JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 93, NO. B11, PAGES 12,995–13,004, NOVEMBER 10, 1988

Thermal Regime of the State 2-14 Well, Salton Sea Scientific Drilling Project

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Temperature logs were made repeatedly during breaks in drilling and both during and after flow tests in the Salton Sea Scientific Drilling Project well (State 2-14). The purpose of these logs was to assist in identifying zones of fluid loss or grain and to characterize reservoir temperatures. At the conclusion of the active phase of the project, a series of logs was begun in an attempt to establish the equilibrium temperature profile. Initially, we were able to log to depths below 3 km, but beginning in late May of 1986, it was impossible to log below about 1.8 km owing to casing failure. Our best estimates of formation temperature below 1.8 km are 305° ± 5°C at 1890 m and 355° ± 10°C at 3170 m. For the upper 1.8 km the latest temperature log (October 24, 1986), using a digital "slickline" (heat-shielded downhole recording) device, was within a few degrees Celsius of equilibrium, as confirmed by a more recent log (July 31, 1987) to a depth of ~1 km. As in most other wells in the Salton Sea geothermal field, there is an impermeable, thermally conductive "cap" on the hydrothermal system; this cap extends to a depth of more than 900 m at the State 2-14 well. Thermal conductivities of 19 samples of drill cuttings from this interval were measured at room temperature. The conductivity values were corrected for in situ porosity as determined from geophysical logs and for the effects of elevated temperature. Thermal gradients decrease from about 250 mK m^{-1} (same as degrees Celsius per kilometer) in the upper few hundred meters to just below 200 mK m^{-1} near the base of the conductive cap. Using one interpretation, thermal conductivities increase with depth (mainly because of decreasing porosity), resulting in component heat flows that agree reasonably well with the mean of about 450 mW m^{-2}. This value agrees well with heat flow data from shallow wells within the Salton Sea geothermal field. A second interpretation, in which measured temperature coefficients of quartz- and carbonate-rich rocks are used to correct thermal conductivity, results in lower mean conductivities that are roughly constant with depth and, consequently, systematically decreasing heat flux averaging about 350 mW m^{-2} below 300 m. This interpretation is consistent with the inference (from fluid inclusion studies) that the rocks in this part of the field were once several tens of degrees Celsius hotter than they are now. The age of this possible disturbance is estimated at a few thousand years.

Introduction

A major goal of the Salton Sea Scientific Drilling Project was the acquisition of a complete set of wire line geophysical logs. To this end, a standard set of commercial logs was obtained (limited by the temperature capabilities of the tools) and both complementary and redundant logs were obtained by the Water Resources Division of the U.S. Geological Survey (USGS) using tools that were designed to operate at higher temperatures than conventional, commercial tools [*Paillet*, 1986]. A description and analysis of both sets of logs are the subject of a companion paper [*Paillet and Morin*, this issue]. The characterization of formation and reservoir temperatures at the site and of the thermal regime of the surrounding region was of particular importance. Regional heat flow and its tectonic implications were discussed by *Lachenbruch et al.* [1985], and models for the Salton Sea hydrothermal system have been developed by *Younker et al.* [1982] and by *Kasameyer et al.* [1984, 1985]. The thermal regime of the unconsolidated sediments of the Imperial Valley as a whole (Figure 1) was presented by *Sass et al.* [1984]. Recently, *Newmark et al.* [1986, this issue] have conducted a thorough study of the shallow (~100 m) heat flow from the Salton Sea geothermal field (Figure 1).

Temperature logs were obtained during the active drilling and testing phase to identify fluid loss zones, to predict equilibrium formation temperatures, and (during flow tests) to provide information required for characterization of the thermodynamics of the flow zone. These temperatures were measured with a platinum resistance thermometer (RTD) attached to four conductors of a logging cable, or with modified oil field downhole recording devices ("Kuster" gauges) which employ mechanical transducers (bimetallic strips). The latter devices are less accurate and have a much longer time constant (minutes as opposed to seconds) than the RTD, but with the recorder section in a dewar, are capable of withstanding much higher temperatures (400°C as opposed to 250°–300°C) than the RTD with surface readout. A selection of temperature logs obtained during the drilling phase of the project is shown as Figure 2 of *Paillet and Morin* [this issue].

The dewared mechanical temperature tools were available as backup and as an independent check for temperature measurements postdrilling; however, all of the temperature

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Paper number 7B7141.
0148-0227/88/007B-7141$05.00

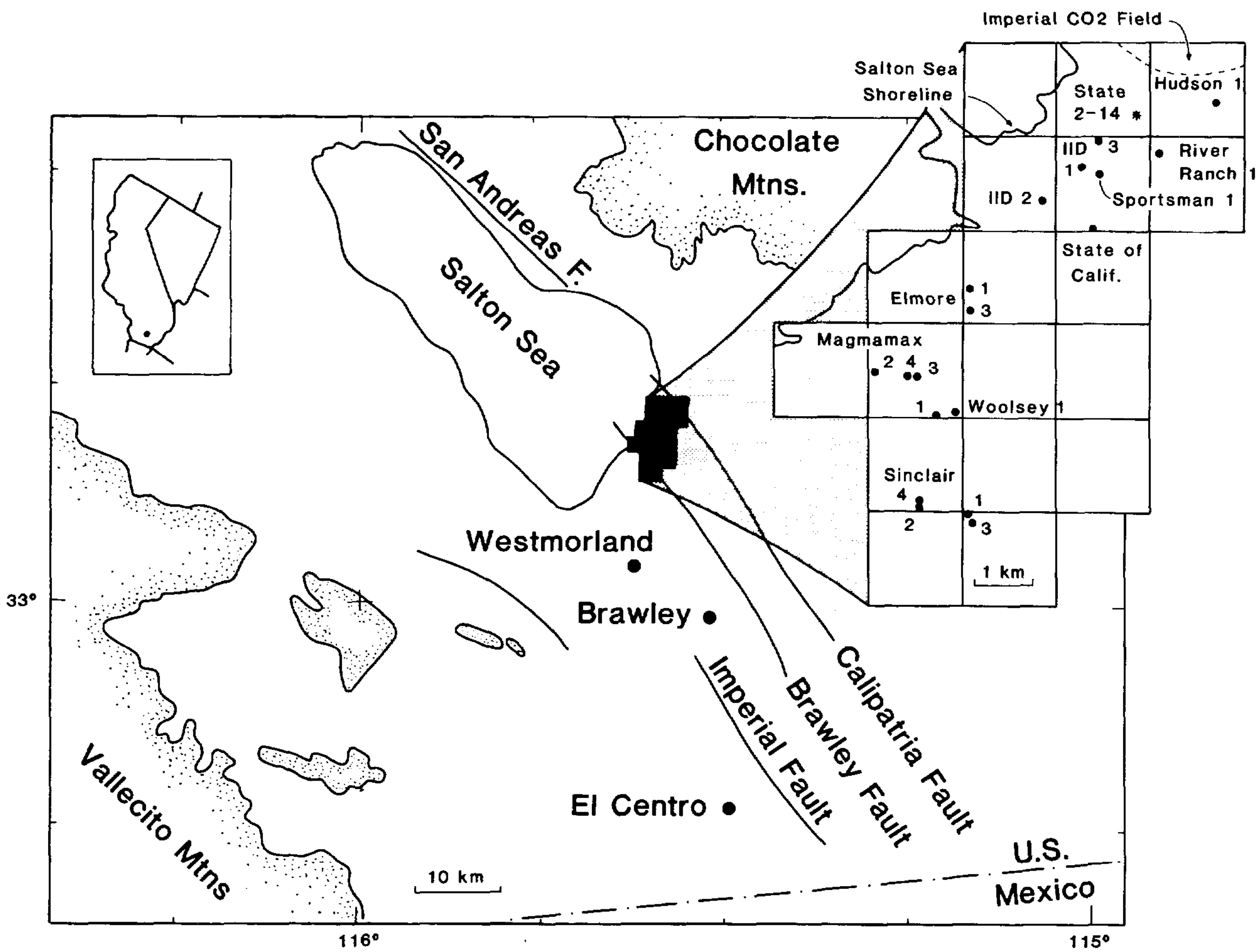

Fig. 1. Map of the Imperial Valley, adjoining ranges, and major faults [after *Younker et al.*, 1982]. Inset shows location of State 2-14 and other wells in the Salton Sea geothermal field.

data presented here were obtained with platinum RTDs deployed either in a conventional mode (surface readout via four electrical conductors) or in a downhole recording digital "memory tool" either of which allows a much higher density of data than do the conventional tools with mechanical transducers. The memory tool is described briefly in the following section. A time series of temperature logs is then presented together with an analysis of equilibrium temperatures, some measurements of thermal conductivity from the upper 1 km of the well, and some estimates of conductive heat flux.

Digital Temperature-Pressure Tool

The downhole memory tool, which measures both the temperature and pressure of the well bore fluid, was developed to provide a smaller and more accurate alternative to the electromechanical tools. This tool is a self-contained instrument operating on a "slickline"; that is, there is no direct communication of the tool with the surface. The temperature and pressure measurements are recorded in internal digital memory. The electronic circuitry is protected from the hot well bore fluid temperatures by a vacuum heat shield or "dewar." The tool was built to Sandia National Laboratories' specifications by Service Systems Engineering of Mansfield, Texas (now Madden Systems, Inc., of Odessa, Texas). Tool specifications are summarized in Table 1 and a sketch of the tool is presented as Figure 2.

The temperature is measured with a platinum resistance temperature detector (RTD) mounted inside a tube that extends into the well bore fluid. The resistance of the RTD is converted to frequency by a voltage-controlled oscillator. Pressure is measured by a quartz crystal transducer mounted inside the dewar and uses a stainless steel capillary tube to transmit the pressure from the well bore fluid. The output of the transducer is a frequency. Both frequency components are then stored in the internal memory. The heat shield is an integral pressure housing and vacuum dewar built by PDA Engineering of Santa Ana, California. The pressure housing was built from corrosion resistant Inconel 718. This heat

TABLE 1. Summary of Specifications for the Downhole Memory Tool

	Value
Diameter	89 mm
Mass	45 kg
Length	1.5 m
Temperature range	0°–600°C
Accuracy	2°C
Sensitivity	0.02°C
Pressure range	0–15,000 psi (0–100 MPa)
Accuracy	3 psi (20 kPa)

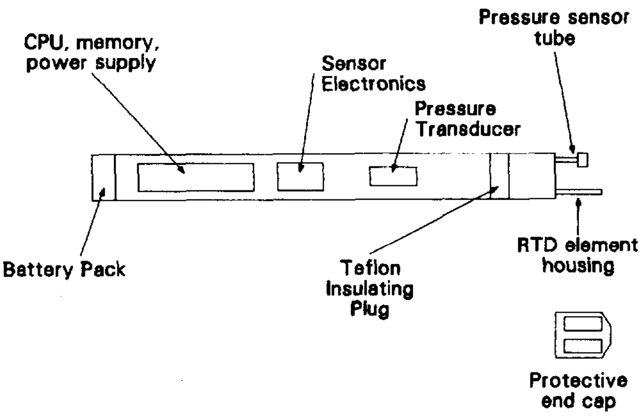

Fig. 2. Diagram of the downhole memory tool showing the arrangement of the sensor head, electronic circuitry, and battery power. The Inconel pressure vessel is not shown.

shield will allow the tool to operate for 10 hours at an ambient temperature of 400°C.

The data acquisition in the tool is controlled by an RCA 1802 microprocessor. Prior to a logging run, the microprocessor is programmed at the surface, using a personal computer, for up to 20 fixed-time interval readings. For example, the first reading may be delayed 20 min for tool and wireline makeup. Then 800 readings at 15-s intervals may be obtained as the tool is lowered in the well bore. The on-board memory consists of 8 kbyte of RAM which allows for the storage of up to 1000 temperature and pressure readings. When the tool is returned to the surface, the data are transferred to a personal computer via an RS-232 interface for further processing.

The design specifications outlined in Table 1 were essentially those used in soliciting bids from potential suppliers. As such, they constituted minimum standards by which to judge contractor performance. From internal consistency of repeated calibrations and repeated log segments, we are confident that the accuracy of the temperature portion of the tool is significantly better than the 2°C design specifications, more of the order of a few tenths of a degree Celsius.

TEMPERATURE DATA

The memory tool was used on four occasions between April 8 and October 24, 1986, to obtain temperature profiles in State 2-14. A fifth attempt on July 31 to August 1, 1987, was unsuccessful owing to failure of the metal-to-metal seal between dewar and instrument housing. These logs are shown in Figure 3, together with an earlier log obtained using the USGS logging unit and a profile labeled "Equil. Temps." A profile obtained with the USGS logging unit on July 31, 1987, is discussed later.

The log of April 1, 1986, was begun immediately upon cessation of the injection of nearly 10^4 m^3 of cooled brine (which had earlier been produced during a flow test) and irrigation water. There are several striking features in this profile (see also log of March 27, 1986, from *Paillet and Morin* [this issue]). There is a pronounced temperature reversal just below

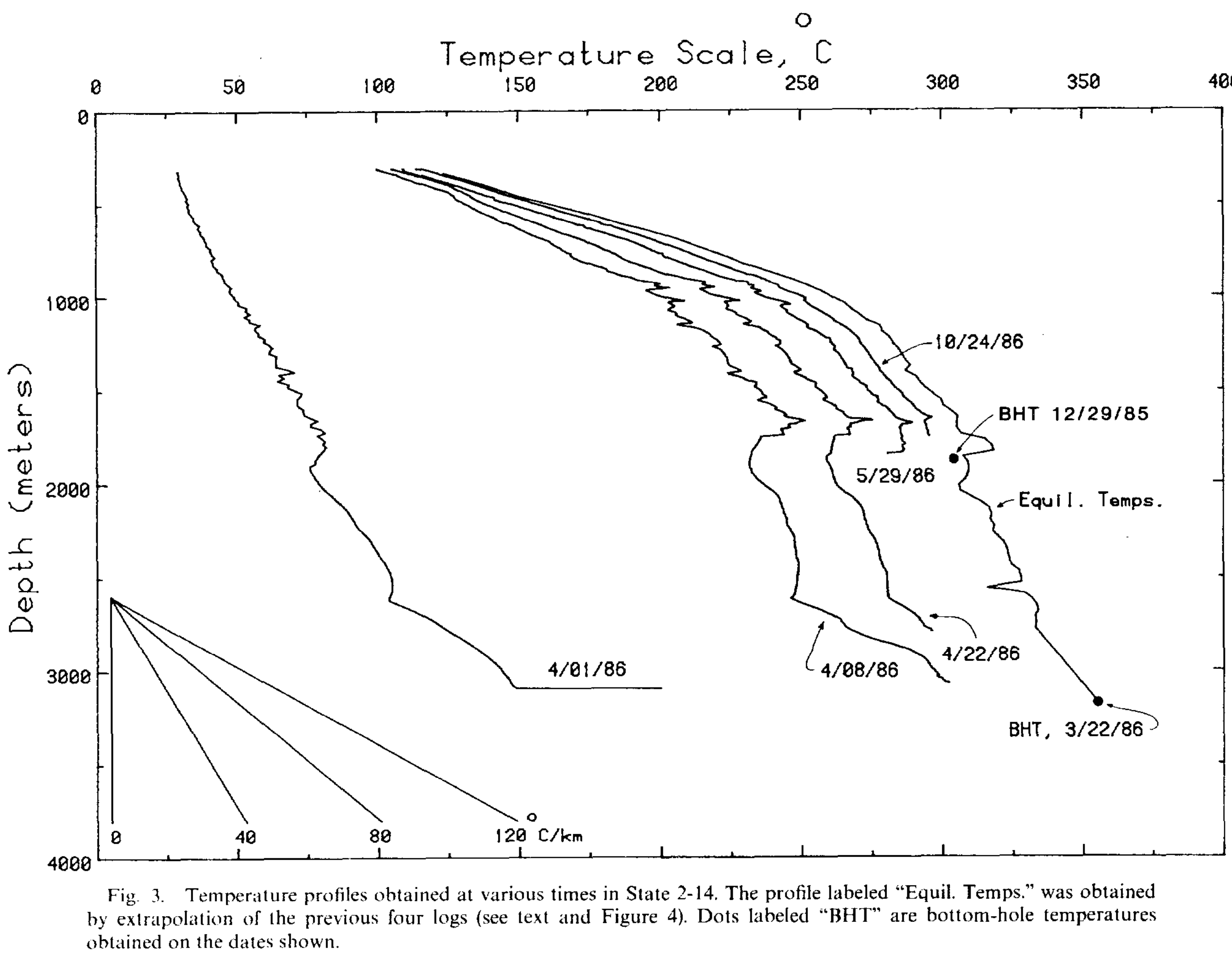

Fig. 3. Temperature profiles obtained at various times in State 2-14. The profile labeled "Equil. Temps." was obtained by extrapolation of the previous four logs (see text and Figure 4). Dots labeled "BHT" are bottom-hole temperatures obtained on the dates shown.

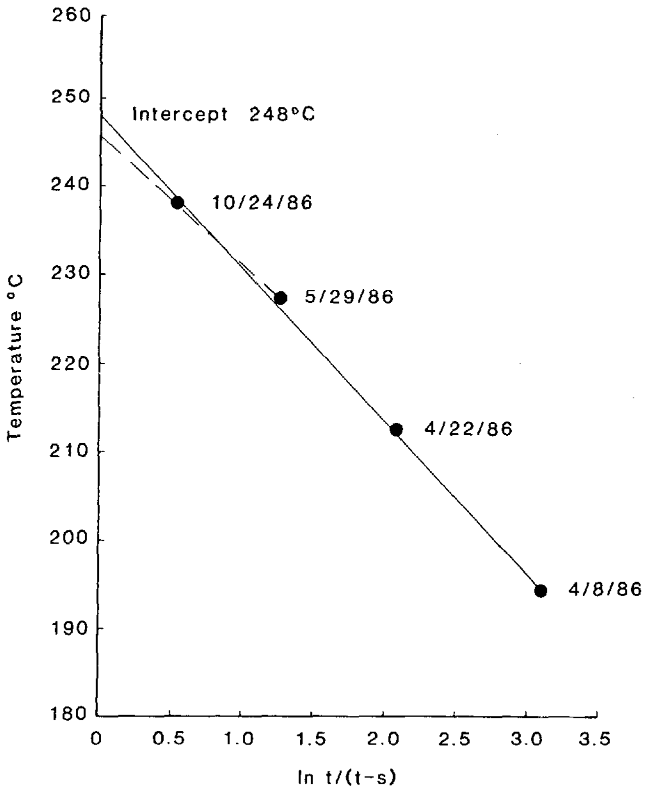

Fig. 4. Extrapolation to infinite time of temperatures measured on the dates shown at a depth of 914 m in State 2-14.

the $9\frac{5}{8}$-inch (244 mm) outside-diameter casing at a depth of ~1830 m. This was the depth of the first production zone tested in December 1985, and it evidently accepted a large amount of injected fluid. There was recurrent loss of fluid while drilling between this depth and ~2800 m. The strong reversal at about 2620 m (Figure 3) indicates a very permeable zone there also. The casing configuration after drilling to total depth included a 7-inch (178 mm) o.d. liner hung (but not grouted) from a depth of 1725 m to a total depth of 3093 m [see *Sass and Elders*, 1986, Figure 3]. At this depth there is a sharp increase in temperature in the April 1 log, indicating that (1) very little, if any, fluid was injected below the liner, and (2) most, if not all, of the fluid injected flowed around the bottom of the liner, up the annulus and entered the formation at various depths, mostly around 1850 and 2620 m. Profiles on subsequent dates retain the memory of the massive injection below 1830 m and show a conductively decaying spike at about 1670 m, most likely the result of injection of a large quantity of cement behind the $9\frac{5}{8}$-inch casing at that depth. This spike and another one at about 1 km depth, are misaligned by about 15 m in the October 24 profile, indicating some slippage of our depth counter at shallower depths on that date. The temperature data support the hypothesis that most of the fluid produced in the March 1986 flow test was coming from behind the liner, not from the expected flow zone between 3093 and 3220 m.

Disturbances to the thermal regime caused by circulation of drilling fluids persist for several drilling periods [*Bullard*, 1947; *Lachenbruch and Brewer*, 1959; *Jaeger*, 1956, 1961]; however, when the drilling history is relatively simple and no loss of drilling fluid occurs, it is a relatively straightforward matter to predict equilibrium temperatures. Also, equilibrium thermal gradients are approached within a few percent much more rapidly than are equilibrium temperatures [*Lachenbruch and Brewer*, 1959; *Jaeger* 1961]. In the case of State 2-14, the drilling-testing history is very complex. During the 160-day active phase of site operations [see *Sass and Elders*, 1986, Figure 2; *Elders and Sass*, this issue, Figure 5] there were two flow tests; we experienced recurring episodes of lost circulation, during which times both mud and cement were introduced at various levels (mostly below 1800 m); and we twice deliberately disposed of about 10^4 m^3 of cold fluid during the periods December 28–29, 1985, and March 26-31, 1986. The task of measuring or predicting equilibrium temperatures was further complicated by failure of the liner which limited access to a depth of ~1.8 km from May 1986 onward.

Notwithstanding the difficulties mentioned above, we were able to estimate equilibrium temperatures in the upper 1.8 km with some confidence, using the four logs that reached that depth. Following *Lachenbruch and Brewer* [1959], we plotted measured temperature versus the quantity ln $t/(t - s)$, where t is the time, in days, since the drill bit first reached a given depth and s is the duration of the drilling process at that depth. An example from a depth of 914 m is given in Figure 4. In oil field parlance this is a "Horner" plot. Extrapolation to ln $t/(t - s) = 0$ (infinite time) yields the equilibrium temperature. The example shown in Figure 4 is typical of this data set in that while the four points define an excellent straight line ($R^2 = 0.998$), extrapolation of the two most recent logs (dashed line, Figure 4) results in an estimate some 2° or 3°C lower than the least squares line for all four points. In Figure 5 a profile obtained on July 31, 1987, to 980 m using the USGS Water Resources Division's logging unit is juxtaposed against the October 24, 1986, log and the extrapolated "equilibrium" temperatures. There is reasonable agreement between 300 and 500 m, but the most recent log diverges from the "equilibrium" temperatures below that depth. It is likely that the divergence is due, at least in part, to the tendency of the slope of the temperature-log time line to decrease as the abscissa approaches zero. This tendency can, in turn, be attributed to the departures from theoretical assumptions of the source term for the temperature disturbances. We thus conclude that our extrapolated equilibrium temperatures are most likely overestimated but are probably reasonable upper limits to a depth of about 1800 m (Figure 3) where there is a proboscislike computational artifact. Below that depth, the extrapolations are controlled by only the two early logs and also reflect the large amounts of fluid lost to the formation. Even so, they are in reasonable accord with the two bottom-hole temperatures obtained during the flow tests in December 1985 and March 1986 using a platinum resistance transducer (RTD) and the heat-shielded Kuster tool, respectively (Figure 3).

From the logs of the upper portion of State 2-14 (Figure 5) we may make some additional observations. From the October 24, 1986, logs a measure of the thermal inertia of the "bullnose" (protective metal housing) and other metal immediately adjacent to the temperature transducer can be obtained. The RTD tracks temperature to within ~0.1°C while moving, but when the tool is stopped (to check synchronization with the surface depth-time log), a complex interaction between RTD and sorrounding metal takes place. For all logs the logging speed was 0.15 m s^{-1}, and the duration of each stop was 5 min. When the tool began moving again, there was an inertial lag of a few tenths of a degree Celsius which took

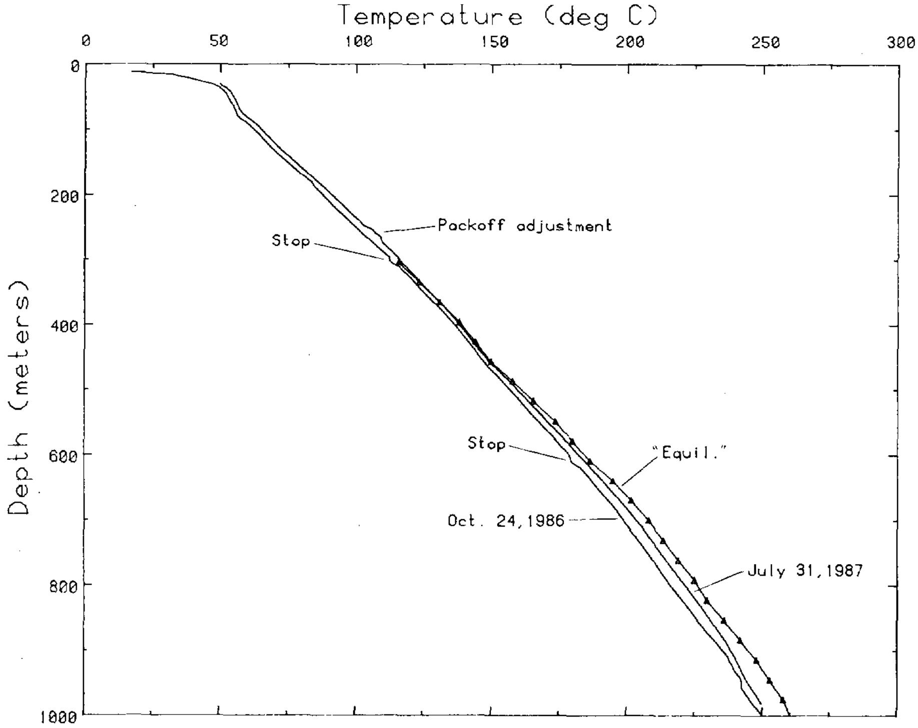

Fig. 5. Temperature profiles from the upper kilometer of State 2-14. Curve labeled "Equil." was calculated from the four logs between April 4 and October 24 (see text).

about a minute (~ 10 m) to stabilize, hence the slight dip in the temperature profile. A redesign of the bullnose in which its diameter and weight is reduced would alleviate this problem. The profile of July 31, 1987, exhibits some irregularities at about 250 m. The well was under pressure of 300 psi (~2 MPa). During the early part of the log we were adjusting the hydraulic pack-off to allow free movement of the logging line with a minimum of leakage. At about 250 m the cable stopped moving owing to excess pressure on the pack-off. While the slack in the line was being taken up, the tool dropped suddenly and several liters of fluid leaked past the pack-off causing the disturbance observed. Apart from this "glitch" we consider the profile of July 31, 1987, to be the best estimate of formation temperatures in the upper 1 km of the well, and that data set is used in our estimates of conductive heat flux.

Representative temperature profiles from the Salton Sea geothermal field (Figure 1) are shown for comparison with the "equilibrium" temperatures from State 2-14 in Figure 6. Indi-

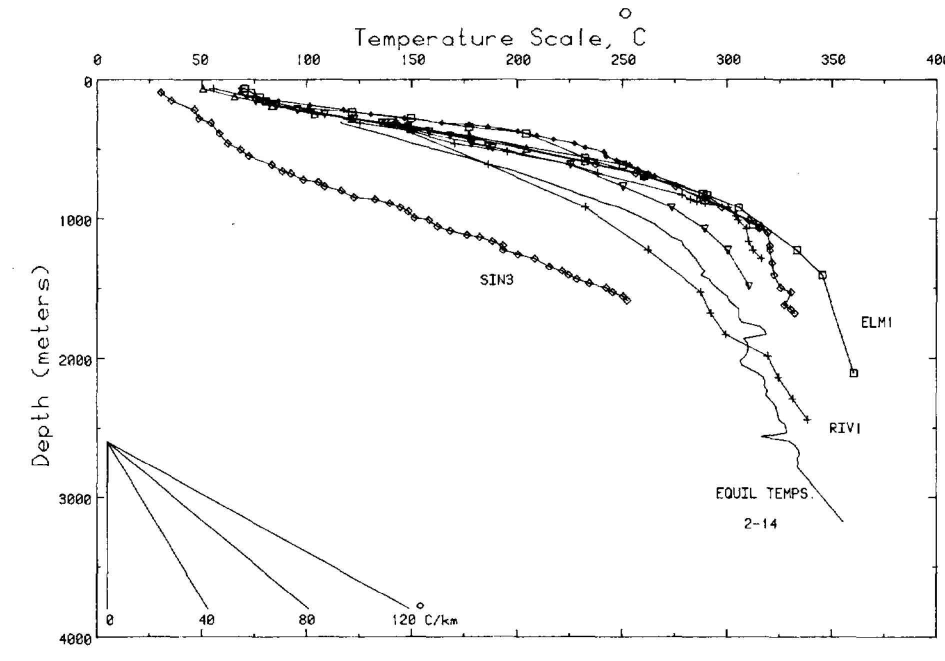

Fig. 6. Representative temperature profiles from the Salton Sea geothermal field (Figure 1) and calculated equilibrium temperatures for State 2-14.

TABLE 2. Thermal Conductivity of Drill Cuttings From State 2-14

Depth, m	Lithology	λ_s*	ϕ,†%	λ_{corr}‡ a	λ_{corr}‡ b
306	claystone	2.62	40	1.53	1.45
373	siltstone	3.80	45	1.77	1.56
401	claystone	3.51	28	2.22	1.88
437	sandstone	4.31	35	2.27	1.93
474	claystone	2.29	30	1.60	1.33
510	sandstone	3.90	20	2.75	2.21
529	claystone	3.74	23	2.53	2.04
556	claystone	3.45	21	2.45	1.95
584	sandstone	3.26	20	2.38	1.87
625	claystone	2.99	19	2.25	1.74
643	claystone	3.36	17	2.56	1.98
674	claystone	2.77	17	2.18	2.65
710	siltstone	3.11	15	2.47	1.84
735	sandstone	3.38	15	2.65	1.96
765	claystone	3.40	13	2.75	2.01
796	sandstone	2.97	22	2.13	1.59
828	siltstone	3.07	30	1.93	1.48
895	sandstone	3.24	17	2.46	1.77
917	sandstone	3.16	15	2.48	1.75

Thermal Conductivity, W m^{-1} K^{-1}

*Thermal conductivity of the solid component of drill cuttings at ~25°C.

†Percentage porosity, based on borehole-compensated density log, assuming a grain density of 2.65 g cm^{-3} (see Figure 7).

‡Corrected thermal conductivity: a, corrected for porosity only; b, corrected porosity and temperature. The temperature coefficient used is that for carbonate- or quartz-rich rocks [*Birch and Clark*, 1940].

vidual profiles were presented by *Palmer* [1975] and were analyzed in some detail by *Younker et al.* [1982]. With the exception of Sinclair 3 (SIN3) which is near the southern edge of the field, temperature profiles have a characteristic shape, with a linear or gently curved high-gradient upper portion a few hundred meters to 1 km thick, below which the temperature gradient drops off sharply, indicating nonconductive heat transfer. Elmore 1 (ELM1) is the hottest well in the field for which data are available, and there are some many others hotter than State 2-14 at comparable depths. The profile from State 2-14 most resembles that from River Ranch 1 (RIV1) which is located about a kilometer to the southeast (Figure 1). A comparison of the widely spaced measurements (obtained using conventional mechanical transducers) of Figure 6 with the profiles of Figure 3 emphasizes the advantage of being able to resolve details of the thermal profile using the dewared, downhole-recording electronic tool.

THERMAL CONDUCTIVITY

In common with many other wells in the Salton Sea geothermal field (Figure 6) the temperature profile from State 2-14 appears to reflect mainly conductive heat transfer to a depth of about 900 m (see also Figures 3 and 5); below that depth, there is a sharp drop in temperature-gradient, suggesting that convection dominates the heat transfer process. As a check on the validity of the inference of conductive heat flux and in an attempt to estimate a heat flow value, we sampled drill cuttings at ~30-m intervals between 300 and 900 m (no cutting samples were available above 300 m). Grain thermal conductivities were determined at a temperature of about 25°C using the "chip" method of *Sass et al.* [1971].

The grain conductivities do not vary systematically with depth (Table 2, column 3), despite a ~20% decrease in gradient over the same interval (Figure 7). To a depth of nearly 800 m, the decrease in gradient is matched by a decrease in porosity (Table 2, column 4, and Figure 7). Assuming little or no variation of the solid component of thermal conductivity with temperature (assumption a, Table 2), in situ conductivities increase by about the same percentages as the gradient decreases to about 800 m (Figure 8).

Unfortunately, it was not possible to evaluate the temperature variation of thermal conductivity experimentally. Over the range 0°–250°C, the most authoritative source of data remains *Birch and Clark* [1940]. Their careful and systematic experiments established that the thermal conductivity of rock-forming materials can increase (glasses) by ~10% per 100°C, have practically no variation (feldspars and mafic minerals) or decrease with temperature (quartz, carbonates) over this temperature range. Clay minerals and their metamorphic derivatives are the most difficult to evaluate because of the myriad chemical reactions and physical changes of state that occur over the temperature range 0°–250°C. Such data as do exist indicate that the decrease with temperature, if any, is modest relative to quartz and carbonates [*Birch and Clark*, 1940; *Clark*, 1966]. In Table 2 we show the porosity-corrected thermal conductivity with no temperature adjustment, assumption a, and with a correction, assumption b, equivalent to that established by *Birch and Clark* [1940] for quartz- and carbonate-rich rocks. The effect of making the extreme correction of assumption b is to lower the overall conductivity and to make it constant over the interval 300–900 m.

HEAT FLOW

Conductive heat flow was calculated over the interval 91–152 m (Table 3) by combining the least squares temperature gradient from the July 31, 1987, log with the average of 93 "in situ" conductivity determinations made at various locations in the Imperial Valley over the same depth range [*Mase et al.*, 1981; *Sass et al.*, 1984]. Between 305 and 884 m, component heat flows were calculated using the least squares gradients and thermal conductivity values from Table 2 calculated according to two different assumptions: assumption a, that thermal conductivity does not vary with temperature; and assumption b, that the temperature coefficient of thermal conductivity obtained by *Birch and Clark* [1940] for carbonate- and quartz-rich rocks apply. There is considerable scatter, but the component heat flows calculated according to assumption a generally agree to within their combined standard errors and with the independent determination between 91 and 152 m.

For either assumption there is a suggestion of a systematic decrease in heat flow below 457 m, but particularly for assumption a the scatter is such that we can just as well assume constant heat flow with depth. For assumption b on the other hand, there is a definite systematic difference between the shallow heat flow of 462 mW m^{-2} and the mean of 351 ± 24 for the lowermost ~600 m.

The implication of heat flows calculated by assumption a is that the system is in equilibrium on a time scale of 10 kyr (the thermal time constant of a ~1-km-thick layer; see Figure 9 of *Lachenbruch and Sass* [1977]). On the other hand, if the thermal conductivity structure of assumption b is appropriate, we must conclude that the hydrothermal system is cooling on this time scale. Alternatively, all or part of the observed curvature could reflect a thermal boundary layer separating the vigor-

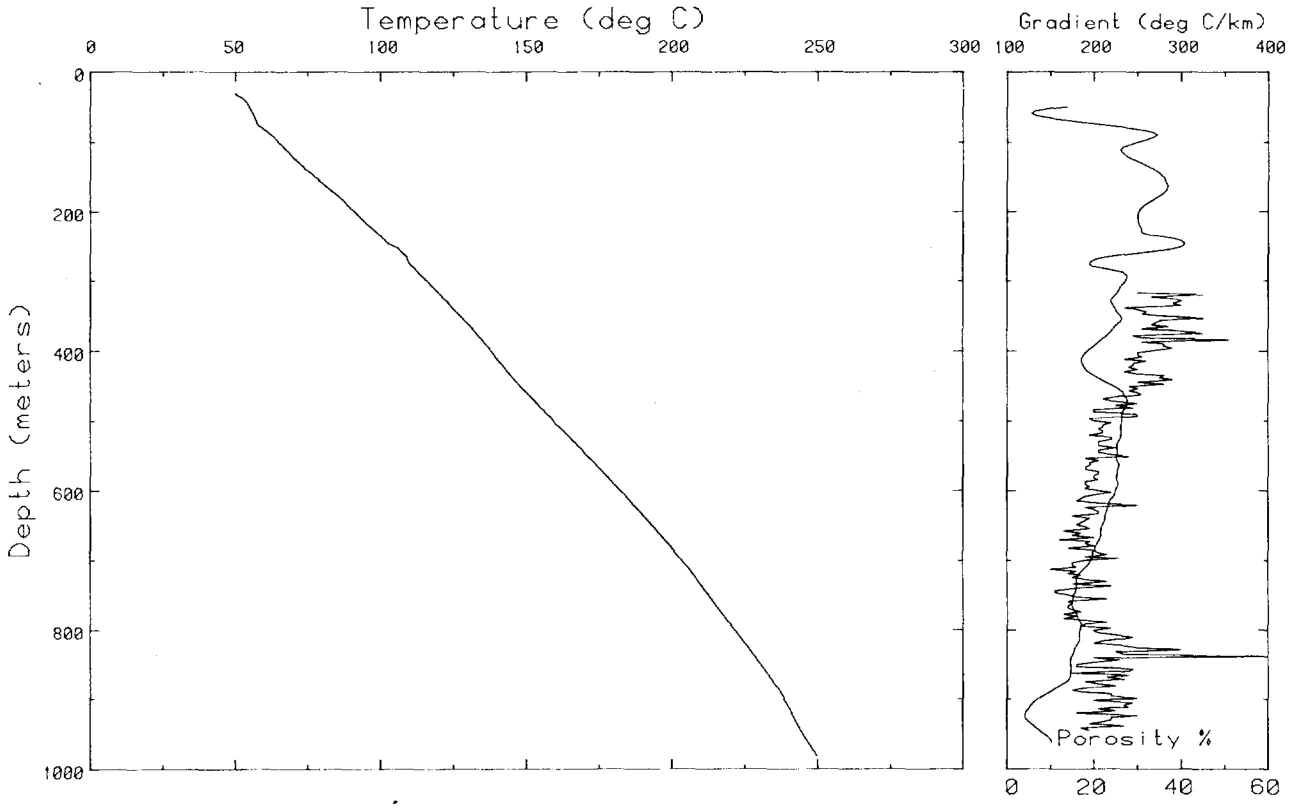

Fig. 7. Temperature, gradient, and porosity profiles for the uppermost kilometer of State 2-14. The porosity profile begins at a depth of 310 m.

ously convecting zone deeper than 1 km from the impermeable cap in the upper few hundred meters. There is, however, no evidence for the distribution of vertical permeability that would be required to sustain such a layer.

Under assumption b, conductivities are approximately constant with depth (Table 2). If we further assume a conductive thermal regime above ~900 m, then the departure from the extrapolated shallow temperatures (dashed line, Figure 8) will

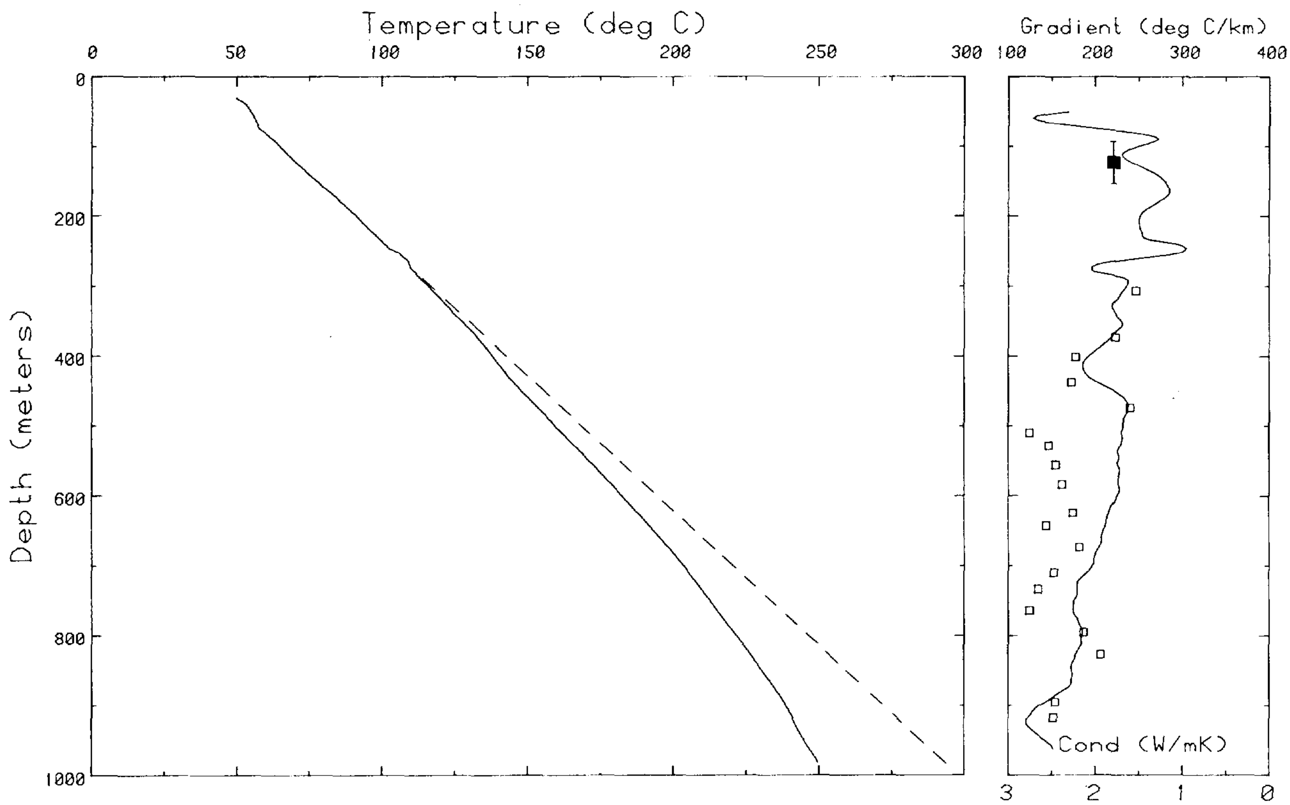

Fig. 8. Temperature, gradient, and thermal conductivity (discrete squares) profiles for the uppermost kilometer of State 2-14. Dashed line extrapolated from near-surface gradients (see discussion of heat flow).

TABLE 3. Heat Flow Over Discrete Depth Intervals in State 2-14

Depth Interval, m	Temperature Gradient, °C/km^{-1}	Number of Conductivities	Harmonic Mean Conductivity† W m^{-1} °C^{-1} a	Harmonic Mean Conductivity† W m^{-1} °C^{-1} b	Heat Flow mW m^{-2} a	Heat Flow mW m^{-2} b
91–152	246 ± 2	93*	1.88 ± 0.04		462 ± 14	
305–457	209.8 ± 1.1	4	1.90 ± 0.18	1.68 ± 0.12	399 ± 40	352 ± 27
457–610	229.4 ± 0.3	5	2.26 ± 0.23	1.82 ± 0.18	518 ± 53	418 ± 42
610–762	198.3 ± 1.0	5	2.41 ± 0.09	1.82 ± 0.06	478 ± 20	361 ± 14
762–884	178.9 ± 0.4	5	2.31 ± 0.15	1.70 ± 0.09	413 ± 28	304 ± 17
Average					454 ± 22	351 ± 24

*In situ determinations over this depth range in other parts of the valley.
†Assumption a, corrected for porosity only; assumption b, corrected for porosity and temperature (see Table 2).

be a measure of departure from equilibrium. As pointed out by *Lachenbruch et al.* [1976] (see also *Lachenbruch and Sass* [1977]), the effective age of the resulting disturbance depends, among other things, on whether the disturbance is a sudden temperature change or a change in source strength. A step change in temperature will be felt at a distance much more quickly than the gradual waning of the hydrothermal system due to cooling of its heat source, even though the ultimate effect might be the same. In any event, if conductivity assumption b is correct, we have a heat sink whose effect is propagating upward from the base of the conductive cap, whose age is the order of a few thousand years, that has caused a lowering of temperature of about 30°–40°C at the base of the conductive cap. This degree of cooling is consistent with estimates of homogenization temperatures from fluid inclusion studies [*Roedder and Howard*, 1987, this issue; *Andes and McKibben*, 1987; *McKibben et al.*, 1987]. These paleotemperatures are higher on average by several tens of degrees Celsius between 600 and 1700 m than the inferred equilibrium temperatures (Figure 3).

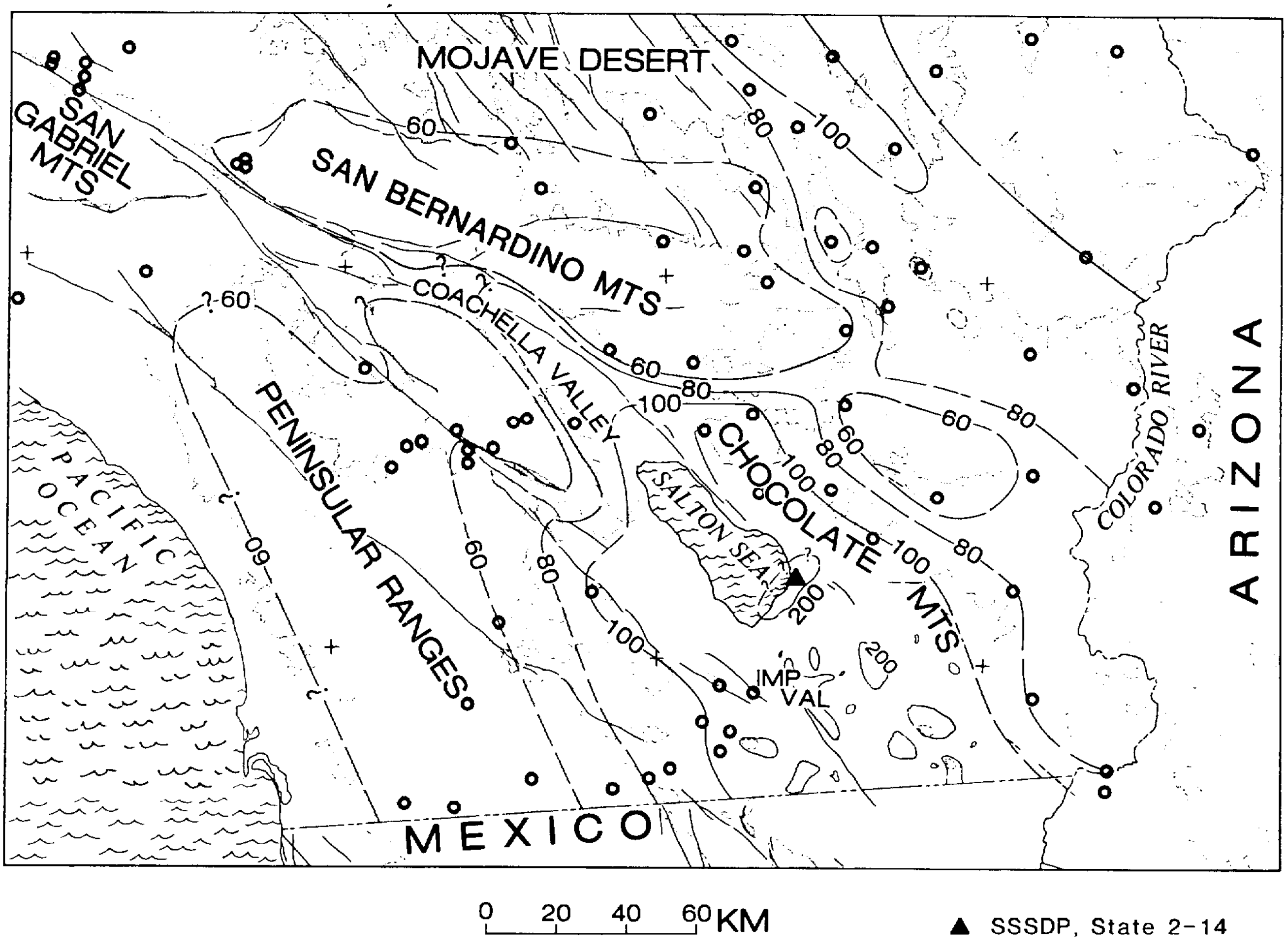

Fig. 9. Regional thermal setting of the Salton Sea system [after *Lachenbruch et al.*, 1985]. Contours are milliwatts per square meter. Control is shown as open circles.

From the internal consistency of component heat flows and the expected correlation between porosity and thermal gradient over a substantial depth interval (Figure 7), taken with fragmentary evidence for a very low temperature coefficient of thermal conductivity for clays and alteration products, we favor the interpretation of assumption a (Tables 2 and 3) and abopt a rounded mean conductive heat flux of 450 mW m^{-2} at this site. This is within the range of shallow heat flow estimates for this area [*Sass et al.*, 1984; *Newmark et al.*, this issue]. We should emphasize, however, that conductivity assumption b, while less plausible, is also possible; in which case, heat flow is presently decreasing with depth and reflects either a thermal transient introduced by recent cooling of this part of the field or a thermal boundary layer. At present, we have no high-temperature thermal conductivity data or data on vertical permeability, which would better allow us to discriminate among the three interpretations.

SUMMARY AND CONCLUSIONS

The regional thermal setting of the Salton Sea geothermal system [*Lachenbruch et al.*, 1985] is illustrated in Figure 9. The entire Imperial Valley and surrounding ranges are enclosed by the 100 mW m^{-2} contour, and there are several large areas of heat flow greater than 200 mW m^{-2} within the region. Of these, the zone at the southeast margin of the Salton Sea (Figure 9) is by far the hottest, with some conductive heat flows greater than 1 W m^{-2}.

Detailed temperature logs to depths of 3 km or greater and to temperatures of 350°C or greater are extremely rare. Mechanical transducers can provide individual, widely spaced points to these depths and temperatures (Figure 6). Temperature-sensitive electrical resistors with surface readout are limited to between 250° and 300°C in the case of Teflon-insulated cables or to depths of a few hundred meters (because of strength limitations and high conductor resistance) in the case of oxide-insulated conductor cables. We have shown that a downhole-recording electronic tool with sensitive components housed in a dewar container, is capable of repeated logs with high data density, rapid response, and good reproducibility. This configuration provides the best combination of characteristics for obtaining reliable, detailed temperature data in the hostile environment of a geothermal field.

The State 2-14 well was sited off the main axis of volcanic centers and its temperature-depth profile lies significantly below those of many wells in the field. Temperatures in State 2-14 were also lower than those predicted before the well was spudded, based on extrapolation of contoured isotherms from other wells. On the most plausible interpretation, heat flow within the upper ~900 m does not vary significantly with depth, suggesting a relatively stable thermal regime over the past 10,000 years or so. There remains the possibility, however, based on ambiguities in the interpretation of thermal conductivities measured at room temperature, that the hydrothermal system deeper than 900 m has cooled from a maximum temperature several tens of degrees Celsius higher than the present profile, a possibility also suggested by fluid-inclusion studies.

Acknowledgments. We thank Jack Kennelly and Walt Wendt for their help in obtaining the data. David Blackwell, Jim Combs, Wilfred Elders, Art Lachenbruch, and Paul Morgan provided constructive criticism of earlier versions of the manuscript.

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(Received October 1, 1987;
revised April 22, 1987
accepted April 25, 1988.)