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MELT ZONES BENEATH FIVE VOLCANIC COMPLEXES IN CALIFORNIA: AN ASSESSMENT OF SHALLOW MAGMA OCCURRENCES

N.E. Goldstein and S. Flexser

December 1984

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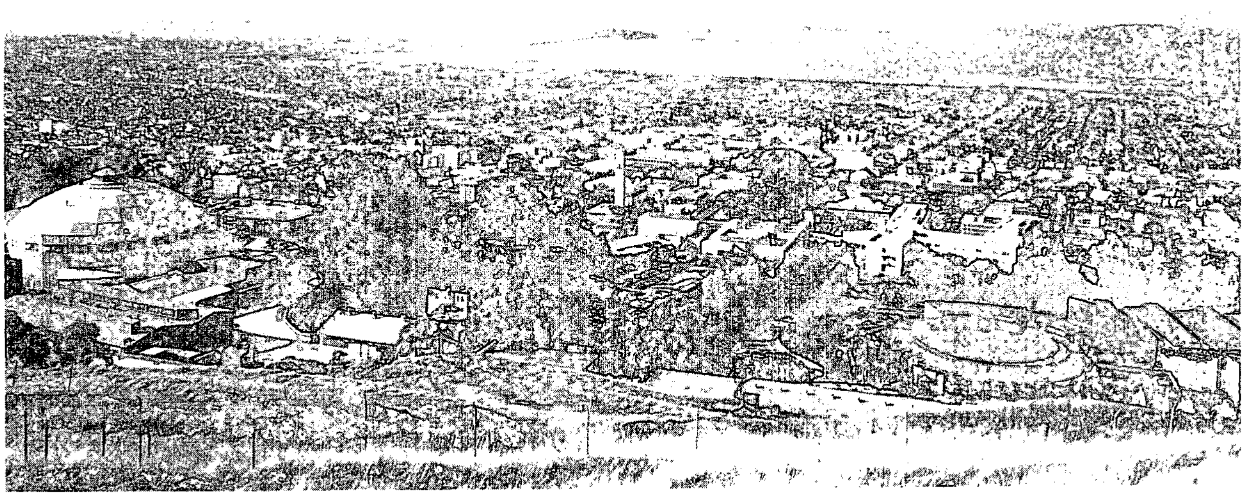

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MELT ZONES BENEATH FIVE VOLCANIC COMPLEXES IN CALIFORNIA: AN ASSESSMENT OF SHALLOW MAGMA OCCURRENCES

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ABSTRACT

Recent geological and geophysical data for five magma-hydrothermal systems were studied for the purpose of developing estimates for the depth, volume and location of magma beneath each area. The areas studies were (1) Salton Trough, (2) The Geysers-Clear Lake, (3) Long Valley caldera, (4) Coso volcanic field, and (5) Medicine Lake volcano, all located in California and all selected on the basis of recent volcanic activity and published indications of crustal melt zones.

INTRODUCTION

An evaluation of recent geological and geophysical data for five magma-hydrothermal systems was made to assess the shallowest depth at which partially or completely molten rocks might be encountered by a deep drill hole. The areas studied were (1) Salton Trough (Imperial Valley), (2) The Geysers-Clear Lake, (3) Long Valley caldera, (4) Coso volcanic field, and (5) Medicine Lake volcano, all located in California (Fig. 1). These areas were selected for this study on the basis of earlier published data, which indicate evidence for a magma body or melt zone beneath each area. Recent information has permitted us to develop more refined estimates as to the probable depths and configurations of the melt zones.

For each of the areas discussed in this report, the geological and geophysical data are presented separately, preceded by a unifying summary of conclusions. As this report is intended to serve as a bibliographic reference as well as an evaluation of magma characteristics, the listing of references

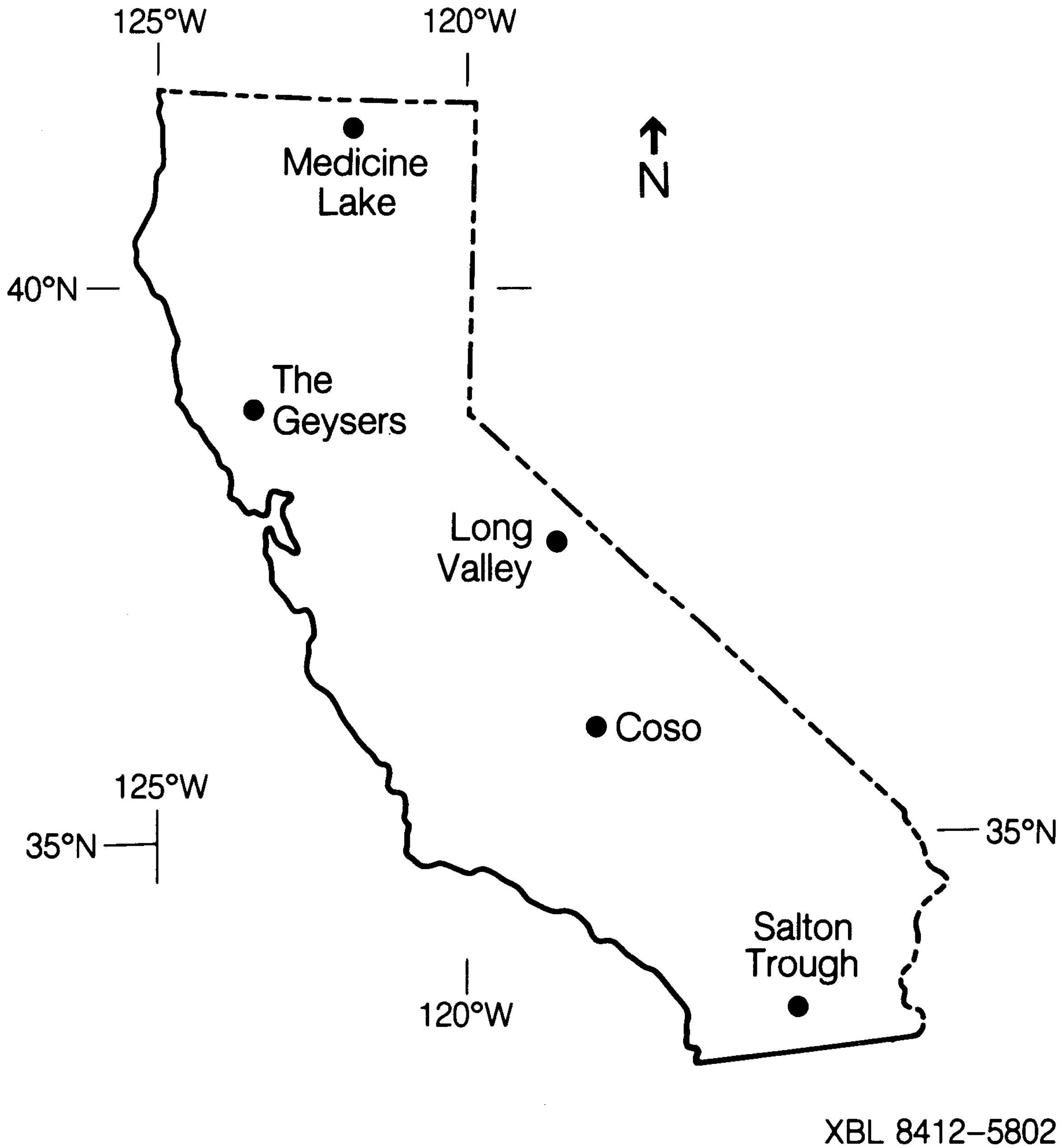

Fig. 1. Index map showing the five volcanic centers discussed in this study.

cited in the text has been combined with a bibliographical listing for each of the five areas discussed. An additional general bibliographic and reference listing, containing items relevent to more than one of the five areas, has also been included.

The areas discussed in this report are all recently active volcanic centers with several features in common. In each area, the most recent eruptive phases are predominantly bimodal assemblages of rhyolite and basalt. Pyroclastic eruptions are of subordinate importance in all the areas with the exception of the Long Valley caldera, and there voluminous pyroclastic eruptions represent an earlier phase (~1 m.y.b.p.) of the caldera's evolution. Each of the areas is also characterized by an extensional tectonic environment.

The occurrence of melt zones in the crust may take a variety of forms, reflecting the amount and rate of heat input and the tectonic environment, among other factors. The schematic diagram in Fig. 2 (reproduced from Hildreth, 1981) is included as a guide to conceptualizing possible melt zones and their associated volcanic and tectonic settings. For example, in terms of the areas discussed in this report, the Salton Trough probably corresponds best to model A; Coso may be intermediate between A and B; and Long Valley may be closest to model B.

A framework for estimating the dimensions and heat contents of magma chambers underlying recently-active silicic volcanic centers was provided by Smith and Shaw (1975, 1978), and their figures serve as benchmarks against which other estimates can be compared. In their model, magma chamber area is based mainly on vent distribution or, if a caldera is present, on caldera area. (In some cases, volume of pyroclastic ejecta is also considered.)

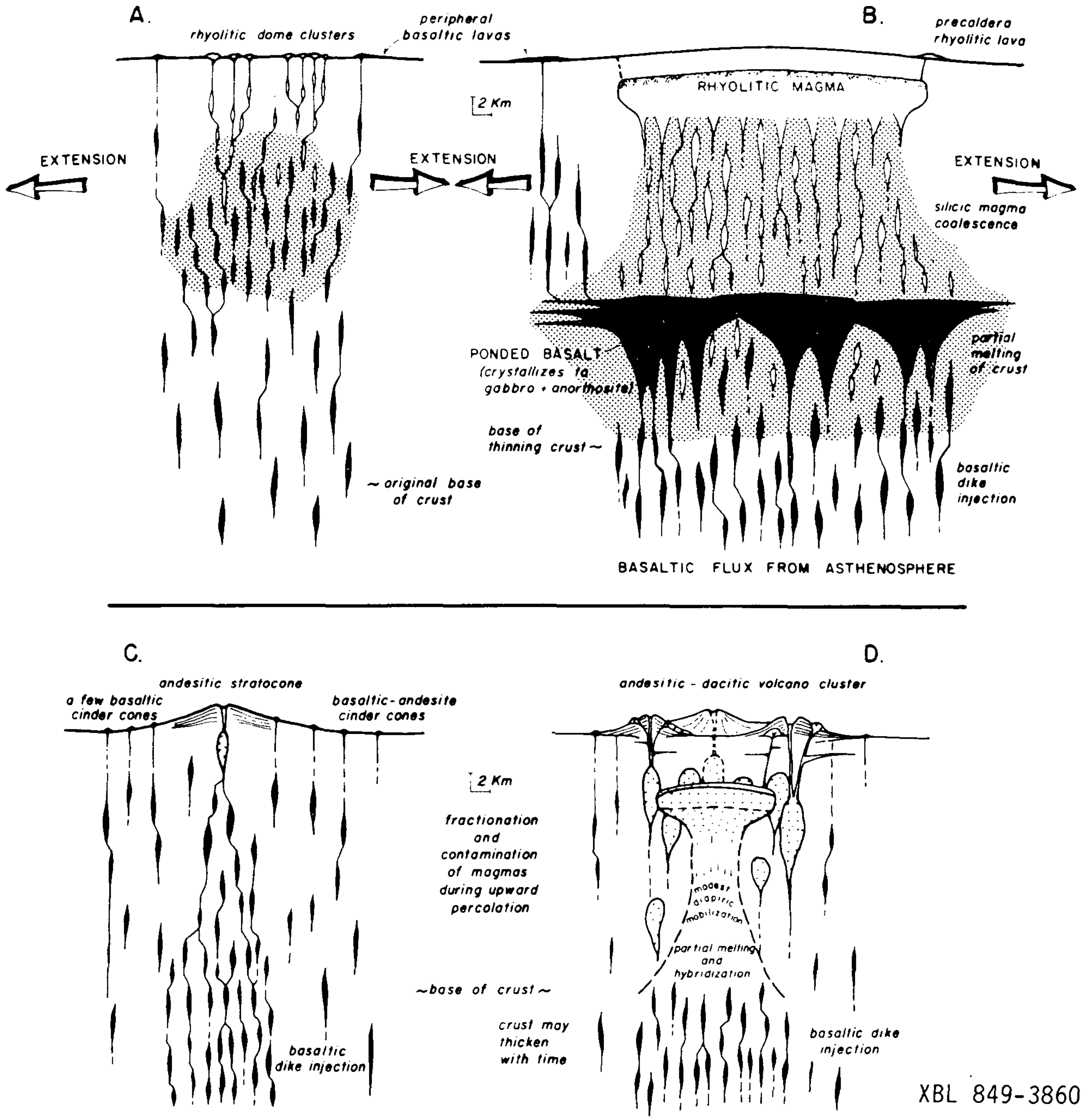

Fig. 2. Some contrasting styles of lithospheric magmatism. Upper pair of cartoons depicts rhyolitic-basaltic magmatism under stress conditions favoring marked crustal extension: (A) modest power input; (B) large power input, advanced stage. Shaded regions indicate partial melting of crustal rocks sufficient to permit separation of rhyolitic magmas as gash veins and dikes. Lower pair of cartoons depicts two possible stages in development of volcanic systems where tectonic extension, if any, is subordinate and shallow: (C) early stage; (D) intermediate stage. This model applies to island arc, continental margin arc, and continental interior systems that produce abundant intermediate magmas. All four sketches are idealized and refer to no particular systems. The models are independent of the mode or site of generation of basaltic magma, but basalt is thought to provide the power supply for virtually all other magmatism. (From Hildreth, 1981.)

Chamber volume is then estimated by choosing a figure for chamber thickness, in the range of 2.5 to 10 km, that is scaled to the magnitude of the area relative to other systems. This approach allows significant uncertainty in the volume estimates, an uncertainty described by Smith (1979) as one order of magnitude to "half an order of magnitude if data on area and ejecta volume are reliable." In cases where estimates of area are uncertain, further error may be introduced. For example, where the distribution of volcanic vents is influenced greatly by fault-controlled dikes (as at Coso), vent distribution may be a poor guide to chamber area. In addition, the model may not be applicable to volcanic systems in which a large, long-lived magma chamber is absent and eruptions are fed by dispersed, short-lived magma bodies (as is likely at the Salton Trough). For these systems, it is necessary to apply the estimates of Smith and Shaw with caution.

In this report we have attempted to refine prior estimates of magma chamber volumes for the five areas studied, and to provide estimates of depths to magma, based on available geological and geophysical evidence. The types of geological evidence best suited to interpreting the location, configuration, and longevity of crustal melt zones are the eruptive histories (ages, volumes, compositions) of the recent volcanics, and the structural, chemical, and hydrological data associated with them. Radiometric dating is invaluable in deciphering eruptive histories, but the commonly used K-Ar method is frequently not reliable for dating very young eruptions. In those cases, ^{14}C dating, obsidian hydration rind dating, or geomorphic criteria have been useful. Structural data can be valuable in interpreting the geometry and possibly the depth of a melt zone; for instance, where a large, shallow magma chamber is present and its configuration is reflected in surface features, such as arcuate patterns of fractures or volcanic vents. Chemical data are essential in interpreting the origin and evolution of magma, and they too are especially useful where a continuously-evolving, stable chamber is present. The most problematical aspect of this study from the geological perspective is estimating quantitatively the depth to magma. In rare cases, this is possible where pressure-dependent geological phenomena or processes can be discerned. For example, pressure-dependent phenocryst equilibria may be present (as observed in Long Valley); or vesiculation during mixing of two magmas (as observed at Medicine Lake) may allow limits to be placed on the depth at which mixing occurred.

Geophysical evidence most applicable for inferring the depth to melting, as well as the location and dimensions of a crustal melt zone, are conductive thermal gradients in deep wells, thermal models, and seismic velocity and/or attenuation anomalies. In the areas studied, most of the deep drilling has been or is being done for geothermal exploration, and these holes are generally shallow (<6000 feet in most cases). Unfortunately, the data from the recent wells are still proprietary. The deep conductive thermal gradient at the Salton Sea geothermal field is better known than that in any other of the five areas studied, since temperature data are available from deeper wells there (and will be augmented when the Salton Sea Scientific Drill Hole is drilled in 1985).

Indirect geophysical evidence for the depth to and location of melt zones has been derived mainly from seismological studies. The maximum depth of earthquake activity is assumed to be the depth at which rocks cease exhibiting brittle behavior, and thus provides a lower limit on the depth to melt. Other important seismological evidence is derived from seismic wave velocity and attenuation anomalies. Using information from regional and teleseismic events, geophysicists have applied ray tracing and inversion techniques to discern zones of shear-wave attenuation and velocity decreases that may be caused by partial melt beneath several of the areas studied in this report.

Other indirect information on deep thermal conditions has been sought from electromagnetic, electric, gravity and magnetic surveys. However, as all these methods are strongly influenced by near-surface inhomogeneities and/or may have a limited depth of investigation (e.g., electromagnetic-electric), it is a difficult process to separate the effects due to shallower features from those that might indicate the thermal state at depth. Curie isotherm analyses of aeromagnetic data have been attempted in a few areas but the analysis can lead to improbable results (Salton Sea); and thus the results by themselves are not generally considered to be a reliable guide to subsurface temperature. Gravity lows that might suggest the presence of a large volume of low-density silicic melt have not been observed in any of the areas studied, with the notable exception of The Geysers. Even at The Geysers other geophysical evidence casts serious doubt on whether the gravity anomaly there is in fact due to a large volume of silicic melt. Deep probing electromagnetic methods are believed by many to be a promising means for discerning large, highly conductive melt zones within a resistive host. However, conductivity inhomogeneities near the surface strongly affect the data and create a major interpretation problem. In addition, conductive zones at depths of 2 to 20 km may result from sources other than magma bodies (e.g., at Long Valley). Deep electromagnetic methods may thus provide permissive evidence for a magma body at depth, but are not definitive.

SUMMARY AND CONCLUSIONS

On the basis of the geological and geophysical data reviewed and discussed in this report, we have attempted to place numbers on the probable depth to melt beneath the five areas studied. The best estimates of the minimum depth to a magma chamber or other melt zone for each area are given in Table 1, along with the criteria used for the estimates.

There remains enough uncertainty in these depth estimates that any or all could have an error of at least 20 percent. Magma chamber dimensions given by Smith and Shaw (1975) are listed and compared in Table 1 to our analyses of the data.

The shallowest melt zone, based on information available in the open literature, unpublished technical reports, and papers in preparation, is judged to occur beneath the Salton Sea Geothermal Field, located at the south shore of the Salton Sea. Our estimate of >6 km depth to melt agrees with the depth to melting of water-saturated granite as suggested by Elders et al. (1972). Chemical data on the rhyolite domes do not suggest the presence of a large, continuous magma chamber beneath the Salton Buttes. Small, discontinuous high-level silicic partial melt zone(s) are more likely to be found. It is possible that the Salton Sea Scientific Drill Hole, to be drilled in 1985, will provide important new information on thermal conditions and magma generation for the area.

There is abundant geophysical evidence for silicic magma at depths of 6 to 8 km beneath the resurgent dome and the western sector of the Long Valley caldera. Although the Long Valley caldera is one of the best studied volcanic complexes in the U.S., scientists are not in total agreement as to the location of the shallowest melt. Additional intermediate-depth drilling

TABLE 1. Estimates of minimum depths and probable volumes of melt zones.

SITE	ESTIMATED DEPTH (km) AND LOCATION OF MAGMA	CRITERIA	SILICIC MAGMA VOLUME (km^3) (from Smith & Shaw, 1975)	COMMENTS ON MAGMA VOLUME
SALTON SEA	>6.	Thermal.	200	Probably much smaller, dispersed bodies.
LONG VALLEY	>5 and probably >7; beneath area enclosed by resurgent dome.	S-wave attenuation. P-wave velocity. Seismic reflection-refraction. Crustal deformation.	2400	Probably significantly smaller unless connected to Mono Craters chamber.
THE GEYSERS-CLEAR LAKE	>6 and probably >7.	Gravity and P-wave velocity. Magma not corroborated by seismic reflection or electromagnetic soundings.	1500	Probably smaller, dispersed bodies.
COSO	>8 and probably 10 to 12; beneath center of rhyolite dome field.	Earthquake focal depths. P-wave attenuation. Thermal and P-wave velocity suggest shallower source, >5 km.	650	Smaller chamber, perhaps 200 to 300 km^3.
MEDICINE LAKE	>7, possibly <10; beneath central caldera.	Petrologic.	300	Significantly smaller bodies, possibly dispersed.

and more detailed seismic reflection and electrical studies are recommended to help define a target for deep drilling. Some of this work is planned or in progress by the U.S. Geological Survey, National Laboratories and university groups.

On the basis of geophysical anomalies (gravity, magnetics, magnetotellurics and teleseismic P-wave delays) a great deal of attention has focused on The Geysers-Clear Lake area as a place to find a shallow silicic magma. However, recent reflection seismic and deep sounding electrical survey data have not confirmed the original hypothesis of a large silicic magma chamber beneath Mount Hannah. The gravity, magnetic and P-wave velocity anomalies could be caused by a complex combination of geological features not related to magma. Geological data are largely inconclusive, and do not support a large magma chamber. However, magma may occur as smaller, discontinuous volumes to the northeast of Mt. Hannah, near Clear Lake and beneath the area of the most recent eruptions.

The Coso volcanic field and the Medicine Lake Volcano have been less thoroughly investigated than the other three areas. Without question, both areas need to be studied in more detail before one could feel confident about intersecting a magma body at a reasonable depth. On the basis of the eruptive histories, eruptive volumes, geochemical studies and indirect geophysical information, there is much stronger evidence for a sizeable magma chamber beneath the Coso volcanic field than beneath Medicine Lake. However, considerably more drilling and geophysical investigations have been done at the Coso volcanic field than at the Medicine Lake volcano. The Coso chamber is probably deep (>10 km), and although the Medicine Lake chamber might be shallower, it is still likely to be ~10 km in depth.

SALTON SEA-IMPERIAL VALLEY

SUMMARY

The strongest direct indications of magma beneath the Salton Sea geothermal field (SSGF) are the young age of the rhyolite domes on the south shore of the Salton Sea, and high subsurface temperatures encountered in several deep geothermal exploration wells. Neither geological nor geophysical data support the concept that a large, high-level magma chamber fed the rhyolite domes at the Salton Buttes. It appears more likely, instead, that the domes erupted from small, dispersed bodies of silicic melt.

GEOLOGICAL SUMMARY AND EVIDENCE FOR MAGMA

Chronology, Composition, and Magnitude of Volcanism

The only extrusive rocks present, the Salton Buttes, are five small-alkali rhyolite domes aligned along a 7-km northeast-trending line at the south shore of the Salton Sea (Fig. 3) (Robinson et al., 1976). The cumulative volume of the domes is ~2 km^3. Associated flows and pyroclastics are minor, and eruptions were subaqueous. The sole published K-Ar age date obtained on the domes is ~16,000 years. Although there may be a large error in this date, the domes are surely <50,000 years old (Muffler and White, 1969). Hydration-rind ages of 8400, 6700, and 2500 years were obtained on two of the domes (Friedman and Obradovich, 1981).

Abundant basaltic and silicic dikes have been intersected by drill holes at 1 to 2 km depth. Basalts are tholeiitic, typical of East Pacific Rise basalts and those encountered elsewhere in Salton Trough drill holes (e.g., Cerro Prieto). Silicic dikes, like the rhyolite domes, have primitive isotopic ratios and resemble the rhyolites at islands of the East Pacific Rise

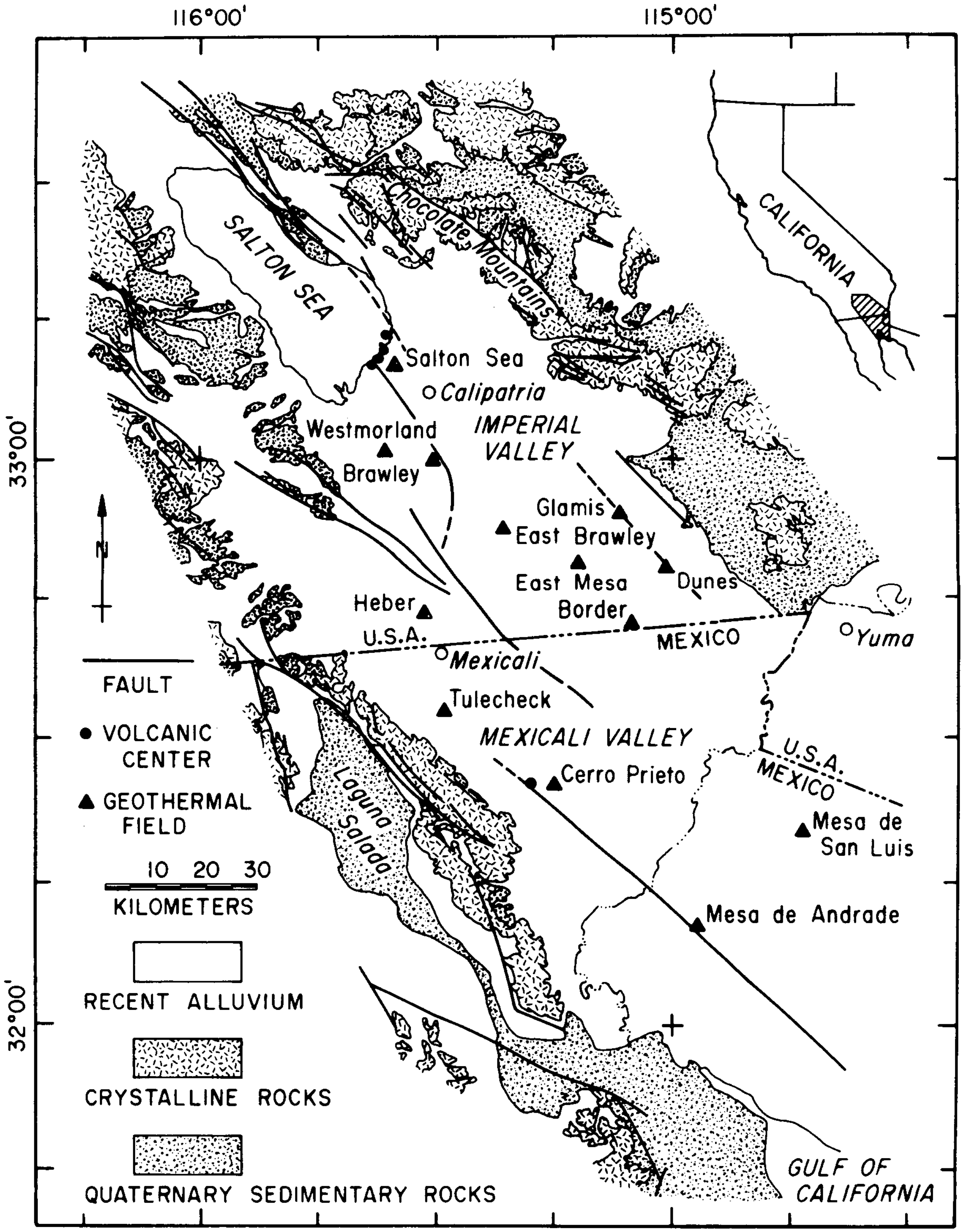

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Fig. 3. Geothermal fields and volcanic centers of the Salton Trough. (From Elders and Cohen, 1983.)

(Robinson et al., 1976). The composition of the igneous rock suite as a whole (domes, dikes, and xenoliths) is distinctly bimodal (basaltic and rhyolitic), as is characteristic of magmatism in extensional environments (Martin and Piwinski, 1972).

Structural and Tectonic Setting

The Salton Trough is a sediment-filled rift valley, tectonically active since at least Pliocene time, in which sedimentation has almost kept pace with tectonism. Formation of the Colorado River delta perpendicular to the rift has isolated the Salton Basin from the Gulf of California, creating a closed basin approximately 200 by 90 km (Elders et al., 1972). Within the basin are numerous areas of high heat flow; the Salton Sea and Cerro Prieto areas are the two foremost on land and the only ones with silicic (rhyolite to dacite) eruptives. Both areas are associated with successive en echelon pairs of northwest-trending strike-slip faults between which crustal spreading has created rhomb grabens or rhombochasms (Mann et al., 1983). These structural features, as well as the high heat flow, are strongly masked by sedimentation. It is believed that dike injection into these extensional basins is responsible for the high heat flow and the geothermal systems delineated by drilling. The five rhyolite domes at the Salton Sea field may be the surface expression of the "leaky" transform zone along which magmas ascended (Robinson et al., 1976).

Hydrothermal System

Thermal manifestations are scarce in the Salton Trough because of the thick sedimentary cover, an effective caprock, and cold groundwater underflow from the Colorado River. Although the area around the Salton Buttes is characterized by high heat flow, surface manifestations at the domes are confined to warm ground and hydrothermal alteration on Mullet Island (the northeasternmost dome) and to adjacent hot springs aligned along a northwest-trending lineament and covered by the Salton Sea (Muffler and White, 1969; Robinson et al., 1976). Temperature logs indicate a temperature of 300°C at 4000 ft, and up to 360°C at 6000 to 7000 ft, below the hottest area (Fig. 4).

Thermal manifestations at Cerro Prieto, Mexicali Valley, Baja California were numerous prior to the commencement of geothermal production in 1973. A large area of mud boils, mud volcanoes, hot springs, and steaming ground have all but dried up. These manifestations occur in a discharge area displaced laterally several kilometers from the main thermal zone (Lippmann et al., 1984).

Evidence Relating to Magma Origin and Magma Chamber Presence

Chemical data for the Salton Sea rhyolite domes show they could not have formed predominantly by partial melting of granitic basement rock (Robinson et al., 1976). Bulk composition and $^{87}Sr/^{86}Sr$ ratios suggest an origin by fractionation of a more primitive mantle-derived basaltic parent (Elders et al., 1972). One possible model for such an origin, which could also explain the close association of basaltic and rhyolitic rocks and the strictly bimodal nature of the suite, has been discussed by Yoder (1973). In this model rhyolitic and basaltic magmas are generated in separate pulses by partial melting in the mantle of a single quartz-normative parent. Some contamination of the rhyolites by crustal material has occurred by assimilation of granitic xenoliths, but Robinson et al. (1976) estimate that at most 20% of the rhyolite could be accounted for in this way. This argues against the possibility of a large, high-level silicic magma chamber beneath the Salton Buttes.

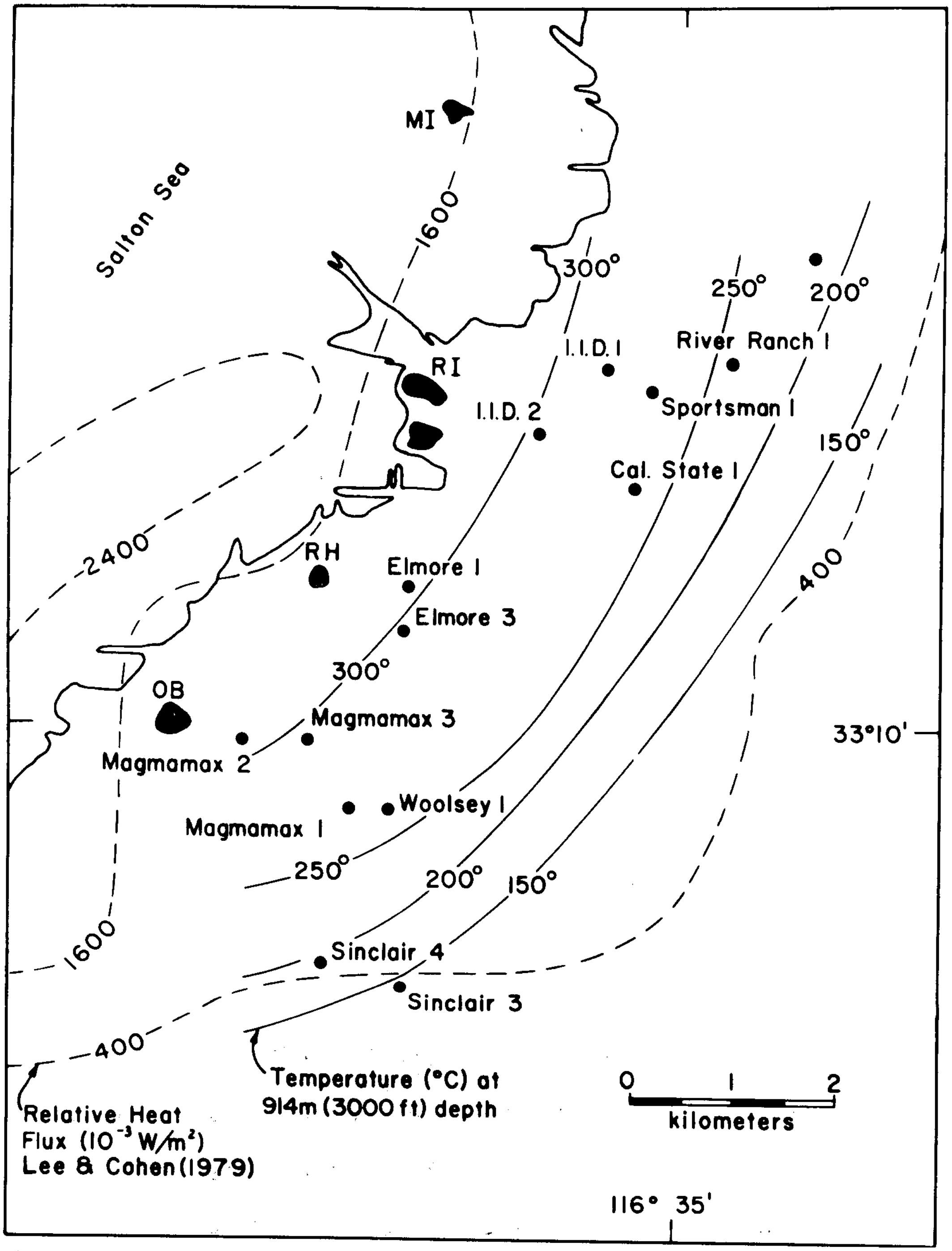

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Fig. 4. Location of boreholes and isotherms at 914 m (3000 ft) depth in the Salton Sea geothermal field. Solid lines are temperature isotherms (°C), modified after Palmer (1975) and Randall (1974). Broken lines are relative heat flux contours in 10^{-3} W/m^2 from Lee and Cohen (1979). Rhyolite extrusives: OB = Obsidian Butte; RH = Rock Hill; RI = Red Island; MI = Mullet Island. (After McDowell and Elders, 1980.)

On the other hand, thermal data support the possibility of "granitic" melt at a relatively shallow depth. Extrapolation of measured thermal gradients (Elders et al., 1972) suggests that water-saturated granitic rocks could begin to melt at or above the average depth to basement (6 km) in the trough. There is no other geological or geophysical evidence for this melt. Magnetic data (Griscom and Muffler, 1971), for example, indicate that the rhyolite domes are underlain by a complex of feeder dikes and plutons, but Curie isotherm analysis of the magnetic data (Kam, 1980) has yielded only perplexing and inconclusive results. It seems possible that small, isolated pockets of silicic melt may exist at >6 km depth and not yield a recognizable geophysical anomaly.

In contrast to our assessment for limited melt, Smith and Shaw (1975) estimated a magma chamber 50 km^2 in area on the basis of the spatial distribution of the rhyolite domes. From this they have inferred a magma volume of approximately 200 km^3, which we believe is much too large a figure.

GEOPHYSICAL EVIDENCE FOR MAGMA AND ESTIMATED MAGMA DEPTHS

Heat Flow

Shallow heat flow data (Lee and Cohen, 1979) indicate that there is a conductive heat flow >200 mW/m^2 (4.8 HFU) over a 560 km^2 area around the Salton Sea rhyolite domes, and high subsurface temperatures have been encountered in several deep geothermal exploration wells (Elders et al., 1972; Robinson et al., 1976; Humphreys, 1978; Lee and Cohen, 1979). Extrapolating temperature data observed in one deep, hot well by assuming a conductive gradient is maintained in the subhydrothermal and impermeable part of the sedimentary-plutonic section, Lee and Cohen (1979) suggest that a granitic melt is possible at ~6 km (650°C) and that a basaltic melt is possible at a depth of >7 km (~875 to 950°C). For the sake of comparison it should also be noted that Elders et al. (1984) modeled the Cerro Prieto thermal regime and were able to match the known subsurface temperatures in Cerro Prieto wells (some drilled to 3.5 km) to a large plutonic heat source at a comparable depth (6 km).

The remainder of the geophysical data available and interpretations thereof cast little light on the question of magma depth or the dimensions of a possible melt zone.

Gravity

Gravity highs are associated with most of the geothermal systems in the Salton Trough, and gravity is considered one of the more reliable prospecting techniques (Meidav, 1970; Biehler and Combs, 1972; Robinson et al., 1976; Fonseca and Razo, 1979). This is because hydrothermal fluid circulation and water-rock chemical reactions produce a densification of the host sediments. At relatively shallow depths, a self-sealed caprock zone has formed (Younker et al., 1982). At greater depths, hydrothermal metamorphism of the sediments has produced a denser set of calcium-silicate minerals typical of green-schist facies metamorphism (chlorite + calcite + amphibole + epidote + sphene) (Muffler and White, 1969; Elders et al., 1979; Keskinen and Sternfeld, 1982). Intrusion of mafic dikes and larger plutons into the sedimentary section would also be expected to contribute to the gravity highs (Elders et al., 1972). Moreover, the gravity anomalies may be influenced by regional structures and locally variable depth to basement, as at Cerro Prieto (Zhou, 1984). Under these conditions it is highly unlikely that the gravity effects of a small silicic magma body can be identified.

Magnetics

A positive, elliptical magnetic anomaly is centered over the SSGF. Griscom and Muffler (1971) interpret this high as due in small part to the exposed rhyolitic volcanoes but mainly to subvolcanic intrusive rocks at depths >2000 m. The subvolcanic intrusives are believed to be dike-sill swarms that probably coalesce at depth into small plutons. The dikes, predominantly basalt-diabase in composition, have been intersected by wells at several geothermal fields: Salton Sea (Robinson et al., 1976; Elders, 1979), East Brawley (Keskinen and Sternfeld, 1982), Heber (Browne, 1977), and Cerro Prieto (Elders, 1979). K-Ar age dates for three samples of igneous rocks intersected in East Brawley wells are in the range 8.1 to 10.5 m.y. (Keskinen and Sternfeld, 1982) and indicate that rifting and dike injection have been going on since at least Pliocene time.

Although there is other geophysical and geological evidence that rifting and dike injection continue to the present, the magnetic data are inconclusive, at best, as to whether a high-level melt zone might exist. Indirect information on this point may sometimes be derived from a Curie isotherm depth analysis of the magnetic data. However, the results can be biased by the processing techniques, by an erroneous assumption for the source geometry, and by the presence of interfering sources. In addition, there is also the unresolved question of what Curie temperature best represents the compositions of Salton Trough intrusives and basement rocks. For example, although magnetite (T_C ~575°C) is usually assumed to be the magnetic mineral present, Goldstein et al. (1984) found that the magnetic mineral in tholeiitic dikes intersected by deep Cerro Prieto wells is titanomagnetite with a Curie temperature of <200°C. This mineral should thus be paramagnetic, not ferrimagnetic, at in situ temperature conditions.

Kam (1980) obtained a surprisingly large, and seemingly erroneous, depth of 20 km to the magnetite Curie isotherm (575°C) beneath the SSGF and Brawley areas. This is the approximate Moho depth according to deep seismic refraction data (Kasameyer, 1980). Goldstein et al. (1984) found a somewhat more realistic magnetite Curie isotherm depth of 6 km for the Cerro Prieto area. This finding led them to estimate that basalt melt occurs at about 10 km. The apparent error in Kam's analysis may have been a result of the fact that Curie depths are poorly resolved from spectral analysis when the magnetized layer is thin, which could be the case in parts of the Imperial Valley.

Electrical and Electromagnetic

Numerous electrical surveys have been made on both regional and detailed scales within the Imperial Valley for geothermal exploration purposes (e.g., Meidav et al., 1976; Harthill, 1978; Humphreys, 1978). Because of the highly conductive nature (~0.5 to 2.0 ohm·m) of the sediments that blanket the trough to depths of several kilometers, the depths of published electrical investigations have also been relatively modest. There is no published information regarding a possible melt zone discerned from electromagnetic sounding techniques. A resistive basal layer was found beneath the SSGF (Meidav et al., 1976) and beneath the Cerro Prieto field (Gamble et al., 1981).

Seismological Investigations

Refraction Survey Results

A seismic refraction profile crossing the SSGF gave evidence for a denser, high-velocity region beneath the field. This has been interpreted as due to metamorphosed sediments and igneous rocks.

Earthquake Activity

Seismic activity in the Imperial Valley has been monitored over several geothermal fields. These measurements have served to define several active fault zones and have provided scientists with a better basis for understanding the regional tectonics (Elders et al., 1972). The patterns of small earthquakes and first-motion studies of larger events with clear first motions have provided a picture of northwest-southeast-trending, en echelon strike-slip faults of the San Andreas-Imperial fault system connected by short segments of northeast-southwest-trending faults that exhibit extensional behavior (Hill et al., 1975; Schnapp and Fuis, 1977; Gilpin and Lee, 1978). Earthquake foci in the SSGF area are unusually shallow (1 to 3 km) compared to foci elsewhere in the Imperial Valley (Gilpin and Lee, 1978). This has been explained as a temperature-related effect; metasedimentary rocks in the upper part of the SSGF system are brittle enough to reflect the crustal spreading by normal faulting and the shearing by strike-slip faulting. However, rocks in the deeper, high-temperature regions deform by aseismic creep.

Areas where one might look for ongoing magma injection are the northeast-southwest-trending seismic zones associated with crustal spreading. One such zone lies 6 km southeast of the SSGF rhyolite domes and connects segments of the en echelon strike-slip faults (Fig. 5). Because this spreading center lies off the main parts of the gravity and magnetic highs and because seismicity is shallow, it is not altogether clear whether the seismicity is related to the processes of dike injection. Careful seismic monitoring over the Cerro Prieto geothermal field area also provided evidence for a leaky transform fault zone (Reyes and Razo, 1979). However, earthquakes there were deeper, and first motion studies were consistent with extensional faulting (conducive to possible magma movement) at depths of 6 to 11 km.

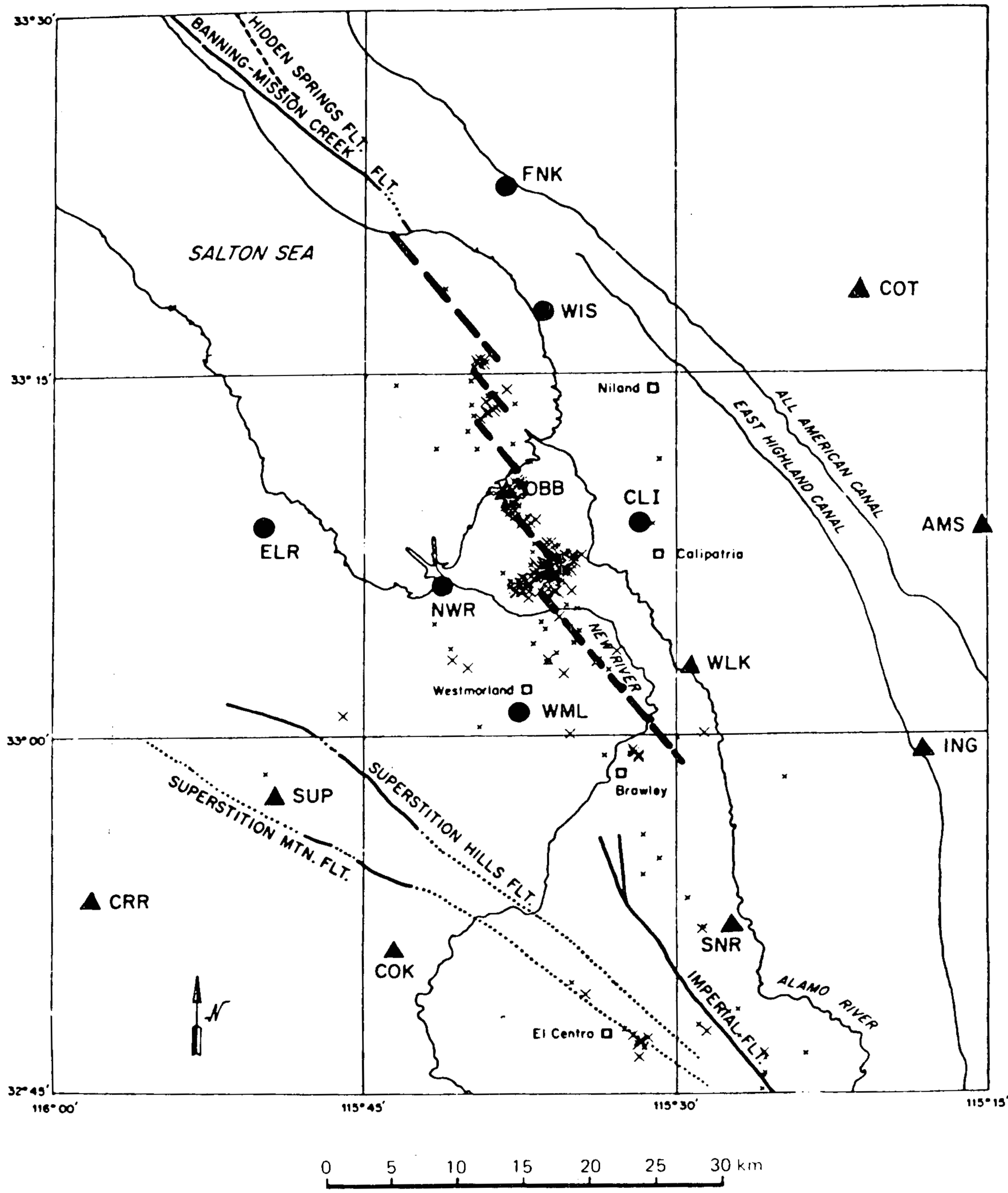

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Fig. 5. Earthquake epicenters at the Salton Sea geothermal field from October 1 through December 31, 1976. Tentative fault locations are shown as heavy broken lines. Seismograph stations are shown as the solid circles and triangles. (From Schnapp and Fuis, 1977.)

LONG VALLEY

SUMMARY

The Long Valley area has become one of the more exhaustively studied areas in recent years for reasons of its geothermal potential, as a possible site for Continental Scientific Drilling Program (CSDP) Thermal Regions drilling and because of strong earthquake activity and possible magma movement. Because much has already been written on the Long Valley caldera and its magma-hydrothermal system, that material will be discussed here only in outline. For more detailed descriptions the reader is referred to Bailey et al. (1976) and Emerson and Eichelberger (1980) for geological summaries and detailed descriptions of the volcanic evolution of the caldera, and to Lachenbruch et al. (1976a, 1976b), Sorey and Lewis (1976), and Sorey et al. (1978) for data pertaining to the present hydrothermal system. In addition, a DOE/OBES-sponsord workshop was recently held for the purpose of discussing the results of many new geophysical studies completed since the area was last reviewed for CSDP drilling (Kasameyer, 1980; Luth and Hardee, 1980) and since the renewal of earthquake activity, and the reader is referred to the Workshop Proceedings (Goldstein, 1984) for a discussion of geology-geohydrology and geophysics. The Workshop Proceedings also provide the basis for the geophysical discussion below.

Although volcanic activity in the Long Valley caldera and vicinity has persisted through at least the past million years, all recent phases of activity--most notably the very young eruptions of the Inyo chain--point to the western portion of the caldera as the location of the youngest magma sources. Studies of the hydrothermal system also suggest the presence of young heat sources in this area.

Various independent lines of geophysical study have produced intriguing anomalies that provide strong evidence for one, and perhaps several, melt zones at depths >6 km beneath the resurgent dome area in the western part of the caldera. The various anomalies differ, however, with respect to specific locations and depths of the inferred melt zones. They also leave open the question of whether the melt zones represent residual or resurgent magma from the main Long Valley chamber, or are related to another magma chamber such as may be present beneath the Mono Craters to the north. The causes of present-day seismic activity and crustal deformation--e.g., dike intrusion or magma chamber inflation, or both--also remain to be resolved.

GEOLOGICAL SUMMARY AND EVIDENCE FOR MAGMA

Hydrothermal System

Most evidence from the hydrothermal system within the Long Valley caldera points to the west and possibly the southwest moat areas as the most likely locations of heat sources. In the conceptual model proposed by Sorey et al. (1978), the ground water system in the caldera is composed of a deep subsystem, comprising the principal hot-water reservoir, in fractured, densely welded Bishop Tuff; and a shallow subsystem of cooler waters in caldera fill above the Bishop Tuff. The shallow subsystem is characterized by dominantly lateral flow and is separated from the deeper hot reservoir except where recently active normal faults provide permeable conduits for upward fluid movement. Stable isotope data and thermal and hydraulic modeling (Sorey et al., 1978), indicate that recharge to the hydrothermal system, for all but the easternmost portion of the caldera, occurs mainly around the western rim, and subsurface flow is eastward toward the main hot spring discharge areas in the Hot Creek gorge and at Casa Diablo (Fig. 6).

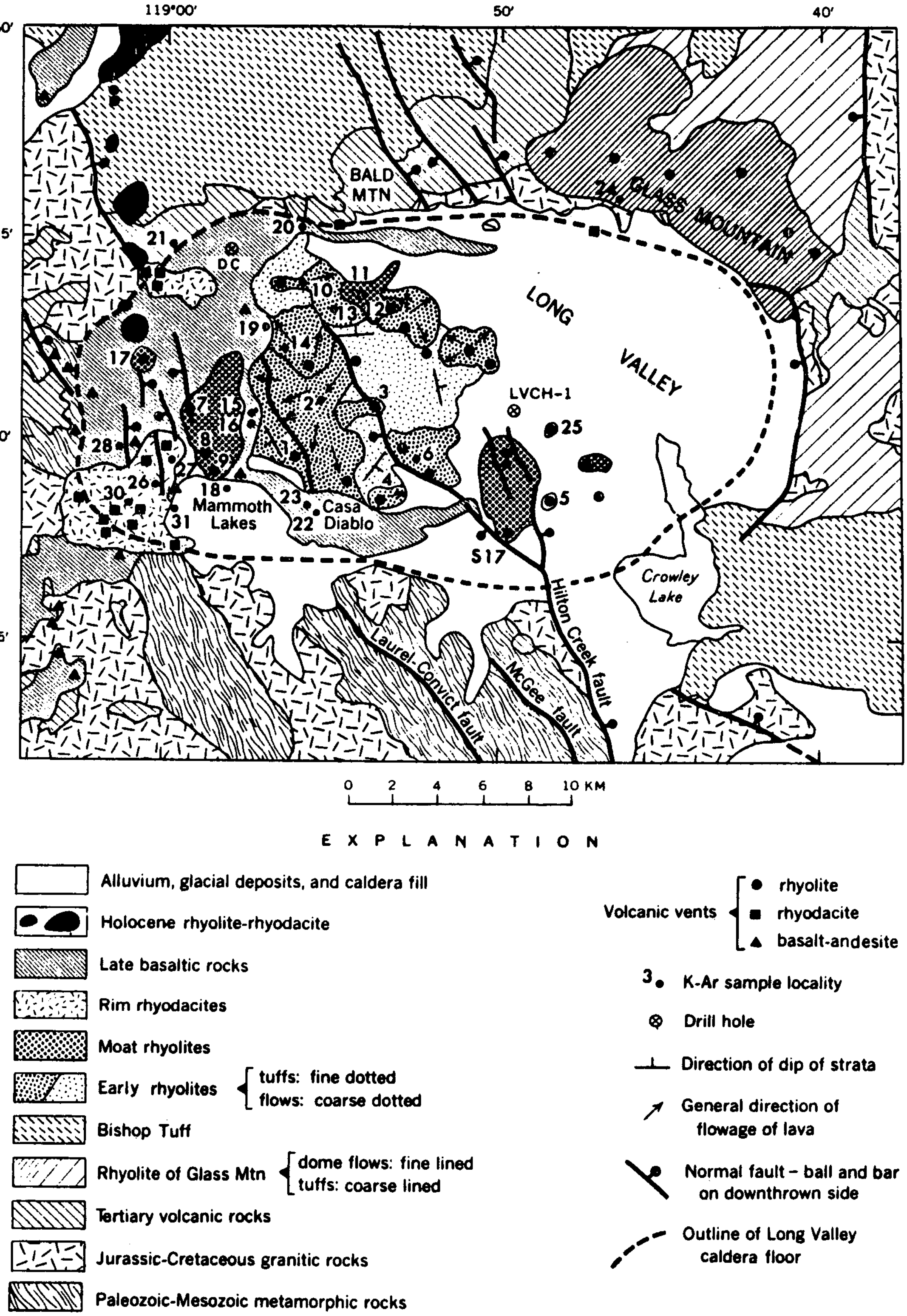

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Fig. 6. Generalized geologic map of Long Valley caldera. (From Bailey et al., 1976.)

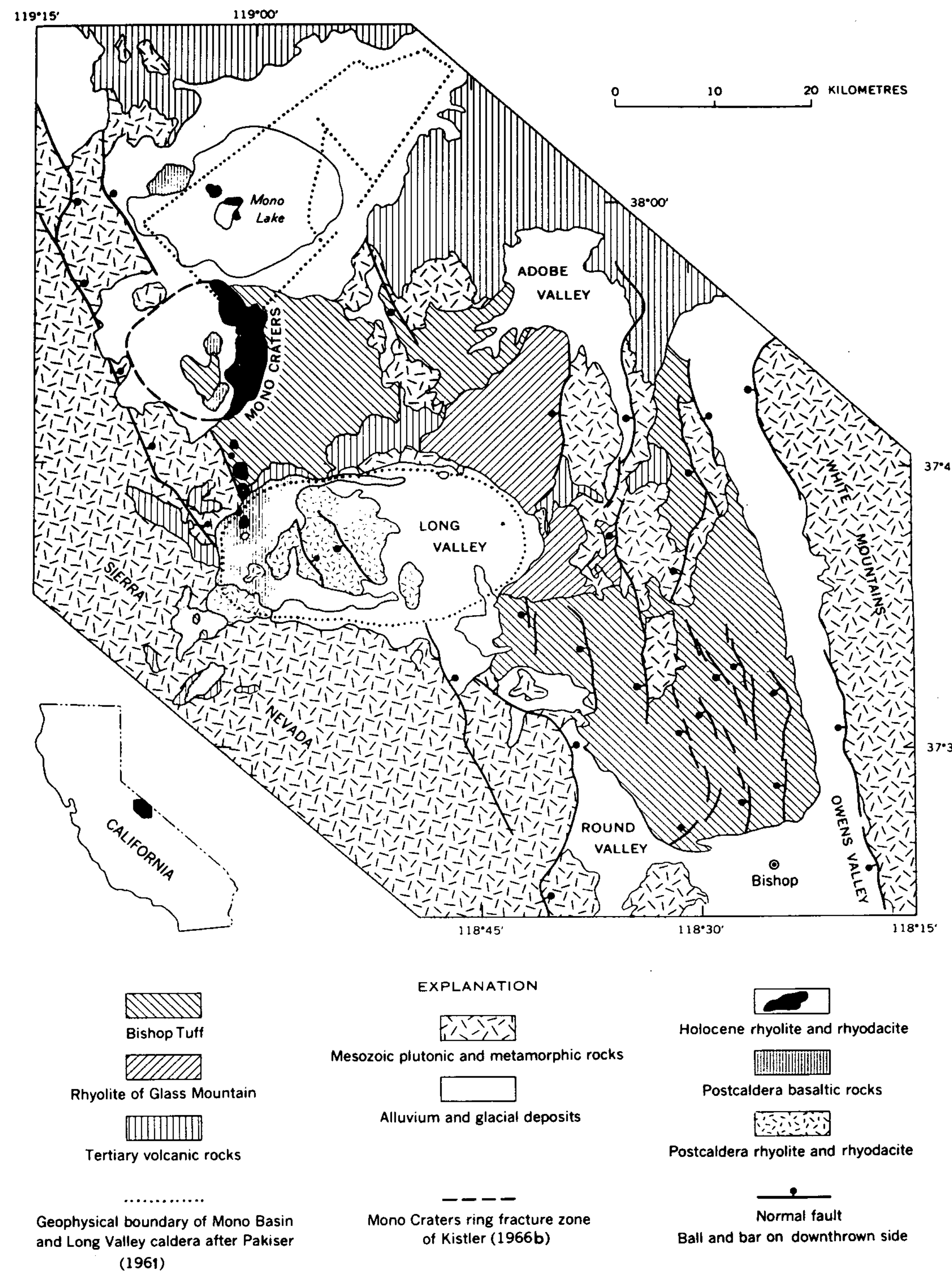

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Fig. 7. Generalized geologic map of the Long Valley-Mono Basin area. (From Bailey et al., 1976.)

Chemical geothermometer temperatures of inferred reservoir fluids range from ~220 to 270°C for Casa Diablo waters and diminish to ~150 to 190°C for waters which surface in the Hot Creek gorge (Sorey et al., 1978). This decline in reservoir temperatures is compatible with a model in which the Casa Diablo and Hot Creek waters are related to a single hot-water reservoir by progressive eastward cooling due to steam loss and dilution (Fournier et al., 1979). Decline in B and Cl concentrations eastward from Casa Diablo is also compatible with this model. (The data do not, however, exclude the possibility that the Casa Diablo and Hot Creek waters are fed by separate and distinct hot-water reservoirs.)

Direct magmatic contribution to thermal fluids is suggested by several types of evidence, but that evidence is generally open to other interpretations as well. Owing to extremely low solubility in magma, CO_2 is likely to degas readily from rising magma, and values of $\delta^{13}C$ in vapor-phase CO_2 obtained at Casa Diablo are consistent with a magmatic source (Taylor and Gerlach, 1983). However, leaching of Paleozoic carbonate rocks, which are likely to be abundant in the basement underlying the Bishop Tuff, may also account readily for the observed ^{13}C concentrations. Soil Hg concentrations have increased in the western moat between 1975 and 1982 (Varekamp and Buseck, 1984), and further increases in 1983 were reported by Williams et al. (1983). While this may reflect present-day magmatic contributions, it is more likely that the Hg anomaly is more directly related to recent seismic activity and crustal deformation in the vicinity of the west and south moats, which may have opened pathways for shallow Hg to reach the surface. Neither is a magmatic source required to account for observed B, Cl, K, Li, and As deposition in evaporites in Searles Lake, downdrainage from Long Valley, which could have been derived from leaching of reservoir rocks (Sorey et al., 1978). Similarly, deuterium and ^{18}O compositions of Long Valley waters do not require significant contributions of magmatic water (Sorey et al., 1978).

Evidence from hydrothermal alteration, supported by analysis of the evaporite deposits mentioned above, suggests that hydrothermal activity was present ~0.3 m.y. ago and was probably more extensive then than today. The general distribution of hydrothermal alteration in the caldera occurs in an arcuate zone peripheral to the resurgent dome, suggesting that the heat source for this older system was residual magma of the resurgent main chamber, and that the rise of heated fluids was controlled by the ring fracture system surrounding the dome (Bailey et al., 1976). In the present-day hydrothermal system, despite the inferred presence of the main source of heat in the vicinity of the western moat, hot spring and fumarole activity in the western moat is much less widespread than it is in the east. This is probably due to the youth of magmatic activity in the western moat and western caldera rim areas in combination with the masking effect of cold ground water infiltrating along the western rim. The lack of any present-day thermal anomaly associated with the 12,000- to 600-y.-old rhyolites of the Mono Craters north of the caldera (Fig. 7) lends support to this interpretation.

Evidence Relating to Magma Origin and Magma Chamber Presence

Although many aspects of the evolution and present state of the Long Valley magma-hydrothermal system have yet to be worked out, the general history of the system is fairly well understood. Following the caldera-forming eruption of the Bishop Tuff (0.7 m.y.), and resurgence of the magma chamber within the next 0.1 m.y., the sequence of eruptions in the caldera records a progressive solidification of the underlying zoned chamber. As outlined by Bailey et al. (1976), this is evidenced by the trend toward more mafic and crystal-rich eruptions, from the early rhyolites contemporaneous with resurgence, through the rhyolites of the peripheral moat (0.5 to 0.1 m.y.) and the rim rhyodacites (0.2 to 0.05 m.y.). The eruption of late basaltic lavas in the west and south moat areas 0.2 to 0.06 m.y. ago also shows that by this time the areal extent of the silicic chamber had diminished significantly (Fig. 6).

Thickening and strengthening of the magma chamber roof over time, due to solidification of underlying magma, is also consistent with fault configurations and displacements inside and outside the caldera borders. The major Sierra Nevada frontal faults to the northwest and southeast change from single, continuous faults to branching and en echelon faults where they transect the caldera, and displacements inside the caldera are much smaller than outside, and are relatively recent. This suggests that the roof of the cauldron block had only recently thickened enough to allow tectonic stresses to be transmitted through it (Bailey et al., 1976).

Constraints on the depth to the top of the Long Valley magma chamber have been worked out for several periods in its history. The Bishop Tuff erupted from the top of the chamber, which progressively deepened from ~6 km (earliest ash falls) to ~10 km (latest ash flows) during the course of eruption, according to mineralogical and geochemical studies by Hildreth and Spera (1974). (An estimate of ~15 km, however, was given by Stormer, 1983.) Bailey et al. (1976) estimated that the roof of the chamber had risen to <5 km and probably to 2-3 km depth by the end of resurgence ~0.1 m.y. later, based on the geometry of the resurgent dome. They also speculated that by ~0.2 m.y.b.p.

the roof had receded to 6-9 km depth, based on the degree of encroachment of the basaltic vents of that age into the caldera moat.

The temporal and spatial distributions of recent eruptive units strongly suggest that the youngest heat sources are located beneath the western moat and western caldera rim areas. The youngest eruptions probably related to the main magma chamber (the 0.1-m.y. moat rhyolites and the 0.05-m.y. rim rhyodacites) were vented from the west moat, and the most recent intra-caldera basalts (0.1 to 0.06 m.y.) from the west and southwest moats. But the Inyo eruptions--a chain of rhyolitic to rhyodacitic domes and flows, with associated phreatic explosion craters, arrayed on a north-trending line intersecting the northwest rim of the caldera (Figs. 6,7)--provide the most direct evidence for a contemporary heat source in this area. Miller (1983) obtained an age of ~550 y. for the youngest group of Inyo eruptions, based on radiocarbon dating of associated organic material. He also suggested that they were erupted when an 8-km vertical dike was emplaced at shallow depth, and the existence of a dike connecting at least the two southernmost of these eruptions has now been confirmed by drilling in October 1984 (Eichelberger, pers. comm., 1984).

The Inyo eruptions, the late basalts, and the late caldera moat and rim eruptions are probably all interrelated. The basalts, first of all, are part of a chain of mafic volcanic rocks extending and decreasing in age from southwest of Mammoth Mountain northward into Mono Basin (Fig. 7), (Bailey et al., 1976). On a regional scale, they are similar to Quaternary basaltic rocks erupted along much of the eastern margin of the Sierra Nevada, including those in the Coso area. The mafic magmas must have played a critical role in sustaining recent upper crustal silicic melt in the Long Valley chamber and in the Inyo system, by providing heat input from the lower crust. Direct mixing of mafic magma with silicic magma probably also occurred, as demonstrated by the rim rhyodacites of Deadman Creek (Bailey et al., 1976), and Mammoth Mountain (Koeppen, 1983).

The Inyo eruptions must also be understood in a wider context. They are closely related to the rhyolites of the Mono Craters, immediately to the north of the Inyo chain, which range in age from ~12,000 to ~600 y. (Dalrymple, 1967; Sieh et al., 1983). The youngest of the Inyo domes erupted a mixture of two distinct silicic magmas: rhyolite obsidian similar to Mono Craters material, and rhyodacite similar to late Long Valley caldera rim and moat extrusions. Bailey at al. (1976) note that the rhyolite obsidian component in this group of eruptions decreases progressively southward approaching the caldera, and infer that these eruptions represent mixing of residual magma of the Long Valley chamber with magma of the Mono system. Stockman et al. (1984) suggest, however, that these compositional trends in the Inyo eruptions may be only apparent, as they have observed comparable variations in drill core from a single dome.

It is thus unclear how large or continuous a body of magma may now be present in the western moat area, or whether the dike intrusion represented by the Inyo eruptions is related to a large shallow silicic chamber. Any estimates of volume or depth of a magma body will depend greatly on what model of magma distribution is envisioned, and will suffer from the same uncertainties. Even assuming that the Inyo and other recent eruptions are directly related to main Long Valley magma chamber, there is still no assurance that the previously discussed estimates of the depth to the top of that chamber, made for various stages in its evolution, are reliable guides to its present depth.

Smith and Shaw (1975) estimated the volume of magma present beneath the caldera at 2400 km^3, but this figure should similarly be interpreted with caution. Their estimate, derived from the area of the caldera, does not account for the progressive solidification of the magma chamber recorded in the sequence of post-caldera eruptions, and is therefore likely to be too high.

GEOPHYSICAL EVIDENCE FOR MAGMA AND ESTIMATED MAGMA DEPTHS

Heat Flow

The total (conductive plus convective) heat flow in the Long Valley caldera is about 15 μcal $cm^{-2}s^{-1}$ (HFU). In their conceptual thermal model of the caldera, Lachenbruch and Sass (1977) picture a convecting hydrothermal system in fractured caldera fill and basement rocks overlying a relatively unfractured basement through which heat is conducted from magma at ~6 km depth (Fig. 8).

The measured bottom-hole temperatures in deep wells on the resurgent dome are consistent with this model if the total heat flow is regarded as having been supplied to a deep circulation system by a large magma chamber for a long enough time (~300,000 y) to reach thermal equilibrium. Alternatively, the temperature log data could represent thermal equilibrium with a deeper magma and no deep fluid circulation.

The bottom-hole temperatures and borehole temperature profiles do not provide an unequivocal picture of the thermal regime beneath Long Valley. Recent studies by Sorey (1984) and Blackwell (1984) indicate a complicated circulation pattern of deep and shallow flow controlled by subvertical faults and shallow aquifers, snowmelt recharge from the Sierra Nevada, and possible shallow and recent heat sources in the western moat area, such as the dike(s) that produced the Inyo domes.

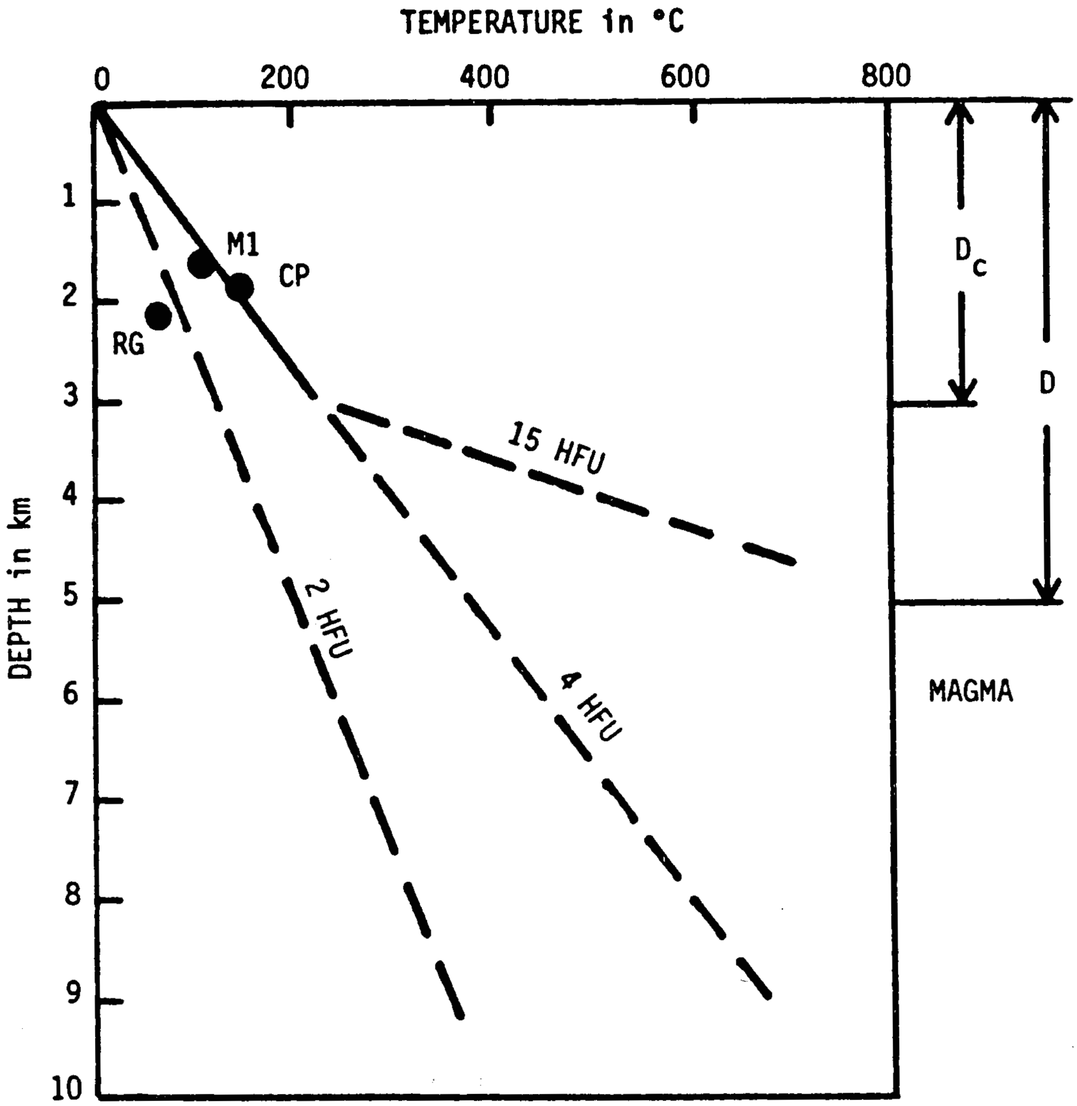

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Fig. 8. Plots of temperature versus depth for conductive heat flow at various rates (thermal conductivity = 5 mcal/(s°C cm) and bottomhole temperatures in deep wells in Long Valley caldera. Solid line represents the inferred background temperature gradient beneath the resurgent dome, based on data for well M1 and CP. D is the depth to magma at 800°C that would produce a heat flux of 15 HFU under steady-state conditions beneath a zone of fluid convection of thickness D_C. (From M. Sorey, U.S. Geological Survey.)

Seismological Investigations

Earthquake Mechanisms

The spatial-temporal distribution and mechanisms of the numerous earthquakes that occurred in 1982-1983 beneath the south moat have been studied by several researchers. No clear and simple picture of earthquake pattern and mechanism has emerged. Most workers (Peterson et al., 1983; Savage and Cockerham, 1984; Smith, 1984) seem to agree that there is evidence for swarm activity with mechanisms consistent with northeast-southwest tension. These events have been hypothesized to be associated with dike injection. In partial support of this hypothesis, it was noted that following the large 1980 earthquake, intensive swarms occurred in the same area, and these had the appearance of spasmodic tremors.

Velocity Anomalies

Cockerham (Savage and Cockerham, 1984) reported that a three-dimensional inversion method was applied to data from 7000 earthquakes and 50,000 P-wave arrivals recorded at USGS seismic stations in and around the caldera. A tomographic reconstruction of these results indicates a zone of low velocity (10% decrease) at 3 to 7 km depth, located roughly beneath the medial graben of the resurgent dome. This finding is in general agreement with the earlier finding by Steeples and Iyer (1976) that teleseismic P-wave delays reveal a NW-SE zone of delays at a depth >7 km beneath the resurgent dome.

No evidence has been found, however, for a velocity anomaly beneath the south moat area.

Attenuation Anomalies

Several types of P- and S-wave attenuation anomalies have been reconstructed on the basis of waves propagating under the caldera. The most complete study has been made by Sanders (1984), who found "massive shear wave attenuating bodies beneath the central and northwest caldera and small anomalous areas in the southern caldera and beneath Crowley Lake" (Fig. 9). The best defined of these anomalies occurs at the south end of the medial graben, and it is ~4.5 km deep at its shallowest point. This anomaly can also be roughly correlated with the P-wave velocity decrease defined by Savage and Cockerham (1984).

Sanders' attenuation anomaly at the northwest corner of the resurgent dome encloses the approximate area of the deep "magma roof" reflection reported by Hill (1976).

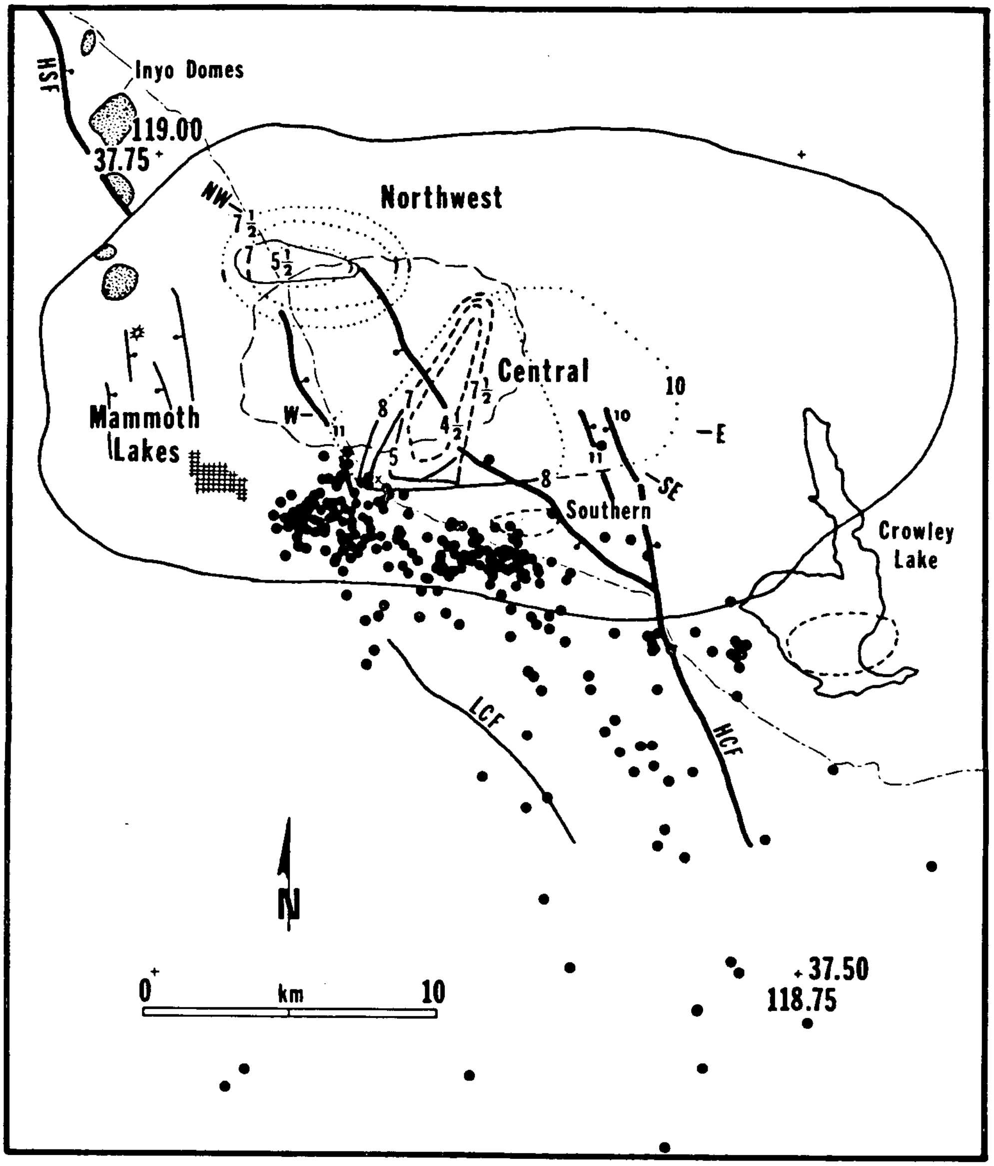
Possibly Reflected Waves

Two intersecting seismic refraction lines within the caldera show strong secondary arrivals that have been interpreted as a reverse-polarity reflection from the "top of a magma chamber" at a depth of 6-7 km beneath the northwest corner of the resurgent dome (Hill, 1976) (Fig. 10). More recent work by Luetgert and Mooney (1984), based on reduced travel times of major phases in earthquake records, suggests a reflection at 16 km in the same area. Together, both reflections may define the top and base of a large melt zone (Fig. 11).

Electrical and Electromagnetic Investigation

Recently, electrical conductors in the basement rocks underlying the Bishop Tuff have been identified.

1. Hermance et al. (1984) found a broad, concave-north, arcuate conductor south of the resurgent dome on the basis of the orientations and dimensions of telluric ellipses. This feature is believed to be due mainly to the conductive surface hydrothermal zone in relation to the resistive Sierran block.



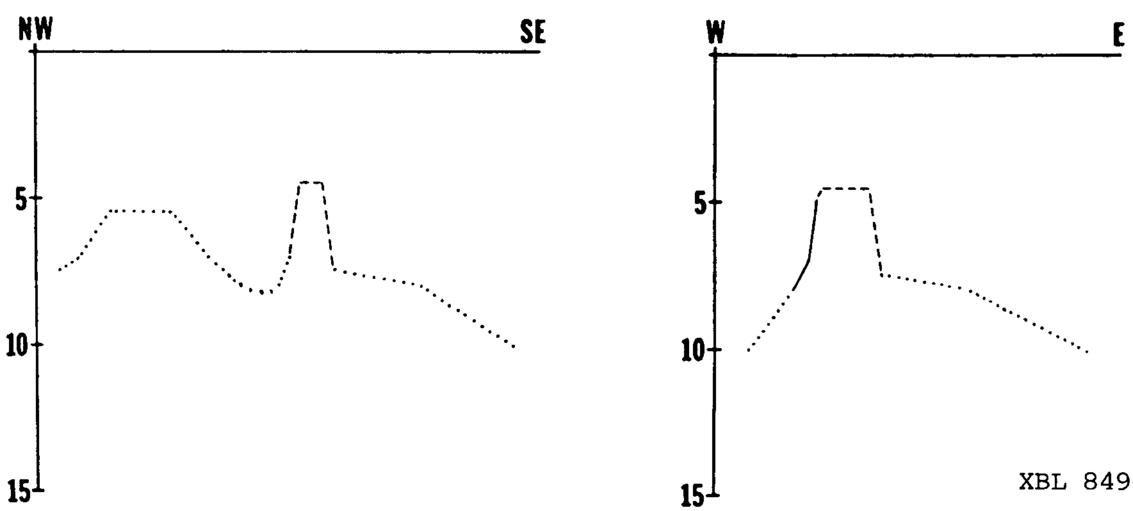


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Fig. 9. Locations and approximate depths to S-wave attenuation zones within the Long Valley caldera. (From Sanders, 1984.)

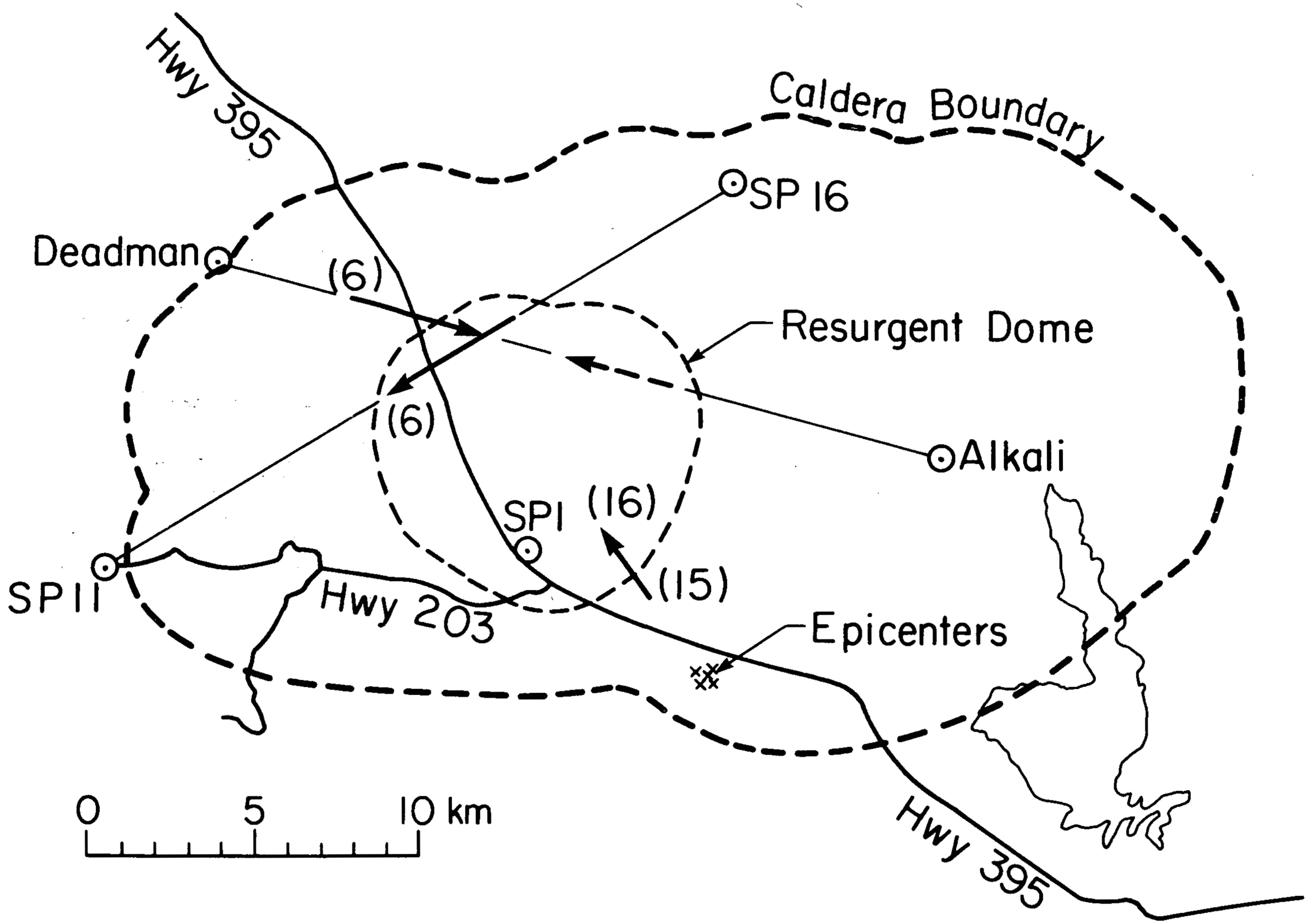

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Fig. 10. Location of refraction profiles and loci of subsurface reflection points (heavy lines). Arrows indicate shot-to-receiver direction. Parenthetic numbers indicate approximate depth in kilometers to reflecting boundary. Cluster of crosses near southern boundary of caldera are epicenters of earthquakes used in profiling. (After Hill, 1976.)

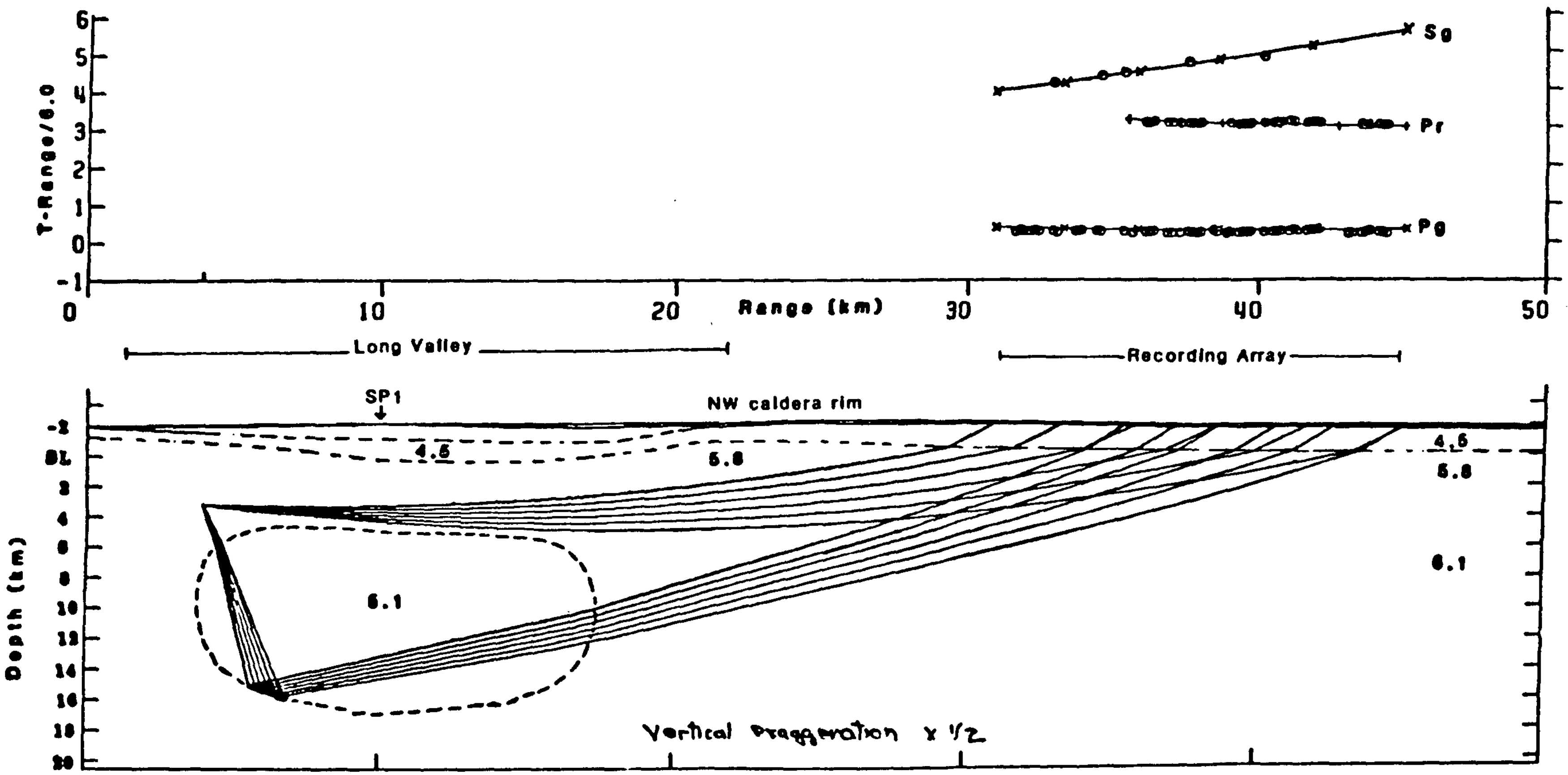

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Fig. 11. Reduced travel-time curve of major phases in earthquake record section and cross section showing ray traces for Pg and Pr phases, assumed to be reflected from the base of a magma chamber. Numbers indicate P-wave velocity in kilometers per second. Broken line indicates approximate basement geometry based on the 1982 refraction profile. (From Luetgert and Mooney, submitted to Seis. Soc. Am. Bull.)

2. They have also found magnetotelluric evidence for a deep conductor at a depth of at least 7 km and perhaps as deep as 10 km beneath the southwest moat and part of the resurgent dome.

3. Using both magnetotelluric and controlled-source EM, workers at Lawrence Berkeley Laboratory (LBL) have defined two basement conductors. One lies below the south moat area at about 3 km and plunges to the east-southeast. The other lies below the northeast corner of the resurgent dome and plunges to the southeast (Goldstein, 1984).

4. During August 1984, LBL made a series of MT soundings along an east-west line across the caldera. Figure 12 shows a preliminary resistivity section calculated from the smoothed apparent resistivity curves using the Bostick 1-D inversion technique. Although this is a crude method for processing MT data taken in geologically complex areas, at least some of the results seem consistent with geology and resistivity results from other electrical surveys. The high resistivity zones extending from near-surface to depths of 3 to 5 km beneath the west end of the line are associated with a moat rhyolite plug-dome (age ~0.1 m.y.) and the early rhyolites of the resurgent dome (age ~0.6 m.y.). The high resistivities suggest that the upper part of the rhyolite plugs are unfractured and unaltered. A third resistive zone is observed at the eastern end of the profile beneath stations 1 and 2. This may represent a thick section of Sierran granite.

Of more interest to us are the conductive zones flanking the resurgent dome. The broad surface conductor on the east has been well documented by earlier surveys made by the USGS, and the narrow surface conductor on the west, described by Hermance et al. (1984) on the basis of their telluric ellipse study, is an extension of the low resistivity zone (LRZ) mapped in the

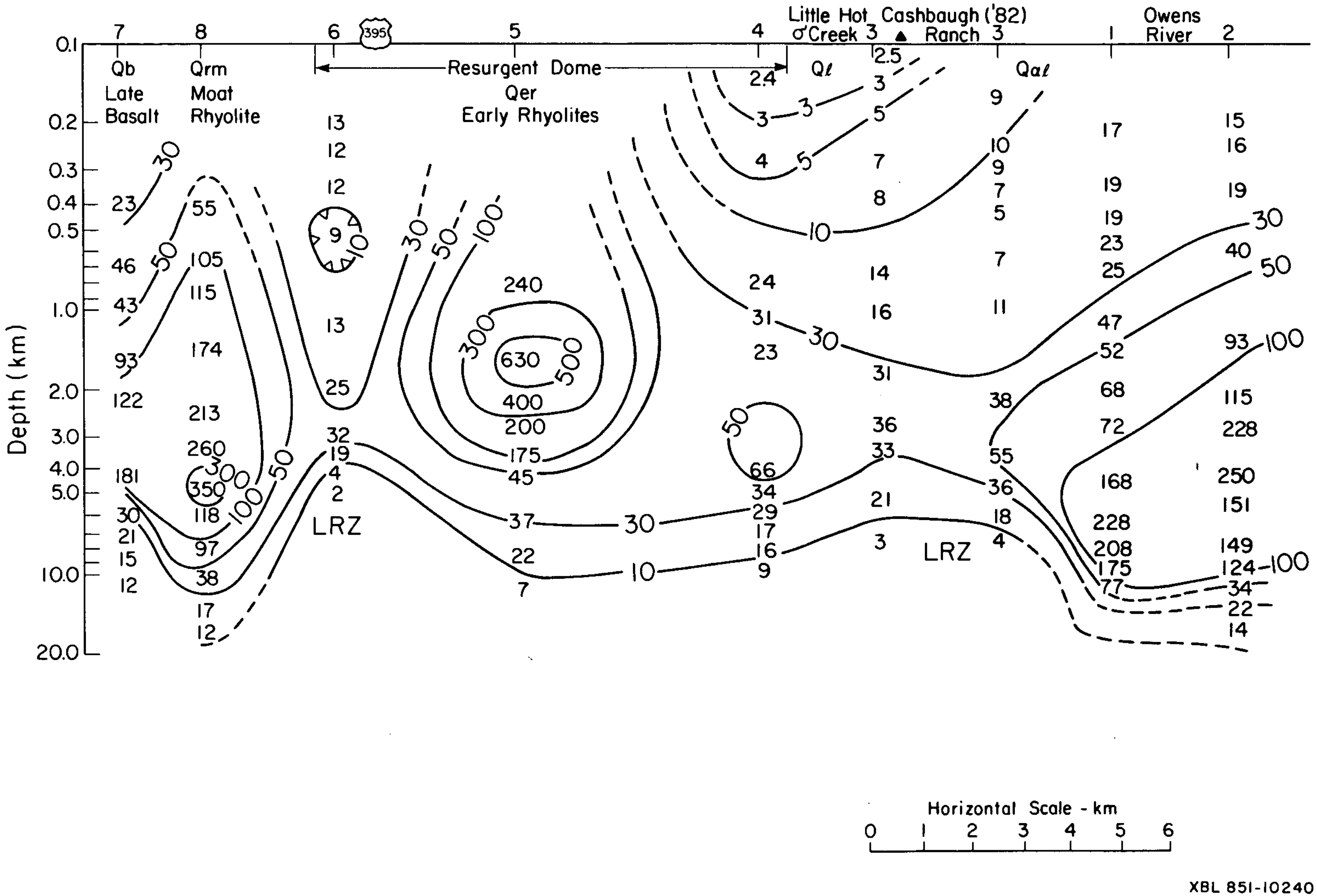

Fig. 12. Preliminary interpretation of MT depth soundings (Bostick inversions), Long Valley MT profile (east-west; units are Ohm-m).

south moat and at Casa Diablo Hot Springs. Both of these flanking surface conductors occur above regions where a deeper LRZ comes to within 5 to 8 km of the surface. While it might be speculated that the shallow LRZ's are associated with high temperatures and partial melt conditions, it is more likely, particularly in the case of station 6, that the shallowness (5 km) of the conductor is an artifact of data bias caused by current channeling in the surface conductor. To resolve the subsurface resistivity distribution, the MT phase data will be studied after the appropriate instrument phase corrections are applied to the data, and the data set will be reexamined in terms of a 2-D model. The LRZ at a depth of ~10 km beneath the central part of the caldera is probably a real feature.

There do not seem to be any serious points of conflict between the various electrical and electromagnetic interpretations at this time. It is generally agreed that more stations are needed, particularly in the areas of the seismic velocity and attenuation anomalies.

Without the benefit of deep drill holes, the relationship between the electrical conductors and possible magmatic heat sources is ambiguous. The relatively shallow conductors found at depths of ~3 km could easily be related to hydrothermal activity in fractured basement rocks or in porous sections of pre-caldera volcanics. However, it is also possible that graphitic schists in Paleozoic roof pendants underlying the caldera account for some portion of the anomalies. Regarding the deep conductor at a depth of 7 to 10 km, it is not unusual for some thermal areas in the Basin and Range geomorphic province to exhibit broad conductive zones (≤10 ohm·m) at depths of around 20 m (Lienert and Bennett, 1977). There are also reported cases of the conductive "layer" rising to within 10 km of the surface. Some of these anomalies have been associated with elevated isotherms, low seismic velocities, and possible occurrences of melt zones. A low-velocity zone beneath Long Valley was not confirmed by a seismic refraction study (Hill, 1976), but one was found by the teleseismic P-wave work of Steeples and Iyer (1976).

Gravity and Magnetic Interpretations

Neither gravity nor magnetic data have yielded conclusive information on subcaldera conditions and structures.

Gravity and Deformation Studies

Several models have been proposed to explain the recent changes in gravity and the concurrent vertical and horizontal deformation of the surface observed between August 1982 and August 1983. Deformation models proposed by Rundle (Rundle and Whitcomb, 1984) and by Savage (Castle et al., 1984) require a volume increase of 0.02 to 0.03 km^3, due to either dike injection or magma chamber inflation.

Rundle proposed a model with two point-source "magma" chambers: a deep chamber at 8 km below the center of the resurgent dome, and a chamber at 5 km below a point east of Casa Diablo Hot Springs. Savage proposed a model with a 30°-dipping slab from 8 to 10 km beneath the resurgent dome and a vertical dike intrusion from the slab to within 3 km of the surface. Neither model fits all the other geophysical data, and both are nonunique.

THE GEYSERS-CLEAR LAKE

SUMMARY

The presence of a large, shallow magma chamber beneath the Clear Lake volcanic field has been inferred on the basis of various geophysical surveys and interpretations performed mainly by scientists of the USGS. An extremely complete review of this subject was presented by Hearn et al. (1981) and Isherwood (1981), and we borrow heavily from those papers in this report.

Although it is still widely believed that a relatively shallow magma chamber underlies the Clear Lake volcanic field, the geophysical data on which this inference is based do not provide a consistent and unique model for the location and depth to a magma body or a partial melt zone. Some of the more recent survey results do not seem to support the long-held view that a large silicic chamber exists beneath Mount Hannah.

Geological and geochemical evidence also do not support the presence of a magma chamber located approximately beneath Mount Hannah. Magma chambers of significant size have probably existed beneath the Clear Lake volcanic field in the past, but there is no solid evidence for a chamber anywhere in the area at the present time. If one were present, a likely location would be well to the north of Mount Hannah, near Clear Lake. This is the area from which the most recent eruptions were vented, including those at Borax Lake, which were probably erupted from a magma chamber of some form.

An alternative model consistent with available geologic data is that magma may be present in the crust in the form of relatively small, deep bodies. Fault movements and/or movement of an underlying heat source may prevent these bodies from coalescing, rising, and evolving into a larger, shallow chamber.

GEOLOGICAL SUMMARY AND EVIDENCE FOR MAGMA

Chronology, Composition, and Magnitude of Volcanism

The Clear Lake volcanic field comprises ~100 km^3 of volcanic rocks erupted in 100 to 200 separate eruptions and ranging in age from 2.1 m.y.b.p. to ~10,000 y.b.p. As a group, the volcanics span a complete range of composition from basalt to rhyolite, with a ratio of silicic rocks (dacite and lesser rhyolite) to mafic rocks (mostly basaltic andesite) of ~3:2. In detail, however, four periods (possibly five) of major eruptive activity have been recognized, each beginning with one or more silicic eruptions; the oldest and youngest periods were dominated by mafic lavas and the intermediate periods by silicic lavas (Donnelly-Nolan et al., 1981).

The Clear Lake Volcanics are generally younger to the north, although the oldest group of eruptions is widely dispersed geographically and does not conform to this trend. Each of the subsequent three groups was much more localized in space than the first, and each was located to the north of the previous group (Fig. 13) (Hearn et al., 1981). The oldest of these three groups (1.1 to 0.8 m.y.b.p.) encompasses the Mount Hannah area beneath which the existence of a present-day magma chamber has been proposed (Fig. 14). The youngest group (0.1 to 0.01 m.y.b.p.) was erupted along the eastern and southeastern arms of Clear Lake, and consists almost entirely of basaltic andesites. The only silicic eruptions in the youngest group were the 0.09 m.y.-old rhyolite and dacite of Borax Lake (Donnelly-Nolan et al., 1981).

Structural and Tectonic Setting

The Clear Lake Volcanics are situated within the San Andreas fault system in the northern California Coast Ranges. The geologic structure is character-

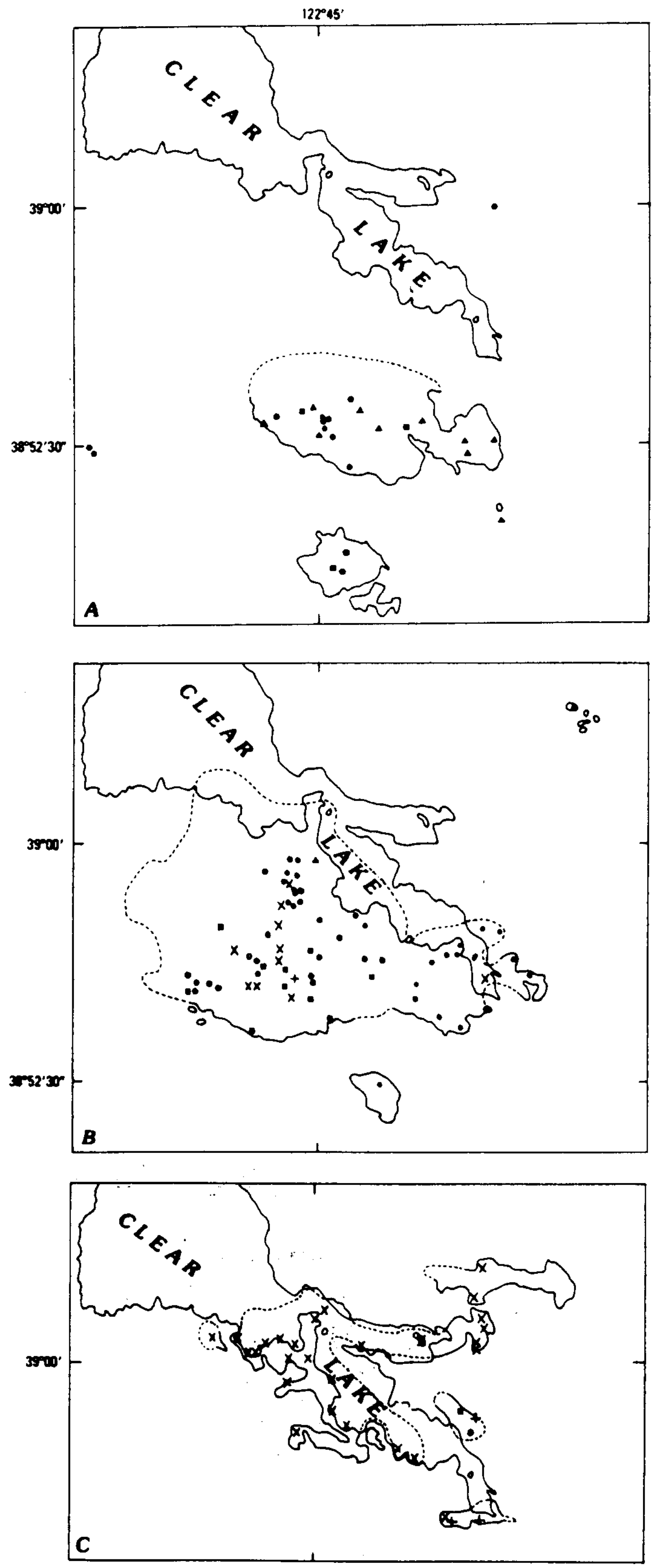

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Fig. 13. Distribution of vents of the latest three eruptive groups of the Clear Lake Volcanics. The second (0.8 to 1.1 m.y.), third (0.30 to 0.65 m.y.), and fourth (0.01 to 0.1 m.y.) groups are represented in A, B, and C, respectively. ■ = Rhyolite, ● = Dacite, ▲ = Andesite, x = Basaltic andesite, + = Basalt. (From Hearn et al., 1981.)

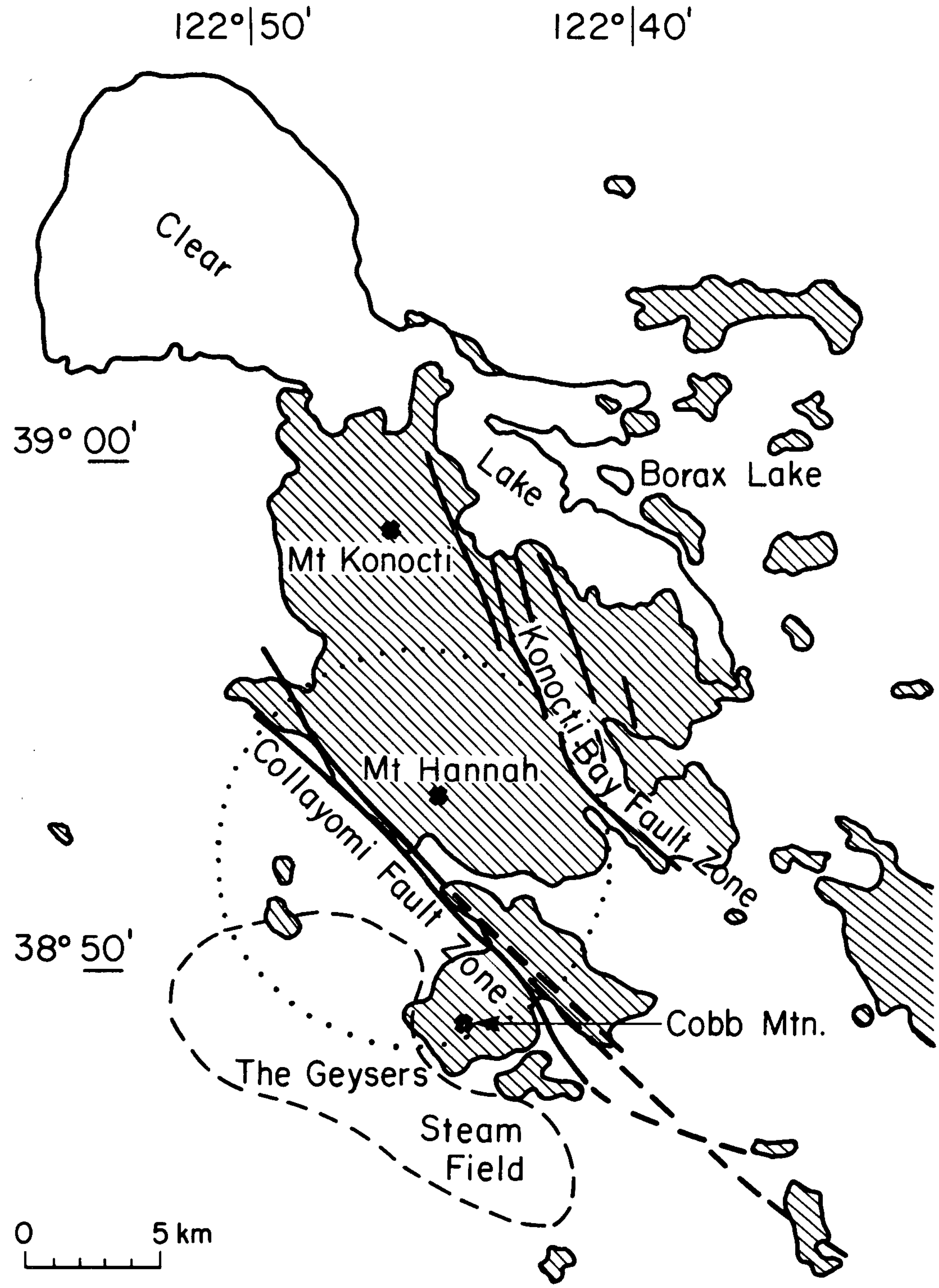

Fig. 14. Generalized map of the Clear Lake volcanic field (crosshatched), showing major fault zones and the approximate boundaries of The Geysers steam field. The outline of the present-day magma chamber inferred from gravity data is approximated by the dotted line. (After Hearn et al., 1981.)

ized by northeast-dipping imbricate thrust slices of rocks of the mainly Mesozoic Franciscan assemblage and the coeval Great Valley sequence. These have been further disrupted by the predominantly strike-slip motion of the San Andreas system (McLaughlin, 1981).

The main faults in the area are the northwest-trending Collayomi fault zone and the north-northwest-trending Konocti Bay fault zone (Fig. 14), both of which are probably active. Fault displacements in these zones and on other faults in the area are mainly normal dip-slip and right-lateral strike-slip (Hearn et al., 1981). Faults provide major conduits for movement of hot water and steam to the surface (Goff et al., 1977), and they have similarly provided avenues for magma ascent at various periods in the eruptive history of the volcanics. Vent alignments of some earliest mafic lavas and of some later dacites follow the dominant northwest to north-northwest fault trends. In addition, a zone of basaltic andesite vents and a zone of young mafic cinder cones both follow north-to-northeast-oriented lines; this is the probable direction for tensional fractures to develop in response to stresses generated between parallel northwest-trending strike-slip fault zones (Hearn et al., 1981).

Speculations concerning the relation of the Clear Lake Volcanics to regional Coast Range volcanism have centered around two main observations: the progressive decrease in age, from Miocene to Holocene, of Coast Range volcanic rocks in a north-northwestward direction toward Clear Lake; and north-northwestward movement of the triple junction between the North American, Pacific, and Farallon plates corresponding to the cessation of subduction and initiation of strike-slip movement along the San Andreas fault zone. The initiation of volcanism in the Clear Lake field followed closely the passage of the triple junction at that latitude, and similar correlations in timing occurred at Coast Range volcanic centers to the south-southeast. This suggests that magma emplacement at the Clear Lake field may be related to crustal extension near the propagating end of the San Andreas transform fault (McLaughlin, 1981). An alternate model proposes that the migration of volcanism is due to movement of the North American plate in a south-southeasterly direction over a stationary mantle hot spot or possibly that a hot spot is tied to the Pacific plate, with which it has moved relative to the North American plate (Hearn et al., 1981).

Hydrothermal System

The producing steam field at The Geysers is located in Franciscan rocks adjacent to the Clear Lake Volcanics, but it is entirely offset from them across the Collayomi fault zone (Fig. 14). It is also offset from the vertical projection of the hypothesized magma chamber beneath Mount Hannah. There remains a debate as to the nature of The Geysers hydrothermal system. It is popularly believed that it began as a hot-water system, evolving into a vapor-dominated system as a result of the low recharge rate associated with low permeability of the Franciscan reservoir rocks (McLaughlin, 1981). However, it may yet remain fundamentally a hot water system with an ever-increasing zone of boiling and superheated steam around the wells. A hot-water-dominated system is observed northeast of the Collayomi fault zone (Sternfeld et al., 1983), where the volcanic rocks and vents provide likely conduits for the deep mixing with cold meteoric water which may sustain a hot-water-dominated system (Goff et al., 1977).

Noncondensible gases in steam at The Geysers, including H_2S and CO_2, could indicate a magmatic origin, but Brook (1981) suggests that thermal degradation of organic matter in Franciscan rocks is a more likely source.

Nehring (1981) comes to the same conclusion regarding gases in springs and wells in the Geysers-Clear Lake area, and ^{13}C isotope values obtained from the same locations suggest no appreciable contributions from magmatic sources (Huebner, 1981).

On the other hand, Steinfeld et al. (1983) propose that magmatic sources contributed to the hydothermal deposition of tourmaline, together with biotite and actinolite, observed in cuttings from a 3-km-deep hole drilled ~5 km north of Mount Hannah. They suggest, based on evidence from associated fluid inclusions, that this mineralization resulted from a hot-water-dominated hydrothermal system related to nearby intrusion of magma during the third period (roughly 0.5 m.y. old) of Clear Lake volcanism. The temperatures indicated by the fluid inclusion data for this older hydrothermal system (330 to 400°C) are significantly higher than present-day temperatures (up to 260°C) encountered in the well.

Evidence Relating to Magma Origin and Magma Chamber Presence

Various lines of evidence point to the present and past existence of magma chambers of the Clear Lake Volcanics. The existence of a present-day magma chamber beneath Mount Hannah has been proposed on the basis of geophysical studies, but whether such a chamber bears any relation to earlier chambers, or to the youngest erupted magmas, is a major and unresolved question.

Of the four clearly recognized periods of eruption of the Clear Lake field, the earliest is both too mafic and too dispersed geographically to have erupted from a high-level crustal magma chamber. The second and third periods, however, were mainly silicic and relatively localized in space (Fig. 13), and some eruption from one or more chambers during these periods is more likely. The possibility that crystallized subjacent lower portions of magma chambers are present in this area is consistent with gravity data (Isherwood, 1981) and with evidence for an older hydrothermal system as inferred from drill cuttings by Steinfeld et al. (1983). Moreover, structural evidence in the form of arcuate and collapse-related features suggests eruption from high-level chambers, particularly for the large-volume eruptions of the third period (Hearn et al., 1981). Lavas erupted during the previous two periods were predominantly mafic and probably not associated with a high-level magma chamber. A chamber may, however, be developing there in response to the new input of heat at depth, of which the mafic lavas are evidence. The Borax Lake eruptions discussed below--the only silicic eruptions of this latest period--show strong evidence of differentiation in a magma chamber, which at present may either be solidified, or be too small or too deep to be detected geophysically.

Hearn et al. (1981) and Futa et al. (1981) provide detailed descriptions and interpretations of the chemical and isotopic compositions of the volcanics; the following summary of the chemical evidence is derived mainly from those sources. The mafic lavas, from which the silicics must in part have evolved, show by their variability in major and trace elements, their REE patterns, and their $^{87}Sr/^{86}Sr$ ratios that they could not have originated from one parent magma. At least two source magmas in the mantle or lower crust are indicated by the chemical data (several lavas are chemically and isotopically very primitive and must have been mantle derived). Furthermore, an overall correlation between geographical location and chemical characteristics of the eruptions holds for all the eruptive periods, suggesting that subjacent regions at depth determined chemical composition to some degree and that similar magmas may have been produced from the same region at more than one time. Upper crustal contamination of the basaltic magmas by assimilation of Franciscan serpentine or ultramafic wall rocks is judged to have been likely as well, on the basis of variability in Mg, Co, and Cr. O-isotopic and Pb-isotopic ratios also suggest that crustal contamination was significant.

In the silicic rocks, the correlation of trace-element and major-element variations, as well as moderate to strong negative Eu anomalies, indicates that these rocks evolved through various degrees of crystal fractionation (or liquid-state differentiation processes, as described by Hildreth (1979) for the Long Valley eruption of the Bishop Tuff). This, together with the fact that they are generally higher in $^{87}Sr/^{86}Sr$ ratios than associated mafic lavas, suggests that assimilation of wall rock Sr also occurred in the silicic rocks. Magma mixing was a factor in their evolution as well, as suggested both by petrographic observations and by observations of feldspar phenocrysts whose $^{87}Sr/^{86}Sr$ ratios are markedly out of equilibrium with those of their host rocks. Bowman et al. (1973) also document chemical evidence (consisting of a closely linear coherence of trace and major element variations) for mixing of rhyolite and a more mafic magma in the Borax Lake rhyolite-dacite eruptions.

Despite the complexity of sources and the varied influences of crustal assimilation and magma mixing, several factors suggest that many silicic units erupted from magma chambers. All of the rhyolites, and many dacites, show the influence of upper crustal differentiation processes consistent with evolution in a magma chamber, such as crystal fractionation or liquid-state differentiation. In addition, rhyolites erupted in each period are enriched to the same degree in several upward-concentrating trace elements and REEs, and these enrichments, while well below the extreme enrichments seen in the Bishop or Bandelier Tuffs, may nonetheless indicate differentiation in the upper level of a magma chamber; these common maxima may represent the maximum magma chamber enrichment attainable given the source magmas and the instability of the continental-border tectonic setting. (Several rhyolites of the Sonoma Volcanics, an adjacent Pliocene field to the south, also show about the same degree of enrichment.)

At least two sequences of silicic eruptions show progressive compositional changes that are consistent with eruption from a common magma chamber. The Cobb Mountain sequence (~1.1 m.y. old), from the second period of volcanism, shows such a progression (Donnelly-Nolan et al., 1981). The Borax Lake sequence (90,000 years old) of the latest period, mentioned previously as an example of magma mixing, is another case; a single magma chamber probably erupted lava of continuously varying chemical composition ranging from dacite to rhyolite. The hypothesized present-day magma chamber beneath Mount Hannah is located ~15 km (laterally) south-southwest of the Borax Lake lavas, and at this distance such a chamber is not likely to be directly related to the earlier chamber that erupted those lavas.

The presence of active faults may have had pronounced effects on magma chamber development and evolution. We noted earlier that faults control the alignment of vents in a number of units. Several series of vents that erupted primitive deep-source magmas are aligned in a north-to-northeasterly direction along probable extensional fractures, suggesting that faults guided magma ascent from deep levels. Given this tectonic setting, it is very plausible that faulting may have interfered with the development of magma chambers, particularly if new fractures developed continually in connection with propagation of a transform fault zone. (Movement of an underlying heat source relative to the crust, as proposed by the hot-spot model, would also likely interfere with magma chamber development and probably propagate new fractures as well.) Faulting may have caused repeated tapping of magma chambers, thereby inhibiting the buildup of volatiles necessary for large ash-flow eruptions and obstructing development of zones of extreme differentiation at the tops of chambers--neither of which occurs in the Clear Lake Volcanics. The Pliocene Sonoma volcanics to the south are located in a similar tectonic setting, and despite major differences in chemical composition and eruptive style, they are similar to the Clear Lake Volcanics in the abundance of eruptions, the lack of large ash flows (although small ash flows are common), and a similarly moderate maximum in trace-element enrichment in the most silicic units (Donnelly-Nolan et al., 1981; Hearn et al., 1981). Tectonic setting thus appears to have influenced, or hindered, the development of high-level crustal magma chambers in both volcanic fields.

Smith and Shaw (1975) estimated the volume of a silicic magma chamber underlying the Geysers-Clear Lake area at 1500 km^3, based on vent distribution and on the area defined by the gravity anomaly in the Mount Hannah vicinity. Our interpretation of the geological and geochemical evidence, however, suggests that any shallow magma bodies now present are likely to be much smaller, and that the criteria used by Smith and Shaw may not apply well to the present-day distribution of magma. Large high-level silicic chambers have played a significant role in earlier periods of Clear Lake volcanism, but vent distribution is more likely to reflect those earlier chambers than any present-day chamber. The association of the gravity anomaly with a present-day magma chamber is also far from certain.

GEOPHYSICAL EVIDENCE FOR MAGMA AND ESTIMATED MAGMA DEPTHS

Thermal Models

By all accounts, none of the deep production wells (to nearly 4 km) at The Geysers field has penetrated below the isothermal, convective steam zone

(~240°C). Thus, there is no solid information by which one can extrapolate temperatures to depth and develop a subsurface thermal model that doesn't depend on a priori assumptions, such as the depth and temperature of a heat source. Jamieson (1976) attempted to model available data and concluded that the surface heat flow was consistent with hot (>700°C) intrusive rocks at a depth of about 8 km over a wide area, and with conduction providing the primary means of heat transport.

Due to the competitive nature of geothermal leasing and energy development, we suspect there exists a great deal of proprietary subsurface information for the Clear Lake Volcanic Field in company files. The eventual release of these data will go a long way toward developing more accurate thermal models for the region.

Gravity and Aeromagnetics

Chapman (1975) and Isherwood (1975) have both interpreted the 25-mGal gravity low over the volcanic edifice of Mount Hannah as due to a low-density silicic differentiate of a magma chamber between about 6 and 14 km depth and centered beneath the Collayomi fault (Fig. 15). The gravity low to the southwest--called the "production low"--coincides with the steam production field and is believed due to shallower low-density rocks. On the basis of the gravity data alone one cannot make an unequivocal statement regarding the cause of the Mount Hannah low. Because drill holes to depths of >3 km have not penetrated low-density rocks, the source must be deeper and the density contrast must be large to account for the observed anomaly. It has been argued that only a silicic melt would have the proper density relative to the Franciscan assemblage to cause the anomaly. Sediments of the Great Valley

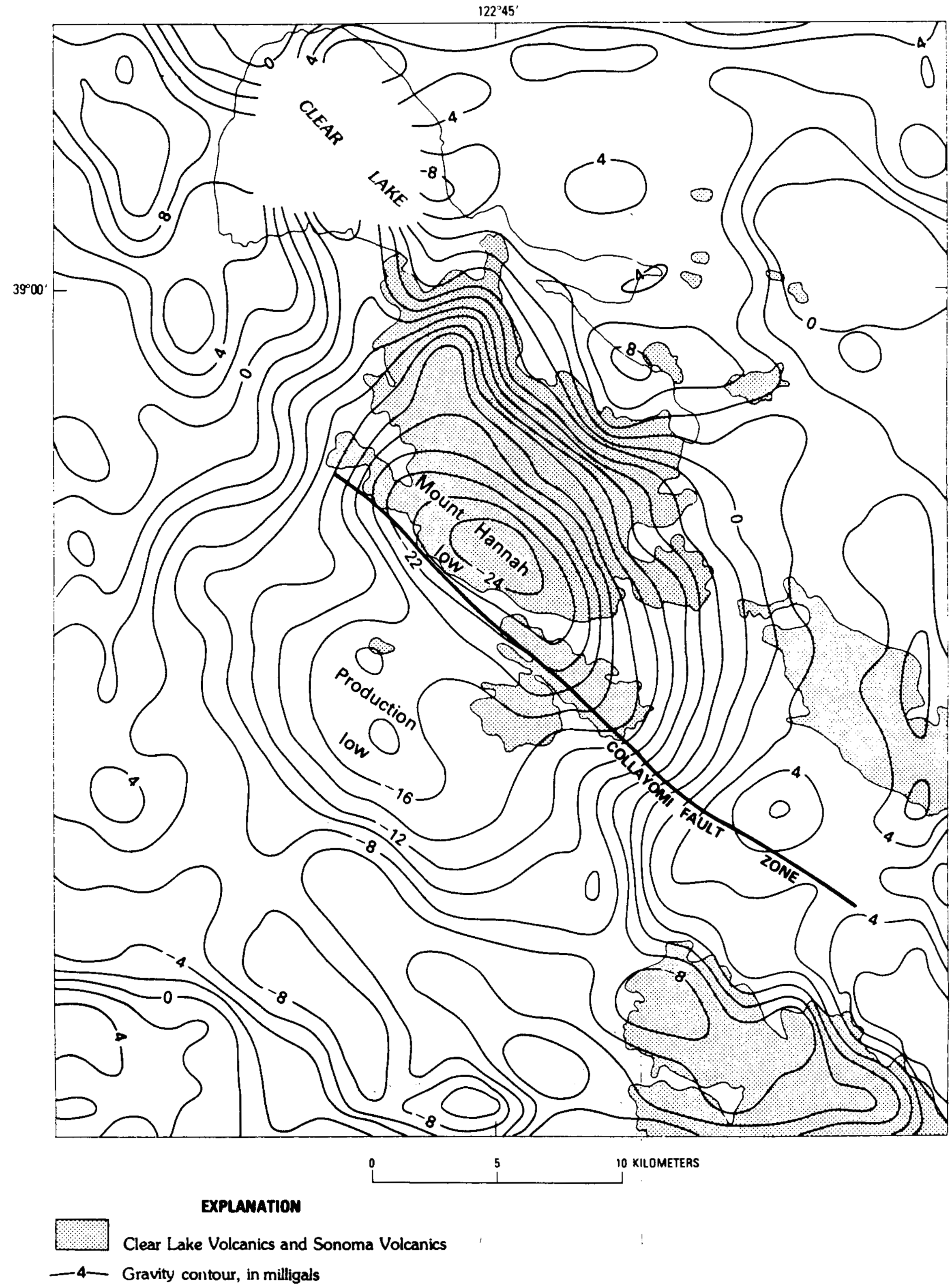

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Fig. 15. The Geysers area, Calif., showing residual gravity based on reduction densities of 2.76 g/cm^3. Contour interval, 2 mGal. (From Isherwood, 1975.)

sequence, however, could also have a sufficiently low density, although not where these rocks are densified by high-pressure-temperature conditions. Some of the gravity low may be caused by a large volume of rhyolite, dikes of which are reported to have been intersected in a few deep wells.

The aeromagnetic anomalies in the area are unrelated to the gravity anomalies. A pseudogravity map derived from the filtered aeromagnetic map indicates different sources (Isherwood, 1975). Because no magnetic source deeper than about 6.5 km can be identified, this depth is believed to coincide with the Curie temperature (~500°C), the temperature at which common ferrimagnetic minerals become paramagnetic.

Seismological Investigations

Seismic Velocities

Teleseismic P-wave delay studies by Iyer et al. (1981) outlined a broad region of 0.5-s delay centered at Mount Hannah and extending southwestward into the production area. Comparisons of the teleseismic velocity results with the results of a seismic refraction survey over the same area rule out the possibility that a low-velocity zone associated with complex near-surface geology is the cause of the P-wave delays. The authors attribute the observed delays to a deep zone of partially molten rock in which velocities are reduced by about 25%. They estimate that the zone extends downward some 30 km from a depth of about 5 km beneath the surface and has a horizontal extent roughly equal to the diameter of the gravity anomaly source (Fig. 16).

Earthquake Focal Depths

Earthquakes in the Geysers-Clear Lake area have been monitored continuously since 1975 (Bufe et al., 1981), and the hypocenters are no deeper than

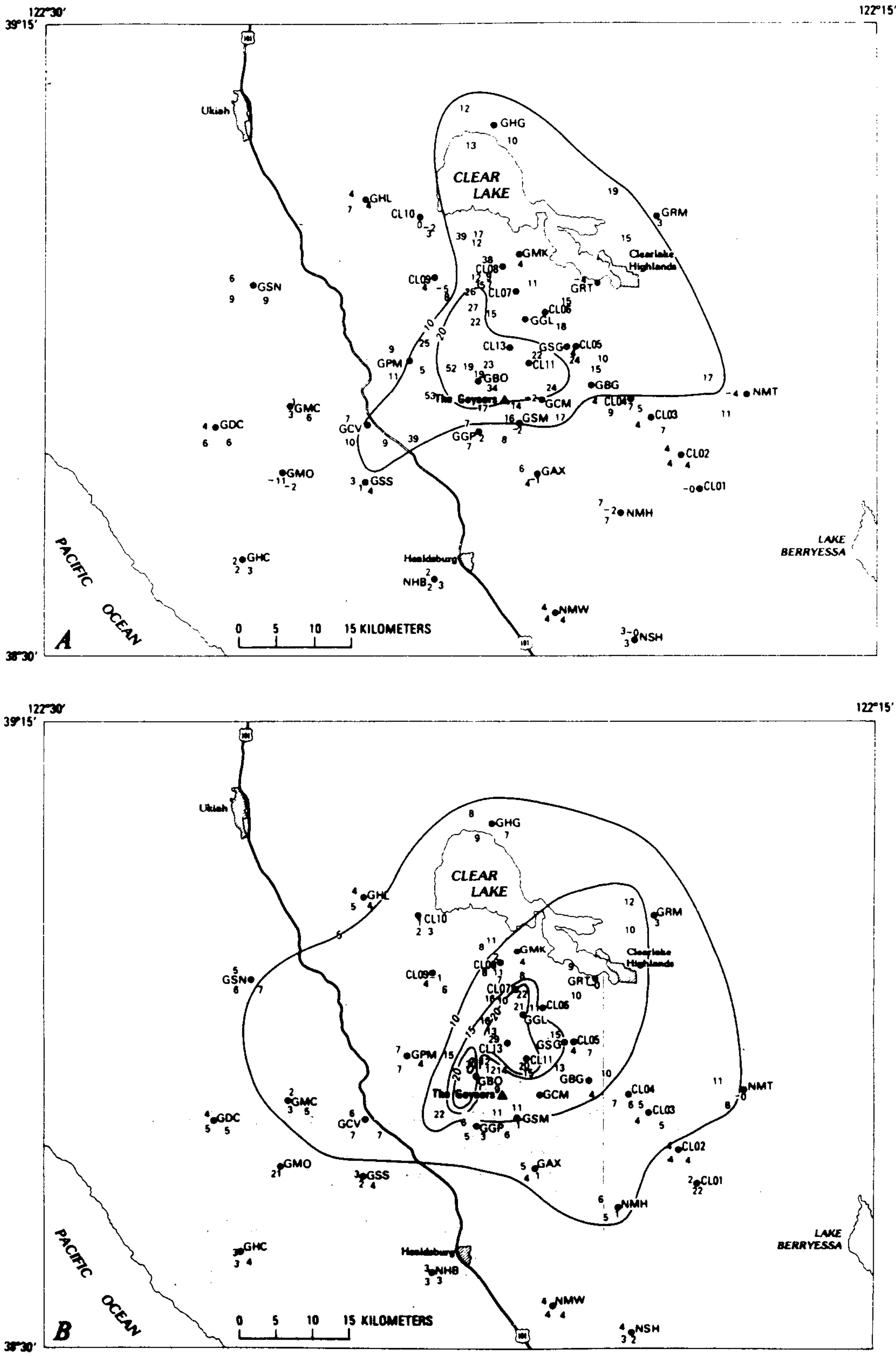

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Fig. 16. Calculated depth to bottom of anomalous body required to account for observed delays. Top of body is considered flat and assumed to be at a depth of 4 km. Numbers near station locations indicate depth in kilometers to bottom (+) or top (-) of body. Normal seismic velocity outside body is 6 km/s. (A) 15% velocity decrease; contour interval, 10 km. (B) 26% velocity decrease; contour interval, 5 km. (From Iyer et al., 1981.)

6 to 7 km. The reduced thickness of the seismogenic zone above the heat source implies a zone of relative weakness in the crustal plate (Bufe et al., 1981).

Attenuation Anomalies

Young and Ward (1981) conducted a brief field experiment to determine the location and extent of zones with high seismic wave attenuation. Their analysis of 22 teleseismic events recorded at 13 seismographs enabled them to define a shallow zone of high attenuation extending an undetermined distance from the production area to the northeast. The zone is approximately 15 km wide and is generally centered in the area of Mount Hannah and the gravity low (although the shapes of the anomalies are quite different). Young and Ward (1981) believe that the value of Q (the quality factor) is "low enough to indicate a partial melting zone or a magma chamber...." However, as the top of the low-Q zone near Mount Hannah is very close to the surface, a magma source is not a reasonable explanation for this anomaly. Another small, low-Q anomaly exists 5 to 10 km northwest of Mount Hannah, toward Clear Lake, and is at an interpreted depth of 8 km.

Majer and McEvilly (1979) obtained evidence for attenuation anomalies associated with the production area by deploying a tight array of geophones over the field and studying the refracted waves from two explosions. They found that Q within the steam field was high at shallow depths (low attenuation at depths <1 km). Attenuation seems to increase with depth, but whether local attenuation increases more than regional values could not be determined from the experiment.

Electrical Resistivity Studies

Electrical resistivity and magnetotelluric methods have been applied to The Geysers area by various workers. Vertical electrical soundings indicate a low-resistivity region near Mount Hannah that also corresponds roughly to the lowest part of the Mount Hannah gravity low and shows steep gradients near the Collayomi fault zone (Isherwood, 1981). Interpretation of these data suggests that the low-resistivity rocks probably represent the Great Valley sequence (marine rocks thrust over the Franciscan), which is overlain by the resistive Clear Lake Volcanics. Figure 17 shows an idealized geoelectric section of the Clear Lake area based on dc resistivity survey data from various sources (Keller et al., 1984). The underlying "resistive" basement that has been inferred from several soundings is the Franciscan assemblage at depths of 1.5 to 5.0 km.

A number of deep MT soundings were conducted by the USGS (Isherwood, 1981) and Group Seven (Kaufman and Keller, 1981) to determine whether a conductive basement exists anywhere in the area of the hypothesized magma chamber. The USGS MT results did not indicate a conductor at depths less than 7 km below the recording sites, but the coverage was sparse. Kaufman and Keller (1981) found evidence for a deep conductor (>10 km) at several sites. Because there seemed to be no correlation between the location and depth to this conductor and the surface conductance effects and because the deep conductor did correlate reasonably well in plan with the gravity low, Kaufman and Keller (1981) felt that the MT data supported the probable existence of rocks heated to their melting point. There are no electrical results, however, that support the gravity model that calls for a large volume of conductive melt within 7 km of the surface.

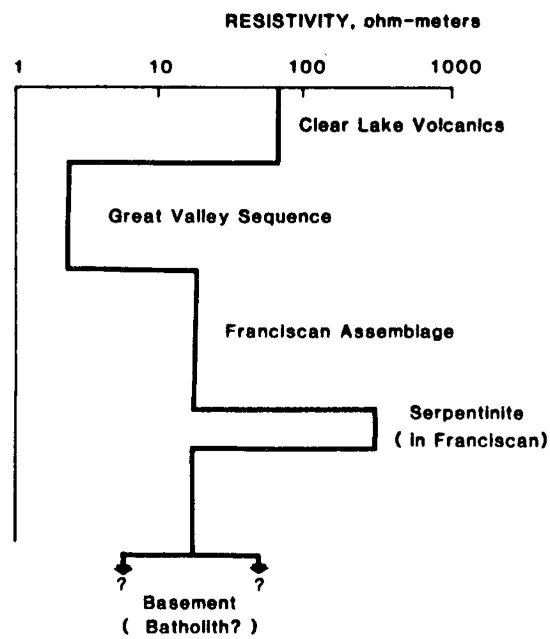

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Fig. 17. Geoelectric sequence in the Clear Lake area based on dc resistivity surveys from many sources. (From Keller et al., 1984.)

Recent Geophysical Studies

Although there remains the possibility of a melt zone south of Clear Lake, results of recent geophysical surveys conducted over the hypothesized magma body are inconclusive to negative with respect to the presence of a large, shallow magma chamber.

1. Combined compressional- and shear-wave seismic surveys were conducted northeast of The Geysers geothermal field, and the velocity sections indicated Poisson's ratios <0.25 for depths between 5 and 11 km (Rossow et al., 1983). High values for Poisson's ratio (such as might be expected if a partial melt existed) were not found. The low values indicate partial liquid saturation, such as would be found in rocks with steam-filled fractures.

2. The Colorado School of Mines operated its "Megasource" time-domain EM system in the area around Clear Lake, but northeast of the producing area, and obtained 245 soundings (Keller and Jacobson, 1983; Keller et al., 1984). The single source, a 1-km-length of AWG 4-0 wire, was located in a marshy area at the southeast corner of Clear Lake. Keller et al. (1984) state that "the most conductive rocks occur at relatively shallow depths [1 to 2 km] to the south [of the transmitter] and at greater depths to the north...In many cases the inversions do not indicate...resistive rock, even to the greatest depths interpreted...[The conductive] zone with resistivities of 3 to 10 ohm·m...is surprisingly consistent...and is assumed to belong to hotter portions of the Franciscan assemblage. Where adequate penetration was obtained, rocks at depths beyond 10 to 12 km appear to be resistant. No conductive rock...was found...beyond 10 to 12 km in the area south of Clear Lake where the gravity data suggest the presence of a [silicic melt]."

The absence of a highly conductive zone doesn't necessarily refute entirely the silicic chamber model that has been established from gravity and seismic studies, but it may indicate that the melt fraction and/or the total volume of partial melt regions are too small to produce a recognizable conductivity anomaly.

COSO VOLCANIC FIELD

SUMMARY

From the geological and geochemical evidence at Coso, the conclusion that can be stated with most assurance is that a magma chamber of significant size and longevity is present, and that it is most likely located beneath the central part of the rhyolite field. The depth and dimensions of the magma chamber are much more open to question, but it may be in the early stages of its development and it is probably deeper and smaller than the large chambers of continental caldera-forming systems. Rapid crustal extension has caused the topmost level of the chamber to be tapped for periodic small eruptions, thereby reducing the potential for larger explosive eruptions and perhaps also preventing the migration of the chamber to shallower depths. Crustal extension at lower levels has allowed basaltic magmas to penetrate the crust and provide the heat to sustain the magmatic and geothermal systems. The tectonic setting at Coso thus resembles that of a spreading center or rift zone. The available geological evidence suggests that the system has not cooled or diminished in size since the latest series of rhyolites were erupted <40,000 years ago.

Geophysical evidence--primarily the results of several independent seismological investigations--also strongly suggests the presence of magma beneath the Coso volcanic field. Anomalous seismicity, P-wave velocities and P-wave attenuations appear to be closely related to the heat flow anomaly that encloses the Coso geothermal field. Interpretations indicate a magma that is deeper than 8 km and probably deeper than 10 to 12 km. The seismological interpretations do not, however, resolve the depth or location of the magma very well. A magma beneath the general vicinity of Sugarloaf Mountain is consistent with most of the data, but it is not clear whether a single or several melt zones exist.

GEOLOGICAL SUMMARY AND EVIDENCE FOR MAGMA

Chronology, Composition, and Magnitude of Volcanism

Eruptions in the Coso Range have taken place in two periods (Fig. 18). The younger one is of Pleistocene (and possibly Holocene) age, and extruded roughly equal volumes of rhyolite and basalt totaling 4 to 5 km^3. The older group is of Pliocene age (~4.0 to ~2.5 m.y.b.p.), is more voluminous, (~31 km^3), and is largely mafic in composition although spanning a broad range of composition from basalt to rhyolite (Duffield et al., 1980).

The Pleistocene rhyolite field comprises at least 38 steep-sided domes and short flows, with associated pyroclastics. Total volume is ~2 km^3, of which <20% is pyroclastic material. All of the rhyolite eruptions consist of high-silica, crystal-poor to aphyric rhyolite, and they are very similar chemically, particularly with respect to major-element composition. Seven episodes of rhyolite eruption are recognized from trace element and chronological criteria. They range from ~1.0 to <0.04 m.y. old, but most are less than ~0.15 m.y. old (Bacon et al., 1981).

Within each of the seven groups of rhyolites, the various eruptions are indistinguishable in terms of both trace and major elements, and they are very close in age despite their geographic spread. The volume of rhyolite erupted has increased with decreasing age, the latest group being the most voluminous. The youngest K-Ar age is ~40,000 years, but the error due to low radiogenic Ar is ~50%. Evidence from the chronology and geomorphic effects of recent pluvial periods suggests that several rhyolites may be much younger, perhaps <10,000 years old (C.F. Austin, pers. comm., 1984). Ages calculated from

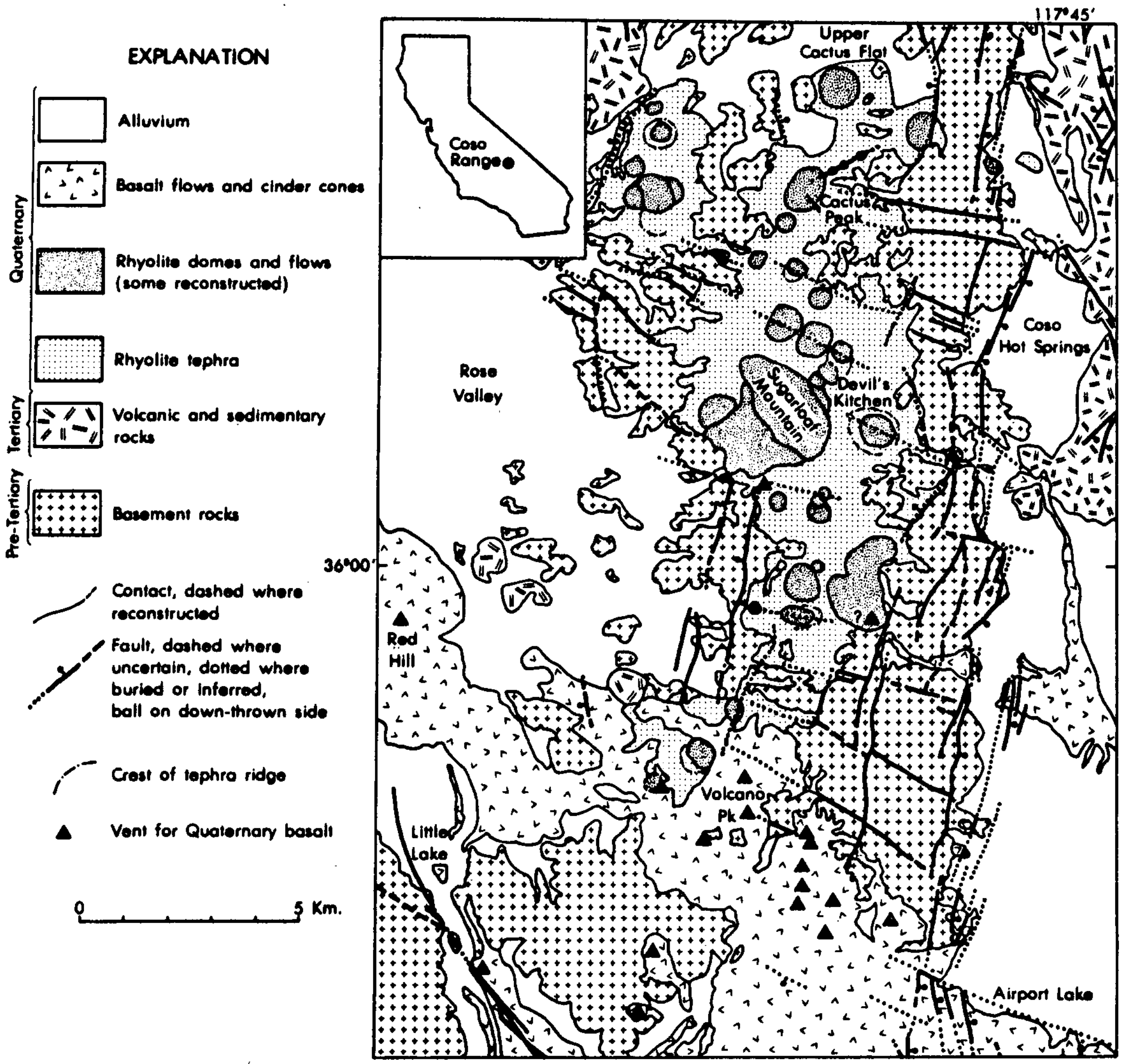

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Fig. 18. Generalized geologic map of the Coso rhyolite field and vicinity. (From Bacon et al., 1980.)

hydration rind thicknesses of obsidian (Friedman and Obradovich, 1981) include three <40,000 years and one of 17,000 years (these dates also have large uncertainties).

Basalt has erupted throughout the span of Pleistocene (or younger) rhyolite eruptions. During at least the past ~0.24 m.y. basaltic vents have consistently been peripheral to the central part of the rhyolite field. Nineteen Pleistocene basaltic vents are recognized, and each fed one or more flows. The cumulative volume of these basalt flows is ~1 km^3. The most recent basalts are located to the south and southwest of the rhyolite field, and the youngest K-Ar age on basalt is similar to the youngest rhyolite age.

Structural and Tectonic Setting

The Coso Range is situated at the boundary between the Basin and Range and the Sierra Nevada geological provinces and close to the Mojave province located across the Garlock fault to the south. The rhyolite domes and flows are located atop an approximately 5-by-8-km horst in pre-Cenozoic basement rocks, on the margins of which considerable uplift has taken place since Pliocene time (Duffield et al., 1980).

A high level of seismicity in the Coso area shows it to be tectonically active, and three dominant sets of faults are recognized (Fig. 18) (Duffield et al., 1980):

1. Northwest- to north-northwest-trending, steeply dipping faults, with linear traces and of regional extent. They are mainly strike-slip faults and have been active since perhaps Mesozoic time. Recent earthquakes show right-lateral displacement.

2. North- to north-northeast-trending faults, dipping steeply to the west, with linear traces. Both field evidence and data from recent earth-

quakes indicate that they are mainly normal faults related to the regional Basin and Range extensional regime. Several major faults in the Coso Range belong to this group, including the bounding faults of the central horst, the fault associated with Coso Hot Springs, and the series of en echelon faults forming a westward-facing staircase in Pliocene basalts on the northeast side of the range.

3. Faults with arcuate traces, of local extent, in the north and northeast part of the range. They are steeply dipping and concave toward the center of the rhyolite field. These faults may be related to a similarly oriented arcuate fault zone cutting Sierran granitic rock just west of the Coso Range, and it has been suggested (Duffield, 1975; Moore and Austin, 1983) that they are part of a ring fracture zone associated with a large underlying magma chamber. However, evidence for 360° continuity for these faults has not been found, and they do not offset rock younger than Pliocene (Duffield et al., 1980). Roquemore (1980) believes they are not volcanogenic but are instead the splayed ends of the regional right-lateral strike-slip fault system terminating against the Garlock fault zone to the south.

Fault control has been a factor in the distribution of both rhyolites and basalts. Vent orientation has been influenced mainly by the north-northeast-trending normal faults and to a lesser extent by the older north-northwest-trending structures. Vent orientation thus reflects the late Cenozoic and present-day Basin and Range tectonic regime, in which the axis of maximum horizontal compression is oriented north-northeast and extension is in a west-northwest direction.

Volcanism in the Coso Range is related to regional tectonics; it began concurrently with the onset of regional mafic volcanism and normal faulting along the eastern margin of the Sierra Nevada, ~3-4 m.y.b.p. Rhyolite volcanism near Coso also occurred in the Pliocene and Pleistocene on the Kern Plateau in the Sierra Nevada, where it was associated with basaltic eruptions of similar ages. One small dome, located at Long Canyon ~40 km northwest of the Coso field, is composed of high-silica rhyolite very similar in major- and trace-element composition to Coso rhyolites, and has a K-Ar age of ~0.2 m.y.b.p., similar to the age of the first major group of Coso rhyolites (Bacon and Duffield, 1981).

Hydrothermal System

The Coso volcanic field is characterized by high heat flow, and the central part of the field is located within, but offset 1 to 2 km west of the center of, the triangular area defined by the 15-HFU contour (Fig. 19) (Combs, 1980). Within this area of high heat flow, an east-northeast-trending zone ~6 km long encompasses nearly all sites of active thermal emissions and associated clay-opal-alunite alteration (Hulen, 1978). The main thermal manifestations are the intermittently discharging Coso Hot Springs, located along a north-northeast-trending active fault defining the eastern border of the main heat flow anomaly, and the acid-sulfate fumaroles at Devil's Kitchen, located near the center of the anomaly (Fig. 19).

The highest published temperatures encountered by drill holes are 195°C at a depth of 1100 m (well CGEH-1, Fournier and Thompson, 1980) and 213°C at a depth of 388 m (Moore and Austin, 1983). The maximum calculated geothermometer temperatures indicate a reservoir at 240 to 250°C (Fournier and Thompson, 1980). Recent drilling by the California Energy Company, 1 to 2 km south of Devil's Kitchen, has encountered temperatures of ~250°C at a depth of ~850 m (R.H. Adams, pers. comm., 1984).

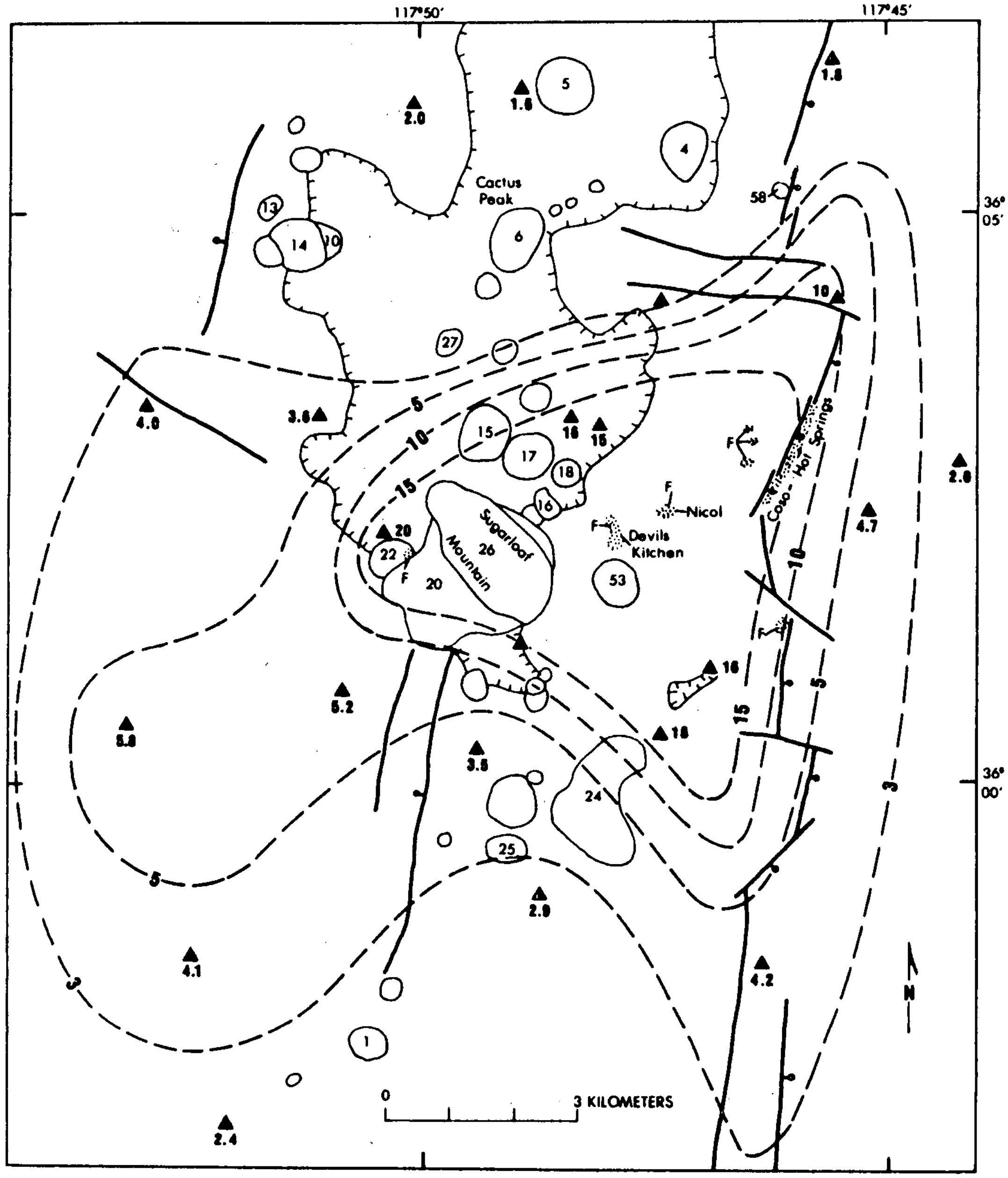

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Fig. 19. Generalized geologic map showing principal geothermal areas and Pleistocene rhyolite. Heat flow (triangles) in heat flow units (HFU) (from Combs, 1980). A 1477-m-deep drill hole is adjacent to heat flow site 15. F = fumarole area; heavy solid lines = faults, with bar and ball on downthrown side; broken lines = heat flow contours; hachures = outline for areas of internal drainage; 53 = rhyolite locality. (From Duffield et al., 1980; after Moyle, 1977.)

The presence of hot spring sinter below a dated basalt flow suggests that vigorous hydrothermal activity has probably occurred near the central part of the rhyolite field since at least 0.3 m.y.b.p. (Duffield et al., 1980). To date, the system has been characterized by a high ratio of rock to water on the basis of stable isotope analyses (Fournier and Thompson, 1982). One model that has been proposed for the present-day geothermal system (C.F. Austin, pers. comm., 1984) consists of a chloride-rich brine at depth, overlain by a zone of steam above condensate, which is overlain in turn by a 1000 to 1400-ft-thick argillized caprock.

Evidence Relating to Magma Origin and Magma Chamber Presence

The Coso rhyolites, although volumetrically small, appear to have erupted from a single magma chamber of considerable size and longevity. The argument in favor of this model is comprehensively presented by Bacon et al. (1981), and what follows largely summarizes their work.

The rhyolite domes and flows are composed of highly differentiated, high-silica rhyolite, with extreme enrichments in elements that tend to concentrate in the uppermost layer of large silicic magma chambers (e.g., Na, Rb, Th, U, and volatiles) and corresponding depletions in downward-concentrating elements (e.g., Mg, P, Ba, Sr, Eu). The rhyolitic rocks are similar to, and even more differentiated than, the uppermost layers inferred to be present in the Long Valley and Valles magma chambers prior to their eruptions as ash flows. Seven rhyolite groups are distinguished on the basis of their trace element compositions and ages. Two are older and differ compositionally from the other groups, suggesting that they may have had a different source. But the five youngest groups are very close in composition, and the trace-element variations between them match variations expected from varying degrees of differentiation at the top of a magma chamber. (In comparing variations between any two groups, the upward-concentrating elements in each group vary together in one direction, while the downward-concentrating elements vary in the other direction.) These five groups thus represent leaks of magma at different times from the topmost layer of the same vertically-zoned chamber.

The five groups of rhyolites span an age range from >0.24 to <.04 m.y., and their similar compositions indicate that the chamber from which they erupted was in existence and was compositionally zoned for at least that period of time (possibly much longer if the earlier rhyolites were also associated with it). This is consistent with an age of >0.3 m.y. for the geothermal system, as well as with the absence of basaltic eruptions younger than ~0.24 m.y. within the central portion of the rhyolite field.

The extreme uniformity of composition within each group of rhyolites shows that their eruption occurred from the small uppermost fraction of the vertically zoned chamber only, and that the main body of the chamber was not tapped. The volume of the chamber therefore must always have been far greater than that of any of the groups of rhyolites (all were <0.7 km^3). A quantitative estimate of the size of the magma chamber is difficult to make, but several have been attempted. By analogy with the rate of growth of large continental silicic magma systems, and using an estimate of the duration of volcanic activity in the Coso Range, Bacon et al. (1981) estimated the volume of magma presently in the system at several hundred km^3. Smith and Shaw (1975) estimated a volume of ~650 km^3 by deriving chamber area from the distribution of rhyolite vents, and chamber thickness by a very approximate relation between thickness and area (see Introduction). Bacon et al. (1980), however, found that the vent distribution is consistent with a magma chamber underlying the central part of the field only, and with outlying extrusives fed from the chamber by dikes propagating along tensional fractures. According to this model, then, deriving the area of the magma chamber from the total areal extent of the vents would lead to an overestimation. This would explain why Smith and Shaw (1975) came up with the much larger estimate of chamber volume.

Other models for the geometry of the magma chamber have been proposed; if valid, they could imply a larger chamber than so far discussed. If the arcuate fractures in and near the Coso Range do in fact represent an elliptical ring fracture zone some 40 by 50 km in size, as Duffield (1975) and Moore and Austin (1983) suggest, the associated magma chamber could be on the order of 1000 km^2 or more in area. Furthermore, such a ring fracture zone could imply an extremely deep chamber (30 to 40 km) if the generally conceived model relating ring fractures to underlying magma chambers proved valid at such depths [see Heiken's (1978) discussion of this model vis-à-vis the Medicine Lake caldera]. However, the ring fracture concept probably does not apply, at least to the Pleistocene rhyolite field, because of structural reasons discussed earlier, because basalts have erupted repeatedly from within the area vertically above the presumed magma chamber, and because no eruptions or hydrothermal manifestations have occurred along the ring fractures.

Another possibility is that the center of the rhyolite field, from which rhyolite has vented in each of the latest five periods, overlies only the margin of a larger magma chamber, instead of encompassing an entire smaller chamber as Bacon et al. (1980) suggest. Such a chamber could be centered west of the rhyolite field, perhaps bounded by recent basaltic eruptions on the south and southwest and by the Sierran front on the west. Although the lack of eruptive or heat flow evidence for this model makes it seem unlikely, it remains plausible by analogy with Glass Mountain, adjacent to the Long Valley caldera. The Glass Mountain rhyolites are very similar to the Coso rhyolites in composition, style of eruption, and near absence of phenocrysts (Noble et al., 1972, in Long Valley reference list). They erupted from vents along a probable incipient ring fracture system and represent early leakage from the Long Valley magma chamber (Bailey et al., 1976, in Long Valley reference list).

Any attempt on strictly geological grounds to assess the depth to the top of the Coso magma chamber must be only qualitative. It is likely that the chamber is comparatively deep, since the eruptions tapped the volatile-enriched upper fraction of the chamber, yet explosive activity was relatively minor. High Cl and F contents in the rhyolites and the presence of micro-phenocrysts of hydrous minerals also point to a high magmatic volatile content. Thus much of the water originally present in the magma may have been lost during a long ascent from a deep reservoir (Bacon, 1982).

Eruption from a shallower magma chamber (<5 km?) would be likely to proceed differently, either (1) explosively, in predominantly pyroclastic eruptions, with rapid bubble growth resulting from exsolution of volatiles once the confining pressure is released at the start of eruption; or (2) nonexplosively, but guided by ring fractures which open in response to the shallow buildup of magmatic pressure (Bacon, 1982). The latter mechanism describes ring fracture eruptions originating from large, shallow magma chambers that ultimately (as at Glass Mountain, Long Valley), or previously, formed calderas (Smith and Bailey, 1968). At Coso, however, small leaks from the magma chamber did not occur along arcuate faults, suggesting (1) that the chamber had not migrated to depths sufficiently shallow to permit the geometry of the chamber to exert a greater influence than regional structures on vent distribution; and (2) that crustal extension played a greater role than magmatic pressure in triggering eruptions.

Extension can thus be seen as a safety valve on explosive eruptions, preventing an extreme buildup of volatiles by causing periodic bleeding off from the upper level of the chamber. The rate of crustal extension at Coso may be fairly constant, as argued by Bacon (1982), who finds that the time interval between any two successive groups of eruptions is proportional to the volume of the first of the two groups. This seems to be true of the Quaternary basalts as well as the rhyolites. This pattern is consistent with a model in which extensional strain, accumulating in roof rocks at a constant rate, causes eruptions when it reaches a critical value. It contrasts with the opposite pattern, in which erupted volume correlates with the preceding repose time between eruptions. The latter pattern is consistent with a system in which a steady buildup of magmatic pressure causes eruptions when it reaches a critical value or in which the volume erupted is a large fraction of the total reservoir that is being supplied at a constant rate (Bacon, 1982), as is observed in some large central volcanoes (Smith, 1979). Thus the pattern observed at Coso suggests that the magmatic system responds relatively passively to crustal extension, erupting in small leaks from the top, and is consistent with a deep chamber in which internal pressure buildup is not the critical factor in causing eruptions.

The maintenance of an erupting magma chamber and a convecting hydrothermal system for over 0.2 m.y. requires that heat be added continuously to the magmatic system by intrusion of basalt (Lachenbruch and Sass, 1978). Assimilation of crustal rocks due to this influx of heat would contribute greatly to the development of the silicic magmatic system, and studies of Sr and Pb isotopes at Coso support a mid- to upper-crustal source for the rhyolites and a much deeper source for at least some of the basalts (Bacon et al., 1981). Once the silicic system was well established, any basaltic eruptions would likely be confined to the periphery of the silicic volcanic field, and such a "shadow" is observed for basaltic eruptions younger than ~0.24 m.y. Supporting evidence of a basaltic heat source for the system (possibly in the form of discrete pulses that trigger rhyolitic eruptions from the magma chamber) comes from the close association of basalt with the rhyolites. Each of the groups of rhyolite eruptions either is associated with approximately coeval eruptions of basalt or contains inclusions interpreted to be quenched blobs of mafic magma (Bacon et al., 1980; Metz and Bacon, 1980).

The most recent eruptions at Coso could be several tens of thousands of years old, and this raises the question whether the magma chamber may have cooled and largely solidified since the last rhyolites were extruded. Several lines of evidence suggest it has not. The volume of rhyolite erupted in the latest group (<70,000 years) is greater than that of any of the earlier groups, and the present interval of dormancy is no longer than several earlier intervals. The latest rhyolites also show no significant growth of phenocrysts, as might be expected if the chamber were in a waning stage. Finally, there is no evidence that the processes ultimately responsible for the development of the silicic magmatic system--crustal extension and emplacement of basaltic magma into the lithosphere--have ceased or slowed. Indeed, the latest basalts are similar in age or perhaps younger than the latest rhyolites, and normal displacement on steep north- to northeast-trending faults continues to occur.

GEOPHYSICAL EVIDENCE FOR MAGMA AND ESTIMATED MAGMA DEPTHS

Heat Flow

On the basis of 25 shallow drill holes and one of intermediate depth, Combs (1975, 1980) outlined a heat flow anomaly whose peak is centered approximately between Sugarloaf Mountain (one of the <40,000-year-old rhyolite domes) and the Devil's Kitchen area of fumaroles and intense argillic alteration (Fig. 19). The heat flow anomaly is sharply bounded on the east by the north-northeast-trending normal faults that define the irregular eastern margin of the central horst of the Coso Range. Elongation of the heat flow contours along this fault zone suggests a fault-controlled, convective component in the heat flow. The heat flow anomaly is also elongated to the west. The anomaly shape suggests that the present heat source could be centered a few miles to the southwest of Sugarloaf Mountain. There is also seismic evidence for a magma west of Sugarloaf Mountain.

The depth to the heat source was estimated by Hardee and Larson (1980) to be about 5 km, assuming a spherical source and steady-state conductive heat flow. This could be a minimum figure because of uncertainties introduced by convective heat transport in the highly fractured and permeable rock (Combs, 1980) and because the thermal system is probably not in the steady-state condition. Most of the seismic evidence (discussed in a later section) favors a deeper zone of magma or partial melt.

Temperature Gradients

Recent geothermal development drilling by the California Energy Company (CEC), under contract to the U.S. Navy, has been done near the center of the heat flow anomaly. The holes have been located southward from the Devil's

Kitchen area a distance of approximately 1.5 miles. According to R. Adams (pers. comm., 1984) the bottom-hole temperatures beneath Devil's Kitchen are approximately 210°C at 500 m; temperatures of 250°C at 850 m were encountered in a deeper well 2 km south of Devil's Kitchen. The CEC hole temperatures and gradients have been larger than those encountered in the DOE Coso Exploration Well (CGEH-1) (Galbraith, 1978), which was drilled 2 km north of Devil's Kitchen but also within the zone of maximum heat flow. Temperature logs made after the thermal equilibrium in the DOE well was restored indicate a maximum temperature of about 190°C at 670 m and a negative temperature gradient below 900 m to the total depth of 1200 m.

Gravity

The gravity data give no indication of a gravity low that one might associate with a shallow, silicic magma chamber beneath the Coso volcanic field. However, a deep zone of partial melt producing a small density contrast ($\sim$0.1 g/cm^3) would produce a small anomaly (peak value $\leq$2 mGal) that would be difficult to discern in the background of other density inhomogeneities (Plouff and Isherwood, 1980).

Magnetic

Fox (1978b), Plouff and Isherwood (1980), and Roquemore (1984) have examined high- and low-level aeromagnetic and ground magnetic survey data taken over the area. It is generally agreed that most of the individual magnetic anomalies are associated with specific rock types mapped at the surface, with large structural features (such as the alluvial valleys), and with possible concealed mafic plutons (such as those inferred both south and north of the heat flow anomaly). There are, however, no inferences that can be drawn relating the aeromagnetic anomalies to a magma body or heat source.

To our knowledge there is no published information on Curie temperature depths beneath the area.

Seismological Investigations

Earthquake Focal Depths

Seismic studies of Walter and Weaver (1980) have shown an abundance of seismic activity of magnitude 0.5 to 3.9. During two years of seismic monitoring (1975 to 1977), 4216 local earthquakes were located in a 2000-km^2 area around the Coso geothermal field. Plotting the distribution of $M > 1.5$ earthquakes with respect to depth, Walter and Weaver (1980) show a sharp increase in events from 1 to 6 km depth, and then a sharp decline in earthquake activity at depths >8 km. The depth distribution has the same character for earthquakes beneath the Coso volcanic field as for earthquakes occurring in the region. For this reason Walter and Weaver (1980) conclude that there is no evidence for rocks "near liquidus at shallow depth beneath the rhyolite field." It also seems possible to state that the earthquake focal depths do not support the presence of any significant melt fraction at depths less than 8 km.

Plotting <u>all</u> earthquakes located in the area along north-south and east-west sections through the area, Walter and Weaver (1980) show that there seems to be a definite loss of earthquake activity at depths >8 km beneath the Coso volcanic field relative to the regional picture. This general picture of earthquake focal depths had earlier been reported by Combs (1975).

Focal Mechanisms and Swarms

Overall, the first-motion studies of earthquakes occurring in the area show mechanisms consistent with north-south compression and east-west extension (Walter and Weaver, 1980). Within the geothermal area, there is mainly normal faulting and a small component of strike-slip motion along north-northeast faults. The normal faulting indicated by the earthquakes is roughly parallel to the alignment of rhyolite domes and to the alignment of surface faults (Duffield and Bacon, 1981).

Walter and Weaver (1980) reported that four of the six earthquake swarms monitored were within the rhyolite field, and two of those occurred directly below Sugarloaf Mountain. Other than these earthquakes, there was no evidence for unusual activity beneath the heat flow anomaly.

Combs (1975) reported that strain release in the Coso geothermal area seemed to occur primarily in swarm-type sequences of nearly continuous occurrences of microearthquakes. On the other hand, he reported that earthquakes outside the area exhibit the more usual mainshock-aftershock sequence.

Earthquake Magnitude Distribution

Statistically significant high-b values for earthquakes were found in the upper 5 km of the crust beneath the volcanic field relative to the surrounding rock. This implies shorter than average fault lengths beneath the Coso rhyolite field and, thus, more highly fractured rocks.

Attenuation Anomalies

An attenuation analysis based on 44 teleseisms observed at one 16-station and one 26-station array of geophones was made by Young and Ward (1980). Performing a three-dimensional inversion of the differential attenuations between each station and a reference station, Young and Ward (1980) characterized the relative attenuations in three zones beneath the Coso volcanic field:

1. A high-attenuation zone from the surface to 5 km depth beneath most of the Pleistocene rhyolite domes and flows. This is consistent with the picture of water-saturated and highly fractured rocks discussed earlier.

2. A low-attenuation zone from 5 to 12 km depth.

3. Three high-attenuation zones in the depth range 12 to 20 km. These might be related to the presence of magma. One of these zones is located west of Sugarloaf Mountain, another beneath Coso Hot Springs.

Velocity Anomalies

The three-dimensional velocity structure under the Coso Range was studied using the P-wave phases of steeply incident teleseismic waves. Lateral variations in the velocity structure beneath the range were determined by Reasenberg et al. (1980) by observing the relative arrival times of the P phases across a geophone array and by employing both ray tracing and inversion techniques to model the P-wave delays. The compressional waves studied had wavelengths of approximately 5 km, the geophone array was ~25 km in diameter, and the average geophone separation was ~5 km in the center of the array.

An intense, low-velocity zone was discerned in the middle crust. Ray tracing (forward modeling) gave an interpretation of a spherical zone, radius 5 km, buried beneath Devil's Kitchen at 15 km depth. The zone has a 10% lower velocity than the enclosing medium. However, this model is highly nonunique and served only as a guide to the three-dimensional inversion model. Three-dimensional inversion produced a more accurate fit to the data and revealed a low-velocity body between 5 and 20 km deep under Devil's Kitchen. The zone is approximately 5 km wide on top, becoming increasingly elongated in the north-south direction and more intense with depth. A maximum velocity contrast of 8.4% was estimated at between 10 and 17.5 km depth.

Combs (1975) also reported a low V_p in the surface on the basis of calibration shots provided by personnel of the Naval Weapons Center. The refraction study showed a zone of essentially constant P-wave velocity (4.75 km/s down to a depth of 5 km) underlain by a high-velocity zone

(6.0 km/s). Analyzing the S-P times in terms of a Wadati diagram, Combs (1975) found that the upper layer under the field is characterized by a low Poisson's ratio of 0.16 instead of the more normally observed values in the 0.25 to 0.30 range. The low value of the Poisson's ratio suggests fractured rocks with steam or a two-phase fluid.

Electrical and Electromagnetic Soundings

Reconnáissance Surveys

Areas of high and low resistivity were outlined by means of a telluric survey (Jackson and O'Donnell, 1980). Micropulsation energy in the 0.02- to 0.01-Hz bandwidth was used to estimate the variation in the longitudinal conductance, S over the area ($S = H/\rho$, where H is the thickness of conductive rocks above basement and ρ is the average resistivity). Stations were located 5 to 8 km apart. Resistivity lows were delineated over Rose Valley and the Coso Basin. A subsidiary low was found over the Coso Hot Springs-Devil's Kitchen area and the general area of the heat flow high. Resistivity highs were located southwest of the rhyolite field near Volcano Peak (Fig. 18), west of the field and in a broad arc from north to east. All these highs are presumed to be caused by higher-resistivity basement rocks at or near the surface, and there is a rough correlation between these highs and aeromagnetic highs.

A 7.5-Hz audiomagnetotelluric (AMT) survey was also made in a somewhat smaller area around the geothermal area (Jackson and O'Donnell, 1980). Station separations of 2 to 5 km were employed, and the AMT resistivity pattern agrees very well with the telluric pattern. A prominent resistivity low (6.3 ohm·m) was mapped in the east-central part of the rhyolite field; the low extends northeastward beyond Coso Hot Springs and into the valley. This low is centered between Devil's Kitchen and Coso Hot Springs and may be fault-related, but no mapped faults with the same orientation occur in the area.

The reconnaissance electrical surveys did not find any other areas of possible geothermal interest in the Coso Range, and the depth of exploration was too shallow to reveal anything about resistivity conditions at mid-crustal depths.

Electrical Surveys

Detailed dc electrical soundings were made by Furgerson (1973) and by the USGS (Jackson and O'Donnell, 1980). Fox (1978a) reported on the results and interpretations of several lines of dipole-dipole dc resistivity run across the geothermal field. The dc resistivity studies better define a bedrock resistivity low measuring 10 to 15 km^2 in extent associated with the geothermal system. This low-resistivity zone (10 to 20 ohm·m) is presumed to be caused by a combination of high fracture density, saline pore fluids, high temperature, and hydrothermal alteration. The zone probably extends to depths greater than 750 m and shows a north-south orientation at depth (Fox, 1978a).

The University of Oregon (H. Waff, pers. comm., 1984) has conducted 42 remote-reference magnetotelluric stations for CEC. The stations are located southeast, south, west, and northwest of Sugarloaf Mountain, and many soundings show pronounced splitting of the apparent resistivities in the principal directions. This is indicative of inhomogeneous subsurface conditions, necessitating a two-dimensional interpretation. Until a proper interpretation of the data is completed, it would be very difficult and premature to offer an assessment of the MT results.

MEDICINE LAKE VOLCANO

SUMMARY

Arguments have been made on a variety of geological grounds that a large silicic magma body may underlie the caldera at the summit of the Medicine Lake Highland volcano. The existence of a shallow (<15 km and probably <10 km deep) body of magma of uncertain size can probably be claimed with some assurance on the basis of vesiculation of basaltic magma during mixing with a reservoir of silicic magma prior to one of the recent rhyolite eruptions. As a pressure-dependent process, the vesiculation provides constraint on the depth at which mixing took place. The volume and longevity of the silicic reservoir, however, is more difficult to evaluate. Most of the geological evidence suggests that it, or any other shallow reservoir of silicic magma which may be present at Medicine Lake, would be a relatively small and short-lived body. However, the possibility that a sizeable magma is present cannot be ruled out, mainly because two strikingly similar rhyolite eruptions, very close in age, are located on either side of the summit caldera, 15 km apart.

In general, geophysical evidence for the existence and depth of magma beneath the Medicine Lake volcano is inconclusive. This may be due, in part, to the fact that there have been few published geophysical studies on this area compared to the other areas considered in this report.

Geophysical evidence does point strongly to a dense, resistive, high velocity zone of rocks directly below the caldera and extending from near the surface to depths >4 km below the surface. The probable cause is an assemblage of mafic dikes and larger plutons. The existence of high-temperature conditions is suggested by the large number of geothermal wells being permitted and drilled within the caldera by commercial geothermal developers.

GEOLOGICAL SUMMARY AND EVIDENCE FOR MAGMA

Chronology, Composition, and Magnitude of Volcanism

The Medicine Lake Highland is a broad Quaternary shield volcano, measuring 25 to 50 km in diameter, depending on whose estimate one wishes to accept. It is composed of basaltic andesite and andesite flows and tuffs that have accumulated to a thickness of ~1 km on a dominantly volcanic, late-Tertiary plateau (C.A. Anderson, 1941). At its summit is an 8-by-6-km, 100- to 200-m-deep caldera that formed prior to the end of glaciation (Eichelberger, 1975). The caldera is ringed by a rampart of mainly andesitic cones and domes contemporaneous with or slightly younger than the caldera. Since formation of the caldera, volcanism has been essentially bimodal. Basalt and basaltic andesite have erupted mainly on the flanks of the shield volcano, and rhyolite and dacite obsidian and tephra have erupted within and near the ring of the caldera (C.A. Anderson, 1941).

The most recent silicic eruptions include the Glass Mountain rhyolite-dacite, the Little Glass Mountain-Crater Glass rhyolites, and the Medicine Lake dacite, with a cumulative volume of ~1.5 km^3 (Fig. 20). Organic material associated with the flows and tephra of Glass Mountain have been dated by ^{14}C, yielding dates of 130 to 140 years and 1100 to 1400 years. Heiken (1978) reviewed geochronological data in light of observed field relations and concluded that all of the above-mentioned silicic eruptions, as well as the most youthful basaltic eruptions on the shield's flanks, appear to be about the same age, perhaps all younger than 1100 years.

Structural and Tectonic Setting

The Medicine Lake volcano is located to the east of the line of High Cascades volcanoes on the western margin of the Basin and Range. Normal

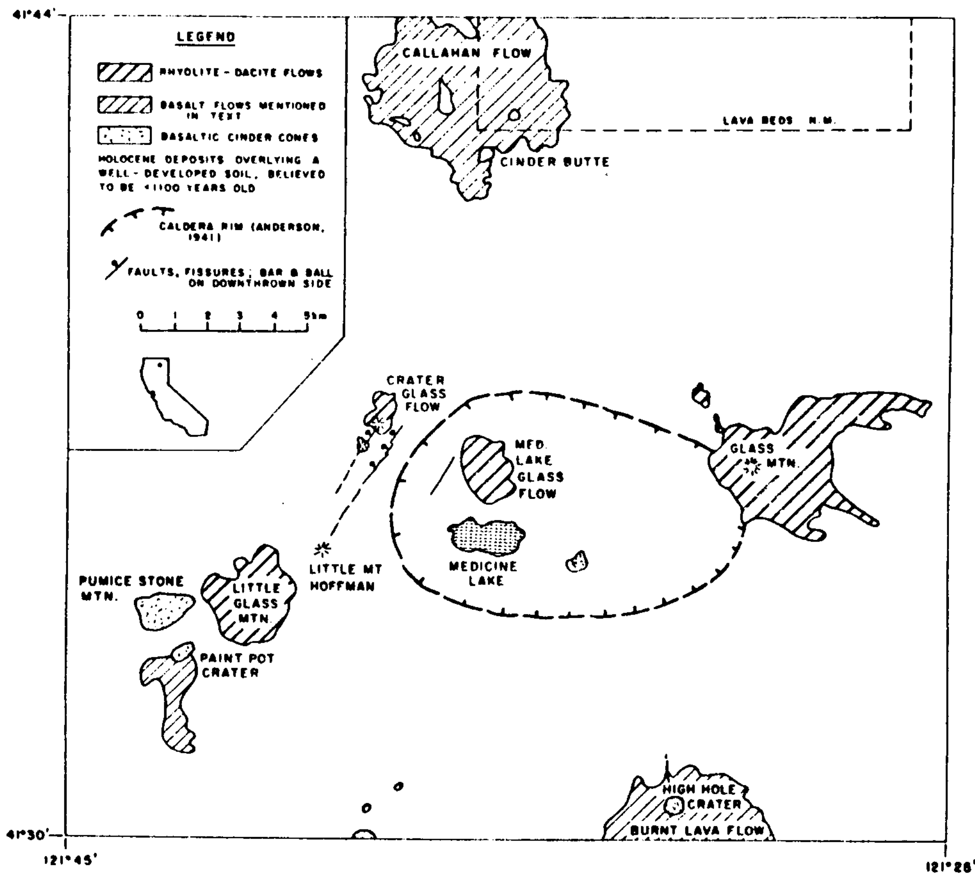

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Fig. 20. Distribution of Holocene deposits believed to be <1100 years old in the Medicine Lake Highland. (Directly from Heiken, 1978.)

faults trending approximately north-south, including the western margin of the Tulelake graben, cut the entire surface of the volcano (Ciancanelli, 1983), and numerous small scarps attest to the recency of faulting (C.A. Anderson, 1941). A major northeast-trending regional lineament cuts across the volcano near the northern rim of the caldera (Ciancanelli, 1983; Donnelly-Nolan, 1983), and northwest-trending normal faults are also present (Heiken, 1978).

Alignment of vents along faults is common in many of the young silicic and basaltic eruptions. Most prominent is the alignment of the Little Glass Mountain and Crater Glass flows along a northeast-trending zone that includes open fissures up to 3 m in width and which may connect with a concentric ring fracture system. Fink and Pollard (1983) studied this zone and concluded that the fissures overlie silicic dikes possibly as shallow as 30 m. The vents of the Glass Mountain flow are also aligned, in this case parallel to the northwest-trending fault set, and much of the youngest basaltic volcanism has occurred along normal faults of the same trend, as well as a north-south trend (Heiken, 1978). Heiken believes that the controlling factor in the formation of the caldera and in the eruption of the Holocene silicic rocks is the caldera's location at the intersection of two major sets of normal faults, one trending north-south, the other northwest. In this, as well as in other respects (C.A. Anderson, 1941), the Medicine Lake Highland resembles Newberry Volcano in Oregon.

Hydrothermal System

There are few surface manifestations in the Medicine Lake Highland of a geothermal reservoir at depth. One present-day site of fumarolic activity is the Hot Spot, located 0.5 km west of Glass Mountain, where temperatures >80°C occur at 0.5 m depth in an area of ~1 acre (Eichelberger, 1975). Two relict fumaroles, within the caldera and near the northern rim, have also been recognized on the basis of altered pumice and tuff (J.M. Donnelly-Nolan, pers. comm., 1984).

The near-absence of surface hydrothermal manifestations may be due to influx of large volumes of meteoric water. This would occur through a combination of high rainfall and very high permeability of the young volcanic rocks, resulting in thorough dilution of geothermal fluids by large aquifers, as suggested by Mase et al. (1982). They report zero heat flow for the Medicine Lake Highland area, although their measurements were taken on the flanks of the volcano and may not represent locally higher heat flow in the area of the caldera at the summit. Donnelly-Nolan (1983) suggests that hot water from the Medicine Lake volcano may migrate northward 30 to 50 km toward the structural low at Klamath Falls, Oregon, where a geothermal system exists that is unrelated to any recent extrusions.

Evidence Relating to Magma Origin and Magma Chamber Presence

Geochemical studies of the Medicine Lake volcanic suite indicate that crustal assimilation, fractional crystallization, and magma mixing were all factors in the origin of the silicic rocks. Magma mixing and its possible role in triggering eruptions have been documented (A.T. Anderson, 1976; Gerlach and Grove, 1982), most notably in the case of the Glass Mountain dacite-rhyolite flow (Eichelberger, 1981). $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ data are consistent with partial melting of continental crust, and depletions in Sr and Eu are consistent with shallow fractional crystallization (Peterman et al., 1970; Condie and Hayslip, 1975; Grove et al., 1982). But despite the evidence for crustal assimilation and fractional crystallization, the recent silicic rocks do not show evidence of prolonged evolution in a stable magma body, such as the extreme differentiation observed in the Coso rhyolites or in ash-flows common at large calderas. Gerlach and Grove (1982) state that "none of our petrologic observations allow us to infer the characteristics of or even the existence of a magma chamber beneath Medicine Lake."

Several other arguments have been made in favor of a magma chamber as the source of the recent obsidian and tephra eruptions. Heiken (1978) presents a case for a small chamber centrally situated below the caldera. His argument is based mainly on two lines of evidence: (1) the striking similarity between Glass and Little Glass Mountains (located 15 km apart on either side of the caldera) with respect to age, eruption sequence, major- and trace-element compositions, and tephra characteristics, suggesting they were erupted from the same body of magma; and (2) recent rhyolite vents and fissures on the west and northeast sides of the caldera, which suggest alignment along inward-dipping ring fractures. By analogy with the geometry of ring fracture systems studied in other intrusive-volcanic complexes, Heiken infers an intersection of the ring fractures at the top of a chamber at a depth of 4 to 8 km (Fig. 21). Smith and Shaw (1975) also use the configuration of the caldera to infer characteristics of an underlying magma chamber. They hypothesize a chamber equal in area to that of the caldera (~75 km^2) and, on the basis of a very approximate relationship between chamber thickness and area (see Introduction), infer a chamber volume of 300 km^3.

There are several problems, however, with basing magma chamber models on the size of the caldera. One problem, which Heiken recognizes, is that a small basaltic cinder cone of Holocene age is present within the caldera (Fig. 20), suggesting that a silicic magma body either does not exist beneath the caldera or that it is very narrow. Another problem relates to the nature of the caldera itself. C.A. Anderson (1941) interprets the caldera as a col-

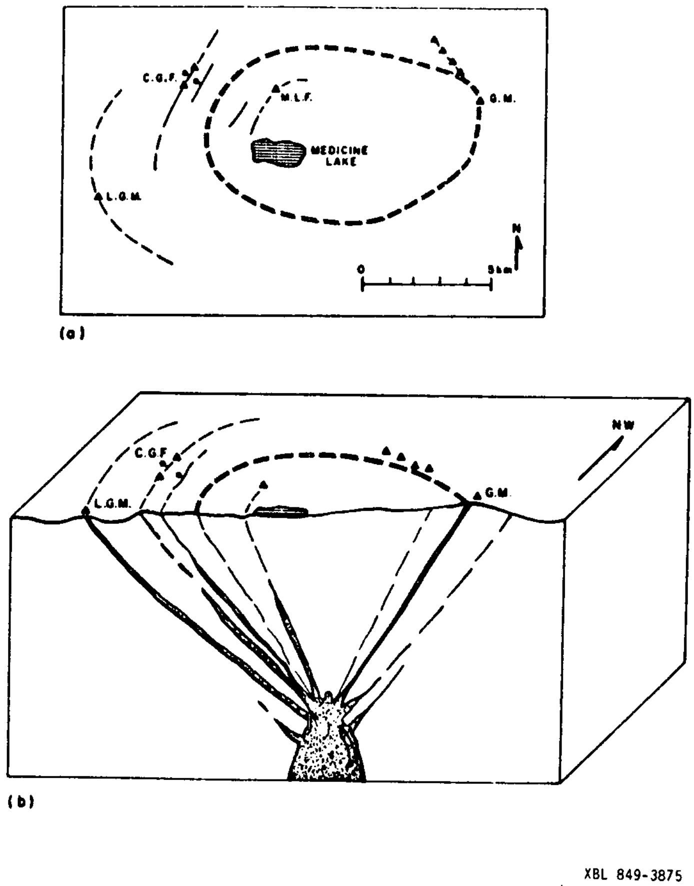

Fig. 21. Sketch map and block diagram of cone sheets proposed as conduits for Holocene rhyolites and dacites erupted in the Medicine Lake Highland. (a) Sketch map of the upper region of the Highland. L.G.M. = Little Glass Mountain; G.M. = Glass Mountain; M.L.F. = Medicine Lake Flow; C.G.F. = Crater Glass Flow. The double broken line is the location of Anderson's (1941) buried caldera rim. Solid lines are faults; bar and ball on downthrown side. The fault immediately east of the Crater Glass Flow is an open fissure, limited in extent, that may have been opened above a rhyolite dike that did not reach the surface. (b) Block diagram, illustrating a possible cone sheet geometry developed over a relatively small rhyolite body below the highland. This geometry could explain the remarkable similarity of tephra and lavas at Little Glass Mountain and Glass Mountain, located 15 km apart. (From Heiken, 1978.)

lapse feature, and that view has found general acceptance. But the original walls of the caldera, which would have resulted from collapse, are not visible; Anderson attributes this to the andesitic flows along the rim, which built up until they overran and obliterated the walls. There is, however, considerable disagreement on whether the caldera is a collapse feature at all. Noble (1969) believes it is, and he and others (e.g., A.T. Anderson, 1976) suggest that its collapse may correlate with a unit known as the Andesite Tuff, the only major ash-flow unit in the region. Mertzman (1981) disagrees, and interprets the eruption of the Andesite Tuff as a much earlier event than the formation of the caldera. Donnelly-Nolan (pers. comm., 1984), in recent mapping at Medicine Lake, sees no evidence of collapse following eruption of the Andesite Tuff. But even if the caldera formed by subsidence following withdrawal of underlying magma, there is no evidence directly linking the recent eruptions of rhyolite and dacite magmas to that earlier magma body.

The interpretation of the fractures controlling the alignment of rhyolite vents on either side of the caldera is also open to dispute, given the abundance of active faults cutting the volcano. For example, Donnelly-Nolan (1983) suggests that the northeast trend of faults and fissures near Little Glass Mountain may be related to a major regional normal fault. In the end, perhaps the best of the arguments favoring a sizeable body of magma as the source of the recent rhyolites is that based on the similarity of the two widely separated Glass Mountains.

Eichelberger (1980, 1981) takes a different approach in making a case for a high-level magma chamber. His model is derived from a detailed study of the rhyolite-dacite eruption of Glass Mountain, which he interprets, as have others (e.g., C.A. Anderson, 1941; A.T. Anderson, 1976), as an example of the mixing of two magmas. Eichelberger proposes that preceding the Glass

Mountain eruption, basaltic magma was injected into a silicic magma chamber, and the evidence for this mixing is found in abundant basaltic xenoliths. The observation that is central to Eichelberger's interpretation is that the xenoliths are vesicular, despite the virtual lack of vesicularity of the rhyolitic host rock. Although the basaltic magma was significantly denser than the rhyolitic, its density was reduced sharply by vesiculation as it cooled and crystallized in the cooler host magma. Blebs of basaltic magma that separated from the main body of intruding magma were able to float upward to form a layer of frothy basaltic xenoliths at the top of the silicic chamber. The cause of vesiculation in the basaltic blebs was exsolution of water from the basaltic liquid phase when its concentration in that phase was increased drastically during crystallization; water could not diffuse out of the basaltic liquid, because the rate of cooling of the blebs was far greater than the rate of diffusion of water into the rhyolite magma.

The critical variables in determining whether vesiculation and flotation can occur during mixing are water content of the basaltic magma, and pressure. Using the available data on original water content of basalts, Eichelberger estimates that to produce the observed vesiculation, the mixing must have occurred in the range of 7 to 15 km depth, and probably <10 km depth. This interpretation provides the best estimate, on purely geological grounds, of the depth to a magma reservoir at Medicine Lake (although considerable uncertainty is introduced through lack of adequate constraints on the water content of basaltic magma). It is the only interpretation that is based upon a process that is pressure and depth dependent.

Eichelberger goes further in his interpretation to propose that the magma body is large as well; he cites the following evidence: (1) similar mixed eruptions located near Glass Mountain and elsewhere near the summit of the Highland suggest that intrusion of basalt into a large, long-lived chamber is a continuing process; and (2) a similar mixture erupted from Mount Mazama just prior to the climactic Crater Lake eruption, which clearly did originate from a large shallow magma chamber. These arguments are difficult to assess, for in neither case does Eichelberger describe in detail the eruptions to which he is referring or in what respects they are similar to the Glass Mountain mixed magma eruption. The model he proposes (Fig. 22) still relies heavily on the geometry of the caldera in its estimate of the dimensions of the underlying magma body; the pitfalls of this approach were discussed earlier. He also neglects the presence of the aforementioned basaltic cinder cone (Fig. 20) within the boundaries of the caldera.

An alternative view concerning magma at Medicine Lake is that any silicic magma bodies are small, disconnected, and short lived (J.M. Donnelly-Nolan, pers. comm., 1984). Such a view is consistent with most of the geological and geochemical evidence and with a model that Eichelbrger (1978) proposes for areas in which crustal extension plays a major role, as it may at Medicine Lake. Possibly the best argument in favor of a magma body of significant size remains the strong similarity between the widely separated Glass Mountain and Little Glass Mountain eruptions.

One further point that runs counter to the notion of a large, high-level magma chamber at Medicine Lake is that made by Mase et al. (1982) on the basis of heat-flow measurements, referred to earlier. They suggest that large volumes of cold water circulating at depth within the volcanic pile may have "serious implications for the development and thermal longevity of 'shallow' crustal magma chambers. In such an environment, the enhancement of cooling by convective over conductive heat loss will have a profound effect on the cooling rate of the magma chamber...." Their views conflict, however, with

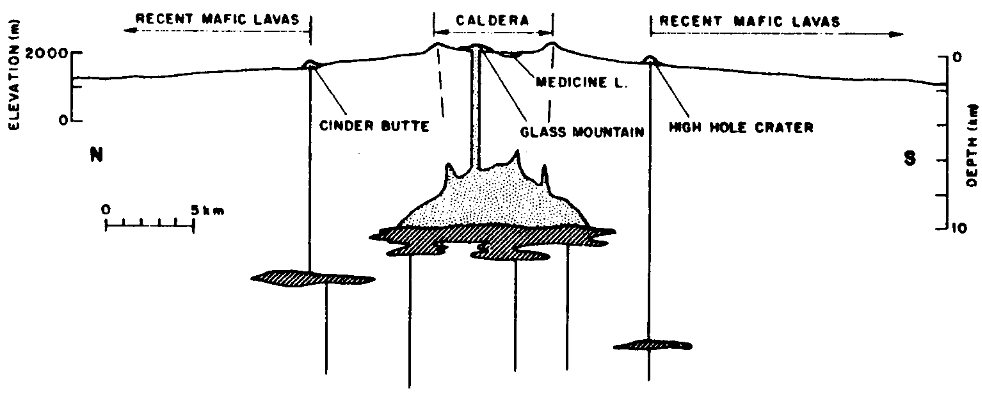

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Fig. 22. A north-south cross section through the center of the Medicine Lake Highland volcano, with major features projected onto the profile. Vertical exaggeration of surface topography is 2X. Silicic magma chamber is shown at depth discussed in text and intermediate in size betweeen the minimum case, a small body at intersection of cone sheet caldera fractures (Heiken, 1978), and the maximum possible extent marked by position of flanking mafic vents. (From Eichelberger, 1981.)

those of Smith and Shaw (1978), who state that "calculated times for solidification of magma by either conduction or magma convection models are affected negligibly by the presence and amount of water in the surroundings," and that water "cannot ... greatly influence the interior cooling regimes that govern the overall duration of magmatic crystallization."

GEOPHYSICAL EVIDENCE FOR MAGMA AND ESTIMATED MAGMA DEPTHS

Heat Flow and Subsurface Temperatures

Accurate estimates of temperature gradient and heat flow have been difficult to obtain in many parts of the Cascade Range because of the cold-water, or "hydraulic," masking effect that exists to depths of 300 m. This is generally, and probably correctly, attributed to the high rates of infiltration and percolation by rainfall and snowmelt (e.g., Black et al., 1982). This effect also explains why the Medicine Lake Highland and the Quaternary volcanic field northeast of Mount Lassen are depicted as areas of zero heat flow (Mase et al., 1982). Below the cold water surface layer the subsurface temperatures are likely to be similar to those encountered by deep holes in other areas of the Cascades. For example, a bottom-hole temperature of 265°C at 932 m was recorded in the Newberry 2 well (Sammel, 1981, 1983). On the basis of their heat flow studies in the Cascades, Blackwell and Steele (1983) argue that it is not inconceivable to expect temperatures of about 800°C and partial melt at 10 km depth.

Exploration and temperature-gradient drilling is currently in progress at the Medicine Lake volcano by several geothermal developers. Union Geothermal is the operator of a unit agreement that includes Union, Occidental, and Phillips. Republic Geothermal and California Energy are also active in the area. To date, several wells have been completed, two to at least 4000 ft.

Approximately another 40 wells have been permitted or are in the process of being drilled. One has a planned depth of 8000 ft. According to a BLM verbal communication, all of the wells are located within a 3-mile radius of Medicine Lake and within the boundary of the caldera. Exact hole locations, depths, and temperature data have not been made public.

Gravity

The Bouguer gravity map of the Alturas sheet (Chapman and Bishop, 1968) shows a positive anomaly associated with the Medicine Lake volcano. The gravity high occurs within a large (60-mile diameter), roughly circular gravity low caused in part by low-density volcanic rocks (LaFehr, 1965). The low may also be partially related to subsidence (Blakely et al., 1983). After separating the Medicine Lake gravity high from the regional field and the field from a nearby interfering gravity high, Finn and Williams (1982) interpreted the 27-mGal Medicine Lake residual anomaly as a dense zone consisting of mafic plutons and dikes extending from 1.5 to 4.0 km below the caldera. The roughly cylindrical anomaly has a diameter of 9 km at its top, expanding to a 20-by-36-km ellipse at its base. There is no evidence in the gravity data to suspect a low-density silicic chamber below the caldera. The maximum gravity low from a small felsic chamber at a depth of 6 to 10 km would be <2 mGal, and thus it would be obscured by the larger shallow effects.

Aeromagnetic Studies

Although interpretation and analysis of the aeromagnetic data are not complete, Blakely et al. (1983) report that the Medicine Lake volcano occurs in an area of a broad magnetic low. This they feel may indicate a shallowing of the Curie-point depth. Connard et al. (1983) found Curie-point depths as shallow as 9 km beneath parts of the central Oregon Cascades, between Crater

Lake and Three Sisters. While that study has little direct bearing on the Medicine Lake caldera area, it is mentioned because the Curie-point depth implies temperature gradients of >60°C/km and heat flow >100 mW/m^2. These values are consistent with gradient and heat flow in the California Cascades (Mase et al., 1982) and in various wells around Mount Hood, Oregon (Priest and Vogt, 1983).

Seismological Investigations

Catchings et al. (1983) reported on a seismic refraction experiment they conducted to investigate the deep structure beneath and adjacent to the Medicine Lake volcano. The shape of the conical, high-velocity body (6.1 km/s) is consistent with the gravity interpretation.

Differences in crustal structure around the caldera are attributed to variations in flow thicknesses due to fault offsets of up to 5 km. A prominent reflector was observed at 21 km depth beneath and southeast of the volcano. It could not be determined whether the reflector was caused by a positive or a negative impedance contrast. The former could be interpreted as local doming of the mantle. The latter might imply a magma chamber.

Amplitude attenuation patterns suggest that there may be partial melt below the volcano from the near surface down to the 21-km depth reflector.

Electrical and Electromagnetic

Regional Surveys

Stanley (1982, 1984) reported on the results of 10 regional magnetotelluric (MT) profiles conducted at widely spaced stations along east-west lines across the California, Oregon, and Washington Cascades. Line number 1 passed through the Medicine Lake Highland. On that line, MT sounding station 1-3 was located near Fourmile Hill, 10 km north of Medicine Lake and on pre-caldera olivine andesite (Anderson, 1941). A one-dimensional inversion of an average of the two apparent resistivity curves rotated to the principal directions (Table 2) gave no indication of a conductor that one might associate with a partial melt within 10 km of the surface. The low-resistivity second layer is probably Tertiary volcanics.

Table 2. MT resistivity layering, Medicine Lake (Station 1-3) (after Stanley, 1982).

Layer	Resistivity (ohm·m)	Depth to layer bottom (km)
1	>100	.3
2	7.1	1.7
3	216.	17.5
4	7.4	infinite

Table 3. TDEM resistivity layering, Medicine Lake (Station 8) (after Anderson et al., 1983).

Layer	Resistivity (ohm·m)	Depth to layer bottom (km)
1	1630	0.6
2	42	0.8
3	15	0.9
4	17	undetermined

dc Resistivity Surveys

Zohdy and Bisdorf (1982) conducted 50 Schlumberger soundings in the Medicine Lake area as part of the USGS geothermal research program. One of their stations (station 37) was close to Stanley's (1982) MT station 1-3. MT and dc resistivity compare reasonably well, although the dc resistivity soundings have much less depth of investigation. As part of the same USGS program, Anderson et al. (1983) conducted time-domain EM soundings using both single and central loop configurations with a SIROTEM MKII instrument. Table 3 shows the result of a one-dimensional data inversion at a station close to Stanley's MT station 1-3. The major point of difference is that TDEM seems to reveal a thicker first layer.

So that we could examine the results of the dc resistivity and TDEM inversions in cross section, we plotted the results for several soundings in a N85°E alignment passing through Medicine Lake (Fig. 23). The eastern end of the line is at a point south of Glass Mountain. The two techniques show various differences in layer thicknesses and layer resistivities that will not be discussed in this report. The major similarities are a resistive first layer, grading down to a conductive second layer at a depth of around 500 m. Notice that the conductor is best resolved close to Medicine Lake, but the region immediately below the lake appears to be resistive. The conductive second layer does not extend below 1 km depth anywhere except at Schlumberger sounding 2. From the figure one could infer a resistive plug beneath the lake, with zones of fractured rock and possible hydrothermal circulation flanking the plug. The resistivity cross section seems consistent with the gravity interpretation given by Finn and Williams (1982) and with the seismic refraction interpretation of Catchings et al. (1983).

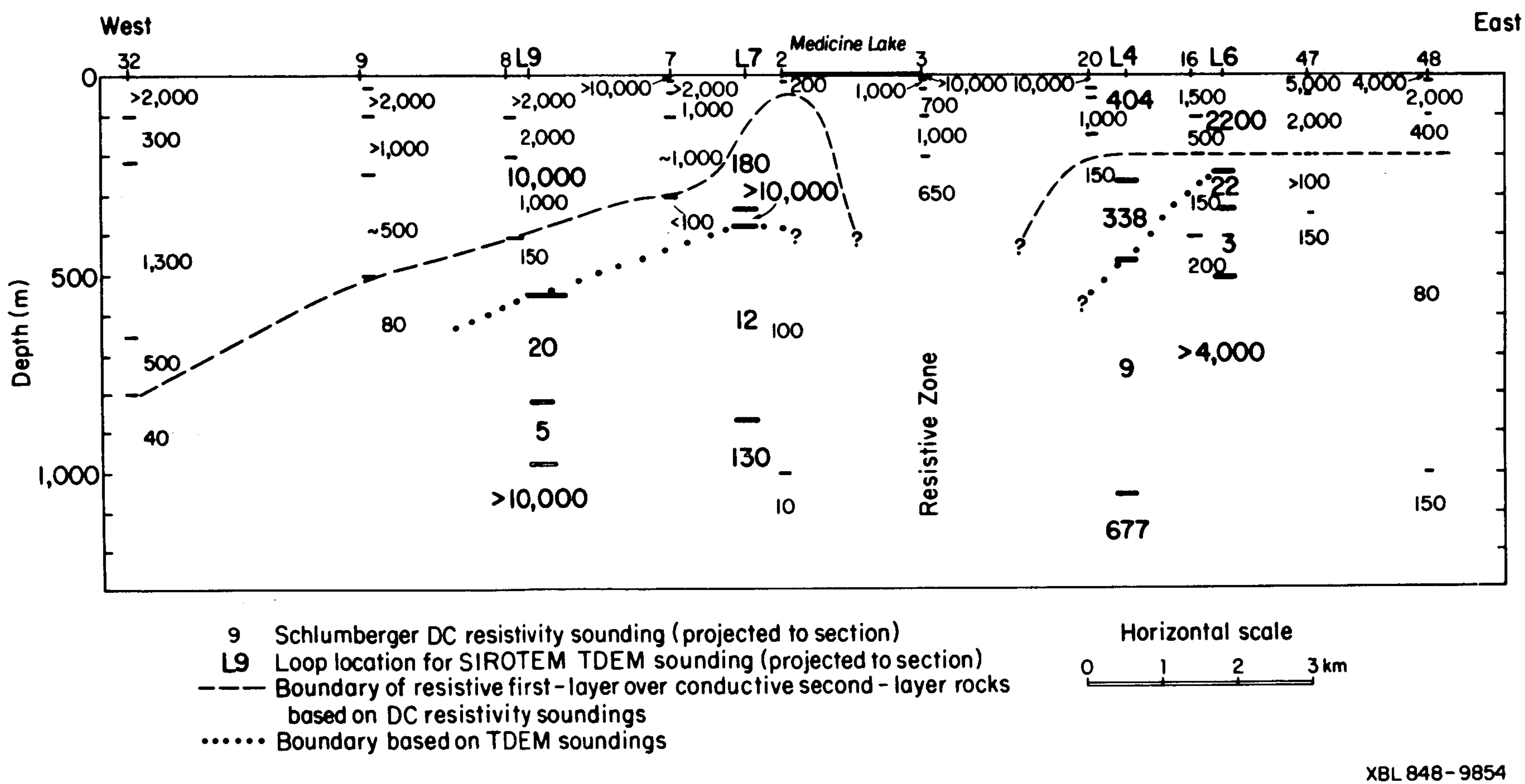

Fig. 23. Resistivity cross section through Medicine Lake based on dc resistivity and time domain EM soundings (values in ohm·m).

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