GUIDEBOOK

for

CALIFORNIA STATE UNIVERSITY, NORTHRIDGE

DEPARTMENT OF GEOLOGICAL SCIENCES

40TH ANNIVERSARY
ALUMNI REUNION
FUN FIELD FROLIC
AT ZZYZX

JANUARY 29-31, 1999

PARTICIPANTS AT 40TH ANNIVERSARY ALUMNI REUNION FIELD TRIP TO ZZYZX

NAME	STATUS	BS YR	MS YR
Allen, Mark	A	88	
Ballmer, Matt	O		
Brown, Howard	A	79	
Caceres, Carmen	A, G	98	
cao Hsieh, Hsiu-Chin	O		
Cao, Aiguo	A		98
Cao, Charles	C		
Carroll, Edward	D		
Chatman, Mark	O		
Clark, Sharon	U		
Collins, Ellen	U		
Delis, Dave	U		
Devlahovich, Michael	C		
Devlahovich, Vince	G		
Dunne, George	F		
Elliott, Burt	O		
Elliott, Steve	A, G	95	
Field, Leni	G, F		
Fischer, Pat	A	87	
Francuch, Dean	A	87	
Freitag, George	A	86	89
Fritsche, Gene	F		
Geraci, Jeff	A	87	
Hanna, Frank	A	66	
Hanna, Marilyn	O		
Harding, Perry	A		
harma Williams, Angelique	C		
harma Williams, Catherine	C		
Harma, Roberta	A, F	78	
Hill, Kim	O		
Holt, Jake	A	97	
Howard, Bob	F		
Jewett, Sandy	F		
Kennedy, Cameron	U		
Klassette, Charles	A	87	
Kofoed, Jeff	A	88	
Loeb, Kim	A	88	
Lukesh, David	A	73	
Mack, Jeff	A	81	
Mansour, Sherief	A, G	97	
Miller, Jonathan	A	97	
Nussrallah, Charles	U		
Nussrallah, Judy	O		
Oborne, Mark	A	78	82
Palmer, Jody	A, G	97	

KEY: A = alumnus, C = child, D = dean, F = faculty, G = graduate student, O = other (spouse or guest), U = undergraduate student

PARTICIPANTS AT 40TH ANNIVERSARY ALUMNI REUNION FIELD TRIP TO ZZYZX

Panaro, Dave	A		
Panaro, Ryan	C		
Pedone, Vicki	F		
Phillips, Diane	A	78	
Ponek, Heather	O		
Ponek, Mike	A	79	
Puckett, Cindy	U		
Real, Charles	A	69	
Reid, Tom	G, F		
Reid, Tony	A	76	79
Richins, Jane	O		
Richins, Layne	A	85	
Roug, Ralph	A	86	
Schroeter, Chuck	A	67	
Sexton, Chris	A	83	
Shapiro, Shawn	U		
Sloan, Jon	F		
Stolla, Gordon	O		
Syms, Harold	A	73	
Syms, Steve	C		
Thompson, Gregory	C		
Thompson, Jennifer	C		
Thompson, Terry	A	80	
Thun, Roy	A	88	
Trembly, Butch	A		87
Weigand, Peter	F		
Werner, Mike	A	76	79
wood Le Page, Tina	A	87	
Wood, Lauren	C		
Wood, Michael	A	87	

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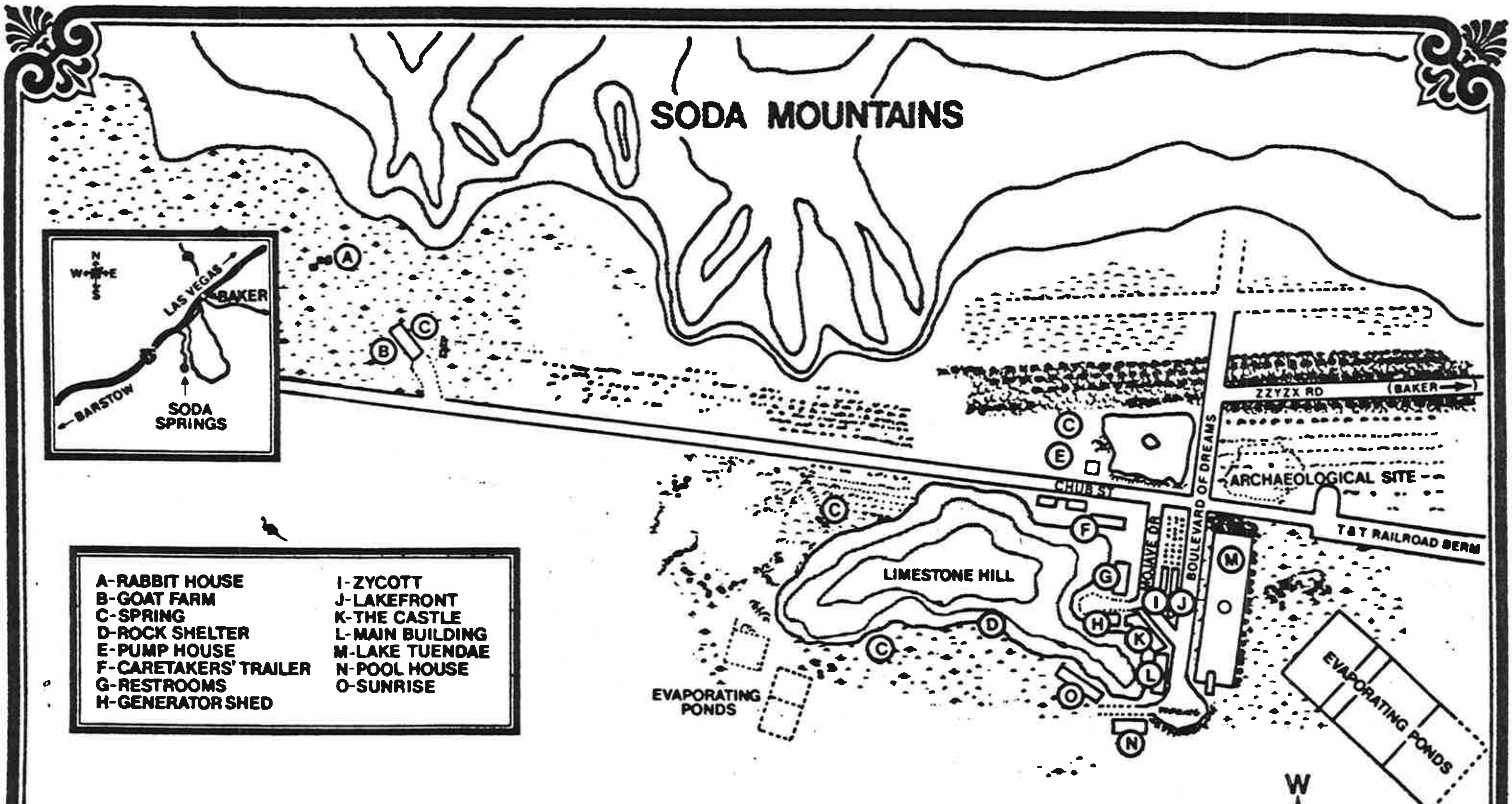

PLAN OF SODA SPRINGS
MOJAVE DESERT, CALIFORNIA

Second Symposium on Salt, Northern Ohio Geological Society, vol. 1, p. 133-150, 1966

Stratigraphic and Structural Evolution of the Kramer Sodium Borate Ore Body, Boron, California

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ABSTRACT

The Kramer sodium borate ore body consists of a lenticular sedimentary facies of borax and kernite crystals together with varying amounts of interstitial and interbedded claystone. This sodium borate facies has been divided into seven stratigraphic units. The sodium borate facies is successively enveloped by a ulexite facies, a colemanite facies, and a barren claystone facies. Together, these four facies constitute the Shale member of the Kramer beds. The term Kramer beds is locally used to designate all of the conformable Miocene strata, including the borates, between the base of the Quaternary alluvium and the base of the Saddleback basalt. The borax is believed to have been precipitated in a permanent shallow lake, fed from nearby thermal springs containing anomalous amounts of sodium and boron. AsS and Sb_2S_3 were most abundant in the lake during a late stage of borax deposition. Successive beds of borax crystals protected by layers of mud were progressively deposited as the lake remained structurally low due to movement along a fault scarp at its south edge. The ore body has been deformed by faults and associated folds which developed in several stages. Minor faulting and folding took place in the borax lake contemporaneously with borax deposition. Some of the borax was altered to kernite after burial beneath a great thickness of fluvial sediments. Later, some of the kernite reverted back to borax when regional uplift and erosion and a renewal of local faulting and folding resulted in a reduction of temperature and redistribution of excess water within the ore body.

INTRODUCTION

The Kramer borate deposit is in the northwestern Mojave Desert of California, approximately 100 miles northeast of Los Angeles, immediately north of the town of Boron (Fig. 1). The deposit derives its name from the mining district in which it lies. The deposit is presently being mined from the Boron open pit, which supplies a major portion of United States borate production.

This paper deals with previously unpublished stratigraphic and structural studies of the borate deposit, based primarily on mapping undertaken during the last three years. This basic work is still in progress and will be completed before mineralogic and geochemical studies of a detailed nature are undertaken. The work completed to date has resulted in a series of detailed geologic maps of the ore body which are presently being utilized in open pit design and other mining operations.

The ore body consists of a roughly lenticular crystalline mass of borax ($NA_2B_4O_7 \cdot 10H_2O$), and kernite ($NA_2B_4O_7 \cdot 4H_2O$), containing interbedded claystone, and completely enveloped by ulexite-bearing shales (Gale, 1946). Stratigraphic and structural studies indicate that the Kramer borates were deposited in a small structural, nonmarine basin, elongated in an east-west direction and limited on the south by a fault scarp.

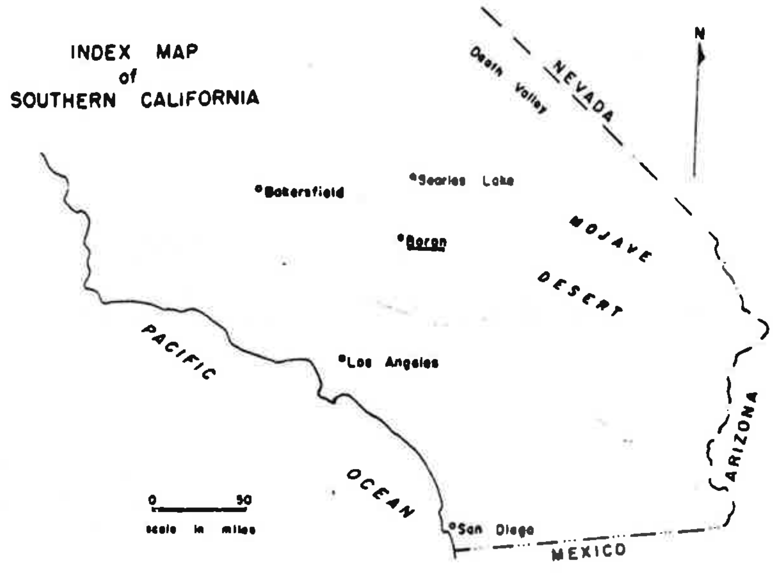


Figure 1. Location map.

Gale (1946) assigned the Kramer borates and associated shales, together with the underlying basalt and overlying arkose, to the Kramer Lake Bed member of the Pliocene Ricardo Formation. This correlation was made on the basis of lithologic similarities and geographic proximity to the Ricardo sediments and volcanics which crop out about 29 miles northwest of the Kramer deposit. The Kramer borates and associated sediments were later placed in the Pliocene (?) upper part of the Tropico Group (Dibblee, 1958).

Recently, however, well-preserved mammalian remains have been uncovered in sediments overlying the borate deposits in the Boron open pit mine. These fossils have been identified as a pre-Ricardo fauna, with an age no younger than early Middle Miocene (Whistler and Tedford, 1964). The term Ricardo can therefore no longer be used for any of the beds associated with the Kramer deposit, and Dibblee's age correlation within his Tropico Group must be revised.

Figure 2. Exposure of the Kramer beds on the north side of the Boron open pit.

GENERAL STRATIGRAPHY OF THE KRAMER BEDS

The term Kramer beds is being used locally by U.S. Borax geologists to designate all the similar dipping Miocene strata, including the borates that lie between the base of the Quaternary alluvium and the base of the Saddleback basalt in the structural basin. The saddleback basalt is unconformably underlain by older Tertiary arkoses, tuffs, and shales.

We have divided the Kramer beds into three distinct members, the Saddleback basalt member, the Shale member, and the Arkose member, in ascending order. Figure 3 is a generalized stratigraphic section of the Kramer beds near the central part of the ore body.

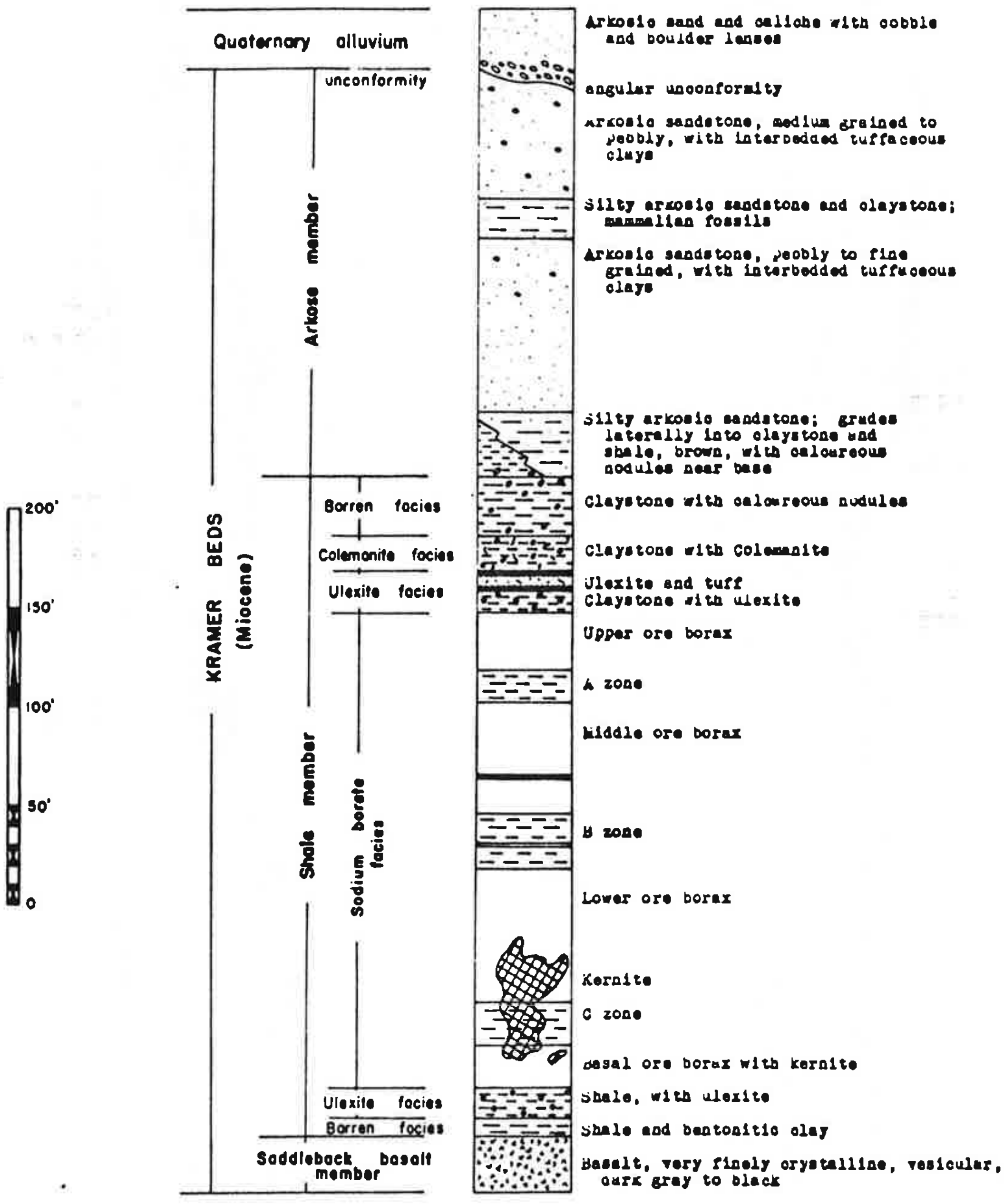


Figure 3. Generalized stratigraphic section of the Kramer beds.

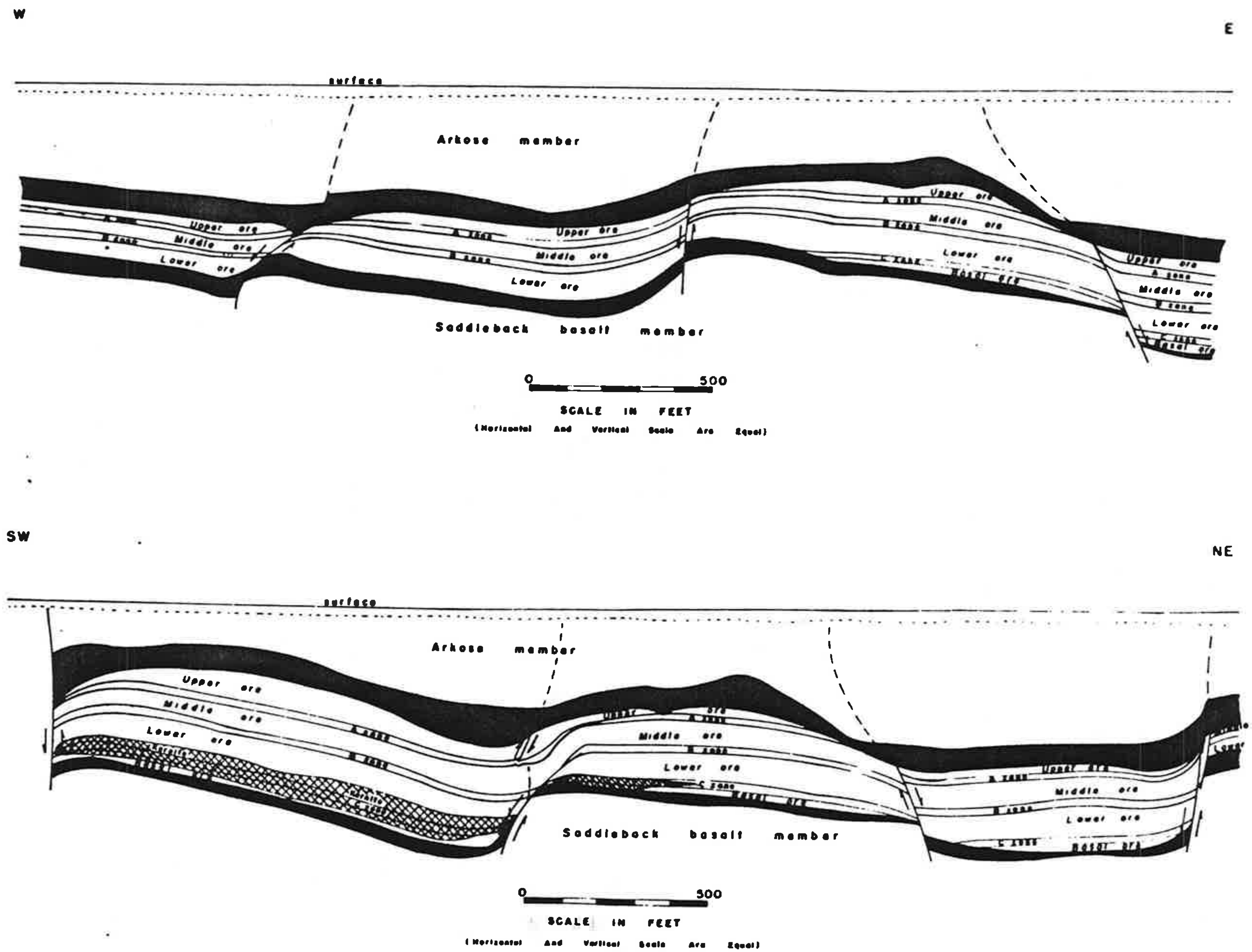


Figure Cross sections of parts of the Kramer borate deposit. Lines of section shown on Figure 8. Darkened areas above and below the ore body represent the ulexite, colemanite, and barren claystone facies.

STRUCTURAL AND DEPOSITIONAL HISTORY AND ITS EFFECTS UPON THE SODIUM BORATE FACIES

Our detailed stratigraphic and structural studies to date have led to the following conclusions regarding the history of the Kramer beds, and particularly the sodium borate facies:

Events immediately prior to and during the deposition of borax.

No later than early Middle Miocene, the Saddleback basalt was extruded on older Tertiary sediments in the area of the future borate basin. This area was already topographically low due to downward movement along the "Western Borax" Fault immediately to the south. Soon after the extrusion of basalt, the topographically low area became a playa basin. The deepest part of the basin was close to its southern border, which was formed by the fault scarp. To the north, east, and west the rise out of the basin was more gradual.

In the structural setting described above, boron and sodium, derived from local thermal springs, combined with calcium from surface water and formed ulexite in the playa muds. (A local thermal spring source for the borate deposits at Kramer has been suggested by Gale, Muessig, and others.) This belief, which we support, is based primarily on the presence of volcanic rocks in the Kramer beds, the association of syngenetic arsenic and antimony sulphides with the borax, the limited suite of evaporitic minerals, and the observation of recent borax deposition at other localities.

A permanent shallow lake soon developed in the central portion of the playa basin. The lake, fed almost entirely from the local thermal springs, contained a great abundance of sodium and boron. The cooling lake brine became saturated with respect to borax, which was precipitated in great abundance in the muds. Muessig (written communication, 1956) has suggested that borax, whose solubility is much more sensitive to temperature changes than other evaporites, probably crystallized at times of decreased temperatures. The solubilities of other salts such as NaCl are not much affected by small temperature changes, and if present in the lake, they remained in solution in the surface waters until they were removed by overflow at a low outlet. The primary precipitation of borax in the lake probably occurred at temperatures in the range (25°-35° C (Christ & Garrels, 1959, p. 517).

During the period of borax, ulexite, and lake mud deposition, the local basin remained structurally low as a result of continued downward movement along the "Western Borax" Fault to the south. Thus, successive beds of borax crystals protected by layers of mud were progressively built up as the basin sank. In the outer portions of the playa, the lake water mingled with calcium-bearing surface and ground waters, resulting in the precipitation of ulexite rather than borax along the margins of the central borax lake.

At three different times during the deposition of borax, temporary slight climatic changes or slight decreases in available boron resulted in the formation of the low-grade A, B, and C zones. During deposition of the B zone, and again during the early deposition of Middle ore, layers of arkosic to silty volcanic ash settled in the borax lake.

The center of borax deposition shifted and the size of the lake varied as the basin was filled with sediments and borax. The first borax deposition was restricted to the southern and eastern portions of the ore body. The borax lake probably reached its maximum development in area in "Lower ore time." During "Middle ore time," the lake retracted slightly from the south and east and shifted slightly farther north. The area covered by the lake was considerably reduced in "Upper ore time," having retracted from the east, south, and north.

From time to time anomalous amounts of AsS and Sb_2S_3 were precipitated in the lake along with the borax. The greatest precipitation of AsS and probably of Sb_2S_3 occurred during the late stages of borax deposition. The local thermal springs were probably the source of arsenic and antimony.

Minor faulting and folding along northwest trending lineaments took place in the borax lake contemporaneously with borax deposition. These vertical movements resulted in less deposition of borax adjacent to faults, especially on the upthrown side.

Waterman Hills (north of Barstow) Field Trip Stop

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

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

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

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

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

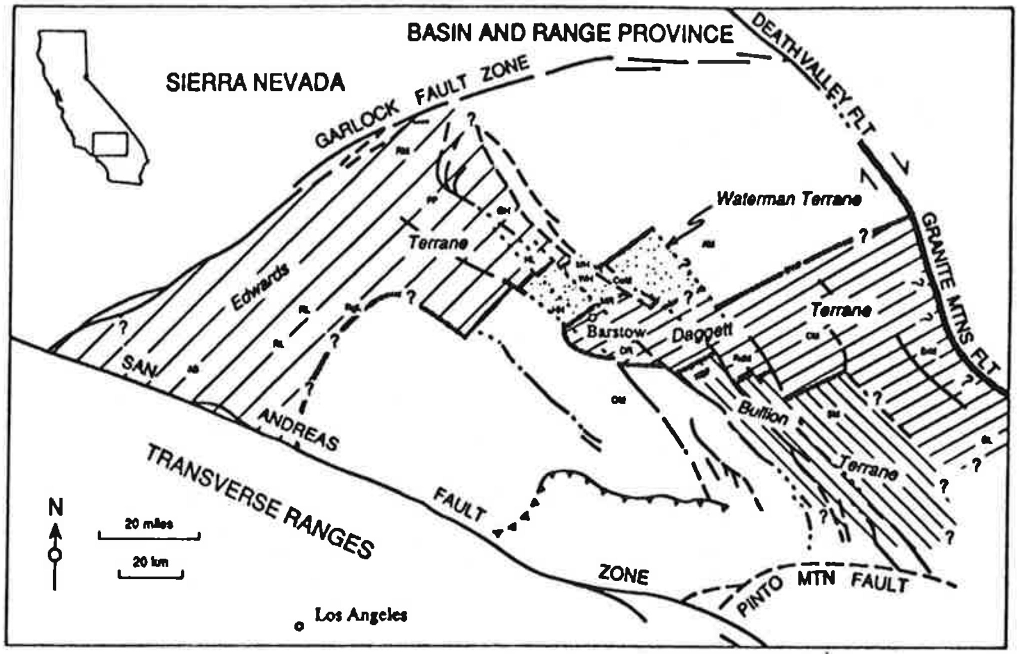


Figure 5. Index map of the Mojave Extensional Belt showing location of structural domains (modified from Dokka, 1989a). Localities and features mentioned in text: AM, Alvord Mountains; AB, Antelope Buttes; BM, Bullion Mountains; BrM, Bristol Mountains; BL, Bristol Lake; CM, Cady Mountains; DR, Daggett Ridge; FP, Fremont Peak; GH, Gravel Hills; HL, Harper Lake; HH, Hinkley Hills; KH, Kramer Hills; MR, Mitchel Range; MH, Mud Hills; NM, Newberry Mountains; OM, Ord Mountains; RM, Rand Mountains; RdM, Rodman Mountains; RgL, Rodgers Dry Lake; RH, Rosamond Hills; RL, Rosamond Dry Lake; WH, Waterman Hills. KSF, Kane Springs fault; LMF, Lane Mountain fault; BWF, Baxter Wash fault.

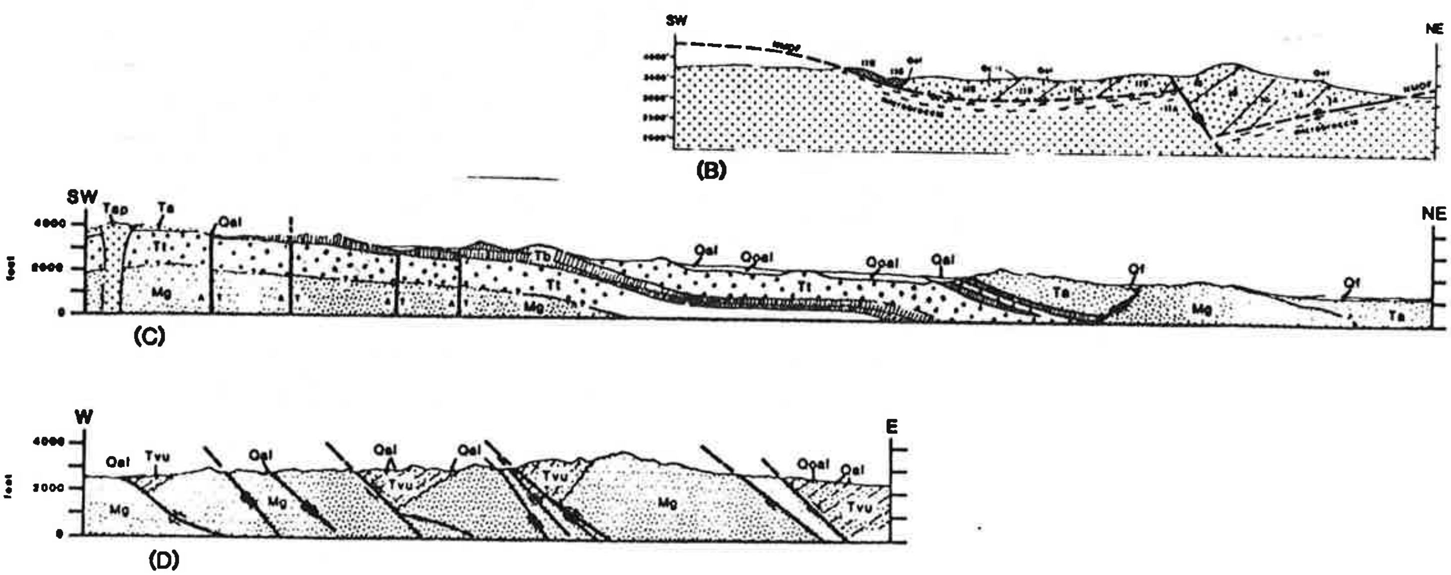


Figure 3. Examples of the architecture of extensional domains of the Mojave Rift. (A) Block diagram of the western Mojave Desert (Waterman terrane) based on COCORP seismic reflection data (Cheadle and others, 1986). Note reflection B and the inferred high-angle normal faults that lie above it. Reflection B is interpreted to be a detachment fault. This fault projects to the surface at STOP #5. (B) Structure section across the north-central Newberry Mountains (Daggett terrane) (based on map presented in Dokka [1986]). (C) Structure section across the Bullion Mountains (Bullion terrane) (after Dibblee, 1967). (D) W-E cross-section of of the central Cady Mountains (Daggett terrane) (from Mathis, 1986).

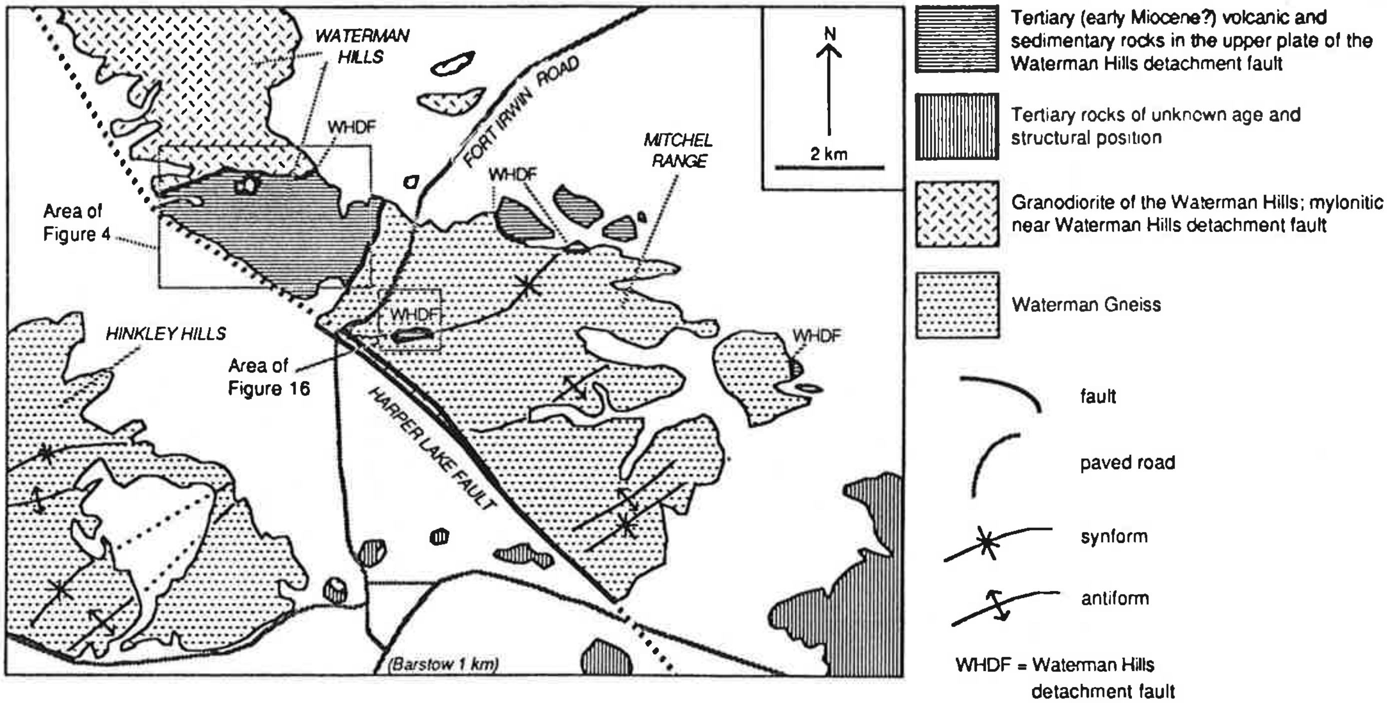


Figure 3. Geology around the Waterman Hills detachment fault. From Dibblee (1960, 1970), with modifications in the Waterman Hills and Mitchel Range. WHDF - Waterman Hills detachment fault.

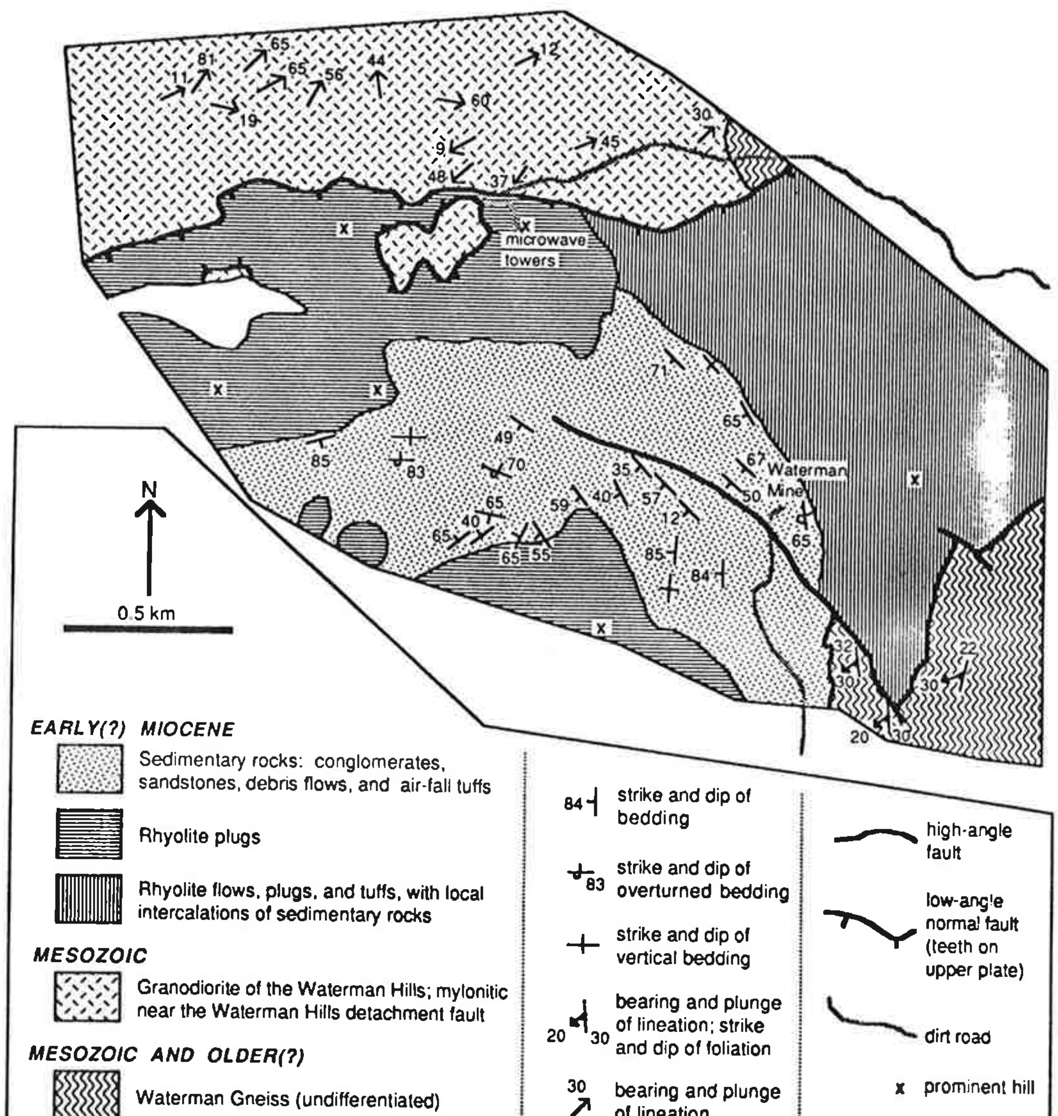
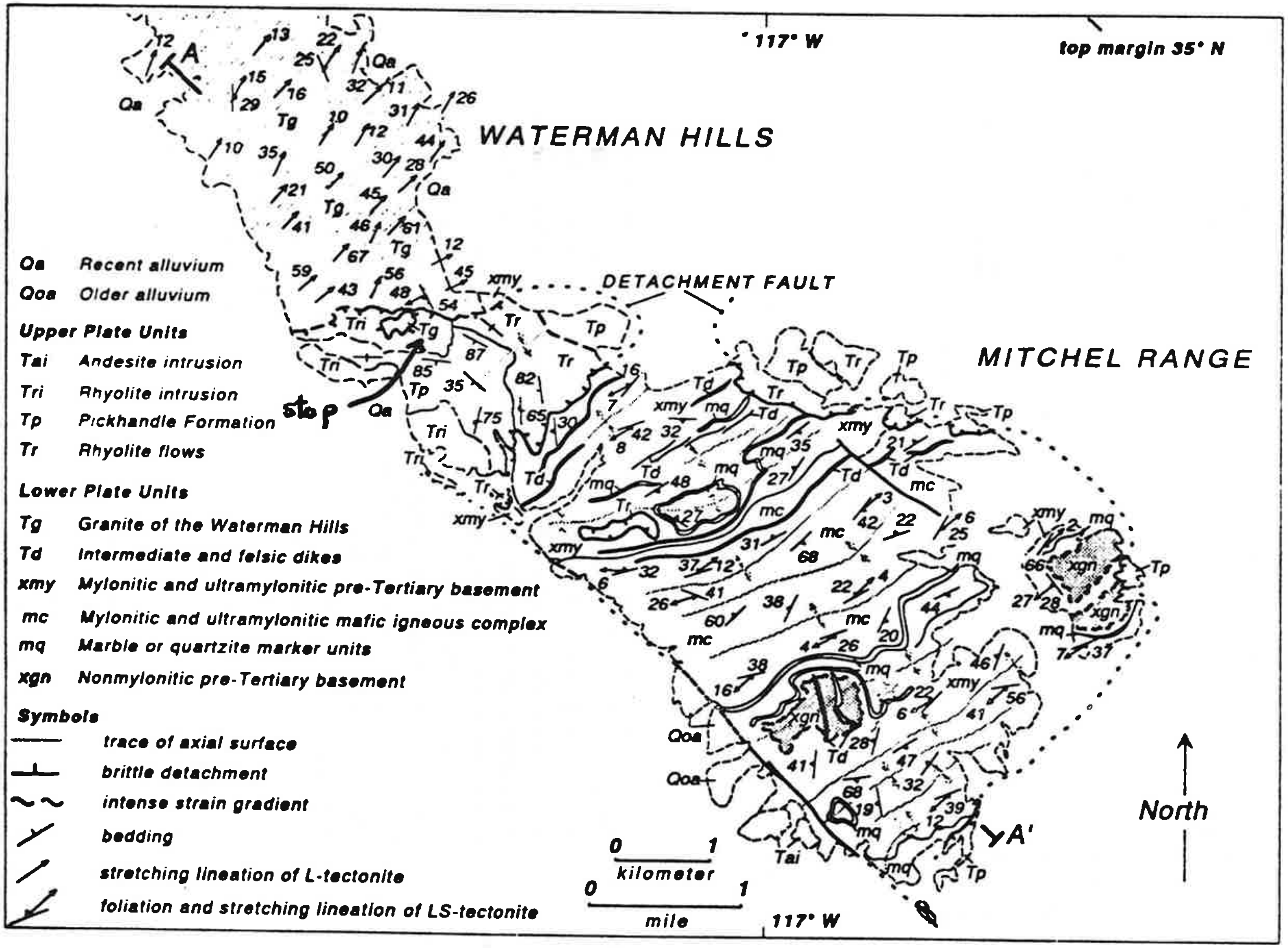
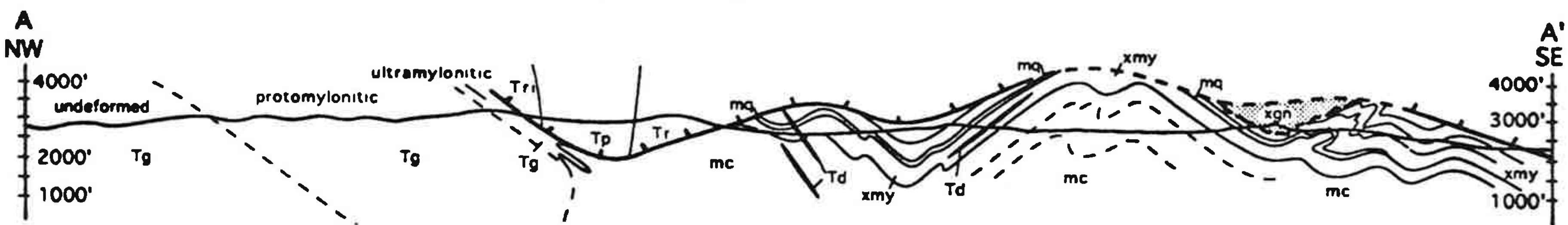


Figure 4. Preliminary geologic map of the Waterman Hills detachment fault. Mapping by the authors, S. B. Dent, and M. W. Martin.


117° W
top margin 35° N
WATERMAN HILLS
DETACHMENT FAULT
MITCHEL RANGE
North
stop
A
A'
0 1 kilometer
0 1 mile
117° W
Qa Recent alluvium
Qoa Older alluvium
Upper Plate Units
Tai Andesite intrusion
Tri Rhyolite intrusion
Tp Pickhandle Formation
Tr Rhyolite flows
Lower Plate Units
Tg Granite of the Waterman Hills
Td Intermediate and felsic dikes
xmy Mylonitic and ultramylonitic pre-Tertiary basement
mc Mylonitic and ultramylonitic mafic igneous complex
mq Marble or quartzite marker units
xgn Nonmylonitic pre-Tertiary basement
Symbols
trace of axial surface
brittle detachment
intense strain gradient
bedding
stretching lineation of L-tectonite
foliation and stretching lineation of LS-tectonite


A NW
A' SE
4000'
3000'
2000'
1000'
undeformed
protomylonitic
ultramylonitic
Tg
Tri
Tp
Tr
mc
mq
Td
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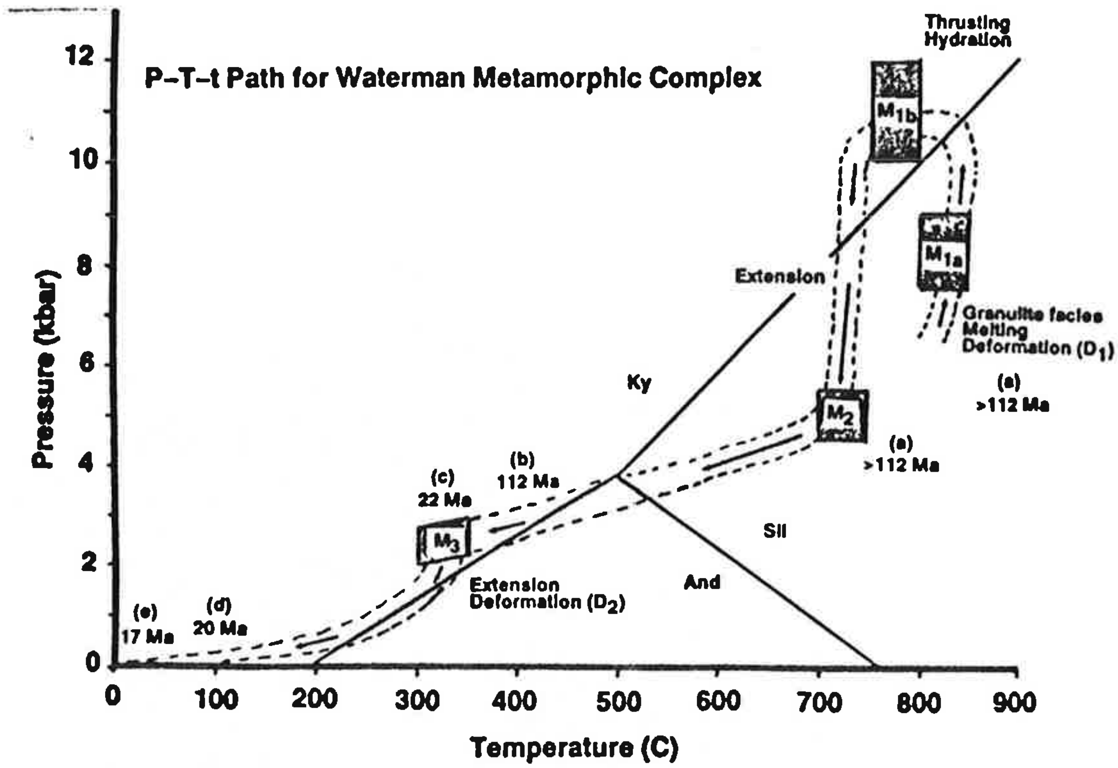
xgn



The Waterman Metamorphic Complex is polymetamorphic having been affected by an initial series of high grade metamorphic events and a localized lower grade overprint (Fig. 3): the first metamorphism attaining granulite/upper amphibolite facies conditions (M_1), the second metamorphism amphibolite facies conditions (M_2) and the third metamorphism (M_3) lower greenschist facies. The initial metamorphism (M_1) can be separated into two stages along its high grade PT path: M_{1a}, a granulite facies metamorphism at 800-850°C and 7.5-9 kbar and M_{1b}, an upper amphibolite facies overprint at 750-800°C and 10-12 kbar. M_{1a} developed generally anhydrous mineral assemblages and textures consistent with granulite facies conditions at a reduced activity of H_2O and is associated with an intense deformation (D_1) and local anatectic melting. M_{1b} overprinted the granulite assemblages to varying degrees in the different lithologies with a series of hydrous phases under conditions of increasing pressure and H_2O activity and was accompanied by little or no deformation. M_2 developed at lower pressures and somewhat lower temperatures (700-750°C; 4.5-5.5 kbar) and is distinguished by a second local overprint of hydrous phases that reflects an input of aqueous fluids probably associated with the intrusion of a series of granitic dikes and veins. M_3 developed locally near the early Miocene detachment zones and is characterized by local ductile/brittle deformation (D_2) and aqueous fluid influxes and superimposed greenschist facies conditions (300-350°C; 2-3 kbar) on the core complex rocks. The influence of the M_3 overprint is a function of proximity to the detachment zone, the depth of the faulting and the rock type; the marbles were the most susceptible to deformation.

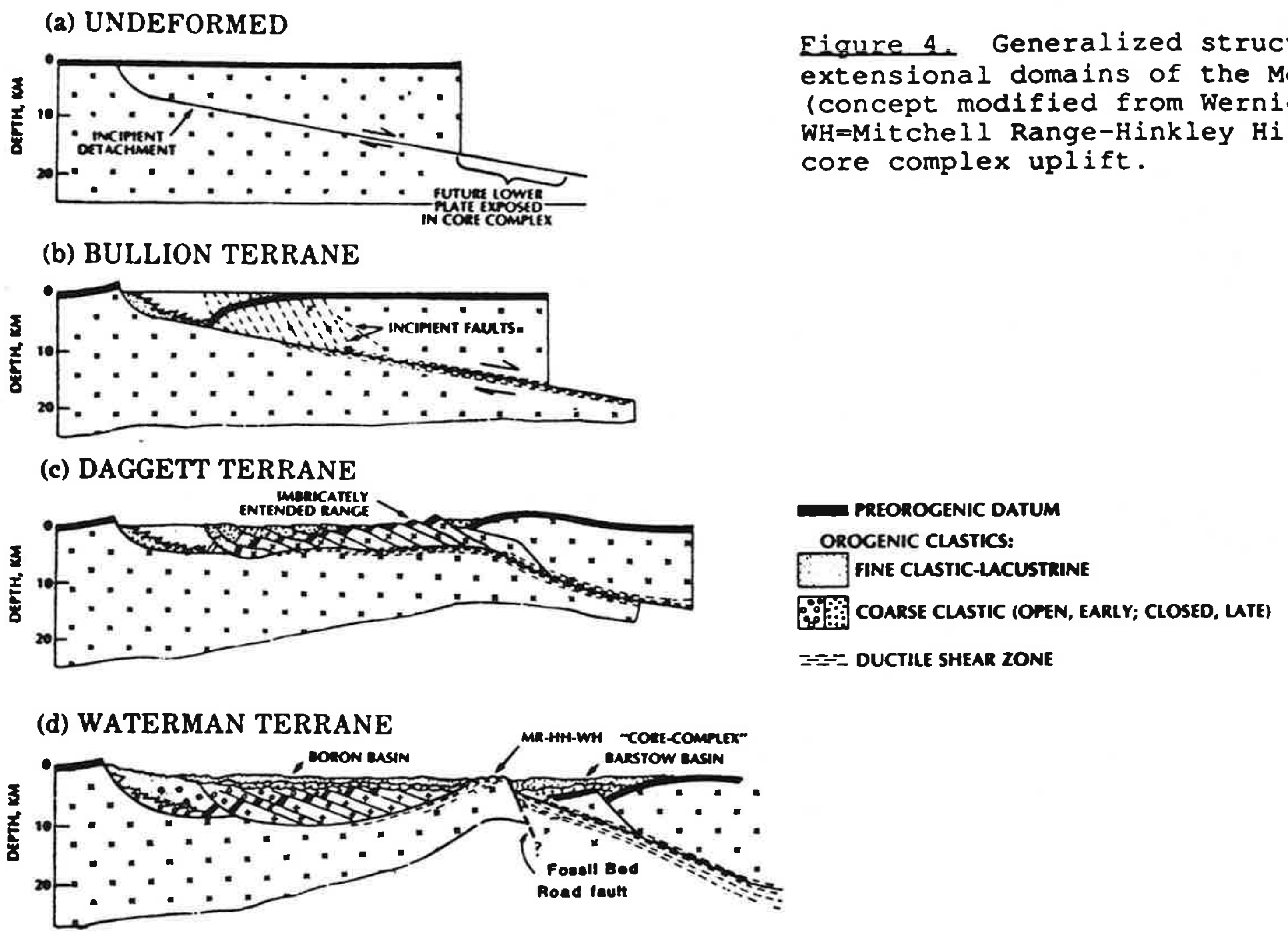


Figure 4. Generalized structure of extensional domains of the Mojave Rift (concept modified from Wernicke, 1985). MR-HH-WH=Mitchell Range-Hinkley Hills-Waterman Hills core complex uplift.

-Dokka

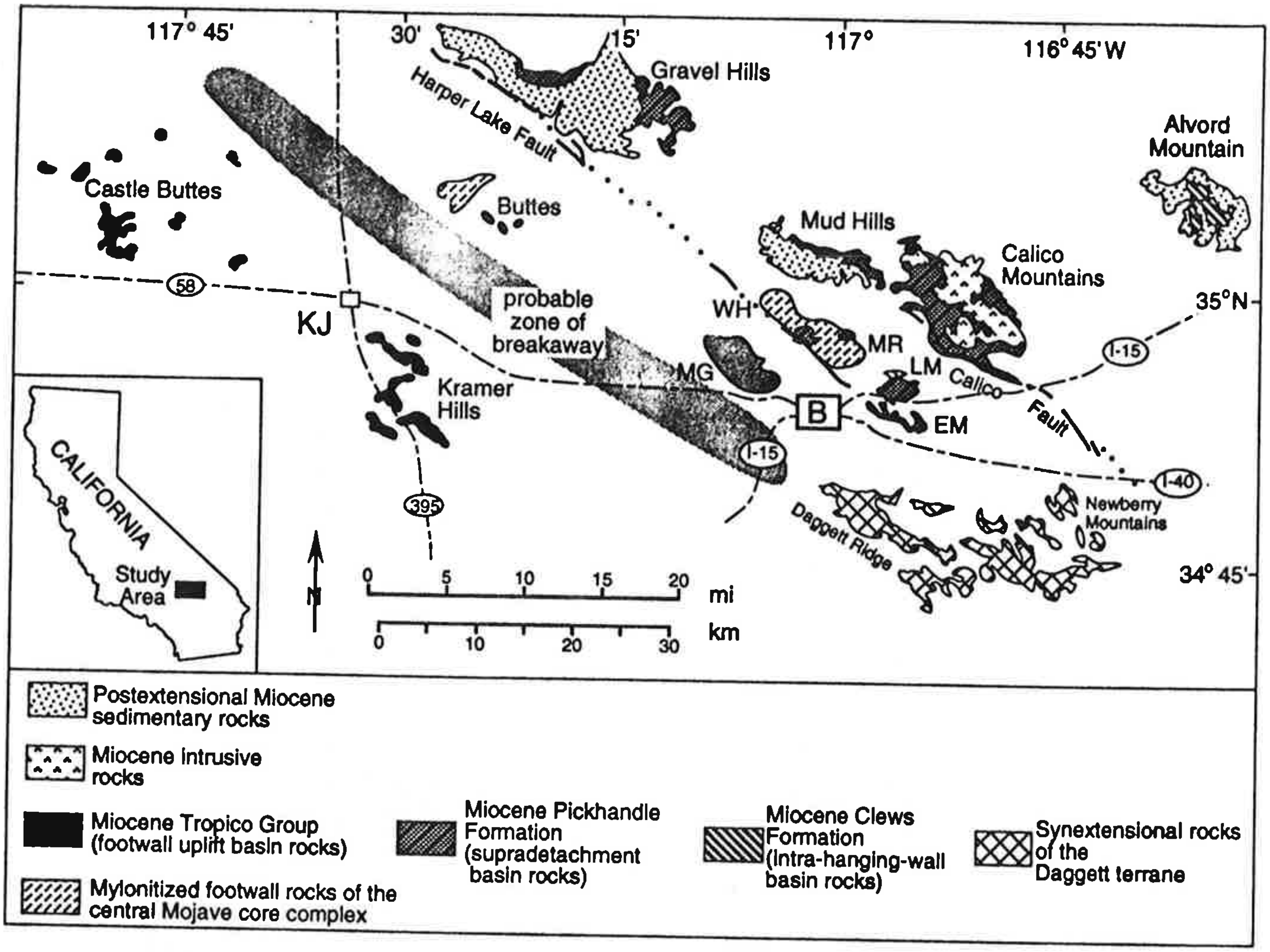


Figure 2. Simplified geologic map of central Mojave Desert showing distribution of lower Miocene rocks and footwall rocks of central Mojave metamorphic core complex. EM—Elephant Mountain, LM—Lead Mountain, MG—Mt. General, MR—Mitchel Range, WH—Waterman Hills, B—town of Barstow, KJ—Kramer Junction.

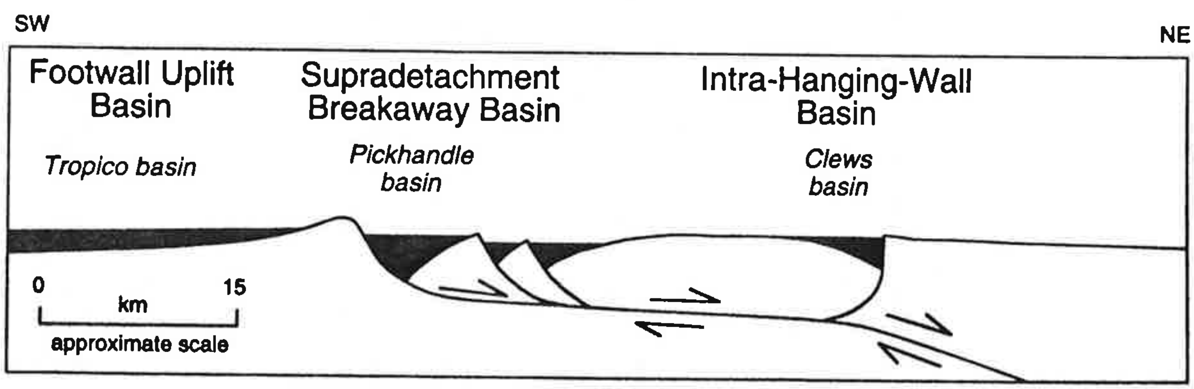


Figure 1. Schematic cross section of basins formed early in development of metamorphic core complex. This is referenced to central Mojave extensional system and shows relative locations of basins and related tectonic elements. Continued deformation may significantly disrupt these basins.

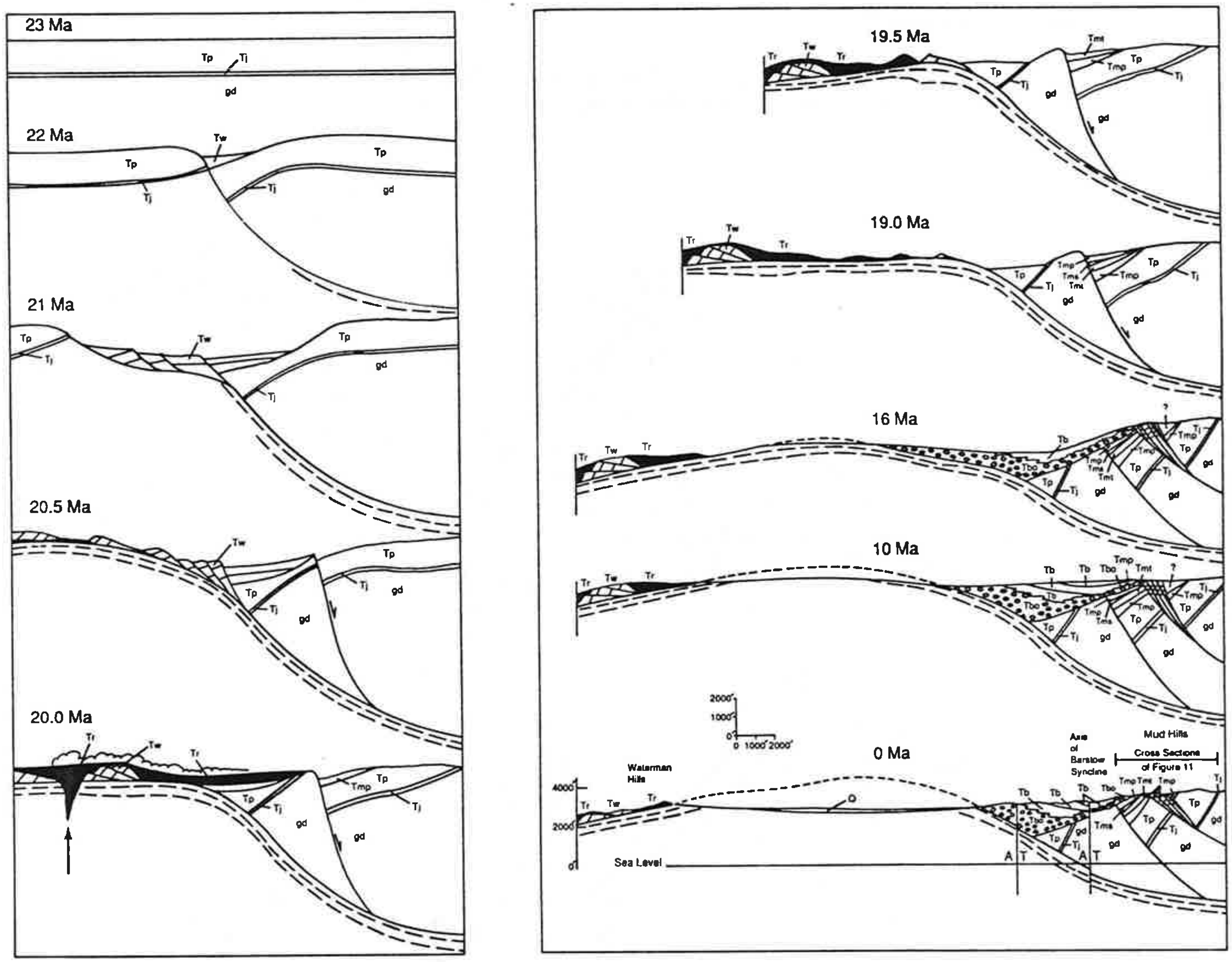


Figure 15 (on this and facing page). Sequential development of Waterman Hills detachment and Mud Hills area (see Fig. 1 for location). Palinspastic reconstruction is at true scale, with right side (north) held fixed. Some thin units (e.g., distal Tms) are omitted because of scale. Topography and geologic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

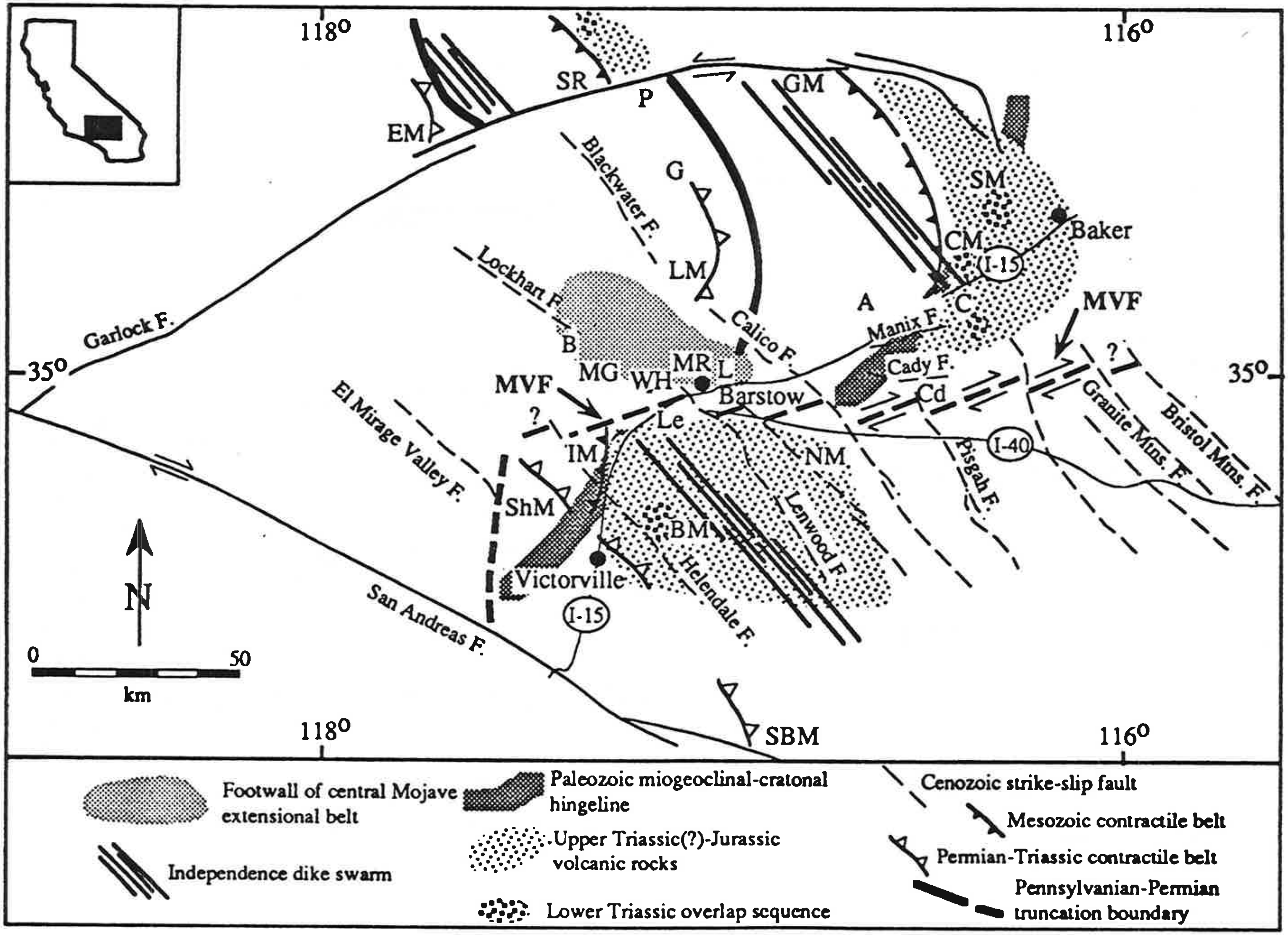


Figure 1. Regional tectonic map of Mojave Desert showing proposed location of Mojave Valley fault (MVF). EM—El Paso Mountains, SR—Slate Range, GM—Granite Mountains, P—Pilot Knob, B—Buttes, G—Goldstone, A—Alvord Mountain, CM—Cronese Mountain, SM—Soda Mountains, C—Cave Mountain, Cd—Cady Mountains, LM—Lane Mountain, MG—Mount General, MR—Mitchel Range, NM—Newberry Mountains, L—Lead Mountain, Le—Lenwood, IM—Iron Mountain, ShM—Shadow Mountains, BM—Black Mountain, SBM—San Bernardino Mountains, WH—Waterman Hills. Jurassic and Permian-Triassic contractile belt symbols do not necessarily imply vergence or dip direction. Paleozoic miogeoclinal-cratonal hingeline is used as defined by Martin and Walker (1991, 1992). Mojave Valley fault is schematically shown offset by northwest-trending late Cenozoic strike-slip faults, which have individual offsets of 1–10 km (Dokka, 1983).

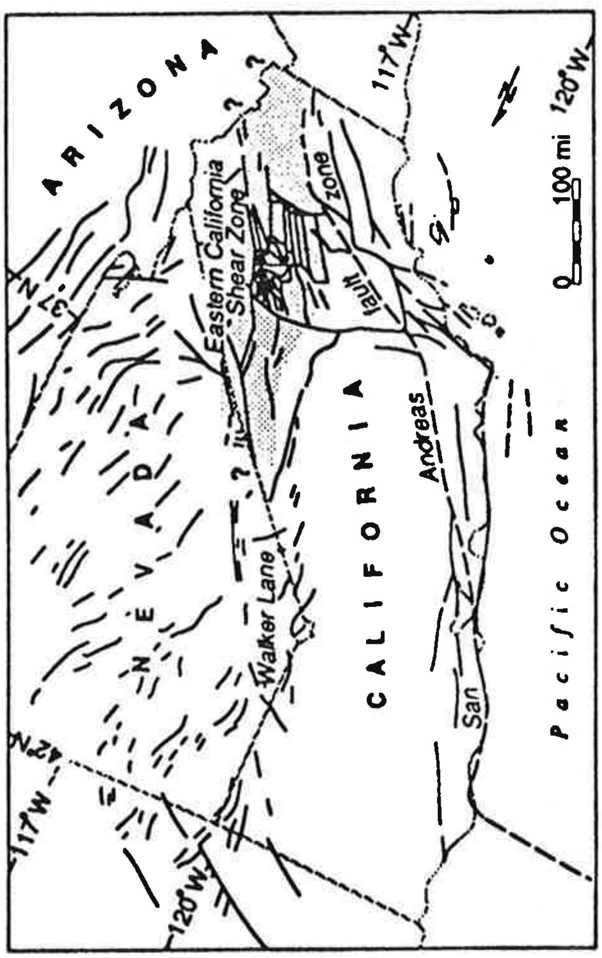


Figure 6. (A) The Pacific-North American transform boundary in the western USA highlighting the location of the Eastern California shear zone (Dokka and Travis, 1990b). (B; following page) Fault map of the Mojave Desert highlighting the location of late Cenozoic faults and associated features (from Dokka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

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

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

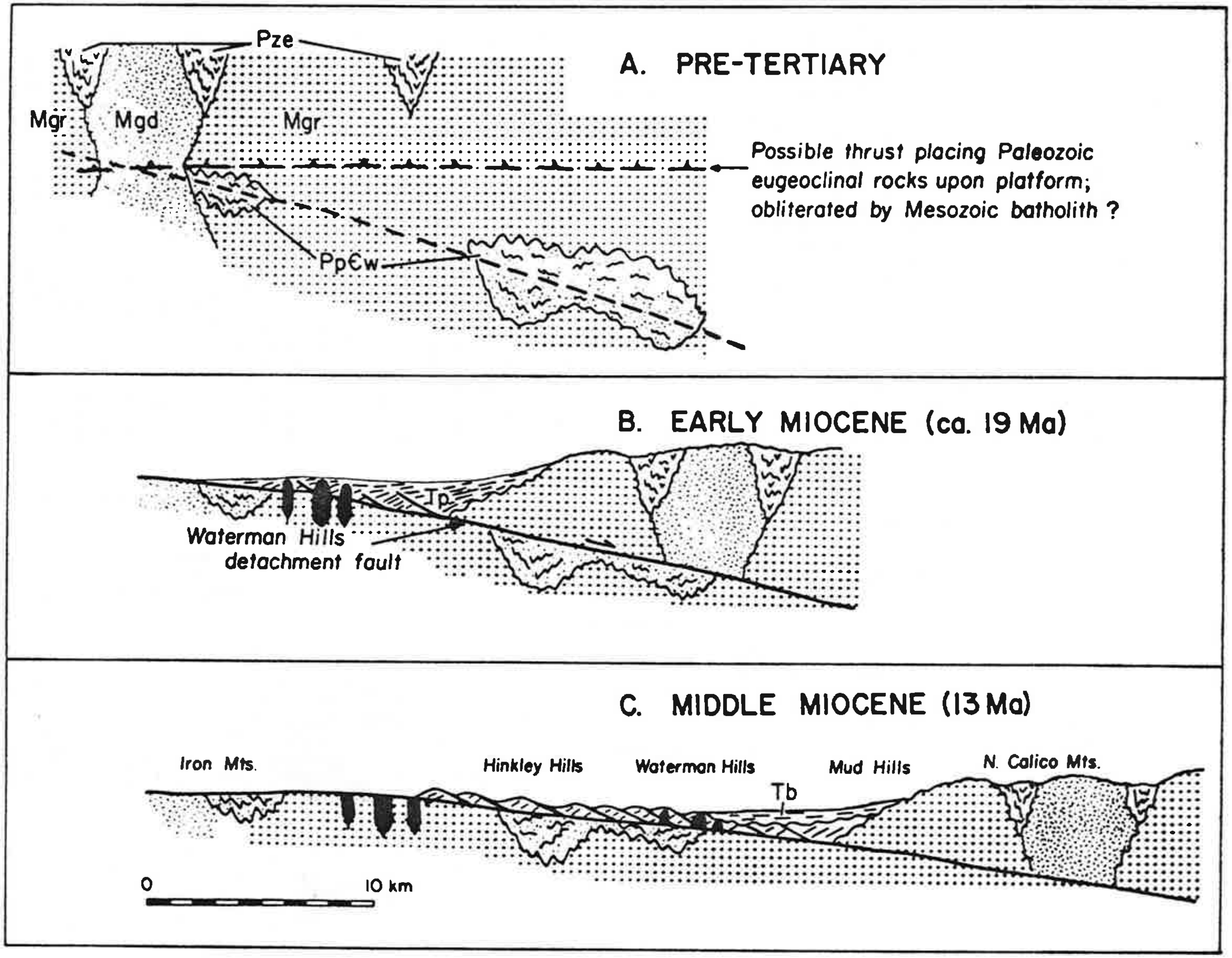


Figure 3. Conceptual model for evolution of Waterman Hills detachment fault (WHDF). Neogene folding related to right-slip Calico fault (Dibblee, 1968) has been removed. A: Geometry with 40 km of displacement on WHDF restored. Eugeoclinal Paleozoic rocks (Pze) lie structurally above miogeoclinal/cratonal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

STOP #6 - CALICO MOUNTAINS

Objectives

1. Observe middle Miocene Barstow Formation.

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

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

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

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

* * * * * * * *

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

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

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

STOP #8 - AFTON BEACH RIDGE

Objectives

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

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

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

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

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

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

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

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

Rancholabrean Age

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

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

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

* * * * * * * *

Return to I-15 north

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

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

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

About one mile beyond Rasor Road, look at 3

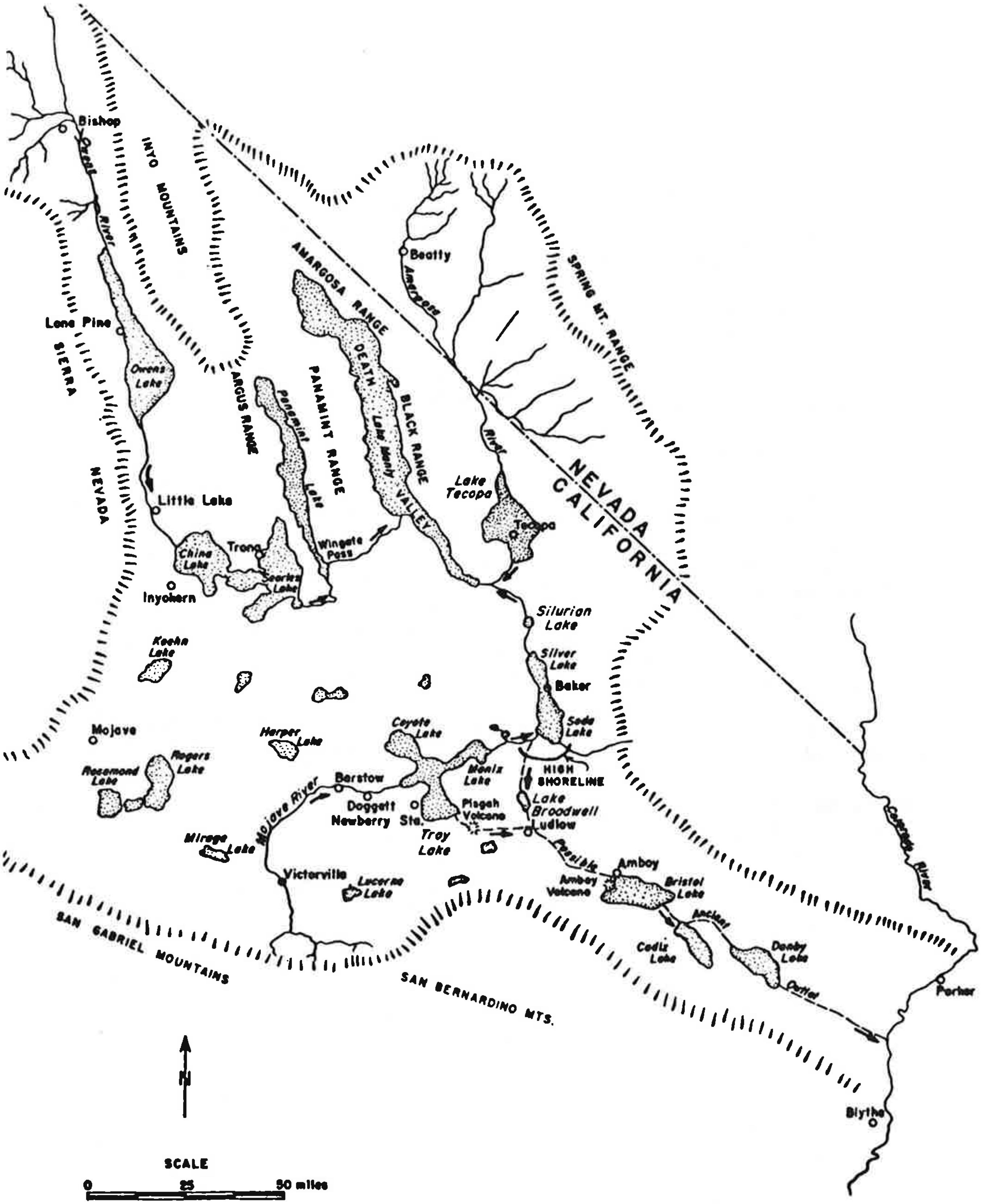


Figure 17. Pleistocene drainage in the Mojave Desert (modified from Blackwelder, 1954).

AFTON CANYON GEOLOGY

Prepared by Ken Schulte, BLM

INTRODUCTION:

During the California Desert Conservation Area (CDCA) planning process (1976 - 1980), Afton Canyon was nominated as an Area of Critical Environmental Concern to protect natural and scenic values in the area. The nomination noted that Afton Canyon is one of the few places where the Mojave River surfaces and sustains extensive riparian vegetation and diverse wildlife. It also noted that Afton Canyon contains outstanding scenic quality due to the unique vegetation and spectacular erosional strategical. Afton Canyon itself is a notable geologic feature, even without considering associated geology (Wachter, et al, 1976, p.244). BLM is preparing a report in support of a protective withdrawal petition for a total of 8,835.26 acres.

PRE-PLEISTOCENE HISTORY:

The basement complex, consisting of Mesozoic and pre-Mesozoic-age metamorphic and igneous rocks is overlain by Cenozoic-age volcanic and sedimentary deposits. The oldest rocks are in the central and eastern part of Afton Canyon. Marble, exposed at the downstream end of the canyon, is believed to be Paleozoic in age. An orthoquartizite unit in the metamorphic-igneous complex is tentatively correlated with the Triassic-Jurassic Aztec Sandstone on the basis of lithologic similarity and age constraints (Cameron et al, 1979, p. 397).

During and since the Mesozoic era, the region has undergone uplift and erosion which has exposed large areas of intrusive and metamorphic rocks. During the period of uplift and erosion, igneous activity continued with the intrusion of numerous dikes. Volcanism became particularly active during the Miocene epoch with extrusive and, in part, explosive phases. Pyroclastic rocks are exposed on the south side of Afton Canyon in Sections 20 through 24 and Sections 28 through 30. Erosion of the pre-Tertiary rocks continued during the Miocene epoch. Drainage was partly into internal basins with deposition of non-marine sediments (Collier & Danehy, p. 18). Faulting and folding were active during Tertiary time and have continued up to the present.

PLEISTOCENE LAKES

More than a century ago it was recognized that many of the valleys in the Great Basin states of Utah, Nevada, Oregon and California were completely enclosed and lacked outlets to the sea (Snyder, et al, 1964). Throughout many of these valleys the early investigators recognized land forms as having formed in a lake environment.

Manix Lake existed during the late Pleistocene, or Wisconsin stage, probably representing the pluvial period equivalent to the Tioga stage of Sierra glaciation. Water that filled Manix Lake came from the north slope of the San Gabriel Mountains before complete capture by drainage through Cajon Pass, and from the San Bernardino Mountains, which formed the headwaters of the Mojave River in Manix time. If the present rainfall had been doubled during a pluvial (rainy) period, it would have represented little actual increase in the desert areas, which might have had 8 or 10 inches, but would have represented an increase to something like 50 inches in the mountains, enough to supply a good-sized perennial river and to keep a large desert lake filled. Overflow from Manix Lake filled the pluvial Lake Mohave which covered both the present Silver Lake and Soda Lake Basins, respectively north and south of Baker, and overflowed that lake into Death Valley to fill Lake Manly (Bassett, A.M., 1967, p. 2).

MANIX LAKE:

In late Pleistocene time, a large area encompassing Coyote Lake, Troy Lake and the Mohave River eastward to Afton Canyon, was covered by Manix Lake. Fossils have been found mainly in the upper lake clay facies of the Manix Beds of Buwalda (1914). These include bones, largely those of birds (some extinct, but most still living), mammals (including horse and camel), and locally abundant shells of clams and snails.

Blackwelder and Ellsworth (1936) examined exposures of lacustrine sediments in the Afton Canyon area, and inferred two moist epochs of extensive lakes. Radiometric dates and field evidence suggest that Manix Lake #1 rose to 1780 feet, formed the lower of two shorelines, deposited a thick section of sediments containing Rancholabrean fauna, and may have dried up prior to 47,000 years ago based on carbon 14 age-dating results from the Yale Radiocarbon Lab according to Jefferson (1982, p.11). Manix Lake #2 rose to 1800 feet, carved an upper shoreline in alluvium, buried the earlier lake beds, and was drained perhaps 19,000 years ago (Jefferson, 1982). <u>Anodonta</u> clam shells collected from the 1800-foot bench near Afton Canyon yielded an age of 20,050 years. A more recent date for the draining of Manix Lake is provided by Meek (1989) who proposes that Manix basin was intact at least until 14,230 years ago. A new age estimate on <u>Anodonta</u> found in situ on the highest shoreline indicates that the Manix basin was intact at least until 14,230 years ago, plus or minus 1,325.

The draining of Manix Lake was probably accompanied by erosion of its outlet near the Manix fault. Subsequent downcutting and erosion by the Mojave River has resulted in badlands geomorphology with excellent exposures of the lake and pre-lake sediments as well as the fanglomerate in Afton Canyon. Afton Canyon is described as the finest such water gap of the Mojave Region (Wachter, et al, p. 244). It is probable that the Manix fault was responsible for the zone of weakness that has been eroded to form Afton Canyon, and it is possible that a break along this zone may have taken place while Manix Lake existed, creating a new outlet for the lake which could have cut rapidly into the shattered rock draining Manix Lake rather quickly (Bassett, 1967). On April 10, 1947 movement along the fault produced an earthquake of magnitude 6.2 centered near NE1/4 Section 4, T.10 N., R.4 E., eight miles west of the study area.

MANIX FAULT:

The most important structural feature of the area is the Manix fault zone which is well exposed in Afton Canyon. It is considered to be a major active fault in the east-central Mojave Desert (Bassett & Kupfer, p. 40). Geophysics and surface mapping indicate that the total length of the Manix fault zone is at least 22 1/2 miles with the western terminus at the Calico Mountains and trending eastward, disappearing beneath alluvial cover east of the Afton Canyon (Hamilton, 1982). The left-lateral Manix fault system terminates at the northwest trending right-lateral Calico fault, somewhat as the Garlock fault terminates at the San Andreas fault. Strike-slip movement is suggested by the braided or anastomosing tendency of the fault trace and the presence of high blocks on both sides of the fault (Bassett & Kupfer, p. 41). Left Lateral displacement is suggested by the direction of the fold axes in the Miocene rocks adjacent to the fault zone. The fault zone strikes slightly north of east. South of the canyon the fault separates thick gravel deposits on the north from brecciated slivers of Tertiary volcanic rocks on the south. In Afton Canyon, a zone of intensely fractured, sheared, and altered rock over 1,000 feet wide separates thick Tertiary sedimentary rocks on the south side of the canyon from basement granitic and metamorphic rocks on the north side. The Manix Fault Zone has been described in detail by Keaton (1977).

SPECIAL FEATURES

Beach Bar:

The Afton Road turnoff traverses a beach bar of Pleistocene Manix Lake in the S1/2 Section 1 (T.11 N., R.5 E.). Rounded pebbles can be found just west of the road edge. The bar trends northwest, and another bar in Sections 8 and 17 (T.11 N., R.6 E.) trends north-northwest. Both were formed along the eastern edge of the basin, due to wave action generated by winds from the west.

Palisades:

Vertical walls have been cut into the Pleistocene gravels of the Manix Formation and eroded into impressive palisades. These are especially prominent on the south side of the canyon in the south one half of Section 14 (T.11 N., R.6 E.). The Pleistocene gravels have been consolidated into "fanglomerate" derived from the Cady Mountains to the south. As this alluvial fan was being deposited where Afton Canyon exists today, lacustrine clays were being deposited to the west during the early phase of Lake Manix.

CAVE MOUNTAIN MINERALIZATION:

Both limestone and iron have been produced from deposits just downstream from Afton Canyon and along the south slope of Cave Mountain.

Limestone was mined from the Cave Canyon deposits about 500 to 1,000 feet east of the Afton Canyon ACEC (Area of Critical Environmental Concern). Technically, the limestone is actually marble and is part of the unit mapped by Southern Pacific geologists as "undifferentiated metamorphic-igneous complex". The marble is probably Mississippian in age based on fossils found in similar hills farther east (Bassett, p. 3). From 1906 to the early 1920's "limestone" for use as sugar rock was quarried from the Baxter and Ballardie deposits, then owned by D.F. and D.A. Baxter and A.W. Ballardie. Adjoining quarries were operated during part of the same period by Sugar Lime Rock Company of Los Angeles (Wright et al, p. 173). These old properties, together with adjoining land owned by the Southern Pacific Company, were acquired by California Portland Cement Company (CPC) in 1930. CPC is now owned by Calmat.

Iron mineralization occurs in or associated with shear zones as replacement bodies in either marble or the metamorphic-igneous complex rocks.

Iron ore (hematite and Magnetite) has been mined east of the Afton Canyon ACEC from the Cave Canyon iron deposits (SE1/4 Sec. 11 & SW1/4 Sec. 12, R.11 N., R.6 E.) owned by Calmat. CPC has mined the deposits intermittently since 1930 (Wright, et al, p. 93) and used it as one of the essential ingredients for making Portland cement at Mojave and Colton. The iron oxide minerals, are used to form calcium aluminum ferrite, which helps to reduce kiln temperatures for production of clinker and to increase sulfate-resistance of the finished cement. Clinker refers to the temperature at which partial fusion occurs. Workings include several shafts, adits, and trenches. In 1957 (Gay, p.247) production was given as 50,000 tons every 2 years. In 1984 CPC produced 15,000 tons of which 5,000 tons were trucked to Mojave (Rains, 1985). In 1957 the reserves were estimated as 3,500,000 tons of 52-57 percent iron.

MAGNESITE OUTCROP:

Magnesite occurs in the "Cliffside deposit" in SW1/4 SE1/4 Section 21, T.11 N., R.6 E. The deposit is described by Collier and Danehy (1958 p. 13) as being about 14 feet thick, striking northeast, and dipping 45 degrees to the northwest. Wright et al (p. 159) described the deposit as 30 to 75 feet thick and cropping out for a horizontal distance of 400 to 500 feet (from southwest to northeast). The white carbonate bed was probably initially a lacustrine limestone that has been altered to dolomite and in part to magnesite, the commodity for which the bed was mined (Bassett, 1967). Basalt, perhaps as a sill, according to Bassett, is associated with the magnesite bed, which is separated by a thin mudstone unit from the pre-Tertiary hornblende biotite schist upon which the Tertiary section was deposited nonconformably.

A few carloads of magnesite were shipped in 1917-1918 from the Cliffside deposit (Wright, p. 159). Ore was carried by a 1900-foot aerial tram across the river to the railroad, and shipped to International Magnesite Company, Chula Vista. Resources of 100,000 to 200,000 tons of ore lie above the canyon floor with 30 percent MgO content, but the silica and lime content are high and may present quality problems (Wright, p. 159). Seawater and natural brines are the principal raw materials used for U.S. magnesium compound production (Kramer, p. 104).

CRONESE LAKES:

Cronese Valley is occupied by two playa lakes known as West Cronese and East Cronese. The present environment in Cronese Valley seems to be duplicating, on a smaller scale, an identical environment that existed at a slightly higher elevation quite recently. Present playas sit on and in dissected older playa deposits; present sand dunes blow across eroded dune deposits now partly cemented; present playa lakeshores lie below older shorelines, and older fans are being dissected to yield modern fan deposits. A contributing factor to this situation is the whimsical nature of the Mojave River which can, in flood, drain northward from Afton Canyon into Cronese Valley or eastward into Mojave Sink. Cronese Valley has been flooded in historic times and certainly must have held varying amounts of water since the Pleistocene. Base level changes may be accounted for by such water level changes and erosional changes in basin outlets, but tectonic base level controls are also suggested (Watchter, et al, 1976, p. 271).

Although Mojave River flood water occasionally flows into East Cronese Lake, it only rarely overflows into West Cronese lake. Ground water underflow from the Mojave probably continues to move into East Cronese Valley throughout the year (Burnham, 1955, p. 10). The clays that floor West Cronese Lake have a high salt (surface) content and in other ways seem to be quite different from those flooring East Cronese Lake (Anctil, et al, 1957, p. 7). A well in Section 19, near the center of East Cronese Valley, penetrated 252 feet of sediment, mostly clay and silt. A 67-foot-thick zone of dark blue clay was logged, beginning at a depth of 91 feet. Possibly this clay is correlative in time with Lake Manix (Anctil, et al, 1957, p. 7).

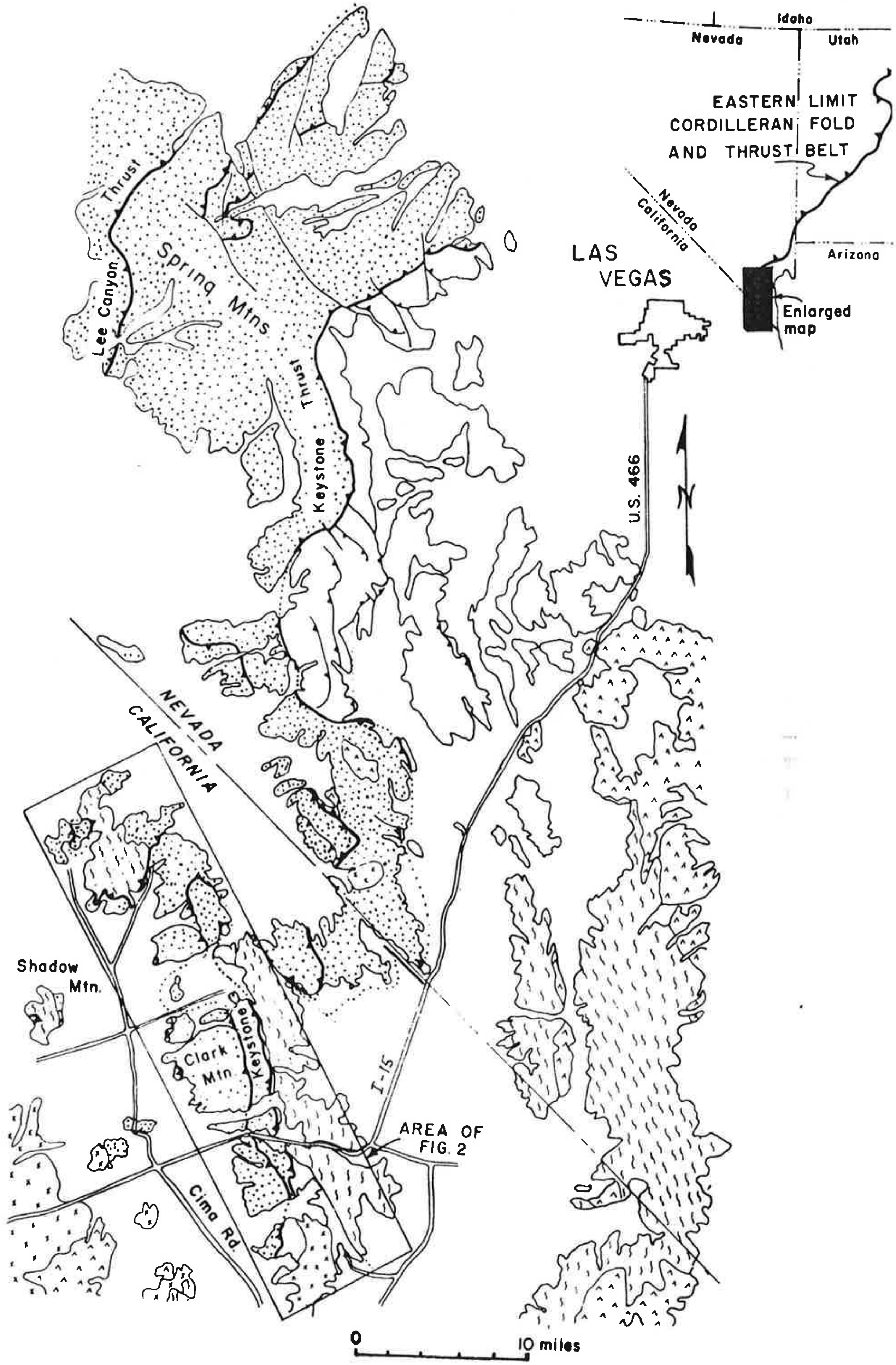
Clark Mountains Thrust Complex Field Trip Stop

The eastern or 'frontal' part of the Cordilleran "foreland fold/thrust belt" passes through the Clark Mountains/Mescal Range area of this field trip stop. This area was mapped in reconnaissance by Hewett of the USGS in the 1920's (not published until 1956), then in much greater detail by Clark Burchfiel (Rice Univ., now at MIT) and Greg Davis (USC) and their students in the 1970's and 1980's.

This thrust complex is noted for its exceptional exposures of Precambrian crystalline basement carried in the higher thrust sheets. These, however, are time consuming to reach. Our stop will be at the more readily accessible lowest, easternmost thrust in this imbricate thrust system exposed at Mountain Pass. This thrust was originally correlated with the Keystone thrust farther northeast, but later recognized to be a separate fault now called the Keaney/Mollusk Mine thrust.

Points of interest that will be discussed are:

- Interplay of thrust fault development with slip on ~vertical faults, leading to emplacement of younger thrust plates over footwalls of varying rock units
- Challenges in dating thrust faults
- The challenge of looking through prevailing hypotheses to objectively assess the field evidence, as exemplified by the reinterpretation of some Mesozoic thrust faults as late Cenozoic normal faults.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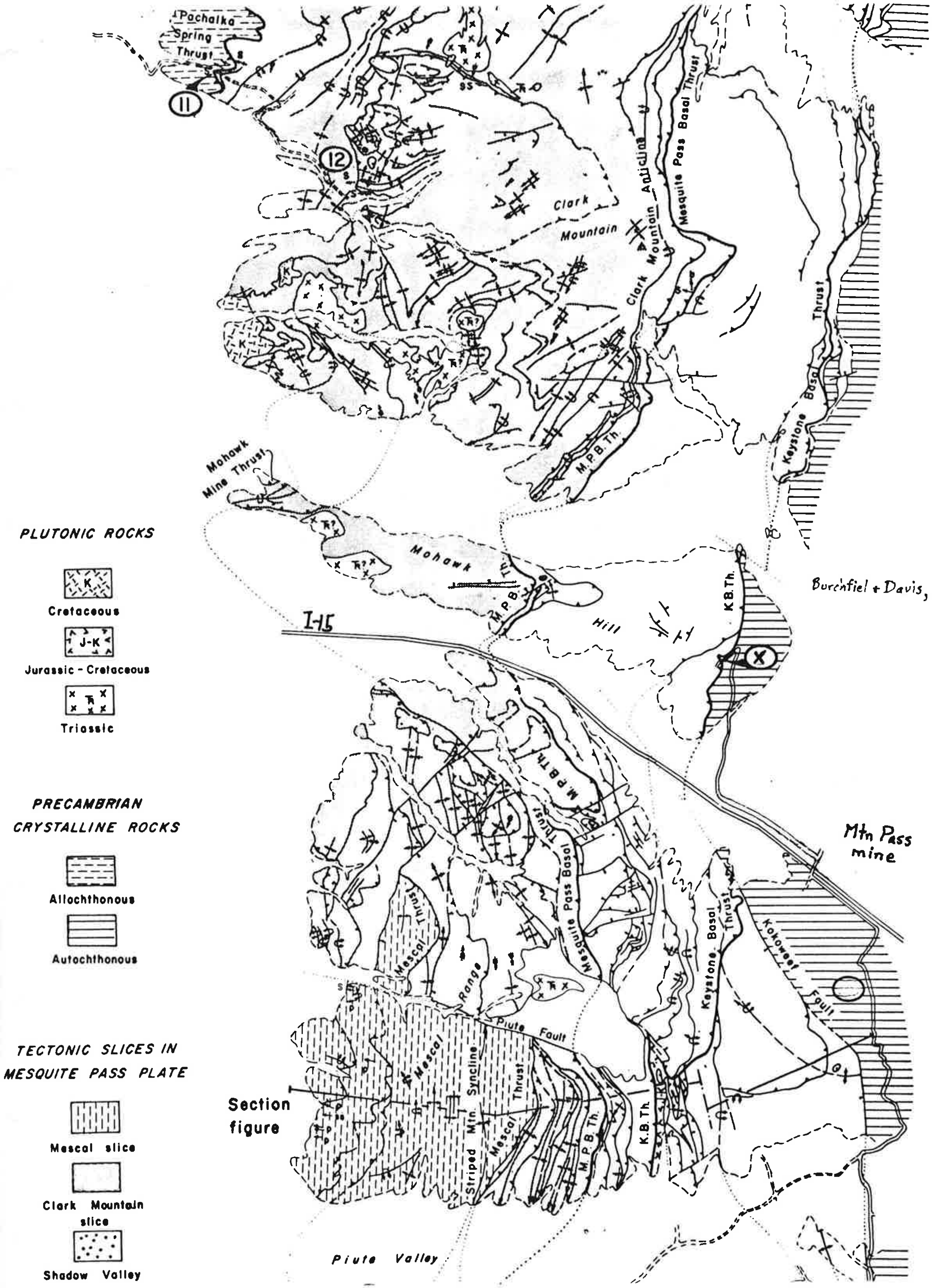
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TECTONIC SLICES IN MESQUITE PASS PLATE
Mescal slice
Clark Mountain slice
Shadow Valley
Mtn Pass mine
M.P.B.Th.
Mesquite Pass Basal Thrust
Mescal Thrust
Mescal Range
Keystone Basal Thrust
Kokoweef Fault
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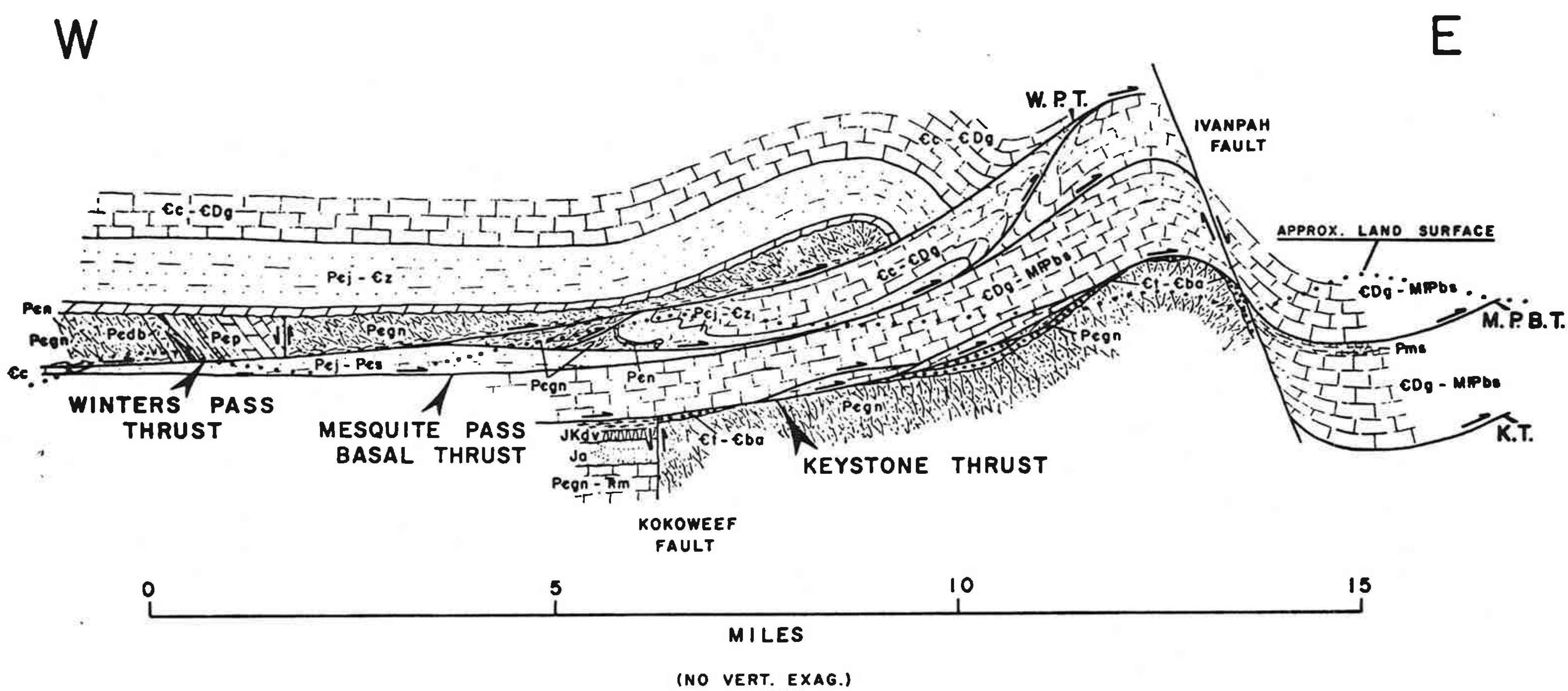
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WINTERS PASS THRUST
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(NO VERT. EXAG.)

Burchfiel + Davis, 1988

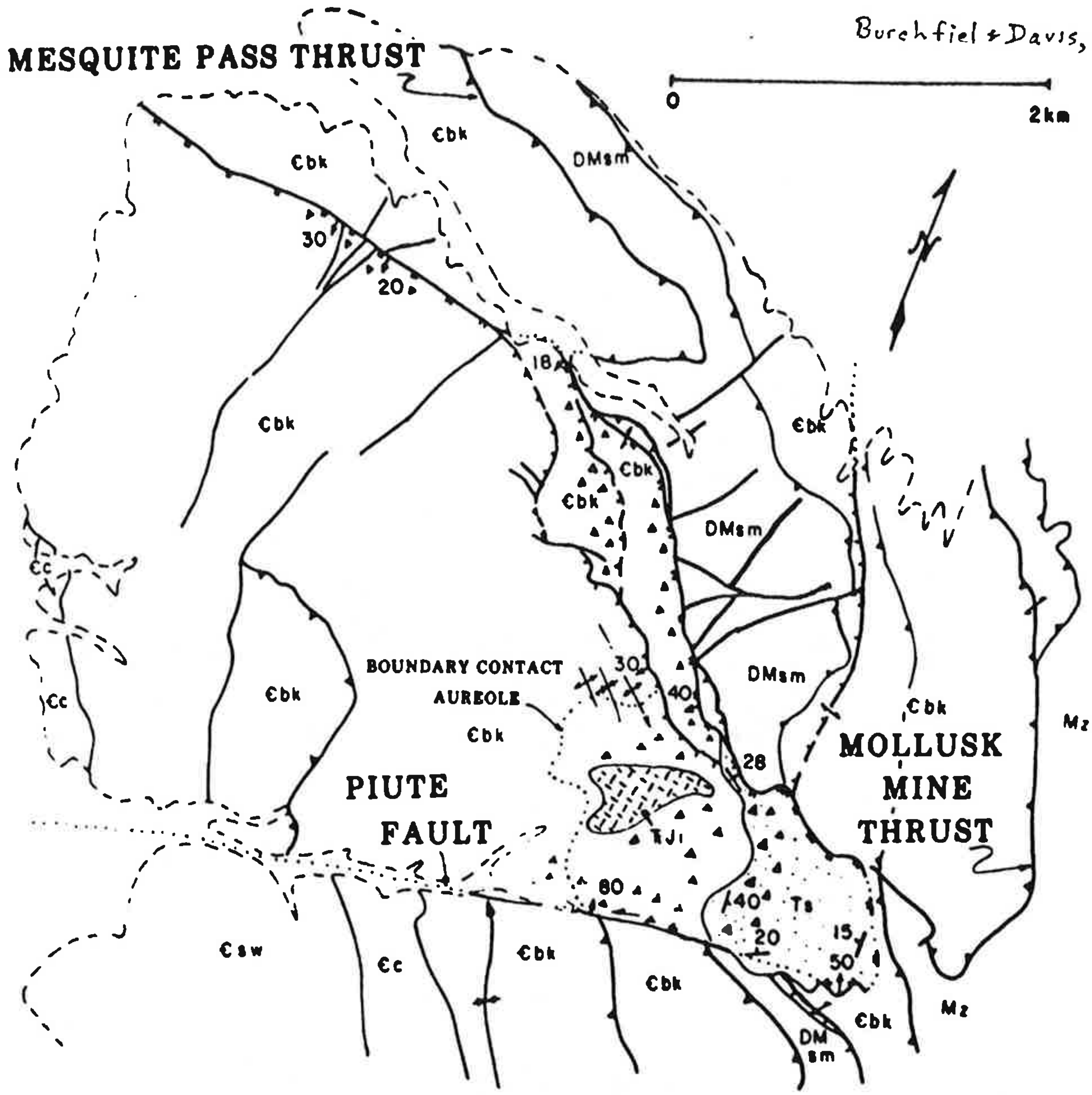

Figure 13. Highly simplified tectonic map of northern Mescal Range, Clark Mountains thrust complex. Interstate Highway 15 lies directly north of the map area (cf. Fig. 9). Most of map area lies within an extensional Cenozoic allochthon that has developed across the Mesozoic thrust belt. The allochthon is bounded on the south by the steep Piute fault (a tear or transfer fault), on the east by a complex, west-dipping breakaway zone, and on the north by a single, low-angle oblique slip extensional fault. The breakaway zone is characterized by (1) the eastward tilting and extreme shattering and brecciation of Bonanza King carbonate rocks (€bk), and (2) by Tertiary conglomerates and sedimentary breccias (Ts) deposited in a restricted breakaway basin. Other rock units (oldest first): undifferentiated Sterling, Wood Canyon, and Zabriskie formations (€sw); Carrara Formation (€c); undifferentiated Devonian and Mississippian sedimentary rocks (DMsm); Triassic/Jurassic intrusion (TrJi); and undifferentiated Mesozoic rocks (Mz).

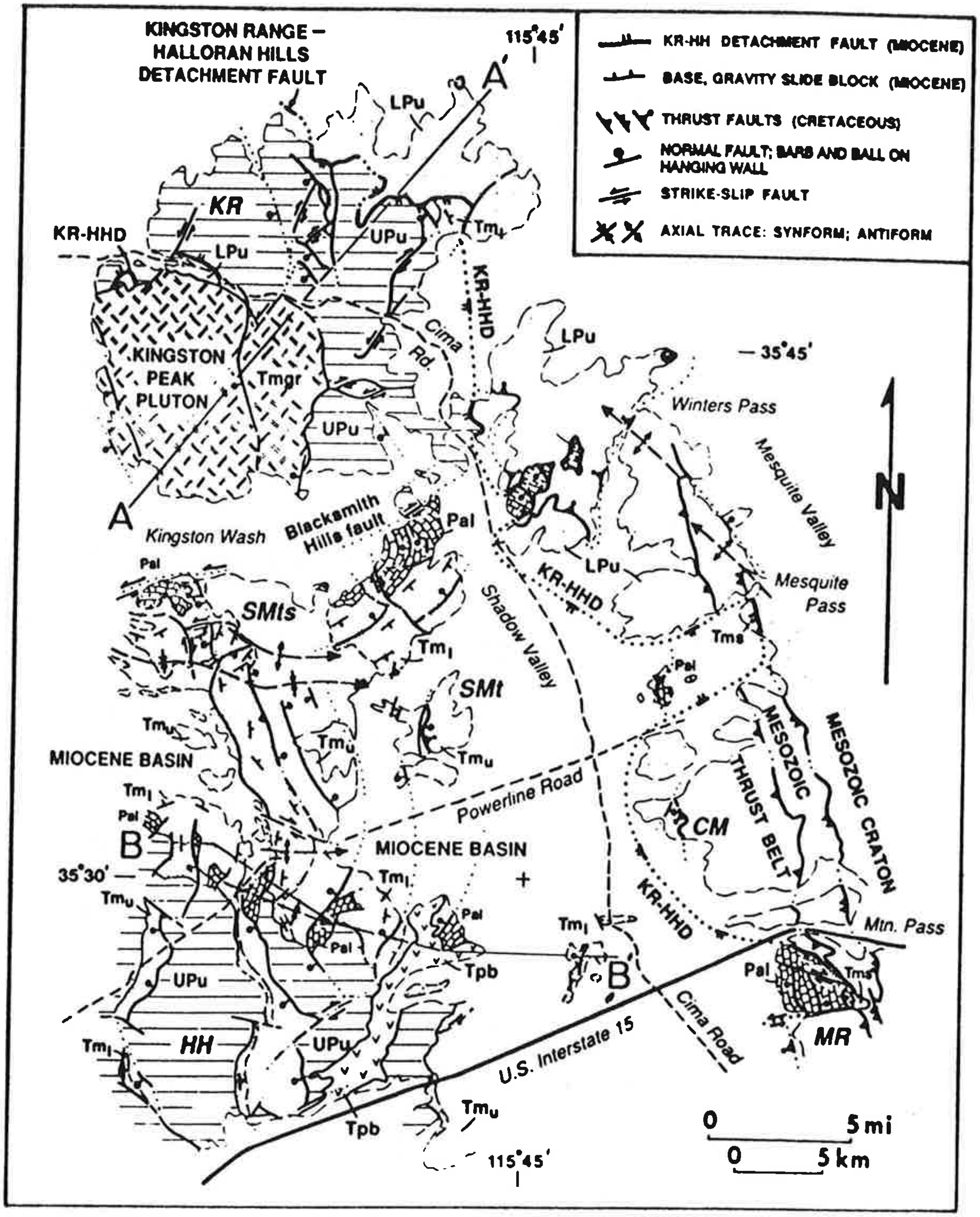

Figure 2. Tectonic map of Kingston Range–Halloran Hills (KR-HHD) detachment-fault system, eastern Mojave Desert. CM = Clark Mountains; HH = Halloran Hills; KR = Kingston Range; MR = Mescal Range; SMt = Shadow Mountain; SMts = Shadow Mountains. UPu (horizontal ruled pattern) = undifferentiated upper plate basement rock units for Tertiary section; LPu (aligned-dot pattern) = undifferentiated lower plate units; Pal (brick pattern) = Paleozoic miogeoclinal carbonate units (Cambrian Bonanza King Formation and younger); Tm_l (dense stippled pattern) = Miocene sedimentary and volcanic rocks of Shadow Valley basin, lower members 1 and 2; Tm_u (sparse stippled pattern) = Miocene sedimentary rocks, upper members 3 and 4; Tms = undifferentiated Miocene sedimentary rocks; Tmgr = Miocene granitic rocks (Kingston Peak pluton); Tpb (v pattern) = latest Miocene–Pliocene basalt. Geologic relations in Kingston Peak area from Calzia et al. (1993); HW = hanging wall.

GEOLOGY OF THE CIMA VOLCANIC FIELD
SAN BERNARDINO COUNTY, CALIFORNIA

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GEOLOGIC SETTING

The Cima volcanic field (Fig. 1), located at the crest of the Ivanpah upland (Hewett, 1956), comprises more than 50 vents and associated flows of small volume in an area of about 300 km^2. Eruption of alkaline basaltic magmas occurred over a period of at least 7.5 Ma with a hiatus between 3 and 1 m.y. ago (Turrin and others, 1985; Wilshire and Noller, 1986). Xenoliths in the basalt contain an exceptional record of upper mantle and lower crustal processes.

The basaltic volcanic rocks overlie Proterozoic metamorphic rocks, lower Paleozoic sedimentary rocks, Cretaceous granitic rocks of the Teutonia batholith (Teutonia Quartz Monzonite) (Beckerman and others, 1982; DeWitt and others, 1984), and weakly consolidated Tertiary continental sedimentary rocks that are locally intercalated with tuff tentatively correlated with the Miocene Peach Springs Tuff of Young and Brennan (1974).

Proterozoic rocks consist predominantly of granitic and mafic gneiss and smaller amounts of schist, amphibolite, and marble. Lower Paleozoic limestone and sandstone crop out only locally within the volcanic field, but more extensive exposures of Paleozoic rocks are found in the adjacent Old Dad Mountains (Barca, 1966; Dunne, 1977). The Proterozoic and Paleozoic rocks are intruded by Cretaceous granitic rocks of the Teutonia batholith. Probably at least three separate Cretaceous plutons occur within the volcanic field: coarse-grained porphyritic granite with abundant phenocrysts of K-feldspar, medium-grained, nonporphyritic granite, and medium-grained granite crowded with mafic inclusions. Moderately abundant apophyses of granite, aplite, and pegmatite intrude older rocks adjacent to the plutons. The Cretaceous granitic rocks are intruded by a northwest-trending swarm of Tertiary andesitic to dacitic dikes (Fig. 2).

Overlying all of the above-described rocks are weakly consolidated, locally derived continental deposits with thicknesses ranging to about 300 m. These deposits are composed predominantly of coarse to very coarse boulder beds, but also include a gravity glide block of brecciated dolomite and, near the base of the section, a rhyolite tuff interbedded with fine-grained sandstone and siltstone. The tuff is tentatively identified as Peach Springs Tuff on the basis of its paleomagnetic orientation (J. Hillhouse, oral communication, 1987). If the correlation is correct, the overlying continental deposits are 18 Ma or younger, the mean age of the Peach Springs Tuff (Glazner and others, 1986). The boulder beds are dominated by granite clasts of the Teutonia that reach dimensions of 13 m. In the thickest known section of Tertiary rocks in the Cima volcanic field (immediately south of highway I-15 beneath capping basalt flows), the gravels are composed almost exclusively of Teutonia clasts with less than 5 percent of clasts representing the dike swarm rock types. However, in the southern and northern parts of the volcanic field, Proterozoic and Paleozoic rocks contributed substantially to the clast populations of the gravel beds. The large, slab-like mass of dolomite breccia is clearly interbedded with the boulder beds in the Cima volcanic field; it may have been derived from the Old Dad Mountains, to the southwest, where similar glide blocks of dolomite breccia, reaching lengths of nearly 2 km, are common (Barca, 1966; Dunne, 1977).

CENOZOIC ALKALINE BASALT VOLCANISM

Basaltic volcanism began at least 7.5 m.y. ago (Turrin and others, 1985). Like other Cenozoic Basin and Range basalt fields, the Cima basalts range from moderately ne-normative to hy-normative compositions (Wilshire and Noller, 1986). Where the basalt magma rose through the continental detrital deposits, phreatic eruptions generally occurred so that initial ejecta consist of maar beds with abundant boulders and cobbles of Teutonia granites mixed with basaltic tephra. Maar beds are associated with vents with ages as disparate as 6 Ma and 0.35 Ma indicating availabil-

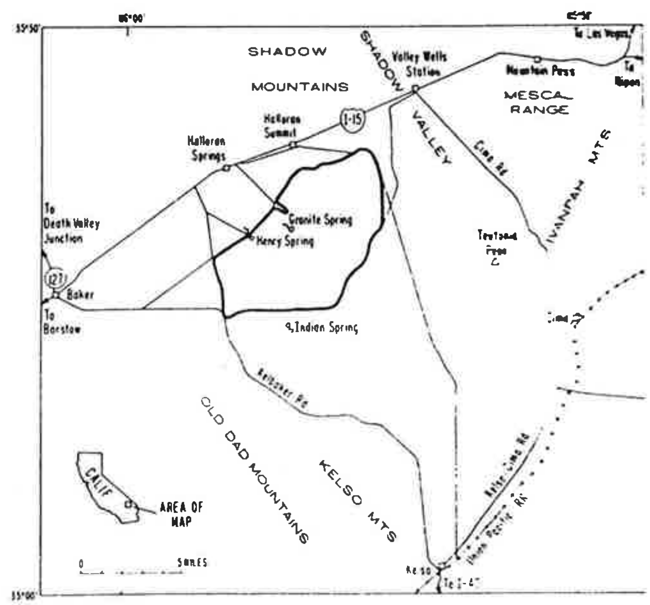

Figure 1. The Cima volcanic field (dark outlines), south of I-15 and east of Kelbaker Road, San Bernardino County, California.

ity of ground water in the Tertiary sediments over a substantial period of time. Other distinctive features of the volcanism are the very long period of time (comparable to the time of development of any one sector of the Hawaiian Ridge [Shaw, 1973]) over which quite small volumes of magma were erupted; the hiatus between 3 and 1 m.y. ago, which was followed by vigorous eruptive activitiy that has not necessarily ceased; and the exceptional record of what was happening in the lower lithosphere before and during this period of magmatism as revealed by xenoliths in the basalts.

RELATION OF STRUCTURE TO TERTIARY MAGMATISM

Widespread coarse-grained terrigenous sediments intercalated with large gravity glide blocks of Precambrian(?) and Paleozoic(?) rocks (Suneson and Lucchitta, 1983) are associated with Miocene extension in the Basin and Range Province (Eberly and Stanley, 1978). Miocene age (equal to or younger than 18 Ma) of these deposits in the Cima volcanic field is tentatively based on identification of Peach Springs Tuff within the sedimentary sequence. A minimum age of Pliocene (older than 4 m.y.) is based on the age of basalts that overlie the sediments with angular unconformity. The Tertiary dike swarm (Fig. 2) in the Teutonia granitic rocks may map the local stress field during extension. When the ages of the dikes are known (they are presently being determined), they will give a minimum age of the onset of extension.

Angular discordance between the Tertiary sedimentary rocks and the alkalic basalts indicates that deformation and substantial erosion occurred between the time of sedimentation and alkaline volcanism. The lag time between the extension recorded by the continental sediments and the eruption of alkaline basalts (about 10 m.y.) is about equal to the duration of the basaltic volcanism (more than 7.5 m.y.).

The xenoliths contained in the basaltic rocks (Katz, 1981; Breslin, 1982; Wilshire and Noller, 1986) consist of mantle peridotite representing the spinel facies (stable at pressures of about 9 to 20 kb); spinel peridotites partly reequilibrated to plagioclase peridotite (stable at pressures below about 10 kb); mantle spinel websterite (Cr-diopside group) and the equivalent rocks partly reequilibrated to plagioclase and olivine assemblages; olivine-free 2-pyroxene pryoxenite, gabbro, and microgabbro; 1-pyroxene clinopyroxenite, gabbro, and olivine microgabbro; and sparse inclusions of the granite country rock.

Composite xenoliths of spinel peridotite (and spinel-plagioclase peridotite) and all mafic rock types indicate that the mafic rocks form dikes in the peridotite. Crosscutting relations and mineral assemblages of the mafic rocks indicate a sequence of emplacement of dikes from spinel websterite to olivine microgabbro, a sequence which also represents decreasing pressure of dike crystallization.

Partial melting phenomena are found in all medium- to coarse-grained mafic and ultramafic xenolith rock types found at Cima, and some microgabbros did not complete crystallization before entrainment in

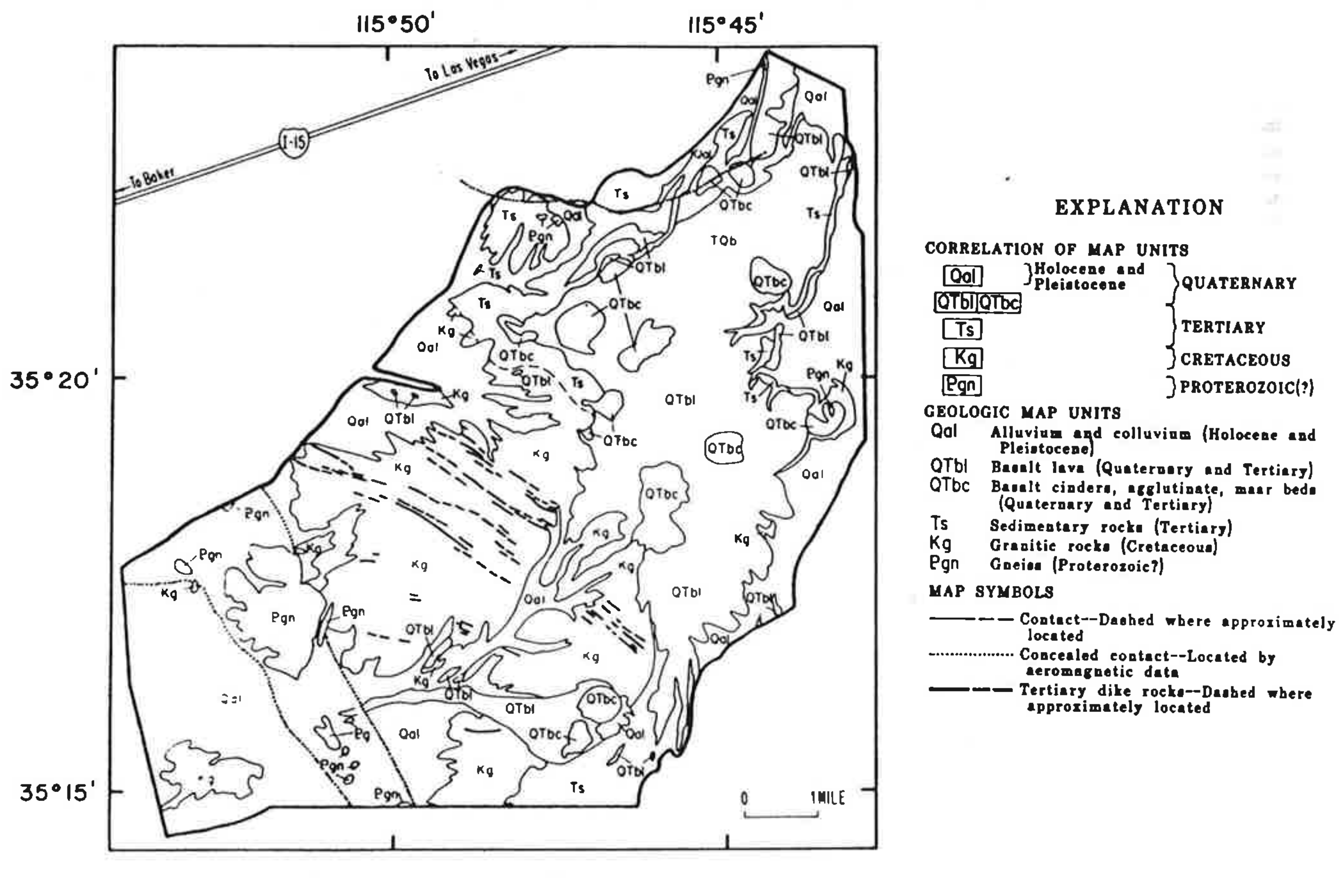

Figure 2. Geologic map of the Cima volcanic field.

the host basalt. The partial reequilibration of spinel facies peridotite and Cr-diopside websterite to plagioclase facies rocks was accomplished by melting and crystallization of the melts before entrainment in the host basalt. The same phenomena are seen in the 2-pyroxene and 1-pyroxene pyroxenite-gabbro suites, in which partial melts (affecting up to 73 percent of the volume of individual xenoliths) crystallized in varying degrees before entrainment in the host basalt. The mineral assemblages in crystallized partial melts of mafic xenoliths represent either the same or lower pressures as those of their parent rocks.

These relations indicate a long history of melting, crystallization of melts trapped in the mantle, and remelting in a situation in which the upper mantle was progressively depressurized. A possible explanation of these phenomena is that the Cima xenolith population reflects the extensional thinning of the crust. Early melting, perhaps of lower crustal rocks, resulted in emplacement of the dike swarm of andesitic and dacitic rocks in fractures directly reflecting the local direction of extension (NE-SW). This was followed by a period of continental sedimentation associated with upper crustal extensional faulting, and finally by a protracted period of time in which the upper mantle responded to the extensional thinning of the crust by melting and probably by upwelling. The magmas generated by this process were mostly trapped in the lower lithosphere, creating new crust (Lachenbruch and Sass, 1978) and, by fractionation, creating the materials necessary for generation of the alkaline basalts of the Cima field (Wilshire, 1987). According to this model, the Cima volcanic field represents the ongoing responses of the upper mantle to extensional thinning that began early in the Miocene.

ACKNOWLEDGMENTS

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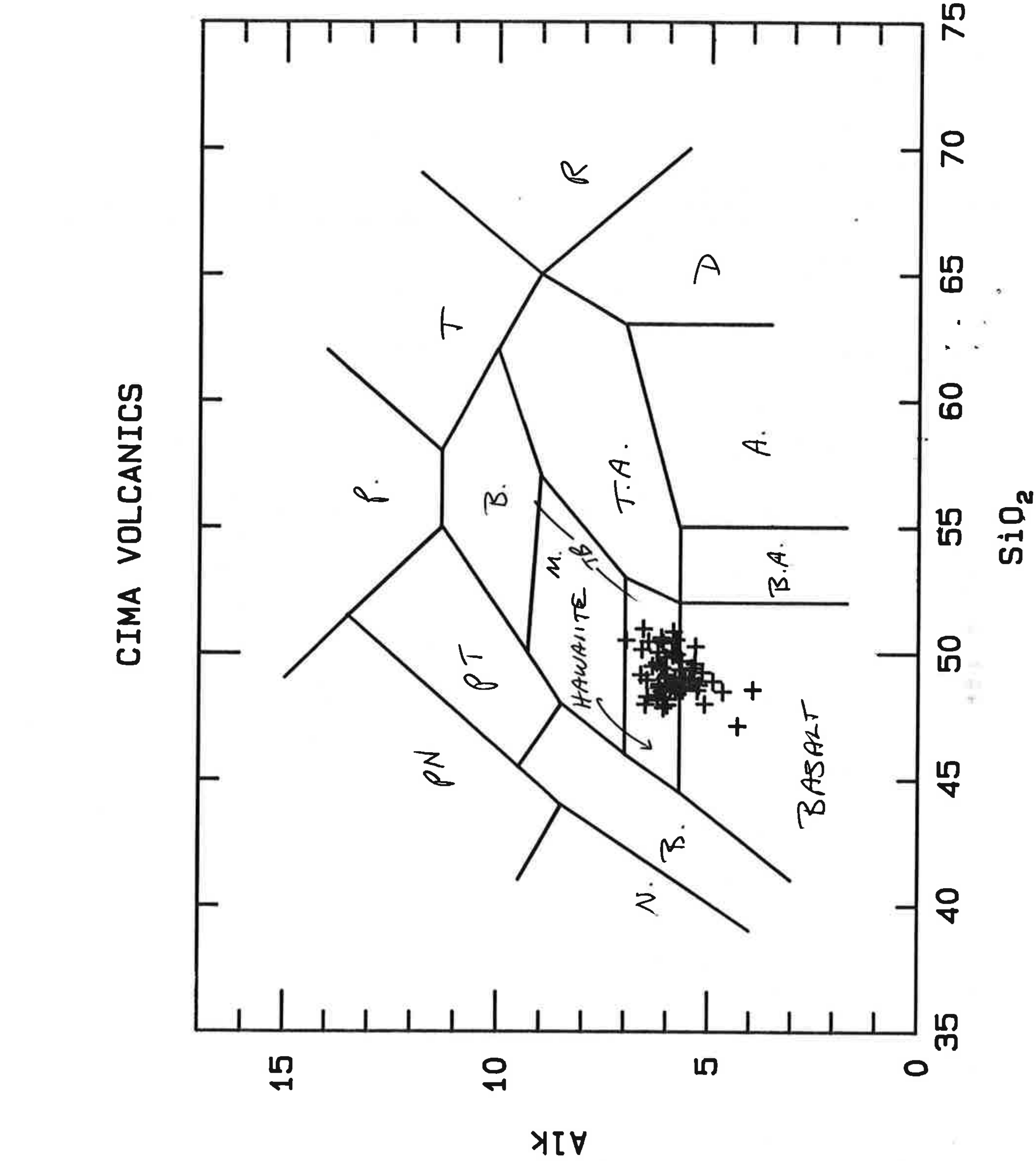

CIMA VOLCANICS
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KELSO DUNES

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INTRODUCTION

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

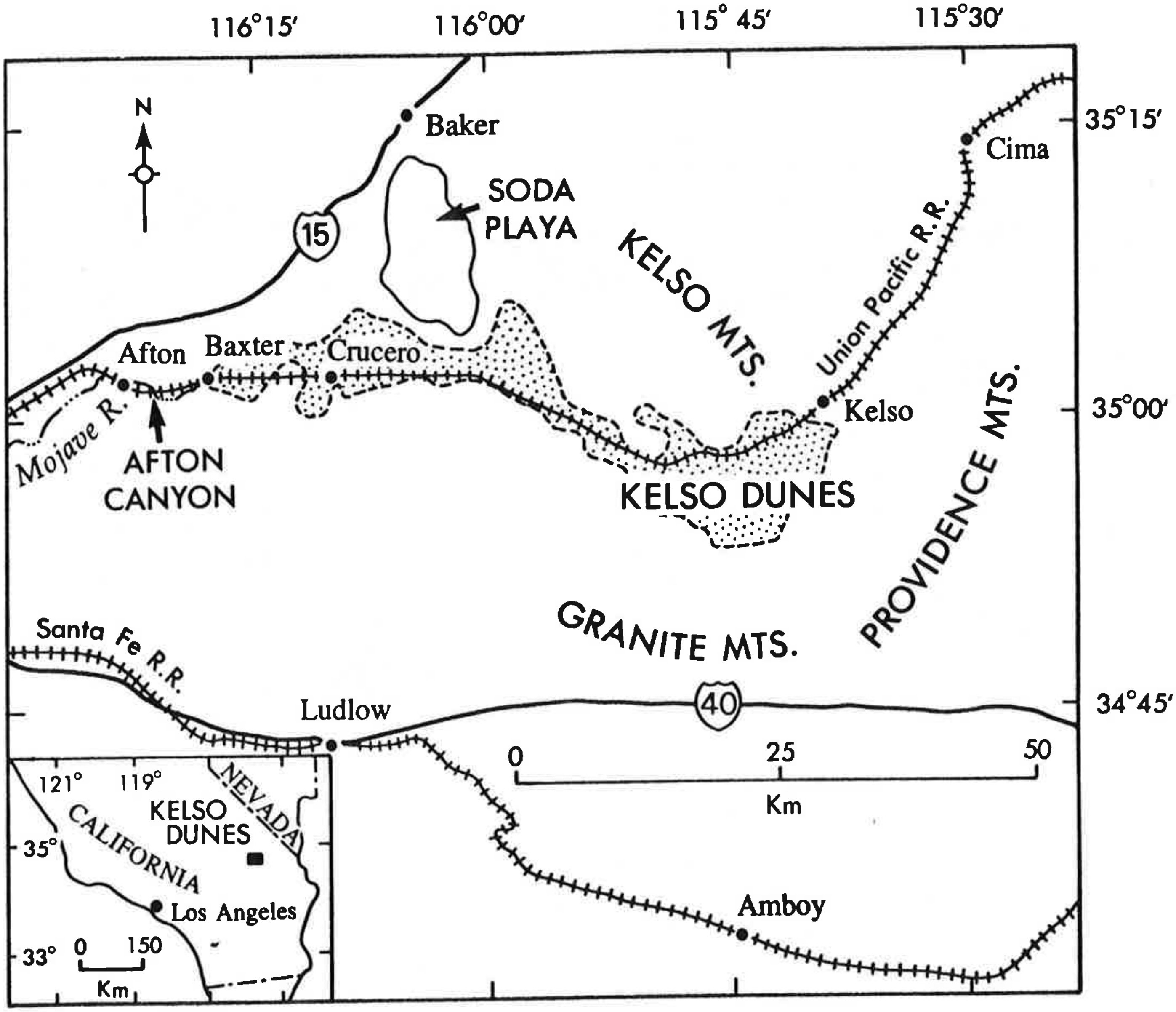


Figure 1. Location map for Kelso Dunes, Mojave Desert, California

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

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

DUNE MORPHOLOGY

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

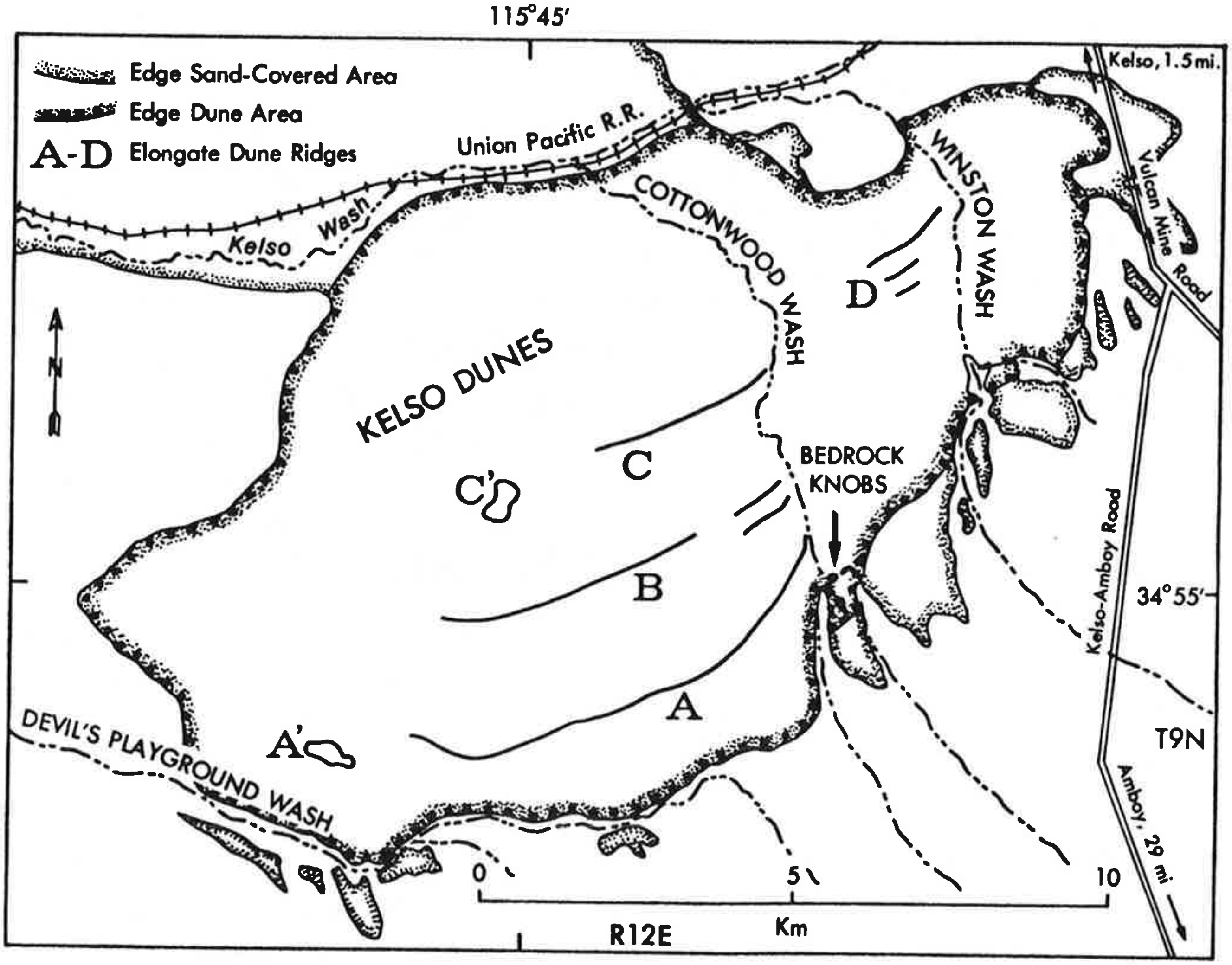


Figure 2. Geographic details of Kelso Dunes.

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

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

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

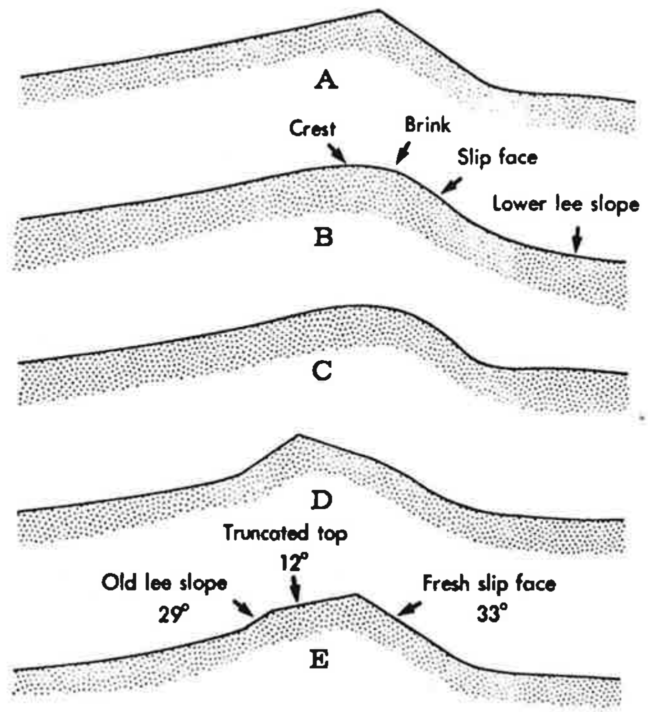


Figure 3. Cross sections through transverse dune ridges.

TEMPORAL CHANGES IN TRANSVERSE DUNES

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

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

INTERNAL BEDDING AND LEE-SLOPE ORIENTATION

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

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

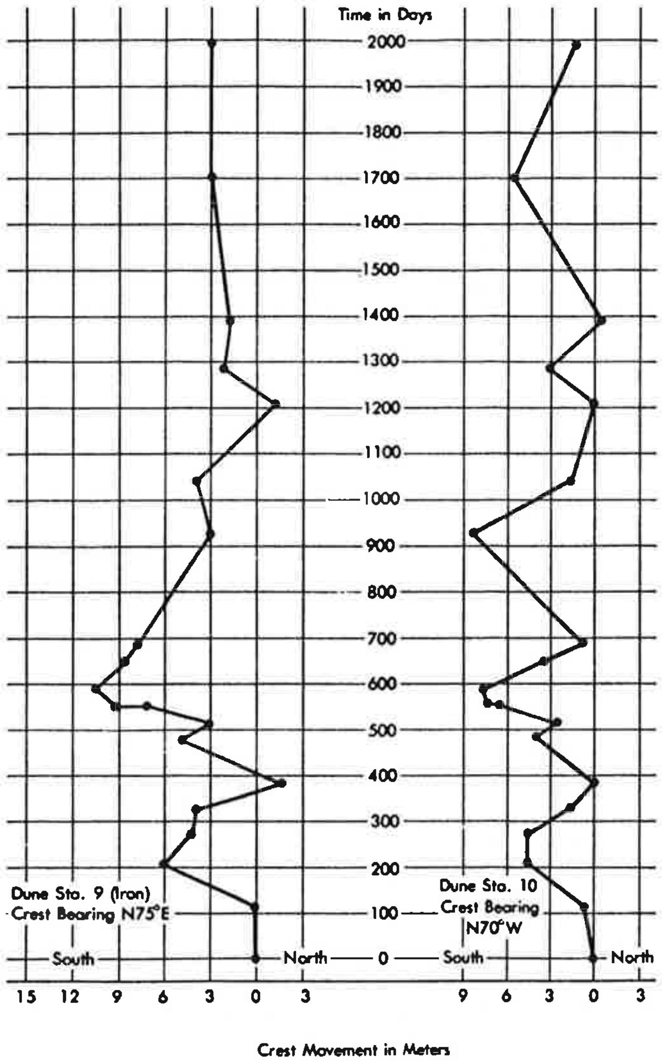


Figure 4. Plot of shifts in position of dune crests at stations 9(iron) and 10.

THINGS TO LOOK FOR

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

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

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

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

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

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

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

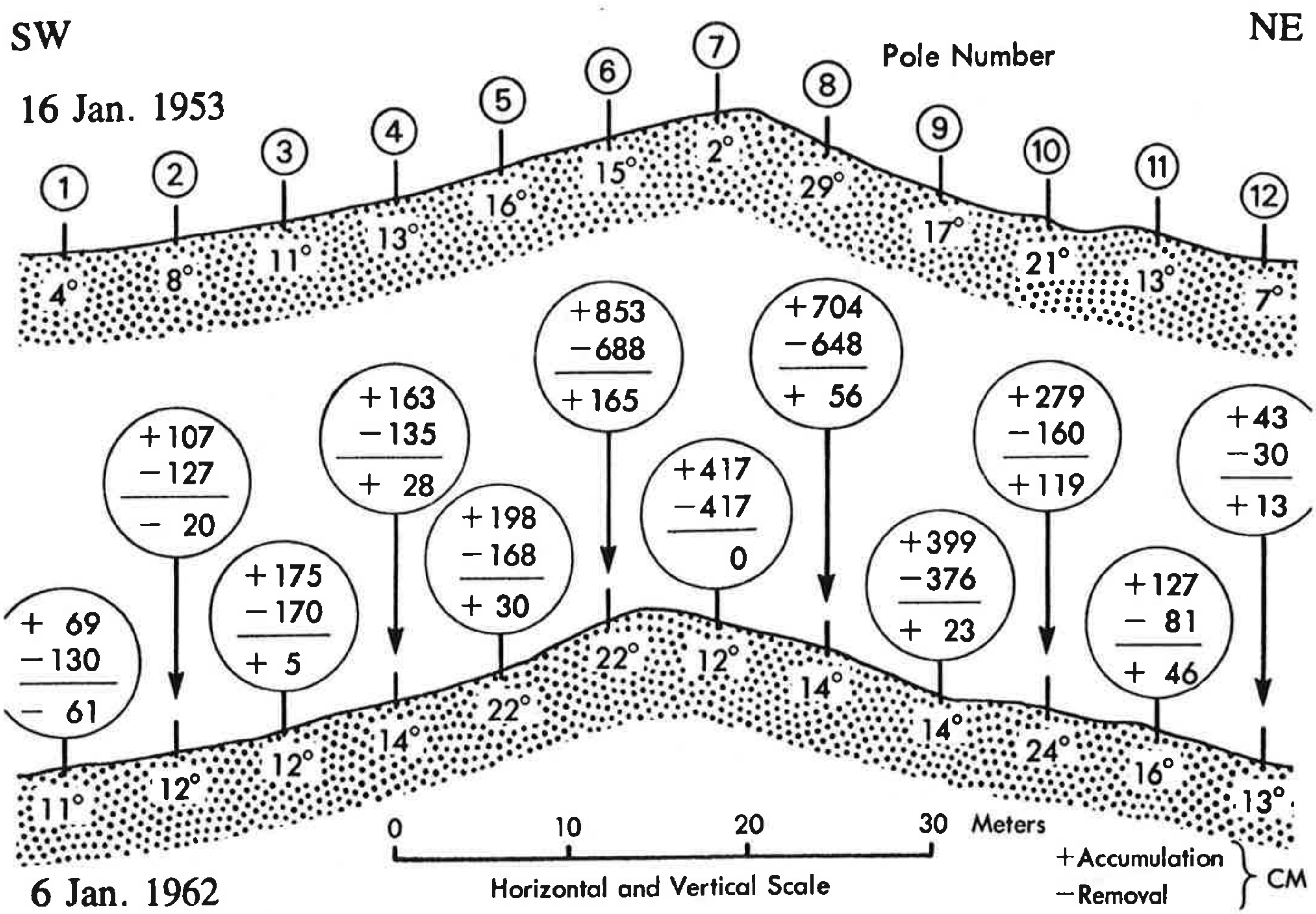


Figure 5. Summation of changes in centimeters recorded at poles of dune station 4 over a 9-year period. Because of intervals between measurements, gross changes recorded are probably only one-third to one-fourth of the absolute changes.

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

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

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

Localities

CAMBRIAN TRILOBITES OF THE MOHAVE

(A Classic Collecting Locality)

by George L. Kennedy

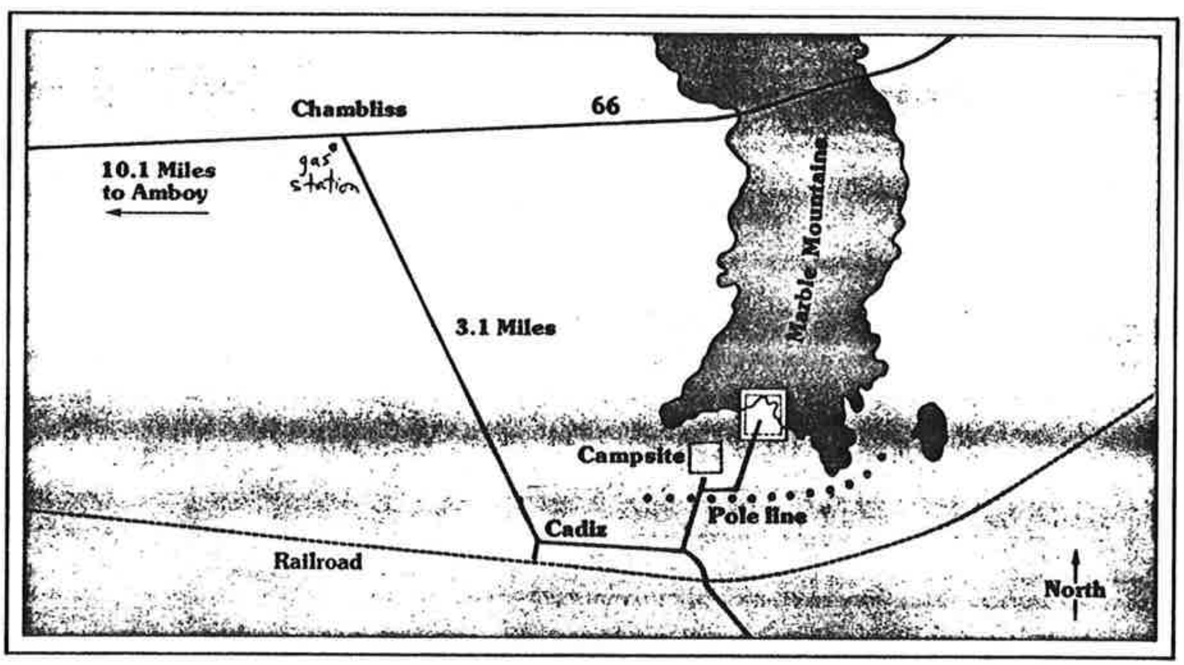
One hundred miles northeast of Los Angeles, California, lies the great wasteland of the Mohave Desert. Today, blistering summer heat in the Mohave and sometimes bitter winter cold can make it hard to imagine that this area was ever anything but a barren moonscape. But 570 million years ago, during the Cambrian Period, there was a warm and shallow sea here, which swarmed with life. The proof of this is to be found at the southern end of the Marble Mountains, where trilobites and other fossils have been collected in abundance for almost 70 years. This area remains one of the best trilobite collecting localities in the western United States, and time spent here is sure to add some fine Cambrian trilobites to your collection.

Getting There

The staging area is Barstow, California, which may be easily reached by interstate highway from Los Angeles, Las Vegas, Bakersfield and other points. At Barstow, take Route 66 (Interstate 40) east to Ludlow, where Routes 40 and 66 separate. Stay on 66 as far as Chambless (10.1 miles beyond Amboy), then turn right (south) on the paved road to Cadiz. After you have gone 3.1 miles south, and just before you come to the railroad siding at Cadiz, the road takes a jog to the left (east). Stay on this road for another mile. Then, just before the paved road takes a jog to the right, turn off to the left on a dirt road that heads northeast. Follow this dirt road 7/10 of a mile, until you reach a line of power poles. At the poleline, turn right (east) on the road that parallels the line. (Note: if you want to camp before starting to hunt fossils, do not turn right at the poleline. Continue north 3/10 of a mile to the best campsite in the area.)

As you follow the poleline east, the road will come to a fork. Take the left (northern) branch of the fork and continue on for another 3/4 mile. When you reach a turnaround area, park your car and proceed for another 1/8 mile on foot. You have entered the only south-facing canyon at this end of the Marble Mountains, and as you walk forward you will see the site directly ahead of you — an outcrop of greenish shale, lying beneath a knob of gray limestone, with peaks on either side.

[A precise description of where you are is the center of the southeast 1/4 of the northeast 1/4 of Section 11, Township 5 North, Range 14 East, San Bernardino Base and Meridian, as shown on the United States Geological Survey, Danby, California quadrangle (1956 edition, 15′ series, scale 1:62,500). Access routes from the west are shown on the neighboring Cadiz quadrangle (1956).]



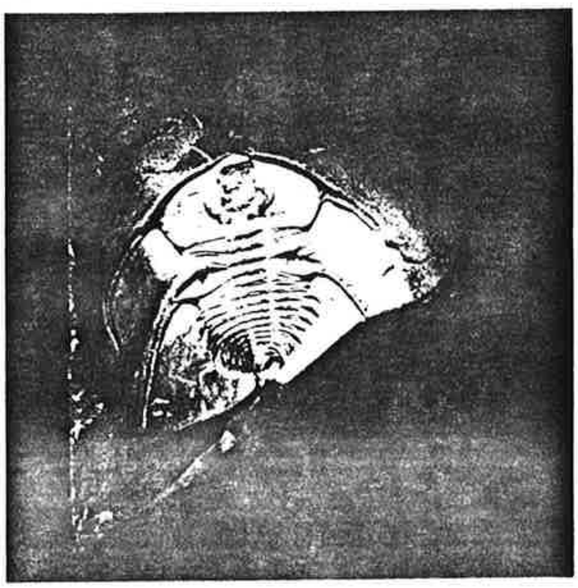
How to Collect

There are two ways to collect here. The easiest is to sit on the talus piles left by others and randomly turn over every piece of shale within reach, looking carefully for fossils that others may have missed. This often turns up small cephalons, as well as an occasional brachiopod (*Paterina prospectensis*), many of which are overlooked by most people as they search for trilobites.

The most productive way to collect, and therefore the most labor, is to work your way down through the bedding surfaces of the shale, layer by layer. Begin by clearing the loose talus from the slope in front of you, as well as the highly weathered top few inches of shale from the rock face itself. Clear an area approximately one and a half feet by two and a half feet square.

Next, take a hammer and a sturdy thin-bladed tool — a broad chisel, putty knife, or paint scraper — and start removing pieces of the platey shale by placing the tool along a bedding plane and striking it with a hammer. Work down your whole surface by removing thin flat slabs of rock one at a time. Examine each side carefully for trilobites, and when you find a fossil, mark the bottom of the slab containing it with a grease pencil or a piece of masking tape, to be sure you don't forget it. Keep pairs together. Newspaper is an excellent wrapping medium for transporting your fossils safely back to the car, camp and home.

Two or more people using this second method of collecting can actually *improve* the locality, by opening a new bench that others can work on later.



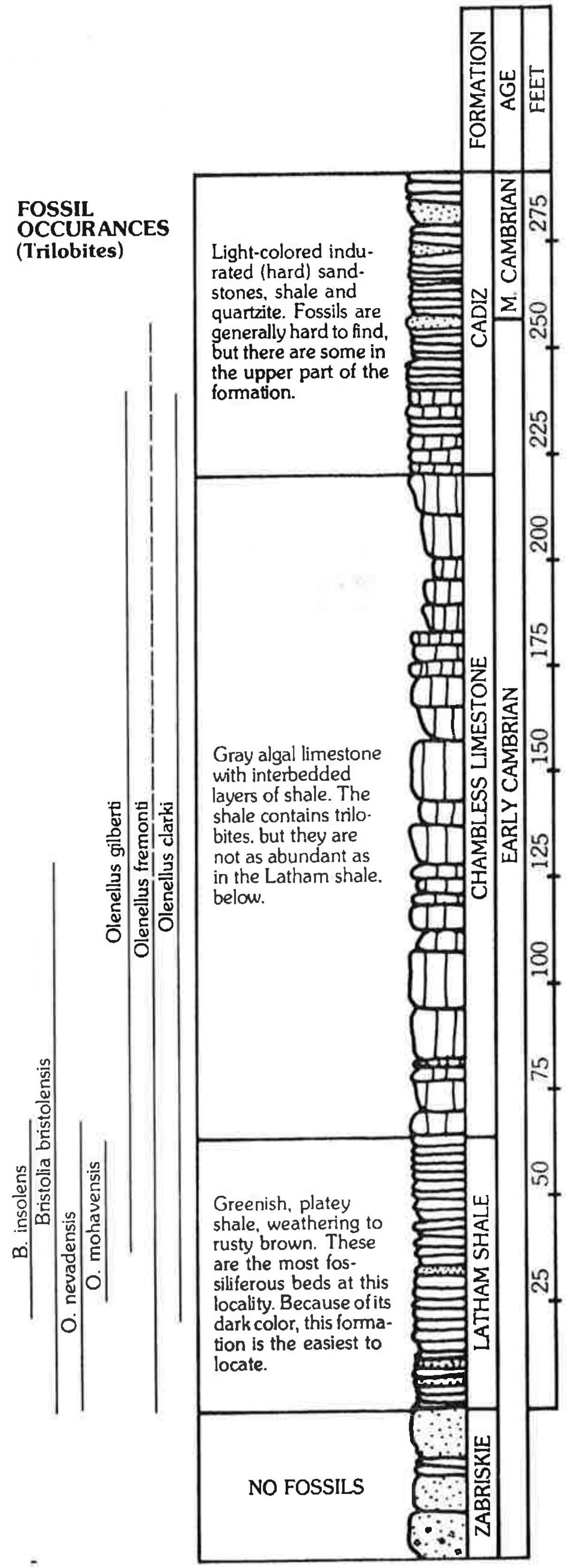
This unusually fine specimen of Bristolia bristolensis (Resser) was recently found at the Marble Mountains locality. It is now part of the collection of the University of California (Riverside). (UCR 10 7)

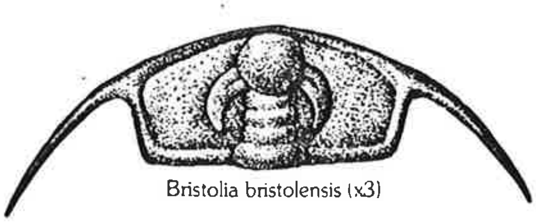
Where to Look

The most fossiliferous beds at this locality are the Latham Shale — a greenish, platey shale that is weathered in places to a rusty brown. The Latham lies directly below a gray algal limestone called the Chambless Limestone, and the green is so easy to detect against the gray that the stratigraphic relationship between the two can be used to locate fossil-bearing beds anywhere in the southern Marble Mountains.

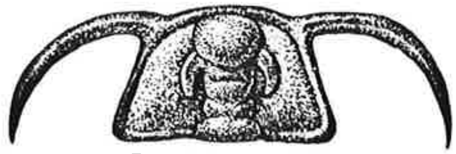
You can also find some trilobites in shale beds within the gray Chambless Limestone, but not in such great numbers. (The dark splotches on this limestone, by the way, are algae and the red patches were once oxidized limy mud.)

Above the Chambless Limestone is the Cadiz Formation, made up of lightly colored indurated (heat-hardened) sandstones, shale, and quartzite. Fossils are rather difficult to find in the Cadiz, but the upper part of the formation does contain some Middle Cambrian trilobites.

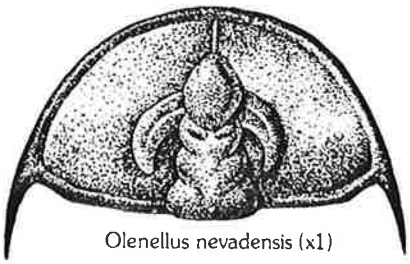




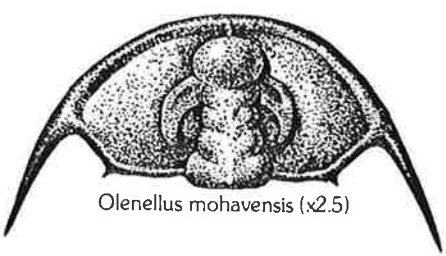
Bristolia bristolensis (x3)



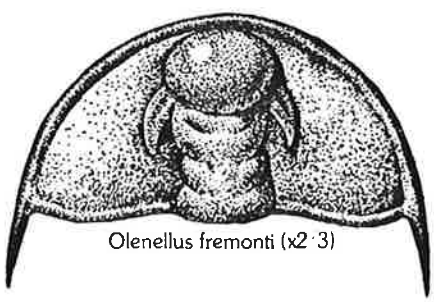
Bristolia insolens (x2)



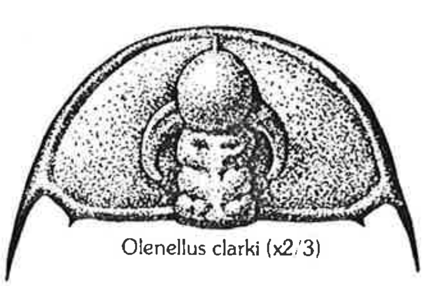
Olenellus nevadensis (x1)



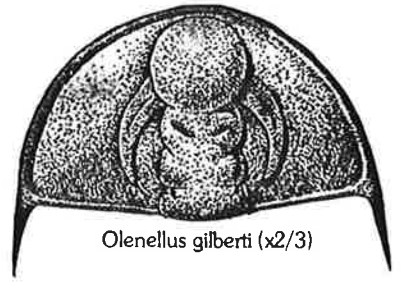
Olenellus mohavensis (x2.5)



Olenellus fremonti (x2/3)



Olenellus clarki (x2/3)



Olenellus gilberti (x2/3)

Amenities and Precautions

In addition to fossil collecting, there are several interesting places nearby to see, including Amboy Crater, Bristol Dry Lake, and Mitchell Caverns. The southern end of the Marble Mountains offers pleasant hiking and photogenic views, especially in the early morning or late afternoon, and there are many interesting rock types for the rock collector. Camping in the spring and fall can be delightful, but the summers are hot and in winter even deserts can get cold. Be prepared for some wind as this is part of any desert climate. This is a dry camp with no facilities, and water and firewood must be brought in. Other things to remember are dark glasses, hat, lip balm, and perhaps gloves. Please carry all of your trash out with you. Shooting is frowned upon. The last half mile to the locality is rocky, but standard sedans will have no problems. Trailers should be left in the flat below, or in the previously mentioned camping area. A gas station and small store are available at nearby Chambless. A small cafe and motel, for the less adventuresome, can be found at Amboy.

References

Darton, N.H. 1907. *Discovery of Cambrian rocks in southeastern California.* Journal of Geology, 15(5): 470-473, fig. 1.

Fritz, W. H. 1972. *Lower Cambrian trilobites from the Sekwi Formation type section, Mackenzie Mountains, northwestern Canada.* Geological Survey of Canada, Bulletin 212: (ix)+1-90, fig. 1-3, pl. 1-20.

Mason, J. F. 1935. *Fauna of the Cambrian Cadiz Formation, Marble Mountains, California.* Bulletin of the Southern California Academy of Sciences, 34(2): 97-119, pl. 15.

Mount, J. D. 1974. *Early Cambrian faunas from the Marble and Providence Mountains, San Bernardino County, California.* Bulletin of the Southern California Paleontological Society, 6(1): 1-5, fig. 1-23.

Resser, C. E. 1928. *Cambrian fossils from the Mohave Desert.* Smithsonian Miscellaneous Collections, 81(2): x-14, pl. 1-3.

Riccio, J. F. 1952. *The Lower Cambrian Olenellidae of the southern Marble Mountains, California.* Bulletin of the Southern California Academy of Sciences, 51(2): 25-49, fig. 1-10, pl. 5-9.

Over twenty different species, from tracks and trails to trilobites, have been reported from the Marble Mountains (see Resser, 1928; Mason, 1935; Riccio, 1952; and Mount, 1974, for lists and figures). The most abundant and well known are the seven species of trilobites shown at left, and one species of brachiopod (not shown), which are found in the Latham Shale.

The three species at bottom are large, and cephalons (or head shields), which are usually all that is found, may be four inches across. Specimens of Bristolia *are smaller, usually less than an inch across, and can be distinguished from* Olenellus *specimens by the long genal spines which protrude laterally from the cephalon.*

Olenellus fremonti *was a recent unsuccessful candidate for state fossil of California. The brachiopod found at this site is* Paterina prospectensis *(Walcott). It is less than half an inch across, round, flattened, and shiney black. A child once aptly described it as a baby phonograph record.*

PLEASE NOTE: unusual specimens of unknown affinity are sometimes found at this locality. These, and any complete trilobite specimens should be taken to a local museum, such as the Natural History Museum of Los Angeles County in Los Angeles, for examination. (Who knows — you may have discovered a new type of fossil.)

STOP #4 - NEWBERRY SPRINGS KLIPPE

Objectives

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

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

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

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

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

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

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

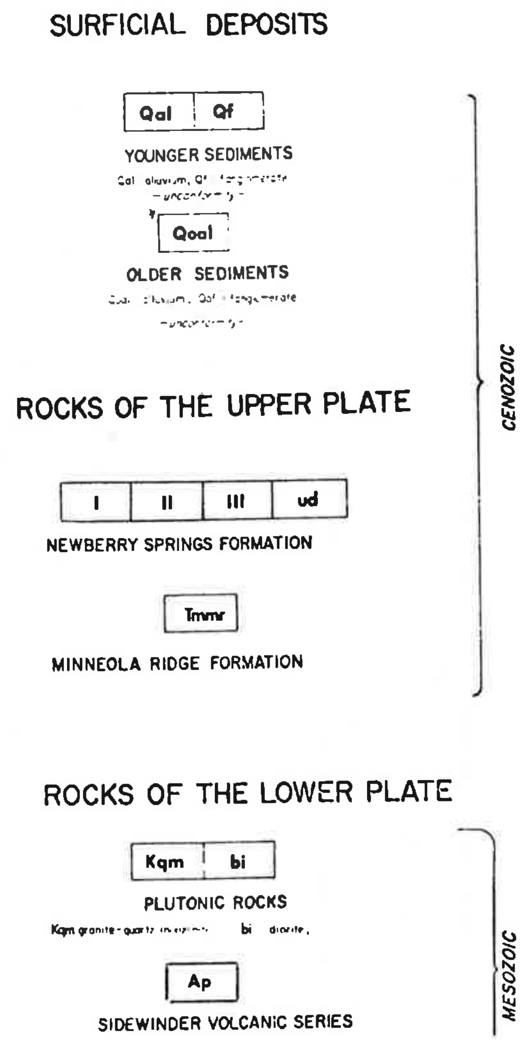
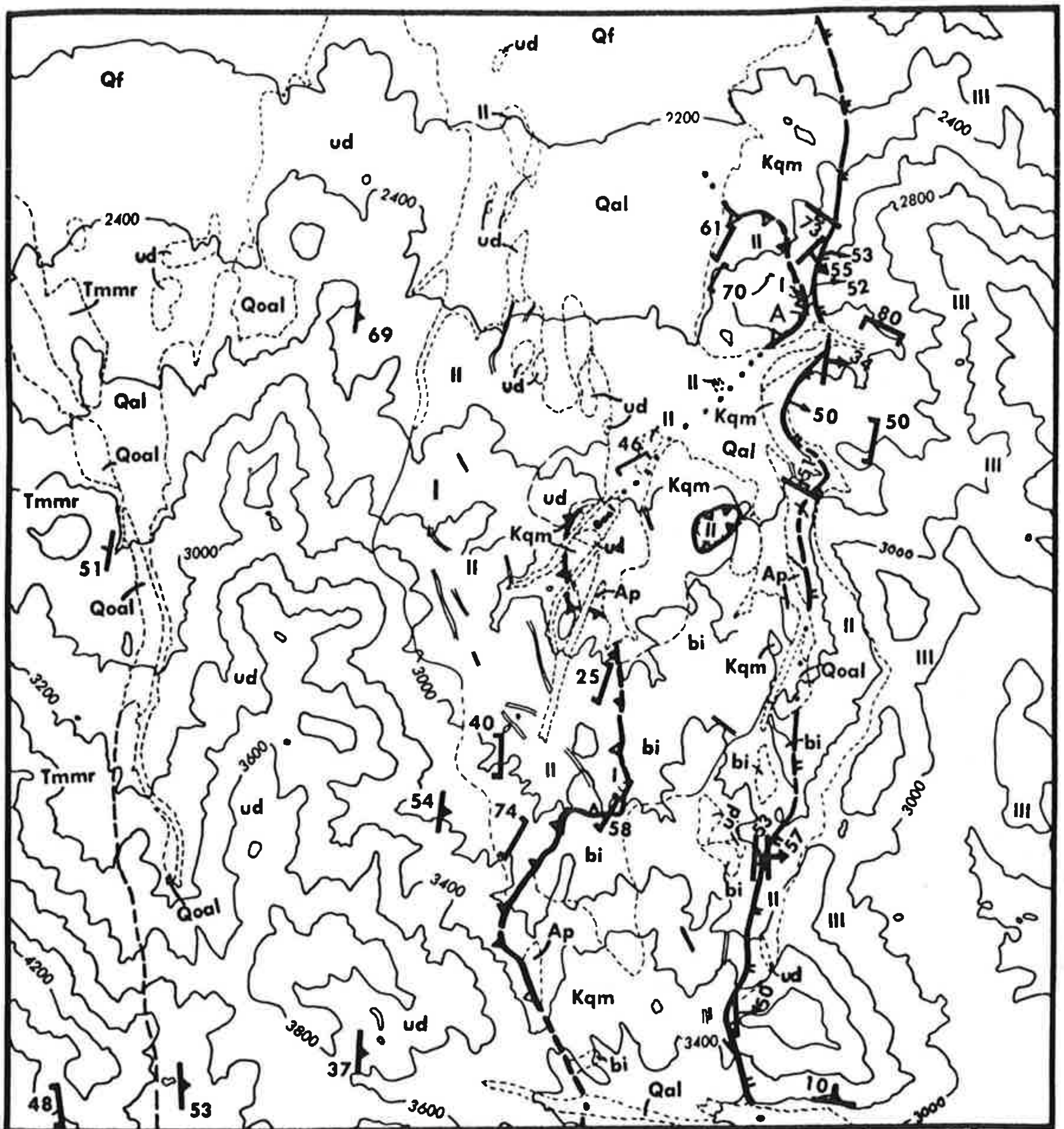


Figure 9. Geologic map of the Newberry Springs klippe area (from Dokka, 1980). Formation names are informal.



Dokka '80

STRIKE AND DIP OF STRATA

CONTACT

dashed where approximate, questioned where queried

STRIKE AND DIP OF FOLIATION

HIGH-ANGLE FAULT

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

STRIKE AND DIP OF JOINTING

LOW-ANGLE FAULT

barbs on upper plate; lower plate generally cataclasized

DIKES

1.5 km

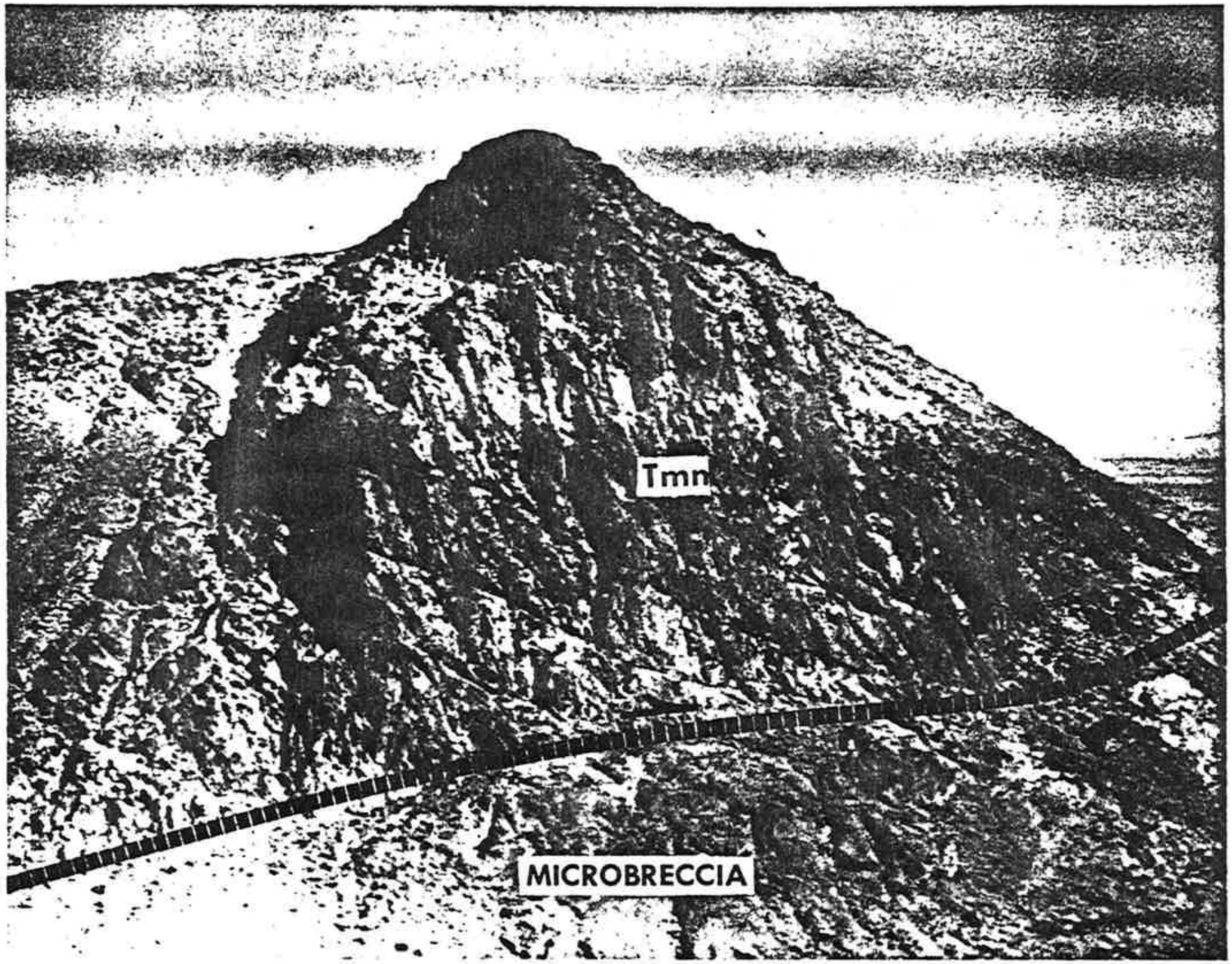


Figure 10. Low-angle fault located near Newberry Springs. View is to the north. Upper plate rocks are dacite flows and tuffs. Lower plate rocks are Mesozoic granite and volcanic rocks that have been cataclasized.

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A GREAT MIDDLE TERTIARY BUTTRESS UNCONFORMITY IN THE NEWBERRY MOUNTAINS, MOJAVE DESERT, CALIFORNIA, AND ITS PALEOGEOLOGIC IMPLICATIONS

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The Newberry Mountains near Daggett in the central Mojave Desert form an east-trending range about 20 miles long. The range is eroded from a sequence some 15,000 feet thick of volcanic rocks and intercalated alluvial deposits of probable early middle Tertiary age. The thick sequence strikes northwest across the range, dips about 30 degrees southwestward, and all but the uppermost part buttresses out abruptly westward against an erosion surface of very high relief cut into a basement complex of plutonic and metavolcanic rocks of Mesozoic age. Volcanic flows and most of the tuff-breccias of the Tertiary sequence are of nearly even thickness and were erupted from many vents and fissures. Fanglomerate forms intercalated wedges and is composed of very coarse alluvial detritus of basement complex exposed immediately south. This sequence accumulated rapidly on lowlands adjacent to a mountainous terrane to the south that must have once been more than 10,000 feet higher than the area of deposition.

The Tertiary sequence is totally absent south of the Newberry Mountains but is extensive in other parts of the Mojave Desert. The great thickness of this sequence of volcanic flows and coarse fanglomerates in the Newberry Mountains and its very abrupt buttressing against a once-mountainous terrane indicates the local severity of the middle Tertiary orogenesis that affected much of southern California.