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Comparative Assessment of Five Potential Sites for Magma–Hydrothermal Systems: Geophysics

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Lawrence Livermore National Laboratory

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ABSTRACT

As part of a comparative assessment for the Continental Scientific Drilling Program, we used geophysical data to characterize and evaluate potential magma–hydrothermal targets at five drill sites in the western United States. The sites include Roosevelt Hot Springs, Utah, the Rio Grande Rift, New Mexico, and The Geysers–Clear Lake, Long Valley, and Salton Trough areas, California. This summary discusses the size, depth, temperature, and setting of each potential target, as well as relevant scientific questions about their natures and the certainty of their existence.

INTRODUCTION

In November 1979, a joint proposal to conduct a comparative assessment of five potential magma–hydrothermal sites was prepared and submitted to the Geoscience Program of the Department of Energy's (DOE's) Office of Basic Energy Sciences. Four DOE laboratories were involved in the proposal: Los Alamos National Laboratory, Lawrence Berkeley Laboratory, Lawrence Livermore National Laboratory, and Sandia National Laboratories. This proposal was a logical consequence of the Workshop on Continental Drilling for Scientific Purposes held at Los Alamos, New Mexico, July 17–21, 1978. The workshop resulted in a National Academy of Sciences report entitled *Continental Scientific Drilling Program* (CSDP Report, 1979). The workshop and report identified major scientific objectives in four areas: basement structures and deep continental basins, thermal regimes, mineral resources, and earthquakes. The Department of Energy has a particular interest in thermal regimes.

The five sites selected were Geysers–Clear Lake, California; Long Valley, California; the Salton Trough, California; Roosevelt Hot Springs, Utah; and the Rio Grande Rift, New Mexico. Their locations are shown in Fig. 1. Each site has indications of hydrothermal and magmatic activity, hot springs, high heat flow, recent volcanism, and a commercial interest in the possibility of producing geothermal fluids.

The four DOE laboratories have prepared a joint report (Luth and Hardee, 1980) that discusses the five sites, important scientific questions to be resolved by drilling, and recommended future research. Four detailed background reports, each of which examines the sites according to a particular discipline and reviews pertinent literature, are also being prepared. This report reviews interpretations of existing geophysical data that bear on the selection of one or more of these sites as candidates for further study; the other three reports cover geology, geochemistry, and energy transport.

Geophysical measurements can be used to study many characteristics of a site, including its regional setting and local structure (in three dimensions) and the location of specific objects such as magma or ore bodies, hydrothermal systems, aquifers, or alteration zones. The use of geophysical data to locate and characterize two classes of targets to be studied by the CSDP is emphasized in this review. The first class of targets consists of possible zones of molten or partially molten rock, the second of possible hydrothermal systems. The reader is assumed to be familiar with the regional setting of each location or to have access to summary papers listed under References. The geophysical data

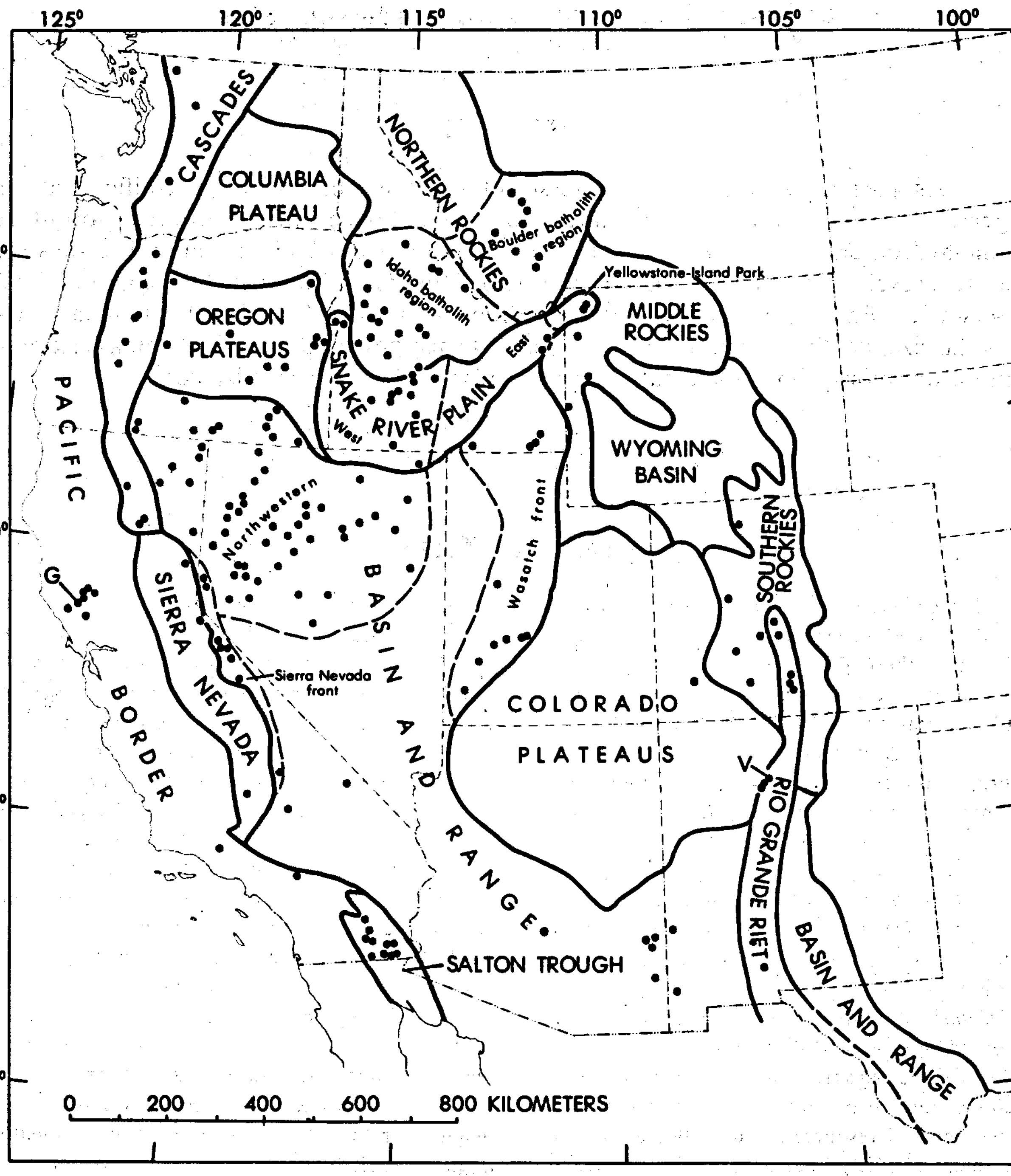

FIG. 1. **Map of the physiographic provinces and high-temperature hydrothermal systems of the western United States, showing the locations of the five potential magma-hydrothermal sites, after Brook, et al., 1979. The Salton Trough and Rio Grande Rift provinces are marked by their names. The other sites are represented by letters: R = Roosevelt Hot Springs, G = The Geyers–Clear Lake, L = Long Valley, and V = Valles Caldera.**

leading to that regional characterization are not discussed unless they bear directly on the existence of a magma–hydrothermal system.

Conclusions from geophysical data at each site are summarized in this report, first for potential magma targets, then for hydrothermal targets. The following questions are addressed for each class of target:

- What has geophysics told us about potential targets in the area?
- What additional questions remain to be answered about those targets?
- How could the questions be answered using geophysics?
- In what ways are these attractive targets for deep drilling?

SUMMARY OF THE WAYS GEOPHYSICAL DATA ARE USED TO STUDY MAGMA–HYDROTHERMAL SYSTEMS

Several geophysical methods have been used at each site discussed in this report. Although the manner in which data are used depends very specifically on the nature and setting of the target, general aspects of these geophysical methods and their applications can be described.

It is useful to remember that geophysical surveys may be used to infer such physical properties as electrical conductivity or density. In the applications discussed here, the objective is to locate zones of molten rock or zones with a convecting hydrothermal system. Consequently, two separate steps are involved in interpretation when geophysical data are used to study these objectives. Because both steps involve assumptions and are often nonunique, they are easily confused.

First, the observed data must be used to infer a distribution of physical properties within the ground. For example, one can determine the seismic velocities for a series of horizontal layers from the seismic-refraction arrival times. This part of the interpretative process is called "inverting" geophysical data, and the resulting model must represent the real world only if the data are good, the model is appropriate (in this case, the real world might have dipping beds or reflections from lateral features), and no other models fit the data. Since it is difficult in a summary article to assess whether all these requirements are met, the original researcher's judgment is usually trusted.

After the data are inverted, the physical property distribution can be interpreted in terms of a process or geological history. For example, a certain layer velocity in the seismic-refraction model might be interpreted to indicate the temperature at the base of the crust or to indicate whether the material is molten or solid. This part of the interpretive process can be very subjective, and the results of the interpretation can be influenced strongly by the geophysicist's desire to find what he is looking for. For example, a low-resistivity anomaly seen in a profile taken to study a geothermal reservoir is apt to be interpreted as a hot, saturated, porous rock even though clay layers may be equally conductive. At this stage, interpretations are often aided by laboratory measurements of the physical properties of materials under appropriate conditions.

When using geophysical interpretations, it is important to remember the two steps involved in those interpretations. For example, a statement that geoelectric surveys "found evidence for magma at 10–15 km depth" has to be treated as a convenient shorthand for "data were collected and inverted, using a particular subset of possible models (which may have been nonunique or inappropriate), to show low resistivity at depth, and that low resistivity was further modeled as being caused by partially molten rock."

The inversion and interpretative approaches used to evaluate targets at the five sites are described briefly. The details of measurement techniques and the routine methods used to determine structure and to aid a general geological interpretation are not discussed here.

Seismic-Refraction Data

Seismic-refraction data along a profile are inverted to cross sections of velocity and geometry of a small number of layers as a function of depth and position. If the profile is shot in two directions (i.e., reversed), both the dip and velocity of the layers can be determined; otherwise, inversions are made assuming horizontal layers unless dip or velocity can be inferred independently. Inverted models can be inappropriate if signals are refracted or reflected by lateral features, if velocity changes with distance in a layer, or if low-velocity zones exist and are not accounted for.

The models are interpreted to indicate several physical phenomena. Traditional applications include inference of the depth to basement and structures within sedimentary basins. Large-scale experiments may detect arrivals with high velocities (above 7 km/s). If those arrivals are interpreted as P_n signals, which travel along the crust–mantle boundary, they can be used to estimate crustal thickness. Some applications are specific to magma–hydrothermal regimes. High velocities detected in a sedimentary section may be interpreted as intrusions or as zones of hydrothermal alteration and deposition. Low P_n velocity might be interpreted as indicating anomalously high temperatures in the mantle.

Seismic-Reflection Data

Seismic-reflection data along a profile are inverted to determine the geometry of reflecting boundaries in a cross section beneath the profile. With multiple coverage of subsurface reflecting points, velocity profiles can also be determined. Inverted models will be inappropriate if signals were reflected from features to the side of the profile.

Seismic-reflection cross sections are generally interpreted as indicating structures within the crust, particularly in sedimentary areas, or depth to the crust–mantle boundary. In thermal regions, zones of anomalously high reflectivity may be interpreted as indicating zones of partial melt, particularly if there are indications of low velocity.

Seismicity Data

Seismicity is estimated either from historical records of felt events or from seismograms caused by local earthquakes. These data are inverted to produce maps showing the number and size of events in different areas. If enough accurate data are collected and a velocity structure is known, magnitudes and earthquake mechanisms can be plotted in three dimensions. Locations determined this way will be inappropriate if the velocity structure is wrong. In addition, local events are also used as sources for reflection and refraction studies. Zones of abnormally low seismicity within regions of high seismicity are interpreted as indicating the presence of a hot intrusion that is close enough to the surface for strain accumulation to be relieved by aseismic creep rather than by brittle fracture.

In this report, seismicity is interpreted as reflecting geological processes. Infrequent large earthquakes or "normal" distributions of smaller events are interpreted as reflecting major tectonic activity, such as rifting or shearing. Abnormally large numbers of associated small events and highly variable rates of seismicity are called swarms. Swarms are interpreted to reflect intense localized deformation, such as might be caused by the movement of magma within the crust or the fracture of rocks by pressures in the hydrothermal fluids.

Observation of Teleseismic Arrivals

Differences in the arrival times and amplitudes of signals from a distant earthquake can be observed at several nearby stations. These differences are caused by different delays or attenuations along ray paths beneath the receivers. By assigning a value for anomalous velocity or attenuation and by observing signals from earthquakes at different locations, one can invert the data to define a three-dimensional anomalous body below the receivers. This inversion can be nonunique since bodies with different velocities and thicknesses can produce the same data. With a sufficient variety of paths, one can make a three-dimensional map of velocity.

The interpretation of inverted teleseismic data depends on the results. Anomalously high velocities below a depth of 20 to 40 km are interpreted as indicating oceanic crust, rather than continental crust. Anomalously low velocities or high attenuation in the crust or upper mantle are interpreted as indicating partially molten material.

Gravity Data

Observations of the gravity field on the earth's surface are inverted to determine the three-dimensional distribution of relative density. (Observations are repeated in units of mgal where 1 gal = 10^{-2} m/s^2.) Simple shapes are often used because of the inherent nonuniqueness of the inversions. Sphere or disk models can be used to determine the maximum possible depth to the center of an anomalous mass, but many shallower distributions will fit the data as well.

The spatial extent of a gravity anomaly often determines the geological interpretation derived from it. Anomalies more than 100 km across are interpreted as reflecting the thickness of the lithosphere. Medium-scale (50-km) anomalies are interpreted to determine basement depth and structure in sedimentary areas. Sphere and disk models of localized anomalies are interpreted several ways.

In a sedimentary section, high relative density is interpreted as indicating deposition, alteration, or igneous intrusion. In an igneous or metamorphic region, low densities are interpreted as indicating hot or molten intrusions.

Magnetic Data

To determine the distribution of induced or remnant magnetization, aeromagnetic data are inverted in the same ways that gravity data are. The ambiguities of the inversions are increased by the vector nature of magnetization. Since rocks lose magnetization above the Curie temperature (about 600°C), anomalously highly magnetized zones in sedimentary areas are interpreted as representing cooled intrusions, and anomalously low magnetization zones in igneous or metamorphic terrains are interpreted as indicating hot or molten rock. The absence of long-wavelength magnetic anomalies in igneous terrains is interpreted as indicating the depth to the Curie temperature is usually shallow.

Gravity and magnetic data can be interpreted together if one assumes that the same target area has anomalous density and magnetization. This would be the case for an intrusive cooled below the Curie temperature. The magnetic field can be modified to produce a pseudogravity field. If that field looks like observed gravity, the resulting models can be interpreted as showing igneous rock.

Geoelectric Surveys

Several different methods involving active sources and the passive measurement of natural fields can be used to study electrical conductivity within the earth. Point measurements of field due to a source (i.e., roving-dipole methods) can produce a reconnaissance map that shows regions with generally high conductivity but yields little information about depth. Soundings, measurements over a range of frequencies (i.e., magnetotellurics) or spacings (i.e., Schlumberger soundings), are inverted to produce a conductivity–depth profile at a single location. Such interpretations are inappropriate if lateral inhomogeneities influence the soundings.

Electrical-conductivity structures are interpreted to indicate a variety of geological features. Water can provide the main conduction path, and temperature and salinity increase water conductivity dramatically. In sedimentary rocks, which are relatively conductive, zones with unusually low conductivity are interpreted as being dry or as being saturated with relatively cool, pure water. In igneous rocks, which are relatively resistive, low-conductivity zones are interpreted as fracture zones that have unusual amounts of fluid or as zones of partially molten rock. All the listed causes for high-conductivity anomalies, with the exception of clay, might be present in a magma–hydrothermal system. Consequently, high-conductivity anomalies are often assumed to represent the probable locations of such systems.

Heat-Flow Data

Heat-flow measurements determine the temperature equilibrium in drilled holes. For convenience, heat-flow data are measured in heat-flow units (1 HFU = 41.8 $mW\text{-}m^2$), and values of 1 to 2 HFU are common in normal areas. Inversion usually involves estimating the temperature at greater depth by assuming that the temperature field results from steady-state conduction, meaning that temperature gradients are very uniform. As the results described in this paper demonstrate, the temperature field can be quite wrong if that assumption is violated.

The derived temperature field can be interpreted as showing the depth or temperature of the heat source—either a hydrothermal system or a deeper intrusion. Three characteristics of heat-flow data are used to infer that convection is an important heat-transport mechanism: high heat flow, the variability of heat flow over short distances, and isothermal zones or temperature inversions. If a wide enough distribution of measurements has been made in a convecting system, models of its development can be made to determine its age and history and to allow extrapolation of temperature data below the hydrothermal system to find the heat source.

EVIDENCE FOR MAGMA AT FIVE SITES

OVERVIEW OF GEOPHYSICAL EVIDENCE FOR MAGMA AT DEPTH

Because the tectonic settings and modes of occurrence of intrusive rocks are different at the five sites, the means by which we could infer the existence of magma from geophysical data differs from site to site. Consequently, the way in which the data are discussed depends strongly on the general character of each site.

The sites can be placed in two groups: (1) those for which the occurrence of distributed bodies of magma is inferred from a regional tectonic model, and (2) those where geologic and geophysical evidence indicates directly the presence of a localized magma body. The Rio Grande Rift and the Salton Trough are in the first group; Long Valley, Roosevelt Hot Springs, The Geysers–Clear Lake area, and the Valles Caldera fall in the second group. The interpretation of geophysical data for sites in each group is discussed below.

Sites Associated with Regional Rifting

The sites associated with regional rifting share many characteristics. Most of the geophysical evidence for magma in the Rio Grande Rift and the Salton Trough comes from data supporting a regional model of tensional rifting, which produces zones where magma can intrude the shallow crust. Tensional deformation produces a thinner crust with elevated regional heat flow. Molten rock can migrate—perhaps passively to fill cracks or actively into zones of weakness in the crust. The inferred distribution of intrusion depends on the regional deformation at each site.

Tensional deformation in the Salton Trough is combined with a regional shear to produce local "leaky-transform areas." The regional strain in the Rio Grande Rift is more completely tensional, allowing the possibility of more extensive zones of magma emplacement. In both cases, much of the relevant geophysical evidence relates to the regional picture rather than to specific locations of magma emplacement. Such evidence includes gravity, large-scale seismic refraction and reflection, and regional heat flow. In addition, the areas share an unusual seismicity distribution—a mixture of major earthquakes (related to the regional tectonics) and swarms of small earthquakes—that some investigators interpret as being related to magma emplacement.

Sites Related to Single Magma Bodies

The sites related to single magma bodies share many characteristics. Most of the relevant evidence at these sites pertains directly to the inferred bodies and can be summarized more succinctly than the regional data required for the other sites. The most direct evidence relates to the molten nature of the rock and to seismic attenuation and delays, gravity lows, high local heat flows, and reduced seismicity. At some sites (e.g., Roosevelt Hot Springs), there is little direct geophysical evidence for the inferred magma body; at others (e.g., Long Valley), several geophysical methods indicate the body exists.

EVIDENCE FOR MAGMA AT THE RIO GRANDE RIFT

The Rio Grande Rift is a series of connected north–south-trending basins extending from Mexico to Colorado (see Fig. 2 for locations). The occurrence of young volcanism throughout the region makes the Rio Grande Rift a potential target for drilling toward magma.

Geophysical evidence for magma at the Rio Grande Rift can be divided into two sets. The first set pertains to a model of Rift formation and development, one of whose consequences is the intrusion of magma into the crust throughout the region. The second set contains those data that indicate directly the presence of magma at specific locations within the Rift.

Data Supporting the General Model of the Rio Grande Rift

The Rio Grande Rift is a major continental rift that consists of a series of interconnected north–south-trending grabens and associated flanking uplifts stretching from central Colorado across New Mexico to Chihuahua, Mexico, and west Texas (Tweto, 1979; Chapin, 1971; Woodward *et al.*, 1978; Kelley, 1979; Seager and Morgan, 1979).

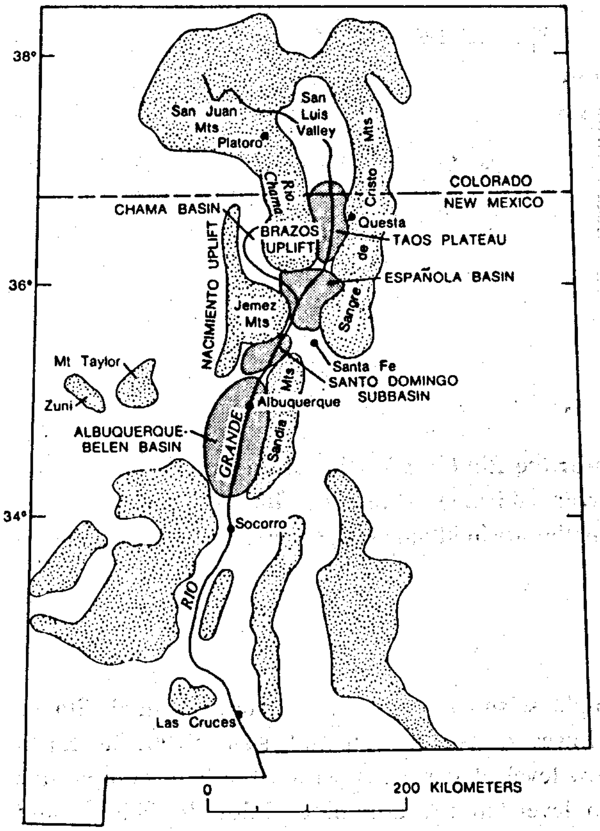

FIG. 2. Location map of the Rio Grande Rift, after Manley (1976).

In central and southern Colorado and northern New Mexico, the Rift lies between two salients of the Southern Rocky Mountain physiographic province that, in turn, separates the Great Plains to the east and the Colorado Plateau to the west. From central New Mexico southward, the Rift merges in a complex, and as yet poorly understood, way with the Basin and Range Province and the Chihuahua tectonic belt (Cordell, 1978; Seager and Morgan, 1979; Gries, 1979). The present phase of rifting began between 32 and 27 million years ago (Chapin, 1979), apparently as one of the earlier episodes in the widespread, late-Cenozoic, east–west extension that now affects a large part of the western United States (Eaton, 1979).

This regional model of crustal thinning and rifting provides paths for the upward mobility of magma. Several geophysical measurements support this model.

Seismic-Refraction and -Reflection Data. Most of the information on large-scale crustal structures near the Rio Grande Rift is derived from four approximately north–south-oriented unreversed refraction profiles (Stewart and Pakiser, 1962; Roller, 1965; Toppozada and Sanford, 1976; and Olsen *et al.*, 1979). Their interpretation is summarized in Fig. 3. Profiles outside the Rift show a relatively thick crust and high mantle velocity. Stewart and Pakiser (1962) used the "Gnome" nuclear explosion near Carlsbad, New Mexico, as a source, and their profile is confined to the Great Plains Province along the eastern side of New Mexico. Their results indicate a thick crust (50 km deep) and normal P_n velocity in the Moho (between 8.0 and 8.2 km/s). Roller's (1965) Hansville/Chinle profile is the only profile that was reversed; it indicates Moho depths of 40 to 45 km and a P_n velocity of 7.8 km/s in the Colorado Plateau. These values for the Colorado Plateau are also supported by surface-wave dispersion studies and by gravity data in the area (Keller *et al.*, 1979).

The north–south "Gasbuggy" profile interpreted by Toppozada and Sanford (1976) lies along a transitional zone between the Colorado Plateau and the western margin of the Rio Grande Rift, about 50 km west of Albuquerque. The model derived from the profile has a P_n velocity near 7.9 km/s under a 40-km-thick crust that can be resolved into two layers—a 19-km-thick, 6.1-km/s upper layer and a 21-km-thick, 6.5-km/s lower layer. There is evidence that, within the Rift boundaries where deep alluvial fill exists, the P velocity of the upper crustal basement rocks may decrease to about 5.8 km/s.

The 350-km south–north "Dice Throw" profile of Olsen *et al.* (1979) follows closely the axis of the Rio Grande Rift from central New Mexico to the Colorado–New Mexico border. The measured values of 34 km for Moho depth and 7.6 km/s for P_n velocity indicate a significant thinning of the crust along the Rift when compared to adjacent provinces. The low P_n velocity of 7.6 km/s is typical of values found in the Basin and Range Province and other active continental rift zones. Thus, these data support the concept that continental rift zones are regions of crustal thinning and spreading underlain by relatively shallow, hot mantle material.

Since 1976, the Consortium for Continental Reflection Profiling (COCORP) group (Brown *et al.*, 1979) has done extensive deep-reflection profiling near Socorro, and their results are consistent with the general model of rifting. Six Vibroseis lines, totaling 155 km, have been run across and parallel to the rift structures in this southern part of the

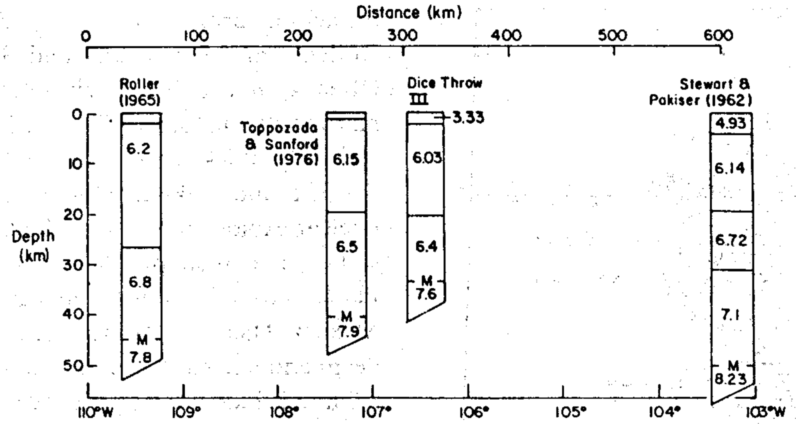

FIG. 3. Crustal models within and near the Rio Grande Rift, after Olsen *et al.* (1979). The profiles are arranged from west to east, with the westernmost profile at left. All velocities are in kilometers per second.

Albuquerque Basin. The processed seismic sections reveal thick alluvium overlying a basement with extensive normal faults and a large buried intergraben horst separating two arms of the Albuquerque Basin. The crust–mantle transition (Moho) near 34 km depth consists of a complex set of discontinuous (possibly laminated) reflection units that were not well resolved by the longer-period refraction studies. Much of the crustal section exhibits zones of complex, discontinuous reflections that probably indicate the presence of deformed and intruded metamorphic terrains. In recent processing of the shallow parts of the sections, Jurdy and Brocher (1980) have been able to derive compressional wave velocities for layers overlaying the Precambrian basement and to correlate these velocities with stratigraphic data obtained from nearby wells. Their analysis indicates strong and complex relief at the top of the basement.

Seismicity. Two classes of seismicity have occurred within the Rift: swarms and tectonic events. The frequent swarms of moderate events are interpreted as being associated with magma movement within the crust. These swarms are discussed in the "Data Indicating Magma at Particular Locations" section of this review, where data from specific sites are presented.

The record of tectonic earthquakes is ambiguous. Historical records of earthquakes felt in New Mexico from 1840 to 1961 were thought to indicate that the Rio Grande Rift zone had unusually high seismicity for the western United States. However, instrumental studies since 1962 indicate a low level of seismicity within the Rift, comparable to levels in the adjoining Colorado Plateau and Great Plains Provinces (Sanford *et al.*, 1979, 1980). Abundant geological evidence for major fault scarps indicates that large earthquakes have occurred within the last million years. The rate of seismic events measured over the last 20 years is inadequate to account for these major offsets. The low seismicity in the Rift and comparable seismicity in nearby tectonic–physiographic provinces may indicate a hiatus in the tectonism that is manifest by the presence of the Rio Grande Rift.

Other Geophysical Data. Gravity data coverage of the Rio Grande Rift zone in New Mexico is reasonably extensive (Cordell, 1976, 1978; Sanford, 1968; Kleinkopf *et al.*, 1970; Ramberg *et al.*, 1978; Williams, 1979A, 1979B). The gravity readings are fairly well corrected for terrain, which is important in mountainous topography. The gravity data have been used to map major graben border faults, low-density sediments, basin fill material, structural grain, and relief of the basement (Cordell, 1976, 1978). East–west gravity profiles across the Rift show broad lows (200 to 400 km in length and 75 to 100 mgal in amplitude) resembling analogous profiles across the East African rifts (Cordell, 1978). Various workers (Cordell, 1978) have interpreted this type of broad feature as being caused by thinning of the lithosphere over an asthenospheric bulge and/or by thermal expansion and intrusion-related density variations within the crust. Such interpretations are supported by the mantle velocity data derived from seismic-refraction profiling. Superimposed on the broad gravity low is an axial positive anomaly that is closely related to the main-rift graben structures. This axial high may be indicative of mantle material dynamically intruded into the crust. Aeromagnetic survey data are also available (Cordell, 1976, 1978) and were used, in conjunction with gravity data, to map trends in the basement structure and to estimate the depth to the Curie isotherm.

Geoelectric measurements indicate high temperatures in the crust and upper mantle beneath the Rio Grande Rift. Most of the published geoelectric work has used geomagnetic variation studies and telluric–magnetotelluric techniques that are capable of penetrating depths between 5 km and the upper mantle (Jiracek, 1980). Pioneering geomagnetic variation studies in the southwestern United States (Porath, 1971 and Gough, 1974) showed that a broad band of relatively high conductivity (about 200 km wide) follows the Rio Grande Rift through New Mexico and Colorado. Conductivity values in this zone are comparable to or greater than those found over much of the Basin and Range Province and considerably greater than those found in the adjacent Great Plains Province and the Colorado Plateau.

Reiter and others (1976, 1978, 1979) have carried out extensive heat-flow measurements in available drill holes throughout New Mexico, eastern Arizona, and southern Colorado. Their results support earlier conclusions of Decker and Smithson (1975). A zone of high heat flow (generally greater than 2.5 HFU) follows the trend of the Rio Grande Rift throughout most of New Mexico. From the Colorado–New Mexico border at 37 deg north latitude to about 35.5 deg north latitude just south of the Jemez volcanic field, the ribbon-like band of high heat-flow values is approximately 50 km wide and follows the center of the Rift. From 35.5 deg north latitude to south of Socorro, the band narrows to about 20 to 30 km and shifts to more closely follow the western boundary of the Rift. The high heat-flow band broadens again near Socorro to about 5 km and becomes more complex and variable. At four or five sites along this band within New Mexico, small regions of high heat flow (greater than 4.5 HFU) are found; these local regions are discussed in the following section. The heat-production data of Edwards *et al.* (1978) indicate that the high heat flow is not caused by anomalously high radioactivity in the crust. Analysis of the amplitude and width of the regional thermal anomaly suggests that partially molten rock must underlie the Rift zone at a depth of 15 to 30 km.

These geophysical data are all interpreted to support a regional model of the Rio Grande Rift that makes intrusion of magma likely. Additional data indicate that three sites probably had recent intrusions.

Data Indicating Magma at Particular Locations

Although the present-day trends of uplifts, grabens, and extensional faulting occur mainly along a north–south axis, several major magmatic and topographic features, as well as geophysical discontinuities at depth, suggest the inherited Precambrian structural grain (trending northeast–southwest) has played an important role in the evolution of the Rift (Cordell, 1976, 1978). At least three parallel northeast-trending fracture (or shear) zones intersect the Rift obliquely and have been important in controlling both structure and magmatism. These lineaments have, from north to south, been named the Colorado Lineament (Warner, 1978), the Jemez Lineament (Mayo, 1958), and the Morenci Lineament (Chapin *et al.*, 1978).

Mineralized districts in Colorado have long been known to follow a northeast-trending belt that runs diagonally across the state (Warner, 1978). The alignment of these localized ore deposits (the Colorado Mineral Belt) forms an important part of the evidence for the Colorado Lineament. Laramide and younger plutons in Colorado are most concentrated near the intersection of the Colorado Lineament and the northern "tip" of the Rio Grande Rift near Leadville.

A similar apparent close association of magmatism with the intersection of northeast-trending Precambrian fracture zones (the Jemez and Morenci Lineaments) with the Rio Grande Rift occurs in New Mexico near the Socorro and Jemez–Espanola areas. Many varied geological and geophysical investigations are currently underway along the Rift in New Mexico, and a large fraction of this work is being concentrated in these two general areas.

Socorro Area. Instrumental recording of regional earthquakes and local microearthquakes in New Mexico and the Rio Grande Rift began in 1960 when a high-gain, short-period instrument was installed at Socorro (Sanford *et al.*, 1979). Since about 1973, microearthquake networks have been installed and operated by (a) Los Alamos National Laboratory for the north-central New Mexico area near the Jemez Mountains and Espanola Basin, (b) the U. S. Geological Survey Albuquerque Seismological Laboratory for the Albuquerque–Belen Basin area of the Rift, and (c) the New Mexico Institute of Mining and Technology mainly for the area within 100 km of Socorro.

Studies of both historic reports of felt earthquakes and recent instrumental data indicate the area of the Rift between Socorro and Belen is the most seismically active segment of the Rift or, indeed, of the entire state. The largest known earthquakes within the state occurred near Socorro between mid-1906 and early 1907 as part of a protracted swarm-like sequence (Ried, 1911). Based on noninstrumental reports of felt areas and peak intensities, the three largest events of this series are estimated to have Richter magnitudes near 6.

Since about 1960, Sanford and his students/coworkers have used temporary networks of microearthquake recorders near Socorro to map and study details of local microearthquake activity (Sanford *et al.*, 1977). They have found that most of the activity is confined to depths above 20 km and that seismograms exhibit characteristic reflection phases in the codas that indicate a strong localized reflector about 20 km beneath the surface. The strong reflections come from an area of about 1700 km^2, shown in Fig. 4. These reflection signatures, along with data on screening and attenuation of S waves from other small local events, are interpreted as arising from seismic-wave interactions with a midcrustal sill-like body. Shadowing data indicate the presence of several smaller "pockets" of low-rigidity material lying above the main sill-like body.

The "magma-body" or "partial-melt" layer explanation for the microearthquake observations is strongly supported by other geophysical data. An important additional result of the "Dice Throw" measurements is that strong, wide-angle reflections are observed from a midcrustal discontinuity (about 20 km deep) near Socorro. The large amplitudes of these wide-angle reflections imply that a layer of low-rigidity material exists at the top of the lower

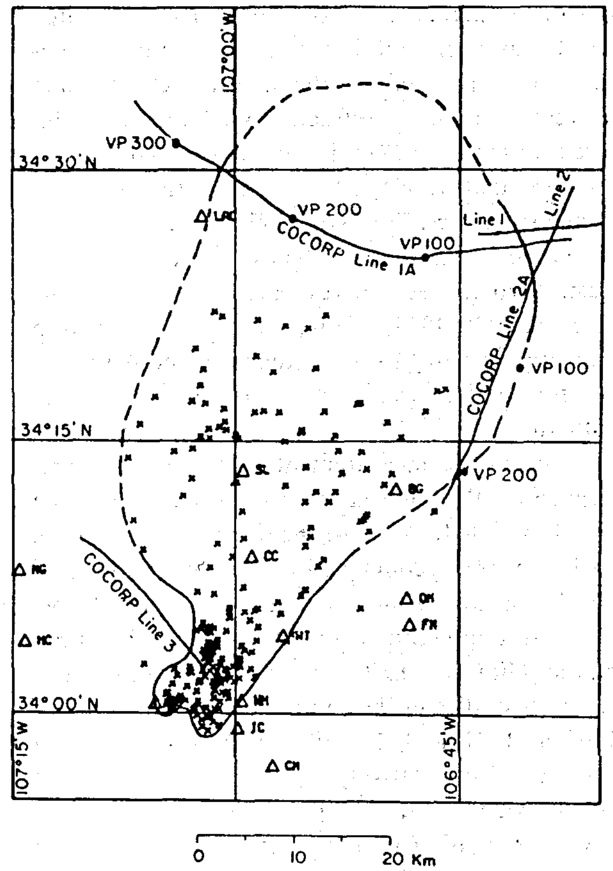

FIG. 4. Map of well-located reflection points for the midcrustal body near Socorro, after Rhinehart *et al.* (1979). The inferred extent of the body is indicated by solid and dashed lines. Reflections were absent outside the solid line; stations were not placed to sense the area outside the dashed line.

crust. Since low rigidity implies abnormally low shear-wave velocities, the reflecting layer at 20 km depth is most likely a continuation or closely related segment of the sill-like magma body that has been partially mapped by Sanford *et al.* (1977). Brown *et al.* (1979, 1980) report that a strong reflector is observed at about 20 km depth on the COCORP lines, also shown in Fig. 4, corresponding closely with the magma body inferred by Sanford *et al.* (1977). This anomalous seismic reflector coincides with one of the small heat-flow highs reported by Reiter *et al.* (1979). One COCORP reflection survey northwest of Socorro detected a bright reflector less than 3 km wide and at a depth of about 6.4 km. (Brown *et al.* 1980). The stacking velocity associated with this reflector is anomalously low, and the preliminary interpretation is that it is a small, shallow magma body.

Until recently, geodetic survey data precise enough to reveal contemporary tectonic movements in New Mexico have been confined to a few first-order, level-line surveys. Reilinger *et al.* (1979) discuss the results of repeated leveling surveys in three sections of the Rift. Perhaps the most outstanding of these is a zone of rather symmetrical uplift about 25 km north of Socorro. A maximum of 20 cm of relative uplift occurred between the 1911 and 1951 surveys. The uplifted region is dome-shaped and is closely centered over the sill-like body detected by Sanford *et al.* (1977). Modeling indicates that the observed surface deformation could reasonably result from magma movement during an inflation episode in a 20-km-deep magma chamber.

Savage *et al.* (1980) used geodetic techniques recently at a small geodetic network near the Socorro uplift area in an attempt to measure horizontal deformations in this seismically active section of the Rift. No horizontal strain accumulation significantly above the random standard error of measurement (±0.5 microstrain/year) has been found over the 1972–1979 measurement interval. These results, although consistent with horizontal components expected from the observed vertical rates of magma-body doming, appear to indicate a lull in the rate of tectonic crustal spreading across the Rio Grande Rift in this area.

Magnetotelluric measurements within the Rio Grande Rift also indicate the midcrustal target and suggest that its extent may be considerably greater than that determined by seismic studies. Jiracek *et al.* (1979) conducted detailed telluric–magnetotelluric measurements near the Socorro magma body. Their results show a decrease in crustal conductivity that could be interpreted as a highly conductive layer near the depth of the presumed magma body. However, there is considerable uncertainty in this interpretation because of the inappropriateness of one-dimensional models.

Hermance and Pedersen (1980) have recently examined the two-dimensional nature of the Rift and concluded that one-dimensional models can be used to interpret magnetotelluric data. They find a crustal conductor at depths of 10 to 17 km near Santa Fe and at a depth of 21 to 28 km near El Paso. The conductivity of the body is appropriate for 15 to 100% molten basaltic material. These results suggest that midcrustal bodies such as that detected at Socorro may be common throughout the Rift.

Espanola. The second most active seismic zone in the state during 1970–1980 is a 50-km-long belt northwest of Espanola, where the western flank of the Rift intersects the northern margin of the Jemez Lineament (Sanford *et al.*, 1979). Again, the apparent association of several geophysical anomalies with the intersection of two major tectonic features suggests a cause-and-effect relationship. Reilinger and York (1979) have analyzed the repeated leveling data along a line extending from Espanola northwest across the southern margin of this earthquake zone and find a pronounced area of subsidence (maximum 4.9 cm over 5 years) close to it. The presence of Tertiary dikes and high heat flow (Reiter *et al.*, 1975) tend to support a hypothesis that the seismic activity and surface subsidence feature are caused by deflation of a midcrustal magma chamber analogous to the hypothesized inflation near Socorro.

Jemez Volcanic Area. The Jemez volcanic zone is the dominant tectonic feature at the intersection of the Rift and the Jemez Lineament. The Jemez volcanic field is one of the largest and youngest silicic caldera complexes in the continental United States (10 to 100 million years old); only the Yellowstone and Long Valley (California) Caldera complexes are of comparable age and size. In the Jemez field, the most recent and possibly more extensive volcanic activity occurred in two major caldera-forming episodes 1.4 and 1.1 million years ago. Subsequent to the latest collapse—which formed the present Valles Caldera (about 22 km in diameter)—continued influx of magma created a central resurgent dome that is uplifted about 1 km above the Caldera floor.

Shallow heat-flow measurements near the Caldera are anomalously high, ranging from 4.5 to 6 HFU (Reiter *et al.*, 1976). Kolstad and McGetchin (1978) summarize the published deep-well temperature data near the Caldera and propose a model of magma emplacement that would produce those temperatures. They report that Union Oil observed temperatures of 229–279°C at depths of 1500–1700 m within the Caldera. The Hot Dry Rock holes 3 to 4 km west of the Caldera had lower temperatures, about 200°C at 3 km. The conductive heat flow in the deep holes outside the

Caldera is 3 to 4 HFU, slightly less than that measured at the surface. Linear extrapolation of these data, ignoring transient heat flow and convection, would place the 600°C isotherm at 6 to 10 km depth.

Kolstad and McGetchin determined the range of models of magma emplacement that could produce the observed temperature and heat flow at the bottom of one of the Hot Dry Rock holes. They modeled the pluton as a cylinder emplaced instantaneously at 1000°C one million years ago and calculated the cooling of that pluton by conduction. The best models require the pluton to have a radius of at least 12 km, with its top no deeper than 3 km. They indicate that, under these assumptions, 10 to 30% of the pluton would still be molten. They emphasize that the assumptions that the massive body was emplaced instantaneously and that convection can be ignored are significant simplifications.

Other geophysical data collected near the Jemez volcanic zone may indicate a large volume of hot material close to the surface. The Jemez volcanic zone is ringed by moderate earthquake activity in the upper 20 km of the crust but is notably aseismic in a large elliptical area (50 by 100 km) centered on the Valles Caldera (Olsen, 1979; Sanford *et al.*, 1979). It is not unreasonable that the contemporary microearthquake belt to the north may be related to magma movements in a rift-zone dike or sill complex related to the main magma chamber. The midcrustal conductor, detected using magnetotellurics east of the Jemez Mountains, has been interpreted as molten material that may be related to the source for the magma chamber (Hermance and Pedersen, 1980).

The inferred magma body near Socorro was detected on seismograms. Although several seismograph stations are located near the Jemez Mountains and the microearthquake belt just to the north, a search of seismograms shows magma-layer reflections, similar to those seen near Socorro, have not been detected (Olsen and McFarland, private communication, 1979). If a magma body exists here, the few stations in the area are not well placed for good reflection paths from it.

CONCLUSIONS ABOUT MAGMA BODIES AT THE RIO GRANDE RIFT

The regional model of tensional deformation for the Rio Grande Rift is generally accepted even though recent seismicity and deformation measurements fail to confirm it. Seismic refraction and reflection studies indicate complex grabenhorst structures in a sedimentary basin overlying a thin crust (34 km thick) and a mantle with low P_n velocity. Regional magnetic and gravity lows and inferred high conductivities for the upper mantle support this picture. Locally high and variable heat-flow zones are superimposed on a regionally high heat flow.

A logical consequence of that model is that regional crustal tension produces pathways for magmatic intrusion, and distributed, potentially molten intrusions are expected to be found in many parts of the Rift. Passive and active seismic-reflection methods have detected a particular target area near Socorro. This extensive target (at least 1700 km^2) is interpreted as being formed of thin, sill-like intrusions at a depth of 20 km and of smaller intrusions at shallower depths. Apart from their locations, as determined by seismic reflection, little is known about these bodies. Near-surface heat flow is high above the 20-km-deep body near Socorro, but the temperature of the body and the thermal structure between the body and the surface have not been studied. The vertical deformations observed above these targets can be interpreted as resulting from inflation or deflation of magma chambers at 20 km depth, but more studies are needed before that interpretation is accepted as conclusive.

Geophysical methods could improve our understanding of these features. More detailed heat-flow studies could better define the thermal regime above them and perhaps lead to an estimate of their temperature. On the other hand, teleseismic arrival time and attenuation studies would probably not detect thin, sill-like bodies. Further geoelectric studies could be useful (if the bodies are actually anomalously conductive), but they would have to be designed to deal with the three-dimensional nature of the bodies.

These bodies are attractive targets for deep drilling for several reasons. First, a reasonably certain target can be determined by geophysical studies before drilling begins. Second, the nature of the bodies—e.g., their age, temperature, composition, and relation to regional stress—is not understood; therefore, much can be learned about them by drilling. Third, there are many questions of general interest relating to the augmentation of continental crust by interactions in the rifting zones. Finally, a considerable body of geophysical data has been taken near similar features throughout the Rift. Consequently, a drill hole in one well-characterized part of the Rift would lead to verification and modification of interpretations of much of that geophysical data from all parts of the Rift.

On the other hand, these bodies have several drawbacks as targets for deep drilling. First, they are relatively small (particularly those at 4 to 10 km depth) and might easily be missed by drilling. Second, since small magma bodies cool rapidly, we cannot be certain what stage in the cooling process our target has reached. Third, the environment above the bodies must be well-characterized before a deep hole is sited; there are no other reasons for that characterization in this area (such as the potential for geothermal production). Therefore, there are few possibilities for "piggyback" experiments in shallow exploration holes near the target.

The Valles Caldera presents a different type of target. Extensive volcanism (as recently as one million years ago), high temperatures in the geothermal field, and the absence of seismicity suggest that a massive intrusive may be cooling beneath the Caldera. Unfortunately, current geophysical evidence does not define a target, much less indicate its depth or extent.

The Valles Caldera is an attractive site for deep drilling for several reasons. First, even though geophysical data do not define a magma body, it is easy to decide where to drill. A deep drill hole within the Caldera would determine the presence of a cooling body and help answer many questions about the eruption path and mechanics and about the history since the eruption. Second, having this knowledge about the Valles Caldera could increase our knowledge of eruptive volcanic systems in general. Finally, the near-surface thermal regime may be well-characterized as a result of industry efforts to produce geothermal power.

The Valles Caldera also has several drawbacks as a target. First, poor definition of what a drill hole might encounter reduces the effectiveness of scientific planning and drilling. This drawback might be removed by additional geophysical measurements aimed at characterizing the target more fully. Second, this type of volcanic system can be studied by examining older exhumed systems without the expense of drilling.

EVIDENCE FOR MAGMA AT THE SALTON TROUGH

Much of the geophysical data from the Salton Trough is not used to detect the presence of magma directly. Rather, it is used to develop and support a regional model of crustal thinning and extension. One implication of this model is the intrusion of magma into the crust. Data supporting the general model will be discussed first, followed by data related to the existence of magma at specific sites in the Trough.

Data Supporting the General Model of the Salton Trough

The Salton Trough (see Fig. 5) is a transitional region between the compressional and shear tectonics of the Transverse Range and central California and of the spreading centers in the Gulf of Mexico (Biehler *et al.*, 1964; Elders *et al.*, 1972). Several major strike–slip fault systems pass through the area. South of San Gorgonio Pass, which marks the northern limit of the Trough, the right-lateral San Andreas fault system splits into three subparallel strands, the Banning–Mission Creek, San Jacinto, and Elsinore fault systems. South of the Salton Sea, the three fault systems lose their distinct characters and merge into a series of smaller subparallel faults that continue down into the Gulf. The most important member of this smaller series of faults is the Cerro Prieto–Imperial–Brawley–Calipatria system. This group of right-stepping *en echelon* faults extends from the head of the Gulf north to the Salton Sea, where it joins the Banning–Mission Creek fault.

The Trough also has elements of a zone of crustal spreading (Biehler, 1964; Biehler *et al.*, 1964; Elders *et al.*, 1972). The spreading centers in the Gulf of California, which are essentially the

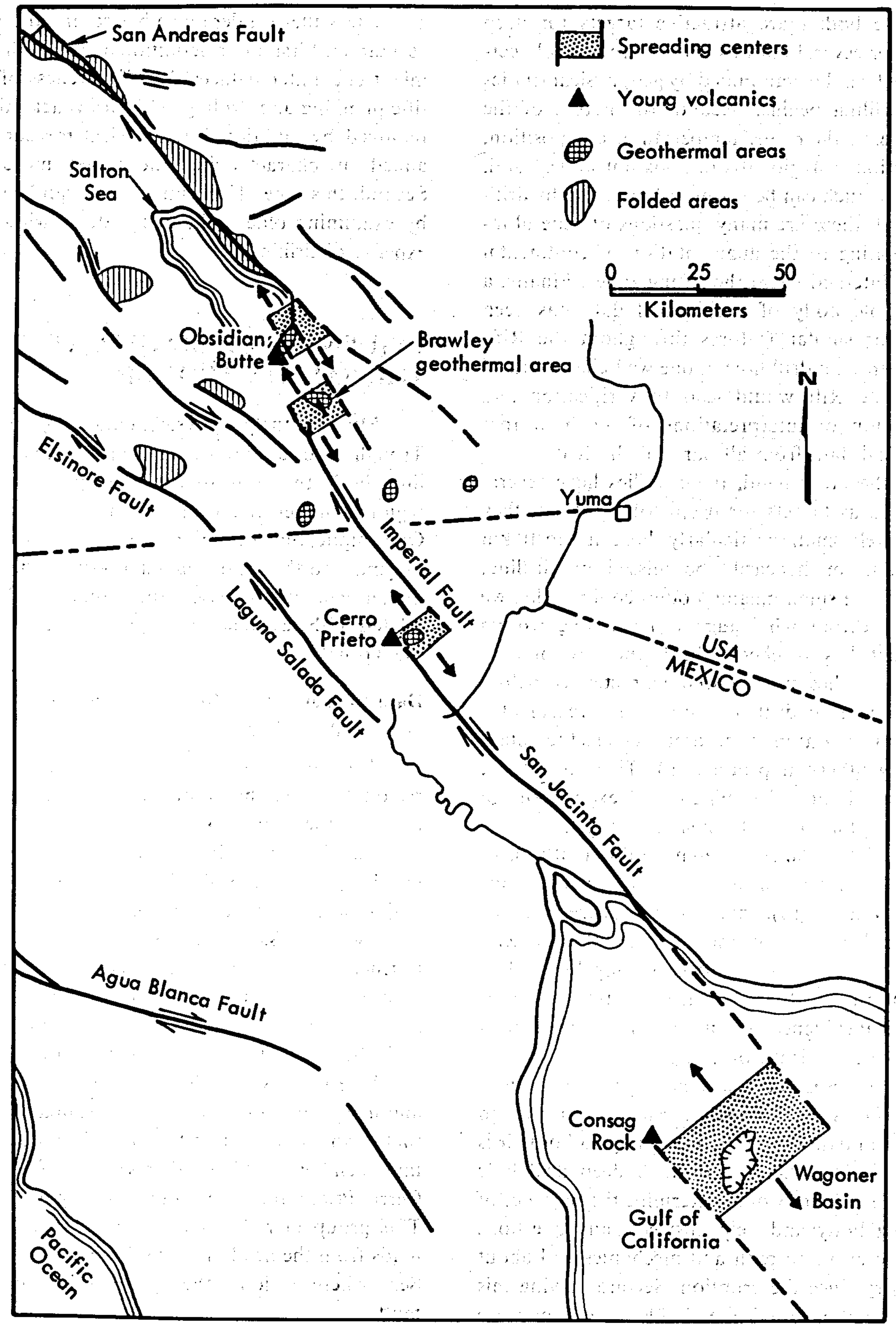

FIG. 5. Possible relationship between pull-apart basins and slip-strike faulting in the Salton Trough. Postulated spreading centers or tensional zones, young volcanics, geothermal areas, and zones of intense folding and compression in Tertiary sediments are indicated, after Elders *et al.* (1972).

northern extension of the East Pacific Rise, are evidently offset by right-stepping *en echelon* faults in a manner similar to that displayed on a smaller scale in the Imperial Valley (Larson *et al.*, 1968; Sykes, 1968). Lomnitz *et al.* (1970) suggested that the tectonic framework in the northern Gulf of California and the Salton Trough could be understood by considering these strike–slip faults as transform faults connected by short spreading centers. They postulated that active ridge segments account for the geothermal anomalies near Cerro Prieto and Salton Buttes. Elders *et al.* (1972) expanded and refined the model on the basis of the following geological and geophysical evidence:

- Anomalously high heat-flow areas scattered throughout the Trough.
- Active right-lateral strike–slip faulting, resulting in the formation of the Trough.
- Gravity models suggesting the crust is thinning.
- Geodetic measurements supporting the idea of steady right-lateral creep and dilation.
- Inclusion within rhyolite buttes of low-*K* tholeiitic basalt similar to material extruded on the East Pacific Rise. Rhyolite buttes may be a product of partial melting of the continental basement or of a two-stage melting process in the underlying mantle (Robinson *et al.*, 1976).

Using this evidence, Elders *et al.* (1972) suggest that active spreading centers occur in tensional zones (or rhombochasms) between *en echelon* strike–slip faults. They postulate spreading centers near the Salton Buttes, Brawley, and Cerro Prieto. Elders and Biehler (1975) and Hill *et al.* (1975) label these areas leaky-transform faults to emphasize that the dominant movement in the valley is strike–slip, with "spreading" taking place in a rather diffuse zone of offset strike–slip faults. Vonder Haar and Cruz (1979) have described a refined application of the model to Cerro Prieto.

The extensional nature of the region is responsible for the structure of the Trough: a deep depression filled with 2 to 6 km of sediments along the axis (Elders *et al.*, 1972). A reasonable tectonic model for Trough formation starts with uplift and lateral extension, followed by intrusion of magma into the lower crust (Elders *et al.*, 1972). About the same time, metamorphism of the lower sedimentary layers and gravitational slumping of the walls occur. Finally, melting of basement rocks leads to extrusion of magma at the surface near the regions of fault offset. Lachenbruch (1976) and Lachenbruch and Sass (1977) have calculated heat and mass budgets for volcanic centers similar to those associated with leaky-transform systems and have proposed that rapid local extension controls the passive rise of basalt through the lithosphere, thereby controlling the location of volcanic centers.

Geophysical Evidence for Crustal Thinning

Seismic, gravity, and leveling data all indicate crustal thinning has occurred in the Salton Trough.

Seismic Structure. General results show a gradual thickening of sedimentary layers from basement outcrops at the edges of the valley to as much as 6 to 7 km along the axis of the valley. Along the axis of the Trough, the basement is shallowest in the north (1.3 to 2.3 km), deepest in the Brawley–Westmorland area (5.6 to 6.6 km), and rises again to 3.4 km near the Mexican border (Biehler, 1964; Frith, 1978). All studies found a higher basement velocity in the center of the Trough, and Frith (1978) found the basement velocity near the Salton Sea Geothermal Field (SSGF) to be 15% higher than elsewhere. The crustal thickness (depth to Moho) was found to be approximately 21 km.

Extensive seismic refraction work by the USGS to determine the structure in the Imperial Valley will be summarized by a report in press (Fuis *et al.*, 1980). That work is based on the interpretation of over 3000 seismograms from 1300 stations. The results agree with previous interpretations and add considerable detail. Fuis finds relatively high velocities (5.0 km/s) interpreted as indicating metamorphosed sediments in a "transition zone" on the last kilometer above basement with velocity appropriate for granitic rocks (5.65 km/s). The newest result is the detection of a high-velocity crust (7.2 km/s) at a depth of 10 to 16 km under two-thirds of the Imperial Valley. These results are consistent with the previous description of the Salton Trough, with the crust becoming thin due to tension, new oceanic crust forming at the leaky transforms, and conductive heat transfer and hydrothermal activity metamorphosing the overlying sediments.

Gravity. As of 1971, approximately 1400 gravity stations existed in the Imperial Valley and

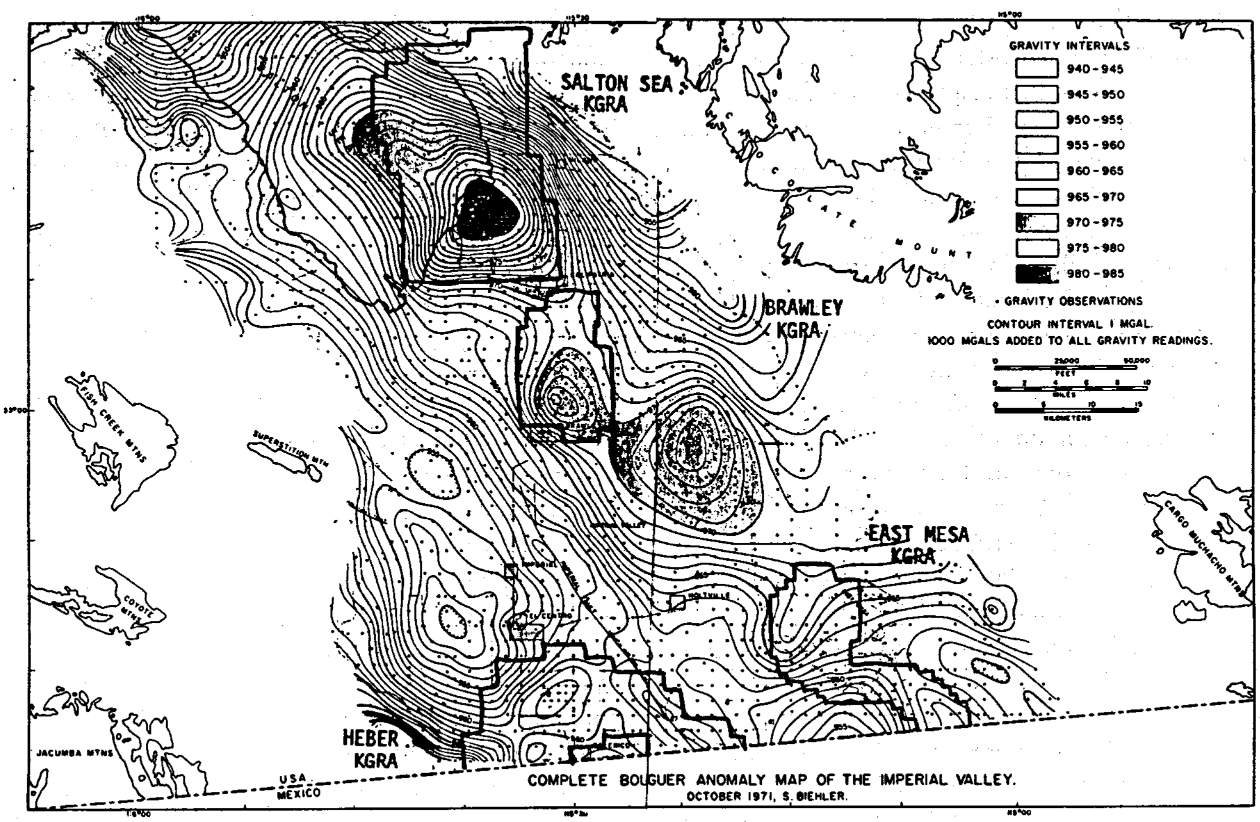

FIG. 6. Complete Bouguer anomaly map of the Imperial Valley, after Biehler (1971).

surrounding area. Figure 6 shows the station locations superimposed on the complete Bouguer anomaly map (Biehler, 1971). Kovach *et al.* (1962) conducted a combined regional-gravity–seismic-refraction survey in the Salton Trough. They report depths to basement ranging from less than 1 to over 4.5 km, with significant discrepancies between depths determined by the two methods. They attribute these discrepancies, which are not accounted for in the interpretations, to complexities in the lower stratigraphic section.

More generally, from San Gorgonio Pass south to the Salton Sea, the regional horizontal gradient is about 6 mgal/km. From there to the border, the gradient is 0.2 mgal/km. These regional gradients suggest that the crust thickens gradually from the border north to the Salton Sea and rapidly from there to San Gorgonio Pass. The total increase in crustal thickness is 6 to 8 km. The gravity maximum in the center of the Valley can be explained by a crust which is 8 km thinner than that near the edges, by a density excess of 0.1 g/cm^3 along the Trough axis, or by a combination of these effects. These latter results are consistent with both igneous intrusion and contemporaneous metamorphism of the sedimentary rocks.

Vertical Deformation. The history of vertical deformation in Southern California is dominated by continuing subsidence of the Salton Sea (Lofgren, 1978A) and by the well-known Southern California uplift (Castle *et al.*, 1976, 1977). Vertical leveling data have been collected since 1900. It has recently been shown that the observed Southern California uplift could be an artifact of systematic surveying errors in hilly terrain (Jackson and Lee, 1979), but the measured subsidence in the Salton Trough exceeds the error by an order of magnitude. The observed on-going deepening of the Salton Trough (as much as 3.5 cm/year) suggests that crustal thinning continues.

Geophysical Evidence of the Appropriateness of a Leaky-Transform Model for Crustal Thinning

Seismicity. The most striking aspect of the seismicity of the Imperial Valley is the dual occurrence of both large earthquakes (magnitudes near 7) and swarms of small-magnitude earthquakes (e.g., Ulrich, 1941; Hill *et al.*, 1975; Sylvester, 1979;

Johnson, 1979). The large earthquakes are an obvious consequence of the relative motion between the North American and Pacific plates.

Direct observations of surface deformations following large earthquakes illustrate the combination of regional shearing and local extension required by the leaky-transform model. Regional shearing has occurred because of several moderate-to-large earthquakes since 1940. Perhaps the most famous photographic example of strike–slip deformation is the orange grove whose straight rows were offset by the 1940 Imperial Valley earthquake (magnitude 6.4 to 7.1, Trifunac and Brune, 1970; Richter, 1958). Measured displacements across the fault were large—up to 5.8 m at the international border (Ulrich, 1941), with perhaps an additional 0.35 m of afterslip (Brune and Allen, 1967). Recent events have resulted in observed strike–slip deformation. On October 15, 1979, an earthquake (M_L about 6.4 to 6.6) occurred; its epicenter was on the Imperial Fault, and surface displacement was again 0.2 to 0.3 m. Comparable afterslips were detected in the U.S., but no slip was detected in Mexico (Kerry Sieh, personal communication, 1979). The smaller 1966 Imperial Fault event (M_L about 3.4) also had a detectable associated surface faulting (1.5 cm). In the Salton Trough, moderate-to-large earthquake-related tectonic deformations are common, the most recent being along the Cerro Prieto Fault in early 1980.

A detailed field examination of the Imperial and Brawley faults following the 1979 event detected a faulting pattern that closely mirrors both the shear and tensional aspects of the regional tectonics (Rundle and Krumhansl, unpublished field notes, 1979). From the border north about 30 km, the style of faulting was distinctly right-lateral strike–slip. In the next 10 km, scarp appeared, with about 5 to 10 cm of relative vertical displacement up on the west side. Similarly, the Brawley fault, slightly to the east, displayed 5 to 10 cm of relative vertical displacement up on the east side. The subsided area was located precisely on an enclosed structural depression known as Mesquite Dry Lake. Apparently, the normal faulting and crustal thinning due to a short spreading center connecting the two major strike–slip faults is causing observable subsidence at the surface.

Of course, other earthquakes have occurred in the area along other members of the San Andreas system. The most notable recent event was the Borrego Mountain earthquake of 1968, which occurred on the Coyote Creek Fault. The event (M_L about 6.8) caused sympathetic slip on the Banning–Mission Creek, Superstition Hills, and Imperial Faults (Allen *et al.*, 1972). Thus other faults in the San Andreas system have played a role in the regional tectonics of the Salton Trough since 1940, but their role has been relatively minor.

The second distinct style of strain release in the Salton Trough is by earthquake swarm activity. With the advent of an increasingly dense network of seismographs in the Imperial Valley since about 1972, interest in studying swarm events has increased dramatically (Johnson and Hadley, 1976; Sharp, 1976; Hill, 1977; Hartzell and Brune, 1977; Gilpin and Lee, 1978; Weaver and Hill, 1978; Johnson, 1979). The first well-studied swarm occurred on the Brawley Fault during January 1975 (Johnson and Hadley, 1976; Sharp, 1976). In fact, this swarm conclusively established the Brawley Fault as a structure distinct from the Imperial Fault. The method used to analyze the swarm data relies on relocating the epicenter of a "master event" accurately within the swarm and using relative travel times to locate the other members of the swarm. This technique can establish the relative locations of swarm events to within 100 to 300 m.

The most interesting aspect of the January 1975 swarm was the apparent activation of smaller fault structures transverse to the trend of the Brawley Fault (Johnson and Hadley, 1976; Johnson, 1979). One of the structures lies along a line striking northeast, intersecting the Brawley Fault in almost a direct line with Mesquite Dry Lake. The other structure strikes northeast from Brawley until it intersects the Brawley Fault some 3 km away. Focal mechanisms of events on the latter lineation display thrust, normal, and left-lateral mechanisms on nearly vertical, north-, and south-dipping planes (Johnson and Hadley, 1976). Other swarm activity (Johnson, 1979) was observed from 1973 to 1979 on transverse structures striking northeast across Mesquite Dry Lake, between the Brawley and Imperial Faults, along much of the Imperial Fault down to the international border, along the Calipatria Fault, and in the area of the Buttes (Gilpin and Lee, 1978).

The cause of earthquake swarms has not been established, but they are believed to be related to a hydraulic fracturing process (Johnson, 1979) or to a magmatic process (Hill, 1977). The latter view is supported by the relatively common occurrence of swarms worldwide in regions of magmatic activity (Sykes, 1970) and crustal spreading (Thatcher and Brune, 1971; Ward, 1971). However, their relatively shallow hypocenters in the Imperial Valley support the argument for a process related to hydrothermal activity.

Any proposed cause of the activity must explain the observed spatial and temporal correlations among the earthquakes. Much of the activity is displaced both vertically and horizontally with a velocity of 0.5 km/h. Other spatiotemporal trends are apparent in the data. Evidently, some kind of "communication" exists between swarms over distances of as much as 10 to 20 km (Johnson, 1979). Possible triggering mechanisms are associated fault creep, pore-fluid migration, and large-scale magma intrusion.

Although swarm activity and large-scale faulting are generally observed as being distinct processes, they may have some connection (H. Kanamori, personal communication, 1980). A pattern of gradually increasing seismicity followed by an abrupt cessation of activity was observed along the Imperial Fault before an earthquake. Whether this behavior argues more for a pore-fluid or magma-intrusion mechanism is unknown.

Horizontal Deformation. Periodically since 1900, various agencies have reobserved selected triangulation nets in the area (Thatcher, 1979). Additional data have, since about 1973, been collected with one of the modern generation of laser distance-measuring instruments (Savage *et al.*, 1978, 1979; Prescott *et al.*, 1979). As expected, the triangulation data, which have an accuracy of about 1 part in 10^6 over distances of about 100 km, show principally the effects of strain accumulation and release across the major right-lateral faults in the Trough (Thatcher, 1979). It is interesting to note that deformations measured over shorter time intervals differ significantly from the long-term trend. The early data from 1934 to 1941, an interval that spans the time of the 1940 earthquake, show little right-lateral slip along the fault trend. From 1941 to 1954, however, the strain field was better organized into a right-lateral field trending with the major faults. During this time, the right-lateral motion across the Imperial Valley was about 8 cm/year, decreasing to about 2 cm/year during 1954 to 1967. A significant amount of afterslip on the Imperial Valley faults may be responsible for the initial high strain rate.

Geophysical Evidence for Magma at Depth at Specific Locations in the Salton Trough

Several local thermal anomalies found in the Trough have features that suggest they have been local spreading centers long enough to be sites of substantial igneous intrusions. The intrusions probably consist of distributed swarms of small dikes. The sites include the Salton Sea Geothermal Field (also known as the Buttes), the Brawley, Mesa, and Heber areas, and Cerro Prieto, all of which share several geophysical features: positive gravity anomalies, high heat flow, microearthquakes, and substantial hydrothermal convection systems.

The geophysical data that pertain to the possibility of magma at each site are covered in this section. As discussed earlier, the gravity field shows a regional low across the Salton Trough because of the thick accumulation of low-density sediments. Local gravity maxima, superimposed on that regional low, were the first means of locating several geothermal fields in the Salton Trough. The association of a gravity high with geothermal fields is unusual, occurring only in regions like the Salton Trough, where the heat source and associated hydrothermal system occur within low-density sedimentary rocks or, if the heat source is basaltic, within a lower-density granitic basement. For individual sites discussed below, the maximum possible depth for the source of the gravity anomaly is estimated by calculating the depth to the center of a sphere of anomalous density.

Since near-surface heat flow is strongly influenced by fluid convection, it does not describe the location of the heat source well without extensive modeling; this is discussed in the "Evidence for Hydrothermal Systems" section of this paper. The presence of cooled intrusives in a region with nonmagnetic sedimentary rocks would produce a magnetic anomaly. The Salton Sea Geothermal Field (SSGF) is the only such locality having published data showing a magnetic anomaly. Black (1975) has discussed five possible reasons for the absence of a magnetic anomaly at the other sites: (1) there are no igneous intrusions, (2) the intrusions are too deep to have surface manifestations, (3) the intrusions are still above the Curie temperature, (4) geothermal fluids have leached magnetite from the intrusions, or (5) permanent and induced fields cancel each other.

Salton Sea Geothermal Field. The most detailed magnetic data on a known thermal anomaly in the Salton Trough are those of Griscom and Muffler (1971) at the SSGF (see Fig. 7). On the basis of an analysis of the large northwest–southeast ridge magnetic anomaly, they estimate an average depth of about 2.5 km to the dike–plutonic complex. They interpret the elliptical magnetic anomaly centered at the rhyolite buttes as being due to 10 to 20% dike material less than a kilometer from the surface. Evidence from measured tectonic activity in the area suggests that dike injection may be continuing in areas where previously emplaced magma has cooled below the Curie temperature. As an example, Fig. 8 shows the location of earthquake epicenters for October 1 through December 31, 1976 (Schnapp and Fuis, 1977). One group of epicenters correlates spatially with the offshore magnetic anomaly, the other with the onshore Buttes anomaly.

A particularly large gravity maximum is centered in the Salton Sea geothermal area, indicating the intrusion of higher-density igneous rocks amid the sediments (Biehler and Combs, 1972). The increased density accompanying the active metamorphism undoubtedly occurring in the area is another likely contributor. The broadness of the anomaly, together with reasonable assumptions about the density contrast of the metamorphosed sediments, leads to a maximum-depth estimate of 5 to 6 km for the center of the anomalous mass and roughly the same distance for the lateral extent.

Robinson *et al.* (1976) report that samples of subsurface igneous rock were recovered from at least four wells within the SSGF. Basaltic and silicic rock—interpreted to be thin dikes and sills—were present at depths ranging from 1 to 2 km. Seismic-refraction data in the area are consistent with a high velocity near the surface, presumably caused by a mixed basalt-sediment layer (Frith, 1978).

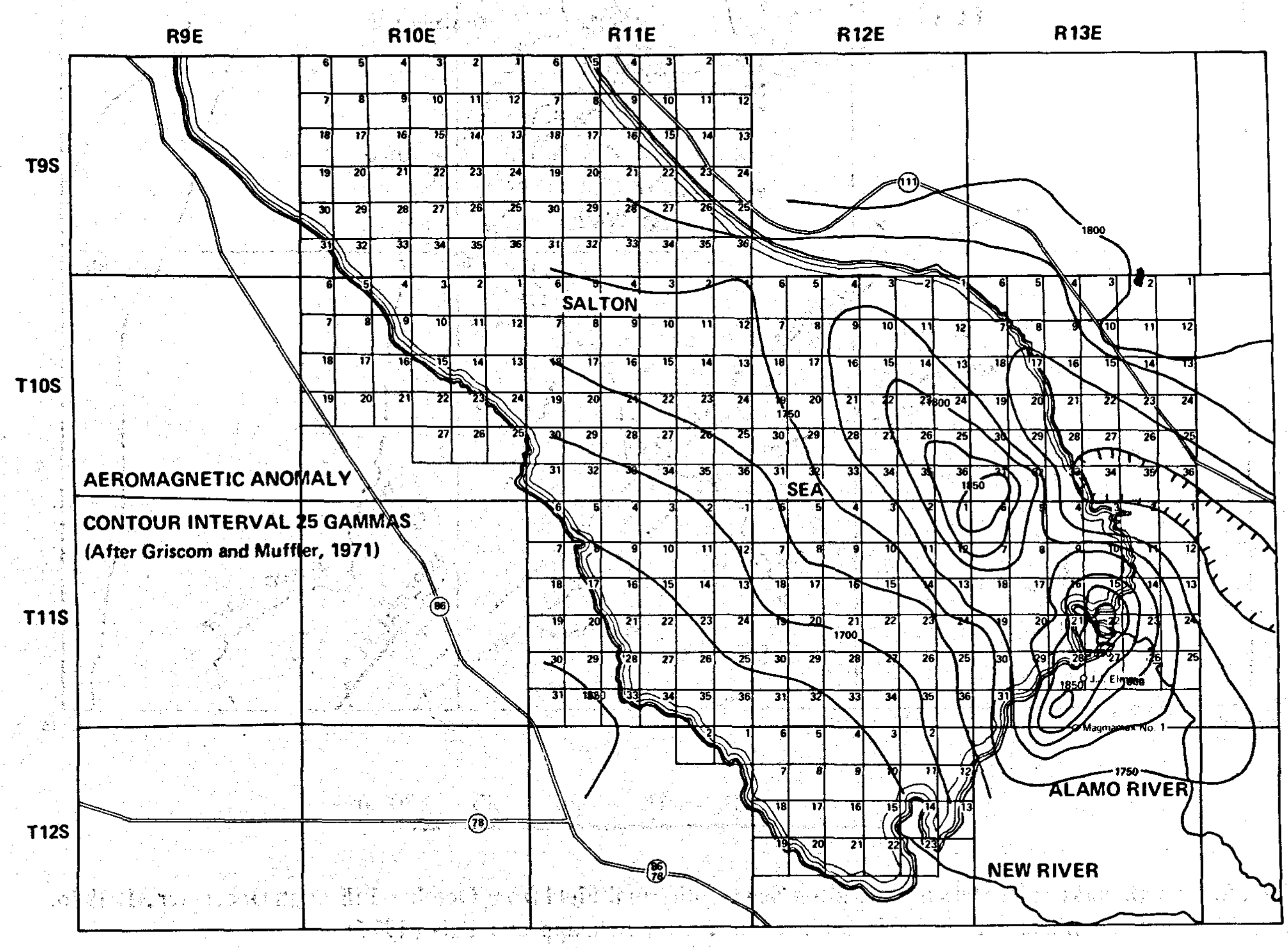

FIG. 7. Aeromagnetic anomaly map of the Salton Sea Geothermal Field, after Griscom and Muffler (1971).

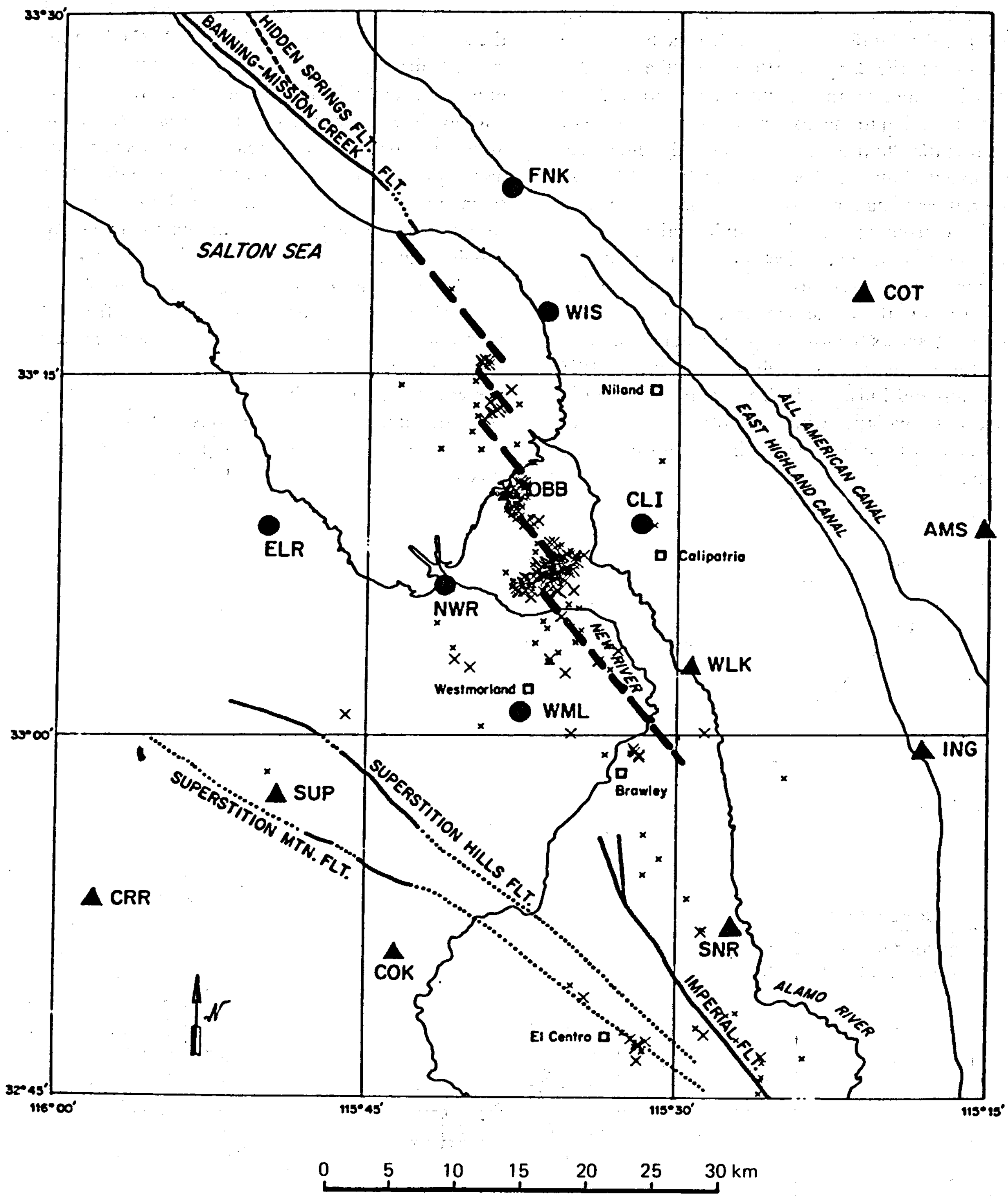

FIG. 8. Earthquake epicenters at the Salton Sea Geothermal Field from October 1 through December 31, 1976. Tentative fault locations are shown in broken lines, after Schnapp and Fuis (1977).

From these data, it is reasonable to conclude that a considerable volume of material has been intruded into the upper 5 km of the SSGF and that intrusion is still occurring. This area is the lowest spot in the Trough and is subsiding the most rapidly, suggesting very active crustal spreading. The location, by seismic methods, of specific targets of molten material at the Salton Sea is inhibited by the water that covers the area, making instrument siting prohibitively expensive.

Brawley Area. Kovach *et al.* (1962) have examined the large Bouguer gravity anomaly about 15 km east of Brawley. Using the width-at-half-amplitude method as an indicator of depth, they find that the source of the anomaly has a depth of no more than 6 to 7 km. They suggest that the most likely cause of the anomaly is density variation in the basement since their seismic-refraction data for this region indicate a depth to basement of 3.2 km.

Mesa Area. Biehler (1971) interpreted the gravity maximum in the Mesa Area as being caused by a spherical anomalous mass at approximately 4 km depth, with a 2100-m radius and an assumed density contrast of 0.15 g/cm^3. The depth to basement in this area is on the order of 6 to 7 km, implying that the mass is within the sediments and not a density variation in the basement. Subsequent studies at the Mesa anomaly have not found significant silica deposition (Combs, 1972). Because there is no magnetic anomaly, the gravity high in this region is assumed to be due to increased density resulting from thermal metamorphism by rising geothermal fluids (Black, 1975) with a little-known driving force.

CONCLUSIONS ABOUT MAGMA BODIES AT THE SALTON TROUGH

The Salton Trough has a combination of tensional and shear tectonic activity, which is well explained by the "leaky-transform" model. Long *en echelon* strike–slip faults are connected by short active spreading centers in a region of thinning crust and rapid sedimentation. Several possible spreading centers in the Trough have been identified and are good candidates for studies of continental-crust augmentation by distributed intrusions.

Several types of geophysical data are used to characterize the regional model. Gravity and seismic-refraction data define the structure: up to 6 km of sediments lying in a trough over a thin crust (about 20 km thick) with velocities appropriate to oceanic crust at depths of 10 to 16 km. High rates of regional subsidence indicate that crustal thinning is active, and seismicity and observations of horizontal deformation define the transform faults and spreading centers that are the locus of the thinning.

Geophysical methods have been used to study the spreading centers. Local gravity maxima indicate cementation of sediments at depths of less than 4 km at East Mesa, a dense basement 6 to 7 km deep at Brawley, and intrusions and/or metamorphism shallower than 6 km at the Salton Sea Geothermal Field. A positive magnetic anomaly at the SSGF is interpreted to show that cooled intrusions occupy 10–20% of the volume of rock as shallow as 1 km depth, and that a substantial dike complex exists below 2.5 km. Local heat flow is high at all the spreading centers, but accurate definition of the heat sources will not be known until the natural flow paths in the convecting hydrothermal systems are modeled.

Additional geophysical work in the Salton Trough could be directed toward identifying specific target bodies at each spreading center and determining whether they are still molten. Seismic-reflection and teleseismic delay studies might provide more information in areas where the Salton Sea does not interfere with instrument siting. Enough temperature data are available at several locations to model the convectively driven flow and to learn more about the characteristics of the heat sources.

The Salton Trough is attractive for deep drilling toward magma for several reasons. First, we are confident that intrusion is occurring there: the regional model described here is well accepted, igneous rocks have been detected in geothermal wells at three fields (Cerro Prieto, Heber, and the Salton Sea), and there is young volcanism. Second, conclusions drawn for one site in the Trough can be used to understand several other sites in the Trough. Third, ground "truth"—determined from one deep hole in the Trough—can be used to interpret extensive geophysical data throughout the region. Fourth, targets may be as shallow as 5 km, and some regions of intrusion are large. Finally, there is active industrial interest in several areas that can provide extensive data for characterizing sites and picking targets.

The Salton Trough has two drawbacks as a target for deep drilling toward magma. First, although specific targets of intrusion can be identified, we have no direct evidence about the extent or location of molten material. Second, the molten targets are probably small.

Two sites within the Trough probably provide the best potential targets for drilling toward magma: the Salton Sea Geothermal Field and Cerro Prieto (which has not been discussed in this review of American sites). The Salton Sea target may be within a reachable depth (probably shallower than 6 km), the area of intrusions is large, the spreading center is probably the longest, and the hydrothermal-system models give a reasonable target area for the heat source (see the "Evidence for Hydrothermal Systems at the Salton Trough" later in this report).

EVIDENCE FOR MAGMA AT LONG VALLEY

General Setting of the Long Valley Caldera

Many of Long Valley's characteristics indicate the presence of magma (see Fig. 9 for locations). A caldera, it has been the source of volcanic eruptions from 700,000 to as recently as 450 years ago. Hot springs indicate subsurface temperatures of 180°C, and several geophysical anomalies suggest the existence of a magma chamber.

Much of the work of the USGS at Long Valley was summarized in 13 articles in a special issue of the *Journal of Geophysical Research.* Most of the information in this review comes from those summary articles.

Geophysical data relating to the regional setting of Long Valley are not discussed here because geophysical data point to an area of less than 450 km^2 where a target might be located. However, Long Valley's regional setting makes the interpretation of geophysical data difficult. Lachenbruch *et al.* (1976B) state:

> "The Long Valley Caldera lies nearly astride a major fault system that separates the tectonically active Basin and Range Province to the east from the relatively stable Sierra Nevada tectonic province to the west. The Basin and Range Province is characterized by Cenozoic normal faulting and volcanism, high seismicity, thin crust, and low upper-mantle seismic velocity; none of

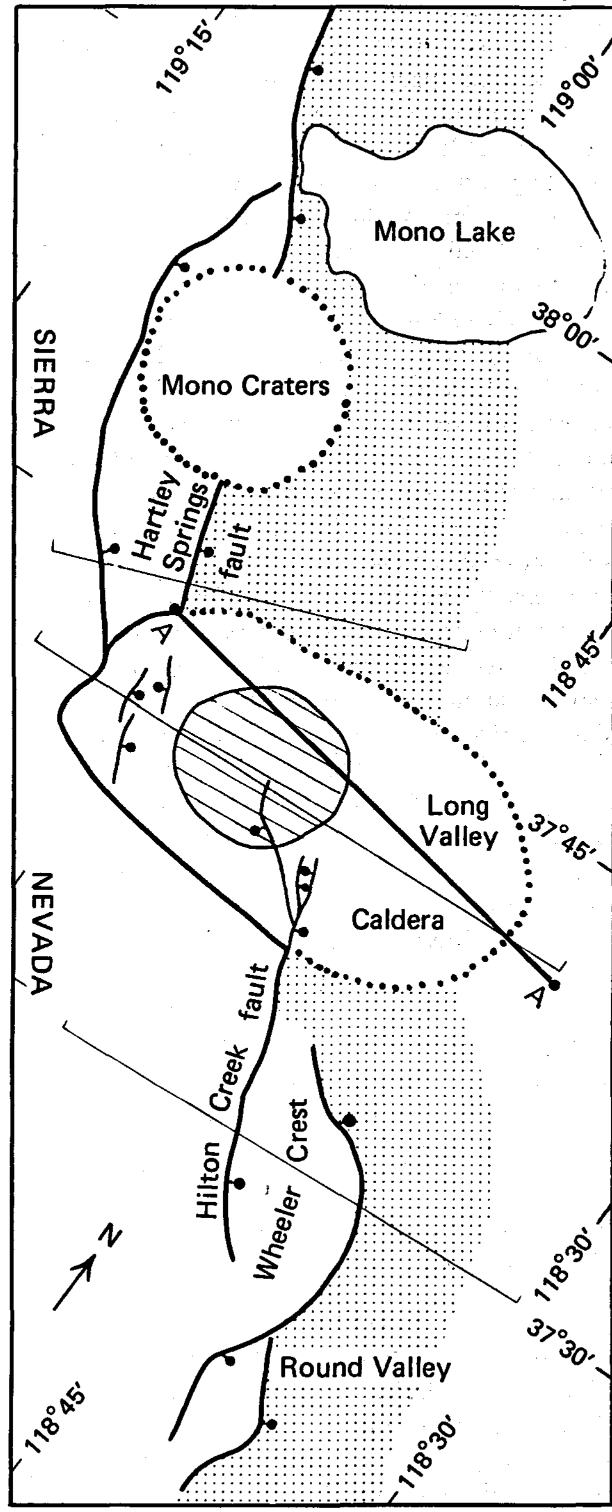

FIG. 9. Tectonic map of Long Valley–Mono Basin. Heavily dotted area is zone of "reverse drag" on Sierra Nevada front. Lined area is resurgent dome in Long Valley Caldera. Solid lines represent faults (ball on downthrown side). Profile A-A' shows location of seismic line in Fig. 10, after Bailey (1976).

these characteristics obtain in the Sierra Nevada. Consistent with these tectonic indicators, the contrast in regional heat flow across this boundary is one of the most abrupt known on the North American continent. While this location makes Long Valley an interesting subject for study, the transition zone compounds the problem of establishing the regional heat flow on which any local anomaly might be superimposed."

The last sentence also applies, to a greater or lesser extent, to other geophysical parameters like crustal velocities and attenuation, magnetic signature, and electrical resistivity. The saving characteristic of the Long Valley Caldera is that it provides a local and relatively small target; this target can be more readily examined within the transition zone than could an extensive, distributed intrusion system were it to occur at the boundary between tectonic provinces.

Evidence for a Magma Body at Depth

The direct geophysical evidence for a magma body existing below the Long Valley Caldera comes from active and passive seismic observations and from a gravity survey. Electrical methods have yielded little information below 2 km depth, apparently a limitation imposed by the conductive conditions occurring near the surface of much of the Caldera and by the practical length of dipole that could be used. With the published data base, seismic refraction and teleseismic data provide the most convincing case for at least a major, temperature-related discontinuity at depth. Gravity data and the microbearthquake survey provide confirmatory evidence.

Refraction Survey. Two seismic refraction lines—trending east-west and north-south—extend across and beyond the limits of Long Valley. Hill's (1976) summary interpretation of the east-west line is shown in Fig. 10. Most of the complexity occurs above 3 km, where Hill correlates the various interpreted velocity intervals with the known geology. Below 2 to 3 km depth, the velocity is interpreted as being a generally uniform 6.0 km/s.

The dashed line at 7 km depth in Fig. 10 indicates a possible interface with an underlying zone of lower velocity. Evidence for this zone stems from prominent secondary arrivals on the east-west line and from a similar but weaker set of arrivals on the north-south line. Hill indicates the velocity contrast is consistent with either a solid-melt interface or with a deep, low-velocity solid, perhaps a crystallized residual melt fraction. Other possible explanations are lateral reflections or refractions from a steeply dipping boundary or multiple refractions from the basement, including one reflection from the earth's surface.

Gravity. Kane *et al.* (1976) show that the major features of the gravity map of Long Valley can be satisfied by a low-density fill of density contrast 0.45 g/cm^3 that has an undulating lower surface with "deeps" projecting downward to 5.5 km and domes projecting upward to 1.0 km. The irregularities, they point out, could also be explained by variations in the density of the Caldera fill.

Disk models presented by Kane *et al.* show that the effect of a second, more-deeply-buried, low-density body representative of a magma chamber

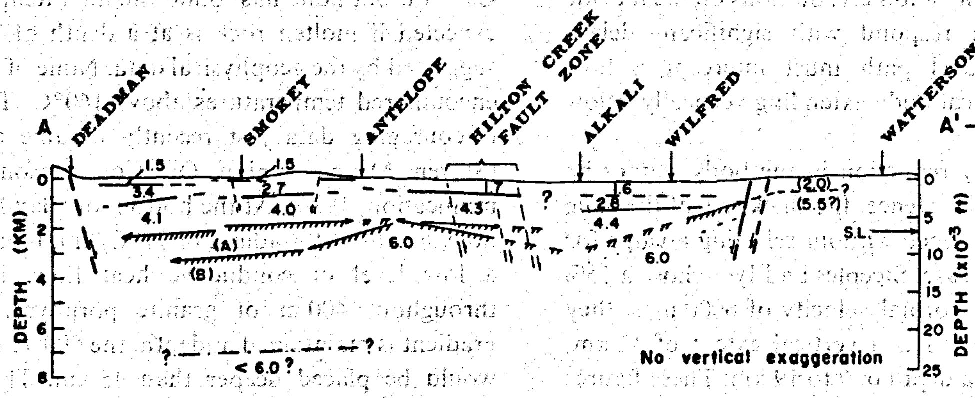

FIG. 10. Inferred P-wave velocity structure under an east-west profile across Long Valley, after Hill (1976). The low-velocity region on the west end of the cross section was detected by reflections and may represent molten material.

can only be manifested in the gradients in the gravity profiles outside the Caldera. The gravity gradient due to a shallow source (less than 3 km depth) cannot exceed 1 mgal/km in the interval 2 to 4 km beyond the boundary of the disk source. However, one of the measured gradients in this region is 1.5 mgal/km. The fact that this gradient is greater than what can be contributed by the shallow low-density source leads them to conclude that there is a deep, low-density source of undetermined thickness and density centered at a depth of 8 to 16 km.

Teleseismic Data. The best evidence for an extensive body of hot rock–partial melt at a depth greater than 7 km comes from Steeples' and Iyer's analysis (1976A) of the teleseismic data gathered in and around the Caldera. P-wave residuals relative to a station outside the Caldera were recorded on a net of portable seismograph stations operated for nine weeks in the Long Valley area. The probable uncertainty in the average relative residual of three or more events recorded from a particular azimuth was judged to be 0.1 s. Residuals recorded in the Caldera ranged as high as 0.57 s; those recorded outside the Caldera ranged as high as 0.18 s, but most were less than 0.10 s. Near-surface contributions to the P-wave delay were established at 0.2 s by three independent sources. This delay was subtracted from the computed residuals; Fig. 11 shows the results for the two predominant approach azimuths.

The largest residuals in Fig. 11 shift from one side of the Caldera to the other as the approach azimuth changes from northwest to southeast. Stations at the edge of the Caldera closest to the arriving teleseisms generally show residuals free of any significant velocity anomaly. Stations on the far side of the Caldera respond with significant delays because the travel path must intercept a low-velocity cylindrical body extending vertically below the Caldera.

The velocity reduction in the body cannot be measured directly. Hence, the thickness of the zone cannot be constrained without selecting a value for the velocity decrease. Steeples and Iyer chose a 15% decrease from a normal velocity of 6000 m/s; they estimate the body has a vertical extent of 12 km, and is placed at a depth of 7 to 19 km. These figures are obviously speculative and only serve to illustrate one possible interpretation. Moreover, at these temperatures and pressures, a 15% velocity decrease cannot be interpreted directly in terms of the degree of partial melt. It is sufficient to say that this unique data set nicely defines a low-P velocity zone at depths greater than 7 km below Long Valley and that the decrease is best attributed to an effect of high temperature.

Microearthquake Activity. Steeples and Pitt (1976) recorded microearthquake activity near Long Valley during 1973, using 20 different station locations during the recording period. To locate events, they used a horizontally layered crustal model established from a refraction profile. Their event locations, and those from a previous survey, are summarized in Fig. 12. Note that the microearthquake activity at depths of 5 to 20 km follows trends along the major valleys but stops abruptly at the southern edge of the Long Valley Caldera. Steeples and Pitt offer several explanations for the striking absence of activity within the Caldera. They suggest possible observational limitations, such as a short recording period coinciding with an aseismic period within the Caldera, or insufficient seismometer sensitivity. Neither of these possibilities accounts for the large number of events located elsewhere during two recording periods, and high temperatures or partial melting must be considered a possible cause. Interpretation of 1980 earthquakes in the Long Valley–Mono Lake area may modify the conclusions discussed here.

Temperature Data. Considerable shallow-hole heat-flow data have been collected in Long Valley, but the circulation of hydrothermal fluids makes it impossible to determine the deep thermal structure from those data (see "Evidence for Hydrothermal Systems at Long Valley"). Three moderately deep (1550–2100 m) holes have been drilled in the Caldera, but none has found the high temperatures expected if molten rock is at a depth of 7 km, as suggested by the geophysical data. None of the wells encountered temperatures above 160°C. The most discouraging data just recently became available (Steven Mione, Union Oil Co., personal communication, 1980). At the bottom of Clay Pit No. 1 hole, a constant gradient of 37°C/km, representing a low level of conductive heat flow, is found throughout 400 m of granite porphyry. If that gradient is maintained at depth, the 600°C isotherm would be placed deeper than 15 km. These data must be considered further before final conclusions are drawn, but they appear to weigh heavily against the presence of magma within drilling range.

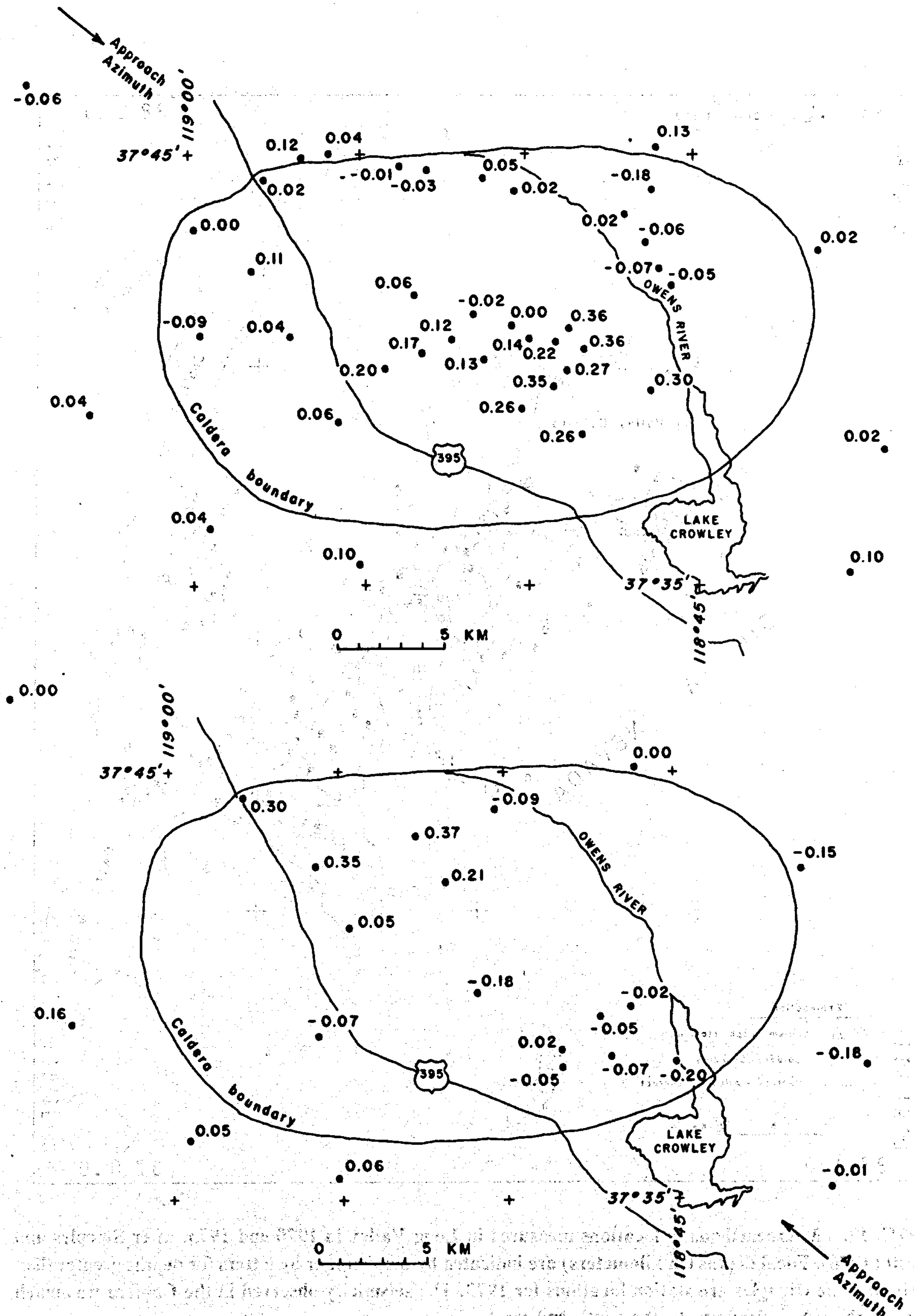

FIG. 11. Residual teleseismic P-wave delays measured in Long Valley, after Steeples and Iyer (1976). All relative arrival times have been corrected for near-surface effects; the differences should represent structures below about 5 km. The different patterns of delay for different approach azimuths give information about the depth of the low-velocity body that causes the delays.

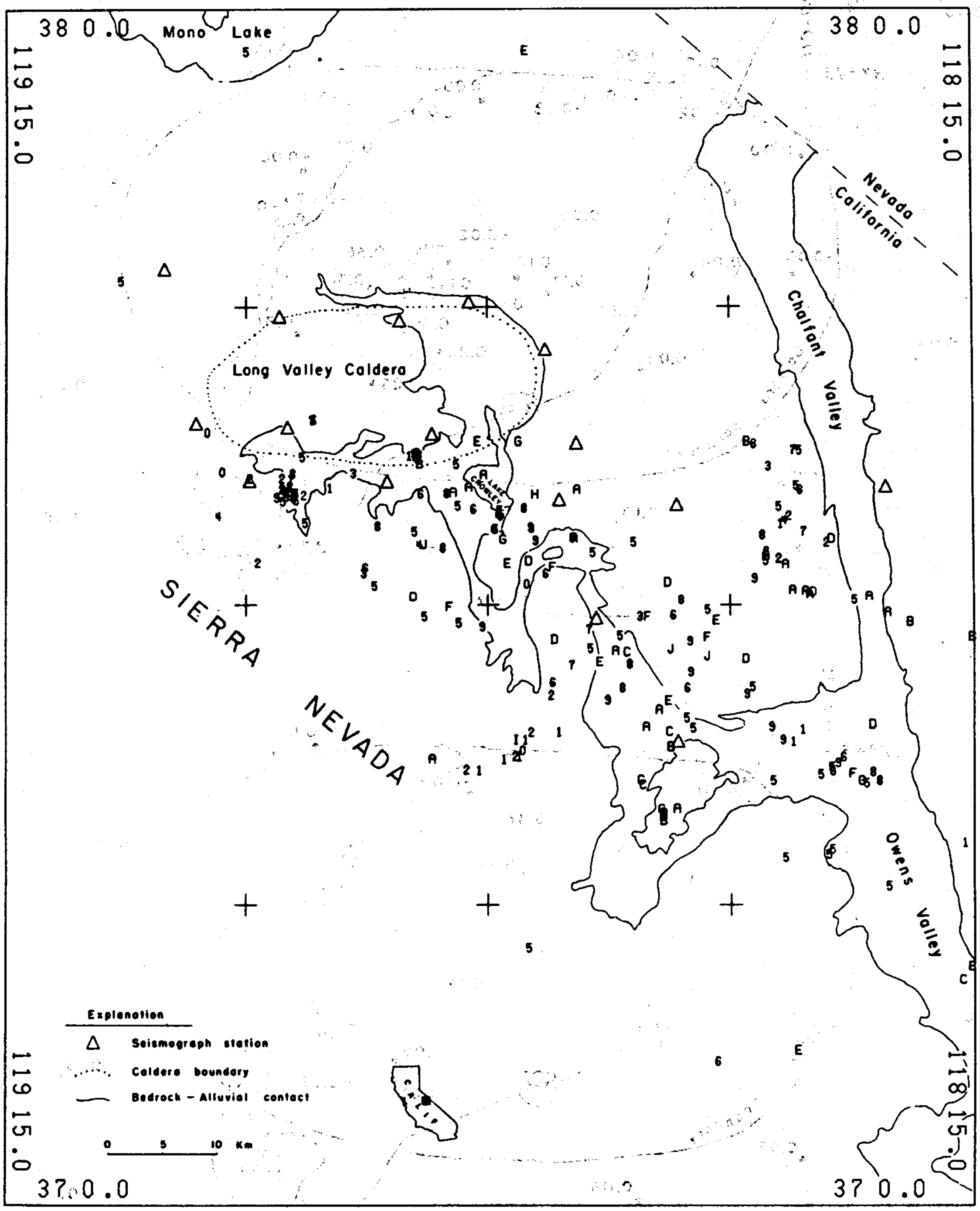

FIG. 12. Microearthquake locations measured in Long Valley in 1970 and 1973, after Steeples and Pitt (1976). Focal depths (in kilometers) are indicated by numbers, or by letters for depths greater than 10 km. The triangles are station locations for 1973. The seismicity observed in the Caldera was much lower than the seismicity to the south and west.

CONCLUSIONS ABOUT A MAGMA BODY AT LONG VALLEY

Long Valley may provide a location for drilling toward a silicic magma chamber associated with recent, extensive volcanism. Several geophysical methods support the interpretation that a zone of magma or partial melt as large as 10 km diam could lie only 7 km below the surface. Seismic reflections indicate a low-velocity region 5 to 8 km wide at 7 km depth. A gravity low is interpreted as indicating a low-density body of uncertain size at 8 to 16 km depth. One interpretation of teleseismic data indicates a low-velocity body 8 to 10 km wide and 7 to 12 km deep. Confidence in these interpretations will be strongly influenced by evaluation of data recently released from two drill holes that did not show the steep conductive gradients expected above such a body and thereby indicate that molten rock is deeper than 15 km.

The major question at Long Valley is how to resolve the apparent contradictions between interpretations based on geophysics and on the data from three deep holes. An understanding of the specific paths of fluid flow and the geometry of convective heat transport is essential to reconcile those interpretations. The near-surface heat-flow regime is well studied. Perhaps a reevaluation of models based on those data, motivated by the low gradients in the deep holes, will lead to a better understanding. Otherwise, the only method that will provide new information in that complicated regime is more deep drilling.

Long Valley is an attractive target for drilling toward magma for several reasons. First, if there is a target, it probably lies below the Caldera, a relatively small area. Second, the choice of a drilling location could be based on the considerable effort that has been spent evaluating the shallow subsurface. Third, a deep hole can provide ground truth to resolve conflicting interpretations from geophysical and thermal data. Finally, a better understanding of the magmatic processes at a caldera could be applied to Cenozoic volcanism throughout the western United States.

Long Valley also has several drawbacks as a site for deep-hole drilling toward magma. Geophysical data do not define a target, and suggest that magma could lie at a considerable depth. The system may be late in its development, and it might be easier and as useful to study analogous systems exposed by erosion.

EVIDENCE FOR MAGMA AT ROOSEVELT HOT SPRINGS

General Setting of Roosevelt Hot Springs

Roosevelt Hot Springs is a fault-controlled hydrothermal system; it is located in Utah, on the eastern boundary of the Basin and Range Province (Fig. 13). The geophysical data collected there have recently been summarized by Ward *et al.* (1978). The Hot Springs lie on the eastern edge of Milford Valley—a graben bounded by the Mineral Mountains, a fault block of Tertiary granite intruded and overlain by young rhyolitic volcanism (0.5 to 0.8 million years old). Faults bounding the Milford Valley control the flow of hydrothermal water upward from a heat source presumed to lie under the Mineral Mountains. Geophysical measurements have been used to delineate the regional setting, to locate faults, and to detect the hydrothermal system. Although many techniques have been used to locate the ultimate heat source, little has been determined about it.

Evidence for Magma at Depth

Although there is no direct evidence of magma beneath the Mineral Mountains, such a body is inferred as a possible heat source for the Hot Springs system. The only quantitative constraints on the location, depth, and size of this body come from an analysis of the system's total heat flow. Ward *et al.* (1978) use a first-order, steady-state conduction model to put the magmatic source below 7 km depth if it is as large as the Mineral Mountain pluton (approximately 10 by 30 km).

Other geophysical data reported by Ward *et al.* (1978) give only tenuous indications of a zone of magma. Positive P-wave delays and "qualitative estimates" of S-wave attenuation were observed across the Mineral Mountains, but the station spacings were too sparse to provide interpretable data. Seismicity studies for two field seasons found considerable activity in the Cove Fort area, about 15 km from Roosevelt Hot Springs, diffuse moderate

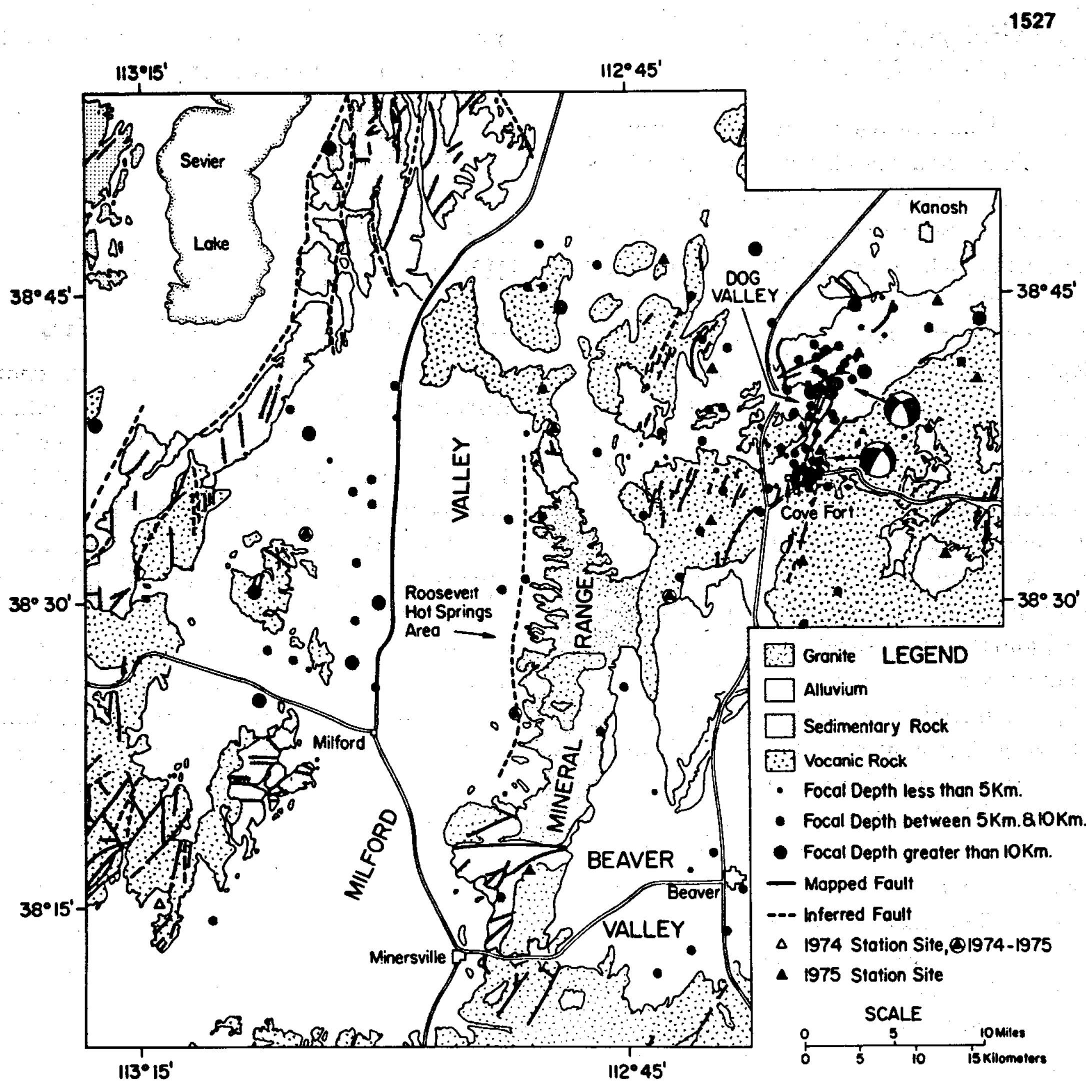

FIG. 13. Location map for Roosevelt Hot Springs, after Ward *et al.* (1978). Epicenters from seismicity studies and generalized geology are also shown.

activity along the west side of Milford Valley, and "minor" activity adjacent to the Mineral Mountains. Ward *et al.* consider the Cove Fort events as potentially related to a magma body, but do not discuss the source of the activity near Roosevelt Hot Springs. Gravity and magnetic surveys produce data that can be explained by geologic structures without invoking a magma body. Magnetotelluric signals might be sensitive to the hypothesized magma body, but near-surface conducting anomalies make interpretation of these data difficult.

Brief papers summarize the conclusion that additional seismic work is needed near Roosevelt Hot Springs. Young *et al.* (1979) report that seismic-attenuation values detected at Roosevelt are intermediate to those at Yellowstone and Coso Hot Springs, but they do not give quantitative results. Robinson and Iyer (1979) report that the preliminary interpretation of travel-time data confirm the presence of a 10% reduction in velocity centered at 10 to 25 km depth beneath the Mineral Mountains. In addition, they indicate that a small, shallow, "anomalous" region 5 km southeast of Roosevelt Hot Springs distorts the seismic waves. They speculate that the large, low-velocity area is caused by the presence of basaltic magma and that the distorted wave forms are caused by a zone of rhyoline magma in the upper 5 km of the crust.

CONCLUSIONS ABOUT MAGMA BODIES AT ROOSEVELT HOT SPRINGS

There is little direct geophysical evidence for a magma body at Roosevelt Hot Springs. Neither is such a body precluded by the published data. Attractive targets may lie beneath the Mineral Mountains, but additional work is required before the depth, size, or nature of those proposed targets can be discussed.

EVIDENCE FOR MAGMA AT THE GEYSERS-CLEAR LAKE AREA

General Setting of The Geysers–Clear Lake Area

The Geysers–Clear Lake area is situated in the Coast Ranges of northern California (see Fig. 14). The Clear Lake volcanic series is a result of the continuing interaction of the North American plate with the remnants of the subduction zone, which has been replaced by the strike–slip motion of the San Andreas fault system. The volcanics lie above the intensely deformed Franciscan melange and the Great Valley sequence—two distinctly different marine deposits brought together by the subduction in a series of slice-like thrust sheets, and extensive serpentinites lie mixed between the Franciscan and Great Valley rocks. Because this geologic setting, common to the Coast Range Province of much of California, provides little insight into the occurrence or extent of magma at a particular site, the geophysical data that help one understand that setting are not discussed here. However, the nature of the setting does have a strong influence on both the geophysical data and its interpretation.

The coincidence of active hydrothermal alteration, steam at shallow depths, and Quaternary volcanic rocks led many early geologists to speculate on the presence of an underlying magma body as the heat source for The Geysers geothermal system. Rodger Chapman (1966) first gave this idea geophysical credence by publishing a gravity map showing a major negative Bouguer anomaly centered near Mount Hannah and by suggesting that the anomaly could be due to low-density molten rock. Subsequent research includes analyses of more-refined gravity data, aeromagnetic and ground-magnetic surveys, heat-flow measurements, microearthquake surveys, studies of teleseismic P-wave delays and local seismic reflection and refraction, various methods of determining electrical resistivities, and miscellaneous investigations by remote sensing. Each of these techniques contributes to our ideas about the heat source and its relation to the structure of the region.

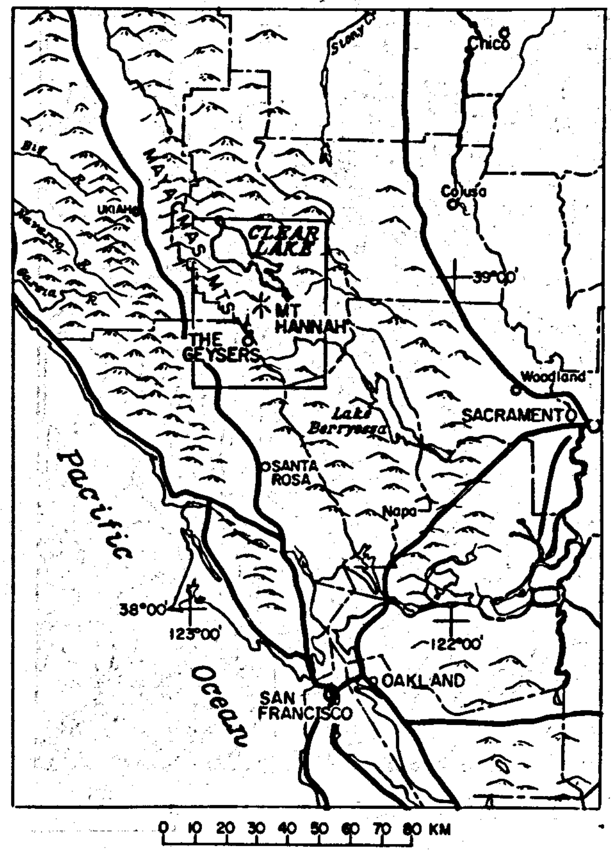

FIG. 14. Map showing the location of The Geysers–Clear Lake area, after Isherwood, (1975).

Evidence for Magma at Depth

Gravity. A recent map of residual gravity (Fig. 15) shows the salient features of the gravity field discussed by both Chapman (1975) and Isherwood (1976). The major feature of the field is a two-part, nearly circular gravity low. The deepest part—the Mount Hannah low—lies northeast of the producing steam field and appears to coincide with the volcanic edifice of Mount Hannah. The other part—the production low—coincides with the producing steam field. Although part of the overall gravity low must be the result of shallow-density contrasts relating to the thickness of near-surface volcanic rocks and the reservoir structure, the major

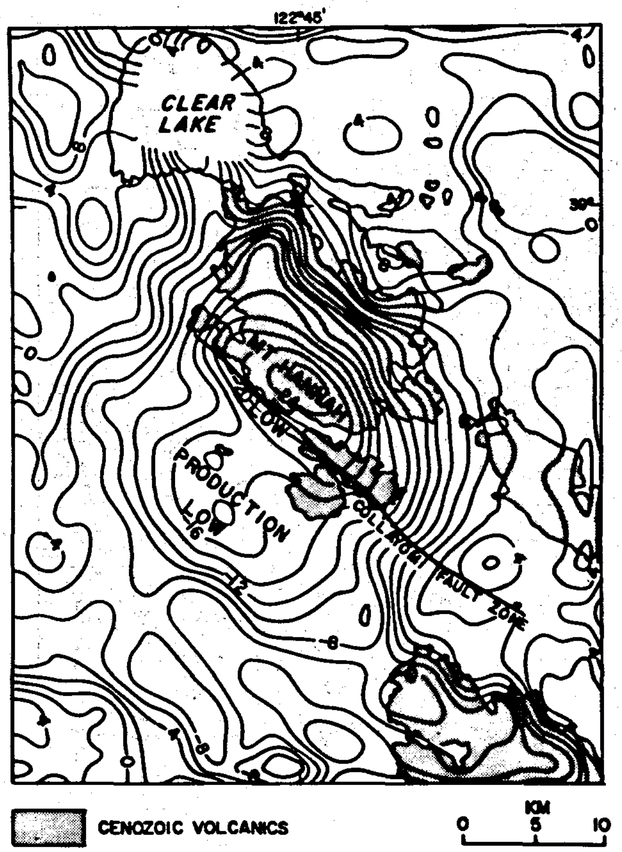

FIG. 15. Residual Bouguer anomaly map of The Geysers–Clear Lake area, after Isherwood (1975). The assumed density of the near surface was 2.67 g/cm^3.

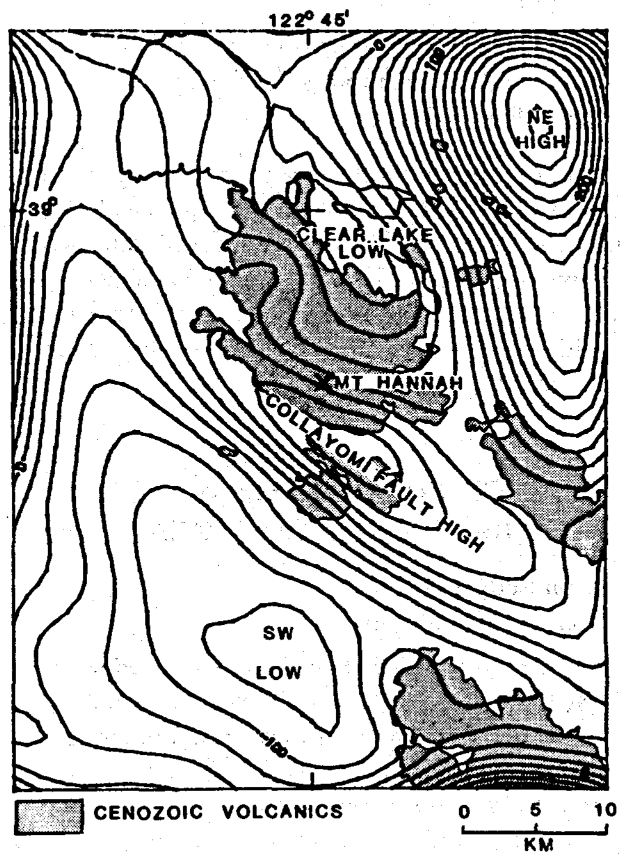

FIG. 16. Magnetic anomaly map for The Geysers–Clear Lake area, after Isherwood (1975). Data have been continued upward to 3 km. The contour interval is 20 gamma.

low is interpreted to reflect the low-density upper silicic differentiate of a magma chamber centered between about 6 and 14 km depth (Isherwood, 1976). The negative gravity anomaly extends farther to the southwest than could be explained merely by the low densities of the Clear Lake volcanics and the Great Valley sequence, located primarily northeast of the Collayomi fault zone. The spatial wavelength of the nearly circular gravity anomaly corresponds to an equivalent sphere centered at about 13.5 km depth. Rocks within 2 or 3 km of the surface have been reached by numerous steam wells without finding any low-density mass. Thus, to account for the large mass deficiency indicated by the gravity anomaly, the density contrast between the low-density body and the surrounding rocks must be large. The only likely material that can give a large enough contrast of density compared with that of typical rocks of the Franciscan assemblage is partially molten silicic rock, or magma.

Magnetic. Magnetic anomalies are difficult to interpret in regions with highly magnetic surface rocks, such as volcanic flows and serpentinite. The most obvious topographic and other near-surface effects can be subdued, however, using filtering provided by analytic upward continuation of the digitized magnetic field. No magnetic anomaly corresponds to the major gravity low. The 3-km continuation illustrated in Fig. 16 shows only a major magnetic high along the Collayomi fault zone, indicating that the serpentinite seen at the surface along the zone may extend to a depth of several kilometers. Furthermore, no magnetic sources can be identified below a depth of about 6.5 km. The low-density body is nonmagnetic, which is consistent with it being above both the Curie temperature and its melting point.

Seismic. Seismic studies provide data that constrain the limits of models of The Geysers geothermal system. The teleseismic P-wave delay technique provides an independent test of the magma-chamber hypothesis. Initial measurements (Steeples and Iyer, 1976) suggested that seismic body waves did, indeed, show delayed arrivals in a region roughly corresponding to the gravity low. Recent, more-detailed studies (Iyer *et al.*, 1980) show the location of the observed delays more accurately. A refraction study by Majer and McEvilly (1979) shows that this low-velocity material is not in the upper three kilometers. Consequently, Iyer and others attribute the delays to a magma chamber that extends from 4 to below 30 km in depth and is centered between Mount Hannah and The Geysers.

Theory also predicts attenuation of seismic body waves by rock materials with any degree of partial melt. Recording earthquake signals at an array of stations throughout a region can detect seismic attenuation on a relative basis. Ward and Young (1980) mapped a region of high attenuation coinciding roughly with the gravity low at The Geysers, which they too attribute to underlying magma.

The occurrence of microearthquakes near The Geysers has been known for some time (Lange and Westphal, 1969; Hamilton and Muffler, 1972). Continuous monitoring since 1975 (Marks *et al.*, 1978) has recorded substantial shallow seismicity, much of it induced, in the geothermal region. The absence of earthquake hypocenters deeper than about 5 km in the region of the gravity low (Marks *et al.*, 1980) is consistent with the hypothesis of elevated temperature associated with a magma body at depth.

Denlinger (1979) has shown that recent microearthquakes are confined to the portion of the reservoir undergoing fluid depletion and strain. This pattern of seismicity (and geodetic measurements described below) suggests that some seismic activity is caused by the release of regional tectonic stress triggered by the production of steam. Most fault-plane solutions for local seismic events suggest extensional motion along short faults trending more northerly than the geologically well-defined Maacama, Mercuryville, or Collayomi fault zones.

Resistivity. Electrical-resistivity measurements by Stanley, Jackson, and Hearn (1973) provide ambiguous support for the model of a magma body beneath Mount Hannah. A composite apparent-resistivity map, based on data from an assortment of techniques (Fig. 17), shows a distinct, low, apparent-resistivity anomaly coinciding with the major gravity low, which could possibly support the

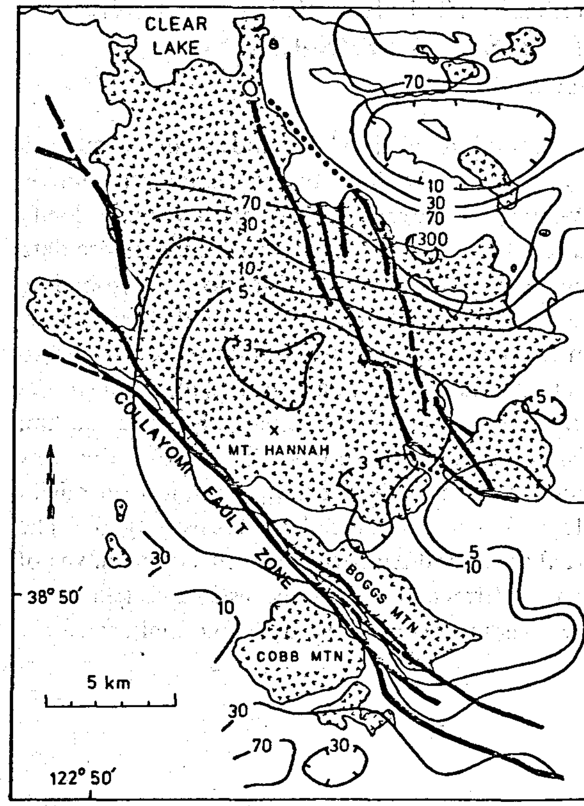

FIG. 17. Electrical resistivity map of The Geysers–Clear Lake area, after Stanley *et al.* (1973). Contours are in ohm-meters.

interpretation of magma at depth. Unfortunately, ambiguities from possible lateral inhomogeneity, as well as uncertainties about the depth being sampled and the correlation with rock type and condition, make the quantitative interpretation of these data impossible.

Heat Flow. Heat-flow measurements are difficult to extrapolate through regions of known hydrothermal convection. Only one model for the heat source has been published. Jamieson (1976) attempted to model the heat source for a region of the southwest flank of the gravity anomaly. He interpreted the heat flow as being consistent with hot intrusive rocks (over 700°C) at a depth of about 8 km over a wide area, with regional heat flow from this heat source to the surface being primarily by conduction. Superimposed on the regional heat flux, Jamieson found high-temperature gradients in the producing area. These gradients reflect the convective transfer of heat from depth by local steam and hot-water systems along high-angle fractures.

CONCLUSIONS ABOUT MAGMA AT THE GEYSERS–CLEAR LAKE AREA

Several geophysical methods point to an area of approximately 200 km^2 of highly attenuative, low-strength material of low density and velocity that could be as shallow as 7 to 10 km. Those data include a gravity low with no corresponding magnetic anomaly, interpreted as indicating a low-density (partially molten) body in the crust at 8 to 14 km depth; P-wave delays and attenuation studies indicating a low-velocity zone in the upper 30 km; seismic-refraction studies that show the low velocities must be deeper than 4 km; low seismicity below 5 km; and an unexplained resistivity low. The actual depth and temperature of this target have not been confirmed by heat-flow estimates from drill holes passing beneath the hydrothermal system.

The Geysers–Clear Lake area is an attractive site for deep-drilling toward magma for several reasons. First, the target area is well-defined and all evidence to date supports the existence of a magma chamber at reachable depths. Second, if the target model is correct, this area will give us an opportunity to study the environment of a moderate-sized silicic batholith while it is still developing. Third, the strong industry interest in the area provides the means for detailed characterization of the shallow zones and the potential for piggyback studies. Finally, a deep drill hole would provide ground truth to add certainty to the interpretation of several methods of deep-sounding geophysics.

The Geysers–Clear Lake area also has two drawbacks as a drilling site. First, the nature of the target is not confirmed. Second, the geological setting is unusual and complex, possibly making interpretation of results difficult and very site specific.

EVIDENCE FOR HYDROTHERMAL SYSTEMS

Hydrothermal systems are usually discovered because of such surface manifestations as hot springs, seeps, or altered surface rock. Geophysical measurements are used in three ways to study these systems. First, geophysical exploration techniques are used directly to measure the thermal field, the density differences due to alternation or deposition, and the low resistivity in saline hydrothermal systems or high resistivity in steam fields. These measurements allow detection of "hidden" systems and help determine their extent and location. Second, the flow patterns in such systems are influenced by the driving force (i.e. the heat source) and by permeability patterns caused by the geologic structure and fault patterns. Consequently, any geophysical exploration technique that senses the geologic structure is useful for understanding hydrothermal systems. Third, direct temperature measurements in deep wells have proved invaluable in understanding hydrothermal systems. For example, downward extrapolation of heat-flow measurements into regions where fluid is moving is unreliable, and deep-well measurements provide the most certain temperature information.

The next section of this review discusses the geophysical data relating to hydrothermal systems at the five potential sites.

EVIDENCE FOR HYDROTHERMAL SYSTEMS AT THE RIO GRANDE RIFT

Key evidence for the existence of hydrothermal systems in the Rio Grande Rift zone comes from surface geological features, spring occurrences, and petrographic–geochemical data rather than from geophysics. The observation of local, extremely high heat-flow values indicates directly that hydrothermal systems exist. Reiter *et al.* (1978) describe the unusual variability of heat-flow data within the Rift, where five localities with heat flows of 4.5 to nearly 10 HFU were found. The large variability of heat-flow values implies a dynamic geothermal environment at shallow depths; hydrothermal systems contribute to this variability throughout the Rift. Unfortunately, heat-flow measurements are spaced too far apart to evaluate any particular hydrothermal system at this time.

One system has been drilled and verified to have high temperatures. The surface heat flow is quite high near the Valles Caldera (4.5 to 6.0 HFU), and deep wells within the Caldera have encountered temperatures as high as 330°C (Brook *et al.*, 1979). Dondanville (1978) describes the results of geophysical surveys near the geothermal field. The Caldera itself is marked by a gravity low of 25 mgal, which is interpreted as indicating the subsidence following major eruptions. Other geophysical data apparently indicate structural features within the Caldera, and Dondanville gives no information to indicate that geophysics were useful for determining the volume of fluid in the hydrothermal system.

CONCLUSIONS ABOUT HYDROTHERMAL SYSTEMS AT THE RIO GRANDE RIFT

The most extensive, high-temperature hydrothermal system in the Rio Grande Rift is situated in the Valles Caldera. It has been drilled and found to have temperatures as high as 330°C. This system is an attractive target for study because it is hot and presumably large and because industrial interest in the system could result in a wealth of available data. Unfortunately, the extent of the system and the details of its "plumbing" are not known, and it is difficult to recommend geophysical methods to resolve those uncertainties.

EVIDENCE FOR HYDROTHERMAL SYSTEMS AT THE SALTON TROUGH

Several isolated hydrothermal systems in the Salton Trough have been located by geophysical methods and verified by drilling. The most significant of these are the Salton Sea, Brawley, Heber, East Mesa, and Cerro Prieto systems, each of which is large enough and contains enough hot fluid to have an industrial interest in power production. They cover areas from 18 to 54 km^2 and have estimated maximum temperatures ranging from 160 to 360°C (Renner *et al.*, 1975).

Use of Geophysical Data in the Salton Trough

The use of geophysical data to study hydrothermal systems in the Salton Trough is discussed below. Because the systems are of a similar nature, geophysical data are used the same way at each site. Consequently, the discussion is organized by the exploration method.

Gravity. Most of the Salton Trough hydrothermal systems have only minor surface manifestations; several have none. The discovery of several isolated gravity highs (Biehler, 1971), later determined to pinpoint the location of hydrothermal systems, led to further geophysical exploration (see Fig. 6). Because these gravity highs may be caused by intrusive activity or by hydrothermal alteration and deposition, the gravity data were described earlier in this report (see "Geophysical Evidence for Magma at Depth at Specific Locations in the Salton Trough").

Heat Flow. A considerable amount of shallow temperature and heat-flow data have been taken in the Salton Trough (Rex, 1966; Rex *et al.*, 1971; Lee and Cohen, 1979). Combs' results illustrate the general features (Fig. 18). Areas with more than five times normal conductive heat flow were found to coincide with the gravity highs. Isolated gradient observations near rare thermal manifestations are as much as a factor of 50 above background (Lee, 1977).

Temperature measurements in deep holes at all sites in the Salton Trough show a similar thermal structure. A thermal cap with a constant gradient (and inferred low permeability) lies above a more nearly isothermal zone, where fluid circulation is the dominant heat-transfer mechanism.

Electrical–Telluric. Electrical-resistivity measurements also show the hydrothermal systems. A total of 58 resistivity soundings—distributed throughout the Imperial Valley—were carried out between 1968 and 1971 (Meidav and Furgerson, 1972). From the southeast corner of the Valley, resistivities were found to decrease gradually northwestward. This decrease is attributed to the increase in the salinity of the groundwater with increasing distance from the Colorado River. Local resistivity lows associated with enhanced salinity and temperature effects due to thermal anomalies are superimposed on the regional trend. Maas and Humphreys (1977) report a strong correlation between telluric electric-field anomaly and thermal anomaly at both the SSGF and Mesa anomalies.

Interpretation. Models are often used to interpret these geophysical data. Most of the Salton Trough heat- and mass-transfer models incorporate the same components—fluid heating at depth, percolation of the fluid up a high-permeability fault or fault zone, lateral spreading of the hot water beneath a vertically impermeable cap of rock or interbedded shales, and subsequent conduction-dominated heat transfer from the hot fluid through the cap to the surface (Bird and Elders, 1975; Black, 1975; Kasameyer *et al.*, 1980; Kassoy and Goyal, 1979; Lau, 1980; Noble *et al.*, 1977; Riney *et al.*, 1979; Swanberg, 1975; Tansev and Wasserman, 1978). All existing thermal models are vague about the depth and temperature at the heat source; this information is usually inferred from other geophysical or geochemical indicators.

Data from Specific Hydrothermal Systems

Of all the Salton Trough hydrothermal systems in the United States, only the Salton Sea and East Mesa systems have been well characterized in published analyses of geophysical data. The Salton Sea field, in particular, is a very well characterized hydrothermal system. Temperature data up to 360°C from at least 11 deep wells have been published and studied (Helgeson, 1968; Randall, 1974; Palmer, 1975; Younker, Kasameyer, and Tewhey, 1980). On the basis of these data, all authors agree that the hydrothermal system is large—the inferred area is over 50 km^2 (Younker and Kasameyer, 1978; Brook *et al.*, 1979)—and actively convecting (Helgeson, 1968). The spatial distribution of temperature is determined well enough so that the development of the convecting system can be modeled. Models of the system must explain two observed features: that the heat flow in the cap is very uniform over 20 km^2 and that the transition to normal heat flow occurs over a very short distance (Younker, Kasameyer, and Tewhey, 1980).

Resistivity data are used to map the fluid from the hydrothermal system. A high-resolution, deep electrical-sounding survey has been done at the SSGF by Geonomics, Incorporated (Meidav *et al.*, 1976). This survey revealed six significant northwest–southeast-trending faults and indicated

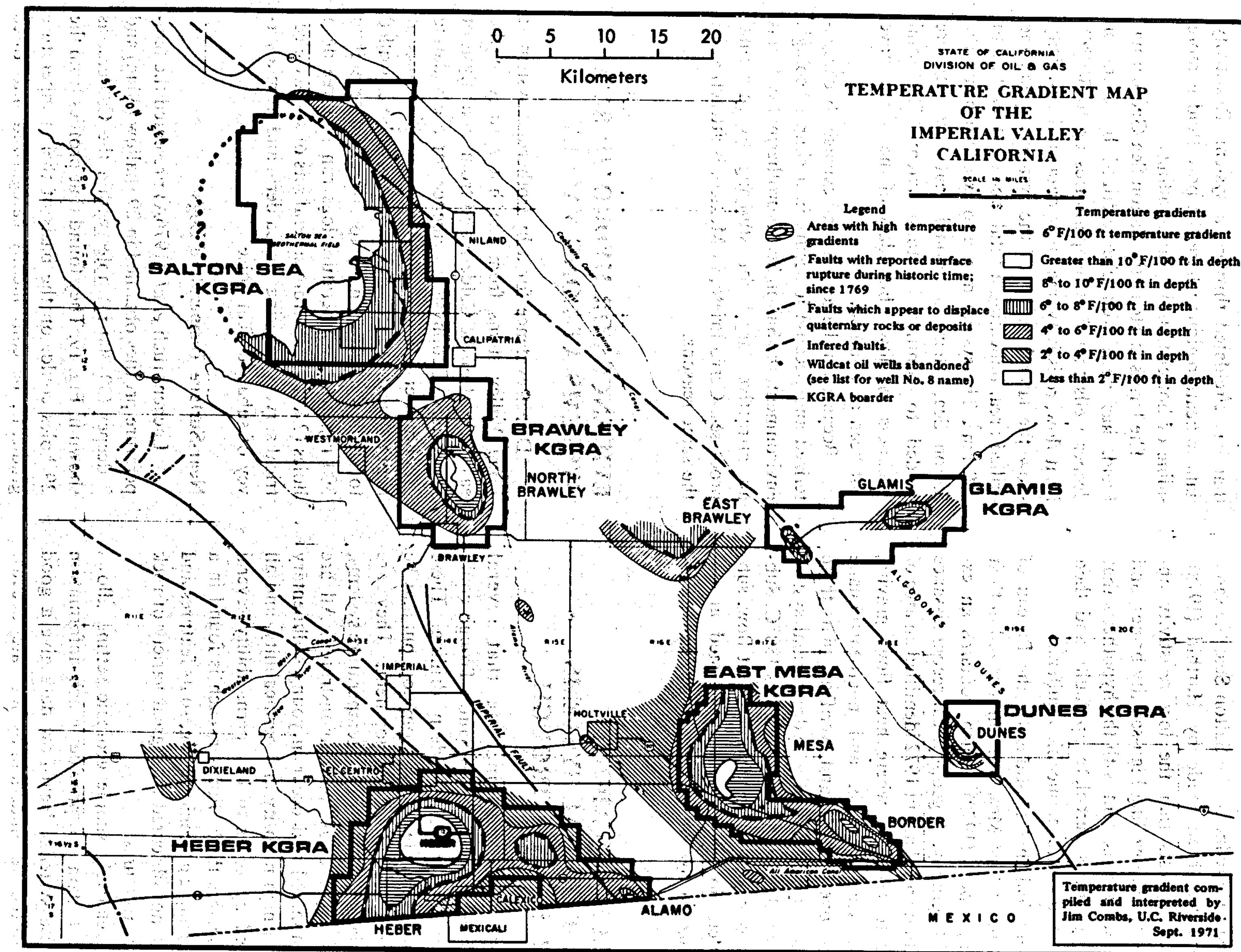

FIG. 18. Temperature-gradient map of the Imperial Valley, after Combs (1972).

a rather complex hydrological system. In general, all of the soundings indicated a resistivity low trending northwest–southeast, with the low axis going through the Obsidian Butte–Rock Hill area on the southeast shoreline of the Salton Sea; the same observation has been made by Harthill (1978). Geonomics suggests that the low resistivity observed is due to upwelling high-salinity brines, and their data set is consistent with the hypothesis that high-salinity brines percolate up faults and then migrate laterally into permeable aquifers.

Geophysical investigations by the Bureau of Reclamation at East Mesa are summarized in their concluding report (1979), which contains an excellent bibliography. A geophysical reconnaissance program that ended about 1972 identified an area of about 30 to 50 km^2, where heat flow, gravity, resistivity, seismic noise, and microearthquake anomalies overlapped, and verified that these anomalies were related to a hydrothermal system with measured temperatures as high as 180°C. Subsequent geophysical work (detailed heat flow, subsidence detection, etc.) has been aimed at understanding the system's production potential.

CONCLUSIONS ABOUT HYDROTHERMAL SYSTEMS AT THE SALTON TROUGH

Geophysical data indicate that several hydrothermal systems in the Salton Trough could provide good targets for deep drilling. These systems have many features in common, but may be at different stages of development. They are generally large, relatively high-temperature systems, and their existence is well-demonstrated. All are liquid-dominated and occur in young sedimentary rocks with high porosity and permeability, and all are capped by zones of low permeability several hundred meters thick. Further, each is found in a relatively simple geologic setting, so its features and effects are not obscured by a great variety of rock types and structures.

The relatively simple geologic setting allows geophysical data to provide much information about these systems. Heat-flow data give a good measure of the area underlain by hot water because the shallow thermal regime is conductive. And the young sediments in the Trough have, except where increased by hydrothermal circulation, relatively uniform densities. Consequently, gravity highs indicate system locations and, with several assumptions, provide a way to estimate the amount of fluid that has circulated through a system. Resistivity data indicate regions where the pore fluid is more saline or hotter than its surroundings.

The largest system, the Salton Sea Geothermal Field, has been well characterized by 2-km or deeper wells that find temperatures up to 360°C. Heat flow, resistivity, magnetic data, and seismicity show the lateral extent of the system beyond the drilled area, indicating a total area of at least 50 to 120 km^2. The system may also extend beneath the Salton Sea, where additional heat-flow data are needed to confirm its total extent. The total thickness and lower boundary of the system have not been determined.

Because it is large, hot, and extensively characterized and because it occurs in a simple geologic environment, the Salton Sea hydrothermal system is an attractive target for studying hydrothermal processes. Cerro Prieto, in Mexico, is attractive for the same reasons.

EVIDENCE FOR HYDROTHERMAL SYSTEMS AT LONG VALLEY

Direct evidence for a hydrothermal system at Long Valley stems from the hot springs in the eastern part of the Caldera and from an analysis of the waters. Geophysical evidence for a circulating groundwater flow system driven by a heat source at depth stems from temperature logs in holes drilled to shallow depths (many to 30 m and several to 300 m), from temperature logs run in three deep holes (2 km deep), from several sets of surface electrical surveys covering the Caldera, and from a survey of subacoustical noise.

Near-Surface Temperature Logs

Lachenbruch *et al.* (1976A) presented the temperature profiles obtained in 29 shallow (30 m) and 7 deeper (30–300 m) holes within the Caldera, as shown in Fig. 19. They found that profiles from the 29 holes could, on the basis of their temperature gradients, be conveniently categorized into three groups (see Table 1).

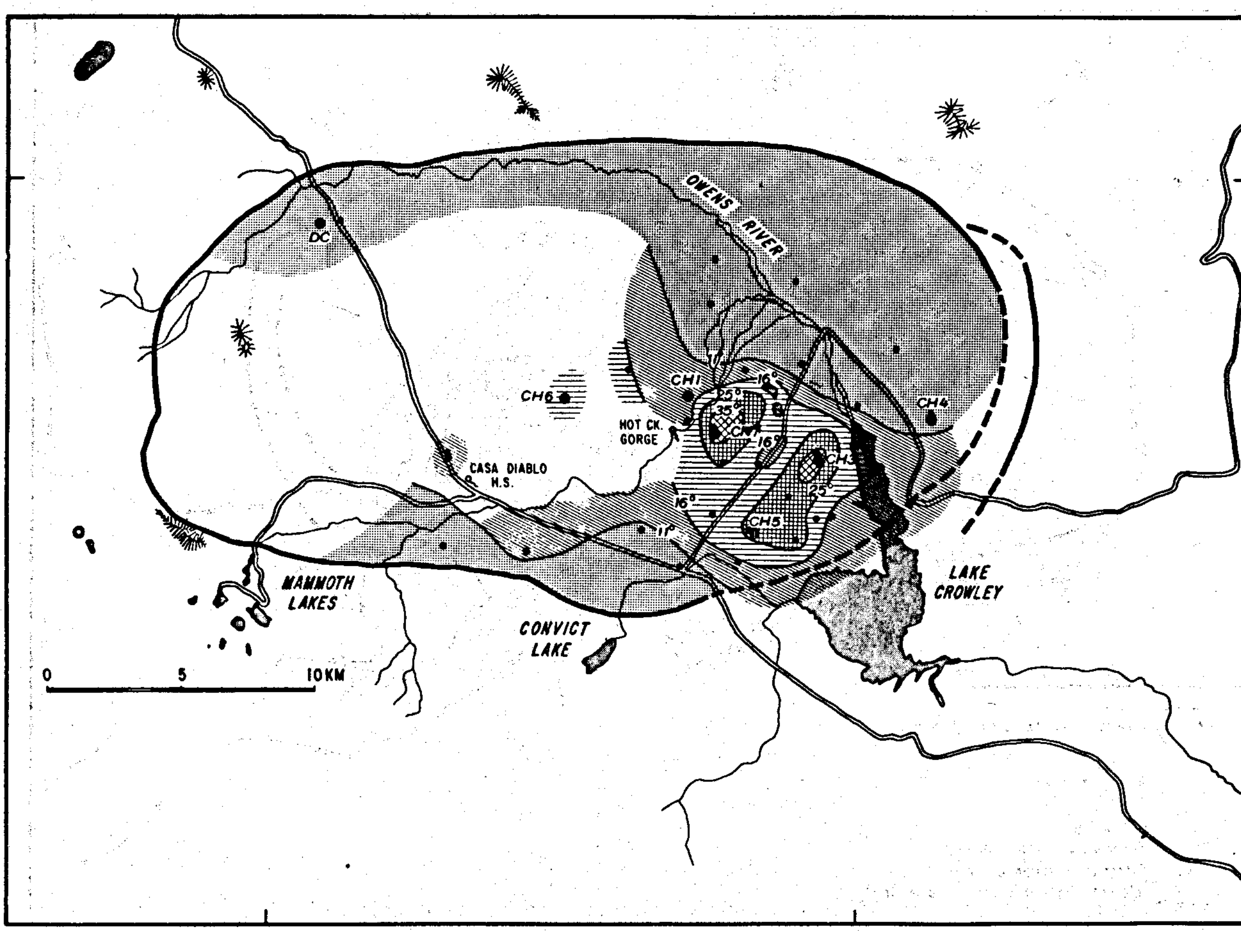

FIG. 19. Shallow temperatures at Long Valley in June 1974, after Lachenbruch *et al.* (1976A). Temperatures measured at 10 m depth in shallow holes (indicated by dots) are contoured.

The zone of convective heat transport covers more than 40 km^2. The temperature at 10 m depth, which Lachenbruch *et al.* (1976A and B) found indicative of the thermal gradient, is contoured in Fig. 19. The high-temperature, high near-surface heat-flow area coincides well with the eastern lobe of the locus of hot springs activity displayed in Fig. 20. Hence, the near-surface temperature data appear to contain few surprises. However, three of the deeper holes, which were drilled into areas deemed to be convective, showed an overturn of thermal gradients, indicating considerable complexity in the distribution and mixing of hotter and colder waters. Lachenbruch *et al.* warn that extrapolation of near-surface temperatures even 100 m downward is hazardous.

TABLE 1. Three categories of temperature profiles from 29 holes in the Caldera, based on their temperature gradients.

Temperature at 10 m, °C	Gradient, °C/km	Heat flow HFU	Nature
11	0	—	Zone of recharge, generally toward Caldera periphery
11–16	200–400	4–8	Zone of conductive heat flow
16	Irregular	To 50	Zone of convective heat transport

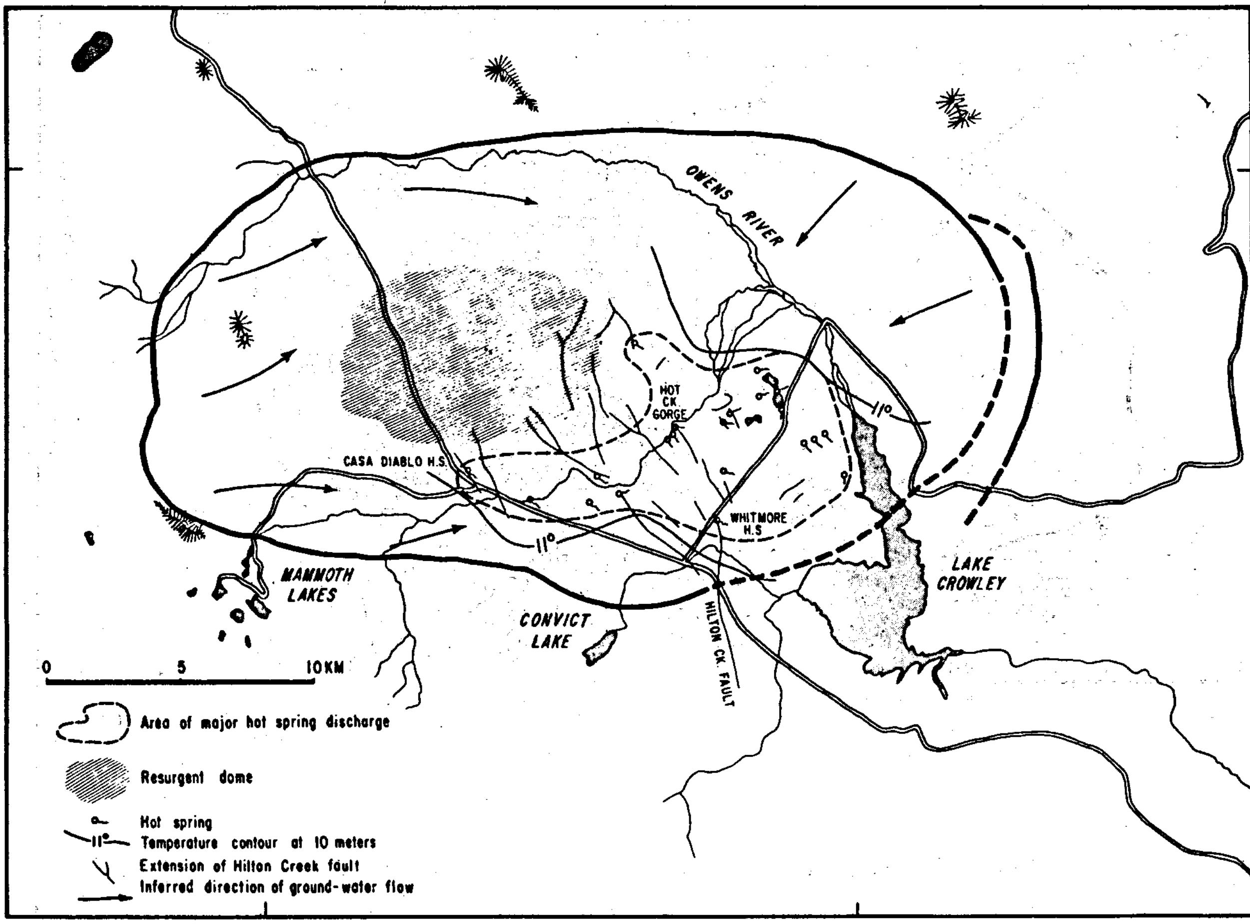

FIG. 20. Map of Long Valley Caldera, showing relations among groundwater flow, hot spring discharge, 10-m temperatures, and the resurgent dome, after Lachenbruch *et al.* (1976A).

Temperature Logs from Deep Holes

Smith and Rex (1977) presented the drilling history, the geological section, and the porosity and temperature logs obtained in a 2109-m-deep hole drilled near the center of the near-surface thermal anomaly. Those data are shown in Fig. 21. The upper 250 m of the temperature logs displays a high gradient indicative of convective transport. Below about 250 m the temperature declines in a fashion that is, to a great extent, controlled by permeability variations of the underlying tuffs (as inferred from the porosity). Figure 21 furnishes dramatic evidence that the near-surface temperature gradients, especially in the convective regime, cannot be extrapolated to depth.

Well-log data from two additional exploratory geothermal holes (Steve Mione, Union Oil Company, private communication, 1980) within the Caldera have just become available. Clay Pit No. 1, collared in the north-central part of the Caldera, extends to almost 2 km depth, where the maximum temperature of 146°C is attained. The gradient in the upper 900 m of hole is perturbed by the hydrological system. Below 900 m, the temperature increases monotonically except for a disruption caused by the loss of drilling fluid.

The third hole, Mammoth No. 1, is collared in the Casa Diablo area. The maximum temperature of 157°C is attained near-surface at a depth of 120 m. A secondary temperature peak of 142°C occurs at 680 m depth. The bottom-hole temperature at 1600 m depth is only 102°C.

If the temperature profiles represent the undisturbed temperature field, several conclusions can be drawn about the hydrothermal system. All wells show the effects of fluid transport of heat from the upper 100 m to depths of up to 1 km. Those effects include temperature inversions and isothermal

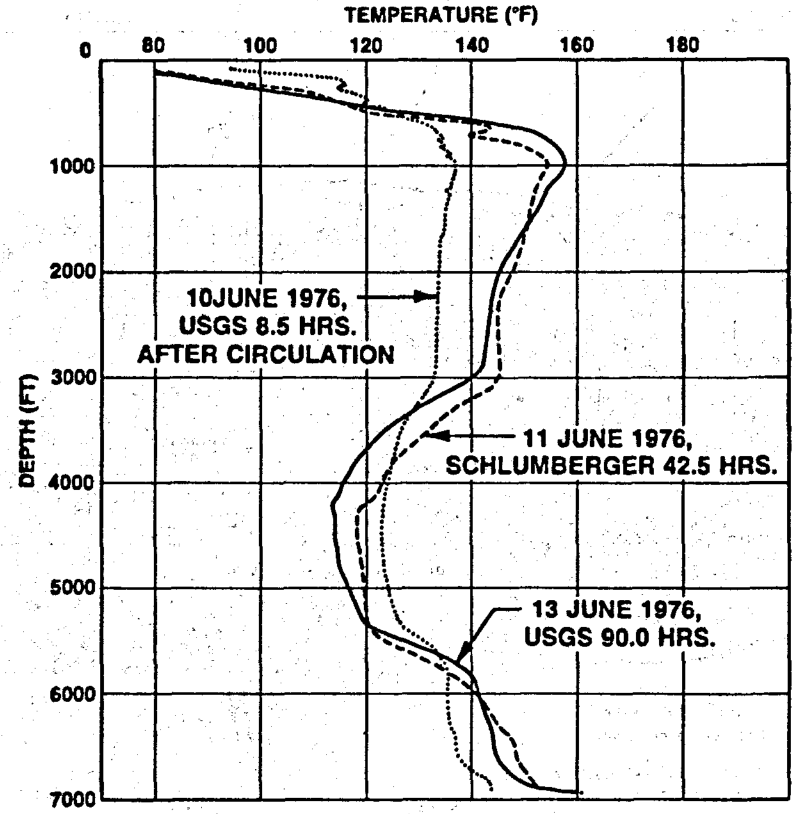

FIG. 21. Temperatures in a deep well near the center of the heat-flow anomaly in Long Valley, after Smith and Rex (1977). Data are from the Republic Well, Long Valley.

zones. The temperature inversions indicate substantial lateral flow of fluids as hot as 140 to 160°C over an area of at least 50 km^2, but the direction of flow and the location of the heat source are not known.

The Mammoth No. 1 and Republic holes imply that significant lateral heat transport is taking place. Moreover, the locus of a high-temperature hydrothermal system has now been further constrained by the release of new well data. The temperature profiles and accompanying well logs from these three 2-km-deep holes should be analyzed further to advance our understanding of the hydrothermal regime.

Electrical Surveys

Information on the electrical-resistivity structure of the Long Valley Caldera was obtained by four survey techniques. Hoover *et al.* (1976) report apparent resistivity values obtained by the audio–magnetotelluric technique of sampling the impedance of the earth's surface at a range of frequencies where sufficient electromagnetic energy is generated by thunderstorm activity. The penetration depth of electromagnetic waves depends on both frequency and earth resistivity. It varies from a few meters at high frequencies and low resistivity values to 2.5 km at several hundred ohm-meters and 8 Hz, the lowest frequency. The contour patterns are remarkably similar to the locus of hot-spring activity depicted in Fig. 20, but are displaced to the west of the high-temperature anomaly of Fig. 19. The station spacing used for the audio–magnetotelluric data was rather coarse, and additional measurements could quickly produce a higher resolution map of the near-surface conductive zone.

A more-comprehensive resistivity map of the Caldera is presented by Stanley *et al.*, 1976 (Fig. 22); it is based on a transmitter-bipole-roving-dipole survey, using three of the four bipole sites. Here again, there are few surprises when the low-resistivity areas of Fig. 22 are compared with the locus of the major hot-spring discharge. The bipole–dipole data were refined by Schlumberger

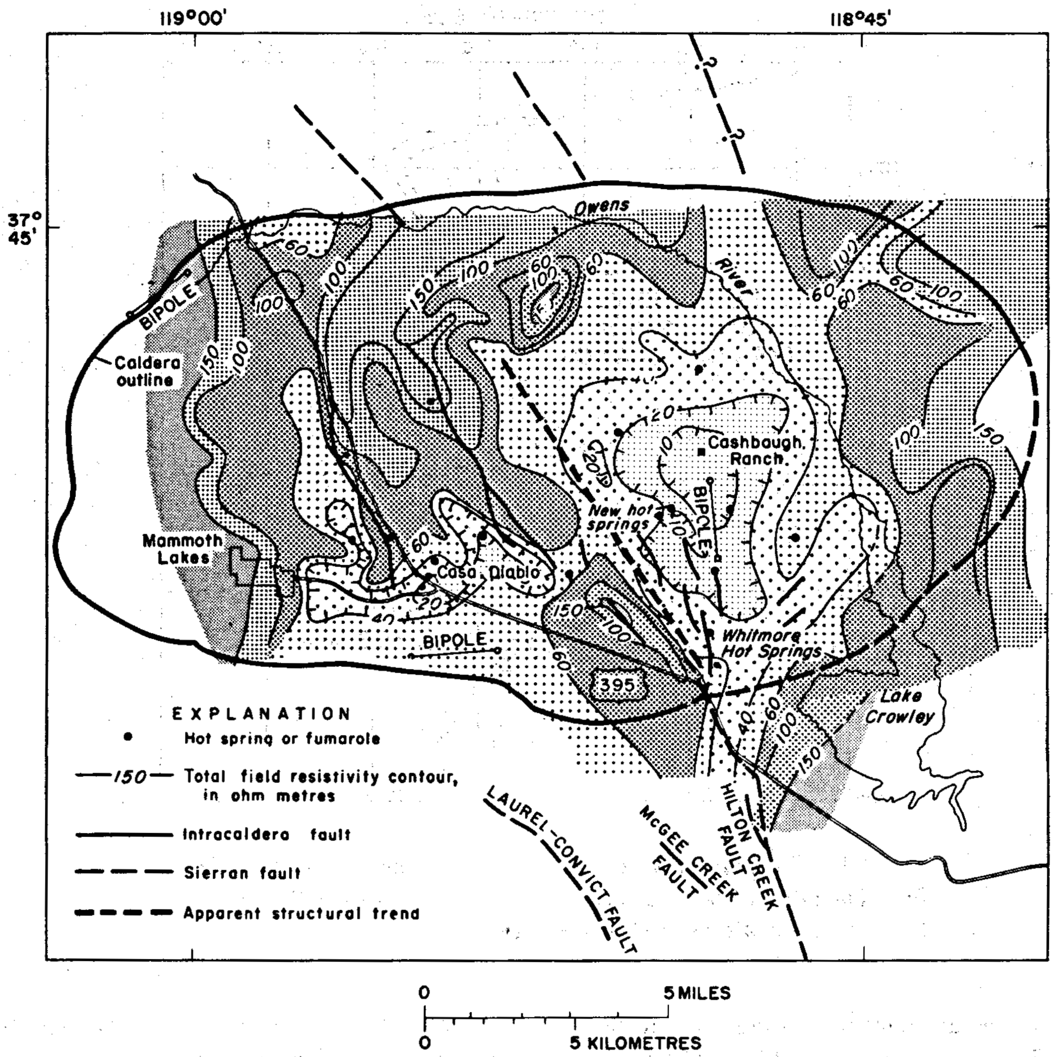

FIG. 22. Composite total field resistivity map for Long Valley, using data from three source locations, after Stanley *et al.* (1976).

soundings and transient magnetic soundings from a number of sites within the Caldera. Stanley *et al.* (1976) present the interpreted results in the form of a fence diagram that shows areas of about 50 km^2 underlain by substantial near-surface volumes of rock (1 km thick) with resistivities of less than 10 ohm-meters. These results can be interpreted by examining the well-log data from Mammoth No. 1. They probably map the presence of the highly porous upper Bishop tuff and early post-Caldera tuffs, with the hotter groundwater possibly contributing to the reduced resistivity.

CONCLUSIONS ABOUT HYDROTHERMAL SYSTEMS AT LONG VALLEY

Geophysical data map an area of about 50 to 200 km^2 in Long Valley where hydrothermal convection is important in the upper 600 to 1000 m depth. Measured temperatures in that system are up to 160°C, and geochemical estimates of reservoir temperature are much higher. The hydrothermal system has not been characterized well enough to determine the thickness of the convecting systems, the directions of fluid flow, or the amount of fluid involved. Geophysical evidence for substantial volumes of porous Bishop tuff within the Caldera suggest that the hydrothermal system could be large and extensive. Low temperatures and low gradients at depth in deep wells indicate that the location of part of the system is not known and that it could be quite deep.

Geophysical techniques have defined many features of the hydrothermal system at Long Valley. Heat-flow measurements define the area where the near-surface thermal regime is influenced strongly by the hydrothermal system. An area of 40 km^2 shows high and variable heat flow, indicating very shallow convection. A larger area, about 200 km^2, includes the region with high conductive heat flow (4–8 HFU), indicating the top of the hydrothermal system is at a greater depth. Deep holes indicate that lateral flow is important, that the hydrothermal system is at least 600 m thick, and that we don't know the fluid flow patterns in the system. Resistivity measurements map out a 1-km-thick layer of more than 50 km^2 that might be an aquifer containing the hydrothermal system.

The Long Valley hydrothermal system is an attractive target for deep drilling for two reasons: it is large and extensively studied. Until we better understand the relatively low-temperature portion of the system, however, we cannot ensure that drill holes will find the high temperatures indicated by geothermometry.

EVIDENCE FOR HYDROTHERMAL SYSTEMS AT ROOSEVELT HOT SPRINGS

Data

Both the accepted model of the hydrothermal system at Roosevelt Hot Springs and the geophysical data leading to that model are summarized by Ward *et al.* (1978). The general model involves the influx of meteoric water through fracture systems in the Mineral Mountains, the heating of the meteoric water by an unspecified deep heat source, and the rising of the heated water through a zone of high permeability associated with the Opal Mound and Hot Springs fault.

Figure 23 is a heat-flow map of the area. Ward *et al.* suggest that the 100-$mW\text{-}m^{-2}$ (2.4 HFU) contour represents the boundary of the zone where surface heat flow is influenced by the convecting system. The low heat-flow values over the Mineral Mountains (less than 80 $mW\text{-}m^{-2}$) indicate that recharge occurs there. The control of upward flow by fractures is indicated by the shape of the anomalous zone. An area of over 40 km^2 has surface heat flow above 400 $mW\text{-}m^{-2}$.

Other geophysical methods are used to map the faults that control the hydrothermal system. In

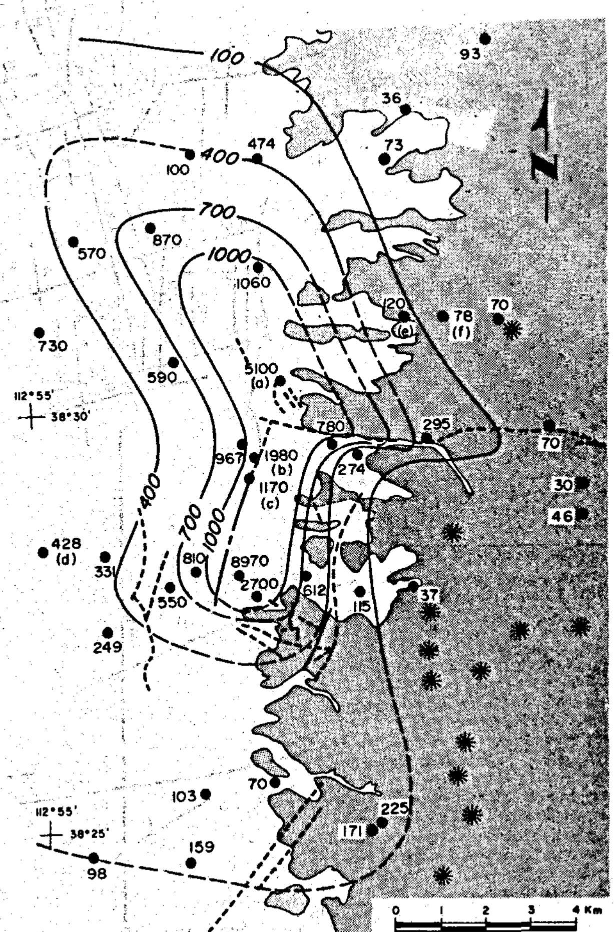

FIG. 23. Heat-flow map of the Roosevelt Hot Springs area. Heat-flow values are reported in milliwatts per square meter at the sites indicated by black dots. Major fault patterns are indicated by heavy solid and broken lines, after Ward *et al.* (1978).

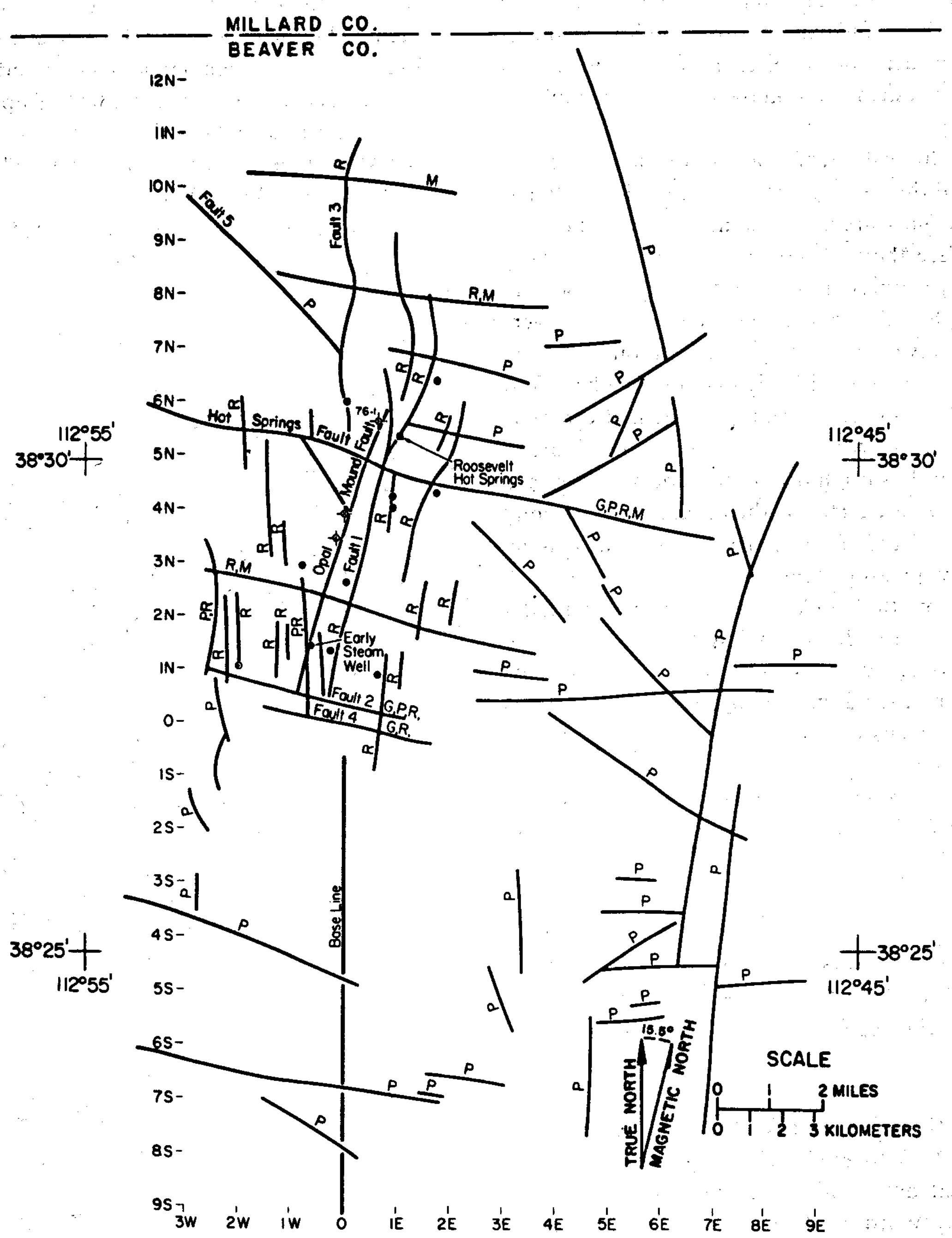

FIG. 24. Summary of detected faults in the Roosevelt Hot Springs area, after Ward *et al.* (1978). P, G, M, and R indicate faults detected by photographs, geologic observation, aeromagnetics, and resistivity.

Fig. 24, the inferred faults are labeled with letters indicating the methods used for detection. An aeromagnetic survey shows faults in this area because they displace or truncate volcanics or rocks of the pluton. The resistivity within the fault zones is very low because of clay.

CONCLUSIONS ABOUT HYDROTHERMAL SYSTEMS AT ROOSEVELT HOT SPRINGS

Geophysical measurements provide information about the hydrothermal system at Roosevelt Hot Springs. High heat flow is found in an area of more than 40 km^2. The faults and fractures that control the fluid flow are mapped by aeromagnetic surveys that sense the truncation of volcanic beds and by resistivity surveys that sense clay in fault zones. Reported well temperatures are as high as 269°C.

The details of the Roosevelt system's plumbing are not known; the depth of water circulation, the volume of water involved, and the actual flow paths are of interest. The hydrothermal system apparently has few inherent physical properties, only high temperatures and low electrical conductivity in relatively localized fault zones. Consequently, geophysical data are used mainly to study the geologic setting of this system.

The hydrothermal system at Roosevelt is an attractive target for deep drilling for two reasons: drilling might help us understand the "plumbing" of the system, and the knowledge might be applied to many Basin and Range systems. On the other hand, fracture zones where the system is active provide small targets that might easily be missed, and considerable understanding would have to be developed before a hole could, with certainty, be sited toward high temperatures.

EVIDENCE FOR HYDROTHERMAL SYSTEMS AT THE GEYSERS–CLEAR LAKE AREA

The Geysers hydrothermal system has three important characteristics. First, as the resource for extensive commercial power production, it has been explored and verified by many techniques and wells. Second, it has produced enough fluid to make the perturbing effects of production useful for understanding the dynamics of the natural system. Third, it is the only vapor-dominated geothermal field in the United States. Consequently, The Geysers hydrothermal system provides the opportunity to study a well-characterized system.

Geophysical Data About the Hydrothermal System at The Geysers–Clear Lake Area

Geophysical methods have been used in two ways at The Geysers: to delineate the natural system and to observe the effect of production on it. The means of interpreting geophysical data here is unique because of the unusual and complex geology. For example, Isherwood (1976) finds a magnetic anomaly corresponds to the steam field when the magnetic data are converted to pseudogravity. He interprets this effect as the result of destruction of magnetite by alterations in the hydrothermal system.

Active seismic techniques could shed light on the structure of the geothermal reservoir. Standard oil-field interpretations of reflection and refraction seismic sections are of marginal use in many geothermal areas because of the complex metamorphic and volcanic terraces involved. At The Geysers, present production comes primarily from fractures in the Franciscan graywacke. From a relatively shallow-penetration Vibroseis survey, Denlinger and Kovach (personal communication, 1980) recognized reflecting horizons, which may be shear zones, within the Franciscan complex southeast of The Geysers. Although the equipment they used severely limited resolution, their results raise the possibility that some fracture zone may be mappable from the surface.

The difficulties of interpreting electrical methods in The Geysers region make accurate delineation of the hydrothermal system by these methods impossible. For example, much of the data of Stanley, Jackson, and Hearn (1973) can be modeled by a conductive unit that is 1 to 3 km thick, that is buried no deeper than 1.1 km, and that extends over 60 km^2. The unit could be a hot saturated sandstone representing the top of the convecting system, or it could be extensive, relatively impermeable shales.

Heat-flow and equilibrium temperature data from wells are the most direct way to study a hydrothermal system. Unfortunately, few data from wells at The Geysers have been released. At The

Geysers, convection may be limited to the reservoir itself, with conductive regions above and below. For two drill holes within The Geysers steam field, away from active steam fumaroles, Urban and others (1976) found the interval between the surface and the steam reservoir to be mainly conductive. They calculate the steam field must have existed for at least 10,000 years for the cap to reach thermal equilibrium. None of the deep holes (up to 3.5 km depth) has penetrated the region beneath the essentially isothermal region of the steam zone.

Geophysical measurements can be used to measure the response of the hydrothermal system to extensive fluid production. Seismicity studies are not useful in delineating the natural system because they began after steam production started and only show strong seismicity from the production zone. Isherwood (1977) used precise measurement of small changes in the gravity field to estimate the amount of mass lost from the hydrothermal system. He found, within the uncertainty of his estimate (about 20%), that none of the produced mass is being recharged. Furthermore, he calculated that the mass decrease, presumably due to boiling, takes place throughout a depth range that is too shallow to be the boiling of a water surface below the steam zone. Consequently, any model that describes the development of The Geysers steam field must account for relatively impermeable boundaries and for distributed and intermixed zones of steam and liquid.

CONCLUSIONS ABOUT THE HYDROTHERMAL SYSTEM AT THE GEYSERS–CLEAR LAKE AREA

The vapor-dominated hydrothermal system at The Geysers is large (50 to 100 km^2) and hot (approximately 240°C). Commercial production of over 1000 MWe of electricity testifies to the certainty of that system's existence. Several features of the system are not known: its depth, what lies beneath it, and its history of evolution. Most information about the hydrothermal system has come from drilling. Because of the complex geologic setting, the physical properties of the hydrothermal system could not be predicted without measurements from drill holes. Consequently, geophysical data have only served to supplement the drill hole data. Gravity and magnetic anomalies map out the steam field. Seismic-reflection data indicate structures within the Franciscan rocks that may be shear zones controlling the fluid paths.

The hydrothermal system at The Geysers is attractive as a deep-hole drilling site for several reasons. The target exists, and there are a lot of data to characterize it extensively. The system is large and involves considerable energy. Many questions concerning its evolution are unanswered. Finally, commercial interest may allow piggyback experiments in wells planned for nonscientific purposes.

The Geysers system also has drawbacks as a target. Because it is a unique, vapor-dominated system, results from experiments done here may not be applicable elsewhere.

Comparatively little is known about the extent of the hydrothermal system associated with the Clear Lake volcanic field. The physical properties of the hot-water system near Clear Lake are not known from drilling. Two drill holes have apparently encountered temperatures of 180 to 200°C in the middle of an extensive resistivity low. If the resistivity low represents the size of the hot-water system, its area could be as great as 60 km. Heat-flow studies could define the extent and nature of the system. Additional geophysical measurements are difficult to recommend in this complicated setting.

SUMMARY

A considerable body of evidence exists for magma bodies and hydrothermal systems at the five target sites. The sites provide potential magma targets that range in composition from basaltic to silicic in a variety of settings. They also provide a range of potential hydrothermal targets that vary from massive, steam-dominated systems to relatively low-temperature, fracture-dominated systems. Further exploration or drilling in these areas could be used to study scientific questions that range from understanding specific details of the heat-transfer mechanisms to improving models of the evolution of caldera and rift-intrusion systems.

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