

Bob,
Best wishes,
Frank

Neogene evolution of the Indian Wells Valley, east-central California

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ABSTRACT

The Indian Wells Valley located in east-central California has been an active sedimentary basin off and on throughout Tertiary time. During the Paleogene, the Indian Wells Valley was part of the Goler basin, and in the Miocene, the valley received volcanoclastic and terrigenous sediment correlated with rocks of the Ricardo Group. In the Pliocene, the Indian Wells Valley became the locus of deposition in a half graben formed by low-angle normal faulting along the eastern front of the Sierra Nevada.

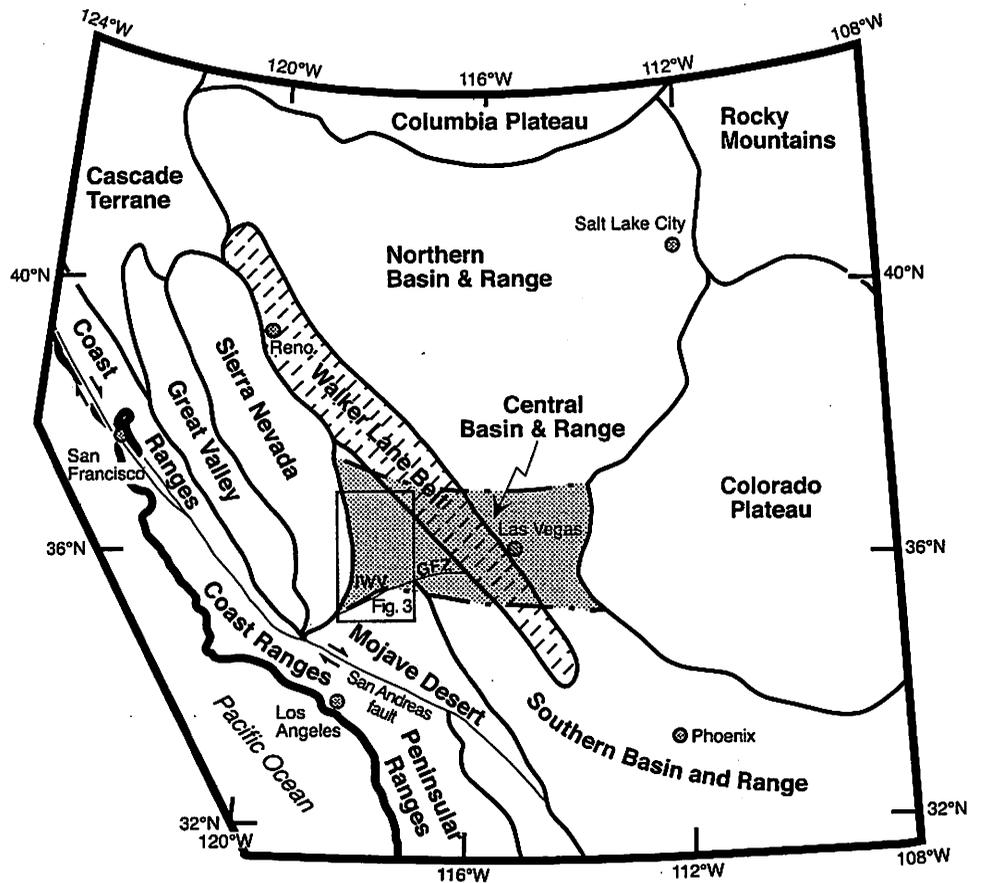
Two significant structural and tectonic changes are recorded in the stratigraphy of the Indian Wells Valley: one at the close of the Miocene and the second at the end of the Pliocene. The late Miocene transition occurred sometime between 7 and 5 Ma and resulted in the uplift of the nearby Sierra Nevada and the formation of a half graben. Transition from a regime of east-west extension to one dominated by transtensional dextral faulting occurred in the late Pliocene, marking a change in regional stress, the cause of which is, as yet, undocumented. Dextral strike-slip faulting dominates the modern structural setting and controls the sparse sedimentation in the valley. This pattern of extension followed by transtension appears to be part of a progression that began in Death Valley ca. 16 Ma and moved westward. Today the Indian Wells Valley accommodates a component of integrated transtensional shear and is part of the evolving margin between the North American and Pacific plates.

INTRODUCTION

The Basin and Range province has been the focus of studies on continental extension since 1975 (e.g., Wernicke, 1992). One of the most intensely studied parts of this region is the central Basin and Range, which encompasses much of the area from the Colorado Plateau to the Sierra Nevada at the latitude

of Las Vegas (Fig. 1). Because of the excellent exposure and numerous geologic studies here, the timing and magnitude of Tertiary extension in this area is fairly well understood (Wernicke et al., 1988; Serpa and Pavlis, 1996). One important area that remains poorly known and not well integrated into the regional picture for the central Basin and Range province is the Indian Wells Valley, located at the southwesternmost corner of

Figure 1. Map of the western United States showing the major tectono-physiographic provinces. The central Basin and Range designation acknowledges the uniqueness of the highly extended area around, and to the west of, Las Vegas. The Walker Lane belt is closely tied to physiographic and structural features; for a summary discussion of the Walker Lane belt, see Stewart (1983). The area covered by Figure 3, and the location of Indian Wells Valley (IWV) and the Garlock fault zone (GFZ), are shown in the box. Adapted from Wernicke (1992).



the central Basin and Range (see Fig. 1). Understanding the evolution of the Indian Wells Valley is important because of its location in the broad zone of transtensional shear in eastern California. The valley is bounded on the south by the Garlock fault, long recognized for its key role in the accommodation of differential extension between the Basin and Range and the relatively unextended Mojave Desert region (Davis and Burchfiel, 1973). Unraveling the geologic history of the Indian Wells Valley provides valuable information about the Cenozoic evolution of the Pacific–North American plate boundary and the partitioning of strain along the eastern front of the Sierra Nevada (Wesnousky and Jones, 1994).

Modern seismicity is abundant in the Indian Wells Valley; an average of 5000 events occur there annually. Between August 1995 and January 1996 there were three earthquakes of >5.0 magnitude. First motion for these events is consistent with the differential movement between the Pacific and North American plates. Epicenters define a north-northwest–trending belt of seismicity through the area. Nur et al. (1993) speculated that the 1992 Landers earthquake sequence, and related earlier dextral events in the Mojave Desert, represented the latest evidence of an eastward shift in the plate boundary. The seismic activity in the Indian Wells Valley fits well with this model.

This paper presents many new data and interpretations for

the structure, stratigraphy, and tectonic setting for the Indian Wells Valley and surrounding areas throughout the Neogene. We first review the geologic and tectonic setting of the central Basin and Range, and then we present results of geophysical and geological work we have conducted over the past several years. We conclude with an integrated model of the evolution of the Indian Wells Valley from Miocene to the present and speculate on the future of the basin in the context of continued large-scale extension and transtension in the southwestern Basin and Range.

BACKGROUND

The southwestern part of the central Basin and Range is a triangle-shaped area bounded on the west by the Sierra Nevada batholith, on the south by the Garlock fault, and on the east by the Walker Lane belt (Fig. 2). Greensfelder et al. (1980) recognized the uniqueness of this area as shown in the obvious topographic features, which they attributed to a greater extensional strain rate than affected surrounding areas. They referred to it as the Owens Valley block although that is a misnomer because it covers a wider area than just the Owens Valley. In this paper we refer to it as the “Owens Valley–Death Valley extended terrane” because the term both encompasses the most

widely recognized geographic landmarks found there and describes the most salient tectonic feature. It should be emphasized, however, that the area described includes the Indian Wells, Panamint, Owens, Saline, and Death Valleys and the intervening mountain ranges.

The Owens Valley–Death Valley extended terrane exhibits unique topographic features that set it apart from the rest of the Basin and Range. Figure 2 shows that topographic elements in this area are irregularly spaced, and there is a distinct north-northwest to northwest orientation to the ranges. Basins and ranges are wider by a factor of two than their counterparts to the east and north, and relief is much greater. Although the average elevation within the northern Basin and Range is higher, both the highest (4418 m) and the lowest (–70.7 m) points in the contiguous 48 states are found within the southwestern area. These two points are separated from one another by only 120 km, attesting to the dynamic nature of the forces currently acting on the area.

Regional structural and tectonic setting of the Owens Valley–Death Valley extended terrane

For the most part, present-day deformation within the Owens Valley–Death Valley extended terrane is dominated by

Figure 2. Digital topographic map of the southwestern United States. The Walker Lane belt separates the "traditional" Basin and Range from either the Basin and Range itself or any of the areas to the west. is characterized by physiographic elements that are distinctly different border. The Owens Valley–Death Valley extended terrane (OVDVET) west approximately along the northwest-trending Nevada–California Northern Death Valley–Furnace Creek fault zone, OVF–Owens Valley Mountains, HMFZ–Hunter Mountain fault zone, NDV–FC FZ–Funeral Mts. Valley, BM–Black Mountains, CW–Coso Wash, EPM–El Paso Mesa. The area covered by Figure 4 is shown by the box.

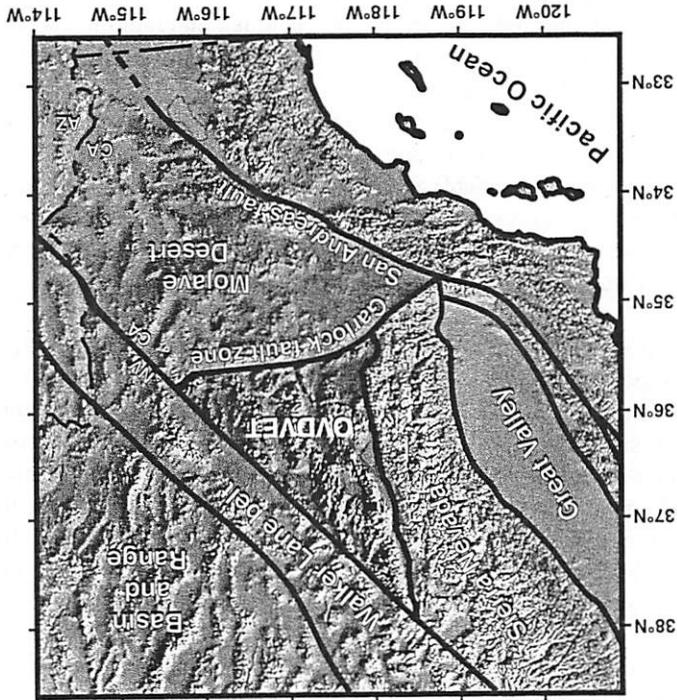
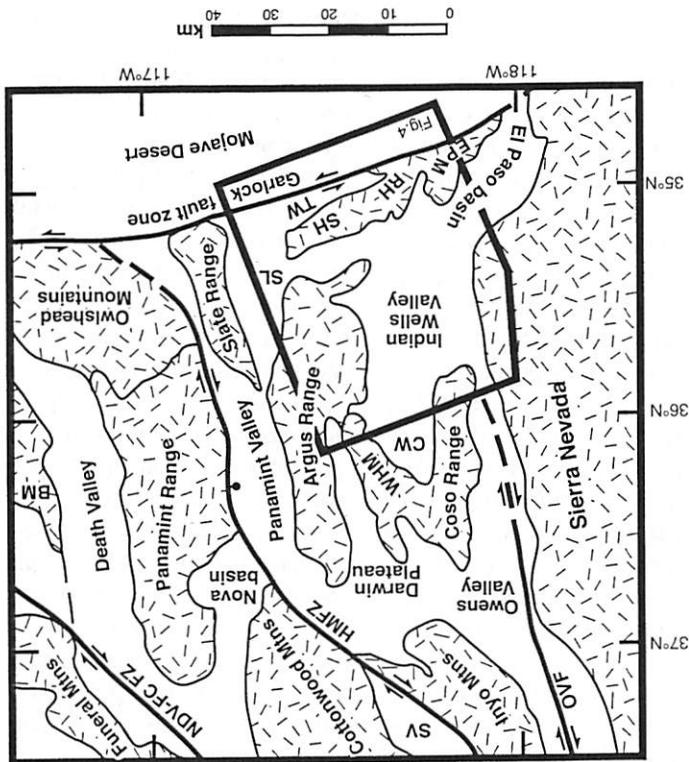


Figure 3. Map of the study area and the surrounding geographic features. BM—Black Mountains, CW—Coso Wash, EPM—El Paso Mesa. The area covered by Figure 4 is shown by the box.



Owens Valley–Death Valley extended terrane in the tectonic framework of the region by proposing that it was an accommodation zone that reconciled deformation in the Walker Lane belt with northwest-directed movement of the Sierra Nevada batholith. This concept of accommodation has been supported by the work of other investigators (e.g., Wernicke et al., 1982; Stewart, 1983; Hodges et al., 1989; McKenna and Hodges, 1990; and Walker and Glazner, 1999) and will be further tested in this paper.

Wernicke and Snow (1998) proposed that in middle Oligocene time, the Sierra Nevada lay ~320 km closer to the Colorado Plateau than at present. Snow and Wernicke (2000) presented a detailed kinematic model of intraplate deformation across the central Basin and Range showing that most of the Neogene deformation occurred in the Owens Valley–Death Valley extended terrane and the Lake Mead terrane to the east. This area of intraplate deformation has been called the eastern California shear zone by Dokka and Travis (1990), who proposed that between 10% and 15% of the total Pacific–North American plate-boundary deformation was accommodated by right-lateral shearing in a broad zone extending from the Mojave Desert eastward into Nevada. Although it is clear that this dextral shearing is distributed over the central and northeast Mojave Desert eastward to the Death Valley fault zone, it is not clear how it is accommodated at the Garlock fault nor how shear is transferred northward into the Owens Valley–Death Valley extended terrane.

Westward progression of extension in the Owens Valley–Death Valley extended terrane is also well documented (Wernicke, 1992) with onset occurring in Death Valley itself at ca. 16 Ma (Fitzgerald et al., 1991; Gans and Bohrsen, 1998). By ca. 6–5 Ma, extension had begun to wane in southern Death Valley (Davis et al., 1993), and the locus had shifted westward to Panamint Valley and the northern Argus Range (Fig. 3, Hodges et al., 1989; Schweig, 1989; Bacon et al., 1982). Deformation in Panamint Valley was accommodated along a low-angle normal fault that makes up the western slope of the Panamint Range (Fig. 3, Conrad et al., 1994). Strata in the Nova basin, located at the north end of the Panamint Range, were deposited as a result of this extension. Schweig (1989) documented the onset of extension west of the Darwin Plateau at 6.0 Ma, and Bacon et al. (1982) proposed that extension was initiated in the Panamint and Owens Valleys by 6–4 Ma. By ca. 4 Ma, tectonism was focused in the Coso and Saline Ranges (Schweig, 1989). Transtension began in the Saline and Panamint Valleys after 3.7 Ma (Zhang et al., 1991; Conrad et al., 1994; Hodges et al., 1989) with onset of right-oblique displacement along the Hunter Mountain fault (Burchfiel et al., 1987) and vertical movement on the east face of the Inyo Mountains.

Farther west, Bachman (1978) suggested that the eastern Sierran escarpment began to form between 3.4 and 2.3 Ma on the basis of stratigraphic correlation of lacustrine sediments in the central Owens Valley. Step faulting of the Wild Horse Mesa area of the Coso Range, reflecting the onset of major extension

of Coso Wash, began between 3 and 2 Ma (Duffield and Bacon, 1980; Duffield et al., 1981; Whitmarsh et al., 1996). This extension coincided with a hiatus in volcanism that separates Pliocene calc-alkalic (mostly intermediate) volcanism from later bimodal volcanism that began at ca. 2.0 Ma (Duffield and Bacon, 1980). On the basis of the foregoing, the locus of extension was in the Indian Wells Valley and the Owens Valley during the late Pliocene and into the Holocene. Regional stresses during this time were accommodated by a series of faults reaching from the Garlock fault on the south to Bishop in the north and eastward to the Walker Lane belt. The pattern of distributed strain in this area is complex and is the subject of other, neotectonic studies (e.g., Reheis and Sawyer, 1997; Humphreys and Weldon, 1994).

The Indian Wells Valley

The focus of this paper is on the development of the Indian Wells Valley, which is situated at the southeastern terminus of the Sierra Nevada. It is located directly north of the Garlock fault at the extreme southwest corner of the Owens Valley–Death Valley extended terrane (see Fig. 3). The valley is nearly as wide (30 km) as it is long (35 km), making it anomalous for the central Basin and Range. Adjacent to the Indian Wells Valley and extending along the southern front of the Sierra Nevada is a small northeast-southwest-oriented valley, referred to locally as the El Paso basin.

The Indian Wells Valley is bounded by the Sierra Nevada on the west, the Argus Range on the east, the Coso Range on the north, and the relatively low-relief Spangler and Radermacher Hills on the south. For the most part, these higher-elevation areas are composed of Mesozoic plutonic basement rocks typical of the Sierra Nevada. The Coso Range has an areally restricted, but significant volcanic cover consisting of basaltic and rhyolitic flows and associated pyroclastic rocks. Basalt flows extend south from the Coso Range into the northwestern part of the valley.

Modern drainage from ranges surrounding the Indian Wells Valley is internal; there are no perennial streams feeding the valley. For the most part, the surface of the valley is covered with recent alluvium of either lacustrine (playa) or fluvial origin (Fig. 4) with minor eolian deposits. Sedimentation is dominated by alluvial fans emanating from canyons of the Sierra Nevada on the west and the Argus Range on the east. Two small playas occupy the southeast part of the valley, and there are small inliers of older alluvial material that protrude through the modern sedimentary cover on the northern, northwestern, and eastern sides of the valley.

Previous geologic studies of the valley have primarily focused on understanding the hydrologic conditions that control groundwater availability and recharge (e.g., Berenbrock and Martin, 1991; Berenbrock and Schroeder, 1994; Kunkel and Chase, 1969; Moyle, 1963). The only combined geological

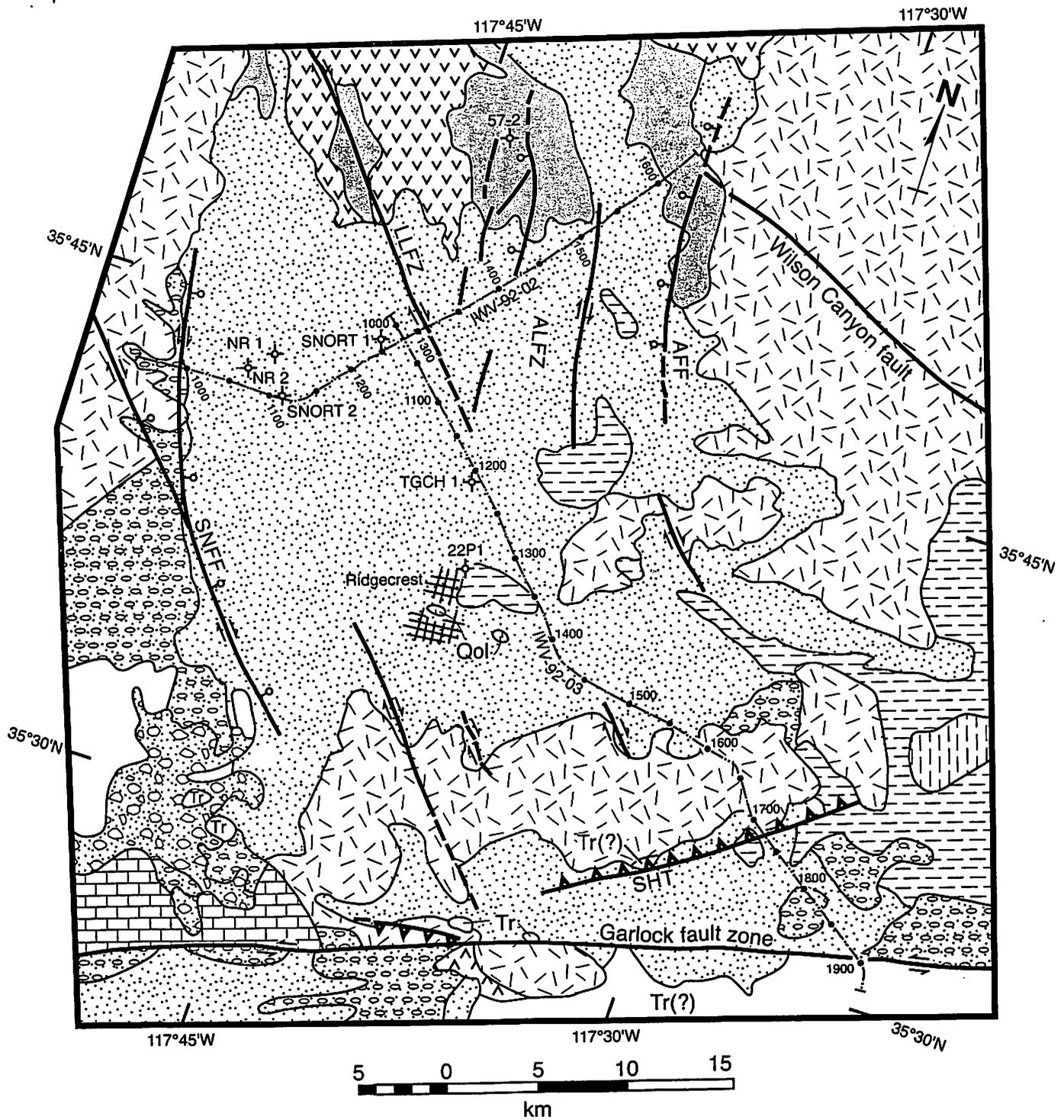


Figure 4. General geologic map of the Indian Wells Valley and surrounding areas. Heavier lines represent faults, dashed where approximate. Dash and double-dot lines are locations of seismic reflection lines discussed in the text and shown in Plates 2, 3, and 4; numbers represent every hundredth shot 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; some are unpatterned on the map): Pzm—Paleozoic metamorphic rocks, Mzp—Mesozoic plutonic rocks (undifferentiated), Tg—Paleocene-lower Eocene Goler Formation, Tr—Ricardo Group consisting of lower Miocene Cudahy Camp Formation and middle to upper Miocene Dove Spring Formation, Tv—Miocene Lava Mountains volcanic rocks, Tal—Pliocene White Hills sequence, Qpv—Pliocene-Pleistocene volcanic rocks of the Coso Range, Qoa—older alluvium for which there are no conclusive data to permit assignment to a specific formation, Qol—older lacustrine rocks for which there are no conclusive data to permit assignment to a specific formation, Qal—Quaternary alluvium, Ql—Quaternary lacustrine deposits. Basic geology from Jenkins (1963) with some areas remapped by Monastero.

and geophysical study of the Indian Wells Valley was conducted by von Huene (1960), who examined the structural geology of the basin by using gravity data and a limited amount of refraction seismic data. Zbur (1963) acquired seismic refraction data over seven lines in the Indian Wells Valley that provided valuable constraints for the subsurface geology of the valley. However, these data have three limitations. First, the lines are not all interconnected; second, not all of the profiles were shot in both directions; and, finally, two of the profiles were quite short. Roquemore (1981) and Roquemore and Zellmer (1987) mapped surface features associated with neotectonic activity along two major fault zones in the Indian Wells Valley: the Little Lake and Airport Lake fault zones. Both have been active since 1980 and accounted for a significant percentage of the total seismicity recorded in California in 1995 and 1996.

Monastero and Katzenstein (1995), and Monastero et al. (1995) discussed geophysical, petrophysical, and geological data supporting the hypothesis that there has been substantial extension and crustal thinning in the Indian Wells Valley. The work presented in this paper expands on those previous efforts and provides a more definitive interpretation for timing of tectonic and sedimentary events related to the formation of the valley.

Local stratigraphy. The documented stratigraphy for the area in and around the Indian Wells Valley includes deposits ranging in age from Paleocene to Holocene (Fig. 5). The El Paso Mountains rise nearly 1600 m above the southwest corner of the valley floor and are the site of the most significant outcrops of Paleogene and Neogene rocks. The core of this range consists of Paleozoic metasedimentary and metaigneous rocks and Mesozoic plutonic rocks, which are overlain by Cenozoic sedimentary and volcanic rocks of the Goler Formation and the Ricardo Group. The total measured section for these rocks is ~4000 m (Loomis, 1984; Cox, 1982; Cox and Diggles, 1986).

The Goler Formation was first described in detail by Cox (1982) and ranges in age from Paleocene to early Eocene (Cox and Diggles, 1986). It rests disconformably on highly eroded, irregular, and weathered plutonic basement and, for the most part, consists of nonmarine conglomerate, sandstone, siltstone, and shale, with characteristic reddish-brown to dark brown color. Rocks were deposited in an east-west elongate basin whose axis was more or less coincident with the present location of the Garlock fault (Cox and Diggles, 1986). Coarser facies consist of alluvial-fan- and alluvial-plain-type deposits that formed on a southward-sloping piedmont adjacent to a mountainous area to the north (Cox and Diggles, 1986). The principal environment of deposition was an internally draining alluvial basin with marginal fans and broad outwash plains covered at times by meandering braided streams.

The Ricardo Group consists of two formations: the lower to middle Miocene Cudahy Camp Formation and the middle to upper Miocene Dove Spring Formation. These rocks were deposited in the El Paso basin (Loomis, 1984; Loomis and Burbank, 1988) and crop out today in the El Paso Mountains, al-

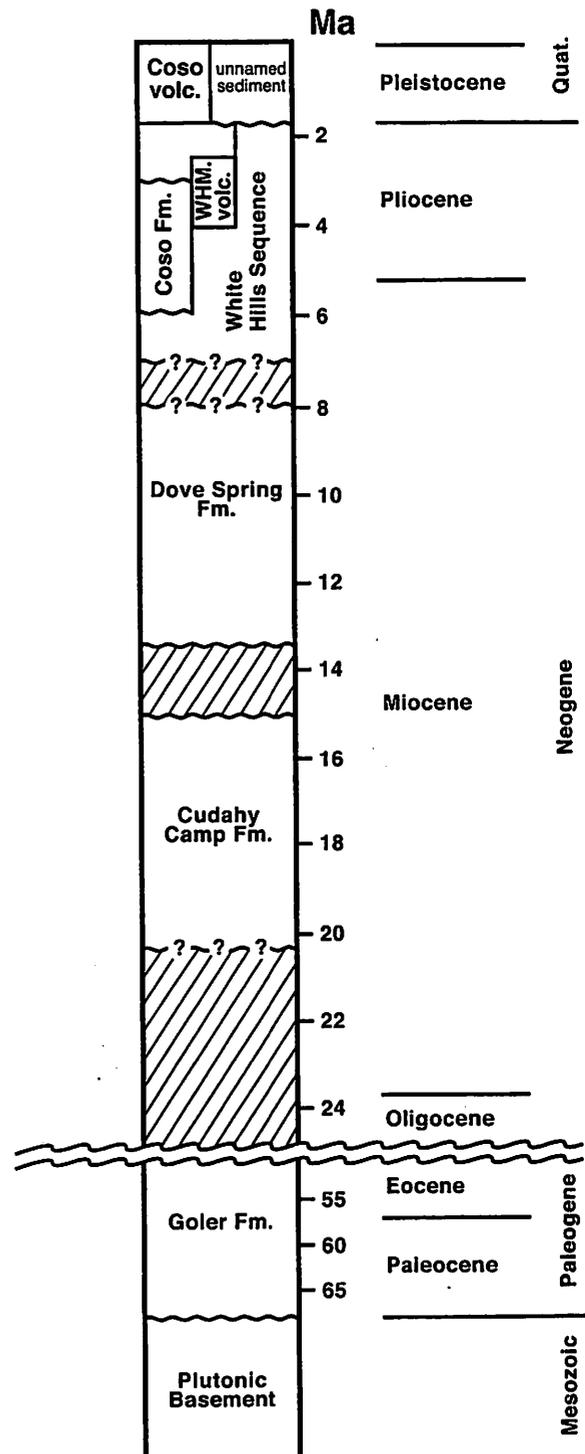


Figure 5. Composite stratigraphic column for the Indian Wells Valley area. WHM refers to the Wild Horse Mesa calc-alkalic volcanic suite. The Coso volcanic rocks are the bimodal suite that constitutes the youngest volcanic rocks in the Coso Range. Age ranges for the Goler Formation are from Cox (1982); for the Cudahy Camp and Dove Spring, they are from Loomis and Burbank (1988) and Whistler and Burbank (1992). Note the change in the time scale between the Eocene and Oligocene.

though remnants of the Cudahy Camp Formation are exposed in Teagle Wash south of the Indian Wells Valley (Figs. 3 and 4).

The Cudahy Camp Formation ranges in age from 20 to 15 Ma (Loomis and Burbank, 1988). It consists primarily of volcanic tuffs, lava flows, and epiclastic rocks that were deposited in an east-trending basin whose axis, like that of the Goler basin, approximated the present location of the Garlock fault. Monastero et al. (1997) showed that the source of these rocks was a volcanic field that was situated south of the El Paso basin in early Miocene time. These authors showed that the full 64 km of offset on the Garlock fault must be restored in order to properly juxtapose correlative rocks in the volcanic field with those of the Cudahy Camp Formation and that there was no evidence at 17 Ma of the existence of the Garlock fault.

In the middle Miocene (ca. 16 Ma) there was a major change in regional tectonics with the advent of significant northwest-directed crustal extension (Wernicke and Snow, 1998). If the Garlock fault is an intercontinental transform as suggested by Davis and Burchfiel (1973), then movement on the fault had to have commenced at that time. This movement may have been the cause of a hiatus in deposition from 15.1 to 13.6 Ma in the El Paso basin that is reflected in an angular unconformity between the Cudahy Camp Formation and the overlying Dove Spring Formation (Loomis, 1984). The Dove Spring Formation was deposited from early Miocene to Pliocene (?) (Loomis and Burbank, 1988; Whistler and Burbank, 1992) in a basin that first subsided, then was simultaneously rotated and translated, and finally was tilted and uplifted.

Deposition in the lower part of the Dove Spring section was dominated by alluvial-fan and alluvial-plain sediments that were derived from the south and southeast (Loomis and Burbank, 1988). Several tuffs and basalt flows that have been radiometrically dated by us as well as others (Cox and Diggles, 1986; Whistler and Burbank, 1992) occur in the section. The upper parts of member 4 and member 5 of the Dove Spring Formation reflect a shift of depositional style and provenance to fluvial-lacustrine with a source area to the north and possibly northeast (Loomis, 1984; Loomis and Burbank, 1988). Younger units of the Dove Spring Formation have a high proportion of angular and subangular alkali feldspar and quartz grains, indicating a nearby granitic source, probably a rejuvenated Sierra Nevada. Biostratigraphy and radiometric dating of tuffs by Whistler and Burbank (1992) established an age of ca. 7 Ma for the youngest Dove Spring Formation rocks (member 6). The top of the Dove Spring Formation is an angular unconformity; the Dove Spring is capped by a Holocene nonmarine sedimentary unit that has yielded mammal bones with radiocarbon ages between 19 000 and 16 000 yr (D. Whistler, 1999, personal commun.).

Pliocene rocks of the Indian Wells Valley are limited to the White Hills sequence. These rocks were described by Kunkel and Chase (1969), who did a detailed measured section through the upper part of the unit. They determined that the rocks were lacustrine in origin from the occurrence of freshwater diatoms,

very fine grained clastic sediments, and thin ostracod-bearing limestone beds. They observed that the White Hills sequence was older than the unnamed Pleistocene and Holocene volcanic rocks that flow south from the Coso Range to partially cover the outcrops in the northwest corner of the valley.

Kunkel and Chase (1969) also described outcrops in the northwest corner of the Indian Wells Valley that they referred to as Quaternary older alluvium (Tal; Fig. 4). These outcrops are undeformed to moderately deformed lenticular deposits of semi-indurated silt, sand, gravel, and boulders. They have an open framework structure, are poorly sorted, and are matrix supported with angular to subangular clasts derived from nearby plutonic and metamorphic sources (Fig. 6). There is no evidence of volcanic clasts in the outcrops. The total exposure is more than 240 m thick, but the maximum thickness is unknown. These deposits project eastward ~1 km to a position beneath the same basalt flows that cover the White Hills sequence. Outcrops beneath the basalt flows measure up to 50 m in thickness. Because of the stratigraphic position relative to the basalt flows, it is likely that these alluvial deposits are a facies of the White Hills sequence and are grouped as such.

GEOPHYSICAL AND GEOLOGICAL DATA

The rocks just described project into the subsurface of the Indian Wells Valley with a limited number of outcrops on the periphery of the valley. We have used a combination of geological and geophysical techniques to image these rocks and to relate those images to specific formations. Two principal types of data are presented in this study: reflection seismic data collected in 1992 and geophysical and lithologic logs from four deep exploration holes drilled on, or in close proximity to, the seismic lines. Other sources of data are regional gravity measurements augmented by finer-scale surveys to verify local features, surface geologic mapping, and seismicity data.

The reflection seismic data referenced in this study were acquired by using the parameters listed in Table 1. The survey design was based on the objective of imaging the Moho, so the far offsets were nearly 9200 m, and the sweep range was gauged for optimal penetration to the desired maximum record length of 8 s two-way traveltime (TWTT). The data are two-dimensional and are subject to all of the pitfalls associated with processing and interpreting such data, e.g., noise from out of the plane of the section, multiples, etc. They are, however, high-quality 180-fold data with a good signal-to-noise ratio.

Reflection seismic data were processed in two separate ways to achieve different objectives (Table 2). Initial processing focused on clearly imaging the basement contact, gross structural features, mid-crustal reflectors, and the top of the Moho, if possible. In the first round of processing, the final band pass was a 12–48 Hz time-invariant filter that was selected to preserve and enhance deep reflectors. By using this low-frequency band pass, we were able to more clearly image the basement contact, but sacrificed shallow detail in the predominantly sed-

Figure 6. Outcrop of the debris-flow facies (Tal) of the White Hills sequence in the northwest Indian Wells Valley along Highway 395. These rest nonconformably on Sierran basement and are representative of the very coarse clastic deposits found in the bottom of SNORT 2. Note the open framework, poor sorting, and matrix support of clasts. Hammer (center) is 28 cm long.



TABLE 1. FIELD DATA ACQUISITION PARAMETERS FOR INDIAN WELLS VALLEY REFLECTION SEISMIC DATA

Shot By	Halliburton Geophysical Services, Party 1776
Date Shot	October 1992
Recorder/Format	MDS-18/SEG-D
No. Channels	362
Field Filter	7.5/12-125/90 Hz
Geophone Array	24 phones/158 feet, in-line
Energy Source	Vibroseis
Source Array	4 Vibes, in-line, move-up over 161 feet
Sweep Freq/Type	12-72 Hertz, linear
No. Sweeps/Length	10 sweeps/10 seconds
Field Sample Rate	2 milliseconds
Data Length	8 seconds
Shot Point Interval	165 feet
Group Interval	165 feet
Spread Geometry	30,112.5-412.5-X-412.5-30,112.5

imentary section. The second round of processing was designed to better image shallow stratigraphic features. Long-offset data were muted to eliminate normal-moveout stretch effects, two rounds of residual statics analysis were performed, a shot-domain dip moveout was applied, and the final band-pass filter was time varying (see Table 2). These processing steps resulted in preservation of a significant amount of shallow stratigraphic detail that is the underpinning of our interpretation.

The basic gravity data set was assembled from government, university, state, and local sources. We acquired additional data by using a Scintrex model CG-3M gravimeter that provides repeatability to $\pm 1 \mu\text{gal}$. These new data were especially helpful in addressing specific questions associated with interpretation of seismic line IWV-92-02.

Since 1993, four exploratory holes have been drilled in the Indian Wells Valley (see Fig. 4 for locations) by the Geothermal

Program Office (GPO) of the U.S. Navy for the purpose of gaining information about the subsurface geology of the valley and to test the potential for exploitable geothermal resources. The holes are designated SNORT (Supersonic Naval Ordnance Research Track) 1, which was drilled to 2437 m; SNORT 2, drilled to 3050 m; TGCH 1, drilled to 749 m; and White Hills 57-2, which reached 914 m. SNORT 1, SNORT 2, and TGCH 1 all penetrated plutonic basement rocks; White Hills 57-2 bottomed in older lacustrine sedimentary rocks.

Resistivity, self-potential, gamma, sonic, and Formation MicroScanner (FMS) geophysical logs were acquired for each hole. Continuous coring of the entire length of TGCH 1, and coring of selected intervals of the other three holes, provided valuable material for detailed stratigraphic and geochemical analysis and geochronology. Nearly 1200 m of FMS images from SNORT 2 were interpreted, resulting in identification of more than 2000 separate interfaces in the interval from 658 to 1865 m (Fig. 7). Approximately ninety percent (90%) of these were judged to be bedding features, and the rest were fractures or faults.

STRUCTURE AND STRATIGRAPHY OF THE INDIAN WELLS VALLEY BASIN

The Indian Wells Valley is filled with sedimentary rocks that are equivalent to all of the formations found in nearby surface outcrops. Seismic reflectors have great lateral continuity and relatively high amplitude, indicating relatively uniform depositional conditions over large areas (Plates 2, 3, and 4). The average depth to basement in the central Indian Wells Valley, as measured from the seismic data and the drill holes, is

TABLE 2. PROCESSING PARAMETERS FOR INDIAN WELLS VALLEY REFLECTION SEISMIC DATA

Low frequency processing
Zero-Phase Correlation
Minimum Phase Conversion
Data Initialization—Generation of Spatial Attributes
Mute
True Amplitude Recovery
Correction for Spherical Divergence and Inelastic Attenuation
Time Variant Deconvolution
Time Variant Equalization
Velocity Analysis
Surface Consistent Statics
Velocity Analysis
NMO Correction
Mute
CDP Stack
Wave Equation Migration
Time Invariant Filter (12–48 Hz)
Time Variant Trace Equalization
High frequency reprocessing
Geometry File Generation
Correction for Spherical Divergence and Inelastic Attenuation
Spectral Whitening
Source Record Predictive Deconvolution
Time Variant Equalization
Common Depth Point Gather
Velocity Analysis
Residual Statics Analysis
Velocity Analysis
Residual Statics Analysis
Application of Weathering Statics
Shot Domain Dip Moveout
Removal of NMO Correction
DMO Velocity Analysis
Application of Datum Statics
Space Variant First Arrival Mute
Common Depth Point Stack
Finite Difference Migration
Space and Time Variant Bandpass Filter (12–70 Hz/12–50 Hz)
Time Variant Equalization

2 km although it is likely as deep as 3 km or more in the northwest corner on the basis of gravity anomalies (Black et al., this volume). The sedimentary section thins to a depositional basin boundary toward the south, but is terminated on the northeast end of line IWV-92-02 by a fault on which there is at least 300 m of vertical offset (see Plate 2). Modern trans-tensional faulting, evident in flower structures seen in the seismic data, overprints the entire stratigraphic section.

Throughout this paper, reference will be made to four key horizons in the reflection seismic sections that have been used in interpretation of the data. From oldest to youngest, they are as follows: A—top of crystalline basement, B—top of older, more lithified rocks including the Goler Formation and the Ricardo Group, C—middle Pliocene basalt, and D—base of a clay layer in Pleistocene (?) sediments. Other local reflectors that highlight important features germane to the interpretation are shown on Plates 2 and 3, but do not have time-stratigraphic

significance, e.g., clinof orm reflectors. In Figure 8, we use a sonic log from SNORT 1 to correlate time in the reflection sections with depth in the stratigraphic section.

IWV-92-02 (Fig. 4 and Plate 2) is the principal seismic line used for describing the structure and stratigraphy of the Indian Wells Valley. The section originates within the Sierra Nevada, extends eastward 7 km where it turns northeast, traversing the width of the valley, and terminates near the western front of the Argus Range. Total length of the line is ~35 km. We use the higher-resolution version of this line because it more clearly depicts key stratigraphic features. The lower-resolution version was used to verify basement picks and to crosscheck fault locations. Key horizons on IWV-92-02 have been tied to IWV-92-03 (Fig. 4 and Plate 3), which extends southward across the valley and provides information on the stratigraphy of the southern margin of the basin.

Gravity data and Indian Wells Valley basin structure

Gravity data indicate that there is a very large (15 mgal), slightly north-south-elongated Bouguer gravity low that occupies the west-central part of the Indian Wells Valley (Fig. 9). This low is bounded by steep (~3 mgal/km) gradients on three sides, and the reflection seismic data indicate that it correlates with a ~3-km-thick sedimentary section. On the south, it opens gradually at a rate of ~1.5 mgal/km into a broad north-north-east-trending trough that merges to the southwest with the El Paso basin. The presence of the Sierra Nevada frontal fault on the west of this low undoubtedly enhances that gradient on that side. The reasons for the steep gradients on the north and east margins are not as obvious but also appear to be fault controlled. The fault located at SP (shotpoint[s]) 1328–1350 on IWV-92-02 (Plate 2) and SP 1057–1086 on IWV-92-03 (Plate 3) has ~400–425 m of down-to-the-west normal offset and appears to contribute to the gravity gradient on the east side of this low.

We conducted a gravity survey directly over seismic line IWV-92-02 as a means of independently verifying the reflection seismic interpretation and to shed further light on the nature of the eastern boundary of the gravity low. These data were modeled to approximate the stratigraphy and structure of the basin (Fig. 10) as defined by the interpretation of the seismic section and the lithologic data from drill holes. Gravity station spacing was 161 m, resulting in 210 data points along the 33.8 km line. Sedimentary units were assigned average density values based on results from drill-hole data and assumptions about correlation with units exposed in surface outcrops. Depths to the tops of units are calculated from root-mean-squared stacking velocities and from formation velocities derived from borehole sonic logs. Locations of faults are taken from the interpreted reflection seismic line. The mafic basement rock shown in the gravity model is necessary to make the calculated gravity measurements best fit the observed data. On the basis of outcrops of

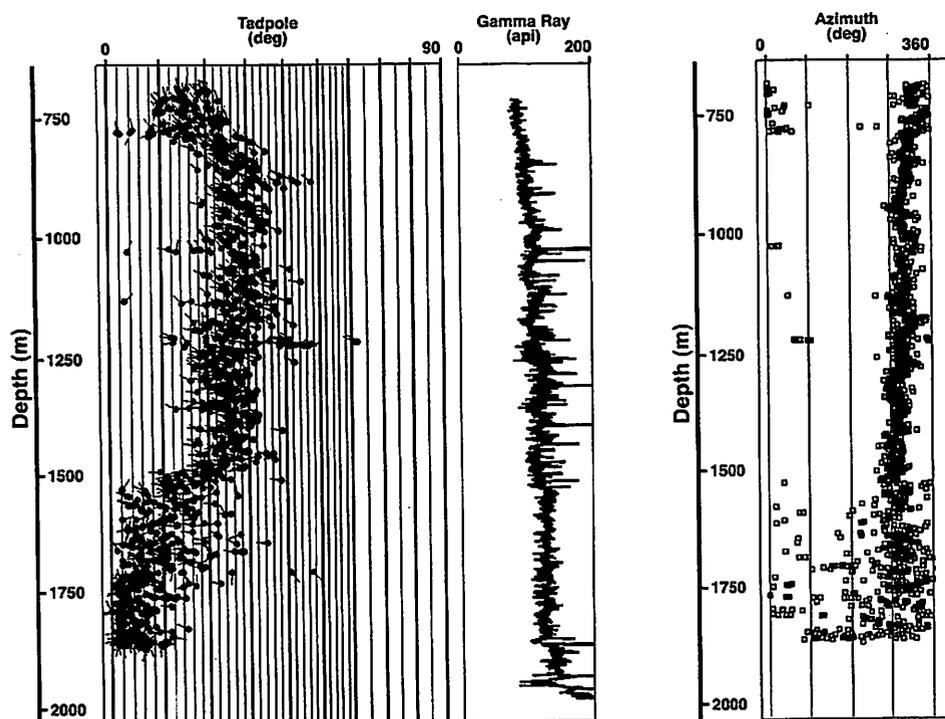


Figure 7. Results of interpretation of Formation MicroScanner images of the SNORT 2 hole. Images are based on electrical resistivity. The plots show the orientation of sedimentary bedding in the interval 683 m to 1856 m, which is the top of the plutonic basement. There are 1380 separate images that have been picked in this interval. The left-hand plot consists of bedding tadpoles with dip represented by the tail and strike by position on the horizontal axis. The curve in the center plot is the natural gamma-ray log. The right-hand plot shows bedding azimuth versus depth. Note the systematic rotation of bedding with depth. The lower part of the section has dips in the 0° – 15° range, and with decreasing depth, the dips increase to 20° – 35° and then go back to the 10° – 20° range. Azimuths are consistently in the 270° – 290° range, but there is a distinctive rotation of dip azimuth to the north with decreasing depth. This type of feature is characteristic of bedding orientation in a growth-fault setting (Hatcher, 1990).

Mesozoic basement in the Sierra Nevada and the Argus Range that consist of plutonic rocks ranging in composition from leucogranite to gabbro, it is reasonable for a mafic rock body to exist in the subsurface of the valley. For instance, the basement core from SNORT 1 is an altered hornblende diorite. The collocation of a strong positive anomaly in our aeromagnetic data (AEROMAG on the CD-ROM accompanying this volume) also supports the assertion that there is a more mafic body buried there. For purposes of modeling, we have assigned a density of 2.9 g/cm^3 to this body, which is reasonable for a hornblende diorite. Gravity models are in excellent agreement with the observed gravity values (green circles in Fig. 10) along IWV-92-02, and the percentage error of calculated versus observed values is less than 1%. Despite the fact that gravity solutions are nonunique, we think that this modeling lends substantial strength to our interpretation of the structure and stratigraphy of the basin.

Sierran basement

Basement in the Indian Wells Valley consists of plutonic rocks of Cretaceous and Jurassic age. Basement was encountered in SNORT 1, SNORT 2, and TGCH 1 where propylitically

altered plutonic rocks were penetrated at depths of 2208, 1856, and 702 m, respectively (Fig. 11). An easily discernible 18–20 Hz doublet reflector characterizes the top of basement on the seismic sections, marking a velocity contrast between the overlying sedimentary rocks (3000–3700 m/s) and the underlying plutonic basement ($\sim 4900 \text{ m/s}$). The basement throughout the Indian Wells Valley has a highly irregular surface, as evidenced on the seismic sections, with a maximum of 1500 m relief.

Sierra Nevada frontal fault. Evidence of the Sierra Nevada frontal fault that resulted in the relief on the eastern Sierran escarpment is found on IWV-92-02 (Plate 2). East-dipping reflectors at SP 990 (300–400 ms) and SP 1054–1080 (1600–1800 ms) have been interpreted as splays of the frontal fault. These reflectors project to the surface at the eastern front of the Sierra Nevada and at a prominent northwest-trending lineament within the Sierran basement rocks (western end of IWV-92-02, Plate 2). The average dip of these features is 25° to the east, although they become more steeply dipping near the surface. Where the reflectors project to the surface within the Sierra Nevada, there is a lineament that separates competent, relatively unweathered granitic basement on the west from less competent, highly fractured basement on the east (Fig. 12). Kunkel and Chase (1969) and Jenkins (1963) mapped faults within the southern Sierra Nevada that are subparallel to the eastern front

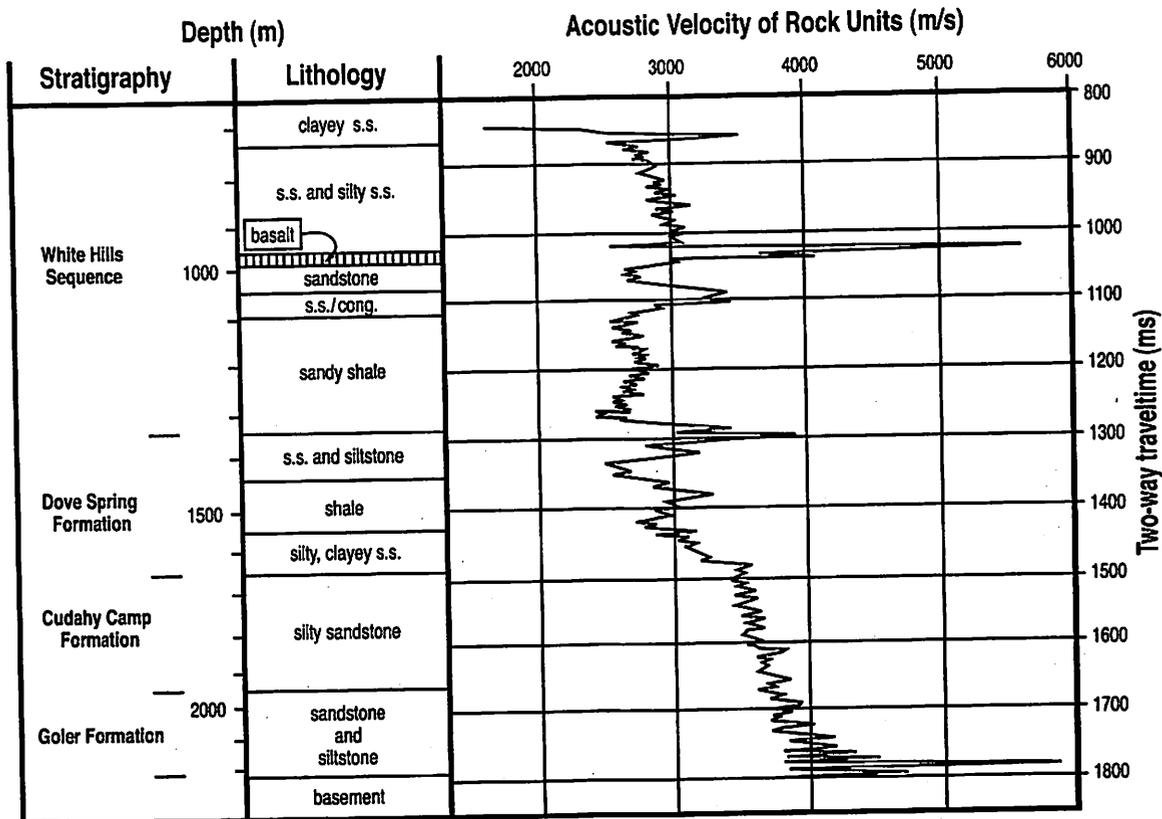


Figure 8. Acoustic velocity profile derived from sonic log of SNORT 1 showing correlation with stratigraphic intervals derived from lithologic log. The steady increase in velocity beginning at ~1550 m and continuing to ~1658 m marks the top of the Ricardo Group rocks. The relatively high velocities of the rocks from that point to the bottom of the log indicate a significant amount of lithification. Horizons A-D correspond to reflection seismic sections; s.s.—sandstone.

of the range coinciding with this lineament. Soil cover on the east side of this lineament makes it difficult to ascertain the precise geologic relationships at this location. However, to the north of IWV-92-02, we found that monolithic megabreccia blocks measuring tens to hundreds of meters on a side are exposed on the east side of the lineament. These blocks consist of locally derived Sierran basement and are interpreted as fault-proximal breccia blocks related to the Sierra Nevada frontal fault (see, for instance, Miller and John, 1999).

When traced along strike to the south, the lineament emerges from the Sierra Nevada and turns to the south-southeast where it joins with the Sierra Nevada frontal fault and continues beneath the sedimentary cover of the Indian Wells Valley (Fig. 4). The presence of the Sierra Nevada frontal fault buried in the southwestern part of the valley was first proposed by Zbur (1963) on the basis of his refraction survey. Later, Kunkel and Chase (1969) inferred a fault in the same location to explain a steep groundwater gradient in local wells. This latter assertion has been substantiated by recent measurements of the groundwater surface that show a > 150 m difference over a distance of 8 km along the location of this buried fault (Indian Wells Valley Water District, 1996, personal commun.). This fault is also the northeast boundary for outcrops of older sedi-

mentary rocks of the Goler, Cudahy Camp, and Dove Spring Formations found in the El Paso Mountains.

Older rocks in the Indian Wells Valley

From the similarity of lithology, degree of induration, and acoustic velocity of rocks in SNORT 1 and the El Paso basin, we infer that horizon B marks the top of a section of sedimentary rocks equivalent to parts of the Goler, Cudahy Camp, and Dove Spring Formations. This horizon—interpreted to be time varying because of differences in depositional conditions and subsequent erosion of parts of the section—occurs over much of the Indian Wells Valley. The basis for selecting horizon B for interpretation was lateral continuity and evidence of an angular unconformity seen in pinching out of reflectors against the bottom of the horizon (e.g., SP 1240–1280, IWV-92-02). This horizon has a strong coefficient of reflectivity in conjunction with a 1400 m/s velocity increase (see Fig. 8), which usually is the result of an increase in lithification and/or change in rock type. To date, no outcrops of Miocene or older rocks have been identified in the Indian Wells Valley.

Goler formation. Evidence of the Goler Formation occurs in both SNORT 1 and TGCH 1, but not in SNORT 2 (Fig. 11).



Complete Bouguer Gravity of the Indian Wells Valley

Countour Interval: 2.5 mgal

Reduction Density: 2.67 g/cm³

Figure 9. Gravity map compiled from more than 20000 individual measurements made in and around the Indian Wells Valley. Location of the gravity profile for seismic line IWV-92-02 (see Fig. 15) is shown in red. The large gravity low in the northwest part of the Indian Wells Valley corresponds to the deepest part of the sedimentary basin.

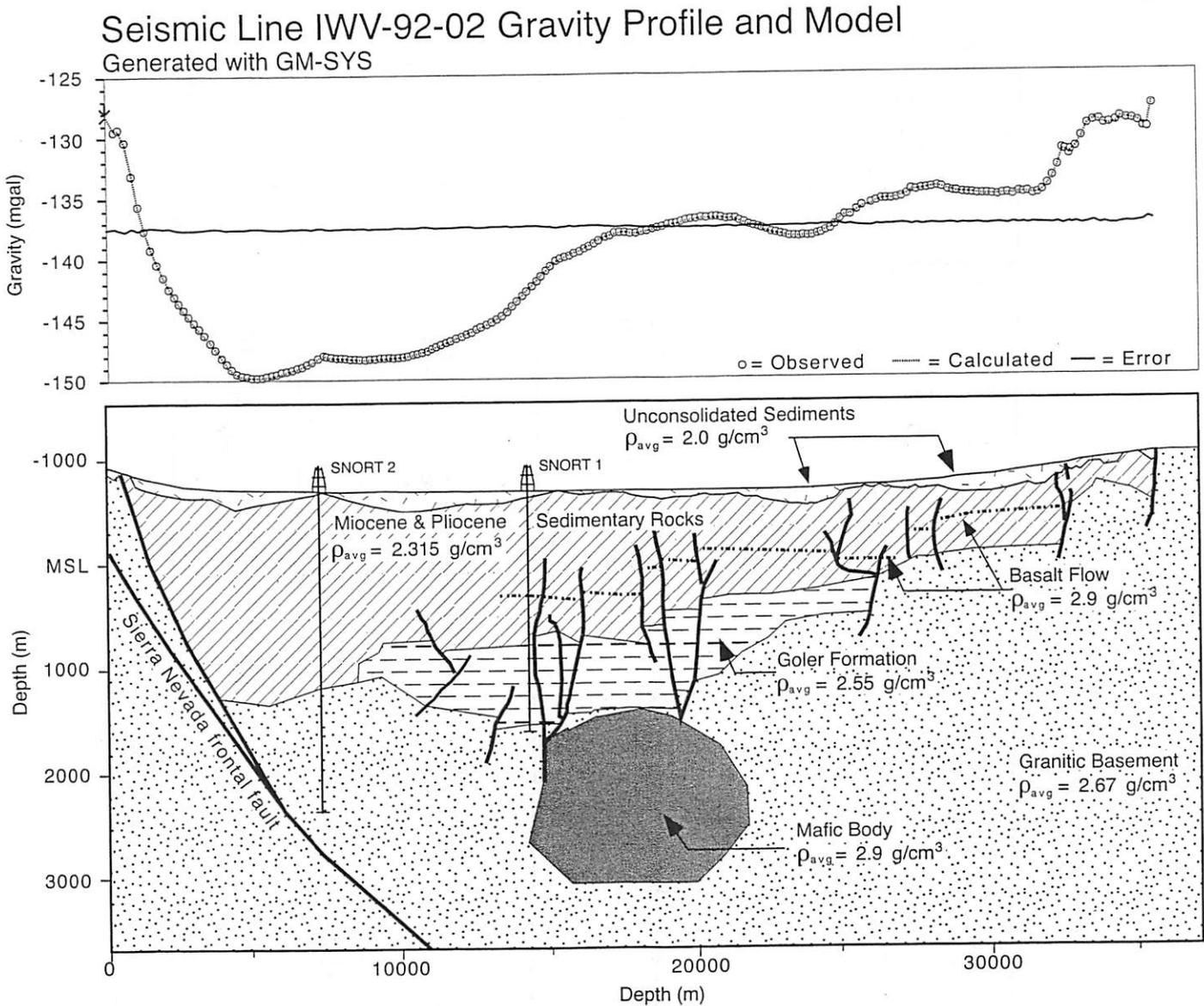


Figure 10. This gravity profile coincides with the location of seismic line IWV-92-02. Modeling was done with GM-SYS software. Densities were derived from samples taken in cores from TGCH 1, SNORT 1, and SNORT 2 and are, in some cases, average densities for thicker sequences. Locations of faults and approximate thickness of sedimentary sections were taken from the interpreted reflection seismic profile. Circles in the upper part of the figure represent actual gravity measurements; the solid line running through the circles is the resulting gravity profile derived from the model (bottom cross section).

On the basis of degree of induration and the fact that the rocks are almost exclusively red beds, the lower 227 m (1981–2208 m) of the sedimentary section in SNORT 1 is assigned to the Goler Formation. Most of the rocks in this interval are sandstone with small and varying amounts of admixed silt. They are red and reddish brown, and shape of the constituent grains is more angular to subangular than those in the units immediately above. Silica cement is prominent, and there are persistent trace occurrences of epidote, calcite, chlorite, pyrite, and plagioclase; black lithic fragments occur in the lower part of the section. There is a steady increase in acoustic velocity from 3350 m/s

at the top of the interval to 3810 m/s at the bottom. These types of acoustic velocities are characteristic of well-indurated sedimentary rocks, an observation that agrees with a reduced drilling rate of penetration in the interval. The stratigraphic section between basement and horizon B varies in thickness with distance northeastward along IWV-92-02 owing to basement relief, distance from source, erosion, and strike-slip displacement.

The lower 100 m of core from TGCH 1 consists of red to red-brown conglomerates with angular to subangular plutonic, metamorphic, and occasional volcanic clasts in a matrix of

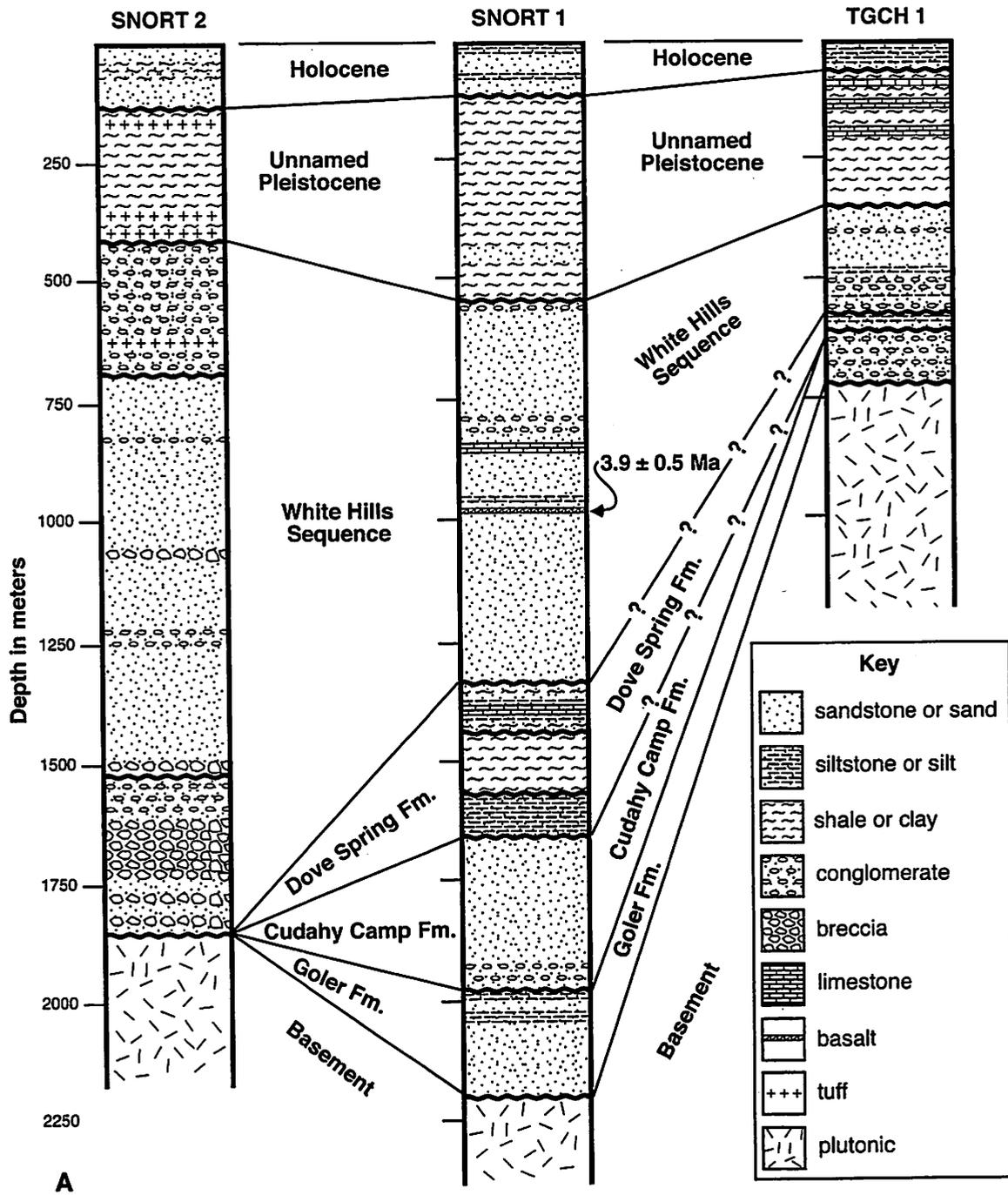


Figure 11. Caption on facing page.

sand- and silt-sized grains. These match outcrops of the Goler Formation in the El Paso Mountains (B. Cox, 1993, personal commun.). In his refraction study of the Indian Wells Valley, Zbur (1963) assigned the stratigraphic unit just above basement to the Ricardo Group because the average formation velocity was in the range of 2774 to 2938 m/s, which he thought was too slow for the better indurated Goler. Core samples from

TGCH 1, which is coincident with one of Zbur's profiles, lead us to conclude that the unit just above plutonic basement is a facies of member 4 of the Goler Formation (Cox, 1982). Additional evidence of Goler rocks in the Indian Wells Valley was reported by Berenbrock and Martin (1991), who described 140 m of lithified continental deposits overlying plutonic basement in well 25/40-22P1 located 5 km south of TGCH 1 (Fig. 4). On

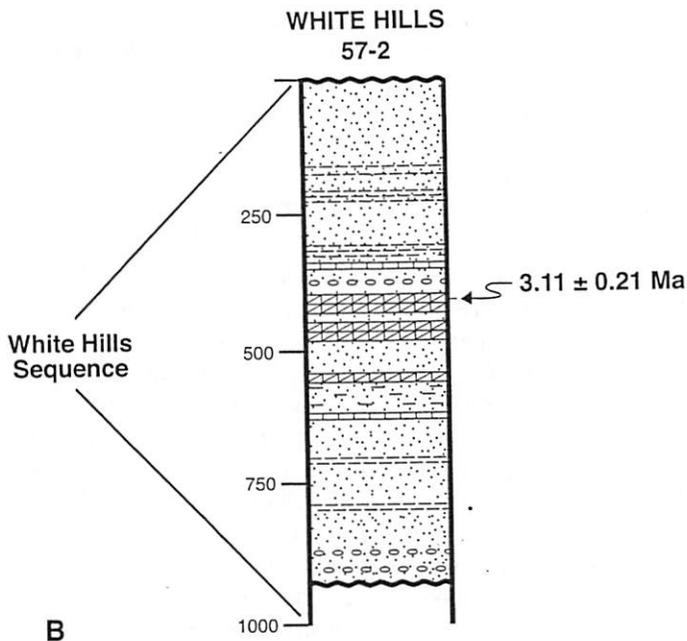


Figure 11. (A) Stratigraphic columns for SNORT 1, SNORT 2, and TGCH 1. Symbols are for both unconsolidated sediments and consolidated sedimentary rocks, i.e., sandstone and sand. The unconsolidated sediments generally occur above the Dove Spring–White Hills contact. The basalt at 1000 m in SNORT 1 is the 3.9 ± 0.5 Ma flow that is horizon C on the seismic sections (Plates 2–4). (B) Stratigraphic column for White Hills 57-2. The hole was spudded in the White Hills sequence and was terminated at 914 m in the same unit. The 3.11 ± 0.21 Ma basalt is located at 418 m.

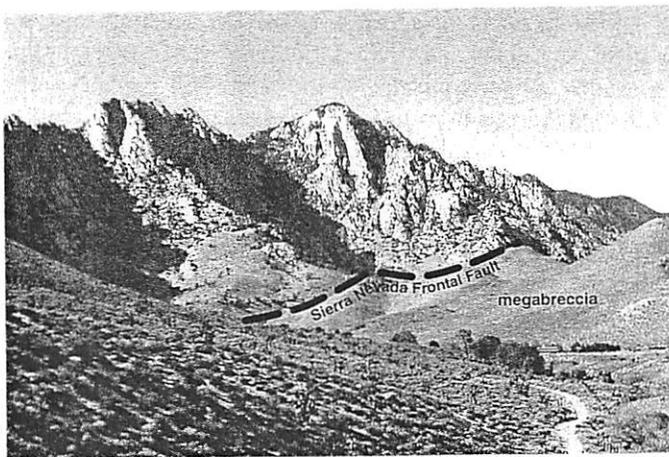


Figure 12. Photograph of southern Sierra Nevada ~2 km west of the end of seismic line IWV-92-02, showing relationship between exposed basement in the background and soil-covered megabreccia blocks in the foreground.

the basis of lithologic descriptions, and close proximity of these continental deposits to TGCH 1, we infer that they are probably correlative with Goler and/or Cudahy Camp Formation rocks.

It is noteworthy that in SNORT 2 there are no occurrences of red or reddish-brown rocks, nor any well-lithified rocks at all, that could be correlated with rocks of either the Goler or Cudahy Camp Formations. In fact, other than the basement itself and basement clasts in the breccia layers there was little competent rock encountered in the hole during drilling. This circumstance indicates that older rocks were either never deposited in this location or they were eroded subsequent to deposition.

Ricardo Group. Evidence of Ricardo Group rocks in the subsurface of the Indian Wells Valley is found in SNORT 1 and TGCH 1. These rocks appear to be distal facies of both the Cudahy Camp Formation and the Dove Spring Formation that crop out in the El Paso Mountains southwest of the valley.

Petrology, color, and texture of rocks in the interval 1341–1981 m in SNORT 1 likely make them distal facies of the Cudahy Camp and Dove Springs Formations although there is no datable material in this section to verify this interpretation. We place the boundary between the two formations at a depth of 1658 m where there is a break in acoustic formation velocity in SNORT 1 that correlates favorably with the 'greater degree of lithification seen in outcrops of the Cudahy Camp Formation (compared to the Dove Spring Formation) in the El Paso basin. From a lithologic point of view, the rocks are fine-grained sandstone, siltstone, and shale throughout most of the section with an abundance of lithic fragments and scattered volcanic detritus. There is no evidence of a granitic source that is common in latest Dove Spring rocks. The sedimentary section beneath horizon B thins in a southeastward direction along IWV-92-03 (Plate 3 and Fig. 4). Core from TGCH 1 (Fig. 11) shows that there is an absence of section between well-lithified red beds at the bottom of the hole, which we assign to the Goler Formation, and overlying poorly lithified rocks that appear to be Dove Spring. This stratigraphic feature leads us to the conclusion that thinning of the section between horizons A and B is due to the disappearance of the Cudahy Camp from the section.

Between 1658 and 1981 m in SNORT 1, most of the rocks are poorly to well-sorted, fine- to medium-grained sandstone, with subangular to round, often frosted, grains. The average formation velocity is ~3810 m/s. There is an increase in lithification and calcite cement in this interval as well as a pronounced increase in iron oxide stain, mica, pyrite, chlorite, and epidote when compared to the next higher interval. This mineral suite is representative of detritus from altered and unaltered plutonic rocks found in the nearby mountain ranges. From 1570 to 1658 m, sand proportion increases, but for the most part there is little change in either the mineral contents or the physical properties of the intervals. There is a uniform 575 m/s upward decrease in acoustic velocity in this interval, which probably reflects decreasing lithification. The interval 1448 m to 1570 m

consists predominantly of shale and siltstone that are mostly green, gray, and less commonly, tan with varying amounts of silt, locally abundant calcite cement, and traces of pyrite and chlorite. From 1341 to 1448 m in SNORT 1, the section consists of a thick sandstone and silty sandstone sequence with local occurrences of limestone throughout (Fig. 11). Drill cuttings show that sand grains are subrounded to subangular quartz with trace amounts of biotite. Grains are typically well sorted and clear to frosted; they range in color from pink to tan to red and red-brown. These rocks compare favorably with units of the Cudahy Camp and Dove Spring Formations described by Loomis (1984).

East of SP 1340 on IWV-92-02, the stratigraphic section between horizons A and B thins, suggesting approach to the depositional margin of the basin. However, at SP 1485, the section abruptly terminates against a high-angle fault, thus preventing that determination. The fact that horizon B is relatively flat in this part of the seismic section indicates that there was no significant syntectonic basin growth on the northeast side of the Indian Wells Valley during this period. Furthermore, the well-developed internal reflectivity in this section suggests that the rocks are likely fluvial or lacustrine, as opposed to alluvial-fan- or delta-type deposits. The shallow depression centered on SP 1400 at 880 ms depth appears to be a broad fluvial channel. This feature, in combination with the absence of clinofold reflector sequences, supports the interpretation of an alluvial-plain depositional environment in the eastern Indian Wells Valley at this time.

Subsurface Miocene rocks in the Teagle Wash. Additional evidence for subsurface occurrence of rocks of the Cudahy Camp and Dove Spring Formations is found on the southeast end of seismic line IWV-92-03 (Fig. 4 and Plate 3). Coring in Searles Lake, 10 km to the northeast of this line, found 915 m of Pliocene and younger sedimentary rocks resting on quartz monzonite basement (Smith et al., 1983). There was no indication of Miocene rocks in that section. Data from the seismic line, however, show evidence of a sedimentary section that is more than 4000 m thick. There are at least two angular unconformities, at 550 ms and 700 ms depth (SP 1820), separating rocks with relatively high interval velocities (3300 to 3500 m/s) in the lower part of the seismic section (700 to 2000 ms) from those with lower interval velocities (1980 to 2440 m/s) in the 550–700 ms part of the section. Loomis (1984) described angular unconformities between the two formations and at the top of the Dove Spring Formation in the El Paso basin. These unconformities, in combination with outcrops of Cudahy Camp rocks a few kilometers southwest of the seismic line in the Teagle Wash, lead us to conclude that this previously unrecognized buried section is correlative to the Cudahy Camp and Dove Spring Formations. It is reasonable, therefore, to assign the lower unconformity (700 ms at SP 1820) to the top of the Goler and the upper one (550 ms at SP 1820) to the top of the Cudahy Camp Formation on the basis of the fact that interval velocities in the lower part of the section are consistent with

well-indurated Goler rocks, whereas those in the upper interval are similar to Dove Spring Formation rocks found in SNORT 1.

White Hills sequence

The stratigraphic unit between horizons B and D is assigned to the White Hills sequence. The age of this sequence is established on the basis of a whole-rock $^{40}\text{Ar}/^{39}\text{Ar}$ date of 3.11 ± 0.21 Ma on a basalt encountered at 418 m in White Hills test hole 57-2 (Fig. 4; Figure DR1¹). An attempt to date a deeper flow found at 469 m was unsuccessful. We also successfully recovered a 3.9 ± 0.5 Ma date on a basalt flow found at 1000 m in SNORT 1 (Fig. 11; Figure DR2 [see footnote 1]) that we have designated horizon C. Correlation of the SNORT 1 basalt flow sequence with the seismic section was accomplished by comparing the acoustic velocity profile (Fig. 8) developed from a sonic log with the lithology log and seismic line IWV-92-02. The depth of the basalt sequence in SNORT 1 corresponds to a large spike in the reflection coefficient and an increase in acoustic velocity of nearly 2500 m/s. By virtue of the similarity in ages of horizon C and the basalt flow within the White Hills section, we are able to correlate the stratigraphic section from horizon D to horizon B on seismic line IWV-92-02 with the rock outcrops in the White Hills anticline. This correlation is the foundation for the subsequent comprehensive discussion of the half-graben depositional setting in the Indian Wells Valley during the Pliocene.

Beneath the basalt flows in White Hills 57-2 there is a 445-m-thick section of sandstone with some limestone, siltstone, and claystone. For the most part, sand grains are rounded to subangular quartz with minor occurrences of feldspar and locally common fragments of basalt. They range in color from gray to yellow orange and are moderately well sorted. In all, the rocks appear to have been deposited in a shallow-lake environment or on alluvial plains.

Western subsurface facies. Besides the lacustrine rocks described by Kunkel and Chase (1969), we define several other facies that are age equivalent to the White Hills sequence. The following discussion focuses on a combination of the nature of seismic reflectors and lithology from drill holes to describe five facies of the White Hills sequence found within the subsurface of the Indian Wells Valley: debris flow and avalanche, alluvial fan, fan delta, lacustrine, and alluvial plain or sheet flood. These facies are associated with the Sierra Nevada frontal fault, and their depositional characteristics appear to be tied to the structure.

Evidence of synsedimentary deposition associated with movement on the low-angle Sierra Nevada frontal fault is seen on the west end of IWV-92-02 between SP 1045 and 1120 (700 to 1400 ms) where reflectors at the bottom of the interval dip

¹GSA Data Repository item 2002104, Figure DR1 and Figure DR2, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA, editing@geosociety.org, or at www.geosociety.org/pubs/ft2002.htm.

to the west, and those at the top of the interval dip eastward. This type of dip reversal is characteristic of growth-fault-type deposition associated with low-angle normal faults (Christensen, 1983). Reflectors are absent or have limited lateral continuity in the seismic section immediately above horizon A (SP 990 to 1120, IWV-92-02). However, reflectors increase in both strength and continuity with distance from the Sierran front and at shallower depths. When this pattern is correlated with rock units found in SNORT 2, we find that reflectors correlate with crudely bedded, coarse breccia admixed with sand and pebbles (Fig. 13A), which dominates the stratigraphic section from horizon A to 1527 m depth (see Fig. 11). Breccias in the lower part of the section are more massive, have larger, angular clasts (i.e., up to 70 cm in diameter), are less well sorted, and are more common than they are in the upper part of the section. There are also numerous examples of scoured surfaces in the FMS images (Fig. 13C). Lithologically these correlate with older alluvium (Kunkel and Chase, 1969) that crops out in the northwest corner of the valley. Megabreccia deposits lie in fault contact with the Sierran basement on the west (Fig. 12), and they are separated from landslide deposits (Fig. 14) by down-to-the-east normal faults. Landslide deposits grade eastward into debris flows that are covered by the Quaternary basalt flows

mentioned earlier. Megabreccia blocks range up to 50–60 m in diameter and are completely engulfed by pulverized, poorly sorted debris and slump-type deposits.

The combination of seismic reflection characteristics, data from the drill hole, and outcrops leads us to conclude that the stratigraphic sequence in the western part of the Indian Wells Valley represents the landslide, slump, and debris-flow facies characteristic of syntectonic deposition proximal to a low-angle normal fault in a half-graben setting (Leeder and Gawthorpe, 1987; Crowell, 1982). The pattern of seismic reflectors on the west end of IWV-92-02 can be interpreted as follows. Slump blocks closest to the normal faults have few internal reflectors because they are the same lithology as the underlying bedrock and there is no internal structure that results in acoustic impedance contrast. Overlying the acoustically transparent slump blocks is a sequence of laterally limited reflectors that pinch out in an eastward direction (SP 1000 to SP 1045). On the basis of the length of the reflectors and their location relative to the sediment source to the west, they probably represent debris flows. These offlap the Sierran basement (see, e.g., 580 ms beneath SP 1017) and are truncated abruptly on their upper surfaces. Weak internal reflectivity is the result of stacking of debris flows and crude layering caused by hydrodynamic effects

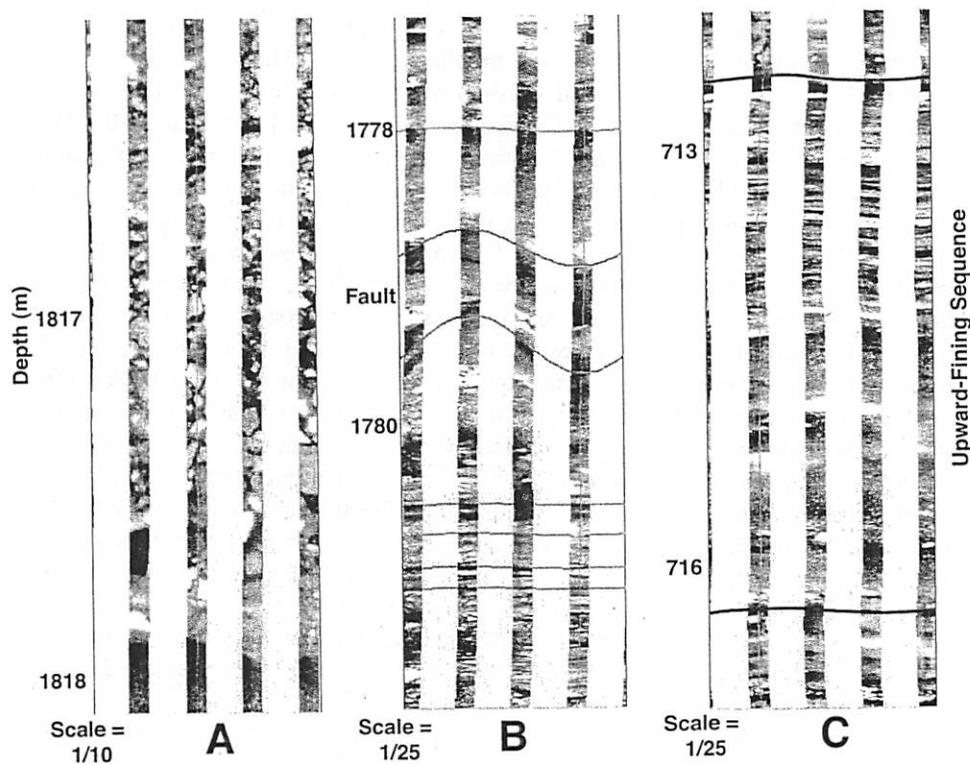
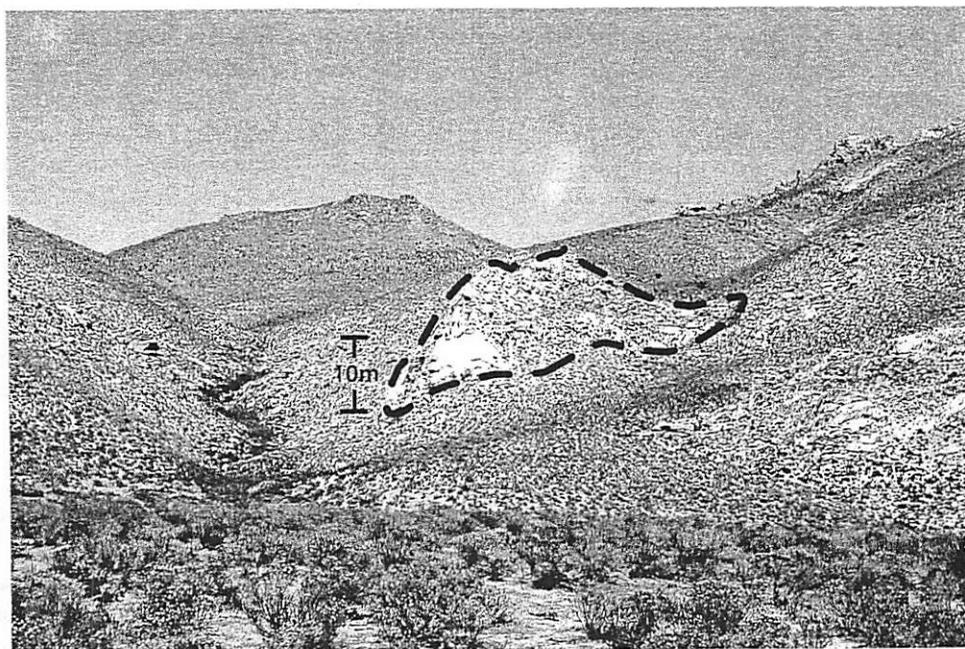


Figure 13. Formation MicroScanner (FMS) images from the SNORT 2 hole. Lighter tones indicate greater electrical resistivity of the material being imaged. Note that the scale of A is 1:10 whereas B and C are 1:25. (A) Breccia of the type shown here is common throughout the lower part of the SNORT 2 hole. Such breccia units become less commonplace with decreasing depth. These are interpreted as landslide-dominated alluvial fans related to early movement on the Sierra Nevada frontal fault. (B) High-angle fault surface located at a depth of 1779 to 1780 m. The fault appears as a high-amplitude sigmoidal curve because each of four sensor pads is 90° away from the one next to it and thus any dipping surface will appear as a sigmoid. This fault dips ~55°–60° to the west and correlates with one of the antithetic faults interpreted on the seismic section. (C) This upward-fining sequence from just below 716 m to just above 713 m is typical of what is seen throughout the upper part of SNORT 2. The sequences consist of laminated sandy, silty, and clayey material that exhibits cross-bedding and cut-and-fill structures. These are interpreted as facies (proximal, medial, and distal) of fan deltas and are related to the axial drainage system that fed into the Indian Wells Valley from the north. There are scoured surfaces at 712.8 and 714.3 m.

Figure 14. Large block of plutonic basement rock fully involved in the landslide debris. These types of clasts are common in the landslide deposits found in the northwest part of the Indian Wells Valley and are indicative of proximity to the Sierra Nevada frontal fault.



within the flows. The zone of poor reflectivity directly above the basement extends eastward to SP 1135 and correlates with massive debris-flow and avalanche-type deposits found in the lower part of SNORT 2. Reflectivity increases in the upper part of the section until there is a continuous sequence of reflective beds from the surface to basement east of SP 1135.

This lateral and vertical pattern of reflectivity is consistent with a progressive time shift in depositional style from a fault-proximal- to an alluvial-fan-type environment and is supported by data from SNORT 2. The stratigraphic interval 1524–686 m consists primarily of sand and gravel units with local, interspersed breccia (Fig. 11). This clast-supported material is composed of angular to subangular granitic detritus including feldspar, quartz, and biotite grains and lithic clasts of granite and metamorphic rocks in a very fine grained clay-rich and locally oxidized matrix. The poorly sorted, ungraded character is commonly observed throughout the sedimentary interval of SNORT 2 and is the product of alluvial-fan-dominated depositional processes with a local plutonic source. There is abundant evidence of cut-and-fill structures and cross-bedding in the FMS images from this interval, and there are upward-fining sequences (Fig. 13B) that range in thickness from a few meters to as much as 10 m. These sequences usually have scoured lower contacts that are immediately overlain by pebble conglomerate grading upward into finer-grained, typically cross-bedded, sandstone and siltstone. We interpret these features as progradational alluvial-fan-type deposits (Leeder and Gawthorpe, 1987; Kerr, 1984) whose source was in the Sierra Nevada.

In the stratigraphic interval 686–411 m, the rocks become much finer-grained with almost no evidence of breccia. There are local pebble conglomerates, cross-beds, and cut-and-fill structures, although they are much less abundant than in the interval below. We interpret these strata as the product of sed-

imentation in a lower-energy, fluvial-dominated alluvial-fan depositional environment. Because horizon C projects westward from SNORT 1 into the middle of this reflective sequence, we interpret these rocks to represent the fault-proximal basin-margin facies of the Pliocene White Hills sequence.

Central subsurface sequence. Reflectors associated with the White Hills sequence in the central part of the Indian Wells Valley are a mixture of strong, laterally continuous features interspersed with ones that are alternately strong and weak. This pattern is usually indicative of sand-shale sequences characteristic of fluvial or lacustrine depositional environments (Sangree and Widmier, 1977). The laterally continuous reflectors generally represent more-or-less synchronous events such as erosion surfaces or tuffs. The lithology in SNORT 1 (Fig. 11) shows that the lower part of the sequence (1341–1005 m) from bottom to top consists mostly of claystone, sandy claystone, and sandstone. The sand grains are poorly to well-sorted, subangular to subrounded, clear to frosted white quartz with varying amounts of hematite stain and scattered mica. The 3.9 Ma basalt flows that occur at 1005–975 m are overlain by a 416-m-thick section of claystone, silty sandstone, and sandstone with local, thin limestone beds. For the most part, the characteristics of the sand grains are little different from those in the lower sedimentary section. The top of the section is marked by horizon D.

These lithologies, grain features, and vertical variability are consistent with deposition in lacustrine and fluvial environments. The amount of rounding and frosting of sand grains combined with the nearly monomineralic character of the intervals indicates a substantial amount of fluvial and eolian transport. Varying amounts of clay and intermittent layers of limestone record lower-energy environmental conditions when there was little sediment flux into the basin.

The fact that the thickness of the White Hills sequence from the top of horizon C to horizon D does not vary more than a few milliseconds between SP 1330 and SP 1290 on IWV-92-02 indicates a relatively stable sedimentary basin during that time. The small variation is probably as much the result of resolution of the seismic record as it is a real phenomenon. Reflectors within that interval consistently dip to the northeast except in the immediate vicinity of SP 1330 where they are flexed downward toward the southwest. This flexure may be due to postdepositional compaction or possibly to fault drag. The fact that horizon C is offset vertically along that fault indicates that movement occurred after 3.9 Ma. Horizon D is only slightly warped, however, indicating that most of the movement on that fault must have occurred before the beginning of the Pleistocene (?) (see later discussion).

Eastern facies. A change in the depositional conditions associated with the White Hills sequence in the eastern Indian Wells Valley is manifested in the difference in reflection character above and below horizon C. Pre-3.9 Ma deposition appears to have been controlled by basin subsidence whereas post-3.9 Ma conditions were dominated by westward progradation of alluvial fans and/or fan deltas. The earlier phase of deposition resulted in a local wedge of sedimentary rocks that thins toward the basin. Reflectors at 800 ms appear to define an angular unconformity at SP 1480. A small basin beneath that horizon appears to have been dropped downward, resulting in an eastward dip of internal reflectors. Subsequent deposition resulted in overlapping of rocks to the west. This unit also was down-dropped prior to emplacement of horizon C, followed by onset of the deposition of the alluvial fans and fan deltas.

Duffield et al. (1981) described alluvial fans cropping out on the western margin of the Argus Range, which they divided into younger (Qya) and older (Qoa) units primarily on the basis of their state of dissection. We found that there are also distinct clast differences in the two types of fans. The older fans have a predominance of rounded to subrounded volcanic clasts, and the younger fans have mostly plutonic and very few volcanic clasts. The presumed sources for the volcanic clasts are Pliocene basalt flows that occur in the Argus Range directly east of the older fans. We attribute this change in clast type to erosion of the flows and subsequent exposure of underlying plutonic rocks. The fans were later uplifted by post-Pliocene faulting of the Argus Range, and a younger set of alluvial fans developed on the top. The older fans are likely age equivalent to the White Hills sedimentary sequence seen on the eastern end of IWV-92-02.

Between SP 1370 and SP 1420 at 500–700 ms there are at least two stacked sets of distinctive westward-prograding, fluvial-dominated fans or fan deltas (Leeder and Gawthorpe, 1987; Ballance, 1984; and Kerr, 1984) in the White Hills depositional basin. These grade westward into the lacustrine and fluvial deposits of the central basin. East of SP 1420, reflectors in this stratigraphic interval are both laterally continuous and broken features consistent with an alluvial-plain depositional environment where braided streams dominated, but occasional

total submergence and/or sheet flooding was possible. Syntectonic sedimentation associated with faulting is suggested by the increase in thickness of the White Hills sequence evidenced by the sag in horizon D from SP 1314 to SP 1344. The short reflectors between 500 and 600 ms from SP 1325 to SP 1337 dip toward the sag and are possibly foreset beds associated with small-fan construction into the depositional basin adjacent to an active fault.

Southeastern extent. The White Hills sequence thins to the south along IWV-92-03 (Plate 3) until horizon D appears to pinch out at SP 1260. Analysis of microfossils from TGCH 1 indicates that the section from 377 to 567 m is a Pliocene lacustrine suite deposited in a highly alkaline lake environment (R. Forester, 1999, personal commun.). This interval is significantly thinner (~190 m) than in the northern part of the valley, but its stratigraphic position on the seismic section appears to make it correlative with the White Hills sequence. If it is assumed that this correlation is correct, the lacustrine depositional basin during the time of White Hills deposition extended at least 20 km in a north-south direction.

Evidence of a lacustrine sequence that is possibly age equivalent to the White Hills sequence is also found in well 22P1 (Fig. 4). According to Kunkel and Chase (1969), this well penetrated a thin veneer of fine sand, then went through a 151-m-thick section of lacustrine deposits primarily consisting of clay, calcareous clay, and fine silt with minor amounts of fine gravel, some sand, and scattered fossils. The next 123-m-thick section consisted of an alternating sequence of unindurated sand, silt, and clay with one gravel zone that Kunkel and Chase (1969) characterized as older alluvium. The lower interval best correlates with the White Hills sequence, whereas the upper interval correlates with the unnamed Pleistocene rocks discussed next.

There are several locations in the southern Indian Wells Valley where older lacustrine deposits (Kunkel and Chase, 1969) crop out (Tal, Fig. 4). These consist of poorly to well-indurated fine sands, silts, and clays with locally dense limestone layers. Small pebbles (<2 cm in diameter) of granite, vesicular basalt, and diorite are present in areally limited beds of conglomerate. Microfossils are scarce in surface outcrops, but where present, indicate the same type of depositional environment indicated by the White Hills rocks. Local occurrences of conglomerate are likely due to sheet-flood deposition from low-lying basement outcrops to the south. It is noteworthy that these lacustrine beds are exposed as broad, low-relief (<2 m high) warps on an otherwise flat valley floor and they occur only in the southern part of the valley.

Pleistocene–Holocene features of the Indian Wells Valley

During the Pleistocene and into the Holocene, the Indian Wells Valley appears to have been an intermontane basin. Besides clastic input, lava flowed southward from the Coso Range into the valley. It is during this period of time that there was a major transition in structural style from low-angle normal fault-

ing along the Sierran front to throughgoing, north-northwest-oriented dextral faulting in the Indian Wells Valley itself.

Determination of the precise point in the sedimentary section that marks the transition from Pliocene to Pleistocene in the Indian Wells Valley is not possible on the basis of available data. However, we think that a reasonable estimate can be inferred from lithology and sedimentation-rate data from the drill holes in the valley. The thickness of the sedimentary section above horizon D, which we believe marks the Pliocene–Pleistocene boundary, on IWV-92-02 (Plate 2) reaches a maximum of ~600 m at SP 1320 and thins to a basin margin at the ends of both seismic lines. In SNORT 1, SNORT 2, NR 1, and NR 2 (see Fig. 4 for location) there is a clay layer that ranges in thickness from 450 m in the central part of the valley to ~100 m on the east side of the valley. In SNORT 2, the section from 411 m upward to ~135 m (i.e., 276 m thick) consists of a blue-green to dark gray clay that is also found in the nearby NR 1 (450 m thick) and NR 2 (311 m thick) wells. This same layer is ~450 m thick in SNORT 1 and thins to ~100 m in TGCH 1. The bottom of the interval is defined by a marked upward decrease in electrical resistivity coinciding with a change from sand below to clay above. This lithologic transition gives rise to the strong, laterally pervasive seismic reflector that we refer to as horizon D.

On the basis of work done in nearby basins, we infer that the base of this clay layer is very close to the Pliocene–Pleistocene boundary at 1.8 Ma. Thick clay layers in the subsurface of Owens Valley and Searles Lake (Fig. 3) have been attributed to large lakes (Smith et al., 1983, 1997) that formed during pluvial (interstadial) periods of the Pleistocene. Using radiocarbon-dating methods, Bischoff et al. (1997) determined an average sedimentation rate of 3×10^{-2} cm/yr for lacustrine clays and 80 cm/yr for coarser sediments in a 322-m-long Pleistocene–Holocene core from the Owens Valley. By using these average sediment-accumulation rates, we find that the 450-m-thick clay in SNORT 1 and NR 1 represents ~1.5 m.y. of deposition and the 125 m of sand and silty-sand overlying the clay in SNORT 1 represents an additional ~138 k.y. The combination of the two being 1.64 m.y. justifies a reasonable estimate for the base of the clay layer being very near the Pliocene–Pleistocene boundary.

Lithologic information from water wells drilled in the Indian Wells Valley reflects the fact that the valley was an internally draining basin throughout most of the Pleistocene and Holocene. Smith and Pratt (1957) analyzed core material from China Lake playa that consisted of fine sand, clay, and some freshwater limestone and evaporite. They correlated these deposits with similar material from Searles Lake (Fig. 3) and discussed the Indian Wells Valley in the context of an interconnected Pleistocene lake system that originated in ancient Lake Lahontan. We accept this interpretation, and we apply their chronology to the valley rocks.

Smith et al. (1983) dated old lake deposits from nearby Searles Lake (Fig. 4) by using paleomagnetic stratigraphy from

a 930 m core. They concluded that the core spanned the late Pliocene and the entire Pleistocene, forming an almost continuous depositional record of pluvial lake sedimentation. The following summary of depositional conditions is taken from Smith et al. (1983). Prior to 3.2 Ma, sedimentation in Searles Valley consisted of prograding alluvial fans from the ancient Argus Range on the west. At ca. 3.2 Ma there was a change in sedimentation pattern to a deep, freshwater lacustrine environment. This pattern lasted until near the beginning of the Pleistocene (2.0 Ma) when circulation in the lake became more restricted, and the lake environment became more saline. These conditions persisted throughout the Pleistocene except for a period of 250 k.y. beginning at 1.25 Ma when the lake reverted to a more freshwater condition. From ca. 1.0 Ma until 0.6 Ma, Searles Valley was the site of a moderately saline lake in which clays and saline layers were alternately deposited with interspersed air-fall tuffs. At 0.6 Ma there was yet another change in depositional environment resulting in Searles Lake becoming a dry salt flat for the next ~300 k.y. The final episode recorded in the Searles Lake core lasted from 0.3 Ma until ca. 0.13 Ma during which time there was deposition of large thicknesses of evaporites precipitated from a highly saline lake. This condition is consistent with a climate in which there was a lot of rainfall during certain times of the year and a significant amount of evaporation at other times. Fluvial and alluvial processes dominated the basin margins forming alluvial fans, fan deltas, and sheet-flood plains.

A resurgence of volcanism in the Pleistocene resulted in basaltic rocks being erupted in the Coso Range as part of a bimodal suite of flows and tuffs. Duffield and Bacon (1980) mapped four separate flows, ranging in age from 0.49 ± 0.11 Ma to 0.14 ± 0.09 Ma that originate in the southern Coso Range and extend into the northwest corner of the Indian Wells Valley. Prior to eruption of the Coso Pleistocene basalts, the Owens River flowed south into the valley (Duffield and Smith, 1978). On the basis of geophysical profiles, Duffield and Smith (1978) postulated that the flows changed the course of the ancient Owens River by forcing it westward into a narrow channel in the extreme northwest corner of the valley where it flowed until the mid-1900s.

Lithology of cores and drill cuttings from Indian Wells Valley holes records an environment that is consistent with the findings from Searles Lake (Smith et al., 1983). Sedimentation in the central part of the valley during this time was either in lakes or playas with some likely deltaic input from the ancient Owens River that emptied into the northwest part of the valley (Duffield and Smith, 1978). We presume there was a relatively consistent inflow to the valley, normal rates of evaporation, and a more or less continuous overflow from the lake in the Indian Wells Valley to Searles Lake because there is only limited evidence of evaporites in any of the Indian Wells Valley holes during this period. On the basis of the interpreted environment of deposition, the occurrence of air-fall tuffs, and the similarities in lithology, we correlate the clay-rich interval in the Indian

Wells Lake drill holes with the lake deposits from Searles Lake (Smith et al., 1983) that were emplaced throughout the Pleistocene.

Overlying the entire basin sequence is a relatively thin (0–30 m) layer of modern alluvial-fan, lacustrine, and playa deposits with scattered patches of eolian sand. At the present time, deposition in the Indian Wells Valley is limited to alluvial-fan progradation on the extreme western and eastern margins of the basin, and aerially limited sheet-flood deposition during heavy rainfall periods. There are several small playas in the modern valley, mainly in the eastern and southern parts.

Structural features. Right-lateral faults cut through the entire Indian Wells Valley stratigraphic section in three narrow north- to northwest-trending zones. From west to east, these are the Little Lake fault zone, the Airport Lake fault zone, and the Argus frontal fault zone (Fig. 4), all of which are currently active. In the seismic sections, they form typical flower structures with fault traces cutting through to the surface for the Little Lake (SP 1250–1295) and the Airport Lake (SP 1496) faults.

DISCUSSION

The Cenozoic history of the Indian Wells Valley is recorded stratigraphically only for the Paleocene–early Eocene and Miocene–Holocene intervals. The former interval is represented by the Goler Formation; the latter, by a sequence of rocks that includes the Ricardo Group (Cudahy Camp and Dove Spring Formations), the White Hills sequence, and unnamed Pleistocene–Holocene rocks. We briefly describe the Paleogene history of the valley, followed by a more comprehensive discussion of its Neogene history.

Paleogene history of the Indian Wells Valley

The earliest record of deposition in the Indian Wells Valley is contained in the Goler Formation. We have identified rocks that are equivalent in age to the Goler in the lower parts of TGCH 1 and SNORT 1 on the basis of clast composition, degree of lithification, and physical characteristics. These rocks are principally coarse-grained, reddish-brown to dark brown conglomerates with local red to reddish-brown sandstone layers. Locating the top of the Goler in the Indian Wells Valley seismic profiles is somewhat more problematic. We have chosen a significant increase in acoustic formation velocity coupled with lithologic characteristics as the most significant criteria in making that call. Cox (1982) found that the Goler–Cudahy Camp unconformity is as much as 35°, which should make it easily recognizable on the seismic sections; however, the interface between the Cudahy Camp and the Dove Spring Formations is also an angular unconformity so this criterion alone is insufficient.

Cox (1982) noted that the Goler Formation rests nonconformably on a highly irregular, eroded, and weathered base-

ment, which appears to be the case in the Indian Wells Valley as seen in the seismic sections. The depositional edge of the Goler basin can be distinguished only on the southeast end of line IWV-92-03 (Plate 3) where it rises very near the surface; elsewhere, the margin is faulted. Given that there are only four points available to estimate the extent of the basin and owing to structural complications discussed in subsequent sections of this paper, it is difficult to make a definitive statement about the extent of the basin during Goler time. However, on the basis of data already reviewed, we conclude that the Goler basin existed during the Paleocene at the present location of the Indian Wells Valley, was more or less elongate in an east-west direction, and was the site of alluvial-fan and sheet-flood deposition.

During the remainder of the Paleogene, general uplift and erosion occurred in the Mojave Desert region and the southwestern part of the Basin and Range, as evidenced by the fact that there are no documented occurrences of rocks of middle Eocene–early Miocene age. The cause of this regional uplift has been attributed to a worldwide adjustment in plate motions and a change from convergence to divergence along the Farallon and North American plates. This change was accompanied by a buoyancy effect resulting from subduction of young, hot lithosphere beneath western North America (Ward, 1991; Crowell, 1982; Glazner and Loomis, 1984). Active marine and nearshore nonmarine deposition proceeded during this interval in the southern San Joaquin basin (Goodman and Malin, 1992), but there is no evidence of deposition in the area of the Indian Wells Valley.

Neogene record of the Indian Wells Valley

The Miocene saw renewed deposition in the Indian Wells Valley and the adjacent El Paso basin. Rocks of the Ricardo Group attest to the fact that the area was once again a negative topographic feature receiving sediments. Throughout the next 20 m.y., the valley was for the most part a depocenter that recorded passive-basin sedimentation during the early to middle Miocene, low-angle normal faulting and uplift of the Sierra Nevada in the latest Miocene and Pliocene, and transtensional faulting from the latest Pliocene to the present.

Miocene time. Throughout the Miocene, the Indian Wells Valley was a depositional basin for Cudahy Camp and Dove Spring Formation rocks. Rocks in the lower parts of SNORT 1 and TGCH 1 appear to be distal facies of units of both formations. It does not appear that there is a complete section of either formation buried in the Indian Wells Valley, which leads to one of two conclusions. Either the valley was at the most distal edge of the depositional basins during that time, or the rocks were deposited and subsequently eroded. In Teagle Wash, located 20 km south of the Indian Wells Valley (see Fig. 4), there are outcrops of a distinctive andesite breccia (Trab of Loomis, 1984), tuffs, and basalt flows that we have correlated with units of the Cudahy Camp Formation. Although these exposures are not areally extensive, they are clearly in place, attesting to the fact

that the Cudahy Camp depositional basin extended at least that far. These outcrops and the ~3000-m-thick buried section seen on the south end of line IWV-92-03 (Plate 4), which we interpret to consist of rocks of the Goler, Cudahy Camp, and Dove Spring Formations, lead us to conclude that in Miocene time, the depositional basin extended 25 km east of present outcrops in the El Paso Mountains (Fig. 15A). However, because the Miocene section in the Teagle Wash (IWV-92-03) had a source in the Eagle Crags (Monastero et al., 1997), the Indian Wells Valley must have been located northeast of the volcanic edifice. This position suggests that the valley was at the margin of the basin and the thickness of the rocks was never very great. The absence of members of the Cudahy Camp and Dove Spring Formations in the drill holes and seismic sections could then be explained simply on the basis of variation of the northern margin of the depositional basin.

During the middle to late Miocene, the Indian Wells Valley was also the locus of sedimentation at the distal margin of the Dove Spring Formation basin (Fig. 15B). The setting was predominantly lacustrine with occasional fan deltas and marginal flood-plain and sheet-flood environments. Again, because the basin was distal to the source, which was primarily the ancient Lava Mountains, the stratigraphic record for this interval is discontinuous in the Indian Wells Valley.

Abrupt termination of more than 3000 m of Goler, Cudahy Camp, and Dove Spring Formation rocks at the northeast end of the El Paso Mountains must be reconciled. It seems unlikely that the basins for all three of these formations simply stopped receiving deposits without any evidence of basin-margin features in the outcrops. It is much more likely that the section is terminated by faulting, resulting in the older rocks being dropped down and subsequently buried by younger sediments. We infer that this faulting occurred in the latest Miocene or early Pliocene in conjunction with low-angle faulting along the Sierra Nevada frontal fault (as discussed in the next section).

Latest Miocene–Pliocene low-angle faulting. At the end of the Miocene and throughout the Pliocene Epoch, the Indian Wells Valley was the locus of faulting along the Sierra Nevada frontal fault with related clastic deposition in a half graben. Coarse debris was shed into the basin along the rugged western margin, alluvial fans prograded westward from the Argus Range, and lacustrine, outwash plain, and fluvio-deltaic sedimentation dominated deposition in the central part of the valley (Fig. 15C). Volcanic rocks periodically extended into the basin from the north and northeast and are preserved in subsurface basalt flows as far south as SNORT 1.

Onset of faulting on the Sierra Nevada frontal fault is estimated to be post-7 Ma on the basis of the Dove Spring Formation stratigraphy. The top of that formation is an unconformity that is both erosional and angular. By using fossil faunas and radiometric dating of tuffs, Whistler and Burbank (1992) determined that the youngest unit in the Dove Spring Formation was deposited at 7 Ma. Therefore, it must be assumed that the

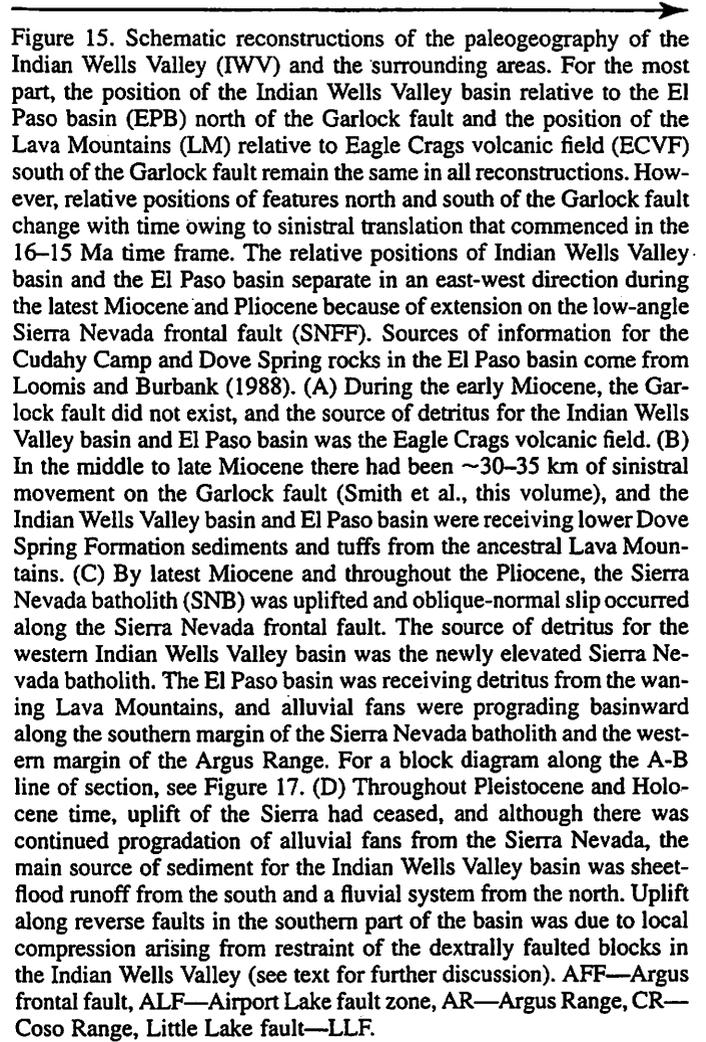
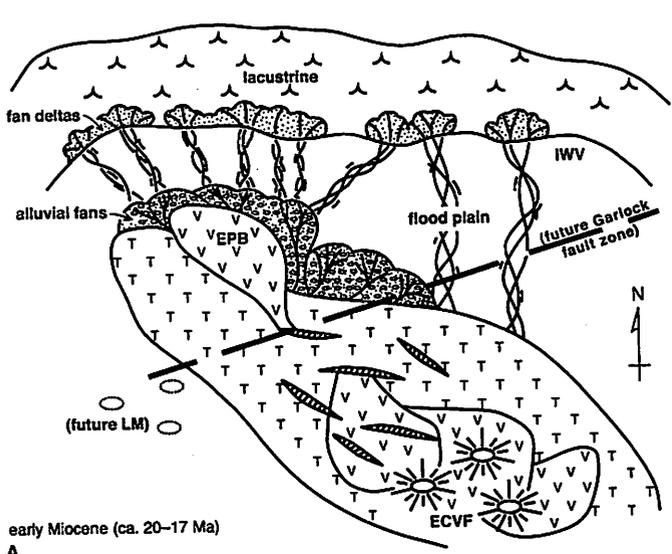


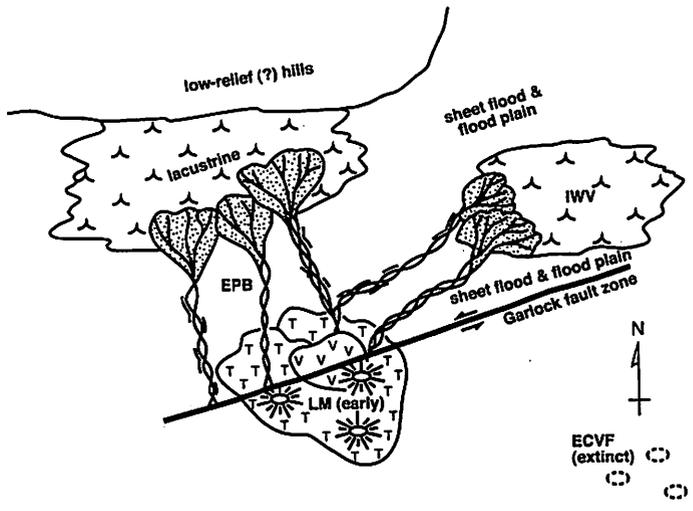
Figure 15. Schematic reconstructions of the paleogeography of the Indian Wells Valley (IWV) and the surrounding areas. For the most part, the position of the Indian Wells Valley basin relative to the El Paso basin (EPB) north of the Garlock fault and the position of the Lava Mountains (LM) relative to Eagle Crags volcanic field (ECVF) south of the Garlock fault remain the same in all reconstructions. However, relative positions of features north and south of the Garlock fault change with time owing to sinistral translation that commenced in the 16–15 Ma time frame. The relative positions of Indian Wells Valley basin and the El Paso basin separate in an east-west direction during the latest Miocene and Pliocene because of extension on the low-angle Sierra Nevada frontal fault (SNFF). Sources of information for the Cudahy Camp and Dove Spring rocks in the El Paso basin come from Loomis and Burbank (1988). (A) During the early Miocene, the Garlock fault did not exist, and the source of detritus for the Indian Wells Valley basin and El Paso basin was the Eagle Crags volcanic field. (B) In the middle to late Miocene there had been ~30–35 km of sinistral movement on the Garlock fault (Smith et al., this volume), and the Indian Wells Valley basin and El Paso basin were receiving lower Dove Spring Formation sediments and tuffs from the ancestral Lava Mountains. (C) By latest Miocene and throughout the Pliocene, the Sierra Nevada batholith (SNB) was uplifted and oblique-normal slip occurred along the Sierra Nevada frontal fault. The source of detritus for the western Indian Wells Valley basin was the newly elevated Sierra Nevada batholith. The El Paso basin was receiving detritus from the waning Lava Mountains, and alluvial fans were prograding basinward along the southern margin of the Sierra Nevada batholith and the western margin of the Argus Range. For a block diagram along the A-B line of section, see Figure 17. (D) Throughout Pleistocene and Holocene time, uplift of the Sierra had ceased, and although there was continued progradation of alluvial fans from the Sierra Nevada, the main source of sediment for the Indian Wells Valley basin was sheet-flood runoff from the south and a fluvial system from the north. Uplift along reverse faults in the southern part of the basin was due to local compression arising from restraint of the dextrally faulted blocks in the Indian Wells Valley (see text for further discussion). AFF—Argus frontal fault, ALF—Airport Lake fault zone, AR—Argus Range, CR—Coso Range, Little Lake fault—LLF.

Dove Spring Formation was uplifted and tilted sometime thereafter.

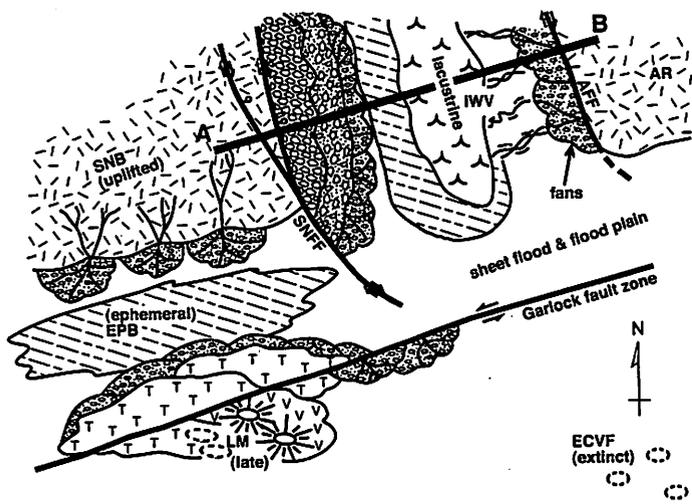
Loomis and Burbank (1988) found that between 8 and 7 Ma, the source area for the Dove Spring Formation shifted to the north from the south and southeast, and detritus from Sierra Nevada plutonic rocks began to dominate the section. It appears then, that the Sierra Nevada was already being uplifted in the latest Miocene. On the basis of stream-gradient changes on the west side of the Sierra Nevada, Unruh (1991) suggested that at ca. 5 Ma, the range was uplifted and tilted westward. Lueddecke et al. (1998) determined that westward tilting of the Sierra Nevada and simultaneous opening of the Owens Valley occurred between 6 and 3 Ma. This coincides with deposition of the fluvio-lacustrine Coso Formation that filled a narrow, internally draining basin between the Coso Range and the Sierra Nevada (Kamola and Walker, 1999). Additional evidence of a sweeping tectonic change at this time is provided by Schweig (1989) and Bacon et al. (1982), who documented the onset of extension in



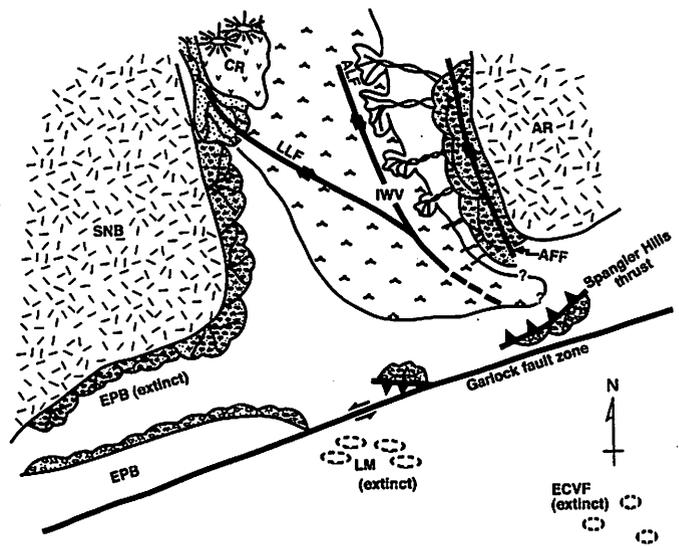
early Miocene (ca. 20-17 Ma)
A



middle to late Miocene (ca. 15-7 Ma)
B



latest Miocene - Pliocene (ca. 7-3.5 Ma)
C



Pleistocene - Holocene (ca. 2-0.02 Ma)
D

- | | | | | | |
|--|--|--|------------------------------------|--|--------------------------------------|
| | strike-slip fault | | extinct volcanic center | | megabreccia |
| | oblique-slip fault, dash and ball on downthrown side | | tuff | | sheet-flood and flood-plain deposits |
| | reverse fault, teeth on upthrown side | | lava flow | | lake |
| | volcanic dike | | fan delta | | ephemeral lake |
| | active volcanic center | | alluvial fan | | plutonic basement |
| | future volcanic center | | debris-flow and avalanche deposits | | |

this area between 6 and 4 Ma. The coincidence of these events leads us to conclude that uplift of the Sierra Nevada began in the latest Miocene or earliest Pliocene, between 7 and 5 Ma, and was accommodated by normal faulting along the eastern front of the Sierra Nevada.

Why the Sierra Nevada was uplifted during latest Miocene–early Pliocene time is uncertain. However, on the basis of analysis of mafic and ultramafic xenoliths, Ducea and Saleeby (1998) proposed that somewhere between 8 and 4 Ma there was delamination of an eclogitic keel from the southeastern Sierra Nevada. This process could have resulted in uplift of the range because of either isostatic rebound from loss of the dense keel or thermal forces resulting from addition of lighter, hotter asthenospheric mantle on the underside. Kay and Kay (1993) showed that such delamination is usually accompanied by rapid uplift, stress change, and changes in magmatism—conditions that were prominent in the southern Sierra Nevada–Indian Wells Valley region in the late Miocene–early Pliocene. The proposed timing of delamination under the southern Sierra Nevada by Ducea and Saleeby (1998) is permissive as a driving mechanism for movement on the Sierra Nevada frontal fault. Calc-alkalic volcanism that began at ca. 5 Ma in the Wild Horse Mesa area and was essentially over by 3.5 Ma may be related to a second period of delamination (Manley et al., 2000) or may be part of the same event.

Although normal displacement on the Sierra Nevada frontal fault is easily demonstrated in the relief of the Sierra Nevada, a component of lateral movement may also be present. Analysis of FMS logs from SNORT 2 shows a counterclockwise rotation of bedding azimuths with increasing depth (Fig. 7) that can be achieved if there is simultaneous down-to-the-east, right-oblique normal movement and east-west extension on the Sierra Nevada frontal fault. Note that in this case, rotation of the bedding planes is related to local fault drag and does not imply rotation of the entire Indian Wells Valley. East-west extension can be achieved if the axis of maximum tension is oriented as shown in Figure 16. On the basis of a geologic reconstruction, Wernicke and Snow (1998) found that the direction of extension in the southwestern Basin and Range changed from east-west to northwest-southeast between 10 and 8 Ma. This is older than our estimate of earliest onset of extension at 7.5 Ma, but is permissive, given the uncertainties in the two methods.

Over the next 3.5–4 m.y., movement on the Sierra Nevada frontal fault resulted in a classic half-graben depositional setting (see Fig. 17) as described by Leeder and Gawthorpe (1987). Rocks that are proximal to the Sierra Nevada frontal fault consist of megabreccia and coarse, poorly sorted avalanche and debris-flow deposits, all of which are represented in outcrop in the northwestern Indian Wells Valley as well as being found in the subsurface in SNORT 2. Syntectonic sedimentation along the Sierra Nevada frontal fault resulted in a thick wedge of eastward-prograding coarse clastic units that give way to sheet-flood deposits, alluvial fans, and ultimately to fluvial, deltaic, and lacustrine deposits in the central part of the valley. At the

same time, alluvial fans and fluvial-deltaic deposits were prograding westward into the valley from the Argus Range, resulting in formation of fan and fan-delta complexes (Fig. 15C). The whole of these deposits constitutes facies of the Pliocene White Hills sequence.

We have estimated the amount of extension in the Indian Wells Valley during this half-graben phase on the basis of where horizon B terminates against the basement on IWV-92-02 (Plate 2). First, we assume that the basin edge, represented by the pinch-out of horizon B against a basement high at SP 1157, was at or near the eroded surface of the Sierra Nevada in the past. The southern Sierra Nevada has a plateau at an altitude of ~2300 m. We presume that this plateau is an eroded surface that was either a low, positive feature or a slightly negative feature throughout the Miocene. When the subsurface contact between the basement and the low-angle fault is restored updip to the west such that it is at the 2300 m altitude, the amount of horizontal extension that is accommodated by the low-angle fault is on the order of 10 km. This amount represents ~40% of the total width of the modern Indian Wells Valley.

Although there is some uncertainty regarding when movement on the Sierra Nevada frontal fault ceased, we will argue subsequently that it was at ca. 3.5 Ma, which means that the duration of low-angle normal displacement was a maximum of 3.5 m.y. If correct, then the rate of extension during the early and middle Pliocene was 3.1 mm/yr. Smith et al. (this volume) have shown that since 10 Ma there has been ~30 km of sinistral displacement on the Garlock fault, a rate of 3 mm/yr. Assuming that all of this displacement is parallel to the trace of the Garlock fault, the predicted amount of east-west extension along the Sierra Nevada frontal fault would be 3.1 mm/yr with a north-northwest component of displacement of ~0.8 mm/yr (see Fig. 18). The total amount of extension accommodated by the Sierra Nevada frontal fault during this period would then be ~11 km, and the amount of north-northwestward translation would be ~2.8 km. Given the uncertainties in these rate calculations, the westward component agrees remarkably well with our calculated offset on the Sierra Nevada frontal fault. The dextral offset on the frontal fault that we have documented may represent accommodation of the north-northwest-directed translation, although we cannot document that with our data.

Pleistocene–Holocene dextral faulting

The final stage in the Neogene evolution of the Indian Wells Valley is dominated by dextral shearing evidenced by right-lateral offset in surface features (Roquemore, 1981; Roquemore and Zellmer, 1987), the large number of modern seismic events, and development of flower structures in the Little Lake, Airport Lake, and Argus Frontal fault zones. The throughgoing shear zone is oriented in a north-northwest direction consistent with modern geodetic measurements of relative crustal velocity in the area (e.g., Bennett et al., 1997).

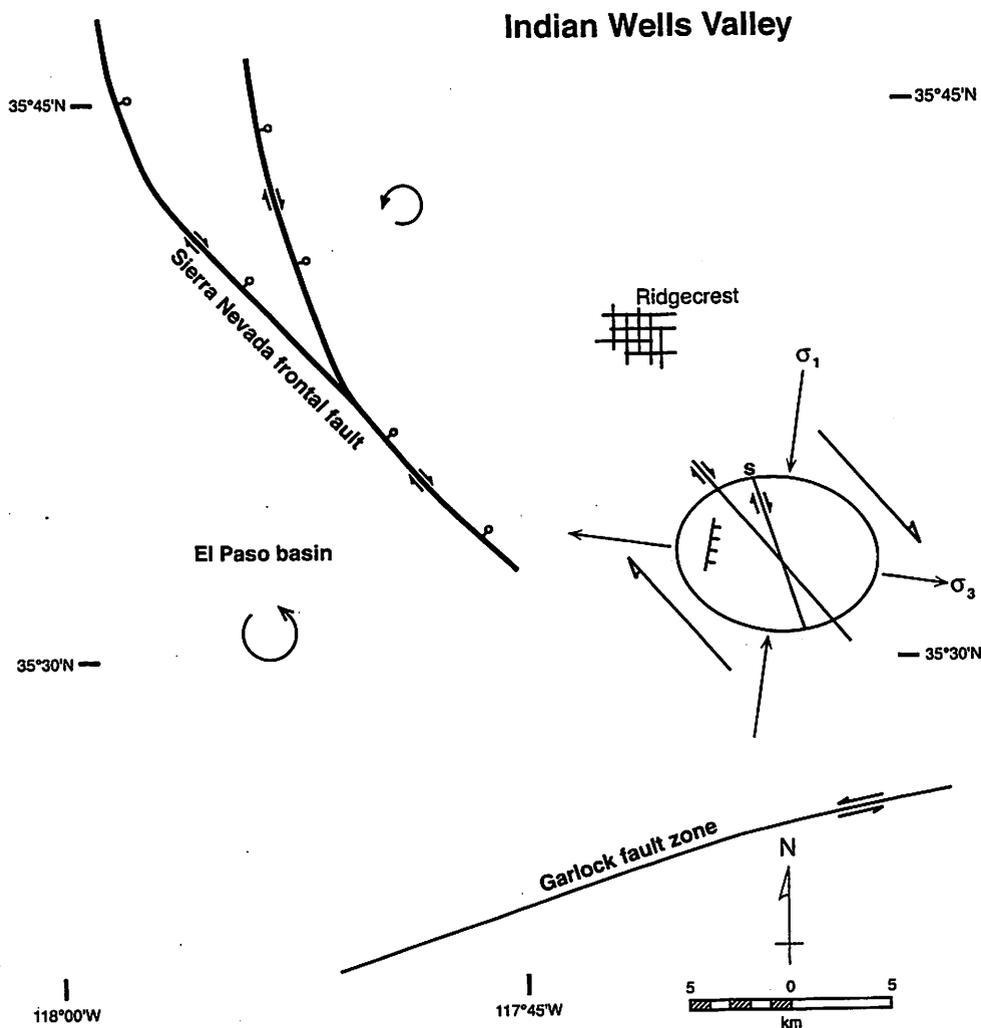


Figure 16. Simple kinematic model to explain the tectonic regime of the Indian Wells Valley during the latest Miocene-early Pliocene interval (ca. 7–3.5 Ma). All of the structures can be explained by a single stress field in which σ_1 is oriented northeast-southwest and σ_3 is oriented northwest-southeast. Simultaneous rotation of the El Paso basin block and east-dipping low-angle normal faulting can be achieved if the western component of extension is greater than the northern one. Fault symbols as in Figure 15. S—synthetic strike-slip fault orientation. Hachured lines inside strain ellipse denote ideal preferred orientation of normal faults.

During the past 2 m.y., the Indian Wells Valley has been the site of a sedimentary basin in which lacustrine deposition dominated (Fig. 15D). Alluvial fans built out from the eastern and western margins, and at varying times, depending on whether it was a glacial or an interglacial period, a delta developed where the ancient Owens River debouched into the northwest part of the valley (Duffield and Smith, 1978).

It is unclear when the transition to north-northwest-directed dextral faulting took place, but we conclude that it likely occurred after 3.5 Ma and before 2 Ma. Whitmarsh et al. (1996) have reported the occurrence of a faulted 3.5 Ma basalt at the southwest side of Coso Wash (Fig. 3) that matches in age with basalts from the east side of the wash (Tbp of Duffield and Bacon, 1980). At least one strand of the modern right-lateral fault passes through the Coso Wash. So, when drilling that we conducted between the two outcrops revealed no basalt of that same age, we assumed that the Coso Wash opened sometime after 3.5 Ma. This interpretation coincides with the findings of Zhang et al. (1991), Conrad et al. (1994), and Hodges et al.

(1989), who determined that right-oblique transtensional deformation in the Saline and Panamint Valleys began after 3.7 Ma. On the basis of the fact that there is ~1000 m of section in the half graben above the 3.9 Ma basalt (horizon C) in SNORT 1, we think that the transition was substantially later than that time. At ca. 2 Ma, step faulting was initiated on the Wild Horse Mesa (Duffield and Bacon, 1980), perhaps indicating the onset of transtension. From 2.5 Ma until 1.0 Ma there was little volcanism in the Indian Wells Valley area. Two lava flows and associated pyroclastic rocks were extruded between 2.0 and 1.75 Ma (Duffield and Bacon, 1980), but these were extremely small in volume and separated areally by more than 20 km. At 1 Ma there was an outbreak of bimodal volcanism in the Coso Range; since then, basalt flows and pyroclastic rocks were emplaced in the northern Indian Wells Valley, and rhyolite domes began forming in the Coso Range itself.

There is no report in the literature of a region-wide tectonic event that took place at this time that would explain the coincidence of these events. In reconstructing the Neogene exten-

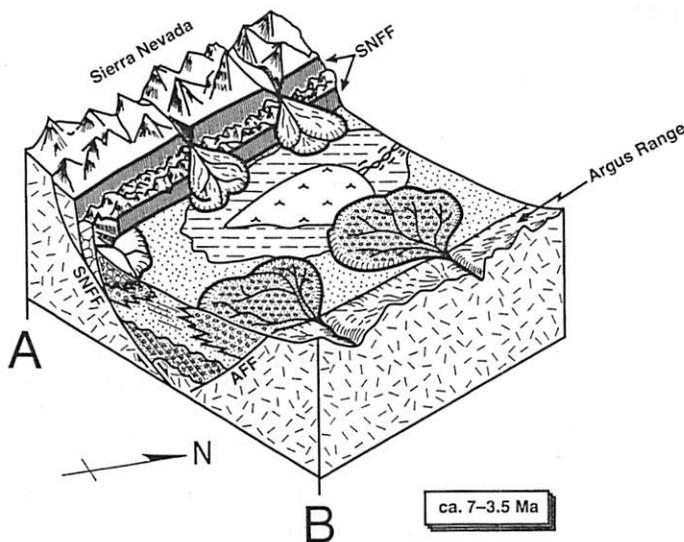


Figure 17. Block diagram of the paleogeography of the Indian Wells Valley during the latest Miocene and Pliocene; patterns as in Figure 15, and location shown in Figure 15C. Uplift has commenced on the Sierra Nevada frontal fault (SNFF), resulting in the formation of megabreccia blocks, avalanche deposits, and coarse-debris flows. Alluvial fans originate from the Argus Range along the antithetic Argus frontal fault (AFF). The axis of the basin is dominated by sheet flooding from the south, fluvial deposition from the northwest, and lacustrine sedimentation.

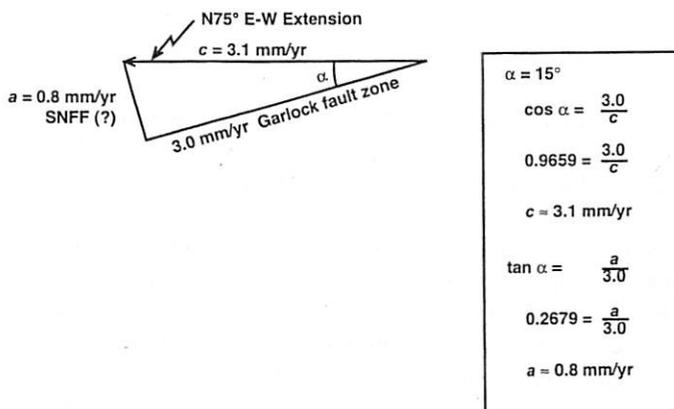


Figure 18. An average of 3.0 mm/yr of westward-directed left-lateral displacement on the Garlock fault produces 3.1 mm/yr of movement along a west-striking vector and a 0.8 mm/yr northward component. If such movement on the Sierra Nevada frontal fault for a period of 3.5 m.y. is assumed, 11 km of west-northwest-directed extension and 2.8 km of north-northwest-directed displacement would be produced. The 11 km value is in good agreement with our calculated displacement on the Sierra Nevada frontal fault (SNFF). The 2.8 km of north-northwest-directed displacement may have been accommodated by dextral movement on that fault as well although we cannot substantiate this possibility with our data.

sion direction in the Owens Valley–Death Valley extended terrane from geologic data, Wernicke and Snow (1998) showed a clockwise shift in the extension vector in the Owens Valley at ca. 2 Ma, although there is no discussion of the probable cause of that change.

Horizon D on IWV-92-02 (Plate 3) appears to represent a stratigraphic break that is manifested by progradation of reflectors over this horizon from the west and northeast where the younger unit forms a cap on the fan-delta reflectors. The section above horizon D is thicker in the middle of the valley, which may be related to downwarping and syntectonic sedimentation associated with the strike-slip faulting. We interpret the fact that this horizon represents the base of a clay layer in SNORT 2, SNORT 1, and nearby wells as evidence of a tectonic event that influenced the entire basin. As already noted, the relatively thick (300–450 m) and areally extensive clay layer indicates deposition in a deep, quiescent lake environment. This setting could be the result of a broad downwarping of the valley, analogous to that proposed by Smith et al. (1983) for Searles Lake, or could simply represent a major pluvial or interstadial period.

A consequence of dextral shearing through the Indian Wells Valley is localized compression in the southern end of the basin. Figure 19 shows the orientation of the major modern right-lateral faults that traverse the valley as well as associated structures to the south. The strain ellipse shows that orientation of the dextral faults is consistent with an extension direction of approximately N60°W, but orientations of folds and reverse faults in the southern part of the valley do not fit this stress field. This mismatch can be explained if the Garlock fault represents a significant structural barrier whose present orientation is not favorably aligned with the modern regional stress field. Therefore, dextral translation along the north-northwest-trending faults results in strain accumulation at the southern end of the valley. Several noteworthy examples of north-south shortening are found in the southern valley that cannot be related to sinistral strike-slip on the Garlock fault. In Teagle Wash there are anomalous northeast-trending folds that straddle the Garlock fault and are being actively folded (Smith, 1991). Smith (1991) recognized that the orientation of these folds was kinematically inconsistent with fault-induced folding resulting from sinistral movement on the Garlock fault, but was consistent with north-northwest-directed compression. Seismic reflectors on the southeast end of IWV-92-03 (SP 1810; Plate 4) show evidence of a reverse fault with ~800 m of throw. The absence of any remnants of the Goler, Cudahy Camp, or Dove Spring Formations in the Rademacher and Spangler Hills south of the Indian Wells Valley, but their presence in the valley and Teagle Wash, indicates that they have been uplifted and eroded. Finally, within the valley itself, outcrops of older lacustrine deposits (Qol, Fig. 4) are restricted to the southern end of the basin where they form broad, low upwarps on an otherwise flat valley floor. We suggest that all of these are the result of more or less north-south contraction related to modern right-lateral shear through the valley.

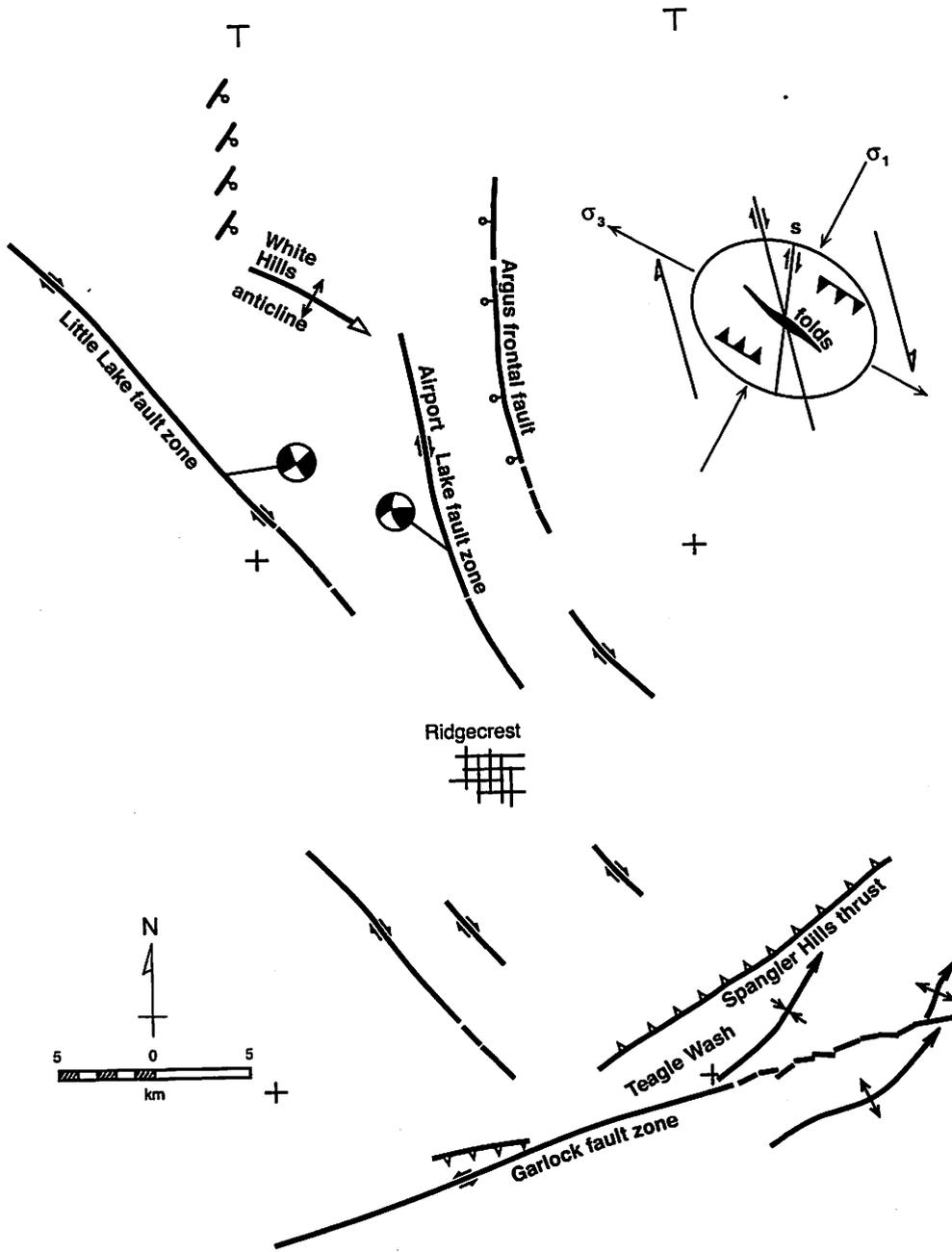


Figure 19. Orientation of modern, active structures in the Indian Wells Valley. These are forming in response to a stress field in which σ_1 (maximum stress) is oriented northeast-southwest with σ_3 (minimum stress) oriented northwest-southeast. The Airport Lake fault zone, the Argus frontal fault, and the Little Lake fault zone have all supported historical seismicity ranging from magnitude 3.0 to as high as magnitude 5.8 events. Focal mechanisms are representative of composite movement on the Airport Lake and the Little Lake fault zones. Structures shown in this plot are the result of the throughgoing right-lateral fault system, including these two fault zones and others that are unnamed, that originates near the present trace of the Garlock fault and extends into the Coso Range. The stress-strain determination is consistent with independent geodetic measurements of crustal velocity data for this area (Monastero et al., 2000). Fault symbols as in Figure 15.

Because of the absence of significant accumulations of modern sediment, the Indian Wells Valley does not appear to be downwarping rapidly at the present time. However, as noted previously, horizon D on IWV-92-02 does bow downward in the area of active strike-slip faulting (Plate 2, SP 1250-1350). Without major bends or stepovers in the plane of the faults, one could logically expect that deposition, if there was any, would be localized along the active zone of deformation. There are some relatively inconsequential deposits of alluvial-fan material that have extended over the valley floor from the surrounding ranges, but these appear to be confined to the extreme outer margins of the basin. Similarly, there are a few scattered eolian

and playa deposits throughout the valley, but they are thin and patchy in their distribution.

SUMMARY AND CONCLUSIONS

The Indian Wells Valley is situated in a key location with regard to unraveling the history of the northeastern Mojave Desert and the southwestern Basin and Range. It is bounded on the south by the Garlock fault, the major accommodation structure in the development of the southwestern Basin and Range (Davis and Burchfiel, 1973), and on the west by the Sierra Nevada. Geologic information from the Indian Wells Valley doc-

uments processes that have been active over the past 65–70 m.y. at the margin of two major physiographic and tectonic provinces in the southwestern United States.

The Indian Wells Valley has been the locus of deposition during four intervals: the Paleocene–early Eocene, the Miocene, the Pliocene–Pleistocene, and the Holocene. Our data indicate that deposition was virtually all nonmarine in basins that were limited to dimensions on the order of tens of kilometers. Alluvial-fan, alluvial-plain, fluvial, and lacustrine environments dominate the stratigraphy in the Indian Wells Valley; volcanic rocks provide a relatively small, but geologically significant, percentage of the total section.

At least three major periods of deformation are recorded in the Indian Wells Valley. The first change from northeast-directed extension to west-northwest-directed extension in the Death Valley area (Fitzgerald et al., 1991; Gans and Bohrsen, 1998) at ca. 16 Ma resulted in the onset of sinistral movement on the Garlock fault. This event is recorded as an angular unconformity between the Cudahy Camp and Dove Spring Formations that represents a hiatus of ~1.5 m.y. Beginning at ca. 13.5 Ma, and over the succeeding 6 m.y. period, the Indian Wells Valley was at the margin of a passive intermontane basin that received deposits now forming the distal facies of the Dove Spring Formation. The absence of parts of the Miocene stratigraphic section in the Indian Wells Valley indicates that it may also have been an intermittent topographic high where there was no sedimentation or that it was undergoing erosion.

Between 7 Ma and 5 Ma there was a significant pulse of uplift of the Sierra Nevada that was accommodated by low-angle normal faulting along the Sierra Nevada frontal fault and formation of a classic half graben in the Indian Wells Valley. This second phase of deformation may have been due to delamination of the lower crust or lithospheric mantle beneath the southern Sierra Nevada (Ducea and Saleeby, 1998). The western margin of the Indian Wells Valley was deformed at this time in conjunction with rotation and westward translation of the adjacent El Paso basin (Burbank and Whistler, 1987; Loomis and Burbank, 1988).

Finally, sometime after 3.5 Ma, but before 2 Ma, there was a regional shift in deformation to northwest-directed extension that resulted in cessation of movement on the Sierra Nevada frontal fault and initiation of simple right-lateral or right-oblique strike-slip faulting on a north-northwest-oriented system located in the central Indian Wells Valley. Sedimentation patterns changed from being basin-wide to highly localized, relatively low energy deposition in close proximity to the faults, i.e., marginal alluvial fans with central alluvial plains and restricted playas. Modern deformation of the valley is the result of this dominantly dextral faulting.

As the Pacific–North American plate boundary continues to evolve, it is likely the Indian Wells Valley will undergo further dynamic responses in sedimentation patterns. If the trends of the past 16 m.y. continue, the locus of extensional deformation will shift westward, and the Indian Wells Valley basin,

like Death Valley and Panamint Valley before, will eventually become a passive record of a small segment of this dynamic plate boundary. For the time being, however, it is one of the most tectonically active sites in the southwestern United States and the world.

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