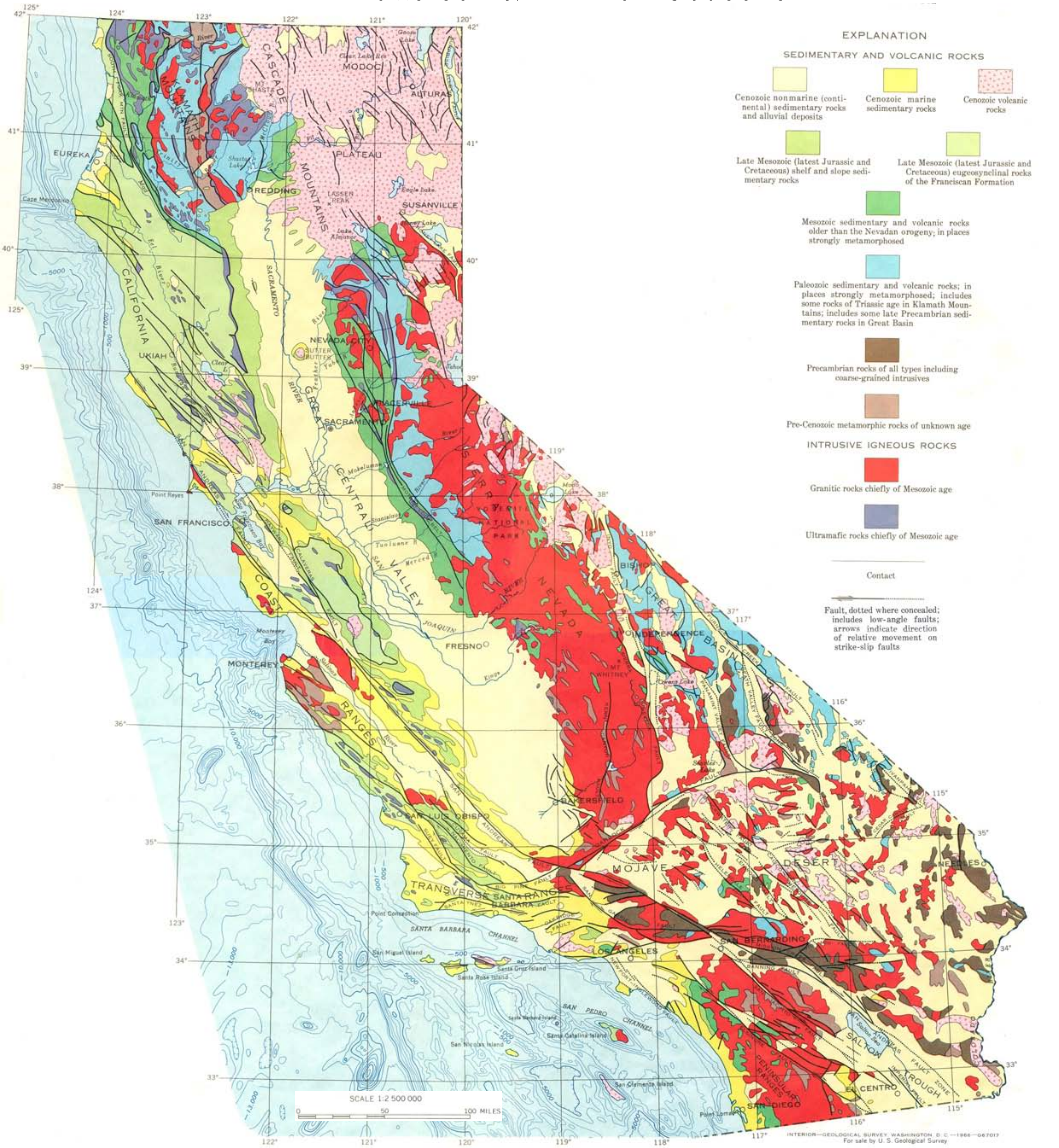
California 2006

Advanced Field Studies - ERTH 4807/ 5903

Dr. RT Patterson & Dr. Brian Cousens



Dept. of Earth Sciences
Carleton University
Aug. 18 - Sept. 4, 2006

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Marking Scheme

Notebooks 30%
Participation 30%
Final Project 40%

DAY	DATE	LOCATION	ADDRESS	TELEPHONE	ROOMS/CAMPSITES	RESERVATION NUMBERS
1	Friday 18 August	Courtyard by Marriot LAX Segundo	2000 E. Mariposa Ave El Segundo, CA 90245	310-322-0700	Seven rooms (booked by Expedia) Cousens-Kerrei-Mount-Cornejo-Carter-Patterson-Gupta	
2	Saturday 19 August	Upper Oso Campground	3505 Paradise Rd Santa Barbara, CA 93105	805-967-3481	10, 11, and 13	1.4617427NRRS 1.4617437NRRS 1.4617411NRRS (all under Cousens)
3	Sunday 20 August	Same as Day 2				
4	Monday 21 August	Same as Day 2				
5	Tuesday 22 August	San Simean State Park	Van Gordon Creek Rd at San Simean Creek Rd Cambria, CA 93428	805-927-2020	Three group sites, assigned on arrival	1.4805862CA (Cousens) 1.4805832CA (Cousens) 1.4812635CA (Patterson)
6	Wednesday 23 August	Same as Day 5				
7	Thursday 24 August	Monterey Downtown Days Inn	850 Abrego St. Monterey, CA 93940	831-649-6332	Seven rooms (booked by Expedia) Cousens-Kerrei-Mount-Cornejo-Carter-Patterson-Gupta	
8	Friday 25 August	June Lake Campground	Hwy 158, South Junction June Lake, CA	760-647-3045	004, 005, 006, 007	1.4622344NRRS 1.4622318NRRS 1.4622347NRRS

						(all under Cousens)
9	Saturday 26 August	Same as Day 8				
10	Sunday 27 August	Same as Day 8				
11	Monday 28 August	Same as Day 8				
12	Tuesday 29 August	Lone Pine Campground	Whitney Portal Rd Lone Pine, CA	760-873-2500	012, 013, 014, 017	1.4622376NRRS 1.4622383NRRS 1.4622390NRRS 1.4622398NRRS 1.4806889NRRS 1.4806929NRRS 1.4806952NRRS 1.4806967NRRS (all under Cousens)
13	Wednesday 30 August	Same as Day 12				
14	Thursday 31 August	Same as Day 12				
15	Friday 1 September	Joshua Tree National Park	Cottonwood Group 2 Site Cottonwood Spring, CA	760-367-5500		4990766 (Cousens)
16	Saturday 2 September	Same as Day 15				
17	Sunday 3 September	Ramada Plaza Hotel LAX	Hawthorne, CA 90250	310-536-9800	Seven rooms (booked by Expedia) Cousens-Kerrei-Mount-Cornejo-Carter-Patterson-Gupta	
18	Monday 4	Return to Ottawa				

Reservations and Travel

Budget Van Rental at LAX
Confirmation Numbers: 33273690CA2
33273852CA3
33274049CA4
33274180CA2

Day 1 Friday 18 August
Arrive in Los Angeles (LAX airport)
Meet at Budget Car Rental booth (take shuttle to off site booth) ASAP (around 12:30)

Hotel
Courtyard by Marriot LAX El Segundo
2000 E. Mariposa Ave
El Segundo, CA 90245
Tel: 310-322-0700
Seven rooms (booked by Expedia)
Cousens-Kerrei-Mount-Cornejo-Carter-Patterson-Gupta

Days 2-4 Saturday 19 August - Monday 21 August
Upper Oso Campground
3505 Paradise Rd
Santa Barbara, CA 93105
Tel: 805-967-3481
Camp sites 10, 11, and 13
Reservation ID:
1.4617427NRRS (Cousens)
1.4617437NRRS (Cousens)
1.4617411NRRS (Cousens)

Days 5-6 Tuesday 22 August – Wednesday 23 August
San Simean State Park
Van Gordon Creek Rd at San Simean Creek Rd
Cambria, CA 93428
Tel: 805-927-2020
Three group sites, assigned on arrival
Reservation ID:
1.4805862CA (Cousens)
1.4805832CA (Cousens)
1.4812635CA (Patterson)

Day 7 Thusday 24 August
Monterey Downtown Days Inn
850 Abrego St.
Monterey, CA 93940
Tel: 831-649-6332
Seven rooms (booked by Expedia)
Cousens-Kerrei-Mount-Cornejo-Carter-Patterson-Gupta

Days 8-11 Friday 25 August – Monday 28 August
June Lake Campground
Hwy 158, South Junction
June Lake, CA
Tel: 760-647-3045

Camp sites 004, 005, 006, 007
Reservation ID:
1.4622344NRRS (Cousens)
1.4622318NRRS (Cousens)
1.4622347NRRS (Cousens)
1.4622350NRRS (Cousens)

Days 12-14 Tuesday 29 August – Thursday 31 August
Lone Pine Campground
Whitney Portal Rd
Lone Pine, CA
Tel: 760-873-2500
Camp sites 012, 013, 014, 017
Reservation ID:
1.4622376NRRS (Cousens)
1.4622383NRRS (Cousens)
1.4622390NRRS (Cousens)
1.4622398NRRS (Cousens)
1.4806889NRRS (Cousens)
1.4806929NRRS (Cousens)
1.4806952NRRS (Cousens)
1.4806967NRRS (Cousens)

Days 15-16 Friday 1 September – Saturday 2 September
Joshua Tree National Park
Cottonwood Group 2 Site
Cottonwood Spring, CA
Tel: 760-367-5500
Reservation ID:
4990766 (Cousens)

Day 17 Sunday 3 September
Los Angeles
Ramada Plaza Hotel LAX
Hawthorne, CA 90250
Tel: 310-536-9800
Seven rooms (booked by Expedia)
Cousens-Kerrei-Mount-Cornejo-Carter-Patterson-Gupta

Day 18 Monday 4 September
Return to Ottawa!

Geologic Time Scale

Era	Period	MYA	Life at the Time
Cenozoic Era Age of Recent Life or The Age of Mammals	Quaternary	0.01 - 5	Many mammals vanish during a vast ice age. Modern humans emerge and spread world wide. Species of plants and animals are similar to what we see today
	Tertiary	5 - 145	The continents have moved to positions near where they are today. Flowering plants thrive and diversify. Vast forests exist in the tropical and temperate environments. Mammals spread and diversify.
Mesozoic Era Age of Medieval Life (Time of the Ruling Reptiles)	Cretaceous	145 - 200	North and South America begin to split apart. India is a separate continent, In the north are Euramerica and Asiamerica, with differing plant and animal species. Dinosaurs diversify and rule until the end of the Cretaceous period when they die out in a mass extinction along with many other marine and terrestrial species. Flowering plants appear, and insects begin to pollinate.
	Jurassic	200-250	The supercontinent, Pangea, splits and the Atlantic Ocean appears separating Asia from the Americas and Africa. The age of the Ruling Reptiles is in full swing. Dinosaurs rule the earth and pterosaurs rule the skies. Dinosaurs are much larger and include giant herbivores. The first birds appear.
	Triassic	250 - 295	The first dinosaurs and mammals appear; they tend to be small and quick predators who run on their back legs. Cycads abound, seed ferns go extinct. Ammonites are common. Dinosaurs, crocodiles and pterosaurs emerge and diversify.
Paleozoic Era Age of Ancient Life	Permian	295 - 362	All land is in one giant continent, Pangea. Ferns, seed ferns and conifers abound. Insects diversify and spread. Many marine animals, including the trilobites, go extinct.
	Carboniferous	362 – 418	All land is in two great continents, in the north is Euramerica and Gondwanaland is in the south. The first reptiles appear on land, as do the first conifers and cycads. Giant ferns, horsetails, and club mosses are common. Trilobites become less common, graptolites go extinct.
	Devonian	418 – 439	Amphibians and insects begin to invade the land. The earliest ferns and plants with seeds appear. Fish abound.
	Silurian	439 - 490	The first life emerges on land, as plants and invertebrates. Fish develop and split into the bony fishes (teleosts) and the cartilaginous fishes (sharks and rays). Marine invertebrates continue to thrive.
	Ordovician	490 - 543	Corals, bryozoa and graptolites thrive along with the marine life of the Cambrian. The first true vertebrate fish discovered.
	Cambrian	543 - 2500	Marine life abounds, including red and green algae, brachiopods, gastropods, trilobites, sponges. The earliest fishes appear.
Precambrian Era (Time before Life)		2500 – 4600	Time from the birth of the planet approximately 4.6 billion years ago, until the first simple life forms appeared about 3.6 bya, including early bacteria and blue-green algae. The first multicellular animals, such as worms and jellyfish, appeared near the end of this era.

From: http://www.nps.gov/akso/ParkWise/Students/ReferenceLibrary/Paleontology /GeologicTimeScale.htm

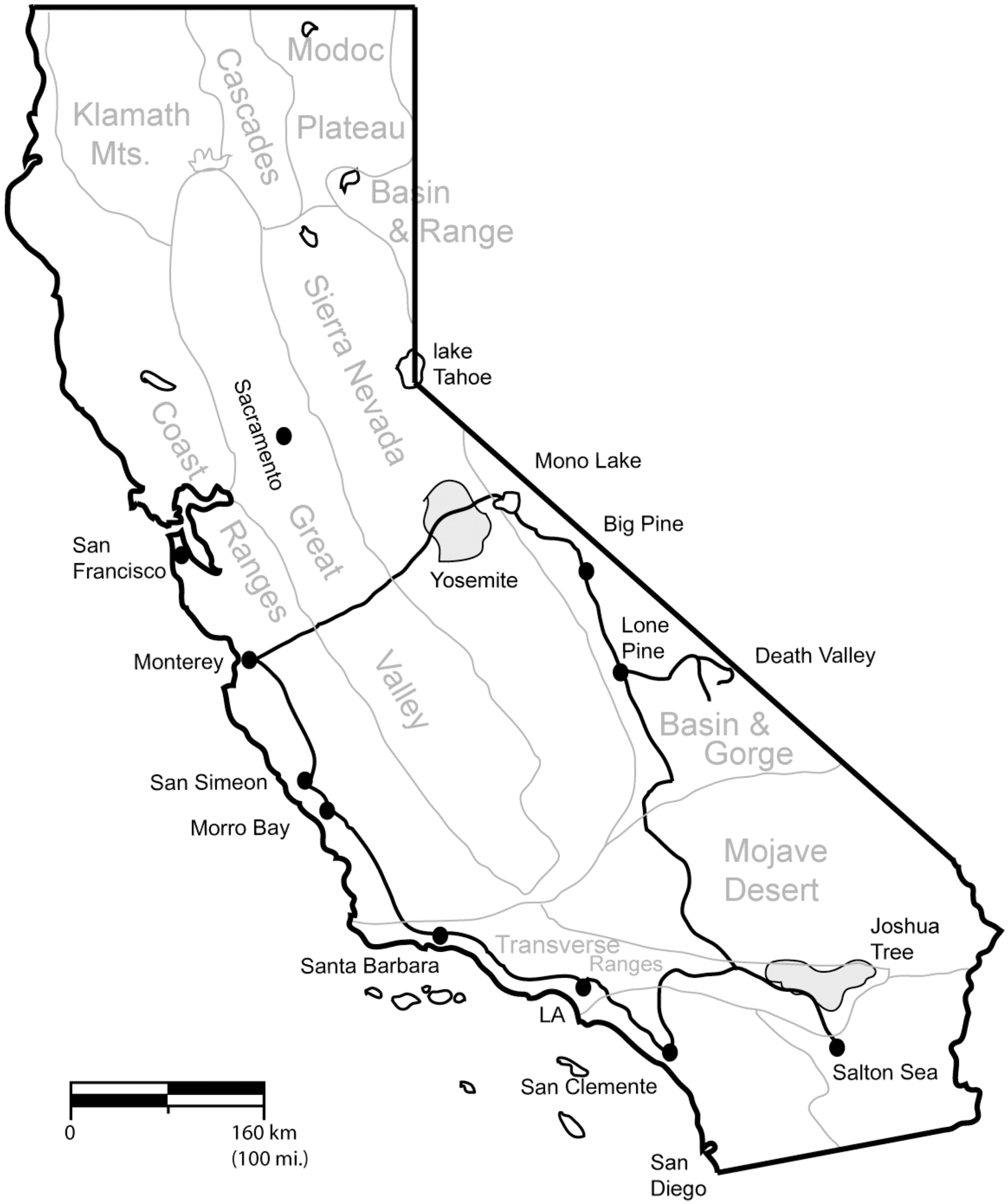

Figure 1 - Geomorphic Provinces of California (after Norris and Webb, 1989, Fig. 2.3, p. 54)

Day 1 - Friday, 18 August 2006

Travel from Ottawa to Los Angeles
Overnight in Los Angeles (Courtyard Marriott LAX)

Day 2 - Saturday, 19 August 2006

Summary: Leave Marriott
La Brea Tar Pits
Venice Beach
Travel to Upper Oso Campground via the San Fernando Valley

Directions: Courtyard Marriott LAX To La Brea Tar pits

Start out going West on E Mariposa Ave toward N Sepulveda Blvd / CA-1 N. 0.1 miles.
Turn right onto N Sepulveda blvd / CA-1 N. 0.5 miles.
Turn right onto E Imperial Hwy 0.2 miles.
Merge onto I-105 E. 0.7 miles.
Merge onto I-405 N toward Santa Monica 3.5 miles.
Take the La Tijera blvd exit- exit 48 0.2 miles.
Turn Right onto La Tijera Blvd. 0.6 miles.
Turn slight left onto S La Cienega Blvd. 3.5 miles.
Turn slight right onto S Fairfax Ave 2.4 miles.
Turn right onto Wilshire blvd. 0.3 miles.
End at La Brea Tar Pits 5801 Wilshire Blvd, Los Angeles, CA 90036 323-857-6311.
Reservations for 24 people made under Carter, $3.50 each, profs free. Parking behind the museum.

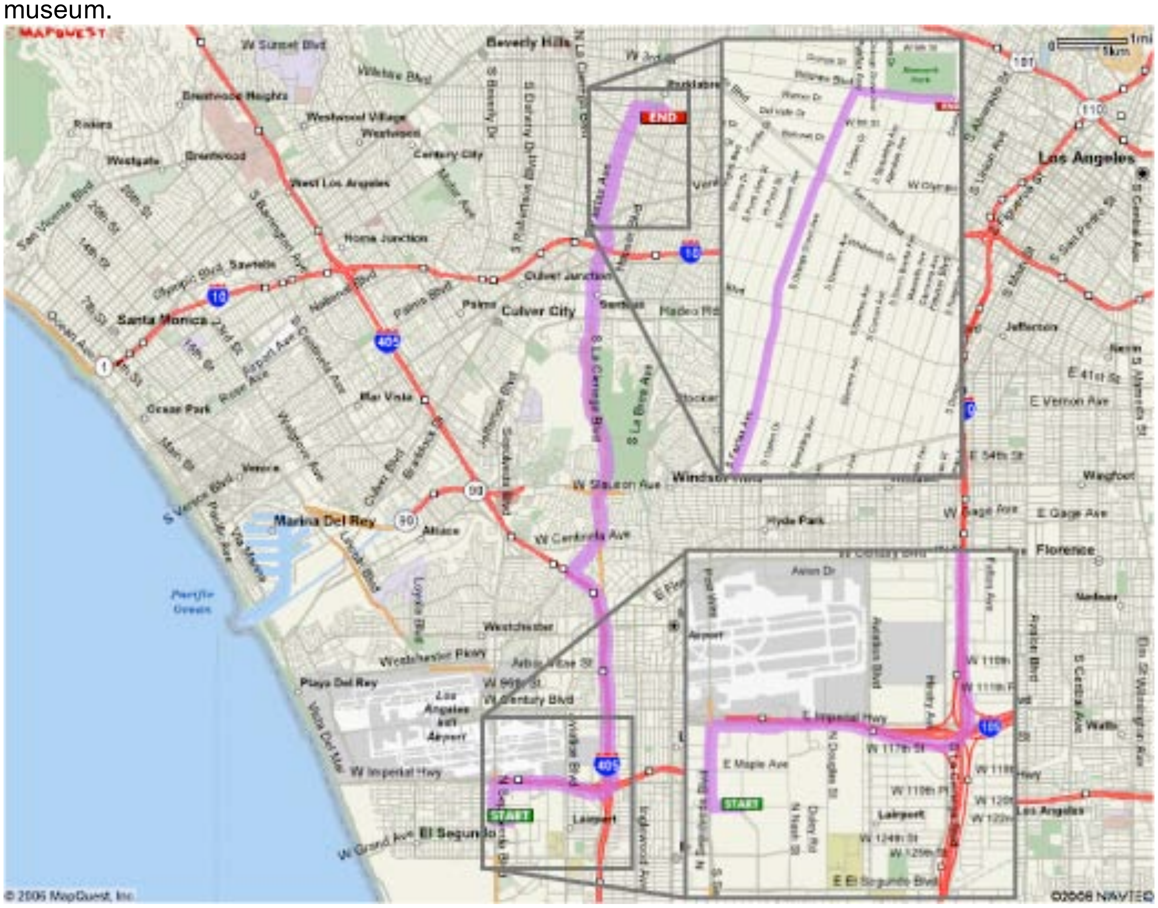

Figure 2: Marriott to La Brea Tar Pits

2 - 1 La Brea Tar Pits and Page Museum -

The Rancho La Brea Tar Pits are among the richest sources of ice age fossils in the world. These sticky asphalt beds trapped and preserved prehistoric plant and animal life. The viewing station and observation pit show how specimens appeared when they were discovered. Findings are displayed in the Page Museum. The George C. Page Museum of La Brea Discoveries located adjacent to the tar pits exhibits reconstructed fossils of such ice age animals as wolves, birds, horses and saber tooth cats. In the La Brea Story Theatre's 15-minute documentary, films and slides depict prehistoric life in the area. Visitors can watch fossils being cleaned, identified and catalogued in the paleontological laboratory.

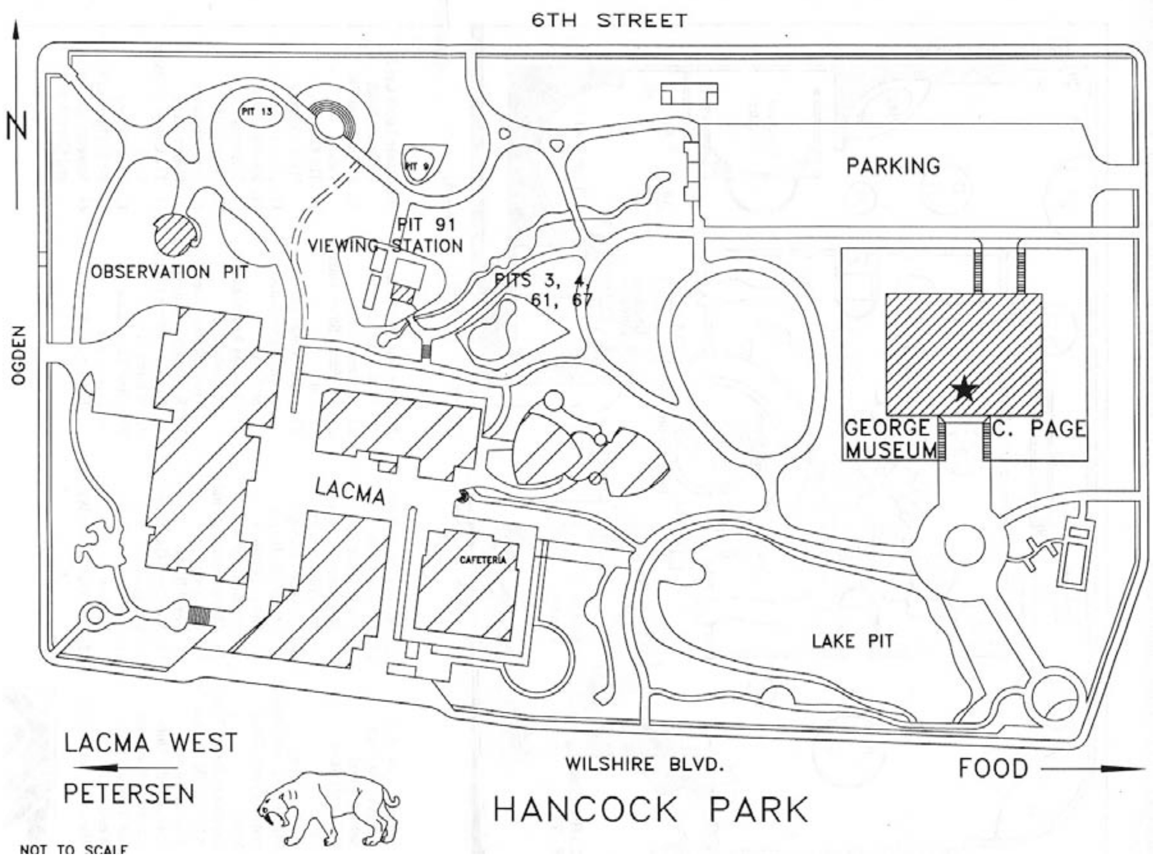

Figure 3: Map of La Brea

Background and Significance

Discovered in 1905, the La Brea tar pits are located on the Santa Monica Plain in the northwest part of the Los Angeles Basin, southern California. The pits are one of the most famous Late Pleistocene fossil deposits in North America (LaDuke, 1991). The array of fauna here not only represents the diversity of large mammals that inhabited North America, but also provides us with a picture of what life was like between 10,000 and 40,000 years ago as the Ice Ages drew to a close. Furthermore, they are also a key in vertebrate paleontology as the basis for the Rancholabrean Land Mammal Age (Akersten *et al*., 1983; Valkenburgh, 1994). Radiocarbon dates obtained from the pits range from 4450±200 for recent Indian artifacts to over 40,000 years for the tar itself as well as for a wood fragment (Valentine and Lipps, 1970). The collagen dates of bones for some of the fauna range from >36,000 years to approximately 11,000 years. Certain mammals are better represented in some pits than others. This may be due to several factors including environmental or climatic variations at varying times. For example, deer and timber wolves found in one pit represents forest conditions and therefore reflecting an environment different than a pit occupied by fauna such as camel, bison, and horse.

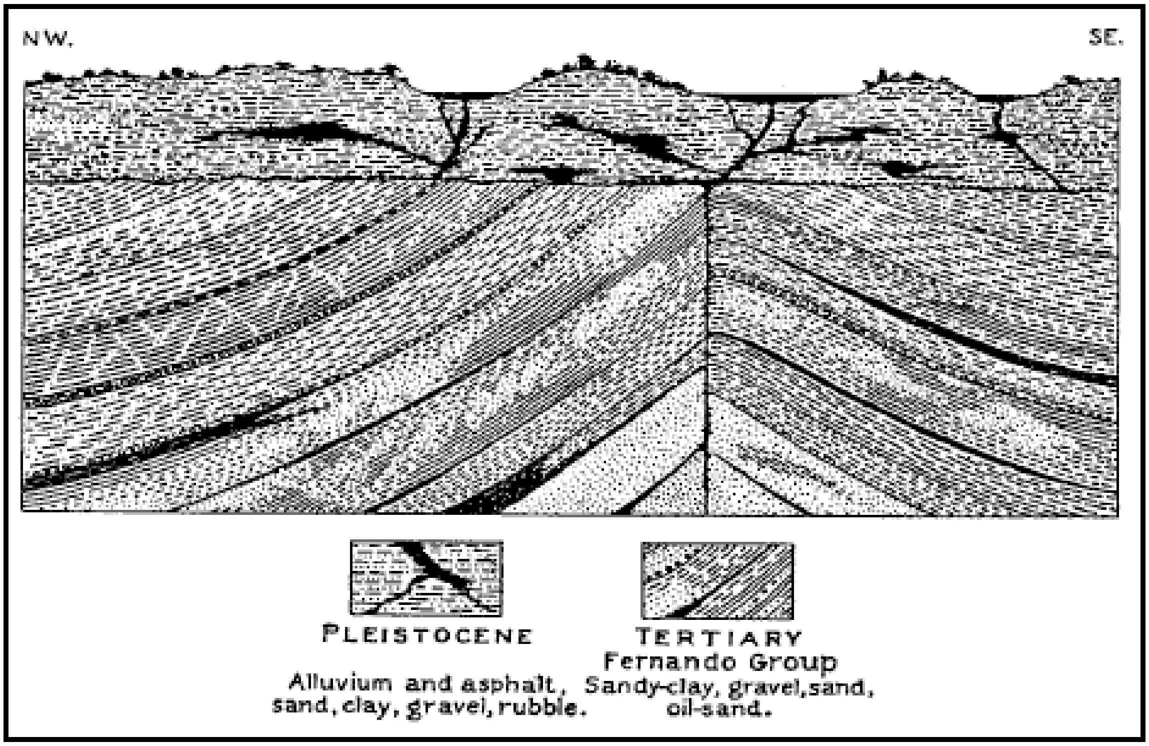

Figure 4 - Generalized cross-section showing geologic structure and relationships of formation at Rancho La Brea (Stock and Harris, 1992, p. 19)

Stratigraphy and Deposition

Late Pleistocene sediments at Rancho La Brea consist of clays, sands, gravels, and cobbles derived from the southern slopes of the Santa Monica Mountains (Reynolds, 1985; Stock and Harris, 1992). These sediments, which were laid down by streams, cover a much older structurally controlled petroleum reservoir that occurs at a depth of 350m within late Tertiary marine shales and sandstones (Akersten *et al*., 1983; Whitmer, 1987). As a result of movement in the earth's crust, these older marine strata are deformed and folded (figure 5). Heat and pressure forced the oil up through the cracks. In moving toward the surface, oil tends to lose its more volatile constituents, becoming more viscous and asphaltic. Normally, the tar pools were covered in thin sheets of slightly salty water, creating effective traps for the Pleistocene animals. (Norris and Webb, 1990).

A typical stratigraphic column at Rancho La Brea can be found on the wall at the Page Museum.. The section of interest where the fossil bones are found is within the late Pleistocene Palos Verdes Sand. The Palos Verdes Sand has been subdivided into three members A (marine), B (marine), and C from older to younger, respectively. The fossiliferous stratum is widespread throughout the region of the tar pits, although the thickness may vary.

Radiometric dates from member B provides an age of about 100,000 years or more for the marine biozone (Valentine and Lipps, 1970). The marine biozone represented by member B is believed to represent the Sangamonian Interglacial and is evidence for the latest known Pleistocene marine submergence of the northwestern Los Angeles Basin (Woodard and Marcus, 1973). Hence, this suggests that the Rancholabrean fauna were deposited no earlier than the Wisconsin age (Valentine and Lipps, 1970; Marcus and Berger, 1984). The bone-bearing beds are contained largely in the member C that makes up the upper 28 feet. This member is further subdivided in to three submembers as depicted in figure 6. The uppermost bone-bearing strata are largely of fluviatile origin as evident by the fabric of the sediments, abundance of coarse-grained sand, gravel, and conglomerate lenses, and the presence of freshwater limestone and molluscan faunas (Woodard and Marcus, 1973). Accumulation of these sediments, hence, followed a high sea-level stand representing the Sangamonian Interglacial period, as represented by member B (Woodard and Marcus, 1973).

The Fauna

In today's ecosystems, herbivores are more abundant than carnivores, but at La Brea, only 9% of the fauna remains from the pits were of herbivores and the remaining 91% were of carnivores (Figure 6) (Linze, 1948; Norris and Webb, 1990). The dominance of carnivores in the pits is believed to be due to larger packs assembling together in the pit to kill and consume a single herbivore (Linze, 1948).

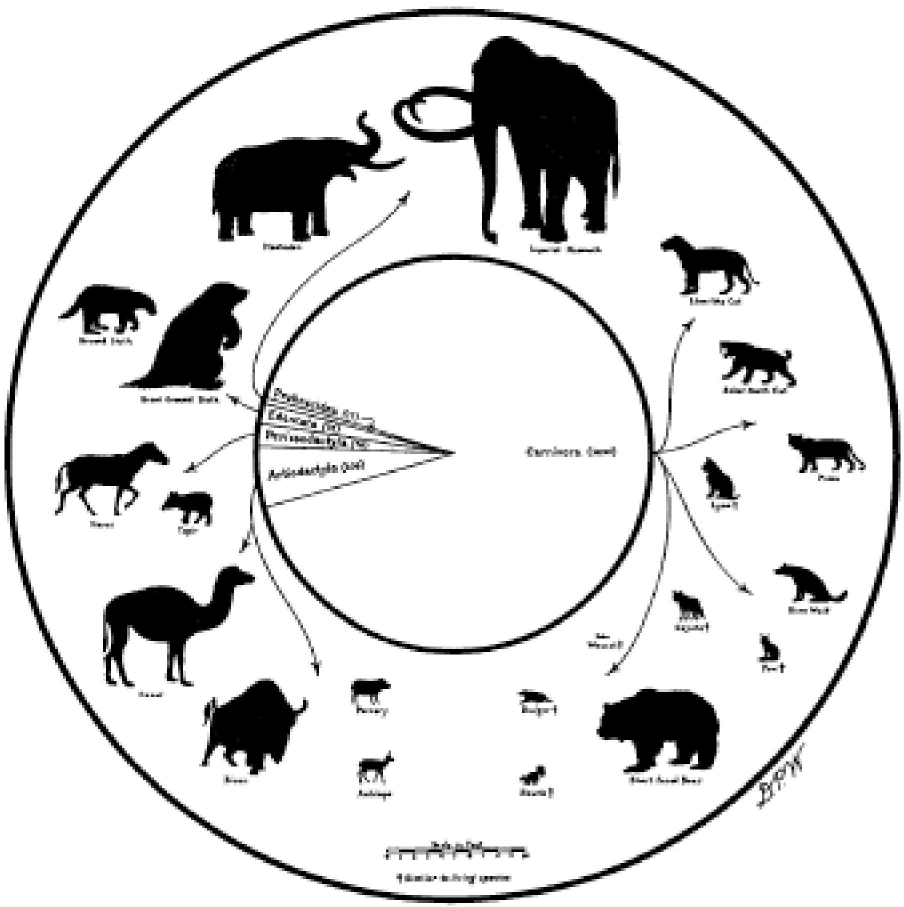

Figure 6 - Diagram illustrating relative number of individuals in the mammalian orders occurring in the Rancho La Brea Pleistocene fauna (Stock and Harris, 1992, p. 29)

Since 1906, more than a million bones have been recovered representing over 231 species of vertebrates, 159 kinds of plants and 234 kinds of invertebrates. Dire wolves are the most common of the large mammals recovered, followed by the saber-tooth cat. Together, both extinct and living species have been collected from the site.

In 1914, a human skull and partial skeleton dating at 9,000 years old was discovered from Pit 10. It was from a young woman (La Brea Woman) about 22-24 years old, standing 4'10" tall. A few artifacts and the skull and mandible of a domestic dog were also found. Evidence suggests that she was a victim of a violent blow to the head (Reynolds, 1985). The material appears to represent a secondary burial (reburial) rather than an accidental entrapment event (Marcus and Berger, 1984; Reynolds, 1985).

Entrapment and Preservation

The Rancho La Brea fossils were preserved by a unique combination of sedimentation and asphalt impregnation. Summer's heat dried the streams and warmed the semi-solid asphalt. Hence, it was during summer months that the fauna became trapped. Winter and spring rains filled the streams with sand and silt resulting in the burial of the fossils. The bones are believed to have lain on the surface for some time prior to burial. This is supported by the abraded and weathered nature of the bones (Woodard and Marcus, 1973). The tissues of the animals have all decayed but the skeletons remain perfectly intact. Teeth and very delicate bones have all been preserved. Although the pits at La Brea are known to have trapped thousands of animals over a period of about 30,000 years, many believe that only about 10 animals were trapped every decade, all of them victims of the same entrapment episode (Thompson, 1990). Hence, an average of one major entrapment every ten years, over a period of 30,000 years, would be sufficient to account for the number of fossils found at La Brea. Entrapment of organisms is still occurring today. About 8-12 gallons (32-48 litres) of asphalt seep to the surface today, trapping invertebrates (insects and worms), reptiles (lizards), birds (pigeons, hawks, ducks, doves, sparrows), small mammals (rodents and rabbits) and occasionally large mammals.

Extinction

Of the large mammals from Rancho La Brea, 40% (24 species) are now extinct. The cause(s) for extinction of the late Pleistocene fauna is still under debate. The arrival of humans to North America may have led to extinction through what is known as the overkill hypothesis. The most favored cause is a due to climate change. The fact that a colder and wetter climate existed in Los Angeles 11,000 years ago is evident from La Brea snails. Some of the snail recovered occurs today in surrounding colder lakes. Others are found today further to the north where precipitation is greater. Also, coast redwoods and other cool, moisture-loving trees lived in the area as evident in the deposits at La Brea. Today, they have vanished from the area and are only found in higher latitudes and altitudes (Thompson, 1990). Perhaps extinction was the result of a combination of these factors.

Late Pleistocene Environment at Rancho La Brea

A record spanning more than 25,000 years is present at La Brea tar pits. This allows for a complete paleoenvironmental reconstruction encompassing periods of Wisconsin glacial maxima and interstadials. Evidence for paleoenvironmental reconstruction comes mostly from the analysis of the plant fossils, as they are sensitive indicators of climate (Stock and Harris, 1992). No other fossil site has produced a complete representation of an ancient terrestrial ecosystem as at Rancho La Brea. Many of the preserved plants are from Pit 91 and record a variety of communities in and around a stream-fed marshy channel (Marcus and Berger, 1984). Six fossil plant associations and groups (closed-cone pine forest, chaparral and foothill woodland, freshwater aquatics and moistureloving herbs, and herbs of drier situations) have been identified (Akersten *et al*., 1983). Some of these represent habitats proximal to the site of deposition whereas others may have been brought in by distal habitats. Based on the vegetation studies, the flora appears to indicate that the climate was similar to that of the present with dry summers (subtropical) and strong seasonal rainfalls. However, precipitation was probably greater then and the overall temperature cooler (Akersten *et al*., 1983).

Suggested Readings

Akersten, W.A., Shaw, C.A., and Jefferson, G.T. 1983. Rancho La Brea: status and future. Paleobiology, v. 9, no. 3, pp. 211-217.

LaDuke, T.C. 1991. The Fossil Snakes of Pit 91, Rancho La Brea, California. Contributions in Science, no. 424, pp.

Linze, D.W. 1948. Now They Are Gone: California Tar Pits Supply Paleontologists with Data on Ice Fauna. The Earth Science Digest, v. 2, no. 6, pp. 7-11.

Marcus, L.F. and Berger, R. 1984. The Significance of Radiocarbon Dates for Rancho La Brea. In: Quaternary Extinctions, Tucson: The University of Arizona Press, pp. 159-180.

Norris, R.M. and Webb, R.W. 1990. Geology of California. New York: John Wiley & Sons, Inc., p.

353.
Reynolds, R.L. 1985. Domestic Dog Associated with Human Remains at Rancho La Brea. Bulletin: Southern California Academy of Sciences, v. 84, no. 2, pp. 76-85.
Stock, C. and Harris, J.M. 1992. Rancho La Brea: A Record of Pleistocene Life in California. Science Series (Los Angeles), no. 37, 113p.
Thompson, S.E. 1990. Tales from the Tar Pits. Lapidary Journal, v. 43, no. 10, pp. 81-88.
Valentine, J.W. and Lipps, J.H. 1970. Marine Fossils of Rancho La Brea. Science, v. 169, pp. 277-278.
Valkenburgh, B.V. 1994. Tough Times in the Tar Pits. Natural History, v. 103, pp. 84-85.
Whitmer, J.H. 1987. Rancho La Brea: A New Interpretation. Geological Newsletter, v. 53, no. 1, p. 3.
Woodard, G.D. and Marcus, L.F. 1973. Rancho La Brea Fossil Deposits: A Re-Evaluation From Stratigraphic and Geological Evidence. Journal of Paleontology, v. 47, no. 1, pp. 54-69.
www.ucmp.berkeley.edu/index.html
www.tarpits.org/exhibits/fossils/index.html

2 - 2 Venice Beach and Long Shore Transport

Directions: La Brea Tar Pits to Venice beach

Start out going South on S Curson Ave toward W 8th St 0.3 miles.
Turn right onto W Olympic Blvd 0.3 miles.
Turn left onto S Fairfax Ave 1.3 miles.
Turn right onto Venice Blvd 0.3 miles.
Turn right onto David Ave 0.1 miles.
Merge onto I-10 W / Rosa Parks Fwy 2.8 miles.
Merge onto I-405 S via exit 3B toward LAX airport / Long Beach 3.5 miles.
Merge onto CA-90 W toward Marina Del Rey 2.8 miles.
Turn slight right onto Lincoln blvd / CA-1 N 0.5 miles.
Turn left onto Washington blvd 1.3 miles.
End at Venice Fishing Pier 1 Washington Blvd, Marina Del Rey.

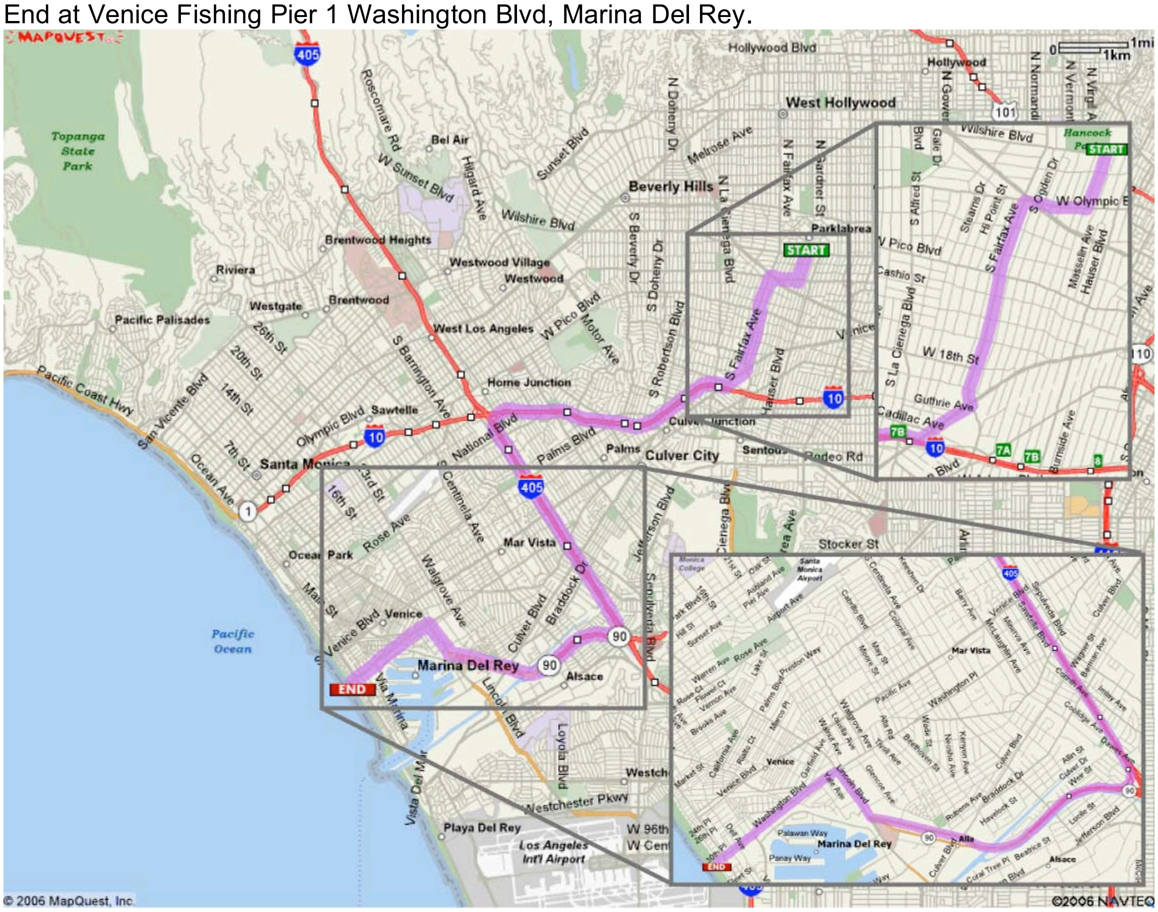

Figure 7: La Brea to Venice Beach

Venice beach is a unique cultural experience. It is characterized by a variety of interesting denizens and attractions including the famed Muscle Beach, turbaned guitar playing roller skating troubadours, skate board grandmothers, chain saw jugglers, tee-shirt and sun glass hawkers all set against the back-drop of the Pacific Ocean. Skimpily clad sunworshipers stroll up and down the mile long boardwalk, seeing and being seen, while others either lie on the expansive beach or splash in the pounding surf.]

Analysis of Longshore Sediment Transport

Ocean shorelines, such as Venice Beach, California, are constantly being pounded on by processes such as waves, currents, and tides, which continually modify existing shoreline features and structures. Longshore currents carry sediment along shorelines and are also responsible for the deposition of spits and baymouths, which occur between the breaker zone and the beach (Brown *et al*., 1989) (Figure 8). As one part of a wave enters shallow water and breaks as it encounters wave base, that part of the same wave still in deep water races ahead until it too encounters wave base. The net effect is called wave refraction where the wave bends so that they more nearly parallel the shoreline. Hence, when waves approach a straight coastline at an oblique angle, a longshore current is formed. These longshore currents are best developed along straight coastlines and are an important way of moving bars (Figure 9), both of which are variations of the same feature. A *spit* is a continuation of a beach forming a point that projects into a body of water, commonly a bay. A *baymouth bar* is a spit that has grown until it completely closes off a bay from the open sea.

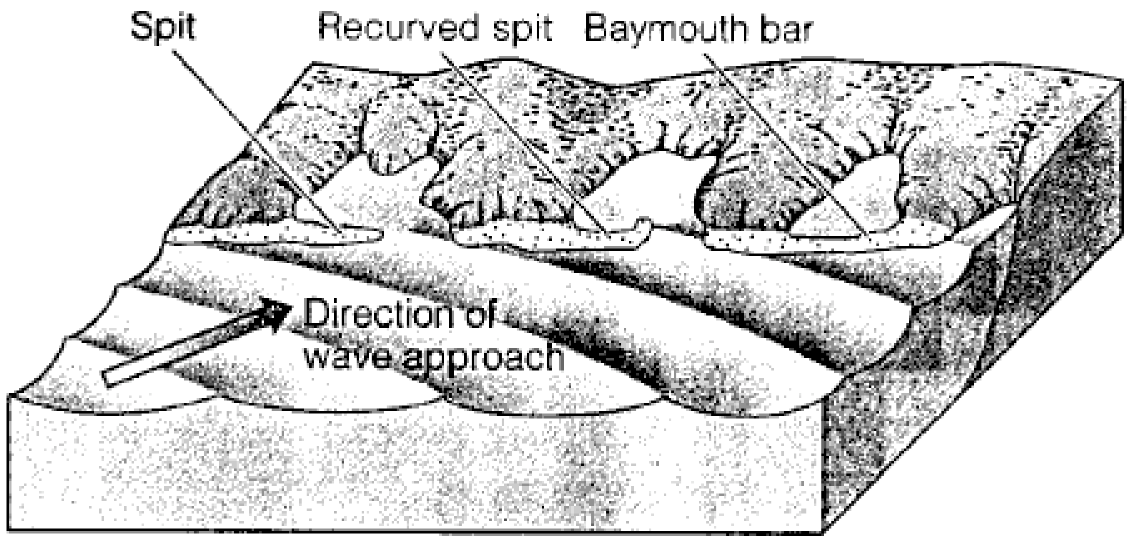

Figure 8 - Spits form where longshore currents deposit sand in deeper water as at the entrance to a bay. A baymouth bar is simply a spit that has grown until it extends across the mouth of the bay (Wicander and Munroe, 1992, p. 584)

Since the early 1900s, there has been much interest in the study and analysis of long-shore sediment transport along California's southern coast (Busen, 1983). Attention has increased with increasing urbanization and coastal development. Analysis of long-shore sediment transport requires a clear understanding of the various parameters, which influence the movement of sediment. This includes gathering necessary information about the physical environment (tides, currents, wind and wave conditions, bathymetry, sand sources, and grain size) as well as sediment transport obstacles such as manmade structures (groynes, piers) and offshore bedrock outcrops (Busen, 1983). Field study methods include the use of sand traps. Indirect or lab studies may also be applied and range from sediment budget studies, aerial photo analyses, and hydraulic modelling.

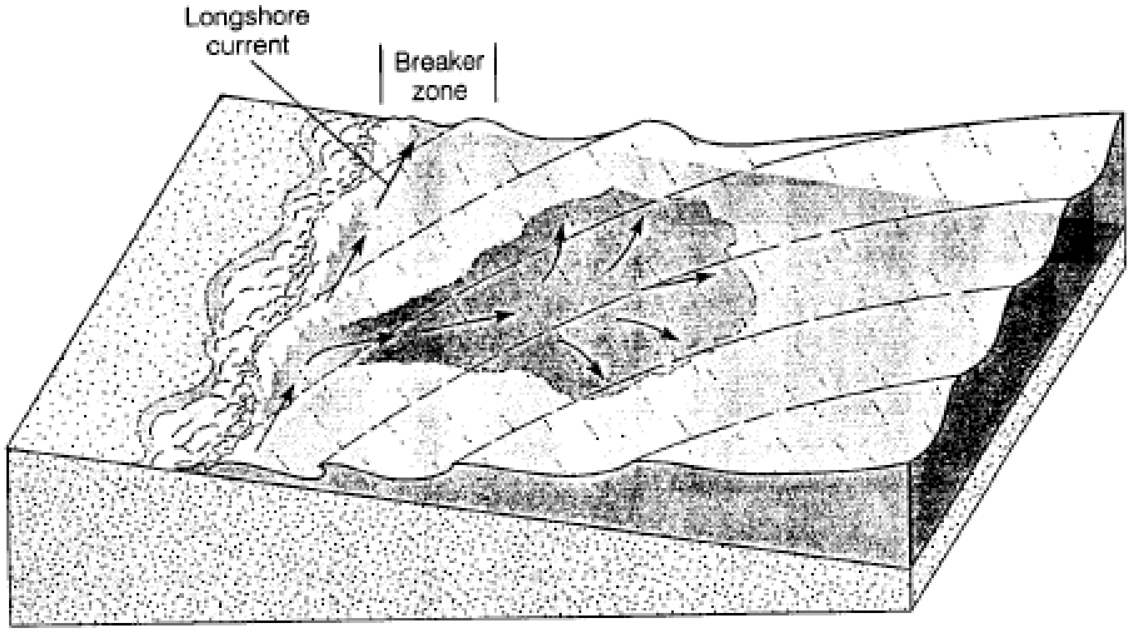

Figure 9 - Migration of Longshore currents parallel to the shoreline (Wicander and Monroe, 1992, p. 581)

Field Studies

1) Sand traps

The primary force in longshore sediment movement along the southern coast of California is the longshore currents located between the breaking waves and the shoreline. If a barrier were placed across this zone, the net effect would be deposition of littoral material on the updrift side of the barrier until an equilibrium profile is reached. Once equilibrium is established, net longshore transport of the material will continue around the end of the barrier. Such barriers can be used to obtain a direct indication of the net littoral drift rate by measuring the rate of the deposition of littoral material on the updrift side of the barrier. This holds true only if the barrier extends across the breaker zone and that no material moves around its seaward end. Groynes, breakwaters, and jetties are examples of barriers to longshore transport. This method is often expensive and also has an adverse impact on loss of beach property due to erosion. A simpler method is using sand traps to estimate sediment movement in the littoral zone. The technique involves placing arrays of sand traps across the width of the littoral zone and to measure the amount of sediment that is entrapped over a span of time. The greatest problem associated with the traps is the influence of the trap structure on the water flow characteristics.

Lab Studies

1) Sediment budget studies

Sediment budget studies focus on littoral cell boundaries. A littoral cell is defined as a length of coastline where the supply and loss of littoral material within the cell can be estimated to give a net gain or loss over a period of time. Sediment budget studies require long-term steady state conditions of the littoral cells. In other words, the sediment transported into each cell (via longshore transport of upcoast material, sediment runoff from streams within the cell, onshore movement of material from offshore sources, cliff erosion) must equal the sediment transported out (via longshore transport of material downcoast, sediment loss down submarine canyons, off-shore movement of sediment, sediment loss into inlets, bays, and harbors) if no accretion or erosion occurs. This condition is only reached when a true steady state is reached.

2) Aerial photo analysis

Aerial photos taken of the same area of shoreline many years apart can be used to estimate the rate of erosion or accretion of the shoreline. The basis for determining shoreline erosion or accretions and the associated long-shore sediment transport rate in an aerial photo study is by measuring the distance to the standardized shoreline from a stationary object such as a roadway or railroad that has remained in place over the time period being studied.

3) Hydraulic modeling

These studies involve the use of wave-generating tanks and scaled-down models of desired beach configuration. Hydraulic models are very valuable in evaluating the impact of manmade structures such as groyne fields on the shoreline configuration. Disadvantages to this method include not only cost but also the availability of proper facilities. Also, the reliability of this study is dependent on coefficient and parameter values that must be obtained through field measurements of the area being modeled.

Suggested Readings

Brown, J., Colling, A., Park, D., Phillips, J., Rothery, D., and Wright, J. 1989. Waves, Tides and Shallow-Water Processes. Oxford: The Open University, pp. 1-187.

Busen, K.L. 1983. Various Methods for Determining Long-Shore Sediment Transport and Associated Coastal

Erosion Along The Southern California Coast. Proceedings of the Symposium on Coastal Sedimentology, v. 6, pp. 71-90.

Monroe, J.S. and Wicander, R. 1992. Physical Geology: Exploring the Earth. St. Paul: West Publishing Company, 639p.

Slope Instability

The steep cliffs of the California coast are susceptible to land slides. Balancing the increasing pressures of urbanization and development against risk from potentially catastrophic land movements remains a challenge. There are three main factors that control the type and rate of mass wasting that might occur at the Earth's surface: 1.) Slope gradient: The steeper the slope of the land, the more likely that mass wasting will occur. 2.) Slope consolidation: Sediments and fractured or poorly cemented rocks and sediments are weak, and more prone to mass wasting. 3.) Water: If slope materials are saturated with water, they may lose cohesion and flow easily. Whenever a mass of slope material moves as a coherent *block* , we say that a slide has taken place. There are several types of slides, but one of the most common is a slump. A slump occurs when a portion of hillside moves down slope under the influence of gravity. A slump has a characteristic shape, with a scarp or cliff at the top of the slump, and a bulge of material (often called the toe of the slump) at the base of the slump shown in Figure 10. (http://seis.natsci.csulb.edu/basicgeo/landslid.html)

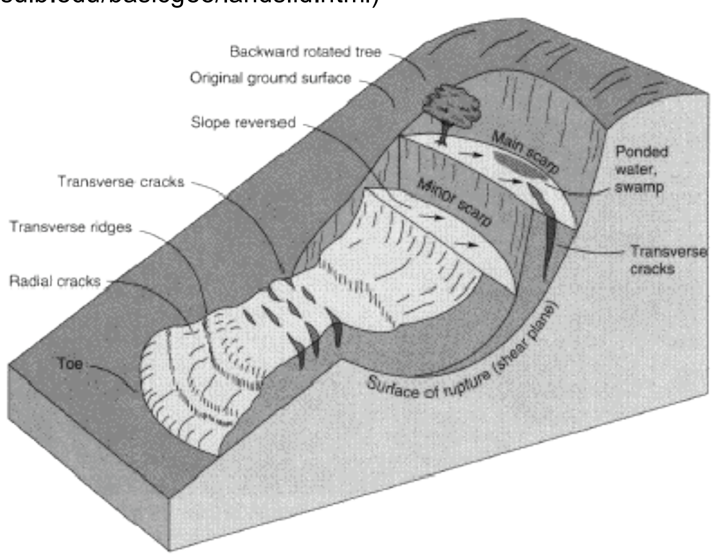

Figure 10: Schematic of a slump (http://www.ussartf.org/landslides.htm)

As Figure 11 shows, housing developments along the highly desirable coastline are built at the top of cliffs, on the slopes and at the foot, that has lead to substantial property damage and loss of life.

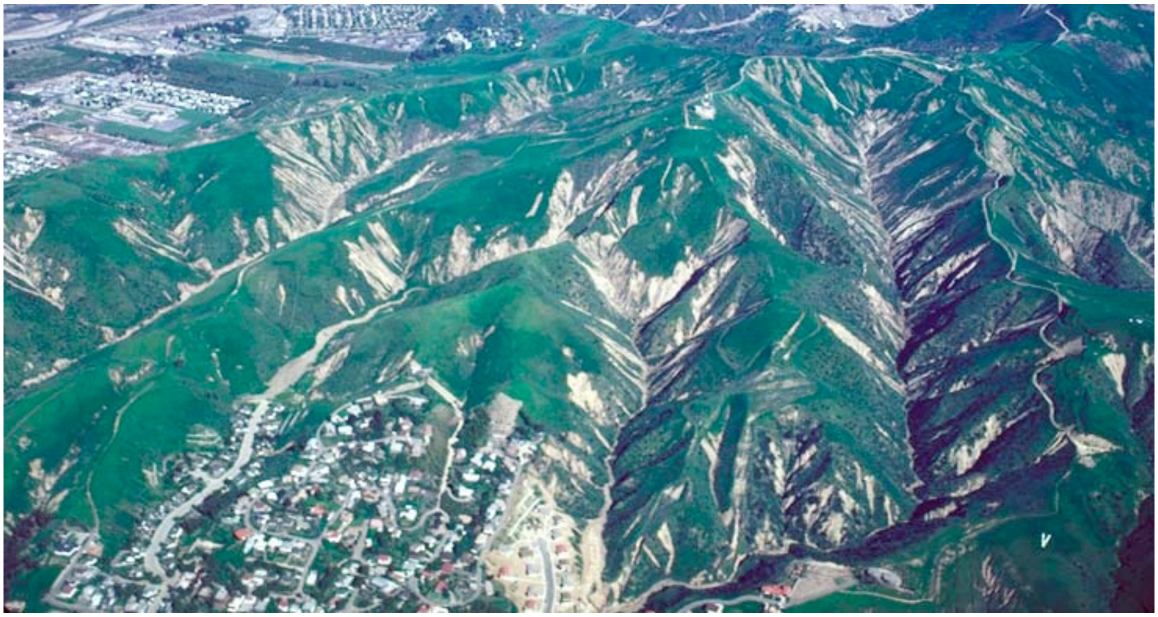

Figure 11. Extensive residential development on unstable slopes. (http://geology.wr.usgs.gov/wgmt/elnino/scampen/examples.html)

On January 10, 2005, a landslide struck the community of La Conchita in Ventura County, California, destroying or seriously damaging 36 houses and killing 10 people. The failure of the cliff face was covered extensively by the North American media, but it was a story that had been told before. Ten years earlier a similar landslide occurred in the same location. Half hearted efforts to safeguard the community below were made. Steel–and–timber wall built after the 1995 landslide but was overtopped and tilted forward in places by the 2005 landslide, as seen in Figure 12. A steel–and–timber wall was built after the 1995 landslide was overtopped and tilted forward in places by the 2005 landslide. The tight knit affluent community has collected insurance monies and had re-built their quiet community in the same location, minus the 10 who died.

Figure 12: Aftermath of the La Conchita landslides (http://landslides.usgs.gov/learningeducation/images/laconchita/2005/IMG_0127a.jpg, http://pubs.usgs.gov/of/2005/1067/508of05-1067.html#conchita03)

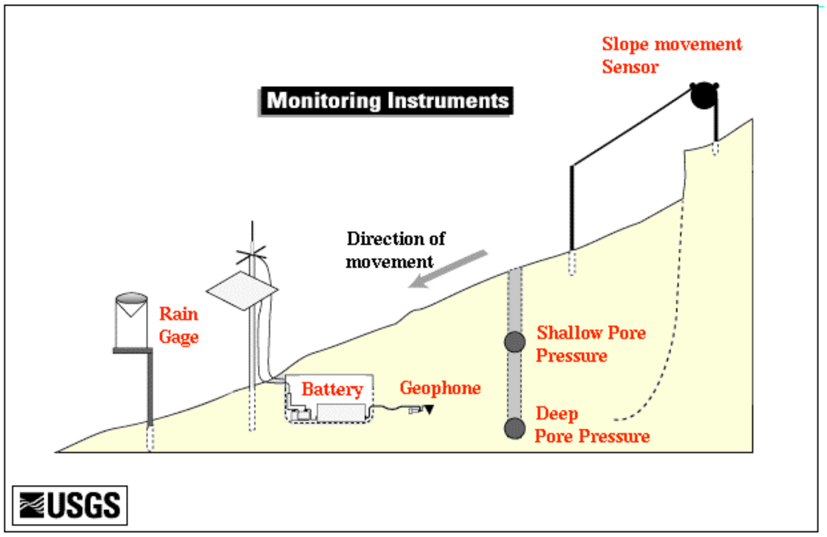

Figure 13: The USGS is using remote sensing to monitor areas of concern along the California coast. The automated sensors send alarms to USGS regional centres.
http://landslides.usgs.gov/monitoring/hwy50/rtd

Directions: Venice Beach Fishing pier To Upper Oso Campground, Start out going Northeast on Washington blvd toward Speedway. 1.3 miles.
Turn Right onto Lincoln blvd / CA-1 S. 0.5 miles.
Turn left onto CA-90 E / Marina Expy. Continue to follow CA-90 E 2.3 miles.
Merge onto I-405 N toward Sacramento 13.2 miles.
Merge onto US-101 N toward Ventura 82.9 miles.
Take the State st exit toward CA-154 / Cachuma lake 0.1 miles.
Turn left onto State st. 0.3 miles.
Turn right onto San Marcos Pass rd / CA-154 10.7 miles.
Turn right onto Paradise rd / NF-5N18 4.9 miles.
End at 3505 Paradise Rd, Santa Barbara

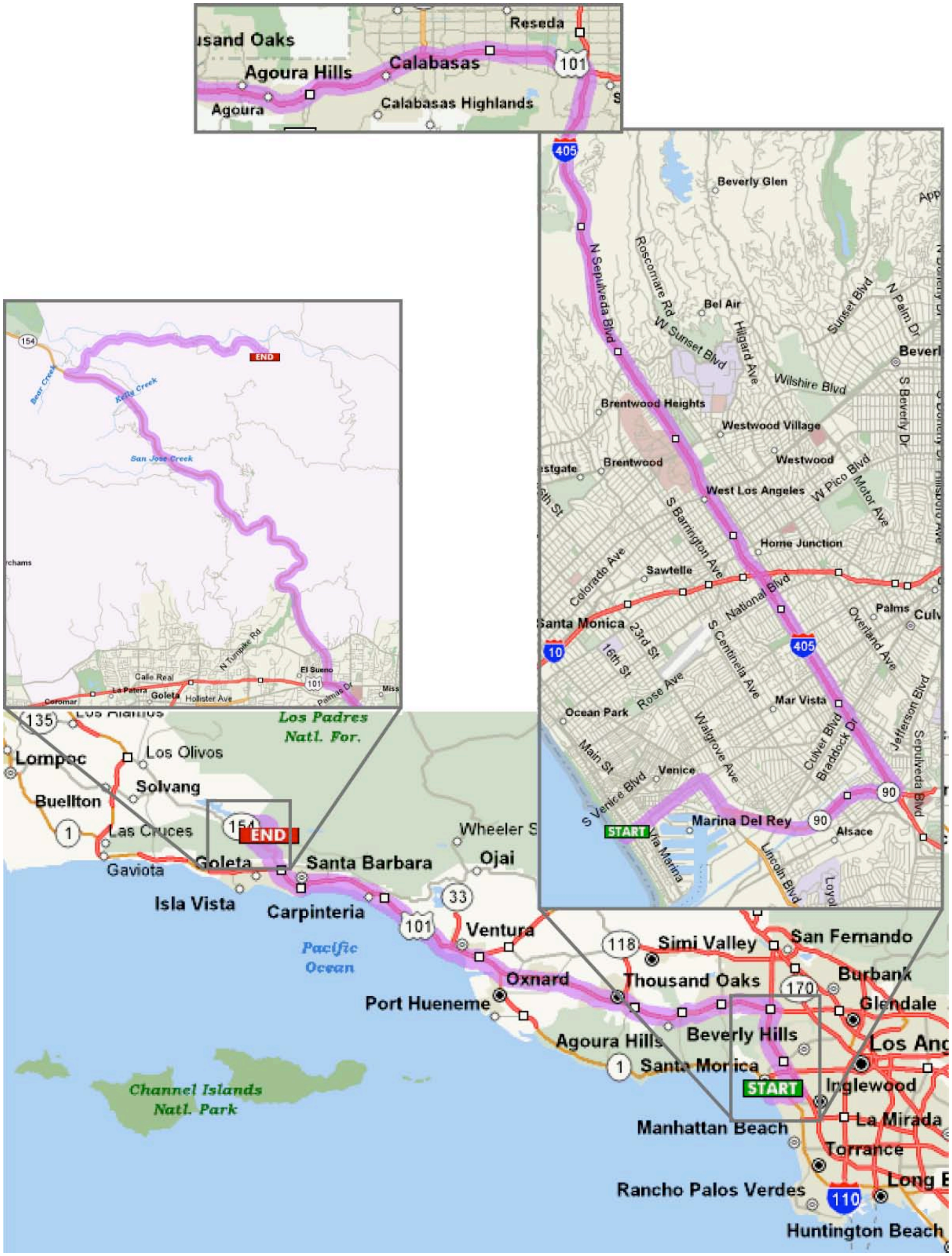

Figure 14: Venice Beach to Oso Campground

2-3 Introduction to Transverse Range Province and San Fernando Valley

San Gabriel Mountains

The San Gabriel Mountains are a high, rugged block located between the Los Angeles Basin and the Mojave Desert. They form a continuous feature 96 kilometers (60 miles) long and up to 39 kilometers (24 miles) wide. The Sierra Madre fault zone forms the range's southern boundary. The eastern boundary is the San Andreas fault zone, which crosses through Cajon Pass and separates the similar but higher San Bernardino Mountains. The San Gabriel Mountains face the Soledad Basin on the northwest and the San Fernando Valley on the west (Figure 15).

Because the San Gabriel Mountains have experienced considerable uplift in recent geologic time, the range has become a deeply dissected, rugged horst. Stream canyons are steep-sided and up to 900 meters (3,000 feet) deep. Many peaks exceed 2,100 meters (7,000 feet) in elevation, the highest being Mount San Antonio (Old Baldy) at 3,074 meters (10,080 feet). The southern and western flanks are very steep and abrupt where they face the lowlands of the Los Angeles Basin and the San Fernando Valley. The north face is less dramatic although almost as steep. The difference results from the greater elevation of the Mojave Desert, with its floor nearly 1,200 meters (4,000 feet) high at the base of the range.

A feature of the San Gabriel Mountains that is critical to understanding the geologic history of North America is the presence of ancient crystalline rocks, particularly in the northwest part of the range. Included are extensive exposures of anorthosite which is known almost exclusively from pre-Phanerozoic terrains (and from lunar sites). The only other California exposure is in the Orocopia Mountains. The San Gabriel anorthosites have been dated as 1,022 million years old. Other ancient rocks are the Mendenhall gneiss (1,045 million years old) and the augen gneisses (1,700 million years old). These rocks are not as old as the Archean rocks of the Lake Superior region, but they are earliest Proterozoic nonetheless.

Other large exposures of metamorphic rocks also exist in the northeast San Gabriel Mountains. These rocks are generally assigned to the Pelona schist, which is thought to have either a Jurassic or Cretaceous protolith and to have undergone metamorphism in either the late Cretaceous or Paleocene. In addition, another thick sequence of metamorphic rocks is located in the eastern San Gabriel Mountains, probably originally Paleozoic sedimentary rock. Mesozoic granitic rocks dominate the San Gabriel Mountains and constitute perhaps 70 percent of the exposed rocks. Cenozoic beds are located only along the range's margins. The oldest of these is the marine Paleocene assigned to the San Francisquito (formerly Martinez) Formation, which occurs near Devils Punchbowl and near Cajon Pass.

Los Angeles Basin

The Los Angeles basin is a name applied to both the modern physiographic coastal lowland as well as the 30 km wide by 80 km long remnant of a late Miocene to Holocene sedimentary basin. Some geologists limit the Los Angeles Basin to the coastal plain between the Santa Monica Mountains on the north, the Puente Hills and Whittier fault on the east, the Santa Ana Mountains and the San Joaquin Hills on the south, and the Palos Verdes Peninsula and the shoreline on the west. The U.S. Geological Survey, conversely, takes a broader view, dividing the basin into four blocks that contain both uplifted portions and synclinal depressions (Figure 16).

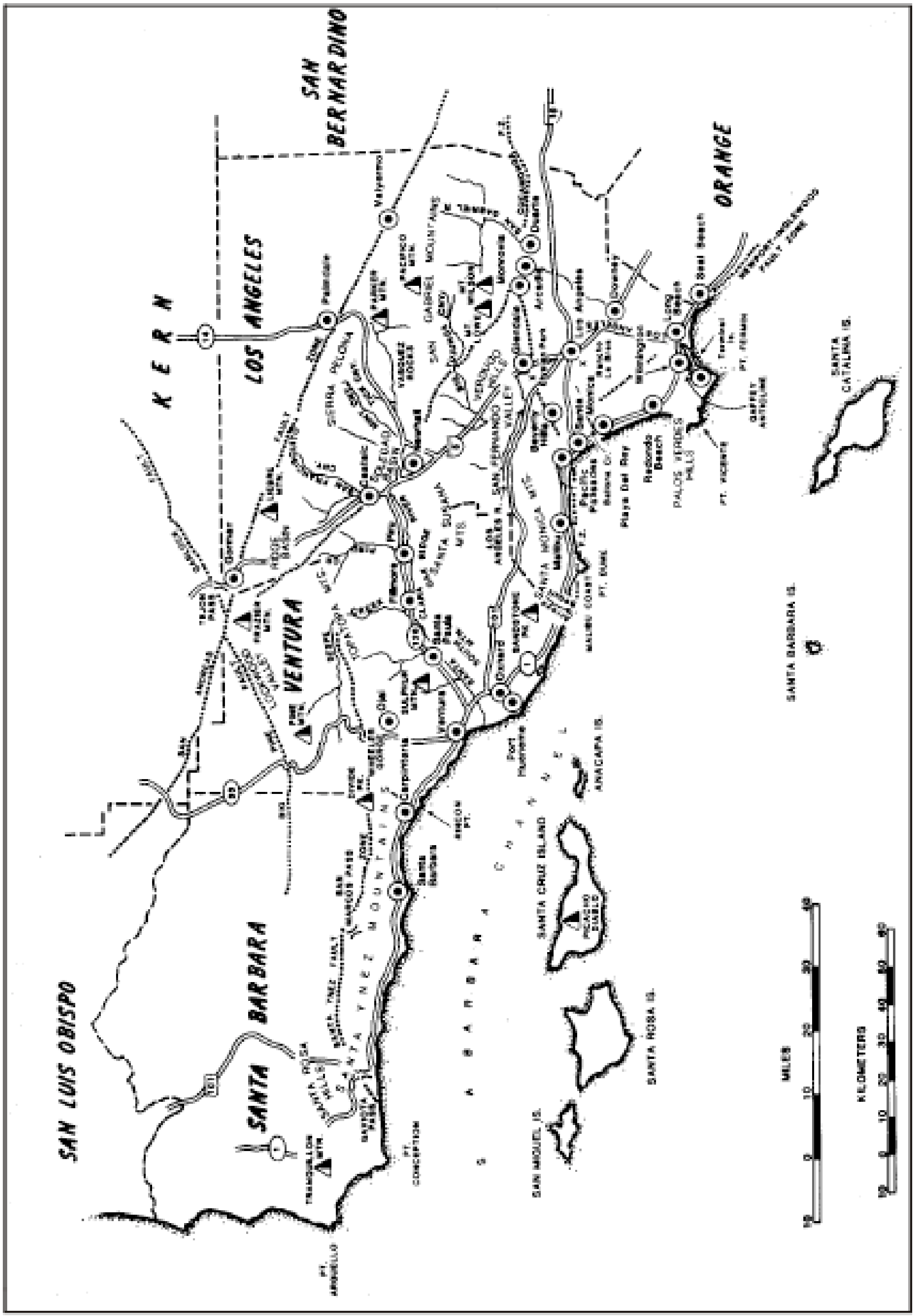

Figure 15 - Place names, western Traverse Range Province (From Norris and Webb, 1989, p. 302, Fig. 10-1a

Southwestern Block The southwestern block is the seaward part of the Los Angeles Basin. It is bounded on the east by the Newport-Inglewood zone of deformation, which can be traced from Beverly Hills to Newport Bay where it strikes offshore. This structural trend, a combination of folds and faults, is expressed as a chain of low, en echelon (overlapping staggered) anticlinal hills. The distinguishing feature of the southwestern block is its basement. Although actually exposed only in the Palos Verdes Hills, it has been encountered in numerous oil wells at depths of 1,500 to 4,250 meters (5,000- 14,000 feet) below sea level. These basement rocks belong to the Catalina schist facies of the Franciscan Formation and are chiefly green chlorite and blue glaucophane schists. They have no known base, are always in fault contact with other basement rocks, and are of undetermined age. The oldest rocks in depositional contact with them are Miocene. Based on lithologic affinities with dated Franciscan in the Coast Ranges, a late Jurassic to late Cretaceous or even younger age is probable.

The main structural features of the southwestern block are the anticlinal Palos Verdes Hills that have been raised along a steep reverse fault, several anticlinal ridges in the basement rocks over which younger sediments have been draped, and intervening broad synclines. The anticlinal structures of the younger rocks have formed important traps for petroleum and natural gas. For example, the Wilmington field is the most productive field in California and the second most productive in United States. The sedimentary blanket filling this block is quite thick, up to 6,250 meters (20,500 feet). It is all post-Oligocene and almost entirely marine.

Displacements on the Newport-Inglewood fault have both vertical (to 1,200 meters or 4,000 feet) and horizontal (at least 1,500 meters or 5,000 feet) components. The upper surfaces of the basement rocks typically have about 1,200 meters (4,000 feet) of separation across the fault zone, but the overlying sediments show less vertical separation the younger they are. Movement seems to have begun in the Miocene and is still progressing, as indicated by arching of late Pleistocene and younger strata and by recent seismicity--the Newport-Inglewood fault zone caused the 1933 Long Beach earthquake. The zone displays mainly right slip like that observed on the San Andreas, although of smaller magnitude. This movement probably produced the en echelon wrinkles that are reflected on the surface as the Baldwin, Dominguez, and Signal Hills (Figure 16). The surface expression of folding is more prominent than the expression of faulting. Several of these anticlinal hills overlie up-faulted blocks of basement rocks, however, so their origin may he more complex than first believed.

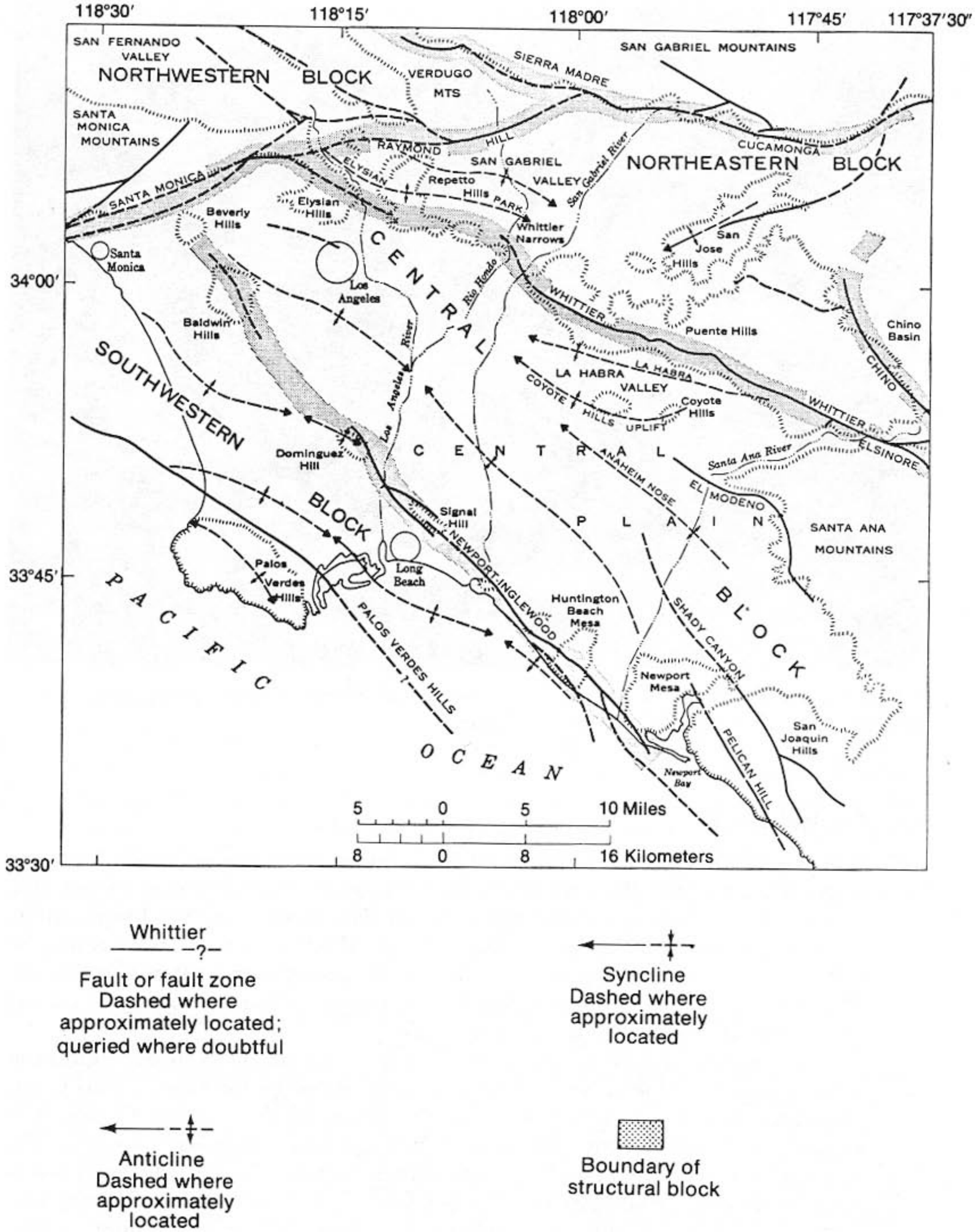

Figure 16 - Los Angeles Basin (From Norris and Webb, 1989, p. 325, Fig 10 - 13)

Central Block The central block of the Los Angeles Basin includes the low portions of the Los Angeles coastal plain from Beverly Hills southeast to central Orange County (the Downey Plain), the Coyote Hills uplift, the La Habra Valley, the San Joaquin Hills, and the Newport and Huntington Beach mesas. The Santa Ana Mountains may be included also, but they are more conventionally placed in the Peninsular Ranges. The block's main portion is occupied by the Downey Plain, a broad synclinal sag about 16 to 22 kilometers (10-14 miles) wide.

There are several folds within the coastal plain. One that lacks surface expression is the anticlinal Anaheim Nose. Another is represented by the Coyote Hills uplift, extending lion, the low hills near Sante Fe Springs southeast to the Coyote Hills proper, which stand nearly 160 meters (500 feet) above the adjacent lowland. The synclinal trough of La Habra Valley lies northeast of the Coyote Hills uplift and is bounded on the northeast by the Whittier fault zone. The Whittier fault zone also forms the eastern side of the central block from Montebello to near Corona, where it merges with the Elsinore fault. Northwest from Montebello, the presence of he Whittier fault is uncertain. Here the block's eastern boundary is marked by the Elysian and Repetto hills.

Basement rocks are not known from the deep part of the central basin, although they are encountered in some oil wells around the block's margin and are exposed in the Santa Ana Mountains. They are probably equivalent to those seen in the eastern Santa Monica Mountains. They consist of slightly metamorphosed sedimentary Jurassic rocks that have been intruded by late Cretaceous granitic rocks of the southern California batholith. No transition is discernible between the basement rocks of the central and southwestern blocks. This suggests that basement rocks of quite different origin have been brought into contact with one another, probably by appreciable right slip on the fault zones, as well as by about 90 degrees of rotation of the adjacent mountain blocks suggested by paleomagnetic studies.

Younger rocks resting on the basement are best exposed on the western slopes of the Santa Ana Mountains. At least 9,700 meters (32,000 feet) of marine and nonmarine Cretaceous to Pleistocene sedimentary rocks occur here, plus some Pliocene volcanic rocks. The older rocks in this sequence are missing from the central part of the block, although Miocene, Pliocene, and Pleistocene beds alone total more than 6,700 meters (22,000 feet) thick. Where older sediments occur, near the junction of the Rio Hondo and Los Angeles River, total thickness is 9,700 to 10,700 meters (32,000-35,000 feet). The basement surface upon which the younger rocks lie is bowed downward from the edges of the central block. Sedimentary rocks in the deepest part of the basin may lie on some sort of volcanic or oceanic crust. In any event, the basement lies about 4,000 meters (13,000 feet) below sea level at the ends and 4,250 meters (14,000 feet) along the Newport-Inglewood fault zone. In the deepest part of the Los Angeles Basin, the probable crystalline floor lies more than 9,150 meters (30,000 feet) below sea level and is known only from geophysical data.

Northeastern Block The northeastern block is situated between the Whittier fault zone and the base of the San Gabriel Mountains and is separated from the northwestern block by the Raymond Hill fault. This block is a deep synclinal basin that contains mostly marine Cenozoic sedimentary rocks, but includes some thick Miocene volcanic rocks in the east. Basement lies as much as 3,650 meters (12,000 feet) below the surface in the central part of the San Gabriel Valley, and in the eastern Puente Hills more than 6,700 meters (22,000 feet) of Cenozoic sedimentary rock covers the basement.

Northwestern Block The northwestern block embraces the eastern Santa Monica Mountains and the San Fernando Valley. It is bounded on the south side by the Santa. Monica and Raymond Hill faults, on the east and northeast by the San Gabriel Mountains, and on the west and north by ranges usually included in the Ventura Basin portion of the Transverse Ranges. The San Fernando Valley is a broad syncline with the eastern Santa Monica Mountains an adjacent anticline. No faulting of consequence separates the Santa Monica Mountains from the San Fernando Valley, but the Santa Monica block has been appreciably uplifted with respect to the other blocks of the Los Angeles Basin.

Geologic History Unraveling the geologic history of the Los Angeles Basin is a complicated process. It involves not only vertical movements of great magnitude (such as more than 6,100 meters [20,000 feet] of subsidence in the central basin since the middle Miocene), but also includes substantial strike-slip movement on the Newport- Inglewood zone and boundary faults like the Malibu Coast-Santa Monica. This latter fault zone may have undergone as much as 80 kilometers (50 miles) of left-slip since the Eocene. Further, paleomagnetic evidence indicates that some of the blocks adjacent to the Los Angeles Basin may have been rotated as much as 90 degrees in the last 30 million years or so. Our understanding remains incomplete.

At the beginning of the late Cretaceous, an extensive erosional surface developed across the older rocks and was subsequently covered by later Cretaceous marine sediments from an advancing sea. Rocks of this age are known in the Santa Ana and eastern Santa Monica mountains. A similar pattern probably applies to Paleocene and Eocene rocks, although they are buried so deeply that they have not been encountered in wells drilled in the central basin. Late Eocene, Oligocene, and early Miocene non-marine sedimentary beds occur in both the Santa Ana and Santa Monica mountains, and the presumption is that they also extend across the basin.

The great relief of the present basin floor began to evolve in early Miocene time, when sizable vertical movements began to exert control on the pattern of later deposition. About this time, approximately 22 million years ago, the area was stretched and broken as the result of the earlier subduction of the East Pacific Rise beneath this part of the continent. Between about 15 and 13 million years ago, volcanic activity was widespread in the basin and rotation of the crustal blocks is believed to have begun. During the late Miocene, the sea advanced over the Los Angeles Basin from south to north, eventually covering basement highs at Palos Verdes and other parts of the southwestern block. By the close of the Miocene, the sea had reached the base of the San Gabriel Mountains and flooded most of the Los Angeles Basin, although a shoal existed at the site of today's Anaheim Nose. There is evidence that the Los Angeles and Ventura basins were connected at this time and that marine conditions prevailed over the intervening area. It is important to remember that modern ranges and valleys did not exist until near the end of the Pliocene. Furthermore, although the great thickening of marine Miocene in the Central Los Angeles Basin is reflected in well records, drills have not yet penetrated to the probable crystalline rocks below the sedimentary column, which is often more than 6,100 meters (20,000 feet) below the surface.

During the Pliocene, the rate of sinking accelerated in the central basin, and some sediments were deposited in as much as 1,800 meters (6,000 feet) of water. Concurrently, the basins margins were undergoing marked uplift, and rocks from the surface and from oil wells show many unconformities that record continuing tectonism. The central basin continued to receive large volumes of sediment from the Northeast, and the basin's sea became steadily shallower. By the close of the epoch, more than 3,000 meters (10,000 feet) of deposits had been laid down. All unconformity at the top of the Pliocene section shows that deposition was interrupted by deformation about this time. As the Pliocene closed, land areas included an island formed by the Palos Verdes Hills, much of the Santa Ana and Santa Monica mountains, the Puente Hills, and probably some small islands along the Inglewood fault zone.

In early Pleistocene time, the Palos Verdes Hills sank below sea level, the central basin continued to receive marine deposits as did the San Joaquin Hills and the San Gabriel Valley, and the San Gabriel Mountains and the Puente-Repetto Hills were rising. The Santa Monica Mountains seem to have persisted as a lowland. Deposition began to outpace subsidence, and by middle Pleistocene the shoreline probably lay along the southern margin of the Santa Monica chain and along the Whittier fault zone. Exceptions were two embayments: one at Whittier and one at the base of the Santa Ana Mountains and north of the San Joaquin Hills. Near the center of the basin are more than 900 meters (3,000 feet) of Pleistocene land-laid beds, some deposited in near-shore lagoonal environments along a low-lying coast. To accommodate such a thickness with today's low elevations, there must have been substantial subsidence until the present. The region experienced its last deformational episode, the Coast Range orogeny, by the middle Pleistocene. This was expressed in the central basin by more subsidence, but in the surrounding areas by considerable uplift.

By late Pleistocene, the Palos Verdes Hills began to rise along the Palos Verdes fault, producing marine terraces. Lowering of sea level, which was partially caused by continental glaciation, made the entire coastal plain emerge and allowed streams to cut channels as much as 76 meters

(250 feet) deep at the present shore (The land area at that time was thus more extensive than today's). The hills along the Newport-Inglewood uplift were developed, and some of the hills in the eastern basin were uplifted. For example, the San Joaquin Hills rose as much as 300 meters (1,000 feet) during the late Pleistocene and Holocene. The Whittier fault zone is thought to have about 4.8 kilometers (3 miles) of right separation; 1.6 kilometers (1 mile) of which appears to be late Pleistocene and Holocene. This last orogenic episode is apparently continuing, as indicated by historical earthquakes and folding of young deposits in such areas as Signal Hill, the central basin, and the Coyote Hills. There seems to be good evidence too that the Gaffey anticline, north of San Pedro, is still rising.

The thick marine sediments of the Los Angeles Basin are richly fossiliferous, especially within Pleistocene sections. The San Pedro and Timms Point beds from Palos Verdes Hills which we visited earlier in the trip are particularly rich. In one study, a single exposure of a 30-centimeter (1-foot) layer of Palos Verdes sand near Playa del Rey yielded more than a million shells. Nonmarine deposits are less fossiliferous, but do contain vertebrate remains. Particularly remarkable is the array of animal remains recovered from the tar pits at Rancho Brea.

The Los Angeles Basin is one of the world's most prolific petroleum-producing areas and has a long history of petroleum exploitation. In 1543, the Portuguese explorer Juan Cabrillo reported that the local Indians used tar and asphaltum collected from the La Brea tar pits for both medicinal and sealing purposes. Oil was collected from seeps and hand-dug pits beginning in 1855, oil was being shipped from Wilmington by 1873, and over 1000 holes had been drilled by 1900. It is one of the most prolific hydrocarbon-yielding basins in the world for its size. Its cumulative production of nearly 8,000,000, 000 barrels of oil and over 7,000,000,000 MCF of gas, when measured as volume/acre/foot, exceeds that of the Persian Gulf. The 62 Los Angeles basin oil fields are structurally controlled along four major zones of wrench faults: Palos Verdos, Newport-Inglewood, Whittier-Elsinore, and Malibu Coast- Santa Monica-Raymond Hill. The major reservoir rocks accumulated as sediment-gravity flow deposits in rift and wrench-fault created basins during Miocene and Pliocene time. These reservoir rocks were deposited within and above organic-rich, silled basin deposits of Miocene Monterey Formation source rocks. The oil-fields of the basin continue to be important producers, with nearly 140 million metric tons (1,000 million barrels) of reserves still in the ground.

San Fernando Valley West of Pasadena we proceed into the San Fernando Valley (Figure 22). The geologic history of the valley was described in detail in the previous section. As we cross valley we pass many famous cities including Burbank, Glendale, North Hollywood, Van Nuys, Reseda, and Tarzana. Fifty years ago this valley was an agricultural paradise filled with orange groves but the spread of suburbia and industry from the expanding Los Angeles has resulted in the valley becoming crowded and very smoggy. Shortly after passing the city of Thousand Oaks in the western part of the valley we will be passing through a belt of Miocene volcanic rocks of submarine origin erupted within the Los Angeles basin. Shortly thereafter we will be entering the Ventura basin.

Ventura Basin The Ventura basin is a highly folded synclinorium that contains a maximum of about 50,000 feet of Tertiary and Quaternary strata, and possibly as much as 8,000 feet of Cretaceous strata (Figure 17). The synclinorium is broken by a number of large thrusts or reverse faults, some of which dip south and others north. Except for its northern margin, the western half of the basin is submerged beneath the Santa Barbara Channel. Most of the larger interior valleys, such as Ojai Valley, Simi Valley, and the Santa Clara River Valley above Saticoy, are synclinal, and the intra-basin ranges or hills or mountains, such as Red Mountain, South Mountain-Oak Ridge, and the Camarillo-Las Posas Hills, are anticlinal. The most extensive lowland area, the Oxnard Plain, is gently folded but considerably faulted beneath its thick alluvial cover.

The central part of the Ventura basin has been subjected to direct north-south compression, resulting in overturning of beds and the development of thrusts or reverse faults on one or both flanks of many of the anticlinal ranges. The Santa Clara and Ojai Valleys are deep fan synclines with both limbs overturned, and the limbs are broken by thrusts that represent movements toward the valleys from both sides. Although Oak Ridge and the Santa Susana Mountains are parts of the same anticlinal uplift, Oak Ridge has been thrust northward along the southdipping Oak Ridge fault, whereas the Santa Susana Mountains farther east have been thrust southward along the Santa Susana thrust, which dips northward at low angles.

The Ventura Avenue oil field and several other good fields farther west are located on the 16-mile long Ventura anticline, the axis of which lies 3 miles north of Ventura. This anticline has fairly regular limbs that dip at angles of 40 to 50 degrees, but it is severely broken by thrusting, toward both the north and the south, in the subsurface Pliocene beds. These thrusts die out surfaceward into zones of steep dips.

The northwest margin of the Ventura basin includes the southern foothills of the Santa Ynez Mountains and the narrow coastal plain and hills around Santa Barbara, Goleta, and Carpinteria. The foothill belt is a south-dipping homocline that is interrupted by a few anticlines and synclines and is cut by many nearly vertical faults. These intersecting faults trend northeast `and northwest, have had oblique-slip movements, and commonly show displacements of a few hundred to a few thousand feet. Santa Barbara and Goleta lie in alluviated valleys that are synclinal grabens, and the Carpinteria alluvial plain is structurally a syncline in Oligocene to lower Pleistocene strata that is cut by faults south of the axis.

The east end of the Ventura basin is a series of closely-spaced anticlines and synclines whose moderately to steeply dipping flanks are broken by the Holser reverse fault. They are cut off diagonally by the San Gabriel fault. Oil fields, surprisingly numerous for such a small area, are present in the vicinity of Piru, Newhall, Castaic, and Saugus. Most of these are on domal anticlines or faulted anticlines, but some represent stratigraphic traps on the flanks or plunging noses of anticlines.

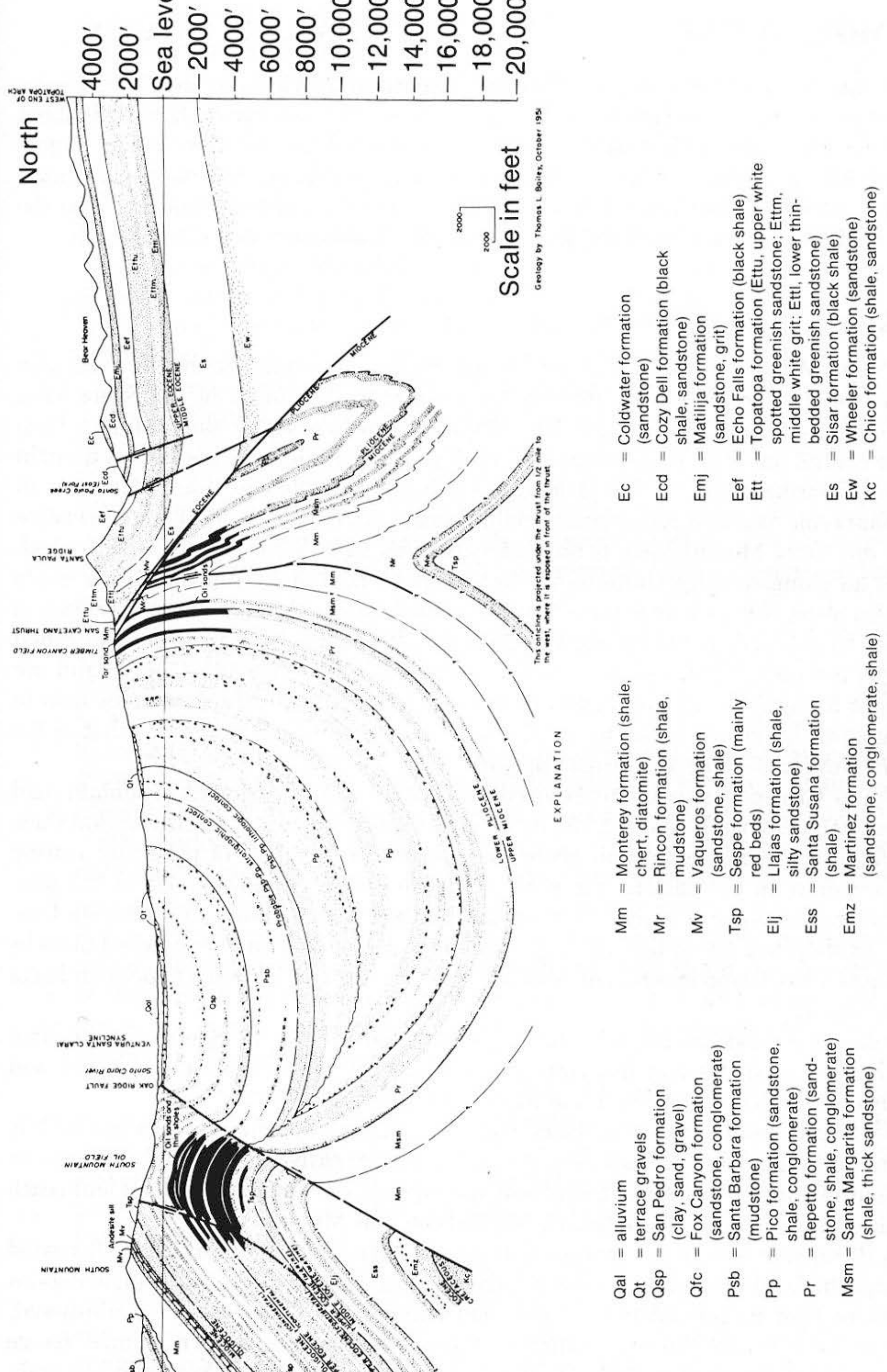

Figure 17 - Structure section across part of Ventura Basin, showing extraordinary thicknesses of the strata, particularly the Pliocene. Rock types and structures are typical of much of the western Traverse Range sedimentary pattern (From Norris and Webb, 1989, p. 313, fig. 10-7)

Day 3: Sunday, 20 August 2006

Summary: Leave Oso Campground
Ventura Avenue Anticline and Rincon Oil Field
Wheeler Gorge
Return to Oso Campground

Directions from Oso Campground to Ventura Avenue Anticline:

Start going west on Paradise Rd/NF-5N18 toward El Gaucho LN for 4.9 miles
Turn left onto CA-154/San Marcos Pass Rd for 10.5 miles
Merge onto US-101 via the ramp on the left 31 miles
Merge onto CA-33 toward Ojai for 7 miles
End at anticline that crosses road.

Figure 18 Oso to Anticline

INTRODUCTION TO THE SANTA BARBARA-VENTURA BASIN

The Tertiary sequence of the Santa Barbara-Ventura basin reflects a highly structured depositional environment that is the result of a tectonically dynamic setting of the southern California continental margin. The sequence, which locally exceeds 10 km (30,000 feet), is characterized by an alternation of mud- and sandstone units of submarine fans and interlobe deposits, gradually filling the east-west trending Paleogene basin with continent derived clastic sediment. By late Eocene time in the Ventura basin and early Oligocene time in the central Santa Ynez Mountains, marine deposition was followed by non-marine alluvial deposition of the Sespe Formation, whereas shallow marine deposition of the Gaviota Formation continued in the western Santa Ynez Mountains. Tectonic reorganization of the Pacific-North American plate boundary initiated a phase of renewed subsidence and deposition of marine strata during late Oligocene-early Miocene time. Separation of sub-basins from continental clastic input by intrabasinal sills provided isolated conditions for the deposition of organic-rich siliceous hemipelagic mudstone of the Monterey Formation between about 18 and 6 Ma. An increase in terrigenous detrital input marks the onset of Sisquoc deposition in late Miocene time, followed by the locally more than 3 km (10,000 feet) thick turbidite sequence of the Plio-Pleistocene Pico Formation. Uplift and exhumation of the Santa Ynez Mountains at about 1 Ma is marked by the deposition of coarse clastic units in the Ventura, Carpinteria, and Santa Barbara coastal area.

West of Ventura, Hwy. 101 follows a 1-5 Ka old marine terrace, cutting obliquely across the E-W trending anticlinorium of the Ventura Avenue oil field and its western continuation, the San Miguelito, Rincon, and Dos Cuadras oil fields. The combined production of these oil fields exceeds 1.5 billion barrels (Conservation Committee of California Oil and Gas Producers, 1991), produced from the turbidite sands of the Plio/Pleistocene Pico Formation. Radiometric ages from interbedded ash beds and amino-acid ages from uplifted marine terraces indicate that folding of the Ventura Avenue anticline started about 0.2 Ma ago, with initial uplift rates as high as 15 mm/yr and an average horizontal shortening rate of about 2 cm/yr.

The high rate of tectonic shortening across the Rincon-Ventura anticlinorium is presumably responsible for tilted oil-water contacts in the reservoirs and elevated pore fluid pressure, reaching pore pressure gradients of up to 0.9 psi/ft (McCulloh, 1969). Elevated pore fluid pressure in the offshore Dos Cuadras field contributed to the blowout of platform A in 1969.

Striking evidence for rapid uplift is provided by a 45 Ka old marine terrace deposit, well exposed along Hwy. 101 between Sea Cliff and Punta Gorda. The terrace deposit lies in angular unconformity on steeply dipping Pico Formation and sits about 120 ft above sea level at Punta Gorda, rising to an elevation of about 800 ft near Ventura. The artificial Rincon island marks the crest of the Ventura Avenue-San Miguelito-Rincon-Dos Cuadras anticlinorium. The north limb of this anticlinorium is truncated by the east-west trending Red Mountain fault which crosses Hwy. 101 between the little town of La Conchita and Rincon Point. The north-dipping Red Mountain fault thrusts the Paleogene sequence over the Plio-Pleistocene Pico Formation. A subparallel fault, the Carpinteria fault, forms a north-facing fault scarp south of Hwy. 101 just east of Carpinteria, apparently offsetting the 45 Ka old marine terrace (Eichhubl and Behl, 1998).

3-1 Ventura Avenue Anticline-

Sharp and Glazner, Geology Underfoot in Southern California, 1993, pp 37-46.

Southern California has so many geologic faults that we tend to forget the crust can also deform by simple tilting, warping, or folding. Folds form best in well-layered, weak sedimentary rocks, of which Ventura County has a great thickness. About 2.5 miles north of Ventura, along California 33, is a truly beautiful, large anticline that extends east and west for fifteen miles. Dubbed the Ventura anticline, it is the site of a highly productive oil field, one of California's largest with a cumulative production approaching one billion barrels of oil and several trillion cubic feet of gas. Oil, gas, and water can collect in anticlines, if those fluids, and porous beds to hold them, exist within the rocks. As these fluids rise to the crest of an anticline, they arrange themselves in order of density: gas on top, then oil, then water—all trapped beneath impervious layers of rock. Many anticlines are shaped like an inverted canoe with the central axis plunging down at both ends. That form will securely trap oil and gas. A few anticlines are cracked so they leak oil and gas; that is the case of the natural oil pollution in the Santa Barbara Channel and the tar blobs along the shoreline near Goleta and Carpenteria.

The Ventura anticline is very young (<200,000 years). This is amazing; in most regions, folds are tens or hundreds of millions of years old. But terraces along the Ventura River that are only 16,000 years old tilt where they cross the fold. That suggests the fold may still be rising. The ages of the tilted rocks and the amounts of their tilting show that the Ventura anticline has risen at an average rate of nearly 0.6 inch per year for the last 200,000 years, rising roughly two miles in that time.

Marine terraces on the south limb of the anticline are old wave-cut benches that originally sloped about one degree seaward, the usual slope of such terraces. They are two to three times older than the stream terraces on the north flank of the fold. The marine terraces now slope about ten degrees seaward and stand hundreds of feet above sea level. Part of that rise is due to the folding, the rest to movement along faults.

The thick section of folded sedimentary beds in the Ventura anticline contains several thin layers of volcanic ash. The well-known Lava Creek ash layer lies within steeply tilted beds on the south limb of Ventura anticline. It exactly matches ash erupted during a gigantic volcanic explosion in Yellowstone Park about 600,000 years ago.

3 - 2 Rincon Oil Field, Seacliff –

Rincon Oil Field is located along Hwy CA-101.

The formation exposed in the steep sea cliffs is mainly of upper Pliocene (Pico) age. It is part of the south flank of the Ventura anticline, the crest of which finally passes into the ocean at Seacliff. It is illegal to stop along the freeway in California. We will exit into a parking lot along the freeway. The Rincon oil field is located on this fold and is partly on shore and partly in the sea. The fold here is asymmetric and somewhat complicated by reverse faulting on its north flank, owing to the influence of the Red Mountain fault, which is less than a mile (1.6 kilometres) from the coast at this point. Here, as in the Ventura Avenue field, the oil is derived from the lower Pliocene (Repetto formation), though at shallower depths. The top of the main zone lies at a depth of about 3,000 feet (914 meters) on the crest of the anticline. At first development took place on the shore, but in 1930, after a geologic survey of the outcrops on the floor of the ocean, a long pier was built and wells drilled over the water. The depth of the water here is not more than 35 feet (10.7 meters).

3 - 3 Wheeler Gorge –

Wheeler Gorge is located along Hwy CA-33

Wheeler Gorge is the narrow part of the North Fork of Matilija Canyon that cuts through the Santa Ynez- Topatopa Range of the western Transverse Ranges, in Ventura County. This canyon is within the Los Padres National Forest. At Wheeler Gorge the canyon is so narrow and steep-walled that three short tunnels had to be cut and three bridges built for California Hwy 33. Dibblee (1987) provided a very comprehensive overview of this locality. The following description is from that field guide.

Significance of the Site

The western Transverse Ranges south of the Santa Ynez Fault in western Ventura County expose the entire Upper Cretaceous - Cenozoic sedimentary series of the north flank of the Ventura Basin, in a generally simple structure (Figure 19). The total aggregate thickness of this varied sedimentary series, including one of the thickest Cenozoic sequences in the world, is about 42,000 ft (12,800 m). The lower part of this enormously thick and nearly continuously deposited up ended sedimentary series is a highly indurated marine clastic turbidite series of Eocene-Upper Cretaceous age that forms the east-striking Santa Ynez-Topatopa Range (Dibblee, 1982; Fig. 1). The Upper Cretaceous part of this series is superbly exposed along Wheeler Gorge. This part, exposed along the creek adjacent to Calilornia 33, has been much visited and studied by geologists. Their interpretations of the source and modes of transportation and deposition of these upper Cretaceous sedimentary rocks provide an important basis of the late Cretaceous paleogeography of this region.

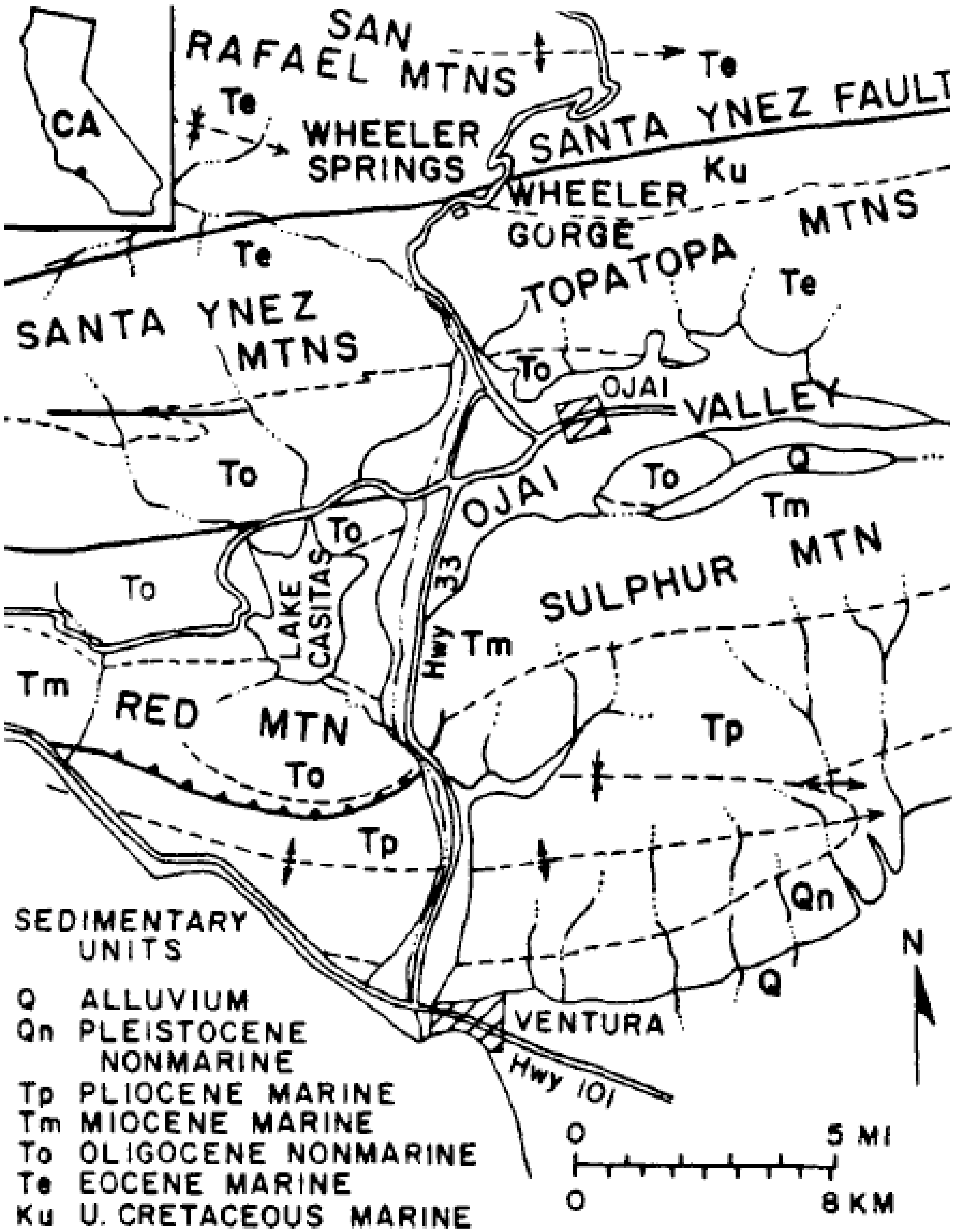

Figure 19 - Index map showing location of Wheeler Gorge with respect to highways, towns, physiographic, and geologic festures (From Dibblee, 1987, p. 227, fig. 1)

Regional Geomorphology and Geology

Wheeler Gorge is a deeply incised canyon eroded by the North Fork of Matilija Creek through the west end of the Topatopa Mountains. This fork joins Matilja Creek (main fork) below Wheeler Springs and drains southward via the Ventura River wash to the coast. The Santa Ynez-Topatopa mountain uplift was elevated on the south side of the 90mi (145 km) long Santa Ynez fault as a southward-tilted strip (Dibblee, 1982). Both forks of Matilija Creek, which drain the somewhat higher eastern San Rafael Mountains to the north, are antecedent to this uplift. Wheeler Gorge is the most narrow part of North Fork Canyon.

The stratigraphy of the rock units of the Upper Cretaceous-Eocene marine turbidite sequence exposed along the North Fork of Matilija Canyon, including Wheeler Gorge, south of the Santa Ynez fault are shown on Figures 20 and 21, and the sedimentology of the Upper Cretaceous section exposed along Wheeler Gorge site is briefly described.

Unnamed Cretaceous marine strata. An unnamed marine clastic turbidite formation of Late Cretaceous age is the oldest formation exposed in the Topetopa Mountains in which about 4,600 it (1,400 m) crops out. An unknown amount of additional lower strata of this formation are concealed. At Wheeler Gorge the uppermost 1,560 ft (475 m) of this formation crops out south of theSanta Ynez fault (Figs. 20 and 21).

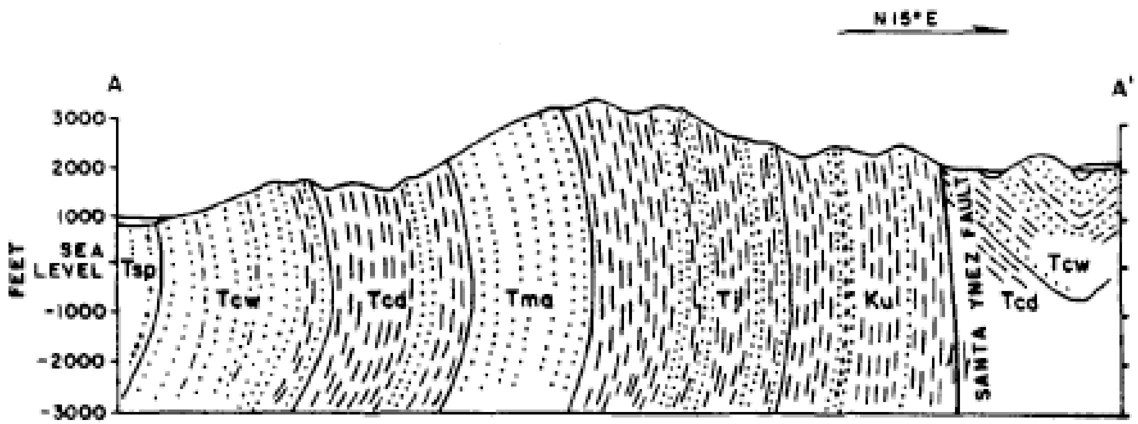

Figure 20a - Cross section through the Topatopa Mountains near Wheeler Gorge, Line of section shown on Figure 28 (From Dibblee, 1987, p. 229, fig. 3)

At Wheeler Gorge this unnamed formation yielded a few marine fossils, including a large ammonite (Desmoceras sp.) and pelecypods (Inoceramus subundatus Meek) diagnostic of late Campanian-Maestrichtian age (Rust, 1966) or very late Cretaceous. The Upper Cretaceous Formation was deposited essentially by turbidity currents probably on a shelf or fore-arc trough under a moderately deep sea of the eastern Pacific and derived from a continental basement terrane to the east.

Sedimentology of the Upper Cretaceous strata exposed at Wheeler Gorge

Along Wheeler Gorge most of the upper 1,560 ft (475 m) of the Upper Cretaceous Formation of the Topatopa Mountains is superbly exposed between the Santa Ynez fault and the overlying Juncal Formation, This section is composed of two shale sequences separated by a conglomerate member.

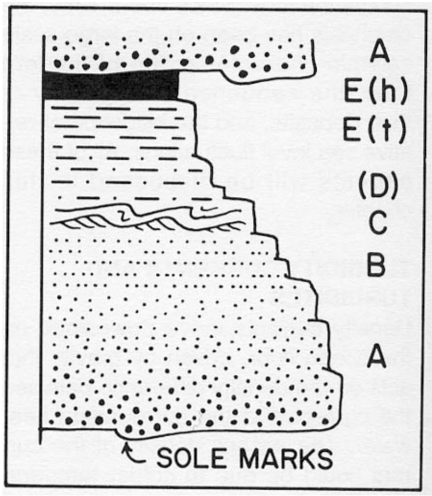

Figure 20b

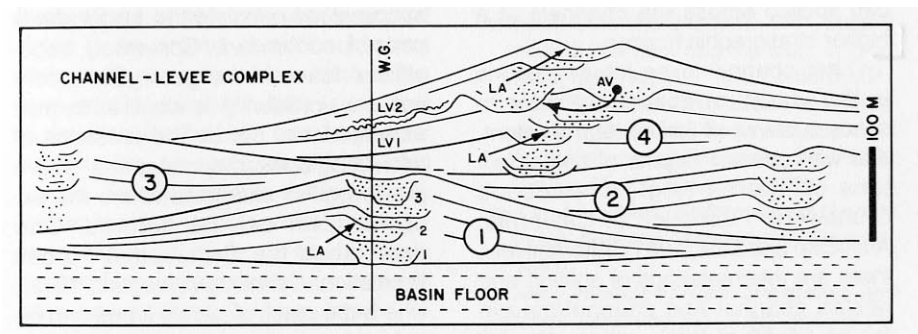

Figure 20c

Figure 20b: Bouma (1962) sequence; division A is structureless, B is parallel-laminated sand, C is rippled and/or convoluted, D is hard to see in weathered or tectonized outcrops, and consists of parallel-laminated silt and mud. The politic interval E is partly of turbidite origin (t) and partly hemipelagic (h).

Figure 20c: Interpretation of the Wheeler Gorge section. WG indicates the vertical profile of Wheeler Gorge. Circled numbers suggest the order of channel-levee system development, and uncircled numbers 1-3 indicate the stacked channels at Wheeler Gorge. LA = lateral accretion and AV = avulsion. Thus the thinning –and fining-upward levee sequences are interpreted to reflect lateral migration of the channel away from Wheeler Gorge. Channel-levee systems 2 and 3 are necessary in order that the levee sequences of system 4 can build on top of the stacked channel deposits of system 1. No horizontal scal is implied; systems 2 and 3 could equally well be on opposite sides of system 1, and system 4 could be a mirror image of what is shown, on the left of Wheeler Gorge. Stipple indicates sandstone and comglomerate within channels.

Figure 20d:Stratigraphic section through Cretaceous rocks at Wheeler Gorge, California. Base of section is faulted (Satna Ynez Fauld), and overlain by a small slump. The lower part of the section consists of 250 m of mudstones. CM – channel margin deposits, LV = levee. Arrows indicate thinning- and fining-upward facies successions.

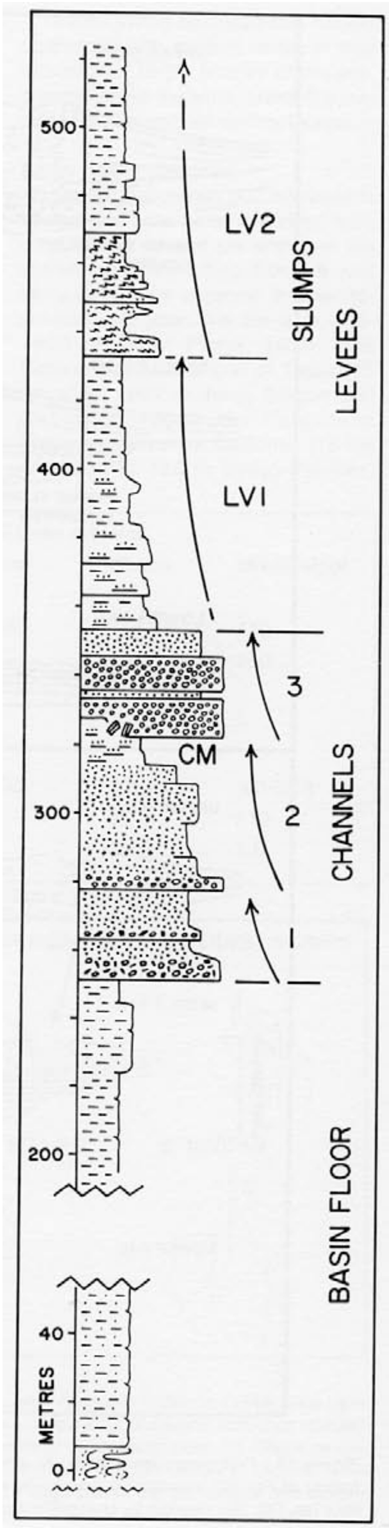

Figure 20d

The sedimentology of these three exposed units and their paleogographic implications are described in detail by Rust (1966), fisher and Mattison (1968), and Walker (1975b, 1985), from which the following paragraphs are summarized.

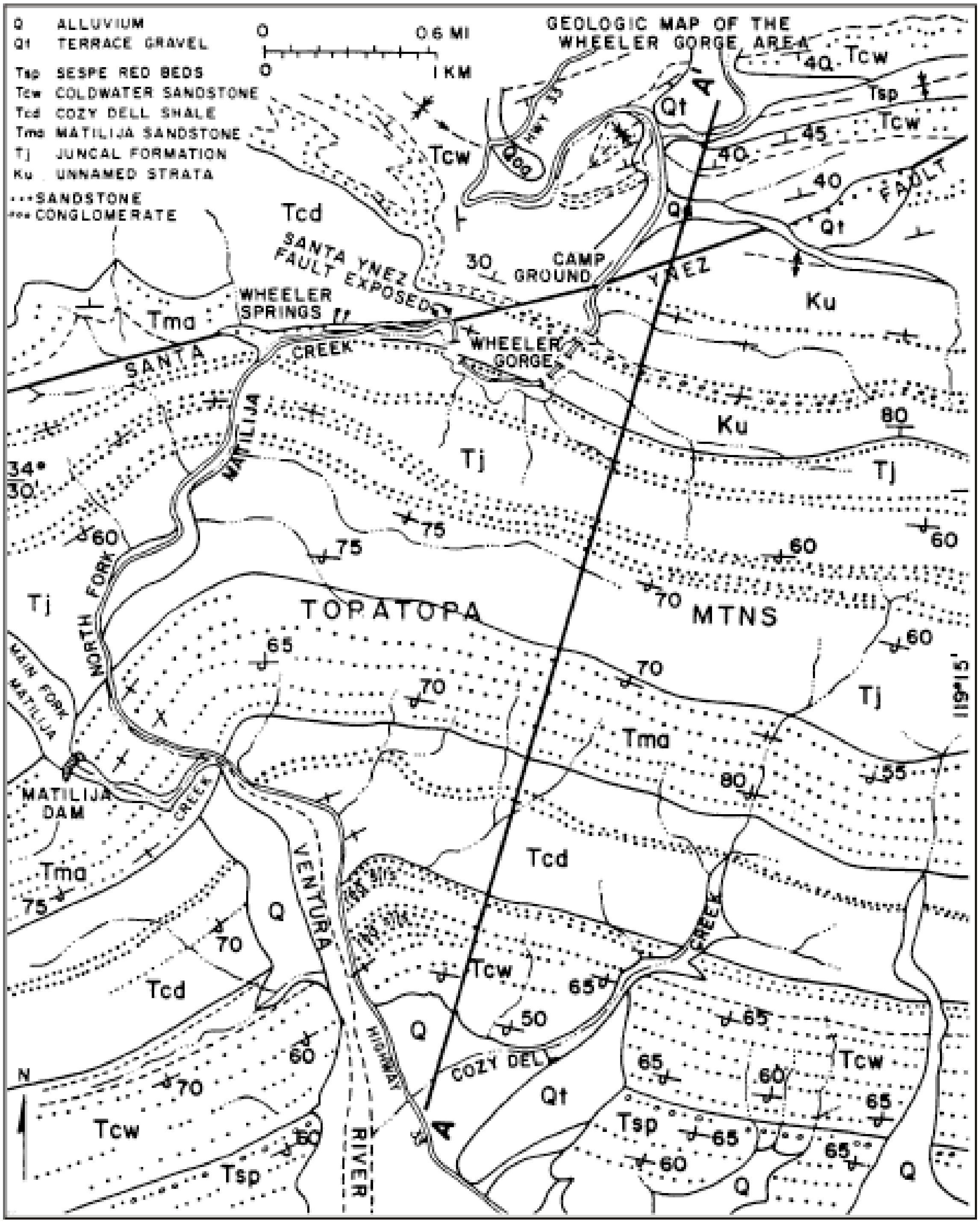

Figure 21 - Geologic Map of the Wheeler Groge Area (From Dibblee, 1987, p. 228. Fig. 2)

The shale or mudstone sequence exposed below (north of) the conglomerate member at Wheeler Gorge northward to the Santa Ynez is about 770 ft (235 m) thick. Walker (1985) divided this shale sequence into eight turbidite units. Each of these is composed of gray to black micaceous shaly mudstone that contains many very thin intercalations averaging 1 cm thick of light gray very fine grained sandstone that grades upward into siltstone. This sequence differs from that above the conglomerate member in that the sedimentologic features generally present in the upper shale sequence are absent; the sandstone intercalations are thinner and more sharply defined; and only the lowest exposed turbidite unit contains slumped beds.

Walker (1985) interprets the lower shale or mudstone sequence to have been deposited quietly on a generally smooth submarine basin-plain floor, probably far from any submarine channels.

The conglomerate member at Wheeler gorge is about 350 ft (107 m) thick, and ranges from coarse cobble conglomerate to massive light gray arkosic sandstone and minor gray mudstone. The basal contact with underlying shale sequence, well exposed at the north end of the north tunnel, is sharp and contains numerous flute-casts. Ripped-up angular mudstone fragments occur locally near the base and elsewhere in this conglomerate member. Walker (1985) divides the conglomerate of this exposure into three turbidite sequences, each of which grades upward from coarse, massive cobble conglomerate at the base through massive sandstone into fine-grained sandstone and interbedded mudstone at the top. The conglomerate is composed of smooth rounded pebbles and cobbles from 0.5 to 15 in (1 to 40 cm) in diameter in a matrix of arkosic sandstone.

Most are composed of gray quartzite, gray andesitic to dacitic porphyries, and hard granitic rocks, mostly quartz monzonite-granodiorite. Uncommon to rare clasts are of white quartz, gneiss, schist, gabbro, anorthosite, syenite, mylonite, gray chert, hard sandstone, and argillite (Rust, 1966, p. 1395).

The conglomerate member appears to be a submarine fan, deposited in a channel partly cut into the underlying shale-mudstone. The hard, durable rock types of the majority of clasts and their roundness indicates transport over great distances, possibly on land, or aggradation by wave action near shore, or both. The occurrence of this conglomerate in deep marine mudstone and its eastward pinch-out (east of Fig. 19) indicate it is not part of a deltaic fan. This condition suggests that this conglomerate may have been redeposited from a shallow near-shore deltaic fan, possibly by gravity, into its deep-sea environment, as suggested by Rust (1966), through deep-sea channels. Most paleocurrents and flute casts at Wheeler Gorge suggest the conglomerate was deposited under currents that flowed generally westward: the clast types indicate a continental basement source terrane to the east. The nearest sources of the quartzite and porphyry clasts are in the Mojave Desert region and the San Bernardino Mountains far to the east. Both of these areas are east of the San Andreas fault and may have been shifted by right slip on that fault farther southeast with respect to the conglomerate than they were in late Cretaceous time. The rare clasts of gneiss, gabbro, anorthosite?, syenite, and mylonite reported by Rust (1966) were probably derived from the rocks now exposed in the San Gabriel Mountains east of the San Gabriel fault. Right slip on the San Gabriel fault since late Cretaceous time has probably shifted the San Gabriel Mountains farther southeast with respect to the Wheeler Gorge area.

The shale or mudstone sequence exposed at Wheeler Gorge from above (south of the conglomerate member to the base of the overlying Eocene Juncal Formation is about 760 ft (232 m) thick, of which about 520 ft (158 m) is exposed. It is composed of gray to black micaceous mudstone containing many thin intercalations of light gray fining-upward siltstone/very fine-grained sandstone. Walker (1985, p. 281) divides this sequence into two megasequences, each with a gradual upward decrease in siltstone /sandstone intercalations; he divides the lower one, 262 ft (80 m) thick, into eight turbidite units, each thinning and fining upward. Most of these contain thin Bouma - A, AC, and AB sandstone types, with ripped-up mud clasts in some A-types, convoluted laminae, scouring bedlenticularity, and multiple slumped beds. From these sedimentology features, Nelson and others (1977) and Walker (1985) interpret the turbidite units as submarine channel-margin or levee deposits.

Return to Oso Campground via CA-101.

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Material Safety Data Sheet Number: D1288. http://www.jtbaker.com/msds/d1288.htm

Day 4: Monday, 21 August 2006

Summary: Leave Oso Campground
San Marcos Pass
Bath House Beach and Santa Barbara Harbour
Goleta Beach
Return to Oso Campground

Directions from Oso Campground to San Marcos Pass:

Start out going West on Paradise rd / NF-5N18 toward Fremont Tract 3.0 miles. Turn LEFT onto CA-154 / SAN MARCOS PASS RD. 2.7 mile

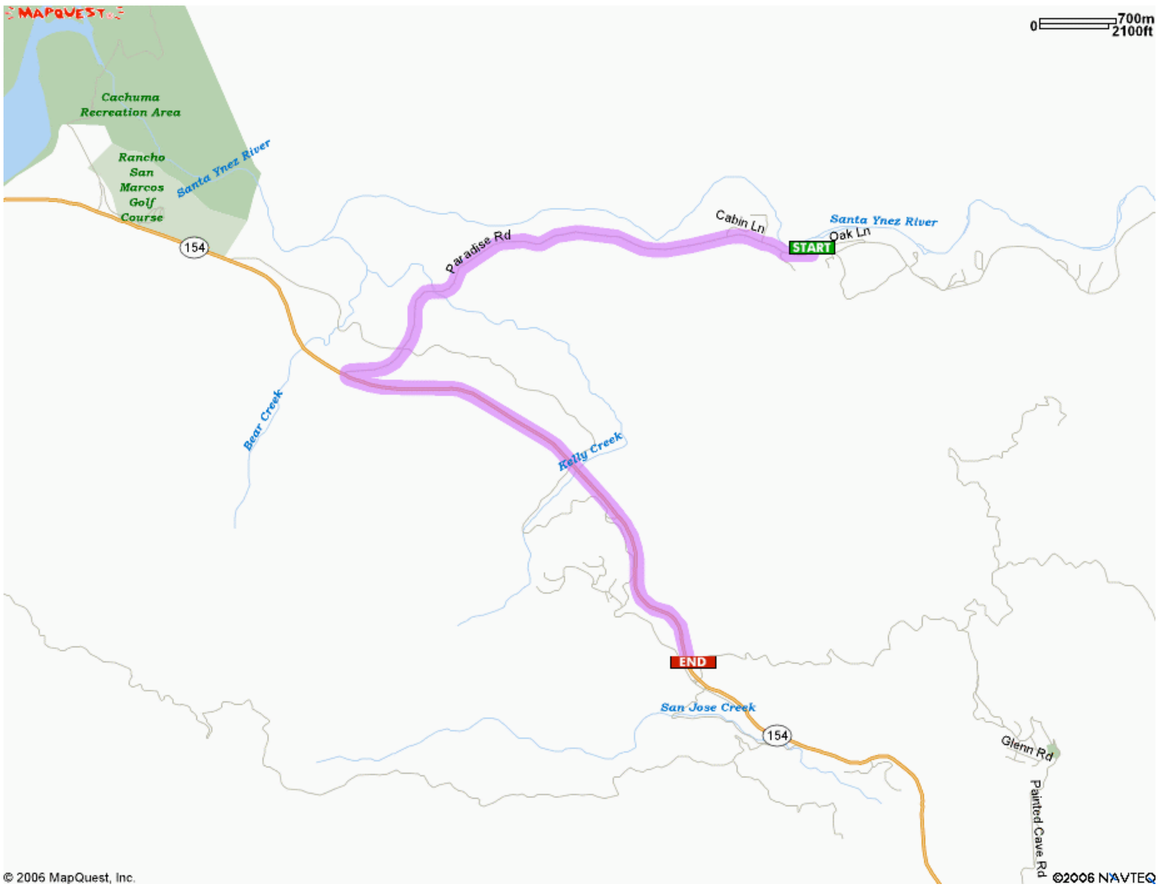

Figure 22: Oso Campground to San Marcos Pass

4-1 San Marcos Pass

The Transverse Ranges, of which the Santa Ynez Mountains form the most westerly part, is one of the few ranges in the United States that runs in an east-west direction. Forming a continuous crest from Point Arguello to Ojai, a distance of 70 miles, the Santa Ynez Mountains are tilted steeply to the south at an angle of nearly 50 degrees. From Point Arguello to Gaviota Pass, the range is generally less than 2,000' high. East of Gaviota, however, the mountains gain height rapidly, reaching 4298' at Santa Ynez Peak, before dropping gradually to San Marcos Pass which has an elevation of 2250'. San Marcos Pass occupies a low saddle formed by a synclinal (V-shaped) fold that crosses the main axis of the range diagonally. East of San Marcos Pass, the mountains rise once again, averaging 3,500' behind Santa Barbara, with La Cumbre Peak measuring 3985' in height. The range reaches its apex at 4,690' Divide Peak, near the Santa Barbara-Ventura county line.

Red-colored rocks are exposed along the base of the Santa Ynez Mountains, especially visible on San Marcos Pass Road several miles above Cathedral Oaks. These are the Sespe “red beds”, a series of rock layers composed of shale, sandstone, and a mixture of pebbles and larger cobbles called conglomerate. The reddish color is the result of iron oxides within the shales and sandstones, a vivid celebration of Santa Barbara’s rise from its primeval depths. It also leads geologists to conclude that this also was a period of tropical or subtropical climate, since red soils similar to these are being formed in the Tropics today.

<u>Photo copy from Brian</u>
The oldest rocks in the San Marcos Pass Area are parts of the Franciscan Formation. The Franciscan consists of a highly mixed assemblage of deep water sedimentary rocks including cherts and graywackes, plus altered basalt and serpentinite. The Franciscan is thought to be derived from oceanic crust, and is Jurassic to Cretaceous in age.

Middle and upper Eocene rocks are all marine and are well represented in the area, attaining a thickness of over 3,000 meters. Nearshore conditions are indicated in this time by oyster beds and other marine fossils, and occasional streaks of red shale, presumably washed into the shallow marine waters from nearby land.

In Oligocene time, the sea had withdrawn from the San Marcos Pass area, leaving a broad coastal plain on which were deposited the conglomerates, sandstones, and silts of the Sespe Formation. The Sespe is coarsest near its base, where conglomerates are interbedded with sandstones. These beds represent a fluvial or deltaic depositional environment. Several lines of evidence suggest that the Sespe is a non-marine unit. Its red color comes from the oxidation of iron within the sediment, which generally takes place in terrestrial environments where sediments are exposed to alternating wet and dry seasons. The arrangement of sedimentary structures in the Sespe is best interpreted as representing a series of non-marine environments. Finally, marine fossils, such as those seen in the Coldwater, are conspicuously lacking in the Sespe.
In Pleistocene time, the Coast Range orogeny elevated the Sa Marcos Pass area. Through a series of faulting and folding events, the formations described above were arranged into their current homoclinal geometry. They have since been eroded, forming the alluvium at the top of the section.

<u>4 - 2 Bathhouse Beach and Santa Barbara Harbour –</u>

Directions from San Marcos Pass to Bathhouse Beach
Go SE on N San Marcos Rd 0.3 miles
Turn right to stay on N San Marcos Rd 0.5 miles
Turn right onto Cathedral Oaks Rd 0.7 miles
Turn left onto N Patterson Rd
Merge onto US-101 N 3 miles
Take Glen NE exit toward Storke Rd 0.5 miles
Turn left onto N Glen NE Rd <0.1 mile
Stay straight to go onto Storke Rd 1.1 miles
Storke Rd becomes El Colegio Rd 0.1 miles
Turn right onto Camino Corto 0.5 miles
Turn right onto Del Playa Drive
Proceed to the end of the street and park near the Del Playa- Camino Majorca Road intersection. Walk to the shore. A stair will take you to the beach.

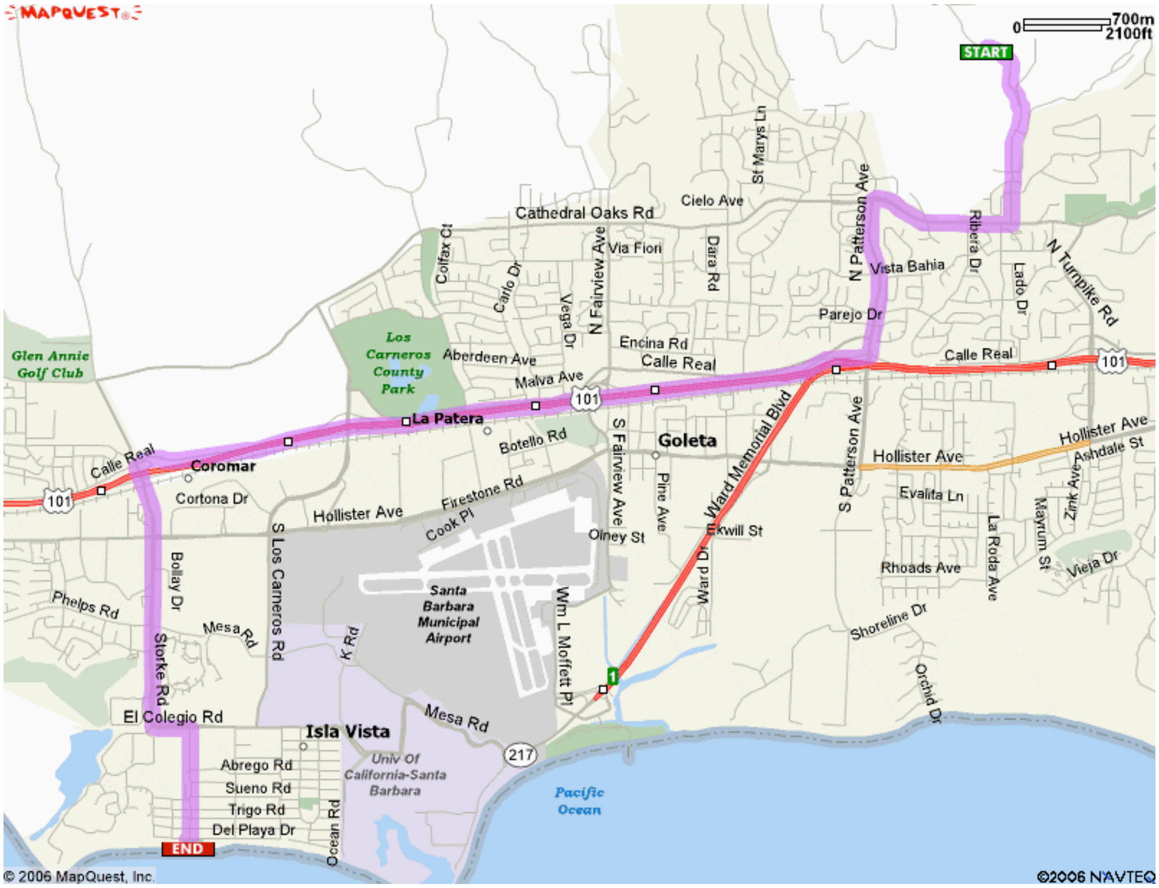

Figure 23: Map of San Marcos Pass to Bathhouse Beach

Pleistocene Terrace Near Goleta

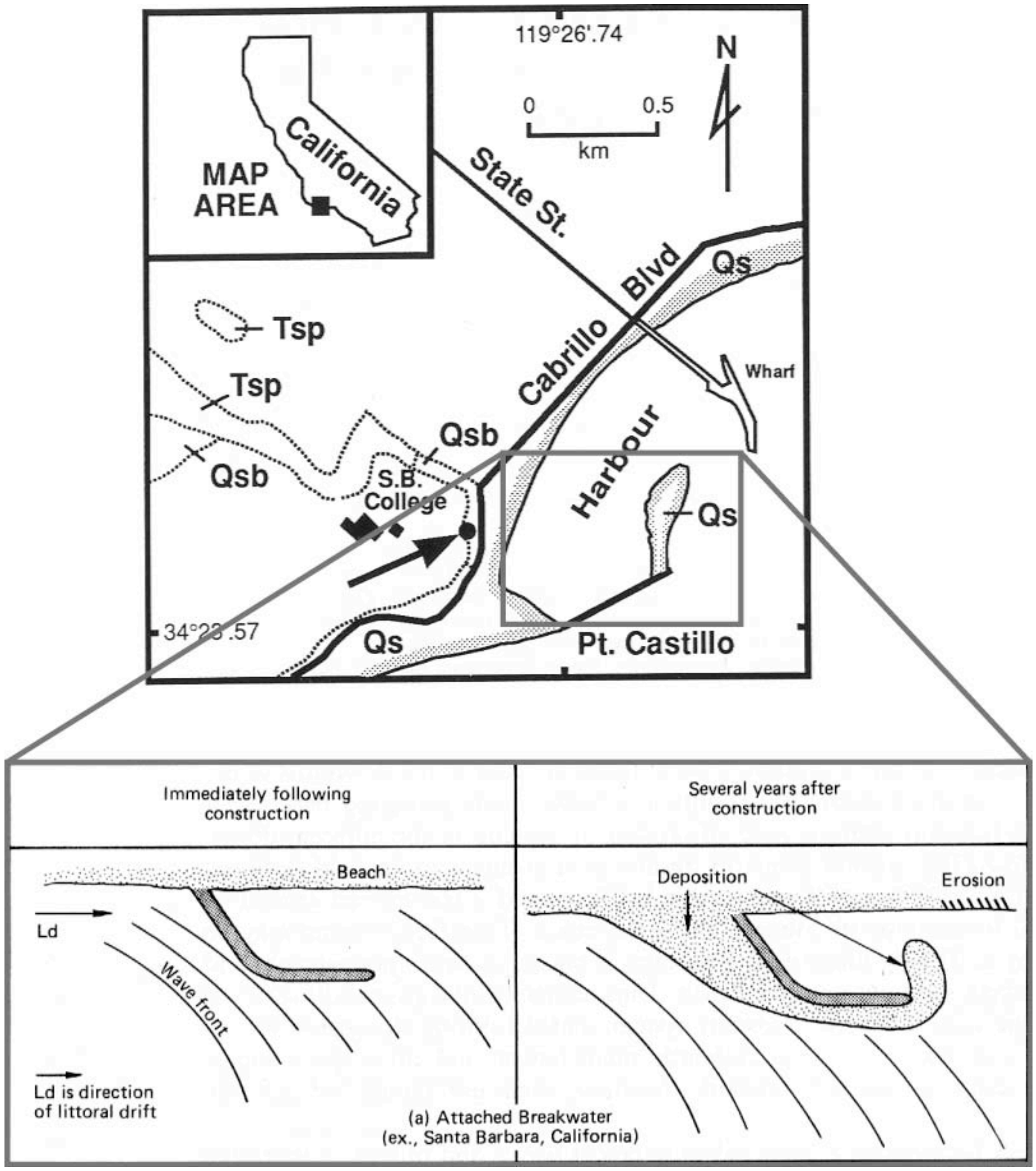

Figure 24: Map of Pleistocene Bathhouse locality (top) and the effect of breakwaters and jetties on deposition patterns in Santa Barbara Harbour (Patterson, 1990 & Keller 1979).

Figure 24 shows the long term effects of construction of a break water in Santa Barbara Harbour, resulting in the need for continual dredging of sediment in the harbour. Breakwaters are designed to intercept waves, providing a calm moorage for boats, but in doing so, breakwaters block the natural littoral (near-shore) transport of sediment resulting in new areas of deposition and erosion. While a clam moorage was achieved in Santa Barbara harbour, substantial sediment deposition has occurred, hindering the journey of the vessels that the breakwater was designed to aid. This example illustrates how the planning and implimentation of engineering structures in coastal systems must be carefully thought out to limit unforeseen consequences. (Keller 1979)

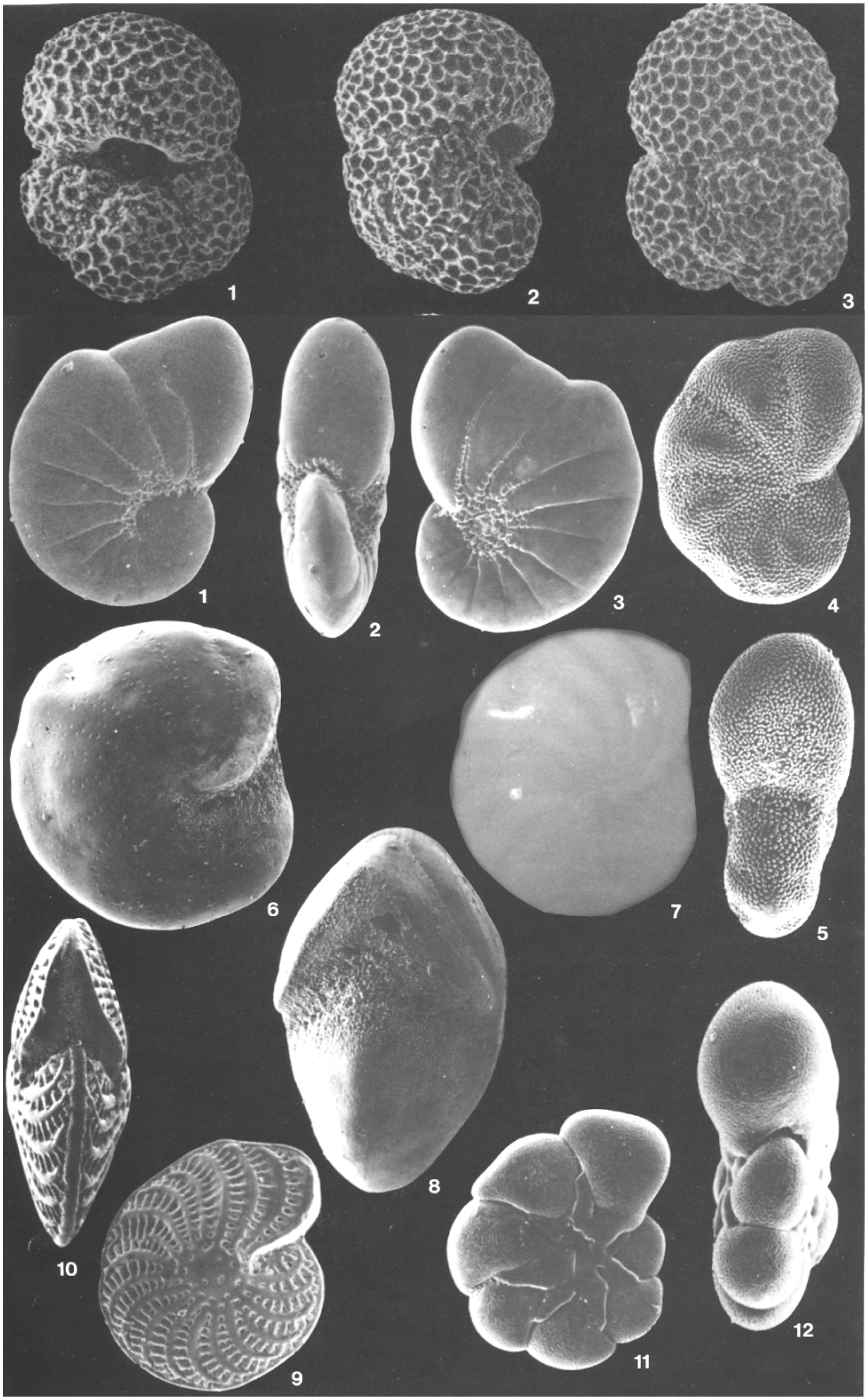

Figure 25: Local Foraminifera of the Bathhouse formation. The top three are various views of dextral (warm water) Neogloboquadrina pachyderma. (Patterson, 1990)

A one to three foot thick fossiliferous marine deposit is exposed almost continuously along the coastal bluffs of the elevated coastal block. This terrace can be identified for hundreds of miles along the Pacific Coast. The deposit rests with a sharp angular unconformity on a marine terrace 7 to 20 feet above the modern beach and eroded into the under-lying north-north-northeastward dipping diatomaceous shale of the Pliocene Sisquoc Formation. The deposit is conformably overlain by unfossiliferous alluvium-colluvium. The deposit was studied for its molluscan (Wright, 1972) and foraminiferal fauna (Barrick et al, 1989).

The Santa Barbara Formation at Bathhouse Beach contains up to 130 foraminiferal species (Young, 1981). The formation was deposited in waters of less than 100 m depth (Woodring 1957). Deposition may have begun during the late Pliocene, as indicated by the first appearance of the foram *Globorotalia truncatulinoides* (d'Orbigny) some 2 m from the base of the formation.

The megafossil molluscan fauna consists of three elements: a rock boring pelecypod element, an epifaunal gastropod element, and a sandy infaunal pelecypod element. The rock boring element consists predominantly of *Platyodon, Penitella*, and *Parapholas*, and many articulated specimens occur in living position in the bedrock surface of the terrace. The sandy-clay overlying the terrace contains a variety of epifaunal (live on the sediment surface) and infaunal mollusks (living within a soft sediment), most of which are preserved undisturbed in their living position. Many of the infaunal pelecypods are articulated and have ligamental material still attached to the valves. A thin layer (up to 2 inches thick) of tar occurs neat the center of the deposit (evidence of a late Pleistocene tar seep offshore).

The foraminiferal fauna and molluscan fauna are interpreted to have inhabited a mostly cool, shallow (<12 m, but usually 0-5 m) subtidal meters. Such sublittoral water temperatures are found presently only north of Point Conception, California, indicating slightly cooler temperatures than presently found in the Goleta area. Paleotemperature was reconstructed based on ratio of sinistral to dextral morphs of the foraminifera species, Neogloboquadrina pachyderma that made up 25-60% of the forams found. Statistical analysis of the relative amounts of sinistral to dextral morphs of Neogloboquadrina pachyderma, indicate that a warmer interval is overlain by a layer that was laid down in cooler times. In comparison to modern bathymetric and temperature ranges for species found at Goleta, the upper portion of the facies, late Pleistocene, was laid down at a max water depth of 0-10m during cooler temperatures (11-20°C) than present (13-15°C) (Patterson, 1990).

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4-3 Goleta Beach Erosion and 1969 Oil Spill

Since 2000, the sand at Goleta Beach Park off of Sandspit Road has been eroding. In order to prevent further erosion, Santa Barbara county workers have placed 1,000 linear feet of rock revetment - made up of large boulders - along the beach. While this revetment has temporarily halted the erosion of the beach, some criticize its installation. Bob Keats, vice chair of the I.V. chapter of the Surfrider Foundation, said the wall will have a negative effect on the environment. "Structures designed to trap sand, such as the rock revetment, are not environmentally sound," Keats said. "They essentially steal sand from other beaches. This may halt the erosion at Goleta Beach, but it's only going to worsen the problem at other beaches along the coast."

Keats said the problem has only worsened and that a redesign is necessary if for no other reason than to accommodate the large amount of tourists that fuel the Santa Barbara County economy. Keats said that in 1998, the county conducted a research project that concluded Goleta Beach Park was filled to capacity every day during summer months.

Santa Barbara Parks have said they don't believe the revetment will negatively impact the environment. Consultants hired by the county concluded that although an improperly built revetment could damage the environment, the one at Goleta Beach is structurally sound and is therefore not an environmental hazard.

Before the revetment was put in place, a large amount of sand was hauled to the beach in order to replace sand being washed away. The sand brought to the beach last winter cost approximately $1.5 million and was washed away after two storms. (http://www.dailynexus.com/news/2005/8764.html) The gradient of the beach changes throughout the year, the gradient is steeper in the winter when more energetic storms carry sand off shore, while in the calmer summer time, sand is washed and deposits onshore.

1969 Oil Spill

On the afternoon of January 29, 1969, an environmental nightmare began in Santa Barbara, California. A Union Oil Co. platform stationed six miles off the coast of Summerland suffered a blowout. Oil workers had drilled a well down 3500 feet below the ocean floor. Riggers began to retrieve the pipe in order to replace a drill bit when the "mud" used to maintain pressure became dangerously low. A natural gas blowout occurred. An initial attempt to cap the hole was successful but led to a tremendous buildup of pressure. The expanding mass created five breaks in an east-west fault on the ocean floor, releasing oil and gas from deep beneath the earth.

For eleven days, oil workers struggled to cap the rupture. During that time, 200,000 gallons of crude oil bubbled to the surface and was spread into a 800 square mile slick by winds and swells. Incoming tides brought the thick tar to beaches from Rincon Point to Goleta, marring 35 miles of coastline. Beaches with off-shore kelp forests were spared the worst as kelp fronds kept most of the tar from coming ashore. The slick also moved south, tarring Anacapa Island's Frenchy's Cove and beaches on Santa Cruz, Santa Rosa and San Miguel Islands.

Ecological Impact Animals that depended on the sea were hard hit. Incoming tides brought the corpses of dead seals and dolphins. Oil had clogged the blowholes of the dolphins, causing massive lung hemorrhages. Animals that ingested the oil were poisoned. In the months that followed, gray whales migrating to their calving and breeding grounds in Baja California avoided the channel —their main route south.

The oil took its toll on the seabird population. Shorebirds like plovers, godwits and willets which feed on sand creatures fled the area. But diving birds which must get their nourishment from the waters themselves became soaked with tar.

The Santa Barbara Zoo was among three emergency bird treatment centers established during the disaster. Volunteers were recruited to pluck oiled birds from local beaches. Grebes, cormorants and other seabirds were so sick, their feathers so soaked in oil that they were not difficult to catch. Birds were bathed in Polycomplex A-11, medicated, and placed under heat lamps to stave off pneumonia. The survival rate was less than 30 percent for birds that were treated. Many more died on the beaches where they had formerly sought their livelihoods. Those who had managed to avoid the oil were threatened by the detergents used to disperse the oil slick. The chemicals robbed feathers of the natural waterproofing used to keep seabirds afloat.

In all 3686 birds were estimated to have died because of contact with oil. Aerial surveys a year later found only 200 grebes in an area that had previously drawn 4000 to 7000.

Cleanup Efforts It took oil workers 11 1/2; days to control the leaking oil well. Workers pumped chemical mud down the 3500 foot shaft at a rate of 1500 barrels an hour. It was then topped by a cement plug. Residual amounts of gas continued to escape and another leak sprung up weeks later, releasing oil for months to follow.

Skimmers scooped up oil from the surface of the ocean. In the air, planes dumped detergents on the tar covered ocean in an attempt to break up the slick. On the beaches and harbors, straw was spread on oily patches of water and sand. The straw soaked up the black mess and was then raked up. Rocks were steamed cleaned, cooking marine life like limpets and mussels that attach themselves to coastal rocks.

What Went Wrong?Union Oil's Platform A ruptured because of inadequate protective casing. The oil company had been given permission by the U.S. Geological Survey to cut corners and operate the platform with casings below federal and California standards. Investigators would later determine that more steel pipe sheating inside the drilling hole would have prevented the rupture.

Because the oil rig was beyond California's three-mile coastal zone, the rig did not have to comply with state standards. At the time, California drilling regulations were far more rigid those implied by the federal government.

Aftermath In the spring following the oil spill, Earth Day was born nationwide. Many consider the publicity surrounding the oil spill a major impetus to the environmental movement.

Only days after the spill began, Get Oil Out (GOO) was founded in Santa Barbara. Founder Bud Bottoms urged the public to cut down on driving, burn oil company credit cards and boycott gas stations associated with offshore drilling companies. Volunteers helped the organization gather 100,000 signatures on a petition banning offshore oil drilling. While drilling was only halted temporarily, laws were passed to strengthen offshore drilling regulations. Union Oil suffered millions in losses from the clean-up efforts, payments to fishermen and local businesses, and lawsuit settlements. But maybe worse, the reputation of the oil industry was forever tarnished.

In Their Own Words . . . Fred L. Hartley, president of Union Oil Co.: "I don't like to call it a disaster," because there has been no loss of human life. "I am amazed at the publicity for the loss of a few birds."

Santa Barbara NewsPress Editor Thomas Storke: "Never in my long lifetime have I ever seen such an aroused populace at the grassroots level. This oil pollution has done something I have never seen before in Santa Barbara – it has united citizens of all political persuasions in a truly nonpartisan cause."

U.S. President Richard Nixon: "It is sad that it was necessary that Santa Barbara should be the example that had to bring it to the attention of the American people. What is involved is the use of our resources of the sea and of the land in a more effective way and with more concern for preserving the beauty and the natural resources that are so important to any kind of society that we want for the future. The Santa Barbara incident has frankly touched the conscience of the American people."

Many credit the 1969 oil spill with igniting the environmental movement. For eleven days, 200,000 gallons of crude oil spilled into the channel from a disabled oil rig. In the aftermath, 3600 birds were dead along with ten seals and dolphins and countless fish and marine invertebrates

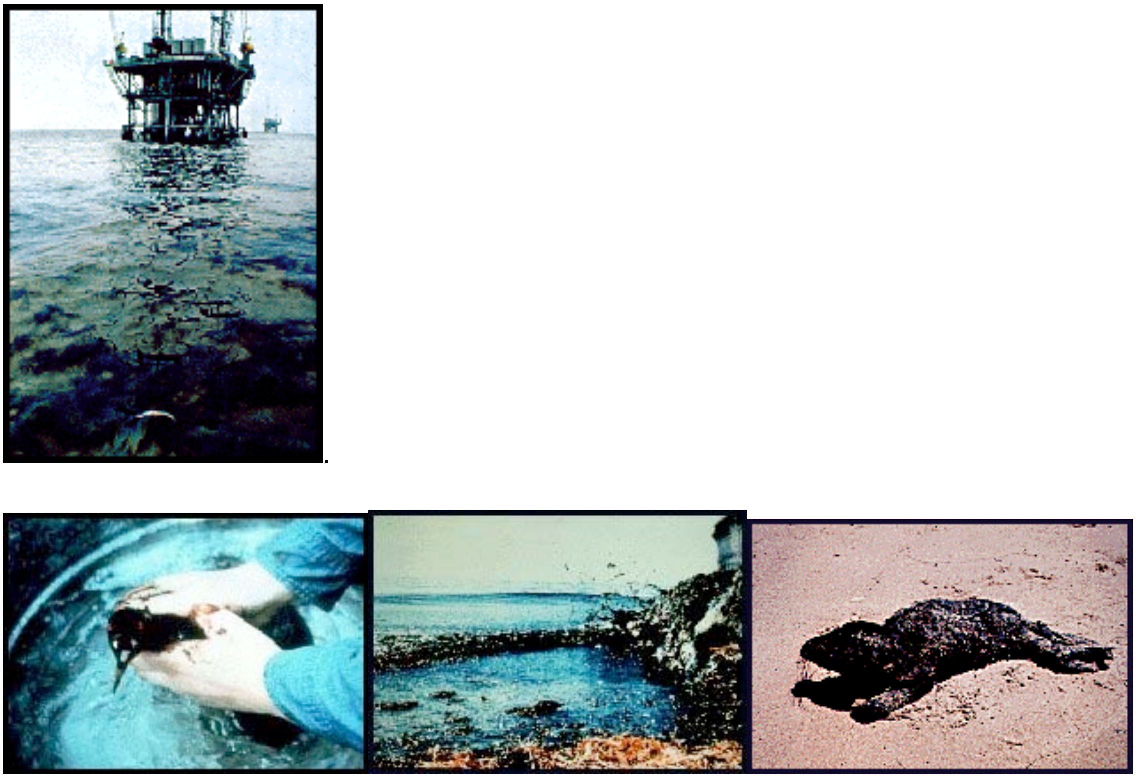

Figure 26: Photos of the oil platform and aftermath of 1969 spill. (http://www.geog.ucsb.edu/~jeff/sb_69oilspill/69oilspill_articles2.html)

Day 5: Tuesday, 22 August 2006

Summary: Leave Oso Campground
Guadalipe-Nipomo Dunes
Morro Bay
Travel to San Simeon Campground

Directions from Oso Campground to Guadalipe-Nipomo Dunes:

Start out going west on Paradise rd / NF-5N18 toward Fremont tract 3.0 miles.
Turn right onto CA-154 / San Marcos Pass rd Continue to follow CA-154. 21.8 miles.
Merge onto US-101 N. 18.9 miles.
Take the Clark ave exit- exit 164- toward Orcutt 0.1 miles.
Turn left onto E Clark ave 3.5 miles.
Turn RIGHT onto Casmalia rd / CA-1 / Cabrillo hwy.
Continue to follow CA-1 / CABRILLO HWY. 9.8 miles.
Turn left onto W Main st / CA-166 3.1 miles.
End at Guadalupe-Nipomo Dunes Preserve Guadalupe

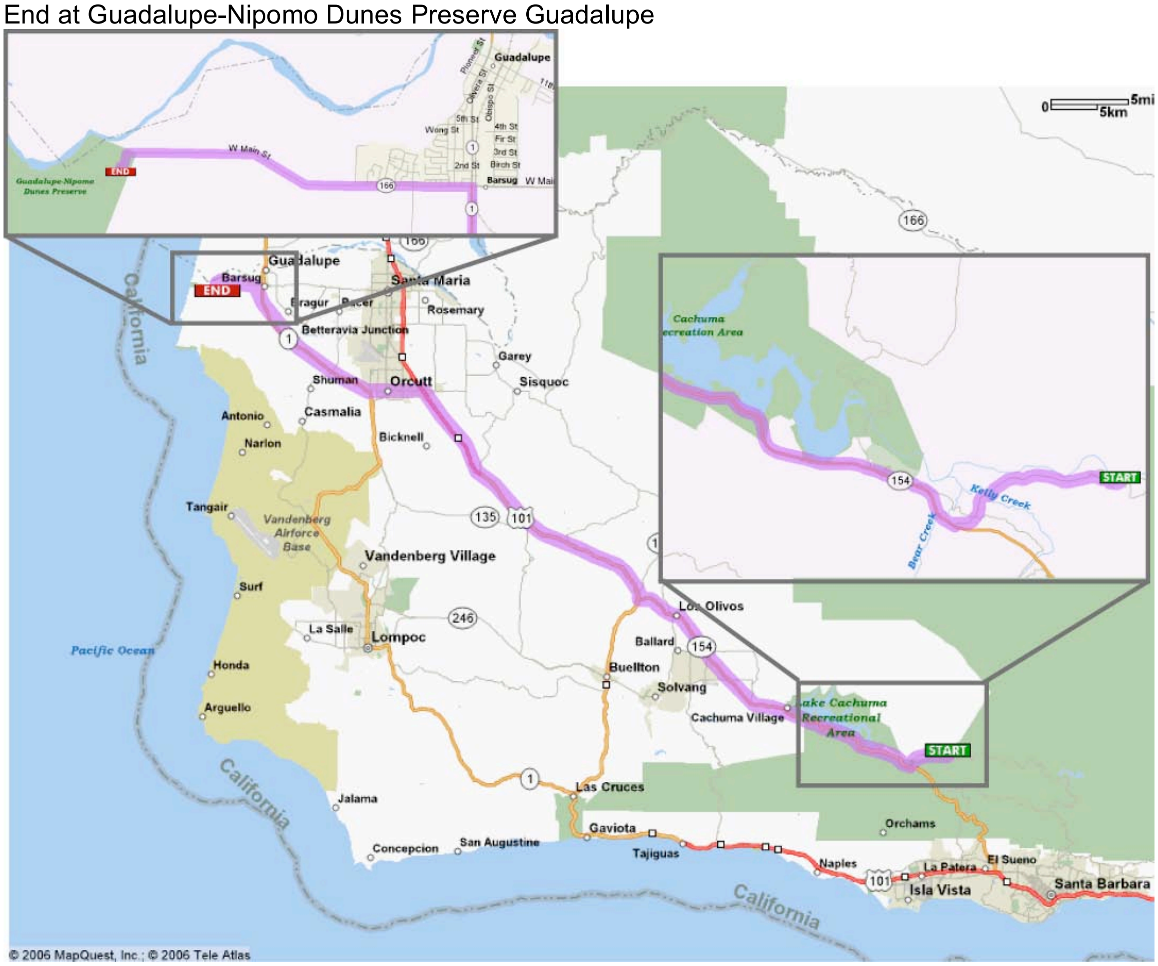

Figure 27: Oso to Dunes

5 - 1 Guadalipe-Nipomo Dunes south of Pismo - Oceano Dunes SVRA Coastal Dunes

Oceano Dunes is recognized by scientists, conservationists, government agencies, and the public as the finest, most extensive coastal dunes remaining in California. Most of the materials which form the dunes has been carried down to the ocean by various rivers and creeks, or deposited by ocean currents and then shaped by the wind into the dunes that we see today.

The prevailing winds that blow in from the ocean push sand particles up into wave-like crests that run north and south. On the west, or the "windward" side, the slope is gentle. On the east, or the "leeward" side, the slope is quite steep. Sand grains, as they are blown over the dune crest, tend to accumulate high on the leeward slope, slip down the slope as thin tongues of sand. For this reason the leeward slope is called a "slipface."

Coastal Dunes Shaped by wind into curving ridges, coastal sand dunes are among the most dynamic and fragile natural formations. Their contours shift over time until hardy dune pioneer plants take hold in the drifting sand and create a stable landform. Even then, dunes can change form rapidly under the stress of storm waves and wind, or the traffic of human activity.

Offshore sandbars and sediment deposited at the mouths of rivers are the most important sources of material for dune building; sediment is carried by longshore currents until a projection landform traps the particles and they are deposited on the beach by wave action. Dune formation begins when wind blows dry sand particles landward from the beach. Drifts accumulate around object, such as plants and logs, that interrupt the wind flow. With steady winds, the sand drift acts as a barrier to moving sand, and the drift gradually grows into a sizable mound. Until a dune is completely veiled and stabilized by plant cover, sand may be borne away by winds.

Coastal dune fields form characteristic patterns. A common pattern along the Northern California coast is a series of parallel ridges perpendicular to the prevailing winds, called "transverse ridges." The parabola-type dune field is a series of U-shaped dunes with the concave side facing the prevailing wind direction; this dune type is found at Pismo Beach in Central California. All dune fields consist of two or three sets of parallel dunes, with the most recently formed foredunes nearest the beach, and the older, usually vegetated and stabilized dunes farthest inland; the inland dunes may by as much as 18,000 years old.

Deep-rooted succulent, matted plants such as beach strawberry, silver beachweed, and yellow sand verbena grow on the foredunes along with various dune grasses. The aggressive European beach grass, planted on California dunes in the 1930's, rapidly extends new shoots when half buried by sand drifts. The new shoots snag more sand, building and stabilizing the dune, and crowding out native species. The globose dune beetle, which cannot survive under European beach grass, is restricted to areas where native dune plants persist.

Protected by the foredunes from salt spray and wind, wild buckwheat, yellow bush lupine, and purpleflowered beach lupine grow on the richer soils of the back dunes. At least tree rare insect species, the San Francisco tree lupine moth, the Pheres blue butterfly, and the Morro blue butterfly, lay eggs on the lupines. Wild buckwheat is a favorite food of the endangered Smith's blue butterfly larvae.

Deer mice, California voles, and black legless lizards burrow into the sand dunes, seeking cover from predators such as the northern harrier. This white-rumped hawk hovers a few feet above land; concave disks on the sides of its head funnel sound into the ears, enabling the hawk to detect the slightest movement of its prey. Gray foxes and striped skunks eat insects and dune plants. Mule deer wander over the dunes, browsing on shrubs.

Dunes shield low lying inland areas from violent storm waves. Sand eroded from dunes and beaches during winter storms usually forms a sandbar a short distance offshore. This sand is gradually returned to the beach during the calm summer season. A stable dune system can undergo some wave erosion without permanent damage.

California's dunes were formed over thousands of years, yet today, dune erosion is outstripping sand deposition. Dams trap river sediments, depleting the sand supply, and coastal protective structures, such as seawalls, disrupt the natural recycling of sand from sandbar to beach. Coastal development has disturbed dunes at many points along the coast. Off-road vehicles, foot traffic, and horses cam damage dune plants, loosening the sands and leaving the dunes vulnerable to wind erosion and blowouts.

Of the 27 dune fields in coastal California, the largest are the Monterey Bay dunes, covering about 40 square miles, and the 18 square mile Nipomo Dune complex, north and south of the Santa Maria River. Other major dune fields are located at Humboldt Bay and San Diego Bay.

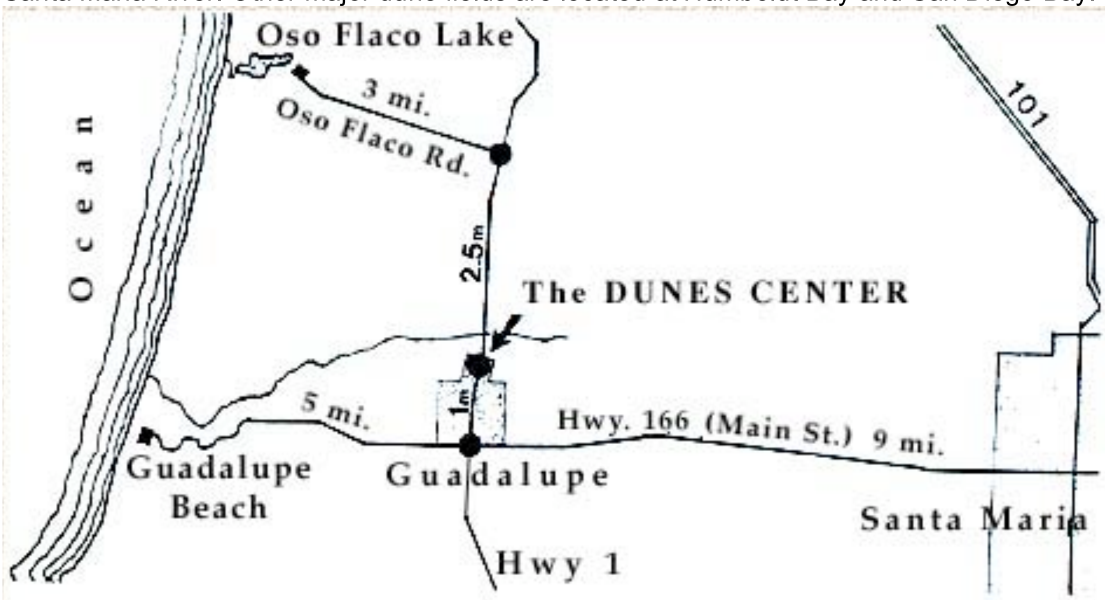

Figure 28 - Map showing location of Dune Preserve
http://santalucia.sierraclub.org/osoflaco.html

Figure 29: Dunes as viewed northward from Mussel Rock Dune
(http://santalucia.sierraclub.org/osoflaco.html)

5 - 2 Morro Rock, Morro Bay –

Directions from Dunes to Morro Rock:

Start out going east on W Main st / CA-166 toward Calle Cesar E Chavez 3.1 miles.
Turn left onto CA-1 / Cabrillo hwy / Guadalupe st Continue to follow CA-1 / Cabrillo hwy. 11.5 miles.
Turn right onto Valley rd 1.3 miles.
Turn right onto Fair Oaks ave 0.4 miles.
Turn left onto Traffic Way 0.3 miles.
Turn left onto CA-227 / W Branch st / E Grand Continue to follow CA-227 / E Grand ave 0.1 miles.
Merge onto US-101 N. 15.9 miles.
Take the CA-1 N exit toward Morro Bay / Hearst Castle <0.1 miles.
Turn slight right onto Toro ST / CA-1. <0.1 miles.
Turn right onto CA-1 / Walnut St <0.1 miles.
Turn right onto CA-1 N / Santa Rosa St Continue to follow CA-1 N. 12.9 miles.
Take the Main St ramp 0.6 miles.
Turn right onto Beach St. 0.2 miles.
Turn right onto Embarcadero rd 0.4 miles.
Turn left onto Coleman Dr (Gate access required) 0.4 miles.
End at **Morro Rock Natural Preserve** Coleman Dr, Morro Bay

Figure 30: Dunes to Morro Rock

Collectively, these plugs are often termed “The Morros” or “The Nine Sisters” (not Fourteen???). The plugs form a string of picturesque hills with bold, rocky upper slopes and grassy, oak-covered lower slopes. The peaks separate the Los Osos and Chorro Valleys. The most prominent peaks are those between San Luis Obispo and Morro Bay. Morro Rock itself rises more than 160 m above sea level. These features are interpreted as eroded Miocene (25-21 Ma) volcanoes, probably emplaced along the West Huasna fault. The volcanic edifices themselves have long since been eroded, leaving only the stumps of the more resistant lavas that once congealed in the throats of the volcanoes. These volcanic centers are thought to be the source of the Obispo tuff, a pyroclastic deposit widely distributed in the area. The composition of the volcanic rocks preserved in the volcanic necks varies from basalt to dacite. The larger peaks in the chain are andesite, but Morro Rock is composed of dacite. Three of the Morros are found in Morro Bay State Park: Morro Rock (el. 576 ft; Figure 31), Cerro Cabrillo (el. 911 ft), and Black Hill (el. 665 ft). The remaining plugs range in height from 780 to 1560 ft above sea level.

Figure 31 - Morro Rock, a dacite plug in Morro Bay. Morro Rock is a tombolo or land-tied island that resulted from beach and dune deposition on its landward side (From Norris and Webb, 1989, p. 402, Fig. 11-20) Morro Rock has long been a focal point of human interest. The distinctive rock was a prominent landmark long before Juan Cabrillo named it “el Moro” in 1542. Between the late 19th and mid-20th centuries, Morro Rock was an important quarry site, supplying dense, wave-resistant dacite for central coast harbor construction. The Rock to the mainland, forming a major recreational site. Morro Rock is also a protected home for the endangered peregrine falcon, so access to the Rock is prohibited.

Directions from Morro Rock to San Simeon Campground:

Start out going Northeast on Coleman Dr toward Embarcadero rd 0.4 miles.
Turn right onto Embarcadero rd 0.4 miles.
Turn left onto Beach St 0.2 miles.
Turn left onto Main st 0.5 miles.
Merge onto CA-1 N / Cabrillo hwy via the ramp on the left 23.7 miles.
Turn right onto San Simeon Creek rd <0.1 miles.
End at San Simeon State Park Camp 300 San Simeon Creek Rd, Cambria.

Figure 32: Morro Rock to San Simeon

Day 6 - Wednesday, 23 August 2006

Summary: Leave San Simeon Campground
Herst Castle
San Luis Harbour
Return to San Simeon Campground

Directions: San Simeon to Hearst Castle

Start out going west on San Simeon Creek rd toward CA-1 / Cabrillo hwy <0.1 miles.
Turn right onto CA-1 / Cabrillo hwy 4.6 miles.
Turn right onto Hearst Castle Rd (Gate access required) 0.4 miles.
End at Hearst Castle Historical Mnmnt 750 Hearst Castle Rd, San Simeon.

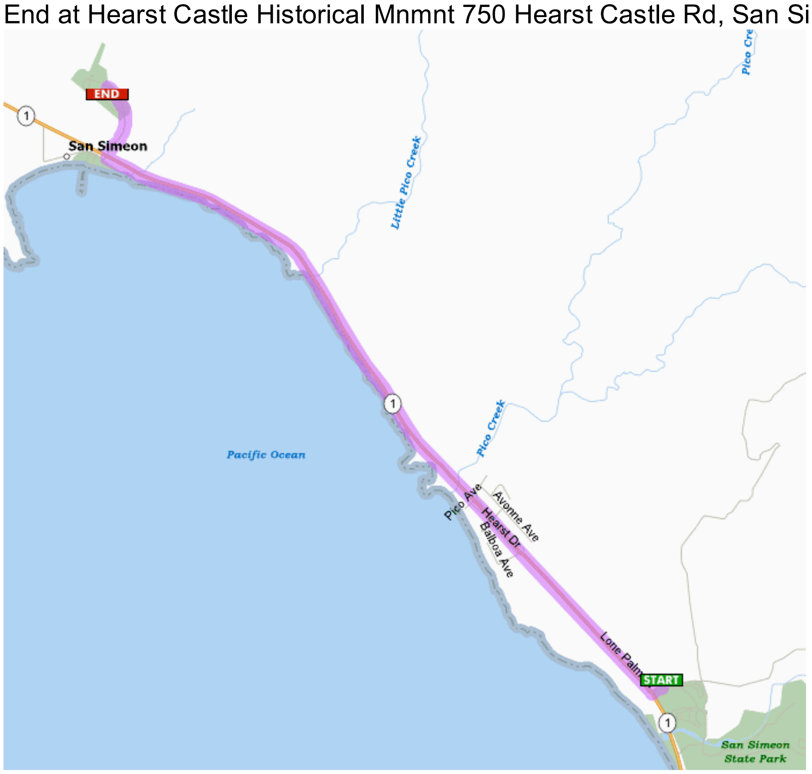

Figure 33: San Simeon to Hearst Castle

6-1 Hearst Castle

While on tour, still and video photography is allowed for personal use only. We do not allow flash photography or tripods. The tour guides are paid state employees and do not accept tips. All visitors are asked to leave any packages or backpacks in their vehicles while on tour

Figure 34: Aerial photo of Hearst Castle.

Hearst Castle was the palatial estate of newspaper magnate William Randolph Hearst. It is located near San Simeon, California, on a hill overlooking the Pacific Ocean, halfway between Los Angeles and San Francisco. Donated by the Hearst Corporation to the state of California in 1957, it is now a State Historical Monument and a National Historic Landmark, open for public tours. Hearst formally named the estate 'La Cuesta Encantada' ('The Enchanted Hill'), but he usually just called it 'the ranch'. (http://en.wikipedia.org/wiki/Hearst_castle)

Figure 35: William Randolph Hearst (1863-1951)
(http://www.hearstcastle.com/history/william_r_hearst.asp)

William Randolph Hearst, the man behind Hearst Castle, is an important figure from the twentieth century whose influence extended to publishing, politics, Hollywood, the art world and everyday American life. His power and vision allowed him to pursue one of the most ambitious architectural endeavors in American history, the result of which can be seen in magnificent grounds and structures of Hearst Castle.

Throughout his life, Hearst dreamed of building a dwelling similar to those he had seen on his European tour as a boy. Hearst Castle was to become the realization of this dream as he and architect Julia Morgan collaborated for 28 years to construct a castle worthy of those he saw in Europe.

The estate's magnificent main house, "Casa Grande," and three guest houses are of Mediterranean Revival style, while the imposing towers of Casa Grande were inspired by a Spanish cathedral. The blending of the architectural style with the surrounding land, and Hearst's superb European and Mediterranean art collection, was so seamless that world-renowned architectural historian, Lord John Julius Norwich, was moved to say that "Hearst Castle is a palace in every sense of the word."

Figure 36: The Neptune Pool.

Three swimming pools were built on this site, each successively larger. Initial plans for the site called for a "Temple Garden" with an ornamental pool and temple structure. The Roman Pool at Hearst castle is a tiled indoor pool decorated with eight statues of Roman gods, goddesses and heroes. The pool appears to be styled after an ancient Roman bath such as the Baths of Caracalla in Rome c. 211-17 CE. The mosaic tiled patterns were inspired by mosaics found in the 5th Century Mausoleum of Galla Placidia in Ravenna, Italy.

Hearst furnished the estate with truckloads of art, antiques, and even whole ceilings that he acquired en masse from the great houses of Europe. Hearst Castle was like a small self-contained city, with 56 bedrooms, 61 bathrooms, 19 sitting rooms, 127 acres of gardens, indoor and outdoor swimming pools, tennis courts, a movie theater, an airfield, and the world's largest private zoo. Zebras and other exotic animals still roam the grounds. Morgan, an accomplished civil engineer, devised a gravity-based water delivery system from a nearby mountain.

William Randolph Hearst created the largest private zoo in the world on his ranch at San Simeon. Traveling the winding ranch road to Hearst Castle guests passed through fenced fields populated with many species of exotic wild animals freely roaming over the hillsides as though they were native to this land. It was an amazing sight. The ever-changing collection of animals was established in 1923 when American bison, Rocky Mountain elk, and European white fallow deer were acquired.

Hearst Airport Many of Mr. Hearst's guests at the ranch began their visits from the same spot that today's visitors do. The original airport/airstrip was located where the current Visitor's Center is today. The building that today houses the ticket office and snack bar is where the hangar was located. In addition to the stars, politicians and other influential members of society that were continually being flown in for extravagant parties. Famous aviators that visited Hearst's ranch included; Sir Charles Kingford-Smith, Howard Hughes, Amelia Earhart, and Charles Lindbergh.

"Yellow Journalism" Though the term was originally coined to describe the journalistic practices of Joseph Pulitzer, William Randolph Hearst proved himself worthy of the title. Today, it is his name that is synonymous with "yellow journalism." William Randolf Hearst hated minorities, and he used his chain of newspapers to aggravate racial tensions at every opportunity. Hearst especially hated Mexicans. Hearst papers portrayed Mexicans as lazy, degenerate, and violent, and as marijuana smokers and job stealers. The real motive behind this prejudice may well have been that Hearst had lost 800,000 acres of prime timberland to the rebel Pancho Villa, suggesting that Hearst's racism was fueled by Mexican threat to his empire. (http://www.redinkwhitelies.com/book.htm)
Ernest L. Meyer wrote: "Mr. Hearst in his long and not laudable career has inflamed Americans against Spaniards, Americans against Japanese, Americans against Filipinos, Americans against Russians, and in the pursuit of his incendiary campaign he has printed downright lies, forged documents, faked atrocity stories, inflammatory editorials, sensational cartoons and photographs and other devices by which he abetted his jingoistic ends."(http://www.brasscheck.com/seldes/lords17.html)

6-2 SAN LUIS HARBOUR

Directions from Hearst Castle to San Luis Harbour:

Go NW on Hearst Castle Rd 0.1 miles
Turn left, stay on Hearst Caslte Rd 0.5 miles
Turn left on CA-1 S/CABRILLO Hwy 41.2 miles
Turn right onto CA-1/Olive ST 0.1 miles
Merge onto US 101 S/CA-1 S 7.6 miles
Take Avila Beach Drive Exit 0.2 miles
Turn right onto Avila Beach Drive 2.6 miles
Turn left onto San Miguel St 0.1 miles
Turn right onto First St <0.1 miles
Turn left onto San Francisco Street

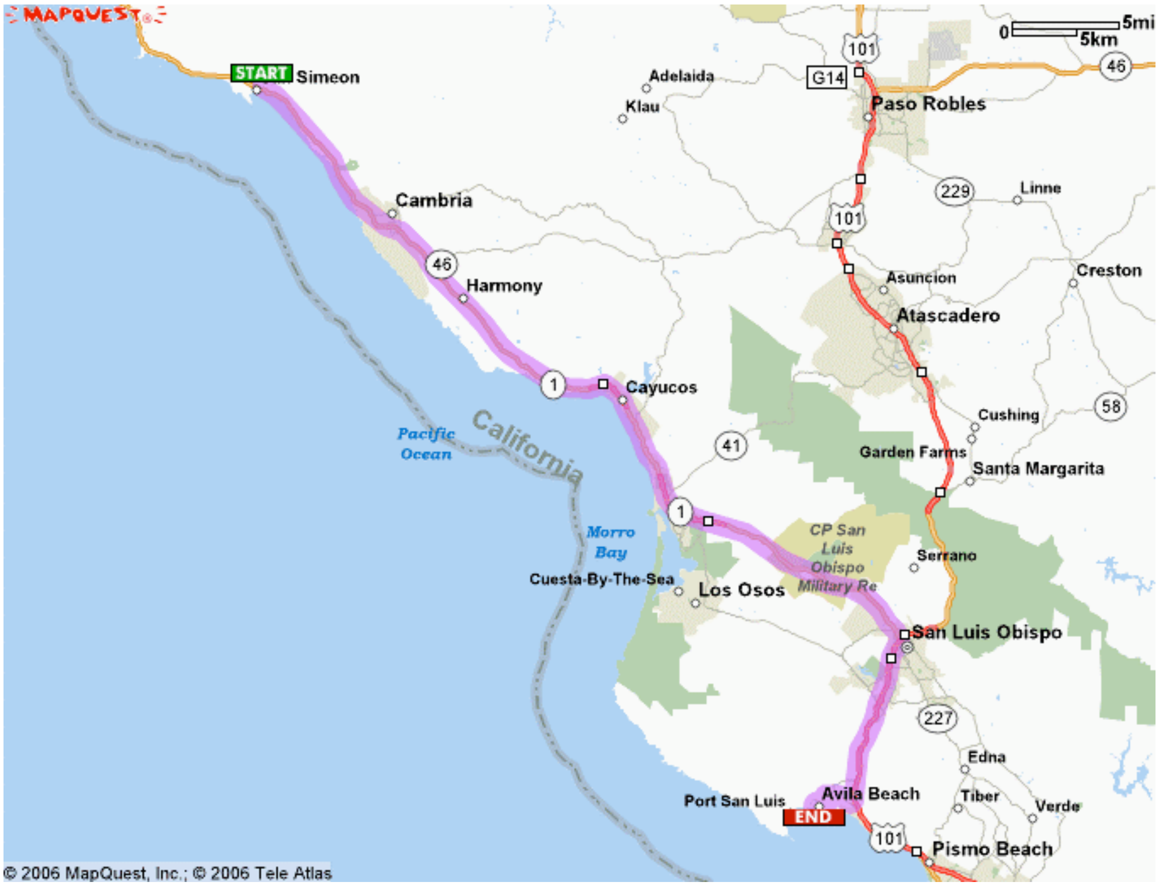

Figure 37: Hearst Castle to Port San Luis Harbour

COAST RANGE AND FRANCISCAN ASSEMBLAGE

(From Mary Leech, 2006)
The Coast Ranges extend from the southern and western edge of the Klamath Mountains near the Oregon border, to the Transverse Ranges in southern California, a distance of more than 1000 km; in addition, the geology found in the Coast Ranges continues as basement through the western Transverse Ranges, then offshore from the Peninsular Ranges and Baja California, finally to reappear on land at Cedros Island, offshore Mexico. The Coast Ranges got their name because of their proximity to the Pacific Ocean, and the interface between shoreline processes and the geology of the province dominates the scenery found in the province, as seen in the view of a wave-carved sea tunnel near Mendocino in northern California, eroded into Franciscan Complex sandstones.

In the Coast Ranges Province proper, three linear belts are recognized: (1) an eastern belt composed of Franciscan Assemblage (Franciscan Complex)rocks (graywacke and shale in coherent units, often metamorphosed to blueschist facies, together with melanges composed not only of sheared shale and graywacke blocks but also exotic blocks of some or all of the following rock types: serpentinite, chert, greenstone (as in the photo to the right), eclogite, and blueschist, overlain by the Jurassic Coast Range ophiolite (in thrust contact), which is in turn depositionally overlain by the Jurassic to Cretaceous Great Valley Sequence of turbidites and fan deposits; (2) a central belt composed of Mesozoic plutons that had intruded metamorphic basement rocks, overlain by younger sediments, and called the Salinian Block; and (3) a western belt which consists once again of the sequence of Franciscan Assemblage, Jurassic Coast Range ophiolite, and remnants of the Great Valley Sequence, but with some differences – less blueschist metamorphism and more age restriction of Franciscan sediments (Jurassic to Paleocene in the northern part of the eastern belt, but lack of Paleocene-age sediments in the western belt), and more restricted Jurassic age for Great Valley sequence equivalent rocks. The three belts are separated by major faults, the eastern belt from the Salinian Block by the active San Andreas strike-slip system, and the Salinian Block from the western belt by the more enigmatic Sur-Naciemento fault system, a complex system of apparent strike-slip faults (e.g., Rinconada fault), but with some faults that have reverse components and are thrusts (Page, 1981). These boundary fault systems have been responsible for several hundreds of kilometers of right lateral motion in the Neogene, but paleomagnetic and micropaleontologic evidence in studies of Franciscan cherts suggests that these rocks have travelled from low latitudes, more than 1000 km (Champion, 198?; Wahrhaftig, 1987). While the San Andreas system of faults is clearly active, and has caused numerous historical earthquakes in the Coast Ranges (the most famous being the large 1906 San Francisco event), the Sur-Nacimiento is considered to be inactive (Alt and Hyndman, 2000).

The origin of the Franciscan Complex has been assumed to be related to subduction processes. A simplified version of the geological history was synthesized by Blake and Jones (1981):

"The close temporal and spatial relations of Franciscan rocks with rocks of the structurally overlying Great Valley Sequence and the volcanoplutonic rocks that lie to the east in the Klamath Mountains and the Sierra Nevada have led to the widely accepted hypothesis that rocks of the Franciscan assemblage were deposited in an active trench contemporaneously with deposition of the Great Valley sequence in a forearc basin adjacent to the active Klamath-Sierra arc complex. These three elements comprise the arc, arc-trench gap, and trench setting . . . typical of convergent plate boundaries (Dickinson, 1971). A large component of subduction generally has been thought to be eastward normal to the trend of the continental margin (e.g., Hamilton, 1969)." Blake and Jones (1981) go on to explain that, although meritorious in a general sense, the hypothesis is too simplistic to explain some anomalous relationships, including the nearby presence of additional arc terranes, and sedimentological problems (compositional characteristics and correlations).

Basaltic Lavas at Port San Luis

Pillow lavas and sheet flows exposed at the harbour in Port San Luis are fragments of seafloor rocks of late Jurassic age that are part of the Franciscan complex. To the east and north are more complete ophiolite sequences of the middle to late Jurassic Coast Range Ophiolite (CRO) that are exposed over a 700 km-long stretch from Santa Barbara to north of San Francisco. Rocks of the Franciscan Complex were underthrust to the east beneath the CRO during the late Jurassic.

The mafic lavas of the CRO and the Franciscan differ chemically (Shervais and Kimbrough, 1985). Lavas and intrusive rocks of the CRO have the geochemical signature of ocean island tholeiitic rocks, and are thought to represent remnants of island arc crust (back-arc basalts?) that were added to the California margin ans a small back-arc basin closed. In contrast, lavas from the Franciscan Complex appear to be fragments of true oceanic crust that were added to the California margin during subduction of an oceanic plate.

Figure 38: Pillow lavas at the Port San Luis Pier.
www.cuesta.edu/deptinfo/geology/pillow_lava.htm

PILLOW LAVAS AND BLUESCHISTS AT PORT SAN LUIS HARBOUR

(Personal communication from Cilff Hopson, UCSB)

The Port San Luis pillow lavas and sheet flows are the next best thing to peering out the observation port of Alvin, along any mid-ocean ridge! Walk ~50 feet to the steep rocky bluff along the waterfront, where the parking lot ends. Traverse west along the foot of the bluff, on the narrow ledge at the waterline.

Those magnificent pillow lavas, with interpillow pelagic limestone, are exposed in the steep face of the bluff (~40-50 feet long), as well as beneath your feet. The greenish stuff in the pillow interstices is "chloritic" material, altered from the original glassy selvages spalled off the pillow surfaces; locally this intermixes with the interpillow lms.

Keep going beyond the end of the bluff, where you turns right and jump onto a narrow sandy beach; located here are the most gorgeous thin sheet flows this side of the East Pacific Rise. The upper surface of one sheet flow, with pahoehoe-type ropystructure, slopes toward you. And the sheet flows, and their relationships to the pillow lavas, are well displayed in cross section. The lavas here (Franciscan, probably Jurassic) have MORB-type chemistry (Shervais et al., ~1990).

As an extra dividend, check out the quarried blocks on which the parking lot is constructed. These are Cretaceous blueschists, barged over from Santa Catalina Island. The dark blue, very fine-grained rock is chiefly lawsonite-blue amphibole, + minor sphene, aragonite, etc. But the conspicuous greenish veins in them are pumpellyite (light green) and actinolite (dark green), ± white calcite, etc. These actinolite-pumpellyite veins are Miocene, formed during uplift associated w/ Miocene core complexing.

Directions from San Luis Harbour back to San Simeon Campground:

Go north on San Francisco St toward First Street <0.1 miles
Turn right onto First ST <0.1 miles
Turn left onto San Miguel St 0.1 miles
Turn right onto Avila Beach Drive 2.7 miles
Turn left to stay on Avila Beach Drive 0.2 miles
Merge onto US-101 N/CA-1 N for 7.8 miles
Take CA-1 N turnoff to Morro Bay/Hearst Castle <0.1 miles
Turn a slight right onto Toro St/CA-1 <0.1 miles
Turn right onto CA-1/Walnut St <0.1 miles
Turn right onto CA-1 N/Santa Rosa St
Follow CA-1N for 37 miles
Turn right onto San Simeon Creek Road

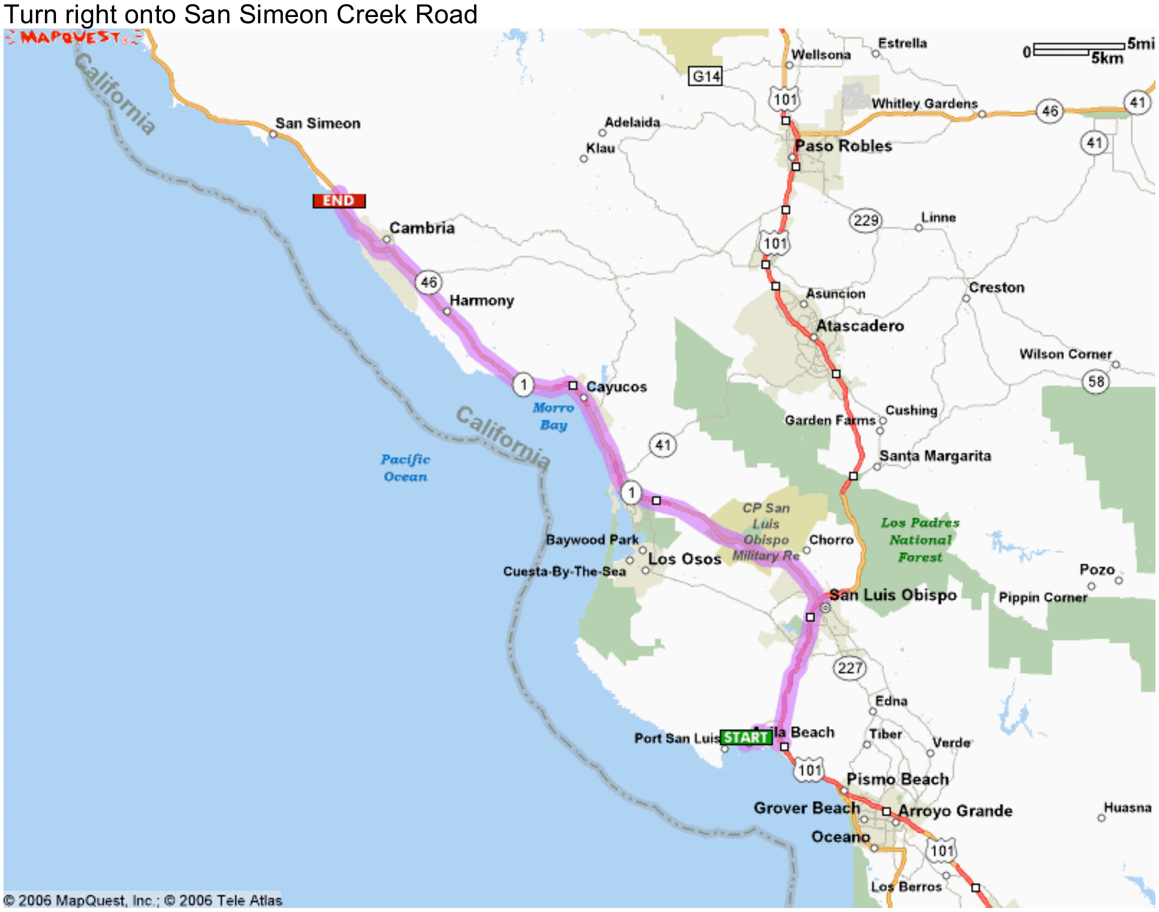

Figure 39: San Luis to San Simeon

Day 7 - Thursday, 24 August 2006

Summary: Leave San Simeon Campground
Monterey Bay Aquarium
Travel to Day's Inn in Monterey Bay

Directions: San Simeon to Monterey aquarium

Start out going west on San Simeon Creek rd toward CA-1 / Cabrillo hwy <0.1 miles.
Turn right onto CA-1 N / Cabrillo hwy Continue to follow CA-1 N 96.6 miles.
Take the exit toward Monterey 0.3 miles.
Merge onto Munras ave 0.8 miles. Stay straight to go onto Abrego St 0.4 miles.
Turn slight right onto Washington St 0.1 miles.
Turn left onto Lighthouse ave 1.3 miles.
Turn right onto David ave 0.1 miles. David ave becomes Cannery Row <0.1 miles.

Figure 40: San Simeon to Monterey Bay Aquarium

7-1 Monterey Bay Aquarium

History Monterey Bay Aquarium is located on historic Cannery Row in Monterey. The city of Monterey was named by explorer Sabastian Vizcaino after Count Monte Rey, viceroy of Mexico. The city was named capital of Alta California (Old California) in 1775, was fortified and became a port of entry and center of Spanish culture. Monterey became a retreat for artists and writers over the years, (e.g., Robert Louis Stevenson) and is the backdrop for several of John Steinbeck's novels (e.g. *Cannery Row*). At the turn of the century, the sardine industry became important for the local economy and many people relied on the bounty of fish that lived off the central Californian coast. However, by the 1940s, the decline of the sardine population led to the eventual closing and abandonment of the canning houses on Cannery Row. Monterey fell into economic despair soon after. Only until recently, in the last 25 years, has Monterey made a comeback as a major tourist area and site for scientific research. The establishment of a world-class aquarium by David Packard in 1984 is part of that success story.
The aquarium is situated on the southern edge of Monterey Bay, whose waters nurture dense kelp forests, rocky tide pools and an abundance of vertebrate and invertebrate marine life. The exhibits in the aquarium focus on the life found exclusively in the Monterey Bay area.

Directions from Monterey Bay Aquarium to Day's Inn

Go NW on Cannery Row toward David Ave <0.1 miles
Cannery Row turns into David Ave 0.1 miles
Turn left onto Lighthouse Ave 1.2 miles
Turn slight right onto Tyler St 0.1 miles
Turn left onto E Franklin ST <0.1 miles
Turn right onto Washington ST 0.1 miles
Washington turns into Abrego St 0.3 miles
Arrive 850 Abrego St

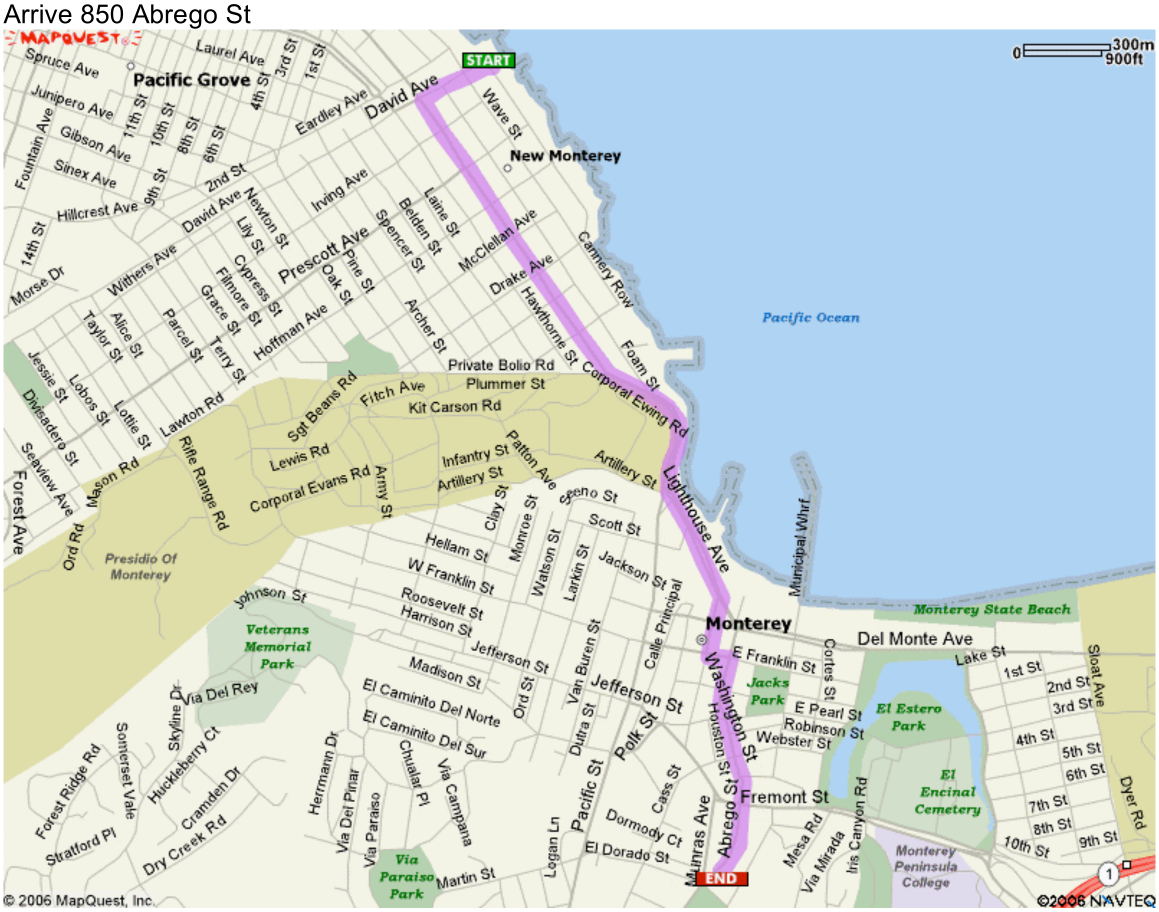

Figure 41: Aquarium to Day's Inn

Day 8 - Friday, 25 August 2006

Summary: Leave Day's Inn (bye bye nice soft bed)
Travel to June Lake Campground
View Yosemite along the way
End day at June Lake Campground

Directions: Days Inn to June Lake

Start out going Northeast on Abrego St toward Fremont St 0.1 miles.
Turn right onto Fremont St 0.5 miles.
Merge onto CA-1 N. 13.4 miles.
Merge onto CA-156 E via exit 414B toward Castroville / US-101 / San Jose 6.4 miles.
Merge onto US-101 N toward Hollister / San Francisco 16.6 miles.
Take the CA-25 S ramp toward Hollister 0.1 miles.
Turn left onto CA-25 / Bloomfield ave 0.6 miles.
Turn left onto Bloomfield ave / CR-G7 3.2 miles.
Turn right onto Pacheco Pass hwy / CA-152 E. Continue to follow CA-152 E 62.7 miles.
Take the CA-59 ramp toward Merced 0.2 miles.
Stay straight to go onto CA-59 15.0 miles.
Merge onto CA-99 S / CA-140 E 0.6 miles.
Take the CA-140 E exit toward Mariposa / Yosemite 0.2 miles.
Turn left onto Yosemite pkwy / CA-140 / CA-99 BR.
Continue to follow CA-140 35.7 miles.
Turn slight left onto CA-140 / CA-49. Continue to follow CA-140 E 37.0 miles.
Turn left onto Big Oak Flat rd 9.5 miles.
Turn right onto CA-120 / Tioga Pass rd (Portions may be closed seasonally) 58.6 miles.
Turn SLIGHT RIGHT onto US-395 S / CA-120 E. Continue to follow US-395 S 10.4 miles.
Turn SLIGHT RIGHT onto CA-158 / JUNE LAKE LOOP. 2.6 miles.
End at **June Lake Loop Recreation Area** Ca-158, June Lake, 6 hours 274.16 miles

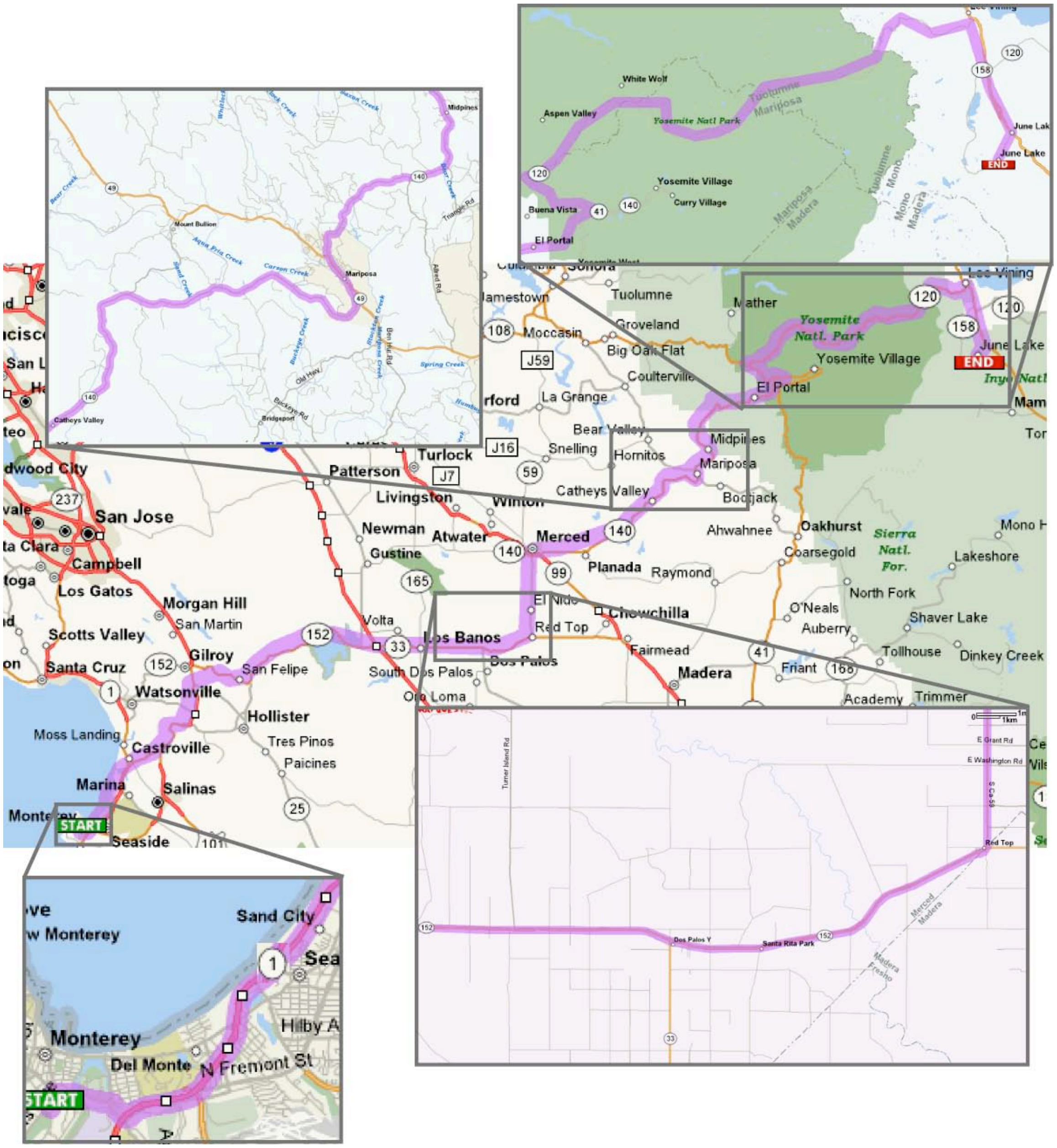

Figure 42: Days Inn to June Lake

8 - 1 Geology of Yosemite National Park -

Yosemite Valley (7 miles long and nearly 1 mile wide) is carved into the west slope of the Sierra Nevada. The park (figure 37) straddles both the physiographic Sierra Nevada and the Sierra Nevada batholith and enclosing metamorphic wall rocks (Huber *et al*., 1984). Immense cliffs, domes, and waterfalls tower over the area and the meandering Merced River below (figure 38). The formation of the valley only began during the late Cretaceous Period as the Sierra Nevada mountains were starting to build up (Matthes, 1924). The forces of erosion that have occupied the valley are evident in the highly complex assemblage of granitic rocks seen today. Many of the geological features seen today in Yosemite are still fairly young. The major landforms of the Yosemite Valley are due to erosion - first by streams and later by glaciers. The huge granite domes and deep gorges were shaped by glacial action, weathering, mass-wasting and stream erosion (Harris, 1980).

Figure 43 - Map of Yosemite National Park
z.about.com/d/gocalifornia/1/0/z/e/yosmap2.gif

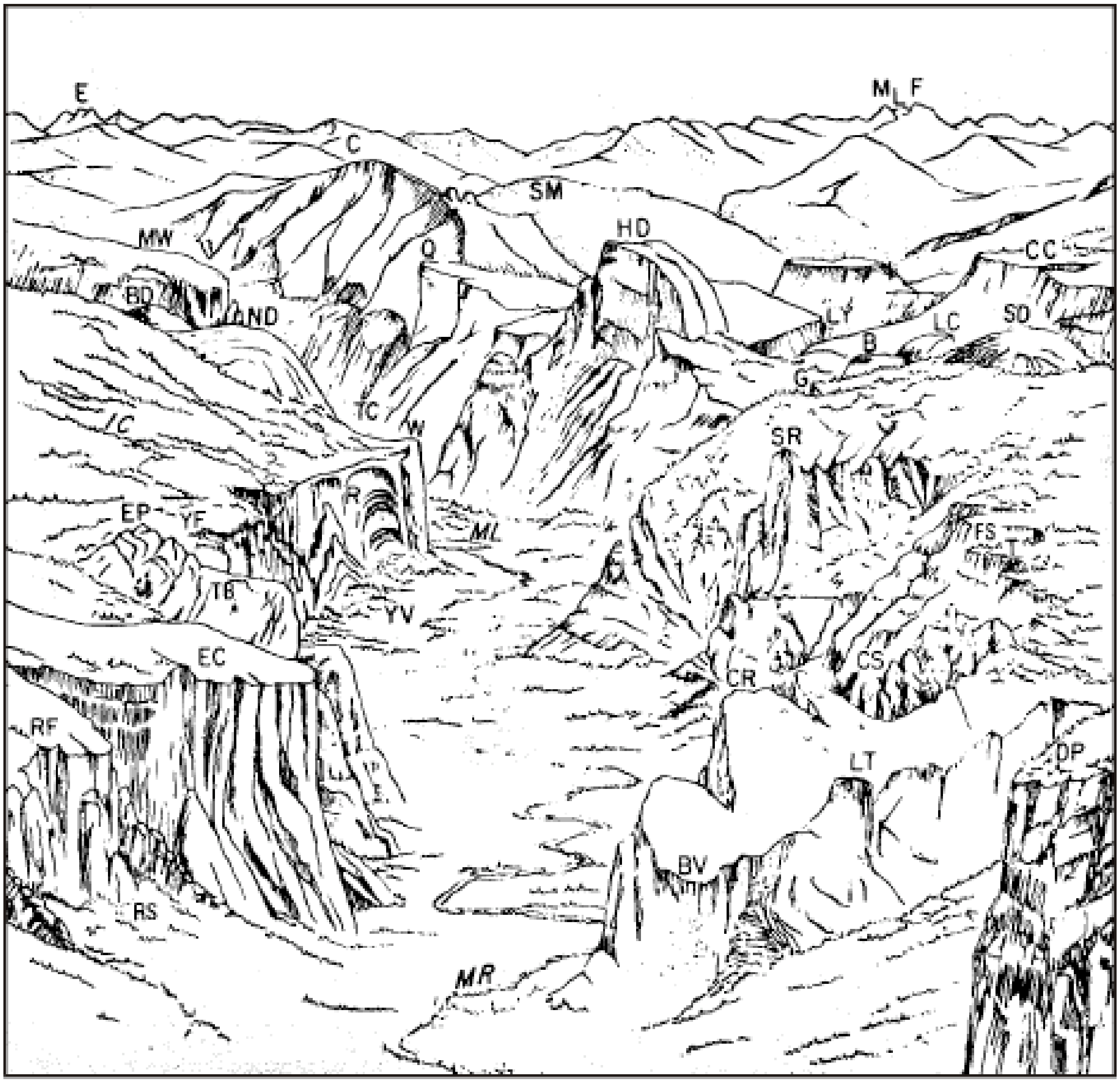

Figure 44 - Bird's-eye view of Yosemite Valley, with identification of selected land forms (Calkins, 1985, p. 2)

RS Rockslides
RF Ribbon Fall
EC El Capitan
TB Three Brothers
EP Eagle Peak
YF Top of Yosemite Falls
YV Yosemite Village
IC Indian Creek
R Royal Arches
W Washington Column
TC Tenaya Canyon
ML Mirror Lake
ND North Dome
BD Basket Dome
MW Mount Watkins
E Echo Peaks
C Clouds Rest
SM Sunrise Mountain
Q Quarter Domes
HD Half Dome
M Mount Maclure
L Mount Lyell
F Mount Florence
CC Cascade Cliffs
LY Little Yosemite
LC Liberty Cap
B Mount Broderick
SD Sentinel Dome
G Glacier Point
SR Sentinel Rock
FS Fissures
T Taft Point
CS Cathedral Spires
CR Cathedral Rocks
BV Bridalveil Fall
LT Leaning Tower
DP Dewey Point
MR Merced River

Formation of Yosemite Valley

The numerous plutons that compose the Yosemite Valley area and the Sierra Nevada, together are called the Sierra Nevada batholith (Peck *et al*., 1966). The complex history of the batholith was only deciphered when individual plutons were recognized as separate units and not variations in one huge rock mass. The plutons exposed in the walls of Yosemite Valley were intruded over a period of about 30 million years (approximately from 120 to 90 million years ago) during part of the Cretaceous Period. Because the Sierra Nevada batholith is composed of hundreds of individual plutons, construction of this batholith may have taken as much as 130 million years (Calkins, 1985). Table 1 provides a summary of the events involved in the formation of Yosemite Valley.

The great diversity of erosional features seen today has resulted from the varied composition and structure of rocks from which they were carved. In the case of plutonic rocks, the most significant structure that affects their erosional form consists of joints (sets of parallel fractures or cracks in the rocks). Generally the more siliceous, or quartz-rich rocks (granite and granodiorite) have more widely spaced joints than the less siliceous rocks (tonalite and diorite). Also, the finer-grained rocks generally have more closely spaced joints than the coarser-grained rocks. For example, the diorite of the Rockslides is the least siliceous of the plutonic rocks in the valley. Because of this extensive jointing, enormous talus piles of blocks of diorite have accumulated throughout the area of the Rockslides and at the base of Bridalveil Fall. In contrast, the siliceous rocks of El Capitan makes it massive and resistant to collapse.

The type of jointing that has most influenced the form of Yosemite's monuments is the broad shell-like unloading joints, or sheeting, referred to as *exfoliation* (e.g. Half Dome). As the plutonic rocks were uplifted into mountains and the overlying rocks eroded, the release or unloading of the previous pressure caused the rock to expand outward. The outermost layer, being exposed to the weather, gradually disintegrates and the layers fall off, similar to onion skins (figure 45) (Calkins, 1985). The process of sheeting eliminates sharp corners and angles and replaces them with curves. As sheeting continues, a smoother, more rounded surface evolves.

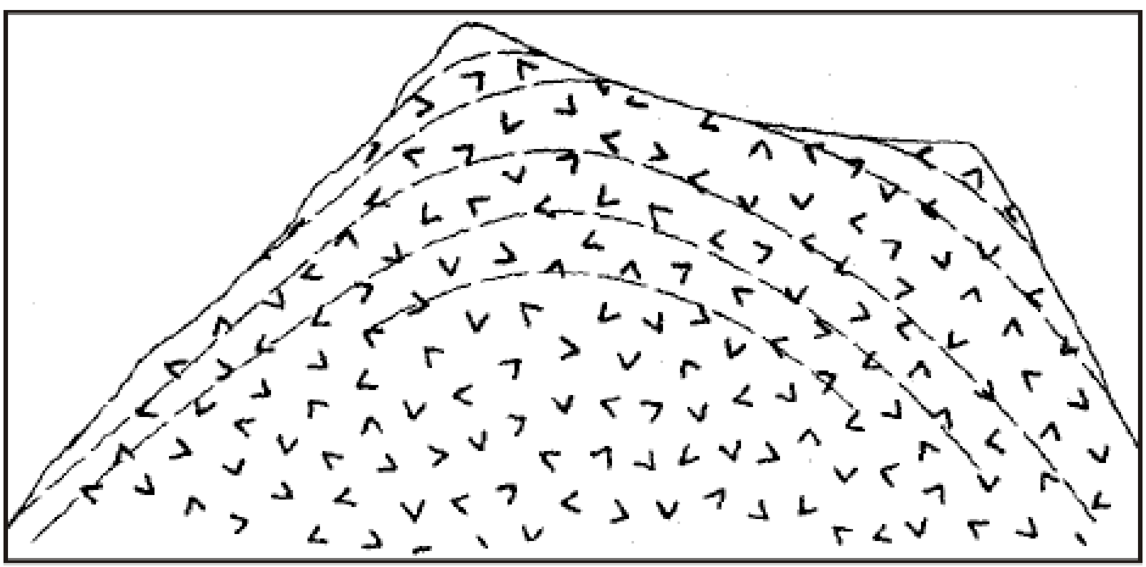

Figure 45 - Diagram illustrating progressive rounding (exfoliation) of massive granite as successive sheets peal off (Calkins, 1985, p. 6)

There are also several domes (e.g. Lembert Dome and Fairview Dome), which are elongated and asymmetrical and not formed by exfoliation but rather by glaciation. These have steep ends on the west and polished and striated eastern slopes (Harris, 1980). These domes were overridden and reshaped by glaciers. The up-glacier end was ground down by abrasion and the down-glacier end was quarried leaving a steep, stair-stepped cliff. These asymmetrical domes are called *roches moutonnées*. There are many other evidences of glacial action in the valley. Terminal moraines (e.g. from El Capitan across the valley to Cathedral Rocks) may be found throughout (Scharff, 1967). Other evidences include polishing or glacial striae and glacial boulders left by the melting glaciers.

Rocks of the Yosemite Valley From a distance, all of Yosemite's granite rocks look the same. But it actually consists of individual rock bodies, each with their own mineral composition and texture. It is all these variations, which affect the rock's resistance to abrasion, fracturing, and weathering. The plutons making up Yosemite Valley are composed chiefly of quartz-bearing granitic rocks (Peck *et al.*, 1966). Figure 40 shows a classification scheme for plutonic rock types found within the valley. The mineralogy of some of these intrusions is seen in table 2. Most of the granitic rocks are grouped into two series, the older Western Intrusive Series, and the younger Tuolumne Intrusive Series . Based on potassium-argon ratios of biotite from the granitic rocks, Cretaceous isotopic ages ranging from 95-84 million years have been determined for the plutons (Peck *et al.*, 1966).

Glaciers in Yosemite

Huber, King N., The Geologic Story of Yosemite Valley
http://www2.nature.nps.gov/geology/usgsnps/yos/topobk.html

The Yosemite landscape as we see it today strongly reflects the dynamic influence of flowing ice that long ago covered much of its higher regions. Geologists are still uncertain how many times ice mantled Yosemite, but at least three major glaciations have been well documented elsewhere in the Sierra Nevada. In the higher country, icefields covered extensive areas, except for the higher ridges and peaks. Lower down the western slope, at middle elevations, glacial tongues were confined to pre-existing river canyons, such as those of the Merced and Tuolumne Rivers.

In contrast to the sinuous V-shaped valleys of normal streams in unglaciated mountainous terrain, glaciated valleys tend to be straighter and have U-shaped profiles. Whereas a stream erodes the outsides of bends preferentially and makes its course more sinuous, glacial erosive force is concentrated on the insides of bends, removing the protruding spurs of the original stream valley and leaving a wider, straighter valley.

The resulting modification, in detail, depends on the nature and structural integrity of the bedrock over which the glacier is flowing. For granitic bedrock, the dominant structure of concern is jointing, which controls the ease of removal of rock that is otherwise highly resistant to glacial erosion.

A glacier tends to straighten a valley and smooth its walls as it grinds past them. But the walls of Yosemite Valley are extremely ragged, with many pinnacles and spires projecting upward from them—Leaning Tower, Cathedral Spires, Sentinel Rock, and Lost Arrow stand out strikingly.

Little doubt exists that Yosemite Valley indeed represents a profound, glacially-driven modification of the Merced River canyon, as no other erosive agent could have accomplished such excavation. A glacier filling the valley to its rim created the basic broad shape of the valley and gouged out a deep bedrock basin whose bottom locally, in its eastern part, lies more than 1,000 feet below the present valley floor. Today we correlate that glaciation with the Sherwin glaciation, defined from studies along the east side of the Sierra Nevada. The Sherwin was the most extensive, and longest-lived, glaciation documented in the Sierra. It may have lasted almost 300 thousand years and ended about 1 million years ago. A Sherwin-age glacier was almost surely responsible for the major excavation and shaping of Yosemite Valley within the Merced River canyon.

Later glaciations in the Sierra Nevada were of lesser aerial extent and briefer than the Sherwin. The best documented are the Tahoe and Tioga glaciations, which probably peaked about

130,000 and 20,000 years ago, respectively. The last glacier in Yosemite Valley—Tioga in age—advanced only as far as Bridalveil Meadow. At this location the forward movement of the glacier was balanced by the melting of ice at its front, or terminus. A “terminal” end moraine—a low ridge crossing the valley—was constructed with rock debris transported by the glacier and deposited at its terminus. Since the original excavation of Yosemite Valley by a Sherwin-age glacier, no subsequent glacier has filled the valley to its rim, a conclusion that has important consequences for the scenery.

From its terminus at Bridalveil Meadow, the ice surface of the Tioga glacier would have sloped upward toward the east end of the valley with the ice reaching a thickness of perhaps a little over 1,000 feet at Columbia Rock west of Yosemite Falls, 1,500 feet at Washington Column, and 2,000 feet in Tenaya Canyon below Basket Dome, as reconstructed by Matthes. Thus the Tioga and similar Tahoe glaciers could do very little to further modify or smooth the walls of Yosemite Valley. Above the ice surface of those glaciers, the valley walls have had a million years to weather: joints widened, rock fractured and crumbled, and waterfalls and cascades eroded back into alcoves and ravines. Thus the pinnacles and spires that seem so anomalous for a glacial valley actually had a million years to form and, being above the level of later glaciers, remain to amaze us today. In Tenaya Canyon, Tioga ice was thicker and reached farther up the walls, smoothing them and removing irregularities; no pinnacles and spires are found there.

Waterfalls leaping out from a valley’s walls far above the valley floor have long been considered evidence of a glacial origin for the valley. The enormous Sherwin-age glacier that shaped Yosemite Valley was able to excavate the central chasm to a greater depth than smaller glaciers in side-entering tributaries. The result was that some of the side valleys were left “hanging” with waterfalls at their brinks. Since Sherwin time, most of the tributaries have eroded their channels back into the walls to leave little more than steep ravines with minor falls interrupted by chains of cascades, such as those at Sentinel Fall. Bridalveil Fall is the single exception, although it also has receded back into an alcove from its original position further out on the valley wall.

While the Tioga glacier was constructing its terminal moraine at Bridalveil Meadow, the climate apparently warmed slightly. The ice at the front of the glacier began to melt faster than the ice was moving forward, and the ice front, or “snout,” of the still-flowing glacier began to “retreat” up the valley. The climate cooled again; the ice front paused and temporarily stabilized just west of El Capitan Meadow. Here the glacier began to construct a new moraine, known as a “recessional” moraine because the glacier had receded from its terminal position. It remained at this location longer than it had at Bridalveil Meadow and the resultant El Capitan Moraine is larger in both volume and height. Eventually, the climate warmed abruptly, and the Merced glacier’s snout retreated toward the head of the valley with no more recessional pauses, probably leaving Yosemite Valley by 15,000 years ago.

When the Tioga-age glacier departed from Yosemite Valley it left behind a lake, which Matthes christened Lake. It is likely that the advancing Tioga glacier had excavated some of the pre-existing valley fill east of the El Capitan Moraine, creating a shallow lake basin. The lake was in part dammed by this moraine, with the Merced River flowing over a low spillway through the moraine near the south valley wall. As the separate arms of the Tioga glacier retreated up the Merced and Tenaya canyons, the melt-water-swollen, debris-laden rivers issuing from their snouts delivered large quantities of sediment to the lake basin. The lake was soon filled in with this sediment, creating the relatively level valley floor we see today. The resulting gentle slope allowed the Merced River to develop a sinuous meander pattern across this broad flood plain. A low-gradient, meandering stream is particularly susceptible to over-bank flow during high water, and its flood plain is naturally destined for periodic flooding.

TUOLUMNE INSTRUSIVE SUITE and TENAYA LAKE

(From Coleman, et al., 2004)

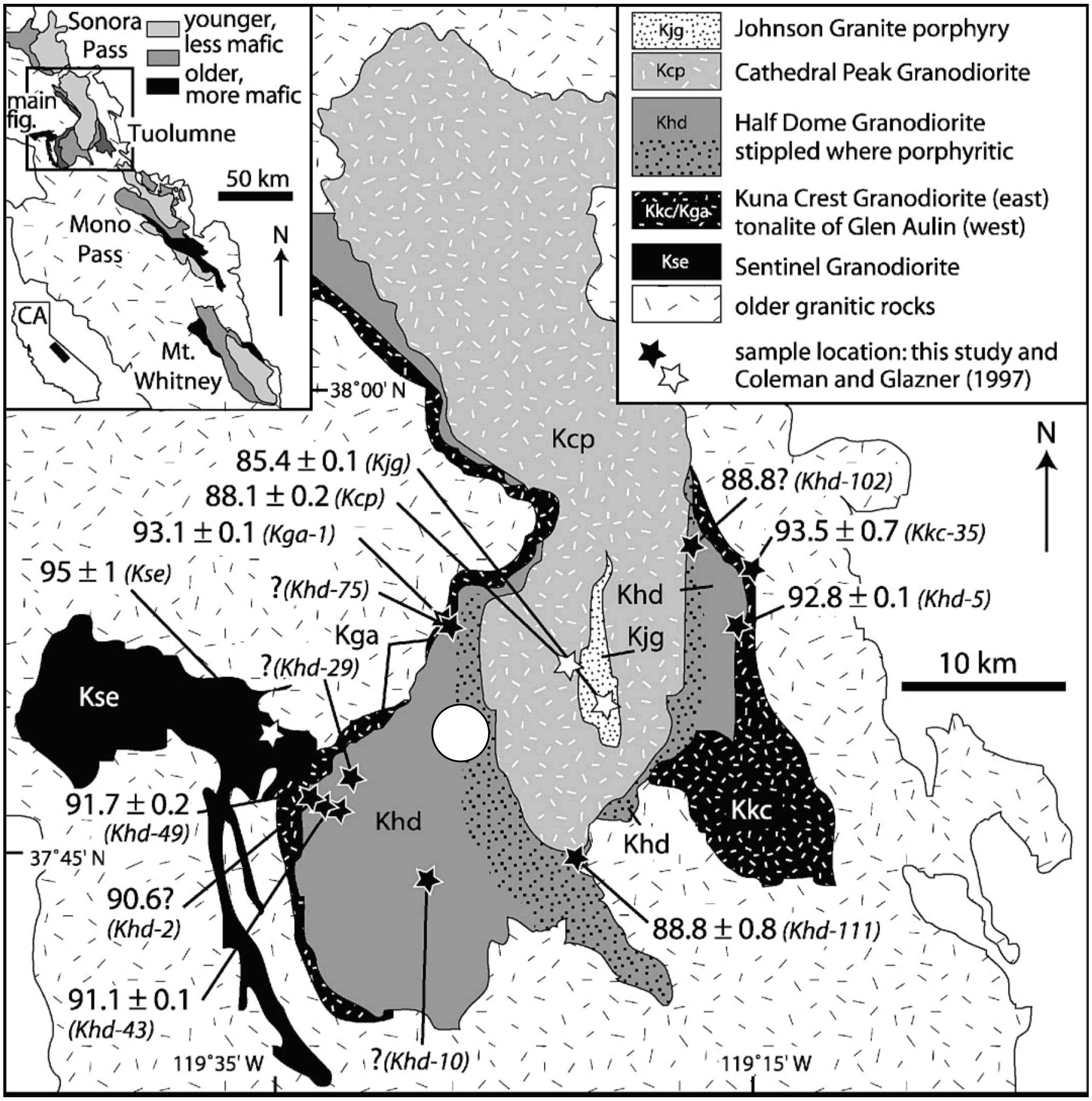

Figure 46. Simplified geology of Tuolumne Intrusive Suite after Bateman (1992). Inset shows location of Tuolumne and other Sierra Crest zoned intrusive suites. Tenaya Lake is located at the western contact between the two Half Dome units, marked with the large white circle (not to scale).

The 1200 km2 Tuolumne Intrusive Suite is one of several large-volume zoned intrusive suites emplaced in the Sierra Nevada in the Late Cretaceous. These suites all have mafic granodioritic outer phases that grade progressively inward to granodioritic or granitic cores (Bateman, 1992), and were intruded at pressures of 1–3 kbar (Ague and Brimhall, 1988). From rim to core, the Tuolumne Intrusive Suite includes the granodiorite of Kuna Crest and the tonalite of Glen Aulin (assumed to represent the eastern and western parts of the same intrusion; Bateman and Chappell, 1979), the Half Dome Granodiorite, the Cathedral Peak Granodiorite, and the Johnson Granite Porphyry (Fig. 1). Kistler and Fleck (1994) also considered the Sentinel Granodiorite on the west side of the suite to be part of the Tuolumne. Rocks of the Tuolumne Intrusive Suite are commonly interpreted to be cogenetic. Both fractional crystallization of exposed units (Frey et al., 1978; Bateman and Chappell, 1979) and mixing between them (Reid et al., 1983; Kistler et al., 1986) have been invoked to explain the chemical evolution of the suite.

New U-Pb geochronologic data indicate that the Tuolumne Intrusive Suite, California, was assembled over a period of at least 10 m.y. between 95 and 85 Ma, and that the Half Dome Granodiorite intruded over a period approaching 4 m.y. Simple thermal considerations preclude the possibility that a magma chamber the size of the Half Dome pluton could have existed as a liquid at shallow crustal depths for that long. Rather, field evidence for sheeting along the margins of the suite, the range of ages, and the regular decrease of ages toward the center of the suite and within individual plutons suggest incremental assembly. Geochronologic evidence for incremental assembly is consistent with the failure of geophysical methods to detect large magma chambers with more than 20% melt, even in active volcanic areas. Because it is unlikely that the individual plutons composing the Tuolumne ever coexisted as liquid-rich magmas, the chemical evolution of the suite cannot be the result of simple fractionation and/or mixing between exposed units, but instead must reflect processes occurring during magma generation.

Tenaya Lake is located within the Half Dome Granodiorite of the Tuolumne Intrusive Suite, a medium to coarse grained and contains well-formed plates of biotite and crystals of hornblende. This rock characteristically disintegrates by sheet jointing (exfoliation) and it forms nearly all the domes of the valley area. Tenaya Lake was created by the Tenaya branch of the Tuolumne Glacier as it passed through Tenaya Canyon. The outflow of the lake is Tenaya Creek, which runs through Tenaya Canyon into Yosemite Valley. Glacial activity created some of the most spectacular polished rock on earth! (Huber, 1987).

Period	Epoch	Event	Approximate Duration in Years
Quaternary	Recent	Postglacial time. Return to normal climatic conditions. Lake Yosemite formed and filled in, forming present level valley floor.	20,000
	Pleistocene	The Great Ice Age. Second series of uplifts pushed Sierra Nevada up to its present height of over 14,000 ft. Yosemite Valley invaded three times by glaciers.	1 to 2 million
Tertiary	Pliocene	Period of relative stability in which the Merced River developed a rugged mountain valley more than a 1000 ft in depth.	10 million
	Miocene	Volcanic eruptions begin, major series of uplifts caused the Sierra Nevada to stand out as a block range as eastern edge rose and western edge steepened. Merced River accelerated as a result.	12 million
	Oligocene/ Eocene	The region is slowly upwarped. Volcanic activity in the north part of the range. Continued land erosion. Birth of Merced River.	40 million
Cretaceous		Mountain ranges gradually worn down and bulk of sedimentary rock carried away by streams, uncovering the underlying granite.	75 million
Jurassic		Continued deposition of sediments on ocean bed, followed by folding of strata into northwestward-trending mountain ranges. Intrusion of molten granite into the folds from below.	40 million
Triassic		Mountains worn down to hills and land sinks below the sea. New sediments deposited.	40 million
Carboniferous	Permian	Sediments uplifted and folded into mountain ranges of slate, shale, sandstone, and limestone.	415 million
Pre-Carboniferous		Sediments accumulate on the floor of the Pacific Ocean.	

Figure 47: Sequence of Events in the Yosemite Region (modified from Beatty, 1960, p. 34)

Series	Name	Lithology
Tuolumne Intrusive Series	Johnson Granite Porphyry	Light-gray, fine-grained quartz monzonite porphyry containing a little biotite but no hornblende
	Cathedral Peak Granite	Light-gray quartz monzonite containing abundant large phenocrysts of K-feldspar in a medium-grained groundmass containing both biotite and hornblende
	Half Dome Quartz Monzonite	Light- to medium-gray granodiorite and quartz monzonite containing well-formed crystals of biotite and hornblende. Includes both a nonporphyritic and a porphyritic facies, the latter resembling Cathedral Peak Granite but containing better formed biotite and hornblende.
	Sentinel Granodiorite	Medium-dark-gray granodiorite and quartz diorite, variable in color and texture, typically well foliated
Minor Intrusive Series	Diorite of the 'Map of North America'	Very dark-gray diorite, similar to the diorite of the Rockslides but finer grained
	Quartz-mica diorite	Medium-dark-gray, medium-fined-grained quartz mica diorite
	Leaning Tower Quartz Monzonite	Medium-gray, medium-grained granodiorite, clusters of biotite and hornblende give the rock a speckled appearance
Western Intrusive Series	Taft Granite	Very light-gray, fine-grained quartz monzonite, finer-grained and more uniform than the El Capitan Granite
El Capitan Granite		Light-gray, medium-coarse-grained biotite quartz monzonite and granodiorite. Vaguely porphyritic in part
Granodiorite of the Gateway		Dark-gray, medium-grained quartz diorite and granodiorite
Granite of the Arch Rock		Medium-light-gray, medium-grained, biotite quartz monzonite and granodiorite, contain characteristic poikilitic K-feldspar grains
Diorite of the Rockslides		Very dark-gray, coarse- to medium-grained diorite, quartz diorite, and gabbro, very variable

Figure 48: Granitic rocks of the Yosemite Valley region (Peck et al., 1966, p. 489).

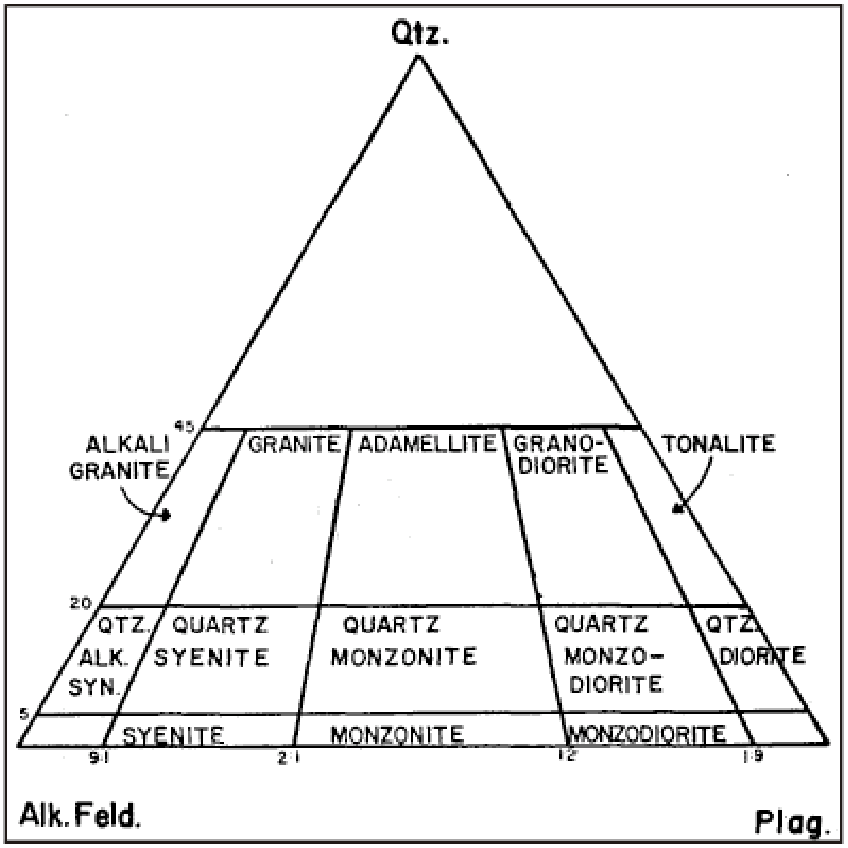

Figure 49 - Classification of Plutonic Rocks (modified from Robinson, 1979, p. 39)

Suggested Readings

Beatty, M.E. 1960. A Brief Story of the Geology of Yosemite Valley. Yosemite Nature Notes, v. 22, pp. 33-40.

Calkins, F.C. 1985. Bedrock Geologic Map of Yosemite National Park, California. Miscellaneous Investigation Series - U.S. Geological Survey, 1 sheet, 7p.

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Matthes, F.E. 1924. The Story of the Yosemite Valley. New York: American Museum of Natural History, Guide Leaflet No. 60, pp. 1-21.

Peck, D.L., Wahrhaftig, C., and Clark, L.D. 1966. Field Trip: Yosemite Valley and Sierra Nevada Batholith. In: Geology of Northern California, pp. 487-502.

Robinson, J.R. 1979. Petrology and Petrochemistry of Granitic Intrusives of the Cima Dome - Southern Ivanpah Mountains Area, Southeastern California. Los Angeles: University of Southern California (Masters), 125p.

Scharff, R. 1967. Yosemite National Park. New York: David McKay Company, Inc., 213p.

Day 9 - Saturday, 26 August 2006

Summary: Leave from June Lake Campground
Conway Overlook
Mono Lake Visitors Center
Mono Lake Tufa Spires
Panum Crater
Tertiary basalts at Gaspipe Spring
Taylor Canyon obsidians (Glass Mtn)
Aeolian Buttes
Return to June Lake Campground

Volcanism at Long Valley Caldera, Eastern California

Synopsis of Cenozoic Volcanic Activity in the Long Valley Area: Long Valley caldera is located at the western edge of the Basin and Range Province, straddling the eastern escarpment of the Sierra Nevada (Figure 50). Here, the bounding fault of the Sierra Nevada is offset, forming the Mammoth Embayment. Volcanism has been focused in the Long Valley area for the past 3.6 Ma, probably as a result of this embayment. Prevolcanic basement rocks include Mesozoic granitoids of the Sierra Nevada plus Paleozoic metasedimentary and metavolcanic rocks trapped between plutons (roof pendants).

Initial volcanic activity in the Long Valley area began with widespread eruptions of trachybasalt-trachyandesite lavas between 3.6 and 2.2 Ma. Erosional remnants of these lava flows are scattered discontinuously over a 4000 km2 area around the caldera. The distribution of these initial mafic eruptions suggests a laterally extensive mantle source. Slightly younger rhyodacite domes and flows associated with these mafic eruptions were emplaced along the northwest rim of the present caldera between 3.2 and 2.6 Ma. These domes probably represent the first products from the Long Valley magma chamber from which subsequent more silicic eruptions originated. The first eruptions from this highly differentiated chamber were on the northeast rim of the present caldera at Glass Mountain, where 1000 m of high-silica rhyolite domes, flows and tuffs accumulated between 2.1 and 0.8 Ma.

Catastrophic rupturing of the roof of the magma chamber at 0.73 Ma expelled at least 600 km3 of rhyolite magma as Plinian ash falls and ash flows. This partial emptying of the chamber caused collapse of its roof to form the 2-3 km deep oval depression of Long Valley caldera. The resulting ash flow deposits, the Bishop Tuff, inundated 1500 km2 around the caldera and accumulated to thicknesses as great as 200 m. Plinian ash clouds drifted thousands of kilometres downwind and deposited an ash layer as far east as Kansas, and south throughout California and into the eastern Pacific Ocean.

After collapse of the chamber, continued volcanism on the caldera floor produced 100-500 m of tephra deposits and fluid, aphyric rhyolite flows termed the Early Rhyolite. Simultaneously, pressure from below arched and faulted the caldera floor, forming a resurgent dome with a northwest-oriented medial graben. The resurgent dome formed within 100,000 years of caldera collapse, based on the 730,000-650,000 year ages of the Early Rhyolites. A quiescent period of 100,000 years ended with the eruption of the coarsely plag-qtz-san-bio-hbl-phyric Moat Rhyolites, which form thick, steep domes and flows that ring the resurgent dome. These rhyolites, erupted at 500, 300, and 100 ka in clockwise succession around the resurgent dome, probably signal the onset of cooling and crystallization of the magma chamber. Several studies indicate that there is still a body of partially molten magma beneath the resurgent dome.

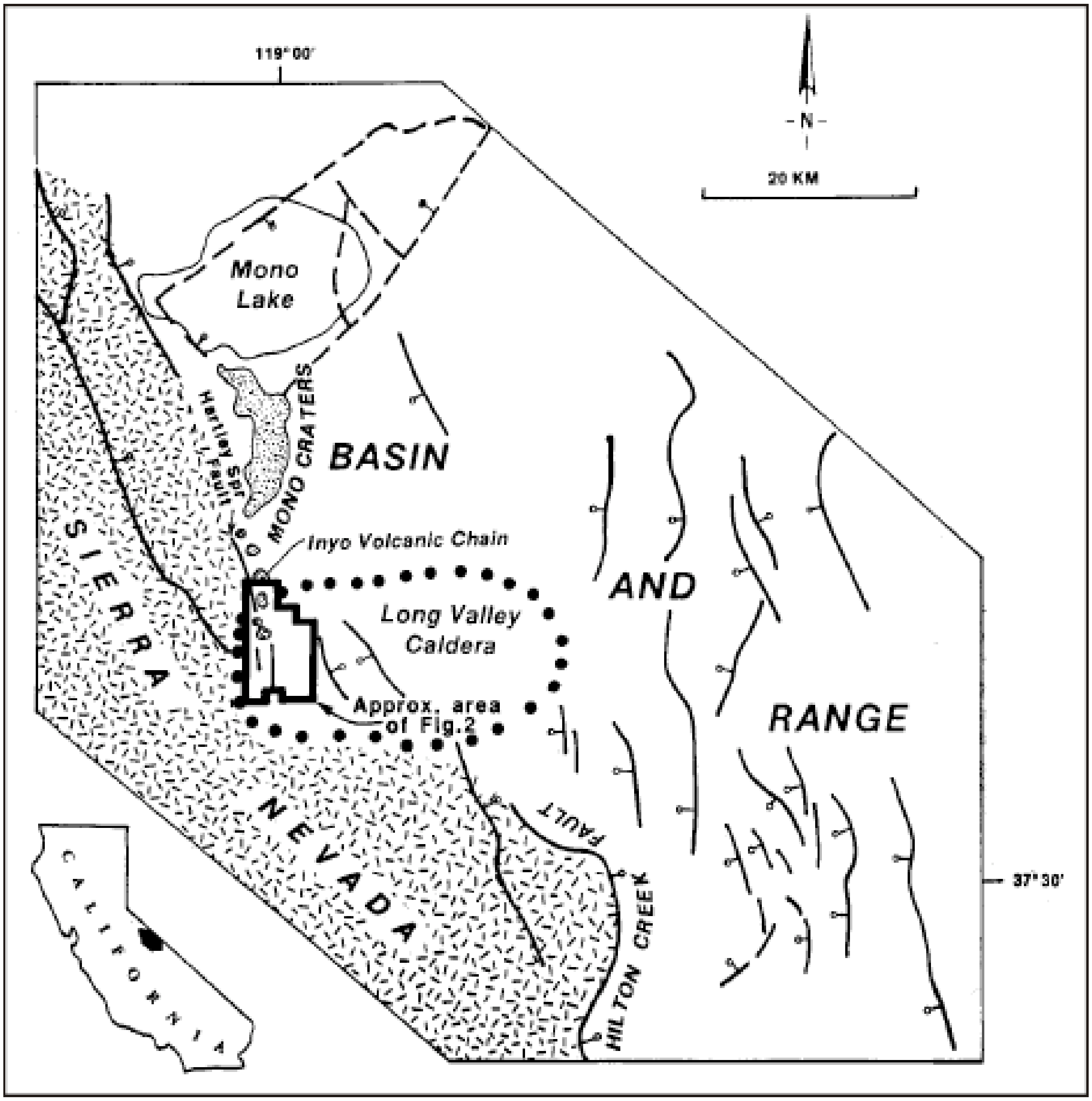

Figure 50 - Map of Long Valley Region (Suemnicht and Varga, 1988)

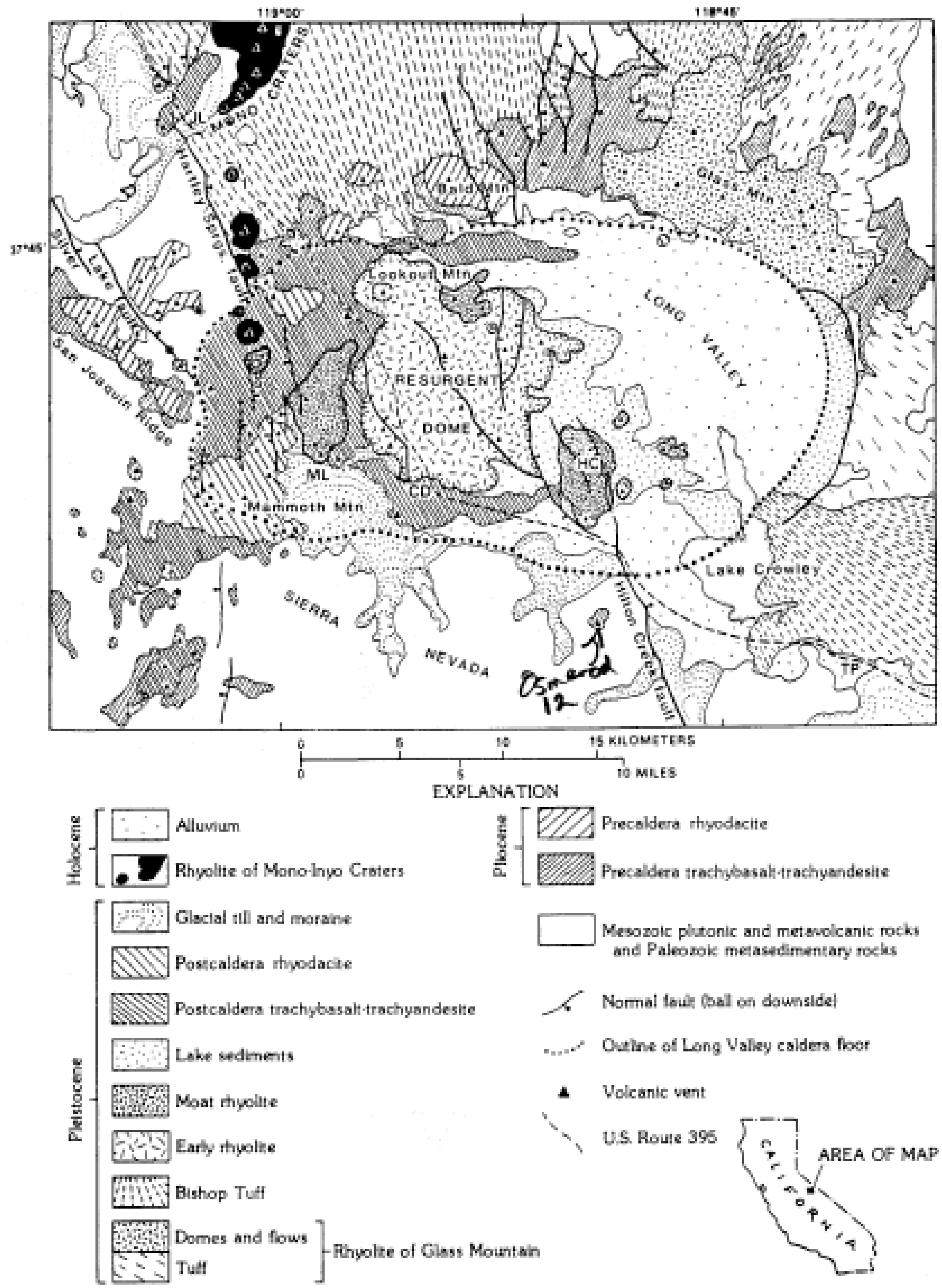

Figure 51 - Geological map of Long Valley Caldera. JL, June Lake; ML, Mammoth Lakes; CD, Casa Diablo; HC, Hot Creek; TP, Toms Place (Chapin and Zidek, eds., 1989)

Mafic magmatism resumed around the margins of the caldera around 450,000 years ago. Vents along the west edge of the caldera sent trachybasaltic-trachyandesitic lavas into the west moat which accumulated to at least 250 m in thickness. Tongues of mafic lava flows extend into the north and south moats. Mafic volcanic rocks and a rhyodacite were erupted within the drainage of the San Joaquin River southwest of the caldera, in what is now the Devils Postpile National Monument. Mafic magmas also erupted in the June Lake area between 40,000 and 20,000 years ago, at Black Point (Mono Lake) approximately 13,000 years ago. Concurrent with mafic activity, rhyodacite domes were emplaced sporadically within the western part of the caldera, including the Mammoth Mountain.

The locus of volcanism then shifted to the Mono-Inyo fissure system, which extends from the west moat north to Mono Lake. The Mono Craters consists of a 17 km-long, arcuate chain of 30 or more coalesced rhyolite domes, flows and craters ranging in age from 35,000 to 600 years in age. They are composed largely of high-silica most recent activity is concentrated at the north end of the chain, at Panum Crater and Negit Island. The Inyo Craters chain forms a 12 km-long, discontinuous line of mainly low-silica rhyolite domes, flows and craters erupted between 6000 and 500 years ago. Obsidian, Glass Creek, and Deadman Creek domes were fed by a shallow, 8 km-long rhyolite dyke, within which 2 distinct magma types mingled: a light-coloured, coarsely porphyritic, pumiceous rhyolite and a more silicic, sparsely-phyric rhyolite. Rocks from Glass Creek Dome are spectacularly banded in a “marble-cake” fashion. Inyo volcanic activity was accompanied by phreatic eruptions on the north flank of Mammoth Mountain and near Deer Mountain. Note that both the Inyo and northern Mono systems were simultaneously active around 600 years ago.

Recent Seismicity and Ground Deformation. Based on the moat-rhyolite periodicity of 200,000 years and the most recent eruption of 0.1 Ma, future eruptions from the residual Long Valley magma chamber would seem only a remote possibility, not to be expected for another 100,000 years. However, an unusual sequence of magnitude-6 earthquakes in May 1980 and another of magnitude-5 earthquakes in January 1983, accompanied by 20 in (50 cm) uplift of the resurgent dome, suggested that new magma was being injected into the chamber and possibly into the south moat ring fracture zone. While the intensity of seismicity and the rate of uplift declined through 1985 and 1986, the potential for future activity remains (Hill and others, 1985). However, statistically, the more likely site for future eruptions is the Mono-Inyo Craters volcanic chain, where, on the basis of 14C and obsidian hydration-rind dating (Wood, 1977), eruptions have occurred at about 500 year intervals for the past 3,000 years, and where the last eruption was 550 years B.P. Miller and others (1982) have outlined the nature of the potential hazards associated with possible future eruptions in the area.

See Figure 51 for a description of the geological features of the Long Valley area.

<u>9-1 Conway Summit Overlook and the Mono Basin</u>.

June Lake to Hwy 395, north to Conway Summit Overlook

From this vantage point, the entire Mono Basin and most of its geological features can be seen. In the centre of the basin lies Mono Lake, the saline remnant of a much larger Pleistocene glacial lake known as Lake Russell. The surface elevation of Mono Lake is at about 1941 m, and up until recently was dropping at a rate of about 1 m per year as a result of diversion of most of its freshwater inflow to the city of Los Angeles. At its high point of 2188 m, Lake Russell once lapped against the foot of the granite slope immediately below the overlook. To the west are the subdued, flat, terraced lava flows of the Adobe Hills, the earliest volcanism associated with the Long Valley system, and older Tertiary andesites. On the near shore of the lake is Black Point, and in the centre of the lake are Negit (dacite) and Paoha (uplifted sediments) Islands that are part of the Mono Craters chain that continues south from the opposite shore. To the left of the Mono Craters domes, the skyline is formed by the precaldera basalts, rhyodacites, and rhyolites of the north rim of Long Valley caldera. Beyond the Mono Craters are the high peaks that form the south rim of the caldera, and the steep escarpment immediately west of Mono Lake is the Lee Vining Fault, one of the Sierran bounding faults.

Mono Craters Mono Craters is a misnomer. They are not true craters, but plug domes. Composition of the pumice, lapilli, and flow material is rhyolitic to dacitic, remarkably similar to the flow material in the Big Obsidian Flow in Newberry Crater and the Rock Mesa Flow on the slope of South Sister. Chunks grade from black, glossy obsidian to gray, cellular pumice, whipped to a froth by the highly charged gasses that were contained in the flows. Flow banding is a prominent feature of the solid material. The large piles of flow material that escaped the domes and spread out over the surface are known as coulees. The Forest Service has erected a sign along the highway behind the domes to help explain the phenomena.

Figure 52 - Looking southwest over Mono Craters. Note the plug-in-crater pattern shown clearly crater (left center) (From Norris and Webb, 1990, p. 105, Fig. 3-27; Photo by John Shelton)

9-2 Lee Vining Mono Lake Visitors Center

Hwy 395 south ~ 10 miles to Lee Vining , pull into Mono Lake Visitor Center parking lot.

The Visitors Centre includes many displays concerning the Mono Lake ecosystem and the geology of the Long Valley area.

9-3 Mono Lake Tufa Towers Preserve

Return to Hwy 395, ~5 miles south to Hwy 120, turn left, proceed 4.8 miles, turn left to Mono Lake Tufa Towers Preserve

Mono Lake Tufa Towers: The delicate, knobby white towers along the shore of Mono Lake (Figure 53) are calcareous sublacustrine spring deposits formed where freshwater springs percolate through lake-bottom sediments and flow up into the highly saline water. Calcium in the freshwater combines with carbonate in the saline waters to precipitate calcite or aragonite. The towers formed entirely underwater, and are now exposed due to the dramatic decline in lake level since the 1940's. Many towers have fallen after being exposed. Signs along the trail discuss many other aspects of the ecology of Mono Lake.

Mono Lake: Mono Lake is 13 miles long, 9 miles wide, and approximately 170 feet deep. The lake has no outlet, and its present level is about 700 feet below the Tahoe stage drainage into Adobe Valley. Mono Lake water has about 6% dissolved solids with the composition given in Figure 54. The tufa towers of Mono Lake are porous limestone pinnacles located at present and former lake levels. The tufa forms about the orifice of sublacustrine springs. The dominant mechanism for tufa formation is the chemical difference between the spring and lake waters. However, a other studies show that algae play dominant role in the precipitation of the tufa. Photosynthetic withdrawal of carbon dioxide lowers the solubility of calcium carbonate close to the plants. Their appearance above water demonstrates the recession of the water level of the lake, a matter of serious concern to environmentalists and conservationists.

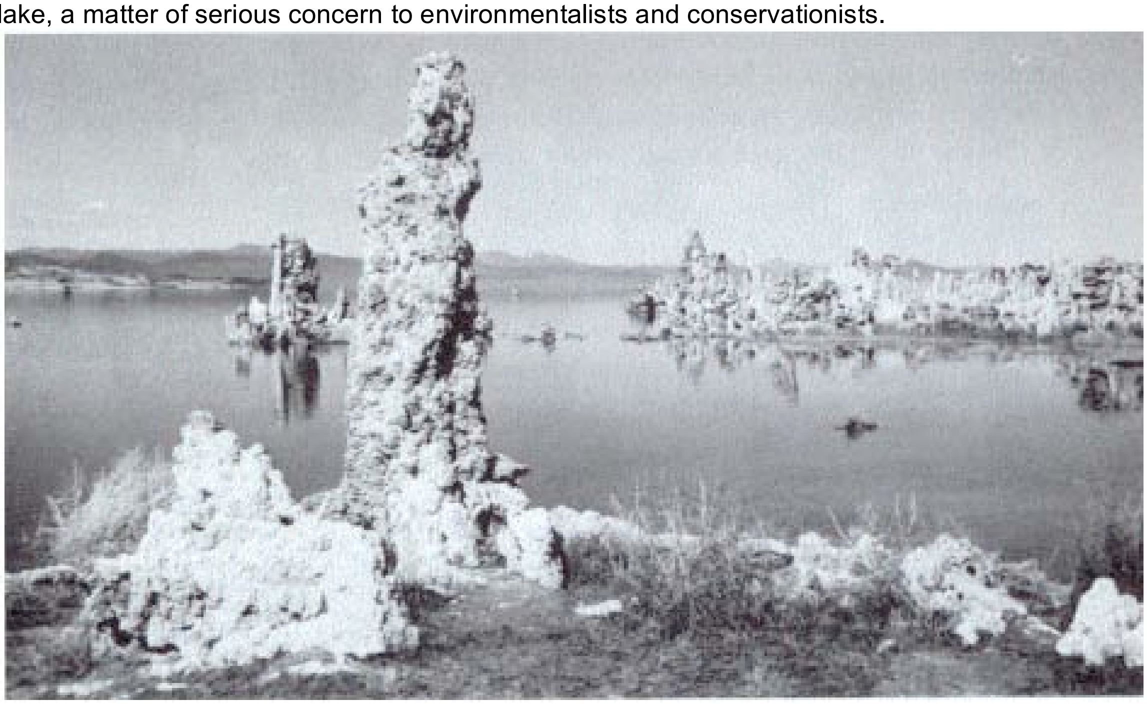

Figure 53 - Tufa Towers on the southwestern shore of Mono Lake, which were produced by lime-secreting cyanobacteria (blue-green algae), from springs bubbling up from the lake bottom (Norris and Webb, 1990, p. 122, Fig. 3-39; Photo by R. Norris)

Composition	mg/L	Composition	mg/L
SiO_2	14	HCO_3	5230
Fe	0.6	CO_3	11200
Ca	4.3	SO_4	7870
Mg	37	Cl	14500
Na	21700	F	44
K	1150	Br	35
NH_4	0.9	I	6
		NO_2	0.04
		NO_3	16
PH = 9.7		B	350

Figure 54: Chemical composition of Mono Lake; From Whitehead and Feth, 1961 (as listed in Hildebrand and Gall, 1983).

Another unique product of the lake is thinolite. It is found on the lake's northwest shore in abundance. It is a pseudomorph, or replacement, of some former completely unknown crystal, caused by some ancient chemical reaction. The crystals look somewhat like marine tree coral with spiny irregular points that divide and redivide again. They vary in color from beige to gray, with an occasional yellow specimen. Thinolite is known in but few places throughout the world. Investigations of the bathymetry of Mono Lake show that its eastern half is a smooth lake plain and the western part has two deep basins with much irregular topography, due in part to faulting, volcanism, glaciation, and submarine slumping. Most of the lake floor relief, including the islands, was formed in Holocene times. Mono Basin is a roughly triangular depression about 30 miles long in a northeasterly direction and 20 miles wide along its southwest base, which is part of the eastern front of the Sierra Nevada. The floor of the basin rises from 6400 feet at Mono Lake to 7000 feet around the margins. Its northern wall is a low escarpment 1000-2000 feet high in the late Tertiary volcanic rocks of the Bodie Hills. Granitic rocks are exposed in part of this escarpment at Conway Summit. Its southwestern wall is the rugged east face of the Sierra Nevada, as much as 6500 feet high. This wall has several steps and one offset where the southwest end of the Mono Basin extends about 4 miles west into the Sierra. Furthermore, this wall is cut by several deep glaciated canyons. The southeast border of the basin is indistinct, the land rises gradually from the basin floor to the Glass Mountain Bald Mountain ridge. Locally, as on the north side of Cowtract Mountain, this border is marked by a fault scarp 1000 feet high (Wahrhaftig et al., 1965). The Mono Basin has been the site of volcanic activity periodically from mid-Pleistocene to the Recent. The mountains bordering the basin on the north, east, and south are predominantly Tertiary and Quaternary volcanics. On the north shore of Mono Lake, directly below is Black Point, a subaqueous cinder eruption whose internal layering is nearly flat. The two islands in the lake both have cinder cones and flows (the northern of the two, Negit, consists entirely of volcanics). The lack of shore features or calcareous tufa on their surfaces indicates that they erupted after the lake fell to its present level or lower (Wahrhaftig et al., 1965). The escarpments bordering the Mono Basin are fault scarps, and most of the faults are probably still active Terminal moraines of the Mill Creek (Lundy Canyon) glacier, at the northwest corner of the Mono Basin, are offset as much as 50 feet along the Sierra Nevada boundary fault, while the moraines of the Lee Vining Creek glacier, which cross this fault about 7 miles farther southeast, are not offset (Wahrhaftig et al., 1965). During the Pleistocene, glaciers advanced beyond the mouths of the canyons of the Sierra Nevada as narrow tongues which were bordered by lateral moraines as much as 500 feet high. Contemporaneous with the glacial maxima were high lake stands in Mono Basin. The highest stand, correlated with the Tahoe glaciation, is marked by indistinct and eroded beaches at an altitude of 7170-7180 feet, when the lake was more than 900 feet deep. At this altitude, the lake could have overflowed through a low pass on the east side of its basin into Adobe Valley and thence into the Owens River system. The narrowness of the floors of canyons along the overflow route suggests that the discharge of water that reached the Owens River was probably small. The little-eroded beaches corresponding to the high lake level of the Tioga glaciation reach an altitude of 7070 feet and the lake did not overflow. These shorelines cut

Tahoe moraines, and Tioga meltwater built deltas into the Pleistocene lake to correspond to them. Putnam has recognized about 36 younger shorelines and has correlated the more prominent (at altitudes of about 6800 feet) with groups of prominent recessional moraines in the valleys of Lee Vining and Rush Creeks (Wahrhaftig et al., 1965). Recent geophysical studies suggest that the central part of the Mono Basin is a structural graben. However the amount of sedimentary and volcanic fill in the basin is debated. Some researchers contend from geologic data that the basin is merely a shallow downwarp, whereas others consider it to be a deep volcano-tectonic depression.

9-4 Panum Crater

Return to Hwy 120, turn right, go ~2 miles to gravel road beyond the one marked Panum Crater, and turn right, drive 0.8 miles to parking lot

Panum Crater is the youngest vent in the north Mono chain. The deposits emplaced during this activity include (oldest to youngest): (1) a vent-clearing breccia, (2) four pyroclastic flow and surge deposits, (3) a tephra ring deposit, and (4) a composite, crater-filling dome. Two episodes of activity are suggested. The initial vent-clearing eruption produced an early tephra ring and crater, from which the pyroclastic flows were erupted. An early dome rose into the crater and then apparently collapsed, breaching the northwest side of the tephra ring and sending a block avalanche into Mono Lake. Subsequent less-energetic pyroclastic events healed the breach in the tephra ring, and finally the modern dome was extruded. The dome consists of three subunits, the eastern and southwestern of which are thinly mantled with tephra. The northern part of the dome is the most recent extrusive event (Figure 52).

9-5 Tertiary Basalts at Gaspipe Springs

Drive to Hwy 120, turn left, proceed past Panum Crater Mono Mills, Big Sand Flat to Gaspipe Spring

Exposed here are late Tertiary basaltic lavas that pre-date the formation of the felsic Long Valley magmatic system. These lavas are well exposed north of Mono Lake, forming the Adobe Hills. The lavas are mildly alkaline, trachybasalts that are thought to be derived by melting of the lithospheric mantle during initial stages of Basin and Range extension. If so, these lavas are relatives of the Big Pine volcanic field basalts that we will see later in the trip.

9-6 Taylor Canyon Obsidians at Glass Mountain

Continue east on Hwy 120 to Adobe Valley, look for Taylor Canyon Rd. Proceed SW towards Symos Road (~3miles), at intersection turn left and proceed up Tayler Canyon.

Exposed in Taylor Canyon are obsidian lavas of Glass Mountain. Glass Mt represents the initial extrusive volcanic activity of the Long Valley caldera system. The lavas are high silica rhyolites (75% SiO_2), similar to the Bishop Tuff and the Mono Craters. Glass Mt lavas are approximately 2 Ma in age, although dating of K-feldspars and the glassy obsidian matrix has shown that these magmas may have an extensive history. These obsidians are unique to the area in that they have streaks of brown obsidian mixed in with black obsidian, a so called mahogany obsidian texture. This locality is also special in that the obsidians are beginning to devitrify and form spherulites. Pay particular attention to where the spherulites are forming in the obsidian, and can you suggest an origin for spherulites?

9-7 Aeolian Buttes

Retrace Hwy 120 to Hwy 395, turn left. Proceed ~ 1.0 mile to dirt road on left, drive 1.5 miles toward Aeolian Buttes

The Aeolian Buttes are the eroded remnant of a northward-oriented lobe of the Bishop Tuff. At this locality, the Bishop Tuff is densely welded, and glassy fiamme have begun to devitrify. Note the colour of the Bishop Tuff where it is densely welded, and compare it to the colour of the Bishop Tuff where we see it elsewhere on Day 11. This is a great sampling locality for the Bishop Tuff.

Return to Hwy 395, turn left, return to June Lake Campground

Day 10: Sunday, 27 August 2006

Summary: Leave June Lake Campground
Minaret Vista
Devils Postpile Hike
Post-caldera Quaternary mafic basalts at CA203-Hwy395 Intersection
Inyo Craters
Return to June Lake Campground

10-1 Minaret Vista

June Lake to Hwy 395, turn right.
At CA-203, turn right towards Mammoth Lakes.
Follow CA-203 ~ 6.4 miles to Minaret Vista parking area

Minaret Vista: Sierran Cretaceous basement and Tertiary vocanic rocks

The view west from Mineret Vitsa overlooks the deep, glaciated canyon of the middle fork of the San Joaquin River and the Ritter Range beyond, the most prominent peaks of which, viewed from south to north, are the jagged Minarets, massive Mt. Ritter, and Banner Peak. The rocks of the Ritter Range include coarse metavolcanic breccias, welded tuffs, and associated granitic plutons that are part of a deeply dissected Cretaceous caldera probably comparable in size to Long Valley Caldera but formed in a Cretaceous island arc rather than a Plio-Pleistocene continental setting.

San Joaquin Canyon is an ancient drainage that predates uplift of the Sierra Nevada. The River was initially established on the gradually rising western slope of the Sierra Nevada in Tertiary time, possibly as early as 25 Ma. It formerly drained a large area east of the present Sierran crest, but at about 3 Ma, during initiation of Sierran frontal faulting, the upper reaches were beheaded by faulting and dammed by eruption of ballistic lavas. Blockage of this drainage formed ephemeral lakes, as many canyon-filling lavas are thick pillow complexes. These Tertiary lavas and associated overlaying rhyodacite domes and breccias are part of the 3.6 2.2 Ma precaldera sequence that heralded the more silicic eruptions from the Long Valley magma chamber. They are exposed along San Joaquin Ridge, which extends north from Minaret Vista and forms the present drainage divide. Two Teats and San Joaquin mountain, the two most prominent peaks along the divide, dissected remnants of large rhyodacite domes and their surrounding pyroclastic cones.

During early Pleistocene time the Middle Fork was re-excavated by repeated glaciations by repeated glaciations and acquired its broad, U-shaped form. Intracanyon remnants of the Bishop Tuff indicate that the valley was temporary filled by the caldera forming eruptions but has been re-excavated by more recent glaciations.

10-2 Devil's Postpile Hike

Continue west on CA-203 ~ 6 miles to Devils Postpile entrance.

Devil's PostPile Overview

The Postpile is a fine example of columnar "basalt" (Figure 55). Actually the flow is reported to be andesite in composition. The flow erupted in the glaciated valley of the San Joaquin about 3.0 miles north of the Postpile. It filled the valley and was probably more than 400 feet deep in the vicinity of the Postpile. This eruption occurred almost 1,000,000 years ago. The flow was very fluid and pooled so quickly that for awhile it was molten throughout. Because its top was exposed to air and its bottom to the granitic rocks below, it cooled to solid rock inward from these surfaces. While cooling it shrank. Tension developed as the flow shrank. Stresses imposed in a vertical direction only caused the mass to settle in response to gravity, but stresses imposed in a horizontal direction were relieved only by the cracking of the solidifying rock. Ideally, this cracking would tend toward a hexagonal section. Rapid and even cooling rates in a homogeneous mix would do this. However, cooling stresses in rocks are rarely completely uniform and regular hexagonal jointing is not apt to occur and columns are irregular shaped polygons with variable numbers of sides. Ideal cooling conditions are hardly ever reached at the surface of a flow, and jointing there is usually irregular or poorly columnar -- of brickbat texture. However, as cooling progresses into the interior of the flow a point may be reached where the shrinkage may be uniform enough to permit the formation of more or less columnar jointing. This is what happened at the Postpile. The upper, poorly jointed flow material has been eroded away, exposing the well jointed columns of the lower part of the flow. The glaciation that has occurred in the valley of the San Joaquin subsequent to the eruption of the Postpile flow has almost obliterated the flow because the abundant jointing in the flow left it very susceptible to glacial plucking. A hike to the top of the Postpile reveals the top of the piles, with glacial striae still evident. The Devils Postpile area was included within the boundary of Yosemite Park when it was set aside as a National Park in 1890. Pressure from mining and timber interests led to the return of this portion of the park to the Public Domain in 1905. In 1910, a request for a permit to blast the Postpile into the San Joaquin River to form a dam for the use of mining operations stirred up protest from influential persons and on July 6, 1911, President William Howard Taft proclaimed the area Devils Postpile National Monument.

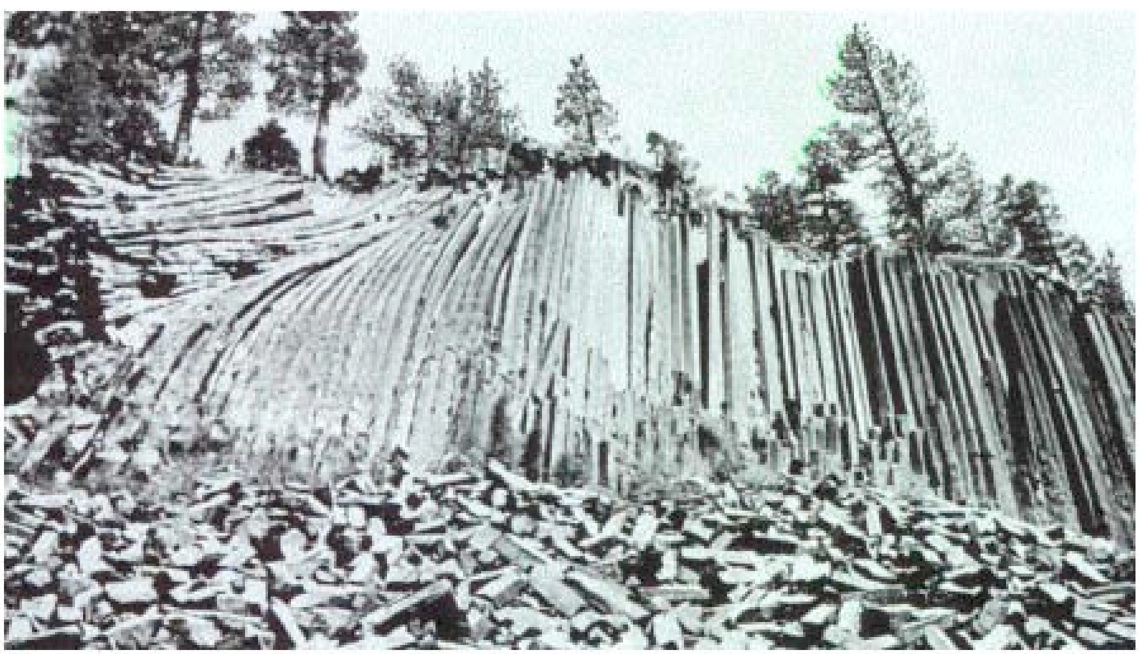

Figure 55 - Devils Postpile National Monument, Madera County (Norris and Webb, 1990, p. 120, Fig. 3-36)

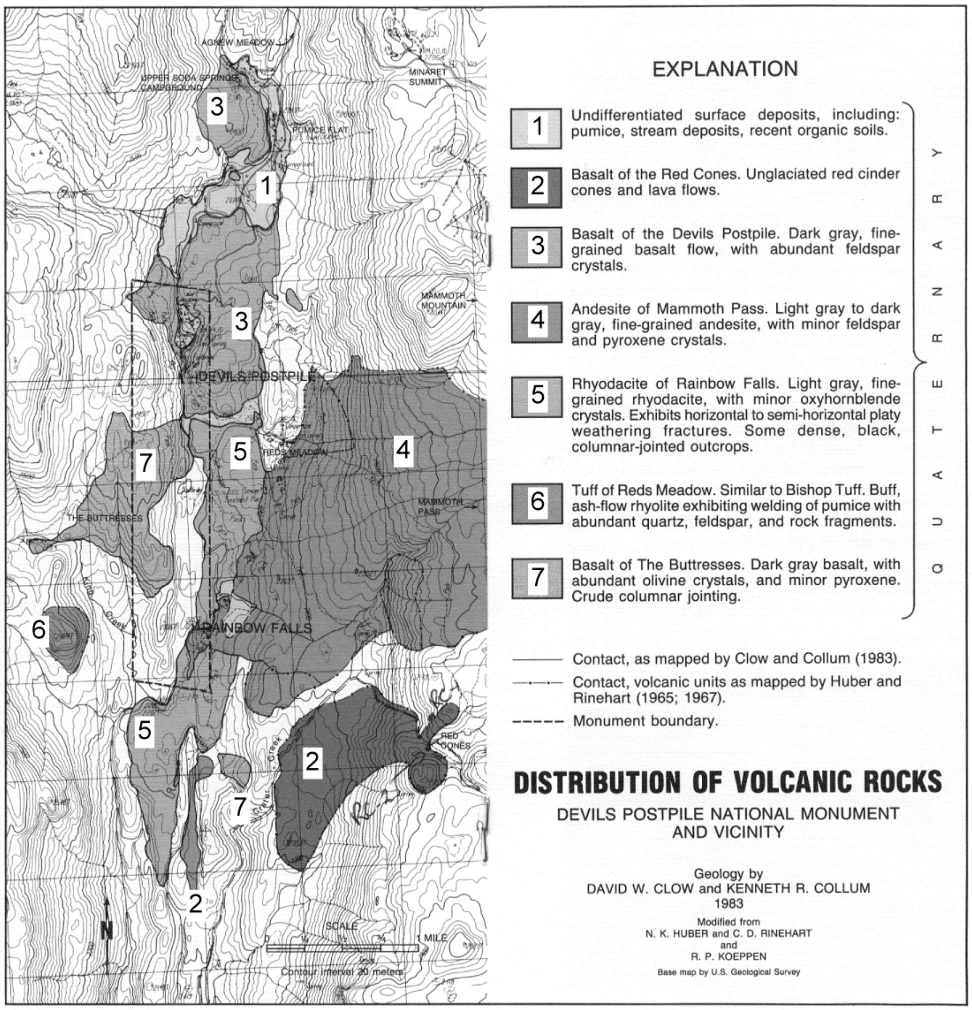

Figure 56: Map of Devils Postpile Area (Huber and Eckhardt)

Other Volcanic Rocks In and Near Devil's Postpile Monument

(From the Devils Postpile Story, Huber and Eckhardt, available at: http://geology.wr.usgs.gov/parks/depo/dpgeol9.html)

Not all of the volcanic rock in the Monument area belongs to the Postpile basalt flow, although much of it has been grouped into a single geologic unit on most published maps of the area. This has led to confusion both as to the composition of the Postpile flow (called andesite on some maps) and as to its source (said to be Mammoth Pass in some reports). Recent studies have given us a better understanding of the variation in rock types and the complex relationships between the different volcanic units. Individual volcanic units previously grouped with the Postpile flow, but here considered separately, are: andesite of Mammoth Pass; rhyodacite of Rainbow Falls; and basalt of The Buttresses (basalt of the Devils Postpile is the youngest unit; the others are listed in order of increasing age).

Lava that erupted from a vent near Mammoth Pass once was thought to be part of, and indeed the source of, the Postpile basalt. This was simply because this lava was erupted from an obviously higher elevation and cascaded into the Middle Fork valley toward the Postpile. The Mammoth Pass rock, however, is andesite, a rock with somewhat more silica than basalt, and resulted from a separate volcanic event. In appearance the andesite lacks the small, but visible, crystals of feldspar that characterize the Postpile basalt.

The oldest volcanic unit formerly included with the basalt of the Devils Postpile is a basalt that makes up The Buttresses, southwest of the Postpile. In appearance this basalt superficially resembles the Postpile basalt, but the abundant visible crystals are mostly pyroxene rather than feldspar, a subtle but important difference between the two rock types. The basalt of The Buttresses was erupted from one or more vents west of the Monument and flowed eastward into the Middle Fork valley.

A totally different type of volcanic rock that occurs near the Monument is a pyroclastic rock or welded-tuff. This tuff formed from volcanic ash that was hot and plastic when it fell and fused into a generally cohesive mass. Exposures of this tuff can be seen behind the old ranger cabin east of Reds Meadow. At this locality the tuff has columnar joints, but the columns are nearly horizontal rather than vertical. The exact source of the volcanic ash is not known, but it is probably related to the extensive deposits of the Bishop Tuff, a welded tuff that erupted from the Long Valley caldera, and can be seen in road cuts near Tom's Place on Highway 395.

The most recent volcanic event in the Devils Postpile area, the only one younger than that which produced the Postpile flow, built the Red Cones about one and one-half miles southeast of the Monument. The two basalt cinder cones, with well-preserved summit craters, illustrate what the eroded pile of cinders near the Upper Soda Springs Campground might once have looked like. A basalt lava flow that issued from the base of the southern cone extends down Crater Creek to within one mile of the Middle Fork, perhaps similar to the way that the Postpile flow might have issued from the ancient cinder cone that once existed near the Upper Soda Springs Campground. The Red Cones and their lava flow are so well preserved because they are less than 10,000 years old, and thus escaped the powerful excavating force of glacial erosion that so drastically modified the older volcanic rocks.

Pumice, a frothy volcanic glass, is a porous lightweight material of pastel shades that covers the ground at various places in the area. Its unweathered, loosely compacted nature indicates it was deposited recently. This pumice was formed in postglacial time when molten rock of high silica content and abundant dissolved gasses erupted from the Mono and Inyo Craters northeast of the Monument.

The recency of the Red Cones and widespread blanket of pumice, the presence of hot springs here and to the east, and the recent volcanic eruptions and earthquakes in Long Valley provide dramatic evidence that volcanoes in this region cannot be labeled "extinct." Within the framework of geologic time, which we tend to think of in terms of millions of years, the dormancy of local volcanism is a barely measurable pause that might, at any time, be abruptly terminated by the onset of new eruptions.

10-3 Post-Caldera Lava Flows

Retrun to CA203-Hwy395 intersection area

Find safe place to park along CA203 road to view post-caldera basalt flows

Post-caldera Mafic Volcanism: The lava flows exposed along this stretch of CA-203 are typical of post-caldera mafic volcanism in the Long Valley area. Most of the flows probably originated from vents along the margin of the caldera and flowed into the caldera moat. Mafic magmas were extruded where feeder dykes passed outside of the still molten felsic magma chamber that produced the Bishop Tuff. The oldest unit is olivine-pyroxene phyric, as are the pre-caldera mafic volcanic rocks. The younger unit, exposed in the road cuts, most commonly includes plagioclase phenocrysts with only minor olivine and pyroxene. Most of the post-caldera mafic rocks have been contaminated by upper crustal rocks, and xenocrysts of quartz and biotite can be found in the most SiO_2-rich flows.

10-4 Mammoth Scenic Loop and Inyo Craters

Return to Hwy 395, turn left, drive to Mammoth Scenic Loop, turn left, proceed 2-3miles to Inyo Craters turnoff (signed), turn right, follow unpaved road to Inyo Craters parking lot. Park and follow trail 0.3 miles west to craters.

The Inyo craters are three northerly aligned, phreatic explosion craters on the south flank and summit of Deer Mountain, a 115 ka porphyritic dome in the west most of the caldera. The two southernmost craters (informally designated as north crater and south crater) are about 200 m in diameter, about 60 m deep, and contain small lakes; the crater on the summit of Deer Mountain (informally referred to as summit crater) is smaller, irregular in outline, breached on its south side, and dry. The lake in the south crater is yellowish green, suggesting the presence of suspended sulfur, but the water is cold (11°C) and other evidence of thermal influx is lacking. The lake in the north crater is colored brown with organic matter. Despite apparent differences in morphology and vegetation, all three craters formed at nearly the same time, probably within hours or days. They erupted in succession from north to south, as the light-colored deposits from summit crater underlie the darker ones north of north crater- which in turn underlie the darkest ones south of crater- relations that are well exposed in the Northeast wall of north crater and on the east flank of Deer Mountain. The light-colored debris around summit crater is comprised primarily of pulverized hornblende-biotite rhyolite of Deer Mountain Dome, whereas the darker debris around north and south craters consists largely of fragmented trachyandesite derived from flows like those exposed in the in the walls of the south crater. In the north wall of the south crater, exposed in succession above the trachyandesite flows, are: (1) trachyandesitic cinder from nearby vents to the northwest and southeast (10 m), (2) pumiceous rhyolite tephra from the Inyo domes magmatic eruptions centered a few kilometers to the north (1 m), and (3) coarse, crudely bedded phreatic explosion deposits mainly from the south crater (13 m). The latter, which extend as far as 1 km from the crater, consist mainly of trachybasalt-trachyandesite blocks in a grayish-brown, compact semi-indurated, fine-grained matrix; blocks and boulders of granite and metamorphic rocks, probably derived from subsurface glacial till or possibly Sierran basement, also are included in the debris. A battered log incorporated in the deposit of south crater yielded a radiocarbon age of 710 +/- 60 yrs B.P., which together with dendrochronlogical data gave a calendar age of between 1340 and 1460 A.D. for phreatic eruptions.

The terrain around the craters is broken by many north-trending faults and fissures, which show up to 20 m displacement individually and define a graben 0.6 km wide and 2.5 km long. The graben probably formed as a consequence of uplift and distension above a rising dike, which probably generated the phreatic explosion craters. To test this hypothesis, Sandia National Laboratories slant-drilled an 865 m scientific corehole eastward at an angle of 68° from a site on top of the fault scarpe just west of south crater. It passed through a 320 m sequence of post-trachyandesite and trachybaslt flows, about 50 m of gravel, 360 m of tuffs and flows of the early rhyolite, 63 m of glacial till, and terminated in 30 m of Paleozoic quartzite presumed to be the Sierran basement rock. At a depth of about 600-650 m, within the early rhyolite section and directly beneath the center of the south crater, the hole penetrated an apparent vent breccia consisting primarily of pulverized early rhyolite and postcaldera trachybasalt. Included within the breccia were small pumiceous fragments of a distinctive high-silica rhyolite, apparently juvenile and considered to be the magma that generated the phreatic explosions. Presumably a dike or conduit of this magma raising from depth encountered water-saturated tuffs of the early rhyolite, causing explosive flashing of the water to steam, which in turn reamed a vent through the overlying postcaldera trachybasalt sequence, producing the Inyo explosion craters and surrounding phreatic deposits. Surprisingly little or no juvenile rhyolite magma reached the surface during eruptions.

Return to June Lake Campground

Day 11: Monday, 28 August 2006

Summary: Leave June Lake
Bishop Tuff Pumice Quarry
Owens River gorge, Bishop Tuff
Owen's River Gorge View
McGee Faults
Summit of Lookout Mountain
Obsidian Dome
Glass Creek Dome
Hot Creek
Return June Lake

11-1 Bishop Tuff Pumice Quarry

**June Lake to Hwy 395, turn right and proceed south through Bishop to Hwy 6
Turn left onto Hwy 6. Drive 8.5 miles north to Rudolph Road, turn left into Insulating Aggregate Company pumice quarry. Pumice Fall**

Bishop Pumice Quarry: Bishop Tuff, basal pumice fall and distal ash flows

Exposed in the quarry are 4 m of basal pumice fall and 4-6 m of distal ash flows of the Bishop Tuff. The pumice fall is roughly divisible into: (1) a lower third composed of moderately well-bedded ash and fine pumice lapilli; (2) a middle third that is darker-colored, more distinctly bedded, with numerous fine-to-coarse alterations containing pumice lapilli up to 4 cm in diameter. Discontinuous dark manganiferous streaks are present in the upper third of the deposit. The lack of distinct discontinuities in the lower two-thirds suggests continuous deposition over a relatively short time, with activity becoming more variable or intermittent in the upper third. The tendency for lithic clasts as well as pumice lapilli to increase upward suggests generally increasing eruption intensity. Very subtle low-angle crossbedding together with the poor sorting in the lower part suggest deposition from very dense, laterally drifting, eruption clouds. The overlaying ash flows are generally massive and poorly sorted, but the basal few centimeters consist almost entirely of fine ash, which locally also shows faint low-angle cross-bedding suggestive of pyroclastic surge. The lowermost 2 m of the ash flows contain many discontinuous swarms of relatively coarse pumice lapilli and blocks, indicating complex multiflow deposition near this distal part of the formation.

Interbedded within the pumice-fall sequence at the north end of the quarry is a 1-2 m thick, unsorted, tongue-shaped deposit with an irregular upper surface; it is buff-colored and consists of unusually coarse pumice blocks and lapilli in an ashy matrix; it resembles a pyroclastic flow but is more likely a mudflow, possibly formed by choking of local streams with tephra during plinian episode.

At one or two localities in the quarry, low angle reverse faults with as much as 1 m displacement can be seen. The fault planes incline eastward with decreasing dip and become bedding-plane faults that are difficult to trace laterally. In this structural setting, near the western boundary fault of the White Mountains and on the eastern edge of the volcanic Tableland, which is broken by numerous normal faults, reverse faults seems anomalous. Possible they are related to local compressional wedging or bending within down fualted blocks in an otherwise extensional region.

11-2 Owen's River Gorge, Bishop Tuff

Return to Hwy 6, turn right. Drive to Hwy 395, turn right onto 395 for 12 miles to Gorge Road. Turn right. Drive 0.7miles to T-junction, turn left.
Drive 6.0 miles to Y-junction, bear right. Drive 0.2 miles to locked gate. Park.

Owens River Gorge: Section of Bishop Tuff: rosette jointing

The Los Angeles Department of Water & Power (LADWP) upper powerhouse road descends the west wall of Owens River Gorge and affords spectacular close-up views of a 150 m thick section of Bishop Tuff on the east wall. At the same time, the road permits close-hand examination of road cuts and outcrops that show vertical changes in density and texture of the entire tuff section.

On the east wall of the gorge, the Bishop Tuff displays a remarkable variety of columnar jointing. In the upper half of the cliff the joint columns are remarkably well formed five- to six- sided columns 1-3 m in diameter. Locally these columns curve and converge downward toward common foci, forming joint rosettes. These rosettes are the loci of large fossil fumaroles. At the top of the gorge directly above many of these joint rosettes are fumarolic mounds like those seen on the Bishop Tuff surface near Round Valley. The distribution of these mounds close to the present gorge suggests that the volatiles responsible for their formation were derived in part from the ancestral Owens River, which was overrun and vaporized by Bishop ash flows.

In the lower half of the gorge, in dark-grey, densely welded tuff, the joint columns are much larger, 10-20 m in diameter, and are relatively crudely formed as a consequence of slower cooling around more widely spaced cooling centers. Locally, some of these joint columns have secondary horizontal joint columns developed perpendicular to their primary surface. Between the upper and lower tiers of joint columns is a broadly undulating ill-defined parting that marks the contact between two sub-cooling units of the Bishop Tuff. This parting and another like it at the top of the gorge near the gate represents brief hiatuses in deposition of the tuff and show that the tuff consists of multiple cooling units. However, the partings probably do not represent intervals of more than a few days or weeks or at most 2-3 yrs. The prominent crystal-rich parting near the gate is the contact between Tableland and Gorges cooling units; the Tableland (upper) unit contains two pyroxenes and has Fe-Ti-oxide temperatures of 737-763°C whereas the Gorges (lower) unit lacks pyroxenes and has lower oxide temperatures of 725-736°C

Exposures along the left (west) side of the powerhouse road display almost the entire range of lithologic variation in the Bishop Tuff. Near the upper gate the rock is light gray to pinkish or purplish, porous, poorly to moderately welded, vapour-phase tuff. Downward, the tuff grades through brown to dark gray, becomes progressively less porous and more densely welded, and displays conspicuous eutaxitic texture. In the most densely welded, darker facies eutaxitic is almost entirely obscured by devitrification. At the first major left bend in the road, however, eutaxitic texture becomes more prominent over a 5 m interval where scattered obsidian fiamme occur, suggesting a minor eruptive/cooling interval. The tuff below this level is monotonously dark gray, densely welded, and devitrified to the bottom of the gorge, but the powerhouse foundation was excavated in dense eutaxitic vitrophyre, indicating that the base is not far below.

11-3 Owen's River Gorge View

Return 0.2 miles to Gorge Rd, turn right and drive 1.8 miles to Y-junction.
Bear right, follow paved road 0.3 miles, turn left on dirt road and drive 0.1 miles to rock ring fireplace. Walk to gorge.

Upper Owens River Gorge overlook: Gorge startigraphy and structure

Owens River Gorge is incised into the surface of the Volcanic Tableland and here crosses a north-trending fault zone that has been intermittently active for the past 3-4 m. y. The bottom of the gorge is cut in the Triassic Wheeler Crest Quartz Mononite, which is overlain by a 20 m thick sequence of 3.2 Ma precaldera trachybasalt flows that that erupted from a vent about 1 km north of the gorge. Locally overlying the trachybasalt are a few meters of outwash gravel of the Shewin glaciation. The upper 60 m of the gorge exposes partly welded ash flows of Bishop Tuff. Down-gorge, the tuff thickens too a maximum of about 200 m, gradually becoming more densely welded and intensely jointed. Up-gorge, the tuff thins over the quartz monzonite high formed by the north-trending fault zone, multiple strands of which can be seen on the opposite canyon wall. Increasingly greater offset down-to-the-east in successively older units-the bishop Tuff, trachybasalt and quartz monzonite- demonstrates the growth nature of the fault zone.

11-4 McGee Faults

Proceed 0.2 miles on dirt road to aqueduct suge tanks and junction of US Forest Service Rd. Continue straight ahead 2.5 miles to Hwy 395. Turn right, proceed ~7 miles to McGee Canyon Rd on left. Park 0.4 miles from Hwy.

McGee Canyon: McGee Canyon provides a spectacular view into the heart of the Sierra Range, heading in the Big and Little McGee Lakes at the foot of Red and White Mountain, the canyon is cut through a fine assortment of ancient and beautiful rocks. The sculpturing of water, frost and ice have carved the landscape into rugged and beautiful land forms.
The best way to see it is, of course, to hike into the back country but enough can be seen by following the road to its end to gain an appreciation of the wonders that the world of nature has unfolded for your view. Perhaps, when you reach the end of the road, you might feel inclined to stroll just a wee bit further, to let the mighty mountains close in a bit about you, to revel in the feelings that such a journey can instill within you: awe, humility, love, peace. Listen with your body to the song that nature sings. Be one with the spirit of the great outdoors. And with it, gain awareness of the irresistible forces that have moved their inexorable way through the last several hundred million years to set the present stage. As you exit Hwy 395 McGee Mountain, elevation 10,871 feet, is directly in front as you. South of McGee Canyon is Nevahbe Ridge, reaching to 12,553 feet. The road starts-up the slope of the north lateral moraine immediately. Pull off and park 0.4 miles from highway.

Recent Faults in McGee Canyon: Note small moraine at right. Last 100 feet or so at toe of moraine is down dropped and covered with green grass. Both of these phenomena are indicative of recent faulting. Another fault along face of hillside approximately at top of moraine can be seen to cut moraine ahead one-quarter mile. Road traverses valley between two lateral moraines emanating from McGee Canyon. Both laterals are on north side of glacier, evidence of two distinct episodes of glaciation, Road traverses ground moraine left by earlier ice flow.
On the shoulder of McGee Mountain, stretching north toward Tobacco Flat, can be seen a boulder field. This is a remnant of the McGee Glacier moraine and is an important detail in the story of the development of the Sierra scarp.
How was the imposing front of the range produced? Was it by uplift of the Sierra block with the Owens Valley remaining at its present level, or was it by sinking of the Owens Valley with the Sierra block maintaining its previously attained height? Or was it a combination of the two?
The answer is found in the positions of the deposits of ice-borne material, moraines, built by the glaciers which occupied these canyons during the Ice Age. The moraines indicate the faulting movements. They also indicate the approximate time of the faulting. The Ice Age consisted of four successive glacial periods, or stages, separated by long intervals of normal climate. There are, therefore, four sets of moraines differing greatly in age and degree of preservation. The moraines of the older glaciation are poorly preserved and relatively obscure, and the older they are, the more so. The oldest are mostly destroyed, being recognizable only in spots. The sharp crested moraines of the later glaciation extend outward from the mouths of the canyons, into the lowland to the east. They show by their positions that when the last glaciers advanced, the canyons had already attained their present depth. Some of these latest moraines 'step down' abruptly 50 feet or more where they cross the fault line at the foot of the range. In those places, only one small dislocation has occurred within, perhaps, the last 25,000 years. Few of the older moraines extend into the lowlands. Most of them were cut off at the fault line, in some instances

at heights of 1,000 to 1,500 feet. Thus we know that faulting movements of that magnitude have taken place since these moraines were deposited -- that is, during the last 500,000 years. The oldest moraines, which may be from 750,000 to 1,000,000 years old, lie not in the canyons or at their mouths, but high on their shoulders, thousands of feet above the lowlands. A fine example of this is the partially disintegrated moraine on the northern shoulder of Mt. McGee. It terminates on the brink of the escarpment, 3,000 feet above Long Valley. To judge from its gentle slope, moreover, this ancient moraine was built by a glacier that lazily wended its way through a rather flat, shallow valley high in the Sierra block. McGee Canyon had evidently not yet been cut. But this implies that there was as yet, no escarpment. It is then, an inescapable conclusion that the major faulting movements did not begin until after the first glaciation, or, roughly, less than 750,000 years ago. The great Sierra scarp appears to be, then, less than 750,000 years old. But how was it formed? Did the Sierra rise 3,000 feet, or did Long Valley drop 3,000 feet? The ancient McGee Moraine is of shout the same dimensions as the moraines of later glaciations, and that fact, along with similar evidence from other parts of the range, warrants the inference that during the first deglaciation of the Sierra Nevada, it was about as extensively mantled with glaciation as it was in later glacial periods. That could not have been true unless the range was as high as it was during the later glacial periods. It might have been a little lower but it could not have been 3,000 feet lower, for then its summit would have lain below the snow line and it would have borne no glaciers. The three main peaks at the head of McGee Canyon - Red and White Mountain (12,850 feet), Mount Crocker (12,457 feet), and Mount Stanford (12,651 feet) -- would only have risen to between 9,000 feet and 10,000 feet, and-yet, from detailed geologic studies, it has been determined that the snow line during glacial times never stood such below 11,000 feet. It follows then, that the Sierra Nevada could not have risen 3,000 feet since the first glaciation. The 3,000 foot escarpment below the ancient McGee Moraine could not have been produced by uplift of the Sierra block. It must then, have been produced by subsidence of the adjoining valley block. And so, that little field of boulders, high on the shoulder of Mt. McGee, becomes an important piece in the puzzle of the geologic story of the Sierra Nevada.

McGee Moraine Continue up road. Road cuts around toe of moraine and onto floor of last glacial stage. Proceed 0.4 miles and Stop. Note wide streak of green on face of moraine on south side of valley where it exits from canyon. Also note down drop of top of moraine. This is fault line, probably same fault noted on north side of canyon as you came in. Very recent faulting, post glacial. Proceed 0.5 mile. As you round curve you get superb view of upper McGee Canyon with Mt. Baldwin at head. The geology here is very complex with formations folded and faulted and turned back upon themselves, The thick light band on left side of cirque, dipping steeply to right, is known as Mt. Aggie Formation, an Ordovician calcareous quarts sandstone. This calcareous sandstone, or quartzite, was deposited in a shallow sea. It was cemented by calcium carbonate precipitated out of lime rich water. Note smooth face of cliff indicating great resistance to erosion. The jagged, vertically inclined surface above it is Round Valley granodiorite. Light material on skyline is Mt. Baldwin marble. It is of grayish-blue crystalline composition, remarkably similar to the Devonian crystalline limestone in the Suplee area. Below it and to right, the reddish material is metacherts of the Bright Dot Formation. Both formations are of Pennsylvanian age, the Bright Dot chert being the oldest. Brachiopods and crinoid columns are found in the marble. Note also the tight wedge of lighter Mt. Baldwin marble on skyline ridge of reddish Bright Dot metacherts to right of Mt. Baldwin Peak (high point on skyline). Wedge is tightly folded contact between two formations. Note dikes in face of marble and calcareous quartzite, probably related to granodiorite intrusion. Rounded dome on skyline above marble is grano-diorite of the Cretaceous intrusion that formed the great Sierra batholith. Jagged cliffs in foreground showing much deformation are older rocks, Ordovician and Silurian metasediments. Terracing of slope of moraine across valley is probably result of stillstands during retreat of glacier and do not represent sequence of separate glacial-episodes. Road ends just beyond campground entrance. A train follows McGee Creek up the canyon, past McGee lakes, and crosses the divide through McGee Pass. This trail has been closed due to the danger of falling rocks because of renewed earthquake activity. Should you have the time and inclination however, the first nine or so beyond road's end is relatively safe from falling rock and will open bright new vistas to your eye.

11-5 Summit of Lookout Mountain

Return to Hwy 395, turn left. Drive ~20 miles north to Mammoth Scenic Loop (on left). Turn right (east) onto gravel road, follow it 3.1 miles to summit of Lookout Mtn.

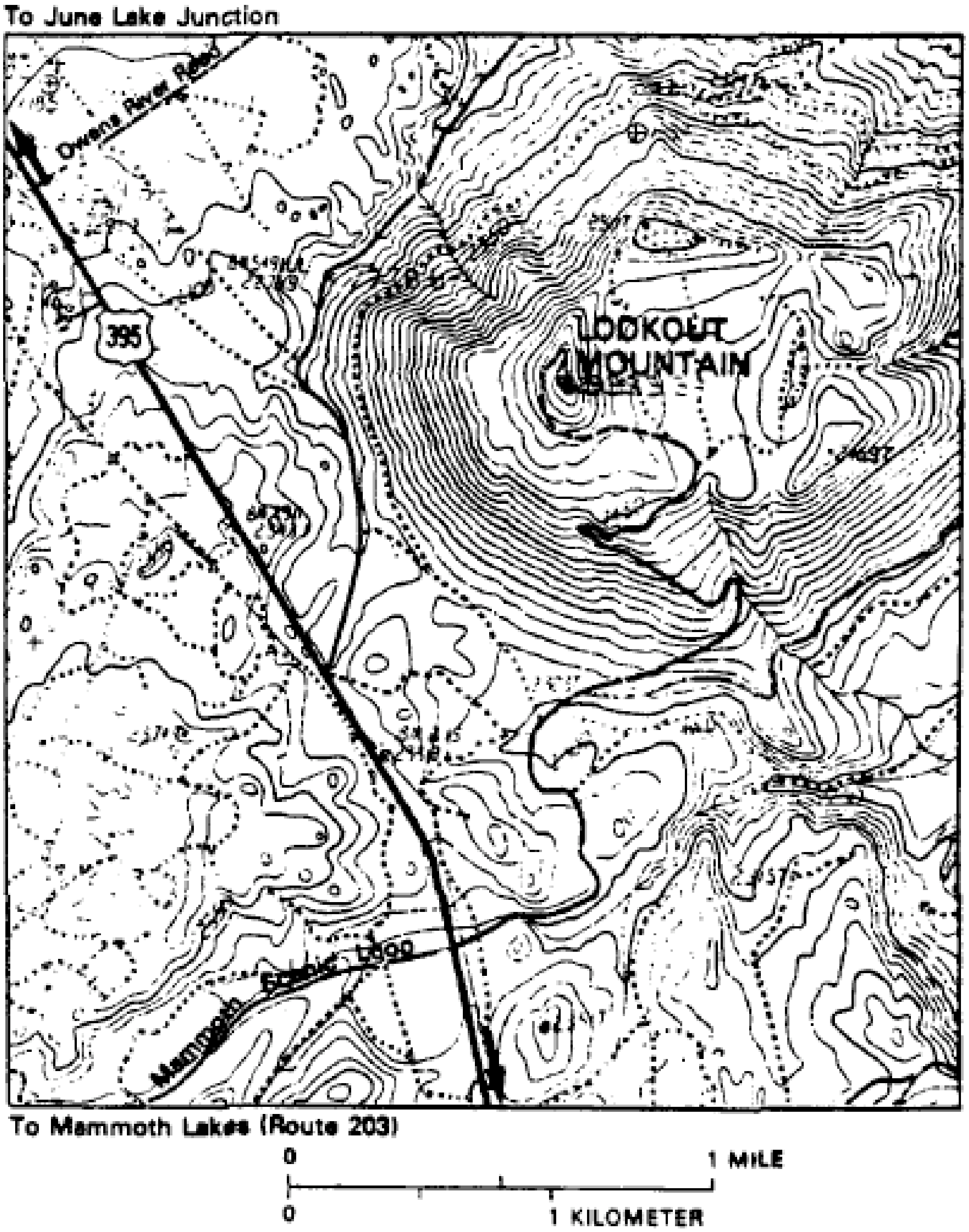

Figure 57 - Location map showing access route to Lookout Mountain. Elevations in meters; contour interval 10 m.

Lookout mountain is a small rhyolite stratovolcano with a summit crater 2 km in diameter, which forms the tree-covered depression immediately west of the viewpoint. The volcano is in the northwest moat on the northern edge of the resurgent dome, and, although not part of the resurgent dome, it is constructed of intensified 0.69 Ma flows and tuffs of the early rhyolite. The aphyric obsidian underfoot is typical of the early rhyolite.

To the west, is the west moat of the caldera, underlain mainly by 150-60 ka postcaldera trachybasalts and trachyandesites, and the west and northwest wall of the caldera, underlain by Sierran granitic and metamorphic rocks and capped by Pliocene precaldera volcanic rocks. In the middle distance are the barren, craggy, Holocene rhyolite dome-flows of the Inyo Craters chain, including from north to south: Wilson Butte, Obsidian Dome, Glass Creek Dome, and Deadman Creek Dome. Wilson Butte is the oldest of the four (1350 yrs B.P.) and is chemically and petrographically similar to high-silica rhyolites of Mono Craters, visible on skyline to north. Obsidian Dome is composed of commingled sparsely and coarsely porphyritic low-silica rhyolites, and Glass Creek and Deadman Creek domes are younger and all about the same age (650-550 yrs B.P.). Activity at these three younger vents occurred within a very short interval, progressed through from initial phreatic explosions through pyroclastic eruptionsn and concluded with passive lava extrusion, the latter occurring only after pyroclastic activity had ceased at all three vents. Pyroclastic eruptions began at Deadman Creek vent and shifted to the Obsidian Dome vent and then to the Glass Creek vent. The minimum calendar age of the tephras, based on cross-dating of three rinfs of Jeffery pines growing on the deposits, is 1369, 1433, and 469 A.D., respectively. The distribution of the tephras was mainly to the northeast and southwest. North-south alignment of the Inyo phreatic and magmatic vents and their association with subparallel grabens and fissures suggest that the eruptions were fed by an 8 km long dike that trapped commingled rhyolite magmas from at least two sources. This feeder dike transects the caldera margin os Obsidian Dome erupted on the caldera rim and flowed down the northwest wall, and Deadman Creek Dome erupted well within the caldera.

11-6 Obsidian Dome

Return to Hwy 395, turn right (north) and drive 2.5 miles to Glass Creek Rd, turn left. Drive 2.4 miles to Glass Creek Park amongst pine trees. Walk up bulldozer road that ascends Obsidian Dome.

Glass Creek: Obsidian Dome margin and sqeeze-up

The steep, south facing talus escarpment composed of obsidian blocks is the southern edge of Obsidian Dome, which vented 1 km to the north. The morphology of the dome edge is virtually unmodified since it formed 600 yrs ago. The road up the escarpment leads to a commercial pumice prospect and provides access to a scientific corehole drilled through the distal end of the domeflow. The corehole revealed details of the internal lithology and structure not evident from surface exposures and, together with data from another corehole through the conduit of the dome, has provided a better understanding of the role of volatiles during a lava extrusion.

Near the drill site on the flow surface is an unusual feature commonly referred to as a squeeze-up, or crease structure, composed of very coarse vesicular obsidian. The feature is roughly circular in plan, about 100 m in diameter, slightly depressed below the general level of the flow surface, and bisected by a central level of the flow surface, and bisected by a central “crease,” about which flow foliation and lineation symmetrically fan outward. The crease in such structures is usually oriented perpendicular to the flow margin, subparallel to the direction of flow. It appears to form as a result of lateral or tangential spreading of the flow carapace, allowing more fluid, gas rich-lava from the flow interior to well up to the surface. Such features most commonly form where flows move onto gentler slopes and where forward movement slows and lateral spreading increases.

11-7 Glass Creek Dome

Walk backdown bulldozer road, then proceed to north edge of Glass Creek (gulch). Walk upstream (west) along trail ~500 m to road crossing the creek and follow road south (left) to base of Glass Creek Dome.

Glass Creek Dome: Commingled rhyolite lava

Directly south of Glass Creek is the steep, talus- and block-mantled, north margin of Glass Creek Dome. The dome is composed of two distinctive rhyolites: (1) sparsely porphyritic rhyolites, typically consisting of black obsidian or vitrophyre and occurring mainly on the dome margins, and (2) light-gray, pumiceous, coarsely porphyritic, hornblende-biotite rhyolite, occurring mainly in the center. Between the margin and centre is an intervening zone where the two lavas types have commingled, producing fascinating "marble cake" structures that illustrate the contrast in viscosity between the relatively fluid obsidian and the more viscous porphyritic rhyolite. Petrologic and chemical studies indicate that the two contrasting rhyolites originated and evolved in separate chambers, respectively, and that they commingled along an interconnecting fissure just prior to eruption

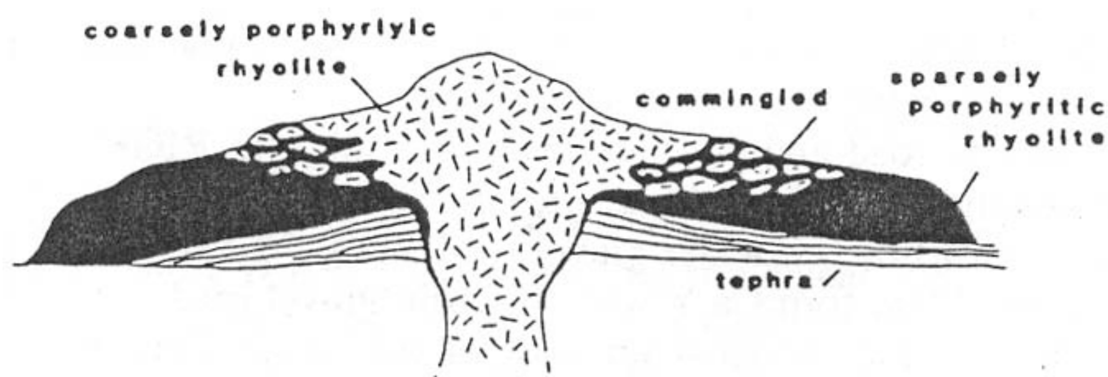

Figure 58: Schematic cross section of Glass Creek Dome.

Inyo Domes (Glass Creek Dome): Directly south of Glass Creek is the steep, talus- and block-mantled margin of Glass Creek Dome. The dome is composed of two distinctive rhyolites, a sparsely porphyritic obsidian and a grey pumiceous porphyritic rhyolite. Between the core of the dome and the margin, these two magmas have mingled (not mixed) to produce spectacular marble-cake textures. Chemically, the two rhyolites are quite different and likely came from two sources: the sparsely porphyritic unit from the Mono Craters chamber and the coarsely porphyritic unit from the Long Valley chamber. Mingling of these two magmas occurred within an interconnected fissure just prior to eruption.

11-8 Hot Creek

Return to Hwy 395, turn right . proceed ~11 mils to Hatchery Rd, turn left. At Hatchery, bear right, follow signs to Hot Creek swimming area.

Hot Creek Gorge (Swimming): Hot Creek is the lower reach of Mammoth Creek where several vigourous hot springs issue from its bottom, forming this popular swimming area. The gorge is incised into the 288,000 year-old Moat Rhyolite that erupted on the north shore of Pleistocene Long Valley Lake. The flow coursed northwards and eventually became entirely subaqueous. The hot glassy rhyolite was pervasively altered and hydrated by interaction with lake water and hydrothermal springs. Outcrops of altered flow breccia are exposed along the trail leading to the bottom of the gorge. The Hot Creek flow is a sparsely porphyritic, san-px-bearing rhyolite of the Moat Sequence. Chemically, it resembles the Bishop Tuff, and its eruption may have been triggered by influx of fresh mafic magma into the base of the Long Valley magma chamber (coincides with uplift of resurgent dome, eruption of trachyandesites in west moat). Thermal

springs issue from the stream banks all along Hot Creek Gorge, but the hottest springs are localized on two north-trending faults that crosscut Hot Creek. The boiling pools in the bottom of the gorge change in vigor and configuration in response to local earthquakes: the pools are now much hotter than they were in 1986, probably as a result of increased seismic (and magmatic??) activity in the 1990's. The white deposits ringing the pools are travertine.
Return to Hwy. 395 by retracing route to Hot Creek. Turn right towards Mammoth Lakes, continue north over Deadman Summit out of the caldera. At sign for Glass Creek, 0.4 mile north of Deadman Summit, turn left. Follow dirt road 2.4 km to Glass Creek. Park amongst the large Jeffrey pines at the south edge of Obsidian Dome and walk up the bulldozed road to the top of the dome.

Return to Hwy 395, turn right, return to June Lake Campground

Day 12 Tues August 29, 2006

Summary: Leave June Lake Campground
Bristlecone Pine Forest
Archeocyathid Reef
Mt Whitney Ranger Station
Travel to Lone Pine Campground

Directions from June Lake Campground to Bristlecone Pine Forest

Go NE on CA-158/Boulder Dr/June Lake Loop toward Lake View Drive
Continue to follow CA-158/June Lake Loop 2.5 miles
Turn slight right onto US-395 S 53 miles
Turn a SHARP left onto N Main Street/US-6
Follow US-6 for 4 miles
Turn right onto Silver Canyon Rd 6.5 miles
Straight 4 miles
Turn left onto Whitemountain Rd

ALTERNATE DIRECTIONS ONCE AT BISHOP: A visit to the Ancient Bristlecone Pine Forest is an hour drive from Bishop on paved roads (to Schulman Grove). Take US Hwy 395 south to Big Pine and turn east onto State Hwy 168 just north of Big Pine. Follow Hwy 168 east 13 miles to White Mountain Road. Turn left (north) and drive 10 miles to the Schulman Grove Visitor Center.

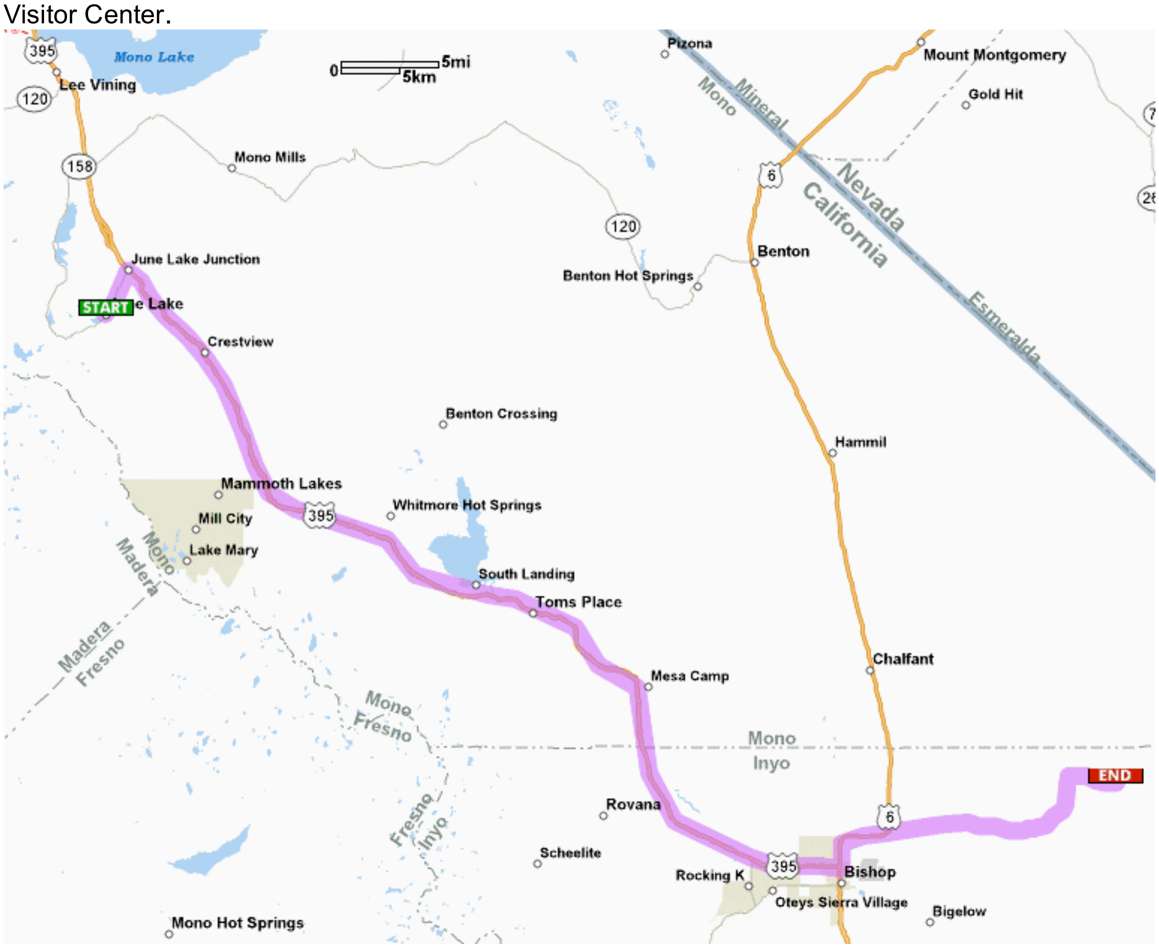

Figure 59: June Lake to Bristlecone Pine Forest

12- 1 Ancient Bristlecone Pine Forest in White Mountains

Drive to Visitor Center at Schulman Grove, then a reasonable (though rough and steep) dirt road another 13 miles to Patriarch Grove. Beyond Schulman Grove lies the Patriarch Grove. This second grove is a 12-mile drive north of Schulman Grove on a good quality dirt road. Near tree line, the grove is the home of the world's largest Bristlecone Pine, the Patriarch Tree.

From here a trail leads to the Methuselah Grove where the oldest living tree yet discovered is located. The Methuselah tree is in excess of 4,600 years old. The round trip covers over four miles and contains many steep gradients. Because of the altitude -- in excess of 10,000 feet, this is a hike to be undertaken by only the physically fit.

A shorter trail winds through the nearby trees, affording an -- intimate glimpse into the life style of the Bristlecone. A trail guide is furnished to enable the visitor to better understand the interplay of natural forces in the creation of this spectacular display. Located within the Schulman Grove is “Alpha”, the first tree found to exceed 4,000 years of age.

A dirt road continues north from the Visitor Center to the Patriarch Grove, a distance of approximately twelve miles, The road is narrow and very steep. Road conditions should be checked before attempting the drive. The Patriarch Grove is 1,000 feet higher than the Schulman Grove, This increase in elevation permits the capture of an infinitesimal increase in moisture from the winds that surrender the bulk of their water content to the Sierra Nevada to the west, This added moisture is reflected in the more rapid growth and larger size of the trees at this elevation, is within this grove stands the Patriarch, the world's largest bristlecone pine, with a girth of 36 feet 8 inches. The additional moisture produces deleterious side effects also, while faster growth and larger size are promoted, longevity is decreased. No tree in this area has yet been found to exceed an age of 1,500 years.

Studies of tree ring growth initiated by Dr. A. E. Douglass at the University of Arizona have developed today into the science of dendrochronology, by which means a system of absolute dating has been extended back in time for a period of over 8,000 years.

The work of Dr. Douglass was continued in the University's laboratory of Tree Ring Research by Dr. Edmund Schulmam, dendroclimatologist, and since his death, by Dr. C. W. Ferguson, professor of dendrochronology.

An important development of their studies has been the calibration of dates derived from radiocarbon analysis. By matching dates derived by the two systems, it was discovered that Carbon 14 dates previous to 1,000 B.C. were too recent. In 1967, the first tree ring calibration chart was worked out by Professor Hans E. Suess of the University of California at San Diego. Analysis of Carbon 14 content in annual growth rings of trees showed a definite change in isotope production over the hundreds of years, apparently due to the bombardment of cosmic rays.

The concept of modern man in Europe has been radically changed by the reevaluation of Carbon 14 analysis. Northern European civilization has now been determined to be much older than anyone had dreamed.

Outstanding views of the Sierra are obtained from many vistas within the Bristlecone Forest area. Because of the altitude, a different perspective is gained, one looks into, not up to, the mighty Sierra peaks. Mt. Whitney, highest mountain in the conterminous United States, is readily seen from vantage points throughout the Forest area.

The Ancient Bristlecone Pine Forest is located in the Inyo National Forest of the White Mountains about 30 miles east of Bishop in east-central California. This part of the Inyo National Forest is located east of Owens Valley and the Sierra Nevada along the California-Nevada border

(Cermak, 1966). They are also found in Nevada, Arizona, Utah, New Mexico, and Colorado, but the oldest trees are those from White Mountain (Smitter, 1974).

The bristlecone pines (*Pinus longaeva* and *P. aristata*) are a small tree, usually 30 feet high. Trunks and branches of the tree have often been twisted, gnarled, and sculptured by windblown sand and ice (figure 60 Many of the trees in the Inyo National Forest are more than 4,000 years old and many are centuries older than the largest of the giant sequoias or the redwoods of California. The oldest bristlecone pine, “Methuselah”, was discovered here in 1957 and has been dated at 4,723 years.

The trees seem to be adapted to an environment that is harsh and difficult - most of the older trees are found along north-facing mountain slopes at elevations of 9,500 to 12,000 feet. These older trees are characterized by large areas of die back (deadwood) and thin strips of living bark. The near absence of competitive plants in such a harsh environment is an important factor in the successful growth of the bristlecone pines. Precipitation on the slopes is slight and range from 9 to 14 inches/year. This precipitation contributes to only 1/100th of an inch in the increase in girth of the trees. Temperatures are cold and annual averages are 30 to 35/F and winter winds are fierce and laden with heavy loads of ice and snow.

The trees are found to grow in areas of exposed dolomite (alkaline calcareous substrate) (figure 61) having a high alkalinity (pH 7.9 to 8.4), low soil nutrient yield, and high moisture content than the surrounding sandstone. Therefore, the dolomite is capable of reflecting more sunlight than other rocks, contributing to cooler root zones and saving moisture. To live under such harsh conditions, the bristlecone has established several strategies. Firstly, because bristlecone needles can live 20-30 years, adding new foliage takes very little energy. The needles provide astable photosynthetic capacity to sustain the trees over years of severe stress. Secondly, many of the older trees occupy sites with considerable spacing between trees. The longevity of the needles and the inability of other plants to grow in the dolomite soil make for little leaf litter. Therefore, the spacing and lack of ground cover allow the trees to sustain lightning and fires.

Climatic Reconstruction: The bristlecone pines have a significant value as a record of climate in the Southwest. Trees such as these, which grow in arid climates, show response to wet or dry years by changes in the growth-ring characteristics such as ring width (Cermak, 1966).

A long δ13C chronology record is present from bristlecone pines in the White Mountains of California. Measurements of δ13C (representing 13C/12C ratios) of tree rings have been used to reconstruct changes in δ13C of atmospheric CO2 and to reconstruct past climates (Leavitt, 1994). Furthermore, stable hydrogen isotopic studies have also been conducted on tree rings in determining past climates (Feng and Epstein, 1994). The deuterium to hydrogen ratios (δD) in the cellulose of the trees is related to the δD of the source water used by the tree. The higher the δD in tree rings, the higher the climatic temperature (Feng and Epstein, 1994). Tree rings from three dendrochronologically dated bristlecone pines were analyzed. These trees gave a continuous time series from 8000 years ago to the present indicating the presence of a postglacial climate optimum 6800 years ago and a continuous cooling since then (figure 62).

Figure 60 Bristlecone pines near the summit of White Mountain (Smitter, 1974, p. 280)

Suggested Readings

Cermak, R.W. 1966. Ancient Bristlecone Pine Forest. National Parks and Conservation Magazine, v. 20, no. 226, pp. 4-8.

Feng, X. and Epstein, S. 1994. Climatic Implications of an 8000-Year Hydrogen Isotope Time Series from Bristlecone Pine Trees. Science, v. 265, pp. 1079-1081.

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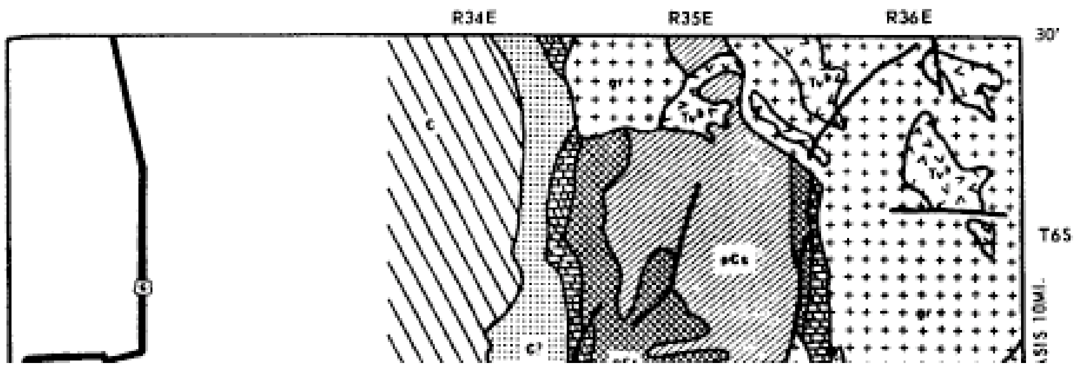


Figure 61 Geology of the Ancient Bristlecone Pines area (Smitter, 1974, p. 281)

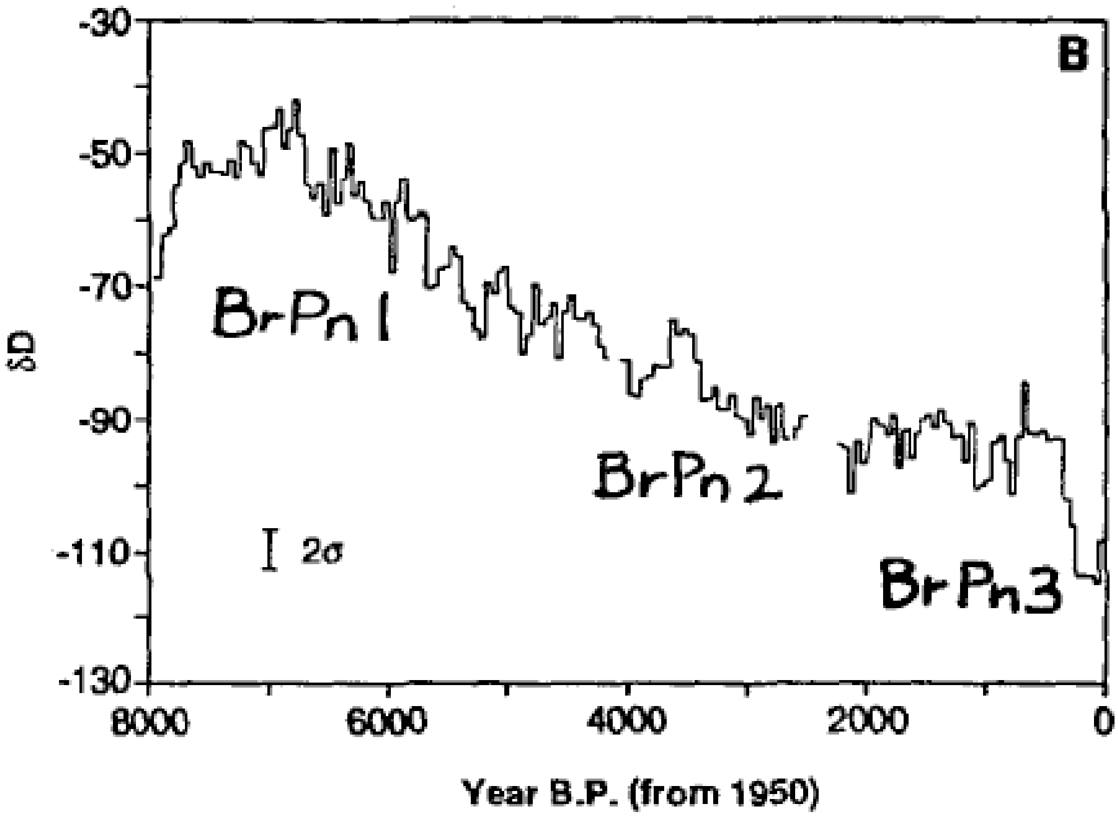


Figure 62 - Continuous time series derived from three Bristlecone pines (Xiahong and Epstein, 1994)

<u>12 – 2 Archeocyathid Reef</u>
It is upslope a 100 meters or so at the gate to the Bristle Cone Pine location. We will stop there on the way down from the Bristlecone pines.

Alternate location?: To reach the Waucoba district, first travel to the intersection of Hwy 395 and State Route 168 in Big Pine. Turn east on route 168 and proceed 2.4 miles to Death Valley Rd. Turn right here. At the 2.3 mile mark from the SR 168, look to the north of the road (left) and you will begin to see the impressive badlands carved through the Pleistocene sedimentary rocks deposited in ancient Lake Waucoba. These calcareous silts and sands, so prominently exposed, accumulated during the Sherwin Glacial Stage of the Pleistocene Epoch, approximately 760,000 years ago.

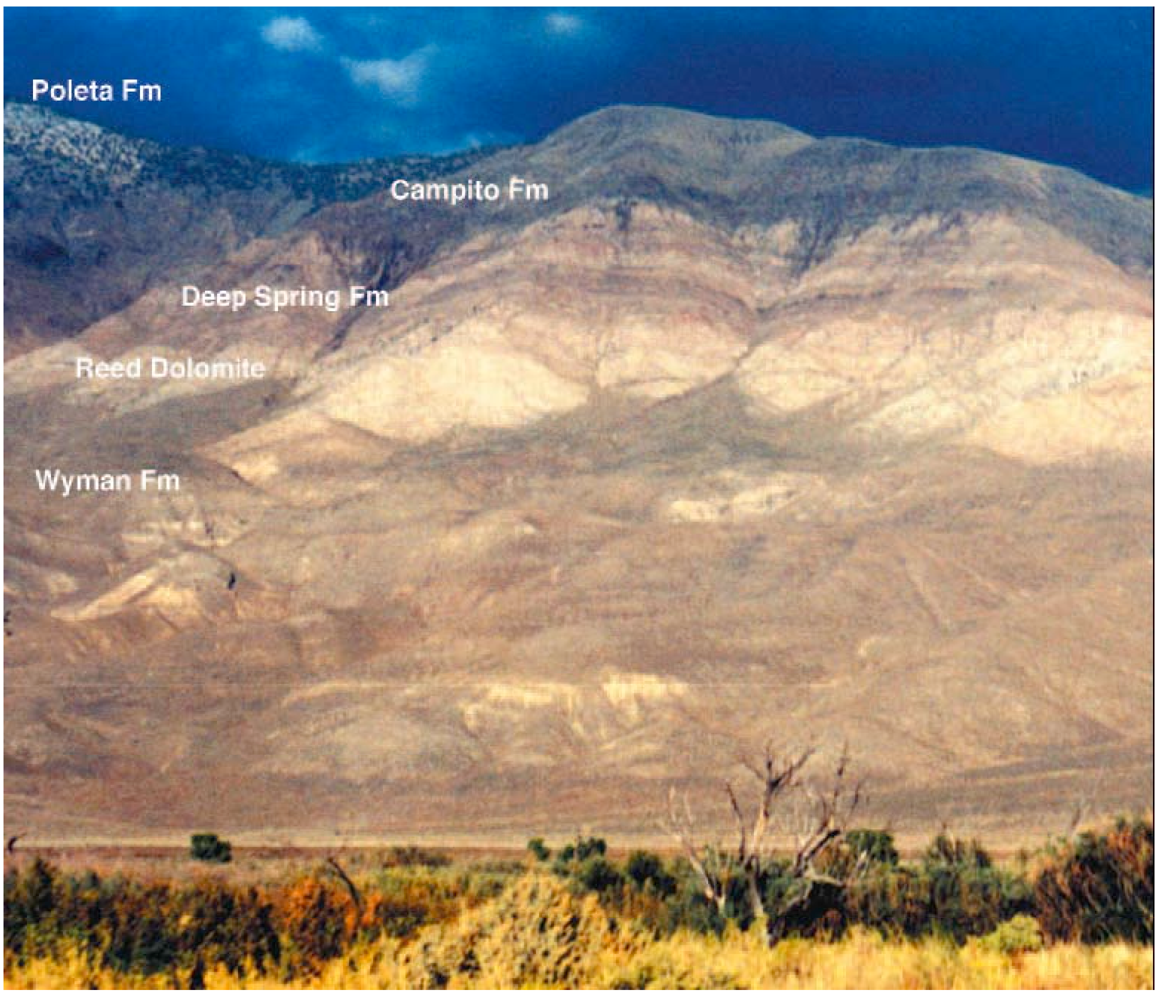


Figure 63: Precambrian-Cambrian boundary interval in the White-Inyo succession, White Mountains, view to the east from Highway 395 between Bishop and Big Pine, California

Stratigraphic sections in White-Inyo Mountains, California-Nevada, provide a well-exposed and easily accessible PC-C boundary interval through a mixed siliciclastic- carbonate succession. In this succession, δ13C chemostratigraphy has been combined with biostratigraphy to provide well constrained correlations, and these correlations have application for correlating between siliciclastic- and carbonate-dominated successions globally. The ubiquitous negative δ13C excursion near the base of the Cambrian is confirmed to coincide with the first occurrence of *T. pedum* in multiple sections across the southern Great Basin. From the time of Walcott to the present day, the Neoproterozoic-Cambrian succession in the southwestern United States continues to provide important data on one of the most interesting intervals in Earth history.

The Precambrian-Cambrian (PC-C) transition records one of the most important intervals in the history of life, because it encompasses the appearance and diversifica-tion of metazoans, the invasion of the infaunal realm, the advent of biomineralizationand predation, as well as dramatic isotopic and atmospheric changes.

The Wyman Formation consists of interbedded mudrock, siltstone, and quartzite, with lensoidal oolitic, pisolitic, and oncolitic carbonate layers that increase in number upsection. Near Andrews Mountain in the Inyo Range, the formation exceeds 3000 meters in thickness. The section primarily represents shallow marine deposition. The base of the formation is not exposed, so the nature of any underlying contact in not know.

The Reed Dolomite rests unconformably on the Wyman at most localities and is divided into three members: the Lower Member, Hines Tongue, and the Upper Member. The Lower Member is characterized by coarsely-crystalline pink dolostone with cross-bedded oolitic horizons, and minor domal stromatolite horizons. This suggests subtidal to intertidal marine deposition. The Hines Tongue is a southward-thickening siliciclastic unit consisting of hummocky crossbedded sandstone and minor siltstone with minor carbonate interbeds. Thus suggests deposition below normal wave base but above storm wave base.

The Hines Tongue is thickest in the Hines Ridge area of the Inyo Range, and thins dramatically to the north in the White Mountains and Esmeralda County, Nevada. The Upper Member is characterized by massive dolostones. Minor karstification is present at the contact with the overlying Deep Spring Formation at some localities.

The Deep Spring Formation is formally divided into the Lower, Middle, and Upper Members, each consisting of a siliciclastic-carbonate couplet. The siliciclastic halfcycle of each member contains green, ripple cross-laminated siltstones, and quartzites with hummocky cross-stratification, indicating deposition in relatively shallow water above storm wave base. The boundary between the siliciclastic and carbonate halfcycle is transitional at most localities. The carbonate half-cycle is commonly characterized by rhythmically interbedded carbonate wackestone and siliciclastic-rich siltstone, crossbedded oolite, and intraclastic grainstone; a high-energy, shallow-water depositional environment is indicated. The top of each carbonate half-cycle is commonly dolomitized, and often shows minor karstification periodic emergence is indicated. Overall, sedimentary stacking patterns, sedimentary structures, and facies associations suggest that each member represents a shallowing- upward parasequence.

The Campito, Poleta, Harkless (and Saline Valley), and Mule Spring formations record similar shallow marine, mixed-siliciclastic carbonate strata, and have similar sedimentary origins.

Archaeocyaths are an extinct group of sponges that had a very brief (geologically speaking) and spectacular history. The first archaeocyaths appear roughly 530 million years ago, during the Lower Cambrian. Archaeocyath species were very important members of Lower Cambrian communities. They diversified into hundreds of species during this time period and some of these species contributed greatly to the creation of the first reefs. Reef ecosystems tend to support a wide variety of organisms both in the present and in the past. Despite their great success in terms of numbers, the archaeocyaths were a short-lived group. They were almost completely non-existent by the middle Cambrian, some 10 to 15 million years after their first appearance. (http://www.ucmp.berkeley.edu/porifera/archaeo.html)

Figure 64: http://www.ucmp.berkeley.edu/porifera/archaeo.html

12 - 3 Mt. Whitney overview, Ranger Station

The Mt Whitney Ranger Station is located in Lone Pine, on US Hwy 395, next to Lone Pine High School.

Mount Whitney (14,495.9 ft) is the highest mountain in coterminous United States (Figure 56 and 57). At the base of the Sierra are the Alabama Hills, which extend about 16 kilometers (10 miles) and rise 90 to 120 meters immediately to the west of Lone Pine. The hills are a series of fault slivers raised in the Sierra Nevada fault zone. During the nineteenth century, a myth persisted that the Alabama Hills were the oldest in the world, because it was assumed that weathered features were of great age and that granite was always an old rock. In addition, in 1872, an earthquake that was possibly the strongest in California history, destroyed Lone Pine and produced significant ruptures at the base of the Alabama Hills. Finally, many westerns have been filmed here, among the boulder-piles, that is weathering along numerous joints in the Mesozoic granites (Norris and Webb, 1990).

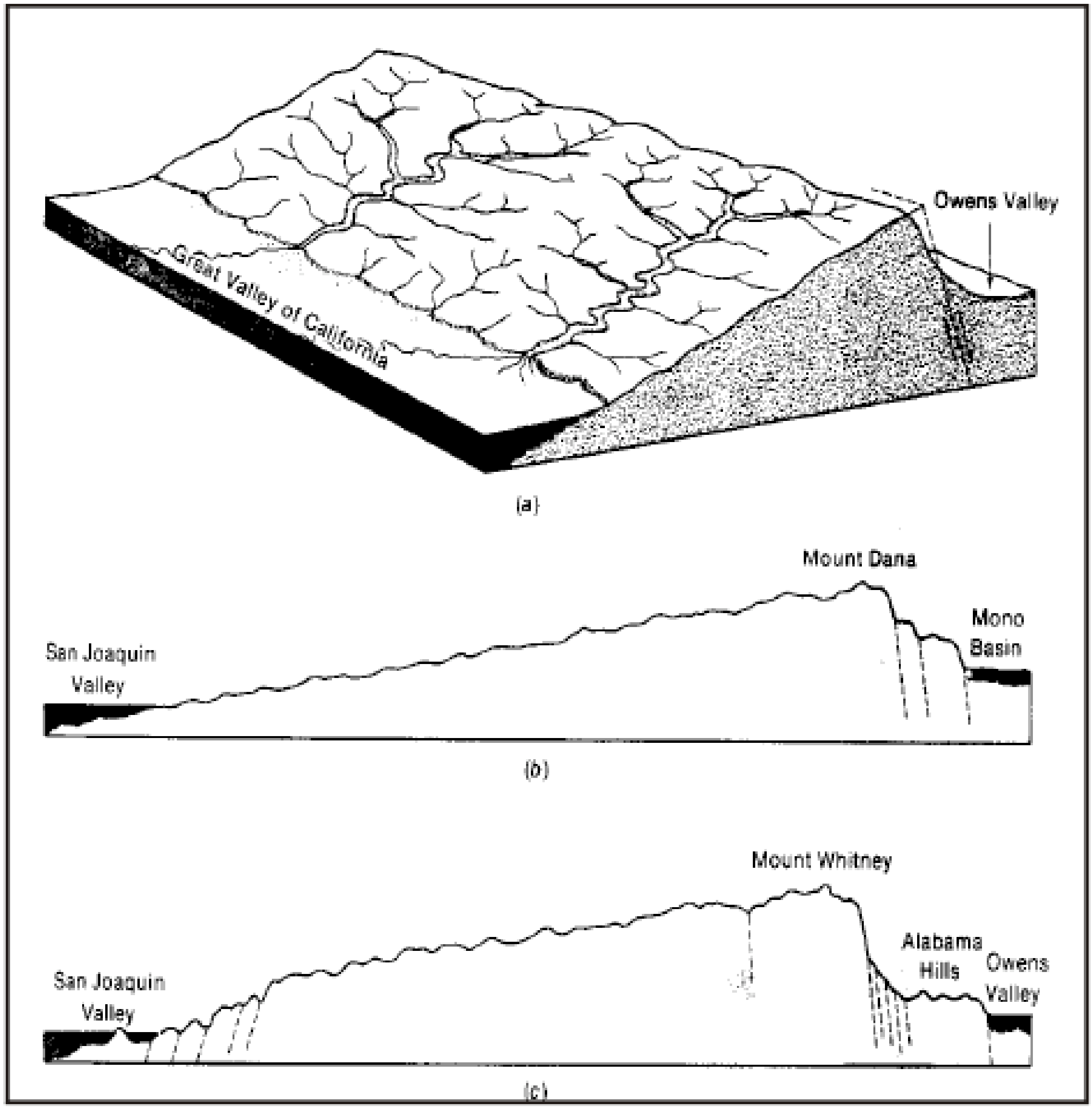


Figure 65 - (a) Generalized diagram of part of the tilted Sierran and Owens Valley blocks. Relative directions of movement are indicated by arrows. Height and slant of the Sierran blocks are much exaggerated. The strip is the Great Valley, with thick layers of sand and silt derived from the elevated Sierran block. At the back is a strip of the Owens Valley, veneered with a thin layer of sediment. (b) An idealized section across the Sierra Nevada about the range in the latitude of Yosemite Valley. (c) An idealized section across the Sierra Nevada about the latitude of Sequoia National Park. (From Norris and Webb, 1990, p. 89, Fig. 3-17)

Figure 66 - Whitney (Muir) Crest, with Mount Whitney in the center. Alabama Hills and Lake Diaz are in the foreground (Norris and Webb, 1990, p. 117, Fig. 3-34; Photo courtesy of US Geological Survey)

Travel to Lone Pine Campground: turn west onto Whitney Portal Road from US-395 in Lone Pine, continue on Whitney Portal Road going west.

Day 13: Wednesday, 30 August 2006

Summary: Leave from Lone Pine Campground
Big Pine Volcanic Field
Poleta Folds
Alabama Hills
Fossil Falls and Little Lake
Return to Lone Pine Campground

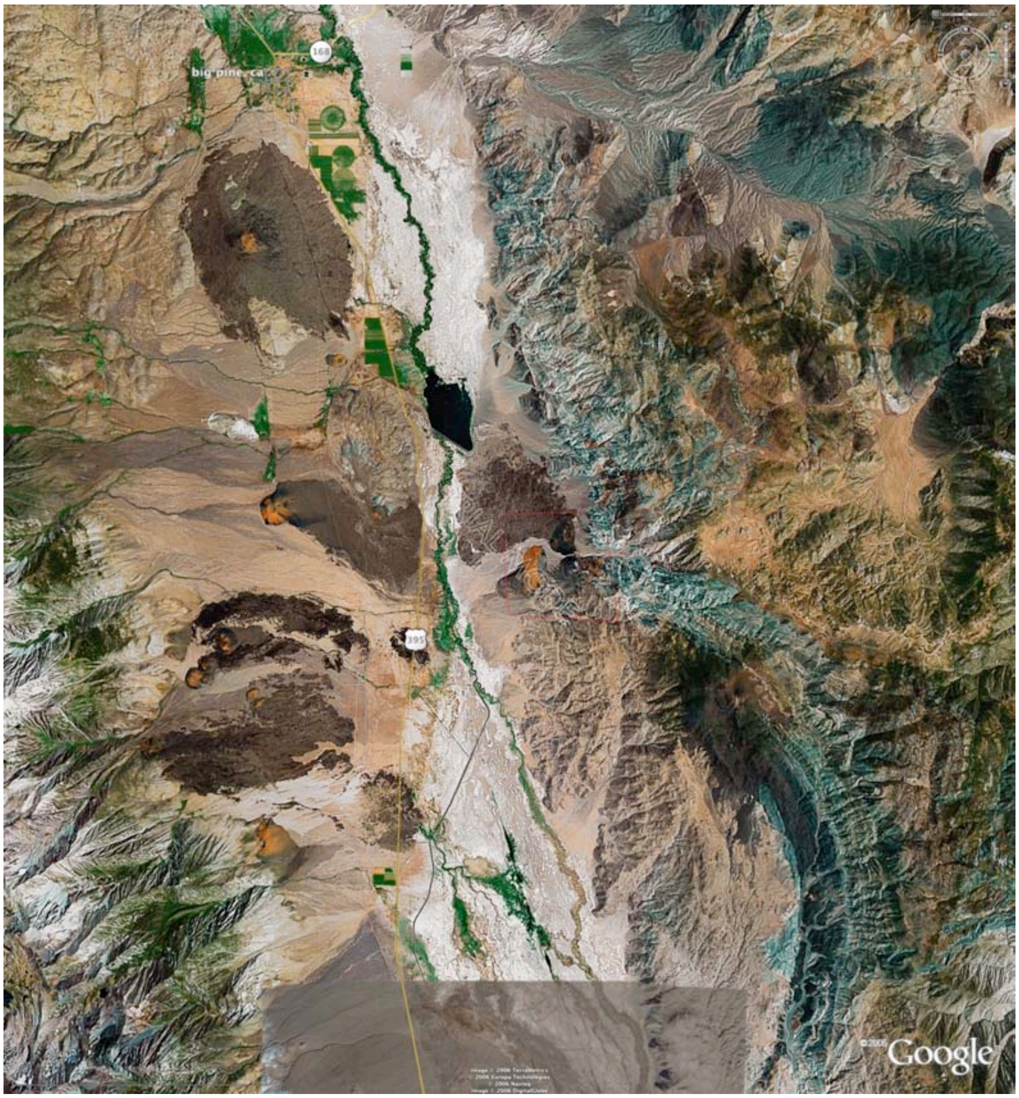
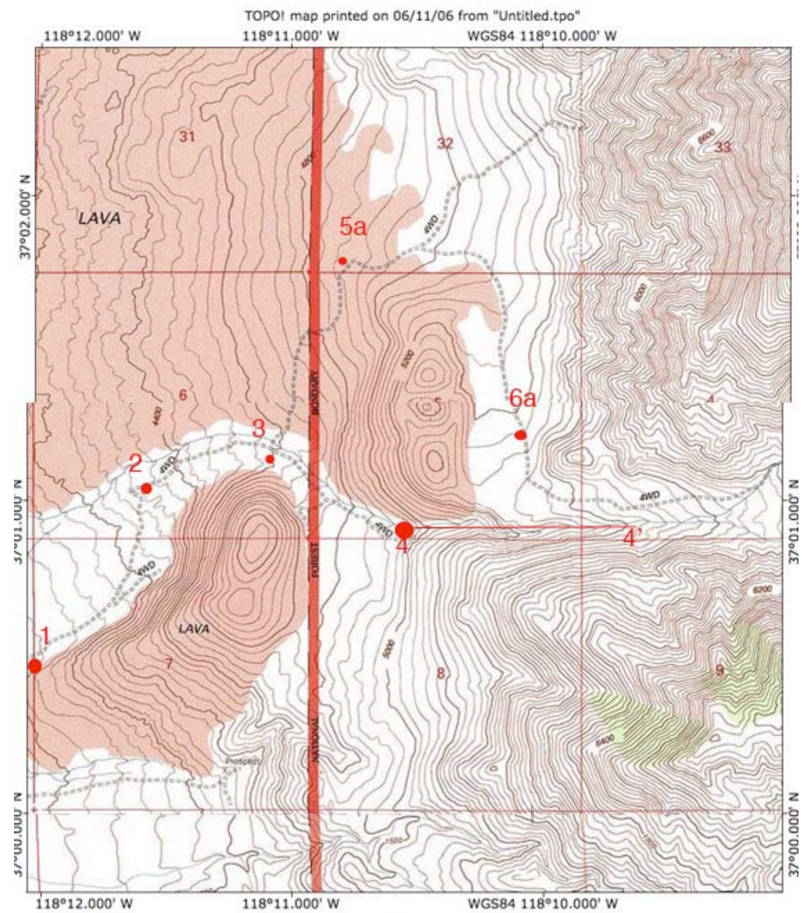


Figure 68. Google Earth view of the Big Pine Field, south of the town of Big Pine. Most volcanoes are west of HW395. The Papoose Canyon region is the only prominent set of volcanic products on the Inyo Mountain side of Owens Valley.

The Big Pine Volcanic Field (BPVF) comprises over thirty basaltic centers and one small rhyolite dome that formed during the past million years in response to extension in Owens Valley. In detail, magmatism, which is primarily mantle-derived, formed most likely in response to transtension along the Owens Valley shear zone, which is a component of the eastern California shear zone. The Papoose Canyon area exposes a rich variety of volcanic and neotectonic features. The proposed stops are shown in Figure 69. Try to follow 1 to 4 with 5 and 6 if time permits. Also the road that goes across the lava into the back side of Papoose Volcano requires 4WD.

Basalts. Most are olivine-phyric basanites and trachybasalts and are very primitive; these are some of the most primitive lavas in the western US. The amount of fractionation in magma chambers must have been small to none; many of the flows have mantle-xenoliths. You will see peridotites, and a variety of pyroxenites as well as gabbroic and other crustal xenoliths in the purple and green units mapped by McGraw in Papoose Canyon. A nicely exposed sequence of flows coming from the unnamed volcano in the SE corner of the area shows there is a chemical and isotopic temporal trend from early to later eruptions (Figure 70) – we don't have a full explanation for why exactly those trends yet. Basalts come in a variety of volcanologic flavors, ash flows, scoria cones, a'a' flows.



Key: "a" stands for alternate - if time permits
4-4' is a 1 hour long hike; other stops are short, see description in text

Figure 69. Proposed stops on the 1/24000 topo base. 4 to 4' is a one hour hike through the narrows of the Papoose Canyon.

Faults. Many faults are mappable in the area. The timing of magmatism and faulting is overlapping and there is good reason to suspect that some faults acted as magmatic conduits. The fault system that stands out is outlined in orange in figures 5 and 6 and is the active releasing bend on the Owens Valley fault (Figure 7). Major historic earthquakes along this fault were recorded in 1872, 194X and 1970. This fault system is referred below to as the “Arizona fault”.

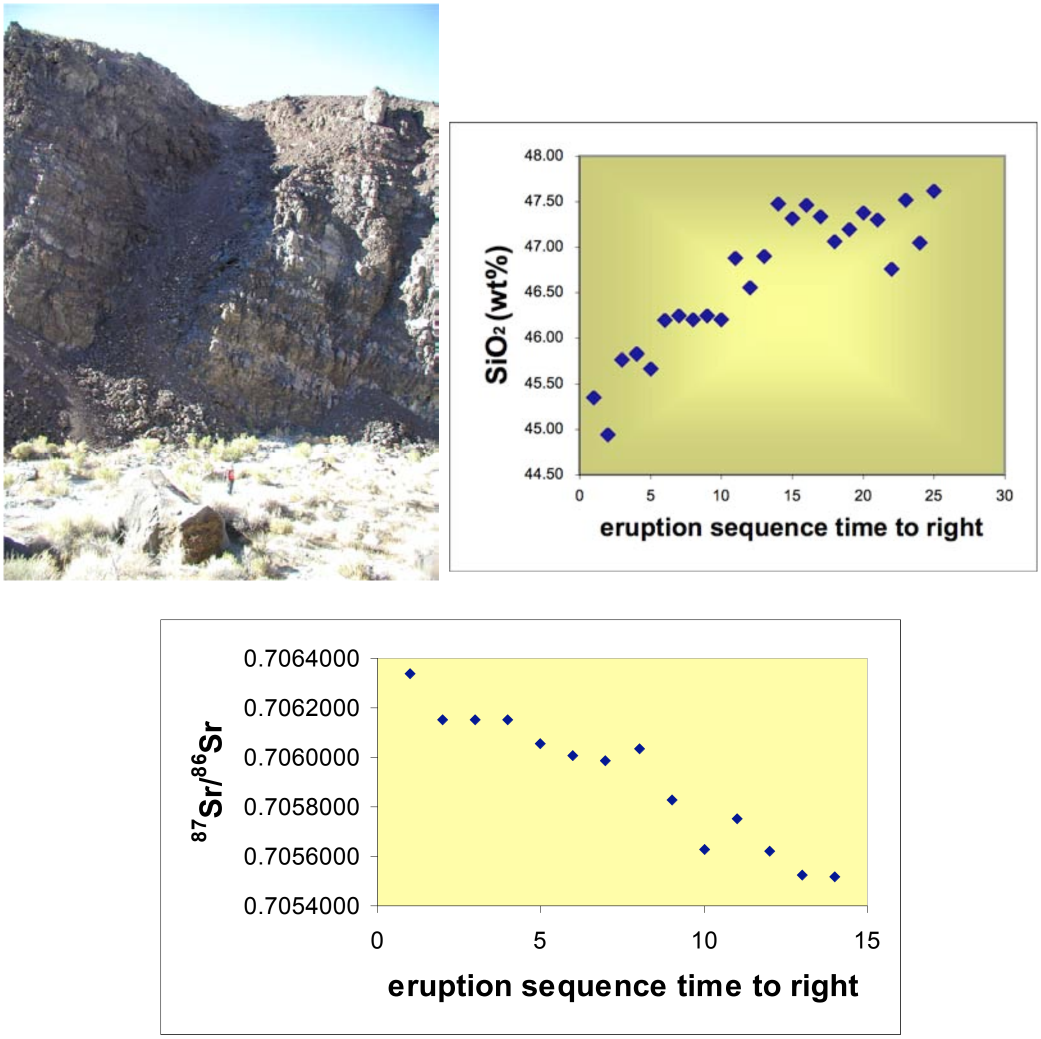


Figure 70. A Photo showing part of the volcanic stratigraphy in Papoose canyon (humanoid for scale); B, chemical and C, isotopic time sequences (Blondes et al., in labor).

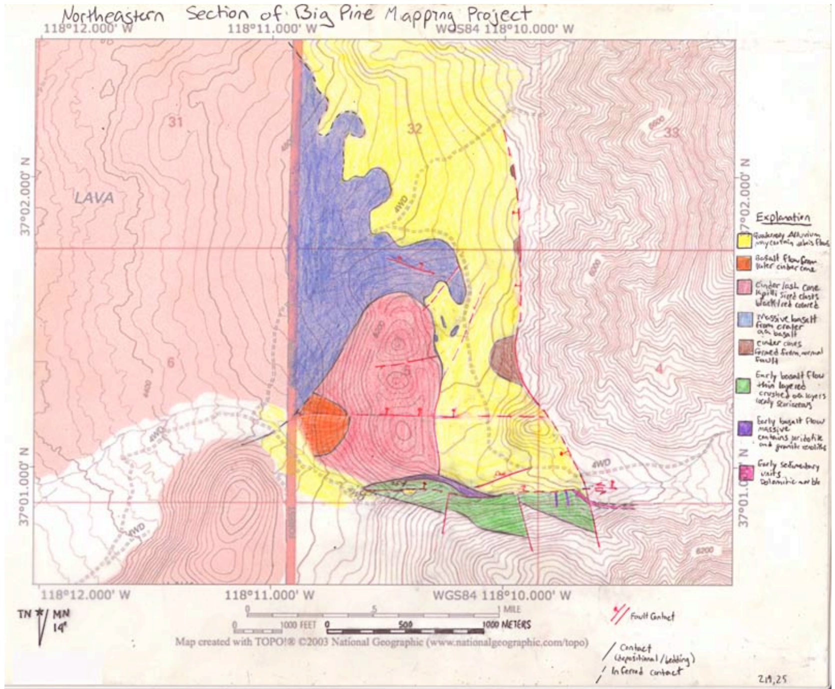


Figure 71. Preliminary (and not very good) geologic map of the Papoose Canyon area (Jen McGraw, 6/06).

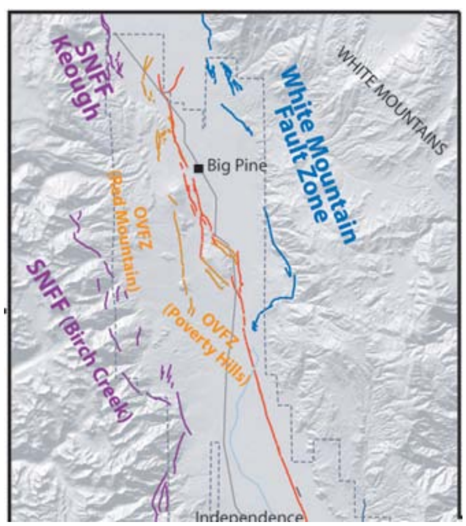


Figure 72. Active fault in the BPVF area (from Jayco in press, 2006). Note the releasing bend from the Owens Valley system (red) to the White Mountain fault system (blue).

13-1 Big Pine Volcanic Field

Travel east on Whitney Portal Road to Hwy 395 in Lone Pine. Turn left. Proceed ~60 miles north to Fish Springs Road on the left, opposite the Tinemaha Reservoir (on the right). Proceed north on Fish Springs for approximately one mile, just past the cinder cone on the left. As we cross over a culvert (culvert may have been removed), we will see a dirt road on the left that will take us to the cinder cone. The Fish Springs cinder cone is approximately 314 thousand years old. It is partially quarried. Break open the volcanic bombs and you will notice that many are cored with granite and contain obvious inclusion of granite.

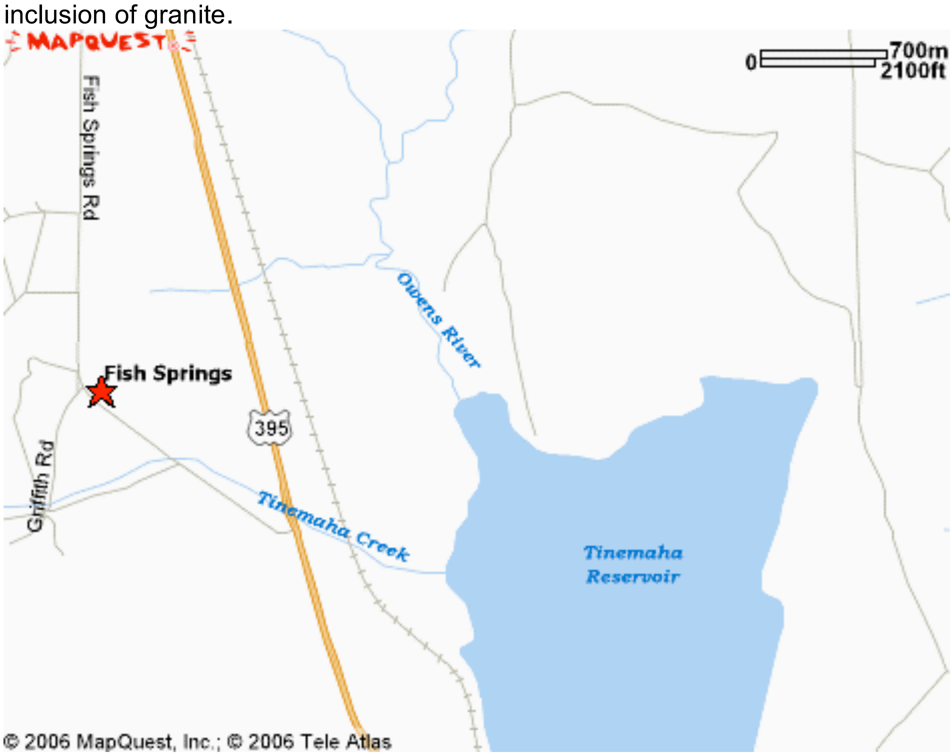


Figure 67: Map of Fish Springs Area

Next, Return to Hwy 395, turn right. Proceed south to Aberdeen Station Road, turn left. Go over the Owens River Bridge, and take the dirt road that goes to the north of Split Mountain. Then follow directions for each stop.

STOP 1. Western edge of Split mountain – quick stop to observe the normal fault that splits… Split Mountain. Source of the magnitude 6.1 earthquakes of early April 1872 (and numerous aftershocks) was close to this spot.

STOP 2. Quick stop after the only major switchback in the road. Use this to overview BPVF features on the Sierra side, including the 1 Ma rhyolite dome several cinder cones and associated flows, as well as Crater Mountain.

STOP 3. Active fault scarp (Figure 73) down-dropping dirt relative to the two 90ky a'a flows (both exposed here) that crop out throughout the area (the purple unit in Jen McGraw's map Figure 71). Great place to examine a cross section through a lava flow (… or two).

Figure 73. Normal offset along the Arizona fault at stop 3. The black flows (there are two) in the fault's footwall are 90ky old (Peter Reiners, supersecret unpublished He age data not to be shared with anyone or else we'll all be killed). Minimum slip rate can be determined by students in the field.

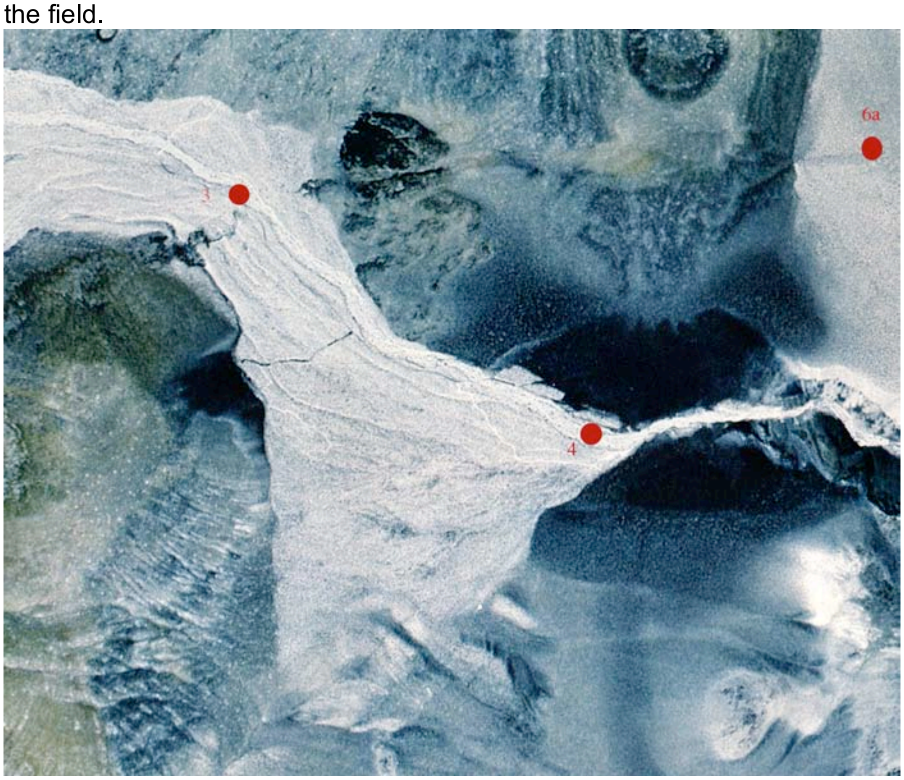


Figure 72. Orthophoto of stops 3, 4, and 6. One can see clearly the active "Arizona" fault (connecting points 3 and 6a), as well as the narrows of Papoose Canyon beyond stop 4.

Stop 4. End of the dirt road. Park near the abandoned equipment used by the Caltrans to mine cinder here (long time ago). The black colored half cinder cone north of the stop is a distinctive mappable unit and was suggested by volcanologist Liz Holt to represent products of sub-aqueous eruption (a.k.a. plinian), when a lake was present. Whatever, you be the judge. Figure 9 above shows that the end of the road is just west of where the canyon becomes narrow for about a mile or so. Walk this mile; it is pretty (Figure 10), exposes the volcanic stratigraphy of over 50 flows from unnamed SE cone on the south side of the canyon and the sequence of massive 90 ka flows on the north side of canyon (Figure 11). The narrows themselves follow a normal fault. You will see plenty of mantle and some crustal xenoliths during the second half of the narrows. There are lots of olivine pyroxenite, wehrlites and things that people classify as cumulates. Lherzolites are present too.

Figure 73 View of Papoose Canyon narrows from the cinder cone to the north.

Figure 74. Basaltic flows in the Papoose Canyon narrows – the Sierran crest in the background. Right - more of the same; excited students look for xenoliths.

Figure 75. Peridotite and pyroxenite xenoliths from Papoose Canyon. (Right picture missing)

13-2 Poleta Folds

Directions from Big Pine Volcanic Field to the Poleta Folds
Proceen north on Hwy 395 to Big Pine. At the intersection of Hwy 395 with Hwy 168, turn right (east) onto Hwy 168. Proceed north to Westgard Pass and the Poleta Folds area.

Located in the White-Inyo Mountains in the western Deep Springs Valley, the folded shelf sediments of the Cambrian Poleta Formation are found to occur at various sites on an oval, nearly flat-floored valley called Papoose Flat (figure 76). This flat is dotted with knobs rising from the valley floor and is part of the Papoose Flat pluton (body of quartz monzonite Cretaceous in age), which includes the Cambrian marine sedimentary rocks. These strata comprise a sequence of shallow-water sediments including quartz arenite, slaty siltstone, and limestone, which were deposited in a subtidal to tidal environment flanking the Cambrian continental margin. The structure of the area is locally that of an elongate dome with long axis oriented N40E, which is cut by at least two generations of faults. Stereogram analysis of these structures reveals principal stress orientations involved in their formation came from at least three separate directions at different times; the folding occurred first, followed by E-W faulting and then NW-SE faulting. Although there is considerable uncertainty as to the origin of stresses, which gave rise to the folding, intrusive bodies, which transect the folds and are not deformed, constrain the age of deformation to before 150 Ma. The Paleozoic Antler and Sonoman orogenies may have caused the folding exposed. Stresses associated with the Mesozoic formation of the Last Chance thrust seem a more likely candidate as they more accurately match the geometry and orientation of the Poleta Folds area. Forced intrusion of plutonic bodies to the north and south of the area may have caused some of the faulting.

The nature of the wide distribution of the Poleta folds is due to the fact that during the intrusion of the pluton, the cutting across of the sedimentary rocks caused many of these rocks to spread apart. On the western margin of the pluton, Cambrian rocks have been metamorphosed, stretched, and thinned dramatically (figure 76) so that they are a tenth as thick as they are elsewhere in the range. Stratigraphy of the unmetamorphosed Poleta Formation of normal thickness can be traced into the stretched, thinned, and metamorphosed equivalents that wrap around the western side of the pluton.

Suggested Readings
Norris, R.M. and Webb, R.W. 1990. Geology of California. New York: John Wiley & Sons, Inc., 541p.

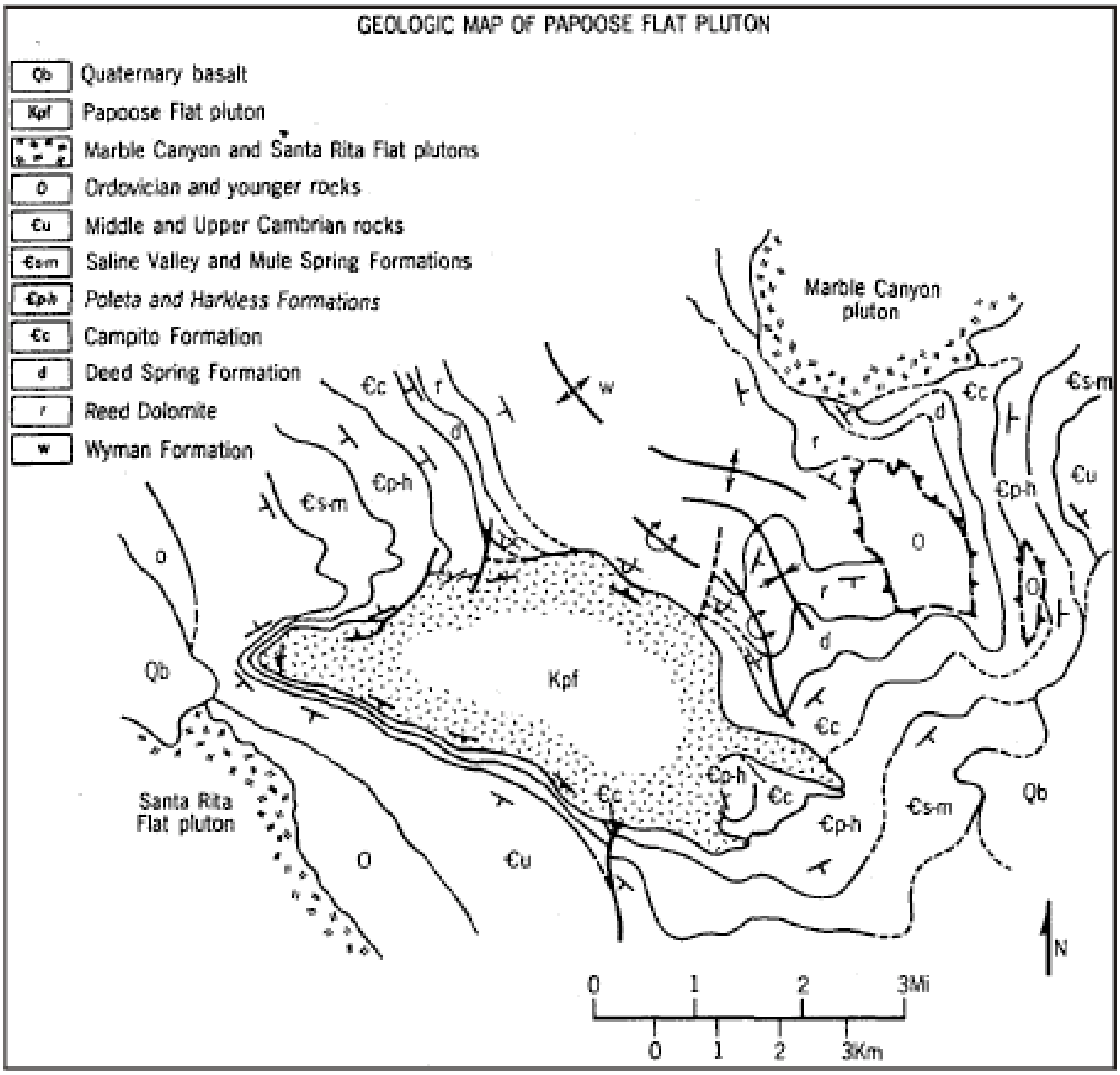
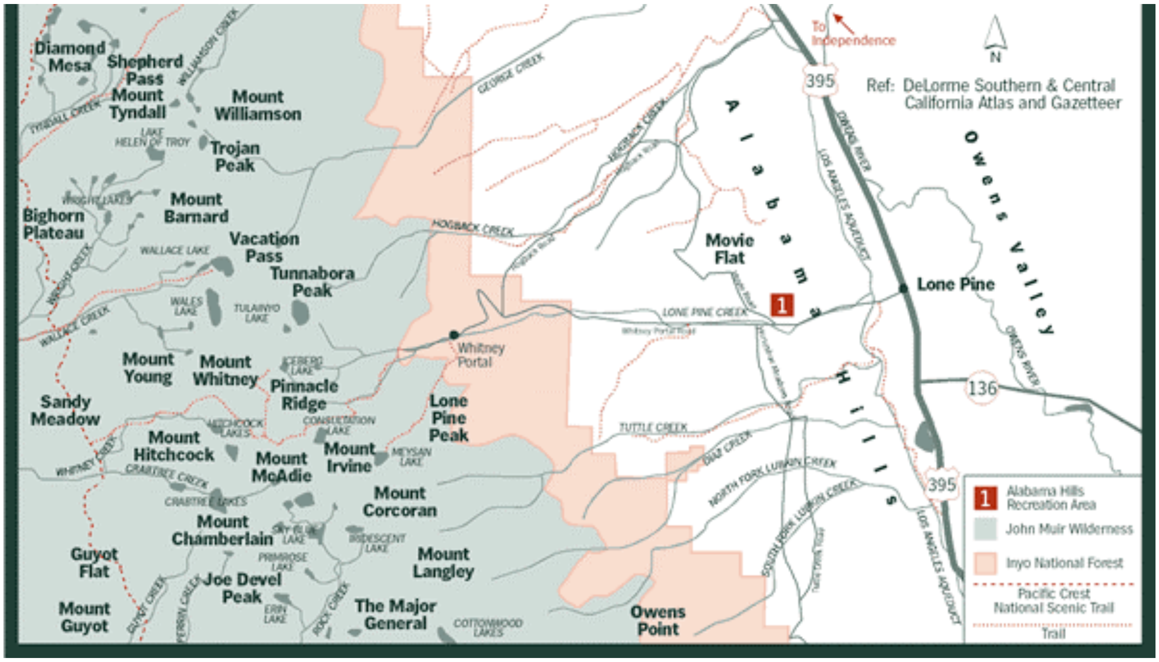


Figure 76 - Geologic map of Papoose Flat area showing Poleta Folds around Papoose Flat pluton (Norris and Webb, 1990, p. 217, Fig. 6-20a)

13-3 Alabama Hills

Directions from Poleta Folds to Alabama Hills: Return to Hwy 395, turn left. Proceed back to Lone Pine. Turn right on the Whitney Portal road and proceed approximately 3 miles west of Lone Pine. Follow signs for Alabama Hills Movie Road.



Map from http://www.cityconcierge.com/travel/alabama.html

Figure 77 Diagram of the Alabama Hills (from Jessey and Wall).

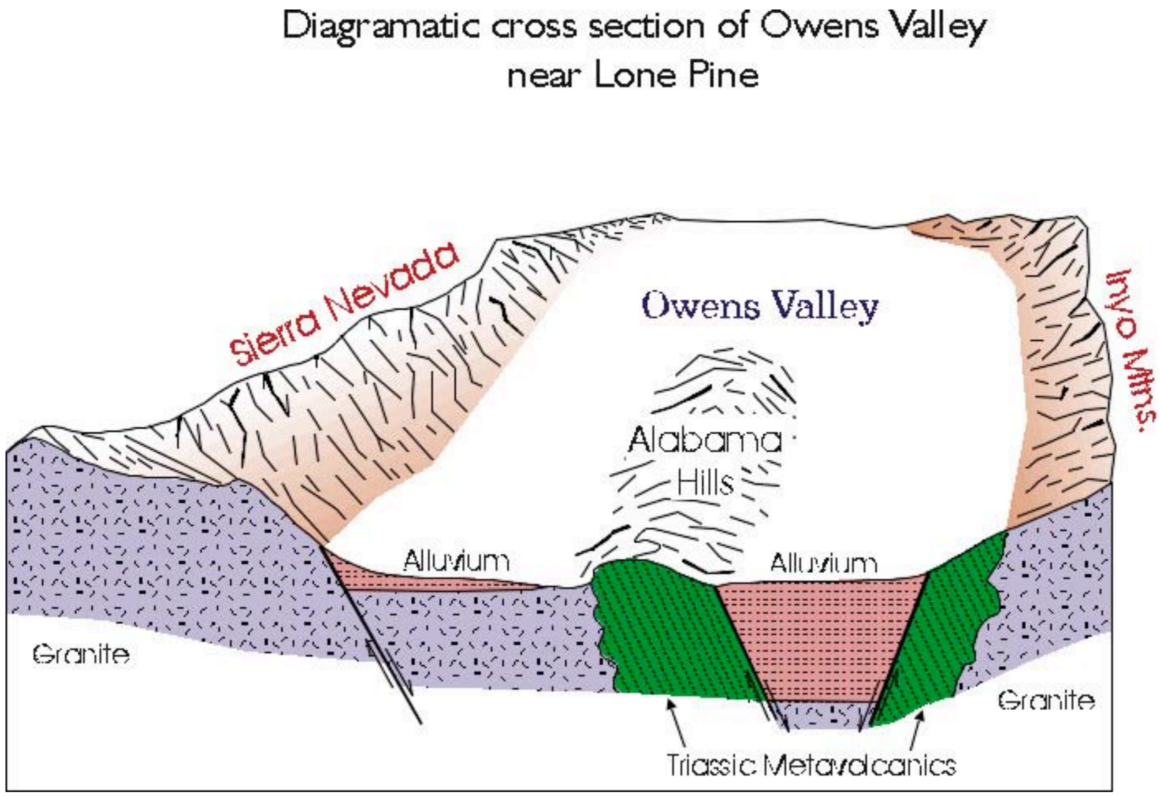


Figure 78 Diagram of the Alabama Hills (from Jessey and Wall).

GeologyThe Alabama Hills are a block of Triassic/Jurassic metavolcanics (150-200 Ma), orange and drab weathered, intruded by Mesozoic granite (90 Ma), which weathers to potato-shaped large boulders, many of which stand on end due to spheroidal weathering acting on many nearly vertical joints in the rock. The cartoon sketch below illustrates the structural relationship of the Alabama Hills to the Sierra Nevada. On the eastern side of the Alabama Hills geophysical surveys suggest that the depth to bedrock is approximately 9,000', about the same as the elevation difference between the floor of Owens Valley and the summit of Mt. Whitney (14,495'). The base of the graben block thus sits nearly four miles below the crest of the Sierras. The fact that the Alabama Hills outcrop at all indicates fault motion must have been a combination of uplift of the Sierra Nevada Mountains with concurrent down dropping of the Owens Valley. (Jessey and Wall).

Normally, one would not expect such a feature in the middle of a rift valley. But also puzzling is that the granites of the Alabama Hills have a different composition than the surrounding granites. The granites in the boulders on either side of the Alabama Hills contain pink feldspars and are characterized by large chunks. These are the same granites that can be found in Mount Whitney and the Sierra Nevadas to the west. In the Alabama Hills, however, the plagioclase feldspars are more weathered, have smaller chunks, and have turned more to clay. Clearly, the boulders on the valley floor did not come from the Alabama Hills. It is reasonable to assume that the boulders on the valley floor, on either side of the Alabama Hills, originally came from Mount Whitney and the Sierra Nevadas. But how could rocks from Mount Whitney to the west get to the eastern side of the valley? The Alabama Hills are in the way, and boulders from Mount Whitney could not have crossed over them. The most likely explanation for this apparent inconsistency is that the Alabama Hills have only recently been thrust upwards, and that the granitic sediments east of the Alabama Hills were laid down there *before* the Alabama Hills were uplifted. (Meltzner, 1998).

Filming

The Alabama Hills are a popular location for television and movie productions (especially Westerns) set in an archetypical "rugged" environment. Since the early 1920s 150 movies and about a dozen television shows have been filmed here including Tom Mix, Hopalong Cassidy, Gene Autry, and the Lone Ranger. Classics such as *Gunga Din*, *Springfield Rifle*, and *How the West Was Won*, as well as more recent productions such as *Tremors* and *Joshua Tree* were filmed at sites known as *Movie Flats* and *Movie Flat Road*. In *Gladiator*, actor Russell Crowe rides a horse front of the Alabamas, Mount Whitney in the background, for a scene presumably set in Spain. (Wikipedia-Alabama Hills).

<u>13-4 Fossil Falls and Little Lake</u>

Directions from Alabama Hills to Fossil Falls: Return to Hwy 395 and proceed back to Line Pine. Continue south on Hwy 395 approximately 90 miles. Take the Cinder Road exit. Once you reach the parking area, there is a trail for you to follow, though not difficult, sturdy shoes are recommended due to its uneven surface.

Geology of the Area

To the south, a prominent V-shaped valley, Owens Valley, cut by the Owens River is present between the Coso Range to the east and the Sierras to the west (figure 79). Pleistocene and Holocene cinder cones and flows are exposed throughout the Owens Valley, particularly near Little Lake where basaltic flows and cinder cones dominate the landscape. The basalts occupying the valley have been age dated allowing a rough stratigraphy to be established. Three ages of volcanic activity are evident in the Little Lake area. The oldest of these, about 400,000 years old, is a prominent lava flow on the east side of the valley south of Little Lake. This flow extends for several kilometers and was erupted from a cinder cone south of Little Lake as the base of the Coso Range. The next youngest flow forms a low cliff on the east side of the lake and has an age of about 130,000 years. The youngest flow, a little over 10,000 years old, is related to Red Hill, a cinder cone north of Little Lake. This flow is the one cut by glacial Owen River at Fossil Falls (Norris and Webb, 1990). Although no thorough petrographic studies have been done on these basalts, they appear to be olivine tholeiites. Basalts of this composition occur at several other localities in the Owens Valley/Mammoth Lakes area.

Fossil Falls (figure 79) and surrounding area tells of interplay between climate change and volcanic activity. The falls are about 30 m high and show old plunge pools and potholes formed during the latest pluvial episode of the Pleistocene (Norris and Webb, 1990). The Falls formed when Owens River was dammed by a lava flow about 0.5 mi (0.8 km) east of the falls. Water accumulated in a shallow lake, now a playa (figure 80). An abandoned channel, 100 ft (30m) wide, is cut 6 to 10 ft (2 to 3 m) into the lava flow (Saint-Armand, 1987).

At Fossil Falls, a narrow canyon incises the riverbed and descends steeply to the west. It reaches a depth of 40 ft in the deepest part of the canyon. After 300 ft, the canyon widens into a smooth, flat-bottomed channel. Below this flat area, the canyon floor drops for about 30 ft. Water once flowed over the lip to a depth of about 3 ft. Evidence for this is seen by the scour marks on the sides of the canyon. In the flat riverbed, water has dissolved and undercut the sides of many of the boulders. This carving has resulted in the large potholes now visible; one vertical pothole is 2-3 ft in diameter and 12 ft deep. The lava of the falls is traversed by many fractures through which water descends. Water rarely runs over the top of Fossil Falls but following rains, it often descends through fissures in the basalt. A large mass of water did course through Fossil Falls between 2,000 and 4,000 years ago.

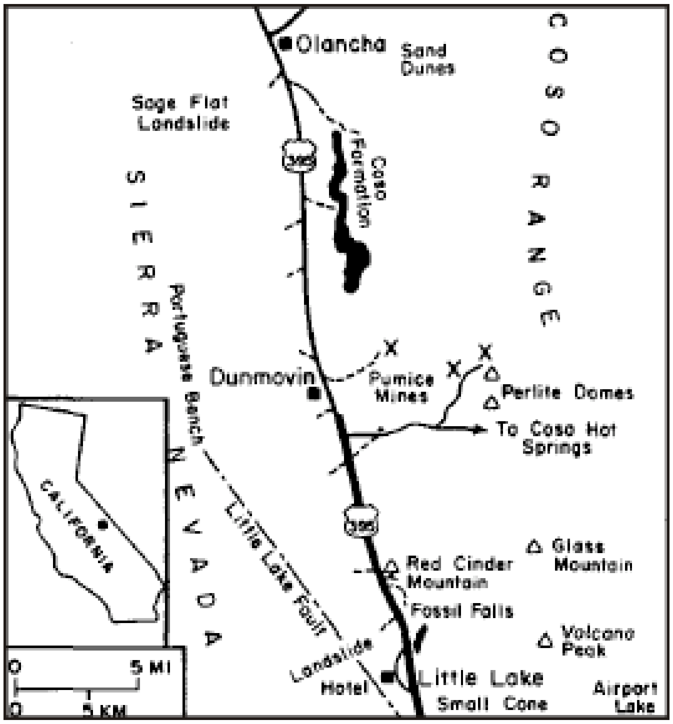


Figure 79 - Fossil Falls and Red Cinder Mountain are found on the east side of US-395 about 3 miles north of Little Lake (Saint-Amand, 1987, p. 143)

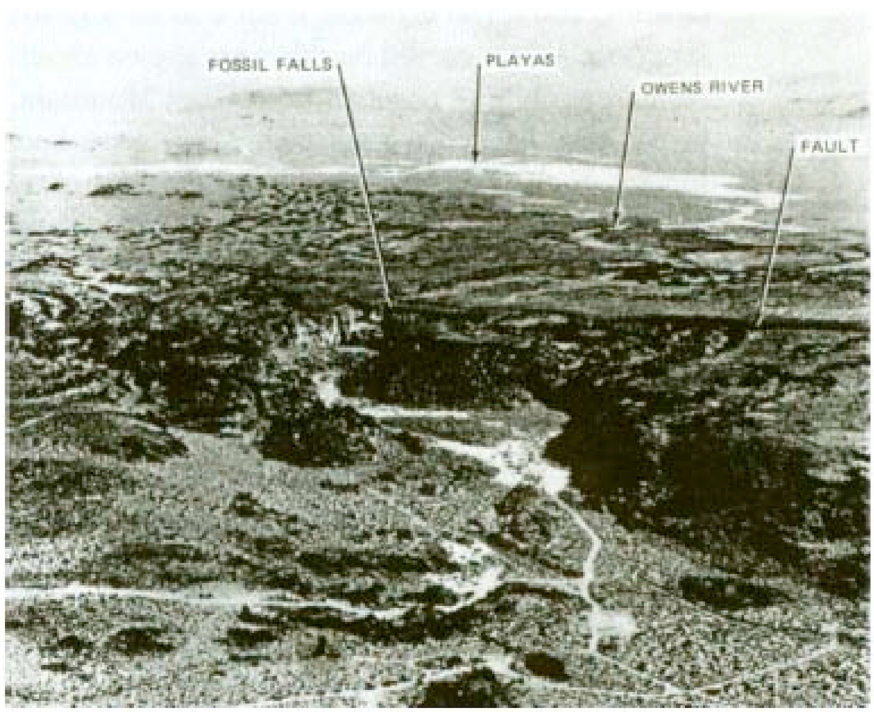


Figure 80 - Fossil Falls and the course of the Owens River, looking east. The river drained from the playas in the background. (Saint-Amand, 1987, p. 143)

Return to Lone Pine Campground via Hwy 395 and Whitney Portal Road.

Day 14 - Thursday, 31 August 2006

Summary: Leave Lone Pine Campground
Owens Lake
Death Valley
Return to Lone Pine Campground

Fig 81: Map from Lone Pine to Owens Lake to Death Valley

14-1 Owens Lake

Travel down Whitney Portal Road to Hwy 395 South. Turn left onto Hwy 190 going around the east side of Owens Lake. We will stop along the road at some point and discuss Owens Lake.

The Owens Lake area lies in a well known geologic and paleoclimatic setting. In the geological past, it was the first in a series of Pleistocene lakes that at times extended south and east to Indian Wells, Searles, Panamint, and Death Valleys, the floors of which are now dominated by playa lakes. The number of perennial lakes in that succession primarily reflected the amounts of precipitation falling in their collective basins, which included the high eastern slopes of the southern Sierra Nevada which drain into the Owens River, as well as the slopes of lower-elevation ranges that adjoin those lake's basins. Variations in wind, relative humidity, temperature, and other climatic variables that influence evaporation rates were also factors in determining lake sizes, but changes in them were less important than variations in precipitation. Published studies of exposed lacustrine outcrops, cores, and landforms have helped reconstruct the past histories of lakes in the downstream basins that were part of this formerly-extended drainage Owens Lake today lies in a well-known hydrologic setting. Its drainage area is one of the most thoroughly studied in the United States as a result of more than a century of measurements by scientists and engineers concerned with the water supply for the City of Los Angeles.

Geophysical studies have shown that more than 1.8 km of low-density sediments underlies Owens Lake's surface, meaning that a long record of valley-filling sediments of late Cenozoic age is likely to be preserved.

Holocene climates in the American Southwest have been substantially drier than Pleistocene climates during the past 30 k.y or more, yet historical records (up to about 1912, when the Owens River was diverted to Los Angeles) consistently describe Owens Lake as a perennial body of water. A perennial lake would have accumulated a lacustrine record that was both continuous and unaffected by subaerial erosion.

Before the end of the last great ice age of the Pleistocene Epoch (over 11,000 years ago), huge snow packs and glaciers covered the Sierra Nevada. Melting of this snow and ice sent enormous quantities of water down Owens River, filling the deep valleys and basins along its path to overflowing. Remnants of ancient beaches at the southern end of Owens Valley indicate that glacial Owens Lake was over 200 feet deep and covered nearly 200 square miles. Glacial Owens Lake ran south to China Lake, where it overflowed into vast Searles Basin and Panamint Valley, forming lakes estimated to be more than 600 feet deep. Some geologists believe that glacial Lake Panamint may have overflowed into Death Valley, where it joined forces with the Amargosa and Mojave Rivers to form ancient Lake Manly, over 600 feet deep. During thousands of years of evaporation the lakes gradually dried up, as enormous quantities of salts precipitated out in vast salt flats.

Before the Owens River was diverted into the Los Angeles Aqueduct in 1913, Owens Lake was a large, blue salt lake covering 100 square miles. During the late 1800s, a steamship crossed Owens Lake to carry lumber, mine timbers, charcoal and other supplies to the east shore, where it was packed up to the Cerro Gordo Mine near the crest of the Inyo Range. In 1913 a tramway was completed across the rugged 9,000 foot crest of the Inyo Range, east of Owens Lake. During its peak operation, the tram bucket brigade carried 20 tons of salt per hour from isolated Saline Valley on the east side of the range. Remnants of the ingenious salt tram can still be seen along the Owens Lake Loop

Even when the lake appears dry, a layer of brine occurs beneath the salt crust. It is fed (in part) by the Owens River and the tributaries that drain the snow-covered Sierra Nevada. Owens Lake had been gradually drying up for thousands of years, and was already saline when the Owens River was diverted to supply Los Angeles with water. Brine fly pupae (Ephydra), common insects of saline ponds and lakes, were an important food in the diet of local Paiute Indians. The pupae, which look like grains of rice, occur in enormous numbers and can still be found around the shoreline where there is standing water. They can also be found by the thousands, embedded in the salty crust.

The reddish coloration of Owens Lake is caused by microscopic, salt-loving bacteria, called halobacteria. A single drop of the brine contains millions of rod-shaped bacterial cells. The bacteria produce a red carotenoid pigment which is similar to that found in tomatoes, red peppers, pink flamingos, and in many colorful flowers and autumn leaves. Carotenoid pigments are also the source of Beta-carotene, an important antioxidant and the precursor of vitamin A. In fact, in some parts of the world, B-carotene is extracted from salt ponds containing red salt-living bacteria and algae. In the case of the halobacteria living in Owens Lake, the red pigment may protect their delicate cells from the intense desert sunlight.

If samples of the red brine from Owens lake are spun in a high speed centrifuge at 5,000 rpm, the water becomes clear as the red bacterial cells are forced to the bottom under about 3,000 g's. The bacteria may then be grown in a special nutrient agar containing at least 25 percent sodium chloride and incubated in a warm oven.

The exact chemical explanation for the extreme salt tolerance of these bacteria, and their need for salinity at least three to four times that of sea water, is very complicated. The cells themselves contain a very high internal salt concentration (primarily potassium and sodium), equal to or higher than their environment, otherwise, they would be rapidly dehydrated (plasmolyzed) in the brine. It has also been shown that the highly saline environment is essential for normal enzyme function within the cells, and to maintain the fragile protein coating or "wall" around the delicate cell membrane. In fact, if the salt concentration drops too low, the outer protein "wall" actually dissolves and the inner cell membrane disintegrates, thus destroying the cell

Halobacteria can thrive in concentrated brine nine times the salinity of sea water, and can even remain alive in dry salt crystals for years. In fact, their extreme tolerance for ordinary table salt (sodium chloride) makes them a nuisance to companies using solar evaporation ponds for the production of solar salt. Freshly produced solar salt is often contaminated with these organisms, and they occasionally cause spoilage of fish, meats, vegetables and hides when salt has been used in the preservation process. They may also cause an unsightly, pinkish discoloration of pickled foods known as "pinkeye" in salted fish and "red heat" in salted hides.

Halobacteria are placed in the "Archaebacteria," a group of unusual bacteria that survive under some of the most extreme conditions on earth. In fact, some biologists feel that these bacteria should be placed in their own Kingdom Archaebacteria, separate from the Kingdom Monera that contains most of the true bacteria. Heat-loving (thermophilic) Archaebacteria have been found thousands of feet deep at the bottom of the ocean, near steam vents where the water temperature is three times that of boiling water. They can live in this black world of boiling water without oxygen. It has been suggested that if any bacteria could survive on the surface of Mars, it might be a form similar to the Archaebacteria.

The salt crust and brine of Owens Lake is sometimes greenish, due to the abundance of another organism called Dunaliella. This is a unicellular green alga, much larger than the bacteria, though visible only under high magnification. Each individual oval or pear shaped cell has two whip-like tails or flagella at its anterior (head) end. The moving flagella propel Dunaliella through the water in a spiral motion. Under high magnification, numerous Dunaliella can be seen swimming among the gleaming, geometrically shaped crystals of salts. Dunaliella is clearly a green alga because of a distinct, green, cup-shaped chloroplast that occupies most of the cell. Except for coloring salt lakes red, the salt-loving bacteria probably seem insignificant to most people; however, they have been studied extensively in recent years by biologists and biochemists. A pigment has been discovered in the cell membrane of Halobacterium that is remarkably similar to the light sensitive pigment (rhodopsin) in the rod cells of human eyes which enables us to see in dim light. When we enter a dimly lighted room, it takes several minutes for our eyes to adjust as the pigment rhodopsin gradually increases in concentration. In fact, during World War II night-flying aviators sometimes wore special goggles just before the start of a mission. The goggles enabled the pilots to see and carry on normal activities while stimulating rhodopsin production in the eye for maximum night vision. The pigment in salt-loving bacteria (called bacteriorhodopsin) enables them to utilize sunlight for energy, just as green photosynthetic plants are able to capture the sun's energy. Future studies of these amazing solar-powered bacteria may lead to new and more efficient uses of the sun as a source of energy.

http://www.desertusa.com/mag98/april/owens/owenslake.html
http://pubs.usgs.gov/of/1993/of93-683/1-intro/intro.html

14-2 Death Valley

Continue along Hwy 395 east toward Death Valley National Park. Use Figure 82 to help you find the stops. Our first stop will be at the sand dunes and Stovepipe Wells near the center of the map.

Death Valley National Park (Figure 82) has more than 3.3 million acres of spectacular desert scenery, interesting and rare desert wildlife, complex geology, undisturbed wilderness, and sites of historical and cultural interest. Bounded on the west by 11,049 foot Telescope Peak and on the east by 5,475 foot Dante's View, Badwater is the lowest point (-282 feet) in the western hemisphere.

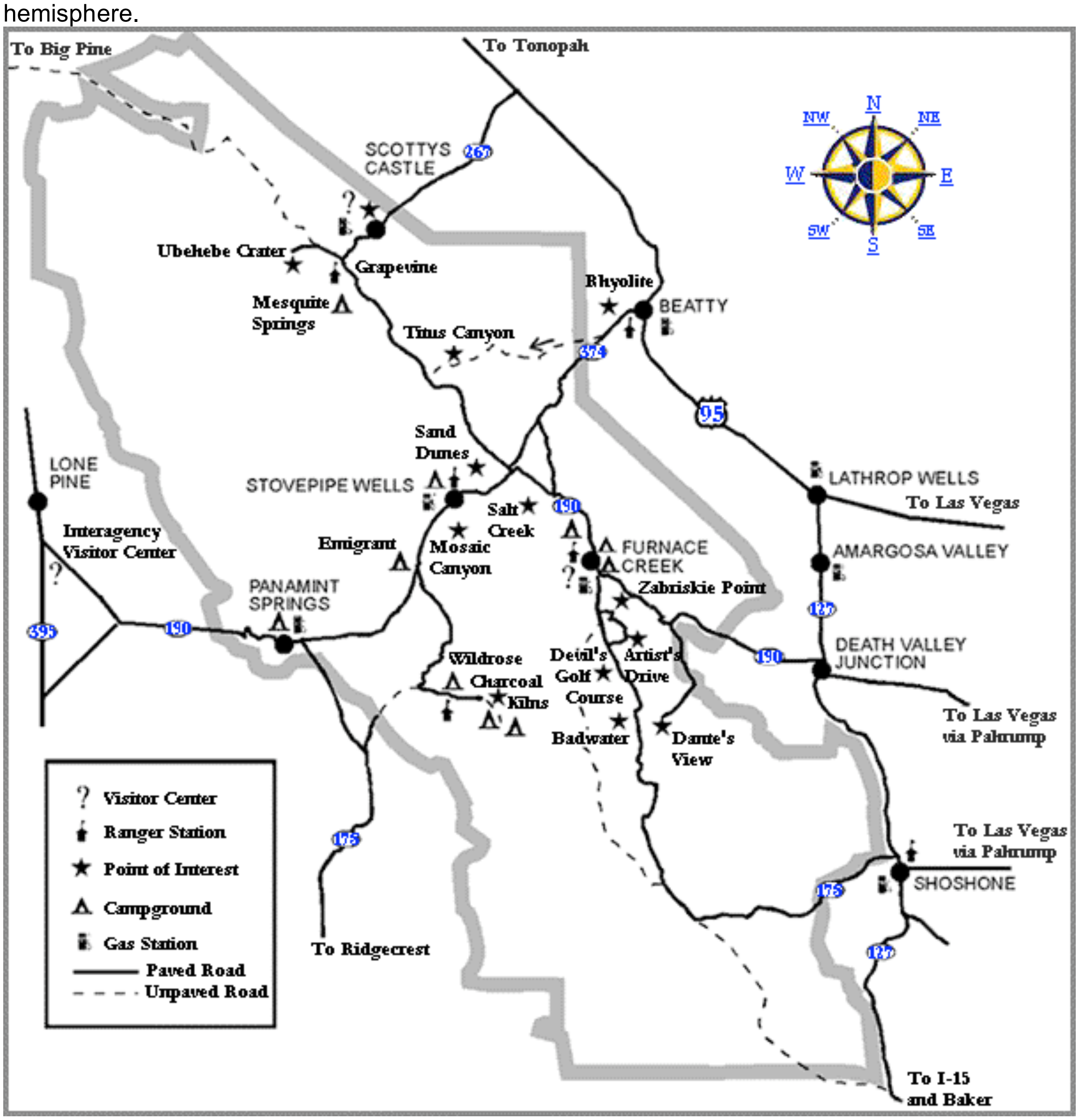


Figure 82- Death Valley National Park (From http://www.reviewjournal.com/images/recreation/map_deathvalley_area.gif)

Geology of the Basin and Range:

The oldest Basin and Range rocks are a complex of early to middle Proterozoic shists and gneisses, with associated igneous rock. Good exposures of these rocks form the western Black Mountains (the steep eastern wall of southern Death Valley; dated at 1.8 billion years).

Metamorphosed sedimentary rock (Proterozoic) ranges to a few thousand metres thick and consists of regularly bedded sandstone, siltstone, shale, limestone, dolomite and conglomerate. The rocks are easily seen in Funeral, Nopah and southern Panamint mountains (Figure 61).

The Pahrump group, estimated from 800 to 1200 million years old, contain microfossils (cyanobacteria or blue-green algae) and *stromatolites*. Unconformably overlying the Pahrump Group is a thick sequence of late Proterozoic carbonate and clastic rocks consisting of Noonday dolomite (300 m), Johnnie Formation (750 m) and the Stirling Quartzite (600 m). Above these is a thick sequence of Cambrian rock (5200 m). In the White-Inyo Mountains, this sedimentary package is the boundary between the Proterozoic and the Phanerozoic; and, the Cambrian contains the first occurrence of invertebrates with hard shells (Stirling Quartzite). A composite geologic column is included as Figure 83, and a geologic column with events is included as Figure 84.

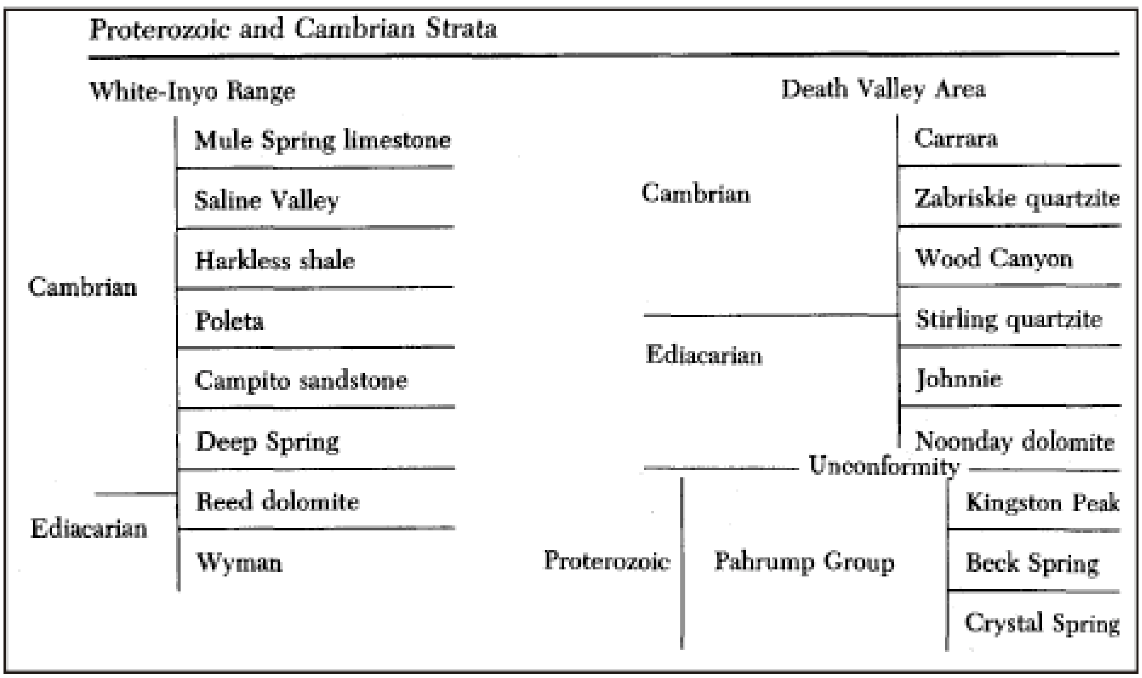


Figure 83 - Proterozoic and early Paleozoic strata, Basin and Range province (Norris and Webb, 1990, 6-6)

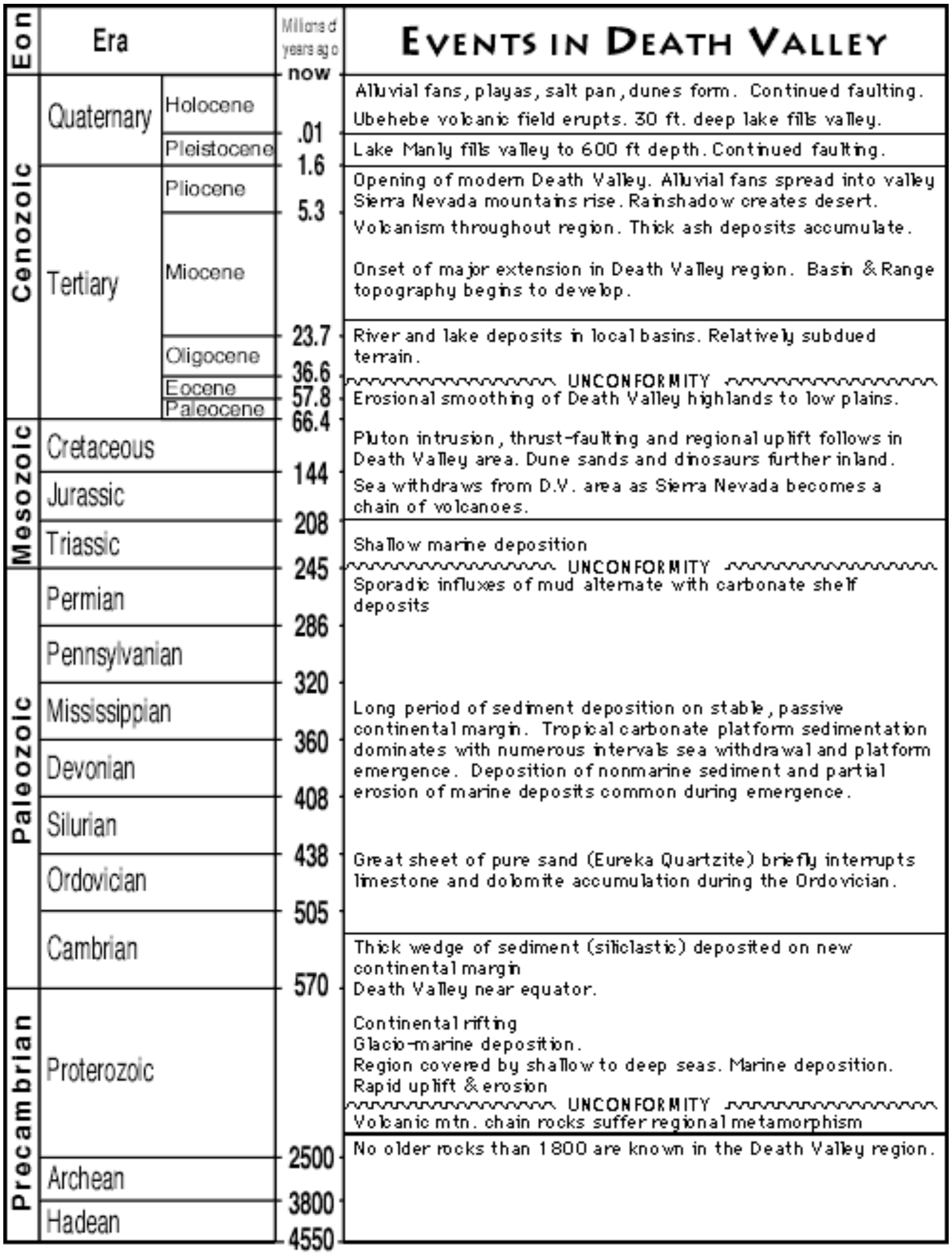


Figure 84 - Events in Death Valley (From
http://geology.wr.usgs.gov/docs/usgsnps/deva/time8.html

14 - 3 Devil's Cornfield/Sand Dunes, Stovepipe Wells Village

The Devil's Cornfield near Stovepipe Wells has isolated high tufts of the arrow-weed plant separated by stretches of empty sand. Deflation of sand dunes and silt left these clumps of arrow-weed atop hummocks of sand and silt which have been anchored by roots (figure 85).

Figure 85 - Devil's CornField (www.americansouthwest.net/california/death_valley/south.html)

14 – 4 Furnace Creek Visitor Center

Located in the center of the park, the Furnace Creek Visitor Center houses museum exhibits, a visitor information desk, and the Death Valley Natural History Association book store. The Furnace Creek Visitor Center is open daily from 8 a.m. to 6 p.m.

Death Valley-Furnace Creek Fault System

A predominately strike-slip fault zone, which is known as the North Death Valley fault zone to the north of Furnace Creek. The Furnace Creek portion extends from the vicinity of Furnace Creek Ranch, eastward to the Resting Spring Range (Figure 86). Although the Furnace Creek Portion is often concealed by alluvial gravels, the North Death Valley portion displaces alluvial gravels to the south and east of Stovepipe Wells, and northward along the Grapevine range. It is marked by alluvial fans (as much as 15 m high) and prominent lines of springs from Scotty's Castle road to Ubehebe Crater.
The Southern Death Valley fault zone joins the Garlock fault zone at the southern end of Death Valley and helped form a pull-apart-basin in the Central Death Valley fault zone, which is well exposed along the western base of the Black Mountains. Large vertical movement has occurred here, producing a precipitous west-facing escarpment near the deepest part of the Valley (Norris and Webb, 1990).

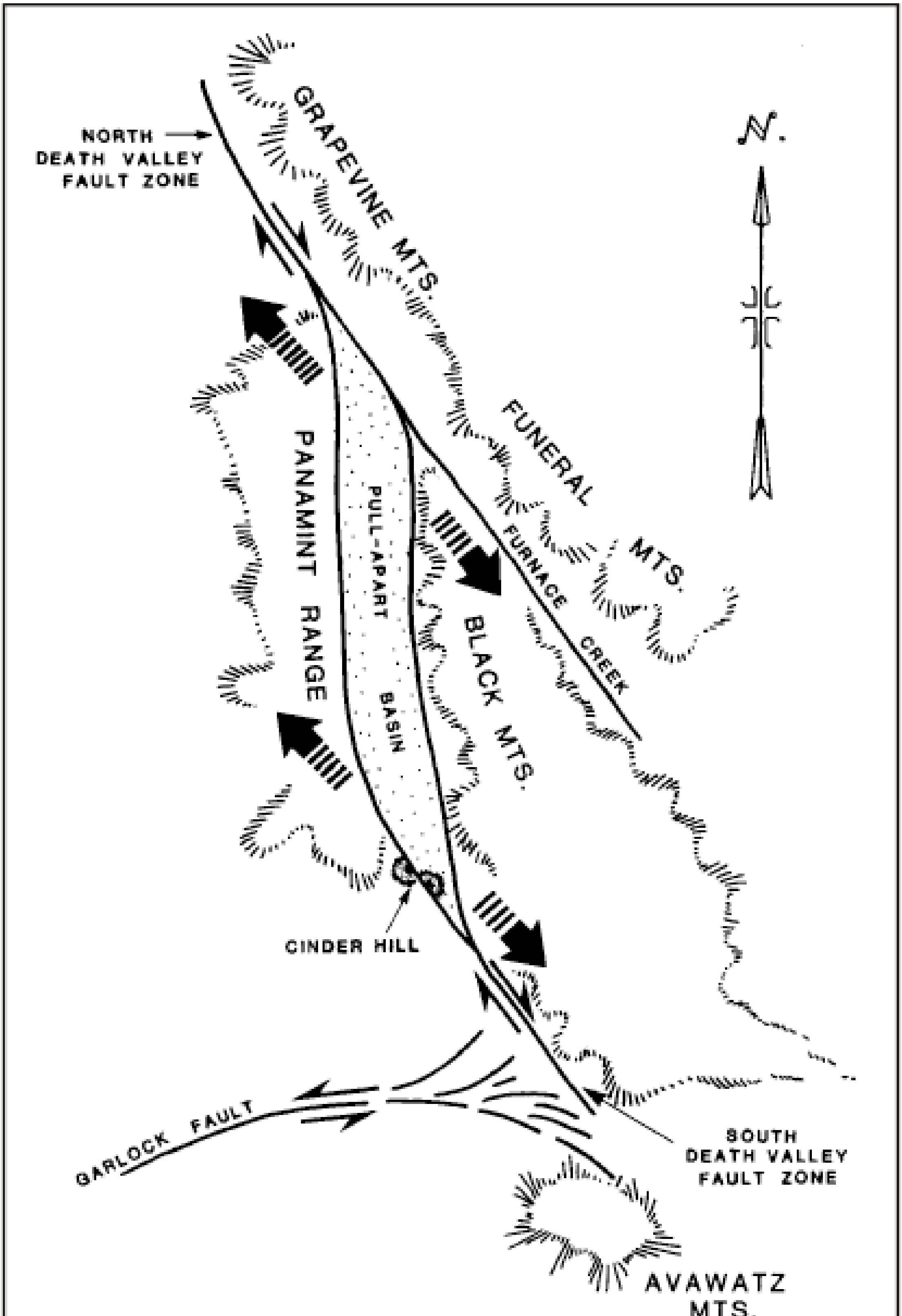


Figure 86 - Main structural features in the Death Valley area (Norris and Webb, 1990, p. 195, Fig. 6-14; after Wright and others, 1974)

Desert oasis

Sitting on top of a remarkably symmetrical alluvial fan lies Furnace Creek (Figure 87), the hub of Death Valley National Park. Furnace Creek is also at the center of a controversy over water in this parched environment. Death Valley averages less than 2 inches of rainfall each year. Yet even here, in the hottest, driest spot in North America one can find oases of life. The Furnace Creek area is such an oasis — one of those rare spots in the desert where springs rise out of the rock, providing life-sustaining water for desert plants and animals. Surface water is at a premium here, so it's no wonder that wetlands are among the rarest habitat types in the valley. The scarce springs and surrounding lush oases support thriving plant communities and attract a wide variety of animals.

The same water sources that provide lush habitats for plants and animals also attract people. Many of the springs, streams, and marshes within Death Valley National Park have already been developed to support human activities. In the early 1900's people flocked to resorts built around natural springs thought to have curative and restorative properties. The spring at Furnace Creek was harnessed at that time to develop the Furnace Creek Ranch resort. As water was diverted for resort use, the marshes and wetlands around Furnace Creek began to wither and shrink.

Figure 87 - Furnace Creek alluvial fan
(From http://geology.wr.usgs.gov/docs/usgsnps/deva/ftfur1.html; Photo by Martin Miller)

14 - 5 Artist Drive; Dante's View

More unusual and varied colours are visible in the rocks and canyons along Artist's Drive (Figure 88) - a narrow paved one-way loop on which travel is possible south to north. Often the road bends sharply and crosses steep gullies, resulting in a rather tortuous journey with a top speed of only 15 mph, but this is a very rewarding trip with much unusual scenery - particularly striking is **Artist's Palette**, with curved bands of clayish rock in hues of green, white, pink and black. There are plenty of parking places, to allow for taking photographs and exploring on foot, and the drive should be passable by all but the largest of vehicles. The route is only 9 miles long but takes at least half an hour to drive. As it climbs into the hills, there are great views across the salt flats below (www.americansouthwest.net/california/death_valley/south.html).

Figure 88 - Black Mountains along Artist's Drive is made up of the multicolored rock of the Artist Drive Formation (Fromhttp://geology.wr.usgs.gov/docs/usgsnps/deva/ftfur1.html; Photo by Tom Bean, NPS)

The curvy, one-way, one lane Artist's Drive leads you up to the edge of the Black Mountains. Artist's Drive rises up to the top of an alluvial fan fed by a deep canyon cut into the mountain. As you make your way up to the mountain face you'll dip up and down, roller coaster-like as the road dips into ravines carved into the fan by Death Valley's occasional, but intense flash floods. The narrow road runs high up onto the fan, with views of the strikingly white salty floor of Death Valley in the distance

14 - 6 Zabriskie Point

"Like the Garlock, the Death Valley and Furnace Creek zones are composed of *en echelon* faults that incorporate elongate, narrow slices of crust. Where right-slip faults step to the right, the crust is stretched between the fault strands and a pull-apart-basin is formed. Where right-slip faults step left, the opposite occurs and compression produces ridges more or less at right angles to the faults. Such geometries produce varied topography and, in the relatively down-dropped blocks, preserve younger rocks that elsewhere are largely destroyed by erosion. Colorful examples occur at Zabriskie Point and Artists Drive, where thick sections of mostly yellow, late Tertiary non-marine Furnace Creek Formation are well exposed between strands of the Death Valley and Furnace Creek fault zones." (Norris and Webb, 1990, p. 197)

As well as being the title of a curious film by Antonioni (1969), Zabriskie Point (Figure 89) is an elevated overlook of a colourful, undulating landscape of gullies and mud hills at the edge of the Funeral Mountains, a few miles from Death Valley - from the viewpoint, the flat salt plains on the valley floor are visible in the distance. In the past it was possible to drive right to the edge of the overlook, and several minutes of the film was set there, but since then a new larger car-park has been constructed lower down and visitors now have a short walk uphill. Most people do little more than briefly admire the scenery, which is best at sunrise, but it is possible to climb some of the adjacent hills to get a better overall view, or wander down amongst the variegated dunes. A foot-path leads through the mounds, down a ravine and into Golden Canyon after 2 miles (www.americansouthwest.net/california/death_valley/south.html).

Figure 89 - Zabriskie Point (www.americansouthwest.net/california/death_valley/south.html)

14 - 7 Central and Southern Death Valley

The oldest rocks - Relics of the Precambrian world in Death Valley

The steep face of the Black Mountains(Figure 90) is made up of some of the oldest rocks in Death Valley. These 1.7 billion-year-old Precambrian rocks are the remnants of an ancient volcanic mountain belt with flanking deposits of mud and sand. About 1.8-1.7 billion years ago, these volcanic and sedimentary rocks were severely metamorphosed — altered, recrystallized, and partially remelted by the Earth's internal heat and by the load of overlying younger rocks. The original rocks were transformed to contorted gneiss, making their original parentage almost unrecognizable.

11 million years ago, these venerable rocks were injected with magma that solidified to form the Willow Spring pluton. The diorite to gabbro composition of the Willow Spring pluton blends well with the dark Precambrian gneiss, so you'll have to look carefully to see the contact between the two rock types.

14 - 8 Badwater

The low, salty pool at Badwater, just beside the main park road is probably the best known and most visited place in Death Valley. The actual lowest point (-282 feet) is located several miles from the road and is not easily accessible, but a sign in front of the pool proclaims it too to have an elevation of -282 feet, and it is here that everyone comes to take photographs.

Sea Level: There is not much else to see apart from an orientation table, identifying many of the surrounding mountains. High in the rocky cliffs above the road, another sign reads 'SEA LEVEL', giving a good indication of just how low the land is. Far above this, the overlook at Dante's Peak has imposing views over Badwater and the surrounding desert.

Salt Pools: Several salt trails and shallow seasonal streams lead towards other pools out across the valley. During occasional rainy periods, a large shallow lake forms, several miles across and only a few inches deep, but most of the water soon evaporates or sinks below ground. Badwater never dries out completely, and even manages to support a unique species of fish - the Death Valley pupfish, a small bluish creature which has evolved to survive in the hot saline conditions.

South of the salt pools, the seasonal Amargosa River meanders for 30 miles via several routes towards the mouth of the valley, before sinking into the sand.
Heat: Apart from the high temperatures, one unusual feeling is the heaviness of the air - all movement seems more laboured and difficult than usual. The shade temperature here can be 125 °F. It is an unforgettable experience to wander a little way out onto the salt flats, and just stand for a while in the stifling heat.

Figure 90 - The steep face of the Black Mountains rises from the valley floor. Few visitors realize that these mountains are made up of some of Death Valley's oldest rocks. (http://geology.wr.usgs.gov/docs/usgsnps/deva/ftfur1.html; Photo by Ray Nordeen, NPS)

Ancient rocks - youthful mountains: The deep, narrow gorge of Titus Canyon (Figure 91) cuts into the steep face of the Grapevine Mountains. Although the mountain range was uplifted quite recently, geologically-speaking, most of the rocks that make up the range are over half a billion years old.

Tropical seas The gray rocks lining the walls of the western end of Titus Canyon are Cambrian age (570-505 million years old) limestones. These ancient Paleozoic rocks formed at a time when Death Valley was submerged beneath tropical seas. By the end of the Precambrian, the continental edge of North America had been planed off by erosion to a gently rounded surface of low relief. The rise and fall of the Cambrian seas periodically shifted the shoreline eastward, flooding the continent, then regressed westward, exposing the limestone layers to erosion.
The Earth Shook, the Sea Withdrew

Mesozoic Death Valley summary: During Late Paleozoic and Mesozoic time (225 - 65 million years ago), the Death Valley landscape changed dramatically. To the west, the collision of tectonic plates changed the quiet, seacovered continental margin into a zone erupting volcanoes, uplifting mountains, and intense compression. A deep trench formed when the oceanic Pacific plate began to sink (subduct) beneath the more buoyant continental rock of the North America plate. A chain of volcanoes rose through the continental crust parallel to the deep trench. Thousands of feet of lavas erupted, pushing the ocean over 200 miles to the west. The Death Valley region was no longer coastal real estate, as it had been for the previous billion years.

Most of the volcanic activity was centered just to the west of Death Valley. The deep magma chambers feeding the volcanoes eventually cooled and solidified, forming the granitic rocks now exposed in the Sierra Nevada Mountains. A few of these granitic bodies intruded Death Valley's Cottonwood Mountains, but these rocks are not easily accessible. One of the small granitic plutons emplaced near the end of the Mesozoic Era created one of the more profitable precious metal deposits in Death Valley. Skidoo was one of the rough-and-tumble towns that sprang up near the Death Valley region gold deposits. Death Valley itself was a broad, mountainous region during this time. While some of the Mesozoic rocks that formed in Death Valley are preserved, most were eroded as the region uplifted.

The rise and fall of Death Valley’s mountain ranges and valleys (Figure 92)
Dante's View provides a spectacular look at Death Valley's remarkable scenery. From this vantage point you can view distant 11049 ft. Telescope Peak to Badwater, the lowest point (-282 ft.) in the Western Hemisphere. This difference in elevation is a staggering 11,331 feet (3455 m) — the greatest topographic relief in the conterminous U.S. This striking topography is a product of Death Valley's very active fault system. Death Valley National Park lies in one of the youngest and most active parts of the Basin and Range province. The term "Basin and Range" is taken from the unique character of this province's landscape. Here, steep, elongate mountain ranges alternate with flat, dry, desert valleys in a pattern that extends from eastern California to central Utah, and from southern Idaho into the state of Sonora in Mexico.

Figure 91 - Titus Canyon can be seen as a deep gash cut into the Grapevine Mountains. (http://geology.wr.usgs.gov/docs/usgsnps/deva/ftfur1.html; Photo courtesy of Martin Miller)

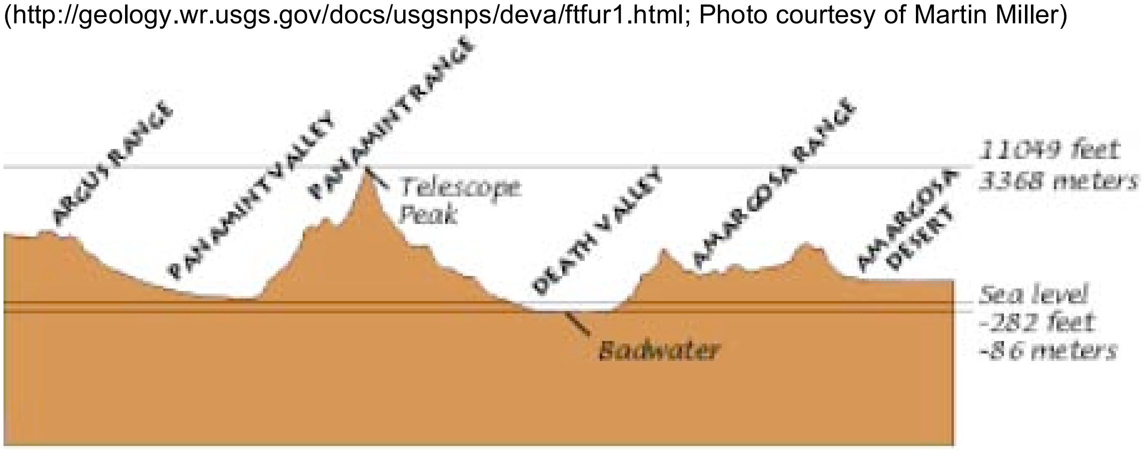


Figure 92 - A slice through the highest and lowest points in Death Valley National Park. Death Valley is the lowest "basin" of the Basin and Range Province (From http://geology.wr.usgs.gov/docs/usgsnps/deva/ftfur1.html)

Stretched to the breaking point The forces that created this distinctive topography begin deep beneath the Earth's surface. Tension created by movements of Earth's tectonic plates have stretched the rocky crust Basin and Range province to the breaking point. The entire region has been pulled apart, fracturing the crust and creating large faults. Along these roughly north-south-trending faults mountains have uplifted and valleys down-dropped, producing the distinctive alternating pattern of linear mountain ranges and valleys of the Bas
in and Range province.

Figure 93 - Dante's View Dante's View is a perfect place to see the results of Death Valley's very active fault system. (From http://geology.wr.usgs.gov/docs/usgsnps/deva/ftfur1.html)

14 - 9 Devil's Golf Course

Mineral music; It's an early summer morning. The temperature is rising fast. The air is completely still and the quiet is profound. But, listen carefully and you'll hear a sounds like tiny pops and pings. Bend your ear to the ground and the sound grows louder. The musical sound of literally billions of tiny salt crystals bursting apart as they expand and contract in the heat provides the backdrop for this salty story.

Not long ago, about 2000-4000 years ago during the Holocene, the climate was quite a bit wetter than today. It was so wet that water gradually filled Death Valley to a depth of almost 30 feet. The ancient peoples of Death Valley must have enjoyed centuries of abundant food in their lakeside homes.

The desert returns

These good times didn't last, however. The climate warmed, rainfall declined, and the shallow lakes began to dry up. Minerals dissolved in the lake became increasingly concentrated as water evaporated. Eventually, only a briny soup remained, forming salty pools on the lowest parts of Death Valley's floor. Salts (95% table salt - NaCl) began to crystallize, coating the muddy lakebed with a three to five feet thick crust of salt (Figure 94).

While the saltpan at Badwater periodically floods, then dries, Devil's Golf Course lies in a part of the Death Valley salt pan that is several feet above flood level. Without the smoothing effects of flood waters, the silty salt at Devil's Golf Course grows into fantastic, intricately detailed pinnacles. The pinnacles form when salty water rises up from underlying muds. Capillary action draws the water upward where it quickly evaporates, leaving a salty residue behind. The pinnacles grow very slowly, perhaps as little as an inch in 35 years. Wind and rain continually work to erode and sculpt the salty spires into an amazing array of shapes.

14 - 10 Shoreline Butte

Ice age Death Valley

During the Pleistocene ice ages, climate cooled and became wetter, glaciers grew in the Sierra Nevada Mountains. Rivers flowed into what are now dry deserts and lakes formed in many of down-dropped valleys of the Basin and Range. Shoreline Butte reveals evidence of a large lake, Lake Manly, that filled what is now the driest desert of the United States. Imagine a time during the ice age, between 186,000 - 128,000 years ago, when Shoreline Butte (Figure 95) was an island in a lake nearly 100 miles long and 600 feet deep!

Figure 94 - The lumpy salts you see here are the residue of Death Valley's last significant lake, which had evaporated by 2000 years ago (From http://geology.wr.usgs.gov/docs/usgsnps/deva/ftfur1.html Photo by Martin Miller)

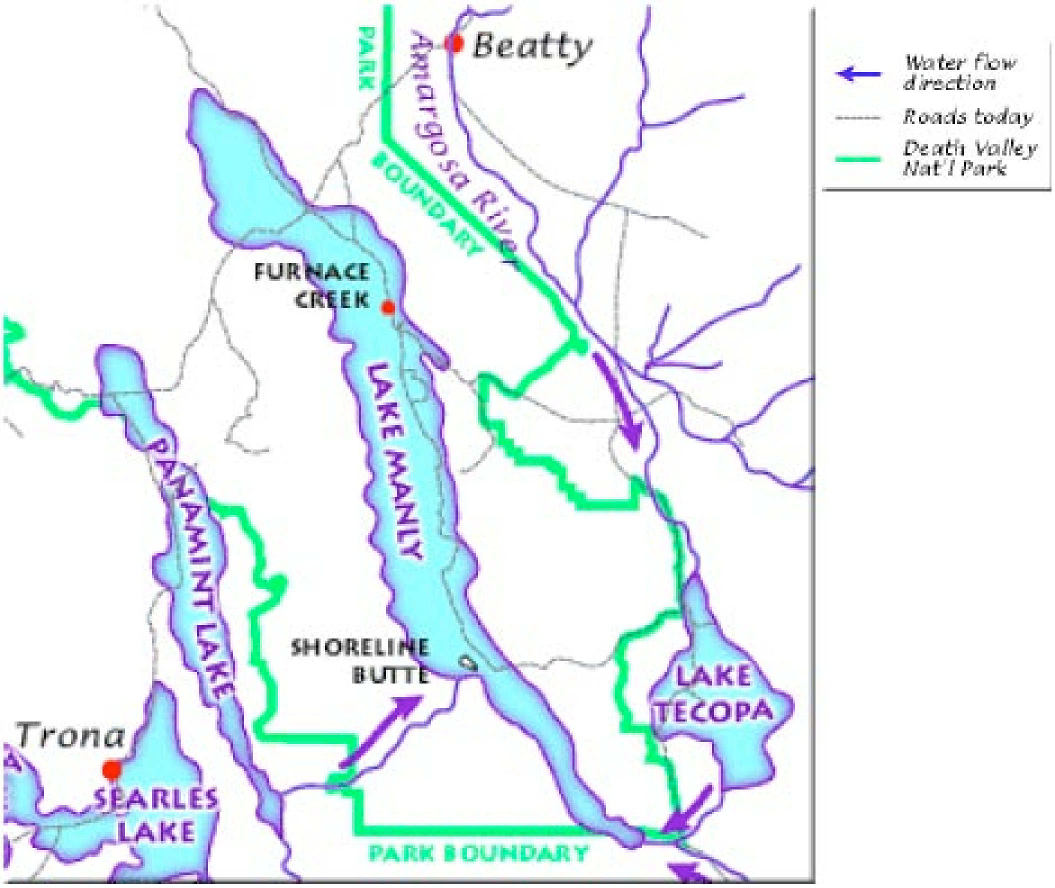


Figure 95 - Lake Manly and other Pleistocene lakes of the Death Valley region. Note the location of Shoreline Butte at the south end of the Valley (From http://geology.wr.usgs.gov/docs/usgsnps/deva/ftfur1.html)

Waves left their mark
Look carefully and you can see several horizontal lines carved into the northeast flank of Shoreline Butte. These lines are actually flat terraces called strandlines that are cut into the hillside by waves battering the shore. It takes some time for waves to gnaw away terraces like these, so these benches provide records of times when the lake level stabilized long enough for waves to leave their mark on the rock. The highest strandline is one of the principle clues that geologists use to estimate the depth of the lake that once filled Death Valley. Shorelines of ancient Lake Manly are preserved in several parts of Death Valley, but nowhere is the record as clear as at Shoreline Butte (Figure 96).

Not so long ago
Several lakes have occupied Death Valley since the close of the Pleistocene Epoch 10,000 years ago, but these younger lakes were quite shallow compared to Lake Manly.


Figure 96 - Shorelines etched into northeast flank of Shoreline Butte (From http://geology.wr.usgs.gov/docs/usgsnps/deva/ftfur1.html; Photo by Martin Miller)

Return to Lone Pine Campground via Hwy 395

Day 15 - Friday, 1 September 2006

Summary: Leave Lone Pine
San Andreas Fault at Palmdale
Palm Springs Gondola and San Gorgonio Pass
Travel to Joshua Tree National Park Campground

Directions from Lone Pine to Fault at Palmdale

Take US395/S Main St and continue to follow US395 S for 65 miles
Stay straight to go on CA-14 S for 48.5 miles
Follow CA-14 to Palmdale and take County Hwy N2 (Avenue S). The locality is along I-14 (Antelope Valley Freeway) freeway just north of the Avenue S exit. As it is illegal to stop along a freeway we will view the locality from outside the fence on Avenue S.

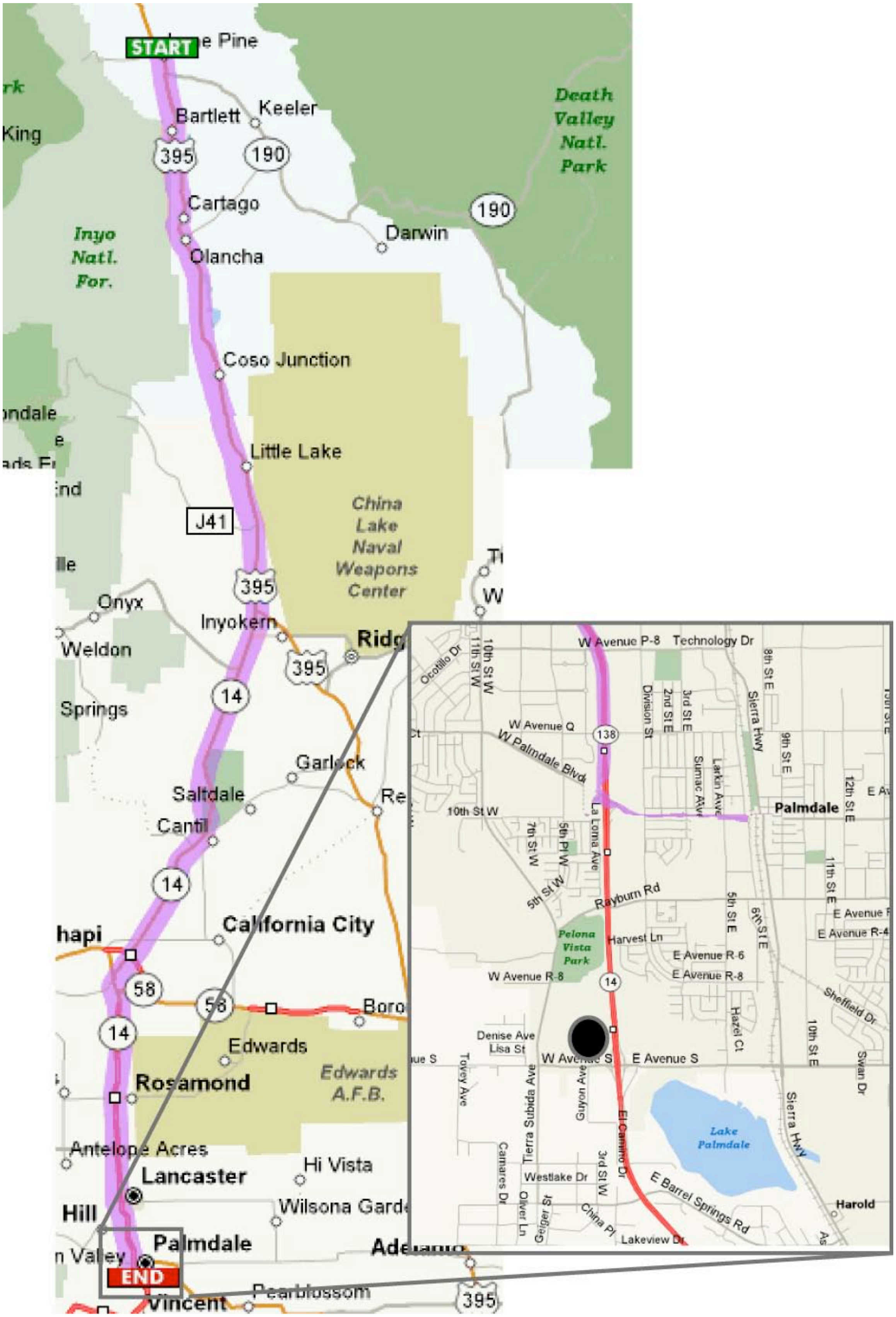


Figure 97 Lone Pine to Fault Location

15 - 1 San Andreas Fault at Palmdale

From Twentynine Palms take CA-62 West to I-10. Take I-10 West towards Los Angeles and merge onto I-215 North towards Riverside/San Bernardino/Barstow. Continue north to CA-138, follow northwest to CA-18, then proceed west to CA-14.

Significance of the Site

The great 1857 Fort Tejon earthquake that occurred along the south-central stretch of the San Andreas fault resulted in surface faulting across northern Los Angeles County where the fault passes within 35 mi (56 km) of downtown Los Angeles. Intensive geologic studies of the fault in recent years have led to the conclusion that a similar great earthquake, potentially much more catastrophic due to vast growth in the region, is likely to recur before the end of the twentieth century (Davis and others, 1982). Widespread publicity of this anticipated earthquake has, naturally, stimulated concern about its effects and curiosity about the nature of the San Andreas fault. The roadcut exposure of contorted rocks within the fault zone provides visual proof of the existence of the fault and dramatically displays the effects of powerful forces that have acted along it for a long time. The following description of this locality is taken from Barrow (1987).

Site Information The San Andreas fault, in the vicinity of Palmdale (Figures 98, 99), consists of a modern "main trace" along which the most recent surface rupture has taken place. The fault is expressed as an alignment of youthful geomorphic features such as scarps, troughs, and linear ridges that comprise a belt typically 50 to 100 ft (15 to 30 m) wide. The modern fault lies at the center of the San Andreas fault zone, which includes all the subparallel fault strands that are spatially and tectonically associated with the main trace proper. The San Andreas fault zone is 1 mi (1.6 km) wide where the Antelope Valley Freeway crosses it west of Lake Palmdale (a reservoir developed within a closed depression along the fault). Approximately 2 mi (3.2 km) to the east, the fault zone is 2 mi (3.2 km) wide. Excavation of the roadcut across the uplifted area north of the San Andreas fault exposed complexly folded and faulted strata of the Anaverde Formation for 2,400 ft (730 m) (Figure 100) between the main trace and another major strand called the Little Rock fault (Smith, 1976). With minor exceptions, the entire exposure consists of interbedded buff, arkosic sandstone, and dark brown, gypsiferous, clay shale of the lower-middle Pliocene, nonmarine, Anaverde Formation. A map of the distribution and generalized structure of the Anaverde Formation near Palmdale was first published by Wallace (1949), who also presented a geologic section that, fortuitously, coincides with the freeway alignment. Dibblee's map of the Lancaster Quadrangle (1960) also covers the area surrounding the roadcut.

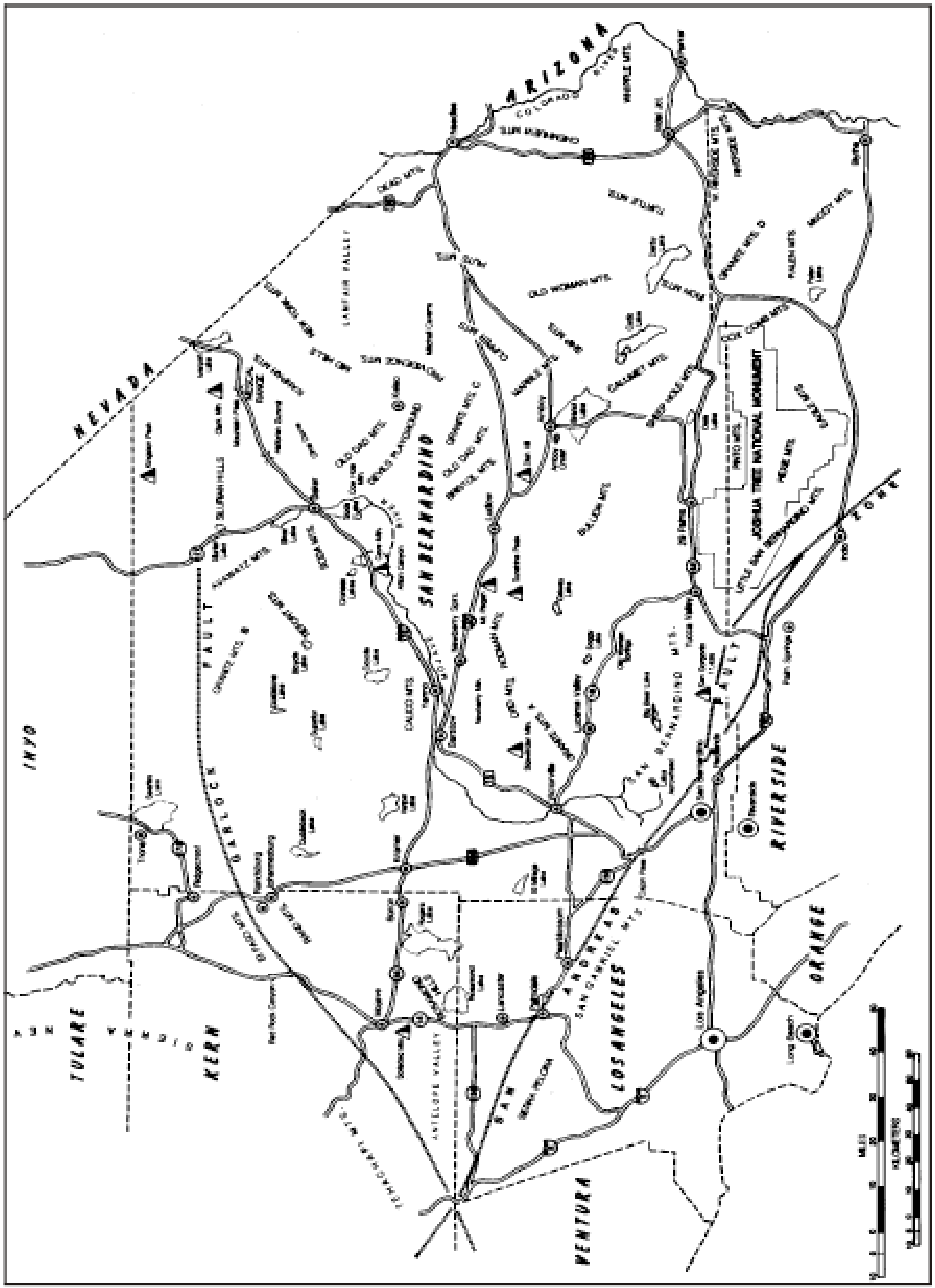


Figure 98 - Mojave Desert Province (From Norris and Webb, 1989, p. 221, Fig. 7-1)

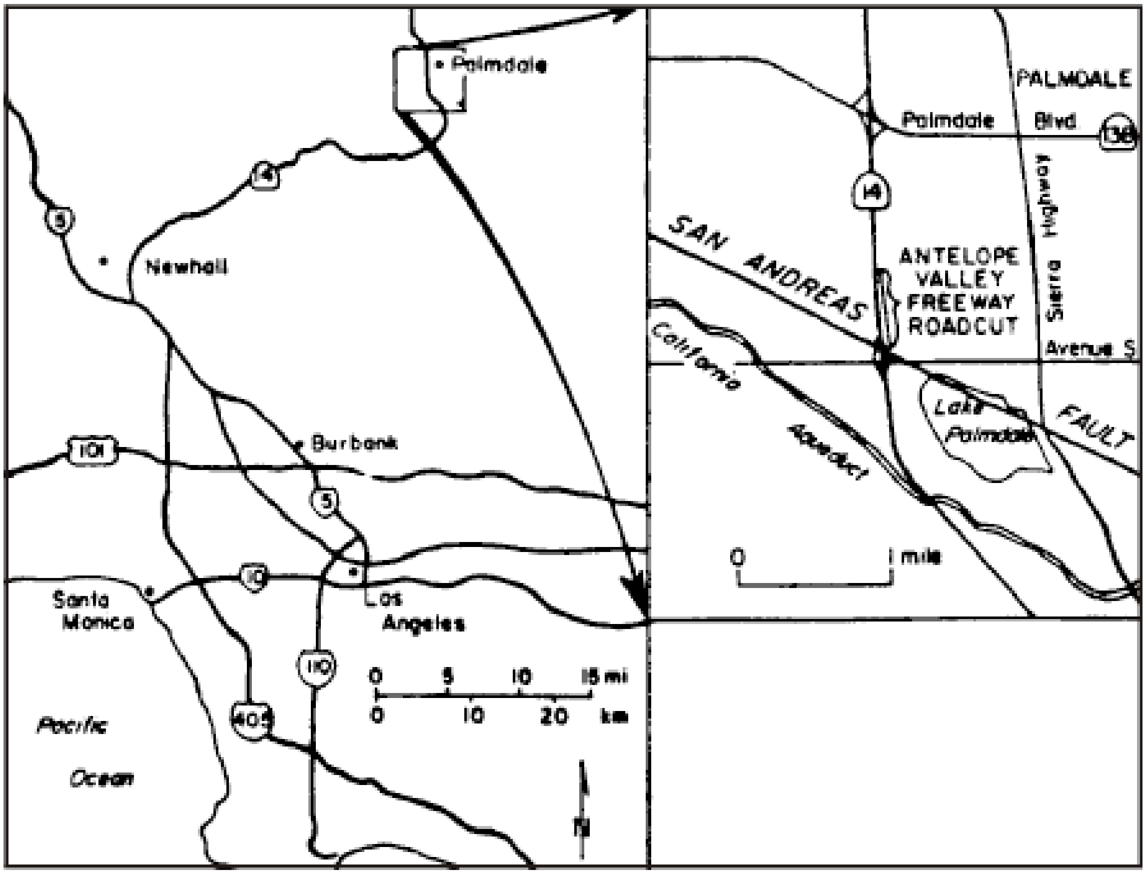


Figure 99 - Location Map, San Andreas fault zone near Palmdale, CA (From Barrows, 1987, p. 211, Fig. 1)

As many as eight members of the Anaverde Formation have been mapped in the 30 mi (48 km) stretch between Elizabeth Lake and Juniper Hills along the north side of the San Andreas fault (Barrows and others, 1976, 1985), but only the clay shale (Tac) and buff arkose (Tab) members are present in the roadcut. At its northern boundary the Anaverde Formation is juxtaposed against granitic rocks along the Little Rock fault, which is not exposed in the roadcut but is visible along the base of the northern slope of the hill. The Little Rock fault, a possible ancestral trace of the San Andreas fault, parallels the San Andreas for more than 37 mi (60 km) and has been the site of substantial lateral slip, in excess of 13 mi (20 km), since the deposition of the Anaverde Formation. A distinct bump in the freeway pavement, presumably a result of swelling of the clayey gouge beneath the roadbed, reveals to motorists the location of the San Andreas fault 400 ft (120 m) north of Avenue S. At the southern end of the roadcut, the San Andreas fault manifests itself as a 50 ft (15-m) wide zone of dark gray gouge. Buff, arkosic sandstone is folded into a possible, although minor, anticline. Internal faulting disrupts the syrnmetry of the fold, which extends along the cut for about 600 ft (l80 m) north of the San Andreas fault. Proceeding northward, a prominent fault separates the sandstone from an assemblage of intricately folded shale and sandstone that extends for 250 ft (75 m) to another prominent fault. The faults bounding this section have several inches of gouge and may be sites of important lateral movement. The remainder of the cut, beyond the fault that is 850 ft (260 m) north of the San Andreas, consists of alternating, internally faulted shale and sandstone sequences that have been deformed into an asymmetrical synclinal fold whose axis lies within the tightly folded and faulted area near the deepest part of the cut. Plastic deformation is graphically demonstrated within this area by the complex folding of dark clay shale and interbedded resistant, thin, light-colored layers of nearly pure gypsum. Numerous pits on the surface of the hillside nearby attest to the mining of the abundant gypsum during the early part of this century.

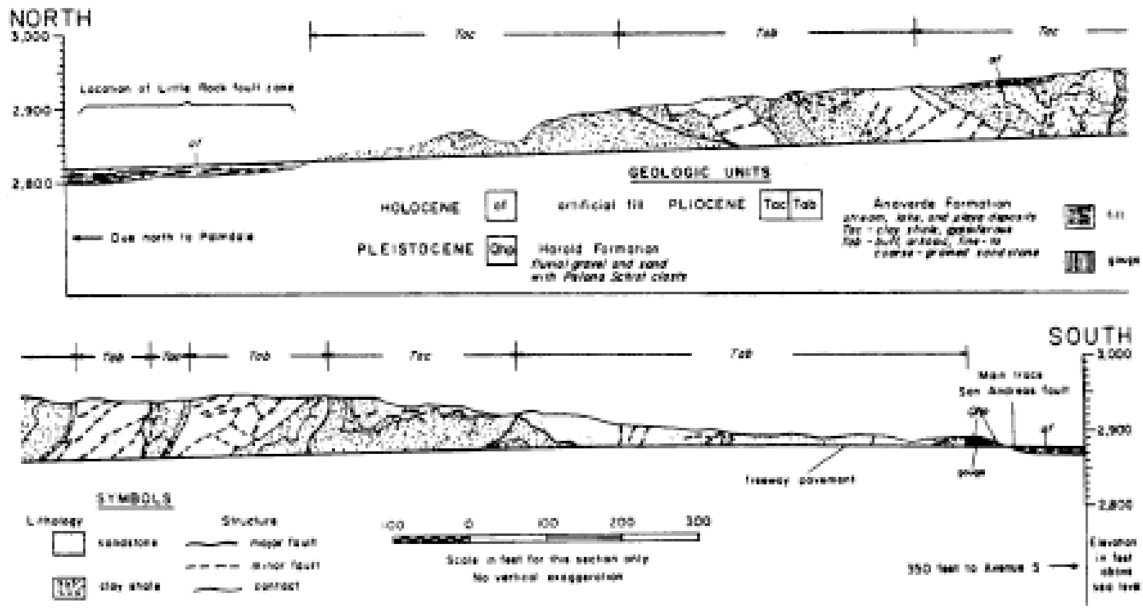


Figure 100 - Geologic sketch of part of the San Andreas fault zone exposed in the east wall of a roadcut along the Antelope Valley Freeway (California 14) north of Avenue S near Palmdale, California (From Barrows, 1987, p. 211, Fig. 1)

15-2 Palm Springs Gondola and San Gorgonio Pass

Directions from Palmdale to Palm Springs Gondola

East on Palmdale Blvd/CA-138, Stay on CA-138 for 45 miles
Merge onto I-15 S toward San Bernardino 7.5 miles
Keep left and take I-215 S toward San Bernadino/Riverside 13.5 miles
Merge onto I-10 E toward Redlands/Indio 40.5 miles
Merge onto CA-111 S toward Palm Springs 3.5 miles
Turn a slight right onto Tramway Road

Figure 101: Map of Palmdale to Palm Springs

Palm Springs

A tourist destination since the late 19th century, Palm Springs had already caught Hollywood's eye by the time of the Great Depression. It was an ideal hideaway: celebrities could slip into town, play a few sets of tennis, lounge around the pool, attend a party or two, and, unless things got out of hand, remain safely beyond the reach of gossip columnists. But it took a pair of tennis-playing celebrities to put Palm Springs on the map. In the 1930s actors Charlie Farrell and Ralph Bellamy bought 200 acres of land for $30 an acre and opened the Palm Springs Racquet Club, which soon listed Ginger Rogers, Humphrey Bogart, and Clark Gable among its members.

During its slow, steady growth period from the 1930s to 1970s, the Palm Springs area drew some of the world's most famous architects to design homes for the rich and famous. The collected works, inspired by the mountains and desert sands and notable for the use of glass and indoor/outdoor space, became known as Palm Springs Modernism. The city lost some of its luster in the 1970s as the wealthy moved to newer down-valley communities.

Palm Springs Tramway/Gondola:

A trip on the Palm Springs Aerial Tramway provides a 360-degree view of the desert through the picture windows of Rotair rotating tram cars. The 2½-mi ascent through Chino Canyon, the steepest vertical cable ride in the United States, brings you to an elevation of 8,516 feet in less than 20 minutes. On clear days, which are common, the view stretches 75 mi -- from the peak of Mt. San Gorgonio in the north, to the Salton Sea in the southeast. At the top, a bit below the summit of Mt. San Jacinto, are several diversions. Mountain Station has an observation deck, a restaurant, a cocktail lounge, apparel and gift shops, picnic facilities, and a theater that screens a worthwhile 22-minute film on the history of the tramway.
(Fodors.com
http://www.fodors.com/miniguides/mgresults.cfm?destination=palm_springs@116&cur_section=sig&property_id=85884)

The Palm Springs Aerial Tramway constructed in rugged Chino Canyon on the north edge of Palm Springs. For years, it was the dream of a young electrical engineer named Francis F. Crocker to "go up there where it's nice and cool".
In the late 1930s, Crocker enlisted the aid of desert pioneer and co-manager of the famed Palm Springs Desert Inn, O. Earl Coffman, in his tramway project. Coffman eventually was named chairman of the committee to plan the attraction.

Even though the enthusiasm for the tramway idea was high locally, political roadblocks caused numerous disappointing setbacks. Twice, a tramway enabling bill passed the California State legislature, only to be vetoed by then Governor Culbert Olson. With the outbreak of World War II, the plans were shelved.

However, Crocker's vision of a tramway to scale those cliffs to the coolness of the San Jacinto mountains never died. Years after the original plans were shelved, they were dusted off and the battle enjoined.

In 1945, a new tram bill was passed and Governor Earl Warren signed the measure creating the Mount San Jacinto Winter Park Authority. Coffman, who had labored long and hard to see the vision realized, was named the Authority's first chairman.

By 1950, technicians were moving ahead on designs for the Tramway, spending more than $250,000 solving riddles of road and tower construction. Funds for the construction of the Tramway were raised by the sale of $8.5 million in private revenue bonds. Not one cent of public funds was used for either the construction or operation of the Tramway. The 35-year bonds paid 5½ percent interest and were paid off in 1996. The Korean conflict was to cause yet another delay, but the ambitious project began to take form in July 1961.

Construction of the Tramway was an engineering challenge and was soon labeled the "eighth wonder of the world." The superlative was earned because of the ingenious use of helicopters in erecting four of the five supporting towers. Some 20 years later, the Tramway was designated an historical civil engineering landmark.

The first tower is the only one that can be reached by road. The helicopters flew some 23,000 missions without mishap during the 26 months of construction, hauling men and materials needed to erect the four other towers and the 35,000 sq. ft. mountain station.

Francis Crocker's dream was completed in 1963; the inaugural ride occurred on September 14th with scores of local and state dignitaries and celebrities on hand.

More than 12 million people have been safely transported by the Tramway into the majestic mountains overlooking the Coachella Valley since the attraction opened in September 1963. (http://www.pstramway.com)

San Gorgonio Pass

The San Gorgonio Pass (elevation 2600 feet) cuts between the San Bernardino Mountains on the north and the San Jacinto Mountains to the south. Like the Cajon Pass to the northwest, it was also created by the San Andreas Fault. The San Gorgonio Pass provides the main link between the Los Angeles Basin and the Coachella Valley and points further east. The pass itself is not as steep as the Cajon Pass or the Grapevine, but it is still one of the deepest mountain passes in the 48 contiguous states, with the mountains to either side rising almost 9,000 ft (2,750m) above the pass. Mount San Gorgonio is at the pass' northern end, and Mount San Jacinto is at the pass' southern end. Mount San Jacinto also has the fifth-largest rock wall in North America and the peak is only six miles south of Interstate 10

Interstate 10 and California State Route 60 (former U.S. Highway 60) intersect at the western end of the pass in Beaumont and it climbs through Banning before reaching Cabazon, where it begins its descent towards the Coachella Valley. The eastern end of the pass is at the junction of Interstate 10 and California State Route 111 near Whitewater Canyon. The Southern Pacific Railroad (now the Union Pacific) laid down tracks through the pass in 1875, and in 1952, an expressway was built through the pass, carrying U.S. Highway 99, and U.S. Highway 60. There are still portions of the old US-99 route between Whitewater Canyon and Cabazon, and Ramsey Street in Banning and 6th Street in Beaumont are actually old US-99.

The most famous sight on San Gorgonio Pass is the wind farm on its eastern slope, as it marks the gateway into the Coachella Valley. The pass is one of the windiest places in Southern California, and it is one of three major wind farms in California, with the others being at Altamont Pass and Tehachapi Pass. (http://en.wikipedia.org/wiki/San_Gorgonio_Pass)

Figure 102: A small segment of the San Gorgonio Pass wind farm. This digital photograph, facing due South, was taken near I-10 in White Water, California on 2006-04-22. Snow-covered Mt. San Jacinto is seen in the background. The wind currents in the Pass are especially strong, which are desirable for the wind turbines, but can make photographing and driving through this area a bit tricky. (http://en.wikipedia.org/wiki/San_Gorgonio_Pass)

Directions from Palm Springs to Joshua Tree Campground

Start out going east on Tramway Rd toward North Palm Canyon Drive/CA-111 for 0.1 miles
Tramway Rd becomes W San Rafael Rd for 0.7 miles
Turn left onto N Indian Canyon Dr for 1 mile
N Indian Canyon Dr becomes Indian Canyon Dr for 10 miles
Turn right onto CA-62 E/Twenty nine Palms Hwy 35.5 miles
Turn slight right onto National Monument Drive/National Park Drive
Continue for 1 mile.

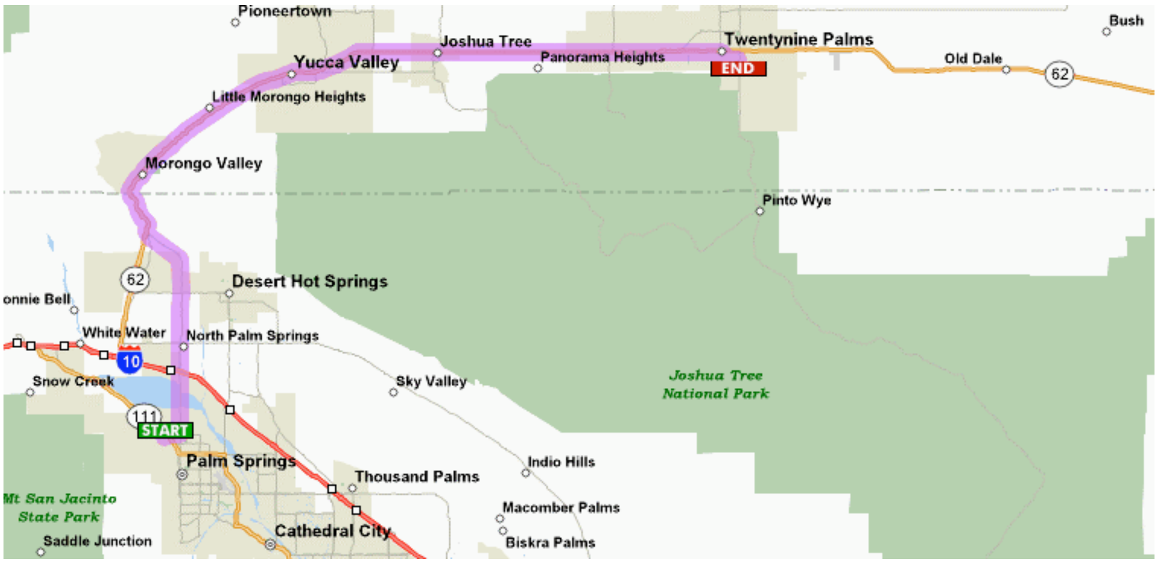


Figure 103 Palm Springs To Joshua Tree

Day 16 – Saturday, 2 September 2006

Summary: Leaving Joshua Tree Campground
Joshua Tree National Park
Salton Sea
Salton Barchans Dune Field
Return to Joshua Tree Campground

16-1 Joshua Tree National Park

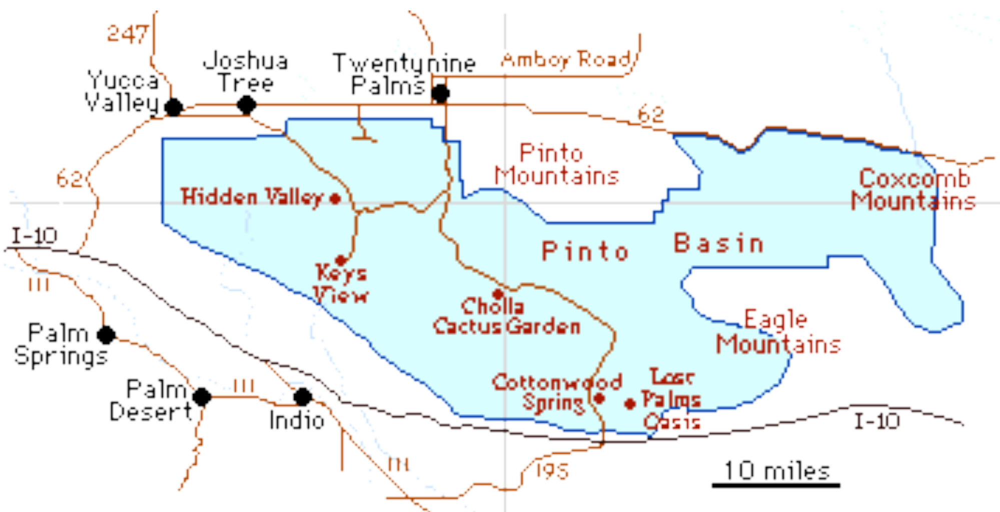


Figure 104- Joshua Tree National Monument

Joshua Tree National Monument

The Desert is immense and infinitely variable, yet delicately fragile. It is a land shaped by sudden torrents of rain and climatic extremes. Rainfall is sparse and unpredictable. Streambeds are usually dry and waterholes are few. This land may appear defeated and dead, but within its parched environment are intricate living systems, each fragment performing a slightly different function, each fragment depending upon the whole system for survival. Two deserts, two large ecosystems primarily determined by elevation, come together at Joshua Tree National Monument. Few areas more vividly illustrate the contrast between high and low desert. Below 910 metres (3000 ft), the Colorado Desert, occupying the eastern half of the monument, is dominated by the abundant creosote bush. Adding interest to this arid land are small stands of spidery ocotillo and jumping cholla cactus. The higher, slightly cooler, and wetter Mojave Desert is the special habitat of the undisciplined Joshua tree, extensive stands of which occur throughout the western half of the monument (Figure 104) Standing like islands in a desolate sea, the oases, a third ecosystem, provides dramatic contrast to their arid surroundings. Five fan-palm oases dot the monument, indicating those few areas where water occurs naturally at or near the surface, meeting the special life requirements of these stately trees. Oases once serving earlier desert visitors now abound in wildlife.

The monument encompasses some of the most interesting geologic displays found in California's deserts. Rugged mountains of twisted rock and exposed granite monoliths testify to the tremendous earth forces that shaped and formed this land. Arroyos, playas, alluvial fans, bajadas, pediments, desert varnish, granites, aplite and gneiss interact to form a giant desert mosaic of immense beauty and complexity. As old as the desert may look, it is but a temporary phenomenon in the incomprehensible timescale of geology. In more verdant times, one of the Southwest's earliest inhabitants, Pinto Man, lived here, hunting and gathering along a slow moving river that ran through the now dry Pinto Basin. Later, Indians travelled through this area in tune with harvests of pinyon nuts, mesquite beans, acorns and cactus fruit, leaving behind rock paintings and pottery ollas as reminders of their passing. In the late 1800s, explorers, cattlemen and miners came to the desert. They built dams to create water tanks and dug up and tunnelled the earth in search of gold. They are gone now, leaving behind their remnants, the Lost Horse and Desert Queen Mines and the Desert Queen Ranch. In the 1930s, homesteaders came seeking free land and the chance to start new lives. Today, many people come to the monument’s more than 200,000 hectares (500,000 acres) of open space seeking clear skies and clean air, peace and tranquillity, and the quietude and beauty that only deserts can offer.

Directions from Joshua Tree to Salton Sea

Start out going NW on Quail Springs Rd toward Rock Haven Rd 4 miles
Quail Springs Rd turns into Park Blvd 1 mile
Turn left onto Twenty-nine Hwy/CA-62 West 27 miles
Merge onto I-10 East toward Indio 28 miles
Keep right to take CA-86 past Thermal Airport on your right and cross CA-111 junction
Continue south on CA-86 along the west side of Salton Sea
From here, follow the leader ☺

Figure 105: Map from Joshua Tree to Salton Sea

16-2 Salton Trough Overlook and Salton Sea Overlook

Ignore references to road locations (previous field trip came from a different direction), but pay attention to the geological information mentioned here:

Roadcuts between the overlook and the floor of Elsinore Valley expose Bedford Canyon schist (Early Mesozoic) are intruded by spheroidally weathered gabbro and tonalite characterized by abundant rounded inclusions. Physiographic details of the Elsinore fault zone are easily observed along the southwest 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 southeast of Lake Elsinore, where fault scarps border some of these low hills on both the southwest and northeast sides. We cross this zone of faulting on Corydon Rd. Additional fault features along the Elsinore zone are in view as we drive along it to Temecula and Rancho California. The Rainbow Canyon Golf Resort is located within the fault zone in the region where alluvium and terrace deposits have been dissected and removed by the capture of inland drainage by the Santa Margarita River. This river, which flows into the Pacific Ocean along a direct course to the southwest, has formed the deep Temecula Gorge, and is now capturing drainage northeast of the Santa Ana Mountains and the Elsinore fault. The entry of this gorge lies west and adjacent to the freeway (I-15) at the Hwy 79 interchange. Between the Elsinore fault and the Salton Trough overlook, we drive through the northern Peninsular Ranges, underlain by granitic and gneissic rocks of various types (Sylvester and Bonkowski, 1979). The region is also transected by several northwest-southeast-striking fault zones, and our route takes us along the course of the Agua Caliente fault on our way to Warner Springs (Figure 106). Vertical separations along the faults of this zone show up in the topography and have helped form inter-montane basins in which shallow alluvium has accumulated. The direction and amount of slip on these faults, however, is unknown because the basement terrane has yet not been investigated in sufficient detail to establish correlations across them. Their strike and pattern suggests, however, that they are mainly right-slip faults with variable vertical separations. Hot springs, such as those at Warner Springs, are present along the aptly named Agua Caliente fault zone. Alluvium consisting of disintegrated granitic debris (grus) makes up much of the sedimentary fill in several of the small basins crossed along our route. At places these deposits are undergoing dissection so that unconformable contacts, including buttress unconformities, are exposed in highway cuts. In fact, the whole region

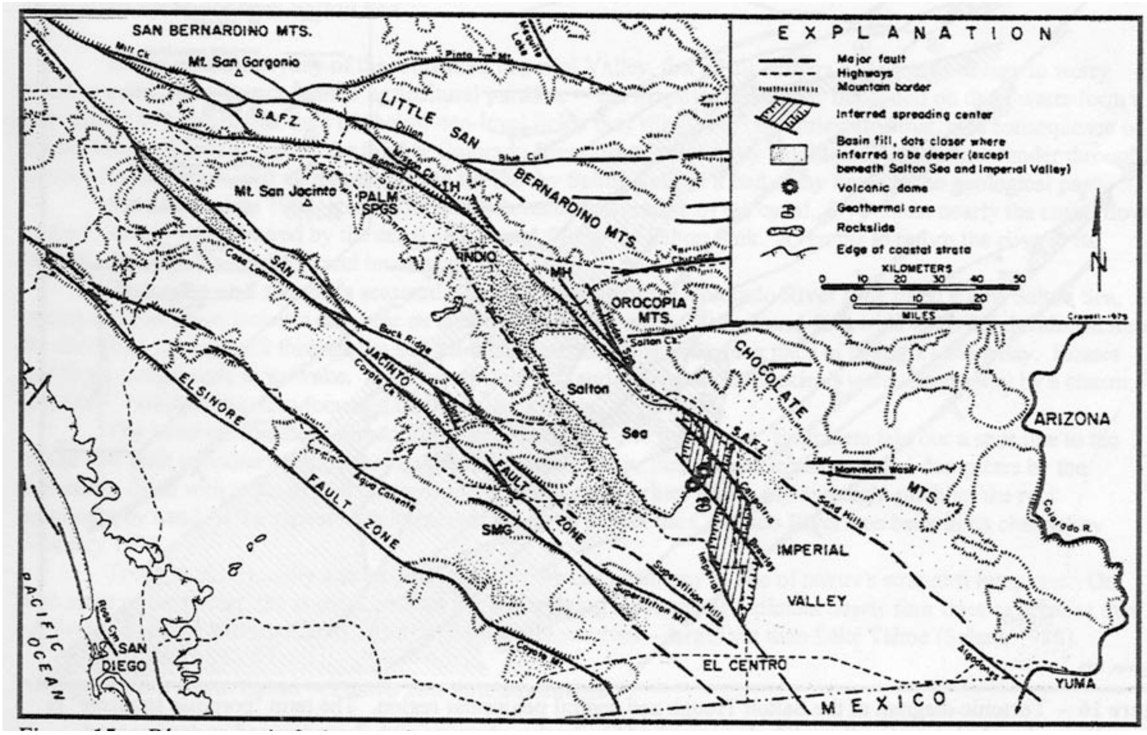


Figure 106 - Diagrammatic fault map of central Peninsular Ranges and Salton Trough region. The map shows the major faults of the San Andreas system near their juncture with the divergent plate boundary in the Salton Trough. Basin fill is indicated by stippling, and inferred active spreading centers by the line pattern southeast of the Salton Sea. Abbreviations: IH, Indio Hills; MH, Mecca Hills; SMG, Split Mountain Gorge. (from Crowell and Sylvester, 1979, p. III)

constitutes a veritable sand-grain factory. Deep weathering breaks down the basement gneisses and granitic rocks to disaggregated grains which in time make their way through a succession of continental basins and temporary resting places to the beaches and to the sea. In Miocene and Pliocene times such regions surrounded the Los Angeles basin, as well as several basins in the near offshore, and provided ideal sources for the sand within turbidity currents. These currents carried sand into deep-basin fans, channels, and sheets that now form the reservoir rocks for many of the prolific oil fields, especially those in Late Miocene and Pliocene strata of the Los Angeles basin and environs.

Salton Trough Overlook

View eastward into the Salton Trough with Borrego Valley in the foreground. Note the fault-controlled topography, the granitic and gneissic rocks of the Peninsular Ranges basement in the vicinity, and the sedimentation pattern now prevailing in the trough below. Stream courses into the Borrego Valley are much shorter and steeper than those flowing westward into the Pacific Ocean behind us. Distally the large alluvial fans grade into playa deposits. With good visibility we can see beyond the Salton Sea to the crest of the Chocolate Mountains, about 96 km (60 mi) due east. The Salton Trough or graben is complex, and its width between the marginal rims at this latitude is about 100 km (62 mi). Faults with vertical separations, including those with normal slip, are in part responsible for the difference in topographic relief in the vicinity of the overlook. For example, aligned notches in the basement terrane below us to the southeast are along a fault, but one within apparently homogeneous granite. Some of these faults are interpreted as related to collapse under gravity as the graben widened, but others may be related to the formation of the proto-Gulf of California (Crowell and Ramirez, 1979), or even older basin-range faulting (Figure 107).

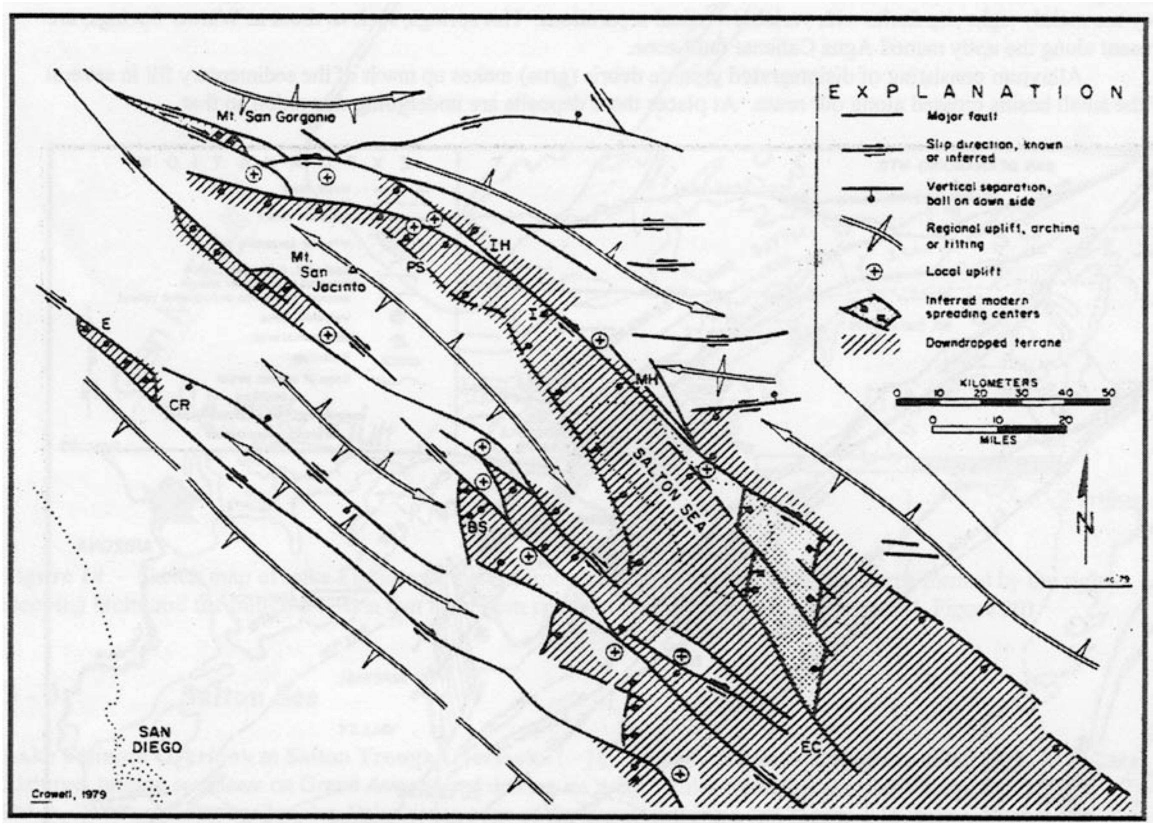


Figure 107 - Tectonic diagram of the Salton Trough and central peninsular region. The term "porpoise structure" is informally used to designate locally uplifted or uparched blocks that alternate with downdropped or downsagged blocks betweeen braided fault strands, such as along the San Jacinto fault zone northeast and east of Borrego Springs (BS). CR, E, Lake Elsinore; PS, Palm Springs; IH, Indio Hills; I Indio; MH, Mecca Hills; EC, El Centro (from Crowell and Sylvester, 1979, p. 8)

Displacements along braided strands of the San Jacinto fault zone are largely responsible for the topography within the Borrego Valley and at its northwest and southeast margins. Note the faceted spurs along the Coyote Creek fault and beyond Clark Valley along the southwest face of the Santa Rosa Mountains. Alluvium, folded into a complex broad arch, constitutes the Borrego Badlands lying southeast of the plunging end of Coyote Mountain. Borrego Mountain lies in turn along trend farther southeast. Structural details in this region were mapped and studied intensively following the Borrego Mountain Earthquake of April 9, 1968 (magnitude 6.4) (Sharp and many others, 1972). Many of these structural details fit a simple-shear scheme (Crowell and Ramirez, 1979).

Salton Sea Overlook

We are standing on a pediment surface armored with desert pavement extending southeastward from the Santa Rosa Mountains, and underlain by beds of the Pleistocene Ocotillo Conglomerate. Unconformably beneath these beds are sandstone and siltstone layers, folded and faulted, consisting of the non-marine Plio-Pleistocene Palm Spring and Borrego Formations (Crowell and Baca, 1979). Note our position with respect to the rims of the Salton Trough, the pattern of sedimentation taking place around us and the tectonics responsible for the topography.

From Hwy 86 south, drive west along Hwy S-22 toward dunes (Fig 105).

16 - 3 Salton Barchans Dune Field

Formation of the Modern Salton Sea

In the youthful heyday of the reclaimed Imperial Valley, few of the settlers had time or energy to worry about one disturbing aspect of their agricultural paradise -- the irrigation canal they depended on drew water form a river nearly 400 feet higher than the below-sea-level fields they tilled. No one anticipated that, as a consequence of human folly and natural forces, the mighty Colorado River, forty miles away would jump its banks, wander through Mexico for a distance, and then turn north to fill the dry Salton Sink as it had many times in the geological past. Starting in early 1905, a series of floods breached the intake of the canal, By August nearly the entire flow of the Colorado was captured by the canal and began filling the Salton Sink. Attempts to return the river to its original channel with sandbags and brush mats were futile. During the next summer's seasonal floods, the whole of the Colorado River flow filled a new Salton Sea, raising its level seven inches a day over an area of 400 square miles. A thousand-foot wide waterfall developed near the shoreline, cutting back through the soft alluvial deposits of the valley at a pace of nearly a mile a day. Homes and farms disappeared in its wake. A corner of Calexico and nearly half of Mexicali were sliced away by a chasm 50 feet deep. National attention focussed on the valley's plight. The Southern Pacific Railroad Company finally came to the rescue. Engineers laid out a spur line to the intake site, built temporary bridges across the flooding channels, then moved in flat cars and dump cars by the thousands, filled with loads of rock and gravel. Hundreds of workers toiled day and night pitching the rock overboard by hand. After repeated failures, success came at last: the Colorado River was back in its channel by February 1907.
Today, a permanently altered landscape bears mute testimony to one of nature's strangest rampages. On both sides of the border, the cutback chasms are still missing a volume of sediment nearly four tines as great as that excavated from the Panama Canal. And the Salton Sea survives -- larger yet than Lake Tahoe (Schad, 1988).

Ancient Salton Basin Lakes -- Fossils and Paleo-salinities

Ancient shorelines, alternating lacustrine and subaerial deposits, and fossil data indicate several floodings of the Salton Basin in the Holocene. The fossils found in the lacustrine beds include fresh and salt water types (van de Kamp, 1973). Hubbs and Miller (1948) suggested freshwater flooding of the basin by diversion of the Colorado River into the Salton Basin and marine flooding from the Gulf of California in order to explain the occurrence of fresh and salt water fauna. Flooding from the Gulf of California would now require raising the water to about 40 feet (12 meters) above sea level to pass over the Colorado River delta barrier between the Gulf and the Salton Basin, or dropping the basin by fainting. An alternate explanation of fauna distributions is outlined below. Van de Kamp (1973) suggested an alternate hypothesis and suggested a correlation between the salinity indicated by fossils, (clams, fish, gastropods, ostracods, and forams) and their position with respect to elevation. Saltwater fossils are found in the lowest parts of the basin, generally below the- 200 foot (-61 meters) contour. Fossils indicating brackish water are found below -150 foot (-45 meters) elevation and freshwater fauna are found to the exclusion of others between -100 foot (-30 meters) and +40 foot (+12 meters) elevations, the latter being the high shoreline of Recent Lake Cahuilla. Similar observations were reported by Hubbs and Miller (1948). The vertical distribution of fauna from saltwater types at the bottom of the basin to freshwater types at the ancient lake shore, and the fact that the saltwater fauna is a restricted one, prompted the calculation of possible paleo-salinities for various stages of filling in the basin. The approximate volume of the basin for several levels was computed, and, based on the assumption that the present Salton Sea still contains all the salt that was in solution in any previous lake in the basin van de Kamp (1973) suggested that the larger that any of these paleo-lakes the lower the paleo-salinity. Foraminiferal genera, such as *Cribroelphidium* and *Ammonia*, capable of surviving fluctuations in salinities are common in the modern Salton Sea and in fossil sediments (Arnal, 1961; van de Kamp; 1973). Foraminifera were probably brought from the Gulf of California by birds or animals into an ancient lake in the Salton Basin. Cretaceous foraminifera as described by Merriam and Bandy (1965) were also found in the samples. These are reworked from formations in the Colorado Plateau region.

The Salton Barchan.

A brief stop will be made to see the swarm of 47, small, rapidly moving barchan dunes described by Long and Sharp (1964). They found that movement of the dunes varied from about 100 m to about 280 m during the seven years between 1956 and 1963, and average of 25 m per year. The dunes vary considerably in size, with the smaller ones only about 10 m from horn to horn, with heights of 1.5 to 2 m; the largest are more than 100 m across and have heights of 8 to 12 m.

Return to Hwy 86 and go north back to Joshua Tree Campground.

Day 17: Sunday, 3 September 2006

Summary: Leave Joshua Tree Campground
Lake Elsinore
San Clemente Beach
Travel to Ramada Plaza Hotel at LAX in Los Angeles!

Directions from Joshua Tree Campground to Lake Elsinore

Start going NW on Quail Spring toward Rock Haven 4 miles
Quail Springs Rd turns into Park Blvd 1 mile
Turn left onto Twenty-nine Palms hwy/CA-62 west 26.8 miles
Merge onto I-10 west toward LA 23 miles
Merge onto CA-60 west via the exits on the left towards Riverside 18 miles
Merge onto I-215 south via exit 58 toward San Diego 12miles
Take the D St exit 0.2 miles
Merge onto North D St 0.5 miles
Turn right onto West 4th St/CA-74, follow CA-74 10 miles
Turn right onto Collier Ave/CA-74 0.5miles
Turn left onto Riverside Drive/CA-74 4 miles
Continue on CA-74, becomes Grand Hwy 1.1 miles
Hwy 22 is on your right, southwest of Grand Ave.

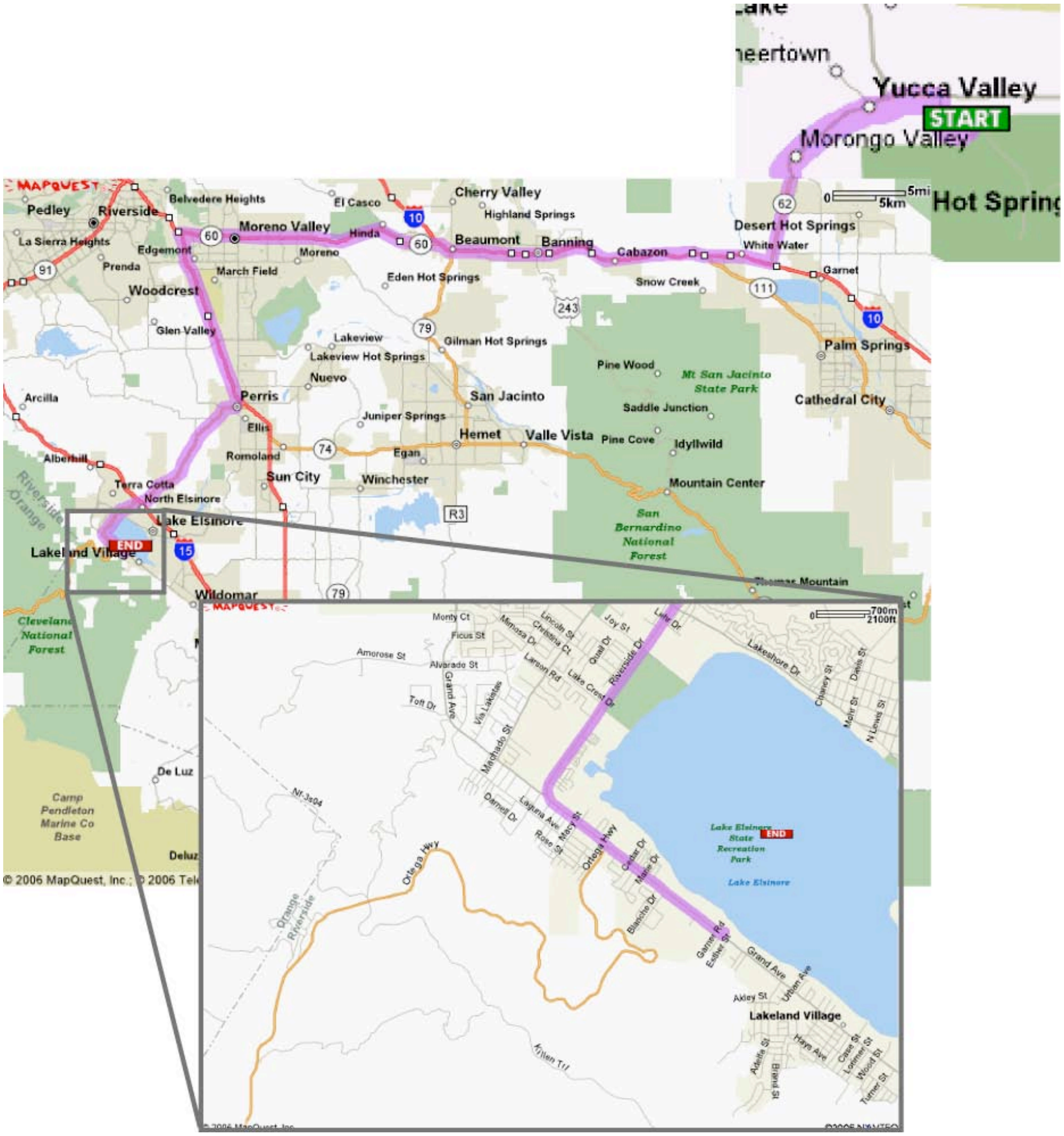


Figure 108: Joshua Tree to Lake Elsinore Area

Rock Exposures along Highway 74

Rock exposures are poor between San Juan Capistrano and the basement terrane of the Santa Ana Mountains. The stratigraphic section here dips gently toward the southwest and consists primarily of Cretaceous, Eocene and Miocene strata. Thus, our route proceeds down-section from Miocene strata to pre-Cretaceous crystalline basement rocks. The Cretaceous and Eocene beds are interpreted as part of a forearc-basin sequence now broken by younger faults and locally folded, as well as tilted. Middle and Late Miocene strata, assigned to the Monterey Formation, crop out near San Juan Capistrano and for the first 4 miles or so along Hwy 74. These beds consisting largely of siliceous shale with intercalated sandstone, were laid down off-shore in a basin formed during plate-margin fragmentation when the Pacific lithospheric plate met the North American plate. The Cristianitos fault, which we cross at the contact between Miocene and Eocene strata, has a north-north-west strike and is interpreted as having played a role in the formation of the Los Angeles basin in Miocene time. This fault is parallel to other faults in the San Joaquin Hills, W of San Juan Capistrano, that are intruded by Middle Miocene volcanic rocks.

They are therefore somewhat older than later stretching of the basin floor when the lava was intruded. East of the Cristianitos fault the Eocene Santiago Formation is the first exposed, consisting of about 975 m of marine buff sandstone with interbedded dark siltstone (Fife, 1972). The Paleocene beds along Hwy 74 are faulted against marine Cretaceous strata assigned to the Williams and Ladd Formations, units much better exposed in the northern Santa Ana Mountains. Outcrops in the hills west of Hwy. 74, visible from our vans, consist mainly of gently southwest-dipping, thick, buff sandstone with interbedded thin siltstone. The marine Cretaceous section in this region grades downward into the nonmarine Trabuco Conglomerate that in turn lies nonconformably upon the basement complex of the Santa Ana Mountains. This formation, composed of coarse red-brown conglomerate, crops out for about a mile along the highway beyond the bridge over San Juan Creek where the road follows along the NW side of the canyon. Stone types and imbrications show that the conglomerate was derived from basement terrane exposed to the east. The basement complex below the Upper Cretaceous Trabuco Conglomerate first consist of Santiago Peak volcanic rocks composed of interbedded rhyolite, andesite, diabase, and tuffaceous breccia that have been weakly metamorphosed. Along our route these exposures are succeeded by those of the Southern California Batholith consisting of gabbro, tonalite, granodiorite, quartz monzonite, and related dikes and veins. Isotopic ages based on Rb-Sr and U-Pb methods for these granitic rocks range form 135 to 95 m.y. K-Ar apparent ages range between 115 m.y. and 50 m.y., with a gradient from southwest to northeast from older to younger (Krummenacher and others, 1975). Near the summit of Hwy 74, the hillsides south of the highway are littered with light-colored exfoliation boulders of gray tonalite, distinguished from most other granitoids of the Southern California Batholith by the presence of abundant well-oriented inclusions of gabbro, metavolcanic and metasedimentary rocks.

17-1 Lake Elsinore pull apart basin overlook

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 (Abbott et al., 1989). This 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. 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 14 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 kinematic mechanisms for the Elsinore pull apart are probably very similar to those that created many of the California Continental Borderland basins during Neogene time, e.g. the basin of deposition for the Capistrano Formation. 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 regions.

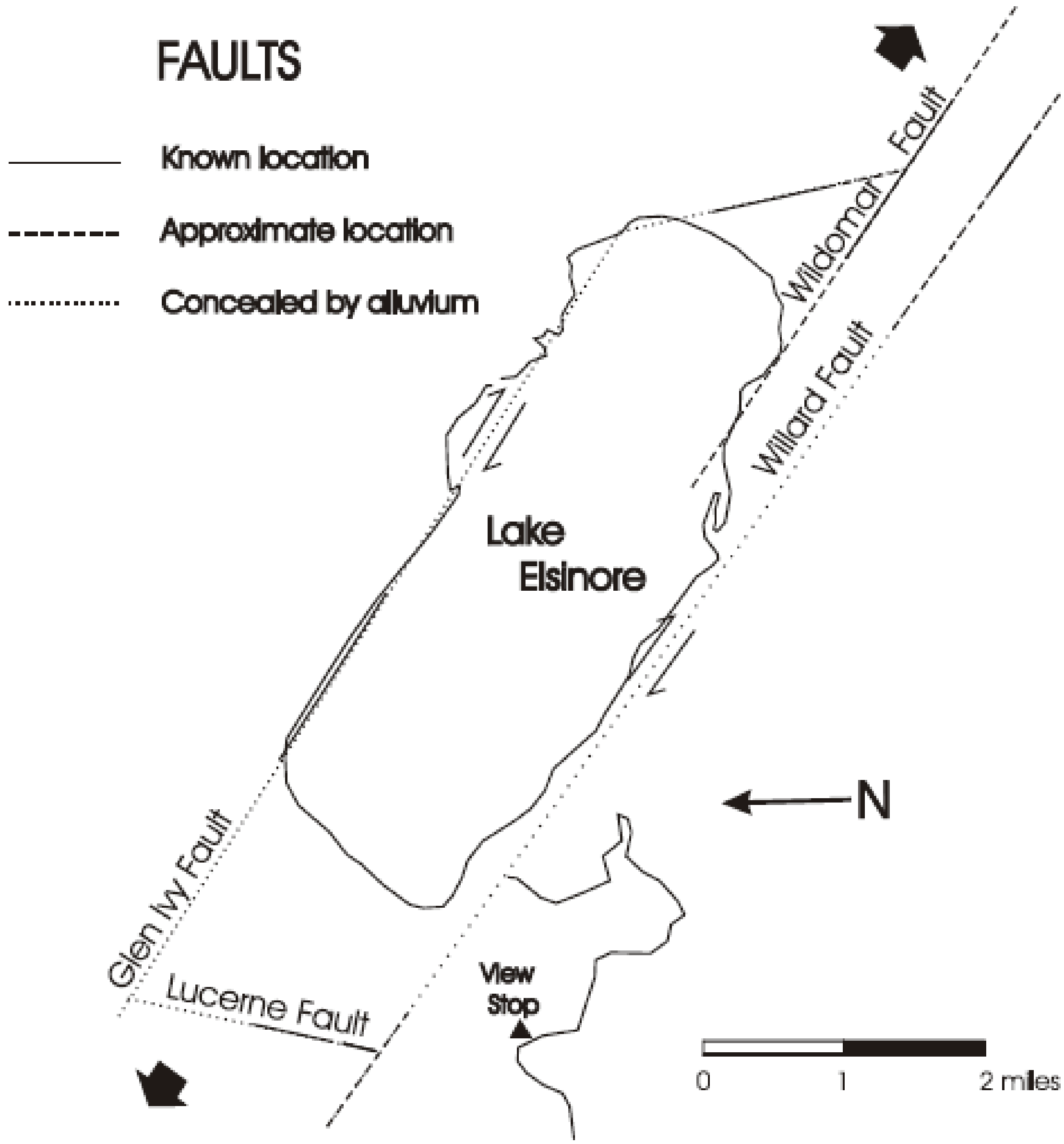


Figure 109 - Sketch map of Lake Elsinore as viewed from Hwy 74 overlook. Note the scarps formed by the rightstepping faults and the pull-apart basin that have been created (from Abbott et al., 1989, p. 34, Figure 20)

Directions from Lake Elsinore to San Clemente Beach

From Grand Avenue, going NW to Koves Rd 4 miles
Turn left onto Ortega Hwy/CA-74 28 miles
Merge onto I-5 S via the ramp on the left toward San Diego 6 miles
Take the Ave Pico exit 0.3miles
Turn right onto E Avenida Pico 0.6 miles
Turn left onto N El Camino Real <0.1 miles
Turn right onto Boca del la Playa <0.1 miles
Turn left onto Calle Las Bolas <0.1 miles
Turn a slight right onto Calle Colima 0.1 miles
Turn left onto Buena Vista 0.2 miles

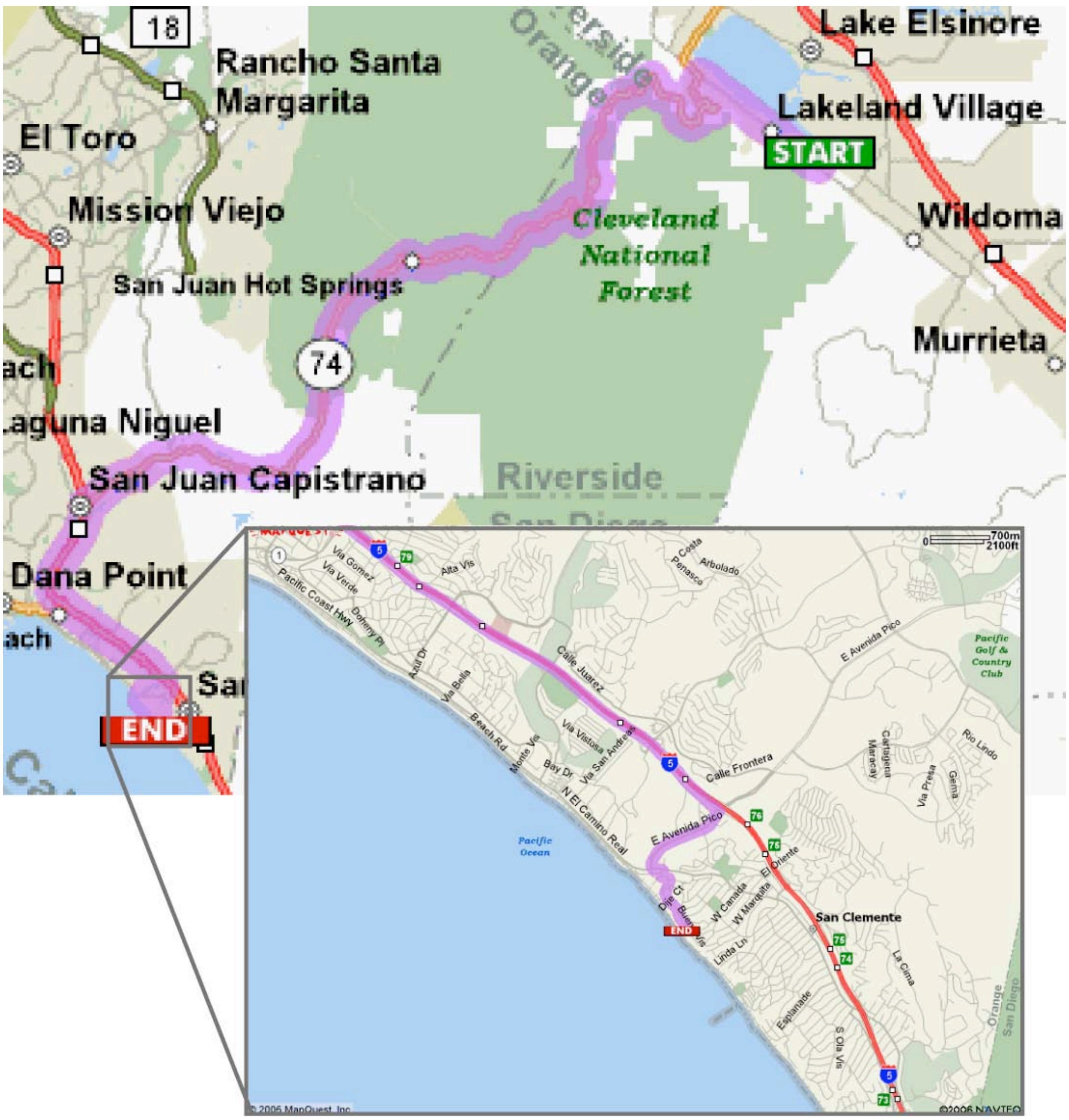


Figure 110: Lake Elsinore to San Clemente Beach

17-2 Turbidites at San Clemente Beach

Miocene - Pliocene Depositional Systems at San Clemente State Beach

We will be making a transect along the lower part of the 30 m high sea cliffs along the railroad tracks. Figure 111 illustrates the superb cross-section exposures of nested submarine channels in an upper Miocene portion of the Capistrano Formation. For 550 m to the south and 250 m to the north of the parking area, there are channels cut into mudstone strata that contain a few thin, west- to southwest-directed, turbidite sandstone beds. An almost 3- dimensional view of the facies is obtained because of the stream canyons cut perpendicular to the sea cliff. The rocks were interpreted as a braided suprafan or suprafan depositional lobe by Walker (1975). Hess (1979) regarded the rocks as pre-channel sediments on a lower suprafan that were prograded over and cut into by upper suprafan channels.

The section is best viewed beginning with the southern exposures because they contain the oldest channels. Walker (1975) described eight, nested channels numbered 1 (oldest) through 8 (Figure 111). Channel trends range from 230 to 300°. Their gently dipping channel walls are draped with mudstone in 7 of the 8 channels. Sandstone beds deposited in the channels pinch out against the walls or slope up the wall a little ways. Mudstone beds in the channels appear to merge with the mudstone drapes and become part of them higher up the walls. Some channel-fill sediments are thickening- and coarsening- upward; others are thinning- and fining upward. Several of the channel complexes have well preserved levees on their southern flanks. Northwest of the parking lot the section is composed of amalgamated, irregularly bedded, pebbly sandstone beds that exhibit some grading, gravel imbrications, rip-up clasts, and present in some of the laminated sandstone beds are cross bedding and flame structures. Hess (1979) reported that the margin of the coarse channel-filling sandstones with the pre-channeling mudstones is exposed 0.5 km inland as stepped margins similar to those observed in the Doheny channel. The best channel geometries are seen a little farther up the beach where sandstone masses are completely encased in mudstone.

Trace Fossils in the San Clemente Fan Section

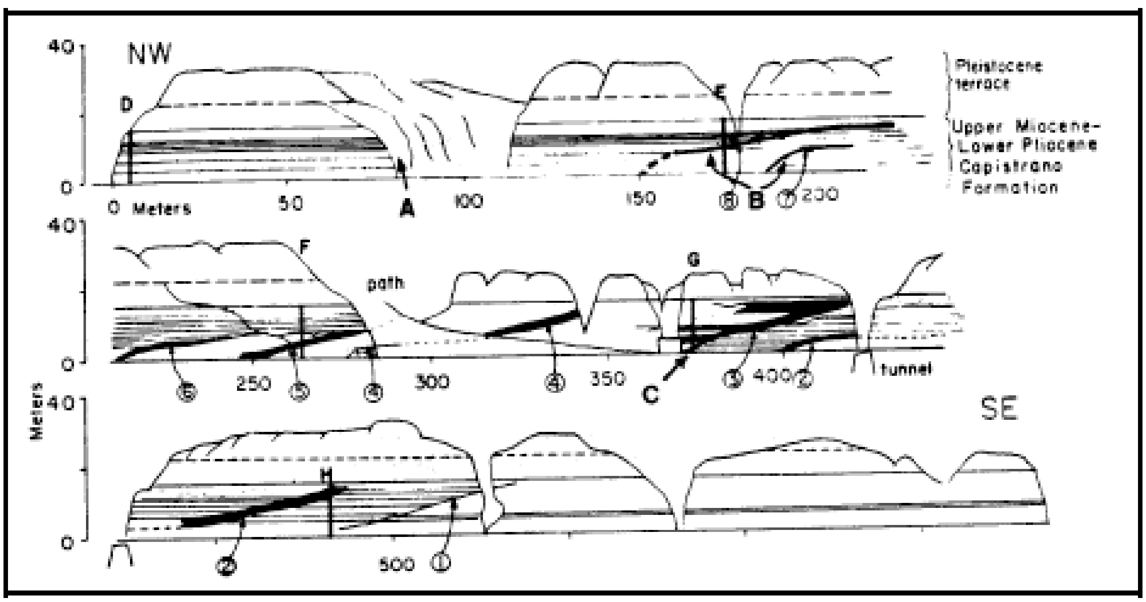


Figure 111 - Nested submarine channels exposed in cliffs along the railroad tracks at San Clemente State Beach (from Walker, 1975). Letters on section are after Barrick (1987), p. 44.

Trace fossils are not abundant or diverse in the San Clemente fan outcrop. Only three ichnogenera were found, *Ophiomorpha* Lundgren, 1891, *Thalassinoides* Erenberg, 1944, and *Skolithos* (Figure 112). *Thalassinoides* burrows are the most abundant trace fossils found. These burrows are relatively abundant in both sections A and B (Fig. 111). In section B they occur

predominantly in the clay silty clay layers. Most of the burrows are found in cross-section and they range in diameter from 0.5 to 2.5 cm. Several burrows are in positive semi-relief although a full traces are not exposed. Partial lengths are as great as 6 cm. More rare are the pellet-lined *Ophiomorpha* traces which are found in the intervening fine sandstone. All of the traces appeared to have been made by the same type of animal although clear connections between the two burrow types could observed. All of the burrows are sand filled and the *Thalassinoides* are outlined with a dark stained burrow wall. The burrows commonly appear horizontal to bedding.
Thalassinoldes is nearly as abundant in the lower part of section A. They occur in the siltstone sections between the thick debris sandstones and along the soles of the sandstone beds. Here cross-sections range between 1.0 and 4.0 cm in diameter. Higher in the section, as the siltstone beds become thinner, the *Thallaslnoides* burrows become smaller and scarcer until they disappear completely. This probably indicates that the time between deposional events decreased, not allowing larger communities or larger individuals of the tracemakers to occur. *Ophiomorpha* were not found in this section.
Vertical *Skolithos* Haldeman, 1840, tubes occur almost exclusively in section B, although several individual tubes may be observed in the basal silt layer of section A. The burrows and sections of burrows range from 2.5 to 10.5 cm in length and average 1.0 cm in width. All tubes are sand filled. Skolithos is less abundant than *Thalassinoides* and comprises less than 25% of the burrows in section B and is very rare in section A.
These types of vertical burrows have also been classed as *Tigillites* (Hayward, 1976), the distinction from *Skolithos* being that *Tigillites* occurs singly while *Skolithos* occurs in groups. Those found at San Clemente were
spaced widely enough that they need not be considered as clustered into groups however, they have been classified as *Skolithos* after Alpert (1975) who combined most *Tigillites* species into the genus *Skolithos*. No distinct trace fossils are recognized in section C, although some bioturbation appears to be present, which was probably responsible for some of the clay disruption. Turbidity currents appear to have destroyed any distinct individual traces and to have disrupted the layering.

Figure 112 - San Clemente ichnofossils (From Treatise of Invertebrate Paleontogy).

Directions from San Clemente Beach to Hotel in Los Angeles

Start out going NE on Buena Vista toward Avenida Aragon 0.2 miles
Turn right onto Calle Colima 0.1miles, Calle Colima becomes Calle Las Bolas <0.1 miles
That turns into Boca de la Playa 0.1 miles
Turn right onto West Avenida Pico 0.8 miles
Merge onto the I-5 North via the ramp on the left toward LA 8.7 miles
Merge onto CA-73 N toward Long Beach **TOLL**, 18 miles
Merge onto I-405 N toward Long Beach 13 miles
Keep left and take I-405 N via exit 23 toward Santa Monica/I-605 N for 21 miles

Figure 113: San Clemente to LA Hotel

Day 18 - Monday, 4 September 2006

Travel from LA to Ottawa!

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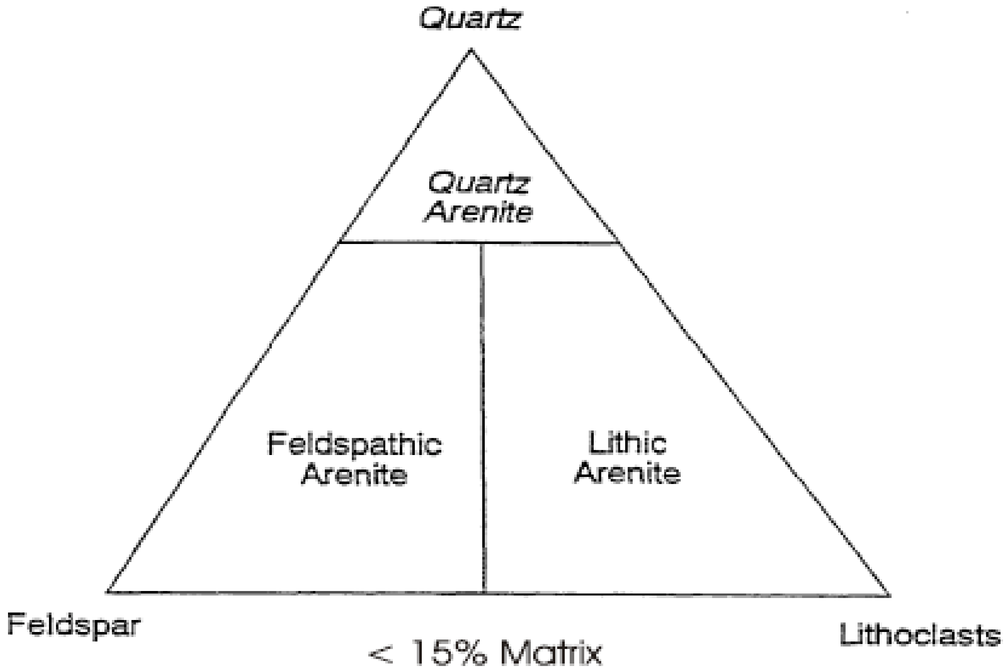
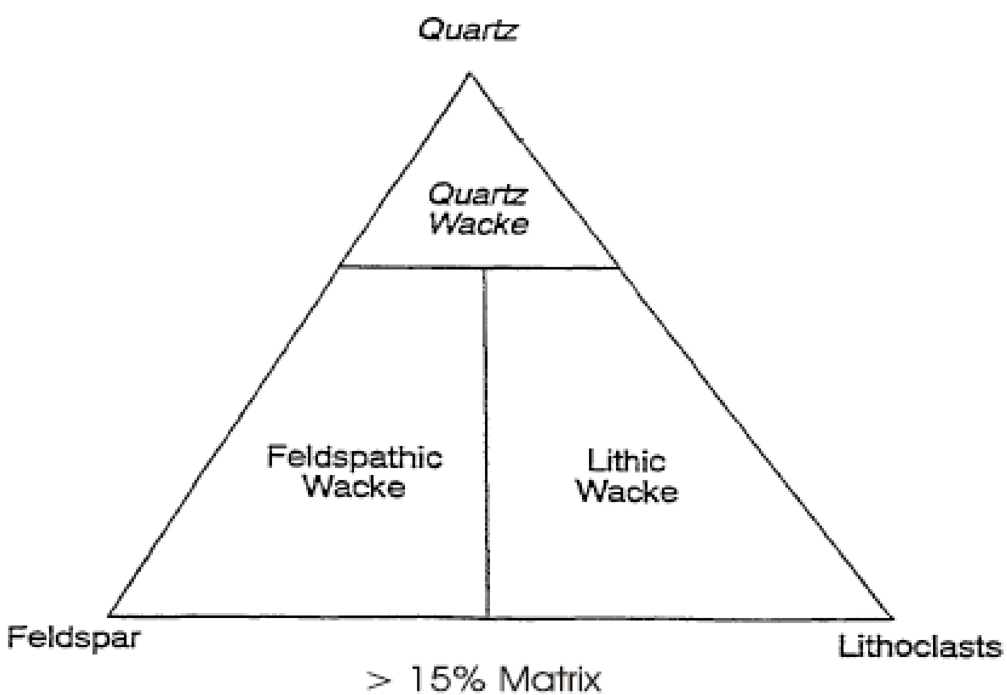
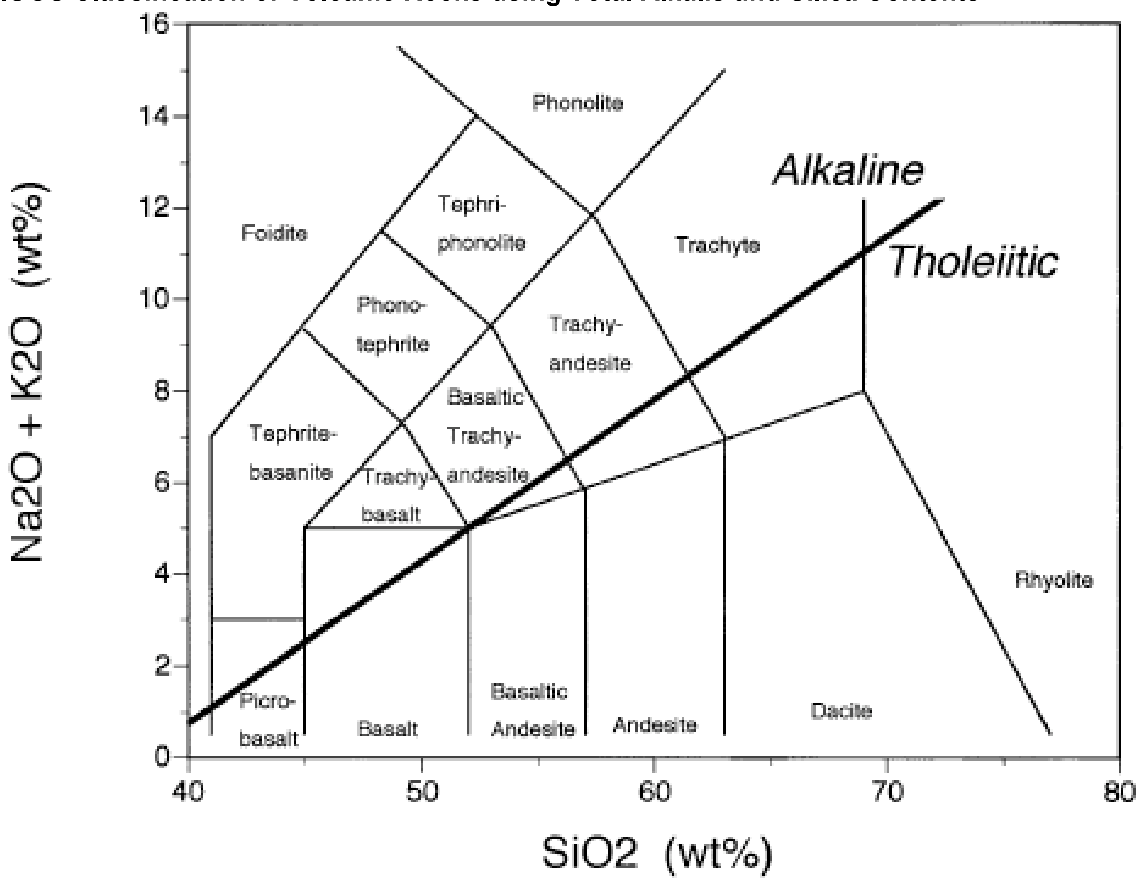
Classification of Siliciclastic sandstones





Grain size* (log scale)		Sediment	Siliciclastic Sedimentary Rocks		Carbonate** Sedimentary Rocks: Framework-supported	Carbonate** Sedimentary Rocks: Matrix-supported
	Boulders	Gravel	Conglomerate Breccia (framework- or matrix-supported for both)		Rudstone	Floatstone
256 mm	Cobbles					
64 mm	Pebbles					
2 mm	Very coarse sand	Sand	Sandstone Arenite (<15% matrix) Wacke (>15% matrix)		Grainstone (no matrix) Packstone (matrix present)	Wackestone
1 mm	Coarse sand					
500 µm	Medium sand					
250 µm	Fine sand					
125 µm	Very fine sand					
62 µm	Silt	Mud	Siltstone	Mudrock (if fissile = Shale)	Mudstone	
4 µm	Clay		Claystone			

IUGS Classification of Volcanic Rocks using Total Alkalis and Silica Contents



IUGS Classification of Volcanic Rocks Using Total Alkalis and Silica Contents

Classification of Igneous Rocks

APHANITIC	Rock Colour		Light	Medium	Dark	
	typical phenocrysts in parentheses below rock name	> 10%* QUARTZ	RHYOLITE (quartz, K-feldspar)	DACITE (plagioclase, biotite)		
		< 10%* QUARTZ	TRACHYTE (K-feldspar)	ANDESITE (plagioclase, hornblende, pyroxene)	BASALT (olivine, plagioclase, pyroxene)	KOMATIITE ultramafic extrusive
PHANERITIC	Typical Colour Index		0 - 40		40 -90	90 -100
	ESSENTIAL MINERALS		POTASSIC FELDSPAR	PLAGIOCLASE FELDSPAR	PLAGIOCLASE+ PYROXENE	OLIVINE +/or PYROXENE
	Common Varietal Minerals		Plagioclase Hornblende Biotite-Muscovite	K-Feldspar Hornblende Biotite	Olivine Hornblende	Plagioclase
		> 10% QUARTZ	GRANITE	GRANODIORITE		
		< 10% QUARTZ	SYENITE	DIORITE ANORTHOSITE (>90% Plagioclase)	GABBRO (Diabase is a fine-grained variety)	PYROXENITE (Pyroxene) PERIDOTITE (Pyroxene+Olivine) DUNITE (Olivine)

Identification of Metamorphic Rocks