GEOPHYSICS, VOL. 34, NO. 3 (JUNE 1969), P. 463-478, 11 FIGS., 2 TABLES

A COMPARISON OF TERRESTRIAL HEAT FLOW AND TRANSIENT GEOMAGNETIC FLUCTUATIONS IN THE SOUTHWESTERN UNITED STATES†

ROBERT E. WARREN,* JOHN G. SCLATER,‡ VICTOR VACQUIER,‡ AND ROBERT F. ROY**

Magnetic time-variations between Tucson, Arizona and Sweetwater, Texas indicate that a zone of high electrical conductivity underlies the southwestern United States. The interpretation of this zone by Schmucker as a rise of the isotherms in the upper mantle is supported by six heat flow observations along the line of the geomagnetic profile. These and other observations indicate a high but variable heat flow in the Basin and Range Province which contrasts strongly with the uniform values of 1.1 μcal/cm^2 sec reported for the Texas Foreland. The width of this high heat flow anomaly, which may extend across the entire Basin and Range Province, suggests anomalously high temperatures in the upper mantle. This interpretation is further supported by magnetotelluric data between Phoenix, Arizona and Roswell, New Mexico and by the low seismic P_n velocity and negative gravity anomaly. It is suggested that the "anomalous mantle" may be related to the tectonic evolution of the western United States and the late Cenozoic fault system.

INTRODUCTION

The southern Arizona and New Mexico Cordillera is an area of great tectonic instability with an eventful and diversified igneous history which began in the Cretaceous and has continued until the present. The major part of this activity occurred during the late Cretaceous and early Tertiary and is particularly noticeable along the Rio Grande rift belt. To the east of the rift belt lies the stable Texas Foreland, which has been relatively stable since the Precambrian (Eardley, 1962) and contrasts strongly with the tectonically active Cordillera. To investigate whether the surface differences between the regions were reflected in the subcrustal mantle, Schmucker (1964) analyzed the transient geomagnetic variations from a closely spaced network of temporary three-component magnetometer observatories (Figure 1), along an approximate east-west line across the two regions. He suggested the presence of a zone of anomalously high electrical conductivity under southern Arizona and New Mexico. More recently, Swift (1967) has analyzed data from four magnetotelluric stations between Phoenix, Arizona and Roswell, New Mexico (Figure 1) and the magnetotelluric data support Schmucker's interpretation. Both Schmucker and Swift have interpreted this zone to mean that higher temperature isotherms are shallower in it than in the area to the east and have suggested that the different igneous and tectonic histories of the Basin and Range Province and the Texas Foreland might be related to thermal imbalance in the upper mantle.

In order to investigate whether the suggested irregularities in the deep isotherms are reflected

† Contribution of the Scripps Institution of Oceanography, new series. Manuscript received by the Editor August 24, 1968.

* Ocean Science Engineering, Long Beach, California 90802.

‡ University of California, San Diego, Marine Physical Laboratory of the Scripps Institution of Oceanography, La Jolla, California 92037.

** California Institute of Technology, Seismological Laboratory, Division of Earth Science, Pasadena, California 91105.

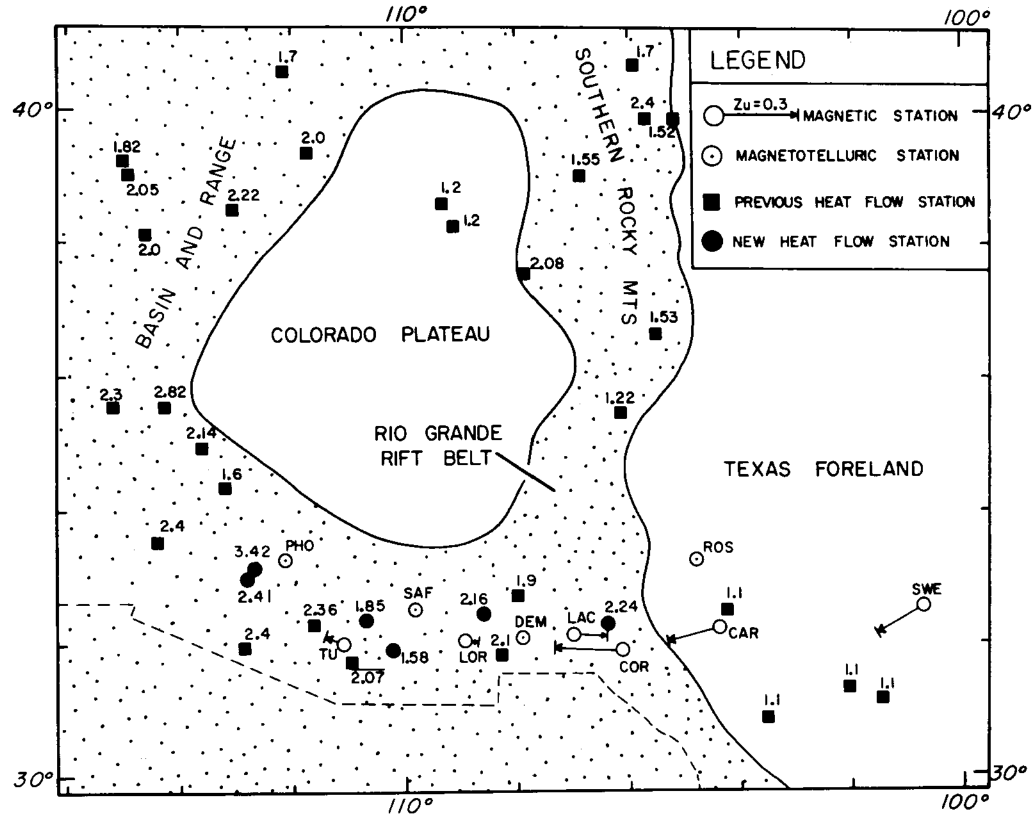

FIG. 1. Location of geomagnetic and magnetotelluric observatories and heat-flow measurements in the southwestern United States. The anomalous Z variations are represented by in-phase induction arrows Z_u for 1 cph (cycles per hour) from Schmucker (1969). Additional heat-flow values from Roy et al (1968b), Herrin and Clark (1956), Birch (1950), Spicer (1964), and Costain and Wright (1968).

as increased heat flow at the surface, a set of heat flow observations was made along the line of the geomagnetic stations. Six boreholes were drilled between Buckeye Hills, Arizona and Carlsbad, New Mexico; heat flow values were obtained on all the holes. In addition there are available heat flow measurements reported by Herrin and Clark (1956) in west Texas and by Roy et al (1968b) in Arizona. The position of each heat flow station on the profile and others in the area is shown on Figure 1.

GEOLOGICAL ENVIRONMENT AND DRILLING

As the drilling of special holes for the measurement of terrestrial heat flow is both time consuming and expensive, sites were selected only after careful reconnaissance of the surrounding topography and geology. Important desirable features of a drill hole are absence of groundwater circulation, relatively uniform thermal conductivity, and low surface relief. Each of the sites was selected with these criteria in mind.

The holes were drilled in granite outcrops of low topographic relief. The location, drilling history, and other data for each of the holes are summarized in Table 1. They were diamond drilled and continuously cored with BX wireline equipment which recovers cores 36 mm in diameter from a 60 mm hole. A brief description of the local geology of each of the locations is given in Table 2.

THERMAL CONDUCTIVITY

Thermal conductivity of representative samples of the core was measured with both the transient needle probe apparatus described by Von Herzen and Maxwell (1959) and with the steady-state

Table 1. Location and geothermal data for six drill holes in Arizona and New Mexico. Depth range is the portion of the hole used for the heat-flow measurement. N_c is the number of conductivity measurements by divided bar (DB) and needle probe (NP) methods. N_t is the number of temperature measurements used in the gradient calculation. The error limits are standard errors.

Location	Latitude	Longitude	Altitude, meters	Drilling started completed	Temperatures measured	Depth range meters	N_c	Mean thermal conductivity $\times 10^3$ cal / cm sec °C	N_t	Temperature gradient $\times 10^4$ °C / cm	Heat flow $\times 10^6$ cal / cm² sec
UCSD-1, Buckeye Hills, Arizona	33°17′00″N	112°37′37″W	280	4-9-64 5-1-64	5-10-65	143 161	NP 32 DB 6	6.75 ± 0.09	7	5.07 ± 0.08	3.42 ± 0.07
UCSD-2, Rainbow Valley, Arizona	33°11′28″N	112°39′00″W	271	5-5-64 5-26-64	5-10-65	46 107	DB 11	7.89 ± 0.06	5	3.06 ± 0.04	2.41 ± 0.04
UCSD-3, Oracle, Arizona	32°37′10″N	110°48′33″W	1295	6-1-64 6-15-64	5-1-65	180 277	DB 23 NP 32	8.23 ± 0.08	33	2.25 ± 0.01	1.85 ± 0.02
UCSD-4, Dragoon, Arizona	32°02′41″N	110°04′06″W	1463	6-17-64 7-7-64	5-2-65	210 290	DB 14 NP 20	7.20 ± 0.07	27	2.20 ± 0.01	1.58 ± 0.02
UCSD-5 Tyrone, New Mexico	32°40′30″N	108°29′15″W	1853	7-9-64 7-23-64	7-25-64 7-27-64 8-4-64	183 290	DB 32 NP 16	7.52 ± 0.05	8	2.87 ± 0.03	2.16 ± 0.03
UCSD-6 Orogrande, New Mexico	32°26′30″N	106°05′44″W	1341	7-28-64 8-6-64	8-12-64 8-21-64	107 274	DB 22 NP 10	5.59 ± 0.03	12	4.01 ± 0.02	2.24 ± 0.02

divided bar apparatus described by Roy et al (1968b). The absolute values of quartz and silica glass given by Ratcliffe (1959) were used to calibrate both techniques.

The divided bar measurements were made on samples 38 mm thick which had been saturated with water under pressure (200 bars), after being degassed in a vacuum for 24 hours. Water saturation increased the conductivity from 1.3 percent in samples from UCSD-6 to as much as 6.5 percent in those from UCSD-3. All measurements were made under an axial stress of 50 bars and within 5°C of the in-situ temperature. Measurements of conductivity with the divided bar apparatus were reproducible to ± 1 percent.

Needle probe measurements were made in 1.5 mm diameter holes drilled through the sample normal to the core axis. At first, to insure good thermal contact, the hole was filled with mercury. As this was both difficult and time consuming, an alternate technique was developed. After the small hole was drilled, porous samples were soaked in water and then encapsulated in American Cyanamid AM-9 chemical grout. The gelled

Table 2. Geological description of drill holes

Location	Description	Total depth (meters)	% Core recovery
UCSD-1, Buckeye Hills, Arizona	Shattered monzonite stock of Laramide age intrusive into Precambrian gneiss and schist. Drilling extremely difficult due to heavy ground.	163	75
UCSD-2, Rainbow Valley, Arizona	Precambrian granite. Groundwater circulation encountered from water table to total depth. Hole abandoned.	242	95
UCSD-3, Oracle, Arizona	Precambrian granite with very old erosional surface. Hole collared in the Oracle granite about 1.5 miles north of Mogul Fault which may be expression of the Texas Lineament. Numerous shear zones encountered to total depth.	283	95
UCSD-4, Dragoon, Arizona	Quartz monzonite stock of Laramide age intrusive into Paleozoic sediments and metasediments.	294	99+
UCSD-5, Tyrone, New Mexico	Precambrian granite of the Big Burro Mountains.	297	99
UCSD-6, Orogrande, New Mexico	Composite stock of Post Eocene age intrusive into Paleozoic sediment. Hole collared and drilled in diorite to total depth.	293	95

grout gave good thermal contact between the needle probe and the hole wall and also retained the interstitial water in the sample. All measurements were made at atmospheric pressure between 25° and 30°C. For a few samples repeated measurements were made using mercury, AM-9, and silicone grease as the hole filler. No detectable difference was found in the apparent conductivity of the sample. Since the cores are granular in character and only a small volume of the material is actually sampled during one measurement, a single reading is not a reliable sample. Consequently four equally spaced measurements were taken in the hole and the conductivity value used was the mean of the four determinations. The degree to which the values of conductivity remained constant for any particular core sample was dependent upon grain size of the rock, and varied from ± 1 percent to ± 10 percent for a medium grained (2–4 mm) monzonite. The great majority of the values varied less than ± 5 percent.

Twenty-five samples were measured with both techniques, and after correcting for differences in temperature and water saturation, both methods agreed to within 5 percent. The average conductivity values listed in Table 1 are harmonic means with equal weight given to the transient and steady state measurements.

TEMPERATURE MEASUREMENTS

The borehole temperatures were logged with a thermistor probe in one arm of a Wheatstone bridge. The change of resistance versus temperature of the thermistor was calibrated to a relative accuracy of ± 0.01°C over any 3°C interval against a Beckman mercury-in-glass differential thermometer. An absolute accuracy for the thermistors of ± 0.05°C was obtained by calibrating them against a National Bureau of Standards

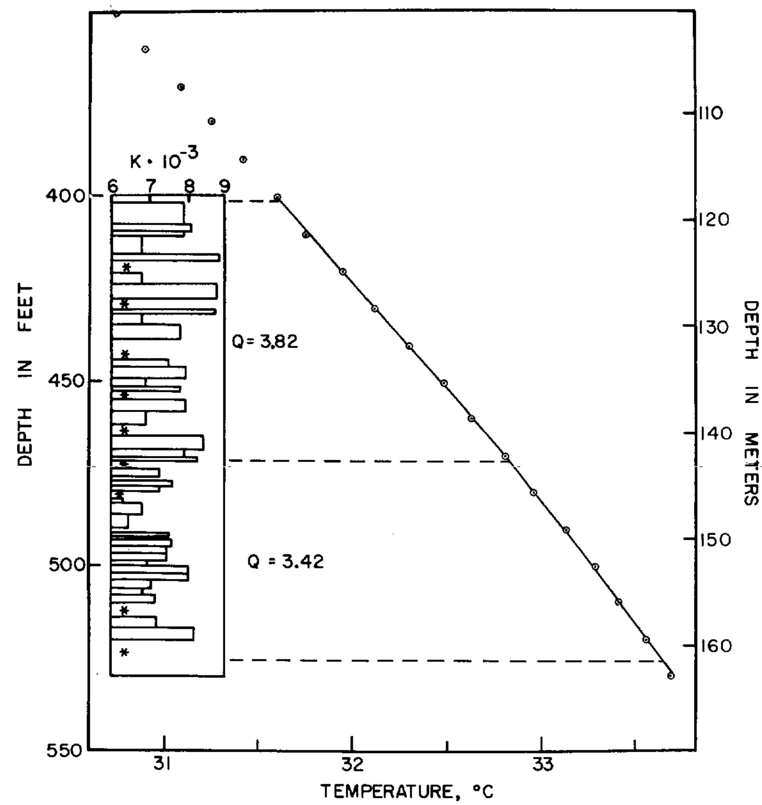

FIG. 2. Temperature and thermal conductivity data for UCSD-1. K given in cal/cm sec °C. Asterisk indicates core not recovered. A conductivity of 5.90×10^{-3} cal/cm sec °C was assumed for these sections. Heat flow values for the intervals 122 m–143 m and 143 m–161 m are shown on the diagram. If the two intervals are combined, the heat flow is 3.66 ± 0.06 μcal/cm^2 sec.

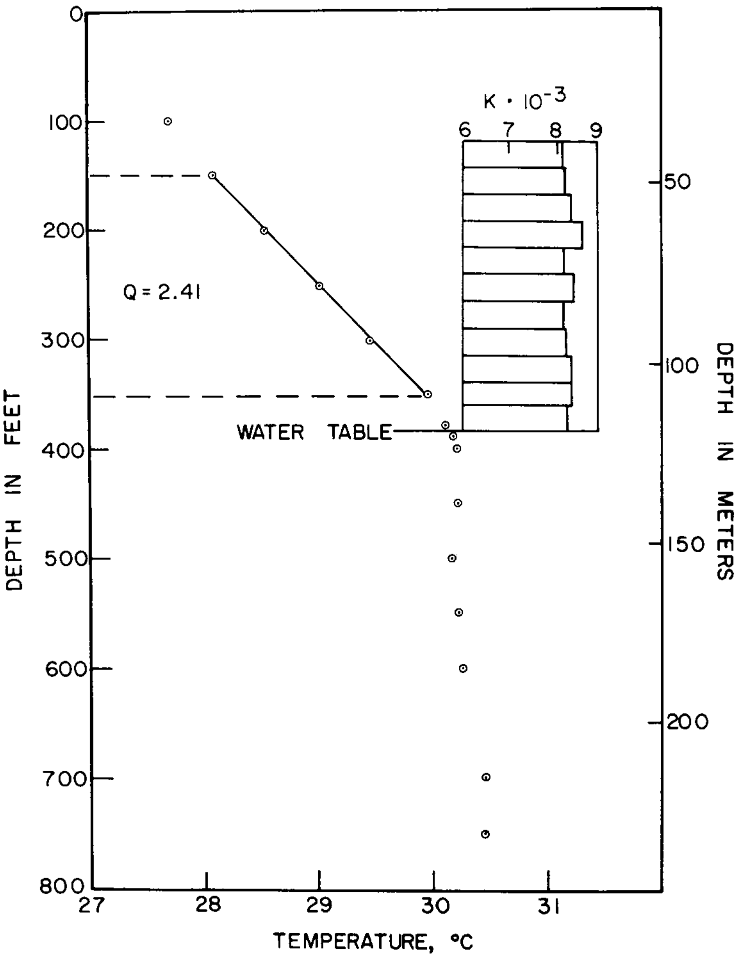

Fig. 3. Temperature and thermal conductivity data for UCSD-2.

certified mercury-in-glass thermometer. In order to minimize errors caused by self heating, the thermistor resistance was measured with the same equipment and the same bias for both the calibrations and the hole logs.

The observed temperature profiles for each borehole are presented in Figures 2–7. For UCSD-1, UCSD-3, and UCSD-4 the profiles are of "near equilibrium" temperatures obtained at least nine months after drilling had ceased. Unfortunately it was not possible to obtain equilibrium values for UCSD-5 and UCSD-6 because within two months of drilling, each hole was bridged over near the surface—UCSD-5 by vandals and UCSD-6 due to hole cave. Since repeated measurements were made at given depths at known time intervals shortly after drilling had ceased, equilibrium temperatures can be determined for each of these holes.

Lachenbruch and Brewer (1959) have shown that the equilibrium temperature $\theta(\infty)$ at a given depth z can be determined from the equation

$$\theta(t) - \theta(\infty) = C \ln\left(\frac{t}{t - s}\right), \quad (1)$$

where s is the time interval between the time the drill bit first reached the depth z and the time drilling ceased, and t the time measured from the time the drill bit first reached the depth z, and $\theta(t)$ is the temperature observed at time t. C is a constant which depends on the diameter of the hole, the temperature, the depth and the thermal

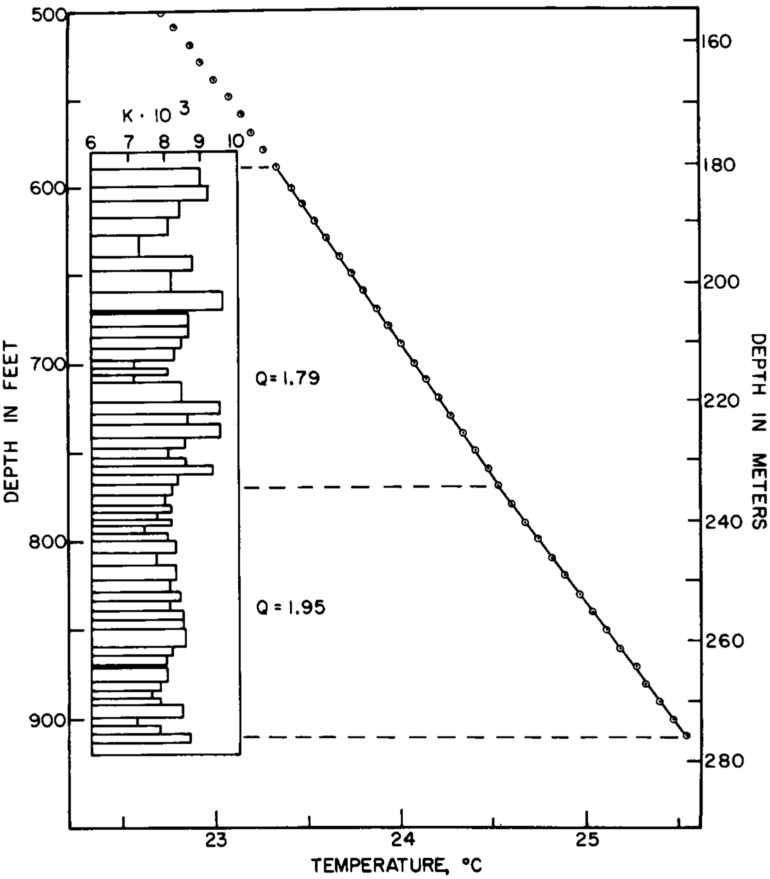

FIG. 4. Temperature and thermal conductivity data for UCSD-3. Heat-flow values for the intervals 180 m–235 m and 235 m–277 m are shown on the diagram. Combining both intervals (180 m–277 m) the heat flow is 1.85±0.02 μcal/cm² sec.

properties of the drilling fluid and of the surrounding rock.

A mean value of C, for UCSD-5, was determined by plotting $\ln(t/t-s)$ against $\theta(t)$ for the temperature data logged on July 25, July 27, and August 4, 1964. $\theta(\infty)$ for each depth was evaluated from equation (1) using this value of C. The resulting equilibrium temperatures are plotted along with the logged temperatures on Figure 6. It is interesting to note that even though the temperature measurement at the 600-ft depth on August 4 was made within 12 days of the cessation of drilling, it is within 0.1°C of the equilibrium value. For the deeper depths the deviation from the equilibrium temperature is even less. This small difference between the observed and the equilibrium temperatures 12 days after drilling stopped is a direct result of the short time required for drilling. The last 122 m of rock, all solid granite, were drilled in less than four days.

A similar analysis was carried out on the temperature logs of UCSD-6. Drilling ceased on August 7, and the borehole was temperature logged on August 12 and 21. The mean value of C was determined and the equilibrium temperature for each depth computed as for UCSD-5. The change in temperature, at the 183 m depth, between the August 21 logging and the equilibrium temperature is less than 0.05°C. As 0.05°C is of the same magnitude as the error in the absolute temperature measurement and $\theta(t)-\theta(\infty)$ at deeper depths will be even less than this, equilibrium temperatures were not plotted for this station. The August 21 temperatures were used as if they were equilibrium values. The error, introduced by this assumption, is, of course,

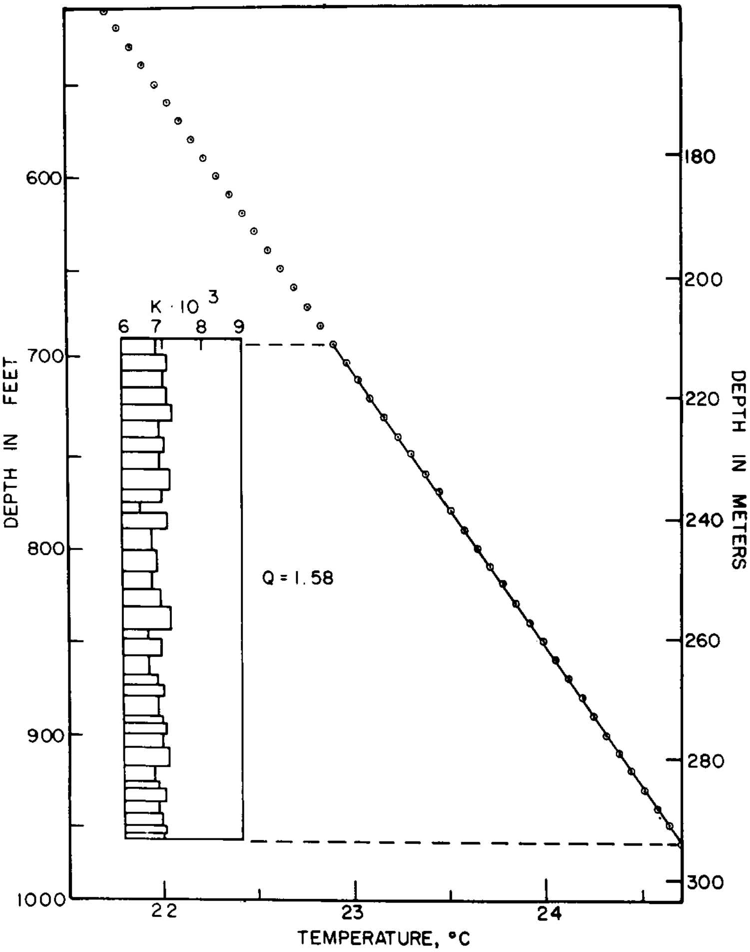

FIG. 5. Temperature and thermal conductivity data for UCSD-4.

systematic, but it is unlikely to decrease the actual heat flow value by more than two percent.

The validity of using equation (1) to determine $\theta(\infty)$ is dependent upon the thermal properties of the rock, the variation of source strength with depth, the diameter of the hole, and the proximity of the surface. Each of these effects is treated at length by Lachenbruch and Brewer (1959). In both UCSD-5 and UCSD-6 the rock is uniform and the conductivity does not change with depth. Below 170 m the drilling was continuous and the rate was relatively uniform, consequently the variation of source strength with depth is unlikely to be large. Because the diameter of the hole is small, 60 mm, it will have a negligible effect upon the linear source assumption of equation (1). The measurements used to determine temperature gradient were made below 152 m, consequently the effect of the surface can be ignored.

Other factors which might influence the temperature measurements are the stability of the water column, the effect of pressure on the thermistors, groundwater movement and the surrounding terrain. The first problem has been treated by Diment and Robertson (1963), who showed that for a diameter of less than 76 mm the effect of convective overturn in the drill hole can be ignored. The second problem was avoided by keeping the thermistors at atmospheric pressure for all measurements. The effect of groundwater movement is extremely important, but fortunately it only affected UCSD-2, the least reliable of the heat flow measurements.

UCSD-1 penetrated the shattered contact

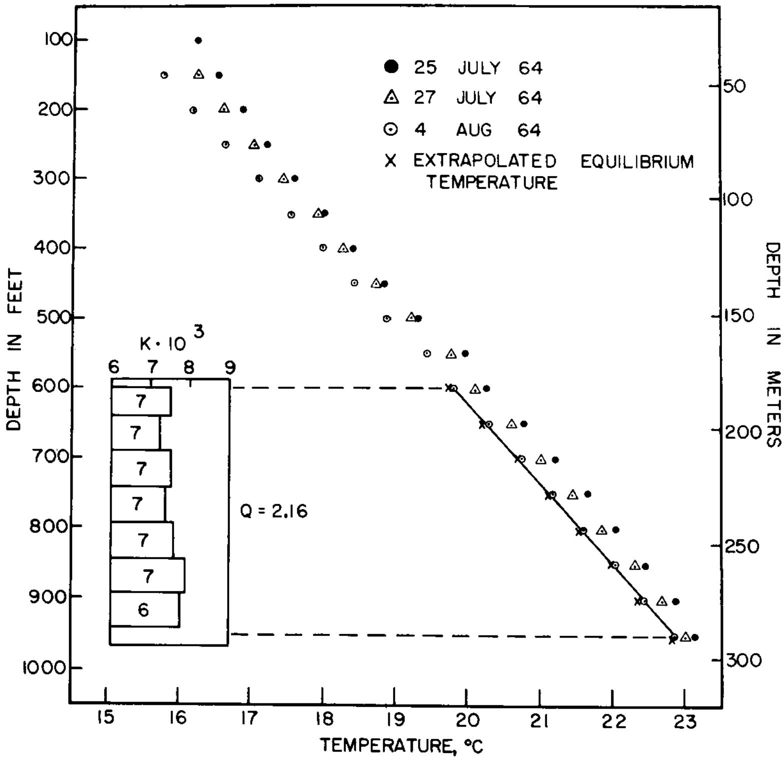

FIG. 6. Temperature and thermal conductivity data for UCSD-5. The numbers in the conductivity blocks are the number of samples that determined the mean conductivity of the interval.

zone between the intrusive monzonite and the Precambrian gneiss. In an effort to find more favorable conditions the rig was moved about 15 km, and UCSD-2 was drilled in Precambrian granite only to find circulating water below 115 m. Measurements in these holes are not considered as reliable as the others. The unreliability in UCSD-1 is caused by the difficulty in measuring the conductivity of the shattered monzonite stock and the shallowness of the hole. The heat flow obtained from UCSD-2 is questionable because circulating water was encountered below 115 m. However, temperatures measured above 115 m give a uniform gradient which extrapolates to a surface of 26.5°C, a nearly perfect agreement with the mean annual temperature of 26.4°C.

The effect of terrain upon the temperature gradient measurements is barely significant and then only for UCSD-6. All the other holes were drilled in subdued topography where the relief is less than ±60 m within 3 km of the drill hole. Figure 8 shows the surrounding relief and the position of UCSD-6. The hole was drilled in a valley between two east-west trending spurs. Consequently the lines of heat flux will concentrate preferentially in the valley and raise the effective heat flow. The effect of the topography upon a constant heat flow at depth, assuming uniform conductivity throughout the surface layers, was computed with a relaxation solution of Laplace's equation in two dimensions. Below 183 m the topographic correction reduces the gradient by slightly less than 4 percent. A three-dimensional correction was also calculated using the method described by Birch (1950). Assuming steady-state topography the correction for terrain within a radius of 5 km reduces the gradient by about 3 percent, which is in good agreement with the two-dimensional calculation. Extending the correction to a radius of 20 km reduces the correction to −2 percent. As the correction for the terrain (−2 percent) is in the opposite direction to the correction for equilibrium temperature (+2 percent), the net correction for the two

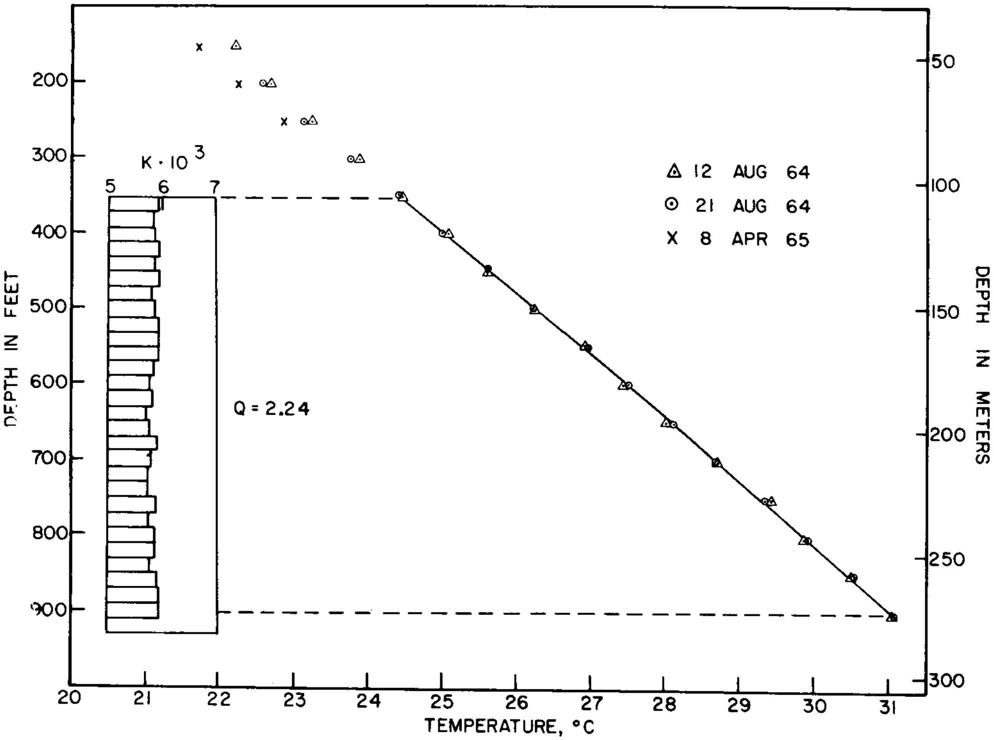

FIG. 7. Temperature and thermal conductivity data for UCSD-6.

effects is small and was neglected in determining the temperature gradient of this station.

INTERPRETATION OF HEAT FLOW OBSERVATIONS

The location and value of the six observations are shown on Figure 1. Values for four stations in West Texas (Herrin and Clark, 1956), 23 in the Basin and Range and Southern Rocky Mountain provinces (Roy et al, 1968b), and two in the Colorado Plateau (Spicer, 1964; Wright, 1966) are also presented on Figure 1. To simplify this figure, only the mean position and value ($\overline{2.07}$ μcal/cm^2 sec) of the four closely spaced stations just south of Tucson have been given. (The individual values are shown on Figure 11a.)

The mean and standard deviation of the values in the Basin and Range Province south of 36° N is 2.21 ± 0.43 μcal/cm^2 sec. This clearly indicates that the major heat flow feature in the southwestern United States is a high but variable flux in the Basin and Range Province which contrasts strongly with the uniform 1.1 μcal/cm^2 sec heat flow on the Texas Foreland. The considerable recent igneous activity, as evidenced by the large number of hot springs and wide extent of Tertiary extrusives in southern Arizona and New Mexico, is another indication of high temperatures at depth. There are many possible explanations of this contrast. The difference in heat flow between the two regions could be attributed to different concentrations of radioactive elements in the crust and upper mantle beneath the two provinces. The high heat flow anomaly might also be caused by a rise in the upper mantle isotherms beneath southern Arizona and New Mexico. Though, if this be the case, the sharp change of the heat flow anomaly between Roswell and Orogrande (UCSD-6) suggests that the major temperature changes are at quite shallow depths.

It is difficult to differentiate between an explanation based on a rise of the subcrustal isotherms and one assuming an excess concentration of radioactivity in the upper 50 km.

Preliminary measurements of the radioactivity of the surface rocks at heat flow boreholes in the Basin and Range Province by Roy et al (1968a) indicate that the heat flow Q may be related to the heat production A (pcal/cm^3 sec) by an

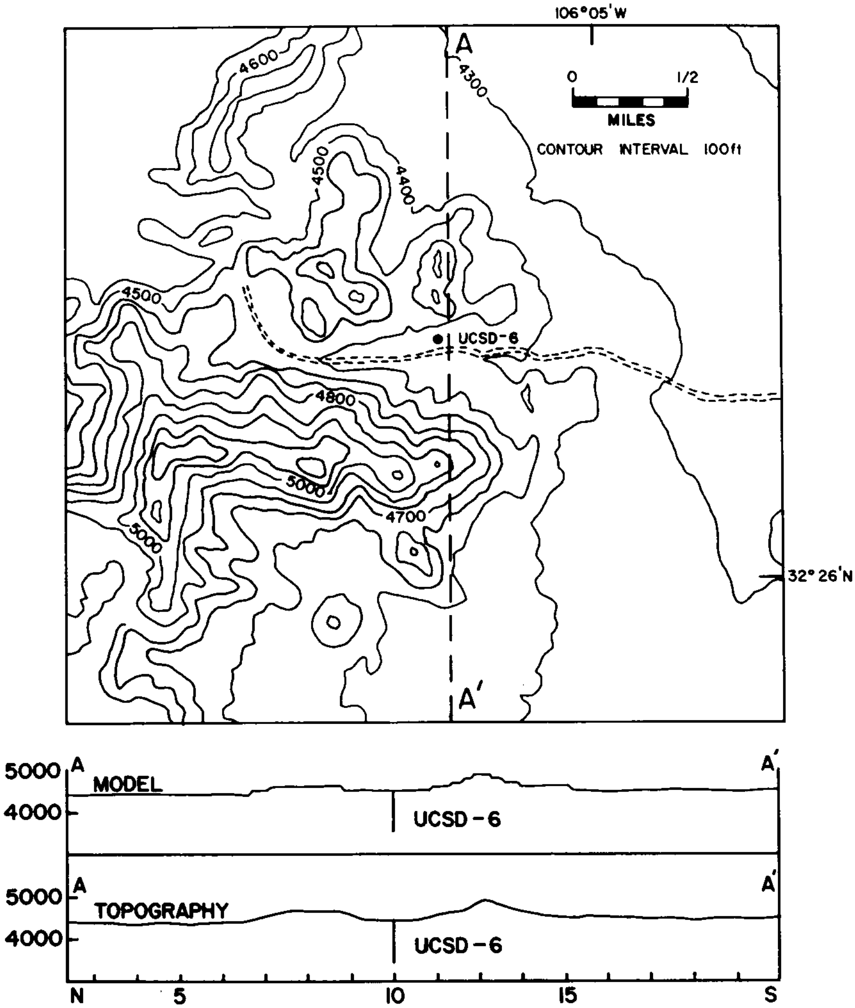

Fig. 8. Generalized topographic map showing location of UCSD-6, the profile A-A′, and the model used for relaxation solutions.

equation of the form

$$Q = a + bA. \tag{2}$$

This suggests that the fraction of heat a, from the lower crust and upper mantle, remains constant within the Province, but that the variable upper crustal radioactivity generates the variable surface heat flow. For the Basin and Range Province Roy et al (1968a) obtain a value of 8 km for b, the thickness of the crustal radioactive layer, and 1.5 μcal/cm^2 sec for a. As the average heat flow in southern Arizona and New Mexico is approximately 2.1 μcal/cm^2 sec, this suggests that an average heat flow due to heat production in the crustal rocks above 8 km is of approximately 0.6 μcal/cm^2 sec. This is only one-fourth of the total heat flow and indicates that the broad heat flow anomaly is not related to excess heat production in the upper crust. On the other hand, the high lower crust and upper mantle heat flow suggests an upper mantle source for the anomaly.

COMPARISON OF HEAT FLOW AND TRANSIENT GEOMAGNETIC FLUCTUATIONS

We shall now compare the depth of the isotherms inferred by downward continuation of the

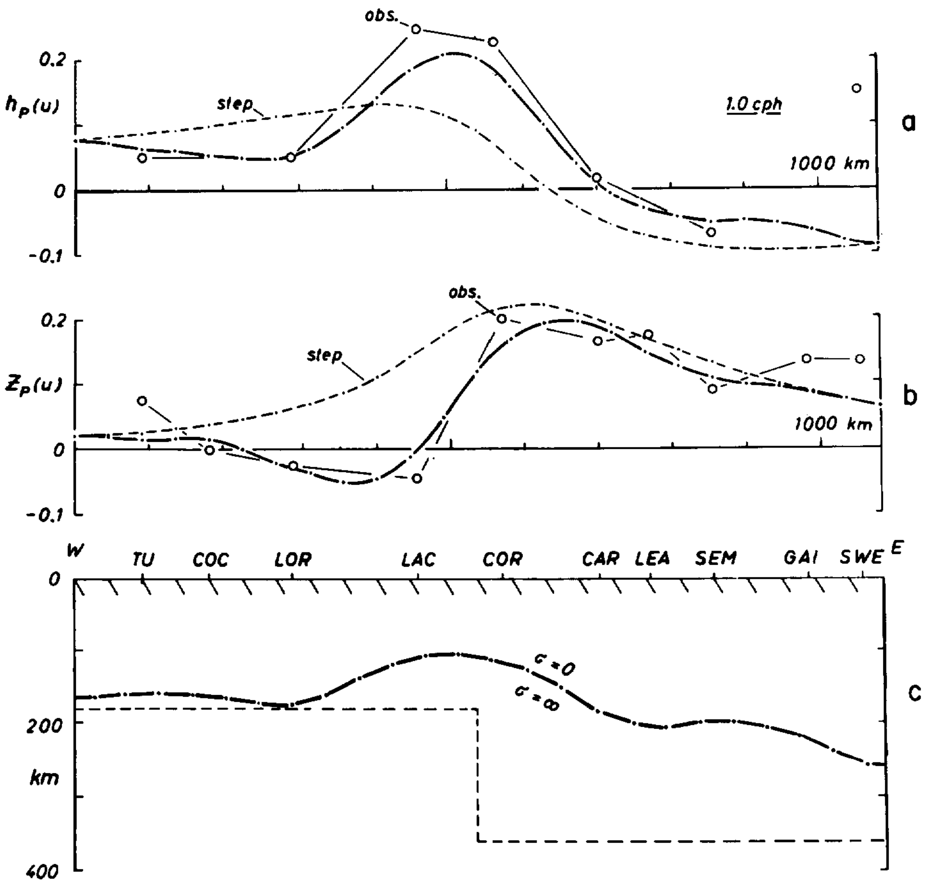

FIG. 9. The Rio Grande anomaly for 1 cph in-phase bays and its interpretation as a deep conductivity anomaly (Figure 48 from Schmucker, 1969). Figure 9a gives the anomalous horizontal variations h_P computed from the data and projected in the direction of the profile. Figure 9b presents the anomalous Z variations z_P computed from the data. The observed h_P and z_P profiles (open circles) are compared with profiles for the step and undulating surface models shown in Figure 9c.

surface heat flow with that determined from Schmucker's (1964) analysis of the transient geomagnetic fluctuations. Figure 9 [Figure 48 from Schmucker (1969)] presents the results of the analysis of magnetic records for bays (1 cph, in-phase) and their interpretation as a deep conductivity anomaly. Figure 9a gives the anomalous horizontal variations computed from the data and projected in the direction of the heat profile and Figure 9b the anomalous Z variation. The observed h_P and z_P profiles (open circles) are compared with computed profiles for a step model and an undulating surface model of a highly conductive mantle which are shown in the lower portion of Figure 9c. For bays with a period of one hour Schmucker suggests that the amplitude of the anomaly can be explained by a fictitious core of infinite conductivity at a depth of 160 km under southern Arizona and a depth of 320 km under West Texas. However, the depth of this core is strongly dependent upon the frequency of the bays. For instance, corresponding estimates for a period of two hours yield 250 km under southern Arizona and 400 km under Texas. The step model, however, does not explain the sharp increase of anomalous Z variations (Figure 9b) between Las Cruces (LAC) and Cornudas (COR); nor does it explain the rapid changes indicated by the length and direction of the arrows originating from these magnetic stations in Figure 1. The length of an arrow designates the magnitude of the anomalous Z variation of one

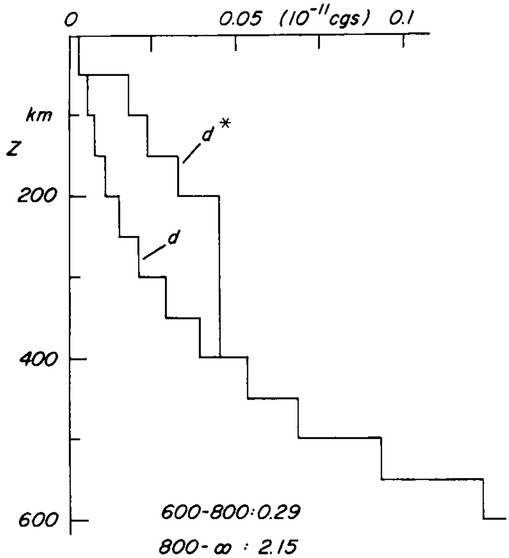

FIG. 10. Conductivity profiles for the earth's upper mantle according to model "d" from Lahiri and Price (1939) and a modified model "d*."

hour period, and its direction is normal to the lateral electrical conductivity change. Provided the ocean is far away, they point towards the region of higher electrical conductivity. The oppositely directed arrows at Las Cruces and Cornudas indicate a very narrow region of high conductivity between the two stations. The undulating model shown in Figure 9c takes this region into account and represents the surface of a highly conducting mantle that will explain the observed z_P and h_P anomalies. (See comparison of curves in Figures 9a and 9b.)

The actual increase of conductivity is not a single step from zero to infinity but rather a continuous increase with depth. Figure 10 (Schmucker 1969, Figure 49) presents a possible distribution that is compatible with the empirical h^* (depth to the infinitely conductive mantle) values as a function of frequency. The computed h^* values east of the Rio Grande are compatible with model "d" of Lahiri and Price (1939) while the h^* values to the west can be explained by a modified model labeled "d^*" on Figure 10. As the electrical conductivity of a dry rock is strongly dependent upon its temperature, it should be possible to infer from these distributions the increase of temperature with depth beneath both southern Arizona and Texas.

In the past few years a considerable number of measurements (Coster, 1949; Hughes, 1955; Bradley et al, 1962; Akimoto and Fujisawa, 1965; Hamilton, 1965) of the variation of electrical conductivity with pressure and temperature of a number of possible upper mantle minerals have been made. These measurements have indicated that the electrical conductivity of the silicates has a temperature dependence given by

$$\sigma = \sigma_0 e^{-E/k\theta}, \tag{3}$$

where σ_0 is a constant independent of temperature, E is an activation energy, and k is Boltzman's constant. Unfortunately, the inversion of mantle conductivity into ambient temperature depends critically upon the assumed composition and the postulated influence of pressure. For instance, Hamilton (1965) has shown that for olivine a ten-fold increase in conductivity can be produced by a 200°C increase in temperature, a 10 percent increase in Fayalite, or a 40 kilobar increase in pressure.

Assuming Ringwood's pyrolyte model (Ringwood, 1966) with an iron concentration of between 10 and 13 percent, we can extrapolate the electrical conductivity as a function of temperature from Figure 8 of Hamilton (1965). This gives an electrical conductivity between 0.01 and 0.03 mhos/m (or 0.01 and 0.03×10^{11} cgs units) for a temperature of 1200°C. By using these values for the electrical conductivity and applying them to Figure 10, the 1200°C isotherm would lie between 50 and 200 km in depth under southern Arizona whereas under Texas it would be between 150 and 400 km. To simplify the argument, the effect of pressure has been ignored.

In order to compare these isotherms with those inferred from the surface heat flow, the heat flow observations between Tucson, Arizona and Sweetwater, Texas were projected onto 32° N, the latitude of the geomagnetic profile. The projected heat flow profile is shown in Figure 11a. The curve through the observations is the assumed surface heat flow field. If the thermal conductivity is taken as a constant, the structure of the near-surface isotherms, assuming steady-state conditions, can be determined by extending the surface temperature gradient downwards.

In two dimensions, with x the horizontal coordinate and z positive downwards, steady-state conditions, and a constant thermal conductivity

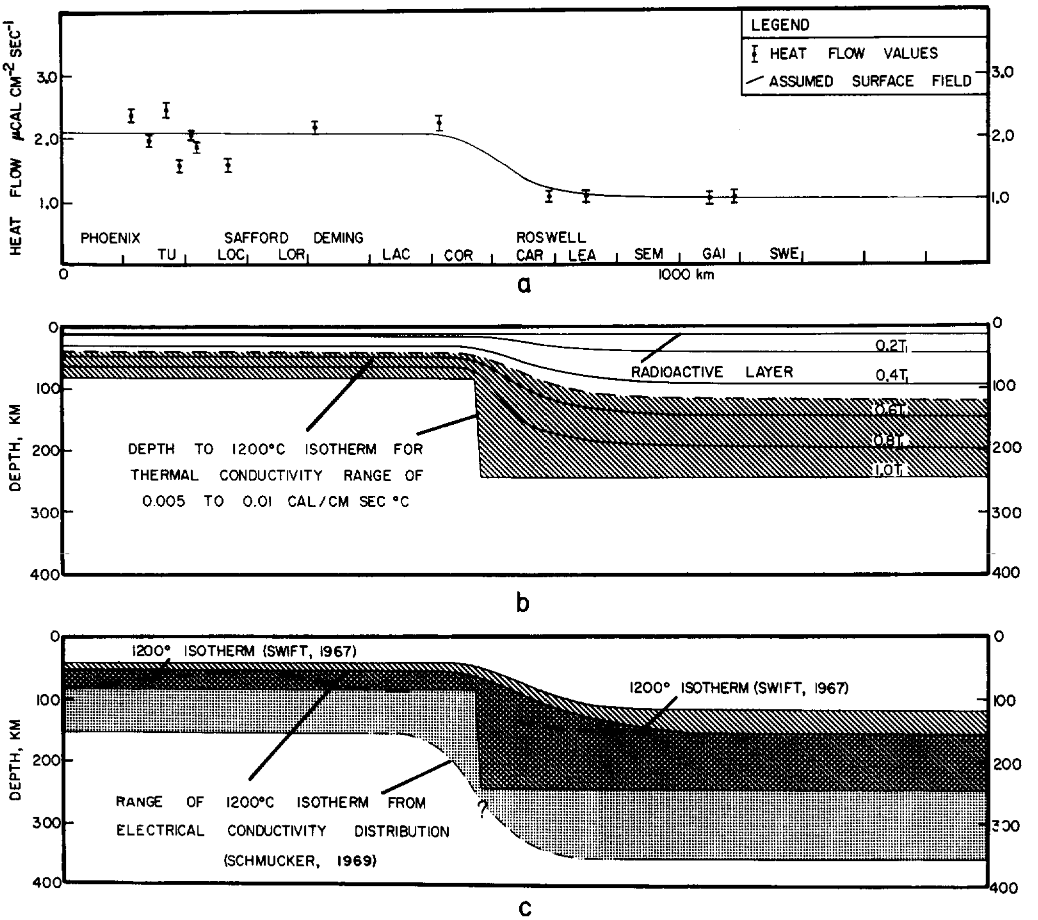

FIG. 11. (a) The surface heat-flow observation between Tucson, Arizona and Sweetwater, Texas. (b) Isotherms at depth assuming a heat production of 0.6 pcal/cm³ sec in a 10-km layer. The hatched region indicates the possible depth of the 1200° isotherm for a thermal conductivity range of 0.005 to 0.01 cal/cm sec °C. (c) Depth of the 1200° isotherm inferred from geomagnetic and magnetotelluric observations.

K, the temperature in the upper mantle is governed by Poisson's equation

$$\frac{\partial^2\theta}{\partial x^2} + \frac{\partial^2\theta}{\partial z^2} = \frac{A(x, z)}{K}, \tag{4}$$

where θ and A are the temperature and the heat production at the point (x, z). If the surface of the earth represents an isotherm, the surface heat flow can be extended downwards by setting up an isotherm at depth, assuming vertical flow at a large distance from the anomaly, and solving equation (4) between the boundaries. If l is the maximum depth of the lower boundary and the upper and lower temperatures are T_0 and T_1 respectively, substitution of

$$\theta = (T_1 - T_0)\theta' + T_0$$

and

$$A = A_1 A'$$

into (4) gives

$$\frac{\partial^2\theta'}{\partial x'^2} + \frac{\partial^2\theta'}{\partial z'^2} = B'A, \tag{5}$$

where B′ is given by

$$-\frac{l^2 A_1}{(T_1 - T_0)K}. \tag{6}$$

A relaxation method of solution was applied to equation (5). The standard two dimensional five-point difference approximation to Poisson's equation (Southwell, 1949) was used. A mesh size of 10 km by 10 km was taken and a width of 1500 km was assumed for the model. An initial temperature was given to each nodal point within the relaxation net, and equation (5) was solved by an iterative process. The isotherms were extended downwards by trial and error until a step function gave the observed surface temperature gradient. Because the standard difference approximation does not converge fast enough, the principle of over relaxation (Forsythe and Wasow, 1960) was used. The program was tested against the analytical solution of the problem of heat conduction through a uniform box with the potential along the bottom boundary a periodic function of x (Morse and Feshbach, 1953).

In the absence of any radioactive data for the Texas Foreland, a uniform heat production of 0.6 pcal/cm^3 sec (equivalent to 0.6 μcal/cm^2 sec heat flow from a 10 km thick layer) was taken throughout both regions. Figure 11b presents the distribution of the isotherms in nondimensional temperature units. On the assumption that $T_0=0$°C, T_1 would be 1200°C for a conductivity of 0.01 cal/cm sec °C, and 2400°C for a conductivity of 0.005 cal/cm sec °C.

Though the sharp change of the surface heat flow anomaly determines how far the surface gradient can be extended downwards, the probability of partial melting in the mantle rocks puts a further constraint on this depth. The formation of a partial melt would increase the effective thermal conductivity by a few orders of magnitude and would reduce the horizontal variation in the isotherms. On the assumption of Ringwood's (1966) pyrolite model for the upper mantle and a conductivity of 0.005 cal/cm sec °C, the basaltic fraction would start to melt at a temperature of 1500°C and a depth of 100 km. For a conductivity of 0.01 cal/cm sec °C it would start at a temperature of 1300°C and a depth of 50 km. In order to avoid the effect of partial melting, the 1200°C isotherm was chosen for the comparison between the surface heat flow and the geomagnetic data. The lined area in Figure 11b represents the possible range of the 1200°C isotherm inferred for the surface heat flow field on the assumption of a conductivity variation of 0.005 to 0.01 cal/cm sec°C.

For Figure 11c the dotted area which corresponds to the possible spread of the 1200°C isotherm inferred from the electrical conductivity distribution is superposed upon the lined region of Figure 11b. The regions overlap considerably and the sharpest variations in the isotherms occur at exactly the same place—between Las Cruces and Cornudas. It is obvious from this figure that the deep temperature structure inferred from the heat flow observations is consistent with the electrical conductivity distribution suggested by Schmucker (1964). This suggests that the rise of the upper mantle isotherms under southern Arizona and New Mexico may be the cause of both the transient magnetic anomaly and the high heat flow. It is difficult to be more specific than this as there are so many variables associated with the determination of both sets of isotherms.

OTHER GEOPHYSICAL EVIDENCE AND SPECULATIONS

The interpretation of the high heat flow values in terms of high subcrustal mantle temperatures in southwestern United States is supported by the magnetotelluric data of Swift (1967). The magnetotelluric stations are shown on Figure 1, and their projected position on the heat flow and magnetic profile are shown above the magnetic stations on Figure 11a. The thick dashed line on Figure 11c [taken from Figure 4.3 of Swift (1967)] represents the 1200°C isotherm assuming Ringwood's pyrolyte model for the upper mantle. This isotherm indicates a large horizontal temperature difference, at a depth of 50 km, between Safford and Roswell, the major change occurring between Deming and Roswell. This is entirely consistent with the horizontal temperature differences suggested by the geomagnetic anomaly but is about 100–200°C higher than that suggested by the heat flow observations. Considering the assumptions made in the derivation of both sets of isotherms this difference is not significant.

The postulated rise in the isotherms under the southwestern United States is supported by seismic observations of the velocity of compressional Pn waves in the mantle. Herrin and Taggart (1962) and more recently Herrin (1966) in a paper (Figure 2) by James and Steinhart (1967) have shown that variations in the subcrustal velocity of Pn were regionally significant. High velocities ranging from 8.3 to 8.1 km/sec are found under the Texas Foreland while values ranging from 7.8 to 8.0 km/sec are found under Arizona and western New Mexico. Though the accuracy of the spatial distribution of the apparent Pn velocities in the United States can be questioned, the observations do suggest that the subcrustal velocities under the Basin and Range Province are significantly lower than those under the Texas Foreland. Herrin and Taggart have suggested a rise of the isotherms in the subcrustal mantle to explain these lower velocities. Further confirmation for the proposed rise of the isotherms comes from a recently completed analysis by Woollard (1967) of the regional isostatic relations in the United States. He notes that negative isostatic anomalies characterize the Colorado plateau and the adjacent Basin and Range area and suggested that they could be interpreted as anomalously high temperatures at quite shallow depths.

If the heat flow anomaly in the Basin and Range Province is entirely from the upper mantle and the heat is transported only by conduction, the time dependence of the heat flow equation indicates that these high mantle temperatures may have existed since late Cretaceous and early Tertiary. This fact and the continuing igneous activity since the Cretaceous suggest that the different igneous and tectonic histories of the Basin and Range and the Texas Foreland may be directly related to large-scale temperature differences in the upper mantle.

Eardley (1962) and others have shown that the crustal tectonics of the whole region is dominated by the late Cenozoic fault system. Swift and Madden (1967) have suggested from present tensional forces, seismicity, and high temperatures that this fault system may be associated with an extension of the crest of the East Pacific Rise. On the other hand, if the San Andreas fault is a transform fault as proposed by Wilson (1965) and by the tectonic model of McKenzie and Parker (1967) it seems possible that the Cenozoic tectonic history of the western United States can be attributed to the continent overriding and partially reabsorbing the East Pacific Rise (Vine, 1966).

ACKNOWLEDGMENTS

This research was supported by the National Science Foundation under grants NSF GP 425 and NSF GA 715. We would like to thank Dr. R. P. Von Herzen for encouragement in the early phase of the project, and Mr. D. Ryan for help with the drilling and the detailed core logging. We would also like to thank Miss Uta Ritter for her assistance in reduction of the data, and Miss Linda R. Meinke and Miss Francie B. Hodge for drafting the figures.

REFERENCES

Akimoto, S., and Fujisawa, H., 1965, Demonstration of the electrical conductivity jump produced by the olivine-spinel transition: J. Geophys. Res., v. 70, p. 443–449.

Birch, F., 1950, Flow of heat in the Front Range, Colorado: Bull. Geol. Soc. Am., v. 61, p. 567–630.

Bradley, R. S., Jamil, A. K., and Munro, D. C., 1962, Electrical conductivity of fayalite and spinel: Nature, v. 193, p. 965.

Costain, J. K., and Wright, P. M., 1968, Heat flow and geothermal gradient measurements in Utah (abstract): Trans. Am. Geophys. Union, v. 49, p. 325.

Coster, H. P., 1949, The electrical conductivity of rocks at high temperatures: Mon. Nat. Astronom. Soc. Geophys. Suppl., v. 5, p. 193–199.

Diment, W. H., and Robertson, E. C., 1963, Temperature, thermal conductivity, and heat flow in a drilled hole near Oak Ridge, Tennessee: J. Geophys. Res., v. 68, p. 5035–5047.

Eardley, A. J., 1962, Structural geology of North America, second edition: New York, Harper and Row.

Forsythe, G. E., and Wasow, W. R., 1960, Finite difference methods for partial differential equations: New York, John Wiley and Sons, Inc.

Hamilton, R. M., 1965, Temperature variation at constant pressure of the electrical conductivity of periclase and olivine: J. Geophys. Res., v. 70, p. 5679–5691.

Herrin, E. T., 1966, Travel time anomalies and structure of the upper mantle (abstract): Trans. Am. Geophys. Union, v. 47, p. 44.

——— and Clark, S. P., 1956, Heat flow in west Texas and eastern New Mexico: Geophysics., v. 21, p. 1087–1099.

——— and Taggart, J., 1962, Regional variations in P_n velocity and their effect on the location of epicenters: Bull. Seis. Soc. Am., v. 52, p. 1037–1046.

Hughes, H., 1955, The pressure effect on the electrical conductivity of peridotite: J. Geophys. Res., v. 60, p. 187–191.

James, D. E., and Steinhart, J. S., 1967, Structure beneath the continents: A critical review of explosion studies 1960–1965, *in* The earth beneath the continents, edited by J. S. Steinhart and T. Jefferson Smith: Geophys. Monogr. 10, Am. Geophys. Union, p. 293–333.

Lachenbruch, A. H., and Brewer, M. C., 1959, Dissipation of the temperature effect of drilling a well in arctic Alaska: U. S. Geol. Surv. Bull., no. 1983-C, p. 73–109.

Lahiri, B. N., and Price, A. T., 1939, Electromagnetic induction in a non-uniform conductor and the determination of the conductivity of the earth from terrestrial magnetic variations: Phil. Trans. Roy. Soc. London, A, v. 237, p. 509–540.

McKenzie, D. P., and Parker, R. L., 1967, The north Pacific: An example of tectonics on a sphere: Nature, v. 216, p. 1276–1280.

Morse, P. M., and Feshbach, N., 1953, Methods of theoretical physics: New York, McGraw-Hill Book Co., Inc.

Ratcliffe, E. H., 1959, Thermal conductivities of fused and crystalline quartz: Brit. J. Appl. Phys., v. 10, p. 22–25.

Ringwood, A. E., 1966, Mineralogy of the mantle, *in* Advances in earth science, edited by P. M. Hurley, Cambridge, M. I. T. Press, p. 357–399.

Roy, R. F., Blackwell, D. D., Decker, E. R., and Birch, F., 1968a, Heat generation of surface rocks and heat flow provinces (abstract): Trans. Am. Geophys. Union, v. 49, p. 319.

Roy, R. F., Decker, E. R., Blackwell, D. D., and Birch, F., 1968b, Heat flow in the United States: J. Geophys. Res., v. 73, p. 5207–5221.

Schmucker, U., 1964, Anomalies of geomagnetic variations in the southwestern United States: J. Geomagn. and Geolectr., v. 15, p. 193–221.

——— 1969, Anomalies of geomagnetic variations in the southwestern United States, 1959–1962: in preparation.

Spicer, H. C., 1964, Geothermal gradients and heat flow in the Salt Valley anticline, Utah: Bull. Geophys. Teorica Appl., v. 6, p. 263–282.

Southwell, R. V., 1949, Relaxation methods in theoretical physics: Oxford, Clarendon Press.

Swift, C. M., 1967, A magnetotelluric investigation of an electrical conductivity anomaly in the southwestern United States: Ph.D. thesis, M.I.T.

——— and Madden, T., 1967, A magnetotelluric investigation of the electrical conductivity anomaly in the upper mantle in the southwestern United States (abstract): Trans. Am. Geophys. Union, v. 48, p. 210.

Vine, F. J., 1966, Spreading of the ocean floor: New evidence: Science, v. 154, p. 1305–1415.

Von Herzen, R. P., and Maxwell, A. E., 1959, The measurement of thermal conductivity of deep sea sediments by a needle-probe method: J. Geophys. Res., v. 64, p. 1557–1563.

Wilson, J. T., 1965, A new class of faults and their bearing upon continental drift: Nature, v. 207, p. 343–347.

Woollard, G. P., 1967, Regional isostatic relations in the United States, *in* The earth beneath the continents, edited by J. S. Steinhart and T. Jefferson Smith: Geophys. Mon. 10, Am. Geophys. Union, p. 293–333.

Wright, P. M., 1966, Geothermal gradient and regional heat flow in Utah: Ph.D. thesis, University of Utah.