The Near-Surface Hydrothermal Regime of Long Valley Caldera

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Temperatures at the 5- to 10-m depth from 29 shallow holes in Long Valley caldera can be contoured systematically; they correlate well with the character of the thermal gradient to 30 m. Where the temperature at a depth of 10 m is less than 11°C (group I), the gradients to 30 m are practically zero; where the 10-m temperature is between 11°C and 16°C (group II), the gradients are 200°–400°C/km and uniform, corresponding to conductive heat flows of 4–8 HFU (1 HFU = 1×10^{-6} cal/cm^2 s). Where the 10-m temperatures exceed 16°C (group III), gradients are larger and irregular, with local heat flows to 50 HFU. Thermal considerations suggest that the first group is characteristic of regions of hydrologic recharge, that the second group is probably characteristic of regions with conductive régimes to substantial depth, and that the third group is characteristic of regions of hydrologic discharge. This interpretation is supported by limited drilling to depths up to 300 m. Regimes in group I occur in the peripheral portion of the caldera, suggesting that this is an area of recharge. The hot springs discharge in a fault zone characterized by near-surface regimes in groups II and III; chemical evidence indicates that their source reservoir is at about 200°C. Evidently, the springs are fed by local fractures; if the background regime is conductive, their reservoirs are probably less than 1 km deep. Hydrologic and isotopic data indicate that gross circulation in the hydrothermal system is from west to east, suggesting that the hot springs gain their heat in the western caldera. The large estimates of heat being removed from the caldera by flowing water and the geologic inference that hydrothermal activity was more intense in the past support the view that the Long Valley system was resupplied with heat from deep magmatic sources during its eruptive history.

INTRODUCTION

As part of an investigation of the thermal state of the Long Valley geothermal area, we have measured temperatures in 29 holes drilled to depths up to 30 m and in 7 deeper holes (up to 300 m) within the caldera (Figure 1). A limited number of thermal conductivity measurements were made to permit rough estimates of conductive heat flow, but the detailed measurements of thermal and hydrologic parameters needed to calculate local heat and mass flux were not undertaken. Our aim was to see if inexpensive near-surface measurements could be used to obtain useful information about the hydrothermal system at greater depth.

Geothermal areas are generally characterized by large and locally variable surficial thermal gradients and by various forms of hot spring activity indicating that heat is crossing the earth's surface at an anomalous rate. At some points beneath these areas, we can expect anomalously high temperatures which might represent exploitable concentrations of geothermal energy. Such concentrations are generally at depths of ½–3 km; deeper sources would be costly to exploit, and shallower ones have insufficient pressure to permit the high enthalpy desired. The process by which heat is transferred to the surface from these depths can be complex. The farther beneath the surface we extend our observations, the more we can expect to learn about the hydrothermal system associated with the potential resource, but the more costly and time consuming each observation becomes. Hence, it is of interest to examine the near-surface hydrothermal regime in a geothermal area in which more comprehensive studies are anticipated; inferences about conditions at greater depth eventually can be tested, and the role of shallow observations in geothermal exploration can be evaluated.

MEASUREMENTS OF TEMPERATURE

Temperature measurements, particularly in geothermal areas, can be sensitive to details of the construction of observation wells. Our measurements were made in two types of wells: (1) 'shallow holes' to a maximum depth of 30 m and (2) 'core holes' to a maximum depth of 300 m. Their locations are shown by the small and large dots, respectively, in Figure 1.

The shallow holes were drilled with a hollow stem auger rig and completed with 5-cm PVC pipe. In those wells where the depth to the water table was greater than about 3 m, a second 1.9-cm pipe was placed inside the 5-cm pipe and sealed at the bottom. It was filled with water and allowed to equilibrate to facilitate temperature measurement above the water table. The core holes were drilled to 14.3-cm diameter with hydraulic rotary equipment and completed with 3.2-cm black iron pipe that was sealed at the bottom. The annulus outside the 3.2-cm pipe was filled with cement to the surface to prevent vertical water circulation. Cores were obtained at selected intervals where feasible; their thermal conductivities were measured in the laboratory [*Sass et al.*, 1974].

Temperatures were measured repeatedly to millidegree precision with equipment described by *Sass et al.* [1971]. Representative values are shown in Figures 2, 3, 4, and 5 (for a preliminary compilation of the data, see *Sass et al.* [1974]). Five of the core holes were drilled at the sites of shallow holes, thereby permitting a comparison of effects of the two different types of hole construction on measured temperature. System-

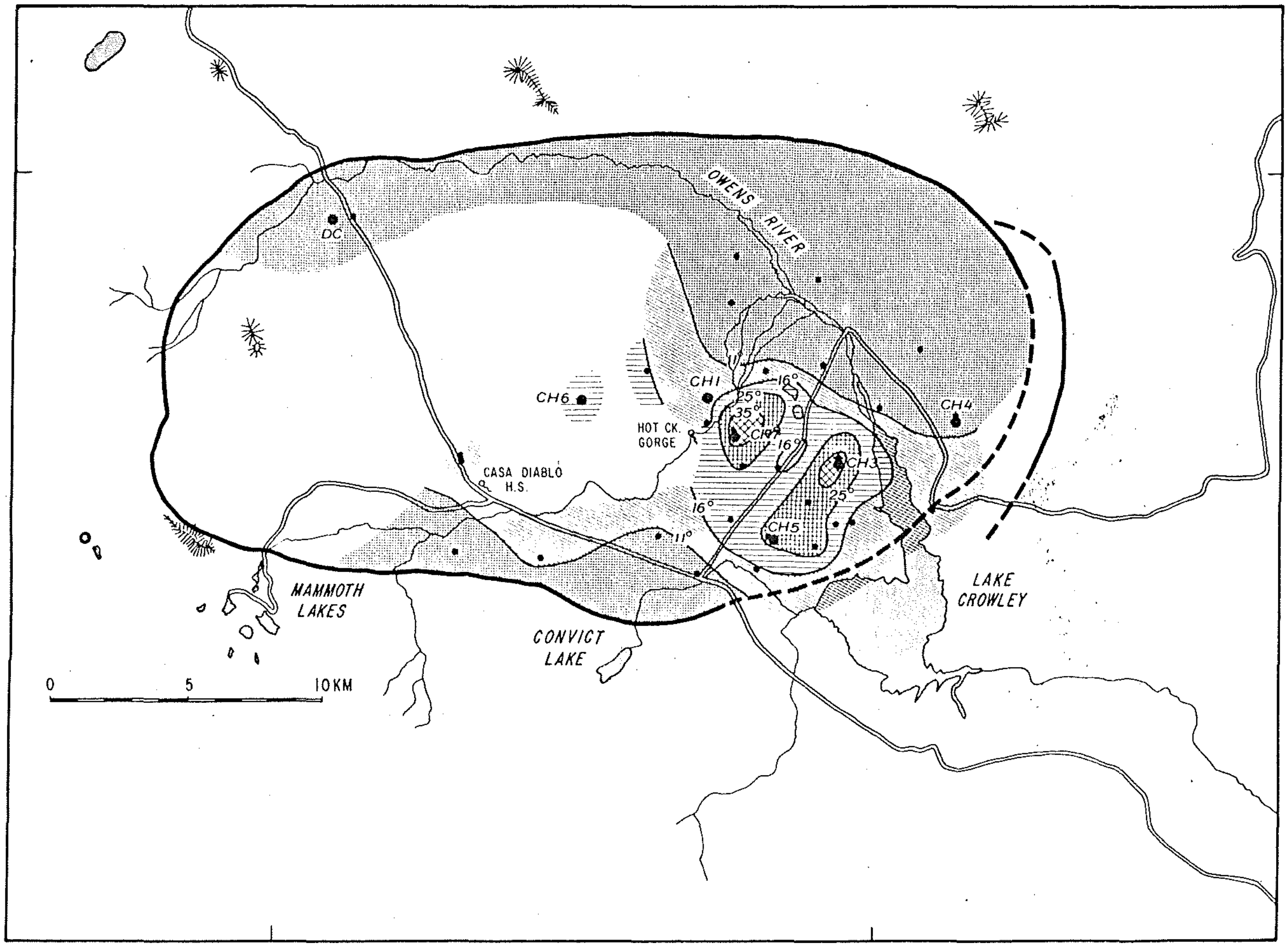

Fig. 1. Map of Long Valley caldera. Contours and patterns represent temperature at depth of 10 m in June 1974 at shallow hole locations indicated by small black circles. Large black circles indicate core hole locations.

atic differences were observed at all sites, but at only one of them, CH-3, were the differences large enough to affect the gross interpretation. At CH-3, temperatures above 45°C persist to within 3 m of the surface (Figure 4), whereas those in the adjacent shallow hole (LV-6) diverge from the core hole temperatures and fall rapidly in the upper 15 m. The difference is probably caused by relatively cool water leaking downward outside the pipe at LV-6 and then moving laterally into a permeable zone known to occur there. This disturbance would result in an underestimation of conductive flux to the surface by an order of magnitude. The vertical flow is blocked in CH-3 by cement in the annulus. In CH-1, slow upward water movement produced a systematic temperature disturbance in the upper 200 m until the annulus was cemented a year after completion. Subsequently, the disturbance decayed conductively; the equilibrium condition is represented by the data for CH-1 shown in Figure 5.

Temperature Patterns

We shall consider the shallower temperature observations first and then proceed to the deeper ones, investigating what types of information might be obtained from each. Figure 1 shows that temperatures in the shallow holes at a depth of 10 m can be contoured in a fairly simple way. We should like to know what this systematic pattern might mean in terms of the thermal regime at greater depths and whether the same information might have been obtained from temperature observations at smaller depths. Partial answers and a useful perspective are provided by the complete temperature profiles shown for two dates (June and October) in Figure 2*a*. (The individual curves can be identified with their respective hole locations from the compilation of *Lewis* [1974].) Figure 2*b* indicates the gradients that would be associated with steady conductive heat flows from 1 to 40 HFU (1 HFU = 10^{-6} cal/cm^2 s = 41.8 mW m^{-2}) for an assumed conductivity of 2 mcal/cm s °C (0.84 W/m °K).

With some notable exceptions (probably due to local water movements), the observed seasonal variation extends only to a depth of about 10 m (Figure 2*a*), roughly consistent with what we might anticipate from a simple model of conductive damping. At greater depth the profiles can be subdivided somewhat arbitrarily into three groups, denoted by the roman numerals in Figure 2*a* and illustrated by examples in Figure 3. In group I the steady gradients in the upper 30 m are small, generally representing heat flows less than 1 HFU. In group II, gradients lie typically in the range of 200°–400°C/km, representing conductive heat flows of about 4–8 HFU. In group III the gradients are generally larger, ranging to over 1000°C/km and representing conductive heat flows up to 50 HFU locally. The gradients in group III tend to be quite variable with depth, and although it is not obvious from the selected data in Figure 2*a*, they are also subject to more variation with time than gradients in the other groups. Figure 2*a* shows that these groups are distinguishable at the 10-m depth; in fact, the 11°C contour in Figure 1 separates groups I and II, and the 16°C contour separates groups II and III. As long as synoptic obser-

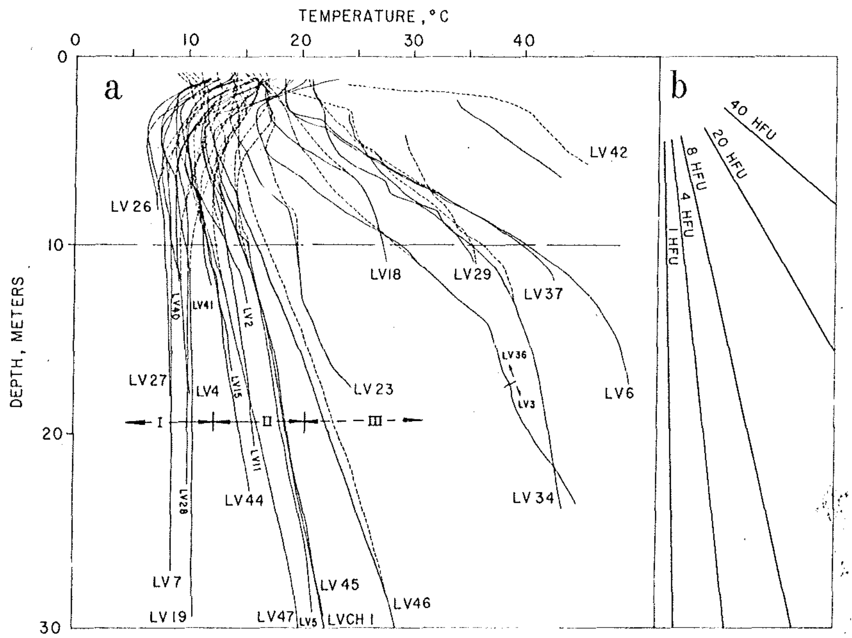

Fig. 2. (*a*) Temperatures to 30 m in Long Valley caldera in June 1974 (solid curves) and October 1974 (dashed curves). Roman numerals denote the three groups discussed in the text. Symbols (e.g., LV-34) are hole designations [see *Lewis*, 1974]. (*b*) Theoretical temperature profiles for steady conductive heat flows from 1 to 40 HFU (1 HFU = 10^{-6} cal/cm² s = 41.8 mW/m²). Assumed thermal conductivity is 2 mcal/cm s °C = 0.84 W/m °K.

vations are used at these sites, essentially the same pattern emerges for contours at the 6-m depth, and much of it persists at 3 m. Hence at this locality, temperature observations at depths of a few meters contain some information on conditions to depths of a few tens of meters.

What is the nature of this information? The general seasonal uniformity of temperatures below 10 m in group I suggests that the local heat transfer in the upper 30 m is predominantly by conduction. The vanishingly small gradients suggest that heat from greater depth is being absorbed by moving groundwater, a characteristic condition in or near zones of hydrologic recharge. Hence the peripheral part of the caldera, where 10-m temperatures are less than 11°C (Figure 1), may be an area of general recharge. The actual downward flow from the surface is probably localized; at least it does not generally occur near the holes in group I, or we should have seen seasonal temperature variation throughout their depth (Figure 2*a*). Similarly, we infer from the large and variable gradients that near-surface thermal conditions in group III are characteristic of areas of regional discharge in the sense that the groundwater beneath them is giving up heat as it moves generally to regions of lower ambient rock temperature. The magnitude and uniformity of gradients at group II sites suggest that they overlie regions less disturbed by water flow. The heat flows at group II sites are comparable to the hydrologically undisturbed value of about 4 HFU measured in granitic rock at site DP just beyond the western caldera rim (Figure 1; *Lachenbruch et al.* [1976]), although this agreement is probably fortuitous. Figures 1 and 6 show that the region in which the hot springs discharge is characterized by near-surface thermal regimes in groups II and III. Group II sites occur within 1 km of Casa Diablo Hot Springs and Hot Creek Gorge Springs, suggesting that the upflow is confined to local fracture systems.

The core hole observations expand our depth range another order of magnitude (Figure 5) and provide additional information on the meaning of the near-surface measurements. Two of the core holes, DC and CH-4, were drilled at opposite ends of the caldera (Figure 1) in the region in which the shallow hole profiles are in group I. They indicate that the low temperatures and small conductive heat flows persist at depth (Figure 5). A temperature minimum near the 100-m depth in DC implies that heat is being absorbed by lateral water flow there, but the small conductive heat flow (1.5 HFU) in the bottom of DC suggests that hydrologic heat sinks occur also at greater depth. Hence the fragmentary data from all holes taken collectively suggest that beneath the region around the periphery of the caldera delineated by 10-m temperatures less than 11°C (Fig-

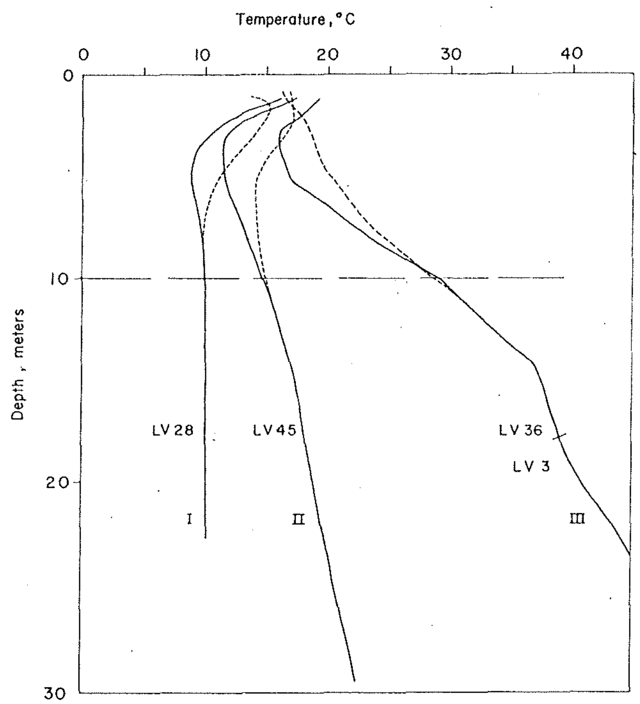

Fig. 3. Typical examples of the three groups indicated in Figure 2.

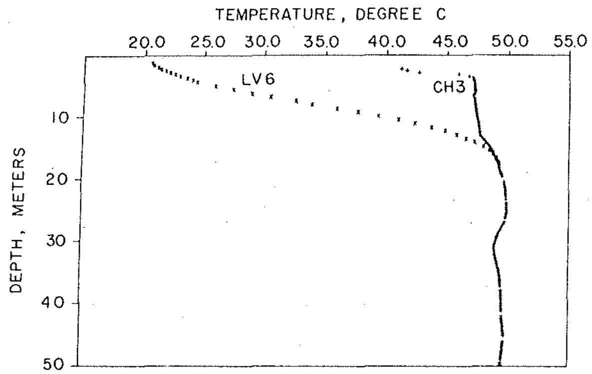

Fig. 4. Comparison of temperatures in LV-6 with uncemented annulus and CH-3 with cemented annulus. Holes are 20 m apart.

ure 1), heat is being absorbed by recharging groundwater to depths of at least 200 m or so.

Three of the core holes, CH-3, CH-5, and CH-7 (Figure 5), were drilled in the region in which shallow profiles were in group III. They show the influence of lateral and vertical circulation of hot water in permeable layers or fractures. As it liberates heat to the surrounding rock, the moving water causes high conductive heat flow above and sometimes gradient reversals below. Data from CH-5 suggest that water discharged at or near the surface in the hot springs is moving laterally in a shallow aquifer overriding colder recharge water, this suggestion being consistent with one by *Mariner and Willey* [1976]. One core hole, CH-6, was drilled in Little Antelope Valley, where no local shallow temperature information was previously available. The near-surface regime is in group III; the nearly isothermal part of the temperature profile below 165 m indicates that circulation of hot groundwater is influencing the thermal regime in the upper 200 m. It is clear from this group of results that deepening any of these holes by 50 or 100 m could lead to surprises and that downward extrapolation of near-surface temperatures is hazardous.

One of the core holes, CH-1, was drilled in a region in which the nearby shallow temperature was in group II, and it shows little or no effect of hydrothermal circulation. The change in gradient from about 250°C/km in the upper half to about 175°C/km in the lower half could be caused by a systematic variation in thermal conductivity or by very slight vertical water movements; the thermal conductivity sampling was insufficient to permit a distinction. (Limited conductivity data from CH-1 [*Sass et al.*, 1974] indicate a conductive heat flow of 3–5 HFU.) In any case, temperatures in CH-1 seem consistent with the speculation that regions in Figure 1 in which the 10-m temperature lies between 11°C and 16°C might be underlain by predominantly conductive regimes to depths of hundreds of meters.

DISCUSSION

Some tentative generalizations regarding the Long Valley geothermal system are suggested by combining the foregoing results with findings of related studies reported in this volume.

It has been pointed out that the present area of major hot spring discharge lies within the region in which the 10-m temperatures were above 11°C (Figure 6). This region includes areas (in group III) where thermal profiles in the upper 200 m are strongly influenced by lateral and vertical circulation of water at temperatures up to at least 110°C. It also includes areas (in group II) where the thermal regime is largely conductive with intermediate gradients to 30 m, and in at least one location (CH-1), to 300 m. The hot springs are probably fed by fractures, as they are located on or near the intracaldera extensions of the Hilton Creek fault (Figure 6) [*Bailey et al.*, 1976]. From chemical evidence it has been estimated that the reservoirs feeding these springs have a temperature of about 200°C [*Mariner and Willey*, 1976; *Sorey and Lewis*, 1976]. Even if we allow for an increase in conductivity with depth, this temperature would occur at a depth of about a kilometer or less according to extrapolation of group II gradients. Some justification for the extrapolation is provided by the fact that Casa

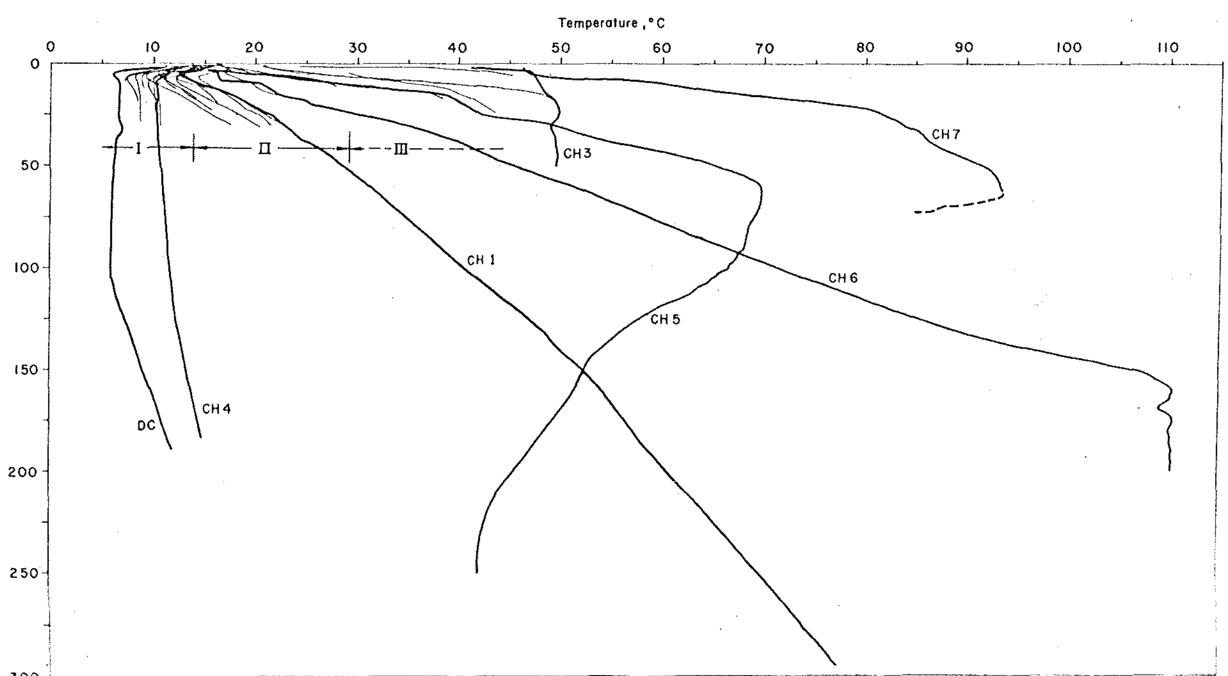

Fig. 5. Temperature profiles from the core holes (heavy lines) and shallow holes (light lines). Dashed portion of CH-7 represents approximate temperatures (±2°C) obtained before caving.

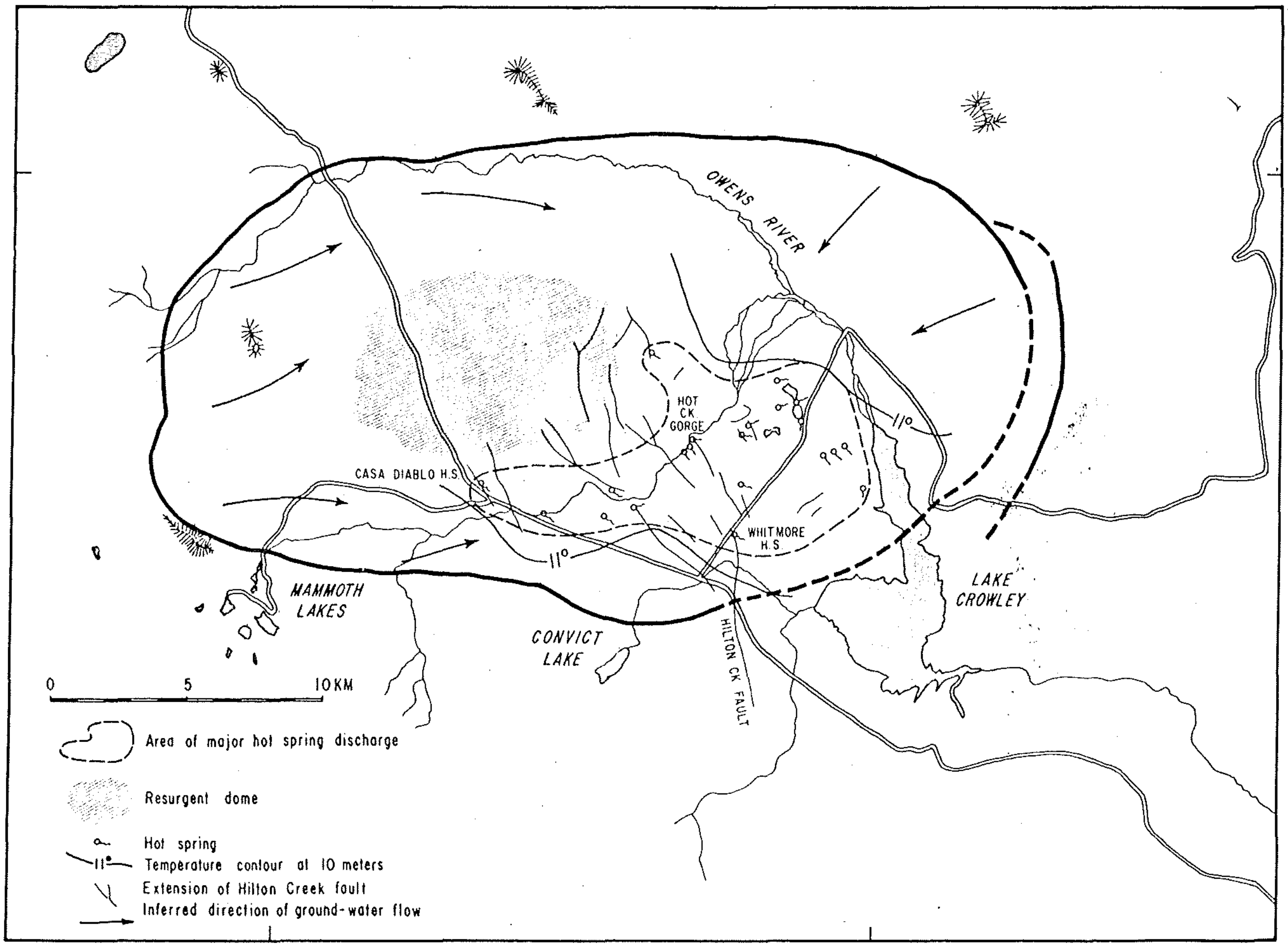

Fig. 6. Map of Long Valley caldera showing relations among groundwater flow, hot spring discharge, 10-m temperatures, and the resurgent dome (crosshatched).

Diablo Hot Springs and the Hot Creek Gorge Springs lie very close to shallow holes with regimes in group II (Figure 1), implying that flow in the conduits of the springs may produce only local perturbations to a predominantly conductive regime. However, the limited information in Figure 5 clearly indicates that many measurements to greater depths will be needed before the circulation beneath the hot springs can be delineated with confidence.

The topography indicates that surface drainage in the caldera is primarily from the Sierra Nevada Mountains in the west towards Lake Crowley in the east, interrupted locally by higher elevations on the resurgent dome. Limited data on water table elevations [*Lewis*, 1974] indicate that flow in the shallow groundwater system follows the topographic slope. Moreover, the supply of water available for recharge to the groundwater system is considerably larger in the mountains around the western and southern rims than around the eastern rim of the caldera. Thus it is reasonable to expect that the gross circulation of groundwater in the caldera is from west to east, towards the areas of hot spring discharge and Lake Crowley.

The limited information on ground temperatures indicates that cool recharge water occurs around the rim of the caldera to depths of at least 200 m. Isotopic data on concentrations of ^{18}O and D, as presented by *Mariner and Willey* [1976], suggest that water which emerges in the hot springs originated in the Sierra Nevada Mountains to the west; shallow groundwater from the peripheral regions evidently flows laterally towards the hot springs, mixing in varying proportions with the upflowing hot water. For this reason and the ones given in the previous paragraph we have indicated in Figure 6 that the direction of flow in the hydrothermal system is generally from west to east with discharge in the hot spring area in the southeastern part of the caldera. If this is so, the input of heat required to raise the temperature of the circulating meteoric water from approximately 10°C to temperatures above 200°C occurs at depth in the western half of the caldera.

The above inference is consistent with studies of heat flow and magmatic history, which suggest that the greatest heat at depth is likely to occur in the western part of Long Valley caldera [*Lachenbruch et al.*, 1976]; the magnitude of the mean conductive flux there is still unknown. From geochemical mixing models and measured spring discharges, *Sorey and Lewis* [1976] have estimated that water is presently removing heat at a rate greater than 10 HFU integrated throughout the entire caldera. A similar calculation by *White* [1965, p. 8] based on an estimate of the total boron discharged by the Long Valley thermal waters leads to an even larger value of about 16 HFU. The significance of such heat loss to the thermal budget of the heat source beneath Long Valley depends, of course, upon how long it has been going on at the measured rate.

If we assume that the heat removed by water flow comes ultimately from a magmatic source, it must be supplied to the surface by conduction and circulatory hydrothermal convection through the overlying roof rocks and possibly in addition by some direct contribution of magmatic fluids. A mean flux of 10 HFU could be supplied by steady state conduction from a molten magma chamber through about 5 km of crystalline roof rocks (conductivity ~6 mcal/cm s °C). Circulatory hydrothermal convection would increase the required roof thickness by some fraction of the depth to which it was effective. If 15% of the hypothetical 200°C water contributing convective heat to the springs were supplied directly by magmatic fluids, less conductive flux would be required, and the estimate of roof thickness might be doubled [see, e.g., *White*, 1957]. However, if the magma underlies only half of the caldera [*Lachenbruch et al.*, 1976], the above estimate of roof thickness would be halved.

The time constant for propagation of a temperature disturbance by conduction through 5 km or so of roof rocks above a magma chamber is $\sim 10^5$ yr [see *Lachenbruch et al.*, 1976, Table 2]. Hydrothermal flow events of shorter duration could be removing heat stored at shallower depths. As an extreme example an additional 10 HFU could be supplied for 10,000 yr by convectively flushing the top 1 km of an otherwise conductive rock column; time would be insufficient for the gradient disturbance to propagate to the roof of a magma chamber a few kilometers below [see *Lachenbruch et al.*, 1976, curve I, Figure 10*a*]. Thus the present-day estimates of 10 HFU or greater need not constrain the depth to magma if they represented a flushing of the caldera by such short-term effects on flow as might be caused by post-Pleistocene deglaciation or new fracture openings from recent faulting. However, according to *Bailey et al.* [1976], hydrothermal activity in the caldera began at least 0.3 m.y. ago, and it was formerly much more extensive than it is today. As *White* [1968] has pointed out, heat loss associated with such prolonged activity would place large demands on a magmatic source; evidence for this activity supports the conclusion [*Lachenbruch et al.*, 1976] that the Long Valley system was resupplied with heat from deep magmatic sources throughout its eruptive history. This resupply could be achieved by any combination of (1) hydrothermal circulation in progressively deepening fractures, (2) convection in a magma chamber extending downward through the lower crust, or (3) repeated upper crustal intrusions.

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