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Death Valley & the Mojave:

An informal FieldGuide

Your Tour Guides: Drs. Howard, Hanna, and Dunne

THE BULLION KINGS

Out of the dust along Sixth Street of Los Angeles on June 19, 1868, rolled several light wagons groaning with a cargo that gleamed in the sun. Mortimer W. Belshaw, expert mining man, was bringing the first shipment of bullion through Los Angeles from Cerro Gordo, a new silver camp two hundred miles northward in the "Owens River Country." A few weeks earlier Belshaw had arrived there to find scores of Mexicans mining silver and lead ores, smelting them in crude rock ovens, and producing the kind of glistening bars he was now bringing through El Pueblo.

As early as 1865 Mexican prospectors headed by Pablo Flores had discovered galena deposits at what they called Cerro Gordo, the Spanish equivalent of "Fat Hill." But the new development was not really exploited until May 1867, when a Mexican prospector from Owens Valley rode into Virginia City, Nevada, with tales of Cerro Gordo's riches. Known as a man of reticent nature, he astounded the bustling mining capital with his shouts of laughter and his eager display of silver quartz ore samples. After a careful assaying their value became known on the streets, and Virginia City echoed with the news. Silver quartz veins, far richer than the original galena deposits, had been discovered at Cerro Gordo!

Soon the strike was known in every mining center on the Coast. When the word reached San Francisco, Mortimer W. Belshaw was one of the first to move. Reaching Inyo County in April 1868, he found Cerro Gordo situated two miles above sea level at the crest of the Inyo Range, where a swarm of enthusiastic miners were digging into the side of Buena Vista Peak.

The new quartz mines were so rich, Belshaw found, that the impatient silver seekers were throwing away as useless any ore worth less than two hundred dollars per ton.

Belshaw began regular shipments on December 1, 1868, sending his bullion to San Francisco by a roundabout wagon and boat trip through Los Angeles. Freighters with eight-mule teams loaded at Cerro Gordo for sixty dollars per ton, lurching down Belshaw's toll road with brake blocks burning, wallowing across the sandy Mojave, doubling trips over rugged San Fernando Pass with half loads, and rolling into San Pedro from three weeks to a month later. Under consignment by the John J. Tomlinson forwarding house, the metal was transferred to the steamer *Orizaba*, which carried fifteen tons per trip and unloaded its cargo three days later at the San Francisco wharves. Here it was delivered to the smelting works of Thomas H. Selby, one of San Francisco's earliest hardware merchants.

Under the energy of M. W. Belshaw, Cerro Gordo was bursting with activity by the end of 1868. By an increase in miners' wages to attract new labor, Belshaw had made Cerro Gordo a "four-dollar camp," traditional mark of a full-fledged silver town. During the fall its population had jumped from two hundred to seven hundred, and by the spring of '69, as one resident reported, "roads are being constructed, town lots staked off, buildings going up, shafts sunk," and Belshaw's furnace "turning out the bullion faster than it can be carried away."

The camp's main street was filled with grizzled sourdoughs and fast-talking promoters, merchants and vagabonds, men who had turned up at every new camp from Coloma to Austin. Here in the rude street on Buena Vista Mountain they were hailing old comrades, clasping the hands of arriving rainbow hunters.

Cerro Gordo's rise as a roaring silver camp brought red-painted stagecoaches to her streets by July 1870. A weekly four-horse line began carrying passengers down the tortuous "Yellow Grade," around the lake, and through the adobe village of Lone Pine. Its terminus was Independence, county seat of Inyo, which was connected by stage with the Central Pacific Railroad in Nevada. Traffic was so heavy that a year later Owens Valley was served by a semi-weekly stage from Nevada, a tri-weekly from San Francisco via Walker Pass, and a weekly from Los Angeles across the Mojave Desert through Willow Springs and Little Lake. Between the valley and Cerro Gordo two competing stages were carrying full loads every day.

Deep within Buena Vista's bowels an army of miners toiled by candlelight with pick and blasting powder, working in shifts to extract the ore twenty-four hours a day. By tubs and pulleys they hoisted the precious rock to the heads of shafts, or trundled it to the mouths of tunnels by iron carts running on miniature rails.

Dominating the entire system was Belshaw's furnace. By day its smoke clouds darkened the sky and filled the street with black cinders; by night its vermilion flame cast a glow over the town and lighted the miners across the divide at the changing of the shifts.

But already Belshaw and Beaudry had found their silver-lead bars piling up around the smelters while the wagons lagged farther behind the output. They needed a responsible freighter to haul the bullion on contract. Probably as early as December of '68 Remi Nadeau, owner of a string of ten-mule teams, agreed to deliver Cerro Gordo's output to San Pedro Harbor.

With characteristic vigor Nadeau entered the Cerro Gordo trade. His teams were already falling behind production, however, when Beaudry's improved furnace started in 1870. He increased his outlay to thirty-two teams, most of them with twelve mules and two wagons, and agreed to haul an unheard-of 130 tons a month. But the figure was still only half the capacity of the furnaces, and Belshaw and Beaudry found their camp frustrated by isolation. They had made themselves masters of Cerro Gordo, the bullion kings of Southern California, but the bars of lead and silver were piling up on the mountainside instead of at the refinery in San Francisco.

It was little wonder, however, that Nadeau's teams had trouble over the two-hundred-mile road to Los Angeles. Starting with the eight-mile descent down the toll road from Cerro Gordo, wheels were chained in place and several spans of mules were tied behind out of harm's way in case the wagons lurched out of control. The teams reached Owens Lake near what is now Keeler and followed its shore line to the adobe village of Lone Pine. They then rolled southward, between the Sierras and the lake, whose sparkling waters at that time stretched twenty miles to the southern end of Owens Valley. Along its shores the teamsters could bathe and wash their clothes in the lathery brine, provided a supply of pure rinse water was handy. Two days' drive below the lake took the caravans past Haiwee Meadows Station to the rock-walled outpost at Little Lake, another alkaline body originally designated "Little Owens Lake." Dropping into the Mojave Desert, they traveled to the west of the modern highway in order to camp near the streams of water in Sand and Grapevine canyons. At the board-and-shingle stations at Indian Wells and Coyote Holes the teamsters watered and fed their mules.

Leaving the shadow of the gray Sierras, they toiled onward for the spring at Red Rock Canyon, where the sandstone cliffs were washed with bright red and white contrasts and sculptured by wind and rain into the weird figures of a giant toadstool or a huge wax-dripped candle, or the finely chiseled walls of a Gothic cathedral. But through this canyon the teams were doubled in strings of twenty mules while the wagon wheels sank to the hubs and the brake blocks dragged the sand. Southward lay a three-day, thirty-eight-mile stretch of waterless, sand-rutted road, always the most dreaded portion of the trip. But at the end stood Willow Springs, an adobe tavern eleven miles east of the present Rosamond, where the Cerro Gordo teamsters met about their campfires and broke the stillness of the desert nights with their boisterous songs.

Then on they pushed through forests of spiny Joshua trees, and in springtime among fields of orange poppies and purple lupine, frightening heards of antelope that bounded gracefully across the desert. After twenty-eight miles the mules dipped their heads in the pond at Barrel Springs. At their backs stood the conquered Mojave, before them the brush-covered Soledad Pass.

Along the dry river bed of Soledad Canyon, marked only by a sleepy, one-saloon mining camp, wheels crunched through deep sand that closed over the rims and half buried the turning spokes. At length, swinging southward opposite the mouth of San Francisquito Canyon, the teams pulled up to the stage stop and tavern at Lyon's Station.

Here the teamsters, traveling in twos and fours, unhitched one set of wagons and doubled their teams for the grueling climb over San Fernando or Fremont Pass. Upward they lurched, the chock blocks dragging after each right hind wheel, ready to hold the wagon when the mules lost momentum and stopped for a breath. Over this grade the stage passengers got out and walked — and sometimes pushed. At the summit a deep cut had been carved in the late 1850s for the Butterfield stages; by 1870 the traffic was so heavy that a sprinkling cart was employed to patrol the narrow slit and dampen the dust.

Down the south slope creaked the wagons, rolling into San Fernando Valley and stopping at Lopez Station, now under water near the dam of the San Fernando Reservoir. Across the barren valley they crawled, stopping at the Eight Mile House, a station near the summit of Cahuenga Pass, and swinging into Los Angeles along Sixth Street. Then with lead bells jingling, wood and leather creaking, blacksnake popping, mule skinner shouting and cursing, mules snorting and coughing, Nadeau's teams turned up Spring Street and raised dust through the business district. At the Commercial Street platform of the railroad depot they unloaded their cargoes, repairing then to Los Angeles Street's wholesale houses to be loaded with return merchandise. Bales of hay, casks of wine, sacks of potatoes — everything from a frying pan to a crate of live chickens — headed for Owens Valley behind Nadeau's teams.

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Indeed, the simultaneous arrival of the land boom and the Inyo trade had brought a sudden prosperity to Los Angeles. The farmers swarming into Southern California found a ready-made market for their surplus produce in the high-sided wagons bound for the silver mines of Inyo. Los Angeles was in the best position to supply that region with life's necessities. Most important item, however, was the feed bill for the 500-odd freight mules which hauled the silver bullion. Remi Nadeau and the other teamsters were buying Los Angeles County's entire surplus feed crop, thus establishing barley as one of its staple products. No longer were Los Angeles farmers saddled with the hauling charges and middlemen's fees that characterized export to the San Francisco market.

By 1870 silver bullion was a common sight in El Pueblo. Specimens of Cerro Gordo silver ore and bars of base metal were displayed at the Bella Union Hotel, in the *News* office, in the local bank, at blacksmith shops and jewelry stores up and down Main and Spring streets. Like Sacramento in '49, Los Angeles was now a bustling mining center. Scarcely a citizen but could give detailed information on matters in far-off Cerro Gordo. The camp's fame, in fact, had inspired an army of prospectors to comb the local Soledad and San Gabriel mountains, making El Pueblo their base of operations. On its street corners "quartz talk" could be heard any time of day, and rumors were frequent of rich strikes somewhere on the horizon. Bearded miners circulated about town, talking mysteriously of fabulous ledges, their pockets bulging with rock samples.

Almost every day long trains of Cerro Gordo mule teams, containing up to twenty wagons and $50,000 worth of silver and lead, swung through Spring and Main Streets between Sixth and Commercial, raising dust in the summer and splashing mud in the winter. Only some 340 tons of bullion passed through Los Angeles in the year and a half between Belshaw's first shipment and the end of 1869, but more than 700 tons rolled southward during 1870. "Silver coming, and goods being sent as return freight," observed the Los Angeles *Star*, "seem to be at present the order of the day." Though Belshaw himself would have laughed at the idea, he was taking his place with the men who were making a city of El Pueblo.

12. THE SILVER SENATORS

A stocky, sandy-whiskered prospector rode into Los Angeles in March 1874, and on the twenty-fourth the directors of the Chamber of Commerce were called to the courthouse to hear his story. Veteran of California's Mother Lode, Nevada's Comstock, and the silver mines of Mexico, Richard C. Jacobs was a recognized mining man. He had just arrived from a rock-ribbed canyon in the Panamint Range, western bulwark of Death Valley, where he and his comrades had struck copper-silver ore in January 1873. Rock samples assayed from $300 to $3000 per ton, and already the excitement had set prospectors on the march from every corner of Inyo County. What the region now needed was a wagon road from the outside world. Jacobs knew the Chamber of Com-

At present travelers reached Panamint Valley by branching eastward from the bullion trail at Little Lake and taking a burro path across the Coso and Argus ranges. The most feasible location for a wagon road, however, would be a branch out of Indian Wells which would swing south of those formidable mountains and reach Panamint Valley by way of Borax Lake and the north end of the Slate Range. The last steep pull in Surprise Canyon was already being improved for wheel traffic by a group of Inyo enterprisers headed by Barton McGee, the resolute frontiersman who took issue with the bandits. The only improvements necessary to permit heavy schooners to roll from Los Angeles to Panamint would be along the six-mile haul over the Slate Range.

Two days later the Los Angeles *Star* swung behind the campaign by printing a letter from Panamint describing the richness of the ores that waited only for a wagon road to allow exploitation. "Work will commence as soon as provisions arrive from your city," the writer declared. "Your business men are not alive to their own welfare, if they do not take some interest in this newly discovered Golconda."

By the end of the month the necessary funds had been raised and Jacobs had ridden for Panamint. By mid-June of 1874 his laborers, toiling with shovels and blasting powder on the Slate Range cliffs, had completed the road to Panamint Valley; a few days later Bart McGee and his partners finished the Surprise Canyon section, and Panamint was at last connected by road with Los Angeles.

Thus through the fall of 1874 the adventurers of California swarmed through Los Angeles, Bakersfield, and San Bernardino, converging along dust-marked trails to Panamint. In Los Angeles the strike had revived talk of local riches, had sent men combing the nearby mountains for the elusive metal. The city's streets were alive with miners displaying rock samples and extolling the richness of their own claim "in the hills."

"Almost every other man we meet has a chunk of ore," observed the *Star*. "Prospectors are coming and going daily."

But it was Panamint itself, looming as another source of Inyo silver, that fired Angelenos. "The talk is Panamint, and nothing but Panamint," recorded the *Herald*, adding that "the exodus to Panamint continues in undiminished numbers." In San Bernardino, whose location at the foot of Cajon Pass placed it at Panamint's back door, excitement ran still higher. "The air is redolent with Panamint," exclaimed the San Bernardino *Argus*. "Everybody talks Panamint — the young and old, men and women . . . " By October the editor was complaining, "The Panamint excitement has carried away so many men from this valley that laborers are in great demand."

From Los Angeles the Argonauts followed the Bakersfield stage road as far as Elizabeth Lake, then pushed on across the Mojave to join the contingents from northern California at Walker Pass. The fortunate ones rode horseback; other trudged along under the protection of

the Cerro Gordo mule teams. East of Indian Wells the traffic was so great that a traveler moving against the tide was always in sight of another silver seeker. At the foot of Surprise Canyon they joined an even greater horde of prospectors who had tramped the length of Inyo County — the Nevada boys from turbulent Columbus, Austin, Eureka, Pioche, and Virginia City.

Early in November a wagon road from Owens Valley through the Coso and Argus ranges to Panamint Valley was finished by John Shepherd, a pioneer settler who had brought his family to Inyo during the Indian wars. Without delay Panamint's first stagecoach, a four-horse Concord, lurched over the route and was soon making regular trips twice a week, loaded inside and topside with silver seekers. By the second week in November another six-horse line was bringing Argonauts eastward three times a week from Indian Wells, where it connected with the Bakersfield - Lone Pine stages and eventually with San Francisco and the Coast. On November 15 a four-horse stage began a weekly run through Cajon Pass from San Bernardino, and nine days later another weekly line opened from Los Angeles to Panamint via San Francisquito Canyon, Indian Wells, and Borax Lake. Within a month the bustling camp had been converted from an isolated locality, reached only by horseback and buckboard, to an established terminus receiving seven stages a week along three converging roads.

Meanwhile, a row of stores and saloons, many of them built by merchants from Lone Pine and Independence, lined Panamint's main street by October. Between 700 and 800 men filled the canyon, living in canvas tents, rude cabins, and in some cases even caves in the mountainsides. Lots were selling from $500 to $1,000, title resting not so much in a written deed as in a well-loaded shotgun. Senator John P. Jones had arrived to take a hand in the operations. When buxom Martha Camp arrived from Nevada with a bevy of frilled femininity and the nights soon stirred to the harp and fiddle, the boys hailed Panamint as a full-fledged camp.

By the end of November, Panamint's main thoroughfare was a mile in length, extending from Jacobs's mill at the lower end to the Surprise Valley company's store at the other. The muddy street was swarming with pack burros and mule teams, jostling each other in constant passage. Altogether some fifty buildings filled the canyon, half of them constructed of finished Owens Valley lumber. The rest were log or rock shanties, whose inhabitants were determined to shiver through the approaching winter in order to be on hand for the expected feverish activity of the coming spring. Newer arrivals simply rolled up in their blankets in the stores and saloons at night, grateful to their obliging owners for a space on the floor.

Two drugstores, three barbershops, three bakeries and restaurants, a livery stable, a meat market, and a boot shop flanked the main street. Half-a-dozen general merchandise stores were supplying necessities at such prices as $2.00 per dozen for eggs and $200 per ton for hay. A dozen saloons, most of them lining the south side of the street, were jammed day and night while the poker pots regularly ran into the thousands of dollars. Foremost was the elegant Oriental Saloon, whose billiard table, black walnut bar, and eight-by-six-foot mirror were advertised as "the finest on the Coast outside of San Francisco."

In another section of the same building was an institution of which even Cerro Gordo, queen camp of Inyo, could not boast — the Bank of Panamint. It was soon the business center of the town and the point of departure for the stage lines.

Other metropolitan necessities were also being established in Surprise Valley. In the center of town a log-cabin brewery was erected, while up in Sourdough Canyon a cemetery was improvised when two luckless combatants, killed by bullet and knife, required simultaneous burials. Finally a tri-weekly newspaper appeared on November 26, 1874; a Sacramento newsman named T. S. Harris had arrived and, while the first snowfall blanketed the camp, set up his hand press and type cases in a canvas tent and founded the robust Panamint *News*.

By the end of spring Panamint was almost ready for full scale production. A cable tramway, supported by great wooden arches and equipped with iron ore buckets, was completed from the Wyoming mine to the reduction works below in June 1875. On the twenty-ninth the new mill started up, its two red brick chimneys pouring out smoke and its twenty stamps dancing in measured succession. Two months later the furnace began turning out silver bullion worth some thirty thousand dollars per ton.

The temptation was too great for several hard-looking Panaminters, who arrived for the occasion heavily armed, ready to make off with the first silver ingots. Months before they had sold Senator Stewart one of the Panamint mines. But their frequent inquiries on the progress of mill construction indicated more than the customary interest of former owners. When the first silver was tapped from the smelter, Stewart was on hand — and prepared. Arrogance turned to chagrin as the desperados found the silver molded in pigs weighing four hundred to five hundred pounds apiece. Turning on the grinning senator, they called him the meanest man in the Panamints.

Throughout Inyo County, in fact, new camps were sprouting in the middle 1870s — seemingly wherever the irrepressible miner struck his pick. South of Owens Lake in the Coso hills rich silver-lead deposits were discovered late in October 1874, and by the year's end the town of Darwin had sprung up at the foot of Mount Ophir. Heading the first arrivals from Owens Valley was Abner B. Elder, early Cerro Gordo enterpriser, who was promptly elected recorder for "New Coso" mining district. Victor Beaudry, close at his heels, made haste to buy a nearby spring, lay pipes into Darwin, and become "water king" of the new camp. In December lawyer Pat Reddy rode in from Independence and for ten thousand dollars bought the Defiance mine, hailed as the real bonanza of the district.

Near the end of the year the Panamint contingent began to arrive. From that lofty camp, where snow was deep and thermometers were sagging, some three hundred men trudged down Surprise Canyon for the new Silverado within three days. In January 1875 the nearby Owens Valley—Panamint stage route was modified to include Darwin, and by April the main stage line entering Owens Valley from the south was diverted at Indian Wells to avoid Little Lake and take in the new camp.

While Darwin was flourishing, all Inyo was fairly sprouting with silver strikes. Besides the established camps of Cerro Gordo, Panamint, and Darwin, hundreds of miners were digging silver ore at Wauboba District opposite Big Pine in the Inyo Range, at Wildrose Canyon ten miles north of Panamint, and at Lee District on the trail between Owens Lake and Death Valley.

Starting in May 1875, Lookout District sprang to life on the east side of the Argus Range, between Darwin and Panamint. Here the Lookout, Modoc, and Minnietta were among a dozen silver mines yielding ore averaging two hundred to four hundred dollars per ton — so rich that at first it was packed by muleback to Panamint for reduction in the Surprise Valley mill and furnace. In the unlikeliest place for a town — on top of Lookout Mountain's east summit — the camp of Lookout soon grew to more than forty rock and wooden buildings, including two general stores and three saloons. Early in 1876 a group of San Francisco financiers headed by George Hearst bought the Modoc and other mines. The first machinery for the two furnaces was hauled in by Nadeau's teams over a rugged new toll road across the Argus Range from Darwin. By the fall of 1876 Hearst's two Modoc furnaces were shipping more than 300 bars of silver-lead bullion (similar to those produced at Cerro Gordo and Darwin) every day. The countless broken champagne bottles scattered today over the site of Lookout give evidence of the rich and lively times that ruled this mountaintop camp in the mid-seventies.

Gold in the California Desert
Past, Present, and Future

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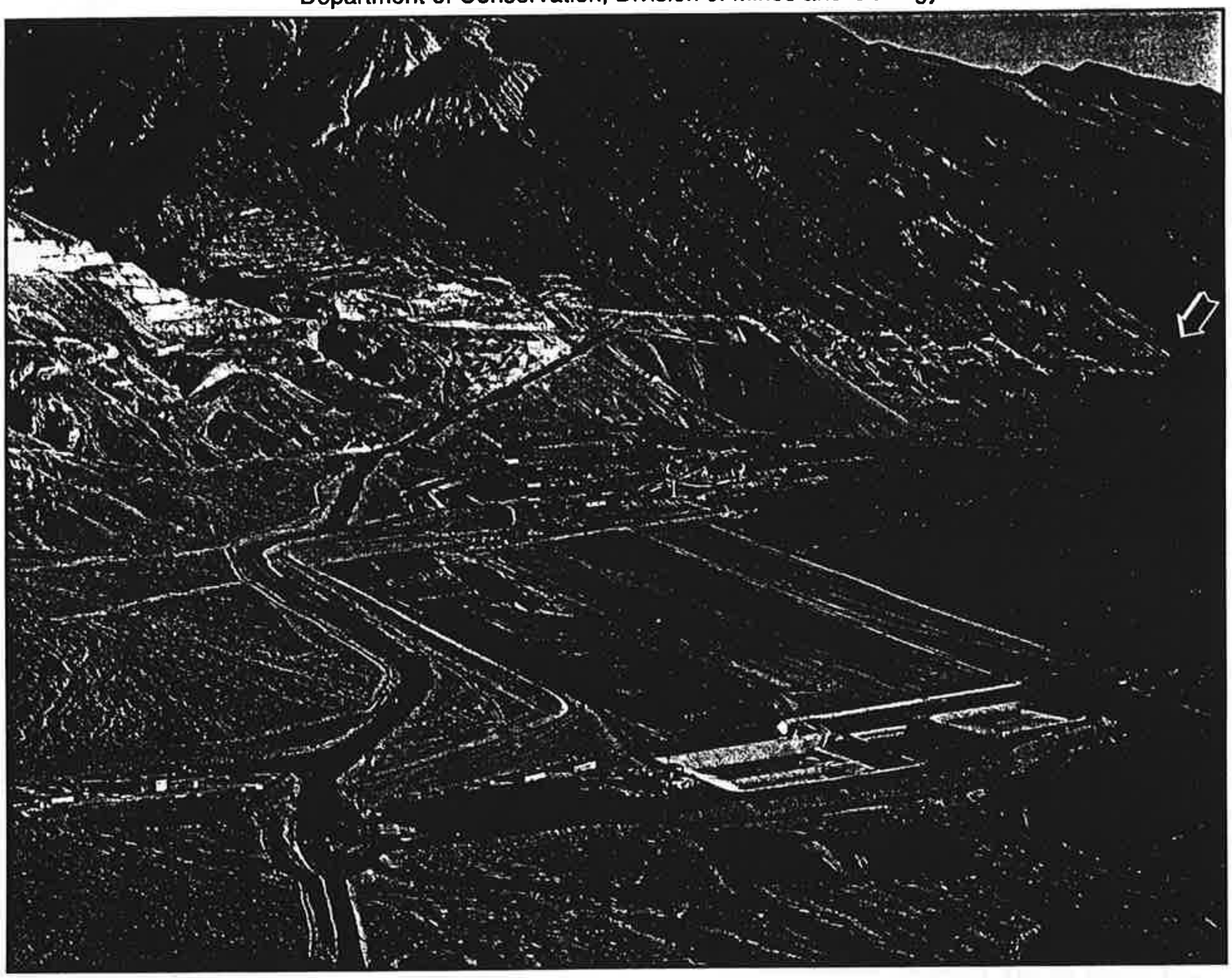

Photo 1. Oblique aerial view of the Briggs Mine, Inyo County, looking southeast. It is at the base of the Panamint Range in Panamint Valley about 8 miles south of Ballarat. Ore is mined from the benches cut out of the mountain slope. Gold is recovered through heap leach technology. Panamint Valley Fault Zone scarp (arrow) can be seen to the right of the mine tailings. *Photo taken 1997. © J.L. Christman, 1998.*

Recounting the history of gold mining in the California desert one can only observe that a dramatic surge in gold production occurred in the last 15 years, surpassing estimates from all previous years. It is considered the gold rush of the 1980s. The modern, large, open-pit mining operations (Photo 1) that developed are in stark contrast to mines that operated in these same mining districts prior to the 1980s. This article not only illustrates this contrast between post-1980 and pre-1980 mining activity, but also offers a future outlook for gold mining in this region...*editor*

HISTORIC OVERVIEW

An overview of gold mining activity from 1775 to 1980 in the California desert districts shown in Figure 1, as well as a summary of gold occurrence, was presented by Clark in 1980 (reprinted 1985). This article is an update to that 1980 report and offers some insight into the cause and extent of the gold rush of the 1980s in this region.

Gold production in the desert region was a cyclical phenomenon driven by gold prices and labor costs and reached peak levels during several time periods: the 1860s through 1890s, the early 1900s, and again during 1930s to early 1940s. Gold mining essentially stopped in 1942 when, on

October 8, the War Production Board Limitation Order L-208 closed gold mines throughout the United States. Although this order was lifted on July 1, 1945, large-scale gold mining in the California desert did not resume primarily due to the low price of gold coupled with the high cost of mining. Gold mining in the desert remained relatively insignificant during the mid-1940s to the early 1980s. Clark's 1980 article was still considered current when it was reprinted in 1985 (Clark, 1985) because the status of gold mining in the desert had not changed much between 1980 and 1985. He mentioned the increase in exploration for gold in the desert region that began in the late 1970s and early 1980s, and it appears that he anticipated its significance. In his introduction, Clark wrote:

> *"One can only imagine what the value of the desert's gold production would be at today's price"* (approximately $660 per ounce in July, 1980). In the closing remarks of his article, Clark also wrote: *"With the very large recent increase in price, there has been renewed interest in many of California's long-dormant gold mines* (Photo 2). *Over the years there has always been some intermittent prospecting in the desert regions, mainly by desert rats or small-scale placer miners and pocket miners. However, several mining organizations, including a number of the major*

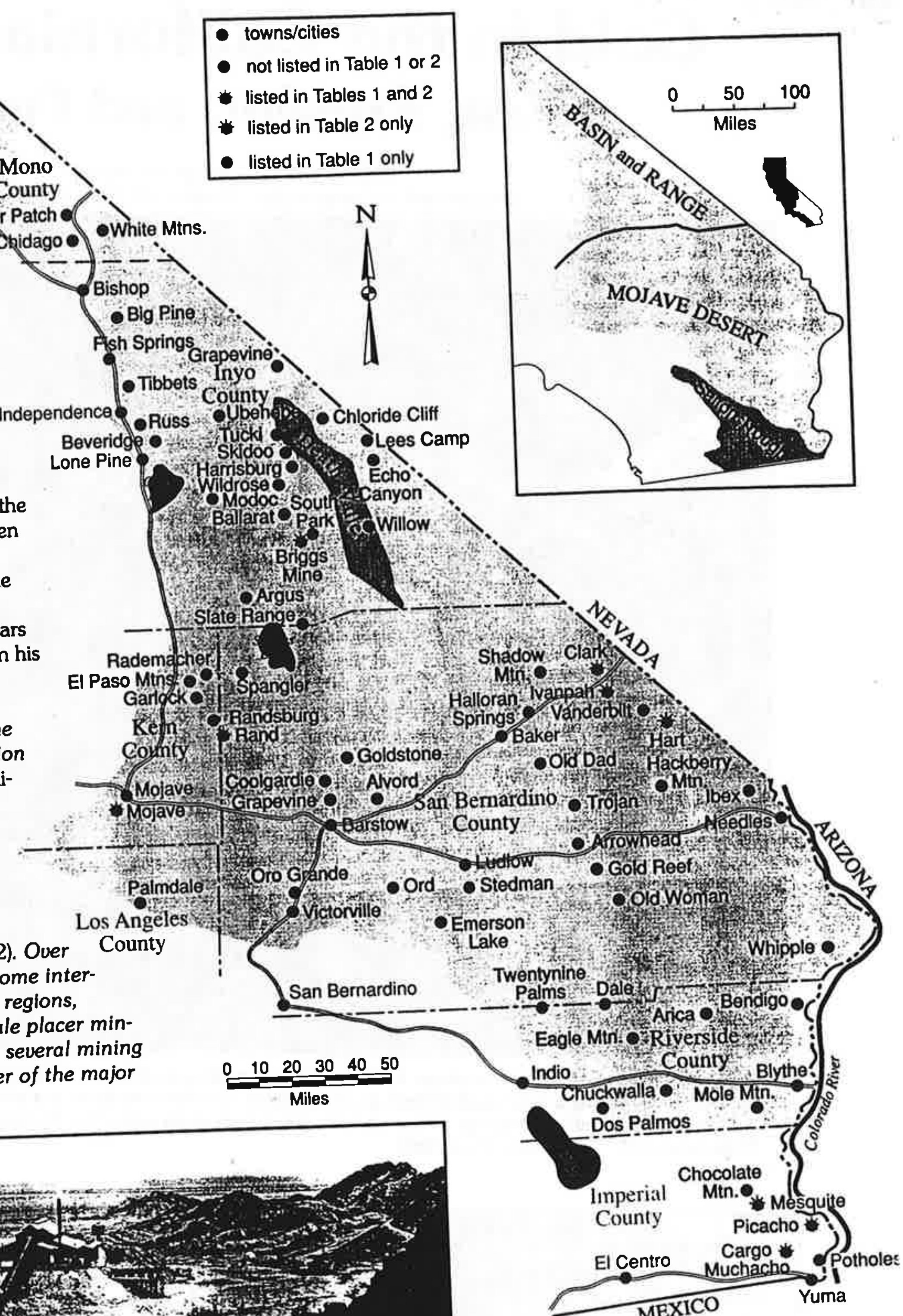

Figure 1. Map showing general location of the gold-mining districts in the desert portion of California. Inset shows the Basin-Range and Mojave geomorphic provinces. Major highways are also shown. Red circles denote towns. *Modified from Clark, 1985.*

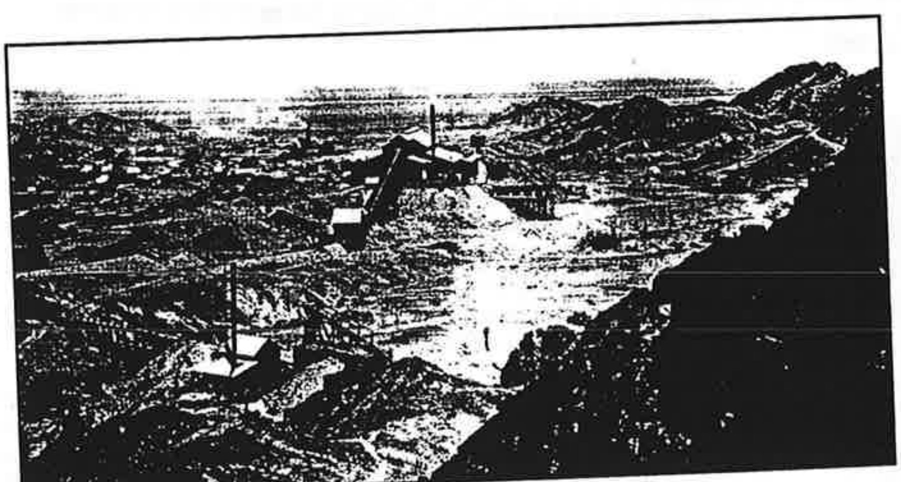

Photo 2. Golden Cross Mine, Imperial County in the Cargo-Muchacho Mining District, 1900. View is toward the west. *DMG archive photo A1017a.*

mining companies, are now actively engaged in prospecting for gold in this region. Although there is considerable activity, the details of many of these prospecting and development operations are not known.

There has been exploration work in most of the important districts (Table 1). This work has consisted of diamond drilling, detailed geologic mapping, and systematic sampling. The general geology of most of the districts is known, but detailed geologic studies of many of the important gold-mining districts and the deposits in them have not been done. One of the results of this so-called rush of the 1980s will be a greater knowledge of the geology and distribution of the gold deposits in the desert region of California."

As Clark anticipated, the rush of the 1980s did result in a greater understanding of gold deposits in the California desert region (Photo 3). What we have learned about the geology of the gold deposits in the desert region is the subject of a number of articles (Davis and others, 1989; Linder, 1989; Burnett and Brady,1990; Higgins, 1990; Pridmore and Frost, 1992) and will not be discussed in detail here. However, Table 2 has been added to summarize the location and extent of gold mining in the California desert since 1980.

Gold Production in the Desert Pre-1980

The record of early gold production in the desert region is incomplete, making it difficult to estimate total gold produced. The problem arises from the fact that gold production from many mines in operation during the 1860s to 1940s was never reported and of those reporting mines the amounts that were reported are incomplete. However, using Clark's 1980 gold production values (and knowing the price of gold at the time of the gold production) it is estimated that a minimum of 3.4 million troy ounces of gold were produced from the **major** mines during 1860 to 1942. If gold production from the smaller mines is included, total gold production for this period is moreclosely estimated to be in the range of 3.5 to 4 million troy ounces. Although gold production did continue after this time, the amount produced between 1942 to 1980 was insignificant and not considered in these estimates.

Clark (1980) presented gold production in terms of the dollar amount of gold (Table 1) based on the **value of gold at the time it was mined**. Except for the estimate of the total amount of gold produced (troy ounces) prior to 1980 just mentioned, no attempt has been made here to estimate the amount of gold produced from each of the mining districts.

The price for gold was set by the U.S. Treasury in 1834 at $20.67 per ounce and this price continued until 1934. From 1934 to 1968, the price was fixed at $35 per ounce. Since 1968 the U.S. Treasury no longer controls the price of gold, and it has risen greatly since then. Part of the production value of some mines listed in Table 1 was calculated at $20.67 per ounce and part was calculated with gold valued at $35 per ounce, particularly mines in the Rand, Mojave, Dale, and Cargo Muchacho districts. Those districts had appreciable gold producers during the 1930s and 1940s.

Post 1980

By comparison, gold production from 1980 to 1997 is well documented and amounts to approximately 5 million troy ounces. Table 2 shows the amount of gold (troy ounces) produced from the mines during 1980-1997. Totals are given in ounces, a better indicator of production due to the large fluctuations in gold prices that reached a high of $850 per ounce in 1980 and a current low of about $287 on December 8, 1997. This modern production came from nine historic mining districts that were also active at various times during the late 1700s to the early 1940s (Photos 4 and 5). Five mines continue to actively mine in the desert region (Photos 6, 7, and 8) with exploration and development of new gold deposits continuing.

1980s Gold Rush

The gold rush of the early 1980s was the direct result of increased gold prices. This heightened interest in gold stimulated the application of large-scale (open-pit) mining techniques that targeted large tonnage, low-grade (on the order of 0.05 troy ounces of gold per ton of ore), gold deposits. Concurrently, highly efficient metallurgical extractive methods developed to recover highly disseminated gold from ores that were previously unminable, not only allowed an economic way to recover gold from this region of California, but it led to a worldwide surge in gold production during the 1980s.

MINING ACTIVITY AND GOLD OCCURRENCE IN THE DESERT REGION

The following discussion of gold mining activity and geology of some of the important gold mining districts is modified from Clark (1980; 1985). Gold is widely distributed throughout the California desert region. There are many gold mining districts in the area; a substantial amount of gold has also been produced as a by-product of lead, zinc, silver, and copper mining. Gold is also found associated with several tungsten and iron deposits. Although primary or lode deposits have been the largest sources of gold, appreciable quantities of placer gold have been produced from the Picacho, Potholes, Rand, Collgardie, Tumco and El Paso Mountains gold mining districts. The first known production of gold in California in the 1770s—some 80 years before the state's first gold rush—was from the Potholes placer-mining district in the area that is now Imperial County.

The California desert gold-bearing region includes both the Mojave Desert and the Basin-Range geomorphic provinces. Prior to 1980 the three most productive and best-known gold-mining districts were the Bodie District in eastern Mono County; the Mojave-Rosamond District in Kern County; and the Rand or Randsburg District, in both Kern and San Bernardino counties. Other important sources of gold have been the Stedman, Cargo Muchacho, Skidoo, Picacho, Potholes, Dale, and Masonic districts.

Table 2 shows that since 1980, the three most productive districts of the desert region have been the Mesquite District in eastern Imperial County, with a total production of 2.3 million troy ounces of gold from the Mesquite Mine; the Hart District in northeastern San Bernardino County, with a total of 755,000 troy ounces of gold produced from the Castle Mountain Mine; and the Tumco-Cargo Muchacho District in southeastern Imperial County, with a total of 479,880 troy ounces of gold produced from the American Girl Mine.

Type of Gold Deposits Sought in the Early Days of Mining

The geology of gold deposits in the California desert varies considerably (Figures 2, 3, and 4). The primary lode deposits occur in a variety of rocks ranging in age from Pre-Cambrian to Tertiary and hosted in diverse geologic structures. Also, placer gold has been produced from several different types of placer deposits. Prior to the gold rush of the 1980s the most common primary deposits mined in the desert were the gold-bearing quartz veins, but the primary deposits that yielded the largest amounts of gold were extensive zones of silicified breccia. Much of the production in these early years that came from the three large desert gold districts, Bodie, Mojave, and Rand, were produced from silicified breccia.

Silicified breccia deposits are tabular or vein-like bodies that range from a few feet to 50 or more feet in thickness and have been mined to depths exceeding 1,000 feet. These deposits occur in schist, quartzite, granitic rocks, and several types of volcanic rocks including rhyolite, andesite, and latite. Typically, they are associated with granitic intrusions. These deposits are composed mainly of quartz and chalcedony, but opaline material and calcite often are present. The angular breccia fragments are siliceous material but there may also be wall rock fragments and fault gouge present. The ore may be banded or have an open porous appearance with small cavities lined with euhedral quartz crystals. The ore bodies contain native gold in fine disseminated grains. Varying amounts of sulfides are present, of which pyrite (FeS_2) is the most abundant. Also chalcopyrite ($CuFeS_2$), galena (PbS), and arsenopyrite (FeAsS) may be present.

Some of the silicified breccia deposits have been the sources of substantial amounts of silver, and silver-bearing minerals such as cerargyrite (AgC), argentite (Ag_2S), proustite (Ag_3AsS_3), and pyrargyrite (Ag_3SbS_3). Although much of the gold occurs as very fine grains that are disseminated throughout the ore bodies, in places, high-grade ore has been found where the gold is coarse. Secondary iron oxide is commonly present and stains the ore and enclosing rocks various shades of red. Dark manganese oxide staining also may be present.

The most common, and also the most widely distributed, primary desert deposits are gold-bearing quartz veins. A typical deposit of this type is a vein or a series of milky-white to light-gray quartz stingers that range from 2 to 10 or more feet in thickness and are steeply dipping. They most commonly are found in various types of metamorphic or granitic rocks; a few are found in volcanic rocks. The metamorphic rocks usually are mica schists, quartzites, and gneiss, but in places, veins occur in metamorphosed dolomitic limestones.

The gold occurs in the native state in fine to coarse grains, and it often is associated with sulfides of which pyrite is most common. The sulfides also may contain gold. Fine-to medium-grained aplite, diorite, and quartz-diorite dikes are commonly associated with gold-quartz veins, and they may play a role in the localization of ore bodies. The veins usually extend to depths of several hundred feet, but several extend 1,000 or more.

Other primary deposits in the California desert include gold-bearing seams or fractures that have no vein filling, gold-quartz-calcite veins, gold-quartz-barite veins, and gold-bearing quartz-chalcedony-jasper veins. Also, gold has been found in tungsten-bearing quartz veins and skarn. Significant amounts of gold have been recovered as a by-product from lead-silver and lead-silver-zinc replacement deposits at such districts as Darwin, Cerro Gordo, and Tecopa and from a number of copper deposits.

Placer gold deposits are found in most desert gold districts. In a few districts, such as the El Paso Mountains and Potholes, virtually all the gold produced was from placer deposits.

Gold placer deposits occur in sand and gravels on bedrock in the present washes and canyons and in older terraces and benches above the present canyons. The gold in these deposits was concentrated by the action of running water, probably much of it from desert cloudbursts. Much of the placer gold that has been recovered from the California desert was from small, one-man or two-man operations. When a deposit was exhausted, the miners would move on to another one, usually in the same district. The most productive dry placer-mining period was from the 1860s to the early 1900s.

Gold Deposits Mined Today

The gold rush of the early 1980s focused on the historic mining districts and on large, low-grade lode gold deposits containing on the order of 0.05 troy ounces of gold per ton of ore. Today's gold ore was considered waste prior to development of the cyanide heap leach method of gold recovery during the early to mid-1960s. The relatively large gold-bearing quartz veins mined by underground methods during the 1800s and early 1900s were not the focus of exploration during the 1980s. Instead, host rocks containing gold associated with sulfide minerals (commonly pyrite, FeS_2) disseminated in the host rock, or free gold in small quartz veinlets disseminated in the host rock were the preferred exploration targets. With sufficient tonnage and grade of ore, gold could be mined by open-pit methods, piled in heaps, treated with a dilute sodium cyanide solution to dissolve the gold (and silver), and recovered by passing the gold-bearing (pregnant) solution through activated charcoal filters. The cost of recovering the gold by this method varies depending on the nature of the ore and other factors, but is generally \$180 to \$270 per troy ounce. At a grade of 0.05 troy ounces of gold per ton of ore, it takes 20 tons of ore to produce one troy ounce of gold. The cut-off grade (the lowest grade of ore mined along with higher grade ore at a particular deposit) can be as low as 0.01 troy ounces per ton.

In some of the gold mining districts, the geology and ore-forming systems were not well understood prior to the exploration and mine development of the 1980s and 1990s. Modern exploration is based primarily on geologic and geochemical models developed and refined from ever-increasing understanding of geologic processes and ore-forming systems. New deposit types have been recognized in the desert region, including deposits associated with low-angle faulting (detachment faulting), such as the Picacho District in eastern Imperial County (Pridmore and Frost, 1992), and gold breccia pipe deposits such as the Colosseum Mine in eastern San Bernardino County (Davis and others, 1989).

FUTURE OUTLOOK

Some modern gold mines in the desert region, such as the Colosseum Mine (San Bernardino County), the Standard Hill and Cactus Queen-Middle Buttes mines (Kern County), and the American Girl Mine (Imperial County) have closed or have ceased mining but continue gold recovery from leaching operations or are undergoing reclamation. Others, such as the Picacho and Mesquite mines (Imperial County), the C.R. Briggs Mine (Inyo County), the Castle Mountain Mine (San Bernardino County), and the Glamis Rand Mining Company operations (Kern County) are still mining reserves, and it is anticipated they will continue for many years depending on remaining gold reserves, development of additional gold reserves, gold price, and other factors.

With the passage of the California Desert Protection Act in October 1994 (CDPA; PL103-433), 10,634,869 acres of federal wilderness, national parks, wildlife refuges and some state parks were created or expanded in the California desert (R. Waiwood, U.S. Bureau of Land Management, written communication,1997). Military reservations such as Fort Irwin near Barstow (San Bernardino County) are undergoing or are being considered for expansion; although many of these areas contain a variety of metallic and industrial mineral resources, including gold resources, mining is not permitted. However, mineral exploration and mining operations with valid existing rights at the time the CDPA passed can continue.

Approximately 13,781,700 acres of the California Desert remain open to mineral exploration and mining subject to various federal and state regulations (R. Waiwood, U.S.Bureau of Land Management, written communication,1997). However, because many areas with favorable potential for future development have been restricted as a result of the CDPA, the level of future exploration and gold mine development is expected to be substantially less than during the "rush" of the 1980s.

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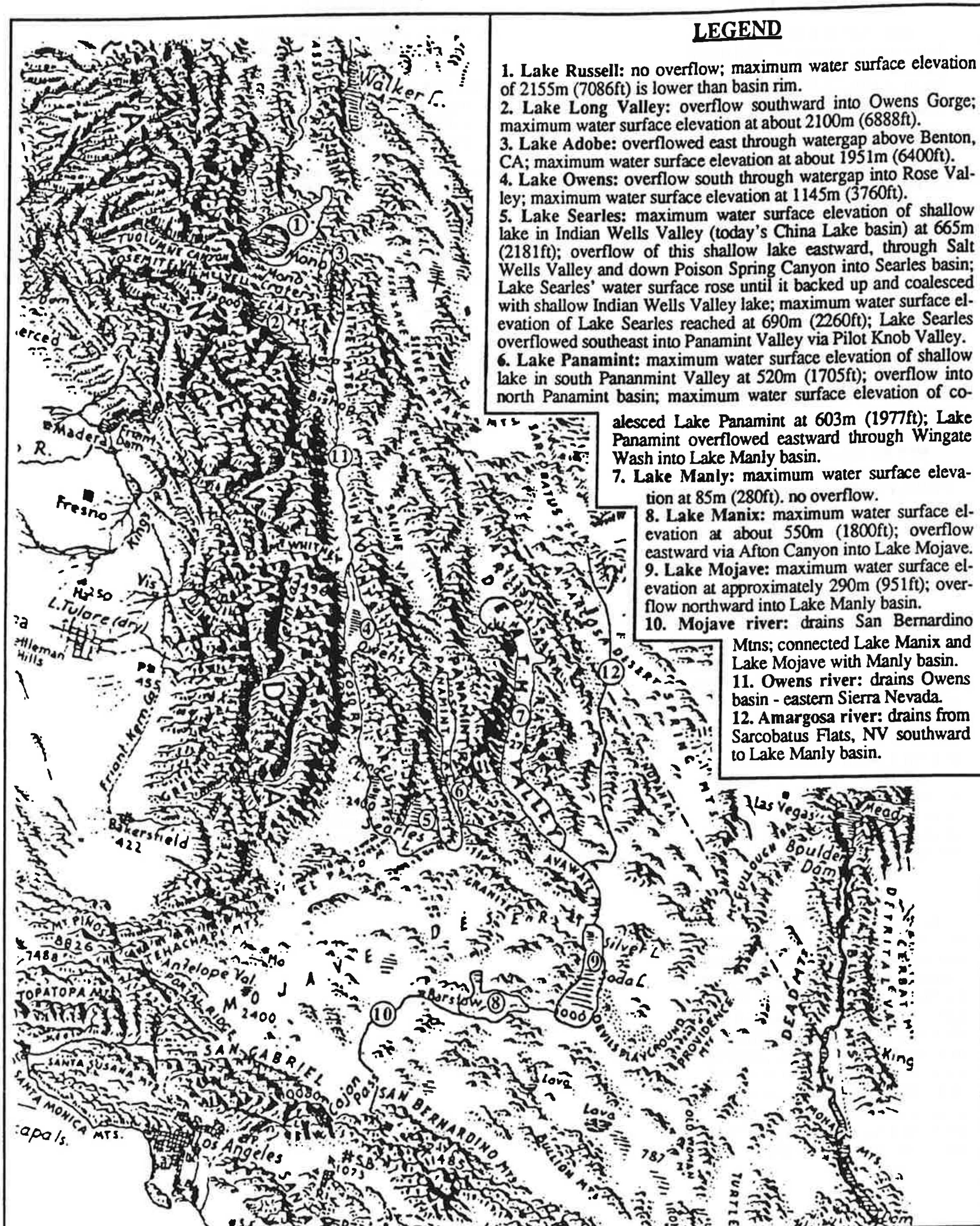

LEGEND

1. **Lake Russell:** no overflow; maximum water surface elevation of 2155m (7086ft) is lower than basin rim.
2. **Lake Long Valley:** overflow southward into Owens Gorge; maximum water surface elevation at about 2100m (6888ft).
3. **Lake Adobe:** overflowed east through watergap above Benton, CA; maximum water surface elevation at about 1951m (6400ft).
4. **Lake Owens:** overflow south through watergap into Rose Valley; maximum water surface elevation at 1145m (3760ft).
5. **Lake Searles:** maximum water surface elevation of shallow lake in Indian Wells Valley (today's China Lake basin) at 665m (2181ft); overflow of this shallow lake eastward, through Salt Wells Valley and down Poison Spring Canyon into Searles basin; Lake Searles' water surface rose until it backed up and coalesced with shallow Indian Wells Valley lake; maximum water surface elevation of Lake Searles reached at 690m (2260ft); Lake Searles overflowed southeast into Panamint Valley via Pilot Knob Valley.
6. **Lake Panamint:** maximum water surface elevation of shallow lake in south Pananmint Valley at 520m (1705ft); overflow into north Panamint basin; maximum water surface elevation of coalesced Lake Panamint at 603m (1977ft); Lake Panamint overflowed eastward through Wingate Wash into Lake Manly basin.
7. **Lake Manly:** maximum water surface elevation at 85m (280ft). no overflow.
8. **Lake Manix:** maximum water surface elevation at about 550m (1800ft); overflow eastward via Afton Canyon into Lake Mojave.
9. **Lake Mojave:** maximum water surface elevation at approximately 290m (951ft); overflow northward into Lake Manly basin.
10. **Mojave river:** drains San Bernardino Mtns; connected Lake Manix and Lake Mojave with Manly basin.
11. **Owens river:** drains Owens basin - eastern Sierra Nevada.
12. **Amargosa river:** drains from Sarcobatus Flats, NV southward to Lake Manly basin.

LATE NEOGENE EVOLUTION OF THE INDIAN WELLS VALLEY AND THE COSO RANGE

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INTRODUCTION

This trip is to examine the late Neogene tectonic development of the Indian Wells Valley and part of the adjacent Coso Range. The Indian Wells Valley (IWV) located in east-central California (Fig. 1) has been an active sedimentary basin throughout most of Tertiary time. During the Paleogene it was part of the Goler basin, and in the Miocene it received volcaniclastic and terrigenous sediment, strata of which are correlated with rocks of the Ricardo group. In the Pliocene it became the locus of deposition in a half-graben formed by low-angle normal faulting along the eastern front of the Sierra Nevada. The Coso Range to the north of the IWV was apparently a positive tectonic feature over most the Cenozoic time. Basin development began during latest Miocene time related to uplift of the Sierra Nevada.

Two significant structural/tectonic changes are recorded in the late Neogene stratigraphy of the Indian Wells Valley and Coso Range: one at the close of the Miocene, and the second at the end of the Pliocene. The late Miocene transition lead to east-west extension in the region. This occurred sometime between 7 and 5 Ma and resulted in the uplift of the nearby Sierra Nevada. Transition from an east-west-directed extensional regime to one dominated by transtensional dextral faulting occurred in the late Pliocene marking a change in regional stress. Dextral strike-slip faulting dominates the modern structural setting and controls sparse sedimentation in the IWV. This pattern of extension followed by transtension appears to be part of a progression that began in Death Valley approximately 16 Ma. Today the IWV and Coso Range accommodate a component of integrated transtensional shear, and are part of the evolving margin between the North America and Pacific plates.

The aim of this field trip is to exam the late Miocene to recent structural development of the Indian Wells Valley and western Coso Range. Most of the constraints on development of the IWV come from subsurface data (reflection seismic lines, refraction and gravity surveys, and drill holes). Development of the western Coso Range over this period are recorded in sedimentation of the Coso Formation. These relations are relatively well exposed. We consider the Coso Formation to provide an exposed look at rocks similar to those that fill the IWV.

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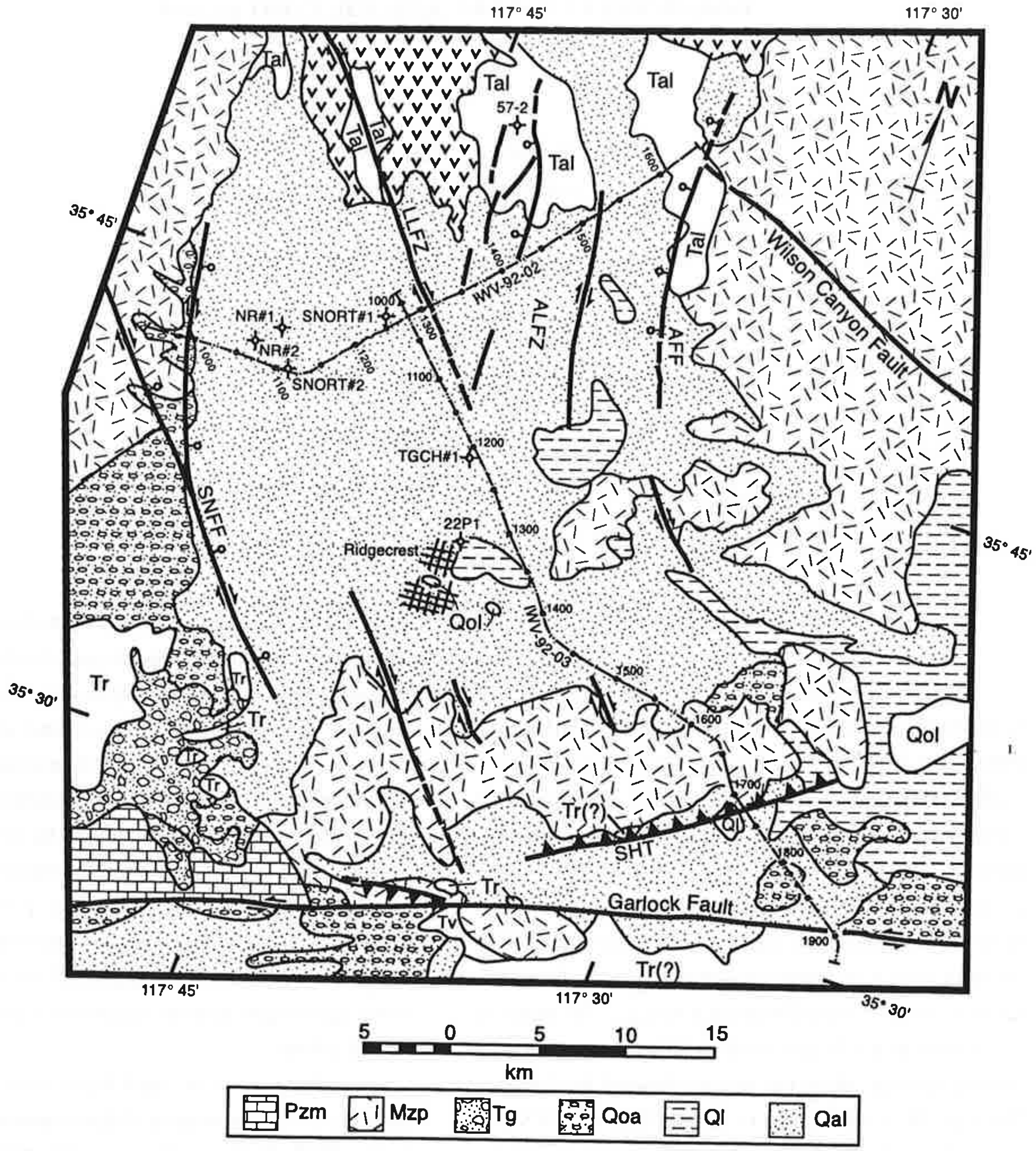


Figure 4. General geological map of the Indian Wells Valley and surrounding areas. Heavier lines represent faults, dashed where approximate. Dash and double dot lines are locations of seismic reflection lines discussed in the text and shown in Plates 1 and 2; numbers represent every 100th shot point. ALFZ = Airport Lake fault zone, AFF = Argus frontal fault, LLFZ = Little Lake fault zone, SHT = Spangler Hills thrust, SNFF = Sierra Nevada frontal fault. Geologic units (oldest to youngest) are: Pzm = Paleozoic metamorphic, Mzp = Mesozoic plutonic (undifferentiated), Tg = Paleocene to early Eocene Goler Formation, Tr = Ricardo Group consisting of early Miocene Cudahy Camp Formation and middle- to late-Miocene Dove Spring Formation, Tv = Miocene Lava Mountains volcanic, Tal = Pliocene White Hills sequence, Qpv = Plio-Pleistocene volcanic rocks of the Coso Range, Qoa = Older alluvium for which there is no conclusive data to permit assignment to a specific formation, Qol = older lacustrine for which there is no conclusive data to permit assignment to a specific formation, Qal = Quaternary alluvium, Ql = quaternary lacustrine. Basic geology from Jenkins (1963) with some areas remapped by Monastero.

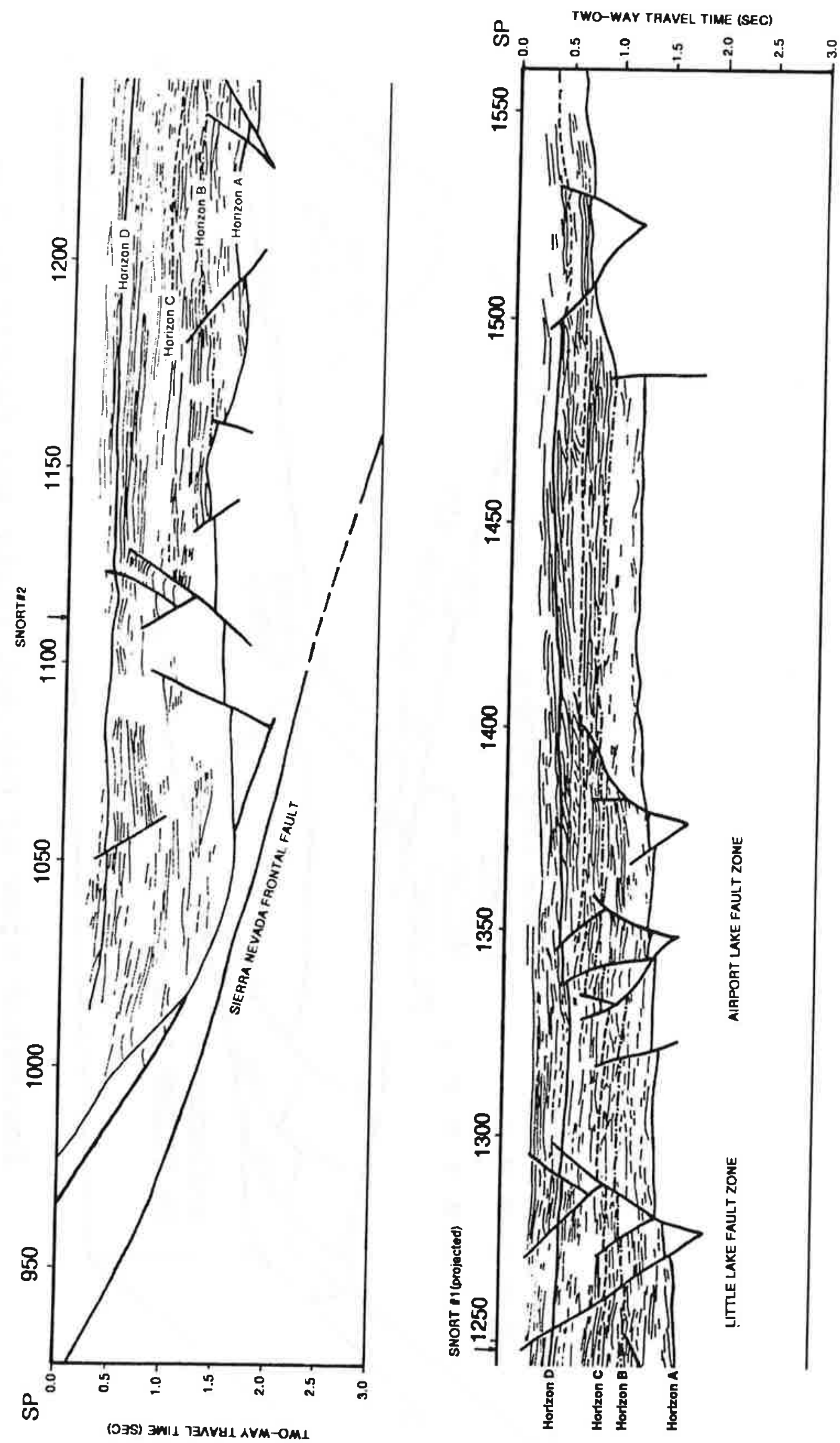
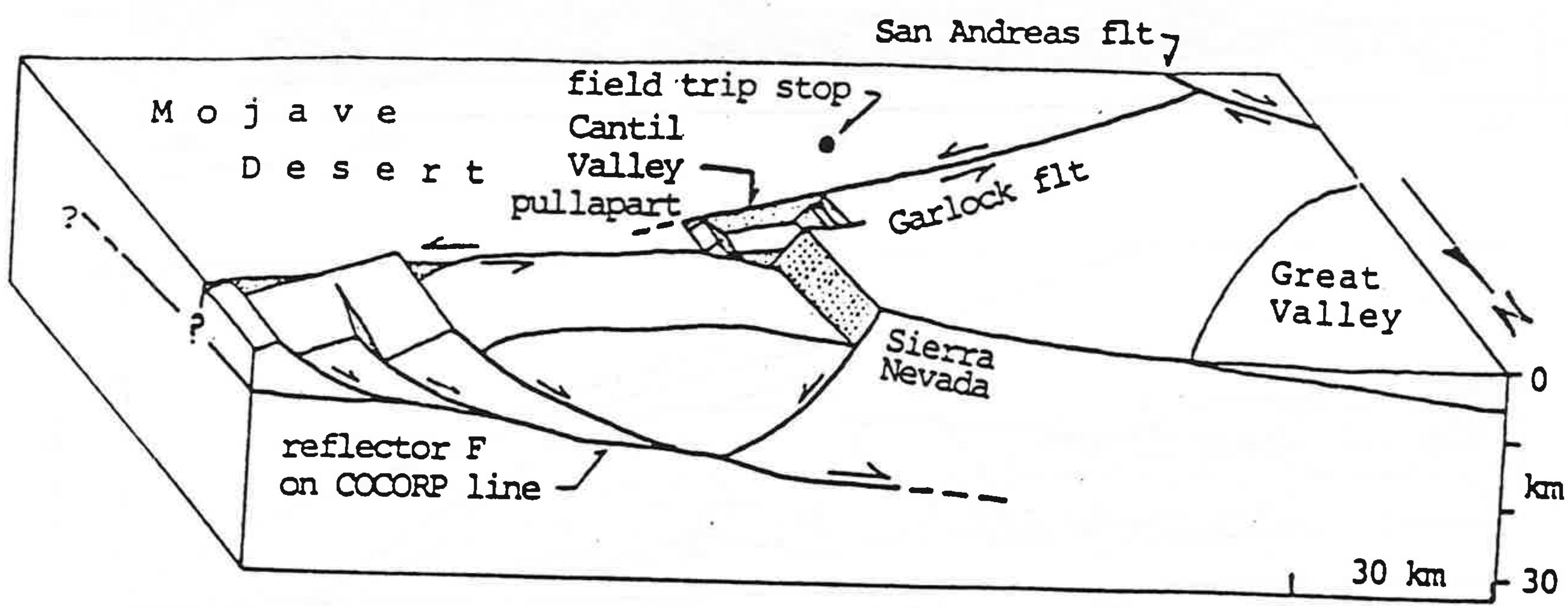


Figure 5. Line drawing of seismic section IWV-92-02.



Southward view of Garlock fault and Cantil Valley.

EL PASO MOUNTAINS TO SEARLES LAKE

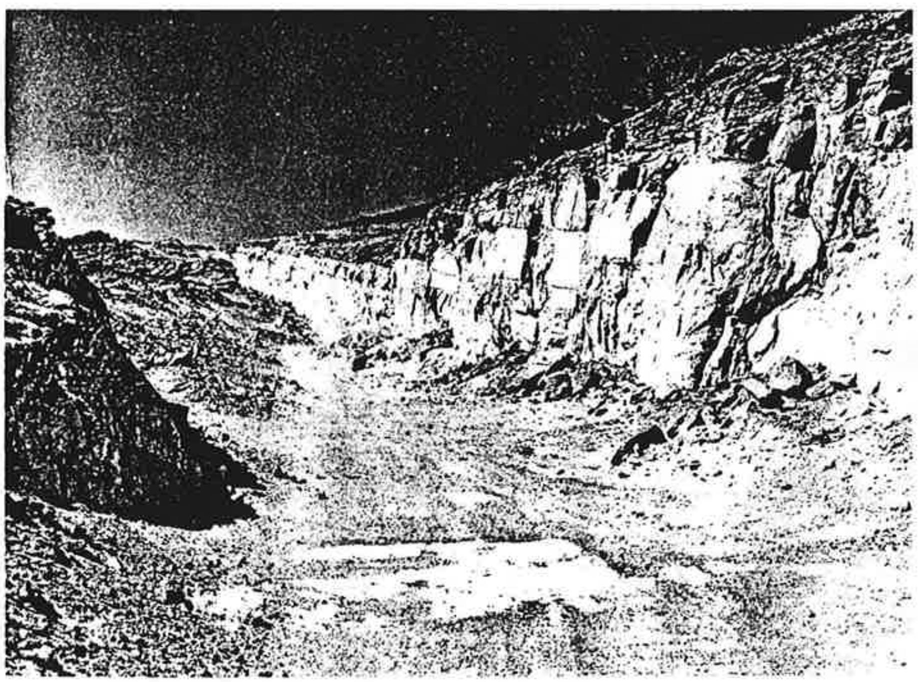
Map 4 (Figure 17-12) shows geology along the highways from El Paso Mountains to Searles Lake. In the eastern part of El Paso Mountains, 35,000 feet of Paleozoic chert, limestone, shale, and lesser volcanic rocks (*lP*—the Garlock Series) stand on end and strike north. Permian fusulina in the upper half of the section indicate its age, but strata in the lower half of the section may be much older. Paleocene nonmarine rocks (*Tem*) exposed in these mountains have yielded the oldest mammalian fossils in California.

A secondary road continues on to Randsburg, Johannesburg, and Atolia (the latter two on U.S. Highway 395)—all old mining towns. Randsburg yielded at least $12 million in gold during the period 1895 to 1918. The gold came from quartz veins in Mesozoic granitic rocks. Nearby Red Mountain is famous for silver, and Atolia for exceptionally rich tungsten deposits (scheelite) in quartz monzonite.

Turn off U.S. Highway 395 to the northeast, at Johannesburg, onto a secondary road toward Searles Lake. On the northern side of the low Summit Hills, the road crosses the Garlock fault, which makes the sharp contact between granite and valley alluvium. The road then crosses the low Spangler Hills, which lie at the south end of the Argus Range—one of the north-trending Great Basin ranges.

Approaching the town of Trona, the broad, white, salt-encrusted floor of dry Searles Lake lies on the right. During the several stages of glaciation in the Pleistocene Epoch the cool, humid climates caused freshwater lakes to form in the basins of the Great Basin and Mojave Desert (Figure 17-13). Death Valley—the lowest and largest basin—was the terminal lake for a series of drainage systems including the Owens, Amargosa, and Mojave Rivers and chains of freshwater lakes along their courses.

Searles Lake was the terminus during the last glacial epoch when it was probably a freshwater lake close to 400 feet deep and 16 miles long. Increasingly dry climate caused desiccation; and the lake today consists of a huge body of crystallized salts of sodium, boron, and potassium, with a dry surface and a saline-mud mush beneath. The lake is a major commercial source of



17-11 *Quarry in stratified Quaternary perlite deposits in the El Paso Mountains of Eastern Kern County. (Charles W. Chesterman photo.)*

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borates, lithium salts, potash, soda, trona (sodium carbonate-bicarbonate), common salt, and other salts. Patches of the Pleistocene lake deposits are visible above the present valley floor, as are also well-preserved shore-line terraces. A wonderful group of 100-foot high, calcareous tufa pinnacles can be seen in the lake basin 5 miles south of the highway. Searles Lake deposits yield 30 percent of the annual production of borates in California and all the potash produced in the state.

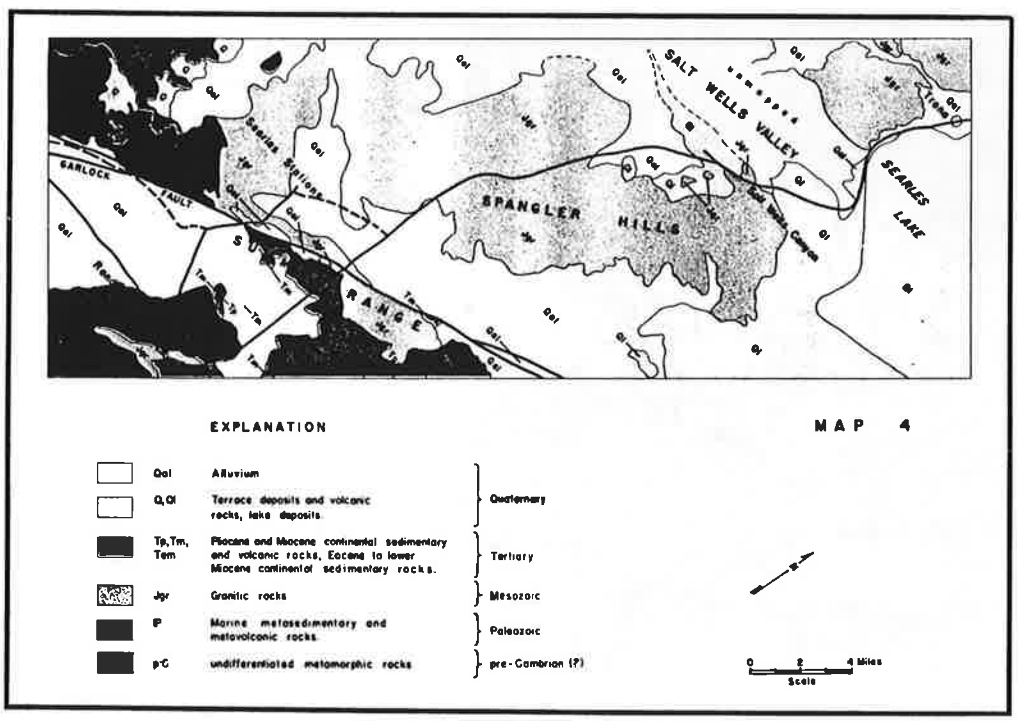
SEARLES LAKE TO THE CREST OF THE PANAMINT RANGE

From Searles Lake to the Panamint Range the geologic features may be followed on Map 5 (Figure 17-15). About 12 miles north of Searles Lake, the road crosses over a low pass between the Argus and Slate Ranges, typical Great Basin ranges. In the pass itself are exposures of stratified Paleozoic rocks intruded by granite. Straight ahead is a marvelous panorama across the Panamint Valley to the great fault scarp that marks the western base of the Panamint Mountains. This is an active fault scarp, along which the valley was downdropped and the range elevated in very late geologic time. The low benches seen are Pleistocene terraces and shore lines representing fluctuating levels of the freshwater lake that once occupied Panamint Valley.

Exposed along the western side of the Panamint Range are 8,000 feet of Cambrian, Ordovician, Silurian, and Devonian strata (Illustration 7-2). These strata lie on Precambrian metasedimentary rocks seen in the lower part of the range.

Evidence of the recency of movement in the Panamint fault zone is a mile-wide, downfaulted trench in alluvium—the Wildrose graben at the mouth of Wildrose Canyon. The road crosses the graben with its two bounding faults 200 feet high. A couple of miles beyond the graben are

FIGURE 17-12 *Map 4 showing geology along paved secondary roads from Randsburg to Trona on Searles Lake.*



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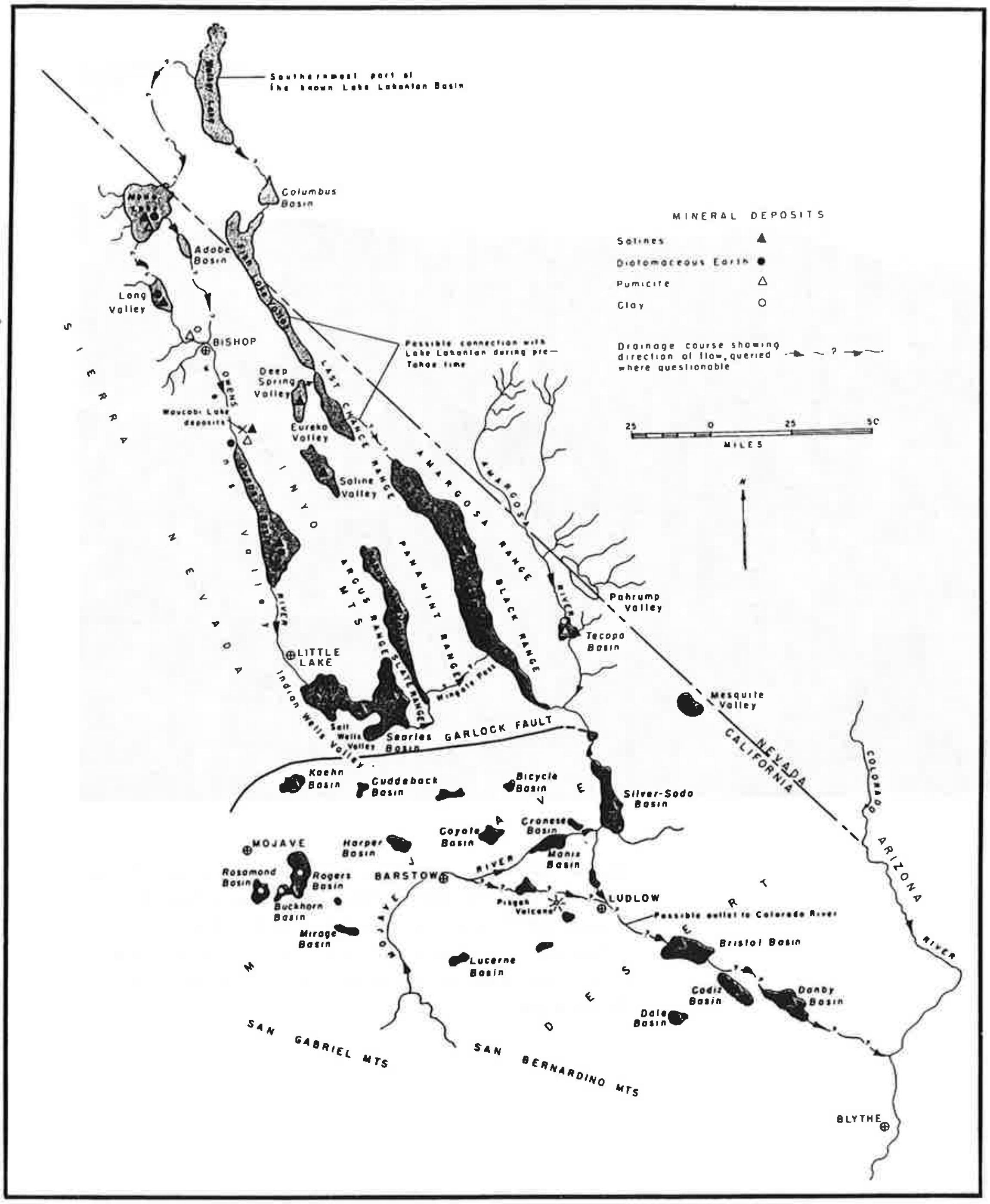

FIGURE 17-13 *Late Pleistocene freshwater lakes and rivers in the Great Basin and Mojave Desert. (California Division of Mines and Geology,* Mineral Information Service, *April, 1961.)*

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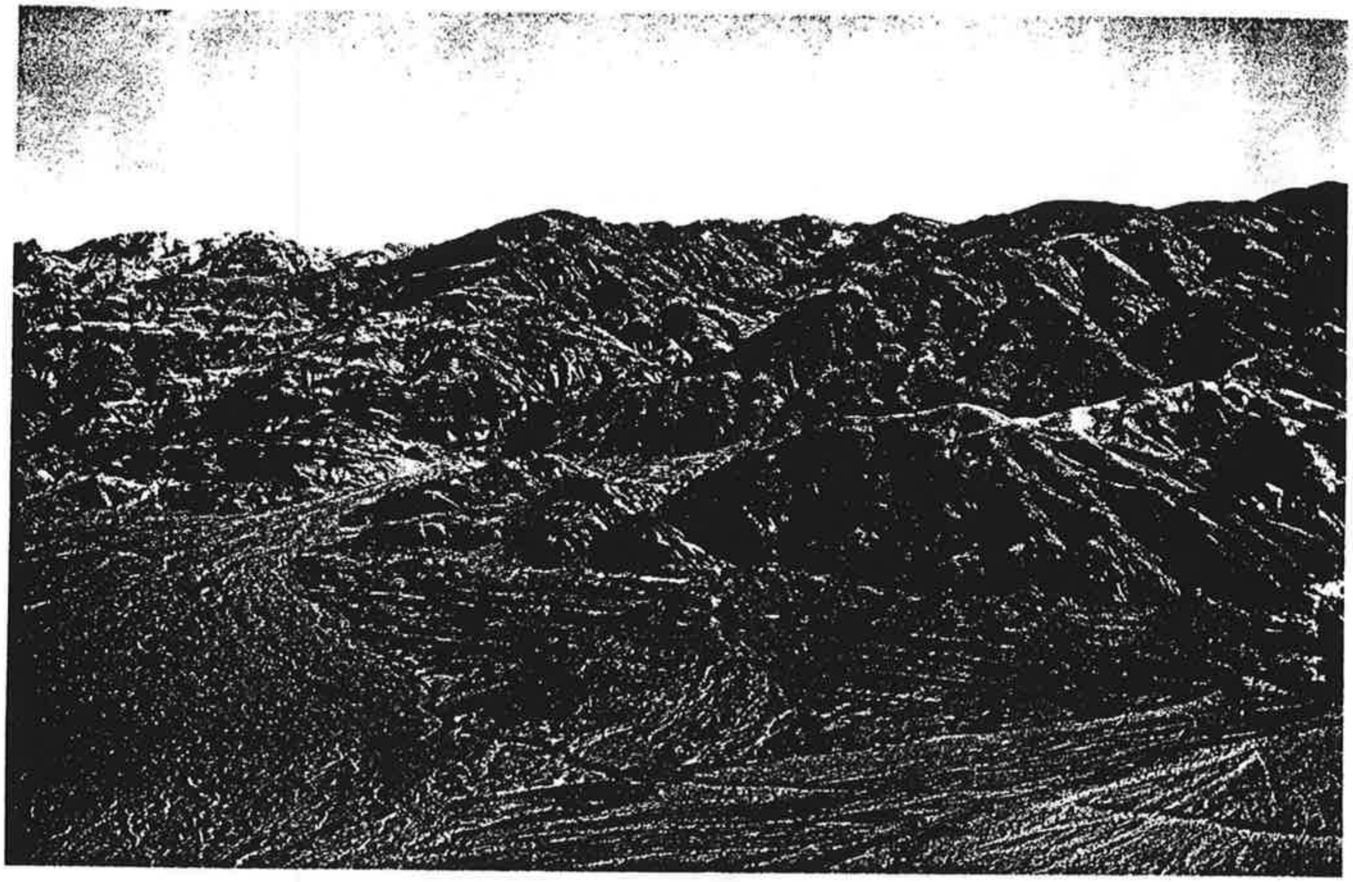

17-14 *Looking east at Paleozoic strata in the Slate Range (see Figure 17-15). Slightly tilted shore lines of late Pleistocene Lake Searles are near center of the photo. Just below the shore lines is a curved fault scarp so recently formed that it has cut deposits of the alluvial fans. (B. W. Troxel photo.)*

exposures of tilted beds of a coarse fanglomerate—the late Tertiary Nova Formation (*T*)—which is seen lying unconformably on folded older Tertiary gravels and Precambrian rocks.

WILDROSE TO THE FLOOR OF DEATH VALLEY

Map 6 (Figure 17-16) shows roadside geology from Wildrose Canyon to Death Valley. Up Wildrose Canyon the road passes through earlier Precambrian (Archean) granite gneisses, overlain by later Precambrian (Algonkian) Noonday dolomite (Photograph 6-6). Six miles past Wildrose ranger station (entrance to Death Valley National Monument) are several beehive charcoal kilns that were built in the 1880s to make charcoal for a smelter at the Modoc lead-silver mine in the Argus Range. Strata of the Noonday dolomite cross the canyon at the kilns. The flat here offers the first spectacular view of Death Valley and the Funeral and Black Mountains across the valley (Photograph 3-12).

North of Emigrant Pass, the road crosses Harrisburg Flat, an old, uplifted erosion surface. A well-marked turnoff leads 8 miles east to Aguereberry Point, noted for its spectacular view of Death Valley. This great structural and topographic, fault-bounded trough stretches north-south for 150 miles and east-west for only about 10. Stratified rocks on the road are part of the lower Cambrian section. Looking eastward from the point, successively higher formations in the Paleozoic section are seen.

Below Emigrant Spring, after the road joins State Highway 190, the steep western face of Tucki Mountain exposes fanglomerate beds of the Nova Formation and cliff-forming interlayered basalt flows. Tucki Mountain consists generally of east-dipping later Precambrian and Paleozoic strata that form an unusually complete section.

Stovepipe Wells hotel and resort were named after an early-day well, concealed by sand but kept open by a stovepipe driven through the sand. Downhill from Stovepipe Wells and extending

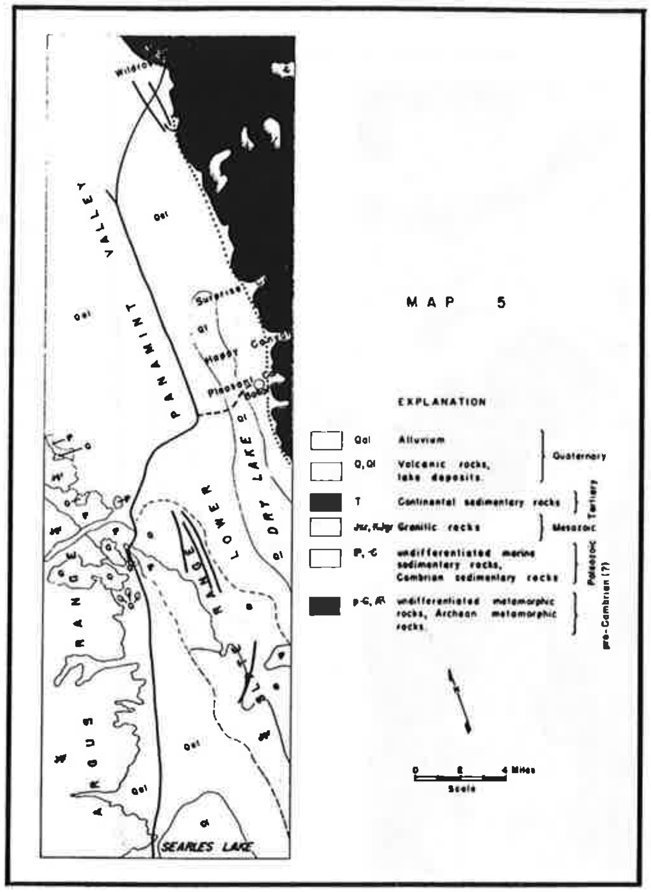

FIGURE 17-15 *Map 5 showing geology along paved secondary road from Searles Lake across the Argus and Slate Ranges to Wildrose Station in the Panamint Range.*

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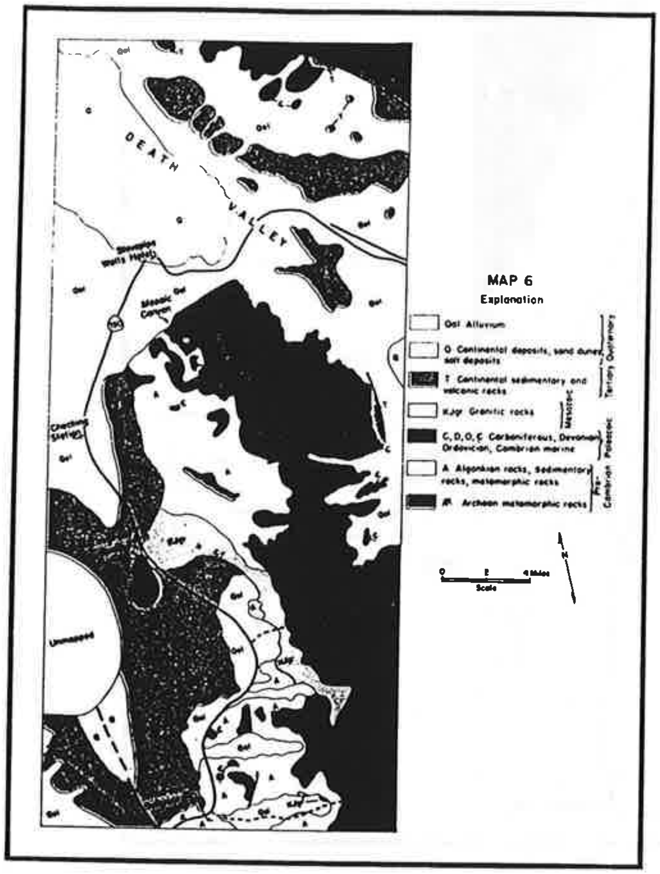


FIGURE 17-16 *Map 6 showing geology along secondary road and State Highway 190 into Death Valley by way of Stovepipe Wells.*

onto the floor of Death Valley are the famous, much-photographed Death Valley sand dunes. A marvelous view of the valley may be seen looking eastward from Stovepipe Wells; southward are seen the faulted later Precambrian, Cambrian, and Ordovician strata on Tucki Mountain. An interesting side trip involves walking into Mosaic Canyon where water-polished yellowish-brown dolomite is exposed on the canyon floor. On Highway 190, a couple of miles east of Stovepipe Wells, Cambrian and Ordovician formations can be seen in the face of Tucki Mountain.

the iron oxide that forms the tiny magnetite crystals is dissolved in the pumice glass rather than segregated into crystals. Second, the pumice, with its gas bubbles, contains myriad reflective surfaces that scatter incoming light; these surfaces disappear when the pumice is remelted. This is why foams are generally white, and even the foam of a dark beer is white. If this rock had crystallized slowly in the earth, forming a granite, its dominant color would have been light gray or pink instead, with the iron segregated into discrete, dark iron-bearing minerals.

Normal volcanic glass, called obsidian, forms by rapid chilling of molten lava, mostly of rhyolitic composition, as it is extruded (vignettes 29 and 30). That origin is not consistent with the relationships seen here. The interior of an ash body could not possibly cool more quickly than the top or bottom. The black glass layer must have formed by localized sintering and re-fusion of glass particles within the ash. We can confirm this through microscopic examination of ultrathin sections of our black glass layer, which reveal remnants of the fragmental texture of the original ash. Obsidian formed by chilling of molten lava displays no such textures.

Formation of this striking glass layer probably proceeded as follows. The eruption blew a thick pile of hot, gas-rich pumice over the countryside. The pumice fragments were hot enough (about 1,000 degrees Fahrenheit or more) that the glass particles began to compact and weld together. This process stopped early at the top and bottom of the layer because these parts cooled quickly, but in the center of the deposit the process continued until nearly all the porosity was squeezed out of the pumice, which fused into black glass. Hot gas from the crushed pumice filtered up, creating vesicles in the upper part of the glass layer and in the immediately overlying tuff, as well as depositing mineral coatings on the walls of those vesicles.

The densely welded, black glass layer is clearly not horizontal, but inclines about 15 degrees to the west. This could be a result of tectonic tilting of an originally horizontal layer, but the flattened pumice blocks and vesicles tell a different story. You have probably noticed that they lie at an angle to the densely welded black glass layer. Pumice compacts in response to gravity, so the little pumice disks were originally horizontal; they are now inclined eastward about 13 degrees. The cooling that controlled the location and size of the densely welded glass layer, however, was controlled by the orientations of the upper and lower surfaces of the tuff. Therefore, we can tell that the tuff was deposited on a sloping surface and later tilted by faulting or folding.

This big roadcut is a favorite stop for geological field trips. You can now understand why. It taxes the imaginations of students and challenges their instructors.

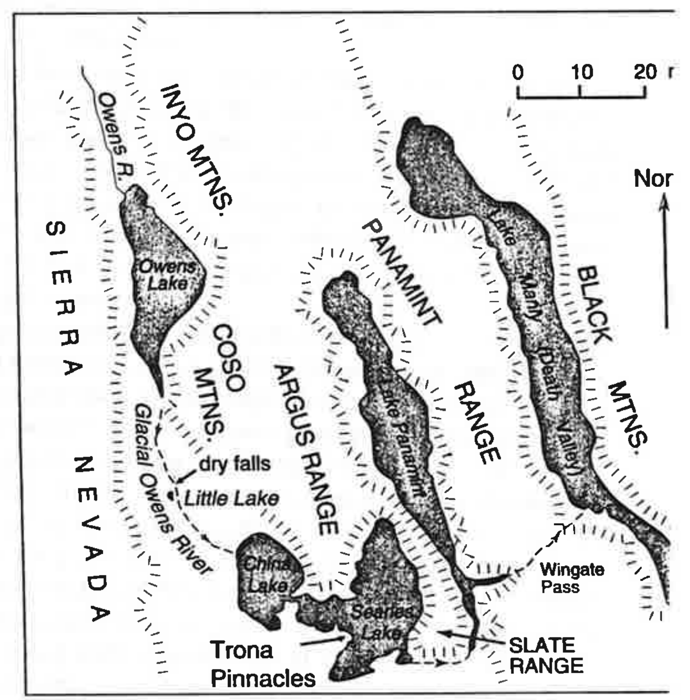
4

A Lunar Landscape

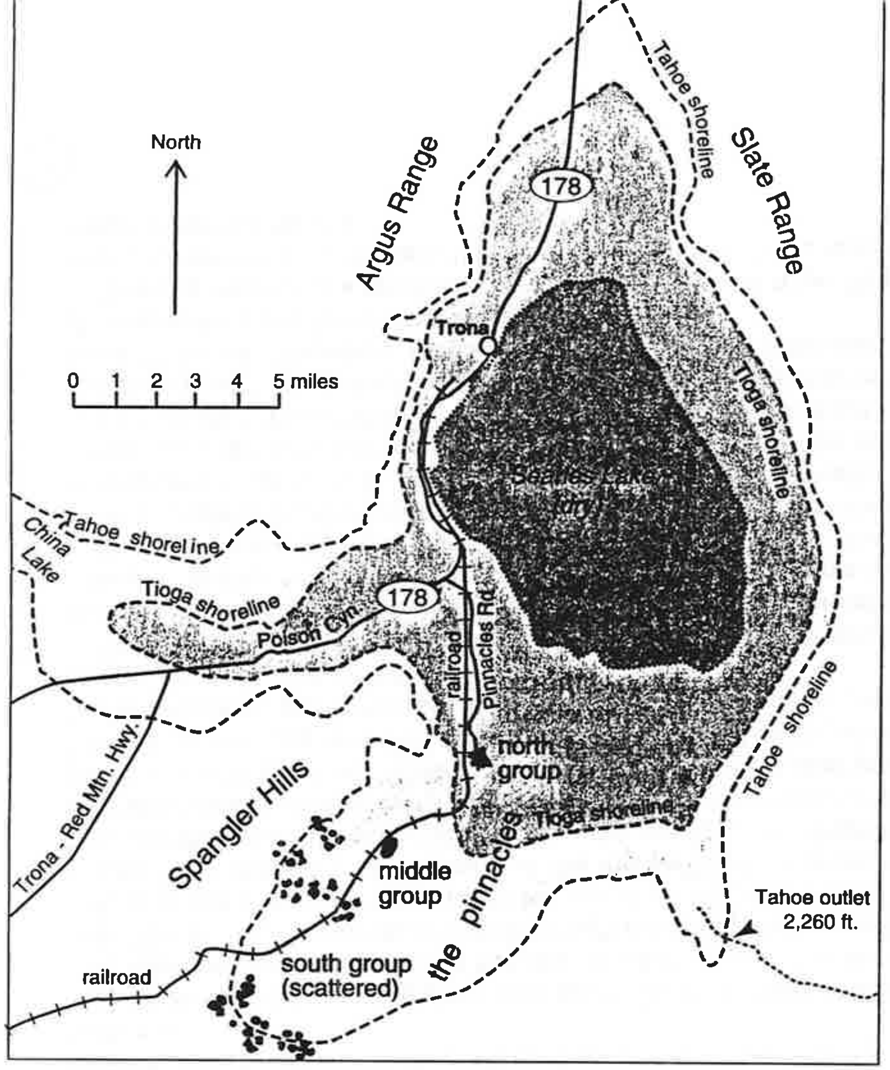
— THE TRONA PINNACLES OF SEARLES LAKE —

One of the routes to and from central Death Valley at Stovepipe Wells is by way of California 178 through Indian Wells, Searles, and Panamint Valleys. Many Death Valley visitors, especially those homeward bound on California 178, have caught distant views of the towering tufa pinnacles at the southwestern edge of the Searles Lake basin. Improvement of a secondary dirt road in the 1990s has made these startling features more accessible. The 7-mile detour is well worthwhile, for this is one of the most unusual landscapes in all North America. It resembles a cartoonist's rendition of the moon's surface before the Apollo astronauts showed otherwise. The pinnacles have appeared in movies, TV shorts and shows, and as background in printed advertisements.

Searles is the third lake, after Owens and China, in a string of five major water bodies nourished mostly by glacial Owens River (vignettes 5 and 18), which formerly carried water from the melting ice and snov



The five pluvial lakes fed by Glacial Owens River from the Sierra Nevada to Death Valley.



Searles Lake, its Pleistocene shorelines, and the pinnacle groups.

GETTING THERE: California 178 traverses the western shore of Searles Lake in the last part of a 25-mile drive from Ridgecrest to Trona. The dirt road to the pinnacles turns southeast off California 178 about 16.6 miles west of the center of Ridgecrest, where California 178 turns east, and 7.7 miles south of Trona High School. Signs mark the junction, and initially the road is wide and well graded. Follow it and the signs about 7 miles southeast and south to the visible pinnacles. After wet weather, soft, slippery lake clays can make the road beyond the railroad crossing impassable. The road, a little bumpy approaching the pinnacles, is easily traversed by slowly driven passenger cars as far as the northern group of pinnacles.

The flat-topped, linear gravel ridge you cross upon entering the pinnacle area is a wave-built beach ridge. Do not turn east on the wheel-track road along its crest but continue over and down onto the flat the pinnacles occupy. There you have a choice of many wheel-track roads that network the area. The Federal Bureau of Land Management monitors the pinnacles as an Area of Critical Environmental Concern. Abuse of the pinnacles in any form and collection of tufa samples are prohibited.

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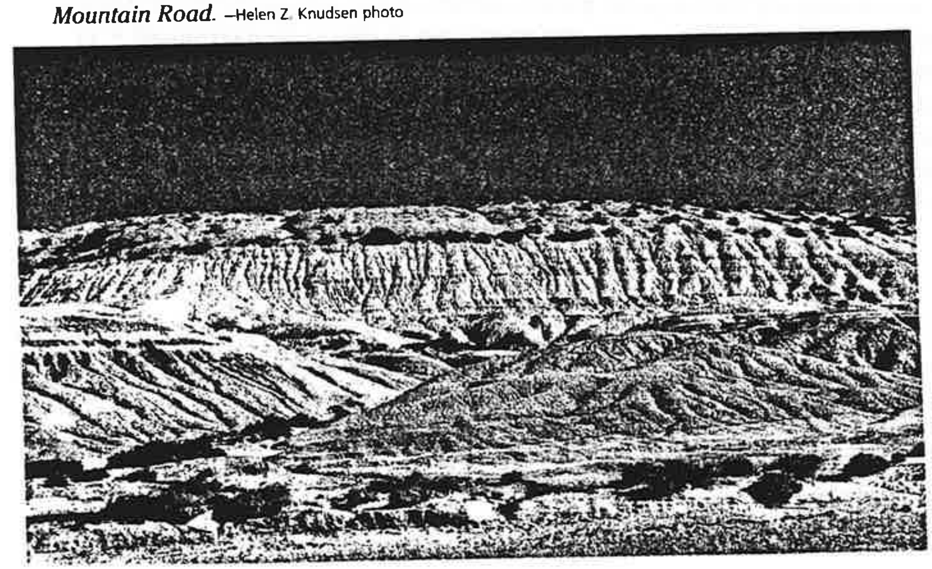
of the Sierra Nevada to Death Valley. Searles was not the largest, areawise, except when it joined with China Lake at the time of its highest water level. Nor, at something over 800 feet, was it the deepest; that distinction goes to 1,000-foot-deep Lake Panamint, the next downstream. The final lake in the string, Lake Manly in Death Valley, was the longest, 100 miles (vignette 5). At its maximum, Searles overflowed its southern margin into Leach trough, along which water flowed east to the southern end of Panamint Valley and then north into Lake Panamint. Shoreline features such as wave-cut cliffs and benches, gravel bars, beaches, tufas, and lakebed deposits identify two principal phases in the late history of Searles Lake. These have been correlated with the last two major glaciations in the Sierra Nevada, the Tioga (about 20,000 years ago) and Tahoe (about 75,000 years ago).

Radiocarbon (^{14}C) ages of lakebed sedimentary layers exposed by erosion around the margins of Searles Lake, especially in lower Poison Canyon along California 178, and as recorded in cores from countless drill holes on the lake flats, confirm this correlation. They also suggest a brief lake phase between Tioga and Tahoe times, possibly correlative with a debatable short-lived glaciation in the Sierra Nevada. The shoreline features of this brief phase could have been wiped out by the more robust and deeper lake of Tioga times.

The five large, interconnected pluvial lakes functioned as a succession of huge decanting vessels. Much mud settled out in the first two, and the water became progressively enriched in soluble salt compounds down the chain, as evaporation from preceding basins concentrated the

Dissected lakebed deposits of the combined Searles and China Lakes, exposed in the headwaters of Poison Canyon just north of California 178 east of Red Mountain Road. –Helen Z. Knudsen photo



minerals. Searles Lake lay in just the right place to accumulate an unusually rich and varied deposit of such salts, including substances of commercial use. Searles hosts a large operation reclaiming and processing various valuable chemical compounds, some containing the element boron that came via the Owens River from springs in Long Valley, well north of Bishop. Boron, a rare element in the earth's crust, is usually associated with volcanic activity, such as that in the Long Valley caldera during the last 760,000 years. Ownership of the resources and facilities at Searles Lake, mainly in Trona, has passed through several hands and in the mid-1990s rests with the North American Chemical Company. The company obtains salts mostly by drilling shallow wells and pumping brines to the chemical plants rather than by scraping the lake floor. Cores from a huge number of drill holes provide valuable subsurface geologic information, as well as organic substances that can be dated by radiocarbon. Whenever the lake dried up, evaporation concentrated a layer of salt on the lake bottom. When incoming water poured into the basin and made a new lake, it laid down a layer of mud. The resulting deposits are interbedded layers of salt and mud.

Approximately 500 pinnacles are clustered into three groups, north, middle, and south. North and south each have about 200, whereas the middle has only 100, including, however, the tallest pinnacle of all at 140 feet. Only the northern group is comfortably accessible by passenger cars. A road continues to the middle and southern groups, but to traverse it, you had best drive a high-clearance four-wheel-drive vehicle.

The pinnacles formed in relatively shallow water along the west shore of a large bay on the southwest side of Searles Lake. They are crowded close to the east and southeast flanks of the Spangler Hills, a mass of hard, jointed granitic rock that extends under the dry lake and is covered by a relatively thin mantle of lake-bottom deposits. A succession

A cluster of tufa knobs and pinnacles within the north group. —Helen Z. Knudsen photo

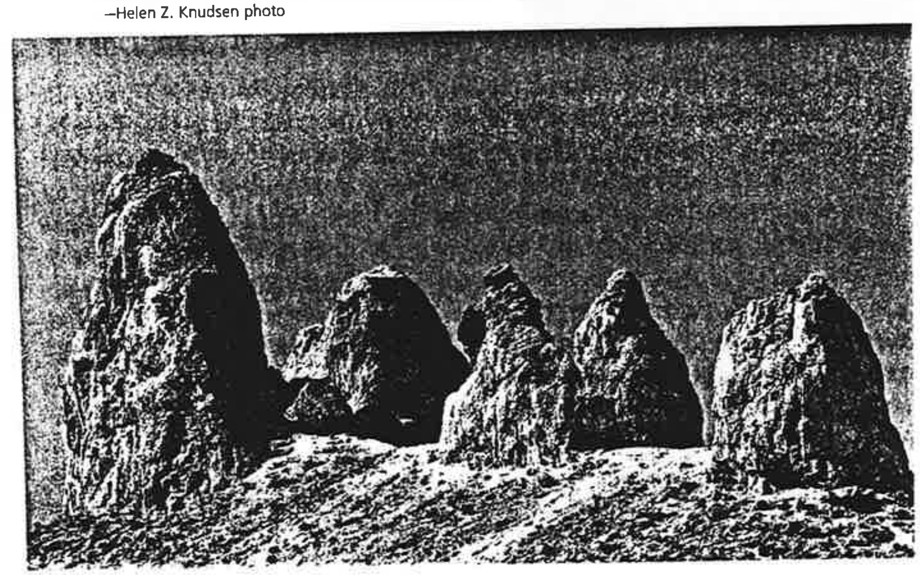
of strandlines on the east face of the Spangler Hills, at elevations well above the tops of north-group pinnacles, show that the pinnacles must have been deeply submerged by rising lake levels, if they existed when the strandlines were made. You can best see these former shorelines from the north pinnacles area when a low western sun backlights the strandlines and creates shadows on the wave-cut cliffs.

Pinnacles are made of a form of calcium carbonate called tufa and take on a variety of forms. Tufa towers are circular in cross section, reasonably symmetrical, and mostly 30 to 40 feet high, with some that are more than three times as tall. Tubby structures, 20 to 30 feet high with an elongated elliptical shape, are called tombstones. Small, dumpy tufa cones, mostly 10 to 15 feet high, and long linear tufa ridges with serrated crests are common. North group's most prominent tufa ridge is 500 feet long, highly serrate, and includes a 120-foot-tall pinnacle. Most pinnacles rise from a broad, gently sloping base of fallen tufa blocks mixed with beds of lake sediment containing lenses and layers of tufa.

The pinnacles did not all form at the same time. Tioga and Tahoe glacial waters submerged the middle and north groups, but pinnacles of the south group stand at higher elevations and could have formed only in a pre-Tahoe-age lake. Much deteriorated by weathering and erosion, they are clearly older. The cores of pinnacles in the middle and north groups, as exposed by erosional stripping of a mantle of cavernous tufa, are composed of hard, solid stony tufa. These core-mantle relationships may indicate two stages of formation.

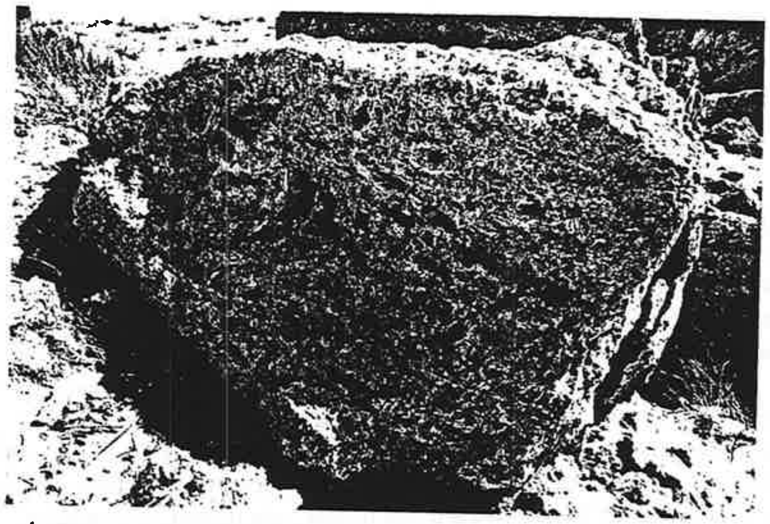
Searles Lake tufas are not restricted to pinnacles. Small deposits form encrustations on rocks, knobby mantles on the ground, low stubby

A small cluster of 20- to 30-foot-tall towers and cones in the north group.
—Helen Z. Knudsen photo



cones, and irregular mounds and ridges. Look for examples 1 to 2 miles south of California 178 along Red Mountain Highway where it crosses several old strandlines.

How did these incredibly large tufa towers and ridges form? Tufa is a chemical or biochemical deposit of calcium carbonate ($CaCO_3$) precipitated from carbonate-rich waters of various origins, commonly lakes and springs. It forms solid stony masses and porous, cavernous, friable jumbles of intertwined calcium carbonate filaments resembling something run through a spaghetti machine. The most thorough study yet made of Trona Pinnacles, by David Scholl, identifies seven varieties of tufa distinguished largely by structural and textural characteristics.



A separated 30-inch fragment of cavernous tufa that mantles most pinnacles in the north group. –Helen Z. Knudsen photo

The nature of tufas varies considerably on and within individual pinnacles and between the three groups. Many variable factors influence tufa formation, including microclimate, changes in water levels, water temperature (as well as its chemistry and depth), wave action, spray, organisms, and even the underlying substrate. Tufa can precipitate chemically from carbonate-bearing water by temperature changes, evaporation, and loss of gas, principally carbon dioxide (CO_2) from solution in the water. Lowering the carbon dioxide concentration in water decreases the solubility of calcium carbonate. Algae living in the water cause calcium carbonate to precipitate by removing carbon dioxide from the water as they photosynthesize. Most tufas contain considerable organic matter, much of it probably of algal origin, and they also contain casts that replicate individual algal forms and structures that suggest

A good example of a tower pinnacle about 50 feet high with a mantle of cavernous tufa (north group). –Helen Z. Knudsen photo

algal colonies. Scholl states firmly that Trona Pinnacles consist princ pally of algal tufa. Algae require sunlight to function and go out of bus ness in water much deeper than 100 feet, less if the water is turbic Since many Trona pinnacles have at times been deeply submerged, the must have experienced intervals of little or no growth.

Tufa towers in the north group have a core of solid, rocklike tufa e cased within a 10- to 20-foot-thick surficial mantle of cavernous tufa wit a variety of open textures. On some towers, part of the cavernous mant has peeled away, exposing the core. The mantle was almost certain deposited from waters of Tioga-age Searles Lake. The inner solid core may have been deposited during the rising phase of the Tioga-age lak or possibly during the preceding Tahoe stage, and were exposed durin the Tioga-Tahoe interval when the lake evaporated to dryness.

Viewed from the air, the pinnacles clearly align in directions beari N65°W, N50°W, N30°E, N55°E, and N65°E. In the north group, N30°E a N55°E are the dominant alignments for both pinnacles and tufa ridge Alignment and localization of pinnacles exist because algae prefer fre spring water to brackish lake water, so they cluster in colonies arou spring outlets. You can see this today at growing Mono Lake pinnacl and elsewhere. At Mono Lake, springs continue to flow from the tops some towers, making them ever higher. The floor of Searles Lake alo the flank of Spangler Hills could have had many underwater springs f by abundant water from as far away as the Sierra Nevada. Many of the

Record of the Rocks

Evolution of the Death Valley area through time

wo thousand years ago Death Valley looked ıuch as it does today. There were a few differnces: a cooler and more moist climate, more egetation, a small lake covering part of the val-'y floor, and of course no roads or buildings.

The lake was brackish, twenty feet deep, nd about two hundred square miles in area. he glistening white salts and jagged salt pinacles of the Devils Golf Course had their rigins in this lake.

Death Valley lies in earthquake country. wo thousand years ago a massive earthquake, 'erhaps more powerful than the San Francisco uake of this century, struck Death Valley. hock waves rolled through the solid rock in the shallow lake, generating waves which sped across the water and hurled themselves against the foot of the Black Mountains.

When rock and water had quieted, stillness again prevailed. But the valley had changed dramatically, within minutes perhaps. Now the waters lapped closer to the mountains, and a ten-foot cliff appeared where no cliff had been before. Today the road from Furnace Creek to Badwater lies at the foot of this cliff, this fault scarp which traces back to one specific event in the creation of Death Valley.

Earthquakes continue to this day in Death Valley; about one every two years occurs somewhere in the three thousand square miles of the monument. Most pass unnoticed, even by the residents. But sensitive seismographs in nearby locations feel and record these quakes; they tell us that the building of Death Valley is not yet finished.

The geologic story here stretches back over the last four billion years, the first two billion of which are lost in time. Fortunately, we have been able to piece together the last two billion—but only because the rocks of this period are exposed and thus available for geologists to read and interpret.

Chapter I—Early Precambrian

The Early Precambrian is by far the longest chapter in the history of the earth and it is the one about which we know the least. It encompasses all of the time from the birth of this planet to perhaps more than a billion years ago.

Death Valley has gneisses and schists, marbles and metadiorites—all metamorphic rocks from this ancient time. These rocks were once sedimentary rocks—sandstones, shales, conglomerates, and limestones—and also igneous diorite. They were altered to their present metamorphic forms miles beneath the surface, under great pressures and temperatures over long periods of time.

The oldest rocks form the so-called "turtlebacks," northwest-southeast trending ridges lying at an angle to the north-south orientation of the Black Mountains. The turtlebacks (the form of these older metamorphic rocks has been compared to that of a turtle's carapace) form the most prominent ridges along the Black Mountain front. The younger layered rocks lie at steep angles against the sides of the turtlebacks.

These ancient rocks were the building blocks of Death Valley's "first" mountains. As with all mountains—past, present, and future—they were attacked and eventually eroded away. That the ancient rocks are still in existence is evidence that they were buried deep in the roots of those first mountains. Erosion reduced the topography to rolling hills, or plains, and the stage was set for Chapter II.

Chapter II—Late Precambrian

The Late Precambrian lasted from about one billion to 600 million years ago, and the record is slim. Rock types are varied, and they include dolomites, rocks that had their origins as lime precipitated from the ocean water that now covered the Chapter I rocks.

In the Ibex Hills and the southern Panamint Range, white mine-tailings are easily seen against the darker mountain slopes. Visible for great distances, they indicate the location of talc mines. The talc is found in the *Pahrump* series, as the rocks of Chapter II are called in Death Valley.

Late in Chapter II, molten rock welled up from deep within the crust. Some of this magma spread along the bedding planes of the sediments. Heat and superheated gases permeated the nearby rock. Where this host rock was dolomite, it was metamorphosed into the commercial product *talc*, from which talcum powder is derived.

Other rocks of the Pahrump series are found in the Panamints: at Wildrose Canyon, Telescope Peak, Harrisburg Flat, and Skidoo. These layers were intruded by gold-bearing granitic rocks in a later chapter.

Another interval of uplift and erosion locally removed much of the record.

Chapter III—The Paleozoic Era

During the Paleozoic era, which began about 600 million years ago, a shallow sea covered much of the American Southwest, and vast thicknesses of layered sedimentary rocks accumulated. Here, and elsewhere, these multicolored layers record the evolving life of the oceans: trilobites, brachiopods, gastropods, corals, and crinoids.

As the layers were being deposited, the sea floor steadily sank, continuously making room for new layers at the top. The great *Cordilleran geosyncline* was being formed, a trough that extended from Canada to Mexico.

Had the Death Valley layers been left undisturbed through time, they would now form a sequence over five miles thick from base to top. But Death Valley has been the scene of several later episodes of mountain-building which have bent, broken, and sheared these layers into a crazy-quilt pattern. The building of mountains also provides renewed energy for erosion, and the topmost portion of the record is continuously being removed.

Paleozoic layers, the cumulative record of over 400 million years, form the walls of Death Valley. The individual bands tell of a conflict as old as the planet: the endless struggle between the forces of uplift and the forces of erosion.

They tell of the edge of the sea encroaching on the continent, of the leveling agents of erosion inevitably defeating the mountains along the retreating shoreline. They tell of mountains that were reduced to pebbles, sand, and silt.

Clastic rocks, such as conglomerate, quartzite, and shale, record the advance of the sea over the land. These rocks, fragments eroded off the land, predominate in the older part of the Paleozoic sequence. As the sea advanced to the east, and debris from the land no longer washed into it, carbonate rocks such as limestone and dolomite formed younger layers.

The ancient rocks of Aguereberry Point, which at 6,443 feet overlooks Death Valley from the west, are bathed in the golden glow of a desert sunrise.

The Paleozoic rocks, and the older basement rocks, were the building blocks for Death Valley as it presently is. Those rocks rest against the turtlebacks of the Black Mountains and the Chapter II rocks on Tucki Mountain. The Grapevine, Funeral, southern Panamint, and Cottonwood mountains consist almost solely of Paleozoic rocks.

At the close of Chapter III, some 180 million years ago, great Paleozoic seas were beginning to retreat from the continent.

Chapter IV—The Mesozoic Era

The Mesozoic era is known world-wide as the "Age of Dinosaurs," but no fossils of these immense reptiles have been found in Death Valley. Marine sediments found in Butte Valley tell us that the Paleozoic seas persisted into Mesozoic time. Interbedded volcanic rocks tell of the lavas that flowed out over the sea floor only to be covered by other sedimentary layers. Still other volcanic rocks were extruded onto dry land.

The presence of these volcanic rocks heralds the turning point in the geologic history of Death Valley and the Southwest. (Up to this point the dominant theme had been the deposition of flat-lying marine sediments on the subsiding Precambrian basement complex.) This volcanism provides the first hint of the crustal upheavals that followed, lifting the land above the sea and ultimately building the landscape of today.

Igneous rocks—magmas that cooled and hardened at depth—form the bulk of the Chapter IV rocks. The injection of these molten rocks into the overlying sedimentary rocks was largely responsible for the minor concentrations of gold, silver, and other metals which have been found in this part of the Great Basin.

Outcrops of granitic rocks are found in the Panamint Range in upper Hanaupah Canyon, Wildrose Canyon, and the Harrisburg-Skidoo area; they are also found in the Hunter Mountain area of the Cottonwood Mountains. Granitic rock forms the "Grandstand," that unlikely appearing mass of shadowy rock that juts up from the yellow flat of the Racetrack playa between the Cottonwoods and the Last Chance Range.

In Mesozoic time the dominant geologic theme throughout most of the West was uplift, in four separate episodes. The major mountain-building episode has been named the *Nevadan orogeny*. Its major axis extended from southern California through Nevada north into Canada. The folding and faulting that accompanied this orogeny have left a complex and confusing story, not only in Death Valley but over the entire Great Basin.

A feature recognized in several parts of the Great Basin is the *thrust fault*. These are low-angle faults along which older rock layers have been thrust over younger layers, in a position more or less parallel to the earth's surface.

In Death Valley, evidence of west-to-east thrust faulting has been found in the Last Chance Range, the Cottonwood Mountains, and the Grapevine Mountains. The origin of the thrusting is uncertain, but it seems likely that it is connected with *plate tectonics*—the rafting of huge blocks of the earth across the face of the globe by subsurface convection currents.

The crustal unrest in the Mesozoic continued into the next chapter, or even to the present. Much of the record has been stripped away by massive erosion and covered by debris from the erosion, or further confused by the superposition of younger structures on older ones.

Chapter V—The Cenozoic Era

At the beginning of the Cenozoic era sixty-five million years ago, the building blocks for Death Valley had been assembled—Precambrian crystalline rocks buried beneath thousands of feet of layered Paleozoic sediments. Much of the record of the Mesozoic era had already been eroded away, and more would be lost in the colossal earth movements of the Cenozoic.

There is no record of the first third of the Cenozoic era in Death Valley; rocks of the Paleocene and Eocene epochs are not found here.

Forty million years ago this part of California and Nevada may have been a gently rolling plain, a grassland with streams and wooded areas. Grazing mammals cropped this grass. Present were the *eohippus* (a three-toed ancestor of the modern horse), rodents and birds much like those of today, and *titanothere* (an immense distant relative of today's rhinoceros).

Earthquakes and volcanism, widely dispersed in time and space, gave warning of approaching change. There were cracks in the apparently solid crust, and the rocks were beginning to move along them.

Oligocene—Miocene—Pliocene

During the time interval from forty to three million years ago, the Death Valley of today was formed. Rubble eroded from the rising mountains and accumulated in the basins. Lavas, tuffs, and ash were extruded from volcanoes, some located along cracks bounding the shifting mounting blocks. The brightly colored Cenozoic rocks provide some of the valley's most colorful, most spectacular, and most accessible scenes.

Titus Canyon formation. In upper Titus and Titanothere canyons, and on the crest of the Grapevine Mountains lie multicolored stream and lake deposits interbedded with explosive volcanic fragments. These rocks form the red, maroon, buff, and gray Titus Canyon formation.

The fossils of the grazing mammals who lived here forty to thirty-five million years ago were found near the present Titus Canyon Road, in Upper Titanothere Canyon. Their presence tells of a climate colder and wetter than the present one. The sands and muds tell of erosion of an uplifted area higher than the area of deposition. The volcanic rocks—rhyolitic ash and tuff, welded tuff—tell of the unrest in the Oligocene epoch that was beginning to build today's landscape.

Artists Drive formation. The color-splashed Artists Drive lies at the foot of the Black Mountains between Furnace Creek and Badwater. Muds and gravels which washed off the emerging Black Mountains lie in layers separated by more volcanic materials. Some of the volcanic rocks are ancient lava flows; others were blown into the air by violent eruptions; some forced their way between the sedimentary layers and cooled underground, to be exposed by erosion millions of years later.

The layers are about four thousand feet thick, and they provide a record of the time from thirty to ten million years ago. The bottom layers are probably part of the Oligocene epoch. Most of the Artists Drive formation is younger, Miocene in age.

The faulting which accompanied the uplift of the Black Mountains fragmented the layers, and today Artists Drive presents a kaleidoscope of broken layers and brilliant colors extending from the valley floor to the crest of the northern Black Mountains. The harsh, dark front of the mountains extending to the south presents a striking contrast.

Furnace Creek formation. In Death Valley the tracks of mastodons and camels—animals long extinct in North America—are preserved in hardened muds, lake sediments. The tracks and lake sediments together indicate a land of grass and fresh-water lakes.

In Furnace Creek Wash, between the Funeral and the Black mountains, the Furnace Creek formation provides the yellow, orange, gray, and mustard badlands that are one of Death Valley's most popular attractions. Zabriskie Point and Golden Canyon provide vantage points overlooking these badlands and within them. The playa sediments have been faulted, tilted, and cut into uncounted canyons by running water. The Furnace Creek formation is Pliocene in age.

The Pliocene Death Valley was already basin-and-range country. The lakes provided the final resting place for the muds washed out of the mountains, and alluvial fans extended up the slopes around the intermittent lakes.

A single basin extended from north of Stovepipe Wells south to Furnace Creek and beyond, almost to the present location of Dantes View. The lake beds record a portion of the Black Mountain uplift. The oldest layers, muds interbedded with volcanic rocks, are found highest on the northeast slopes of the Black Mountains. Progressively younger beds lap against the mountain slope like shingles. As the mountains were raised higher, the south shore of the lake retreated northward.

Borax has been found in the oldest beds of the Furnace Creek formation. The borate minerals are water soluble. As the lakes dried up, borates were precipitated out, often forming lenses large enough for commercial mining. Furnace Creek, Ryan, Twenty-Mule-Team Canyon, Monte Blanco and Bennetts Well are all names linked with borax, the "white gold of the desert."

Funeral formation. The building of Death Valley continued. The mountains reached higher and their slopes steepened. Water flowing off the slopes rushed faster, carrying more of the abrasive rock fragments needed to cut away mountainsides. The debris moved steadily downhill, to accumulate as huge alluvial fans surrounding each uplifted mountain block.

Ground water carried cementing materials in solution and bound the fragments of ancient rock layers into a new rock, a *fanglomerate*, composed of materials deposited in alluvial fans and then cemented.

The Funeral fanglomerate was, in turn, uplifted, tilted, folded, faulted, and eroded to form the steep walls, chimneys, and canyons of Furnace Creek Wash and Emigrant Canyon. Similar in origin and composition, the *Nova* formation was elevated with the Panamint Range, to form the surface of Harrisburg Flat at the head of Emigrant Canyon.

The layers of mud, sand, gravel, and boulders form part of the Pliocene and Pleistocene record. The oldest layers of the Funeral formation are perhaps nine million years old; the youngest were laid down less than a million years ago.

Today the young rocks of the Furnace Creek and Funeral formations have themselves been broken into blocks, raised or lowered, and tilted. The *Texas Spring* syncline between the Funeral and Black Mountains formed when Pliocene and Pleistocene rocks were bent into a *U*-shaped fold between the rising mountain blocks.

Greenwater volcanics. The Black Mountains are a block wedged between the Furnace Creek and Death Valley fault zones. As this block was being raised and rotated to its present position, basaltic lava—its volume measured in cubic miles—poured out from volcanoes located along fault zones. The black mass of basalt is evident in the black cap it has provided for the Greenwater Mountains, just east of the Black Mountains. Other lavas surround portions of the road to Badwater and appear as disconnected black mounds surrounded by the lighter-colored rubble of Furnace Creek Wash and the Greenwater Valley.

Pleistocene—Holocene

The Pleistocene epoch lasted from about three million to ten thousand years ago. The last ten thousand years constitute the Holocene epoch, in which we live. The building of Death Valley has continued to, and continues through, the present.

The last major earthquake occurred only about two thousand years ago. More recently, Ubehebe Crater "blew its top." Its age may be reckoned in hundreds rather than in thousands of years; some of its satellite craters are younger still.

The Pleistocene was the time of glacial ice. Four times in the last million years vast ice sheets covered much of North America. Five major advances of the ice have been recorded in the Sierra Nevada. The ice formed and flowed in response to climatic changes. There were ice caps to the north, valley glaciers in the Rocky Mountains and the Sierra Nevada, and lakes across the Great Basin, Mojave Desert, and Death Valley.

Evidence provided by living fish proves that connections linking Owens Valley, Death Valley, the Amargosa and Mojave rivers, and the Colorado River existed during at least one of the Sierra glacial stages. Confirming landform evidence has either eroded away or has not yet been found.

A salt marsh now covers much of the floor of Death Valley below sea level. On the slopes above this salina, the waters of a Pleistocene lake cut shorelines and deposited gravel bars. The highest terrace, on Shoreline Butte, indicates that the lake (Lake Manly) may have been six hundred feet deep. It was about 116 miles long, ten to twelve miles wide, and probably dried up completely about twenty thousand years ago.

There are other marks left by Lake Manly, but their general scarcity indicates that the lake had a very short life—perhaps only a few thousand years—and had many rapid fluctuations of level.

It is possible, even probable, that lakes formed in the basins in and around Death Valley in each of several glacial advances in the Sierra Nevada. The fragile evidence left by the waters of the first stages has, again, either eroded away or has not been found.

Events of the Tahoe glacial stage of about sixty to seventy-five thousand years ago linked three rivers and many lakes together in what is called the Death Valley system, in which Sierra Nevada water reached Death Valley.

The Owens River filled Lake Owens. The waters cut a notch through a lava flow across southern Owens Valley and spilled into Indian Wells Valley (Ridgecrest). This created a shallow lake which soon spilled into Searles Valley (Trona). When Lake Searles reached a depth of over six hundred feet, its waters flowed into Panamint Valley, Death Valley's nearest neighbor to the west. Shorelines on the west slope of the Panamint Mountains indicate the existence of a lake over nine hundred feet deep, deep enough to allow glacial waters from the Sierra Nevada to reach Death Valley over Wingate Pass.

At this same time, waters from the San Bernardino and San Gabriel mountains may have reached Death Valley by way of the Mojave River, connecting another chain of lakes. And the Amargosa River, draining a larger area than either the Owens or the Mojave rivers, contributed its waters to the Death Valley sump.

Today pupfishes are found in the Owens Valley, Death Valley, the Amargosa Valley—and in the Colorado River system! Apparently, during at least one of the glacial stages, waters from the Death Valley system were joined to waters of the Colorado system.

During the Tioga, the most recent important glacial stage, the Searles and Panamint lakes never gathered enough water to enable them to overflow. Thus, any water reaching Death Valley had to come from the Mojave and Amargosa rivers. The Tioga-stage lake in Death Valley may have been a hundred miles long and four hundred feet deep.

Today all that remains of this system of lakes is the flat, barren salinas and playas that mark the lowest elevations in each of the now separate desert basins.

Cenozoic Block Faulting

Death Valley was constructed on a truly massive scale. Telescope Peak stands 11,331 feet above the lowest point on the surface of Death Valley. Below the peak, the bedrock floor of the valley may be 8,000 feet below the present surface, a structural difference in elevation of 19,000 feet. The Death Valley trough is five to twenty miles wide and about 220 miles long. The valley itself, between drainage divides, is about 150 miles long, of which about 120 miles are included within the monument.

Death Valley is a part of the *Basin and Range* province. Basin and range succeed one another in endless rows from California to Utah and from Idaho south to Mexico. Death Valley is a basin, a block of the earth's crust depressed in relation to the blocks around it. The surrounding ranges are elevated blocks.

In 1874 G. K. Gilbert, an eminent geologist of the time, theorized that the basins and ranges had their origins in "normal" faulting, a process in which blocks slide past one another in a predominantly vertical direction, the uplifted blocks forming ranges and the lowered blocks forming basins.

After a century of study and many other varied theories, it appears that Gilbert's simple explanation may be indeed correct. The blocks appear to be bounded by steep normal faults that flatten with depth. The faults are products of extension toward the east and west, a stretching of the earth's crust.

The block mountains of Death Valley are highest on their western sides, having been tilted down to the east. As the great mountain blocks were being elevated, the block that is Death Valley settled deeper into the crust. The Panamint and Black mountains continue to rotate up on their western sides and down to the east, continuously lowering the bedrock floor of the lowest place in the Western Hemisphere. The movement is similar to that of a seesaw. As the Panamints are elevated, the valley floor is lowered.

The existence of pupfish (genus Cyprinodon) in Death Valley is an evolutionary miracle. Survivors of Ice Age lakes, they found refuge where they could and evolved into several species that live in fresh and salt waters, ponds and streams. The Cottonball Marsh pupfish live on the saltpan, a habitat saltier than the sea, which they share with terrace-building algae.

This process has continued intermittently over the last twenty-five million years. The Artists Drive fault scarp, between Furnace Creek Inn and Badwater, traces but a single instant in the continuing creation of Death Valley.

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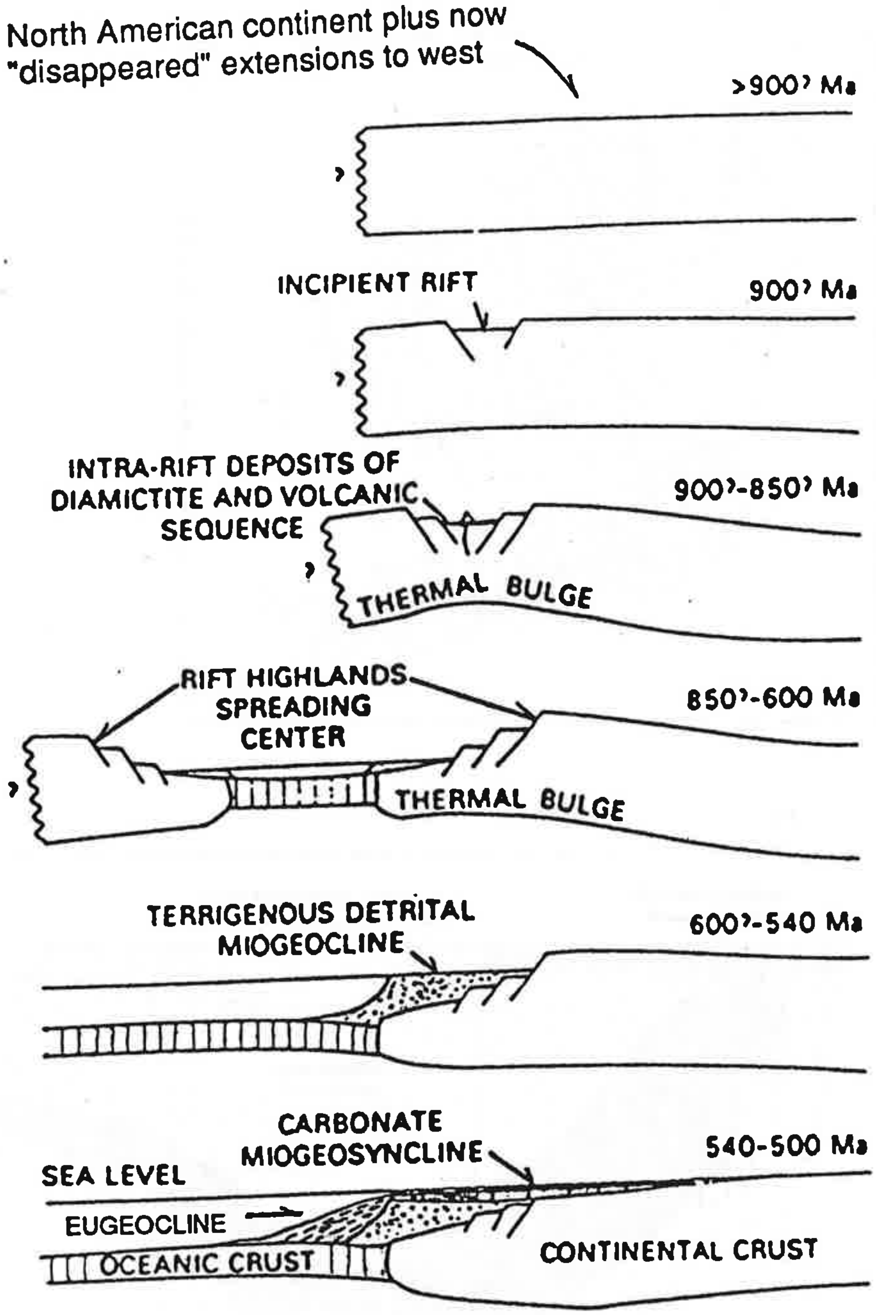
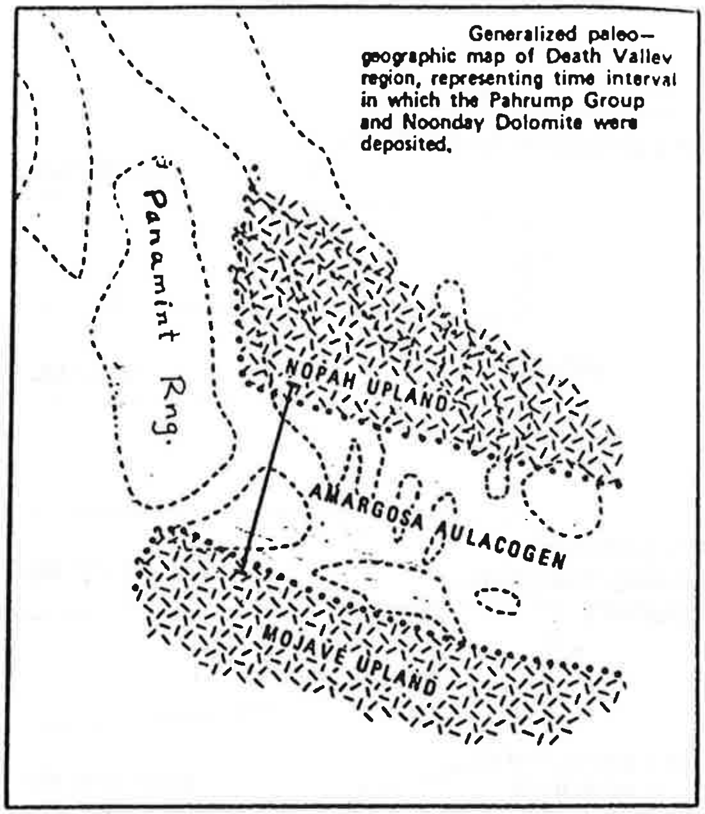
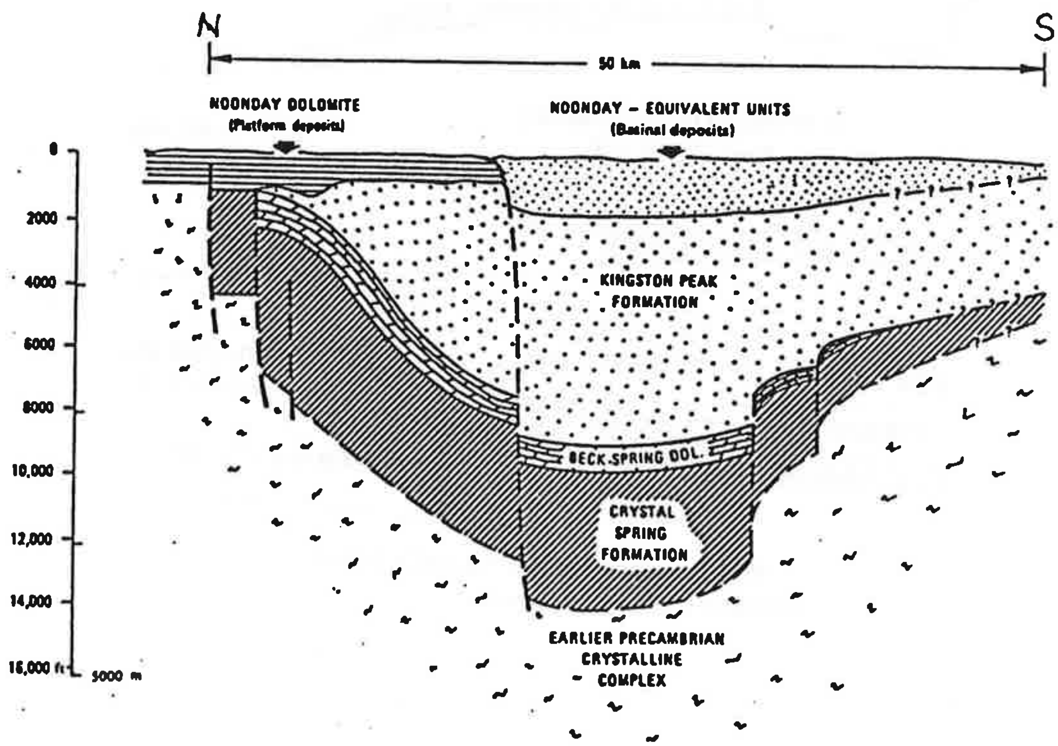


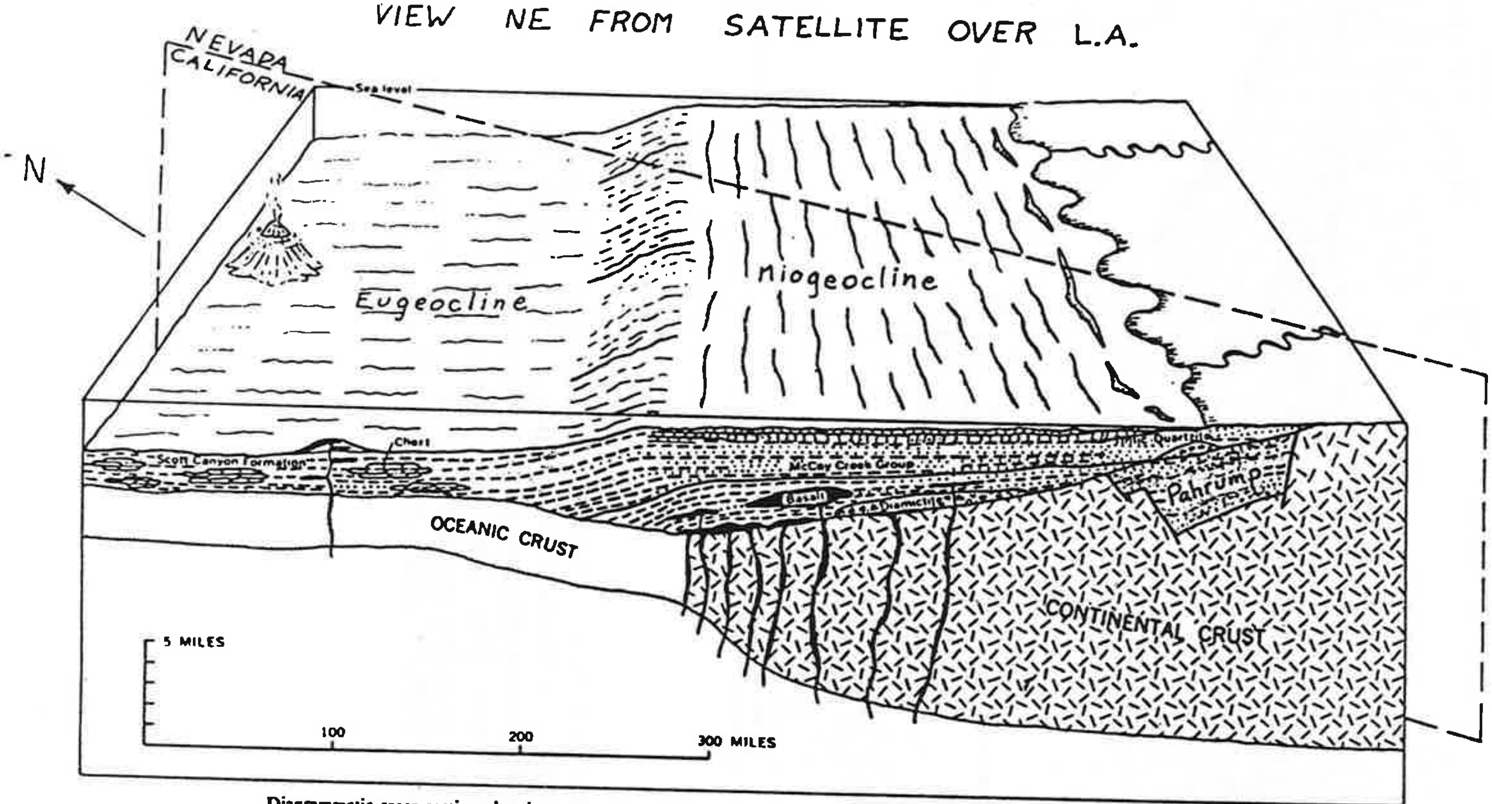
newly created western continental margin for North America



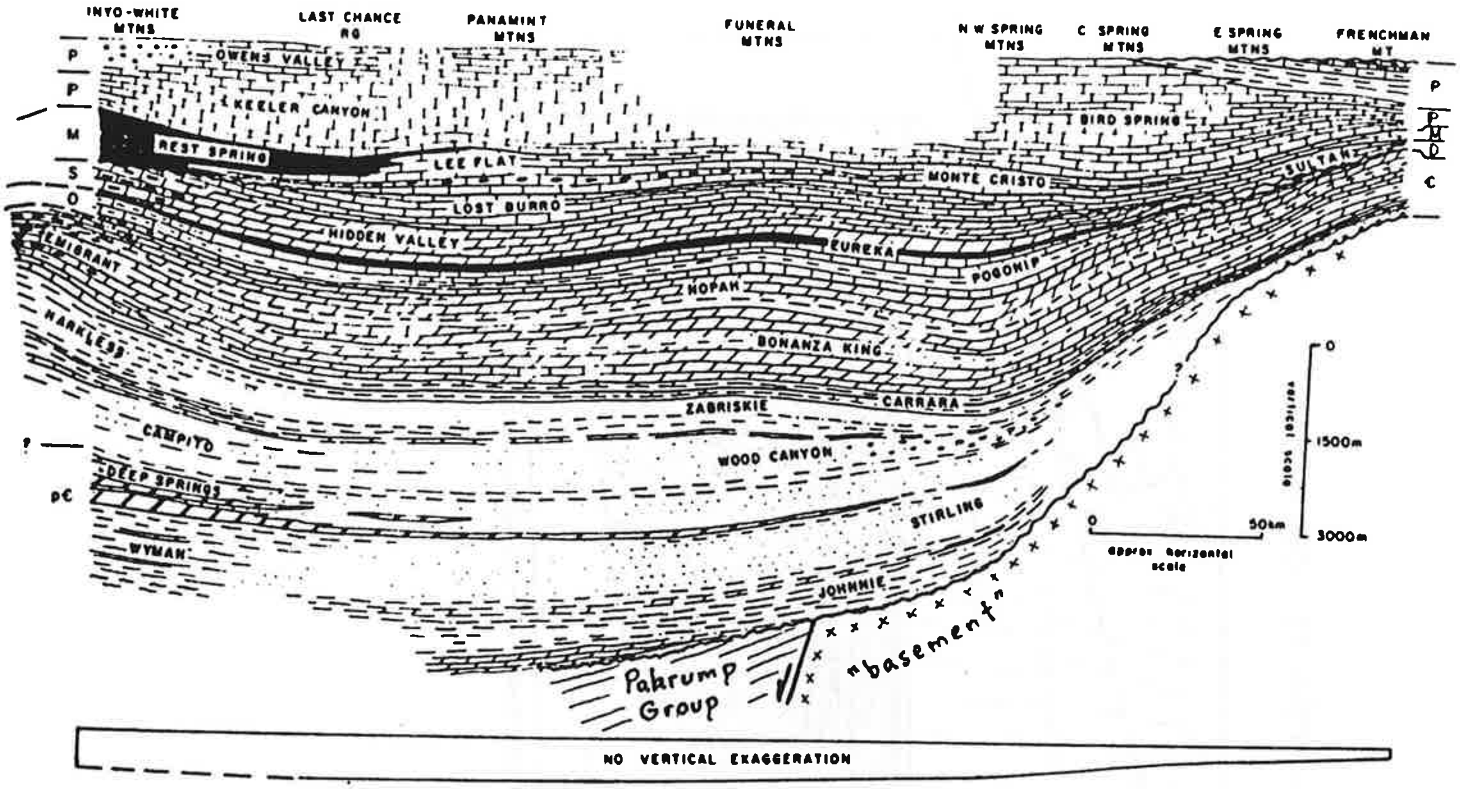
Generalized paleo-geographic map of Death Valley region, representing time interval in which the Pahrump Group and Noonday Dolomite were deposited.



VIEW NE FROM SATELLITE OVER L.A.



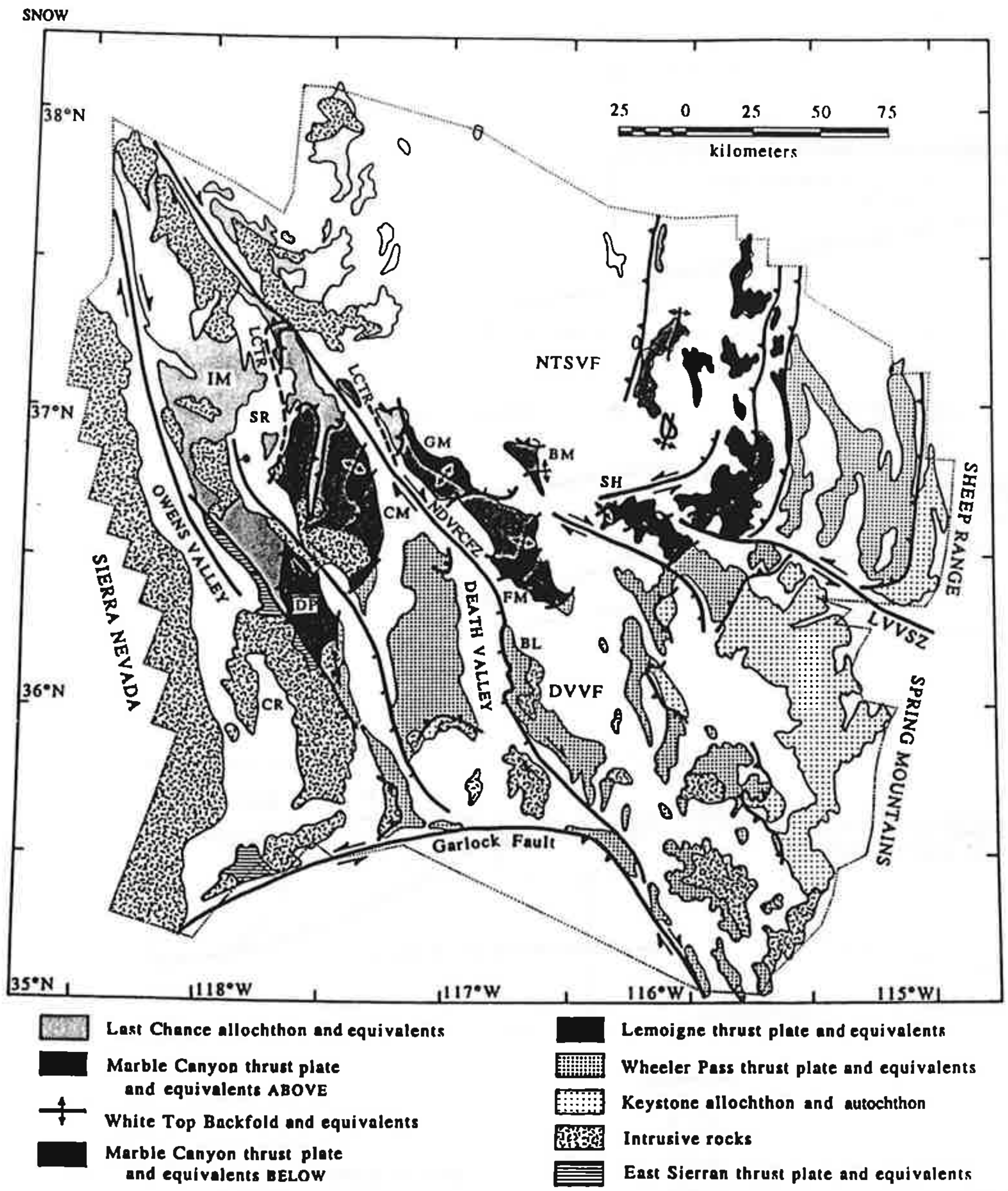
Diagrammatic cross section showing upper Precambrian and Lower Cambrian rocks in the northern Great Basin, Nevada and Utah.

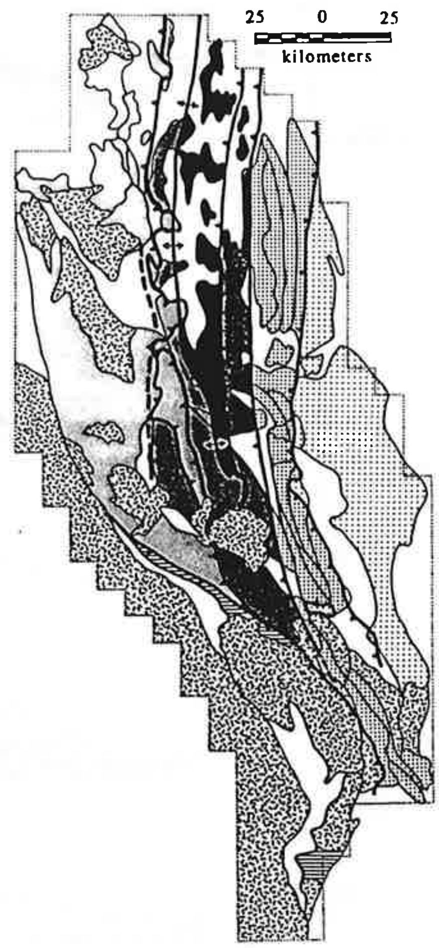


. Schematic stratigraphic cross section from vicinity of Las V Vegas (Frenchman Mtn.) to White-Inyo Ranges. This wedge-like sequence is the Cordilleran miogeocline, bounded to the east by the 'craton' (sections like that at Frenchman Mtn. and the Grand Canyon) and grading westward into the "eugeocline', which represents continental slope and rise strata.

The Death Valley Region Now and Then: A palinspastic reconstruction of what it would have looked like prior to ~250 km of NW-SE Basin & Range extension

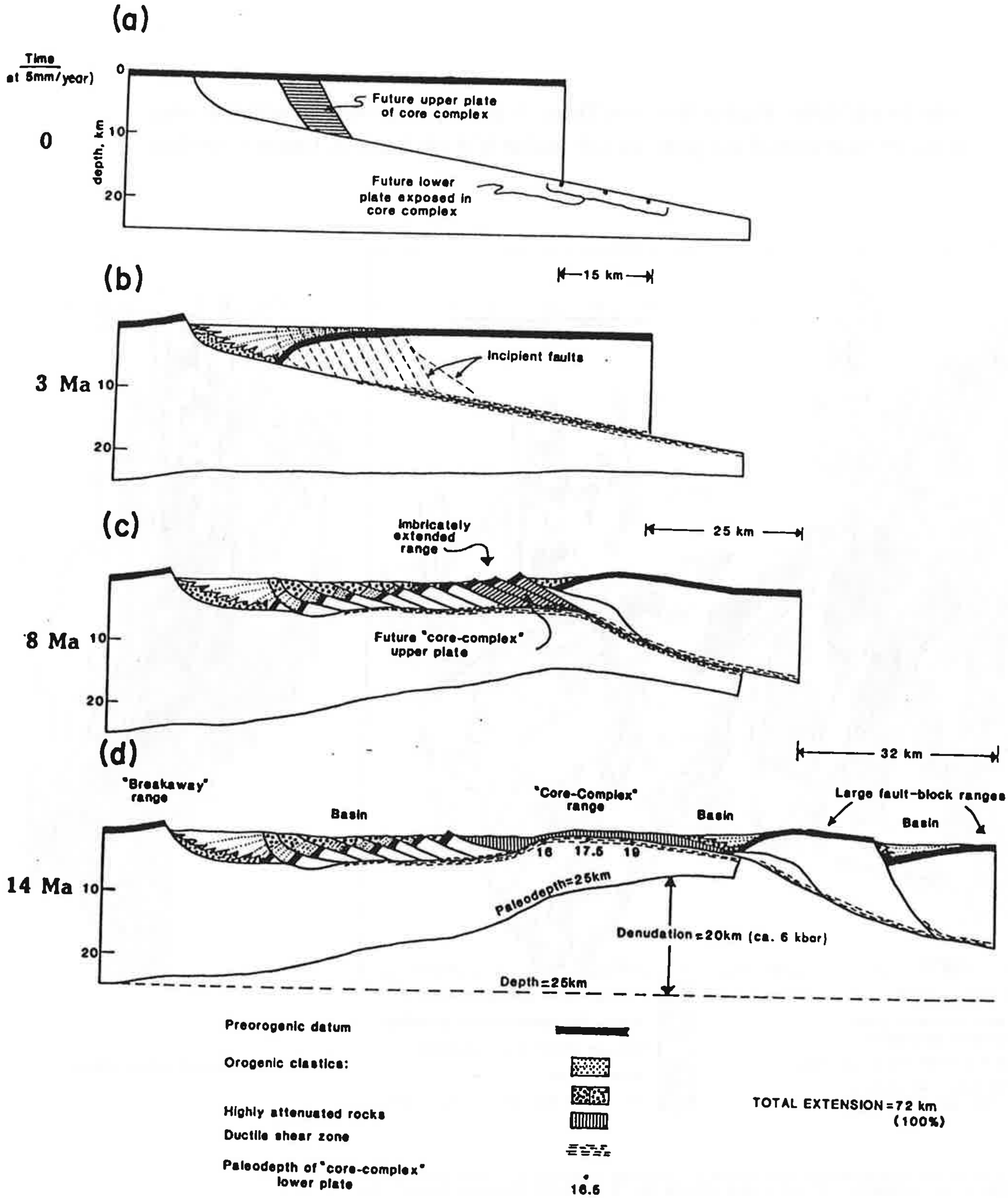
SNOW





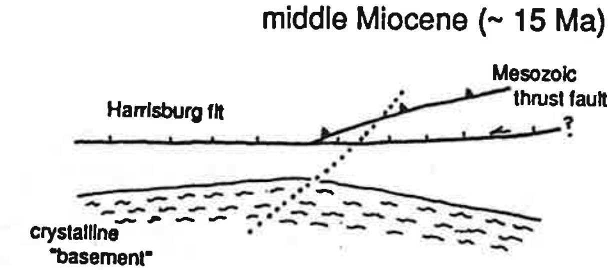
PALINSPASTIC RECONSTRUCTION

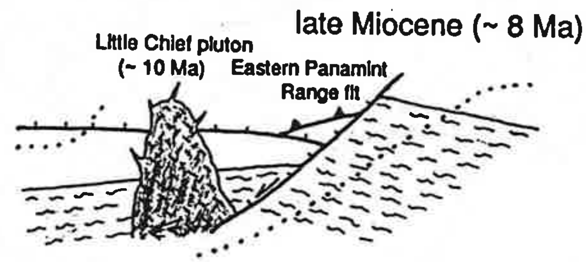
Figure 3. (A, this page) Tectonic map showing correlations and distributions of major pre-tertiary contractile structures used to reconstruct the Death Valley extensional system and selected features discussed in text (after Saow, 1992a; Wernicke et al., 1988). (B, opposite page) Palinspastic reconstruction of Death Valley extended terrane.

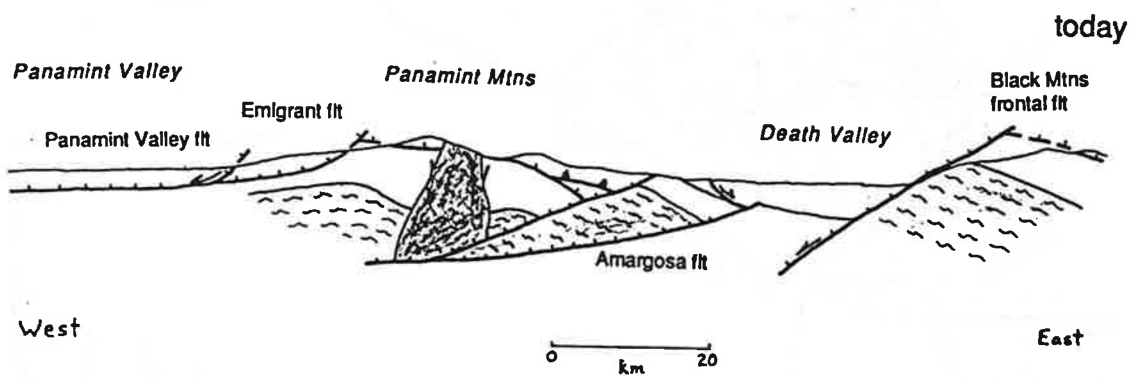


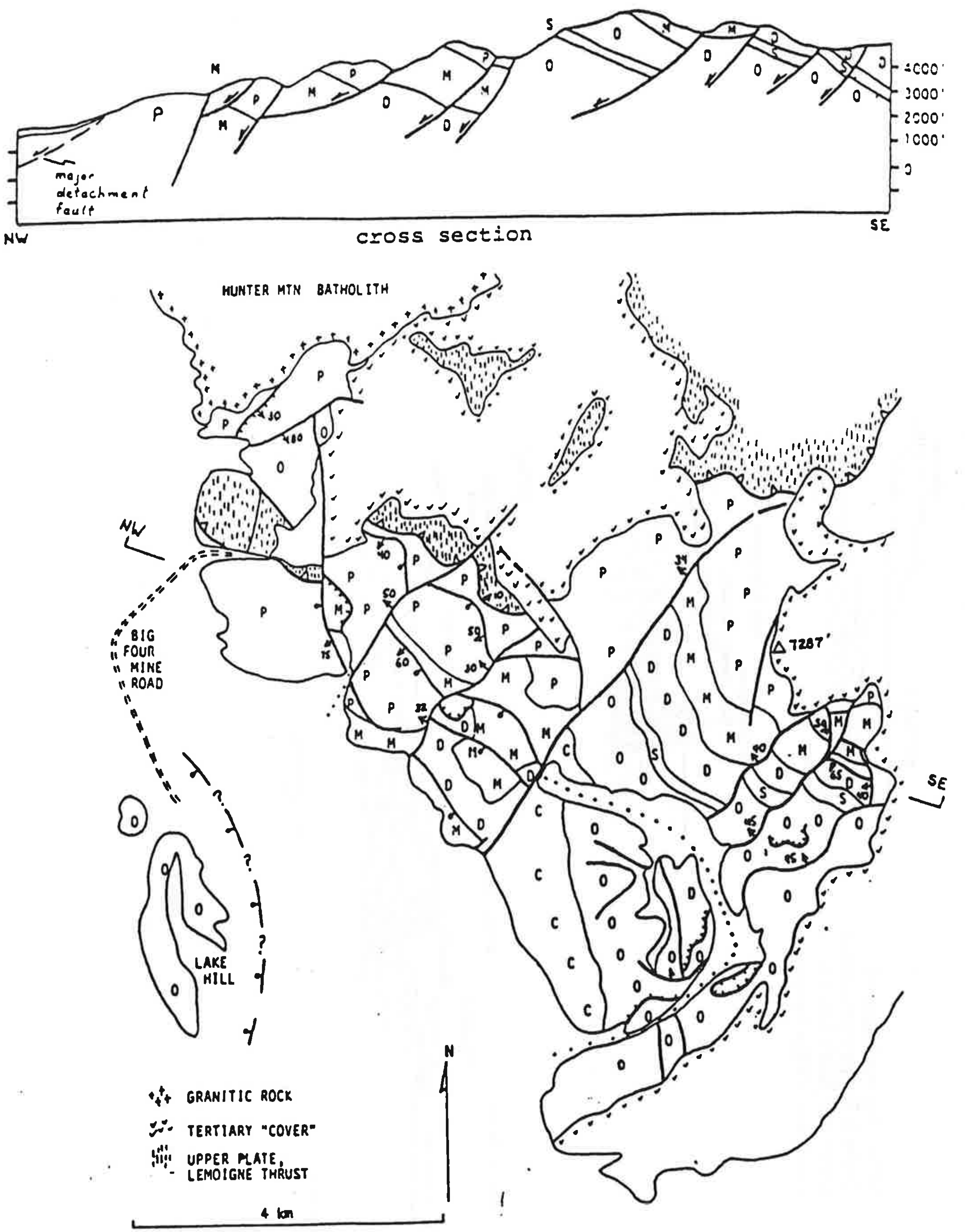
(a)
Time
at 5mm/year)
0
depth, km
10
20
Future upper plate
of core complex
Future lower
plate exposed in
core complex
(b)
15 km
3 Ma
Incipient faults
(c)
Imbricately
extended
range
25 km
8 Ma
Future "core-complex"
upper plate
(d)
32 km
"Breakaway"
range
Basin
"Core-Complex"
range
Basin
Large fault-block ranges
Basin
14 Ma
16
17.5
19
Paleodepth=25km
Denudation=20km (ca. 6 kbar)
Depth=25km
Preorogenic datum
Orogenic clastics:
Highly attenuated rocks
Ductile shear zone
Paleodepth of "core-complex"
lower plate
16.5
TOTAL EXTENSION=72 km
(100%)

Schematic evolution of Basin & Range extensional faults in a Black Mtns to Panamint Valley transect

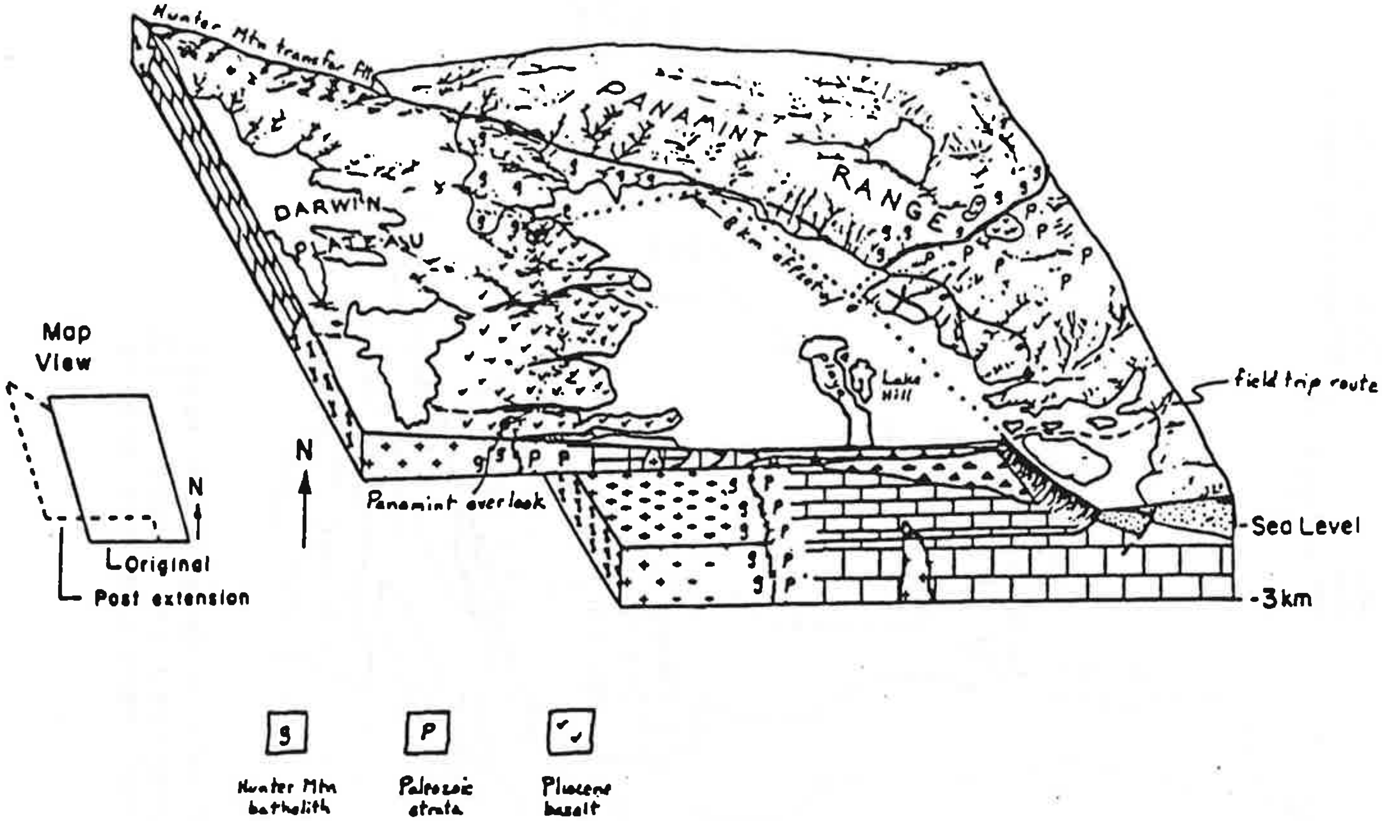


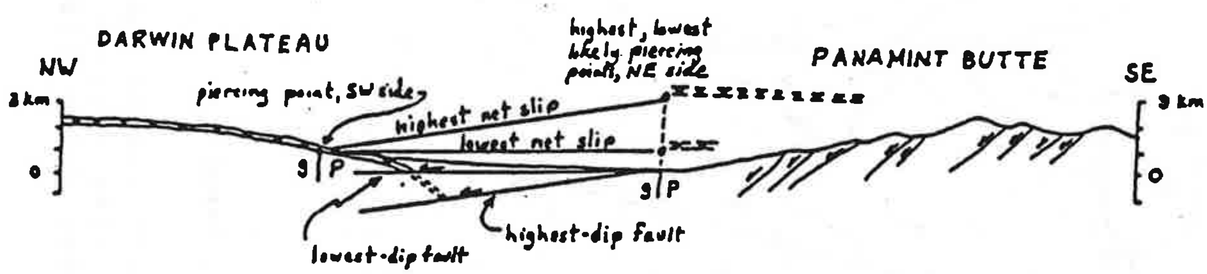




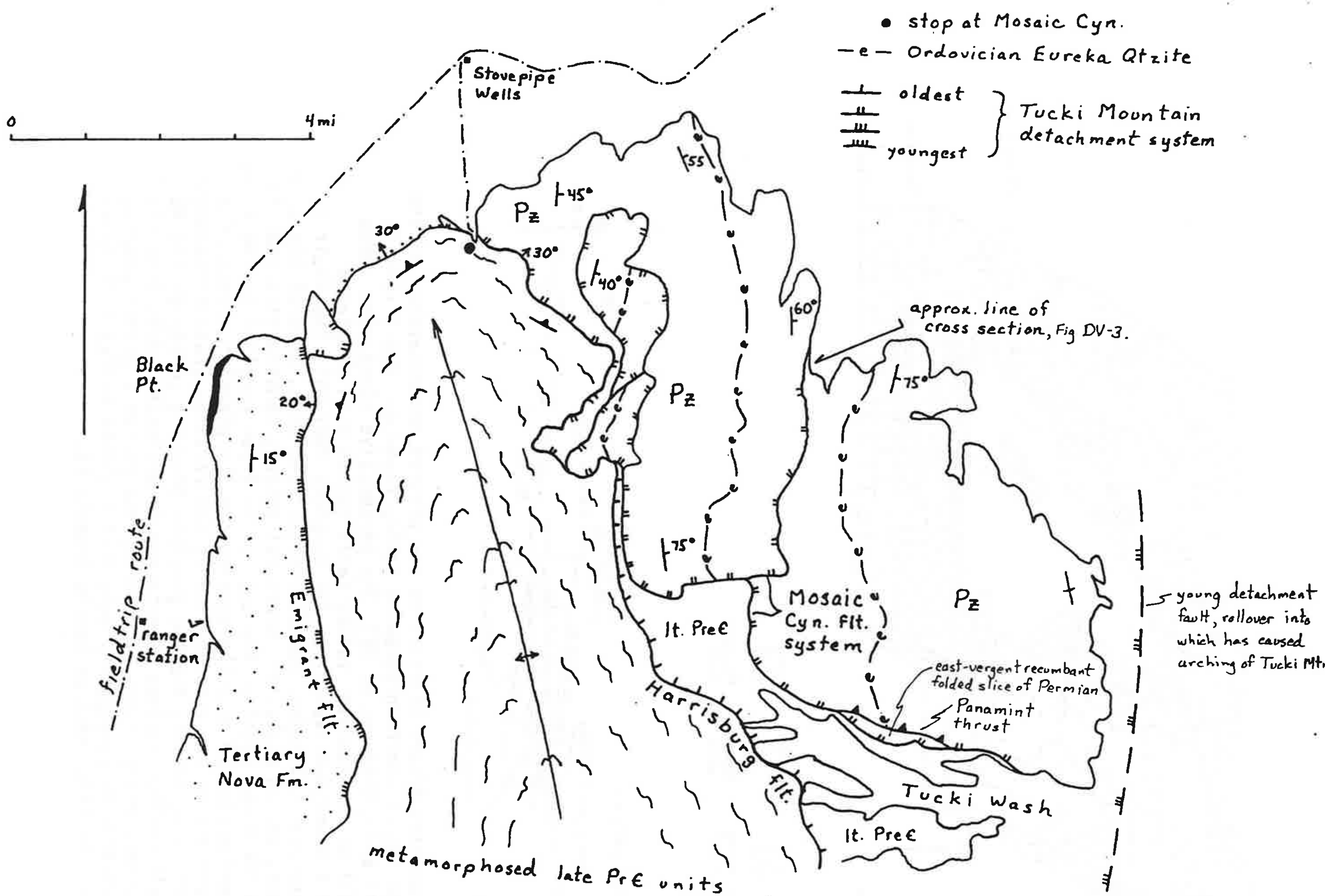


Geologic sketch map and cross section of PanamintButte, as derived from mapping of W. Hall (1971). Capitol letters designate rock systems (C = Cambrian, etc.).

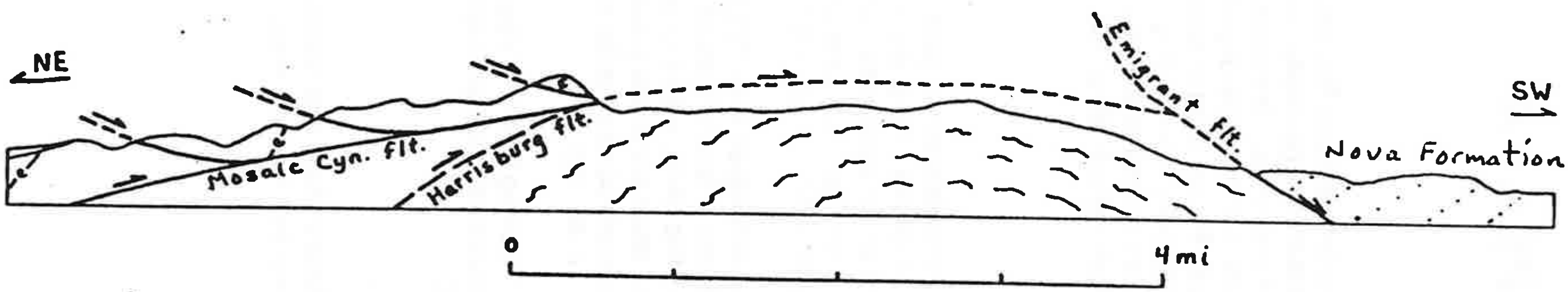




. This two-part figure illustrates our present understanding of the extensional opening of northern Panamint Valley based on recent work by Clark Burchfiel and colleagues at MIT. The perspective block view shows the hanging wall of the Panamint Valley fault having been displaced obliquely down and northwestward from its in-place footwall. The fault is shown to be very gently dipping, based on drill hole data, several kinds of geophysical data, and on a classic structural geology construction that is illustrated in the cross section. The section is drawn parallel to the trace of the fault, and projected onto the section are the piercing points of a geologic line created by the intersection of the vertical granite/Paleozoic contact with overlying basalt. The two possible piercing points on the NE side of the fault reflect maximum uncertainty in projecting data into the cross section plane. Using either of these points yields a gently dipping fault, in agreement with geophysical data that indicate valley fill to be between 0.5 and 3 km thick with no basalt beneath it.



Geologic sketch map of Tucki Mountain detachment fault complex. Schematic cross section is provided in figure 30. Geology after Hunt and Mabey (1966) and Wernicke and others (1986).



Schematic cross section of Tucki Mountain detachment complex, derived from map relations shown in Fig. DV-2. Note that view direction of section has been reversed from map so as to show the complex as it would appear to us as we look south from Mosaic Canyon.

THE PLAYA

The Playa Basin

Playas are extant arid-climate landforms consisting of fine-grained horizontal sediments in undrained basins. Playa basins are generally structural depressions in rocks of differing lithologies located in arid lands. The lower bedrock slopes are mantled by alluvial, colluvial, landslide and mudflow deposits, and the basin floor contains a salt or clay-surface playa. Playas form in direct correlation to the basin size and only if the basin is hydrologically closed. (Fig. 30a) Because of the interfingering nature of the distinct depositional processes, there is a sharp slope and sediment-size break at the playa margin. (Fig. 16)

Basin Processes

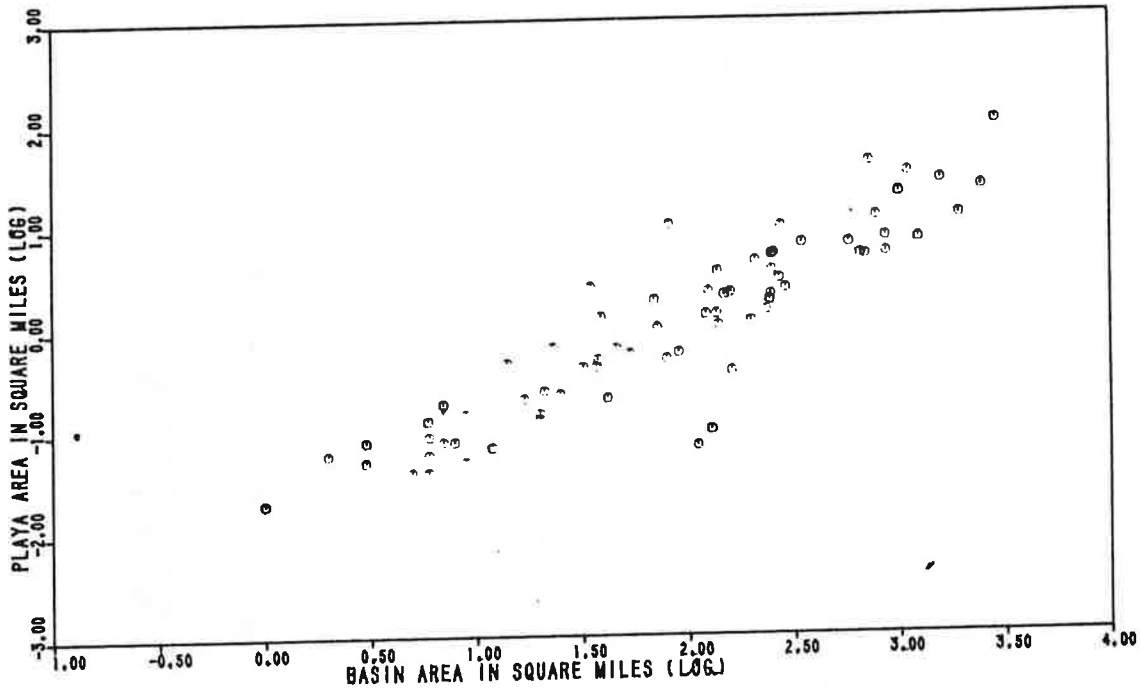
Within these arid basins are sediments deposited on the playa and lower fan by laccustrine processes, fluvial alluvial fan deposits on the fan slopes and playa margin, mudflows deposited over fan and playa sediments, eolian dune and windshadow deposits on the lower fans and playa, and local eolian deflation and reworking of the mudflow lobes and playa surface. (Figs. 88, 3, 16, 30b)

There are several interpretations of basin processes (Fig. 72), but ultimately, the basin fills, the playa migrates toward the lowest limit of closure (drain), and further playa sediments are carried out of the basin, with erosion of the sill and dissection of the playa surface. The filled basin becomes covered by aggrading fans resulting in a smooth sloping alluvial plain across the former basin, interrupted by inselbergs. (Figs. 73, 76, 77, 80, 78)

Subsurface investigation shows the interfingering layers of clay playa deposits, buried armored or desiccated surfaces, coarse alluvial fan material and deflated very coarse flow lobes. (Figs. 28, 33, 39)

Stop XX: Silver Dry Lake, (north end)

We will look at the playa/fan interface, playa surface and subsurface, flow lobes and general morphology of a clay-surface playa near the outlet (drain) in a medium-sized basin.



30a

Figure 30b

Coarse Mudflow Margin,
Lost Lake

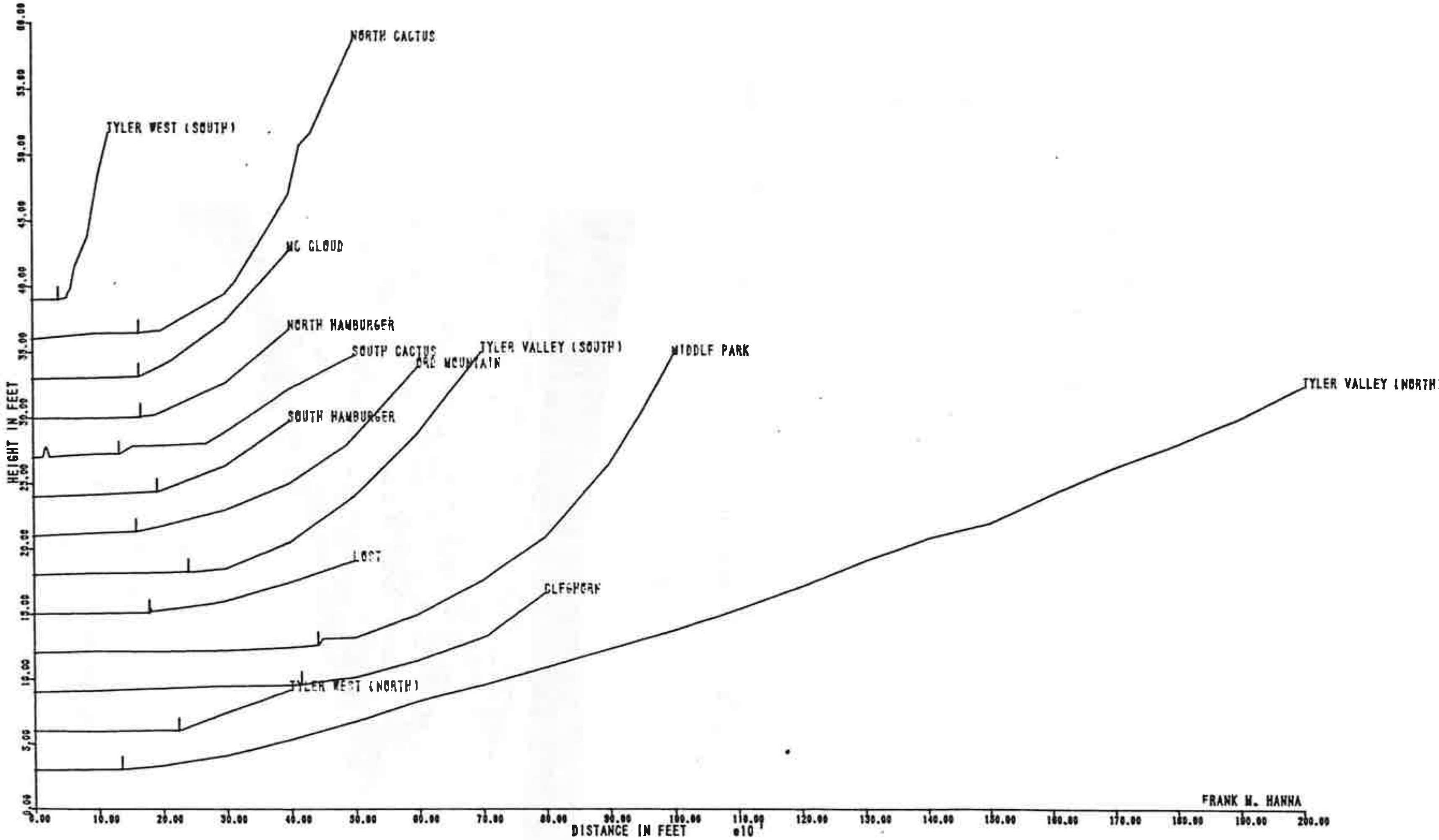


FIGURE 16

BASIN SLOPE PROFILES

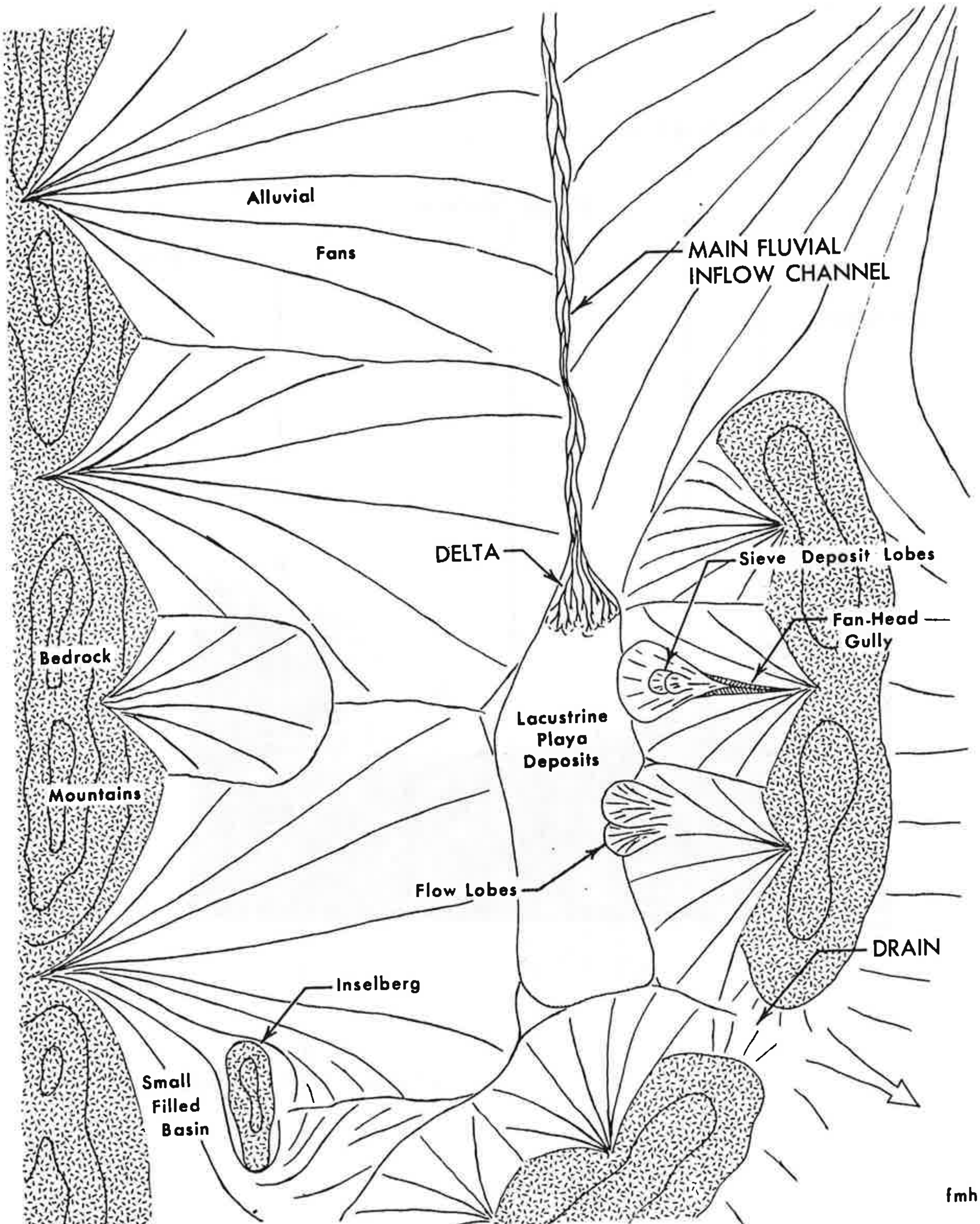


Figure 88
Schematic of Basin Process Elements

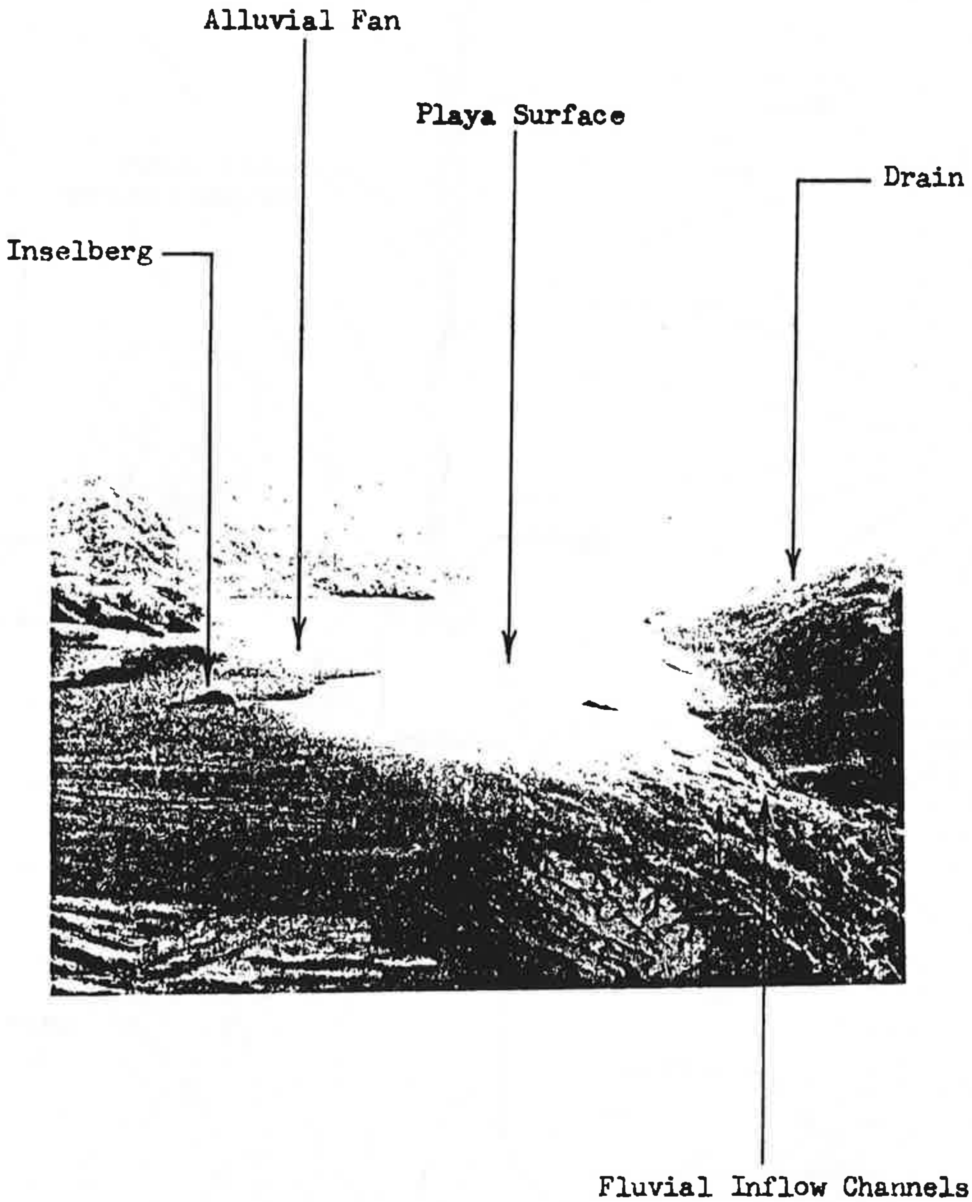
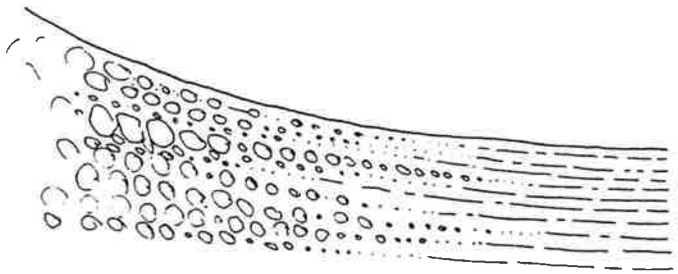
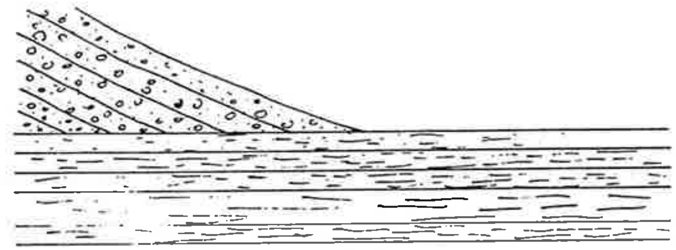


Figure 3

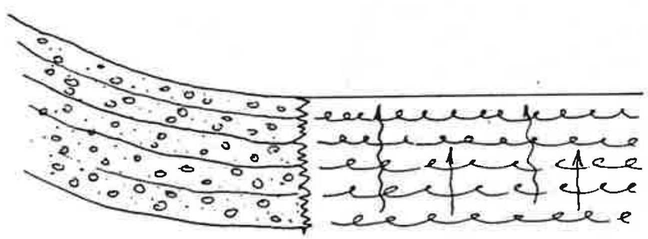
Elements of a Typical Playa Basin,
Racetrack Playa



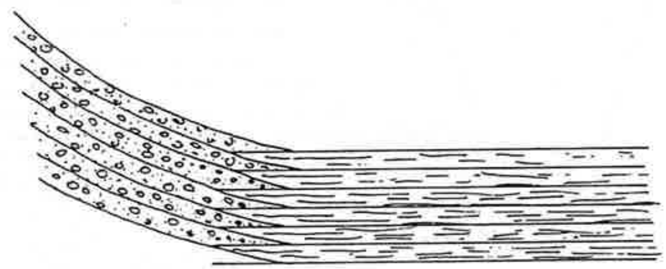
Diminishing Energy Fluvial System



Alluvial Burying of Old Lake Beds



Capillary Rising Groundwater



Alternating Mass-Wasting and Lacustrine Deposition

fmh

Figure 72

Some Interpretations of the Playa-Fan Interface

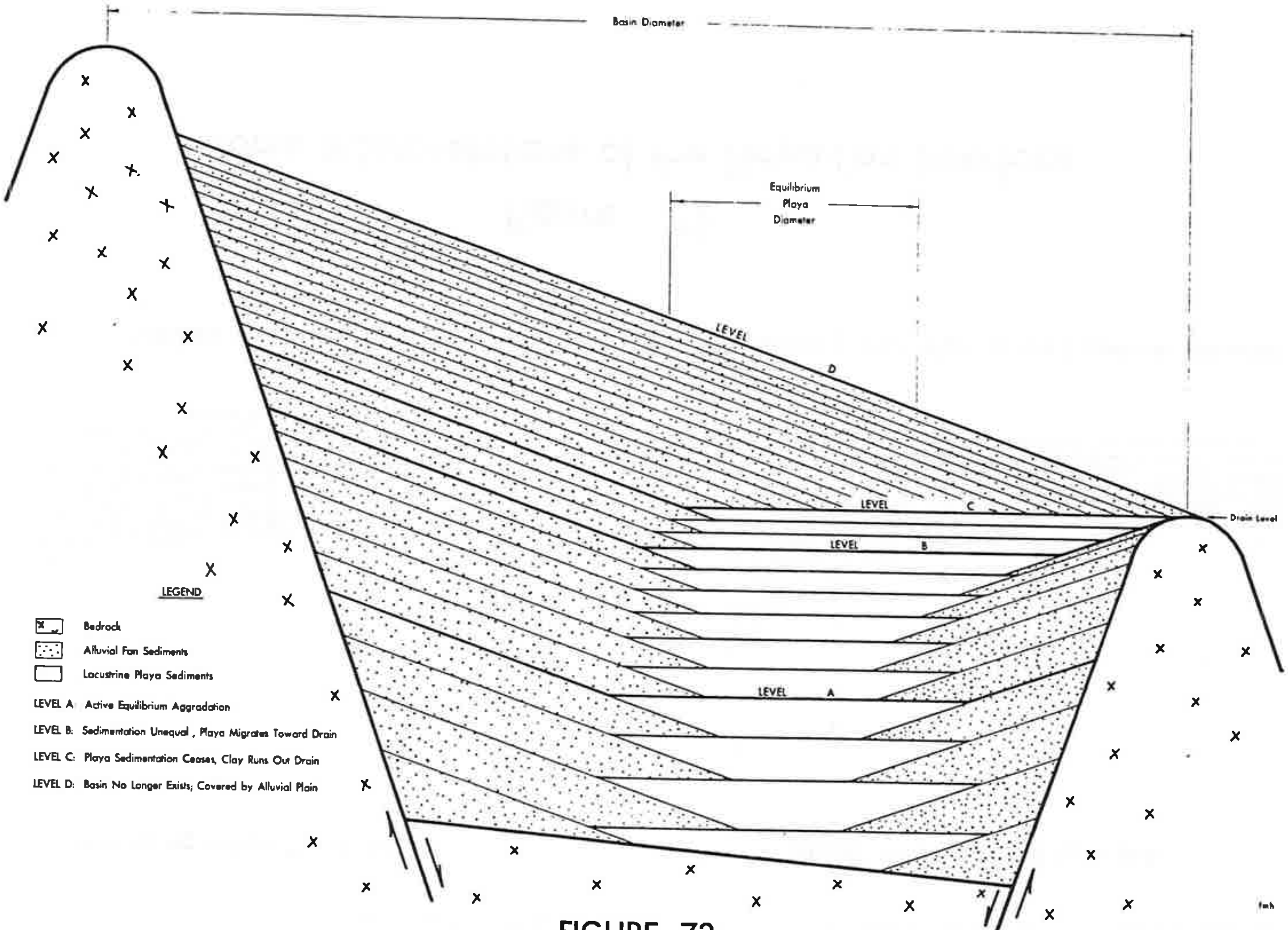


FIGURE 73

BASIN PLAYA SCHEMATIC CROSS SECTION

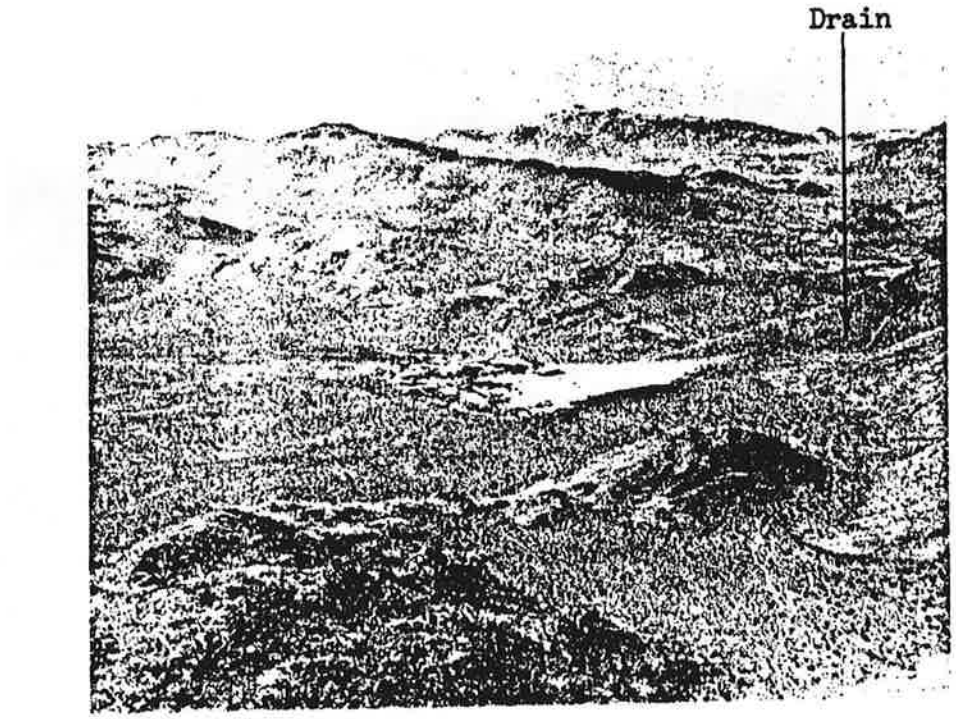


Figure 76

Playa Migrating toward Drain,
McCloud Flat

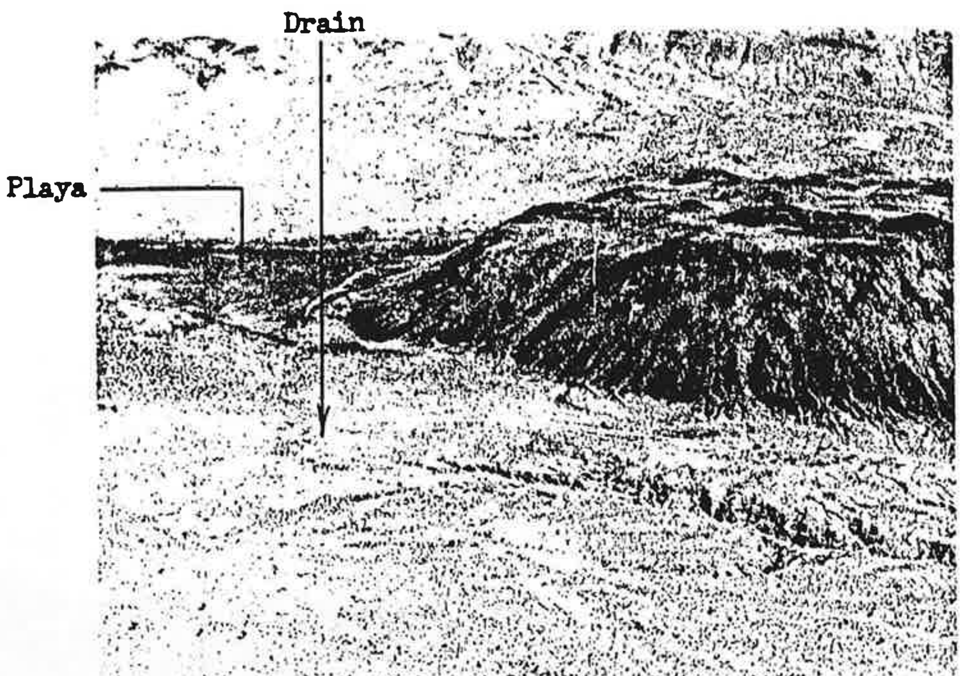


Figure 77

Playa at Drain Level,
North Cactus Flat

48

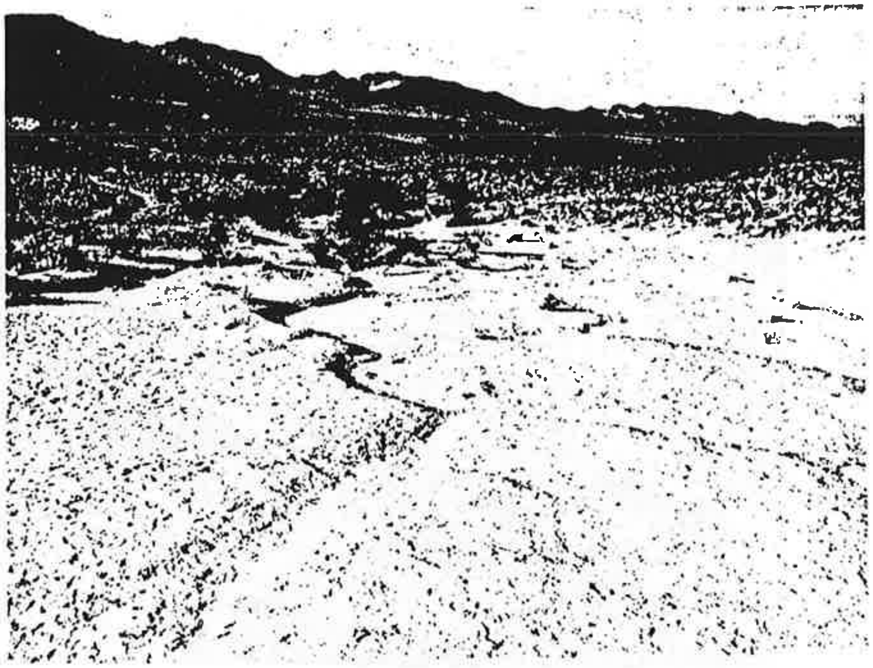


Figure 80

Playa Filled and Eroding down Drain,
Wingate Playa

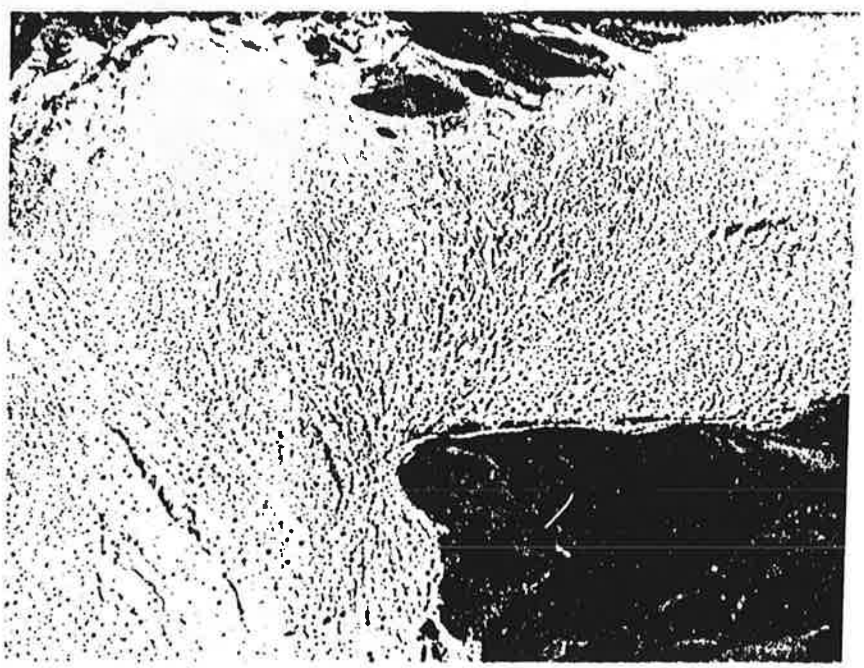


Figure 78

Filled Basin Overflowing down Drain,
Panamint Valley

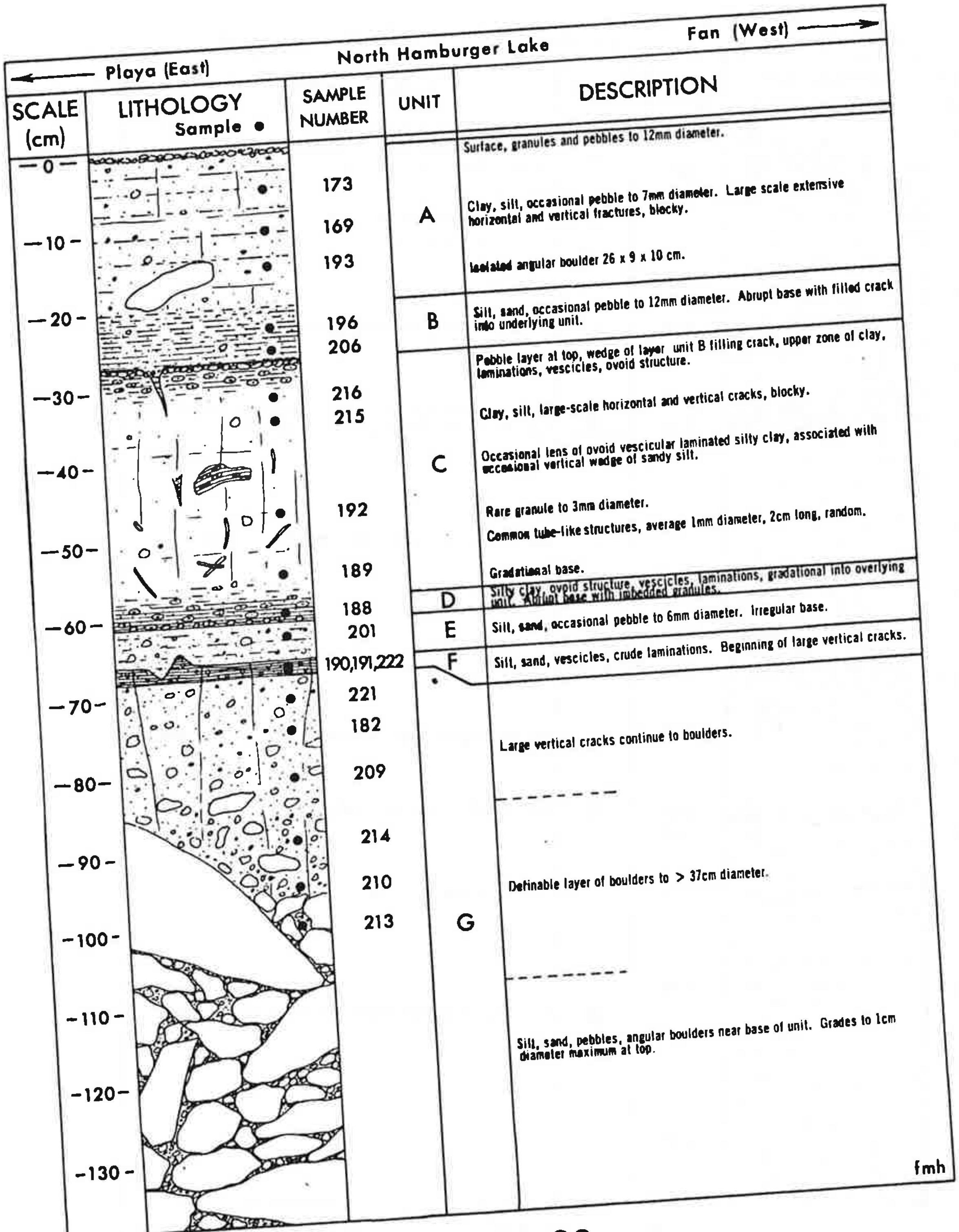


Figure 28
North Hamburger Trench Stratigraphy

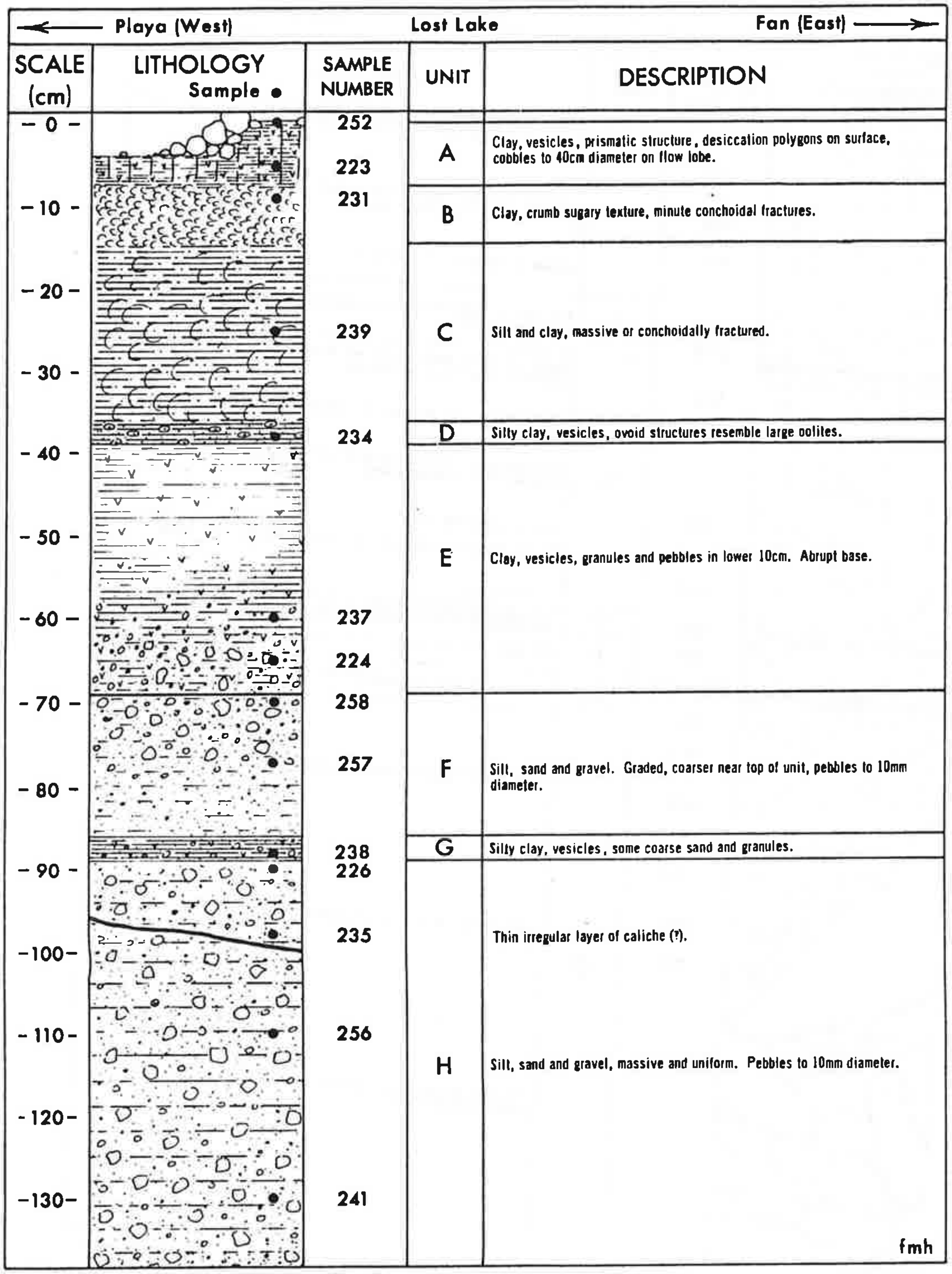
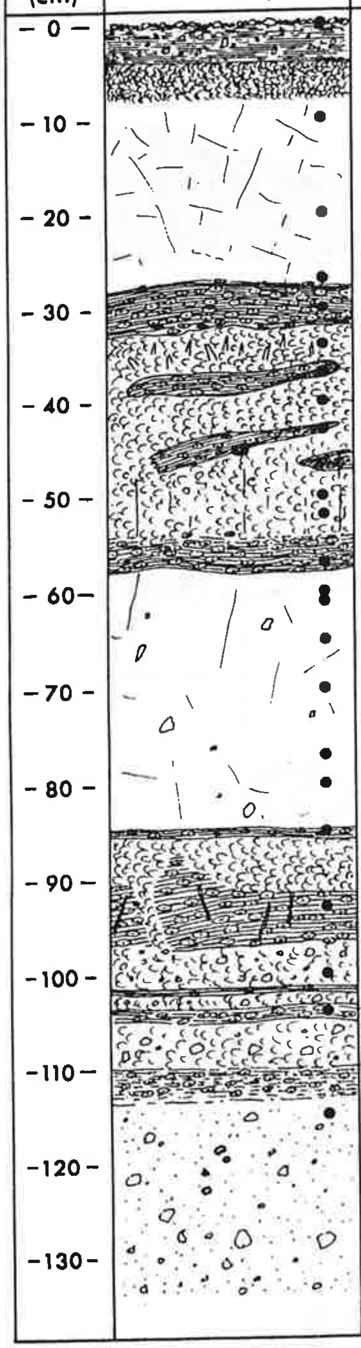


Figure 33
Lost Trench Stratigraphy

← Fan (West) McCloud Flat Playa (East) →



SAMPLE NUMBER	UNIT	DESCRIPTION
254	A	Surface granules and pebbles 4-5mm diameter. Clay with horizontal and vertical cracks, silty, fine laminations, occasional granule.
	B	Clay, crumb sugary texture, minute conchoidal fractures.
236, 233	C	Blocky, uniform massive clay. Abrupt uneven base.
261, 251, 227	D	Silty clay, fine laminations, compaction (?) structures, very coarse öolitic texture with small vesicles on the oolites. Small horizontal and vertical cracks. Abrupt base.
229, 250, 240, 248, 249, 243, 246	E	Clay, crumb sugary texture, minute conchoidal fractures, small angular voids or crystal impressions (?). Silty clay, fine laminations, coarse oolitic texture, vesicles, discontinuous lens. Clay, crumb sugary texture, minute conchoidal fractures. Silty clay, fine laminations, coarse oolitic texture, vesicles, discontinuous lens, unidentified spherical structure. Clay, crumb sugary texture, minute conchoidal fractures, larger vertical cracks.
260	F	Silty clay, fine laminations, coarse öolitic texture, vesicles.
230, 255, 218, 225, 259, 247	G	Blocky, uniform massive clay, occasional pebble.
153	H	Silty clay, fine laminations, coarse oolitic texture, vesicles.
	I	Clay, crumb sugary texture, minute conchoidal fractures.
245	J	Silty clay, fine laminations, coarse öolitic texture, vesicles, occasional vertical tube impression (?). Abrupt base.
228, 242	K	Coarse silt with common granules, occasional layer of vesicular ovoid structure, laminated silty clay.
	L	Very coarse silty clay, fine laminations, oolitic texture, vesicles.
244	M	Uniform non-graded coarse conglomerate with silt matrix, pebbles to 8-10cm maximum diameter.

fmh

Figure 39
McCloud Trench Stratigraphy

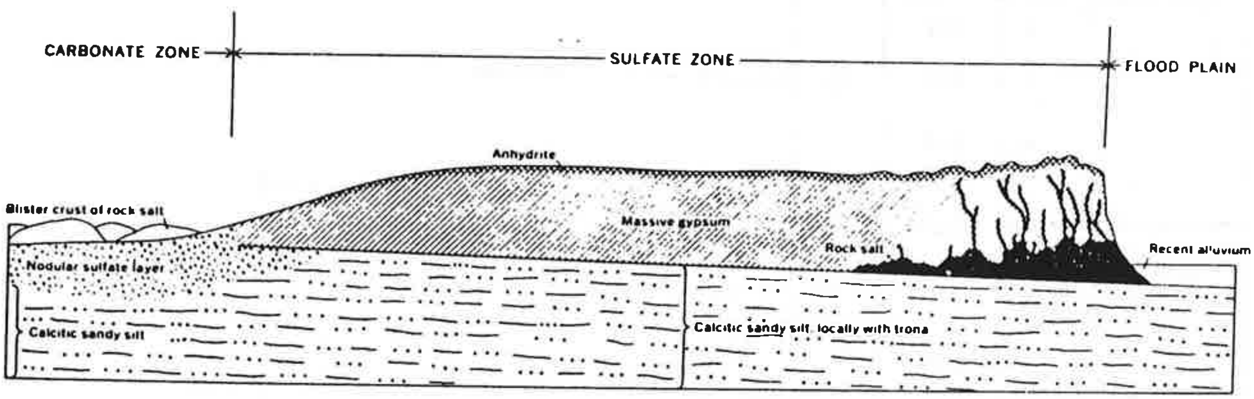


FIG. 27. Smooth, silty rock salt opposite foot of Furnace Creek fan. It has been subject to frequent washing by water running off the fan, as shown by rounded cobbles of highly porous, lightweight, vesicular lavas washed down onto surface.

FIG. 28. Diagrammatic section of massive gypsum deposit. Gypsum is 2 to 5 feet thick; uppermost layer (6 inches thick) is dehydrated to bassanite or anhydrite. Under gypsum is calcitic sandy silt. Where gypsum ends panward at a wash, as in this section, rock salt is deposited on silty sand and rises into gypsum in irregular veins, producing a rough surface analogous to rough, silty rock salt. Relationship between massive gypsum and nodular sulfate layer in carbonate zone is not known. (From U.S. Geol. Survey Prof. Paper 494-B.)

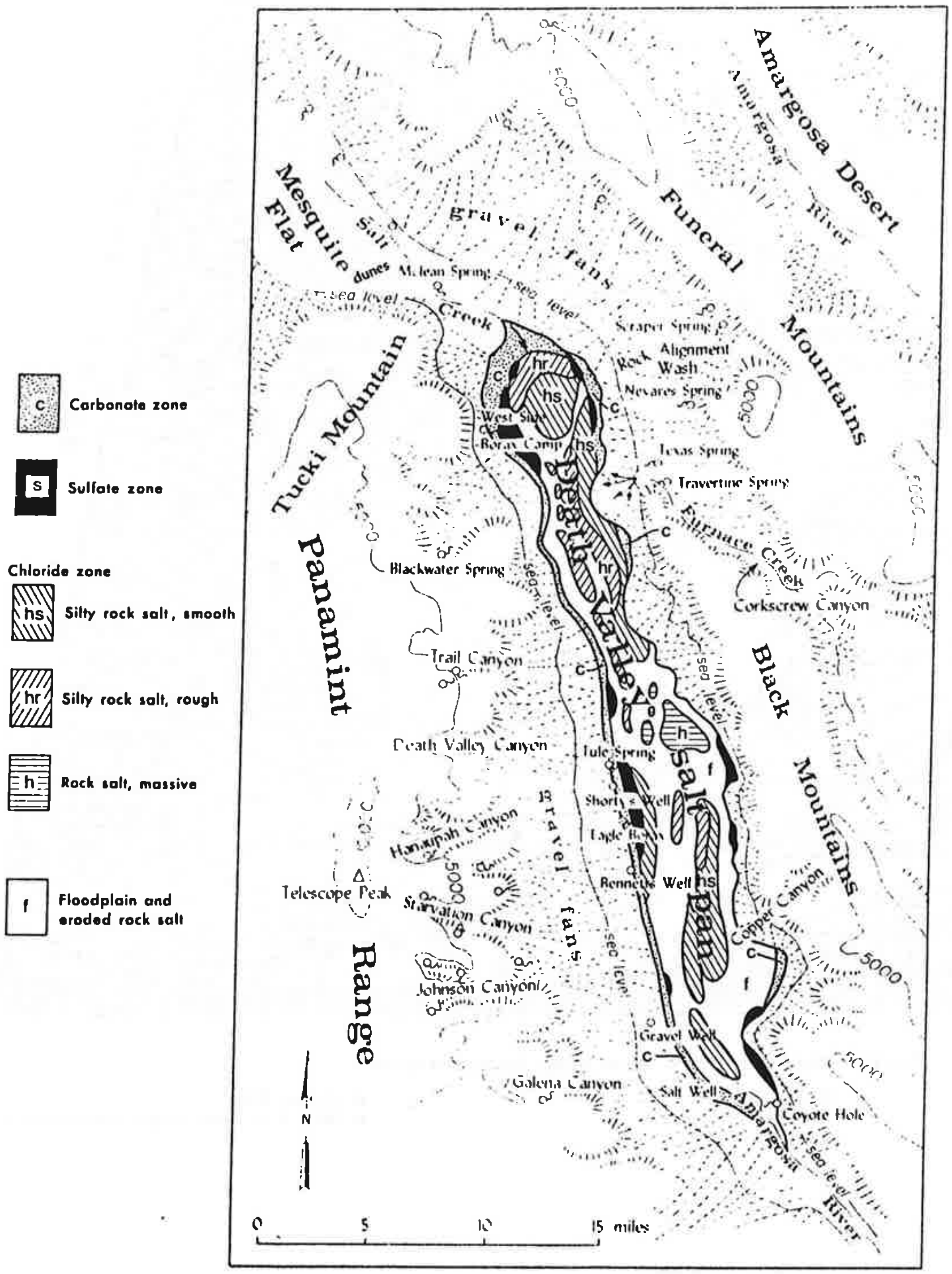


FIG. 22. Map of Death Valley salt pan.

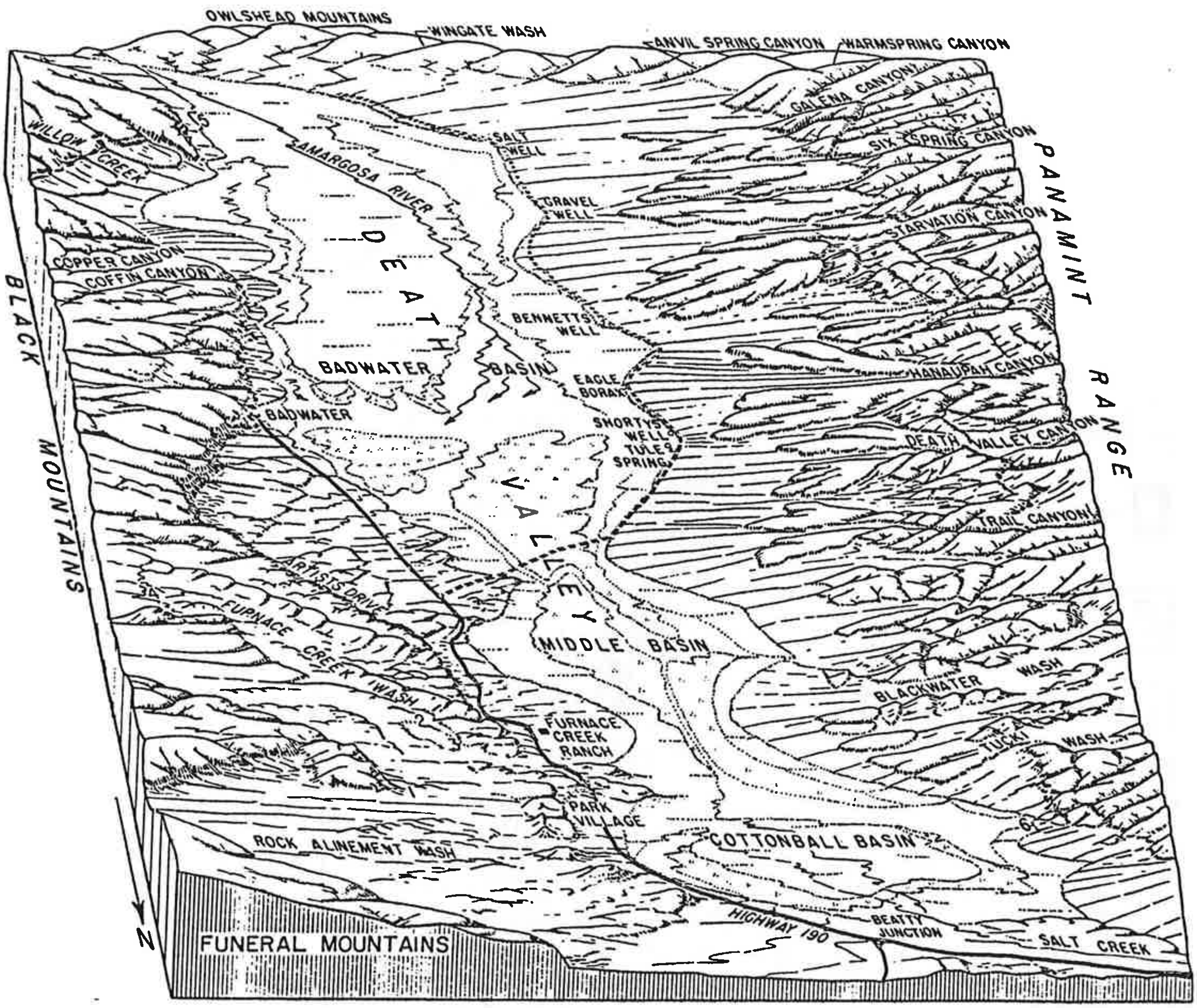


Figure 2.—Block diagram of Death Valley, Calif., looking south.

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

54

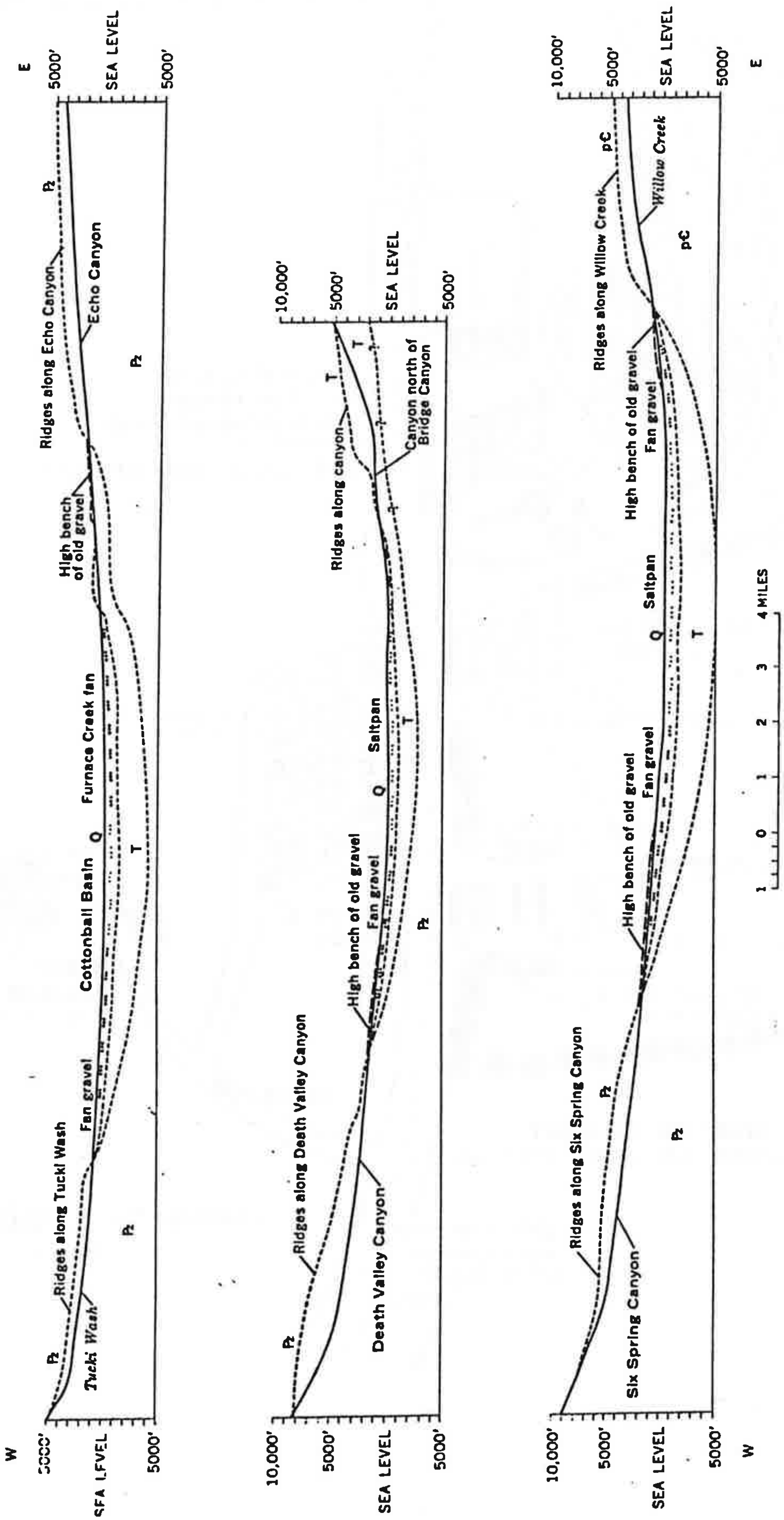
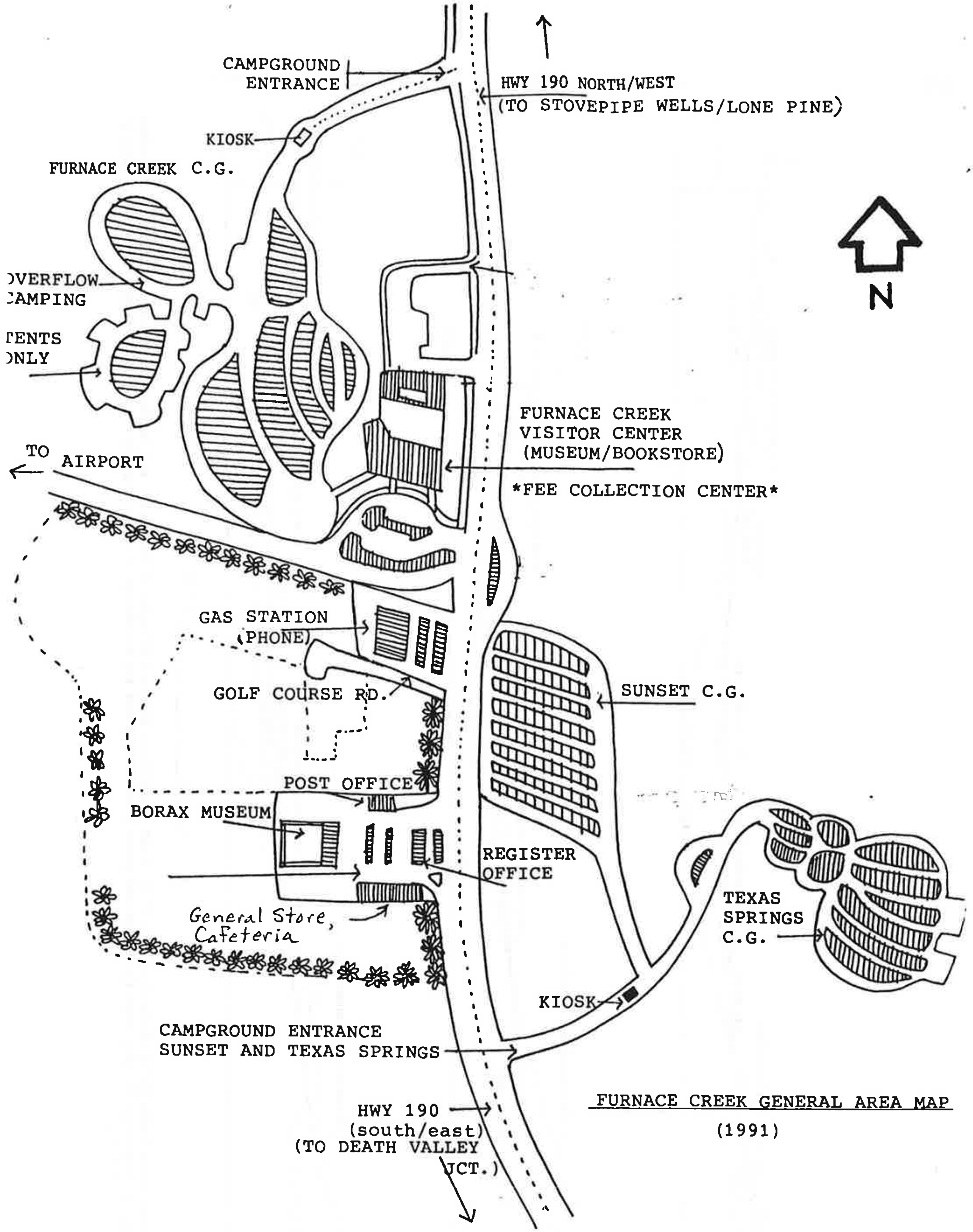
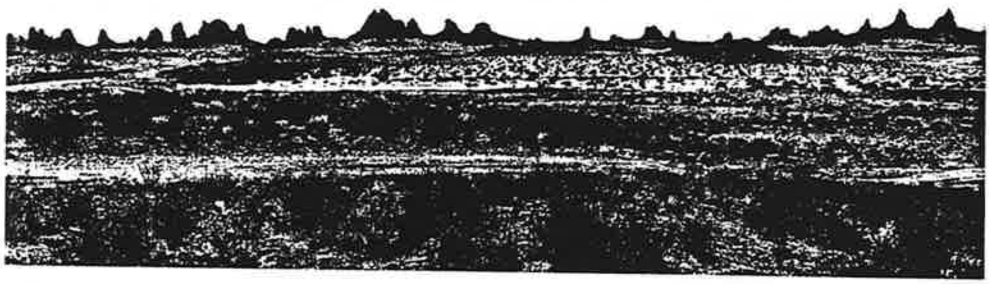


FIGURE 67.—Sections across Death Valley illustrating depth of fill under the saltpan, slope and thickness of fan gravels, and the depth and gradient of canyons in the adjoining mountains. Q, Quaternary fill (silt under the saltpan and gravel between there and the mountains); T, Tertiary fill (in large part volcanics); Pz, Paleozoic rocks; p€, Precambrian crystalline rocks.



FURNACE CREEK GENERAL AREA MAP

(1991)



A 5-mile-long tentacle projecting from the north-group pinnacles along a linear fracture. Most pinnacles in this group are 10 to 50 feet tall. —Helen Z. Knudsen photo

springs probably lined up along fractures in the Spangler Hills granitic rock that underlies the nearshore part of the Searles Lake floor. This could explain the alignment of the Trona towers and ridges and the elongated shape of other tufa features. Disintegration of the Spangler Hills granitic rock produces coarse, permeable lake-bottom deposits through which spring waters could easily feed lake-floor springs. Such springs presumably existed when the climate was much wetter than it is today.

Pick up and study fragments of tufa that have fallen from pinnacles, but return them to their resting place. A hand lens will help you examine the surface features more closely. Look at some of the larger fallen blocks to see the textures within cavernous tufas. Explore as much of the area as you like by car and foot. Travel slowly on the plexus of wheel-track roads, and watch for high centers and accumulations of windblown sand. Please resist the urge to scale the sides of pinnacles; it's hard on them.

Under a full moon, the pinnacles cast enchanting shadows that invite us to imagine this desert landscape during wetter glacial times: an expansive lake with bubbles surfacing from underwater springs and a linear chain of tufa islands that breaks the water's surface like the ridges on a dragon's tail.

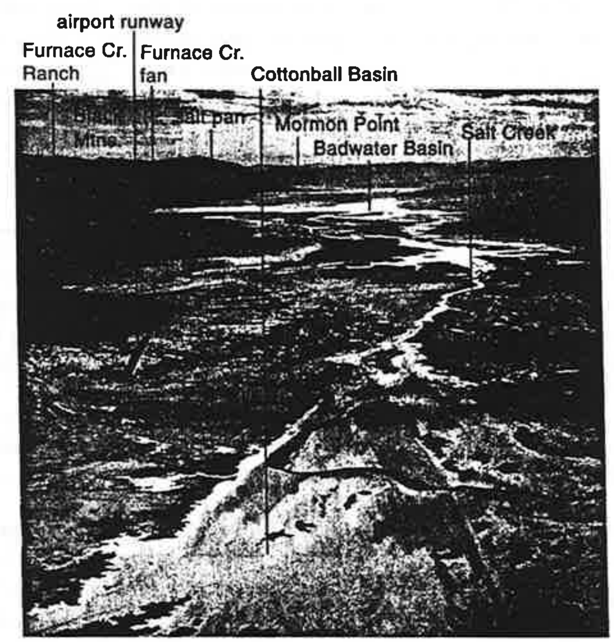
57

5

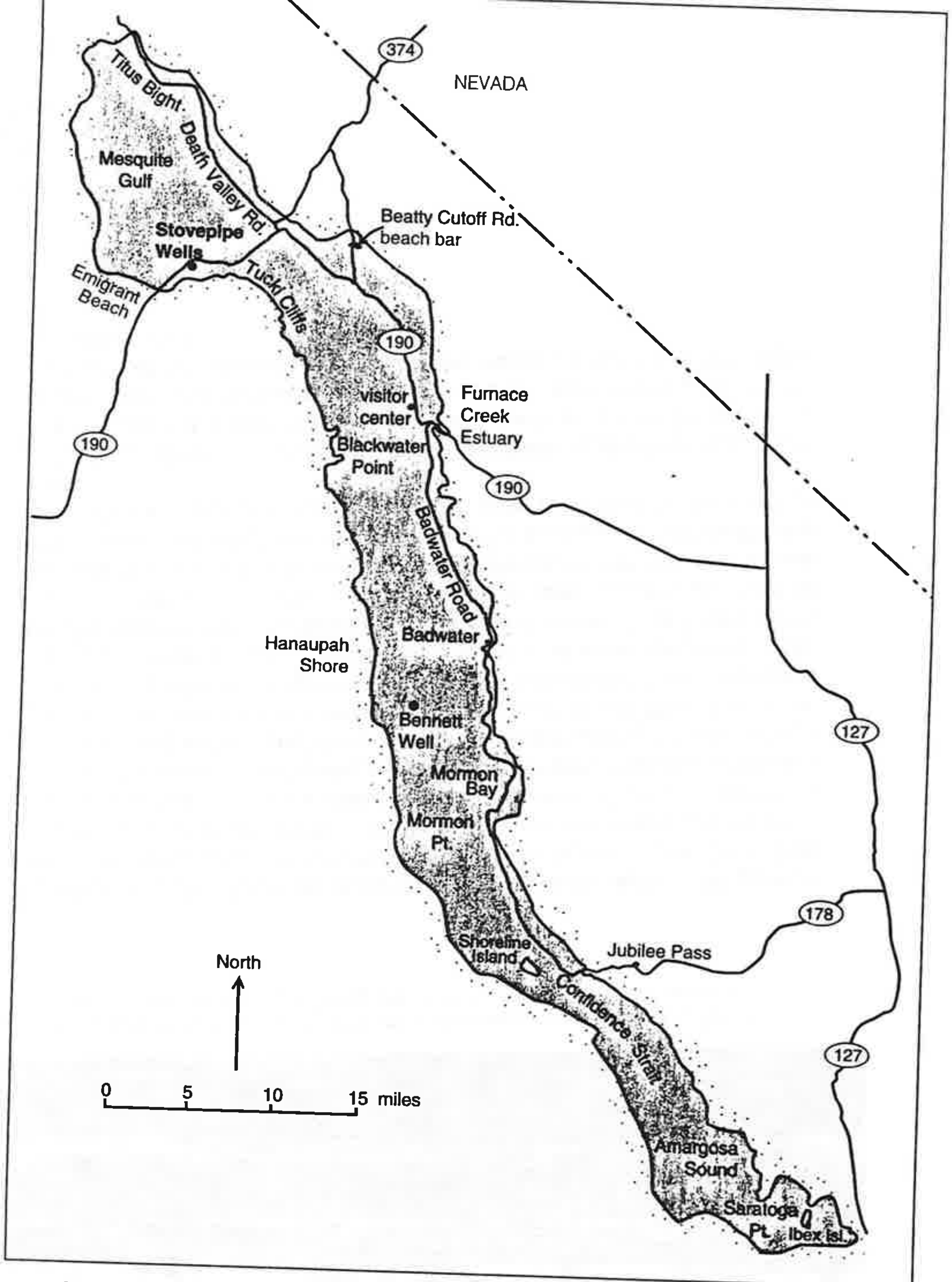
A Huge Bathtub without a Drain

— PLEISTOCENE LAKE MANLY AND THE SALT PAN —

At 282 feet below sea level, Death Valley is the lowest dry land in North America. The basins of some of the Great Lakes, Lake Chelan in Washington State, and large lakes in Canada's Northwest Territories are lower than Death Valley, but filled with water. Death Valley once held such a lake, about 600 feet deep and close to 100 miles long. That was during the Great Ice Age when the climate was colder and wetter and glaciers occupied mountains within an expanded Death Valley drainage system. Lakes formed and disappeared repeatedly in Death Valley over a period possibly exceeding a million years. Lake Manly is the name usually applied to all late Pleistocene water bodies older than 10,000 years in the valley. Their various phases are called stands. We see surface evidence for only late Pleistocene stands that existed within the last 240,000 years, the youngest of which disappeared about 10,500 years ago. Younger, much smaller, short-lived lakes may have occupied Death Valley during



View looking south down Death Valley from 3 miles north of Furnace Creek Ranch.
—William and Mary Lou Stackhouse oblique air photo



Map reconstruction showing the extent of Pleistocene Lake Manly at maximum 600-foot depth. We added some fictitious names to help you visualize the lake.

GETTING THERE: Features attesting the former existence of a large lake in Death Valley are mostly faint, fragmental, and widely separated. In the southern part of Death Valley, just east of Ashford Mill, Shoreline Butte's many strandlines on its basaltic slopes provide convincing evidence of the lake's former presence. The strandlines are most distinct on the butte's north and northeast flanks. The next best set of strandlines is in the cove just east of Mormon Point, where the Black Mountains front is offset more than a mile to the east.

The steep western front of the Black Mountains from this cove north, nearly to Artists Drive, retains scattered remnants of horizontal strandlines, some more than 300 feet above the base of the mountains. Especially good places to look are along Badwater Road on the flank of Copper Canyon turtleback (vignette 9) opposite milepost 42, between mileposts 47 and 48, at Badwater, and for 0.5 mile north of Badwater. Faint strandlines cross the west face of low hills north of the Park Service's residential complex. By far the best evidence of Lake Manly farther north are beach ridges along the Beatty Cutoff Road to Daylight Pass. Badwater Road bisects the largest and best ridge in a double-walled roadcut 1.8 miles northeast of its junction with California 190.

the last few thousand years, especially one known only as the Recent Lake that started about 5,000 years ago and possibly lasted for 3,000 years. If water inflow was unlimited, Death Valley would be part of a huge lake 1,950 to 2,000 feet deep covering a large part of the Mojave Desert.

Water levels in your household bathtub can leave rings, which remain until you scour them away with a bit of elbow grease. Remnant rings on Death Valley's tub are called strandlines. They consist of wave-cut cliffs and benches, deposits of calcareous tufa, or accumulations of shoreline gravels. Such gravels commonly contain smooth, well-worn, tabular pebbles, flattened by sliding up and down beaches in the swash of breakers. In some deposits, these stones overlap each other like shingles on a roof. Geologists regard shingled gravels as good evidence of lakeshore conditions. Like any good housekeeper, Mother Nature busily scours away the rings on her bathtub every time a stand of a lake dries up. Consequently, rings have been fully erased in many places and are only faintly visible in others.

The best-preserved succession of Lake Manly strandlines lies on the flanks of Shoreline Butte in the southern part of Death Valley. The butte, formerly an island in Lake Manly, was battered by wind and waves on all sides. Especially large waves generated by storm winds with a long over-water fetch vigorously attacked the butte's north and northeast flanks, creating well-defined strandlines there. Strandlines in the relatively tough 1.5-million-year-old basaltic lavas of the butte have resisted nature's scouring rather well. Strandlines are more obvious when partly shadowed under low sun angles. See how many you can count. Geologists who have carefully explored these slopes have counted more than

View looking south from the West Side Road to the northeast flank of Shoreline Butte. Linear streaks on the butte are Lake Manly strandlines.

dozen. Some strandlines consist of beach deposits, which may not have obvious topographic expression.

The strandlines tell us several things about Lake Manly. Their number indicates that water depth fluctuated frequently, seldom stabilizing long at any level. Lakes with stable overflow sills, or outlets, and sufficient inflow to continually overflow at the outlet form few but strong strandlines, but such was not the situation at Lake Manly. Lake Manly had no outlet; it was a bathtub without a drain. During periods of wet climate the water level rose, and during dry times the water evaporated and the level fell.

A bathtub is a closed depression. If we plug the bottom outlet and close the overflow drain, the difference between the bottom and rim is its closure depth. In most domestic tubs, the closure depth is between 12 and 20 inches; in Death Valley it is something greater than 600 feet. Even if the water filled the basin, water would not escape to the ocean from the southern California desert region until water was 2,000 feet deep in Death Valley. The region enclosed by the shoreline of that lake would drain into Death Valley—the ultimate sump—as the water level fell, except for small temporary bathtubs from which the water would evaporate.

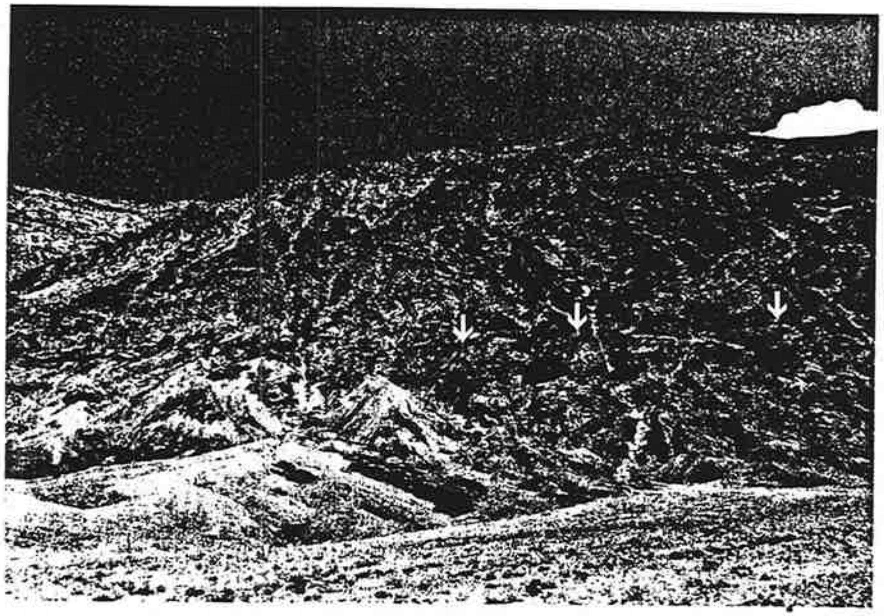
Strandlines on the butte also tell us something about the maximum water depth in the late stands of Lake Manly. The butte's summit is 663 feet above sea level. The highest confidently identified strandline is at 285 feet above sea level, although tantalizingly faint suggestions of still higher strands exist. The highest strandline on Shoreline Butte is therefore about 570 feet above the lowest point on the valley floor, 282 feet below sea level. Over the lowest part of the valley near Badwater, strandline elevations on the western front of the Black Mountains also suggest a lake close to 570 feet deep.

That depth seems reasonable, but faulting has complicated the estimates of maximum lake depth. Major strike-slip faults flank Shoreline Butte's northeast and southwest sides. The northeast fault has experienced significant lateral displacement within the last 600,000 years. Its southeastward trace in lavas along the butte is marked by remnants of a west-facing scarp up to 40 feet high, which suggests at least a modest component of vertical uplift. This means that movement on faults may have changed the height of the Shoreline Butte strandlines at least a few tens of feet. Also, the floor of Death Valley may not yet have sunk to its present 282 feet below sea level by Lake Manly time. You can understand, then, why geologists hedge by saying that the maximum depth of Lake Manly was probably between 500 and 600 feet.

View looking east to a fault scarp of basalt along a segment of the Death Valley fault zone just northeast of Badwater Road and 1.4 miles north of Shoreline Butte. Scarp is up to 40 feet high.

Another good place to see strandlines is near Mormon Point in the cove formed by the 1.3-mile eastward offset of the Black Mountains front. High waves driven by strong storm winds from the north with a long over-water fetch cut these well-developed strandlines into weakly consolidated fanglomerates.

Travelers on Badwater Road between Mormon Point cove and California 190 see many places where the steep bedrock face of the Black Mountains retains faint horizontal marks of Lake Manly strandlines, 200 to 300 feet above the valley floor. Most of the marks are carbonate-cemented gravels, but not all gravel remnants adhering to the bedrock face are related to the lake; some are remnants of the upper reaches of alluvial fans left by the subsiding Death Valley floor. To identify strand-



Remnants of a horizontal, dark, cemented gravel deposit trace a Lake Manly strandline along the western face of the Black Mountains on the flank of the Badwater turtleback. Debris cone left foreground. —Helen Z. Knudsen photo

lines, watch for strictly horizontal features, for as you can see, modern fans contact the mountain face at various elevations. Features made by standing water are always horizontal unless the land has tilted since their formation. Remnants of young (post-Manly) fault scarps 75 feet high along the base of the Black Mountains indicate the lake may have been at least 75 feet shallower than the Black Mountains strandlines suggest.

Strandlines on Death Valley's west side are harder to see. Possible deltaic deposits sit at about 165 feet above sea level on the alluvial apron below Wingate Pass, through which water may have flowed from Pleistocene Lake Panamint. Anvil Spring, Warm Spring, and Hanaupah fans at the base of the Panamint Mountains reportedly bear strandlines. Some of the best westside shoreline features lie on a basaltic knob between Blackwater and Tucki washes, across from Furnace Creek Ranch. Obvious westside strandlines are sparse for at least two reasons. They lie in the Panamint Mountains' shadow under prevailing westerly winds, so waves on the western shore were more gentle than those on the eastern shore. The west side also has few surfaces of the right age on which good strandlines could form and be preserved.

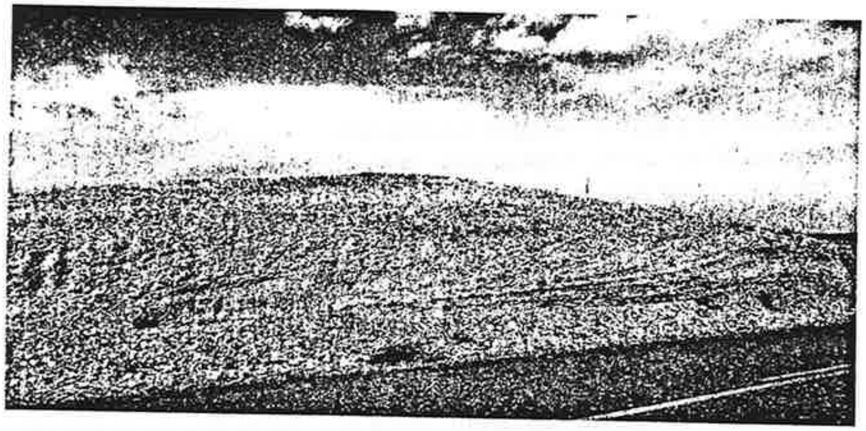
North from Furnace Creek Ranch, the best features along the eastern shore are shingled gravels and narrow terraces on low hills east and

The Daylight Pass (Beatty Cutoff) Road crosses a Lake Manly beach ridge at the double-walled roadcut, 1.8 miles from California 190. Darkly varnished desert pavement, lower right, mantles a smaller and lower beach ridge. —Helen Z. Knudsen photo

north of the Park Service residential area, plus some well-developed beach ridges along Beatty Cutoff Road to Daylight Pass. The largest of these is 1.8 miles northeast of the junction with California 190, about 9 miles north of the visitor center. Essentially a spit, it projects 400 to 500 yards east from low hills. It is several hundred feet wide at most, round topped, a maximum of 30 feet high, and steeper on its south, lake-facing side. The beach ridge consists of clean, well-sorted, and smoothly worn tabular beach gravel, seen in roadcuts to be cross-bedded and shingled. Slopes veneered by beach gravels are paler than adjacent alluvial deposits, but some stones on the beach ridge crest are darkly varnished.

The top of this beach sits 150 feet above sea level, which suggests a maximum lake depth around 430 feet here, ignoring possible fault-related complications. Longshore currents swinging out from hills to the west, where they picked up stones, probably helped waves build this ridge. You can see other smaller beach ridges east of Beatty Cutoff Road as you approach the California 190 junction. Watch for an especially obvious one, 0.6 mile from California 190, with a stream-cut face on its upslope side.

Let's be generous and assume a maximum lake depth of 600 feet. The lake's north end, then, would have been near the northern end of Mesquite Flat about opposite Titus Canyon, and the south end about 100



Beach gravels with cross-bedding in the Lake Manly beach ridge at the roadcut along the Daylight Pass Cutoff Road to Beatty. —Helen Z. Knudsen photo

miles away in the broad flat valley north of the Avawatz Mountains and south of the Ibex Hills. In central Death Valley the lake's width ranged between 7 and 8.5 miles, but to the north and south the lake was wider and its shoreline more irregular.

Any lake 100 miles long and 600 feet deep contains a lot of water. Where did it all come from, considering the aridity of the surrounding country? The climate around Death Valley was generally cooler and wetter during the Great Ice Age, when huge ice sheets covered large parts of North America, than it is today. Although none of the mountains immediately adjacent to Death Valley bore glaciers, they certainly had considerable snow, and glacier-bearing mountains lay within the expanded Death Valley drainage area. The Sierra Nevada harbored huge ice streams, and even the San Bernardino Mountains had a few small glaciers on the highest peaks. Geologists call such cooler, wetter conditions in desert areas pluvial, rather than glacial.

At least two, and probably three, large pluvial rivers emptied into Death Valley: the Amargosa, rising from the east in the lofty Spring Mountains of western Nevada (Charleston Peak 11,919 feet); the Mojave River (vignette 1) from the south, born among 8,000-foot peaks in the San Bernardino Mountains; and the Owens River from the west, draining the ice- and snow-covered 13,000- to 14,000-foot peaks of the eastern Sierra Nevada. Water arrived from the Sierra Nevada after passing through four large pluvial lakes: Owens (vignette 18), China, Searles (vignette 4), and Panamint. Besides runoff from local mountains, groundwater rising to the surface in springs, mostly on the east side of Death Valley and along the Amargosa River, contributed significantly to the nourishment of Lake Manly. This slower and more sustained groundwater flow helped moderate Manly's water-level fluctuations.

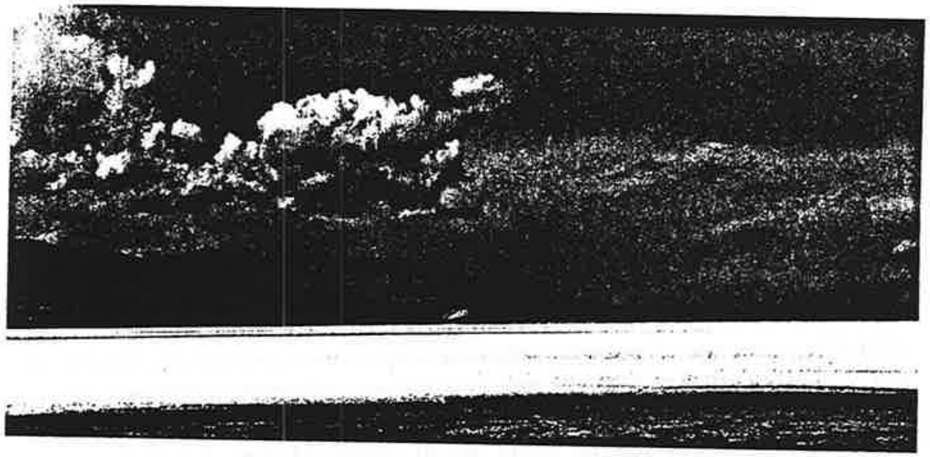
Today, despite arid conditions, groundwater makes Death Valley one of the best-watered parts of southern California's deserts. Large springs, including Nevares, Texas, and Travertine, grace its eastern side, and Travertine Springs sustain Furnace Creek, one of the two perennial streams in the valley. The other perennial stream, Salt Creek, is maintained by groundwater discharge from the Mesquite Flat drainage area, which relatively impervious strata in the Salt Creek Hills forces to the surface. Much of the meager surface water of the current salt pan comes from groundwater seepage. Large floods of the Amargosa River still flow, albeit infrequently, all the way to Badwater Basin. Today, Mojave River water (vignette 1) travels on occasions only as far as Silver Lake playa, just north of Baker, and the Owens River has flowed only as far as Owens Lake in historical times.

Long-enduring uncertainty concerning discharge of pluvial Owens River water from Lake Panamint over Wingate Pass and into Lake Manly remains, although considerable evidence, including the Wingate delta, indicates that such a discharge indeed occurred. A modern study of pluvial Lake Panamint strandlines suggests that the lake attained a depth of 1,000 feet, high enough to send overflow through Wingate Pass. As in Death Valley, however, faulting in Panamint Valley introduces some uncertainty as to the maximum water depth.

Conditions attending the Ice Age lasted for more than a million years. The stages of Lake Manly addressed here existed only at the tail end of that period, probably the last 240,000 years. Scientists recognize two stages of the lake, the older with a water depth around 600 feet and the younger with maximum depth around 300 feet.

The last Pleistocene lake deposited 25 to 50 feet of sediment in Death Valley. Two drill holes on the valley floor penetrate 1,000 feet of alternating salt and mud layers without reaching bottom. At the accumulation rate of the last stand of Lake Manly, only 200,000 to 400,000 years would be required to form 1,000 feet of lakebeds. Records of western mountain glaciations suggest that larger and longer-lived lakes possibly occupied Death Valley in pre-Manly time, provided the valley had by then become a sump. Unfortunately, we do not know when that occurred. The story of lakes in Death Valley does not, however, end 10,000 years ago with evaporation of the youngest Lake Manly. Read on.

The present salt pan is not a leftover from Pleistocene Lake Manly. Rather, judging from native campsites and food storage structures of known age, a smaller lake or lakes no more than 30 feet deep and only a few thousand years old occupied the valley floor. As the water evaporated, the salt pan formed. The salt pan occupies three separate but interconnected basins: Badwater, by far the largest, to the south; Cottonball, the



View looking west at the salt pan near Badwater. —Helen Z. Knudsen photo

next largest, to the north opposite the Harmony Borax site; and long, narrow Middle Basin connecting them. Small, perennial Salt Creek and its floodplain traverse Middle Basin.

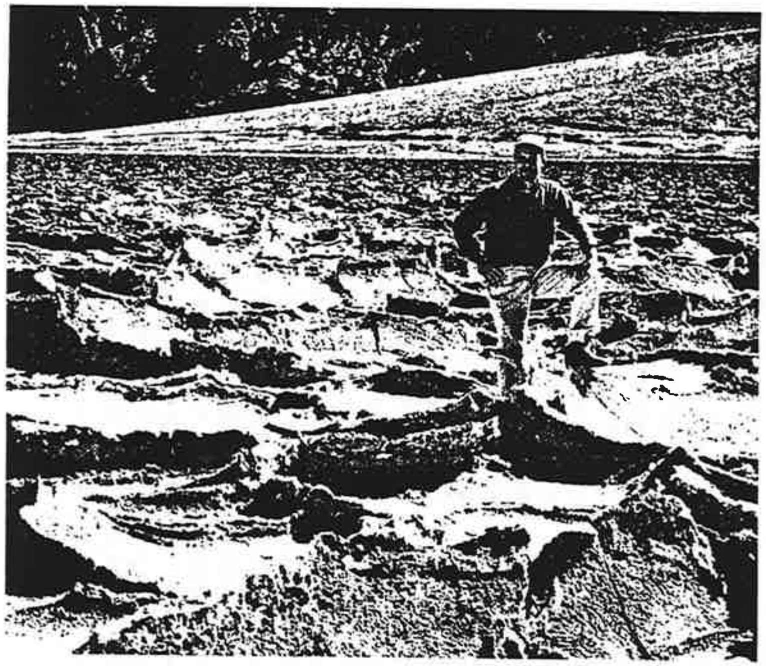
Evaporation rates in Death Valley are among the highest in North America, measured at 120 to 150 inches per year from evaporation pans. The loss from lakes is certainly lower, probably in the neighborhood of 65 to 75 inches, owing to higher humidity over water bodies and lower ambient temperatures. Wind plays an important yet unevaluated role, increasing evaporation by removing humid air.

Viewed from a distance, Death Valley's salt pan looks uniformly flat, smooth, and pristine white. Close-up inspection shows, however, that it is anything but uniform. Its chemical composition is complex and zoned, its small-scale surface features such as knobs and hollows at Devils Golf Course are rugged. Areas not recently flooded develop polygonal cracks that evolve into salt saucers. A stream course 10 feet deep locally scars the pan's surface, and many shallower stream channels border higher ground. Gentle folds and faults of small displacement deform the pan. The surficial solid salt layer, 1 to 6 feet thick, lies on top of salty mud. Where silt and clay adulterate the surface salt, as along the pan's edge or in stream channels and floodplains, the pan surface has an uneven puckered crust, like a rich cookie or pie.

To most people, salt is the condiment we sprinkle on food, the compound sodium chloride (NaCl), but other compounds are also salts. Waters flowing into Death Valley also carry sulfate and carbonate salts, mostly of the element calcium. Sodium chloride is the most soluble, followed respectively by sulfates and carbonates. Upon evaporation, carbonates are deposited first, followed by sulfates, and last by sodium chloride—each salt forming a layer in the deposit. The dominant salt in each layer is contaminated to some degree by the other salts. As a lake shrinks and its shoreline regresses, bands of the dominating salt form around the basin's edge, corresponding to the layers laid down within the basin. The outermost band is predominately carbonate, then comes a sulfate-rich band, and finally sodium chloride makes a veneer over the rest of the lake floor, commonly covering 50 percent or more. Because of the proximity of volcanic-derived sedimentary rocks, Cottonball Basin also has a concentration of borate minerals. In the 1880s, borate nodules (then referred to as "cottonball") were mined and hauled out using the famous twenty-mule teams.

View looking west at the rough salt surface of Devils Golf Course. Snow on Telescope Peak in upper left. —Helen Z. Knudsen photo

As the Recent Lake evaporated, it left salt bands of different composition and width along its shores. If evaporation had been the only factor causing a drop in water level, the bands would be roughly the same width in all directions because of the smooth and symmetrical configuration of the lake basin. The bands of the Recent Lake salt pan, though, are wider on the west side of the salt pan. If the basin was tilted gently down to the east while the water evaporated, the lake would recede more from the western shore than from the eastern shore. That would make the salt bands wider along the western shore than along the eastern shore. Tiltmeter and leveling surveys show that eastward tilting is occurring today; bands in the Recent Lake salt pan demonstrate that the basin was tilting up to 5,000 years ago.



Salt saucers on a long-unflooded part of the salt pan a little south of Badwater. Growth at the edges causes the upturn.

Areas of the salt pan that are regularly flooded by streams or by precipitation runoff and spring seepage tend to be smooth. Some salt dissolves with each flooding and precipitates again as a smoothing veneer when the water evaporates. Contraction of the salt pan during intervals of drought creates polygonal fractures in the salt. These polygons grow at their edges as salt water seeping up the cracks evaporates. The growing polygons push against each other, causing the edges to bend up to form salt saucers. Extensive flooding of the pan's surface, as happens in wet years such as 1969, can completely erase a field of saucers. In that year, a lake 1 to 3 feet deep lay over the Badwater Basin, and one adventuresome soul rowed across Death Valley in a boat.

You'll find a good place to walk onto a pristine pan surface near the toe of the first fan south of Badwater. Don't be intimidated by the narrow zone of moist and squishy mud at the pan's edge. Unless flooded, the pure salt pan is easy walking. The beaten path extending out from the Badwater pond is less satisfactory for viewing because foot traffic has trampled the salt pan.

Distributary channels of the Amargosa River describe broad, gently curving, semicircular patterns on the pan around the northwestward projections of the Mormon Point and Copper Canyon turtlebacks (vignette 9). The river is arching around areas of higher elevation. Ongoing deformation of a partly buried turtleback could create a higher area on the salt pan above the buried turtleback. Another possible explanation is that sediments overlying a buried turtleback are thinner, and thus less compacted, than the sediments elsewhere. The difference in compaction could create an area of slightly higher elevation. Faults displace the pan's surface by a few feet along the southwest projection of the fault ridges near the entrance to Artists Drive. An 18-inch-high sill, possibly formed by upwarping, separates the Badwater and Middle Basins.

Let your imagination picture Death Valley filled by a lake hundreds of feet deep. Take an imaginary boat cruise from end to end. The north, south, and west shores would be irregular with inlets, peninsulas, and offshore shallow water. The east shore along the Black Mountains would be more linear with minor indentations, save for the Mormon Point cove, and water would be mostly deeper. You could almost reach out and touch the Copper Canyon and Badwater turtlebacks. Furnace Creek Wash would form an interesting inlet. But if you see such a lake in Death Valley today, it is probably just a heat-shimmering mirage on the dry desert floor.

Paleogeographic reconstruction of the Death Valley extended region: Evidence from Miocene large rock-avalanche deposits in the Amargosa Chaos Basin, California

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ABSTRACT

Large rock-avalanche deposits are extremely useful tools in paleogeographic and tectonic reconstructions. These deposits are large (>0.1 km³), semicoherent "snapshots" of a source area that were emplaced instantaneously in a sedimentary basin. A rock-avalanche deposit with a known age in a sedimentary basin requires the restoration of that basin to a position adjacent to the rock-avalanche source area at the time of emplacement. The Miocene sedimentary section of the Amargosa Chaos and Sperry Hills basins in the southern Death Valley region contains deposits of large rock-avalanches from both spatially and temporally separated source terranes.

The rock-avalanche deposits in the Amargosa Chaos basin require the restoration of the basin to within 10 km of the Kingston Range prior to 7.8 Ma. West-derived alluvial fan sediments in this basin can be confidently tied to the southern Panamint Mountains, with early east-derived alluvial fan sediments (pre–7.8 Ma) tied to the Kingston Range. Moreover, the pre–7.8 Ma stratigraphy of the Amargosa Chaos basin is equivalent to the pre–7.8 Ma stratigraphy of the Miocene sedimentary section in the Sperry Hills. Prior to ~7.8 Ma, the Amargosa Chaos and Sperry Hills basins were both parts of a larger sedimentary basin. Sedimentologic constraints on the geometry of the Amargosa Chaos basin indicate that throughout its history it was a half-graben opening along a west-dipping normal fault with substantial vertical relief in the uplifted footwall. Evidence for this topography exists in the form of multiple large rock-avalanche deposits derived from footwall bedrock.

The axis of this basin trends northwest with the active basin-bounding normal fault on the northeast margin. At ~7.8 Ma, the extension direction in the southern Death Valley region changed from southwest to northwest. Younger sediments, including three younger rock-avalanche deposits, record the motion of the Amargosa Chaos basin away from the Sperry Hills basin. Top-to-the-northwest displacement of 30 km along the Amargosa detachment system is required between 7.8 Ma and 4.9 Ma between the Sperry Hills basin and the location of Amargosa Chaos basin in the southern Black Mountains. Continued, post–4.9 Ma, northwest-directed extension is required between the Amargosa Chaos basin and the southern Panamint Mountains. In addition, 15 km of right-lateral slip on the southern portion of the Grand View fault is required from 7.8 to 3 Ma to move the Sperry Hills basin away from its pre–7.8 Ma position adjacent to the Kingston Range.

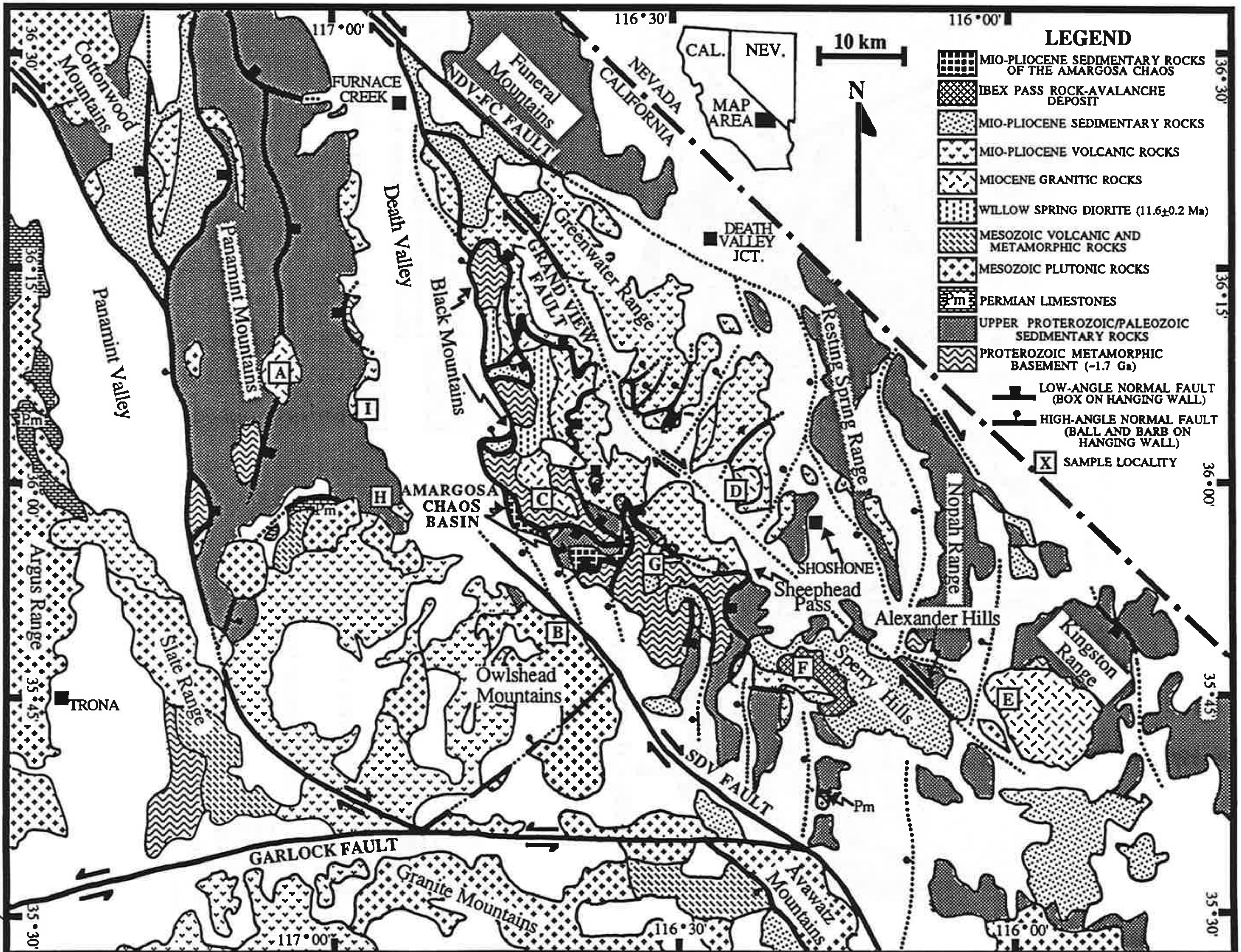
INTRODUCTION

In recent years, the goal of many studies in the Basin and Range province has been to attempt geologic reconstructions across portions of the province to determine the amount of extension across this region. The focus of much of this research has been the Death Valley region (Wright and Troxel, 1967, 1973; Stewart, 1967, 1983; Wernicke and others, 1982, 1988; Cemen and others, 1985; Burchfiel and others, 1987; Hodges and others, 1989; Snow and Wernicke, 1989; McKenna and Hodges, 1990; Wright and others, 1991; Holm and others, 1992). Various methods of correlation of geologic units and structures have been used, but rarely have studies correlated laterally restricted deposits, for example, rock-avalanche megabreccias, with point sources. Correlation of a "point deposit" with a "point source" provides relatively tight constraints on both magnitude and direction of tectonic transport of that deposit away from the source. On a map scale, rock avalanches produce essentially point deposits but are usually composed of sedimentary rocks from nonunique sources. Rock-avalanche deposits composed of rocks that are unique, for example, igneous rocks from a laterally restricted intrusion, would be ideal for paleogeographic reconstructions. In the southern Death Valley region, such rock-avalanche deposits occur in the Amargosa Chaos basin.

The purpose of this paper is to restore the Amargosa Chaos basin, located in southern Death Valley, California (Fig. 1), to its various source terranes through time by analysis of the deposits of large rock avalanches and interbedded alluvial fan sediments. The composition and textures of the rock-avalanche deposits, field relations, petrology, geochemistry, and geochronology were used to determine the provenance of the rock-avalanche deposits. Paleocurrent, paleoslope, and clast-compositional evolution of the alluvial fan conglomerates was studied.

In order to test the hypothesis of large-scale extension in the Death Valley region, I reexamined the Jubilee phase of the Amargosa

Figure 1. Simplified geologic map of the southern Death Valley region showing major structural features and selected stratigraphic and intrusive units. The letters *A–I* refer to sample localities for geochemical and/or geochronologic analyses: (A) Little Chief Stock, (B) Owlshead Mountains granite, (C) Smith Mountain granite, (D) southern Greenwater granite, (E) Kingston Range granite, (F) lowermost granitic rock-avalanche megabreccia (at Ibex Pass) in the Sperry Hills Basin, (G) metamorphic basement at Rhodes Hill, (H) latite flow and overlying lapilli tuff north of Warm Springs Canyon, and (I) uppermost ash-flow tuff north of Johnson Canyon. Modified after Noble and Wright (1954), Johnson (1957), Jennings (1961), Jennings and others (1962), Chesterman (1973), Wright (1974), Jennings and others (1977), Albee and others (1981), Wright and Troxel (1984), Hillhouse (1987), Hodges and others (1989), and Wright and others (1991).



LEGEND
MIO-PLIOCENE SEDIMENTARY ROCKS OF THE AMARGOSA CHAOS
IBEX PASS ROCK-AVALANCHE DEPOSIT
MIO-PLIOCENE SEDIMENTARY ROCKS
MIO-PLIOCENE VOLCANIC ROCKS
MIOCENE GRANITIC ROCKS
WILLOW SPRING DIORITE (11.6±0.2 Ma)
MESOZOIC VOLCANIC AND METAMORPHIC ROCKS
MESOZOIC PLUTONIC ROCKS
Pm PERMIAN LIMESTONES
UPPER PROTEROZOIC/PALEOZOIC SEDIMENTARY ROCKS
PROTEROZOIC METAMORPHIC BASEMENT (~1.7 Ga)
LOW-ANGLE NORMAL FAULT (BOX ON HANGING WALL)
HIGH-ANGLE NORMAL FAULT (BALL AND BARB ON HANGING WALL)
X SAMPLE LOCALITY
10 km
N
CAL.
NEV.
MAP AREA
NEVADA
CALIFORNIA
117 • 00'
116 • 30'
116 • 00'
36 • 30'
36 • 15'
36 • 00'
35 • 45'
35 • 30'
Cottonwood Mountains
FURNACE CREEK
Funeral Mountains
NDV-FC FAULT
Death Valley
Panamint Mountains
Panamint Valley
Greenwater Range
GRAND VIEW FAULT
Black Mountains
DEATH VALLEY JCT.
Resting Spring Range
Nopah Range
AMARGOSA CHAOS BASIN
SHOSHONE
Sheephead Pass
Alexander Hills
Kingston Range
Sperry Hills
Owlshead Mountains
Argus Range
Slate Range
TRONA
SDV FAULT
GARLOCK FAULT
Granite Mountains
Avawatz Mountains
A
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Pm

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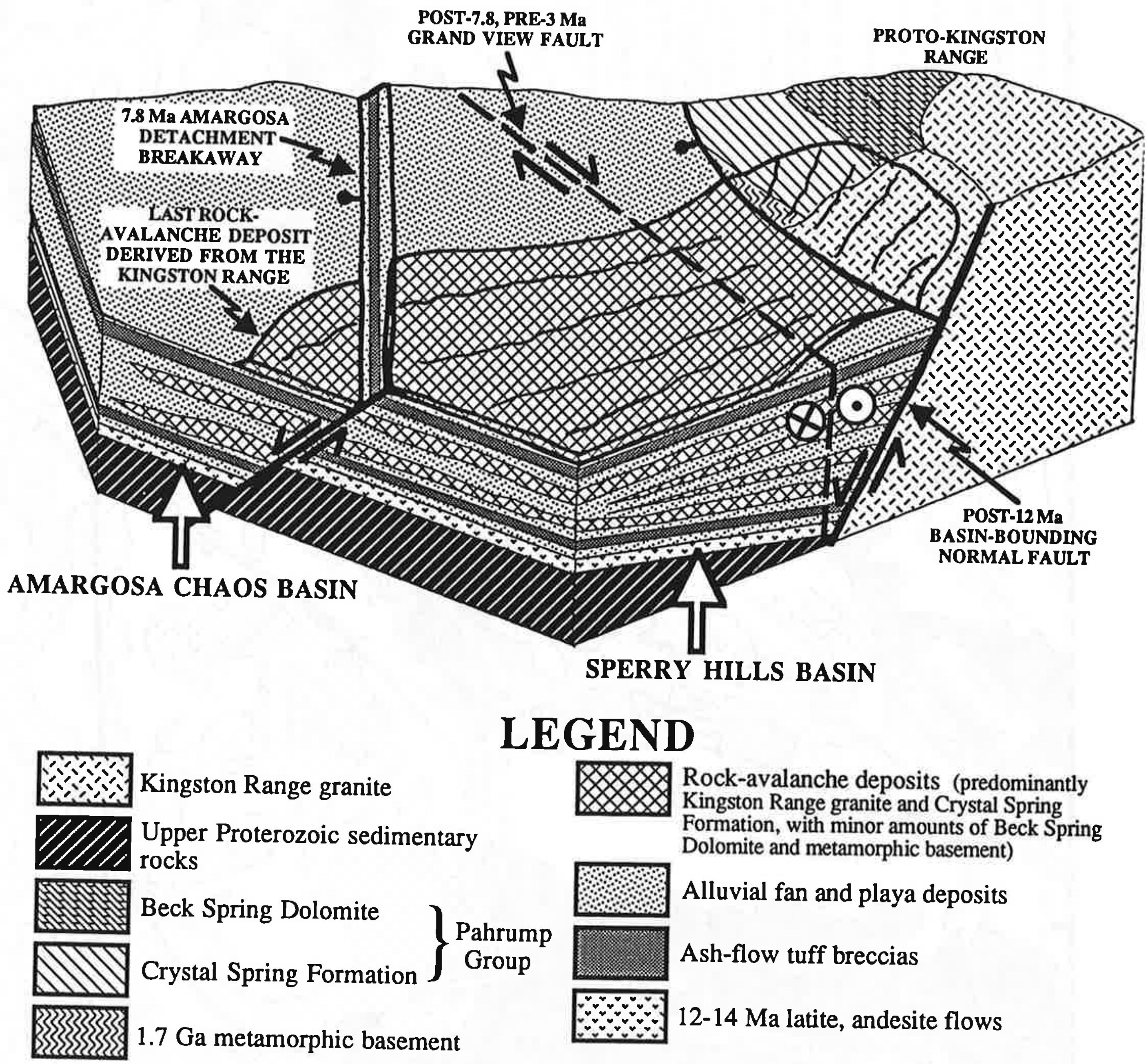

Figure 12. Block diagram of the proto–Kingston Range area at ~7.8 Ma showing the relative positions of the Amargosa Chaos and Sperry Hills basins.

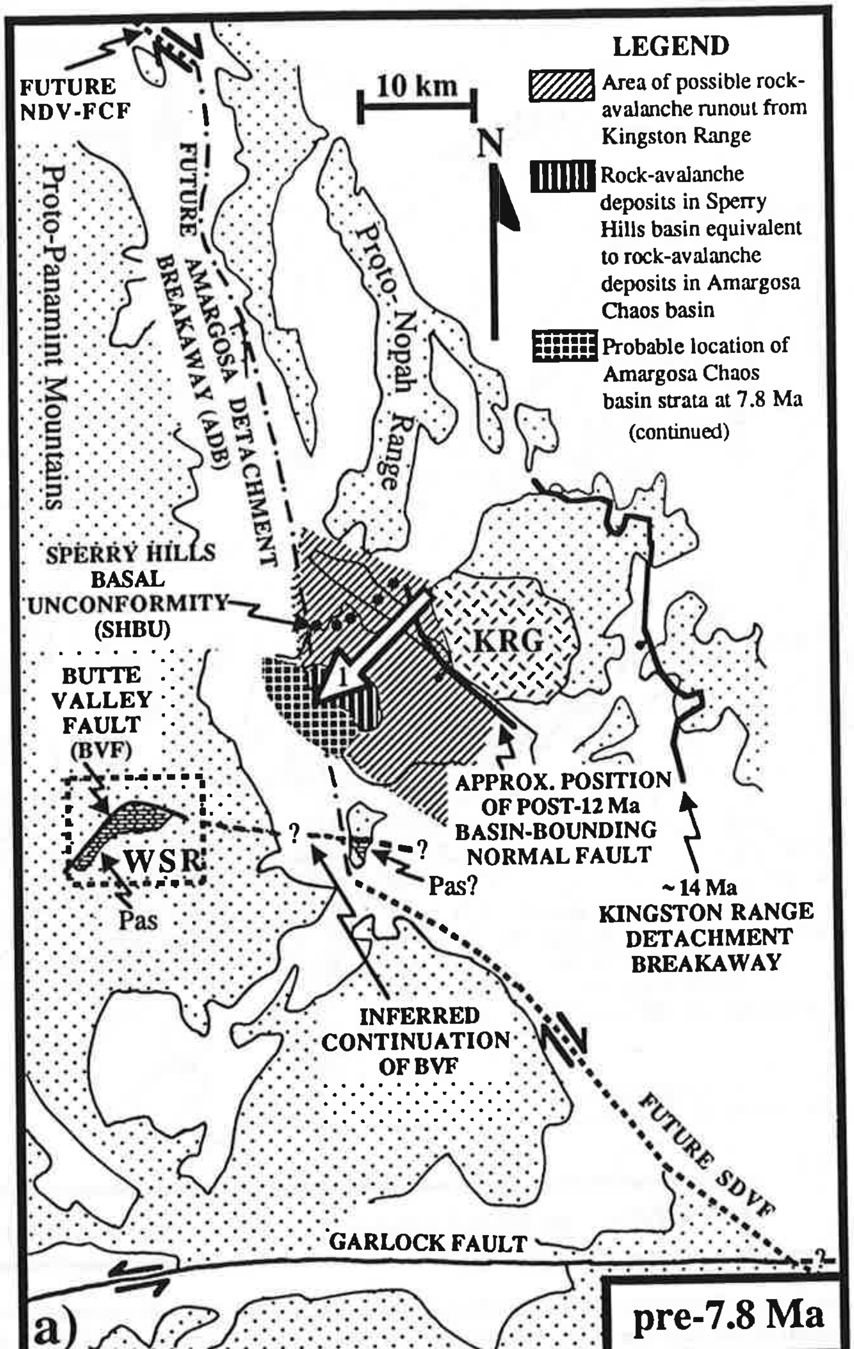

Figure 13. Model reconstruction of the position of the Amargosa Chaos basin through time. The legend for all of Figure 13 is spread across parts a, b, and c. (a) Inferred configuration of the Death Valley region at 7.8 Ma. Vector 1 shows the probable pre–7.8 Ma extension direction between the proto–Kingston Range and southern Panamint Mountains and is inferred from the extension direction for the Kingston Range detachment system (McKenna and Hodges, 1990). The combined Amargosa Chaos–Sperry Hills basin is a northwest-trending half graben with an active basin-bounding fault on the northeastern margin. The Kingston Range–derived rock-avalanche deposits in the combined Amargosa Chaos–Sperry Hills basin are still within 10 km maximum landslide runout. Abbreviations: Pas = Permian Anvil Spring Formation; WSR = western source region; KRG = Kingston Range granite; NDV-FCF = Northern Death Valley–Furnace Creek fault; SDVF = Southern Death Valley fault.

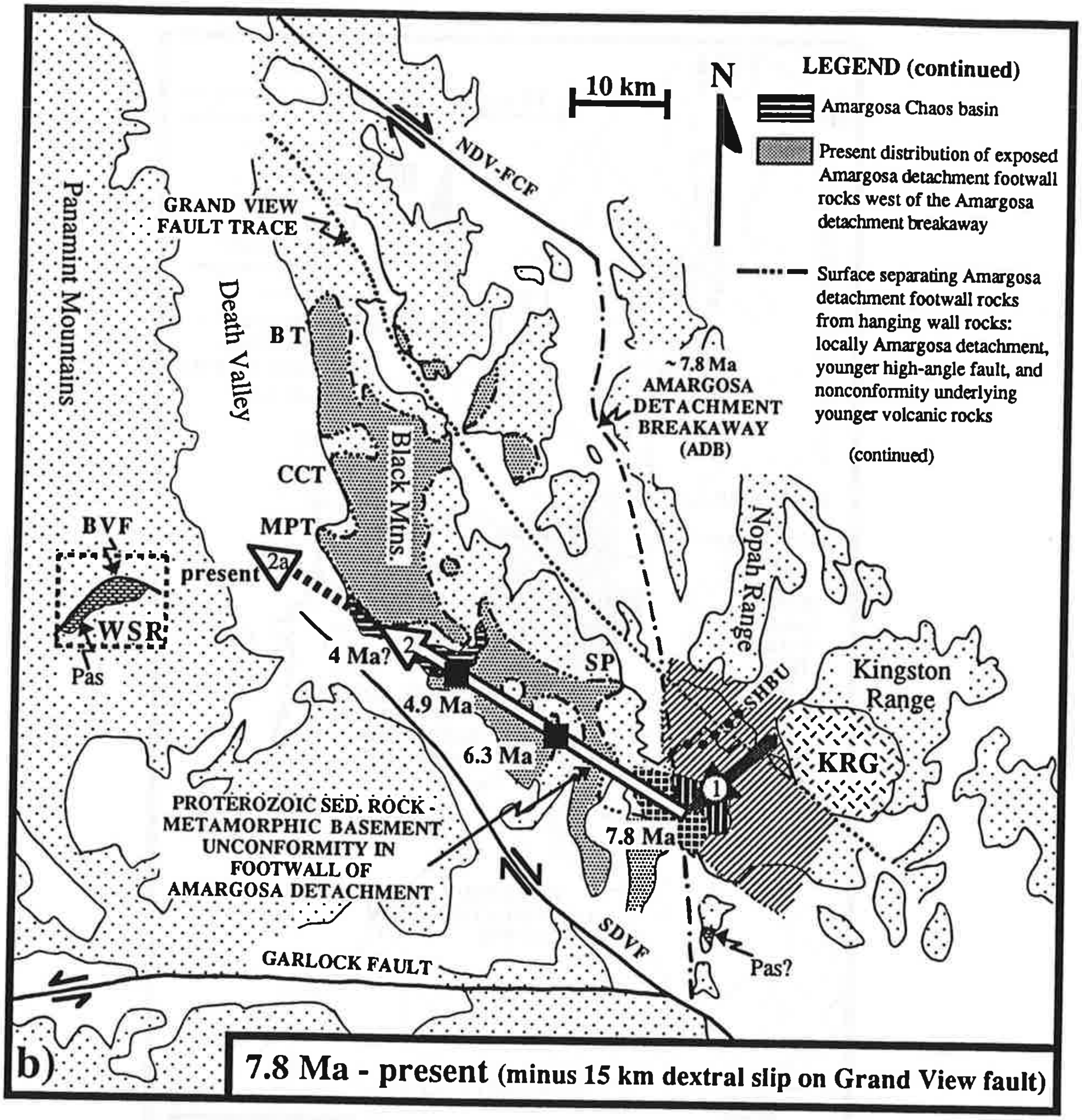

Figure 13. (*Continued*). (b) The present configuration of the Death Valley region *with all strike-slip motion on the Grand View fault restored*. Vector 2 shows the northwest transport of the Amargosa Chaos basin above the Amargosa detachment from 7.8 Ma to ~4 Ma. Extension of 30 km is required between the Jubilee Wash Area section of the Amargosa Chaos basin and the Sperry Hills basin from 7.8 Ma to 4.9 Ma; an additional 5 km of extension is required between the Ashford Mill Area section and the Jubilee Wash Area section of the Amargosa Chaos basin from 4.9 Ma to ~4 Ma. The 6.3 Ma box indicates the position (assuming constant extension rates along vector 2 from 7.8 to 4.9 Ma) of the Amargosa Chaos basin at the time of emplacement of the metamorphic basement rock-avalanche deposit. Vector 2a shows the post-4 Ma transport of the Panamint Mountains away from the Ashford Mill Area section of the Amargosa Chaos basin. Abbreviations: BT = Badwater Turtleback; CCT = Copper Canyon Turtleback; MPT = Mormon Point Turtleback; SP = Sheephead Pass.

Black Mountains Crustal Section, Death Valley Region, California

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INTRODUCTION

The Black Mountains of Death Valley, California contain features which have long held the interest of geologists since they were first described in the 1940's and early 1950's (Noble, 1941; Curry, 1954; Noble and Wright, 1954). With the initial discovery and continued documentation of large-magnitude extensional tectonism in the region (Anderson, 1971; Wright and Troxel, 1973; Davis and Burchfiel, 1973; Wernicke and others, 1988; Snow and Wernicke, 1989), there has been an enthusiastic rejuvenation of interest in this range block. The Black Mountains have in a short time become one of the best studied metamorphic core complexes in the western U.S. The purpose of this trip is to evaluate the idea that the Black Mountains may represent a nearly coherent section of the pre-extensional crust unroofed by extensional tectonism. Participants will critically investigate important structural relations, syntectonic intrusive chronologies, and synextensional depositional systems and integrate these observations with published geochemical, thermochronologic, geobarometric and paleomagnetic data. The results allow us to address important questions regarding the timing, nature, and amount of both footwall and hanging wall deformation in a superbly exposed normal fault system.

GEOLOGIC OVERVIEW

The Black Mountains lie in the center of the Death Valley extended region and along the western margin of the Central Death Valley volcanic field (Wright and others, 1981). The region is dominated by west-dipping normal faults developed within the Cordilleran miogeocline, a westward thickening sedimentary prism of Eocambrian through early Mesozoic age that was shortened by east-vergent folding and thrusting during the late Paleozoic and Mesozoic. Extension at the latitude of Death Valley since ~15 Ma has resulted in the exposure of east-tilted sections of Late Precambrian to Paleozoic sedimentary and Tertiary volcanic rocks which constitute the geology of the ranges surrounding the Black Mountains (Figure 1). Overall, in this region, the cessation of major extension has migrated from east to west (Wright and others, 1984) with the main intervals of range-scale tilting and unroofing having occurred as follows: Nopah-Resting Spring Range, 12-9 Ma; Black Mountains and Funeral Range, 10-6 Ma; overprinting of the Black Mountains and ranges to the west from 4 to 0 Ma (e.g., Wernicke and others, 1988).

In contrast to the thick sequence of miogeoclinal rocks present in surrounding ranges, the Black Mountains core consists mainly of three antiformal folds (historically known as the Death Valley "turtlebacks") of metamorphosed Precambrian

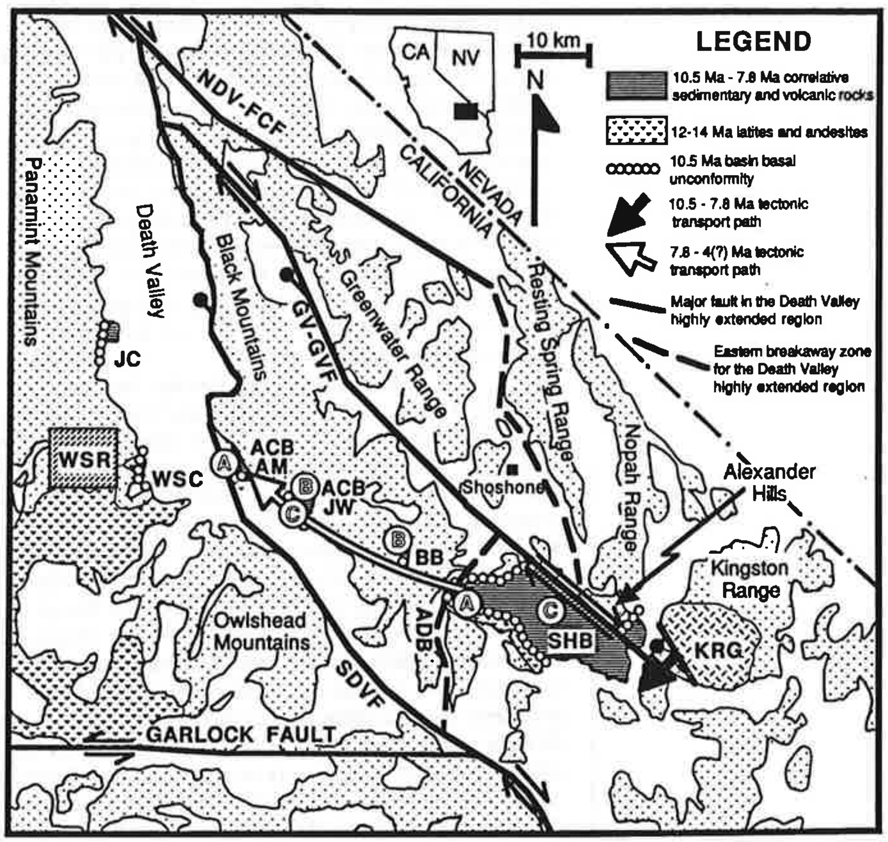


Figure 1. Regional setting of the Black Mountains in the Death Valley region. Large arrows depict a three stage tectonic transport path of the Amargosa Chaos basin. (1) Southwest- west-directed extension between the Kingston Range block and the Ibex Hill-Alexander Hills-Panamint Mountains block from 10.5 to 7.8 Ma. (2) Northwest-directed extension between the Ibex Hills block and the Panamint Mountains block from 7.8 Ma to the present. (3) Fifteen kilometers of dextral strike-slip along the Greenwater Valley-Grand View fault from 7.8 to 3 Ma. Outlined letters A, B, and C refer to the locations of stratigraphic sections in Figures 12 and 13. Abbreviations: NDV-FCF = Northern Death Valley-Furnace Creek fault; GV-GVF = Greenwater Valley-Grand View fault; SDVF = Southern Death Valley fault; JC = Johnson Canyon; WSR = western source region for the Amargosa Chaos basin; WSC = Warm Springs Canyon; ACBAM = Amargosa Chaos basin, Ashford Mill area; ACBJW = Amargosa Chaos basin, Jubilee Wash area; BB = Buckwheat basin; SHB = Sperry Hills basin; ADB = 7.8 Ma brittle Amargosa detachment breakaway zone; KRG = Kingston Range granite pluton. See text by Topping (below) for further discussion.

Eocambrian outcrops. The unmetamorphosed stratified rocks are overlain in mild angular unconformity by 10.5-8 Ma volcanic and sedimentary strata that dip moderately to steeply to the east. Approximately 15-20 km to the northwest, in the middle of the range, these same strata are complexly faulted and extended, forming the Amargosa chaos (Noble, 1941). Northwest of the chaos, Miocene magmatic rocks, intruded into amphibolite facies Eocambrian metasedimentary rocks and middle Proterozoic crystalline schist and gneiss, are variably faulted and ductilely sheared. Most of the crystalline rocks in the Black Mountains, and in particular the northwestern part of the range, are well away from basal unconformities of 10.5-8 Ma strata.

At Ashford Canyon (loc. B, Figure 2), the Precambrian nonconformity lies about 5 km away from the Miocene plutonic rocks and metamorphosed Eocambrian strata (unit mp, loc. A, Figure 2). However, Holm and Wernicke (1990) interpreted this nonconformity as having been translated 15-20 km northwestward away from tilted equivalent strata in the southeastern Black Mountains (loc. C, Figure 2) along a system of low-angle faults (the Amargosa chaos) mapped by Wright and Troxel (1984). The crystalline basement rocks beneath the chaos both to the northwest and southeast appear unbroken by any major detachment system (although late normal imbrication has occurred to the southeast) and these rocks are thus interpreted to be a structurally contiguous, northwest-deepening section of the pre-extensional crust (Holm and Wernicke, 1990).

Exposure of the Crustal Section.

Juxtaposition of the Panamint and Nopah-Resting Springs blocks (Stewart, 1983; Wernicke and others, 1988; Snow and Wernicke, 1989; Topping, 1993a), places the Panamint Range above the intervening Black Mountains prior to Miocene extension. Exposure of the crustal section occurred by westerly-directed tectonic denudation as the Black Mountains footwall pulled out from underneath the relatively rigid, scoop-shaped hanging wall of the Panamint Range. Denudation of the Black Mountains occurred during the mid- to late Miocene (10-6 Ma). On the eastern flank of the crystalline core, volcanic rocks deposited over the time interval 14-4 Ma become progressively less tilted and faulted with decreasing age (Wright and Troxel, 1988). Strata ~8-9 Ma old are locally intensely faulted and steeply rotated, and are overlain in angular unconformity by relatively undisturbed basalts and fanglomerates that are about 4-5 Ma (Wright and others, 1983, 1984). In addition, a southeast to northwest progression of cooling (from temperatures above 300°C to below 100°C) associated with unroofing of the crystalline core occurred at ~8.5-6.5 Ma (Holm and others, 1992; Holm and Dokka, 1993). The current predominantly low, eastward dip of the detachment fault on the eastern flank of the range (Figure 2) reflects eastward tilting of the range as the crustal section was progressively denuded to the west.

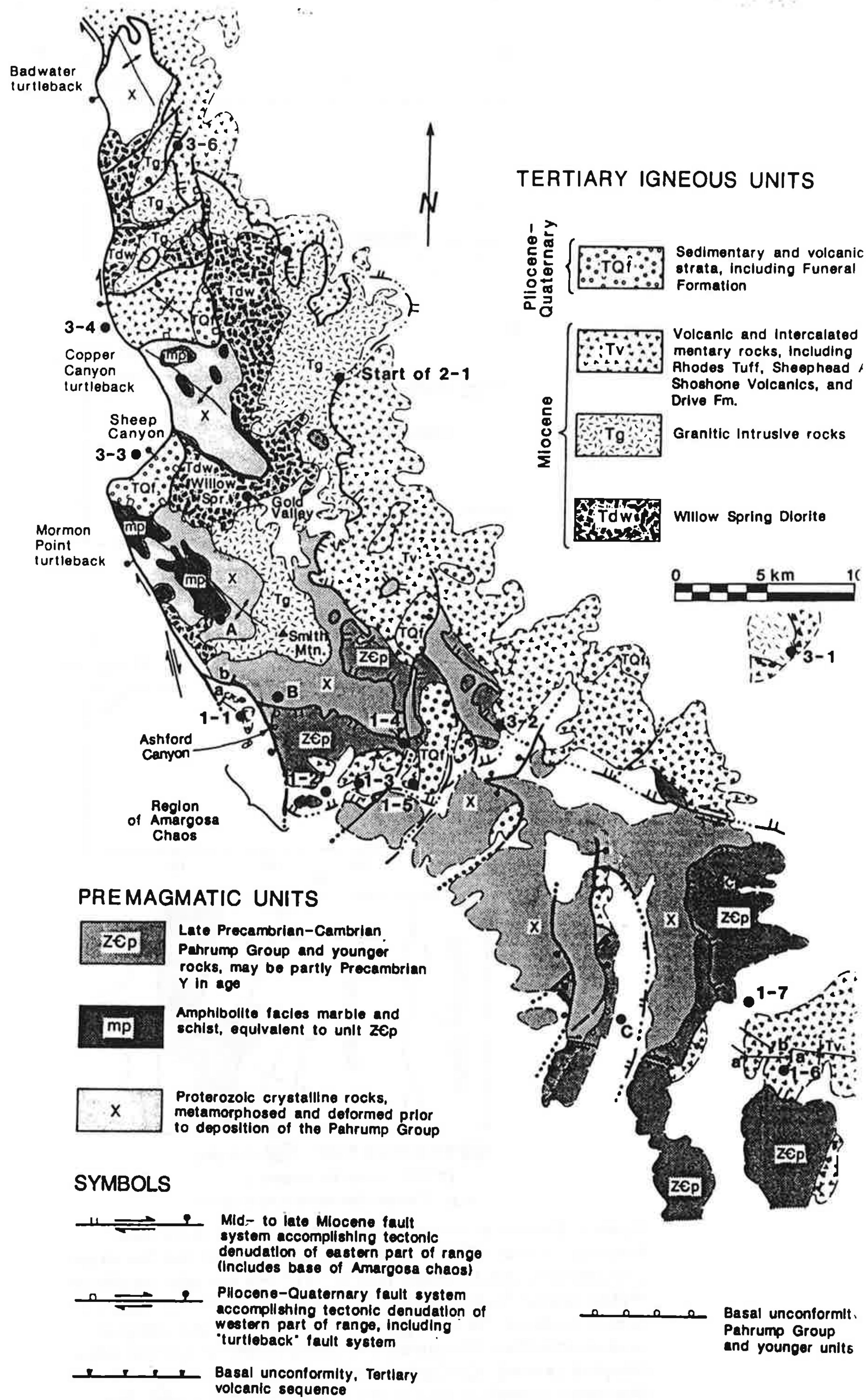


Figure 2. Simplified geologic map of the Black Mountains (modified after Holm and Wernicke, 1990) w localities depicted for this fieldtrip.

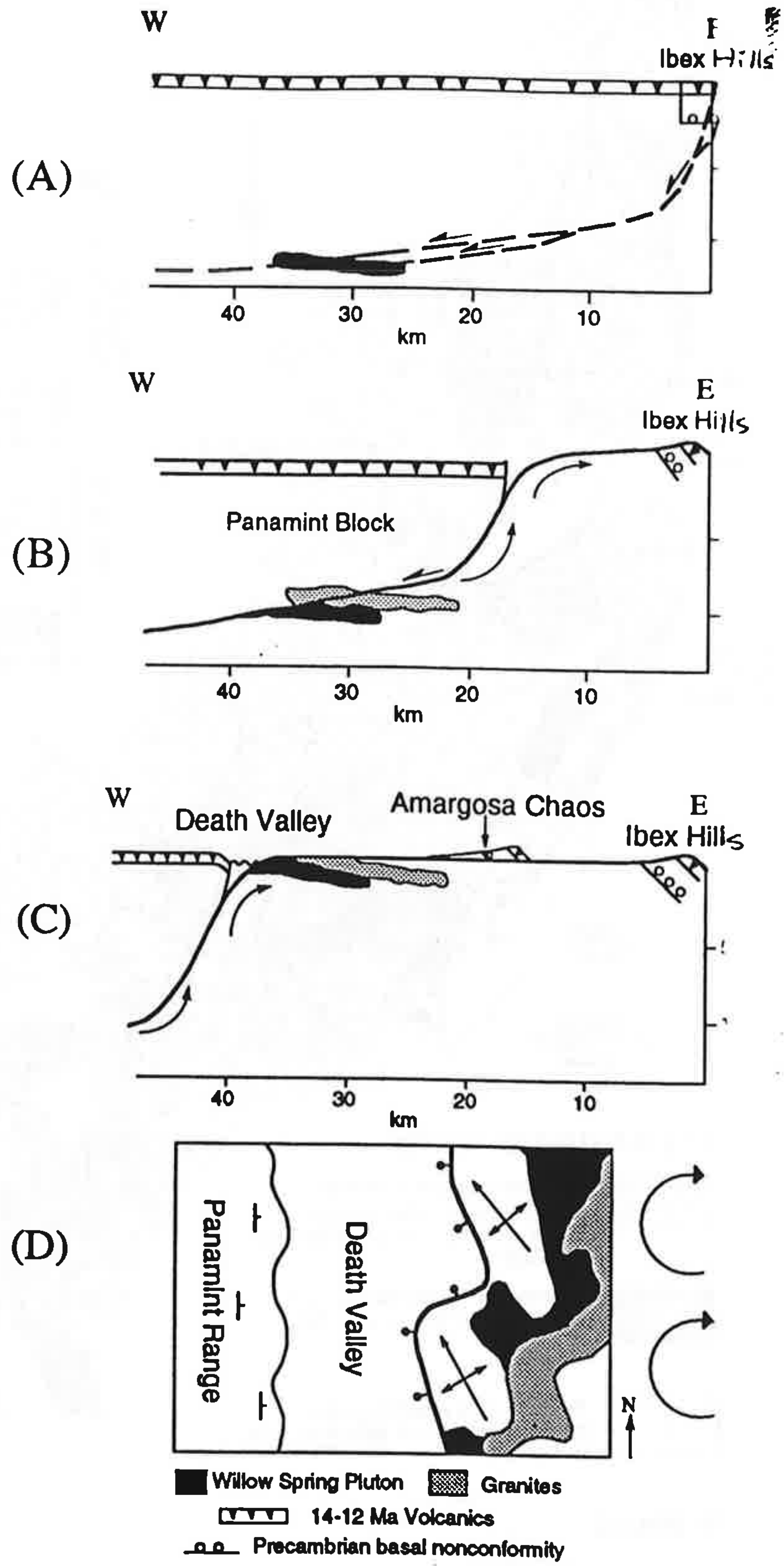


Figure 4. Schematic interpretation of deformation in the Black Mountains (modified after Holm and others, 1993). (A) Pre- extens configuration of detachment fault at ~11.5 Ma just after emplacem Willow Spring Pluton (shaded) at midcrustal levels into future Bla Mountains block. (B) Configuration after onset of unroofing in southeastern Black Mountains and emplacement of granitic intrusi (stippled pattern). (C) Configuration just after unroofing of the Bla Mountains crystalline core by the migrating hinge model. (D) Large-scale clockwise rotation of the crystalline core after ca. 6.5 M resulting in present configuration of the range block (map view).

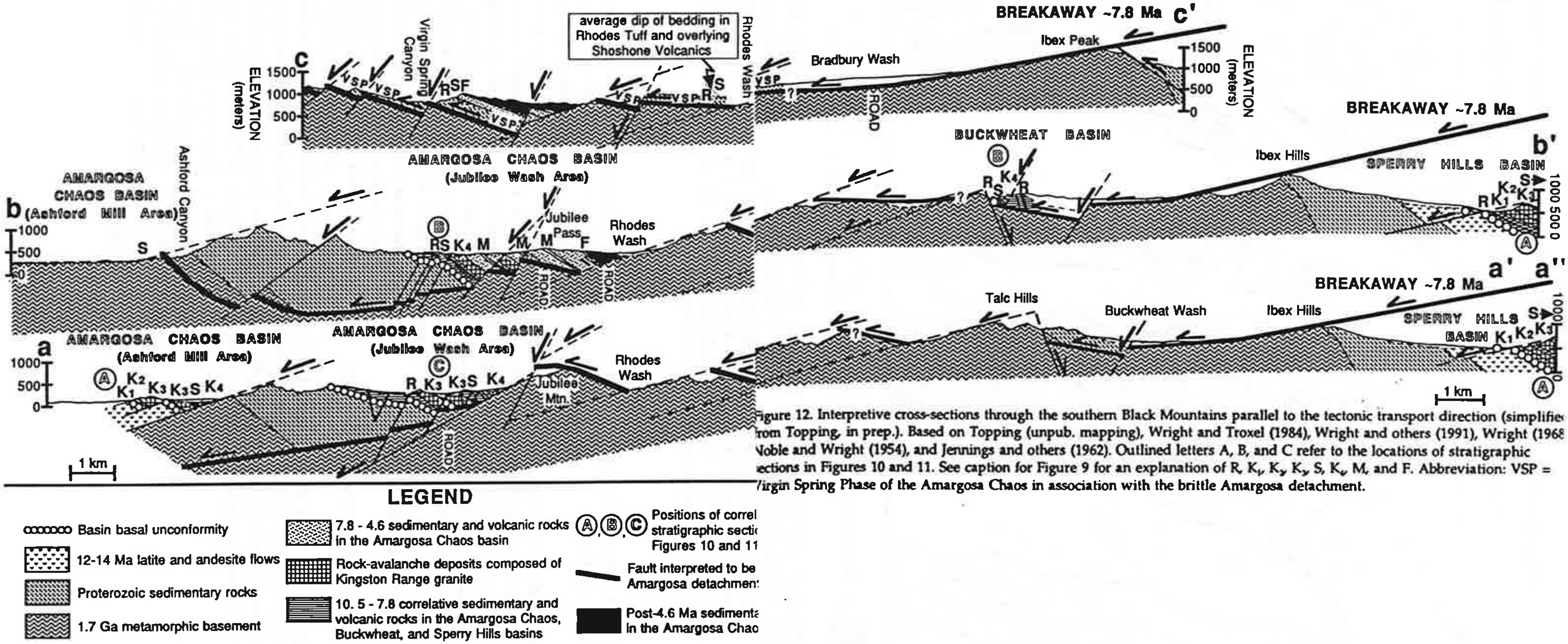


Figure 12. Interpretive cross-sections through the southern Black Mountains parallel to the tectonic transport direction (simplified from Topping, in prep.). Based on Topping (unpub. mapping), Wright and Troxel (1984), Wright and others (1991), Wright (1968), Noble and Wright (1954), and Jennings and others (1962). Outlined letters A, B, and C refer to the locations of stratigraphic sections in Figures 10 and 11. See caption for Figure 9 for an explanation of R, K_1, K_2, K_3, S, K_4, M, and F. Abbreviation: VSP = Virgin Spring Phase of the Amargosa Chaos in association with the brittle Amargosa detachment.

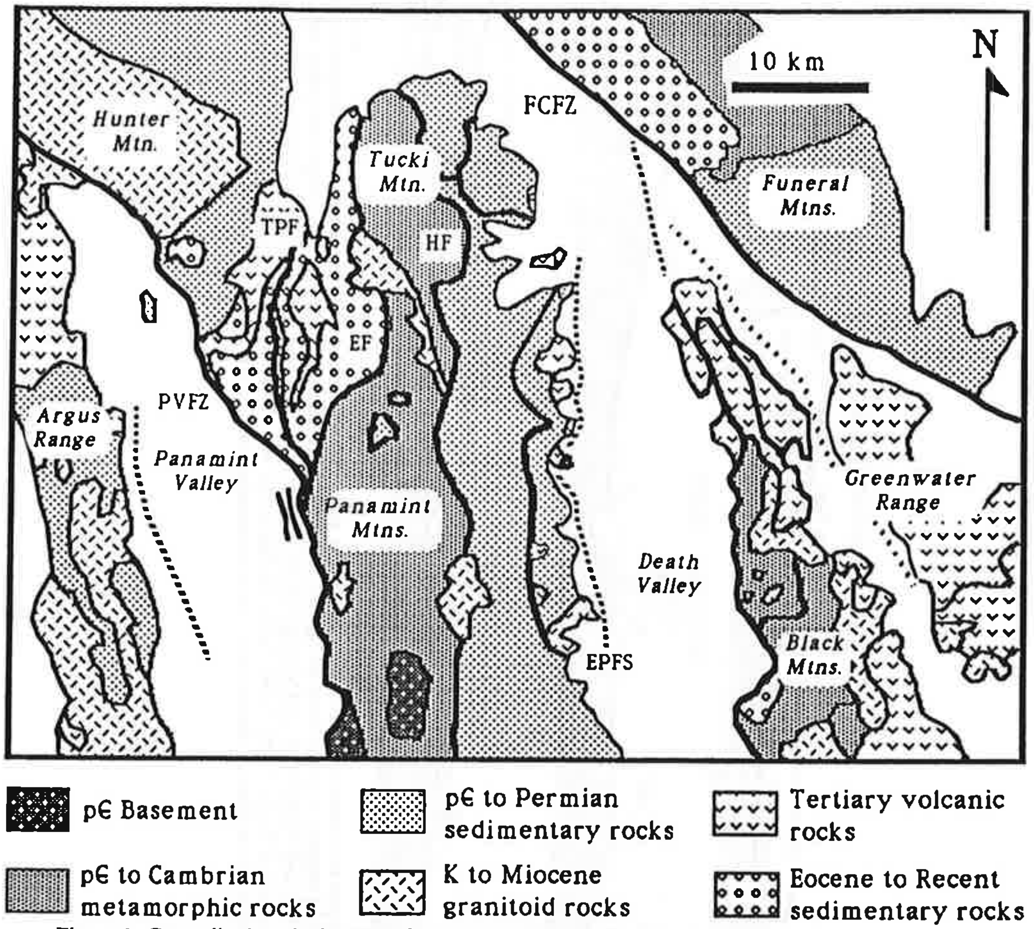


Figure 1. Generalized geologic map of the Death Valley area. EF—Emigrant fault; EPFS—Eastern Panamint fault system; FCFZ—Furnace Creek fault zone; HF—Harrisburg fault; PVFZ—Panamint Valley fault zone; TPF—Towne Pass fault. On this and other figures, solid lines represent well-located features, and dashed lines represent buried structures.

KINEMATIC MODEL FOR UNROOFING OF THE PANAMINT CORE COMPLEX

Structural relations suggest that greenschist to amphibolite facies metamorphic rocks in the core of the Panamint Mountains were brought to the surface as a consequence of movement along several temporally distinct normal fault systems. Our interpretation of the unroofing sequence is illustrated in Figure 9 as a series of cartoon cross sections. For ease of discussion, we arbitrarily subdivide the sequence into four stages.

Stage I: Development of the Harrisburg subsystem

Prior to intrusion of the middle Miocene Little Chief stock, the Harrisburg detachment and related structures juxtaposed greenschist to amphibolite facies rocks of the Lower Allochthon and greenschist facies rocks of the Middle Allochthon. Following models proposed by several others (e.g., Wernicke, 1981; Davis and others, 1986), we believe that movement on this detachment was responsible for substantial upward transport of the footwall toward the surface. We have no real constraints on the magnitude of this transport because we cannot match hanging-wall and footwall cutoffs. Apparent stratigraphic throw across the fault commonly exceeds 1 km, but this must be considered an absolute minimum estimate of the true throw because it is insufficient to explain the observed metamorphic discontinuity across many segments of the detachment and the fact that synmetamorphic mesoscopic structures in the footwall (Hodges and others, 1987) have no clear correlatives in the hanging wall. Although pressures equivalent to 8 to 11 km depth were attained in the footwall during amphibolite-facies metamorphism (Labotka, 1981), no assemblages appropriate for quantitative thermobarometry occur in the hanging wall, and it is thus impossible to quantify the difference in Mesozoic paleopressures across the Harrisburg detachment. Even if it were possible to do so, the result probably would not accurately reflect throw on the detachment because of the likely differential uplift of hanging wall and footwall subsequent to Mesozoic metamorphism but prior to Harrisburg faulting.

Stage II: Development of the Eastern Panamint fault system

The Eastern Panamint fault system developed subsequent to intrusion of the Little Chief stock (10.6 Ma) and during late Miocene eruption of the Trail Canyon volcanic sequence. Near Hanaupah Canyon (Fig. 3), the sole fault of the system juxtaposes upper Precambrian Stirling Quartzite in the hanging wall and Proterozoic crystalline basement in the footwall. Stratigraphic throw across the sole fault here clearly exceeds 1.5 km (the approximate thickness of omitted Johnnie and Noonday strata). The cumulative throw on the entire Eastern Panamint system must have been close to the thickness of the entire upper Precambrian through lowest Silurian-Devonian section (McKenna and Hodges, this volume). Because the major unconformity developed beneath the Noonday Dolomite locally excised substantial Pahrump Group stratigraphy in the Death Valley region (Cloud and others, 1974; Wright and others, 1978), we cannot assume an omitted thickness of the Pahrump Group without qualification, but the total throw on the Eastern Panamint system is likely to have been in excess of 3 km and perhaps as much as 7 km. We infer that the most important consequence of movement on the Eastern Panamint system was the upward transport of the Parautochthon relative to the Lower and Middle Allochthons.

Stage III: Range-scale folding

The anticlinal geometry of the Panamint Mountains has been interpreted in a variety of ways. Hunt and Mabey (1966) suggested that the structure was in part related to doming over intrusions of plutons such as the Skidoo monzogranite but noted that some of the rotation of the range must have postdated eruption of the late Tertiary volcanic rocks on the eastern side of the range. Labotka and others (1980, 1985) also ascribed the folding to a late Cretaceous deformational event associated with plutonism. Wernicke and others (1986) and Hodges and others (1987) noted that late Tertiary structures of the Tucki Mountain detachment system are clearly involved in the anticlinal structure, and they attributed the folding to reverse-drag flexure above the Eastern Panamint fault system. We still believe that at least some reverse-drag might have occurred during Eastern Panamint movement; however, recent mapping of the Eastern Panamint system (McKenna and Hodges, this volume) reveals that these structures also have been rotated roughly 40° eastward during development of the range-scale anticline, and we must appeal to a different mechanism for the folding.

Recently, several workers have ascribed the domal form of metamorphic core complexes to isostatic rebound of a footwall denuded by movement along a detachment (e.g., Spencer, 1984; Buck and others, 1988; Davis and Lister, 1988). Although we concur that this mechanism probably was important in producing some of the broader domal morphologies of detachments in areas such as the Whipple Mountains, California, we are doubtful that it was the primary cause of the much tighter anticlinal structure of the Panamint Mountains. Block and Royden (1990) have shown that the wavelength/amplitude ratio of isostatically induced flexures is a sensitive indicator of the effective elastic thickness of the lithosphere. For the Panamint Mountains, this ratio has an approximate value of 10.5, implying an effective elastic thickness of less than one-fourth of that permissible in other highly extended terrains like the Whipple Mountains. Even in a region of relatively high heat flow, such an elastic thickness seems unreasonably low.

A more plausible mechanism involves reverse-drag flexure above a detachment system lying below the Panamint Mountains that developed after movement on the Eastern Panamint system and before initiation on the undomed Emigrant subsystem. A likely candidate is the west-dipping, low-angle fault system that involves rocks ranging in age from Precambrian to Tertiary in the west-central Black Mountains (Fig. 1). Noble (1941) first described the complex geometric relations in this system, naming it the "Amargosa chaos." He believed that the Amargosa system was regionally extensive and included structures that we group as the Eastern Panamint system. Although Wright and Troxel (1973, 1984) abandoned the idea of a regionally extensive Amargosa system, Stewart (1983) postulated that the low-angle structures at the eastern foot of the Panamint Mountains and the western foot of the Black Mountains are part of a single detachment system that accommodated 80 km of northwest transport of the Panamint block relative to the Black Mountains.

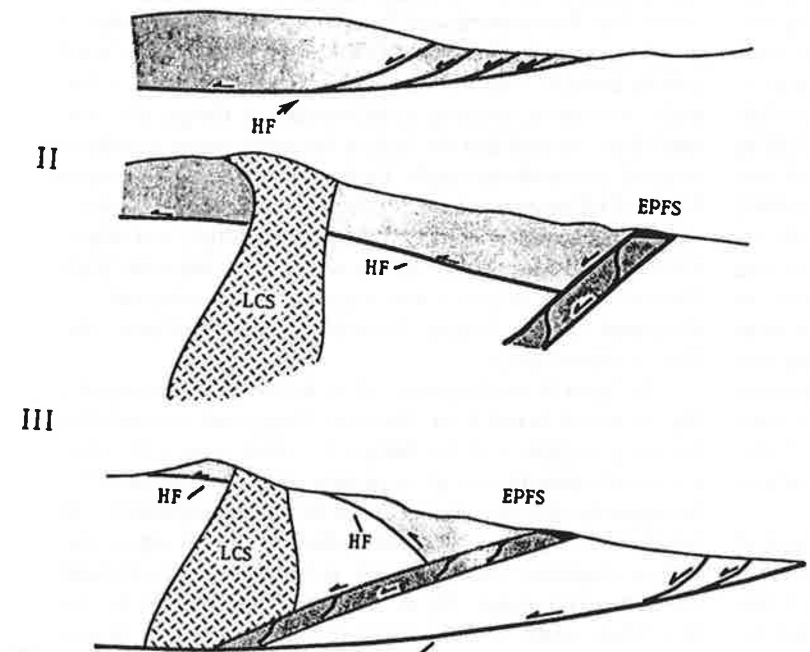
Despite similarities between the Eastern Panamint structures and the type exposures of the Amargosa chaos in the Black Mountains, two lines of evidence suggest that the systems are not correlative. Reconstruction of Tertiary volcanic flow boundaries based on mapping by Noble (1941) and Wright and Troxel (1984) indicates that the Amargosa system initiated at a low angle; in contrast, mapping by McKenna and Hodges (this volume) demosntrated that the Eastern Panamint system must have initiated at a moderate angle. Furthermore, the upper Miocene Rhodes Tuff appears to have largely predated development of the Amargosa chaos (Wright and Troxel, 1984; Wright and others, 1984), whereas the apparently correlative tuffs within the Trail Canyon volcanic sequence were erupted synchronously with development of the Eastern Panamint system (McKenna and Hodges, this volume).

In Figure 9, we illustrate our bias that the Amargosa system dips westward beneath the Panamint Range and truncates the down-dip projection of the Eastern Panamint system. We infer that reverse-drag flexure as a consequence of movement on the Amargosa system was responsible for the modern eastward dip of the Harrisburg detachment and shallow westward dip of the Eastern Panamint system, as well as for the overall anticlinal structure of the range. Up to 1.2 km of uplift of the Lower Allochthon could have occurred during development of the flexure.

Stage IV: Development of the Emigrant subsystem and subsequent structures

Correspondence between the position of the Emigrant breakaway and the hinge zone of the range-scale anticline suggests that the high topography induced during Stage III flexure may have influenced initiation of the Emigrant subsystem. We believe that the Emigrant structures developed in order to minimize the thickness of the Amargosa hanging wall. Although the total displacement on the Emigrant subsystem could have been greater than 20 km (Hodges and others, 1989), palinspastic reconstructions of the breakaway zone in the Panamint Mountains (e.g., Fig. 8) indicate that Emigrant structures were responsible for no more than about 1.5 km of structural unroofing of the metamorphic core of the range. Substantial (but unquantified) erosion of the denuded core is recorded in the conglomerates of the Nova Formation.

After movement ceased on the Emigrant system in early Pliocene time, structures to the west, such as the Towne Pass fault and the Panamint Valley fault zone (Fig. 1), permitted further development of the Nova–Panamint Valley extensional basin (Burchfiel and others, 1987; Hodges and others, 1989). East of the range, the Death Valley basin continued to open as a consequence of pull-apart motion on the Southern Death Valley and Furnace Creek strike-slip fault zones (Burchfiel and Stewart, 1966) and movement on west-dipping faults that are now exposed as "turtleback" surfaces along the western front of the Black Mountains (Wright and others, 1974). Collectively, these structures have caused further eastward tilting and uplift of the Panamint Mountains relative to their surroundings.



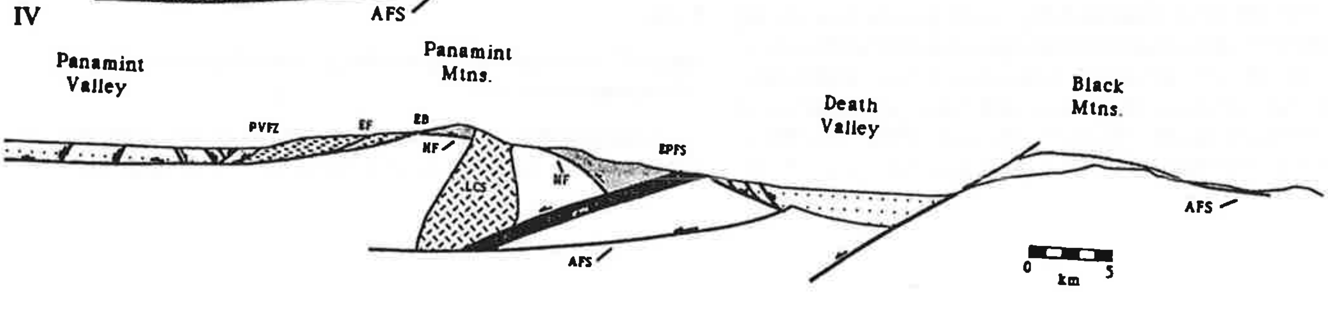


Figure 9. Interpretive cartoons illustrating the Tertiary-Quaternary evolution of the central Panamint Mountains. Frames progress in time from Stage I to Stage IV. Widely spaced dot pattern—Death Valley and Panamint Valley basins; black fill—Miocene volcanic rocks; random dash pattern—Little Chief stock; dark shade pattern—Eastern Panamint fault zone; horizontal dash pattern—Upper Allochthon; light shade pattern—Middle Allochthon; unpatterned—footwall of Harrisburg fault. AFS—Amargosa fault system; EB—breakaway of the Emigrant fault system; EF—Emigrant fault; EPFS—Eastern Panamint fault system; HF—Harrisburg detachment; LCS—Little Chief stock; PVFZ—Panamint Valley fault zone.

Segment B—U. S. 395 to Barstow, 46 miles, Figure 3-3

● From the U. S. 395 separation, continue north-northeasterly on Interstate 15 toward Victorville. Mojave River, draining from high, relatively well-watered parts of the San Bernardino Mountains, flows north in a wide sandy bed about 10 miles to the east. We are slowly converging to a rendezvous with it at Victorville.

● Beyond the Hesperia-Phelan exit, the west end of a broad trough extending from Victorville east-southeasterly to the Twenty-Nine Palms country makes the low skyline at 2 o'clock (*see* Mojave Desert province). Ahead are the bedrock hills of Victorville, composed of granitic and metamorphic rocks. Some of the white spots are carbonate-rock quarries (limestone and marble) which provide material for the cement plants of Victorville and Oro Grande. At 10 o'clock are low peaks and ridges of the Shadow Mountains. They contain extensive exposures of metamorphic and igneous rocks like those near Victorville.

● Beyond the Lucerne Valley exit, plumes of dust and smoke from cement plants at Victorville and Oro Grande are often visible. The smoke dead ahead beyond Victorville is from the city dump. At 11 o'clock are granitic rocks of Silver Mountain, and the conical hills at 1 o'clock are erosion residuals carved mostly in metamorphic rocks.

● As the freeway starts to curve around Victorville you will become aware of greater gullying and dissection. This reflects the influence of Mojave River which flows in a course cut about 175 feet below the alluvial surface which we have been traversing.

● Shortly we drop down and cross the ▶ river. (Note odometer reading at crossing.) A good stream of water flows here the year around, although the stream bed a few miles upstream is bone dry except during floods. What happens is this: About 1.5 miles upstream at the Apple Valley bridge (Upper Narrows, Figure 3-3) the Mojave River, as it cut down its bed, uncovered a westward projecting ridge of granitic rock buried in the alluvium. The river had no choice but to cut a narrow, steep walled gorge through the granite, creating the Upper Narrows (Photo 3-1). Bedrock close

Photo 3-1. Upper Narrows of Mojave River at Victorville formed as river cut down through a buried granitic rock ridge, now exhumed, as viewed southeastward. San Bernardino Mountains in background. (Photo by John S. Shelton, 3402).

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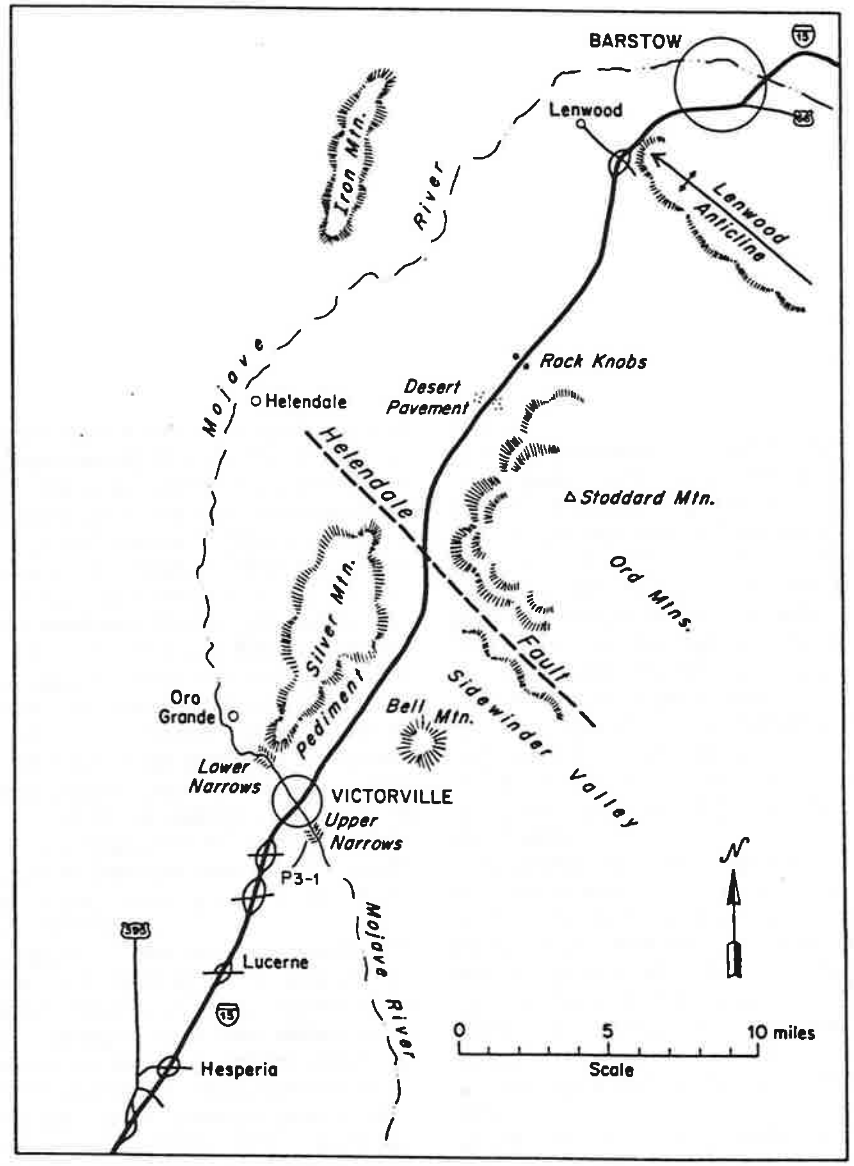

Figure 3-3. Segment B, U. S. Highway 395 separation to Barstow.

under the channel floor at this point brings subsurface water seeping through the porous river sands back to the surface. The water disappears again a mile or two below the Lower Narrow (Figure 3-3), which were formed in the same fashion.

● Across the river we start climbing out of the valley, and good exposures of jointed granitic rock are seen at 1-5 o'clock. They are discolored gray by a mantle of cement dust that has been converted to a hard surface coating. To the left at 9-10 o'clock we see again the granitic rock in Silver Mountain. The gently sloping bedrock surface extending outward (south and southeast) from the mountain base has been formed by erosion and is called a *pediment*. Within 2 miles beyond the river, quarries, roads, and other workings in metamorphosed carbonate rocks (light colored) are seen among hills to the east. At about 4 miles we begin to traverse the broad alluvial surface of Sidewinder Valley.

● Beyond the Speedometer Check sign all but the driver can look back at about 4:30 o'clock for a view of the graceful symmetry of Bell Mountain on the near skyline. The smooth concavity of its slopes is due to the fact that its summit is capped by metavolcanic rocks that yield particles of all sizes upon breakup by weathering. These particles are distributed down the flanks of Bell Mountain in such a manner (coarser near the top and finer near the bottom) as to produce a slope grading smoothly from steep to gentle, hence concave. In another mile or two the rugged Stoddard Mountain mass (4894 ft.) in the Ord Mountains begins to dominate the scene at about 12 o'clock. The rocks therein display considerable variation in color and appearance being largely metavolcanics of possible Triassic age (200 m.y.).

● Shortly the highway curves west; we pass under some large powerlines (note odometer,) ◀ and in about 2 miles we start a gentle descent into a shallow valley. We are approaching a crossing of the Helendale fault, one of a dozen major northwest trending faults slicing this part of the Mojave Desert like a loaf of bread (*see* Mojave Desert province). We are here traveling over metavolcanic rocks and debris derived therefrom, and about the only visible indication of a fault is the abrupt change to light colored knobby granitic rock seen a little before crossing the bridge over the shallow wash about 3.5 miles from the powerlines. A few miles to the southeast, and also in Lucerne Valley, Helendale fault is marked by scarplets breaking alluvial fans, an indication of fairly recent movements. (Note ◀ odometer reading at bridge.)

● Within the next few miles we get good views of the western Mojave, a region of low relief, broad domes, and small residual rocky peaks and ridges. The Mojave River is in a course about 8 miles west, and we intersect it again at Barstow. The black rocks seen at about 10-10:30 o'clock are part of Iron Mountain. They are dark igneous intrusive rocks relatively rich in iron and magnesium.

● For several miles beyond the bridge, note the blocky, angular character of rock outcroppings and of rock debris on the hill slopes east of the freeway. This is characteristic of the metavolcanic rocks of the Sidewinder Series (Triassic?) largely composing these hills. Locally there are small bodies of intrusive granitic rocks which you may be able to spot by the more rounded nature of their outcrop exposures.

● Between 5 and 6 miles beyond the bridge we come into areas where alluvial surfaces on both sides of the road have

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patches of desert pavement (smooth, stony, vegetation-free areas). Stones within the pavement are blackened by desert varnish. Desert pavement is a concentration of stones left on the surface of alluvial deposits as finer materials have been carried away. Slight dissection by gullies draining to the intrenched course of Mojave River is slowly destroying this pavement leaving only remnant patches.

● A little short of 8 miles beyond the bridge we pass what looks like piles of huge rounded boulders on both sides of the highway. These are actually the outcrops of jointed granitic bedrock rounded by weathering and granular disintegration, a typical behavior for uniform, coarse-grained igneous rocks in a desert environment.

● Within another mile look dead ahead about 5 miles to a low, gray, skyline ridge with many small gullies. The ridge is a geologically young anticline (an upfold) within relatively unconsolidated alluvial fan materials (fanglomerate). This is the Lenwood anticline, and we will cross its western plunging nose as the freeway curves around into Barstow beyond the Lenwood exit. To complicate matters another northwest trending fault, similar to the Helendale, passes along the southwest margin of the Lenwood anticline.

● Soon the highway bends westward, and buildings at the Lenwood exit are seen ahead. The smooth subdued skyline at 12:30 o'clock is the area of Rainbow Basin and the Barstow syncline (a downfold) formed in the famed Barstow Formation, a Miocene (10-15 m.y.), terrestrial basin deposit rich in fossil remains of extinct vertebrate animals (*see* Mojave Desert province, special features.)

● Beyond Lenwood exit, the freeway crosses the nose of the Lenwood anticline, and road cuts and hillsides expose some of the gently dipping beds of unconsolidated materials composing this structure.

● Approaching and beyond the first exit to Barstow, at 10-10:30 o'clock, is a prominent reddish-brown rock knob with microwave relay towers and a white letter B. This is a small, cylindrical, intrusive plug of Tertiary igneous rock (rhyolite) formed at a time when the surface volcanics of the Barstow region were being extruded. It has subsequently been etched out by erosion.

Segment C—Barstow to Baker, 61 miles, Figure 3-4

● Just beyond Main Street exit at Barstow, the broad, sandy, and normally dry bed of Mojave River is crossed. During exceptionally wet winters, a good stream of water flows under this bridge. (Note your odometer reading on the bridge.)

● The low hills across the river, traversed for the next 4-5 miles, consist of a mixed and strongly deformed assemblage of Tertiary volcanic and, locally well-bedded, terrestrial sedimentary rocks. At the Bakersfield exit, another large northwest trending fault crosses our route but without recognizable expression at the freeway.

● Descending to the Yermo plain beyond Meridian Road, the Calico Mountains are in view at 9-11 o'clock. The name comes from the variegated appearance produced by complex structural relationships (Photo 1-1) between highly colored sedimentary and volcanic rocks of Tertiary age. The cream and bright-green colors are sedimentary and volcano-sedimentary beds, and the dark reddish-browns are principally volcanics.

● Approaching Ghost Town exit, the large black knob just off the highway at 3 o'clock is Elephant Mountain. This is another near-surface intrusive plug, in this instance of a darker rock than the reddish plug seen in Barstow. (Note odometer at the Ghost Town Road underpass crossing.)

● Beyond Ghost Town Road, look to the mountains at 9:15 o'clock to see the ghost town of Calico, now operated as a tourist attraction by San Bernardino County. Calico was a silver camp from 1882-1896, with a claimed production of 86 million dollars, and then a borax producer until about 1907. The silver deposits were rich, but they pinched out at shallow depths.

The south front of Calico Mountains is bounded by another member of the family of northwest trending faults.

● At the Yermo Exit 1 mile sign, note the highly colored beds in the canyon cut into the mountains at about 9 o'clock.

● About 3.5 miles beyond Ghost Town Road, good bedding is seen in tilted sedimentary and fragmental volcanic deposits of the Calico Mountains face.

● Skip ahead to ① and read about the archeological diggings. Anyone wishing to visit the site should turn off on Mineola Road, about a mile beyond the agriculture inspection station, turn north across the freeway, drive 0.5 mile east on the paved road, go north on the dirt road leading to the County Refuse Disposal Site for 1.1 miles, and then turn east on the one-track labeled road. The public is welcome, and you should find the experience both fascinating and educational.

● Beyond the agricultural inspection station one looks at 2 o'clock down the large trough extending from Barstow to beyond Amboy. This is the route followed by U. S. Highway 66 and the Santa Fe Railway to Needles, and perhaps in earlier times by the Mojave River, which now takes a course closely parallel to the one we travel.

① Just before passing under the large powerlines ahead, look 2 miles north at 9 o'clock. The light area on the hillface about ¼ mile east of the county disposal area (smoke) with trailers and small buildings is near the site of the "Calico Digs," an area excavated by the San Bernardino County Museum of Natural History. At this spot some very primitive "artifacts" have been recovered from deep pits dug in well-cemented fan gravels. If accepted as genuine, they establish occupation of North America by ancient man much earlier than classically pictured, at least 50,000 and perhaps many more years ago.

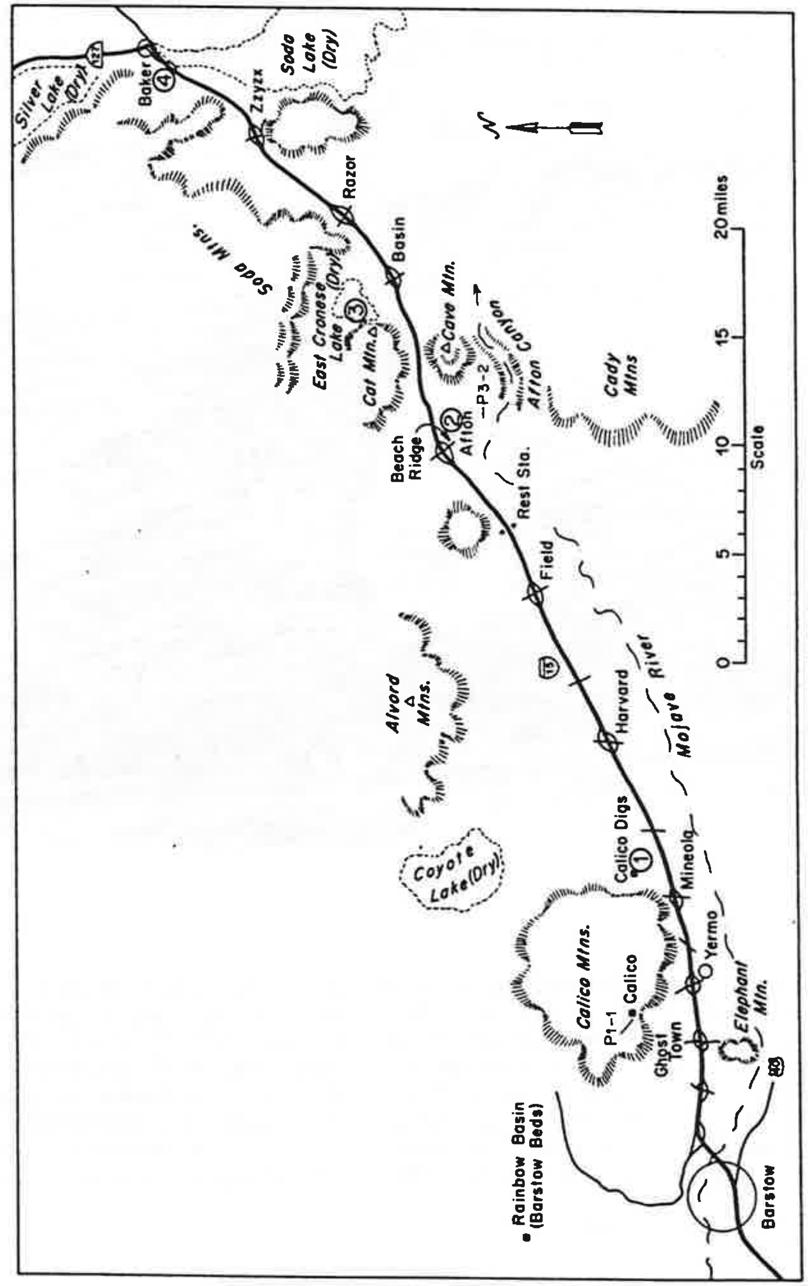

Figure 3-4. Segment C, Barstow to Baker.

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● A mile beyond the powerlines, the low hills close to the freeway on the north side expose gently dipping, light-colored, mid-Miocene Barstow beds capped by younger dark alluvial gravels. About 2 miles farther, beyond the next overpass, a view is had of the west end of the Cady Mountains at 1-2 o'clock. The part seen here is composed of dark Miocene volcanic rocks locally mantled by light-colored, windblown sand and silt picked up from the Mojave River alluvial plain by prevailing westerly winds. Little knobs on both sides of the freeway in the next several miles are composed mostly of Barstow beds (light) and volcanic rocks (dark).

▶● (Note odometer at the Harvard Road overpass.) About 1.5-2 miles beyond, Alvord Mountains can be seen about 10 miles away at 9-10:30 o'clock. The lighter-colored materials on the skyline near the tower at the east end are young, uplifted, gently deformed fanglomerates. They rest upon much darker Tertiary volcanic rocks of complex structure, which in turn overlie older granitic rocks that have intruded a still older metamorphic sequence exposed mostly toward the west end of the mountains. The irregular light areas within the dark west-end rocks are surficial deposits of wind-blown sand and silt.

● Nearly 5 miles beyond Harvard Road approaching a southward curve, the freeway dips into a broad shallow swale. Here are some dissected soft silty beds deposited in a body of water formerly covering 200-300 square miles in this basin, named Lake Manix. Large bodies of water covered this area at least twice and possibly several more times within the last 15,000-75,000 years when the San Bernardino Mountains had more snow and when the runoff to Mojave River was greater. Wet intervals in desert regions are called pluvial periods, a handy term. Pluvial Lake Manix was a little over 200 feet deep. Its basin was probably created by deformation, perhaps faulting, but it was eventually breached by overflow and downcutting along Afton Canyon.

Animal life was abundant along the shores of Lake Manix. There were shellfish, turtles, beetles, and larger animals including early dogs, bears, cats, mammoths, horses, camels, antelopes, bison, and sheep. Their fossil remains are found in the lake-bed and shore deposits. Among the abundant birds were pelicans and flamingos. The picture of a pink flamingo standing stiffly at attention on one leg in the Mojave Desert is a bit incongruous.

● Approaching the rest area, the prominent mountain at 10 o'clock is a mixture of old granitic and metamorphic rocks.

● Leaving the rest area, the high sharp skyline peak at 1 o'clock is Cave Mountain (3585 ft.). Extensive small-scale gullying of the hills across the valley at 2:30 o'clock suggests relatively unconsolidated deposits, presumably in large part fanglomerates. At a mile beyond the rest area, the commercial plants (1 o'clock) along the railroad (Dunn and New Dunn) are engaged in processing materials trucked out of the surrounding desert.

● In another mile one looks down Afton Canyon at 2 o'clock. As noted, it was cut by the outlet flow from Lake Manix and is now followed by Mojave River and the Union Pacific Railroad. Bedrock in the channel floor forces water to the surface, and short reaches of flowing stream usually exist in the canyon throughout much of the year.

● In 2 more miles, just beyond the Afton Road Exit 1 mile sign, is another and more extensive area of dissected Lake Manix beds. These pale-green, fine silts are capped by younger, brownish gravels.

(2) Approaching the overpass at Afton Road, the freeway rises and passes through a magnificent gravel beach ridge formed along one of the higher levels of Lake Manix (Photo 3-2). This is a good place to turn off for a look around. The ridge is hard to see from the entrenched freeway route, but after rising to the surface beyond the overpass, look back at 3-4 o'clock to see the backside of the beach ridge and the little playa flat enclosed by it. If you turn off, these relations are easily seen, and the road to Afton Canyon campground at the boulevard stop runs southeast

Photo 3-2. Beach ridge of ancient Lake Manix near Afton Canyon, 2 miles south of Interstate Highway 15, as viewed southward. (From **Geology Illustrated** by John S. Shelton, W. H. Freeman and Company, Copyright © 1966).

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▶ along the crest of the beach ridge. (Note odometer reading here.)

- Back on the freeway, starting a mile beyond Afton Road, blocky granitic rock with a modest coating of desert varnish makes up the hill slopes on both sides for the next few miles. Much wind-blown sand and silt from the west have accumulated locally on these slopes.
- In about 5 miles from Afton Road, the freeway starts a descent to Cronese Valley (East Cronese Dry Lake) between Cave Mountain on the south and Cat Mountain on the north. These mountains are composed of an older and tougher type of granitic rock, which is why they are higher, more rugged, and more darkly varnished. Note the very rough, rocky fans along the base of Cave Mountain. The light areas thereon were recently bulldozed during highway construction.
- Nearing the floor of Cronese Valley about 7 miles from Afton Road, note 'hat rock exposures close to the road on the south are lighter than the rocks higher up the slope. Their desert varnish has been removed by the blasting of wind-blown sand which now partly mantles these lower slopes.
- In crossing the concrete bridge about 8.5 miles from Afton Road, don't be surprised to see a small flowing stream in winter. This is Mojave River water which comes to the surface in Afton Canyon. Upon leaving Afton Canyon the Mojave can flow east to Soda Lake or north to Cronese Lake. At times of flood it often does both. In 1916 flood waters accumulated in East Cronese Lake to a depth of 10 feet.

The abrupt mountain face east of Cronese Valley is the west face of Soda Mountains, an irregular mass extending to Baker. The part viewed here is mostly granitic rock with a mantle of lava on the south flank.

(3) Between 1 and 1.5 miles beyond the Basin Road overpass, passengers can look back to 8 o'clock to see a large dune on the east face of Cronese Mountain. It consists of sand blown over the top of the mountain from the west, and the shape viewed in favorable light strongly resembles a cat lying on its tummy with ears and tail visible, hence the names "cat dune" and Cat Mountain.

- Rocks in the mountains around Rasor Road exit are mostly Tertiary volcanics, locally mantled by wind-blown material. About a mile beyond Rasor Road is a good view of Kelso Dunes some 25 miles to the southeast at 3 o'clock. These dunes, fully 500 feet high, lie at the east end of Devil's Playground, a barren sandy windswept plain across which sand is driven for 35 miles from the mouth of Afton Canyon by prevailing westerly winds. The dunes have accumulated at a site where the local topography allows strong storm winds from the north, south, and east to counterbalance the prevailing westerly wind. These beautiful dunes are accessible by road from either Baker or Amboy.
- Looking ahead within the next 2-3 miles, the hills south of the freeway at 12:15-3 o'clock are part of the Soda Mountains, and the rocks are granitic. North of the freeway, at 10:30-12 o'clock, are subdued, lighter-colored hills composed of uplifted young fanglomerates within which are local patches of dark, more highly colored rocks. These dark areas are mostly slices of Jurassic-Triassic (150-200 m.y.) metamorphic rocks inserted into the fanglomerates by faulting within the wide, complex, north-trending Soda-Avawatz fault zone.
- After passing Zzyzx Springs Road, Kelso Dunes are seen again at 2 o'clock in the far distance, and the floor of Soda Lake appears in the mid-foreground. Soda Lake is a remnant of a much larger pluvial water body, Lake Mojave, which lay in this basin 10,000-15,000 years ago.
- In less than a mile the freeway passes through deep cuts in the fanglomerates mentioned above. From the elevated position beyond the cuts one gets good views of Soda Lake and the Devils Playground beyond it at 4 o'clock.
- Coming down the long straightaway toward Baker, the even skyline at 1-1:30 is capped by lava flows, and the dark conical peaks at 2 o'clock are just a few of the twenty-six volcanic cones making up an extensive volcanic field in that area.

(4) Take the first turnoff into Baker, and as the highway starts to climb toward the freeway overpass, look to the far (northwest) side of the highway to see a pit in fine bedded gravels of a Lake Mojave beach.

- The dark rock knob just north of the road coming into town is Baker Hill. It is composed of badly broken and deformed upper Paleozoic (Permian) limestone about 250 m.y. old. Note the many small niches and caverns formed by weathering.
- Baker is a good place to gas up and to get supplies before heading north to Death Valley on State Highway 127.

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Segment D—Baker to Shoshone, 57 miles, Figure 3-5

● (Note your odometer reading in the center of Baker.) Going north on State Highway 127 toward Shoshone, rocks in the mountain front immediately to the west are early Precambrian metamorphics, probably between 1 and 2 billion years old, and much younger Mesozoic (150 m.y.) granitic intrusives. The light spots mark outcrops of carbonate rock which are not able to preserve a coating of desert varnish. The high Avawatz Range looms on the sky-line at 11 o'clock. Rocks in the mountains to the east are also largely early Precambrian metamorphics and local Mesozoic igneous intrusives.

● About 3.5 miles from Baker the floor of Silver Lake playa appears close by on the west and continues for the next 5.5 miles. This is a remnant of ancient Lake Mojave, fed and nourished largely by an expanded discharge from Mojave River. Even now, during very wet winters, Mojave River waters get this far. The lake flat was flooded in the winter of 1968/69, and in 1916 the basin was filled to a depth of 10 feet by Mojave River floods.

Keep watching the base of the hills along the far (western) shore as you travel north. In places you should be able to make out disconnected parts of a horizontal line (a wave-cut cliff) marking a Lake Mojave water level (Photo 3-3). From 6.5-7 miles out of Baker is a good place to look. Indians favored the shore of this lake, and evidences of their occupation, perhaps 10,000 and more years ago, are found in many sites along this abandoned strand line.

● The piles of light colored material near the road, about 8 miles from Baker, are talc, dumped by trucks hauling from mines to the north. Look west for shoreline remnants as you continue north. Read ahead in (1).

(1) Approaching the powerline crossing at the north end of Silver Lake, the horizontal mark of the old shoreline is clearly

Photo 3-3. View west across Silver Lake playa. Horizontal line along base of hills marks ancient Lake Mojave water level.

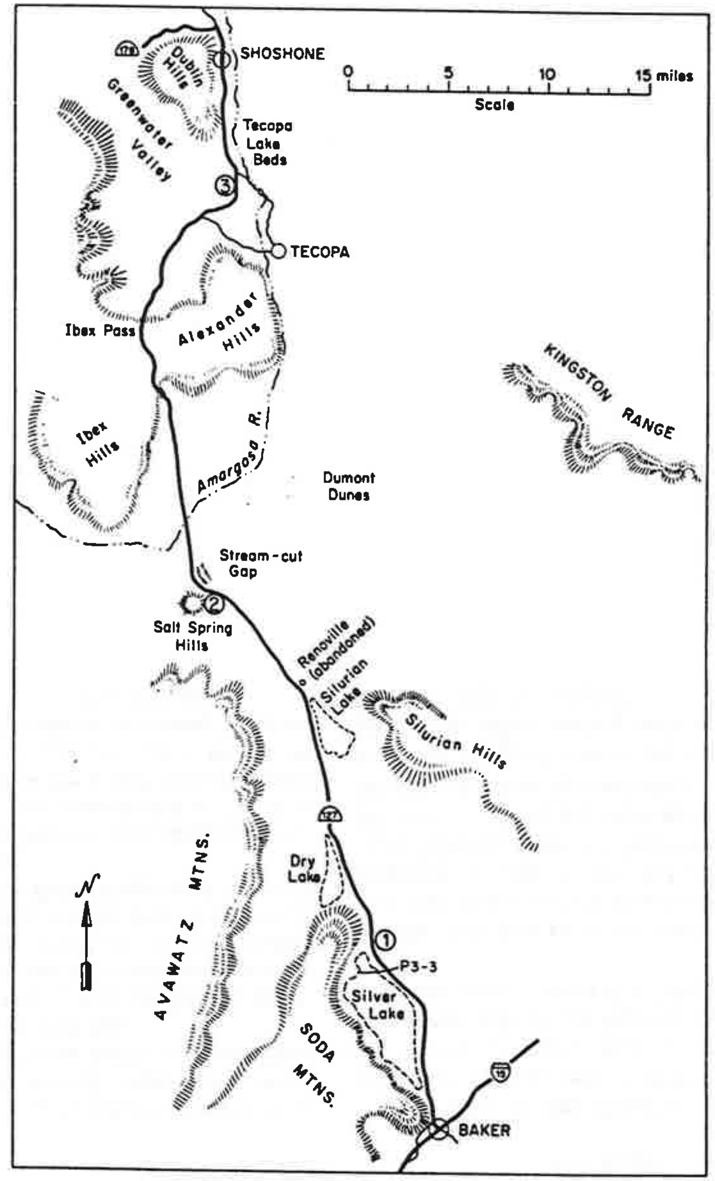

Figure 3-5. Segment D, Baker to Shoshone.

(82)

visible in good light along the base of the hills across the lake. The light-colored rock knobs here and at the northwest corner of Silver Lake are composed of a carbonate rock named dolomite, which is like limestone except for a higher magnesium content. Immediately west of the highway, 0.3 mile before passing under the powerlines, is a low beach ridge closing off a small shallow basin on its north side. A good view of relationships is seen by looking back at 7 o'clock from the first powerline crossing. The broad low ridge under the powerline is another slightly older beach. (Note odometer at the powerlines.)

• Within a mile beyond the powerlines, after the first curve west, one sees the Avawatz Mountains at 10 o'clock. Their less rugged southern flank, at 9:30 o'clock, is underlaid by the Avawatz Formation, a lower Pliocene (8 m.y.) fossil-bearing, terrestrial, sedimentary accumulation lying partly within the complex Soda-Avawatz fault zone seen earlier west of Baker. The darker rocks making up most of the range front are early Precambrian metamorphics.

• At 3 to 4 miles beyond the powerline, the Silurian Hills begin to attract attention at 2-2:30 o'clock. They feature an extremely complex structural arrangement of old-old rocks (early Precambrian), old rocks (late Precambrian), medium-old rocks (Paleozoic), not-so-old rocks (Mesozoic), and young rocks (Cenozoic) of igneous, metamorphic, and sedimentary varieties. They will be in view for a good many miles, and occasional glances to the east will give you a sense of their lithologic and structural complexity.

• After taking the second curve, this one to the east, about 4 miles from the powerlines, look at the large alluvial fans built eastward from the base of Avawatz Mountains at 9-11 o'clock. If light is good, several generations of fan surfaces can be distinguished on the basis of differences in degree of darkness, desert pavement development (vegetation-free, smooth areas), and dissection (gullying). Especially good views of these relations are seen at 9:30 o'clock just short of the next curve.

• Beyond this third curve, good views of the Silurian Hills are obtained across the valley at 2 o'clock. On the distant skyline at 1-1:30 o'clock, the high rugged peaks of the Kingston Range loom up. The Kingstons are the home of 7000 feet of weakly metamorphosed, late Precambrian sedimentary rocks, largely shale, sandstone, quartzite, limestone, dolomite, and conglomerate, collectively termed the Pahrump Group, of which we will see something later.

• About 10 miles from the powerlines, Silurian Dry Lake lies immediately east of the highway. Now dead ahead, about 10-12 miles away, are the Dumont dunes (Photo 3-4). They consist of sand piled up by winds blowing from several different directions. The raw alluvium deposited by the Amargosa River is one source of sand. The often snowcapped, distant, skyline mountain behind the north end of the Kingston Range is Charleston Peak (11,918 ft.), located in the Spring Mountains of Nevada.

• Beyond Silurian Lake the road curves west again, and at 11 o'clock, about 5 miles ahead, are the Salt Spring Hills composed of Cambrian (500-600 m.y.) quartzite beds. When closer to these hills note the two different shades of desert varnish, brown and black. The brown variety forms on white to pink quartzite layers; the black occurs on very dark quartzite layers.

(2) In another 5 miles (17.3 miles from the powerlines), the highway swings west once again, and ahead at 12:15 o'clock is a narrow gap through a rock ridge east of the highway (Photo 3-4). This gap was cut

Photo 3-4. Vertical high-altitude air photo of Salt Spring Hills area, north to top, scale in lower right. (U. S. Air Force photo taken for U. S. Geological Survey, 744V-074).

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by overflow from Lake Mojave in pluvial times when Mojave River water ran all the way to Death Valley along this route, joining the Amargosa River ahead. The light-colored, fine-grained deposits and the shoreline features (low bank and horizontal markings) at the base of Salt Spring Hills seen within the next mile indicate that some ponding of waters occurred here before cutting of the gap was completed. At the north end of Salt Spring Hills the highway makes two or three little curves, and when it straightens out, the green trees seen at 1 o'clock mark the location of the gap.

If you would like to have a closer look at this gorge, a rocky but easily passable desert road turns off just short of the little granitic knob west of the highway less than 0.5 mile ahead (19.7 miles from the powerlines). The turnoff is roughly 75 feet beyond the State Highway 127 sign, and the distance to the gap is 0.3 mile. It is interesting to imagine what this spot was like when it harbored a stream powerful enough to cut a gorge into granitic rocks. Even at present, the good growth of salt cedar and cane grass make this spot a bit unusual. East of the gap is a mass of dark, variegated, Cambrian sedimentary rocks, and farther to the east-northeast are exposures of the same Cambrian quartzites seen in Salt Spring Hills.

● Back on State 127 and rounding the corner, we get a good view of granitic rocks at 3 o'clock and of the Cambrian quartzites at 2 o'clock. The southern end of the Death Valley depression lies to the west. A dirt road takes off into it within a mile, at the Harry Wade historical monument, but we will enter from the side some 30 miles north. Small sand dunes are seen east of the road at 1 o'clock from this intersection (Photo 3-4). We come to the Amargosa River (sign) in two miles. Don't be surprised to see water running here in winter or spring. The river rises in the high Spring Mountain Range in Nevada and at times flows all the way to the salt flats of central Death Valley. (Note odometer reading at the Amargosa River crossing.) ◀

● After crossing the Amargosa River, the southern end of Ibex Mountains (Saddle Peak Hills) lies at 10-11:45 o'clock. The dark reddish-brown rocks are part of the late Precambrian Pahrump Group of sedimentary beds, and the lighter rock capping the tops of ridges at 11:30 o'clock is the slightly younger Noonday dolomite.

● In another mile the Dumont sand dunes are again in view at 3:15 o'clock.

● At 5.5 miles from Amargosa River, the Noonday dolomite is close by on the west. The subdued hills at 1-3 o'clock are composed largely of fanglomerate. In places this deposit contains lenticular beds composed of large angular fragments of a single type of rock. These probably represent landslide and rockfall accumulations and, being made up of just one kind of rock, are called *monolithologic breccias*. This fanglomerate is similar in character, and possibly equivalent in age, to a late Pliocene-early Pleistocene (2-4 m.y.) deposit of the Death Valley area, the Funeral fanglomerate.

● About 7 miles beyond Amargosa River a microwave relay station west of the highway is passed, and the ascent to Ibex Pass begins. The dark rocks just east of the highway here are volcanics. The deep roadcuts 1.5-2 miles up the grade are in badly fractured granitic rocks, but near the Inyo County line we pass into uplifted fanglomerate deposits containing largely cobbles of granitic and volcanic rocks.

● (Note odometer at the summit just beyond the county line.) ◀ Descending from Ibex Pass, the northern Ibex Mountains, largely a complex of early and late Precambrian rocks are on the west. The white spots are talc workings. The well-bedded rocks of complex structure at about 10:30 o'clock are Cambrian. Within 1.5 miles from the pass we get good views of the dissected badlands formed in light colored Tecopa lake beds on the valley floor ahead, at 11-1 o'clock. Our old friends the dark Cambrian quartzite beds of Salt Spring Hills are seen in the near ridge at 1 o'clock.

The high ranges beyond the lake basin appear striped because of layering within the thick section of early Paleozoic sedimentary formations composing them. The nearer range is the Resting Spring, and the far skyline range is the Nopah. The combined thickness of beds exposed in these mountains is 23,000 feet, ranging from Cambrian (600 m.y.) to Pennsylvanian (300 m.y.).

● In another mile the Dublin Hills, with beautifully layered Cambrian sedimentary formations, loom up dead ahead. The wide valley at 11 o'clock is Greenwater Valley which we enter farther north. The variegated peak at 10:45 o'clock at the north end of Ibex Mountains is Sheephead Mountain.

● In another 2 miles we are down within the low hills of dissected Tecopa lake beds. A thin layer of younger, dark gravel laid down on top of the soft lake beds, before they were dissected, locally drapes down over the slopes, partly masking the deposits beneath. In places, where the layers of lake silt have considerable coherence, some steep castellated cliffs have developed.

(3) About 2 miles beyond the first turnoff (paved) to Tecopa, low, crumbled remains of the adobe walls of the old Amargosa borax works are seen close to the highway on both sides. This site was used during summers from 1882 to 1890 when the heat on the floor of Death Valley prevented crystallization of solutions at the Harmony borax plant.

● Within 0.3 mile beyond the second turnoff to Tecopa (also paved), you can begin to see the faint remains of old, narrow, hand-dug trenches extending up the crests of ridge spurs close to the road on the west. Don't confuse them with the fresher, much wider bulldozer scars. These old, partly infilled trenches, about 2 feet wide and now 1-2 feet deep, were dug during World War I in search of nitrate deposits, the supply from Germany having been cut off. It is amazing that they have survived so long.

● Shoshone is the jumping off place to Death Valley; food, gasoline, water, and a motel are available.

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Segment E—Shoshone to Death Valley Floor (Ashford Mill) via Salsberry and Jubilee Passes, 29 miles, Figure 3-6

● The drive from the crest of Salsberry Pass to Furnace Creek Ranch is magnificent; plan to do it at a leisurely pace if you possibly can. In the first mile going north out of Shoshone, the chunks of black rock on slopes immediately west of the highway are derived from a group of lava flows known as the Funeral basalt (1-2 m.y.),. and the brownish ridge of well layered rocks 2 miles to the east consists of late Tertiary (3-8 m.y.) volcanic and sedimentary materials. The ridge is part of a fault block lying in front of the darker Resting Spring Range of Cambrian beds (500-600 m.y.). Approaching the Salsberry Pass turnoff (Highway 178) about 1.5 miles out, the abandoned Gertsley borax mine (white spot) is visible near the base of the hills at 2:45 o'clock. (Note odometer reading at turnoff.)

● After turning west on Highway 178, we start to circle the north end of Dublin Hills. In about 1.5 miles the Greenwater Range fills the skyline from 11:30 to nearly 3 o'clock, the part seen here being largely Tertiary volcanics. The far-away sharp peak at 3 o'clock is Eagle Mountain, along the Death Valley Junction road, which is composed of Cambrian sedimentary beds.

● Within another mile we begin to see that the east flank of the Greenwater Range is locally mantled by a thin layer of black rock tilting eastward and resting on top of the more highly colored Tertiary volcanics. These are lava flows of the Funeral basalt, and we see them frequently in the Death Valley country.

● In about 5 miles from the turnoff we are descending gently into the wide Greenwater Valley. Ahead at 11 o'clock is Sheephead Mountain, largely volcanic. In another 2 miles we will be near the center of the valley, a broad synclinal downwarp. We earlier crossed its southern end between the two turnoffs to Tecopa. To the north dark layers of Funeral basalt mantling both its sides are inclined inward toward the valley axis. This is evidence for structural downwarping of geologically recent date, for the Funeral basalt is no more than 1-2 m.y. old.

● In another 2 miles the highly irregular color pattern in the hills at 1:30-2:30 o'clock suggests a complex mixture of rocks. Those exposures are indeed part of a structurally jumbled mass, appropriately named the Amargosa chaos. More specifically they belong to the Calico phase of the chaos, an obviously appropriate name. Other less colorful phases are the Virgin Spring and Jubilee, to be seen ahead. In simplest terms, the chaos phases are breccias formed from shattered and jumbled blocks or sheets of rock shoved out over the ground surface along a series of very gently inclined fractures, called thrust faults. Landslides and rock falls occurring along the freshly created thrust fronts contributed to the complexity of the resulting deposits.

● Volcanic flows and tuffs are seen south of the road ascending to Salsberry Pass. (Note odometer reading at the pass.)

(1) Within a mile after crossing the pass, we come out onto a fairly smooth alluvial surface. About 2.5 miles from the pass, you might stop to survey the country; geologically, it's a little complex. The dull, dark rocks in the near hills to the south at 9-10 o'clock are early Precambrian metamorphics. The closest rugged dark rock mass at 1 o'clock (Rhodes Hill, Figure 3-6) is early Precambrian gneiss. The more colorful outcrops near its northern base are

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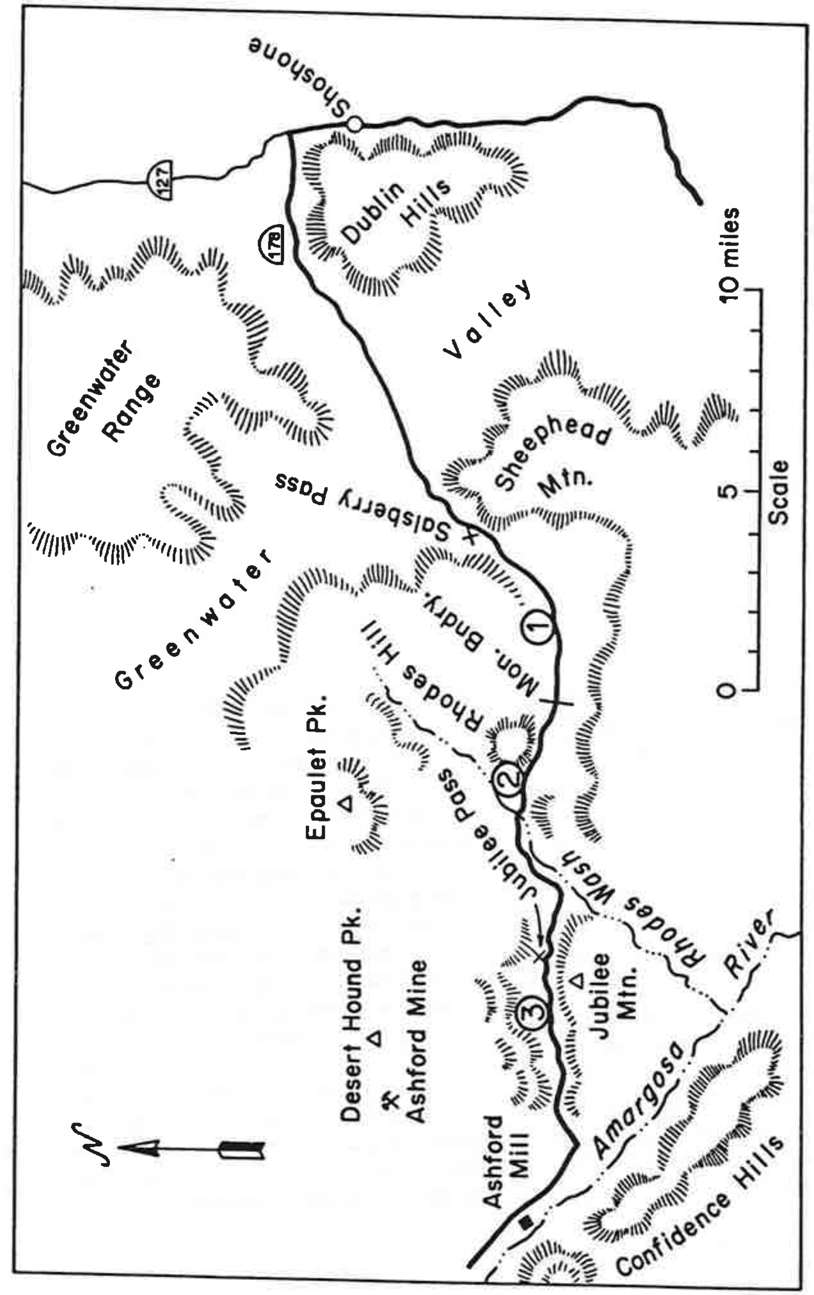

Figure 3-6. Segment E, Shoshone to Ashford Mill (Death Valley).

part of the Virgin Spring phase of the Amargosa chaos. Structurally, the Virgin Spring chaos rests in fault contact upon the gneiss. At 2-2:30 o'clock on the slopes farther back are the highly colored Tertiary rocks of the Calico chaos. The smooth topped skyline hill at 2:15 o'clock (Epaulet Peak) is capped by our friend the Funeral basalt. The highly colored rocks of the sharp skyline peak (Salsberry Peak) at 3:15 o'clock are also part of the Calico chaos.

(2) About 5.5 miles from Salsberry Pass, beyond Rhodes Hill and past the Monument Boundary, stop short of the little rock knobs just north of the highway. At 2:45 o'clock on the skyline is Epaulet Peak, and on its slopes is Calico chaos. The darker but variegated rocks, still lower at 2:30 o'clock, are Virgin Spring chaos. They rest on some more uniformly dark, greenish-gray, early Precambrian rocks at 2 o'clock which are in the center (core) of an anticline. Low down at 1 o'clock is some more Virgin Spring chaos on the south limb of the anticline. The highly colored rock knob alongside the road ahead is also Virgin Spring chaos. Little knobs and patches of the same chaos are seen resting on early Precambrian rocks just south of the highway along here.

- About 6.4 miles from Salsberry Pass, just south of the highway, is a rocky knob displaying much cavernous weathering. If you stop and walk over, you will find it is composed of broken up (brecciated) carbonate rock (dolomite), probably of Cambrian age. It is part of the Virgin Spring chaos and rests in fault contact with early Precambrian rocks on its west side. In this region, cavernous weathering is a characteristic of chaos rocks.
- In just over 7 miles from Salsberry Pass, where the road first curves south, is a high reddish cliff close on the south side of the road. It exposes chaos resting on weathered, rust-stained, Precambrian rocks. The contact is one of the Amargosa thrust-fault surfaces.
- Swinging around the curve beyond the cliff, the subdued near hills at 2-3 o'clock are Funeral fanglomerate, a late Pliocene-Pleistocene (1-4 m.y.) deposit which is younger than the chaos. The high, pointed, dark peak ahead at 1 o'clock is Jubilee Mountain, composed of a coarse-grained, early Precambrian gneiss.

Looking down the wash ahead (Rhodes Wash), one sees the floor of Death Valley and the rounded whale-back surface of Confidence Hills, a faulted anticlinal structure in soft late Tertiary sedimentary rocks.

- Shortly the highway abandons Rhodes Wash, curves west, and ascends a grade for a mile to the summit of Jubilee Pass. From here the high point on the skyline at 2 o'clock is Desert Hound Peak. The brown to reddish rocks on its lower slopes and to the left are Virgin Spring chaos. The upper part of the peak consists of early Precambrian metamorphics which compose the center (core) of the Desert Hound anticline, a major structure extending north to Mormon Point. (Record odometer reading here.) ◀

(3) In about 1.3 miles from Jubilee Pass, close to the road on the north, is a knob of Jubilee phase of the chaos. The large amount of cavernous weathering seen here is characteristic of brecciated rocks. The ridge just beyond, on the south side of the highway, is part of a striking series of tilted red sandstone, fanglomerate, and volcanic layers of Tertiary age. The pink and white smooth areas seen ¼-½ mile south of the highway on this ridge are deposits of volcanic tuff within this sequence.

- Beyond the red sandstone-fanglomerate ridge and just north of the highway, the knobs with good cavernous weathering consist of Jubilee chaos. However, the last big rock knob south of the road, about 3.5 miles from Jubilee Pass, is composed of limestone and dolomite breccia of the Virgin Spring phase of the chaos.
- About 4 miles from Jubilee Pass the southern part of Death Valley is fully in view. Confidence Hills are seen again at 9-12 o'clock, and Shoreline Butte is at 2 o'clock. The horizontal shoreline markings on its slopes become more apparent after we turn north, and they are best seen in late afternoon light. A sharp eyed observer should make out at least a dozen levels.

These strandlines were cut by a lake, nearly 600 feet deep and well over 100 miles long, which lay in Death Valley between 10,000 and 75,000 years ago, named

Lake Manly. It was fed by a greater pluvial discharge from the Amargosa-Mojave rivers system and by water that flowed through Wingate Pass (Photo 2-13) into Death Valley from a deep lake in Panamint Valley which was fed largely by runoff from the Sierra Nevada (*see* Basin Ranges, province, special features).

- As we turn north at the road intersection, 4.7 miles from Jubilee Pass, the steep front of the Black Mountains lies to the east. This is a fault scarp composed largely of Virgin Spring chaos as far north as Ashford Canyon (at 1:30 o'clock). Beyond, it consists of early Precambrian rock on the flank of the Desert Hound anticlinal core.
- The ruins of Ashford Mill lie west of the road 2 miles north from this intersection.

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Segment F—Ashford Mill to Furnace Creek Ranch, 42 miles, Figure 3-7

● (Note odometer mileage opposite the Ashford Mill turnoff.) Starting 0.5 mile to the north, a succession of Funeral basalt knobs rise above the fan surface just west of the highway. Their north-south alignment suggests that they mark the trace of a fault. To the east, forelying rocky knobs scattered outward from the base of Black Mountains at 2-3 o'clock consist of Jubilee chaos. Horizontal shorelines cut into the northeastern face of Shoreline Butte are usually visible in the first mile or two north from Ashford Mill.

● In a little less than two miles the West Side road (dirt) takes off. This is a good place to stop and look around. At 10:30 o'clock near the center of Death Valley floor is a small cinder cone located on a branch of the Death Valley fault system. On the 9 o'clock skyline is Wingate Pass (Photo 2-13) through which water flowed, perhaps 75,000 years ago, from the large, 1000-foot deep pluvial lake in Panamint Valley. Immediately east of the road to the north, a scarp in black Funeral basalt extends for more than a mile. It marks the trace of another fracture within the Death Valley fault system.

(1) About 3 miles north from Ashford Mill, with good light, one can see how the little cinder cone, now at 9 o'clock, is sliced apart by right-lateral fault displacement.

● Some 4 miles north of Ashford Mill, at 1-3 o'clock, is a much dissected body of Funeral fanglomerate at the base of the Black Mountains. The contact between fanglomerate and early Precambrian gneiss composing the mountains is determined by the Black Mountains frontal fault. This is a structure consisting of individual fault segments steeply inclined to the west, which are individually linear in trend but which locally diverge from the general north-northwest bearing. Individual fractures or segments of fractures within this zone have recently been active, and we will see many fault scarplets breaking fan surfaces along the mountain base farther north. Some of these scarps are probably no more than a few hundred to a few thousand years old.

● Now is a good time to compare the huge alluvial fans on the west side of Death Valley with the much smaller cones and fans (Photo 3-8) over which we will be driving along the east side. This difference in size reflects in part the greater amount of water and debris discharged by the larger canyons of the higher Panamint Range. However, it is also a product of an eastward tilting of the Panamint-Death Valley block. Geological relationships suggest that eastward tilting has occurred in the immediate past, and tiltmeters measurements on the valley floor show that it continues today. This is the reason the Death Valley salt pan lies so close to the base of the Black Mountains near Badwater. The effect is similar to tilting a saucer partly filled with water.

Tilting has allowed fans on the west side of Death Valley to grow large by extending themselves outward onto the valley floor. At the same time, tilting depresses fans on the east side allowing them to become partly buried by valley-floor deposits, thus reducing their size. The Black Mountains frontal fault marks the eastern edge of the tilting block, and the fault scarplets cutting Black Mountain fans are partly an expression of this movement.

Death Valley is a structural depression. This means that its form has been determined by deformation, probably both warping and faulting. Any closed depression like Death Valley is geologically suspect. Nature

(87)

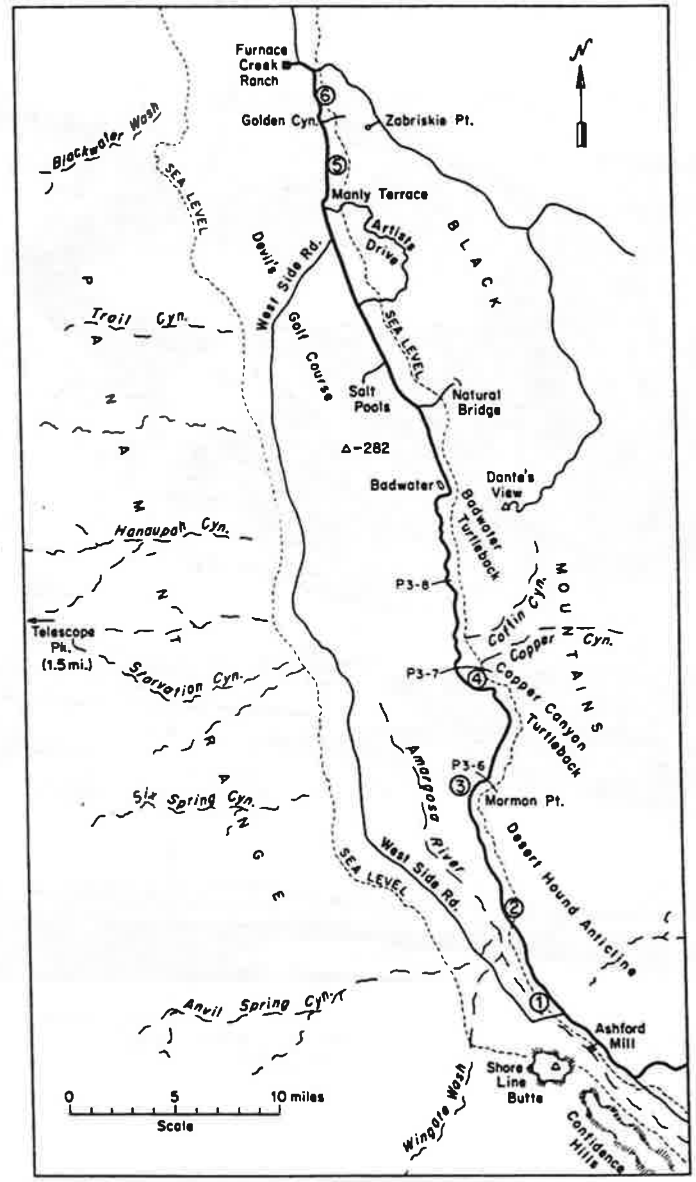

Figure 3-7. Segment F, Ashford Mill to Furnace Creek Ranch.

abhors a vacuum, and she despises closed depressions. She tries to fill them with anything available. In humid areas the initial filling is water, succeeded later by sediment. In dry regions sediment is the principal filling. Even though parts of the Death Valley floor are underlain by as much as 3000 feet of young alluvium resting on top of another 6000 feet of Tertiary sediments and volcanics, nature has not yet been able to complete the filling of this depression, indicating that it is a very young feature and that it has been formed rapidly. With a -282 feet elevation, Death Valley is the drainage sump for a large area in southeastern California and adjacent parts of Nevada. If the climate were more humid, streams would run to Death Valley from all directions, just as they did in pluvial times, and it would harbor a large lake. Death Valley is much better "watered" by springs and streams than most of the surrounding desert because of its topographically low setting.

(2) The west face of Black Mountains is a youthful fault scarp. Such scarps are locally characterized by wine-glass canyons, and a good example of one is approached at 6 and passed at 7 miles north of Ashford Mill (Photo 3-5). The base of the wine glass is the fan at the foot of the mountain, the stem is the narrow steep-walled gorge cut through the mountain front, and the bowl is the open area of dispersed headwater tributaries.

This reach of the mountain front is made up of Precambrian gneiss and carbonate rocks which yield large fragments. The fan surfaces are therefore rough, irregular, and composed of good sized boulders. At and just north of the wine-glass canyon, watch the toes of some of these fans where they come down close to the salty flats west of the highway. There the boulders develop a decrepit appearance because they are disintegrated by the growth of salt crystals within their pores.

● About 7.5 miles north of Ashford Mill the road curves back toward the mountain front, and directly ahead is a little alluvial cone with a steep, uneven, boulderly surface displaying patchy areas with different shades of desert varnish, from gray to dark brown. This and other cones along the base of Black Mountains have been built largely by a succession of rocky debris flows, and are best termed debris cones. Different flows have inundated parts of this particular cone at widely separated intervals, as indicated by variations in degree of brown varnish development on the surface stones. The grayish lobe in the south central part of the cone marks the most recent flow, possibly less than 100 years old. Note the old, faint, game or Indian trail which crosses the lower part of the fan. It is best preserved in the older and more heavily varnished parts of the cone and has been obliterated by the most recent flows. We don't know much about the rate of desert-varnish formation but suspect it is highly variable, depending upon materials and environment. In some places, perhaps hundreds to a few thousands of years may be required to produce a dark dense varnish under the climatic conditions of the last few thousand years.

● About 9 miles from Ashford Mill, just beyond a spot where the road is crowded against the mountain front by a salt pond, is a little fan with a fault scarp 7-8 feet high breaking its surface. The scarp parallels the mountain front about 20-30 feet from its base. Watch for similar scarplets in fans on up the valley; another one is about 1.5 miles ahead.

Coarse, moderately cemented fan gravels adhere to the base of the mountain front north from here. They have been elevated

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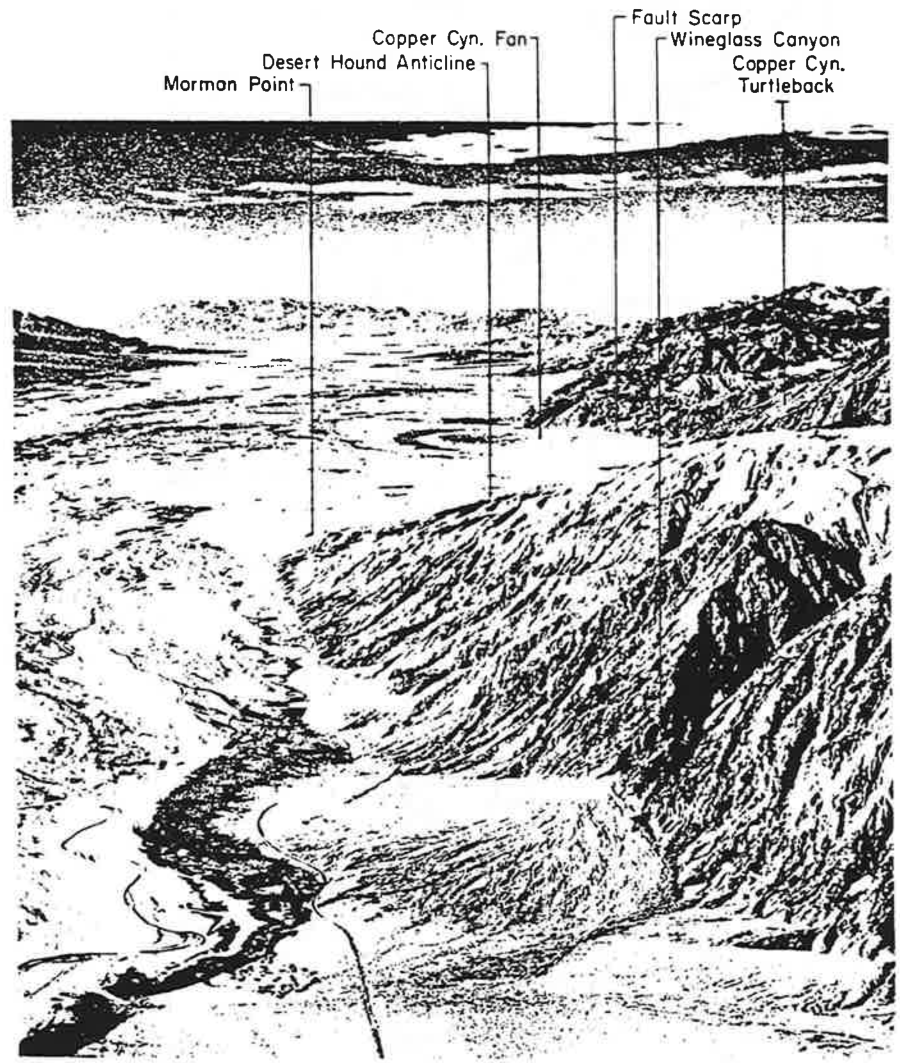

Photo 3-5. View north up central Death Valley. (Photo by John S. Shelton, 4238),

(relatively) by movement on the Black Mountains frontal fault and are sharply incised by little stream courses draining from the mountains.

● Looking west at 9:30-10 o'clock across the valley, you can see that the upper parts of fans there are dissected. That dissection is caused by eastward tilting which steepens fan surfaces causing streams to cut down.

③ At 11 miles from Ashford Mill we approach Mormon Point (watch for small sign on left). Mormon Point marks the northern end of the Precambrian core in the Desert Hound anticline earlier seen from Jubilee Pass. At Mormon Point the range front is set back to the east owing to complexities in the frontal fault system. (Stop at the Mormon Point sign and record odometer reading.) At about 12 o'clock a look northward along the face of the mountains reveals a change in appearance and character where Precambrian rocks give way to less resistant, less homogeneous, and more highly colored Tertiary volcanic and sedimentary rocks.

● As the road curves back toward the mountains, 0.5 mile from Mormon Point, a mass of Funeral fanglomerate straight ahead lies in fault contact with early Precambrian schist, gneiss, and marble which compose the high mountain face behind. Younger fan and lake-shore gravels locally mantle the north slope of the Mormon Point peninsula (Photo 3-6). Keep your eyes peeled

Photo 3-6. Horizontal lake shorelines cut in gently inclined Pleistocene fanglomerate deposits behind Mormon Point, view to south. (Photo by John S. Shelton, 3459).

for small fault scarplets cutting across alluvial fans at the head of the Mormon Point re-entrant.

● In about 1.5 miles from Mormon Point, the road straightens out on a north-northeast course. Here we look dead ahead to the southwest limb of the Copper Canyon turtleback. The nose of the structure is at the Precambrian-Tertiary contact ahead at about 11:30 o'clock.

A turtleback is an unusual geological structure. It consists basically of a mass of Precambrian rock in the core of a plunging anticline which has been exposed by erosional removal of overlying deposits. As a topographic feature it has some resemblance to the shell of a turtle. Where remnants of the overlying deposits remain, they are seen to be in fault contact with the Precambrian core of the fold. According to some interpretations, faulting occurred before the folding; according to others, it is considered essentially a contemporaneous event. The turtleback is composed of the Precambrian rocks underlying the anticlinally folded fault surface. These structural relationships are best seen a little south of the mouth of Copper Canyon ahead, and we view another turtleback north of Badwater. Mormon Point is also a turtleback, but its geological relationships are not as clearly seen.

● The bouldery fans seen along the straightaway here are composed of fragments of early Precambrian gneiss.

● About 5 miles from Mormon Point, just where the road starts to curve west around the large Copper Canyon fan, several little debris cones against the Black Mountains base display prominent patches with different degrees of desert varnish (Photo 3-7.)

④ Stop at something less than 0.5 mile out onto the gently sloping Copper Canyon fan and look around. In the mountain front to the east are gray Precambrian metamorphic rocks in the core of the Copper Canyon turtleback. They are overlain by brownish and red beds of the Pliocene (10 m.y.) Copper Canyon conglomerates (Photo 3-7). If the light is right as you continue north, you will be able to see layering in these conglomerates dipping directly into the Precambrian rocks, indicating a structural discontinuity (fault) between these units. At 3:30-4:30 o'clock at the mountain base are the varnished debris cones, one of which has patches of four different degrees of varnish on its surface. The horizontal lines on the mountain face, a little above the cones, especially at 4:30 o'clock, mark old lake shorelines. The narrow slot at the mouth of Copper Canyon, at 2 o'clock, is partly obscured from this view by a fault scarp about 75 feet high in gray fan gravels. A look back toward Mormon Point should show, in reasonable light, lake shorelines cut into fanglomerates (Photo 3-6).

● About 8.7 miles from Mormon Point, the highway heads westward again as it swings out to round the next fan (Coffin Canyon). On the skyline at 11:30 o'clock is Telescope Peak (11,049), often snowcapped in winter. It is composed of late Precambrian, weakly metamorphosed sedimentary rocks.

From here to Badwater the road swings out and in over five alluvial fans, some so youthful and perfect in form as to look almost artificial (Photo 3-8). Their size varies with the area of drainage within the mountains and the ease with which the bedrock therein is eroded. Their steepness strongly reflects the coarseness of debris (stone size) composing them. Copper and Coffin canyon fans have gentle slopes because the Tertiary rocks in their drainage basins yield much fine material. Closer to Badwater we

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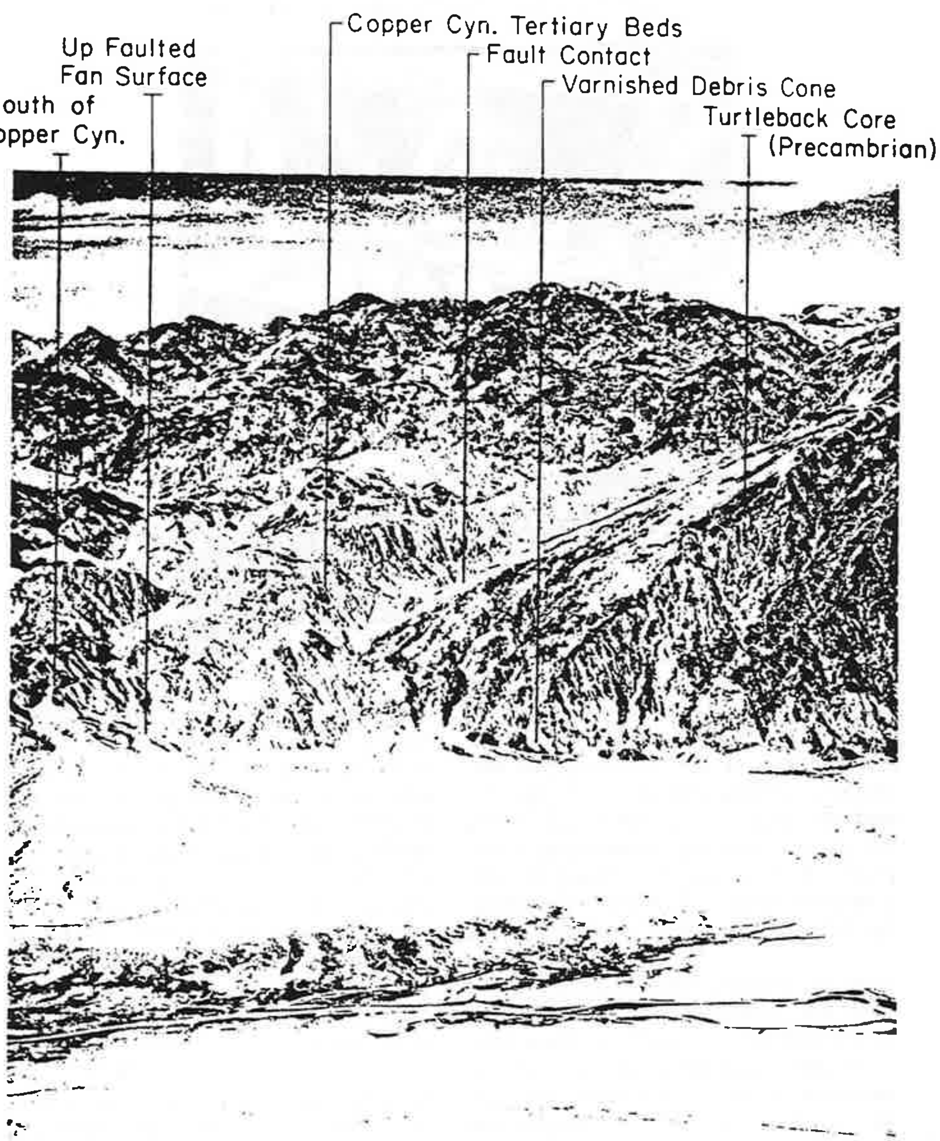

Photo 3-7. Copper Canyon turtleback, viewed from west. (Photo by John S. Shelton, 4241).

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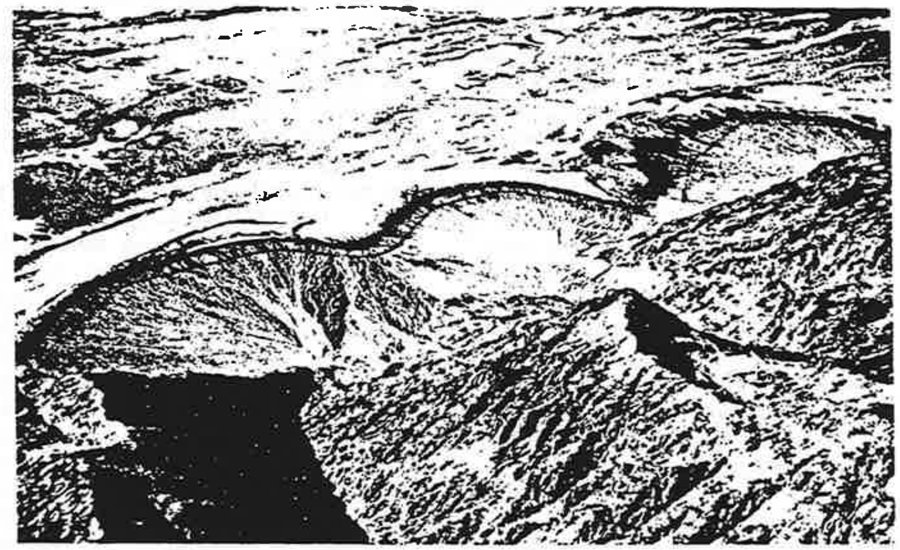

Photo 3-8. Small alluvial fans at base of Black Mountains north of Coffin Canyon (Photo by John S. Shelton, 3441).

cross rougher, steeper, more bouldery fans derived from tough Precambrian metamorphic rocks.

● Between fans where the highway closely approaches the mountain front, for example at 10.1 miles from Mormon Point and also a mile farther north, you can see that the rocks within the frontal fault zone are badly chewed up. Frankly, they are a mess.

● If traveling in the late afternoon, look across the valley to the base of fans on the west side. You may be able to see the shadowed face of a long, low fault scarp at about 11 o'clock cutting across the lower part of Hanaupah Canyon fan (Photo 2-12). Don't mistake lines of vegetation for the scarp.

● At 12.2 miles from Mormon Point, a segment of steep mountain front has broken loose in the form of a rock slide. You may be able to recognize it from the jumble of huge blocks on the lower part of the mountain face.

● In another mile, remnants of fan surfaces at two levels above a fault scarp, at the mouth of a canyon, indicate two successive uplifts. A mile beyond in a similar setting, gravels on a single uplifted surface are heavily covered with desert varnish, which indicates that they have long been undisturbed compared to those on the present fan surface.

● By the time you have reached 15 miles from Mormon Point, you have probably become aware that the salt flats has begun to hug the base of the Black Mountains. This occurs because the eastward tilting of Death Valley is particularly marked here. The character of the pan surface changes with the supply of water, but in places it is often broken into irregular, polygonal fragments 5-15 feet in diameter with turned up edges. They look something like the ice floes in the classical painting of Washington crossing the Delaware. These polygonal plates seem to grow at their edges by crystallization of salt in cracks. The breakup of boulders at the toes of fans by salt-crystal growth is again well seen just east of the road in this reach.

● Badwater is a good place to stop. North along the range front at about sea level height (*see* sign on the mountain face) are little remnants of cemented gravel adhering to the mountain front. At least some of them mark the shoreline levels of a pluvial lake. Since Lake Manly was a good 600 feet deep at maximum, these deposits must represent a lower and later stage when the water depth was only about 300 feet. Small remnants of well-worn shoreline gravels lie at higher levels, but they cannot be identified from the road.

If you walk out on the salt pan, you will see the shorelines and other relationships more clearly. Structurally, we are viewing the west flank of the Badwater turtleback sliced off longitudinally by the Black Mountains frontal fault. The nose of the turtleback is seen better a little farther north. The surfaces of fans near Badwater are broken by fault scarplets, especially the fan to the south. (Before leaving Badwater note odometer reading.)

● Leaving Badwater we move north along the west flank of the Precambrian core of Badwater turtleback. Within the first mile, considerable salt disintegration of stones has occurred just east of the road. Remnants of gravels cemented by calcareous deposits are seen at about sea level height on the mountain face.

● In little over 2 miles the nose of the turtleback becomes more apparent at about 2 o'clock. The light-colored rocks are the Tertiary deposits lying above the anticlinally folded turtleback-fault surface. Looking to about 3:15 o'clock you may be able to see a little conical knob of Tertiary rocks still resting on the crest of the turtleback. It is a remnant of the former covering of Tertiary deposits.

● In about 3.5 miles is the turnoff to Natural Bridge, a 50-foot span in sedimentary rocks of the Tertiary Artists Drive Formation. A good view of Badwater turtleback is seen from here. Northward the mountains are composed of Tertiary volcanic and sedimentary rocks aggregating a thickness of 13,000 feet. The Black Mountains contain a surprisingly small amount of *in situ* Paleozoic rock, in view of tens of thousands of feet of Paleozoic sedimentary beds exposed in adjacent areas. It appears that the Black Mountains block, lying between two large fault systems, the Death Valley on the west and the Furnace Creek on the east, has been strongly uplifted and deeply eroded sometime in the past. The total cumulative uplift and erosion must have amounted to at least 25,000 feet in order to remove the thick Paleozoic section which almost surely once covered this area.

● North of Natural Bridge turnoff is the beginnings of a group of low hills lying in front of the main Black Mountains mass. They are made up largely of highly colored Tertiary volcanic rocks with lesser amounts of associated sedimentary deposits, all rather highly deformed. These hills are part of a subsidiary fault block lying between the Black Mountains block and the Death Valley block (Photo 3-9).

● About 4.5 miles north of Badwater our highway starts across an alluvial fan of no-

Photo 3-9. High-altitude vertical photo of central Death Valley, north toward upper right corner, scale at lower right. (U. S. Air Force photo taken for U. S. Geological Survey, 374V-192).

tably gentle slope. Although large stones locally dot its surface, the fan contains much fine material derived from the Tertiary rocks composing the Artists Drive block. This is the principal reason for its gentle slope and smoothness.

● In less than a mile is the Devils Golf Course (salt pools) turnoff. A visit there will give you a look at the extremely rough surface of a relatively pure salt pan. This relief is a product of solution and of localized crystallization along fractures forming ridges and spires of solid salt.

● In another 2.5 miles is Artists Drive turnoff. Artists Drive is a one-way, 9.5-mile road through a highly dissected badland with narrow washes cut into highly colored sedimentary and volcanic rocks of the Funeral fanglomerate and Artists Drive Formation.

Just west of this turnoff are two low ridges of Funeral fanglomerate warped up by stresses developed through lateral displacement along fractures of the Death Valley fault system. If you feel a need to stretch your legs, get out and walk to the top of the ridge opposite the turnoff. Here you will find volcanic boulders deeply scoured by the blasting of wind-blown sand.

● Within the next mile look ahead at 1 o'clock to black hills of Funeral basalt at the base of the mountains. A close inspection should reveal faint horizontal scars of lake shorelines on their slopes. The nearly flat top of this projecting point is also a lake formed feature called Manly Terrace (Photo 3-9). It may have been the site of very early Indian occupation of Death Valley.

● About 10 miles north of Badwater and a mile short of the junction with West Side road, excellent stratification can be seen in the rocks composing the Black Mountain front. The great variety of colors displayed (tan, cream, lavender, green, brown, gray, and white) suggest that these deposits contain much fragmental volcanic debris. West Side road comes in about 10.6 miles from Badwater.

● (Note your odometer reading at Artists Drive exit point.) The black rocks at the base of the Panamint Range (10-11 o'clock) are Funeral basalt. ◀

● We round the point of Manly Terrace, pass Mushroom Rock, and about 1.5-2 miles from Artists Drive exit we get a good view of Furnace Creek beds which here make up the Black Mountains front. The Furnace Creek is a Pliocene (3-8 m.y.) formation, consisting mostly of light-colored, soft, silty lake beds with intercalated conglomerate and volcanic layers, 5000 feet in aggregate thickness. They are beautifully seen from Zabriskie Point off the Furnace Creek Wash road. These beds contain considerable borax, and these hills are dotted with old abandoned borax mines and prospects.

(5) About 2.1 miles from Artists Drive exit we come to a Dip sign and a 100-foot section of concrete pavement. The fan surface here is scoured, dotted with large boulders, and locally spotted with polygonally cracked deposits of dry mud, all suggestive of flooding. West of the highway, a vertical-walled gulch cut into the fan deepens to more than 20 feet near the mountains. This fan is fed by Gower Gulch which heads at the Zabriskie Point overlook. The flooding and dissection are the result of artificial diversion of Furnace Creek drainage into the head of Gower Gulch. This was done to protect buildings in lower Furnace Creek Wash and on its fan. However, it would have been only a matter of time until the diversion would have occurred naturally, for headward working Gower Gulch was on the verge of

capturing Furnace Creek. Stream capture is a common geological phenomenon. It produces anomalous drainage patterns and effects changes in the regime of both the captured and capturing stream.

● Roughly 3 miles from Artists Drive exit is Golden Canyon, an interesting little drive up a steep walled gulch cut in Furnace Creek beds.

(6) At 3.7 miles from Artists Drive exit, and extending for 0.4 mile, we begin to pass a beautiful little fault scarp, 2-7 feet high, breaking the fan surface 20-100 feet east of the road. Some of the gravels in the face of the scarp look different from those on the fan surfaces because the fault displacement has brought up gray gravels of the larger Furnace Creek fan which here underlie a thin deposit of brownish debris derived from the hills immediately to the east.

● Approaching Furnace Creek Inn, the steeply tilted, greenish beds seen are fanglomerates in the Furnace Creek Formation. We turn left on Highway 190 and proceed down the surface of Furnace Creek fan to Furnace Creek Ranch.

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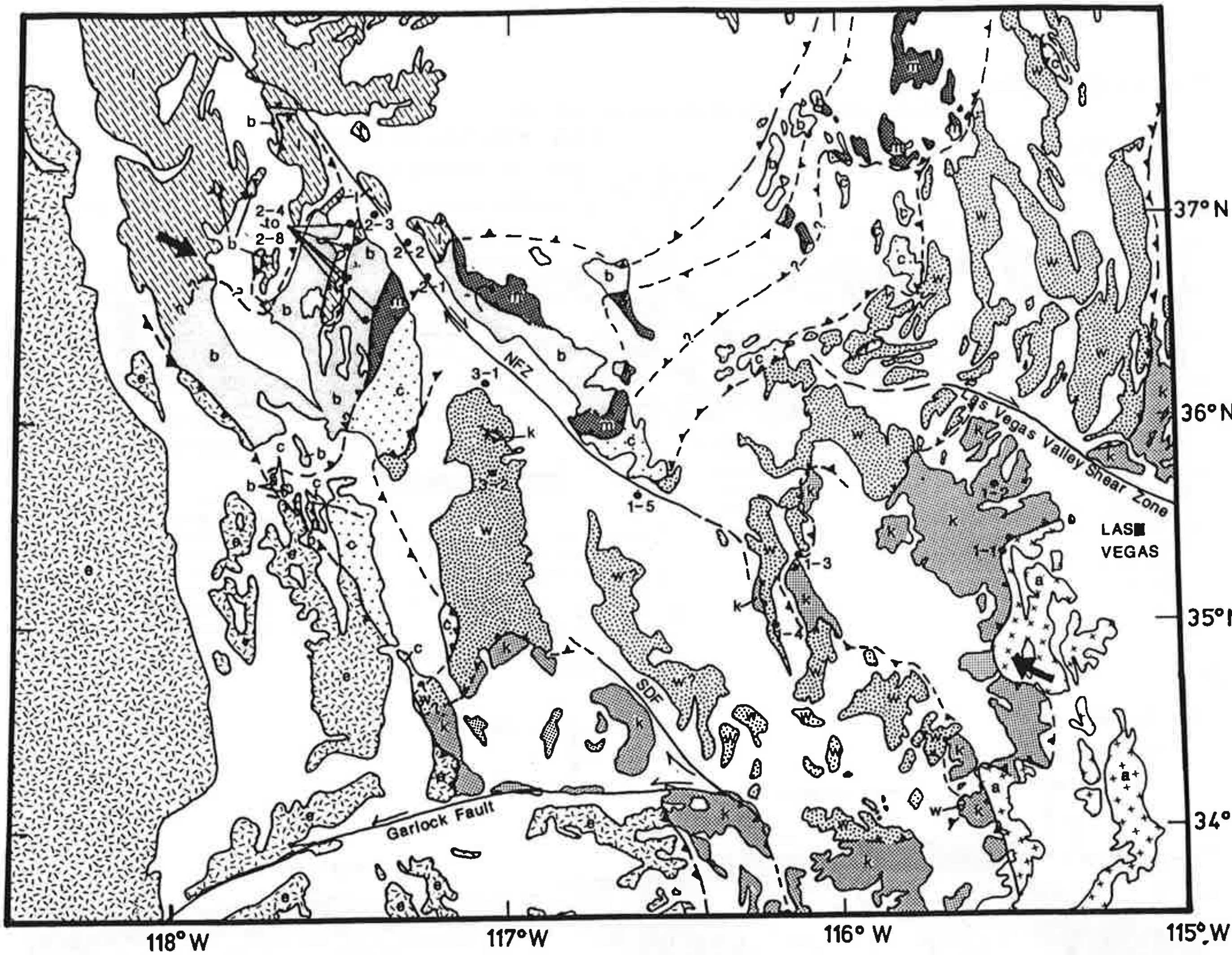


Figure 2. Interpretive tectonic map of pre-Cenozoic rocks showing major Mesozoic thrust plates, field trip stops, and major Cenozoic strike-slip fault zones. From structurally lowest to highest, the thrust plates include: a, autochthonous and parautochthonous rocks below the Keystone thrust and equivalents; k, rocks between the Keystone and Wheeler Pass thrusts and equivalents; w, rocks between the Wheeler Pass and Clery thrusts and equivalents; c, rocks between the Clery and Marble Canyon thrusts and equivalents; m, rocks between the Marble Canyon and White Top thrusts and equivalents; b, rocks structurally below both the White Top and Last Chance thrusts and equivalents; l, rocks structurally above the Last Chance thrust and equivalents; e, rocks structurally above the East Sierran thrust system of Dunne (1986). Boundaries to thrust plates are not necessarily thrust contacts, and are often intruded out or overprinted by younger faulting. Bold arrows show approximate line of section used to construct Fig. 3. Arrow in Grapevine Mountains shows 90-degree clockwise rotation about a vertical axis apparently needed to restore structures there. SDF, Southern Death Valley fault zone; NFZ, Northern Death Valley-Furnace Creek fault zone.

another 200 m.

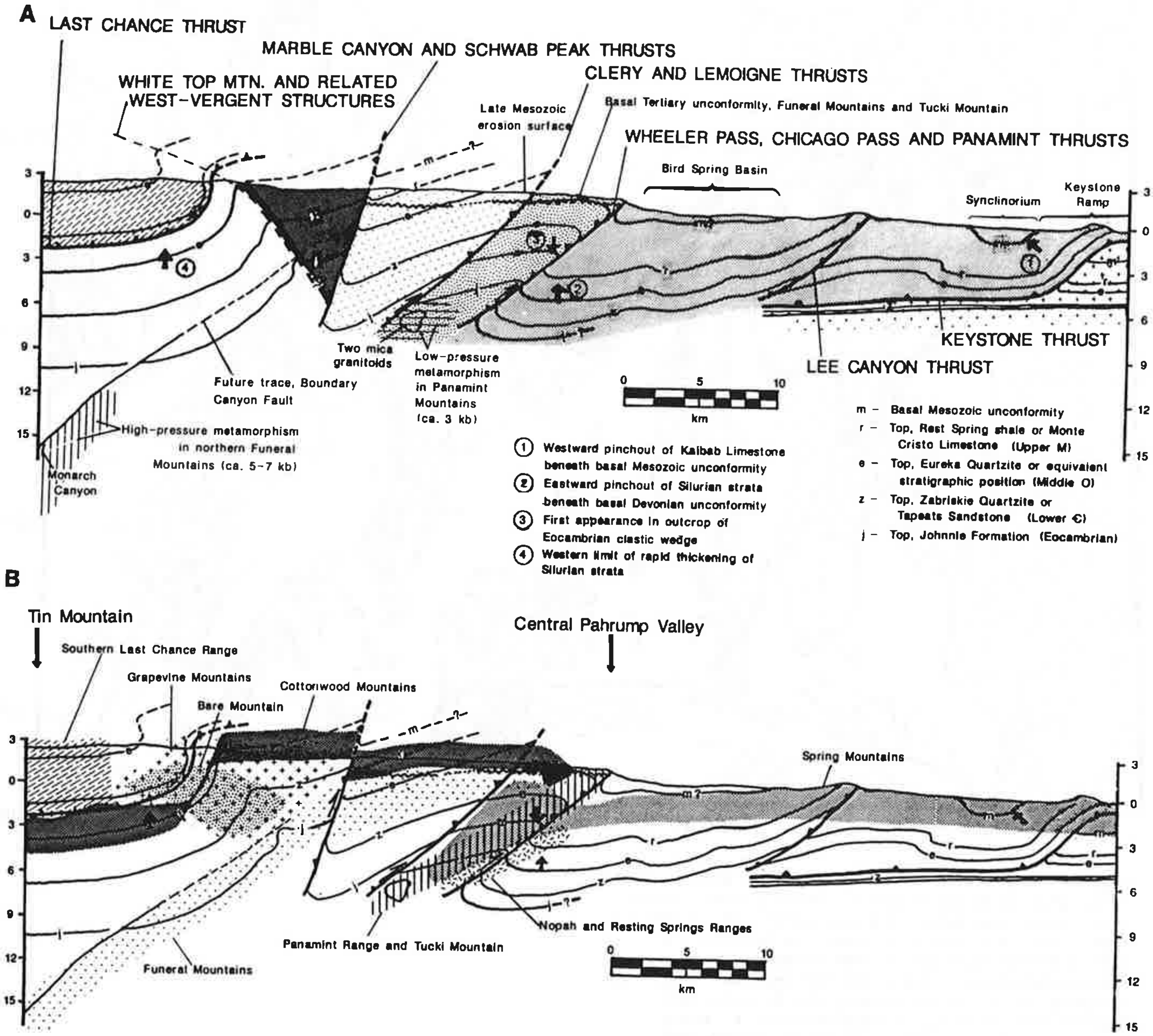
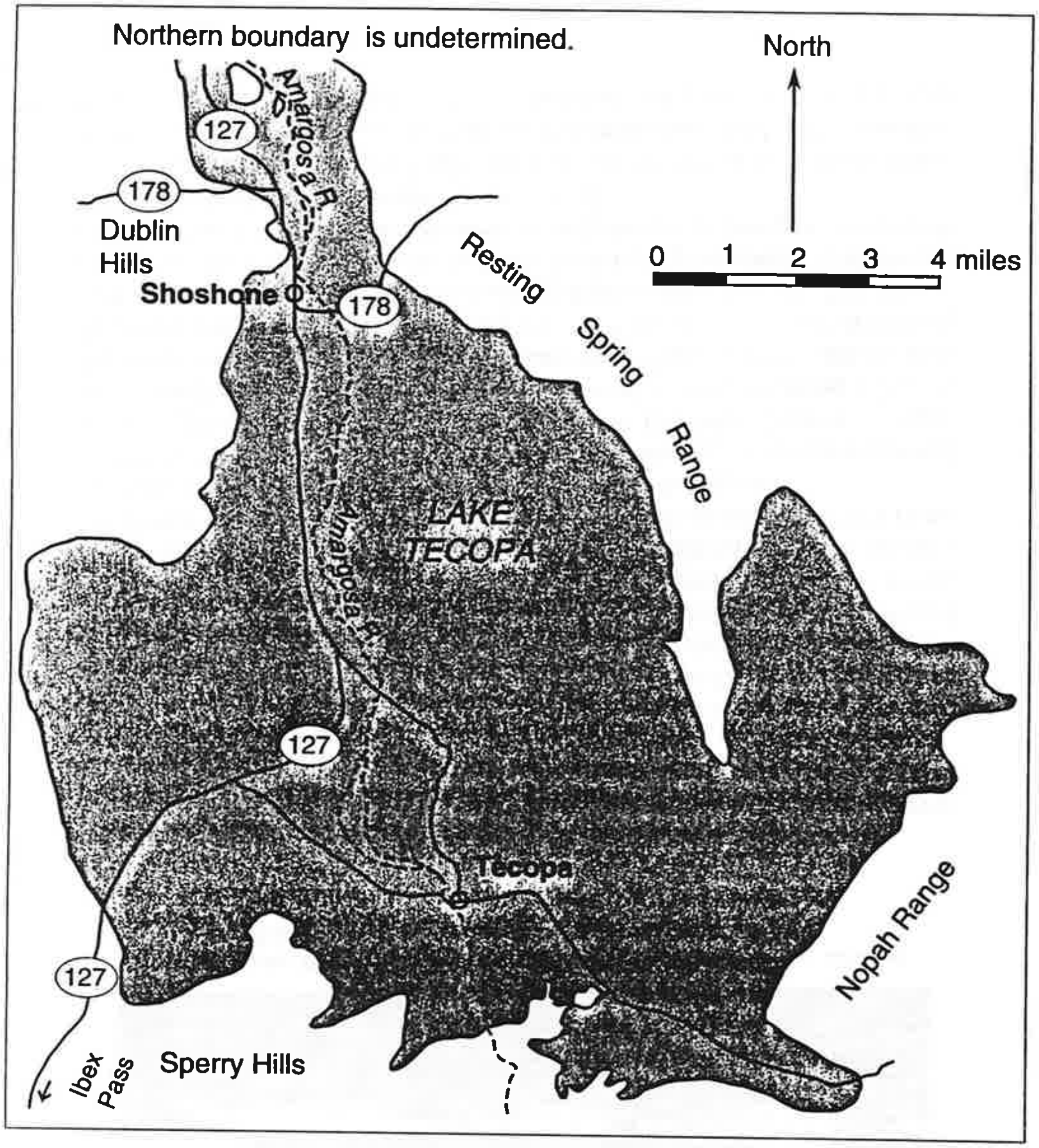


Figure 3. a) Palinspastic reconstruction of Mesozoic orogen along line of section in Fig. 2. Data compiled from many sources. Shading as in Fig. 2. see text for discussion. b) Same as in a), but shaded according to current erosion levels in the various ranges along the transect. Distance from Tin Mountain (see Fig. 4) to central Pahrump Valley is about 35 km in the reconstruction, but is currently about 185 km, giving about 150 km of extension.



Extent and configuration of paleo–Lake Tecopa.

GETTING THERE: Travelers cross the Tecopa basin going between Death Valley and Baker on California 127 and between Death Valley and Las Vegas on California 178. Watch for good views of the Tecopa basin as you descend the slopes 2 to 3 miles north of Ibex Pass northbound on California 127 and after crossing the Resting Spring Range westbound on California 178. You can see topographic details within miniature badlands in the eroded lakebeds from both highways as you approach their intersection at Shoshone.

2

A Collector of Volcanic Ashes

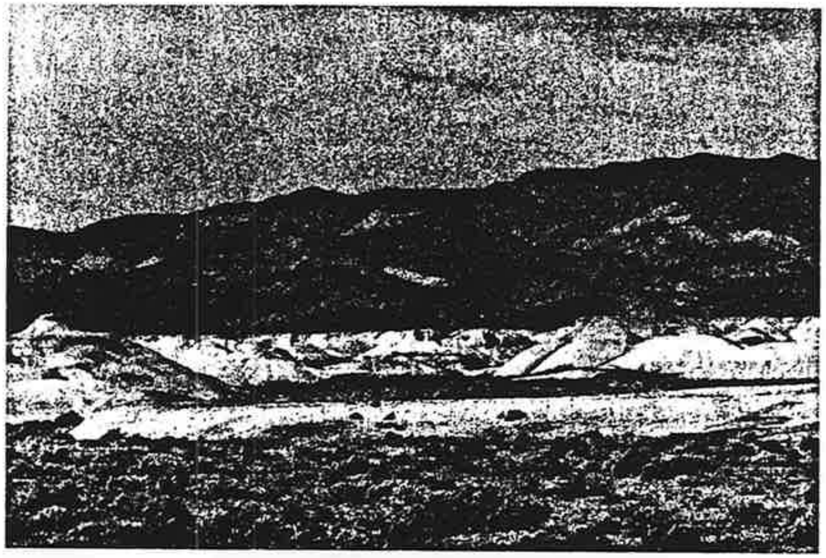
— ANCIENT LAKE TECOPA —

Visitors to Death Valley anticipate unusual scenery, but most would b surprised to crest a ridge and behold a large, deep blue lake. Yet ge logic evidence indicates that such a lake was part of Death Valley's lan scape until about 500,000 years ago—unfortunately no humans we here then to appreciate it. Around Shoshone and Tecopa Hot Springs, dozen miles east of the southeast corner of Death Valley National Par lie white, fine-grained, eroded lakebeds. These are sediments from t floor of ancient Lake Tecopa.

Lying at the terminus of the relatively large Pleistocene-age Amargo River, Lake Tecopa was a full-blown, permanent water body, not ephemeral playa. The lake covered about 85 square miles and was ov 400 feet deep. The Amargosa River presumably flowed right throu the Tecopa valley before faulting within coarse, gravelly deposits at t valley's south end created a lake basin. The Amargosa drainage regi which covered an area more than forty times the expanse of the la supplied most of the lake's water. Some water came from as far away the lofty (near 12,000 feet), often snow-capped Spring Mountains of N vada. Springs in Ash Meadow, near Death Valley Junction, augment the river's discharge. The climate when Lake Tecopa existed had to cooler and moister than it is today because the discharge of the mo ern Amargosa could not maintain a lake of Tecopa's size under t present-day evaporation rate of about 6 feet per year.

Lake Tecopa may have self-destructed by overflowing its sill at t south end. The overflow rapidly cut a channel through unconsolidat poorly sorted gravelly deposits, called fanglomerates, which accum late as part of alluvial fans. A climatic shift to wetter conditions m have contributed to the rising water levels, but the lake was bound overflow anyway. Along with water, the Amargosa River brought a lo sediment into the Tecopa basin. The sediment accumulated on the sin floor to a thickness of several hundred feet, decreasing the hold capacity of the basin and causing the water level to rise and eventu overflow.

The story of the lake is recorded in its deposits, now extensively sected by the Amargosa River and its tributaries. Erosion has not



Dissected Lake Tecopa beds west of California 127 about 3 miles south of Shoshone. Dublin Hills in background. —Helen Z. Knudsen photo

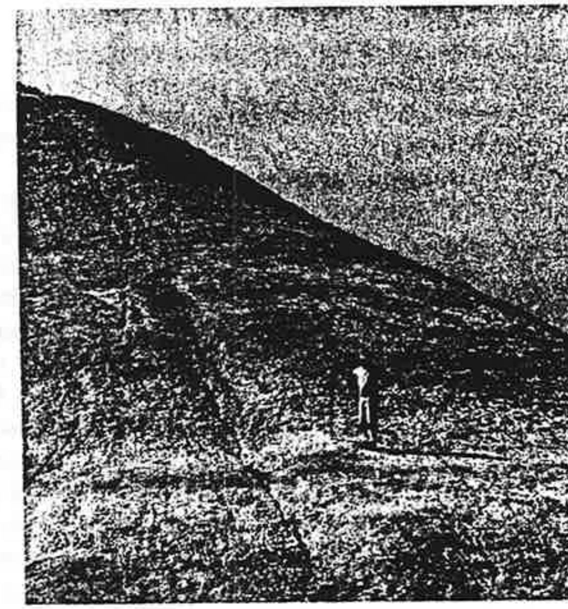
cut to the base of the lake deposits, so their total thickness is not known. Scientists have measured a cumulative column of at least 236 feet in the walls of gullies.

The lake-floor sediments are mainly mudstones of clay, silt, and fine sand, as well as shoreline conglomerate, sandstone, calcareous tufa, and numerous layers of volcanic ash. The lakebeds are not tectonically tilted or folded, but high-angle faults of small displacement cut them here and there. Greater compaction of the sediments toward the basin's center, where deposits are thickest, inclined the layers about 1 degree inward. Local settling over irregularities in the basin's floor created small compaction structures with beds tilted as much as 8 degrees.

Before the lake existed, streams from surrounding mountains carried alluvial gravel into the Tecopa valley. Once the lake drained, similar deposits of gravel once again extended into the basin, forming a cap on the lakebeds. Today you can see remnants of these gravel caps as gently sloping gravel layers composing the flat tops of ridges, small buttes, and mesas within the dissected lakebed topography. Locally, this gravel creeps down steep slopes eroded into underlying lakebeds, completely masking them. Undisturbed gravels on flats form a desert pavement with a dark coating of rock varnish (vignette 12).

The uppermost Tecopa beds contain sparse remains of fossil mammals such as horses, camels, mammoths, muskrats, and some rodents. Miners prospected the lakebeds for borates, nitrates, pumice, and vol-

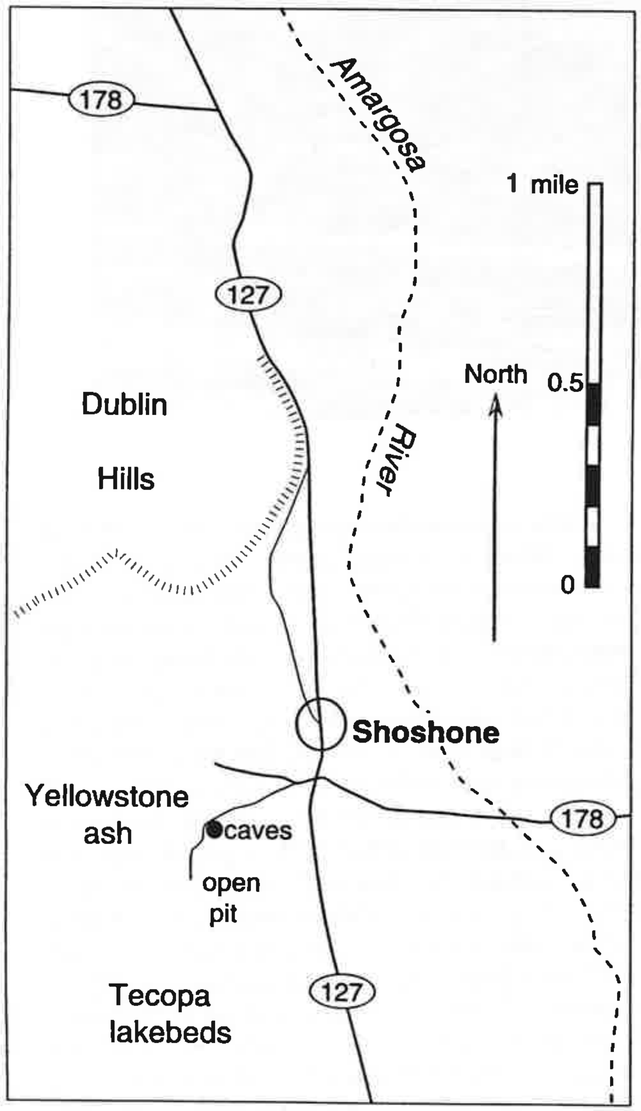
Trace of 1915, hand-dug prospect trench for nitrate deposits during World War I. West of California 127, 4.5 miles south of Shoshone. —Helen Z. Knudsen photo



canic ash. Nitrates, used in explosives, became a critical strategic mat rial for the United States during the First World War, because Chile ha been our principal prewar supplier, and German U-boat activity cou disrupt the nitrate supply line. Extensive searches for nitrate deposi were made in the arid Southwest, without much success. Death Valle including Lake Tecopa, was one of the regions explored. West of Califo nia 127 within the first few miles south from Shoshone, you can still s faint traces of trenches hand-dug in 1915 on some spurs and steep slop in the lakebeds. Don't confuse them with younger fresh trenches or bac hoe and bulldozer workings.

Of great interest are twelve layers of volcanic ash within the lake d posits, three of which are particularly thick and continuous. These thr layers and their sources and ages are: Lava Creek ash from Yellowstor 620,000 years old; Bishop ash from Long Valley, 30 miles northwest Bishop, California, 760,000 years old (vignette 23); and Huckleberry Rid ash from Yellowstone, a little over 2 million years old. Geologists est lish the source for each ash by its distinctive chemical and mineral cal composition and by analyzing its trace-element content, a proc known as chemical fingerprinting. Ages are determined by measur radioactive elements (such as potassium) and their decay produ (such as argon) in rocks at the eruptive center that produced the a Each of these three eruptions created a huge caldera and an ash umn that rose so high into the atmosphere that the ashes are wid distributed.

Ashes help bracket the age of Lake Tecopa. Since the Lava Creek ash is below the top of the uppermost lakebeds, and the Huckleberry Ridge ash is above the bottom of the deposits, the lake must have existed from more than 2 million years ago to less than 620,000 years ago. Using rates at which lakebeds accumulated between the three ash dates and the measured thickness of beds above and below the two Yellowstone ashes, a simple calculation suggests that Lake Tecopa lasted until about 500,000 years ago and came into existence at least 3 million years ago. It could be still older, as the lowest beds are not exposed. These dates are consistent with paleomagnetic measurements on lakebed samples, which record the polarity of the earth's magnetic field. A chart of reversals of



Location of the Yellowstone Lava Creek ash exposures.

magnetic polarity over the last 2.5 million years, established by worldwide measurements, fits well with the Lake Tecopa paleomagnetic data and helps establish the validity of the local record. The Lake Tecopa beds seen here are much older than the glacial-age Lake Manix and Lake Manly (vignette 5) deposits.

Travelers on California 127 and 178 can easily inspect exposures of the Lava Creek ash by making a small detour. Opposite the junction of these highways at the south edge of the town of Shoshone, near a large road sign pointing north to Shoshone and Death Valley and south to Baker, a gravel road extends westward. In 100 feet it forks. Continue straight ahead on the left fork for 0.25 mile to a vertical cliff in which there are caves. Homeless people excavated and lived in these caves during the Great Depression of the 1930s. The basal 6 feet of this cliff is Lava Creek ash, pure white beneath the stain and mud washed down from overlying beds. The lower part of the ash layer is relatively massive, but the upper part is thinly bedded, which indicates reworking by surface runoff. A thickness of 6 feet is abnormal for an air-fall ash bed deposited 650 miles from its source. Much of this ash accumulation probably washed into the lake from surrounding slopes. It grades upward into typical Lake Tecopa mudstone layers.

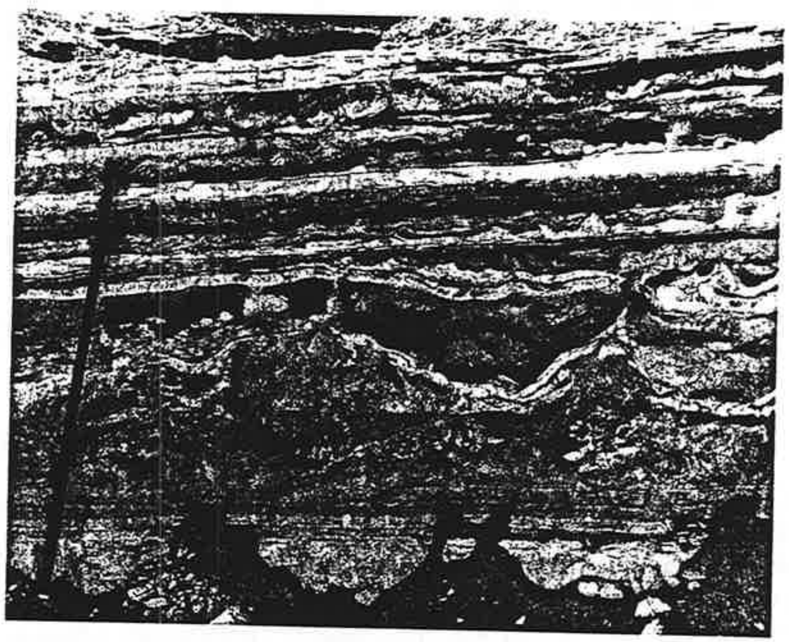
Lava Creek ash consists mainly of tiny particles of clear volcanic glass. You can see them easily under a ten-power magnifier, and even to the naked eye they glisten in the sun. Most Lava Creek ash blew eastward because of prevailing westerly winds. Deposits are extensive in Kansas


A doorway to a rough home cut into a 6-foot layer of largely reworked, 620,000-year-old Lava Creek ash from Yellowstone in Lake Tecopa beds near Shoshone. —Helen Z. Knudsen photo

are known as far east as Mississippi, and were recently recorded in surficial materials near Washington, D.C. Considering this easterly drift, how did Lava Creek ash end up as much as 700 miles southwest in southern California, where it exists in many desert lake deposits and in marine beds at Ventura? Some people speculate that the ash column from the Lava Creek caldera rose so high it got caught in the high-altitude jet stream, which is notoriously erratic and characterized by large oscillations that probably whipped some ash southwestward.

To see more ash, proceed a bit farther on this road to an open pit. If possible, avoid stepping into wind-drifted deposits of fine ash near the bottom of the pit walls—you can sink in over your shoe tops. Also be careful about rubbing your eyes after handling chunks of ash. The small glass particles may adhere to your skin and are most uncomfortable in your eyes. Avoid the pit in strong wind because of blowing ash.

A commercial operation occasionally mines the pit for ash, which is used primarily as an absorbing medium, for example, to mop up oil spills on airport runways. Fresh exposures in the pit walls show many thin layers that in places are complexly convoluted, possibly by slumping or more likely earthquake jiggling of soupy lake-floor deposits. You will seldom, if ever, see better exposures of volcanic ash than here.



Contorted layers of Lava Creek ash, jiggled by sliding or seismic shaking while in a soupy state, lying between undeformed horizontal beds of ash. Exposed in a quarry near Shoshone. Staff is 54 inches long. —Helen Z. Knudsen photo

Geologists prize reliably dated ash layers whose source can be ider tified. The beds are useful time-marker horizons. The technique of ider tifying and dating such layers is dignified by the impressive terr tephrochronology. Tephra is volcanic debris, blown into the air by a explosion, that settles back to the ground directly out of the atmospher Chronology refers to time.

Lake Tecopa deposits harbor much information about events an conditions in the greater Death Valley region and certainly have not ye revealed all their secrets. By dissecting the lakebeds, the Amargosa Rive exposes the story of the Tecopa basin for all to appeciate.

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Kingston Range detachment fault, southeastern Death Valley region, California: Relation to Tertiary deposits and reconstruction of initial dip

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ABSTRACT

The Kingston Range detachment fault, the northern part of the Kingston Range–Halloran Hills detachment fault system, dips gently southwest and places highly extended Middle and Late Proterozoic Pahrump Group, Late Proterozoic and Cambrian miogeoclinal deposits, and middle Miocene Resting Springs Formation over relatively unextended Early Proterozoic gneiss, folded miogeoclinal deposits, and Resting Springs Formation. Rocks in the upper plate of the detachment fault are cut by numerous northwest-trending planar and listric normal faults and northeast-trending tear faults, and were transported as much as 6 km to the southwest. South of the Kingston Range, a correlative detachment fault cuts a 13.4-Ma hypabyssal sill. In the Kingston Range, the detachment faults cuts 16.0-Ma clastic and lacustrine carbonate rocks of the lower Resting Springs Formation and unconformably overlying <13.3- to 12.5(?)-Ma volcanic and sedimentary rocks of the upper Resting Springs Formation. Upper plate faults and, possibly, the detachment fault itself, are cut by the 12.4-Ma granite of Kingston Peak. These relations indicate that the Kingston Range detachment fault formed between 13.4 and 12.4 Ma.

Syn- and pretectonic strata in the upper and lower plates of the Kingston Range detachment fault now dip 32° ± 3°NE, indicating that the fault has been tilted. Limited paleomagnetic data suggest that the granite of Kingston Peak is tilted 15°NE. When middle Miocene and disconformably underlying rocks are restored to their original orientation, the fault/bedding cutoff angle with upper and lower plate rocks indicates that the average initial fault dip was 38° ± 9°SW between the earth's surface and a depth of 4 km. Tilting of the detachment fault entailed rotation of a structural block consisting of the Kingston Range and Mesquite Mountains. Tilting occurred after, and probably also during, detachment fault displacement. We suggest that tilting resulted from flexural isostatic uplift of the detachment fault lower-plate, with a possible contribution from rotation above a structurally-deeper detachment fault.

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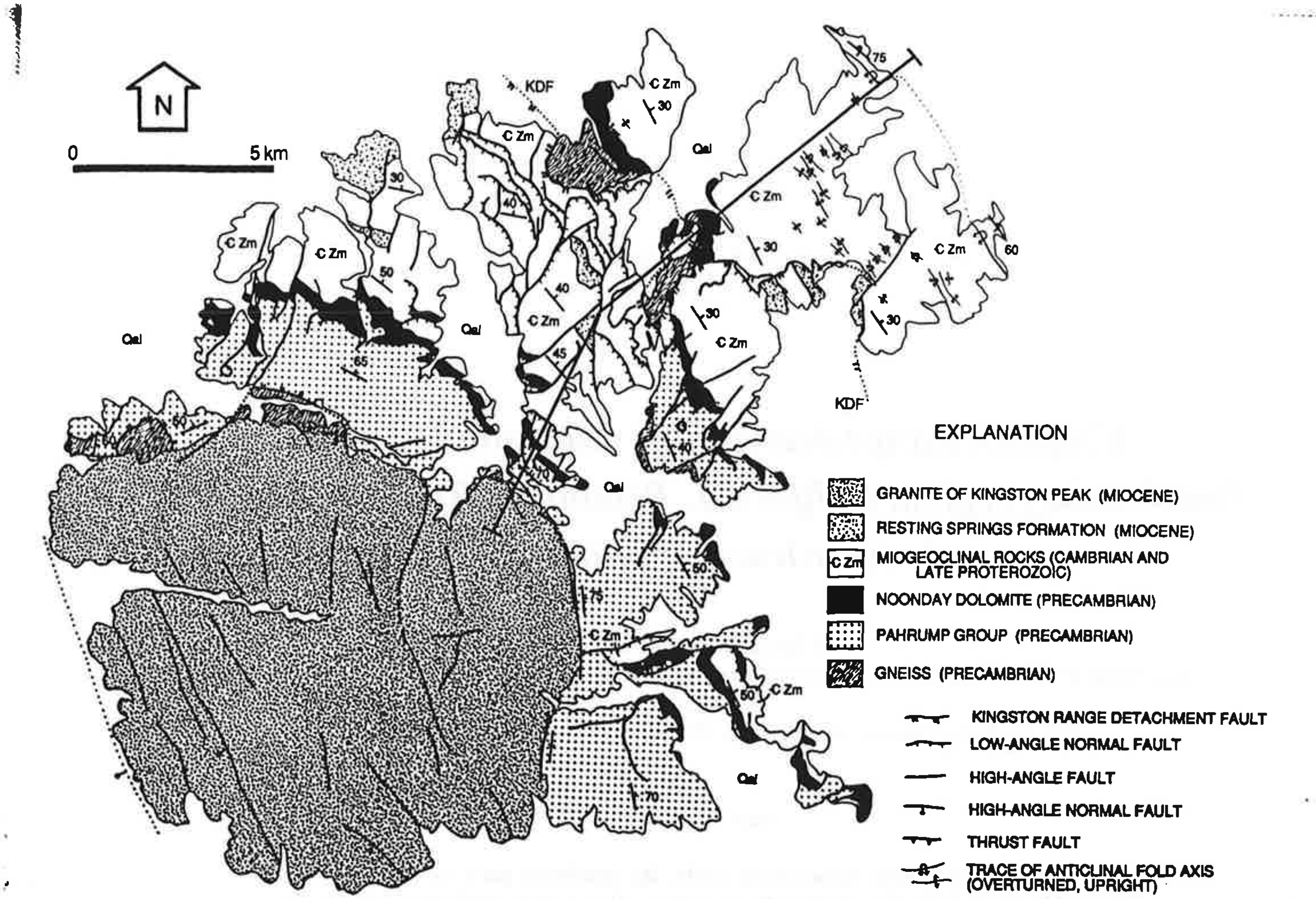


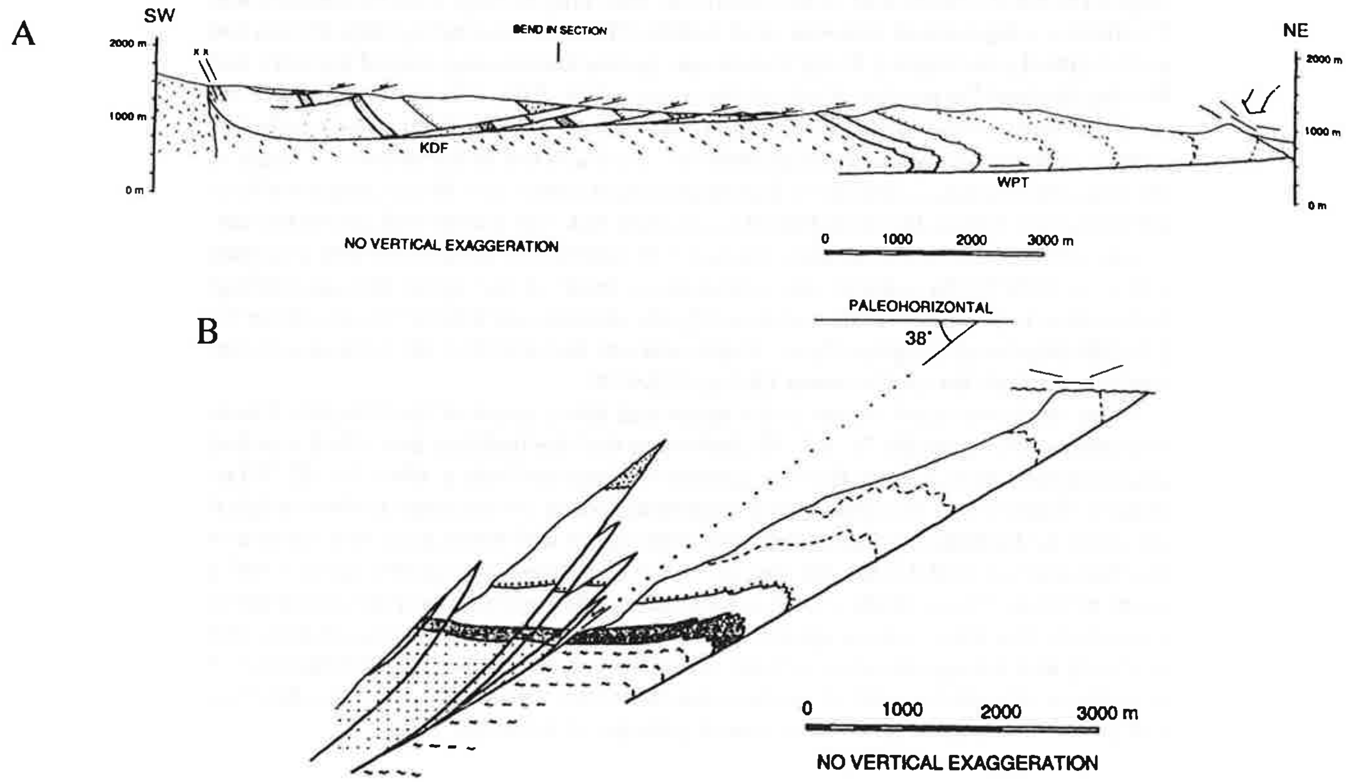


Figure 4A, Simplified geologic cross section of the northeastern Kingston Range. Symbols same as in Figure 3. Representative dips of lower plate Tertiary strata have been projected approximately 500 m onto the line of section (indicated below open arrow). Solid line with dots above is marker horizon in the Johnnie Formation. Dashed lines in lower plate are form lines showing geometry of folds. WPT = Winters Pass thrust fault; KDF = Kingston Range detachment fault. 4B, Structural reconstruction of northeastern Kingston Range immediately prior to inception of Miocene detachment faulting. See text for constraints.

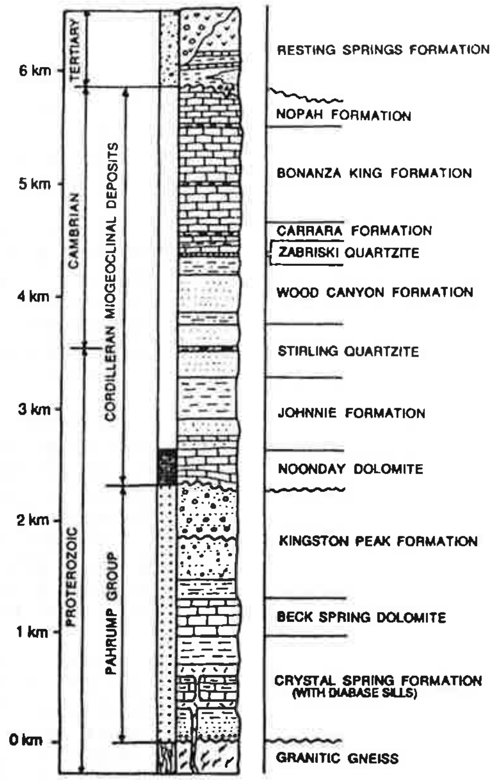


Figure 2. Simplified stratigraphic column of the Kingston Range (Calzia, 1990).

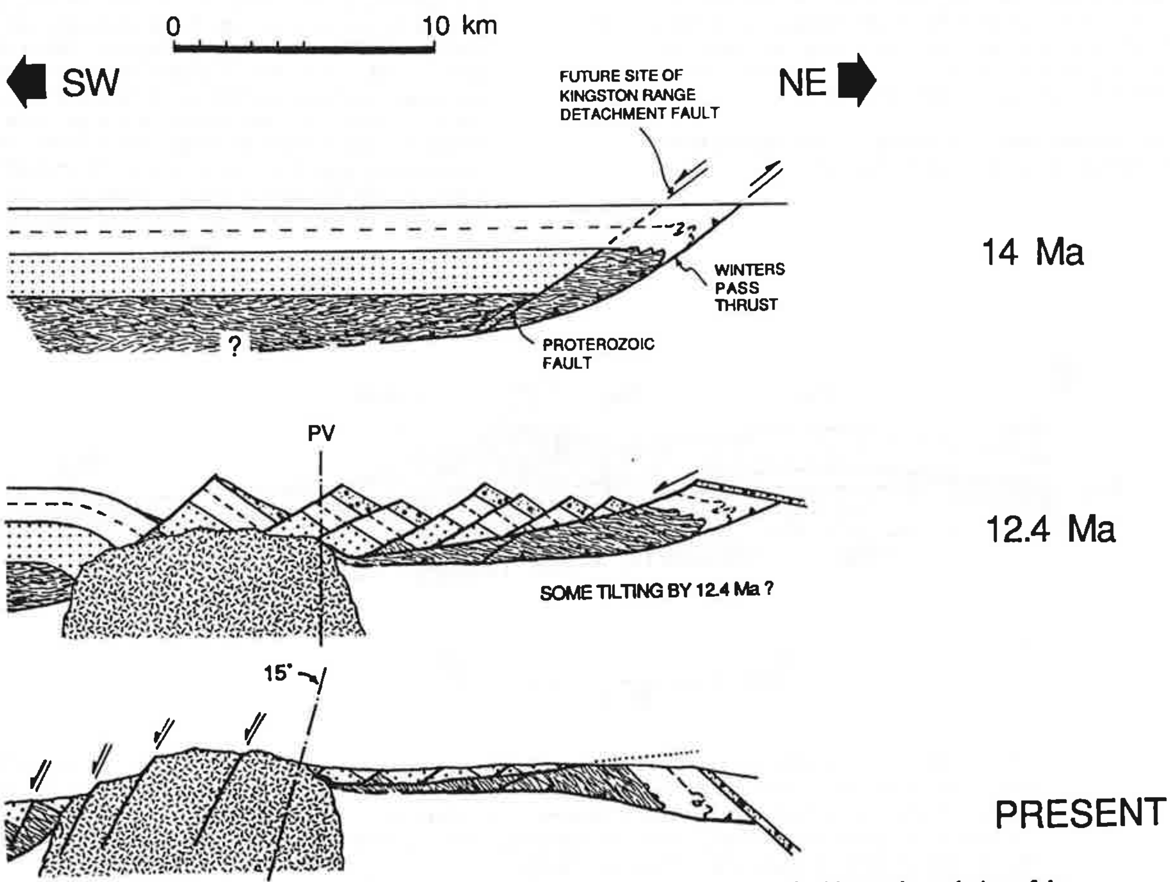


Figure 8. Cartoon cross section, approximately to scale, showing model for kinematic evolution of the Kingston Range detachment fault. Unit patterns are the same as in Figure 3. Actual Pahrump/miogeocline relations are more complex than indicated in the figure. PV = paleovertical from paleomagnetic data of Jones (1983).

Third Day

<u>Introduction</u>

Return to the northern part of the Clark Mountain thrust complex, north of Interstate 15. Having established the Mesozoic tectonic framework of the complex yesterday, we now focus on Cenozoic extensional modifications of it. We have recognized four examples of such modifications:

(1) extensional faulting in the west-central Mescal Range (stop 2-7 and Fig. 13);

(2) normal faulting, down to the west, within the base of the Keaney/Mollusk Mine thrust plate from the powerline road north of Clark Mountains (stop 3-3 and Fig. 9) to the northern Mescal Range;

(3) low-angle detachment faulting near the base of the same plate near Mesquite Pass, 3-4 miles north of stop 3-2 (Fig. 9); Mesquite Pass and Winters Pass plate rocks above this detachment have been extended along southeast-dipping normal faults during an episode of probable Cenozoic extension unrelated to (2) above; and

(4) major low-angle detachment faulting in the eastern Kingston Range and at structurally high levels of the Mesquite Mountains (between the Kingston Range and Clark Mountains); this domain of southwest-directed extension (relative to lower-plate rocks) constitutes the southeastern breakaway margin of a major, probably composite, late Tertiary extensional province that extends as far to the west as the Sierra Nevada (Fig. 3).

Of the four examples of Cenozoic extension, only (4) appears at this time to have regional implications. The trace of a major west-dipping, low-angle normal fault, the Kingston Range detachment, is present in the northeastern Kingston Range (Burchfiel, Hodges, and Walker, in prep.). Most of the range consists of complexly faulted, east- and northeast-tilted Precambrian, Cambrian, and upper Cenozoic strata (Figs. 14, 15). The Kingston Range detachment (Fig. 3) separates a region to the east and south, including the Mesquite and Clark Mountains, that has been little affected by Cenozic extension, from a region (as far to the west as the Sierra Nevada) that has been strongly affected by such extension. For this reason, the Kingston Range detachment forms the eastern breakaway zone for the extended regions to the west (Burchfiel and others, 1983). South of the Kingston Range, the continuous trace of the Kingston Range detachment is largely covered by young alluvium, but it probably lies just west of the Mesquite Range close to Cima Road. Isolated blocks of upper Precambrian, Cambrian, and

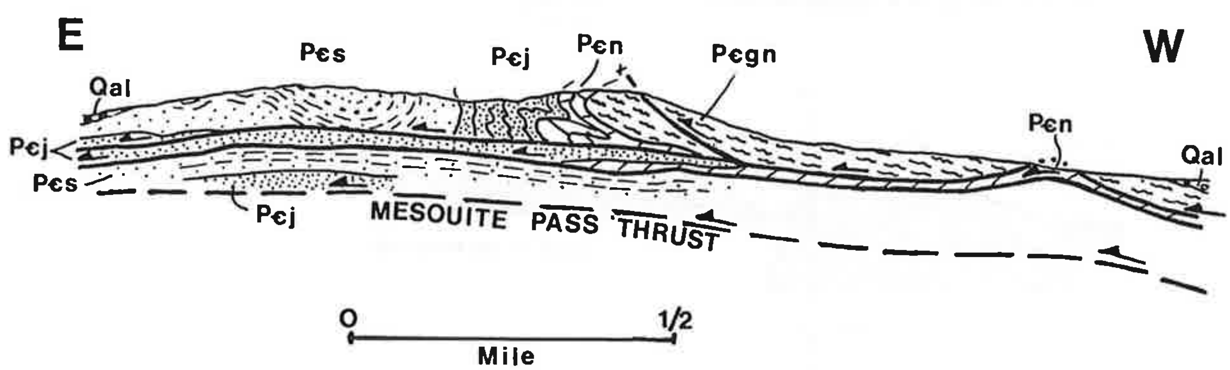
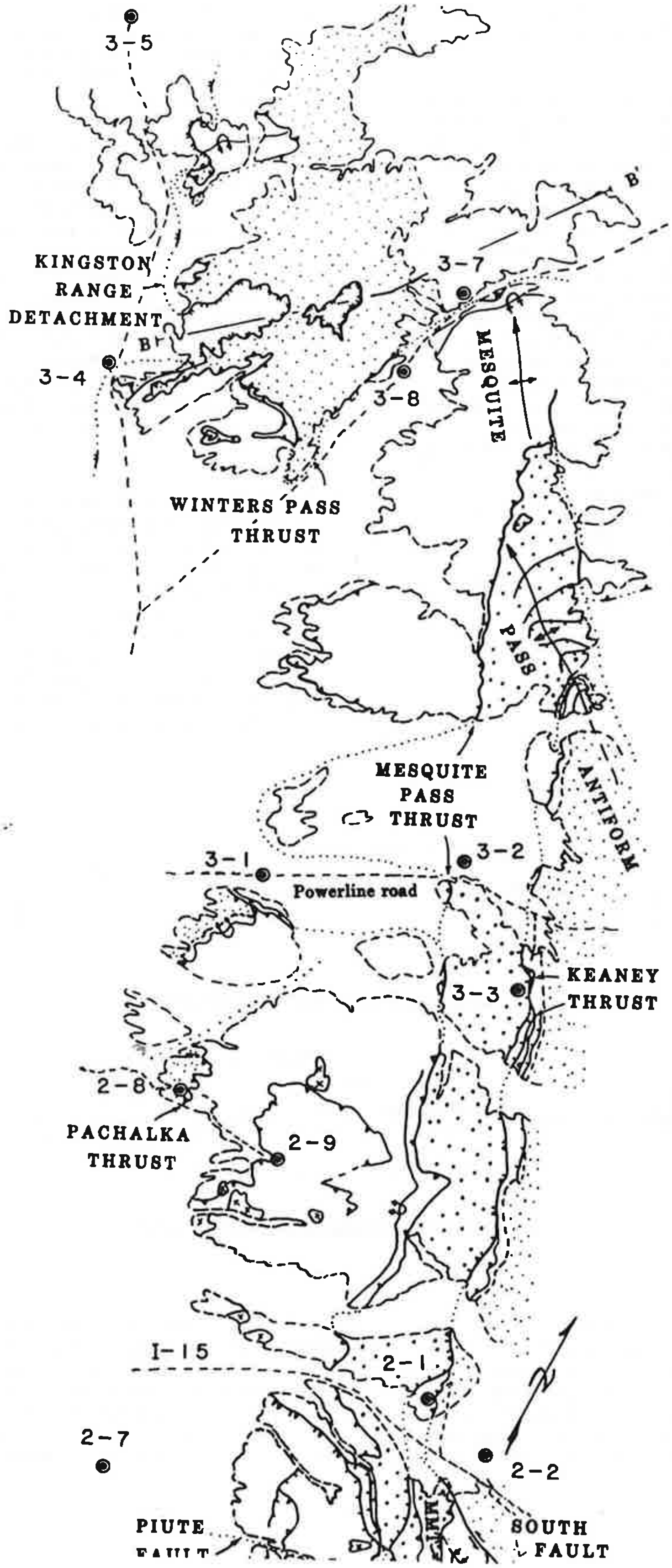


Figure 14. Cross-section through the Mesquite Pass thrust plate, hill south of powerline road (cf. Fig. 9), illustrating involvement of basement crystalline rocks in the thrust belt. Geologic relationships are explained at stop 3-1. Rock units (oldest first): Precambrian gneisses (P€gn); Noonday Dolomite (P€n); Johnnie Formation (P€j); Sterling Quartzite (P€s); Quaternary alluvium (Qal).


3-5
B'
KINGSTON
RANGE
DETACHMENT
3-7
B
3-4
3-8
MESQUITE
WINTERS PASS
THRUST
PASS
MESQUITE
PASS
THRUST
ANTIFORM
3-1
3-2
Powerline road
KEANEY
THRUST
3-3
2-8
PACHALKA
THRUST
2-9
I-15
2-1
2-7
2-2
PIUTE
SOUTH
FAULT

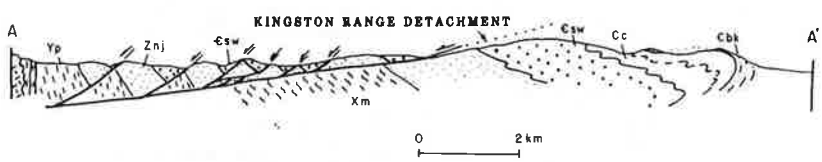
Cenozoic rocks that rest on the detachment fault in the western Mesquite Mountains are erosional klippen of its hanging wall (stop 3-4, Fig. 15). The detachment fault is locally exposed beneath these klippen and dips only an average of 3 degrees to the southwest, the direction of relative upper-plate displacement along it.

Upper Precambrian and Cambrian rocks are folded along northwest-trending axes in the footwall of the Kingston Range detachment (Figs. 15, 16). The folds are overturned to the northeast and are well exposed in the eastern foothills of the Kingston Range. These folds are on strike with similar folds in the Mesquite Mountains to the southeast that are in the hanging wall of the Mesozoic Winters Pass thrust. We suggest that these folds formed during an early episode of northeast-directed thrusting along the Winters Pass thrust. Because the older rocks were folded by Mesozoic deformation it is difficult to assess how much of their tilting within the upper plate of the Kingston Range detachment occurred during Cenozoic rotation. Upper Cenozoic strata rest unconformably on folded Cambrian rocks at one locality on the northeast slope of the Kingston Range in the footwall of the detachment. They dip 10 to 20 degrees northeastward, suggesting some Cenozoic rotation of the footwall rocks. It is not clear, however, whether this rotation is related to detachment faulting or to warping along the northeastern flank of the Mesquite Pass antiform (Fig. 9).

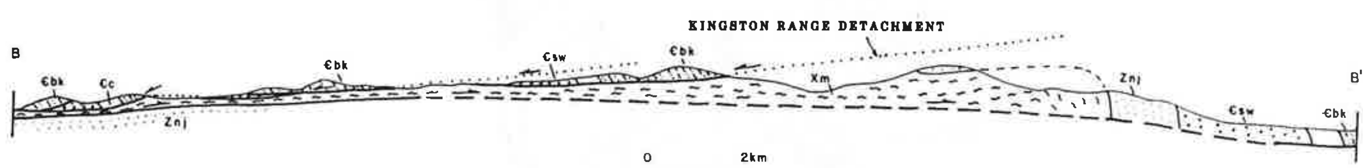
Rotation of upper-plate rocks above the Kingston Range detachment is clearly recorded by tilted upper Cenozoic strata. Conglomerate, sandstone, lacustrine limestone, and volcanic rocks dip 30 to 35 degrees northeastward into the detachment fault in the easternmost part of the hanging wall. In some fault blocks in the central part of the Kingston Range they dip vertically into Precambrian crystalline rocks of the lower plate.

The geometry of upper-plate faulting is very complex and consists of numerous, closely spaced, southwest-dipping, planar and listric normal faults and associated northeast-striking tear or transfer faults. Some of the normal faults have clearly been rotated into shallower dips. Many of the faults are strongly curved in plan view and, presumably, in cross section. This suggests that some of the hanging wall faults are spoon-shaped. Extensional duplexes can be seen locally in erosional windows. The matching of hanging-wall to footwall cutoffs of the northeast-dipping Noonday Dolomite indicates that the easternmost and lowest fault of the detachment complex has about 3 to 4 km of displacement (stop 3-6).

The Noonday Dolomite rests unconformably on crystalline metamorphic and igneous rocks in the footwall of the Kingston Range detachment with only thin (a few tens of meters), intervening local deposits that may be equivalent to the upper Precambrian Pahrump Group. However, the hanging wall contains a thick sequence (several kilometers) of Pahrump Group rocks below the Noonday Dolomite. The dolomite rests unconformably on rocks of the Pahrump Group and progressively rests toward the north on older Pahrump units until it lies across crystalline



NORTHEAST KINGSTON RANGE



NORTHWEST MESQUITE MOUNTAINS

Figure 15. Cross-sections through the northeast Kingston Range and the northwest Mesquite Mountains that illustrate the shallow dip and upper-plate structure of the Kingston Range detachment fault. The locations of sections AA' and BB' are shown on Figures 16 and 9 respectively. Unit designations not explained for previous figures are: Xm = Precambrian metamorphic rocks; Yp = Precambrian Pahrump Group; Znj = Precambrian Noonday Dolomite and Johnnie Formation. The patterned unit at A, section AA', is the Miocene Kingston pluton. The heavy dashed line near the bottom of section BB' is the Mesozoic Winters Pass thrust fault.

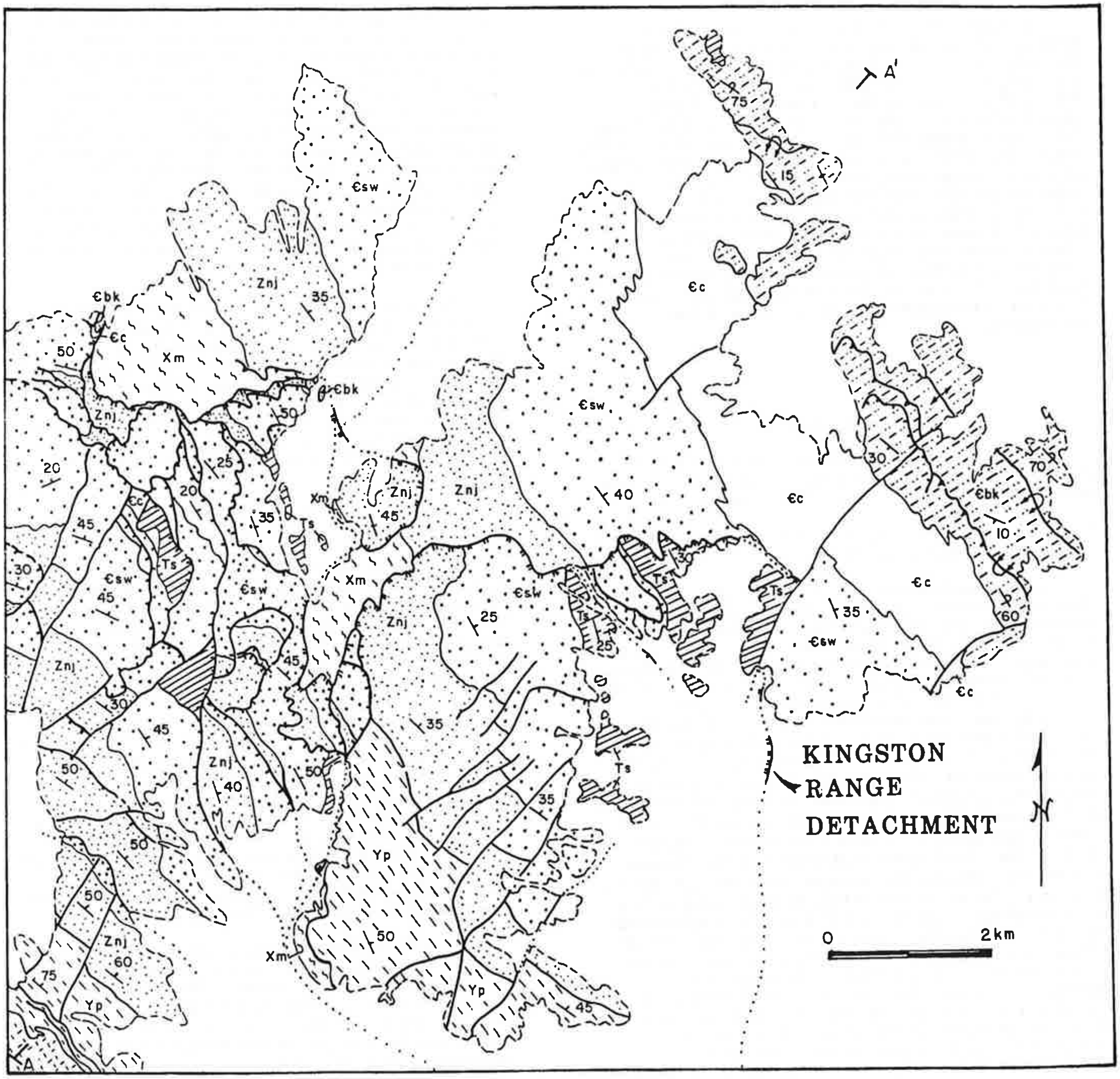


Figure 16. Simplified geologic map of the northeastern Kingston Range (cf. Fig. 4) showing location of the Kingston Range detachment fault and its upper-plate structure. The unlabeled patterned unit at the southwestern corner of the figure is the Miocene pluton. Abbreviated designations for rock units have been explained in previous figures.

basement rocks in the northern Kingston Range. These relations indicate that the Pahrump Group was faulted and tilted prior to deposition of the Noonday Dolomite. It also suggests that the Kingston Range detachment fault either reactivated or followed closely the old eastern boundary fault of the Pahrump Group.

Volcanic rocks from two localities in the hanging wall of the detachment have been dated at 12.5 and 12.1 Ma (J. Spencer, person communication to BCB, 1983; P. G. Tilke, personal communication to BCB, 1985). These volcanic rocks were clearly deposited prior to the onset of detachment faulting. The large Miocene Kingston Range pluton (12 to 13 Ma, Armstrong, 1970) is intruded into Precambrian strata of the Kingston Range, but whether the pluton is older or younger than the detachment faulting is not yet known. Paleomagnetic data indicate that the pluton has been rotated (Jones, 1983), but it is not clear whether this rotation is the result of detachment faulting or post-detachment strike-slip faulting. At present there is no good upper age limit to displacements on the Kingston Range detachment.

Fieldtrip Stops

Return to Cima Road and drive north along it from Interstate 15. Turn east on the powerline road approximately 9 miles north of the freeway.

Stop 3-1 (optional) Style of deformation in Mesquite Pass thrust plate (Fig. 14)

In the westernmost hill south of the powerline road are exposures of several thrust slices that characterize the structural style in the lower part of the Mesquite Pass thrust plate (Fig. 14). At this view stop, look south toward the hill and the three thrust-bounded slices within it. At the base of the hill is a tectonic slice containing, in normal sequence, the upper Precambrian Johnnie Formation and the middle and lower parts of the Stirling Quartzite. An important feature of this lowermost slice is its lack of internal folding. Above it is a slice of very thin, essentially unfolded Noonday Dolomite seen as the prominent brown ledge near the middle of the slope. Eastward, this second slice also contains Johnnie Formation rocks separated from the Noonday by a minor(?) thrust; the Noonday rocks are cut out to the east. Rocks in the third, highest slice belong to a complete sequence from Precambrian crystalline rocks to the Zabriskie Quartzite. The style of this slice is very different from the other two, because its rocks are highly folded. The folds are overturned to the east, exhibit a well-developed axial plane cleavage, and are truncated by the thrust fault at the base of the slice. Precambrian crystalline rocks appear in the cores of the westernmost folds and, like the sedimentary rocks above them, also possess an axial plane cleavage. The Noonday Dolomite, which directly overlies the basement rocks, has been squeezed into the cores of folds and is tectonically thinned or missing along their flanks farther to the south.

Stop 3-2 Contact between the Mesquite Pass and Keaney thrust plates

Stop and look northward from the intersection of the powerline road and the Mesquite Pass road leading north. The contact between the Mesquite Pass and Keaney thrust plates is clearly evident in the southern Mesquite Mountains across the broad open valley. Brownish, upper Precambrian clastic rocks in the Mesquite Pass thrust plate (Stirling Quartzite, Johnnie Formation) form the large, dark hill to the west of a small valley along the Mesquite Pass thrust. The well-bedded black, gray, and white carbonate rocks east of the valley lie in the Keaney thrust plate and include the Bonanza King, Sultan, and Monte Cristo formations. Northeast-striking, southeast-dipping, high-angle faults in this plate lie in the hanging wall of a Cenozoic detachment fault. The low-angle fault is located just below the Keaney thrust in Bright Angel shales and carbonates, but it cannot be seen from this stop.

Continue driving east on the powerline road. Turn right, approximately one mile past stop 3-2, onto a dirt road that leads southward. This road lies variably within and below the basal portion of the Keaney thrust plate. Drive through a complicated footwall zone of Mesozoic parautochthonous thrust slices in Tapeats Sandstone and Bright Angel Shale before stopping near the base of the plate approximately 1.35 miles south of the powerline road.

Stop 3-3 Low-angle Cenozoic normal faults, basal Keaney plate

The Keaney thrust fault juxtaposes Cambrian carbonate rocks over older units (Precambrian basement, Tapeats Sandstone, Bright Angel Shale) from the Mesquite Pass area (cf. stop 3-2) southward to Mohawk Hill (stop 2-1). We believe that this is the consequence of the uplift and erosion of a Cambrian-Jurassic cratonal section from the northern wall of the South fault (stop 2-2), prior to emplacement of the Keaney/Mollusk Mine thrust plate across the lowest part of this section; the complete cratonal section is still preserved south of that fault (stop 2-3). Sharp (1984) was the first to recognize that the thrust zone at the base of the Keaney/ Mollusk Mine plate has been overprinted by down-to-the-west normal faulting of Cenozoic age. His studies in the Colosseum Mine area, 1 mile south of this stop, established that a zoned, mineralized aureole around a Cretaceous granitic stock had been downdropped approximately 1500 feet.

At this stop we examine a shallow-dipping zone of brittle faults that is presumably of Cenozoic age and normal fault displacement. The fault zone is exposed in several prospect pits and small mine workings above and west of the dirt road. Copper mineralization typical of the normal fault zone (malachite, azurite) is seen in these workings, as is the characteristic presence of sheared fluorite. Just below one of the prospect pits we will contrast the brittle Cenozoic deformation with an exposure of the Keaney thrust. This sharp, foliated thrust contact separates upper-plate Bonanza King carbonates from lower-plate Bright Angel(?) carbonates. The foliated nature of this contact, in contrast to the brittle fault seen at stop 2-1, may indicate that lower-plate rocks in this area had a considerably higher temperature due to nearby igneous intrusion (Colosseum Mine area) than in areas farther south. Our remapping of the Keaney thrust in areas to the south indicate that the base of the Keaney plate is now locally defined by a shallow-dipping (<35°) normal fault or faults. This relationship is perhaps best seen in the bulldozed pit of the Pacific Fluorite Mine west of the Colosseum Mine.

Return to Cima Road and turn right (north) along it. Stop approximately 3 miles north of the northeast-trending side road to Winters Pass.

Stop 3-4 (optional) Complexity of slicing below Winters Pass thrust plate

At this stop we will walk through upper Stirling Quartzite in the Mesquite Pass thrust plate, and climb upward across several thin tectonic slices of Noonday and Stirling rocks below the Winters Pass plate. Mylonitic gneisses at the base of the plate have the shallow dipping mylonitic foliation and southwest-plunging stretching lineation seen at stops 2-6 and 2-8. The black layers in some of the mylonitic gneisses at the top of the hill are not mylonitic in origin, but are composed of magnetite.

To the north are klippen in the northeast-southwest-trending train of klippen above the shallow (3°) southwest-dipping Kingston Range detachment (Fig. 15). Cambrian strata dip eastward into the detachment surface and are repeated again and again by southwest-dipping upper-plate normal faults (one of which can be pointed out from this locality). Massive Bonanza King rocks overlie Carrara Formation in the nearest klippe. The major dark gray ridge west of Cima Road, surrounded by alluvium, is composed largely of tilted Bonanza King carbonates, but a thick, tilted Tertiary section of sedimentary and volcanic rocks lies concordantly above it.

It is clear from this stop that the Kingston Range detachment fault developed across crystalline rocks in the Winters Pass plate in this area. It did not reactivate the Winters Pass thrust fault which lies only a hundred meters or so below the detachment fault in this area. The close parallelism of the Cenozoic detachment fault and the subhorizontal older thrust fault throughout the northern Mesquite Mountains should dispel arguments by some geologists that low-angle detachment faults cannot have a primary, shallow-dipping geometry. The 2 to 3 degrees southwest dip of the Kingston Range detachment fault in this area, across a distance of at least 15 km, must be very close to the original dip of the detachment fault in this, its breakaway area.

Continue driving north on Cima Road. Turn right (eastward) onto a dirt road not far south of the white talc tailings in the Kingston Range. Stop 3-5 and 3-6 are along this road.

Stops 3-5, 3-6 View stops of Kingston Range detachment fault relations as seen from the road between the eastern Kingston Range and the northern Mesquite Mountains (Figs. 15, 16).

Upon reaching Mesquite Valley, turn right (south) and drive to the well-graded road between the settlement of Sandy, in the valley, and Winters Pass in the Mesquite Mountains. Turn right (west) and drive into Winters Pass.

Stop 3-7 Winters Pass thrust fault, Winters Pass area

The Winters Pass road follows the trace, mostly concealed, of the Winters Pass thrust fault. The northeast strike of the fault here is due to its position on the northwest-plunging nose of a large Cenozoic(?) antiform (Mesquite Pass antiform) that warps all three major thrust plates in the area. The Winters Pass plate northwest of the road consists of a well-exposed, northeast-dipping, miogeoclinal section approximately 3200 m in thickness. This section extends from crystalline basement into the Cambrian Bonanza King Formation. The basal unit of the section, the Noonday Dolomite, rests unconformably on Precambrian gneiss and granitic rocks. Locally, a thin (several m) basal conglomerate is developed on the irregular erosion surface beneath the Noonday. The contact is only locally and mildly deformed, in marked contrast to the highly deformed contact discussed at stop 3-1. A thick sequence of generally brownish clastic rocks overlies the Noonday Dolomite (including Johnnie, Stirling, Wood Canyon, Zabriskie, and lower Carrara formations). The thin Zabriskie Ouatzite is strongly small folded. Several ledges of thick grey limestone, some containing numerous algal structures (Gervinella sp.) are interbedded in the lower Carrara Formation with greenish phyllitic shales. We have seen one such limestone (highly deformed) in the overturned syncline below the Pachalka thrust plate (stop 2-8).

Southeast of the road in Winters Pass is a small exposure of the Precambrian and Cambrian clastic sequence that lies above the Winters Pass thrust; this is the only segment of the thrust in Winters Pass that is not buried beneath alluvium. If time permits, we will walk to an exposure of the foliated thrust contact. The footwall of the thrust here is a section of overturned Bonanza King Formation. The hinge of the overturned synclinal fold trends northeast-southwest, parallel to the trace of the thrust fault. Several thin, tectonic slices of orange-weathering silty carbonates of the upper Carrara Formation lie exposed below the Winters Pass thrust, but the Carrara beds are not in stratigraphic continuity with the overturned Bonanza King carbonate rocks.

Stop 3-8 (optional) Kingston Range detachment fault and klippe

Time permitting, we will drive southeastward on the Winters Pass road to Winters Pass at the summit of the small grade. From here we can look back down along the obvious trace of the Winters Pass thrust. The prominent isolated gray hill on the skyline to the northwest is largely underlain by Bonanza King carbonate strata dipping 20 to 40 degrees northeastward. The hill is the most northeasterly of the klippen above the Kingston Range detachment fault, here seen as the planar, shallow-dipping (2 to 3 degrees) contact between upper-plate carbonate rocks and Precambrian gneisses which form the low rolling terrane surrounding the hill.

Reconstruction of the Mojave Block

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INTRODUCTION

The Mojave Desert region is one of the geologically most diverse and complex regions of the southwestern North American Cordillera. We define the Mojave Desert as that area between the Garlock fault to the north, the San Andreas fault to the southwest, and the Colorado River to the east (Fig. 1). The Mojave block is that part of the Mojave Desert that lies west of the southern extension of the Death Valley fault zone (Glazner, 1990). This definition is primarily tectonic in scope.

This area forms the southwesternmost extent of Precambrian continental North America and lies at the present plate edge formed by the San Andreas fault. Because the western part of North America has faced an oceanic plate since late Precambrian time (Burchfiel and Davis, 1972), the Mojave Desert and environs have seen active tectonism over much of the Phanerozoic. This leading edge position has resulted in superposition of most continental tectonic styles and regimes in the Mojave Desert over the last 600 million years.

It has been a long-term goal of the authors to decipher this history of superposition. The tectonic history of the Mojave Desert divides rather nicely into Cenozoic events overprinting Paleozoic and Mesozoic tectonism. Unraveling this story necessarily involves using both young and old markers to understand the strain significance of various deformational events. We hope, in our process of reconstruction, that this is done in an iterative rather than circular manner. Below we discuss important tectonic, magmatic, and depositional events that have affected the Mojave Desert. This is followed by a road log for a field trip that works backward in time, undoing successive tectonic events.

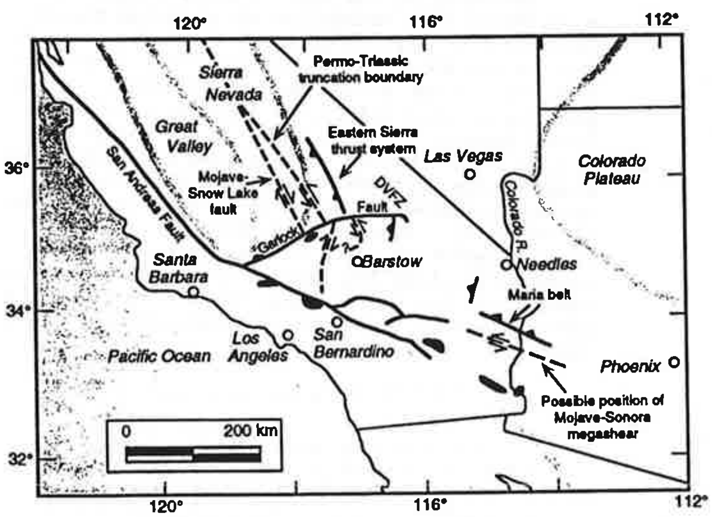

Fig. 1. Location of the Mojave block. Also shown are several important faults that are either observed (solid lines) or inferred (dashed lines). Black blobs are outcrops of Pelona, Rand, and Orocopia schists.

SUMMARY OF GEOLOGIC HISTORY

Geologic Setting

Rocks of Precambrian to late Cenozoic age are exposed across the Mojave Desert (Fig. 2). Precambrian basement rocks are generally either ~1700 Ma or ~1400 Ma (Anderson et al., in press), and are widely distributed across the Mojave Desert (Martin and Walker, 1992; Anderson et al., in press). Precambrian rocks of other ages, such as ca. 1.1 Ga diabase sills, are present as well, but are of relatively minor exposure. Upper Precambrian to Lower Permian miogeoclinal to cratonal strata overlie the basement rocks (Fig. 2; Burchfiel and Davis, 1972, 1975; Martin and Walker, 1992; Stewart and Poole, 1975) The miogeoclinal/cratonal strata generally have southwest- to south-trending facies patterns. Deposits related to the Devonian-Mississippian Antler Orogeny are not present in any of the Mojave miogeoclinal/cratonal sections (Walker, 1988).

Paleozoic rocks of eugeoclinal affinity (e.g., continental slope and rise deposits) are present in the northern Mojave Desert from the Alvord Mountain area to Pilot Knob Valley, and across the Garlock Fault in the El Paso Mountains (Fig. 2; Burchfiel and Davis, 1975). There is a facies mismatch, therefore, across the central Mojave Desert, with eugeoclinal rocks to the north and west and miogeoclinal/cratonal rocks to the south and east. As we will see below, the region of mismatch has been a site of repeated tectonic activity.

Mesozoic strata consist of Early Triassic marine sedimentary rocks and Middle(?) Triassic to Late Jurassic volcanic and epiclastic strata (Schermer, 1993; Walker, 1987, 1988). In addition, the Jurassic-Cretaceous magmatic arc crosses the Mojave Desert (Fig. 2; Barton et al., 1988; Kistler, 1974). Facies and magmatic patterns in Mesozoic strata generally follow northwest trends across the Mojave Desert region (Burchfiel and Davis, 1981; Hamilton, 1969; Hamilton and Myers, 1966; Walker, 1987, 1988), indicating that the continental margin had a northwest to north trend during that time. Orientation of the continental margin probably changed in Late Paleozoic time due to truncation of the continental margin accompanied by late-stage contractional deformation and igneous activity (Burchfiel and Davis, 1981; Davis et al., 1978; Walker, 1988).

Paleogene rocks are not known from the Mojave Desert, whereas Neogene to Recent rocks are widely exposed. Most Neogene sections include earliest Miocene (or possibly latest Oligocene) volcanic rocks overlain by lower to middle Miocene sedimentary, volcanic and epiclastic sections. These rocks were deposited in basins formed by extensional faulting in and around areas of metamorphic core complex development. Extensional activity began at about 23 Ma from areas around the Colorado River into the central Mojave Desert (Dokka,

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(Carr et al., 1984).

Late Permian deformation and magmatism are recorded in ranges in the western Mojave Desert, with the best-understood ages known from El Paso Mountains and the Victorville area. In the El Paso Mountains, Upper Permian volcanic and volcaniclastic rocks as young as ~260 Ma are deformed by west-vergent folds and faults and cut by plutons as old as 247 Ma (Carr et al., 1984; Walker, unpublished; Miller et al., in review). In the Victorville area, rocks as young as Mississippian are deformed and intruded by a pluton dated at 242 Ma (Miller, 1981; Walker, 1988; Miller et al., in review). Rocks inferred to be Pennsylvanian/Permian continental borderland sediments in the Shadow Mountains are capped with undated arc-derived strata similar to those present in the El Paso Mountains (Martin and Walker, 1991; Martin, 1992).

From this evidence, it seems that eugeoclinal strata of the northern Mojave were juxtaposed against miogeoclinal/cratonal strata to the south by latest Permian time. The igneous activity present in these areas, but absent elsewhere in the southern Cordillera, serves to stitch these sequences together, as do Early Triassic rocks (Walker, 1987, 1988).

The mechanism of juxtaposition has been controversial. Left-lateral, strike-slip truncation of the continental margin and some portion of the craton has been proposed for the western Mojave Desert for late Paleozoic time (Davis et al., 1978; Stone and Stevens, 1988; Walker, 1988). Alternatively, extreme telescoping along a preexisting continental margin has been considered as the mechanism of juxtaposition (Poole, 1974; Snow, 1992). We prefer the former mechanism and timing for several reasons: 1) late Paleozoic shortening in the Mojave Desert is west-directed rather than east-directed as proposed for the Death Valley area; 2) Late Permian igneous activity in the western Mojave Desert probably indicates that this area shifted into a leading tectonic position during late Paleozoic time; and 3) isotopic data for Permian plutons in the eugeoclinal sections lack evidence for involvement of ancient continental crust which would be necessary for the telescoping model (Miller et al., in review). Hence, we interpret truncation of the western margin of the Mojave Desert and associated juxtaposition of miogeoclinal and eugeoclinal sections to be the first major event to modify the region.

Triassic to Middle Jurassic: Overlap and Arc Initiation

Shallow marine to nonmarine sedimentation across much of the region returned during Triassic time. Lower Triassic rocks consist of calcareous shales, limestone, and local conglomerate and sandstone, and their metamorphic equivalents. Facies patterns for these rocks trend northwest (Walker, 1987, 1988). Early Triassic plutonic rocks are also locally present (see above). The lower Triassic rocks rest paraconformably to unconformably on upper Paleozoic rocks in the eastern Mojave and unconformably on deformed Paleozoic rocks and Permo-Triassic plutons in the western Mojave Desert.

There are no dated Middle Triassic to Middle Jurassic (ca. 230 to 175 Ma) stratified rocks in the Mojave Desert. Early Triassic sequences give way upward into quartzose rocks which are overlain by volcanic and volcaniclastic strata. These volcanic rocks were interpreted by Walker (1987, 1988) to be Middle Triassic in age and to indicate that the area was the site of a magmatic arc of that age. Where dated, however, these rocks consistently yield Middle Jurassic ages (Busby-Spera, 1988; Graubard et al., 1988).

Triassic plutonic rocks are present in the central and southwestern Mojave Desert and surrounding areas (Miller, 1978; Barton et al., 1988; Barth et al., 1990; Miller et al., in review). These plutons range in age from 245 Ma to about 22 Ma and have been interpreted to represent early magmatic activity in the Cordilleran arc (Barth et al., 1990). Similar age plutonic rocks are exposed northward into the Death Valley area and into the Walker Lane region (Dilles and Wright, 19 Snow et al., 1991).

No deformational events have been identified in the Mojave block as Triassic to Early Jurassic in age. To the east, deformation in the Clark Mountains, previously considered t be pre-200 Ma (Burchfiel and Davis, 1971, 1981, 1988) is now considered to be Late Jurassic in age (Walker et al., unpublished data). Other structures, such as a fault in the Cowhole Mountains (Burchfiel and Davis, 1981), are not definitively dated. Although structures in the region may tu out to be Middle Triassic to Early Jurassic age, none have ye been clearly identified.

Early and Middle Jurassic: Plutons and Volcanic Rocks

Middle Jurassic (175 to 165 Ma) magmatic arc activity a associated sedimentation are widespread in the Mojave Des Rocks of the lower Sidewinder volcanic series are exposed f the Victorville area to the Cady Mountains and consist of intermediate to felsic volcanic rocks with intercalated quartz sandstone (Schermer and Busby, in press). This sequence is dominated by intracaldera sections >4 km thick and local exposures of thin (~1 km) sections. The age of these rocks is constrained by a ~170 Ma date at the base of the sequence (U-Pb zircon; Graubard et al., 1988) and ages of around 166 in the upper part of the lower series (Schermer, Mattinson, Busby, unpublished data). Similar age rocks, grouped with Delfonte Volcanics whose type section is in the Mescal Ran (see below), are present in the Cowhole Mountains (Busby-Spera et al., 1987). Possibly coeval rocks are also pre in the Soda Mountains (Grose, 1959), Cronese Hills area (Walker et al., 1990b), Providence Mountains (Hazzard, 195 and New York Mountains (Burchfiel and Davis, 1977). Volc rocks are commonly associated with quartzose sedimentary rocks in these areas. Walker (1987, 1988) interpreted the quartzose sedimentary rocks at the base of the volcanic sequences to be Middle Triassic in age, but it appears that age assignment of Middle Jurassic is more consistent with known ages and regional stratigraphic relations (Marzolf, 1 Older volcanic rocks are exposed to the east in southern Arizona, with some sequences dating back into the Early Jurassic (Tosdal et al., 1989). These older sequences seem to absent, however, across most of the Mojave Desert.

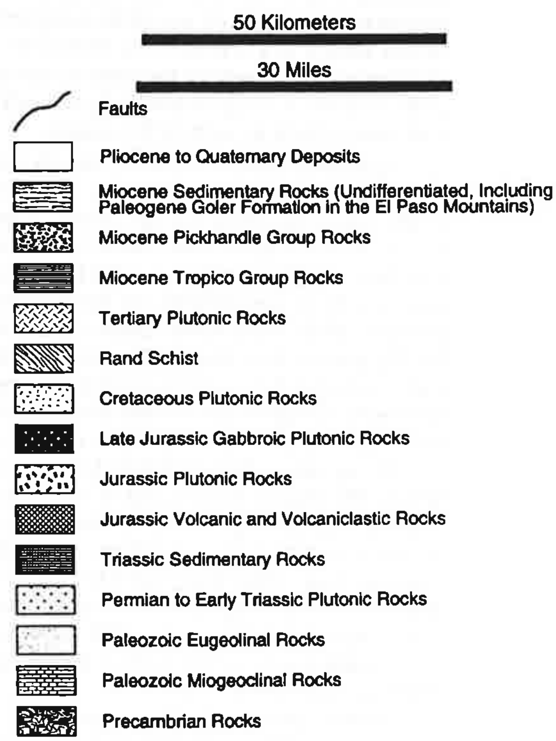
Middle Jurassic plutons are also common throughout t region. A significant fraction of the Jurassic plutons in the region cluster with ages of 155 to 170 Ma (Karish et al., 19 Walker et al., 1990b; Miller et al., in review). The plutons f the southern continuation of the Sierran batholith into the Mojave Desert. This belt continues southeastward into southeast California and Arizona.

Middle to Late Jurassic: Extension and Shortening

The structural development of the Jurassic arc is comp and of both contractional and extensional character. Norm faults cut the Lower Sidewinder volcanic series, resulting i about 15% north-south extension across the area where the Lower Sidewinder section is exposed (Schermer, 1993). Tin of extensional activity is constrained to be between ca. 166 the end of Sidewinder volcanism, and 150 Ma, the age of which crosscut the extensional faults (Schermer, 1993). Busby-Spera (1988) interpreted the Lower Sidewinder seri

(109)

Figure 2. Geologic map of the Mojave Desert. Modified, based on our work, from Dibblee (1968a) and other sources.



1989; Walker et al., 1990a). Starting in late Miocene time, the Mojave Desert was transected by northwest-striking, right-lateral strike-slip faults with local areas of east-striking, left-lateral strike-slip faults (Garfunkel, 1974; Dokka and Travis, 1990). Cumulative slip on these faults has been small. This deformation was accompanied by contractional faulting and folding across most of the region (Bartley et al., 1990).

Late Paleozoic: Truncation of the Margin

Major changes occurred in the tectonic setting and continental margin of the western Cordillera in late Paleozoic time. The Mojave Desert was dramatically affected by these changes because it went from an inboard tectonic position in middle Paleozoic time to a plate-margin position in latest Paleozoic time. Stratigraphic sequences across much of the region record Pennsylvanian to middle Permian platform sedimentation conformable with older miogeoclinal/cratonal strata. This pattern is broken for sections in the western Death Valley area and into the central and northeastern Mojave Desert (Martin and Walker, 1991; Walker and Wardlaw, 1989) where Pennsylvanian to middle Permian rocks consist of turbidites and other deep water strata apparently deposited in a continental borderland setting (Stone and Stevens, 1988; Walker, 1988). Mojave eugeoclinal sections, which were part of the Antler belt of Nevada and eastern California during Mississippian time, also show evidence for similar active basin formation in Pennsylvanian and especially Early Permian time have been deposited in an arc-graben depression that ran the length of the Early and Middle Jurassic arc. Little direct evidence for graben structures is present and the Sidewinder sequence in the Mojave Desert owes its preserved thickness largely to caldera activity (Schermer, 1993). Hence, extensional deformation is post-extrusion rather than synextrusion, and occurred during the interval 166 to 150 Ma.

Regional contractional deformation is present to the north of the Sidewinder exposure belt. Deformation is typified by mylonitic shear zones, thrusts faults, and probable thrusts that verge to the east and southeast (Boettcher and Walker, 1993; Davis and Burchfiel, 1973; Miller et al., 1991; Stephens et al., 1993; Walker et al., 1990b). The shear zones and thrusts generally place Jurassic plutonic rocks over Mesozoic sedimentary and volcanic strata. Hence, the zones place structurally lower rocks over structurally higher ones. Deformational fabrics developed under prograde metamorphic conditions to greenschist or amphibolite facies (Boettcher and Walker, 1993; Walker et al., 1990b; Miller et al., 1991). Timing relations for this contractile event are best bracketed in the Cronese Hills area. Prekinematic plutonic rocks are dated at 166 ± 3 Ma and postkinematic granite is dated at 155 ± 1 Ma, giving a possible age range of 169 to 154 Ma (Walker et al., 1990b). Timing relations in other areas are consistent with this age, and are bracketed between ca. 170 to 148 Ma (Boettcher and Walker, 1993; Miller et al., 1991; Stephens et al., 1993). Walker et al. (1990b) considered this deformational belt to be continuous with the east Sierran thrust belt exposed north of the Garlock fault.

Reconciling the evidence for essentially coeval contractional and extensional deformation across this region (bracketed between 169 and 154 Ma, and 166 and 150 Ma, respectively) is crucial to understanding the late Middle Jurassic development of the area. Perhaps one of the most critical relations is that exposed from the Iron Mountain area southward to the northern part of Silver Mountain (Stops 8 and 9). At Iron Mountain, rocks equivalent to the Lower Sidewinder volcanic series were deformed under greenschist facies metamorphic conditions during an episode of contractional deformation (Boettcher and Walker, 1993). The deformational fabrics are cut by a 151 ± 10 Ma pluton. Lower Sidewinder rocks at Silver Mountain, only about 10 to 15 km to the southwest of Iron Mountain, are affected only by extensional deformation. Hence, it is clear that Lower Sidewinder strata experienced both the contractional and extensional events, and that these events may not be completely isolated in areal extent.

As described above, Permo-Triassic plutonic rocks that intrude the eugeoclinal sections show no evidence for the involvement of Precambrian continental crust. Middle Jurassic plutons (as old as 175 Ma), on the other hand, have Pb, Sr, and Nd isotope ratios and zircon systematics that show strong involvement of Precambrian crust in their genesis (Miller and Glazner, 1994a; Miller et al., in review). This indicates that stacking of eugeoclinal rocks onto miogeoclinal rocks and their Proterozoic basement may have occurred between Early Triassic and Middle Jurassic time (235 to 175 Ma). No structural features in the Mojave Desert have yet been associated with this time period. In the east Sierran thrust system to the north, there is abundant evidence for some pre-180 Ma deformation (Dunne, 1986; Dunne et al., 1978). If these structures continue into the Mojave, they have since been obliterated by later deformation and plutonism.

Late Jurassic: Independence Dike Swarm

The late Middle to early Late Jurassic deformation was followed by limited plutonism and volcanism of Late Jurassic age. Late Jurassic granitoids are exposed in the Cronese Hills, at Iron Mountain, and in the Shadow Mountains (Boettcher and Walker, 1993; Walker et al., 1990b). Other such granitoids, if present, await identification.

The most widespread Late Jurassic igneous rocks are those related to the Independence dike swarm (IDS). The IDS consists of a suite of north-northwest-striking, mafic to felsic dikes that run from the eastern wall of the Sierra Nevada southward into the Eagle Mountains of the Eastern Transverse Ranges (Chen and Moore, 1979; James, 1989). These dikes commonly have a quoted age of 148 Ma. Similar NNW striking dikes at Cronese Mountain are dated at 152 ± 1 Ma (Walker, unpublished data) and at Black Mountain at about 150 Ma (Schermer, 1993). Hence, some age range is implied for these rocks. Coeval mafic plutons are locally exposed in the Shadow Mountains, Goldstone area, and probably at Iron Mountain. The IDS apparently fed lava flows of the Upper Sidewinder volcanic series (Karish et al., 1987; Schermer and Busby, in press).

Cretaceous Arc

Regionally voluminous plutonism and important deformation occurred across the Mojave Desert during Cretaceous time. Cretaceous plutonic rocks are common in the western and eastern parts of the Mojave Desert and are scattered across the rest of the region. These rocks are typically alkali feldspar and muscovite-garnet granites (Miller and Glazner, in prep.) and differ significantly from the more tonalitic Cretaceous rocks of the southern Sierra Nevada (Saleeby et al., 1987; Ross, 1989). Most of the dated Cretaceous rocks are Late Cretaceous in age. Cretaceous volcanic rocks are only locally exposed in the eastern Mojave Desert where they are preserved beneath thrust faults. As in the Sierra Nevada, earliest Cretaceous (145-125 Ma) plutons are scarce to absent (Glazner, 1991).

Cretaceous deformation is present across the Mojave Desert. Mylonitic shear zones of latest Jurassic to Cretaceous age are exposed from Ord Mountain to Tiefort Mountain . The zones strike north-south at Ord Mountain and have a shallowly-plunging lineation (Karish, 1983). At Iron Mountain these zones are steep and strike northeast, again with a shallow lineation. Shear sense indicators yield both dextral and sinistral results, with dextral being somewhat more common. These shear zones are present in rocks as young as Cretaceous biotite granite and are crosscut by 83 Ma dikes (Boettcher and Walker, 1993). At Tiefort Mountain, similar steep shear zones cut a ~105 Ma pluton but are intruded by ~82 Ma dikes (Stephens, Schermer, and Walker, unpublished data). In addition, Sidewinder rocks are folded about northwest-trending, upright folds.

High grade metamorphic rocks and migmatites of Cretaceous age are exposed from the Buttes to the Fremont Peak area, where mineral assemblages record temperatures up to 650°C and pressures on the order of 6 kb (Fletcher, unpublished data). Monazite and zircon from leucosomes of synkinematic migmatite in the Buttes yield ages around 95 Ma (Martin and Fletcher, unpublished data), and zircon from synkinematic garnet-muscovite granite at Fremont Peak gives an age of 98 Ma (Miller et al., 1992). Hence, deformation and high-grade metamorphism occurred in the area during early Late Cretaceous time.

The Rand thrust is exposed in the northwestern part of the

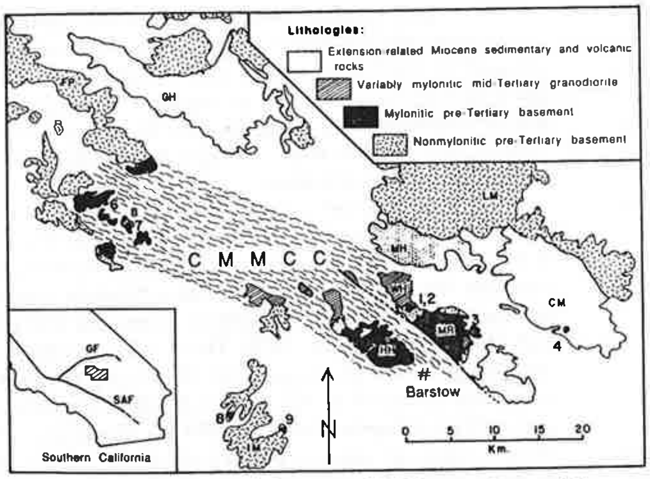


Fig. 3. Tectonic map of the central Mojave metamorphic core complex (CMMCC). The zone of ductile deformation is largely coextensive with outcrops of the Waterman Hills granite. HH = Hinkley Hills; WH = Waterman Hills; MR = Mitchel Range; CM = Calico Mountains; LM = Lane Mountain; MH = Mud Hills; GH = Gravel Hills; FP = Fremont Peak; B = Buttes; IM = Iron Mountain; SAF = San Andreas fault; GF = Garlock fault. Modified from Fletcher and Bartley (in press).

Mojave Desert . This thrust places North America miogeoclinal and related rocks (the Johannesburg Gneiss) over the Rand Schist, a complex of Mesozoic forearc strata. The timing of this structure is best estimated as late Cretaceous. The Rand Schist is cut by a postkinematic pluton dated at 79 ± 1 Ma (Silver and Nourse, 1986). The hanging wall of the thrust contains the Buttes Cretaceous migmatite complex (95 Ma). Hence, we interpret the thrusting to have preceded the postkinematic pluton, but postdated migmatization in the hanging wall in the adjacent Butte/Fremont Peak area.

Early Tertiary: Quiescence and Cooling

The early Tertiary was a time of quiescence in the Mojave block. No rocks younger than Late Cretaceous and older than late Oligocene have been identified, although Eocene and Oligocene sedimentary rocks are known north of the Garlock fault. This indicates that the region was a highland undergoing external drainage (Hewett, 1954; Nilsen, 1977). Hewett estimated uplift of over 4 km, based on the thickness of pre-Mesozoic strata outside the block and their absence within. Coney and Reynolds (1977) proposed that the lack of magmatism in spite of active subduction was a result of rapid, flat subduction, with analogy to part of the Andean margin today, and Damon (1979) proposed that uplift resulted from subduction of progressively younger lithosphere.

Late Oligocene to Early Miocene: Magmatism, Extension, and Sedimentation

The Oligocene-Miocene boundary marked a dramatic return of magmatism and tectonism to the Mojave block. At about 22-24 Ma, volcanic rocks were erupted along an east-trending belt that stretched from the westernmost Mojave Desert inland to the Whipple Mountains and beyond (Glazner, 1990; Glazner and Bartley, 1984). The onset of magmatism was accompanied by the onset of extensional faulting, as both swept northwestward out of Arizona in concert with the Mendocino triple junction (Glazner and Bartley, 1984). Isotopic ages indicate that volcanism began abruptly across this belt at about 23 ± 1 Ma (Glazner, 1990; Glazner and Bartley, 1984). Magmatism was accompanied in the central Mojave Desert by locally intense but areally restricted extension.

Volcanism, which locally produced piles up to several km thick, was a combination of intermediate, Cascade-type calc-alkaline magmatism and bimodal accumulations of high-Ti, mildly alkaline basalts and basaltic andesites with silicic tuffs (Glazner, 1981; Glazner, 1990). Isotopic data indicate that magmatism involved significant recycling of preexisting crust; for example, $^{87}Sr/^{86}Sr$ ratios correlate positively with both SiO_2 and distance inland (Glazner and O'Neil, 1989). The areal correlation probably reflects both an eastward increase in the proportion of Proterozoic rocks in the crust, as well as changes in the underlying mantle lithosphere. New data (Keith et al., 1994; Miller and Glazner, 1994b) indicate that rocks from the north-central Mojave Desert, where eugeoclinal strata are exposed, have uniformly low $^{87}Sr/^{86}Sr$ and high ε_{Nd}, consistent with derivation from oceanic lithosphere.

Near Barstow, magmatism and intense crustal extension were synchronous. The oldest volcanic rocks) are about 23-24 Ma, as is the synkinematic Waterman Hills granite pluton in the footwall of the Waterman Hills detachment fault (Fig. 3). Extensional basin development and accumulation of the Pickhandle Formation began about the same time (Fillmore and Walker, 1993a,b; Walker and Fillmore, 1993; Bartley and Glazner, in prep.). Field observations from the central Mojave metamorphic core complex indicate that magmatism and mylonitic deformation were closely linked. Mylonitization is intense near synkinematic plutons and dikes (see Stops 1 and 3), and weak to absent elsewhere. The most intensely extended rocks (as indicated by extreme distension and tilting of upper-plate rocks) are only found in the area from the Mitchel Range to the Buttes, roughly coincident with the areal extent of the Waterman Hills granite.

The areal extent of extension is a topic of some controversy. Dokka (1989) and Tennyson (1989) proposed that much of the Mojave Desert, including virtually all of the western ("Edwards terrane") and most of the eastern ("Daggett terrane") Mojave block, was extended in the Miocene. However, as we show below (see Road log), much of the area included in these regions was unaffected by extension and the dominant form of deformation was actually crustal shortening. For example, in the Kramer Hills, Dokka (1989) proposed that steep Miocene strata reflect tilting above a shallow detachment fault, but field studies (Bartley et al., 1990; Dibblee, 1967a; Linn, 1992) demonstrate that the strata and their basal nonconformity are tightly folded, as are strata in surrounding ranges. There is no structural or stratigraphic evidence for Miocene extension in the western Mojave Desert.

In the eastern Mojave block, many ranges included in the "Daggett terrane" are essentially undeformed (Glazner and Bartley, 1990). Closer to Barstow, modest amounts of extension are expressed by homoclinally tilted strata in the Newberry and Cady Mountains (Dokka, 1986; Glazner, 1988), but the aggregate extension in these areas is likely to be small (a few km or less), and field relations in the Rodman Mountains indicate that homoclinal tilting there may be a result of transpression (Dibblee, 1990; Sanner, 1985; Bartley and Glazner, in prep.). We conclude that significant extension in the Mojave Desert was restricted to the area northwest of Barstow. Extension was probably accommodated laterally by a variety of

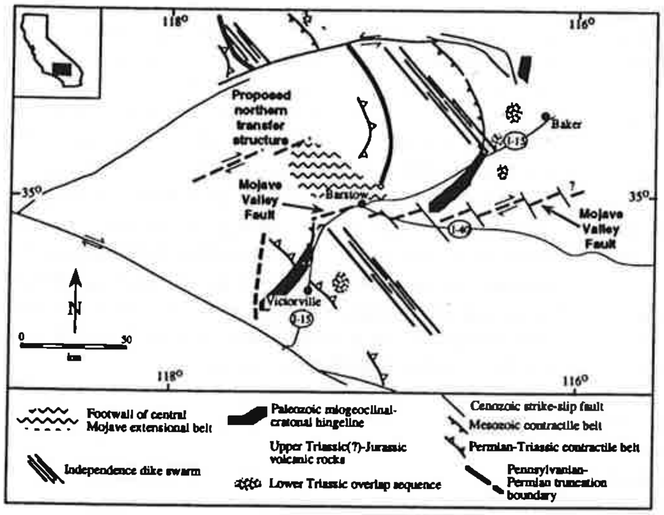


Fig. 4. Map showing pre-Tertiary structural and stratigraphic features that are offset along the Mojave Valley fault (MVF), an inferred transfer fault that bounds the CMMCC on the south. From Martin et al. (1993).

transfer mechanisms (Fig. 4; Bartley and Glazner, 1991; Glazner et al., 1992; Martin et al., 1993).

Large-scale extension (tens of km) near Barstow caused a significant rearrangement of pre-Cenozoic structure and stratigraphy. Glazner et al. (1989) showed that removing this extension greatly simplifies the geology of the Mojave Desert; for example, removing extension aligns the Independence dike swarm and many other pre-Cenozoic markers (Glazner et al., 1989; Martin et al., 1993).

Sedimentary rocks deposited during extension vary greatly depending on their position relative to the extensional basin. Fillmore et al. (in review) show that three main basin types were present: (1) the intrarift Pickhandle basin, which received a thick section of coarse clastic and volcanic detritus; (2) the extrarift Tropico basin, which lay to the southwest and probably formed by flexure of the footwall during extension; and (3) intra-hanging wall basins to the east, including the Clews basin at Alvord Mountain and the Hector basin in the Cady Mountains.

Various paleomagnetic studies have produced a confusing array of possible Miocene vertical-axis rotations, with studies in adjoining areas commonly giving contradictory results (Ross et al., 1989; Valentine et al., 1993). These studies are typically forced to rely on single stratigraphic units, where secular variation may not be adequately averaged. In addition, many of the units sampled were affected by intense noncylindrical folding which makes standard structural corrections inadequate (Bartley et al., 1990). Thus, further work is needed to sort out the inferred rotations. Data from Wells and Hillhouse (1989) indicate no rotation in much of the Mojave block since eruption of the 18.5±0.2 Ma Peach Springs tuff.

Dokka (1989) proposed, on the basis of these data, that extension in the Mojave Desert originally occurred with the hanging wall moving to the north, and that the current northeast orientation of extension vectors resulted from clockwise rotation. We view this as unlikely for several reasons, including the consistent orientation of many structural markers (e.g., Independence dike swarm and Late Cretaceous dikes), the parallelism of the present extension vector to those in much of the rest of California and Arizona (Bartley and Glazner, 1991), and the observation that vertical-axis rotations in the Colorado River extensional corridor are commonly restricted to the hanging walls of normal fault blocks (Wells and Hillhouse, 1989).

Middle Miocene to Recent: Sedimentation, Faulting, Transpression, and Volcanism

Following early Miocene extension, which was over by Peach Springs tuff time (18.5 Ma), the central Mojave Desert was the site of fluviolacustrine deposition of the lower Barstow Formation, upper Tropico group, and upper Hector Formation. The Barstow Formation sits in angular unconformity upon Pickhandle strata in the Mud Hills (Dibblee, 1968a), and records quieter deposition and less volcanism. A similar transition is recorded at Alvord Mountain to the east, where the Barstow Formation overlies deformed lower Miocene Clews Formation strata that were deposited in a hanging-wall basin

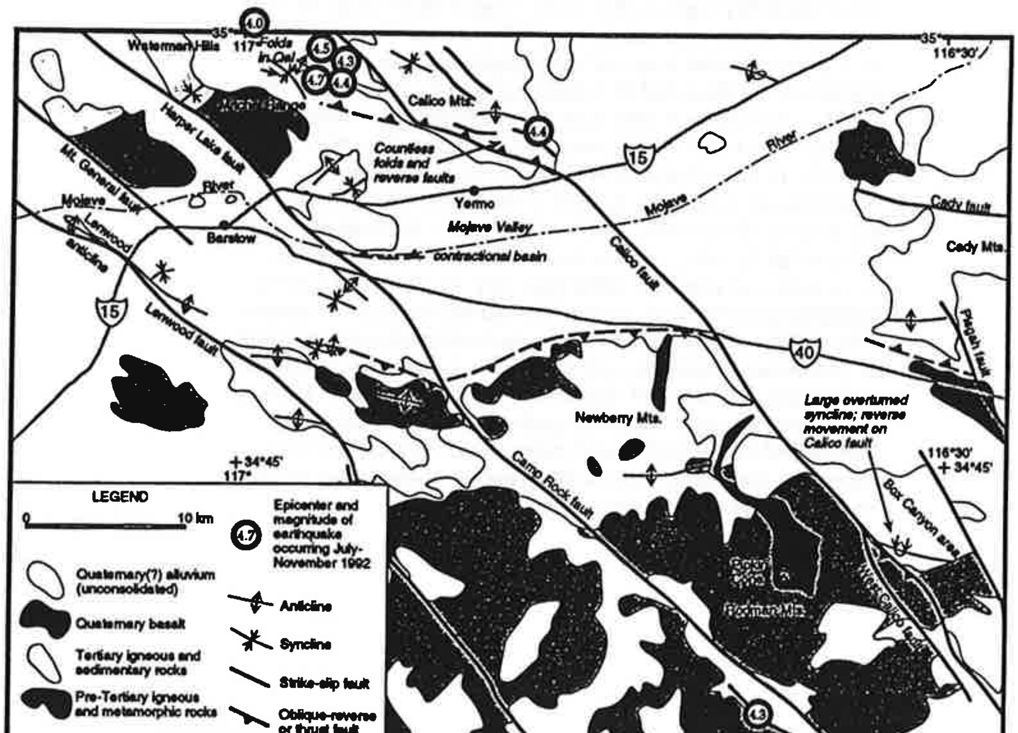


Fig. 5. Interpretive tectonic map of the area around Barstow, emphasizing late Neogene contractional and strike-slip faults. Based in part on relations described in Bartley et al. (1990) and Bartley et al. (1992). Note the abundance of contractional structures, many of which are currently active; these continue to the northwest.

(Fillmore et al., in review). We attribute this sedimentation to filling of extensional and flexural basins, coupled with thermal subsidence.

Volcanism continued its northward migration. Volcanism at the latitude of Barstow shut off about 18 Ma, although some of the undated silicic plugs around Barstow could be younger. Post-18 Ma volcanism was concentrated in the northern Mojave Desert, northwest of Barstow (Burke et al., 1982), on the China Lake and Fort Irwin military bases (Schermer, in prep.), and in the Lanfair Valley area (Turner and Glazner, 1991). This volcanism is more clearly bimodal than that east of Barstow (Keith et al., 1994; Schermer, L. Keith, and Glazner, in prep.), sits nonconformably on pre-Tertiary basement, and was not affected by extension.

Southeast of Barstow, volcanism resumed after a hiatus of 10-20 Ma as isolated alkali basalt cinder cones and lava flows (Wise, 1969). These lavas sit in angular unconformity on tilted early Miocene strata. K/Ar dating indicates that most of these volcanic rocks were erupted within the last 10 Ma (Glazner and Farmer, 1993). They bear little relationship to current structure in the region, and some were erupted through areas undergoing active crustal shortening (Glazner and Bartley, in press).

The dominant post-early Miocene deformation comprises strike-slip faulting and related transpression (Bartley et al., 1990). Although it is commonly assumed that strike-slip faulting began when the Gulf of California opened, 4-5 Ma ago, there is evidence that it began at least locally in the early Miocene (Bartley et al., 1990). Transpressional structures are ubiquitous across the Mojave block, and are superimposed on extended rocks in the eastern part of the area.

Transpression and strike-slip faulting have shaped much of the present-day topography of the Mojave block (Fig. 5). This view is in sharp contrast to that of Dokka and Travis (1990), who proposed that much of the topography is controlled by transtension, which led to several large pull-apart basins. We see several problems with their interpretation, the most important of which is that many of the basins that they consider to be pull-aparts are bounded by or contain contractional structures. Specific examples of this include: (1) the Mojave River valley south of the Calico Mountains, which is bounded by steep dextral-reverse faults which carry the mountains out over the valley (see Stop 4); (2) valleys in the southern Mojave block, east of the San Bernardino Mountains, where Dibblee (1967b, 1968b) mapped contractional structures in Quaternary alluvium; and (3) Coyote Lake basin, northeast of Barstow, where Meek (1990) showed that the basin is being contracted. We thus reject the pull-apart interpretation of Dokka and Travis (1990) and note instead that contraction caused by transpressional faulting has produced much of the current topography in the region.

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Waterman Hills (north of Barstow) Field Trip Stop

About 1980, geologists began to recognize areas of very large crustal extension, one expression of which was the development of low-dip normal faults called extensional detachment (or just detachment) faults. Where large (>10 km) slip occurred on these detachments, removal of the hanging wall led to resultant isostatic upbowing of the footwalls. As a consequence, we commonly observe in such areas a domal footwall of high-grade metamorphic footwall rocks structurally (separated by the detachment) overlain by (1) thin, extensionally deformed (brecciated to shattered) remnants of hanging wall rocks, and (2) sedimentary rocks deposited in local basins formed during the extensional process. This entire assemblage of structures and rocks have come to be called ***metamorphic core complexes.***

Metamorphic core complexes and other signs of great extension were recognized in the early 1980's in the Las Vegas Valley, Death Valley and Colorado River areas. These are all considered expressions of the Basin-and-Range orogeny. Geologists working in the Mojave Desert were initially puzzled as to why so little apparent Basin-and-Range extension occurred in that region, considering all the extension that had occurred to the north, northeast, and southeast. Roy Dokka (a CSUN alumnus) was among the first to recognize evidence of detachment faults in the Mojave, based on his work in the Newberry Mountains in the mid-1980's. He and his students have continued their research on the Mojave region. Another group of researchers led by Allen Glazner at the University of North Carolina began studying the late Cenozoic evolution of the central Mojave Desert about the same time as Dokka et al. Their interpretations agree with and build upon Roy's work in some instances but are in direct conflict with other aspects of his extensional models. In recent years the two research groups have "butted heads" on a number of occasions at meetings and in published research articles.

Our stop north of Bastow is located in a metamorphic core complex underlying the Waterman Hills and surrounding areas. This complex has come to be called the Central Mojave Metamorphic Core Complex by many workers. Interpretations of this complex indicate that 50 ±10 km of ENE-directed extension occurred ~19 to 21 Ma, resulting in elevation of a thoroughly metamorphosed footwall of midcrustal rocks being elevated to the surface.

Points of interest that will be discussed at this stop include:

- The geology of the core complex
- How early Miocene sedimentary formations in the region have been interpreted as being influence by this extensional event.
- How the present distribution of pre-Cenozoic rock units and structures across the Mojave might be explained by the location and kinematics of this core complex.

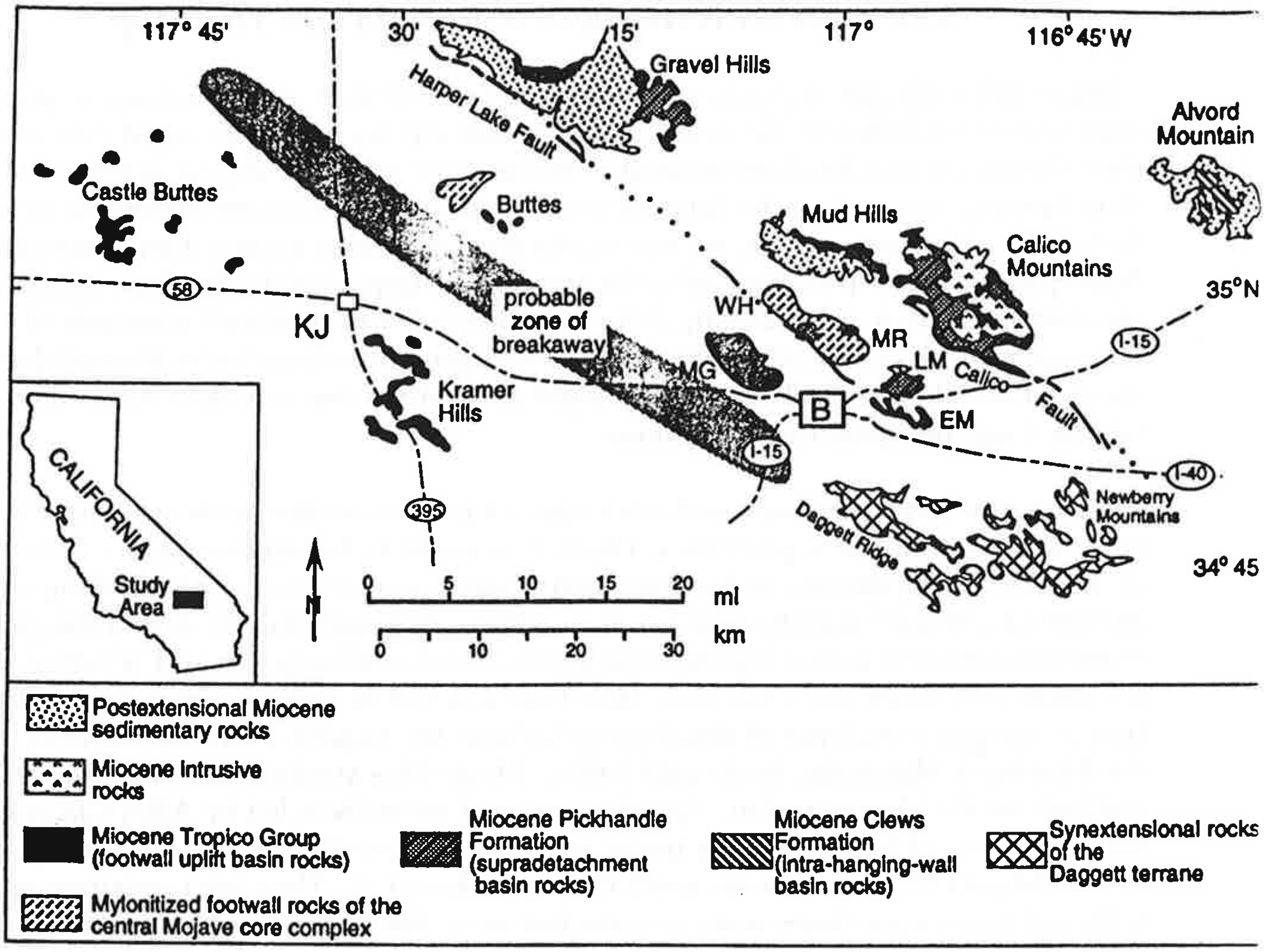


Figure 2. Simplified geologic map of central Mojave Desert showing distribution of lower Miocene rocks and footwall rocks of central Mojave metamorphic core complex. EM—Elephant Mountain, LM—Lead Mountain, MG—Mt. General, MR—Mitchel Range, WH—Waterman Hills, B—town of Barstow, KJ—Kramer Junction.

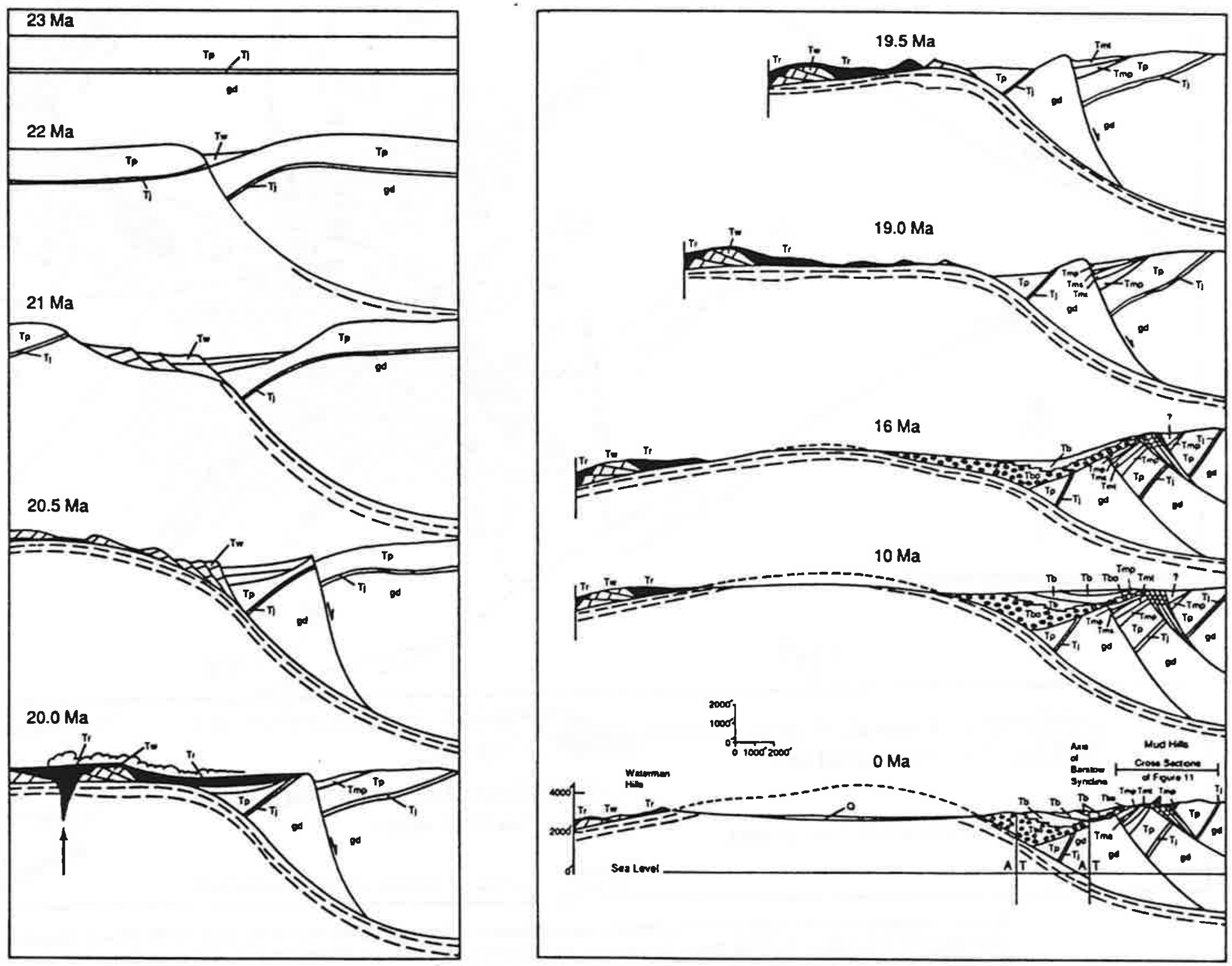


Figure 15 (on this and facing page). Sequential development of Waterman Hills detachment and Mud Hills area (see Fig. 1 for location). Palinspastic reconstruction is at true scale, with right side (north) held fixed. Some thin units (e.g., distal Tms) are omitted because of scale. Topography and geologic cross section of study area in Mud Hills in modern cross section (bottom right) are well constrained (e.g., cross sections of Fig. 11); other aspects of reconstruction are speculative, but consistent with all observations and data. Arched detachment fault after Walker et al. (1990). Two right-slip vertical faults are shown for modern cross section, after Dibblee (1967, fig. 55). Ages are interpolated between well-constrained times. Extension may have begun earlier than shown in upper left (e.g., Fillmore and Walker, this volume); 21 Ma is latest possible time in Mud Hills area. Symbols: gd = Mesozoic granodiorite; Tj = Jackhammer Fm.; Tp = Pickhandle Formation; Tmp = plutoniclastic breccia of Mud Hills Formation; Tmt = rhyolite-tuff breccia of Mud Hills Formation; Tms = sandstone of Mud Hills Formation; Tbo = Owl Canyon Conglomerate Member of Barstow Formation; Tb = other members of Barstow Formation; Tr = rhyolite of Waterman Hills; Tw = sedimentary strata of Waterman Hills; Q = Quaternary units. Long-dashed lines indicate mylonitic gneiss of lower plate of Waterman Hills detachment. Short-dashed lines indicate inferred eroded detachment fault.

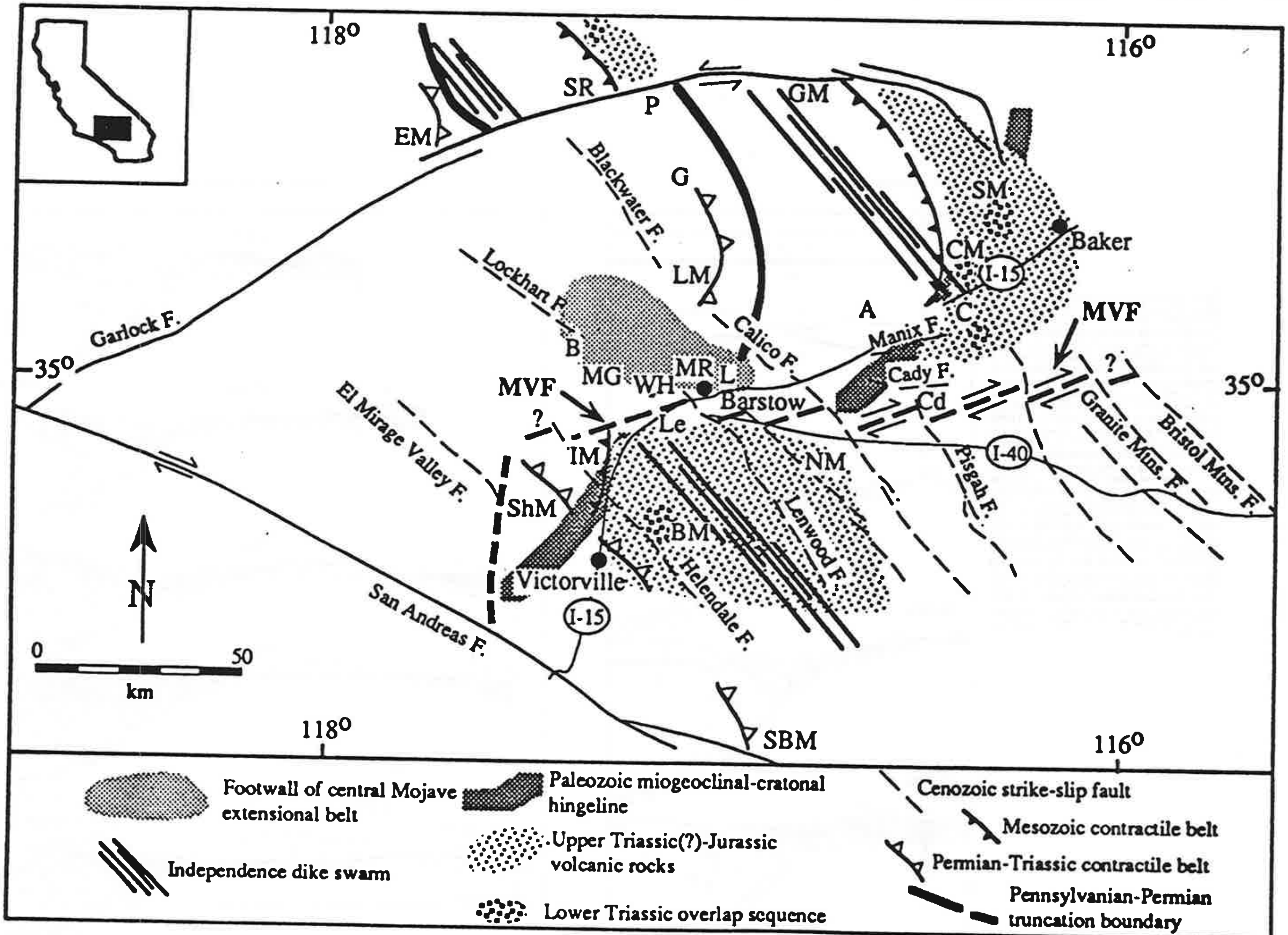


Figure 1. Regional tectonic map of Mojave Desert showing proposed location of Mojave Valley fault (MVF). EM—El Paso Mountains, SR—Slate Range, GM—Granite Mountains, P—Pilot Knob, B—Buttes, G—Goldstone, A—Alvord Mountain, CM—Cronese Mountain, SM—Soda Mountains, C—Cave Mountain, Cd—Cady Mountains, LM—Lane Mountain, MG—Mount General, MR—Mitchel Range, NM—Newberry Mountains, L—Lead Mountain, Le—Lenwood, IM—Iron Mountain, ShM—Shadow Mountains, BM—Black Mountain, SBM—San Bernardino Mountains, WH—Waterman Hills. Jurassic and Permian-Triassic contractile belt symbols do not necessarily imply vergence or dip direction. Paleozoic miogeoclinal-cratonal hingeline is used as defined by Martin and Walker (1991, 1992). Mojave Valley fault is schematically shown offset by northwest-trending late Cenozoic strike-slip faults,

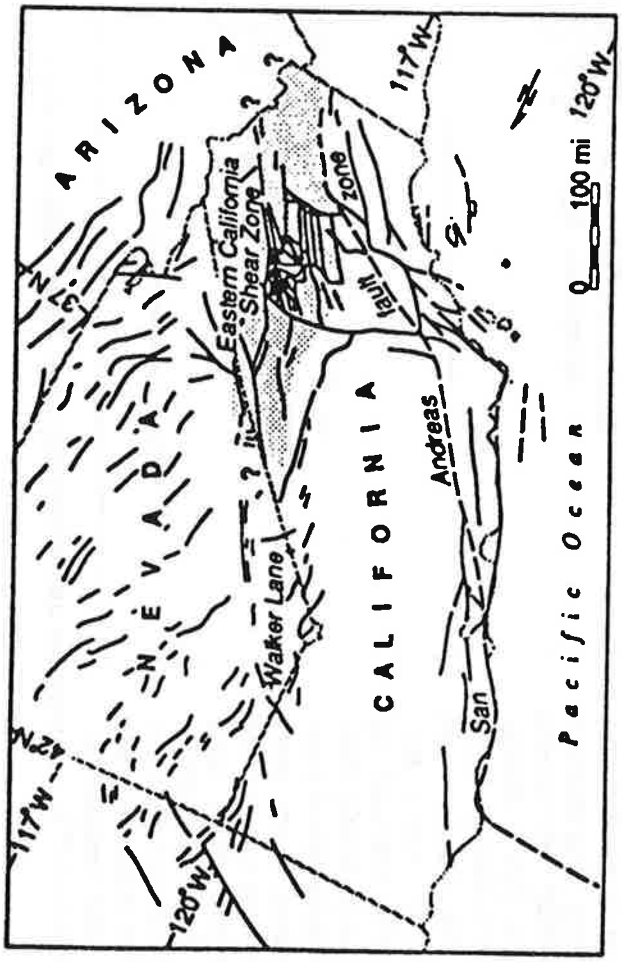


Figure 6. (A) The Pacific-North American transform boundary in the western USA highlighting the location of the Eastern California shear zone (Dokka and Travis, 1990b). (B; following page) Fault map of the Mojave Desert highlighting the location of late Cenozoic faults and associated features (from Dokka and Travis, 1990a). AM, Alvord Mountains; AW, Avawatz Mountains; BM, Bristol Mountains; CM, Calico Mountains; CdM, Cady Mountains; CP, Cajon Pass; GM, Granite Mountains; MH, Mud Hills; MM, Marble Mountains; NM, Newberry Mountains; OM, Ord Mountain; PR, Paradise Range; RM, Rodman Mountains; SBM, San Bernardino Mountains.

Eastern California Shear Zone. Most syntheses dealing with the evolution of the Pacific-North American transform boundary have emphasized the part played by the San Andreas fault system and subparallel faults to the west (Fig. 6), even though the existence of similar late Cenozoic faults in the Mojave Desert and east of the Sierra Nevada has been well established [e.g., Longwell, 1960; Dibblee, 1961; Burchfiel and Stewart, 1966; Hamilton and Myers, 1966; Hill and Troxel, 1966]. Recent studies by Dokka and Travis [1990ab] were initiated to better understand the kinematic development of NW-striking right slip faults of the Mojave Desert Block and their role in the tectonic framework of southern California (Fig. 7). Previously, most tectonic models for southern California have treated the entire Mojave Desert Block as the site of distributed simple shear (i.e., homogeneous strain) during post-middle Miocene time [Garfunkel, 1974; Carter et al., 1987]. In contrast, analysis of the region indicates that strain is regionally heterogeneous and is partitioned into six domains that are separated by major strike-slip faults and extensional zones [Dokka and Travis, 1990a]. Tectonic rotation of these domains as well as their internal deformation by strike-slip faulting have occurred as the result of broadly distributed regional right shear; sixty-five km of total right slip is reckoned to have occurred along faults of the southern half of the province. This broad network of faults, along with kinematically and temporally similar strike-slip faults of the Death Valley region (Furnace Creek and Southern Death Valley fault zones) and intervening extensional zones, constitute a regional, throughgoing zone of right shear named by Dokka and Travis [1990ab], the Eastern California shear zone (ECSZ). This zone of intracontinental shear likely continues to the north where it may include the Walker Lane belt of western Nevada (Fig. 6).

Regional and local dated cross-cutting relations constrain the time of onset of right shear across the ECSZ. At the regional scale, all late Cenozoic strike-slip faults of the central and eastern Mojave cut and fully displace ~20 Ma elements of the early Miocene Mojave Extensional Belt [Dokka, 1983, 1989ab]. Locally, the best evidence for the age of initiation of an individual fault in the central Mojave can be seen in the Calico Mountains-Mud Hills area along the Calico-Blackwater fault [Dokka, 1989a]. Here, the youngest dated rock that is displaced the full amount (10 km) is a 13.4±0.2 Ma tuff from near the top of the Barstow Formation [MacFadden et al., 1990]. The Garlock fault (northern boundary of the Mojave Desert Block) was initiated near 10 Ma [Burbank and Whistler, 1987] and is truncated at its extreme eastern end by faults of the ECSZ (southern Death Valley fault zone [Davis and Burchfiel, 1973; Plescia and Henyey, 1982]). Age relations described by Stewart [1983] in the adjacent Death Valley region imply that faulting in that part of the ECSZ may have begun as recently as late Miocene (~6 Ma). Paleomagnetic data from rocks south of the Pinto Mountain fault in the adjacent eastern Transverse Ranges suggest that regional deformation there and, by inference, deformation of the Mojave Desert Block began after ~10 Ma [Carter et al., 1987]. Based on the relationships presented above, we conclude that the ECSZ became active no earlier than 20 Ma and no later than 6 Ma. An initiation time between ~10 Ma and 6 Ma (late Miocene) is considered to be most likely.

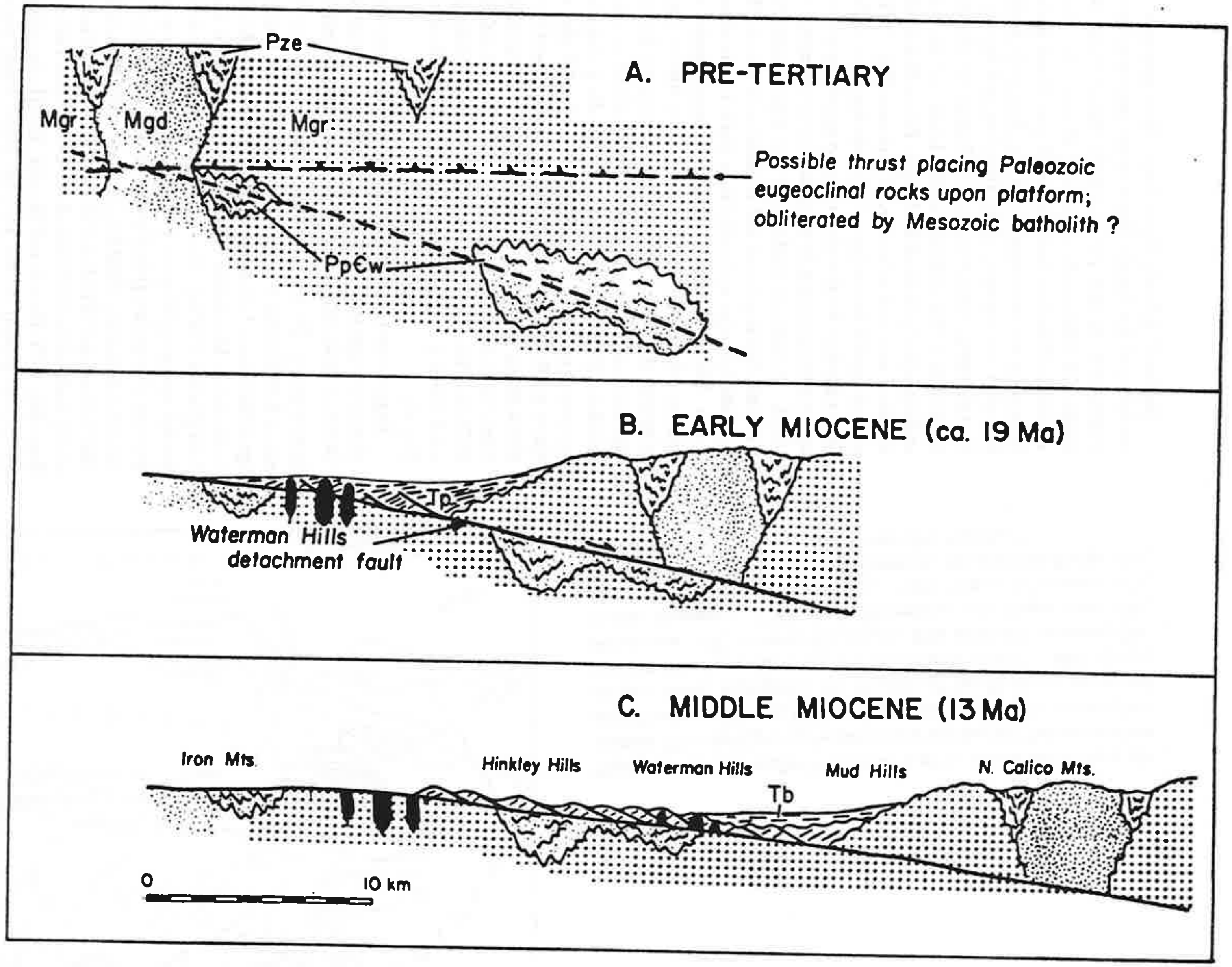
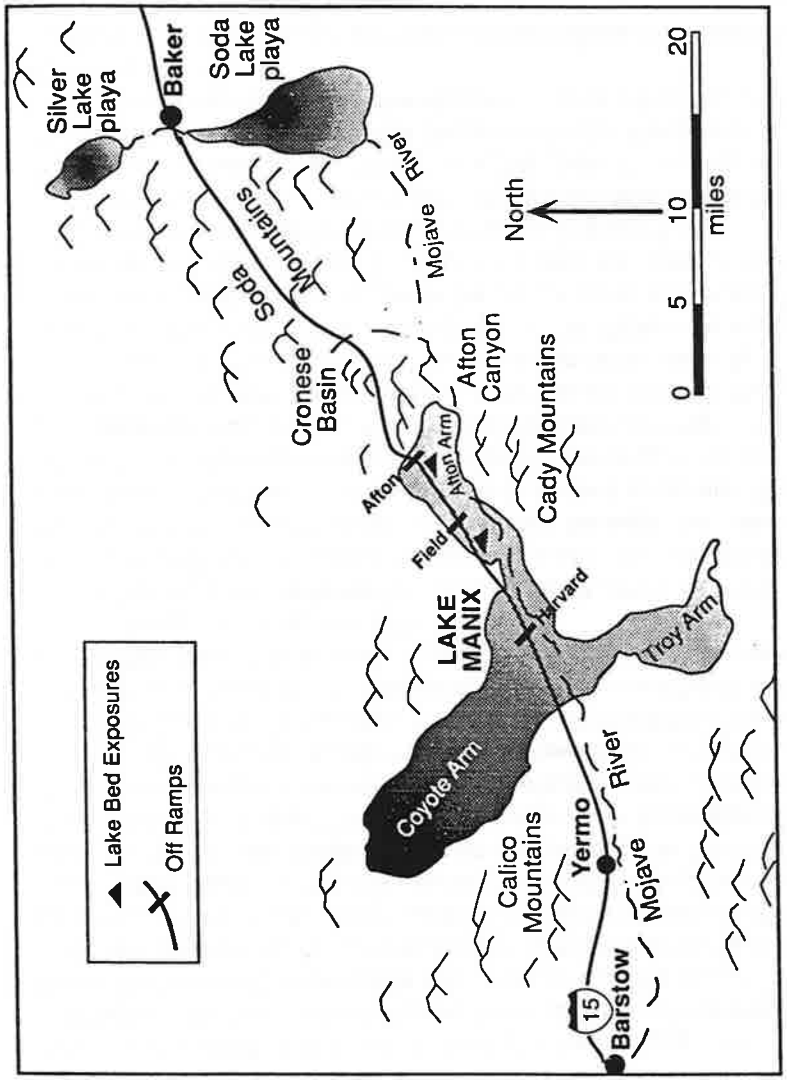


Figure 3. Conceptual model for evolution of Waterman Hills detachment fault (WHDF). Neogene folding related to right-slip Calico fault (Dibblee, 1968) has been removed. A: Geometry with 40 km of displacement on WHDF restored. Eugeoclinal Paleozoic rocks (Pze) lie structurally above miogeoclinal/cratonal Paleozoic strata in Waterman Gneiss (PpЄw). These strata are engulfed by Mesozoic batholith, including gabbro-diorite complex (Mgd) and more widespread granodioritic intrusions (Mgr). B: Geometry during displacement along WHDF. Pickhandle Formation (Tp) is deposited in extensional basin formed by displacement along WHDF and is syntectonically intruded by rhyolite plugs (black). Continued displacement truncates plugs, upper parts of which now are exposed in Waterman Hills; roots of plugs have not been located. C: By mid-Miocene time, after movement has ceased, post-tectonic Barstow Formation (Tb) accumulates unconformably upon Pickhandle Formation in topographic depression formed by extension.



Lake Manix area. Modified from Meek, 1989.

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VIGNETTE 15

FLAMINGOS IN THE DESERT
Pluvial Lake Manix

SAN BERNARDINO COUNTY

The Mojave Desert was not always as desolate and dry. About 15,000 years ago, not long ago geologically speaking, streams flowed through the arid and semi-arid terrain nourishing blue-water lakes in the broad valleys. The cooler and wetter climates of the ice ages were pluvial periods of increased rain in southern California. Pluvial water from the Sierra Nevada ran all the way to Death Valley through the Owens River and a string of four large lakes. The Mojave River, flowing out of the San Bernardino Mountains, first north, then east, and finally north again, also reached Death Valley through a succession of at least three large lakes. Death Valley, its floor 280 feet below sea level, was the ultimate sink for much of our southwestern desert, including the lofty (12,000 feet) Spring Mountains of western Nevada, which still feed water to Death Valley through the Amargosa River. Not surprisingly, Death Valley harbored a pluvial lake more than 100 miles long and 600 feet deep.

Those desert lakes must have been oases. Those with through-flowing streams were fresh, but lakes with no outlets gradually turned brackish and eventually saline as evaporating water left its dissolved salts behind. The pluvial lakes dried up with the change in climate after the last ice age ended some 10,000 years ago, and they all became brackish or saline in their final stages. Those desert lakes must have been beautiful in their prime; their placidness and deep blue color contrasted with the harshness of the surrounding rough and relatively barren terrain. Their shorelines, lush with bulrushes and other plants, provided rich habitats for birds, mammals, reptiles, amphibians, fish, insects, and shelled creatures. Native Americans lived along the lakeshores. Mono Lake, east of the Sierra Nevada, and Walker Lake, in west-central Nevada, are two modern desert lakes that give some sense of what the vanished pluvial lakes were like.

Much of the water for desert lakes came from melting snow and ice in nearby high mountains, as well as from local rain and runoff. Such water fed the pluvial Mojave River, which nourished Lake Manix. Fifteen thousand years ago, the San Bernardino Mountains received heavy snows that created a handful of small glaciers. Spring and summer melting sustained a perennial Mojave River all the way to Death Valley. Today, the Mojave River is an ephemeral stream that fills much of its river bed when the snow melts. Even so, people who live along it must reckon with floods when an unusually warm spring follows a winter with heavy snow in the mountains. Such floods carry water all the way to Silver Lake playa north of Baker, flooding that community en route. Several feet of water can accumulate in Silver Lake and linger for the better part of a year. The historical record is ten feet during the winter of 1916, an abnormally wet year. No historical records tell of the Mojave River having flowed all the way to Death Valley, but the path it followed north from Silver Lake during pluvial times is clear. Water in Silver Lake must be 36 feet deep to overflow the sill at its north end and spill into the old channel the Mojave River once followed to Death Valley.

The Mojave River is deceptive. Even in the driest time of year, when no rain has fallen for months and the surface of its bed is bone dry from the mountains to Victorville, water flows through the narrows at Victorville. California 18 crosses it there, as does Interstate 15 a little farther north. The coarse sand in the river's bed is so pervious that water percolates into and through it easily. The river goes underground in dry periods, and its water moves slowly downstream through the porous sand in its bed. This is hard on fish, but it conserves water by slowing runoff and reducing loss from evaporation. At Victorville narrows, a nearly impermeable mass of granite forces the water to surface. Within a mile or two downstream, it soaks back into the streambed, except during floods when surface flow extends far into the desert.

Beyond Yermo, about 15 miles east of Barstow, both Interstate 15 and the Mojave River, which follow roughly parallel courses north of Victorville, enter the broad valley that pluvial Lake Manix once flooded. The highway crosses the abandoned lake bed eastward for 19 miles, to the Afton exit.

At its maximum filling, Lake Manix covered about 85 square miles and had three large bays. Coyote Arm was north of the present location of Interstate 15, tucked in behind the Calico Mountains. Troy Arm was farther south, along where the Santa Fe Railroad and Interstate 40 are today. Afton Arm was where Interstate 15 is today. People rushing to the gaming tables of Las Vegas travel through the heart of the Lake Manix basin and Afton Arm.

What evidence tells of a lake that once occupied this basin? The extremely flat valley floor is suggestive, but many desert basins, never occupied by lakes, have flat floors. Fortunately, the Mojave River has cut 100 to 200 feet into the floor of Afton Arm exposing the underlying deposits. You can see that the uppermost beds are mostly fine sand, silt, and clay—typical lake deposits. They rest, locally unconformably, on older lake beds and stream-laid sand and gravel, deposits that are not part of the Lake Manix sequence, although they accumulated in the same basin. These pre-Lake Manix beds contain at least three ash layers, of which only the lowermost—the 2.2 million-year-old Huckleberry Ridge ash from Yellowstone—is correlated and dated with confidence.

The one identifiable ash layer within the Lake Manix beds is correlated with a 185,000-year-old tuff in the southern Sierra Nevada. Its position near the bottom of the sequence indicates Lake Manix came into existence about 190,000 years ago.

The riverbank exposures are not visible or easily accessible from Interstate 15. But gullies draining to the Mojave River did erode through the lake beds in two places along the highway. Eastbound travelers get the better view. One exposure is just south of the highway and north of the parallel railroad tracks 5.2 miles east of the Harvard Road exit, where the

Oblique air view south of large beach ridge (center) along east end of Afton Arm of Manix, near Afton exit on Interstate 15. —From *Geology Illustrated*, by John S. Shelton, W. H. Freeman Company, copyright © 1966.

Platy beach pebbles in a small excavation near the south end of Afton beach off Interstate 15. Knife is 5½ inches long.

highway dips gently into a broad swale. The lake beds are soft, fine-grained, and slightly greenish. Westbound travelers should watch for this site 2.2 miles west of the Field Road exit, as they start down into the swale. A larger exposure of lake beds, clearly visible to motorists going in either direction, is 3.5 miles east of the rest stop and 0.8 mile west of the Afton Road exit. You can reach it on the frontage road west from the Afton exit.

Other souvenirs of pluvial Lake Manix include lake shoreline features such as wave-cut cliffs, terraces, and beaches. To see the very best beach, take the Afton exit, and then drive a hundred yards or so south onto a wide and level ridge. The ground along the ridge's western edge and the face of the ridge are littered with smoothly worn, flat pebbles, some nicely circular. These are typical beach pebbles. They would make good skipping stones, if we had some water. The ridge is a magnificent beach built by large waves driven east up Afton Arm by the powerful westerly

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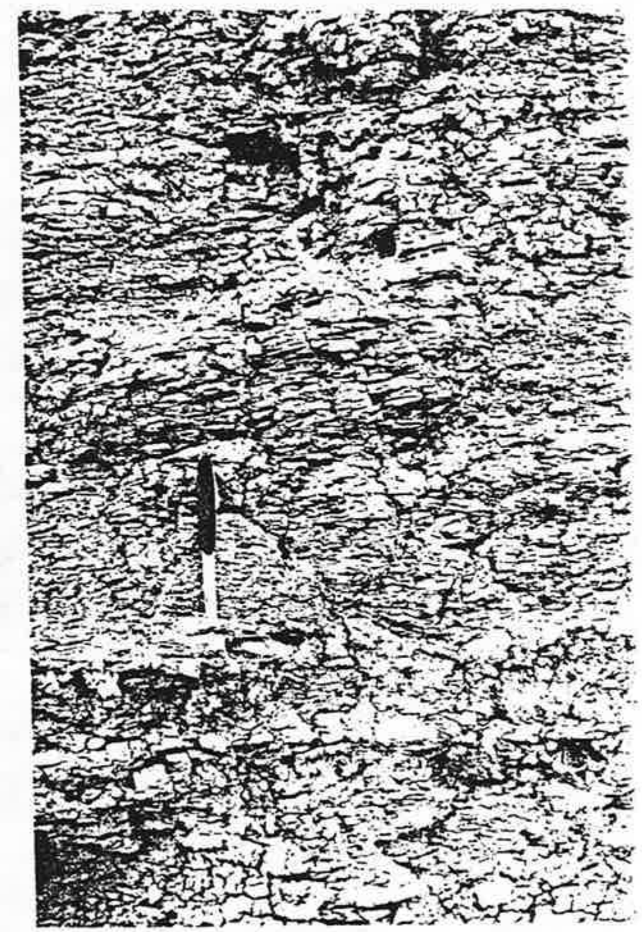
winds. This beach ridge is part of the highest young shoreline of Lake Manix. Its size indicates that the water stayed at that level for a long time, presumably because the outlet stream flowed over hard rock that it could not erode. That high outlet may have been through Troy Arm. About 15,000 years ago, an outlet developed through Afton Arm, a little east of here. Water pouring across soft rock quickly carved Afton Canyon, putting the presumed Troy Arm outlet out of business and sending water through Soda and Silver Lake playas and Silurian Dry Lake to Death Valley.

We have discussed only the latest events in the history of Lake Manix. If you hiked into gullies off the frontage road, you probably noticed layers of brownish sand and gravel within the deposits of lake clay. They are stream deposits, which show that Lake Manix shrank dramatically at times, perhaps disappeared. Age dates on the ash layers within the older lake beds, indicate that water flooded the basin at least 185,000 years ago,

Lake clays in gully bank just south of frontage road, west of Afton exit from Interstate 15. Pocketknife is 5½ inches long.



long before the beach ridge at Afton exit formed. The record in the landscape tells only of the last chapter in a long and complicated history.

The fossil remains of animals and birds that lived along the shores of Lake Manix are preserved within shoreline deposits. Animals represented include dogs, cats, bears, horses, camels, antelope, bison, sheep, and mammoths. Most were grazing animals. Shoreline birds were abundant, including storks, pelicans, cormorants, grebes, ducks, geese, eagles, cranes, and even two species of flamingos. Imagine them standing on one slim leg, solemnly surveying the Lake Manix scene. It really happened. Under the flamingos' feet were snails, beetles, fish, and an occasional turtle. This was a thriving community.

The Mojave Desert must have been a rather inviting place during pluvial times, 15,000 and more years ago. It was a land of lakes, streams, and wildlife, a land that locally supported grass, flowers, bushes, and even small trees such as piñon pines and junipers. Its size and the strong westerly winds would have made Lake Manix a great place for windsurfing—had any of the local denizens been so inclined.

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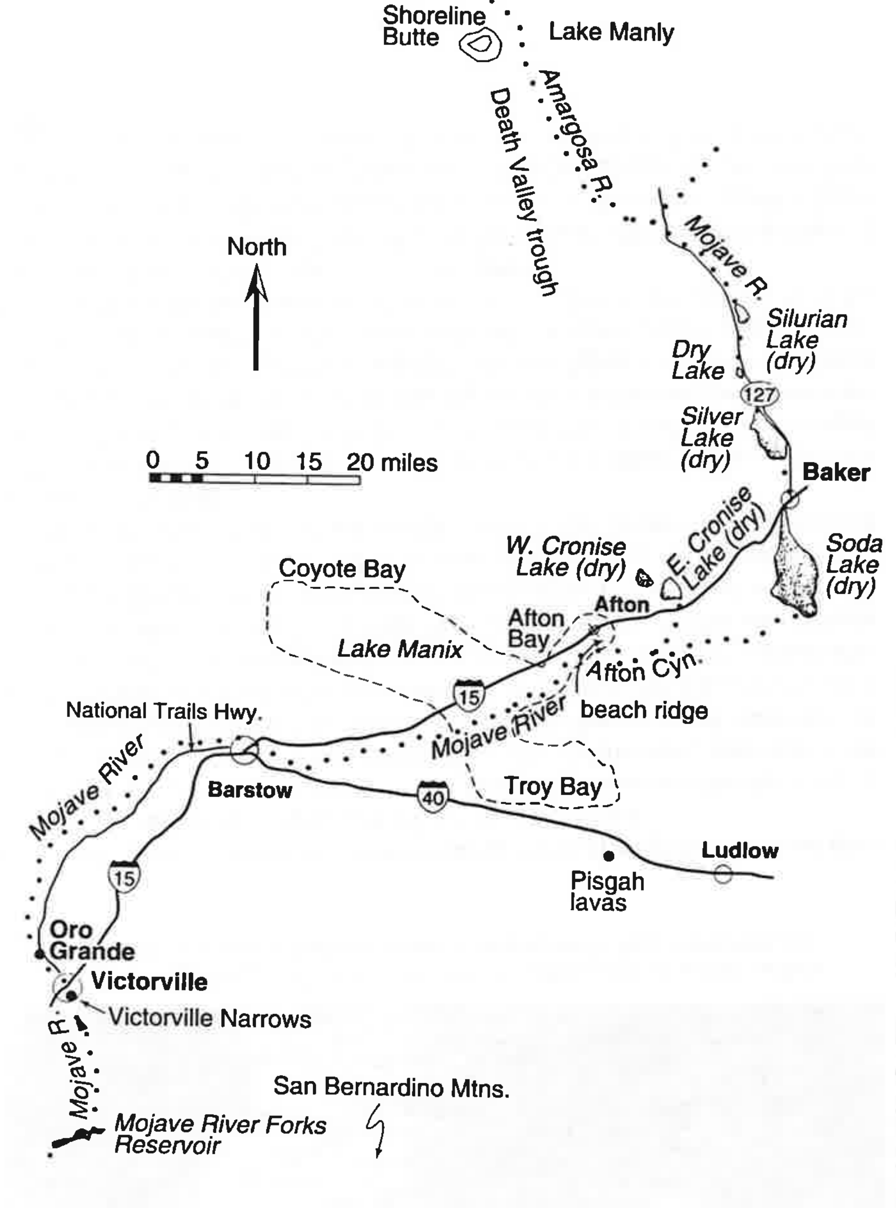


The dotted line traces the course of the Mojave River to Death Valley.

ETTING THERE: Travelers to Death Valley on Interstate 15 proceed to Baker and then go ıth on California 127. The route follows a course closely parallel to the channel of the ıcial-age Mojave River, when it flowed from snow-covered 8,000-foot peaks in the San rnardino Mountains to Death Valley, the ultimate sump of California deserts at 282 feet low sea level. Features along the river's former route can be viewed along I-15 and ılifornia 127 on the way to Death Valley, some from main highways and others by way of ort detours.

1

An Intrepid Explorer

— THE MOJAVE RIVER —

Southern California's highest mountains, the San Bernardinos (with 11,499-foot San Gorgonio Peak), spawn two good-sized rivers, the Santa Ana and the Mojave. The Santa Ana has the larger discharge but flows only 65 miles southwest from the mountains into the ocean near Newport Beach. Today, the Mojave flows, during times of heavy flooding, north and northeast 135 miles from the San Bernardino Mountains to Silver Lake playa, just north of Baker. In glacial times, roughly the last million years, ice and deep snows in the San Bernardino Mountains fed the Mojave River, and it filled Silver Lake to overflowing before continuing to Death Valley, a total journey of 225 miles. Along the way, the river's water cut deep rock gorges, languished for months in placid lakes, breached the sills of lake basins, filled other basins with sediment, and provided food and attractive lakeshore campsites for Native Americans as well as habitats for fish, birds, and mammals.

Let us follow the course of the Mojave River, beginning where it emerges from the Mojave River Forks Reservoir, at the north base of the San Bernardino Mountains. This reservoir captures floodwaters for controlled release to users downstream. The riverbed beyond the reservoir is exceptionally wide and sandy, so water flowing along it, unless in flood quantity, generally percolates into the pervious sand. Underground flow is an efficient mode of water transport, especially in deserts, where evaporation is high and surface runoff often catastrophic and wasteful. Most downstream water users drill wells to tap into the slow-moving groundwater of the river's bed. The subsurface course of the river thus serves as a storage system far less expensive and more efficient than any artificial reservoir. All this is hard on fish, and beyond the foot of the mountains, the Mojave River is not a favorable home for trout.

About 50 miles north of the San Bernardino Mountains by river course, I-15 crosses the Mojave River's channel on a long bridge at the north edge of Barstow. Before the river's impoundment at the reservoir, travelers on I-15 were occasionally treated to the spectacle of the normally dry riverbed filled to the brim with a roaring torrent of muddy water with high waves. Heavy rains or a winter of exceptionally heavy snow in

Narrows of the Mojave River at Victorville, looking upstream where surface water flows year-round because a granite spur dams the flow of groundwater.

the mountains followed by unseasonably warm spring weather can generate large and enduring floods of the Mojave River.

During summer when most of the riverbed between the mountains and Victorville, 15 miles downstream of the mountains, appears bone dry, surface water flows continually through the bedrock narrows on the east edge of Victorville. You can best approach the narrows by exiting from I-15 at Stoddard Wells Road on the east edge of Victorville. Follow Stoddard Wells Road less than a mile south to the old bridge across the Mojave River. Viewing the downstream (northern) end of the narrows from the old bridge is more safe than viewing it from the heavily traveled modern Apple Valley bridge, which lies between the narrows and the old bridge.

At these narrows, the river flows in a steep-walled gorge—140 feet wide, 150 feet deep, and 1,000 feet long—cut into a rocky granitic spur projecting west from the mountains. Upon seeing the narrows, you may wonder why a river would cut a gorge into hard granite when it appears that the river could have detoured around the spur's west end and easily carved a channel through alluvial gravel. The river simply had no choice but to cut the gorge; it was trapped.

In a depositional period, the Mojave River and other streams flowing north from the San Bernardino and San Gabriel Mountains built a huge alluvial apron far into the Mojave Desert, to and beyond Victorville. The apron banked against the slopes of bedrock hills east of Victorville, completely burying spurs that projected west from them. When this depositional stage of the Mojave River ended, a geologic event, perhaps a climatic change that increased the river's discharge or tectonic tilting that increased the slope of its bed, initiated an erosional stage of the Mojave River. As it eroded, the river deepened its channel. Soon, the river could no longer escape from its banks. When the entrenched river encountered the top of a buried granite spur, it had no choice but to continue cutting down. Following the path of least resistance, the river carved as narrow a channel as possible, which required the least amount of work and involved carrying the least amount of debris. Meanwhile, up on the alluvial apron, erosion continued to eat away at the gravel that buried the spur, eventually uncovering the spur and leaving it as a ridge, which increased in relative height and extent as the erosional stripping progressed. The downcutting of the channel and the erosion of the gravels continue today.

Why does water flow on the surface in the narrows but not upstream from it? Along the upper Mojave River, water-saturated sands lie close to the riverbed surface. The subsurface water flows slowly through the sands until it meets the impervious granite spur, which forces the water to the surface. The thin veneer of channel sands in the narrows cannot accommodate all of the water so it flows on the surface year-round for a mile or so downstream, passing under the I-15 bridge. By looking sharply right or left as you cross the I-15 bridge, you can see the stream. The water percolates back into the sandy bed farther downstream of the

Railroads use the gap cut in a formerly buried granite spur by the superimposed Mojave River at Victorville.

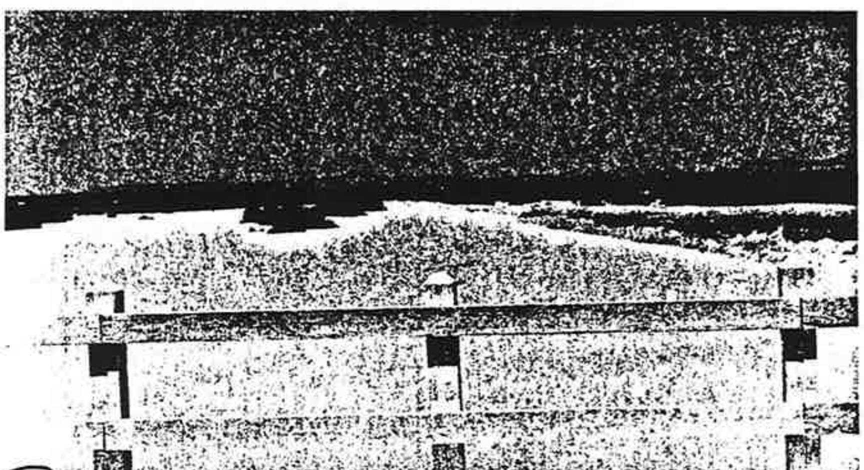
spur but returns to the surface 3.5 miles above the Oro Grande cement plant. There, the river flows through a second bedrock narrows cut into a formerly buried spur of granitic rocks that projects west from Quartzite Mountain. The water usually disappears again within a mile or two, but it flows close to the surface in places, such as Palisades Lake and Silver Lake Roads, 4.5 miles downstream from the Oro Grande cement plant. The groundwater table is shallow all along the riverbed, as little as 20 feet below the surface in places, but deeper farther from the main channel. Mojave River water, mainly pumped from wells, sustains a broad belt of farmland and habitation from Victorville to beyond Barstow.

Between Victorville and Barstow, 35 miles by river, the streambed is mostly wide, sandy, and dry. Interstate 15 follows a straighter course and rejoins the river on the north edge of Barstow. Old Route 66 (National Trails Highway) closely parallels the river, and carries less traffic.

At Barstow, both the river and I-15 turn northeastward, the highway following a path several miles to the north of the river. In glacial times, the river encountered its first ponded water here in Lake Manix, which covered about 85 square miles to a maximum depth of nearly 200 feet. The lake had three large arms: Coyote to the northwest, Troy to the southeast, and Afton to the northeast. Much Mojave River water lies underground in the western and central parts of the Manix basin, where numerous wells tap it for irrigation of fodder crops. Airline passengers are startled by the green fields lying in the midst of arid desert terrain east of Barstow.

Geologists speculate that Lake Manix's initial outlet drained east-southeastward from the Troy arm down the wide trough extending from

The 250-foot-wide, sandy bed of the Mojave River at the Hinkley Road bridge, about 14 miles upstream from Barstow.



Barstow to Bristol Lake playa near Amboy. From Bristol Lake playa, water probably flowed southeast through Cadiz and Danby Lake playas to the Colorado River near the present-day town of Blythe. Currently, the Santa Fe Pacific Railroad and the National Trails Highway closely parallel the river's route as far as Amboy, and Interstate 40 does likewise as far as Ludlow. At one time, the Santa Fe Railroad capitalized on the old stream-graded segment of the Mojave River between Cadiz and Danby by operating a spur line from Chambless southeast into Arizona, now a decommissioned route.

Rivers are adept at seeking out low areas within rough terrain and have a neat way of finding the lowest path around an obstacle. They simply form a lake, and the water level within this impoundment rises until it reaches the lowest outlet. The weakness of this procedure is that the lowest point does not necessarily lead to the lowest areas within the terrain beyond. In the case of the Mojave River, the lowest point nearby was Death Valley, not the Colorado River and the Gulf of California. The Mojave River received one benefit, however, from its detour to the Colorado: an enriched fish fauna.

About 15,000 years ago, the Mojave River's course changed dramatically. The Troy outlet was blocked either by lava flows, possibly from Pisgah volcano or, more likely, by movement on the Pisgah fault. The water level in Lake Manix rose until it overtopped its rim at the east end of the Afton arm. The large outflow quickly cut a narrow, 5-mile-long gorge at what is now Afton Canyon through which the Mojave River extended its course toward Death Valley. Railroads prefer water-level routes, and the Union Pacific Railroad route currently follows Afton Can-

View looking south at Afton Canyon, the outlet channel of Lake Manix. Light, horizontal lines across midphoto are high tension wires. Near-horizontal lineation on foreslope is the surficial expression of layers in the underlying sedimentary deposits. —Helen Z. Knudsen photo

yon on its way to Las Vegas. Water flows year-round through parts of Afton Canyon for the same reason that it does so at Victorville Narrows: impermeable bedrock in the channel floor.

With the carving of Afton Canyon, Lake Manix emptied, and the Mojave River cut into the sediment accumulated on the lake's floor. You can see such eroded beds of lake sediment southwest of the Afton off-ramp. To see a magnificent Lake Manix beach ridge, take the Afton off-ramp and stop in the parking area at the top of the ridge just to the south of I-15. The level, rounded crest of the ridge and the many flat, smooth, beach pebbles on its flanks reflect its origin. Strong winds from the west created huge waves on Lake Manix, especially near Afton where winds have a long over-water fetch. The breakers picked up and reworked sand and stones along the shoreline. The waves carried away most of the fine particles, leaving behind worn, smooth stones flattened by sliding back and forth in the wash. The heavy surf flung many of these stones above the normal water level, where they accumulated as a broad, rounded beach ridge extending parallel to the shoreline.

Upon emerging from Afton Canyon, the Mojave River flows east on a broad, gently sloping alluvial plain, much of which the river built by depositing its sediment load. There, its channels are shallow, braided, and constantly shifting. Some floodwater occasionally runs north into Cronise basin, passing under I-15 at the Basin off-ramp. This water can accumulate in East Cronise Lake playa and also, at times of highest flooding, on West Cronise Lake playa. You can go to the alluvial plain at the

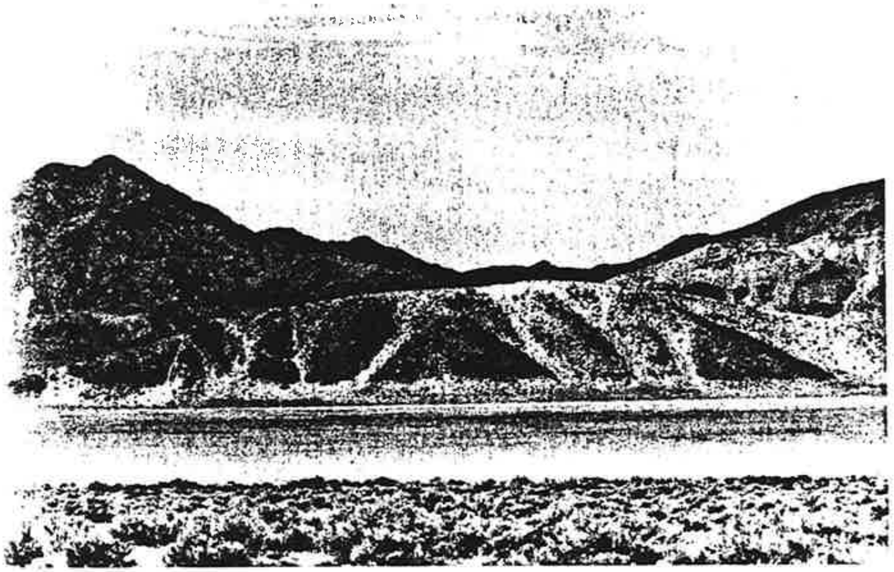
Beach pebbles of a Lake Manix beach ridge at Afton. —Helen Z. Knudsen photo

View looking northeast to flat-topped, gullied Lake Manix beach ridge at Afton off-ramp, south of Interstate 15.

mouth of Afton Canyon from the Basin off-ramp, but four-wheel driv and high clearance are advisable owing to road roughness and accumu lations of wind-drifted sand.

The largest floods on the Mojave River send water farther east int Soda Lake playa, which connects northward to Silver Lake playa by channel through the center of Baker. Baker has suffered considerabl discomfort from Mojave River floods because at least once a decade th water goes all the way to Silver Lake. The greatest historical water dept recorded at Silver Lake is 10 feet in 1916, an unusually wet year, bu several feet of water have accumulated in other years, for example abou 3 feet in 1969. Water depth in Silver Lake must be above 35 feet for th lake to overflow, which it did in glacial times. Then, Baker would hav been largely submerged, except for the upper part of its giant thermon eter, and Silver and Soda Lakes would have joined to form a larger bod Lake Mojave. The 10 feet of water in Silver Lake in 1916 came uncon fortably close to inundating Baker, thereby alerting residents to the po tential danger of large floods. Volunteer workers made an informal e fort to deepen the outlet channel at the north end of Silver Lake, but th sill there consists of hard bedrock, and the volunteers abandoned th excavation before attaining an effective length and depth.

Native Americans found the shores of Lake Mojave an attractive plac to live, and professional and amateur archaeologists have collected mar artifacts along the now-abandoned shorelines. Today, collecting is pr hibited for the amateurs. From California 127 along Silver Lake's ea side, you can see traces of old shorelines cut into the base of hills we of the lake, especially when shadowed in the late afternoon.



A wave-cut cliff near the base of the hill west of the north end of Silver Lake playa. Windblown silt emphasizes the strandline and fills in gullies. –Helen Z. Knudsen photo

After glacial Lake Mojave overtopped the sill at Silver Lake, it continued north toward Death Valley. The river soon ponded in two areas along the east side of the Avawatz Mountains: at a small playa off the north tip of the Soda Mountains, simply named Dry Lake, and farther north in the larger Silurian Lake playa (that's its name, not its geologic age), west of the Silurian Hills. Water in these lakes was shallow, and since neither basin has a stable, enduring bedrock outlet—necessary for the formation of strong shoreline features—strandlines of these lakes are faint and poorly preserved.

From Silurian Lake, water flowed north into a minor pond nestled against the east base of the Salt Spring Hills. You can see fine-grained, pale pond sediments and a faint shoreline cliff there. Again, water was shallow but deep enough to overflow a saddle in a narrow ridge of granitic rock that projected west from some higher hills. The water cut a narrow gorge, draining the pond and leading to dissection of pond sediments. A cluster of green salt cedar trees, which currently marks the head of this gorge, benefits from ponding of groundwater behind the impervious spur, 0.5 mile east of California 127.

Once across the last spur, the Mojave River had an easy run of about 3 miles across alluvium to a junction with the Amargosa River, which still flows into Death Valley. Waters of the combined rivers contributed greatly to glacial Lake Manly (vignette 5). As the level of that lake rose, the river's debouchment into the lake moved south down the Death

The Mojave River cut this gap in a granite spur of the Salt Spring Hills. View northwestward. –Helen Z. Knudsen photo

Valley trough. By the time the lake was about 250 feet deep, it had reached the vicinity of Shoreline Butte. Strandlines on the butte show that the lake subsequently became deeper, and its shore receded much farther south. Lake Manly seldom, if ever, had an outflow across a stable bedrock sill, so its level rose and fell with variations in the climatically controlled water supply. A small change in water level caused a large shift in where the Amargosa–Mojave River entered Lake Manly. The level of the lake seemingly never stabilized long enough to allow the combined rivers to construct a significant delta; at least none has been recognized for certain.

It is no mean feat for a river of modest size and limited water to forge a 225-mile-long channel across a desert. The Mojave River persevered in its journey, obeying the law of gravity that governs the flow of water. The river took advantage of a relatively lofty, 8,000-foot origin and simply traveled in search of lower ground—a journey that culminated at the lowest dry land on the North American continent.