Mesozoic metamorphic rocks. The broad white stripe is a marble layer which does not varnish darkly in the desert climate as do the enclosing quartz-feldspar rocks.

At the Field Road exit, the sand-mantled mountains on the right are the northern end of the Cady Mountains. The Cadys are dominantly Cenozoic volcanic and sedimentary rocks, but on the northern end some Mesozoic metavolcanics and granitoids are exposed. The mountains were named for Fort Cady, an army post used in the mid 1800s to protect travelers along the San Bernardino-Fort

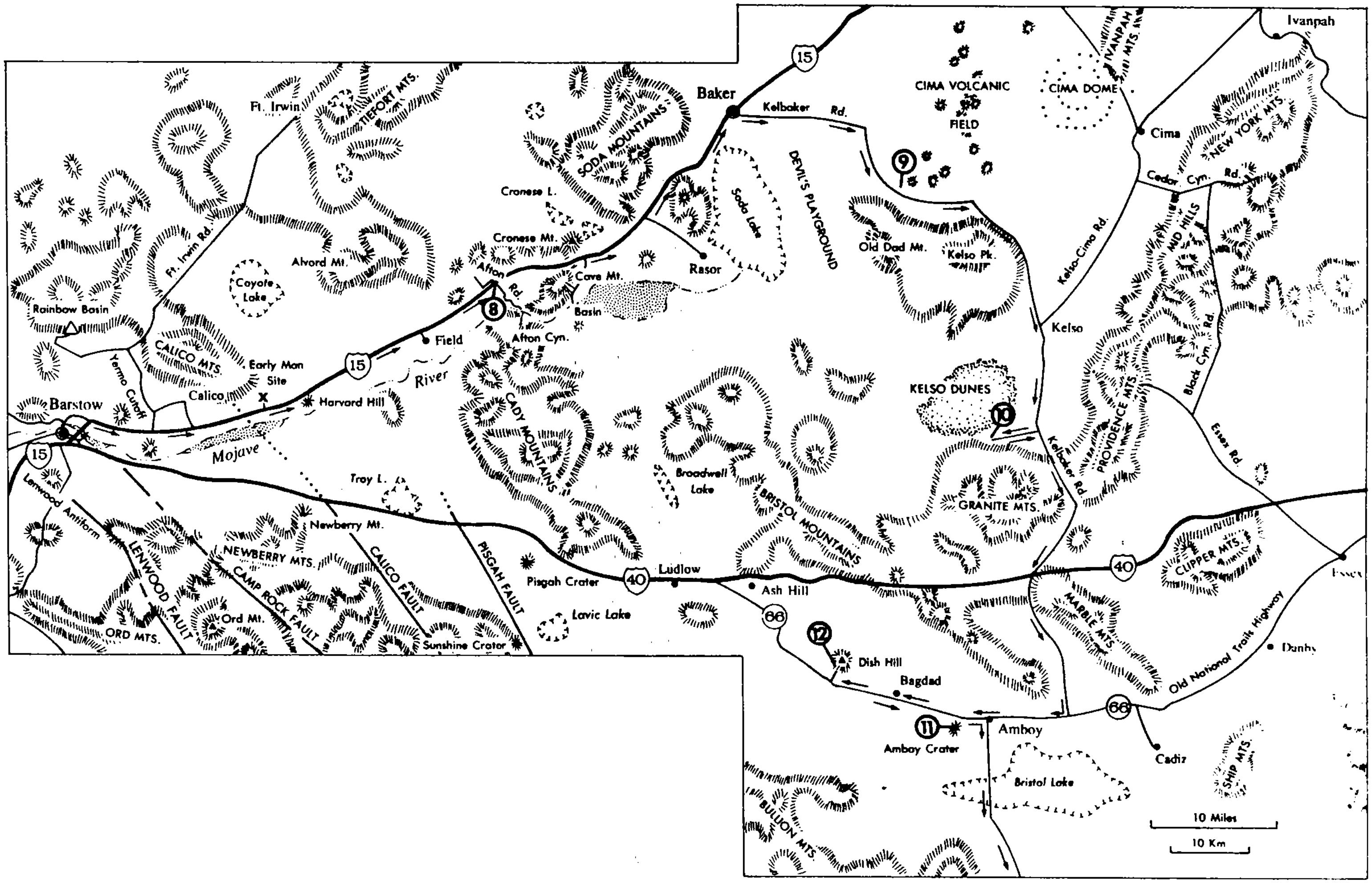

Figure 16. Day 2 - Field trip route. Numbers refer to stops described in text.

Mojave road. The fort was located at the western end of the range.

About a mile past the rest area, the low, horizontally bedded green hills on the right are Quaternary Manix Lake sediments.

Exit at Afton Road. Turn right and park on the dirt road.

STOP #8 - AFTON BEACH RIDGE

Objectives

1. To view a gravel beach deposit formed during an early stand of Pleistocene Lake Manix.

2. To discuss the Pleistocene climate and wildlife of the Mojave.
3. To discuss the significance and seismic history of the Manix fault.

4. To discuss the Tertiary geology of the northern Cady Mountains.

For a detailed discussion of the history of the Mojave River in Manix basin, see Weldon (this volume).

Lake Manix was the largest of the Pleistocene lakes of the Mojave Desert block. During its highest stand it covered much of the basin between Afton and Barstow (Blackwelder and Ellsworth, 1936). It consisted of 3 lobes, the remnants of which now form Coyote, Troy, and Manix Lakes (Fig. 17).

Lake deposits in the Afton area give evidence for 3 separate stands of Lake Manix. The highest of these, which filled the basin to the 1880-foot level (Clements, 1979), is correlated with the Tahoe glacial stage, about 60,000-75,000 years b.p. (Blackwelder and Ellsworth, 1936; Clements, 1979). This stand of the lake produced the gravel bar upon which we are now standing. Lower stands are tentatively correlated with the Tenaya (45,000 years b.p.) and Tioga (20,000 years b.p.) glacial stages.

Two lines of evidence suggest that the Mojave River did not flow through Manix basin before the first lake formed. First, clays of the first lake are not underlain by river deposits; they rest upon fan gravels. Second, there is no evidence in Afton Canyon for a steep gorge that predates the earliest lake deposits. Such a gorge would have been cut by the Mojave River during the long drop (315 m in 22 km) from Manix basin to Soda Lake.

Blackwelder and Ellsworth (1936) offered the following model for the evolution of Lake Manix. A damp climate during the Tahoe glacial stage caused the Mojave River to pond in Manix basin, forming lake no. 1. During the interpluvial period the lake dried up completely and was covered by alluvial fans. Lake no. 2 formed during the Tioga stage and eventually overflowed its basin on the east, cutting the deep gorge that is now Afton Canyon. Lake no. 3 formed at an unspecified later time. The mechanism by which the river was dammed, after having cut Afton Canyon, is not known; Danehy (1954) proposed that movement on local faults blocked the river.

The Manix beds have yielded a diverse group of fossils reflecting Pleistocene life in the Mojave during pluvial times. The fauna included dogs, bears, cats, mammoths, horses, camels, antelopes, bison, sheep, shellfish, turtles, beetles, pelicans, and flamingos (Sharp, 1972). We agree with Sharp (p. 72) that "the picture of a pink flamingo standing stiffly at attention on one leg in the Mojave Desert is a bit incongruous."

The mountain front about 6 km south of here is uplifted along the east-striking Manix fault. In 1947, a magnitude 6.4 earthquake shook the northeastern Mojave Desert. The epicenter was located on the Manix fault, about 15 km southwest of here, and inspection of the epicentral area showed cracks with left-lateral slip of 5-8 cm. Portable seismometers were set up to record aftershocks. To the surprise of seismologists, the aftershocks fell on a north-south line, at right angles to the Manix fault. First-motion studies are consistent with either left-lateral slip on the Manix fault or right-lateral slip on a buried basement fault. Richter (1958, p. 518) supports the second possibility and considers movement on the Manix fault to be an indirect consequence of basement faulting.

East-west faults are abundant in the northeastern portion of the Mojave block although little mapping has been done. East-west faults are rare in the rest of the block, where northwest-striking faults dominate. East-west faults are important to recent models for the late-Cenozoic tectonic evolution of the Mojave, because they are apparently conjugate to the northwest set and help accommodate block rotations (Garfunkel, 1974; Luyendyk and others, 1980). Garfunkel's model predicts that the northeastern quadrant of the Mojave did not rotate, whereas Luyendyk and others predict that it rotated clockwise 70-80 degrees. Current paleomagnetic work in the area will test these hypotheses.

* * * * * * * *

Return to I-15 north

About one mile past Afton Road, Cave Mountain (pre-Tertiary igneous and metamorphic rocks; elevation 3585 feet) looms up on the right. Cave Mountain is one of the landmarks of the Mojave; the extremely steep northern face of the mountain is atypical of granitoids in the Mojave. It may be fault controlled.

Passing Cave Mountain, the road descends into Cronese Valley. In times of flood the Mojave River may split after passing through Afton Canyon, and drain into both Cronese Valley and Soda Lake, which lies ahead. During wet winters, water often stands deep on East Cronese Lake; in 1916, floodwaters accumulated to a depth of 10 feet (Sharp, 1972, p. 74).

A few miles past Basin Road, as we climb out of Cronese Valley, look back at 8 o'clock to see Cat Mountain dune, a large sand dune on the east face of Cronese Mountain that bears an uncanny resemblance to a cat. The dune consists of sand blown eastward over the top of the mountain.

About one mile beyond Rasor Road, look at 3

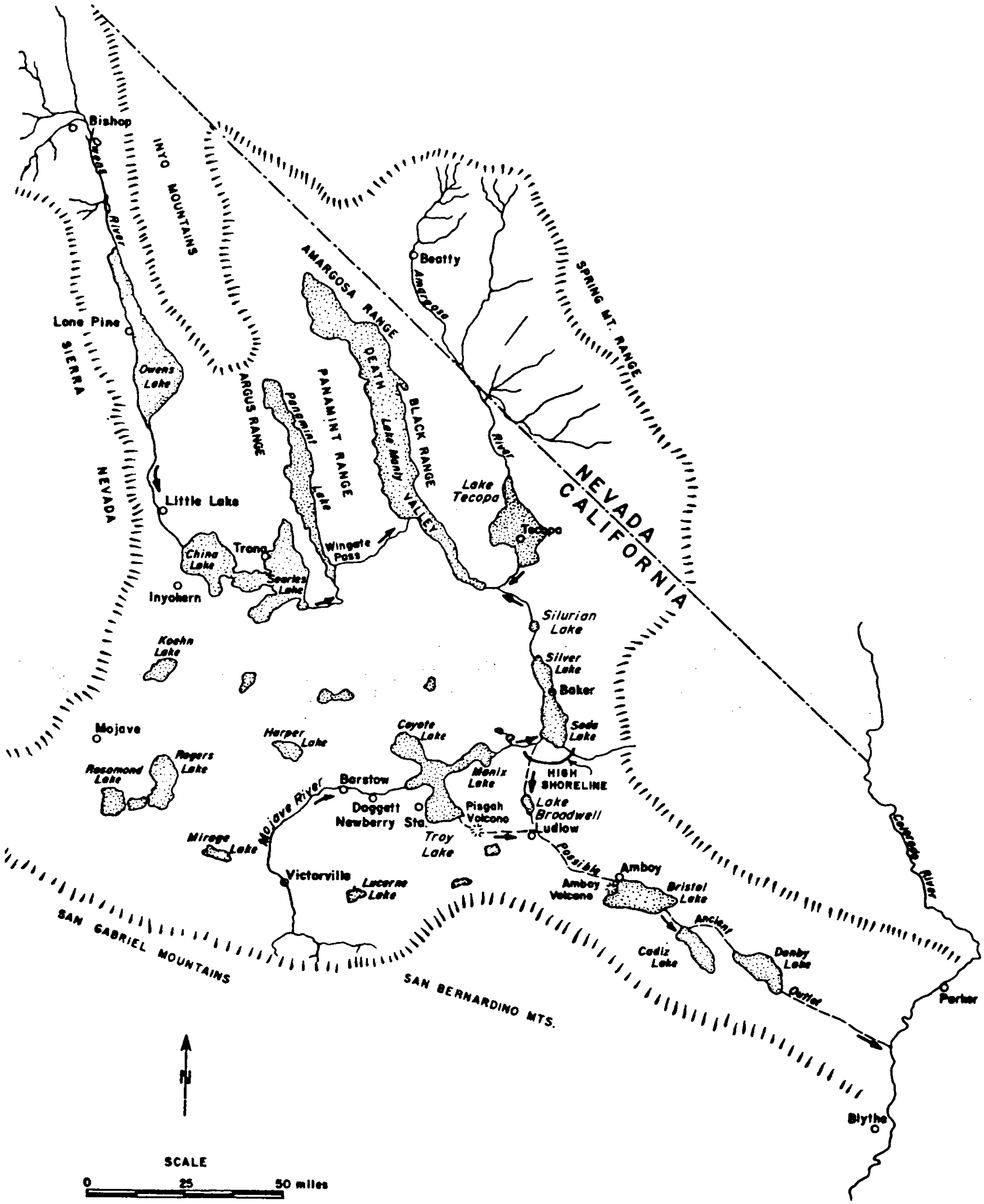

Figure 17. Pleistocene drainage in the Mojave Desert (modified from Blackwelder, 1954).

o'clock to see the Kelso Dunes, 40 km to the southeast. We will visit the dunes later today.

Past Zzyzx Road, the road enters the valley of Soda Lake, which is currently the sink of the Mojave River. Soda Lake and its companion to the north, Silver Lake, are remnants of pluvial Lake Mojave. During floods the Mojave River fills Soda and Silver Lakes as well as Cronese Lakes. Flood waters may persist in these lakes for several years after the storm. After the floods of 1916, remains of many fish were found at the edge of Silver Lake. They were identified as species known to live in pools along the Mojave River and were probably swept into the lake during the flood, where they later multiplied. Such "fish bursts" in otherwise dry lakes can lead to problems in paleoenvironmental interpretation; some future stratigrapher might find these remains and decide that Holocene Silver Lake was perennial.

Silver Lake attracted attention some years back when the Los Angeles Times reported that its surface was suddenly found to be covered with hundreds of small pyramids, a few inches on a side, laid out in geometrical patterns. The display turned out to be the art project of a student at a southern California university.

Coming down the long straightway toward Baker, cinder cones of the Cima volcanic field are in view at 2 o'clock. Our next geologic stop is at a young flow from one of these cones.

* * * * * * * *

<u>Take the first Baker exit and gas up. This will also be a stop to replenish supplies.</u>

Baker, originally named Berry, was named after R. C. Baker, president of the Tonopah and Tidewater railroad, in 1908.

Baker sits astride a low ridge between Soda and Silver Lakes. The road north leads past Silver and Silurian Lakes to the southern end of Death Valley. The impressive mountain range northwest of town is the Avawatz Mountains, a large mass of Precambrian, Paleozoic, and Mesozoic metasediments and plutonic rocks with a thin cover of Tertiary sediments. The Avawatz lie at the junction of the Garlock and Death Valley fault zones, and form the northeastern corner of the Mojave block.

<u>Turn right (east) at the stop sign, onto Kelbaker Road.</u>

The low hills to the right, just south of town, are Paleozoic limestone faulted against Mesozoic granitoids. The Cima volcanic field (see below) lies just ahead.

The mountains at 2 o'clock, about 15 km southeast of Baker, are the Old Dad Mountains. Dunne (1977) described the geology of this range, and recognized several Paleozoic and Mesozoic lithologies, including the Carrara Formation, Goodsprings Dolomite, Bird Spring Formation, and Aztec Sandstone. The section is cut by the Playground thrust fault, which moved Paleozoic rocks eastward or southeastward at least 2.5 km. The fault probably formed during the late Mesozoic Sevier orogeny.

About 11 miles from Baker, the road makes a sharp bend to the right, heading south-southwest to skirt the edge of the Cima field. <u>Note mileage at the bend.</u>

About 3.5 miles past the bend, the road squeezes between a young basalt flow and the northeastern part of the Old Dad Mountains. This flow comes right down to the road and offers good exposures of the morphology. The flow overlies sand, and is not as thick as it looks. It has been dated at 38,000 +/- 37,000 years by R. W. Kistler (K/Ar; A. L. Boettcher, personal communication, 1981). Boettcher feels that it is significantly closer to 0 than to 38,000, based on morphology and weathering. Across the road, the Old Dads are cut by shafts and tunnels of the Oro Fino and Brannigan mines (gold).

About 6-6.5 miles from the bend, the road passes near the youngest cone in the field. Park by the side of the road and walk northeast to the young flow (a 5-10 minute hike).

STOP #9 - CIMA VOLCANIC FIELD

Objective

1. View and discuss Quaternary alkali-basalt volcanism.

The Cima volcanic field comprises about 30 late-Cenozoic alkali-basalt cinder cones and associated flows (fig. 18, 19). It is divisible into an older northern part (up to about 10 m.y. old) and a younger southern part. The southern part was mapped by Katz (1981) and is the subject

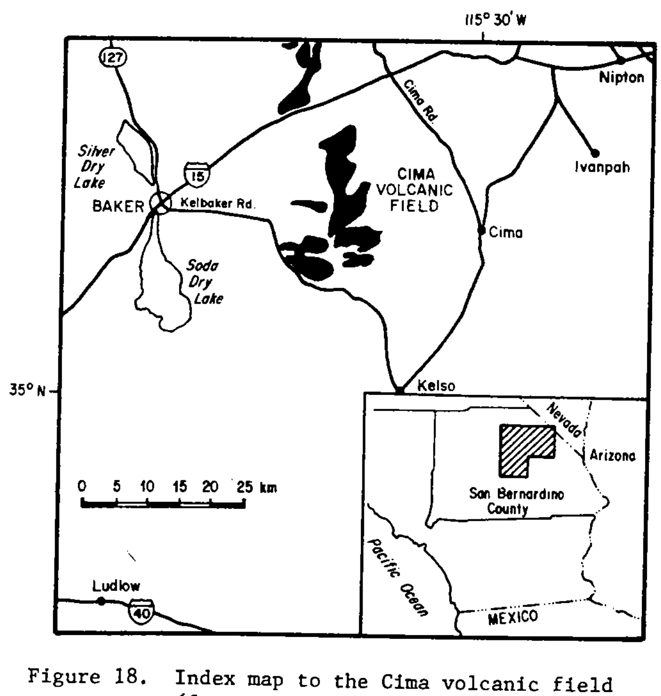

Figure 18. Index map to the Cima volcanic field (from Katz and Boettcher, 1980)

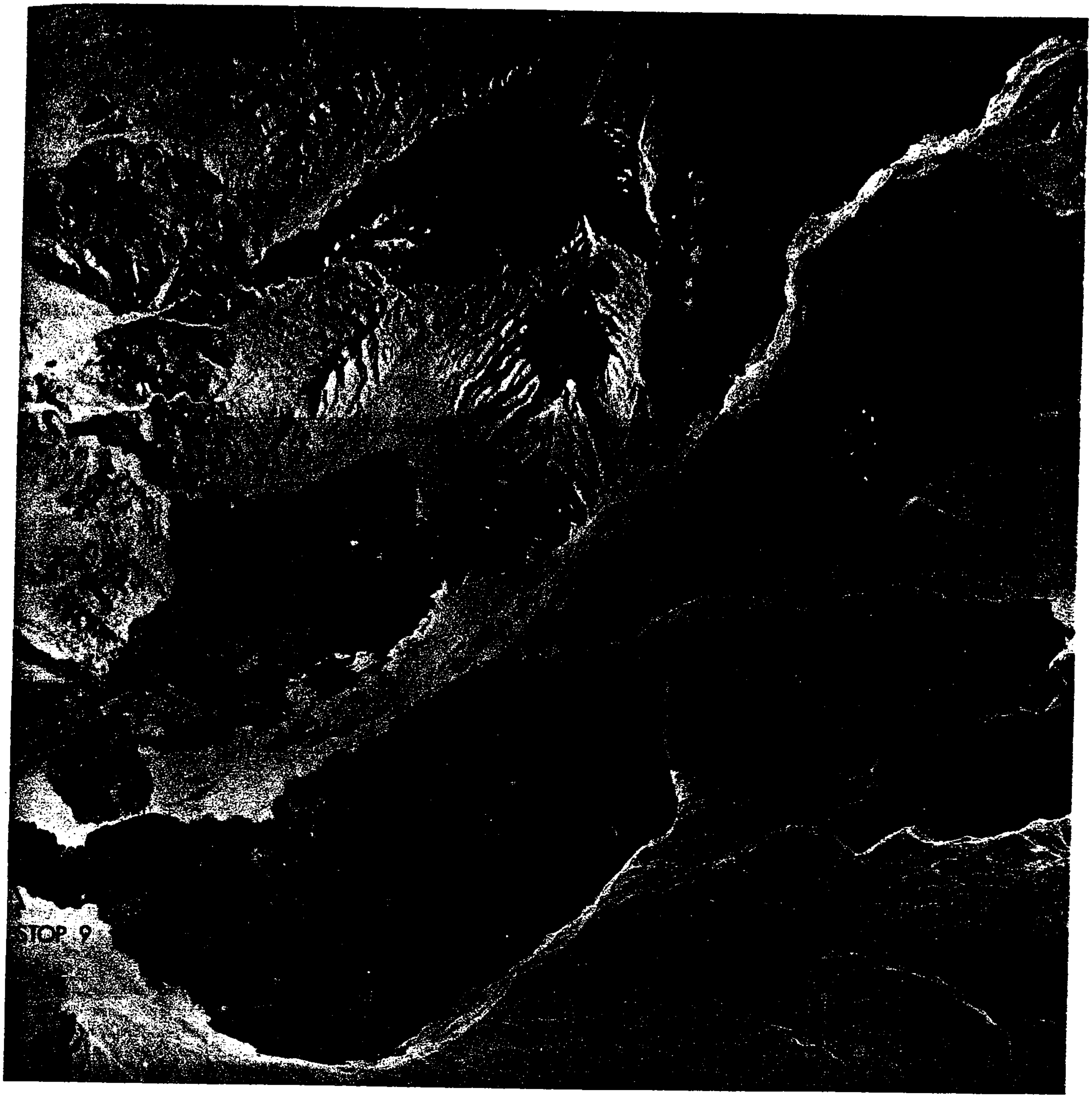

Figure 19. Airphoto of the Cima volcanic field (from Katz and Boettcher, 1980).

of ongoing studies by A. L. Boettcher of UCLA. Much of the following discussion is based on Katz and Boettcher (1980).

The flow we are standing on is the youngest in the field, and its cinder cone is probably the youngest volcano in the continental United States outside of the Cascades. Carbon-14 (from charcoal in the base of the flow) and glass-hydration methods yield a semi-concordant age of about 390 years (A. L. Boettcher, personal communication, 1981). The flow issued from a breach in the northwest wall of the cone. Note the lack of vegetation, soil, and weathering products on the flow surface, compared to surrounding flows. It is also instructive to note the difficulty in separating flows from different cones, once they have flowed together.

Most flows in the field are alkali basalts or basanites, and contain 2-8% nepheline in the norm. Na_2O+K_2O averages 4-6 wt %. Basalts have megacrysts (phenocrysts and xenocrysts) of plagioclase and clinopyroxene (with or without olivine and oxides) in a matrix of plagioclase, clinopyroxene, oxides, and apatite.

Nearly all the flows contain ultramafic and mafic (gabbroic) xenoliths with occasional partially fused granitic xenoliths. The most fertile cones are in the southeastern part of the field, about 10 km east of here. Ultramafic xenoliths are spinel lherzolite (clinoproxene, orthopyroxene, olivine, and aluminous spinel, with traces of pargasite and plagioclase), with xenomorphic granular textures. Gabbroic inclusions have igneous (commonly cumulate) textures and consist of plagioclase, clinopyroxene, orthopyroxene, olivine, and oxides. Mineralogical geothermometers and geobarometers for the gabbros yield temperatures of about 1000°C and pressures up to 8 kb. These temperatures are considerably hotter than normal geothermal gradients, and approach the anhydrous solidus of basalt (Ernst, 1976, p. 151-153). The calculated pressures indicate derivation from near the base of the crust. Katz and Boettcher (1980) suggest that the high temperatures can be explained as relic igneous temperatures preserved by the low activity of water.

The Cima field is lacking in amphibolites, granulites, and other rocks typical of the lower crust. This suggests that the lower crust in this portion of the Mojave Desert is dominantly gabbroic.

A remarkable feature of the Cima area is Cima Dome, a smooth, symmetrical, alluvium-fringed dome cut in Mesozoic granitoids. Cima Dome lies northeast of here, on the far side of the volcanic field (Figure 16). Sharp (1957) considered possible origins for the dome. He concluded that it most likely formed by gentle warping of the crust. Main evidence for this theory consists of gently tilted remnants of Quaternary basalt that fringe the dome.

* * * * * * * *

Return to vehicles and continue south on Kelbaker Road. As we make the long descent to Kelso, the Providence Mountains are ahead at 9 to 12 o'clock. The Providence Mountains comprise Paleozoic sediments that rest on Precambrian metamorphic rocks. The core of the range is intruded by a large plug of Tertiary rhyolite, which is presumably related to the extensive ash-flow fields northeast of the range (McCurry, 1980). From the Kelso area, the range presents a magnificent series of west-facing, nearly vertical cliffs cut in Mississippian limestone. The mountains probably gained their name from early travelers who appreciated the many springs found there.

Kelso is a stop on the Union Pacific Railroad. From Kelso, the rails head west to Afton Canyon across the sandy wastes of the Devil's Playground and the sink of the Mojave River. Note the quaint railroad building in Kelso.

<u>Note mileage in Kelso when crossing the railroad tracks. In 3.6 miles, bear right at the "Y" intersection with Vulcan Mine Road.</u>

<u>At 6.3 miles from Kelso, turn right on the dirt road. Drive approximately 2.8 miles and park.</u>

STOP #10 - KELSO DUNES

Objectives

1. Observe and climb the Kelso Dunes.
2. Appreciate the scenery of the Mojave Desert.

The Kelso Dunes (Fig. 20) are one of the scenic wonders of the Mojave Desert. Their geology and geomorphology are described in an accompanying article by Sharp (this volume). We will spend 2-3 hours at this stop.

<u>Return to the Kelbaker Road and head south (right).</u>

From Kelso Dunes, our route takes us south over Granite Pass, which separates the Providence Mountains from the Granite Mountains. From Granite Pass we make a long drop (1050 m) into the Bristol-Danby trough, which is the current expression of the Barstow-Bristol trough.

The mountains at 12 to 3 o'clock are the Granite Mountains (beware -there are several "Granite Mountains" in the California desert), a large mass of Mesozoic and older (?) granitic and metamorphic rocks. The Providence Mountains span the horizon at 7 to 12 o'clock. Rugged Foshay Pass (elev. 1586 m) separates the northern part of the range, which contains the Paleozoic section, from the southern half, which consists of Mesozoic granitoids.

About 6.5 miles from the turnoff to Kelso Dunes we cross Granite Pass (elev. 1227 m). At this point Kelbaker Road bends to the right to head for its junction with Interstate 40 at the north end of the Marble Mountains.

<u>Continue south on Kelbaker Road, crossing I-40.</u>

South of I-40, Kelbaker Road travels between the Marble Mountains (on the left) and the southeastern tip of the Bristol Mountains (on the right). The Marbles comprise Tertiary volcanic rocks resting on Mesozoic and Precambrian granitoids; Cambrian sediments from the southern tip of the range, south of old highway 66, have yielded trilobites. The southeastern Bristol Mountains comprise pre-Tertiary igneous and metamorphic rocks. See the accompanying paper by Miller and others (this volume) for more information on the geology of this region.

<u>Turn right on the Old National Trails Highway (old U.S. Highway 66), toward Amboy. Bristol Lake is the playa on the left.</u>

Most of the railroad sidings between Amboy and the Colorado River were named in alphabetical order by the railroad in the late 1800s. These stations are Amboy, Bristol, Cadiz, Danby, Edson (now Essex), Fenner, Goffs, Homer, Ibis, and Java.

<u>Entering Amboy (elev. 195 m). Bear right after crossing the railroad tracks. Drive approximately 2 miles and park by the side of the road.</u>

Figure 20. Kelso Dunes.

STOP #11 - AMBOY CRATER AND BRISTOL-DANBY TROUGH

Objectives

1) Discuss features of the Bristol-Danby trough and the southeastern Mojave Desert.

2) Study lava-flow features at Amboy Crater.

We are now standing on the floor of the Bristol-Danby trough. This trough is the southeastward extension of the Miocene-to-Recent Barstow-Bristol trough, which has been the locus of volcanism, downwarping (or downfaulting), sedimentation, and ore deposition since at least the early Miocene. The Bristol-Danby trough extends from Ludlow (40 km westnorthwest of here) to the Colorado River (140 km to the southwest).

The floor of the trough contains three large playas. From northwest to southeast these are Bristol, Cadiz, and Danby Lakes. Cadiz is the lowest, at about 180 m; Bristol Lake lies at 189 m and Danby Lake lies at 186 m. All of the lakes are or were mined for salts, which crystallize on the walls of trenches dug in the lake surfaces. Bristol Lake brines have an unusual composition; they are quite rich in calcium chloride and lithium.

Sediments in the lakes extend at least a few hundred meters below sea level (Bassett and others, 1959). This proves that the trough is, at least in part, tectonic in origin, because stream erosion (perhaps by the ancestral Mojave River - see paper by Weldon, this volume) could not have cut below local base level, which was the Colorado River. However, the tectonic process responsible for downwarping is poorly understood. The playas do not lie in a large graben that parallels the trough because they are separated by ridges of basement. Rather, the playas appear to lie in a series of en echelon grabens or troughs that trend northwest. It is possible that they are pull-apart grabens produced by steps in northwest-trending, right-lateral faults (Glazner, 1981). Evidence for bounding faults is slim, but some of the mountain ranges bordering the playas, such as the Calumet and Sheep Hole Mountains, are linear and may be fault controlled.

Amboy Crater is the southeasternmost in a string of Quaternary alkali-basalt volcanoes that dot the axis of the Barstow-Bristol trough. The cinder cone rises about 75 m above the surrounding lava flows, and is about 450 m in basal diameter. Lavas extruded from the base of the cone cover a circular area of about 50 square km. The flow was extruded onto the surface of Bristol Lake, dividing it in two. Lake sediments stand about one meter higher under the lava than where uncovered. Bassett and Kupfer (1964) suggest that this disparity results from lowering of the playa surface by wind erosion (deflation).

The surface of the flow is covered with a thin layer of sand except leeward (east) of the cone. Interestingly, this patch of barren lava roughly coincides with the limit of volcanic bombs found on the surface, as mapped by Parker (1963). However, Parker believes that bombs cover a large area of the surface, but are covered by later flows.

An interesting feature of Amboy Crater is the low profile of the cinder cone. Such low "ash cones" are often produced by shallow hydro-magmatic eruptions when basaltic lava encounters wet sediments or a body of water (MacDonald, 1972, p. 191). Amboy has a much lower and broader profile than other cinder cones in the Mojave, such as Pisgah, Dish Hill, or those in the Cima area. Perhaps there was water in Bristol Lake during part of Amboy's explosive phase.

Parker (1963) suggested the following sequence of events in the formation of the crater (fig. 21):

1) early explosive eruptions which formed the main cone,

2) eruption and deposition of an agglutinated aggregate of basaltic blocks on the rim and west flank,

3) formation of the outermost inner conelet by a mild explosive phase,

4) breaching of the western wall of the cones by a sideways-directed explosion or by a flow which now occupies the breach,

5) formation of another inner conelet,

6) possible late quiet eruptions of fluid basalt.

Like other Quaternary basalts of the Mojave Desert, Amboy is built of alkali olivine basalts with phenocrysts of olivine and plagioclase in a fine-grained matrix of plagioclase, clinopyroxene, and magnetite. It is undated, although Parker (1963) lists evidence that suggests that it postdates the last glaciation, and is thus younger than 6000 years.

As you examine the flow surface, note the various features displayed, arches, collapsed depressions, blisters, pahoehoe, collapsed lava tubes, breached pressure ridges, and the like. Consider how complicated simple flows like this can appear when their irregular surfaces are covered by later units and then exposed for the geologist to ponder.

* * * * * * * *

<u>Return to vehicles and continue west on highway 66.</u>

About 6 miles past Amboy Crater, look left to see the western portion of Bristol Lake. The low cinder cone just south (left) of the road is considerably more eroded than Amboy and is breached on the north side. This cone, known as hill 1068, is made of several layers of xenolith-bearing basaltic ash.

<u>Note mileage at hill 1068</u>.

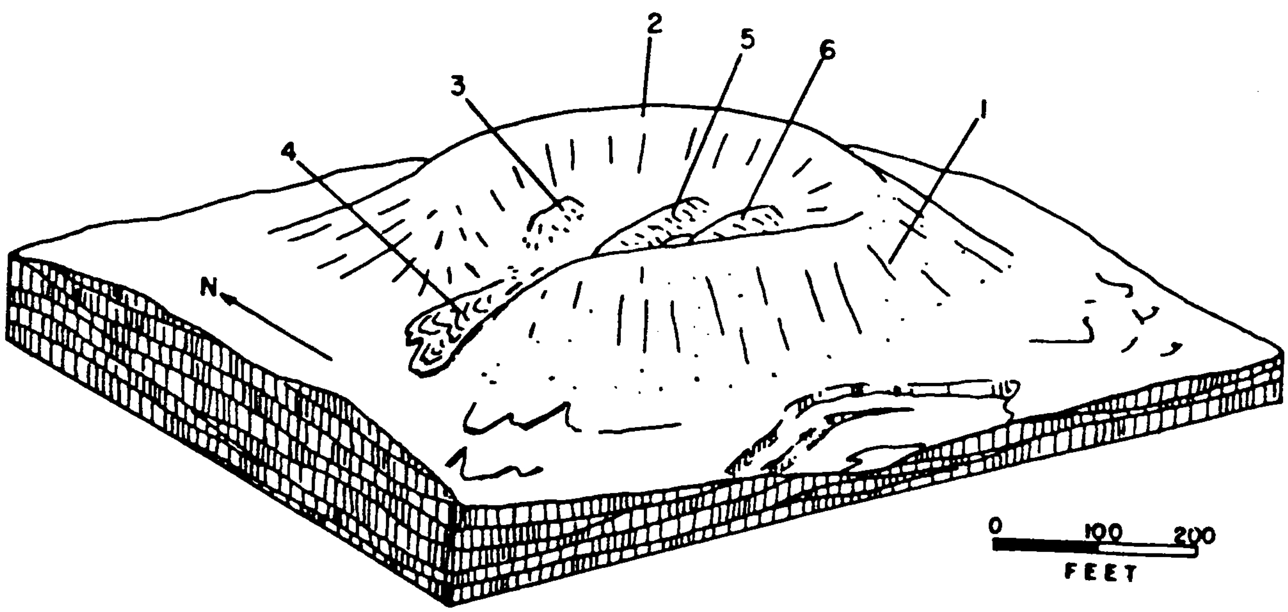

Figure 21. Sequence of events in the formation of Amboy Crater (from Parker, 1963). Numbers refer to phases described in text.

After passing hill 1068, look at 9 o'clock to see a large cinder cone in the Bullion Mountains and a small basalt hill in the valley floor.

Turn right 1.1 miles past hill 1068, on the dirt road that leads under the railroad tracks at the diagonal yellow sign. Park near the railroad tracks.

STOP #12 - DISH HILL

Objectives

1) To view an older (Pliocene) alkali-basalt cinder cone.

2) To observe and collect mantle xenoliths.

Dish Hill is one of the most famous of the late Cenozic cinder cones of the Southwest, and justly so, because of the wealth and compositional range of xenoliths found in its ejecta (see Boettcher, this volume). It is almost true that one cannot walk around the cone without stubbing a toe on a xenolith. Most xenoliths are olivine-rich, with olivine in various states of weathering, from fresh dull-green grains to weathered, orange pits. Bright-green chrome diopside is a common associate. The following description of Dish Hill and its companions is a short paper by Art Boettcher taken verbatim from Glazner and Bilodeau (1980, p. 49-50).

"About 10 million years ago, flows and pyroclastic cones of alkali basalt composition began to erupt throughout the eastern Mojave Desert and elsewhere in the southwestern United States and Mexico. This activity continues through the Holocene in the Cima area near Baker. Pisgah and Dish Hill are easily accessible examples of previous activity.

"Dish Hill is actually two cinder cones (Dish Hill proper and Siberia Crater). Two other cones occur nearby, Hill 1068 (S.22, T.6N., R.10E., weathered and uninteresting, just south of Highway 66), and Hill 1933 (S.11, T.6N., R.10E., similar to Dish Hill, about a 3 km walk northeast). These cones are probably part of the Deadmen Lake volcanic field farther south in the Twentynine Palms Marine Corps Firing Range and are also similar to those of the Cima volcanic field just east of Baker.

"All of these cones and flows are of basaltic compositions, specifically alkali basalts (0-5% normative nepheline) and basanites (more than 5% normative nepheline). Both generally have normative olivine greater than 10%. Rocks of these compositions are of interest because they commonly contain xenoliths and megacrysts, which are fragments of mantle and crust captured by these volatile-rich magmas as they are forcefully erupted at velocities perhaps greater than the speed of sound from depths as great as 70-80 km.

"Dish Hill is older and more dissected than most of the cones that you will see north of the road between Baker and Kelso. C.W. Naeser (U.S.G.S.) obtained an age of 2.1 $\pm$ 0.2 million years by fission track analyses of apatite in a granitic xenolith. M.A. Lanphere (U.S.G.S.) obtained 1.9 $\pm$ 0.2 million years with the K-Ar method on an amphibole megacryst.

"Dish Hill (including Siberia) and Hill 1933 are particularly pregnant with mantle xenoliths (e.g., spinel lherzolites, commonly amphibole-bearing) and megacrysts (amphibole, clinopyroxene, olivine, chrome diopside, orthopyroxene, spinel, and others). Amphibole-bearing xenoliths are most abundant in the saddle between the craters."

* * * * * * * *

Return to highway 66 and turn right (west).

After Dish Hill the road makes a broad turn to the right followed by a broad turn to the left, and traverses hills of Tertiary (probably lower Miocene) basalts, andesites, and silicic tuffs. The road is climbing out of the Bristol-Danby trough. This is the heart of the great northwest-trending belt of volcanic rocks that defines the Barstow-Bristol trough. Desert pavement is well developed on both sides of the road.

After the second broad turn, the road passes just south of low, sand-covered basalt hills. The railroad station here is known as Ash Hill and was named not for pyroclastic rocks but for Ben Ash, a railroad employee who died of thirst while surveying here in the late 1800s.

Past Ash Hill we descend toward Ludlow. Dead ahead are the southeastern Cady Mountains, otherwise known as the Sleeping Beauty area. They were mapped by one of us at 1:10,000 for dissertation studies (Glazner, 1981). The playa at 1:30 o'clock is Broadwell Lake. Geologic relations suggest that this basin is a pull-apart related to termination of the right-lateral Ludlow fault, which heads north through Ludlow and then loses expression in the eastern Cady Mountains.

Entering Ludlow. Ludlow got its name in the 1870s from William B. Ludlow, master car-repairer for the railroad. Turn right in Ludlow and cross under the freeway. Turn left onto I-40 west toward Barstow. A more leisurely route is to stay off the freeway, turn left just past the onramp, and travel down the frontage road.

The Stedman gold district is in the northern Bullion Mountains, about 10 km south of Ludlow. The Bagdad-Chase mine, which was the main producer, operated from 1903 to 1952 with a total production of about 400,000 tons of ore and a recovery of more than 7 million dollars in gold, silver, and copper (Wright and others, 1953, p.71-72).

The Bagdad-Chase area was mapped by Polovina (1980), who found that the ore occurs in a hydrothermal breccia "sill" between two rhyodacite flow units. He interpreted the contact between the flows as an unconformity; we suggest that it may be a low-angle fault, with analogy to other areas in the Mojave Desert (Dokka and Glazner, this volume).

The Sleeping Beauty area, which makes up the mountains on the right, contains one of the thickest sections of lower Miocene volcanic rocks in the Mojave. A generalized section is given in figure 22. From the highway, the long, low, southeastern spur comprises a basalt-rhyodacite unit capped by dacites; the high mountain on the left is the lower andesite stratovolcano unit. Near the southeasternmost tip of the area, several deep barite prospect pits are visible. The dark red dikes visible in the low ridge to the left of the pits are siliceous breccia dikes that are surrounded by potassium-metasomatized aureoles. About one third of the outcrop area in the Sleeping Beauty area has been affected by strong potassium metasomatism (Glazner, 1979, 1980, and in preparation).

Eight miles from Ludlow, Sunshine Crater, another Quaternary cinder cone, is visible 10 km distant at 9 o'clock. Associated flows are cut by the northwest-trending, right-lateral Pisgah fault. The fault produced a large scarp in the basalt, which is easily visible from the highway. (It is possible that the basalt poured over a preexisting scarp, and only looks faulted; Dibblee (1966) mapped the fault as cutting the basalt.) Sunshine Crater is poor in xenoliths.

The playa at 9 o'clock is Lavic Lake. This basin is another example of a possible pull-apart basin related to the right-lateral fault system; the squared-off outline, bounding faults, and presence of several cinder cones all suggest a pull-apart origin.

The large cinder cone near the road to our left is Pisgah Crater. The crater is currently being dismantled for cinders. It was named after Mt. Pisgah in Palestine, from which Moses saw the promised land. It is unclear what promised land

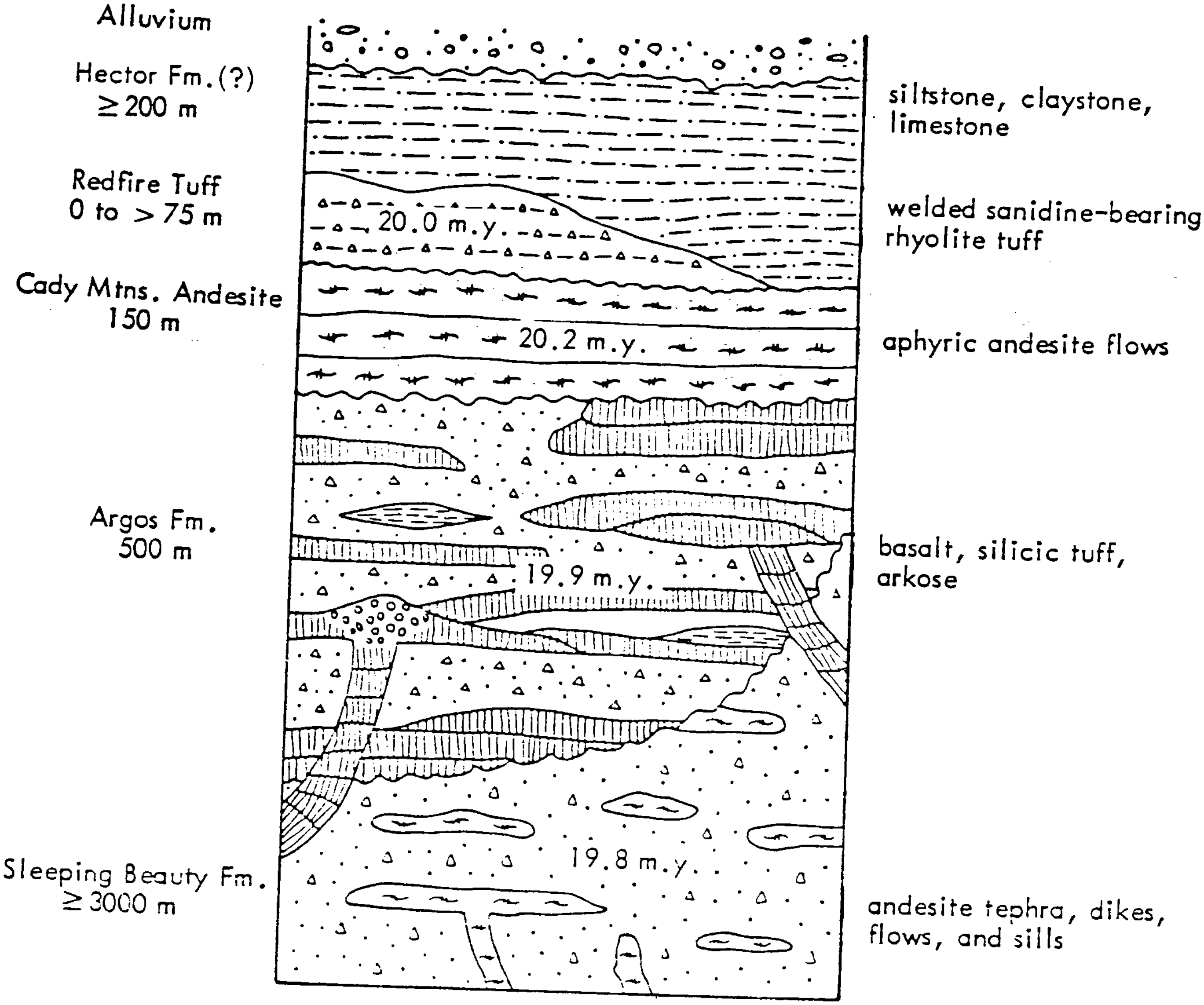

Figure 22. Schematic stratigraphic section of the Sleeping Beauty area, Cady Mountains, California (from Glazner, 1980).

can be seen from Pisgah Crater. Pisgah is similar to Amboy in many respects, although its cone is steeper and its lava flows are more hummocky because they flowed onto a hilly surface.

If you plan to make the optional stop at the Pisgah fault, exit at Hector Road (unless you are already on the frontage road). If not, continue on I-40 past Hector Road.

Note mileage at Hector Road. If stopping, proceed 1.7 miles west and park on the side of the road.

OPTIONAL STOP - PISGAH FAULT

Objective

1. View the Pisgah fault where it cuts flows of Pisgah Crater.

Recent activity on the Pisgah fault is evidenced by the linear ridge cutting the basalt flow on both sides of the highway. Although Pisgah Crater has not been dated, it is probably similar in age to Amboy, and thus Holocene or latest Pleistocene. Thus, this ridge, and the much larger one in the Sunshine Crater flow to the southeast, attest to significant movement on the fault in the last 10,000-20,000 years. Dates on the faulted basalt flows will help constrain movements.

The low sand-covered hills to our southeast are Quaternary sediments that were apparently uplifted by the fault. Pisgah Crater lies on the far side of these hills. A long, narrow arm of the Pisgah flows extends northwestward from the crater, along the fault, to Troy Lake. Perhaps this long arm followed a fault-controlled trough at the base of the sandy hills.

If you are on the frontage road, continue west and rejoin the freeway ahead in Newberry Springs.

* * * * * * * *

After leaving the Pisgah flows, the road traverses the floor of Troy Lake, a remnant of the southern lobe of Lake Manix. The mountains at 9 o'clock are the Rodman Mountains, a mass of Mesozoic granitoids flanked on the north (near) side by early Miocene sedimentary and volcanic rocks. A cinder cone perched high in the Rodman Mountains sent a long tongue of lava down Kane Springs Wash to the northwest. The tip of the cinder cone is visible at 9 o'clock as we cross Troy Lake, and the snout of the lava flow is visible far below, in the wash at 10 o'clock.

About 3 miles beyond the edge of the dry lake we reach the Calico fault, stop #5 of yesterday.

* * * * * * * *

END OF ROAD LOG

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ASPECTS OF EARLY MIOCENE EXTENSION OF THE CENTRAL MOJAVE DESERT

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INTRODUCTION AND GENERAL GEOLOGY

The Mojave Desert block of southern California is a wedge-shaped structural province bounded on the north by the Garlock fault, on the southwest by the San Andreas fault system, and to the east by a NNW-trending line defined by a group of present-day geophysical anomalies (fig. 1) (Dokka, 1980). The region has experienced a complex tectonic history during the Cenozoic, marked by deformations that were similar in style and kinematics to deformations that affected surrounding provinces. In this paper we briefly describe the geometric, kinematic, and timing aspects of the structural association produced by a short interval of crustal extension during the early Miocene. It is also our goal in this brief synthesis to identify problem areas where additional studies are needed.

The greater Mojave region is composed of the following lithologic groups: 1) Precambrian crystalline complex (metamorphic and igneous rocks) ranging in age from 1.87 b.y. to 1.2. b.y. (from synthesis, Burchfiel and Davis, 1980); 2) Upper Precambrian-Paleozoic sequence of miogeoclinal and cratonal rocks; 3) Mesozoic back-arc and intra-arc shallow marine and continental sequences; 4) Magmatic arc terrane (mostly granitoid rocks along with minor amounts of related (?) volcanic cover); 5) lower Miocene calc-alkaline and bimodal volcanic sequences and continental sedimentary rocks; 6) Quaternary alkalic volcanic rocks; and 7) Quaternary alluvium. In terms of surface exposure, the province is dominated by Mesozoic granitic rocks which are part of the arc-related batholithic belt that rings the Circum-Pacific region (Hamilton, 1969). Three intervals of plutonism have been recognized in the Mojave Desert based on geochronologic studies: 1) Early Triassic alkalic monzonites (C. Miller, 1977; E. Miller, 1977); 2) Late Triassic- Early Jurassic (210-180 m.y.) quartz monzonite-granodiorite (Kistler, 1974; Kistler and Peterman, 1978); and 3) Cretaceous granite-quartz monzonite (120-90 m.y.) (Kistler, 1974; Kistler and Peterman, 1978). Roof pendants in the batholithic complex consist of scraps of metamorphosed carbonate and clastic sedimentary rocks, volcanic rocks, and crystalline rocks. These rocks are probably equivalent to thick sequences of Upper Precambrian-Paleozoic miogeoclinal and cratonal rocks, and Mesozoic arc-related volcanics and associated sedimentary rocks

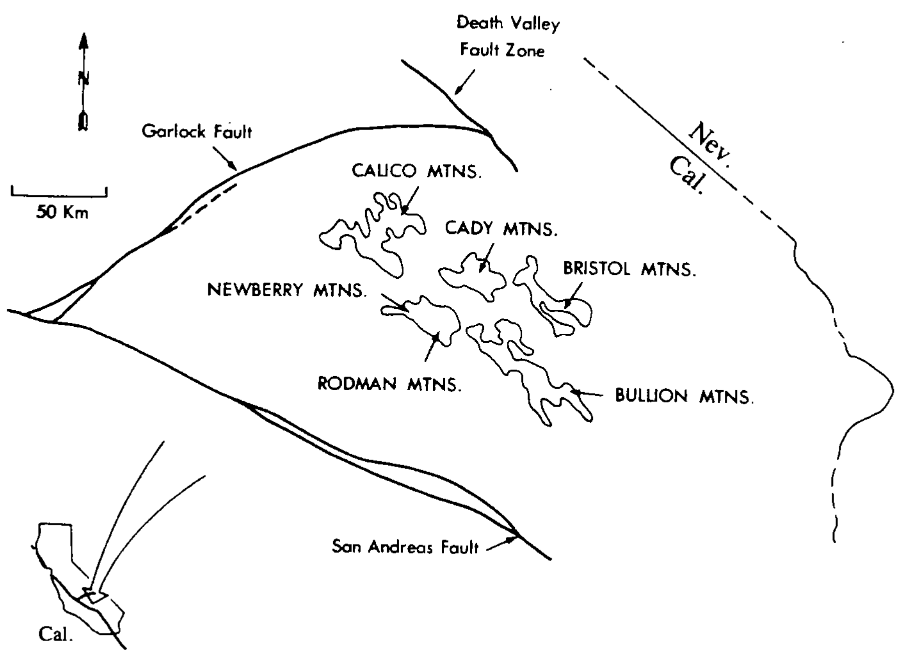

Figure 1. Index map to localities in the Mojave Desert, California

that lie to the south in the San Bernardino Mountains and in the eastern Mojave in the Providence Mountains (Stewart and Poole, 1975; Tyler, 1979; Burchfiel and Davis, 1980). Although these data strongly indicate that Late Precambrian-Paleozoic paleogeographic trends are continuous across the Mojave Desert, the Precambrian crystalline basement upon which these sedimentary rocks rest is not present. This is truly one of the more puzzling aspects of the geology of the Mojave Desert and deserves more attention.

Information regarding the early Cenozoic history of the Mojave Desert is extremely limited because of the lack of lower Cenozoic rocks. Hewett (1954) was first to recognize this and concluded that the Mojave Desert stood high during the early Cenozoic and provided sediments to surrounding provinces. Approximately 3 km of erosion may have occurred in the central Mojave area during the early Cenozoic. This rough estimate is based on the absence of thick, complete sections of sedimentary rocks that were probably deposited across the entire Mojave region during the Late Precambrian, Paleozoic, and Mesozoic. (Stewart and Poole, 1975; Tyler, 1979; E. Miller, 1977; Burchfiel and Davis, 1980).

Volcanism commenced in the central Mojave Desert at the beginning of the Miocene (ca. 23 m.y.) (Armstrong and Higgins, 1973). These initial deposits are sometimes underlain by a thin (0-5m) basal arkosic sandstone-gravel that probably represents short-traveled weathering products of the previously exposed granitic batholith. Glazner (1981) suggests a regional lower Miocene volcanic stratigraphy that consists of the following volcanic succession (oldest to youngest): 1) andesites; 2) high-Ti basalts and silicic tuffs; and 3) dacites and sediments. The thickness of the sequences varies locally (e.g. 3.7 km, Cady Mountains; Glazner, 1981). The lower andesite unit records a period of calc-alkaline, Cascade-type volcanism and is similar to subduction-related andesites found in other parts of the western U.S.A. The bimodal suite corresponds to post-subduction assemblages that are spatially related to extensional complexes. The upper dacite unit is unexplained. Areally the bimodal suite is the most extensive, whereas the lower andesite is present in only a few areas (Glazner, 1981).

Lacustrine and basin-edge continental upper lower Miocene Hector Formaton and the Barstow Formation and their time equivalents overlie with angular unconformity the lowermost Miocene section. Additionally, local exposures of Miocene—Pliocene intermediate volcanic rocks and Quaternary alkali basalts dot the landscape. The remainder of the central Mojave Desert is blanketed by Quaternary sediments of many different ages and depositional environments.

EARLY MIOCENE EXTENSION

General Aspects

The central Mojave Desert underwent a profound change in physiography, tectonics, magmatism, and sedimentation patterns during the early Miocene. Crustal extension is considered to have produced a structural association consisting of the following elements: 1) high-angle normal faults; 2) low-angle faults; 3) rotated upper plate middle Tertiary strata and older crystalline rocks; 4) regional long wavelength folds oriented parallel to the direction of extension (?), and 5) strike-slip tear faults bounding structures. The area distributed by extension was in excess of 5,000 km^2 and affected the following ranges: Newberry Mountains, Rodman Mountains, Bristol Mountains, Bullion Mountains, Cady Mountains, Calico Mountians, Kramer Hills, and Hinkley Hills (Fig. 2).

Faults

Faults related to early Miocene extension in the central Mojave can be divided into three groups based on their position within the terrane. The first group are those faults that occur along the perimeter of the extended terrane and served to accommodate the differences in strain between the extending complex and adjacent areas. A second group of shears are those that occur within the terrane (high-and low-angle faults). The third group includes high-angle normal faults that cut shears of the second group.

Terrane Boundary:

The northern margin of this terrane is unstudied. Geometric aspects of part of the southern margin of the terrane can be seen or inferred in the Newberry and Rodman Mountains (Dokka, 1980). East of the Rodman Mountains in the Bullion Mountains, this bounding structure becomes intra-terrane, and separates two tilting domains with opposite senses of rotation (Fig. 2). This fault has been named the Kane Springs fault, for the locality in the Newberry Mountains where it was first discovered (Fig. 3). Geometry and displacement along the fault appear to vary systematically as a function of the local strike of the boundary and the direction of extension of the terrane. For example, in the Rodman Mountains, the Kane Springs fault is a strike-slip fault because its strike is nearly parallel with the inferred direction of extension. Conversely, in Stoddard Valley, where the fault becomes more northwesterly in strike, the fault displays mainly down-dip displacement. Along Kane Springs Wash, movement has been oblique (mainly strike-slip with a component of normal dip-slip).

The shape of the Kane Springs fault may be inferred by integrating information regarding the shape of its trace, type of movement, and the nature of the associated terrane-edge deformation. The arcuate surface trace of this fault (Fig. 4) suggests that it has a scoop-shaped form that flattens with depth. This geometry may also be inferred from the development of reverse drag folding (edge foundering) along the flank of the allochthonous terrane (Fig. 5a). Reverse drag is

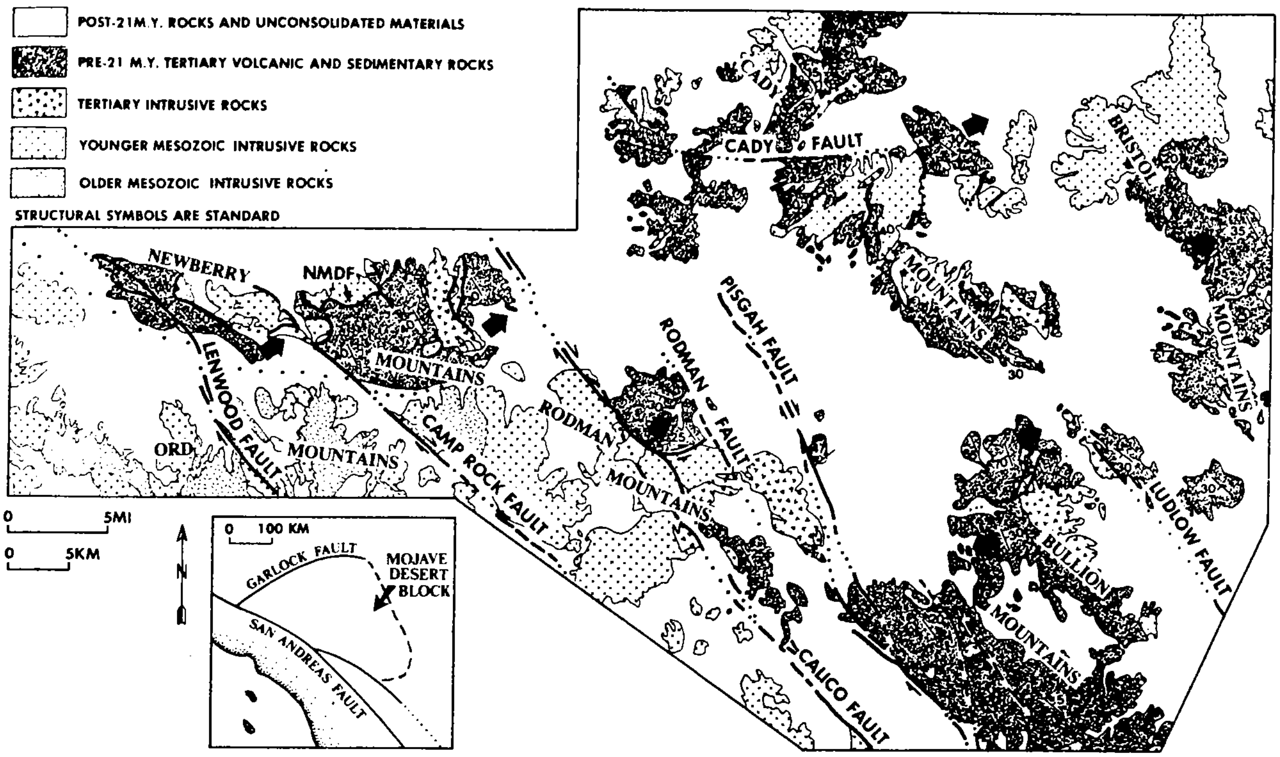

Figure 2. Generalized geologic map of the central Mojave Desert. Large black arrows indicate inferred direction of transport of areas affected by early Miocene detachment faulting.

a geometric consequence of normal slip movement along a curviplanar fault (see Hamblin, 1965).

Movement along the Kane Springs fault produced a narrow basin(s) that accumulated disorganized packages of coarse breccia and conglomerate (Fig. 5b). Most of the sedimentary packages are monolithologic and were derived from exposed Mesozoic crystalline terranes that lay adjacent to the extending terrane. Less extensive assemblages of terrane-derived sediments (mainly lower Miocene volcanic clasts) locally interfinger with the external deposits. The Kane Springs fault is a growth fault in the sense that sediments filled structurally-controlled basins adjacent to the fault. Older sedimentary rocks are more steeply tilted than younger strata. These relations are depicted on figures 5a and 5b.

Intra-Terrane Normal Faults:

The second group of faults is responsible for much of the tilting and extension of lower Miocene and older rocks throughout the central Mojave Desert (Fig. 2). These shears can be grouped into 2 classes based on their geometry: 1) two families of north-northwest striking high angle normal faults with opposing dips; and 2) a regional low-angle detachment fault. The normal faults merge with the detachment and appear to be geometrically and kinematically related to it.

The two families of early Miocene normal faults occur in separate fault domains that are coextensive with zones of uniformly tilted packages of lower Miocene volcanic strata (Fig. 6). The areally most extensive group includes faults that are north-northwest striking, east dipping, and high angle. A major effect of this faulting has been the repetition of the basal unconformity that separates the lower Miocene and crystalline basement rocks. Mapping shows that this contact is repeated in the Bristol Mountains (Casey, 1980), central Cady Mountains (Dokka, work in progress), Newberry Mountains (Dokka, ms), and probably in the Kramer Hills (Dibblee, 1967a). A second family of normal faults is located south of Ludlow in the Bullion Mountains (Fig. 2). The proposed extension of the Kane Springs fault separates these two domains which extended simultaneously but in opposite directions. These faults are not well-known because most lie within Twenty-Nine Palms Marine Corps Training Center. However, reconnaissance mapping in unrestricted areas nearby has shown the occurrence of several southwest dipping, down to the south faults (Dokka, unpublished mapping). Strata in this domain are locally steeply tilted (70°) to the northeast (Dibblee, 1967b).

A family of low-angle normal faults crops out in the Newberry Mountains (Dokka, 1980). The most spectacular of these shears is the Newberry Mountains detachment fault (NMDF) which has juxtaposed tilted lower Miocene and upper plate crystalline rocks against lower plate crystalline rocks (Fig. 7). This surface is thought to pass laterally and merge with the Kane Springs fault. The contact was originally interpreted by Dibblee (1971) as the "Great buttress unconformity" of the Tertiary of California. Detailed mapping by Dokka, (1976,

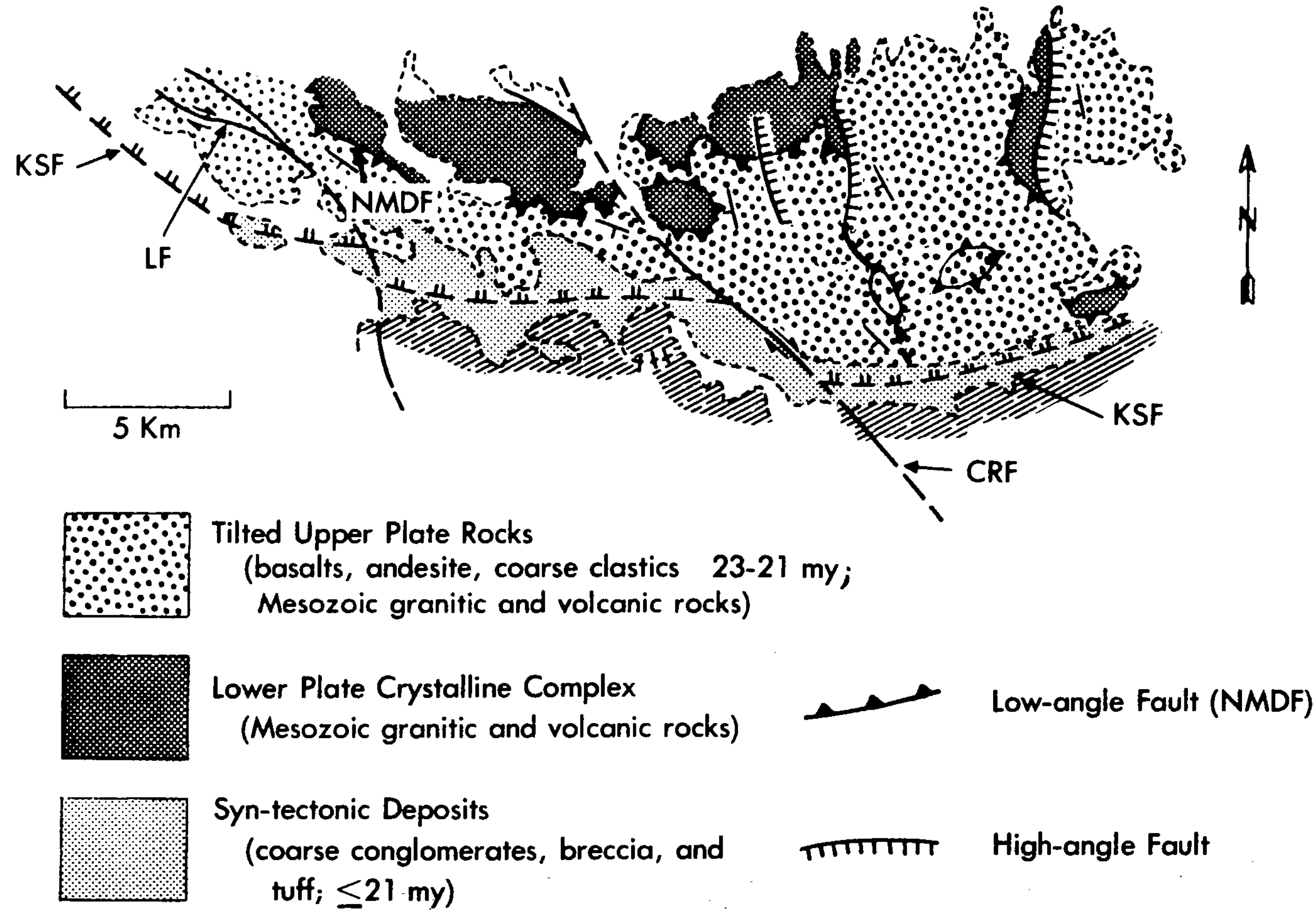

Figure 3. Generalized geologic map of the Newberry Mountains. NMDF = Newberry Mountains Detachment Fault; KSF = Kane Springs Fault; LF = Lenwood Fault; CRF = Camp Rock Fault.

1979, 1980) of all basement-cover contacts in the Newberry Mountains (type area for Dibblee's buttress unconformity) demonstrates the dislocational nature of this surface. Only one short (0.5 km) segment of the basal unconformity between lower Miocene volcanic rocks and crystalline basement can be seen in this range (it lies within a fault sliver). Evidence that this surface is not an unconformity include, for example, sediments and volcanics juxtaposed across the contact do not contain any clasts of the basement (i.e., cataclasized granite-quartz monzonite or Sidewinder Volcanic Series), nor do they onlap the basement as might be expected if this surface was one of deposition. Instead, this boundary is marked by extreme shearing of lower-plate rocks, comminution of materials near the surface, development of fault formed kinematic indicators (slickenside striations, etc.), and the abrupt truncation of bedding and foliations (Fig. 7). Strata above this surface display a consistent angular relation with the local fault-surface, in that the acute angle formed by this intersection is between 40 and 70 degrees. The high-angle normal faults of the upper plate are considered to merge with the NMDF; movement linkage is suggested by the similarity of kinematics (Dokka, 1980). Cross-sections of the Newberry Mountains are shown on figure 8.

Basement shear zones parallel the local orientation of the NMDF and occur as deeply as 100 m below that surface (Fig. 9). These zones (0-10 cm wide) are composed of anastomosing faults that border thinner regions of cataclasized rocks. These basement shears are best observed along the north-central portion of the Newberry Mountains of the Camp Rock Road.

A puzzling aspect of the geometry of the NMDF is that it is not flat but is, rather, a curviplanar surface. The eastern two-thirds of upper plate lower Miocene volcanic rocks and older crystalline rocks in the Newberry Mountains rest in a large synform that is oriented ~N50E. This curved character of the NMDF is depicted on figure 8b.

Lower-plate rocks beneath the NMDF are intensely shattered to depths of 5 to 100 m (and locally, perhaps more). These rocks can be related genetically to the development of the NMDF by observing the following: 1) the zone of cataclasis only occurs adjacent to the NMDF (Fig. 7, 8); 2) the unit contains several generations of shears that are geometrically and kinematically similar to the NMDF; and 3) the time of cataclasis and faulting was coincident.

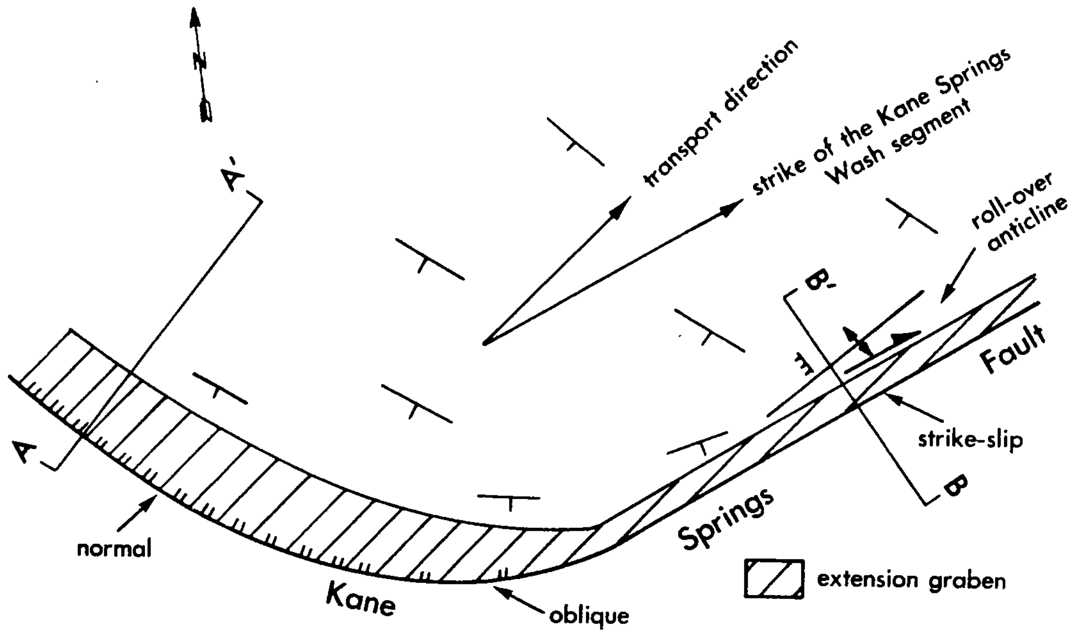

Figure 4. Geometry of the Kane Springs fault (map view). At depth the KSF is inferred to flatten and to become the Newberry Mountains detachment fault. Cross section lines indicate location of figures 5a and 5b.

According to the guidelines presented in Higgins (1971) these rocks are microbreccias and cataclasites, based on their possession of primary cohesion and a lack of fluxion structure. An introduction to the petrography of these rocks was presented in Dokka (1976) and is reiterated below.

The granite microbreccia of this area is composed of porphyroclasts of perthite and quartz set in a more finely ground matrix of the same mineral (Fig. 10) (this is referred to as mortar texture, a term coined by Tornebohm, 1880). Porphyroclast to matrix ratio is about 2.0. Diameter of the porphyroclasts ranges from 0.2 mm to 0.5 m and averages 25 mm. Typically the porphyroclasts are angular polygons and are easily discerned on the outcrop owing to resistant weathering ability relative to the matrix. The rock is cemented by quartz and, to a lesser degree, by hematite. There is little evidence for any neomineralization (i.e., metamorphism), except for the cements and small, local occurrences of chlorite. Quartz grains commonly show undulose extinction, probably due to deformation. Secondary minerals such as opalline silica and calcite are common vein constituents near shears. These minerals occupy joints of variable orientation and are probably due to ground water action.

Precise geometric relations of the microbreccia unit are poorly understood; however, it is probably tabular because it generally conforms to the configuration of the Newberry Mountains detachment fault and associated shears (fig. 7) (Dokka, 1976, 1980). The structure of this rock is dominated by subhorizontal shears, along with fractures oriented at low angles (north-striking sets, dipping 10-30 degrees to the east and west) to the NMDF. Granite-quartz monzonite shows the most pronounced effects of cataclasis, whereas the Sidewinder Volcanic series and hornblende diorite are less-affected, perhaps owing to their greater shear strengths.

Later Normal Faulting

Two areas of the central Mojave Desert have been recognized to contain normal faults that clearly post-date the major phase of major upper plate distension and rotation and associated detachment faulting (Fig. 8a). Orientation and kinematics of these faults suggest that some of these structures may be related to the earlier extension, perhaps as part of a continuum of deformation. In other areas, younger normal faulting cannot be linked to earlier deformations.

In the Newberry Mountains, north-northwest striking high-angle normal faults cut the Newberry Mountains detachment fault and display a consistent down-to-the-east displacement pattern. The trends of slipline indicators along these faults are subparallel to those of the older normal faults. The amount of dip separation along these faults is generally small(<100m), but one shear located in the eastern of the range displays as much as 1000m of displacement. In contrast with the normal faults of the earlier extension, this phase of faulting only produced a minor amount of stratal rotation (10°-15°). Another example of an area affected by later normal faulting is the northern Cady Mountains. Here, S. Miller (1978) has demonstrated that alluvial and lacustrine sediments of the 21-17 m.y. Hector

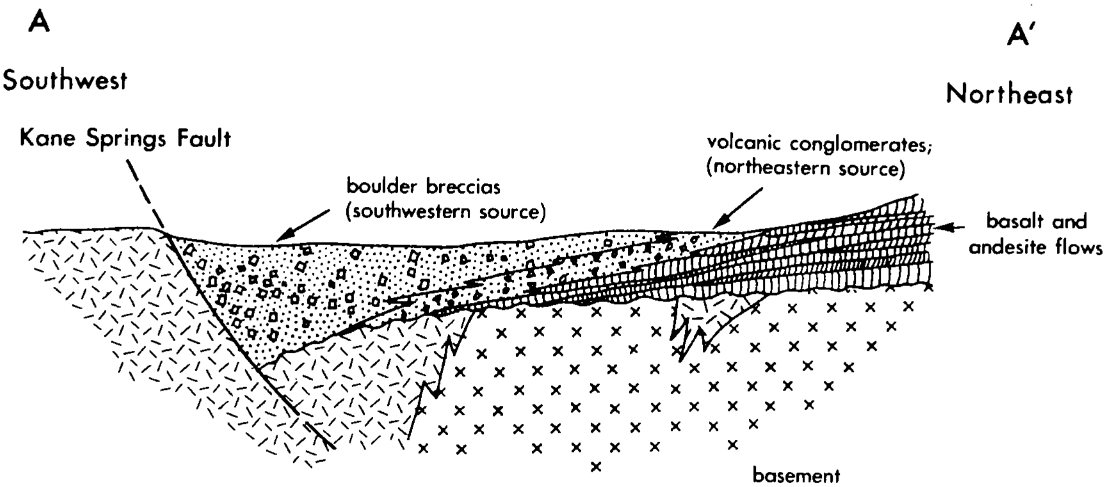

Figure 5a. Paleogeologic cross section across the southwestern margin of the Newberry Mountains (from Dokka, 1980). Rocks deposited adjacent to Kane Springs fault show sedimentary growth. See figure 4 for location.

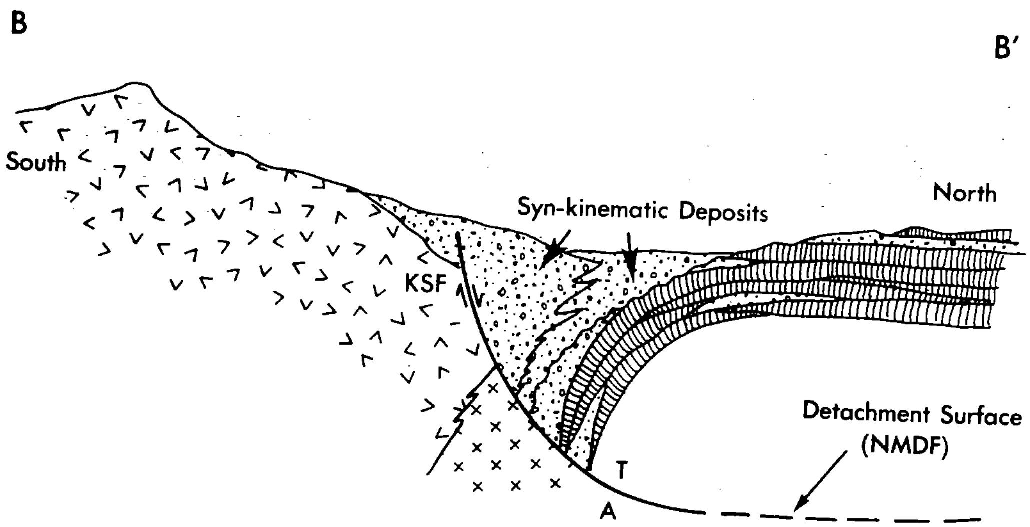

Figure 5b. Tectonic and stratigraphic relations along terrane edge in Kane Springs Wash. View is to the southwest. Synkinematic clastic deposits are composed of two distinct clast populations. Clasts in coarse breccias and conglomerates derived form the south comprise Mesozoic granitic and pendant material. Detritus derived from the north are exclusively composed of lower Miocene volcanic clasts. Folding is the result of reverse drag along the Kane Springs fault. See figure 4 for the locaton of the section.

Formation are cut by north-striking, east-and west-dipping normal faults. Displacement along these faults is small (100m), as is associated tilting(<15°). According to Miller, faulting was not coincident with deposition of Hector sediments.

Kinematics

Timing of regional high-angle normal faulting, detachment faulting, and tilting of lower Miocene strata and older crystalline rocks

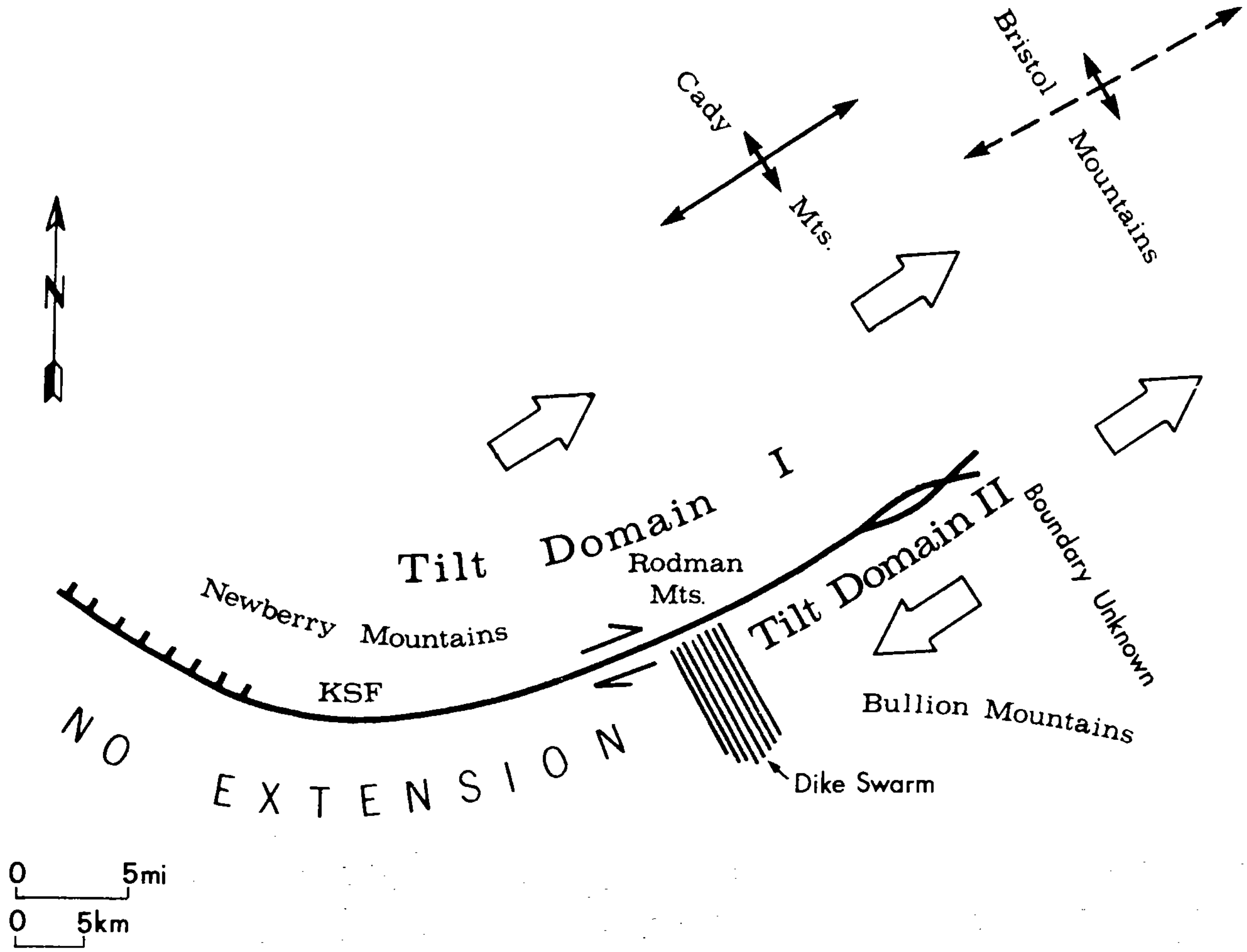

Figure 6. Tilt domains related to early Miocene extension of the central Mojave Desert. Arrows indicate direction of tectonic transport of upper plate rocks relative to stable regions to the southwest. Each domain contains tilted lower Miocene volcanic strata and older rocks. Tilting in domain I is to the southwest, whereas rocks of domain II dip to the northeast.

is bracketed as 23-20 m.y., based on the ages of the youngest strongly tilted upper plate rocks and the relatively flat-lying strata that overlie them. This was determined through analysis of cross-cutting relations from each of the following ranges: 1) Newberry Mountains, 23-21 m.y. (Dokka, 1979, 1980); 2) western Cady Mountains, 23-22 m.y. (Otton and Dokka, ms); 3) southern Cady Mountains, early Miocene, but pre-20 m.y. (Glazner, 1981). Similar relations (pre-18 m.y. southwest tilting and normal faulting) are reported in the Bristol Lake region where structural and age bracketing studies are now in progress (D. Miller and others, 1982, this volume).

Three types of kinematic indicators were used in this analysis and include: 1) striations of slickensided surfaces along faults; 2) sense and direction of rotation of upper-plate strata; and 3) the strike of intra-terrane tear faults. The method of obtaining the direction of stratal rotation of upper plate rocks was determined from the average strike of bedding in several areas that were unaffected by later deformations. The direction of rotation was considered to be the heading opposite the dip azimuth.

Analysis of kinematic indicators produced by the early Miocene deformation suggests that extension was northeast (N.50°E.) southwest (S.50°W.) directed. Movement was apparently bidirectional, however, as indicated by adjacent tilt domains with opposite rotational characters. The "transform" fault between them (the Kane Springs fault near Ludlow) suggest that extension of the two domains was synchronous. Relations along the Kane Springs fault in the Newberry Mountains suggests that southwest-dipping rocks north of the fault were transported to the northeast relative to stable areas to the southwest. If these relations are correctly interpreted, it follows that northeast dipping rocks located south of the proposed extension of the Kane Springs fault must have been transported to the southwest. These relations are depicted on figure 6.

EARLY MIOCENE FOLDING

One of the most enigmatic groups of structures of the central Mojave Desert is a group of long wavelength regional antiforms and synforms that are oriented subparallel to the direction of early Miocene extension (east-northeast). The

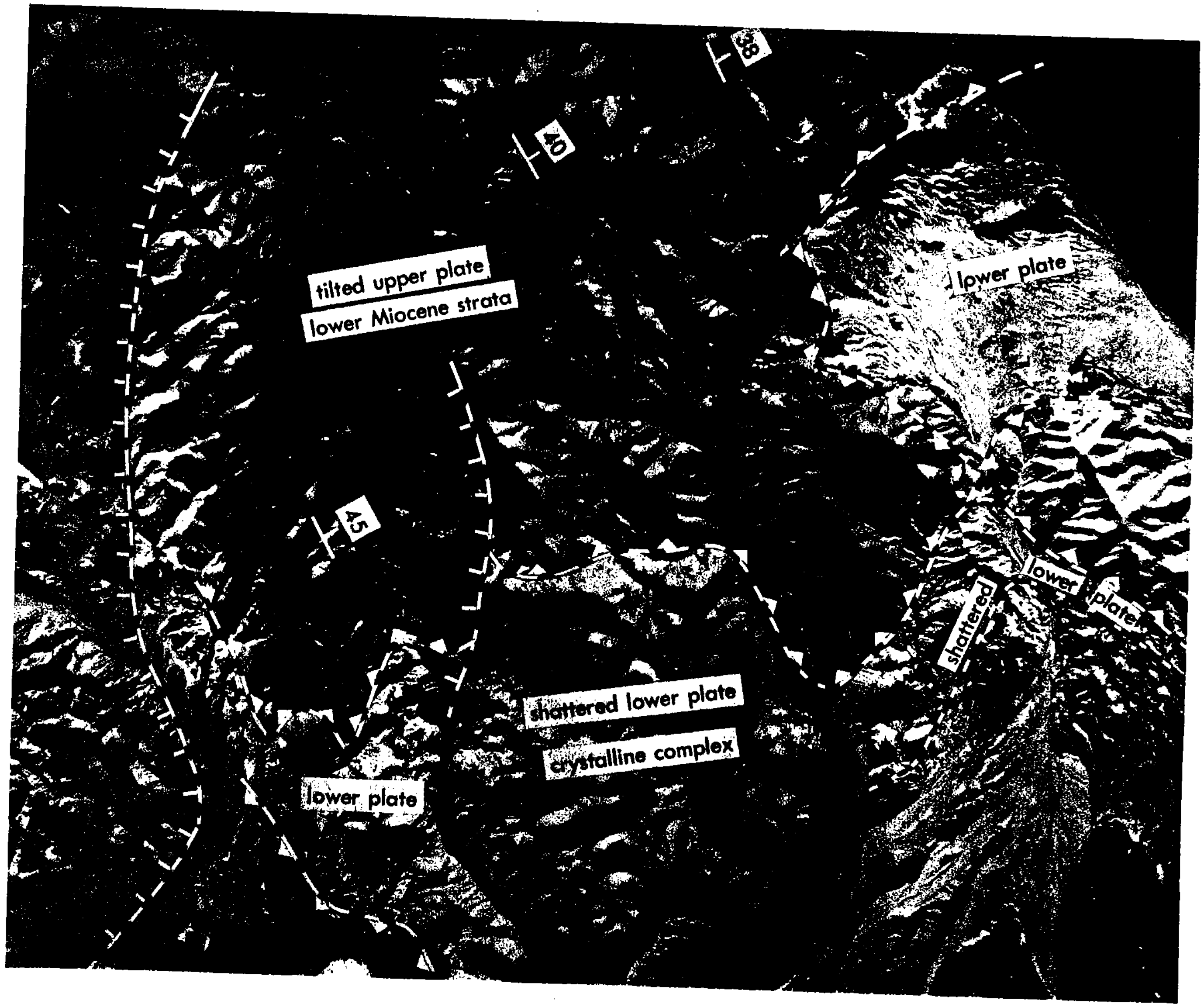

Figure 7. Airphoto of a portion of the northwestern Newberry Mountains (east of the Camp Rock Road). The Newberry Mountains detachment fault (NMDF) and associated upper plate high angle faults are highlighted. Note the curviplanar character of the NMDF and the uniformly tilted package of lower Miocene volcanic and sedimentary strata that lie above it.

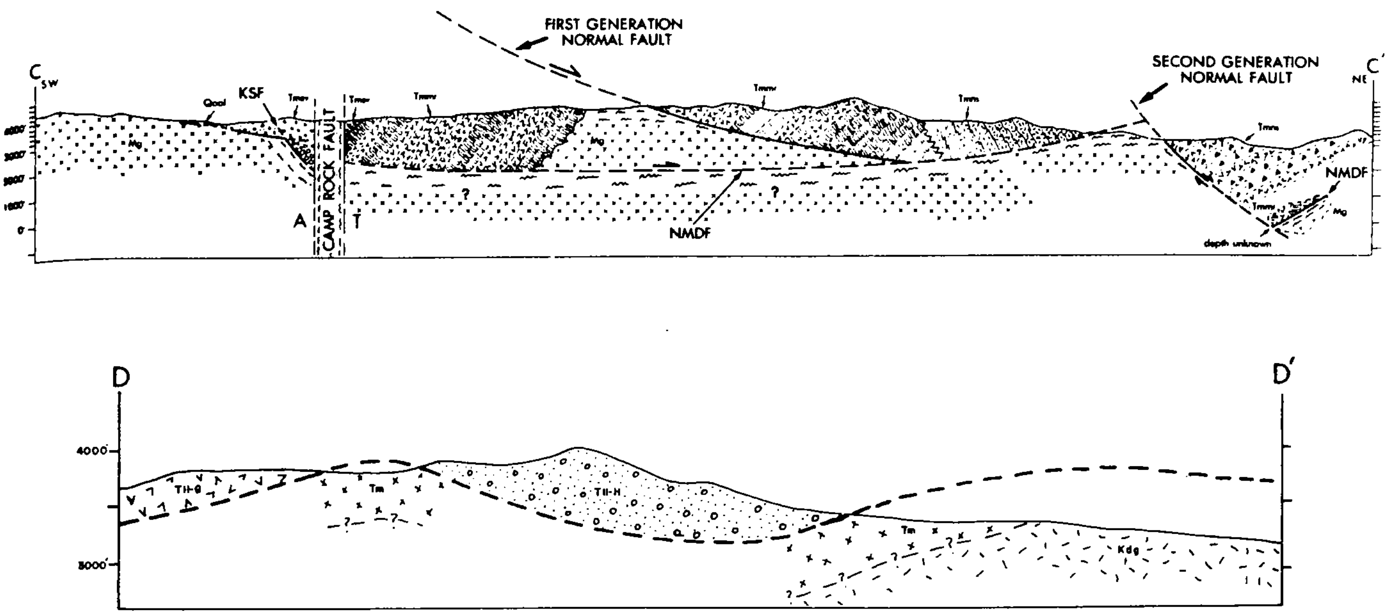

Figure 8. Cross sections across the Newberry Mountains. C-C' is a section drawn parallel to the inferred direction of transport of rocks lying above the NMDF (from Dokka, 1980). D-D' is a section drawn perpendicular to the transport direction (from Dokka, 1976).

Figure 9. Subhorizontal shears in the basement structurally below the Newberry Mountains detachment fault (near Daggett Ridge).

Figure 10a. Outcrop photograph of the tectonic microbreccia (granite protolith), 1m below NMDF.

Figure 10b. Photomicrograph of the tectonic microbreccia.

reason for discussing these structures is that they apparently formed synchronously with extension. We do not offer an explanation for their origin.

Three areas of the central Mojave Desert have been recognized thus far as containing these structures: Newberry Mountains, Cady Mountains and Bristol Mountains. This brief description is based on preliminary studies of these folded areas and should not be taken as the final word. As described above, the NMDF is an undulatory surface where it crops out in the Newberry Mountains. The fault is a wavy surface that can be divided into three segments based on its local shape. In the eastern half of this range, upper plate rocks reside in a broad synform that has a half-wavelength of 3.5 km. In the central part of the range, the NMDF is a complexly curved surface composed of shorter wavelength antiforms and synforms. Still farther to the west in the range, a large doubly-plunging antiform (half-wavelength = 10 km) is the dominant fold. A baffling aspect of the Newberry terrane is that whereas the NMDF is curviplanar in form, upper plate rocks adjacent to the fault (save areas near the Kane Springs fault) are not folded (Fig. 7). This suggests that the NMDF may have formed as a curved surface.

Two large basement-cored arches have recently been found in the central Cady Mountains and northern Bristol Mountains (Dokka, unpublished mapping). The Cady Mountains arch is currently the object of close study in order to document its geometry, associated fractures, and relation to the other structures. Preliminary observations suggest that this fold is a northeast-trending, doubly-plunging, upright, broad antiform whose geometry is defined by the now-warped basal unconformity between the underlying Mesozoic crystalline basement and lower Miocene volcanic strata (Fig. 11). The half-wavelength of this antiform is approximately 10 km but its amplitude

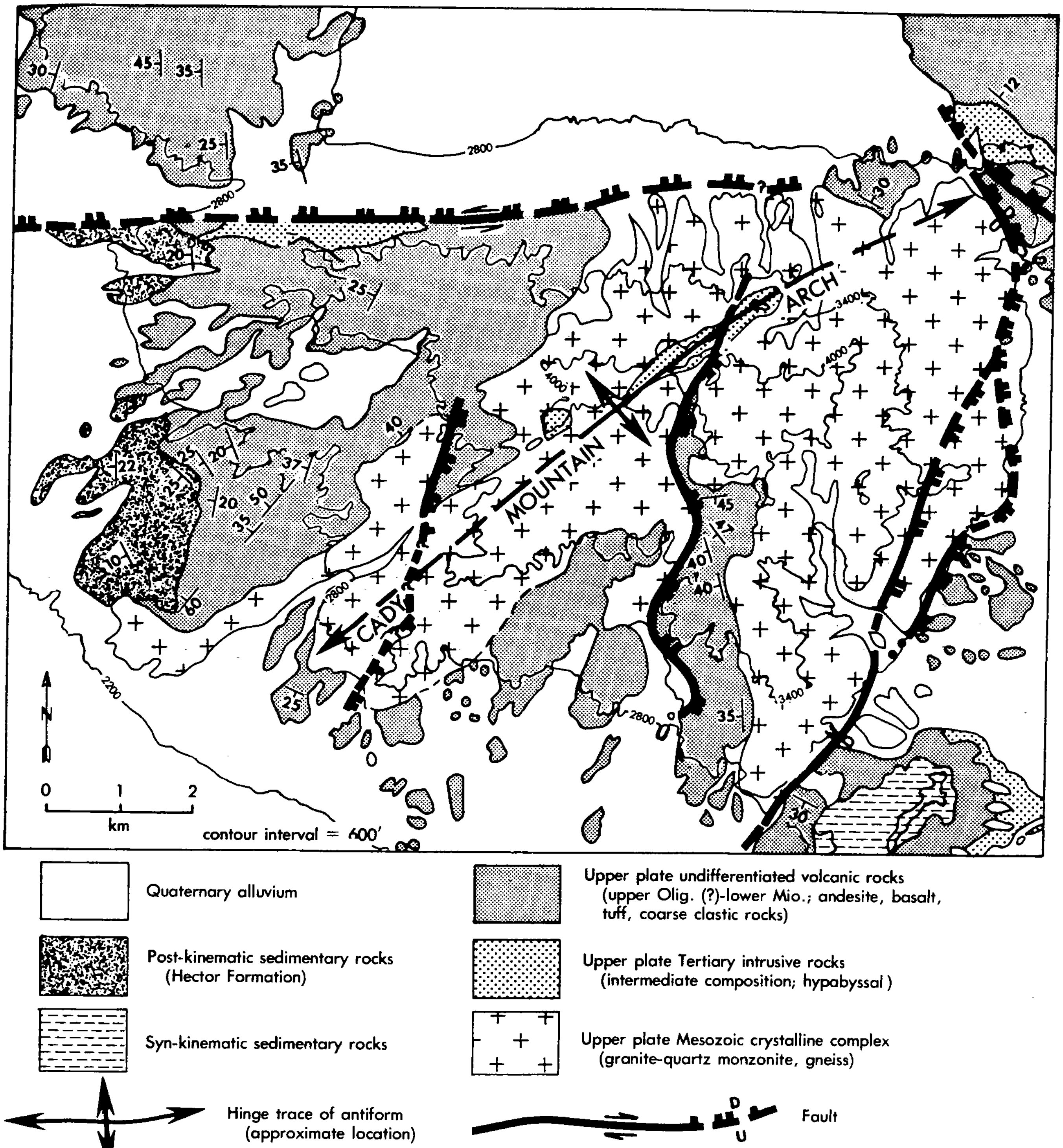

Figure 11. Geologic map of the west-central Cady Mountains (modified from Dibblee and Bassett, 1966). Normal faulting relations are more complex than shown.

is not resolvable from surface exposures. Less is known about the arch located in the northern Bristol Mountains. Recognition of this large northeast-trending (?) fold is based on the observation that the basement-lower Miocene unconformity in the Bristol Mountains changes orientation near the north end of the range. The contact to the south strikes northwest and dips to the southwest, whereas near the fold it gradually turns and strikes northeast and dips to the south. The northern limb of the fold is not observable.

Timing relations for the development of the Cady Mountain Arch can be inferred from the geologic map of figure 11. The youngest rocks that are folded are lower Miocene volcanic rocks (tuff and dacite). Overlying these rocks in angular unconformity is undeformed Hector Formation that is 21-17 m.y. (Woodburne and others, 1974). Thus, the age of fold formation is bracketed as 23-21 m.y. This age bracket coincides with the time of northeast-directed major extension of the central Mojave Desert.

DISCUSSION

In this section we will outline our interpretation of the intense deformation that so severely disrupted the central Mojave Desert during the early Miocene. This deformation changed the physiography of the region from a south and west-facing, low relief surface to a high relief terrain, composed of a main northwest-trending structural trough (Barstow-Bristol trough) and lesser parallel basins. This deformation produced a synkinematic structural association consisting of high-angle normal faults, detachment faults, moderate to extreme tilting of upper plate rocks, and regional long wavelength folds. In addition to describing a geometric model proposed for these structures, we will also discuss a number of working hypotheses that satisfy existing data. We will also comment on how this newly recognized terrane fits current tectonic models for the development of the Mojave Desert.

Earlier, one of us proposed a kinematic model that integrated high-angle normal and low-angle faulting within the context of upper crustal detachment (Dokka, 1980, 1981). Slightly curved to planar normal faults merged with a subhorizontal surface of detachment at a shallow level (2-7 km). Tilting of originally horizontal strata and upper plate crystalline basement was accomplished by simultaneous block rotation and faulting on multiple, subparallel, high-angle faults. As deformation progressed both faults and upper plate rocks rotated, with movement at depth being accommod by detachment along the NMDF. The thick microbreccia zone that lies below the NMDF is considered to be the result of tectonic erosion of upper plate rocks due to the imperfect merger of intra-terrane high-angle faults with the detachment. In short, the high-angle faults were not listric when they formed. The model for this terrane is depicted on figure 12 by way of a series of sequential cross-sections.

The origin of the detachment complex in the central Mojave Desert cannot be fully understood until the existence of, or lack of, extension below the detachment can be determined. The only range in the central Mojave Desert that contains lower plate rocks is the Newberry Mountains, and there no evidence for extension can be found. However, a group of hypotheses have been identified that satisfy the existing data.

Packages of rotated normal faults and strata and accompanying subhorizontal detachment zones are well-represented in the late Cenozoic history of the Cordillera, and generally occur in areas that experienced subsequent or concommitant extension. This strongly suggests that a genetic relation exists between this style of deformation and extensional tectonics, either as a primary manifestation or as a second effect. By a primary effect, we mean that the detachment faulting is also accompanied by lower-plate extension of an equivalent amount. This type of mechanism will be referred to below as crustal extension. The role of the detachment, then, is to act as a surface of decoupling between crustal levels that are extending by differing modes. An example of a secondary effect is denudation faulting, which results from gravitational instabilities caused by crustal extension. The following discussion is a brief comparison of the central Mojave terrane with several, well-documented examples of both primary and secondary types. However, before proceeding further the following point should be made. Although crustal extension models would seemingly require equivalence of strain above and below the detachment, this relation has yet to be demonstrated in any study. Further, it would not be surprising to find additional upper-plate extension as a result of down-gradient gravity sliding.

Several models have been proposed that consider detachment faulting as a mode of upper crustal extension. Each model calls upon a different mechanism to extend the crust below detachment. Longwell (1945) was the first to describe late Cenozoic low-angle normal faulting in southern Nevada. He viewed these structures as subordinate features, developed on the flanks of a rising anticlinorium that was produced in a regionally-distributed compressional stress field. Normal faults were thought to be developed in a rigid carapace above a region of plastic flow. Anderson (1971), working south of Lake Mead, proposed a different model where granitic plutons of middle Miocene age were emplaced into a dilating lower plate in response to a laterally spreading magma below. Detachment occurred within the basement (Precambrian schist, gneiss pegmatite) and at a depth probably less than 6 km (R. E. Anderson, 1982, pers. comm.). He viewed this area as one in which one could see how basin-range structures give way downward into horizontal laminar flow, as proposed by Hamilton and Myers (1966). The shallow level of deformation was the result of the high geothermal gradient produced by concurrent magmatism. Dokka (1980) argues that because the detachment is a zone of accommodation between regions extending by differing modes and amounts, lower plate extension need not be uniformly distributed beneath the detachment. Instead, lower plate extension could also be accomplished by concentrated distension within a narrow zone. Wernicke (1981) suggested that lower plate extension is even further removed, by considering detachment faults as major low-angle surfaces of crustal separation. Several other authors (G.H. Davis, 1975; Davis and Coney, 1979; Rehrig and Reynolds, 1980) have related upper-plate brittle extension to concurrent lower-plate ductile stretching in the context of the so-called

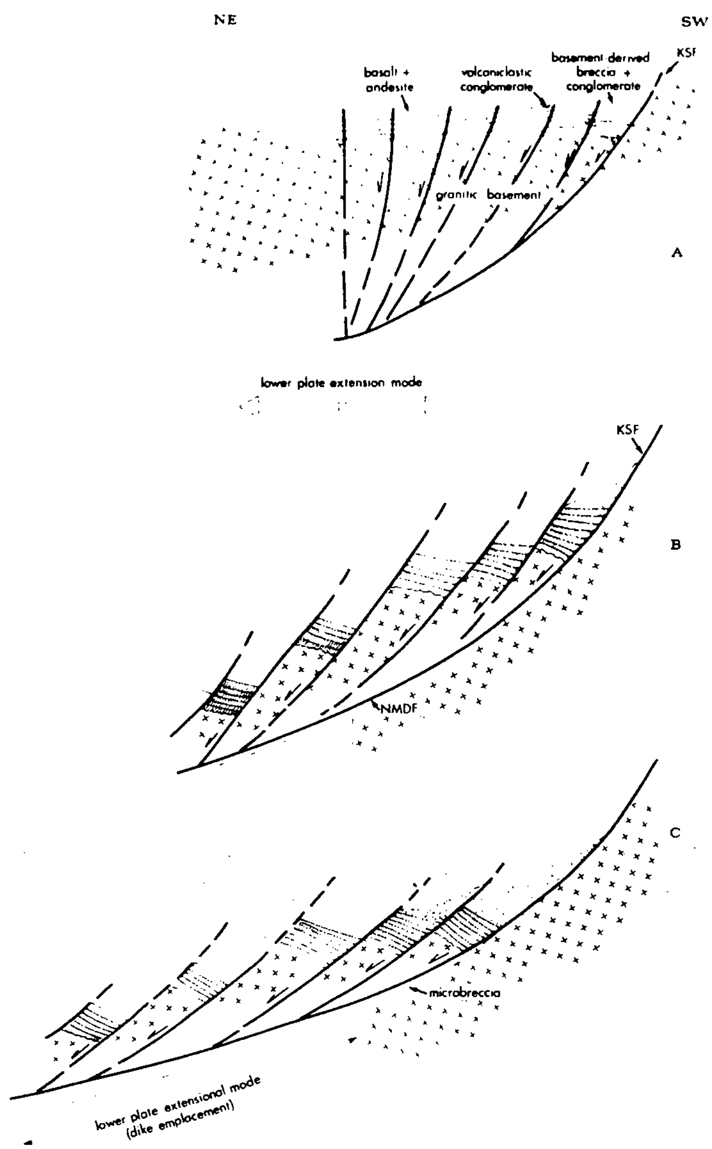

Figure 12. Developmental model for detachment faulting in the central Mojave Desert. See text for discussion.

metamorphic core complexes. These complexes occur from the Yukon to Sonora and are characterized by the development of lower-plate mylonitic rocks that contain a subhorizontal lineation that is generally perpendicular to the trend of this regional belt.

Viewing the central Mojave detachment complex as a secondary response to crustal extension, we mean that upper plate extension formed passively in response to more active processes that occurred nearby. For example, Armstrong (1972) cited several examples from the eastern Great Basin of denudation faults that formed in respense to relief produced by concurrent basin-range normal faulting. Crustal extension localized in one area is thought to have produced a gravitational instability and resulted in material in-filling from surrounding areas. This style of deformation should not be confused with local landsliding in that it is a large-scale response of the crust to lithospheric separation.

In summary, data suggest that the central Mojave Desert experienced a short, but remarkably intense, interval of upper crustal extension during the early Miocene. Extension developed within a northwest-trending belt of calc-alkaline volcanism that slightly predated deformation. The interior of the extensional zone was low and received alluvial fan debris and lake sediments. This same belt has continued act as a sediment and drainage sink and as the locus for volcanism since the early Miocene. Moderate to extreme upper plate tilting of lower Miocene strata and older crystalline rocks occurred simultaneously with movement along nearly planar, originally high-angle intraterrane normal faults. These faults merged with a regional low-angle detachment fault that coordinated movements. The existence of adjacent oppositely tilted domains suggests bimodal transport along a N.50E. - S.50W. extension direction. Recently discovered concomitantly developed long wavelength folds whose axes are subparallel to the extension are not understood by us. Minor extension continued locally after the main phase of extension and probably ended no later than 16 m.y. (deposition of Barstow Formation).

These new data are clearly in conflict with syntheses that have considered the Mojave Desert a stable (i.e., no crustal extension) block throughout most of the Cenozoic (Hewett, 1955; Dibblee, 1967; Garfunkel, 1974; Gardner, 1980). Clearly, a new synthesis is needed. Regional models must also be adjusted to consider the Mojave's mobile past.

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STRATIGRAPHY AND GEOCHRONOLOGY OF MIOCENE STRATA IN THE CENTRAL MOJAVE DESERT, CALIFORNIA

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Abstract

Strata of early and medial Miocene age comprise a major element of the Tertiary geological sequence in the Mojave Desert, California. The rock units discussed here occur in, and reflect the geological evolution of, several outcrop districts in the central part of this region, known as the Central Mojave Basin. The stratigraphic, paleontological, and radiochronological evidence contained therein is discussed so as to determine the general pattern in, and historical geological relationships between the several districts. From this, a framework can be developed to guide additional studies, as well as to serve as a reference of comparison with other rock sequences in the Mojave Desert province and elsewhere.

The following broad depositional and extrusive history seems plausible (categories are not necessarily mutually exclusive). 1) 23 - 24 Ma. Major extrusion of andesitic to rhyodacitic flows, associated with breccias, lahars, and local quartz latite ash-flow tuffs, and fluviatile/lacustrine sedimentation. A subsequent regional unconformity and change in volcanism separates this from; 2) 22 - 14 Ma. Initiation, and locally sporadic continuation of, basaltic extrusion, and production of air-fall tuffs. 3) 19 - 18 Ma. Locally violent rhyolitic volcanism resulting in tuffs, tuff breccias, and a major ignimbrite. 4) Lacustrine deposition is an important component of the record in the northern Cady Mountains-Afton Canoyn area from 22 - 16 Ma (and no younger Miocene rocks preserved). In contrast, lacustrine units are most common in the western part of the Central Mojave Basin from about 16 to 12 Ma (although locally present earlier). 5) Based on rock sequences of early and medial Miocene age in the districts studied here, it is possible to infer the general location of the northern and western, but not the eastern, and but little of the southern, limits of the central Mojave Basin.

Introduction

The purpose of this report is to summarize the geology, stratigraphy, biostratigraphy, and geochronology of sequences of rocks and strata of early and medial Miocene age in the central Mojave Desert. The study extends from the Gravel Hills on the west to the Cady Mountains on the east, and includes the following intervening districts, from west to east: Mud Hills, Calico Mountains, Yermo Hills, and Alvord Mountain. The stratigraphic and facies relationships of the rock units in these districts suggest that they are generally related and were deposited in parts of an essentially continuous basin. Even though the location of the southern limit of the basin is not well represented as yet, we use the new term, Central Mojave Basin, to refer to the regionally integrated setting in which the deposits discussed here were accumulated. This area includes the Barstow Basin of Dibblee (1967, fig. 71) at its western end, but encompasses districts 60 to 75 km to the east of that basin as cited by Dibblee (1967) or by Burke et al. (In press), respectively.

In general, deposition in the several districts began about 23 - 24 Ma ago with coarse-grained fluviatile deposits and/or volcanic (andesitic to dacitic) flows. This is followed generally by continued alluvial and minor lacustrine deposition with locally important eruptions of rhyolitic to dacitic ash. Miocene basaltic volcanism in this area occurred as early as 22 - 23 Ma and continued until about 14 Ma; lacustrine and locally prominent fluviatile sedimentation continued, with contributions of tuff and minor ignimbrite units until about 12 Ma, the upper age limit of the bulk of the Tertiary rock units in this area.

Subsequent geological evolution of much of the Mojave Desert Province is recorded in districts that are peripheral to the area discussed here. Later geological features of the central Mojave Desert reflect a different structural/tectonic regime which resulted in a different pattern of sedimentation, in contrast to the more regionally integrated setting within which rocks and deposits of early and medial Miocene age can be evaluated.

History of study

The Mojave has long been a desert of geological knowledge as well as in terms of the present-day climate. Evidence regarding the Tertiary stratigraphy began to accuumulate at the beginning of the present century (Hershey, 1902; Baker 1911; Merriam, 1919), but major advances have taken place in the interval since Hewett (1954) published the first general summary of a number of rock units discussed here. Other important works include McCulloh (1952), Byers (1960), and Dibblee (1967, 1968). In addition to these we draw from the regional considerations developed by S.T. Miller (1980), and an unpublished manuscript compiled by R.H. Tedford and colleagues. Other important works, of more local significance, will be cited under the appropriate sections.

Acknowledgments

We are grateful to the late T. Galusha, as well as to Morris F. Skinner and Beryl E. Taylor, of the Frick Laboratory, American Museum of Natural History, New York, for their help, consultation, and discussion during present and past years on the general topic of Mojave Desert paleontology and stratigraphy, and for developing some of the most important fossil mammal collections from this region. We also wish to thank D.E. Savage, University of California Museum of Paleontology, D.P. Whistler, Natural History Museum, Los Angeles County, for their support and assistance in increasing our knowledge of the Tertiary geology of the Mojave Desert during the past several decades. The paleontological work of R.M. Alf, Webb School, Claremont, and R.H. Reynolds, San Bernardino County Museum, also are to be recognized and appreciated here. The work on the stratigraphy of the Barstow Formation in the Mud Hills by R.P. Steinen, then a student at UCR, but now at the Department of Geology, University of Connecticut, has been a fundamental aid to our studies in that area; the studies of E.H. Lindsay, Department of Geosciences, University of Arizona have similarly aided our work in this district. Of the many students of the Department of Earth Sciences who particiapted in class field trips during the past 23 years and whose efforts contributed to our understanding of the stratigraphy of this region, especial recognition is given to Craig E. Moseley, Donald A. Williamson, and Brian E. Casey, for the contribution they have made to our understanding of the geology of the Cady and Bristol mountains. We also have benefitted from support and consultation with, and field programs of, Paul T. Robinson, of the Department of Earth Sciences. C. Rader, Lucerne Valley, and R.W. Daily, Riverside, California, have maintained a continuing interest in Mojave Desert paleontology and have found and contributed a number of important fossil specimens. Karoly Fogassy and Robert Hicks, Department of Earth Sciences, graciously prepared the illustrations used here. Grants from the University of California, Riverside, Adademic Senate Committee on Research defrayed some of the expenses associated with these studies; this support is gratefully acknowledged, as well.

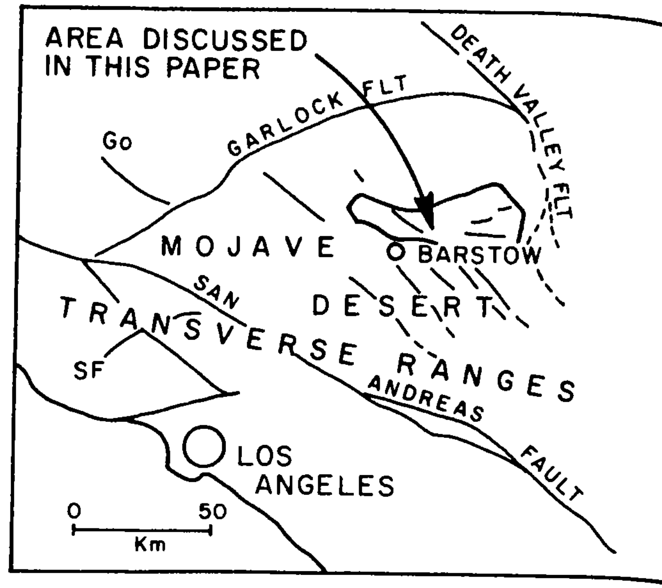

Figure 1. Index map of the Mojave Desert and adjacent provinces, showing the central Mojave Desert, discussed in this paper. SF = San Francisquito Formation (Paleocene, marine) of the Ventura Basin district. GO = Goler Formation (Paleocene, nonmarine) of the El Paso Mountains. Both reflect early Tertiary erosion of the Mojave Desert province.

Conventions

Stratigraphic and areal dimensions are given in metric notation. Taxonomic names that appear in " " indicate general use of a name that probably should be restricted to other species. All K-Ar dates cited have been recalculated when necessary to conform to the current IUGS decay constants (Steiger and Jaeger, 1977).

DES refers to field sample number of D.E. Savage.

Regional setting

The Mojave Desert Province (Fig. 1) is bounded on the north by the Garlock Fault, and on the southwest by the San Andreas Fault. Its eastern boundry is more diffuse, but for the purposes of this study can be taken at the point where basin-and-range structures of the Death Valley fault zone break southeastward past the eastern end of the Garlock Fault.

The Mojave Desert displays large expanses of alluvium and playa lake deposits of Quaternary age that lie at elevations of about 600 m, and are pierced locally by hills and small mountain ranges that reach elevations of 1,400 m or more. Many of these ranges contain plutonic rocks of later Mesozoic age, composed of quartz monzonite, granodiorite, or quartz diorite (Dibblee, 1967, 1968; Miller and Morton, 1980). Roof pendants of older rock are rare in area of study, although present regionally (Rogers, 1967; Jennings, Burnett, and Troxel, 1962). In most of the districts discussed here, a thick sequence of Tertiary volcanic rock and volcaniclastic and fluvio-lacustrine strata unconformably overlies the basement terrane. The Tertiary rocks are of early to medial Miocene age (ca 24 - 12 Ma), and record the onset of internal drainage in the Mojave Desert Province (e.g. Hewett, 1954; Byers, 1960). Rocks or deposits of earlier Tertiary age are unknown in this region, although their presence has been suspected (Byers, 1960). Early Tertiary sediments are present on the northern and southwestern margins (Goler Formation, Paleocene nonmarine, El Paso Mountains; and San Francisquito Formation, Paleocene marine, Transverse Ranges; Go and SF respectively, Fig. 1). Both instances reflect drainage patterns away from the Mojave Desert province (B. Cox, pers. commun. to M.O. Woodburne, 1981; Kooser, 1981) and contribute to the long-held (Reed, 1951; Hewett, 1954) interpretation that the desert area was a high-standing feature at that time.

Subsequent to, and locally synchronous with, their deposition and extrusion, the early and medial Miocene units have been variously tilted, folded, and cut by faults that generally trend northwest and show right-lateral separation or - more rarely - trend east-west and show at least recent left-lateral separation as well as more evident dip-slip motion.

Models proposed recently to relate the geological features of the Mojave Desert to the general tectonic regime of western North America include (1) plate margin subduction reflected by the onset of Tertiary volcanism (Atwater, 1970; Lipman et al., 1972; Christiansen and Lipman, 1972), (2) inter-plate strain reflected in counterclockwise rotation of fault-bounded blocks (Garfunkel, 1974,

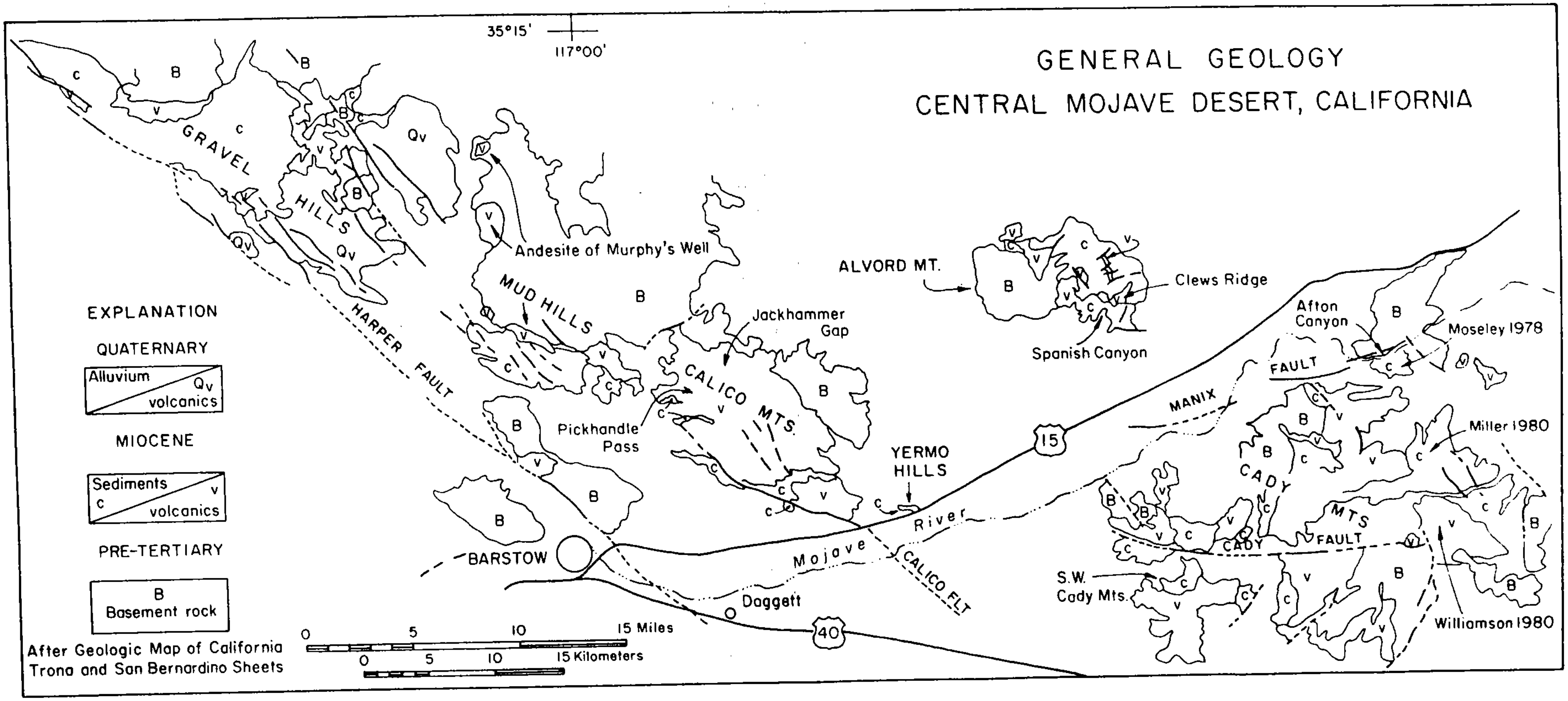

Figure 2. Generalized geologic map of the central Mojave Desert, showing the areas and some of the locations discussed in this report. Moseley, 1978; Miller, 1980; Williamson, 1980, refer to areas studied by these workers, cited in the text.

Cummings, 1976), (3) or by clockwise rotation of fault-bounded blocks (Kamerling and Luyendyk, 1979; Luyendyk et al., 1980), and (4) compression denoted as late Oligocene or early Miocene detachment faulting (Dokka, 1979; Dokka and Glazner, 1982); Miller and others, 1982).

The summary below sets out the stratigraphic and temporal relationships of these rocks irrespective of the above models so as to determine the general depositional and extrusion pattern in, and historical geological relationships between, the several areas. The emphasis here is upon the basic stratigraphic, paleontological, and radiochronological data to be found in the rock sequences of early and medial Miocene age, with evidence and interpretations focused on physical stratigraphic and other age-significant information. We hope that this will form a framework within which to evaluate the geological evolution of this part of the Mojave Desert, and will serve as a reference for comparisons to rock units in other areas. The treatment begins in the Gravel Hills on the west and progresses to the Cady Mountains on the east, a distance of about 85 km (Fig. 2).

Gravel HLills

The Gravel Hills (Fig. 2), located about 35 km northwest of Barstow, California (Figs. 1, 2), have been discussed most recently by Dibblee (1968). Most of what follows is taken from that source, modified where necessary by information contained in Burke et al. (In press). The Miocene sequence in the Gravel Hills begins with the Pickhandle Formation (including the Opal Mountain Volcanics of Dibblee, 1968 = Opal Mountain Volcanic Member; Burke et al., In Press), and the overlying Barstow Formation (Fig. 3). The Barstow Formation is unconformably overlan by the Black Mountain Basalt, formerly (Dibblee, 1968) thought to be of Pleistocene age, but more recently (Burke et. al., In Press) shown to be about 2.5 $\pm$ 0.6 Ma old, or Pliocene in age.

The Pickhandle Formation in the Gravel Hills (Dibblee, 1968, p. 21) is composed largely of pyroclastic materials. As much as 600 m of light gray breccia and tuff is interbedded locally with (1) basalt and limestone overlain by beds of white lithic and lapilli tuff; (2) local lenses of conglomerate that contain cobbles of granitic and dioritic rock; and (3) rhyolitic flow breccia. The Pickhandle Formation unconformably overlies pre-Tertiary quartz monzonite, and is conformably overlain by red rhyolitic breccia. This breccia was assigned to the Opal Mountain Volcanics by Dibblee (1968, p. 21) who thought that this unit was interbedded with the Pickhandle Formation. Burke et al. (In press) relegate the Opal Mountain Volcanics to member status within the Pickhandle Formation, and report a K-Ar age of 18.9 $\pm$ Ma on quartz latite welded tuff in the top of the unit at Opal Mountain (Fig. 3). According to Burke et al. (In press), the welded tuff and other materials of the Opal Mountain Volcanic Member pertain to a final explosive phase of volcanism in the Barstow Basin area that initially generated the large volume of lithic tuff and tuff breccia of the Pickhandle Formation.

The Opal Mountain welded tuff establishes the upper limit of the Pickhandle Formation in the Gravel Hils. This date is consistent with the age of the Barstow Formation in the eastern Gravel Hills, where a local unconformity separates the Pickhandle from the Barstow Formation. In this area (section 18, T. 32 S., R. 45 E.), an age of 16.5 $\pm$ 0.4 Ma was obtained from a basalt flow interbedded in a sequence of tuffaceous sandstone, limestone and chert, considered to be Barstow Formation by Burke et al. (In press); Dibblee (1968; pl. 1) previously had assigned this sequence to the Pickhandle Formation.

Dibblee (1968, p. 22) suggests that the Pickhandle Formation was derived from volcanic centers in the Gravel Hills and Opal Mountain areas as well as from sites in the Calico Mountains, 35 km to the southeast. He indicates (op. cit. p. 23) that the Pickhandle Formation may have been deposited in a large valley that extended between these sites, and notes the absence of these rocks in wells drilled south of Harper Valley (to the south of the Gravel Hills) to suggest that they may never have been deposited in that area. The Pickhandle Formation, then, was of relatively local origin and probably was derived in part from a rugged topographic margin (see below) that served to limit the northern boundary of the basin in which the succeeding Barstow Formation was deposited.

The Barstow Formation is exposed more widely in the central and western parts of the Gravel Hills, than in the east (c, fig. 2); in these areas, the Barstow Formation consists of a sequence of lacustrine sediments, overlain (Fig. 3) by basalt, beds of more coarse-grained, fluviatile, sandstone, and thick units of fanglomerate that contain granitic and volcanic detritus (Dibblee, 1968, pl. 1, p. 30-31). The total sequence ranges up to about 1,500 m in thickness. The basalt in Black Canyon is undated and may not be the same age as the basalt flow dated by Burke et al. (In press). Sparse fossil mammals in the deposits below the basalt in Black Canyon suggest the presence of elements of the Green Hills Fauna (early Barstovian) of the Mud Hills, calibrated there as between 16.3 - 15.5 Ma old, whereas mammal fossils from the beds above the basalt appear to be comparable to elements of the younger, Barstow Fauna of the Mud Hills (ca 15 - 13 Ma old; see below). The mammal remains are fragmentary, however, and geochronological conclusions based on them are tentative, at best (see also Lewis, 1968b, p. 35).

The units that underlie the fanglomeratic deposits are exposed only locally, but the two facies of coarse-grained units - granitic on the west and southwest, volcanic debris on the north (Fig. 3) - reflect deposition from two different sources. Dibblee (1968, p. 32) suggests that the fanglomerate of granitic and quartz latite clasts probably accumulated as eastward-sloping alluvial fans, derived from rising basement terranes to the west and southwest. The fanglomerate of volcanic detritus apparently accumulated at the same time as southward-sloping alluvial fans derived from the northwest. The two-sided basin postulated for the deposition of some parts of the Barstow Formation in the Gravel and Mud Hills (see below) differs from the reconstruction of a more localized linear source for the underlying Pickhandle Formation. Sparse fossil mamals from these coarse-grained strata suggest that they could be as young as late Barstovian in age (Fig. 3). Although deposits of generally similar character interfinger on the west and southwest with fossiliferous strata of Barstovian age in the Mud Hills, there is no physical continuity between these and the fanglomeratic units of the Gravel Hills.

STRATIGRAPHY OF EARLY AND MEDIAL MIOCENE EXTRUSIVE AND SEDIMENTARY ROCKS, CENTRAL MOJAVE DESERT.

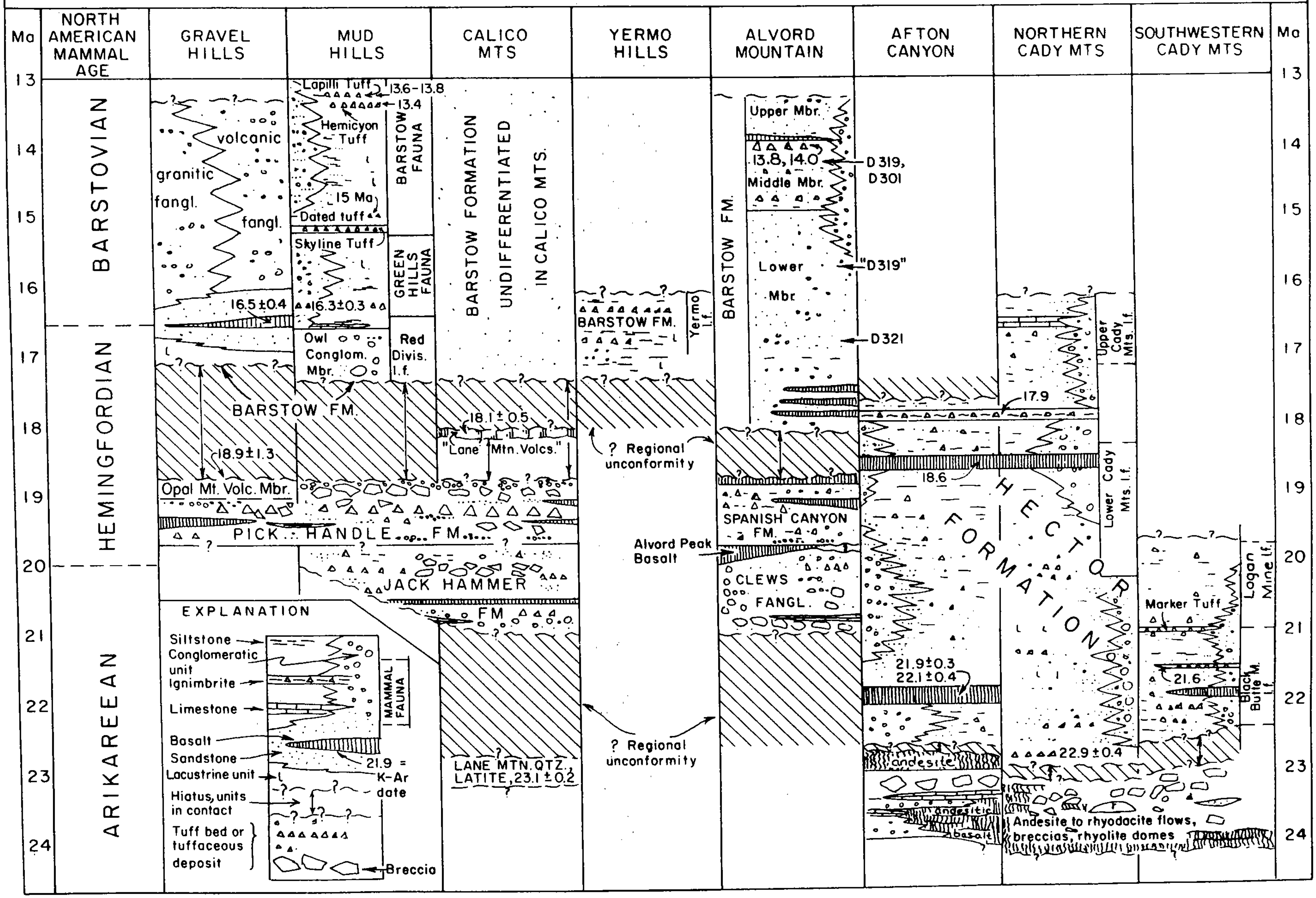

Figure 3. Stratigraphic diagram of rock sequences discussed in this report, showing distribution of major lithologic types, stratigraphic position of important fossiliferous and radiometrically dated units.

The youngest unit of Miocene age in the vicinity of the Gravel Hills is the "Lane Mountain Andesite," mapped (Dibblee, 1968, pl. 1) in the vicinity of Murphy's Well about 10 km north of the western end of the Mud Hills (Fig. 2). This unit (not shown on Fig. 3), is designated as Andesite of Murphy's Well by Burke et al. (In press), and separated from the Lane Mountain Andesite of the Calico Mountain district. The Murphy's Well andesite (Dibblee, 1968, p. 37) is a brown to dark gray massive, porphyritic, unit that unconformably overlies pre-Tertiary quartz monzonite, but is otherwise stratigraphically unconfined except by Quarternary alluvium. Burke et al. (In press) report a K-Ar age of 13.5 $\pm$ 0.1 Ma for this andesite and note that it is very similar in age to tuffaceous units in the upper part of the Barstow Formation of the Mud Hills.

Mud Hills

A relatively complete sequence of deposits of early and medial Miocene age crops out in the Mud Hills, north of Barstow, California (Figs. 2, 3). The sequence consists of the Jackhammer, the Pickhandle, and the Barstow formations, in ascending order.

The type locality of the Jackhammer Formation is in Jackhammer Gap (Fig. 2), in the northwestern Calico Mountains (McCulloh, 1952, in Dibblee, 1968), but the unit is preserved in its western-most occurrence on the north limb of the Barstow syncline in the Mud Hills, at the head of Owl Canyon (Fig. 2; Dibblee, 1968, pls. 1, 3; Woodburne and Tedford, 1982, Fig. 3). There the unit unconformably overlies pre-Tertiary quartz monzonite, is about 30 m thick, and is unconformably overlain by the Pickhandle Formation. The Jackhammer Formation of the Calico Mountains reaches a thickness of about 240 m and includes beds of conglomerate, basalt, tuff, and tuff breccia, and tuffaceous sandstone. In the Mud Hills, a lower unit of arkosic sandstone and conglomeratic sandstone (23 m thick), is overlain by a lenticular sequence of white bentonite and arkosic tuff about 2 m thick, overlain in turn by 10 m of black amygdaloidal basalt (Fig. 3). The basalt and the sandstone unit pinch out westward between the pre-Tertiary basement and the overlying Pickhandle Formation.

The age of the Jackhammer Formation is unknown. Its conformable relationships with the overlying Pickhandle Formation suggest that the two units are not greatly different in age. Burke et al. (In press) suggest on this basis that the Jackhammer Formation may be younger than the Lane Mountain Quartz Latite of the Calico Mountains district, dated by them at 23.1 $\pm$ 0.2 Ma.

The Jackhammer Formation apparently was one of the earliest units to be deposited in the basin that extended from the Calico Mountains on the southeast toward the Gravel Hills on the northwest, but is of lesser regional extent, as preserved, than the immediately overlying Pickhandle Formation.

Like the Jackhammer, the Pickhandle Formation in the Mud Hills crops out only on the northern limb of the Barstow syncline. The Pickhandle is about 1,360 m thick, and is comprised of a basal unit of beds of gray conglomerate, overlain by greenish to reddish-gray arkosic to tuffaceous sandstone. A middle unit is composed largely of lithic tuff and andesitic tuff breccia. This is followed by an upper unit of granitic and rhyolitic breccia and of megaconglomerate that is unconformably overlain by the Barstow Formation (Dibblee, 1968, p. 19-23).

The age of the Pickhandle Formation given in the Gravel Hills is compatible with its stratigraphic setting in the Mud Hills, where fossil-bearing beds low in the Barstow Formation are of late Hemingfordian age (see below, and Woodburne and Tedford, 1982).

The general depositional setting of the Pickhandle Formation has been mentined above. The spectacular, thick lenses of monolithologic granitic and rhyolitic breccia exposed in the upper part of the Pickhandle Formation in Owl Canyon may have been deposited as landslides or debris flows. The size of the blocks suggests near-by rather than more distant provenance, and the fact that they overlie tuffaceous units of volcanic origin in the Pickhandle Formation suggests that the breccia units were the result of local tectonic activity that produced a rugged terrance of granitic mountains on the northern margin of the Pickhandle, and later, of the Barstow, basins in this district (Dibblee, 1968, p. 22, 23).

The stratigraphy and geochronology of the Barstow Formation of the Mud Hills is reviewed by Woodburne and Tedford (1982). The unit is comprised of about 1,300 - 2,000 m of fluviatile, lacustrine, and locally tuffaceous sedimentary rocks that have been deformed into a syncline and cut by faults that generally trend northwestward and display right-lateral separation.

The formation ranges in age from 17 Ma or older at its base to about 12 Ma at its top, and reflects a depositional regime in which coarser-grained fluviatile sandstone and conglomeratic sandstone beds become finer-grained eastward and northeastward and give way to thick sequences of beds of sandstone, lacustrine mudstone and claystone, and water-laid air-fall tuff.

The base of the Barstow Formation is composed of the Owl Conglomerate Member, about 65 m thick, that unconformably overlies the Pickhandle Formation. Owl Conglomerate Member is composed largely of conflomerate and conflomeratic sandstone beds that contain granitic and lesser amounts of volcanic clastic constituents. The unit generally is limited to the northeastern and southern parts of the Mud Hills and was deposited from source areas located both to the north and south of the Barstow Basin. Sparse fossil mammal remains of the Red Division Local Fauna (Fig. 3); "Merychippus" carrizoensis, Merychyus [Metoreodon] from near the top of the unit are of late Hemingfordian age, calibrated elsewhere as older than about 16.1 Ma and younger than about 17.6 Ma. On this basis, the Owl Conglomerate Member should be at least 17 Ma old (Woodburne and Tedford, 1982).

The middle part of the Barstow Formation is gradational with the Owl Conglomerate Member, and reaches a thickness of about 700 m. On the southern limb of the syncline, this interval of the Barstow Formation is composed of a western facies of fluviatile conglomeratic sandstone beds that grade eastward into finer-grained beds of generally lacustrine origin. Similarly fine-grained strata predominate in the middle part of the Barstow Formation on the north limb of the syncline.

A bed of tuff above the middle of this part of the Barstow Formation has been dated at about 16.3 ± 0.3 Ma (Fig. 3). Fossil mammals found a short distance stratigraphically below the tuff, and in beds that extend upward to the top of the unit, contain elements of the Green Hills Fauna (Fig. 3; "Merychippus" stylodontus, Peridoimys, Eupolocyon, Amphicyon cf. ingens, Brachycrus buwaldi, Rakomeryx, Merriamoceros, Hesperhys, Aepycamelus, and Hemicyon (Plithocyon), of early Barstovian age, as well as the Cupidinimus nebraskensis Assemblage Zone of Lindsay (1972).

The upper part of the Barstow Formation begins with the Skyline Tuff (Fig. 3), or "lower marker tuff bed," of Dibblee (1968), and extends upward for about 400 m. The exposures are most complete in the western part of the Mud Hills. Overall, the unit is composed dominantly of beds of lacustrine claystone, mudstone and water-laid air-fall tuff, but - as for the middle part of the formation - these interfinger westward with coarser-grained fluviatile deposits.

The tuff beds in the upper part of the Barstow Formation contribute importantly to its biostratigraphy and geochronology. Fossils from the Skyline Tuff to the top of the formation represent the Barstow Fauna (Fig. 3; "Merychippus" intermontanus, "M." sumani, Archaeohippus mourningi, Megahippus mckennai, Paramoceros, Meryceros, Mediochoerus, Parapliosaccomys, Aelurodon, and Gomphotherium), of late Barstovian age. The Pseudadjidaumo stirtoni Assemblage Zone of Lindsay (1972) occurs in the deposits that begin at the Skyline Tuff and continue upward stratigraphically for about 80 m. A tuff known as the Dated Tuff occurs 17 - 30 m stratigraphically above the Skyline Tuff, and yielded K-Ar dates of about 15 Ma (Woodburne and Tedford, 1982). The Cupidinimus longidens Assemblage Zone of Lindsay (1972) extends from 80 - 150 m above the Skyline Tuff; the interval includes the Hemicyon Tuff - ca 100 m above the Skyline Tuff and dated at 13.4 ± .7 Ma, and the Lapilli Sandstone tuff - ca 15 m above the Skyline Tuff and dated at 13.6 - 13.8 Ma (see Woodburne and Tedford, 1982). The Copemys russeli Assemblage Zone follows the C. longidens Assemblage Zone and continues to a level about 250 m above the Skyline Tuff. As noted above, the Andesite of Murphy's Well (not shown on Fig. 3) is comparable in age to tuffs in the upper part of the Barstow Formation, but the genetic relations, if any, between these units, is not known.

As noted above, the lower part of the Barstow Formation was deposited from sources to the north as well as the south or southwest, and reflects the prsence of apparently high-standing source terranes in those areas. In contrast, the deposits of the middle and upper parts of the Barstow Formation in the Mud Hills reflect deposition from the southern or southwestern margin of the basin, with lacustrine deposition recorded by beds in the eastern and northern parts of the Mud Hills. Input from sources in the northern part of the Barstow basin apparently is not recorded in these deposits.

Calico Mountains

This range, well known for silver mining activity in the past, is located immediately east and southeast of the Mud Hills (Fig. 1). The Calico Mountains contain a thick sequence of eruptive and intrusive volcanic rock of early to medial Miocene age, with minor intercalated sedimentary deposits. These units, which locally are complexly folded and are cut by northwest-trending faults, unconformably overlie both intrusive rocks of Mesozoic age and Paleozoic carbonate rocks. The units of Miocene age (Fig. 3) include the Lane Mountain Quartz Latite, the Jackhammer, the Pickhandle, and the Barstow formations (Dibblee, 1968, McCulloh, 1952, Dibblee, 1970; Burke et al., In press).

Dibblee (1968, p. 36) interpreted the Lane Mountain Andesite (=Lane Mountain Volcanics of McCulloh, 1952) to be younger than the Barstow Formation of this district. Burke et al. (In Press) indicate that this is not the case; a K-Ar date of 23.1 ± 0.2 Ma was obtained from a basal tuff in the unit. As such, the Lane Mountain Quartz Latite (nomenclature follows Burke et al., In press) is the oldest Miocene unit in the Calico Mountains, and in the general area of the Barstow basin that extends northwestward to the Gravel Hills. It is not equivalent to units mapped by McCulloh (1952) as overlying the Jackhammer and Pickhandle formations in the Calico Mountains, nor to the isolated andesite at Murphy's Well (K-Ar date of 13.5 ± 0.1 Ma), mapped by Dibblee (1968, pl. 1) as Lane Mountain Andesite.

According to Dibblee (1968, p. 36) the Lane Mountain Quartz Latite of Lane Mountain is a large extrusive body that lies unconformably on pre-Tertiary basement rock. McCulloh (1952; in Dibblee, 1968) indicates that the units that he included in the "Lane Mountain volcanics" are nearly horizontal flows of calcic hornblende andesite, dacite, and quartz-bearing latite; the rocks occur as remnants of multiple flows that lie on basement rocks in the northwest, and on tilted sediments of the Jackhammer and Pickhandle formations to the south. Burke et al. (In press) note that the latter rocks are much younger than the flows at Lane Mountain; that the so-called Lane Mountain 'volcanics' (Fig. 3) that unconformably overlie the Jackhammer and Pickhandle formations in Pickhandle Pass and in Jackhammer Gap are 18.1 ± 0.5 Ma old, in contrast to the ca 23 Ma date obtained from the unit in its type area.

As restricted by Burke et al. (In press), the Lane Mountain Quartz Latite (Fig. 3) is of limited areal extent, and represents remnants of a number of welded quartz latite ash-flow sheets that locally contain unwelded tuff and vitrophyre at the base. The rocks apparently were derived from a vent of unknown location, but may have been part of a number of composite ash flow units that formerly were more widespread. If so, the areal restriction of the Lane Mountain Quartz Latite suggests considerable erosion subsequent to its eruption and deposition in the Barstow basin. As indicated on Fig. 3, this erosional hiatus may have been regionally important, and separated the sequence of dominantly andesitic to rhyodacitic extrusive rocks from that represented dominantly by basaltic and tuffaceous volcanism and fluvio-lacustrine sedimentation.

The Jackhammer Formation in the Calico Mountain-Lane Mountain district probably is younger than the Lane Mountain Quartz Latite, although the two units are nowhere in contact. Evidence in favor of this is the conformable contact between the Jackhammer and Pickhandle formations. An upper age-limit for the Jackhammer and Pickhandle sequence is provided by dates of 18.1 Ma on a dike that intrudes the Jackhammer and on a flow that unconformably overlies both the Jackhammer and Pickhandle formations (Burke et al., In press; see also Fig. 3 of this report; "Lane Mtn. Volcs."). This is consistent with the ca 19 Ma date obtained from the Opal Mountain Volcanic Member of the Pickhandle Formation in the Gravel Hills (Fig. 3).

Although Dibblee (1968) formally recognizes the Jackhammer Formation in the northwestern Calico Mountains, he does not extend the nomenclature to rocks of similar lithology and stratigraphic position in the southern part of that range (Dibblee, 1970). Dibblee (1968, p. 18) quotes from McCulloh (1952) to indicate that the Jackhammer Formation in its type locality in the northwestern Calico Mountains consists of about 50 m of basal beds of arkosic conglomerate and tuff, overlain by 60 m of black to reddish-black massive amygdaloidal olivine basalt, and 5 m of white thin bedded to massive tuff; the tuff in turn is overlain by a 115 m-thick sequence of green-gray, light gray, and white dacitic tuff, tuff breccia, and tuffaceous sandstone, along with white hornblende biotite dacite tuff and tuff breccia, and interbedded brown tuffaceous sandstone. McCulloh (1952) notes that the thickest and best exposures of the Jackhammer Formation occur in and around Jackhammer Gap (Fig. 2) between one mile east and one mile west of the Barstow-Camp Irwin Road. Throughout the Calico Mountains, however, the Jackhammer as mapped by McCulloh (1952, 1965) crops out only sporadically.

McCulloh (1952, p. 109) recognizes that the Jackhammer displays lateral variation both in lithology and thickness. Of the various jackhammer lithologic units present in the Calico Mountains, the basal arkosic conglomerate and the overlying olivine basalt are the most persistent (McCulloh, 1952, p. 113). Both the conglomerate and the basalt, however, pinch out within one kilometer west of the type locality. At the same time, a lens of coarse volcanic breccia from the northwest is interposed between the olivine basalt and the overlying white Tuff (McCulloh, 1952, p. 109). Locally, the limestone beds up to 1 m thick also occur near the base of the unit. According to McCulloh (1952, p. 109, 110) "presumably all of the members of the Jackhammer Formation wedge out in the same manner as the arkose and basalt members." From a thickness at the type locality of approximately 230 m, the Jackhammer locally pinches out, leaving the overlying Pickhandle Formation resting on the crystalline basement rocks.

Neither McCulloh (1952) nor Dibblee (1968) speculate on the mode of deposition of the Jackhammer Formation. Several interpretations may be inferred from the data presented, however. (1) The basal arkosic conglomerate contains well-rounded clasts of the basement terrane and is poorly stratified. This unit probably represents alluvial deposition on the flanks of the pre-existing crystalline basement. (2) Concomitant with alluvial deposition, the region experienced burst of volcanic activity as represented by the olivine basalt flow and the tuff units. (3) Deposition of limestone displaying algal laminations requires the presence locally of either intermittent small ponds and lakes, or intermittent incursion of arms of larger standing body of water. (4) Lastly, a landslide, possibly generated by seismic activity, probably was responsible for the extremely coarse-gained volcanic breccia lens mentined above (McCulloh, 1952, p. 109). Thus, rocks of the Jackhammer Formation in the Calico Mountains probably were deposited on the flanks of the developing Barstow basin, where alluvial fans interfingered with lacustrine sediments and where sedimentary deposition was interrupted by volcanic and tectonic activity.

The type locality of the Pickhandle Formation is in Pickhandle Pass (Fig. 2), in the northwestern Calico Mountains. The unit is about 1,300 m thick there; it conformably overlies the Jackhammer Formation, and is unconformably overlain by the Barstow Formation (Fig. 3). In the type area, the Pickhandle consists of about 150 m of basal beds of tuff, and tuff breccia, followed upward by 430 m of light pink to tan or green dacite tuff breccia, 100 m of brown quartz monzonite breccia, 70 m of light gray tuff, red sandstone, and red brown mudstone, 120 m of brown, massive quartz monzonite breccia, 30 m of light brown to pink hornblende andesite tuff breccia, and 40 m of dark green to brown granitic conglomerate.

The Pickhandle Formation in the Calico Mountains "crops out in a discontinuous northwesterly- to westerly-trending belt" that passes southeastward through the southwestern corner of the Lane Mountain quadrangle and into the northern and northwestern parts of the Yermo and Nebo 7.5-minute quadrangles (McCulloh, 1952, 1965). Dibblee (1970) does not employ the term Pickhandle Formation in the southern Calico Mountains for rocks assigned to that unit by McCulloh (1965).

The Pickhandle Formation in the Calico Mountains undergoes rapid lateral changes in both lithology and thickness (McCulloh, 1952, p. 119). For example, the formation thickens and becomes more pyroclastic as it is traced from the Mud Hills eastward to the type section in Pickhandle Pass. Within approximately 9 km, the formation increases in thickness by roughly 610 m, and the pyroclastic content increases by fifty percent (McCulloh, 1952, p. 121). In addition, the unit in the east contains lava flows that are not present in the western exposures.

In contrast to the sections in the Gravel Hills and Mud Hills, the Pickhandle in the Calico Mountains in intruded by numerous plugs, dikes, and sills which complicate the stratigraphic and the structural relationships. Despite these complexities, however, McCulloh (1952, p. 122) was able to correlate several units, including a biotite dacite mudflow breccia and an overlying dacite tuff breccia, for a distance of 14.5 km.

Both McCulloh (1952, p. 123) and Dibblee (1968, p. 20, 22) suggest that this formation reflects a violent interval of volcanic activity following the deposition of the Jackhammer Formation. In the Calico Mountains, the data indicate that the violent "eruptions were separated by intervals of relative quiescence, during which tuffaceous sandstone, conglomerate, and mudflow deposits accumulated" (McCulloh, 1952, p. 123). This pyroclastic deposition was interrupted also by flows of basalt and by landslide deposits of granitic megabreccias (ibid.).

The Barstow Formation is exposed in the northwestern, and more extensively in the southern and southeastern parts of the Calico Mountains. The unit reaches a thickness of about 1,000 m and consists of a folded sequence of beds in which lacustrine sandstone and shale predominate. Basal beds of limestone are locally present, and beds of granitic conglomerate occur at intervals throughout the sequence. Fossil mammals are rare in the Barstow Formation of the Calico Mountains, but Dibblee (1970, explanation sheets) notes the local presence of "Merychippus" intermontanus, an element of the Barstow Fauna in the upper part of the Barstow Formation of the Mud Hills. "Merychippus" carrizoensis occurs in strata exposed in Sunrise Canyon in the southern Calico Mountains. These fossils indicate that rocks as old as late Hemingfordian age are included in the Barstow Formation there.

According to the somewhat limited information, the Barstow Formation of the Calico Mountains apperas to reflect mostly the lacustrine depositional conditions inferred for much of the unit in the eastern and northern outcrops of the Mud Hills. The deposits in the Calico Mountains appear to have been formed in the interior of this part of the Central Mojave Basin.

Yermo Hills

These limited and stratigraphically unconfined deposits of the Barstow Formation crop out in small hills north of Interstate 15, about 6 km east of the Calico Mountains (Figs. 2, 3). The sequence is about 200 m thick, slightly to strongly tilted toward the northwest, and overelain by Quaternary alluvium. The base of the unit is not exposed (see Dibblee and Bassett, 1966a).

The sedimentary rocks consist generally of drab-colored fine-grained claystone, with minor beds of sandstone, tuff, and limestone. Fossil mammals of the Yermo Local Fauna (Fig. 3) are of late Hemingfordian and early Barstovian age. "Merychippus " carrizoensis occurs throughout the unit, but about 10 m below its stratigraphic top, taxa characteristic of the Green Hills Fauna of the Mud Hills are associated with "M." carrizoensis. These taxa include "Merychippus" stylodontus, Amphicyon cf. ingens, Merriamoceras, and Rakomeryx yermonensis.

The Barstow Formation of the Yermo Hills probably is equivalent in age to the oldest fossiliferous levels of the formation in the Mud Hills; a correlated position below the tuff in the Mud Hills dated at 16.3 $\pm$ 0.3 Ma seems likely. Taxa of the Green Hills Fauna are well represented in association with what is probably the latest occurrence of "M." carrizoensis preserved in this region (see Fig. 3).

The fine-grained nature of the clastic units exposed here, coupled with the presence of limestone beds and water-laid tuffs indicate that the Barstow Formation of the Yermo Hills was deposited in a largely lacustrine environment, comparable to the conditions recorded for much of the formation in the Calico Mountains and eastern Mud Hills.

Alvord Mountain

A relatively complex sequence of deposits of early and medial Miocene age is exposed on the eastern flank of a terrane of Mesozoic plutonic and pre-Cretaceous metamorphic rock in Alvord Mountain, located about 20 km northeast of the Yermo Hills (Fig. 2). The Tertiary sequence was studied by Byers (1960). Major reference to the fossil mammals from the Barstow Formation there is Lewis (1964; 1968a; 1968b). The rock units of early and medial Miocene age are: the Clews Fanglomerate, the Alvord Peak Basalt, the Spanish Canyon Formation, and the Barstow Formation (oldest to youngest).

The base of the sequence is represented by the Clews Fanglomerate, the type locality of which is Clews Ridge (Fig. 2). The Clews is about 200 m thick and unconformably overlies plutonic basement rock. The unit was deposited on a former erosional surface with about 300 m of relief, and pinches out westward beneath the Alvord Peak Basalt. This basalt is only of local occurrence, and pinches out to the east, where the Clews Fanglomerate is overlain with possible unconformity by the Spanish Canyon Formation (Byers, 1960, p. 15; Fig. 3 this paper).

Ninety per cent of the Clews Fanglomerate consists of red-brown fanglomerate with blocks of mafic plutonic rock up to 3 m wide; the red color of the unit is considered to have been derived from pre-Clews soils that were stripped off and contributed to the sediment. A lower unit consists of fine-grained tuffaceous sediment, arkosic sandstone, limestone, siltstone, chert and claystone that overlie granitic breccia. An upper unit consists of beds of sandstone and tuff that locally include pebbles of leucocratic quartz monzonite porphyry and banded argillite. The pebbles probably had a more distant, northwestern, source because they lack a red-colored mantle that presumably was washed off during transport, according to Byers (1960, p. 16).

Most of the unit appears to have been derived from tectonic events that affected the local basement complex. Clastic constituents are very similar to, and were derived from, pre-Tertiary rocks exposed in Alvord Mountain. According to Byers (1960, p. 17-18), "...gently northward dipping Clews Fanglomerate rests on diorite cut by a 7-foot vein of white quartz. Angular fragments of this same white quartz are strewn through the basal 10 feet of the fanglomerate for several hundred feet north of the quartz vein. The blocky, angular character of the debris is possibly due to local fault movements that brecciated the basement rock and prepared it for transport, for much of the coarser breccia is close to faults that appparently were active during deposition of the fanglomerate."

The age of the Clews Fanglomerate is unknown. It is the oldest unit of this district, and probably is of Miocene age. Two other rock units (Fig. 3) intervene stratigraphically between the Clews and the Barstow Formation which - as elsewhere - contains fossil mammals of Hemingfordian and Barstovian age. If the pyroclastic episodes recorded in the Spanish Canyon Formation were synchronous with those of the Calico Mountains (as proposed below), the Spanish Canyon Formation could be a correlative of the Pickhandle Formation, and all or part of the underlying sequence at Alvord Mountain, including the Clews Fanglomerate, could correlate with the Jackhammer Formation to the west.

The Alvord Peak Basalt, about 100-125 m thick, is located mainly west of Spanish Canyon (Fig. 2), and thins eastward. The Alvord Peak Basalt conformably overlies the Clews Fanglomerate, and dikes and pipes of basaltic rock locally cut the Clews. The basalt also overlies plutonic rocks and granodiorite prophyry dikes. This unit of multiple flows is overlain by the Spanish Canyon Formation; the contact is unveven but few, if any basalt particles occur in the overlying beds. This suggests that little, if any, erosion intervened between the extrusion of the basalt and the deposition of the overlying sediments. In places where the Spanish Canyon Formation is missing, the basalt is overlain by the Barstow Formation. Indeed, the basalt is overlapped to the northwest by progressively younger parts of the Barstow Formation, suggesting downwarping of the basin prior to or synchronous with the deposition of the Barstow sediments.

The basalt is mainly dark gray to grayish black, nonporphyritic rock, with subparallel andesine-laboradorite laths in hematite-charged glass. Rare iddingsite phenocrysts reveal the presence of reclict olivine. As indicated above, the basalt apparently was derived from pipes or vents near Alvord Mountain; it is of limited areal extend..

The age of the Alvord Peak Basalt is unknown. The development of a down-warped basin prior to the deposition of the Barstow Formation that locally overlies the basalt (during which time the Spanish Canyon may have been deposited as well), suggests that a distinct hiatus preceded the deposition of the Barstow Formation with its fossil mammals of Hemingfordian and Barstovian age. On the other hand, the in part (locally) apparently conformable contacts between the units involved suggests to Byers (1960) their close temporal association. Because of its local extent, the Alvord Peak Basalt probably has no regional significance in terms of the timing of basaltic volcanism in the other districts discussed in this report. The position shown for this unit on Fig. 3 is derived from its stratigraphic position below the Spanish Canyon Formation.

The Spanish Canyon Formation is about 100 m thick. It conformably overlies the Alvord Peak Basalt and the Clews Fanglomerate; it is locally conformably overlain by the Barstow Formation, but erosional unconformities elsewhere (Byers, 1960, p. 22) indicate the presence of a hiatus between the two units. The formation is of limited areal extent, and wedges out to the west. A lower unit is composed of lenticular beds of white and grey tuff and interbedded sandstone and granitic boulder conglomerate, mafic boulder conglomerate, rare basalt fragments, pink aplite granite and light-colored granodiorite clasts all probably locally derived. An upper unit is composed mainly of two flows of olivine basalt and interbedded sandstone and conglomerate.

According to Byers, (1960), the tuffs of the Spanish Canyon Formation record a major explosive volcanic event or events in the central Mojave Desert. Inasmusch as they are fine-grained and apparently wind-borne, the events from which they originated should have been located to the west of Alvord Mountain, if past winds prevailed from that direction as they do now. Microscopic study (Byers, 1960, p. 26) suggests a correlation with the pre-Barstow Formation volcanic sequence of the Calico Mountains where the Pickhandle Formation consists of dacitic tuff, tuff breccia and agglomerate. If this is accurate, the units that underlie the Spanish Canyon Formation could correlate with the Pickhandle Formation of the Calico Mountains-Mud Hills area, as suggested above.

The Barstow Formation of this district crops out in the eastern part of Alvord Mountain, and consitsts of "clastic and tuffaceous beds that are characterized by lithology and vertebrate fossils similar to or identical with those found in the Barstow Formation at the type locality [and] are assigned to [that formation]...(Byers, 1960, p. 26.). The lithic features characteristic of the Barstow Formation in the Mud Hills, Gravel Hills, Calico Mountains, and Yermo Hills, however, do differ in many ways from the rocks correlated to the Barstow Formation in Alvord Mountain, and the diffeneces are especially striking when these are compared to the primarily lacustrine deposits of the Yermo Hills, the geographically closest sequence. We do not belabor this point here, but emphasize that the Barstow Formation of Alvord Mountain probably pertains to a more marginal depositional facies than commonly encountered in the typical deposits in the Mud Hills, Calico Mountains, and Yermo Hills.

The sequence in Alvord Mountain is about 400 m thick and is divided into three members. It overlaps the Alvord Peak Basalt, is basically unconformable on the Spanish Canyon Formation as well as on older units; the upper contact is locally gradational and conformable with overlying granitic fanglomerate (Byers, 1960, p. 27), but in places there is a slight angular unconformity and local channeling.

The lower member of the Barstow Formation is about 250 m thick: 175-200 m of interbedded sandstone and pebble conglomerate, are overlain by 50 - 175 m of tuff, tuffaceous sandstone, siltstone, and volcanic pebble conglomerate. Conglomeratic beds in the lower part contain rectangular fragments of banded siliceous pre-Tertiary argillite, presumably derived from the Paradise Range, 19 km to the northwest. Eastward, these are replaced by pebbles of grayish red-purple lava, which increase in grain-size and frequency in that direction. Southeast of Clews Ridge (Fig. 2), the most common clasts are locally(?) derived dark schist, gnieissoid quartz monzonite, granitic pegmatite, and white quartz. Three flows of olivine basalt occur in the lower 60-100 m of the unit, adjacent to Clews Ridge. In other parts of the area, in upper Spanish Canyon, a single basalt flow occurs about 100 m above the base of the formation.

The tuffaceous middle member of the Barstow Formation here ranges in thickness from 7-30 m, and consists of two or three beds of white granular to lapilli tuff and intervening clastic sediments; variations in thickness depend largely on the amount of interbedded clastic material. Beds above, below, and within the tuffs are among the most fossiliferous parts of the Barstow Formation in Alvord Mountain. The tuff beds generally increase in thickness and grain size to the west, suggesting a source in that direction. An intrusion and flow of nonporphyritic basalt occurs west of Alvord Peak in the northern part of the mapped area, west of the main outcrops of the Barstow Formation. Where it forms a flow, the basalt rests on the tuffaceous member of the Barstow Formation, and although the field relations are not entirely clear, Byers (1960, p. 30) suggests that the basalt is coeval with the Barstow Formation. A grayish-blue basaltic ash is locally conformable on the tuffaceous member of the Barstow, which, with the proximity of the flow, suggests that the eruption of the tuffs was followed by the extrusion of the basalt, which would then pre-date some or all of the upper member of the formation.

K-Ar dates of 13.8 (DES 7017) and 14.0 (DES 7015) Ma have been obtained on tuff beds of the middle member of the Barstow Formation (Pers. commun. to M.O. Woodburne from D.E. Savage, 1970). These ages are comparable to those obtained from the Hemicyon Tuff and Lapilli Sandstone tuff of the upper part of the Barstow Formation of the Mud Hills (Fig. 3). Additional commments on the age of the Barstow Formation at Alvord Mountain are given below.

The upper member of the Barstow Formation is about 170 m thick as exposed northeast of Alvord Peak, where it is in local (?)gradational contact with the overlying granitic fanglomerate. Elsewhere, the upper member is much thinner, having been removed by erosion prior to the deposition of the fanglomerate. This suggests to us that a distinct hiatus separates the Barstow Formation and the granitic fanglomerate, even though the contact between them appears to be gradational (reworked) in some localities.

At the head of Spanish Canyon, the lower 30 m of the upper member consists of tuffaceous siltstone and sandstone, and is locally fossiliferous. About 5 km east of these sites, the equivalent stratigraphic position is represented by a cobble conglomerate of andesitic debris. Above the basal 30 m, the upper member consists of pebble conglomerate and interbedded salmon-colored sandstone. Most of the clastic consituents apparently were deposited on alluvial fans, with source areas to the northwest (grayish red-purple Tertiary andesite-latite, pre-Tertiary banded siliceous argillite, and pale-orange quartz monzonite.

From the above summary, the Barstow Formation at Alvord Mountain was largely locally derived; the lower member contains clastic constituents apparently derived from sources to the northwest as well as to the northeast; the tuffs of the middle member probably came mostly from the west, and the source of clasts in the upper member seems to be mostly in the northwest. We believe that most of the lower and upper members are comprised of alluvial fan, channel, and overbank deposits and that lower engery, possibly in part lacustrine, environments are recorded chiefly by sediments and tuff beds in the middle member. Neither of these units show the extensive drab colored beds of lacustrine claystone, mudstone, and limestone so typical of the Barstow Formation in the Yermo Hills, the Calico Mountains, and much of the Mud Hills. The Barstow Formation at Alvord Mountain appears to represent deposits that accumulated near the northern margin of their depositional basin; if linked lithically with the sediments of the districts listed above, the connecting strata must be now buried beneath Quaternary alluvial and lacustrine beds south and southwest Spanish Canyon.

As shown in Fig. 3, fossil mammals occur in all members of the Barstow Formation at Alvord Mountain, and are comparable to the succession found in the Mud Hills (see also Lewis, 1968b). "Merychippus" carrizoensis (= M. tehachapiensis listed in Lewis, 1968b) and Merychyus (Metoreodon) relictus have been recovered from USGS Loc. D321 (Fig. 3), found in the lower member of the formation, stratigraphically about 170 m below the middle member. As disucssed for the Mud Hills, these taxa are characteristic of faunas of late Hemingfordian age, regionally calibrated at about 17 Ma.

Remains of Brachycrus buwaldi have been obtained from a site about 50 m stratigraphiclly below the middle member, in the general area of USGS Loc. D319. This species was listed (Byers, 1960, p. 33; pl. 2; and Lewis, 1968b, p. C78) as having been obtained from Loc. D319, in the middle member of the formation, and in association with "Merychippus" cf. stylodontus. R.H. Tedford, who collected the specimen in question, attests to the stratigraphic separation of the two species, as given here. The level is shown as "D319" on Fig. 3.

Based on samples from USGS Locs. D319 and D301, both from the tuffaceous middle member of the Barstow Formation, this unit contains Brachypsalis cf. pachycephalus, "Merychippus" cf. stylodontus, Merycodus, and Protolabis barstowensis.

These data suggest that the strata of the lower member of the Barstow Formation in Alvord Mountain that contain "M." carrizoensis and M. (M). relictus correlate best with the top of the Owl Conglomerate Member of the Barstow Formation in the Mud Hills, where similar taxa of late Hemingfordian age occur.

Brachyrus buwaldi, from the upper part of the lower member in Alvord Mountain, is one of the characteristic elements of the Green Hills Fauna, from the middle part of the Barstow Formation in the Mud Hills that contains the tuff dated at 16.3 ± 0.3 Ma near its base and the Skyline Tuff at its top.

Based primarily on the presence of "Merychippus" cf. stylodontus from the tuffaceous middle member of the Barstow Formation at Alvord Mountain, a correlation to the Green Hills unit in the Mud Hills also would be suggested. On the other hand, the specimens from the Alvord deposits are larger and appear to be evolutionarily more advanced than specimens from the Green Hills Fauna in the Mud Hills. It seems likely that the fauna from the tuffaceous middle member of the Barstow Formation in Alvord Mountain correlates best with the Barstow Fauna of the Mud Hills, even though specimens referable to "M" cf. stylodontus are not represented in the upper part of the Mud Hills sequence, and specimens referable to "M." intermontanus and "M." sumani - characteristic of the Barstow Fauna in the Mud Hills - are not present in the Alvord sequence. The ca 14 Ma age of the tuffaeous middle member of the Barstow Formation in the Mud Hills supports the above interpretation.

Cady Mountains

Dibblee and Bassett (1966a, b) mapped the regional geology of the area, and indicate that the Tertiary succession here (Fig. 2) is largely arranged peripheral to a central core of pre-Tertiary plutonic rock and metamorphic rock of limited areal extent. Major outcrops of Miocene sedimentary rock occur on the northern and western flanks of the range, but toward the center, extensive units of volcanic rock (Tb, Tbb, Tfb, Ta, Tah, Tab, and Tt of Dibblee and Bassett, 1966a, b) are shown. As discussed below, these rocks probably are mostly older than the sedimentary succession, and pertain to a volcanic center that was active as early as 23 - 24 Ma ago. Nonmarine fluviatile, volcaniclastic, and pyroclastic depostion began locally about 23 Ma ago, and persisted until at least 17 Ma ago, prior to a major hiatus. Subsequently, only local sedimentary and volcanic sequences record episodes of later Tertiary and Quaternary geologic events. The Cady Mountains are traversed in the cental and again in the northern parts by two of the east-west trending faults (Cady and Manix faults, respectively; Fig. 2), that are relatively rare in the Mojave Desert (Fig. 1). Such faults are thought to have experienced left-lateral slip (Garfunkel, 1974). As mapped by Dibblee and Bassett (1966b) and Williamson (1980), however, the Cady Fault dies out in the east-central Cady Mountains. Similarly, the Manix Fault to the north, does not extend beyond the eastern margin of these mountains (Dibblee and Bassett, 1966b; Moseley, 1978). Also, in contrast to areas to the south or west, the Cady Mountains are not strongly disrupted by major northwest- trending faults such as those that cut the Calico Mountains and Mud Hills (Fig. 2). As in Alvord Mountain (Fig. 2), the northwesterly oriented structural grain, but perhaps not the westerly, of much of the western Mojave Desert seems to die out in the Cady Mountains. These patterns may be related to the observation that there has been little if any Recent seismic activity on faults that occur east of a line that passes extends southeastward from the Death Valley Fault Zone, past the Avawatz, Soda, Cady, and Bristol mountains (Fig. 1).

The following discussion begins with the Tertiary succession of the southwestern Cady Mountains, which contains the geologically oldest biochronologic information, and is followed by comments on successions exposed on the northern and eastern flanks of the range.

Southwestern Cady Mountains.--Woodburne et al. (1974) indicate that the base of the Tertiary succession in the southwestern Cady Mountains begins with andesite lahar and agglomerate with interbedded tuffaceous sediments that rest unconformably on pre-Tertiary basement rock. As currently mapped (mainly Ta, Tab, and Tt of Dibblee and Bassett, 1966 b), this succession is not named, but appears to be stratigraphically and in part lithologically comparable with pre-Hector units mapped by Wiliamson (1980) in the northeastern Cady Mountains.

Miocene strata of the Hector Formation of Woodburne et al. (1974) are about 500 m thick; they unconformably overlie the agglomerate units and are unconformably overlain by Quaternary alluvium. The Hector Formation can be divided into a largely tuffaceous and volcaniclastic lower sequence (below a marker tuff bed in Fig. 3) and an upper succession that, although of tuffaceous matrix, contains many fewer beds of tuff. A local flow of basalt is, interbedded with the lower sequence, and a tuff near the upper part of that sequence has been dated at about 21.6 Ma.

Based on facies changes (coarser-grained clastic constituents more prevalent to the east) and sparse paleocurrent data, the Hector Formation appears to reflect alluvial fan and locally lacustrine sedimentation. The alluvial material was derived from the topographically higher parts of the present range to the east.

Fossil mammals designated as the Black Butte Mine Local Fauna occur in the lower part of the succession, below the marker tuff (Fig. 3), and are of late Arikareean age. The Logan Mine Local Fauna, of latest Arikareean age, occurs in strata above the dated tuff. [The Black Butte Mine Local Fauna originally was designated as the Black Butte Local Fauna in Woodburne et al. (1974). The name is changed here to prevent confusion with a unit known by the same name in Oregon.]

The Black Butte Mine Local Fauna consists of remains of an oreodont, Merychyus calaminthus (stratigraphically restricted to strata below the dated tuff) and a camel, Stenomylus cf. hitchcocki, with a stratigraphic occurrence above and below the dated tuff, but below the marker tuff that separates the two local faunas. The oreodont also occurs in the Tick Canyon Local Fauna of the Ventura Basin in deposits that unconformably overlie the Vasquez Formation, with basal basalt flows dated at 25.6 $\pm$ 2.9 Ma and 24.6 $\pm$ 0.8 Ma (Crowell, 1973) and 20.8 $\pm$ 0.8 Ma (Woodburne, 1975). M. calaminthus also occurs in the Diligencia Formation of the Orocopia Mountains (Crowell, 1975; Woodburne and Whistler, 1973) stratigraphically above (but with imprecise Locality data) an apparently altered basalt dated at 23.1 $\pm$ 2.9 Ma and 20.7 $\pm$ 8.0 Ma (Crowell, 1973) and 19.2 $\pm$ 1.9 Ma (Spittler, 1974). In Nebraska, a closely related, and possibly conspecific species, M. crabilli, occurs in the Harrison Formation and Marsland Formation in Nebraska and Wyoming (Schultz and Falkenbach, 1947, p. 189). Stenomylus hitchcocki is otherwise known from the Harrison Formation; the type of this species comes from Stenomylus Quarry, associated with an ash dated at 21.9 Ma (Evernden et al., 1964). Taking all of the above into account, including the alteration of some of the basalts and precision of the stated ages, it appears that taxa of the Black Butte Mine Local Fauna existed in North America about 21 - 22 Ma ago. The pre-Hector terrane of volcanic lahars and agglomerate that unconformably underlies the fossiliferous sediments, should be older than that and a pro-rated age of 23 - 24 Ma for these older rocks is consistent with data developed elsewhere in the range, as discussed below.

The Logan Mine Local Fauna is composed of the camels, Protolabis, and Michenia cf. M. agatensis, an oreodont, Phenacocoelus cf. P. leptoscelos, the amphicyonid carnivore, cf. Daphoenodon, and the weasel, Promartes. All these taxa are presently considered to be of latest Arikareean rather than of early Hemingfordian age, as stated by Woodburne et al. (1974, p. 16). Recent biostratigraphic studies in western Nebraska are reflected here in the revised correlations of the Hector assemblages in terms of the North American Mammal Ages.

Some of the taxa from the Black Butte Mine Local Fauna are represented in the Hector Formation of the northern Cady Mountains, but interestingly, none of the Logan Mine Local Fauna taxa are, even though rocks of equivalent age are present there.

Northern and Eastern Cady Mountains.--The geology, paleontology, and radiochronology of the Hector Formation in the northern Cady Mountains have been described recently by Miller (1980), Moseley

(1978), and Williamson (1980). Williamson (1980) studied the dominantly extrusive volcanic succession adjacent to the eastern end of the Cady Fault, within the eastern part of the range (Fig. 2). The area studied by Miller (1980) begins at the northern base of the range (Fig. 2), about 2 km north of the rocks mapped by Williamson (1980), and contains a succession of sedimentary, extrusive, and pyroclastic rocks that together comprise the thickest single lithic, and fossiliferous sequence in this area. Moseley (1978) worked a few km farther north, adjacent to the eastward trace of the Manix Fault, in eastern Afton Canyon (Fig. 2), and investigated strata equivalent to the older part of the succession studied by Miller (1980).

Eastern Cady Mountains.--As mentioned above, Dibblee and Bassett (1966b) show that an extensive terrane of largely pre-sedimentary volcanic units (Tb, Tbb, Tfb, Ta, Tah, Tab, and Tt) are present in the northwestern part of the range and partially intervene between the two areas of major sediment accumulation on the southwest and north. Detailed studies of the northwestern volcanic terrane are lacking, but the area studied by Williamson (1980) in the eastern Cady Mountains seems to contain a similar suite of rocks (Tbb, Tfb, Ta, Tah, Tab, and Tfb of Dibblee and Bassett, 1966b). The units in the eastern Cady Mountains (and if correlated to them, those of the northwestern part of the range (Dibblee and Bassett (1966b), and those that unconformably underlie the Hector Formation to the southwest; Woodburne et al., 1974), probably represent part of a major volcanic center of ?Oligocene and Miocene age.

Williamson (1980) described a sequence of volcanic flows, flow breccia, domes, and lahars about 1100 m thick north and south of the eastern end of the Cady Fault, in the eastern Cady Mountains. The volcanic sequence unconformably overlies pre-Tertiary quartz diortie and quartz monzonite and is unconformably overlain by fanglomerate of Tertiary or Quaternary age.

The volcanic suite is designated as calc-alkaline in composition and ranges from basaltic andesite to rhyodacite. Most of the rocks are rhyodacites and dacites, however.

A thin (8 m), discontinuous unit of bedded arkose, probably transported only a short distance, occurs locally between the volcanic sequence and the basement rocks, and clearly has been derived from weathering of the underlying rock. This is stratigraphically similar to weathered debris reported at the base of the Tertiary section in Alvord Mountain by Byers (1960).

The stratigraphically lowest unit of the volcanic sequence of the eastern Cady Mountains is a sequence of monolithologic hornblende dacite breccia about 60 m thick. It is overlain stratigraphically by a unit of black augite andesite and interfingers with light gray dacite breccia. The augite andesite breccia is a flow of autobrecciated lava with a thickness of only 15 m, and of limited areal extent. It is overlain by an areally more widespread unit of light gray augite hornblende dacite breccia (75 m), capped by a thin flow of greenish gray to brownish gray hornblende biotite rhyodacite. This is followed upward by a relatively widespread unit of dark augite hypersthene dacite breccia, and locally, by a white dome of rhyodacite. The breccia is about 50 m thick. A black monolithologic breccia of augite hypersthene lahar about 25 m thick probably is the next youngest unit, but field relations are obscure. It is overlain by a unit of hypersthene augite andesite breccia (85 m) of limited areal extent, that is followed upward by bluish gray to brown laharic dacite breccia (105 m), a rather widespread unit of brown laharic breccia of andesite and dacite (225 m), a thin (15 m) breccia of black olivine basaltic andesite, and a unit of augite hypersthene rhyodacite domes and breccias, one of the most widespread sequences in the area.

This volcanic succession is unconformably overlain by the upper part of the Hector Formation as mapped by Miller (1980) which contains fossil mammals of the Upper Cady Mountains Local Fauna of late Hemingfordian age. This indicates that the volcanic terrane described by Williamson (1980) is older than at least the upper beds of the Hector Formation (see also below).

The range of volcanic units exposed in the eastern Cady Mountains, and the complexity of their field relations, suggest that they are part of a major, formerly active, volcanic terrane that may be similar to the volcanic units mapped to the northwest by Dibblee and Bassett (1966a). The northernmost units mapped by Williamson (1980) crop out within 2 km of the southernmost units of the Hector Formation mapped by Miller (1980) and, indeed, all of the units of Williamson (1980) can be found within 6 km of the southernmost exposures of the Hector Formation. Nevertheless, the sediments of the Hector Formation are not so strongly charged with clastic materials of the types cited by Williamson as to suggest that the two units were coeval. In fact, extensive volcanic terranes in this and other parts of the Cady Mountains appear to be on the whole older than the oldest sediments of the Hector Formation (Woodburne et al., 1974; Moseley, 1978; Miller, 1980). For the moment, the volcanic terrane described by Williamson (1980) is most plausibly considered to be largely if not completely pre-Hector Formation in age, and thus older than about 23 Ma (Fig. 3; included in column for Northern Cady Mountains).

The Cady Fault, one of the few high-angle east-trending faults in the Mojave Desert is about 19 km long, and contained totally within the boundaries of the Cady Mountains (Fig. 2). Dibblee and Bassett (1966a) and Williamson (1980) have shown that fault to reflect mainly dip-slip movement, with the south side up. The fault dies out in the eastern Cady Mountains, strain possibly having been absorbed along a zone of faulting that trends northwestward along the eastern margin of the range and extends into the area mapped by Miller (1980) immediately to the north. Garfunkel (1974) suggested that east-treding faults in the Mojave Desert, including the Cady Fault, should show left-lateral strike slip motion. The Cady Fault does not appear to fit this model.

Northern Cady Mountains.-- The sequence studied by Miller (1980) contains a succession of tuffaceous and volcaniclastic sedimentry rock, interbedded with a flow of olivine basalt and a distinctive ignimbrite unit. These deposits have been gently deformed into a shallow syncline that trends east-west, and plunges shallowly to the east. This has been cut by a number of faults that trend generally northward and show dip-slip throw of about 100 m at the most.

The two volcanic units are used to subdivide the Miocene sequence into three intervals. The lowest of these begins with an interval of tuffaceous sediment, tuff, and derivative clastic rock that is about 100 m thick. Most of the unit is comprised of beds of fine-grained sandstone, siltstone, and claystone, but toward the southwest, the frequency of conglomeratic sandstone and lenticular units of conglomerate increases. These beds contain pebbles, cobbles, and rare boulders of mostly volcanic rock apparently eroded from the terrane mapped by Williamson (1980), as well as minor amounts of granitic and metamorphic rock clasts. The relationships suggest that a basin-margin facies was present mostly to the south, whereas a more distal, lacustrine sequence prevailed to the north. Tuffs near the local base of the lacustrine succession have been dated at 22.9 $\pm$ 0.4 Ma (Miller, 1980). A basalt flow that overlies this interval has been dated at 18.6 Ma (Miller 1980).

The middle part of the Hector Formation in the northern Cady Mountains begins with the above-mentioned basalt, which is about 20 m thick, and continues upward stratigraphically for about 85 m to a distinctive ignimbrite dated at 17.9 $\pm$ 0.3 Ma (Miller, 1980). The rocks of this interval, which are generally similar to those discussed for the lower unit, comprise a succession of white to green tuffaceous sediments, with sparse units of conglomerate. Most of the sediments are relatively thin-bedded, and plausibly of lacustrine origin. These become distinctly coarser-grained, and contain an increased frequency of conglomeratic channel deposits as traced to the southwest. This, in conjunction with sparse but major units of strongly cross-bedded tuffaceous sandstone, indicate the dominance of fluviatile as opposed to lacustrine conditions in that direction.

The Lower Cady Mountains Local Fauna, of approximately late Arikareean to early Hemingfordian age, is represented by sparse but significant fossil mammals that occur in this part of the section, from about 30 m below the basalt to about 30 m above it. In the southwestern part of the area mapped by Miller, (1980) a number of faults repeat basalts that are apparently parts of the same flow as the unit dated at 18.6 $\pm$ 0.2 Ma. A nearly complete skull and jaws of Merychyus cf. M. calaminthus was recovered from a tuffaceous sandstone that lies 28 m stratigraphically below one of these fault-bounded basalts. Sparse remains of a fossil oreodont that probably are attributable to the same species also were found a few feet stratigraphically above another fault-bounded block of what appears to be the same basalt. As noted above, Merychyus cf. M. calaminthus is an important member of the Black Butte Mine Local Fauna, of late Arikareean age in the southwestern Cady Mountains, and elsewhere in California. In the present instance, we take the fact that this taxon and closely allied forms are found to stratigraphicallly bracket a basalt dated at 18.6 Ma to indicate that the occurrence of Merychyus cf. M. calaminthus in this part of the Hector Formation probably represents the upper part of the range of that species.

Sparse fossils occur in deposits that lie 16 - 33 m stratigraphically above the basalt, prominently exposed near the northern edge of the study area. These fossils, also attributed to the Lower Cady Mountain Local Fauna, consist of merychyine oreodonts, as well as aletomerycine cervoids that first occur in deposits of early Hemingfordian age in the Great Plains. Their presence in the Cady Mountains indicates that the Lower Cady Mountains Local Fauna extends into early Hemingfordian time.

Thus, the fossil mammals from sites about 30 m stratigraphically below the dated basalt appear to represent the same interval of time as do those that occur a comparable distance stratigraphically above that basalt. Taken together, the taxa identified as Merychyus cf. M. calaminthus, Merychyinae and Aletomerycinae, comprise the Lower Cady Mountains Local Fauna, of latest Arikareean to early Hemingfordian age.

The upper part of the Hector Formation in this area begins with a brown to brownish gray to orange-brown, and locally bluish gray, ignimbrite about 14 m thick. The unit displays varying degrees of welding, but usually shows conspicuous eutaxitic texture caused by compresson-induced flattening of pumice fragments during the welding process. Because of the generally low dip (to the northwest) of the beds and slight shuffling by faults, this ignimbrite occurs in only a few places in the main fossiliferous district, and it is not yet be possible to document a complete stratigraphcic thickness of the overlying units. The ignimbrite is important in that it forms a distinctive marker bed and beacuse it has been dated at 17.9 Ma. (Miller, 1980).

The upper unit that rests conformably upon the ignimbrite, is on the order of 130 m thick, and consists of tan to greenish gray and tan tuffaceous siltstone and mudstone, with local lenses of conglomerate. Beds of limestone have proven to be valuable marker units for mapping and biostratigraphic purposes. About 65 m stratigraphically aboved the ignimbrite (thickness not completely certain because of the presence of some faults that cut the section) occur the major fossil localities of the northern Cady Mountains (Upper Cady Mountain Local Fauna). These occur over a map distance of about 2 km, but within a stratigraphic range of about 100 m, and collectively yield a horse, "Merychippus" carrizoensis, the beaver, cf. Anchitheriomys?, a rhino, cf. Diceratherium, the dog, Tomarctus cf. hippophagus, a rodent, Proheteromys sulculus, the camels, "Miolabis" cf. tenuis, and cf. Aepycamelus, and the antelope, Merycodus. The assemblage as a whole suggests a late Hemingfordian age for the Upper Cady Mountains Local Fauna (Miller, 1980). The Upper Cady Mountains Local Fauna is the most diverse and one of the most important late Hemingfordian fossil mammal assemblages from the Mojave Desert Province.

Afton Canyon District.--Farther to the north, Moseley (1978) described the geology and stratigraphy of the area south of Afton Canyon that includes the eastern end of the trace of the Manix Fault (Fig. 2). Most of the rocks pertain to a 570 m thick sequence of the Hector Formation. The succession is comprised of gently to steeply dipping units that are deformed into a broad anticline in the east-centeral part of the mapped area. The fold is oriented northeastward, and plunges in that direction. A northern and eastern suite of faults are oriented approximately east-west, whereas a southern set is oriented mostly northward. The largest structural feature of the area, exclusive of the Manix Fault, is the fault that separates the northern from the southern set. It displays a vertical throw on the order of 125 m; other faults in the area display much lesser amounts of throw.

The Miocene sequence near Afton Canyon probably represents some of the oldest, if not the oldest, sediments of the Hector Formation. The Hector Formation unconformably overlies pre-Tertiary basement rock composed mostly of quartz monzonite, and is unconformably overlain by an unnamed sequence of conglomeratic rocks of Tertiary or Quaternary age, that is in turn unconformably overlain by coarse-grained basin margin facies of the Manix Lake beds of Pleistocene age.

The Hector Formation near Afton Canyon includes a sequence of alluvial and lacustrine stata interbedded with laharic breccia, and a sequence of volcanic rock, airfall tuff, a major ignimbrite, and flows of basaltic to andesitic lava. Based on clastic and textural features, the basin was bounded on the north by a terrane composed primarily of granitic rock, whereas a source area to the south included a terrane of active volcanism. The eastern and western margins of the basin are unknown, at present.

Beds of tuff and a number of andesitic and basaltic flows aid in differentiating the several stratigraphic units of the Hector sequence. A basal interval of four rock units is separated by an erosional unconformity from an overlying sequence of seven units. As discussed below, the lower quartet is unique to the Afton Canyon area, whereas the rock units above the unconformity are correlative with much of the section described 4 km to the south by Miller (1980).

The lowest portion of the Hector succession begins with an alluvial fan deposit composed of conglomerate and sandtsone. It unconformably overlies pre-Tertiary basement rock, and is overlain by, and interfingers with, the higher flows of a largely contemporaneous sequence of andesitic basalt. The alluvial unit is of limited areal extent, best exposed in and adjacent to a small horst in the vicinity of Afton Canyon. The unit reaches a thickness of at least 35 m, and is composed of beds of red boulder and cobble conglomerate composed largely of platy muscovite schist, and minor granitic, metavolcanic, and marble clasts; beds of limestone also are interstratified with the volcanic units.

These deposits interfinger upward for about 30 m with flows of andesitic basalt which range in thickness up to about 60 m. The andesitic basalt is areally widespread in the mapped area, and locally unconformably overlies pre-Tertiary basement rock. K-Ar dates of 14.2 ± 3.0 and 13.2 ± 3.0 Ma (E.H. McKee, U.S. Geological Survey, Menlo Park, pers. commun. to Moseley, 1978), were obtained from those andesitic basalts. As indicated below, these ages are clearly too young. Based on the regional correlations and stratigraphic relationshps discused here, the rocks should be at least 22 Ma old.

The andesitic basalt and conglomeratic sequence apparently is overlain by a 60 m thick sequence of light green to orange, buff, gray, and marroon andesitic laharic breccia, with a basal unit of poorly bedded red cobble conglomerate (5-10 m) and ostracode-bearing freshwater limestone (3-5 m). The breccia also is areally widespread, but generally is separated by faults from the units discussed above.

The upper unit of the sequence - below the minor erosional unconformity - consists of a single flow of hornblende pyroxene andesite about 60 m thick that is exposed mostly on the southern part of the mapped area. The unconformity within the Miocene sequence described by Moseley (1978) separates a terrane of generally andesitic extrusive volcanic rock - and associated sediments - from a superjacent succession of largely basaltic and tuffaceous rock and associated sediments. The comparable stratigraphic position of what appear to be geologically/chronologically similar units on Fig. 3 represents our interpretation that this unconformbity is regionally significant, and reflects a major geological change in the region. The stratigraphic record of this change may be most closely documented in the Afton Canyon District, but also is reasonaly well shown in the southwestern Cady Mountains.

The base of the next higher sequence - above the unconformity - is comprised of a 60 m thick succession of red conglomerate, green to red to light pink sandstone and siltstone, reddish brown mudstone of alluvial lacustrine origin, with minor beds of air-fall tuff. Conglomeratic units become coarser-grained northward and include an extensive debris flow. Most of the clasts are schistose metamorphic rock, but metavolcanic rocks and clasts of marble are locally abundant.

The basal sediments are overlain by an areally extensive flow of greenish black olivine basalt (50 m) with a well developed baked zone beneath it. A K-Ar date of 21.9 ± 0.3 Ma was obtained from this basalt by Armstrong and Higgins (1973), and an age of 22.1 ± 0.4 Ma is reported from the same unit by Moseley (1981; pers. commun. from E.H. McKee).

The basalt is followed upward by green to brown lacustrine tuffaceous sandstone, siltstone, and mudstone with interbedded fluviatile conglomerate (94 m). The conglomeratic units are composed mostly of granitic rock fragments with lesser amounts of marble, metaquartzite, and schist. Volcanic rocks are rare or absent. This conglomeratic unit is best developed in the northwest part of the mapped area, and thins and pinches out into finer-grained units to the southeast. Based on the K-Ar ages reported on units above and below, these sediments are correlative with the lower part of the Hector Formation mapped by Miller (1980) to the south. The ca 22 Ma old basalt described by Moseley (1978) apparently does not crop out farther to the south. In contrast to the sequence described by Miller (1980), as adjacent to the southern margin of the depositional basin, those mentioned here seem to have been adjacent to its northern margin.

A flow of greenish black olivine-augite basalt (107 m) is the next youngest unit in the sequence. It also has a well developed basal baked zone. The unit pinches out against flows of the basal andesitic basalt to the northwest, and is correlated by Moseley (1978) with the basalt mapped by Miller (1980) to the southwest and dated at 18.6 Ma.

The olivine basalt is overlain by a sequence of coarse-grained alluvial fan and fine-grained lacustrine deposits that crop out mostly in the central and western part of the mapped area. The sequence is about 70 m thick. The coarse-grained units occur chiefly immediately adjacent to Afton Canyon in the northwest part of the mapped area, and are composed of gray to greenish gray medium to coarse-grained sandstone and pebble to cobble conglomerate. Clasts are mostly fragments of plutonic igneous and metamorphic rocks. The finer-grained constituents of this unit occur mostly in the cental part of the mapped ara, and are composed of tuffaceous light green to gray and buff sandstone, siltstone, and mudstone, and local dark brown mudstone. This succession appears to be correlative with the middle part of the Hector Formation as mapped by Miller (1980). Facies relationships mentioned above suggests that the sediments were deposited near the northern margin of the basin, in contrast to those described by Miller (1980) as adjacent to its southern side.

A distinctive ignimbrite, that also is a prominent marker bed in the area to the southwest studied by Miller (1980; dated at 17.9 Ma), crops out in the northwestern to the north-central part of the area mapped by Moseley (1978). This is a pink to reddish-violet and bluish-gray welded tuff (although the lower few meters may be unwelded). and appears to represent the northern part of the unit mapped by Miller (1980).

The youngest unit of the Hector sequence in this district is composed of white to green lacustrine tuffaceous sandstone, siltstone, and claystone exposed in a few fault-bounded blocks in the northwestern part of the mapped area. The unit is only sparsely represented. The upper part of the Hector Formation as mapped by Miller (1980), appears to be mostly lacking near Afton Canyon, possibly due to erosion.

As mentioned above, the Manix Fault forms much of the northern boundary of the area mapped by Moseley (1978). This linear, steeply dipping feature strikes nearly east-west, and first-motion studies of the 1947 earthquake, which had a magnitude of 6.2 on the Richter scale, showed left-lateral slip on a northward dipping fault plane (Richter, 1947; Richter and Nordquist, 1951), with left-lateral displacements at the surface on the order of 5 - 7 cm (Buwalda and Richter, 1948). Moseley (1978) also suggested that about 10-15 m of dip slip took place on the fault in Quarternary time (during and subsequent to the deposition of Quaternary gravel beds, but prior to the deposition of local exposures of the Manix lake beds, of Pleistocene age).

SUMMARY

Geological events in nonmarine volcanic terranes produce stratigraphic sequences that are notoriously variable in character. In general, the districts surveyed are a case in point. We do believe, however, that the rock sequences of early and medial Miocene age discussed here display a certain unity in stratigraphic pattern. The evidence is far from perfect, and it is at least conceivable that several of the rock sequences mentioned accumulated in its own local depositional basin. We prefer, however, to treat these rocks as having acculumated in contiguous parts of a larger basin that occupied the central Mojave Desert, designated as the Central Mojave Basin. The following remarks summarize the geologic history of that basin and comment on the evidence as to its regional extent and bounaries.

To the degree that it can be considered as a single geo-topographical unit, the Central Mojave Basin experienced the extrusion and accumulation of an extensive and varied sequence of andesitic to rhyodacitic lahars and breccias in the eastern Cady Mountains, and several ash-flow sheets (Lane Mountain Quartz Latite) in the Calico Mountains ca 22 - 24 Ma ago, and local deposition of conglomeratic units and andesitic basalts in the Afton Canyon Area (the K-Ar "ages" of ca 14 Ma yielded by some of these units in Afton Canyon are thought to be spuriously young).

This primary episode of andesitic volcanism and tectonism was followed by what appears to have been a regionally significant change. Deposition of fluvio-lacustrine volcanic sediments, local air-fall tuffs, and minor but eventually regionally widespread basaltic volcanism is recorded in the lower part of the Hector Formation in the Cady Mountains from about 22 to 21 Ma ago.

By 21 to 19 Ma ago, fluvio-lacustrine deposition and tuffaceous volcanism continued in the Hector Formation of the Cady Mountains; a generally similar, but somewhat coarser-grained, suite of deposits is represented by the Clews Fanglomerate and the Spanish Canyon Formation of Alvord Mountain which, with the Alvord Peak Basalt, record basaltic volcanism, as well. This pattern is approximately replicated by the still coarser-grained (and locally tectonically triggered) deposits of the Jackhammer and Pickhandle formations in the Gravel Hills - Calico Mountains area. The tectonism culminated in intervals of violently explosive tuffaceous volcanism, also reflected in the Spanish Canyon Formation, to the east.

A local final phase of this explosive volcanism may be represented by the important and relatively widespread ignimbrite unit of the Hector Formation, in the northern Cady Mountains, and in Afton Canyon. Local basaltic extrusions continued here during the 19 to 17 Ma interval, as well as elsewhere (basalts in lower member of the Barstow Formation in Alvord Mountain and, slightly later, in the Gravel Hills).

About 17 Ma ago, largely lacustrine deposition is recorded in the Cady Mountains, in contrast to the more fluviatile deposits of the lower member of the Barstow Formation in Alvord Mountain and the Owl Conglomerate Member of the Barstow Formation in the Mud Hills.

A major unconformity generally is prevalent beneath the Barstow Formation, and may be reflected in the fact that Miocene units younger than about 18 Ma are almost absent in the Hector Formation of Afton Canyon. This interval also could correspond to the period of downwarping that in part preceded the deposition of the Barstow Formation in Alvord Mountain.

The interval from 16 to 12 Ma is not represented by deposits in the Cady Mountain-Afton Canyon area. Alvord Mountain sequences of this age show a dominantly fluviatile regime in contrast to the major lacustrine and tuffaceous sequence with marginal fluviatile units represented on the southwest and west (Yermo Hills to Gravel Hills.).

Based on the data summarized here, we believe that in early and medial Miocene time the Central Mojave Basin extended from the Gravel Hills on the west to the Cady Mountains on the east. The facies interpretations suggest that in most instances only the northern limits of the basin may be inferred. In the Gravel Hills, the younger parts of the Barstow Formation were deposited from the north as well as the west and south-southwest, and may reflect the location of the western terminus of the basin. A dual - southern and northern - origin can be inferred for the older, Owl Conglomerate Member, of the Barstow Formation of the Mud Hills, and proximity to the southern basin margin is indicated by overlying units of the Barstow Formation exposed on the southern limb of the syncline in that district. All older units in these districts and in the Calico Mountains to the east, however, reflect either very local deposition (Jackhammer Formation) or suggest only the position of the northern margin of the depositional basin (Pickhandle Formation).

At the eastern end of the Central Mojave basin, deposits of the Hector Formation, and the underlying extrusive volcanic units, must have pinched out ultimately against a margin that lay somewhere to the east of the exposures in the southwestern Cady Mountains and south of the units found in the northern and eastern part of the range. Evidence suggests that predominantly lacustrine conditions prevailed in the area between Afton Canyon and the northern base of the Cady Mountains; the units coarsen northward in the vicinity of Afton Canyon and suggest that the northern margin of the depositional basin was located relatively nearby.

The succession developed in Alvord Mountain, about 23 km to the northwest, was derived dominantly from local sources, with much of the materials having been shed off Alvord Mountain itself. The Barstow Formation of that district, however, appears to have been in part developed from source areas that lay farther to the northwest than Alvord Mountain, as well as from areas to the northeast. The basin margin here might have been continuous ultimately with that interpreted to have occurred north of Afton Canyon, although the deposits that relate to it in Alvord Mountain are distinctly younger than the Hector Formation in Afton Canyon.

The Barstow Formation sediments of Barstovian age in Alvord Mountain are clearly a much more marginal facies that the equivalent (and older rocks) of the Yermo Hills and much of the equivalent rocks in the Calico Mountains and the Mud Hills. The sequences of Barstovian age in those areas are considered to have accumulated much nearer the center of the depositional basin, resulting in substantial successions of tuffaceous lacustrine sediments which, in the Mud Hills, interfinger with units derived from the southern rather than northern basin margin. In the Mud Hills, probably in the Calico Mountains, and certainly in the Yermo Hills, the northern edge of the basin can be considered to have been well removed from sites of sediment acculumation in Barstovian time.

Except for the limited evidence, cited above, the southern limit of the Central Mojave Basin is not well known. Rock sequences in the northern Rodman Mountains, southwest of the Cady Mountains, and in the Newberry Mountains and in Daggett Ridge (in progressively more eastward districts; Dibblee, 1964, 1970, Dibblee and Bassett, 1966a ; Dokka and Glazner, 1982) seem broadly similar to the units summarized in this report, and conceivably could represent the southern component of Central Mojave Basin deposition during early and medial Miocene time.

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LITHO- AND BIOSTRATIGRAPHY OF THE BARSTOW FORMATION, MOJAVE DESERT, CALIFORNIA

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Abstract

The Barstow Formation, of approximately medial Miocene age, crops out discontinuously in a northwardly concave belt in the central western Mojave Desert, north of Barstow, California. The Gravel Hills form the northwestern end of this belt, the west Cronese Basin, the northeastern. The exposures of the stratotype, in the Mud Hills, or Barstow Syncline, represent a sequence of clastic and pyroclastic deposits that subsequently have been deformed into a generally synformal pattern, and cut by faults that trend northwest-southeast and generally display right-lateral separation. The long recognized abundance of fossils of the Barstow Formation of the Mud Hills resulted in it being chosen as the typical fauna of the Barstovian Land Mammal Age in North America. The excellent exposures, its abundant fossils, and the presence of numerous pyroclastic and air-fall volcanic units that are actually or potentially datable by radiometric means combine to make the Barstow Formation one of the most important rock units to consider when studying the Miocene geological and paleontological history of the Mojave Desert.

Introduction

The Barstow Formation crops out discontinuously in a northwardly concave belt about 83 km long in the westen Mojave Desert, north of Barstow, in southern California. The exposures extend from the Gravel Hills on the west to the West Cronese Basin on the east. Between these end points, the Barstow Formation occurs in the Mud Hills, Calico Mountains, Yermo Hills, and Alvord Mountain (Fig. 1).

In its type area, in the Mud Hills, or Barstow Syncline, north of Barstow, the formation is about 1,300 - 2,000 m thick and consists of a sequence of coarse- to finer-grained fluvial and lacustrine sediments interbedded with water-laid air-fall tuffs. The unit unconformably overlies the unfossiliferous Pickhandle Formation, of largely pyroclastic lithology, and is unconformably overlain by Plio/Pleistocene basalt and Quaternary alluvium.

Prominent tuff beds can be used to divide the formation into at least three superposed units, and the upper two of these are locally highly fossiliferous. The abundance of fossils resulted in the faunas of the Barstow Formation being chosen to typify the Barstovian Land Mammal Age in North America (Wood, et al., 1941). As such the unit is an important cornerstone of the biochronologic correlation network based on fossil mammals that is used by vertebratae paleontologists and nonmarine stratigraphers when assessing the age of continental deposits in North America. Barstow Formation strata are locally fossiliferous in the other districts mentioned above, but those of the Mud Hills are by far the richest.

Radiometric calibration of tuffaceous units indicates that the Barstow Formation is about 16 Ma old near its base, and about 13 Ma old near its stratigraphic top in the Mud Hills. Calibrations in other parts of its regional extent are contained within those ages.

The excellent exposures of the Barstow Formation in the Mud Hills and in adjacent districts permit the analysis of a wealth of geologic phenomena in a reasonably detailed chronologic framework. In addition to stratotypic significance for lithostratigraphic and biochronologic units in its type area, the regionally extended Barstow Formation deserves to be an important standard of reference for regional studies of the Tertiary geological history of the Mojave Desert, and adjacent regions.

The purpose of this report is to summarize the litho- and biostratigraphy of the Barstow Formation in its type area so as to set out the framework within which these deposits can be viewed and within which further studies may be placed.

History of Study

The term, Barstow [Series], was first applied to deposits of Quaternary age adjacent to the Mojave River between Barstow and Daggett, California (Hershey, 1902). Hershey (1902) also coined the term, Rosamond Series, for essentially all strata of known or presumed Tertiary age on and adjacent to the western part of the Mojave Desert. Hershey (1902) included deposits on the north side of the Mojave River between Barstow and Daggett in the Rosamond Series. Subsequent field work by parties from the University of California, Berkeley (Baker, 1911; Merriam, 1911, 1900, 1900, 1915, 1919) soon discovered the fossiliferous strata in the Mud Hills, north of Barstow. At first, Baker (1911) utilized the name, Rosamond Series, for the deposits of the Mud Hills; he divided them into five units, of which three are shown in Fig. 2, and noted that the uppermost of these (Fossiliferous Tuff Member) contained abundant fossils. Baker (1911, p. 344) also noted that fossils were present in the next lower unit (Resistant Breccia Member). Merriam (1915; 1919) discussed the possible ways in which the Rosamond Series might be related to the deposits in the Mud Hills that contained fossils pertaining to the Barstow 'Fauna;' he intended to use the term, Barstow beds or Barstow Formation, for those deposits that contained elements of the Barstow 'Fauna.' Merriam (1915) initially intended the term Barstow beds to be restricted to the Fossiliferous Tuff Member of Baker (1911), but later (Merriam,

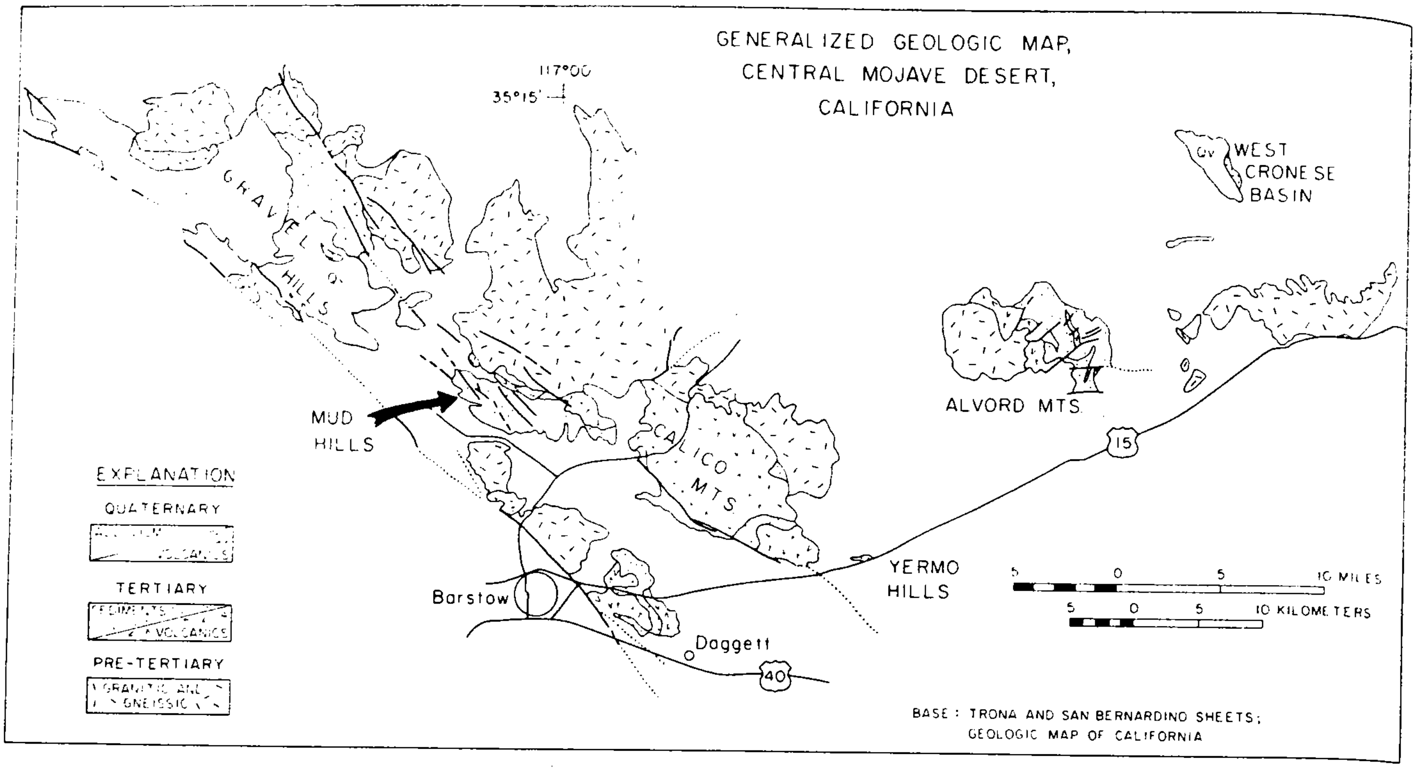


Figure 1. Index map of a portion of the Mojave Desert, California, showing districts and sites discussed in the text. The stratotype of the Barstow Formation is located in the Mud Hills, north of Barstow.

1919) indicated that the formational term could include units both laterally and vertically that were not necessarily restricted to that member. Later, Merriam (1919) recorded fossil mammasl from the Resistant Tuff Member of Baker (1911), and indicated that this unit could be included in the Barstow beds or Formation. Although procedurally incorrect (Hedberg, 1976), use of the term, Barstow Formation, for deposits of Tertiary age in the Mud Hills was stabilized in 1924 by the U.S. Geological Survey (see Lewis, 1964).

The studies of Merriam (1915, 1919) remain the basic references for the vertebrate paleontology of the Barstow Formation in the Mud Hills. Schultz and Falkenbach (1947, 1949) published studies on oreodonts that included specimens from the Barstow Formation. Frick (1926, 1937) presented a study of hemicyonine bears and horned ruminant artiodactyls, respectively, that similarly included Barstow taxa. These later authors published little of the stratigraphic data accompanying their material, which was contained in collections of the American Museum of Natural History, New York. Lewis (1964, 1968) made stratigraphically important, if taxonomically limited, additions, based on collections of the U.S. Geological Survey. Other, taxonomically restricted, studies that concerned specimens from the Barstow Formation include Hall (1930; rodents and lagomorphs), Stirton (1930 - an insectivore), Wood (1936 - rodents), Stock, (1900 - Dyseohyus fricki), Tedford and Alf (1962 - Megahippus mckennai). An informal preliminary summary of the biostratigraphy of the Barstow Formation was made by personnel of the American Museum of Natural History, to accompany a field trip of the Society of Vertebrate Paleontology in 1966.

Lindsay (1972) introduced an important biostratigraphic zonation of the upper part of the Barstow Formation based on rodents, but very little work that is both a general taxonomic as well as biostratigraphic synthesis has been published on the vertebrate fossils of the Barstow Formation since Merriam (1919). A manuscript that includes a brief, but stratigraphically and taxonomically modern, discussion of the biostratigraphy of the Barstow Formation has been prepared by R.H. Tedford and colleagues, and pertinent aspects of that work are included here.

HERSHEY (1902)		BAKER (1911)		MERRIAM (1915)		DIBBLEE (1968)		THIS PAPER
Rosamond Series (Part)	ROSAMOND SERIES	Fossiliferous Tuff Member	BARSTOW FORMATION	Fossiliferous Tuff Member	BARSTOW FORMATION	No Subdivisions	BARSTOW FORMATION	Upper
No Local Subdivisions		Resistant Breccia Member		Resistant Breccia Member				Middle
		Fine Ashy and Shaly Tuff Member				Owl Conglomerate Member		Lower
						Pickhandle Formation		

Figure 2. Summary of stratigraphic nomenclature of the Barstow Formation in the Mud Hills employed in past and present works.

The most recent studies of the geology of the Barstow Formation in the Mud Hills are Steinen (1966), Dibblee (1968) and, (through Dibblee, 1968), McCulloh (1954). Durrell (1953) however, published a discussion of the general geology of the Tertiary deposits and the mineralogy of strontianite-bearing beds in the vicintiy of Solomon Canyon, east of Owl Canyon, in the Mud Hills (Fig. 3). As noted by Dibblee (1968), this discussion pertains to a reasonably undeformed sequence of deposits, and the cross sections and measured section offered by Durrell (1953) include not only the stratotype of the Barstow Formation as redefined by Dibblee (1968), but also show its relationships to the underlying Pickhandle Formation. Sheppard and Gude (1969) discuss diagenesis of tuff beds in the Barstow Formation in the Mud Hills, and present a modern version of the nomenclature of units and beds that enable subdivision of the upper third of the Barstow Formation in that area. A more regional perspective, which includes comments on the Barstow Formation, is given by S.T. Miller (1980) in an open file report for the U.S. Geological Survey. Personnel of the Department of Earth Sciences, University of California, Riverside, have maintained a steady, if intermittent study of the stratigraphy and vertebrate paleontology of the Barstow Formation (as well as of the Mojave Desert in general) since 1959 and, since about 1920, geological and paleontological studies have been undertaken by personnel of the Frick Laboratory of the American Museum of Natural History. What follows below is presented as a reasonable synthesis of the results of those programs, incorporating elements of the works cited above.

Acknowledgements

We are deeply grateful to the late T. Galusha, American Museum of Natural History, New York, and S.T. Miller, U.S. Geological Survey, Menlo Park, for their help, consultation, and discussion during the present and past years on the general topic of Mojave Desert paleontology and stratigraphy, and with regard to the Barstow Formation, in particular. Morris F. Skinner and Beryl E. Taylor, American Museum of Natural History, generously shared the unpublished results of their studies of Barstow horses, camels, carnivores, and horned ruminants, and participated in disucssions of the biochronological significance of thee mammals. The work on the stratigraphy of the Barstow Formation in the Mud Hills by R.P. Steinen, Department of Geology, University of Connecticut, was of particular value in preparation of this summary. We also wish to thank D.E. Savage, University of California Museum of Paleontology, D.P. Whistler, Natural History Museum, Los Angeles County, and E.H. Lindsay, Department of Geosciences, University of Arizona, who introduced Woodburne to the Mojave Desert in the rainy winter of 1966, and who have continued their interest in, and contributions to the study of, this region in subsequent years. The paleontological work of R.M. Alf, Webb School, Claremont, and R.H. Reynolds, San Bernardino County Museum, also are to be recognized and appreciated here. Also to be cited but un-named are the many students of the Department of Earth Sciences who particpated in class field trips during the past 22 years and whose efforts contributed to our understanding of the stratigraphy of the Barstow Formation. We also wish to acknowlege grants from the University of California, Riveside, Academic Senate Committee on Research that defrayed some of the expenses associated with these studies.

Conventions

Stratigraphic and areal dimensions are given in metric notation. Taxonomic names that appear in " " indicate general use of a name that probably should be restricted to other species. All K-Ar dates cited have been recalculated when necessary to conform to the current IUGS decay constatns (Steiger and Jaeger, 1977).

Stratigraphy of the type Barstow Formation

General Statement.--As summarized in Steinen (1966) and Dibblee (1968), the Barstow Formation in the Mud Hills crops out as a sequence of generally steeply dipping beds that have been deformed into a syncline with antiformal features on the limbs. The unit is cut by several faults that usually trend northwestward and show right-lateral separation. The Barstow deposits are fluviatile and lacustrine sediments that reach a thickness of about 1,300 to 2,000 m. The unit unconformably overlies the Pickhandle Formation, and is unconformably overlain by Plio/Pleistocene Basalt and Quaternary alluvium. The numerous beds of water-laid air-fall tuff aid in subdividing the unit into members for more precise litho- and biostratigraphic analysis, and contribute to radiometric calibrations. The deposits vary widely in color, with hues of brown, green, orange, and red being common. The beds of tuff commonly are vitreous white in color, but may be yellow or brown.

Fossil mammals of the Barstow Formation are mostly of Barstovian age, although sparse remains near the base of the unit are considered to be of late Hemingfordian age. Based upon the radiometric dates from the Barstow rocks themselves, the unit ranges in age from about 16 Ma to about 13 Ma (Evernden et al., 1964; Lindsay, 1972; this paper). Based on assignments generally attributed to faunas of late Hemingfordian age in California (Evernden, et. al., 1964; Miller, 1980), the base of the Barstow Formation could approach an age of about 17 Ma.

Stratotype of the Barstow Formation.--Dibblee (1968, p. 26, 27) discusses the history of past useage of the term, Barstow Formation, and redefines the unit to be essentially equivalent to the upper two members of Baker (1911): "that sequence of deformed, stream-laid conglomerates, sandstones, lacustrine clays, and several thick tuffs, which lies unconformably above granitic breccia and tuff of the Pickhandle Formation, unconformably below flat-lying older alluvium of Pleistocene age, and which contains a mammalian fauna of late-middle and upper Miocene age" (Dibblee, 1968, p. 27). The type section is taken at the "most complete and unbroken sequence of the Barstow Formation [in] ... the south-dipping section at the east end of the Mud Hills, just west of Solomon Canyon" Dibblee (1968, p. 29). This section was measured by Durrell (1953), and is designated as the stratotype of the Barstow Formation. An abridged version of this measured section is given below. See also the road log, this volume, by Dokka and Glazer.

We utilize this stratotype section to demonstrate the physical continuity of the three-fold subdivision of the Barstow Formation employed in this report, a treatment which follows that of Lindsay (1972). In the present instance our work plus that of Steinen (1966) is utilized to reconcile questions or uncertainties in other studies.

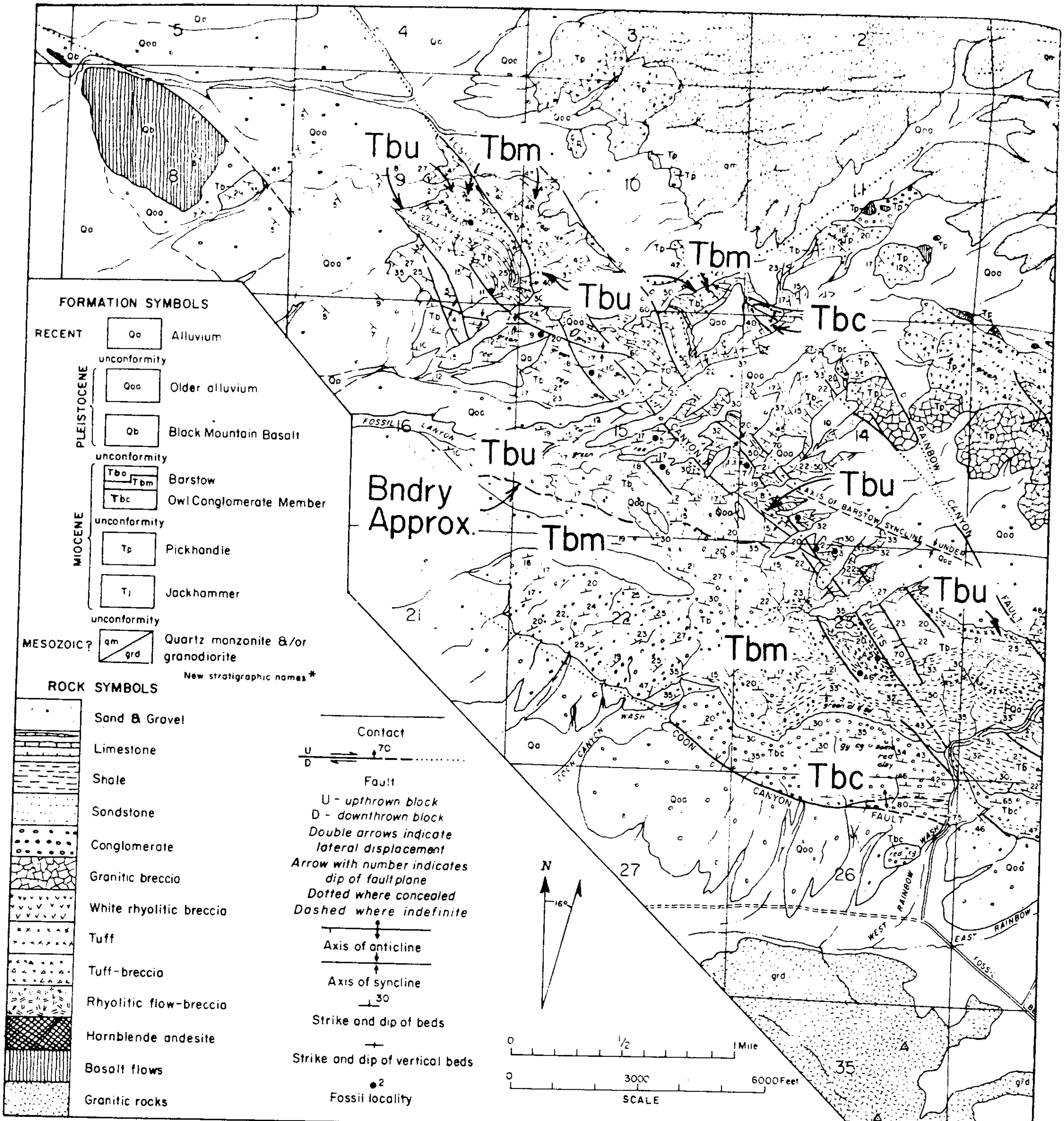
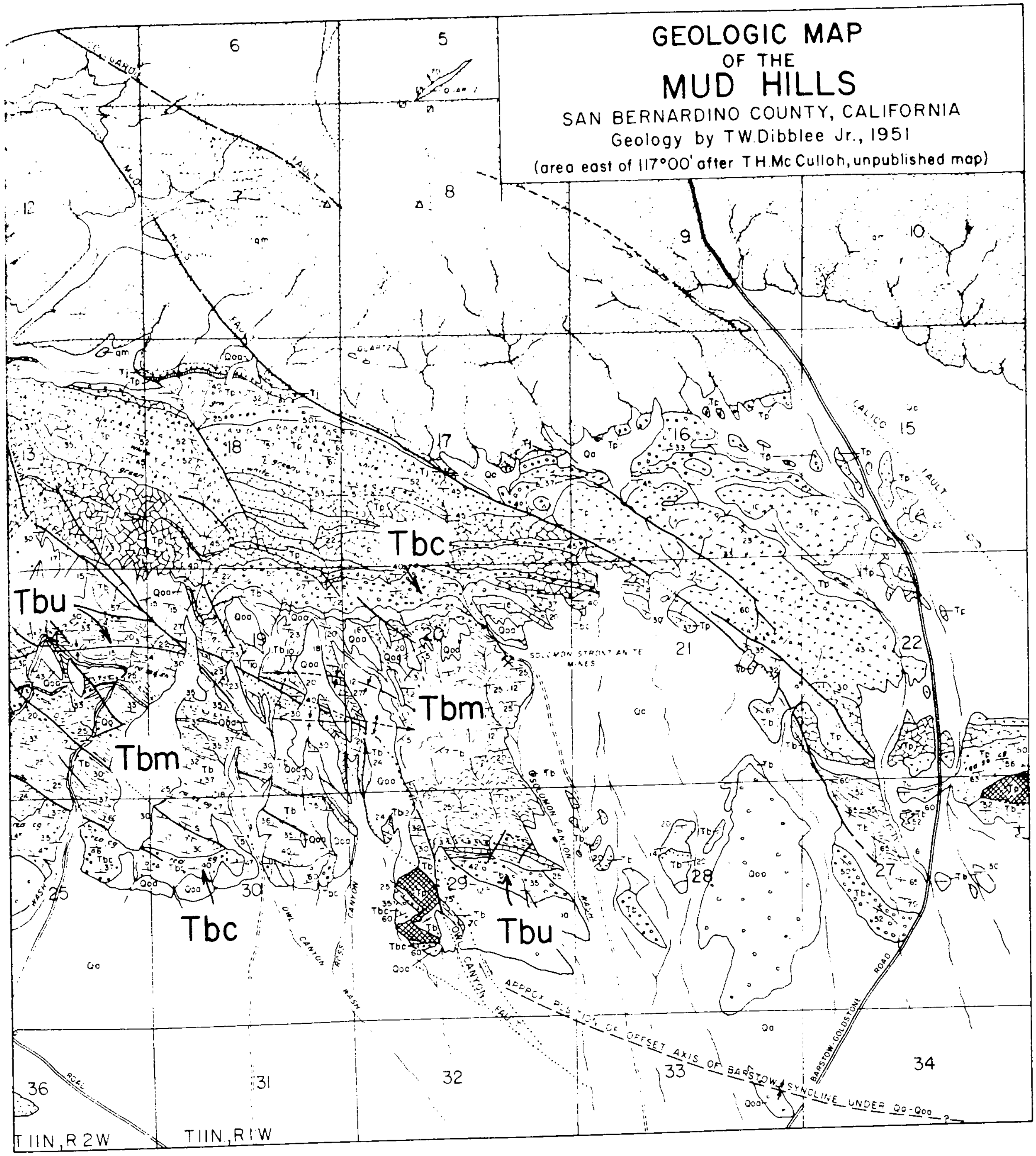


Figure 3. Geologic Map of the Barstow Formation in the Mud Hills, including the three-fold subdivision employed in the present work. The boundary between the middle and upper parts of the formation in the southwestern portion of the mapped area is approximate, as explained in the text. Modified from Dibblee (1968, pl. 3).


GEOLOGIC MAP
OF THE
MUD HILLS
SAN BERNARDINO COUNTY, CALIFORNIA
Geology by T.W. Dibblee Jr., 1951
(area east of 117°00' after T.H. McCulloh, unpublished map)
Tbc
Tbu
Tbm
Tbm
Tbc
Tbu
CALICO FAULT
SOLOMON STRONTIANITE MINES
SOLOMON CANYON
OWL CANYON
OWL CANYON WASH
BARSTOW-GOLDSTONE ROAD
APPROX. POSITION OF OFFSET AXIS OF BARSTOW SYNCLINE UNDER Qa-Qoa ?
T11N, R2W
T11N, R1W

Quaternary Alluvium

- Unconformity -

Barstow Formation

Upper Part of the Barstow Formation (subdivision employed for this paper, not in Dibblee [1968]).

Light gray brown to rusty yellow, poorly consolidated and sorted conglomerate and sandstone, of mostly granitic and volcanic composition (150 m), preceded by gray, algal limestone (10 m), light gray, fine- to medium- grained arkosic sandstone and minor limestone (10 m), and a basal unit of white, massive, fine-grained tuff (3 m). Although queried by Dibblee (1968), this is, in fact, the "lower marker tuff" of Dibblee (1968) and the Skyline Tuff of Sheppard and Gude (1969). It forms the base of the upper unit of the Barstow Formation as used here.

The total thickness of this unit in the stratotype section is about 173 m.

Middle Part of Barstow Formation

Sandstone; light gray, and green gray clay (66 m), overlies green claystone, with white limestone beds and nodules (88 m), brown, tuffaceous bentonite (28 m), brown arenaceous limestone (22 m), green claystone with white calcareous nodules (15 m), brown bentonite (10 m), green claystone and white nodular limestone (48 m), gray and green sandstone and green and buff claystone (112 m), yellow and greenish gray claystone, minor sandstone, limestone, and strontianite nodules, and tuff beds (40 m), green and gray arkosic sandstone (8 m), white and gray calcareous, rhyolitic tuff (0-1 m), light gray fine- and coarse-grained pebbly, tuffaceous sandstone (47 m), and a basal algal limestone (10 m). This is the base of the middle part of the Barstow Formation in the stratotype section. The unit is about 495 m thick.

Lower Part of Barstow Formation

Conglomerate (Owl Conglomerate Member of Dibblee [1968]); gray to green gray, crudely and locally cross-bedded, moderately to poorly consolidated, composed of subrounded cobbles and bouldes of quartz monzonite, some of aplite, pegmatite, and rarely of brown andesitic porphyry.

The lower unit of the Barstow Formation in its stratotype is about 46 m thick.

Total thickness of the Barstow Formation stratotype..................713 m

Subdivisions of the Barstow Formation.--As stated above, we use here a three-fold division of the Barstow Formation, as did Lindsay (1972). In an unpublished report, Steinen (1966) recognized six members of the Barstow Formation, which are generally consonant with the scheme followed here, but inasmuch as his work remains unpublished the names for his units cannot be utilized. Steinen's (1966) nomenclature is generally similar to that used by Sheppard and Gude (1969; Fig. 5) for tuffaceous units in the upper part of the Barstow Formation. We do use some of Steinen's data, however, in order to fill out our own information on the stratigraphy of the various intervals of the Barstow Formation, inasmuch as Dibblee (1968) focuses most details on the Owl Conglomerate Member, and treats the rest of the formation in a more cursory manner.

Lower part of the Barstow Formation - Owl Conglomerate Member.--As noted above in the summarized version of the type section of the Barstow Formation, as redefined, the stratigraphically lowest unit is the Owl Conglomerate Member (Dibblee, 1968). Fig. 3, which is based on Dibblee (1968, pl. 3), shows that the Owl Conglomerate Member crops out on both limbs of the Barstow Syncline in the Mud Hills.

The main outcrops of the Owl Conglomerate Member the north limb of the syncline (Fig. 3), occur between Solomon and Owl canyons. The unit unconformably overlies more steeply dipping granitic breccia and tuff of the Pickhandle Formation. To the west, the Owl Conglomerate Member pinches out between the overlying Barstow Formation and the subjacent Pickhandle deposits, as further evidence of the unconformable contact between the two formations. Minor outcrops occur to the west, in section 14, T. 11 N., R. 2 W., in the vicinity of the Rainbow Canyon Fault. On this side of the syncline, the Owl Conglomerate Member is composed of about 65 m of gray, subrounded to rounded clasts of quartz monzonite, aplite, pegmatite, quartz, and brown andesitic porphyry, in a matrix of arkosic sandstone; local finer-grained, occasionally tuffaceous, units are interbedded in the upper part of the unit.

Faults or alluvial cover obscure the base of the Barstow Formation on the south limb of the syncline, except locally, near Owl Canyon (Fig. 3). On this limb of the syncline, the Owl Conglomerate Member crops out from Ross Canyon on the east to Coon Canyon on the west, where it is cut out by the Coon Canyon Fault. The deposits consist of about 300 m of dull to vivid red-brown and maroon to pale brown coarse-grained sandstone and breccia, according to Steinen (1966) and Dibblee (1968). Color and bedding are variable laterally. The dominant clasts are mainly of granitic and volcanic rock, but some metamorphic rock clasts can be found. These beds are distinguished from the gradionally superjacent parts of the Barstow Formation by the the generally redder or more yellow and orange color of the beds of the former versus the more gray to yellowish brown color of the superjacent units (Steinen, 1966).

Steinen (1966) and Dibblee (1968) suggest that the deposits of this member were derived from the north or northeast as well as from the south and southeast. Source terranes from the north possibly include a unit known as the Paradise Mountain Quartz Monzonite that crops out a short distance to the north, and from the Jackhammer and/or Pickhandle formations of the Calico Mountains (volcanic rock constituents). Source terranes to the south apparently include Mesozoic granitic rocks, Tertiary volvanic rock, and ?Precambrian Waterman Gneiss.

Barstow Formation above the Owl Conglomerate.--Dibblee (1968) proposes no other members of the Barstow Formation, but shows (op. cit., pl. 1) that map units in the Mud Hills designated as Tbfg (fanglomerate of granitic detritus), Tbl (lacustrine clay shale), and Tbs (terrestrial sandstones) interfinger and that "the rest of the Barstow Formation above the Owl Conglomerate Member is composed of sedimentary rocks that range from coarse fanglomerate to clay shales, with no orderly sequence for its entire aerial extent...[and that] ...Several layers of white tuff are fairly persistent in the fine-grained clayey and intermediate sandy facies" (Dibblee, 1968, p. 29). In this report, however, we follow a three-fold division of the Barstow Formation of which the Owl Conglomerate Member is the lowest.

The Middle Part of the Barstow Formation.--This is taken as that part of the formation between the Owl Conglomerate Member and the "lower marker tuff" of Dibblee (1968), or the Skyline Tuff of Sheppard and Gude (1969). The middle part of the Barstow Formation, as used here, is nominally equivalent to the Resistant Breccia Member of Baker (1911). It reaches a thickness of about 700 m and generally is gradational with the Owl Conglomerate Member.

As mapped (Dibblee, 1968, pls. 1, 3) the middle part of the Barstow Formation of our terminology is composed largely of Tbfg (fanglomerate of granitic detritus) on the west, in contrast to consisting mostly of Tbs (continental sandstone) and Tl (lacustrine deposits) on the east. These relationships indicate that a thick unit of granitic fanglomerate is extensively developed in the western part of the southern limb of the Barstow Syncline. These deposits thin eastward, and interfinger with finer-grained deposits, composed of tuffaceous sandstone, siltstone, and lacustrine claystone. Steinen (1966) indicates that the fanglomeratic deposits consist of pale brown to yellowish brown coarse- to fine-grained deposits with many conglomeratic channels. Although composed of generally angular fragments of granitic rock, volcanic rock clasts also are conspicuous. Metamorphic clasts are rare. Scour and fill featurs coupled with pebble imbrications indicate that these sediments were deposited in a generally northward direction (in fact these units are not found on the north limb of the syncline), and eastward as well (where they are superseded by finer-grained deposits (Tbs, Tbl).

In the eastern part of the outcrop area (Rainbow Basin Loop east to Solomon Canyon), the deposits of the middle part of the Barstow Formation resemble those summarized in the type section (above, and Dibblee, 1968). These are comprised of abundant green, yellow, greenish gray, or brown claystone, with white nodules and lenses of limestone, beds of gray, locally green arkosic sandstone, yellow-brown conglomerate, brown bentonites, minor beds of white tuff, and a basal algal limestone.

A major tuff bed (which has been dated radiometrically - see below) occurs about 420 m above the base of this unit between West Rainbow Wash and Coon Canyon (sections 23, 24, T. T. 11 N., R. 2 W.; Fig. 3). In general the deposits above this tuff bed are coarser-grained than those below (Steinen, 1966), as approximately indicated by the presence of Tbs stratigraphically above Tbl in this area (Dibblee, 1968, pl. 1). Overall, the beds of the eastern outcrop area in the middle part of the Barstow Formation can be regarded as a distal basinal facies of the coarser, marginal conglomeratic unit of Tbfg.

The Upper Part of the Barstow Formation.--This unit, which crops out over much of the Mud Hills, but is most prevalent in the west, is defined as those strata that conformably include, and continue upward from, the "lower marker tuff" of Dibblee (1968), or the Skyline Tuff of Sheppard and Gude (1969). As such, these deposits are essentially equivalent to the Fossiliferous Tuff Member of Baker (1911). Steinen (1966) has utilized another tuff bed, the Hemicyon Tuff of Sheppard and Gude (1969), or "Hemicyon stratum" of Lewis (1968) to further subdivide this interval (see also Lindsay, 1972). For the present we do not propose formal lithic subdivisions of the upper part of the Barstow Formation, but will make use of marker beds within it to discuss relevant aspects of fossil distribution in the section dealing with Biostratigraphy.

As shown in the type section (above, and Dibblee, 1968), the upper part of the Barstow Formation is at least 173 m thick. Above the "lower marker tuff" or Skyline Tuff, the unit is composed - in ascending order - of light gray fine- to medium-bedded arkosic sandstone (10 m), gray, crudely laminated algal limestone (10 m), and light gray-brown to rusty yellow beds of conglomerate and sandstone, with subrounded clasts of granitic rock, andesite, and minor hornfels and quartzite (150 m). In most areas to the west of the type section, the basal unit of the upper member of the Barstow Formation can be recognized as containing the commonly thick (.3 - 1 m, or more) Skyline Tuff at its lower boundary. This is followed upward stratigraphically by about 10 - 30 m of beds commonly composed of dark brown mudstone, but including fine grained sandstone, siltstone, minor conglomerate and limestone (Steinen, 1966), capped by a thinner (3 - 6 cm) brown, biotite crystal tuff (Dated Tuff of Sheppard and Gude, 1969). Westward, this tuff grades locally into a biotite sandstone (Steinen, 1966). This persistent pair of tuffs provides a useful stratigraphic marker for mapping and biostratigraphic purposes.

In Rainbow Basin (area of the Rainbow Loop road), and in Owl Canyon to the east, about 100 m of deposits overlie the Dated Tuff. These deposits are mostly red brown to brown, but locally green to greenish gray beds of siltstone, claystone, and minor sandstone mapped by Dibbee (1968, pl. 1) as Tbl. Westward, in Coon Canyon, some stratigraphically equivalent rocks are coarser-grained beds of conglomerate and sandstone, but green-gray and brown claystone beds predominate. Beds in this area also grade into and interfinger with granitic conglomerate units (Steinen, 1966) mapped by Dibblee (1968, pl. 1 as) as Tbfg.

The Hemicyon Tuff ocurs about 100 m stratigraphically above the Skyline Tuff (Sheppard and Gude, 1969, fig. 5). This unit underlies a sequence of generally fine-grained beds of siltstone, tuffaceous sandstone, and limestone of usually greenish gray, but locally gray and brown color; its total thickness is about 130 m (Steinen, 1966). As was the case with the Skyline Tuff, another tuff occurs about 10 - 18 m stratigraphically above the Hemicyon Tuff, but in the latter case the upper tuff is white, vitric, and fine-grained in contrast to the dark brown, crystalline Dated Tuff associated with the Skyline Tuff (Steinen, 1966). Lindsay (1972, p. 4; Fig 2), notes the presence of a lapilli sandstone unit about 30 m stratigraphically above the Hemicyon Tuff on the north limb of the Barstow Syncline in section 15, T . 11 N., R. 2 W. This tuff, and the Hemicyon Tuff, have been dated by the K-Ar method, as discussed further below.

The generally fine-grained units are overlain in the northwestern part of the mapped area by a sequence of brown to yellowish brown coarse-grained sandstone which can be distinguished from the generally greenish gray beds below. Both units contain interbedded green tuffaceous siltstone and fine-grained sandstone beds. The upper unit is about 130 m thick and represents the stratigraphically highest sequence of the Barstow Formation in the Mud Hills. As such it is important from a biostratigraphic standpoint. These and the comparably coarse-grained but differently colored underlying units are mapped by Dibblee (1968, p. 1) as Tbs in section 9 and 16, T. 11 N., R. 2 W., and include the green beds that overlie red colored units shown in this area on Fig. 3 of Dibblee (1968) and of this report.

The above description of the deposits of the upper part of the Barstow Formation is taken mostly from sediments on the north limb of the Barstow Syncline. On the south, especially in the districts west of Coon Canyon, rocks that pertain to the same interval are mostly coarser-grained fluviatile, rather than basinal, lacustrine materials, and consequently both the lithic characteristics and the distinctive tuff beds are absent, the latter probably having been washed from topographically higher areas into the lacustrine basin. For this reason, is difficult to identify the boundary between the middle and upper parts of the Barstow Formation in this area. The dashed line shown between these parts of the formation on Fig. 3 is based on along-strike projections of what is believed to be the approximately equivalent stratigraphic boundary.

Based on the generally fine-grained lithology, the presence of numerous beds of tuff, some of which preserve mud cracks and tracks of mammals on their lower surfaces, a lacustrine origin is suggested for a majority of the deposits above the Skyline Tuff. The presence in areas to the west of increasing amounts of interbedded coarser-grained units suggests that the local basin margin lay in that direction (see also Dibblee, 1968, p. 32).

Geochronology of the Barstow Formation

Some of the tuff beds so abundantly exposed in the Barstow Formation have been utilized for radiometric age analysis. The results of these determinations have outlined the general age parameters of the unit. As background perspective, the following comments on the nature of the tuff beds is summarized from Sheppard and Gude (1969, p. 8). According to these authors, most of the tuffs in the Barstow Formation were deposited in lakes; most occur in the upper half of the formation; they range in thickness from .3 - 2.3 m, but most are less than .3 m thick. The thicker tuffs generally are more continuous laterally. Most contain vitric shards, and are even-bedded (beds commonly fining upward). Some tuffs (i.e., the Dated Tuff) show casts of mudcracks or animal tracks. A number of the thicker units show reworking of their upper portions. Rock fragments in the tuffs are chiefly volcanic and granitic materials, but may include quartzite, chert, and schistose and metamorphic constituents. Acccretionary lapilli - signifying that they were formed by accretion of moist ash in eruptive clouds, and then fell as mud-pellet rains (Moore and Peck, 1962), are common in the bases of the Skyline and Hemicyon tuffs.

Inasmuch as most of the tuff beds of the Barstow Formation appear to have been water-laid, many have been zeolitized and some show evidence of reworking in their upper portions, the question of origin of dated materials must be a factor to be considered when evaluating the accuracy of the radiometric ages.

The stratigraphically lowest tuff to be dated as yet is the massive unit near the middle of the middle part of the Barstow Formation, about 270 m below the Skyline Tuff. A K-Ar age of 16.3 $\pm$ .3 Ma was obtained on as a average of two determinations from this unit (R.H. Tedford, unpubl. data).

The Dated Tuff of Sheppard and Gude (1969; KA 449 of Evernden et al., 1964), that occurs about 17 m stratigraphically above the Skyline tuff in Rainbow Basin, yielded a K-Ar date of 15.5 Ma (Evernden et al., 1964, p. 176). R.H. Tedford (unpubl. data) initiated additional K-Ar analyses of the same unit, which gave an age of 14.8 Ma (average of two determinations). In view of the fact that the Dated Tuff preserves mud cracks and animal trackways in its base, and clearly is water-laid, and grades laterally into tuffaceous sandstone, the possibility must be entertained that some of the biotite crystals utilized in the K-Ar determinations were reworked from older rocks. For the present, we believe that 15 Ma is a reasonable age for the Dated Tuff.

R.H. Tedford also initiated K-Ar determinations of the Hemicyon Tuff that occurs about 100 m stratigraphically above the Skyline Tuff (Sheppard and Gude, 1969). This sample (average of two determinations) yielded an age of 13.4 $\pm$ .7 Ma.

Lindsay (1972, p. 8) reports K-Ar ages of 13.6 and 13.8 Ma on biotite and plagioclase, respectively, on minerals obtained from the Lapilli Sandstone. This unit occurs about 150 m stratigraphically above the Skyline Tuff and about 30 m stratigraphically above the Hemicyon Tuff (op. cit., Fig. 2). Inasmuch as both minerals yielded essentially the same age, the date seems reliable and is consistent with the age determined for the stratigraphically proximal Hemicyon Tuff.

The dated range of the Barstow Formation in the Mud Hills appears on present evidence to extend from about 16 to about 13 Ma. Inasmuch as about 600 m of section are present below the lowest date and about 200 m remain above the highest, the actual age range of the formation certainly is greater than its radiometric calibration. Evernden et al. (1964; KA 1368) indicate an age of about 12.7 Ma for beds in the West Cronese Basin district (Fig. 1) that correlate paleontologically with the upper part of the Barstow Formation in the Mud Hills. Skinner, Skinner, and Gooris (1977); p. 299, 362) disucss fossiliferous units in Nebraska that are of late Hemingfordian age and calibrated at somewhat older than 16.1 Ma, and Evernden et al. (1964; KA 478) show that faunas of similar age in California are younger than about 17.6 ma. The Owl Conglomerate Member of the Barstow Formation in the Mud Hills - which contains strata with fossils of late Hemingfordian age - is likely to be at least as old as 17 Ma.

Biostratigraphy of the Barstow Formation

Baker (1911) and Merriam (1919) noted the presence of locally abundant fossil mmammals in the Resistant Breccia and Fossiliferous tuff members of the Barstow Formation. These units are essentially equivalent to the middle and upper parts of the formation, as used here. Wood et al., (1941, p. 12) proposed the Barstovian Land Mammal Age, "based on the Barstow Formation...and specifically on the fossiliferous tuff member in the Barstow syncline and its fauna. Index fossils noted by Wood (et al., 1941) were Amblycastor, Dyseohyus fricki, Hemicyon, Monosaulax, Peridiomys. Taxa first appearing in this age were Aelurodon, Calippus, Hypolagus, Proboscidea, Prosthennops, and Teleoceras. Taxa making their last appearance in this age are Archaeohippus, and Parahippus. Characteristic forms include Alticamelus, Amphicyon, Aphelops, Blastomeryx, Cynodesmus, Hypohippus, Merychippus, Merycochoerus, Merycodus, Mylagaulus, Procamelus, and Ticholeptus. Taxonomic revisions have changed the name of some of these genera, and new forms have been added since 1941. Some of the taxa cited in 1941 or revised since that time, however, are not restricted to the Fossiliferous Tuff Member of Baker (1911), or to the upper part of the Barstow formation of this report.

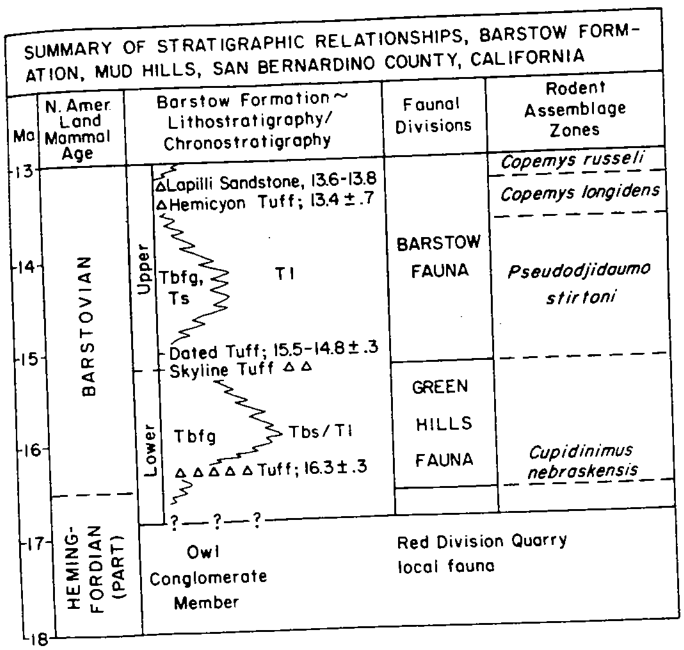


Figure 4. Generalized stratigraphic column of the Barstow Formation in the Mud Hills, showing the three-fold lithic subdivision of the unit, and the stratigraphic disposition of the major tuff beds, radiometric determinations, determinations, the Green Hills and Barstow Faunas, and the rodent zonations of Lindsay (1972). The ages derived from the Lapilli Sandstone (Lindsay, 1972) are equal to or slightly older than that from the Hemicyon Tuff, which is in fact about 30 m lower stratigraphically. All of these ages are nominally the same, within the limits of precision of this technique. Tbfg, Ts, and Tl follow Dibblee (1968, pl. 1).

Most of the fossil studies since 1941 concerned specimens from the upper part of the Barstow Formation, notable exceptions being Schultz and Falkenbach (1947, 1949), Frick (1926, 1937), Lewis (1968). Beginning in the 1930's personnel of the Frick Laboratory, American Museum of Natural History, began a series of paleontological studies first under the direction of J. Rak, then J. Wilson, and later (1950's), T. Galusha. A map generally similar to that of Dibblee (1968, pl. 1) was prepared by Galusha, in 1951, and biostratigraphic information obtained. A similar study was inititated in 1964 by R.H. Tedford, and R.P Steinen, then at the Department of Geological Sciences, University of California, Riverside. One aim of this project was to locate and integrate into a stratigraphic framework the localities made by various institutions during the past decades, so as to develop a detailed biostratigraphy for the Barstow Formation in the Mud Hills. This study is still in progress, but certain aspects of the biostratigraphy of the Barstow Formation have been cited in recent years in studies on rock units in other districts (e.g. Woodburne and Golz, 1967; Miller, 1980), as well as on the Barstow Formation, itself (Lindsay, 1972).

The present review of the biostratigraphy of the Barstow Formation is drawn from an unpublished manuscript by R.H. Tedford and colleagues, which integrates the previous work. We also add here the rodent assemblasge zonation proposed by Lindsay (1972).

As seen in Fig. 4, fossil mammal remains have ben recovered from three major intervals in the Barstow Formation of the Mud Hills. The oldest of these, from the Red Division Quarry of the Frick Laboratory occurs in the southwestern portion of the Mud Hills, in the unit mapped by Dibblee (1968, pl. 1) as Tbc, in the upper part of the Owl Conglomerate Member of the Barstow Barstow Formation, about 700 m below the Skyline Tuff, and 440 m below the tuff dated at 16.3 Ma. These sparse remains of fossil horse ("Merychippus" carrizoensis) and oreodont (Merychyus [Metoreodon]) are comparable to taxa found elsewhere in California and the Great Plains in faunas of late Hemingfordian age, locally calibrated at older than 16.1 Ma (Nebraska, Skinner, Skinner, and Gooris, 1977), and younger than 17.6 Ma in the Tehachapi Mountains (Evernden, et al,, 1964).

Fossils are more abundant in the upper half of the middle part of the Barstow Formation (Fig. 4). These taxa are grouped together under the name, Green Hills Fauna, and occurs in the rocks that extend downward to about 300 m below the Skyline Tuff. The tuff dated at 16.3 Ma occurs stratigraphically about 30 m above the base of this fossil-bearing unit. Taxa of the Green Hills Fauna that are important for correlation to other assemblages of early Barstovian age, including the Lower Snake Creek Fauna from the Olcott Formation, Nebraska, include Peridiomys (rodent), Euoplocyon and Amphicyon cf. ingens (carnivores), "Merychippus" stylodontus, (equid) Brachycrus buwaldi (oreodont), Rakomeryx, and Merriamoceros (cervoids), Hesperhys (peccary), Aepycamelus and protolabine camels. The Green Hills Fauna also contains the earliest North American occurrence of the bear, Hemicyon (Plithocyon), which can be used to define the beginning of the Barstovian Land Mammal Age.

Mammal remains also are abundant in the upper part of the Barstow Formation, just below the Skyline Tuff (Figs. 3, 4), and are found from that unit to the stratigraphically highest parts of the formation exposed in its western outcrops. In addition Alf (1970) has described remains of oak, palm, juniper, and poison oak from these deposits, indicating the presence of chapparal vegetation in the areas in which the mammals flourished.

The mammal fossils are gouped into the Barstow Fauna, to reflect the fact that this originally was (Baker, 1911; Merriam,, 1919) and still is the most fossilifeous part of the formation, and that the fossils from it were specifically cited by Wood et al. (1941) when naming the Barstovian Land Mammal Age. Taxa of the Barstow Fauna include "Merychippus" intermontanus, "M." sumani, Archaeohippus mourningi, and Megahippus mckennai (equids), Paramoceros (cervoid), Meryceros (antelope), and Mediochoerus mohavensis (oreodont), and the first North American appearances of Parapliosaccomys, Aelurodon and Gomphotherium. The presence of the mastodont, Gomphotherium, is especially significant, as this represents the lowest occurrence of the Proboscidea in the Barstow section, and by correlation, close to the first occurence of this important group of Old World immigrants in mid-lattitude North America. The first appearance of proboscideans forms a useful datum by which to define the beginning of late Barstovian time across North America. Evidence at Barstow and elsewhere fixes this datum at about 14.5 Ma.

We refer both the Green Hills and Barstow faunas to the Barstovian Land Mammal Age. Even though the taxa of the Barstow Fauna were specifically cited by Wood et al. (1941) as typical of the age, the genera listed by them also included forms found in, and characteristic of, the Green Hills Fauna. We exclude the sparse remains from the upper part of the Owl Conglomerate Member of the Barstow Formation from the Barstovian Land Mammal Age. They appear to be correlative with taxa typical of late Hemingfordian faunas elsewhere.

Fig. 4 also shows the distribution of the rodent assemblage zones nominated by Lindsay (1972) from the Barstow Formation. These were described from strata of the western part of the Barstow Formation so that thicknesses reported by Lindsay (1972) almost certainly are not strictly applicable thoughout the lateral extent of the formation. The oldest unit is the Cupidinimus nebraskensis Assemblage Zone. It occurs in the middle part of the Barstow Formation about 300 m below the Skyline Tuff, and extends upward to the stratigraphically next highest unit. One of the more important taxa of the Cupidinimus nebraskensis Assemblage Zone is Peridiomys sp., a genus cited in the original characterization of the Barstovian Land Mammal Age by Wood et al. (1941).

The next highest of these zones is the Pseudadjidaumo stirtoni Assemblage Zone. It contains taxa that range from the Skyline Tuff upward for about 80 m. The Dated Tuff (ca 15 Ma) occurs near the base of this zone.

The Copemys longidens Assemblage Zone occurs in an interval that begins about 80 m above the Skyline Tuff and extends stratigraphically upward to a level about 150 m above that tuff. This unit straddles the Hemicyon and Lapilli Sandstone tuffs, and is thus calibrated at about 13 Ma.

The Copemys russeli Assemblage Zone is the stratigraphically highest of these units. It follows the C. longidens Assemblage Zonme, and extends upward to a level about 250 m above the Skyline Tuff.

In terms of the biochronologic framework used here, the Cupidinimus nebraskensis Assemblage Zone is found in the interval that contains the Green Hills Fauna. The Pseudadjidaumo stirtoni, Copemys longidens, and Copemys russeli assemblage zones collectively are found in strata that represent about the lower half of the interval to which the Barstow Fauna pertains. At present, the rodent assemblage zones of Lindsay (1972) are largely descriptive; comparative studies in other sections of the Barstow Formation or its correlatives have yet to be accomplished; the chronological significance of the rodent assemblage zones of the Barstow Formation is yet to be determined.

Summary

The highly fossiliferous Barstow Formation, of medial Miocene age, crops out in the Mud Hills, north of Barstow, in the northwestern part of the Mojave Desert, California. The Barstow Formation is composed of a sequence of fluviatile and lacustrine sediments, and water-laid air-fall tuff beds; it is about 1,330 - 2,000 m thick, unconformably overlies the Pickhandle Formation, and is unconformably overlain by Plio/Pleistocene basalt and Quaternary alluvium. The sediments have been folded into a syncline and broken by several faults that generally trend northwest-southeast and show right-lateral separation.

In this report, the Barstow Formation has been divided into three parts. The lower unit, the Owl Conglomerate Member of Dibblee (1968) is about 300 m thick. It is exposed on both limbs of the syncline and is composed mainly of conglomerate and conglomeratic sandstone, gray in color on the north, of reddish hues on the south; clasts in the unit indicate deposition from the north and south sides of the basin. Although the actual limits of the basin are not known, present outcrops of potential source terranes suggest that it may have been about 8 km wide from north to south. Fossil mammals from near the top of the Owl Conglomerate Member are of late Hemingfordian age.

The middle part of the Barstow Formation is comprised of the fluviatile and lacustrine sediments that stratigraphically overlie the Owl Conglomerate Member and are overlain by the Skyline Tuff. The unit is about 700 m thick; it is gradational with the Owl Conglomerate Member, and is composed of a sequence of dominantly fluviatile conglomerate and conglomeratic sandstone beds on the west and southwest, in contrast to a sequence of dominantly lower-energy sandstone and claystone beds on the east. Facies distribution and sparse paleocurrent data suggest that the conglomeratic sequence was derived from sources to the south and southwest, and that the finer-grained beds were deposited in the more distal part of the basin to the east and northeast. A tuff about 420 m above the base of the unit has been dated at about 16.3 Ma. The upper 300 m of this sequence, beginning about at the position of the dated tuff, yielded fossil mammals of the Green Hills Fauna, of early Barstovian age. The Cupidinimus nebraskensis Assemblage Zone begins a few meters stratigraphically below that tuff.

The upper part of the Barstow Formation begins at the base of the Skyline Tuff, and extends to the top of the formation, a stratigraphic interval of about 400 m. In the eastern half of the outcrop area and on the north limb of the syncline in the western, the deposits are largely beds of fine-grained lacustrine shale, mudstone, and claystone, with lesser amounts of interbedded sandstone and conglomeratic sandstone. To the west and southwest the equivalent stratigraphic interval is represented by beds of sandstone and conglomeratic sandstone. This part of the Bartstow Formation also seems to have been deposited in a basin that was filled on the west and southwest by fluviatile deposits, whereas lacustrine sedimentation took place mainly on the north and east.

Numerous beds of tuff are present in the finer-grained units in the northern and eastern ourctops. These commonly are only a meter or less thick, but locally range up to 3 m in thickness; most are white, fine-grained vitric tuffs that have been deposited in bodies of standing water, some have been reworked in their upper portions, and many have been zeolitized.

A few of these tuff beds have been dated radiometrically. A unit known as the Dated Tuff, that occurs stratigraphically 10 - 30 m above the Skyline Tuff, has been dated at 15.5 and 14.8 Ma, or about 15 Ma. The Hemicyon Tuff, that lies stratigraphically about 100 m above the Skyline Tuff, yielded a radiometric age of 13.4 Ma, and the Lapilli Sandstone tuff which lies about 30 m above the Hemicyon Tuff yielded ages of 13.6 and 13.8 Ma.

Fossil mammals from the upper part of the Barstow Formation pertain to the Barstow Fauna, of late Barstovian age. Three rodent assemblage zones (Pseuadjidaumo stirtoni; Copemys longidens, and Copemys russeli, from lowest to youngest), have been described from the interval that begins just below the Skyline Tuff and extends upward for about 250 m above it.

The excellent exposures of the Barstow Formation in the Mud Hills, and the wealth of geologic, paleontologic, and radiochronologic data contained within them, combine to make this unit important not only for the development of further work, but also as a major reference for historical geological and paleontological studies in correlative strata in the Mojave Desert and elsewhere in southern California.

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PLEISTOCENE DRAINAGE AND DISPLACED SHORELINES AROUND MANIX LAKE

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INTRODUCTION

Late Pleistocene drainage in the Mojave Desert was dominated by the Mojave River. The Pleistocene Mojave River, like the current Mojave, originated in the San Bernardino Mountains and flowed north and east past Victorville and Barstow (fig. 1). During the late Pleistocene it filled Manix Lake and then spilled into Soda basin, forming a lake that included present-day Soda and Silver Lakes (see Figure 1). Water from this lake spilled out to the north where it joined the Amargosa River, which drained into Death Valley's Pleistocene Lake Manly. The similarity between the pupfish (Cyprinodon) living in this system and the Colorado River has led to the speculation that these two systems were once connected (Hubbs and Miller, 1948; Turner, 1974). Investigations into the possible Pleistocene link between the Colorado and Mojave Rivers have yielded no positive evidence that suggests the link actually existed. However, discovery of a high shoreline in southern Soda Basin (fig. 1) that was probably rapidly uplifted during the latest Pleistocene and Holocene by the Manix fault, makes it impossible to dismiss a Mojave-Colorado link until the evolution of the topography in the central Mojave is better understood. Recognition of the deformation responsible for the high shoreline also provides insights into the history of interaction of Manix Lake and Afton Canyon. Uplift of the sill between Manix and Soda Lakes caused Manix Lake to deepen until it abandoned its previous sill and overflowed along the Manix fault. Downcutting into the fault zone formed Afton Canyon and drained Manix Lake.

Appropriately, studies that utilize the Pleistocene Mojave's record to establish level lines within basins and time control between basins offer the best hope for extrapolating the topography into the past. Integrating the development of the Mojave drainage with the active structure in the area yields valuable insights into both the history of the Mojave River and the neotectonics of the central Mojave Desert.

BACKGROUND

Cenozoic Drainage

Little is known about the Cenozoic drainage until the late Pleistocene because the Mojave block has generally been a topographically high region subject to erosion throughout the Cenozoic (Hewett, 1954; Dibblee, 1968). Nothing is known about the early Tertiary except that previously adjacent provinces received sediments from Mojave sources (Hewett, 1954). This suggests that the Mojave was already a high region or the stratigraphic record was later removed (Dibblee, 1968). During the early and middle Miocene the Mojave and Sonoran Deserts were shedding sediments to the west toward the Pacific Ocean (Dibblee, 1968; Woodburne and Golz, 1972; Ehlig and others, 1972). By the late Miocene this east-west flow was disrupted by the formation of northwest-trending basins and an internal, basinal drainage system developed (Hewett, 1954; Bassett and Kupfer, 1964; Dibblee, 1968). However, these early basins bear no resemblance to the current set that began to form in the Pliocene. A major unconformity separates the Miocene and the Pliocene throughout the Mojave region (Dibblee, 1968). Westward drainage of the Mojave River in the Pliocene is suggested by the following: 1) sedimentary rocks deposited on the edge of the Mojave were derived from within the block (Foster, 1980); and 2) these rocks of the west coast contain fossils of the fish, Fundulus, an inhabitant of Death Valley during the Pliocene (Miller, 1950). Empetrichthys, a descendant of Fundulus, still lives in the Death Valley drainage (Miller, 1950). During the late Pliocene, accelerated basin formation and the uplift of the Sierra Nevada and the Transverse Ranges began to give the Mojave Desert its modern form (Hewett, 1954; Sharp, 1954; Dibblee, 1968).

Pleistocene Drainage

The drainage in the Mojave during the Pleistocene can be divided into isolated basins and basins forming a part of an integrated system. There were three integrated systems: 1) Mojave River; 2) Owens River; and 3) Amargosa River (fig. 1). All of these rivers delivered water to Death Valley during the late Pleistocene. The Owens and Amargosa Rivers barely enter the Mojave proper and will not be discussed in detail here. Lakes along these rivers have been studied in much greater detail than those along the Mojave River (e.g., Hooke, 1972; G.I. Smith, 1978; R.S.U. Smith, 1978).

There has been a great deal of speculation about another possible drainage that may have flowed out of Manix or Soda Lake past Ludlow and cascaded down through Bristol, Cadiz and Danby Lakes into the Colorado River (see fig. 1). This proposal is based on the string of basins that extends from Barstow to the Colorado River. This relation was first recognized and named the "great trough" by Thompson (1929, p. 625). The speculation that this trough was carved by a river was based on the similarity of pupfish (Cyprinodon) in the Colorado and Mojave River systems (Hubbs and Miller, 1948) and the bias of

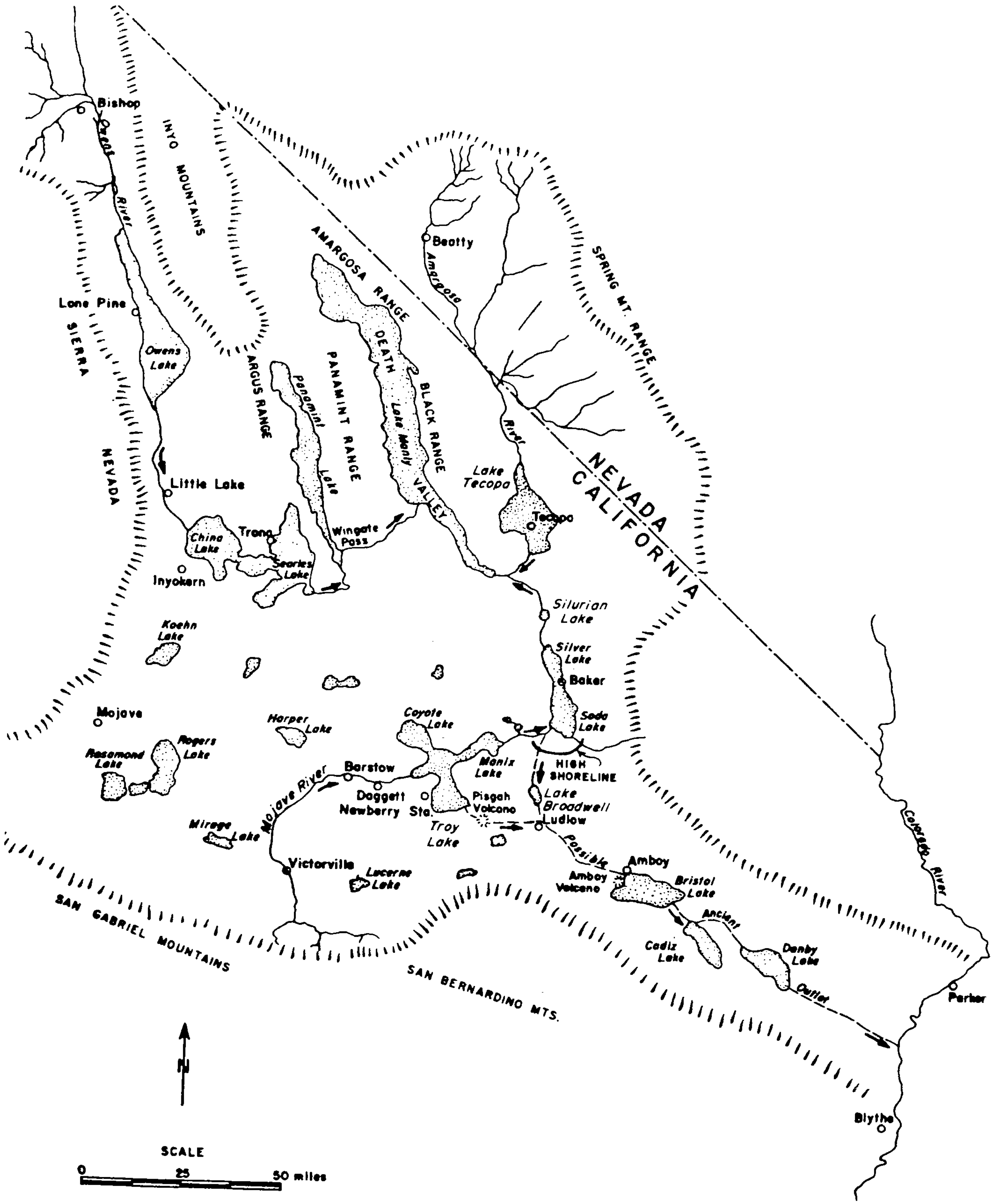

Figure 1. Pleistocene drainage in the Mojave Desert (modified from Blackwelder, 1954).

eastern-trained geologists that troughs must be carved by rivers. This "great trough" is still poorly understood but it is a structural and not an erosional feature. Even if the Mojave did not actually carve this trough, it is, given the current topography, the only possible link to the Colorado and the Mojave may have followed it. A quick examination of the topography of the central Mojave Desert reveals that the critical pass is about 5 miles east of Ludlow. If the Mojave River could have topped this 1900-foot pass it would have easily made it to the Colorado. Since Manix Lake has shorelines as high as 1800 feet, topping this pass does not seem, on first examination, too difficult. However, this pass is actually between Lake Broadwell and Bristol Lake, so Broadwell would also have to be filled up to 1900 feet, which would require filling Soda Lake and all the other lakes downhill as far as Death Valley. Thus, before water could spill over the pass into Bristol Lake a huge lake occupying much of the Mojave Desert would have to be formed. Evidence for such a lake has not been recognized.

Another problem with the proposed Mojave-Colorado link is the lack of an integrated system south of Ludlow. Lakes that are part of a river system in the Mojave typically have well-developed shoreline features around the basin at the level of a recognizable sill or spillway into the next basin in the system. This is seen in the Manix and Soda Lakes on the Mojave's route and in the basins along the Owens River route to Death Valley. In isolated basins, where water never reaches a sill, the water level fluctuates rapidly, depending on runoff and evaporation. Distinct, well-formed shoreline features and associated deposits do not develop to the degree that they do when a sill forces the water level to remain constant for long periods of time. Lakes in the "great trough" to the Colorado appear to be isolated systems; though if the link was old enough the traces could have been eliminated.

The inference that the Mojave-Colorado link did not exist depends critically on the assumption that the topography has not changed drastically during the late Pleistocene. While it is clear that the topography must be changing in the central Mojave, we do not know how fast or even in what manner. If we can decipher the changes in the topography, perhaps the link between the Mojave and Colorado will be better understood. Alternatively, if the link can be found or proven, inferences about the nature and rate of change necessary to allow the link will help us understand the active tectonics in the central Mojave.

SODA BASIN

High shoreline features were found at the south end of Soda basin (fig. 1) while investigating the link between the Mojave and Colorado rivers in the central Mojave. Wave cut benches, tufa deposits, gravel and sand bars and lacustrine sediments ring the southern end of Soda basin at approxmately 1100 feet above sea level. The existence of the 1100-foot shoreline south of Soda Lake suggests that there was a much larger and deeper lake at some point in the past or that the shoreline has been relatively uplifted from the 946-foot shoreline associated with the 946-foot sill at the north end of the basin. Shorelines at 946 feet are best expressed near Baker and Silver Lake (fig. 1). The first possibility would require an extremely large lake stretching from Afton Canyon to Death Valley. However, no evidence has been found to support the presence of a lake above the 946-foot level north of the southern end of Soda basin. Considering how well the 1100-foot shoreline features are preserved, it seems unlikely that all traces could have been removed from the north end of the basin.

In an attempt to prove that the 946 foot shoreline of northern Soda basin was the same age as the 1100-foot level of the south end of the basin, radiocarbon dating was performed on Anodonta shells from a sand bar on the 946-foot shoreline and on tufa from the 1100-foot shoreline. The Anodonta shells yielded an age of 11,970 ± 160 C-14 years BP and the tufa yielded an age of 5,900 ± 100 C-14 years BP (Beta Analytic, written comm., 1981). The Anodonta age is consistent with the end of the last pluvial, but the tufa age is obviously spurious. Unfortunately, no other datable material has been found associated with the 1100-foot shoreline. Whereas the geologic evidence strongly suggests that the 1100 level has been uplifted from the 946 level, the 1100-foot level need not be the same age as the Anodonta shells. Since the only shoreline in the north end of the basin corresponds with the sill level, it is possible that several pluvial lakes of different ages are superimposed at the same level and the shells simply date the end of the last one.

Because the 946-foot level extends about 19 miles from Silver Lake to Fort Mojave (now called Zzyzx) without showing any suggestions of deformation and the 1100 foot shoreline is only 8 more miles to the south, faulting seems more likely than regional warping or tilting to explain their separation. The most likely candidate is the Manix fault, which strikes N.70E. from near Calico through Manix Lake to Afton Canyon where it disappears beneath the Mojave River's wash and Soda playa. A continuation of this fault to the east would separate the two levels in Soda Basin. Although there are some suggestive air-photo lineaments, sedimentary processes in the Mojave wash and Soda playa are probably too active to reveal the fault east of Afton Canyon.

The Manix fault is a left slip fault where it cuts through Manix Lake (Buwalda and Richter, 1948; Keaton and Keaton, 1977). In the Afton Canyon area the fault appears to have a component of dip-slip (south-side-up) motion based on geologic relationships (Keaton and Keaton, 1977) and geomorphology. However, Dibblee and Bassett (1966) reported vertical separation of a north-side-up sense at a locality about 15 miles to the west. Movement associated with the April 10, 1947 earthquake on the Manix fault was pure left-slip (Buwalda and Richter, 1948). Surface rupture only occurred for 2-3 miles about 15 miles west of Afton Canyon (Buwalda and Richter, 1948).

If the Manix Fault has lifted the 1100-foot shoreline in Soda Basin, rates can be calculated if we assume that the shorelines formed during one of the last pluvial maxima during the late Pleistocene. This assumption appears to be reasonable from the preservation of the features and the presence of what appears to be Indian artifacts associated with the 1100-foot level

(Indian artifacts are commonly associated with the 946 level in north Soda Basin). Rates between 1 and 4 mm/yr are calculated, depending on which of the late Pleistocene maxima are chosen. This is extremely high, considering that the Manix fault displays dominantly left-slip to the west. This may suggest that the shorelines are older than suspected.

This rapid, south-side-up component can be used to explain certain inconsistencies in the history of Manix Lake. The youngest Pleistocene shoreline (Tioga?) in Manix Basin is at 1800 feet, whereas the older shoreline (Tahoe?) is only at 1780 feet (Blackwelder, 1954; Keaton et al., 1979). Previous workers have used this to infer that the older lake did not spill over whereas the younger one did. This seems unlikely because the older shoreline features are better-developed than the younger. The excellent wave cut benches and massive gravel bars associated with the older, 1780-foot level must have required a sill to control the water level. Also, there appears to be an old spillway south of Afton Canyon. Erosion associated with the incision of Afton Canyon has destroyed the lip of this spillway but it would appear to be slightly higher than 1800 feet. Using this observation that the Manix fault has a south-side-up component in this area allows for a much more reasonable history. The lake associated with the 1780-foot shorelines (north of the fault) flowed out south of Afton Canyon and the Manix fault. Uplift of this sill between the time of the 1780-foot and 1800-foot lakes caused the later lake to be moderated by a different sill, probably a pass associated with the Manix fault. Since this sill would be less resistant than the previous, uplifted sill, overflow associated with the 1800-foot level eventually carved Afton Canyon, which starts exactly where the Manix fault crossed the rim of the lake. This solution is much better than making Manix Lake the terminal sink for the Tahoe-age Mojave River which should have been, and appears to have been from the shoreline features, much more significant than the Tioga-age Mojave River. It also explains why Afton Canyon was not cut until the Tioga, when earlier pluvials must have been at least as significant and would have cut Afton Canyon if they had flowed out of Manix Lake at the same place that the Tioga river did.

CONCLUSIONS

As the late Pleistocene Mojave River flowed from the San Bernardino Mountains to Death Valley it left a record of lakes and spillways through each of the basins it passed. Since the shoreline features of these now dry lakes originally rimmed the basins at the level of the spillway, they provide excellent level lines within individual basins. Also, since the Mojave flowed continuously as far as Death Valley only during discrete pluvial maxima, correlation between basins is possible.

A study of the record of shorelines in Soda Basin revealed that the southern end of Soda Basin is rising with respect to the northern end at between 1 and 4 mm/yr. If this uplift is associated with the Manix fault, we can solve some of the problems of the interaction of Manix Lake and Afton Canyon. Uplift along the Manix fault caused the Mojave River to abandon its old sill south of the Manix fault. Downcutting into the fault zone cut Afton Canyon and drained Manix Lake.

An examination of the central Mojave Desert to determine if the Mojave River ever flowed south into the Colorado yielded no positive evidence. However, if the topography is changing as fast as the record in Soda Basin suggests, perhaps some time in the past a route could have existed that would have allowed pupfish (Cyprinodon) to swim from the Colorado to the Mojave.

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KELSO DUNES

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INTRODUCTION

Kelso Dunes comprise one of the larger dune fields of the California desert. They lie in the east-central Mojave Desert (lat. 34° 48' N., long. 115° 43' W.) roughly midway between the respective freeways to Las Vegas and Needles (Fig. 1) and are now easily accessible by well-graded and largely surfaced desert roads from both north (Baker) and south (east of Amboy). The dunes are part of a larger sand sea extending S.80° E. from the sand-source area; a broad alluvial apron periodically flooded by the Mojave River where it debouches from Afton Canyon (Fig. 1).

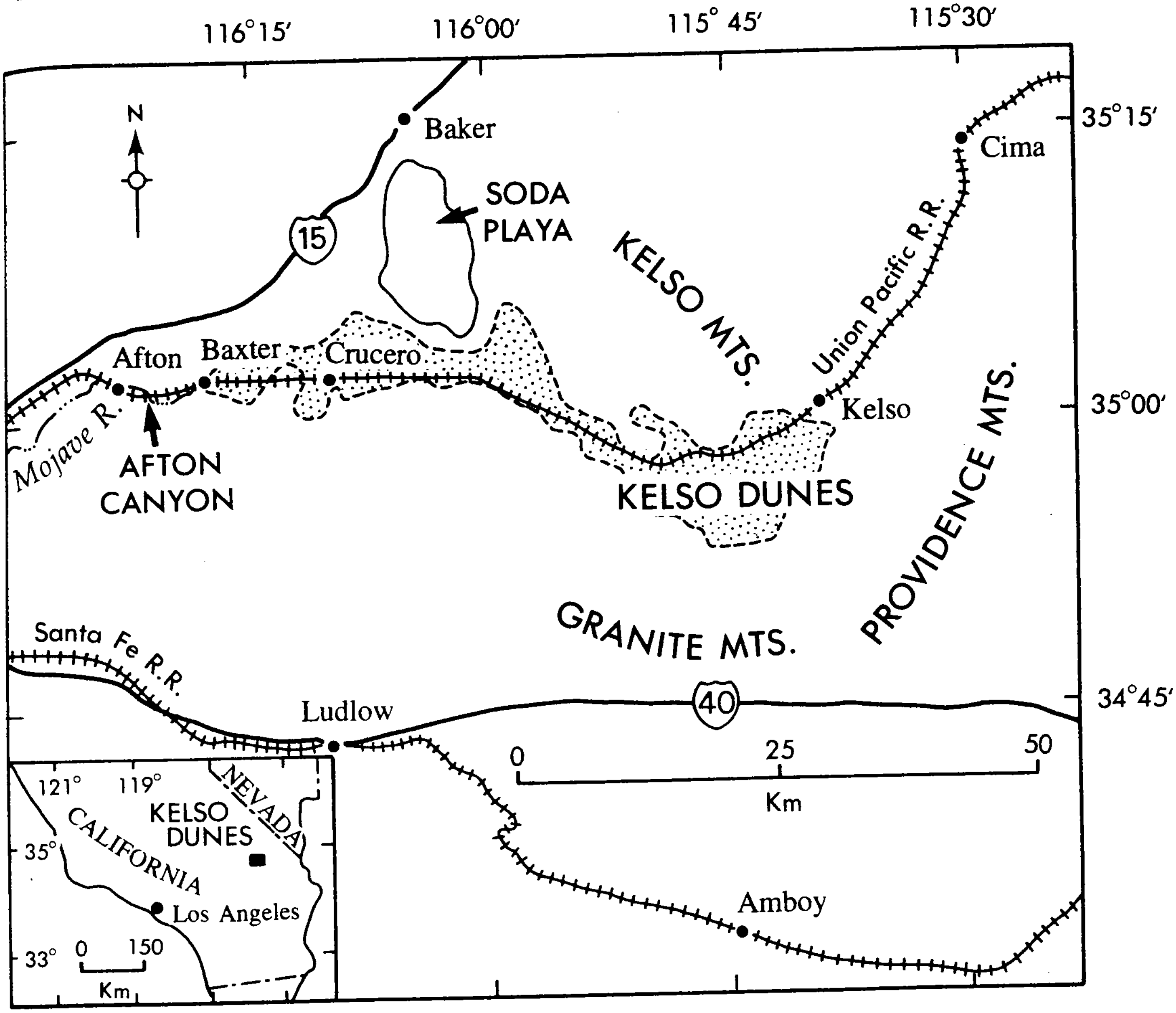

Figure 1. Location map for Kelso Dunes, Mojave Desert, California

The Kelso Dunes field covers 115 km^2, and maximum sand thickness approaches 215 m. The dunes have been accumulating for thousands of years, perhaps as much as 10,000 - 20,000, and vestigial dune patterns and vegetation remnants growing on the dunes indicate they have experienced significant environmental changes. The dunes lie within a broad mountain-rimmed valley, well out on the alluvial apron sloping north from the Granite Mountains. They are not plastered up against some topographic barrier, although their localization may have been initiated by a few low bedrock knobs, exposed on the east side of Cottonwood Wash (Fig. 2). They appear to be fixed in their present location by orographically controlled, conflicting wind patterns. Prevailing winds from westerly quadrants are locally counterbalanced by strong sand-transporting winds from northerly, easterly, and southerly quadrants. As a result, although much of the dune field is extremely active, the field itself is not going anywhere. Patterns of transverse dune ridges in areas partly stabilized by vegetation suggest the possibility of an earlier wind regime with an effective component from the southwest.

Material composing the dunes is a typical eolian sand in terms of grain size (90 percent between 0.25 and 0.50 mm), sorting, and rounding. In making the 56 km transit from the source area, ascending about 300 m in the final 37 km, individual grains are worn from subangular to progressively more rounded forms during the entire journey, but the well-sorted characteristic appears to become established within the first 16-19 km of travel. The sands are mineralogically complex, with quartz and feldspars, both potassic and plagioclase, predominant, but with subsidiary hornblende, pyroxene, sphene, tremolite, epidote, biotite, zircon, apatite, ilmenite, and considerable magnetite. Magnetite is abundant enough to justify establishment of placer claims within the dunes and to support attempts at magnetic separation to produce commercial iron ore. Much of the magnetite may come from a bedrock deposit near the mouth of Afton Canyon at the source area.

DUNE MORPHOLOGY

The prevailing dune forms of the Kelso complex are typical, small, irregular transverse

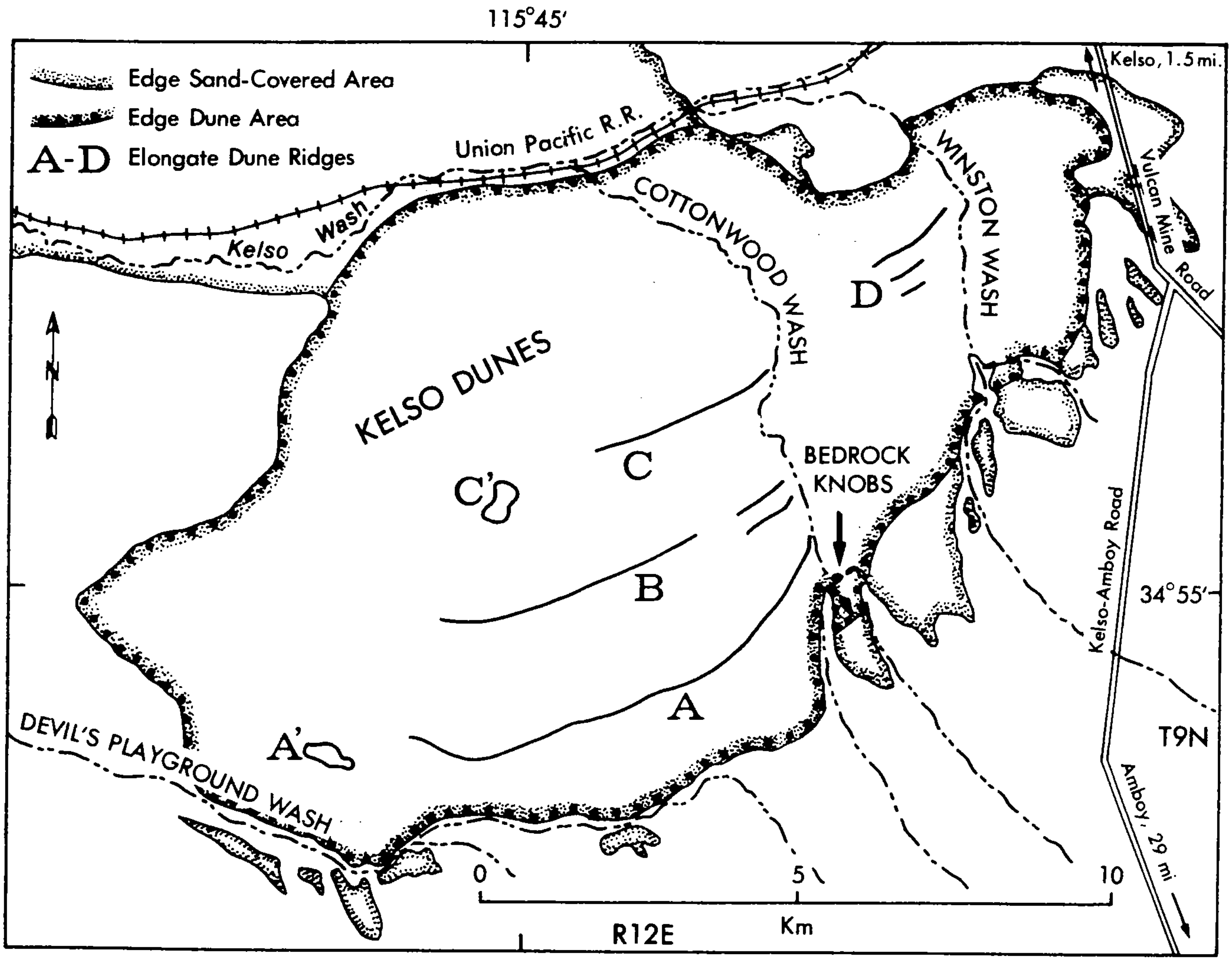

Figure 2. Geographic details of Kelso Dunes.

dune ridges. Though, in the currently most active areas, the orientation of many transverse dunes reflects the prevailing westerly wind regime, transverse dunes of other orientations, consistent with winds from more northerly and southerly quadrants, are also numerous. Some of these divergent dune ridges may reflect patterns initially established by earlier wind regimes. The combination of dune ridges of different trends locally produces waffle-like patterns, as viewed on airphotos.

These modest-sized dunes, lee faces to 10 m high, are superimposed on a coarser pattern of four much larger linear sand ridges (Fig. 2), trending about N. 65° E. The largest of these ridges (A, Fig. 2) is an imposing feature rising fully 170 m above the south base of the dune mass. The crest of these linear ridges is irregular, with numerous peaks and saddles, and abrupt departures from the prevailing linear trend. The linear ridges are not parallel to any existing dominant wind, but they are coherent with earlier transverse-dune ridges in partly stabilized areas of the field, suggesting that they may have been longitudinal to earlier winds from the southwesterly quadrant. The morphology of their active crests is currently shaped by oblique winds from both sides. Locally, large sand ridges projecting laterally from peaks on the linear ridge crests create a star-dune configuration.

Essentially all currently active transverse dunes in the Kelso field are subject to frequent reversals in crestal symmetry owing to the complex wind regime of the area. Some of the cross-sectional forms and changes commonly seen are graphically depicted in figure 3, wherein A represents the ideal transverse dune profile, and B to E show more typical forms.

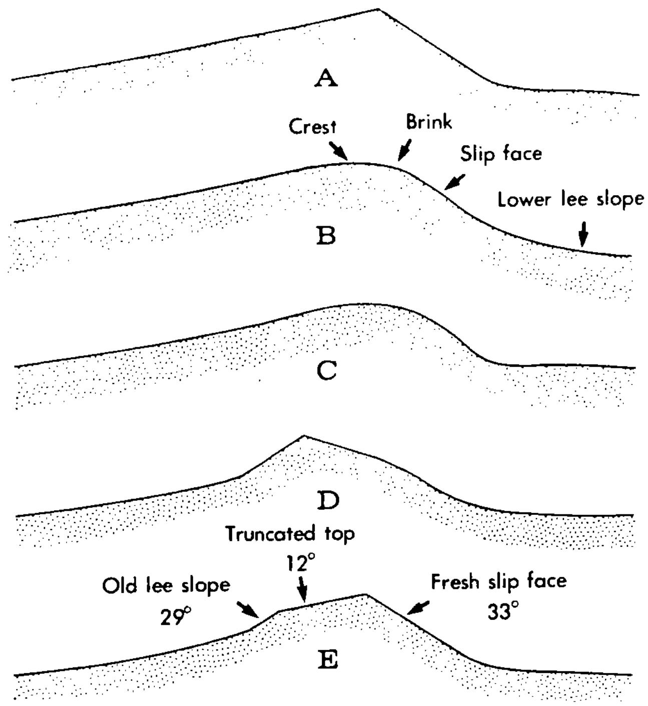

Figure 3. Cross sections through transverse dune ridges.

TEMPORAL CHANGES IN TRANSVERSE DUNES

Accumulation and removal of sand on dunes and changes in form and facing direction of individual transverse ridges have been monitored and measured over a 15-year interval at 10 separate stations within the dune field. These observations show that changes in shape, orientation and sand thickness are greatest in the crestal areas of transverse dune ridges, and that over a period of years, under the current complex wind regime, changes tend to cancel out. The crest (or brink) of a dune may shift back and forth many times, moving a cumulative distance measured in many tens of meters, and yet end up just about where it started (Fig. 4). Cumulative values for sand accumulation at a specific point on a dune tend to be just about balanced by cumulative episodes of removal at the same point.

At one dune station (Fig. 5) over a 9-year interval, the accumulation of 417 cm of sand on one side of the dune's crest was exactly balanced by 417 cm of sand removal at the same point. On the opposite side of the crest, 853 cm of sand accumulated and 688 cm of sand were removed, leaving a net accumulation of 165 cm, which was probably eliminated by a subsequent episode of removal. Measurements of this type support the conclusion that even though many individual transverse dunes within the Kelso field are highly active, they are not going anywhere at any detectable speed. The same behavior seems to hold for the entire dune field.

INTERNAL BEDDING AND LEE-SLOPE ORIENTATION

The literature contains many examples of attempts to determine paleo-wind directions from the attitude of cross laminations within ancient sandstones of presumed eolian origin. Large sweeping cross beds, inclined at angles within a few degrees on either side of 30°, are assumed to represent lee-side bedding. Excavations into the Kelso Dunes reveal that most of the cross lamination dips at angles less than 25° and is so inconsistent in orientation that detemination of effective wind directions in this area from bedding attitudes in the sand deposits would be an essentially hopeless task.

Measurement of current lee-slope orientation on hundreds of transverse dune ridges likewise gives a picture of wind directions considerably at variance with data derived from direct wind observations. If the Kelso Dune sand deposit were to be fossilized and incorporated into a stratigraphic sequence, it appears that the orientation of cross laminations in these sands would not yield a very reliable indication of the prevailing wind regime of the eastern Mojave Desert. Short-lived powerful storm winds from aberrant directions, complexity of dune forms, and local orographic setting are factors complicating the record provided by cross-lamination attitudes. These are factors that would be hard to evaluate in a fossilized setting exposed largely in cross section.

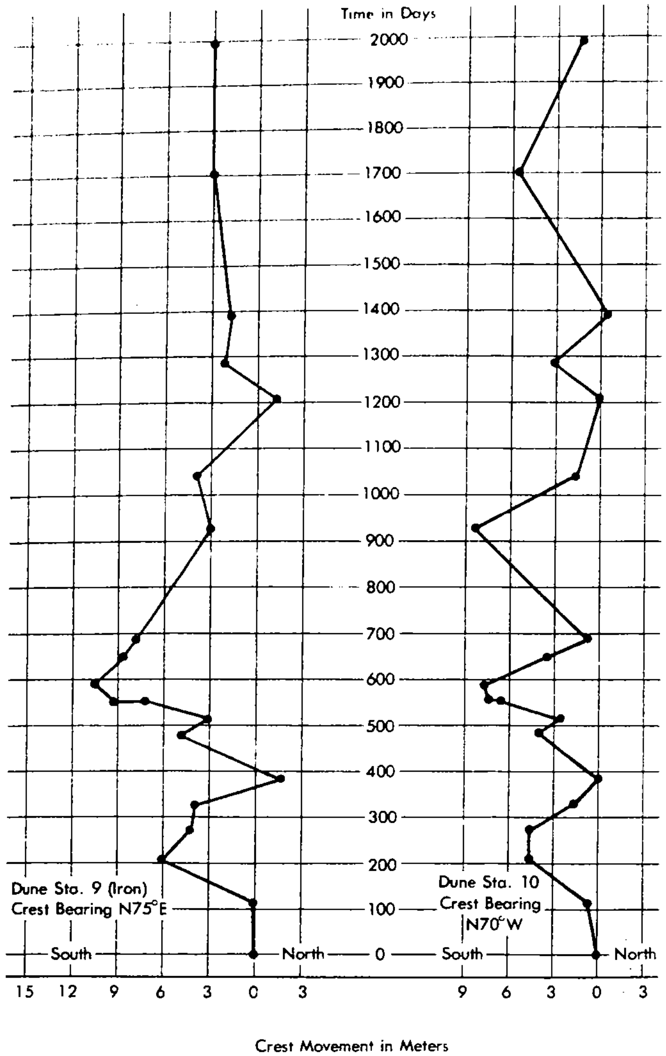

Figure 4. Plot of shifts in position of dune crests at stations 9(iron) and 10.

THINGS TO LOOK FOR

The usual and easiest access to Kelso Dunes is from the south side. Approaching the semi-stabilized (by vegetation) south edge of the sand deposits, dead "bird-cage" bushes (dune primrose) are often seen. Indians loved the dunes, and remnants of Indian campfire sites and some artifacts used to be seen and found in the lower, partly vegetated areas of the dunes. These have been largely disturbed or removed by visitors to the dunes over the last 15 years.

Deposits of dune sand preserve moisture effectively, and digging or exposures by wind scour often reveal areas of wet sand within the dunes many months after any precipitation has occurred and when the surrounding desert is bone dry. Vegetation, burrowing animals, and bugs are well aware of this reservoir of moisture, and take advantage of it in their habitat. Undisturbed sand surfaces record a variety of interestig animal and crawling bug tracks. The well camouflaged, very fleet lizards that burrow into and boil up out of the sand are a constant source of surprise to foot travelers on the dune surface.

Depending upon immediately antecedent wind conditions, sand ripples of various sizes, orientations and ages may be available for inspection. One can get a good feel for the divergence of sand transporting wind currents, at the ground level, from the prevailing wind direction aloft by observation of ripple orientations of contemporaneous origin. Winds at the ground surface in dunes can move sand in directions orthogonal to the prevailing wind aloft. The directions of ground-surface winds are strongly influenced by small and subtle topographic configurations, so, to a large degree, the wind conforms to the ground-surface rather than the reverse.

Differences in ripple wave-lengths and amplitudes can be seen to be strongly influenced by the grain size of the sand involved. This becomes particularly apparent on the floors of some intra-dune hollows which have been subjected to deflation. Accumulations of larger grains there, in excess of 1 mm diameter, lead to the local development of granule (as contrasted to sand) ripples which may have wavelengths of more than a meter and amplitudes of 10 cm. Such coarse grains move by saltation-impact creep.

If a visit happens to coincide with an interval of strong wind action, many interesting little experiments with wind ripples are possible. One is to mark the crests of a series of ripples with toothpicks, and then measure the rate of ripple movement. Under high winds (50 km/hr) some ripples are capable of moving a full wavelength in one minute. One can also smooth out an area of rippled sand by hand and then watch the ripples reform. As long as one keeps to windward dune slopes and avoids areas to the lee of dune crests, dunes are not unpleasant in a high wind. The saltating sand curtain across a firm smooth windward sand surface seldom exceeds a height of 30 cm.

If powerful winds have recently been at work, lee faces of transverse dunes will also have been active and will probably be scarred by the marks of recent sand avalanches. The lee faces will also be unstable, and one can easily start a sequence of sand avalanches on such slopes by walking along the brink of the face, or jumping from the brink onto the upper part of the lee face. Watching a sand avalanche flow is like eating peanuts. It's an activity difficult to abandon. One can learn a great deal about the mechanism of grain flow by watching this phenomenon. Larger avalanches will generate the low pitched noise of singing or booming dunes.

If strong winds are blowing, a visitor will have opportunity to inspect the fallacy, or reality, of the so-called fixed lee-side eddy of transverse dunes, long ago enunciated by Vaughan Cornish and later demolished by Wm. S. Cooper. Stand on the brink of a transverse dune ridge and watch the movement of sand or light fragments of dry grass or leaves along the lee side of the dune. If such material is lacking, toss a few scraps of crumpled paper or cellophane onto the lee face. It will soon become apparent that no powerful fixed lee-side eddy is undercutting the base of the lee slope. Sand avalanches occuring on the slope are taking place because of

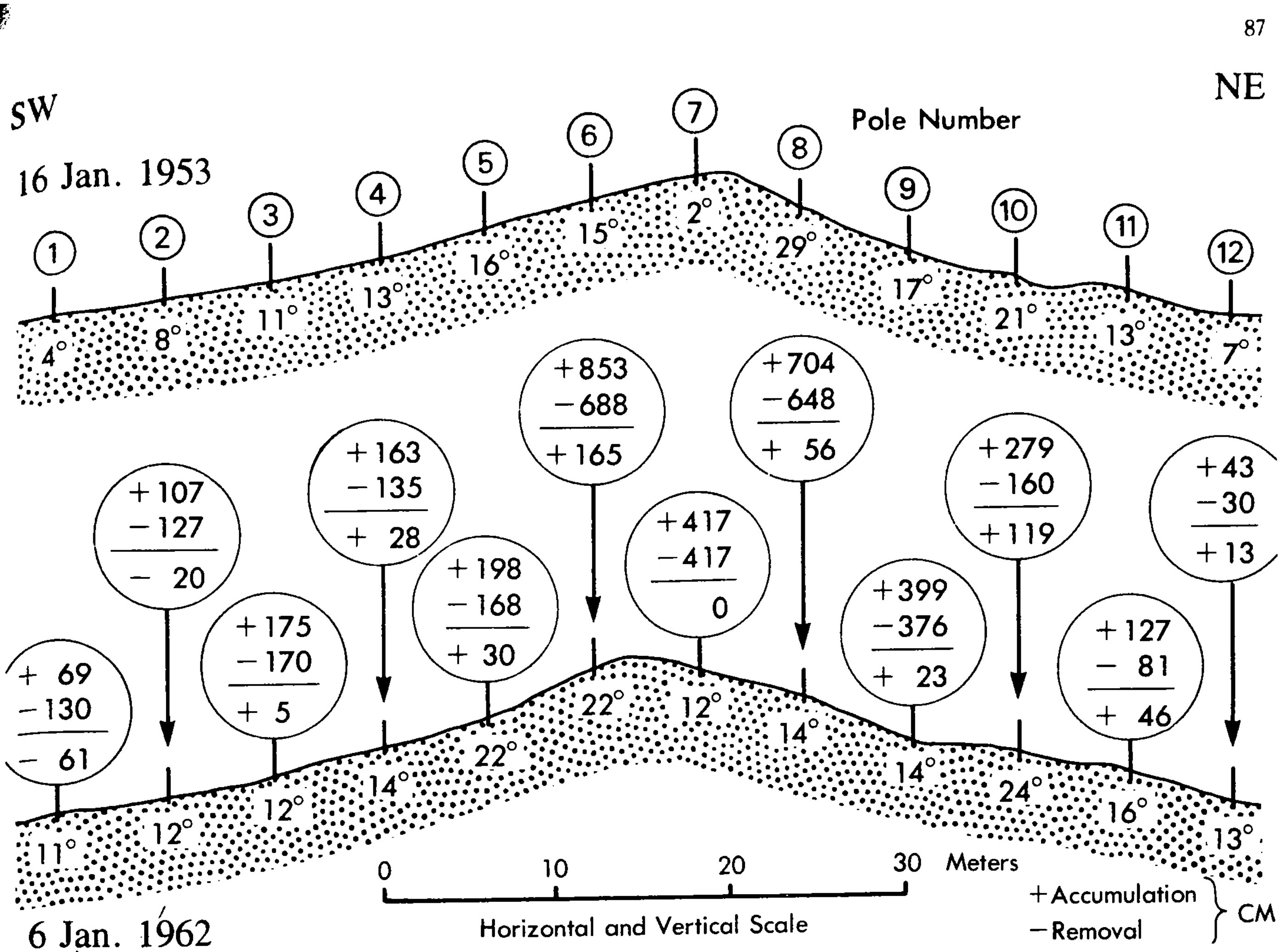

Figure 5. Summation of changes in centimeters recorded at poles of dune station 4 over a 9-year period. Because of intervals between measurements, gross changes recorded are probably only one-third to one-fourth of the absolute changes.

oversteepening by deposition on the upper reach of the slope, not because of undercutting at the bottom. Sand is not being swept up the slope, rather, if any sand moves at all, the drift is more nearly longitudinal along the slope or obliquely up or down it. Movement is only intermittent, caused by an occasional traveling eddy. Experiments with smoke bombs at various positions on the lee face amply confirm this behavior. Cornish's extrapolation from observation of fixed eddys to the lee of aqueous ripples in a flume is not applicable to sand dunes. If the wind becomes strong enough (80 km/hr) an observer standing on a firm windward sand surface sees that ripples disappear and indeed one is hard put to distinguish, visually, between firm sand surface on which one's feet rest and the very dense curtain of moving sand above.

Well up on the south flank of the eastern part of the highest linear ridge (A, Fig. 2) are remnants of still living large desert willow trees (catalpa), worth a quick inspection. All have experienced episodes of partial or complete burial, and some have not survived the experience. Trunks to 35 cm diameter are seen. These trees must be relics from some earlier interval of greater sand stability in this part of the dunes. They are certainly out of phase with the present regime of wind and sand activity.

The witching time in any dune field is the hour just after sunrise or before sunset, when low light emphasizes the exquisitely beautiful curves and shapes of dune forms. Deep shadows and highlighted dune slopes make fascinating collages, and the delicate pastel colors of the desert environment penetrate the dunes. People who have not been within a dune field under such conditions have missed one of nature's greater beauties.

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The Dish Hill complex (Field trip stop #12) consists of two volcanic cones, Dish Hill Crater and Siberia Crater (Wilshire and others, no date) (fig. 1). In addition, there are two neighboring craters, Hill 1068 to the southeast near U.S. 66, and Hill 1933, which can be reached by walking about 3 km northeast from Dish Hill Crater. Numerous other cones lie to the south in the Twenty-nine Palms Marine Corps Firing Range.

Dish Hill is well known because of the abundance of mantle-derived xenoliths and megacrysts in the lavas and tephra. Hill 1933 is similarly endowed, although less accessible. Hill 1068 contains only sparse, altered xenoliths.

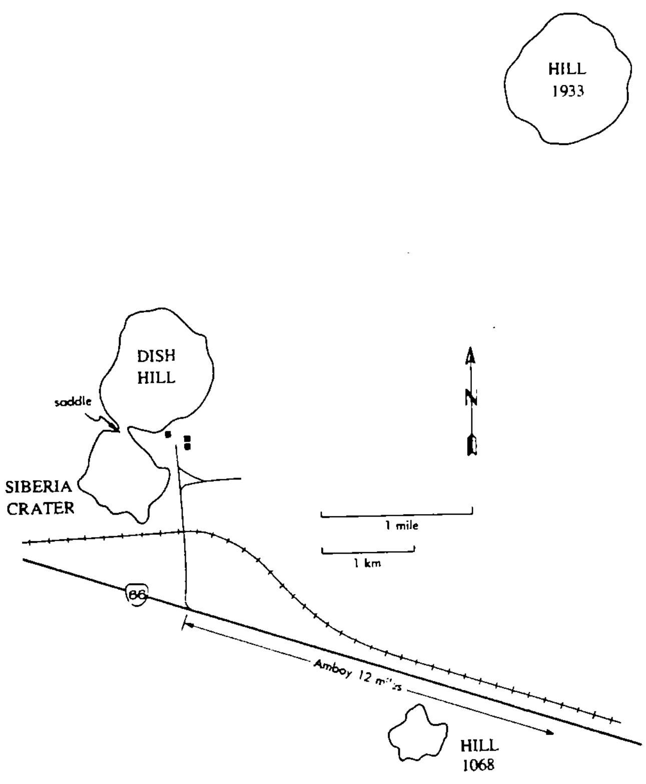

Figure 1. Index map to the Dish Hill complex.

The flows and tephra can be classified as alkali basalts in terms of their chemistry, but because they contain more than 5% normative nepheline, they technically are basanites.

The craters intrude Mesozoic granitic rocks similar to those in the Sierra Nevada. Dish Hill Crater has been dated by fission tracks in granitic xenoliths (2.1 ± 0.2 m.y.) and by the K-Ar method on an amphibole megacryst (1.9 ± 0.2 m.y.) (Wilshire and others, no date).

Xenoliths consist of spinel lherzolites, dunites, and other peridotites and pyroxenites. Megacrysts, which may be phenocrysts or, more likely, disaggregated dikes from mantle peridotite, consist of black amphibole (kaersutite), black clinopyroxene, chrome diopside, olivine, spinel, magnetite, apatite, and phlogopite. The kaersutite always has well-defined cleavage, commonly contains magnetite and/or apatite inclusions, and is always anhedral. The black pyroxenes never display cleavage, commonly contain spinel rather than magnetite inclusions, and are commonly euhedral.

The narrow canyon on the southeast side of Dish Hill Crater provides ready access to xenoliths and megacrysts, and it exposes well the volcanic stratigraphy. Black pyroxenite xenoliths are particularly abundant on the northwest slope of this crater. Many of the xenoliths in the saddle between the two Dish Hill craters display rinds (selvages) or dikes (veins) rich in kaersutite and/or phlogopite. Shervais and others (1973) describe a single garnet-bearing clinopyroxenite xenolith from Dish Hill; most or all of the garnet may have exsolved from the host pyroxene.

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PRELIMINARY GEOLOGY OF THE BRISTOL LAKE REGION, MOJAVE DESERT, CALIFORNIA

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INTRODUCTION

Clues for resolving several problems in Cordilleran geology are to be found in the central Mojave Desert. Mesozoic magmatic arcs, the Cordilleran thrust belt, and the western margin of the North American Precambrian basement all converge toward this region. Cenozoic volcanism and deformation overprint the earlier formed features and are related to both crustal extension and strike-slip regimes. The complex geology of this important region has until recently received surprisingly little attention. Only sparse modern data from the central Mojave Desert were available for the regional syntheses of Dibblee (1980a, 1980b, 1980c) and Burchfiel and Davis (1981). Several new studies, many of them summarized in Fife and Brown (1980) and in Howard and others (1981), further contribute to the geologic framework of the region. Our reconnaissance investigations in the Bristol Lake region (Fig. 1), summarized by this report, provide new data about styles of Mesozoic plutonism and deformation in the central Mojave Desert and about other aspects of the geological evolution since middle Proterozoic time.

The Bristol Lake region was previously mapped in reconnaissance by Kupfer and Bassett (1962; Bassett and Kupfer, 1964) and Southern Pacific Land Company geologic teams. These studies outlined expanses of Quaternary deposits and desert ranges that expose extensive Mesozoic plutonic terranes, Tertiary volcanic rocks, and scattered Precambrian and Paleozoic rocks (Bishop, 1963). This paper presents brief accounts of the findings from ongoing U.S. Geological Survey reconnaissance studies of the western part of the Needles 1° by 2° quadrangle (Fig. 2). More detailed results of our studies are being prepared for publication as U.S. Geological Survey maps. The geologic setting is outlined below, followed by descriptions and interpretations of rock units and structures.

GEOLOGIC SETTING

The Bristol Lake region lies in the central part of the Mojave Desert, which is bounded on the north and southwest by the Garlock and San Andreas faults and on the east by the Colorado River (Fig. 1). A trough variously called the Bristol-Cadiz-Danby trough and the Barstow-Bristol trough crosses the Mojave Desert on a west-northwest trend as a conspicuous physiographic feature (Gardner, 1980; Glazner, 1981a). In the Bristol Lake region the trough is occupied by the Bristol and Cadiz Lake playas at elevations of 181 m (593 ft) and 166 m (545 ft), respectively. Northwest trending mountain ranges in the Bristol Lake region rise to 1428 m (4685 ft) south of the trough and 2070 m (6790 ft) north of the trough.

Northwest trending mountain ranges in the western part of the Mojave Desert (Mojave tectonic block of Fuis, 1981) are in many cases bounded by faults active into the Quaternary, whereas ranges of generally northerly trend in the eastern part (Sonoran tectonic block) resulted from Tertiary tectonism with little or no younger faulting. The seismically active Mojave tectonic block has a thicker crust and intermediate heat flow values, as compared to the seismically inactive Sonoran tectonic block, which has a thinner crust and higher heat flow (Fuis, 1981; Lachenbruch and Sass, 1981). The nature of the transition between the Mojave and Sonoran tectonic blocks is poorly known. The Bristol Lake region approximately straddles this boundary.

Tertiary volcanic strata and local continental sediments lie unconformably on older rocks over a wide part of the southwestern United States, including the Mojave Desert. Complex faulting and rotation of the strata above low-angle detachment faults are now recognized in many areas in southern Arizona and along its border with southern California (Coney, 1980; Davis and others, 1980; Davis and Hardy, 1981). Detachment terrranes in the central Mojave Desert west of the Bristol Lake region were recently described by Dokka (1981) and Glazner (1981b), raising the possibility that much of the eastern Mojave Desert underwent similar deformation.

Mesozoic plutonic belts of probable Jurassic and Cretaceous age trend northwest across the Mojave

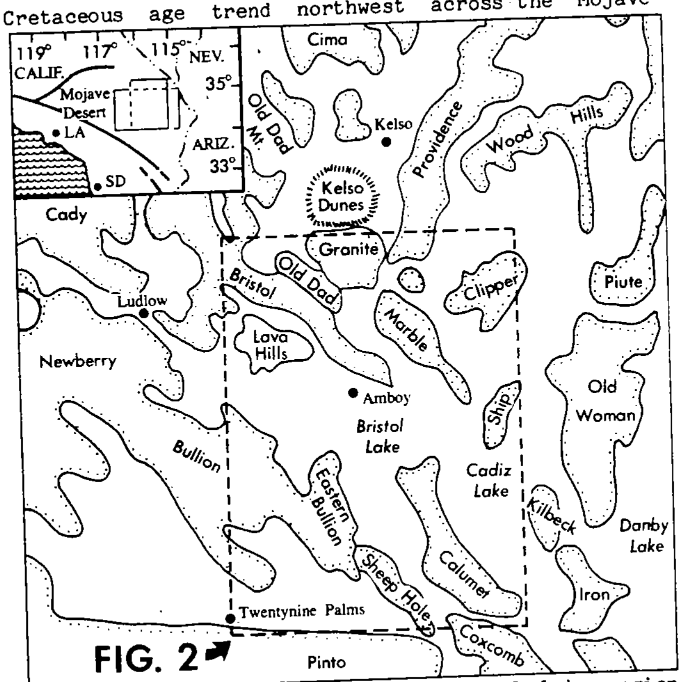


Figure 1. Location map of Bristol Lake region showing mountain ranges and lakes discussed in the text. The Needles 1° by 2° sheet is outlined by dashed lines on the inset map. LA, Los Angeles; SD, San Diego. The San Andreas and Garlock faults are shown on the inset map.

Desert (John, 1981). They link the Sierra Nevada batholith with plutonic terranes in Arizona and Sonora, Mexico, and represent subduction-related magmatic arcs (Hamilton, 1969; Coney and Reynolds, 1977). The Bristol Lake region lies near the eastern margin of the Jurassic magmatic belt (Burchfiel and Davis, 1981; John, 1981), whereas Cretaceous plutons occur throughout the Mojave Desert and farther east into central Arizona (Coney and Reynolds, 1977). As yet, there are few data reported on the composition, mode of emplacement, and age of plutonic terranes in the Mojave Desert.

Epicontinental Paleozoic strata thicken westward across the southwest Cordillera. A cratonal sequence was deposited on Precambrian crystalline rock through much of the southeastern Mojave Desert and Arizona, whereas a thicker, miogeoclinal sequence was deposited in the northwestern Mojave Desert and much of Nevada and Utah. Thrust faults and folds of the Cordilleran thrust belt deformed these sequences at several times during the Mesozoic, resulting in mostly eastward overriding of allochthonous strata (Burchfiel and Davis, 1981). North of the Mojave Desert the thrust belt lies east of the Mesozoic plutonic belts but in the Mojave Desert the belts intersect (Burchfiel and Davis, 1981), resulting in still poorly understood changes in the geometry and mechanics of deformation.

Aspects of the geologic history of the central Mojave Desert are revealed by rocks exposed in the Bristol Lake region, including Precambrian crystalline basement, Paleozoic and Mesozoic strata, Mesozoic plutonic and volcanic rocks, and Cenozoic volcanic and continental deposits. These rocks are described below.

ROCK UNITS

Prebatholithic Rocks

Precambrian rocks in the Bristol Lake region (Fig. 2) include gneiss in the Calumet, Bullion, Ship, and Bristol Mountains, granite in the Marble and Bristol Mountains, and diorite in the Marble Mountains. The Precambrian exposures are scattered and as yet few consistent patterns of lithologic types have been recognized. Augen gneiss, with alkali feldspar augen as large as 10 cm, in the Calumet and eastern Bullion Mountains resembles the Fenner Gneiss of Hazzard and Dosch (1937), widespread to the east in the Piute and Old Woman Mountains and Kilbeck Hills (Fig. 1). Silver and McKinney (1963) and Lanphere (1964) dated porphyritic granite in the Marble Mountains at 1.4 to 1.5 b.y. using U-Pb and Rb-Sr methods, respectively.

Paleozoic strata unconformably overlie Precambrian crystalline rocks in the Ship and Marble Mountains. The Marble Mountains contain about 500 m combined thickness of Cambrian strata consisting of the Tapeats Sandstone, the Latham Shale, Chambless Limestone, and Cadiz Formation of Hazzard (1933), and the Bonanza King Formation (Hazzard, 1931; Hazzard and Mason, 1936; Stewart, 1970). These rocks are deformed by bedding-plane faults in the Marble Mountains and are contact metamorphosed in the Ship Mountains, but are mostly homoclinal sequences. Similar, but more strongly metamorphosed, strata occur in the southern Bristol Mountains (Brown, 1981), where the lowest units are highly deformed and possibly separated from overlying strata by a bedding-plane fault. Pendants of metamorphosed Cambrian carbonate and clastic rocks also occur in granodiorite in the Calumet Mountains. In the Bristol Mountains metamorphosed Paleozoic strata form a nearly continuous section including the Cambrian Bonanza King and Nopah Formations, Devonian Sultan Formation, Mississippian Monte Cristo Limestone, and the lower part of the Pennsylvanian and Permian Bird Spring Formation(?) (Brown, 1981). Metamorphosed Bird Spring Formation containing Early Pennsylvanian (Morrowan) conodonts (Anita Harris, written commun., 1980) also occurs in the Marble Mountains, and an unmetamorphosed section 750 m thick crops out in the Ship Mountains.

The presence of a sub-Devonian unconformity in the Paleozoic strata of the Bristol Lake region indicates that they are cratonal, as defined by Burchfiel and Davis (1981). The North American craton, including Precambrian basement, therefore once extended southwest from eastern Utah to the Bristol Lake region.

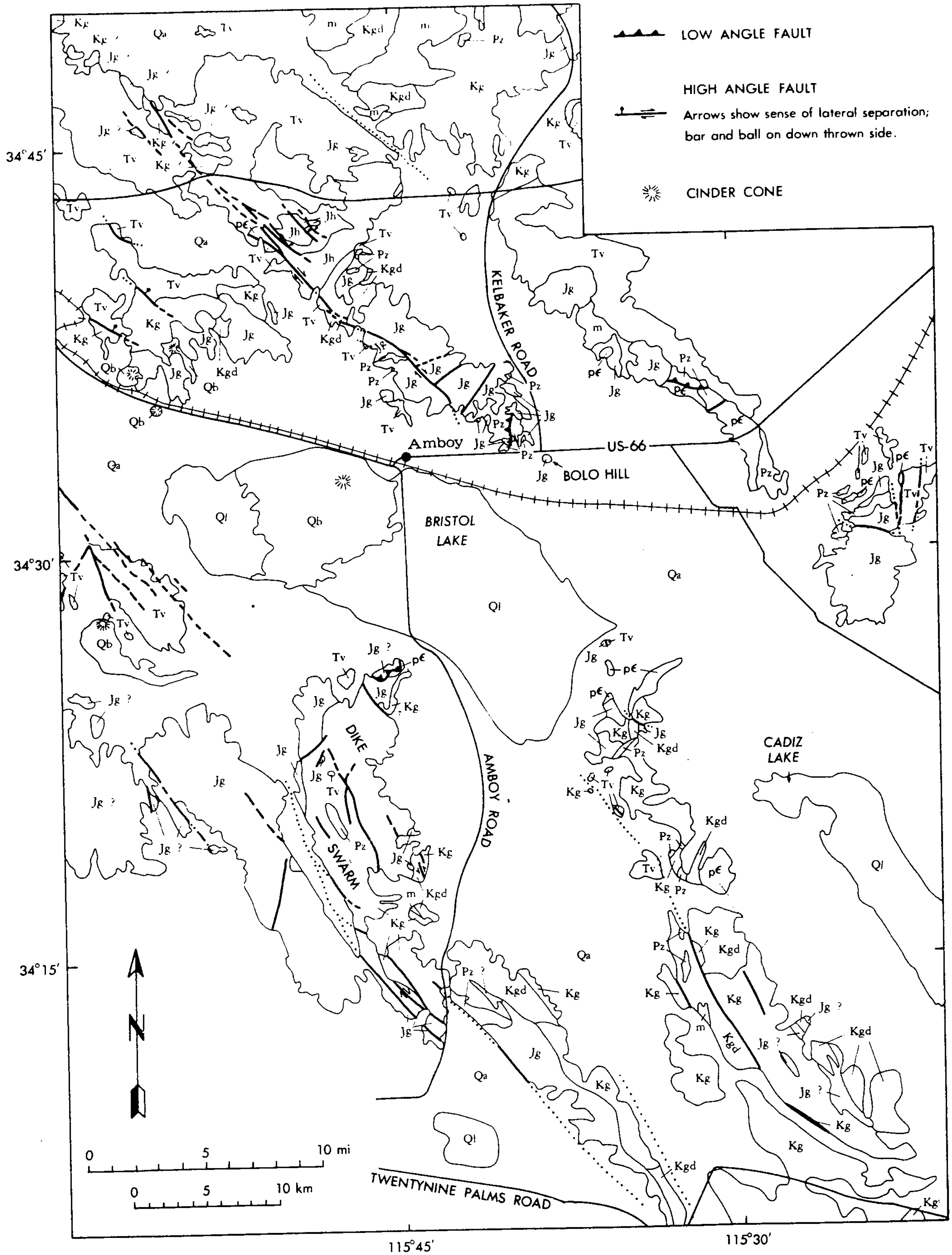
Banded gneiss and calc-silicate rock in the Sheep Hole, eastern Bullion, and Calumet Mountains possibly correlate with the metamorphosed Lower Triassic Moenkopi Formation(?) as exposed in the Old Woman and Little Piute Mountains to the east. Strata of probable Late Triassic(?) and/or Early Jurassic age crop out at Bolo Hill, southeast of the eastern tip of the Bristol Mountains (Fig. 2). Here, cross-bedded, arenitic metaquartzite, probably correlative with the Aztec Sandstone, is unconformably overlain by massive metavolcanic rocks of intermediate composition. These rocks are intruded by Middle(?) Jurassic quartz syenite. To the north and west of the Bristol Lake region similar relations involving Aztec Sandstone overlapped by and interfingered with volcanic rocks (Marzolf, 1980) are considered by Burchfiel and Davis (1981) to represent transgression of the Jurassic magmatic arc over cratonal rocks.

Metamorphism of Paleozoic and Mesozoic strata was apparently related to widespread emplacement of plutonic rocks during the Mesozoic. Skarns are developed along some contacts of Jurassic plutons.

Mesozoic Plutonic and Related Rocks

Mesozoic plutonic rocks constitute northwest-trending belts of arc magmatism that cut across Paleozoic depositonal trends; they are divided into two broad suites. The older suite commonly has high color index, contains lavender, grey, or pink alkali feldspars, and contains clots of the mafic minerals biotite, hornblende, pyroxene, and magnetite. This suite is intruded by the younger suite typified by low color index, white to flesh-colored feldspars, and disseminated mafic minerals. Available isotopic age data suggest that the older suite as we have identified it is Jurassic and the younger suite is Cretaceous. Refinement, and possible revision, of this classification will result from U-Pb isotopic geochronology in progress.

The Jurassic suite crops out extensively in the Bristol, Marble, Ship, Providence, and eastern Bullion Mountains and the Lava Hills. Intrusive contacts are discordant. Highly deformed equivalents of these rocks occur as screens in Cretaceous plutons in other mountain ranges in the region. The rocks exposed in the Bristol Mountains and Lava Hills provide a good example of the wide variety of rock types in the Jurassic suite. Here coarsely crystalline, variably porphyritic, biotite-bearing rocks range from monzodiorite and quartz diorite to quartz syenite and syenogranite (IUGS classification, Streckeisen, 1973). Less compositionally variable biotite-hornblende diorite and coarsely porphyritic syenogranite occur as plutons having mutually cross-cutting relations with


LOW ANGLE FAULT
HIGH ANGLE FAULT
Arrows show sense of lateral separation; bar and ball on down thrown side.
CINDER CONE
34°45′
34°30′
34°15′
115°45′
115°30′
KELBAKER ROAD
Amboy
US-66
BOLO HILL
BRISTOL LAKE
CADIZ LAKE
DIKE SWARM
AMBOY ROAD
TWENTYNINE PALMS ROAD
0 5 10 mi
0 5 10 km
N

ROCK UNITS

QUATERNARY	Qa	Surficial Deposits
QUATERNARY	Ql	Playa Sediment
QUATERNARY	Qb	Basaltic Rocks
TERTIARY	Tv	Volcanic and Sedimentary Rocks
CRETACEOUS	Kg	Granite to Granodiorite
CRETACEOUS	Kgd	Granodiorite
JURASSIC	Jg	Granitic Rocks
JURASSIC	Jh	Hypabyssal and Metavolcanic Rocks
PALEOZOIC & MESOZOIC	Pz	Paleozoic and Mesozoic Strata
PRECAMBRIAN	p€	Crystalline Rocks
	m	Metamorphic Rocks undivided

Figure 2. Generalized geologic map of the Bristol Lake region based on Bishop (1963) and Sabine (in Stein and Warrick, 1979); unpublished mapping by Howard and John, 1980, in the Calumet and Sheep Hole Mountains and the northeastern Bullion Mountains, and by Miller, 1980-1981, in the Bristol Mountains and Lava Hills south of highway I-40; and reconnaissance by John and Howard, 1980, in the Granite Mountains, by Howard, John and P. Stone in the Marble Mountains, and by P. Stone and Howard in the Ship Mountains.

the variably porphyritic rocks. This entire group of heterogeneous, coarse-grained rocks intrudes finer grained plutonic rocks in the central Bristol Mountains. The finer grained rocks occur as small bodies, each of which is compositionally diverse. The finer grained plutons range widely in composition, but are generally more silica-rich than the coarse-grained group. Rocks belonging to both groups intrude hypabyssal and metavolcanic rocks in the northern Bristol Mountains (unit Jh, Fig. 2). This third group includes a wide variety of dark green to dark brown or black, highly altered, pervasively fractured rocks that are consistently fine grained. Rare outcrops show flow-banding suggestive of volcanic textures. These relations suggest that Jurassic plutonic rocks intrude related hypabyssal plutons and their volcanic ejecta. Hypabyssal rocks of the Jurassic suite also occur in the southern Ship Mountains and the Bullion Mountains, in the Pinto Mountains to the south, and possibly in the Providence Mountains to the northeast (Fig. 1).

Our work in the Jurassic suite suggests a general trend of decreasing quartz content to the northeast, perpendicular to the trend of the magmatic belt (John, 1981). Monzogranite is common in the Sheep Hole, Bullion, and Bristol Mountains and the Lava Hills. In contrast, quartz monzonite prevails in the Calumet and Providence Mountains, and in parts of the Bristol Mountains; and quartz syenite and/or leucoquartz syenite occur in the Ship, Bristol, and Marble Mountains. However, quartz-rich granite and leucogranite in the Marble Mountains complicates the trend of decreasing silica to the northeast.

K-Ar dates for the Jurassic suite are 68 to 158 m.y. in the Bristol Lake region (Armstrong and Suppe, 1973; Calzia and Morton, 1980) and represent various degrees of argon loss during cooling or later reheating (Miller and Morton, 1980). The youngest dates, 68 to 74 m.y., come from rocks adjacent to the Cretaceous Cadiz Valley batholith and near Quaternary volcanic centers in the Lava Hills. Two Jurassic U-Pb ages reported by L. T. Silver (in Bishop, 1963; Calzia and Morton, 1980) suggest crystallization 151 m.y. ago for the Ship Mountains pluton, and 167 m.y. ago for a pluton in the Marble Mountains for which biotite K-Ar ages are 145 and 158 m.y. (Armstrong and Suppe, 1973). Rocks containing distinctive lavender phenocrysts of alkali feldspar, which are widespread in the western half of the Bristol Lake region, resemble porphyritic granodiorite in the Pinto Mountains and elsewhere in the central Mojave Desert that yields maximum K-Ar dates of 160 to 185 m.y. (Bishop, 1963; Calzia and Morton, 1980; Miller and Morton, 1980). We therefore infer that most plutons of the older suite in the Bristol Lake region are of Early(?) to Middle Jurassic age. The plutons and the volcanic rocks they intrude represent part of an Andean-type magmatic arc along the Mesozoic margin of the North American craton. The Jurassic magmatic arc in the Bristol Lake region is largely epizonal.

Plutonic rocks of the Cretaceous suite are extensively exposed in the southern part of the region as the Cadiz Valley batholith (John, 1981), and crop out as smaller bodies in the north. The Cadiz Valley batholith, which underlies much of the eastern Bullion, Sheep Hole, Calumet, northern Coxcomb, and Iron Mountains, is made up of a number of probably cogenetic plutons (John, 1981). The batholith is in sharp contact with the Jurassic plutonic suite and metamorphosed Triassic(?) strata in the western Sheep Hole, eastern Bullion, and northern Calumet Mountains, and intrudes Precambrian rocks in the Calumet Mountains. Much of the northern boundary of the batholith lies buried in the Bristol-Danby trough. The oldest plutons in the Cadiz Valley batholith are hornblende-bearing, biotite-sphene granodiorite with a color index of 7 to 17%. These plutons adjoin and contain inclusions of prebatholithic wall rocks. Younger plutons are leucocratic, subequigranular to coarsely porphyritic granodiorite to monzogranite, containing 2-11% biotite and as much as 3% muscovite. Rare late-stage dikes of garnet-bearing, two-mica aplite and pegmatite cut the plutons. The batholith is compositionally restricted (used in the sense of Pitcher, 1979) to a relatively narrow range of leucocratic granodiorite to granite. The Cadiz Valley batholith has steep, generally concordant walls except along a subhorizontal roof zone in the Iron Mountains.

K-Ar dates on biotite and muscovite from the Cadiz Valley batholith range from 55 to 69 m.y. (Armstrong and Suppe, 1973; Calzia and Morton, 1980). John (1981) suggested that emplacement occurred perhaps 75 to 85 m.y. ago based on estimated cooling rates.

Cretaceous(?) plutons in the Bristol Mountains and the Lava Hills range from monzogranite to leucocratic granodiorite and contain biotite but not muscovite. Plutonic rocks in the Granite Mountains include hornblende granodiorite and younger two-mica monzogranite, both of which intrude possibly related tonalite, quartz diorite, and diorite country rocks (Sabine, 1971; Stein and Warrick, 1979). This range in composition contrasts with the narrower range observed in the Cadiz Valley batholith. The wider range of composition is typical of the Sierra Nevada batholith where older, hornblende-rich quartz diorite and tonalite plutons have been intruded by younger granodiorite and monzogranite bodies. The Sierra Nevada plutons are commonly rich with mafic inclusions, in contrast to most of the Cretaceous rocks in the Bristol Lake region. Further contrasts are the rarity of plutonic muscovite in the Sierra Nevada and that plutonic rocks in the Bristol Lake region, unlike the Sierra Nevada, demonstrably intrude Precambrian crust.

Tertiary Rocks

A thick section of Miocene(?) volcanic and sedimentary rocks occurring in the northern Marble and Bristol Mountains and the Lava Hills was briefly described by Dibblee (1980b). Similar rocks crop out in smaller areas of the Calumet, Ship, and Bullion Mountains. In most areas the Tertiary volcanic and sedimentary rocks are tilted. Dikes of probable Tertiary age occur in both crystalline basement rocks and Tertiary strata, and may represent feeders for some of the extrusive rocks.

Tertiary strata in the Marble and Ship Mountains unconformably overlie the crystalline rocks and dip 10° to 30° east or northeast (Southern Pacific Company, 1959; Kilian, 1964). The lower part of these sections includes varied types of silicic and intermediate volcanic rocks and continental sedimentary rocks. A welded tuff overlain by an olivine basalt flow occurs at the top of the volcanic sequence. Based on its lithology and stratigraphic position, this tuff may correlate with similar tuff(s) containing blue sanidine in the Clipper Mountains and the Little Piute Mountains to the east, and possibly with a blue sanidine bearing welded tuff to the west in the Bristol Mountains. The tuff in the Little Piute Mountains is dated at 18.3 ± 0.6 m.y. by K-Ar on sanidine (R. F. Marvin, written commun., 1981) and may correlate with the Peach Springs Tuff of Young and Brennan

(1974), widely exposed to the east, northeast and southeast (Young and Brennan, 1974; Carr, 1981).

Tertiary volcanic and sedimentary rocks in the Bristol Mountains and the Lava Hills exhibit complex facies changes, caused in part by faulting coeval with deposition. In general, the base of the section contains fanglomerate overlain by water-lain tuffs, although in the western Lava Hills a thick section of probable nonmarine sandstone, siltstone, and limestone occurs. Stratigraphically higher are felsic tuffs and widespread andesite or dacite breccia and lava flows. Perlite, flow-banded rhyolite, and olivine basalt appear to lie in the upper part of the section. Sill-like aphanitic rhyolite(?) intrusions may be related to the perlite and flow-banded rhyolite. Along the southwestern margin of the Bristol Mountains, blue-sanidine bearing welded tuff forming the youngest volcanic unit is overlain by fanglomerate, sandstone, and fine-grained lake sediments possibly correlative with the Hector Formation of Woodburne and others (1974) described by Glazner (1981b) in the southern Cady Mountains. The welded tuff has a provisional K-Ar sanidine age of about 20 $\pm$ 4 my (M. A. Pernokas, personal communication, 1981). Because the tuff is similar in lithology, age, and stratigraphic position to tuff west of the Bristol Mountains (Glazner, 1981b) and to the Peach Springs Tuff of Young and Brennan (1974) to the east, it may be a distinctive marker within the Miocene volcanic sequences of the region.

A dense swarm of moderately northeast-dipping dacite dikes cuts Cretaceous granite in the eastern Bullion Mountains and constitutes approximately half of the rock across an outcrop width of 9 km. Porphyritic dacite and aphyric felsite dikes also cut Cretaceous granite in the Calumet Mountains. Similar dikes, as well as lamprophyric (Brown, 1981) and diabase dikes, commonly dip steeply and strike northeast in the Bristol Mountains and the Lava Hills.

Remnants of three breccia deposits of probable Late Tertiary age occur along the west sides of the Calumet Mountains and the adjacent range to the south. The southern two deposits contain exotic clasts of foliated two-mica granite gneiss and the northern one consists of volcanic clasts. A similar deposit in the Bristol Mountains 9 km northeast of Amboy contains exotic unmetamorphosed limestone clasts.

Quaternary Rocks and Deposits

Alkali-olivine basalt and basanite of latest Tertiary(?) and Quaternary age form composite cinder cones and flows west and northwest of Bristol Lake (Wise, 1966; Wilshire and Trask, 1971). Most flows are covered by Pleistocene(?) alluvium, but the youngest Amboy Crater flow is covered only by Holocene deposits of Bristol Lake (Bassett and others, 1959; Parker, 1963). These young volcanic rocks, and similar rocks to the northwest, are located in or near the axis of the regional trough that contains Bristol and Cadiz Lakes.

Pleistocene or older (Pliocene?) coarse alluvial deposits in a broad area north of the Lava Hills contain abundant flow banded rhyolite clasts that are rare in the Pleistocene and Holocene sediments of the area. Widespread younger Pleistocene fanglomerate and alluvium are typically capped by desert pavement composed of clasts coated with "desert varnish". Active alluvial systems are shedding sediment toward the Bristol Lake and Cadiz Lake depressions. The divergent trends of the ranges with respect to regional slopes result in damming of the sediments on the upslope sides of the ranges, creating as much as 300 m difference in elevation of the Quaternary surfaces on opposite sides of the ranges.

Playa sediments in Bristol and Cadiz Lakes extend below sea level in the subsurface; in Bristol Lake they are dominated by clay and salt to a depth greater than 300 m and those of Cadiz Lake are mainly clay, silt, and sand to at least 150 m depth (Bassett and others, 1959). Fossil-bearing beds in the lower part of cores recovered from Cadiz Lake suggest that it was a lake, rather than a playa, for part of its early history. Subsurface $CaCl_2$ brines are being extracted commercially from both lake basins. The surface of Cadiz Lake has been deflated 1 to 2 m, as indicated by remnant gypsum-capped pedestals (Basset and Kupfer, 1964). Large tracts of wind-blown sand lie in the basins and are banked against the northwest sides of mountains in the region.

STRUCTURE

The structural evolution of the Bristol Lake region is best recorded where stratified rocks serve as structural markers and is poorly recorded in the widespread massive batholithic rocks. Precambrian gneiss in the Ship Mountains was deformed and metamorphosed prior to deposition of Paleozoic sediments, and probably before the intrusion of undeformed 1400 m.y. old granite in the nearby Marble Mountains. Paleozoic metasedimentary rocks record folding, low-angle and high-angle faulting, and metamorphism, all of Mesozoic age. Tertiary strata, extensively exposed in the northern part of the region, are tilted and cut by northwest-striking high-angle faults with both dip- and strike-separation.

Detailed studies by Brown (1981) in the Bristol Mountains have shown that rocks as young as Pennsylvanian are deformed by westward-overturned folds and related bedding-plane faults that formed prior to intrusion of Middle(?) Jurassic rocks having a minimum K-Ar age of 151 m.y. Nearby small outcrops of the Early Jurassic Aztec(?) Sandstone contain no folds, but it is unclear whether the lack of folds is caused by the small outcrops or by the absence of the folding event (compare with Brown, 1981). Two sets of folds are cut by Jurassic intrusives which locally are themselves folded and faulted, indicating that plutonism was late-tectonic. In the Marble Mountains, marble of the Bird Spring Formation records Early Mesozoic metamorphism and deformation. Steeply dipping, folded marble here is in fault contact with unmetamorphosed Cambrian strata and Precambrian diorite along a zigzag, northwest-dipping fault having steep segments alternating with flatter (30° north dip) segments. Slivers of Cambrian strata displaced downward and relatively westward along the fault suggest low-angle, normal displacement. Burchfiel and Davis (1981, p. 141), who instead interpret the fault as a south-directed thrust(?), report that the fault is cut by a Middle Jurassic pluton (U-Pb isotopic age of 167 m. y. by L. T. Silver, reported in Bishop, 1963) and therefore records Jurassic or Triassic folding and faulting.

Early Mesozoic bedding-plane faulting, folding, and metamorphism is therefore bracketed between deposition of the Bird Springs Formation of Pennsylvanian and Permian age and lower Middle Jurassic plutonism. These early Mesozoic low-angle faults are west- or northwest-directed, in contrast to the southeast-directed thrusts and nappes of early Mesozoic age 40 km north at Old Dad Mountain (Dunne, 1977) and those of late Mesozoic age in ranges to the east and northeast (Howard and others, 1980; Burchfiel and Davis, 1981). Moderately east-dipping reverse-separation faults in the Bristol Mountains record post-Middle Jurassic (Cretaceous?) deformation that is contrary to the Cretaceous deformation styles observed to the east and north (Brown, 1981).

A major metamorphic and deformational event that postdates emplacement of the Jurassic plutonic suite and predates intrusion of Late Cretaceous plutons is recorded in the Kilbeck Hills, just east of Cadiz Lake (Fig. 1). Metamorphosed, folded and foliated Paleozoic strata and Jurassic(?) sills there are intruded by undeformed Late Cretaceous granite and granodiorite.

The Cretaceous Cadiz Valley batholith and its country rocks show evidence of forceful intrusion under mesozonal conditions. Metamorphosed wallrocks, screens, and enclaves contain rocks such as amphibolite, hornblende-epidote gneiss, wollastonite marble, and sillimanite-bearing gneiss. Foliation in the metamorphic rocks is steep and concordant with the plutonic contacts. In map view, the foliations wrap around the plutons, suggesting that the wallrocks were forced aside during intrusion. Wallrock deformation is strongest in the Sheep Hole Mountains, where Jurassic porphyritic granite is converted to schistose augen gneiss. Ptygmatic folding of satellitic(?) aplite dikes in the wall rocks suggests that deformation occurred during pluton emplacement.

The batholith itself is ductilely deformed in the south, where tectonite fabrics suggest an increase in late-batholithic deformation eastward and upward. This deformation is manifested by subhorizontal mylonitic foliation and lineation that we attribute to flattening. Along the eastern contact of the batholith in the Iron Mountains, undeformed granite grades structurally upward into mylonitic gneiss 1.3 km thick which bears a well-defined west-southwest plunging lineation defined by elongate quartz grains (Miller and others, 1981). Aplite dikes bearing a similarly oriented mylonitic lineation in the northern Coxcomb Mountains cut megascopically undeformed granodiorite. Granodiorite in the Sheep Hole Mountains is locally lineated. Mineralized joints dated at 61.3 $\pm$ 1.4 m.y. by K-Ar on muscovite (R. F. Marvin, written commun., 1981) cut the mylonitic fabric in the Iron Mountains. This age is close to the presumed batholithic age (as described above) and suggests that the joints formed during cooling and were mineralized by late-stage emanations from the batholith. The mylonitic deformation preceeding joint development was related to batholithic emplacement and probably resulted from flattening of the roof (Miller and others, 1981). The mineralized joints, which dip moderately to the northeast, were utilized by younger faults in the Calumet, Sheep Hole, and Bullion Mountains.

Widespread steeply tilted and faulted Miocene and Oligocene(?) strata indicate widespread Tertiary tectonism. Apparently the oldest Tertiary structure is a north-dipping low-angle fault younger than Cretaceous granite but older than Miocene(?) dacite dikes in the northeastern Bullion Mountains. Tertiary strata dip 65 to 75° southwest in the Calumet Mountains and 25 to 55° southwest in the Bristol Mountains and Lava Hills. Northwest-striking normal(?) faults in the Lava Hills separate rotated Tertiary strata on the northeast from crystalline rocks. The direction of tilting and the orientation of normal faults are similar to those caused by Miocene northeast-directed transport and rotation on detachment faults and listric(?) normal faults in ranges to the west (Dokka, 1981; Glazner, 1981b) and to the east (Carr and others, 1980; Davis and others, 1980; Dickey and others, 1980). Tertiary dikes dipping northeasterly 45 to 60 degrees in the eastern Bullion Mountains may have been similarly rotated from an original steep inclination. Their exposed volume suggests 4 to 5 km of northeast-southwest inflation at this crustal level. Growth faults in the Bristol Mountains indicate that Tertiary basins were tectonically active early in the volcanic history. In contrast to much of the Bristol Lake region, Tertiary strata in the Marble, Ship, and Clipper Mountains dip gently northeast or are flat-lying, suggesting that these ranges represent a terrane relatively little affected by the same style and/or sense of Tertiary faulting and rotation.

Younger tectonism is indicated by latest Tertiary and/or Quaternary breccia deposits and gravity slide blocks in the Bristol Mountains (Brown, 1981) and the southern Calumet Mountains. These deposits suggest rapid uplift and denudation, perhaps during the initial blocking out of the present ranges. Whether the present northwest-trending ranges were blocked out by Miocene(?) normal faults or by later strike-separation faults of similar orientation is unclear. Right-separation faults occur to the west in the Bullion Mountains (Dibblee, 1967; Miller and Morton, 1980; Jennings, 1975). Our studies have demonstrated at least 30 m of right separation for a fault 2.5 km west of Sheep Hole Pass, and more than 6 km for the fault system along the southwest border of the Bristol Mountains. The Bristol Mountains fault system cuts the lower beds of the Pleistocene(?) alluvium. However, a small amount (15 m or more) of left separation can be demonstrated for en echelon north-northwest-striking faults cutting dacite dikes 7 to 8 km north of Sheep Hole Pass, indicating that all faults of that orientation cannot be assumed to be right-separation. Northwest-trending faults that cut lower Pleistocene(?) alluvium along the Sheep Hole Mountains and in the southern Calumet Mountains dip steeply or moderately northeast parallel to older joints, and may have large dip-slip components. A possibly Quaternary fault encountered in core at about 78 m depth in Cadiz Lake places horizontally bedded lake sediments on similar, but moderately dipping, sediments (Bassett and others, 1959). This fault and the faults adjacent to the Bristol, Calumet, and Sheep Hole Mountains are among the easternmost Quaternary faults in the Mojave Desert. The region to the east is tectonically and seismically quiet (Carr and Dickey, 1976).

Superposed on the northwest trending mountain blocks is the west-northwest trending Barstow-Bristol-Danby structural trough that contains Bristol and Cadiz Lakes and extends far to the west and east beyond the region shown in Figure 2. Thick accumulations of playa sediments, tilting and faulting of the playa sediments, Quaternary and latest Pliocene(?) alkali basalt volcanism (Gardner, 1980; Glazner, 1981a), and divergent trends of mountain blocks south and north of the trough (Fig. 1) indicate that it is a major crustal structure. In the Bristol Lake region the trough coincides approximately with regional changes in neotectonics and seismicity (Carr and Dickey, 1976), depth to moho (Fuis, 1981), upper mantle seismic velocity (Hadley and Kanamori, 1977), and heat flow (Lachenbruch and Sass, 1981).

SUMMARY

Precambrian granite and gneiss in the Bristol Lake region forms a basement on which were deposited cratonal Paleozoic sediments of Cambrian, Devonian, Carboniferous, and Permian age. These rocks and locally preserved strata of probable Triassic and Lower Jurassic age were intruded in the Early(?) to Middle Jurassic by epizonal plutons, synchronous with or closely following folding and faulting marked by westward transport. This transport direction contrasts with the southeastward transport indicated by thrusts of both Early and Late Mesozoic age east and north of the region.

Plutons assigned to the Jurassic suite range widely from diorite to monzogranite and quartz syenite. Coarsely crystalline plutons discordantly intrude Precambrian and Paleozoic rocks and also finer grained rocks that evidently represent their hypabyssal roofs and volcanic ejecta.

Plutons of a younger, probably Cretaceous, suite of leucocratic granodiorite, monzogranite, and two-mica monzogranite invade much of the region. In contrast with typical Sierra Nevada batholith rocks, the Cretaceous plutons of the Bristol Lake region exhibit a restricted compositional range, a sparsity of mafic inclusions, and they intrude known Precambrian crust. The Late Cretaceous(?) Cadiz Valley batholith includes several plutons, and is a mesozonal forceful intrusion marked by deformed walls and local internal deformation. These features contrast to the shallow level, undeformed Jurassic plutons. Unseen overthrusts or thick Cretaceous sediments possibly could account for the different levels of emplacement.

Tertiary strata, largely volcanic, unconformably overlie older rocks. The lower part of the sequence is dominated by clastic rocks and rhyolitic tuffs and breccias, and is succeeded upward by andesitic, dacitic, and basaltic rocks, and welded rhyolite tuff. The welded tuff is possibly correlative with the widespread Peach Springs Tuff of Young and Brennan (1974). Tertiary rocks are tilted southwesterly over much of the Bristol Lake region, and may have been rotated in blocks above detachment faults similar to those recognized in contiguous terranes to the west and similar terranes to the southeast. A Tertiary dike swarm suggests 4 to 5 km of northeast-southwest extension. Tertiary strata in the northeastern part of the region are gently northeast-tilted, suggesting that they belong to a separately rotated or unaffected terrane.

Lower Pleistocene gravels are cut by northwest-trending right-separation faults in the western portion of the region, along the eastern margin of the seismically and tectonically active Mojave tectonic block. Whether the present ranges were blocked out by these faults or by earlier normal faults is uncertain. The west-northwest trending Barstow-Bristol structural trough obliquely transects the physiographic and structural grain of northwest-trending mountain blocks. This major crustal structure is marked by thick accumulations of playa sediments, latest Pliocene and Quaternary alkali basalt extrusion, and divergent trends of the mountain blocks north and south of the trough.

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