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Finding Faults in the Mojave

The 1998 Desert Research Symposium Field Trip Guide and Volume

James P. Calzia and Robert E. Reynolds, editors

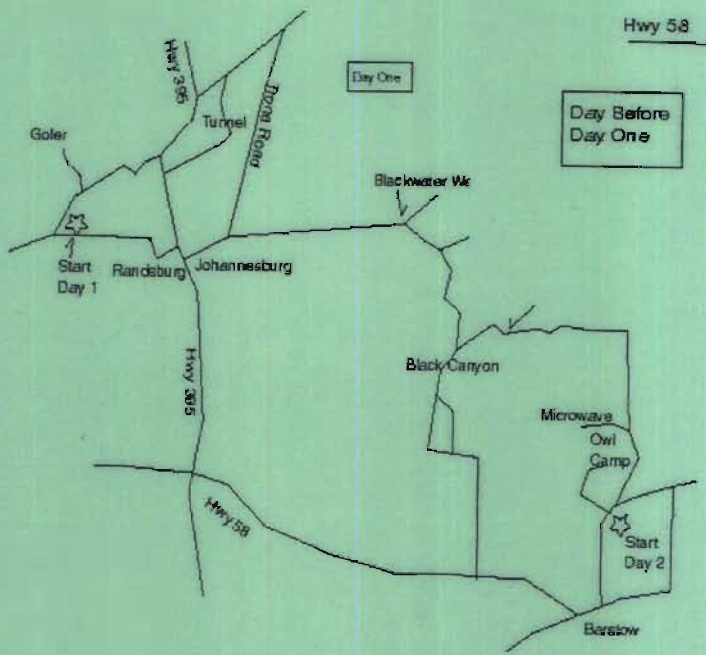
and

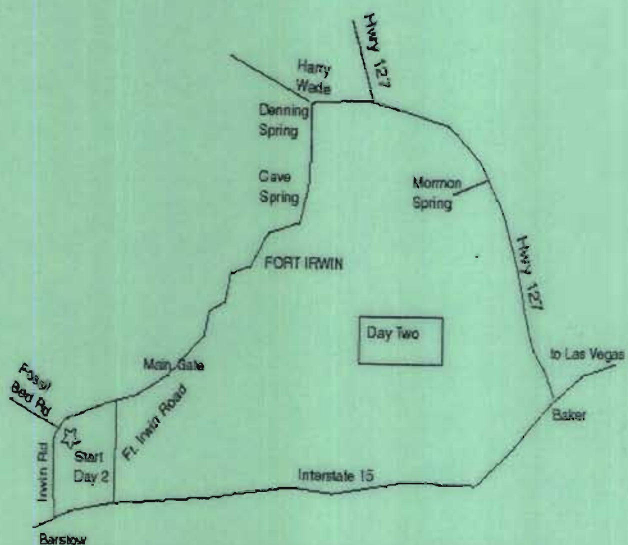
Abstracts of Proceedings: 1998 Desert Research Symposium

Jennifer Reynolds, editor

Finding Faults in the Mojave

Field trip sketch maps not to scale





Finding Faults in the Mojave
J.P. Calzia and R.E. Reynolds, editors

2024 Orange Tree Lane
Redlands, CA 92374

Front cover: Garlock fault scarp along southwest corner of the Slate Range, view to the southwest. Relative displacement along this part of fault is south-side (left) up, along base of slopes in shadow. Lava Mountains in middle distance (just beyond end of straight road leading away and to left of photo center) nearly to the skyline; Rand Mountains form most distant skyline. Tower and buildings at left, in the northwest corner of Pilot Knob Valley, are part of U.S. Naval Weapons Center. *G.I. Smith photo.*

Back cover: Garlock fault scarp along south-central side of the Slate Range, view to the northeast. Fault line is at base of scarp; to its left, Quaternary gravels overlie strongly deformed Tertiary lake deposits. South-draining alluvial fan to right of scarp is north edge of Pilot Knob Valley. Quail and Owlshead Mountains on middle skyline. *G.I. Smith photo.*

The 1998 Desert Research Symposium

Mojave Desert Quaternary Research Center, San Bernardino County Museums, Redlands, CA 9237

Friday, April 24

Registration Welcoming Remarks	8:30 a.m. 9:20 a.m.	
Session 1	9:30 a.m.	Does the Greater Roadrunner Hibernate? (James Cornett, Palm Springs Desert Museum)
	9:50 a.m.	A Systematic Review of the North American Iguanid Genus *Callisaurus* (Reptilia: Sauria): The Status of *Callisaurus draconoides* and *Callisaurus ventralis* (Benjamin Banta, Southwestern Biological Associates)
	10:10 a.m.	Geographic Variation and Environmental Determinants of Reproductive Output in the Desert Tortoise (Jeff Lovich, USGS Riverside; Phil Medica, USGS Las Vegas; Hal Avery, USGS Riverside)
Break	10:30 a.m.	
Session 2	10:50 a.m.	Fort Irwin Archaeology: A Preserved Past for the Mojave Desert (Mark W. Allen, Fort Irwin)
	11:10 a.m.	The Bodie Bowl Mining Claim Validity Examination (Gregg Wilderson, BLM Bakersfield)
	11:30 a.m.	Arid Unsaturated Zones and Groundwater Contamination Barriers for Radioactive Waste Disposal (H.G. Wilshire, Public Employees for Environmental Responsibility, Washington D.C.)
Lunch break	Noon	
Session 3	1:30 p.m.	Vulnerability and Recoverability of the Mojave Desert Ecosystem (Leonard J. Gaydos, Ames Research Center)
	1:50 p.m.	Unburned Fuels in an Arizona Upland Saguaro-Shrub Community (R.C. Wilson, CSUSB; M.G. Narog, USDA Forest Fire Laboratory; B.M. Corcoran, CSUSB
	2:10 p.m.	Flamingo Egg from the Miocene Sediments of the Calico Mountains, San Bernardino County, CA (R.E. Reynolds, Earth Sciences section, San Bernardino County Museum)
Break	2:30 p.m.	
Session 4	2:40 p.m.	Carnivore Tracks in the Miocene Horse Spring Formation, Lake Mead, Nevada (M.M. Kissell and S.M. Rowland, Dept. of Geosciences, Univ. of Nevada Las Vegas)
	3:00 p.m.	A Closer Look at *Mammuthus hayi* Barbour (1915), the Long-jawed Mammoth (G.E. McDaniel, Jr. and G.T. Jefferson, Anza-Borrego Desert State Park)
	3:20 p.m.	Announcements and Current Research
Poster Session and Symposium Social	4:00 p.m.	Publications table, San Bernardino County Museum Association; Publications Table, California Division of Mines and Geology; Desert Studies Consortium (Willliam Presch); Dogs of Robbins Quarry (Tom Howe and R.E. Reynolds); Structure Contour Map, San Bernardino Mountains (M. D. Kenney); Rock Art of Baja, California (Lowell Lindsay); Paleogeology of the Nopah Range (Judy Morris); Defining the Hemphillian-Blancan LMA Boundary in the Temecula Arkose, Riverside Co., CA (W.L. Rader); Wild Cats of the Barstow Formation (R.E. Reynolds and Mary Aruta); Unburned Fuels in an Arizona Upland Saguaro-Shrub Community (R.C. Wilson, M.G. Narog, and B.M. Corcoran)
Symposium Supper	6:00 p.m.	Reservations required.
Evening Lecture Public welcome	7:30 p.m.	Unrest in the Long Valley Caldera (Dr. David Hill, U.S. Geological Survey, Menlo Park)

Saturday, April 25

Registration Welcoming Remarks	8:30 a.m. 9:20 a.m.	
Session 1	9:30 a.m.	The Early Pliocene Caribbean Bryozoan *Acanthodesia savarti moilifera* in the Imperial Formation near Whitewater, Riverside County, California (E.C. Wilson, Rancho Mirage, and R.J. Cuffey, Dept. of Geosciences, Pennsylvania State University)
	9:50 a.m.	Desert Survival in the Nineties (Brian Brown, Shoshone Museum Association)
	10:10 a.m.	Status of the Northern and Eastern Colorado Desert Coordinated Management Plan (R.E. Crowe, BLM California Desert District, Riverside)
Break	10:30 a.m.	
Session 2	10:50 a.m.	Petroglyphs of Renegade Canyon (R.F. Hilburn, Mojave River Valley Museum, Barstow)
	11:10 a.m.	The Culture of the People of Black Canyon (B.M. Bjorkman, Twin Peaks)
	11:30 a.m.	Petrology of Early Proterozoic Granitoids from the Southwestern United States: Implication for Genesis and Tectonics of the Mojave Crustal Province (E.E. Bender, Dept. of Geology, Orange Coast College, Costa Mesa)
Lunch	Noon	
Session 3	1:20 p.m.	Analysis of Rock Varnish from the Whipple Mountains, Southeastern California, Using Time-of-Flight Secondary Ion Mass Spectrometry (R.C. Anderson, JPL; P.A. Zimmerman, Dept. of Chemistry, Univ. of Pittsburgh; K. Beratan, Dept. of Geology, Univ. of Pittsburgh)
	1:40 p.m.	Tectonic Control on Development of Old Alluvial Surfaces in the Whipple Mountains, Southeastern California (Kathi Beratan, Dept. of Geology, Univ. of Pittsburgh)
	2:00 p.m.	Microgeographic Differences in Rainfall Affect the Nutrition and Survivorship of Desert Tortoises (*Gopherus agassizii*) in the Mojave Desert (Harold W. Avery, USGS Riverside; P.A. Medica, USGS Las Vegas; Jeffrey Lovich, USGS Riverside)
	2:20 p.m.	Saltdale, Kern County, California (Larry Vredenburgh, Alan Hensher, and Gregg Wilkerson, BLM Bakersfield)
	2:40 p.m.	Current research and announcements
Adjourn	3:00 p.m.	Participants in field trip should form car pools and prepare to depart.

The San Bernardino County Museum
2024 Orange Tree Lane (California Street exit from Interstate 10)
Redlands, California 92373
1-888-BIRD EGG • (909) 307-2669

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James P. Calzia
and
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Compiler

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Finding Faults in the Mojave: Field Trip Guide

Robert E. Reynolds, *Earth Sciences Section, San Bernardino County Museum, 2024 Orange Tree Lane, Redlands CA 92374*
James P. Calzia and **Brett Cox**, *U.S. Geological Survey, 345 Middlefield Road, Menlo Park, CA*

Foreword

The Mojave Desert in southern California is riddled with faults. Northwest-trending right-lateral strike-slip faults dominate the geology of the northern Mojave Desert from the San Andreas fault to the Blackwater fault near the west side of the China Lake Naval Weapons Center. East-west-trending left-lateral faults are common from the weapons center to the Eastern California Shear Zone along the east side of Fort Irwin National Training Center. This broken maze of desert is bounded by the left-lateral Garlock fault to the north and Pinto Mountain fault to the south.

This field trip, associated with San Bernardino County Museum's 12th Annual Mojave Desert Research Symposium and California State University's Desert Studies Symposium, will focus on the Garlock fault. We are especially interested in the age, geophysics, offset, and kinematic history of the Garlock fault. Hopefully, information obtained and shared on this trip will lead to a lively discussion on the significance of the Garlock fault relative to crustal dynamics of the Mojave Desert at our last stop. Along the way, we will learn a few things about the mineral resources and archaeology of this most interesting region. So keep your eyes and ears open, watch for faults, and have a wonderful trip.

Good hunting!

James P. Calzia, U.S. Geological Survey, March 1998

On the Afternoon Before Day One

0.0 0.0 Leave parking lot at San Bernardino County Museum in Redlands and take Interstate 10 west to I- 215.

3.7 3.7 Take I-215 north toward Barstow. We cross the trace of the San Jacinto fault at the junction of I-10 and I-215.

18.7 15.0 Junction of I-215 and I-15. Continue north.

21.8 3.1 The view west (left) is along Lone Pine Canyon, the San Andreas fault rift zone. Lost Lake is a sag pond developed on the fault trace (Crowell, 1975; Dibble, 1980).

25.2 3.4 Highway 138 junction. Continue on I-15.

33.6 8.4 Exit I-15 at Highway 395. Follow Highway 395 to Kramer Junction.

74.5 40.9 Kramer Junction. Proceed north on Hwy 395.

74.7 0.2 Pass federal prison.

77.5 2.8 Saddleback Buttes at 10:00 appear as small hills. This is the type locality for the Saddleback Basalt, a basal member of the Kramer Beds.

80.5 3.0 The radar facility to the west tracks the space shuttle on its approach to Earth.

81.5 1.0 Cross the trace of the south Lockhart fault as we pass through granitic rocks into Pleistocene alluvium.

82.9 1.4 Cross the trace of the right-lateral Lockhart fault. The Helendale and the Lockhart faults may merge to the northwest. Fremont Peak, to the east at 2:00, sits on the lower plate of the Waterman Hills detachment fault. Red Mountain is at 12:30. The Gravel Hills and Barstow Fm are in the distance at 3:00.

96.5 13.6 Cross the projected trace of the right-lateral Harper Lake fault. Cuddeback Lake and the Gravel Hills are on the north side of the fault, to our southeast. The fault runs on the north side of Fremont Peak, the prominent summit also to our southeast.

98.7 2.2 Red Mountain is to the northeast at 1:00. The "Spud Patch," post-World War II scheelite diggings, are around the head frame at 9:00.

101.1 2.4 The Kelly Mine and the historic Rand district are straight ahead. The Kelly Mine was located in 1919 and peak production was in the 1920s when it was the richest silver mine in California (Wright and others, 1953; Hulin, 1925).

103.1 2.0 Trona Road exits to the right. Continue on Hwy 395.

104.0 0.9 Pass through Johannesburg and proceed downhill. Look for the approaching left turn to Randsburg.

105.0 1.0 TURN LEFT (south) off Hwy 395, towards Randsburg.

105.6 0.6 Pass Goler Road and proceed to Randsburg.

106.1 0.5 TURN RIGHT (westerly) onto Randsburg/Red Rock Road. The Randsburg Museum is to our left (southwest).

112.3 6.2 Pass fenced well on the right. A left turn onto a BLM road is in 0.9 miles.

113.2 0.9 Junction on left with BLM road running south toward Rand Mountains.

114.4 1.2 STOP at junction.

End of Day Before Day One.

DAY ONE: Goler Gold

Robert E. Reynolds and *Brett Cox*

Today we will visit the Goler Fm at the northern margin of the Mojave Desert. Sediments filling this Paleocene basin give us an idea of early Mojave Desert history and pre-Garlock fault structure. We will follow sections of the central and eastern Garlock fault. This left-lateral fault zone marks the northern boundary of the Mojave Desert Province. Fault movement may have started in the middle Miocene (Carter, this volume; Monastero et al, 1997) and has bisected distinct lithologies of volcanic rocks that were once part of the Eagle Crags volcanic field. Offset within the last 17 million years may total 40 miles (64 km.). This area is seismically active. Earthquakes of magnitudes 1.6 and 1.6 were recorded in March, 1998 on the Garlock fault zone in the Lava Mountains and the Helendale fault zone near Boron. A magnitude 2.5 event occurred in January 1998 on the Lockhart fault zone south of Harper Lake (see Hauksson, this volume).

0.0 0.0 CONVENE at junction of Red Rock/Randsburg Road and Garlock Road. Ahead to the northwest is an uplifted escarpment that sits between the Garlock fault, the trace of which is close to us at the base of the hill. The El Paso fault is farther north and at this point out of sight. The escarpment consists of interfingering fanglomerates from Mesquite Canyon and lacustrine sediments deposited much like those being deposited today in Koehn Dry Lake, which we can see due west at 10:30. The southern face of the El Paso Mountains is formed by the two faults. From this point to the east, Mesquite Canyon, Iron Canyon and Goler Gulch, respectively, cut through the El Paso Mountains and deposit fans which contain distinctive gravel lithologies (Carter, 1987). PROCEED EAST (turn right) on Garlock Road.

0.8 0.8 Pass through historic Garlock (Hensher, this volume). Garlock flourished in the late 1800s when gold was discovered in

Goler Gulch. The scarp of the Garlock fault was first recognized here (Hess, 1909). The El Paso Range rises over us to the north. At this point it consists primarily of metamorphosed Paleozoic sediments of the Garlock Group that appear to dip northeasterly.

2.7 1.9 40 MPH sign. Prepare to turn left into Iron Canyon. Watch for oncoming traffic and large trucks.

2.8 0.1 TURN LEFT and proceed north up once-graded dirt road with BLM signs into Iron Canyon. You will need 4WD in this canyon.

3.6 0.8 Flat-lying, well-cemented Tertiary? gravels overly steeply-dipping and contorted Paleozoic siltstone.

3.7 0.1 Drive up waterfall.

4.0 0.3 TURN LEFT up tributary out of Iron Canyon.

5.4 1.4 BLM plastic slat sign on the right side of the road. Take the good road which bears left (northwest).

5.5 0.1 Second saddle and second BLM plastic post between 2 forks. We **take the right fork** due north. **STOP** in saddle for overview of the metamorphosed Paleozoic rocks of the Goler Group, the Goler Fm and the early Miocene Cudahy Camp Fm.

5.6 0.1 BLM plastic sign on the left and 3 metal fence pipes on the right. This is a 3-way intersection. Do not take the left (reverse) turn. Do not take the right fork. **Go straight ahead** toward Mormon Flats.

6.2 0.6 TURN RIGHT at Mormon Flat junction, a T intersection with a BLM sign to the left and a BLM sign to the right.

6.7 0.5 TURN LEFT (northeast) at T-intersection toward alternating red beds of the Goler Formation (Fig. 1).

6.8 0.1 Proceed north-northeast at intersection. Do not take right turn to mine.

7.0 0.2 Proceed left (northwest) at reverse Y intersection..

7.2 0.2 Pass left turn. Proceed north and drop into Mormon Gulch.

7.5 0.3 Note interbedded fluvial sediments and lignites.

7.6 0.1 TURN SHARP LEFT out of wash and proceed up terrace.

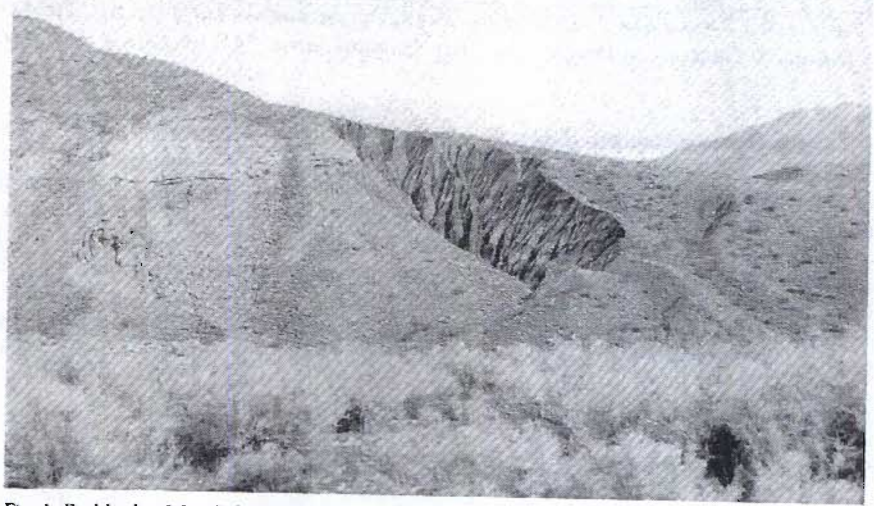

Fig. 1. Red beds of the Goler Formation north of Mormon Mesa. *R.E. Reynolds photo.*

7.7 0.1 STOP on top of terrace at complex intersection. Mormon Gulch is to the south, Goler Gulch is the deep canyon to the east (Cox, this volume). RETURN to vehicles and RETRACE route to Iron Canyon.

8.2 0.5 TURN LEFT and climb terrace on south side of Mormon Gulch.

8.3 0.1 Proceed south on top of terrace; do not take right turn.

8.4 0.1 TAKE RIGHT FORK to southwest at complex turn.

8.6 0.2 Join reverse junction and proceed southwest.

8.8 0.2 TURN RIGHT (northwesterly) at T intersection towards Mormon Flats.

9.3 0.5 TURN LEFT between BLM plastic posts at T intersection.

9.8 0.5 Proceed on left fork at complex intersection. There is a BLM plastic sign on the right and 3 metal posts on the left.

10.0 0.2 Saddle with view. BLM plastic sign on right.

10.1 0.1 Saddle with plastic sign on left and quail guzzler on left.

11.3 1.3 TURN RIGHT into Iron Canyon proper. As you proceed down Iron Canyon, note calichified Quaternary sediments perched high on the walls of the canyon.

11.7 0.4 Waterfall.

12.5 0.8 Stop at junction with paved Garlock Road. Watch for cross traffic. TURN LEFT onto Garlock Road.

12.7 0.5 Pass tamarisk trees at well site. The community of Goler Heights is to the northeast at about 10:00.

13.0 0.3 The Yellow Aster mill is at 10:00 on the south flank of a large depression (Fig. 2).

13.8 0.8 Watch for cross traffic and TURN LEFT. Proceed northwest up graded dirt road toward the track of the Garlock fault scarp.

14.0 0.2 **STOP** and PARK at the intersection of the graded road and a northeast-trending crossroad. We

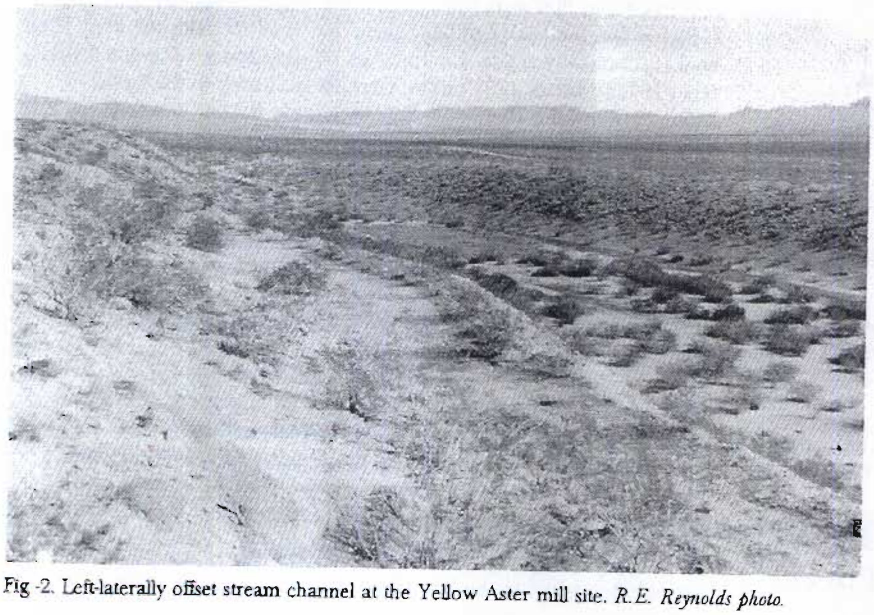

Fig -2. Left-laterally offset stream channel at the Yellow Aster mill site. *R.E. Reynolds photo.*

are parked on the trace of the Garlock fault. Walk to the west and look down the depression south of the Yellow Aster Mill. This depression on the south side of the Garlock fault was formed when Goler Wash was offset left-laterally. Walk easterly to the north side of the ridge and look east along the trace of the Garlock fault. Notice the graben on the north side of the fault, east of where we parked. Look southeast across Cantil Valley. We can see coarse old alluvial linear ridges above the recent alluvium. These mark the trace of the Cantil Valley fault, which parallels the Garlock fault on the south side of Koehn Lake. Return to vehicles and RETRACE to pavement.

14.2 0.2 Stop at pavement, look both ways, TURN LEFT onto Garlock Road and proceed EAST to power line road.

14.5 0.3 Straight ahead at 12:00 is a shutter ridge of gravel crossed by a set of power towers. Low on this ridge are boulders which could only have come from Paleozoic Garlock Group sources in Mesquite Canyon, 7 mi (11 km) to the west. High on the ridge are lithologies that come from the Black Mountain Basalt and the Goler Formation in Goler Wash, the mouth of which is 4 mi (6.4 km) to the west (Carter, 1987 and this volume). We will drive over this ridge to look at the scarp of the Garlock fault on the north side.

15.6 1.1 Notice the two large piles of gravel ahead on the right side of the road. We will turn left using these piles as a landmark.

16.3 0.7 TURN LEFT across from the gate on the right, past the first pile of gravel. If you cross under the metal power line, you have gone too far.

16.4 0.1 The powerline road bears westerly up the dissected gravel ridge.

17.0 0.6 Intersection at top of gravel ridge. Proceed north on powerline road and wind downhill.

17.3 0.3 PARK past low ridge on left. We are at junction with a west-trending road with a good view down the scarp on the north side of the gravel ridge. Walk to explore the brecciated Garlock fault zone. Return to vehicles RETRACE to highway near gravel pits.

20.0 0.7 Stop at pavement of Garlock Road. TURN LEFT (east) towards Highway 395.

21.9 1.9 Stop at the intersection of Hwy 395 and Garlock Road. TURN RIGHT onto Highway 395 but be prepared to turn left before you reach the upcoming railroad tracks.

22.4 0.5 TURN LEFT across from yellow railroad sign onto dirt road. Pass the placer diggings of Hard Cash Gulch.

23.1 0.7 Proceed through complex junction. This is the Dusenburg Gold Mining Association at Phoenix Gulch.

23.5 0.4 Proceed straight, past left turn to Dusenburg Gold Diggings.

23.6 0.1 Stop at railroad tracks. Proceed across tracks and parallel tracks on south side. After 0.1 miles the road drops away but we continue to parallel tracks. Note calichified surface of fanglomerate on both sides of the tracks.

23.7 0.1 Road drops downhill towards wash and culvert under railroad. Prepare to turn left into culvert and then sharp right on far side of culvert. Culvert is one lane wide so watch for oncoming traffic.

23.8 0.1 TURN LEFT into culvert. Pass through culvert and TURN SHARP RIGHT. This is sandy; don't slow down. Take railroad access road parallel to railroad tracks.

24.4 0.6 Road curves slightly away from railroad at junction. **Take the right turn** which leads back toward the railroad tracks.

24.8 0.4 Watch for washouts on left side of road.

24.9 0.1 Junction; stay right, toward railroad tracks.

25.4 0.5 Approach tunnel. At the mouth of the tunnel, the Garlock

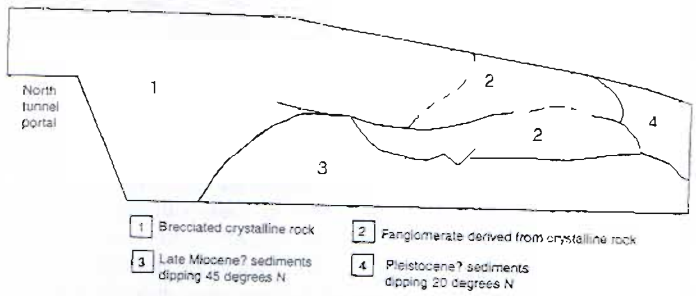

Fig. 3. Sketch of north tunnel portal in Summit Range, view west. 1: brecciated crystalline rock. 2: fanglomerate derived from crystalline rock. 3: Late Miocene? Sediments dipping 45ºN. 4: Pleistocene? Sediments dipping 20ºN.

fault places the Miocene Cudahy Camp Fm of the Summit Range on the south against Mesozoic granite on the north. The tunnel was built in 1908. The granitic ridge may have been offset left-laterally from its counterpart 4 miles east in the Summit Range. Proceed to north end of tunnel on a sinusoidal road.

26.2 0.8 **STOP**. Exposed in the westerly wall of the north tunnel portal is a very fractured granitic breccia that has pushed northward over steeply-dipping gravels derived from the same rock source. This overlies brown sediments that dip northerly. The relationship suggests compression along the Garlock fault which created north-vergent thrust faulting. Thrusting deformed older sediments so that they dip 45ºN. These were covered by crystalline rock debris and younger sediments with a shallow dip (Fig. 3). Return to vehicles and PROCEED NORTH to the siding of Searles and Searles Station Road.

27.3 1.1 TURN RIGHT onto the pavement of Searles Station Road.

27.8 0.5 BLM signs for Spangler area.

28.4 0.6 We are crossing light-colored sediments east of granitic outcrops. These were deposited by ponding against a northwest-trending fault which runs into Indian Wells Valley.

29.3 0.9 Cross railroad tracks.

32.3 3.0 **STOP** at the junction of Searles Station Road and Trona Road. To the south-southeast we can see the crest of the Summit Range and the trace of the Garlock fault. The sediments south-southeast are granitic and presumably left-laterally offset from the granites at the tunnel stop. Follow the trace of the fault until you are looking southeast. You can see the Bedrock Springs Fm which consists of volcaniclastic sediments, lacustrine sediments, and coarser fluvial sediments. To the east-southeast you can see Pilot Knob and slightly to the north Robbers Roost. Further east is the Eagle Crags volcanic field which has been offset 64 km (40 miles) in the last 17 million years from distal flows that we saw at the start of this day (Monastero et al., 1997). The Late Tertiary volcanic and clastic rocks of the Lava Mountains were presumably deposited in a trough formed along the Garlock fault at the north edge of the Mojave block. The overlying Quaternary volcanics were deposited unconformably on uplifted and deformed Tertiary basin fill. Limited exposures of questionable middle Miocene sediments and volcanics occur on Cretaceous Atolia quartz monzonite at the base of the section. These are overlain by the Bedrock Spring Formation that is Late Hemphillian in age. This formation is unconformably overlain by the Almond Mountain volcanics and the Lava Mountain andesite. These volcanics are in turn overlain by the Christmas Canyon Fm which lies at the far east end of the Lava Mountains (Smith, 1964). Searles Lake once filled the valley to our east and left well-developed shorelines at elevation 2,250'. These shorelines and associated tufa may be as old as 50,000 ybp and have been developed across the Garlock fault. This constrains the last major Garlock movement to older than 50,000 ybp (Smith, 1964). PROCEED SOUTH into the Spangler Hills/Teagle Wash area.

33.9 1.6 Cross railroad tracks

34.4 0.5 Pass left turn to Teagle Wash OHV area.

36.0 1.6 **STOP** at the junction of Trona Road and a bypass to the tunnel stop. From here, you can look southeast to the trace of the Garlock fault on the skyline, follow it to the east-southeast. We can see Robbers Roost but not Pilot Knob. A bit north of due east we can see dark mountains of the Slate Range. The magnitude of displacement on the Garlock fault was first measured by comparing disjunct outcrops of the Independence Dike Swarm (Smith, 1962). The dike swarms have since played an important part in interpreting offset patterns and block rotation throughout the Mojave Desert (Ron and Nur, 1996; Howard and Hopson, 1997; Moore and Hopson, 1961; James, 1989; Schermer and others, 1996). The white patch at 2:00 is Searles Lake and looking closely you can see the Pinnacles. At 1:30 in the foreground are the Spangler Hills.

36.9 0.9 BLM sign marks the turnoff to the Spangler OHV area.

37.1 0.2 Top of rise. We cross the Garlock fault, marked by the change from granitic rocks to volcaniclastic sediments and volcanic rocks.

37.7 0.6 Directly ahead south are red beds and white ash-rich sediments of the Bedrock Spring Fm.

39.1 1.4 Leave the Summit Range. Due south at 12:00 is a perched playa on the right side of the road. This playa may be developed on the south side of the eastern trace of the Cantil Valley fault.

41.0 1.9 At 2:00 are the microwave towers situated above Johannesburg. At 11:00 is Red Mountain. At 10:00 is Steam Well Valley. Klinker Mountain and Dome Mountain are at 9:00.

42.9 1.9 Road bends right and goes over a slight rise. Prepare to turn left across traffic at the dirt road we see about 1000 feet away on the east side of the road.

43.6 0.7 TURN EXTREMELY SHARP LEFT onto Steam Wells Road, historically the route to Pilot Peak (Mendenhall, 1909). Proceed northeast.

43.9 0.3 Pass right turn to mine.

44.1 0.2 Cross pipeline and road forks. Stay on right fork.

44.4 0.3 Proceed past turn to right. We are near wells that supply Johannesburg and Randsburg with water. They are described as Old Wells, City Well, and Mountain Wells (Mendenhall, 1909) and some were powered by steam pumps.

45.4 1.0 Y turn. Stay to the left. Keep the BLM plastic post on your right. Watch for dropoff.

45.9 0.5 Stay left in Y junction.

46.4 0.5 Pass through crossroads and mine dumps to right.

46.7 0.3 Junction to left goes to Steam Well, drilled as a mercury prospect in 1920, when it encountered water at 96°C (Smith, 1964). Now a wilderness area, the roads have been blocked.

48.1 0.4 Pass second junction to Steam Well.

50.1 2.0 Leave Steam Wells Valley and enter broad, open Golden Valley (Thompson, 1929) which contains Cuddeback Lake. Granite Mt is the low ridge due east. Cuddeback Lake (Willard Lake, Mendenhall, 1909) is east-southeast. Almost due east to the north end of Cuddeback Lake is Slocum Mountain. The sharp mountain to the southeast is Fremont Peak. Almond Mountain is to the east-northeast and is made up of Almond Mountain volcanics intruded by domes of Lava Mountains andesite. The base of Almond Mountain is cut by high shorelines of Cuddeback Lake. A prominent shoreline is developed at 10 feet above the current playa surface of 2,553'. A higher shoreline is seen in two places on southern Almond Mountain at 2,660'.

50.5 0.4 Pass through junction marked by thin steel pipe. This road goes south to the mill and tailings of the Blackhawk Mine.

51.8 1.3 Structures and tamarisks mark the site of Brown's Ranch. We are in a northeast-trending valley created by the Browns Ranch fault zone on the west side of Almond Mountain. Proceed westward to the southern tip of Almond Mountain.

52.6 0.8 Pass reverse Y turn road. Stay to the right. Road bears southeasterly around the southern tip of Almond Mountain.

52.8 0.2 Road bends due easterly. We are near the surface of Cuddeback Dry Lake.

53.0 0.2 Stay left at Y junction.

53.5 0.5 We are at the southern tip of Almond Mountain.

54.3 0.8 Pass BLM sign at junction to right (south).

54.8 0.5 Pass large wood frame over cattle guard and junction with road to right (south).

56.1 1.3 Pass right turn to water tanks marking the site of Willard Well on the Death Valley Borax Road. Note that the water tanks sit at a distinct line demarking a change in vegetation. This change is at the contact of playa silts and arkosic sands on the granitic bedrock slope. The playa margin is poorly drained; the slopes are well drained.

57.3 1.2 The plastic post on the left marks a crossroads. The left road runs to the southern Slate Range and then forks toward Wingate Wash and Death Valley Borax Works or to Leach Lake and Saratoga Springs. We **proceed on right crossroads** going easterly. On the

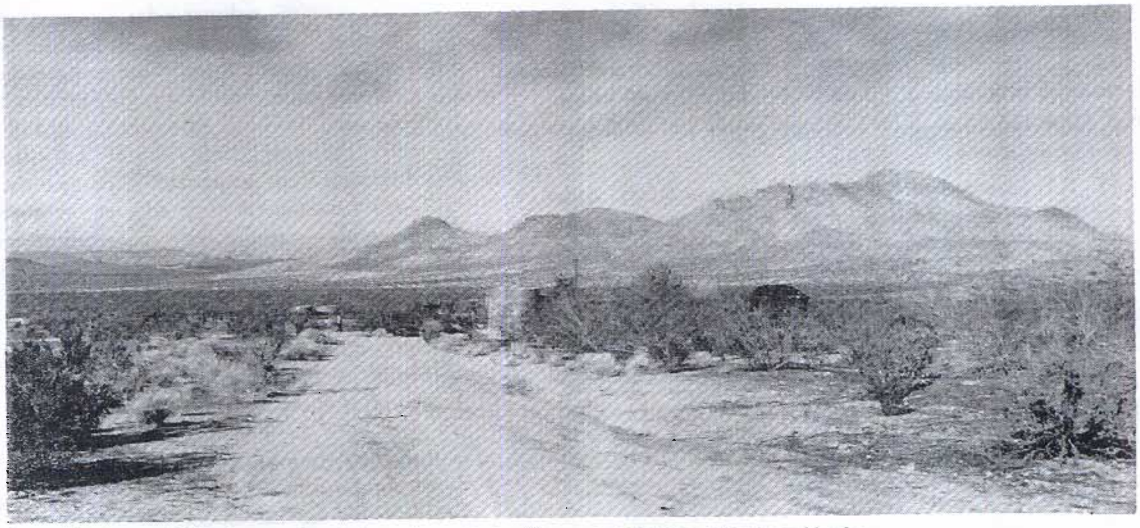

Fig. 4. Blackwater Well, view northwest toward granitic inselbergs on pediment. *R.E. Reynolds photo.*

horizon at 10:30 is Pilot Knob, a butte with a dark reddish top and white ash beds underneath. Straight ahead are the Black Hills basalts sitting on Mesozoic granitic rocks. To our right at 1:30 to 3:00 are Tertiary intrusive rhyolites, a row of plugs that run southeasterly past Grass Valley to Black Canyon. Proceed eastward towards Blackwater Well.

Our route to Blackwater Well takes us up an erosional surface developed of Mesozoic quartz monzonite ("mountain pediments" of Thompson, 1929, p. 225) (Fig. 4). Mountain pediments are eroded rock surfaces beveled across hard rocks and covered with alluvium only to a comparatively slight depth (Thompson, 1919, p. 225). These surfaces are seen throughout the Mojave Desert and are often overlain by dated early Miocene volcanics including the Peach Spring Tuff (18.5 Ma, Nielson et al., 1990). The age of the rocks on this surface suggests that the surface is pre-Miocene in age. Round granitic boulders remain exposed in certain areas. These have been interpreted as having formed during deep subsurface weathering (Oberlander, 1972). Since the Miocene, other areas have been stripped of such boulders. Our route crosses four areas of granitic pediment development: on Day One, in the vicinity of Blackwater Well and at Coolgardie Summit; on Day Two at the entrance to Fort Irwin, north of the Coyote Lake fault, and in the eastern Granite Mountains.

61.2 3.9 Complex of roads joins at the white gate to Blackwater Well. Stop, open gate, and proceed through and last person closes gate. We have been traveling eastward up a large domelike erosional surface formed on granitic rocks. To our north there appears to be retreating inselbergs of granite. We should also expect retreating granite faces underneath the basalt of the Black Hills, which lie to the northeast. If we look north past the military facility, we will look into the Teagle Wash/Spangler area where we were earlier this day.

61.5 0.3 Proceed right past left turn towards house and tank at Blackwater Well.

61.6 0.1 Junction at windmill. TURN RIGHT past the east side of a small hill. Blackwater Well is located near the junction of roads from Randsburg to Silver Lake, Death Valley borax mines and Resting Spring (Lyman and Walker, 1997). It also lies along historic routes running north from Barstow to Ballarat in the Panamint Valley and to the Searles borax works. The well's importance as a watering place was noted 90 years ago: ". . . The well can be located by the bare ground . . . from which campers have stripped all vegetation. The well, which was dug years ago by government troops, is 15 feet deep . . . water was at one time piped down the slopes for a distance of one-half mile to the old Death Valley Borax Works road . . . a road that turns off at the wells crossed the divide to the east and extends to Copper City, a small mining camp" (Mendenhall, 1909). By 1917, a gasoline engine and pump at Blackwater Well provided water for cattle (Thompson, 1929).

61.9 0.3 Pass through saddle. Notice the spheroidal granitic boulders developed by subaerial weathering.

62.1 0.2 Join the first road and proceed easterly.

62.2 0.1 Proceed up erosional surface on granite. Aplite dikes are seen to the right (south).

63.4 1.2 Cross over saddle. The granitic surface runs uphill to the southeast (right). Notice the spheriodal granitic boulders on this erosional surface. Black Hills basalt can be seen ahead at 11:00.

64.3 0.9 Cross cattle guard and fence line road intersection.

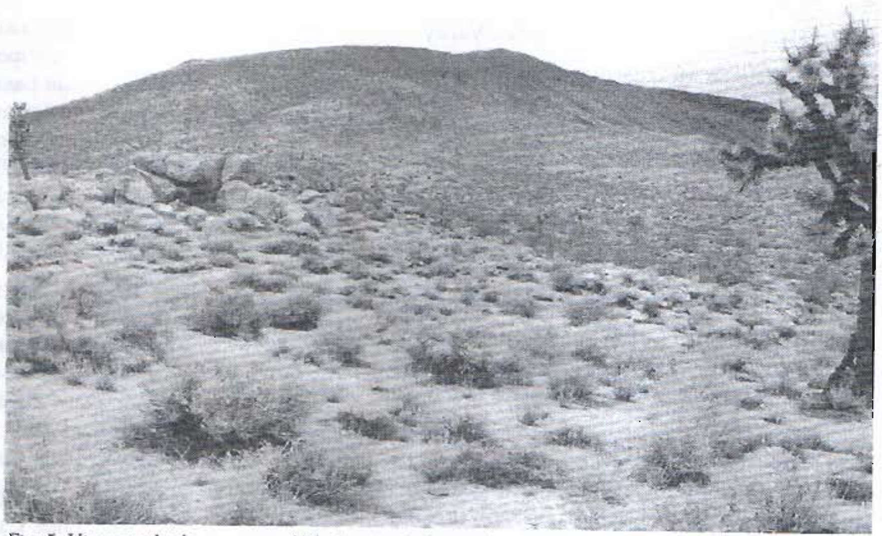

Fig. 5. View north along scarp of Blackwater fault trending toward basalts of the Black Hills. *R.E. Reynolds photo.*

64.7 0.4 TAKE THE RIGHT FORK at Y junction. The road to the left leads to the site of Copper City and then southeast to Coolgardie (Mendenhall, 1909). Copper City is located seven miles east and to the southeast of Pilot Knob. In 1909 it was described as a "small mining camp" (Mendenhall, 1909 p. 53). The Juanita (Dixie) copper prospects were reported east of Pilot Knob near the turn of the century (Aubury, 1902; Aubury, 1908). By 1917, Copper City was abandoned: "..several small houses at this place . . . are now destroyed." (Thompson, 1929.

65.2 0.5 Park and **STOP** for view. The Blackwater fault scarp cuts granite 300 feet to the west. The Black Hills are to the north-northwest (Fig. 5). Pilot Knob is the striped butte to the northeast, sitting on the concordant surface of the Granite Mountain erosional surface. Tertiary volcanic buttes cap the granitic ridge and run south toward the volcanic cap that is Slocum Mountain.

The Mojave Desert is characterized as a geologic province lying between the left lateral, west-trending Garlock fault and the right lateral, northwest-trending San Andreas fault. The western Mojave Desert is crossed by a number of northwest-trending, right lateral faults that parallel the San Andreas fault. The most northerly of these is the Blackwater fault. Interpretation of this fault from geologic maps (Jennings et al., 1962; Dibble, 1968) indicates that the Blackwater fault runs to the north end of Almond Mountain where it is truncated by faults in the southern Lava Mountains. Mapping by Roquemore et al., 1982, shows the fault truncating against the Garlock fault. Microseismicity following the 1992 Landers earthquake also suggests that the Blackwater fault extends to the Garlock fault and possibly beyond to the north. It runs southerly to Opal Mountain where it bifurcates. Projecting the fault southeastward, it aligns with southeast-trending faults that cut the Barstow Formation. These include the Coon Canyon fault which may be a northern projection of the Calico fault (Reynolds, 1992). Topographic features southeast of Blackwater Well suggest right lateral offset of approximately three miles. In western Superior Valley, geologic features ranging in age from Mesozoic, Tertiary, and Quaternary appear to have been offset only 1.5 miles right laterally. The Blackwater fault does not exhibit simple strike-slip movement. In the Blackwater Well area, where the fault bends concave westward, there is a west-side-up component. In the Opal Mountain area where the fault steps to the east, there is an east-side-up component of movement. The youngest rock cut by the fault is the Late Pliocene Black Mountain Basalt, dated at 2.55±0.58 Ma (Burke et al., 1982).

65.5 0.3 Proceed southeasterly along the trace of the Blackwater fault, passing a junction to the left.

66.2 0.7 Pass over the crest into Grass Valley.

67.0 0.8 Slocum Mountain is to the east-southeast at 10:30. Lane Mountain is at 12:00, Coolgardie Ridge is slightly to the rightt. Opal Mountain is about 1:30, and rhyolite intrusions are sharp silhouettes on our right, southwest, at 3:00.

68.2 1.2 Fence with "restricted area" signs. Proceeding south along the military boundary.

68.3 0.1 Cross the approximate trace of the Blackwater fault.

70.5 2.2 TURN LEFT (east) at military fence corner.

71.4 0.9 TURN RIGHT (south) on post road that runs southwesterly to the Arena Mine and southerly towards Black Canyon. We are in the lacustrine silt and silty arkosic sediments of the playa of Grass Valley.

71.5 0.1 Cross the cattle guard and fence line and continue south. Do not take right turn toward the Arena Mine. The pedogenic carbonates, lacustrine sediments, and silty arkosic playa sediments of Grass Valley are ponded on the northeast side of the Blackwater fault.

72.2 0.7 Leave lacustrine sediments and enter granitic terrain as we cross the trace of the Blackwater fault. Road goes up a hill onto Pleistocene gravels that contain volcanic and granitic rocks.

74.4 2.2 We are driving through a gorge that drains Grass Valley. From Grass Valley southward, all of the basins east of the west-side-up Blackwater fault drain southwest to Harper Lake basin.

75.5 1.1 The hills to the left at 9:00 consist of volcanic and granitic gravels.

75.8 0.3 We are entering the north margin of a southwest-dipping basin that received volcanic debris from sources to the northeast as well as from Opal Mountain to the southeast. These gravels were shed southwestward until they interfingered abruptly with granitic gravels derived from Fremont Peak to the southwest. The granitic gravels overlie lacustrine silts and both are considered approximately middle Miocene, Barstovian LMA (Dibblee, 1968).

76.7 0.9 Pass through spectacular light-colored cliffs of ash and perlite of the Opal Mountain volcanics (Dibblee, 1968). The Hicks Perlite Company mined in this area.

77.1 0.4 **STOP** at complex intersection. In Black Canyon, a steeply-dipping bedding plane of Opal Mountain perlite is to the east. Opal Mountain is to the southeast and there are many localities for semi-precious opal, agate, and jasper in the area. On the skyline is the Late Pliocene Black Mountain basalt which unconformably overlies a Late Miocene erosional surface developed on coarse and fine sediments of the Barstow Formation. The basalt defines the surface topography at 2.5 Ma (Burke and others, 1982). Terrace gravels 1000 feet down drainage on the west side of Black Canyon contain clasts of Black Mountain basalt. These Pleistocene terrace gravels reflect the development of drainages from Superior Valley, southwest across the Blackwater fault, across the Gravel Hills and the Black Mountain basalt, and into Harper Basin. Return to vehicles and proceed north-northeast up Black Canyon.

78.3 1.2 Continue north; do not take right turn into basalts.

79.3 1.0 **STOP** at the basalt point. We first crossed the west-side-up trace of the Blackwater fault east of Blackwater Well. At Grass Valley, the west-side-up movement of this lateral fault dammed lacustrine sediments in a playa. At this point, the Blackwater fault cuts the Pliocene Black Mountain basalt (2.5 Ma, Burke et al., 1982). This is the youngest dated rock cut by the Blackwater fault. The west-side-up component of this right lateral fault apparently caused internal drainage in Superior Valley during Pleistocene times. Filling of Superior Valley in the Pleistocene may have produced overflows which cut Black Canyon and left perched terrace gravels.

From this point on the Blackwater fault to southeast, the apparent vertical component on the fault is east side up. We can see the Black Mountain basalt flows dipping at approximately 10º to the northeast. The Quaternary playa surface of Superior Valley is surrounded by silty sands indicating a higher shoreline in the Pleistocene. Mapping (Dibblee, 1968) indicates that the current playas are higher in the west and lower to the east. One mile east of our stop, the playa surface is elevation 3040'. At the east end of Superior Valley, Superior Lake lies at elevation 3,002'.

The east side up component of the southern Blackwater fault may be due to compression east of the Harper fault that relates to an area where there are active east-west trending faults such as the Coyote Lake fault. The west-side-up component and westward curvature on the northern Blackwater fault may be due to its proximity to the left lateral Garlock fault. Return to vehicles and PROCEED SOUTHEAST to Inscription Canyon.

80.0 0.7 **STOP** at the junction of Black Canyon Road and Inscription Canyon Road. Inscription Canyon is marked by a telephone pole fence. The canyon is eroded westerly into the basalt scarp created by the Blackwater fault. Numerous petroglyphs can be seen locally, and one intaglio is on the surface of the basalt flow (see Whitley, this volume; Turner, 1994). Return to vehicles and PROCEED EASTERLY around the south margin of the playas in Superior Valley. *CAUTION: If playa surface appears at all wet at this point, it will get worse as we go east and vehicles will become mired. Return to Black Canyon and drive south. Be sure to turn easterly to Hinkley Road to avoid Harper not-always-dry Lake.*

83.1 3.1 Pass point of Late Pliocene Black Mountain basalt on the right.

85.4 2.3 Pass reverse intersection on the left. This road runs north to exposed Pleistocene dune sand along the south margin of one of the Superior Lake playas. The road south goes to perlite mines near Willis Well.

86.3 0.9 Continue east through crossroads. The low hill to the southeast is Mesozoic hornblende diorite.

86.5 0.2 Watch for dip.

86.8 0.3 Proceed east as our multiple-track road meets a 2-lane, graded road which turns north and leads to the north side of Superior Lake.

87.4 0.6 Slow for bend in road around gravel pit.

87.8 0.4 Pass south of Crutts Well. This well is mentioned by Mendenhall (1909) and by 1917 named for the nearby Crutts post office (Thompson, 1929). The Ballarat/Barstow Road of 1920 ran northwest-southeast at this point. To the north it went to Copper City, Granite Wells, Lone Willow Spring; to the north, it led to through the Panamint Valley to Ballarat.

88.0 0.2 Continue past a northeast-southwest junction. This is the Barstow-Ballarat Road that runs past Crutts post office and Crutts Well. Historically, it ran southeast to connect with what is now Copper City Road.

89.6 1.6 TURN RIGHT at junction with Copper City Road; proceed south. This was the road to Indian Spring (Mendenhall, 1909) and was called the Hidden Spring branch of the Barstow-Ballarat Road (Thompson, 1929). To the north is the Eagle Crags volcanic center. To the northeast are Paleozoic sedimentary rocks that have been intruded by Cretaceous granitic plutons. To the east are basic intrusive rocks. We are heading south up an erosional surface that has been developed on Mesozoic quartz monzonite (Dibblee, 1968). Southwest is the andesite mesa called "Mesa" and south-southwest is the andesite mesa, "Spear." These are composed of Lane Mountain andesite flows, dated at 23.1±0.2 Ma (Burke et al., 1982), the oldest Tertiary rock lying directly on the pre-Miocene erosional surface.

93.5 3.9 We have crossed a diorite ridge dividing the drainage on the north to Superior Valley from a drainage regime that cuts across the Blackwater fault and flows westerly towards Black Canyon and Harper Lake.

97.8 4.3 Lane Mountain is at 9:30.

98.0 0.2 Continue through intersection. To the west lies Coolgardie Camp, described as scattered miners' cabins with dry placer workings (Mendenhall, 1909). Water for Coolgardie was brought from Kane's Well, also the site of a stamp mill. Kane's Well was also known as Lane's Well and Williams Well (Thompson, 1929).

98.8 0.8 Continue past several intersections and a community of collectibles.

100.3 1.5 Prepare to turn right at the pole line road, visible ahead.

100.6 0.3 TURN RIGHT at Y intersection and follow the pole line road to the microwave relay station on top of the hill.

100.8 0.2 PARK at junction 1/10 mile before microwave facility. Walk to view stop. The view to the west includes Harper Lake, and the southern Greenhorn Mountains. Harper Lake sits in a valley south of the Harper fault and north of the Lockhart fault. West-southwest you can see Leuhman Ridge. The Shadow Mountains are west-southwest. Southwest on the horizon is Mt. San Antonio and, in the foreground and a bit north, the Mojave River and Mt. General. Directly over the freeway is the Lenwood Anticline, a compressional feature between the Lenwood/Lockhart fault and the southern Harper fault. Above the anticline is triangular Stoddard Mountain. Below it are the Waterman Hills. We are looking into the valley that contains the Fossil Bed Road fault of Dokka which may be a predecessor to the Owl Canyon fault, a projected trace of the Calico fault (Dokka et al., 1991; Reynolds, 1992). Looking south we can see the Marine Corps base below Daggett Ridge. South-southeast is Ord Mountain, elevation 6270'. Further east from Ord Mountain are the Newberry Mountains and Rodman Mountain. On the skyline we look east-northeast to Lane Mountain, with the oldest dated Tertiary volcanic rocks in the area. On the north side of Lane Mountain isTiefort Mountain on Fort Irwin. To the northeast are the Avawatz Mountains where we end Day Two of this trip. North past Superior Valley are the rugged Eagle Crags volcanic field with dates as young as 17 Ma. That area on the south side of the Garlock fault has been offset 65 km from the Garlock Group and Cudahy Camp volcanics where we started our trip this morning. We are on the granitic erosional surface developed prior to Hemingfordian times (20 Ma). We passed similar erosional surfaces at Blackwater Well and in Grass Valley. Return to vehicles and RETRACE to Copper City Road.

101.2 0.4 Pass left turn to camping area.

101.6 0.4 **STOP** at bend in road for overview. We are at the approximate trace of the Coolgardie fault and north of the trace of the Mud Hills fault (Dibblee, 1968). Southeast, the quartz monzonite block between these faults ends abruptly at a steep contact (or fault?) with the Barstow Formation. From here, we can look easterly at the elevated pediment surface and an abrupt scarp-like margin that may mark the eastward continuation of the Coolgardie fault zone as it trends eastward and meets the east/west-trending fault zones that we will see on Day Two of this trip (Jennings et al., 1962).

102.1 0.5 Road bends easterly at complex junction.

102.7 0.6 Stop, look both ways, TURN RIGHT (south) onto Copper City Road.

105.4 2.7 Pass the green and red volcanics of the Pickhandle Formation and drive through the siltstone of the Barstow Formation.

106.7 1.3 Continue past a right turn to a mine with zeolitized tuff in the Barstow Fm.

108.1 1.4 Stop at intersection with paved Irwin Road. Watch for fast traffic; TURN RIGHT and proceed southwesterly.

109.0 0.9 TURN RIGHT onto Fossil Bed Road.

112.0 3.0 TURN RIGHT onto Rainbow Basin Loop Road.

112.3 0.3 TURN RIGHT at junction with loop road onto Owl Canyon Campground Road.

114.3 2.0 Owl Canyon Campground.

End of Day One

Day 2

James Calzia

Today's field trip takes us across Fort Irwin, over the rugged Avawatz Mountains, and into Death Valley. Along the way, we will visit the recently discovered Richard W archaeological site, cross several major fault zones including the Garlock fault, and ponder the significance of aseismic slip and intensely deformed rocks. Observations and insight gained yesterday and today hopefully will fuel a lively discussion on the significance of the Garlock fault (and other east-west-trending faults) in the Mojave Desert at our last stop. WARNING: This day includes 23 miles of extremely rough roads. If you are concerned about shocks, springs, or other parts of your car's suspension, DO NOT GO on this part of the field trip.

0.0 0.0 Intersection of Fossil Bed Rd and Irwin Rd. Turn north onto Irwin Rd

4.4 4.4 Intersection Irwin Rd and Fort Irwin Rd. Turn north onto Fort Irwin Rd.

6.8 2.4 Cross summit of Calico Mountains at Pickhandle Pass. We are driving through volcanic flows, plugs, and breccias within the late Oligocene and Miocene Jackhammer Formation and Miocene Pickhandle Formation

17.4 10.6 Coyote Lake to right. This dry lake is located in a pull-apart

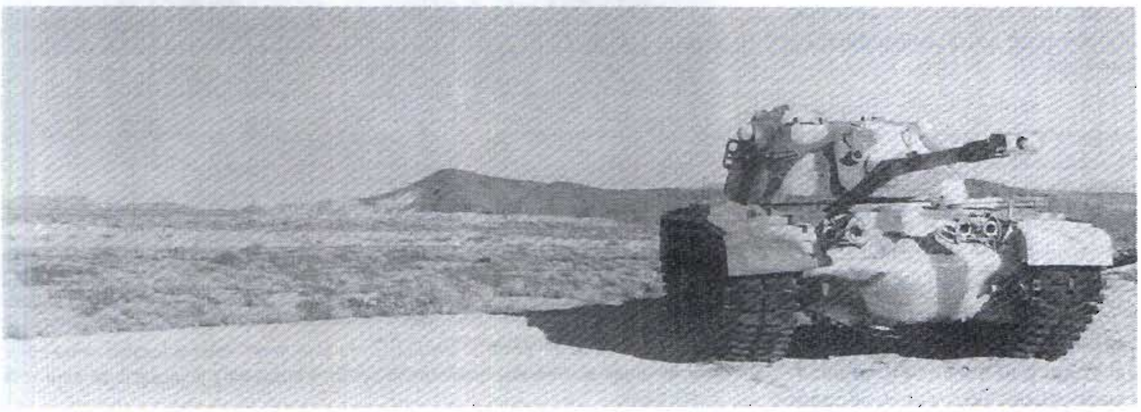

Fig.6. View eastward along the trace of the east-west-trending Coyote Lake fault at Fort Irwin welcome post. *R.E. Reynolds.*

Fig 7. Mike Messersmith (left) and Mark Allen standing in possible living area at the Richard W Site, Fort Irwin National Training Center. *Jim Calzia photo.*

basin created as Alvord Mountain (low peak on skyline to right) was translated eastward from its original position against the northern flank of the Calico Mountains and Paradise Range (Dokka and Travis, 1990).

19.6 2.2 Cross Coyote Lake fault (Fig. 6). This is the first of four east-west trending left-lateral fault zones we will cross as we proceed across Fort Irwin. The Coyote Lake fault juxtaposes Tertiary gravels to the south against Mesozoic granitic rocks to the north.

24.6 5.0 Entrance to Fort Irwin National Training Center. We will meet our military escort here. If you miss this escort, you can not cross Fort Irwin.

27.2 2.6 Turn west at first signal light.. Continue turning and essentially make a U turn onto a dirt frontage road parallel to and just west of Fort Irwin Rd.

28.1 0.9 Proposed Fort Irwin stables. Park and walk to first **STOP** for a lecture and tour of the recently discovered Richard W archaeological site (Fig. 7). Return to signal light on Fort Irwin Rd and turn north toward Fort Irwin. Continue through base complex.

31.4 3.2 Cross Bicycle Lake fault zone. This fault trends generally east-west, but in this area it trends northwesterly to meet west-trending faults on the north side of Tiefort Mountain. fault separation around this bend opened up a depression filled by Bicycle Lake.

42.6 11.2 Entering Granite Mountains. This range consists of Mesozoic granitic rocks characterized by spheroidal boulders. The rounded boulders are caused by erosion of exfoliated granitic rocks and indicate the pre-Miocene erosional surface.

47.4 4.8 **STOP** for discussion of east-west trending faults in Fort Irwin (Fig. 8; Miller and Yount, this volume; Pavlis and others, this volume). Return to vehicles and continue north.

50.9 3.5 Cemetery for several turn-of-the-century mining camps in the Avawatz Mining District (Vredenburgh, 1994).

54.4 3.5 Enter Avawatz Pass. **STOP** for discussion of mylonites and their relation to low-angle faulting in eastern Granite and western Avawatz Mountains (Fig. 9; Wrucke and Stevens, this volume, Pavlis and others, this volume).

56.0 1.6 Entering Cave Spring. **STOP** for discussion of archeological resources of Fort Irwin area (Allen, this volume). Between 1883 and 1887, Cave Spring was a water and rest stop for wagons loaded with borax from Death Valley (Vredenburgh, 1994). Return to vehicles and continue north for spectacular views of southern Death Valley.

58.0 2.0 Leaving Fort Irwin National Training Center. **STOP** for discussion of east-west trending fault zones in western Avawatz Mountains. Two of these fault zones are visible from this stop. The Arrastre Spring fault zone is mapped along base of the hills to east; the left-lateral Garlock fault zone is mapped along base of hills in the foreground (Troxel and Butler, this volume). Return to vehicles and continue north.

60.2 2.2 Turn north at fork and continue down Denning Spring Wash. Left fork goes to Denning Spring and west along the Garlock fault.

61.2 1.0 Mouth of Denning Spring Wash. Mule Spring fault zone along northern front of Avawatz Mountains places Jurassic Avawatz Mountains quartz monzodiorite over lower member of Miocene Military Canyon Formation. Folded strata of the Tertiary Celestite Hills in foreground to the north; granitic rocks of the Cretaceous Teutonia Batholith in the Owlshead Mountains in middle distance. **STOP** for discussion of slip rates along Garlock and Owl Lake fault zones (McGill, this volume). Return to vehicles and continue east and north along dirt road.

63.1 1.9 Enter Noble Hills. These low hills consist of tilted, folded, and faulted Tertiary fanglomerate and evaporite deposits caught between strands of the right-lateral Death Valley fault zone (Troxel and Butler, this volume).

Fig 8. View from the north of the Fort Irwin fault zone. Fault is located at the base of the low hills in the middle distance. Note the linear mountain front suggesting a young fault zone with the south side up. *Terry Pavlis photo.*

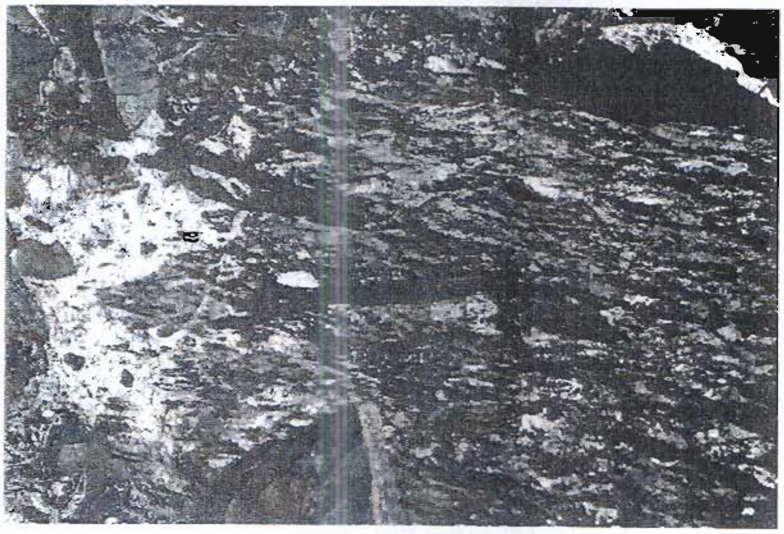

Fig 9. S-C mylonites developed in Mesozoic metavolcanic rocks of western Avawatz Mountains. This outcrop shows one of the sinistral strike-slip shear zones from the area that apparently developed at the same time as thrust-related mylonites that are generally correlated with the Layton Well thrust in the Slate Range .*Terry Pavlis photo*

66.5 3.4 Exit Noble Hills. Deformed gypsum beds on left are near the northern strand of the Death Valley fault zone.

66.8 0.3 Intersection of Denning Spring Wash Road and Harry Wade Road. TURN RIGHT onto Harry Wade Road.

67.6 0.6 Intersection of Harry Wade Road and Pipeline Wash Road. **STOP** for discussion of intersection of the Garlock, Mule Spring, and Death Valley fault zones (Fig 10; Troxel and Butler, this volume). Return to vehicles and continue east on Harry Wade Road. Saratoga Hills to the north consist of the Middle Proterozoic Pahrump Group intruded by 1.1 Ga diabase sills (kelly green rock along west flank of hills). White talc is product of contact metamorphism between diabase and dolomite in lower member of the Pahrump Group.

74.4 6.8 Intersection of Harry Wade Road and Highway 127. **STOP** for discussion of interaction of Mule Spring and Death Valley fault zones along the northern Avawatz Mountains south of the highway (Troxel and Butler, this volume). Return to vehicles and turn south onto Highway 127. Salt Spring Hills to north of Highway consist of Proterozoic to Cambrian sedimentary rocks intruded by Cretaceous

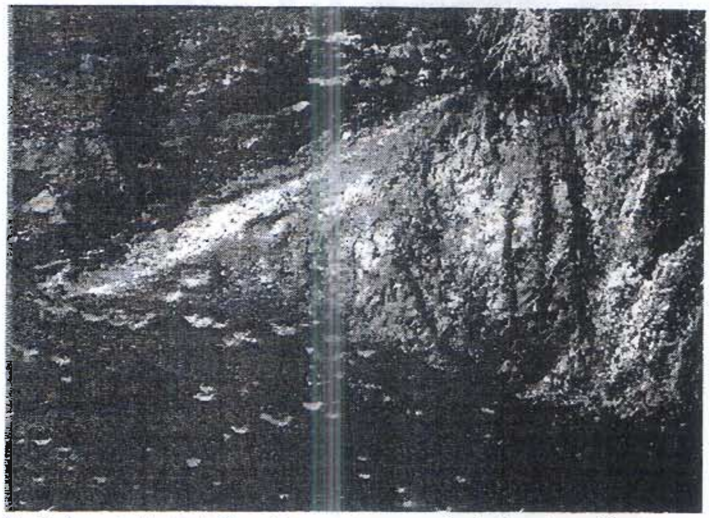

Fig 11. Jurassic Avawatz Mountains quartz monzodiorite thrust over Quaternary gravels near Old Mormon Spring, eastern Avawatz Mountains. *Jim Calzia photo.*.

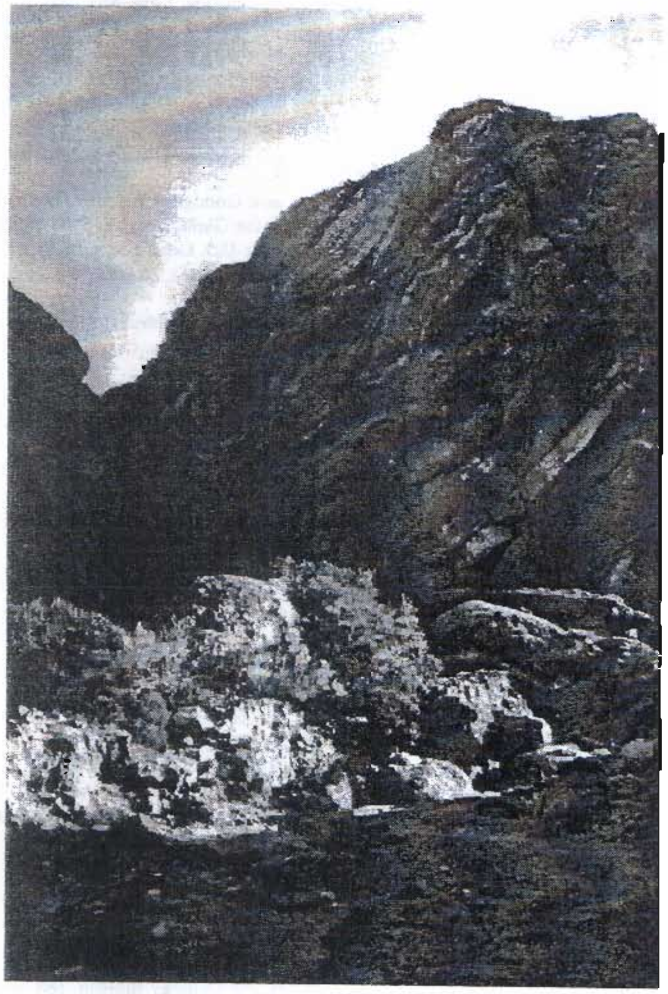

Fig. 10. Deformed gypsum beds in the latest Miocene Noble Hills Formation hear intersection of the Mule Spring and Death Valley fault zones, southern Death Valley. *Jim Calzia photo.*

granitic rocks of the Teutonia Batholith. These granitic rocks are overlain by 12.8 Ma basaltic andesite; the andesite is tectonically overlain by carbonate megabreccia sheets derived from the Spring Mountains. The megabreccias are overlain by 30 ka lacustrine sediments of Lake Dumont.

75.3 0.9 Entrance to BLM's Salt Creek Hills on north side of highway. Salt Spring Hills to north, Avawatz Mountains to south, and Kingston Wash to east of highway. Valjean Hills and Kingston Range to north of Kingston Wash; Silurian Hills to south of Kingston Wash.

84.9 9.6 Intersection of Highway 127 and Old Mormon Spring Road. Turn right on Old Mormon Spring Road and continue west into the Avawatz Mountains. Road is difficult to see from Highway and is approximately 150 yards north of Milepost 19 on Highway 127.

93.9 4.5 Old Mormon Spring. Park and walk north along base of Avawatz Mountains to last **STOP** (Fig 11; Troxel and Butler, this volume). Return to cars for discussion of geology and geophysics of Garlock fault based on what we have seen on this field trip. Return to Highway 127 and turn right to Baker and home.

End Day Two

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The History of Early Mining in the El Paso Mountains

Alan Hensher, *593 Collins Drive, Apt. 3, Merced, Cal. 95348*

PART I: THE PROMOTIONAL BOOM (1863-1866)

The discovery of rich and often massive deposits of silver ore during the late 1850s led to a flurry of strikes east of the Sierra Nevada a few years later. Several discoveries resulted in well-founded booms at Virginia City and Aurora. Other strikes, especially those in the Owens Valley and the Coso Range, tended to be more promotional than mineral.

The first excitement in the El Paso Mountains was one of these promotional flurries. There, near trails leading from Los Angeles to the Owens Valley, mouth-watering pockets of high-grade ore were found during the early 1860s: silver sulphides, argentiferous galena, even native silver. The Los Angeles *News*, Visalia *Delta*, *Mining & Scientific Press*, and *Alta California*, in particular, began spreading the word.

The scene of the strikes, however, was especially bleak, as C.W. Tappan, a persevering promoter, conceded in April, 1863: ". . . Not a bush or tree is in sight larger than the musquit [mesquite] by our camp. All is barren, the mountains appearing like cones of ashes, sharp and precipitous. Not a drop of water is anywhere to be found," except at six widely scattered springs.

Despite the bleakness, the region offered several advantages for prospecting. Two well-traveled trails skirted the hills. Abundant timber was available 25 miles away, in the Sierra Nevada. Mesquite grew very lush in some places – up to 8 feet high – and "greasewood" (creosote) was considered "an excellent fuel." Best of all, abundant water could be obtained at several waterholes: Mesquite Springs (near the later site of Goler); Grape Vine Springs, 8 miles from Mesquite; and "a fine spring" on the side of Laurel Hill, 15 miles from Mesquite.

For a brief, shining moment, the future of the district looked promising. Tappan made the first sale of mining property at his "office" (a tent) in late May, 1863. Several weeks later, a 1,100-pound lot of silver ore, from the Ophir Mine, yielded $1,150 a ton.

Yet the El Paso mines remained fairly isolated. Many miners weren't even sure of which county the district was in. The nearest important post office was at Los Angeles, 147 miles south. Newcomers would have to depend on the stamps, papers, and envelopes that they had brought with them; return mail (brought by a friend) was said to take one or two months. Fortunately, Russell Sackett, a former justice of the peace, began running his Slate Range Express through the El Paso Mountains in July, 1863; he carried mail, packages, and newspapers. Sackett offered to take passengers back to Los Angeles for $10 each and carry freight for 10¢ a pound.

The excitement might have ended that summer, for the pockets of ore were too small to justify further development. By late June, both the American and Mexican laborers had become "disgusted with the excessive labor under the burning sun, at climbing from five to eight miles to their work every morning, and the small pay." A month later, it was so hot – 108° F – that outdoor work had ceased; most men remained idle, awaiting the arrival of tools and more workers.

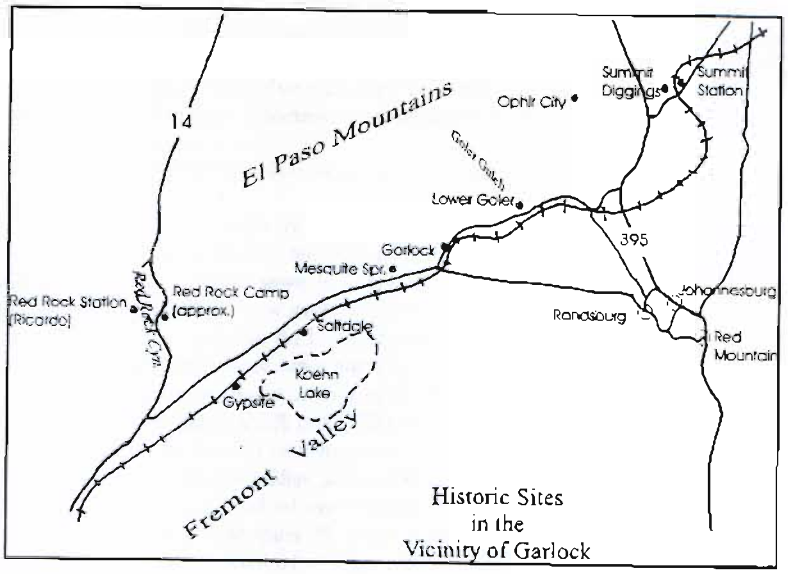
The harsh conditions – and a dubious future – didn't matter. The business of selling mining stocks, boosted by frequent news reports, was flourishing, especially in Los Angeles. The mine owners, "instead of gassing to get stock up, are working to get metal out, which strikes us as being sensible," the *Delta* commented in late July.

Apparently, Laurel Hill was the center of the limited amount of activity. There, near the Ophir Mine, stood Ophir City, consisting of six camps pitched in a one-acre square on a slope; above the camps flowed "a fine spring." The first – and perhaps the only – board house went up in August.

A few companies, meanwhile, continued to drive tunnels. The tunnel of the Yarbrough Company had reached 150 feet by mid-August, when a miner was seriously injured in another tunnel. He soon recovered.

Although Tappan would remain in the district for another year, the bubble soon burst. An intermittent war between intruding settlers and various Indian tribes had driven many miners out of eastern California. After wobbling for a while, the market for mining stocks collapsed in early 1864. Even then, prices remained high: 10¢ a pound for feed, and $100 a ton for hay in May.

A few assays still showed fabulously rich ore during the summer of 1864, and some tunnels and shafts were still being excavated. "We have had all the 'ups and downs' of prospectors – have been here in the sunshine and in the storm, have passed through all the vicissitudes of persons searching for the Eldorado in the form of Quartz Lodes," one correspondent wrote to the *Delta* in late July. "Sometimes we have felt *amazingly* rich, other times have almost come to the conclusion to surrender the ship, but fortunately stuck to it with the tenacity of a Shylock, and are now, we hope, or soon will be, in a condition to reap some of the fruits of our labors. . ."



Historic Sites in the Vicinity of Garlock

But the mines never revived. In July, 1866, another correspondent lamented, "there is not a single soul, at this time, in the district . . . yet the time will come, and that soon, when these hills will teem with busy miners."

PART II: THE MINING BOOM (1892-1897)

Red Rock Mining District

After the bust of the 1860s, the El Paso Mountains hardly seemed like an inviting region for mining. The heat, for one thing, could be insufferable. To the writer Mary Austin, Red Rock was "all desertness, affording no pasture and scarcely a rill of water." Walled in by huge colonnades of reddish sandstone and conglomerate, Red Rock

Canyon lay about 25 miles north of Mojave, on a well-traveled stage road to the Owens Valley.

Major deposits of gold dust and nuggets were found in Red Rock Canyon and Goler Gulch in late 1892 and early 1893. In Red Rock, the gold field lay in Iron Canyon, a northeast-running ravine that branched off the main canyon, several miles from its mouth. At first, the strikes aroused little interest.

But several months later, one of the worst depressions in American history swept in from the East. Banks failed; men and women lost their jobs; and the prices of goods, crops, and metals fell – except for gold.

Near the end of 1893, the gold fields suddenly seemed inviting. Reporters from Los Angeles, San Francisco, and Bakersfield rushed in. In early December, a Lancaster man saw "whole sacks of gold nuggets so rich and shining that they made his eyes water and grow dim. . . ." The sight of a large nugget put on display in Bakersfield was "enough to set the blood of any man tingling . . . ." As Christmas approached, Red Rock supported "a mining camp of no mean dimensions."

Despite the depression, however, Red Rock was no place for a poor man. "Enthusiastic young men who have no knowledge of the discomforts of travel and life on the desert should give the new camps a wide berth, unless they are prepared to undergo without a murmur all sorts of hardships," the Bakersfield *Daily Californian* warned in late December. The Los Angeles *Times*, meanwhile, warned about the many unscrupulous speculators who "all rush off and locate from three to forty claims apiece, and then sit around waiting to sell to some greenhorn."

Enough snow and rain fell in January, 1894, to nearly halt mining. Two miners used a trickle from melting snow to operate a rocker. But operations at Bonanza Gulch shut down for the winter. Several miners soon began stripping the gold-bearing gravel to bedrock and drying it in the sun.

The district did manage to generate some publicity when it sent 2,840 pounds of auriferous gravel to a fair in the East. The Black & Sullivan firm furnished and sacked the gravel. The owner of the Mojave-Keeler stage line hauled the gravel to Mojave without charge. And a manufacturer of dry-washers donated one of his devices.

Several camps grew up in the district, but the main settlement was **Red Rock camp**, the site of an early stage station, where Red Rock and Iron canyons joined. Founded in late 1893, Red Rock consisted of 20 tents scattered along both sides of the main gulch in December, when about 35 men were living there. Two stores, selling supplies at "very reasonable prices," were doing business there by late January, 1894. Then the initial excitement quieted down. By early March, Red Rock contained a store, a saloon, and from eight to 10 tents.

Nearly two miles to the north, at a well once used by a freighting company, stood **Miller's (Red Rock) Station**. A store and saloon were in business there in March, 1894.

At the **Black & Sullivan camp**, the mine owners, in early January, 1894, were preparing to build a barn for their stock and sell hay hauled from Tehachapi. They also generously supplied drinking water. By early March, when the mine operators had 10 men at work, the camp embraced a store and saloon and at least 15 tents.

Two miles northeast of Miller's station, meanwhile, **Bonanza Gulch** contained half a dozen tents. A Los Angeles firm bought up many of the claims that spring and brought in a pumping plant, a steam engine, and pipe to sluice or hydraulic-mine the placers, but the venture soon failed.

About 12 miles northeast of Red Rock stood **Black Hills**, a dry camp. Black Hills flourished after the other places declined. By late May, 1894, Black Hills was considered "the liveliest camp of all. . . . There is more real mining life and excitement at this camp than at all of the others put together . . .", one correspondent boasted. About 10 or 12 men were mining there in June, getting a "fair return."

It could hardly be surprising that the camps were fading away. As early as February, 1894, the miners had voted down a proposal to prohibit a person from staking out an indefinite number of claims. ". . . Times were exciting and talk ran high, but there was no blood shed." The output of gold soon started to slip, and traffic began to fall. By late May, the first excitement was waning, a correspondent for the *Californian* observed, "and the eager ones who rushed in with a hurry and ran all over the country, have most of them rushed out again, leaving behind as their only remembrance a liberal assortment of corner posts, stone monuments and location notices. . . ." Only 40 men remained by late June.

Even so, enough people remained in the El Paso Mountains to induce the Kern County supervisors to form voting precincts at Red Rock and Goler in early September.

The discovery of gold in the Rand Mountains set off small revivals throughout the El Paso Mountains in 1896. When Thomas Jaggers, a Denver capitalist, found a nugget worth over $500 in June, he rushed into Mojave and put it on display in a drugstore. The "wildest excitement prevailed. The nugget . . . caused staid man to lose their heads. Business men closed their places, hitched up their teams and left for the mines, and by 1 o'clock not a horse or vehicle was to be had. At one time it was feared that the employees of the Southern Pacific and Santa Fe would catch the infection and abandon their posts."

Goler Mining District

Goler, the second important district in the mountains, experienced a rocky start. Several miners were working there in October, 1893, but few claims were paying; some men were making just enough to buy "only grub," one correspondent complained. ". . . It might be well to stay away entirely."

But the prospects soon changed. J.S. Reed, the namesake of a rich gully, found a 56-ounce nugget in Goler Gulch in early November. Fifty miners were working there by early December, and many others were on the way, by burro and team. The prices were moderate – 35¢ for meals – as were wages: $2.50 a day and board for good miners. By the end of the month, Reed and Benson gulches alone had yielded up $30,000 in dust and nuggets.

But Goler no longer had any room for newcomers. "The whole country for miles around is covered with location notices," a correspondent for the San Francisco *Chronicle* lamented in late December. "Corner and center monuments and the like are as plenty as greasewood bushes, or almost so. Every one who has been upon the ground, after recording it for himself, seems to have built a lot of monuments in memory of his wife's relations, and in this wise the whole country has been gobbled up . . ." At a meeting, the miners voted to limit prospectors to one claim each.

Some rain and snow in early January, 1894, dampened the excitement. A few enterprising miners dug beneath the damp topsoil to continue dry-washing. The recorder of the district even sank a 100-foot shaft. A Los Angeles man took out $1,000 in a single week, and several rich nuggets were found in February.

Most of the travel to the mines was over by late May, but small-scale mining went on. At **lower Goler camp**, near the mouth of Goler Canyon, an estimated two dozen or more Mexican and American miners were running dry-washers and rockers. Lower Goler was a section of deep gravel, in which several shafts, one of them 174 feet deep, had been sunk. Two miles north, over a rough and rocky road, stood **upper Goler camp**, near Benson and Reed gulches, the richest sections. A varying number of miners, perhaps 25, were working there in late May or early June. Several months later, in early September, the Kern County supervisors created voting precincts at Red Rock and Goler. The Goler precinct apparent included Summit and the newly organized Rand Mining District.

The feverish activity in the Rand district a few years later renewed interest in Goler. Only 15 men were working the placers in late July, 1896. They were doing fairly well – except for a miner who had fallen down a 40-foot shaft; although severely bruised, he suffered no broken bones or internal injuries. Other miners were more fortunate, finding a nugget worth $654 in August.

Another mineral, meanwhile, was attracting attention: water. To prepare for a mill at Randsburg, the owners of the Yellow Aster Mine surveyed a pipeline from Goler to the mine during the early summer. After a crew began sinking a shaft in early August, a family from Garlock opened a boardinghouse at the site. A derrick for a hoist was also set up, but it collapsed while a worker was descending the shaft in a bucket; the worker was jerked 30 feet into the air and landed with near-fatal force. The machinery for a pumping plant arrived in late August; by then, water was entering the shaft faster than it could be bailed dry.

Fig. 1. Summit Camp, early 1893(?). Harold W. Fairbanks, State Mineralogist, Rep 12 (1894). *Larry Vredenburgh collection.*

Summit Mining District

Summit was the third important district in the El Paso Mountains. The placers there were apparently discovered in early or mid-1893, near a station owned by the San Bernardino Borax Mining Company. At the station, teams, travelers, and miners could obtain free water.

Summit turned out to be a somewhat poorer district than Red Rock or Goler, but it still remained attractive. By December, while some miners were making $20 a day, two men from San Bernardino had taken out $17,000 in gold in five months. Several months later, in February, 1894, T.R. Davis, the owner of a Tehachapi hotel, opened a store at the mines. And by April, 200 "permanent residents" were operating 60 dry-washers there.

The most persistent group of miners seemed to be the Van Slyke brothers. After cleaning up $1,000 in five days, they cleaned up again: by jumping the Trix Mine. In April, a jury found one brother and four others guilty of claim-jumping. The jury recommended mercy, however, and each of them was sentenced to a $10 fine or 10 days in jail.

The excitement nearly died out, but the ex-jumpers found a bonanza in late June; the camp enjoyed "a new lease on life." This boom, too, failed to last.

Still, Summit was a legitimate district. Nuggets worth from $3 to $20 were found during the late spring and summer of 1896. Shown at Garlock, the nuggets were said to be as plentiful as marbles. Again, Summit was destined to "become a lively camp." About 15 or 20 miners were working there in early August.

Summit experienced its largest – and last – boom in May, 1897, when 300 men were dry-washing, "all making wages." Also being worked were several lodes, one of them yielding $25 a ton. A mill manufacturer from San Francisco was sinking a well, preparing to put up a 5-stamp mill (probably never built).

Kane (Koehn) Springs

For travelers and miners in the El Paso Mountains, especially the Goler district, Kane Springs served as an important supply center. The nearest town was Mojave, 26 miles away.

Charles Koehn, a native of Germany, was keeping a ranch at the waterhole, "a somewhat brackish spring," where a post office, named Koehn, was established in September, 1893. Two stores were doing business at the spring by early December.

Koehn' s ranch, however, resembled a one-man town. Besides running the post office and his ranch, Koehn kept a bar, delivered letters in the mines (for 25¢ each), provided free water to travelers, and kept a store. He sold hay (for 1½¢ a pound), grain ($1.50 a sack), meat (8¢ to 11¢ a pound), and other provisions. He also made the rounds of

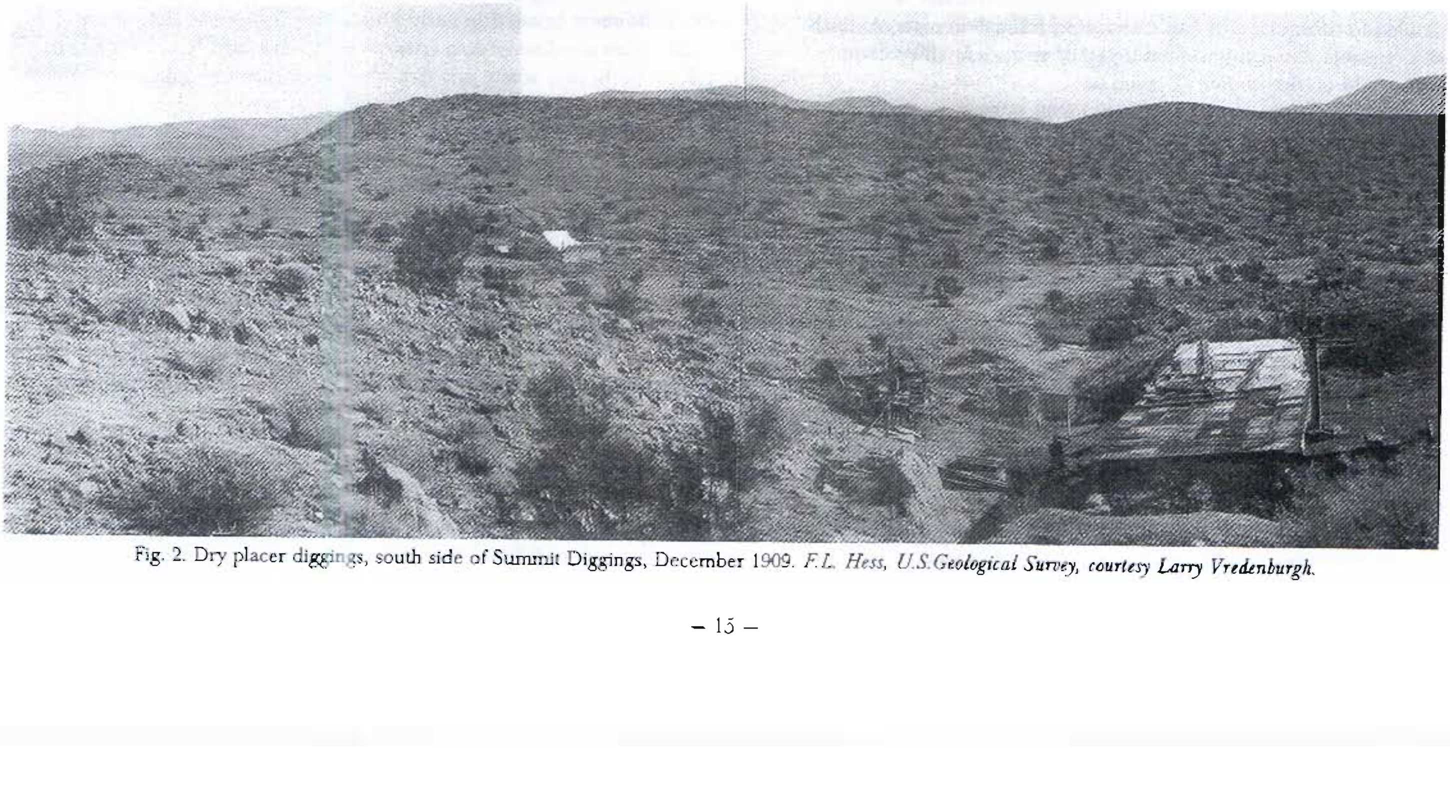


Fig. 2. Dry placer diggings, south side of Summit Diggings, December 1909. *F.L. Hess, U.S.Geological Survey, courtesy Larry Vredenburgh.*

the camps, selling supplies out of a wagon.

Koehn was considered "very much of an accommodation in general." His prices were "reasonable." And, wrote a correspondent for the Los Angeles *Herald*, Koehn "has not the gall so common among settlers on the desert to charge travelers for water."

Despite the decline of the placers, **Koehn Springs**, as it came to be called, remained a "favorite halting place" during the rush to the Rand Mining District. Austin Young, a member of the same Masonic lodge (in Tehachapi) as Eugene Garlock, was managing Koehn's various enterprises in May, 1896. The store, post office, and bar were housed in a stone building; the walls, which were several feet thick, were "warranted to keep out the desert heat." Nearby, a large meadow contained a pond, wells, and springs. ". . . It is a veritable oasis in the desert . . ."

Meanwhile, to process ore from the Rand district, Koehn and a partner, O.B. Stanton, sank a well and had a 5-stamp mill built. The mill started up in late June; a clean-up a month later yielded $1,000.

As an oasis, the ranch thrived. One correspondent came across 72 mules and horses one day in February, 1897. But the mill had shut down. It had proved no more efficient than the mills at nearby Garlock. Two veteran mill operators from the Slate Range soon bought 2,000 tons of tailings from Koehn and Stanton, installed a small cyanide plant, and, in July, began turning out bullion: $2,000 worth in one shipment made in November.

By then, mine operators in the Rand district were sinking their own wells and putting up their own mills. Railroad service to Johannesburg began a few months later. The post office at Koehn Springs was discontinued in January, 1899, although Koehn maintained his ranch for 30 more years.

The stations of the San Bernardino Borax Mining Company

Kane Springs and Cow Wells were not the only oases on the way to the discoveries in the Goler, Summit, and Rand mining districts. Several stations used by the San Bernardino Borax Mining Company also served as important supply points.

The brothers John and Dennis Searles, the founders of the company, began extracting borax from the dry bed of Borax (Searles) Lake in 1872. The loads were huge – up to 30,000 pounds in 1894 – and the teams were long: 20 mules.

By 1894, when E.M. Skillings had replaced Dennis Searles as a partner, the company was maintaining five stations: **Forks**, 6 miles from Mojave; **Mesquite**, 23 miles from Forks; **Summit**, 21 miles farther; **Salt Canyon**, 17 miles; and the home station, 9 miles, at Borax Lake. Forks, Summit, and the home stations were supplied with good water piped from springs in the nearby mountains; water had to be hauled to Mesquite and Salt Canyon in 500-gallon carts. At each station, a stable, "so firmly bolted together as to defy all ordinary blasts," could accommodate 40 animals.

In fact, the stations were "conducted with military precision and order," a correspondent for the *Californian* explained. ". . . It is no slight task to set out across the desert with one team of twenty mules, one driver, one assistant called a 'swamper' and 15 tons of freight . . ."

Apparently, Searles and Skillings had abandoned Mesquite station by May, 1896. Nearby Koehn Springs was well prepared to handle even large teams. A Los Angeles firm was running a 5-stamp mill at Mesquite in early June.

Besides the Rand district, major mines were being developed in the Panamint Range, where a supply center named Ballarat had been laid out. "There is a constant stream of travel through this place, headed for the mining camps to the eastward," the Mojave correspondent for the *Californian* reported had reported in May. "Every train brings passengers bound thither, while teams come in every day, all headed in the same direction. . . ." So much traffic was heading to the camps that Kern County officials surveyed an improved road to the Rand district in December; the survey ran from Warren's station, near Mojave, followed the borax road as far as Garlock, and then made a jog up the Rand Mountains to Randsburg. The road had been so well maintained that the county estimated the cost of improvements at only a few hundred dollars, to the delight of one correspondent in January, 1897. ". . . This is certainly a cheap investment for the county and a very necessary one."

Summit station, near the Summit placer mines, was probably the best developed. Excellent water was being piped 4½ miles from the mountains. Growing near the stables were a lush vineyard with delicious grapes; an orchard with healthy apricot, peach, pear, plum, and fig trees; and a thriving garden protected by a rabbit-proof fence.

John Searles and Skillings started up their borax works again in February, 1894, when two Lancaster men opened a store at Mesquite Springs. A hotel owner from Tehachapi, meanwhile, opened a store in the Summit district.

Apparently, Searles and Skillings reorganized their string of stations. The first station, Forks, had come to be known as **Six-Mile House** by May, 1896, when it comprised a barn and water tank. A correspondent for the *Californian* praised Searles "for the vast amount of money spent by him in building the road, erecting stations and piping water from the mountains. . ."

Sixteen-Mile House was a new station for the borax company. ". . . Here 'refreshment for man and beast' can be obtained, besides all manner of what may be called 'spiritual consolation.'" Cinco, an aqueduct camp and highway stop, was built near Sixteen-Mile House; both sites are now only memories.

Law and order

During the first true mining boom in the El Paso Mountains, in 1893 and 1894, the forces of law were fairly remote. Originally, the court district (or "township") centered in Tehachapi covered the desert districts of Kern County. Rural townships then were entitled to two justices of the peace and two constables. Their pay came out of fees or fines in civil and criminal cases (besides a certain prestige derived from the positions).

This type of thinly-spread judicial organization seemed adequate at first. About the only criminal matter then was the claim-jumping case of Van Slyke and four other defendants in April, 1894. Visiting several months later, a mining engineer called the diggings "the most peaceable ever known; none of the usual adjuncts of mining camps, without which some people cannot imagine them, are to be found; no saloons (with the exception of Red Rock), no gambling, no deaths by violence, no lawless element and no Chinamen – as peaceable a community and as hard working a lot of men as can be found anywhere." During the past three years of mining, a correspondent for the *Californian* boasted in early June, 1896, not one violent death, either in an accident or in a crime, had taken place. ". . . This record cannot be beaten where any like number of men are gathered together under similar conditions."

This near-Edenic state of affairs began to change a few weeks later, when "a considerable fracas" broke out at Charles Koehn's store and bar. He went to Bakersfield to swear out a warrant for battery.

Trouble was also beginning to plague Garlock. The residents there (and at Randsburg) tended to band together to keep order, for "the proverbial tough character so frequent in early-day mining camps when he appears in this vicinity has to behave himself or leave. . . ." According to local lore, early in the history of Garlock, a ruffian who had beaten up "old man Harkins in a shameful manner without provocation was promptly waited upon and told to take his departure – which he did in a hurry," one resident reported.

When a bartender in Garlock hammered one bald man over the head with an ice pick in August, 1896, a crude type of law enforcement followed. The owner of the saloon fired the bartender at once, and then one of the "mill boys" gave him a thrashing. That night, the ex-barkeeper "walked to Mojave a prettily used up man."

By then, serious violence – sometimes a killing a week – was breaking out at Randsburg. To keep order in the desert, the county supervisors created a court district, Township 10, centered at Mojave.

But Mojave was too far away. In response to several petitions, the board of supervisors created Township 11, serving the El Paso Mountains and the Rand district, in December, 1896. The board also voted to have a jail built at Randsburg. Justice courts were organized at Randsburg and Garlock in January, 1897.

As the population of the township increased (to an estimated 800 by early May), so did the crime. At the Summit mines in early April, two prospectors, "accompanied by a jug of whisky," started quarreling. One of them beat his partner severely and then went on to Randsburg, where he told of the fight. Then a traveler reported the victim as dead. The prospector was jailed and charged with murder. A party went out to bring in the body, "which they soon met covered with gore and all bloody red, walking along the road toward camp. . ." The murder charge was reduced to battery.

Worse was to come several weeks later in the Goler district, where three brothers – Ben, John, and William Higgins – were working the placers. Apparently, Ben had long held a grudge against John. When they began feuding one morning, William put an end to the fight. But that afternoon, Ben came up to their claim, threw John a rifle, told him to defend himself, and immediately shot him in the neck. William rushed to John's side, then heard a report and saw Ben fall backward.

The county coroner rushed from Bakersfield and held an inquest. But by the time William arrived in Mojave, the bodies had started to decompose. And without coffins, the railroad refused to ship the bodies. Reluctantly, William had his brothers buried in Mojave "just as they were found," a correspondent lamented. "It was a sad and horrible sight and the living brother has the sympathy of all the people."

At Garlock in early May, a man named W. Bull, who had rented a horse from D.B. Newell in March but failed to return it, was brought before the justice of the peace on a charge of grand larceny. The judge held a preliminary hearing and sent Bull's case to the superior court, setting his bail at $250.

Adults weren't the only malefactors. John Hawthorne, 16, who was mining with his father at Goler, entered a cabin a few weeks later and took $85 in cash and gold dust from the pants pockets of a miner. Apparently, the father found the loot hidden under the grub box of their tent, and the boy confessed. He was taken before the judge in Garlock.

Epilogue

After mining declined during the 1890s, Rudolf Hagen, a prospector, bought the area around Red Rock station, where he kept a store, stable, feed yard, and perhaps a saloon. A post office, named after Hagen's son Ricardo, was established there in January, 1898, although the voting precinct was abolished in 1902. In 1907, Hagen tried to build a water project to hydraulic-mine the old placers and irrigate the Fremont Valley. Hagen probably failed to get financing, but Los Angeles soon built a branch railroad up the canyon to haul in materials for the construction of its aqueduct to the Owens Valley. After the completion of the aqueduct, the railroad was torn up and the post office was discontinued (December, 1917).

But some of Hagen's ideas turned out to be sound. He put in a store, cafe, and gasoline pumps near the site of the mining camp (next to the present highway) and developed an extensive farm below the colonnaded cliffs of today's state park. The highway was paved about 1930, when Hagen retired in Bakersfield.

Like Hagen, Charles Koehn remained in the area. He energetically operated a variety of small mines and ran cattle in the El Paso Mountains. Koehn's long career ended after he was convicted of attempted murder in September, 1928, and sentenced to prison.

The nearby Goler placers, meanwhile, were being worked on and off with some success. During the early 1930s, the Goler Canyon Mining Company and other concerns put in wells, pumps, and gravel-washing machinery. Several hundred men, women, and children lived at several camps there; a building was even put up for a school and church. But no permanent settlement developed.

Like Red Rock, Summit station became an important supply stop. During the rush to Ballarat, the Teagles, a family of Randsburg merchants, bought the property and renamed it **Garden Station**, where they ran a store and telephone station. A post office, named **Searles**, was established in August, 1898. The Southern Pacific Railroad built its Owens Valley branch past the place about 1908. Crews surveying a route for the Trona Railway later made Garden Station their base camp. The construction of the railroad, starting in late 1913, apparently made the station obsolete, for the post office was discontinued in July, 1914.

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Acknowledgments

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Gypsite: A Humble Product from a Humble Camp

Alan Hensher, *593 Collins Drive, Apt. 3, Merced, Cal. 95348*

Several products of the desert were too humble to attract much attention. Gypsite, a mixture of gypsum and clay, was one of those products. The purest grades could be turned into plaster, and lower grades were valued by farmers as a soil amendment.

Charles Koehn, who had figured prominently in the mining boom in the El Paso Mountains during the 1890s, found an unusually large and pure deposit of gypsite on the dry bed of Koehn Lake in late 1909. The discovery was made while the Southern Pacific Railroad was building a branch from Mojave to the Owens Valley.

Even though gypsite was selling for only $10 a ton, the California Crown Plaster & Gypsite Company, of Los Angeles, began leasing Koehn's claim and put up buildings and a calcining (roasting) plant near Kane (Cane) Springs in early 1911. Twenty men worked at the mill. A post office served the milling camp from June, 1911, to March, 1912.

The demand nearly overwhelmed the plant. Running the mill day and night during the summer of 1912, a mere 12 men produced 30 tons of plaster a day.

Reorganized as the California Gypsum Hollow Tile Company during mid-1913, Crown Plaster enlarged and modernized the operation. The company built a 3-mile narrow-gauge railroad on the lake bed, where tests had revealed a 14-foot deposit layer of very pure gypsite. At the mill, the gypsite was dried out in three huge oil-fired kettles, mixed with fiber and retarder to give it firmness, and pulverized to make plaster of "very good quality." Expecting to soon double its force, the company also put up a depot, a hotel, houses, and a post office (which never reopened).

For reasons that are not clear, the operation failed. Koehn, who held a mortgage on the mill, took it over in early 1915 and threatened to ship out gypsite himself if the plant were not redeemed. But the company never returned.

The deposit was worked from time to time by various companies. In late 1926, for example, the Consolidated Clay Company, of San Francisco, installed a huge dryer on Koehn's property; it cost $20,000. Consolidated planned to put up a mill, to cost $40,000, to grind the abundant clay found there. The company planned to grind 80 tons a day for shipment to Los Angeles, where the clay would be sold to oil refineries. The mill would also process gypsite. The operation employed 23 men, which was expected to increase to 50 when a second mill was built. As ambitious as this project was, it probably went no further.

In any case, Koehn soon lost control of his claim. He was accused of trying to kill a Bakersfield judge with a bomb and found guilty in September, 1928. George Abel, meanwhile, bought his gypsite property. Making frequent sales trips to orchard owners – one order totaled 1,500 tons – Abel successfully ran a small operation for several years. After Abel died during the early 1930s, other operators continued small-scale mining well into the 1950s.

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The History of Saltdale

Alan Hensher, *593 Collins Drive, Apt. 3, Merced, Cal. 95348*, **Larry M. Vredenburgh,** *Bureau of Land Management, 3801 Pegasus Dr., Bakersfield, CA 93308*, and **Gregg Wilkerson,** *Bureau of Land Management, 3801 Pegasus Dr., Bakersfield, CA 93308*

Introduction

Koehn Lake is a geologic anomaly: a "moist" playa, in which shallow ground water rises to the surface by capillary action, carrying with it salt, which is deposited in the center of the desert playa lake. This readily available source of salt, close to transportation and to major markets in Los Angeles and the San Joaquin Valley, was also the site of one of the longest running deceptions in the Mojave Desert.

The Saline Placer Act of January 31, 1901 placed a limit of one mining claim for saline minerals per locator. To thwart the intent of this Act, the salt producers on Koehn Lake employed the services of numerous individuals or "dummy locators" to locate mining claims, thereby acquiring large blocks of mining claims. It began in the period between 1909 and 1913 when sixty saline placer claims were located and leased to Thomas Thorkildsen and Thomas H. Rosenberger for a period of forty years. By having individuals locate these claims, and subsequently lease them, Thorkildsen was able to tie up sufficient ground to begin salt production. Thorkildsen during this period began developing borate deposits near Lang in Los Angeles County, at Stauffer in Ventura County, and near Daggett, giving William "Borax" Smith a fright that his borax monopoly was less than secure.

The claim staking by Thorkildsen and Rosenberger did not go unchallenged by feisty Charles Koehn, founding father of the settlement of Kane (or Koehn) Springs. Koehn, who had previously located claims on the lake bed, challenged the claim jumpers in January, 1912, in a "lively" gun battle on the dry lake. Swift justice was meeted out in Randsburg, the Randsburg *Miner* reporting on Feburary 10, 1912, that in the case of *People vs. T. H. Rosenberger and ten others*, the defendants were found guilty of forcible entry and detainer, and Rosenberger was fined $50. Nearly a year later, a happy ending to the dispute was reported by the *Miner*. Koehn sold his claims to Thorkildsen, who in turn sold them to the Diamond Salt Company of Los Angeles. More than likely the Diamond Salt Company actually leased the property.

While the chain of ownership for the *claims* located on Koehn Lake is detailed in government investigations in between 1945 and 1971, the corporate relationship in the early years is confused. The Consolidated Salt Company was incorporated February 11, 1913. It forfeited its charter November, 1913, but the company's interests were not transferred until 1933. In addition, the Randsburg *Miner* continued to report as late as 1915 that the Diamond Salt Company was actively working the deposit.

Early Operations

The Consolidated Salt Company constructed a crushing and screening plant and laid a baby-gauge railroad track onto the playa, from where a gasoline-powered locomotive hauled the salt to the crusher. Consolidated began shipping in 1914 — 240 tons or more a week by October. The output that year totaled 20,000 tons. In January, 1915, the company was shipping about twelve cars of salt weekly.

Business boomed. Employing 30 men, Consolidated was turning out about 720 tons a week by June, 1915. The crew was increased to 65 in April, 1916, while a 4-story mill was under construction. A long-awaited post office was finally established that September. But a chronic problem — the inability of the Southern Pacific to supply enough cars — was delaying shipments by five months.

Consolidated ran an extensive operation. Except during rainy winters, the company pumped well water onto the lake floor. The brine thus produced was then pumped through a 1± mile ditch into several pond-like "vats" — the largest covering 43 acres — where the brine was allowed to evaporate. After two or three months, a 6-inch layer of very pure salt would form. At "harvest" time, a circular saw mounted on a portable platform cut the layer into cakes. The cakes were then cleaned by hand, loaded into small cars running on a temporary track, and hauled by a gasoline-fueled locomotive to the mill. There, the cakes were ground, sized in screens, sacked, and shipped to Los Angeles.

Newcomers to Koehn Lake

In activities which harkened back to the claim staking by Thorkildsen and Rosenberger's crews in 1912, T. Y. DeFoor and Philo H. Crisp located a block of one-hundred eleven mining claims on Kohen Lake between 1916 and 1918. In order to locate this large block of claims "dummy-locators" were paid 5,000 shares of stock with a par value of $1 per share for signing their names as locators. After DeFoor located a claim on the ground, he gave Crisp, an old time prospector in the Garlock area that knew all the section corners in the vicinity, the location notice and a deed with the name of the grantee blank. At the same time, another associate, Paul Greenmore, a resident of Bakersfield, was rounding up the "dummy-locators" described as "just a bunch of widows none of whom could write a check for $50." Greenmore received $2.50 for each signed location notice. Each of the 111 claims was 20 acres, and each claim was "located" by a different individual. These claims were then deeded to the Fremont Salt Company.

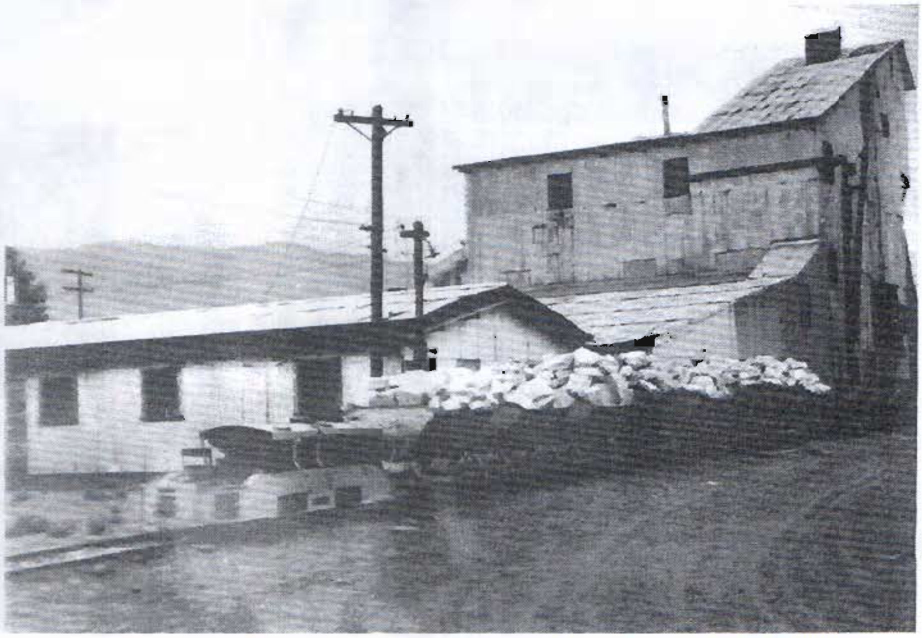

Fig. 1. Mill at Saltdale, February 1955. *William Ver Planck, Calif. Division of Mines, courtesy Larry Bredenburgh.*

With location of these claims, a second producer, the Fremont Salt Company, incorporated December 7, 1916 and built a plant on the east side of the playa in 1917. In 1919, when the Southern Sierras Power Company brought in electricity, the companies produced altogether 17,000 tons.

By then, the operations were becoming somewhat erratic. Enough families were living at the plants to induce the Kern County supervisors to organize the Saltdale School District in February, 1920. But Consolidated was employing only six men, and few pupils showed up at school; in fact, no schoolhouse was built. Even so, the companies managed to produce 22,000 tons. The camps probably remained small, for the school district was absorbed by Garlock's in August, 1921. The output of salt declined somewhat, to about 18,900 tons in 1923.

The year 1922 also saw the transfer of claims held by Thomas Thorkildsen and Thomas H. Rosenberger to the Consolidated Salt Company.

Although Consolidated's operation was being kept in good condition — "as neat as a lady's kitchen" — only six men were working in the mill in July, 1924, besides a handful running the pumping plant and train. A shortage of water and power was holding down production to about 6 to 10 tons a day. Apparently, the school was moved from Garlock to the plant about then. Alas, the building was little more than a shack, and the institution was one of the poorest in the county, suffering from a high rate of absences. Although H.C. Topp, "the rustling superintendent" for Consolidated, called 1925 the best season so far, the companies finished the year with 6,900 tons, their lowest total output.

Slowly, the operations began to recover. The total output reached nearly 15,000 tons during the 1927 season. Even so, the companies were facing another dry year.

Coming onto the scene was Henry Fenton, the owner of the Western Salt Company, based in San Diego. Western Salt had acquired part ownership of the Long Beach Salt Company, which in turn bought out Fremont on November 5, 1927. The Long Beach Salt Company had operated salt ponds and a salt works in the marshes opposite Terminal Island between Wilmington and Long Beach. The salt operations were gradually displaced after discovery of oil.

Long Beach Salt dismantled the Fremont Company's plant and concentrated operations at Consolidated's plant. By then, the camp's "business district" probably included no more than a company store, the post office, the school, and a service station along the Cantil-Randsburg road.

The school, too, began to enjoy better days. Under the guidance of its teacher, Mrs. Ruby Rogers, and H.C. Topp, who also served as the district's clerk, the school began to set records for its high attendance rate. The building was repaired, repainted, and enlarged in late 1927, enough to make it "very attractive and well lighted."

Like many camps then, Saltdale was composed of two groups: managers, skilled workers, and their families, who tended to be Anglo Protestants, and common laborers and their families, who were usually Latino Catholics. The Protestants had their own group, the Ladies' Aid Society, which held weekly meetings, often at Cantil. For the Catholics, many of whom worked at other camps, the center of religious life was St. Mary's Church, in Randsburg.

It was the job of many schools, including Saltdale's, to bring the groups together. To carry out the work of "Americanization," Latino children were encouraged to participate in play activities that demanded "the use of the English language and the finer points of good sportsmanship and cooperation." At a Christmas party held in 1929, the pupils put on a well-received play, after which cake and sandwiches "and some delicious enchiladas made by our Spanish American ladies" were served. Another teacher, Mrs. Caroline Larson, began teaching a night course in English ("Americanization") for Latinos and a Spanish course for Anglos during the fall of 1930. She "deserves a great deal of credit," one correspondent commented.

Although the work at the mill was hot and hard, the residents could enjoy an abundance of humble pastimes during the late 1920s and early 1930s. The Ladies' Aid Society often held parties, bazaars, and fund-raising events. The mill workers put on dances that attracted people from all over Fremont Valley. The schools at Cantil and Saltdale together went on picnics, held Christmas parties, and put on field days. During August, 1928, Mrs. A. Soto invited several friends to "a splendid enchilada dinner" in honor of her husband's birthday; a week later, two Latino youths spent Labor Day "swimming in the 20 per cent brine-solution ditch, and claimed that they liked it. Felipe Hernandez made an eager second for the impromptu swimming party."

Fig. 2. Harvesting salt at Saltdale, April 1953. *William Ver Planck, Calif. Division of Mines and Geology, courtesy Larry Vredenburgh.*

Despite these simple pleasures, Saltdale could suffer from its isolation. Crime was easy to commit since the closest justice of the peace, constable, and jail was in Randsburg, 16 miles away. The company store was robbed of several games one night in March, 1928. Topp "feels sure it was strangers and we feel sure no one around here would commit a felony," one correspondent explained. And rather than go to the nearest hospital, at Red Mountain, some mothers gave birth at home. But this practice could lead to complications: during the same week that a boy was born to one family, in December, 1931, the infant daughter of another family died.

It must have been difficult for Saltdale to weather the Depression. A proposal was made in September, 1931, to consolidate the school districts at Saltdale and Cantil. The construction of a modern campus, the paper in Randsburg predicted, was probably "the best improvement that could be suggested." (The merger had to wait 20 years.)

Even though depressed, Saltdale and other camps still had to be serviced. To handle the shipments of salt, gypsite, and pumice, the Southern Pacific built a modern loading platform at Saltdale in late 1931, and the county graded 10 miles of the Cantil-Randsburg road, which was now oiled. And to increase the flow of brine, Long Beach Salt blasted a 1.7-mile ditch in the mud of the lake. The company, in fact, enjoyed enough good years of production to keep its parent, Western Salt, prosperous through the Depression.

Two important transactions occurred in 1933. On June 3, all the leaseholds held by the Consolidated Salt were transferred to the Long Beach Salt Company. Then in July, 36 association placer claims were filed by the Long Beach Salt Company on Koehn Lake, allegedly for placer gold. The location of these claims continued the legacy of deceptively located claims on Koehn Lake, for with the passage of the Mineral Leasing Act of 1920, salt was no longer a mineral which could be acquired with mining claims.

Where the 1930s were prosperous, the 1940s were another matter. An increased amount of gypsum in the salt limited its sale to farms and factories. The rainfall, meanwhile, dwindled, finally drying up for a few years after January, 1947. Attempts to run the plant on salt shipped from San Diego turned out to be impractical. Only three workers remained in 1949. The post office closed in June, 1950. The school district was dissolved in July, 1951, the same year that Fenton died. The Saltdale operation, a family member recalled, was "one of the few salt ventures that did not support his good judgment."

The mill, however, was kept intact and modernized during the 1950s. It remained a highly mechanized, round-the-clock operation that required only a handful of workers.

The claims from which the salt operations at Saltdale had germinated were known to the United States. An extensive investigation was begun in 1945 by geologists and engineers with the General Land Office (case SF-62514), but apparently died with the creation of the Bureau of Land Management (BLM) in 1946. The case was reopened in 1956 when the president of the Long Beach Salt Company filed a protest with the agency when another company filed a Sodium Prospecting Permit Application under the provisions of the 1920 Mineral Leasing Act. By 1960 the renewed investigation died as well.

The beginning of the end started with a September 3, 1968 letter from D. Livengood, president of the West Coast Salt and Milling Company of Bakersfield. Livengood's letter to Secretary of Interior Udall, Congressman Robert Mathias, and the Bureau of Land Management shook the agency into activity. Livengood was steamed that Long Beach Salt Company was able to undercut his product, for which he paid royalties to the United States from operations on Searles Lake. Earlier investigations focused on the manner in which Consolidated Salt, and Fremont had acquired the claims. The complaint that was filed July 23, 1973, charged that there was insufficient gold on the 36 claims located in 1933 to constitute a valuable mineral discovery. Even though salt was a valuable mineral which had been produced since 1913, production had moved off of the pre-1920 (Mineral Leasing Act) claims to the claims filed in 1933. Incredibly, in a civil case decided July 9, 1972, the court determined that the terms of the Saline Placer Act of 1901 had not been violated in the location of the pre-1920 mining claims. However, the fate of the 36 claims from which Long Beach Salt Company was producing salt was decided by the United States Interior Board of Land Appeals (IBLA) on December 2, 1975, when they declared them null and void.

At the time of the IBLA's decision, only four workers remained — and none of them lived at Saltdale. The plant probably shut down soon afterward. Amid the rubble of buildings, the corrugated-iron shell of the mill still stood in May, 1980. The wind banged the doors eerily in the glow of the setting sun. A year and a half later, even this remnant of mining was gone.

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Mining History of Goldstone

Larry Vredenburgh, *Bureau of Land Management, 3801 Pegasus Drive, Bakersfield, CA 93308*

Gold was reportedly discovered at Goldstone as early as the 1880s. Subsequently there were two periods of activity, 1910-1916, and the 1930s.[1]

In May 1910, the *Pacific Miner* trumpeted: "Gold Stone is the name of a new camp, thirty-five miles north of Barstow, where some phenomenal ore has been found. Quite a stampede followed the discovery. Reports show that it is the richest locality in the Southwest."[2]

This "stampede" seems to have been touched off by the rich returns from the Drumm mine, which had reported that a ton of ore from the surface yielded $97 in gold and five ounces of silver at the Needles Smelter. In August, John Harper came into Barstow showing off some "picture rock" (rock typically shot through with native gold) from Goldstone which he said ran up to $180 per ton in gold. Others working properties at this time included Jack Halford, Joe Goodrich, C. E. Burkhart, W. H. Scott and R. M. Dillingham. George Drumm in November began sinking a 50 foot shaft, and soon Drumm and partner W. M. Clancy had mined a ton of high-grade gold ore which they shipped to Los Angeles for processing. Early in January 1911 Drumm, Clancy, and John S. Cook, a Goldfield banker, visited Goldstone in an auto to look over the property. In February, Cook bonded the claims. The terms of the bond called for $25,000 to be paid within 18 months, with the first payment to be made by August. Cook himself came to the property with two men to work the property, but before the month of February was out, dropped the bond.[3]

After the deal with Cook fizzled, the mine of Halford and son began to attract attention. In February Halford had material and supplies hauled to his camp and by May the two had sunk two shafts 80 and 50 feet deep and had drifted 30 feet in a four foot wide ledge of "shipping ore". Their cabin was built out of yucca "...which is fitted out in good old miners' style."[4]

By February 1912 some half dozen miners were developing properties, and Mitchell and Andrews struck ore peppered with gold. They were developing a three foot wide vein with a pay streak yielding $100 per ton. Also of note is Rinaldo and Durand's 18 inch pay streak that at a depth of 50 feet ran $65 per ton. In November 1913, there was a small stamp mill at the Drumm Mine.[5]

Mining excitement was stirred anew at the camp in October 1915. On October 15, 1915, gold was discovered on the Redfield claim that ran from $1,400 to $3,000 per ton in gold. Soon the Barstow *Printer* announced "Gold Stone camp is attracting many prospectors, and indications are that Barstow will have a camp equal to any in the Southwest."[6]

The rediscovered gold district attracted swarms of prospectors. By March 1916, there were some 150 "permanent" residents, in addition to a lodging house, and daily mail service and delivery of supplies. For $5 one could buy a round trip ticket from Barstow to the nearby camp, or $10 from Los Angeles with service on Tuesday and Saturday. By May, 1916 the camp had "seven operating properties, two of which have already granted five leases." A townsite was surveyed and several lots immediately sold. Residence lots were priced from $50 to $250 and business lots from $150 to $350. There were nine buildings and tenthouses at the site, and the rooming house could accommodate 25 and boasted a separate restaurant capable of serving more than one hundred people a day. Within a month a large general store was doing business. In October a Chilean mill with a daily capacity of about 20 tons was operating. A subscription campaign to construct a telephone line to Barstow was kicked off in April, 1917, but the line apparently never was completed. A few months later, in June 1917 the post office of Goldbridge was opened at the camp, and Malcom Smith sworn in as postmaster. At that time the Goldstone Company's shaft down 210 feet.[7]

But as suddenly as the camp swept onto the mining scene, news abruptly ceased. Perhaps the entry of the United States into World War I on April 6, 1917, had proved too much a draw of men and equipment to the still fledgling mining camp. The post office closed August 15, 1918.[8]

The camp was never completely deserted, nine years later the Barstow *Printer* noted: "A half dozen people are in Goldstone and several have been working their claims for months. One miner is reported as making wages developing his property and using a dry washer."[9]

George Drumm continued to develop a promising vein and in August 1928 had begun construction of a "cyanide plant." Only months later the camp lost its long time promoter, George Drumm died on January 10, 1929.[10]

But with the new decade came new progress. New properties were developed and by 1931 there were two mills in operation, one at the Goldstone mine and the other at the Belmont.[11]

Acknowledgments:

I thank Alan Hensher for liberal use of his source information, his monumental work of copying information pertinent to the Mojave Desert from the *Los Angeles Mining Review*, and above all for his continued research into the history of the Mojave Desert, and allowing *me* to be his sounding board.

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Garlock: The History of a Milling Town

Alan Hensher, *593 Collins Drive, Apt. 3, Merced, Cal. 95348*

If a mine needs anything above all, it needs a mill. And that mill needs water.

So it was with great foresight that Eugene Garlock, the owner of a mine near Tehachapi, saw an opportunity in the discovery of rich and large deposits of gold ore in the Rand Mountains, northeast of Mojave. The Yellow Aster claim, discovered in April, 1895, turned out to be the most important mine.

But the Rand Mining District lacked a supply of readily available water. In the beginning, all of it had to be hauled from Goler and Cow Wells, about a dozen miles up a sandy road from the floor of Fremont Valley. There, in the deep gravels lying below Cow Wells, were heavily mineralized but abundant quantities of the "precious fluid."

Garlock moved an 8-stamp mill from the Tehachapi area to Cow Wells, perhaps in late 1895 or early 1896. As the pioneer mill operator, Garlock readily found customers.

The nucleus of a settlement began to form around Cow Wells during early 1896. A post office, named after Garlock, was established in April. Mrs. I.M. Kelly managed the office and corresponded for the Bakersfield *Daily Californian*; her husband, John, owned the camp's feed yard. (He would be elected sheriff several years later.) Mrs. Archie Martin's boardinghouse was praised for its neat table and excellent food. ". . . One does not look for such creature comforts as she furnishes away out here on the desert." Four stage lines, meanwhile, were running through Garlock, bound for Randsburg. Since Garlock was receiving mail and newspapers daily, the residents "will not be deprived of the comforts of civilization though living remote from the railroad . . . ." In fact, another correspondent noted, in early June, "everything man and beast requires can be obtained here."

Despite the heat of summer, the traffic to Randsburg and elsewhere was so heavy that it reminded one correspondent of "the old days of prosperity — . . . The boom has come again and come to stay . . ." Five saloons were doing business at Garlock by early July. A telephone line was soon extended from nearby Koehn Springs, another milling camp, to Garlock, where messages could be relayed to Randsburg over a telegraph line. In August, a Porterville man put up a 2-stamp mill and Eugene Garlock installed another engine at his mill. That fall, the Visalia Mining and Milling Company began work on a 5-stamp mill.

Although the Kern County supervisors had established voting precincts (Red Rock and Goler) at the desert mines in 1894, the camps remained almost out of reach of local government. The supervisors finally organized a judicial township for the Randsburg-Garlock area, in late 1896 or early 1897. The township was served by two justices of the peace and two constables. Then, after San Bernardino County began pushing a road toward the Rand district, the Kern County supervisors hastily had a road surveyed from Warren's Station, near Mojave, to Garlock and Randsburg, much to the delight of one correspondent in the early 1890s: "By this route one feels as if he had hardly left civilization . . No wonder one is agreeably surprised after making a trip by the short route of Buzzards, Bones and Beer from San Berdoo!"

Meanwhile, by January, when the Visalia mill started up, Garlock was "building like magic." The business district, where lots were selling for only a few hundred dollars, now embraced three general stores, one drugstore, one physician (Dr. William Wright, who worked out of his house), one assay office, four lumberyards, four livery yards, two butcher shops, three boardinghouses, two barber shops, two restaurants, five saloons, the post office (served by a daily mail delivery from Mojave), and a recently opened office of Wells, Fargo & Company. Several businesses went by such colorful names as Cheney's Thirst Emporium, the Desert House, and the Big Barn, where the stages would change teams.

Garlock was also the home of several dozen children (51 by May). Money was collected and a lot set aside for a school. In response to a petition signed by 21 parents, the county supervisors organized the Garlock School District (and one at Randsburg) in early February.

By then, Garlock was developing an enviable range of institutions. A newly organized fire-protection association – reported to be the first in the desert – ordered a dozen fire extinguishers and a dozen leather buckets. A justice of the peace was at work, holding a preliminary hearing for a grand-larceny suspect. And a weekly newspaper, the Garlock *News*, was founded. Published by Charles F. Schmidt, Jr., and H. H. Schmidt, the *News* contained four 11x16-inch pages and cost $2 a year.

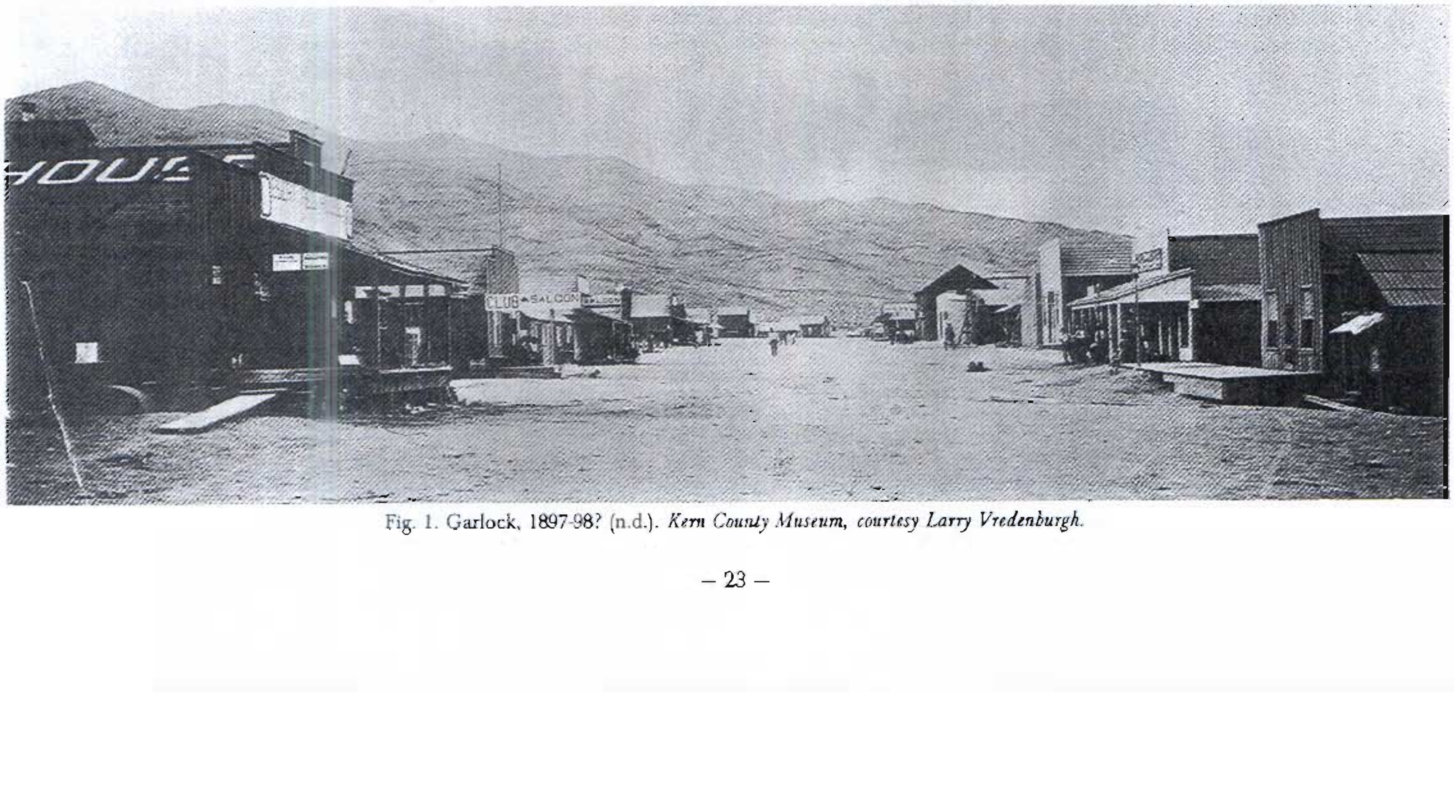

Fig. 1. Garlock, 1897-98? (n.d.). *Kern County Museum, courtesy Larry Vredenburgh.*

Town life

Strangely, a townsite wasn't laid out until June, 1897. But what a townsite! Offering the comforts of civilization were two well-managed hostelries, one owned by A.J. and Sarah Doty, the other by Zeke T. Lillard and his wife.

Doty's Hall, which also served as the stage station, was the most important place in town. Doty's was Garlock's only two-story building and, despite persistent sandblasting by the wind, stood for a certain elegance. Inside were 10 bedrooms and a parlor, lobby, large dining room, and kitchen. Outside stood an ornate street lantern, one of four along the main street.

The cuisine was another matter. Although fresh meat arrived three times a week from a local slaughterhouse, the lack of refrigeration forced Doty's to serve mostly canned or dried food.

Lillard's, in contrast, was famed for its food. Zeke Lillard and his family had emigrated from Los Angeles, hoping to make a quick fortune in the gold fields. But Mrs. Lillard instead built up an outdoor kitchen into a bakery and then the hotel. Working almost round the clock, she would bake pies and bread for 100 lunches before getting breakfast ready. Mrs. Lillard might also have to prepare as many as 300 meals a day for her regular boarders: miners and mill workers. Her two daughters helped out by waiting on tables.

Zeke Lillard, in contrast, tended to drift away from the front desk to the sitting room or porch, smoking a cigar or talking about mining. And whenever hot weather arrived, the Lillards' son, Frank, would visit the tall jar of chocolate-covered cream candies on the front desk. "It was a drippy, sticky mess, but wonderful. I got to eat the whole batch, right down to the liquid chocolate on the bottom."

At the other end of Main Street stood the Garlock School, a building of simple dignity. Inside were two rows of home-made desks, one row for the boys, the other for the girls. Between the desks stood a cast-iron stove. Near the door was a dipper and galvanized pail for water. At the front of the room was the teacher's desk, platform, and blackboard. During the boom, the school was well attended (reaching a peak of 20 in 1898) and served as a dance hall on Saturday nights, a church on Sundays, and a meeting place for the literary society.

Like the school, Dr. Wright's house was improvised to serve a variety of needs. A simple frame structure, the house contained only about 400 square feet, but the back room and lean-to were used as living quarters, bedrooms, a kitchen, and a laundry; the front room served as an office (where Dr. Wright also pulled teeth), drugstore, and assembly room for miners. About 1900, half of the front room was partitioned off for use as the post office.

Two businesses were especially popular, although in different ways. Attached to one saloon were a few cribs and a gambling den. To fend off critics, mainly prohibitionists, the saloon owners organized their own peace-keeping force: the Wirecutters' Association. The other force for cleanliness – literally – was Juan Basarto, a shy young man who ran a hand laundry; he charged a standard 25 cents for jobs of any size.

Juan Basarto was typical of many of Garlock's residents: unpretentious and generous. "It was a friendly little settlement where people helped each other in common need and hardship," recalled Bessie McGinn, one of the daughters of Jim McGinn, a merchant. For example, when an impoverished miner named Becker painfully shuffled in with a strangulated hernia, Dr. Wright summoned a doctor from Randsburg and together operated on Becker on the kitchen table. Afterward, Mrs. Wright cared for Becker until he was able to return home to Los Angeles. Mrs. Wright "was always helping someone who was down on his luck," her son, Sherman, remembered.

Happily, day-to-day life was rarely so grim. The residents could look forward to a variety of homespun activities: the daily arrival of the stage, dances in the school, music, meetings of the literary society, baseball games, Sunday sermons preached by Dr. Wright, and Independence Day festivities, at which Dr. Wright would deliver the oration.

Signs of decline

Although Garlock was prospering, some signs of weakness began to appear. The losses of precious metals during milling became so great that a cyanide plant had to be installed at one mill in May, 1897. Mine owners in the Rand district, in the meantime, were sinking their own wells: at dry Cuddeback Lake, east of Randsburg, and near the St. Elmo Mine, southeast of Johannesburg.

But as long as the Rand district continued to yield rich and abundant ore, life at Garlock went on as usual. Despite the press of

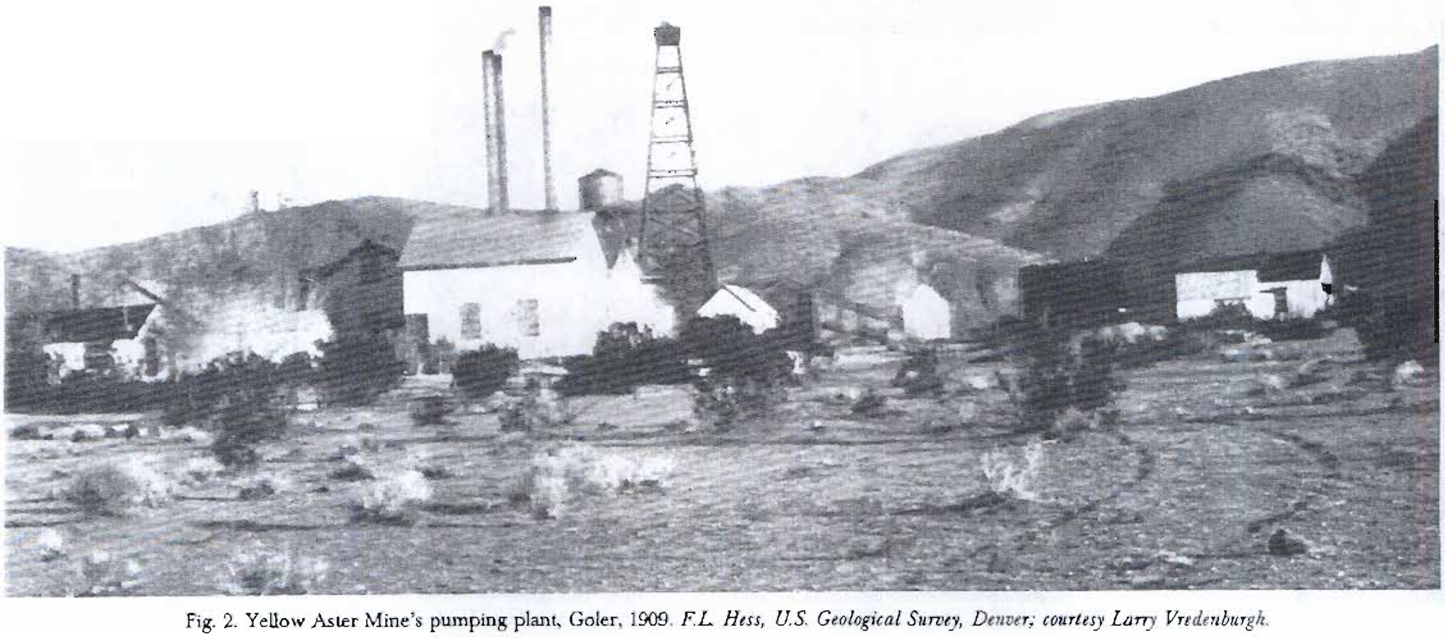

Fig. 2. Yellow Aster Mine's pumping plant, Goler, 1909. *F.L. Hess, U.S. Geological Survey, Denver; courtesy Larry Vredenburgh.*

business, for example, the mills had to shut down in anticipation of the Independence Day festivities in 1897. As before, the "boys" went on a binge, first at Randsburg on July 3 and then at Garlock on July 5, and then took another day to recover. The Garlock Mill had to wait until the morning of July 7 to resume operations. ". . . Now everything is again serene, and the music of the stamps is heard from 'early morn till dewy eve.'"

A more respectable pastime – and the most popular – was baseball. In its first game, the Garlock team beat Randsburg, 11 to 5. But in "the event of the season" in September, the Tehachapi team came from behind and beat Garlock, 4 to 1.

By October, Garlock supported six mills, running day and night almost entirely on large shipments of ore from the Rand district. But appearances could be deceiving. Each of the largest two mills ran only 10 stamps; one mill was equipped with a roller, a cheap but inefficient crushing device. Only one mill was powered by a gasoline engine; the others were equipped with steam engines, which voraciously burned huge stacks of hot-burning creosote ("greasewood"). Two cyanide plants, meanwhile, were processing the tailings.

Garlock continued to lose its monopoly of mills and wells throughout the fall and winter. Small mills were already operating at Mesquite Springs, Koehn Springs, Cuddeback Lake, and Johannesburg, where a 10-stamp mill started up in early December. Eastern investors, in the meantime, were pushing the construction of a 29-mile rail line – the Randsburg Railway – from Kramer, a station on the Santa Fe Railroad, to Johannesburg. They planned to ship ore from the Rand to a 50-stamp mill under construction at Barstow. The prospects for business looked good: in March, 1898, one mine, employing 120 men, was shipping from 30 to 50 tons of high-grade ore a day to Garlock's mills.

The final years

For Garlock, the beginning of the end came in June, 1898, when the mill at Barstow started up. The Yellow Aster soon began shipping its ore to Barstow and closed all but one of the mills at Garlock. Since the Yellow Aster already had a pipeline running from its wells at Goler, it built a 30-stamp at Randsburg; the mill started up in February, 1899. Only 50 people were living in Garlock – and 34 in its environs – by the spring of 1900. When the Yellow Aster built a second, 100-stamp mill – the largest in the state – in 1901, the last families left Garlock. School enrollment fell from 19 that spring to three or four in May, 1903, when the school district was dissolved.

The Wright family left that year. As son Sherman recalled, the Wrights "abandoned the house and gave Juan [Basarto] our horse and buggy, together with an old four wheeler wagon in exchange for driving us to Mojave. There we took the train for Oakland. I never knew how my father got together enough money for the train fare. He was at heart a promoter, quite resourceful, and always managed somehow." The post office closed in March, 1904. Only Juan Basarto remained. And soon, even he was gone.

The revival

Garlock began to reawaken several years later. The construction of a railroad line from Mojave to the Owens Valley from 1907 through 1910 brought a section crew and station to the townsite. A slaughterhouse still operated there. Then, in 1914, Sarah (Granny) Slocum bought up much of the townsite and opened a boardinghouse, which her customers dubbed the "Hotel de puke." She scrapped the mills during World War I.

After the war, J.D. Voss and other operators reopened several nearby gold and silver claims. A small settlement grew up at the townsite. In response to a petition signed by 13 parents, the county supervisors re-established the Garlock School District in January, 1920; Garlock annexed the short-lived Saltdale School District several months later. About then, John D. Norton opened a general store, where the post office was re-established in October, 1923.

A flurry of small-scale mining continued for several years. After shipping out some ore to a smelter in early 1925, in fact, Voss was feeling "very much encouraged."

The revival soon faded away. Norton moved to Cantil and opened a store. Garlock lost its post office in June, 1926. The school was moved to nearby Saltdale a few years later.

The onslaught of the Depression, ironically, led to an increased interest in gold mining. A cyanide plant was built at Garlock in 1931. Major placer-mining operations, employing hundreds of men, also were resumed in the Goler district. But Garlock never revived as a town.

SOURCES

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Thorough day-to-day coverage can be found in the Bakersfield *Daily Californian* and the *Mining & Scientific Press* (San Francisco) for 1896 and 1897; occasional articles can be found in the San Bernardino *Times-Index*.

The actions of the county government can be traced in the Bakersfield *Daily Californian*, especially January, 1897; the minutes of the Kern County board of supervisors (microfilm), 1896 and 1897, in the office of the clerk of the board, Bakersfield; "Annual Report of the Condition of Public [Common] Schools in the County of Kern," 1898-99, 1899-1900, 1900-01, and 1901-02, in the California State Archives, Sacramento; and "School Districts: Formation Change of Boundaries: Garlock," file 19-14, also in the office of the clerk of the board.

Acknowledgments:

I thank Larry M. Vredenburgh, of the Bureau of Land Management, Bakersfield; Jack Peskin, of Reno; and Mark Faull, of Red Rock Canyon State Park, for their help.

Studies of the Garlock Fault, Southeastern California: A Brief History with Commentary

George I. Smith, *Geologist-Emeritus, U.S. Geological Survey, Menlo Park, CA*

Abstract

The 260-km-long Garlock fault is the second longest fault in California. It trends northeasterly near its western end but curves gradually toward the east, eventually trending slightly south of east. Geologic studies of the fault began in 1910 and continue into the late 1990's, leading to a more complete understanding of its geologic and tectonic character, including its earthquake recurrence intervals that are mostly estimated to be between a few centuries and a few millennia. However, there is little agreement on the origin of the fault and its geologic function in the complex tectonic setting of southwestern North America.

The Garlock fault probably was initiated in the middle Tertiary as a consequence of plate tectonic activity, though indirect evidence argues for an earlier origin. Its total left-lateral displacement is near 65 km. However, the lack of geomorphic expression of activity along the easternmost and westernmost few kilometers of the main branches, combined with a host of geomorphic indications of very late displacements along the middle segments, has led many to consider the Garlock an intracontinental transform structure, where the largest lateral displacement is found in the middle segment with little or no lateral-displacements at one or both ends.

Near the middle of the Garlock fault, just west of Koehn (dry) Lake in Fremont Valley, the fault as traced from its western end dies out according to geomorphic criteria, but its apparent continuation is exposed further east in bedrock terrain. There, stratigraphic criteria indicate it to have been inactive since the Pliocene(?), possibly with the stresses responsible for its earlier movement transferred to a south-dipping thrust fault that crops out to its north. About 3 km north of the easternmost geomorphic expression of the west half of the fault, geomorphic indications of a "new" left-lateral Garlock fault materialize which can be traced almost to its east end. There, the Garlock loses its geomorphic expression and is truncated by the Mule Spring fault zone, a reverse fault exposed along the northeast base of the Avawatz Mountains.

Seismic activity along the fault between 1932 and 1981 was concentrated near the western ends of both the western and eastern halves of the Garlock fault, whereas the eastern parts of those two segments have been nearly aseismic. The western half of the entire fault is characterized by continuing aseismic creep, whereas its eastern half lacks measurable creep.

It appears that for some purposes, the Garlock fault could be considered as a fault composed of two halves, with each half having similar but not precisely identical seismic characteristics and having terminations indicating no geologically recent strike-slip activity.

Introduction

The Garlock fault is the second longest fault in California, yet its trace is virtually at right angles to the San Andreas fault, the longest in the state, as well as to the vast majority of shorter faults in California (Fig. 1). The middle part of the Garlock fault was first identified in 1910, and the numerous studies since then have shown that its full length is about 260 km and that it is left lateral, has about 65 km of offset, and results from northeast-southwest compression. First displacements along it probably took place in the middle Tertiary, but estimates of its initiation range from Late Cretaceous almost into the Quaternary. No major earthquakes along the fault have taken place in historic time, but numerous lines of evidence suggest that one will take place in the future.

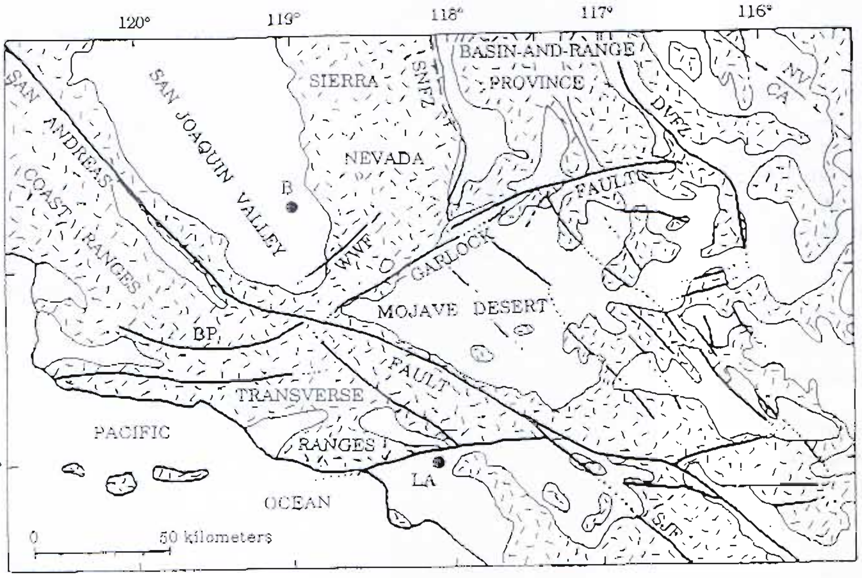

Fig. 1. Part of southeast California showing locations of Garlock fault and selected other faults mentioned in text. Names of major provinces and sub-provinces also labeled. Abbreviations: B, Bakersfield; BP, Big Pine fault; CA, California; DVFZ, Death Valley fault zone; LA, Los Angeles; NV, Nevada; SJF, San Jacinto fault; SNFZ, Sierra Nevada fault zone; WW, White Wolf fault. See Figure 2 for names of geologic and geographic features along the Garlock fault that are mentioned in text.

Discussion of this history of past studies is here divided into groups whose findings were similar, and they are generally discussed in chronological order. This paper, however, can not be a summary of all the literature relevant to the Garlock fault. Some of the omitted studies either confirm earlier studies, seem inconclusive, or have only peripheral relevance to the Garlock fault, and some relevant studies may have been omitted inadvertently.

The Garlock fault is composed of at least two branches whose traces overlap by about 50 kilometers in the vicinity of Fremont Valley (Fig. 2), near the middle of the 260-km-long fault. Throughout this area, the west branch lies several kilometers south of the east branch (Fig. 2).

Identification of The Garlock Fault, Its Extent, Character, and Amount of Displacement

The first recognition of the Garlock fault was by Hess (1910) during his study of the Randsburg mining district, which is 170 km northeast of Los Angeles, 110 km east of Bakersfield, and about 7 km south of the central part of the fault (Fig. 2). He named the fault after the town of Garlock, a mining settlement at the time, located on the fault trace. Hess recognized from the landforms that the fault extended many kilometers east and west of

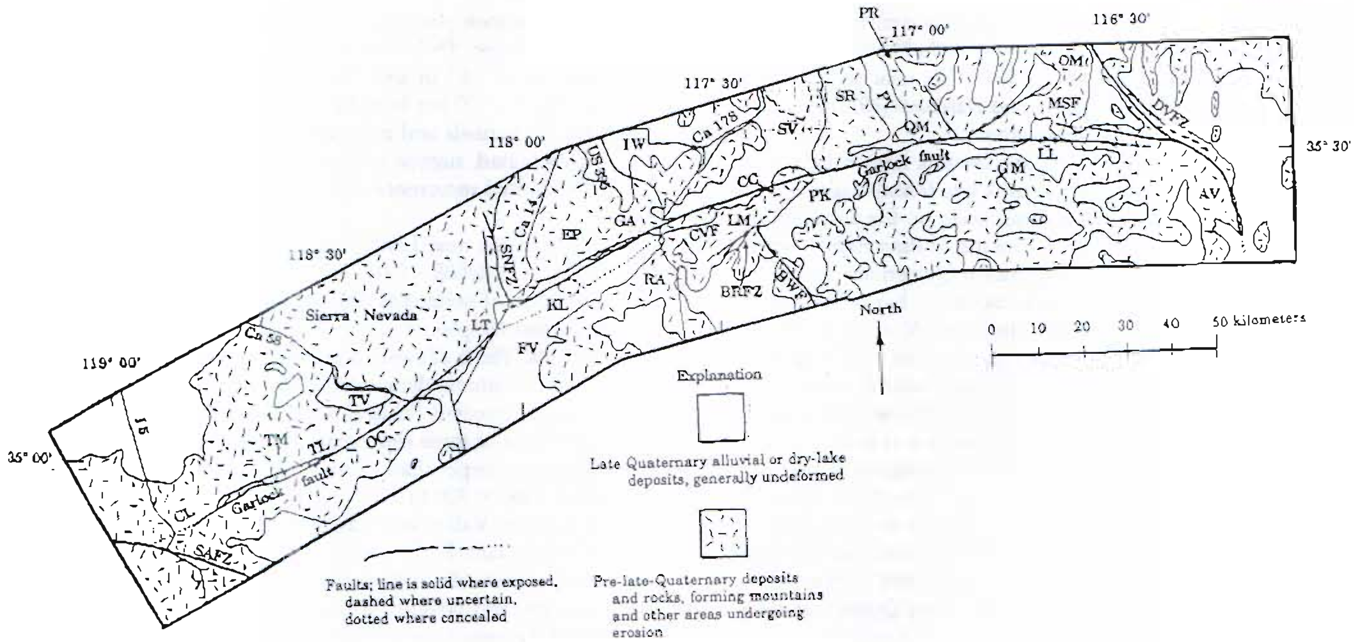

Fig. 2. Strip map along the Garlock fault showing faults, mountains, valleys, dry lakes, and towns or settlements near the Garlock fault and mentioned in text. Abbreviations: AV, Avawatz Mountains; BRFZ, Brown's Ranch fault zone; BWF, Blackwater fault; CC, Christmas Canyon; CL, Castac Lake; CVF, Cantil Valley fault; DVFZ, Death Valley fault zone; EP, El Paso Mountains; FV, Fremont Valley; GA, Garlock town site; GM, Granite Mountains; IW, Indian Wells Valley; KL, Koehn (dry) Lake; LL, Leach (dry) Lake; LM, Lava Mountains; LT, Lone Tree Canyon; MSF, Mule Spring fault; OC, Oak Creek Canyon; OM, Owlshead Mountains; PK, Pilot Knob Valley; PR, Panamint Range; PV, Panamint Valley; QM, Quail Mountains; RA, Randsburg town; SAFZ, San Andreas fault zone; SNFZ, Sierra Nevada fault zone; SR, Slate Range; SV, Searles Valley; TL, Twin Lakes; TM, Tehachapi Mountains; TV, Tehachapi Valley. Abbreviations for highways: I 5, Interstate 5; US 395, U.S. Highway 395; Ca 14, California Highway 14; Ca 58, California Highway 58; Ca 178, California Highway 178.

the Randsburg area, but apparently he did not detect its strike-slip character.

The "Fault Map of California" (Willis and Wood, 1922) showed the Garlock fault correctly located between the Randsburg area and its western end, where it meets the San Andreas fault. To my knowledge, no published work had described the western end of the the fault prior to compilation of that map, but unpublished geologic mapping must have existed, perhaps by the map's compilers. East from the Randsburg, the fault is plotted on the Willis and Wood compilation as a dotted line that extends to a point several kilometers beyond where later studies showed it to curve southeast and apparently join the Death Valley fault along the northeast edge of the Avawatz Mountains (Fig. 2).

A reconnaissance study by Gregory and Noble (1923) was probably motivated in part by the question implicitly raised by the 1922 map: Where is the eastern end of the Garlock fault? As a result of that reconnaissance, Gregory and Noble concluded that the Garlock fault constitutes one of the major structural lines of California, considered it to be the natural northern boundary of the Mojave Desert, and determined that along the northeast side of the Avawatz Mountains, the fault curves to the southeast and apparently joins the Death Valley fault (Fig. 2).

Hulin (1925), in a published report on the Randsburg 15' quadrangle, was the first to conclude that the Garlock fault was left lateral slip; he estimated its offset to be ~8 km. This estimate was based on the correlation of a contact between undifferentiated Paleozoic rocks and the Atolia Quartz Monzonite in the El Paso Mountains, north of the fault, with an inferred contact between small outcrops of Paleozoic rocks (immediately south of the fault and up to 8 km to the east), and the Atolia Quartz Monzonite exposed 5 km further east. Hulin's (1925) sense of displacement was thus correct, but Dibblee (1967, fig. 68) showed that the Paleozoic rocks south of the fault are probably landslide debris from the El Paso Mountains that was shed southward across the fault after most of its present displacement was completed.

Later, as the result of more trips to the area, Noble (1926) published an account of his new observations along the eastern Garlock fault, expanding on those in the earlier paper that he co-authored with Gregory. He described in more detail the topographic contrast between the mountains north of the Garlock fault, now considered to be at the south edge of the Basin-and-Range Province, compared to the much less spectacular mountains of the Mojave Desert (Fig. 1). He also noted the very thick sections of pre-batholith sedimentary, metasedimentary, and metavolcanic rocks exposed in the ranges to the north, and the near absence of similar rocks immediately to the south in the Mojave Desert, concluding that erosion had removed them.

Three papers by Hewett (1954a, 1954b, and 1955) set forth his interpretation of the Garlock fault, based on nearly three decades of observations. By the time these papers were written, a large amount of new topographic and geologic mapping had been published that post-dated Noble's 1926 paper. Also, aerial photographs of most of the involved areas were available, making it possible to produce a photo-interpreted fault map showing very late Cenozoic fault displacements over the entire Mojave Desert region.

The next study of displacement along the Garlock fault was by Smith (1962). He correlated two Mesozoic-age, steeply dipping, northwest-trending dike swarms, one north of the fault and one south of it but well to the east, and he estimated the lateral offset along the fault to be about 65 km. A 25° difference in the trends of the two dike swarms was attributed to clockwise rotation of the terrain south of the fault. Later, Michael (1966) proposed a correlation of two areas, now on opposite sides of the Garlock fault, that were characterized by northwest-trending faults. One set of faults is in the southern Sierra Nevada, and the other set consists of the series of northwest-trending faults in the southwestern two-thirds of the Mojave Desert; Michael estimated ~75 km offset was indicated. Later, Smith and Ketner (1970) correlated sets of steeply dipping eugeosynclinal metasedimentary rocks in the El Paso Range north of the fault, with similar rocks in Pilot Knob Valley (Fig. 2) south of the fault, and concluded that displacement of these rocks indicated 50 to 65 km total offset in the central part of the Garlock fault. Davis and Burchfiel (1973) and Carr and others (1992) supported this correlation.

The amount of displacement along the Garlock fault in Quaternary time (considered here to have lasted from 1.6 to 0 Ma) was inferred by Carter (1982) to be 10 to 20 km. This was based on the discovery that fragments of a distinctive schist exposed in the El Paso Range, north of the fault, are present in sediments of apparent Quaternary age to the east and south of the fault. A Pleistocene lacustrine bar on the east edge of now-dry Koehn Lake has been displaced 80 m by the Garlock fault. If the age of that bar is ~15 ka, as suggested by a ^{14}C date on aquatic shells found near the bottom of a trench located a few hundred meters southwest of the bar (Burke, 1979), extrapolation of this rate to the entire Quaternary period indicates 8 to 9 km displacement during Quaternary time. This calculation is suspect because there is no provable connection between the dated lacustrine horizon and the age of the bar. However, it may be nearly correct inasmuch as tufa from the crest of the Koehn Lake bar was dated as 11.4 ka (Clark and Lajoie, 1974), allowing a calculated displacement rate of 7 m/ky, and ~11 km of displacement during Quaternary time, though ^{14}C ages from some tufas in nearby Searles Valley have been found contaminated and much too young (Smith, in press).

Earthquakes Along The Garlock Fault And Their Recurrency Rates

A study of the seismicity of the Garlock fault by Astiz and Allen (1983), covering the 50-year period between 1932 and 1981, showed that the fault had different characteristics in segments east and west of their "Rand" alignment array that is on the east branch of the Garlock fault at ~117° 40' longitude, about 6 km east of where the fault crosses U.S. Highway 395 (Fig. 2). West of that line, the fault had a low but definitive seismicity during the period of study, well documented aseismic creep, and a relatively complex fault trace; east of that line, the fault had less seismicity, no demonstrated creep, and a relatively simple fault trace. The location of this change in character also coincides aproximately with a division in the maximum depths of microseismicity on and near the fault. The microseismicity has a maximum depth of ~13 km in the western segment and ~10 km in the eastern segment (calculated by McGill and Sieh, 1991, using the USGS/Caltech southern California network earthquake catalog, following principles developed by Sibson, 1984). Comparison by Astiz and Allen (1983) of the seismic energy release during the preceding 50 years shows that it is three orders of magnitude smaller than the seismic moment rate estimated from geologic studies, "indicating that the fault currently represents a temporal seismic gap, and that the potential exists for large earthquakes."

Trenches have been excavated at several places along the Garlock fault for the purpose of establishing the numbers of geologically young fault displacements over measured periods of time. Starting near the western end of the Garlock fault, at Castac Lake (Fig. 2) that is ~5 km east of the apparent junction of the Garlock fault with the San Andreas fault, in the southwest part of the Tehachapi Mountains, LaViolette and others (1980) placed four trenches normal to the projected trace of the fault. No evidence of faulting was found in the top ~3 m of the valley fill; a ^{14}C date from one of the middle horizons indicated an age of ~8 ka, suggesting that deposits in the exposures represent the past ~10 to ~20 ky. However, a few kilometers to the northeast, geodetic measurements across an area that included a 300-m-wide gouge zone along the Garlock fault suggest that Holocene left-lateral movement may be present in this area but distributed over a wide zone (Snay and Cline, 1980) and A. B. Arnold (pers. comm., quoted by Astiz and Allen, 1983).

At Twin Lakes (Fig. 2), ~40 km northeast of the west end of the fault, trenching by Stepp and others (1980) indicated two faulting events, the first took place prior to 2,800 years ago and the second since 890 years ago. Only vertical offsets could be demonstrated, with displacements of ~0.6 m and ~0.9 m, respectively. At Oak Creek Canyon (Fig. 2), ~50 km from the west end of the Garlock, they noted offset stream channels and soil sequences that, though their ages are poorly constrained, suggest average lateral movement 1.6 to 3.3 mm/yr and an "indicated recurrence interval . . . on the order of thousands of years."

At a trench site near Lone Tree Canyon (Fig. 2) in the southern Sierra Nevada, (McGill, 1994a; oral commun., 1994), exposures indicate ~60 m of horizontal offset of a channel in alluvial sediments containing two samples of charcoal that yielded ^{14}C ages of ~19.0 ka and ~18.3 ka. These allowed calculation of a minimum slip rate of 3 mm/yr; other relations allow calculation of maximum slip rate of 10 mm/yr, using corrected ^{14}C ages. One trench across the fault provided evidence of two or three slip events; a corrected ^{14}C age on charcoal from a horizon deeper than the oldest event indicates earthquake recurrence rates of 700 to 2700 years.

The Fremont Valley area has been the site of two trenching studies of the Garlock fault. Burke (1979), working in a trench crossing the fault near the east side of Koehn Lake (Fig. 2), counted 9 distinct fault displacements, and possibly as many as 17. A ^{14}C age on shells near the base of the trench showed that these events took place during the past ~15 ky, indicating a recurrency rate of 880 yrs to 1,670 yrs. A trench across a sag pond along the Garlock fault (McGill and Rockwell, in press) provided convincing evidence of five strong earthquakes in the past 5100 yrs, with less convincing evidence of three more. The average recurrence rate was 700 to 1200 years, close to the rate determined by Burke (1979). McGill and Rockwell conclude that if the two youngest earthquakes are ^{14}C dated correctly at 160 and 350 yrs ago (AD 1790 and AD 1600), a large quake on this part of the fault within the next century or two may not be likely. They note, however, that the fault's recurrence rate is very irregular.

The Garlock fault has been trenched in two places in Searles Valley (Fig. 2). Roquemore and others (1982), in a trench dug across the fault in the south-central edge of Searles Valley, identified 6 displacements in sediments of Holocene age, indicating a recurrence rate of 1,700 yrs. In southeasternmost Searles Valley, McGill and Sieh (1993) dug 20 backhoe trenches across and near the Garlock fault as a means of determining the offset of a shoreline eroded by an indirectly dated high-stand of Pleistocene Searles Lake, as well as the offset of a post-shoreline channel. The shoreline was offset 82 to 106 m (preferred measurement: 90 m); the channel was offset 68 m. These relations indicate slip rates along this segment of the fault of 5 to 11 mm/yr, with the preferred rate, using corrected ^{14}C ages, being 5 to 7 mm/yr.

McGill and Sieh (1991), using large-scale air photographs, selected four areas in a 130-km section along the central and easternmost Garlock fault for study of geomorphic features that are offset by small amounts and are most abundant, presumably representing the youngest event. The studied areas are near U.S. Highway 395, Searles Valley, Pilot Knob Valley, and in the vicinity of Leach Lake (Fig. 2). Measured displacements in these areas ranged from 2 to 7 m/event, calibrated slip rates from 3 to 9 mm/yr, and recurrence intervals ranged from 200 to 1300 yrs. At the easternmost part of the fault, however, fault activity may have been less frequent and intense inasmuch as displacements ranged from 2 to 3 m/event, slip rates from 1 to 9 mm/yr, and recurrence intervals from 200 to 3000 yrs. Future earthquake magnitudes, estimated assuming offsets of 7 m and ruptures ranging from the entire fault (260 km) to 30-km segments, ranged from magnitudes 6.6 to 7.8.

Two field-trip guides describe segments along the east half of the Garlock fault, many of which can be reached using 4-wheel-drive vehicles (McGill, 1994b; McGill 1994c). They are based on both published and (at the time) unpublished work, and they describe evidence indicating earthquake-recurrence intervals ranging from several hundred to a few thousand years. The 1994b paper concentrates on the easternmost part of the Garlock fault, near the Avawatz Mountains, where most geomorphic features indicate ~3 m of left-lateral slip during the most recent earthquake; preliminary ^{14}C dating of desert varnish suggests slip rates of 0.4 to 1.5 mm/yr, much lower than rates determined on more central parts of the fault. The 1994c paper describes several sites in the central and east-central parts of the fault. Trenching and offset geomorphic features along the south side of the El Paso Mountains indicate 2 to 18 m of left-lateral displacements. At a site only 5 km west of the easternmost geomorphic evidence of activity on the west branch of the Garlock, an offset channel indicates 60±5 m of left-lateral displacement. Estimates of slip rates decrease eastward along the east half of the fault; studies near the El Paso Mountains, Searles Valley, Pilot Knob Valley, and Leach Lake areas, are respectively 6, 5, 4, and 2.5 mm/yr.

Collectively, these studies estimate earthquake recurrence rates along the Garlock fault ranging from ~200 yrs to ~3,000 yrs. This is in accord with the low activity observed during historic time, but it also may increases the likelihood of a major earthquake sometime in the near future.

Evidence of The Earliest Activity Along The Garlock Fault

Noble (1926) and Hewett (1954a, 1954b, 1955) inferred that the Garlock fault became active in the late Mesozoic or very early Tertiary. Their reasoning, especially that of Hewett (oral discussions, 1952-1962), was based on two points. First, that the several-kilometer-thick sections of pre-middle-Mesozoic, mostly marine, sedimentary rocks that crop out north of the Garlock fault almost certainly were also were deposited over much of the area south of the fault. Thick sections of these deposits are still preserved <1 km north of the fault, in the El Paso Mountains and Slate Range (Fig. 2), as well as in areas further north (Grose and Smith, 1989). Some of these rocks are also preserved in the Mojave Desert, proving that the region was within the area of their deposition even though most of their sections are thinner (Smith and Streitz, 1989). Between Leach Lake and Tehachapi Valley (Fig. 2), however, the only large areas of such rocks south of the Garlock fault are 30 km or more south of it. Because of the many stratigraphic studies of the pre-Mesozoic sedimentary rocks that had been made during the previous quarter of a century, Hewett also was able to make quantitative estimates of the rocks that were virtually "missing" from the Mojave Desert, with the line of demarcation clearly being along the Garlock fault. He estimated that a minimum of ~4.5 km to ~6.0 km of pre-late Mesozoic sedimentary and metavolcanic rocks had been eroded from the Mojave Desert side of the Garlock fault, yet were preserved in mountains throughout the southwestern Basin-and-Range province to its north. (He also wondered where all of that eroded debris was now!) Hewett's estimate of the age of initial activity along the Garlock fault was late Mesozoic, but he made no estimate of its horizontal displacement.

The second reason for Hewett's conclusion regarding the late Mesozoic or early Tertiary age of first activity on the Garlock fault is that early Tertiary sedimentary deposits known in several areas just north of the Garlock fault are virtually unknown in the Mojave Desert area. He deduced from this observation that the Mojave Desert must have been uplifted along the Garlock fault (and San Andreas fault ?) in very late Mesozoic or early Tertiary time and thus was undergoing continuing erosion instead of sedimentation during all of early Tertiary time. The best dated early Tertiary section north of the Garlock fault is in the El Paso Mountains, where the 2,000-m-thick Goler Formation (Dibblee, 1952) contains vertebrate fossils of Paleocene age (McKenna, 1955; West, 1970). A distinguishing characteristic of these deposits is the abundance of very well rounded quartzite cobbles and boulders. Sediments bearing quartzite cobbles and boulders also crop out northeast of Tehachapi Valley in the early Tertiary Witnet Formation, where they are overlain, with a 30° to 50° angular unconformity, by two vertebrate-bearing Miocene formations (Buwalda, 1954; Quinn, 1987). Gravels containing similar quartzite clasts also are found along the crest and flanks of the Slate Range (Smith and others, 1968); these gravels are undated but must be old because the only known nearby sources of quartzite are the Precambrian Stirling Quartzite and Cambrian-to-Precambrian Wood Canyon Formation that now crop out to the east in the Panamint Range (Fig. 2), indicating that the fragments had to be transported west to the Slate Range area before the two areas were separated by Panamint Valley, now more than 400 m deep in this area.

Counterclockwise rotation of the middle-to-upper Miocene Dove Spring and lower-to-middle Miocene Cudahy Camp Formations in the El Paso Mountains (Fig. 2), formerly, the upper part of the Ricardo Formation of Dibblee (1952), has been reported by Burbank and Whistler (1987) and Loomis and Burbank (1988) on the basis of paleomagnetic data. Three fission-track dates on zircons from volcanic ash beds, plus correlation with global paleomagnetic stratigraphy, show that these formations range in age from 18 Ma to 7.5 Ma. Strata older than ~10 Ma had been rotated clockwise 15° to 20¡ (presumably, by earlier crustal movements), but younger strata show a history of progressive counterclockwise rotation until ~7.5 Ma. These authors attribute the rotation to drag caused by the left-lateral displacement of the Garlock fault, 6 to 11 kilometers south of the sampled section, with the age of the initial rotation between ~10 Ma and ~7.5 Ma representing the initial activity along that fault. The lack of continuing rotation of the Dove Spring Formation sediments is attributed by them either to the fault displacements having ocurred in two (or more?) pulses or to structural isolation of the sediments from the effects of the faulting after that date.

Non-fault Deformation along the Garlock Fault Zone

An anticline and two synclines lie subparallel to the Garlock fault in and near the Lava Mountains (Fig. 2). The folds lie north, south, and astride the Garlock fault, and stratigraphic relations show that they formed since deposition of the Lava Creek B ash bed, whose age is now considered to be 667 ka (Izett and others, 1992). The anticline has a maximum relief, relative to the syncline to its northwest, of more than 250 m, and a length of more than 25 km. Using the revised age of the ash, the calculated minimum rate of crustal shortening to create an anticline with this relief in less than 667 ky is 7.3 mm/ky, with the inferred compression oriented 140°-320°. This orientation is about 60° to 70° from the inferred compresional vectors deduced from earthquake first-arrival sudies of two nearby quakes (017°-197° and 031°-211°; Astiz and Allen, 1983, events 5 and 6), from the strain-rate vector orientation described by Savage and others (1981), and from calculations based on the orientation indicated by this part of the left-lateral-slip Garlock fault itself. Geologic studies of other tectonic features found along the fault, west of the Lava Mountains, includes 10 additional examples of field relations requiring similar "anomalous" compressional directions, ranging in age from post(?) Mesozoic to Holocene(?).

Linear zones of en echelon cracks and fissures in the alluvium-covered southern part of Searles Valley (Fig. 2), at a site about 5 km north of the Garlock fault, were interpreted by Zellmer and others (1985) to be an an expression of stresses built up along the east-central part of the fault. Other mechanisms that are known to produce similar cracks and fissures seem not to apply in this instance. The zone of cracking is only 6 km from the Christmas Canyon seismic-creep-measuring station (Fig. 2), described by Astiz and Allen (1983) and Louie and others (1985), where no creep was detected along or near the fault over a period of several years. This may mean that the Garlock fault in this segment and to the east another 100 km, is locked by some mechanism, yet is accumulating strain that is unreleased along the fault itself but is released by creep along adjoining zones.

Benchmark lines that cross the east-central part of the Garlock fault that have been re-surveyed months to decades apart show no changes that would indicate 20th century vertical displacements on the fault. They do, however, seem to show that some of the anticlines and synclines in the area are still developing, including those described above that cross the fault yet indicate a stress field nearly at right angles to that tbelieved responsible for the Garlock fault (Smith and Church, 1980). Re-leveling surveys of five benchmark lines, established between 1906 and 1975, were used to measure changes in elevations with time. Lines 3 and 4 that cross the Garlock fault show uplift toward the south. The benchmarks in line 3, made only three months apart (February 1943 and April 1943), are spaced closely enough to show whether the uplift was on the fault itself. Although the surveys document gradual uplift toward the south along much of the line, no abrupt increase was found at the fault line. The benchmarks in line 4 also show gradual uplift toward the south, but the two benchmarks nearest the fault are 1 and 2 km from it, making them insensitive to any uplift on the fault. Line 4 does show, however, a 20 mm increase in elevation in the vicinity of the projected axis of the anticline that lies astride the Garlock fault, possibly indicating at least this much increase in anticline elevation in the 29 years between surveys.

Regional Studies And Data Compilations Relevant to The Garlock Fault

A geologic map compiled by Dibblee (1967) of the western Mojave Desert at 1:125,000 scale includes the west half of the Garlock fault. The report containing the map includes 1:62,500-scale geologic maps of 67 small areas, several of which show segments of the Garlock fault. Geologic maps of short segments of the fault, made by Dibblee and by others, are listed in Table 1.

At a scale of 1: 24,000, Clark (1973) mapped all of the scarps, graben, linear features, offset stream channels and ridges, and other geomorphic landforms he attributed to recent displacements along the entire Garlock fault. He plots numerous examples indicating recent activity along most of the length of the fault and shows that the evidence of movement along the fault is extremely variable, that several splays exist, that the fault zone width apparently varies from <1 m to ˜3,000 m, and several left-stepping discontinuities are present, the largest of which is the ˜3-km discontinuity in Fremont Valley (Fig. 2). He was unable to find geomorphic evidence of recent displacements along the fault in the final 7 to 8 km before the main trace of the Garlock fault reaches the San Andreas or the Death Valley fault zones that commonly are cited as being responsible for terminating the Garlock fault.

A gravity survey of the western Mojave Desert by Mabey (1960) shows that all but one of the several playa lakes in that area do not overlie the lowest part of their bedrock basins, showing that the tectonic processes that determined the bedrock configuration do not now control surface topography. The exception is Koehn Lake which lies in Fremont Valley, between the two en echelon strands of the Garlock fault, showing that the valley and its playa are in a graben, as later suggested by Clark (1973), Ayden and Nur (1982), Astiz and Allen (1983), and Louie and Qin (1991).

A reconstructed profile across Fremont Valley (Mabey, 1960, fig. 26) shows up to 3.5 km of low-density sediments between the east branch of the Garlock fault and the west branch, 3 km to its south. He also estimates about 1 km of low-density sediments underlie the 1.5-km-wide area between the north side of the graben and the sub-parallel El Paso fault which Dibblee (1952) determined to be the fault responsible for the uplift of the El Paso Range. On the south side of Fremont Valley, gravity data indicate the approximate location of the west branch of the Garlock fault that shows minimal geomorphic expression within or just northeast of Koehn Lake. Maybe (1960) named this concealed extension of the west branch of the Garlock fault the Cantil Valley fault; Cantil, a small settlement in Fremont Valley, is a locally used term for Fremont Valley.

The west branch of the Garlock fault (or Cantil Valley fault) is tentatively extended east into the Lava Mountains (G. Smith, 1964). Here, it crops out as a steeply dipping fault ˜3 km to the south of the active east branch of the Garlock fault. It has an estimated left-lateral displacement in the Lava Mountains of about 2 km, a north-side-up displacement of ˜200 to ˜300 m, and it is covered in one place by a Pliocene(?) volcanic flow, indicating that this part of the fault has been dormant since the flow solidified. The easternmost exposure of this fault, as mapped, shows it curving toward the north; for the next 1.5 km, it is concealed but may may well flatten and become the 30°- to 55°-south-dipping thrust that places plutonic rock over upper Miocene(?) sediments in an area adjacent to the presently active east branch of the Garlock fault.

The possible structural relations at the easternmost segment of the west branch of the Garlock fault (the Cantil Valley fault) in the Lava Mountains, are similar to those proposed by Hewett (1955) for the easternmost segment of the east branch of the Garlock in the Avawatz Mountains. There, he postulated that the Garlock fault curved southeast, flattened, and turned into the southwest-dipping thrust fault that is exposed in the northeastern Avawatz Mountains, placing Mesozoic plutonic rocks on Quaternary alluvial gravels. More recent detailed mapping in and north of the Avawatz Mountains (Troxel and Butler, 1979; Brady and others, 1980; Brady and Troxell, 1981; Brady and others, 1989) have shown that Hewett (1955) was nearly correct. The only significant variations are that it is the Mule Springs fault, a subparallel, south-dipping fault along the north side of the Garlock fault zone (Fig. 2), that curves to the southeast and grades into the exposed thrust, placing Mesozoic plutonic rocks on Quaternary gravels. Slickensides along its 45°-dipping sole show it to primarily have had reverse movement. This fault becomes the northern boundary of the plutonic mass that makes up much of the Avawatz Mountains, and it truncates the eastern end of the Garlock fault and

TABLE 1. GEOLOGIC MAPS SHOWING SEGMENTS OF THE GARLOCK FAULT. [MAPPING AT SCALES OF 1:125,000 OR LARGER.]

Location or quadrangle name (see Fig. 2)	Bounding longitude	Scale	Reference
Avawatz Mountains	116° 21'- 116° 49'	~1:25,000	Brady and others, 1980
do do	116° 15'- 116° 28.4'	1:24,000	Troxel and Butler, 1979
Quail Mountains	116° 48'-117° 00'	1:48,000	Muehlberger, 1954
Slate Range	117° 03'- 117° 17"	1:62,500	Smith and others, 1968
Searles Valley	117° 14'- 117° 28'	1:48,000	Smith, in press
Lava Mountains	117° 21'- 117° 36'	1:24,000	G. Smith, 1964
Randsburg quadrangle	117° 30'- 117° 45'	1:62,500	Hulin, 1925
El Paso Mountains	117° 39'- 117° 52.5'	1:24,000	Carr and others, 1997
Saltdale quadrangle	117° 45'- 118° 00'	1:62,500	Dibblee, 1952
Mojave quadrangle	118° 00'- 118° 15'	1:62,500	Dibblee, 1959
Tehachapi quadrangle	118° 15'- 118° 30'	1:62,500	Dibblee and Louke, 1970
Neenach quadrangle	118° 30'- 118° 45'	1:62,500	Wiese, 1950
Lebec quadrangle	118° 45'- 118° 52.5'	1:62,500	Crowell, 1952

two of the three strands of the Death Valley fault zone. Minimum reverse movement is 1.2 km. Another significant finding of those studies is that the fragment lithologies in the Owl Hole Spring Formation, a 16.3 Ma unit that crops out in the Noble Hills (in the acute angle between the Death Valley and Garlock fault zones), were derived from the same bedrock sources as modern alluvial gravels, showing that there has been at most only a few kilometers of lateral displacement on this part of the Garlock fault since their deposition in the middle Miocene.

In another regional compilation of gravity data north and south of the Garlock fault, three other bedrock basins, 1- to 2-km deep, lie adjacent to the Garlock fault along the south edge of Searles Valley (Jachens and Calzia, 1998). Two lie north of the fault and one lies to the south. They are nearly equidimensionsal and thus probably not graben, but their depths relative to the surrounding subsurface bedrock terrain suggest some form of pull-apart structures are involved.

Other gravity compilations that together include all of the Garlock fault are shown on the Bouger Gravity Map of California, Los Angeles Sheet (Hanna and others, 1974a), Bakersfield Sheet (Hanna and others, 1974b), and Trona Sheet (Nilsen and Chapman, 1971).

As noted above, gravity data of Mabey (1960) best indicate the location of the concealed eastern extension of the west branch of the Garlock. If the faults in the Lava Mountains (G. Smith, 1964) are correctly interpreted here as being the continuation of the Garlock (or Cantil Valley) fault, the origin of the 3.5-km-thick section of low-density fill in Fremont Valley can be more satisfactorily explained. The mapped relations of this fault in the Lava Mountains indicate that this part of it has been inactive or nearly so since Pliocene(?) time, but the line of crustal weakness it developed earlier probably localized the southern limits of the graben, as easterly displacements along the south side of the active east branch created a pull-apart depression. This depression is almost certainly more pronounced than it would have if the west branch of the Garlock fault had been simultaneously slipping left laterally, counteracting some of the easterly slip of the block south of the east branch. It is, therefore, the distribution of the active and inactive branches of the Garlock fault that accelerated and localized the pull-apart construction of the Fremont Valley graben.

Additional support for the concept that the Garlock fault is composed of more than one fault comes from three other sources. (1) Along the west branch of the Garlock fault in Fremont Valley, Clark (1973) located the easternmost geomorphic evidence of recent faulting at a point ~5 km southwest of Koehn Lake; along the east branch of the Garlock fault in this area, he located its westernmost linear trace about 3 km due north of that point; west of that point, along the southeast side of the Sierra Nevada (an 8-km-long series of sinuous, valley-side-down faults are interpreted here as normal, dip-slip faults, unrelated to the Garlock). (2) Astiz and Allen (1983, Fig. 2) plotted the most closely spaced clusters of epicenters on the entire Garlock fault along the trace of the Garlock north of Koehn Lake, yet they plotted only one or two epicenters further west that might line up with that trace, supporting the premise that the east branch of the Garlock is the active fault in the Fremont Valley area but that it is inactive or nonexistent west of those epicenters. (3) Astiz and Allen (1983, Fig. 4) also calculated the cumulative seismic moment recorded during the 1932-1981 period and plotted them as histograms, each bar representing 25-km-long segments of the fault. In the 75- to 100-km segment, approximately in the middle third of the distance between Tehachapi Valley and Koehn Lake (Fig. 2), the seismic moment is lower than in any of the other segments except near the fault's two ends (that they interpret as signs of the fault activity dying out); the moderate increase in seismic moment they plotted east of the easternmost distance marker appears from their Figure 2 to include much or all of its magnitude from faults that occurred along the Death Valley fault zone. This low seismic moment between the 75 and 100 km markers supports the proposal that the Garlock fault has a discontinuity in this zone. The eastern Garlock fault branch north of Koehn Lake shows little or no seismic activity west of the westernmost scarp located by Clark (1973), and the western branch of the Garlock fault south of Koehn Lake displays a low seismic moment relative to its more western parts.

It seems that the Garlock fault could be considered **two**, somewhat overlapping, en echelon intracontinental transform structures, probably results of the same set of regional stresses but possibly developed at different times. Both the western and eastern branches are characterized by west ends that show little or no evidence of recent activity (LaViolette, and others, 1980; Clark, 1973; Troxel and Butler, 1979), and by east ends that apparently terminate as thrust faults. However, both branches are tectonically active in their middle segments, and there seems to be no way to determine the ages of their first development, if different.

Geologic Explanations For The Garlock Fault in Terms of Western North America's "Big Picture"

Theoretical Studies

Two theoretical studies of the northwest-trending, right-slip faults, found throughout the southwestern part of the Mojave Desert may help explain their origin and the curvatures in the Garlock and San Andreas faults (Fig. 1). Garfunkle (1974) suggested that the terrain between the Garlock and San Andreas faults underwent counterclockwise rotation during Cenozoic time, giving the northwest-trending faults in that terrain the orientation we see now, generating right-lateral slip on them, and creating a ~20° counterclockwise change in the trend of the San Andreas fault for 200 km southeast of the apparent junction with the Garlock fault. The curvature of the eastern part of the Garlock fault was not addressed, possibly because the Blackwater fault, the northeasternmost fault of the northwesst-trending set in the Mojave Desert set, is about at the point where the strongest curvature of the Garlock begins (Fig. 2).

Garfunkle (1974) suggested paleomagnetic studies to test the rotation part of his hypothesis, but the several paleomagnetic studies that have been made of Mojave Desert terrain show a wide range in the amounts and directions of rotation, depending on the area studied (Burke and others, 1982; Ross, 1987; Wells and Hillhouse, 1989), many of which show clockwise rotatation. Another problem is that although Garfunkle's model produces properly oriented northwest-trending faults and could explain the zone of east-trending faults in the northeast corner of the Mojave block (his Fig. 10c), the now-northwest-trending faults would have had to exist prior to rotation, so the blocks would have established lines of weakness allowing them to slip past each other. At the time rotation started, those fault would have been oriented nearly north-south, requiring northeast-southwest-directed stresses to produce right-lateral displacement along them. To continue right-lateral activity into very late Cenozoic time, as field relations demonstrate, the source of this stress pattern would also have had to rotate to avoid having a nearly right-angle relation today. Such a process is problematical.

Cummings (1976, 1980) developed a model based on plasticity theory, using a modified Prandtl cell. Instead of parallel east-west edges to the cell, he modified it so two edges bounding the cell were on the northwest (like the Garlock fault) and the southwest (like the San Andreas fault), with the cell undergoing compression from those two directions; the east edge of this triangular model was unconstrained. To help match the configuration of the Garlock fault, he divided the northwest edge into two parts, approximately in the middle, and rotated the eastern half ~20° clockwise, making each half a slightly different cell. Compression normal to the northwest and southwest sides of the cell theoretically should produce two sets of curving stress trajectories, approximately normal to each other, that on failure would become faults. One set of these trajectories would produce right-lateral displacements and trend northwest; the other trajectories are considered unexpressed for theoretical reasons.

Cummings' model (1980, Fig. 8) would create northwest-trending faults by applying compression from the northwest, normal to the Garlock fault, and this would produce near-maximum shear stresses along the south edge of the fault. However, none of the northwest-trending faults in the Mojave Desert can be traced into the Garlock fault, reducing the probability that their origin depended on compression from that direction. Only in the Lava Mountains do Tertiary rocks crop out within a kilometer of the south side of the Garlock. There, the northwest-trending Blackwater fault, which forms the northeast edge of the northern Mojave-block fault cluster, can be traced into the Bedrock Spring Formation (very late Miocene), where it creates a structurally disturbed zone as close as 9 km south of the Garlock fault (G. Smith, 1964). It may have been offset by the northeast-trending Brown's Ranch fault zone (Fig. 2), but it does not reach and offset the post-very-late-Miocene, 19-km long Dome Mountain anticline that lies 2.5 km northwest of where the Blackwater fault is last well exposed, nor does it displace any of the more northerly post-very-late-Miocene volcanic rocks or the Garlock fault itself, which has a single trace in this area. On the other hand, evidence of compression from the northwest is provided by the Dome Mountain and Christmas Canyon anticlines, and the Teagle Wash and Pilot Knob Valley synclines, mentioned earlier,which are subparallel to the Garlock fault in the Lava Mountains area (G. Smith, 1964, 1991), plus nine other similarly oriented compression structures in areas adjacent to segments of the Garlock fault that lie west of the Blackwater fault (Smith, 1991 [table 2]).

Field Studies

Near the southwestern terminus of the Garlock fault, Hill and Dibblee (1953) suggested that the left-lateral Big Pine fault, on the southwest side of the San Andreas fault and ~10 km to the northwest, was the offset continuation of the Garlock fault, citing their similar trends, types of displacement, and other characteristics as evidence of their similarities. However, their paper was published before the large offset of the Garlock was realized. If their correlation was correct, the large displacements along the Garlock would have had to take place very rapidly or the separation of the two faults by the San Andreas would have had to be much slower than other evidence indicates. Neither explanation seems likely.

Davis and Burchfiel (1973) accepted the estimates of fault displacement by Smith (1962) and Smith and Ketner (1970) for the Garlock fault, and added four more possible correlations indicating approximately the same amount of offset. Because four of those six correlations involving rocks and structures 30 km or less from the eastern end of the fault, where it joins the Death Valley fault zone and deflects it to the east only a few kilometers, they question the correctness of the "traditional" interpretation of faults in which one must offset the other, or be offset, wherever two meet. Instead, they propose that the Garlock fault should be considered an intracontinental transform structure, partly because it divides two terrains characterized by very different structural characteristics, the Basin-and-Range and the Mojave Desert provinces, and partly because transform faults, by definition, turn into non-lateral-slip faults at their ends and have the most displacement in their central sections. The large displacement along the Garlock fault is attributed by them to the westward extension of the block north of the fault caused by block faulting in and west of the Death Valley area, possibly increasing the apparent displacement all the way to the eastern frontal fault of the Sierra Nevada. They also consider the bend in the trace of the San Andreas fault just northwest of its apparent junction with the Garlock to be a result of the extensional nature of the Basin-and-Range province, caused by the east-dipping subduction zone along the western edge of the North American plate.

Glazner and Loomis (1984) and Glazner and Schubert (1985) investigated the relation between faulting and volcanism in the southwestern United States and their underlying controls. Glazner and Loomis concentrated on the possible topographic effects in the North American plate that would have resulted from subduction of the Farallon plate. The segment of the subducted Farallon plate south of the Mendocino transform fault was at a higher elevation than the segment north of the fault, and it would theoretically have produced topography in the North American plate that was 500 to 1,000 m higher than that north of the transform fault. Thus, after the south plate was subducted beneath the North American plate, it created a 70- to 120-km-wide topographic high in the crust above and south of the easterly extension of the Mendocino fracture zone. The Mendicino fracture zone would have passed beneath the latitude range of the Garlock fault between ~19.5 and ~17.5 Ma (early Miocene). That this happened is supported by lithologic evidence from early and middle Miocene sediments that show a transition from west- to north-draining basins and from marine to continental deposits as the "wave" of higher topography approached. The authors also suggest that the east-west fractures caused by this flexuring may be responsible for initiating east-west faulting in southern California.

However, erosion during this period of elevation would not explain the absence of early Tertiary deposits south of the Garlock fault. If the Farallon plate's "northward" movement was at ~3 cm/yr, as Glazner and Loomis (1984) suggest, the Mojave area south of the Garlock fault would have remained a high-elevation area for only ~2.5 to 5 my. This is about the length of time since early Pliocene, yet many sedimentary deposits of Miocene age, representing an additional ~19 to 22 my, are preserved today in this area. To remove the several kilometers of pre-middle Mesozoic metasedimentary and metavolcanic rocks, as well as any rocks deposited in early Cenozoic time, a much longer period of uninterrupted erosion, or a long period of erosion of the Mojave block in earlier Cenozoic time, seems likely.

An appraisal of the contemporary tectonic processes in California and Nevada led Hill (1982) to the conclusion that a continuing stress field generated by the relative motion of the Pacific and North American plates is basically responsible for the tectonic patterns we see today. He concluded that the greatest stress fields nominally are from north-south directions and the the least stresses nominally are from east-west directions, and the modern displacements we see are largely controlled by crustal inhomogenieties or by pre-existing fault pattern regimes. These would include earlier periods of spreading (as in the Basin-and-Range province), converging (as in the Transverse Range province), and rotation (as in the San Andreas and Garlock faults). Maximum shear stresses are required for converging (thrust) configurations and minimum shear stresses are required for spreading (normal fault) configurations; intermediate stresses are required for wrench (rotation) configurations. These basic configurations acted on at least the brittle part of the crust and they became slowly modified through repeated faulting. Many of the presently observed fault displacements can be explained by this general relationship, but inhomogeneities in the properties of the crust probably caused some deviations from their expected character.

Dokka and Travis (1990), in a paper concerned primarily with the northwest-trending faults of the Mojave Desert, explain the curvature in the east half of the Garlock fault by considering it an orocline produced by right-lateral shear along the zone between the Blackwater fault and the "Eastern California shear zone," a loosely defined set of northwest-trending, right-lateral faults in southeastern California. This might explain the 25° of clockwise rotation of the dike rocks in the Granite Mountains relative to the dikes in the Argus Range and Spangler Hills (Smith, 1962), but it seems that the ~60 km of right shear needed to transform an originally straight Garlock fault, trending at N. 45° E., to the present configuration should have created a clearer geologic record of late Cenozoic compression. Admittedly, the bulging outlines of the southern ends of the Slate Range and Owlshead Mountains (Fig. 2), immediately north of the Garlock fault, do suggest compression from the north, but the apparent lack of east-west-trending folds and thrust faults of Tertiary age in the northern part of the eastern Mojave Desert makes oroclinal folding of the Garlock fault less probable; the scattered examples of such structures noted by Smith (1991) all lie west of the most pronounced curvature of the Garlock fault.

Summary

Even though the Garlock fault provides several lines of evidence indicating 40 to 65 km of left lateral displacement, the last several kilometers at both ends of the fault are now tectonically quiescent. Neither of the "terminating" faults, the San Andreas and the Death Valley fault zones, are offset significantly by the Garlock fault. Although the San Andreas fault has a 20° change in trend about 50 km northwest of the apparent junction with the Garlock fault, there is no visible displacement at the projected intersection. The Death Valley fault zone shows minimal change in trend at the termination of the Garlock fault, although the geologic relations in this area are very complex. These characteristics seem best explained by considering the Garlock to be an intracontinental transform structure that separates the Basin-and-Range province from the Mojave Desert province; the Basin-and-Range province is generally considered to be an "extensional" (or "spreading") province, while the Mojave Desert province shares characteristics of both a "rotational" (or "wrench") province and a "compressional" (or "converging") province.

Studies of late Quaternary faulting observed in trenches show that except near the two ends of the Garlock fault, faulting has continued nearly into historic time. Earthquake recurrence rates, based primarily on ^{14}C- dated materials from trench sites, show recurrence rates ranging from 200 to 3,000 yrs and creep rates of 1 to 11 mm/yr, but most minimum rates are nearer 3 mm/yr.

The Garlock fault serves to decouple the stresses that cause the terrain to its north to be seismically more active than the terrain to its south, but along the Garlock fault, the seismicity of the east and west halves differ, as do the depths to the bases of their hypocenters. The western half has small but persistent seismic activity and measurable creep, whereas the eastern half has relatively rare seismicity and no creep. The dividing line between these characteristics lies in Fremont Valley, where geologic mapping shows the fault making a 3-km-wide en echelon step to the left. Gravity surveys show as much as 3.5-km of low-density fill along the south side of the east branch in Fremont Valley, where the east and west halves of the Garlock overlap; this great depth, and the fault geometry of the two branches of the Garlock, indicate that the valley is a pull-apart basin. This area also is just south of the junction of the Garlock fault and the Sierra Nevada fault zone (Figs. 1 and 2), suggesting a possible interaction between them.

Geomorphic evidence shows that late Quaternary displacements end on the west branch and begin on the east branch in Fremont Valley, and 50 years of seismic data indicate a reduced seismic moment in the same area. At their western ends, both branches show no geomorphic evidence of recent activity along their main fractures although they may have transfered existing stresses to other nearby fractures. At their eastern ends, both branches may terminate as thrust faults. For some purposes, the Garlock should be considered as two, en echelon, intracontinental transform structures, probably reflecting the same tectonic stresses, although possibly they were not initiated at the same time.

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A Geophysical Analysis of the Garlock Fault

Robert C. Jachens and **James P. Calzia**, *U.S. Geological Survey, Menlo Park, CA 94025*

ABSTRACT

Analyses of gravity and magnetic anomalies associated with the Garlock fault and surrounding areas are used to infer three-dimensional geometries of the fault and related structures and to constrain the total offset and evolution of slip on the fault. Modeling of a gravity profile across the western part of the fault and a magnetic profile across the central part of the fault indicates that the fault is nearly vertical and extends at least to mid-crustal depths of about 12-16 km. We propose that a total left-lateral offset of 47 km is accommodated across the central part of the fault, based on matching fault-truncated magnetic anomalies north and south of the fault. In some locations, the 47 km of offset is partitioned between the currently active strand of the fault and some earlier, now-abandoned subparallel strands. Our preferred interpretation has the 47 km of offset across the central part of the fault decreasing linearly to 0 km at the eastern end of the fault. At least two pull-apart basins caused by left-stepping fault strands are identified from the gravity data, the well-known one beneath Cantil Valley and another 50 km to the east. Large, deep basins in the westernmost Mojave Desert immediately south of the Garlock fault may reflect large-scale distributed extension of the northwestern Mojave block, extension that could help explain the puzzling possible decrease in total offset across the Garlock fault from 47 km in the central part to no more than about 12 km in the western part.

Introduction

Although the geology and kinematic history of the left-lateral Garlock fault have been studied for many years, information about its subsurface characteristics is limited. Knowledge of three-dimensional fault geometry can contribute to improved understanding of the nature of major faults and their evolution. Such information is available locally along the Garlock fault based on geophysical investigations (e.g. Mabey [1960]; Plescia and Henyey [1982]; Cheadle et al. [1986]; Serpa and Dokka [1992]; Blakely et al. [in press]), but we are not aware of any systematic geophysical study that examined the entire fault. Here we analyze gravity and aeromagnetic anomalies along the entire length of the Garlock fault and use these data to infer information about the deep structure, total offset, and offset history of this important fault.

Geophysical data sets are now available with which to investigate the deep structure of the Garlock fault and to examine its near-surface characteristics in areas where they are concealed beneath young deposits. Regionally consistent compilations of gravity data covering the entire Garlock fault at data spacing of 2-10 km have become available within the past 15 years (Oliver et al., 1980; Roberts et al., 1990). The Garlock fault is covered entirely by aeromagnetic data collected continuously along flightlines spaced 1.6 km apart or less (U.S. Geological Survey, 1970). Disparate aeromagnetic surveys recently were compiled and numerically merged into a coherent, internally consistent database that approximates the magnetic field at a height of 305 m above the ground surface (Roberts and Jachens, 1993). These regional databases and some products derived from them constitute the basis of the present analysis. Our analysis based on geophysical data offers quantitative predictions and explanations as working models that need to be tested quantitatively against detailed geology.

Gravity Anomalies

Gravity data from Roberts et al. (1990) are presented in the form of *isostatic residual gravity anomalies* in figure 1. We chose this form of gravity anomaly rather than the more commonly encountered *Bouguer gravity anomaly* because the isostatic reduction process removes the regionally dominant gravity effects of deep-seated density deficiencies that buoyantly support the topography. Compared to the *Bouguer gravity anomaly*, the *isostatic residual gravity anomaly* more clearly

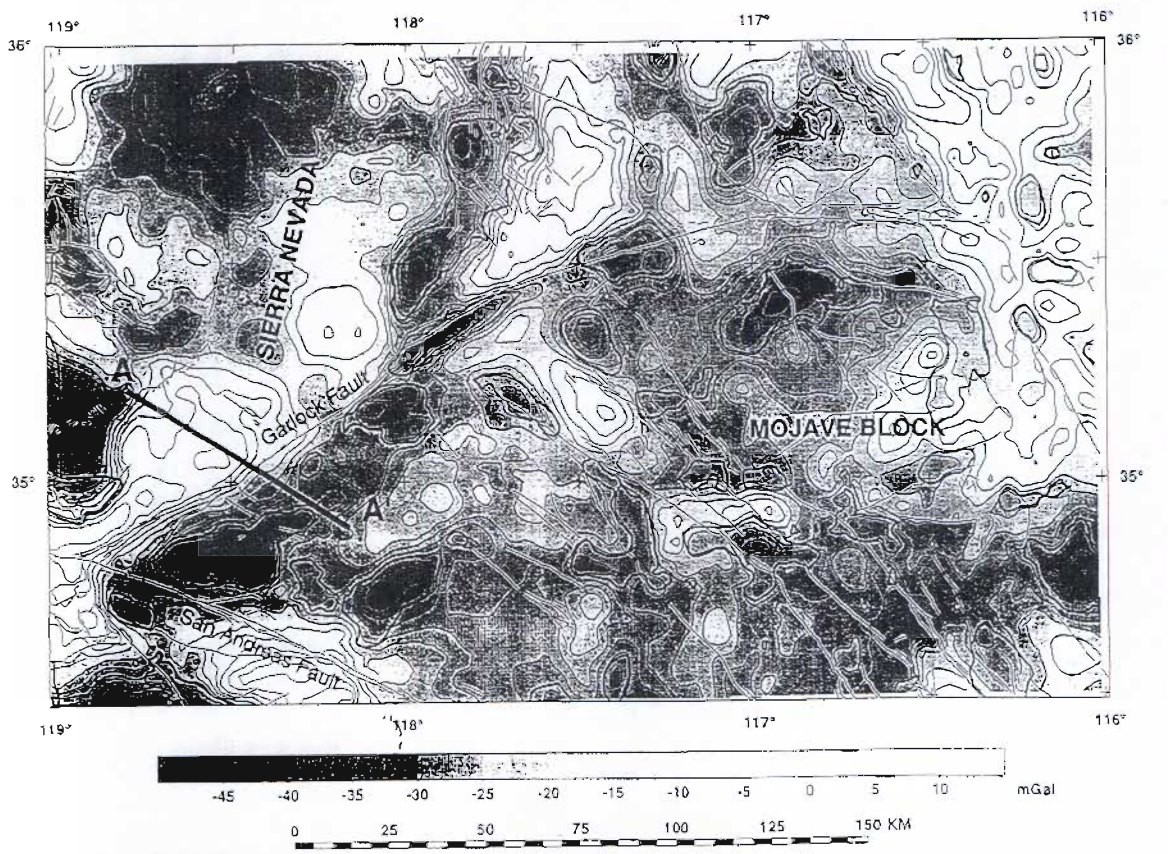

Fig. 1. Map showing isostatic residual gravity anomalies over the Garlock fault and vicinity (after Roberts et al., 1990). Faults (shown in relief-black on white) from Jennings (1994). Heavy lines show 600 m depth contour of basins (from figure 3). Profile A-A' was modeled to estimate depth extent of fault. Contour interval 5 mGal.

reflects density distributions in the middle and upper crust (Jachens and Griscom, 1983), those of primary interest in this study.

On a regional scale, the Garlock fault separates generally low gravity values over the Mojave Desert to the south from higher values over the Sierra Nevada and Basin and Range province to the north (fig. 1). The abrupt gravity transition across much of the western half of the fault reflects a sharp density boundary, in contrast to the implied more gradual transition across parts of the eastern half of the fault. The regional gravity picture is complicated by the presence of some gravity lows caused by basins filled with low-density Cenozoic deposits. However, these basin anomalies tend to complicate the gravity field only locally, and stripping them away does not materially change the regional gravity pattern shown in figure 1 (see the "basement" gravity field in Saltus and Jachens, 1995, which is free of the basin gravity anomalies). Thus the regional gravity in figure 1 mainly reflects lateral density variations of the pre-Cenozoic basement rocks. The juxtaposition of basement rocks with different densities across the Garlock fault could result from 1) strike-slip offset that moved rock bodies with different densities laterally against each other, 2) differential uplift that likely brought originally deeper, denser layers up against shallower, less dense bodies, or 3) the initial rupturing of the Garlock along crustal-scale flaws that are also reflected as density boundaries (e.g. intrusive contacts or ancient sutures). We will examine each of these possibilities later in the paper.

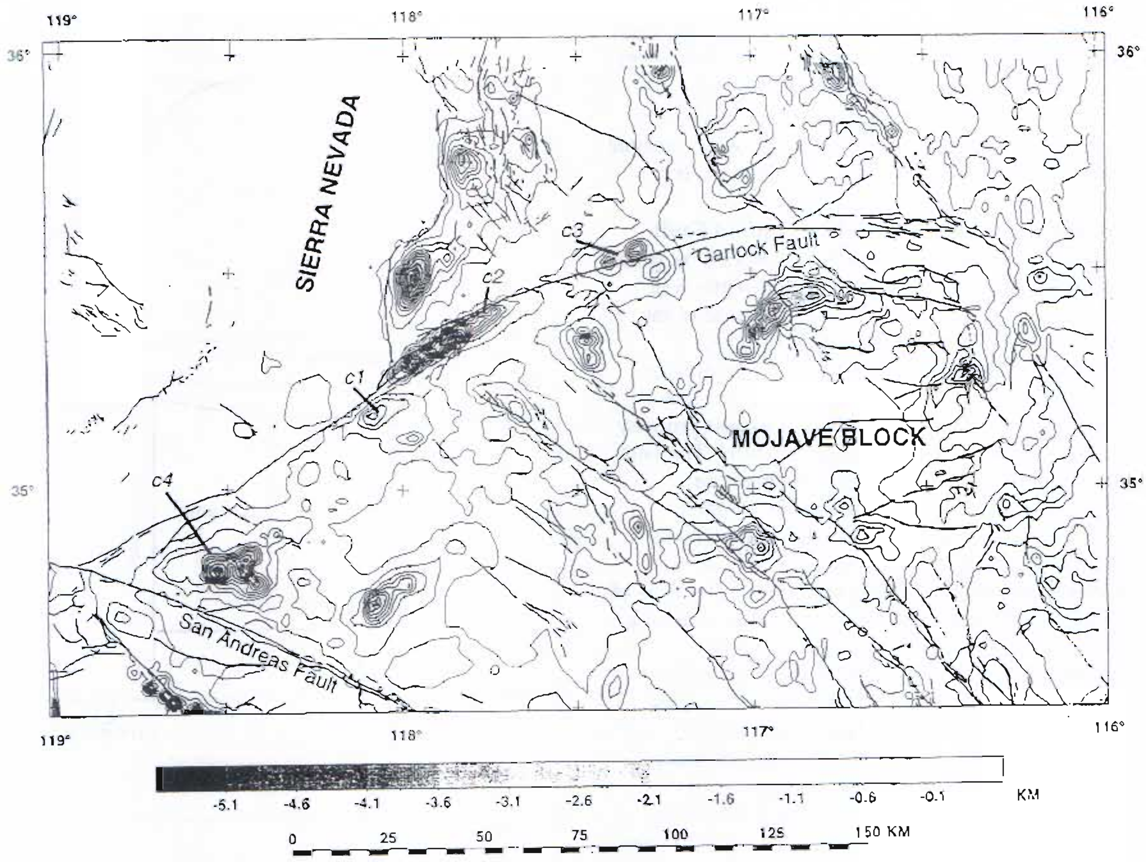

Fig. 3. Map showing the depth of Cenozoic basins in the vicinity of the Garlock fault inferred from gravity data constrained by outcrop geology and drill-hole data (after Saltus and Jachens, 1995). Faults from Jennings (1994). Alphanumeric symbols denote basins discussed in the text. Contour interval 0.5 km.

Magnetic Anomalies

The magnetic anomalies shown in figure 2 (pocket) from the compilation of Roberts and Jachens (1993) are presented here for the first time. These anomalies reflect the distribution of magnetic minerals, mainly magnetite, in the crustal rocks surrounding the Garlock fault. Although some Cenozoic volcanic rocks in the region produce magnetic anomalies, they tend to occur as thin sheets with highly variable magnetic properties and, as such, rarely produce regionally coherent magnetic anomalies. Almost all of the large magnetic anomalies shown on figure 2 are caused by magnetic minerals in pre-Cenozoic igneous or meta-igneous crystalline rocks. Sedimentary rocks of all ages typically are non-magnetic.

The small, aligned black "+" symbols (fig. 2) represent the locations of inferred boundaries across which rock magnetizations laterally change abruptly. These boundaries were determined automatically from the magnetic map by a gradient analysis procedure developed by Cordell and Grauch (1985) and implemented by Blakely and Simpson (1986). We have included these inferred boundaries on figure 2 because magnetization boundaries often coincide with lithologic contacts or tectonic structures.

Along most of the eastern two thirds of the Garlock fault, strongly magnetic bodies of Mesozoic plutonic rock to the north are truncated at the fault. For example, magnetic boundaries ***a2-a6*** (fig. 2) coincide with strands of the Garlock fault. Even at the extreme western end of the fault where only weakly magnetic bodies exist, at least one body to the north is sufficiently more magnetic than the rocks across the fault that it produces a recognizable magnetic boundary (***a1***) coincident with the fault and nearly 20 km in length. Curvilinear magnetic boundaries that parallel but are displaced from the fault (***b1-b4*** on figure 2) probably delineate buried, unmapped, or abandoned fault strands. Magnetic rock bodies truncated at the Garlock fault likely have displaced cross-fault counterparts that can be used to estimate the total offset on the fault, because the Mesozoic rocks are older than the Tertiary age of the fault (see Powell, 1993 for a review of age estimates).

Fault-related Basins

Basins frequently form along strike-slip faults (Christie-Blick and Biddle, 1985) and the basin shapes contain information about the nature of the slip and the slip history of the fault. Figure 3 shows the distribution and shapes of Cenozoic sedimentary basins in the vicinity of the Garlock fault (after Saltus and Jachens, 1995) inferred from an inversion of gravity data constrained by outcrop geology and drill-hole data. In this inversion, the gravity field is iteratively partitioned into a component caused by the low density Cenozoic deposits that fill the basins and a component caused by lateral density variations within the pre-Cenozoic basement rocks. The "basin" component is then interpreted in terms of the thickness of basin fill.

At least three basins (***c1-c3*** on figure 3) are aligned along the Garlock fault. The two western basins lie south of the main strand of the fault whereas the eastern basin straddles the fault and is deepest and best developed north of the main strand. The large, deep basin (*c4*) crudely parallels the westernmost Garlock fault located ~7 km to the north, and possibly is related to the triangular tectonic junction of the San Andreas and Garlock faults.

Attitude and Depth Extent of the Garlock Fault

Gravity or magnetic anomalies produced by large rock bodies truncated at a fault, if sufficiently well defined, can be used to estimate the attitude of the fault and its minimum depth extent by direct modeling. Modeling can only provide a minimum estimate of the depth extent because the anomaly only reflects the depth to which the physical property contrast (density or magnetization) extends. The fault could extend deeper, into a region where the properties are similar on both sides of the fault.

The Garlock fault is best defined in the gravity field (fig. 1) along its westernmost 50 km, and we modeled this anomaly along profile A-A'. Densities of basement rock samples from the region surrounding A-A' fall naturally into three groups, closely corresponding to the character of the local gravity field. Densities of 82 samples from the low gravity region south of the Garlock fault (westernmost Mojave Desert within 30 km of the Garlock fault) average 2.66 g/cm^3. In contrast, densities of 53 samples from the high gravity region of the Sierra Nevada extending ~10 km north of the Garlock fault average 2.76 g/cm^3. 64 samples from the Sierra Nevada region more than 10 km north of the Garlock fault (where the gravity tends to drop off from its maximum values) have an average density of 2.72 g/cm^3. Using these density values and an average density of 2.26 g/cm^3 for the Cenozoic basin deposits (an assumption that does not influence the main features of the model), the gravity model shown in figure 4a was constructed. The important features of this model are that, at least along profile A-A', the Garlock fault is vertical and extends to a minimum depth of about 16 km beneath the land surface. Models with a greater contrast in density across the fault would require a smaller depth extent for the density boundary. However, of the models examined, those that both fit the shape of the observed gravity data and have densities that are in reasonable accord with the sample measurements all produce a near-vertical Garlock fault that penetrates at least to mid-crustal depth. Given the generally consistent shape and amplitude of the gravity anomaly associated with the westernmost Garlock fault, the main features of the model shown in figure 4a probably characterize the western 50 km of the fault.

The Garlock fault is best defined magnetically between 117.5º W and 118º W because here a large, strongly magnetic body north of the Garlock fault is truncated at the fault and juxtaposed against a non-magnetic or, at most, very weakly magnetic body to the south (fig. 2). We modeled the deeper parts of this anomaly along profile B-BŌ (fig. 2) because detailed magnetic data collected at a height of 120 m above the ground surface were available along this profile (U.S. Department of Energy, 1979). We did not have any magnetic property measurements from

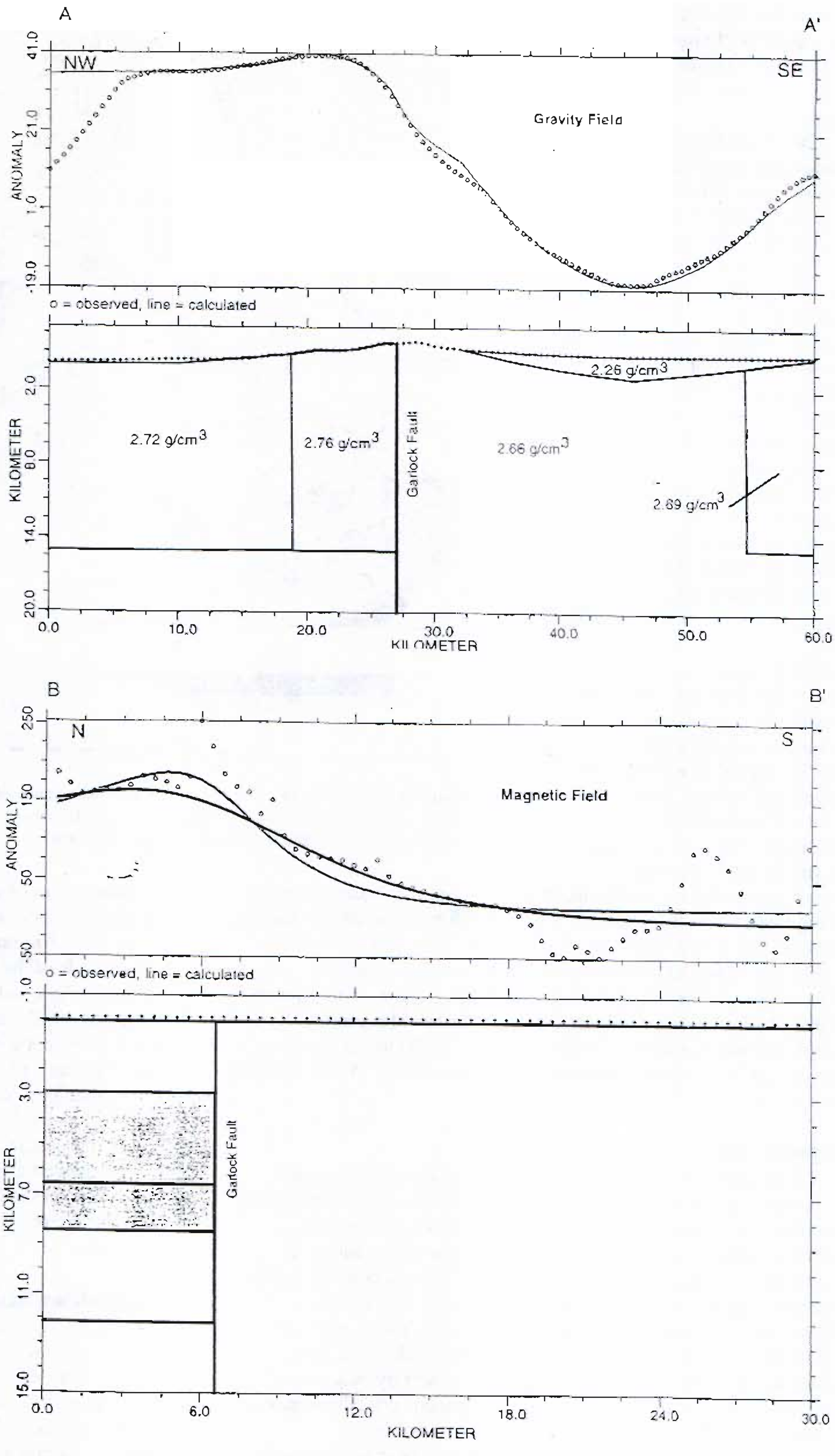

Fig. 4a. Gravity model along A-A' (fig. 1). Profile was scaled to approximate projected distances for a profile normal to the fault trace. Body with density of 2.72 g/cm^3 extends 20 km NW of km 0. **4b**. Magnetic model along B-B' (fig. 2). Profile was scaled to approximate projected distances for a profile normal to the fault trace. The critical part of the observed data that reflect the deep parts of the magnetic body lie between km 9 and km 18. Short wavelength magnetic anomalies between km 6 and km 9 are caused by small near-surface magnetic sources and were not modeled. Gray line is calculated anomaly from shallow magnetic body (shaded), which is too highly curved to satisfy the observed data. Only when the magnetic body gets deep (unshaded) does the calculated anomaly (black line) become straight enough to fit the observed data. Magnetic bodies extend 20 km N of km 0.

outcrop samples on which to base the modeling, and such sampling, even if it were available, tends to be much less reliable than density sampling. Instead, we depended on the characteristic shape of the magnetic anomaly (the long, smooth southward decline of the anomaly amplitude) to control the model results and to provide us with estimates of the fault parameters at this location. Based on numerous models that fit the observed anomaly and retained a magnetization within a reasonable range for Mesozoic plutonic rocks of this region, we conclude that the anomaly is best fit by a flat-lying, buried tabular body 5-7 km thick with a bottom at least 12 km beneath the surface (fig 4b). Bodies with bottoms at shallower depth produce anomalies that are too strongly convex upward to satisfy the observed anomaly (see example in fig. 4b). As was found in the gravity modeling, the magnetic model also suggests that the Garlock fault here is approximately vertical to the depths of the base of the model. Again, this depth estimate is a minimum, both because the fault could extend below the base of the model and because bodies with bottoms deeper than 12 km were found during the modeling that could also satisfy the observed magnetic measurements. The regionally consistent shape of the magnetic anomaly surrounding profile B-B' suggests that the modeling results apply to the entire fault between 117.5° W and 118° W.

Total Offset

Estimates of total left-lateral offset across the central and eastern reaches of the Garlock fault range from 48 km to 74 km, with the most reliable estimates falling between 48 and 64 km according to Powell (1993). The large spread in estimates results from the scarcity of piercing-point features in the geology surrounding the fault, combined with the fact that much of the older geology is covered by young alluvium near the fault which makes it difficult to follow piercing-point features all the way to the fault trace. In situations where geologic constraints can be used to narrow the amount of possible offset, magnetic data often can refine the estimates because alluvial cover is transparent to magnetic anomalies. Therefore, piercing-point features that produce magnetic anomalies can be followed to the fault trace, independent of the presence of alluvial cover.

North of the Garlock fault, the large magnetic body that is truncated at the fault along ***a3*** (fig. 2) has a western magnetic edge (***d*** on figure 2) that trends normal to and terminates at the fault. Geologically, this edge corresponds to the contact between magnetic Mesozoic plutonic rocks to the east and nonmagnetic Paleozoic sedimentary rocks to the west (Jennings et al., 1977). The offset counterpart to this magnetic edge should be found south of the Garlock fault provided that uplift and erosion of the Mojave block has not been so severe since the inception of faulting that the Paleozoic rocks have been eroded away. If found, the southern counterpart to the magnetic edge can be used to estimate the total offset across the central reach of the fault.

South of the fault, east of ***d***, and generally within the offset constraints imposed by the geology, we identify possible counterparts to the magnetic edge ***d*** at offset distances of 47 km (***d'***), 60 km (***d''***) and 66 km (***d'''***) (fig. 2). The sources responsible for all three of these magnetic anomalies are concealed beneath young alluvium and/or volcanic rocks in the vicinity of the fault, so no strong geologic arguments can be invoked to decide which, if any, of these features correspond to ***d***. Although none of the three possible offsets provides an overwhelmingly convincing match of detailed magnetic features north and south of the fault, we will argue that the 47 km offset constitutes the best choice considering both the magnetic anomalies and information resulting from an analysis (in the next section) of the fault-related basins, especially when non-uniform slip is permitted on the eastern part of the fault. This offset for the central part of the fault also is in accord with the conclusions of Smith and Ketner (1970) who correlate the metasedimentary rocks in Pilot Knob Valley near ***d'*** with the Paleozoic rocks immediately west of ***d***.

Basin Analysis — Implications For Offset

The deep, linear basin beneath Cantil Valley (***c2*** on figure 3) long has been identified as a classic example of a pull-apart basin caused by a left step in a left-lateral strike-slip fault (Aydin and Nur, 1982). Movement of the Mojave block eastward past the left step in the trace of the Garlock fault (the west end of the Cantil basin) creates a region of extension and collapse east along the fault that lengthens as movement along the fault continues. For a simple basin, its length reflects the total offset accommodated across the fault segments that define the left step. Thus the length of the Cantil basin implies a left-lateral offset of at least 40-45 km, reasonably close to the estimate of 47 km suggested by the distance between ***d*** and ***d'***.

The 25-km-long basin ***c3*** (fig. 3) probably had a similar origin, even though the current strand of the Garlock lies south of the deepest, most linear part of the basin. For this basin to have resulted from a left-step geometry of the fault trace, at least 25 km of left-lateral movement must previously have been accommodated on a strand located about 5 km north of the current strand, along the north edge of the basin. This strand would have curved south and merged with the current strand at ~ longitude 117.5° W., thus forming the left step. The magnetic data define a linear magnetic boundary parallel to, but 3-5 km north of the current Garlock fault (***b4*** on figure 2). The westward projection of this magnetic boundary lies along the north edge of the basin.

We propose that this magnetic boundary and its westward extension along the north edge of the basin is an early strand of the Garlock fault across which ~25 km of offset took place, thus leaving at least 22 km of offset to be accommodated on some other strands, possibly the current fault. Added support for this proposal can be seen by closing this basin through palinspastic restoration of 25 km of offset which unites the now isolated sliver of magnetic rock located between strand ***b4*** and the current Garlock fault with the large magnetic rock body to the west.

The origin for basin ***c3*** proposed in the preceding paragraph does not readily explain the cause of that part of the basin south of the current Garlock fault, although this part might have resulted from local collapse of northern edge of the Mojave block into the extensional zone formed in the wake of the left-step. If so, then the minimal lateral offset of the south part of the basin relative to the north part (~10 km) suggests that the current strand of the Garlock fault only accommodates 10 km of left-lateral offset, at least during the period following the formation of basin ***c3***. This leaves about 12 km of offset to be accommodated either on the current strand prior to the 25 km of offset on strand ***b4***, or on some other strand not presently recognized.

The shallow, oblong basin (***c1*** on figure 3) might also have originated as a pull-apart basin resulting from ~15 km of movement on a former strand located west of the basin and about 10 km south of the current Garlock fault. The magnetic data show a possible magnetic boundary (***b1*** on figure 2) that may reflect an older south strand of the Garlock fault, but its spatial coincidence with the expected location of an earlier left-step strand is not particularly good. Additional evidence will be required to test the origin of basin ***c1*** as a possible pull-apart basin associated with a left-step in the Garlock fault.

Discussion And Speculation

We now return to the discussion of total offset across the central and eastern reaches of the Garlock fault. Figure 5a (in pocket) shows a palinspastically restored magnetic map based on a total sinistral offset of 47 km (***d*** to ***d'***) and applies only to the fault east of longitude 118° W., that is, east of the Sierra Nevada. For the fault between the east edge of the Sierra Nevada and about ***d'''***, all 47 km is accommodated on the current strand of the Garlock fault. East of ***d'''***, the reconstruction accommodates 25 km of offset on a proposed strand that lies along the north edge of basin ***c3*** and includes magnetic boundaries ***b3*** and ***b4***. The remaining 22 km is placed on the current strand to the south. In this reconstruction, all 47 km of offset east of ***b4*** is accommodated on the Mule Springs fault (***a5***), although recent evidence indicating that this fault is a thrust fault may require us to seek an alternative on which to accommodate the 47 km of strike-slip offset.

This reconstruction of the slip on the Garlock fault brings into alignment a number of magnetic features north and south of the fault and reasonably accounts for the two most important basins immediately adjacent to the fault. As discussed previously, the restored position of the isolated magnetic sliver between ***b4*** and ***a4*** is against the south edge of the large magnetic rock body partly bounded by ***a3*** and ***b3***. Also, the pattern of magnetic highs south of the fault that abut the fault from ***d'*** east for 25 km roughly matches a similar pattern of anomalies north of the fault and east from ***d***. Finally, the magnetic low south of the Mule Springs fault (***a5***) is restored against a similar width magnetic low east of ***b3***.

This reconstruction apparently leaves two strong, fault-bounded magnetic bodies unmatched across the fault–the body bounded on the south by the El Paso fault (***b2***) and the body bounded on the north by the Leach Lake fault (***a6***). The magnetic low south of the El Paso fault probably is the result of down-drop and burial of the magnetic basement sliver beneath basin sediments, rather than a lack of magnetic basement rocks in this area. Thus, this feature is not necessarily in conflict with the present reconstruction. The lack of a magnetic counterpart north of the magnetic rocks bounded by the Leach Lake fault (***a6***) is not as easily explained because there is no significant basin north of the fault at ***a6*** (see fig. 3). Restoring an additional 20 km of left-lateral offset would place magnetic rocks across the fault from this body, but would degrade the cross-fault matches of the other magnetic anomalies. In this reconstruction we can see no reasonable Garlock fault-related explanation for boundary ***a6*** unless the Leach Lake fault originally broke along a pre-existing intrusive boundary that was also a magnetic boundary.

A consequence of the reconstruction shown in figure 5a is 47 km of eastward thrusting of the Mojave block over rocks east of the eastern limit of the Garlock fault, rocks that would occupy the "white" gap in the eastern part of figure 5a. An alternative model and one that avoids the need for such large-scale thrusting is to produce the strike-slip offset across the Garlock fault by having the region north of the fault undergo greater east-west extension than the region to the south, with no extension (or at least equal extension) occurring at the east end of the fault (Hamilton and Myers, 1966; Davis and Burchfiel, 1973). Figure 5b shows the magnetic map palinspastically restored according to the differential extension model with the following assumptions: 1) the east end of the Garlock fault is fixed; 2) 47 km of left-lateral offset on the central Garlock fault is restored by aligning ***d*** and ***d'***; 3) the decrease in slip from 47 km at ***d*** to 0 km at the east end of the Garlock fault is distributed linearly between the two points (equivalent to an east-west contraction to 60% of its pre-restoration width); and 4) all fault strands accommodating offset are the same as those used to construct figure 5a, except at the extreme eastern end where figure 5b uses the Leach Lake fault (***a6***) rather than the Mule Springs fault (***a5***).

Compared to the reconstruction shown in figure 5a, that shown in figure 5b provides a better cross-fault match of the most important magnetic features and leaves fewer problems either unresolved or explained by complex kinematic models. First, no major thrusting is required at the east end of the Garlock fault. Second, the major zone of magnetic rocks between ***d*** and just east of ***b3*** restores opposite a similar width zone between ***d'*** and just east of ***d'''***, and the adjacent magnetic low zone from just east of ***b3*** to ***a4*** lines up opposite an equivalent magnetic low zone between just east of ***d'''*** and ***a6***. The cross-fault match of these features is far less satisfactory in figure 5a. Third, the extensive magnetic zone just east of ***a4*** lines up against the magnetic body bounded on the north by the Leach Lake fault (***a6***), a body that has no cross-fault counterpart in figure 5a. Finally, in this reconstruction the 25 km of offset needed to close basin ***c3*** and the additional 10 km of offset on the current strand of the Garlock fault implied by the relative offset of the southern part of basin ***c3*** (fig. 3) accounts for all the slip at the location of ***c3***, in contrast to the additional 12 km of offset that needs to be somehow accommodated in the reconstruction in figure 5a (see discussion in section "Basin Analysis–implications For Offset").

Powell (1993) reviews the evidence for offset on the western reach of the Garlock fault and makes a reasonable case for total offset of no more than about 12 km. He concludes that if the 12 km estimate is correct, then the decrease in total offset across the fault (from at least 48 km in the east to 12 km in the west) requires extension in the westernmost Mojave Desert to accommodate the difference in displacement. The large, deep basin ***c4*** (fig. 3) and surrounding basins in the westernmost Mojave Desert are consistent with major distributed extension of the western Mojave block and thus provide a reasonable explanation for a westward decrease in offset across the Garlock fault. Furthermore, the 35-40 km along-fault length of basin ***c4*** is quantitatively compatible with the required pull-apart-like distributed extension of the western Mojave Desert.

The reconstructions of slip on the Garlock fault presented above fail to account directly for the marked differences in the regional gravity fields north and south of the fault (fig. 1). No distribution of slip in the range of 48-64 km (or less in the west) can account for the marked change from high gravity north of the fault to low gravity south of the fault. This striking contrast persists and is even enhanced when the low gravity values over basins are discounted (see basin outline on figure 1). For the western Garlock fault, the cross-fault gravity contrast can be explained by major uplift of the southern Sierra Nevada (Ross, 1989) relative to the adjacent Mojave block. This uplift resulted in the juxtaposition of dense, formerly middle-to-lower crustal rocks of the Sierra Nevada batholith against less dense upper-crustal batholithic rocks of the Mojave Desert. In contrast, east of the Sierra Nevada front the presence of isolated patches of pre-batholithic sedimentary and metavolcanic rocks both north and south of the fault suggests that differences in uplift north and south of the fault are small. Therefore, another explanation for the gravity contrast is needed. One possibility is that the cross-fault regional gravity contrast in the vicinity of the central and eastern Garlock fault predates the fault, and that the Garlock fault originally broke along a regional-scale Mesozoic magmatic terrain boundary that was, and continues to be, reflected in the gravity field. Another possibility is that, at least for the differential extension model, dike intrusion or underplating accompanying extension of the terrane north or the Garlock fault may have resulted in an overall increase in the average density of the crust in this region, producing higher gravity values.

Acknowledgments

We wish to thank Keith Howard, Dave Ponce, and Bob Powell whose thorough reviews prompted us to reexamine the palinspastic reconstruction of the evolution of the Garlock fault, and led to our testing of the non-uniform slip model for the eastern part of the fault, a model we now believe satisfies the geophysical data much better than the uniform slip model.

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Seismicity of the Northeastern Mojave Desert, 1981 - 1997

Egill Hauksson, *Seismological Laboratory, Caltech, Pasadena, CA 91125*

The 1981-1997 seismicity of the northeastern Mojave Desert is associated with several late Quaternary faults in the region. To the north, minor levels of background seismicity are associated with the eastern part of the Garlock fault. In the central and southern part of the region, the seismicity is associated with Camp Rock, Calico, and Coyote Lake faults and to a lesser extent with Blackwater, Harper, Lenwood, and Lockhart faults. This seismicity is dominated by aftershocks of the M_w7.3 1992 Landers earthquake.

The Garlock Fault

The Garlock fault, a 265 km long left-lateral strike-slip fault, separates the Mojave Desert from the Tehachapi Range and the Sierra Nevada to the north. Astiz and Allen (1983) describe the seismicity of the Garlock fault from 1932 to 1981. They found that to the southwest of Rand, the fault showed some aseismic slip, has a complex surface trace, and some background seismicity. A significant change in strike coincides with an en-echelon offset at Rand. To the east of Rand the 155 km long segment of the Garlock fault has an east strike, simple surface trace, no creep and less background seismicity (Astiz and Allen, 1983). The seismicity from 1981 to 1997 is very similar to the activity during the previous decades with a low rate of activity (Figure 1). The hypocenters shown in Figure 1 are calculated using a new three-dimensional velocity model of southern California (Hauksson, in preparation). Although the focal mechanisms of small events along the Garlock fault are not well constrained, they are mostly left-lateral strike-slip consistent with the long term geological motion along the Garlock Fault (Astiz and Allen, 1983). The largest earthquake associated with the Garlock fault in the last 60 years occurred 55 km south-southwest of Rand. It was a M5.7 event triggered by the 1992 Landers mainshock on 7 November 1992.

Faults and Seismicity Near Barstow

From 1981 to June 1992 only two M4 events occurred, in June and December 1991, near the north end of the Calico fault and near the central part of the Garlock fault. A total of 37 Landers aftershocks of magnitude greater than 4.0 have occurred in this region since June 1992. The Barstow seismicity extends to 10 km depth. The focal

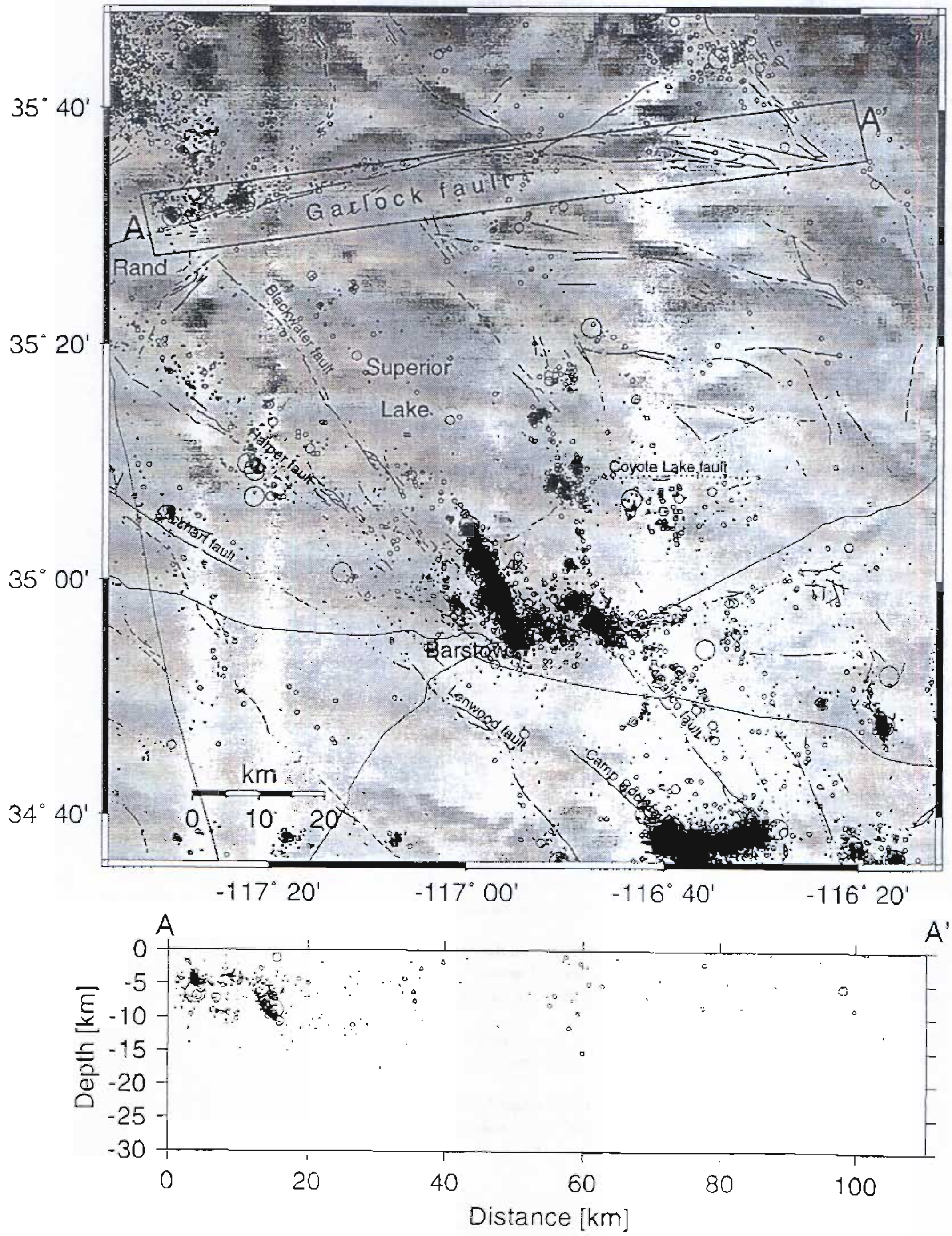

Fig. 1. Seismicity recorded by the Southern California Seismic Network.

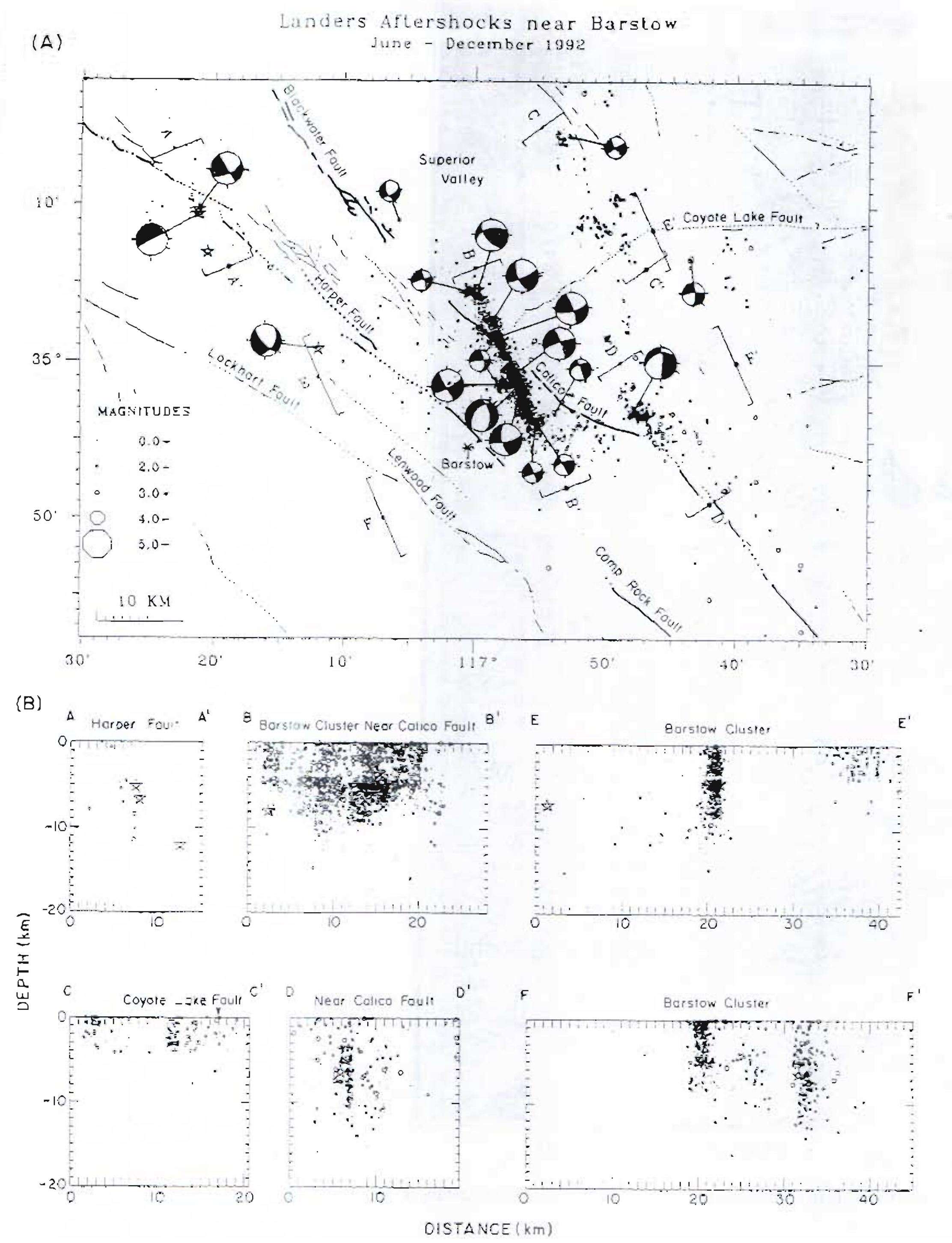

Fig. 2. Aftershocks near Barstow. (a) Map showing major faults (dotted where inferred) and locations of aftershocks from Jun 28 to December 31, 1992. Lower-hemisphere, first-motion, focal mechanisms of typical events are also shown. (b) Cross sections showing the depth distribution of aftershocks. Earthquakes of $M \geq 4.0$ are shown by stars. *From Hauksson et al. (1993).*

mechanisms are mostly strike-slip with some minor amount of normal faulting (Hauksson et al., 1993). The largest earthquake to occur in this region was a M5.3 Landers aftershock of 18 March 1997. It was located 15 km to the northeast of Barstow at the north end of the Calico fault (Figure 3).

The 1992 Landers aftershock sequence illuminated several late Quaternary fault structures in the Barstow area (Hauksson et al., 1993). The Barstow sequence that is still active in 1998 is separated by 30 to 40 km from the Landers mainshock rupture. The initial Barstow cluster was 20 km long, 2 to 3 km wide, with a north-northwest strike (Figure 2). From 1992 to the end of 1997 the Barstow cluster has grown into two major linear seismic zones and several smaller clusters. The main Barstow cluster is located at the juncture of the Blackwater and Calico faults, although it strikes more northerly than the strike of either of these mapped faults.

The second largest cluster, about 10 km long, is located at the north end of the main trace of the Calico fault. The north-northwest strike of this cluster coincides with the strike of the Calico fault and appears to extend its surface trace 5 to 10 km to the north-northwest, beyond the mapped surface trace of the Calico fault.

Several small clusters with a spatial extent of several kilometers are found throughout the region. Some of these clusters appear to form linear trends for distances of 5 to 10 km, and in extreme cases up to 20 km. Some of these longer trends can be seen on the east side of Superior Valley. The strike of these apparent linear trends in the seismicity vary across the region, with almost northerly trends to the east and more north-northwest trends to the west. This is also consistent with mapped late Quaternary faults in the region.

The occurrence of Landers aftershocks in the Barstow area is both related to the magnitude of the Landers earthquake as well as possibly the type of tectonic regime. The Landers mainshock appears to have caused almost complete stress drop along most of the mainshock rupture (Hauksson, 1994). This in turn changed the state of stress on adjacent faults, causing aftershocks over a wide area. In addition, the faults in the Barstow area may have been closer to failure and thus the static stress triggering from the mainshock caused a large number of aftershocks.

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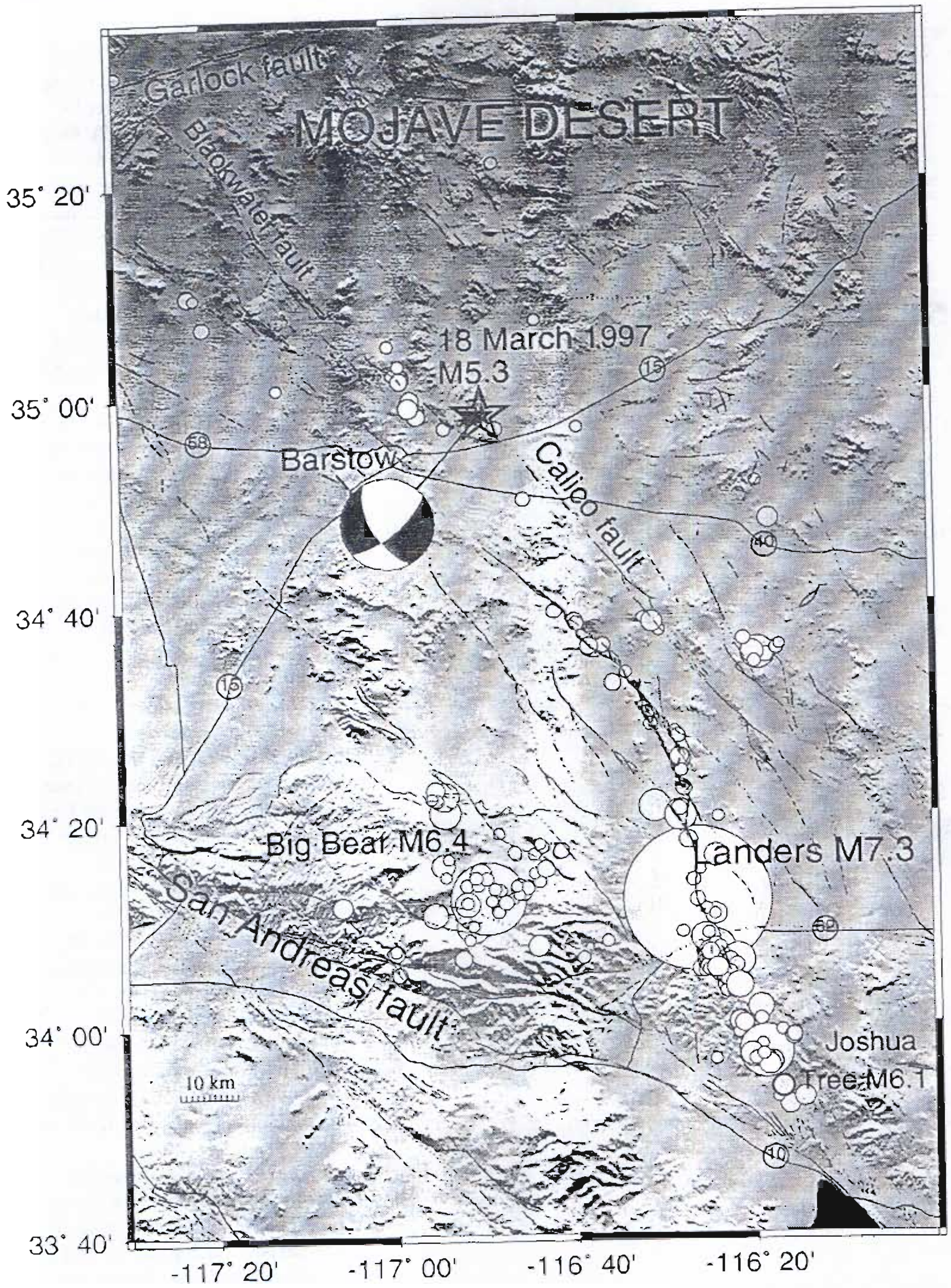

Fig. 3. The M5.3 Calico earthquake and its lower hemisphere focal mechanism. Landers aftershocks of magnitude 4 or greater recorded by SCSN are also shown.

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Late Cretaceous and Early Paleogene Tectonics and Sedimentation near the Garlock Fault, California—did Neogene Strike-slip Faulting Exploit an Older Structural Zone?

Brett F. Cox, *U.S. Geological Survey, 345 Middlefield Road, Menlo Park, CA*

ABSTRACT

Recent studies suggest that the Garlock fault did not exist as a strike-slip structure until the Miocene. However, judging from abundant evidence of earlier large-scale crustal movements along and near the fault, structural inheritance may have played a key role in determining its present configuration. Separate tectonic regimes that produced crustal extension in the Late Cretaceous and Paleocene (100–55 Ma) and contraction in the early Eocene (55–50 Ma) each significantly altered the structure of the crust along the future trace of the fault.

The Late Cretaceous-Paleocene extensional regime may be genetically related to subduction of the Rand Schist. Following the emplacement of a thick wedge of the relatively weak and buoyant schist beneath batholithic upper crust in the Late Cretaceous, the upper crust apparently collapsed and flowed laterally along low-angle normal faults. In consequence, the underlying schist locally was nearly exhumed in the southernmost Sierra Nevada, Tehachapi Mountains, and neighboring areas along the Garlock fault. The physical evidence supporting this image of profound upper-crustal extension includes outliers of Rand Schist lying as much as 100 km inland from the San Andreas fault, detached slabs of upper-crustal granite resting on lower-crustal gneiss and schist, and thick sequences of coarse-grained clastic rocks deposited in continental basins (Goler and Witnet Formations). Uppermost Paleocene marine deposits capping a thick Cretaceous(?)-Paleocene continental sequence in the El Paso Mountains show that the floors of structural basins locally subsided as much as 4 km below sea level in the extended terrane. By the time extension finally waned near the end of the Paleocene, the upper crust along the western and central segments of the Garlock fault was severely thinned and dissected by high-angle and low-angle normal faults. It possibly also was warped into a broad NE-trending basement arch cored at shallow depths by Rand Schist and isostatically compensated at great depths by a bulge in the mantle.

During the early Eocene, regional contraction along a NW-SE axis uplifted and deformed the crust over a broad area spanning the southeasternmost San Joaquin Valley, Tehachapi Mountains, southernmost Sierra Nevada, and northwestern Mojave Desert. Near the Garlock fault, this episode is recorded by a regional unconformity that separates the uppermost Cretaceous(?)-Paleocene Goler Formation from the Eocene Tejon Formation. The contractional deformation destroyed the Goler and Witnet basins through uplift and erosion, combined with NW-vergent tilting, folding, and thrust faulting. Simultaneously, a northeast-trending antiformal welt cored by Rand Schist and capped by slabs of dismembered batholithic rocks rose along the southeastern margin of the disrupted sedimentary basins. This contractional structure may have nucleated along the crest of the hypothesized older extensional arch. In combination, the sequential episodes of extensional and contractional deformation produced a unique set of northeast-trending structures and likely zone of crustal weakness that closely parallels the modern Garlock fault and probably predetermined its location.

Introduction

The Garlock fault extends northeastward for about 255 km from its junction with the San Andreas fault, separating the Mojave block to the south from the Basin and Range and Sierra Nevada provinces to the north (Fig. 1). Total left-lateral displacement of about 48 to 64 km is well established for the northeast half of the fault, based on an offset Late Jurassic dike swarm (Smith, 1962; Chen and Moore, 1979) and an offset steeply-dipping section of Paleozoic strata (Smith and Ketner, 1970; Carr and others, 1992) Monastero and others (1997) concluded that the earliest lateral movement of the Garlock fault was after 17 Ma, because a distinctive sequence of early Miocene continental volcanic and sedimentary rocks lying north of the fault in the western El Paso Mountains is separated left-laterally about 64 km from a similar sequence south of the fault near Pilot Knob Valley. This finding supports the widely endorsed idea that transcurrent movements on the Garlock fault are propelled by east-west extension in the Basin and Range province north of the fault, which began about 16 or 17 Ma (Eaton, 1932; Davis and Burchfiel, 1973). A magnetostratigraphic study of middle and upper Miocene strata in the western El Paso Mountains revealed progressive counterclockwise rotation beginning about 10 Ma, which might correspond to the onset of lateral slip on the Garlock fault (Burbank and Whistler, 1987; Loomis and Burbank, 1988). Quaternary alluvial fans offset from their source canyons along the south front of the El Paso Mountains record at least 18 km of left-lateral displacement within the past 1.5 Ma, implying that a large fraction of the fault's total slip may be post-Miocene (Carter, 1980).

Although the foregoing evidence suggests that horizontal slip on the Garlock fault might be due entirely to Neogene and Quaternary displacements, significant questions remain regarding the origin of the fault and the extent to which its development might have been affected by earlier tectonic events. The NE-trending arcuate trace of the fault is intriguing, as it runs athwart the prevailing northwest structural grain of California. However, the fault now appears to be balanced geometrically and kinematically by the equally anomalous west-northwest-trending "big-bend" segment of the San Andreas fault that lies north and east of Los Angeles (Fig. 1). The origin of this structural combination is problematic (Stuart, 1991): did the Garlock fault initially rupture in response to stresses focused by the bend in the San Andreas fault, or did early left-lateral displacements on the Garlock fault deflect the San Andreas fault, thus determining the position of its big bend? In either case, was the location of the Garlock fault partly controlled by a pre-existing zone of structural weakness? To address these problems, one must consider the pre-Neogene tectonic history of areas bordering on the Garlock fault.

Regional Uplift History and Significance of Pelona-Orocopia-Rand Schists

In a pioneering study of the Mojave Desert region, Hewett (1954, 1955) inferred profound uplift and erosional denudation of the Mojave block between Late Cretaceous and Miocene time, which he suggested

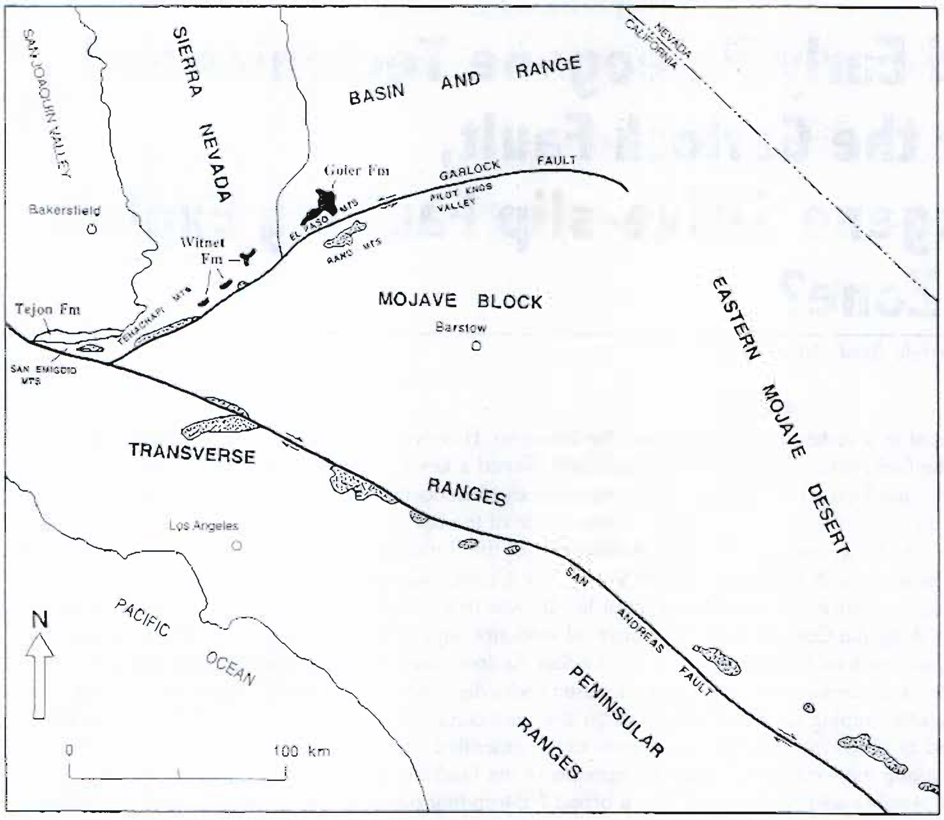

Fig. 1. The greater southern California region showing geologic provinces and generalized major strands of the San Andreas and Garlock fault zones. Numerous exposures of undifferentiated Pelona, Orocopia, and Rand schists (stippled pattern) are distributed along the San Andreas fault and the western half of the Garlock fault. The belt of schist outcrops along the Garlock fault is bordered to the north by several masses of lower and middle Paleogene clastic rocks, including: the uppermost Cretaceous(?) and Paleocene Goler Formation in the El Paso Mountains and its likely contemporary, the Witnet Formation, in the southernmost Sierra Nevada and northeastern Tehachapi Mountains (solid pattern); and a distinctly younger unit, the Eocene Tejon Formation, in the southwestern Tehachapi Mountains and San Emigdio Mountains (unpatterned).

was accomplished partly by vertical movements on an ancestral Garlock fault. This idea was based on the observation that there are few if any Paleogene strata and only scattered eroded remnants of pre-Tertiary strata in the Mojave block, whereas a more complete stratigraphic record is found in neighboring regions to the north and south. Later studies determined that a region much broader than the Mojave block was elevated during the Paleogene (Nilsen and McKee, 1979). It gradually became apparent that local zones of profound uplift and crustal denudation are distributed irregularly across parts of the Mojave block, eastern Mojave Desert, Transverse Ranges, Tehachapi Mountains, and the southernmost Sierra Nevada (Hoisch and others, 1988; Pickett and Saleeby, 1993; Jacobson and others, 1996; Wood and Saleeby, 1997). As the structural boundaries of the Mojave block, including the Garlock fault, lie within this broader region, the hypothesis of simple differential uplift along an ancestral Garlock fault now appears doubtful.

The exhumation of mid-crustal rocks in various parts of southern California is widely thought to have resulted from emplacement of the family of Pelona-Orocopia-Rand schists beneath the region in the Late Cretaceous (for a contrary view, see Barth and Schneiderman, 1996). Petrologic and thermochronologic studies suggest the schist was metamorphosed in the middle to lower crust at depths of 25–30 km or more during the Late Cretaceous and rose to upper-crustal depths of 15 km or less during the Late Cretaceous and Paleocene (Jacobson, 1990; Pickett and Saleeby, 1993; Jacobson and others, 1996). The schist originated as a thick wedge, or perhaps multiple wedges, of relatively weak, low-density metagraywacke subducted beneath continental batholithic rocks. Owing to the underplating of thick masses of relatively weak and buoyant schist, the crust presumably rose isostatically, while simultaneously collapsing and flowing laterally under its own weight (e.g., Malin and others, 1995; Wood and Saleeby, 1997). During this process, the relatively brittle batholithic rocks were extended along low-angle normal faults.

Outcrops of the schist are mainly distributed along or near the Garlock and San Andreas faults (Fig. 1). The outcrops include several masses of Rand schist lying on either side of the Garlock fault in the San Emigdio, Tehachapi, southern Sierra Nevada and Rand Mountains. The spatial association of the schist outcrops with the major strike-slip faults may be interpreted in either (or both) of the following ways (Jacobson and others, 1996): (1) The faults may have ruptured along pre-existing zones of relatively weak crust corresponding to structural culminations or upward bulges in the schist, or (2) the schist may have been uplifted by transpressional stresses associated with Neogene and Quaternary strike-slip faulting.

Unroofing of Gneiss and Schist near the Garlock Fault

Structural, petrologic and geophysical evidence from the Tehachapi and Rand Mountains suggest that some of the basement rocks exposed in these ranges rose from lower-crustal to upper-crustal levels during the Late Cretaceous and Paleocene (Jacobson, 1990; Pickett and Saleeby, 1993; Malin and others, 1995; Wood and Saleeby, 1997). The southwestern Tehachapi Mountains contains exposures of orthogneiss (Tehachapi gneiss complex) and associated plutonic rocks that are thought to have crystallized in the root of the Sierra Nevada batholith (Ross, 1985; Sams and Saleeby, 1988). Geobarometric studies and isotopic dating indicate the gneiss was intruded and metamorphosed in the lower to middle crust (25-30 km depth) between about 115 and 100 Ma, and then was uplifted to upper-crustal depths (10-15 km) between about 100 and 85 Ma (Pickett and Saleeby, 1993). Based on fission-track dating of apatite, the gneiss cooled to 100°C by about 70 Ma, implying it was uplifted to within a few kilometers of the earth's surface before the end of the Cretaceous (Naeser and others, 1990). Unroofing was completed before the late early Eocene (50 Ma), when marine clastic rocks of the Tejon Formation (Fig. 1) were deposited on the gneiss (Nilsen, 1987; Goodman and Malin, 1992).

The uplift history of the Rand Schist is less well constrained, but the schist seems to structurally underlie the Tehachapi gneiss (Malin and others, 1995), and it apparently crystallized at a similar or slightly greater depth (about 30 km; Sharry, 1981; Pickett and Saleeby, 1993). Thus, it seems likely that the schist was uplifted along with the gneiss during the Late Cretaceous. This is supported by evidence from the Rand Mountains, where the schist structurally underlies orthogneiss and paragneiss of the Johannesburg Gneiss. Here the schist is locally intruded by a 79-Ma granodiorite stock fringed by a retrograde metamorphic halo (Silver and others, 1984; Silver and Nourse, 1986). Muscovite from the schist yields $^{40}Ar/^{39}Ar$ ages of about 72 Ma (Jacobson, 1990). These relations indicate the schist was relatively cool (about 350°C or less) and therefore probably ensconced well within the upper crust by about 79 Ma.

Widespread evidence suggests that the gneiss and schist may have been unroofed through extensional collapse of the upper crust by detachment faulting during and following the subduction of the Rand Schist (Malin and others, 1995; Wood and Saleeby, 1997). Possible examples of major Late Cretaceous-Paleocene detachment faults include the Pastoria fault in the Tehachapi Mountains, upper strands of the Rand "thrust" system in the Rand Mountains, and the recently identified Blackburn Canyon and Jawbone Canyon faults in the southern Sierra Nevada (Wood and Saleeby, 1997). Most of these faults crop out along or near the Garlock fault not far from outcrops of the Rand Schist. Extreme extension of the upper crust near the path of the future Garlock fault may have created a zone of crustal weakness that eventually helped to determine the location of the fault.

Large-scale Tilting and Folding of Basement Rocks

The present structural configuration of the Rand Schist and overlying gneissic rocks might provide clues to early tectonic events that potentially influenced the origin of the Garlock fault. In the Rand Mountains, the Johannesburg Gneiss and superposed fault-bounded slabs of upper-crustal plutonic rocks are broadly arched across the schist, forming a large ENE-trending antiformal structure that trends approximately parallel to the Garlock fault (Silver and Nourse, 1986). In the southwestern Tehachapi Mountains, seismic-reflection data suggest that the schist and overlying Tehachapi gneiss complex form a major crustal-scale homocline dipping about 15°-30° to the northwest (Malin and others, 1995). As the Rand Mountains were directly east of the Tehachapi Mountains before being displaced left-laterally along the Garlock fault, the question naturally arises as to whether the antiformal structure in the Rand Mountains might be genetically related to the homoclinal structure in the Tehachapi Mountains.

Malin and others (1995) proposed that the gross homoclinal structure of the Tehachapi Mountains originated during the Late Cretaceous-Paleocene extensional unroofing of the schist. However, they also concluded that the southeast edge of the homocline has been modified by younger northwest-vergent reverse faults that may reflect Neogene transpression along the Garlock fault. Similarly, there may be two reasonable options for the origin of the antiformal structure in the Rand Mountains. The folding may entirely post-date the massive unroofing of the schist, being a product of post-Paleocene regional contraction or local transpression. Alternatively, the early unroofing event, and compensatory upwelling of the mantle, might have induced the growth of a large NE-trending basement arch cored by the Rand Schist. If so, then the antiformal structure of the Rand Mountains may represent the crest of the old arch, modified to a greater or lesser degree by more recent tectonic events, and the large-scale homoclinal structure of the Tehachapi Mountains may represent the northwest limb of the arch.

Cretaceous(?) and Paleocene Clastic Rocks

The belt of Rand Schist outcrops along the Garlock fault is flanked on its north side by a parallel narrow belt of old clastic rocks consisting of the Goler and Witnet Formations (Fig. 1). These units primarily consist of fluvial and alluvial-fan deposits composed of arkosic sandstone, siltstone, and conglomerate.

Goler Formation

The Goler Formation crops out on the north flank of the El Paso Mountains (Fig 1), where it has been mapped and described by several workers (Dibblee, 1952, 1967; Cox, 1982, 1987; Cox and Diggles, 1986). The formation locally is nearly 4 km thick (maximum thickness about 3,910 m), and it overlies a deeply channeled erosion surface cut into metamorphosed Paleozoic rocks and latest Paleozoic to early Mesozoic plutonic rocks. The Goler consists of two superposed stratigraphic sequences (Cox, 1982, 1987). An upward-fining sequence that forms the lower half of the formation was deposited primarily by southward-flowing ephemeral streams and consists of angular detritus derived from granitic and metasedimentary rocks similar to those exposed in the adjacent basement core of the El Paso Mountains and in the nearby southern Sierra Nevada. The upper stratigraphic sequence mainly consists of a westward-fining succession of fluvial sediments, but it also contains a thin (45 m) interval of marine conglomerate, sandstone, and siltstone near its top. Most of the clastic debris in the upper sequence is foreign to the El Paso Mountains region, and it includes abundant rounded pebbles and cobbles of indurated quartzite and silicic volcanic rocks that apparently were transported a considerable distance by streams and deposited along the axis of the sedimentary basin. Cross-stratification and clast imbrication indicate the basin-axis streams flowed toward the west and southwest (Dibblee, 1952; Cox, 1982, 1987).

The upper half of the Goler Formation has produced datable fossils including early and late Paleocene mammals (McKenna, 1955, 1960; McKenna and others, 1987). Although conflicting Paleocene and late early Eocene microfossil ages were initially reported for the marine deposits (Cox and Edwards, 1984; McDougall, 1987), more definitive evidence from calcareous nannofossils indicates they accumulated at or slightly before the Paleocene-Eocene boundary (zone CP8 of Okada and Bukry, 1980) (Squires and others, 1988; Reid and Cox, 1989). The lower half of the Goler Formation is essentially undated, but if the sedimentation rate derived from the overlying fossiliferous strata is projected downward, then Late Cretaceous ages of about 70–80 Ma are predicted for deposits near the base of the formation.

The Goler Formation dips gently northward about 10°-30° and is overlain with pronounced angular unconformity by lower Miocene continental volcanic and sedimentary rocks of the Cudahy Camp Formation (Loomis and Burbank, 1988). Tilting and erosion of the Goler Formation occurred sometime after the latest Paleocene but before 18 Ma, based on the well-dated marine sequence at the top of the Goler Formation (Reid and Cox, 1989) and K-Ar ages from overlying Miocene volcanic rocks (Cox and Diggles, 1986). A correlative lower Miocene volcanic sequence south of the Garlock fault near Pilot Knob Valley is dated at about 20 Ma near its base (Monastero and others, 1997; Sabin and others, 1997), which may more tightly constrain the minimum age of the unconformity in the El Paso Mountains. Syndepositional and postdepositional net vertical movements of the sedimentary basin are recorded by the marine deposits near the top of the Goler, which constitute a sea-level datum. Thus, the floor of the basin subsided at least 4 km between latest Cretaceous(?) and latest Paleocene time and was subsequently uplifted at least 5 km sometime after the Paleocene (Cox and Edwards, 1984; Cox, 1987).

Witnet Formation

The Witnet Formation is exposed in three separate areas in the southernmost Sierra Nevada and northeastern Tehachapi Mountains (Fig. 1). The greatest reported thickness, about 1,220 m, is in the central outcrop area (Dibblee, 1967; Dibblee and Louke, 1970). The Witnet is poorly dated owing to a lack of fossil remains. The sediments generally resemble those of the Goler Formation and are probably correlative (Dibblee, 1967; Cox, 1982, 1987). Field reconnaissance by the author suggests that all three outcrops of the Witnet most closely resemble deposits in the lower stratigraphic sequence of the Goler Formation, which implies a tentative latest Cretaceous(?) to early Paleocene age. Like the Goler Formation, the Witnet is unconformably overlain by lower Miocene volcanic and sedimentary rocks.

Deformational structures associated with the Witnet Formation are more pronounced than those in the Goler Formation and provide evidence of two distinct tectonic events. Recent work suggests that all three outcrops of the Witnet Formation lie within the upper plate of a SE-dipping detachment fault (Blackburn Canyon-Jawbone Canyon fault) that apparently was active during deposition of the formation (Wood and Saleeby, 1997). This major extensional structure apparently is overprinted by younger contractional structures including a large-scale NE-trending, NW-vergent syncline (Wood and Saleeby, 1997). Beds near the base of the Witnet Formation on the southeast flank of this fold generally dip moderately to steeply to the northwest and are locally overthrust from the south by granitic rocks (Smith, 1951; Dibblee and Louke, 1970; Wood and Saleeby, 1997).

Basin models

Several paleotectonic models have been advanced to explain the origin of the Goler and Witnet basins. Based on the substantial thickness of coarse clastic sediments in both formations, and the linear distribution of the outcrops, Nilsen and Clarke (1975) concluded the deposits accumulated in a tectonically active, fault-bounded basin, thus implying the existence of an old zone of structural weakness along the Garlock fault. Similar ideas were stated earlier by Smith (1962, 1964). Nilsen and Clarke (1975) further suggested that deposition might have occurred in a strike-slip basin along a proto-Garlock fault. The latter suggestion has not been confirmed by any field studies. Moreover, the existence of a proto-Garlock fault now seems doubtful, as recent studies suggest that known lateral separations of distinctive basement units across the Garlock fault are accounted for by Neogene and Quaternary faulting (Monastero and others, 1997).

A second model envisioned that the Goler and Witnet Formations accumulated in a contractional basin during differential uplift and northward thrusting of the Mojave block (Cox, 1987). This model enlarges upon the earlier ideas of Hewett (1954, 1955) and Smith (1962, 1964) regarding Paleogene uplift of the Mojave block and was devised in order to explain several critical stratigraphic and structural features, including: (1) the northward-dipping homoclinal structure of the Goler Formation and local northward overthrusting or overturning of strata along the south margin of the Witnet Formation, (2) the large vertical tectonic movements that are implied by the marine deposits at the top of the Goler Formation, (3) the apparent absence of correlative Paleocene clastic deposits south of the Garlock fault, and (4) the present high structural level of the Rand Schist in the Rand Mountains south of the fault (Cox, 1982; Cox and Diggles, 1986).

In retrospect, there are several obvious problems with the foregoing model. The kinematic significance of the Goler homocline is ambiguous, and its interpretation as a north-vergent contractional structure requires confirmation. The absence of Paleocene strata on the south side of the Garlock fault also may be explained in various ways besides northward thrusting and does not necessarily imply syndepositional uplift of the Mojave block. Bodies of Rand Schist are present on either side of the Garlock fault, and reversal of left-lateral displacement on the fault places schist in the Rand Mountains adjacent to another body of schist in the southern Sierra Nevada (Fig. 2); thus, the schist does not imply differential uplift of the Mojave block relative to the El Paso Mountains or southern Sierra Nevada. Moreover, as will be discussed, all of these features might more appropriately be referred to a separate tectonic regime that apparently directly followed the deposition of the Goler Formation.

Most recently, the Goler and Witnet basins have been interpreted as "supradetachment" basins, meaning they are extensional basins that formed in the upper plates of detachment faults (Wood and Saleeby, 1997). According to this model, the basins formed during large-scale extensional unroofing of the southernmost Sierra Nevada batholith during the Late Cretaceous and Paleocene. The Witnet Formation is thought to have formed in a basin above the newly recognized Blackburn Canyon and Jawbone Canyon detachment faults. Structures and mineral fabrics observed along these faults suggest that upper-plate transport was toward the south or southeast, implying NNW-SSE extension (Wood and Saleeby, 1997). The model also proposes that the Goler Formation formed in the upper plate of a similar, presumably coeval, detachment fault whose breakaway may now be concealed beneath Neogene and Quaternary deposits directly northwest of the El Paso Mountains.

The new basin model is provocative and should inspire further study of the Goler and Witnet Formations. The general conclusions of the model are persuasive because they are supported by significant new field observations and are consistent with current regional paleotectonic models for Late Cretaceous and Paleocene time (e.g., Malin and others, 1995). Certain major features of the Goler Formation seem consistent with its reinterpretation as an extensional basin, including the fact that the basin floor subsided at least 4 km below sea level during deposition (Cox and Edwards, 1984; Cox, 1987). Cox (1982) mapped a horst of plutonic rocks that projects northwestward into the Goler Formation in the southwestern El Paso Mountains. This fault block is bounded on either side by steeply dipping normal faults that were active during deposition of undated (uppermost Cretaceous?) strata in the deepest stratigraphic levels of the Goler Formation. The syndepositional faulting apparently records differential subsidence of the floor of the Goler basin in response to tensional stresses, which agrees with the general extensional mechanism endorsed by the new basin model. The northwest trend of the fault block seems inconsistent with the NNW-SSE extension direction proposed by Wood and Saleeby (1997), although this might be reconciled by assuming that the block ruptured along a pre-existing set of northwest-trending fractures oblique to the ambient extension direction.

The westward- and southwestward-flowing paleocurrent trends of the main basin-axis streams in the Goler Formation agree with extension along a NNW-SSE axis, as proposed in the new model, because supradetachment basins typically are elongated perpendicular to extension (Wood and Saleeby, 1997). Tributary streams carrying locally derived detritus entered on the north side of the Goler basin and flowed to the south or south-southeast, approximately parallel to the proposed extension axis. These streams conceivably drained the uplifted footwall of a south-dipping normal fault that has been suggested to bound the north side of the basin (Cox, 1982, 1987; Wood and Saleeby, 1997). However, if such a syndepositional normal fault existed, it apparently did not have pronounced concave-upward listric geometry, for bedding is essentially concordant throughout the thick stratigraphic succession of the Goler Formation, implying that the basin was not tilted significantly during deposition.

The large-scale northward-dipping homoclinal structure of the Goler Formation also might plausibly be ascribed to north-south extension, if one temporarily ignores its possible genetic relation to demonstrably north-vergent contractional structures in the Witnet Formation. The strata in the Goler mainly dip northward about 10°–30°, which might be explained by southward transport and tilting of the basin in the upper plate of a south-dipping listric normal fault. However, as was just noted, the stratal dips do not decrease upsection, so any such listric normal faulting must have postdated the deposition of the Goler Formation. Inasmuch as the Goler Formation has not yielded conclusive evidence of syndepositional extension, and because there is clear evidence of post-depositional north-vergent contractional deformation in the nearby Witnet Formation, there is presently little justification for assuming an extensional origin for the homoclinal structure of the Goler Formation. In fact, the weight of evidence suggests that the Goler was uplifted and tilted northward during a separate tectonic episode of regional crustal shortening and uplift in the early Eocene, immediately following the Late Cretaceous-Paleocene extensional episode.

Early Eocene Crustal Shortening and Regional Uplift along the Garlock Fault

Besides the NW-vergent contractional structures in the Witnet Formation, several other stratigraphic and structural features distributed along and near the Garlock fault provide evidence of an important episode of regional NW-SE contraction that immediately followed the deposition of Paleocene clastic rocks along the Garlock fault. These features include all the items that were formerly marshaled in support of a contractional origin for the Goler basin (Cox, 1982, 1987): the Goler homocline, large-magnitude (5 km) post-depositional uplift of the Goler Formation, the absence of Paleocene clastic deposits directly south of the Garlock fault, and the linear belt of Rand Schist outcrops along the fault. Contractional deformation and associated uplift seem to have affected a broad zone extending northeastward at least 150 km from the San Emigdio Mountains to the El Paso Mountains, a region now occupied by the southwest half of the Garlock fault. Stratigraphic and sedimentologic impacts of regional uplift also can be traced about 100 km northwestward into the southeastern San Joaquin Valley, based on evidence from boreholes (Reid and Cox, 1989). The event may also have affected much of the Mojave block to the southeast of the Garlock fault. However, the available evidence suggests that deformation was especially intense in a narrower belt that straddles the trace of the modern Garlock fault.

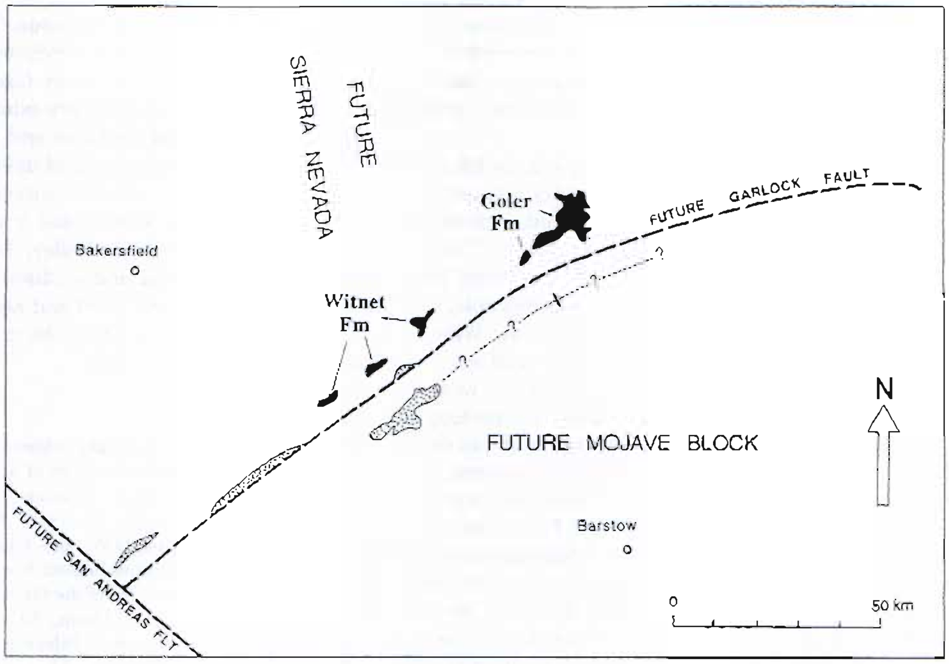

Fig. 2. Palinspastic sketch map of Garlock fault, showing present outcrops of the Rand Schist (stippled pattern), and the Goler and Witnet Formations (solid pattern) restored to their approximate relative positions prior to major Neogene deformation. Crude restoration was accomplished by simply reversing 60 km of left slip on the Garlock fault and by straightening the southwest end of the fault. Note that the outcrops of schist and clastic rocks define two slightly overlapping narrow belts adjacent to the fault. Dotted line with queries represents hypothetical eastward continuation of antiformal basement fold that is attributed to early Eocene contractional deformation (see discussion in text). The sketch does not fully account for paleomagnetically determined clockwise rotations as great as 40°-60° that have been determined for the Tehachapi Mountains (McWilliams and Li, 1985). A palinspastic reconstruction that fully accommodates the paleomagnetic data would restore the southwesternmost part of the Garlock fault to a NNE trend (about 015°; Goodman and Malin, 1992).

Reid and Cox (1989) proposed an early Eocene age for this tectonic event, because stratigraphic features in the southeastern San Joaquin Valley and El Paso Mountains suggest a major regional unconformity of that age; this interpretation superseded a middle(?) Eocene age that was proposed earlier (Cox, 1987; Cox and McDougall, 1988). The unconformity is inferred from the following relations. An arm of the ancestral Pacific Ocean that flooded the Goler basin near the end of the Paleocene presumably entered from the west or southwest. This assumption is supported by westward-flowing paleocurrent trends observed within beds of fluvial sandstone and conglomerate that lie directly below and above the marine sequence in the Goler Formation (Cox, 1987). However, Paleocene marine sediments are now absent from a broad region southwest of the El Paso Mountains, including the southernmost Sierra Nevada, Tehachapi Mountains, and southeasternmost San Joaquin Valley. The next oldest Paleogene marine deposits in the region are found in the Eocene Tejon Formation. The Tejon crops out in the San Emigdio Mountains and southwestern Tehachapi Mountains (Fig. 1), where it directly overlies mid-Cretaceous batholithic rocks (Nilsen, 1987; Goodman and Malin, 1992). The basal deposits of the Tejon are late early Eocene in age (K. McDougall, oral commun., 1998) and are thus distinctly younger than the uppermost Paleocene marine deposits near the top of the Goler Formation. Marine deposits correlative with the Paleocene marine sequence in the Goler Formation–and with much of the underlying uppermost Cretaceous(?)-Paleocene continental succession as well–probably were deposited in the southwestern Tehachapi Mountains and surrounding areas and thus should underlie the Tejon Formation. However, they are absent and evidently were eroded as a result of uplift in the early Eocene.

Intriguing clues regarding the lateral variability of early Eocene contractional deformation along the trend of the Garlock fault are apparent through inspection of a palinspastic sketch map (Fig.2). On this figure, the present outcrops of the Rand Schist, Goler Formation, and Witnet Formation are restored to their approximate relative positions prior to major Neogene deformation by simply reversing 60 km of left slip on the Garlock fault and straightening the southwest end of the fault. It should be noted that this sketch does not fully adjust for paleomagnetically determined clockwise rotations as great as 40°-60° that have been determined for the Tehachapi Mountains (McWilliams and Li, 1985). A more rigorous palinspastic reconstruction accommodating the paleomagnetic data would restore the southwesternmost part of the future Garlock fault to a NNE trend (about 015°; Goodman and Malin, 1992), in contrast to the NE trend shown in the sketch. However, this difference in orientation does not alter the basic conclusions of the following analysis.

With left slip on the Garlock fault restored, the outcrops of clastic rocks and Rand Schist form two overlapping parallel belts along the path of the fault. The Witnet and Goler outcrops form a narrow belt along the north side of the fault, and the Rand Schist forms a belt that straddles the fault. The overlapping pattern, combined with basic differences in the thickness and deformational structure of the Goler and Witnet Formations, seems to reflect a lateral gradient in the severity of deformation and in the magnitude of post-Paleocene erosion. The Goler Formation contains the thickest (3,910 m) section, and it includes a sequence of marine deposits near its top. Homoclinal dips in the Goler Formation are relatively gentle, mostly 10°-30°. By contrast, the Witnet contains only about one-third the thickness of the Goler Formation (max. 1,220 m), and it apparently lacks the entire upper sequence of the Goler, including the marine deposits. The Witnet is prominently folded and is generally more intensely deformed than the Goler, with bedding inclined as steeply as 40°-90° or even overturned, and with basement rocks locally thrust northwestward across its base. The southwestward increase in deformational intensity that is so evident between the Goler and Witnet Formations is also apparent within the Goler itself. Near the northeast end of its outcrop belt, strata near the base of the Goler mostly dip northward about 15°-20°, whereas the basal deposits near the southwest end typically dip about 35°-40° (Dibblee, 1952; Cox, 1982). Continuing southwestward from the Witnet outcrops into the central Tehachapi Mountains, the belt of early Paleogene clastic rocks ends, presumably owing to steadily increasing uplift and erosion toward the southwest, whereas the masses of Rand schist crop out discontinuously to the southwest end of the future Garlock fault.

It should also be noted on the palinspastic sketch that the prominently deformed Witnet Formation lies adjacent to the northeasternmost outcrops of the Rand Schist, including the antiformally folded body in the Rand Mountains. This suggests the folds and other contractional structures in the Witnet might profitably be compared to structures in the schist. For example, it might be worth investigating whether the south limb of the Witnet syncline could be equivalent to the north limb of the antiform in the Rand Mountains. If such a correlation were accepted, it would not rule out continued growth of the Rand Mountains antiform long after early Eocene time. To the contrary, the synclinal fold that deforms the Witnet formation also warps overlying Miocene strata to a lesser degree, implying that contractional deformation resumed during Neogene time (Smith, 1951; Dibblee and Louke, 1970). Stratigraphic studies in the southwestern Tehachapi Mountains indicate that the Rand Schist was not actually exposed as a source of sediments until the middle Miocene (Goodman and Malin, 1992), which suggests that its uplift continued into the Miocene. Nevertheless, the juxtaposition of the Rand Mountains antiform alongside the synclinally folded Witnet Formation raises the possibility that the basic contours of the schist-cored antiform in the Rand Mountains might have been established during an early Eocene contractional event.

Farther east, the relatively mild contractional deformation implied by the gently dipping Goler homocline might reflect a northeastward attenuation in the intensity of basement folding south of the Garlock fault (dotted line on Fig. 2). However, the total absence of the 4-km-thick Goler succession south of the fault still seems to require at least 4 km of uplift along the northern margin of the Mojave block in this region. Furthermore, this uplift apparently occurred before the Miocene, because the area of the Mojave block that lay opposite the Goler Formation prior to left slip on the Garlock fault (Fig. 1, Pilot Knob Valley) is bordered directly to the south by an early Miocene volcanic field that rests directly on granitic rocks (Monastero and others, 1997; Sabin and others, 1997).

Conclusions

Geologic investigations in areas bordering on the southwest half of the Garlock fault indicate the region was strongly impacted by two deformational events in the Late Cretaceous and early Paleogene. The earlier event, which ranged from Late Cretaceous through Paleocene time (about 100–55 Ma) apparently involved the extensional collapse of upper-crustal batholithic rocks following the subduction of the Rand Schist. Extreme extension along a system of low-angle normal faults in the Tehachapi Mountains and adjacent areas near the future Garlock fault may have severely thinned the upper crust, causing rapid unroofing of mid-crustal orthogneisses and raising the underlying Rand Schist to high levels in the upper crust. The extreme extension may also have resulted in the development of deep fault-bounded basins containing uppermost Cretaceous(?) and Paleocene deposits of the Goler and Witnet Formations. The distribution of these deposits along the Garlock fault suggests the upper crust was already significantly fractured and weakened along the trend of the southwestern Garlock fault by latest Cretaceous or earliest Paleocene time.

The second tectonic episode was in the early Eocene, about 55-50 Ma, when the northwestern Mojave block, Tehachapi Mountains, southeasternmost San Joaquin Valley, and surrounding areas were subjected to intense NW-SE oriented compression. This event left a strong imprint in the region straddling the southwestern half of the Garlock fault, possibly because the batholithic upper crust in this region was severely thinned and weakened by the preceding episode of regional extension. The weakened crust evidently buckled under compression, producing an antiformal welt of Rand schist along the trend of the future Garlock fault. The contractional deformation destroyed the pre-existing extensional basins, uplifting, tilting, and folding the Goler and Witnet Formations. The intensity of deformation and magnitude of uplift increased toward the southwest during this event, with the consequence that Upper Cretaceous and Paleocene strata were eroded from a large area including the southeasternmost San Joaquin Valley, Tehachapi Mountains, southernmost Sierra Nevada, and northwestern Mojave block. The development of the welt of Rand Schist and associated structures further weakened the upper crust, preparing the ground for the development of the Neogene Garlock fault.

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Neogene Offsets and Displacement Rates from Offset Gravels, Central Garlock Fault, California

Bruce Carter, *Pasadena City College, Pasadena, CA*

INTRODUCTION

The Garlock fault, one of the major faults of southern California, extends about 250 km from the San Andreas fault near the town of Gorman on the west to the south end of Death Valley on the east. This left-lateral fault separates the Basin & Range and Sierra Nevada provinces on the north from the Mojave Desert on the south (Figure 1), and has been described as an intracontinental transform fault (Davis and Burchfiel, 1973). Total displacement on the Garlock fault is about 65 km (Smith, 1962, Michael, 1966, Davis and Burchfiel, 1973, Carr et al., 1993), and the well-defined fault-line features in the central and eastern parts suggest very recent activity (Clark, 1973, Roquemore, et al, 1981, Carter, 1987). A well-preserved Pleistocene gravel shoreline bar on the northeast side of Koehn Lake has been offset by the fault (Carter, 1971, Clark and Lajoie, 1974). A variety of offset geomorphic features suggest that the past few earthquakes have produced about 3 to 7 m of slip, and that recurrence intervals on different parts of the fault range from a few hundred to a little over a thousand years (McGill and Sieh, 1991).

The central part of the Garlock fault extends through the northern part of Cantil Valley, which is bounded on the north by El Paso Mountains and on the south by the Rand Mountains (Figure 2). Cantil Valley is a closed depression that is most likely caused by the dilational step-over in the Garlock fault that occurs there. The Garlock fault enters the valley from the west along the southern margin of the valley and exits the valley to the east along the northern margin of the valley. This leftward step in a left-lateral fault creates an area in which the crust is pulling apart, leaving a basin behind. Gravity studies (Mabey, 1960) suggest that the deepest part of Cantil Valley lies approximately beneath Koehn Lake where an estimated 4000 m of Cenozoic sediments underlie the valley. El Paso Mountains are abruptly terminated on the south by El Paso fault, a branch of the Garlock fault which dies out toward the western end of Cantil Valley. There is little or no evidence of lateral movement on El Paso fault, but maximum vertical displacement has been estimated to be about 3000 m (Dibblee, 1952). Farther east the Garlock fault separates the Lava Mountains on the south from Searles Valley on the north (Figure 2).

North of the central part of the fault, El Paso Mountains contain exposures of several distinctive bedrock lithologies, including hornblende diorite, Mesquite Schist and metasedimentary and metavolcanic rocks of the Garlock Series (Figure 3). Gravels derived from these lithologies were transported southward across the fault and deposited in Cantil Valley. Several highly distinctive deposits south of the Garlock fault now lie east of their probable areas of deposition (which was commonly near the mouth of Mesquite Canyon) and therefore demonstrate left-lateral displacements on the fault. Striking examples of these offset deposits can be seen just east of Goler Wash, about 5 km east of the town of Garlock. Much greater offsets can be demonstrated using deposits exposed in the Lava Mountains, particularly in the vicinity of Christmas Canyon (Figure 2). The matching of offset deposits of different ages with their bedrock source areas in El Paso Mountains shows that the Garlock fault has averaged about 7 mm/year displacement since latest Miocene, and that strike slip on the Garlock fault probably began about 9 ma.

BEDROCK SOURCE AREAS

Bedrock source rocks are exposed along the south front of El Paso Mountains and erosion of these rocks has shed distinctive clasts directly southward across the Garlock fault into Cantil Valley. These rocks include Jurassic granitic and dioritic rocks, Paleozoic(?) Mesquite schist, and metasedimentary and metavolcanic rocks of the Paleozoic Garlock Series. In addition, distinctive boulders of Tertiary basalt and clasts of quartzite and other lithologies derived from the Tertiary Goler Formation are exposed on the north side of El Paso Mountains and are being transported and deposited to the south across the Garlock fault only at the mouths of Last Chance Canyon and Goler Wash (Figure 3).

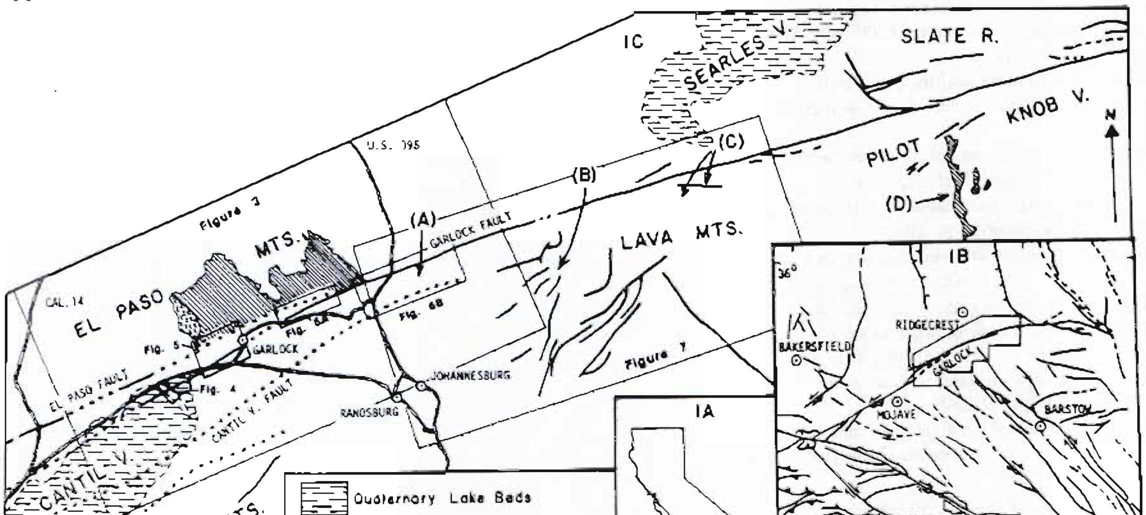

Fig. 1. Index maps to the central Garlock fault, California. Figure 1C. (A) Eastern limit of Pleistocene ans late Pliocene(?) gravels containing Mesquite Schist clasts south of the fault and 19 km east of their probable site of deposition. (B) Mesquite Schist clasts in sediments interlayered with the Pliocene Almond Mountain volcanics about 32 km east of their probable site of deposition. (C) Large blocks of Garlock Series rocks and Mesquite Schist within the latest Miocene Bedrock Springs Formation about 247 km east of their probable site of deposition. (D) Paleozoic Garlock Series rocks displaced a total of about 64 km from similar rocks in El Paso Mountains.

Granitic Rocks

Most of the south face of El Paso Mountains west of Mesquite Canyon is underlain by quartz diorite which is cut by a few small bodies of granite and granophyre. The western part of this area consists of light-colored, massive equigranular biotite quartz diorite and the eastern part consists of dark-gray, medium-grained, equigranular hornblende quartz diorite (Figure 3) which contains much hornblende and less quartz than the biotite quartz diorite. This hornblende quartz diorite contrasts with the generally much more felsic granodiorite, quartz monzonite and granite plutonic rocks elsewhere in the region, including large areas in El Paso Mountains and Spangler Hills north of the fault and the Rand and Lava Mountains south of the fault.

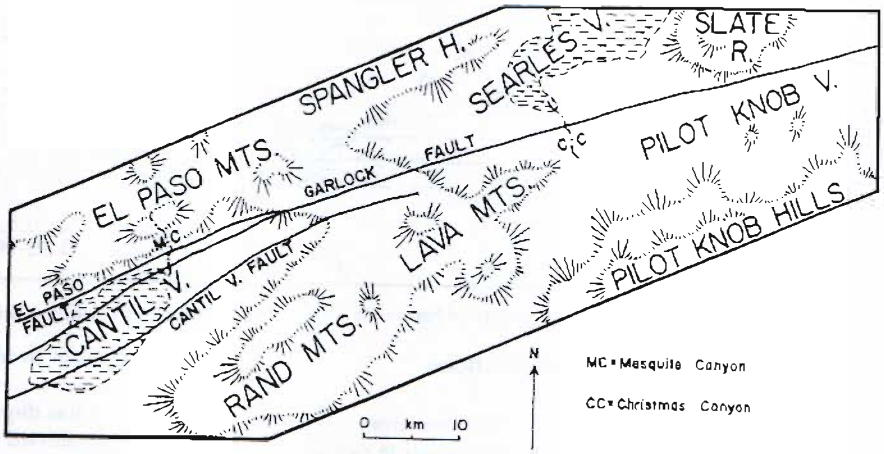

Fig. 2. Physiographic map of the central Garlock fault area showing the location of Mesquite Canyon (MC) draining southward from El Paso Mountains across the fault into Cantil Valley, and Christmas Canyon (CC) draining northward from the eastern Lava Mountains across the fault into Searles Valley.

Mesquite Schist

The Mesquite schist, which crops out in Mesquite Canyon, consists of andalusite schist and schist with interbedded marble (Carr et al., 1984). The schist is a fine-grained quartz-sericite-albite schist nearly always containing porphyroblasts of andalusite (pseudomorphs), chlorite, albite or chloritoid (Christianson, 1961). It is prominently and thinly bedded, and cleaves into thin slabs that have a distinctive silvery sheen. It is readily distinguished from the Rand schist found in the Rand Mountains south of Cantil Valley, which is distinctly coarser and does not exhibit the type of spotted porphyroblastic texture characteristic of the Mesquite schist. Mesquite schist crops out on the south slope of El Paso Mountains in Mesquite Canyon where it covers an area of about 1.5 by 5 km, dips steeply to the east and is overlain disconformably by the Garlock Series to the east.

Garlock Series

The Garlock Series consists of up to 10,000 m of slightly metamorphosed (lower greenschist facies) Paleozoic sediments and volcanics which are exposed on the south side of El Paso Mountains for a distance of about 14 km east of Mesquite Canyon. Most units in this assemblage strike northwest approximately perpendicular to the mountain front and dip moderately to steeply to the northeast. The metasediments in this series are mostly slate and metachert, with smaller amounts of fine-grained marble, quartzite conglomerate and sandstone. The metavolcanics are predominantly basalt-greenstone, andesite porphyry and tuff (Carr et al., 1984). Gravels derived predominantly from the Garlock Series are easily identified, but it may be difficult to determine from which part of the outcrop area they originated.

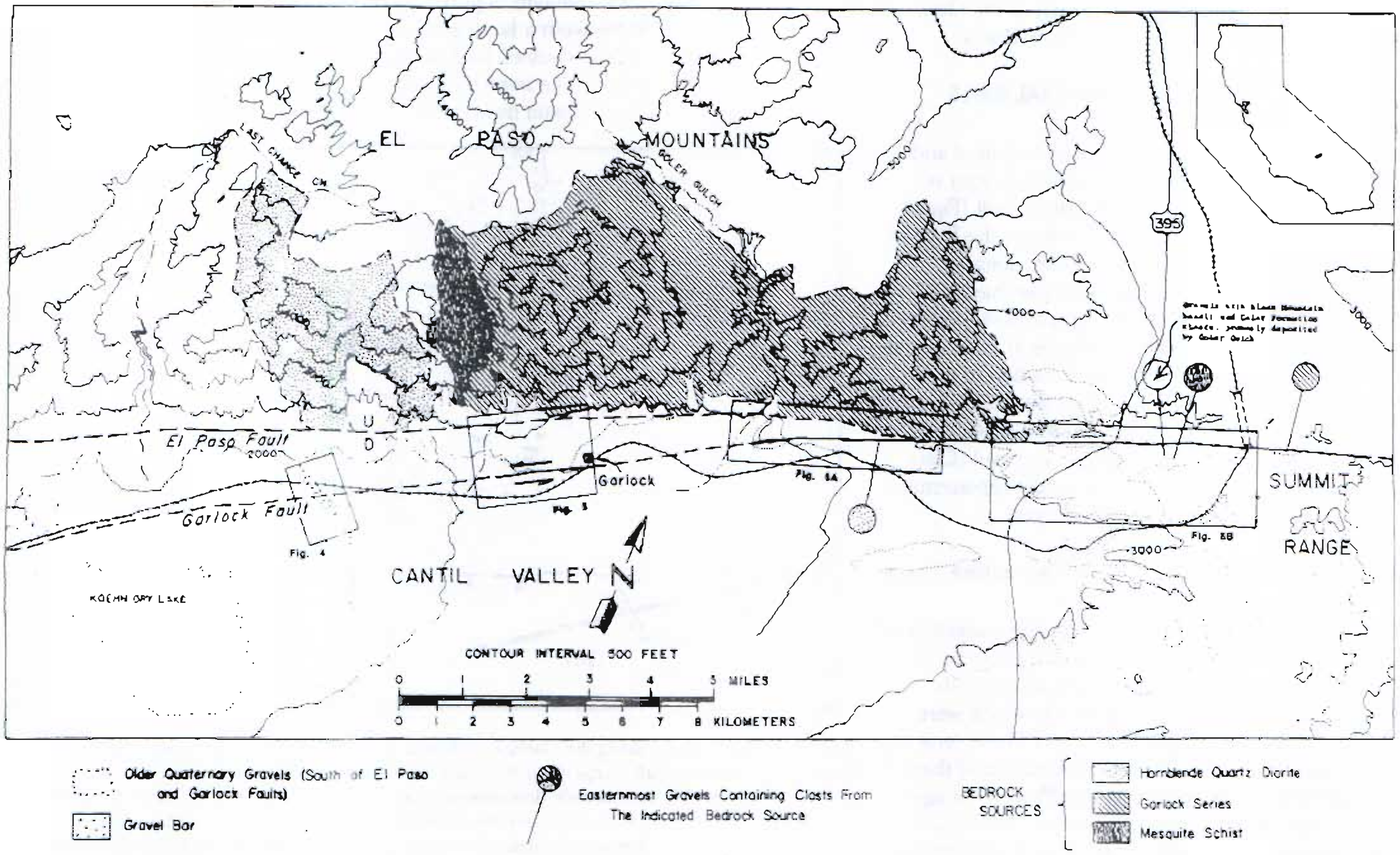

Fig. 3. Geologic map of bedrock source areas in El Paso Mountains and displaced Quaternary to late Pliocene(?) gravels south of the Garlock fault.

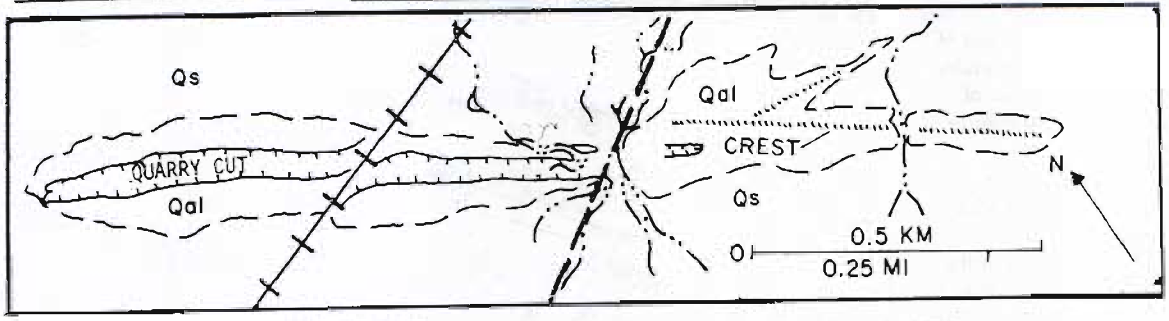

Fig. 4. Well-preserved Pleistocene shoreline gravel bar on the eastern edge of Koehn Dry Lake in Cantil Valley.

Tertiary Formations

Goler Formation

The Paleocene non-marine Goler Formation crops out on the northern flank of El Paso Mountains and consists in part of conglomerate which contains large (up to 60 cm) well-rounded boulders of granitic rocks, quartzite, chert, limestone and aphanitic porphyry (Cox, 1987). These boulders are easily recognized when reworked and deposited as components of younger gravels because of their distinctive lithologies and characteristically well-rounded forms. Boulders derived from the Goler Formation are currently being deposited south of the Garlock fault at the mouths of Last Chance Canyon (sparse) and Goler Gulch (abundant).

Black Mountain Basalt

Black Mountain Basalt (Dibblee, 1952) caps several hills on the north side of El Paso Mountains. It consists of black, fine-grained vesicular basalt which commonly contains small phenocrysts of feldspar and ferromagnesian minerals. Large, moderately rounded boulders of this basalt are very distinctive and form conspicuous components of some gravel deposits. Boulders of Black Mountain Basalt are currently being deposited south of the Garlock fault at the mouths of Last Chance Canyon and Goler Gulch.

OFFSET DEPOSITIONAL UNITS

Holocene Shoreline Bar

About 6 km west of Garlock a prominent sand and gravel shoreline bar along the northeastern margin of Koehn Dry Lake is crossed by the Garlock fault (Figure 3). The crest of this bar lies about 9 m above the floor of Koehn Dry Lake and extends about 6 km along its eastern edge. Although its northwestern part has been quarried for gravel, its geomorphic form is well-preserved about 0.6 km south of the railroad (Figure 4) and suggests that the bar is relatively young. Reconstruction of the original bar morphology requires about 80 m of left-lateral displacement (Carter, 1971). Based on an age of about 11,000 years obtained on tufa deposits on gravel (Clark and Lajoie, 1974), this indicates an average displacement rate on the Garlock fault of about 7 mm/year.

Pleistocene and Upper Pliocene(?) Deposition South of the Garlock Fault

A small area of folded and faulted poorly-consolidated arkosic silt, sand and gravel beds exposed between El Paso and Garlock faults near the mouth of Mesquite Canyon (Figure 5) yields vertebrate fossils which were reported as "Pleistocene or perhaps upper Pleistocene in age" (Dibblee, 1952). More recent examination of these fossils suggests that they may be upper Pliocene in age (David Whistler, personal communication). Of about 100 m of this section exposed, the lower 65 m is primarily silt, but in the upper 35 m of the section, beds of coarse, angular gravel make up as much as half of the section. These gravel beds contain abundant clasts of Mesquite schist and appear to have been deposited as debris flows from the Mesquite Canyon area immediately to the north. This upper Pliocene(?) formation has been tilted 10 to 20 degrees and is unconformably overlain by terrace gravels which also contain abundant clasts of Mesquite schist.

Most displaced gravels south of the Garlock fault are not so well-consolidated as the upper Pliocene(?) gravel beds and are not interbedded with sand or silt and so apparently correlate with the overlying terrace gravels. The characteristics of the displaced terrace gravels suggest that they were deposited near the base of El Paso Mountains at a time when the mountains were elevated and undergoing rapid erosion. Some uplift of El Paso Mountains took place in early Pleistocene time but the major uplift occurred during late Pleistocene time (Dibblee, 1952). Thus most of the displaced terrace gravels south of the fault appear to be Pleistocene in age, although some of the more deeply dissected easternmost gravels may correlate with the underlying upper Pliocene(?) deposits.

Offset Pleistocene and Upper Pliocene(?) Units

Goler Wash Area:

The Garlock fault crosses Goler Wash about 5 km east of the town of Garlock. East of the wash a depression is formed where the southeast bank of lower Goler Wash has been offset about 0.8 km and sealed by alluvium deposited from the west Figure 6A). About 1-4 km east of the wash a large graben is formed between the north-facing scarp of the Garlock fault and the south-facing El Paso Mountains which lie on the north side of the concealed El Paso fault.

The mountain front north of the fault is underlain by distinctive

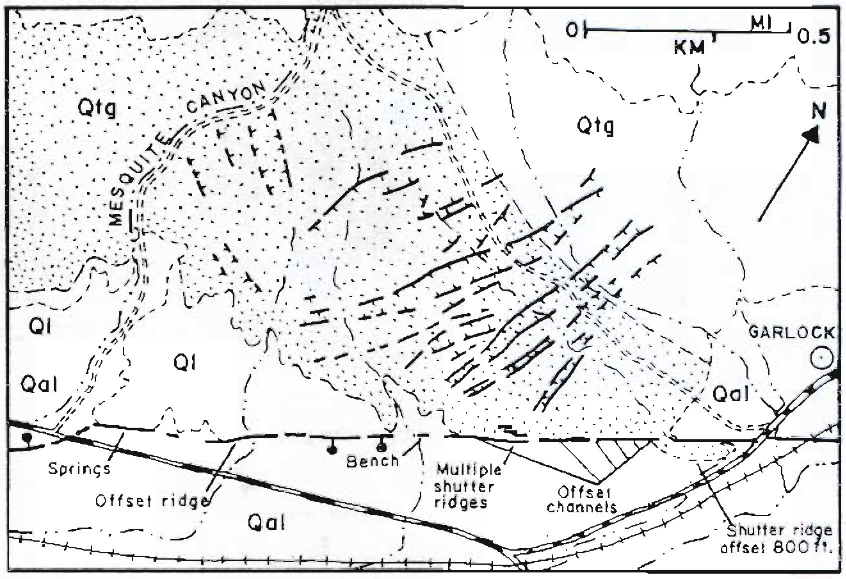

Fig. 5. Fault-line features along the Garlock fault near the mouth of Mesquite Canyon (after Clark, 1973). Normal fault scarps up to 8 m high cut the older uplifted fan surface east of Mesquite Canyon. Ball on relatively downthrown side of scarps along break on which movement has been predominantly horizontal; hachures on relatively downthrown side of scarps along break on which movement has been predominantly vertical. Qal-Quaternary alluvium, Qtg-Quaternary terrace gravels, Ql-Lake beds of Pleistocene or late Pliocene(?) age.

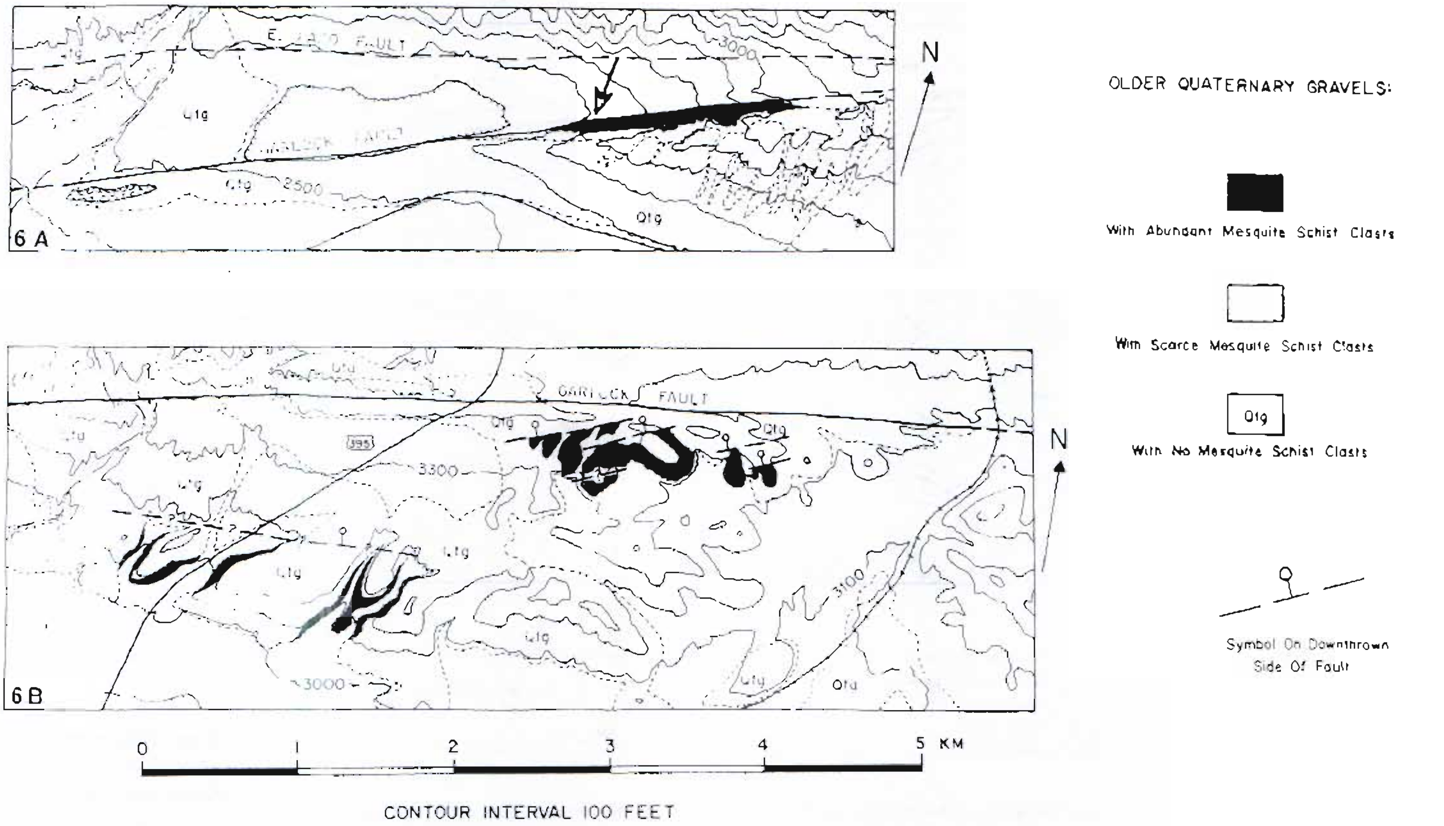

Fig. 6. Geologic maps of Pleistocene and late Pliocene(?) gravels containing Mesquite Schist clasts, Central Garlock fault, California. Arrow shows location of fault slice of monolithologic hornblende quartz diorite gravels faulted up from below Mesquite Schist-bearing gravels along the south side of the fault east of Goler Wash.

rocks of the Garlock series. Gravels south of the fault contain boulders derived from the Black Mountain basalt and from the Goler Formation, which could only have been deposited at the mouth of Goler Wash, and indicate as much as about 6.5 km of left-lateral offset of the gravels. The top of the north-facing fault scarp about 4 km east of Goler Wash exposes surface gravels deposited at the mouth of Goler Wash which are underlain by gravels containing a few small clasts of Mesquite Schist. These gravels are in turn underlain by gravels containing abundant large clasts of Mesquite Schist near the bottom of the scarp. These lower gravels now lie about 11 km east of their depositional area near the mouth of Mesquite Canyon. At about the center of the south wall of the graben a small fault bench exposes gravels (inferred to underlie the schist-bearing gravels) which contain abundant coarse clasts of hornblende quartz diorite (Figure 6A, arrow). This hornblende quartz diorite-bearing gravel is now about 12 km east of its easternmost possible depositional area (Figure 3).

Summit Range-Highway 395 Area:

In the Summit Range, mostly east of Highway 395, many of the surface terrace gravels contain large boulders (eroded from the Goler formation and the Black Mountain Basalt) which are now as much as 11 km east of their probable depositional area near the mouth of Goler Wash (Figure 6B). Garlock series clasts are contained in surface terrace gravels up to about 8 km east of their probable depositional area (Dibblee, 1967) although Quaternary drainage directions from the east slope of El Paso Mountains are less certain than drainage directions on the south flank of the range. Several short, poorly-defined faults south of and sub-parallel to the Garlock fault cut terrace gravels in the Summit Range. Erosion on the uplifted side of these faults exposes Mesquite Schist-bearing gravels which underlie the surface terrace gravels (Figure 6B). These gravels lie approximately 18 km east of their probable area of deposition near the mouth of Mesquite Canyon.

North Side of the Lava Mountains:

A small isolated remnant of Quaternary(?) terrace gravels is surrounded by alluvium on the north side of the Lava Mountains about 3 km east of the Johannesberg-Trona Road (Figure 7, arrow). Alluvium in this area contains clasts of felsic plutonic rocks and volcanic rocks derived from the western Lava Mountains immediately to the south. In contrast, the surface of the terrace gravels contains many clasts derived from the Garlock Series and very few small clasts of Mesquite Schist. These gravels consist of debris eroded out of the Lava Mountains and deposited northward across the Garlock fault in the south end of Searles Valley. In order to contain the Garlock Series and sparse Mesquite Schist clasts, these gravels must have been deposited opposite an area about 1-2 km west of the mouth of Christmas Canyon about 17 km to the east (Figure 8).

Almond Mountain Volcanics

In the Lava Mountains east of the Summit Range, the Almond Mountain Volcanics rest with an angular unconformity on the latest Miocene Bedrock Springs Formation (see below), and are intruded or unconformably overlain by the Lava Mountains Andesite (Smith, 1964). The western volcanic center of this formation lies in the vicinity of Dome Mountain about 7 km south of the Garlock fault in the western Lava Mountains (Figure 7). This western center consists of brecciated intrusive porphyritic andesite intruded into several irregularly-shaped vents that formed domes immediately above the vents, and a much larger area of fragmental volcanic rocks surrounding the domes. Conglomerate, sandstone, and some tuffaceous sandstone, siltstone and claystone are interbedded within the fragmental volcanic rocks of this formation, mostly north of the western volcanic center about 5-6 km south of the Garlock fault. Volcanic lithologies typically constitute 50-90% of the clasts in the coarser sedimentary rocks, but locally layers contain considerable

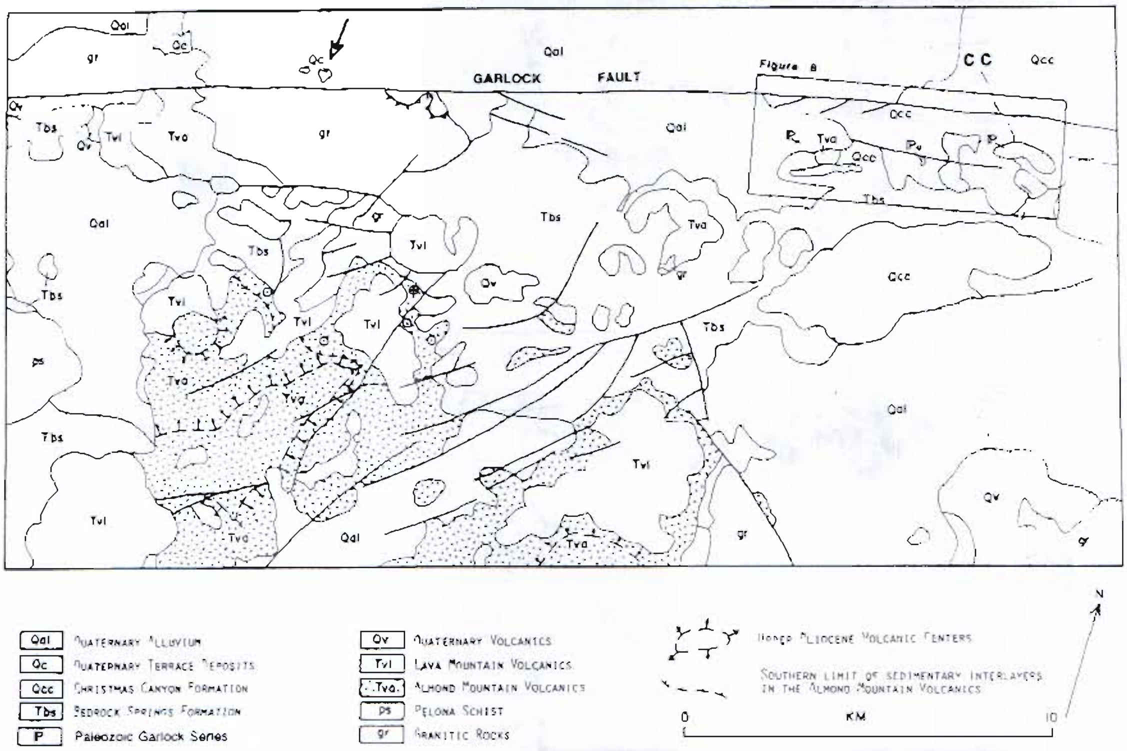

Fig. 7. Generalized geologic map of the Lava Mountains (after Smith, 1964). Volcanic centers of the western part of the Pliocene Almond Mountain volcanics are indicated, as well as the southern limit of sedimentary interlayers in the western part of the Almond Mountain volcanics. Tva (patterned)-Almond Mountain Volcanics, Tva (unpatterned)-other late Pliocene volcanics, circles with dots-conglomerate with no schist clasts, circle with cross-conglomerate containing a few schist clasts which closely resemble Mesquite Schist, CC-Christmas Canyon.

percentages of quartz, schist (nowhere very abundant) and quartz monzonite clasts. These sedimentary interlayers are present only north of the volcanic centers (Figure 7) and therefore non-volcanic clasts must have been derived by erosion of rocks to the north (Smith, 1964).

In a few places northeast of the western volcanic center, a very small percentage of clasts are schist which resembles the Mesquite Schist. These sub-rounded clasts are 1-10 cm in diameter and never constitute more than 0.1% of the clasts in the rock. These clasts differ from Garlock Series lithologies by their coarser average grain size, silvery sheen and spotted texture, and from the Rand Schist by their finer average grain size, silvery sheen and spotted texture. The only occurrences of these clasts are along the north-easternmost edge of the western volcanic center (Figure 7). Apparently a small component of Mesquite Schist clasts were included in the sediment deposited as alluvial fans derived from a source area to the northeast, on the north side of the Garlock fault. There is no possible source area for these clasts south of the Garlock fault. North of the fault the most likely source area from which these clasts could have been derived now lies about 32 km east of the present outcrop area of the Mesquite Schist (Figure 1).

Bedrock Springs Formation

The most extensive sedimentary rock unit in the Lava Mountains is the Bedrock Springs Formation which consists of more than 1500 m of coarse arkosic conglomerate, sandstone, siltstone and claystone, with smaller amounts of limestone, evaporites, tuff, tuff breccia, rubble breccia and lapilli breccia (Smith, 1964). The mammalian fauna, mostly collected from the lower third of the formation, is known as the Lava Mountains Fauna and is early Hemphillian (6-9 Ma) in age (Woodburne, 1987). Most of the younger sediments were probably deposited subaerially, mainly in a closed valley surrounded by alluvial fans that sloped from all directions toward a central lake or playa. Its elongation generally parallel to the Garlock fault suggests that the basin formed by downward faulting between the Garlock fault and the Brown's Ranch fault zone to the south (Smith 1964).

The basin center was apparently occupied by a playa during most of Bedrock Springs time. Surrounding the playa, extensive alluvial fan deposits on the east, south and west contain locally-derived pebbles of plutonic and volcanic rocks. Only in the northeast part of the basin do exposures contain abundant metamorphic clasts.

In this northeastern part of the Bedrock Springs basin, a series of highly brecciated metasedimentary rocks crop out on either side on Christmas Canyon (Figure 8). Most of these rocks consist of gray to yellowish-orange impure marble, and gray or gray-green slate and phyllite. These rocks clearly represent part of the Garlock Series (Dibblee, 1967, Carter, 1982), but since these are common lithologies in the Garlock series, it has not been possible to determine which part of the outcrop area in El Paso mountains they correlate with. In this area the Bedrock Springs Formation contains an abundance of fragments of these metamorphic rocks, showing that they cropped out and perhaps formed the northeast edge of the sedimentary basin. In most places these rocks are overlain by a thin veneer of gravels of the Christmas Canyon Formation. Brecciated metamorphic rocks occur as a discontinuous series of outcrops 700-1500 m south of the Garlock fault along a length of about 3 km. It is inferred that these outcrops represent several slide blocks which slid southward across the Garlock fault into the northern Bedrock Springs basin from a scarp on the north side of the Garlock fault about 1 km to the north and were subsequently buried by Bedrock Springs sediments. Subsequent erosion of Bedrock Springs sediment left these more resistant rocks standing out as hills, around and on top of which gravels of the Christmas Canyon formation accumulated.

About 2 km west of Christmas Canyon two small outcrops of Garlock Series metamorphic rocks are surrounded and overlain by gravels of the Christmas Canyon Formation which contain clasts of Mesquite schist up to 15 cm in size (Figure 8). Although not abundant (less than 1% of the clasts), Mesquite schist clasts are much more common in this area than elsewhere. Christmas Canyon Formation gravels elsewhere in the Christmas Canyon area contain small, extremely rare clasts of schist. It is inferred that a small slide block of Mesquite schist was incorporated in the Bedrock Springs sediments in this area along with more common blocks of Garlock Series rocks. Subsequent to Bedrock Springs accumulation, the schist was either completely removed by erosion or completely buried by gravels of the Christmas Canyon Formation so that now only clasts are found in the local gravels. This suggests that the contact between Mesquite Schist and Garlock Series north of the fault was about 2 km west of Christmas Canyon during emplacement of blocks of these lithologies into sediments of the Bedrock Springs Formation. This location is about 47 km east of this contact in El Paso Mountains today (Figure 1).

TOTAL DISPLACEMENT

Total left-lateral displacement of about 64 km was determined by Smith (1962) based on the offset of the Independence dike swarm and probably represents the total displacement on the Garlock fault.

Offset of a line separating areas of different regional fault trends suggests about 69 km of displacement on the Garlock fault (Michael, 1966) and the Layton Well thrust north of the fault is offset 56 km from the equivalent thrust in the Owlshead Mountains south of the fault (Davis and Burchfiel, 1973). A section of homoclinal east-dipping eugosynclinal metasedimentary rocks in Pilot Knob Valley south of the Garlock fault was correlated with rocks in the Garlock Series north of the fault by Smith and Ketner (1970), and have been described in detail by Carr, et al.(1993). These rocks crop out as isolated small hills that project through the alluvium about 3-12 km south of the Garlock fault. A Lower Mississippian sequence of distinctive argillite-pebble metaconglomerate and meta-agrillite in Pilot Knob Valley is correlated with similar rocks in the western part of the Garlock Series in El Paso Mountains 64 km to the west (Carr, et al., 1993).

DISCUSSION

Offset sedimentary units containing distinctive clast lithologies described in this study include (a) Pleistocene and upper Pliocene(?) gravels south of the fault which now lie as much as about 19 km east of their probable site of deposition near the mouth of Mesquite Canyon, (b) Quaternary(?) gravels north of the fault which now lie about 17 km west of their probable site of deposition near the mouth of Christmas Canyon, (c) conglomerates in the northern part of the Pliocene Almond Mountain volcanics in the western Lava Mountains south of the fault, which now lie about 32 km east of their probable site of deposition opposite the Mesquite Canyon area, and (d) large blocks of Garlock Series lithologies (and probably Mesquite Schist) within part of the late Miocene Bedrock Springs Formation south of the fault, which now lies about 47 km east of its probable site of deposition opposite El Paso Mountains near Mesquite Canyon.

If total offset across the Garlock fault is about 64 km, then the inception of left-lateral faulting on this fault must have been sometime in the late Miocene. The early Hemphillian fossils collected mostly in the lower part of the Bedrock Springs Formation indicate an age of about 6 to 9 Ma. It is not clear what part of the Bedrock Springs Formation contains the Garlock Series blocks near Christmas Canyon, but if these blocks were emplaced between 6 and 9 Ma, then this indicates an average displacement rate since late Miocene of about 5.5 to 8 mm per year. Extrapolating these rates to the total offset of 64 km would indicate that left-lateral faulting on the Garlock fault must have begun between 8 Ma and 12 Ma. This is in good agreement with previous studies (Carter, 1987, Burbank and Whistler, 1987, Loomis and Burbank, 1988).

On the northwestern side of El Paso Mountains Tertiary sediments of El Paso basin crop out about 10 km north and about 15 km west of the mouth of Mesquite Canyon. The Ricardo Group is a 1700 m sequence of Miocene volcanic rocks and continental sedimentary rocks deposited between about 19 and 7 Ma in the basin between El Paso Mountains and the southern Sierra Nevada. Magnetostratigraphic and radiometric studies of these rocks indicate rotational histories which provide constraints for the initiation of strike-slip movement on the Garlock fault (Loomis and Burbank, 1988). Beginning about 10 Ma, El Paso basin began to be rotated counter-clockwise, at the same time that there was a sharp increase in sediment-accumulation rates. These events are interpreted to have occurred as a consequence of the initiation of left-slip motion along the Garlock fault near the southern margin of the basin, along with the beginning of basin-and-range-style extension north of the fault (Loomis and Burbank, 1988).

The data reported in this study are consistent with a generally uniform rate of long-term offset on the Garlock fault since its inception in late Miocene time. Pleistocene to late Pliocene(?) offset of 19 km,. Pliocene offset of 32 km, and late Miocene offset of 47 km are all consistent with a long-term rate of 5.5 to 8 mm per year. These long-term rates, in turn, are consistent with the Holocene rate of 7 mm per year measured on the offset gravel bar on the east side of Koehn Lake. This data is also consistent with measured slip rates of 5-11 mm/^{14}C-yr in southeastern Searles Valley (McGill and Sieh, 1993).

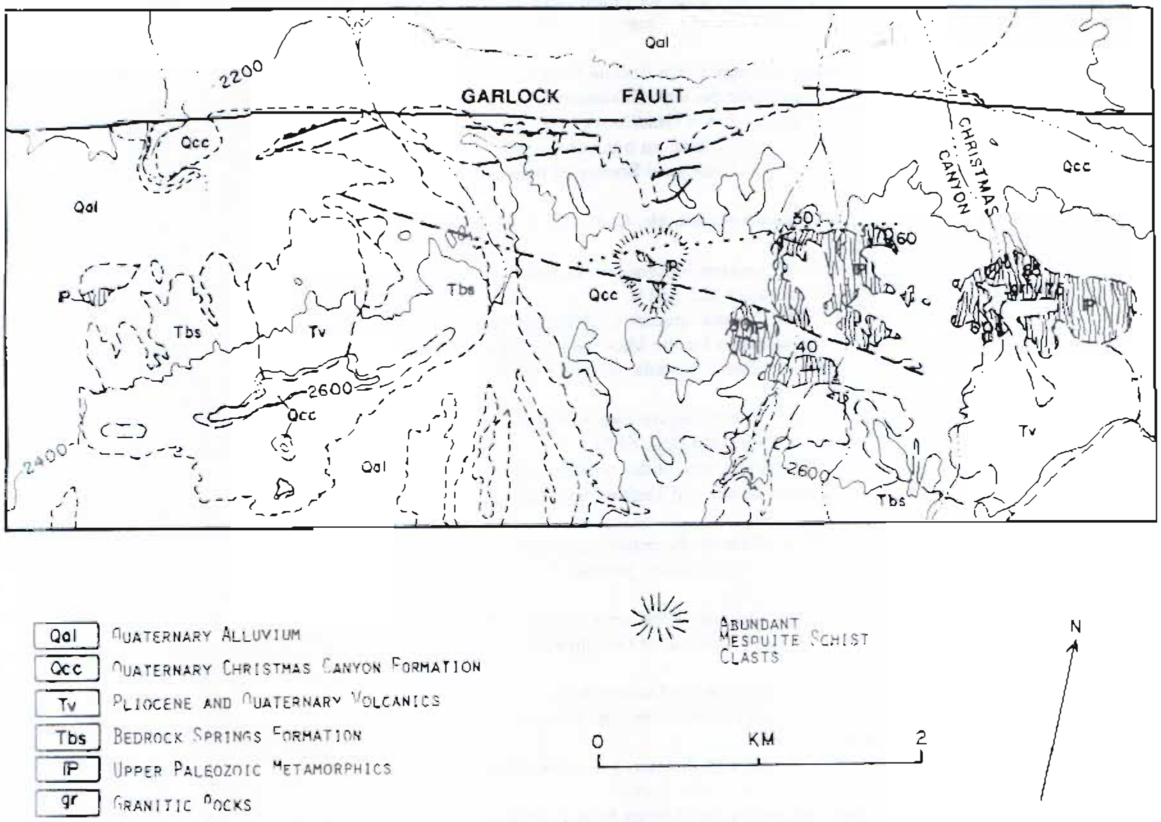

Fig. 8. Geologic map of the Christmas Canyon area in the eastern Lava Mountains (after Smith, 1964). Outcrop areas of Garlock Series rocks probably represent large blocks which slid southward across the Garlock fault into the accumulating Bedrock Springs Formation. Clasts of Mesquite Schist occur in gravels of the Pleistocene Christmas Canyon Formation in a small area about 2.5 km west of Christmas Canyon.

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The Butte Valley Fault in the Panamint Range and the Layton Well Thrust in the Slate Range: Their Relation to Faults South of the Garlock Fault

Chester T. Wrucke, *U.S. Geological Survey, Menlo Park, California*, and **Calvin H. Stevens**, *Department of Geology, San Jose State University, San Jose, California*

Introduction

The Butte Valley Fault and Layton Well Thrust are prominent structural features in adjacent mountain ranges north of the eastern end of the Garlock Fault in southern California (fig. 1). The Butte Valley Fault at the southern end of the Panamint Range can be traced along an arcuate path, convex to the north, for an east-west distance of about 18 km, about 37 km north of the Garlock Fault (Wrucke and others, 1995). To the west, the Layton Well Thrust is well exposed in the southern Slate Range, where it can be followed southward to about 6 km north of the Garlock Fault. The Butte Valley Fault and the Layton Well Thrust have been considered parts of the same fault (e.g. Snow, 1992) but more recently have been interpreted as separate features (Wrucke and others, 1995).

The Butte Valley Fault is poorly known because of lack of exposure and the limited extent of its trace. The Layton Well Thrust is better known because it is well exposed. It is aligned with the Argus-Sterling Thrust in the Argus Range (Moore, 1976) immediately north of the Slate Range and with the Davis Thrust in the Darwin Hills (Hall and Mackevett, 1962) 25 km northwest of the Slate Range. Thus, the Layton Well Thrust apparently is a continuation of the East Sierran thrust system (Dunne, 1986). The Layton Well Thrust and the East Sierran thrust system have been considered by some to extend south of the Garlock Fault (Davis and Burchfiel, 1973; Walker and others, 1990; and Walker and Martin, 1991), though a different interpretation has been postulated by T.L. Pavlis (this volume). Whether or not the Butte Valley Fault and the Layton Well Thrust are present south of the Garlock Fault is an important topic of this field conference and is significant to an understanding of the Garlock Fault.

Butte Valley Fault

The Butte Valley Fault, a major structural feature of the southern Panamint Range, can be traced nearly across the range in the vicinity of Butte Valley and Warm Spring Canyon (fig. 1). The Butte Valley fault that Johnson (1957) mapped in the vicinity of Goler Wash (BVF, fig. 1) is, we believe, not a southward continuation of the Butte Valley Fault farther north because the fault in Goler Wash has a considerably smaller stratigraphic throw and the juxtaposed rocks are much less deformed than those near the Butte Valley Fault in Butte Valley and Warm Spring Canyon. The Butte Valley Fault everywhere has been intruded by granite or covered by younger deposits, though its location is well defined in a narrow zone that separates rocks of greatly different ages and structural styles. At one locality, Paleozoic and Mesozoic rocks to the south of the fault are separated from Proterozoic rocks to the north by a narrow septum of granite 30 m wide. This juxtaposition represents a stratigraphic throw of 10 km.

Rocks to the south of the fault consist of Devonian through Triassic sedimentary strata and Jurassic(?) metavolcanic deposits. The Paleozoic rocks are typical of the miogeocline from the southern Death Valley region to the Inyo Mountains, and the Jurassic(?) rocks consist

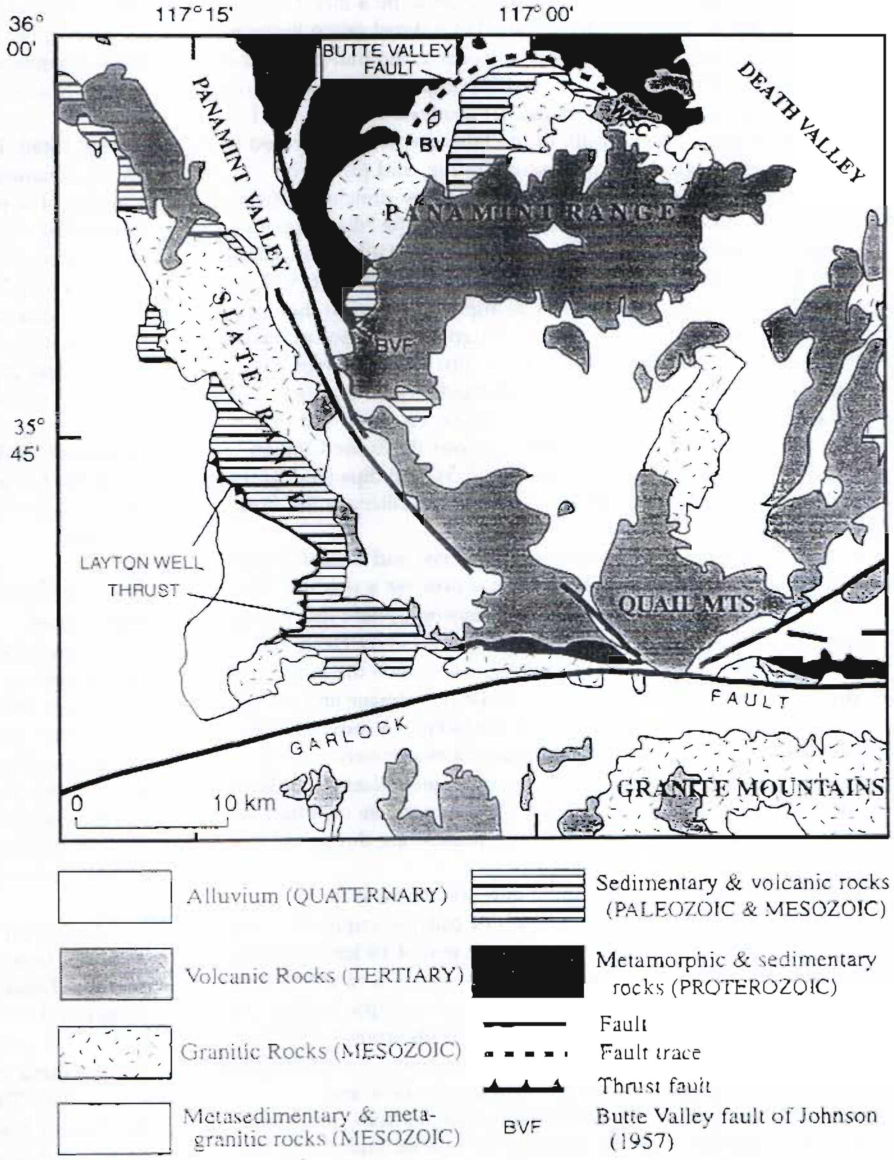

Fig. 1. Generalized geologic map of the Granite Mountains - Panamint Range - Slate Range region. Modified from Jennings and others (1962), Smith and others (1968), and Wrucke and others (1995). **BV**, Butte Valley; **WSC**, Warm Spring Canyon.

principally of andesite flows and less abundant silicic flows and tuffs that are similar to volcanic sequences in the Inyo Mountains, Alabama Hills, and Slate Range. Although the Jurassic(?) metavolcanic rocks in the Panamint Range have not been dated directly, radiometric ages from the volcanic sequence in the Alabama Hills and Inyo Mountains are about 169 Ma (Dunne and Walker, 1993) and are 150 to 148 Ma in the Slate Range (Dunne and others, 1994).

Strata north of the fault trace consist of Early Proterozoic schists and foliated granitic rocks, Middle and Late Proterozoic strata of the Pahrump Group, and the Late Proterozoic Noonday Dolomite, Johnnie Formation, and Stirling Quartzite. The Proterozoic rocks extend to the northern end of the southern Panamint Range and are overlain by Paleozoic formations about 25 km north of the Butte Valley Fault (Hunt and Mabey, 1966).

Granitic rocks that intrude rocks deformed by the fault have yielded recalculated K-Ar ages of 155 to 148 Ma (Armstrong and Suppe, 1973; Stevens and others, 1974). These ages and the tentative correlation of the metavolcanic rocks with rocks in the Alabama Hills, Inyo Mountains, and Slate Range suggest that the Butte Valley Thrust is Middle to Late Jurassic in age and older than about 155 Ma.

The Butte Valley Fault has been considered to be a thrust fault by Wrucke and others (1995). Evidence favoring a thrust origin includes the facts that (1) the Proterozoic rocks adjacent to the mid-part of the fault trace locally are overturned in a south-vergent anticline, as though dragged along a reverse fault or thrust, (2) southward dips in the Late Paleozoic rocks immediately south of the fault are steep compared to those farther south, again as if representing drag, and (3) newly observed strong lineations have been produced by stretched siltstone and limestone pebbles in a debris-flow deposit in the Mississippian Stone Canyon Limestone immediately south of the fault trace. These stretched pebbles have a length to width ratio of 5:2 to 14:1, with a mean of about 5:1, and the lineations plunge 55 degrees to the north. This stretching of pebbles is suggestive of significant tectonism, and the plunge suggests that the Butte Valley Fault dips at a moderate angle to the north. At the westernmost outcrop of middle Paleozoic rocks at the northeastern margin of Butte Valley, a small reverse fault that places the Lost Burro Formation of Devonian age over the Stone Canyon Limestone of Mississippian age strikes N 80° W and dips 55 degrees to the north. This fault is subparallel and perhaps auxiliary to the Butte Valley Fault.

In contrast to the thrust-fault hypothesis, Davis and Burchfiel (1997) have suggested that the Butte Valley Fault represents a segment of a circular fault at the margin of a Jurassic collapsed caldera that is largely hidden beneath younger deposits. Arguments raised by Davis and Burchfiel in support of the collapse hypothesis include the fact that (1) the dip of the fault, now obscured by granite or younger units, could have been steep, (2) the arcuate trace is not likely the result of post-fault folding, (3) they know of no mylonitic fabrics present in rocks associated with the fault, and (4) the generally steep southward tilt of the Paleozoic strata near the fault trace and more gentle dips farther away would be expected consequences of foundering during caldera collapse.

In reply to the Davis and Burchfiel interpretations, Wrucke and others (1997) pointed out that the amount of collapse required would be much greater than is common to calderas and at 10 km would be at the maximum known for calderas (Lipman and others, 1996). The caldera hypothesis also is questioned because no collapse breccia has been found as would be expected near steep walls around a deeply dropped caldera floor (Lipman, 1984), though erosion conceivably has reached below the original floor. In addition, extensive and voluminous silicic outflow that would have resulted from compound collapse and repeated eruptions required for caldera foundering of 10 km is not known. Assuming that the caldera had a diameter of about 18 km, as suggested by the curvature of the fault trace, the eruptive material required to produce collapse of this magnitude would have had a volume on the order of 2,500 km^3. Although the outflow pyroclastics might have been eroded away, the apparent absence of such a large volume of material is a significant problem if the Butte Valley Fault resulted from volcanic collapse.

Layton Well Thrust

The Layton Well Thrust, which presently dips at low angles to the west, is the dominant structural feature of the southern Slate Range. The lower plate consists of a thick section of Jurassic metavolcanic and volcanogenic sedimentary rocks, tuffaceous sandstone, and breccia or conglomerate containing plutonic clasts (Smith and others, 1968). Mesozoic intrusive rocks in the northern Slate Range separate these metavolcanic rocks from Mississippian through Lower Triassic marine sedimentary strata that also are considered to be in the lower plate (Smith and others, 1968; Stone, 1984). Equivalent Paleozoic rocks lie beneath Mesozoic volcanic sequences to the northwest in the Argus Range and in the Inyo Mountains. Volcanic rocks in the lower plate of the Layton Well Thrust have yielded Late Jurassic U-Pb ages of 150 to 148 Ma (Dunne and others, 1994) and are considered to be broadly contemporaneous with volcanic sequences in the Alabama Hills, Inyo Mountains, and the Panamint Range

The upper plate of the Layton Well Thrust consists of locally folded, schistose, and mylonitized metasedimentary, metavolcanic, and metaplutonic rocks (Smith and others, 1968). Upper plate rocks originally were considered to be Precambrian in age (Smith and others, 1968). However, measured $^{87}Sr/^{86}Sr$ ratios in the range of 0.708 -0.709 obtained by Wooden (Wrucke and others, 1995) from two samples of upper plate schist are too low for Precambrian and are compatible with regional values for Mesozoic plutonic rocks. These data, together with the fact that the schists are unlike those in the Death Valley-Inyo Mountains region, suggest that the upper plate rocks are Mesozoic in age.

A granitic pluton at the south end of the Slate Range is interpreted as intruding upper and lower plates of the Layton Well Thrust (Wrucke and others, 1995). Intense shearing locally in the pluton near its margin implies late reactivation of the thrust or that the granite is syntectonic. U-Pb ages of ~147 Ma from this body and a U-Pb age of ~148 Ma for another pluton in the upper plate have been reported by Dunne and others (1994), who interpreted the age of thrusting as about 148 Ma.

Regional implications of the Butte Valley Fault and the Layton Well Thrust

An important question for determining the amount of displacement and the history of the Garlock Fault is, are the Butte Valley Fault and the Layton Well Thrust present south of the Garlock Fault? The Butte Valley Fault has not been recognized south of Butte Valley, either north or south of the Garlock Fault. Perhaps the upper plate of the Butte Valley Fault, if it is a thrust, was completely removed by erosion south of its identified trace before being intruded by granite or covered by younger deposits. If the fault is the margin of a collapsed caldera, it would not be expected to extend south of the Panamint Range.

The Layton Well Thrust closely approaches the Garlock Fault, but it has not been identified unequivocably south of that fault. A thrust fault that Davis and Burchfiel (1973) reported from reconnaissance studies in the Granite Mountains, 65 km east-southeast of the Slate Range, as a continuation of the Layton Well Thrust is now considered to be a much younger fault (T.L. Pavlis, this volume). Thus, the Layton Well Thrust evidently also is not exposed immediately south of the Garlock Fault. An important topic for further consideration is whether the Jurassic thrust in the Cronese Hills south of the Granite Mountains should still be considered part of the East Sierran thrust system, as suggested by Walker and others (1990), and therefore represents a continuation of the Layton Well Thrust.

Acknowledgements

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Rifted Volcano in Wingate Wash, Death Valley Region, Southeastern California

David L. Wagner, *California Department of Conservation, Division of Mines and Geology, Sacramento, CA 95814*

Introduction

Wingate Wash is a northeast-southwest trending topographic trough that separates the Panamint Range from the Owlshead Mountains in the Death Valley region of southeastern California (Figure 1). Famous as an escape route from Death Valley used by 49ers, and later as the route for the 20-mule team borax wagons, the area is now seldom visited because it is nearly surrounded restricted military land (Photo 1). Recently it was designated wilderness, making access even more difficult. The geology of Wingate Wash is poorly known; geologic mapping has only been regional reconnaissance (Wagner and Hsu, 1988) or is unpublished. The northeast trend of Wingate Wash is anomalous to the north-northwest trend typical of Great Basin intermontane basins and its straightness suggests the trough is fault-controlled. A zone of faults, parallel to the trend of Wingate Wash was mapped by Wagner and Hsu (1988), was interpreted to be surface exposures of the Wingate Wash fault observed on seismic profiles by deVoogd and others (1986). On the north side of Wingate Wash there is an eroded volcanic center interpreted by (Wagner, 1988) to be the remains of a volcano rifted by movement along the Wingate Wash fault during a deformational episode 12 to 14 million years ago. This paper presents an update of this interpretation and gives data constraining the time of deformation.

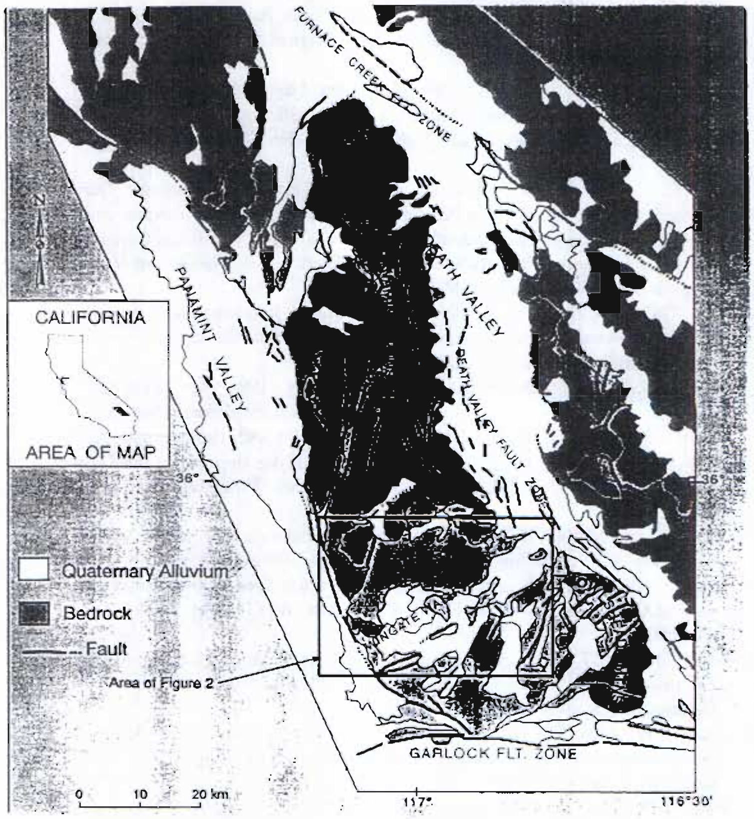

Fig. 1. Generalized map of Death Valley region.

Geology of Wingate Wash

Figure 2 is a geologic map of the Wingate Wash area. Miocene age volcanic rocks overlie a basement of Mesozoic granitic rocks, mostly Cretaceous quartz monzonite, and older metamorphic rock. Lacustrine and fanglomerate deposits overlie the volcanic rocks. Figure 3 is a schematic stratigraphic column of Wingate Wash.

Photo 1. View to the west across Wingate Wash toward Wingate Pass and the Slate Range. Most of this area is restricted land of the China Lake Naval Weapons Center. In the left central part of the photo are the Cyrstal Hills.

Miocene volcanic rocks

An accumulation of andesitic breccia, pyroclastic rocks, lahars, flows, volcaniclastic rocks, dacitic to silicic pyroclastic rock, and basalt flows, more than 2,000 meters thick in places, is exposed throughout Wingate Wash, in the adjacent Panamint Range and in the Owlshead Mountains.

Intrusive volcanic rocks including plugs, dikes and volcanic centers are common, particularly along the north side of Wingate Wash. The most prominent intrusive center is a volcano that has been eroded down to its intrusive root (Photos 2a,b). A basalt collected from a mafic part of the root yielded a whole-rock K/Ar age of 14.0 ± 0.3 Ma (determination by R.E. Drake; L.A. Wright personal communication). Overlying the intrusive core are cream-colored pyroclastic deposits of tuff, tuff breccia, and lapilli tuffs that are probably ignimbrites deposited within the crater itself. Intense alteration is evident throughout this volcanic center as well as the other, less well defined centers, that occur along the northern side of Wingate Wash. Flanking the

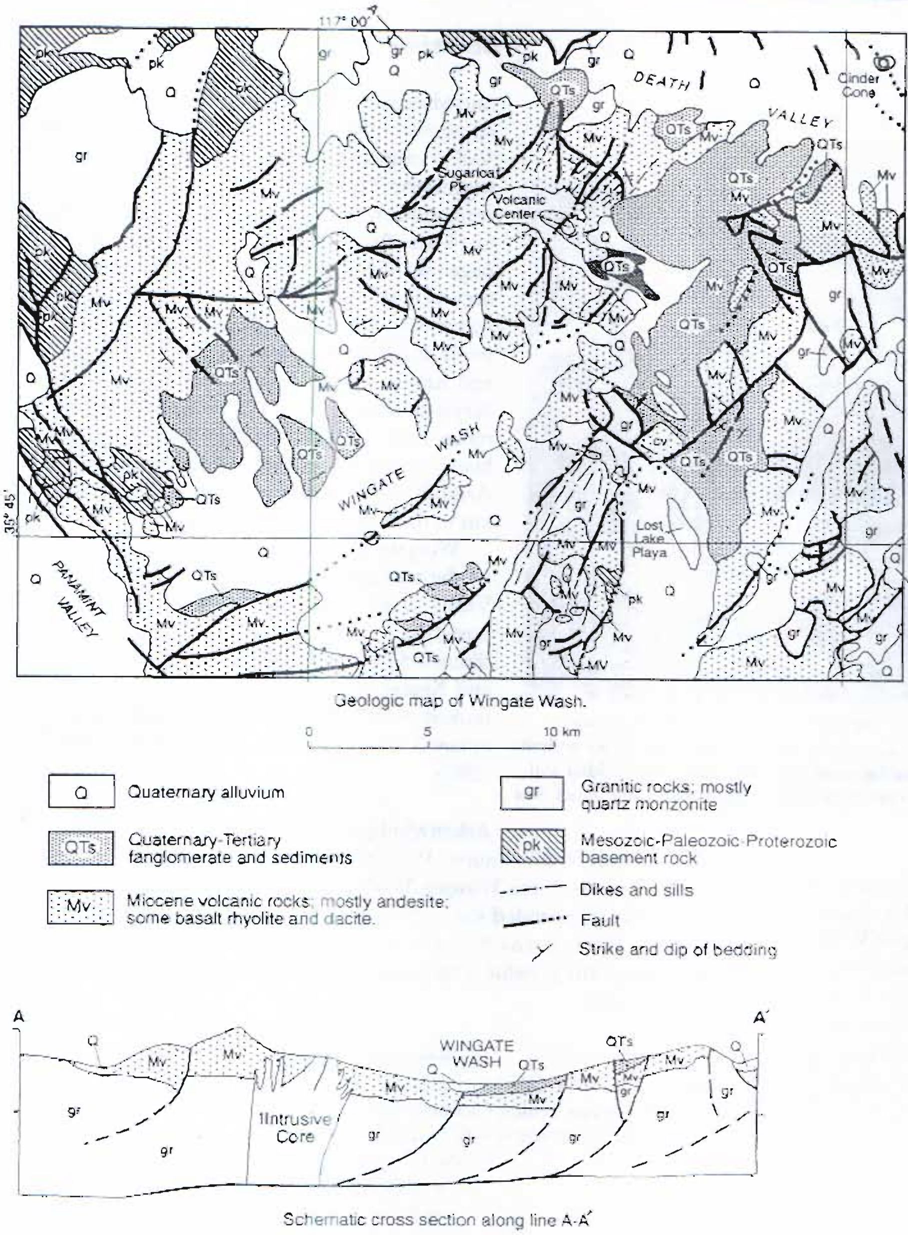

Fig. 2. Geologic map of Wingate Wash and surrounding area.

intrusive root of the volcano are stratified breccia, flows, and tuff that are interpreted to be near-source facies that made up the volcanic cone (Photo 2a). These stratified rocks dip away from the intrusive core in three directions. Significantly, the southeastern flank of the cone is missing (Photo 2b). South of Wingate Wash, in the Owlshead Mountains between the Owl Lake and Lost Lake playas, are andesitic to basaltic breccia, flows, and tuffs that range in age from 12 to 14 Ma (Davis and Fleck, 1977). Wagner(1988)interpreted these rocks as near-source facies that are the missing southeast flank of the volcano in Wingate Wash. This interpretation may be simplistic. Serpa and Pavlis (1996) have presented a reconstruction of the Death Valley region that implies a more complicated history for Wingate Wash and the Owlshead Mountains.

Fanglomerate and lacustrine deposits

Thick deposits of fanglomerate unconformably overlie the Tertiary volcanic rocks. Locally, lacustrine deposits consisting of expansive clay, silt, sand, volcanic ash, chert and limestone occur at the base of the fanglomerate. An ash in lacustrine deposits in the central part of Wingate Wash has been correlated with 12 million year-old ashes using tephrachronologic techniques (Andre Sarna-Wojcicki, 1995, written communication). Earlier, this ash was believed to be about 10 Ma in age (Andre Sarna-Wojcicki, 1988, written communication). Thus, the fanglomerate deposits are considered to be Miocene-Pliocene in age. Gravels of the fanglomerate consist primarily of cobble to boulder-size volcanic detritus although there are local infusions of granitic material from nearby exposures of quartz monzonite. The new correlation suggesting a 12 Ma age for the ash provides support for the inference of Serpa and Pavlis (1996) that the volcanic and sedimentary rocks in this area were deposited in a basin that included the Artist Drive formation.

Faulting in Wingate Wash

The dominant trend for major faults mapped in Wingate Wash is northeast, parallel to the trend of the trough. The northeast trending faults are steeply dipping to vertical and display oblique, dip-slip movement.

A steep escarpment bounds Wingate Wash along the northwest flanks of the Owlshead Mountains. The escarpment is the result of down-to-northwest dip-slip movement along northeast trending faults. The basal contact between Miocene volcanic rocks and Cretaceous granitic basement is exposed on the crest of the escarpment. Miocene volcanic rocks and the overlying fanglomerate deposits have been displaced a minimum of 500 meters along these high angle faults in Wingate Wash.

Deep seismic profiles were interpreted (deVoogd and others, 1986) to show a fault, the Wingate Wash Fault, striking parallel to the trend of Wingate Wash and dipping northward beneath the Panamint Mountains at a angle of 25 degrees. Mapping by Wagner and Hsu (1988) and geophysical work of Grabyan (1974) provides support for the presence of the Wingate Wash Fault Zone. Where exposed at the surface, individual faults within the zone are steeply dipping or vertical, and display down to the north displacement with oblique movement. Faults along the escarpment between Wingate Wash and the Owlshead Mountains are curved indicating they are listric dish-shaped faults that may merge into a master fault zone at depth.

Evidence for detachment

Northwestward extension across Wingate Wash is suggested by alignment of dikes parallel to the N 45 degrees E trending Wingate Wash (Figure 2). Assuming the least principal stress direction is normal to the dikes, a N 45 degrees W extension direction is indicated. In addition to these paleostress indicators, there are several lines of evidence that indicate a structural discontinuity exists across Wingate Wash.

In the central part of the wash, the Miocene volcanic rocks and the overlying lacustrine beds and fanglomerate deposits are folded into a northwest trending, southeast plunging anticline. This fold cannot be traced across the axis of the trough. To fold the rocks of the south side of the wash and leave those on the north unaffected suggests that the two sides of the wash were not continuous during folding.

Photo 2a. The volcanic center on the north side of Wingate Wash. The light-colored rocks in the middle background are pyroclastic deposits associated with the intrusive core of the volcano. Near vent facies andesitic and basaltic breccia, halars and flows make up the ridge in the background. The breccia is well bedded and dips away from the intrusive core. The dark rocks in the middle part of the photo are andesite and basalt flow within Wingate Wash.

Most of the extension must have taken place between 14 and 12 Ma as it did immediately to the south in the Owlshead Mountains (Davis, 1988). The oldest dates on volcanic rocks in Wingate Wash and in the Owlshead Mountains are about 14 Ma(Davis and Fleck, 1977). In Wingate Wash, lacustrine sediments containing the 12 Ma ash unconformably overlie volcanic rocks thereby constraining the age of the volcanism and formation of Wingate Wash between 14 to 12 Ma.

Photo 2b. The rifted volcano. The southeast flank of the volcano is missing.

Discussion

Wingate Wash formed during a major mid-Miocene extensional episode that apparently affected most, if not all, of the Death Valley extended terrane. Data presented here suggest the Panamint Range may have been continuous with the Owlshead Mountains prior to 14 Ma. As northwest-directed extension began at 14 Ma, intense syntectonic volcanism occurred with at least one and possibly more volcanoes forming along a northeast-southwest trend of intrusive centers. By 12 Ma volcanism had largely ceased. A recent reconstruction by Serpa and Pavlis (1997) suggests the volcanic rocks in and around Wingate Wash may have been continuous with the volcanic rocks of the Artist Drive formation now exposed 30 to 40 km to the east across Death Valley.

Wingate Wash is truncated by Death Valley on the east and very abruptly by Panamint Valley on the west. This east-west extension, post-dating the northwest-directed extension discussed here, is related to left lateral (Smith and Ketner, 1970; Davis and Burchfiel, 1973) movement along the Garlock Fault that was initiated about 10 Ma (Loomis and Burbank, 1988).

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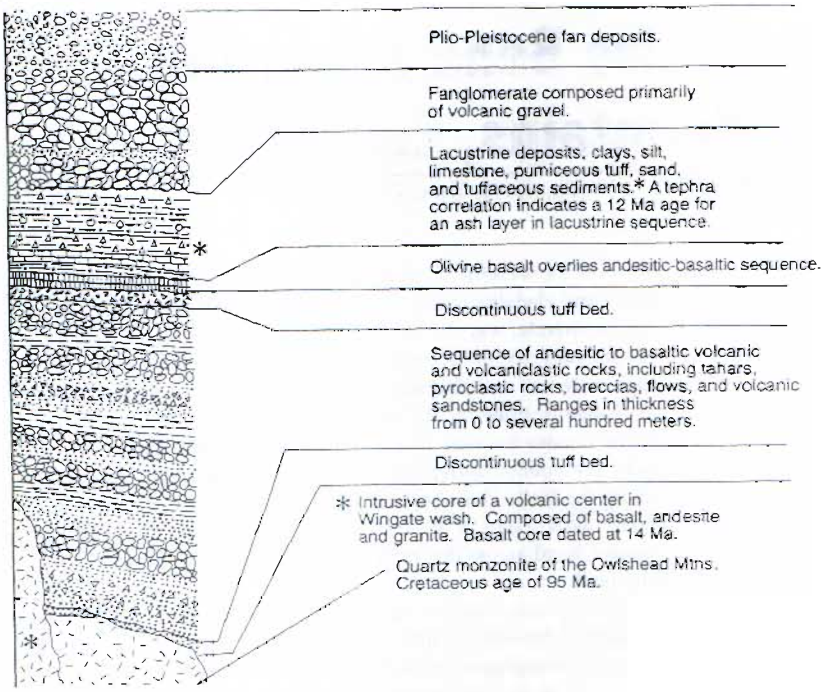

Fig. 3. Stratigraphic column on Wingate Wash area.

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Note on the Major Transpressive Bend along the Garlock Fault at the Quail Mountains, California

W.R. Muehlberger, *Department of Geological Sciences, The University of Texas at Austin, Austin, Texas 78712*

The Quail Mountains in southern California have yet to be studied with care. My three months of field work during the fall of 1950 was terminated by being recalled into active duty in the U.S. Marine Corps. When I was released, this region was under military control with no access being allowed. To record what had been done, I published a map sheet (Muehlberger, 1954) on a planimetric base at a scale of 1:48,000 (mapping was done at a scale of 1:9,600).

At the time of my field work, the only publications near my area were of the Randsburg quadrangle (Hulin, 1925) and southern Death Valley (Noble, 1941), both areas being at least 60 miles distant. The area was selected by D. Foster Hewett (a saint of a man) because it offered the possibility of determining the displacement across the Garlock fault. In addition, Hewett financed the field work and bought and moved a 12-foot-long house trailer to Leach Spring (about 65 miles north of Barstow and a similar distance west of Death Valley) for my use (this trailer was used by many other geology students of the Mojave Desert who also were "foster" children of Hewett).

The correlation and age of units in the Quail Mountains were mostly suggested by Hewett because of his extensive field mapping to the east (Hewett, 1931, 1956) as well as a comprehensive knowledge of the entire Mojave region. None of the units have been independently dated by fossils or isotopic methods. Some shown on the map are mislabeled (pCs and pCv on the map are referred to in the text and should be relabeled pJs and pJv, respectively) and others (with today's knowledge) cannot be what the label proclaims (ex., Archean, for the mylonitized granitic gneiss).

While reviewing the suite of thin sections that I used to illustrate the progressive mylonitization of the granitic gneiss in the Quail Mountains, Sharon Mosher recognized that the feldspars had undergone plastic deformation- something not possible in the present brittle environment. One of her students, Keith Pollman (1983, Pollman and Mosher,1982) studied all of the deformed rocks in the map area using the samples that I had collected. Microprobe analyses showed that the chemistry of the new albite growth was consistently different from the host grain. The granite gneiss exhibits increasing levels of ductile deformation approaching faults, culminating in total recrystallization of all minerals. Deformation of adjacent Mesozoic granitoids and Tertiary rhyolite is purely brittle as seen in thin section. Thus, he proposed a two stage history for the Garlock fault: an earlier deformation of unknown age conducive to ductile deformation of the granitic gneiss, with the present-day Garlock fault zone being due to the reactivation of the earlier fault.

Unfortunately none of my samples were oriented so that the shear fabric in the specimens could not be related to the outcrops. Rachel Burks (1987,1988), in a field study of the granitic gneiss, determined that the ductile fabric, as recorded by S-C planes, asymmetric porphyroblasts, and obliquely recrystallized quartz ribbons indicate top-to-the-northwest shear under greenschist facies metamorphism. This unit is now in fault contact along its southern margin with Mesozoic intrusives that are extensively cut by diorite dikes and along its northern margin by Tertiary volcanic rocks. All units are cataclastically shattered into angular fragments under 5 cm across.

The recognition that the granitic gneiss is the only unit that has been plastically deformed points to the present-day Garlock fault as a younger event that is responsible for the intense cataclastic crushing of the Mesozoic granitic rocks as well as the overlying Tertiary volcanic rocks. The capping stratigraphic unit (labelled QTf on the published map sheet) of the western block of the Quail Mountains contains slightly sheared rhyolite clasts. Their source, as well as the other clasts in the conglomerate, was to the south where the outcrops of relatively intensely sheared rhyolite is exposed. The source areas are now topographically low and drainage from the source areas is to the south.

The change in strike by 40° from west to east along the Garlock fault as one crosses the map area demonstrates that the Quail Mountains are uplifted along the other side of the giant transpressive bend along the fault. This also explains why there is a belt of recent and slightly older traces of the fault zone along the inner side of ther transpressive bend across this map area that are seen nowhere else along the fault.

The set of arcuate scarps south of the principal recent trace of the Garlock fault all have north side up components of slip. The locally exposed older stratigraphic units that lie with an angular unconformity under the capping alluvial fan deposits have consistently steep dips (35°+) that could be interpreted to result from reverse faulting because of the transpressive bend.

The wide belt of cataclastically crushed rock on either side of the mylonitized granitic gneiss must also be a reaction to the transpressive bend. The already mylonitized material being weak allowing for north-south shortening and an upward stretching. This would generate the relief to produce the coarse gravel (QTf) cappping the range that contains slightly sheared rhyolite clasts that presumably is derived from the sheared rhyolite now exposed to the south. The shearing of the rhyolite then must have continued after the deposition of the capping gravel to account for its present state. The wide band of crushed rock with its small fragment size would be easier to erode than the intact units to the north and thus would account for the present inversion of topography across the range.

None of the major faults that outline the Quail Mountains can be traced to an intersection with any of the others. All project toward a common point at the south point of the central block of the Quail Mountains. The Garlock fault as traced eastward across the map area steps right and reappears beyond to continue eastward at a strike that is rotated more than 40° clockwise. Both the Brown Mountain and Owl Lake faults are unmappable to where they project to the step of the Garlock fault. The "intersection" area contains a south-facing monocline and small(?) low-angle thrust faults that appear to override the Garlock fault trace where it steps over.

An interpretation of the fault pattern suggests that the central Quail Mountains are pivoting counterclockwise around a vertical axis. This makes the Brown Mountain fault a right-lateral fault, as well as having a northeast side up component. Farther northwest, the Brown Mountain steps west and becomes the Panamint Mountain fault, which Zhang et al (1990) describe as a right-lateral fault cutting alluvial fans and a dip-slip range-front fault with a slip rate of 2.36±0.79 mm/yr.

The Owl Lake fault along the central Quail Mountains side has an interbedded volcanic and clastic sequence of units that is folded around vertical fold axes. Where the beds approach the fault they turn to vertical to overturned with top toward the fault. This suggests that the Owl Lake fault is a steep reverse fault near where it projects into the Garlock fault. However, the only place I found a fault plane was near the southwest end of the mappable fault and there it is dipping 83° south. The trace of the fault is so linear that it must be steep to vertical. The Owl Lake fault is a left-lateral fault with north side up until it reaches the Owl Lake basin to the northeast of the map area, where it has the north side down. McGill (1993) showed that it is still active with a slip rate of 1-3 mm/yr, a value essentially identical to that for the southern Panamint Valley fault. Both of these rates are less than estimated for the Garlock fault zone west of the map area (3-9 mm/yr, McGill and Sieh, 1991).

These notes emphasize the importance of additional field work to determine ages on the various units and to prove or disprove the interpretations of structural history proposed here.

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The Iconography of the Bighorn Sheep Petroglyphs in the Western Great Basin

David S. Whitley, *W&S Consultants, 447 Third St., Fillmore, CA 93015*

Archaeologists for almost a century denied a direct connection between Mojave Desert petroglyphs and historical Native Americans, insisting that the engravings originated somewhere deeper in the prehistoric past, and that no information could be obtained on their meaning from Native Americans (e.g., Heizer and Baumhoff 1962; Steward 1968). Research conducted over the last decade has proven both these assumptions false, with substantial information on this art uncovered once serious efforts were made to examine the ethnographic record (e.g., Whitley 1992, 1994a). The connection between at least some of the petroglyphs and historical Native Americans has also been confirmed by dating research. Not only does chronometric varnish dating support the recency of some of the motifs, but historical motifs (e.g., horses and riders) have been discovered which at most are only a few hundreds years old (Whitley and Dorn 1987).

The ethnographic data provide a coherent and straightforward explanation for the making and meaning of Mojave Desert petroglyphs, which is supported by numerous Native American informants in a variety of different ways (e.g., direct accounts, word etymologies, place-names). Petroglyphs were made by shamans at the conclusions of their vision quests, and they portray the visionary imagery of their trance states. Shamans underwent these trances to visit the spirit world and to obtain (and manipulate) supernatural power, often in the form of an animal spirit helper. Rock art sites served as vision quest locales even though, in some cases, they were located close to villages (Whitley 1994a, 1996, 1998a).

Because this art is intended to portray the visionary imagery of trance, neuropsychological research on the mental imagery and bodily and emotional effects of altered states of consciousness has also proven useful in understanding the petroglyphs (e.g., Whitley 1994b, 1998b). The basis for this research is the biological fact that all humans are neurally "hardwired" in the same way. The result is that human hallucinations are chemically and structurally similar, although individual variability obviously occurs on more specific levels. A number of important points derive from this neuropsychological research:

(1) Trance results in neurochemical changes in the brain that diminish short-term memory formation; specifically, reductions in the neurotransmitters norepinephrine and serotonin (Hobson 1994:211). Diminished short-term memory was apparently the cause for the creation of the petroglyphs; shamans needed to engrave them on the rocks so that their memory of these supernatural experiences would not be lost (Whitley 1998b).

(2) The mental imagery ("visual hallucinations") of trance typically progresses through three stages (Lewis-Williams and Dowson 1988). First, geometric light percepts, called "entoptics" or "phosphenes", are experienced. These are caused by neural firings in the optical system and brain. Second, an individual attempts to "make-sense" of these entoptics, using the visual recognition systems in the visuo-motor areas of the brain, thereby construing them as iconic or representational forms. Because the visual recognition of forms and shapes employs long-term memory associations, these geometric-entoptic patterns are construed as images that are culturally and personally meaningful to the individual. Third, the entoptic and iconic images may combine, or the iconics may appear alone. In addition to this progression in trance imagery, altered states also involves a series of changes in visual perception; e.g., one image may be superimposed on top or

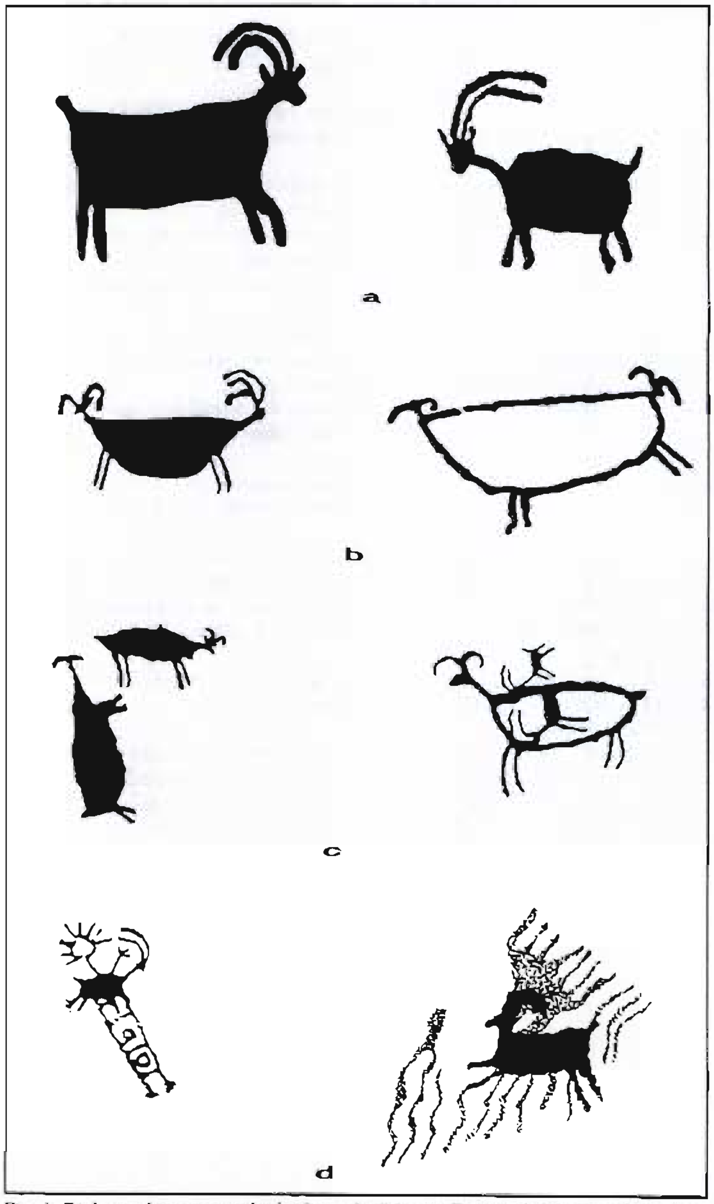

Fig. 1. Bighorn sheep petroglyphs from the Mojave Desert, California, displaying characteristics of trance imagery. **(a)** Many bighorn petroglyphs originated in the construal of a common entoptic form perceived during trance, nested curves, as iconic forms, the horns of the sheep, but the result is an anatomically impossible animal. The example on the left has a head and snout in profile, ears seen from the front, and horns in an unnatural position and perspective, as does the petroglyph on the right. **(b)** "Two-headed" sheep petroglyphs, reflecting the fact that these are not intended to portray "real" animals but instead spirit beings. **(c)** Many petroglyph compositions also incorporate characteristics of trance imagery. The rotation of images off a "normal" horizontal plane is one of these, seen in both examples. **(d)** The superpositioning of one image atop another is a second compositional characteristic of trance imagery, as seen in these motifs and in **(c)** right. Note that, where visible, the bighorns are all shown with pronounced and raised tails, a tail posture that occurs with death.

juxtaposed against another, it may fragment into component parts, or it may rotate off a "normal" (horizontal) visual plane.

(3) The overall experience of hallucinations is widely acknowledged as ineffable – difficult to verbalize, or to portray graphically. The widespread cross-cultural solution to this problem is the use of verbal and graphic metaphors originating in the common bodily hallucinations and emotional reactions to trance (Whitley 1994b). One of these, used by drug-users in our own culture as well as by Native American, southern African and Siberian shamans, is based on a sense of weightlessness, changes in vision, and a dissociative mental state that can occur in trance. Metaphorically speaking, it is called a "trip", "out-of-body experience", or "mystical flight" (Turpin 1994).

Another neuropsychologically-based metaphor, commonly used in Native California, is "death" or "killing". This resulted from the obvious physical analog between death and trance (physical collapse, apparent loss of consciousness, convulsions, etc.) along with particular emotional responses to trance. Although anthropologists refer to the shaman's altered state as "ecstatic trance", laboratory experiments show that about 20% of hallucinating subjects report strong adverse emotional reactions, such as aggressiveness or extreme grief, including the feeling of their own death. Hence, an individual was commonly said "to die" while in a trance. A third metaphor for trance is bodily transformation; typically a change into an animal form. This originated in characteristic tactile and bodily hallucinations and, again, is verbally reported by both Native Americans and modern laboratory subjects while hallucinating (Whitley 1994b). One result is that shamans were widely thought able to transform into their animal spirit helpers.

The interplay of ethnographic and neuropsychological data in the understanding of Mojave petroglyphs is well-illustrated by the bighorn sheep motif, one of the hallmark petroglyphs of the region. Early archaeological interpretations of bighorn petroglyphs linked them to hunting magic – a practice that, notably, all Native Americans systematically denied, and that a variety of lines of empirical evidence cannot support (Steward 1963; Rector 1985; Whitley 1994a). Recent ethnographic analysis shows that the bighorn was not created to aid the hunt but instead was intended to portray a shaman's spirit helper; specifically, the spirit bighorn which was associated with rain-making power. Bighorn petroglyphs, in other words, are the spirit helpers of the rain shaman, while the primary concentration of these motifs, in the Coso Range in the northern Mojave Desert, was itself a nexus for acquiring rain-making power (Whitley 1994a, 1998a, 1998c).

The iconography of the bighorn petroglyphs, including those found in the Black Mountain region, confirms the origin of these motifs in trance imagery, and amplifies our understanding of petroglyph symbolism. Careful examination of the motifs shows that, first, many of them reflect second-stage trance construals of entoptic patterns; specifically, nested curves, one of the seven most commonly perceived entoptic patterns, construed as horns (Figure 1). Second, and in keeping with this first fact, many of these petroglyphs depict anatomically-impossible animals: bodies and snouts in profile, ears seen from the front , horns from a third perspective. Alternatively, sheep are also portrayed with two heads, with front-curving horns, or even with three horns, while some portray extra appendages, which is another common somatosensory hallucination during trance. Third, and further confirming the origin of these motifs in trance, many of the bighorns

Fig. 2. Aspects of the symbolic meaning of the petroglyphs as indicated by Native American commentary are also verified by the Mojave petroglyphs. (a) and (b) An equivalence between sheep petroglyphs, as depictions of shamans' spirit helpers, and the shamans themselves is demonstrated by the fact that both humans and sheep are shown in characteristic "patterned-bodied" forms (reflecting the second stage of trance during which entoptic and iconic images are integrated). Note also the multiple legs on **(a)** top, a characteristic of trance imagery resulting from a common bodily hallucination known as polymelia. **(c)** Some of the bighorns are plantigrade rather than cloven-hooved, suggesting the transformation of a flat-footed human human into a cloven-hooved sheep. Shamans were believed to transform into their spirit helpers. Note that **(c)** left has a projectile sticking out its back. **(d)** The priapic human figure on the left, shown shooting a bighorn, is also probably a conflation of a human and sheep, because Native Americans could not use bighorn racks for ritual headdresses or hunting disguises due to their great weight. Petroglyph (d) right is an obviously dead sheep, emphasizing the fact that essentially all bighorn petroglyphs were iconographically encoded to indicate that they were dead. Because the shaman was a bighorn spirit, and because his entry into the supernatural world of trance was metaphorically described as "dying", the "death of the bighorn" was the metaphor used to describe the shaman's trance experience.

portray the principles by which mental images are perceived during trance, including the juxtaposition of motifs, rotation off the horizontal, and the superpositioning of one atop another. As should be clear, these are not petroglyphs of "real" animals, but instead depictions of trance imagery.

A fourth characteristic of some of the bighorn petroglyphs, further emphasizing the non-realistic nature of the depictions, is the presence of internal body designs on some sheep (Figure 2). This parallels a feature common to certain human petroglyphs in the region, the so-called "patterned body anthropomorphs". Native Americans identify these as self-portraits of shamans, wearing ritual shirts which shamans painted with their "signs of power" (i.e, entoptic visions). These motifs represent the third-stage of trance, where iconic and entoptic patterns combine. That the bighorns are "patterned bodied", like the human figures, implies an equivalence between the shamans/humans and his sheep spirit helper.

The ethnographic record explains the logic of this equivalence: the shaman and his helper were thought indistinguishable, with no words available to distinguish their actions. A rain shaman, thus, *was* a bighorn, and shamans were widely thought able to transform into their spirit helper. This fact is shown in the iconography, where plantigrade rather than cloven-hooved sheep are shown or, alternatively, where the head of a sheep is placed on the upright body of a bipedal human; that is, conflations of human and bighorn features or, in terms of trance metaphors, transformations of the shamans into their bighorn spirit helpers.

The final set of important iconographic features seen in the bighorn petroglyphs concerns the death/killing metaphor for trance. Reference to this metaphor occurs in three ways. First, about 90% of the bighorns are shown with a pronounced and upraised tail. This a tail posture that, in the bighorn, occurs with death. The intent of the bighorn petroglyphs, in other words, was specifically to portray dead sheep. Second, this inference is confirmed by occasional sheep depictions shown more directly as killed: inverted, bloated and with a protruding projectile. Third, so-called "hunter and sheep" motifs show a human-bighorn conflation in the act of shooting a bighorn. We know that these are not true hunters and sheep partly, again, because the sheep are anatomically impossible; partly because petroglyph hunting magic was denied by all Native Americans; and partly because the use of bighorn headdresses was also categorically denied – they were too heavy to wear ceremonially, let alone while actively hunting.

Dead bighorns and "hunter and sheep" motifs reflect the ethnographic fact that the "killing of a bighorn" served as the primary metaphor for the rain shaman's supernatural efforts to manipulate the weather. A shaman transformed into the bighorn upon entering the supernatural world of trance, which was, metaphorically, a kind of death or killing. The rain shaman's vision quest was then an act of ritual auto-sacrifice intended to promote fertility (through rain-making) in the desert.

As is thus clear, Mojave bighorn petroglyphs incorporate a series of specific and coherent iconographic details that were probably obvious to Native Americans, but that have taken archaeologists a century to unravel. Likewise, this art encoded complex metaphors and symbols, and is not simply childish signs or icons, as many archaeologists have assumed. Like our own religious symbolism, the petroglyphs instead imply a Native American religious system that was richly detailed and cognitively complex. But this interpretive conclusion is precisely what the neuropsychological research tells us we should expect, given that we are dealing with the art of anatomically modern humans who were hardwired just as we are, and who maintained the same mental capacities that we enjoy today.

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Fort Irwin Archaeology: a Preserved Past for the Mojave Desert

Mark W. Allen, *Cultural Resource Specialist, National Training Center, Fort Irwin, AFZJ-PW-EV, Fort Irwin, CA 92310-5097*

INTRODUCTION

Fort Irwin has stewardship of nearly 1,000 square miles of the Mojave Desert. As the Army's National Training Center the Fort plays host to large-scale maneuvers of armor and mechanized units nearly each month of the year. It may be a surprise to some, but Fort Irwin nevertheless possesses some of the most significant extant archaeological and historical resources in the Mojave Desert. There are three primary reasons behind the preserved cultural resources of Ft. Irwin: the installation's isolation from populated areas, the protection provided by having a military or National Guard presence more or less continuously since 1940, and the Army's cultural resources program designed to identify, mitigate, and curate archaeological and historic resources. The degree of Ft. Irwin's isolation is readily apparent to anyone who makes the long drive out from Barstow. The deterrent to inadvertent damage, vandalism, and collecting provided by the presence of military units is also easy to grasp. But these factors have also been complemented by active management by the Army. This paper first briefly describes Fort Irwin's Cultural Resources Program. It then outlines the cultural history of the region and the cultural resources present at Ft. Irwin for each major period. This discussion illustrates the great significance of archaeology at Ft. Irwin.

Why is there a Cultural Resources Program at Ft. Irwin?

Aside from other motives, all Federal and Army installations are required by a series of legislation, regulations, and policies to inventory, protect, and mitigate cultural resources. The backbone piece of legislation is the National Historic Preservation Act of 1966. This act created the National Register of Historic Places and the Advisory Council on Historic Preservation. Section 106 of the act requires that proposed actions which may affect sites which are listed, nominated, or determined eligible for the National Register requires consultation with state Historic Preservation Offices, other interested parties, and where appropriate, the Advisory Council. Section 110 of the act requires Federal agencies to inventory, evaluate, and protect historical properties under their jurisdiction. A slough of other legislation has followed, the most prominent being the Native American Graves Protection and Repatriation Act (NAGPRA) which protects the ownership and control of Native American human remains and funerary objects. There are many other important pieces of legislation, but a discussion of these is beyond the scope of this article. They key point is that there are many mandates for cultural resources management on Federal installations.

The Army, too, has its own regulations. AR 200-4 will soon replace earlier versions of regulations which spell out the Army's policy for cultural resources management. The new regulation prescribes the management responsibilities and the required steps for identifying, evaluating, and nominating historic properties for listing on the Register of Historic Places. It also outlines procedures for curating archaeological collections, consultation with Indian tribes, the preparation of agreements, and the preparation of cultural resource management plans.

Responsibilities of the Ft. Irwin Cultural Resources Program

Although archaeological fieldwork at Ft. Irwin dates back to at least the 1920's, the majority of work has been conducted by the Army since the inception of the National Training Center in the early 1980's. Cultural resources management has continued to evolve and make significant contributions at Ft. Irwin. In recent years, the installation's Cultural Resources Program has greatly expanded.

The program's purpose is to support the training mission of the installation by managing cultural resources in compliance with Federal legislation and Army regulations. This responsibility can be broken down into several different efforts:

1) Survey most of the 1,000 square mile installations. The priority is to survey parts of the installation which are more likely to be used for training. Thus, high areas and steep slopes have a lower priority for coverage than playas, gentle slopes, and lower elevations. Survey usually consists of pedestrian coverage of the landscape, with field personnel walking transects and identifying possible archaeological and historical sites. To date, about 40% of the high priority area has been surveyed, mostly in the past 20 years. These recent surveys have been consistently good to excellent in their thoroughness and coverage. Over 400 archaeological and historic sites have been identified by surveys thus far.

2) Conduct assessments of the significance of identified cultural resources. Most importantly the program seeks to identify sites which are eligible to be nominated to the National Register of Historic Places. These are the sites of greatest concern for management and protection efforts. Test excavations are highly desirable in this process to assess the integrity of site deposits.

3) Protection of sites. The program has the responsibility of establishing and monitoring areas which are declared "off-limits" due to potential damage to significant cultural resources. Concertina wire, cyberstakes (designed to be visible with "nightvision" systems), ditches, and other barriers are frequently employed to protect natural and

Fig. 1. Fort Irwin's new archaeological curation facility. This 3,500 square feet building will house a laboratory and provide climate controlled storage for the installation's artifacts and records.

cultural resources. These areas are monitored for breaches and to maintain the barriers and signs.

4) Mitigate significant sites which cannot be avoided due to the training mission. This involves the completion of large-scale areal excavations to recover important data. These produce important information about site occupation, chronology, and structure. They generally generate large numbers of recovered artifacts and voluminous documentation.

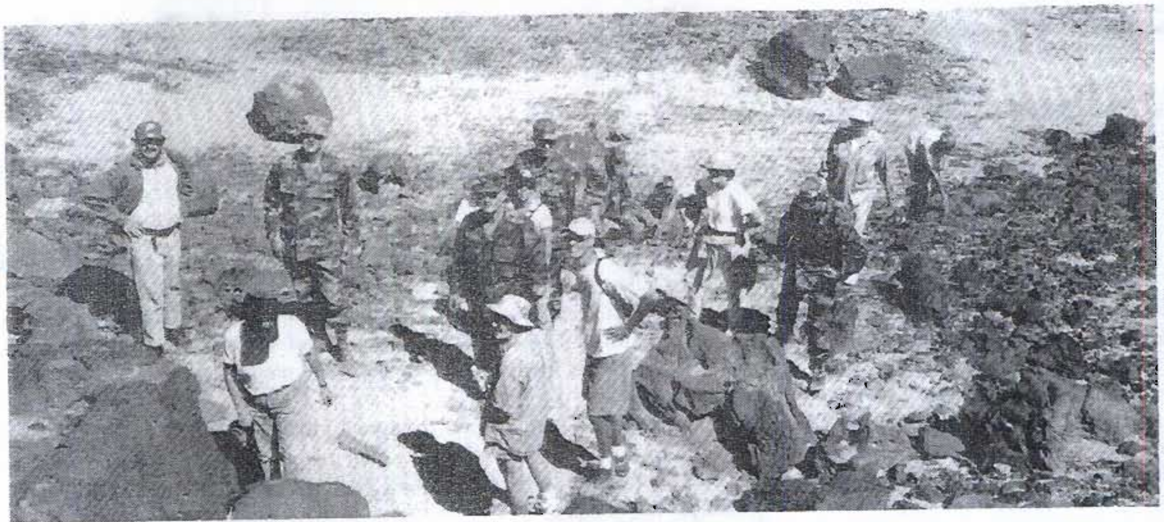

Fig. 2. A site tour conducted by the Ft. Irwin Cultural Resources Program in September, 1997. Members of the "Cobra Team" of Observer-Controllers and their families were shown petroglyph sites, historic sites, and archaeological sites on the installation.

5) Curate artifacts, records, maps, notes and other cultural resources collected and produced by archaeological and historical surveys, excavations, and research. The installation has just completed a 3,500 square foot curation facility with climate control and security systems (Figure 1). This building will permit the installation to store and protect its cultural resources for many years to come. In addition, the building will house a laboratory to permit "in-house" processing and analysis of artifacts.

6) Inform appropriate Native American groups, professional archaeologists, and the public about the cultural resources present on Ft. Irwin, and the Army's efforts to manage and protect them. The Cultural Resources program gives presentations both on and off the installation. The program also gives tours of archaeological and historical sites to interested parties (Figure 2).

A BRIEF CULTURE HISTORY OF THE FT. IRWIN REGION

The antiquity of human presence in North America is a subject of considerable interest to both the public and professional archaeologists. Perhaps the most frequently asked questions posed to archaeologists revolve around the age of artifacts, sites, and cultures. Anytime the issue at hand is the identity or age of prehistoric cultures there is bound to be some disagreement. Fortunately, a fairly widely used and generally accepted cultural history sequence is available for the greater Mojave Desert, and much of the southwestern Great Basin as well. This scheme was developed by excavations in Nevada and California, and has stood the test of time fairly well (Basgall 1995, Basgall and Hall 1994, Bettinger and Taylor 1974, Wallace 1962, Warren 1984, Warren and Crabtree 1986). As such, it permits a relatively uncontroversial reconstruction of cultural change in the Fort Irwin region (key sources of controversy are nomenclature and the exact transitions between periods rather than the overall sequence). This section presents a brief summary of the major periods of culture history in the Mojave (following Warren 1984), and an equally brief assessment of Fort Irwin's resources for each.

The Lake Mojave Period (12,000 to 7500 years ago)

Arguments have been put forth which seek to establish the presence of humans in the Mojave prior to 12,000 years ago. Despite considerable efforts to demonstrate such a presence, the majority of archaeologists feel that the available evidence suggests that humans first entered the area around 12,000 years ago. However, it is important to note that at least one site, Monte Verde in Chile, strongly suggests human migration to the New World by at least 13,000 years ago (Meltzer et. al. 1997). Thus, the possibility does exist that other sites in the Americas may predate 12,000 years ago.

The first cultural period is called Lake Mojave after a series of sites on the margins of Pleistocene lakes in the Baker region (Campbell et. al. 1937). Though frequently regarded as a period when people focused on hunting and gathering around lakes and marshes (Moratto 1984:93-97), at least some of these lakes were already drying out by the beginning of this period. Moreover, Lake Mojave sites and artifacts have also been found in a variety of locations away from the margins of wet areas (Basgall and Hall 1992). The most diagnostic artifacts of this period include "foliate" or leaf-shaped projectile points with long stems, some fluted points, and chipped stone crescents. Other tools have multiple purposes and were frequently reworked. Coupled with the widespread presence of non-local raw materials at sites, multi-functional tools attests to a high degree of mobility (Basgall 1995:22). The majority of points and bifaces (which also served as cores for producing smaller flake tools) were made of igneous rock (basalt, rhyolite, felsite), while formal scrapers are much more likely to be made of glassy (but brittle) silicates (chert, jasper, chalcedony). This pattern eventually changed through time. Finally, groundstone tools for plant processing are present during this period, though in low densities.

Lake Mojave sites are fairly well-represented on Ft. Irwin. In addition, work on these sites has provided important information on the dating of this period, the environmental locations of sites, the presence of non-local stone material, the presence of groundstone, and the variety of animals present in faunal assemblages (Basgall 1995: 17-22).

Pinto Period (7,500 to 4,000 years ago)

Pinto period sites are found throughout the southwestern Great Basin and the Mojave Desert. This name was assigned on the basis of a series of sites from the Pinto Basin in Joshua Tree National Monument (Campbell and Campbell 1935). Interpretations of this period have changed significantly over the past decade or so. Some fifteen years ago Warren (1984:411-413) stated that Pinto sites were "few in number" and limited to small surface scatters. Recent work (much of it at Ft. Irwin) have revealed some key exceptions: the presence of some large and deep sites with hundreds or thousands of artifacts each. These tend to occur near springs and other sources of water in a landscape that had become much drier than that of the Lake Mojave period.

Basgall (1995:24) has developed one explanation for this period:

> the huge Pinto sites probably represent places that were

> occupied repeatedly by small groups over long spans of time, and not major village locations used by large groups for one extended interval. The abundance of small sites implies that populations during the Pinto period were quite mobile, and moved more-or-less constantly across the landscape in response to fluctuations in water and/or food resources.

There are certainly other possible interpretations of these data. It may well be that the larger sites do in fact reflect fairly long-term occupations rather than temporary camps which were repeatedly used. This is certainly a solvable problem, but will require more excavations and analysis of existing collections and fieldnotes.

Pinto period sites are defined largely on the presence of projectile points with shouldered stems with indentations in the base. Other diagnostic artifacts include leaf-shaped bifaces and thick scrapers. The chipped stone tool assemblage, however, is not greatly different from that of Lake Mojave sites. In addition, the preference for basaltic stone for biface tools continues. Silicates, likewise, are still the scrapers of choice in the Pinto period. Recent research at Ft. Irwin and other areas has, however, revealed, an important difference: groundstone tools are more common at Pinto sites. Presumably, these tools indicate greater plant processing than the Lake Mojave period. This shift may well be part of the response to the drier climate which occurred during the early Holocene.

Fort Irwin has numerous significant Pinto period sites. Over a dozen large Pinto sites have been identified thus far. Some of these are unusually deep and well-preserved. The Awl Site (SBr -4562), for example, has been fairly well described by three separate archaeological projects (Basgall and Hall 1993, Byrd 1997, Hanna et. al. 1981, Jenkins and Warren 1986). According to Basgall and Hall (1993:99-100):

> there is little question that the Awl site served as a residential nexus of some importance, a role indicated by the mass and depth of cultural deposits at the locus, and by the abundant array of artifactual and organic debris discarded over the course of its occupations. Chronological resolution is inadequate . . . to identify fine-grained settlement intervals, but there is no reason to believe the location was occupied continually over long periods . . . instead, SBr-4562 was likely inhabited repeatedly, perhaps seasonally, during years or decades when local environmental conditions were sufficiently "good" to attract people . . . then abandoned for prolonged intervals when these conditions deteriorated.

The site also contains well-preserved faunal remains which may prove instrumental in helping to reconstruct the economic changes which occurred during the mid-Holocene. However, more work is needed to better determine the extent and duration of occupation of these large Pinto period sites at Ft. Irwin. An important step in this process will be large-scale excavations at the Awl Site which are scheduled to take place in the spring of 1998 (Byrd 1997). These will provide large amounts of data, diagnostic artifacts, and most importantly intact features (trash pits, fire hearths, structures, and other non-portable archaeological evidence).

In addition to large sites, there is a second type of important Pinto period resources at Ft. Irwin. Three are a number of discrete lithic reduction sites or "chipping stations" which date to this period. These sites contain in situ accumulations of flakes, partially completed tools, and sometimes hammerstones. Some of these are circles of debitage around a clear spot in the middle–presumably where the flintknapper sat or kneeled (Figure 3). Examples include a site in the Goldstone area and a newly discovered Pinto site called Cranch 2 in the vicinity of Garlic Springs. These features are the remains of distinct episodes of human behavior, the product of a few minutes or hours of work thousands of years ago. They present an outstanding opportunity to study the processes of took-making, the materials used, the variation in techniques of flintknapping, the extent of imported material, and the types of tools and flakes made together at the same moment. They are truly remarkable sites with tremendous research potential.

Fig. 3. A remarkable well-preserved lithic reduction station at Fort Irwin. The basaltic material and diagnostic artifacts at this site indicate that the station is from the Mid-Holocene (7,000 to 4,000 years ago). The bare spot in the middle of the circle of debitage is perhaps the space in which the flintknapper sat.

Gypsum Period (4,000 to 1,500 years ago)

The next cultural period in the region again occurs over a wide area in the Great Basin and Mojave Desert. Gypsum (named after a cave site in southern Nevada) sites reflect some noticeable differences from the first two periods. Perhaps most drastic is a great increase in the presence of long-distance trade items, including marine shell beads from the California coast. Second, large sites during this period are rare. Most are small scatters which reflect, "a broad range of tool production and subsistence activities" (Basgall 1995:27). Gypsum sites tend to cluster around and along major travel corridors.

Several technological changes are evident as well. The most diagnostic tools are three different projectile point types: Elko, Humboldt, and Gypsum Cave points. Debitage (waste flakes) are very common in these sites as finely-controlled pressure flaking became common. The stone of choice is now clearly siliaceous (chert, jasper, chalcedony), as basalt flakes and tools drop off drastically. Groundstone use is similar to that of the Pinto period: it is a small constituent of artifact collections, but is nevertheless found in the majority of sites, which would seem to be, "related more to opportunistic exploitation of plant resources wherever they might be encountered than to major seed harvesting episodes at predictable locations" (Basgall 1995:27).

Gypsum period sites are common and widespread across Ft. Irwin. These are small sites which indicate small groups of people in small sites for short periods of time. They also left behind few diagnostic tools, which make these sites somewhat difficult to identify without substantial investigations. This period remains poorly known owing to the short occupations of sites, the difficulty in identifying them, and the low number of large-scale excavations on Gypsum sites. Despite our incomplete knowledge, Warren (1984:420) states that Gypsum in the Mojave seems to be, "a time in which the human populations adapted to the arid desert . . . these adaptations included technological items, ritual activities, and increased socio-economic ties through trade." Without doubt, Ft. Irwin has the potential of significantly increasing our understanding of this period.

Saratoga Springs (1,500 to 700 years ago)

There were at least a few major changes in the next cultural period. First, during the Saratoga Springs (named for sites in Death Valley) period a fairly major technological innovation spread rapidly throughout the Great Basin and Mojave Desert. Indeed, throughout the continent around this time, people quickly adapted the use of the bow and arrow. This is visible archaeologically by the development of small triangular points such as Rose Springs and Eastgate points. However, this change does not seem to have drastically altered tool production or economic practices in the Mojave or southwestern Great Basin (Allen 1986, Warren 1984:420-421). People merely added the bow to their inventory of tools and weapons, and no doubt were at least somewhat more successful in hunting and warfare efforts.

A second major change was the migration or at least influence of Puebloan cultures from the Southwestern United States. Agriculturalists with ceramics apparently exploited turquoise and perhaps other items in the Mojave Desert. It is not yet clear to what extent they actually moved into the area, but their ceramics certainly did. Anasazi and Hakatayan (also called Patayam) ceramics are fairly common in the Mojave Desert. In addition, shell beads from the California coast are widespread. Thus, "trade with the southern California coast may well have been the impetus for the extension of Hakataya influence across the Mojave Desert" (Warren 1984:423).

Stone tools in the Saratoga Springs period are still usually made with chert, jasper, and chalcedony. Apparently, there is also an increased use of local materials, and an increase in the proportion of flake tools, suggesting that, "most artifacts were made on an as-needed basis, and . . . that people worried little about gearing-up for future needs" (Basgall 1995:28). Groundstone is also still present in about the same frequencies as Gypsum sites. Most Saratoga Springs sites are small and indicate short-term use by small groups of people.

Ft. Irwin has a large number of sites which date to the late prehistoric period based on diagnostic projectile points, the prevalence of flake tools, and the presence of ceramics. They occur in a wide variety of environmental settings from springs areas to uplands. Several varieties of ceramics are represented, making the region an ideal area for studying the interaction of various Puebloan cultures with indigenous desert populations. It is also likely that a detailed study of their distribution could shed much light on economic practices and settlement patterns during this period.

Protohistoric/Shoshonean Period (700 to 100 years ago)

This is the period of escalated inter-regional contact and early interaction with European and Euro-American populations. The Mojave Desert contained major travel routes that by at least the last few centuries were main transportation routes. Likely, Europeans following the Santa Fe and Mormon trails were following trails that had been in use for millennia. Brown and buff ceramic wares are found in sites of this period, indicating continued influence from the Southwest, specifically the Lower Colorado River region (May 1976, Warren 1984). Another time-marker is the Desert Side-notched projectile point. Shell beads are very common.

The Mojave River valley seems to have been an important trade route between the coast and the southwest by this time. Apparently, this period saw the entrance of new groups of people into the region. As summarized by Warren (1984:426):

> although the cultural assemblages of the Mojave River appear to have ties with the Yuman peoples of the Colorado River, there seems to be little doubt that these late sites along the Mojave River are the prehistoric remains of the Uto-Aztecan Serrano of the historic period. The reason that the Serrano appear to be more similar to the Yuman groups than their Uto-Aztecan neighbors to the north may be attributed to the Mojave River trade route that, for centuries, brought the Serrano in contact with the cultural developments of the lower Colorado River.
>
> The trade along the Mojave River must have provided the middlemen of the region with opportunities to obtain relatively great amounts of wealth and to develop more complex socioeconomic and sociopolitical organizations.

At contact, these peoples were speaking a variety of Numic (a sub-family of Uto-Aztecan) languages called Shoshonean. However, it is important to note that there were other languages and groups of people in the Mojave region as well (Basgall and Hall 1994). Evidently, there was a long tradition of culture contact throughout the desert. By the mid-1800's, however, the predominant Native American groups along the Mojave River were Paiute.

Basgall and Hall (1994) make two relevant points about the Ft. Irwin area during the Protohistoric or Shoshonean period. First, it is very difficult to assign cultural, linguistic, or ethnic affiliations to the region. They argue that, "ethnolinguistic population centers lay outside the area," and that the region was "exploited sporadically (seasonally, or during especially good years) for select reasons by different populations, none of which probably maintained "permanent" residency. However, even highly mobile hunter-gatherers are known to defend territories (Keeley 1996). If multiple groups of people were trying to utilize the region simultaneously (it is highly likely that the region would be desirable for all groups at the same time) there may well have been some sort of competition given the paucity of dependable water and other scarce resources. This leaves open the possibility that the area was controlled or even occupied for long periods of time by particular groups (which might change through time). More research is certainly required.

Second, they note that "evidence for aboriginal occupation during the protohistoric and historic periods remains scanty, " though "several sites have produced materials suggestive of Native-White interaction" (Basgall and Hall 1994:67). These sites and others may be able to shed light on the nature of culture contact with the Spanish and Euro-Americans, and may reveal evidence which would permit a better understanding of the Protohistoric Native American use and occupation of the region.

Historic Period

Europeans and Euro-Americans have traveled through the Mojave since the late 1700s. Intensive settlement and use, however, did not begin until the 1830's (Basgall and Hall 1994, Jackson et. al. 1987, Quinn 1981, Vredenburgh 1994). Important activities and interests in the nineteenth and early twentieth centuries included trade, mining, ranching, transportation, and tourism. Military activity came to the region with James Carleton's campaign against the Paiute of the Mojave River in 1860 (Casebier 1972).

To date, there are 50 recorded historic sites on Ft. Irwin. Since the majority of these sites are not subjected to severe impacts from training activities, these sites have usually only been recorded and briefly described. One exception is a redoubt built during Careleton's campaign at Bitter Spring. This redoubt was one of three bases of operations during the spring and summer of 1860. It was partially excavated in 1984 (Basgall 1995:31, Hardesty 1988). But the majority of historic sites at Ft. Irwin "relate to either mining activities (beginning after 1860, flourishing between 1896-1910, and reduced thereafter) or military use (1860-1871 and 1940-present)" (Basgall and Hall 1994:67). One rather exceptional historic resource on Ft. Irwin is the Pioneer Deep Space Tracking Station in the Goldstone area. This was the first such station, built in 1958 and operational until 1981. It is a National Historic Landmark and thus on the National Register of Historic Places (Allen 1998).

CONCLUSION

As this brief overview makes clear, the cultural resources of Ft. Irwin are both well-preserved and highly significant for documenting and interpreting human adaptations to a harsh environment. As part of an on-going attempt to collect key data for a better understanding of human history in the Mojave Desert, Ft. Irwin has conducted a great deal of survey and mitigation to record and recover information from prehistoric and historic sites within the installation's boundaries. The Cultural Resource Program has the responsibility of continuing this work as well as the curation of existing resources for future study and appreciation.

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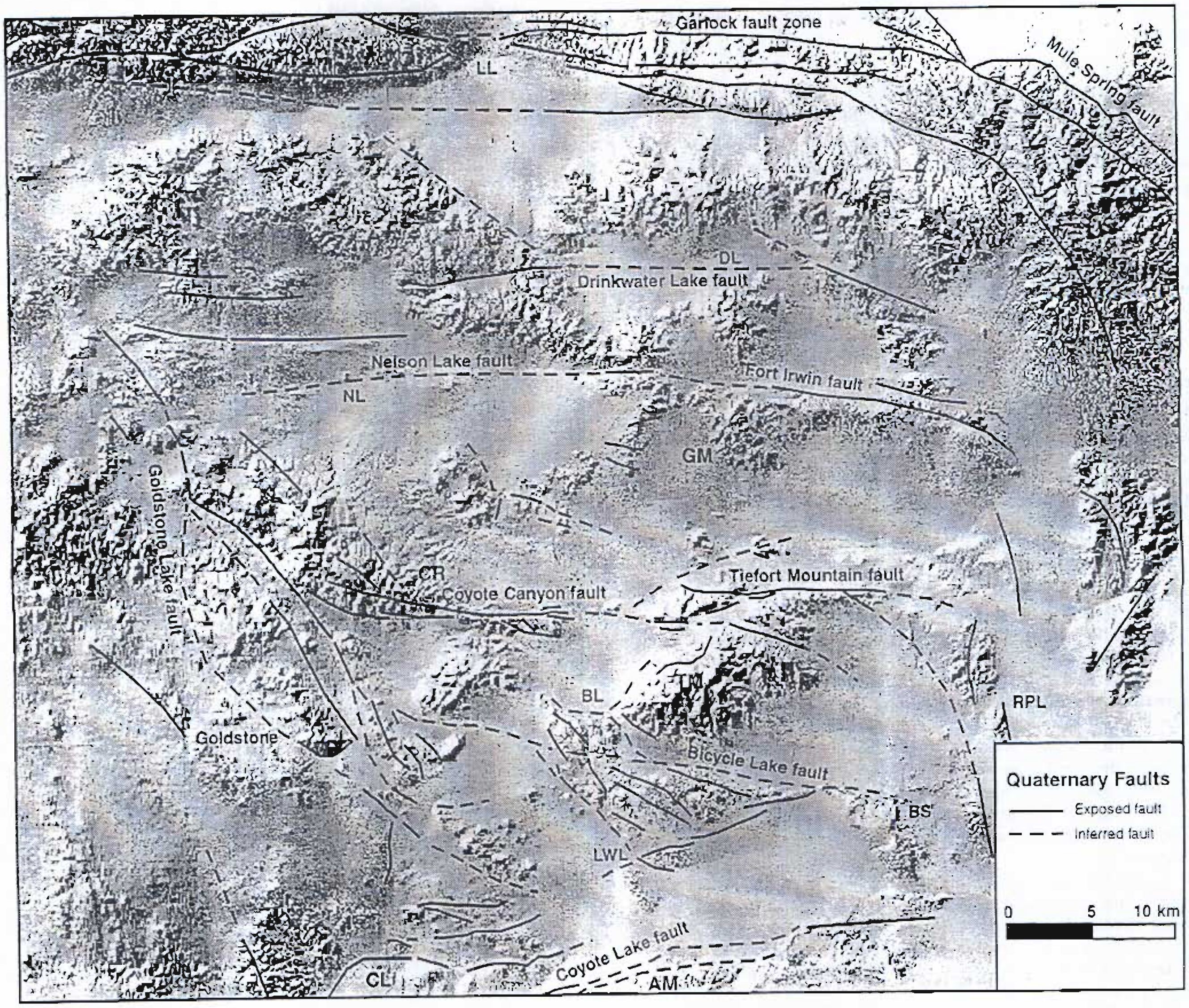
Quaternary Faults and Shaded-relief Topography of the Fort Irwin area, California



Shaded relief illustration of the influence of Quaternary faults on topography. Most east-striking faults exhibit a component of south-side-up vertical offset, seen on all scales from small faults near Coyote Lake to the larger faults such as Bicycle Lake and Fort Irwin faults. The valleys and fault-bounded mountains tend to climb in altitude from Coyote Lake toward the Avawatz Mountains. The Tiefort Mountains, a high mountain mass near Bicycle Lake, are an exception to this general rule; they are bounded on the northwest by a young thrust fault.

David M. Miller and **James C. Yount**

Abbreviations:

LL, Leach Lake; **NL**, Nelson Lake; **RPL**, Red Pass Lake; **BL**, Bicycle Lake; **LWL**, Langford Well Lake; **CL**, Coyote Lake; **GM**, Granite Mountains; **BS**, Blind Spring; **CR**, Coyote Ridge; **AM**, Alvord Mountains.

Late Cenozoic Deformation in Eastern Fort Irwin and its Significance for the Tectonic History of the Garlock Fault System

Terry L. Pavlis[1], **Laura Serpa**[1], **Bennie Troxel**[2], **Michael Dean**[1], **Theresa Hartman**[1], and **Traci Rodosta**,[1] [1]*Dept. of Geology and Geophysics, University of New Orleans, New Orleans, LA 70148*, and [2]*2961 Redwood Rd., Napa, CA 94558*

INTRODUCTION

The deformational history in the vicinity of the Garlock-southern Death Valley fault intersection has long been recognized as a major regional tectonic problem in the Death Valley region. The problem was clearly stated by Davis and Burchfiel (1973) in their classic paper on the Garlock fault when they noted that the piercing lines that define the slip on the Garlock fault all now lie within a few kilometers of the eastern termination of the Garlock. This suggests the Garlock once had an eastward extension that was subsequently cut and offset by younger right-slip on the Death Valley system, or there must be extreme extension to the north of the intersection, contraction to the south, or both, to accommodate the apparent strain incompatibility. The area northwest of the intersection shows evidence of extreme extension but this extension is centered to the north of the Owlshead Mountains, in the central Death Valley region. Similarly, there is no evidence of extreme shortening to the south of the intersection. For this reason, most investigators have assumed an eastward extension exists and several possible sites have been proposed (e.g. Plescia and Henyey, 1982; Davis and Burchfield, 1993) for this extension.

A major handicap to early studies of the Garlock-southern Death Valley intersection was the lack of detailed geologic information within much of the region surrounding the intersection. A particularly prominent gap in the geologic database was the large area southwest of the intersection that lay within the Ft. Irwin military reservation. In the early 90's, however, the DOD lifted many of its security restrictions in Ft. Irwin, and the area became the object of extensive USGS studies (e.g. Schermer et al., 1996) as well as work by our group and researchers from Louisiana State University. In this paper we summarize some of the first order results of our work in eastern Ft. Irwin as it relates to the Garlock system. We begin with a description of the geology of eastern Ft. Irwin. In this description we emphasize the evidence for significant crustal contraction in association with a set of sinistral faults that we interpret as a transrotational system, a conclusion in line with numerous previous interpretations that date back to Garfunkel's (1974) model for the Mojave Desert region. We also note the important role played by the Arrastre Springs fault system which was clearly a left lateral fault during its latest history and we tentatively suggest that it was an extension of the Garlock system; a conclusion that partially explains the conundrum of the eastern termination of the Garlock. We conclude with a brief overview of our interpretation of tectonic history of the system and our inference of the role of transrotation in the late history of the system.

GEOLOGY OF EASTERN FORT IRWIN

The dextral southern Death Valley and sinistral Garlock fault systems meet in the eastern Avawatz Mountains (Figure 1). Both of these fault systems show clear evidence of Quaternary activity, and simultaneous slip on both fault systems produces an extreme strain incompatibility that must be accommodated by subsidiary faulting in the vicinity of the intersection. The best known of these secondary effects is the clear evidence for east-vergent thrusting along the eastern edge of the Avawatz Mountains (e.g. Brady, 1984a, 1984b). This thrusting cannot be the only effect of the fault interaction, however, because the total left slip on the Garlock fault system (55-65 km) is far too large for the apparent shortening recognized by Brady (1984a) along the transpressional fault-fold systems of the Noble Hills-eastern Avawatz Mountains region. Instead, it seems clear that the incompatibility is accomodated over a much larger region as transtensional structures to the north and distributed deformation to the southwest. It is that deformation to the southwest that is the primary object of this paper.

ROCK ASSEMBLAGES OF EASTERN FORT IRWIN

Late Cenozoic deformation in eastern Ft. Irwin has exhumed three major rock assemblages that form the topographic highlands of the region: 1) Pre-Tertiary crystalline rocks; 2) Oligocene to late Miocene terrestrial sedimentary deposits; and 3) Plio-Pleistocene deposits. Major unconformities bound each of these assemblages and the geometry of these unconformities provides critical information on the late Cenozoic deformational history.

Pre-Tertiary rocks

The pre-Tertiary rocks of eastern Ft. Irwin are dominated by Mesozoic rocks that include voluminous intrusive complexes and variably deformed and metamorphosed Mesozoic volcanic rocks (Busby-Spera et al., 1990; Walker et al., 1990). Metamorphic assemblages with pre-Mesozoic protoliths (Figure 1) are limited to roof pendants in the Avawatz Mountains and a metamorphic complex in the Tiefort Mountains (Stephens, 1994).

Mesozoic assemblages of eastern Ft. Irwin can be divided into three major groups: early Mesozoic metavolcanic rocks and associated metasedimentary rocks; Triassic-Jurassic plutonic assemblages; and Cretaceous(?) granites. Geochronologic data from this area is sparse and our tentative assignment of ages for these Mesozoic assemblages is based on correlation to dated units in the Avawatz (Spencer, 1990a, 1990b) and Tiefort Mountains (Stephens, 1994) as well as cross-cutting relationships. A full description of these rocks is beyond the scope of this paper, but a number of first-order features are critical for this paper.

First, one of the most prominent Mesozoic rocks is a series of coarse grained, megacrystic granites that are clearly the youngest plutonic bodies in east-central Ft. Irwin. Based on regional correlations (e.g. Walker et al. 1990; Spencer, 1990b) these bodies are almost certainly Cretaceous in age because: 1) These type of granitoids are typical of the Cretaceous in the eastern Mojave desert; 2) the plutonic rocks are not cut by the Late Jurassic Independence dike swarm and therefore are presumably post-late Jurassic; and 3) they are unfoliated and cross-cut strongly deformed plutonic rocks, indicating a younger relative age. These plutons are important to the Cenozoic geology because they define distinctive piercing lines to constrain slip on structures in Ft. Irwin.

Second, aside from plutonic rocks, the most abundant Mesozoic rocks in eastern Ft. Irwin are variably deformed volcanic assemblages. There are no clear age indicators for these rocks in Ft. Irwin, but rocks of this type are late Triassic to middle Jurassic in age in the eastern Mojave Desert (e.g. Schermer et al., 1996; Busby-Spera et al., 1990). In

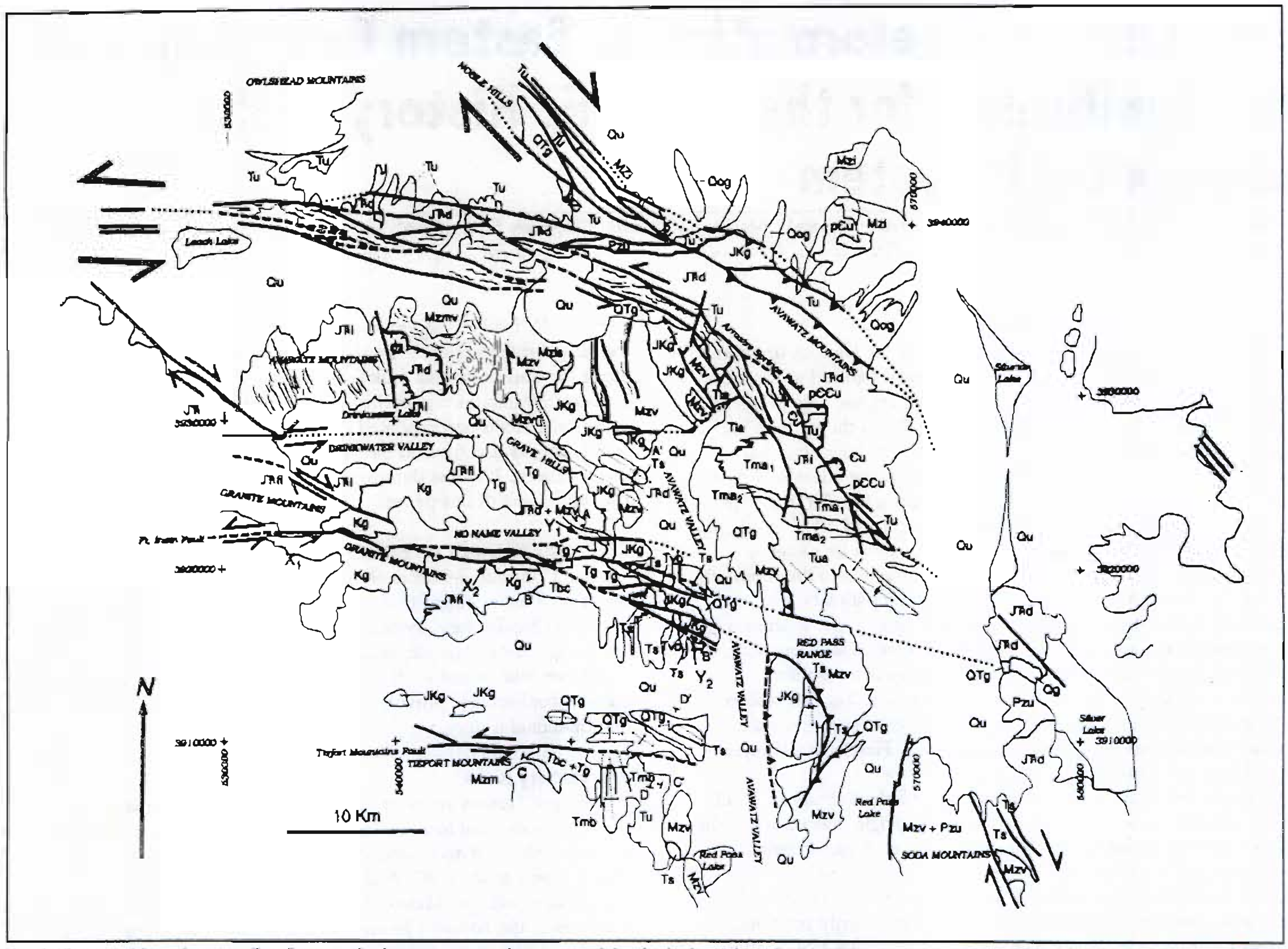


Fig. 1. Tectonic Map of eastern Fort Irwin and adjacent areas in the vicinity of the Garlock-southern Death Valley fault intersection. Map was prepared from 1:100,000 scale Thematic mapper imagery (courtesy of Roy Dokka) as a base map for compilation of our geologic mapping together with mapping by Spencer (1990a) and Brady (1982) from the Avawatz Mountains. UTM grid coordinates are given for geographic reference. Symbols are standard geologic map symbols. Grey lines in western Avawatz Mountains are the Independence dike swarm. Abbreviations (this figure and figure 2): Pzu-Paleozoic undivided; Cu-Cambrian undivided; pCCu-Precambrian undivided; Mzm-Mesozoic plutonic/metamorphic complex undivided; Mzi-Mesozoic intrusive; Mzv-Mesozoic volcanic rocks undivided; Mzmv-Mesozoic mylonitic volcanic rocks; Mzls-Mesozoic limstone; JRi-Jurassic-Triassic intrusive; JRfi–Jurassic-Triassic foliated intrusives; JKg-Jura-Cretaceous granitoids; Kg-Cretaceous megacrystic granite; Tla-Middle Miocene lower Avawatz Formation; Tma_1-fanglomerates of middle Avawatz Formation; Tma_2–upper part of middle Avawatz Formation; Tua-upper Avawatz Formation (presumed equivalent to Tg and Ts of the eastern Fort Irwin basin); Tvb-Tertiary volcanic basalt; Tg-Tertiary fanglomerate and fluvial conglomerate facies of the eastern Fort Irwin basin; Ts-Tertiary sandy and silty facies of eastern Fort Irwin basin; Tmb–megabreccia; Tbc-boulder conglomerate facies of the eastern Fort Irwin basin; QTg-Plio-Pleistocene fanglomerate; Qu-Quaternary undivided.

eastern Ft. Irwin the distribution of these rocks has important implications for the paleogeography of the region because they appear to be restricted to the Avawatz Mountains north and east of the Drinkwater valley. Amphibolites and quartzofeldspathic schists of the Granite Mountains could be derived from these rocks, but the distinction in metamorphic grade implies major distinctions in structural level, paleogeography, or both, across the Drinkwater valley.

An important feature of the metavolcanic rocks of the western Avawatz Mountains is the variable development of a greenschist facies, mylonitic fabric (e.g. Dean, 1995). These mylonitic rocks are important for studies of the Garlock fault because Davis and Burchfiel (1973) correlated these rocks with mylonitic metavolcanics in the Layton Pass thrust in the Slate Range to help define the net slip on the Garlock fault. Our work supports this interpretation because the mylonitic rocks in this sequence clearly show top to the east shear senses consistent with the thrusting recognized along the Layton Pass system (Dean, 1995). The structure along this system is considerably more complex, however, with clear evidence of an interaction between left-slip ductile shear zones and the ductile thrust system as well as younger rejuvenation by extensional structures (see below). Nonetheless, this system remains a critical piercing line for reconstructing the Mesozoic system.

Third, the older crystalline assemblages in eastern Ft. Irwin are variably deformed plutonic rocks and associated metamorphic screens. These rocks are most prominent in the Granite Mountains where they are comprised of a plutonic-metamorphic complex with amphibolite and quartzo-feldspathic schists occurring as screens in highly deformed quartz dioritic to granodioritic plutonic rocks. In the Tiefort Mountains this plutonic/metamorphic complex also contains Paleozoic and older(?) metasedimentary protoliths. In the Avawatz Mountains, however, these older intrusives are not generally foliated and include large areas underlain by mafic and intermediate plutonic rocks

(Avawatz diorite-monzodiorite of Spencer, 1990b) as well as granodioritic plutonic rocks that are cut by the Independence dike swarm in the western prong of the Avawatz Mountains (Figure 1).

Tertiary Rocks

Middle to Late Cenozoic sedimentary rocks nonconformably overlie crystalline assemblages at a number of localities in eastern Ft. Irwin and provide the most clear information on the late Cenozoic history of this area. These rocks include the Avawatz Formation exposed throughout the southern Avawatz Mountains (e.g. Spencer, 1990b) and unnamed rocks exposed in the eastern Granite Mountains and at scattered localities in the Avawatz Valley east of Tiefort Mountain (Figure 1). Spencer (1990b) described the Avawatz Formation in detail and Sobieraj (1994) has described the unnamed assemblages in moderate detail. Thus, the reader is referred to those papers for details on these rocks. A number of observations, however, are critical to this paper.

First, Spencer (1990b) recognized that the Avawatz Formation could be divided into two basic assemblages: 1) a lower, Middle Miocene sequence which shows sedimentary facies and thickness variations that are strongly suggestive of deposition during a period of down to the west motion on the Arrastre Springs fault (Figure 1); and 2) an upper, Late Miocene sequence which overlaps the western branch of the Arrastre Springs fault and is characterized by large megabreccia sheets interbedded with the sedimentary section. Spencer (1990b) showed that the upper Avawatz Formation was moderately folded and contained growth-strata and internal unconformities indicating that the unit was folded during its deposition.

Second, our work together with the studies by Sobieraj (1994) suggests that the upper Avawatz Formation of Spencer (1990b) was deposited in a distinct sedimentary basin that was developed atop the older Avawatz basin in the Avawatz Mountains. This younger basin was developed over most of what is now eastern Ft. Irwin from the eastern Granite Mountains and Tiefort Mountains to the Avawatz Mountains. This "Ft. Irwin basin" also clearly extended across what is now the Red Pass Range (Figure 1) and probably includes rocks now exposed in the eastern Soda Mountains that Grose (1959) described in the context of the Soda-Avawatz fault zone (Rodosta, 1997 and Hartman, 1997). Evidence for this conclusion includes: 1) dating by Sobieraj (1994) that confirms a Late Miocene age for the Tertiary rocks of eastern Ft. Irwin; 2) widespread exposure of the Tertiary unconformity beneath the section defines a basinal area throughout eastern Fort Irwin–the only areas that were not clearly buried by the Tertiary section are the western prong of the Avawatz Mountains and the eastern half of the Red Pass Range (Figure 1); and 3) our studies confirm Sobriej's (1995) inference that this basin had a major source to the north with alluvial fan gravels to the north interfingering with distal fan and lacustrine deposits to the south.

Third, although sedimentary facies indicate that the eastern Ft. Irwin basin had a northern source area and a depocenter in what is now the southern Avawatz Valley, the basin forming mechanism is not clear. Dean (1995) recognized a pair of detachment faults and a dextral high-angle fault in the western prong of the Avawatz Mountains (Figure 1) that apparently represent a NS extensional system in present coordinates. Based on these structures he suggested that the eastern Ft. Irwin basin developed as a pull-apart with a dextral fault to the west and a south-directed detachment fault to the north. Although this interpretation of the basin is consistent with its geometry and the top to the south transport along the detachment faults (Dean, 1995), sedimentary thickness based on our subsequent studies of the basin when combined with gravity data suggest this model is not likely and that the basin is either a post-extensional overlap assemblage or the product of EW extension (present coordinates) along an E-dipping low-angle normal fault (e.g. as inferred in Figure 2). The distinction between these alternatives, however, is not possible from surface observations because the upper part of the sedimentary section onlaps the eastern Granite and Tiefort Mountains, obscuring any older structural contacts that might be present (Figure 1). Thus, unless this basin is imaged with seismic reflection data or the section is drilled, this problem will remain unresolved.

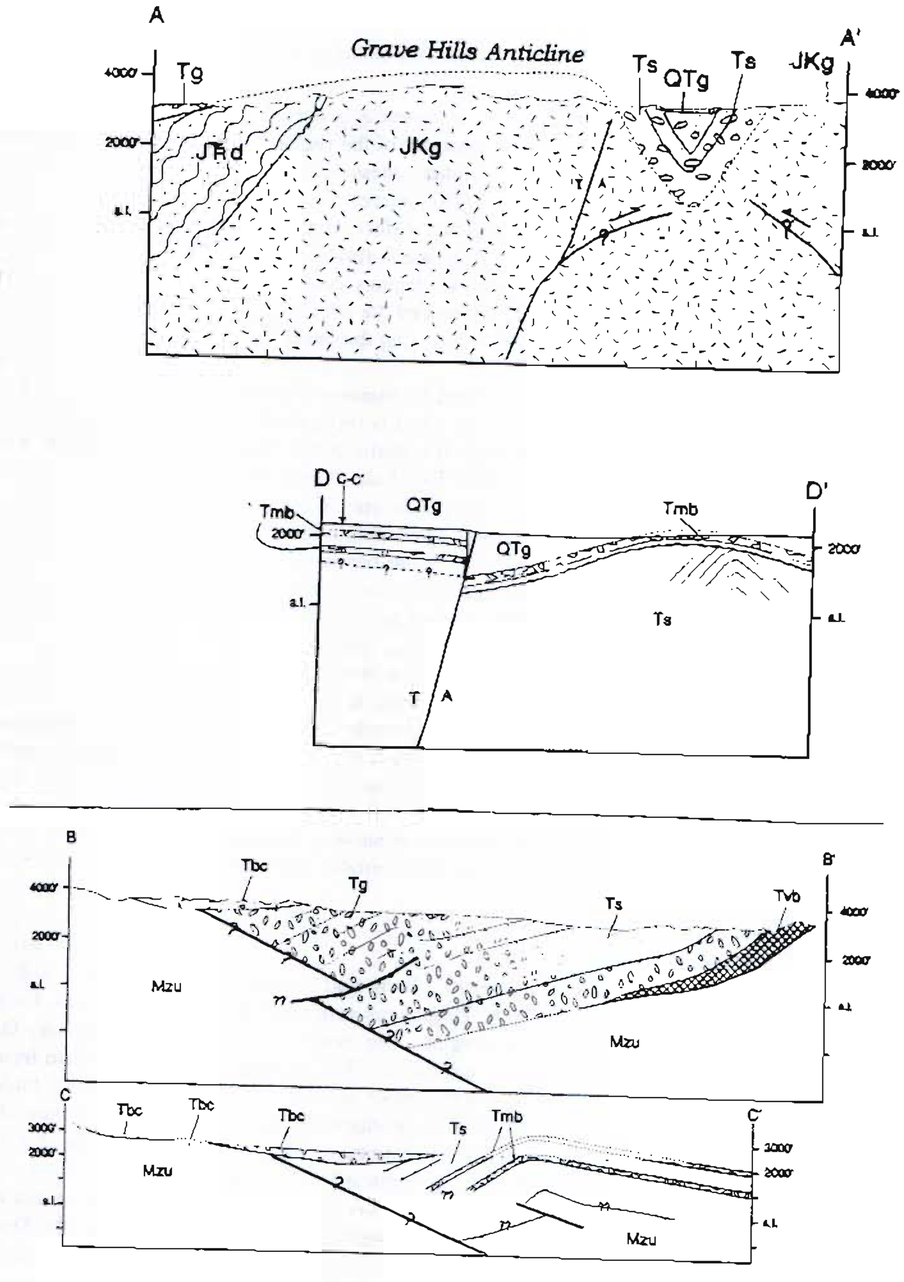


Fig. 2. Cross sections of representative structures in eastern Fort Irwin. See figure 1 for section locations and abbreviations.

Pliocene-Quaternary (?) Rocks

The Tertiary and older rocks are unconformably overlain by a younger section of alluvial fan gravels at several localities in eastern Fort Irwin and this angular unconformity/nonconformity provides important constraints on the latest Cenozoic history of the region. Along the western flank of the Avawatz Mountains these younger gravels form a gently southwest dipping (~15-20 degrees) homocline that is more steeply dipping than the present fan surface, but more gently dipping than the older strata of the Avawatz Formation below the angular unconformity where units dip 30-40 degrees southwest (Spencer, 1990a). Thus, these gravels clearly record the progressive southwest tilt of the Avawatz Mountains block during the latest Cenozoic.

This unconformable relationship recognized by Spencer (1990b, 1990a) is not limited to the Avawatz Mountains and is recognized throughout the eastern Fort Irwin basin. In fold systems of the Grave Hills and in the Avawatz Valley north of Red Pass Lake (Figure 1) folded Tertiary strata are more steeply dipping than unconformably overlying Plio-Quaternary gravels, but these gravels are, in turn, warped into broader folds of the same type as those below the unconformity (e.g. Figures 2); an observation indicating a growth strata relationship. In many areas these warped younger gravels also show geomorphic expressions consistent with the fold–e.g. anticlinal hills–indicating that the folding is very young and is probably ongoing. Younger gravels also cover most of the Tertiary strata in the Red Pass Range (Figure 1), but the entrenchment of these gravels is probably due to stream entrenchment along the Red Pass wash which is a major drainage that is actively downcutting due to lower base level of southern Death Valley. It is not known if this regional unconformity is a distinct time line related to a regional climatic variation or is a series of local features of different age with a coincidental similar appearance.

LATE CENOZOIC STRUCTURE

Sinistral Faults of eastern Ft Irwin

The present day physiography of eastern Ft. Irwin and the Silurian Valley is a manifestation of a series of Plio-Pleistocene structures that dominate the late Cenozoic history of the region. The most conspicuous of these young structures are a series of ~EW striking sinistral strike-slip fault systems. The most well known of these structures are the strands of Garlock fault which produces an elongate, EW ridge along the NW prong of the Avawatz Mountains. Three other sinistral faults in the area produce similar topographic effects including (Figure 1): 1) the Drinkwater Valley fault; 2) the Ft. Irwin fault; and 3) the Tiefort Mountains fault. Of these four sinistral fault systems all but one, the Drinkwater system, show clear evidence of Quaternary motion. Evidence for Quaternary motion include numerous scarps in Quaternary alluvium, disruption of Quaternary surfaces, and offset drainages. Persistence of at least some component of south side up slip is clear for both the Ft Irwin and Tiefort Mountains fault systems as indicated by the linearity of the mountain fronts formed by the faults. Nonetheless, the primary motion on these faults is clearly sinistral based on offset of markers.

Estimates of the total left slip on these structures are handicapped by the paucity of clear markers cut by some of the faults. Nonetheless, net slip on many of these structures is relatively well constrained.

For the Tiefort Mountains fault system (Figure 1) the total slip for the two branches of the fault is apparently >3.4 km (Schermer et al., 1996). Based on surface distribution of Tertiary sediments, however, we conclude that the maximum slip on the north Tiefort fault is <6.5 km and the actual slip is probably ~4 km. This estimate is poorly constrained (+/- 2.5 km), but if this slip is summed with the >2.7km of slip on the south Tiefort fault estimated by Schermer et al. (1996), the net slip on this fault system is probably approximately 7 km.

For the Ft. Irwin fault, Schermer et al. (1996) estimated > 3.7 km of sinistral slip, but our studies indicate the net slip is much greater. Schermer et al's (1996) estimate was based on the offsets of a basalt body that lay at the base of the Tertiary section in the eastern Granite Mountains (Tvb, Figure 1). Examination of this area in greater detail reveals, however, that this basalt is not an intrusive, but is a pair of flows that were deposited directly upon a paleosol developed atop the Mesozoic basement. These basalts are cut out to the north by either erosion or nondeposition, but the nonconformity and the paleosol are clearly recognizable across all but one of the six strands of the Ft. Irwin fault, and the unconformity position can be clearly projected across the sixth strand. Summing these offsets of the unconformity yields a minimum slip across the Ft. Irwin fault of 6.6 km (Y_1 toY_2 in Fig. 1). Note, this offset is a minimum because these splays of the Ft. Irwin fault all appear to have a small component of S-side-up dip-slip, which would reduce the apparent left-lateral shift along individual faults. Probably the best indicator of the net slip on the Ft. Irwin fault is a 9.8 km sinistral offset of a Mesozoic piercing line (labeled X_1 and X_2 in Figure 1) defined by the contact between strongly foliated granodioritic to quartz dioritic plutonic rocks and unfoliated megacrystic granites (Kg, Figure 1). This interpretation of the offset across the fault solves a dilemma posed by more modest estimates of the net slip. That is, the megacrystic granite south of the Ft. Irwin fault at Granite Pass is indistinguishable from the granite immediately north of the fault in the area just south of Drinkwater Lake (Figure 1), yet if these were the same pluton, the contacts would be offset in a dextral sense, a relationship at odds with the unequivocal sinistral offset of the Tertiary unconformity along the eastern end of the Ft. Irwin fault. Thus, we conclude that the lithologic similarity of the granites on both sides of the Ft. Irwin fault is coincidental and they represent different plutonic bodies that have been juxaposed by ~ 10 km of sinistral slip on the Ft. Irwin fault.

The Drinkwater Valley poses a major, unresolved problem in eastern Ft. Irwin. The geometry of the valley is similar to the valleys formed along the Ft. Irwin and Tiefort Mountains faults which has generally led to the conclusion that this valley is formed by another sinistral fault (e.g. Garfunkel, 1974; Schermer et al., 1996; Dokka et al., in press). Unlike the other valleys, however, there is no evidence of Quaternary faulting within the Drinkwater Valley and the irregular mountain front on both sides of the valley is inconsistent with a young strike-slip fault bounding the valley. A distinctive pediment surface in the Granite Mountains to the southwest of Drinkwater Lake projects ~150m above the valley floor indicating some component of south-side up faulting along the southern edge of the valley. Nonetheless, this offset cannot explain the major discrepancy in the pre-Tertiary geology across the Drinkwater Valley. That is, the foliated plutonic rocks cut by a megacrystic granite pluton on the south side of the Drinkwater valley bears no resemblance to the northern side of the valley which contains some very distinctive rock units including the Independence dike swarm cutting the older granitoids of the western Avawatz Mountains and mylonitic metavolcanic rocks. Moreover, two major structures on the north side of the valley–a NS striking dextral fault and two low-angle normal faults–do not appear on the south side of the valley (Dean, 1995). We have no explanation for this discrepancy, but suspect the presence of a major strike-slip fault system that predates the sedimentary rocks of the eastern Ft. Irwin basin. That is, although most maps (e.g. Trona sheet as well as Schermer et al., 1996) project the Drinkwater Valley fault into the east-side of the grave hills (Figure 1), our mapping does not support this interpretation. Instead, we recognize little, if any, disturbance of the Tertiary unconformity by faulting and suggest the dominant structure of the northern Avawatz Valley is due to folding (see below). Thus, if a major fault follows the Drinkwater valley and extends into Avawatz Valley, it would have to be older than the Late Miocene sedimentary section now exposed at the northern end of the basin.

Fold and Thrust Systems of Eastern Ft. Irwin

The EW trending ranges bounded by young sinistral faults of east-central Ft. Irwin contrast dramatically with the eastern edge of Ft. Irwin which is marked by a major NS trending valley, referred to here as the Avawatz Valley. This topographic trough is located between the eastern tips of the EW trending ranges produced by the sinistral faults and the NS trending highland extending from the eastern Avawatz Mountains to the Red Pass Range and western Soda Mountains. This topographic trough is also a structural low, but the origins of the depression are different along the valley. North of the Fort Irwin fault, the valley appears to be a young syncline with west-tilted Plio-Pleistocene gravels as well as Quaternary surfaces to the east of the valley and east-tilted gravels to the west. The west tilt of the eastern Avawatz Mountains is almost certainly the product of back-rotation related to thrusting along the eastern front of the Avawatz Mountains (e.g. Brady, 1984a, 1984b). The origin of the hills to the west, referred to here as the "grave hills", is less clear. Schermer et al. (1996) inferred a contractional-sinistral fault along the eastern side of these hills and they considered this fault an extension of the Drinkwater Valley fault system. Our more detailed work suggests, however, that faulting is minor along this mountain front and the hills are primarily a broad anticline (Figure 2). This interpretation is consistent with the geometry of the Tertiary unconformity on both sides of the hills, but is not well constrained because of limited exposure in the Avawatz Valley.

To the south of the Ft. Irwin fault, the young deformation is more complex and poorly resolved due to limited exposure. In the low hills north of Red Pass Lake (Figure 1) the Tertiary sediments are deformed into a series of open, doubly plunging folds. North of the north Tiefort Mountains fault these folds trend EW to WNW, but south of the fault an open anticline has a NS trending axis nearly perpendicular to the other folds. The origin of these spatial variations in fold axis orientations is unclear but is probably the result of changes through time and variations in shortening directions related to position in the strike-slip system. Temporal variations are suggested by the present geomorphology which suggests that the folds north of the fault are still actively growing because they warp Quaternary surfaces and the youngest unconformity whereas folds to the south lack this geomorphic expression. Thus, the variation could simply be due to variation of shortening direction in time. We suggest, however, that the fold variations reflect actual spatial variations in shortening with the folds north of the fault related to transpressional left slip along the Ft. Irwin fault driving blind thrusts to the north, whereas the folds to the south probably represent transrotational effects at the termination of the Tiefort Mountains fault.

One of the most surprising features recognized in our studies is structural relationships in the vicinity of the Red Pass Range. Mapping by Rodosta (1997) in the Red Pass Range indicated that Tertiary strata correlative with the eastern Ft. Irwin basin are present below a Plio-Pleistocene unconformity at the base of an entrenched alluvial fan sequence. The Tertiary unconformity below the tilted Tertiary strata is exposed locally along the western edge of the Tertiary outcrop area, but the eastern side is clearly a major east-dipping thrust fault that places Mesozoic volcanics above the Tertiary. Gravity data further suggest that the western margin of the Red Pass Range is also an east-dipping thrust fault (Rodosta, 1997). Unlike areas to the west, however, these thrust systems have clearly been inactive for a significant time interval because they are depositionally overlapped by the entrenched Quaternary(?) gravels. Total shortening across both thrusts is not large, but indicates a local EW shortening similar to that indicated by folds immediately to the west.

Relationships at the northern margin of the Red Pass Range provide further indications of the complexity. Hartman (1997) recognized from gravity and magnetic anomaly maps that the deep gravity low produced by the Tertiary Avawatz Formation is offset in a sinistral sense along the northern margin of the Red Pass Range. Although Tertiary Avawatz Formation strata cannot be directly observed beneath cover in the Quaternary trough west of Silver Lake (Figure 1) strata similar to the upper Avawatz Formation are exposed in the eastern Soda Mountains (e.g. Grose, 1959, Rodosta, 1997). Based on this similarity Hartman (1997) interpreted the gravity low west of Silver Lake as a buried segment of the Avawatz Basin. This conclusion together with the results from the central Red Pass Range (Rodosta, 1997) is important because it implies the Fort Irwin fault (Figure 1) does not terminate in the Avawatz Valley, rather it extends into the Silurian Valley were it terminates against the southern Death Valley trough.

Eastern Avawatz Mountains and the nature of the Arrastre Springs fault

Most of the eastern Avawatz Mountains had been examined in considerable detail by Spencer (1990a) and Brady (1984a) and our work in these areas was limited to confirmations of their geologic mapping. Nonetheless, our ties to the geology in eastern Ft. Irwin, together with re-examination of some critical areas suggest three features that have not been previously recognized.

First, gravity data in the central Avawatz Mountains reveal a dramatic gravity high just west of the exposed area of the Avawatz Formation and a deep low over the site of the basin. The gravity high to the west of the basin is consistent with the absence of lower Avawatz Formation strata in eastern Fort Irwin, but begs the question of what structure produces the sudden disappearance of ~5 km of stratigraphic section. Spencer (1990b) clearly demonstrated that the lower Avawatz Formation was deposited during down to the west motion on the Arrastre Springs fault, and thus the western boundary is unlikely to be a normal fault, unless the Avawatz Basin were a true graben. We suggest that the western boundary of the Avawatz basin is a high angle, west side up fault. That is, near Ft. Irwin eastgate, Mesozoic intrusives and volcanic rocks are exposed west of exposures of the Avawatz Formation (Spencer, 1990a, 1990b), implying a major buried fault. Based on structures in the Avawatz Valley, this structure was probably an east directed thrust fault during at least its latest motion because it cuts upper Avawatz Formation. Nonetheless, it could also have pre-upper Avawatz motion, and an intriguing speculation is that structure is some cryptic extension of the apparent structure that follows the Drinkwater Valley or it could be an extension of the Cave Spring fault (Figure 1).

Second, although Spencer (1990b) strongly supported the concept that the Arrastre Springs fault was primarily a Middle Miocene normal fault, it is clear from both Spencer's mapping and our work that the latest motion on the Arrastre Springs system was sinistral-strike-slip. This slip sense is clearly indicated just north of Avawatz Peak in the central Avawatz Mountains where conjugate sinistral and dextral faults offset the Tertiary unconformity along the trace of the Arrastre Springs fault–in this case, the sinistral fault is parallel to the trace of the Arrastre Springs fault. Similarly, there is an ~2.5km sinistral shift of Paleozoic units along a branch of the Arrastre Springs fault in the central Avawatz Mountains which is difficult to reconcile with a pure dip-slip motion on the Arrastre Springs fault (Figure 1). It is unclear how much strike-slip may have been taken up along the Arrastre Springs fault because most makers are ambiguous indicators of slip. From our observations, however, it is a misnomer to describe the Arrastre Springs system as a single fault, or set of distinct faults. Rather, in the northern part of the Avawatz Mountains where we have examined the fault in detail, the Arrastre Springs fault system is a broad zone of cataclastic rock ranging from 500m to 2km in width. Tertiary rocks are clearly caught up and deformed within the fault zone, but local occurrences of an undisturbed unconformity with Tertiary rocks lying on fault rocks clearly demonstrate that at least some of the motion pre-dates the Tertiary sedimentary rocks (e.g. Spencer, 1990a; Brady, 1984a). Nonetheless, this observation could be local remnants of what was clearly an original extensional basin (Spencer, 1990b). Indeed, the faulting in the Arrastre Springs system extends east of all exposures of Tertiary rocks and it is possible that all of the Tertiary rocks lie west of the eastern edge of this fault zone. Thus, all of the Tertiary rocks may been transported along a left slip system. Finally, we note that low-angle faults that place Paleozoic rocks with Tertiary cover onto Mesozoic Avawatz diorite (Figure 1) near Avawatz peak may represent thrust faults. This interpretation is speculative and contrasts with Spencer's (1990b) inference that these faults are low angle normal faults. Nonetheless, the interpretation of these structures as thrusts is more consistent with the dominantly contractional (transpressional) history we recognize in the post-Middle Miocene tectonics of eastern Fort Irwin.

Third, there is a major NS striking fault zone west of the Arrastre Springs fault in the Cave Springs area. This fault, referred to here as the Cave Springs fault, is similar in structural style to the Arrastre Springs fault with a broad (500m to 1km wide) cataclastic zone marking the fault system. Slickenside surfaces and foliated gouge in the Cave Springs fault, like the Arrastre Springs fault, clearly indicate that the last motion on the zone was sinistral(?) strike slip. Unlike the Arrastre Springs fault, however, there is no clear evidence of Tertiary motion on the Cave Springs fault, yet similarity in trend and style to the Arrastre Springs system implies they are related faults.

Together these relationships suggest to us that the eastern edge of the Avawatz Valley has experienced a much more complex deformational history than is generally appreciated. Although this area experienced relatively simple west tilting in the Plio-Pleistocene, the older history is characterized by cryptic NS trending strike-slip faults and younger EW directed contraction. Unfortunately, the relative and absolute ages of these older structures is sufficiently poorly constrained that it is largely speculation to correlate them to distinct events. Nonetheless, there are sufficient gross details to allow a preliminary assessment of this older history once the youngest history is restored.

DISCUSSION

LATE CENOZOIC TRANSROTATION

Our studies in Fort Irwin are consistent with a number of previous studies that have suggested that the northeast Mojave region is dominated by dextral transrotation of crustal panels bounded by sinistral faults (e.g. Garfunkel, 1974, Dokka and Travis, 1990; Schermer et al., 1996, Dickinson, 1996). Our studies provide new constraints on the details of that system, however, and raise some important questions about the three-dimensional interaction through time.

In terms of the Garlock system the most significant result of our studies is the interpretation that the net slip on two of the sinistral faults in eastern Ft. Irwin is significantly greater than previously appreciated. If this slip is, as we infer, due solely to transrotational tectonics these slip estimates imply significantly greater clockwise rotation of the crustal panels in eastern Ft. Irwin and accordingly, imply a greater net dextral shear across the region. That is, using their more conservative slip estimates, Schermer et al. (1996) inferred ~23 degrees of clockwise rotation and ~22km of dextral shear across the region. Using our 10 km slip estimate for the Ft. Irwin Fault as representative of the group, the same model would yield a rotation of 45 degrees and net dextral shear of 55 km.

The EW striking crustal panels involved in this transrotation include one panel with the eastern Garlock as a bounding fault. Thus, if we assume that slip on the faults within Ft. Irwin is representative of the group, then 6-10 km of slip on the Garlock system is solely the product of transrotational tectonics. Although this conclusion does not solve all of the questions on the slip discrepancies along the Garlock fault, it does suggest that a significant fraction of the net slip (up to ~20%) is solely due to transrotation and hypothesized eastward extensions of the Garlock need not exist in the latest Cenozoic.

Perhaps more important, however, is the implication of these slip magnitudes for the southern Death Valley system. It has long been suggested (e.g. Grose, 1959) that the southern Death Valley fault system was relatively continuous along the Silurian Valley-Soda Lake trough and that the Soda-Avawatz Fault system in the eastern Soda Mountains was a southern continuation of that system (e.g. Brady, 1988). Rodosta's (1997) studies suggest, however, that the Soda-Avawatz fault is not active. Moreover, the lack of a clear gravity signature along the assumed projection of the southern Death Valley fault and the dominance of an apparent left slip system projecting as far east as Silver Lake (Hartman, 1997) suggest that the southern Death Valley system is not present south of the Garlock intersection. Indeed, if our estimate of 55km of dextral shear is accurate, virtually all of the Death Valley fault system slip could be absorbed by the left slip faults in Ft. Irwin and there may be no need for a southern extension of the Death Valley fault.

Finally, we note that if the transrotation is as large as we infer, the observations that the latest motion on the Arrastre Springs system was sinistral strike-slip must be considered in the reconstruction of the system. Specifically, the Arrastre Springs fault system clearly merges with the eastern Garlock system in the northern Avawatz Mountains (Figure 1). If this system is restored through 45 degrees of transrotation and the Arrastre Springs system is carried along in that rotation, the present NW strikes of the fault become ~EW after restoration (e.g. Hartman, 1997). Thus, in restored coordinates, the Arrastre Springs system is parallel to the present Garlock system, and by analogy, the Arrastre Springs system may have been an older transrotational extension of the Garlock system. If true, the eastern Avawatz mountains may have experienced as much as 90 degrees of clockwise rotation, a speculation that could be easily tested with paleomagnetic data.

SIGNIFICANCE OF THE LATE MIOCENE HISTORY IN EASTERN FT. IRWIN

Although the latest Cenozoic history of the Garlock-southern Death Valley fault intersection seems reasonably well constrained, the older, Late Miocene history of the system is not clearly resolved. Our work, together with that of Sobieraj (1994) suggests that a distinct sedimentary basin developed over most of the area now occupied by the Avawatz Valley and across the southern Avawatz Mountains/Red Pass Range into the Soda Mountains. The process that formed this basin is not clear but the basin is much more widespread than the basin represented by the lower Avawatz Formation, which is limited to the southern Avawatz Mountains (Spencer, 1990b). The basin could represent a large extensional basin with a low-angle normal fault along the western side of the basin. This hypothesis is supported by the presence of megabreccia sheets in the central part of the basin that appear to contain debris from the western side of the basin in the Tiefort Mountains (e.g. Sobieraj, 1994). Conversely, however, megabreccia sheets in the upper Avawatz Formation are primarily comprised of Mesozoic metavolcanics and Paleozoic carbonates (Spencer, 1990b) which implies an escarpment to the east, south, or both. Moreover, clast and facies distributions in the basin suggest the northern part of the basin represented a large alluvial fan complex feeding into a playa to the south with no clear evidence of major fans along the western side of the basin until near the top of the Tertiary section. Finally, there are no clear structures indicative of an EW directed (present coordinates) extensional system and the only clear extensional structures we recognized in our studies were low-angle normal faults with top to the south (present coordinates) offsets (e.g. Dean, 1995).

Together we believe that these observations imply that the eastern Ft. Irwin basin is probably not an extensional basin. It may represent a post-extensional overlap assemblage developed after the extension recorded by the Avawatz Basin, yet that conclusion seems inconsistent with the extensive development of megabreccia sheets within the basin–implying steep topography, unlikely in a successor basin–and development of complex angular unconformities within the basin.

We speculate that this problem is closely tied to the older, Middle Miocene history of the region and the unresolved problems of the mismatch of geologic structures across the Drinkwater Valley. If the Avawatz Mountains have been rotated 90 degrees the entire Middle Miocene paleogeography is suspect. For example, was this basin developed along a NW striking ancestral Garlock fault that terminated into a transrotational domain that is now entirely fragmented by younger events? Does the mismatch in geology across the Drinkwater Valley imply the existence of cryptic fault(s) that have been overlapped by younger sediments? All of these questions are, as yet, unresolvable, but with additional paleomagnetic data as well as further information from the other areas around the Garlock intersection, this paleogeography may become resolvable.

Acknowledgements

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Preliminary Slip-Rate Estimate for the Owl Lake Fault, California

Sally F. Mcgill, *Department of Geological Sciences, California State University, San Bernardino, CA 92405*

ABSTRACT

Near the southwestern end of the Owl Lake fault, a terrace riser is offset about 80 m left-laterally and 2-28 m south-side down. AMS radiocarbon ages of organic matter collected from 2 boulders on the surface into which the riser incised are about 30 ka. This indicates a slip rate of about 2.5 mm/^{14}C-yr left laterally, and 0.06 to 0.9 mm/^{14}C-yr south-side down. The amount of left-lateral offset is poorly constrained and could be anywhere between 19 and 235 m. Thus the minimum and maximum left-lateral slip rate are 0.5 and 7.8 mm/^{14}C-yr, respectively. The true rate is probably close to the preferred rate of 2.5 mm/^{14}C-yr.

Introduction

The Owl Lake fault diverges from the eastern Garlock fault within the Quail Mountains and extends 19 km northeasterly into the Owlshead Mountains in southeastern California (Figure 1). Muehlberger (1954) mapped the southwestern 7 km of the fault in the Quail Mountains. In this area, he describes the fault as a reverse fault with the north side uplifted, although he notes that farther northeast the fault becomes a normal fault with the south side uplifted. Muehlberger (1954) also noted the presence of scarplets, undrained depressions and offset land surfaces, indicating Recent movement on the fault. Clark (1973) mapped the recently active breaks of the Owl Lake fault, and noted channels and ridges offset left-laterally along the southwestern two-thirds of the fault. He found no evidence for recent lateral displacement along the northeastern one-third of the fault. Left-lateral slip on the Garlock fault is probably partitioned onto the Owl Lake fault and the eastern Garlock fault east of the intersection of the two faults.

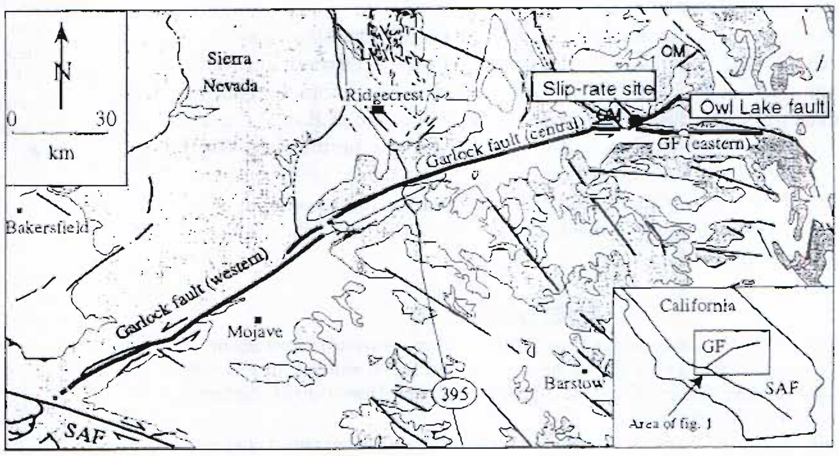

Fig. 1: Reference map showing location of the Owl Lake Fault and the slip-rate site. GF, Garlock fault; QM, Quail Mountains; OM, Owls Head Mountains; SAF, San Andreas fault.

At a site near the southwestern end of the fault (Figure 2), a terrace riser on the eastern side of an incised channel has been left-laterally offset (Figures 3 and 4), providing a piercing line to measure the slip rate of the Owl Lake fault at this location. The channel has incised into late Tertiary or early Quaternary Funeral fanglomerate (TQf) on the northern side of the fault (Muehlberger, 1954), and has incised into late Quaternary alluvial fan deposits on the southern side of the fault. The riser between the Qa3 and Qa4 surfaces could not have been cut in its present position. Some combination of left-lateral and down-to-the-south displacement must have offset this riser from its source: a large, incised channel north of the fault. This report refines the tentative offset measurement and slip rate reported by McGill (1993).

Left-lateral Displacement of the Terrace Riser

The top of the terrace riser between Qa3 and Qa4 is well defined on the south side of the fault. Within about 50 m south of the fault, however, it has been covered by colluvium (Qc) shed from the north side of the fault (Figure 3). This has produced an apparent left bend of the buried riser at the point where it intersects the southern limit of the colluvial wedge. This apparent bend is present both on the riser between Qa3 and Qa4 and on the riser between Qa2 and Qa3 (Figure 3). The schematic cross section shown in Figure 5 illustrates how this effect occurs. To estimate the location of the piercing point on the south side of the fault, I have projected the top of the terrace riser from south of the colluvial wedge to the fault (Figure 3).

The actual intersection of the top of the riser with the fault plane is buried by colluvium. The elevation of this piercing point can be estimated from topographic profile AA' (Figures 3 and 6), which suggests that the top of the Qa4 alluvial fan deposits intersects the fault at an elevation of 1201 to 1207 m. The projected elevation depends on whether the points far from or near to the fault are used to define the slope of Qa4 that is projected to the fault.

For simplicity, I have assumed that the fault plane is vertical. There are no exposures of the fault plane in the study area, but the linear trace of the fault suggests that it is steeply dipping to vertical. About

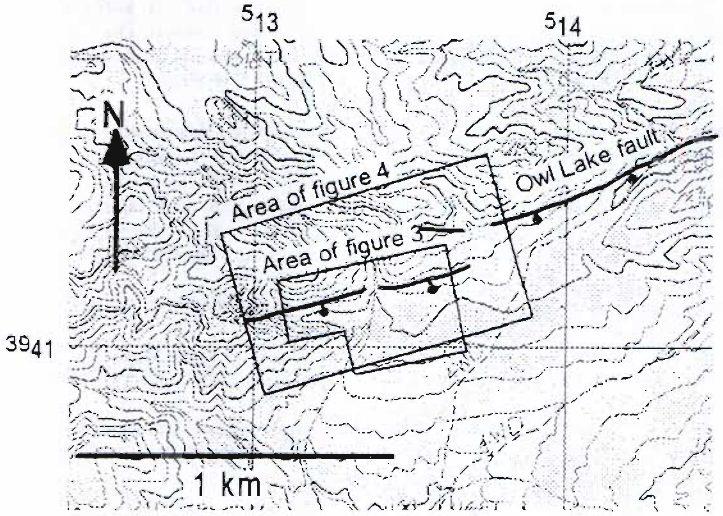

Fig. 2: Location of the slip-rate site on the Owl Lake fault. Base map is Leach Spring, California 7.5' quadrangle. Contour interval, 10 m. Fault location after Clark (1973).

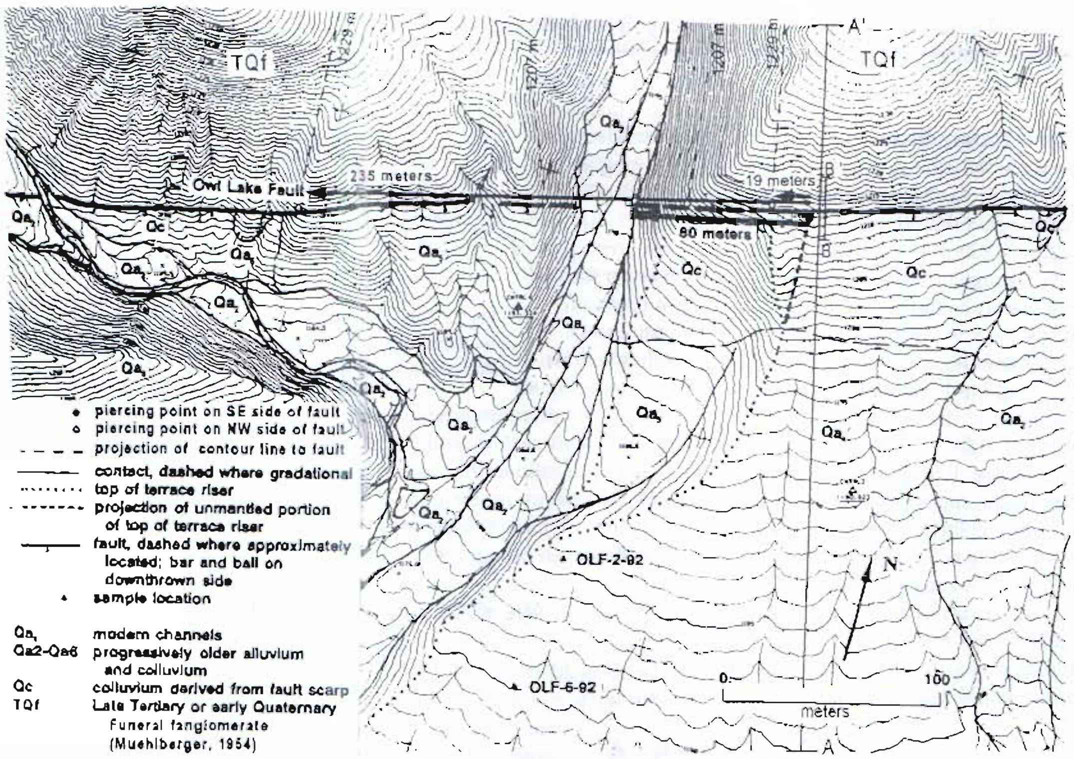

Fig. 3: Preliminary geologic map of slip-rate site on the Owl Lake fault, showing terrace riser between Qa3 and Qa4 offset about 80 meters (minimum =19 m; maximum = 235 m). Contour interval is 1 meter. Map was constructed photogrammetrically by Advanced Digital Maps, Inc., Azusa, Calif.

trace of the fault suggests that it is steeply dipping to vertical. About 1.5 km southwest of the slip-rate site the dip of the fault is 83º S (Muehlberger, personal communication, 1998). Using this dip instead of assuming a vertical fault plane does not significantly change the elevation at which the top of the Qa4 surface projects to the fault.

On the north side of the fault there are no Qa3 or Qa4 terraces. These terraces either did not form, or they have been eroded. The fact that the riser between Qa3 and Qa4 is not preserved on the north side of the fault leads to a large uncertainty in the location of the piercing point on the northwest side of the fault. At the time that it formed, the riser between Qa3 and Qa4 must have been located directly across the fault from some point within the large, incised channel north of the fault. If no vertical slip has occurred, then this point must be located somewhere between the projection of the 1207-m contour lines on either side of the channel to the fault (Figure 3).

The presence of south-facing scarps, however, suggests that some south-side-down displacement has occurred. If so, then the top of the Qa4 fan deposits originally intersected the fault plane at an elevation higher than 1201 to 1207 m., and the range of possible piercing point locations on the north side of the fault is wider.

I use the thickness of the colluvial wedge (Qc) derived from the fault scarp and shed onto Qa4 to estimate the maximum amount of vertical slip that has occurred on the Owl Lake fault since the Qa4 surface formed. The cross-sectional area of the colluvial wedge is 192 to 358 m^2 (Figure 6). I assume that the cross-sectional area eroded from the north side of the fault is approximately equal to the cross-sectional area of the colluvial wedge. The original slope of the surface eroded into TQf north of the fault is not known, but it was probably no steeper than the current slope at the north end of profile AA' (about 10º), and it was probably no less steep than the slope of Qa4 (about 5º). These constraints indicate that prior to vertical displacement, the Qa4 surface intersected the fault at 1223 to 1229 m elevation (Figure 6).

The projection of the 1229-m contour lines on either side of the channel to the fault provide boundaries within which the piercing point on the north side of the fault must lie (Figure 3). These boundaries suggest that the left-lateral offset of the riser between Qa3 and Qa4 is between 19 and 235 m. These bounds are extreme values, however. At the time that it formed, the riser between Qa3 and Qa4 was most likely located near the middle of the large incised channel north of the fault. The walls of the incised channel upstream from the fault have most likely been eroding, so the walls of the channel (and the 1229-m contour lines on each side of the channel) were probably closer together at the time the riser between Qa3 and Qa4 formed.

For my preferred estimate of the location of the piercing point on the north side of the fault, I assume that the channel that created the riser between Qa3 and Qa4 was the same width as and was centered in the same location as the modern channel north of the fault. The preferred location of the piercing point on the north side of the fault is

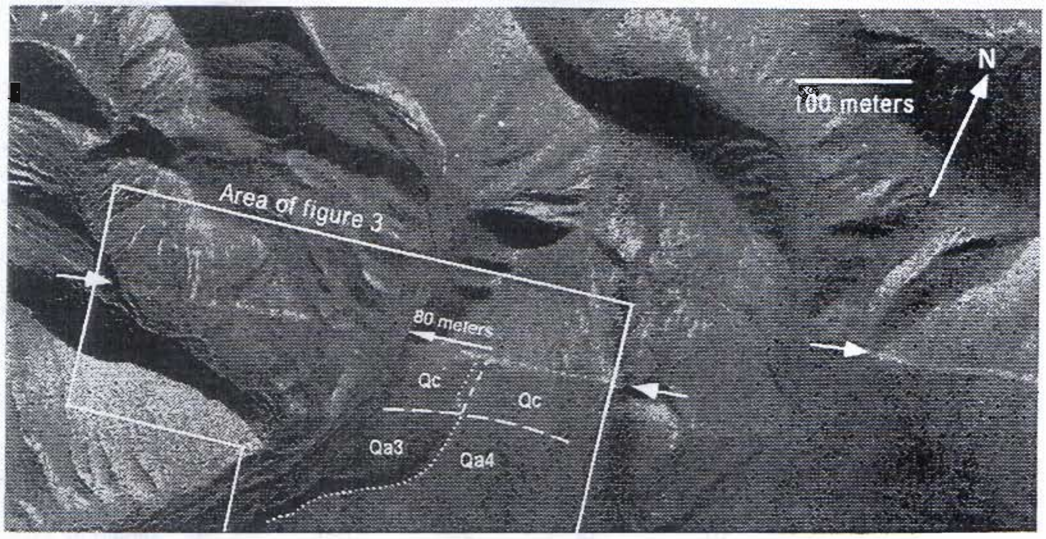

Fig. 4: U.S. Geological Survey aerial photograph of slip-rate site on the Owl Lake fault. Dotted line follows top of terrace riser between Qa3 and Qa4. Short-dashed line shows projection of unmantled terrace riser to fault. Long-dashed line shows toe of colluvial wedge. White arrows mark the location of prominent fault scarps, the north direction, and the preferred left-lateral offset estimate of 80 meters.

thus at the intersection of the eastern edge of the modern channel with fault (Figure 3). The preferred value for the left-lateral displacement of the riser between Qa3 and Qa4 is 80 m.

Dip-slip Displacement of the Terrace Riser

The maximum vertical displacement since the time that deposition of Qa4 ceased can be calculated from Figure 6. As discussed above, the top of the Qa4 deposits may have intersected the north side of the fault plane at an elevation as high as 1223 to 1229 m prior to vertical displacement. Combined with the elevation range at which the top of Qa4 currently intersects the south side of the fault plane (1201 to 1207 m), this indicates that 16 to 28 m of south-side down displacement has occurred since deposition of Qa4 ceased. Assuming a fault dip of 83°S instead of 90° only changes the amount of dip-slip by about 1%.

This is a maximum estimate of the amount of dip-slip, because it is calculated from the cross-sectional area of the colluvial wedge, and no dip-slip is required to produce a colluvial wedge. Pure left-lateral offset of the Qa4 fan would juxtapose it against higher topography on the north side of the fault, which would then shed colluvium onto the Qa4 fan. As mentioned above, however, the presence of south-facing fault scarps all along this portion of the fault (Figure 4) suggests that some amount of south-side-down displacement has occurred.

Topographic profile BB' (Figures 3 and 7) was surveyed across the fault scarp using a total station (a combination of a theodolite and an electronic distance measuring). It provides greater resolution of the steep, fresh fault scarp than is visible on profile AA', which was constructed from the topographic map. Profile BB' suggests that the steep, fresh part of the scarp formed as a result of about 2 m of south-side-down movement (Figure 7). This is a minimum estimate of the amount of dip-slip since the time that deposition ceased on the Qa4 fan. In fact, it may represent the amount of dip-slip in the most recent earthquake alone. The continuous nature of this fresh scarp across minor variations in topography (Figure 4) indicates that this scarp did not form by lateral offset of uneven topography.

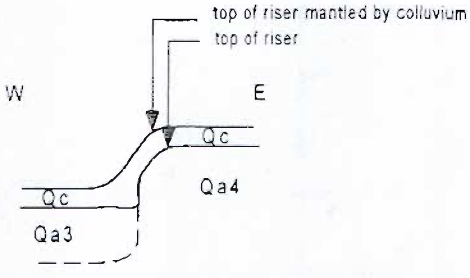

Fig. 5: Schematic cross section illustrating why the mantled top of the terrace riser is located farther west than the actual top of the terrace riser.

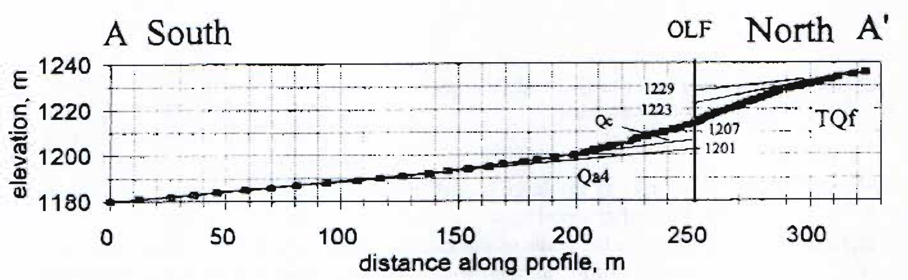

Fig. 6: Topographic profile AA'. No vertical exaggeration. See Fig. 3 for location of profile and definition of geologic units TQf, Qa4 and Qc. The top of Qa4 projects to the fault at an elevation of 1201 to 1207 m. The surface north of the fault probably intersected the fault at an elevation of 1223 to 1229 m, prior to erosion. OLF, Owl Lake fault.

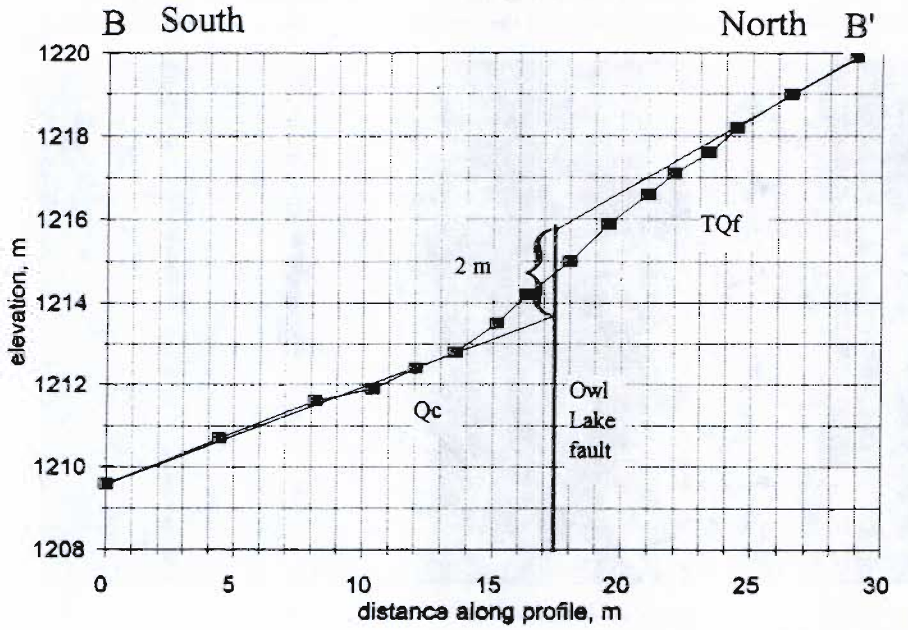

Fig. 7: Topographic profile BB'. See Fig. 3 for location of profile and definition of geologic units TQf and Qc.

Age of the Terrace Riser and Slip Rate of the Fault

Dr. Ronald Dorn and I collected surface samples from six boulders on the surface of the Qa4 deposit. Dr. Dorn extracted organic material from the weathering rind beneath a coating of rock varnish on two of the samples (Figure 8). The organic matter was radiocarbon dated by accelerator mass spectrometry at the Institute of Geological and Nuclear Sciences at New Zealand's Department of Scientific and Industrial Research. The results are shown in Table 1. These ages represent the time at which organisms living within the weathering rind of the rock were killed by being covered with rock varnish (Dorn, oral communication). Varnish begins to form on a surface within about a hundred years after exposure to the atmosphere (Dorn and Whitley, 1984; Dorn and others, 1988), and complete covering of a surface with varnish takes even longer. The reported ages are thus minimum estimates of the age of abandonment of the surface of the Qa4 deposit. Dorn and others (1989) has shown, however, that radiocarbon dates on organic material extracted from the basal layers of rock varnish are usually no more than 10% younger than radiocarbon dates on charcoal, shells and tufa for several different surfaces of Holocene and Late Quaternary age. Thus, deposition of Qa4 probably ceased and permanent exposure of the surface began within a few thousand years before 30,000 ^{14}C-yr B.P. Abandonment of the surface of the Qa4 deposit was most likely caused by incision of the offset terrace riser, so the age of abandonment of the Qa4 surface probably approximates the age of formation of the offset terrace riser between Qa3 and Qa4. If this is the case, then the 19 to 235 m of left-lateral slip occurred over the past 30-33 kyr or so, yielding a preliminary left-lateral slip rate estimate within the range of 0.6 to 7.8 mm/^{14}C-yr. My preferred estimate of the left-lateral slip rate is 2.5 mm/^{14}C-yr (80 m / 31.5 kyr). The rate of dip-slip is 0.06 to 0.9 mm/^{14}C-yr, southeast-side down.

These slip-rate estimates are based on the knowledge currently available. Ideally the age of the Qa4 surface should be corroborated by other dating techniques, and this may lead to revision of the slip rate estimate in the future. Future work may also provide tighter constraints on the vertical displacement of the Qa4 surface, which in

TABLE 1. RADIOCARBON AGES ON ORGANIC MATTER FROM BENEATH ROCK VARNISH

Sample #	Lab #	Graphite yield, mg	δ13C, %	Conventional Radiocarbon Age, years B.P.
OLF-2	R 18089/3 NZ/RD-64	3	-23.2	30,820 ± 280
OLF-6	R 16149/8 NZ/RD-57	2.5	-23.2	29,490 ± 270

turn could substantially tighten the constraints on the left-lateral offset.

The slip-rate estimates reported here may not apply to the entire length of the Owl Lake fault zone. As mentioned above, Clark (1973) found no evidence of recent lateral displacement along the northeastern one-third of the fault. In addition, these estimates only apply to the strand of the Owl Lake fault shown in Figure 3. A parallel strand of the fault is located about 120 m north of the strand shown in Figure 3 (Muehlberger, 1954). In the vicinity of the offset terrace riser, however, the northern strand does not show much evidence for recent movement (Figures 2 and 4, and Clark, 1973). It appears that recent movement on the Owl Lake fault has primarily occurred on the southern strand within the area of Figure 3, and that it steps over to the northern strand to the east of Figure 3.

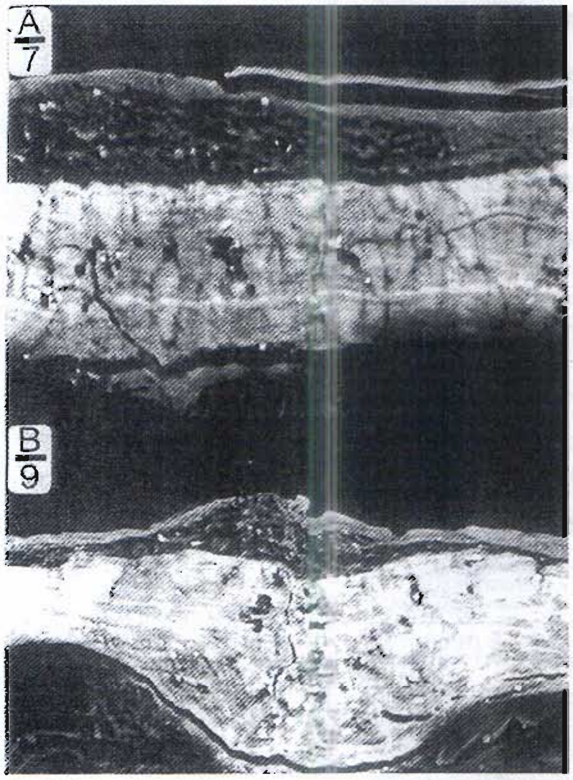

Fig. 8: Backscatter electron microscope images of cross sections through rock varnish on samples OLF-2 (A) and OLF-6 (B). *Courtesy of Ronald I. Dorn.*

Acknowledgments

I thank Joe Grant for assistance with field work. Ron Dorn prepared the rock varnish samples for AMS radiocarbon dating at New Zealand's Department of Scientific and Industrial Research. The Explosive Ordnance Disposal unit at George Air Force base provided the required escort onto this portion of Fort Irwin.

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Summary of Neotectonic Slip-rate Studies of the Garlock and Owl Lake Fault Zones

Sally F. Mcgill, *Department of Geological Sciences, California State University, San Bernardino, CA 92405*

INTRODUCTION

The left-lateral Garlock fault is one of the major tectonic elements in southern California. It separates the Basin and Range Province, the Sierra Nevada and the Tehachapi Mountains on the north from the Mojave block on the south. The left-lateral, northeast- to east-striking Garlock fault is transverse to the dominant pattern of northwest-striking, right-lateral faults in California, and its tectonic role is still not completely understood. Several investigators have proposed that the Garlock fault is a transform fault, with left-slip on the Garlock fault accommodating extension in the Basin and Range province relative to the Mojave block, which is not currently extending (Hamilton and Myers, 1966; Troxel and others, 1972; Davis and Burchfiel, 1973). The location of the Garlock fault at the southern margin of the western Basin and Range province is consistent with this hypothesis, as is the apparent eastward termination of the Garlock fault at the eastern limit of significant Quaternary extension in the western Basin and Range province. However, the northeast- to east-striking Garlock fault is not parallel to the northwestward extension direction in the Basin and Range province, as one would expect a transform fault to be (Stewart, 1983; Burchfiel and others, 1987; Jones, 1987; Minster and Jordan, 1987; Wernicke and others, 1988).

Studies of pre-Quaternary geologic features offset by the fault have provided much information about the total left-lateral displacement across the Garlock fault, as well as some clues to the age and kinematic history of the fault (Bohannon and Howell, 1982; Brady, 1986; Carr and others, 1993; Carter, 1980, 1984; Davis and Burchfiel, 1973; Hill and Dibblee, 1953; Michael, 1966; Smith, 1962; Smith and Ketner, 1970; Troxel and others, 1972). Neotectonic studies of the slip rate of the Garlock fault can also play in role in constraining tectonic models for the Garlock fault. This paper briefly summarizes estimates of the neotectonic (mostly Holocene and late Pleistocene) slip rate of the Garlock fault.

NEOTECTONIC SLIP RATE OF THE GARLOCK FAULT

Published estimates of the slip rate of the Garlock fault are summarized in Table 1 and Figure 1, and are discussed below, beginning at the western end of the fault and progressing eastward. The references cited provide more detailed location maps of the slip-rate sites.

Oak Creek Canyon

Eight stream channels incised into late Pleistocene alluvium at Oak Creek Canyon have been offset about 300 m left-laterally, suggesting a slip rate of 1.6 to 3.3 mm/yr (LaViolette and others, 1980). The Late Pleistocene age is estimated from the degree of soil development. It is not constrained by radiocarbon ages at the site, nor is there any documented correlation with dated soils elsewhere.

Lone Tree Canyon

On the western Garlock fault, near the town of Mojave, an incised channel is offset at least 60 ± 5 meters, left-laterally. Preliminary results were reported by McGill (1994a, b). New radiocarbon dates on charcoal samples collected from trenches excavated at the site indicate that the offset channel incised sometime between 11,470 ± 140 and 7170 ± 140 radiocarbon years ago (McGill, in preparation). This indicates a minimum left-lateral slip rate of about 7.0 ± 2.2 mm/radiocarbon-year for the western Garlock fault. Using the dendrochrono-logically calibrated age ranges for the dates reported above, the slip rate is 6.15 ± 2.15 mm/calibrated year. Additional slip may have occurred on a secondary fault strand about 350 m to the northwest, at the range front of the southern Sierra Nevada.

Koehn Lake

A gravel bar near the late Pleistocene shoreline of Koehn Lake has been offset about 80 ± 5 m left laterally across the Garlock fault (Carter, 1971; 1980; 1987; 1994; Clark and Lajoie, 1974; Clark and others, 1984). The gravel bar has also undergone 7-8 m of north-side-down vertical displacement (Carter and Lauman, 1983). Tufa deposited near the crest of the gravel bar has a radiocarbon age of 11,360 ± 160 radiocarbon years B.P. (Clark and Lajoie, 1974). Ostracods collected from a trench exposure of lacustrine clay presumed to be deposited

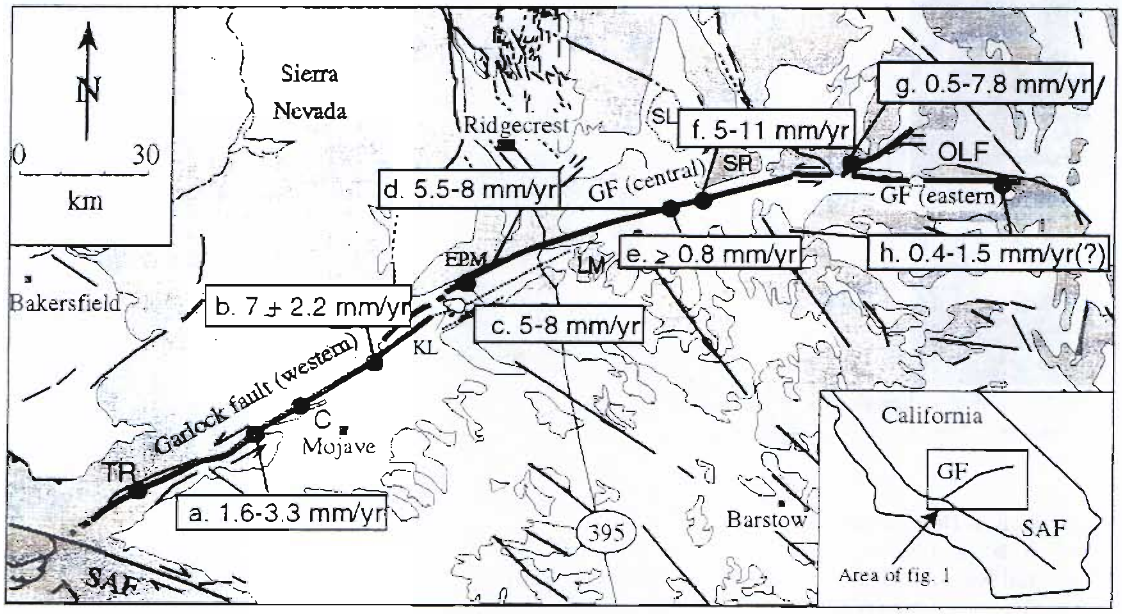

Fig. 1. Locations of slip-rate estimates along the Garlock fault. See Table 1 for references. C, Cameron; **GF**, Garlock fault; **KL**, Koehn Lake; **OLF**, Owl Lake fault; **SAF**, San Andreas fault; **SL**, Searles Lake; **SR**, Slate Range; **TR**, Tejon Ranch.

TABLE 1. LEFT LATERAL SLIP-RATE ESTIMATES FOR THE GARLOCK AND OWL LAKE FAULTS

Location	Left-lateral slip rate mm/^{14}C-yr	Reference
a. Oak Creek Canyon	1.6-3.3	LaViolette and others (1980)
b. Lone Tree Canyon	7 ± 2.2	McGill (1994a and in preparation)
c. Koehn Lake	5-8	Clark and Lajoie (1974); Clark and others (1984)
d. El Paso and Lava Mts	5.5-8	Carter (1994)
e. Christmas Canyon	≥0.8	Smith (1975)
f. Searles Valley	5-11	McGill and Sieh (1993)
g. Owl Lake fault	0.5-7.8	McGill (1993 and this volume)
h. Cave Spring Wash	0.4-1.5	McGill (1994c)

during the latest highstand of Koehn Lake have a radiocarbon age of 14,700 ± 130 radiocarbon years B.P. These data indicate a left-lateral slip rate for the Garlock fault of 5-8 mm/radiocarbon year, with a preferred value of 7 mm/radiocarbon year (Clark and others, 1984).

El Paso Mountains and Lava Mountains

Carter (1994) discusses evidence that the left-lateral slip rate of the Garlock fault has remained between 5.5 and 8 mm/yr since its inception in late Miocene time. This evidence is summarized in Table 2. Carter (1994) also notes that extrapolating the 5.5 to 8 mm/yr slip rate to the total left-lateral offset (about 64 km; Smith, 1962; and other references) implies that left-lateral movement on the Garlock fault began between 8 and 12 Ma, which is consistent with other estimates of the age of inception left-slip on the fault (Burbank and Whistler, 1987; Loomis and Burbank, 1988).

Christmas Canyon

Smith (1975) describes evidence near Christmas Canyon for two horizontal displacements totalling 8 m that have probably occurred within the past 10,000 years, yielding a left-lateral slip rate of at least 0.8 mm/yr.

Searles Valley

The highstand shoreline of Pleistocene Searles Lake crosses the Garlock fault in several places. The intersection of this shoreline and the fault is best preserved near the southwestern corner of the Slate Range, where the lake overflowed through Pilot Knob Valley (see figure 4 in McGill 1994a for a detailed location map). Unfortunately, access to this site is limited because it is located within the China Lake Naval Air Weapons Station. At this site, an abrasion platform and sea cliff have been cut by wave action into older alluvium and older lacustrine sediments. The shoreline angle, the linear feature formed by the intersection of the sea cliff and the abrasion platform, is offset in a left-lateral and vertical sense across two subparallel fault strands. Excavations revealed that the shoreline angle is left-laterally offset a total of 82-106 m, with a preferred value of 90 m. The shoreline is offset about 37 m (36-38m) across the southern fault zone, about 46 m (42-46m) across the northern fault zone, and about 7.4 m (3.6-18.6 m) across minor faults between the two major fault zones. A detailed account of how the offset across each strand was measured is given by McGill (1992), and a briefer summary is given by McGill and Sieh (1993). Constraints on the ages of Pleistocene lakes in Searles Valley are provided by Stuiver and Smith (1979) and by Benson and others (1990). The selection of appropriate dates for the minimum, maximum and preferred age of abandonment of the offset shoreline of Searles Lake are discussed by McGill (1992) and by McGill and Sieh (1993). The resulting slip rate is estimated to be between 5 and 11 mm/radiocarbon-year with a preferred value of 6-8 mm/radiocarbon-year.

TABLE 2. SUMMARY OF OFFSET GEOLOGIC FEATURES IN EL PASO AND LAVA MOUNTAINS (CARTER, 1994)

Deposit	Source	Age	Left lateral offset
Gravels north of fault	near Christmas Canyon	Quaternary(?)	17 km
Gravels south of fault	Mesquite Canyon	Pleistocene and upper Pliocene(?)	19 km
Conglomerates in Almond Mtn. volcanics	Mesquite Canyon	Pliocene	32 km
Slide blocks of Garlock Series and Mesquite schist within Bedrock Springs Formation	Mesquite Canyon	late Miocene (6-9 Ma?)	47 km

Cave Spring Wash

Several Holocene or latest Pleistocene alluvial surfaces and terrace risers are left-laterally offset just east of Cave Spring Wash, near the eastern end of the Garlock fault. McGill (1994c) reports a preliminary left-lateral slip rate of 0.4 to 1.5 mm/yr. This rate is quite uncertain, however, because further field mapping is necessary to substantiate the correlation of surfaces and terrace risers across the fault, and because the age control used in the above estimate was based on cation ratio dates obtained on rock varnish samples. New radiocarbon ages on rock varnish from the site are younger than the cation ratio ages (Alan Watchman, personal communication) and may indicate a higher slip rate, but further work is necessary to obtain more reliable and consistent age control at the site.

Owl Lake fault

The Owl Lake fault diverges northeastward from the Garlock fault in the Quail Mountains. At a site near the southwestern end of the Owl Lake fault, a terrace riser is offset about 80 m left-laterally and 2-28 m south-side down. AMS radiocarbon ages on organic matter collected from 2 boulders on the surface into which the riser incised are about 30 ka. This suggests a left-lateral slip rate of about 2.5 mm/radiocarbon-year and a southeast-side down slip rate of 0.06 to 0.9 mm/radiocarbon-year. The left-lateral offset of the riser is poorly constrained, however, and could be anywhere between 19 and 235 m. Thus the maximum and minimum estimates of the left-lateral slip rate are 0.5 and 7.8 mm/radiocarbon-year, respectively, but the true rate is probably close to the preferred rate of 2.5 mm/radiocarbon-year (McGill, 1993 and this volume). The transfer of left-lateral slip onto the Owl Lake fault may reduce the slip rate of the Garlock fault to the east of the juncture between the two faults.

ASEISMIC CREEP ON THE GARLOCK FAULT

Several authors have reported aseismic creep on the western Garlock fault. Louie and others (1985) report left-lateral, aseismic creep at a rate of 3.7 ± 1 mm/yr from 1971 to 1982 at Cameron. Snay and Cline (1980) report left-lateral creep at a rate of 1.7 ± 0.5 mm/yr from 1964 to 1972 at Tejon Ranch. Rodgers (1979) also reports probable left-lateral aseismic creep in the vicinity of Tejon Ranch.

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Tertiary and Quaternary Fault History of the Intersection of the Garlock and Death Valley fault zones, southern Death Valley, California

Bennie W. Troxel and **Paul R. Butler**, *Dept. of Geology, University of California, Davis CA 95616*

ABSTRACT

The northern front of the Avawatz Mountains, southern Death Valley, is the zone of intersection of the eastern segment of the Garlock fault zone and the southeastern segment of the Death Valley fault zone. Both fault zones appear to die out in this region or are truncated by the Mule Spring fault.

Rocks range in age from Precambrian to Holocene. The oldest rocks are Precambrian gneiss overlain by later Precambrian sedimentary rocks of the Crystal Spring and Kingston Peak Formations. They are succeeded by strata of parts of the Noonday Dolomite and Johnnie Formation of later Precambrian age and undifferentiated strata of Paleozoic age. Mesozoic rocks include plutonic rocks that range in composition from diorite to granite. Tertiary rocks include volcanic and sedimentary rocks. They are overlain by several successive units of Quaternary fan and stream channel gravel.

The Mule Spring fault crosses the northern front of the Avawatz Mountains and separates diorite to the south from all other rocks along the northern margin of the Avawatz Mountains. Rocks in the Avawatz Mountains have been uplifted along the south-dipping Mule Spring fault. Reverse movement has also occurred along south-dipping segments of the Death Valley fault zone in Tertiary and Quaternary time. Lateral slip on the Death Valley fault zone appears to have occurred only during Tertiary time. Small grabens developed near the east end of the Garlock fault involve very young Quaternary gravel.

INTRODUCTION

This mapping project was undertaken to study the complex region of intersection along the Garlock and Death Valley fault zones in southern Death Valley to gain information about the timing and amount of movement along them. Mapping was done during the winter months of 1978 and 1979 on photographs of the GS-CW series enlarged four times to a scale of 1:7,100 and compiled on the north half of the U.S. Geologic Survey 15-minute Avawatz Pass quadrangle, enlarged to a scale of 1:24,000.

The time allotted to this study was directed primarily towards resolving questions about provenance of Tertiary sedimentary rocks, and detailed mapping of fault traces. Regional implications of the nature of movement along the Garlock and Death Valley fault zones are beyond the scope of this study.

Various parts of the area have been mapped and described where mineral deposits of economic or potential economic significance have been discovered. The principal reports are by Noble and others (1922) who discussed a sampling program in search of nitrates during World War I, Durrell (1953) who mapped and described deposits of celestite during World War II, and Ver Planck (1952 and 1957) who described deposits of gypsum and halite. Wright (1968) mapped Precambrian rocks near Sheep Creek where talc deposits occur.

Further work must be done in order to obtain additional information about the rock units in and adjacent to the fault zones. Specifically, data from the Quaternary rocks are especially important if the history of Quaternary faulting is to be fully documented.

Informal names have been added to the map for convenience in referring to specific localities.

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DESCRIPTIVE GEOLOGY

GENERAL FEATURES

The rocks adjacent to or within the fault zone in the area of study range in age from Precambrian to Holocene and include igneous, metamorphic, and sedimentary rocks. Many rock units are sufficiently diagnostic to be useful in making determinations of the type of offset along the faults. In general, they include foliated to massive metamorphic rocks of Precambrian age, and later Precambrian sedimentary rocks of the Pahrump Group and the overlying Noonday Dolomite and Johnnie Formation. Mesozoic intrusive rocks lie within or border the fault zones and include diorite, quartz diorite, granodiorite, quartz monzonite, and possibly granite. Tertiary sedimentary rocks abound between the fault zones. They include coarse conglomerate and sandstone developed in fans, sandstone and siltstone deposited in down-stream basins, evaporite deposits and fine-grained siltstone formed in playa basins. Local patches of Tertiary volcanic rocks occur at the base of Tertiary sedimentary rocks within the fault zones. Tertiary rocks occur mostly in belts within the fault zone. Pre-Tertiary rocks occur in two faulted slices and in the Avawatz Mountains. Quaternary rocks include a succession of coarse fan and stream channel gravel and talus that probably were deposited continuously during the Quaternary Epoch.

In the succeeding description of rock units special attention is devoted to stressing diagnostic characteristics useful for recognition of units and to interpret their significance in measuring offset or lack of offset along the lateral-slip faults. Further work is contemplated to reconstruct stratigraphic sections, obtain petrologic information, and determine realistic environmental settings for the various rocks.

PRECAMBRIAN AND PALEOZOIC ROCKS

Crystalline rocks

Wright (1968, Plate 1D, p 28) mapped and described rocks of the Crystal Spring and Kingston Peak Formations of the Pahrump Group

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and the underlying older crystalline rocks exposed at Sheep Creek Spring on the north slope of the Avawatz Mountains.

The crystalline rocks are mostly quartz-feldspar gneiss, with mica and mafic minerals. In general, the gneiss is medium grained and moderately pale. Dikes of pegmatite, fine-grained mica and mafic minerals (lamprophyre) and diabase, all Precambrian in age, intrude the crystalline rocks. The gneiss forms moderately steep slopes that have a faint brownish weathered color and the slopes have a veneer of equant angular fragments from about one to five inches in diameter that create poor support to the unwary investigator. Foliation is common in the gneiss but locally it is massive. Pink feldspar commonly forms a distinct and closely-spaced augen texture. Elsewhere the rock is marked by dark mica flakes concentrated in closely-spaced wavy foliation planes. The gneiss is distinct but it is indistinguishable from Precambrian gneiss exposed throughout the southern Death Valley region north of the Garlock fault zone.

Later Precambrian Pahrump Group

Only rocks of the Crystal Spring and Kingston Peak Formations comprise the Pahrump Group in the area of study – the Beck Spring Dolomite being conspicuous by its absence. The Crystal Spring Formation is the oldest formation of the Pahrump Group (Hewett, 1940) and has no significant stratigraphic variations from other localities where it crops out in southern Death Valley. Studies by Wright (1968), Kupfer (1960), Roberts (1974), and Maud (oral communication) demonstrate that the stratigraphy of the Crystal Spring Formation, including sills of diabase and the presence of commercial talc bodies along the diabase contacts at Sheep Creek Spring fits a regional pattern and thus precludes significant lateral offset but does permit offset of a few miles. The stratigraphic succession of the Crystal Spring Formation consists of a lower feldspathic quartzite member (arkose), purple argillite, fine-grained quartzite, a dolomite member with interbedded shaly and quartzitic strata, chert, and an upper sedimentary sequence made up of interbedded, carbonate strata (some of which are algal rich), quartzite, siltstone, and fine-grained micaceous rock. The rocks of the Crystal Spring Formation are easily distinguished from other pre-Paleozoic strata in southern Death Valley. Individual clasts or suites of clasts derived from the Crystal Spring Formation can be readily identified in younger gravel deposits.

In southern Death Valley, the Crystal Spring Formation is ordinarily overlain by the Beck Spring Dolomite which, in turn, is overlain by the Kingston Peak Formation. In a regional context, based largely on personal observations by Troxel, the Beck Sporing Dolomite occurs as moderately homogeneous pale gray dolomite that occurs without significant changes in a belt of outcrops extending west-northwest from the Kingston Range to south-central Death Valley. Southwest of this belt of outcrops the Beck Spring, where exposed, becomes much more stratified and contains interbeds of siliceous clastic material and tan dolomite beds. Still farther southwest the Beck Spring is not exposed.

If these variations reflect a facies change from dolomite in the northeast to siliceous clastic sediments and perhaps no clastic sediments or dolomite to the southwest, then the lack of Beck Spring Dolomite along the belt of lateral-slip faults has no structural significance. The section of Pahrump rocks in the Silurian Hills a few miles east of the Avawatz Mountains contains no Beck Spring Dolomite (*Kupfer, 1960*) but there the top of the crystal Spring Formation is overlain by about 150 feet of limestone, which Kupfer suggests may be a stratigraphic correlative of the Beck Spring Dolomite. If so, then the facies change in the Beck Spring Dolomite can be demonstrated in an area not transected by either the Garlock or Death Valley fault zones. The lack of Beck Spring Dolomite at Sheep Creek thus need not be dependant upon large scale lateral transport along the lateral-slip faults.

The Kingston Peak Formation consists of two facies in the southern Death Valley region (Troxel, 1967). A northern facies forms a west-northwest-trending belt about 230 miles wide that extends from the type locality in the Kingston Range to the Death Valley fault zone in south-central Death Valley. The southern facies is readily distinguishable from the northern facies by a discrete difference in stratigraphic sequence, suites of clasts in conglomerate beds, and by directional features preserved in finer-grained sedimentary rocks. The interface between the tow facies lies about 5 miles north of the area of investigation (Troxel, 1967). The best-preserved exposures of the southern facies of the Kingston Peak are in the southern Salt Spring Hills (Troxel, 1967) and the Silurian Hills (Kupfer, 1960; Wright and others, 1974). Only rocks of the southern facies of the Kingston Peak crop out in the Sheep Creek area and elsewhere along the lateral-slip faults.

The exposures of rocks of the Kingston Peak Formation that crop out at Sheep Creek and elsewhere in the lateral-slip fault zones are mainly diamictite composed of isolated clasts of quartzite and coarse-grained granitic rocks supported in a matrix of gray to dark gray medium- to coarse-grained quartzite. In contrast, diamictite in the northern facies of the Kingston Peak Formation contains clasts of earlier Precambrian gneiss, various types of rocks derived from the Crystal Springs Formation, and abundant clasts of Beck Spring Dolomite, which are by far the predominant type of clast. Directional features in the northern facies document transport from a source area to the northeast (Wright and others, 1974) while in the southern facies transport is from the south (Troxel, 1967; Wright and others, 1974).

Diamictite of the southern facies also crops out in the crest of the Avawatz Mountains (Troxel, personal observation). The southern Facies therefore crops out in a belt that overlaps the zone of lateral slip faults and lateral slip cannot be used as a criterion to explain the presence of the southern facies within the lateral-slip fault zone.

Later Precambrian to Cambrian Rocks

The later Precambrian Noonday Dolomite, Johnnie Formation, and Stirling Quartzite and later Precambrian-Cambrian Wood Canyon Formation crop out adjacent to or within the area under study and also are significant to an understanding of the nature of offset along the fault zones. In 1964 Troxel recognized a basin facies of the Noonday Dolomite in the Saddle Peak Hills (in Wright and Troxel, 1966) which was called a clastic wedge. It was later recognized that this facies occurred in its proper stratigraphic position in the Silurian Hills (Wright and others, 1974). Fragments of the basin facies of the Noonday Dolomite have been recognized at various places in the area of study for this report and fit into a predictable stratigraphic pattern. The presence of basin facies of the Noonday Dolomite in the area of the lateral-slip fault zones can be explained without requiring any large lateral slip on either the Death Valley or Garlock fault zones.

The Johnnie Formation, Stirling Quartzite, and Wood Canyon Formation crop out in the southern Salt Spring Hills, Silurian Hills, high in the Avawatz Mountains, and in patches along the lateral-slip fault zones. Sections in the areas outside the fault zone have been measured and described by Diehl (1974), Kupfer (1960), and Troxel (1967). The thicknesses of the various sections fit a regional pattern which, as with the older strata, precludes any significant lateral offset along the Garlock and Death Valley fault zones. Outcrops of Johnnie, Stirling, and Wood Canyon in the lateral-slip fault zones are preserved mainly as pendants in Mesozoic plutonic rocks. However, because they are metamorphosed, they cannot be easily correlated with absolute certainty to specific parts of unmetamorphosed sections exposed nearby.

Paleozoic rocks younger than the Wood Canyon Formation that crop out at various localities in southern Death Valley include the Cambrian Carrara Formation and Bonanza King Dolomite plus younger unidentified metamorphosed Paleozoic carbonate strata. They are not useful at present for measurement of offset along the lateral-slip fault zones because outcrops are widely scattered and regional patterns of facies and changes in thickness have not been documented. Hence they are not described here or discussed further.

MESOZOIC ROCKS

The Mesozoic rocks in the area of study include diorite clearly of Cretaceous age and other intrusive rocks of probably Cretaceous age. The most common intrusive rock in the Avawatz Mountains is diorite. It is distinctive in outcrop as are clasts derived from it. The rock is made up mainly of medium- to coarse-grained hornblende and plagioclase. Quartz is rarely present. It also contains a small proportion of biotite and lessor other minerals. The diorite generally lacks internal planar or linear features except for local areas where internal irregular foliation or flow planes are well developed. The northwestern part of the Avawatz Mountains between Pipe Line Wash and Denning Spring Wash, north of the trace of the Garlock fault zone, is made up mainly of rock that is dioritic in composition but it is extremely well foliated. It is assumed to be of the same age and related in origin to the main mass of the diorite of the Avawatz Mountains even though the two units are separated by the Garlock fault zone. Pegmatite and occasional aplite dikes are common in the diorite.

The age of the diorite in the Avawatz Mountains has been determined by K-Ar methods from one sample collected by Troxel in 1963 in upper Sheep Creek. Biotite from the diorite yielded an age of $1256 \pm 7 \times 10^6$ years B.P., placing its age as early Cretaceous.

The only other masses of diorite in the region are situated on the western flank of the Panamint Range, about 50 miles west-northwest of the Avawatz Mountains, and along the western flank of the Black Mountains about 30 miles north-northwest of the Avawatz Mountains.

Intrusive into the diorite is coarse-grained granitic rock that generally has a pink color due to the pink feldspar phenocrysts it contains as well as faint iron stains. It is herein referred to as red granite.

A third plutonic rock that occurs in the Avawatz Mountains is fine-grained granitic rock that intrudes the diorite and red granite. It characteristically occurs as irregular masses. Some are planar in form and others are circular to semi-circular in plan. Most masses are less than a few tens of feet in maximum width unless they occur as dike-like masses.

Within the bounds of the lateral-slip fault zone are igneous rocks of probably Cretaceous age. One type is coarse-grained pale-colored quartz monzonite. It lies between two belts of Tertiary rocks. The quartz monzonite is medium to coarse grained and is very pale gray to nearly white in color. Elongate pendants of Precambrian gneiss and Later Precambrian Crystal Spring Formation and diabase are preserved.

Fine-grained intrusive rock that is probably granodioritic in composition and Cretaceous in age crops out in the bedrock mass at Sheep Creek Spring. It intrudes the Crystal Spring Formation and Precambrian gneiss and is in contact with pale-colored medium to coarse-grained quartz monzonite, which contains sparse, nearly microscopic ruby-red garnet crystals. The granodiorite and garnetiferous quartz monzonite may be genetically related. Medium- to coarse-grained quartz monzonite of probable Cretaceous age that also contains sparse to rare garnets, crops out in isolated patches in and near the Salt Spring Hills and in an isolated small mass 5 miles west of the Salt Spring Hills, near the base of the fans that extend north from the Avawatz Mountains. The red granite intrudes the diorite and the white granite of the Avawatz Mountains intrudes the red granite. The relative age of the other intrusive rocks is unclear. The garnetiferous quartz monzonite within the fault zone and north of it are correlated on the basis of texture, color, and the presence of garnet. The granodiorite is correlated with the garnetiferous quartz monzonite in the fault zone on the basis of intrusive relations. The pale quartz monzonite between the Tertiary belts is also assumed to be Cretaceous in age. On a regional basis, the various units of the quartz monzonite have a general resemblance to quartz monzonite in the Owlshead Mountains, and to the Teutonia quartz monzonite in the region east of the Avawatz Mountains (Hewett, 1954).

A final unit of Mesozoic (?) intrusive rock is confined to outcrops of the Kingston Peak Formation within the fault zone. It is probably granodioritic in composition and occurs as irregular masses that rarely exceed more than a few inches in width. Most masses are irregular in shape although some of them are planar, especially where they are oriented parallel to bedding planes in the Kingston Peak Formation. The individual masses are not connected by dikes to a source material and are interpreted to be a product of recrystallization of the Kingston Peak Formation. Whatever the origin of the rock, it is limited to the Kingston Peak Formation that lies within the fault zone. Unresolved for the moment is the fact that it occurs in the Kingston Peak Formation, which is essentially unmetamorphosed.

TERTIARY ROCKS

Volcanic Rocks

The oldest Tertiary rocks in the area of study are volcanic. Their age is undetermined but they lie upon or intrude Mesozoic and older rocks and lie beneath the Tertiary sedimentary rocks. Clasts of the volcanic rocks are contained in Tertiary sedimentary rocks.

The most common Tertiary volcanic rock is dark reddish-brown andesite that contains abundant phenocrysts of while feldspar, which form a trachytic texture. In many outcrops the rock has a broken texture suggesting that it was brecciated as it flowed. It may be a lahar type of flow. Locally, it has pale gray to nearly pure white beds of air-fall ash associated with it. The andesite is most abundant along a ridge between Pipe Line Wash and Chaos Wash. It is estimated to average about 50 feet in thickness and occurs in discontinuous elongate outcrops. In upper Sheep Creek it clearly intrudes quartz monzonite but elsewhere it occurs as a flow rock.

Two small patches of rhyolite or rhyo-dacite crop out in one locality east of lower Sheep Creek. The volcanic rocks at this locality lie upon Precambrian gneiss and beneath Tertiary conglomerate. Both patches consist of pale-colored porphyritic rhyolite or rhyodacite.

Sedimentary Rocks

Tertiary sedimentary rocks comprise three belts within the bounds of the lateral-slip fault zones, another succession east of Salt Spring Wash, and a fifth succession south of the Garlock zone. The belts are made up of distinctive rock types or stratigraphic sections and are separated by faults, elongate masses of bedrock, or both. The belts of Tertiary sedimentary rocks are strongly deformed along the fault zones. In places they contain continuous sections of the strata but are incomplete. In other places, however, large segments of the section are cut out or displaced by faults. The thickness of the belts of sedimentary rocks varies along strike and ranges from a few feet to 1,000 feet or more. The partial section of rocks in each belt can be identified as being part of a specific belt.

The ages of the Tertiary rocks have not been determined. They clearly are younger than the volcanic rocks, which they overlie, and are older than Quaternary gravel, which they underlie. The deep maroon color of the lower beds of some of the Tertiary rocks may be an indication of an Oligocene or older Tertiary age, based on similar colors of Oligocene sedimentary rocks elsewhere in the Mojave Desert, but cannot be used as more than an inference of the age. Hence only a Tertiary age is assigned herein. The upper age limit of the Tertiary sedimentary rocks is probably at least late Tertiary, if not early Quaternary as some of the beds are continuous with old deposits of gravel that may be Pleistocene in age. The description of Tertiary and Quaternary age for older gravel deposits is arbitrary, pending further work.

The principal belts of Tertiary rocks are along the fault zones and are designated herein the southern belt, the central belt, and the northern belt. They are described below in that order. Tertiary sedimentary rocks that crop out south of the Garlock fault zone and elsewhere are described separately.

Southern Belt

A succession of coarse gravel to sandstone comprises the southern belt of Tertiary sedimentary rocks. The succession is estimated to be at least 1,000 feet thick. It was not measured nor studied in detail for this report. Neither the base nor the top is exposed. The belt extends from Denning Spring Wash to Chaos Wash.

The lower contact of the succession is fault bounded. The fault gradually and successively cuts out the lower beds from west to east, thus the thickest section with the oldest rocks is in the western part of the belt, west of Pipe Line Wash.

The lowermost beds are deep maroon in color. The color changes gradually; however, in some places, it changes abruptly through dark reddish-brown, to pale red pink and then pale brown to tan and gray. The beds gradually become finer grained stratigraphically upward but locally may coarsen or become finer grained laterally. In general, the beds appear to be a series of fan gravel deposits with less-coarse interfan gravel and sandstone deposits.

The beds strike approximately parallel to the lateral-slip faults and the north front of the Avawatz Mountains. They dip steeply to slightly overturned. The stratigraphic top is to the north as can be readily determined from cross-bedding and graded bedding. Beds range in thickness from a few inches to a few feet. The coarsest beds contain boulders as much as 4 feet in diameter or perhaps larger and the finest beds are composed of siltstone. The coarse beds contain clasts that range in form from angular to well-rounded. The clasts are composed of many rock types, some of which are diagnostic for determining source regions. The clast types do not appear to change significantly either laterally or upwards in the section.

Clasts of diorite appear very near the lowest parts of the section. The diorite is similar in composition and texture to diorite in the Avawatz Mountains which is assumed to be the source area. Clasts of deep-red-stained quartz monzonite, similar in composition and texture to the red granite of the Avawatz Mountains, also are assumed to be derived from there. Coarse-grained carbonate rocks form a modest proportion of the clasts and their source area also appears to be in the Avawatz Mountains, where similar rocks crop out. The presence of clasts of trachytic andesite in the gravel indicates that the andesite once cropped out in the same area as the diorite, red granite, and carbonate rocks and thus is implied to have been deposited in the Avawatz Mountains. An enigmatic type of very dark gneiss forms a modest proportion of the clasts. It has distinctive ptygmatic folds and obvious dark and pale-colored bands. It does not resemble Precambrian gneiss found in southern Death Valley, and therefore is most likely to have come from a source region farther south. It may record the existence at one time of a thrust plate of deep-seated crustal rocks that were emplaced over rocks now exposed in the Avawatz Mountains and were subsequently removed by erosion. Jon Spencer (oral communication, 1979) has reported that similar gneiss crops out at one locality in the Avawatz Mountains.

Some of the clasts in the southern belt are extremely well rounded. They may have more than one cycle of erosion and deposition.

Structural Implications of the Southern Belt. If the clasts of diorite, carbonate rocks, red granite, trachytic andesite, and ptygmatic gneiss were derived from the Avawatz Mountains, then no significant lateral offset has occurred along the faults that lie between the southern belt of Tertiary rocks and the Avawatz Mountains. Of these rock types, the diorite is probably most diagnostic for determining offset because it does not crop out elsewhere except in the southern Panamint Range and central Black Mountains. In both of these localities, other rock types are present that have no apparent counterpart to the clasts of the southern belt. Recrystallized carbonate rocks and red granite clasts could have been derived from the Owlshead Mountains. However, the would not be expected to be mixed with rocks derived from the Avawatz Mountains, which is the case in the southern belt of Tertiary rocks.

Central Belt

The central belt of Tertiary sedimentary rocks is in depositional contact with pre-Tertiary trachytic andesite and in fault contact with the southern belt of Tertiary sedimentary rocks. It forms a belt that extends nearly the full length of the area mapped for this study. In general, it consists of a succession of fine- to medium-grained sedimentary rocks and evaporite deposits. The principal rock types are sandstone, siltstone, and gypsum. It also contains limestone, and monolithologic megabreccia. The beds range in color from maroon, brown, tan, green, and yellow to nearly pure white. Some of the beds may contain volcanic ash.

The oldest rocks are along the north edge of the belt, where they rest in depositional contact upon trachytic andesite or other older rocks. Locally, the contact is disturbed by relatively minor normal- and lateral-slip faults.

The oldest rocks are maroon to brownish-red in color and are composed of thin-bedded siltstone and fine-grained sandstone. The beds average about one-half inch to six inches in thickness and contain abundant cross-stratification markings. Occasional beds composed of pale green very fine-grained material may be reworked air-fall volcanic ash. Within a few tens of feet of the base of the maroon siltstone unit are beds which range in thickness from a few feet to several tens of feet and are many hundreds of feet long. They are composed of angular fragments, block, and coarse grains of a single rock type. These are discontinuous beds of monolithologic megabreccia composed of pale-colored quartz monzonite or Precambrian gneiss. Most of the individual masses are composed on one rock type most commonly of quartz monzonite with no matrix. Some have a minor component of granular trachytic andesite. They occur in at least three stratigraphic positions in maroon siltstone. They occur in at least three stratigraphic positions in maroon siltstone. One mass has deformed siltstone at its leading edge, indicating east or southeast transport. Veins of barite occur in the megabreccia masses, the underlying maroon siltstone, the trachytic andesite and the underlying quartz monzonite. The barite commonly is in the form of open aggregates of bladed white crystals that attain sizes of one or two inches in length. Interstices between barite crystals are filled with very fine-grained, blackish material. Veins are from one inch to several inches wide and some are as long as about twenty feet. Most veins follow fault planes that were formed after the megabreccia masses were deposited.

In a tributary to Amphitheater Wash three sets of hoof prints are preserved in maroon beds which also contain mudcracks, ripple marks, and rare worm burrows. Two sets of prints have cleft hooves. Individual hoof prints are 1-1/2 and 2 inches wide and 6 to 10 inches apart. Another print is about 5 inches wide. Fossil prints were found at a locality farther west in the same stratigraphic position. A further search for hoof prints and other fossil material is anticipated.

Lying above the maroon siltstone is a succession of beds composed of yellow to green siltstone with occasional beds cemented with calcite. The beds also contain limestone concretions that weather out in odd-shaped forms. Most concretions are less than one foot long or 2 inches in diameter. The succeeding unit is made up of beds of gypsum and silty gypsum interbedded with faint maroon to faint green siltstone. These beds lie in fault contact with the southern belt of Tertiary sedimentary rocks and everywhere are highly contorted or highly attenuated.

Nowhere is a complete section of the central belt of Tertiary sedimentary rocks exposed so a stratigraphic thickness cannot be determined. At least half of it, however, is composed of maroon siltstone and the remainder is yellow to green limy siltstone and gypsum-bearing siltstone. The total thickness probably exceeds 1,000 feet but is probably thinner than the southern belt. It crops out nearly continuously from Denning Spring Wash to Salt Spring Wash.

The stratigraphic top of the central belt is to the south. The top is truncated by a branch of the Death Valley fault zone. Cross-bedding, ripple marks, graded bedding, and mud cracks confirm top directions to the south. The strata strike nearly parallel to the belt of outcrops and generally dip steeply to the south. The beds do not appear to vary significantly laterally.

Structural Implications of the Central Belt. The central belt contains clasts of rocks that could have been derived only from the north. All fragments large enough to be recognized by type are of types that occur in the bedrock north of the central belt. The central belt is probably about the same age as the southern belt as they both contain clasts of trachytic andesite and have a similar deep-maroon color in the lowermost units. Each has a different source region for the coarse fragments. The two successions could have been deposited on opposite sides of a common basin and subsequently folded into a syncline. If so, a branch of the Death Valley fault zone extends along the axis of the syncline and little lateral offset has occurred along the fault. In the western part of the area of study however, the southern belt continues to trend westerly and the central belt diverges and trends more northerly. Thus it seems more likely that the two belts have been juxtaposed by lateral slip along a branch of the Death Valley fault zone.

Northern Belt

The northern belt of Tertiary sedimentary rocks crops out in fault contact with the central elongate mass of pre-Tertiary bedrock and in depositional contact with the northern elongate mass of pre-Tertiary bedrock. It forms a nearly continuous belt of outcrops through the mapped area.

In general, it consists of coarse conglomerate that grades upward to siltstone and conglomeratic siltstone; salt and silty salt beds; gypsum and gypsiferous siltstone; and siltstone with celestite, chert, and limestone. It is at least several hundred feet thick but its true stratigraphic thickness cannot be determined because the topmost beds are in fault contact with older rocks and it is complexly folded.

The lowermost beds lie in depositional contact upon parts of the later Precambrian Crystal Spring, Kingston Peak Formation, and Noonday(?) Dolomite. The lowermost beds are conglomerate in places and finer grained sedimentary rocks in other places. Where conglomeratic, the beds fine abruptly way to sandstone and siltstone. The conglomerate appears to have been deposited near the flanks of topographically high local source areas. The conglomerate contains clasts of locally-derived material that are moderately well mixed in terms of variety of rock types. The most diagnostic clasts are diamictite from the southern facies of the Kingston Peak Formation and prominently banded multicolored calcareous mudstone from the basin facies of the Noonday(?) Dolomite. Other clasts are composed mainly of various units of the Crystal Spring Formation. Most of the clasts are angular to sub-angular in form and range in size from pebbles to boulders.

The conglomerate appears to have been deposited in a local water-filled basin because it contains interbeds of gypsum and gypsum-bearing sandstone and siltstone.

The conglomerate unit is overlain by a unit of gypsiferous siltstone and sandstone that includes occasional beds of conglomerate in which are angular to sub-angular clasts of diorite. The presence of diorite clasts implies a source region to the south for the diorite as opposed to a local source region to the north for the clasts lower in the section.

The next younger unit is a poorly preserved section of halite and silty halite. Its stratigraphic position is consistent but in many places it is missing, probably because it has been faulted our or has been squeezed during periods of compression related to faulting and folding. Where present it is marked by a pink to maroon color on the weathered surface, cavernous weathering caused by removal of salt by water, or pinnacle-type erosion of pure halite.

The halite-bearing unit is overlain by an excellent stratigraphic marker unit composed of gypsum and gypsum-bearing siltstone. Gypsum in this unit erodes to locally-prominent outcrops that form a series of small rounded peaks. The gypsum is massive to well-bedded and pale gray to off-white in color. Associated with it are blebs, streaks, and beds rich in manganese oxide. It also contains streaks of red iron oxides. The beds of gypsum and associated manganese and iron oxides are distinct from other gypsum or gypsiferous beds.

The uppermost set of beds is composed of multi-colored siltstone. Within a few tens of feet of its lower contact with the gypsum is a stratigraphic marker bed that contains celestite. The celestite is not continuous and may occur at other stratigraphic positions in this unit. In some outcrops it occurs as beds, occasionally associated with brown-weathering limestone; in others it occurs as nodules as much as two inches across; and in still others it occurs as lenses as much as two inches thick and a few feet long. It is most commonly brownish to tan in color. The remainder of the section is composed mostly of siltstone in beds from half an inch to a few inches thick and which are pink, green, tan, and brown in color. Gypsum is common in the siltstone as stringers and veinlets, most of which cross-cut bedding. In the upper part of the siltstone unit is a bed of limestone that weathers brown. In some places it is a few feet thick but generally is only a few inches thick. It serves as a means to detect faults near the upper part of the siltstone unit.

Structural Implications of the Northern Belt. The rocks of the northern belt diverge at the western edge of the area of study. They crop out farther west at least to the south edge of the Owlshead Mountains and farther northwest to the northeast edge of the Owlshead Mountains (Troxel, personal observations). The rocks appear to have been deposited in an irregular-shaped basin that formerly extended from the northern flank of the Avawatz Mountains into the areas south and northeast of the Owlshead Mountains, much as the topographic lows now exist.

The rocks of the northern belt are probably younger than rocks of the southern and central belts and were juxtaposed by lateral-slip faulting. If so, the amount of lateral slip need not have been more than several miles.

Other Tertiary Rocks

Three other units of Tertiary sedimentary rocks are significant to a discussion of the structural evolution and history of faulting in the area under discussion. One lies south of the Garlock fault. The other two lie north of the northern elongate mass of pre-Tertiary rocks.

The Tertiary rocks south of the Garlock fault consist of poorly-bedded layers of breccia consisting of angular fragments of diorite derived from the Avawatz Mountains and deposited upon the bedrock of diorite. Within the diorite breccia is a prominent cliff-forming breccia layer composed of metamorphosed carbonate rocks of probably late Paleozoic age. It is composed of giant, overlapping blocks that were emplaced as a single sedimentary event. The diorite breccia unit is at least 2,000 feet thick, strikes northwest, and dips southwest. It is truncated to the north at the Garlock fault. Only a fragment of the unit crops out north of the Garlock fault.

The Tertiary sedimentary rocks that lie along the north boundary of the northernmost belt of pre-Tertiary rocks comprise two distinct units. The westernmost unit consists, at the base, of giant blocks of local pre-Tertiary rocks that seem to lie in a jumbled pile very close to the source area — perhaps no more than a few hundred yards from it. They are in their approximate stratigraphic position but are in reverse order as compared to the order in the bedrock units. Above these blocks are intertonging units of monolithologic megabreccia. Each unit is composed of a single rock type and the units are not in stratigraphic order. They consist of masses of breccia of Precambrian gneiss, Kingston Peak diamictite, Mesozoic granitic rock, Later Precambrian carbonate rock and marl, plus other rock types – all common to the nearby Tertiary rocks. From east to west the rocks in this belt gradually change from all bedrock, as at Sheep Creek, to all breccia, as at the mouth of Pipe Line Wash.

The Tertiary rocks east of Sheep Creek Wash lie along the north front of the mountains at Sheep Creek Wash but gradually swing southerly to the Mule Spring fault, where they are in fault contact with diorite. The section is estimated to be a few hundred feet thick. It consists of tan to pink conglomerate, megabreccia, and siltstone. It rests in depositional contact with rhyolitic and older rocks. The stratigraphic top is to the north. The rocks are probably younger than Tertiary rocks in the three belts.

Near Sheep Creek Wash the lower beds of the unit are conglomerate that contain no diorite. Several tens of feet higher in the section the diorite is present as a minor constituent. In the upper beds, diorite is the principal clast type. The lower conglomerate beds give way southeastward to megabreccia beds of granitic rocks.

This unit appears to have been deposited approximately in its present geographic position. The lower beds probably were deposited along the flanks of hills that yielded the clasts contained in them. The upper beds that contain diorite clasts may have attained a sufficient thickness the bury the source for the lower beds.

Structural Implications of the other Tertiary Rocks. The diorite breccia unit probably formed as large sheets that were shed from the west slope of the Avawatz Mountains, perhaps as the mountains were rising along faults along the north flank of the mountains. Subsequently the north edge of the breccia unit was cut by the Garlock fault.

The monolithologic megabreccia sheets on the north flank of the lateral-slip fault zone probably were deposited as the pre-Tertiary rocks south of them were rising along fault zones. The other Tertiary rock unit probably derived most of the diorite clasts from the Avawatz Mountains as it rose along faults between the Tertiary rocks and the Avawatz Mountains.

These three Tertiary units contain locally-derived clasts and they are probably younger than the Tertiary rocks in the three belts. They do not record evidence of significant lateral slip on any of the faults in the area of study.

Tertiary–Quaternary Rocks

Some of the deposits of gravel in the study area were probably deposited in late Tertiary and early Quaternary time. In this report, the deformed gravel that overlay the Tertiary rocks described above and is overlain by relatively undeformed gravel, is considered to be of that general age. In parts of Death Valley farther north, this unit is usually called the Funeral Formation of Pliocene and Pleistocene age.

The gravel is exposed in several places. It is generally unconformable upon older rocks, including some of the Tertiary sedimentary rocks, but locally may be conformable.

The gravel is not described in detail because it lacks definitive features from both older or younger gravel other than its degree of deformation. It is made up mainly of the same type of clasts as the Tertiary gravel, having been partly derived from them and partly from local bedrock sources.

Quaternary Rocks

Many gravel units have been deposited in fans, in stream channels, and as talus along the margins of the Avawatz Mountains. The gravel deposits clearly mark those faults that have not moved since Tertiary time. Landslide masses are also considered to be of Quaternary age as they overlie the Tertiary rocks.

Gravel deposits that are uplifted or down-dropped but are not internally deformed are herein arbitrarily considered to be Quaternary in age. Those that are folded or have steep dips through tilting and rotation are considered to be Tertiary-Quaternary in age.

The gravel deposits are commonly well bedded with individual layers ranging in thickness from a few inches to about two feet. Some are graded and many have cross bedding. Imbrication is common. All are poorly sorted. Clasts are in grain-to-grain contact in some beds and in others they are isolated in coarse sand or fine gravel. Many clasts exceed one foot in diameter but most clasts are in the range of one inch to 4 inches in diameter. Most are angular to sub-angular.

The internal features indicate that most of the gravel was deposited during periods of extremely high runoff in the form of debris flows, mud flows, or flood flows, much as they are deposited at present.

The relative ages of the gravel units can be determined on the basis of superposition, freshness of fan surfaces versus antiquity, color, degree of development of desert varnish or desert pavement, degree of disintegration of clasts on the surfaces of fans, development of soil profiles, and degree of dissection by superposed drainage systems. Gravel deposits can be correlated by comparison of clast content and many of the criteria listed in the preceding paragraph.

STRUCTURAL FEATURES

GENERAL FEATURES

The major faults are identified on the map (in pocket) by name. They are the Garlock, Mule Spring, and branches of the Death Valley fault zone. In addition to these, normal and reverse faults are abundant in most areas. The general sequence of faulting in: pre-Cenozoic thrust(?) faulting; Tertiary lateral-slip, normal, and reverse faulting; and Quaternary lateral-slip, normal, and reverse faulting. Some normal faulting may have occurred during late Precambrian times.

Most of the folds are of limited extent, being most common in gypsiferous and silty Tertiary units adjacent to lateral=slip faults. One large fold occurs in the eastern part of the area of study.

Garlock Fault

The Garlock fault appears to die out in the northwestern part of the Avawatz Mountains near Pipe Line Wash. In the area of study it is expressed as a south-facing scarp south of which are grabens that are mostly in young gravel.

Scarps on the edges of the grabens are from one to several feet high. The grabens are oriented slightly more southwesterly than the main trace of the Garlock fault. The grabens appear to be tensional features south of the trace of the Garlock fault that were formed as a result of left-lateral slip along the main trace. In the vicinity of the grabens are small stream channels offset a few feet in a left-lateral direction. The scarps of the grabens are in young gravel that is next older than modern stream gravel. The faults may have cut the modern stream gravel, but is so, the scarps have been destroyed by erosion.

Mule Spring Fault

The Mule Spring fault was named by Noble and shown on a regional map compiled by Noble and Wright (1954). The fault trace extends across the length of the area of study and for several miles farther west along the north side of a ridge that extends west from the northwestern part of the Avawatz Mountains. The ridge is bounded on the south by the Garlock fault zone. The Mule Spring fault is oriented slightly more northerly than the west-trending Garlock fault, which it truncates.

The Mule Spring fault is a fundamental fault. It is continuous across the main front of the Avawatz Mountains and forms a continuous northern boundary for the diorite mass in the Avawatz Mountains. The diorite and mixed rocks associated with the diorite are pulverized and ground into gouge in a zone that ranges in thickness from a few feet to many tens of feet along the south side of the fault trace. Tertiary sedimentary rocks, by contrast, are much less disturbed at the fault zone. The fault dips steeply southward along most of its trace but flattens gradually from the vicinity of Sheep Creek to 33ºSW near the southeastern end of its trace. As it flattens it also swings to a southeastern strike. Slickensides in several exposures of the fault plane are oriented parallel to the dip, indicting near-vertical movement. Juxtaposition of Tertiary rocks that generally dip steeply northward on the north side of the fault imply reverse movement along the fault plane. Because the Tertiary rocks along the fault contain diorite, there could not have been any significant lateral movement along the Mule Spring fault. Thus movement along the Mule Spring fault is concluded to be reverse with little or no component of lateral movement. As an estimate, the amount of reverse movement is assumed to be at least equal to and probably greater than the vertical distance equal to the difference between the base and crest of the Avawatz Mountains — about 4,000 feet. The Muse Spring fault cuts branches of the Death Valley fault zone.

Death Valley Fault Zone

The Death Valley fault zone, as shown by Noble and Wright (154), consists of several branches in a zone about 2 miles wide that extends several miles north-northwestward from the north flank of the Avawatz Mountains. The branches are well exposed in the group of low hills that extend north-northwestward into southern Death Valley from the Avawatz Mountains (Troxel, unpublished data). In these hills and in the floor of Death Valley, the branches of the fault cut very young gravel and have components of right-lateral slip (Troxel, 1970). Furthermore, the fault traces record a secular migration of movement from branch to branch. The youngest movement is at the northern end of individual branches and on the easternmost branches (Troxel, 1970). As the fault branches extend southeastward along the north face of the Avawatz Mountains, they merge with the Mule Spring fault to a flatter southwesterly dip.

The Death Valley fault zone consists of three branches, two of which extend through bedrock in the area of study and a third that lies beneath Quaternary rocks. The two exposed branches nowhere cut rocks younger than the Tertiary sedimentary rocks. The northernmost trace, however, is marked by series of low-angle reverse faults that cut Quaternary gravel.

It appears that the branches of the Death Valley fault zone in the study area have a secular pattern similar to traces farther north. The southern branches are the oldest and the northeastern branch is the youngest. The southern branches have a component of right lateral slip (discussed below) possibly several miles in extent and are deformed (also discussed below). The northernmost trace probably had modest right lateral slip but it most recent movement is reverse (up on the south). The northernmost segment of this fault has abundant scarps in the gravel, which is the most recent activity of the fault zone.

The two older branches of the Death Valley fault zone possible have had as much as several miles of lateral slip along them. If so, the movement has occurred since the Tertiary sedimentary rocks were deposited as the southern branch juxtaposes two sections of Tertiary rocks. The Tertiary rocks south of the fault contain diorite clasts and other rocks that crop out in the Avawatz Mountains. The Tertiary rocks immediately north of the south branch of the Death Valley fault zone do not contain clasts of diorite.

The next branch of the Death Valley fault zone lies north of the northern mass of pre-Tertiary rocks. The lower part of the Tertiary section contains clasts derived mainly from the pre-Tertiary rocks farther north and must have been transported from that direction. The upper part of the Tertiary section, however, contains clasts of diorite in graded conglomerate beds interbedded with siltstone. Thus it appears that the Avawatz Mountains yielded debris into the Tertiary basin while it was near the Avawatz Mountains. If so, then the Tertiary rocks farther south were not moved into their present position by right-lateral faulting but, because they contain no diorite clasts, were probably shielded by local highlands from receiving sediments from the Avawatz Mountains.

Other Faults

A gently south-dipping fault plane is exposed in a small tributary to Sheep Creek a few hundred feet downstream from Sheep Creek Spring. The fault is underlain by diamictite of the Kingston Peak Formation and by diabase that is older than the Kingston Peak Formation. Vertical beds of the Crystal Spring Formation lie above the fault. Farther north, the Crystal Spring Formation overlies the younger Kingston Peak Formation along a south-dipping reverse fault. The fault can be traced westward to Chaos Wash. East of the mouth of Sheep Creek is another reverse fault that displaces Precambrian gneiss over Crystal Spring and other later Precambrian rocks. In small patches of rocks exposed in canyons west of Chaos Wash are beds of Crystal Spring overlying in fault contact the Kingston Peak diamictite and marly limestone of the Noonday Dolomite. Several thrust planes are exposed at this locality, one of which is intruded by Mesozoic granodiorite.

All of these are thrust faults and may be Mesozoic in age but the fault intruded by granodiorite is clearly older than Tertiary and probably pre-Cretaceous in age. Some may be Tertiary in age and related to the lateral-slip faults. The westward continuation of this belt of pre-Tertiary rocks may contain further evidence of these older thrust faults but exposures are poor at best and Tertiary tectonic and sedimentary events superposed on these rocks preclude an obvious interpretation.

The entire map area abounds in faults with minor offsets and various trends. Many follow bedding planes, especially in Tertiary sedimentary rocks, many are tangential to bedding planes and occur in an en echelon pattern, and other sets of faults trend more obliquely to bedding. Most pre-Tertiary rocks are shattered or cut by closely-spaced faults. Some of the major north-trending stream channels may follow faults. Most of the minor faults are not shown on the map because they cannot be portrayed accurately on it. Minor faults that cut Quaternary deposits were plotted wherever they were observed.

Summary and Significance of Tectonic Events

The oldest tectonic episode is one of thrust faulting followed by intrusion of plutonic rocks during Mesozoic time, probably during the Cretaceous Period. The faults may be thrust faults that moved during the Sevier or other orogenies. They may also be responsible for the anomalous west strike and vertical dip of the later Precambrian strata at Sheep Creek and elsewhere in the local region, and perhaps in the Silurian Hills to the east.

The Mule Spring fault is marked by the most profound zone of gouge in the study area. It dips steeply to the south along its western exposure and gradually flattens along its trace to the east. Slickensides in the fault plane are oriented down-dip and their presence, together with the presence of clasts of diorite and other rock types common in the Avawatz Mountains, in Tertiary rocks north of the fault imply that the principal movement is reverse. Total movement is perhaps on the order of about 4,000 feet, the present elevation difference between the base and the crest of the Avawatz Mountains.

The southeastern segment of the Mule Spring fault displaces diorite over Quaternary gravel on a plane that dips about 45ºSW. Elsewhere the fault is overlapped by Quaternary gravel that has not been cut by the fault.

The Death Valley fault zone consists of three branches. The southern branch juxtaposes two successions of Tertiary sedimentary rocks of different provenance. The Tertiary rocks south of the fault contain clasts that are typical of bedrock in the Avawatz Mountains while those north of the fault apparently do not. Thus movement along the fault branch is probably lateral and may have been several miles, probably in a right-lateral sense. No record of Quaternary movement was found along it.

The central branch of the Death Valley fault zone lies north of the southern branch and separates pre-Tertiary rocks to the south from Tertiary rocks to the north. The upper part of the Tertiary section contains conglomerate beds in which diorite is a common constituent. The lower part of this Tertiary belt contains clasts derived from bedrock exposed at the north base of the section. The most reasonable explanation for the presence of diorite clasts is that they were transported as debris flows several miles northward from the Avawatz Mountains into a local playa basin. The section was subsequently moved towards the source area by right-lateral movement along the central branch of the fault zone.

The north branch of the Death Valley fault zone trends southeastward beneath Quaternary gravel along the north edge of the Avawatz Mountains. The gravel deposits are cut in several places by reverse faults that dip gently towards the Avawatz Mountains. These fault traces form an irregular zone of discontinuous faults that gradually merge with the Mule Spring fault near the southeastern edge of the study area at Mormon Spring. The faults in this area are steepest near the contact with diorite and gradually flatten farther from the mountain front. They appear to be overturned segments of the Death Valley fault zone upon which reverse movement has occurred in Quaternary time.

The clearly identifiable Garlock fault zone extends from the west barely into the southwestern part of the area of study. It consists of a south-facing main fault scarp along which movement has been down to the south. Associated with the main scarp are several minor scarps that trend more southwesterly than does the main scarp. The minor scarps define small grabens that are tangential to the main scarp and which are compatible with left-lateral movement along the main scarp. These scarps appear to mark the east end of this branch of the Garlock fault zone. The grabens are very young as they cut the gravel that is next older than gravel in the modern stream channel.

Many other faults occur throughout the map area. Some branch from the main faults but many do not. A few of them offset Quaternary gravel. They probably are related to lateral-slip on the main faults and to later extension of the area.

CONCLUSIONS

The principal faults in the north front of the Avawatz Mountains are the Mule Spring fault and branches of the Death Valley fault zone. The Mule Spring fault is a reverse fault along which the Avawatz Mountains have been uplifted. The Death Valley fault zone has had right lateral displacement that probably does not exceed a few miles. The Garlock fault dies out near the Mule Spring fault in a series of small grabens. The southern branches of the Death Valley fault zone are truncated by the Mule Spring fault. The northern branch appears to die out southeastward in a series of reverse faults that dip towards the Avawatz Mountains.

The most recent fault movement is along reverse faults associated with the Death Valley fault zone, and small normal faults associated with the Garlock fault. Further movement on any of them is likely.

Three successions of Tertiary sedimentary rocks are juxtaposed between branches of the Death Valley fault zone. Further detailed study of them should provide more precise information about the amount of displacement along individual faults.

Older Quaternary deposits are deformed and cut by faults. Younger Quaternary deposits overlie the principal faults but are cut in many places by other faults.

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Fort Irwin Archaeology: A Preserved Past for the Mojave Desert

Dr. Mark W. Allen, *Cultural Resource Specialist, DPW-Environmental, Fort Irwin/The National Training Center, CA 92310*

The Fort Irwin Cultural Resources Program has stewardship of the archaeological and historical resources of over 1,000 square miles of the Western Mojave Desert. Ft. Irwin's mission as the National Training Center brings large armored and mechanized units to the installation for training ten times a year. The Cultural Resources Program supports this mission by helping the installation meet its cultural resource requirements under Federal legislation and Army regulations. This paper first presents the current status of the program. Second, it summarizes the contributions of archaeology at Ft. Irwin, and briefly summarizes the cultural resources of the installation for each major period of culture history in the Mojave region.

Analysis of Rock Varnish from the Whipple Mountains, Southeastern California using Time-of-flight Secondary Ion Mass Spectrometry

R. C. Anderson [1], **P. A. Zimmerman** [2], and **K. Beratan** [3]
[1] *Jet Propulsion Laboratory, California Institute of Technology, 4800 Oak Grove Drive, Pasadena, CA 91109* randerson@jpl.nasa.gov, [2] *Department of Chemistry, University of Pittsburgh, Pittsburgh, PA 15260,* [3] *Department of Geology and Planetary Sciences, University of Pittsburgh, Pittsburgh, PA 15260.*

Rock varnish, also known as desert varnish, is a natural coating which forms on rock surfaces during exposure in semi-arid and arid regions [1]. Varnishes typically are amorphous translucent films dominated by clay minerals, iron and manganese oxides, silica and aluminum [2]. Although both the chemical properties and field relations of rock varnish have been extensively investigated, much controversy still exists regarding why and how rock varnish forms. Suggested mechanisms for rock varnish formation include microbial activity [3], accumulation of airborne dust [4], and water leaching and transport [5]; all of these processes may play an important role.

Conventional mass spectrometers are not always useful for rock varnish analysis because of the non-volatile nature of the material. Secondary Ion Mass Spectrometry (SIMS) allows the production of secondary ions from nonvolatile compounds and elements on a surface using a focused primary ion beam [6]. The secondary ions are extracted using a high potential field and passed through an ion mirror which compensates for the spread of kinetic energies. The ions are separated by flight time (TOF), with lighter ions reaching the detector first. TOF-SIMS gives complete elemental and molecular information as well as information regarding the spatial distribution of components on the surface.

Preliminary analyses of two rock varnish coated mylonites from the Whipple Mountains of southeastern California illustrates the utility of TOF-SIMS for chemically characterizing rock varnish surfaces. High resolution spectra were obtained which show the presence of and relative quantity of element species on the surface. This technique is particularly useful for studying rock varnish since the spectra acquired are from the top nanolayers of the sample and do not penetrate to the underlying host rock. Identification of the spatial distribution of key elements provides clues to the origin of the varnish. Using TOF-SIMS, three different elemental distribution patterns have been recognized from the two Whipple Mountain samples: 1) uniform distribution of elements across the surface; 2) non-uniform and apparently random distribution of elements across the surface; and 3) non-uniform patterns in which the highest concentration is associated either with the microtopographic highs or lows on the surface. Preliminary analysis suggests that the rock varnish formed as material form outside of the host rock adhered to the rock surface. The likely source of the material is windblown dust form one or more of the local playa lakes.

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Microgeographic Differences in Rainfall Affect the Nutrition and Survivorship of Desert Tortoises (*Gopherus agassizii*) in the Mojave Desert.

Harold W. Avery, *U.S. Geological Survey, Biological Resources Division, Canyon Crest Field Station, Riverside CA 92521*, **Philip A. Medica**, *U.S. Geological Survey, Biological Resources Division, Las Vegas Field Station, Las Vegas NV 89126*, and **Jeffey E. Lovich**, *U.S. Geological Survey, Biological Resources Division, Canyon Crest Field Station, Riverside, CA 92521.*

Annual plants are an important food resource for many desert herbivores, including the federally threatened desert tortoise (*Gopherus agassizii*). Rainfall is an important determinant of variation in annual plant biomass in desert ecosystems, and geographic, elevational and seasonal patterns of rainfall may affect populations of desert tortoises by influencing availability of food and water. Changes in food and water can in turn affect growth, reproduction and survivorship in these and other desert animals. Within Ivanpah Valley, California, rainfall may vary two- to threefold from lower elevations to higher elevations within a single rainfall event, regardless of season. We quantitatively analyzed the diets of tortoises in sections receiving significantly different late summer rainfall in Ivanpah Valley in 1994, by analyzing contents of fresh tortoise scat collected in early October. Tortoises inhabiting the higher elevation received greater amounts of summer rainfall and fed primarily on green perennial grass (*Hilaria rigida*), whereas tortoises at lower elevation that were exposed to significantly less rainfall fed primarily on dry cactus (*Opuntia* sp.). This paucity of water and food at lower elevation is the likely cause of a recent die-off of tortoises, whereas tortoises at higher elevation within Ivanpah Valley have not experienced recent high mortality. Microgeographic differences in rainfall caused by physiographic features of elevation, mountain ranges and/or other unknown factors, may affect the

TABLE 1. SPECIES DIFFERENTIATION, SUMMARIZED FROM VANDENBUGH (1985).

C. ventralis	*C. draconoides*
Large (±>70 mm)	Small (<u>1</u> 55 mm)
Snout short and rounded	Snout longer and less rounded
Supralabials prominent and very convex in lateral outline	Supralabials much less prominent and convex
Males with 2 larege oblique black blotches on each side	Males with 2 smaller, almost vertical black blotches, followed by a small black spot

nutrition and survivorship of tortoises. Such dramatic microgeographic variation in survivorship of tortoises has important implications to wildlife conservation and management, yet present monitoring programs have unreplicated study sites within a small geographic area that are insufficient to detect such variation.

A Systematic Review of the North American Iguanid Genus *Callisaurus* (Reptilia: Sauria): The Status of *Callisaurus draconoides* and *Callisaurus ventralis*

Benjamin H. Banta, *Southwestern Biological Associates, 9847 Willow Lane, Escondido, CA 92029*

The iguanid genus *Callisaurus* occurs in desert areas in the southwestern United States and western Mexico. Prior to 1932, concepts of speciation within the group varied considerably. From 1932 to present time all nominal forms have been recognized generally as belonging to a single species complex, *Callisaurus draconoides*. The examination of several thousand preserved and living specimens in the field and laboratory has convinced me that four well-defined noninsular species groups as well as a number of insular and geographic subentities are diagnosible. What was described as *Callisaurus crinitus* By Edward Drinker Cope in 1896, I consider to belong to the genus *Uma*.

The genus *Callisaurus* Blainville (1835) was based on material obtained in what is now called the Cape Region, Baja California Sur, Mexico. Hallowell (1852) described *Homalosaurus ventralis* from "New Mexico west of the Rio Grande." Smith and Taylor (1950) restricted the type locality to Tucson, Arizona. Baird (1859) considered *Homalosaurus* a synonym of *Callisaurus* and this decision has been supported since.

Cope (1875) placed the then three known nominal entities (*dracontoides* [sic], *gabbi* MS, and *ventralis*) in one species. Yarrow (1882:50-51) followed Cope but recognized *draconoides* and *ventralis* as subspecies of one another. Thus, he recognized two subspecies: "*Callisaurus dracontoides dracontoides* (Blainv.) Cope" listing 32 preserved specimens in the United States National Museum from La Paz obtained by Lyman Belding in 1882, and "*Callisaurus dracontoides ventralis* (Blainv.) Cope" based upon 21 preserved specimens primarily from "Arizona and California."

Stejneger (1893), in a discussion of *ventralis* and *dracontoides*, stated that "the differences are numerous and are found both in structure and coloration. Moreover, after an examination of about 200 specimens, I can affirm that the characters are constant and that the two forms do not intergrade." Not only do morphological criteria apply towards the recognition of these two entities but also zoogeographic according by Stejneger (1893:171):

> . . . *C. draconoides* is restricted to the very southern extremity of the Lower California peninsula . . . that is, to the zoogeographic district which has been termed the Cape Region. . . . *C. ventralis* ranges over a comparatively large area comprising, so far as known, the northern portion of Lower California; the coast of Sonora, Mexico, at least as far south as Guaymas; the desert region of southern California; southern Arizona as far east as Camp Apache ad Fort Buchanan, at least; southern and western Nevada as far north as Pyramid Lake; southern Utah where it is restricted to the Lower Santa Clara Valley.

VanDenburgh (1895:98) supported the position of Stejneger (1893), and presented descriptive information to differentiate these forms (Table 1). VanDenburgh (1895:98) remarked further that "no intergradation of the two forms has yet been shown, but two young females from San Ignacio, and one from Santa Margarita Island are more nearly like *C. draconoides* than are any other specimens of *C. ventralis*, suggesting, but not showing, an instability of character further to the south." The San Ignacio form is *C. plasticus* and the Santa Margarita Island specimen represents a distinct insular species which is probably derived from a *C. draconoides* type ancestor.

Cope's (1900) posthumous tome did not include the decisions rendered by Stejneger (1895) and Van Denburgh (1895) and did not even include the two forms described by Cope in 1896, one of which is not only the most distinct member of the genus, but which shows such extreme similarities to the genus *Uma*.

Dickerson (1917) recognized the unique characteristics of *C. crinitus* Cope (1896) and the basic differences between *draconoides* and *ventralis*. VanDenburgh (1922) supported Dickerson and recognized five species of *Callisaurus*.

Linsdale (1932:259) grouped all nominal forms into one species complex and stated,

> . . . the available facts seem to indicate that the group of forms of *Callisaurus* on the peninsula of Lower California and in Western United States constitutes one species and this species would take the name *draconoides*, which was the first bestowed in the group. The range of characters separating any two of the subspecies in not large but in each case there is good evidence of intergradation where the ranges meet.

Tevis (1944:11) first attempted to more precisely delineate the occurrence of nominal *Callisaurus* in Baja California. In doing so, he followed Linsdale (1932) "in recognizing in Lower California only one species of *Callisaurus* divided into four subspecies: *gabbi* in the north; *draconoides* in the Cape region; *crinitus* in the sand dunes of the Vizcaino Desert, and *carmenensis* in the central coast excluding the Vizcaino Desert sand dunes." Many recent authors have tended to follow Tevis.

Abundant evidence now suggests the recognition of four well-defined species:

1) *C. draconoides* – Cape Region
2) *C. plasticus* – much of eastern Baja California, exclusive of the Cape Region and northern portions (Gulf of California hydrographic basin)
3) *C. ventralis* – southwestern United States and adjacent Sonora and northeastern peninsula Baja California
4) *C. inusitatus* – most of western Sonora extending southward into Sinaloa, Mexico.

Uma crinitus is segregated from *Calllisaurus* occurring in the Vizcaino Desert of central western Baja California.

Most of the taxonomically distinct populations of the *Callisaurus* occur within the boundaries of the Republic of Mexico. A summary of the results of my study will be discussed.

Petrology of Early Proterozoic Granitoids from the Southwestern United States: Implication for Genesis and Tectonics of the Mojave Crustal Province

E. Erik Bender, *Dept. of Geology, Orange Coast College, 2701 Fairview Rd., Costa Mesa, CA 92626*
ebender@lib.occ.cccd.edu

The Early Proterozoic geologic evolution of the eastern Mojave Desert region as defined by the characteristics of its supracrustal rocks, granitoids, metamorphism, structural history and isotopic signature, contrasts sharply with other Proterozoic provinces of the southwestern United States. The oldest supracrustals exposed contain zircons that are over 2.0 Ga, which is evidence for a crust older than elsewhere in the southwestern United States. Granitoids were widely intruded into these supracrustal rocks between 1.76 to 1.64 Ga. The earliest plutons and surrounding supracrustals were metamorphosed to high grade at about 1705 Ma during the Ivanpah Orogeny. Subsequent granitoids, emplaced from 1.69 to 1.67 Ga, were volumetrically extensive along a northeast-trending belt in the middle of the province. Younger plutons were emplaced at about 1.66 Ga in several localities, and later at 1.64 Ga along the extreme southern portion of the province. Based on these and other observations, it is interpreted that these developed in a back-arc setting relative to the main axis of the 1.7 Ga orogen that resided in areas to the east. Whole rock elemental and isotopic data (Rb-Sr, U-Pb, Sm-Nd) indicate that the granitoids of the Mojave crustal province were the product of partial melting (approximately 20%) of amphibolitic lower crustal material.

Tectonic Control on Development of Old Alluvial Surfaces in the Whipple Mountains, Southeastern California

Kathi Beratan, *321 EH, Dept. of Geology and Planetary Science, Univ. of Pittsburgh, Pittsburgh, PA 15260*
kathib@vms.cis.pitt.edu

Alluvial fans in the Whipple detachment terrane mantle major topographic elements that were created during extreme (>100%) crustal extension during Miocene detachment faulting. The last episode of motion on the Whipple detachment fault occurred at ~13 Ma; subsequent uplift driven by crustal adjustment to extension resulted in upwarping of the lower plate, creating the broad domes of the metamorphic core complexes. Almost no Quaternary tectonism has been reported from the region, although there may have been some regional uplift after 8-5 Ma. Development of alluvial surfaces can be generalized into three stages. Stage 1 includes young alluvial-fan deposits with active channels and depositional fan surfaces. Stage 2 includes intermediate-age deposits with well-preserved, inactive surfaces that are incised by modern channels. These surfaces commonly display well-developed desert pavements and rock varnish. Stage 3 includes old alluvial-fan deposits that have been highly dissected, leaving virtually no trace of the original fan surface.

A preliminary map showing the distribution of heavily eroded stage 3 surfaces in the study area has made from analysis of Landsat Thematic Mapper imagery. A striking feature of this distribution pattern is the termination of these old surfaces along the trend of Miocene transfer faults. These faults played an important role during detachment faulting, accommodating differential tilting between adjacent upper plate fault blocks. Motion on the transfer faults was driven by activity on high-angle normal faults that moved in response to motion on the underlying Whipple detachment fault. Gravity-driven flow of ductile lower crust from relatively unextended areas to highly-extended areas is thought to have driven upwarping of the domal lower plate "cores" after ~13 Ma. I hypothesize that this upwarping caused uplift of overlying upper plate blocks, reactivation of the transfer faults, and abandonment of alluvial surfaces. Upwarping of the core complex had ceased prior to deposition of the Bouse Formation at 5-8 Ma. This implies that the abandonment event which was responsible for development of the oldest surfaces in the alluvial apron occurred prior to 5-8 Ma, during the late Miocene, rather than in the Quaternary as previously thought.

The Culture of the People of Black Canyon

Beth McWaters Bjorkman, *P.O. Box 1278, Twin Peaks, CA 92391*

The goal of this presentation is to review anthropologically the overall evidence in the form of artifacts, ecofacts, and features within Black Canyon to show a picture of the types of activities that went on in the canyon and to present the data as to reveal what daily life was like for people who inhabited the area.

This will be achieved by utilizing the disciplines of physical and cultural anthropology by referencing existing data and field work, as well as theorems proposed from scholars. All of the existing evidence needs to be gathered and approached holistically to create a clearer perspective of the cultural belief system. This may answer questions about their rock art.

It is generally accepted that Black Canyon is located in Shoshonian territory, possibly populated by roving bands of the Shoshonean divisions of the Chemehuevi and Serra-Vanyume. These people may have gathered at Black Canyon for the purpose of trade, ceremonialism, or an outpost to replenish supplies before crossing the desert – perhaps for all these reasons. I will review briefly an outline of the Shoshonian cultural beliefs and iconography and their material goods used in various tasks related to daily life. Where applicable, mention of other Mojave Desert sites will be addressed.

Black Canyon is well known for rock art portraying aspects of their cultural belief system and ceremonialism. This will be compared in conjunction with known iconographical evidence since interpretive problems are not new to the issue of rock art. Preservation of the petroglyphs, an issue worthy of mention, is an ongoing problem in the canyon.

With a condensed view of the archaeological evidence and consolidated cultural understanding of the mentioned Shoshonean divisions as well as previous occupation by other groups, activities that may have taken place in the canyon will emerge. I will demonstrate this in a "living history" rendition within the slide presentation. A visual representation showing, for example, how the stone tools were manufactured, or the techniques used to create the petroglyphs, will aid in the attempt to visualize the people of Black Canyon.

Desert Survival in the Nineties

Brian Brown, *Shoshone Museum Association, PO Box 217, Shoshone, CA 92384*

Living and working in the Mojave Desert can be vastly different than visiting on winter weekends for recreation or research. Economic survival in the desert has never been easy, so what happens to modern-day desert towns when the primary industry (mining) pulls out? Must they dry up and die like the boom towns of the last century?

This is a look at how one community is adapting to the new realities of the Mojave. By shifting its focus and treating the environment of the Mojave Desert as the primary resource, towns in the Death Valley area are surviving the exodus of mining, and even providing a little growth.

Changing attitudes and new public interest in California's largest desert has created a new set of customers with different values for those attempting to make a living there. But can this new market of educators, ecotourists and Europeans replace large industry? Stay tuned . . .

Does the Greater Roadrunner Hibernate?

James W. Cornett, *Palm Springs Desert Museum, 101 Museum Drive, Palm Springs, CA 92262*

The Greater Roadrunner (*Geococcyx californianus*) is an uncommon ground-dwelling cuckoo of arid and semi-arid regions of the southwestern United States and northern Mexico (Peterson, 1990). Small animals including insects, reptiles, birds and rodents comprise the bulk of its diet though some plant material is also consumed (Bent, 1940). It is considered to be a non-migratory, resident species (Garrett and Dunn, 1981; Small, 1994).

My own personal observations and conversations with amateur ornithologists over the past twenty-five years suggest the greater roadrunner is observed less frequently during winter than at other seasons. The abundance of anecdotal information on this species' winter rarity has prompted an investigation into its seasonal habits. Two methods of analysis may explain the apparent disappearance of the roadrunner during winter: systematic seasonal fieldwork and the radio tracking of individual roadrunners.

Seasonal observations are being compiled to determine if the anecdotal information is accurate. An evaluation of preliminary field data reinforces that roadrunners are less frequently observed during winter. This could be explained by declining populations due to fall mortality of very old and young individuals. Local migrations to lower elevations by some individuals may be another explanation. The increased foraging of adult birds in late winter, spring and early summer, necessitated by parental feeding of nestlings, could also result in a relative abundance of observations during these times of year.

Another hypothesis for the paucity of winter observations is the possibility that some roadrunners may enter hibernation. This species has the ability to enter torpor on cool nights and conserves energy by doing so. Individual birds studied by Ohmart and Lasiewski (1971) maintained a body temperature of approximately 94º F. when resting in the dark as compared with 101º when active during daylight hours.

Hibernation is known to be practiced by at least one bird species. The late Edmund Jaeger (1948) discovered that the common poor-will (*Phalaenoptilus nuttallii*) has the ability to survive a radically lowered body temperature for more than three months in winter.

In order to obtain definitive information on winter activity, radio telemetry devices will be affixed to three roadrunners in the fall of 1998. The data obtained from these birds will provide information on winter movement patterns and whether the roadrunner, like the poor-will, can enter hibernation.

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Status of the Northern and Eastern Colorado Desert Coordinated Management Plan

Richard E. Crowe, *Bureau of Land Management, California Desert District, 6221 Box Springs Blvd., Riverside, CA 92507-0714*

One of three ecosystem plans in progress that address the recovery of the desert tortoise in the California Desert, the Northern & Eastern Colorado Desert Coordinated Management Plan (Plan) focuses on the Northern and Eastern Colorado Desert Recovery Units and a small portion of the Joshua Tree Recovery Unit. The planning area, 5.5 million acres in size and lying in the Sonoran Desert Ecoregion, is bounded by I-40 (north), the Colorado River (east), the Imperial Sand Dunes and Coachella Canal (south), and the West Mojave Plan (west). The planning area does not have urbanization pressures which characterize other parts of the California Desert. The major cooperating agencies are the Bureau of Land Management (lead agency), Joshua Tree National Park, U.S. Marine Corps Air Station in Yuma for the Chocolate Mountains Gunnery Range, U.S. Fish & Wildlife Service, and California Department of Fish & Game (which has provided the lead wildlife biologist). Additional cooperators to the Plan include other Federal, state, and local agencies as well as many interest groups. Plan scope is ecosystem comprehensive. Plan decisions will amend existing land use plans of the cooperating Federal agencies for the tortoise and other species and habitats and may be of use by other agencies and companies with interests in the planning area.

Work on the plan in the last year was focused on completing the collection of data and the mapping of plant communities. The latter included an accuracy assessment, collection of additional habitat data for species-habitat modeling, and correcting the map with the accuracy information. From the accuracy assessment exercise we found plant communities to be accurate to varying degrees with an overall accuracy of about 50%. Plant communities and overall map accuracy after corrections were made is greatly improved. All spatial data has been GIS digitized to aid in modeling, analyses, and Plan development.

Species-habitat-ecological processes models are currently being run. They will help define and map areas of relative biodiversity importance. The California Wildlife-Habitat Relationship System (WHR), developed by the California Resources Agency, is the modeling technique employed. It has been tailored to the particular species and habitats found in the planning area. Modeling results will be used in analyzing use-biodiversity conflicts and aid Plan development. Model maps will be completed prior to April, 1998. The basis for modeling is plant communities and includes natural and artificial water sources, species accounts and known occurrence of wildlife and plant species of concern, habitat data collected during the accuracy assessment noted above, detailed information on the desert tortoise and bighorn sheep, a characterization of the important ecological processes, and various physical features. Physical features data available include elevation, slope, aspect, landforms, and lithology. Criteria for assigning relative biodiversity value to habitats have been developed for species uniqueness, rarity, or range limits; ecological processes; habitat fragmentation; species richness/diversity; and exotics. A different modeling protocol was developed to predict the occurrence of rare plants. It is based on a limited set of considerations: elevation range, plant community, landform and distance from known occurrences of the same species.

Major scheduled milestones for remaining work on the Plan are as follows:

- June, 1998 – Complete value and conflict analyses; develop Plan and decisions
- November, 1998 – Issue draft Plan/EIS; 90-day public review
- June, 1999 – Issue Proposed Plan/EIS; 30-day public review
- August, 1999 – Cooperating agencies sign Record of Decision

The Plan lead, Dick Crowe, may be contacted for further information at the above address and by calling (909) 697-5216.

Dogs of Robbins Quarry

Tom Howe and **Robert E. Reynolds**, *Earth Sciences Section, San Bernardino County Museum, Redlands, CA 92374*

The radiation of dogs during the Miocene Barstovian Land Mammal Age suggests that they were able to utilize diverse food sources and habitats. Robbins Quarry, located sub-adjacent to the Hemicyon Tuff (14.0 ± 0.1 Ma) contains a varied carnivore taxa including *Amphicyon, Aelurodon*, and *Tomarctus*.

The dogs from Robbins Quarry are distinguished from each other by their size. *Amphicyon* is comparable to a bear in bulk, although its long limbs suggest that it was a swift runner. *Aelurodon* is the size of a small wolf, and *Tomarctus* is the size of a coyote. *Tomarctus* is the North American ancestral borophigine or "bone-crushing" dog. Both the "bone-crushing" *Epicyon* and the hypercarnivore *Aelurodon* are derived from *Tomarctus*. Abundant fossils of *Tomarctus* at Robbins Quarry provide baseline morphometric data for developmental study at that point in time.

Carnivore Tracks in the Miocene Horse Spring Formation, Lake Mead Region, Nevada

Michele M. Kissell and **Stephen M. Rowland**, *Dept. of Geosciences, University of Nevada, Las Vegas, NV 89154-4010*

Two types of carnivore tracks have been found in the Thumb Member of the Miocene Horse Spring Formation in the Lake Mead region in southern Nevada. A probable felid track is known from a natural cast preserved in fine-grained sandstone. There are what appear to be partially retracted claw imprints at the tips of the toes. The second type of track is most likely canid in origin. Sixteen tracks with distinct claw marks are preserved on a slab of fine-grained sandstone that is also covered with numerous bird tracks. Three distinct canid trackways are present, two to three steps long.

These carnivore tracks are part of a track assemblage from the Thumb Member that also includes two types of camel tracks as well as five types of bird tracks. Sedimentological data suggest a shallow, perennial lake paleoenvironment in a fault-bounded basin.

Vulnerability and Recoverability of the Mojave Desert Ecosystem

Leonard J. Gaydos, *U.S. Geological Survey, Ames Research Center, 242-4, Moffett Field, CA 94035-1000*

The Mojave Desert Ecosystem, an area of 125,000 km^2 covering parts of four States, is a land affected by the growing presence of some 40 million people who live within a day's drive. Land managers are challenged with devising policies that allow human use of the desert while also protecting its vulnerable resources. To assist them, the U.S. Geological Survey is conducting an interdisciplinary vulnerability and recoverability study of the entire Mojave Desert Ecosystem. Scientists from the USGS with expertise in geology, hydrology, ecology, botany, geomorphology, geography, and information sciences are studying the issue.

Vulnerability to natural and human-induced disturbances and recoverability of land from these disturbances differs from site to site. While 50-year old tank tracks still scar the landscape in some locations, the most obvious traces of 100-year old former mining towns have disappeared in others. By studying disturbed sites and measuring their rates of recovery compared to control sites, scientists can predict the potential for disturbance of similar sites. They are examining evidence of ecosystem disturbance as measured by soil compaction, wind erosion, water erosion, disruption of overland water flow, and species distribution and density. They are also estimating vulnerability to disturbances are being estimated for specific sites by using a geographic information system containing mapped data describing soil particle size, vegetation cover and cover types, age of geomorphic surface, soil chemistry, ground cover, rainfall frequency and intensity, and topographic roughness. Using remote sensing, the scientists are extrapolating these estimates to larger areas and constructing spatial models that predict vulnerability to and recoverability from disturbance. Working with land managers, the USGS will design a decision support system for planning the appropriate use of resources under its care as well and a monitoring system for assessing long-term results.

Petroglyphs of Renegade Canyon

Robert F. Hilburn, *Mojave River Valley Museum, Barstow, CA 92311*

Petroglyphs are rock art that is pecked, carved, abraded, or scratched into a rock surface. Petroglyphs appear in Australia and Hawaii as well as in the desert of the southwestern United States. Pictographs are painted in color on the rock surface. They appear as cave paintings in France and Baja California. Intaglios or geoglyphs are primarily recessed ground figures produced by scraping away rocky surfaces to show the underlying soil. Such figures occur in England, South America and in the southwest United States just north of Needles, California and outside Barstow, California. This presentation will focus on petroglyphs of Renegade Canyon.

Geographic Variation and Environmental Determinants of Reproductive Output in the Desert Tortoise

Jeff Lovich, *U.S. Geological Survey, Department of Biology, University of California, Riverside, CA 92521-0427*, **Phil Medica**, *U.S. Geological Survey, Las Vegas Field Station, 4765 W. Vegas Drive, Las Vegas, NV 89126*, and **Hal Avery**, *U.S. Geological Survey, Department of Biology, University of California, Riverside, CA 92521-0427*

Three study sites were established in California in the spring of 1997 as part of a multi-year study to examine variation in reproductive output of desert tortoises: Joshua Tree National Park (JTREE), Mojave National Preserve (MOJAVE), and an area of land leased by the Bureau of Land Management (BLM) for wind energy production near Palm Springs (MESA). At MESA, 9 out of 10 monitored females produced a total of 72 eggs. Of these females, 6 produced second clutches and one produced a third clutch. Clutch size ranged from 2-8 eggs. Modal clutch size was 4 eggs. Mean clutch sizes were 4.33 and 5.00 eggs for first and second clutches, respectively. First clutch size was positively correlated with carapace length (R^2=0.29) but not significantly (P=0.14). When all clutches were considered, a significantly positive relationship was detected (R^2=0.26, P=0.04) The earliest date of egg laying occurred between April 18-23. The last clutch was oviposited sometime after July 3. At JTREE, only 1 of 8 females produced a single clutch (5 eggs). At MOJAVE, 12 of 18 monitored tortoises produced 43 eggs in 12 single clutches with no subsequent clutches. Clutch size ranged from 1-7 eggs with modal clutch size of 3-4 eggs. Mean clutch size was 3.58 eggs. Larger females produced larger clutches and the relationship between the two variables was almost significant (R^2=0.32, P=0.055). Most of the variation observed appears to be related to differences among sites in rainfall and associated production of annual food plants.

A Closer Look at *Mammuthus hayi* Barbour (1915), the Long-jawed Mammoth

George E. McDaniel, Jr., and **George T. Jefferson**,
Anza-Borrego Desert State Park, Borrego Springs, CA 92004

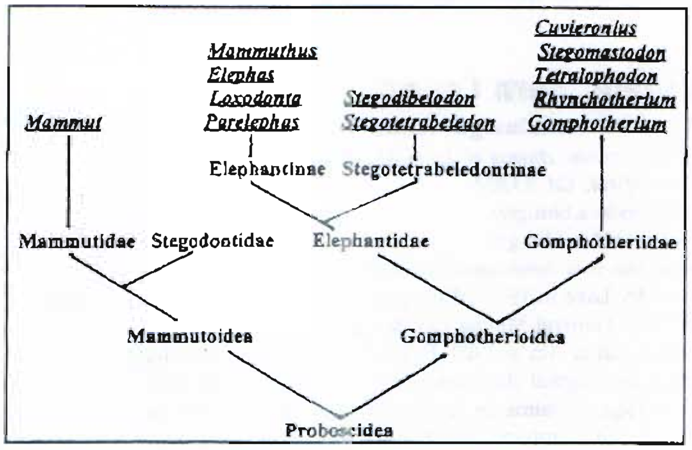
In 1915, Barbour described a new elephantid species, *Mammuthus hayi*, recovered from Aftonian Interglacial (*ca.* 2 Ma) gravels in Hulbert's Sand Pit, Crete, Nebraska. Fragments of an upper tusk were found, along with badly fractured and damaged mandible and M/3s.

> The chief distinguishing characters of *Elephas hayi* are: unusual length of mandible; the last molar small, narrow, and anterior to the coronoid; transverse ridges 10 to 11; angle distinct and sharp posteriorly; coronoids uncommonly prominent, deeply pitted, and set very obliquely (Barbour, 1915).

Most workers agree that *Mammuthus meridionalis* is the earliest (1.8 Ma) species of mammoth in North America (Maglio, 1973; Kurten and Anderson, 1980; Agenbroad, 1984; Graham, 1984; Churcher, 1986). However, Madden (1981) considers *M. meridionalis* an invalid taxon, and refers all primitive North American mammoths, including those previously assigned to *M. meridionalis* by other authors, to *M. hayi*. A review of the type, University of Nebraska State Museum (UNSM) 1315, and close examination of specimens identified as (University of Florida [UF] 86979, UF 67200) (Webb and Dudley, 1995), or referable to *M. hayi* (such as Texas Memorial Museum [TMM] 31055-1,2,3, TMM 43341-1) reveals that these specimens exhibit primitive characters typical of the oldest African Elephantidae.

Mammuthus hayi has a gomphothere-like mandible. The morphology is quite different from that of *Elephas*, *Loxodonta*, or other species of *Mammuthus*. The corpus of the dentary is much longer and thinner than that in *M. meridionalis*. The coronoid processes are prominent, set low, and are posterior to the posterior margin of the M/3s. These processes in all other Elephantinae are high. The ascending ramus of *Mammuthus meridionalis* is inclined posteriorly at an angle of approximately 100° from the ventral plane of the corpus. This angle in *M. hayi* is much more obtuse, the ascending ramus is inclined posteriorly, and is wider than high. The ventral/posterior margin of the dentary and the ascending ramus in *M. hayi* follows a smooth, broad convex arc.

Early in the evolution of the Elephantinae, the dentary became shorter and the molars longer. Available space in the dentary soon was limited to a single tooth (Maglio, 1973). Although the type of *Mammuthus hayi*, UNSM 1315, apparently has only the M/3s in wear (see Barbour, 1915 Figure 3 and Plate 1), UF 86976 (illustrated by Webb and Dudley, 1995 Figure 3-B) and TMM 31055-1,2,3 have both the M/2 and M/3 in wear concurrently. This is not an Elephantinae character, but occurs in the sister subfamily Stegotetrabelodontinae.



The molars in *Mammuthus hayi* resemble those of the Elephantidae, but are intermediate in some characters between those of stegotetrabelodonts and other *Mammuthus* species. They are narrower, and lower crowned than comparable teeth in other species of *Mammuthus*. The M/3s are smaller than those of *M. meridionalis*. The tusks in UF 86979 and UF 67200 are similar to those in the gomphothere *Stegomastodon*, slightly curved, but much shorter than those of *Mammuthus*.

The anterior mandibular symphyseal process of *Mammuthus hayi* (as seen in TMM 31055, this feature in UNSM 1315 appears to have been reconstructed) extends anteriorly, and is much more massive than that of other *Mammuthus* species. This process in *M. meridionalis* extends anteriorly, but is much thinner than in *M. hayi*. In the *M. imperator* - *M. columbi* lineage, the process is downturned at approximately 45°. The condition in *M. hayi* is expected in groups like the stegomastodonts and stegotetrabelodonts. Here the symphyseal process is much more developed in order to support mandibular tusks, and the anterior mandibular symphyseal process remained more massive even after the lower tusks were lost in later genera of stegotetrabelodonts.

Although Madden (1981) claims that *Mammuthus hayi* is the most primitive mammoth in North America, because it has very low coronoid processes and mammoth-like molars, a comparison of the type with *M. subplanifrons*, a very early species of African mammoth (*ca.* 4 Ma), reveals that the coronoids of the latter species are much higher, and the mandible appreciably shorter than in *M. hayi*. It is highly improbable that *M. hayi* is derived from *M. subplanifrons*, or any of the early Elephantinae.

In the Stegotetrabelodontinae, the mandible is more elongate than in the Elephantinae, and has a massive anterior mandibular symphyseal process, which generally supports tusks. *Stegodibelodon*, within the Stegotetrabelodontinae (Coppens, 1972), has no mandibular tusks, but has a protruding mandibular symphseal process (Kalb and Mebrate, 1993). The mandibular morphology of *Mammuthus hayi* is comparable to that in *Stegodibelodon* (*ca.* 3 Ma). Although, the stegotetrabelodonts were the dominant proboscidean group during much of the African Pliocene, there is no evidence that they ranged through Asia or into North America.

In all aspects of the mandible, Barbour's *Mammuthus hayi*, and other specimens referable to this taxon (such as TMM 31055-1,2,3; TMM 43341-1; UF 86864; UF 86979; UF 67200) are more primitive than all but the earliest Elephantinae. Therefore, we argue for removal of *Mammuthus hayi* from the Elephantinae. Specimens presently identified as *M. hayi* should be re-examined, and assigned to an appropriate taxon possibly within the subfamily Stegotetrabelodontinae.

Mammuthus meridionalis, as presently recognized (Maglio, 1973; Kurten and Anderson, 1980; Agenbroad, 1984; Graham, 1984; Churcher, 1986), remains the earliest North American mammoth.

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Defining the Hemphillian-Blancan Land Mammal Age Boundary in the Temecula Arkose, Riverside County, California

William L. Rader, *Dept. of Geology, Pasadena City College, Pasadena CA 91106*

The Temecula Arkose lies within the Temecula and Aguanga basins of the Temecula portion of the Elsinore fault zone of southwestern Riverside County. The formation consists of over 600 feet of white to buff to greenish arkose, brown soil horizons, brown silts, white silicified algal marls, and white tuff. These continental sediments were originally deposited by southwest-flowing streams on a low-relief, down-faulted block of metamorphic and granitic basement complex of Mesozoic age. Vertical movement along northwest-striking, high-angle normal faults in the middle Pleistocene produced tilting of the basement blocks. Arkosic sediments have been consequently removed by erosion from uplifted blocks and preserved on downdropped blocks.

Diagnostic bones and teeth of rabbits and rodents recovered from soil horizons in the Temecula Arkose include indicator taxa for the Blancan and preceding Hemphillian land mammal ages and have yielded an overall Pliocene (Blancan LMA) age for the formation (2.0 to 4.8 Ma). Due to the existence of both Blancan and Hemphillian taxa, the boundary between the two land mammal ages, dated at 4.8 Ma, may be present within the section. Work conducted by Golz, Jefferson, and Kennedy indicates that an abundance of Blancan taxa occur within the upper Temecula Arkose. Fossils recovered low in the section by the San Bernardino County Museum in the Radec and Shamrock areas show a majority of Hemphillian indicator taxa and fewer taxa of Blancan age. Recently sampled localities contain *Cupidinimus bidahociensis*, known from the Hemphillian L.M.A., and *Sigmodon* sp. These taxa reinforce that the localities are probably earliest Blancan in age.

The purpose of the author's work is to attempt to find a lower and earlier portion of the section in which Hemphillian indicator taxa are found exclusively. Structural cross-sections help determine that the NW-dipping beds become older as one moves south and east. Paleosol exploration and matrix collection downsection from Lancaster Valley to Aguanga, together with consequent matrix reduction, sediment sorting, and fossil identification at the San Bernardino County Museum, may yield rabbit and rodent taxa that are solely Hemphillian in age and which may help define the Hemphillian-Blancan Land Mammal Age boundary west of the San Andreas Fault in southern California.

Flamingo Egg From The Miocene Sediments of The Calico Mountains, San Bernardino County, California

Robert E. Reynolds, *Earth Sciences Division, San Bernardino County Museum, 2024 Orange Tree Lane, Redlands, CA 92374*

The interior mold of a duck-size egg was found on the south slope of the Calico Mountains by the William Grubb party. Remains of a small *Merychippus* sp. recovered up-drainage suggest that local fossils are of Miocene Hemingfordian Land Mammal Age (18-16 Ma). Approximately 60 percent of the circumference of the egg is present and has been filled with bedded siltstone. This may suggest that the egg was broken prior to hatching and, subsequently, filled with sediments. Except for size, the morphology of the egg strongly matches that of the common flamingo, *Phoenicopterus ruber*. The Calico specimen is 8 percent smaller in two dimensions than the lesser flamingo, *P. minor*. As described by W.A.S. Sarjeant and R.E. Reynolds, Hemingfordian Land Mammal Age bird tracks, possibly made by flamingos, occur as three ichnomorphs and range from large (length = 110 mm, width = 125 mm) to small (length = 72 mm, width = 87 mm). Extinct flamingos, *Phoenicopterus minutus* and *P. copei* were found in the late Pleistocene Rancho Labrean Age sediments of Manix Lake. A flamingo relative *Ajaia ajaja*, the roseate spoonbill, is known from the Rancho La Brea Tarpits. Presently, the common flamingo is known from southern Florida and the spoonbill from Florida and the south Texas coast.

Wild Cats of the Barstow Formation

Robert E. Reynolds and **Mary Aruta**, *Earth Sciences Section, San Bernardino County Museum, Redlands, CA 92374*

A spectacular carnivore radiation during the Miocene Barstovian Land Mammal Age can be seen in the Barstow Formation in the Mud Hills. Bone-crushing dogs such as *Tomarctus* gave rise to the wolf-size hypercarnivore *Aelurodon* and the hypocarnivore *Epicyon*. Long-legged bear-size dogs, *Amphicyon*, were the largest carnivores of the time. Robbins Quarry, located sub-adjacent to the 14.0 Ma Hemicyon Tuff, has produced a size array of cats that shows diversity similar to that of the dogs. A large cat, *Pseudaelurus sinclairi* is the size of a mountain lion and has round canines. A small cat is inferred from cat-like humeri suggesting an animal the size of a bobcat. An intermediate-sized cat, probably a Pseudaelurine, had a short scimitar-like flattened canine with lingual and labial troughs and posterior denticles. The mandible does not have a flange but has a pocket in the diastema area to receive the upper canine. The lower canine has a flat lingual surface. Measurements of the rostrum and dentition indicate that the face was relatively shorter than *P. sinclairi*.

Saltdale, Kern County, California

Larry Vredenburgh, **Alan Hensher**, and **Gregg Wilkerson**, *Bureau of Land Management, 3801 Pegasus Dr., Bakersfield, CA 93301*
gwilkers@ca.blm.gov
lvredenb@ca.blm.gov

Saltdale was first developed by the Diamond Salt Company on the Koehn Dry Lake in 1911. Mining claims were located under provisions of the 1872 General Mining Law for gold prior to enactment of the Mineral Leasing Act of 1920. Following a common (but illegal) practice, the original claimants used "dummy" locators to block up large acreage of claims on Koehn Dry Lake prior to their sale to Diamond Salt Company. The Consolidated Salt Company purchased the mining claims and began shipping salt in 1914 from a 4-story mill at a production rate of 90,000 tons per year. In 1919 a competitor, Fremont Salt Company, began work on the east side of the lake, a practice which continued to 1927. In 1920 the Federal Mineral Leasing Act removed salt from location under the 1872 Mining Law. After 1919, pumping of salt brine and evaporation in large ponds replaced mining of the ancient lakebed salt. In 1928, the Long Beach Salt Company consolidated operations and produced a moderate supply of salt through the Depression.

An investigation into the legality of the original "dummy" locations was conducted by the Bureau of Land Management (BLM) in the 1970s. A contest over improper locations made prior to 1920 was never issued because mining, by the 1970s, had moved from claims located between 1900 and 1920 to claims located in the 1930s for supposed gold values in clay underlying the salt beds. As a result of a complaint received by a rival salt distributor, Kern Salt Company of Bakersfield, California, BLM performed an investigation of the 193's-era claims and found that salt mining was occurring on claims located for gold after the passage of the Mineral Leasing Act. A validity examination of the claims lead to them being declared null and void by reason of lack discovery of a valuable gold deposit in 1973. No development at Saltdale has occurred from BLM lands since that time. Up to 1981 there was some salt production at Saltdale from private lands.

Rifted Volcano in Wingate Wash, Death Valley Region, Southeastern California

David L. Wagner, *California Department of Conservation, Division of Mines and Geology, Sacramento, CA 95814*

Wingate Wash is a northeast-southwest trending topographic trough that separates the Panamint Range from the Owlshead Mountains in the Death Valley region of southeastern California (Figure 1). Famous as an escape route from Death Valley used by 49ers, and later as the route for the 20-mule team borax wagons, the area is now seldom visited because it is nearly surrounded restricted military land (Photo 1). Recently it was designated wilderness, making access even more difficult. The geology of Wingate Wash is poorly known; geologic mapping has only been regional reconnaissance (Wagner and Hsu, 1988) or is unpublished. The northeast trend of Wingate Wash is anomalous to the north-northwest trend typical of Great Basin intermontane basins and its straightness suggests the trough is fault-controlled. A zone of faults, parallel to the trend of Wingate Wash was mapped by Wagner and Hsu (1988), was interpreted to be surface exposures of the Wingate Wash fault observed on seismic profiles by deVoogd and others (1986). On the north side of Wingate Wash there is an eroded volcanic center interpreted by (Wagner, 1988) to be the remains of a volcano rifted by movement along the Wingate Wash fault during a deformational episode 12 to 14 million years ago. This paper presents an update of this interpretation and gives data constraining the time of deformation.

The Bodie Bowl Mining Claim Validity Examination

Gregg Wilkerson, *Bureau of Land Management, 3801 Pegasus Dr., Bakersfield, CA 93301*
gwilkers@ca.blm.gov

The Bodie Protection Act of 1994 (Public Law 103-433) required the Bureau of Land Management (BLM) to conduct an investigation of 332 lode mining claims, 27 placer claims and 2 mill site claims in the Bodie mining district, Mono County, California. This was the most extensive evaluation of gold mining claims in the history of the U.S. Department of Interior. The BLM evaluation team consisted of 25 persons from BLM, U.S. Bureau of Mines and U.S. Geological Survey. These professionals also worked closely with personnel from Mono County Planning Department and California Department of Parks and Recreation (CDPR) during the investigation. Examination of company, academic and government data and use of the TechBase ore body modeling program lead to an identification of minable gold resouces on private lands and seven unpatented mining claims in the Bodie Bowl. While preliminary estimates of ore occurrences and distribution were being developed by BLM, the private lands, mineral leases and unpatented mining claims in the Bodie mining district were purchased by CDPR and the American Land Conservancy (ALC).

The mineral deposit at Bodie is of the Comstock Epithermal Vein type. The erosional level at Bodie exposes the upper and middle sections of the mineralizing system which is related to boiling of ore-bearing fluids. This unique geology will become available for academic study. When the Bodie Project was purchased by CDPR and ALC, the data collected by various mining companies over the past 30 years (at a cost of over $30,000,000) also came into the public domain. The data collection includes 86,481 feet of exploration drilling data, thousands of feet of underground mapping and sampling, and dozens of geologic and engineering reports made since 1890. The data collection includes 25,000 assay points. Of these, 17,677 composite assays were used to develop a mining plan for Bodie Bluff-Standard Hill. Much of this data is in digital form. The data collection also has 15,000 feet of remnant core and assay pulps from the exploration projects conducted since 1965. This data is now in need of a permanent home. BLM and CDPR seek partners in academia and industry to support the creation of a curation and research facility for the Bodie data collection.

Arid Unsaturated Zones as Groundwater Contamination Barriers for Radioactive Waste Disposal

H.G. Wilshire, *Public Employees for Environmental Responsibility, 2001 S Street, NW, Washington, DC 20009*

Evidence for rapid transmission of radioactive contaminants into groundwater through the overlying unsaturated zone in arid areas (e.g., Nativ, 1991; Gee et al., 1994; Nativ et al., 1995; Striegl et al., 1996; Fabryka-Martin et al., 1996; Conaway et al., 1997) contrasts with evidence of little or no downward movement of precipitation in natural settings. This evidence includes lack of change in moisture content below depths of a few meters and upward water-potential gradients (Fischer, 1992); upward decrease in water vapor density in response to thermal gradients, and decrease in CO2 vapor pressure, interpreted to indicate upward vapor movement from the water table (Prudic and Striegl, 1994); concentration of chloride in the upper 10 m, interpreted to indicate no downward movement of precipitation below ~10 m for 16-33 ka (Andraski and Prudic, 1997); and stable isotope distribution of deuterium and oxygen-18, interpreted as consistent with upward movement of moisture throughout the unsaturated zone (Prudic et al., 1997). In some places, evidence for both downward movement and the long-term absence of downward movement of moisture is found in the same sample sites. For example, elevated tritium (3H), inferred to have moved laterally and downward by saturated flow (Striegl et al., 1996) occurs within the zone of high chloride concentration (Prudic, 1994) near the Beatty, Nevada low-level radioactive waste (LLRW) site.

This conflicting evidence appears to be in part the result of limited sampling in geologically complex environments, lack of understanding of the geologic history of study sites, lack of site-specific data, and limited understanding of the mechanisms of moisture movement in the unsaturated zone. Thus, at the Beatty, Nevada site, among the best-studied sites, interpretations of water-content and water-potential variability are based on measurements made at only three closely spaced locations, and the water-potential data were selectively chosen on the assumption of site homogeneity; interpretation of the chloride data as indicating no downward movement of precipitation below ~10 m for 16-33 ka assumes stability of the Amargosa River floodplain on which the sample sites (four) are located for that period of time. This assumption is not supported by recent flood history of the Amargosa R. (Anderson et al., 1997) and is further contradicted by presence of 3H within the chloride zone. Prudic et al. (1997) interpret dD values in groundwater and the unsaturated zone at the Beatty site as consistent with upward movement of moisture from the water table. This conclusion is derived by postulating that the dD of the groundwater differs from that of modern recharge on the basis of only 6 sampled precipitation events in 18 months at a locality 35 km distant; these isotope data are better interpreted on the basis of long-term measurements of deuterium in regional precipitation, which support modern recharge.

Apparent contradictions among data sets can be reconciled if pathways for contaminant movement are distributed on scales both larger and smaller than sampling intervals. Such preferred pathways, including fractures in both consolidated and unconsolidated materials, are being called on with increasing frequency to explain the fact of deep unsaturated zone contamination in contrast to model predictions (e.g., Striegl et al., 1996; Fabryka-Martin et al., 1996; Conaway et al., 1997). Attempts have been made by proponents of controversial projects, such as the proposed Ward Valley, CA LLRW site, to vitiate evidence of rapid leakage of radionuclides from similar arid sites (Beatty, NV) by asserting that contamination of groundwater is the result of sabotage, or that the problems were caused solely by disposal of liquid wastes, which for unexplained reasons migrated rapidly through the unsaturated zone whereas much larger volumes of natural precipitation are unable to migrate deeper than a few meters. In view of the fact that rapid migration of contaminants in the unsaturated zone has occurred in the absence of liquid waste disposal [e.g., Richland, WA (Kearney, 1987) and Yucca Mountain (Fabryka-Martin et al., 1996) sites], attention is more appropriately focused on the general lack of understanding of rates and mechanisms of contaminant transport in the unsaturated zone.

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The Early Pliocene Caribbean Bryozoan *Acanthodesia savarti monilifera* in the Imperial Formation near Whitewater, Riverside County, California

Edward C. Wilson, *71624 Biskra Road, Rancho Mirage, CA 92270* and **Roger J. Cuffey**, *Dept. of Geosciences, Pennsylvania State University, University Park, PA 16802*

Abundant microscopic fragments of the Caribbean bryozoan *Acanthodesia savarti monilifera* Canu and Bassler, 1919 make up a substantial portion of the Burrobend Member of the Imperial Formation at a locality (Los Angeles County Natural History Museum, Invertebrate Paleontology locality 15386) in an unnamed canyon northeast of the community of Whitewater, Riverside County, California. The bryozoan is a new phylum for the fauna here and a great geographic range extension for the subspecies. It is otherwise known only from the Dominican Republic in Bowden Formation equivalents (Lower Pliocene, Campbell *et al.*, 1975).

The Imperial Formation has been considered to be about 6.5 Ma (Winker and Kidwell, 1996) in this area and represents the northerly ancestral Gulf of California although marine waters had extended into the present central gulf area at Isla Tiburon by 13 Ma (Smith, 1991). Absolute dates were based on associated volcanic units. Biostratigraphic age estimates range from Late Miocene to Early Pliocene for the Imperial Formation near Whitewater (diatoms, nannoplankton, forams, mollusks, whales).

The bryozoan adds further evidence for an Early Pliocene age.

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Unburned Fuels in an Arizona Upland Saguaro-Shrub Community

Ruth C. Wilson[1], **Marcia G. Narog**[2] and **Bonni M. Corcoran**[2]. [1]*Biology Dept., California State University, San Bernardino, CA 92407-2397;* [2]*USDA Forest Service, Pacific Southwest Research Station, Forest Fire Laboratory, Riverside, CA 92507*
rwilson@wiley.csusb.edu
mnarog@rfl.psw.fs.fed.us
bcorcor@rfl.psw.fs.fed.us

Introduction. Wildfires in Arizona's upland saguaro-shrub communities have become larger and more frequent over the past 25 years. During this time, about 30 percent of the saquaro resource on the Mesa Ranger District, Tonto National Forest, Arizona, has been burned. As a result, populations of fire-intolerant giant saguaro cactus (*Carnegiea gigantea)* are being reduced in number. Past research focused on saguaro life history and physiology, but not on community dynamics. Although the significance of fire as a problem in saguaro habitat has been documented (Rogers 1985, 1986; Schmid and Rogers 1988; Thomas 1991), few studies describe the unburned saguaro-shrub vegetation as fuel for fire.

In January 1994, we initiated a study of burned and unburned saguaro-shrub vegetation southwest of Four Peaks on the Mesa Ranger District. We reported over 90 percent injury and 19 percent mortality for saguaro immediately after fire. Long-term mortality may be significantly higher. Nearly 90 percent of the trees and shrubs growing near the bases of the saguaros resprouted after fire. Associated plants may act as nurse plants for saguaro establishment and may also supply fuel during fire. This paper describes the unburned fuels to identify patterns in vegetation distribution potentially important for fire management in this saguaro-shrub community.

TABKE 1. SPECIES ASSOCIATED WITH *Carnegiea gigantea*

Description	Species
small woody trees (300-350 cm)	*Canotia holocantha* *Cercidium microphyllum*
large woody shrubs (125-225 cm)	*Acacia constricta* *A. greggii* *Larrea tridentata* *Lycium* spp.
medium woody shrubs (50-80 cm)	*Ambrosia deltoidea* *Calliandra eriophylla* *Encelia* spp. *Ephedra* spp. *Ericameria* spp. *Eriogonum* spp. *Krameria grayi* *Psilostophe cooperi* *Senna covesii* *Thamnosoma montana*
soft shrubs with woody bases (40-70 cm)	*Baileya multiradiata* *Palafoxia linearis* *Sphaeralcea ambigus*
fibrous woody (137 cm)	*Yucca torreyi*
succulent cacti	*Echinocereus engelmannii* *Mammallaria tetrancistra* *Opuntia engelmannii* *O. leptocaulis* *O. versicolor*
low herbacous cover	grasses and non-woody forbs

Methods. For our study, we collected plant dimension data (height and two diameters) on four 1-ha. unburned sites, using twenty 100-m line transects. In addition, eight points on each of two 350-m point-quarter transects were similarly sampled. These data were used to calculate plant volumes and describe distribution patterns to use for assessing fuels and fire potential in this saguaro-shrub community (Wilson et al., 1996.

Results. Species associated with saguaro cactus were measured and sorted into physiognomic groups based on average height and relative woodiness. Species identified by physiognomic group are listed in Table 1.

Shrubs clustered at the bases of saguaros formed patches of potential fuel. Trees and large shrubs make up about 11 percent of the total plant density and covered more than 55 percent of the area sampled. Medium shrubs were the most common physiognomic group observed. They represented about 63 percent of the total number of individuals and comprised 289 percent of the total plant cover. Tree cover was at least twice that of the other physiognomic groups. Bare ground, dead and downed fuels, and forbs and grasses covered about 58 percent of the total length of the line transects sampled. Trees, shrubs, yuccas, and cacti covered the remaining 42 percent of the line transects. Within this 42 percent of the total length of the line, vegetative cover of the major physiognomic groups was 125 percent. The 25 percent overlap in vegetative cover reflects the dense cover of the living plant fuel clusters that characterize this saguaro-shrub community.

Conclusions. Small trees, woody shrubs, yucca, and cacti form the major physiognomic groups associated with the giant saguaro cactus in this area. Although trees and large shrubs provide greater cover, medium shrubs occurred more often. Associated species often cluster near saguaros forming patches of both dense and diverse vegetation. Bare patches, separating heavy fuels, are bridged by herbaceous material, especially following wet years. This forms a potentially lethal fuel bed for the fire-intolerant giant saguaro cactus. Further experimental study is required to correlate vegetation patterns with saguaro mortality after fire. Land managers should consider this complex saguaro-shrub fuel patchiness in future fire management decisions. Results from this study may help resource managers understand fire management possibilities and problems for this giant saguaro resource.

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San Bernardino County Museum Association
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Format: Two clear copies of the manuscript must be submitted to the Editorial Board with a letter of transmittal requesting that the manuscript be considered for publication and that it is not presently under consideration elsewhere. Manuscripts should be typewritten, double-spaced, on one side only of 8.5 x 11" paper. Ample margins should be allowed for editing comments. The first page should contain the title and author(s) name, address, and telephone number. The author's last name and page number should appear at the top of each following page. Include COPIES of figures, tables, and photographs. Do not send original photographs or figures with your initial submission.

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Attachments: If photographs are used, originals or equivalents will be required. Photographs should be black-and-white or color prints, of good quality and contrast. Figures and drawings should be in India ink or equivalent on white paper or PMTs. Captions should be submitted on separate pages, double-spaced, and referenced to their accompanying figures. Photographs should be marked lightly in pencil on the back border with the author's name and figure number. Authors are encouraged to submit accepted manuscripts on DOS-compatible disks in addition to paper copy. Please contact the editor regarding electronic submission of manuscripts and figures.

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Address queries and manuscripts to:
The Editor
San Bernardino County Museum Association
2024 Orange Tree Lane, Redlands CA 92374
or query by E-mail
jreynolds@ccr.co.san-bernardino.ca.us

Plates in Pocket

Figure 2 in Jachens and Calzia. Map showing magnetic anomalies over the Garlock fault and vicinity (after Roberts and Jachens, 1993). Faults (in red) from Jennings (1994). Aeromagnetic data from a number of separate surveys were numerically filtered and merged to produce a coherent map of magnetic anomalies, reduced-to-pole, at approximately 300 m above the ground surface. Small "+" symbols denote locations of inferred edges of magnetic rock bodies. Alphanumeric symbols denote magnetic features discussed in the text. Profile B-B' was modeled to estimate depth extent of fault. Contour interval 50 nanoTesla.

Figure 5 in Jachens and Calzia

5a. Map showing part of magnetic anomalies given in figure 3, palinspastically restored according to model with offset of 47 km uniform east of Sierra Nevada. Contour interval 50 nanoTesla.

5b. Map showing part of magnetic anomalies given in figure 3, palinspastically restored according to model with offset of 47 km at point ***d*** decreasing linearly eastward to 0 km at east end of Garlock fault. Contour interval 50 nanoTesla.

Plate 1 in **Troxel and Butler**. Geologic map of the North Margin of the Avawatz Mountains, Southern Death Valley, California.

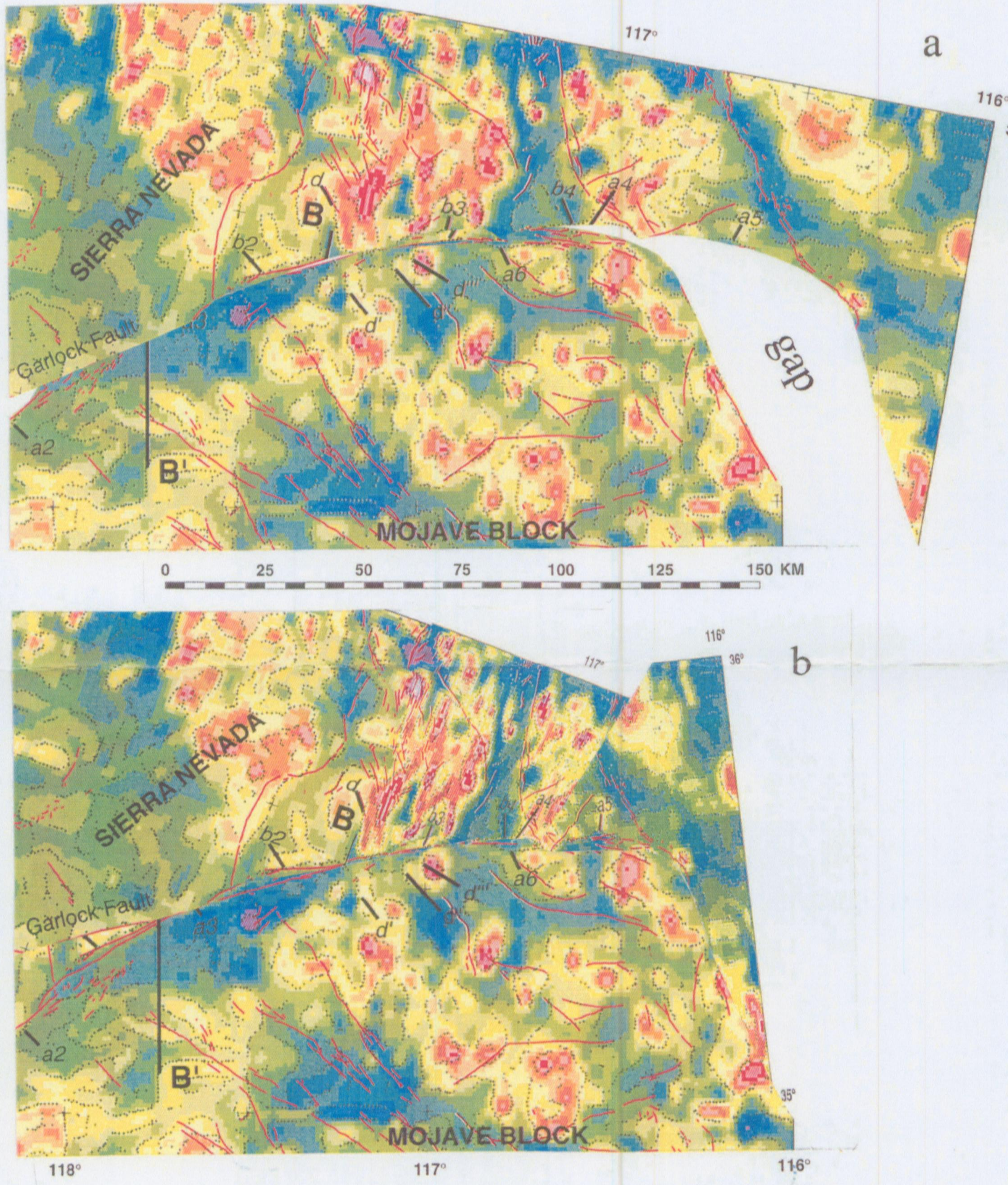


Figure 5. a) Map showing part of magnetic anomalies given in Figure 2, palinspastically restored according to model with offset of 47 km uniform east of Sierra Nevada. Contour interval 50 nanoTesla. b) Map showing part of magnetic anomalies given in Figure 2, palinspastically restored according to model with offset of 47 km at point ***d*** decreasing linearly eastward to o km at east end of Garlock fault. This is equivalent to compressing the region between point ***d*** and the east end of the Garlock fault to 60% of its present width. Contour interval 50 nanoTesla.

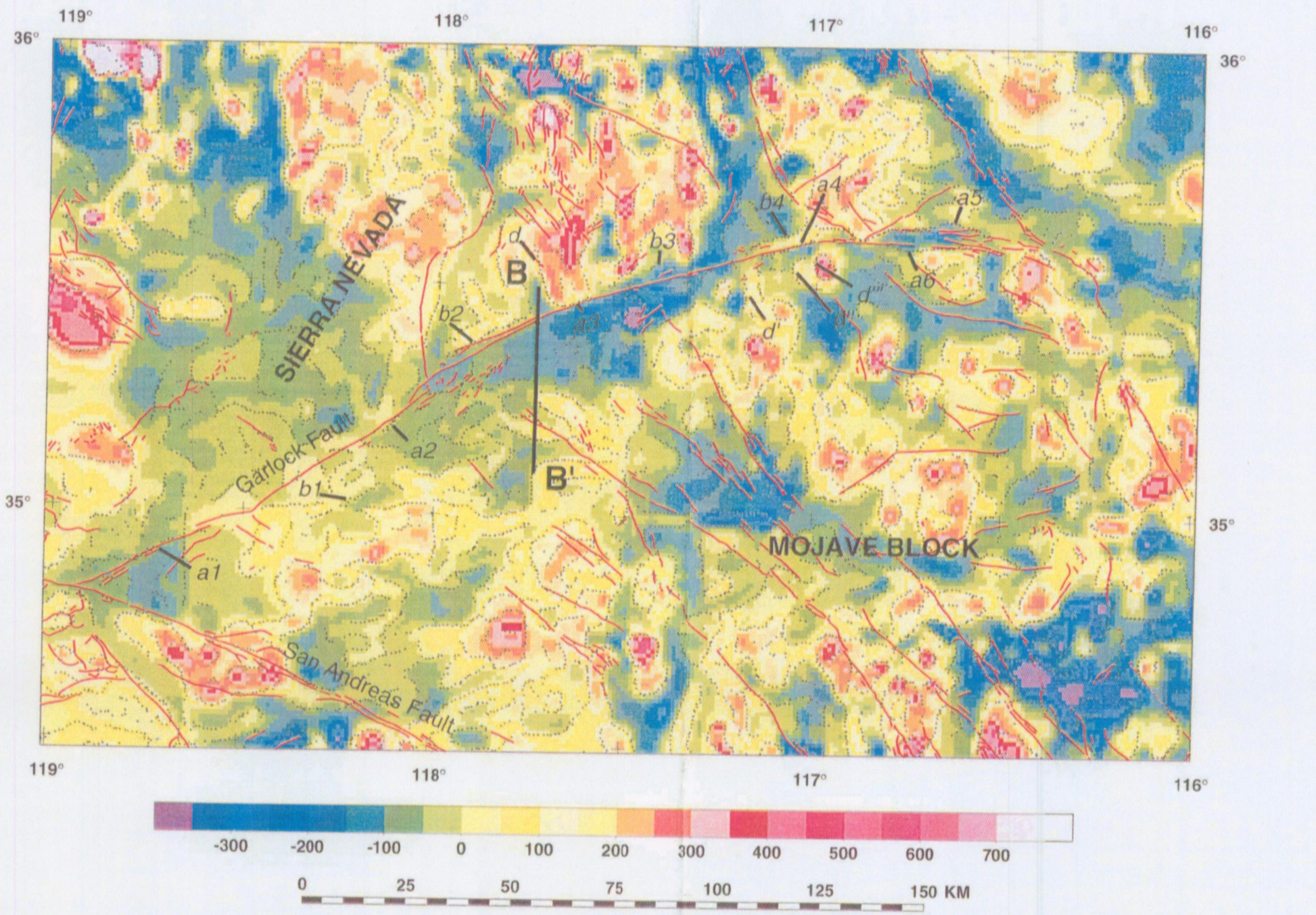
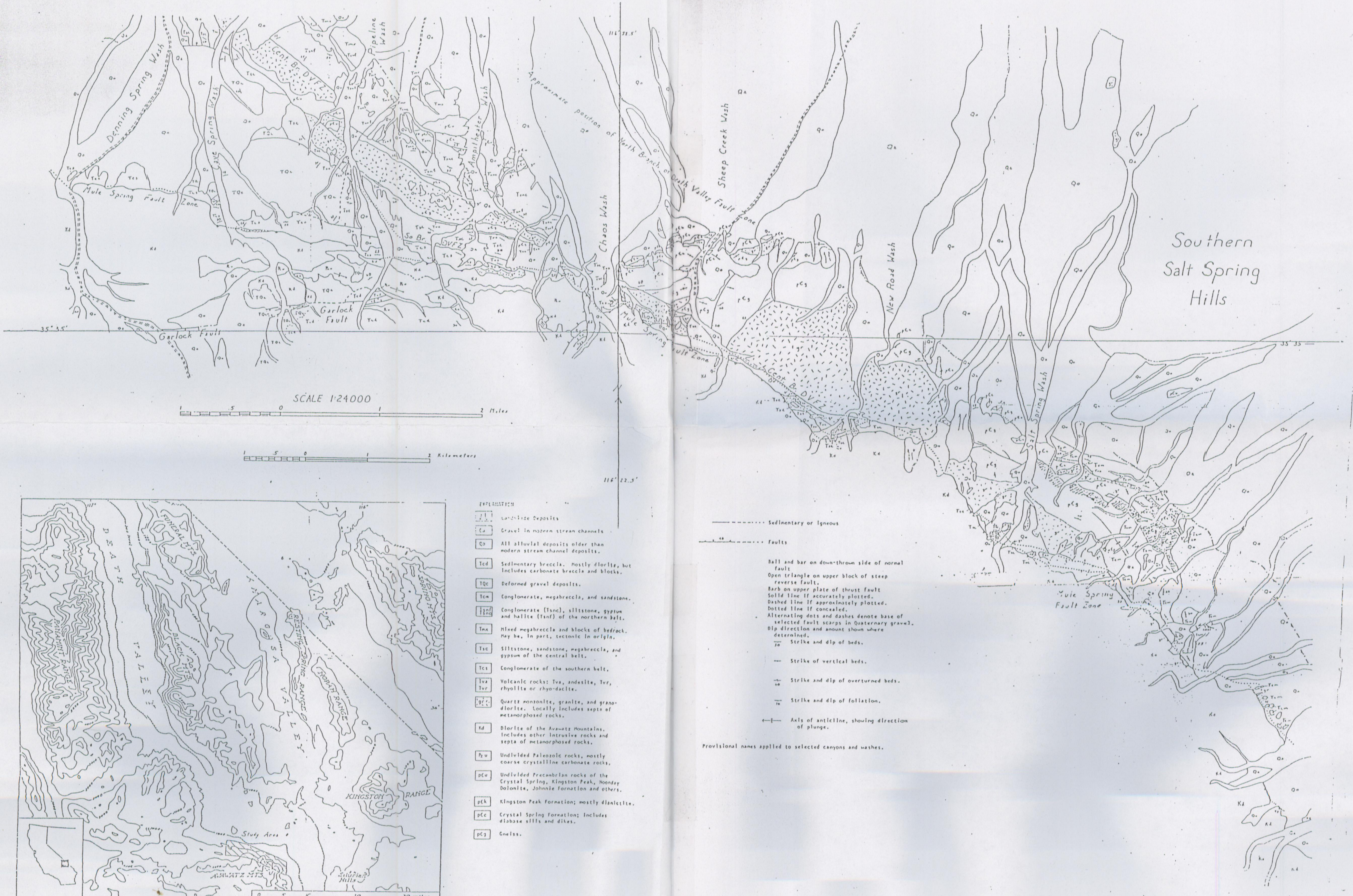


Figure 2. Map showing magnetic anomalies over the Garlock fault and vicinity (after Robers and Jachens, 1993). Faults (in red) from Jennings (1994). Aeromagnetic data from a number of separate surveys were numerically filtered and merged to produce a coherent map of magnetic anomalies, reduced-to-pole, at approximately 300 m above the ground surface. Small "+" symbols denote locations of inferred edges of magnetic rock bodies. Alphanumeric symbols denote magnetic features discussed in text. Profile B-B' was modeled to estimate depth extent of fault. Contour interval 50 nanoTesla.

GEOLOGIC MAP OF THE NORTH MARGIN OF THE AVAWATZ MOUNTAINS, SOUTHERN DEATH VALLEY, CALIFORNIA

BY BENNIE W. TROXEL AND PAUL R. BUTLER



EXPLANATION

	Landslide Deposits
Ca	Gravel in modern stream channels
Qo	All alluvial deposits older than modern stream channel deposits.
Tcd	Sedimentary breccia. Mostly diorite, but includes carbonate breccia and blocks.
TQc	Deformed gravel deposits.
Tcm	Conglomerate, megabreccia, and sandstone.
Tsnf / Tsnc	Conglomerate (Tsnc), siltstone, gypsum and halite (Tsnf) of the northern belt.
Tmx	Mixed megabreccia and blocks of bedrock. May be, in part, tectonic in origin.
Tsc	Siltstone, sandstone, megabreccia, and gypsum of the central belt.
Tcs	Conglomerate of the southern belt.
Tva / Tvr	Volcanic rocks: Tva, andesite, Tvr, rhyolite or rhyo-dacite.
gr	Quartz monzonite, granite, and granodiorite. Locally includes septa of metamorphosed rocks.
Kd	Diorite of the Avawatz Mountains. Includes other intrusive rocks and septa of metamorphosed rocks.
Pzu	Undivided Paleozoic rocks, mostly coarse crystalline carbonate rocks.
pCu	Undivided Precambrian rocks of the Crystal Spring, Kingston Peak, Noonday Dolomite, Johnnie Formation and others.
pCk	Kingston Peak Formation; mostly diamictite.
pCc	Crystal Spring Formation; includes diabase sills and dikes.
pCg	Gneiss.

Sedimentary or Igneous

Faults

Ball and bar on down-thrown side of normal fault
Open triangle on upper block of steep reverse fault.
Barb on upper plate of thrust fault
Solid line if accurately plotted.
Dashed line if approximately plotted.
Dotted line if concealed.
Alternating dots and dashes denote base of selected fault scarps in Quaternary gravel.
Dip direction and amount shown where determined.

- Strike and dip of beds.
- Strike of vertical beds.
- Strike and dip of overturned beds.
- Strike and dip of foliation.
- Axis of anticline, showing direction of plunge.

Provisional names applied to selected canyons and washes.

Geology by Bennie W. Troxel and Paul R. Butler, 1979, [illegible].