Dunne

The 21st Annual Fall Field Frolic
California State University Northridge
Department of Geological Sciences
August 20-23, 2003

—— An Overview——

Your Fearless Leaders: Doug Yule, George Dunne, and Gene Fritsche

This year's frolic consists of an east-to-west transect across the Peninsular Ranges Province between Palm Springs and Dana Point. Our general plan of attack is as follows:

(Aug. 20th) Depart CSUN ~2:30 PM and travel to first night campsite (Black Mountain group camp, at ~ 7000 ft) on the north slope of San Jacinto Mountains, with fast-food dinner enroute. Possibly stop for a short hike and overview of Banning Pass if time permits.

(Aug. 21st) Drive to Palm Springs area and take the Palm Springs tram to "the top". Explore upper reaches of San Jacinto Mountains and the Cretaceous plutons of which it is composed. Return to Black Mountain campsite, either via southward loop route if time permits, or backtracking along morning route if time is short.

(Aug. 22nd) Head west across the Hemet/Perris/Elsinor region along Highway 74, examining (1) various components of the Peninsular Ranges batholith; (2) Mesozoic host rocks of the batholith; (3) San Jacinto and Elsinor fault zones; (4) early to middle Cenozoic strata. Purchase fast-food lunch enroute. Continue west along Highway 74 to crest of Santa Ana Mountains, camping there at Falcon Group Camp.

(Aug. 23rd) Drive west to Dana Point/San Onofre area along coast, examining Cretaceous and Cenozoic strata and 'young' faulting. Purchase fast food lunch enroute. Drive north to San Joaquin Hills to examine evidence of uplift possibly triggered by blind thrusting related to Newport-Inglewood fault zone. From San Joaquin Hills area, proceed home via Interstate 5.

GEOLOGY OF THE SAN JACINTO MOUNTAINS AND ADJACENT AREAS

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INTRODUCTION

The San Jacinto Mountains are among the most impressive of southern California because of their great height and proximity to the Salton trough. They are eroded from a greatly elevated mass of pre-Cenozoic plutonic and metamorphic rocks formed deep in the earth's crust but are adjacent to the deeply subsided Salton trough filled with sedimentary deposits to great depths. These contrasting terrains are within the San Andreas fault system along which irresistable lateral movements over millions of years have elevated and depressed segments of the earth's crust to create and juxtapose contrasting terrains such as those present here.

Figures 1 and 2 show the regional geologic setting of the San Jacinto Mountains in southern California, and Figures 3 and 4 show their overall geology. This guide summarizes the geology of this range as known and interpreted by the author.

PREVIOUS WORK

The earliest geologic investigations in the San Jacinto Mountains resulted from the San Jacinto earthquake of 1899, which was described by Claypole (1900). After the San Francisco earthquake of 1906, Lawson and others (1908) found and traced major faults in the San Jacinto region. Another earthquake in 1918 in this region again stimulated geologic investigations of these faults by Arnold (1918), Rolfe and Strong (1918), and Townley (1918). Vertebrate remains were collected from sedimentary rocks in the western part of the region by Frick (1921).

These early investigations were followed by geologic mapping in the San Jacinto Mountains by many geologists and students, as listed by Rogers (1965). The earliest was a geologic reconnaissance of the San Jacinto 30-minute, (1:125,000) scale quadrangle by Fraser (1931). This was followed in later years by more detailed mapping of certain parts of these mountains, many as thesis projects. Few of these were published. Those published include a detailed geologic map and report on the San Jacinto fault zone from Hemet to Borrego Valley by Sharp (1967); geologic map and report of the Lakeview and Perris 7-1/2-minute quadrangles by Morton (1972), and geologic map and report of the Desert Divide area by Brown (1980).

In 1967-68, the geology of the Perris, Banning, Palm Springs, Hemet, and Idyllwild 15-minute, 1:62,500-scale quadrangles, which include the major part of the San Jacinto Mountains, was mapped by the writer, with incorporation of published material mentioned above, as part of the regional geologic mapping along the San Andreas fault system and Earthquake Research Study of the U.S. Geological Survey. These quadrangles with others were used in a regional geologic map of the San Andreas fault system in southern California compiled at

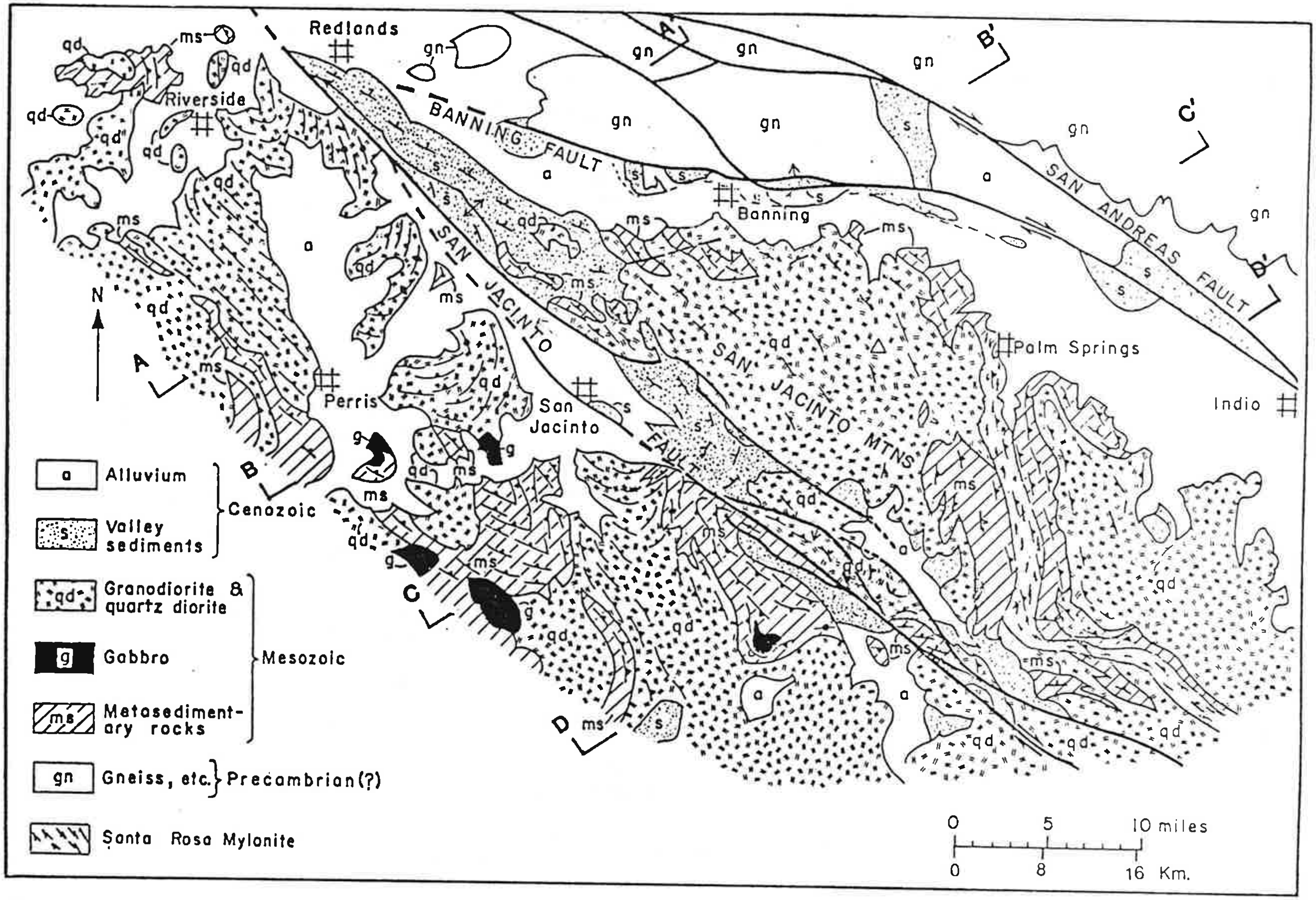

Figure 3. San Jacinto fault zone in the San Jacinto Mountains and vicinity. Modified from Dibblee, 1968.

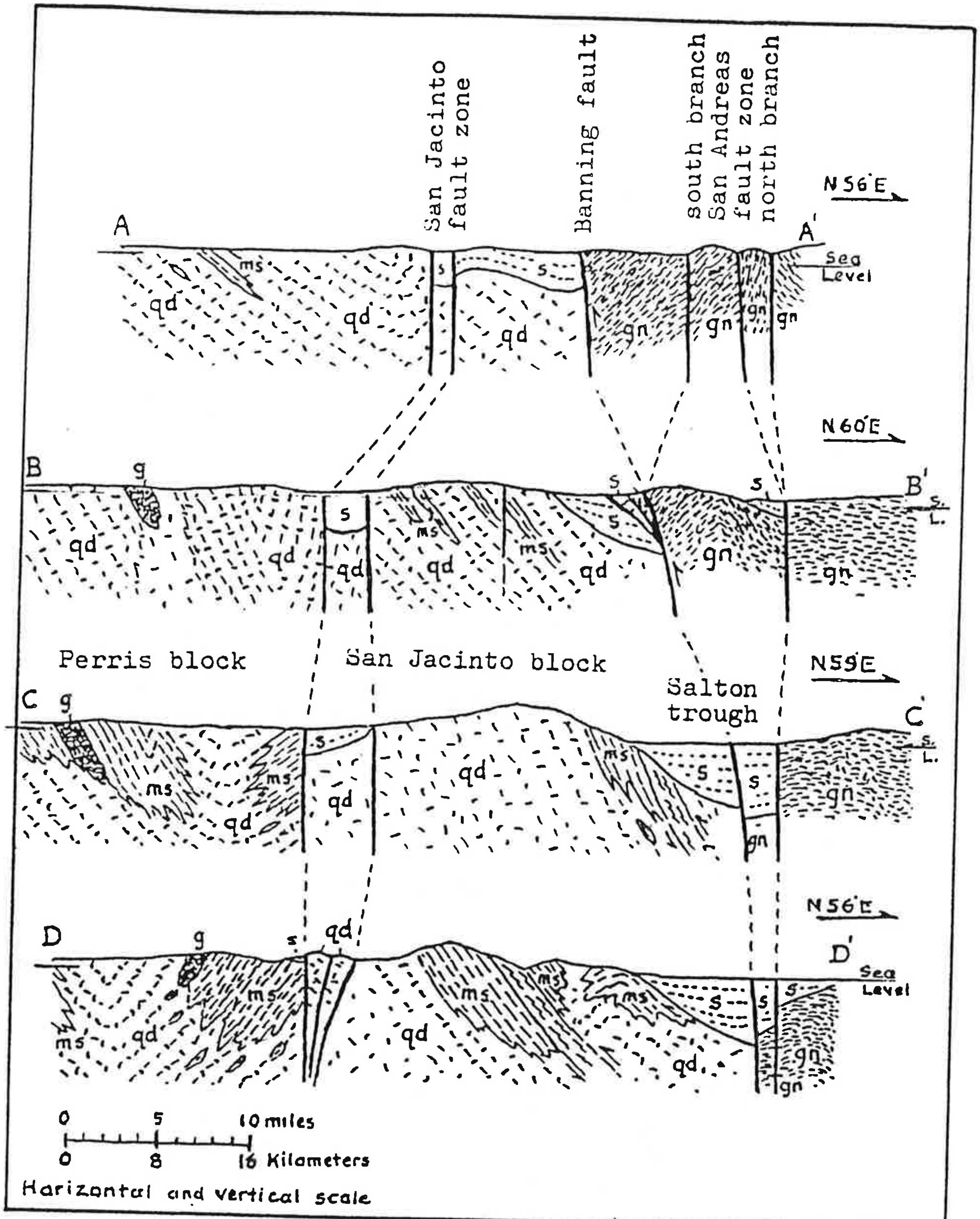

Figure 4. Geologic cross-sections of the San Jacinto Mountains and vicinity. For locations and symbols see figure 3.

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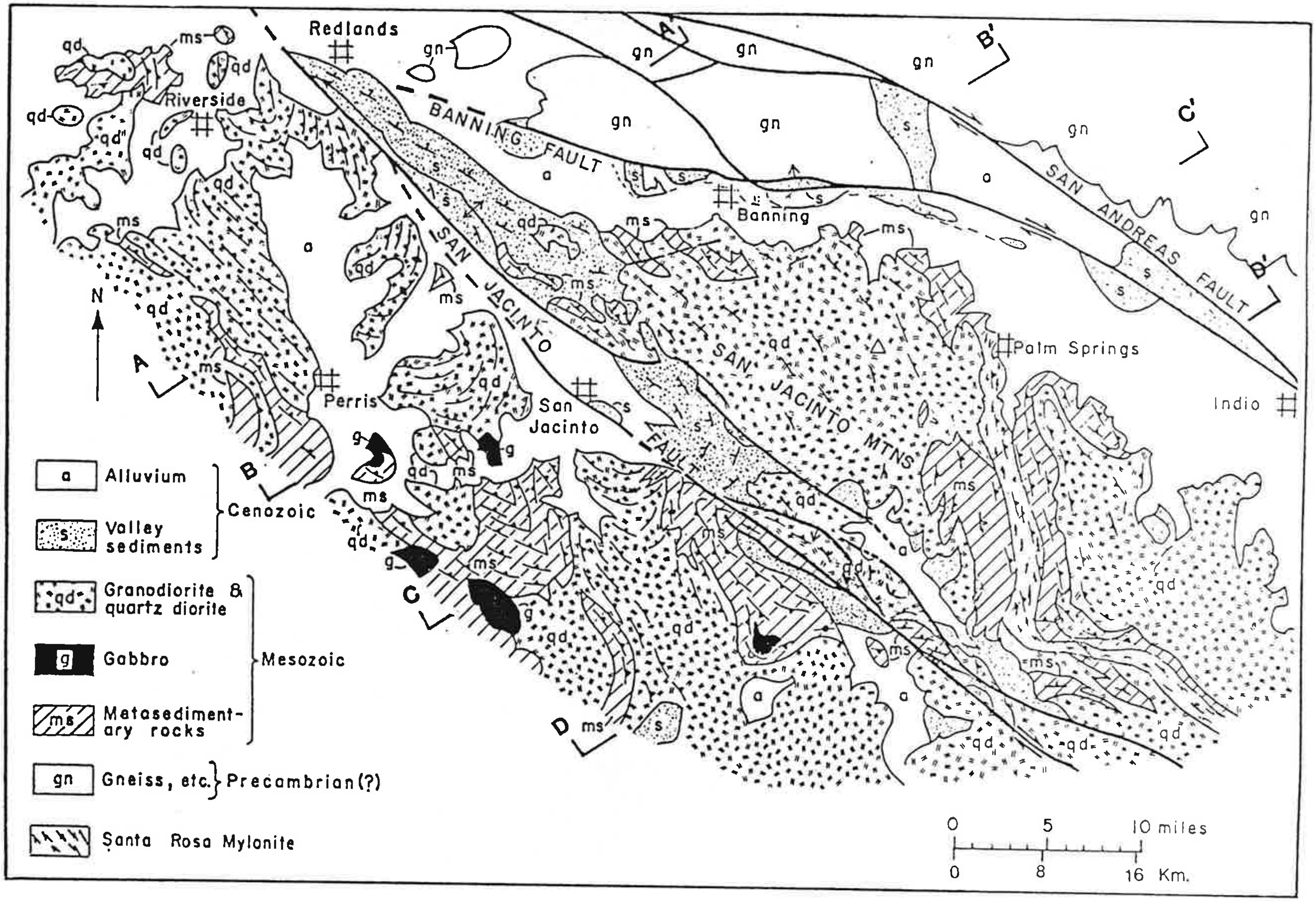

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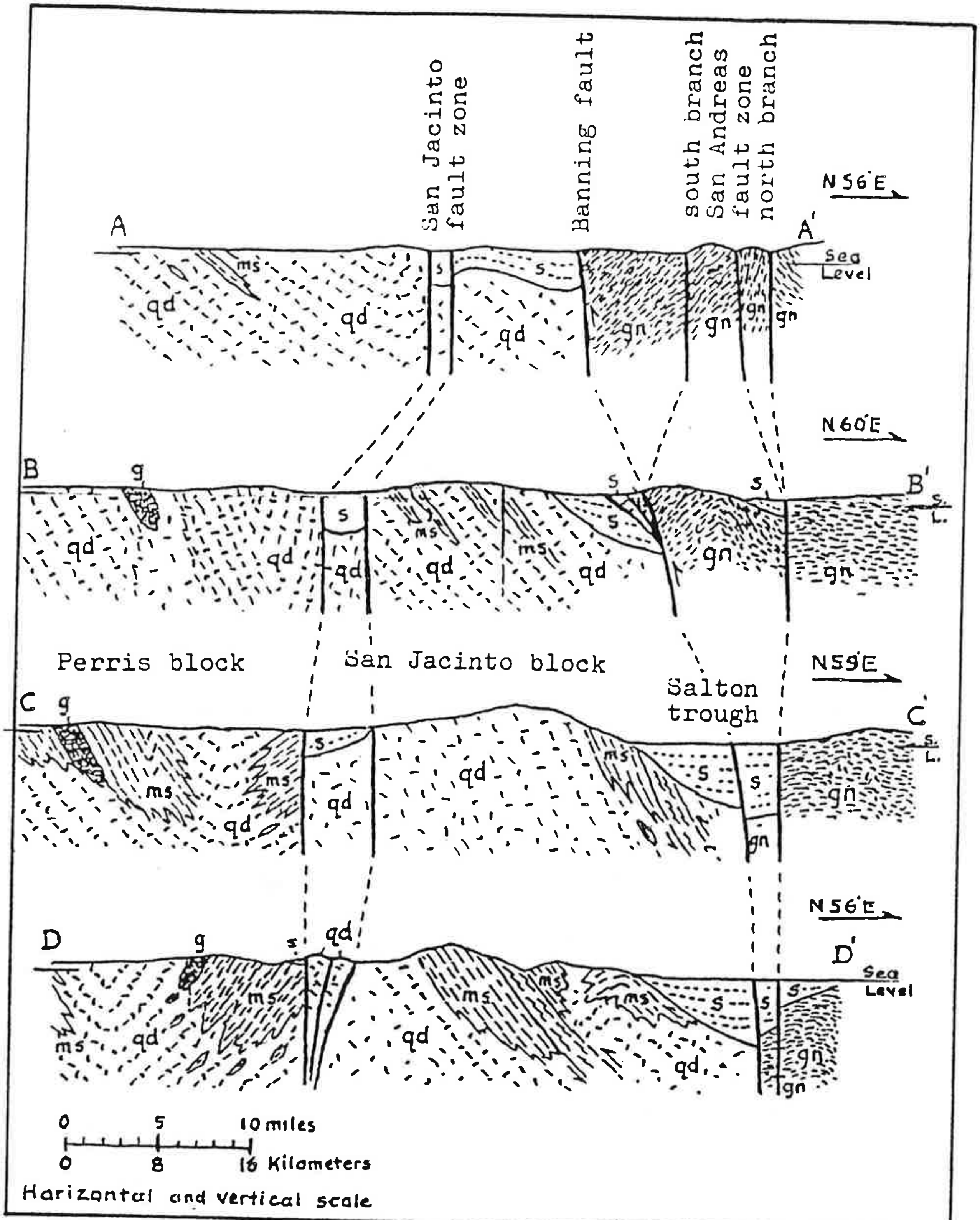

Figure 4. Geologic cross-sections of the San Jacinto Mountains and vicinity. For locations and symbols see figure 3.

scale 1:125,000 and open-filed (Dibblee, 1971). The 5 quadrangle maps, heretofore unpublished, were used in preparation of this guidebook report, and are included with this volume (Dibblee, 1981 a, b, c, d, and e).

PHYSIOGRAPHY

Features

The San Jacinto Mountains form the northernmost and highest of the Peninsular Ranges of southern California, and are composed primarily of a pre-Cenozoic crystalline basement of plutonic and metamorphic rocks.

The San Jacinto Mountains are separated from the Santa Rosa Mountains to the southeast by a dissected alluvial saddle at Santa Rosa Summit. Together the San Jacinto and Santa Rosa Mountains form two segments of a well-defined northwest-trending range of crystalline basement about 60 miles (96 km) long and roughly 15 miles (25 km) wide. Each segment has a single high crest at an average altitude of about 7000 ft (2200 m).

This range rises from the Coachella Valley, the narrow northwestern part of the Salton trough at near sea level which includes Salton Sea at 230 ft (72 m) below sea level. San Gorgonio Pass, which extends westward from Coachella Valley, bounds the San Jacinto Mountains on the north. The San Jacinto-Santa Rosa Mountain uplift is separated from a terrain of low relief, the Perris block, on the southwest by the San Jacinto fault zone, a major splay of the San Andreas fault system (Fig. 2). This mountain block may be defined as the San Jacinto block, with respect to the Perris block on the southwest and the Salton trough on the northeast (Figs. 3 and 4).

It is noteworthy that the San Jacinto block is displaced laterally southeastward along the San Jacinto fault zone from the main mass of the Peninsular Range terrain so that this block protrudes southeastward into the Salton trough (Fig. 2).

Northward from the Santa Rosa Summit the San Jacinto Mountains form a triangular pyramidal highland, as noted by Fraser (1931, p. 502-3), bounded by Coachella Valley and Palm Canyon on the east, San Gorgonio Pass on the north, and the Perris block terrain of low relief on the southwest. San Jacinto Peak, at an altitude of 10,804ft (± 3300 m), forms the center and high point of this pyramidal terrain.

The San Jacinto Mountains are composed of a single high crest of crystalline rocks with a moderately steep southwest flank at higher altitudes. To the southwest and west this steep slope levels off to a lower, more subdued granitic terrain that terminates westward in the San Timoteo Badlands eroded from soft sedimentary deposits.

In contrast, much of the northeast flank of the San Jacinto Mountains is extremely steep and rugged, with as much as 10,000 ft (± 3300 m) of relief facing Coachella Valley in the vicinity of Palm Springs to form the most precipitous mountain front in southern California. This condition suggests that this range was elevated on a fault along its base. However, there are no real surface indications of this inference, because the base of this impressive mountain front, as well as the lower parts of its steep-walled canyons, are buried under the deep alluvial fill of Coachella Valley.

The southern San Jacinto Mountains are partly separated by Garner Valley at altitudes of about 4500 ft (± 1350 m) from the long low ridge of Thomas Mountain elevated as a strip of granitic terrain along the San Jacinto fault zone. The southern San Jacinto Mountains are separated on the east by Palm Canyon from lower mountains that extend northward from the Santa Rosa Mountains.

The Santa Rosa Summit, which separates the San Jacinto and Santa Rosa Mountains, is covered by an old alluvial fan deposit. This and other dissected alluvial fan remnants on both sides of Garner Valley were deposited in this valley when it may have drained out eastward at this location toward the Coachella Valley.

The Santa Rosa Mountains have a single high crest reaching an altitude of 8716 ft (± 2600 m) at Toro Peak. This range has a steep, abrupt southwest flank and a broad, locally rugged but comparatively more gently sloping northeast flank. This range physiographically appears to be a northeast-tilted block of crystalline basement.

The Perris block is a terrain of crystalline basement of generally low relief between the San Jacinto fault zone and the Elsinore fault zone to the southwest (Fig. 2), as defined by Dudley (1936), who gave a detailed account of its geomorphology and genesis.

Part of the Perris block, or that part between the Box Springs Mountains near Riverside and Sunnymead, and the low mountain terrain southeast of Winchester and Hemet, is an alluviated valley terrain that includes isolated small mountains or hills of basement rocks (Figs. 2 and 3). The alluviated area west of these hills is known as Perris Valley, and that east of them as San Jacinto Valley. The alluviated terrain, with an average altitude of about 1500 ft (± 500 m) drains southwest to Lake Elsinore. The hills within it rise to an altitude of about 2500 ft (± 800 m). Some, such as the Lakeview Mountains, have subdued summits. The Box Springs Mountains separate this valley terrain from the San Bernardino Plain to the northwest. Southeastward from Winchester and Hemet the basement terrain rises gradually to the plateau surface of the Peninsular Range terrain.

San Gorgonio Pass is a narrow east-trending alluviated valley between the San Jacinto and San Bernardino Mountains (Figs. 2 and 3). Westward it merges into the Beaumont Plain and San Timoteo Badlands where it is dissected by west-draining San Timoteo Canyon. This valley is highest near Beaumont, with an altitude of about 2500 ft (800 m), and is the lowest part of the major drainage divide between the Pacific Ocean and the Salton Sea - Gulf of California. Eastward from Beaumont, San Gorgonio Pass slopes toward and merges into the Coachella Valley.

Geomorphology

The geomorphology of the San Jacinto block indicates that this basement terrain was elevated in at least three stages. The remnant of an old, subdued erosion surface is evident on the highest part of the San Jacinto Mountains between San Jacinto Peak and Tahquitz Peak, 4 miles (6 km) south at altitudes of more than 7000 ft (± 2200 m), to form a high plateau. Elsewhere this elevated old surface has been destroyed by erosion, but small remnants may occur on the high summit of the Santa Rosa Mountains, and on the summits of some lower mountains on their northeast flank. This old erosion surface was developed during an interval of erosion following the first stage of elevation.

A more extensive, elevated, subdued erosion surface is developed at lower altitudes, mostly below 5500 ft (± 1800 m) on the broad lower southwest flank of the San Jacinto Mountains. It forms somewhat of a bench that terminates northeastward against the moderately steep southwest flank of the high crest of the San Jacinto Mountains, as noted by Fraser (1931, p. 503). This elevated surface of low relief includes the summits of the lower western San Jacinto Mountains and San Timoteo Badlands, summits of dissected high alluvial fan segments around Garner Valley, and summits of the Thomas Mountain uplift. On the northeast flank of the San Jacinto Mountains, this lower erosion surface, if ever developed there, has been destroyed by erosion.

Other extensive remnants of this elevated lower erosion surface are developed on the lower northeast flank of the Santa Rosa Mountains, where it forms broad expanses of low relief cut on basement terrain. This surface slopes toward Coachella Valley around small rugged or flat-topped mountains that protrude above it. This surface of low relief is readily visible along the Palms to Pines highway (State Highway 74) along Pinyon Flat and on both sides of Deep Canyon which cuts a deep gorge through this elevated surface. South of Pinyon Flat part of this surface is covered by thin older alluvium, which is now dissected.

This extensive lower surface of subdued relief was formed during a long interval of erosion that followed the second stage of uplift of the San Jacinto block. This erosion interval was followed by the third stage of uplift during which this terrain was elevated to its present height. As a result, this elevated lower erosion surface is now being deeply dissected and largely destroyed by canyons that are being graded to the present base levels of the lowlands on each side. Because the base level of Coachella Valley is much lower than that of the Perris block, the canyons on the northeast flank are deeply incised. This is evident on the northeast flank of the Santa Rosa Mountains on which the elevated lower east-sloping erosion surface is deeply dissected by Deep Canyon, Palm Canyon, and their tributaries. The Cenozoic sedimentary deposits on the San Jacinto block have been severely eroded to badlands as a result of this latest uplift; for example, in the vicinity of Hemet and in the San Timoteo Badlands.

Garner Valley (Hemet Valley of Fraser, 1931) was formerly filled with alluvial deposits derived mostly from the San Jacinto Mountains. Only a few remnants of these old alluvial deposits now remain. These deposits accumulated when this valley may have drained southeastward, as inferred by Fraser (1931, p. 500), possibly toward Coachella Valley as suggested by the east slope of the alluvial fan at Santa Rosa Summit. If so, the San Jacinto and Santa Rosa Mountains were once separated by an alluviated valley. The latest uplift apparently reversed the drainage of Garner Valley so that it now drains northwest into the San Jacinto River, which may have captured this drainage. This stream is incised in a deep canyon through the Thomas Mountain uplift. This uplift may be the effect of the latest elevation of this strip within the San Jacinto fault zone.

The Perris block was elevated probably in early Tertiary time, and it probably remained elevated throughout the Tertiary, as suggested by the absence of Tertiary sediments. It may have been quite mountainous sometime during the Tertiary but has since been geologically stable so that the mountainous surface was eventually worn down to low relief. Several old erosion surfaces were recognized by Dudley (1936), but the lowest and most extensive subdued surface probably corresponds to the elevated lower erosion surface of the San Jacinto block.

This lower erosion surface appears to include the peneplained surface cut on the basement rocks west of Perris and which slopes eastward under the Perris Valley, and the peneplained surface south of Hemet and which slopes northward under the San Jacinto Valley. This condition suggests that the basement surface beneath the alluvium of this valley terrain is also part of this old peneplain surface. The isolated ridges of basement rocks within this valley terrain are remnants of the former mountainous terrain.

This alluviated terrain is unusual because of 1) its large areal extent within the Peninsular Range Province, and 2) because of the generally transverse northeast trend of the isolated basement hills or ridges within it.

The large extent of this alluviated terrain within the Perris block suggests that this part may have subsided and become partly covered with alluvium derived mainly from the San Jacinto block.

The northeast-trending basement ridges within this alluviated terrain divide part of it into northeast-trending valleys. These valleys must have been carved out of the former mountainous terrain by streams that drained either northeastward (Larsen, 1951, p. 12) or southwestward across it. These alluviated valleys appear to somewhat converge south of Perris, suggesting that the streams that carved them drained southwestward, converged south of Perris, and drained outward near Lake Elsinore. If so, the major source of the streams was probably in the higher terrain northeast, in the San Jacinto Mountains across the San Jacinto fault, when they were directly across the fault. The high San Jacinto Mountain source area has since been shifted southeastward to its present relative position by right slip on the San Jacinto fault zone.

Although San Gorgonio Pass is an alluviated valley terrain, its high part is part of the main drainage divide between drainage westward to the Pacific Ocean and drainage eastward to the Salton trough. Its highest altitude of 2500 ft ($\pm$ 800 m), higher than the alluviated part of the Perris block, indicates that it was involved in uplift by arching of the Peninsular Range terrain. The part west of Beaumont is being severely dissected by tributaries of San Timoteo Wash.

San Gorgonio Pass is composed largely of southward-sloping Quaternary alluvial fans derived from the actively rising San Bernardino Mountains (Dibblee, 1975).

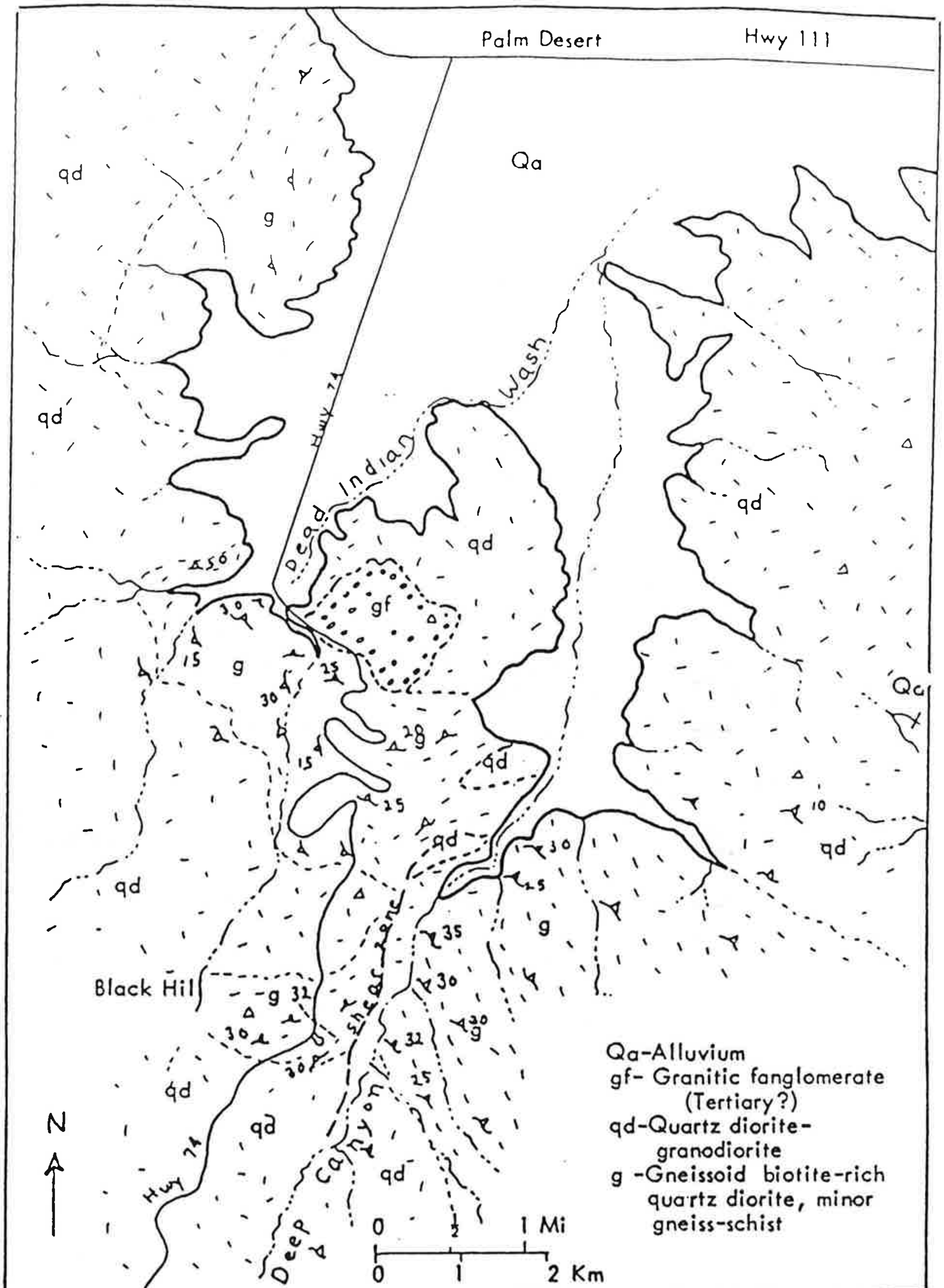

Figure 16. Reconnaissance geology of lower Deep Canyon area south of Palm Desert.

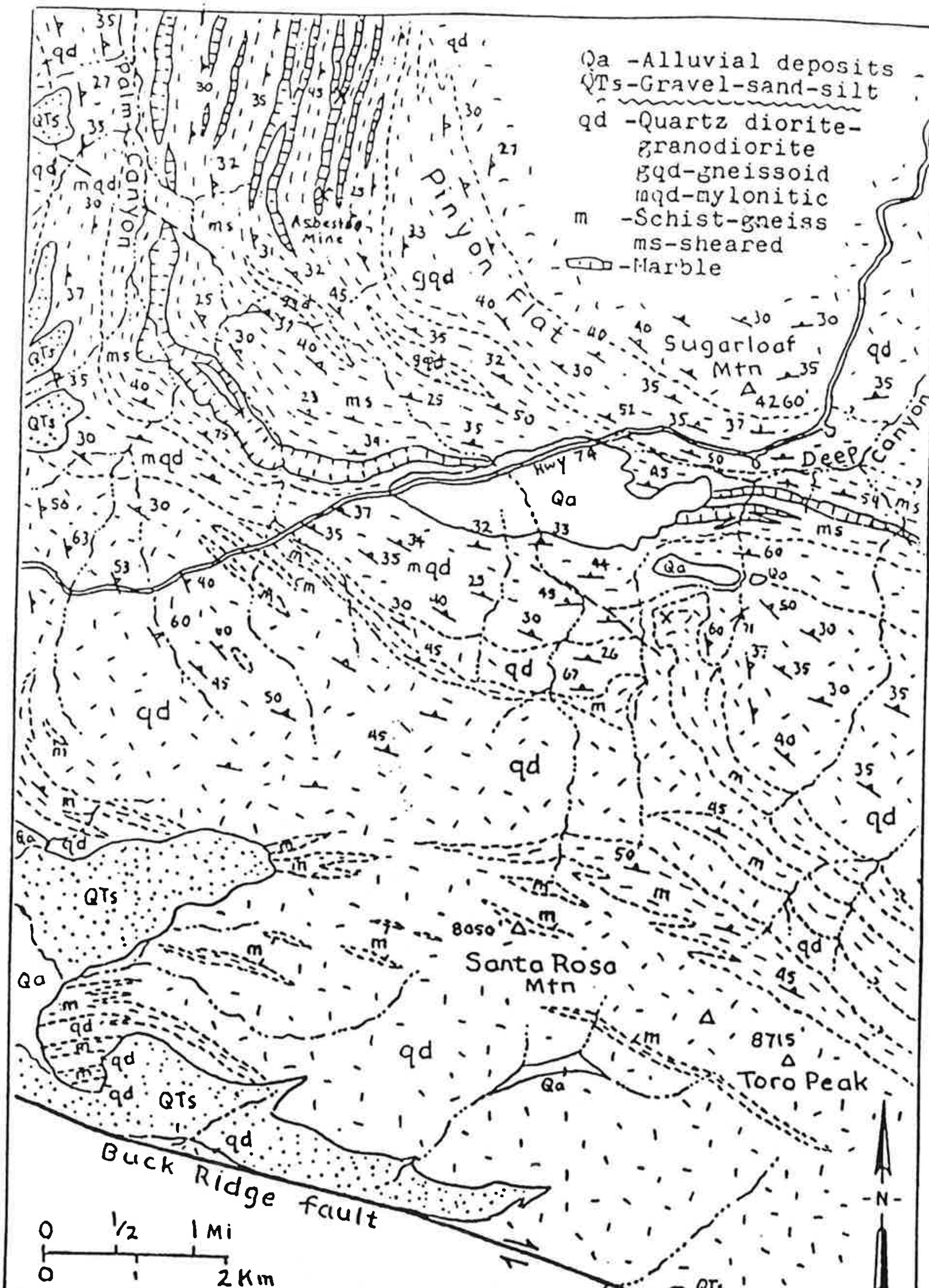

Figure 9. Reconnaissance geology of the Santa Rosa Mountain-Pinyon Flats area. Modified from Wright (1946), Sharp (1967), and Parcel (1972).

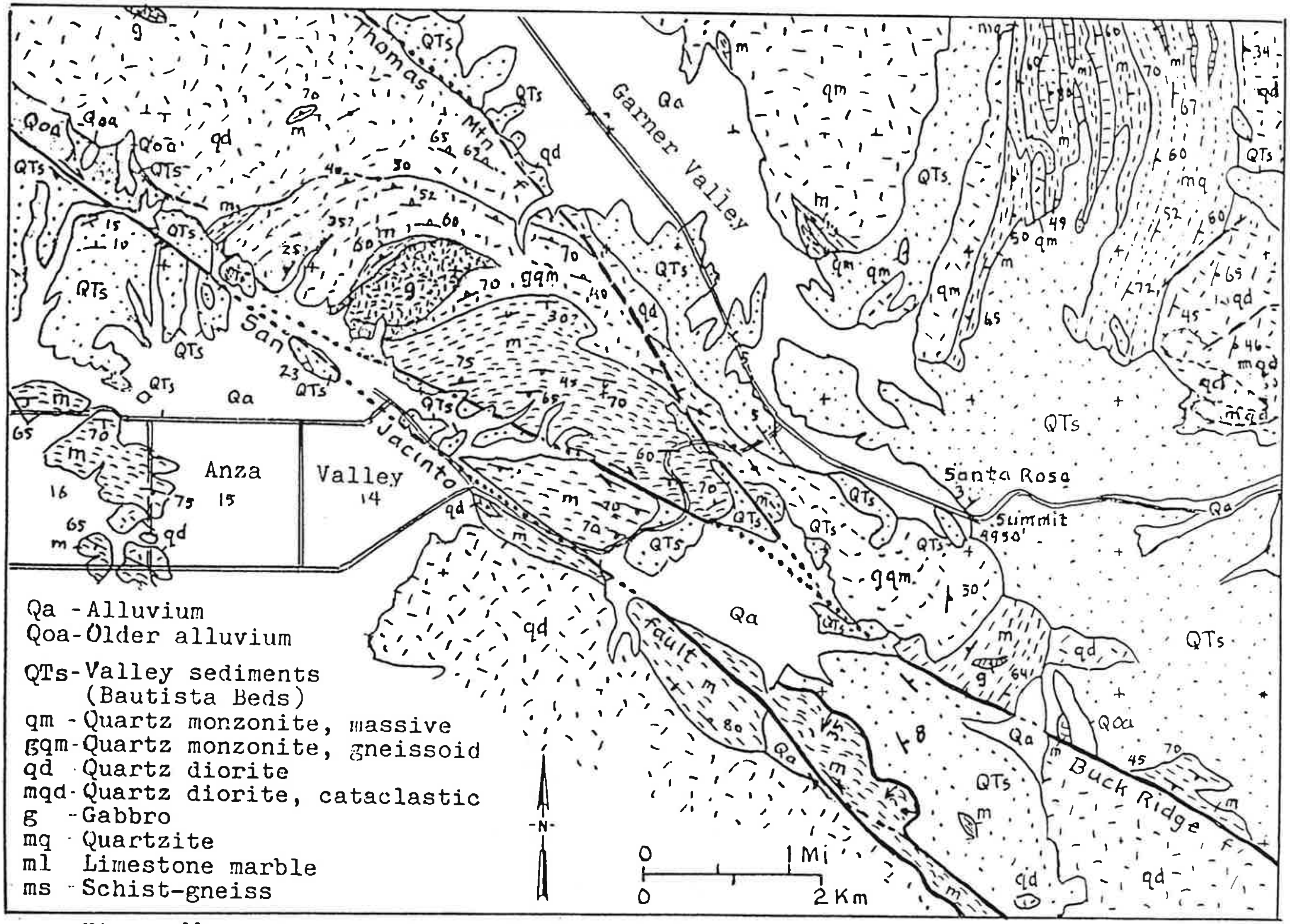

Figure 11. Geology of the San Jacinto fault zone in the Santa Rosa Summit area. Compiled from Sharp (1967) and Brown (1980).

color after

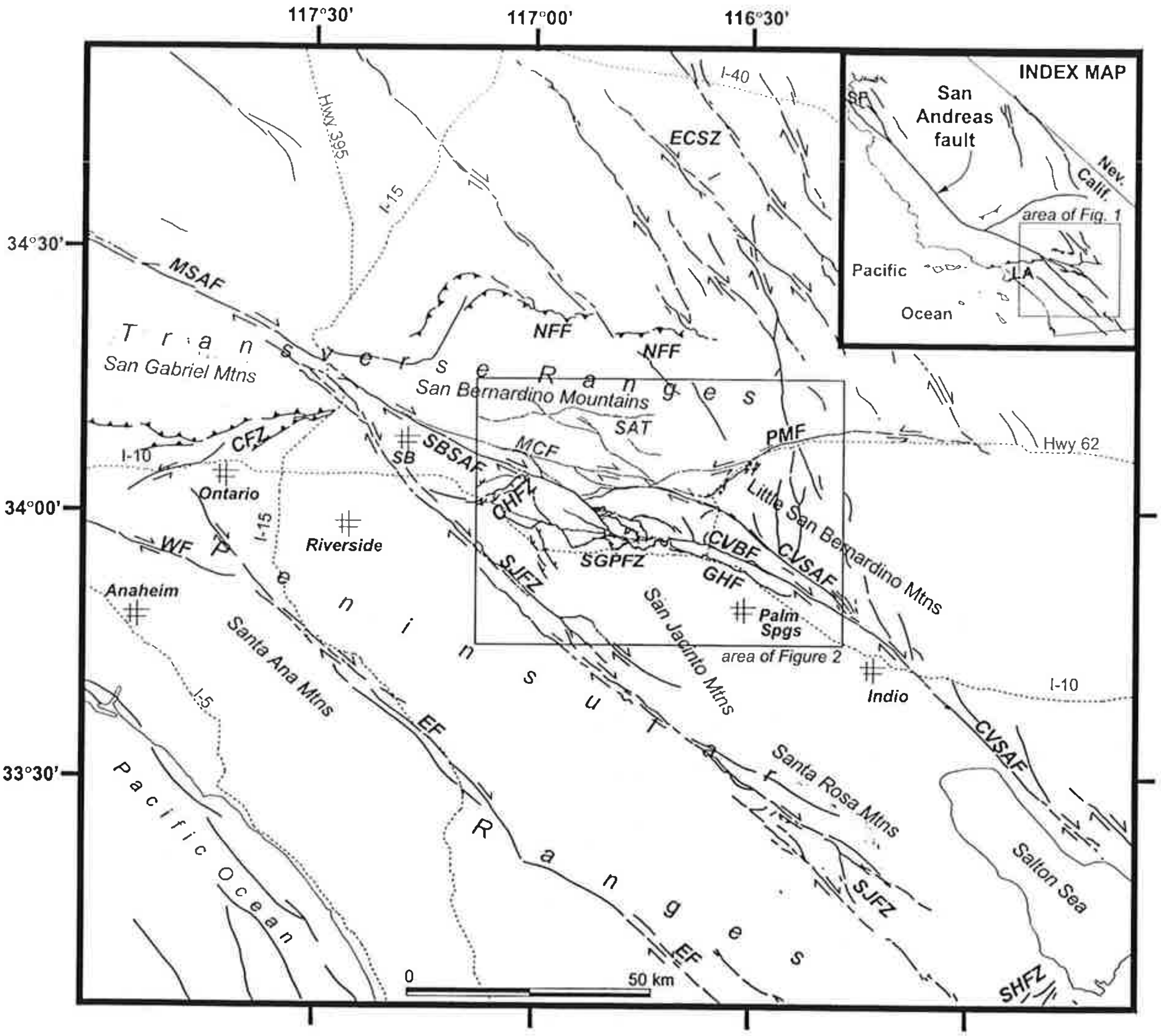

Figure 1.

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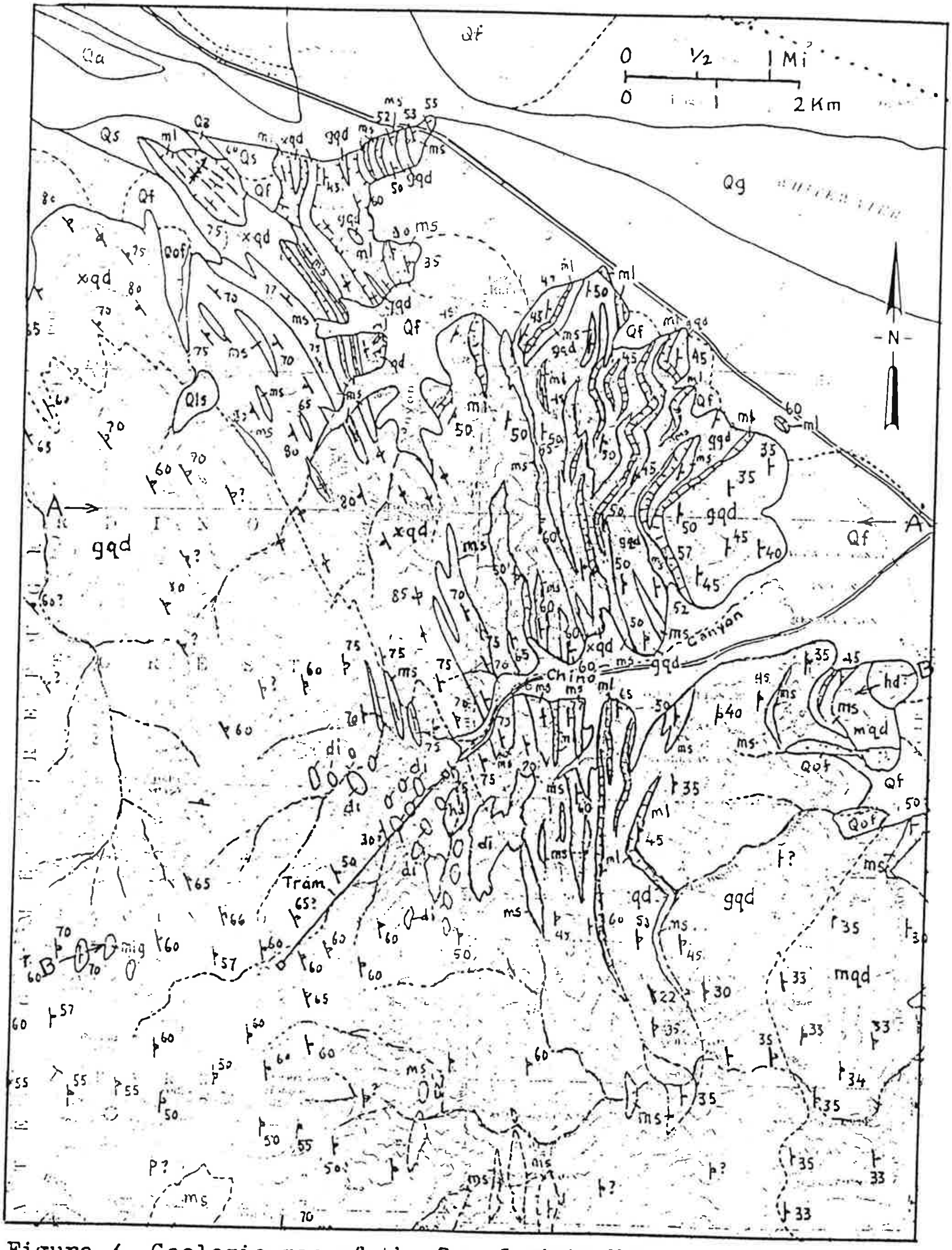


Figure 6. Geologic map of the San Jacinto Mountains in the Chino Canyon area. For symbols see figure 7. Geology by Dibblee (Palm Springs quadrangle) and Sydnor (1975).

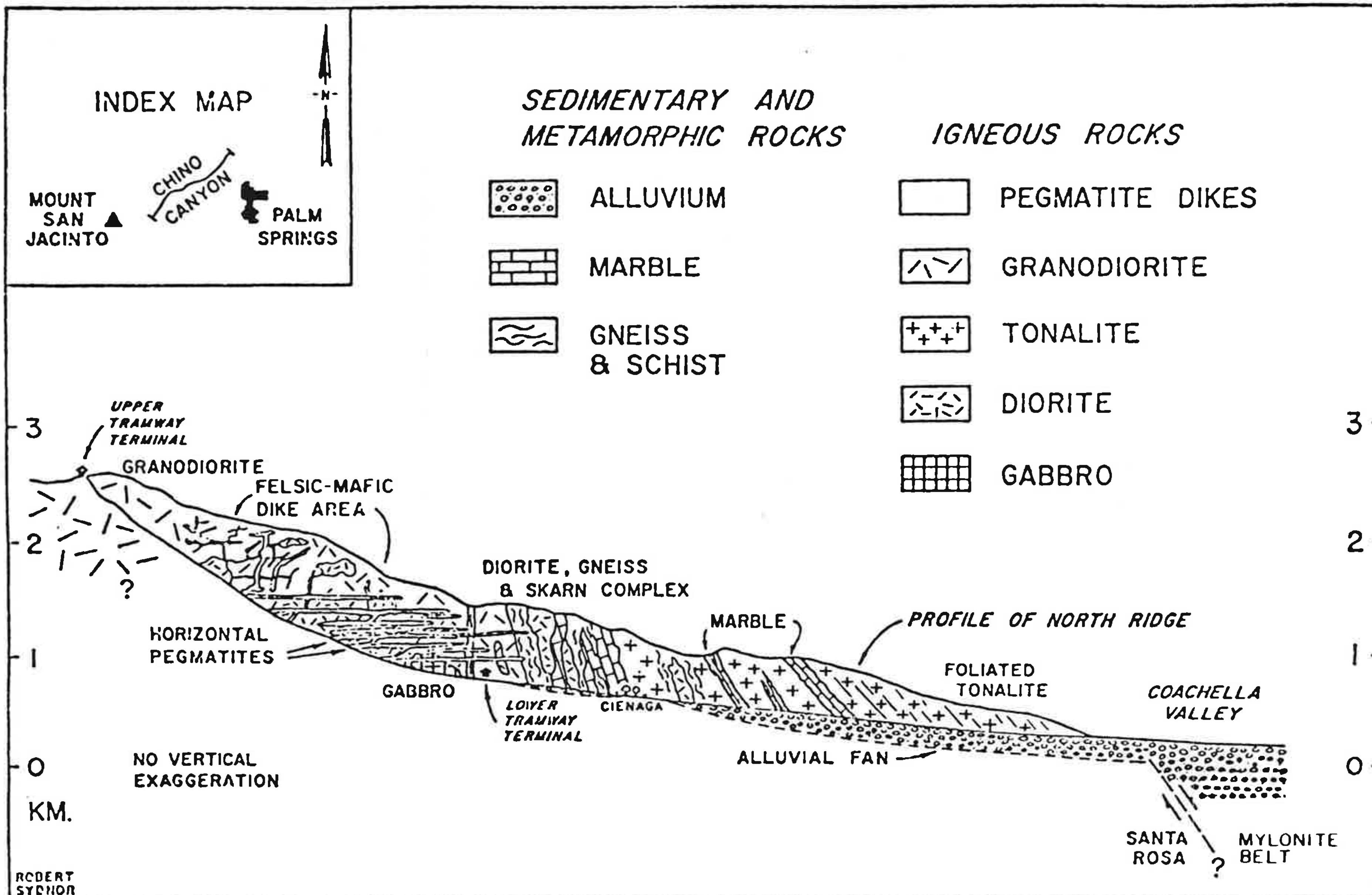


Fig. 14 Schematic cross section of Chino Canyon.

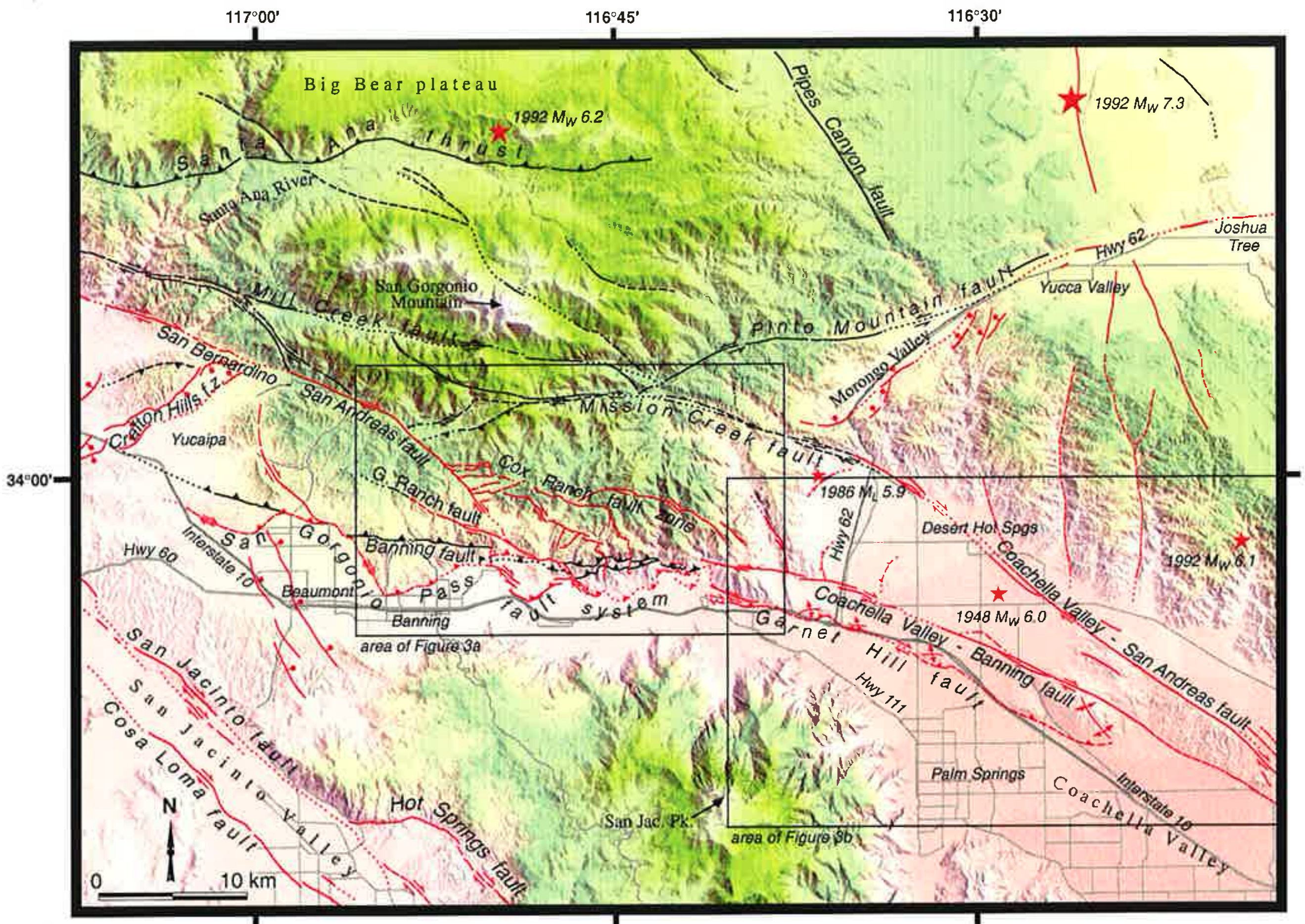


Figure 2a.

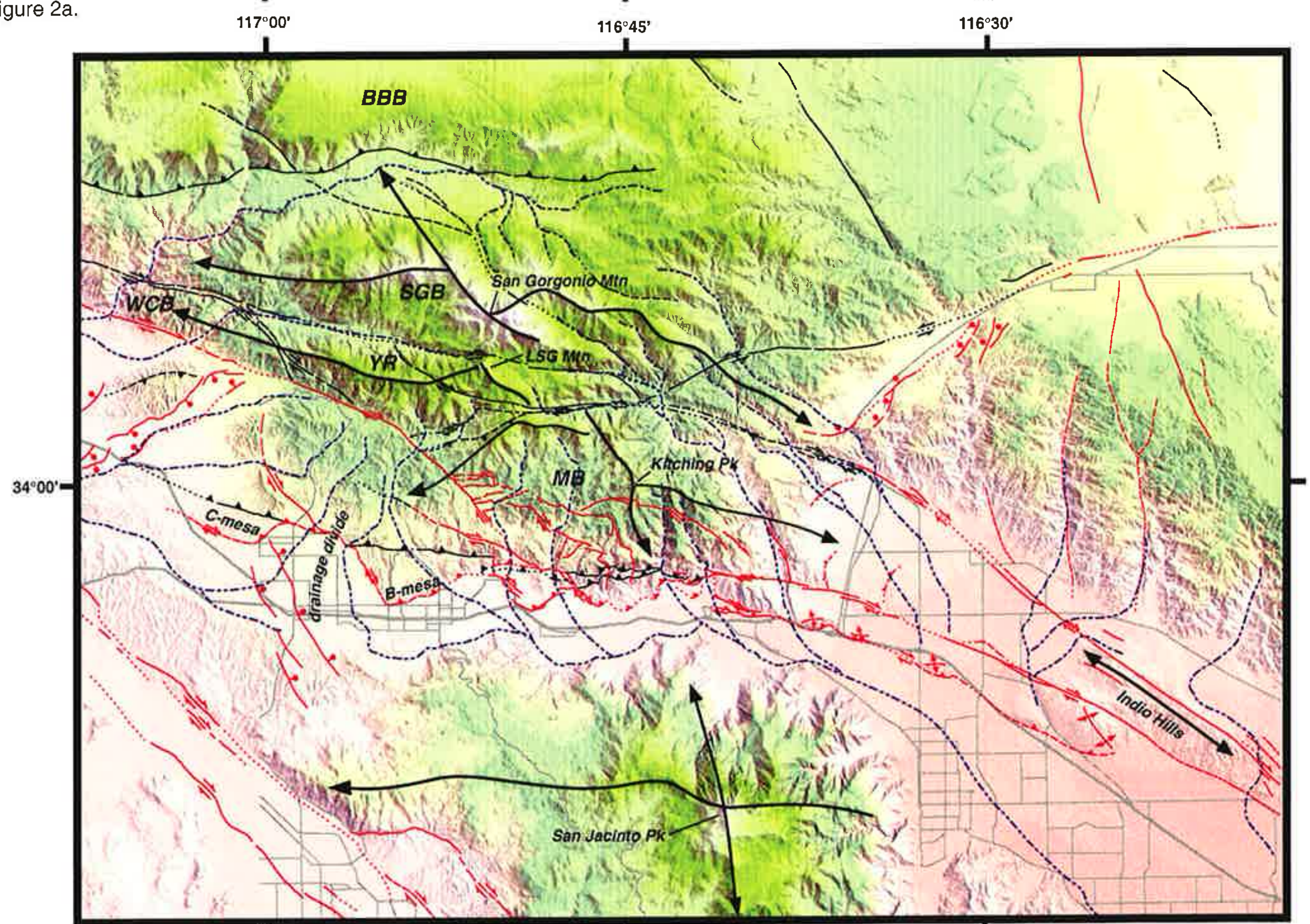
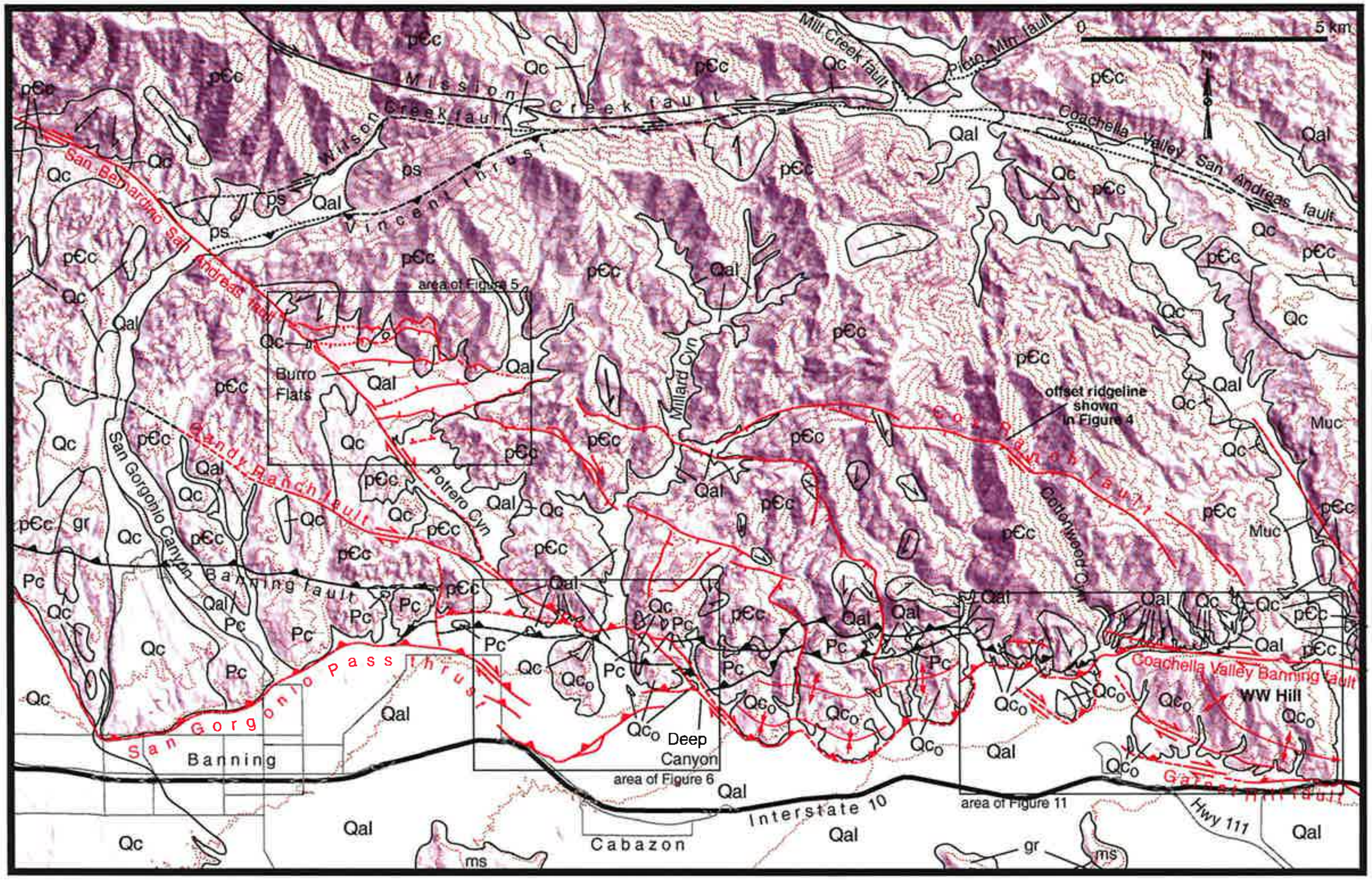


Figure 2b.



0
5 km
Mill Creek fault
Pinto Mtn fault
Mission Creek fault
Wilson Creek fault
Coachella Valley San Andreas fault
Vincent thrust
San Bernardino San Andreas fault
area of Figure 5
Burro Flats
Millard Cyn
offset ridgeline shown in Figure 4
Cox Ranch fault
Cottonwood Cyn
Potrero Cyn
Gandy Ranch fault
San Gorgonio Canyon
Banning fault
San Gorgonio Pass thrust
Coachella Valley Banning fault
WW Hill
Garnet Hill fault
Deep Canyon
area of Figure 6
area of Figure 11
Banning
Interstate 10
Cabazon
Hwy 111

Figure 3a.

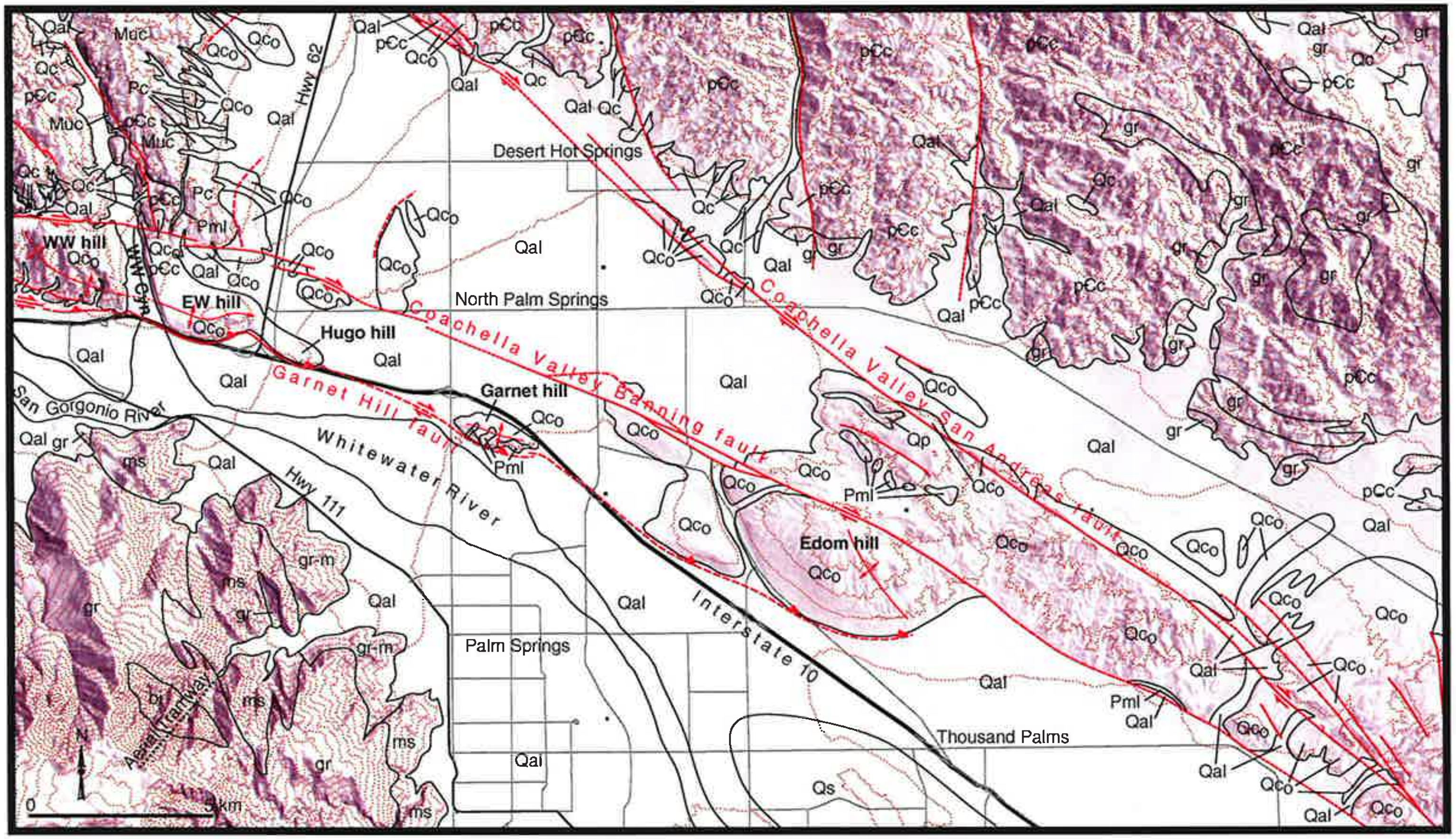
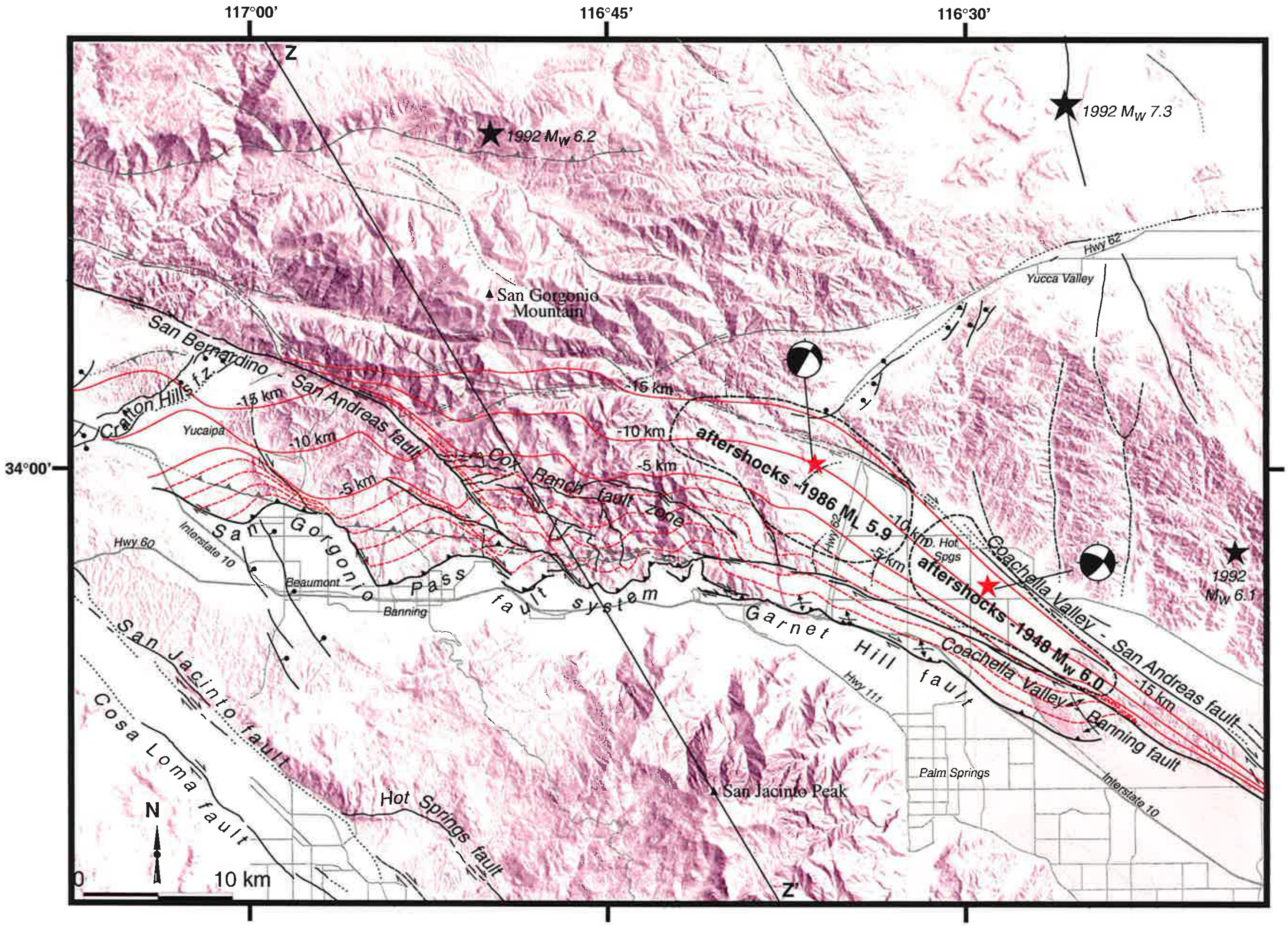

Desert Hot Springs
North Palm Springs
Palm Springs
Thousand Palms
WW hill
EW hill
Hugo hill
Garnet hill
Edom hill
WW Cyn
Hwy 62
Hwy 111
Interstate 10
San Gorgonio River
Whitewater River
Coachella Valley San Andreas fault
Coachella Valley Banning fault
Garnet Hill fault
Aerial Tramway
N
0
5 km

Figure 3b.



117°00'
116°45'
116°30'
34°00'
Z
Z'
1992 M_W 6.2
1992 M_W 7.3
1992 M_W 6.1
San Gorgonio Mountain
Hwy 62
Yucca Valley
San Bernardino - San Andreas fault
Crafton Hills f.z.
Yucaipa
-15 km
-10 km
-5 km
Cox Ranch fault zone
aftershocks -1986 ML 5.9
aftershocks -1948 M_W 6.0
D. Hot Spgs
Coachella Valley - San Andreas fault
Coachella Valley - Banning fault
San Gorgonio Pass fault system
Garnet Hill fault
Hwy 60
Interstate 10
Beaumont
Banning
Hwy 111
Palm Springs
San Jacinto Peak
San Jacinto fault
Cosa Loma fault
Hot Springs fault
N
0
10 km

Figure 15.

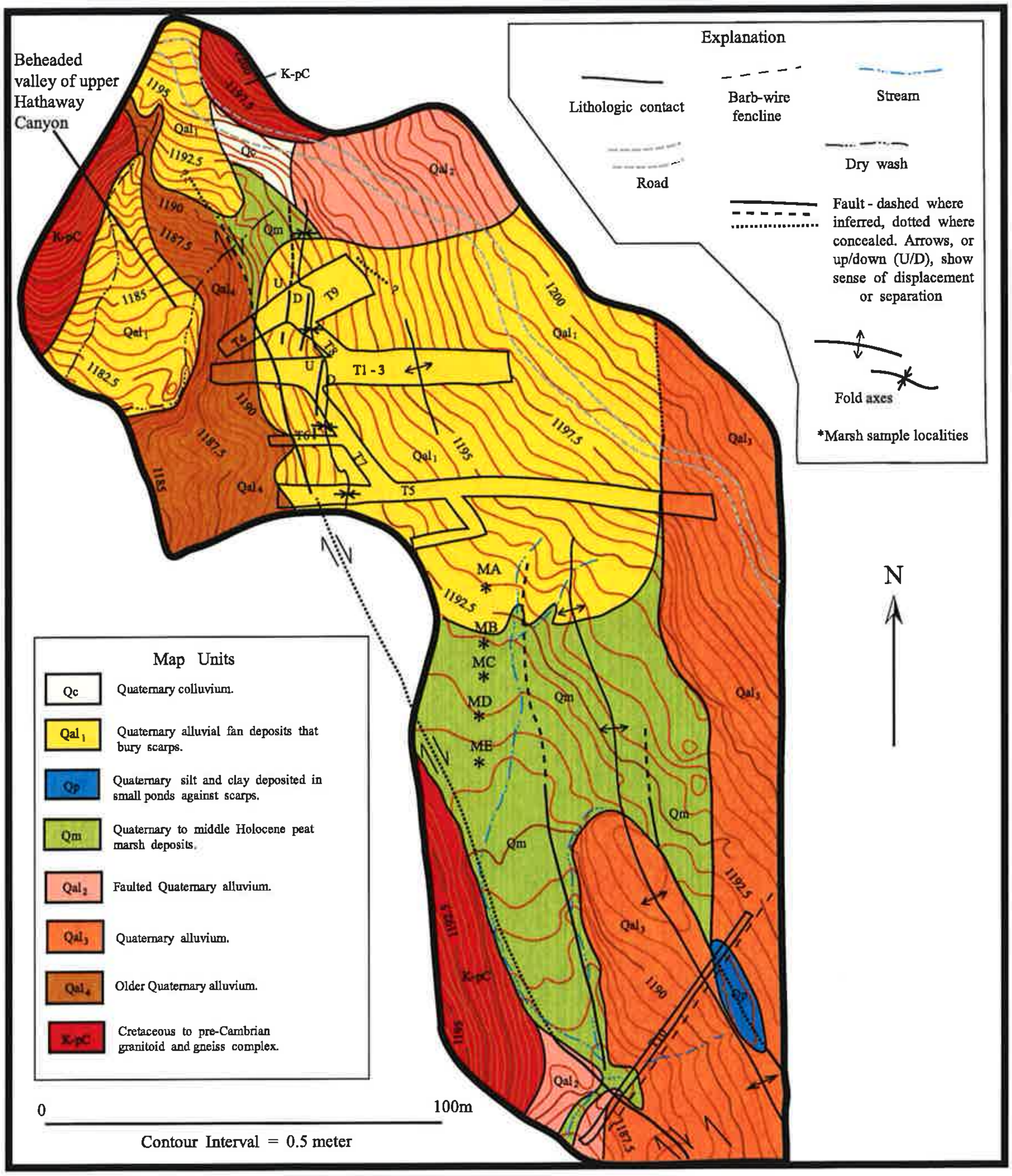


Figure 4. Geologic map of the Burro Flats paleoseismic site. Note location of T5, where much of work was done for this study. Topographic base was produced using a Leica total station. Courtesy of Yule, in preparation.

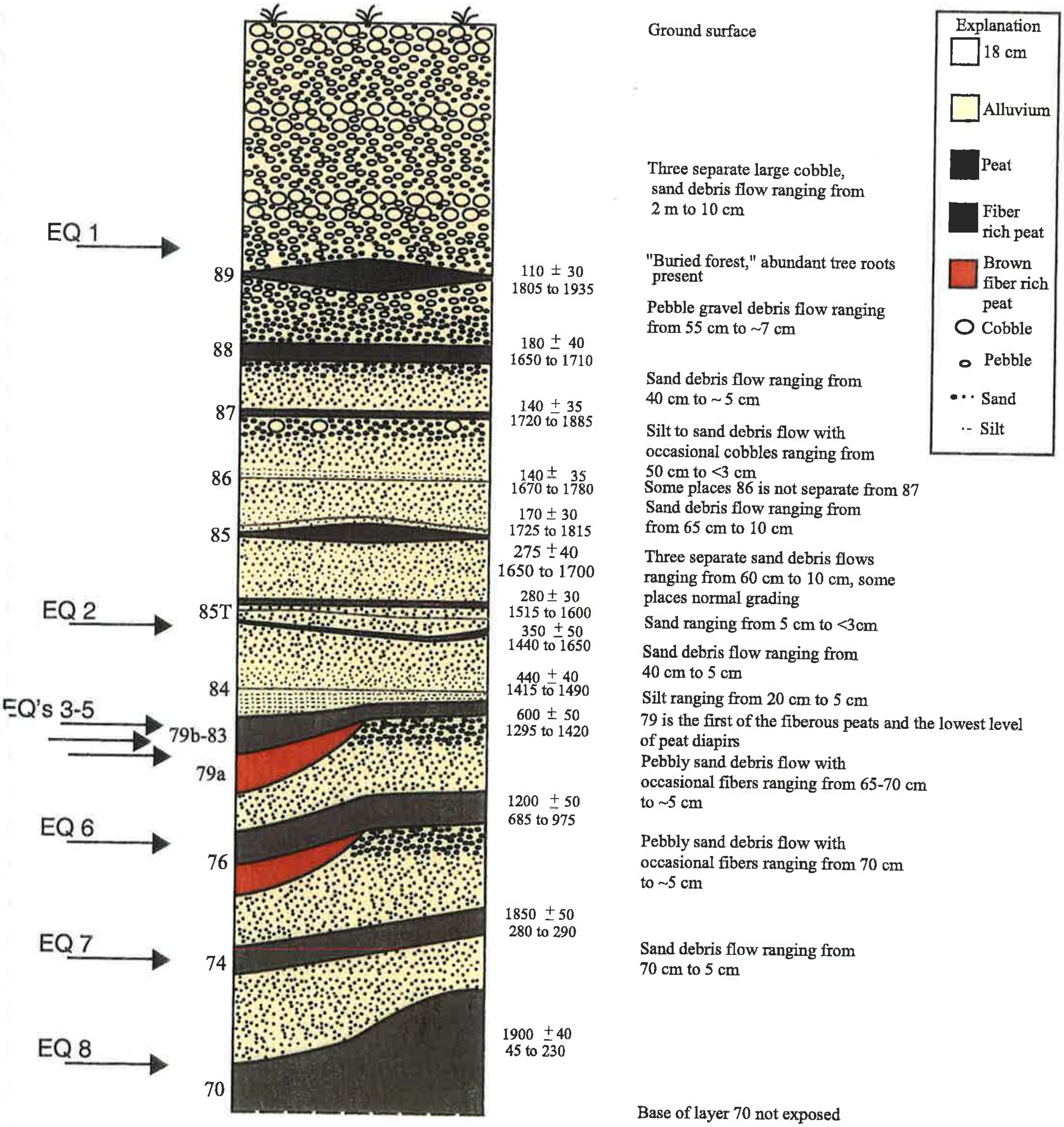


Figure 5. Generalized stratigraphic column of the north wall of trench 5. In general units thicken to the east, and the thickest part of each unit is depicted here. Sediment layers are poorly sorted and poorly indurated. Numbers to the right of the strat column represent the radiocarbon and the preferred calibrated calendar age of the corresponding peat. The first age cited for a given peat layer is the radiocarbon age while the age beneath it is the calibrated calendar age A.D. (see text for explanation). Fiber ages are used for all units except for units 89, 84, (bulk ages) and 76 (bulk humic age) due to a lack of parallel fibers in the samples.

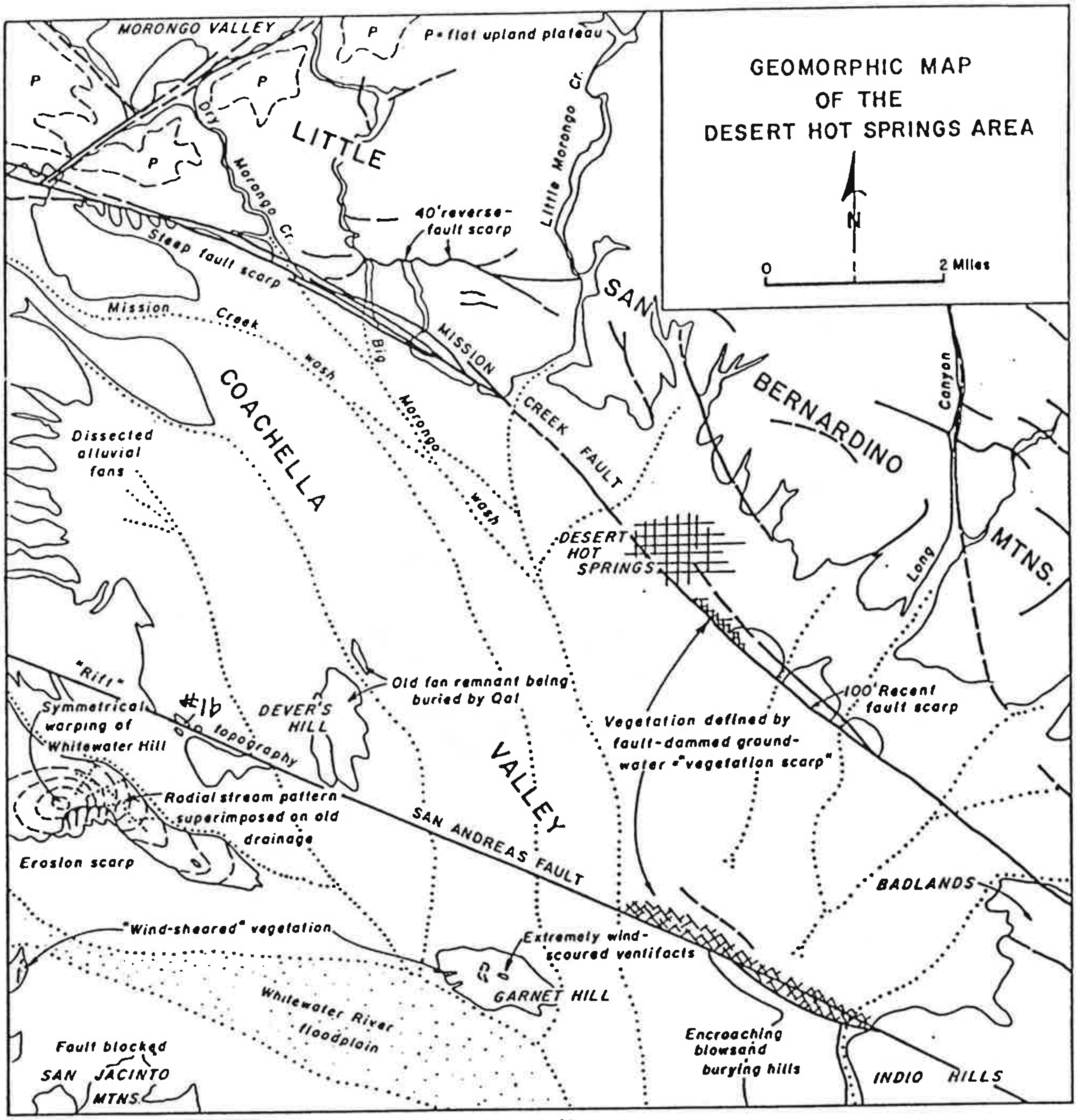
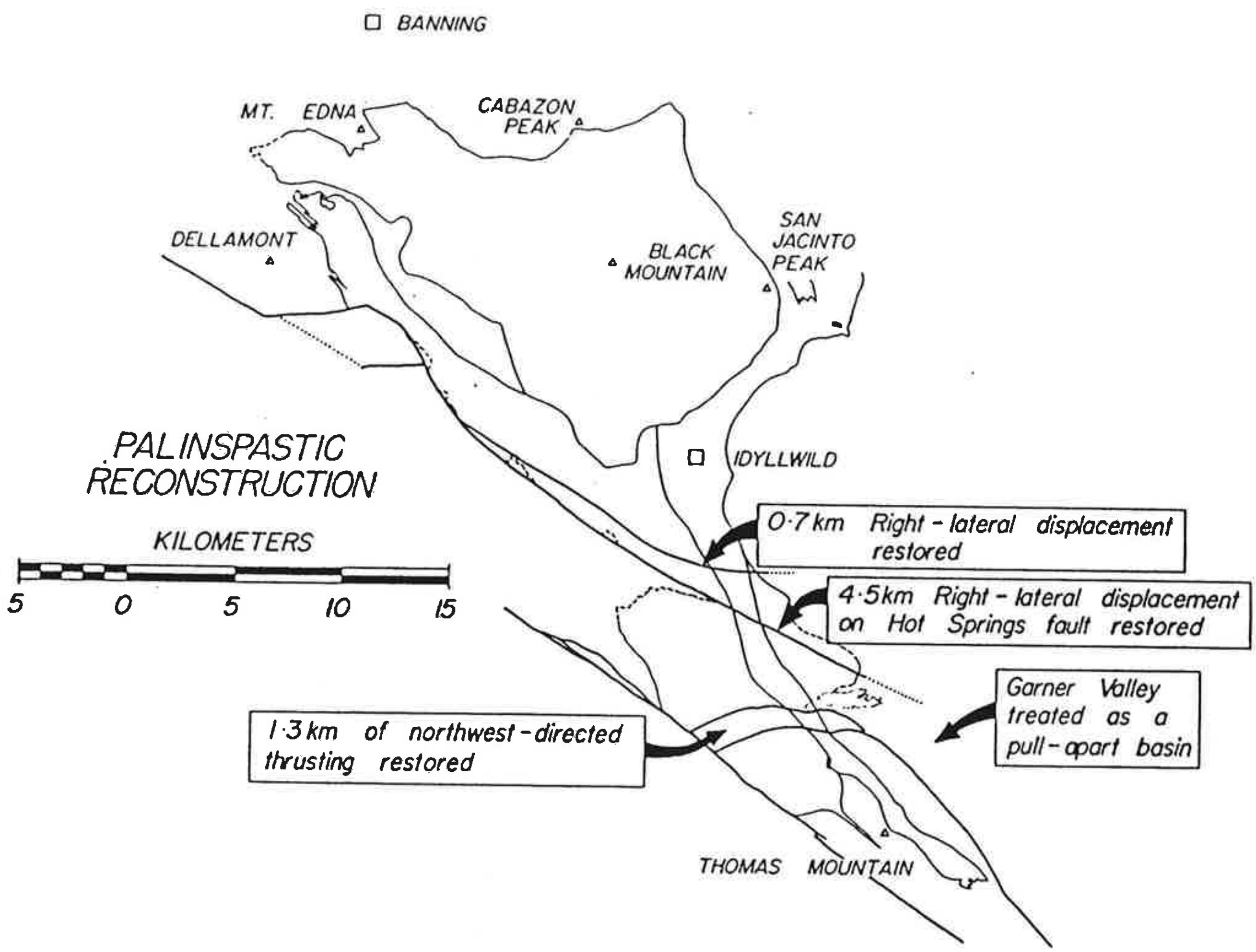


Figure 11.

Richard J. Proctor, 1968, Geology of the Desert Hot Springs, Upper Coachella Valley area, California. California Division of Mines and Geology, Special Report 94.

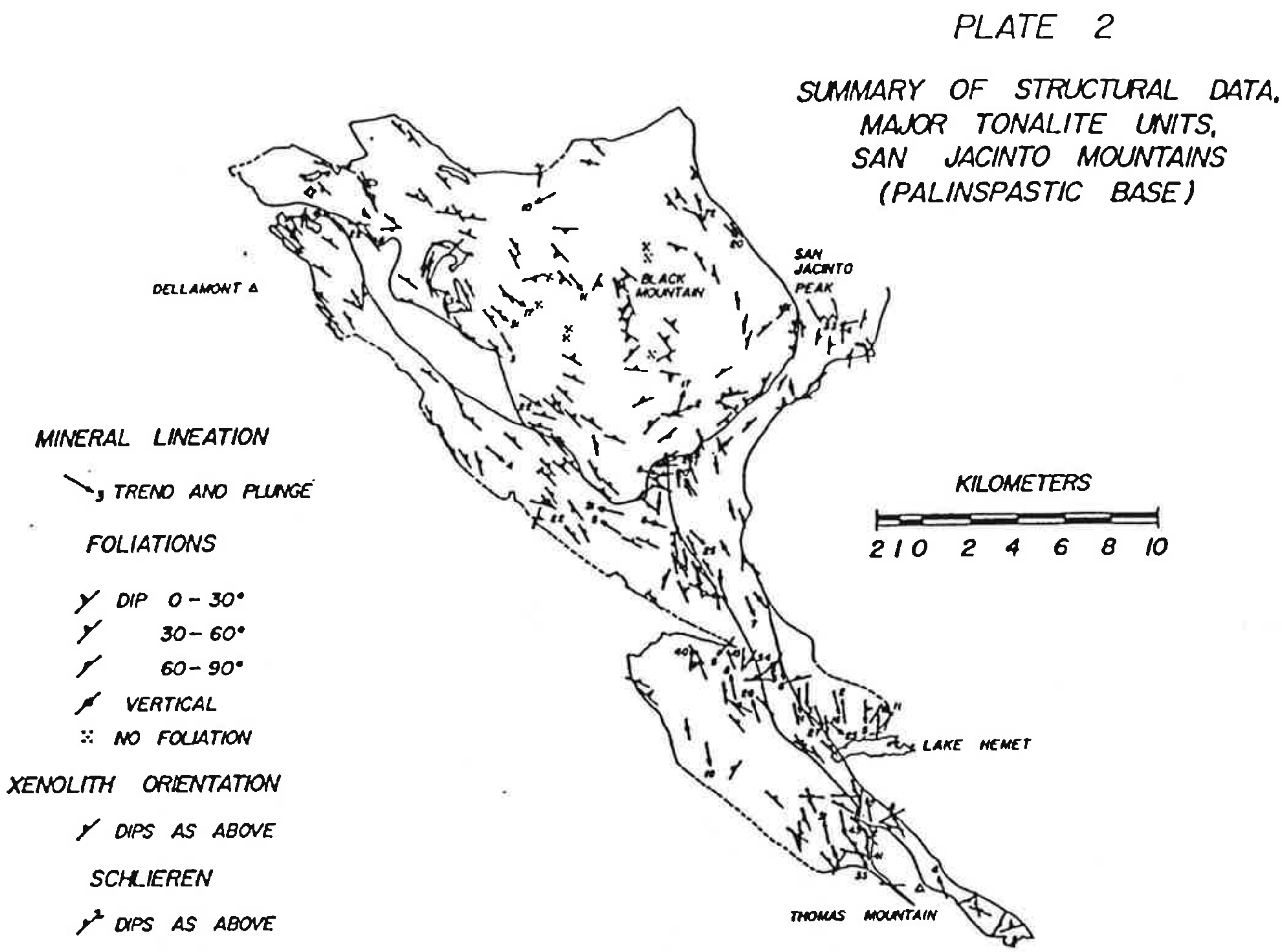
Figure 8 a



Palinspastic reconstruction of the San Jacinto Mountains prior to Tertiary faulting.

Figure 10

Hill, 1984



Summary of structural data (mineral foliations and lineations, schlieren, xenolith orientation) for the major tonalite units.

Figure 11

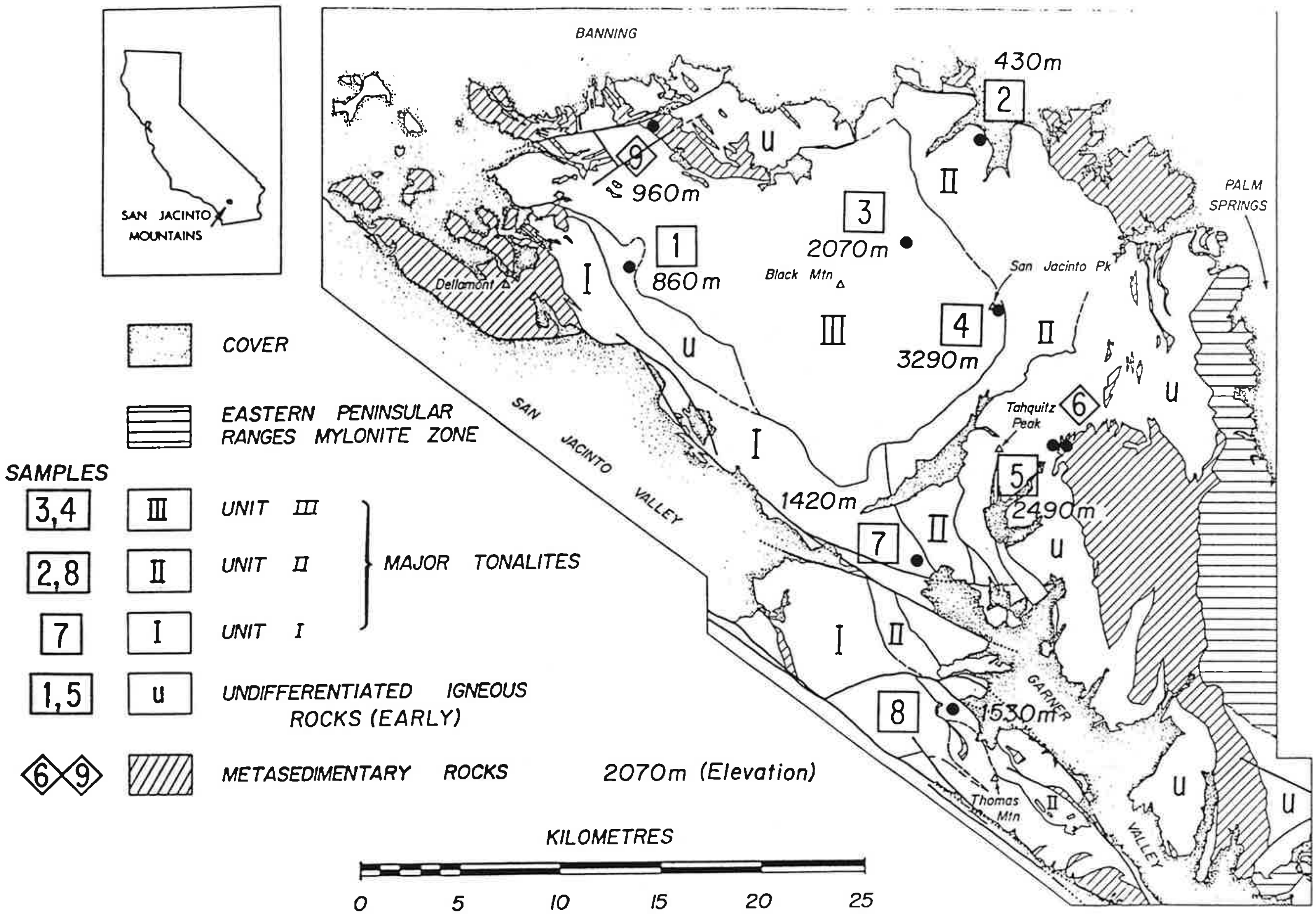
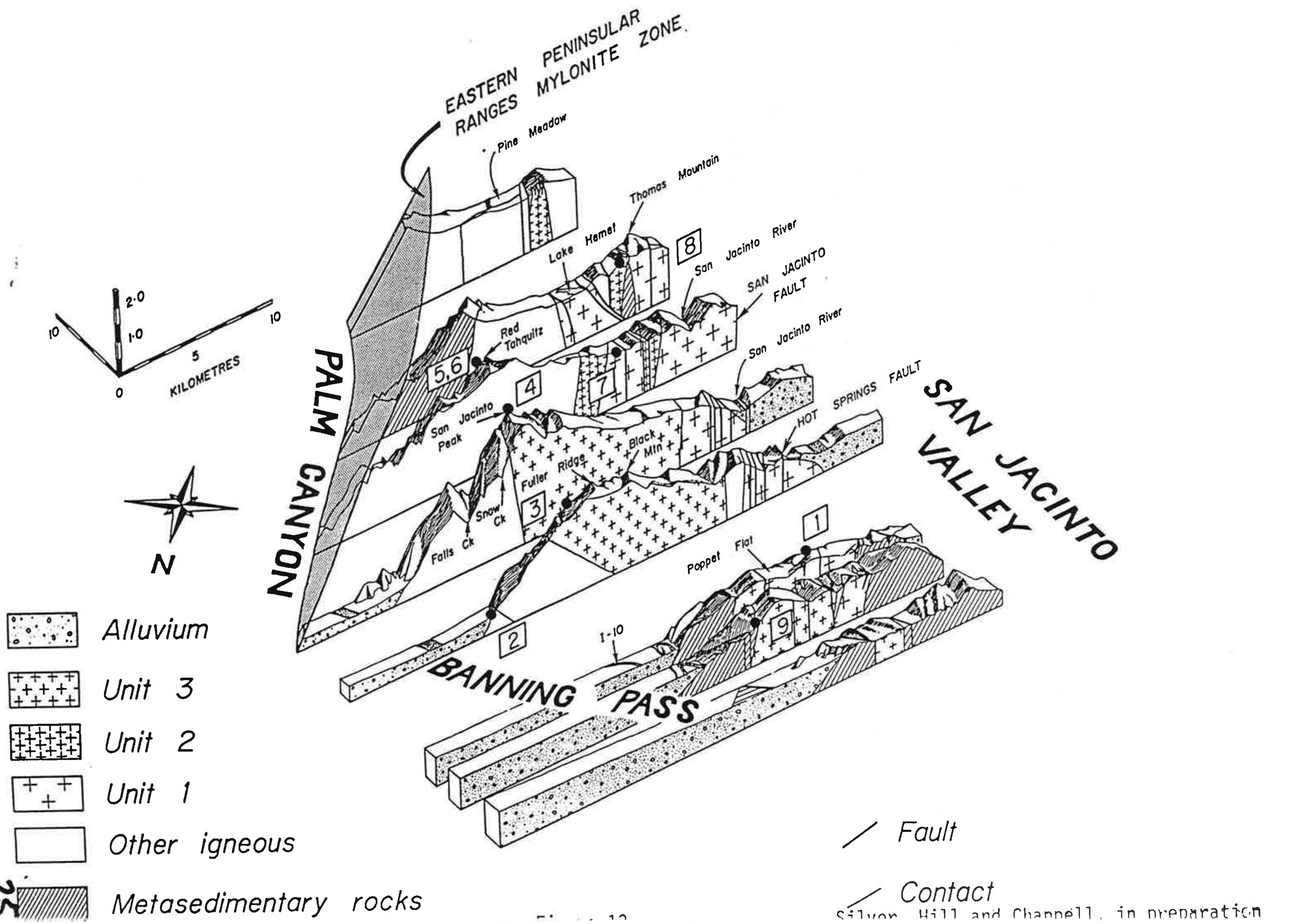
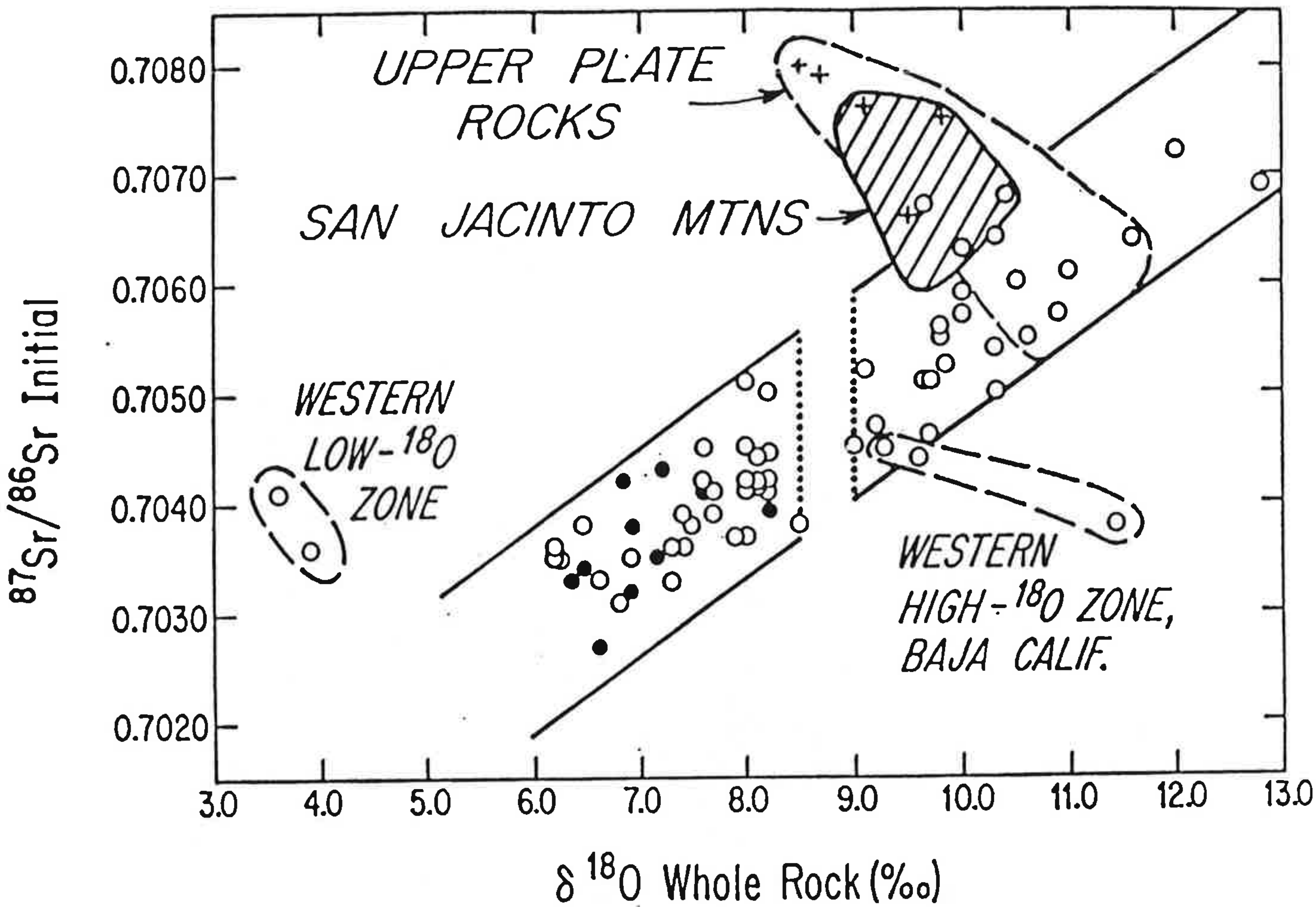


Figure 12

Silver et al., in preparation



Silver, Hill and Chappell, in preparation



Sr_i versus $\delta^{18}O$ for the Peninsular Ranges batholith (from Taylor and Silver, 1978) and the San Jacinto Mountains (this work, Taylor and Silver, 1978). Field for Upper Plate Rocks inferred from Taylor and Silver, 1978, and Silver et al., 1979.

Primary values for rocks from the San Jacinto Mountains largely overlap the field of the Upper Plate Rocks. The majority of the San Jacinto data fall slightly to the low-$\delta^{18}O$, high-Sr_i side of the main batholith trend.

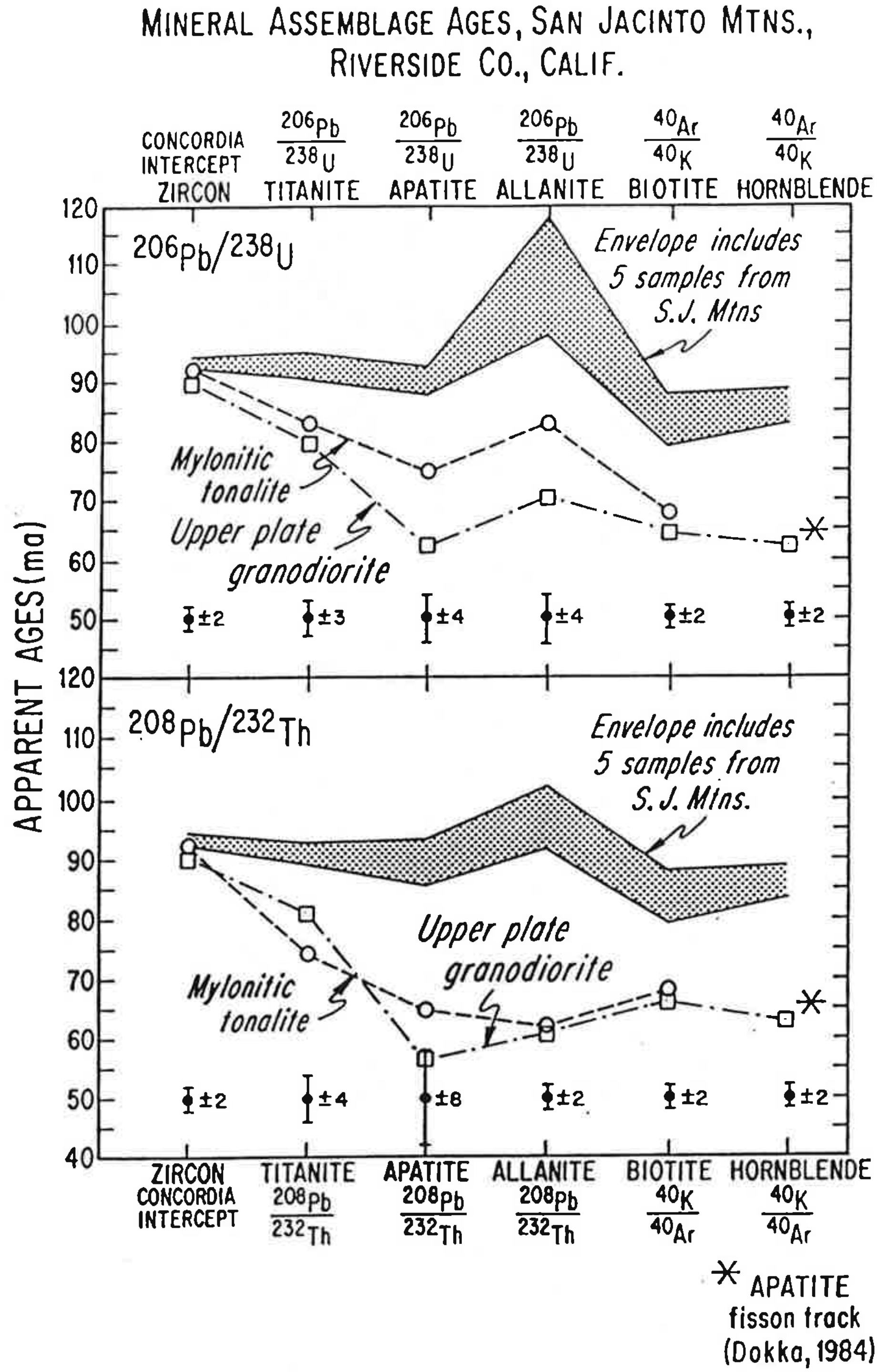


Figure 21 L.T. Silver, in preparation

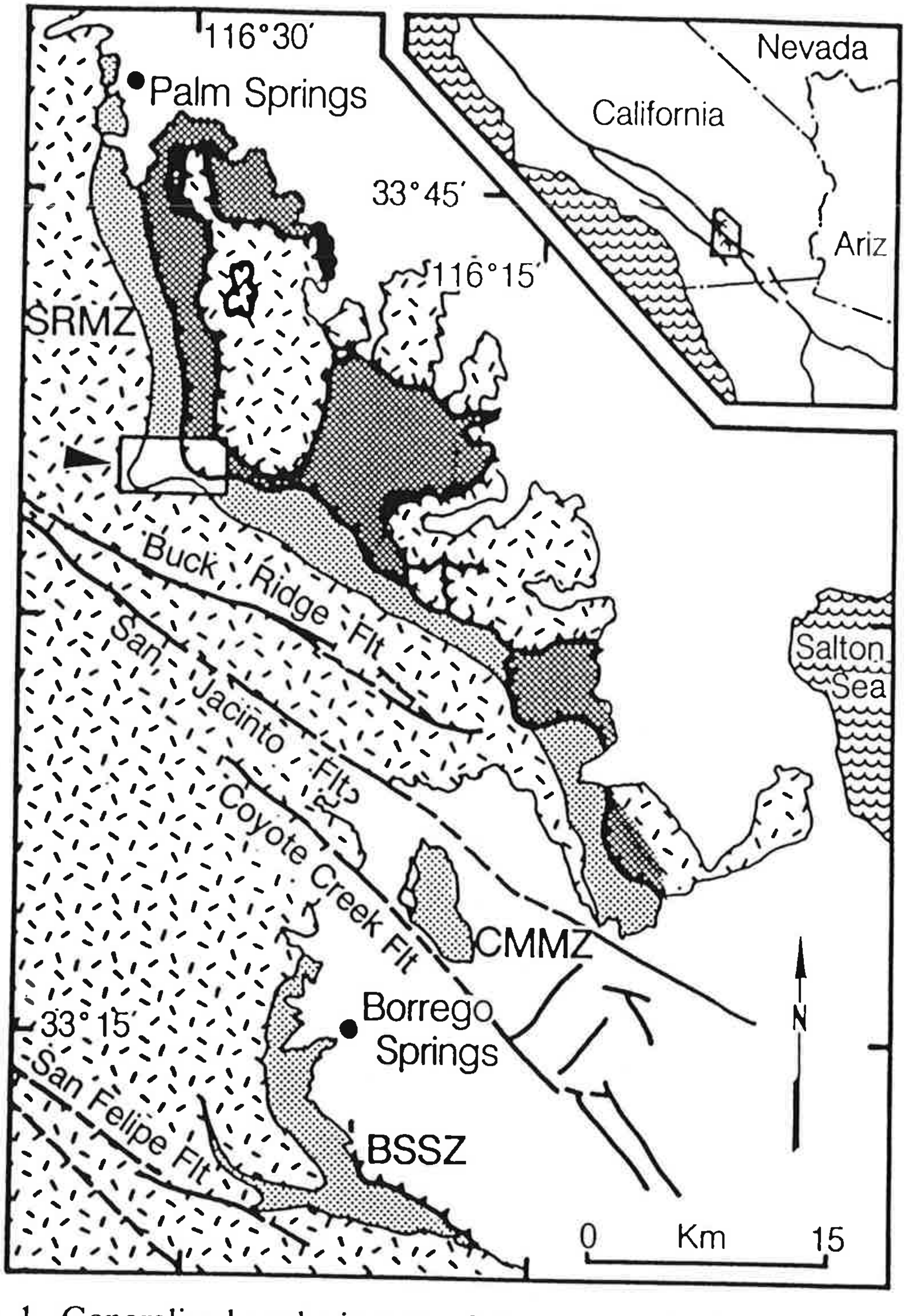


Fig. 1. Generalized geologic map of the eastern Peninsular Ranges mylonite zone (modified after Sharp 1979, Erskine 1985). Stippled pattern represents middle Cretaceous granitic rocks of the Peninsular ranges batholith and older included metasedimentary screens. Shading indicates mylonites. SRMZ = Ranta Rosa mylonite zone, CMMZ = Coyote Mountain mylonite zone, BSSZ = Borrego Springs shear zone. Barbed lines show positions of low-angle faults; barbs face the hanging walls. Bold lines show high-angle, predominantly strike-slip faults. The area of the detailed map shown in

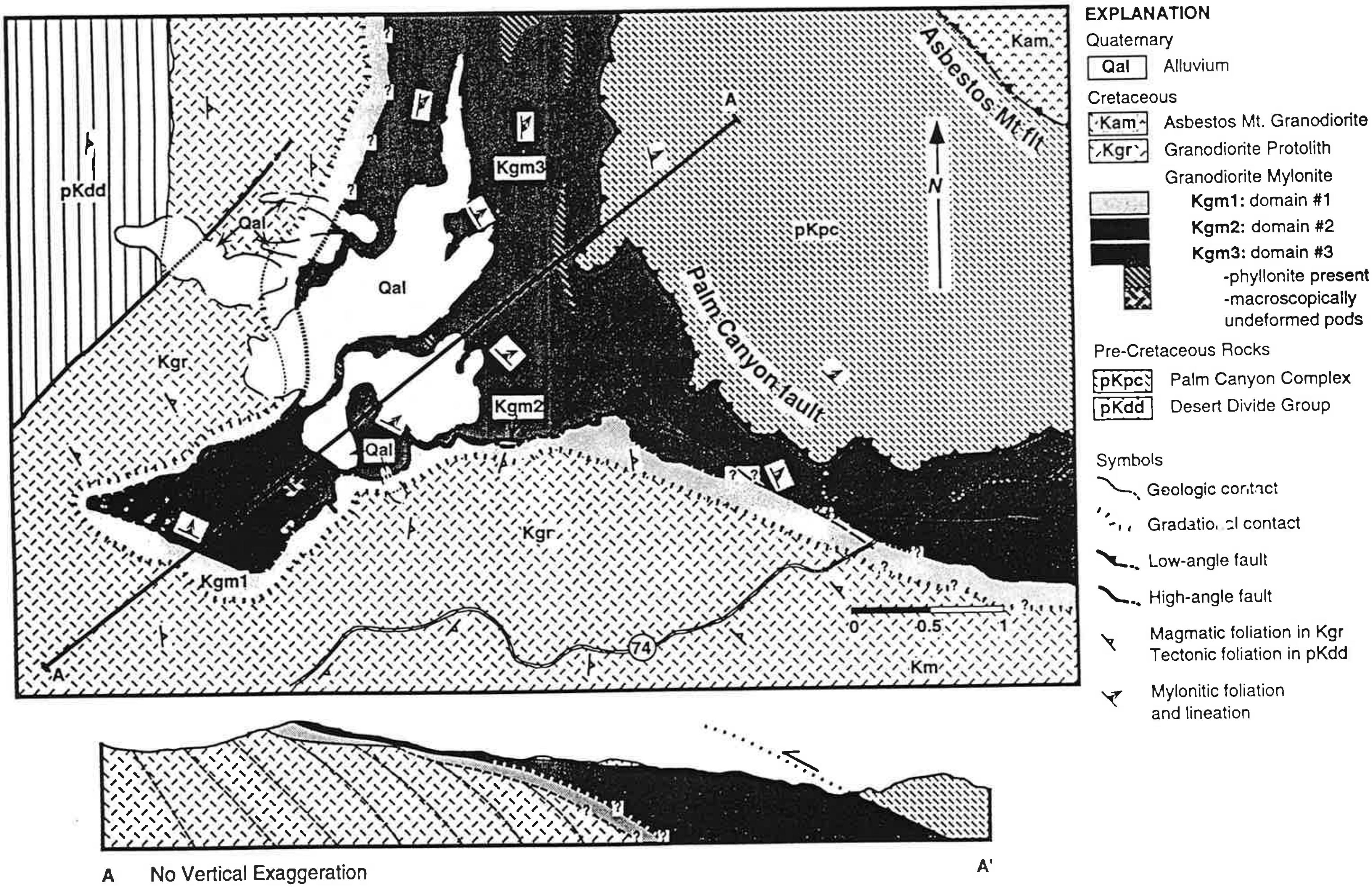


Fig. 2. Geologic map modified from that of Erskine (1985) shows distribution of the three structural domains described in the text. Unpublished mapping by L. Simmons, P. Jordan, K. Rich and Z. Zhong was used in the preparation of the map. Cross-section shows relations between the three structural domains. Dashed lines in granodiorite outside the mylonite zone schematically illustrate the trace of magmatic foliation. Sense of movement in the system is shown by arrow. Subsurface

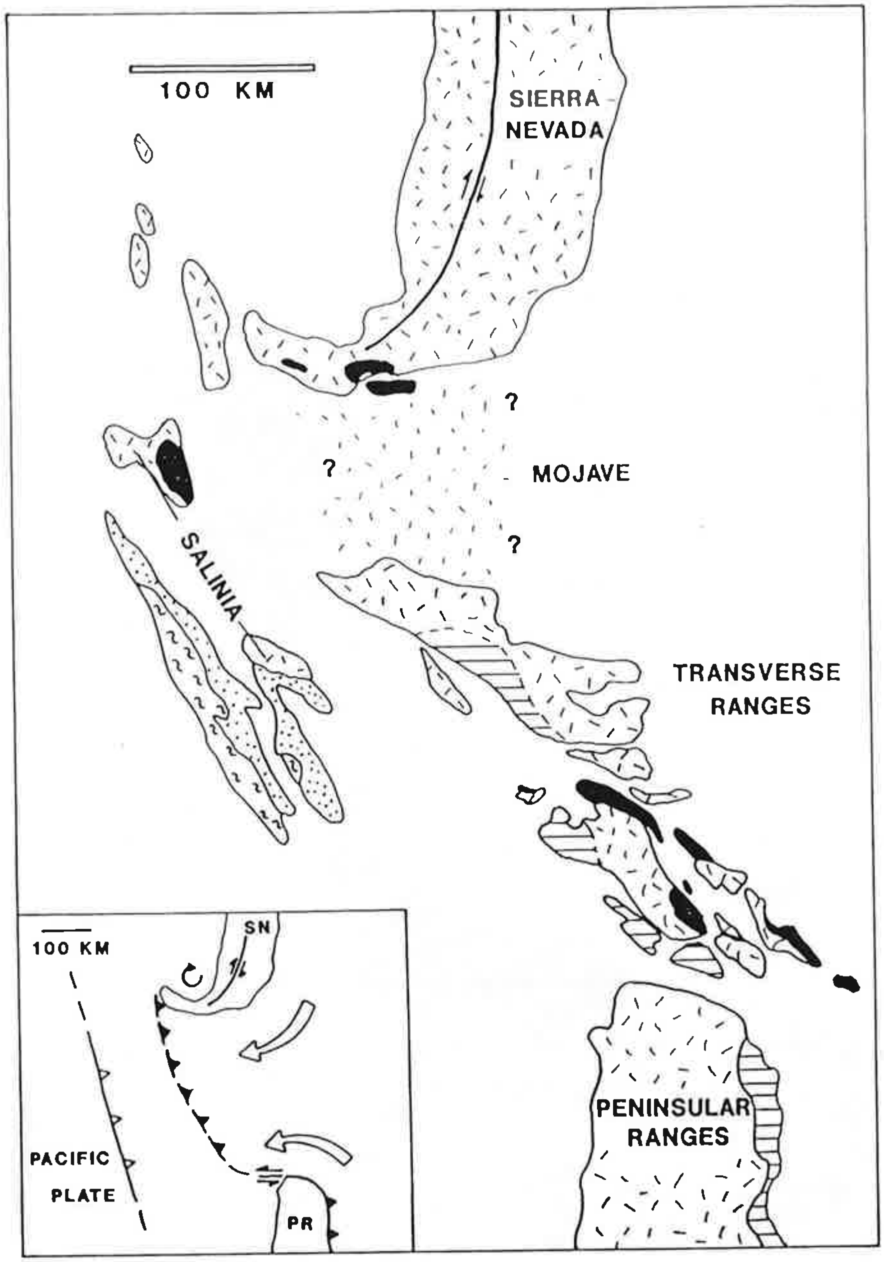


Fig. 5. Distribution of selected Late Cretaceous features and rock types on a palinspastic base for southern California with effects of Neogene and Quaternary faulting removed [after Burchfiel and Davis, 1981]. No attempt has been made to compensate for Neogene rotations in Transverse Ranges and western Mojave or effects of low-angle normal faulting in the Mojave region. Rock symbols are short dashes, Cretaceous and older Mesozoic plutonic rocks, metasedimentary pendants, and Precambrian crystalline rocks (distribution poorly known in Mojave desert); horizontal lines, Late Cretaceous age mylonitic rocks associated with the proposed synplutonic thrust system:

EXPLANATION

Pleist.(?) Recent

d	Debris from mining	QUATERNARY
Qal	Alluvium	
Qf	Fanglomerate, terrace gravels, coarse sand and pebble conglomerates	

Paleocene — Silverado formation[1]

Ts_3ss / Ts_3sh	Interlayered green clay shale (Ts_3sh) and micaceous arkose (Ts_3ss)	TERTIARY
Ts_2	Massive "pink mottle" claystone	
Tsb	Pisolitic claystone ("bone clay")	
Ts_1	White, gray and buff claystone (fire clay)	
Tscg	Clay conglomerate basal unit in places	

[1] (Martinez fm. on Plate I.)

Bedford canyon formation

bcc_2	Red, yellow and gray residual claystone	TRIASSIC
bcc_1	Yellow and gray weathered slate	
Ŧbc	Greenish gray slate, argillite and conglomerate, moderately weathered	

70 Fault showing dip, dashed where approximate

Fault, inferred

Fault, concealed

Contact, dashed where approximate

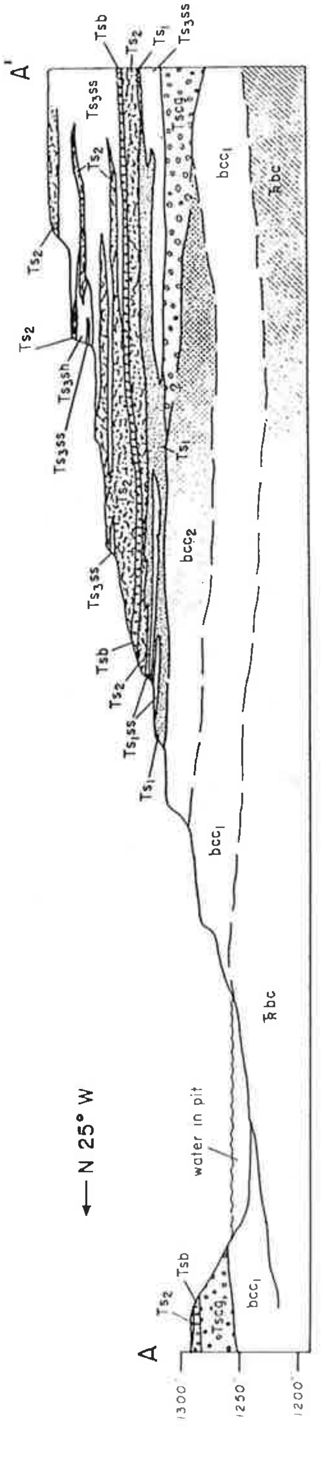
10 Strike and dip of beds

5 Presumed dip from indistinct bedding

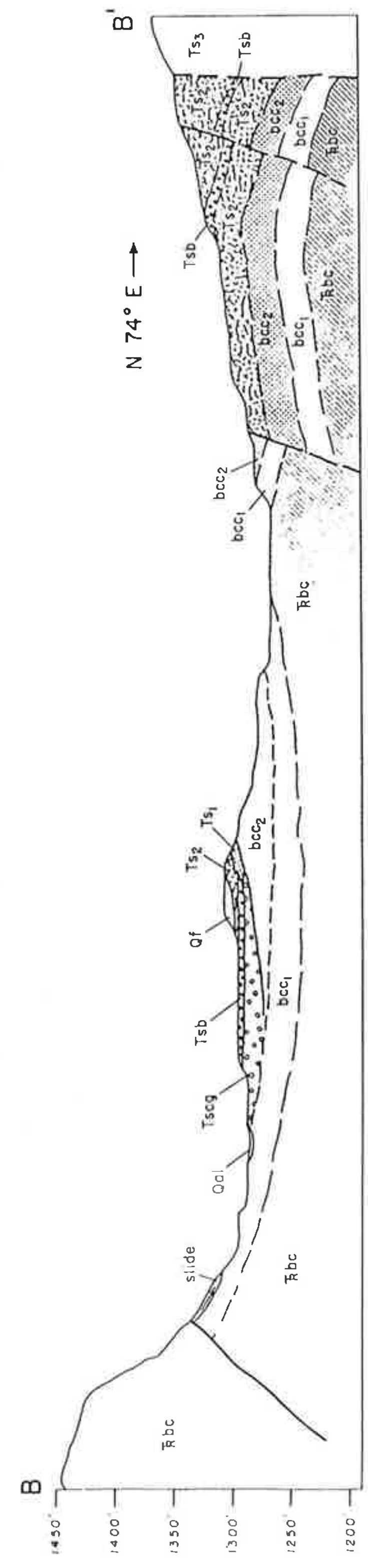
Slide or slump

Geology by Bert H. Rogers, March 1955

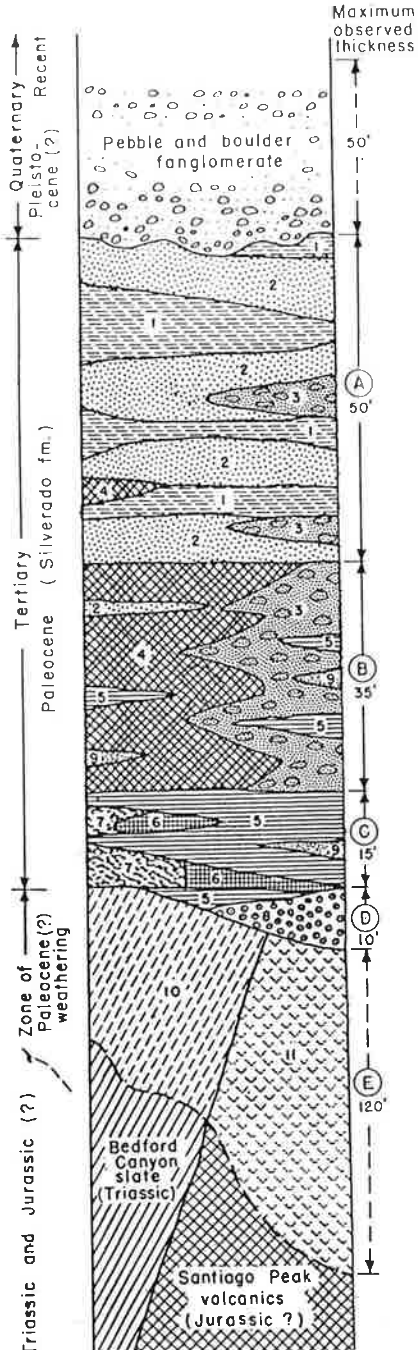
Topography by Bert H. Rogers and C.H. Gray, March 1955



Cross section A-A' across Los Angeles Brick & Clay Products Co. Clay Pit No. I, by Bert H. Rogers



Cross section B-B' across Los Angeles Brick & Clay Products Co. Clay Pit No. I, by Bert H. Rogers



EXPLANATION

(A) Green to gray, waxy clay shale (1) interbedded with arkosic, micaceous, coarse-grained sandstone (2); locally contains 1' to 8' thick layers and lenses of sandy, pink and white mottled claystone (3) and sandy white, yellow and gray claystone (4).

(B) Sandy, white, yellow and gray mottled claystone (4) containing lenses of white to gray claystone (fire-clay) (5) and lenses of coarse-grained, angular, clayey quartz sandstone (9). Pink and white sandy, mottled facies of claystone (3) is most abundant in western part of area.

(C) White to gray claystone (fire-clay)(5) interbedded with lignite (7) and dark gray to black, carbonaceous fire-clay (6); contains lenticular layers of clayey pebble conglomerate, and coarse-grained, quartzose clayey sandstone (9).

(D) White, yellow and red pisolitic claystone (8) with white gray claystone (fire-clay) (5); occurs in lenticular bodies in upper part of residual clay.

(E) White, yellow and red plastic claystone of residual origin, derived from slate (10) and volcanic rocks (11).

Figure 7. Generalized stratigraphic column showing lithologic features of clay-bearing units in Alberhill area.

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COLUMNAR SECTION FOR MAPS 1-5

Holocene	Qal	- Alluvium
Pleistocene	Qt	- Terrace deposits - includes older alluvium
Pliocene	Tr	- Repetto Formation - micaceous sandy siltstone with minor conglomerate
Late Miocene	Tp	- Puente Formation
	Tpsc	- Sycamore Canyon Member - buff sandstone, interbedded siltstone
	Tpy	- Yorba Member - chocolate brown to pink siltstone, locally diatomaceous
	Tps	- Soquel Member - buff sandstone, interbedded siltstone, local conglomerate
Middle Miocene	Tem	- El Modeno Volcanics - basalt
Early Miocene to Late Eocene	Tvs	- Vaqueros and Sespe Formations, undivided - buff, green, red and white clayey sandstone, conglomerate, and sandy clay, in part non-marine
Eocene	Tsa	- Santiago Formation - upper part massive white to buff coarse-grained sandstone and conglomerate; lower part thin-bedded brown sandstone, sandy siltstone and conglomerate
Paleocene	Tsi	- Silverado Formation - upper part greenish-gray sandstone; lower part coarse grained micaceous feldspathic sandstone, grit, conglomerate, clay and rarely lignite
Late Cretaceous	Kws	- Williams Formation, Schultz Ranch Sandstone Member - massive coarse-grained, buff sandstone, conglomerate
	Klh	- Ladd Formation, Holz Shale Member - gray to black silty shale and siltstone
	Klb	- Ladd Formation, Baker Canyon Conglomerate Member - sandstone and gray to greenish gray massive conglomerate
	Kt	- Trabuco Formation - red and white conglomerate and earthy sandstone
Jurassic to Cretaceous	Jsp	- Santiago Peak Volcanics - slightly metamorphosed andesite flows and breccias
	Jbc	- Bedford Canyon Formation - dark gray to black argillite with thin beds of feldspathic quartzite and limestone blocks

Igneous rocks - Kqd-quartz diorite; Kqm-quartz monzonite; Kgd-granodiorite; Kt-tonalite; Jtw-quartz latite porphyry

A section through the Peninsular Ranges batholith, Elsinore Mountains, southern California

Margaret S. Woyski, California State University, Fullerton, California 92634
Astrid H. Howard, California Institute of Technology, Pasadena, California 91125

LOCATION

The Elsinore Mountains in the Cleveland National Forest of southern California are situated in the northwestern Peninsular Ranges. They lie southwest of and overlook Lake Elsinore at elevations of 2,600 to 3,600 ft (790 to 1,100 m). The northeastern front of the mountains is a precipitous escarpment that rises 1,350 to 2,350 ft (410 to 720 m) above the lake at its base. The site is reached by taking the Ortega Highway (California 74) from either I-15 at Lake Elsinore or I-5 at San Juan Capistrano to the Main Divide Truck Trail (Fig. 1) where it intersects the Ortega Highway 0.2 mi (0.3 km) west of the summit. Road cuts along the Main Divide Truck Trail (a graded dirt road) are accessible by passenger car and, with caution, by bus. Two campgrounds are located at the intersection of the Ortega Highway with the Main Divide Truck Trail and group camping sites (limited to 70 people) are located just off the West Extension of the Main Divide Truck Trail. Make reservations with the Cleveland National Forest Trabuco District Office, Corona Ranger Station, 1147 East Sixth Street, Corona, California 91719, (714) 736-1811. There is a daily charge for all campgrounds.

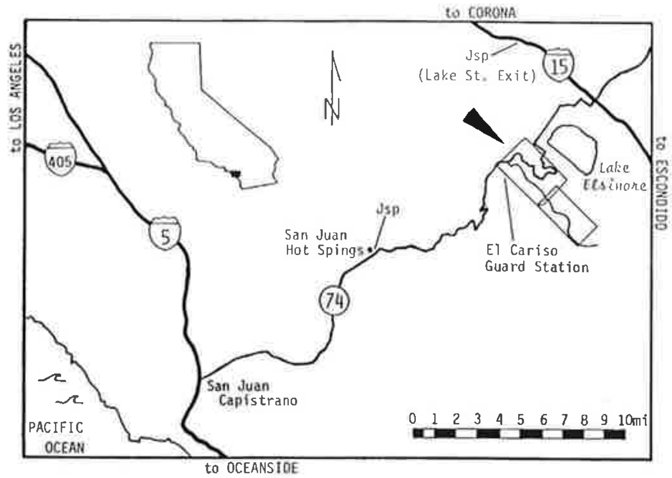

Figure 1. Location of Main Divide Truck Trail. Jsp = Santiago Peak Volcanics.

GEOLOGIC SIGNIFICANCE OF SITE

The major units of the northwestern Peninsular Ranges batholith that Larsen (1948) described are exposed along the Main Divide Truck Trail in the Elsinore Mountains. Pre-batholith country rock of folded Bedford Canyon metasediments is intruded by three units of the western batholith: San Marcos gabbro, Bonsall tonalite, and Woodson Mountain granodiorite. The San Marcos gabbro here shows its two chief facies (calcic hornblende norite and quartz-biotite hornblende norite) and its two major structures (large poikilitic hornblende crystals and rhythmic banding). It is intruded by both the Bonsall tonalite and Woodson Mountain granodiorite. The Bonsall tonalite contains distinctive ellipsoidal inclusions. The Woodson Mountain granodiorite exhibits a wide range of border effects.

Pegmatite and aplite dikes cut all the plutons and porphyry dikes cut the Woodson Mountain granodiorite. Spheroidal weathering is well displayed. Paleocene sediments, which rest on weathered granodiorite, are covered by Tertiary basalts. Remnants of the Perris erosion surface persist on the divide where they have been uplifted by motion on the Elsinore fault. The truck trail parallels the scarp of the active Elsinore fault zone and overlooks Lake Elsinore, a pull-apart basin filled by disrupted drainage.

The Peninsular Ranges batholith is interpreted as the root system of a Late Jurassic–Early Cretaceous volcanic arc off the southwest edge of the North American craton. The extrusive rocks of this arc (Santiago Peak volcanics) were erupted upon the earlier Mesozoic flysch deposits of the Bedford Canyon Formation. From 130 to 105 Ma the island arc system remained static while plutons, represented by those in the Elsinore Mountains, were intruded across the 31 mi (50 km) width of the arc. From 105 to 80 Ma the locus of plutonism migrated steadily eastward to the craton.

PENINSULAR RANGES BATHOLITH

The Peninsular Ranges batholith extends from the Transverse Ranges through southern California and the length of Baja California in a SSE trending belt 995 mi (1,600 km) long and spreads 68 mi (110 km) wide from the foothills along the Pacific coast to the Salton trough along the San Andreas fault. The batholith consists of hundreds of plutons that were intruded side by side, leaving a few screens of country rock between them.

Like many batholiths and volcanic arc systems, the Peninsular Ranges batholith is zoned longitudinally by differences in lithology, chronology, and chemistry. The plutons to the west are more varied and consist of gabbros and quartz gabbros, tonalites, and leucogranodiorites whereas to the east the plutons are mostly tonalites and closely related low K_2O granodiorites; nevertheless, all of the rocks appear to have the chemistry and mineralogy of I-type granites, indicating an igneous mantle source. An igneous mantle source is also indicated by the calcic alkali-lime index of 63% for the batholith. The zircon U-Pb ages of the western plutons range from 122–105 Ma and have no preferred distribution within the belt; on the other hand, the U-Pb ages of the eastern plutons range from 105–89 Ma and become progressively younger to the east. Hornblende and biotite K-Ar ages follow the same general pattern (Silver and others, 1979).

Several geochemical parameters are zoned across the batholith without regard for rock type indicating a common but changing magma source. The initial Sr isotopic ratio becomes increasingly radiogenic eastward; $\delta^{18}O$ values increase eastward from +6 to +12 with a sharp step that closely matches the 105 Ma isochron; Light Ion Lithophile (LIL) elements (e.g., Rb, Sr, Ba, Pb, U, Th) increase in the tonalites eastward (Silver and others, 1979); and Nd isotope ratios decrease eastward +8.2 to −6.4 (DePaolo, 1981). All these trends suggest that the magmas for the western part of the batholith are primarily from a primitive source region, probably the upper mantle, whereas the magmas in the eastwardly migrating belt involved small, but increasing eastward, assimilation of subducted sediments or highly altered upper portions of the oceanic crust. On the eastern edge, in the San Jacinto block, the still lower $\delta^{18}O$ and still higher $^{87}Sr/^{86}Sr$ values are attributed to the involvement of the southwestern edge of the North American craton in the melting process that produced the magmas (Taylor and Silver, 1978). Of the major elements, the alkalis increase eastward and Mg, Fe, and Ca decrease eastward but K varies independently (Baird and others, 1979).

The country rock on the west consists of low-grade metamorphic Triassic-Jurassic flysch-type sediments of the Bedford Canyon Formation and andesitic volcanics of the Santiago Peak Formation, whereas to the east it is medium-grade metamorphic Mesozoic (Paleozoic?) clastic sediments and minor limestones.

L. T. Silver and co-workers consider the Peninsular Ranges batholith to be the result of two stylistically different but related, and essentially continuous, magmatic events that formed a Late Jurassic–Early Cretaceous arc. In Late Jurassic times a plate to the west of the North American plate (possibly the Kula or Farallon plate) began to subduct beneath the North American plate creating a volcanic arc that formed to the southwest of the Precambrian craton (Silver, 1979). Rising magmas were intruded into and erupted onto the sediments of the Bedford Canyon Formation. The spatially random distribution of emplacement ages of plutons of all types on the western side of the batholith implies a static arc system from 130 to 105 Ma. At 105 Ma there was a change in the tectonic regime, possibly an increase in the rate of convergence. The arc then moved progressively eastward from 105 to 89 Ma as demonstrated by the regularly decreasing emplacement ages of the plutons in the eastern half of the batholith (Silver and others, 1979).

By Late Cretaceous time isostatic uplift and erosion, much greater to the east, had exposed the root system of these two related arcs in the Peninsular Ranges. This is deduced from: (1) the increasing grade of metamorphism in country rocks presumedly due to greater depth from west to east (Gastil, 1975); (2) the increasing divergence of the hornblende and biotite K-Ar ages from the zircon emplacement ages to the east that probably resulted from increased cooling time at greater depth; (3) the increase in size and homogeneity and therefore depth of the plutons to the east; and (4) the $\delta^{18}O$ evidence for a roof zone on the western edge (Taylor and Silver, 1978). The Peninsular Ranges batholith was separated from its initial position contiguous to mainland Mexico when the Gulf of California developed as an extension of the East Pacific spreading center, perhaps beginning as early as Miocene times (Silver and others, 1979). The tectonics of the Lake Elsinore region are now dominated by right oblique movement along the en echelon faults of the Elsinore fault zone (Weber, 1983).

ROCK UNITS OF THE ELSINORE MOUNTAINS

The folded and faulted **Bedford Canyon Formation** is the predominant country rock in the western zone of the batholith. It forms large masses, screens between plutons, numerous roof pendants, and abundant inclusions within plutons. The total thickness of the Bedford Canyon Formation cannot be determined because it is overlain unconformably by the Santiago Peak volcanics and the base is not known. It weathers to boulder-free tan soil that supports denser vegetation than the other rock units.

The formation is dark gray argillite or slate interbedded with fine arkosic or lithic quartzite, resulting in its distinctive alternation of dark and lighter beds from 0.4 to 59 in (1 to 150 cm) thick. Rare lenses of dark limestone contain poorly preserved fossils that have been dated as Jurassic (Callovian). No good horizon markers have been found in this thick sequence. Criscione and others (1978) have found two different ages by Rb/Sr isotope analysis, 176 Ma (Middle Jurassic) and 228 Ma (Permo-Triassic). Even close to contacts the Bedford Canyon metasediments are only mildly metamorphosed although inclusions within the plutons have been converted to hornfels. The foliation everywhere parallels the bedding, implying tight isoclinal folds. The beds strike NW, with local deviations to N–S, and dip steeply. Where crossbedding or graded bedding is detectable, it appears that the majority of beds are overturned.

Santiago Peak volcanics and hypabyssal rocks of Late Jurassic age occur along the west flank of the Peninsular Ranges and around the northwestern tip. Outcrops are most accessible in road cuts on I-15 or on California 74. (See Jsp on Fig. 1.) The volcanics resist erosion, producing steep slopes and high peaks, and weather to a rocky clay soil, supporting dense vegetation. The Santiago Peak Formation consists of andesites interbedded with quartz latites and minor rhyolites, as flows, dikes, tuffs, and volcanic breccias. The rocks are aphanitic, altered dark gray or green, with feldspar phenocrysts. Where intensely metamorphosed near contacts, the rock is converted to a dense hornfels with mere traces of phenocrysts (Larsen, 1948).

The **San Marcos gabbro** occurs as small plutons and as screens, roof pendants, and swarms of inclusions in the other plutonic rocks. The gabbro, especially the quartz-bearing facies, is prone to spheroidal weathering, reducing it to a red soil with a few round boulders. On divides the quartz-free gabbro weathers

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to sharp peaks with distinctive conical profiles. The outstanding characteristic of the San Marcos gabbro is its variability even within single outcrops. The grain size ranges from 0.5 to 20 mm although the average is about 1 mm. The mafites range from 20% to 55%. The composition ranges from typical gabbronorite (augite and hypersthene with labradorite, an_{55-70}) to calcic olivine norite (hypersthene and olivine with very calcic plagioclase, an_{89-95}) or else to quartz biotite norite (quartz, biotite, and hypersthene with intermediate plagioclase, an_{46-59}). Hornblende is conspicuously erratic. Some hornblende is an early alteration product, some is a late hydrothermal mineral, but most is a primary mineral that occupies interstices with poikilitic texture. It may constitute as little as several percent of the rock to as much as 50%. Miller (1937) suggests that irregular distribution of water in the magma may explain its erratic character. Miller (1937) grouped all the various gabbros into one unit, "San Marcos," because their ages and occurrences are similar and the variations in mineralogy, although great, show continuous gradation from one extreme to the other.

One common internal structure is rhythmic banding in which layers of hornblende and pyroxene alternate with layers of plagioclase. Another structure is a penetration of finer gabbro by coarser, more hornblende-rich gabbro, as an angular network or uncommonly, as matrix around nodules. Orbicular structures of radiating olivine, hypersthene, and plagioclase occur locally (Miller, 1938).

The **Bonsall tonalite** and other tonalites intrude the gabbro and are intruded in turn by the granodiorites. The gold veins of the Pinacate district are almost exclusively associated with the Bonsall tonalite. This tonalite is noted for its abundant and well-oriented inclusions but few schlieren. Most field geologists have used the nature of schlieren, inclusions, and xenocrysts to distinguish the tonalite plutons. That the size, shape, and abundance of inclusions is distinctive for each pluton, regardless of country rock, suggests that these inclusions are related to the source of magma.

The tonalite is medium gray, coarse-grained from 0.5 to 10 mm, composed of 50%–60% andesine, 20%–25% quartz, 5%–20% biotite, 5%–15% hornblende, and less than 10% orthoclase. The inclusions contain the same minerals as the tonalite but with a greater proportion of mafites.

The **Woodson Mountain granodiorite** forms large plutons that intrude the tonalite and gabbro in the western Peninsular Ranges. Engel (1959) reports that, where vertical sections are distinguishable, contacts dip gently as if the plutons flared outward upward. The Woodson Mountain granodiorite is light gray to tan in fresh specimens with medium-coarse texture. It is composed of 30%–35% oligoclase, 10%–30% microcline perthite, 30%–40% quartz, and 1%–10% mafites (chiefly biotite). The granodiorite appears to have intruded late in the igneous sequence when the surrounding rock was cool enough to induce chilled, streaky, more mafic borders and brittle enough to shatter at the contacts.

The **Paleocene** was an epoch of profound **lateritic weathering** during which alumina-rich clays and iron-rich red soils were developed in situ. These were covered by fresh and brackish water clays and sandstones of the lower **Silverado Formation.** Deep red soil that was developed during the unique Paleocene climatic conditions serves as a time-marker in the Peninsular Ranges. It is well displayed in road cuts at the summit of the Ortega Highway. Paleocene clays in the Elsinore Trough, which are used for ceramics, constitute one of the major mineral resources of southern California.

Four or five horizontal basalt flows of the **Santa Rosa volcanics** form extensive northeast-trending mesas close to the Elsinore fault zone. The basal flow is alkali basalt; overlying flows are tholeiitic. Elsinore Peak and adjacent hills are composed of basalt that weathers to brown soil covered by grassy vegetation. Aphanitic porphyritic dikes with chilled margins that cut the youngest of the plutonic rocks may be associated with the volcanic activity that produced the Santa Rosa basalts.

The **Perris Surface** is one of the oldest and most extensive of the six erosion surfaces on the Perris block, which is bounded by the Elsinore and San Jacinto faults. It lies beneath the Santa Rosa volcanics that were erupted at 8.3 Ma, before the other surfaces were developed (Woodford and others, 1971). On this basis, the erosion surface beneath the volcanics of Elsinore Peak can be correlated with the Perris surface.

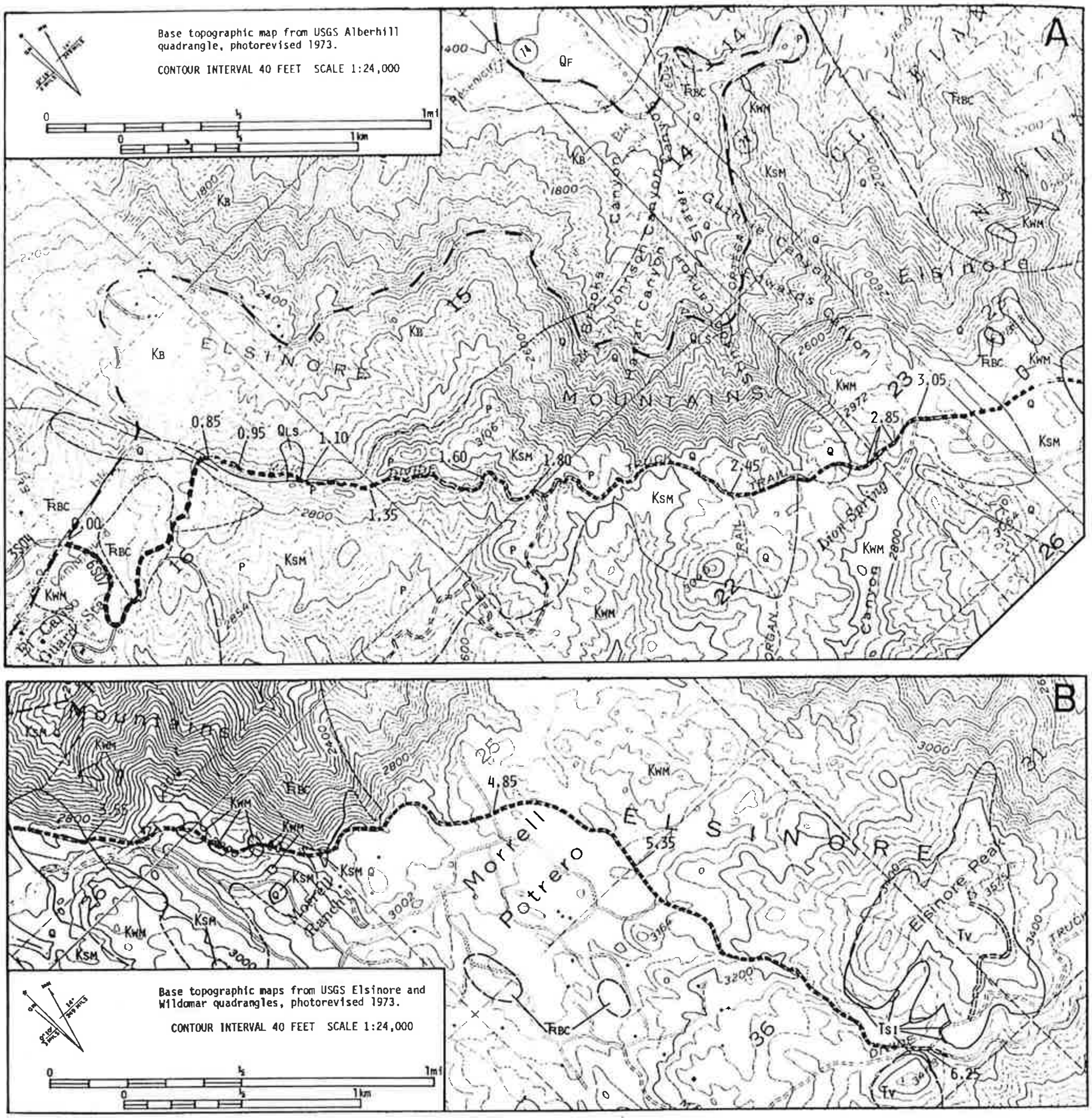
The **Elsinore fault zone** extends along the southwest edge of the Elsinore Trough running from Corona southeastward about 50 mi (80 km). The trough is bordered by low hills to the northeast and by the steep Santa Ana and Elsinore Mountains to the southwest and parallels the more seismically active San Jacinto Fault to the northeast and Newport-Inglewood Fault to the southwest. All along the Elsinore fault zone there is geomorphic evidence of oblique-slip movement through the late Quaternary. Weber (1983) has proposed 5.6 to 6.8 mi (9 to 11 km) of right lateral displacement based on offsets of thin bodies of gabbro, aberrant foliation in the Bedford Canyon Formation, and the contact between the Santiago Peak volcanics and the Bedford Canyon Formation. Uplift of the Elsinore Mountains along the Elsinore fault zone has cut off the San Jacinto River, which is now impounded in a pull-apart basin forming Lake Elsinore.

ROAD LOG

Individual outcrops along the Main Divide Truck Trail (Fig. 2) are described in road-log format with mileages given to the nearest 0.05 mi (0.1 km).

0.0 mi (0.0 km). Ortega Highway California 74. At El Cariso campgrounds, take Forest Service road 6S07 southward; this is the Main Divide Truck Trail.

0.85 mi (1.4 km) Tonalite/Gabbro/Granodiorite Contact. Overlook. On the NE face of the north corner of the truck trail, Bonsall tonalite is in sharp contact with the San Marcos gabbro, with no change in grain size; within a few centimeters of the contact, however, the tonalite shows a very faint foliation



QLS	Landslide	KB	Bonsall tonalite	———	Contact
QF	Fanglomerates & terrace deposits	KSMP	San Marcos gabbro, poikilitic hornblende	-----	Approximate contact within 20'
Tv	Volcanic rocks	KSMQ	San Marcos gabbro, quartz biotite	— —	Contact after Larsen, 1948 and Engle, 1959
TSI	Silverado Formation	ŦRBC	Bedford Canyon Formation	—┴—	Strike and dip of bedding
KWM	Woodson Mountain granodiorite			——	Fault, dashed where approximate

Figure 2. Geologic Map. Main Divide Truck Trail. A. Miles 0.0 to 3.53 (0.0 to 5.7 km); B. Miles 3.53 to 6.25 (5.7 to 10.1 km).

parallel to the contact. An apophysis of tonalite cuts the gabbro. Absence of chilling suggests that the gabbro had solidified but not yet cooled when the tonalite was intruded. (A similar contact can be seen at 1.1 mi [1.8 km]). On the NW face Woodson Mountain granodiorite is in contact with the gabbro. It is coarse and mafic with large inclusions of Bedford Canyon metasediments and gabbro. The San Marcos gabbro here is the calcic poikilitic hornblende facies of this highly variable formation. At the top of the cut, the gabbro shows its typical red clay weathering product. Cutting the gabbro along an aplite dike is a shear zone about 1.6 ft (0.5 m) wide that has been thoroughly altered. This shear zone lies along the trend of a presumed fault valley north of the Ortega Highway.

From this vantage point, one can look northeast over the 1,600-ft (490-m) scarp of the Willard Fault across Lake Elsinore to the scarp of the Glen Ivy North Fault. These two faults are strands of the Elsinore fault zone that defines the Elsinore Trough. In the middle distance are clay pits in Paleocene deposits at Alberhill. Beyond, on the low plateau of the Gavilan, the Steele Valley pluton of granodiorite is outlined by its blanket of large boulders. The Pinacate gold district borders the pluton to the right and behind it. In the distance, the Lakeview Mountains rise from the Perris plain, showing their arcuate structure. On a clear day, Mount San Jacinto, Mount San Gorgonio, and Mount San Antonio can be seen on the horizon.

0.95 mi (1.5 km). Bonsall Tonalite. Here the typical inclusions in the Bonsall tonalite are conspicuous. They are small (3–6 in; 8–15 cm), elliptical and compose several percent of the rock. About half the inclusions pitch to the SE about 30°. A few pegmatite stringers parallel this flow direction also. Toward the gabbro contact, larger angular inclusions become more abundant. Inclusions in the Bonsall tonalite appear to be of two generations: an early ubiquitous set, with granoblastic texture, of uniform size, elliptical shape, and oriented in parallel that apparently came from considerable depth; and a late localized set, of heterogeneous size, irregular shape, without orientation, and composed of identifiable local rock that apparently came from the country-rock roof. The early set of inclusions distinguishes the Bonsall tonalite from other tonalites.

1.35 mi (2.2 km). Poikilitic Hornblende Gabbro. At this wide parking area all three major pluton types of the western Peninsular Ranges batholith can be seen in close proximity. To the north, scattered boulders of gray Bonsall tonalite in poor sandy soil outline an elliptical pluton about 1.9 mi (3 km) long that lies between this overlook and the base of the scarp. Even from a distance, the ubiquitous dark inclusions make the boulders look spotted. To the south, far hillsides are littered with the huge pinkish-white exfoliation boulders in sandy grus by which the Woodson Mountain granodiorite can be recognized. This pluton extends for 6 mi (10 km) to the southwest and 12 mi (20 km) to the southeast. In the road cuts at the parking area is San Marcos gabbro.

At this location, the gabbro contains poikilitic hornblende crystals up to 5 in (12 cm) across that enclose small grains of calcic plagioclase (anorthite) and olivine now altered to brown spots of iddingsite and iron oxides. Most of the original hypersthene has been replaced. Between the hornblende crystals is labradorite and red biotite. The small grains of calcic plagioclase and olivine were probably precipitated in response to assimilation of small amounts of aluminous sediments as predicted by Bowen's reaction principle.

1.6 mi (2.6 km). Hornblende Streaks in Gabbro. This road cut contains narrow pegmatite dikes that cut black dike-like streaks of poikilitic hornblende in the hornblende gabbro. The association of the hornblende streaks with pegmatite dikes and the dike-like character of the streaks suggests that some poikilitic hornblende may have developed in response to introduction of fluids from invading magma.

1.8 mi (2.9 km). Rhythmically Banded Poikilitic Gabbro. Here the gabbro contains a zone of rhythmic banding 23 ft (7 m) wide where dark hornblende-rich layers 0.5 in (1 cm) wide alternate with light plagioclase-rich layers. The composition change is probably due to sequential depletion of constituents as crystallization progressed.

2.45 mi (3.9 km). Quartz-Biotite Gabbro. On the hillside near the top of the rise are light gray spheroidal boulders of quartz-biotite gabbro that contain biotite flakes and quartz as well as hornblende, labradorite, and minor hypersthene. This gabbro, unlike the poikilitic gabbro, contains inclusions of fine-grained hornfels derived from the Bedford Canyon metasediments. Quartz-biotite gabbro occurs close to metasediment contacts both in this area and northwest of the Ortega Highway. The presence of biotite and quartz, in total amounts from 5% to 25%, could well be the result of assimilation of Bedford Canyon silty shales. The finer grain size (suggesting rapid heat loss), the association with the Bedford Canyon formation, the presence of metasediment inclusions, and the increase in minerals (quartz and biotite) that occur in metasediment all point to this origin.

2.85 mi (4.6 km). Granodiorite/Gabbro/Metasediment/Porphyry Contacts. At the head of Morrell Canyon are three road cuts in the space of 0.1 mi (0.2 km). The western cut contains a border facies of the Woodson Mountain granodiorite that is more mafic and finer-grained than the typical granodiorite. (A larger exposure can be seen at mile 2.6 [4.2 km]). It contains small inclusions of gabbro so numerous that it forms an intrusion breccia. Large inclusions of metasediment occur in both the granodiorite and the quartz biotite gabbro on the east end of the cut. Both gabbro and metasediment inclusions have been metamorphosed to fine hornfels. Cutting the border facies granodiorite is a porphyritic dike about 3 ft (1 m) wide. The feldspar phenocrysts range from 1 mm at the chilled margin to 4 mm in the center of the dike. The groundmass is dark gray, fine aphanitic, drawn into platy foliation parallel to and adjacent to the contacts. All these relationships are shown as well in the middle road cut. The east cut contains a very coarse facies of the Woodson Mountain granodiorite with grain size of 5–10 mm. The outcrop is cut by two porphyry dikes 12 and 50 in (30 and 130 cm) wide, faulted at top and bottom. Inclusions of gabbro in the granodiorite are few, rounded, and average 4 in (10 cm) in diameter. Wide-set joints are deeply weathered.

These outcrops show that gabbro intruded metasediments, incorporating large blocks into the magma. Subsequently, granodiorite intruded along the gabbro-metasediment contact. Near the contact the granodiorite was chilled to a more mafic border facies and incorporated inclusions of both gabbro and metasediment. An alternate interpretation would attribute the mafic border to assimilation of gabbro; however, the gabbro inclusions consistently show sharp contacts with no disintegration zones or schlieren. Although the east outcrop is near the contact, the granodiorite has an uncommonly coarse grain size, perhaps due to the addition of water to the magma from the Bedford Canyon Formation, or more likely, due to offset on a fault. Much later, probably during the late Tertiary volcanic activity of the region, porphyry dikes were introduced into the cold rock.

3.55 mi (5.7 km). Granodiorite/Metasediment Contact. At this road cut (Fig. 2B), just past a small gully running northwest, granodiorite is in intrusive contact with the Bedford Canyon metasediments. The Bedford Canyon Formation shows its distinctive bedding of alternating layers 3 to 9 in (8 to 24 cm) thick. One set of layers is fine sandstone, metamorphosed to non-foliated quartzite that resists weathering. The other set is clayey siltstone, metamorphosed to fine-grained mica schist that weathers readily along the foliation. The foliation is parallel to the bedding except where it is deformed around a bend (fault?) in an aplite dike.

5.35 mi (8.6 km). Morrell Potrero. Here and at mile 3.05 (4.9 km) are former river beds on an uplifted erosion surface.

6.25 mi (10.1 km). Basalt. Paleocene Surface. Silverado Formation. At this large parking area, Elsinore Peak, with its microwave relay station, rises to the northeast. Elsinore Peak and the hill to the southwest are composed of basalt with glassy-matrix; some is vesicular at the top of the SW hill. On weathered surfaces, the 1 mm phenocrysts of olivine make brown spots on the otherwise smooth surface. The base of the flows, which were extruded onto the Perris erosion surface, lies at about the level of the parking area. At the foot of the hillside to the southwest is a layer of white medium-grained, moderately-sorted micaceous feldspathic sandstone about 4 ft (1.2 m) thick. It is lithologically similar to the white sands of the Silverado Formation that crop out in the Elsinore Trough. Across the meadow to the north can be seen a ledge of the same flat-lying sandstone. About 0.05 mi (0.1 km) down the road to the southeast is mottled red and white residual clay that grades downward into weathered granodiorite. This clay marks the Paleocene erosion surface and basal unconformity. About 0.2 mi (0.3 km) up the service road to the microwave station is a pull-out from which the basalt is easily accessible.

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Controls on orogenesis along an ocean-continent margin transition in the Jura-Cretaceous Peninsular Ranges batholith

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INTRODUCTION

Mesozoic batholiths emplaced along much of the Cordillera show an increasing continental basement signature from their oceanic to landward positions. In many cases it can be demonstrated that the more outboard assemblages in these batholiths evolved on oceanic to transitional crust. A classic example of one of these oceanic to continental basement transitions occurs in the Jura-Cretaceous Peninsular Ranges batholith (PRb) of the southern Cordillera, which extends >800 km from Riverside California to the 28th parallel in Baja California (Fig. 1). The transition in this batholith between a western zone with strong oceanic basement affinity and an eastern zone of strong continental affinity is narrow and sharply defined by across-strike changes in the petrology and geochemistry of plutons (e.g., Silver et al., 1979; DePaolo, 1981; Gastil et al., 1990; Walawender et al., 1991, Tate and Johnson, 2000). This zone is particularly striking because a number of other geologic features also define it including a belt of distinctive clastic volcaniclastic metasediments (e.g., Gastil et al., 1975; Gastil and Miller, 1984; Gastil, 1993), a discrete belt of contraction structures (e.g., Gastil et al., 1975; Todd et al., 1988; Thomson and Girty, 1994; Johnson et al., 1999a; Schmidt, 2000), and a sharp contrast between western shallow and eastern deep crustal levels (e.g., Todd et al., 1988; Grove, 1994). These across-strike transitions are a hallmark of the PRb, perhaps better developed here than in any other batholith in the world.

Three tectonic models have been proposed for the PRb, in part to explain the characteristics of the transition zone. These include construction of the Mesozoic batholith across a pre-existing ocean-continental lithosphere transition, collision of a fringing oceanic arc by backarc basin collapse, and collision of an exotic island arc (Fig. 2). The first and most fixist model for the batholith, that of an inherited pre-Mesozoic ocean-continent

Figure 1. Map of western North America showing major Mesozoic Cordilleran batholiths and selected Paleozoic-Mesozoic tectonic features of the southwest North American margin discussed in text. Possible continuation of Ouachita-Marathon suture across Mexico from Stewart (1988).

was sutured to the margin in Early Cretaceous time, whereas the southern half (Alisitos arc segment) collided in mid- to Late Cretaceous time. They proposed that the two segments were separated by an ancestral transform fault that has been reactivated to form the presently active Agua Blanca fault.

A third model, built on the ideas of others such as Gastil et al. (1975, 1978; 1981), Todd et al. (1988), and Griffith and Hoobs (1993), is that the western zone is an exotic island arc complex, which was sutured to the North American margin in the Early Cretaceous (Fig. 2C; Sedlock et al., 1993; Johnson et al., 1999a; Dickinson and Lawton, 2001). Intrinsic to this model is evidence that western zone rocks show little influence of continental crustal material including lack of zircon inheritance in both plutonic and volcanic rocks, paucity of quartz and K-feldspar components in volcanogenic sediments, and plutons with strong island arc geochemical and isotopic affinities (Silver and Chappell, 1988; Johnson et al., 1999a). The above authors differ on details of facing of subduction zone(s) that closed the intervening ocean basin and only one possible scenario is shown in Figure 2C.

The range of possible tectonic scenarios discussed above arises partly from researchers working in different areas separated by extensive regions that are poorly known. Most of this previous work has occurred in the northern half of the batholith. Our work has focused on the less studied section in the Sierra San Pedro Martir region of Baja California, Mexico, where relationships in the transition zone of the batholith are particularly well exposed. We have found several significant differences between relationships in this region and those described from farther north, suggesting that the northern and southern parts of the PRb do not share an identical evolution. Moreover, published data as well as our own reconnaissance work suggest that these relationships change across the presently active Agua Blanca fault, and thus, we concur with Gastil et al. (1975, 1981) that the Agua Blanca fault has a Mesozoic history and served as an important tectonic break during PRb evolution.

Below we examine the transition zone in the PRb and its relationship to the western and eastern zones. We further explore the implications of these relationships for tectonic models of the PRb acknowledging that other significant problems beyond the scope of this paper remain with regards to the relationship of the PRb to North America (e.g., paleomagnetic evidence that the batholith evolved in a more southerly location, Hagstrum et al., 1985 and suggestions for a W-northwest-striking Jurassic strike slip fault, the Mojave Sonora Megashear of Silver and Anderson, 1974, or older transform fault, Dickinson, 2000). We argue that the transition zone is a distinct entity in the PRb and has been closely tied to the eastern zone of the batholith throughout the Mesozoic. General aspects of the transition zone appear to remain consistent along-strike between northern and southern parts of the batholith including comparable degrees of Mesozoic basin formation, contractional deformation, and metamorphic and denudation histories. However, specific aspects appear to change across the Agua Blanca fault, including the age of basins and the type of detritus they collected as well as the location and timing of deformation in mylonite shear zones. These differences occur in concert with distinct variation between northern and southern parts of the western oceanic arcs, supporting our contention that the southern, Alisitos arc segment collided with the margin in Early Cretaceous time, whereas the northern, Santiago Peak arc segment evolved as an inherited oceanic terrane, across which the batholith was constructed during Jura-Cretaceous time (Wetmore and Paterson, 2000). Our studies suggest that collision of the Alisitos island arc was only one episode in a long history of orogenesis in the PRb. Other processes must have driven orogenesis along a much larger portion of the PRb during Jura-Cretaceous time masking the effects of local collision.

PREVIOUS WORK ESTABLISHING THE ZONED NATURE OF THE PENINSULAR RANGES BATHOLITH

The Peninsular Ranges batholith traditionally has been divided into distinct western and eastern zones with the boundary defined by a number of different criteria that include: prebatholithic assemblages; pluton lithologies; metamorphic histories; major, trace, and isotopic geochemistry of plutons; geophysically defined crustal structure; and magmatic and cooling histories. Typically, this boundary between western and eastern zones is designated by a single line based on one or more of the above criteria. This practice is problematic because the location of this boundary changes depending on which criteria are chosen to differentiate the two zones (compare for example the location of the $\delta^{18}O$ and magnetite/ilmenite lines on Figure 3). Moreover, some criteria such as the $\delta^{18}O$ step in plutons vary sharply across a distance <5 km, while others, such as initial Sr isotopic values vary across distances >40 km. Thus, in contrast to previous convention, it may be more accurate to define a transitional zone in the batholith, which occurs between western and eastern zones and that encompasses the transitions in criteria listed above. Below we argue that this transitional zone is geologically distinct from the crustal zones to either side, and variation in geochemical, geophysical, and geological parameters by which it is defined along strike are critical to understanding the tectonic evolution of the PRb.

The Peninsular Ranges batholith intrudes a series of northwest-trending, prebatholithic, lithostratigraphic assemblages (Fig. 4) that include from west to east (Gastil, 1993): (1) a Triassic-Cretaceous continental borderland assemblage (not shown in Figure 4); (2) a Jura-Cretaceous volcanic arc and forearc assemblage; (3) a Triassic(?)-mid Cretaceous clastic and volcaniclastic flysch assemblage of uncertain tectonic origin; (4) an Ordovician-Permian slope-basin clastic assemblage; and (5) an upper Proterozoic-Permian miogeoclinal carbonate-siliciclastic assemblage. These have been assigned to distinct belts that include a western zone consisting of the borderland and volcanic arc assemblages and an eastern zone composed of

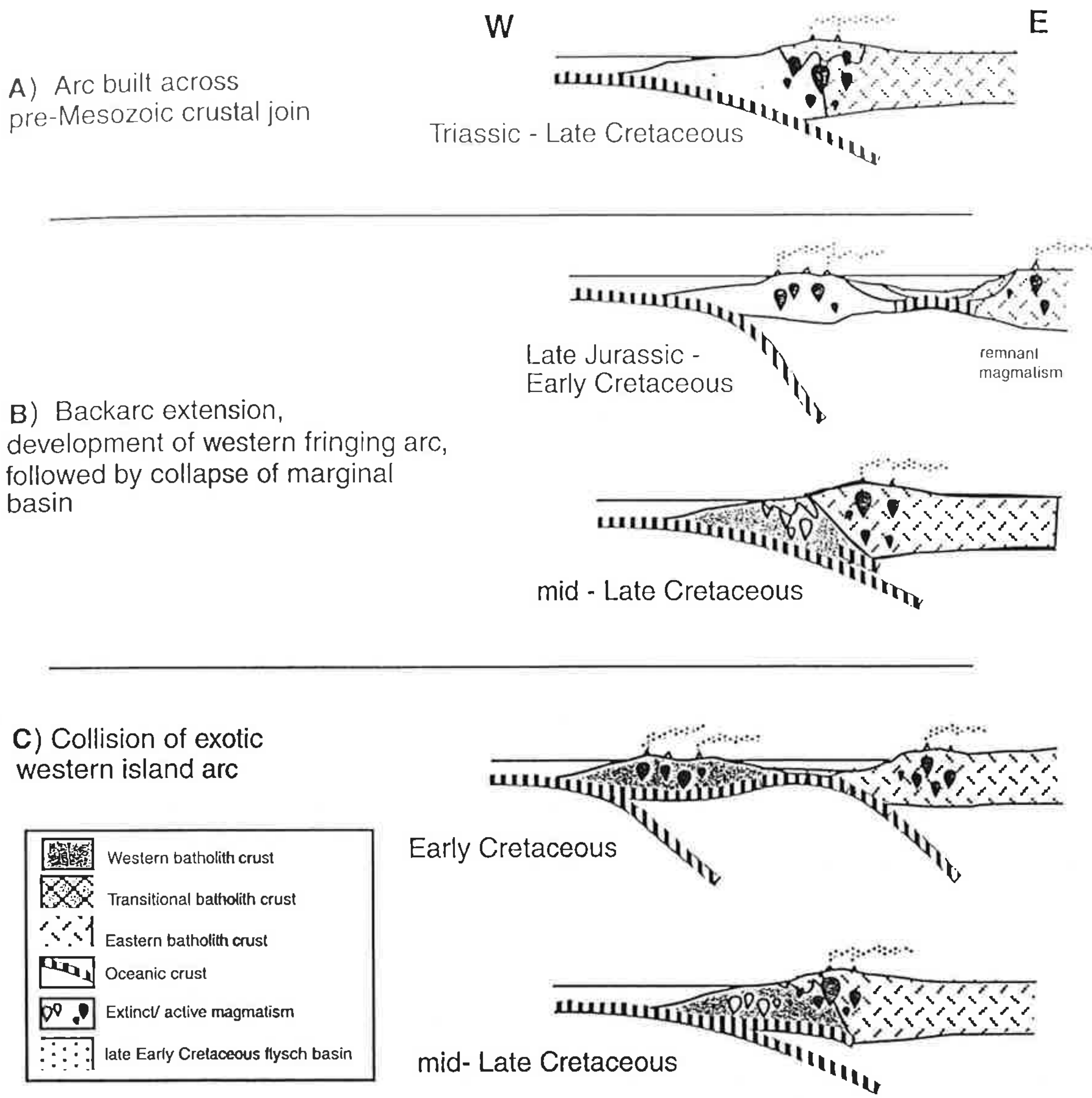


Figure 2. Tectonic models suggested for the Mesozoic evolution of the PRb. See text for historical development of models and corresponding references. (A) Western zone evolved adjacent to the rest of the batholith, separated by an inherited crustal boundary. (B) Backarc basin formation rifted western zone into a fringing arc position in Late Jurassic-Early Cretaceous time followed by collapse of intervening marginal basin and reattachment to North America. (C) Development of the western zone as an exotic island arc during Jurassic and Early Cretaceous time followed by collision and suturing of the exotic arc with North America in the latest Early Cretaceous.

transition, largely originates from work in southern California. Accordingly, Walawender et al. (1991) suggested that the PRb formed above a single subduction zone and developed its zoned character by a shallowing subduction angle and eastward arc propagation beginning at ca. 105 Ma. Thomson and Girty (1994) further suggested an early Mesozoic tie between the western and eastern zones of the batholith from a metasedimentary intra arc basin unit that overlaps both batholith zones and is intruded by a pluton that yielded a discordant Triassic age.

A second model, involving partial evolution of the western zone as a fringing oceanic arc, has been advocated by a number of researchers working in both southern and Baja California (Fig. 2B; Gastil et al., 1975, 1981; Rangin, 1978; Todd et al., 1988; Griffith and Hoobs, 1993; and Busby et al., 1998). They suggested that back-arc extension, initiated in Jurassic time, rifted the western, oceanic part of the batholith to a fringing arc position. Collapse of the marginal basin located between the two arc fragments, either by subduction within the basin (Gastil et al., 1975, 1981) or by crustal shortening (Busby et al., 1998), occurred in the mid-Cretaceous, followed by eastward arc migration. Gastil et al. (1975, 1981) further suggested that the northern half of the western zone (Santiago Peak arc segment)

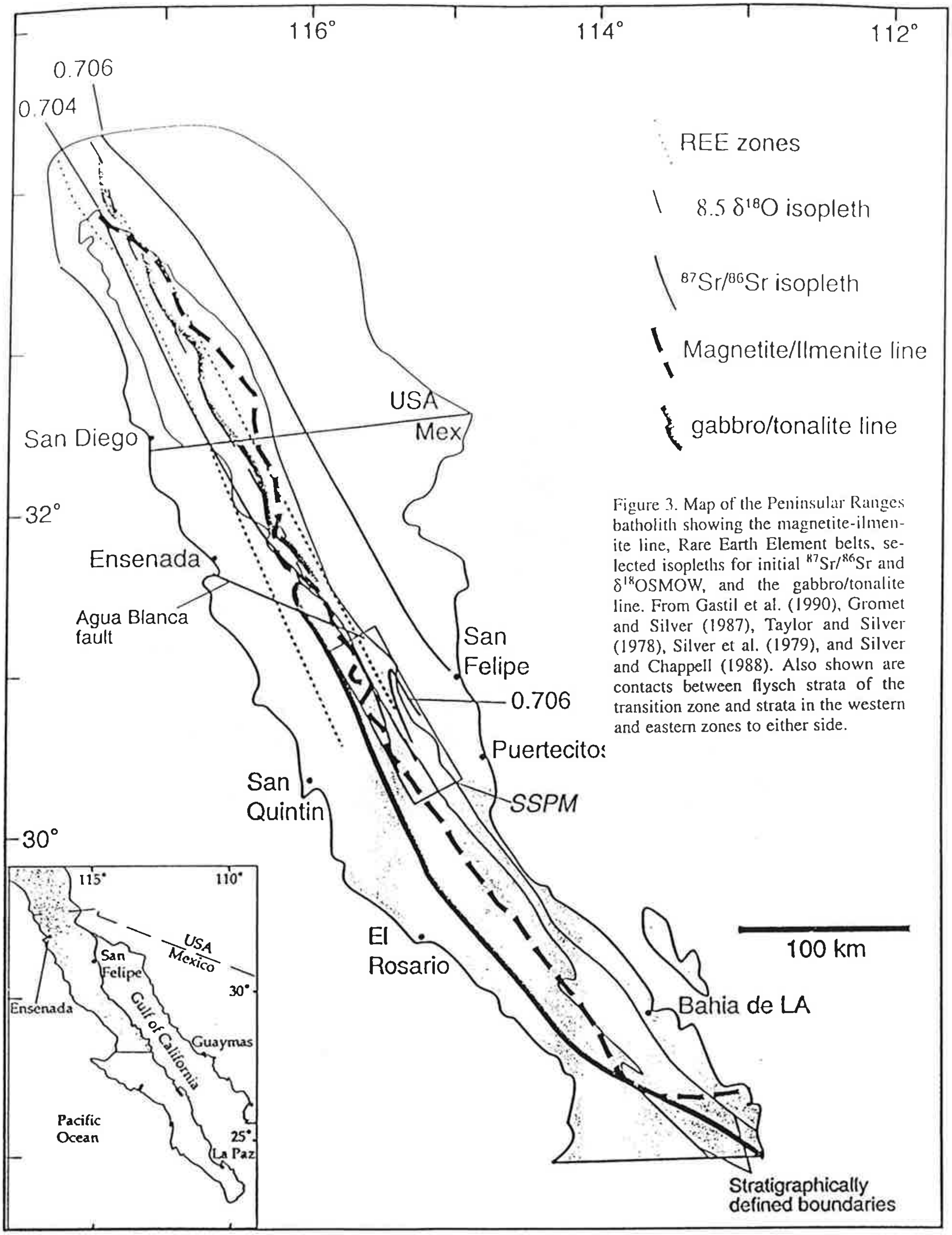
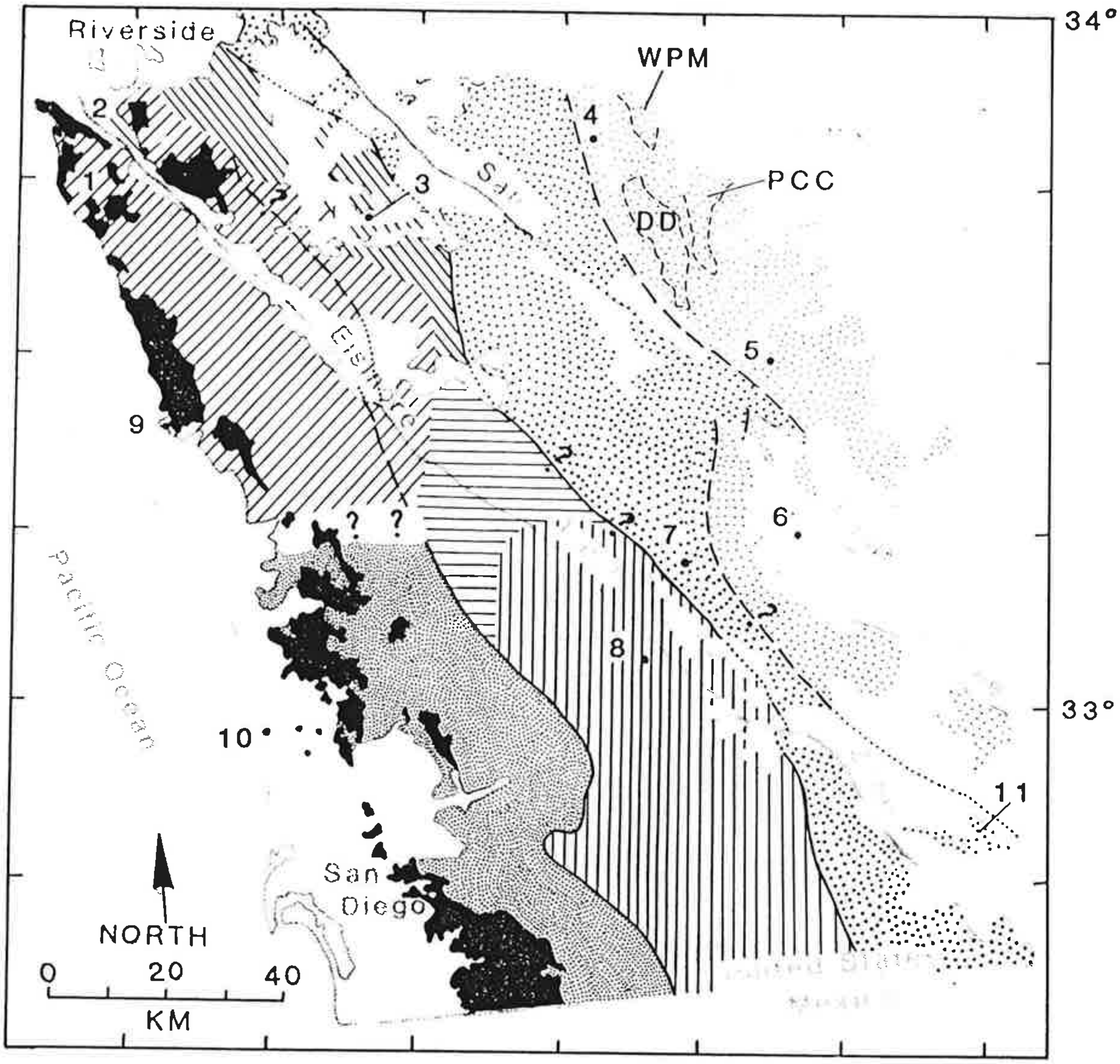


Figure 3. Map of the Peninsular Ranges batholith showing the magnetite-ilmenite line, Rare Earth Element belts, selected isopleths for initial $^{87}Sr/^{86}Sr$ and $\delta^{18}O_{SMOW}$, and the gabbro/tonalite line. From Gastil et al. (1990), Gromet and Silver (1987), Taylor and Silver (1978), Silver et al. (1979), and Silver and Chappell (1988). Also shown are contacts between flysch strata of the transition zone and strata in the western and eastern zones to either side.



ZONE I Metamorphosed volcanic and volcaniclastic rocks

- Santiago Peak Volcanics (latest Jurassic) and quartz latite porphyry including Temescal Wash Quartz Latite Porphyry (Larsen, 1948)
- Metavolcanic rocks of unknown age with scarce metasedimentary layers

ZONE II Metamorphosed submarine fan deposits

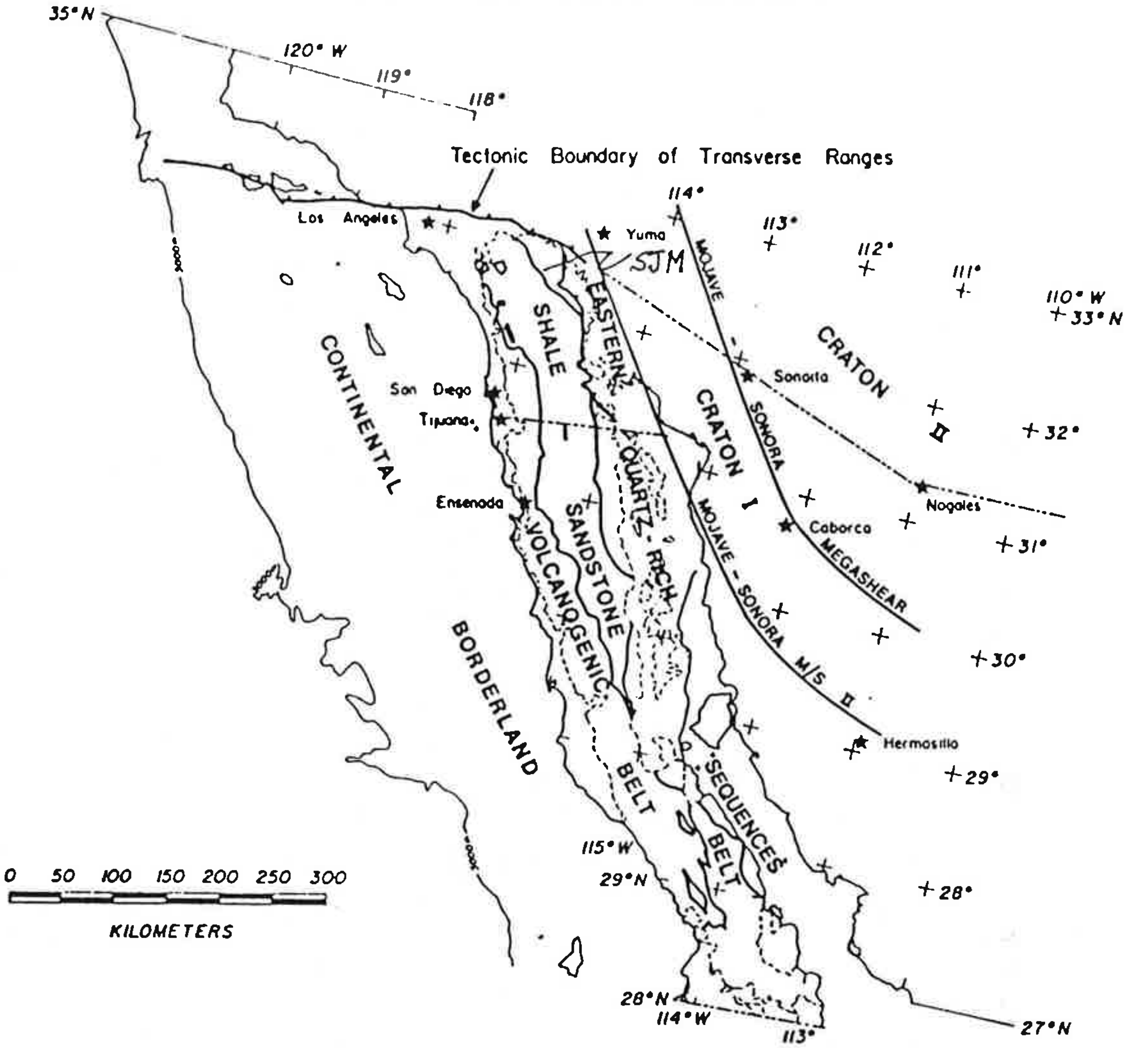
- Bedford Canyon Formation (Jurassic and Triassic) and rocks generally regarded as equivalent; west of dashed line, includes significant metavolcanic component
- French Valley Formation (Triassic?)
- Julian Schist (Triassic?) and rocks generally regarded as equivalent
- Undivided metasedimentary rocks

ZONE III Metamorphosed miogeoclinal and transitional(?) sedimentary rocks

- Metasedimentary rocks transitional between Zones II and III (Mesozoic? and Paleozoic)
- Miogeoclinal metasedimentary rocks (Paleozoic? and late Precambrian?); short dashed lines outline rock units cited in text, DD=Desert Divide Group, WPM=Windy Point metamorphic rocks, PCC=Palm Canyon Complex

FIG. 32-2. Prebatholithic zones in northern Peninsular Ranges batholith generalized from distribution of rock types in screens; plutonic rocks not shown. Solid lines are approximate boundaries between zones, long dashes separate subzones that have different ages, lithologies, or depositional settings. Data from Strand (1962), Rogers (1965), and sources cited in text, modified by unpublished data of authors. Locations mentioned in text: (1) Santa Ana Mountains, (2) Temescal Wash, (3) Winchester, (4) San Jacinto Mountains, (5) Santa Rosa Mountains, (6) Borrego Valley, (7) Ranchita, (8) Julian, (9) Camp Pendleton, (10) Del Mar, (11) Coyote Mountains.

PALINSPASTIC RECONSTRUCTION OF THE NORTHERN PENINSULA OF BAJA CALIFORNIA AND ADJACENT MEXICO



Lithotectonic belts of the northern Peninsular Ranges and adjacent Sonora, palinspastic base. CONTINENTAL BORDERLAND, VOLCANOGENIC BELT, SHALE SANDSTONE BELT and EASTERN QUARTZ-RICH SEQUENCES from Gastil et al., 1975. Position of MOJAVE-SONORA MEGASHEARS from Anderson and Silver, 1979, and L.T. Silver, written communication, 1982. CRATON I AND CRATON II are, in part, diagrammatic, and are meant to show only that there could be considerable complexity in the craton immediately to the east of the Peninsular Ranges.

Restoration of Gulf of California using pole positions of Chase, quoted in Larson et al., 1972, and maximum closure of Gulf. Although this restoration may over-estiamte the amount of separation after rifting commenced, it potentially allows for a small amount of crustal extension prior to the initiation of the active rift system.

Figure 3 Hill and Silver, in preparati

Summary Comparison of West and East Sides of Peninsular Ranges Batholith

FEATURES	WEST SIDE	EAST SIDE
Prebatholithic Country Rocks	Volcanic and volcaniclastic sedimentary rocks; common preservation of primary structures; low grade of metamorphism except near east side.	Dominantly shaly to quartzose clastic rocks with minor marbles; strong metamorphic foliation; primary structures largely destroyed; middle to upper amphibolite metamorphism.
Plutonic Structure	Smaller epizonal plutons commonly isolated within prebatholithic rocks; elongate WNW to NNW; increasingly intense deformation overlay towards the east.	Larger mesozonal plutons mutually contiguous, with prebatholithic rocks minor and confined to isolated pendants and screens; plutons elongate NNW to N-S; deformational patterns tend to be localized along the west side or on pluton margins.
Timing	Static magmatic arc; emplacement ages 105 to more than 130 my; cooling history brief except near east side of belt.	Migrating magmatic arc; emplacement ages near 105 my on west side, becoming progressively younger toward extreme east side, 80 to 90 my; lengthy thermal history, extended up to 25 my.
Lithologies	Diverse lithologies ranging from Gabbros and quartz gabbros to abundant tonalites (relatively mafic and inclusion-rich) with fairly abundant silica-rich leucogranodiorites and rare adamellites. Hornblende prominent throughout the suite.	Limited range of lithologies dominated by sphene-hornblende-biotite-bearing tonalites and low K_2O granodiorites. Gabbros extremely rare and more felsic rocks uncommon.
Chemistry	Calcic suite depleted in LIL trace elements; comparatively moderate REE fractionations except near east side.	Calcic suite with increasing LIL trace element content toward east side. Marked heavy REE depletion and light REE enrichment except near west side.
Isotopic Characteristics	Low initial strontium isotopic ratios, 0.7030-0.7050, tending to be higher toward the east; $\delta^{18}O$ values, +6 to +8.5, increasing toward the east. Evidence of meteoric water or seawater interactions in roof zone of batholith on extreme west side. Lead isotopes are primitive arc-like approaching MORB value on the extreme west side with a radiogenic gradient to the east $^{206}Pb/^{204}Pb$ 18.5 to 19.2.	Marked initial strontium isotopic gradient increasing from 0.7045 to 0.7080 toward east; unusually high $\delta^{18}O$ values, +9 to +12 or more tending to increase toward east but with some reversals. Lead isotopes are generally more radiogenic with values of $^{206}Pb/^{204}Pb$ 19.0 to 19.5.

STOP: LAKE ELSINORE OVERLOOK

From the walled parking area and overlook, we look down on Lake Elsinore, the rectangular alluviated area around it, and the Perris Peneplain beyond to the NE. The lake is the main tourist attraction in the area, but in the early days Elsinore was known for its mineralized hot springs and resorts. Many hot springs issued along the NE side of the lake prior to 1890, but an irrigation canal disrupted the water table and now hot water is obtained only from wells. The City of Elsinore utilized water from thermal wells until a few years ago when 5 ppm fluoride was found in the water, an amount equal to more than five times the recommended limit. The lake is presently filled with water imported by aqueduct from the Colorado River and from runoff from the San Jacinto River.

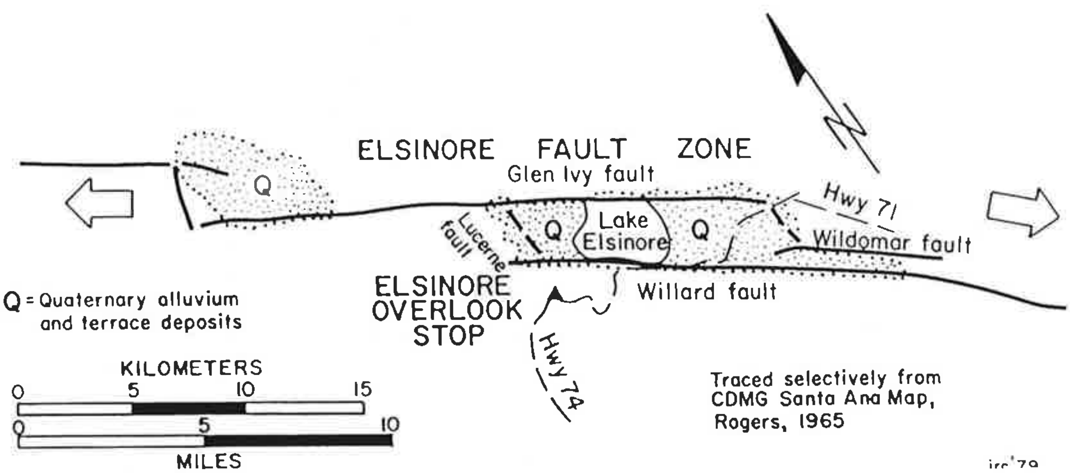
The lake basin is bounded by a rectangular arrangement of active faults (Fig. 30), which are easily distinguished from the overlook by their prominent scarps, and which together constitute a pull-apart depression under right slip between two main en echelon strands of the Elsinore fault system: the Glen Ivy and Willard faults. In this arrangement, the Lucerne fault at the NW end of the lake and the N-striking unnamed fault at the SE end of the lake are dip-slip faults. About 6 km (4 mi) of slip is suggested by the size and shape of the alluviated depression, if the pull-apart is closed up and allowance is made for some crustal stretching and sagging.

Note that there is a smaller en echelon side-stepping of the fault zone to the NW (Fig. 30), resulting in what is probably another pull-apart. Although vertical separations here are great, and probably amount to several kilometers, the right-slip component may amount to very much more (Crowell and Ramirez, Chap. 3, this volume). If the total right slip on the fault zone is indeed as much as the 40 km (25 mi) suggested by the offset Paleocene shoreline facies, the slip required for the Elsinore pull-apart must be a partial slip only, and therefore considerably younger. Confirmation of these interpretations has not yet been made by finding offset geological lines (or strike separation of near-vertical basement contacts).

The kinematic mechanism responsible for the Elsinore pull-apart is inferred to be quite similar to that taking place within the Salton Trough. We have included this stop to observe the geometry of such a feature on a relatively small scale, to appreciate the relations between separation and slip on a strike-slip zone, and to note as well the relations between tectonics and sedimentation. In a flexible crust, such as that envisioned for this region, terrain is arched up at places and stretched and depressed at others within a transform-fault regime.

Roadcuts between the overlook and the floor of Elsinore Valley expose Bedford Canyon schist (Early Mesozoic) intruded by spheroidally weathered gabbro and tonalite characterized by abundant rounded inclusions.

Physiographic details of the Elsinore fault zone are easily observed along the SW edge of the Elsinore Valley (Weber, 1975). Hwy 74 goes over the Willard fault scarp about 1/2 mi before the junction with Grand Ave. Fissures are reported to have formed along this fault during the San Jacinto Earthquake of 1918 (Engel, 1959). Note the notched and faceted spurs at the base of the Santa Ana Mountains. Uplifted fault slices form a chain of discontinuous hills in the broad alluvial area SE of Lake Elsinore, where fault scarps border some of these low hills on both the SW and NE sides. We cross this zone of faulting on Corydon Rd.



LAKE ELSINORE PULL-APART BASIN

From the beaches our journey will next take us through San Juan Capistrano and up to the crest of the Santa Ana Mountains in the northern part of the Peninsular Ranges. From the mountain heights we will look down upon Lake Elsinore which occupies a modern pull-apart basin.

The mission at San Juan Capistrano is famous as "The Jewel of the Missions." It was formally founded by Father Junipero Serra on 1 November 1776. It was named in honor of the fighting priest, Saint John Capistrano (1385-1456), who took a heroic part in the first defense of Vienna against the Turks. The mission was a pretentious establishment with many work shops, loom rooms, tallow vats, and so forth. It had the most important and pretentious stone church of the whole mission chain. In 1812 there were 1,361 Indian neophytes under the "care" of the padres when a significant earthquake affected the mission. The source of the shaking has just been determined last year to have been on a segment of the Elsinore fault we will soon be viewing. Around Richard Henry Dana Jr.'s time in the early 1800's, there were over 31,000 cows, horses, mules, sheep, goats, and pigs on the livestock rolls of the mission. The mission at San Juan Capistrano is famed in song and by yearly media attention for the return of flocks of swallows on March 19, St, Joseph's day. After faithfully departing on October 23 for their travel to Argentina to enjoy the Southern Hemisphere summer, the swallows just as faithfully return to the Mission each March 19.

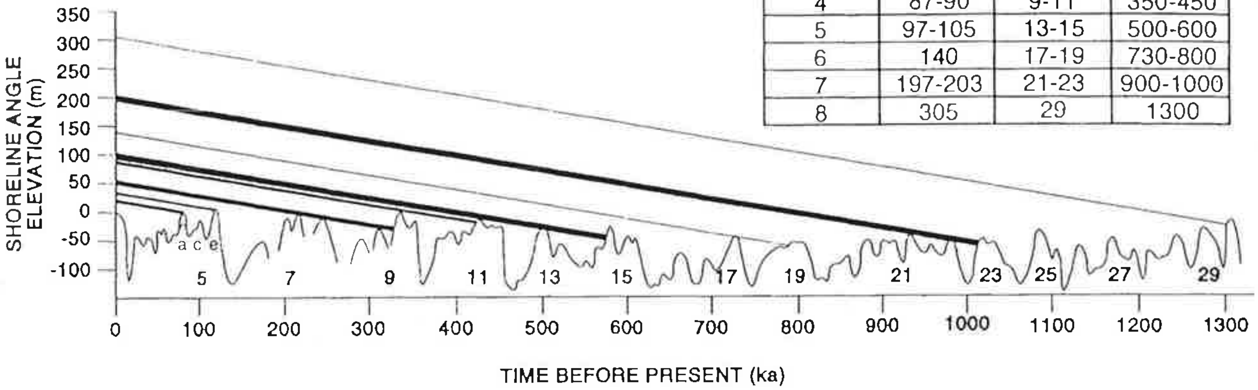
The Lake Elsinore area has held several names in its history. In order to satisfy the U.S. Postal Service in 1884, a unique new name was ordered to be selected. The name chosen was that of the Danish castle made famous by Shakespeare's Hamlet, viz. Elsinore.

This third geologic stop will have us look down upon Lake Elsinore and its fault-bounded rectangular basin which sits between two major, right-stepping, active faults of the Elsinore fault system (Figure 20). These active faults were responsible for the 1812 jolt that rocked the region and for numerous other shakes. The scarps along the Willard and Glen Ivy faults are recognizable from the overlook; they have created a pull-apart basin due to the right slip that steps between them. Figure 20 also shows the north-trending faults between the major strike-slip strands; northwest of the lake is the Lucerne fault and to the southeast is an unnamed fault, both of which are normal faults created to accommodate the pulling apart. Crowell and Sylvester (1979) suggested about 6 km of right slip were necessary to create the alluviated basin. However, this is probably only some of the recent slip as geologic work on offset Paleocene shorelines suggests as much as 40 km of right slip.

The tectonic basin viewed here is quite small but is nonetheless a good example of the wrench-fault created basins in the southern- and Baja-California region.

ıtion of shoreline- of coastal marine oxygen isotope :ic sea-level high- methodology of (1989). Ages of ter- ıted by their corre- -level highstands, ant 0.24 m/k.y. rate of line) based on errace. Sea-level from Chappell and 36) with modifica- s et al. (1994) and ·4). Oxygen isotope nbered. Maximum reported sea-level xygen isotope sub- ic are both shown. ·lation line corre- ꞓ in mapped shore- ions.

Terrace No.	Elevation (m)	Stage	Age (ka)
1	19-22	5a, 5c	83, 105
2	32-36	5e	122
3	49-53	7-9	212-340
4	87-90	9-11	350-450
5	97-105	13-15	500-600
6	140	17-19	730-800
7	197-203	21-23	900-1000
8	305	29	1300



rate a M_w 6.8 earthquake, using ssion of Wells and Coppersmith ively, the magnitude of a maxi- arthquake is estimated by assum- oaquin Hills thrust extends to the nogenic crust at 17 km, dips be- 0°, and extends upward to within ace. In this interpretation the San rust is a backthrust that soles into etachment (Bohannon and Geist, a wedge-thrust structure. The dip iin Hills fault is modeled after the nt of the San Joaquin Hills anti- Dana Point (e.g., Fisher and Mills, nodel, rupture of the entire fault rate a M_w 7.2–7.3 earthquake, us- on of Dolan et al. (1995). The late

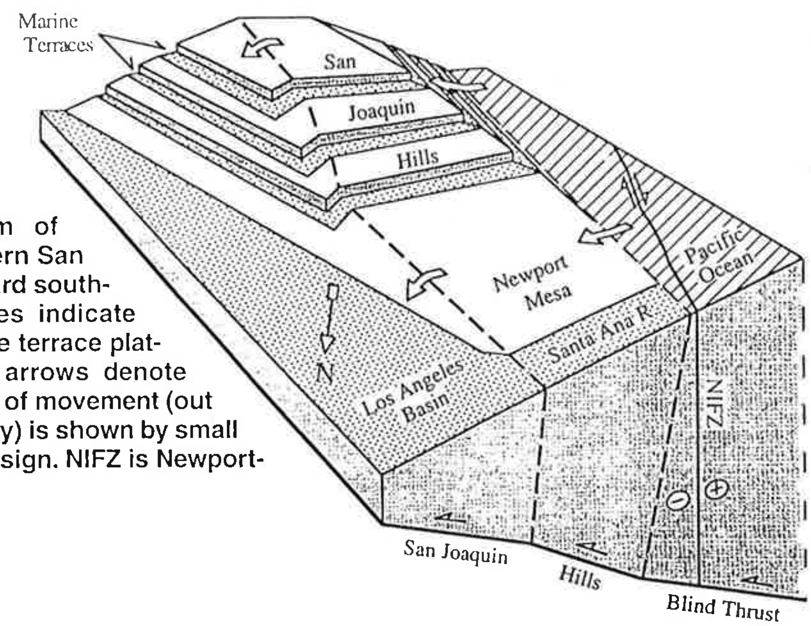


Figure 4. Block diagram of marine-terraces on northern San Joaquin Hills (viewed toward south-west). Heavy dashed lines indicate active axial surfaces where terrace platforms are folded. Large arrows denote sense of folding. Direction of movement (out of or into page, respectively) is shown by small arrows and plus or minus sign. NIFZ is Newport-Inglewood fault zone.

NINE LINES OF EVIDENCE FOR ROTATION OF THE WESTERN TRANSVERSE RANGES

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Evidence Line 1 – Physiography

Geographically, the east-west trend of the western Transverse Ranges has been an enigma for some time. The range was given its name by physiographers because of its generally oblique orientation to all of the other mountain ranges in California (Fig. 1). So the dilemma was recognized by physiographers even before geologists got involved in trying to decipher it. As an historical precedent, Francis Bacon in 1660 (Press and Siever, 2001) noted that the shape of the Atlantic Ocean basin was important long before its geological importance was deciphered.

Evidence Line 2 – Stratigraphic differences

An early indication that something was tectonically anomalous about the Los Angeles basin and southern California Inner Borderland (LAB/IB; defined by Crouch and Suppe in 1993 as the area between the Newport-Inglewood-Rose Canyon fault on the east, the Malibu Coast-Santa Monica-Raymond fault on the north, and the East Santa Cruz Basin fault on the west) was the difference in the stratigraphic sequence in the LAB/IB when compared with the surrounding regions of the western Transverse Ranges and the Peninsular Ranges (WTR/PR). The WTR/PR contain a relatively complete sedimentary sequence from the Jurassic through the Pleistocene. The sedimentary sequence throughout a majority of the LAB/IB contains no Mesozoic or Paleogene strata; where observed, lower and middle Miocene and younger strata rest directly on Catalina Schist basement rocks (Crouch and Suppe, 1993). In order to create this long hiatus in the geologic record in the LAB/IB requires either that the LAB/IB was a long-standing highland throughout the time interval from the Jurassic through the Paleogene (Reed, 1933; Corey, 1954), or that in the early Miocene the area was uplifted and all

Figure 1. Physiographic diagram of southern California that illustrates the transverse nature of the Western Transverse Ranges when compared to the other mountain ranges in California.

earlier deposited sedimentary rocks were eroded. This proposed highland was called "Catalinia" (Reed, 1933; Woodford *et al.*, 1954). If Catalinia had been a long-standing highland, all strata deposited in the surrounding WTR/PR would contain debris eroded from the Catalina Schist (Fig. 2), and current structures in those strata would indicate south-to-north transport in the WTR and west-to-east transport in the PR. If Catalinia had been uplifted and

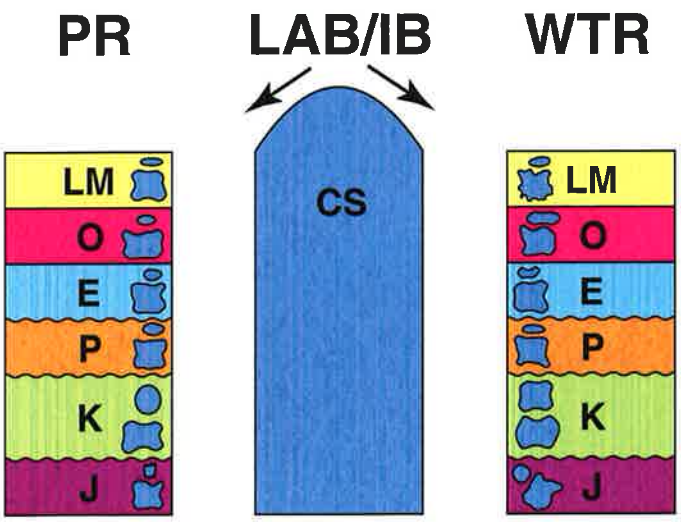

Figure 2. Pre-upper-Miocene stratigraphic columns in the Peninslar Ranges (PR), Los Angeles basin/Inner Borderland (LAB/IB), and Western Transverse Ranges (WTR) areas showing distribution of eroded materials if the LAB/IB (Catalinia) had been a long-standing landmass.

eroded during the early Miocene from a previously depositional setting, then lower Miocene strata in the WTR/PR would contain debris from the Catalina Schist as well as debris from any earlier strata that had been deposited (Fig. 3), and early Miocene current directions would indicate south-to-north transport in the WTR and west-to-east transport in the PR. Of all the above-stated criteria that would demonstrate the long-standing existence or erosion of Catalinia, only the south-to-north transport directions in the WTR have been shown to exist. It is clear that neither of the above hypotheses are supported by geological data and that another hypothesis is needed to explain the differences in stratigraphy in the three areas.

Evidence Line 3 – San Onofre Breccia

In 1925, Woodford described a unique lower to middle Miocene sedimentary unit that he named the San Onofre Breccia that is exposed in several places in southern California. This unit is a marine to nonmarine breccia composed mostly of angular, boulder to sand-size

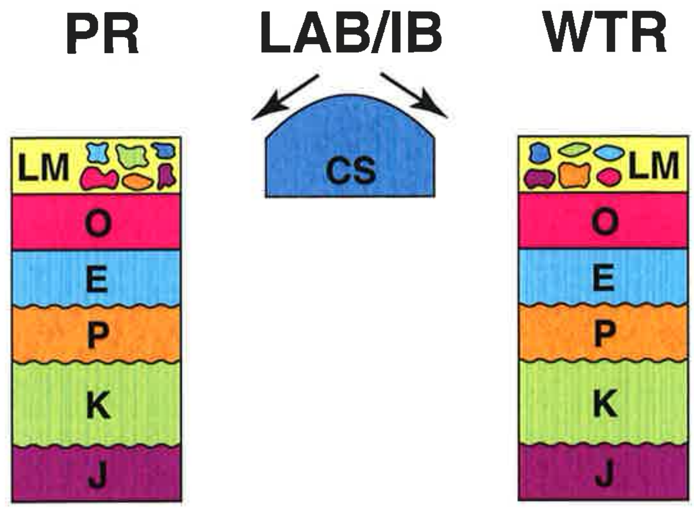

Figure 3. Pre-upper-Miocene stratigraphic columns in the Peninslar Ranges (PR), Los Angeles basin/Inner Borderland (LAB/IB), and Western Transverse Ranges (WTR) areas showing distribution of eroded materials if the LAB/IB are had been uplifted and eroded during the lower Miocene.

fragments of Catalina Schist (Stuart, 1979). The paleogeographic and tectonic importance of this breccia has been recognized ever since Woodford's (1925) paper was published, and the existence of the breccia was one of the best lines of evidence in support of the proposed highland of Catalinia discussed in Evidence Line 2 above. Certainly the provenance of the San Onofre Breccia had to be exposures of almost 100% Catalina Schist. For the PR, therefore, the type area for the San Onofre Breccia, the provenance had to be toward the west because erosion of any rocks in the PR would have produced a deposit that was not nearly 100% schist debris. Current directions in the deposit would be west-to-east directed. Likewise, for exposures in the WTR, the provenance had to be from what is today toward the south and current directions would have been south-to-north directed. If Catalinia had been a long-standing highland, there would be Catalina Schist debris in Paleogene strata in the PR and the WTR, as noted in Evidence Line 2 above (Fig. 2), but there is not. This means that Catalinia had to be uplifted rapidly around the end of the early Miocene. But if Catalinia were not an eroding highland before the early Miocene, it would have been an area of pre-Miocene deposition. So when it became exposed in the early and medial Miocene, these pre-Miocene deposits would have to have been eroded before erosion reached the schist, and this pre-Miocene debris would be found in the breccia (Fig. 3), but it is not. The presence of nearly 100% Catalina Schist debris in the San Onofre Breccia (Fig. 4) is important, therefore, because the only way to create a Catalina Schist source area that is

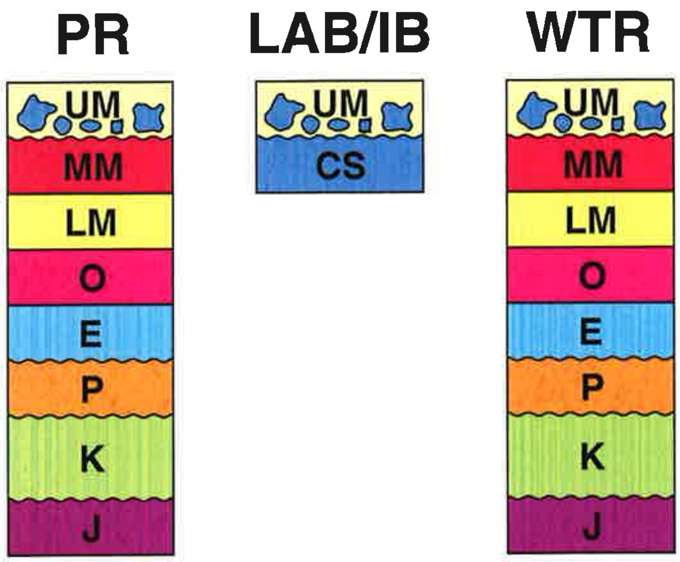

Figure 4. Pre-Miocene stratigraphic columns in the Peninslar Ranges (PR), Los Angeles basin/Inner Borderland (LAB/IB), and Western Transverse Ranges (WTR) areas showing actual distribution of eroded materials that produced the San Onofre Breccia.

devoid of younger rock types is by extensional tectonic denudation of the overlying rocks along low-angle detachment faults. Extension of the crust causes fracturing in the brittle surface rocks, which are then pulled apart in the hanging wall of a detachment fault so that the footwall schist rocks become exposed. The schist is then uplifted isostatically until it reaches the surface of the ocean where it can be eroded and serve as source material for the San Onofre Breccia (Figs. 5-8). Because the San Onofre Breccia is not found everywhere in the LAB/IB (most areas have Monterey Shale on top of the Catalina Schist), it is likely that Catalinia was not a large highland, but rather a series

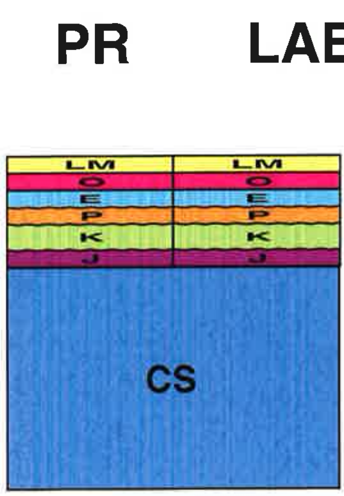

Figure 5. Distribution of stratigraphic units before extension began in the early Miocene.

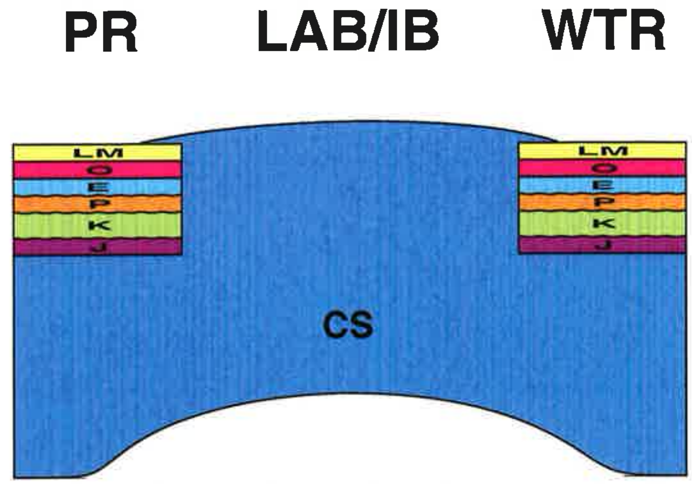

Figure 6. Distribution of stratigraphic units after extension and tectonic denudation during the early Miocene.

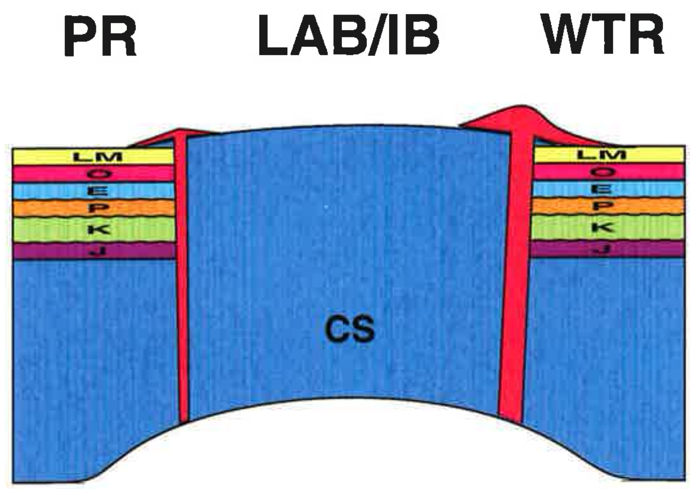

Figure 7. Distribution of stratigraphic units after intrusion and extrusion of medial Miocene volcanics.

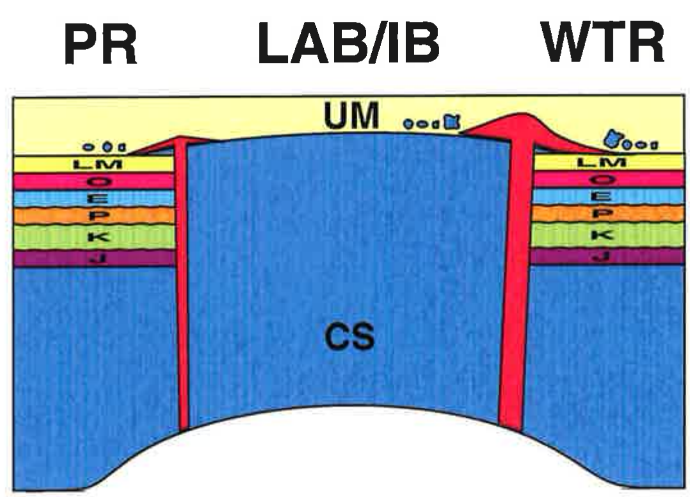

Figure 8. Distribution of stratigraphic units after deposition of the San Onofre Breccia during the medial Miocene.

of scattered islands, providing detritus at local sites of deposition. Note that it took six decades of expanding geologic understanding before we, as professional geologists, were able to understand this key piece of information that has perplexed us for so long. The models for the tectonic history of southwestern California by Yeats *et al.* (1974), Crouch (1979), Kamerling and Luyendyk (1985), and Crouch and Suppe (1993) involve crustal extension and would produce the San Onofre Breccia. The model by Howell *et al.* (1974), involving strike-slip faulting, is not an extensional model and would not produce the San Onofre Breccia.

Evidence Line 4 – Jurassic metasedimentary structural trends

The oldest sedimentary unit in the WTR is the Santa Monica Formation (Hoots, 1931), which is Late Jurassic in age (Fig. 9). The oldest sedimentary unit in the PR is the Bedford Canyon Formation (Allison, 1970), which is Medial Jurassic in age (Fig. 10). Fossils are rare in the Santa Monica Formation, so the age could be extended; fossils in the Bedford Canyon Formation are more common and from several stratigraphic levels (Imlay, 1963, 1964). Although the ages are different, the units are broadly similar in lithology (Durrell, 1954; Dibblee, 1982), both are mildly metamorphosed (the metamorphism coming before deposition of the overlying Cretaceous units), and both have an affinity with the Mariposa Formation of the Sierran western foothills belt of Jurassic rocks in northern California (Imlay, 1963; Jones *et al.*, 1976). Structural trend (bedding, fold axes, etc.) in the Bedford Canyon Formation of the PR and the Mariposa Formation of the Sierran western foothills belt is generally north-south, whereas the structural trend in the Santa Monica Formation of the WTR is east-west. This difference in structural trend prompted Jones and Irwin (1975) to propose post-Late Jurassic rotation for the Santa Monica Mountains.

Figure 10. Exposure of the Middle Jurassic Bedford Canyon Formation in the Santa Ana Mountains in Silverado Canyon.

Figure 9. Exposure of the Upper Jurassic Santa Monica Formation in the Santa Monica Mountains near the intersection of Mulhollnad Drive and I-405.

Evidence Line 5 – Poway clasts

A suite of unique, exotic, and ultrahard rhyolitic clasts has been known from Paleogene strata in the southern California coastal area for some time (Woodford *et al.*, 1968). These clasts are referred to as "Poway" clasts because of their abundance in the middle Eocene Poway Formation near the town of Poway, California, northeast of San Diego (Fig. 11). Their source has been traced to rocks in northwestern Sonora, Mexico (Merriam, 1968; Abbott and Smith, 1989). Discovery of these clasts in conglomerate units on the northern Channel Islands (Merschat, 1971; Parsley, 1972; Yeats *et al.*, 1974; confirmed chemically by Abbott and Smith, 1978) presented a problem because the transport distance from the San Diego area to the present location of the islands is much too great. Yeats *et al.* (1974) shortened the transport distance by suggesting large, eastward, lateral backtranslation for the Santa Monica Mountains and the northern Channel Islands along the Malibu Coast-Santa Monica-Raymond-Cucamonga fault, bringing the Santa Monica Mountains up against the northern end of the Santa Ana Mountains and thus closer to the San Diego area. Howell *et al.* (1974) suggested that the prob-

Figure 11. A Poway clast from the San Diego area.

lem could be solved with left-lateral backslip along offshore right-lateral faults that would bring the northern Channel Islands southward toward the San Diego area. Kies and Abbott (1982), in their summary paper on Paleogene conglomerates, apparently used the Howell *et al.* (1974) model in their paleogeographic reconstructions. Crouch (1979), Kamerling and Luyendyk (1985), Crouch and Suppe (1993), and Fritsche (1998) shortened the distance between the northern Channel Islands and the San Diego area by backrotating the WTR in a counterclockwise direction until the two areas became adjacent.

Evidence Line 6 – Paleogene paleocurrent directions

Paleogene current directions, summarized by Yeats *et al.* (1974), are dominantly east-to-west in the PR and south-to-north in the WTR (Fig. 12). Yeats *et al.* (1974) used this paleocurrent data, along with Poway-clast distributions, to postulate large, eastward, lateral backtranslation for the Santa Monica Mountains along the Malibu Coast-Santa Monica-Raymond-Cucamonga fault. These paleocurrent directions do not as easily support the strike-slip models of Howell *et al.* (1974) and Kies and Abbott (1982). Paleogene paleocurrent data can also be used, however, to support the rotation hypothesis of Crouch (1979), Kamerling and Luyendyk (1985), and Crouch and Suppe (1993) because after ~90° of counterclockwise backrotation, the paleocurrents of the PR and WTR become parallel (Fig. 13).

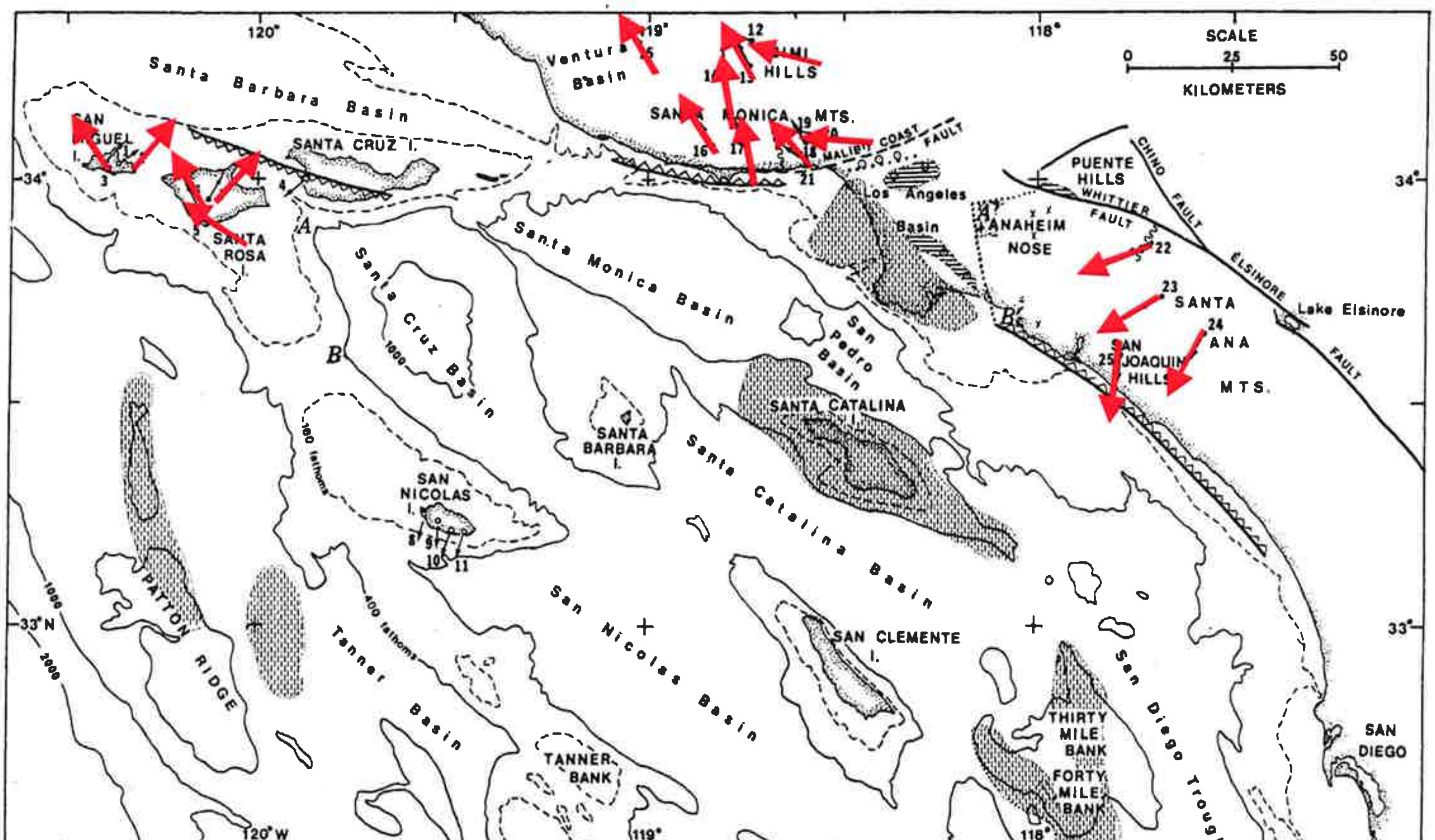

Figure 12. Paleogene current directions as determine by Yeats *et al.* (1974).

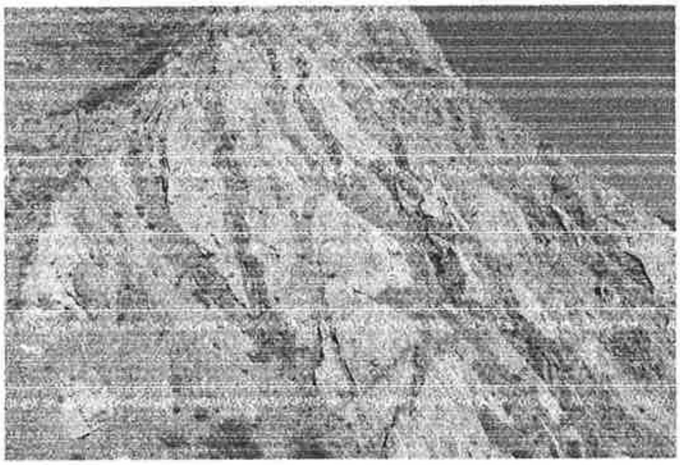

Figure 16. Diabase dikes intruding lower Miocene sandstone on the north side of Highway 1 just east of Point Mugu.

Suppe (1993) over the model of Howell *et al.* (1974), which does not require extension.

Evidence Line 9 – Paleomagnetic data

The single most important discovery in the search for a paleogeographic model to explain Evidence Lines 1 through 8 was evidence produced from paleomagnetic studies of rocks, mostly volcanic, in the southern California region. In 1979, Kamerling and Luyendyk published the first of several studies that made use of the paleomagnetic properties of rocks to indicate the direction to the north magnetic pole at the time the rock was formed. As each new study appeared over the next decade, the evidence mounted that indeed the entire WTR had rotated clockwise about 90° since the time the rocks had either cooled or been deposited (Fig. 17). This discovery proved to be the key that unlocked the solution to the Tertiary tectonic history of southern California. Backrotation of the WTR about 90° into a north-south orientation produces the following results: (1) the physiographic trend of the WTR (Evidence Line 1) and the structural trend of the Santa Monica Formation (Evidence Line 4) are brought to north-south, parallel to the PR and the Sierrra Nevada; (2) the future site of the Catalina Schist basement rocks and the Catalinia islands is covered by WTR strata, bringing the separated Mesozoic and Paleogene sequences of the Santa Monica Mountains and the PR into contact with each other (Evidence Lines 2 and 7); (3) the paleocurrents of the WTR and the PR (Evidence Line 6) become parallel and allow Poway clasts (Evidence Line 5) to be transported to the future northern Channel Islands area in just a short distance; and (4) extension and clockwise rotation of the WTR away from the PR during the early and medial Miocene by a detachment process creates both decompression melting and intrusion and extrusion of volcanic rocks (Evidence Line 8) and the tectonic denudation necessary to produce the San Onofre Breccia (Evidence Line 3).

Soon after paleomagnetic evidence for WTR rotation was discovered, rotation models for the Tertiary tectonic history of southwestern California were introduced because the previous models of Yeats *et al.* (1974) and Howell *et*

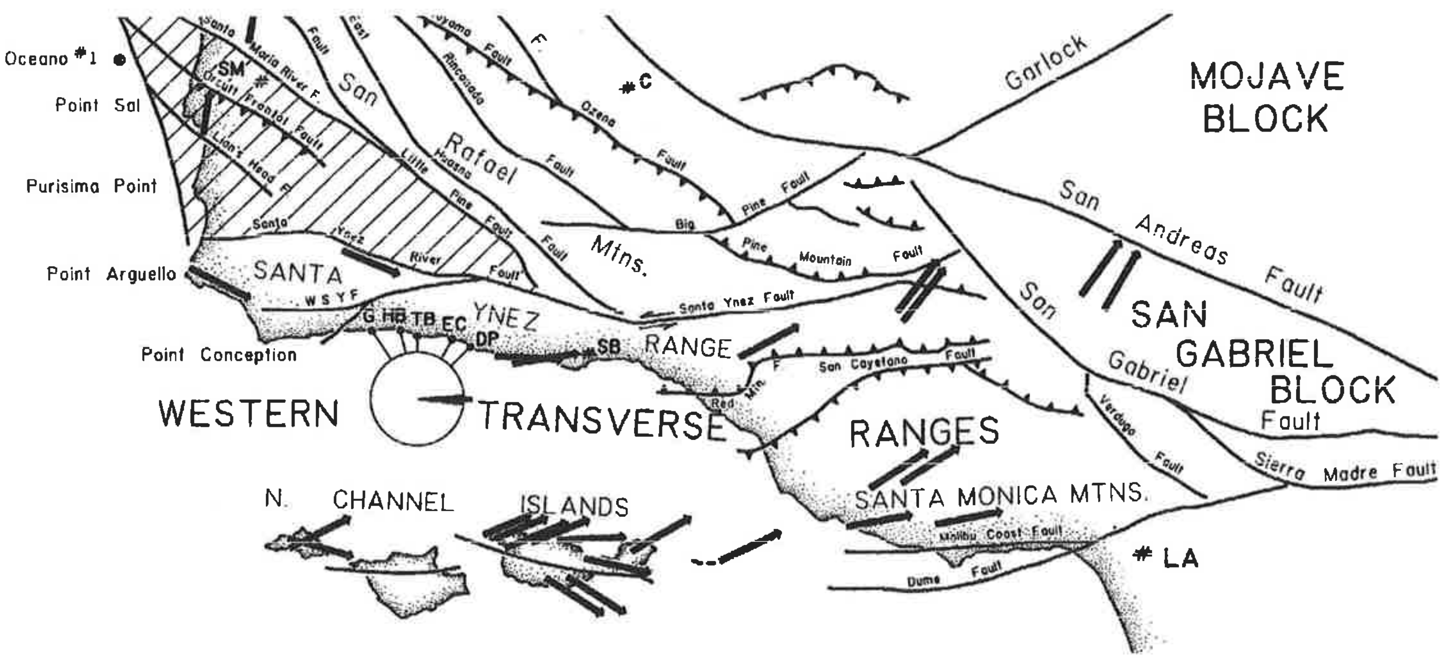

Figure 17. Compilation of paleomagnetic north directions from Luyendyk and others for rocks of the Western Transverse Ranges.

al. (1974), which did not include rotation, were no longer valid. The earliest ones (e.g., Luyendyk *et al.*, 1980; Luyendyk and Hornafius, 1987) used right-shearing shutter panels in the LAB/IB to account for rotation of the WTR. These models did not use a full 90° of rotation and did not fully close up the LAB/IB, thus failing to address the problems associated with Evidence Lines 1 through 4 above. Crouch (1979), Kamerling and Luyendyk (1985), and Crouch and Suppe (1993) combined the concepts of rotation and extension to produce palinspastic models that eliminated the right-shearing shutter panels and restored the western Transverse Ranges to a position parallel with and immediately adjacent to the Peninsular Ranges before rotation and extension. These models satisfy all the constraints of Evidence Lines 1 through 8, but do not suggest a mechanism by which the rotation of the WTR might be accomplished. The rotation-extension model has since received support from Nicholson *et al.* (1994), who suggested that capture of the partially subducted Monterey microplate by the Pacific plate and rotation of the WTR block on top of the Monterey microplate as it was pulled by the Pacific plate provides a mechanism for the operation of the rotation-extension model. A more recent model by Fritsche (1998) is similar to Crouch and Suppe's (1993), but better addresses the details of the paleomagnetic rotation data.

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STOP AT SAN ONOFRE STATE BEACH TO STUDY THE CRISTIANITOS FAULT

Pull into the parking lot at San Onofre Beach State Park and park as close as possible to the sign that says Beach Trail 1. Walk down Beach Trail 1 across a landslide to the beach, then walk northwest along the beach to the Cristianitos fault (Fig. 1). The purpose of this stop is to examine the Cristianitos fault and discuss its origin.

The Cristianitos fault exposed here in the sea cliff strikes ~N32°W and dips ~58°NW (Shlemon, 1987). The upper Miocene to lower Pliocene San Mateo Formation is in the hanging wall on the northwest and the upper Miocene Monterey Formation is in the footwall on the southeast. The stratigraphic relations along the fault, the conjugate shear pattern in the San Mateo Formation (Fig. 2), and the antithetic shears along the fault surface all demonstrate that this is a normal-slip fault. A marine terrace dated at 125 Ka (Shlemon, 1987) overlies and is unbroken by the fault (Fig. 47).

Crouch and Suppe (1993) suggested that the Cristianitos fault is one of the major listric normal faults that join the major detachment fault at depth. They also proposed that the extension and rotation process began in the early Miocene. If extension began in the early Miocene, then sometime soon after that is when listric normal faulting should have begun. If the Cristianitos fault began to move in the early or medial Miocene and was still moving in the Pliocene, there should be clear evidence of the growth fault process, with formations on the hanging wall side of the fault being thicker than those on the footwall side, but no one has suggested that this happened. In addition, the Cristianitos fault dies out about 33 km to the north (Vernon et al., 1970; Shlemon, 1987), which seems to be a rather short distance for a proposed major fault. Crouch and Suppe (1993), however, extend it on to the north on the east side of the Santa Ana Mountains to the Whittier-Elsinore fault and suggest that its continuation on the northeast side of the Whittier-Elsinore fault is the Chino fault. One final point is that most if not all of the extension and rotation ended near the end of the late Miocene (Kamerling and Luyendyk, 1985), yet the Cristianitos fault continued to move into the Pliocene. It seems, therefore, that evidence is insufficient to label the Cristianitos fault as one of the major listric fault players in the extension-rotation model.

References are listed in the Nine Lines of Evidence for Rotation article.

Figure 1. Cristianitos fault at San Onofre State Beach. The San Mateo Formation is in the hanging wall, the Monterey Formation is in the footwall, and both formations are overlain by a 125-Ka marine terrace. Note the antithetic shears in the San Mateo Formation along the fault.

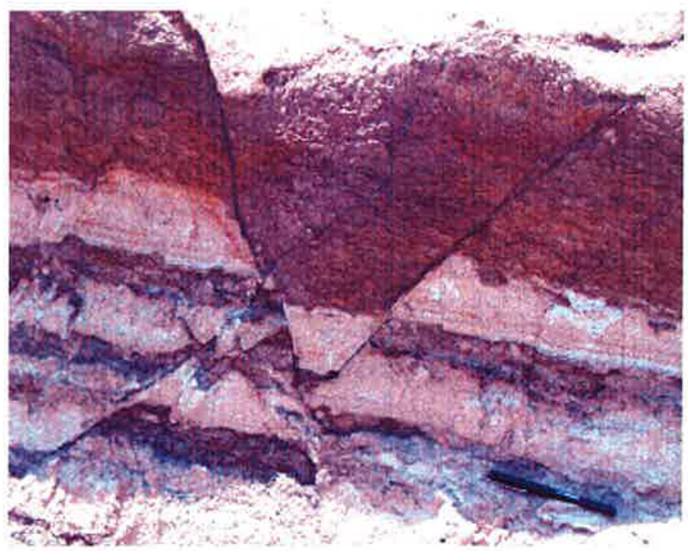

Figure 2. Cross-cutting synthetic and antithetic shears in the San Mateo Formation on the cliff face west of the Cristianitos fault.

STOP AT DANA POINT TO STUDY THE SAN ONOFRE BRECCIA

Park in the Dana Point parking lot as near to the west end of the lot as possible. Walk from the parking lot out to Dana Point. The purpose of this stop is to see the middle Miocene San Onofre Breccia as a product of a different depositional environment and to review the tectonic importance of the breccia in requiring crustal extension as noted in Evidence Line 3.

Clasts in the San Onofre Breccia are nearly 100% Catalina Schist fragments. The tectonic denudation process required to create the very high percentage of Catalina Schist clasts is discussed in Evidence Line 3. The rock contains a red, muddy matrix between the clasts. The beds are thick to very thick, have relatively flat bottoms and tops, and many are clearly reverse-to-normal graded (fine to coarse to fine) (Fig. 1). The color, grading, and thickness all point to deposition as debris flows on an alluvial fan. The Catalina Schist source rocks were southwest of this site (Stuart, 1979), and it is clear that they were not being eroded just by wave action, but were standing sufficiently above sea level to have developed a thick alluvial fan around them. This implies uplift rates of the tectonically denuded submarine schist that were greater than the rate required by ocean waves to keep the exposures eroded to sea level.

At least two possible models exist for the tectonic denudation process (Fig. 2). In the model shown in Figure 2A, the detachment surface is confined to the area between the two outermost listric normal faults. As the WTR block is pulled away from the PR block, extension occurs by stretching the crust below the detachment surface. The amount of crustal extension will be least at the northern end of the basin near the pivot point of the WTR block and will be greatest at the south end of the basin. Crouch and Suppe (1993) estimated the amount of extension in the central part of the basin to be 220 km. As the basin opens, sediment begins to accumulate on the newly exposed Catalina Schist ocean bottom. Because the extension occurs more or less uniformly, there would be no predictable pattern to where the oldest sediment would occur. This model would also suggest greater volcanic activity in the south part of the basin where the extension is greatest.

A second possibility, shown in Figure 2B, was proposed by Platt (1976) based on a suggestion by Ernst (1971) for

Figure 1. San Onofre Breccia outcrop at Dana Point. Note the relatively flat bases of the beds and the inverse-to-normal grading that mark these beds as debris-flow deposits. Hammer in lower center part of photo for scale.

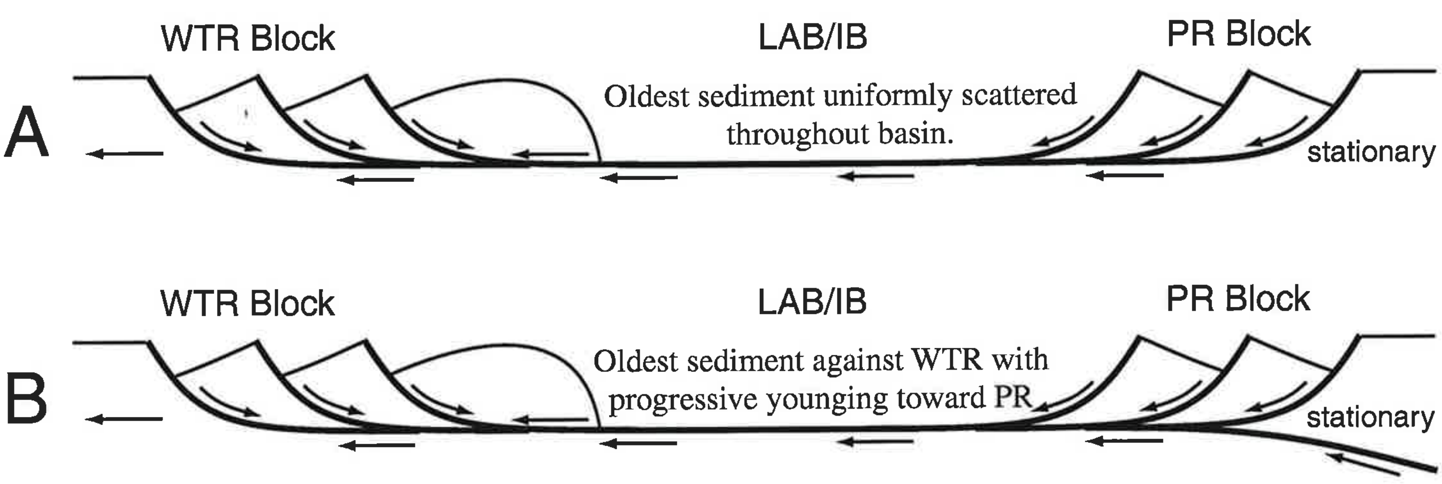

Figure 2. Diagram showing two possible models for how the tectonic denudation process might occur. See text for explanation.

the Franciscan Complex as a whole. In this model, the detachment surface extends beyond the boundaries of the developing basin into the previous subduction zone, and during the extension process the Catalina Schist is pulled back out of the subduction zone. The Catalina Schist crust under the LAB/IB does not stretch at all; it is pulled out of the subduction zone in the direction of Pacific Plate motion, and the WTR block, after separating from the PR block, rotates as it rides along on the back of the extracting schist. Rotation occurs because the northern end of the WTR block, for some presently unknown reason, is pinned in place. This model dictates that newly exposed sea floor will appear first against the WTR block and that this is where the oldest sedimentary rocks should be found on top of the Catalina Schist. In support of this second model, Stuart (1979), in his correlation chart for San Onofre deposits, shows all the deposits in the northern Channel Islands and at Point Dume to be latest early Miocene (Saucesian) in age at the base, the deposits on Catalina Island to be slightly younger, and those deposits along the southern California coast to be medial Miocene (Relizian) in age at the base.

On completion of your study of the San Onofre Breccia, be sure to walk back westward along the sea cliff and past the parking lot to the public toilet building in the park. This toilet building, built entirely of blueschist clasts, should make a lasting impression on the mind of any geologist.

References are listed in the Nine Lines of Evidence for Rotation article.

STOP AT CRESCENT BAY TO STUDY INTRUSIVE DIKES

Turn left (southwest) onto Crescent Bay Drive. This is a circular road that becomes a one-way street after the first left turn. Watch the house numbers as you follow the street and park on either side when you reach the 300 numbers. Walk back to the dead-end street between the 300 block and the 200 block. Follow the dead-end street to the stairs at the end. Descend the stairs and turn right to a second set of stairs that goes down to the beach. Turn right (northwest) at the beach and walk to the outcrops at the point. The purpose of this stop is to look at a Miocene intrusion and its intrusive contact. The existence of the dikes here in the San Joaquin Hills, like the existence of those in the Santa Monica Mountains, favors extensional tectonics, which is the basis for Evidence Line 8.

The first exposures along the beach before the point are of the middle to upper Miocene Monterey Formation, which consists of thin layers of medium gray, laminated shale that is interlaminated with thin, wavy lenses of very fine-grained sandstone and siltstone (Fig. 1). These beds look like Bouma Tcde turbidites that were deposited on the distal fringes of a submarine fan, but they might also have been deposited by contour currents.

From the Monterey Formation exposures move on toward the cliff at the beginning of the point (Fig. 2) and find the contact between the Monterey Formation and an intrusive andesite (Fig. 3). This contact is sharp and cuts at a slight angle across the bedding in the Monterey Formation. There is no obvious evidence of a cooled margin or baking of the Monterey Formation, so this was probably a shallow, cool intrusion. Note that the Monterey beds have been folded into a drag syncline (Fig. 4) along the margin of the intrusion. The andesite is rather homogeneous and exhibits spheroidal weathering (Fig. 5). It is weathered sufficiently to make it look as much like sandstone as andesite.

About a dozen dikes and sill-like bodies of andesite crop out along about 3 km of the coast between lower Moro Canyon and Recreation Point northwest of Laguna Beach

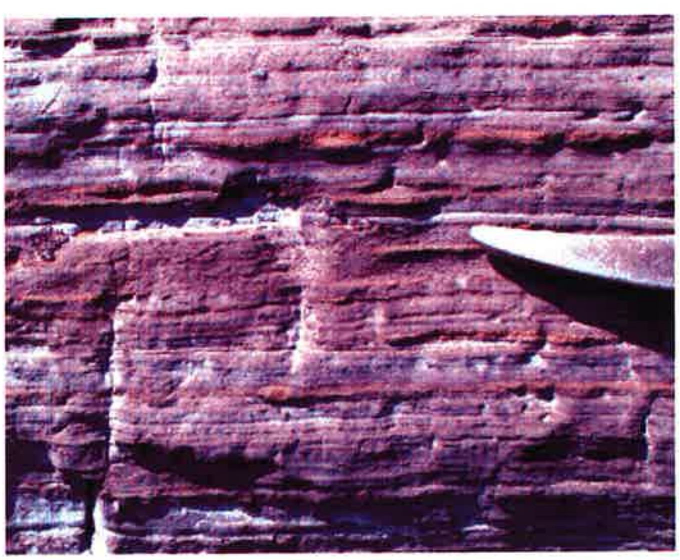

Figure 1. Thin, graded beds with lenticular siltstone bases in the Monterey Formation.

Figure 2. Andesite dike (left side of picture) in intrusive contact with thin shale beds of the Monterey Formation. Shale is folded into possible drag syncline along the contact. Backpack and hammer for scale.

Figure 3. Close-up of sharp intrusive contact between andesite (under the hammer) and shale of the Monterey Formation. Contact is at a slight angle to the bedding.

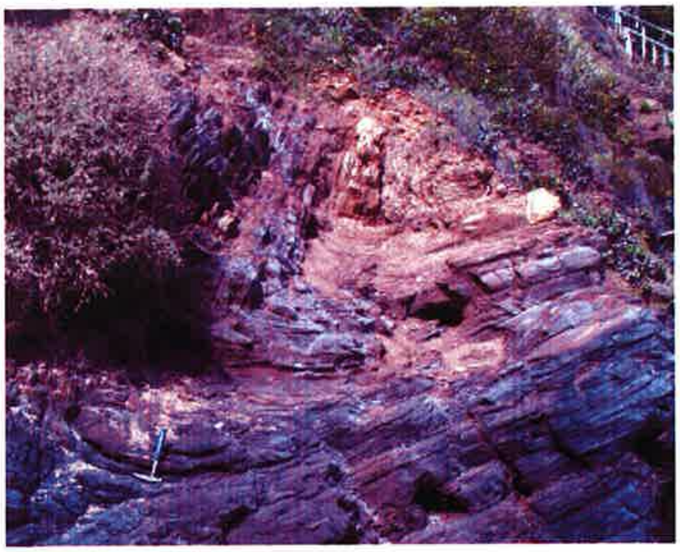

Figure 4. Close-up of the possible drag syncline shown in Figure 2.

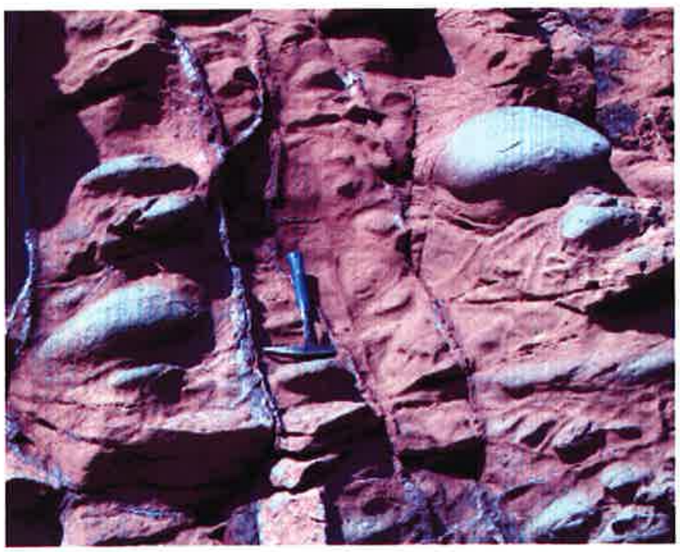

Figure 5. Spheroidal weathering and fracturing in the andesite.

and in the hills immediately north of the city (Vedder *et al.*, 1957). They exhibit trachytic, porphyritic, and pilotaxitic textures and contain plagioclase (An70-50), augite, and occasional hypersthene (Sexton, 1982; Vedder, 1981). The bodies are locally vesicular and locally exhibit crude to well-defined flow alignment (Vedder, 1981). Although no radiometric dates have been determined for these intrusions, Luisian foraminifera in Monterey Shale associated with an intrusive body near Reef Point and upper Mohnian foraminifera in Monterey Shale beneath the sill at Emerald Bay (J. G. Vedder, unpublished notes) suggest that the intrusive episode took place in the late Miocene. Thus, these intrusions are younger than the 15.8 Ma andesite flow in the San Joaquin Hills and younger than the Conejo Volcanics. The few samples of Laguna Beach igneous rocks that have been chemically analyzed (Weigand, 1994) show them to be low- to medium-K calc-alkaline basaltic andesite and andesite (Fig. 6).

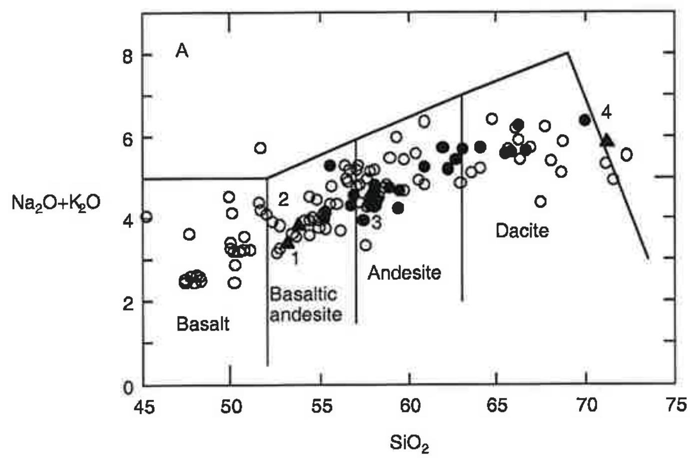

Figure 6. Chemical classification diagram for analyzed samples of volcanic rocks at Laguna Beach.

old-and-thrust belt located offshore between Dana Point and Oceanside (Fisher and Mills, 1991) (Fig. 2A). In contrast to the extensional features, these contractional structures deform Pliocene and younger strata, and are commonly associated with pronounced seafloor fold scarps. These structures do not cut or fold the detachment. Thus, we interpret that they sole into the Oceanside thrust.

The Oceanside thrust is mapped over an area of more than 1800 km^2. In migrated seismic reflection profiles, the thrust is imaged as a coherent set of strong reflections that dip to the northeast between 14° and 25° (Fig. 2). The thrust extends south along the Coronado Banks to the international border near San Diego Bay (Fig. 1). At Coronado Banks, thrusting is reflected by tectonic inversion of the side thrust between Dana Point and Oceanside is the most direct evidence for this tectonic inversion. Folds in this belt generate pronounced seafloor scarps that persist for ~30 km. Although these scarps may reflect recent activity of the underlying Oceanside thrust (Fig. 2), they are not definitive; we lack precise age control on seafloor sediments. However, young contractional folds also occur along the coast and involve dated marine terraces. These structures record recent fault activity that may be attributed to the Oceanside thrust.

The San Joaquin Hills are at the southern margin of the Los Angeles basin, where the mapped part of the Oceanside thrust extends onshore. The hills are formed by a northeast-vergent anticline that uplifts and deforms marine terraces. Grant

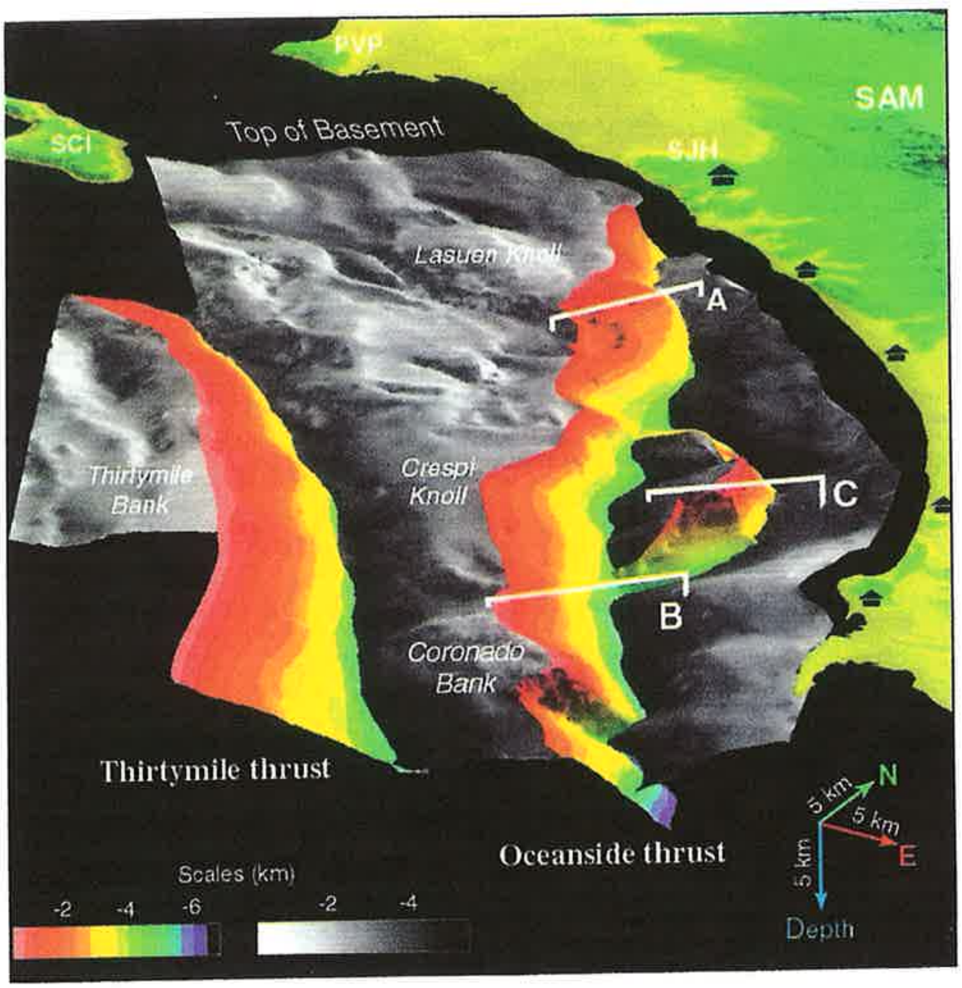

Figure 1. Perspective view of three-dimensional model of Oceanside and Thirtymile Bank blind thrusts. Gray surface is top of basement (Catalina Schist). Small triangles indicate areas of recent uplift (Lajoie et al., 1979, 1992; Barrie and Gath, 1992; Kern and Rockwell, 1992; Grant et al., 1999; Kier and Mueller, 1999). Digital shaded relief map of southern California topography was derived from digital elevation data provided by U.S. Geological Survey. SAM—Santa Ana Mountains; SJH—San Joaquin Hills; PVP—Palos Verdes Peninsula; SCI—Santa Catalina Island.

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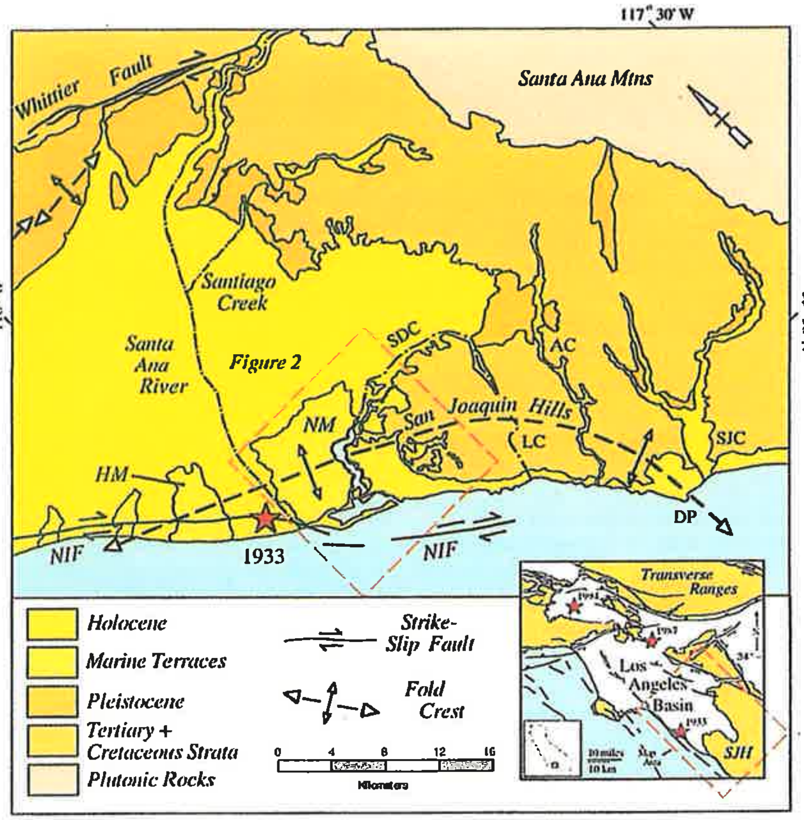

Figure 1. Regional map of San Joaquin Hills (SJH) showing approximate location of fold crest of San Joaquin Hills anticline. Other features: AC—Aliso Canyon, DP—Dana Point, HM—Huntington Mesa, LC—Laguna Canyon, NIF—Newport-Inglewood fault, NM—Newport Mesa, SDC—San Diego Creek, and SJC—San Juan Capistrano. Epicenters of 1933, 1987, and 1994 earthquakes are marked by stars. Modified from Vedder et al. (1957). Inset map shows greater Los Angeles region.

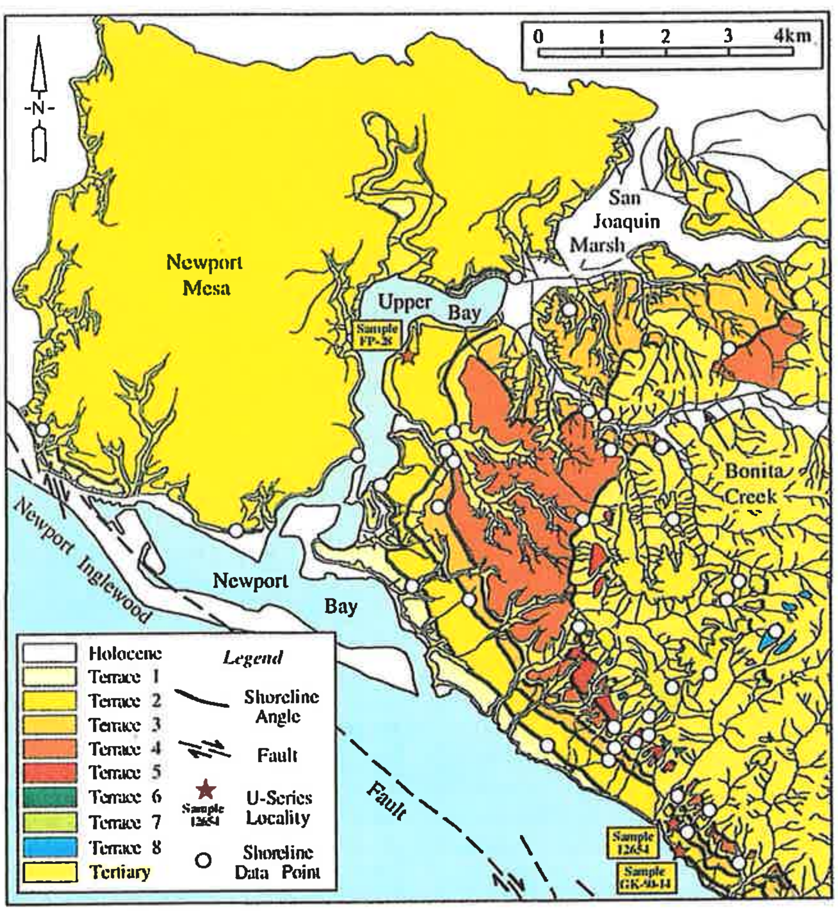

Figure 2. Map of marine-terrace platforms, location of measured shoreline elevations (open circles), and fossil localities (stars). See Figure 1 for location.

mediate terraces were probably occupied by multiple highstands (Fig. 3).

RECENCY OF UPLIFT

The location and thickness of Holocene sediments in the San Joaquin Hills suggest that tectonic uplift continued during the middle to late Holocene. Isopach and structure contour maps of the Holocene Talbert aquifer beneath the Santa Ana River (Sprotte et al., 1980) suggest that the aquifer has been deformed. The distribution of fluvial and estuarine sediments southeast of the Santa Ana River also suggests that Holocene uplift occurred. Newport Bay was incised at least 36 m below present sea level during the last glacial maximum, but rapid sea-level rise at the close of the Pleistocene inundated coastal drainages and induced sedimentary infilling with Holocene sediments (Stevenson, 1954). Stevenson concluded that an elevated bench of former marsh deposits in Newport Bay was created during the late Holocene by emergence. Stevenson speculated that the emergence was due to tectonic uplift and, based on elevation profiles, the uplift reflected anticlinal folding along a northwest-trending fold axis.

DISCUSSION

The late Quaternary uplift rate, anticlinal structure, and indications of Holocene uplift imply that the San Joaquin Hills are the surface expression of an active contractile fold (see Fig. 1), formed above a potentially seismogenic thrust fault (Stein and Yeats, 1989; Lettis et al., 1997; Shaw and Shearer, 1999). Geomorphic analysis of the San Joaquin Hills provides some constraints on the geometry of the proposed blind fault. A fault-bend fold model with movement on a northeast-vergent thrust fault best explains the elevation of marine terraces on the northeast limb of the San Joaquin Hills anticline, as shown schematically in Figure 4. Anticlinal structure of the San Joaquin Hills and northeast vergence of the underlying fault are supported by structural data of Vedder (1975) and our terrace mapping.

The maximum-magnitude earthquake that could occur on the San Joaquin Hills thrust can be estimated from empirical relationships between magnitude and subsurface fault-rupture length, as defined by the areal distribution of Quaternary uplift. This method does not require assumptions about fault geometry. The area of uplift extends at least ~38 km, from northwestern Huntington Mesa southeast to Dana Point, and therefore the

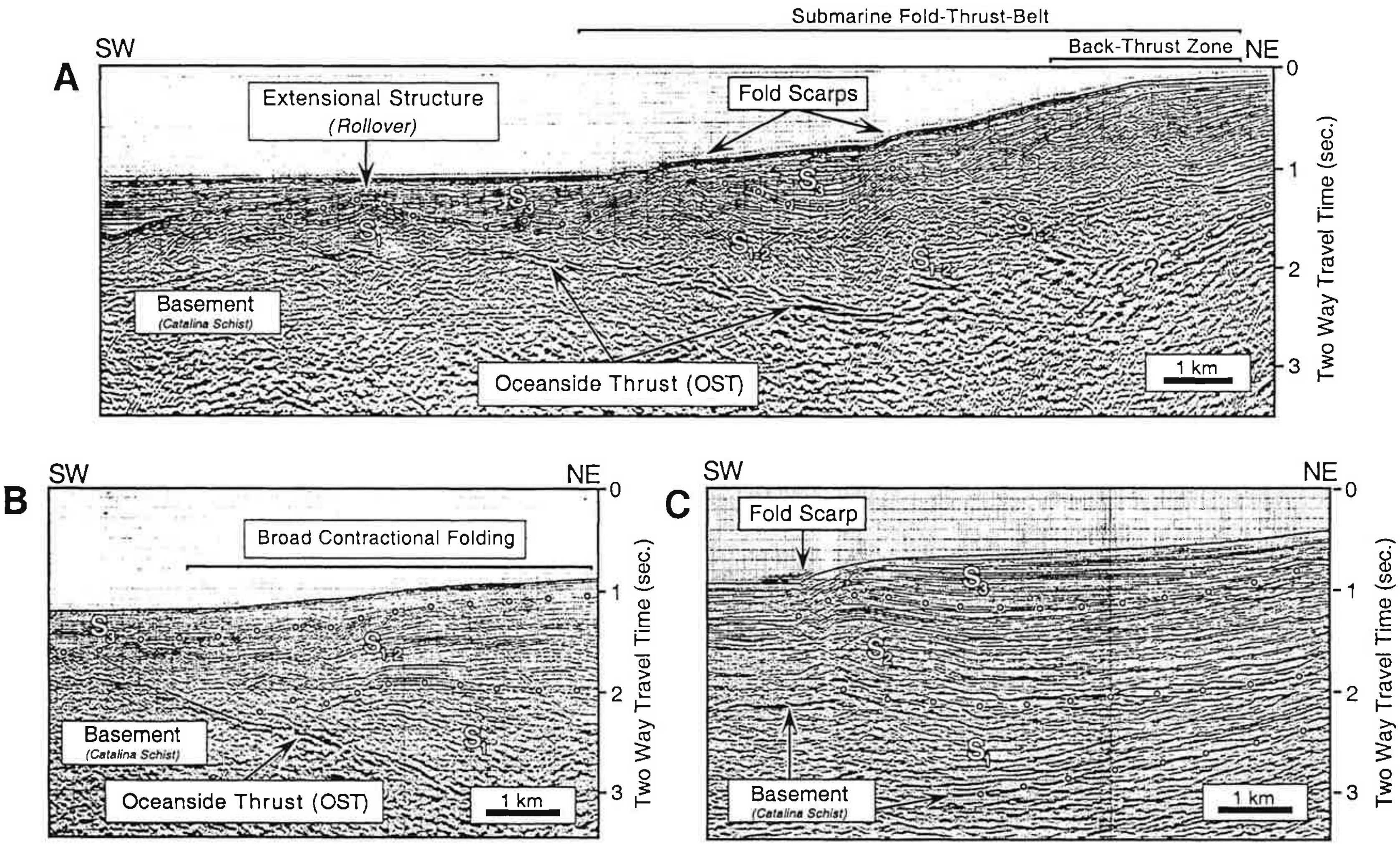

Figure 2. Geometry of Oceanside thrust as imaged in seismic reflection profiles. A: Migrated seismic reflection profile imaging segment of thrust south of Lasuen Knoll. Sharp and continuous reflections dipping to east define location of thrust. Note shallow fold-and-thrust belt above Oceanside thrust that produces seafloor fold scarps. This contractional deformation does not affect thrust; thus, we interpret Oceanside as basal thrust of sequence. Note older (Neogene) extensional rollover structure buried by Pliocene and younger strata preserved on west end of section. B: Migrated seismic image of Oceanside thrust northeast of Coronado Banks. Oceanside thrust motion is reflected by broad, contractional fold involving shallow sedimentary units and forming broad seafloor slope. C: Migrated seismic reflection profile across Carlsbad thrust, which resides in hanging wall of Oceanside thrust east of Crespi Knoll. Fault is defined by offset of top basement reflection, and produces contractional fold with pronounced seafloor scarp. Unit S_1 is Miocene and Oligocene(?) synextensional strata; S_2 is late Miocene–early Pliocene postextensional drape; S_3 is late Pliocene(?)–Holocene syncontractional strata. S_1 and S_2 are grouped where undifferentiated. Vertical scale is ~1:1; datum is sea level; s—seconds. Section traces are shown in Figure 1.

et al. (1999) proposed that the fold is developed above an active, southwest-dipping blind thrust that slips at a rate of ~0.42–0.79 mm/yr, based on the Oceanside blind thrust is responsible for all or part of this coastal uplift, it implies that the thrust is active far to the south of the San Joaquin Hills larly, the slip rate of the Thirtymile Bank thrust should be no more than 0.96 mm/yr, such that the resultant shortening does not exceed the geodetic