GEOPHYSICS, VOL. 42, NO. 1 (FEBRUARY 1977), P. 17–33, 9 FIGS., 4 TABLES

MICROEARTHQUAKE INVESTIGATION OF THE MESA GEOTHERMAL ANOMALY, IMPERIAL VALLEY, CALIFORNIA

JIM COMBS* AND DAVID HADLEY‡

Microearthquakes associated with the Mesa geothermal anomaly were recorded for five weeks during the summer of 1973 using an array of six portable, high-gain seismographs equipped with vertical-component 1-sec natural period seismometers. Background seismicity of the area is thus determined prior to development for geothermal power and water. The local seismicity changed considerably over the recording period. Most daily activity was characterized by only one or two potentially locatable events, while two microearthquake swarms of two- and three-day duration included as many as 100 or more distinct local events per day. Hundreds of small events (nanoearthquakes), some clustered in swarms, were recorded by each seismograph; however, most were not detected on four or more seismograms so that hypocentral locations usually could not be determined. Locations were determined for 36 microearthquakes having epicenters situated in the 150 km^2 areal extent of the geothermal anomaly. Focal depths ranged from near-surface to about 8 km. More than half of the located events have hypocenters greater than the 4.0 km which is approximately the depth to crystalline basement. Stress associated with the Mesa geothermal anomaly is relieved by a combination of continuous microseismic activity and intermittent microearthquake swarms. Based on the results of the present study, a new right-lateral strike-slip fault, the Mesa fault, was defined. First motion studies indicate strike-slip faulting although there is no surface expression of the fault. The northwest-southeast trending Mesa fault is an active fault functioning as a conduit for rising geothermal fluids of the Mesa geothermal anomaly. This investigation is another demonstration that geothermal areas are characterized by enhanced microearthquake activity.

INTRODUCTION

The Imperial Valley of southern California (Figure 1) is an area of considerable interest for the exploration and development of geothermal resources. Geothermal anomalies in the valley have been investigated by a variety of geophysical techniques (Rex, 1966; Helgeson, 1968; Biehler, 1971; Combs, 1971; Biehler and Combs, 1972; Meidav and Furgerson, 1972; Goforth et al, 1972; Douze and Sorrells, 1972; Combs, 1973; Combs and Swanberg, 1977). A fundamental characteristic of most geothermal resource areas is close association with regions of high geological activity manifested most commonly as microearthquakes (Westphal and Lange, 1966; Brune and Allen, 1967; Lange and Westphal, 1969; Ward et al, 1969; Ward and Björnsson, 1971; Ward and Jacob, 1971; Hamilton and Muffler, 1972; Ward, 1972; Combs and Rotstein, 1976). Seismic stress release usually occurs along existing zones of weakness and not in areas of previously unbroken rock. Good correlation has been found between zones of rupture and mapped faults where earthquake locations have been determined accurately and in sufficient number to define a pattern. Consequently, seismically active faults

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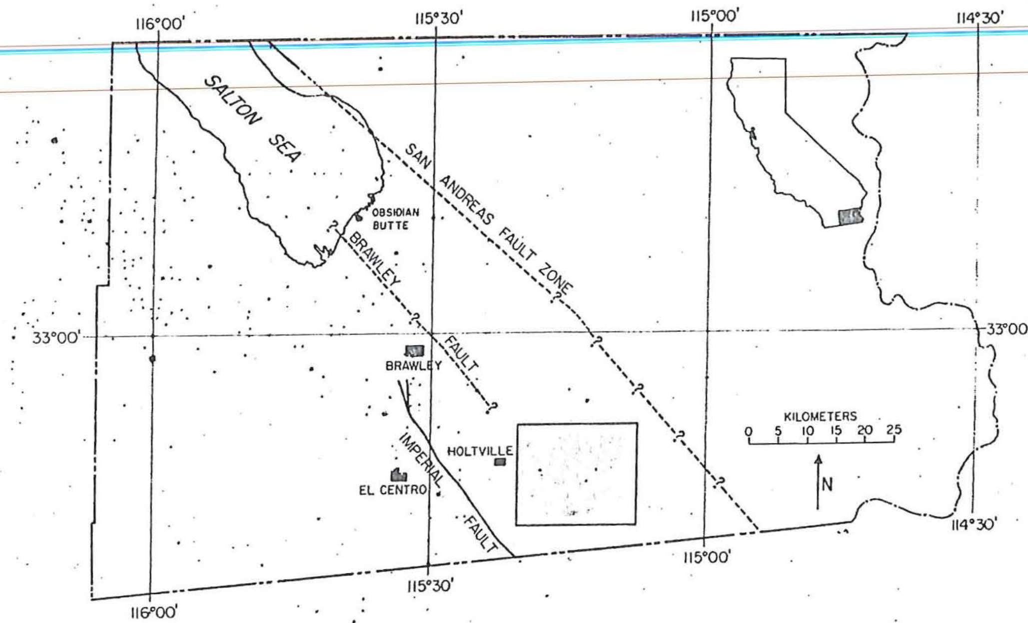


FIG. 1. Location map of Imperial Valley, southern California. Shaded rectangle depicts area of the microearthquake investigation including the Mesa geothermal anomaly and represents the map area in Figures 4, 8, and 9. Spots represent earthquake epicenters from the CIT seismograph network for 1961 to 1971. Size of the symbol denotes number of events with a single cross indicating one event.

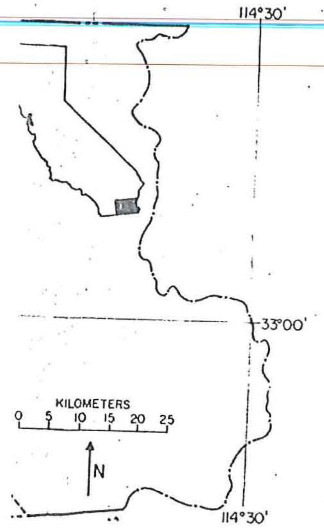
can be mapped in three dimensions using microearthquakes. Since active faults channel hot water from the vicinity of deep heat sources to shallow depths, information on active faults is important for geothermal exploration and production (Facca and Tonani, 1964; Ward and Jacob, 1971).

Studies have shown that faults may vary their normal patterns of earthquake activity if fluid pressures are changed in regions of tectonic stress (Evans, 1966; Healy et al, 1968; Raleigh et al, 1972; Sykes et al, 1972; Gupta et al, 1972; Gupta and Rastogi, 1976). In particular, evidence has accumulated recently showing that shallow earthquakes can be triggered by increases in fluid pressure induced by man. This phenomenon was observed near Denver, Colo., when Evans (1966) accidentally noted a correspondence between injection of chemical waste material into a deep disposal well and the level of earthquake activity in the Denver area. The time-space correlation of seismic activity, compared with the location and rates of injection into this disposal well, clearly demonstrated a causative relation between the earthquakes and fluid injection at this site (Healy et al, 1968). Numerous cases of increased seismic activity associated with the filling of large man-made reservoirs have been documented (Gupta et al, 1972; Gupta and Rastogi, 1976).

The effect of fluid pressure on rock strength to explain large thrust faults was first proposed by Hubbert and Rubey (1959). The physics of their observations is simple. The frictional forces that resist sliding along a fault plane are proportional to a coefficient of friction and the normal stress acting across the fault plane. Fluid pressure reduces this normal stress and therefore weakens the rock. These observations have lead directly to the conclusions that changes in fluid pressure may control the timing of seismic activity, and the possibility that a change in subsurface fluid pressure caused by reinjection of geothermal fluids could lead to a change in seismic activity associated with a geothermal field.

In order to assess the effects of variations in fluid pressure, seismic stations should be established in geothermal areas prior to the onset of the potential effects purportedly caused by production or reinjection. That is, stations should be established near productive geothermal reservoirs to determine if patterns emerge that appear to be related to the removal or reinjection of fluids.

The object of the present study was to delineate



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any seismically active faults and to characterize the background seismicity of the area in advance of withdrawing fluids from or injection of fluids into the Mesa geothermal anomaly. Microearthquakes associated with the area were recorded for five weeks during the summer of 1973 using an array of six portable, vertical-component seismographs. Hundreds of small events (nanoearthquakes which we defined as earthquakes of Richter magnitude ≤ 0) were recorded by each station. However, most events were not detected on four or more seismographs so that hypocentral locations usually could not be determined. Locations were determined for 36 microearthquakes having epicenters situated in the areal extent of the anomaly.

The seismicity changed considerably over the period of recording. The activity was normally characterized by only one or two potentially locatable events per day, while two microearthquake swarms of two- and three-day duration included as many as 100 or more distinct local events per day.

Based on the results of the present study, a new right-lateral strike-slip fault, the Mesa fault, was defined although there is no surface expression of the fault. The Mesa fault is active and functions as a conduit for rising geothermal fluids of the Mesa geothermal anomaly.

With the completion of each new microearthquake study, there is a more obvious correlation between earthquake swarms and subsurface thermal features, i.e., geothermal activity. Causal relationships between swarms and magmatic, volcanic, or hydrothermal activity still remain unresolved. However, by finding these microearthquake swarms, an efficient method of detecting ongoing volcanic, hydrothermal, and/or magmatic processes in potentially valuable geothermal areas may eventually be developed.

GENERAL GEOLOGY AND REGIONAL STRUCTURE

The Imperial Valley of southern California (Figure 1) is an extensive sedimentary basin characterized by high heat flow (Rex, 1966; Helgeson, 1968; Combs, 1971; Combs, 1973; Combs and Swanberg, 1977). The geology of the Imperial Valley, which is part of a structural depression known as the Salton trough that extends from the Gulf of California, has been described by Dibblee (1954). Much of the geologic work has been summarized in the California Division of Mines and Geology 1:250,000 scale geological map sheets—Santa Ana (Rogers, 1965), Salton Sea (Jennings, 1967), and San Diego/El Centro (Strand, 1962).

There are marine deposits of late Tertiary age along the edges of the Salton trough (Woodring, 1931) that contain pebbles derived from older formations in the adjoining mountains. This indicates that the Salton trough was in existence in the late Tertiary and was inundated by the Gulf of California at that time (Merriam and Bandy, 1965; van de Kamp, 1973). In the Salton trough, late Tertiary marine beds are overlain by playa deposits, also of late Tertiary age, but apparently no marine deposits of Quaternary age have been found (van de Kamp, 1973). These data suggest that the Gulf was excluded from the trough at about the beginning of Quaternary time, presumably by the formation of the Colorado River delta. The late Tertiary and Quaternary sediments filling the depression include predominantly unconsolidated sand, silt, and clay with some gravel. The sediments are saturated to within a few meters of the land surface. The accumulated thickness is estimated to be as great as 6 km, based on seismic refraction data (Biehler et al, 1964).

Since the Quaternary, the trough has been sinking, and the neighboring fault blocks, which include Tertiary strata, have been rising. Therefore, the Salton trough is delimited by a framework of mountains consisting of Mesozoic and older granitic and metamorphic rocks. The mountains reach heights above 1500 m.

The trough and the delta have been downwarped during the Holocene epoch. The presence of fresh fault scarps north of the Salton Sea and the frequent occurrence of earthquakes attest to continued structural movement (Allen et al, 1965; Hileman et al, 1973; Hill et al, 1975). Associated geothermal activity is indicated by the many recent or Holocene volcanoes (Elders et al, 1972) occurring along the Gulf and in the Salton trough, by deposits of volcanic ash (pumice) interbedded with lake beds (van de Kamp, 1973), by the occurrence of mud volcanoes that emit sulfurous steam, and by hot springs.

The structure of the Imperial Valley is controlled by numerous strike-slip faults of the San Andreas-San Jacinto fault system (Dibblee, 1954; Hamilton, 1961; Kovach et al, 1962; Biehler et al, 1964) on which considerable right-slip has occurred (Hill and Dibblee, 1953; Sharp, 1967). Some of the faults have been active during historic times (Richter, 1958; Allen et al, 1965; Brune and

Allen, 1967; Hileman et al, 1973; Hill et al, 1975).

The Imperial Valley appears to contain one or more centers of either spreading or crustal extension (Lomnitz et al, 1970; Elders et al, 1972). These centers appear to be bounded by strike-slip faults (Hill et al, 1975). It is important to determine whether the earthquake swarms in the Imperial Valley are located along centers of rifting or along strike-slip faults. Sykes (1970) found that swarm activity was confined to zones of spreading for some 20 earthquake swarms which originated from the mid-Atlantic ridge and from other parts of the mid-ocean ridge. However, this hypothesis (the relation between swarm frequency and transform faults or spreading zones) is not substantiated for earthquake swarms in Iceland (Tryggvason, 1973).

The Mesa geothermal anomaly is on the east flank of the Salton trough. Results of seismic refraction profiling indicate that local basement is at least 4.0 km deep (Kovach et al, 1962). In the immediate area of the Mesa geothermal anomaly, three wildcat oil wells, Shafer Barbara no. 1, American Petrofina no. 27-1, and Texaco Grupe-Engebretson no. 1, were drilled to depths of 2.44, 3.24, and 3.75 km, respectively. West of the anomaly, an exploratory geothermal well, Magma Sharp no. 1, has been drilled to a depth of 1.98 km. All of these wells penetrated predominantly deltaic sediments and none reached basement. Electric logs from these wells indicate interbedded sand and clay with minor silt beds. Two geothermal wells, U.S. Bureau of Reclamation (USBR) Mesa no. 6-1 and USBR Mesa no. 6-2 drilled on the Mesa geothermal anomaly, indicate the same subsurface geology with the upper stratigraphic sequence consisting of more than 50 percent clay which functions as a cap rock for the geothermal reservoir.

PREVIOUS INVESTIGATIONS

Three stations in the microearthquake survey ($0 < M < 3$) of southern California conducted by Brune and Allen (1967) were located within the Imperial Valley. At Obsidian Butte during the first two-day period of operation, a rate of 128 earthquakes per day was obtained; however, later occupation of the site for a period of five days indicated an average rate of about 32 per day. Brune and Allen (1967) postulated that, "although Obsidian Butte is a Quaternary volcanic plug associated with marked local thermal anomalies, the distribution of microearthquakes . . . represents the regional stress system . . . and is not solely the result of localized volcanic or thermal activity at depth."

Two other stations, Painted Canyon and East Mesa, both located on the extension of the San Andreas fault, detected no events with S-P time intervals less than 3 sec during two-day recording intervals. The East Mesa station of Brune and Allen (1967), located at 33°00.4′N and 115°18.0′W, was only 20 km north of the present Mesa geothermal anomaly array.

During the summer of 1971, the U.S.G.S. monitored for three weeks in the East Mesa area, including the region covered by the present array. They did not detect any microearthquakes within their seismic network, and it was suggested that "perhaps a longer recording period would have some success, as microearthquake activity is often sporadic" (R. M. Hamilton, written communication, 1971).

The California Institute of Technology (CIT) catalog of earthquakes in southern California, from 1932 through 1972, shows a dense pattern of magnitude 3 or greater earthquakes within the Imperial Valley (Hileman et al, 1973). Epicenters from the CIT seismograph network for the period 1961 to 1971 are superimposed on the location map of the Imperial Valley (Figure 1). Note that during the ten year period three earthquakes of magnitude approximately 3.0 occurred in the immediate area of the Mesa geothermal anomaly.

In April 1973, a 16-station seismograph network was installed in the Imperial Valley to improve the resolution of earthquake locations since this region has high tectonic activity and known geothermal resources (Hill et al, 1975). This network is part of a cooperative effort between CIT and the U.S.G.S. The Imperial Valley seismograph network recorded a series of earthquake swarms along the Imperial and Brawley faults which occurred in June and July of 1973 (Hill et

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FIG. 2. Representative one-day seismograms with seismic recording system peaked at 20 Hz and recording at 60 mm/minute. (a) Station MGA no. 2 during Imperial Valley swarm of June 19–21, 1973. (b) Station MGA no. 3 for same day; note continuous nanoearthquake activity associated with the geothermal area while no nanoearthquakes were recorded on station MGA no. 2 for the same day. (c) Station MGA no. 3 during normal time; again note the continuous nanoearthquake activity.

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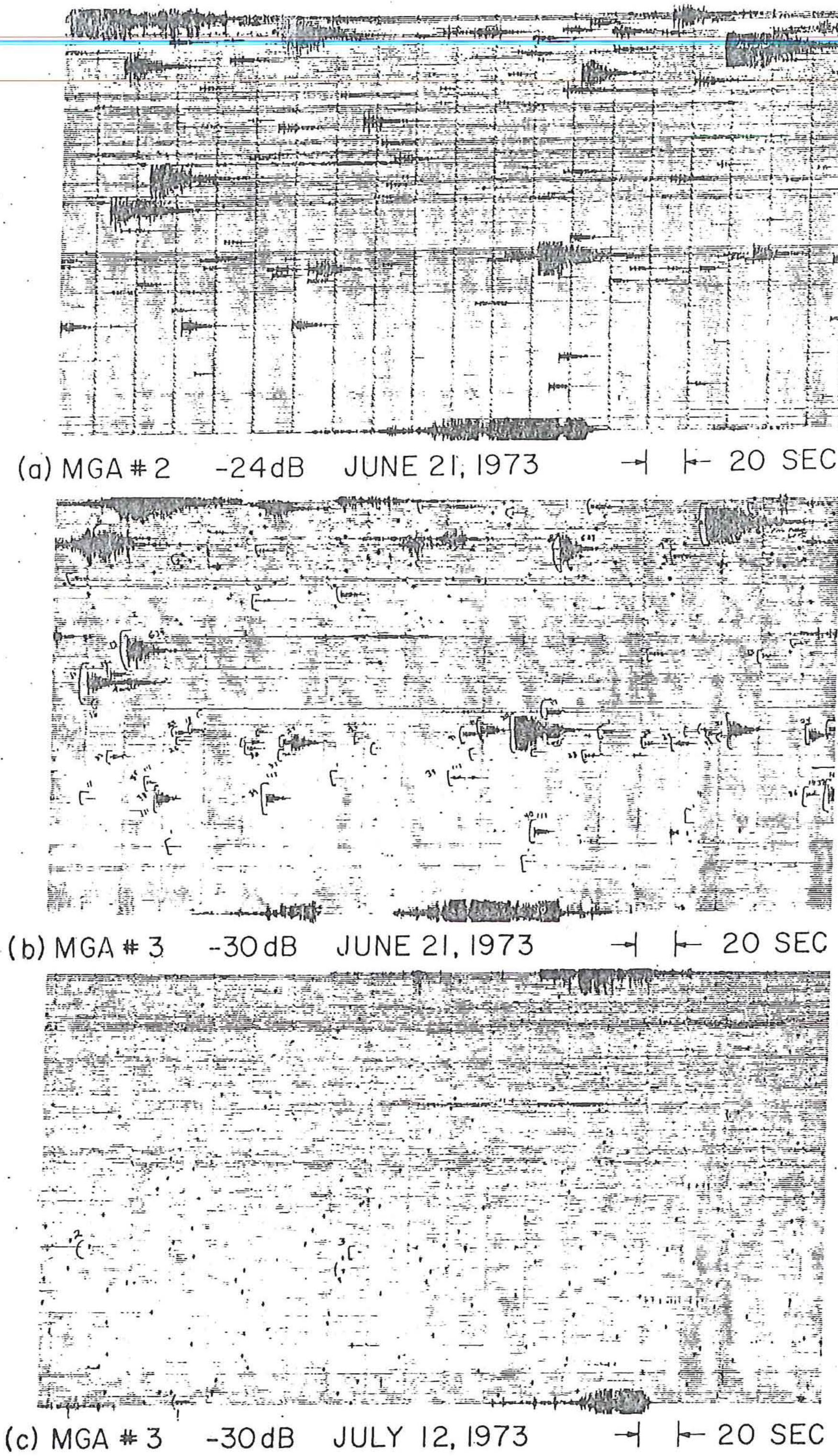

al, 1975). The Salton Sea geothermal field and the Brawley geothermal anomaly are closely associated with the earthquake swarms. Data from the network for these swarm sequences were used in the present study.

Richter (1958) first discussed the occurrence of swarm sequences in the Imperial Valley. In their extensive investigation of an oceanic ridge earthquake swarm which occurred during March 1969 in the area of the Wagner basin of the Gulf of California, Thatcher and Brune (1971) summarized four Imperial Valley swarm sequences. These swarms occurred in 1963, 1965, 1968, and 1969 in the Obsidian Butte area near the south end of the Salton Sea. Two swarms were recorded and examined during the present investigation.

INSTRUMENTATION

The microseismic array consisted of six stations. Each station was equipped with a portable, vertical-component, high-gain Kinemetrics PS-1 seismograph which has been described in detail by Prothero and Brune (1971). Several internal filters provided with the seismographs can be used to shape the system response. Filter 1, peaked at 1 Hz, produced easily discernible first motions, although it probably reduced the number of small local microearthquakes that were recorded since the high frequency waves from these events are mostly filtered out. Higher resolution of events, and well defined *S-P* interval times were obtained by using filter 4 peaked at 20 Hz which is the predominant frequency for very near microearthquakes. Gain was controlled by attenuating the maximum system sensitivity in 6-dB steps. Stations operated on filter 4 were attenuated from 24 to 36 dB, whereas the same stations could be operated on filter 1 with about 12 dB less attenuation. Recording was accomplished at 1 mm/sec with ink systems. Representative one-day records using filter 4 are presented in Figure 2.

Kinemetrics Ranger SS-1 seismometers with a natural frequency of 1 Hz were used. The average generator constant for these seismometers was 100 V-sec/m. External resistors damped the system at 0.7 critical.

Timing for the recording units was provided by a temperature-compensated, crystal-controlled unit with an accuracy of ±0.3 ppm over the temperature range of 0°C to 50°C. Timing was reestablished daily at each station by superimposing the National Bureau of Standards WWV time code with the internal clock on each record. The drum speed and accurate timing permitted *P* arrivals to be picked to 0.1 sec. Without horizontal instruments, *S* phases could not be clearly distinguished from multiple crustal *P* phases (see Figure 2) and, therefore, were not used in hypocentral locations.

HYPOCENTRAL LOCATIONS AND CRUSTAL MODEL

Microearthquakes located during this study were recorded by the array of six stations described in Table 1. Hypocenter locations were determined using HYPO71, a stepwise multiple regression computer program (Lee and Lahr, 1972). A series of iterations is performed with adjustments in the trial hypocenter location until the root-mean-square (rms) error in traveltime is minimized, or the specified number of iterations is exceeded. Greater than 80 percent of the located events have rms residual traveltimes less than 0.1 sec and 90 percent have less than 0.2 sec. The quality of the location for events outside the array is not as good as for those within as the location error increases rapidly with distance from the array.

To provide a velocity model for the shallow subsurface portions of the Mesa geothermal anomaly, a calibration blast of 180 kg of high-velocity seismic gel was exploded. The results of the refraction profile are shown in Figure 3. Location of the shotpoint is plotted in Figure 4. The

Table 1. Station locations and station traveltime corrections.

Station* no.	Latitude (N)	Longitude (W)	Elevation (meters)	Station delay (seconds)	Average attentuation Filter 4 (dB)
MGA 1	32° 49.79′	115° 09.73′	11	0.13	24
MGA 2	32° 51.25′	115° 12.69′	19	0.05	24
MGA 3	32° 48.62′	115° 15.72′	11	0.03	30
MGA 4	32° 46.89′	115° 11.65′	23	0.0	30
MGA 5	32° 44.72′	115° 07.53′	35	0.0	36
MGA 6	32° 44.31′	115° 14.75′	14	−0.21	36

* All microearthquake stations were located in Quaternary alluvium consisting of water-borne and wind-borne unconsolidated to poorly consolidated sediments.

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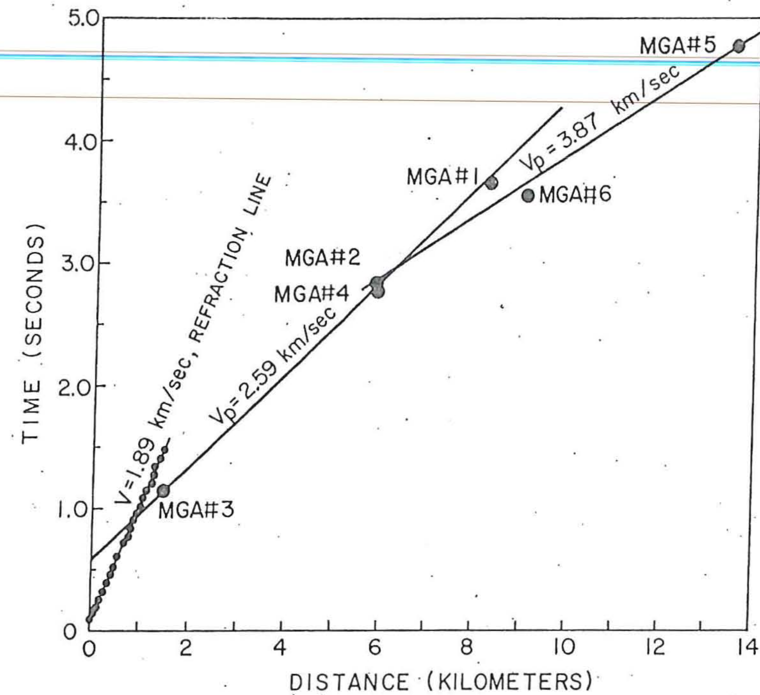

FIG. 3. Seismic refraction profile, Mesa geothermal anomaly, October 26, 1973. Location of shotpoint noted on Figure 4. Near-surface refraction line was west of, and extended from, the shotpoint.

near-surface refraction line was located directly west of and extended from the shotpoint. The recording array for the near-surface velocity data consisted of an approximately 1450-m length cable with 24 geophones spaced at 60-m intervals.

The crustal structure shown in Figure 3 and detailed in Table 2 is an average for the entire area, and does not correspond to the actual crustal structure which is more complicated. For example, small velocity reversals with depth were found on sonic logs for the American Petrofina Salton trough Prospect no. 27-1, the U.S. Bureau of Reclamation Mesa no. 6-1, and the Texas Co.

Table 2. Crustal structure

Velocity in layer (km/sec)	Depth to top of layer (km)
1.50	0.0
2.25	0.40
2.60	0.90
3.10	1.50
3.87	3.20
5.50	3.70
6.00	4.50
7.10	14.00
7.90	25.00

Grupe-Engebretson no. 1 wells. The upper 3 km of this model represents an average structure obtained from the geophysical logs for USBR Mesa no. 6-1. Velocities are similar to those obtained by Kovach et al, (1962) for their seismic profile 3 located less than 10 km north of the Mesa geothermal anomaly. The crustal velocity below 3.7 km is a modified version of Hamilton's (1970) velocity model for the Borrego Mountain area on the west side of the Imperial Valley.

Several different layered velocity models were used to test the uniqueness of the hypocentral locations. However, gross changes in the models seldom changed horizontal locations by more than 2 km. Location of the calibration explosion was made to determine the probable precision of the microearthquake hypocenters. The computed location was within 0.20 km in the horizontal plane and 0.10 km in depth of the actual shotpoint (see Table 3). The typical hypocenter uncertainties for located events within the array are probably less than 1 km.

MICROSEISMICITY

The epicenters of all microearthquakes located in this study are illustrated in Figure 4. Hypocen-

Table 3. Calibration blast on Mesa geothermal anomaly, October 26, 1973

Actual origin time	10/26/73—00:42:05.33 GMT
Calculated origin time	10/26/73—00:42:05.39 GMT
Actual location	32° 48.63′N 115° 14.79′W
Calculated location	32° 48.61′N 115° 14.68′W
Actual depth	0.04 km
Calculated depth	0.14 km
rms error of time residuals	0.02 sec
Error in depth	0.10 km
Error in location	0.20 km
Charge size	180 kg

ters, as well as other pertinent data for each of the 36 events, are presented in Table 4. The location errors for horizontal position and focal depth are less than 1 km for events within the seismometer array, and less than 2 km for events just outside the array.

The epicenters define an elongated zone of seismicity that extends westward from the central part of the Mesa geothermal anomaly. Ninety percent of the microearthquakes are located within a 3-km wide zone that appears to continue toward the west. Since the locational accuracy of events outside the array decreases very rapidly with distance from the array, earthquakes west of 115°20′W were not included. However, almost no events occurred in the central or eastern part of the array or in the area immediately east of the array.

The main concentration of activity is clustered along a fault zone striking N58°W, which passes through the location of the two geothermal wells, USBR Mesa no. 6-1 and USBR Mesa no. 6-2. We have named this fault zone the Mesa fault (see Figure 4). This seismically active fault associated with the Mesa geothermal anomaly provides the main channel along which hot water from depth

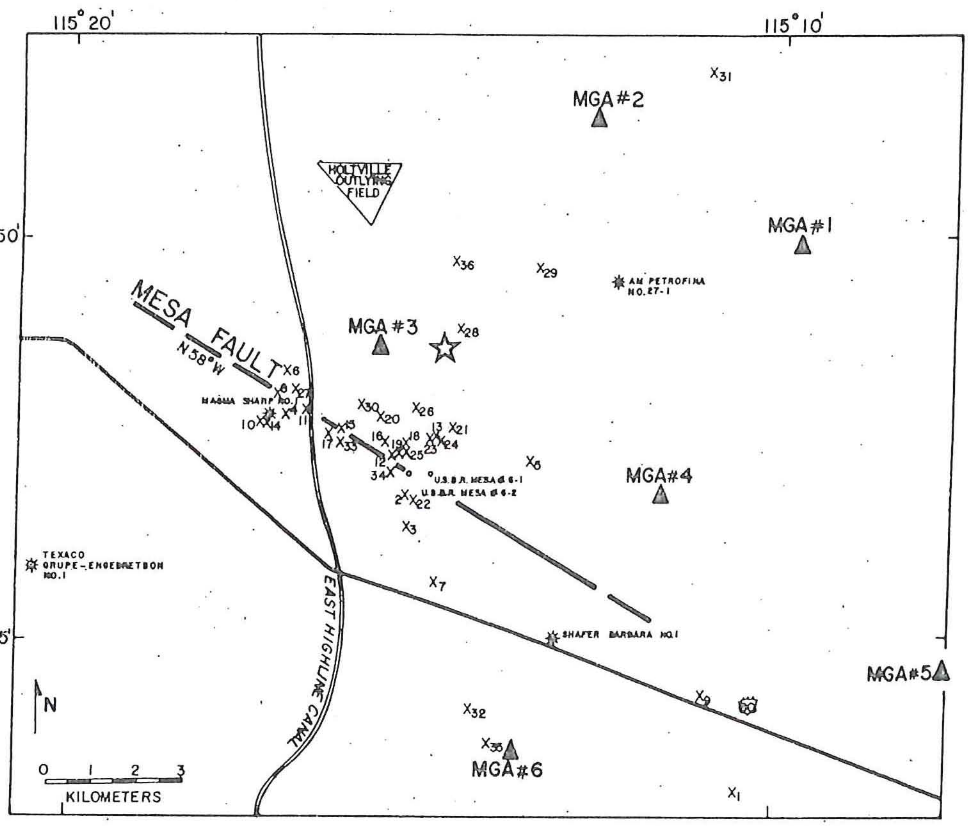

FIG. 4. Location map for Mesa geothermal anomaly with epicenters of present study. The ▲ indicates seismograph recording sites while O indicates U. S. Bureau of Reclamation geothermal wells. The ☼ indicates abandoned dry holes and calibration blast location is identified by ★. The number beside each epicenter (x) is the event number of Table 4.

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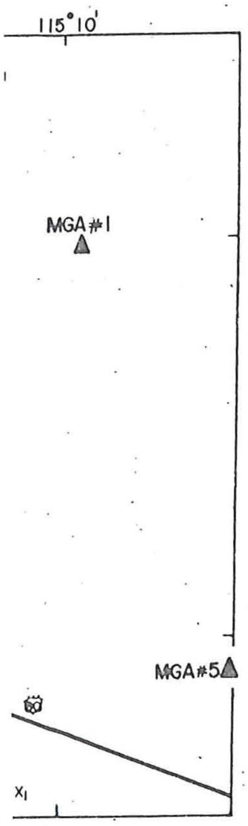

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reaches the permeable horizons in the shallow subsurface. During a similar microearthquake investigation in the Ahuachapan geothermal field of El Salvador, Ward and Jacob (1971) located 17 events that were either situated on or near a plane of seismic activity that intersected fault breccia mapped in a well. Since the fault plane is essentially parallel to other faults in the region, they postulated that the seismically active fault allows hot water to circulate to the surface in the Ahuachapan geothermal area.

The level of seismic activity was typically about one or two potentially locatable microearthquakes per day. During swarms, the activity was sometimes greater than 100 events per day. Swarms account for most of the microearthquake activity observed on the Mesa geothermal anomaly; the events concentrated in the area of highest heat flow (Combs, 1971; Combs and Swanberg, 1977). Two distinct swarms of two- and three-day duration occurred; one from June 19–21, 1973, the other during July 6–9, 1973. The Mesa swarm activity was associated with similar activity on the Brawley and Imperial faults, i.e., the small events (Figure 2) are located in the Mesa area, whereas the larger events cluster along the Imperial and Brawley faults (Hill et al, 1975).

Many of the nanoearthquakes recorded at station MGA no. 3 (Figure 2b and c) are of Richter magnitude less than zero and are not recorded on the other five stations. These events occur at a rate of at least 100 per day. They occur continuously and are not distinctly associated with stress release on the Imperial and Brawley faults. During the last day of recording, all instruments were moved into a tight array around MGA no. 3. Although a few of the nanoearthquakes were recorded on the southern stations in the tight array,

Table 4. Hypocenters of Mesa geothermal anomaly microearthquakes, June 10 to July 15, 1973

Event number[1]	Event code[2]	Latitude (N)	Longitude (W)	Focal depth (km)	Magnitude	Number of stations	
1	70 617	32°42.12′	115°10.45′	2.7	0.8	4	Fair
2	50 619	32°46.58′	115°15.56′	2.4	1.8	4	Good
3	60 619	32°46.16′	115°15.59′	2.3	1.8	4	Good
4	70 620	32°47.37′	115°18.46′	4.6	1.9	5	Fair
5	140 620	32°47.08′	115°13.60′	0.4	1.3	5	Good
6	150 620	32°47.73′	115°18.37′	8.7	1.9	6	Fair
7	70 621	32°44.56′	115°15.37′	3.3	0.8	5	Good
8	140 621	32°47.58′	115°19.00′	6.5	1.6	5	Fair
9	160 621	32°43.76′	115°10.70′	4.3	0.8	5	Good
10	180 621	32°47.24′	115°19.11′	5.4	1.6	5	Fair
11	190 621	32°47.42′	115°18.17′	7.0	1.9	5	Fair
12	210 621	32°46.94′	115°16.31′	3.3	1.9	5	Good
13	220 621	32°47.15′	115°15.51′	3.7	1.4	5	Good
14	230 621	32°47.18′	115°19.20′	6.3	1.6	5	Fair
15	240 621	32°47.22′	115°17.12′	4.1	1.7	5	Good
16	300 621	32°47.10′	115°16.72′	3.6	1.2	5	Good
17	310 621	32°47.14′	115°17.57′	4.5	1.2	5	Fair
18	320 621	32°47.11′	115°15.82′	3.6	1.2	5	Good
19	330 621	32°46.92′	115°16.00′	3.7	1.3	5	Good
20	340 621	32°47.32′	115°16.73′	5.6	1.3	4	Good
21	350 621	32°47.28′	115°15.27′	3.4	1.2	4	Good
22	360 621	32°46.57′	115°15.37′	0.3	1.2	4	Good
23	370 621	32°47.15′	115°15.67′	3.4	1.7	4	Good
24	380 621	32°47.10′	115°15.47′	4.4	1.7	4	Good
25	390 621	32°46.95′	115°15.98′	3.7	1.6	4	Good
26	400 621	32°47.49′	115°16.07′	5.6	1.5	4	Good
27	410 621	32°47.62′	115°19.02′	7.6	1.6	4	Fair
28	10 622	32°48.76′	115°14.56′	2.9	2.3	4	Fair
29	20 622	32°49.46′	115°13.54′	0.6	2.9	6	Good
30	10 623	32°47.62′	115°17.21′	4.1	1.5	6	Good
31	40 703	32°50.99′	115°11.44′	1.8	1.1	4	Fair
32	30 705	32°43.99′	115°14.58′	6.9	1.6	5	Fair
33	40 711	32°47.43′	115°16.28′	1.3	1.3	4	Fair
34	30 712	32°46.95′	115°15.46′	2.5	1.1	5	Good
35	40 712	32°43.10′	115°14.04′	5.0	0.8	5	Fair
36	10 713	32°49.49′	115°14.66′	2.0	1.4	5	Good

[1] The event number corresponds to number on Figure 4.
[2] The event code 70 617 indicates the seventh event recorded on four or more stations from the June 17, 1973 record.

unique locations could not be determined. *S-P* intervals less than 1.5 sec at MGA no. 3 (see Figure 3c) indicate a hypocentral distance of approximately 5 or 6 km, suggesting that these events represent on a microscale continuous activity within the same fault zone i.e., the Mesa fault, typified by swarm activity.

Most of the geothermal earthquakes occur at focal depths of 2 to 5 km. The tectonic events observed during this investigation which were outside the geothermal anomaly appear to extend to as deep as 8 km while Hill et al, (1975) observed focal depths between 5 and 14 km, with an uncertainty of ±5 km, for most of the events in the Imperial Valley that they examined. More than half of the events located in this study have hypocenters greater than the 4.0 km, which is approximately the depth to crystalline basement. Projection of hypocenters into planes transverse and longitudinal to the main trace of the Mesa fault illustrate these phenomena (Figure 5). The high temperature gradients within the geothermal anomaly confine the accumulation of strain to shallow depths and set an upper limit to the magnitude of the microearthquakes (Brace and Byerlee, 1970).

MAGNITUDES

The magnitudes of events located in this study were determined by relating total signal duration to Richter magnitude *M* (Richter, 1958) with an empirical formula derived from a San Jacinto Valley microearthquake investigation using the present instrumentation (Cheatum and Combs, 1975). The method has commonly been used in microearthquake studies (Lee et al, 1972; Real and Teng, 1973; Hadley and Combs, 1974) because of complications arising in determining magnitudes of events recorded on high-gain instruments.

The total signal duration *D* measured for each event was from the *P* arrival to the location on the seismogram where the signal-to-noise ratio became 4:1, which was usually less than 20 sec. The empirical relationship from Cheatum and Combs

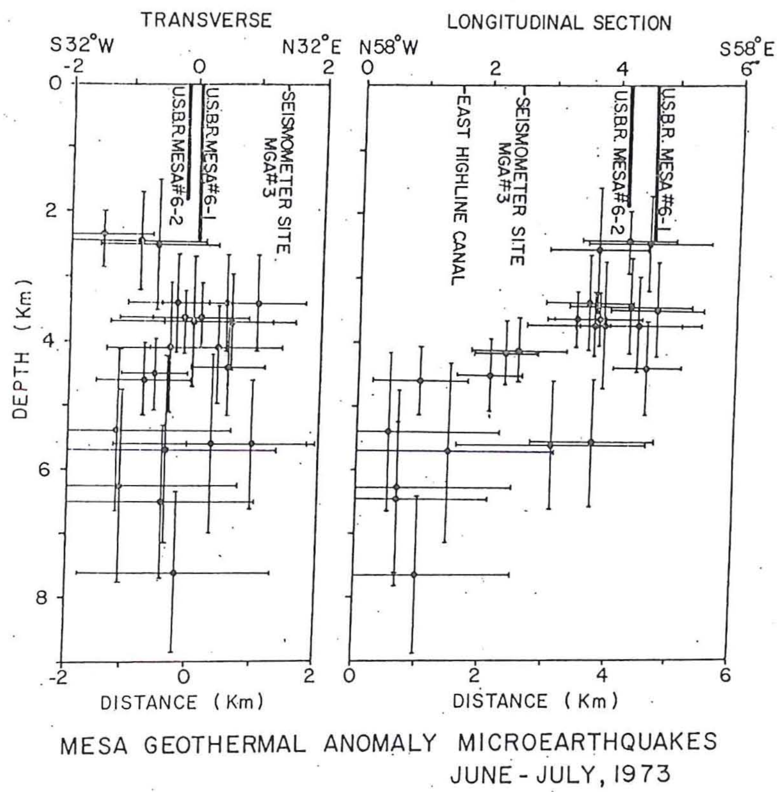

FIG. 5. Projection of hypocenters into planes transverse and longitudinal to the main trace of the Mesa fault, June–July, 1973. The length of lines indicates the relative error base in the hypocenter locations.

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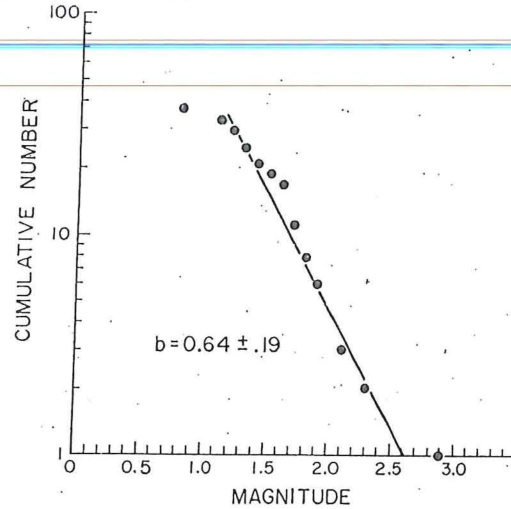


FIG. 6. Frequency-magnitude relation for Mesa geothermal anomaly microearthquakes.

(1975), with the instrument peaked at 20 Hz is

$$M = -2.1 + 2.35 \log D, \qquad (1)$$

where M is the Richter magnitude and D is the signal duration in seconds. Calculated magnitudes for an event recorded at different stations varied by as much as 0.4. Therefore, most magnitudes represent a simple average between two or more stations. Although the calculated magnitudes may not be exact, the relative magnitudes are accurate.

Earthquakes with magnitudes greater than 3.0 do not seem to be located within the Mesa geothermal anomaly, although larger magnitude events (>3.5) occur within 10 or 15 km away (Hill et al, 1975: this study). Ward et al (1969) and Ward and Björnsson (1971) noted that in the geothermal areas of Iceland, earthquakes are of magnitudes less than 4.5 whereas, 10 km away from these geothermal areas events of magnitude greater than 6 occur. These observations can be explained since high-temperature gradients inside geothermal areas set an upper limit on the accumulation of strain and therefore limit the size (magnitude) of the earthquakes (Brace and Byerlee, 1970). The contrast in local geology—basalts in Iceland and sediments in the Imperial Valley—is the primary reason for the difference in maximum magnitude of events associated with both these geothermal areas.

The frequency of occurrence of earthquakes is generally observed to be related to the magnitude of events by the relationship (Gutenberg and Richter, 1954),

$$\log N = a - bM, \qquad (2)$$

where N is the number of earthquakes having magnitudes between M and $M + dM$. The constant a is an index of the rate of earthquake occurrence and the constant b is an index, usually in the range of 0.4 to 1.8, expressing the relative number of small and large events (Allen et al, 1965). Miyamura (1962) and Allen et al (1965) have emphasized the possible tectonic significance of regional variations in b. Francis (1968) summarized the possible interpretations for the different magnitude-frequency relations and postulated that b is a function of the type of faulting, large for normal, and small for strike-slip faulting; however, the relationship between b value and source mechanism is not firmly established.

Using a combination of the techniques of Aki (1965), Wyss and Lee (1973), and Pfluke and Steppe (1973), a value of $b = 0.64 \pm 0.19$ was calculated for the located Mesa events (Figure 6). The value of 0.64 for b is quite similar to a value of 0.65 which Tryggvason (1970, 1973) found associated with the Reykjanes, Iceland, earthquake swarm of September 28–30, 1967. Hill et al, (1975) found a b value of 0.80 for all of the earthquakes of magnitude 2 and greater in the June and July, 1973 swarms in the Imperial Valley.

FIRST MOTION

Focal mechanism solutions can be obtained by plotting the compressional or dilatational first motion, as recorded at each station, on an equal area projection of the upper or lower hemisphere of the focal sphere with the hypocenter at the origin (Stauder, 1962). Polarities of the instruments were checked by using a magnitude of 7.5 earthquake in the Aleutians, as well as several southern Nevada events.

Focal mechanism solutions were obtained for two events from the June 19–21, 1973, swarm (Figure 7) located less than one-half kilometer from the trace of the Mesa fault (see Figure 4). First motions were recorded for five instruments of the Mesa array. In addition, first motions were obtained from several stations of the more widely spaced USGS-CIT Imperial Valley net (Hill et al, 1975).

Although the first motion data cannot constrain the fault plane solution, the northwest-southeast strike of the located events strongly im-

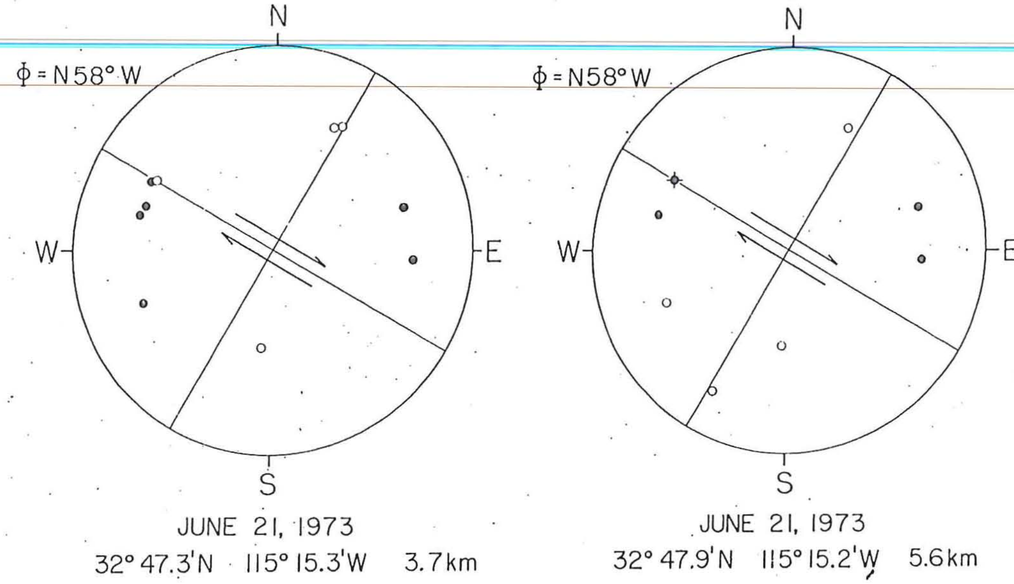


FIG. 7. Focal mechanism solutions for two microearthquakes. Open circle is dilatation; solid circle is compression with stations projected on the lower hemisphere. Data is from present array as well as USGS-CIT Imperial Valley array.

plies a similar trending fault. Thus, the data points require right-lateral strike-slip motion on the Mesa fault. Without a greater station density, the dip-slip component of motion on the Mesa fault cannot be resolved.

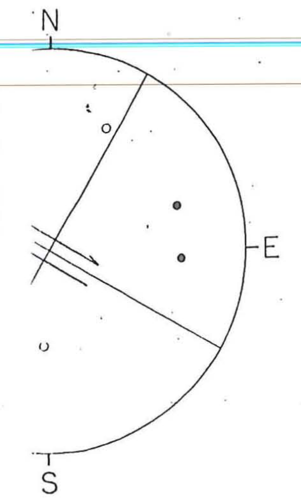
RESULTS AND DISCUSSION

Stress associated with the Mesa geothermal anomaly is relieved by a combination of continuous seismic activity and intermittent microearthquake swarms. Similarly, other geothermal areas are associated with extensive microearthquake activity ranging from continuous activity to intermittent swarms. Lange and Westphal (1969) located 19 microearthquakes during 120 hours of recording, which were "on the fault system associated with the Geysers steam zone in Sonoma County, California". In a more detailed study, Hamilton and Muffler (1972) found 53 microearthquakes in three weeks with epicenters situated in a zone about 4 km long and 1 km wide passing through the Geysers geothermal field along a principal fault zone. Westphal and Lange (1966) observed continuous microearthquake activity associated with hot springs in the Sawtooth Range, Idaho; Socorro Mountain, New Mexico; Bridgeport, California; and Dixie Valley, Nevada.

In the Ahuachapan geothermal field, Ward and Jacob (1971) observed that "local events tended to occur in swarms of 10 to 20 events spaced days and weeks apart." Ward et al (1969) and Ward and Björnsson (1971) also noted that earthquakes in geothermal areas of Iceland occur primarily as swarm type sequences. Thatcher and Brune (1971) located four swarms of earthquakes since 1962 near Obsidian Butte, a geothermal area just south of the Salton Sea (see Figure 1). During the five-week recording interval, two earthquake swarms occurred; the first from June 19–21, 1973, and the second from July 6–9, 1973 (Hill et al, 1975).

Although there does not seem to be a continuous fracture zone, microearthquake swarms in the Imperial Valley appear to migrate from the south to the north. The Imperial fault and the Mesa fault seem to respond simultaneously (see Figures 1 and 4). From the two swarms that were studied, the data were suggestive of a disturbance propagating along a fracture zone at a rate of about 10 to 40 km/day. Thus, the stress release that is initially associated with swarm activity on the Mesa and Imperial faults is transferred in time and space to the Brawley fault (see Figure 1).

Among the several possible mechanisms of swarm generation, Sykes (1970) has pointed out that two are related to volcanism. In one case, magmatic activity acts as a concentrated source of stress; whereas in the other, volcanoes are the loci of heterogeneities in physical properties that lead



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to inhomogeneities in the stress distribution. Mogi (1963) and Scholz (1968) noted during laboratory studies of microfracturing that stress inhomogeneities, related to either inhomogeneous materials or concentrated sources, are correlated with swarm-like sequences. Thus, earthquake swarms associated with geothermal areas may be indicative of magmatic activity in progress. That is, swarms may reflect either magmatic activity that does not reach the surface of the earth as volcanic eruptions, or hydrothermal processes that trigger tectonic strain release.

A third hypothesis for the generation of swarms has been postulated. In particular, swarms may indicate nonmagmatic mechanisms producing concentrations or inhomogeneities in stress. For example, localized sources of high fluid pressure may lower the effective strength of the rocks and act essentially as concentrated sources of stress. Such a mechanism was undoubtedly a causative agent in the series of earthquakes at the Denver Arsenal, Colorado (Evans, 1966; Healy et al, 1968), Rangely, Colorado (Raleigh et al, 1972), and Dale, New York (Sykes et al, 1972). Similarly, magmas or hydrothermal processes are possible sources of high fluid pressures near volcanoes and some geothermal areas.

Energy release in the form of continuous nanoearthquake activity is centered over the zone of highest heat flow (Combs and Swanberg, 1977), i.e., inside the 5.0×10^{-6} cal/cm²-sec isoflux contour of Figure 8. Most of this seismicity was recorded only on station MGA No. 3 (see Figure 2). Although unique locations could not be determined for these nanoearthquakes, a few of them were recorded on the southern stations of a tight (<2 km) array that was centered around station MGA no. 3. The differences in arrival times of shear and compressional waves (*S-P* times) recorded at MGA no. 3 were less than 1.5 sec indicating a hypocentral distance of 5 or 6 km. These data indicate that the Mesa fault is charac-

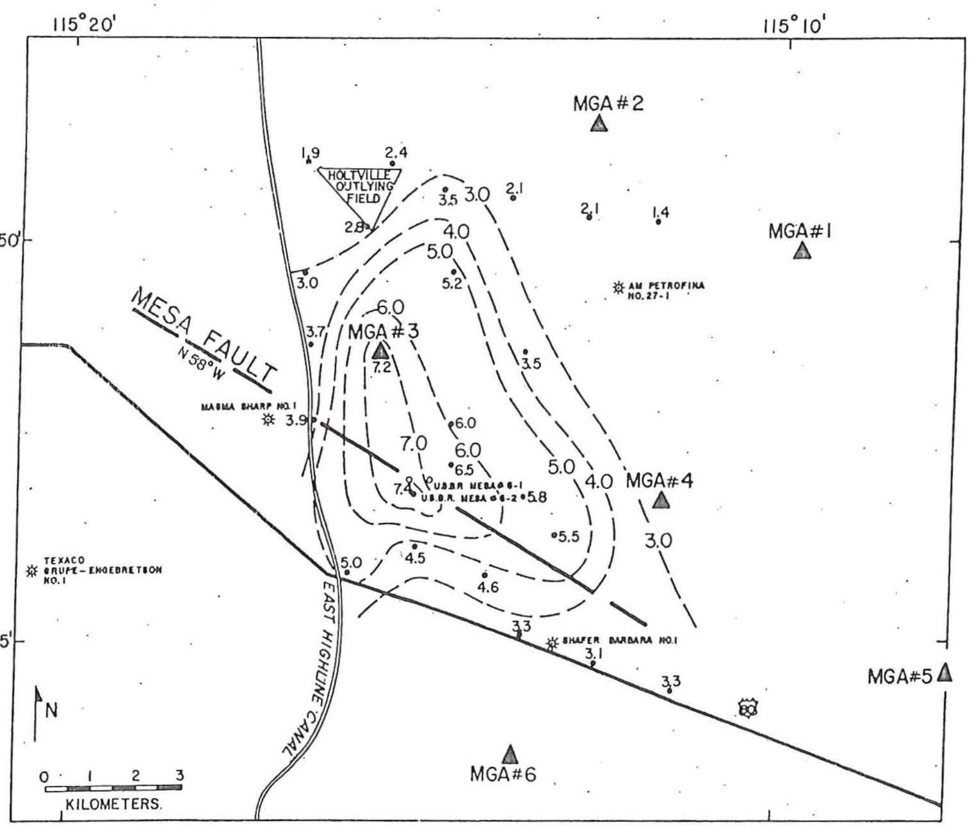

FIG. 8. Heat flow map of the Mesa geothermal anomaly. Contours are isofluxes of 10^{-6} cal/cm²-sec (Combs and Swanberg, 1977). Note the location of the present seismograph network.

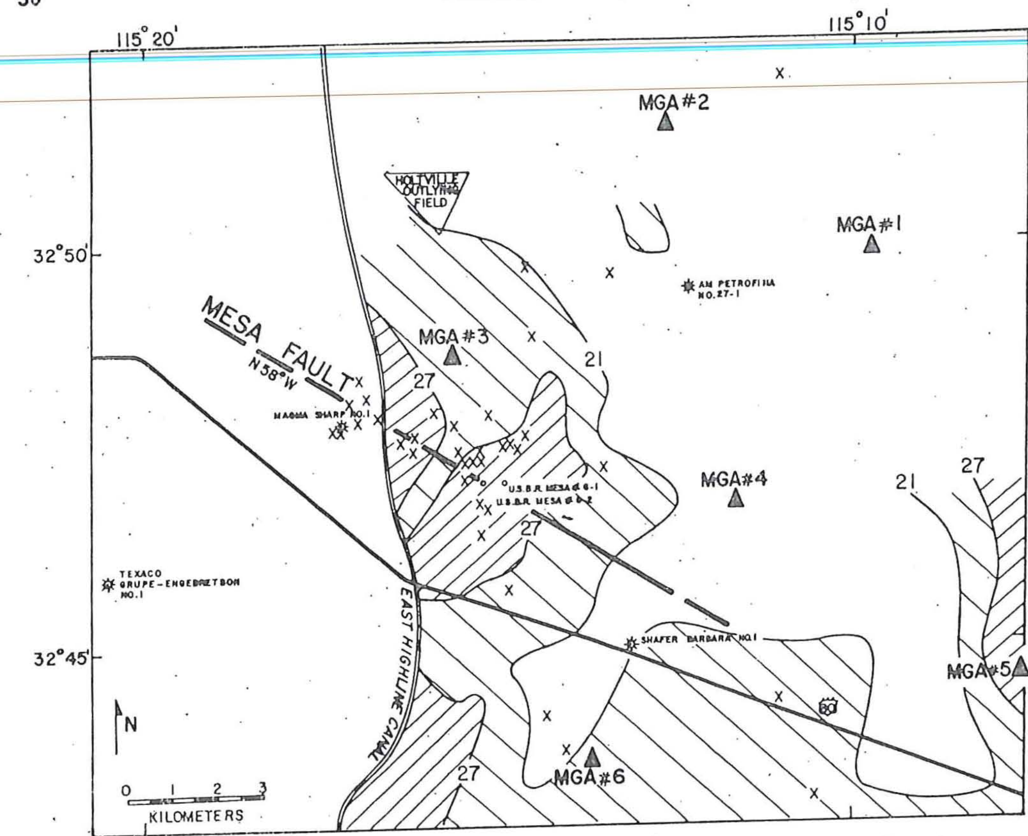

FIG. 9. Seismic ground-noise map of the Mesa geothermal anomaly. Contours of total power are given relative to 1 (millimicron/sec)² per Hertz in the passband of 3.0 to 5.0 Hz (after Teledyne Geotech, 1972). Note microearthquake epicenters plotted as x.

terized by both continuous nanoearthquake activity and intermittent swarm sequences.

There appears to be a coincidence between areas of high seismic noise and areas of microearthquake activity. Results of a seismic ground-noise survey conducted by Teledyne-Geotech (1972) are presented in Figure 9. Contours are total power in decibels relative to 1 (millimicron/sec)² per Hertz in the passband of 3.0 to 5.0 Hz. The contour interval is equal to 3 dB. Epicenters determined using the Mesa geothermal anomaly seismograph network are indicated by an *x*. Enhanced seismic ground-noise is probably caused by surface wave phenomena associated with the continuously occurring nanoearthquakes.

CONCLUSIONS

Recently completed studies in regions of tectonic stress have demonstrated that shallow earthquakes can be triggered by increases in subsurface fluid pressure induced by man. In order to assess the effects of variations in fluid pressure, seismograph stations were established on the Mesa geothermal anomaly prior to the onset of the potential effects purportedly caused by production or reinjection. Recording with an array of six portable, high-gain seismographs for 5 weeks during the summer of 1973, we were able to delineate a seismically active fault, the Mesa fault, and to characterize the background seismicity of the area before the withdrawal of fluids from or injection of fluids into the Mesa geothermal anomaly.

Seismic activity changed considerably during the period of recording. The seismicity was usually characterized by only one or two potentially locatable microearthquakes per day, while two microearthquake swarms of two- and three-day duration included as many as 100 or more distinct local events per day. Focal mechanism solutions were obtained for two events from the June 19–21, 1973 swarm. Although the first motion data cannot constrain the fault plane solution, the northwest-southeast strike of the 36 located microearthquakes strongly implies a similar trending fault. Thus, the data require right-lateral strike-slip motion on the Mesa fault, although there is no surface expression of the fault. Hundreds of nanoearthquakes were recorded. Most were not detected on four or more seismographs so that unique hypocentral locations could not be determined. The majority of the continuous nanoearthquake activity was recorded only on station MGA no. 3. A tight array was established around MGA no. 3 and the recorded *S-P* times on the station were examined. These data indicate that the Mesa fault is characterized by continuous microscale activity as well as swarm activity. The active Mesa fault functions as a conduit maintaining a zone of high permeability for rising geothermal fluids of the Mesa geothermal anomaly.

Stress associated with the Mesa geothermal anomaly is relieved by a combination of intermittent microearthquake swarms and continuous nano- and microseismic activity. Earthquakes with magnitudes greater than 3.0 do not seem to be located within the Mesa geothermal anomaly, although larger magnitude events occur only 10 or 15 km away. In addition, the geothermal earthquakes occur at focal depths of 2 to 5 km; whereas tectonic events outside the anomaly appear to extend to as deep as 8 km. High temperature gradients within the geothermal anomaly confine the accumulation of strain to shallow depths and set an upper limit to the size of the microearthquakes.

This investigation is another demonstration that geothermal areas are characterized by high microearthquake activity. With the completion of each new microearthquake study, there is a more obvious correlation between swarm sequences and subsurface geothermal activity. Although causal relationships still remain unresolved, by locating these microearthquake swarms, an efficient method of detecting ongoing hydrothermal or magmatic processes in potentially valuable geothermal areas may eventually be developed.

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REFERENCES

Aki, K., 1965, Maximum likelihood estimate of *b* in the formula log $N = a\text{-}bM$ and its confidence limits: Bull. Earthq. Res. Inst., v. 43, p. 237–239.

Allen, C. R., St. Amand, P., Richter, C. F., and Nordquist, J. M., 1965, Relationship between seismicity and geologic structure in southern California: Bull. SSA, v. 55, p. 753–798.

Biehler, S., 1971, Gravity studies in the Imperial Valley *in* Cooperative geological-geophysical-geochemical investigations of geothermal resources in the Imperial Valley area of California: Univ. Calif. Riverside, Educ. Res. Serv., p. 29–41.

Biehler, S., Kovach, R. L., and Allen, C. R., 1964, Geophysical framework of northern end of Gulf of California structural province, *in* Marine geology of Gulf of California: T. Van Andel and G. Shor (editors), AAPG Memoir 3, p. 126–143.

Biehler, S., and Combs, J., 1972, Correlation of gravity and geothermal anomalies in the Imperial Valley, Southern California (abstract): GSA Abstracts with Programs, v. 4, no. 3, p. 128.

Brace, W. F., and Byerlee, J. D., 1970, California earthquakes: Why only shallow focus?: Science, v. 168, p. 1573–1575.

Brune, J. N., and Allen, C. R., 1967, A micro-earthquake survey of the San Andreas Fault system in Southern California: Bull. SSA, v. 57, p. 277–296.

Cheatum, C., and Combs, J., 1977, Microearthquake study of the San Jacinto Valley Area, Riverside County, California: Bull. SSA, in preparation.

Combs, J., 1971, Heat flow and geothermal resource estimates for the Imperial Valley, *in* Cooperative geologica-geophysical-geochemical investigations of geothermal resources in the Imperial Valley area of California: Univ. Calif. Riverside, Educ. Res. Serv., p. 5–27.

———, 1973, Heat flow in the Imperial Valley, California: Presented at GSA Cordilleran Section Meeting, Portland, State University, Portland, Oregon.

Combs, J., and Rotstein, Y., 1976, Microearthquake studies at the Coso geothermal area, China Lake, California: Proc., 2nd U.N. Sympos., Develop. use of geothermal Res., v. 2, p. 909–916.

Combs, J., and Swanberg, C. A., 1977, Heat flow on the East Mesa KGRA, Imperial Valley, California: Geophysics, in preparation.

Dibblee, T. W., 1954, Geology of the Imperial Valley region, California: Calif. Div. Mines Bull., v. 170, p. 21–28.

Douze, E. J., and Sorrells, G. G., 1972, Geothermal ground-noise surveys: Geophysics, v. 37, p. 813–824.

Elders, W. A., Rex, R. W., Meidav, T., Robinson, P. T., and Biehler, S., 1972, Crustal spreading in southern California: Science, v. 178, p. 15–24.

Evans, D. M., 1966, Man-made earthquakes in Denver: Geotimes, v. 10, p. 11–18.

Facca, G., and Tonani, F., 1964, Theory and technology of a geothermal field: Bull. Volcanol., v. 27, p. 143–189.

Francis, T. J. G., 1968, The detailed seismicity of mid-oceanic ridges: Earth Planet. Sci. Lett., v. 4, p. 39–46.

Goforth, T. T., Douze, E. J., and Sorrells, G. G., 1972, Seismic noise measurements in a geothermal area: Geophys. Prosp., v. 20, p. 76–82.

Gupta, H. K., and Rastogi, B. K., 1976, Dams and Earthquakes: Elsevier Scientific Publishing Co., Amsterdam, 229 p.

Gupta, H. K., Rastogi, B. K., and Narain, H., 1972, Common features of the reservoir-associated seismic activity: Bull. SSA, v. 62, p. 481–492.

Gutenberg, B., and Richter, C. F., 1954, Seismicity of the earth and related phenomena: 2nd ed., Princeton Univ. Press, Princeton, N. J., 310 p.

Hadley, D., and Combs, J., 1974, Microearthquake distribution and mechanisms of faulting in the Fontana-San Bernardino area of southern California: Bull. SSA, v. 64, p. 1477–1499.

Hamilton, R. M., 1970, Time-term analysis of explosion data from the vicinity of the Borrego Mountain, California, earthquake of 9 April 1968: Bull. SSA, v. 60, p. 367–381.

Hamilton, R. M., and Muffler, L. J. P., 1972, Microearthquakes at the Geysers geothermal area, California: J. Geophys. Res., v. 77, p. 2081–2086.

Hamilton, W., 1961, Origin of the Gulf of California: Bull. GSA, v. 72, p. 1307–1318.

Healy, J. H., Rubey, W. W., Griggs, D. T., and Raleigh, C. B., 1968, The Denver earthquakes: Science, v. 161, p. 1301–1310.

Helgeson, H. C., 1968, Geologic and thermodynamic characteristics of the Salton Sea geothermal system: Am. J. Sci., v. 266, p. 129–166.

Hileman, J. A., Allen, C. R., and Nordquist, J. M., 1973, Seismicity of the southern California region, 1 January 1932 to 31 December 1972: Seism. Lab., Cal. Inst. Tech., 492 p.

Hill, D. P., Mowinckel, P., and Peake, L. G., 1975, Earthquakes, active faults, and geothermal areas in the Imperial Valley, California: Science, v. 188, p. 1306–1308.

Hill, M. L., and Dibblee, T. W., Jr., 1953, San Andreas, Garlock and Big Pine faults, California—a study of the character, history and tectonic significance of their displacements: Bull. GSA, v. 64, p. 443–458.

Hubbert, M. K., and Rubey, W. W., 1959, Role of fluid pressure in mechanics of overthrust faulting, 1. Mechanics of fluid-filled porous solids and its application to overthrust faulting: Bull. GSA, v. 70, p. 115–166.

Jennings, C. W., 1967, Geologic map of California, Salton Sea Sheet: California Division of Mines and Geology, Sacramento.

Kovach, R. L., Allen, C. R., and Press, F., 1962, Geophysical investigations in the Colorado delta region: J. Geophys. Res., v. 67, p. 2845–2871.

Lange, A. L., and Westphal, W. H., 1969, Microearthquakes near the Geysers, Sonoma County, California: J. Geophys. Res., v. 74, p. 4377–4378.

Lee, W. H. K., and Lahr, J. C., 1972, HYPO71: A computer program for determining hypocenter, magnitude, and first motion pattern of local earthquakes: U.S.G.S. Open file rep., Menlo Park, 100 p.

Lee, W. H. K., Bennett, R. E., and Meagher, K. L., 1972, A method of estimating magnitude of local earthquakes from signal duration: U.S.G.S. Open file rep., Menlo Park, 28 p.

Lomnitz, C., Mooser, F., Allen, C. R., Brune, J. N., and Thatcher, W., 1970, Seismicity and tectonics of the Northern Gulf of California region, Mexico. Preliminary results: Geofisica Internacional, v. 10, p. 37–48.

Meidav, T., and Furgerson, R., 1972, Resistivity studies of the Imperial Valley Geothermal Area, California: Geothermics, v. 1, p. 47–62.

Merriam, R., and Bandy, O. L., 1965, Source of Upper Cenozoic sediments in Colorado delta region: J. Sed. Petrology, v. 35, p. 911–916.

Miyamura, S., 1962, Magnitude-frequency relation of earthquakes and its bearing on geotectonics: Proc. Japan Acad., v. 38, p. 27–30.

Mogi, K., 1963, Some discussions on aftershocks, foreshocks, and earthquake swarms—The fracture of a semi-infinite body caused by an inner stress origin and its relation to the earthquake phenomena, 3: Bull. Earthq. Res. Inst., v. 41, p. 615–658.

Pfluke, J. H., and Steppe, J. A., 1973, Magnitude-frequency statistics of small earthquakes from San Francisco Bay to Parkfield: Proc. Confer. Tectonic Problems of San Andreas fault system, A. Nur and R. L. Kovach (eds.), Stanford Univ. Publ. Geol. Sci., v. 13, p. 13–23.

Prothero, W. A., and Brune, J. N., 1971, A suitcase seismic recording system: Bull. SSA, v. 61, p. 1849–1853.

Raleigh, C. B., Healy, J. H., and Bredehoeft, J. D., 1972, Faulting and crustal stress at Rangely, Colorado, *in* Flow and fracture of rocks: Geophys. Monogr. Ser., H. C. Heard, I. Y. Borg, N. L. Carter, and C. B. Raleigh (eds.), AGU, Washington, D. C., v. 16, p. 275–284.

Real, C. R., and Teng, T. L., 1973, Local Richter magnitude and total signal duration in southern California: Bull. SSA, v. 63, p. 1809–1827.

Rex, R. W., 1966, Heat flow in the Imperial Valley of California: (abstract), AGU Trans., v. 47, p. 181.

Richter, C. F., 1958, Elementary seismology: W. H. Freeman and Co., San Francisco, 768, p.

Rogers, T. H., 1965, Geologic map of California, San Diego-El Centro Sheet: California Div. Mines and Geology, Sacramento.

Scholz, C. H., 1968, The frequency-magnitude relation of microfracturing in rock and its relation to earthquakes: Bull. SSA, v. 58, p. 399–415.

Sharp, R. V., 1967, San Jacinto fault zone in the Peninsular Ranges of southern California: Bull. GSA, v. 78, p. 705–730.

Stauder, W., 1962, The focal mechanism of earthquakes: Advances in Geophysics, v. 9, p. 1–76.

Strand, R. G., 1962, Geologic map of California, San Diego-El Centro Sheet, California Div. Mines and Geology, Sacramento.

Sykes, L. R., 1970, Earthquake swarms and sea-floor spreading: J. Geophys. Res., v. 75, p. 6598-6611.

Sykes, L. R., Fletcher, P. J., Armbruster, J., and Davis, J. F., 1972, Tectonic strain release and fluid injection at Dale, N. Y. (ab.): EOS, Trans. AGU, v. 53, p. 524.

Teledyne Geotech, 1972, Geothermal noise survey of the East Mesa area, Imperial Valley, California: Tech. rep. no. 72-19, Garland, Texas, 18 p.

Thatcher, W., and Brune, J. N., 1971, Seismic study of an oceanic ridge earthquake swarm in the Gulf of California: Geophys. J. Roy. Astr., Soc., v. 22, p. 473–489.

Tryggvason, E., 1970, Surface deformation and fault displacement associated with an earthquake swarm in Iceland: J. Geophys. Res., v. 75, p. 4407–4422.

——— 1973, Seismicity, earthquake swarms, and plate boundaries in the Iceland region: Bull. SSA, v. 63, p. 1327–1348.

van de Kamp, P. C., 1973, Holocene continental sedimentation in the Salton Basin, California: A reconnaissance: Bull. GSA, v. 84, p. 827–848.

Ward, P. L., 1972, Microearthquakes: Prospecting tool and possible hazard in the development of geothermal resources: Geothermics, v. 1, p. 3–12.

Ward, P. L., and Björnsson, S., 1971, Microearthquakes, swarms, and the geothermal areas of Iceland: J. Geophys. Res., v. 76, p. 3953–3982.

Ward, P. L., and Jacob, K. H., 1971, Microearthquakes in the Ahuachapan geothermal field, El Salvador, Central America: Science, v. 173, p. 328–330.

Ward, P. L., Palmáson, G., and Drake, C., 1969, Microearthquake survey and the Mid-Atlantic Ridge in Iceland: J. Geophys. Res., v. 74, p. 665–684.

Westphal, W. H., and Lange, A. H., 1966, Local seismic monitoring: Final tech. Rep. SRI Project PHU-5043, Advanced Research Projects Agency, Washington, D. C., 242 p.

Woodring, W. P., 1931, Distribution and age of the Tertiary deposits of the Colorado desert: Carnegie Inst., Wash., Publ. 148, p. 1–25.

Wyss, M., and Lee, W. H. K., 1973, Time variations of the average earthquake magnitude in central California: Proc. Confer. Tectonic problems of San Andreas fault system, A. Nur and R. L. Kovach (eds), Stanford Univ. Publ. Geol. Sci., v. 13, p. 24–42.