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A Guide to Miocene Extension and Magmatism in the Lower Colorado River Region, Nevada, Arizona, and California

by

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A guide for Field trip 3
Eighth International Conference on Geochronology, Cosmochronology, and Isotope Geology
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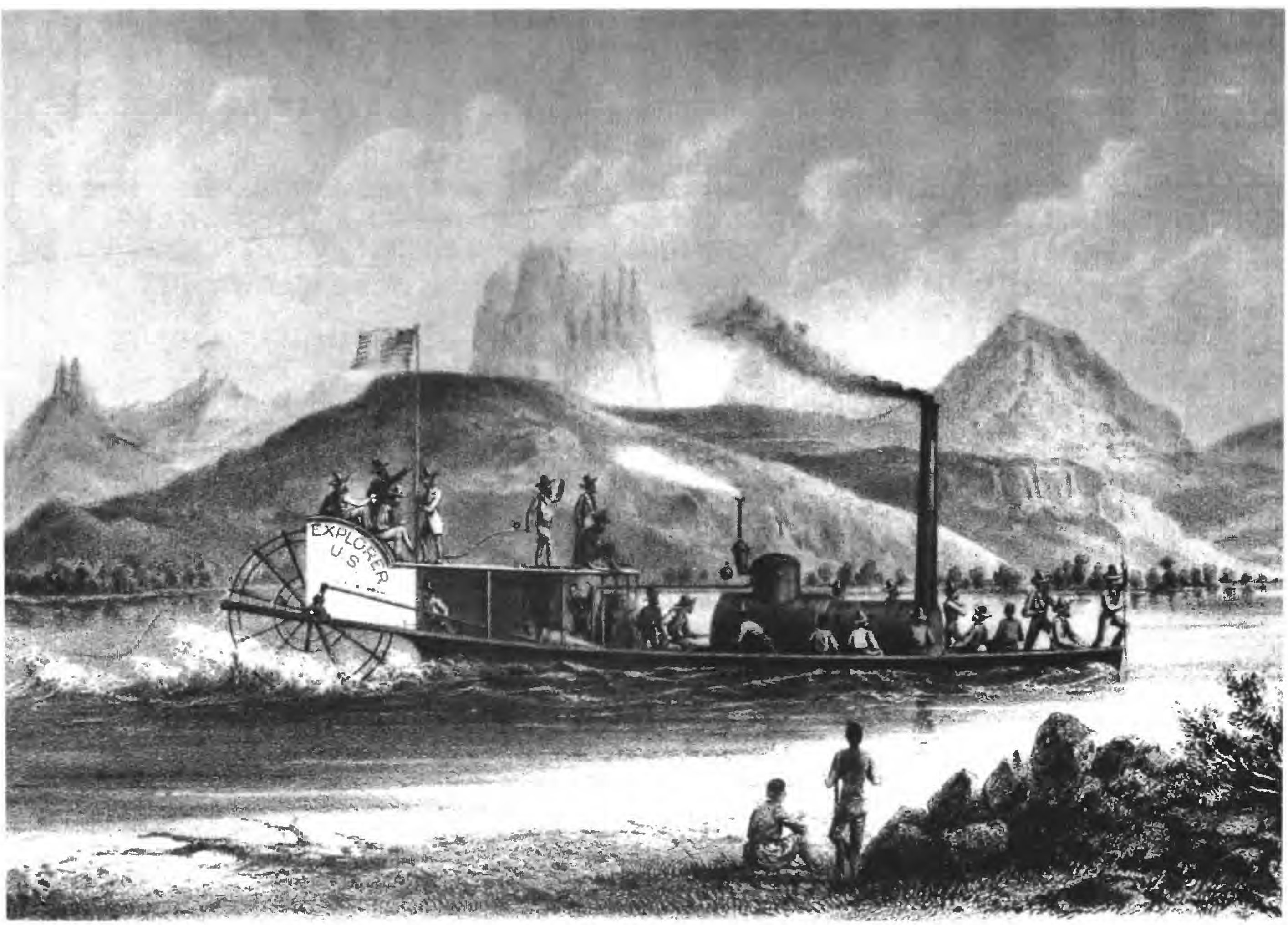


ABSTRACT

Extension and magmatism in the Colorado River extensional corridor occurred between 23 Ma and 13 Ma following Mesozoic crustal thickening and uplift and Paleogene quiescence. Crustal stretching and attenuation (by a factor ≥2) in the Miocene led to tilting of crustal slabs in fault blocks, to tectonic denudation of rocks formerly residing in the middle crust, and to faulting of upper crustal sections down against rocks from deeper levels. Volcanic and sedimentary rocks deposited during this episode were juxtaposed against dike swarms, plutons, and mylonitic gneisses formed at deeper crustal levels. Igneous rocks produced during the extension form a suite of intermediate, mafic, and silicic rocks that are typical of continental magmatic arcs. Low-volume asthenosphere-derived basalts preceded and followed the main magmatic pulse. An appended road log serves as a geologic guide to the extensional corridor.

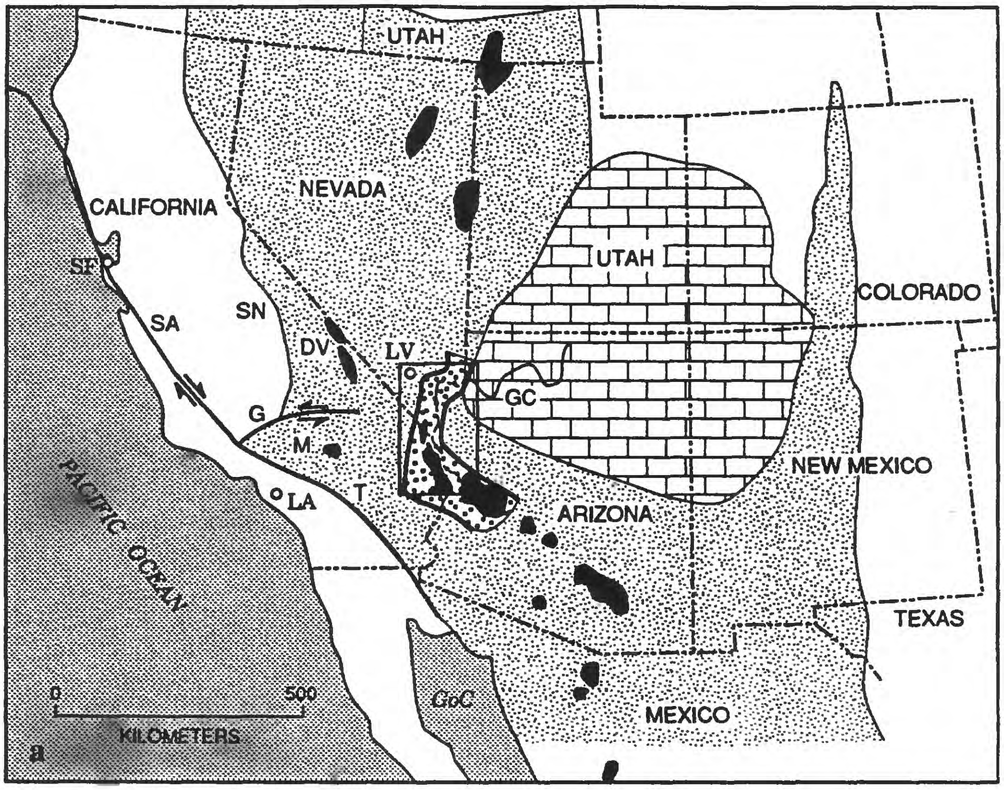
INTRODUCTION

Cenozoic extension has doubled the width of the Basin and Range province of western North America (Fig. 1a) and locally incresed it by a factor of 3-4 (Hamilton and Myers, 1966; Wernicke and others, 1988). In contrast to the northern Basin and Range province, where extensional faulting actively continues today, the southern Basin and Range, south from the latitude of Las Vegas, Nevada, is lower in elevation and now mostly seismically and tectonically inactive.

The Colorado River extensional corridor forms one of several subregions within the Basin and Range that exhibit features associated with extreme extension (Fig. 1a). The corridor exposes regional-scale gently-dipping extensional faults, steeply tilted beds in shingled fault panels, and metamorphic core complexes that were tectonically unroofed and uplifted from mid-crustal levels during Miocene extension. This region has been the fortunate target of seismic, gravity, and other geophysical studies as well as numerous structural, stratigraphic, petrologic, and thermochronologic studies. Superb exposure, a wealth of geologic and geophysical data, and structural juxtapositions that allow comparison of different crustal levels all make this an excellent place to study relationships between magmatism and continental extension. Many studies of this and surrounding regions are contained in a reprint volume of 38 Journal of Geophysical Research papers (1990-1991) available from the American Geophysical Union ("Cactis volume"--California-Arizona Crustal Transect Interim Synthesis).

Pre-Tertiary Framework

The region along the Colorado River in Nevada, California, and Arizona was part of the stable North American craton prior to Mesozoic time. A basement of Proterozoic orthogneisses yields U-Pb crystallization ages concentrated at ~1.7 Ga, which is the age of granulite- to upper amphibolite-facies metamorphism in the Ivanpah orogeny (Wooden and Miller, 1990). Isotopic studies have defined major



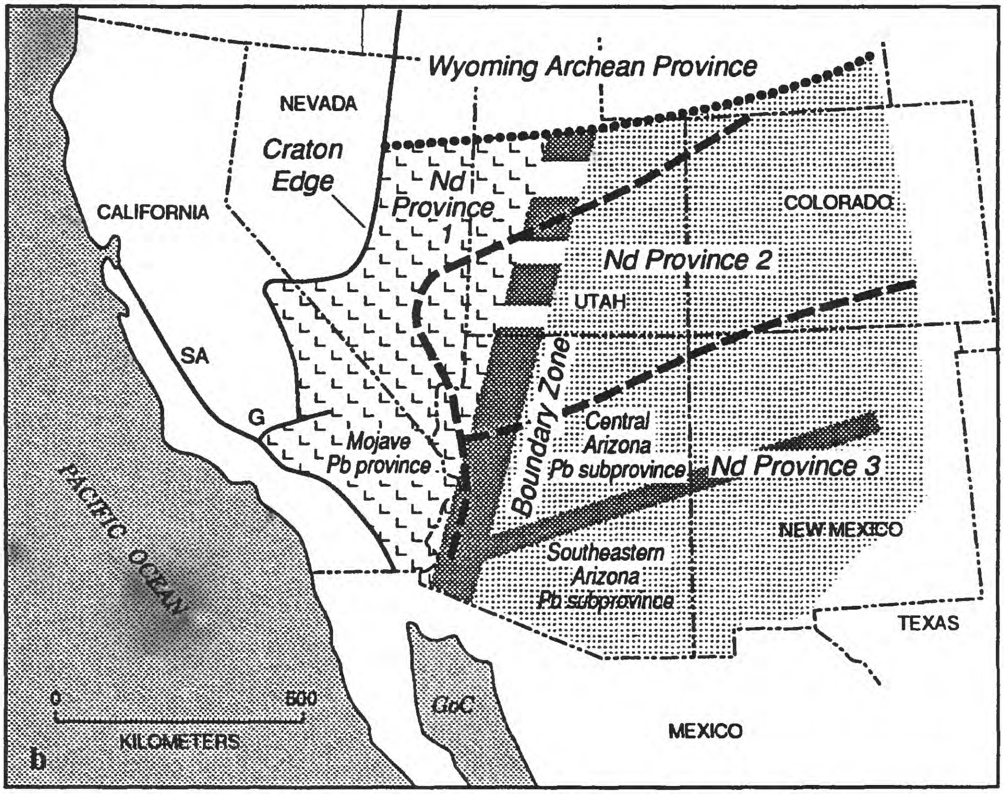


Figure 1a. Map of the southwestern U.S. showing the Colorado Plateau (brick pattern), metamorphic core complexes (black), Basin and Range province (stippled), and Colorado River extensional corridor (coarse dots). DV= Death Valley, G=Garlock fault, GC=Grand Canyon, GoC=Gulf of California, LV=Las Vegas, LA=Los Angeles, M=western Mojave Desert, SA=San Andreas fault, SN=Sierra Nevada, T=easternTransverse Ranges. Box shows area of Figure 2.

1b. Map of the southwestern U.S. showing crustal isotopic provinces (from Wooden and DeWitt, 1991). Patterned stipple is the Mojave Pb province, light stipple is the Arizona-Colorado-New Mexico Pb province, and the thick gray lines are (dashed where uncertain) the boundaries between the Pb Provinces or subprovinces (central Arizona vs. southeastern Arizona). The black dashed lines represent the Nd isotopic provinces of Bennet and DePaolo (1987), in which Nd model ages are 2.0-2.3 Ga for their Nd Province 1, 1.8-2.0 Ga for Nd Province 2, and 1.7-1.8 Ga for Nd Province 3.

crustal provinces in these basement rocks (Fig. 1b) (Bennett and DePaolo, 1987; Wooden and others, 1988; Wooden and Miller, 1990; Wooden and DeWitt, 1991). The Arizona subprovinces exhibit Nd and Pb model ages consistent with the crustal ages of ~1.7-2.0 Ga. To the west, the Mojave province in southern California and Nevada, exhibits older (≥2 Ga) model ages. Although no rocks have been identified as clearly older than Proterozoic in the Mojave province, late Archean U-Pb ages on zircons recovered locally from metaclastic rocks suggest a possible tie to the Archean Wyoming province (Wooden and others, 1994).

Middle Proterozoic (1.4 Ga) plutons intrude the gneisses. These plutons are part of a transcontinental magmatic belt, which extends to Labrador, that is commonly termed anorogenic (Anderson, 1983). A final Proterozoic igneous event was intrusion by 1.1-Ga diabase dikes that were emplaced as horizontal intrusive sheets throughout a region at least 600 km wide in California and Arizona (Hammond, 1990; Howard, 1991). The originally horizontal orientation of most of these Keweenawan-age sheets may indicate intrusion during a compressional or neutral stress regime; in any case they provide a useful structural reference to evaluate younger deformation in the gneissic and granitic basement rocks.

The Proterozoic rocks served as the depositional basement for layers of shallow-marine Paleozoic strata (Stone and others, 1983), exposed spectacularly in the Grand Canyon. During post-Paleozoic orogeny these strata were stripped from a region about 200 km across south of Las Vegas and centered on the Colorado River. The Paleozoic strata thicken dramatically to the northwest into the Cordilleran miogeocline along a hinge line that passes approximately through Las Vegas, recording an early Paleozoic passive margin.

The west coast of southwestern North America became an active convergent margin by late Paleozoic time, leading to a succession of magmatic and orogenic events through Mesozoic time that mobilized the crust eastward as far as the margin of the relatively stable Colorado Plateau province (Fig. 1a). Large-displacement, east-directed, Mesozoic thrusts of the Sevier orogeny form a belt through Wyoming, Utah and southern Nevada, and are spectacularly visible in the Spring Mountains west of Las Vegas (Burchfiel and others, 1974; Burchfiel and Davis, 1988). These thrusts place thick sections of Paleozoic miogeoclinal strata over much thinner cratonal equivalents. Deeply exhumed Jurassic and Cretaceous thrusts south of the latitude of Las Vegas involve the Proterozoic basement rocks and metamorphosed thin Paleozoic strata, and are typically expressed as mylonite belts (Burchfiel and Davis, 1988; Miller and others, 1982; Hamilton and others, 1987; Reynolds and others, 1988; Fletcher and Karlstrom, 1990; Tosdal, 1990). Mylonitic zones also formed during the emplacement of Late Cretaceous granitic plutons (Howard and others, 1987; John and Mukasa, 1990), and during shear associated with Late Cretaceous unroofing (Carl and others, 1991). Rapid cooling and uplift at the end of the Cretaceous has been documented by thermochronologic studies in the Old Woman Mountains 60 km west of the Colorado River (Fig. 2) (Carl and others, 1991; Foster and others, 1991, 1992).

Prior to extension the early Tertiary was a time of gradual erosion in the former thrust belt, and rocks of early Tertiary age are lacking in the lower Colorado River region. Highlands fed Paleogene streams that flowed northeasterly, toward the continent, across the southwest margin of the Colorado Plateau province (Elston and Young, 1991). The opposite, southwestern side of the highlands is marked by Paleocene and Eocene marine shorelines that have been identified in the western Mojave Desert and in the eastern Transverse Ranges (Fig. 1a). Cooling rates of 1 to 10 °C/m.y., based on $^{40}Ar/^{39}Ar$ and fission-track thermochronolgy, suggest that progressive erosion occurred throughout the Paleogene (Foster and others, 1990, 1991). Paleogene cooling totaling 100-150° documented in those studies suggests that perhaps 3-7 km may have been denuded before the onset of Miocene extension. As yet there are no compelling data on the Paleogene crustal thickness or elevation to help constrain whether overthickening of the crust during the Mesozoic orogeny led directly to Neogene extensional collapse (compare Coney and Harms, 1984; Spencer and Reynolds, 1990).

Neogene Setting

In the early Neogene the margin of southwestern North America was in the process of converting from a subduction margin to a transform margin (Dickinson and Snyder, 1979; Severinghouse and Atwater, 1990; Ward, 1991). The Mendocino triple junction between these domains migrated northward relative to North America. South of the triple junction and the Mendocino fracture zone, subduction of the East Pacific Rise (or of zero-age crust) on its eastern flank is commonly presumed to have resulted in a no-slab window under the southern California area. This predicted no-slab window underlay the area of the Colorado River extensional corridor at ~20 Ma (Severinghouse and Atwater, 1990), and thus coincided with onset of extension. Magmatism and extension migrated northward through the region of the southern Basin and Range province, coincident with the northward migration of the Mendocino triple junction (Glazner and

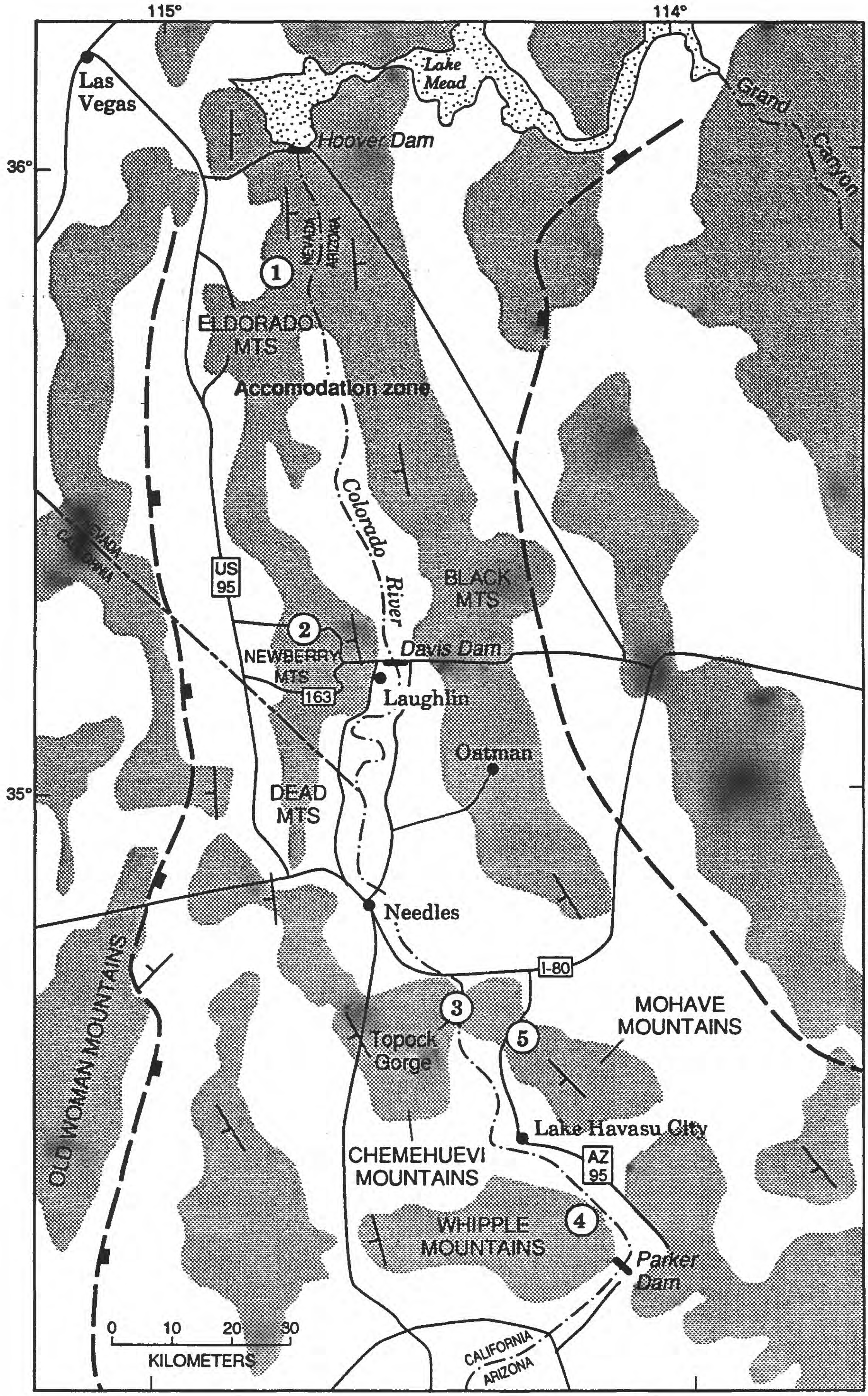


Figure 2. Map of the Colorado River extensional corridor and geographic features. Circled numbers show focus areas of field-trip days. Heavy dashed lines show the margins of the extensional corridor, with teeth along breakaway faults. Strike and dip symbols indicate tilt directions.

Supplee, 1982; Glazner and Bartley, 1984; Gans and others, 1992). Meanwhile magmatism (and probably extension) in the northern Basin and Range migrated southward toward an amagmatic corridor at the latitude of Las Vegas (Armstrong and Ward, 1991). The post-Cretaceous magmatic gap between these converging belts of magmatism was finally extended in the Miocene. Armstrong and Ward (1991) suggested that thermal softening of the lithosphere initiated the onset of rapid core-complex extension when magmatism effectively burnt this bridge (the post-Laramide magmatic gap) binding the Cordillera to North America.

The relationship between magmatism and extension, and the extent to which these two processes are necessary companions in the Basin and Range province, has been a topic of much debate (Gans and others, 1989; Best and Christensen, 1991). In the Colorado River extensional corridor, volcanism overlapped with extension in time but peaked slightly earlier than the highest rates of extension (Gans and others, 1989; Howard, 1993). This observation argues for the idea that crustal heating promoted or allowed the extension, and against the idea that many of the magmas are the result of decompression melting. The bulk of the synextensional igneous rocks resemble intermediate suites in continental volcanic arcs (Smith and others, 1990; Howard, 1993), but these rocks were preceded and followed by low-volume OIB-like (asthenosphere-derived?) basalts (Howard, 1993; compare Bradshaw and others, 1993; Feuerbach and others, 1993). The fundamental causes of the extension and the magmatism remain debated. Discussion continues as to whether extension and magmatism resulted from a failed, propagating slow-spreading ridge, as rifting above hot shallow asthenosphere in a no-slab window, from back-arc spreading, or from some other tectonic situation.

Tertiary Stratigraphy

The Tertiary stratigraphy throughout the Colorado River extensional corridor and adjacent regions was recently compiled by Sherrod and Nielson (1993). Tertiary rock sequences begin commonly with a basal arkose over deeply weathered pre-Tertiary crystalline rocks. Overlying volcanic rocks, in the central part of the corridor near Lake Havasu City and Needles (Figs. 2, 6), are dated between 23 and 18.5 Ma, and consist largely of intermediate, mafic, and silicic flows and breccias; in places they interfinger with landslide megabreccia deposits and other sedimentary rocks. The volcanic rocks reach several hundred meters in thickness, and are capped by the Peach Springs Tuff of Young and Brennan (1974). This widespread rhyolitic ignimbrite is dated at 18.5±0.2 Ma (Nielson and others, 1990). It covers an area 400 km wide from east to west, and originated near Laughlin, Nevada (Hillhouse and Wells, 1991) from a source yet to be identified. Coarse clastic rocks, landslide megabreccias, and interbedded basalt and tuff overlie the Peach Springs Tuff to thicknesses of ~2 km or more and were deposited during extension (Miller and John, 1988).

The Patsy Mine Volcanics form much of the Miocene section in the northern part of the corridor and are mostly younger than the counterpart section of pre-Peach Springs Tuff volcanic rocks in the central part of the corridor. The Patsy Mine Volcanics in turn are overlain by another ignimbrite, the tuff of Bridge Springs, dated at 15.20 Ma (Gans and others, 1994).

Unconformities within dated Miocene sections in the extensional corridor record faulting and tilting beginning ≥20 Ma and ending ≤12 Ma. Alluvial fan deposits and interbedded basalt flows younger than ~9-12 Ma overlie the tilted sections unconformably and document the end of extensional faulting.

Post-extensional deposits also include the Bouse Formation of late Miocene and Pliocene age and Pleistocene deposits of the Colorado River (Buising, 1993). The Bouse represents estuarine and deltaic deposition in a protoGulf of California before the through-going Colorado River was established (Buising, 1993). The Colorado River rises in the central Rocky Mountains of Colorado and Wyoming, cuts through the Grand Canyon in the Colorado Plateaus province and through low deserts of the southern Basin and Range province, then empties into the modern Gulf of California, where Baja California has pulled away from mainland Mexico. The lower Colorado River valley of interest to this guide consists of alluviated valleys interspersed with bedrock ranges that the river incises.

Structural Framework of the Extensional Corridor

The northern part of the Colorado River extensional corridor was the site of R.E. Anderson's (1971) seminal study on listric and low-angle normal faults. He recognized that the steep dips of slivered Miocene units meant they had been rotated along gently dipping normal faults, and his analysis helped trigger a revolution in understanding extension of the continental crust.

The extensional corridor is 50 to 100 km wide and displays a common set of structural elements (Fig. 3). They consist of a breakaway or headwall fault on the updip side of the fault system, numerous fault blocks tilted toward the breakaway, a central zone of domal uplifts (metamorphic core complexes) where the fault system is antiformal, and a zone where the fault system roots downdip under less deformed blocks. The Colorado River extensional corridor (Howard and John, 1987) was originally defined as the domain of southwest or west stratal dips (Whipple tilt domain of Spencer and Reynolds, 1991), but common use has since broadened the term to embrace a domain of east stratal dips north. An accommodation zone (Fig. 2) where the tilt reverses also marks a flip in fault vergence, north of which fault hanging walls slipped westward relative to footwalls (Faulds and others, 1990). Faulds and others (1990) found that the amount of extension decreases at the accommodation zone but the timing of extension was unaffected.

The great domal detachment faults are fundamental elements of the upper crustal structure (Davis and others, 1980; Davis and Lister, 1988; John, 1987a). Mylonitic fabrics in the footwalls represent ductile extensional shearing at deep crustal levels (Davis and others, 1980), and were overprinted successively by cataclasite, breccia, and gouge, as the fault systems were progressively unroofed and rose toward the surface (John, 1987a). This rise is recorded also in geobarometric and thermochronologic studies. For example, rocks exposed in the core of the Whipple Mountains rose from ~30 km depth in the Late Cretaceous through ~16-km depths in the latest Oligocene or early Miocene when extension began, and were tectonically unroofed by the late Miocene (Anderson, 1988; Davis, 1988). The asymmetric shape and slip on the detachment fault system before their doming has been used to argue for a simple-shear model of crustal extension (Wernicke, 1985; Howard and John, 1987).

Geophysical measurements, on the other hand, suggest an overall symmetrical crustal section, the deepest part of which is compatible with McKenzie's (1978) pure-shear model of crustal stretching. This symmetry is shown by elongate gravity and magnetic highs centered over the core complexes (Simpson and others, 1990; Mickus and James, 1991; Campbell and John, 1994), and by a seismically defined laccolith-shaped mid-crustal body of moderate seismic velocity that is centered under the core complexes (Fig. 3)(McCarthy and others, 1991; Wilson and others, 1991). The midcrustal bulge at a depth of 10 km (Fig. 3) is consistent with structural and petrologic evidence that the core complexes were uplifted 10-15 km during extension (Howard and others, 1982a; Anderson, 1988). Rise of the bulge has been likened to diapirism and inward intracrustal flow as an isostatic response to tectonic unroofing during extension (Wilson and others, 1991; McCarthy and others, 1991; Spencer, 1984; Block and Royden, 1990). The seismic studies of McCarthy and others (1991) demonstrated that the Moho is nearly flat and 26-30 km deep across this part of the Basin and Range. Crustal thickness is therefore constant in this part of the southern Basin and Range province despite variations in amount of upper crustal extension. Similar observations for the northern Basin and Range were used to argue that abundant lateral flow in a ductile lower crust accounts for pervasive crustal thinning across regions where the upper-crustal manifestation of extension is markedly variable (Gans, 1987).

Active debate rages over the degree to which magmatism is incidental to, explains details of, facilitates, or drives continental extension. One postulated effect of magmatic intrusion is thermal softening that can allow rapid stretching (Armstrong and Ward, 1991; Lister and Baldwin, 1993) and rise of a domal core complex (Brun and others, 1994). Another important role of magmatism may be its influence on the state of stress and therefore the mode and orientation of rock failure and faulting (Parsons and Thompson, 1993).

The orientation of stress in areas of extreme continental extension has been puzzled over since the discovery in the 1970s that many areas exhibit gently dipping normal-slip faults (Anderson, 1971; Armstrong, 1982). According to the theory of Anderson (1951), normal faults are expected to form with moderately steep dips (45°-70°), assuming that the least and intermediate principal compressive stresses are aligned parallel to the Earth's surface and the greatest principal stress is vertical. Many, perhaps most low-angle extensional faults may have been rotated from originally steeper dips after the faults were active (Proffett, 1977; Spencer, 1984). The rolling-hinge model postulates that footwalls to moderately steep normal faults may rebound isostatically as they are unloaded, thus arching inactive portions of the fault to gentle dips (Buck, 1988; Wernicke and Axen, 1988; Hamilton and Howard, 1991). Geologic and thermochronologic relations have been used to argue that in other cases the faults formed at dips much less than predicted by Anderson's (1951) theory (Wernicke and others, 1985; John and Foster, 1993). If so, other factors such as high fluid pressures (Bartley and Glazner, 1985) or magmatic pressures (Parsons and Thompson, 1993) may be involved that could cause deviations of the maximum compressive stress from vertical. Parsons and Thompson (1991, 1993) argued that magmatic intrusion during extension may influence the orientations and relative magnitudes of the principal stresses, both through the magmatic pressure and by the anisotropic addition of volume upon intrusion. Parsons and Thompson (1993) suggested that intrusive magmatism can have major effects on the local stress field, and may allow the formation of low-angle normal faults within the seismogenic regime.

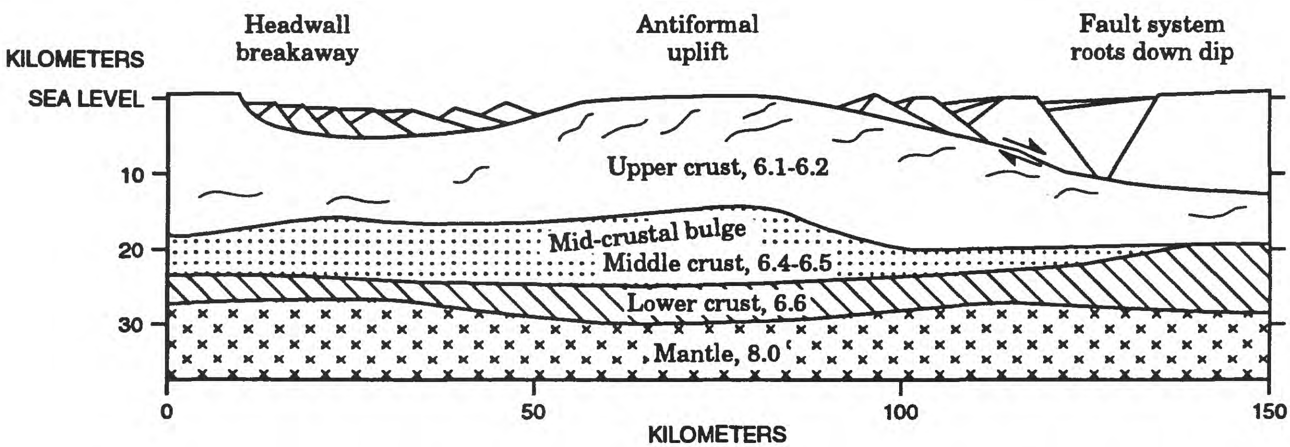


Figure 3. Schematic crustal section of the Colorado River extensional corridor, modified from Spencer (1984) and McCarthy and others (1991). Seismic velocities are given in km/sec tyfor McCarthy and others' (1991) velocity model across the Whipple Mountains area.

Intrusive Rocks

Feeders for the Miocene volcanic rocks in the corridor include necks, dike swarms, rare calderas, and (near Hoover Dam) stratovolcano complexes. The dike swarms commonly strike perpendicular to the extension direction indicated by fault movements (Anderson, 1978; Davis and others, 1982; Nakata, 1982), but other dike swarms are parallel or oblique (Spencer, 1985; John and Foster, 1993). In places dike swarms are sheeted and occupy 15-100% of the the exposed rocks (Davis and others, 1982; Nakata, 1982; Spencer, 1985).

Miocene plutons ranging in composition from diorite to granite crop out in many ranges along the central part of the corridor (Anderson, 1969; Anderson and others, 1972; Weber and Smith, 1987; Metcalf and others, 1993; Hopson and others, 1994; J. Wooden and K. Howard, unpub. data; Campbell and John, 1994; John and Foster, 1993; Nakata and others, 1990; Wright and others, 1986; Anderson and Cullers, 1990; Bryant and Wooden, 1989). Some of the plutonic rocks, within the metamorphic core complexes, exhibit mylonitic fabric produced by extensional shear at mid-crustal levels (Anderson and Cullers, 1990). Concentration of mafic intrusions under the core complexes in the central part of the corridor is suggested by the gravity high (Simpson and others, 1990; Campbell and John, 1994).

Structural attenuation by faulting and tilting within the corridor has resulted in the exposure of a wide variety of crustal depth levels, thus allowing comparison of structures and igneous bodies formed at levels ranging from the surface to deep in the middle crust. Tilted fault panels, now eroded, display oblique or upended cross sections of Precambrian rocks or of Tertiary intrusions reaching paleothicknesses of 15 km (Howard and others, 1982a, in press). Paleomagnetic studies have confirmed the tilting (Faulds and others, 1990; Pease, 1991).

Thermochronology

Thermochronologic studies using $^{40}Ar/^{39}Ar$, K-Ar, and fission-track dating have revealed much about the thermal and deformation history of the region (Dokka and Lingrey, 1979; Davis and others, 1982; Foster and others, 1990, 1991, 1992, 1993; Knapp and Heizler, 1990; Richard and others, 1990; Nakata and others, 1990; Bryant and others, 1991; Fitzgerald and others, 1991; Foster and Spencer, 1992; John and Foster, 1993). In the Whipple Mountains, for example, nearly concordant mineral ages suggested that unroofing rates reached as high as 3 km/m.y. or more (Davis, 1988) (alternatively the concordant ages have been explained as due to synextensional intrusion--Lister and Baldwin, 1993). In the Chemehuevi Mountains, extension began ~23 Ma and mineral ages in the footwall of the detachment fault become younger progressively northeastward, in accord with uplift of deeper rocks in the downdip direction (John and Foster, 1993). That study concluded that slip rates of denudational faults, as measured by the rate of northeastward younging in the footwall, reached 3-8 km/m.y. between 19 Ma and 15.

Before the Twentieth Century

In 1540 the lower Colorado River was reached by two of the early Spanish explorers from Mexico (Darton, 1916). Melchior Diáz, arriving overland near the river's mouth into the Gulf of California, named it Río del Tizón (Firebrand River) after the custom of the natives of carrying firebrands in winter with which to warm themselves. Hernando de Alarcón explored up the river a short way by boat. The Spanish priest, Padre Fray Francisco Garcés, in 1776 found indigenous peoples of several tribes living along the river as he explored farther up the river, to the area covered by this guide. His exploration and that of subsequent ones were nicely summarized by Thompson (1929). Mohave Valley, now occupied by Laughlin, Nevada, and Bullhead City, Arizona, was peopled when Garcés arrived, by the "Jamajabs" tribe now called the Mohave or Mojave, a corruption of the native name Aha-makave, said to mean beside the water (Bahti, 1968) or to be the Yuman name for the three pinnacles called The Needles south of Topock (Darton, 1916). In 1826 a party led by Jedediah S. Smith explored down the Colorado River from what is now Utah. Upon his return later that year, the Mohaves seemed as friendly as usual, but as his party of 19 men was ferrying across the Colorado River on a raft the Indians attacked them and killed 10. In 1829 a party of trappers including Kit Carson passed through the Mohave villages on their way across the desert, and back again, from Santa Fe, New Mexico to the San Gabriel Mission near present-day Los Angeles (Thompson, 1929).

Living in villages along the flood plain of the Colorado River, the Mohaves caught fish from the river and snared a few rabbits in the desert, but lived mainly on mesquite beans, the hearts of yucca plants, and the fruits of the cactus (Powell, 1895). Life in a Mohave village was described in *The Captivity of the Oatman Girls* (Stratton, 1857), the story of two immigrant daughters of the Oatman family who survived a massacre of their family by Yavapai warriors in southern Arizona in 1851. They were subsequently traded as captives to the Mohaves with whom they lived in Mohave Valley for several years before one sister died and the other was eventually ransomed to the whites. Described as nationalistic and warlike, the Mohave fought with neighboring tribes and often traveled great distances to make war on other groups (Stratton, 1857; Bahti, 1968). Yet they maintained friendly relations with some tribes including the nomadic Chemehuevi, who lived in the Mojave Desert west of the river and relied not only on the river but on rare springs found in the desert mountains (Powell, 1895). The Chemehuevis were the original "Digger" Indians so-called by all the other tribes (Powell, 1895). Ives (1861) described the Chemehuevis as having

> ..small figures, and some of them delicate, nicely-cut features.... Unlike their neighbors--who, though warlike, are domestic, and seldom leave their own valleys--the Chemehuevis are a wandering race, and travel great distances on hunting and predatory excursions. They wear sandals and hunting shirts of buckskin, and carry tastefully-made quivers of the same material. They are notorious rogues, and have a peculiar cunning, elfish expression, which is a sufficient certificate of their rascality. One of them tried to cheat me while fulfilling a bargain for a deerskin; but I detected him at it, and, in spite of his denial, proved the fraud upon him. He was highly amused at being fairly caught, and it raised me very much in his estimation.

In the 1840s a wagon trail to the California goldfields ran through Mohave tribe territory. When Ives (1861) visited the villages of the Mohaves, he found that

> Their minds are active and intelligent, but I have been surprized to find how little idea of the superiority of the whites they have derived from seeing the appliances of civilization that surround those whom they have met......In most respects they think us their inferiors. I had a large crowd about me one day, and exhibited several things that I supposed would interest them, among others a mariner's compass. They soon learned its use, and thought we must be very stupid to be obliged to have recourse to artificial aid in order to find our way.

Friction with immigrants culminated in a full-scale attack on a wagon train in 1858. Fort Mohave was built by the U.S. Army in Mohave Valley the following year to maintain peace. The Fort Mohave reservation was set aside in 1865, and Mohave tribe members live and work in the valley today. In 1905 about 900 Mohaves were living in the region (Thompson, 1929).

The Treaty of Higaldo in 1848 following the Mexican-American War ceded the territories of California and New Mexico (including Arizona) to the United States. The discovery of gold in California the same year brought hordes of Americans seeking their fortune to the California gold fields in the Sierra Nevada east of San Francisco. The U.S. Army's explorations of Mohave Valley area of the new territory included an expedition led by Capt. L. Sitgreaves in 1851. An expedition in 1854 led by Lieutenant A.W. Whipple (1856) reconnoitered for a prospective railroad route along the 35th parallel that would cross the Colorado River near present-day Needles and Topock in the southern Mohave Valley. Although a northern route through Utah became the first transcontinental railway to be built, the southern railroad was eventually completed across the Colorado River in 1883 near Whipple's route, and geographic and geologic highlights along it were described by Darton (1916; see also Lee, 1908).

In 1858 an Army party led by Lieutenant Joseph Christmas Ives explored the Colorado River's navigability from its mouth to the Eldorado Mountains using a small paddlewheel steamboat (Frontispiece). The boat *Explorer* had been built in Philadelphia, shipped on steamers from New York and across the Panama isthmus to San Francisco, and then carried by schooner back around the tip of Baja California and to the mouth of the Colorado River. The Ives expedition aimed to explore the potential use of the Colorado River as a supply route for federal troops in Utah. Ives' (1861) accounts of the Indians and the scenery are beautifully prosaic, and drawings from his report (Frontispiece and Figs. 8-10) are artistically lovely (although a few by one artist greatly exaggerate the topography, see Stegner, 1953). Dr. John Strong Newberry was appointed physician and chief naturalist to the expedition and made extensive geological observations. He was later to become Director of the Ohio survey, Professor of geology and paleontology at the the Columbia School of Mines, an influential trigger for establishment of the U.S. Geological Survey, and chief mentor of the eminent geologist Grove Carl Gilbert (Rabbitt, 1979; Pyne, 1980).

In 1861 a ferry was established across the Colorado River at Fort Mohave, and a regular stage line carried freight from California to mines in Arizona (Thompson, 1929). In the early 1860s, the Black Mountains (previously also known as the River Range), where Oatman now lies, was then known as the

Blue Range. Owing to the hostility of the Piute and Hualapai Indians to the east, explorations were confined to a limited district until 1865, when a daring party of miners ventured into the next range east, the Cerbat range, only to be massacred, with the exception of one (Conners, 1913).

Benjamin Silliman Jr., of Yale, son of the first geology professor in America and co-editor of the American Journal of Science with his brother-in-law James Dwight Dana, described a summer journey in 1864 from Los Angeles through Fort Mohave and the San Francisco (Oatman) Mining District. He related coming across the evidence of a 1857 Indian massacre of a party of more than 70 Texan and Arkansan emigrants in Massacre (now Sacramento) Valley on the east side of the Black Mountains (Silliman, 1866). In the 1870s famous or later-to-be famous visitors to the region included the topographer Lt. G.M. Wheeler and geologist G.K. Gilbert (Rabbitt, 1979).

Objectives of Field Trip

This field guide is designed to accompany Field Trip 3 of the Eighth International Conference on Geochronology, Cosmochronology, and Isotope Geology, May 31 - June 4, 1994. The trip will progress southward down the extensional corridor and generally will go from higher structural levels to deeper ones (Fig. 2). Day 1 will be spent in the Eldorado Mountains investigating Miocene volcanic rocks, their rate of tilting, and possible relations between rapid tilting and the presence of underlying magma chambers. Day 2 will address plutonic rocks in the Newberry Mountains and detached volcanic and subvolcanic cover in the Black Mountains, and will introduce the Chemehuevi metamorphic core complex. Day 3 will investigate relationships between tilting and folding of cover rocks to deep-seated detachment faults in Topock Gorge through the Chemehuevi and Mohave Mountains. Day 4 will be spent in the Whipple Mountains metamorphic core complex where we will visit mylonitic intrusions emplaced before and during the extension. Day 5 will address a dike swarm in an upper-plate tilted block that exposes 15 km of crustal section in the Mohave Mountains.

GUIDE AND ROAD LOG

DAY 1

Eldorado Mountains -- P. Gans

The first day of the field trip serves three purposes. The first is to gain an overview of the northern part of the highly extended Colorado River extensional corridor. The second is to examine tilted fault slices exposed in the Eldorado Mountains and discuss evidence for the rate of extensional strain expressed by tilting and faulting. The third is to examine the exposed cross section of a tilted caldera formed during the extension.

Highlights on the Way to Stop 1.1

Go east on Tropicana Avenue in Las Vegas toward U.S. 95. West of Las Vegas behind you, the 11,000-ft-high Spring Mountains magnificently expose Mesozoic thrusts of the Sevier orogenic belt, which place gray Paleozoic limestones eastward over red Jurassic Aztec Sandstone. (You will be able to see this relation clearly when you leave Las Vegas to fly to the Bay Area--see the end of this guide) The allochthonous Paleozoic section in the Spring Mountains is miogeoclinal-facies, and much thicker than an autochthonous time-equivalent section of cratonal-facies strata that is exposed ahead to the east-northeast on Frenchman Mountain on the east side of Las Vegas.

Enter U.S. 95 southbound, and follow it through Henderson toward Boulder City and Searchlight. South of Las Vegas about 12 mi (18 km) the road passes through Railroad Pass. Mountains on the west side of this pass expose a Miocene granite pluton, and lower hills on the east expose Miocene volcanic rocks. South of Las Vegas 15 mi (24 km), exit right to continue on U.S. 95 south toward Searchlight.

Stops 1.1 and others for Day 1 are in the Eldorado Mountains

P. Gans

The Eldorado Mountains expose steeply east-tilted Miocene volcanic rocks on gently west-dipping extensional faults (Anderson, 1971). More than 30 new $^{40}Ar/^{39}Ar$ ages combined with superbly ecposed relations tightly constrain the magmatic and structural evolution of this portion of the Colorado River extensional corridor (Gans and others, 1994). Eruptions began ~18.1 Ma and continued without pause until ~13.0 Ma, resulting in a section ≥4.5 km thick. The section is dominated by mafic lavas. Silicic rocks first erupted ~15.7 Ma, and subsequently every ~150 k.y. for 750 k.y. The 15-km-wide, partly exposed caldera source of the 15.20-Ma tuff of Bridge Spring is tilted 90° east. Rapid large-magnitude east-west extension began immediately after the caldera collapse (Gans and others, 1994).

Highlights at the End of the Day on the Way to Laughlin

Travel south along U.S. 95 through Searchlight, where Miocene intermediate-composition volcanic rocks are exposed in roadcuts. Beyond Searchlight 19 mi, turn left (E) on Nevada highway 163 across the Newberry Mountains about 20 miles to Laughlin and lodging for the next two nights. Along Nevada 163, the road passes first along the dark, coarse-grained, informally named Newberry granite of Anderson and Bender (1989), thought to be 1.4 Ga (top left part of the map area of Fig. 4). Ensuing outcrops of light-colored granite forming the center of the Newberry Mountains are part of the Miocene Spirit Mountain pluton, crossed by swarms of north-trending Miocene dikes. This pluton will be tomorrow morning's destination.

DAY 2

The purposes of Day 2 are to examine the tilted depth section, in the Newberry Mountains, of a plutonic and dike complex emplaced during extension, to visit supracrustal equivalents exposed as Miocene volcanic rocks in the Black Mountains, and to introduce the Chemehuevi metamorphic core complex.

Stops 2.1 and 2.2

Spirit Mountain pluton in the Newberry Mountains --P. Gans

The peak in the Newberry Mountains now called Spirit Mountain (formerly Dead Mountain) was and is still revered by the Mohave people. Olive Oatman, captive to the Mohaves in the 1850s, recounted according to Stratton's (1857) book a Mohave legend of a flood above which only this peak stood, and the peak was afterward descended by red men who went south and whites who went north. A Mohave named Ireteba told Ives (1861) that the mountain is the abode of their departed spirits, and that should any one dare to visit it he would be instantly struck dead.

The Newberry Mountains were considered by Mathis (1982) and Spencer (1985) to constitute a lower-plate core beneath a domed Tertiary extensional detachment fault. Recent studies by Hopson and others (1994) suggest that the core instead forms a block tilted to the west, and bordered only on the east by a major exposed normal fault. Much of this core is occupied by Miocene granite that forms the Spirit Mountain pluton. Hopson and others (1994) interpret the pluton as west-tilted beneath a roof of Proterozoic granite, and cut by now-tilted dike swarms.

Highlights between Stops 2.2 and 2.3

From the intersection with Nevada highway 163 at the east foot of the Newberry Mountains west of Laughlin, drive south on River Road toward Needles. In 5.6 miles (2 miles past El Mirage development), pull off carefully to the right at a dirt path, and park.

Stop 2.3

Mirage pluton and Newberry detachment fault -- K. Howard

A canyon is cut into faulted exposures of a pluton here called the Mirage pluton (Fig. 4). This pluton consists of granite and granodiorite; here it is a porphyritic sphene-biotite monzogranite. The alkali feldspar phenocrysts in this rock are commonly rimmed by plagioclase, a common attribute of Tertiary plutons in southeastern California and southwestern Nevada (Bennie Troxel, oral commun., 1989). The Mirage pluton intrudes the Spirit Mountain pluton, and has yielded a middle Miocene U-Pb zircon age (J.L. Wooden, unpub. data). In this area the Mirage is chopped by many faults that strike subparallel to the overlying Newberry detachment fault. The exposed Mirage pluton is 4-5 km across, varies in igneous texture from porphyritic to equigranular and from fine grained to medium grained, and locally exhibits a mylonitic fabric and ENE-striking lineation.

Walk 200 m north along the base of the outcrop to the exposed Newberry detachment fault of Mathis (1982) (not to be confused with the Newberry Mountains detachment fault of Dokka, 1993, which occurs 200 km to the southwest in California near a different range of Newberry Mountains). The fault dips gently east, and slickenside striae on polished surfaces of the fault strike east-northeast. Work in progress by C.A. Hopson (written commun., 1994) suggests that the fault here does not arch over the range, and "detachment" may be a debatable term. In any case it is a major fault that places red, oxidized rocks of the coarse-grained Middle Proterozoic, informally named Davis Dam granite of Anderson and Bender (1989) (correlative with the informally named Newberry granite of Anderson and Bender, 1989) on rocks of the

Mirage pluton. This superposition omits substantial crustal section and represents many kilometers of displacement (Frost and Martin, 1982). Mathis (1982) described anastamosing minor faults in the footwall.

Highlights between Stops 2.3 and 2.4

Continue south on River Road 2.1 miles. Turn right (W) on a dirt pipeline road. Proceed west 2.5 miles and turn right (N). Proceed north 2.1 miles (past 2 power lines) and stop on a hill crest for views of the Mirage and Spirit Mountain plutons and dike swarms, the most prominent of which is a swarm of south-striking microgranite dikes.

Stop 2.4

Southern Newberry Mountains dikes -- K. Howard

This dike swarm is unique in that it curves markedly in strike where it approaches the youngest major intrusion in the area, the granitic Mirage pluton (Fig. 4). The strike of dikes curves such as to intersect the arcuate edge of the Mirage pluton at high angles rather than tangentially. This pattern of dike curvature resembles the theoretical pattern of greatest principal stress trajectories where a regional stress field is perturbed by a pressurized hole in an elastic plate (Fig. 5), as used to model dike patterns around the Spanish Peaks, Colorado, intrusive center (Odé, 1957; Muller and Pollard, 1977), and Galapagos shield volcanoes (Chadwick and Howard, 1990). If the dike curvature relates to stresses accompanying intrusion of the Mirage pluton, it offers a way to assess directly the influence of synextensional pluton intrusion on the direction of principal stresses. This in turn offers a test of the potential for pluton intrusion to reorient the stress field, which is one proposed explanation for the conundrum of low-angle extensional faulting (Parsons and Thompson, 1993). Dike curvature and the radius of the intrusive center potentially can be used to calculate the far-field differential stress (in two dimensions) and its ratio to the magmatic pressure that perturbs it (Muller and Pollard, 1977).

Field reconnaissance suggests a sequence of cross-cutting dike sets near and in the Mirage pluton (mafic dikes, mylonitized microgranite, quartz porphyry, aplite, quartz veins). These cross-cutting stress indicators and their relation to the pluton offer the potential to track the evolution of paleostress directions, to relate them to geometry and timing of pluton intrusion, and to study possible rotations of the footwall. Rotation elsewhere in the Newberry Mountains block has been suggested by some (not all) paleomagnetic results and by study of dips of dikes and foliations (Faulds and others , 1992; Hopson and others, 1994). Our field reconnaissance farther south near the Mirage pluton indicates that dike dips differ from those described by Hopson and others (1994), so the tilt requires further analysis. Some of the curving dikes near the Mirage pluton are gently dipping, pre-Mirage mafic dikes, and their orientation may indicate wallrock deformation related to Mirage pluton emplacement, or possibly (C. Hopson, written commun., 1994) to drag beneath overlying faults. Other curving dikes are steeply dipping and composed of microgranite that closely resembles the border phase of the Mirage pluton. If they emanate from that pluton, it would build a strong case for curved paleostress trajectories related to pluton emplacement.

Geochronologic studies in progress are expected to lead to a refined chronology for the sequence of Miocene intrusion and structural development in the Newberry Mountains. Current $^{40}An/^{39}Ar$ study of the Spirit Mountain pluton by P.B. Gans and U-Pb zircon dating by J.L. Wooden build on early K-Ar dating by Volborth (1973), Anderson and others (1972), and Spencer (1985). Preliminary results (Wooden and Howard, unpub. data; Gans, unpub. data) suggest that the succession of intrusive events recorded by the dike swarms and plutons in and near the southern Newberry Mountains are 14-18 Ma and therefore span a 4-m.y. period when the Colorado River extensional corridor was undergoing rapid ductile and brittle extensional deformation (John and Foster, 1993; Gans and others, 1994). Mylonitic fabrics are developed in some dikes and locally in the plutonic rocks.

Highlights between Stops 2.4 and 2.5

Return 4.6 mi downhill to River Road and turn right (S) 18 mi to Interstate Highway 1-40 at Needles, California. Along this route the road travels between the Colorado River and the Dead Mountains metamorphic core complex. The Dead Mountains expose undated orthogneisses that are intruded by undeformed Miocene granite and quartz diorite (Fig. 4). Gently dipping mylonitic foliation in the southern Dead Mountains bears a lineation that strikes east to southeast. This direction contrasts with the generally ENE strike of Tertiary lineation in the Newberry, Sacramento, and Whipple Mountains. A lower-plate position of the gneissic core of the Dead Mountains is nevertheless indicated by a dark klippe of red Tertiary sandstone and volcanic rocks on the southeast flank of the range, and by a klippe of Proterozoic granite and diabase sheets on the northwest flank of the range (Howard, 1991). Thick cataclasites mark additional low-angle extensional faults within the lower plate in the southern part of the range. A vertical to steep reverse

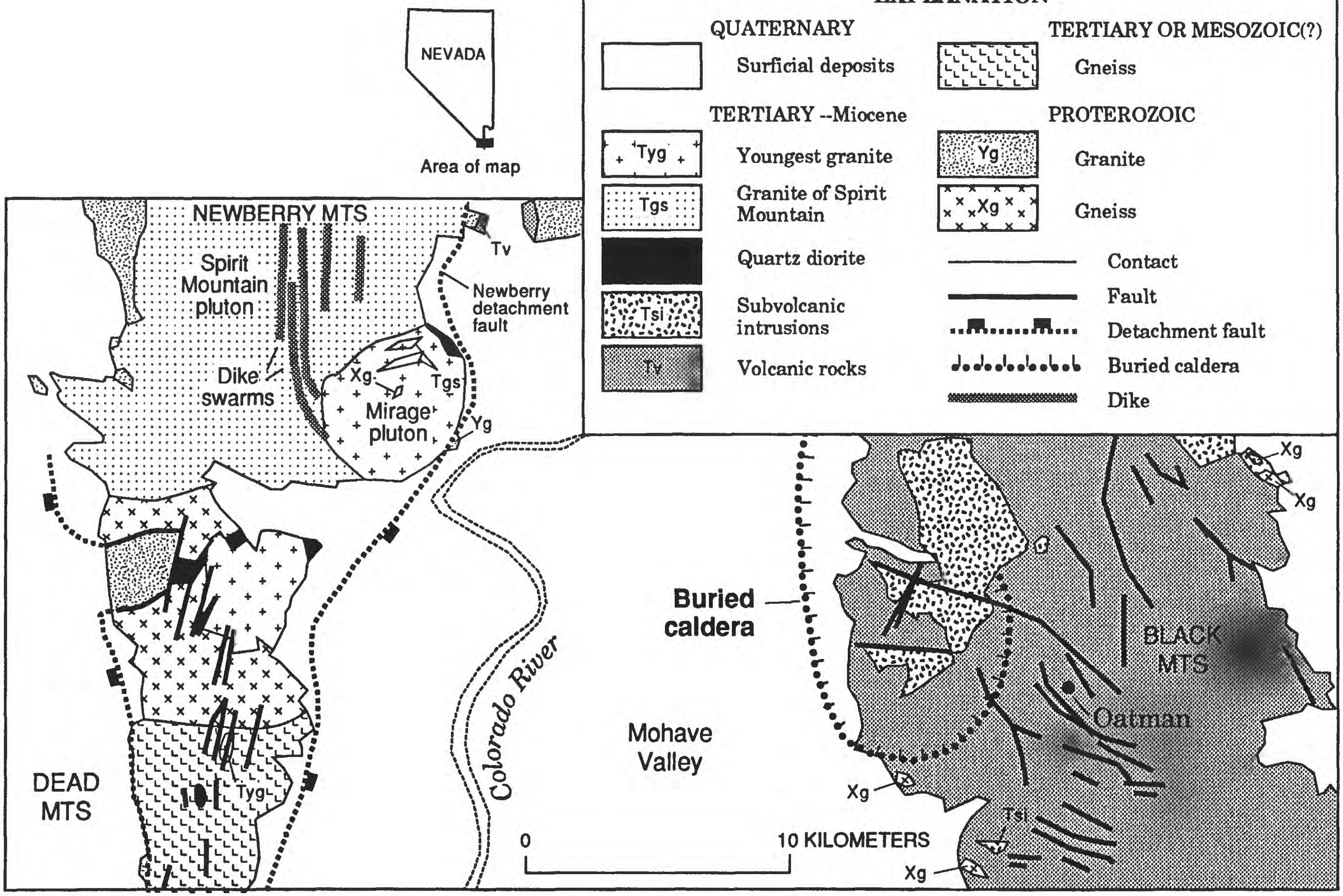


Figure 4. Geologic map of northern Mohave Valley and the adjacent southern Newberry Mountains, northern Dead Mountains, and southern Black Mountains. Mapping in the Black Mountains is after Thorson (1971).

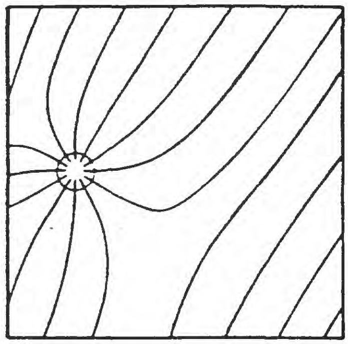


Figure 5. Stress trajectories in an elastic plate, showing the interaction of a pressurized hole on a far-field stress field (Muller and Pollard, 1977).

fault bounds much of the west side of the range against downdropped granites and against sandstone interbedded with 12-Ma basalt (Spencer, 1985). Foliation in the gneisses is dragged sharply down westward at this fault. Its post-12-Ma age indicates that this, and other vertical north-striking faults in the Dead and (farther south) Sacramento Mountains, postdate most of the extensional movement. The faults may reflect adjustments to the doming of the northward-elongate core complexes that they border (Spencer, 1985)

Near the south end of the Dead Mountains, enter Highway I-40 and follow it about 4 mi eastbound (S) through Needles, then exit right on U.S. 95 south toward Parker. The domal Sacramento Mountains (to the west at 3 o'clock) and Chemehuevi Mountains (ahead left) are metamorphic core complexes (Fig. 6). They form domed windows of footwall crystalline rocks surrounded by lowlands exposing red-colored hanging-wall rocks. Synextensional conglomerate, interbedded volcanic rocks, and their depositional basement (plutonic and gneissic rocks) dominate the hanging wall. Mylonitized plutonic rocks in the Sacramento Mountains show a thermochronologic and stratigraphic record of progressive uplift, from ~16 km depth and > 450°C at 22-24 Ma, to exposure and overlap by volcanic rocks at 14 Ma (Anderson, 1988; Foster and others, 1990; Simpson and others, 1991). They are currently under study by Victoria Pease (Oxford University).

In 3.8 miles keep right (southwest) at the junction (following U.S. 95). At mile 6.8, the road begins to climb gradually toward the south and west through exposures of cataclasite in the footwall of the Chemehuevi detachment fault. At 2 o'clock, the footwall in the southern Sacramento Mountains (between Lobecks and Monumental passes) is composed of granitoid rocks including hornblendite and gabbro, diorite, and granite, together informally called the Sacram suite by Campbell and John (1994). These rocks intrude dark colored Proterozoic gneiss and granite (locally with a steeply dipping mylonitic fabric), and contain blobs of mingled mafic rocks as do dikes and small stocks in the northern Chemehuevi Mountains that yielded a U-Pb zircon age of ~19 Ma (S. Mukasa and John, unpublished data). Textural features in the Sacram suite suggest that exposures on the southwest were shallowest so it is tilted down an unknown amount toward the southwest. Westernmost exposures of the suite are dominated by hypabyssal granite and quartz monzonite. Exposures 5 to 8 kilometers to the east may represent the structurally deepest exposures of the suite, and are characterized by mixed mafic and felsic (granite and diorite, gabbro and hornblendite), or purely mafic plutonic rocks. Structural, geophysical, petrologic, and thermochronologic studies of the Sacram suite are in progress by Campbell and John (1994)(University of Wyoming), D. Foster (La Trobe University) and S. Mukasa (University of Michigan).

Continue on U.S. 95 through Lobecks Pass, a narrow pass between the northern Chemehuevi and southern Sacramento mountains. Just prior to entering the pass (just southwest of the South Needles gas compressor station), we will drive over the culmination of the Colorado River gravity high; here the peak is roughly 20 mGal above the background high. This elliptical culmination is part of a gravity and magnetic high that tracks the zone of documented maximum mid-Tertiary extension for >150 km centered along the

Colorado River (Simpson and others, 1990). The culmination is approximately 10 km across; a preliminary model of the source body suggests it is within a few kilometers of the surface (R.W Simpson, written commun., 1990). Despite knowledge of the massive gravity and magnetic high, previous traverses throughout the northwestern Chemehuevi Mountains have found few rocks exposed at the surface of sufficient magnetic and density character to produce the geophysical anomaly. The mafic parts of the Sacram suite in the southernmost Sacramento Mountains seem ideal candidates to generate such an anomaly.

Beyond Lobecks Pass, the Chemehuevi detachment fault dips west off the back side of the core complex and the road crosses into Proterozoic gneiss in the upper plate. This gneiss is capped by steeply SW-dipping Miocene volcanic rocks that form the Sawtooth Range and steep Snaggletooth pinnacle, our destination. Continue south on U.S. 95 from Lobecks Pass southwest for 5.0 miles. Pull off U.S. 95 at 0.2 miles east of Snaggletooth pinnacle on the north (right) side of the highway.

Introduction to the Chemehuevi Mountains

B. John

Our next stop will serve as introduction to core-complex geology in the Chemehuevi Mountains and the central part of the Colorado River extensional corridor (Howard and John, 1987; John, 1987a). Regionally, major mid-Tertiary extension involving the upper and middle crust was accomplished along brittle northeast-dipping, detachment faults. These faults cut gently down-section in the direction of tectonic transport from a headwall breakaway between the Turtle and Old Woman Mountains in the west (Howard and John, 1987). Transport of the upper plate of each fault, where known, was to the northeast (Davis and others, 1980; John, 1987a; Howard and others, 1982a, Spencer, 1985).

The Chemehuevi Mountains lie centered in the Colorado River extensional corridor in the region of maximum crustal stretching (Fig. 6, 7). All rock types in the range, barring Pliocene and Pleistocene deposits and small young Miocene basaltic dikes and plugs, were deformed during extension. The deformation produced three brittle, low-angle normal faults separating at least four plates . The footwall of the Chemehuevi Mountains includes the structurally deepest exposed rocks in the range, below the lowest exposed and small-displacement (≤2 km) Mohave Wash fault. This fault is structurally overlain and cut by the major-displacement (>15 km) Chemehuevi detachment fault, which is in turn overlain by the structurally highest Devils Elbow fault (John, 1987a, 1987b). The two structurally deepest faults are exposed a distance of over 23 kilometers in a down-dip direction, across a total area of greater than 350 square kilometers.

Slip on each of the low-angle normal faults resulted in NE transport of successive hanging walls. At outcrop scale, each of the faults is planar, but when viewed at map scale, both the Mohave Wash and Chemehuevi detachment faults are corrugated parallel to the slip direction. Dips on each fault vary from very gently inclined along the troughs or crests of the corrugations, to as much as 40° on the steeper flanks or strike-slip portions of the faults (John, 1987a). Orthogonal to, and superimposed on, these corrugations are broad north-northwest striking antiformal and synformal undulations of the fault surfaces. Regionally, the Chemehuevi and Mohave Wash faults dip gently (10-15°) toward the southwest along the western flank of the range, and gently toward the northeast (2-15°) on the eastern flank near Topock Gorge.

The upper crust above the Chemehuevi fault system was pulled apart along high-angle normal faults that rotated to more gentle dips through time. Deformation within the footwall was small, accommodating minor extension (<2%) by normal and strike-slip faulting, local ductile shearing, and dike emplacement (John, 1987a).

The footwall to the Chemehuevi detachment fault crops out as a domal exposure of igneous and metamorphic rocks in the central part of the range. These crystalline rocks consist of mylonitized Proterozoic layered gneisses and migmatites, the voluminous Late Cretaceous Chemehuevi Mountains Plutonic Suite (73 ±8 Ma; John and Mukasa, 1990; John, 1988; John and Wooden, 1990), and a swarm of younger Cretaceous (?) and Tertiary dikes.

The Chemehuevi Mountains Plutonic Suite forms a compositionally zoned, calc-alkaline sill- or laccolith-like body. The magma chamber associated with the body evidently grew laterally in stages, with each pulse more fractionated than the previous one (John, 1988). The reconstructed shape of the body shows a flat floor for the lower part of the suite, which was intruded by at least three feeder dikes that provided magma to the chamber. This floor is characterized by lit-par-lit intrusions into subhorizontally foliated mylonitic gneisses (seen on Day 3). The roof is unexposed in the Chemehuevi Mountains but possibly is represented in allochthons in the Mohave Mountains, where the suite discordantly intrudes Proterozoic rocks.

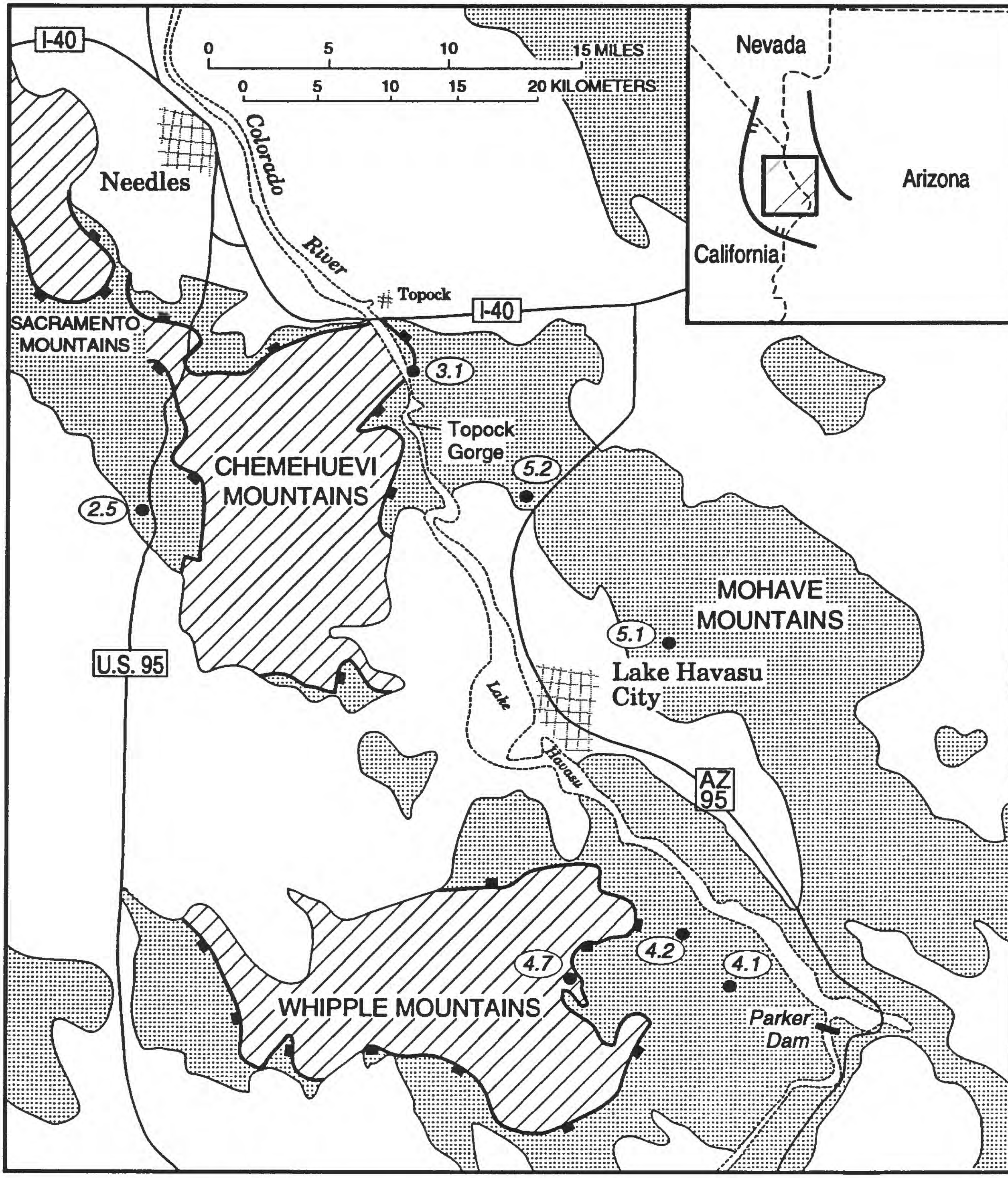


Figure 6. Central part of the Colorado River extensional corridor, showing field-trip stops. Lower plates of metamorphic core complexes have a diagonal pattern. Post-extensional deposits are not patterned. Insert shows position in the Colorado River extensional corridor.

The extensional fault system is domed over the light colored footwall, and crops out again in the eastern Chemehuevi Mountains (see Stop 3.1). The dark weathering dikes that cut the footwall are part of a voluminous dike swarm that intrudes rocks above, below and along the Chemehuevi detachment fault in the western, southern and central parts of the range. The dikes form two subvertical, roughly orthogonal sets oriented ENE, and N to WNW. The orthogonal pattern produced by NW- and NE-striking dikes prevails in the west-central part of the range, where they are hosted by the isotropic Chemehuevi Mountains Plutonic Suite. In the central and northern parts of the range where they intrude older plutonic rocks and Proterozoic country rocks, the dominant strike of dikes changes to E-W.

The dikes range in composition from mafic to felsic, and are of several generations. Composition and relative age appear to correlate little with orientation. The swarm can been subdivided, based on phenocryst mineralogy, texture, and whole-rock geochemistry, into progressively younger types: highly altered, xenolith-bearing hornblende gabbro dikes, diabase and lamprophyres, basaltic andesite and andesite, hypabyssal dacites and rhyolites, a second generation of lamprophyres, and olivine basalt.

Dike intrusion in the Chemehuevi Mountains temporally overlaps extrusive volcanism and spans the timing of extensional deformation. Some dikes predate upper-plate tilting of the 18.5-Ma Peach Springs Tuff as indicated by a K-Ar hornblende age of 20.7±1.3 Ma (John, 1986) for a northwest-striking dacite dike from the northwestern part of the range. Post-detachment basalt, in thin north-striking dikes, plugs, and local flows dated at 11.1±0.3 Ma (K-Ar whole-rock), intrudes and locally fuses cataclasites associated with detachment faulting (John, 1986, 1987a). This relationship implies that the basalt was intruded through, and locally flowed onto, an erosional surface that truncated the central part of the Chemehuevi detachment fault system after movement had ceased.

Structural constraints on the initiation angle of the detachment faults in the Chemehuevi Mountains are based on a wide variety of observations, including fault rock type and associated mineral deformation mechanisms, orientation and cross-cutting relations of syntectonic dikes and faults, and the metamorphic grade of footwall rocks to the regionally developed normal fault system. In each case, the initial dip of the fault is limited to <30°. Application of $^{40}Ar/^{39}Ar$ and fission-track thermochronology to rocks in the footwall of the fault system provides further constraints on the timing and initiation angle of regional detachment faulting. At the onset of extension between 22 and 24 Ma, granitic rocks now exposed in the southwestern and northeastern portions of the footwall were at ~100°C and >400°C, respectively, separated by a distance of some 23 kilometers down the known slip direction. This gradual increase in temperature with depth is attributed to the gentle tilting and warping of originally horizontal isothermal surfaces, and constrains the exposed part of the Chemehuevi detachment fault to have had a regional dip initially of 15° to 30° (John and Foster, 1993).

Stop 2.5

Snaggletooth -- B. John

This stop serves to view and discuss the Chemehuevi Mountains. In addition we will discuss structural and thermochronologic evidence for tilting of the footwall and for the original dip of the detachment faults that are domed over it (see John and Foster, 1993).

Snaggletooth is a group of rock pinnacles of SW-dipping Tertiary andesite and dacite in the hanging wall to the regionally developed detachment fault system. The 18.5-Ma Peach Springs Tuff is at the top of the section here. The Tertiary rocks noncomformably overlie Proterozoic rocks: gneisses, granite and diabase dikes. Toward the east in the Chemehuevi Mountains, the dark Proterozoic rocks are in fault contact above light colored Late Cretaceous granitic rocks of the Chemehuevi Mountains Plutonic Suite (John and Mukasa, 1990; John and Wooden, 1990), along the west dipping Chemehuevi detachment fault that encircles the range (John, 1987a).

Highlights between Stops 2.5 and 2.6

Return to Needles by heading 17.7 mi N along U.S. 95. Turn N on I-40 (westbound toward Barstow) and go 1.2 miles to the "J" Street exit. Turn right on J street for one block, then left at the traffic light on the main street for ~0.5 mi, then right at the "T" for about a mile, crossing the bridge over the Colorado River into Arizona. Keep left (N) on Arizona 95 toward Bullhead City for 9 mi. Turn right (east) on Boundary Cone Road toward Oatman and the southern Black Mountains, which expose a pile of early Miocene volcanic rocks.

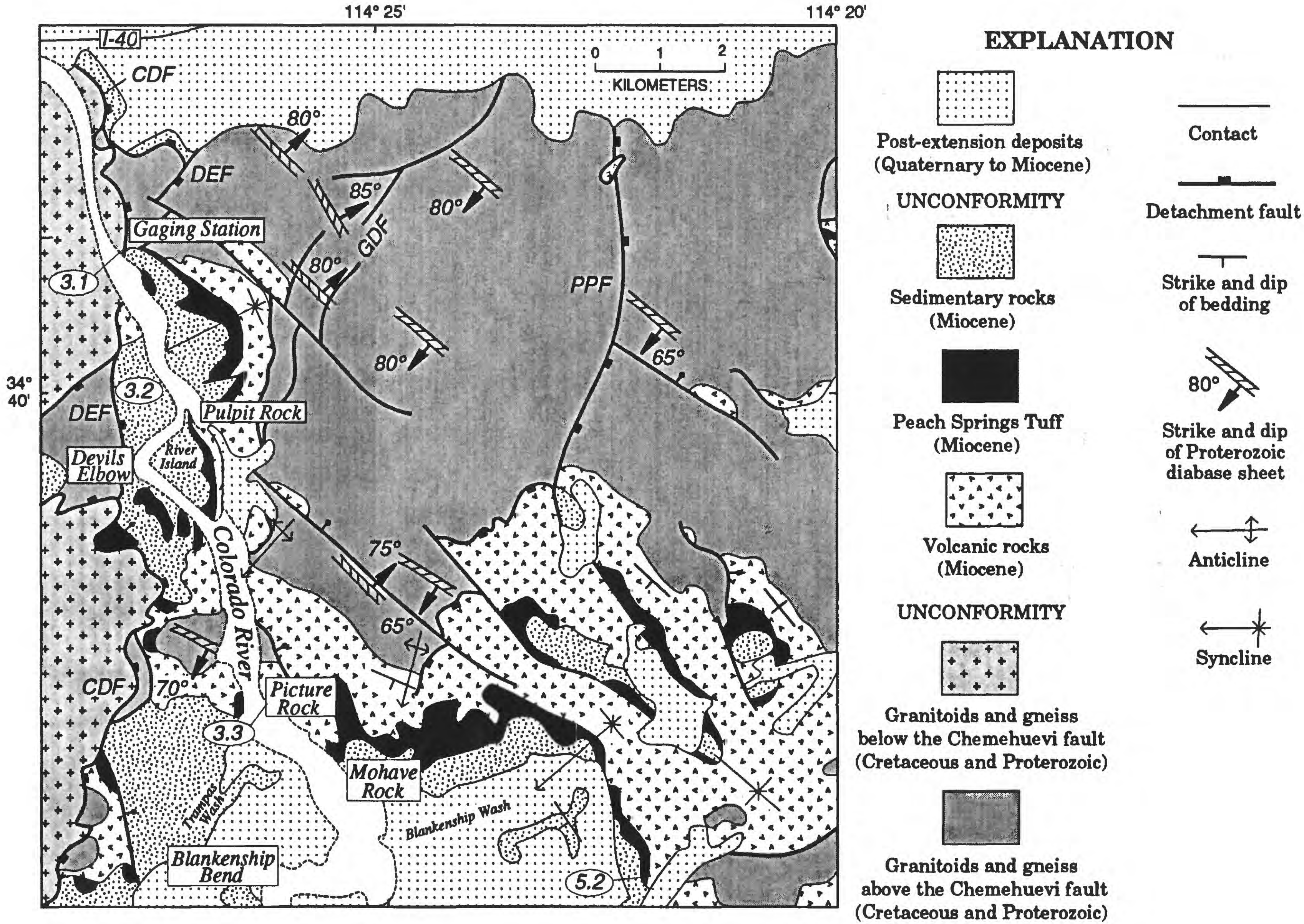


Figure 7. Geologic sketch map of the Topock Gorge area, showing numbered field-trip stops. CDF=Chemehuevi detachment fault. DEF=Devils Elbow fault. PPF=Powell Peak fault. GDF=Gold Dome fault zone.

A sharp peak ahead is Boundary Cone, a rhyolite plug (Fig. 8a). Phenocrysts (now anorthoclase) in a dike emanating from Boundary Cone were dated by K-Ar as 16.8±0.4 Ma by J.K. Nakata (written commun., 1987). This age may date a time of potassium metasomatism, in which Na_2O was mostly replaced by K_2O. A high ratio of K_2O/Na_2O (15:1) in the dated rock is characteristic of potassium-metasomatised rocks common in the southern Basin and Range province (Chapin and Lindley, 1986; Brooks, 1986; Roddy and others, 1988).

10.7 miles beyond Highway 95, park at a pull-out on the right for a view and discussion of the geology of the southern Black Mountains ahead.

Stop 2.6

Black Mountains -- K. Howard

Miocene volcanic and hypabyssal intrusive rocks underlie the southern Black Mountains (Fig. 4). The volcanic rocks are mainly latites, andesites, and trachytes 18-19 Ma (DeWitt and others, 1986b), unconformably overlain by mesa-capping basalt dated as 15.8 Ma (Gray and others, 1990). Thorson's (1971) detailed mapping established that the Alcyone Formation, low within the sequence, includes a caldera-filling tuff interlayered with breccias that represent landslides off the caldera walls. A granitic pluton related to resurgence, the Times Porphyry, intrudes the Alcyone, and in turn is intruded on the north by the younger Moss Porphyry (Thorson, 1971). Dating by U-Pb (zircon) and $^{40}Ar/^{39}Ar$ (hornblende) places both porphyry intrusions as 18-19 Ma (DeWitt and others, 1986b).

Rocks in the Black Mountains lie near the eastern margin of the Colorado River extensional corridor and are much less tilted and extended than in most of the corridor. If slip on the Newberry detachment fault was top to the east as assumed by Spencer (1985), the rocks at Oatman would reconstruct back close to the lower plate in the Newberry and Dead Mountains 20-30 km to the west. If so, Miocene plutons in the southern Newberry and northern Dead Mountains (Fig. 4) may be beheaded lower-plate relatives of the hypabyssal Moss Porphyry and Times Porphyry and the volcanic rocks of the Oatman District.

Highlights between Stops 2.6 and 2.7

The Gold Road Latite, one of the higher units within the volcanic sequence in this part of the Black Mountains, forms prominently layered rocks ahead, overlying the Oatman Latite. Continue 3.4 miles to Oatman.

Stop 2.7

Oatman -- K. Howard

The story of the quaint 19th century mining camp of Oatman in the Black Mountain began when gold was discovered here in 1863 by soldiers stationed at Camp Mohave on the Colorado River. Major discoveries followed at the turn of the century (DeWitt and others, 1986b). Two million troy ounces of gold were produced and a million ounces of silver. Later in the early twentieth century, immigrations to California from the dust-bowl states, immortalized in John Steinbeck's *The Grapes of Wrath,* passed through on historic U.S. Route 66. Clark Gable and Carol Lombard honeymooned in Oatman.

The ore deposits in the Oatman District are epithermal gold-bearing quartz-calcite veins that occupy fault fissures in the early Miocene volcanic rocks and hypabyssal stocks (DeWitt and others, 1986b).

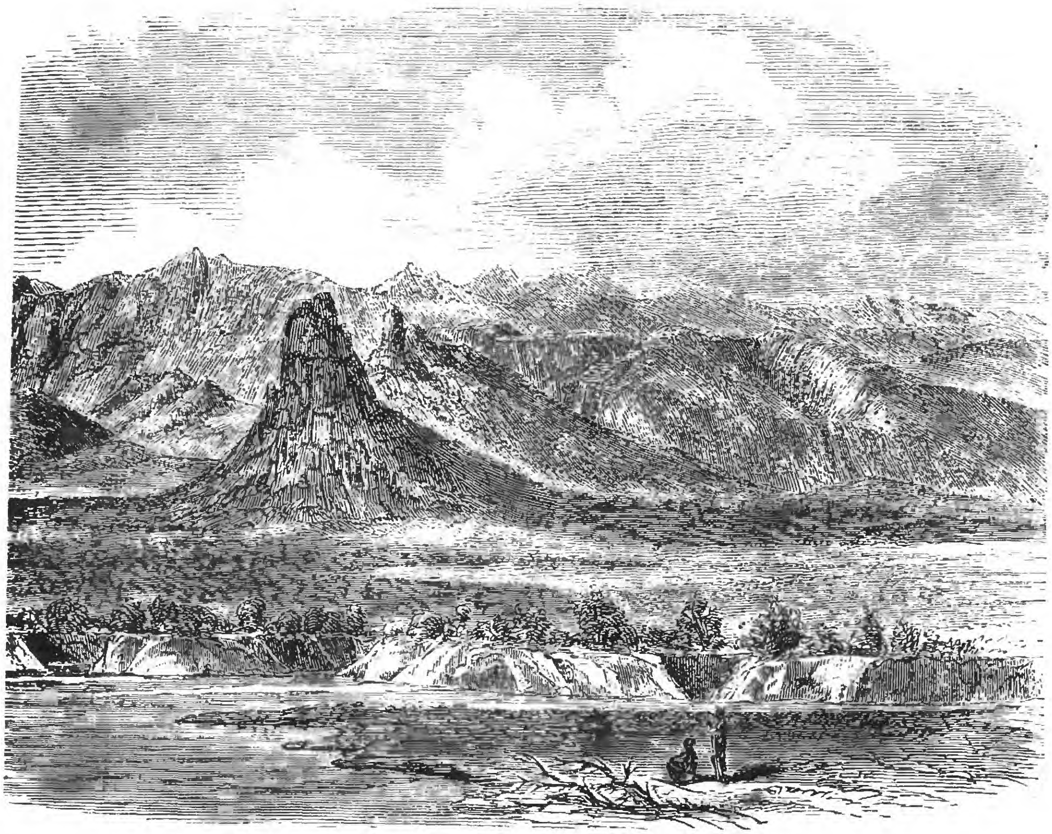
Retrace the route 14.1 miles back to Arizona 95. Turn right (N) 16 miles on Highway 95. At a traffic light, turn left (west) to the bridge across the Colorado River and return to Laughlin.

DAY 3

Topock Gorge -- B. John

Day 3 serves two purposes. The first is to observe two of the major low-angle normal faults associated with mid-Tertiary extension in the Chemehuevi Mountains. Second, we will look in some detail at upended hanging-wall blocks above the detachment fault system, and discuss the syntectonic volcanic succession, and the sedimentary record of unroofing the core complex.

Travel will be by canoe on the calm waters of the Colorado River in rugged Topock Gorge with no access to luggage. It will be necessary to have water, sunscreen and a hat with you in the canoe. We will board the canoes at Golden Shores Marina in Topock, Arizona, and proceed downriver.



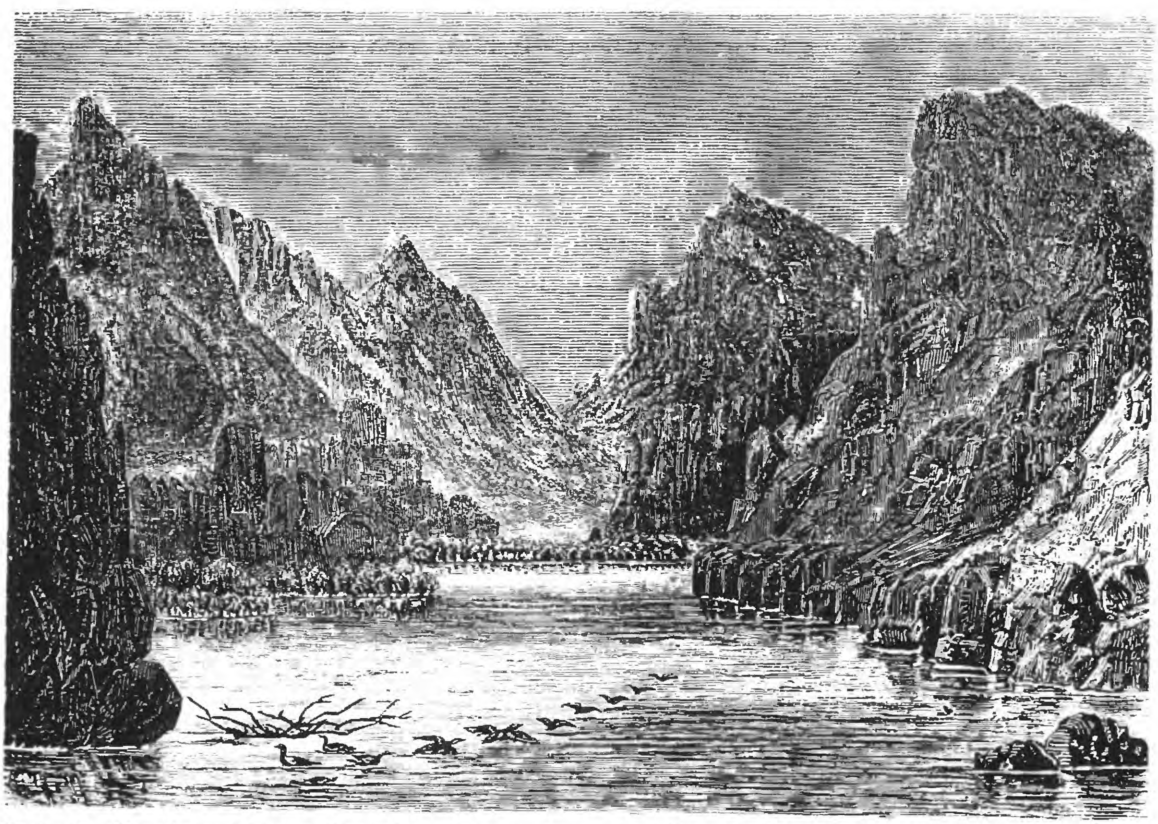


Figure 8. Sketches of (a) the Black Mountains, and (b) upper Topock Gorge, from the Ives (1861) report.
a. (Top) "Boundary Cone." This is a Miocene rhyolite neck seen on the way to Oatman.

b. (Bottom) "Head of Mojave Cañon."

Highlights on the Way to Stop 3.1

Upon leaving Laughlin, proceed west about a mile to River Road, and turn left (S) about 22 mi to Needles and I-40 eastbound (toward Kingman) . Proceed S on I-40 eastbound 7 mi through Needles. Much of the following 5 miles of the highway beyond the agricultural inspection station is built on swelling clays of the Miocene and Pliocene Bouse Formation, deposited in an estuary at the head of the newly forming Gulf of California. Irregularities in the road surface are the result of swelling of the clay after heavy rains in 1992.

At 3 o'clock lies the northern range front of the Chemehuevi Mountains, and the Whale Mountain area. Whale Mountain is underlain by steeply foliated mylonitic rocks, in the footwall to the Chemehuevi detachment fault, that bear a subhorizontal NE-striking stretching lineation. This fabric was interpreted as a product of reorientation of Proterozoic and Mesozoic fabrics during Late Cretaceous sinistral shear along a steep shear zone that links gently-dipping ductile fabrics in the Chemehuevi Mountains with those in the Sacramento Mountains north of Monumental Pass (John and Mukasa, 1990). The Mesozoic mylonitic fabric has been overprinted by both brittle and locally, ductile effects of the Tertiary extensional deformation. Whale Mountain is capped by klippen above the structurally deepest Mohave Wash fault.

In 6.5 miles, cross the Colorado River on I-40. As you cross the river you will see The Needles on the right in the Mohave Mountains. The Needles, named by Captain Whipple, are striking pinnacles of steeply west-tilted, synextensional volcanic, hypabyssal, and sedimentary rocks that are structurally above the Chemehuevi detachment fault you saw yesterday at Stop 2.5. The Needles and spectacular Topock Gorge (Figs. 7, 8b, 9) are today's destination.

Continue 0.8 miles to Exit 1, exit right and head northbound across the interstate highway, following signs for Topock (Mohave for bridge) and old Route 66. Continue west, parallel to I-40, 0.5 miles, under the railroad bridge. Take the first left at Golden Shores Marina, park and prepare for the all-day canoe trip. The field guide is written with stops progressing downstream from the upper end of the gorge, starting from the I-40 highway bridge (Fig. 7).

Lt. Ives' (1861) colorful description of the approach to Topock Gorge ("Mojave Cañon") from the south was was made before Lake Havasu partly flooded the valley behind Parker Dam:

> To-day has been perfectly serene, and the atmosphere indescribably soft and limpid. For several milers the river assumed a new aspect, being straight and broad, having high banks, and presenting a placid unbroken sheet of water - not a bar being visible above the surface. To one viewing the noble looking stream from the bank, it would have appeared navigable for vessels of the heaviest draught, but the depth of water was scarcely sufficient to enable the Explorer to pass without touching.
>
> Entering the foot hills of the Mojave Range, the channel was again tortuous, and after traversing a narrow pass the Needles came in view directly in front. As we approached the mouth of the cañon through the Mojave mountains, a roaring noise ahead gave notice that we were coming to a rapid, and soon we reached the foot of a pebbly island, along either side of which the water was rushing, enveloped in a sheet of foam.
>
> After ascending a few yards a harsh grating noise warned us that we were upon a rocky shoal, and Captain Robinson at once backed the Explorer out and went up in a skiff to reconnoitre....There was danger that the after part of the boat in passing might catch upon a rock, and the bow be swung around by the rapid current against another with such violence as to knock a hole in the bottom. An anchor was carried to a point some distance up stream, and a line taken from it to the bow. This line was kept taut, while, with a high pressure of steam, the Explorer was forced up the rapids, once or twice trembling from stem to stern as she grazed upon a rock, but reaching the still water above without sustaining damage.
>
> A low purple gateway and a splendid corridor, with massive red walls, formed the entrance to the cañon. At the head of this avenue frowning mountains, piled one above the other, seemed to block the way. An abrupt run at the base of the apparent barrier revealed a cavern-like approach to the profound chasm beyond. A scene of such imposing grandeur as that which now presented itself I have never before witnessed. On either side majestic cliffs, hundreds of feet in height, rose perpendicularly from the water. As the river wound

through the narrow enclosure every turn developed some sublime effect or startling novelty in the view. Brilliant tints of purple, green, brown, red, and white illuminated the stupendous surfaces and relieved their sombre monotony. Far above, clear and distinct upon the narrow strip of sky, turrets, spires, jagged statue-like peaks and grotesque pinnacles overlooked the deep abyss.

The upper part of Topock Gorge (Fig. 8b) is along the Chemehuevi detachment fault. Rocks along the right (west) bank lie in the footwall to the fault, and are mostly altered and brecciated gneisses and granitic rocks of Proterozoic and Mesozoic age. The rocks on the left (east) bank lie in the hanging wall and are composed mainly of alluvial-fan and landslide-megabreccia deposits. Stop 3.1 is 1.8 miles downstream from the I-40 highway bridge, along the left (east) shore of the Colorado River by a stream-gaging station.

Stop 3.1
Chemehuevi detachment fault

Disembark at the gaging station, or, depending on river level, at a small beach to the south of the wash (at the gaging station), and climb up over the small hill of Colorado River gravels, to descend into the main wash. Walk 0.2 miles up the main wash from the river. Climb up the small dirt road on the left (N) side of the wash to exposures of the Chemehuevi detachment fault. The fault itself dips 12° toward the east, and superposes biotite granodiorite (Cretaceous Chemehuevi Mountains Plutonic Suite) and numerous (presumed tilted) subhorizontal mafic dikes of Tertiary age in the hanging wall, against hornblende-biotite granodiorite (Chemehuevi Mountains Plutonic Suite) and Proterozoic gneisses cut by subvertical mafic dikes. The fault is characterized as a zone up to tens of meters thick, of altered cataclasite and ultracataclasite derived from both the footwall and hanging wall.

Return to the boat and continue 0.8 miles downstream, crossing the river to Stop 3.2. It is best to land at the tip of the small peninsula just south of the sandbar, and make your way through the salt cedar along small burro trails to the main wash.

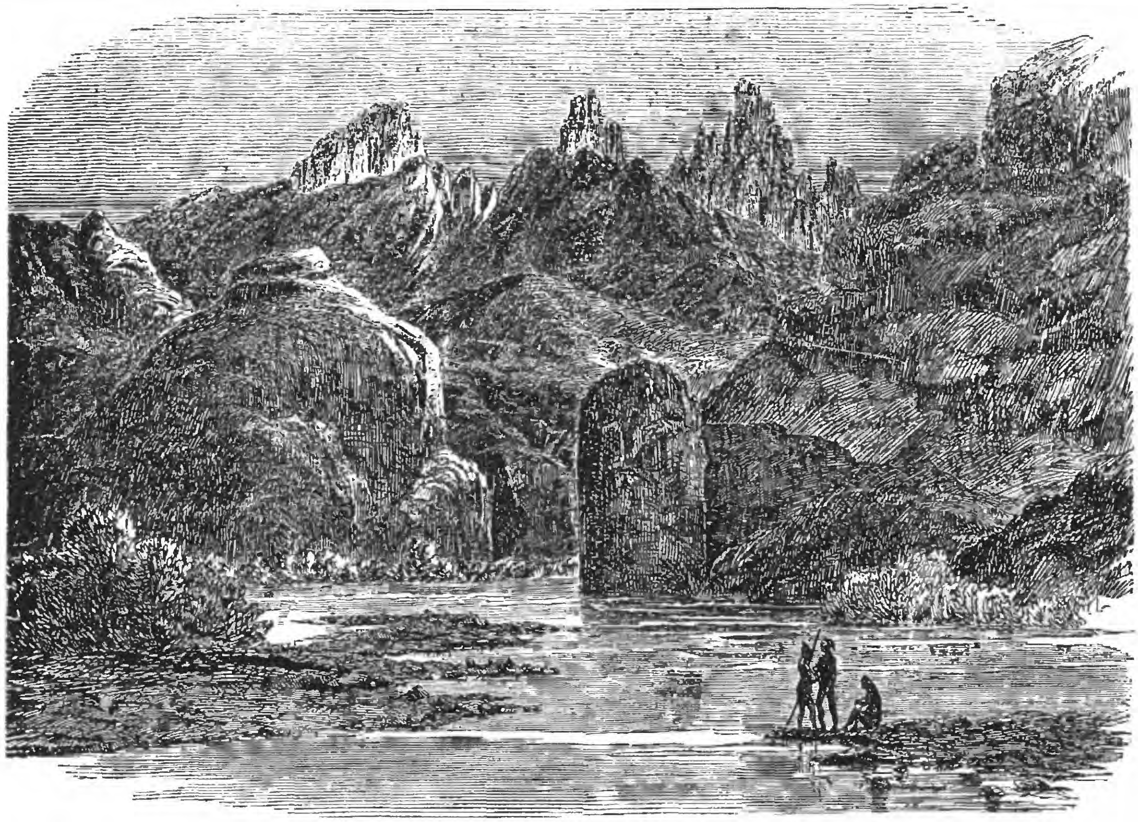
Stop 3.2
Devils Elbow fault

The Devils Elbow fault, our destination, lies structurally above the Chemehuevi and Mohave Wash faults. From the west river bank, walk southwest up the main wash 0.5 miles (0.8 km). Exposures on the right (N) include the Peach Springs Tuff above older mafic volcanic rocks. At the stream fork, keep left and follow the burro trail 0.3 miles (0.5 km) over the low saddle through Colorado River gravels that unconformably overlie altered and fractured granitic rocks in the hanging wall of the Chemehuevi detachment fault. Red rocks to the east are altered Tertiary volcanic rock, with interstratified sedimentary breccia and megabreccia deposits above the Devils Elbow fault. The fault at this exposure dips ~35° east, with well preserved slickenside striae plunging 33° down dip. Although poorly constrained, slip on the Devils Elbow fault is several kilometers to the east-northeast (John, 1987a). Rocks in the footwall to this fault form the hanging wall to the Chemehuevi detachment fault. These granitic rocks are part of the Chemehuevi Mountains Plutonic Suite, with associated mafic and silicic dikes (now gently dipping) derived from a footwall location ~18 km to the WSW.

Return to the boat. Before leaving the beach look east across the river to the high peaks on the skyline, which are a dip slope in the Peach Springs Tuff, the substrate to the syntectonic sedimentary succession. Exposures of the Peach Springs form a spectacular, well defined hanging wall syncline, plunging moderately toward the southwest. The core of the syncline is filled with coarse sedimentary breccia and thick, irregular megabreccia deposits, made up almost exclusively of crystalline clasts.

Highlights between Stops 3.2 and 3.3

Continue down river 1.2 miles to the small bay at the river bend near Pulpit Rock. The bay marks the head of an old abandoned channel of the Colorado River that veers southeast into Arizona and bounds the east side of River Island (Lee, 1908). The gorge downstream between Pulpit Rock and River Island (Fig. 9a) cuts through gently to steeply southwest-dipping volcanic and sedimentary rocks, cut by numerous high-angle and low-angle faults. These Miocene rocks lie nonconformably above light-colored Proterozoic gneisses and granites, all in the hanging walls of the Chemehuevi and Devils Elbow faults, which you have just seen. Beyond Devils Elbow, rocks on the right (W) are dominantly the Peach Springs



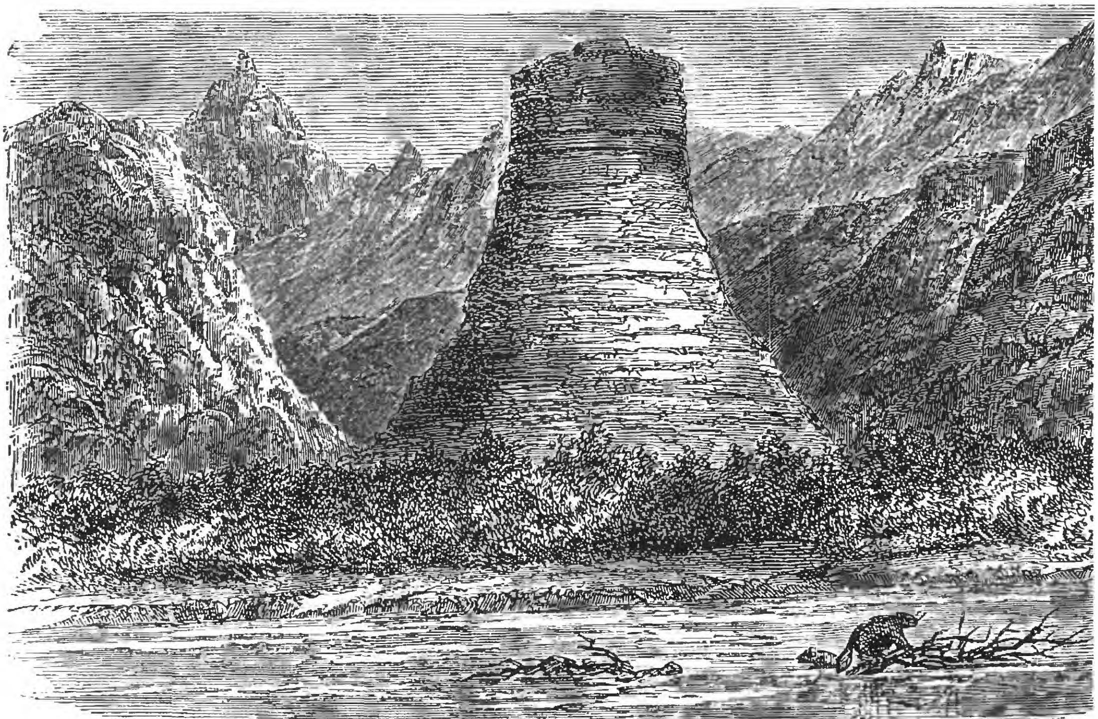


Figure 9. Sketches of Topock Gorge from the Ives (1861) report.

a. (Top) "Mouth of Mohave Cañon." In this view looking upstream, Pulpit Rock can be recognized in the foreground, dip slopes of the Peach Springs Tuff in a syncline keel are on the upper left skyline, and vertical rocks of the lower volcanic sequence form the high Needles pinnacles in the background. Gently dipping middle Miocene conglomerate can be seen on the east bank on the right.

b. (Bottom) "Remains of Grand Mesa in Chemehuevis Valley." This view resembles Castle Rock, except that the background mountains are exaggerated. Castle Rock is formed of horizontally bedded conglomerate laid down after or near the end of the episode of tilting and extension.

Tuff and the lower volcanic sequence, cut by numerous down-to-the-east normal faults. Rocks in the vertical wall along the east bank (left) include post-Peach Springs Tuff sedimentary breccia and megabreccia deposits. At the southern end of River Island, very coarse river gravels are exposed beneath the hogback of Peach Springs Tuff. The large sand dunes banked up against the lower volcanic sequence are reworked (windblown) deposits derived from the Chemehuevi Formation of Longwell (1936).

As we boat down river 4.0 miles to Stop 3.3, we will pass through a series of folds associated with the large, upended tilted blocks of crystalline rock, above the Chemehuevi fault system (similar to that viewed from a distance at Stop 3.2). The stratified rocks drape around the corners of adjoining basement blocks forming the moderately to steeply plunging folds, interpreted as drape (forced) folds, with amplitudes as great as 2 km (John and Howard, 1994). The folds in the stratified rocks contrast with the simply-tilted, homoclinal blocks of crystalline basement rock around which the strata drape. The drape folds here and elsewhere in the southern Basin and Range province demonstrate that major plunging folds can form during extensional tectonism (Howard and John, 1991).

Stop 3.3
Picture Rock

Stop 3.3 is at Picture Rock on the east side of the Colorado River near the lower end of Topock Gorge. Picture Rock is a site of petroglyphs. The Mohave people call it *Hum-Me-Chomp,* "The rock where the river once churned to make this place inaccessible to the living". The illustrations were etched by percussion onto an outcrop of the lower volcanic sequence, and include early representations of bighorn sheep, water, snakes, and the sun, as well as later renditions of horses with riders. Picture Rock is said to have been a trading site between the nomadic Chemehuevi Indians who roamed the area west of the Colorado River, and the Mohave Indians that lived along the east side of the river.

Picture Rock is a part of a dipping sequence of east-striking mafic and intermediate volcanic flows, which locally overlies a very thin basal arkose that rests directly on the Proterozoic basement. The lavas are unconformably overlain locally by the Peach Springs Tuff and succeeding syntectonic sedimentary rocks that are exposed in Trampas Wash directly to the west, and in Blankenship Wash to the south and east.

Highlights between Stop 3.3 and Castle Rock Bay

Continue 1.3 miles downstream to a small island of megabreccia near the east shore of the river. This is Mohave Rock, reputedly used by steamship captains to winch their boats up the rapids described by Ives that were flooded when Parker Dam was built and Lake Havasu formed. Rowboats were sent upriver carrying the ships anchor, which was attached to Mohave Rock and hauled in, moving the ship slowly up the rapids toward smoother water.

Continue down river around Blankenship Bend in the upper reaches of Lake Havasu. At the upstream curve of Blankenship Bend on the Arizona side of the Colorado River are exposures of pinkish sands that are correlated with the Chemehuevi Formation of Longwell (1936) of Pleistocene age. These sands interfinger abruptly with greenish rubble and poorly sorted conglomerate consisting wholly of bouldery clasts derived from chlorite-rich, altered granitic rocks from the footwall exposed in the Chemehuevi Mountains core complex across the Colorado River. The interfingering suggests that this rubble represents a landslide dam across the ancestral Colorado River, behind which the pinkish sands were deposited The presence of the Chemehuevi Formation may therefore record quiet water behind temporary dams in the canyons traversed by the lower Colorado River.

Along the west side of the Colorado River at Blankenship Bend is a gently dipping vesicular basalt flow, one of the last gasps of volcanism associated with extension in the area. The flow is interstratified within the upper part of the post- Peach Springs sedimentary section that makes up the Trampas Wash and Blankenship basins. This flow has been correlated on both sides of the river (based on REE chemistry), tying the two basins together at a late stage in their depositional history. The white cliff exposures along the east shore of the Colorado River are gently dipping Miocene conglomerate derived from the felsic members of the Chemehuevi Mountains Plutonic Suite to the west. The conglomerate is unconformably overlain by the Chemehuevi Formation.

Continue downstream 4.0 miles to Castle Rock Bay, where you must negotiate shallow sloughs on the east (left) side of the river to prominent Castle Rock (Fig. 9b). **The entrance to Castle Rock Bay is marked by a string of widely-spaced buoys across the river, and a large sign on the left (east) side of the river at the junction with a slough. Paddle up the slough heading for Castle Rock, the prominent rock at the east end of the bay, and beach the canoes.**

The canoes must be carried from the beach to the parking lot (0.1 miles east), where they will be loaded on trucks. We will drive 2.2 mi east to Arizona 95 and 8 mi south to Lake Havasu City and lodging near London Bridge.

DAY 4

Whipple Mountains -- G. Davis and L. Anderson

The purpose of today's trip is to visit and study one of the best exposed Tertiary "metamorphic core complexes" in the U. S. Cordillera -- the Whipple Mountains of southeastern California (Fig. 2). The Whipple Mountains are a 50 km-long, east-trending range at the easternmost tip of California just west of the Colorado River (Figs. 6 and 11). The range varies in north-south width from about 8 km in its west-central area to 25 km near its eastern end. Total relief in the spectacularly exposed range is ~1 km; the elevation of its highest peak is 4131 feet and the elevation of the Colorado River at Parker, Arizona is about 400 feet. Our visit to the easternmost part of the range will explore the Whipple detachment fault and its deeply exhumed lower plate, a window into the middle crust.

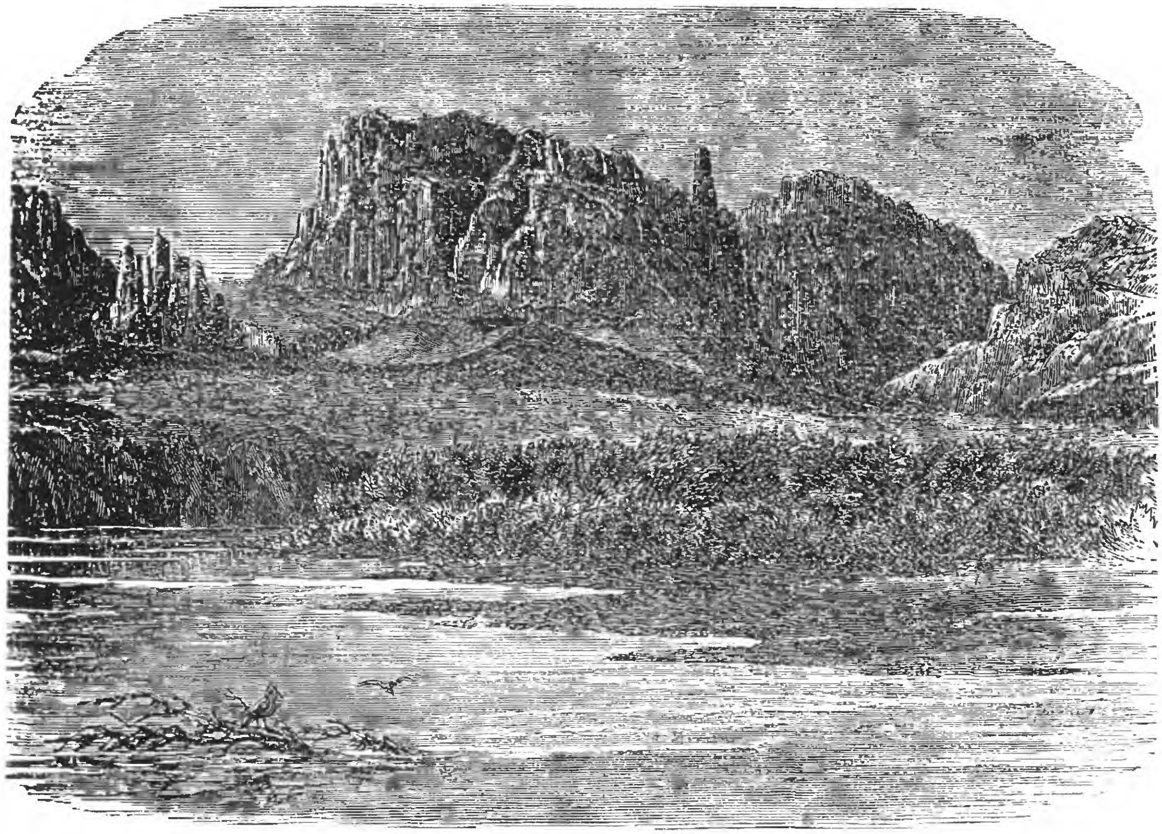
Most of today will be along a 14-km (9-mile) round-trip walking traverse up the rugged, spectacular canyon known as Whipple Wash. **Bring adequate water (4-5 quarts or liters)** because daytime temperatures are likely to be high and there is no drinkable water in the canyon. Highlights on the way to Stop 4.1. are described after the following introduction to the geology of the Whipple Mountains.

Introduction to the Whipple Mountains

The Whipple Mountains exhibit typical characteristics of Cordilleran metamorphic core complexes as originally defined by G. H. Davis and Coney (1979) and Coney (1980). The shallow-dipping and range-wide Whipple detachment fault separates two distinct litho-tectonic domains (Fig. 11): (1) an upper-plate assemblage of supracrustal rocks, exclusively of Tertiary age in the Whipple Mountains proper, and their crystalline basement; and (2) a lower-plate assemblage of mylonitic crystalline rocks of upper greenschist to middle amphibolite facies grade and, above a sharply gradational boundary (the Whipple "mylonitic front"), their nonmylonitized equivalents. As is characteristic of most "core complexes", the low-angle fault between the two domains in the mountain range now has a domed or antiformal geometry (Figs. 11, 12). As a consequence, lower-plate rock assemblages are primarily exposed in the core of the range, whereas upper-plate, typically lower-grade rocks are preserved on its lower flanks, in klippen at higher elevations, and in adjacent sediment-buried basins and synforms. Unmetamorphosed Miocene strata of the upper plate of the Whipple Mountains are now in contact along the Whipple detachment fault with crystalline rocks that were exhibiting ductile deformation deep within the crust during the Miocene. The amount of crustal omission necessitated by bringing the two plates into juxtaposition is impressive. It is likely that the mylonitic gneisses of the lower plate were being recrystallized during the early Miocene at depths greater than 12 km and at temperatures in excess of 440° C (Anderson and others, 1988). The history and rapidity of the uplift by normal faulting of these deep-seated rocks is one of the major aspects of this field trip.

The general geology (Fig. 11) and tectonic setting of the Whipple Mountains have been presented in a number of papers to which the interested reader is referred (Davis and others, 1980, 1982, 1986; Davis and Lister, 1988; Lister and Davis, 1989; Davis, 1988). Especially noteworthy papers by others on the geology and regional tectonics of the surrounding Colorado River extensional corridor include Howard and others (1982a,b), Howard and John (1987), Spencer and Reynolds (1990), Richard and others (1990), Foster and others (1990), and John and Mukasa (1990). The interested reader is referred to three special sections of the *Journal of Geophysical Research* (red) entitled "California-Arizona Crustal Transect, Part 1", v. 95, no. B1 (1990), "CACTIS, Part 2", v. 95, no. B12 (1990), and "CACTIS, Part 3", v. 96, no. B7 (1991) for a wealth of information on the extensional tectonics and geologic history of the region including and surrounding the Whipple Mountains and the Colorado River corridor. Some of the following summary of the geology of the Whipple Mountains is extracted, in some cases copied, from the Davis and Lister (1988) reference.

The geology of the Whipple Mountains is generally similar to that of several other ranges in the Colorado River extensional corridor, but the relatively high relief of the range, and the presence of deep canyons that cut across the Whipple detachment fault, provide some of the most spectacular exposures of detachment fault-related tectonics in western North America. Figure 12, a diagrammatic block diagram viewed to the north of the central and eastern Whipple Mountains, illustrates the major structural and lithologic features of the range. Structural features include: (1) the Whipple Mountains antiform, a major foliation-defined arch in lower-plate mylonitic rocks; (2) the southwest-dipping mylonitic front (MF), which separates lower-plate Proterozoic through Early Miocene crystalline rocks (lpxln) from their deeper, mylonitic counterparts (mgn); (3) the Whipple detachment fault (WDF) and the chloritic breccias (cbr)



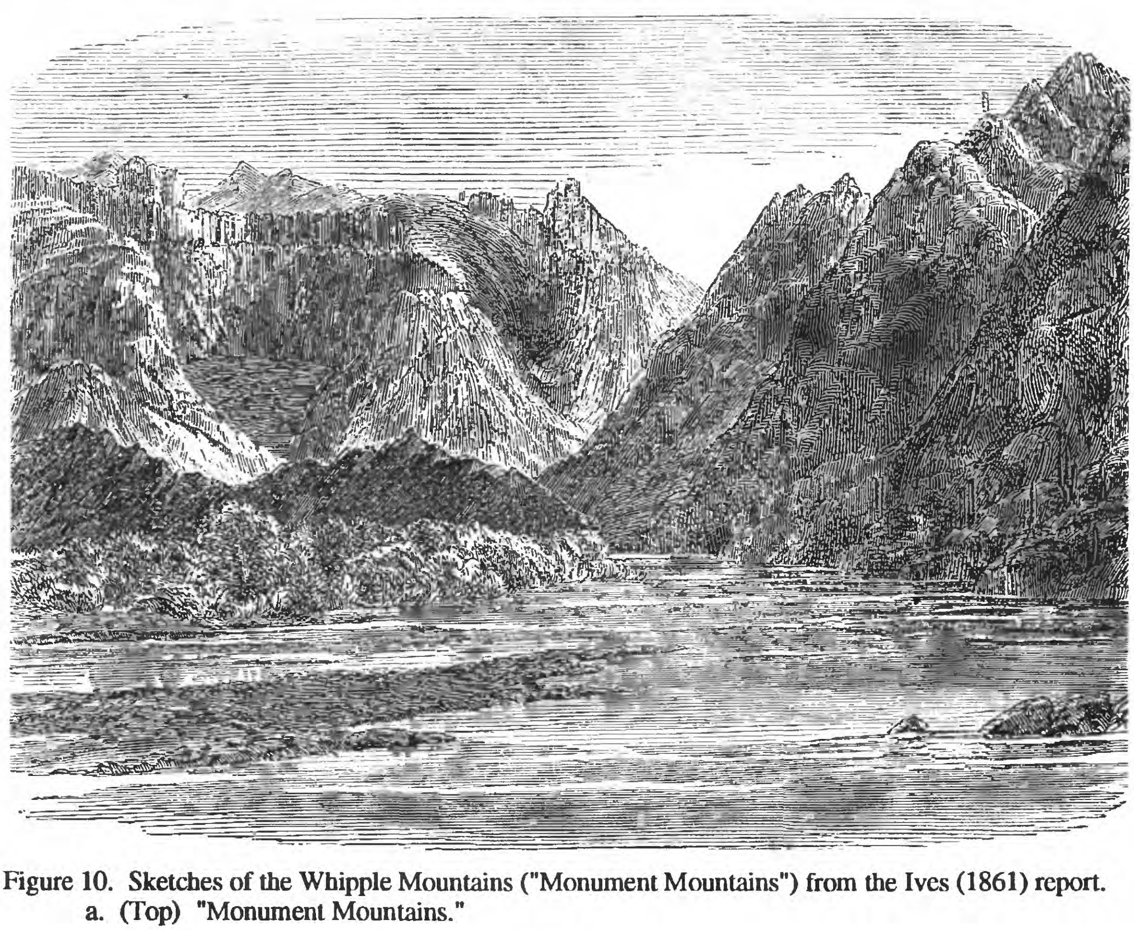


Figure 10. Sketches of the Whipple Mountains ("Monument Mountains") from the Ives (1861) report.
a. (Top) "Monument Mountains."

b. (Bottom) "Monument Range from the north."

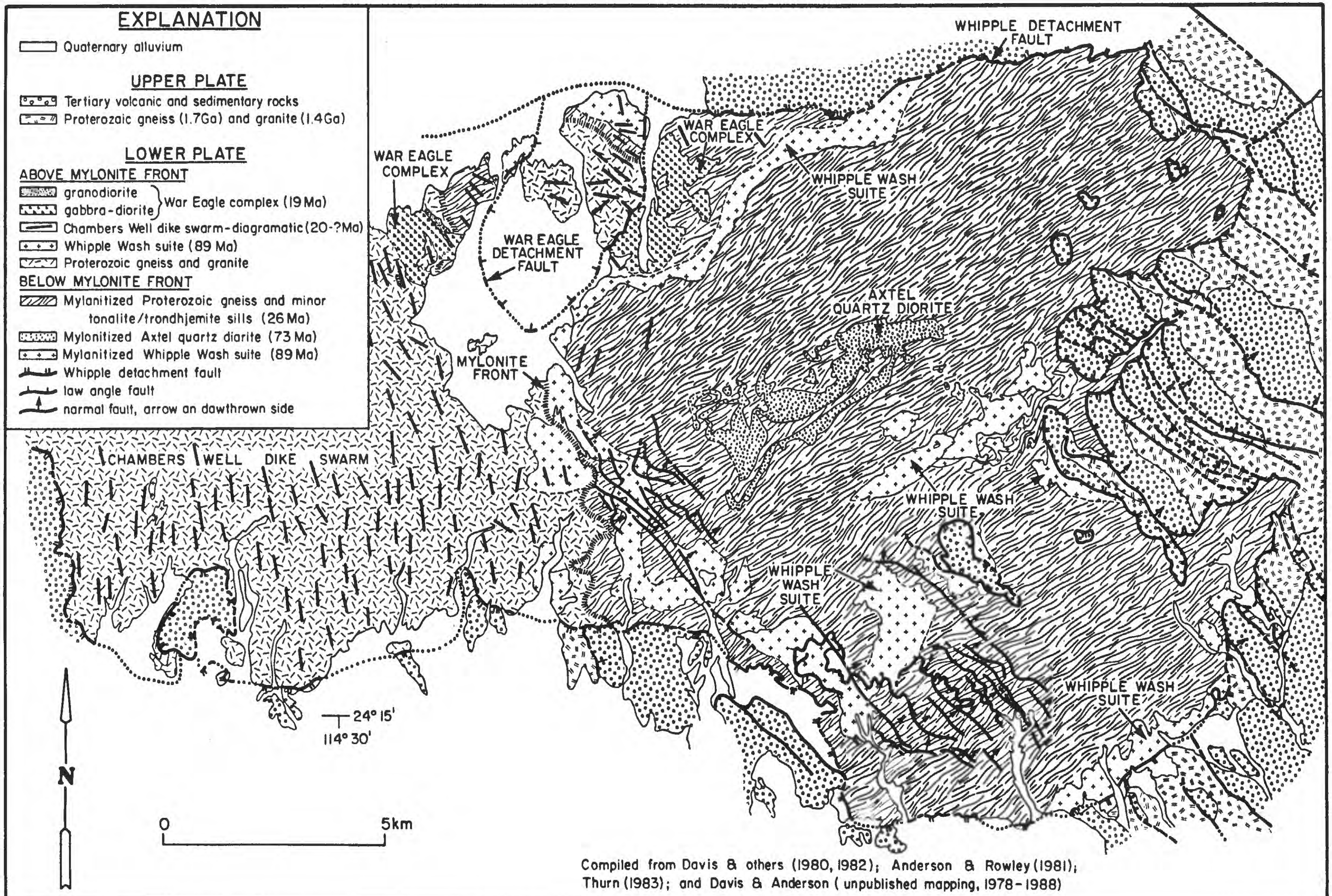


Figure 11. Simplified geologic map of the Whipple Mountains. From Anderson and Cullers (1990).

developed beneath it; and (4) an assemblage of closely spaced (~ 1-2 km), mostly northeast-dipping upper-plate normal faults; they repeatedly offset and rotate to the southwest upper-plate Tertiary strata (sv) and their largely Precambrian crystalline basement (upxln). The figure illustrates both the typical strong discordance seen in the range between the Whipple detachment fault and lower-plate and upper-plate structures, and the presence of a stretching lineation in the mylonitic rocks that parallels the hinge of the foliation-defined antiform and lies perpendicular to the strike of the upper-plate normal faults.

The mylonitic rocks, largely gneissic in character, are given a "mylonitic" designation because they possess compound ductile-brittle fabrics. Although some mineral components of the mylonitic rocks have deformed brittlely (feldspars, amphibole, garnet), quartz and micas exhibit crystal-plastic deformational behavior. Bulk strain during deformation was ductile. The mylonitic rocks are foliated, possess regionally-consistent stretching lineations most typically defined by elongate quartz grains and linear trains of feldspar porphyroclasts, and have a minimum structural thickness of several kilometers. Their base is not exposed, although Anderson (1988) has noted that rock fabrics become increasingly crystal-plastic downwards and with an increase in metamorphic grade. The mylonitic rocks are developed non-uniformly within the lower plate. Tectonic lenses of non-mylonitized rocks, some with thicknesses up to several hundred meters (Fig. 13), occur within the mylonitic sequence. Geometrically, such lenses are analogous to shear-surface-bounded phacoids of undeformed rock seen within brittle shear zones developed at any scale. The physical conditions of mylonitization are discussed below.

The age of the mylonitic rocks in the lower plate of the Whipple detachment fault is controversial. Whereas the co-authors of this section (Davis and Anderson) believe that Whipple mylonitic rocks formed during deep intracrustal shear during the Tertiary (Davis and Lister, 1988; Anderson and others, 1988), John and Mukasa (1990) report that lithologically similar mylonitic rocks in the Chemehuevi Mountains north of the Whipple Mountains (Fig. 6) are largely Late Cretaceous in age. They believe that the Whipple mylonitic rocks are mostly Cretaceous as well, but that extensive mylonitic overprinting of these mylonitic rocks occurred during Tertiary extension. Our case for Tertiary mylonitization is based on evidence for only one generation of mylonitic fabrics in the Whipple rocks and on the widespread deformation of compositionally diverse Tertiary dike rocks below the mylonitic. Pre-mylonitic, ductile foliations do occur in Cretaceous plutons dated at 89 and 73 Ma (Wright and others, 1986) but these fabrics lack a lineation and are not mylonitic. In addition, Bryant and Wooden (1989) report U-Pb dates of 21.6 ± 1.5 Ma (zircon) and 18.5 ± 0.5 Ma (sphene) from a mylonitic granite in the footwall of the Buckskin detachment fault (= Whipple detachment fault) in the nearby Buckskin Mountains. It is noteworthy that the trend and predominant northeast-directed sense of shear within the mylonitic rocks, as revealed from a study of rock fabrics, is identical to the direction and sense of displacement of the upper plate of the Whipple detachment fault relative to the lower plate. For example, the NE-SW stretching lineations in Whipple and Buckskin mountains mylonitic rocks are statistically parallel to striae on the Whipple and Buckskin detachment fault, and to the trend of the fold or corrugation axes that affect the detachment faults themselves. This kinematic unity or coordination between brittle extensional structures of Tertiary age in the Colorado River corridor and the fabrics of mylonitic rocks in structurally adjacent rocks clearly argues for the widespread Tertiary development of mylonitic rocks as well.

The deformational behavior of rocks now exposed in the footwall of the Whipple and numerous other faults in the Colorado River extensional corridor has changed with time, as follows: (1) penetrative deformation (mylonitization) of pre-existing crystalline rocks in low-dipping zones of intracrustal laminar flow (Stops 4.5 to 4.7); (2) retrograde shearing and brittle deformation of mylonitic gneisses directly below the detachment faults to form thick (up to 300 m) zones of chloritic breccias (Stop 4.4); (3) development of pseudotachylite along uncommon narrow (<3 mm) layers crosscutting the breccias, presumably formed as fault-generated melts during intervals of seismic slip (Stop 4.5); and (4) late-stage formation across the chloritic breccias of discrete detachment faults with directly underlying narrow flinty cataclasites or microbreccias (<0.2 m), and, locally, younger fault gouge.

Davis and others (1983, 1986) and Lister and Davis (1989) have interpreted the progressively overprinted sequence of premylonitic crystalline rocks, mylonitic gneisses, chloritic breccias, and flinty cataclasites or microbreccias as indicating that deep-seated rocks have been transported upward from crustal depths of 10 to 15 km along evolving shallow-dipping shear zones of the general type envisioned by Wernicke (1981, 1985). During crustal extension the originally deep lower-plate rock assemblages of the detachment terrane are drawn upward and outward from beneath the brittlely extending upper plate. Lower-plate rocks become exposed at the surface through variable combinations of regional and subregional warping, tectonic denudation, massive landsliding, and erosion. This model accounts for the abrupt contrast in metamorphic grade and deformational history of upper- and lower-plate rocks now juxtaposed along major detachment faults of Whipple type.

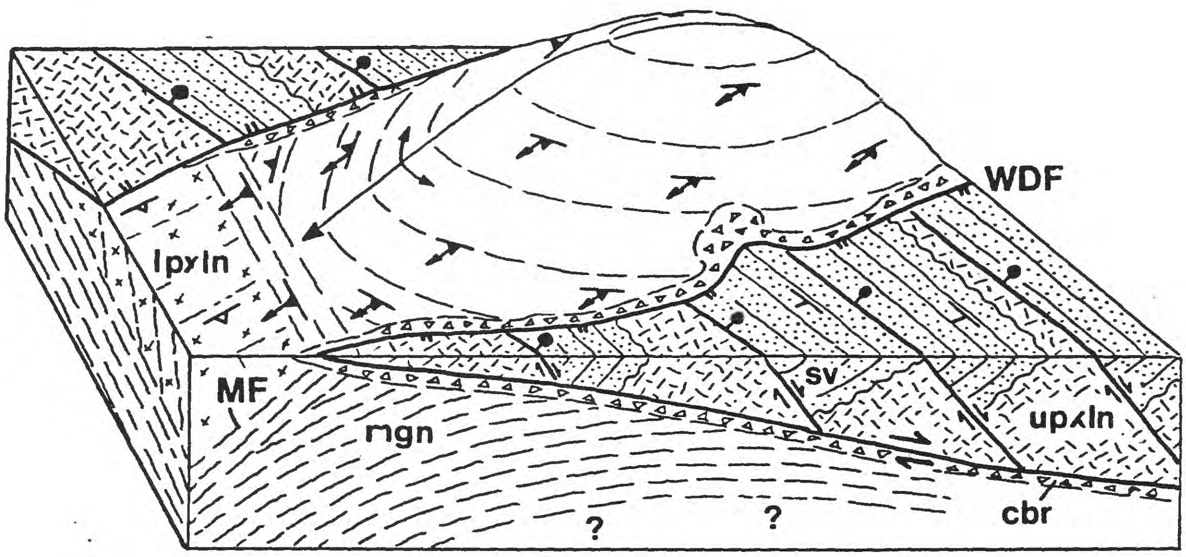


Figure 12. Diagrammatic representation of geologic relations, Whipple Mountains, viewed to the north. Width of block diagram is approximately 30 km; vertical topographic relief, ~1.1 km, is highly exaggerated. Structural features include WDF = Whipple detachment fault and MF = mylonitic front. Rocks units: sv = Tertiary volcanic and sedimentary strata; uplxn = upper plate crystalline rocks; mgn = undivided mylonitic rocks; lpxln = lower plate crystalline rocks above the mylonitic front (not correlative with "upxln" within area of figure; cbr = chlorite breccia zone. From Davis and Lister (1988).

Conditions of Plutonism and Mylonitization

Tectonic decompression of the Whipple Mountains core complex of southeastern California has provided us with a unique window into contrasting crustal levels of magmatic arc development and deformation within this region of the Cordilleran orogen. Previous descriptions of the regional and tectonic setting of the Whipple Mountains core complex have been cited above; the conditions of plutonism and mylonitization have been reported in Anderson (1981, 1985, and 1988), Anderson and Rowley (1981), and Anderson and others (1988); and the origin of the magmatic suites has been addressed by Anderson and Cullers (1990).

While the Mesozoic intrusives are older than core complex mylonitization, Tertiary plutons are both synkinematic and post-kinematic to the mylonitic deformation. The age of mylonitization has been constrained by U-Pb (zircon) dating of synkinematic and post-kinematic plutons (Wright and others, 1986) and $^{40}Ar/^{39}Ar$ dating of the mylonitic gneisses (DeWitt and others, 1986a), with the conclusion that mylonitic deformation was ongoing at 26 ± 5 Ma and had ended by 21 ±1 Ma. Subsequent detachment faulting, which produced the overlying and regionally extensive Whipple detachment fault, is considered to have occurred from 20 - 21 to 14 Ma (Davis and others, 1980, 1982, and 1986). Although the spectacular Tertiary history of crustal extension tends to dominate one's perspective of this core complex, an important record of Cretaceous continental magmatism is preserved in the range's numerous Mesozoic plutons (most, but not all, in the lower plate). Relying on U-Pb ages provided by Wright and others (1986, 1987) and Ar-Ar ages by DeWitt and others (1986a), the succession of Mesozoic and younger events in the lower plate of the Whipple detachment fault (Fig. 11) includes: (1) intrusion of the granitic, informally named Whipple Wash suite of Anderson and Cullers (1990) at 89 ±3 Ma, (2) emplacement of the informally named Axtel quartz diorite of Anderson and Cullers (1990) at 73 ±3 Ma, (3) regional metamorphism and ductile deformation at 69 ±8 Ma, (4) intrusion of numerous biotite tonalite and trondhjemitic aplite sills synkinematic to mylonitization at 26 ±5 Ma, (5) emplacement of the late synkinematic to post-kinematic (with respect to mylonitization) Miocene Chambers Well dike swarm, the youngest members having an age of 21±2 Ma, (6) emplacement of the post-kinematic, informally named War Eagle complex of Anderson and Cullers (1990) of gabbro-quartz diorite at 19 ± 2 Ma, and (7) intrusion of a post-War Eagle complex granodiorite at 16 to 19 Ma.

Thermobarometric studies (Anderson, 1985, 1988; Anderson and others, 1988) have demonstrated that the rocks in the Whipple Mountains have Mesozoic mid-crustal roots, with the older plutons documenting deep-seated emplacement conditions, giving way in time to sequentially shallower crustal levels for progressively younger stages of magmatism and deformation. A basic aspect of the plutonic history of the Whipple lower plate is a provocative range of emplacement depth from deep to shallow with time. Thermobarometric calculations, based on compositions of near-solidus igneous phases, indicate that the Cretaceous plutons were emplaced at middle crustal levels in excess of 28 kilometers, as evidenced by unusually calcic garnet and siliceous muscovite in the peraluminous two-mica granites and aluminous hornblende and probable magmatic epidote in metaluminous quartz diorites. Middle Tertiary intrusive emplacement and concurrent regional development of a low-angle mylonitic fabric occurred at shallower levels, the estimated depth being 16 ± 5 km. Emplacement of post-kinematic Miocene plutons of hornblende-bearing quartz diorite and granodiorite occurred when the same crustal section was at upper crustal levels (~5 to 6 km depth) in route to the near-surface conditions (<5 km depth) that characterize latest detachment faulting at 14-15 Ma (Fig. 14). Thus the lower-plate crystalline assemblages of the Whipple Mountains provide a rare view into middle and upper crustal plutonic members of an evolving magmatic arc. This depth-time sequence of deeper (older) plutons giving way to shallower (younger) stages of plutonism and mylonitization, may be characteristic of metamorphic core complexes. Similar relations have been determined by Anderson (1988) and Anderson and others (1988) for the Sacramento and Santa Catalina Mountains core complexes.

The Cretaceous plutons are composed of two intrusive series as recently described by Anderson and Cullers (1990). The oldest plutons comprise the 89 Ma, informally named Whipple Wash suite of Anderson and Cullers (1990) (Fig. 11), a collection of eight or more granitic intrusives that are marginally metaluminous to peraluminous and include a number of garnet-bearing, two-mica granites. All are low in K and high in Ca and Sr (> 800 ppm) and are concluded to have been derived from a high degree of melting of a continental margin arc-derived, calcic graywacke in equilibrium with an eclogitic residue. Likewise, primitive members of the high-Sr (>1000 ppm), 73 Ma , informally named Axtel quartz diorite of Anderson and Cullers (1990) (Fig. 11) are proposed to have been derived from an eclogitic mafic crust having an original composition of an altered and(or) enriched MOR basalt. Coeval with core complex mylonitization at 26 Ma, the complex was pervaded by swarms of low angle dikes and sills of biotite tonalite and trondhjemitic aplite. The tonalite is compositionally restricted and appears to image the same source as that which formed the Whipple Wash suite some 60 m.y. earlier. The aplites form a coherent group of leucocratic and sodic (Na_2O as high as 7.7 wt. %) intrusives that are strikingly depleted in REE (to subchondritic values) and other incompatible elements.

The 10-km wide Chambers Well dike swarm (Fig. 11) intruded the western portions of the lower-plate rocks during the Oligocene(?) and Miocene. It includes (1) an impressive suite of late kinematic, calc-alkaline andesite to dacite dikes that were affected by late stages of ductile shear (i.e. Tertiary mylonitization), and (2) post-kinematic (~ 21.5 Ma) diabase. The andesites are inferred to have been derived by partial melting of a partly eclogitized amphibolite; the dacites were subsequently evolved by combined fractionation and crustal assimilation processes. The younger diabase dikes represent a separate magma system, one that formed from limited partial melting of enriched mantle.

The last major plutonic event is represented by the 19-Ma War Eagle complex of Anderson and Cullers (1990), a composite intrusion in the north-central Whipple Mountains (Figs. 11 and 13; not seen on this trip), consisting largely of clinopyroxene-hornblende diorite and hornblende-biotite quartz diorite. The most primitive rocks are olivine gabbro having a composition consistent with small degrees of partial melting (~ 5%) of a chondritic mantle, with the remainder of the complex formed by fractionation of a clinopyroxene, olivine, and plagioclase residue. The War Eagle complex was subsequently intruded by a biotite-hornblende granodiorite, the most potassic and least Sr-rich member of the compositionally evolving assemblage of magmas to be emplaced within the core complex.

Rapidity of uplift of lower-plate rock assemblages

Geologic relations, thermobarometric interpretations, and geochronologic data all attest to rapid Miocene uplift of the lower-plate mylonitic rock assemblage from their depths of formation in excess of 10 to 12 km. The age of the end of mylonitization of rocks currently exposed in the lower plate of the Whipple detachment fault is not well constrained. However, the absence of mylonitic deformation in diabase dikes that crosscut mylonitic rocks directly below the mylonitic front indicates that such deformation at highest structural levels in the lower plate had ended prior to 21.5 Ma, an age based on one Ar-Ar age determination for a diabase dike in the Chambers Well swarm (from a locality above the front). Diabase dikes below the front do show limited recrystallization, particularly the development of schistose margins, as well as warping and foliation-parallel offsets (always higher levels displaced to northeast relative to lower). These features indicate that their mylonitic country rocks were still at elevated temperatures and undergoing limited flow and slip following diabase emplacement.

When were mylonitic rocks of the lower plate first exposed at the surface? Unfortunately, the earliest age is not yet known, but cobbles of mylonitic rocks unequivocally derived from the Whipple lower plate occur in abundance in upper-plate Miocene conglomerates in the northeastern Whipple Mountains (Yin and Dunn, 1992). However, the mylonite-clast bearing strata of this area have been offset and rotated to dips as steep as 45^{o} by high-angle normal faults that are truncated downward against the Whipple detachment fault (Dunn, 1986). Hence, lower-plate rocks formed at great depth had been exposed at the surface *during* ongoing crustal extension and were clearly subject to erosion and deposition as sediments prior to the cessation of displacement along the master Whipple detachment fault. The tilted section of mylonite-clast bearing sediments is overlain by a large (~ 70 sq. km), less-tilted, but largely intact gravity slide block of lower-plate mylonitic gneisses. The base of the slide block is truncated by the Whipple detachment fault (Dunn, 1986; Yin and Dunn, 1992). These relations require sediment and mass-wasting transfer of lower plate-derived rocks (mylonitic rocks, chloritic breccias) across an active Whipple detachment system into upper-plate basin(s); Miller and John (1988) describe a similar scenario between sedimentation and tectonics in the Chemehuevi Mountains to the north.

Anderson and others (1988) draw upon their decompression studies showing the progressively shallower intrusion of igneous rocks in the Whipple Mountains in time to conclude that upward transport of mylonitic rocks to the surface occurred at rates of 1.3 to 2.3 mm/yr between ~26 to 14 Ma (Fig. 14). The first geochronologic evidence for rapid cooling of lower-plate rocks, and by inference for rapid uplift, came from fission-track dating of Whipple mylonitic rocks collected just below the Whipple detachment fault. Dokka and Lingrey (1979) found that five fission-track determinations from three mylonitic gneisses (three zircon samples, one sphene, one apatite) yielded ages that varied from 17.9 to 20.4 Ma, with an overlap of error bars between 18.4 and 19.5 Ma. They concluded from their limited data that the mylonitic rocks had experienced a significant temperature drop (>80^{o}, <220^{o}C) between 18 and 20 Ma. Dokka and Lingrey attributed this sudden cooling of footwall rocks to rapid uplift and tectonic denudation, although in 1979 the role of the Whipple detachment fault in accomplishing this uplift had not yet been perceived.

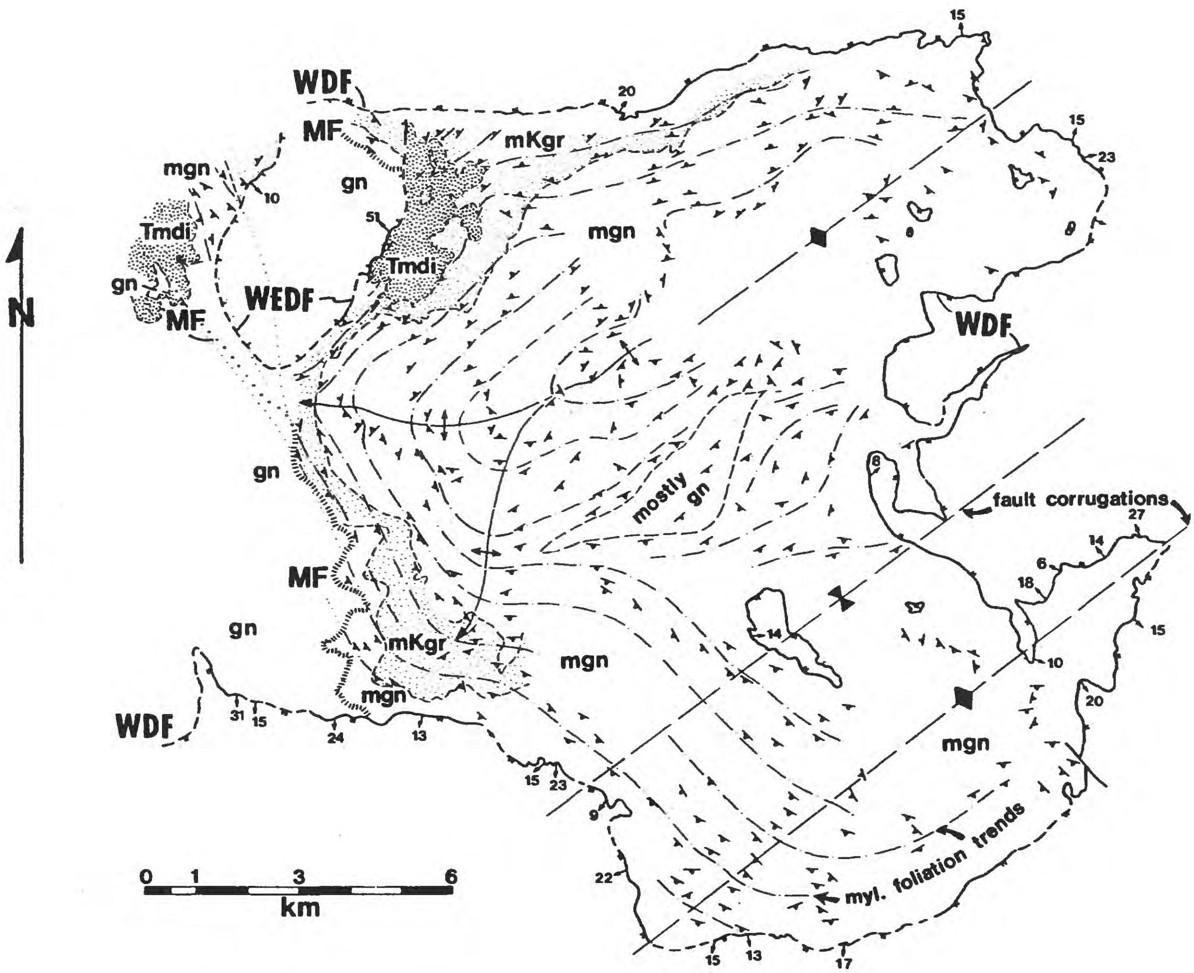


Figure 13. Geologic map of the lower plate of the Whipple detachment fault (WDF), central and eastern Whipple Mountains (from Davis and Lister, 1988). Map emphasizes undulatory or corrugated geometry of the Whipple detachment fault, and foliation trends in mylonitic gneisses (mgn) and a composite Cretaceous granitic pluton (mKgr, light stippled pattern) -- the latter with a sheetlike geometry, now folded. The dips of mylonitic foliation on the flanks of the major northeast-trending antiformal foliation arch are generally steeper than the strike-subparallel dips of the overlying Whipple detachment fault. The major antiformal arch does not appear to be definable directly below the mylonitic front (MF), where structurally higher, northwest-striking mylonitic gneisses lie between northern and southern exposures of the Whipple detachment fault. Nonmylonitized gneisses (gn) overlie the mylonitic front, which is intruded by a composite Miocene pluton of diorite/gabbro, the informally named War Eagle complex of Anderson and Cullers (1990) (Tmdi, heavy stippled pattern). The War Eagle detachment fault (WEDF) offsets the front and the War Eagle complex about 4.5 km to the north-northeast, and is in turn displaced by the higher, younger Whipple detachment fault. Yin and Dunn (1992) suggested that the War Eagle fault may be the base of a scoop-shaped landslide block that originally transferred lower-plate units across the Whipple detachment fault into an upper-plate depositional basin (see Highlights on the Way to Stop 4.1).

Davis (1988; p. 206) concluded that limited $^{40}Ar/^{39}Ar$ and fission track age data from the Whipple Mountains "collectively document rapid cooling of the mylonitic gneisses from above 450°C to below 200° C between 20 and 18 Ma ago." This interpretation has been criticized by Foster and others (1990) based on choice of appropriate closure temperatures for muscovite and orthoclase. Richard and others (1990) report the results of a $^{40}Ar/^{39}Ar$ analysis of a microcline collected in the western part of the Buckskin Mountains and from a mylonitized granitic pluton that all workers would assign to the lower plate of the Whipple (= Buckskin) detachment fault. The K-feldspar records a cooling gradient from 26.6 to 19 Ma. Four K-feldspar samples from a mylonitized Late Cretaceous pluton, which is located near the top of the mylonitic sequence in the Whipple Mountains, were collected by G. Davis with Roger Buck from three localities along a 16 km-long west to east traverse for $^{40}Ar/^{39}Ar$ analysis. Preliminary results by David Foster (R. Buck, written communication, July, 1990) indicate that the feldspars in all samples had reached an argon closure temperature of ~235° ± 10° C by 17 ± 1 Ma ago. Foster and others (1990, p. 20,021) concluded that the Miocene cooling history of the Whipple Mountains is apparently similar to that of the Chemehuevi Mountains, where they found cooling rates of 10-40°/m.y. over the interval 22-16 Ma.

Initial dip of the Whipple detachment fault -- steep or shallow?

The question of the shallow versus steep initial dip of Cordilleran detachment faults is a highly controversial one. Field geologists concerned with the mapping of such faults overs areas of thousands of square kilometers generally opted for an initial, shallow-dipping (<30°) configuration, regardless of differences of opinion regarding their origin. But concerns about the likelihood of existence of extensional stress fields appropriate for the initiation and operation of low-angle normal faults, and about the paucity of seismological evidence for ongoing crustal extension along normal faults dipping less than 30° (Jackson, 1987) have encouraged geologists and geophysicists to consider various models postulating steeper original dips for Cordilleran detachment faults (for example G. H. Davis, 1983; Wernicke and Axen, 1988; Buck, 1988). Nevertheless, most geologists who have studied detachment faulting in the Colorado River corridor continue to believe that the faults of that region propagated to or very near the Earth's surface at shallow angles of dip -- ~40-30° or less (for example, Howard and John, 1987; Miller and John, 1988; Lister and Davis, 1989; Richard and others, 1990; Foster and others, 1990).

The mylonitic front in the lower plate of the Whipple detachment fault had an early Miocene depth in excess of 12 km (that is, 16 ± 4 km) based on mylonitic metamorphic barometric calculations conducted by Anderson and others (1988). That front now lies approximately 27 km east of the likely Miocene surface breakaway of the Whipple detachment fault system in the western Mopah Range. An *average* dip for the Miocene Whipple detachment fault between its surface intersection and the top of the mylonitic front would thus vary from 24° (front at 12 km depth) to 36° (front at 20 km depth).

If the relationship of the Whipple detachment fault to upper-plate geologic units is considered, a very shallow angle of dip in the uppermost three or four kilometers of the crust appears to be required -- an indication that a constant average dip of the fault from surface levels to the mylonitic front is not likely. In the eastern Whipple Mountains, the Whipple detachment fault rises upwards out of Precambrian upper-plate crystalline basement rocks into a gently dipping (≤ 20°) basal Tertiary volcanic and sedimentary section. At a number of localities in this general area, subhorizontal Tertiary strata rest directly on the subhorizontal Whipple detachment fault! The conclusion seems inescapable that upper levels of the Whipple detachment fault were active within several kilometers of the earth's surface during Miocene time (the Tertiary section is no thicker than this) and that the fault crossed a subhorizontal Tertiary-basement unconformity in the eastern Whipple Mountains at a shallow angle of 20° or less. Furthermore, for most of the 50 kilometer distance beneath eastern Whipple exposures of the fault and its probable breakaway zone in the Mopah Mountains, the fault is overlain either directly by Tertiary strata or by basement rocks for that strata that are rarely more than a few hundred meters below the basal unconformity. Given the likelihood that all or most of the basal Tertiary section over this 50 km-long interval is older than the 18.5 -Ma Peach Springs Tuff, an additional conclusion seems required (as confounding as it may be) -- that for many tens of kilometers the Whipple detachment fault lay within a few kilometers (probably less than 3 or 4) of the earth's surface!

The initial configuration of the Whipple detachment fault at depths below the mylonitic front is imperfectly known, but clues are provided by lower-plate rock units. The angle now observed in outcrop between the Whipple detachment fault and the mylonitic front in its footwall, ~10-25°, presumably records the approximate angle of initial discordance between deep, regionally subhorizontal(?) mylonitic gneisses and the somewhat steeper and younger Whipple detachment fault that cut across them. A sheetlike mid-

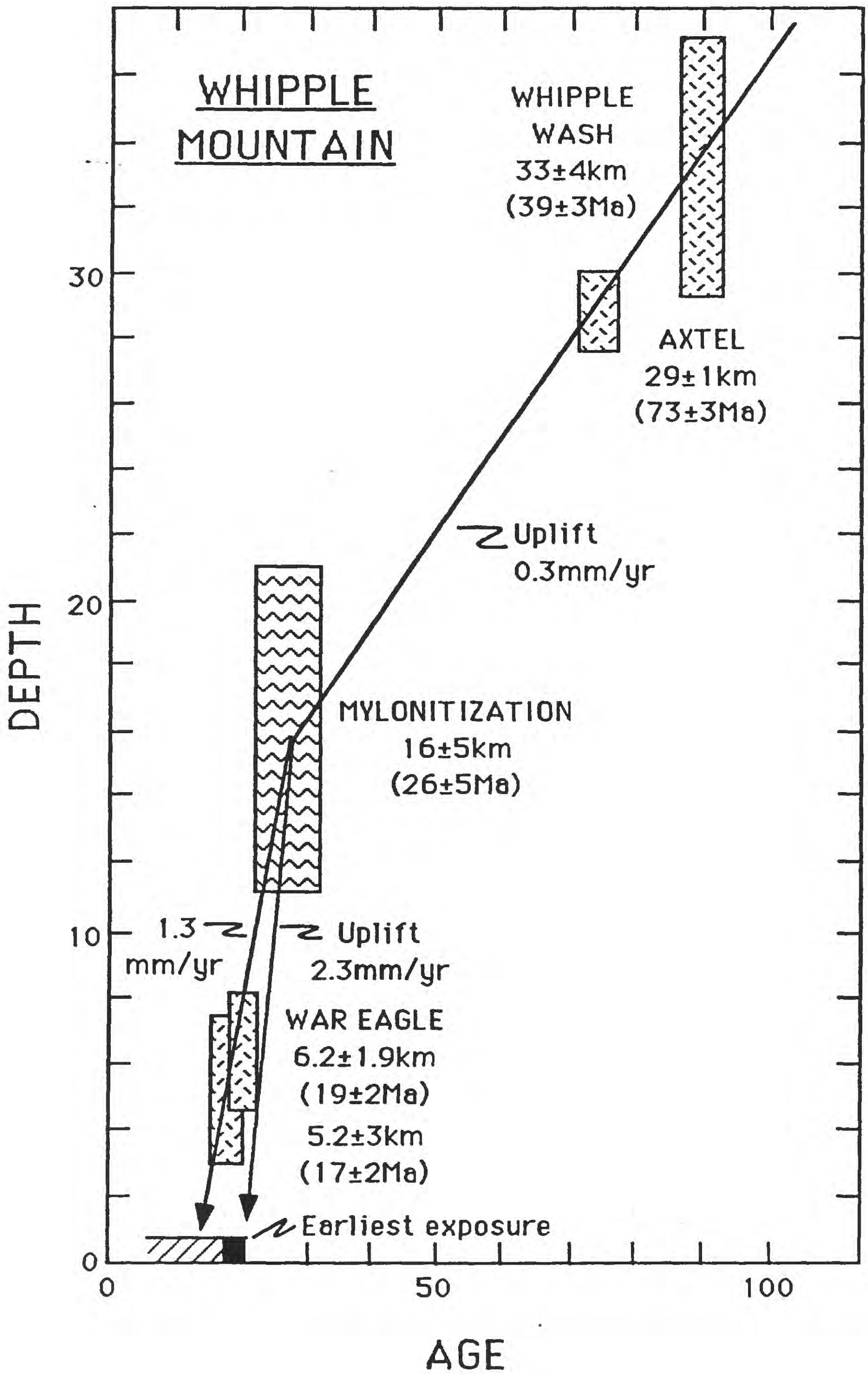


Figure 14. Upward transport of the Whipple crustal section based on mineral barometry of pluton emplacement and mylonitic deformation. Ages are from U-Pb, zircon (Wright and others, 1986). From Anderson and others (1988).

crustal Cretaceous pluton of the mylonitized Whipple Wash suite (Fig. 11) lies just below and parallel to the Whipple detachment fault for a distance of 13 km east of the mylonitic front. If is is assumed that the pluton was initially subhorizontal, then the fault must have flattened to a parallel orientation in the upper middle crust just above the pluton. Similarly, the regionally persistent metamorphic grade of footwall rocks (upper greenschist to middle amphibolite facies) for tens of kilometers downdip within the lower plate argues that the fault that has carried them to the surface sampled only a mid-crustal horizon of persistent metamorphic grade and, by inference, depth.

Collectively, the geologic relationships described above suggest that the Whipple detachment fault may have had the following Miocene eastern dip configuration: (1) < 10^{0} at levels within 4-5 km of earth's surface; (2) 20-40^{0} between 4-5 km depth and the depth of the mylonitic front (16 ± 4 km); (3) a dip of 10-25^{0} at the level of the mylonitic front, assuming that the front was regionally subhorizontal; (4) a subhorizontal (<10^{0}?) dip for eastward distances of tens of kilometers below the mylonitic front. This geology-based fault dip reconstruction is compatible with interpretations of initial fault dip based on $^{40}Ar/^{39}Ar$ analyses of the thermal history of lower plate rocks during their transport to the earth's surface by middle Miocene time (Richard and others, 1990; Foster and others, 1990).

Highlights on the Way to Stop 4.1

From the motel, cross London Bridge, and make three successive left turns at the traffic lights in order to enter southbound onto Arizona Highway 95. We will follow it 23.2 miles before turning right to cross Parker Dam into California. One mile down Arizona 95 from London Bridge, roadcuts expose brecciated gneiss and granitoids that are part of a huge Miocene landslide megabreccia sheet. The exposures are part of a fault block of SW-tilted Miocene volcanic and sedimentary rocks that form the prominent Aubrey Hills to the south, between the highway and Lake Havasu. The megabreccia deposit, several kilometers long and hundreds of meters thick, is intercalated in the Miocene section, above sandstone and basalt dated by J.K. Nakata as 14.1±0.3 Ma by K-Ar on whole rock (Nielson, 1993). Clasts as wide as hundreds of meters in the breccia grade laterally into crackle breccia and mixed breccia in the landslide matrix. The megabreccia sheet was derived as a landslide from lower-plate rocks exposed only in the western Whipple Mountains, and was then evidently transported tectonically 20 eastward to here in the northern Aubrey Hills (see caption for Fig. 13, and the section on Rapidity of uplift of lower-plate rock assemblages).

After about 15 miles, traverse roadcuts in faulted Early Proterozoic gneisses and in granite and quartz monzodiorite correlated with two 1.4-Ga units, the informally named Parker Dam granite of Anderson (1983) and the the informally named Bowmans Wash quartz monzodiorite of Anderson and Bender (1989). These rocks are in upper-plate tilt blocks above the Whipple detachment fault (Whipple Mountains detachment fault of Dickey and others, 1980, and Carr, 1991). The upper-plate crystalline rocks are are typically highly faulted, varicolored from oxidation, and form low rubble-covered hills. Locally, the fault blocks of the Aubrey Hills also expose road cuts of Miocene sandstone dipping steeply southwest.

The Whipple Mountains across Lake Havasu to the right (south and southwest) expose rugged, dark klippen faulted onto lower-plate rocks that are mostly lighter in color. Ives' (1861) name for the Whipple Mountains was the Monument Mountains (Fig. 10a,b). Tabletop mesas (Fig. 10b) in the Buckskin Mountains ahead to the southeast are caps of flat-lying middle to upper Miocene basalt that lie across the tilted rocks of the upper plate and record the cessation of extension .

After crossing a bridge across the Bill Williams River arm of Lake Havasu, large cuts will be seen along the left (S) that expose Miocene clastic rocks and basalt dikes that served as feeders for the overlying meas-forming basalt flows. The cuts are at a pumping station where Colorado River water is withdrawn to service the parched lands of central Arizona. Turn right to Parker Dam, and drive one mile to cross the dam and its abutments of the porphyritic 1.4-Ga informally named Parker Dam granite of Anderson (1983). The following mileages are from Parker Dam.

Past Parker Dam 0.5 mi, turn right past the Parker Dam settlement toward Black Meadow Landing. At mile 0.9, pass an angular unconformity in the lower Miocene section. At mile 1.3, pass exposures of a debris-flow unit within the early Miocene Gene Canyon Formation of Davis and others (1980). Clasts are of mylonitic leucocratic granite that yielded a fission-track age on sphene of 83 Ma (Dokka and Lingrey, 1979). This age is greater than fission-track ages within the lower-plate rocks of the Whipple Mountains, and therefore implies a source of Cretaceous or older mylonitic rocks not now exposed. At mile 2, note red beds of the Miocene Copper Basin Formation of Davis and others (1980), which have gentler dips than the Gene Canyon Formation owing to a general upward fanning of dips. At mile 2.9, keep right on the better road. On the right across Gene Reservoir, red beds of the Copper Basin Formation can be seen dipping at about 25-25° SW toward a fault that separates them from older crystalline rocks. At mile 4.0 note an intrusive contact of a light-colored Cretaceous pluton against Proterozoic gneiss. At 6.2 miles past Parker Dam, park in an open area on the left for Stop 4.1 before heading down major switchbacks.

Stop 4.1

Parker Dam granite and view of upper-plate crystalline complex (Fig. 15)

The informally named Parker Dam granite of Anderson (1983), dated at 1.40 Ga (U/Pb, zircon) by Jim Wright (Anderson and Bender, 1989), is a major intrusion in the eastern and southeastern portions of the upper plate of the Whipple Mountains. The granite is distinctive in having abundant (>34 percent) K-feldspar phenocrysts aligned in a well-defined planar flow foliation. Described by Anderson and Bender (1989), this granite and the informally named Bowmans Wash quartz monzodiorite of Anderson and Bender (1989) of similar age (dark hill 1 km to the north) are part of a transcontinental suite of Proterozoic anorogenic plutons (Anderson, 1983).

The view to the north displays a heterogeneous mosaic of colors representing complex mixture of crystalline units that comprise this upper-plate basement (Fig. 15). The oldest rocks include paragneisses and migmatites (light-colored areas to the northwest), amphibolite (dark-grey areas), and foliated granites and augen gneisses (reddish area west of the road downhill). These rocks all possess a steep foliation that formed during high amphibolite-grade Proterozoic metamorphism. Orrell and Anderson (1987) have estimated peak metamorphic conditions ($660^{o} \pm 40^{o}$ C at 3.5 ± 0.5 kb) and believe that this area is part of a regional low pressure high-amphibolite to granulite terrane that includes western Arizona and extends north to southern Nevada (Thomas and others, 1988).

From the standpoint of Tertiary structure, this stop lies approximately 1 km below the gently east-dipping Whipple detachment fault last exposed some 5 km to the west. The Precambrian terrane below and north of us at this stop is truncated and displaced downward along a prominent east-dipping normal fault that can be seen to the northeast along the west side of the prominent resistant ridge. That ridge is capped by clastic rocks of the Copper Basin Formation of Davis and others (1980) that rest directly on the Precambrian basement assemblage; the lower, Gene Canyon Formation of Davis and others (1980) is thus missing here, although it reappears along strike to the northwest in steeply dipping exposures that are locally overturned. Note that the Copper Basin strata and the underlying nonconformity are folded into a hanging wall syncline just east of the fault. The same stratigraphic units, again resting depositionally on the Precambrian units north of us, can be seen on the skyline ridge to the west. These units dip southwestward, away from us, and are truncated along the western base of the ridge by another major normal fault, the Copper fault. This normal fault repetition of Tertiary strata and Precambrian basement (Figs. 11, 12), first noted by Kemnitzer (1937), is characteristic of the upper plate of the Whipple detachment fault. Thirty-one measurements of fault dip in the eastern Whipple Mountains average 46^{o}, with a range of 30-66^{o} (or $48^{o} \pm 18^{o}$). These faults are restricted to the upper-plate and none has been observed to flatten downward with a listric geometry into the Whipple detachment fault.

Highlights between Stops 4.1 and 4.2

Drive northwest and take the prominent unpaved road to the left (west) after 0.5 mi. Here, we leave the Parker Dam granite of Anderson (1983) and drive into older foliated granodiorite and a large area of gneiss and amphibolite. Swarms of Proterozoic ophitic diabase are notable, particularly on the right side one mile past our last turnoff. In 2.2 mi, turn left and continue on the dirt powerline road; the main road (to Havasu Palms Resort) turns right. Use of passenger vehicles on the powerline road is not recommended. We will pull off on the right shoulder of the road, 1.1 miles past the last turnoff, for a view of both upper and lower Whipple plates.

Stop 4.2

View of Whipple Wash area (Fig.15)

Walk west across the road and follow the burro trail 75 feet to its end. From here we are afforded a spectacular view to the northwest into the low country of Whipple Wash. Rocks on the northern skyline are light-colored mylonitic gneisses of the Whipple lower plate. The Whipple detachment fault is the east-dipping planar surface beneath several dark-brown to reddish klippen containing Tertiary strata that dip steeply (>50°) to the southwest *into* the truncating Whipple detachment fault. The pale greenish tint of the lower-plate mylonitic gneisses is due largely to the development of chlorite in lower-plate rocks -- including chloritic breccias directly below the fault - that have undergone retrograde metamorphism. Note the slight discordance here between the northeast-dipping Whipple detachment fault and the gently southwest-dipping foliation of the mylonitic gneisses. This discordance between mylonitic foliation and the Whipple

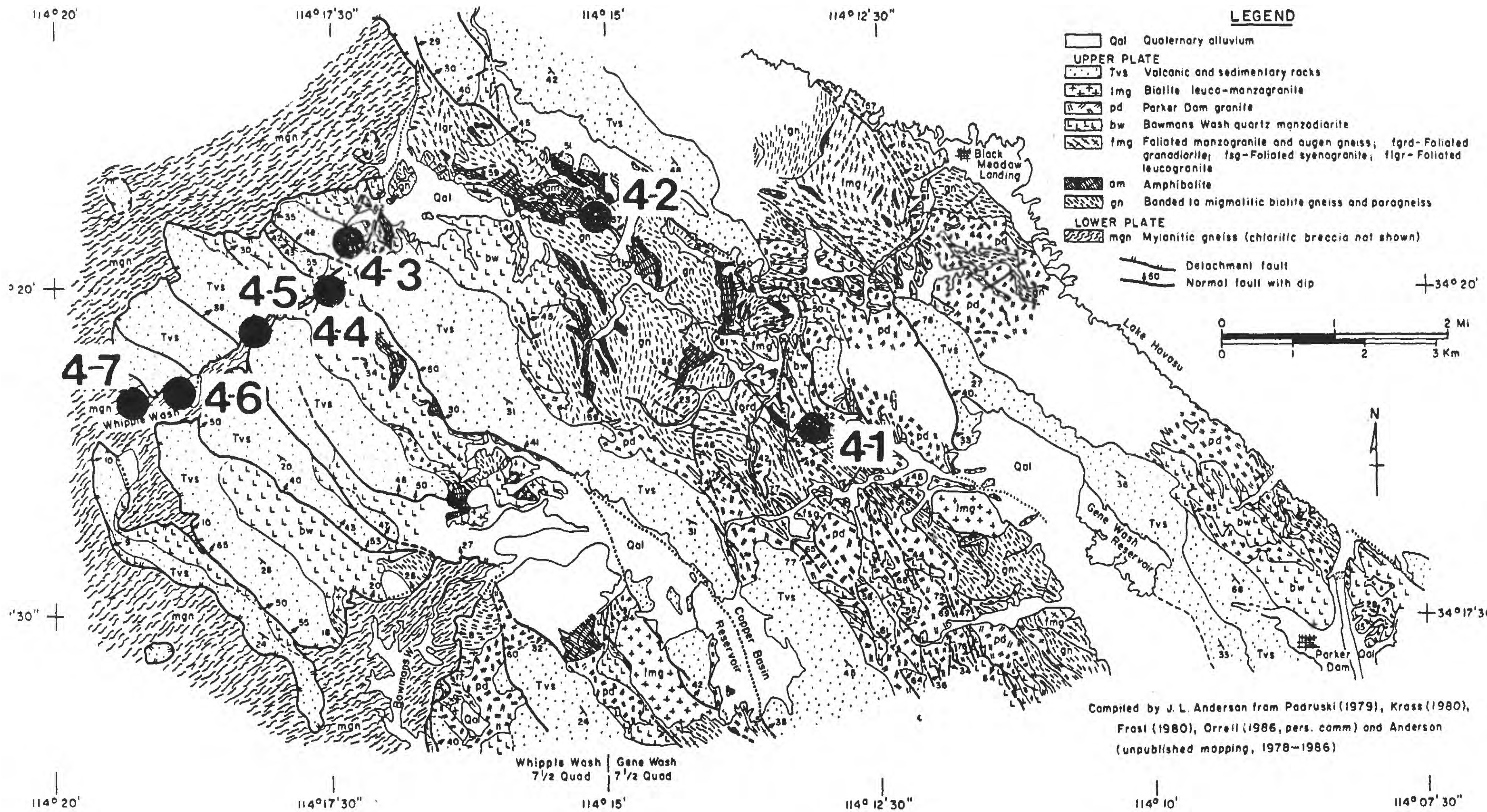


Figure 15. Geologic map of the eastern Whipple Mountains showing location of Stops 4.1 to 4.7.

detachment fault is characteristic of the range and becomes most pronounced near the mylonitic front nearly 20 km to the west.

The contrast in appearance between the upper and lower plates is strikingly evident from this vantage point. We are standing on upper-plate amphibolite; other upper-plate units nearby include quartzofeldspathic gneiss, the Bowmans Wash quartz monzodiorite of Anderson and Bender (1989), diabase, and a few Tertiary basalt dikes. Overlying this basement to the west are cliff-forming Tertiary volcanic and sedimentary rocks. Dark-colored Tertiary rocks to the east are steeply SW-dipping to overturned volcanic rocks that lie with angular unconformity beneath reddish Copper Basin clastic strata. These rocks lie in the hanging wall of a northeast-dipping normal fault (Fig. 16, eastern end).

Highlights between Stops 4.2 and 4.3

Resume driving northward 1.1 mi, descending into Whipple Wash. Park where the powerline road crosses the wash and gather field gear, water, and lunch for a 14 km (9.2 mile) round-trip hike that will occupy the rest of the day. All stops are in the canyon bottom, an area within the 7 1/2-minute Whipple Wash quadrangle. The rest of this day's guide is written for foot travelers and mileages starting from 0.0 at the parking area are estimated. All stop distance measurements are in feet or miles. We will make our first trail stop after approximately 1.6 miles.

Stop 4.3

Bowmans Wash quartz monzodiorite (Fig. 15)

The informally named Bowmans Wash quartz monzodiorite of Anderson and Bender (1989), dated at 1.41 Ga (U/Pb, zircon) by Jim Wright (in Anderson and Bender, 1989), is the dominant Proterozoic unit of the northeastern Whipple Mountains. The "salt-and-pepper" texture visible in this dark outcrop is typical of most of the pluton. The pluton ranges in composition from quartz diorite to quartz monzonite and is one of the more mafic 1.4 Ga intrusions of the region, with SiO_2 ranging from 60.7 to 63.8 wt. percent. It is commonly foliated near its contacts with the younger Parker Dam granite of Anderson (1983).

Continue walking upstream. At 1.9 mi, up on the left bank, is a small upper-plate fault marked by a spring nourishing one of the few native palm trees in this part of the eastern Whipple Mountains. The sheared, gently west-dipping unconformity at the base of the Tertiary volcanic section is about a third of the way up the canyon wall. Many faults in the Tertiary section will be visible on this trek, but none are observed or believed to cut the detachment fault that lies not far below us here (Fig. 16; Gross and Hillemeyer, 1982). At 2.3 mi is an excellent exposure of the exhumed Whipple detachment fault surface.

Stop 4.4

Whipple detachment fault (Fig. 15)

The Whipple detachment fault is the tectonic boundary between the contrasting upper and lower plates. The fault, as seen here in a small prospect pit, is underlain by a 1- to 3-m-thick ledge of dark brown cataclasite or microbreccia, that passes downward into altered lower-plate rocks of the "chloritic breccia zone." This zone, up to 300 m thick in the range, is so pervasively fractured, sheared, and altered by chloritization and epidotization that the foliation and lineation of its protolith mylonitic gneisses are generally not discernible. The low-grade copper mineralization of the microbreccia and the retrogression of underlying lower-plate rocks are regional phenomena.

The next 600 feet of our trek upstream is across exposures within the chloritic breccia zone. The detachment fault is intermittently exposed on the canyon walls below the cliff-forming Tertiary section. Its gentle dip is evident as it gradually climbs in elevation westward above us. Bedding in the Tertiary sequence is shallow dipping (Fig. 16), and the Whipple detachment fault cuts upwards to the west at a low angle across the Precambrian-Tertiary nonconformity. Planar normal faults in the upper plate (Fig. 16) are interpreted by us as once having listric geometries. Reasons for their now-planar geometries will be discussed at the next stop. Gradually, the mylonitic foliation of lower-plate rocks becomes evident within the chloritized gneisses, with dips of 20°-44° SW and a lineation plunging at low angles to the southwest (S 45°-53° W). These mylonitic gneisses, last deformed during the Miocene, are derived from Proterozoic quartzofeldspathic gneiss, augen gneiss, and amphibolite. At 2.9 mi is a ledge of mylonitic gneiss, our next stop.

Stop 4.5

Jutaposition of older gneissic and younger mylonitic foliations; pseudotachylite (Fig. 15)

Giving the outcrop a crossbedded-like appearance, gneiss layers with a steeper, older (Mesozoic?) foliation are preserved within the mylonitic gneisses. This older foliation (N 87° E, 42° SE) is also present

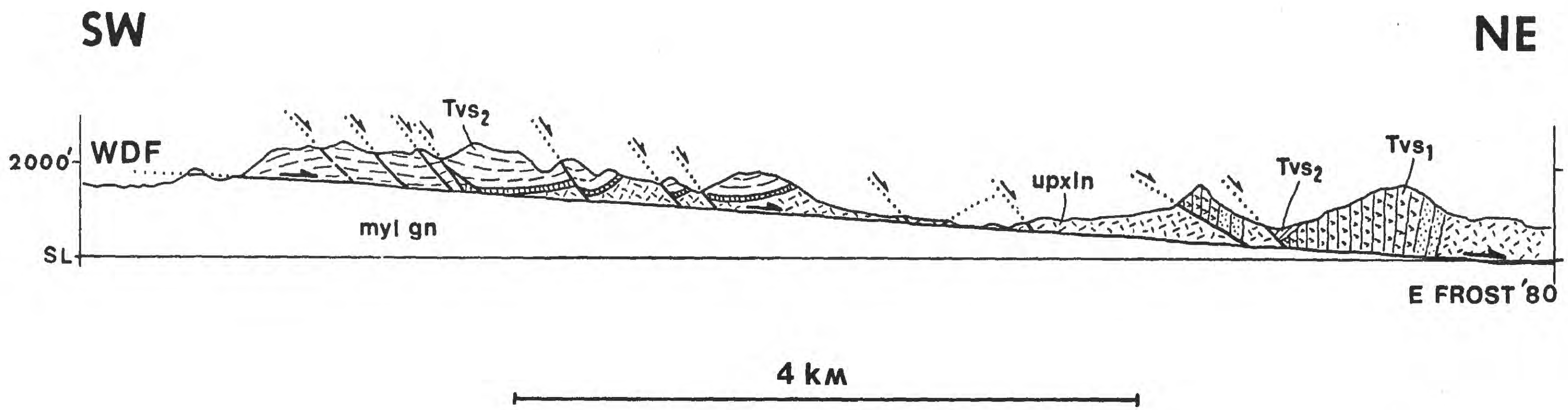


Figure 16. Geologic cross section through the Whipple Wash area, northeastern Whipple Mountains. Tvs1 and Tvs2 are older and younger Miocene strata, respectively; uplxn = upper plate crystalline rocks; myl gn = mylonitic gneisses; WDF = Whipple detachment fault. Chloritic breccias below WDF are not shown. From Frost (1980).

locally in Cretaceous plutons to be seen at Stop 4.7. The attitude of the mylonitic foliation is N 64° W, 31° SW with a lineation plunging 29°, S 52° W. A gray-colored mylonitic tonalite sheet, contained in the mylonitic gneisses, is considered equivalent to one dated at 26 ± 5 Ma (U/Pb, zircon) by Wright and others (1986). As seen here, the transition from nonmylonitic to mylonitic fabrics involves a gradual rotation to lower dips, with the mylonitic fabric becoming increasingly discernible as the dip lessens.

Along the south wall of the canyon at this locality is a splendid view of the Whipple detachment fault with chloritized mylonitic rocks below it, subhorizontal Tertiary strata above it, and several, moderately-dipping planar faults cutting Tertiary units, but ending abruptly downward at the Whipple detachment fault. The lower-plate gneisses at this locality are disrupted by a family of anastomizing, shallow-dipping faults. The resulting lensoidal geometry of fault-bounded slices is clearly related to the presence of the overlying Whipple detachment fault. Older, and presumably once deeper and hotter, fault surfaces transect the mylonitic rocks at this locality as thin (<2-3 mm), subhorizontal pseudotachylite(?) seams. Some of these seams spawn flame-like apophyses that extend out across the bordering rocks. These inferred pseudotachylites are now recrystallized, but are interpreted to have formed as rock melts generated during episodes of seismic slip along shallow-dipping fault surfaces. Such pseudotachylites are not common within the Whipple complex.

At 3.4 mi, a landslide of Tertiary rocks covers the detachment surface, now far above us, on the north side of the wash. At 4.1 mi, we gather for the next stop.

Stop 4.6

Lower-plate low-angle fault and mylonitized augen gneiss (Figs.15, 17)

A gentle ramp of augen gneiss is exposed approximately 500 feet past the landslide and below a planar east-dipping low-angle fault. Chloritic breccias above the fault are more altered and shattered than those below. An east-vergent fold in rocks directly above the fault indicates its sense of displacement. The underlying augen gneiss is a distinctive rock that occurs in the lower-plate gneisses as 2-10 m-thick layers exposed continuously for distances up to 2 km. Ranging from monzogranite to syenogranite in composition, the augen gneiss is a metaigneous rock containing large (to 5 cm) augen of K-feldspar set in a biotitic mylonitized matrix. Where it is not mylonitized, the gneiss contains the Mesozoic (?) fabric seen earlier. Sam Bowring has dated the intrusive at 1.41 Ga (U/Pb, zircon, pers. comm., 1987). Although not correlative with plutons of similar age in the upper plate, the rock is similarly potassic (4.5 to 5.5 wt. percent K_2O) and with an elevated Fe/Mg ratio.

The augen gneiss at this locality is intruded by a two-mica granodiorite pluton of the Whipple Wash suite of Anderson and Cullers (1990). Localized zones of fluid flow and related alteration are visible along bleached conjugate joint sets and tension gashes in the granodiorite. Kinematically, the joint sets and tension gashes should be late kinematic to mylonitization yet the timing and origin of fluids related to the alteration could either be coeval with the mid-crustal deformation or much later. This question has recently been addressed by Morrison and others (1990) and Morrison (1994) in a detailed oxygen isotope study of the role of fluids during ductile and later brittle deformation of lower plate lithologies. Morrison (1994) has analyzed a large number of mylonitic rocks, including samples from this outcrop (note drill holes across bleached zones) and elsewhere. Values of $d^{18}O_{WR}$ for these drill core samples range from 4.4 to 5.5 $^{o}/_{oo}$ (SMOW). These values are strikingly lower than normal igneous $d^{18}O$, including that observed for unaltered portions of the Whipple Wash suite of Anderson and Cullers (1990) (9 to 10 $^{o}/_{oo}$) and document exchange with low $d^{18}O$ meteoric water (Morrison, 1994). The meteoric fluid-rock exchange is estimated to have occurred at ~350^{o}C at depths less than 6km. Unusually deep circulation of surface-derived fluids is not required to explain these low $d^{18}O$ mylonites, but rather they are explained by shallow circulation of fluids related to detachment faulting.

Continue walking upstream past a spring and inviting pools in the augen gneiss. At 4.3 mi, a major valley appears from the left. At 4.6 mi is a broad, stream-washed exposure of a garnetiferous two-mica granodiorite, our final stop of the day.

Stop 4.7

Cretaceous garnet-two-mica granodiorite and Miocene synkinematic tonalite (Figs. 15, 17)

At this outcrop, Proterozoic gneisses and amphibolite are intruded by a garnetiferous two-mica granodiorite and cross-cutting, shallow-dipping dikes of gray biotite, tonalite, and white trondhjemitic aplite. From samples collected at this locality, the granodiorite and tonalite have been dated by U/Pb (zircon) at 89 ± 2 Ma and 26 ± 5 Ma respectively (Wright and others, 1986a). All of the rock types are mylonitized, demonstrating a Tertiary age for this deformation. The mylonitic foliation in the granodiorite

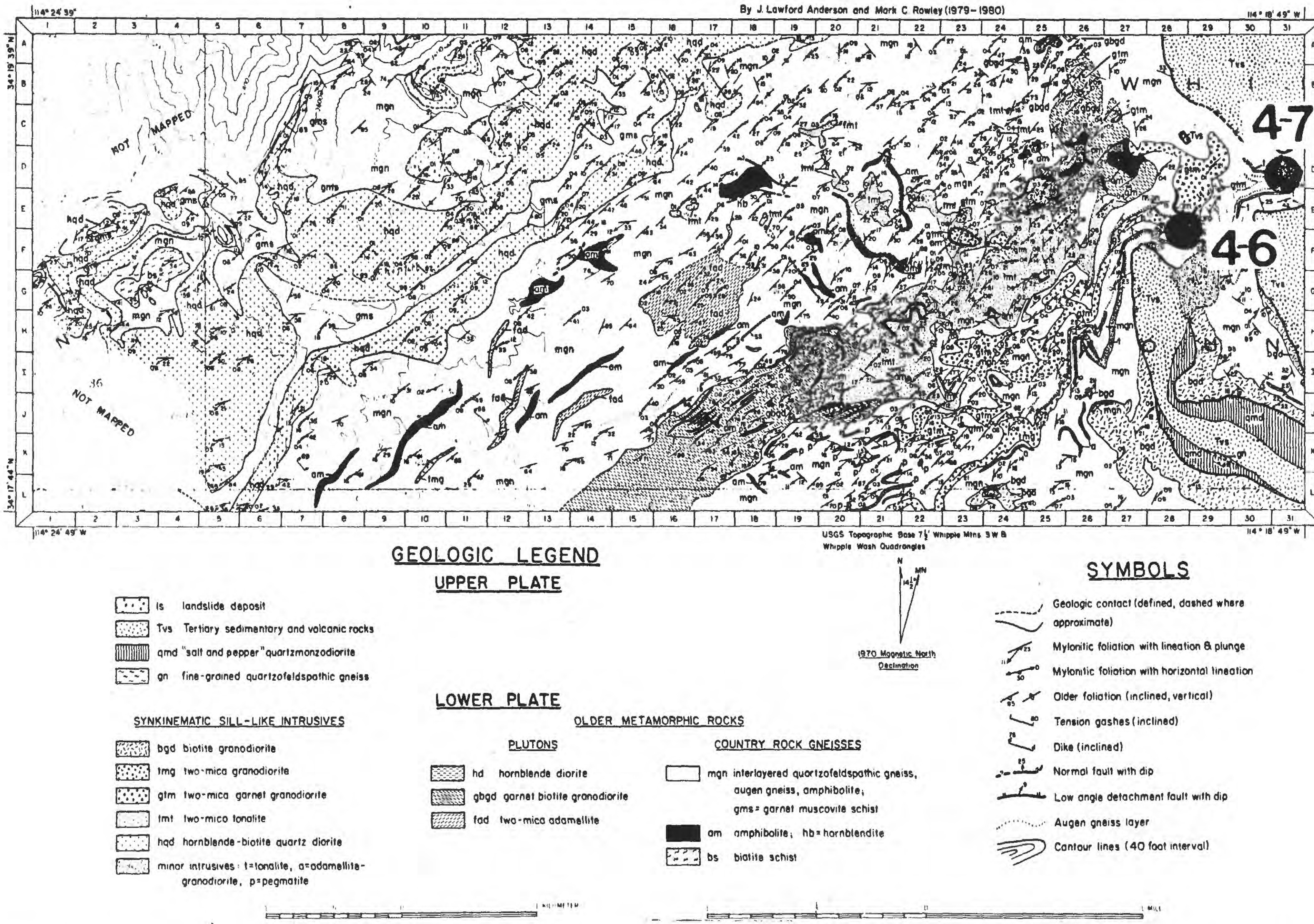


Figure 17. Geologic map of the upper Whipple Wash area depicting location of Stops 4.6 and 4.7. Upper plate units include: ls = land slide; Tvs = Tertiary volcanic and sedimentary strata; qmd = the 1.4 Ga informally named Bowmans Wash quartz monzodiorite of Anderson and Bender (1989); and gn = ~1.7 Ga gneiss. Lower plate units include the following subunits of the informally named Whipple Wash suite of Anderson and Cullers (1990): bgd = biotite-sphene granodiorite, tmg = two mica granodiorite, gtm = garnet, two mica granodiorite, tmt = two mica tonalite, fad = foliated monzogranite, gbgrd = garnet-biotite granodiorite. The informally named Axtel quartz diorite of Anderson and Cullers (1990), shown at deeper structural levels, is labeled hqd. Proterozoic host units include: bs = biotite schist, am = amphibolite, and mgn = quartzofeldspathic gneiss. Modified from Anderson and Rowley (1981).

and surrounding gneisses strikes N 18° E, dipping 19° SE with a lineation plunging 8°, N 44° E. The younger tonalite and aplite dikes cut this mylonitic fabric discordantly by as much as 25° and have mylonitic foliations parallel to their walls. Mylonitic lineations are closely parallel in all rock units. These relations require the tonalite and aplite to have been intruded synkinematically during a protracted period of mylonitization. The older, steeper nonmylonitic foliation viewed earlier at Stop 4-5 is locally present in the granodiorite, indicating a maximum Cretaceous age for that deformation. The older foliation at this locality strikes N 53° E, dips 44° SE, and does not contain a lineation.

The mylonitized granodiorite is a typical intrusive in this area of the core (Fig. 17) and forms a shallow-dipping body with a maximum thickness of 70 m. The rock is light gray, medium grained, and contains porphyroclasts of garnet, feldspar, and two-micas set in a fine-grained mylonitic matrix. Anderson and Rowley (1981) have described the compositional features of this and the other intrusions of the lower plate. Data given by Anderson and others (1979) and Anderson and Rowley (1981) show that the mineral phases in this and other mylonitic gneisses have a wide range of composition, including porphyroclasts of apparent igneous chemistry and matrix grains re-equilibrated during mylonitization conditions of upper greenschist to lower amphibolite grade (Anderson, 1988).

Description of the Whipple Wash suite

Based on field relations, eight marginally metaluminous to peraluminous granitic plutons occur in this area (Figs. 11 and 17). They were informally termed the Whipple Wash suite (Anderson and Cullers, 1990) after the usually dry wash that transects most of their exposures, including that of this stop. All have the equivalent U-Pb isotopic age of 89 ± 2 Ma (Wright and others, 1987) and are sufficiently similar in composition to allow treatment as one magmatic entity. Intruded as shallow-dipping intrusive sheets (up to 0.5 km thick), the plutons are compositionally stacked with leucocratic, metaluminous biotite-sphene granodiorite and tonalite comprising the uppermost structural levels, and peraluminous two-mica granodiorite, garnet two-mica granodiorite (this stop), garnet-biotite granodiorite, and two-mica tonalite occurring at deeper structural levels. All are mylonitically deformed and some of the intrusions contain the earlier (late Cretaceous) metamorphic fabric also observable in this exposure.

Compositionally, the suite spans a silica range from 61.1 to 73.4 wt. %. Although granodiorite is the dominant mode, the rocks range from tonalite to monzogranite. The peraluminous intrusives contain two micas ± garnet with mineral compositions consistent with crystallization at pressures greater than 7.6 kb (perhaps greater than 9 kb by some thermobarometric calibrations) at 746 ± 55°C (Anderson and others, 1988). Such high pressures are suggestive of the occurrence of magmatic epidote (Zen and Hammarstrom, 1984); porphyroclasts of this phase, some containing cores of allanite, are common in most of the plutons. Comparably high pressures of emplacement have also been estimated from hornblende barometry of the 73 Ma Axtel quartz diorite of Anderson and Cullers (1990) that is exposed further up this wash (Fig. 11).

Although mylonitization has affected the composition of mineral phases in these plutons, relict igneous compositions are preserved in most porphyroclasts. Compared to the mineralogy of many two mica granites, that of the Whipple Wash suite of Anderson and Cullers (1990) is unusual in the siliceous nature of the plutonic muscovite and the calcic character of the garnet, both being indications of high pressure crystallization (Anderson, 1988). Primary muscovite, on a 11 oxygen formula basis, contains an average Si content of 3.16 ± .04 atoms leading to an average muscovite mole fraction of 0.58, with the remainder being mostly comprised of Mg-celadonite, ferrimuscovite and a Ti-end member. Garnet, which modally occurs upwards to 4.0 % in three of the plutons, is uniformly enriched in grossular (9 to 24 mole %) and spessartine (21 to 32 mole %) components.

Despite the abundance of relict igneous mineral phases, a second generation of mineral compositions (alkali feldspar, plagioclase, muscovite, biotite, amphibole) provide a basis for quantifying the conditions of the Tertiary mylonitization. Mineral assemblages, coupled with two-feldspar and plagioclase-amphibole thermometry demonstrate that the grade of metamorphism -- upper greenschist to lower amphibolite -- increases with depth over the estimated temperature range of 460 ± 35°C (upper structural levels) to 535 ± 44°C. The average estimated pressure attending mylonitization, 4.4 ± 1.1kb (~16 ± 4km), is demonstrably lower (shallower) than the pressures noted above for emplacement of the Cretaceous intrusives. This conclusion is consistent with the inferred upward rise of the crystalline section between 89 and 26 Ma (Anderson, 1985, 1988; Anderson and others, 1988).

Highlights on the Way to Lake Havasu City

Return to vehicles. Retrace the powerline road 5.0 mi and turn left (NE) 3.2 mi to Black Meadow Landing (Fig. 15). From here we will embark on a nautical tour of lower Lake Havasu which is nestled between upper-plate tilt blocks composed of Miocene and Proterozoic rocks. The giant Miocene landslide, seen this morning as megabreccia roadcrops on the southern outskirts of Lake Havasu City, crops out prominently in the Aubrey Hills on the Arizona side of the lake.

DAY 5

The purposes of Day 5 are to address Miocene dike intrusion and synextensional sedimentation in the Mohave Mountains, and visit the huge tilted Crossman block forming the high Mohave Mountains rising above Lake Havasu City.

Highlights on the Way to Stop 5.1

The high Mohave Mountains to the east represent the large upended Crossman tilt block of Howard and others (1982a). The Whipple Mountains and Chemehuevi detachment fault systems project beneath this allochthonous tilted block, which has been explained as an upended block exposing a crustal section measuring 15 km in original paleothickness (Howard and others, 1982a, 1990 in press; Howard and John, 1987). Because the block has been tilted 90°, a map of the block may be viewed down structure from the northeast as an early Miocene cross section (Fig. 18). Proterozoic gneiss that forms the bulk of the block is overlain by Miocene strata and volcanic rocks (now subvertical, facing SW) and cut by Miocene dikes (now NE-dipping). The time of tilting seems tightly constrained by overlapping K-Ar ages of 19-20 Ma obtained from the upended Miocene volcanic rocks, younger partly(?) tilted dikes, and overlapping gently dipping volcanic rocks (Nakata and others, 1990; Nielson, 1986; Nielson and others, 1993).

Cross east over London Bridge, make two immediate left turns then an immediate right turn onto Arizona 95, and proceed north 2.1 mi. Turn east (right) on Kiowa Blvd. for 4.2 mi, and turn east (left) onto Bison Blvd, continuing past the end of the pavement. Red Hills on the left (north) are made of Miocene rhyolite dipping steeply SW. East beyond Kiowa 1.6 mi, the approximate position of the subvertical, SW-facing Tertiary-Proterozoic nonconformity is marked by Tertiary andesite exposed in the gully on the left (N), and Proterozoic gneiss in the red hill ahead on the left. The nonconformity here is totally obscured by Miocene silicic intrusions. Elsewhere along this contact, basal Tertiary tuffs and locally derived arkose are vertical or locally overturned, and define the rotation of the Crossman block.

One mile ahead to the east in the Mohave Mountains can be seen a mass of light-colored hills formed by sheeted Tertiary dikes. Individual dikes cutting Proterozoic gneiss on the high ridge of the Mohave Mountains beyond are best seen in afternoon light.

Beyond Kiowa 2.0 miles veer left (NE) down into Fall Springs Wash and follow the main road east up it 0.9 mi. Light-colored Miocene dacite dikes and darker mafic dikes along the wash intrude gray Early Proterozoic granodiorite augen gneiss. Similar gneiss in the Bill Williams Mountains was dated by U-Pb on zircon as 1.64 Ga (Wooden and Miller, 1990). Dikes in the dike swarm have yielded biotite K-Ar and $^{40}Ar/^{39}Ar$ ages of 18-21 Ma; K-Ar ages on hornblende, plagioclase, and whole rock range from 11 to 78 Ma and are not considered as reliable intrusive ages (Nakata and others, 1990; Pease, 1991). Felsic dikes average 6 m thick and mafic dikes ~3 m thick (Nakata, 1982).

Stop 5.1

Mohave Mountains dike swarm -- K. Howard

A dike beside the road within the sheeted dikes here exhibits mullion structures that are likely primary flow structures. They plunge gently now, but would restore to a steep plunge if the Crossman block were rotated about a NW axis to its pre-Miocene orientation. Nakata (1982) measured 15-20 volume percent of dikes across the whole Crossman block, and found that most dikes dip moderately NE. This orientation and intrusive relations into the tilted Tertiary section suggest that most of the dike swarm may be less rotated than the vertical Tertiary strata (and the gneisses), and that intrusion, if as originally vertical dikes, occurred while the block was undergoing its rotation. Paleomagnetic measurements of the dikes suggest that they acquired their magnetization after the Crossman block had been partly tilted but before completion of tilting (Pease, 1991; Nielson and others, 1993).

Stream boulders include Proterozoic spotted leucocratic gneiss with granulitic texture and pseudomorphs of retrograded garnet, one of the most distinctive rocks of the Mohave Mountains. It dominates lower parts of the crustal section exposed in the Crossman block (Howard and others, 1990), is also found in the Bill Williams Mountains tilt block to the south, and occurs in beheaded lower-plate position 50 km to the southwest in western parts of the Whipple Mountains core complex.

Gneiss and amphibolite in the Crossman block commonly strike NE and dip steeply. Amphibolite bodies define the limbs and nose of a SW-plunging isoclinal antiform that stretches across much of the width of the Crossman block (Fig. 18). When restored before block tilting, the fold initially plunges NE and conforms to Proterozoic foliation trends elsewhere in the region that are typically NE-SW.

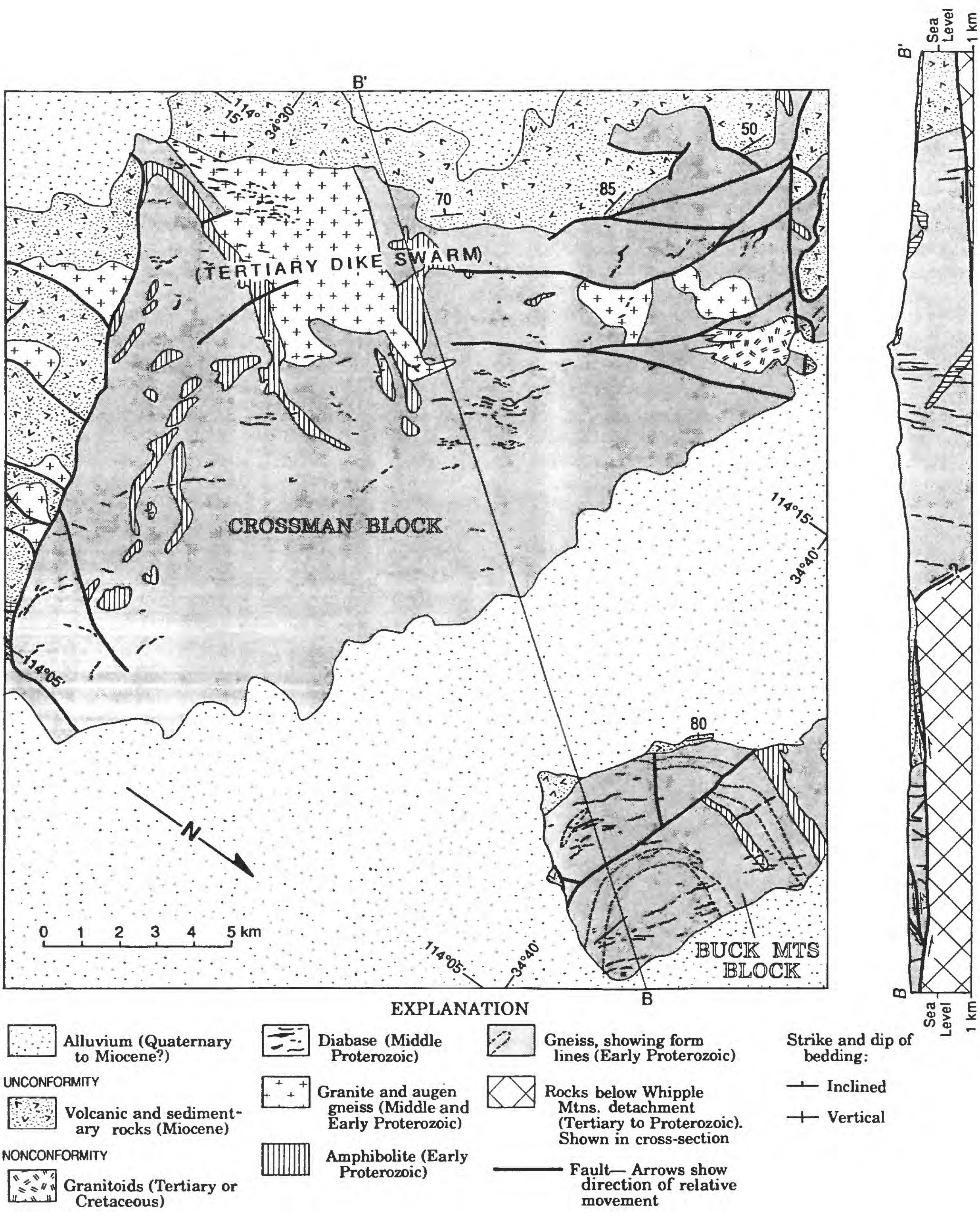


Figure 18 Geologic map of the Crossman block of the Mohave Mountains, viewed from the northeast (Howard, 1991). The eastern outskirts of Lake Havasu City lie near B'. Capping lower Miocene volcanic units in both the Crossman block and the smaller Buck Mountains block dip subvertically and face southwest, the direction of map view. Consequently, the blocks can be viewed on the map as approximate paleo-cross sections, making allowances for faulting that distends each block and Tertiary dikes (not shown) that inflate each. Subvertical Proterozoic diabase sheets restore to subhorizontal orientations through pre-Miocene structural thicknesses of about 4 km in the Buck Mountains block and 11 km in the Crossman block. Younger, less tilted Miocene rocks lap across more steeply tilted rocks.

Highlights between Stops 5.1 and 5.2

Return back 2.9 mi along Fall Springs Wash and Bison Blvd., and right on Kiowa Blvd 4.2 mi to Arizona 95. Turn north (right) on Arizona highway 95. As you drive north, at 3 o'clock, the structurally thick (15 km) Crossman tilt block forms the high Mohave Mountains, above the regionally developed detachment fault system. The main mass of the Chemehuevi Mountains core complex, the footwall to the detachment fault system, lies at 10 o'clock. Continue on Arizona 95 for 5 miles to the beginning of a right turn in Arizona 95. Turn left (west) on a dirt road near mile post 191, proceed 0.2 mile to a broad wash, and park. Walk north (left) 0.3 mile along a dirt road to its end, at the base of a large hill surmounted by a radio tower ("Shangri La" hill).

Stop 5.2

Northern Mohave Mountains -- B. John and K. Howard

The purpose of this stop is to examine evidence for an unconformity and paleochannel in the synextensional Miocene section, and to gain an overview of the detachment terrane expressed by the Chemehuevi, Whipple, and Mohave Mountains. The dirt road exposes gently dipping Miocene conglomerate that contains clasts of the 18.5-Ma Peach Springs Tuff, and here directly overlies tuff and basalt that pre-date the Peach Springs Tuff. The end of the road at the base of a steep hill marks the southern wall of a deep (~90 ft or 27 m) paleochannel cut through the tilted Peach Springs Tuff. The channel is host to large boulders of the tuff (up to 6 m in greatest dimension) and rare gneiss clasts. This is one of several east-west channels cut into and locally through the Peach Springs Tuff, which aided deposition in the Blankenship basin to the north and west as tilting proceeded.

The hill 0.1 mi the the north, topped by a radio facility, is in this conglomerate near the channel wall against the west-dipping Peach Springs Tuff. The tuff in turn forms the more northern and higher of the two radio-facility hills. The tuff exhibits fiamme, large pumice blocks, and adularescent blue sanidine. It dips 10-15° steeper than does the overlying conglomerate.

Views can be had of the detachment terrane in the Mohave, Chemehuevi, and Whipple Mountains. Hogbacks to the N represent a series of tilted slices occupying a synformal trough in the upper plate of the Powell Peak detachment fault. That fault dips 35° W toward us at Powell Peak to the N. The NE-trending trough is 4 km wide between Powell Peak and the high Crossman block, to the SE, and appears to be structurally higher than that block. The upper-plate tilt slices in this trough face SW and are shingled on NE-dipping faults. Several of the tilt blocks exhibit asymmetric synclines (Nielson, 1986) with short SW limbs due to fault drag and long NE limbs reflecting "reverse drag" (block tilt). Sub-Tertiary basement rocks above and below the Powell Peak detachment are unmylonitized Proterozoic rocks cut by Cretaceous granite and granodiorite displaced from the Chemehuevi Mountains lower plate.

From this vantage point we can also see and discuss the geometry of drape folds seen from Topock Gorge on Day 3 (John and Howard, 1994).

Highlights between Stop 5.2 and Las Vegas

Continue driving on Arizona 95 north for 10 mi to highway I-40. As you begin, prominent pinnacles on the E (right) side are remnants of a rhyolite dike dated by K-Ar (plagioclase) as 15.1±0.4 Ma (Nakata and others, 1990). Roadcuts at first are in sub-Peach Springs Tuff volcanic rocks, and finally in broken Proterozoic gneiss, all in allochthons above the Chemehuevi detachment fault. In 5 mi, volcanic rocks 1/2 mi to the W (left) of the road dip steeply SW. The most prominent unit is olivine basalt that yielded a K-Ar whole-rock age of 14.1 Ma (Nakata and others, 1990). If this is the extrusive age, it records one of the youngest indications of steep tilting in this part of the extensional corridor. Elsewhere south of Nevada, much of the extensional tilting was completed before 14 Ma (Spencer, 1985; Simpson and others, 1991). To the northeast, the high distant Hualapai Mountains border the Colorado River extensional corridor on its east side.

Turn W (left) on I-40 toward Needles. The road passes mudstones of the Bouse Formation, representing a late Miocene and Pliocene estuarine proto-Gulf of California (Buising, 1993). To the N (right), Miocene volcanic rocks in the Black Mountains lead toward the Oatman district. To the SW are the Needles made of tilted Miocene volcanic and hypabyssal rocks above the Chemehuevi and Devils Elbow faults. The closest of these pinnacles is an intrusive body that occupies a synformal gap below volcanic rocks draped over a fault boundary between two tilted blocks of basement rocks (John and Howard, 1994).

At 9.6 mi, cross the Colorado River. Retrace I-40 11 mi to Needles and continue another 11 mi. At the Searchlight exit turn north on U.S. 95 toward Las Vegas.

For those leaving Las Vegas by air to central California, the following aerial highlights may be useful.

Aerial highlights between Las Vegas and San Francisco

The southwestern United States are generally cloud free in June and geologic features are easily spotted from the air. Large-displacement Mesozoic thrusts of the Sevier orogenic belt are prominent in the Spring Mountains that overlook Las Vegas on the west. Look at the east base of the range for red Jurassic cliff-forming sandstone, which is structurally overlain by gray Paleozoic limestones of the miogeocline. The route proceeds past two more desert ranges to Death Valley (Fig. 1), the lowest point in North America and site of currently active right-oblique extension including the development of core complexes known as turtlebacks (on the east side of the valley in the Black Mountains). Telescope Peak that overlooks Death Valley from the west exceeds 11,000 ft (3353 m) in elevation. Proceeding northwest, the last big range before reaching the Sierra Nevada is the White-Inyo range, nearly as high as the Sierra Nevada with peaks over 14,000 ft (4267 m) in elevation. At the west foot of the Sierra Nevada, numerous Quaternary rhyolite domes and flows can be spotted between the 15-km-wide circular Mono Lake, and the Long Valley caldera to the south with its resurgent dome and its outflow sheet, the 0.7-Ma Bishop Tuff. Beyond rises the high crest of the heavily glaciated Sierra Nevada, a huge (100 by 500 km) gently west-tilted block of Mesozoic batholithic rocks. The high cirques host a few small glaciers. The forested western side of the Sierra Nevada slopes gradually westward to California's Great Valley. Beyond lie the Coast Ranges which expose a collage of accreted and displaced terranes. On the approach to San Francisco airport on the west side of San Francisco Bay, long and narrow linear lakes that may be spotted in a trough in the hills west of the Bay mark the position of the San Andreas fault.

REFERENCES CITED

Anderson, E.M., 1951, The dynamics of faulting and dyke formation: Edinburgh, Oliver and Boyd, 206 p.

Anderson, J. L., 1981, Conditions of mylonitization in a metamorphic core complex, Whipple Mountains, California: *in* Howard, K.A., Carr, M.D., and Miller, D.M., editors, Tectonic framework of the Mojave and Sonoran Deserts, California and Arizona: U.S. Geological Survey Open-file Report 81-503, p.1-3.

Anderson, J.L., 1983, Proterozoic anorogenic granite plutonism of North America: Geological Society of America Memoir 161, p. 133-154.

Anderson, J.L., 1985, Contrasting depths of "core complex" mylonitization: barometric evidence. Geological Society of America Abstracts with Programs, v.17, p.337.

Anderson, J.L., 1988, Core complexes of the Mojave-Sonoran Desert; Conditions of plutonism, mylonitization, and decompression, *in* Ernst, W.G., editor, Metamorphism and crustal evolution of the western United States (Rubey Volume 7): Englewood Cliffs, N.J., Prentice-Hall, p. 502-525

Anderson, J.L., and Bender, E. E., 1989, Nature and origin of Proterozoic A-type granitic magmatism in the southwestern United States: Lithos, v. 23, p. 19-52.

Anderson J.L. and Cullers, R.L., 1990, Middle to upper crustal plutonic construction of a magmatic arc: An example from the Whipple Mountains metamorphic core complex: Geological Society of America Memoir 174, p. 47-69.

Anderson, J.L., and Rowley, M.C., 1981, Synkinematic intrusion of two-mica and associated metaluminous granitoids, Whipple Mountains, California. Canadian Mineralogist, v.19, p.83-101.

Anderson, J.L., Barth, A. P., and Young, E. D., 1988, Mid-crustal Cretaceous roots of Cordilleran metamorphic core complexes: Geology, v.16, p. 366-369.

Anderson, J.L., Davis, G. A., and Frost, E. G.,1979, Field guide to regional Miocene detachment faulting and early Tertiary (?) mylonitization, Whipple-Buckskin-Rawhide Mountains, southeastern California and western Arizona: *in* Abbott, P. L., editor, Geological Excursions in the Southern California Area: Geological Society of America Guidebook, p. 109-133.

Anderson R.E., 1969, Notes on the geology and paleohydrology of the Boulder City pluton, southern Nevada: U.S. Geological Survey Professional Paper 650-B, p. 35-40.

Anderson, R.E., 1971, Thin-skinned distension in Tertiary rocks of southeastern Nevada: Geological Society of America Bulletin, v. 82, p. 43-58.

Anderson, RE., 1978, Geologic map of the Black Canyon 15-minute quadrangle, Mohave County, Arizona and Clark County, Nevada: U.S. Geological Survey Map GQ-1394, scale 1:62,500.

Anderson R.E., Longwell, C.R., Armstrong, R.L., and Marvin, R.F., 1972, Significance of K-Ar ages of Tertiary rocks from the Lake Mead region, Nevada-Arizona: Geological Society of America Bulletin, v. 83, p.273-288.

Armstrong, R.L., 1982, Cordilleran metamorphic core complexes: Annual Reviews of Earth Planetary Science, v. 10, p. 129-154.

Armstrong, R.L. and Ward, P., 1991, Evolving geographic patterns of Cenozoic magmatism in the North American Cordillera: The temporal and spatial association of magmatism and metamorphic core complexes: Journal of Geophysical Research, v. 96, p. 13,201-13,224.

Bahti, Tom, 1968, Southwestern Indian tribes: Las Vegas, Nevada, KC Publications, Fifth printing, 1975, 72 p.

Bartley, J.M. and Glazner, A.F., 1985, Hydrothermal systems and Tertiary low-angle normal faulting in the southwestern United States: Geology, v. 13, p. 562-564.

Bennett, V.C. and DePaolo, D.J., 1987, Proterozoic crustal history of the western United States as determined by neodymium isotopic mapping: Geological Society of America Bulletin, v. 99, p. 674-685.

Best , M.G. and Christiansen, E.H., 1991, Limited extension during peak Tertiary volcanism, Great Basin of Nevada and Utah: Journal of Geophysical Research, v. 96, p. 13,509-13,528.

Block, L., and Royden, L.H., 1990, Core complex geometries and regional scale flow in the lower crust: Tectonics, v. 9, p. 557-567.

Bradshaw, T.K., Hawkesworth, C.J., and Gallagher, K., 1993, Basaltic volcanism in the southern Basin and Range: no role for a mantle plume: Earth and Planetary Science Letters, v. 116, p. 45-62.

Brooks,W.E., 1986, Distribution of anomalously high K_20 volcanic rocks in Arizona: Metasomatism at the Picacho Peak detachment fault: Geology, v. 14, p. 339-342.

Brun, J-P, Sokoutis, Dimitrios, and Van Den Driessche, Jean, 1994, Analogue modeling of detachment fault systems and core complexes: Geology, v. 22, p. 319-322.

Bryant, B., Naeser, C.W., and Fryxell, J.E., 1991, Implications of low-temperature cooling history on a transect across the Colorado Plateau-Basin and Range boundary, west-central Arizona: Journal of Geophysical Research, v. 96, p. 12,375-12,388.

Bryant, B., and Wooden, J. L.,1989, Lower-plate rocks of the Buckskin Mountains, Arizona: a progress report, *in* Geology and mineral resources of the Buckskin and Rawhide Mountains, west-central Arizona: Arizona Geological Survey Bulletin, v. 198, p. 47-50.

Buck, W.R., 1988, Flexural rotation of normal faults: Tectonics, v. 959-973.

Buising, A.V., 1993, Stratigraphic overview of the Bouse Formation and gravels of the Colorado River, Riverside Mountains, eastern Vidal Valley, and Mesquite Mountain areas, Calif. and Ariz.: U.S. Geological Survey Bulletin 2053, p. 165-170.

Burchfiel, B.C. and Davis, G.A., 1988, Mesozoic thrust faults and Cenozoic low-angle normal faults, eastern Spring Mountains, Nevada, and Clark Mountains thrust complex, California, *in* Weide, D.L. and Faber, M.L., editors, This extended land, geological journeys in the southern Basin and Range: Geological Society of America, Cordilleran Section, Field trip guidebook p. 87- 106.

Burchfiel, B.C., Fleck, R.J., Secor, D.T., Vincellete, R.R., and Davis, G.A., 1974, Geology of the Spring Mountains, Nevada: Geological Society of America Bulletin, v. 85, p. 1013-1022.

Campbell, E. and John, B.E., 1994, Crustal extension and the peak Colorado River gravity high, southern Sacramento Mountains, California: A preliminary correlation: U.S. Geological Survey Open-file Report 94-236, 16 p.

Carl, B.S., Miller, C.F., and Foster, D.A., 1991, Western Old Woman Mountains shear zone: evidence for late ductile extension in the Cordilleran belt: Geology, v. 19, p. 893-896.

Carr, W.J., 1991, A contribution to the structural history of the Vidal-Parker region, California and Arizona: U.S. Geological Survey Professional Paper 1430, 40 p.

Chadwick, W.W. and Howard, K.A.,1990, The pattern of circumferential and radial eruptive fissures on the volcanoes of Fernandina and Isabela islands, Galapagos: Bulletin of Volcanology, v. 53, p. 259-275.

Chapin, C.E. and Lindley, J.I., 1986, Potassium metasomatism of igneous and sedimentary rocks in detachment terranes and other sedimentary basins: Economic implications, *in* Betty, B. and Wilkinson,, P.A.K., editors, Frontiers in geology and ore deposits of Arizona and the Southwest: Arizona Geological Society Digest, v. 16,p. 118-126.

Coney, P. J., 1980, Cordilleran metamorphic core complexes: an overview, *in* Crittenden, M. D., Jr. and others, editors, Cordilleran metamorphic core complexes: Geological Society of America Memoir 153, p. 7-34.

Coney, P.J and T.A. Harms, 1984, Cordilleran metamorphic core complexes: Cenozoic extensional relics of Mesozoic compression: Geology,v. 12, p. 550-554.

Conners, Jo, 1913, Who's who in Arizona: Tucson, Arizona, Press of the Arizona Daily Star, 820 p.

Darton, N.H., 1916, Guidebook of the western United States, Part C., the Santa Fe route: U.S. Geological Survey Bulletin 613, 200 p.

Davis, G. A., 1988, Rapid upward transport of mid-crustal mylonitic gneisses in the footwall of a Miocene detachment fault, Whipple Mountains, southeastern California. Geologische Rundschau, v. 77, p. 191-209.

Davis, G.A., Anderson, J.L., Frost, E.G., and Shackelford, T.J., 1980, Mylonitization and detachment faulting in the Whipple-Buckskin-Rawhide Mountains terrane, southeastern California and western Arizona: Geological Society of America Memoir 153, p. 79-129.

Davis, G.A., Anderson, J.L., Martin, D.L., Krummenacher, D., Frost, E.G., and Armstrong, R.L., 1982, Geologic and geochronologic relations in the lower plate of the Whipple detachment fault, Whipple Mountains, southeastern California; A progress report, *in* Frost, E.G. and D.L. Martin, editors, Mesozoic-Cenozoic tectonic evolution of the Colorado River region, California, Arizona, and Nevada (Anderson-Hamilton volume), San Diego, Calif., Cordilleran Publishers, p. 408-432.

Davis, G.A. and Lister, G.S., 1988, Detachment faulting in continental extension; Perspectives from the southwestern U,S, Cordillera: Geological Society of America Special Paper 218, p. 133-159.

Davis, G. A., Lister, G. S., and Reynolds, S. J., 1983, Interpretation of Cordilleran core complexes as evolving crustal shear zones in an extending orogen. Geological Society of America Abstracts with Programs, v. 15, p. 311.

Davis, G.A., Lister, G.S., and Reynolds, S.J., 1986, Structural evolution of the Whipple and South Mountains shear zones, southwestern United States. Geology, v. 14, p. 7-10.

Davis, G. H., 1983, Shear-zone model for the origin of metamorphic core complexes. Geology, v. 11, p. 342-347.

Davis, G. H., and Coney, P. J., 1979, Geologic development of the Cordilleran metamorphic core complexes. Geology, v. 7, p. 120-124.

DeWitt, E., Sutter, J.F., Davis, G.A., and Anderson, J.L.., 1986a, $^{40}Ar/^{39}Ar$ age-spectrum dating of Miocene mylonitic rocks, Whipple Mountains southeastern California. Geological Society of America Abstracts with Programs, v. 18, p. 584.

DeWitt, Ed, Thorson, J.P., and Smith, R.C., 1986b, Geology and gold deposits of the Oatman District, northwestern Arizona: U.S. Geological Survey Open-file Report 86-0638, 34 p.

Dickinson, W.R. and Snyder, W.S., 1979, Geometry of subducted slabs related to San Andreas transform: Journal of Geology, v. 87, p. 609-627.

Dickey, D.D., Carr, W.J., and Bull, W.B., 1980, Geologic map of the Parker NW, Parker, and parts of the Whipple Mountains SW and Whipple Wash quadrangles, California and Arizona: U.S. Geological Survey Miscellaneous Investigations Map I-1124, scale 1:24,000.

Dokka, R.K., 1993, Original dip and subsequent modification of a Cordilleran detachment fault, Mojave extensional belt, California: Geology, v. 21, p. 711-714.

Dokka, R. K., and Lingrey, S. H., 1979, Fission track evidence for a Miocene cooling event, Whipple Mountains, southeastern California, in Cenozoic paleogeography of the western United States (J. L. Armentrout and others, editors). Pacific Section, Society of Economic Paleontologists and Mineralogists, Los Angeles, p. 141-146.

Dunn, J. F., 1986, The structural geology of the northeastern Whipple Mountains detachment fault terrane, San Bernardino County, California. M. S. thesis, University of Southern California, Los Angeles, 172 p.

Elston, D.B. and Young, R.A., 1991, Cretaceous-Eocene (Laramide) landscape development and Oligocene-Pliocene drainage reorganization of Transition Zone and Colorado Plateau, Arizona: Journal of Geophysical Research, v. 96, p. 12,389-12,406.

Faulds, J.E., Geissman, J.W., and Mawer, C.K., 1990, Structural development of a major extensional accommodation zone in the Basin and Range Province, northwestern Arizona and southern Nevada; Implications for kinematic models of continental extension: Geological Society of America Memoir 176, p. 37-76.

Faulds, J.E., Geissman, J.W., and Shafiqullah, Mohammad, 1992, Implications of paleomagnetic data on Miocene extension near a major accommodation zone in the Basin and Range province, northwestern Arizona and southern Nevada: Tectonics, v. 11, p. 204-228.

Feuerbach, D.L., Smith, E.I., Walker, D.D., and Tangeman, J.A., 1993, The role of the mantle during crustal extension; Constraints from geochemistry of volcanic rocks in the Lake Mead area, Nevada and Arizona: Geological Society of America Bulletin, v. 105, p. 1561-1575.

Fitzgerald, P.G., Fryxell, J.E., and Wernicke, B.P., 1991, Miocene crustal extension and uplift in southeastern Nevada: Constraints from fission track analysis: Geology, v. 19, p. 1013-1016.

Fletcher, J.M. and Karlstrom, K.E., 1990, Late Cretaceous ductile deformation, metamorphism, and plutonism in the Piute Mountains, eastern Mojave Desert: Journal of Geophysical Research, v. 95, p. 487-500.

Foster, D.A., Gleadow, A.J.W., Reynolds, S.J., and Fitzgerald, P.G., 1993, The denudation of metamorphic core complexes and the reconstruction of the Transition Zone, west-central Arizona: Thermochronologic constraints: Journal of Geophysical Research, v. 98, p. 2167-2186.

Foster, D.A., Harrison, T.M., Miller, C.F., and Howard, K.A., 1990, The $^{40}Ar/^{39}Ar$ thermochronology of the eastern Mojave Desert, California, and adjacent western Arizona with implications for the evolution of metamorphic core complexes: Journal of Geophysical Research, v. 95, p. 20,005-20,024.

Foster, D.A., Miller, C.F., Harrison, T.M., and Hoisch, T.D., 1992, $^{40}Ar/^{39}Ar$ thermochronology and thermobarometry of metamorphism, plutonism, and tectonic denudation in the Old Woman Mountains area, California: Geological Society of America Bulletin, v. 104, p. 176-191.

Foster, D.A., Miller, D.S., and Miller, C.F., 1991, Tertiary extension in the Old Woman Mountains area, California: Evidence from apatite fission track analysis: Tectonics, v. 10, p. 875-886.

Foster, D.A., and Spencer, J.E., 1992, Apatite and zircon fission-track dates from the northern Plomosa Mountains, La Paz County, west-central Arizona: Arizona Geological Survey Open-file Report 9209, 11 p.

Frost, E. G., 1980, Whipple Wash Pump Storage Project, California: appraisal design data, structural geology and water-holding capacity of rocks in the Whipple Wash area, San Bernardino County, California. Unpublished report submitted to U. S. Water and Power Resources Service, Boulder City, Nevada.

Frost, E.G. and Martin, D.L., 1982, Comparison of Mesozoic compressional tectonics with mid-Tertiary detachment faulting in the Colorado River area, California, Arizona, and Nevada, *in* Cooper, J.D., compiler, Geologic excursions in the California Desert: Geological Society of America Cordilleran Section, 78th Annual Meeting, Anaheim, California, April 19-21, 1982, p. 113-159.

Gans, P.B., 1987, An open-system, two-layer crustal stretching model for the eastern Great Basin: Tectonics, v. 6, p. 1-12.

Gans, P.B., Darvall, P., and Faulds, J., 1992, Miocene extension in the Colorado River extensional corridor; A comparison with slow-spreading ridges: Geological Society of America Abstracts with programs, v. 24, n. 5, p. 25.

Gans, P.B., Landau, B., and Darvall, P., 1994, Ashes, ashes, all fall down: Caldera-forming eruptions and extensional collapse of the Eldorado Mountains, southern Nevada: Geological Society of America Abstracts with Programs, v. 26, no. 2, p. 53.

Gans, P.B., Mahood, G.A., and Schermer, E., 1989, Synextensional magmatism in the Basin and Range province; A case study from the eastern Great Basin: Geological Society of America Special Paper 233, 58 p.

Glazner, A.F. and Bartley, J.M., 1984, Timing and tectonic setting of Tertiary low-angle normal faulting and associated magmatism in the southwestern United Tates: Tectonics, v. 3, p. 385-396.

Glazner, A.F. and Supplee, J.A., 1982, Migration of Tertiary volcanism in the southwestern United States and subduction of the Mendocino fracture zone: Earth and Planetary Science Letters, v. 60, p. 429-436.

Gray, F., Jachens, R.C., Miller, R.J., Turner, R.L., Knepper, D.H., Jr., Pitkin, J.A., Keith, W.J., Mariano, J., and Jones, S.L., 1990, Mineral resources of the Warm Springs Wilderness Study Area, Mohave County, Arizona: U.S. Geological Survey Bulletin 1737, 24 p.

Gross, W. W. and Hillmeyer, F. L., 1982, Geometric analysis of upper plate fault patterns in the Whipple-Buckskin detachment terrane: in Frost, E. G. and Martin, D. L., editors, Mesozoic-Cenozoic Tectonic Evolution of the Colorado River Region, California, Arizona, and Nevada, Cordilleran Publishers, San Diego, p. 256-266.

Hamilton, W. and Myers, W.B., 1966, Cenozoic tectonics of the western United States:Reviews of Geophysics,v. 4, p. 509-549.

Hamilton, W.B. and Howard, K.A., 1991, Detachment faults -- Steep active sectors are bounded by migrating hinges at both tops and bottoms: Geological Society of America Abstracts with programs, v. 23, n. 5, p. A189-190.

Hamilton, W.B., Tosdal, R.M., Stone, P., and Haxel, G.B., 1987, Mesozoic tectonics of southeastern California, *in* Davis, G.H., and VandenDolder, E.M., editors, Geologic diversity of Arizona and its margins: Excursions to choice areas: Arizona Bureau of Geology and Mineral Technology, Special Paper 5, p. 337-350.

Hammond, J.G., 1990, Middle Proterozoic diabase intrusions in the southwestern U.S.A. as indicators of limited extensional tectonism, *in* Gower, C.F., Rivers, T. , and Ryan, B., editors, Mid-Proterozoic Laurentia-Baltica, Geological Association of Canada Special Paper 38, p. 517-531.

Hillhouse, J.W., and Wells, R.E., 1991, Magnetic fabric, flow directions, and source area of the lower Miocene Peach Springs Tuff in Arizona, California, and Nevada: Journal of Geophysical Research, v. 96, p. 12,443-12,460.

Hopson, C.A., Gans, P.B., Baer, E., Blythe, A., Calvert, A., and Pinnow, J., 1994, Spirit Mountain pluton, southern Nevada: A progress report: Geological Society of America Abstracts with Programs, v. 26, n. 2, p. 60.

Howard, K.A., 1991, Intrusion of horizontal dikes: Tectonic significance of Middle Proterozoic diabase sheets widespread in the upper crust of the southwestern United States: Journal of Geophysical Research, v. 96, p. 12, 461-12, 478.

Howard, K.A., 1993, Volcanic history of the Colorado River extensional corridor: Active or passive rifting?: Geological Society of America Abstracts with programs, v. 25, n. 5, p. 54.

Howard, K.A., Goodge,J.W., and John, B.E., 1982a, Detached crystalline rocks of the Mohave, Buck, and Bill Williams Mountains, *in* Frost, E.G. and D.L. Martin, editors, Mesozoic-Cenozoic tectonic evolution of the Colorado River region, California, Arizona, and Nevada (Anderson-Hamilton volume), San Diego, Calif., Cordilleran Publishers, 377-390, 1982.

Howard, K.A. and John, B.E., 1987, Crustal extension along a rooted system of imbricate low-angle faults, Colorado River extensional corridor, California and Arizona: *in* Coward, M.P., Dewey, J.F., and Hancock, P.L., editors, Continental extensional tectonics: Geological Society of London Special Publication 28, p. 299-311.

Howard, K.A. and John, B.E., 1991, Disharmonic drape folds associated with upper-crustal extension in the southern Basin and Range: Geological Society of America Abstracts with Programs, v. 23, no. 2, p. 36.

Howard, K.A., John, B.E., and Miller, C.F., 1987, Metamorphic core complexes, ductile thrusts, and Cenozoic detachments: Old Woman Mountains-Chemehuevi Mountains transect, California and Arizona: Arizona Bureau of Geology and Mineral Technology Special Paper 5, p. 365-382.

Howard, K.A., Nielson, J.E., Wilshire, H.G., Nakata, J.K., Goodge, J.W., Reneau, S.L., John, B.E., and Hansen, V.L., 1990, Preliminary geologic map of the Mojave Mountains area, Mojave County, western Arizona: U.S. Geological Survey Open-file report 90-684, scale 1:48,000.

Howard, K.A., Nielson, J.E., Wilshire, H.G., Nakata, J.K., Goodge, J.W., Reneau, S.L., John, B.E., and Hansen, V.L., in press, Preliminary geologic map of the Mojave Mountains area, Mojave County, western Arizona: U.S. Geological Survey Miscellaneous Investigations Map I-2308, scale 1:48,000.

Howard, K.A., Stone, P., Pernokas, M.A., and Marvin, R.F., 1982b, Geologic and geochronologic reconnaissance of the Turtle Mountains area, California: west border of the Whipple detachment terrane, in Frost, E.G. and Martin, D.L., editors, Mesozoic-Cenozoic tectonic evolution of the Colorado River region, California, Arizona, and Nevada: San Diego, Cordilleran Publishers, p. 341-354.

Ives, Lt. J.C., 1861, Report upon the Colorado River of the West: Washington, 36th Congress 1st session, Senate, Government Printing Office, 154 p.

Jackson, J., 1987, Active normal faulting and crustal extension, in Continental extensional tectonics *in* Coward, M.P., Dewey, J.F., and Hancock, P.L., eds, Continental extensional tectonics: Geol. Society of London Special. Publication 28, p. 3-17.

John, B.E., 1986, Structural and intrusive history of the Chemehuevi Mountains area, southeastern California and western Arizona: Santa Barbara, California, PhD. thesis, University of California at Santa Barbara, 295 p.

John, B.E., 1987a, Geometry and evolution of a mid-crustal extensional fault system: Chemehuevi Mountains, southeastern California, *in* Coward, M.P., Dewey, J.F., and Hancock, P.L., eds, Continental extensional tectonics: Geological Society of London Special Paper. 28, p. 313-335.

John, B.E., 1987b, Geologic map of the Chemehuevi Mountains area, San Bernardino County, California, and Mohave County, Arizona: U.S. Geological Survey Open-File Report 87-666, (1:24, 000 scale, with explanation).

John, B.E., 1988, Structural reconstruction and zonation of a tilted mid-crustal magma chamber: The felsic Chemehuevi Mountains Plutonic Suite: Geology, v. 16, p. 613-617.

John, B.E., and Foster, D.A., 1993, Structural and thermal constraints on the initiation angle of detachment faulting in the southern Basin Range: The Chemehuevi Mountains case study: Geological Society of America Bulletin, v. 105, p. 1091-1108.

John, B.E. and Howard, K.A., 1994, Drape folds in the highly attenuated Colorado River extensional corridor, California and Arizona, *in* McGill, S.F. and Ross, T.M., editors, Geological investigations of an active margin: Geological Society of America Cordilleran Section Guidebook, 27th Annual Meeting, San Bernardino, California, March 21-23, 1994: San Bernardino County Museum Association, p. 94-106.

John, B.E., and Mukasa, S.B., 1990, Footwall rocks to the mid-Tertiary Chemehuevi detachment fault: a window into the Late Cretaceous middle crust: Journal of Geophysical Research, vol. 95, p. 463-485.

John, B.E., and Wooden, J.L., 1990, Petrology and geochemistry of the metaluminous to peraluminous Chemehuevi Mountains Plutonic Suite, southeastern California: Geological Society of America Memoir 174, p. 71-98.

Kemnitzer, L. E., 1937, Structural studies in the Whipple Mountains, southeastern California: Ph.D. dissertation, California Institute of Technology, Pasadena, 150 p.

Knapp, J.H. and Heizler, M.T., 1990, Thermal history of crystalline nappes of the Maria fold and thrust belt, west central Arizona: Journal of Geophysical Research, v. 95, p. 20,049-20,073.

Lee, W.T., 1908, Geologic reconnaissance of a part of western Arizona: U.S. Geological Survey Bulletin 252, p. 41-45.

Lister, G., and Baldwin, 1993, Plutonism and the origin of metamorphic core complexes: Geology, vol. 21, p. 607-610.

Lister, G. S., and Davis, G. A., 1989, The origin of metamorphic core complexes and detachment faults formed during Tertiary continental extension in the northern Colorado River region, U.S.A.: Journal of Structural Geology, v. 11, p. 65-94.

Longwell, C.R., 1936, Geology of the Boulder Reservoir floor, Arizona-Nevada: Geological Society of America Bulletin, v. 47, p. 1393-1476.

Mathis, R.S., 1982, Mid-Tertiary detachment faulting in the southeastern Newberry Mountains, Clark County, Nevada, *in* Frost, E.G. and D.L. Martin, editors, Mesozoic-Cenozoic tectonic evolution of the Colorado River region, California, Arizona, and Nevada (Anderson-Hamilton volume): San Diego, Calif., Cordilleran Publishers, p. 326-340.

McCarthy, Jill, Larkin, S.P., Fuis, G.S., Simpson, R.W., and Howard, K.A., 1991, Anatomy of a metamorphic core complex: Seismic refraction/wide-angle reflection profiling in southeastern California and western Arizona: Journal of Geophysical Research, v. 96, p. 12,259-12,292.

McKenzie, D, 1978, Some remarks on the development of sedimentary basins: Earth and Planetary Science Letters, vb.40, p. 25-32.

Metcalf, R.V., Smith, E.I., and Mills, J.G., 1993, Magma mixing and commingling in the northern Colorado River extensional corridor; Constraints on the production of intermediate magmas--Part I., *in* Lahren, M.M., Trexler, J.H., Jr., and Spinosa, C., editors Crustal evolution of the Great Basin and Sierra Nevada, Geological Society of America Cordilleran/Rocky Mountains Section Guidebook: Reno, Nevada, University of Nevada Department of Geological Sciences, p. 33-55.

Mickus, K.L. and James, W.C., 1991, Regional gravity studies in southeastern California, western Arizona, and southern Nevada: Journal of Geophysical Research, v. 96, p. 12,333-12,350.

Miller, C.F., Howard,K.A., and Hoisch,T.D., 1982, Mesozoic thrusting, metamorphism, and plutonism, Old Woman-Piute Range, southeastern California, *in* Frost, E.G. and D.L. Martin, editors, Mesozoic-Cenozoic tectonic evolution of the Colorado River region, California, Arizona, and Nevada (Anderson-Hamilton volume), San Diego, Calif., Cordilleran Publishers, p. 561-581.

Miller, J.M.G., and John, B.E., 1988, Detached strata in a Tertiary low-angle normal fault terrane, southeastern California: A sedimentary record of unroofing, breaching and continued slip: Geology, v. 16, p. 645-648.

Morrison, J., 1994, Meteoric water-rock exchange in the lower plate of the Whipple Mountain metamorphic core complex, California: Journal of Metamorphic Geology, v. 12 (in press).

Morrison, J., Anderson, J.L., and Davis, J.L., 1990, Detachment-related fluid movements in the Whipple Mountain metamorphic core complex of SE California. Geological Society of America Abstracts with Programs, v. 22, p. A94.

Muller, O.H. and Pollard, D.D., 1977, The stress state near Spanish Peaks, Colorado, determined from a dike pattern: Pure Applied Geophysics, v. 115, p. 69-86.

Nakata, J.K., 1982, Preliminary report on diking events in the Mohave Mountains, Arizona, *in* Frost, E.G. and Martin, D.L., editors, Mesozoic-Cenozoic tectonic evolution of the Colorado River region, California, Arizona, and Nevada (Anderson-Hamilton volume), San Diego, Calif., Cordilleran Publishers, p. 85-89.

Nakata, J.K., Pernokas, M.A., Howard, K.A., Nielson, J.E., and Shannon, J.R., 1990, K-Ar and fission-track ages (dates) of volcanic, intrusive, altered, and metamorphic rocks in the Mojave Mountains area, west-central Arizona: Isochron/West, n. 56, p. 8-20.

Nielson, J.E., 1986, Miocene stratigraphy of the Mojave Mountains, Arizona, and correlation with adjacent ranges, *in* Cenozoic stratigraphy, structure and mineralization in the Mojave Desert: Geological Society of America, Cordilleran Section, 82nd Ann. Mtg., Los Angeles, CA, March 25-28, 1986, Guidebook and Volume, Field Trip numbers 5 and 6, p. 15-24.

Nielson, J.E., 1993, Stratigraphic and structural correlation of Tertiary strata of the Mohave Mountains and Aubrey Hills, Ariz.: Geological Survey Bulletin 2053, p. 133-138.

Nielson, J.E., Lux, D.R., Dalrymple, G.B., and Glazner, 1990, A.F., Age of the Peach Springs Tuff, southeastern California and western Arizona: Journal of Geophysical Research, v. 95, p. 571-580.

Nielson, J.E., Pease, V.L., and Nakata, J.K., 1993, Paleomagnetic reversals in Miocene dikes and tectonic evolution of the Crossman block, Mohave Mountains, Arizona: Geological Society of America Abstracts with Programs, v. 25, no. 5, p. 127.

Odé, H., 1957, Mechanical analysis of the dike pattern of the Spanish Peaks area, Colorado: Geological Society America Bulletin, v. 68, p. 567-576.

Orrell, S. E., and Anderson, J. L., 1987, Proterozoic metamorphism in the Whipple Mountains, California. Geological Society of America Abstracts with Programs, v.19, p. 438.

Parsons, T. and Thompson, G.A., 1991, The role of magma overpressure in suppressing earthquakes and topography: World-wide examples: Science, v. 253, p. 1399-1402.

Parsons, T. and Thompson, G.A., 1993, Does magmatism influence low-angle normal faulting?: Geology, v. 21, p. 247-250.

Pease, V.L., 1991, The Miocene dike swarm of the Mohave Mountains, Arizona: Petrography, geochemistry, geochronology, and paleomagnetism: San Jose, California, San Jose State University, M.S. thesis, 166 p.

Powell, J.W., 1895, Canyons of the Colorado: Flood and Vincint, republished as The exploration of the Colorado River and its canyons: New York, Dover Publications, 1961400 p.

Proffett, J.M., 1977, Cenozoic geology of the Yerington District, Nevada, and its implications for the nature of Basin and Range faulting; Geol. Society America Bull., v. 88, p. 247-266.

Pyne, S.J., 1980, Grove Carl Gilbert, a great engine of research: London, University of Texas Press, 306 p.

Rabbitt, M.C., 1979, Minerals, Lands, and Geology for the Common Defence and General Welfare, volume 1, before 1879: U.S. Geological Survey, 231 p.

Reynolds, S.J., Richard, S.M., Haxel, G.B., Tosdal, R.M., and Laubach, S.E., 1988, Geologic setting of Mesozoic and Cenozoic metamorphism in Arizona, Englewood Cliffs, N.J., Prentice-Hall, p. 467-501.

Richard, S.M., Fryxell, J.E., and Sutter, J.F., 1990, Tertiary structure and thermal history of the Harquahala and Buckskin Mountains, west central Arizona: implications for denudation by a major detachment fault system. Journal of Geophysical Research, v. 95, p. 19,953-19,972.

Roddy, M.S., Reynolds, S.J., Smith, B.M.,m and Ruiz, J., 1988, K-metasomatism and detachment-related mineralization, Harcuvar Mountains, Arizona: Geological Society of America Bulletin, v. 100, p. 1627-1639. and others, 1988)

Severinghouse, J. and Atwater, T., 1990, Cenozoic geometry and thermal state of the subducting slabs beneath western North America: Geological Society of America Memoir 176, p. 1-22.

Sherrod, D.R. and Nielson, J.E., editors, 1993, Tertiary stratigraphy of highly extended terranes, California, Arizona, and Nevada: U.S. Geological Survey Bulletin 2053, 250 p.

Silliman, B., 1866, On some of the mining districts of Arizona near the Rio Colorado, with remarks on the climate, &c: American Journal of Science and Arts, Second Series, v. 41, p. 289-308.

Simpson, Carol, Schweitzer, Janet, and Howard, K.A., 1991, A reinterpretation of the timing, position, and significance of part of the Sacramento Mountains detachment fault, southeastern California: Geological Society of America Bulletin, v. 103, p. 751-761.

Simpson, R.W., Howard, K.A., Jachens, R.C., and Mariano, J., 1990, A positive gravity anomaly along the Colorado River extensional corridor: Evidence for new crustal material: EOS, American Geophysical Union Trans., v. 71, p. 1594.

Smith, E.I., Feuerbach, D.L., Naumann, T.R., and Mills, J.G., 1990, Mid-Miocene volcanic and plutonic rocks in the Lake Mead area of Nevada and Arizona; Production of intermediate igneous rocks in an extensional environment: Geological Society of America Memoir 174, p. 169-194.

Spencer, J.E., 1984, Role of tectonic denudation in warping and uplift of low-angle normal faults: Geology, v. 12, p. 95-98

Spencer, J.E., 1985, Miocene low-angle normal faulting and dike emplacement, Homer Mountain and surrounding areas, southeastern California and southernmost Nevada: Geological Society of America Bulletin, v. 96, p. 1140-1155.

Spencer, J.E., and Reynolds, S.J., 1990, Relationship between Mesozoic and Cenozoic tectonic features in west central Arizona and adjacent southeastern California. Journal of Geophysical Research, v. 95, p. 539-555.

Spencer, J.E. and Reynolds, S.J., 1991, Tectonics of mid-Tertiary extension along a transect through west central Arizona: Tectonics, v. 10, p. 1204-1221.

Stone, Paul, Howard, K.A., and Hamilton, Warren, 1983, Correlation of Paleozoic strata of the southeastern Mojave Desert region, California and Arizona: Geological Society of America Bulletin, v. 94, p.m 1135-1147.

Stratton. R.B., 1857, Captivity of the Oatman girls: New York, Carlton & Porter, reprinted 1983, London, University of Nebraska Press, 290 p.

Thomas, W. M., Clarke, H. S., Young, E. D., Orrell, S. E., & Anderson, J. L., 1988, Precambrian granulite facies metamorphism in the Colorado River region, Nevada, Arizona, and California: in Ernst, W. G., editor, Metamorphism and Crustal Evolution of the Western United States, Rubey volume VII, Prentice Hall, p. 527-537.

Thompson, D.G., 1929, The Mohave Desert Region, California: U.S. Geological Survey Water Supply Paper 578, 759 p.

Thorson, J.P., 1971, Igneous petrology of the Oatman district, Mohave County, Arizona: Santa Barbara, University of California, Ph.D. thesis, 189 p.

Tosdal, R.M., 1990, Tectonics of the Mule Mountains thrust system, southeast California and southwest Arizona: Journal of Geophysical Research, v. 95, p. 20,035-20,048.

Volborth, A., 1973, Geology of the granite complex of the Eldorado, Newberry, and northern Dead Mountains, Clark County, Nevada: Nevada Bureau of Mines and Geology Bulletin, no. 80, 40 p.

Ward, P.L. 1991, On plate tectonics and the geologic evolution of southwestern North America: Journal of Geophysical Research, v. 96, p. 12,479-12, 496.

Weber, M.E. and Smith, E.I., 1987, Structural and geochemical constraints on the reassembly of disrupted mid-Miocene volcanoes in the Lake Mead-Eldorado Valley area of southern Nevada: Geology, v. 15, p. 553-556.

Wernicke, B., 1981, Low-angle normal faults in the Basin and Range province; nappe tectonics in an extending orogen. Nature, v. 291, p. 645-648.

Wernicke, B., 1985, Uniform-sense simple shear of the continental lithosphere: Canadian Journal of Earth Science, v. 22, p. 108-125.

Wernicke, B. and Axen, G.J., 1988, On the role of isostasy in the evolution of normal fault systems: Geology, v. 16, p. 848-851.

Wernicke, B., Axen, G.J, and Snow, J.K., 1988, Basin and Range extensional tectonics at the latitude of Las Vegas, Nevada: Geological Society of America Bulletin, v. 100, p. 1738-1757.

Wernicke, B., Walker, J.D., and Beaufait, M.S., 1985, Structural discordance between Neogene detachments and frontal Sevier thrusts, central Mormon Mountains, southern Nevada: Tectonics, v. 4, p. 213-246

Whipple, A.W., 1856, Report of explorations for a railway route, near the thirty-fifth parallel of north latitude, from the Mississippi River to the Pacific Ocean, *in* Report of the Secretary of War on the several railroad explorations: U.S. Congress, 33d, 2nd session, Senate Executive Document 78, 136 p. (760).

Wilson, J.M., McCarthy, Jill, Johnson, R.A., and Howard, K.A., 1991, An axial view of a metamorphic core complex: Crustal structure of the Whipple and Chemehuevi Mountains, southeastern California: Journal of Geophysical Research, v. 96, p. 12,293-12,312.

Wooden, J.L. and DeWitt, Ed, 1991, Pb isotopic evidence for the boundary between the Early Proterozoic Mojave and Central Arizona provinces in western Arizona: Arizona Geological Society Digest, v. 19, p. 27-50.

Wooden, J.L., and Miller, D.M., 1990, Chronologic and isotopic framework for Early Proterozoic crustal evolution in the eastern Mojave Desert Region, SE California: Journal of Geophysical Research, v. 95, p. 20,133-20,146.

Wooden, J.L., Nutman, A.P., Miller, D.M., Howard, K.A., Bryant, B., Dewitt, E., and Mueller, P.A., 1994, Shrimp U-Pb zircon evidence for Late Archean and Early Proterozoic crustal evolution in the Mojave and Arizona crustal provinces: Geological Society of America Abstracts with Programs, v. 26, n. 2 (in press).

Wooden, J.L., Stacey, J.S., Howard, K.A., Doe, B.R., and Miller, D.M., 1988, Pb isotopic evidence for the formation of Proterozoic crust in the southwestern United States, *in* Ernst, W.G., editor, Metamorphism and crustal evolution of the western United States (Rubey Volume 7) Englewood Cliffs, N.J., Prentice-Hall, p. 69-86.

Wright, J.E., Anderson, J.L., and Davis, G.A., 1986, Timing of plutonism, mylonitization, and decompression in a metamorphic core complex, Whipple Mountains, California. Geological Society of America Abstracts with Programs, v. 18, p. 201.

Wright, J.E., Howard, K.A., and Anderson, J.L., 1987, Isotopic systematics of zircons from Late Cretaceous intrusive rocks, southeastern California; Implications for a vertically stratified crustal column: Geological Society of America Abstracts with Programs, v. 19, p. 898.

Yin An, and Dunn, J. F., 1992, Structural and stratigraphic development of the Whipple-Chemehuevi detachment fault system, southeastern California: Implications for the geometrical evolution of domal and basinal low-angle normal faults: Geological Society of America Bulletin, v. 104, p. 659-674.

Young, R.A. and Brennan, W.J., 1974, Peach Springs Tuff: its bearing on structural evolution of the Colorado Plateau and development of Cenozoic drainage in Mohave County, Arizona: Geological Society of America Bulletin, v. 85, p. 83-90.

Zen, E-an, and Hammarstrom, J.M., 1984, Magmatic epidote and its petrologic significance: Geology, v.12, p. 515-518.