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SEISMOLOGICAL INVESTIGATIONS OF VOLCANIC AND TECTONIC PROCESSES IN THE WESTERN GREAT BASIN, NEVADA AND EASTERN CALIFORNIA

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ABSTRACT

The western Great Basin is structurally complex, and seismogenic processes vary within it. A significant increase in seismicity around the "White Mountains seismic gap", north of the rupture zone of the 1872 (M = 8+) Owens Valley earthquake, suggests that the potential for a major earthquake in the gap is greater than it has been for several decades. Seismicity within the "Stillwater gap" of northern Dixie Valley (Wallace, 1978) is very low and earthquakes that do occur there are small and shallow, suggesting that the potential for a large shock there in the near future is low.

Earthquakes around the Adobe Hills Tertiary (ca. 3 m.y.) volcanic center show a tendency for temporal clustering, which appears to be characteristic of geothermally active areas in the Great Basin. Teleseismic P-wave traveltime residuals indicate a low-velocity zone, possibly a zone of partial melting, at shallow depth in the crust.

Since October 1978 a major earthquake swarm has been in progress near Mammoth Lakes, California. Shallow earthquakes of this sequence are distributed in in an irregular-shaped zone extending 30 km in a WNW-ESE direction and about the same distance from north to south. At the north end this zone extends 7 km into Long Valley caldera. Within this shallow zone the crust is undergoing complex brecciation, possibly accompanied by the formation of clusters of dikes or fissures. Events deeper than about 10 km are located in a northerly-trending zone 25 km long and about 5 km wide; mechanisms for these events are consistent with oblique or normal faulting along the Sierran front. Some of the Mammoth Lakes earthquakes are characterized by lack of S-waves at regional stations north of the caldera, and P-waves for the same station-event combinations are deficient in frequencies greater than about 2-3 Hz -- effects that can be explained by one or more shallow magma chambers. In July 1980 spasmodic tremor began to occur in one small area inside the caldera. Taken together with the observation of uplift (Savage and Clark, 1982) and new fumarole activity within the caldera, these effects are interpreted as evidence for adjustments within the cauldron block, caused by stress changes related to the earthquake sequence.

More than 150 P-wave fault-plane solutions for earthquakes in western Nevada and the eastern Sierra Nevada show a systematic change in mechanism with depth. Earthquakes shallower than 6 km are characterized by strike-slip motion, while most events deeper than 9 km have a strong normal-slip component. These observations are consistent with a model of lithospheric extension that involves oblique or normal faulting at mid-crustal depth and formation of vertical fissures or dikes in the shallow crust.

INTRODUCTION

From the geologic literature, active faulting in the western Great Basin appears to be rather evenly distributed over much of the region. For example, Figure 1 is from a photogrammetric study by Slemmons (1967) and shows faults in late Quaternary alluvium, glacial deposits or lake sediments of the playa, Bonneville or Lahonton type. Approximately a thousand faults are shown on the figure, ranging in length from about 1 to more than 100 km, striking from NW to N to NE. In contrast to this picture of relatively even fault distribution in the Nevada region, historic seismicity has been concentrated in the western half of Nevada. Figure 2 shows this seismicity from 1969 to 1978 in western Nevada and eastern California, on a generalized map of late Cenozoic structural features (Wright, 1976).

In the first part of this paper we present a general description of seismicity in the western Great Basin, with emphasis on (1) the Dixie Valley area, where the last major earthquakes in this region occurred in 1954 and where a number of industrial groups have focused considerable effort on geothermal exploration; (2) the Excelsior Mountains area where teleseismic P-residuals indicate the possibility of a magma chamber; and (3) Long Valley caldera, where a sequence of earthquakes and associated magmatic activity has been in progress for almost five years. In the second part of the paper we discuss results of a study (Vetter and Ryall, 1983) that indicates a systematic change in focal mechanism with depth in this region and leads to estimates of maximum and minimum principal stresses to mid-crustal depths.

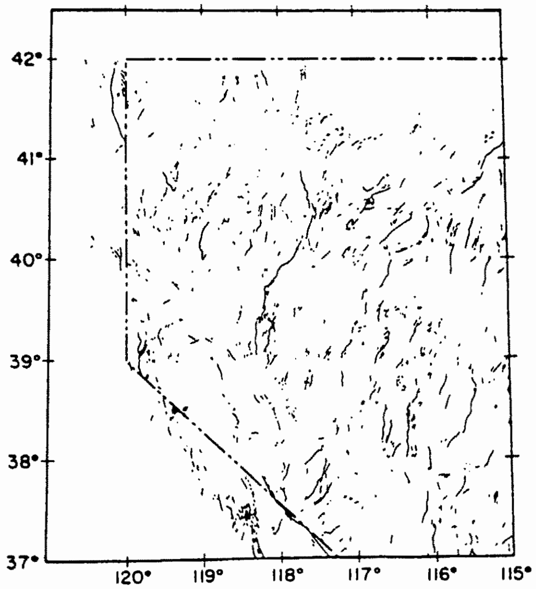

Figure 1. Map of faults that show photogrammetric evidence of Quaternary displacement (adapted from Slemmons, 1967).

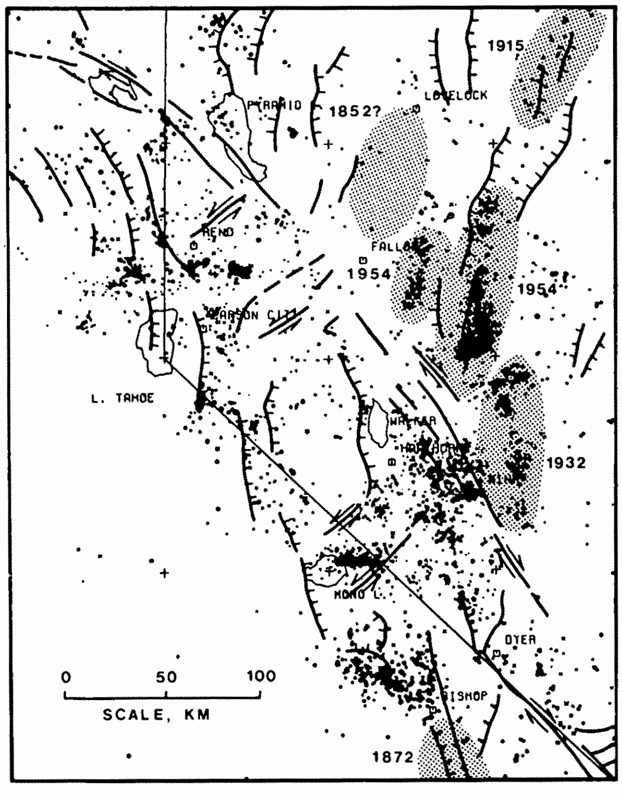

Figure 2. Generalized map of late Cenozoic structural features of the western Great Basin (Wright, 1976), together with epicenters for the period 1969-1978 (dots) and approximate rupture zones of major historic earthquakes (stippled areas with year of main shock). The "Stillwater seismic gap" is between the 1915 and 1954 rupture zones and the "White Mountains gap" is between the 1872 and 1932 zones -- respectively in the upper right and lower right parts of the figure.

SEISMICITY

General Description. From the rupture zone of the M = 8+ Owens Valley earthquake of 1872 (bottom of map in Figure 2), current seismicity spreads to the north through west-central Nevada and to the northwest along the eastern Sierra Nevada, forming two more-or-less separate zones. In the southern part of this region earthquake swarms occur frequently in the area north of Bishop, and a major swarm near Mammoth Lakes has been in progress since 4 October 1978. So far this sequence has produced four earthquakes with M = 6+ and uplift, spasmodic tremor and increased fumarole activity in Long Valley caldera have been associated with probable magma injection (Ryall and Ryall, 1981a, 1983; Savage and Clark, 1982; U. S. Geological Survey, 1982).

As shown by the stippled areas on Figure 2, major earthquakes during the historic period have occurred in a northerly-trending belt from the southern Owens Valley in southeastern California to Winnemucca in north-central Nevada. Wallace (1978) pointed out that three "gaps" within this belt are likely candidates for future large earthquakes. Two of these are the Stillwater gap between the 1954 and 1915 ruptures in the upper right part of Figure 2, and the White Mountains gap between the 1872 and 1932 zones in the lower right part of the figure. The third is the Southern Sierra Nevada gap south of the Owens Valley rupture zone, and will not be discussed in this paper.

In the region around the White Mountains gap since 4 October 1978 more than 50 earthquakes with ML 4.0-5.4 have occurred in the following zones: a NE-trending zone between Mono Lake, California and Luning, Nevada, at the north end and east side of the White Mountains; and in the northern Owens Valley (Figure 3). In addition, more than 100 shocks with ML 4.0-6.3 have occurred in the Mammoth Lakes area. In comparison only 18 events with ML 4.0 or greater were observed in this entire region (including the Mammoth Lakes area) during the previous nine-year period. Thus, the level of moderate seismicity since 1978 is about six times higher than the previous decade if the Mammoth Lakes sequence is excluded, and almost twenty times higher if it is included. The similarity between this pattern and that observed by Mogi (1969) before large earthquakes in Japan suggests that the potential for a major earthquake in the White Mountains gap is substantially higher now than it has been for at least the last two decades (Ryall and Ryall, 1983).

The belt of recent seismicity extends north through the 1932 Cedar Mountains (M = 7.3) rupture zone, the 1954 Dixie Valley-Fairview Peak rupture zone (M = 6.9, 7.1, respectively) and the 1915 Pleasant Valley zone (M = 7.6). A second zone, of frequent moderate earthquakes but no major historic shocks, extends NNW from the Mono basin along the eastern boundary of the Sierra Nevada. In the northern part of this zone near Reno (Figure 4), clusters of earthquakes are observed in the following areas: (1) on the California-Nevada border west of Reno, where an ML 6 event occurred in 1948; (2) north of Truckee, California, where an ML 5.7 shock occurred in 1966; (3) south of Reno in the Steamboat Hot Springs area; and (4) in the Virginia Range SE of Reno.

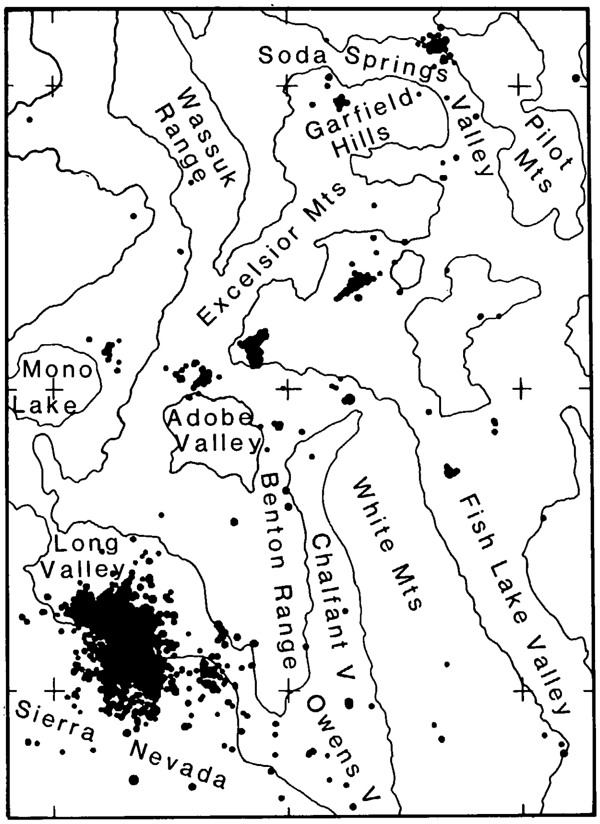

Figure 3. Seismicity north of the 1872 Owens Valley rupture zone, 1978-1982.

Dixie Valley Area. In recent years Dixie Valley, in north-central Nevada, has been the focus of geothermal exploration involving surface and drilling investigations by a number of oil companies. In 1980 and 1981 the University of Nevada Seismological Laboratory operated an eleven-station seismic network in and around northern Dixie Valley, the area of most intensive exploration. Figure 5 shows the seismic network used in the two-year study, together with 1,128 earthquakes analyzed for the period from 1970 to 1981. As the figure indicates, almost all of the seismicity was in a 90 km-long zone extending through southern Dixie Valley and along the east side of Fairview Peak (dense cluster of earthquakes south of station DIX) -- within the rupture zone of two major earthquakes that occurred within four minutes of each other on 16 December 1954 (Slemmons, 1957; Romney, 1957). Detailed analysis of epicenter maps for different periods of time indicates the existence of northerly-trending (NW to NE) fracture zones, with an average trend of about N5-10' E. Since the seismicity is almost entirely confined to the 1954 rupture zone, we conclude that it represents a decaying aftershock sequence. Indeed, a plot of earthquake occurrence in this zone as a function of time (Figure 6) shows a gradual decrease in activity over the 12-year period of observation.

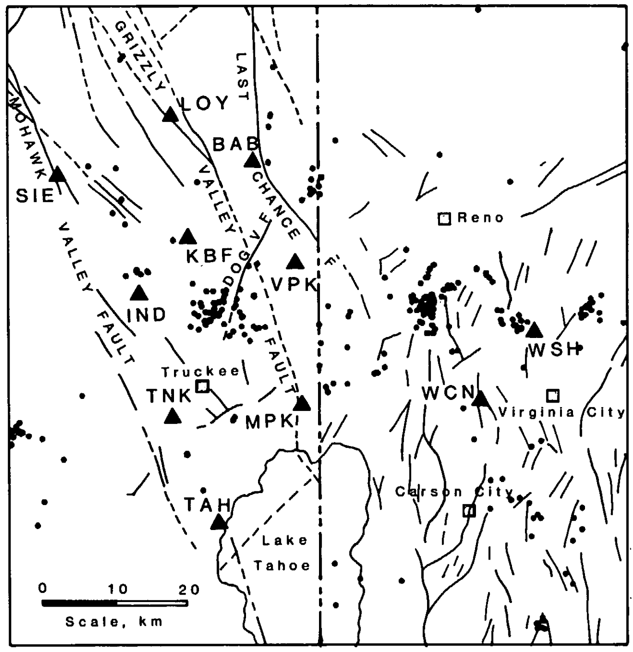

Figure 4. Map of the Reno-Truckee-Lake Tahoe area (39.0° - 39.8°N, 119.5°- 120.5°W), showing seismic stations (triangles), faults, and earthquakes with location quality "C" or better for the period 1975-1982.

A histogram of focal depths for earthquakes in the Fairview Peak-southern Dixie Valley zone is shown on Figure 7. The plot has a peak in the depth range 10-12 km. The few earthquakes that we recorded in northern Dixie Valley all had depth less than 7 km, and all of them were quite small. Meissner and Strehlau (1982) and Sibson (1982) studied earthquake depths on a global scale, based on laboratory tests with quartz-bearing rocks and different assumptions of water content. For a given rock type and water content they predicted the depth of occurrence of earthquakes in a specified region as a function of heat flow. In the western Great Basin the depth distribution of earthquakes agrees well with the theoretical distribution calculated for estimated conditions of water saturation in the crust.

Evidence related to the possibility of a large earthquake in northern Dixie Valley is ambiguous. On the one hand we noted that Wallace (1978) considers that area to represent a seismic gap, with the potential for an M = 7+ earthquake in the near future, and his conclusions are supported by our observation of very low seismicity north of the 1954 rupture zone. On the other, the few earthquakes we recorded in northern Dixie Valley were very shallow and could even be associated with listric faulting in a shallow crustal section overlying a zone in which deformation does not involve brittle fracturing. In a previous study, Richins (1974) noted that earthquakes in NW Nevada, a region characterized by high heat flow and geothermal activity, tend to be shallower and smaller than those in the major earthquake zones of central Nevada. He concluded that the maximum magnitude of

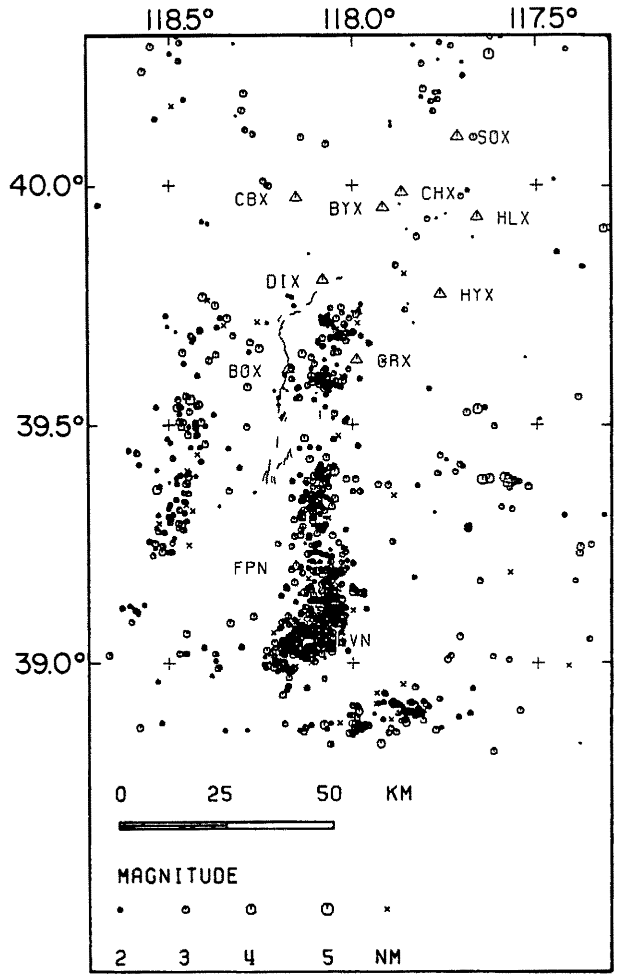

Figure 5. Earthquakes in the Dixie Valley-Fairview Peak area, 1970-1981.

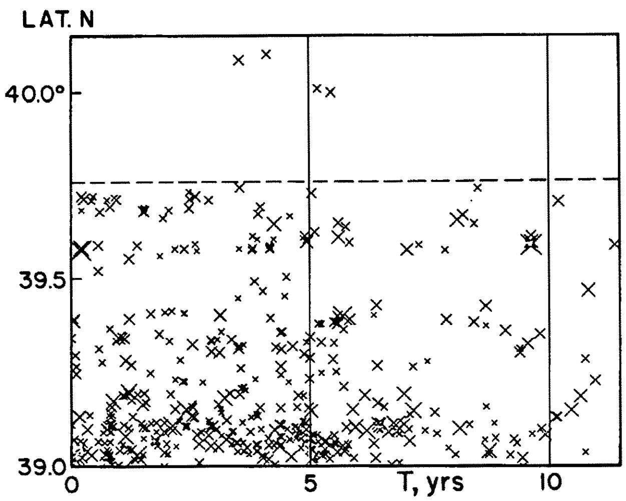

Figure 6. Distribution of earthquakes in the Dixie Valley-Fairview Peak zone as a function of time, 1970-1981. Only events with ML ≥ 2 are shown. Dashed line — northern limit of 1954 rupture zone.

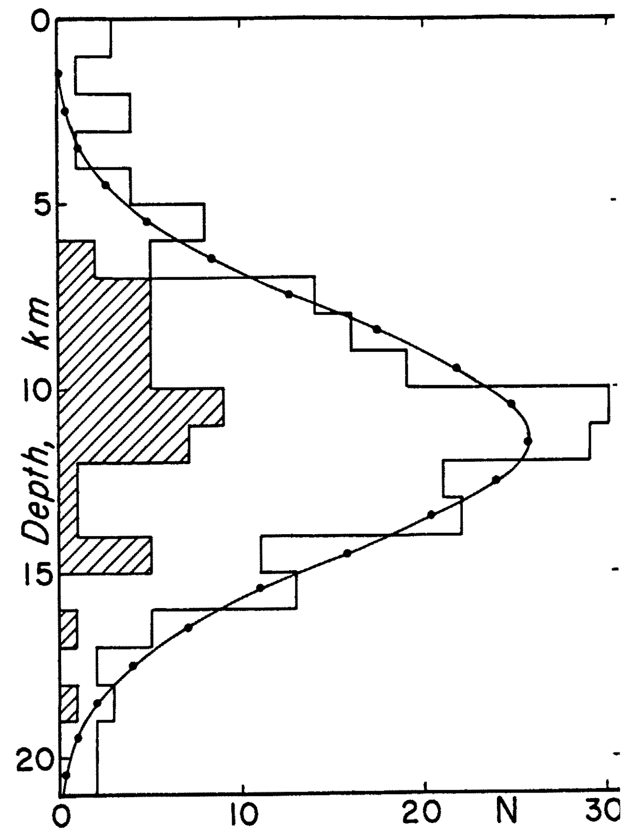

Figure 7. Histogram showing depth distribution fc 217 events with quality "C" or better (unshade area) and 42 events with quality "A" or "B" (shade area). Smooth curve is normal distribution fc mean depth 11.3 km, standard deviation + 3.22 km.

earthquakes in geothermally active regions may b only 5-3/4 to 6, as a result of the weakening c crustal rocks in the vicinity of intrusive bodies or by the effects of stress corrosion and leachin due to geothermal fluids.

To study the stress pattern in the Fairvie Peak-Dixie Valley region, P-wave fault-plane solu tions were determined for eleven earthquakes in th magnitude range 3.0-4.5. These solutions, shown o Figure 8, were based primarily on Pg arrivals, bu clear Pn arrivals were used. We found strike-slip oblique-slip and normal-slip mechanisms, with th axis of minimum compressive stress in all case oriented roughly NW-SE in agreement with the know direction of lithospheric extension in the wester Great Basin. Of particular interest was the obser vation that the deeper events had a strong com ponent of normal slip, while the shallower event were all strike-slip. For the shallow events, th two planes of the fault-plane solution were essen tially vertical, but for the deeper shocks th planes were inclined with dips of 40-60°. Ou interpretation of this change, following a paper b McGarr (1980), was that the change in mechanis could be due to increasing overburden pressure wit depth. This topic is discussed further in the sec tion on focal mechanisms.

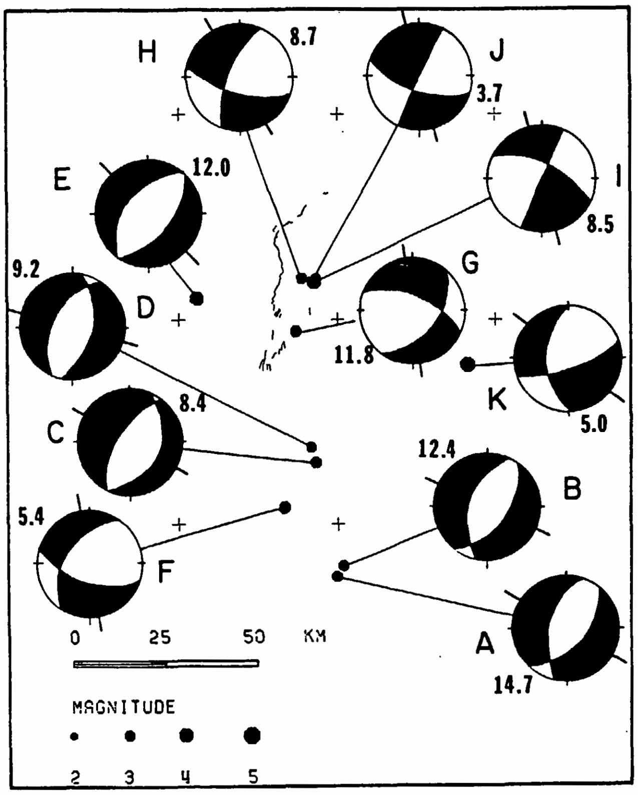

Figure 8. P-wave fault-plane solutions (lower hemisphere, equal-angle projection; compression quadrants shaded) for selected events in the Dixie Valley area. Orientation of T-axis (axis of minimum compressive stress) indicated by heavy line on each mechanism. Numbers indicate depth, km.

<ins>Mono-Excelsior Area</ins>. In the area east of Mono Lake (Figure 2) earthquakes occur in a roughly E-W zone through the Adobe Hills Tertiary volcanic center. Detailed inspection of earthquakes in this area indicates that the E-W zone is made up of a number of short NE-SW segments in an en-echelon arrangement. Gilbert et al. (1968) described the Adobe Hills center as the source of the most voluminous eruptions in the Mono basin during the last 4 m.y., and attributed the eruptions to zones of extension related to left-lateral motion on faults striking about N60°E. According to them, such motion would result in rotation of the blocks between the faults and produce open spaces where north-south faults within the range intersect the transverse faults. Focal mechanisms for this region (Figure 9) are consistent with the observation of predominently left-lateral slip on NE-striking faults by Gilbert et al.

Recent earthquake sequences in this region (Adel, Oregon, 1968; Denio, Nevada, 1973; Mammoth Lakes, California, 1975, 1976, 1978-1983; Mono Basin, California, 1974, 1976, 1978) show a distinct tendency for temporal clustering, which according to Richins (1974) may be characteristic of geothermally active areas in the Great Basin. Another indication of relatively high temperature at mid-crustal depth is the lack of seismic activity at depths greater than about 15 km.

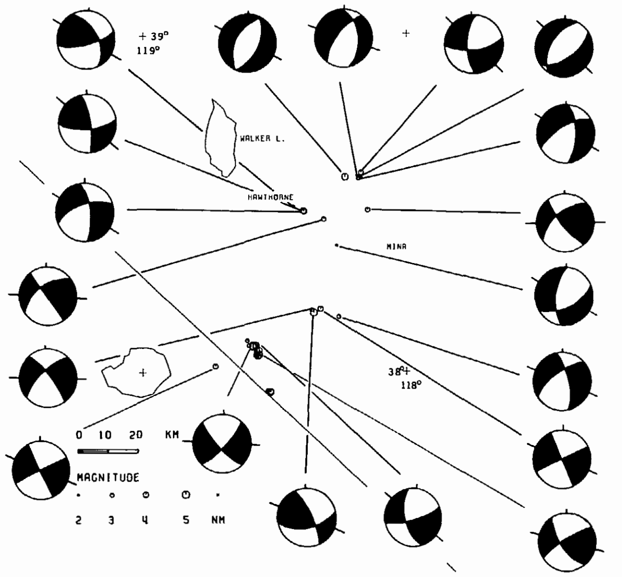

Figure 9. Focal mechanisms for the Excelsior Mountains-Luning area. Town of Luning is 10 km south of event 20 in upper right part of figure.

To investigate the possibility that earthquakes in the Adobe Hills area might be related to a zone of partial melting in the crust, VanWormer and Ryall (1980) analyzed teleseismic P residuals for 22 stations in the area north and east of Mono Lake. These residuals, taken relative to a station (Tonopah) outside the area of interest and corrected for elevation, are shown on Figure 10. For teleseismic sources to the southeast (azimuth 119° to 155°) Figure 10a shows an area of relatively late arrivals east of Mono Lake, elongated to the northeast around the southern edge of the Excelsior Mountains and the eastern side of the Garfield Hills. For sources to the northwest (azimuth 303° to 316°) Figure 10b shows a very similar feature. The amplitude of this traveltime anomaly is about 0.5 second, which is higher than the value of 0.3 second found by Steeples and Iyer (1976) for Long Valley caldera. The size of this this anomaly and the agreement in its location for waves propagating in opposite directions indicates that it is due to shallow structure. However, the plateau between the -0.1 and -0.2 second contours in the upper left part of Figure 10a, and the plateau between the 0.1 and 0.2 second contours in the lower right part of Figure 10b suggest an upper-mantle component of the anomalous zone in this area. Taken together with the volcanic history of the area, the P residuals on Figure 10 would be consistent with a region of partial melting in the shallow crust below the Adobe Hills (centered about 25 km east of Mono Lake), extending to the northeast along a zone of mapped faults and connected to an upper-mantle source below the Excelsior Mountains.

A similar result was obtained by Iyer and Evans (1983) for the Mono and Inyo Craters. Their interpretation of the traveltime anomaly was in terms of a low-velocity body extending from shallow depth to at least 25 km, under most of the volcanic chain.

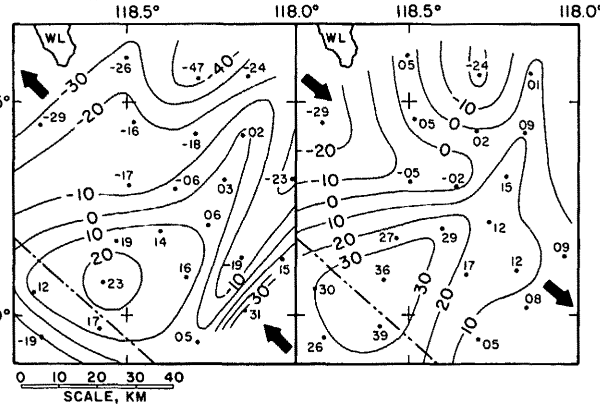

Figure 10. Maps of the area southeast of Walder Lake (WL) showing teleseismic P residuals relative to station TNP. Figure on left is for sources to SE; figure on right is for sources to NW. Arrows show direction of propagation. Numbers are P-wave residuals in hundredths of a second.

Mammoth Lakes Area. Over the last five years, starting in the fall of 1978, a major earthquake swarm has been in progress in the boundary zone between the Sierra Nevada and the Great Basin, near Mammoth Lakes, California. So far this swarm has produced four earthquakes with ML 6 or greater and uplift and spasmodic tremor in Long Valley caldera have been associated with probable magma injection. An Earthquake Hazard Watch was issued by the US Geological Survey for this area in May 1980, and a Volcano Hazard Notice was issued in May 1982. Evidence that led to the prediction of the larger shocks of this sequence is described by Ryall and Ryall (1981b).

Figure 11 shows more than 2,000 earthquakes of the Mammoth Lakes sequence for the period 1978-1982. The epicentral zone for shallow earthquakes of this sequence is irregular in shape, extending more than 30 km in a WNW-ESE direction, about 30 km from north to south, and 7 km into Long Valley caldera. While some lineups of events can be seen, spatial correlations of epicenters with either mapped faults or linear features on Landsat imagery are generally lacking. Instead, the epicentral distribution appears to reflect intense brecciation of the shallow crust. Earthquakes deeper than about 10 km, however, are located in a more restricted zone, 5 km wide and about 25 km from north to south. During this sequence no earthquakes have been recorded in the area between the caldera and Mono Lake.

Six weeks after the ML 6+ shocks in 1980, earthquakes in one small area just east of the town of Mammoth Lakes began to occur as intensive swarms, with a typical swarm lasting 1-2 hours, producing hundreds of microearthquakes and having the appearance of spasmodic tremor (Ryall and Ryall, 1981a; 1983). In volcanic regions spasmodic tremor is considered to represent intensive cracking within the volcanic system, due either to the injection of a tongue of magma or to gas released under high pressure from the magma chamber. In the Mammoth Lakes case, although some of the swarm events have the appearance of very shallow earthquakes (weak P onset, amorphous signature) none have the long-period character associated with magma injection in other areas (e.g., Malone et al., 1983). As a result, while these intensive swarms probably result from magmatic processes within the caldera, it seems less likely that they represent magma injection at very shallow depth. It is interesting to note that swarms with the appearance of spasmodic tremor were not observed prior to the large earthquakes in 1980, suggesting that the swarms reflect adjustments within the cauldron block, caused in turn by stress changes associated with the earthquakes.

In previous work on the configuration of the Long Valley magma chamber, Hill (1976) concluded that secondary arrivals observed on a refraction profile across the caldera could be reflections from the roof of a magma chamber at depth of 7-8 km, and Steeples and Iyer (1976) interpreted a 0.3-second teleseismic P-delay in the west-central part of the caldera as due to anomalously hot rock at depths greater than 7 km and probably less than 25 km. However, Steeples and Iyer also observed that stations inside and outside the caldera recorded teleseismic S-waves and concluded that if true magma was present along the ray paths in question it was either in small pockets or had sufficient viscosity to transmit S-waves. At a US Department of Energy (1980) workshop the participants concluded that low geothermal gradients measured in boreholes in Long Valley caldera suggested a temperature of about 600° C at 15 km depth, and appeared to preclude the existence of magma at depths accessible by drilling.

Figure 11. Mammoth Lakes earthquakes, 1978-1982. Heavy lines show mapped faults and caldera boundary. Town of Mammoth Lakes is shown in SW part of caldera, Lake Crowley is SE of caldera.

During analysis of the Mammoth Lakes earthquakes Ryall and Ryall (1981) noticed that shallow earthquakes around the southwest boundary of the caldera were characterized by lack of S-waves at regional seismic stations to the north, and P-waves for the same station-event pairs were deficient in frequencies greater than about 2-3 Hz. They concluded that these observations were consistent with Hill's (1976) interpretation of a magma chamber at depth of 7-8 km. This study was extended by Sanders and Ryall (1983) based on detailed analysis of more than 200 well-located events of the Mammoth Lakes sequence on recordings of about 30 regional seismic stations. From a comparison of ray paths for signals with anomalous and normal character, they concluded that one or possibly two regions in the upper 13 km of the caldera were responsible for the observed S-wave and high-frequency filtering effects. These regions are shown in map view and cross section on Figures 12 and 13. The analysis suggests that the large magma body in the south-central part of the caldera is relatively massive between depths of about 7 to 13 km. For depths of 5-7 km path effects are less pronounced, suggesting the presence of smaller magma bodies (dikes, sills) alternating with solid rock. From 4.5 to 5 km the attenuating material seems even more diffuse and for depths shallower than 4.5 km paths through the caldera are associated with normal signals. A second magma body in the northwest part of the caldera is less well-defined, but coincides with the reflection tentatively identified by Hill (1976) as evidence for a shallow magma chamber.

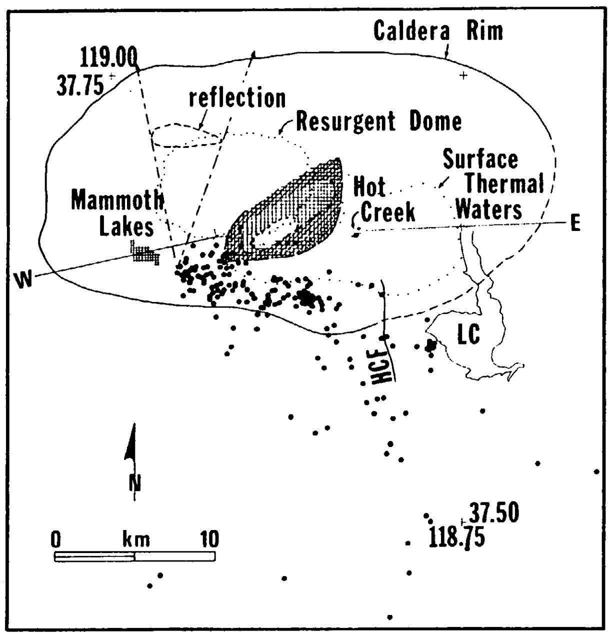

Figure 12. Map showing the location of probable magma bodies in Long Valley caldera. Dots -- epicenters of events used in the analysis; shaded area in central part of caldera -- main magma body; shading patterns correspond to depth ranges shown on Figure 13; dashed lines -- zone in which a second, smaller body may be present in NW part of caldera; dotted lines -- outlines of resurgent dome and hot spring area.

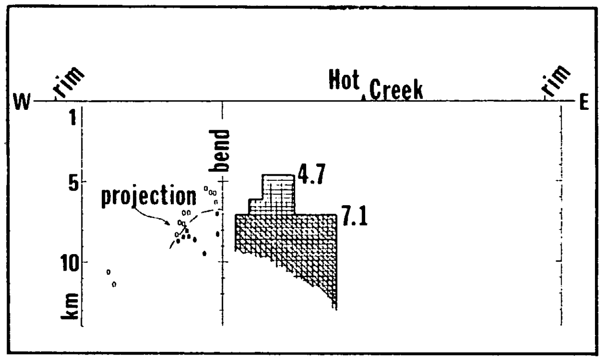

Figure 13. Cross-section along profile W-E on Figure 12, showing approximate outline of magma body in central part of Long Valley caldera. Dashed line -- boundary between normal signals (open circles) and anomalous signals (closed circles) for possible magma body in NW part of caldera. Note that section is bent.

Similar to the pattern observed by Ryall and Vetter (1982) for the Dixie Valley-Fairview Peak zone, P-wave fault-plane solutions for earthquakes in the Mammoth Lakes area show consistent strike-slip mechanisms for depths less than 9 km (Figure 14) and primarily oblique or normal faulting for greater depths (Figure 15). Orientation of the axis of minimum compressive stress (T-axis, Figure 16) for all of the solutions is consistent with crustal extension perpendicular to the NNW-trending Sierra Nevada frontal fault system -- with deeper events resulting from normal or oblique movement on faults striking NNW and dipping east, and shallow events reflecting conjugate right- and left-lateral shear on nearly vertical fractures striking, respectively WNW and NNE (upper right part of Figure 16). Hill (1977) proposed a model in which conjugate shear failures of this type accompany the formation of magma-filled dikes, to explain the predominance of strike-slip mechanisms in volcanic regions prone to earthquake swarms. Comparison of his model with focal mechanisms for the Mammoth Lakes sequence (bottom of Figure 16) suggests that events with strike-slip mechanisms may be associated with the formation of clusters of vertical fissures or dikes at depths less than 9 km.

Taken together, these observations support the suggestion of Lachenbruch and Sass (1978) that lithospheric extension in the Basin and Range province results in a combination of normal faulting and magmatic intrusion of the brittle crust. According to those authors bimodal volcanic centers like Long Valley caldera exist "because they are at places where the lithosphere is pulling apart rapidly, drawing up basalt from below to fill the void." In the case under consideration a major earthquake swarm in the Mammoth Lakes area appears to have started with a complex pattern of strike-slip faulting, possibly associated with the formation of northwest-trending dikes, at shallow depth in a broad area south of Long Valley caldera. This activity reached a crescendo in the spring and summer of 1980, with uplift of the resurgent dome (Savage and Clark, 1982) and the occurrence of several strong, complex ruptures along the caldera and in the crustal block south of it. Following this maximum activity brecciation and possibly associated intrusion of the shallow crust spread rapidly to the south, north and west, with occasional bursts of spasmodic tremor marking an area of rapid crack formation just east of the town of Mammoth Lakes, due either to the injection of a small tongue of magma in the southwest part of the caldera or to gas expelled under high pressure from the shallow magma chamber.

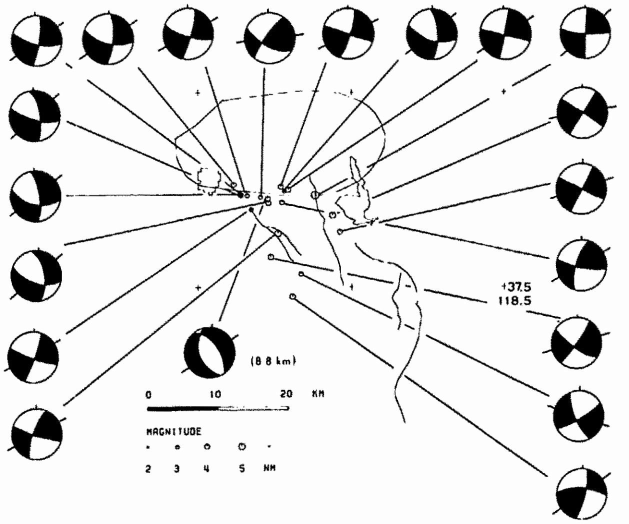

Figure 14. Focal mechanisms for Mammoth Lakes earthquakes shallower than 9 km.

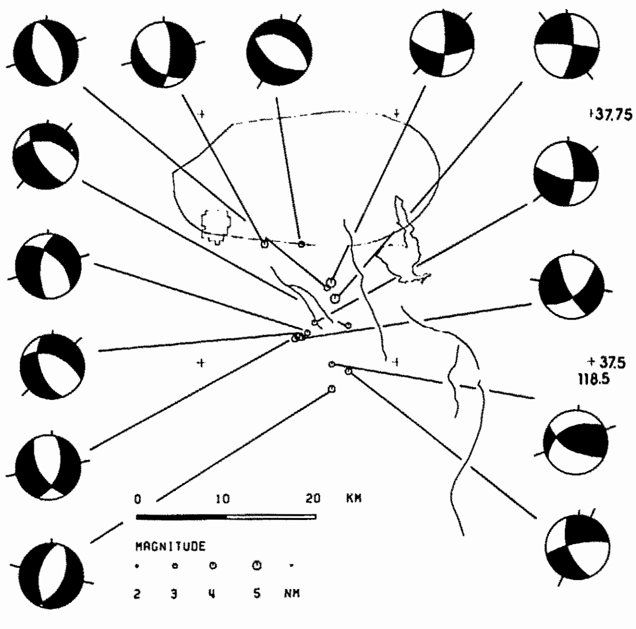

Figure 15. Focal mechanisms for Mammoth Lakes earthquakes deeper than 9 km.

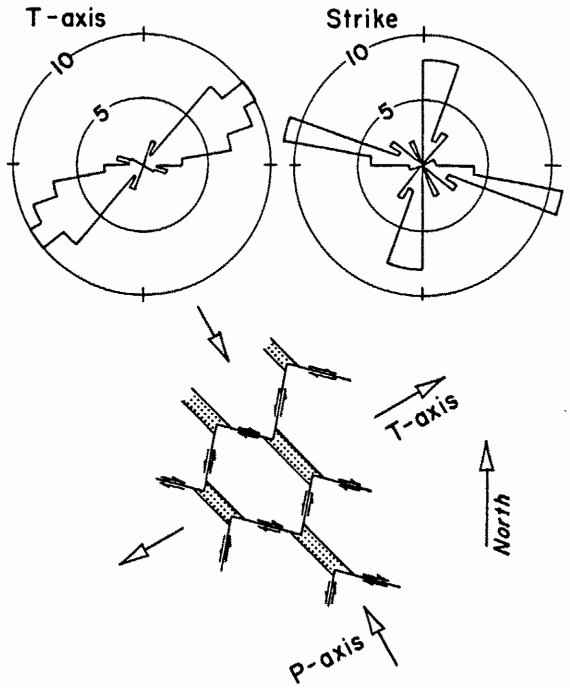

Figure 16. Model to explain strike-slip faulti for shallow Mammoth Lakes earthquakes (adapted fr Hill, 1977). Top left -- rose diagram showi orientation of T-axes; top right — rose diagr showing strike of the two planes of the fault-pla solution; bottom -- model with NW-striking vertic fissures or dikes together with conjugate righ and left-lateral shears on planes shown.

CHANGE OF FOCAL MECHANISM WITH DEPTH AND RELATED STRESS PATTERN

Analysis of P-wave first-motion for more th 150 earthquakes in all active areas covered by c network shows strong indications for a consiste pattern of stress change with depth (Vetter a Ryall, 1983). For the entire region studie earthquakes in the uppermost crust (depth to abo 9 km) are characterized by strike- or oblique-sl mechanisms; purely strike-slip events are strong restricted to the upper 5-6 km. Below 9 km we fi mostly oblique- or normal-slip events. There a some deeper strike-slip earthquakes, but normal-faulting events were found for depths le than about 8.5 km. Similar to the change shown Figures 14 and 15 for the Mammoth Lakes area t deeper events have a significant component of no mal slip for most of the areas studied.

This change in mechanism with depth is we illustrated by Figure 17, which shows fault-pla solutions for two Mammoth Lakes earthquakes th occurred nine minutes apart in time and had almo identical epicenters but different depths. T first, with depth 8.2 km, had a strike-slip mecha ism and the second, at depth 14.2 km, was oblic with a strong normal-slip component.

An exception to this pattern is observed for the Mono Basin-Excelsior Mountains, where primarily strike-slip earthquakes are found, even though a third of the events have depth in the 9-12 km range. This is illustrated by Figure 9, where the only mechanisms with a strong component of normal slip are located north of the Excelsior Mountains.

The observed change in mechanism with depth over much of the western Great Basin can be explained by increasing overburden pressure with depth (Vetter and Ryall, 1983). At shallow depth the predominance of strike-slip mechanisms requires that the greatest principal stress (S_1) is horizontal, while the occurrence of normal faulting at mid-crustal depths requires that S_1 be vertical. We assume that the vertical stress is the overburden pressure, $S_V = \rho g z$, where the mean crustal density $\rho \cong 2.7$ gm/cm .

The observation of changing focal mechanism with depth can be used to estimate the values of the principal stresses as a function of depth in the region, shown on Figure 18. On the figure, P_o is the pore pressure, assumed to be hydrostatic. The least principal stress $S_3 = S_h$ is horizontal at all depths considered, since S_3 represents the axis of extension.

The mean depth of all the strike-slip earthquakes found in our study was about 5 km; for oblique-slip events with about equal strike- and normal-slip components the mean depth is 10.6 km; and for normal faulting it is about 13 km. From this observation we conclude that at about 10.6 km the lines S and S must cross (point 3 on Figure 18). The occurrence of both strike- and oblique-slip earthquakes deeper than about 7 km, and oblique- and normal-slip events deeper than about 9 km indicates that below 7 km the difference between the intermediate and greatest principal stresses is not great, and that local stress variations may determine which type of mechanism occurs. Zoback and Zoback (1980a, b) proposed previously that the intermediate and maximum principal stresses were about equal in the western Great Basin.

Measurements of maximum shear stress $\tau_{max} = (S_1 - S_3)/2$ in the upper few kilometers of the crust have been interpreted by Haimson (1977), McGarr and Gay (1978), McGarr (1980) and Zoback et al. (1977). These papers indicated that shear stress increases approximately linearly with depth, and is lower in soft rock than in hard rock. For the latter the stress change with depth follows a regression line

$$\tau_{max} \approx 5.67 + 6.37\,z\ (MPa), \qquad (1)$$

while a regression line through all the data points measured in an extensional regime is

$$\tau_{max} \approx 1 + 6.7\,z\ (MPa) \qquad (1a)$$

(both curves from McGarr, 1980).

The value of $(S_1 - S_3)$ from equation 1 is shown on Figure 18, extrapolated to mid-crustal depth. The difference between equations (1) and (1a) is greatest at shallow depth where we find strike-slip earthquakes, but it is small at greater depth where oblique- and normal-slip events predominate.

From equation (1) with $S_1 = S_V$ (normal faulting regime) we estimate S_3 at 10.6 and 20 km depth to be about 135 and 260 MPa, respectively (points 4 and 5 on Figure 18). At shallower depth (for a strike-slip regime with $S_H > S_V > S_h$) we can estimate S_1 and S_3 only for the special case where

$$S_V = S_2 \approx \frac{S_3 + S_1}{2} \qquad (2)$$

For a depth of 3 km this relationship gives stress values of 105 and 55 MPa (points 6 and 7 on Figure 18).

For the normal faulting regime at depth $\gtrsim$ 10 km we also use Byerlee's (1977) equation

$$S_3 = \frac{(S_1 - P_0)}{[(\mu^2 + 1)^{1/2} + \mu]^2} + P_0 \qquad (3)$$

with P_o = pore pressure and μ = the coefficient of friction. μ is experimentally determined by Byerlee to be about 0.85 for normal stress less than 200 MPa, representing the upper 6-8 km of the crust, and about 0.6 for normal stress in the range 200-2000 MPa, representing the deeper crust. P is assumed to be hydrostatic.

If we assume, then, that the horizontal stresses increase linearly with depth we obtain from equation (3) that the change in S_h with depth is about 13.4 MPa/km (μ = 0.85), or 15 MPa/km (μ = 0.6), in both cases for the minimum horizontal stress.

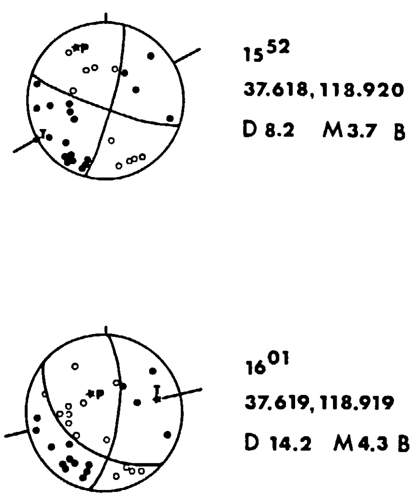

Figure 17. Focal mechanisms for two Mammoth Lakes earthquakes on 9 August 1981. D -- depth, km; M -- magnitude ML; B -- quality.

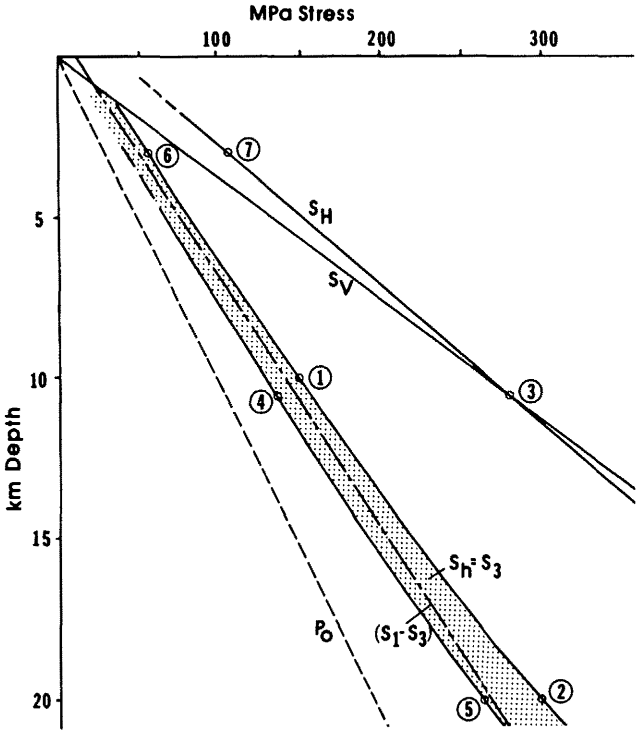

Figure 18. Estimates of the principal stresses as a function of depth. Symbols explained in the text.

For the maximum horizontal stress S_H also increasing linearly with depth we can write (McGarr, personal communication, 1983)

$$S_H \approx A + (\rho g - \delta) z \qquad (4)$$

where A and o are constants. The constant A can be determined at the surface, where for strike-slip faulting $\tau_m = (S \;\; - S \;\;)/2$. According to equations (1a) and (1) we calculate τ_m at depth 0 km as 1 and 5.7 MPa, respectively. These give values for A of 2 and 11.4 MPa, respectively. At depth 10.6 km where the S_H and S_V curves cross, $S_V = S_H$ and we obtain a value of 0.19 MPa/km for the constant δ with A = 2 MPa; with A = 11.4 MPa we find δ = 1.1 MPa/km.

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