Heat flow in Lake Tahoe, California-Nevada, and the Sierra Nevada–Basin and Range transition

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ABSTRACT

Heat-flow measurements made in Lake Tahoe, California-Nevada, demonstrate that the transition from subnormal heat flow in the Sierra Nevada to above-normal heat flow in the Basin and Range province occurs west of the assumed physiographic boundary between these two areas, contrary to earlier belief. In addition, these data, together with data of other workers, clearly reveal the sharpness of this transition, which suggests that the causative thermal sources and (or) sinks must be restricted to depths not greater than the uppermost mantle. The way in which heat-flow data constrain the current hypotheses of crustal structure and evolution of the Sierra Nevada–Basin and Range provinces is illustrated with a tectonic model that employs a post-Cretaceous shallow-dipping subduction zone beneath the Sierra Nevada and an active upper-mantle diapir under the Basin and Range province during late Cenozoic time.

INTRODUCTION

The crust and upper mantle beneath the Sierra Nevada are transitional between the Great Valley and Coast Range provinces to the west and the Basin and Range province to the east (Thompson and Talwani, 1964). The physiographic boundary between the Sierra Nevada and Basin and Range provinces is generally taken to be the first appearance of large-scale normal faulting east of the tilted Sierran fault block. This boundary is well developed along the eastern margin of the southern Sierra (Owens Valley, Long Valley, and Mono Basin) but is more obscure to the north. In the vicinity of Carson City and Reno, Nevada, Feeneman (1931) and later workers (for example, Thornbury, 1965) considered the Carson Valley (Fig. 1) to be the physiographic boundary between the Sierra Nevada and Basin and Range provinces. However, immediately to the west, the Lake Tahoe–Truckee–Sierra Valley depression, together with the Carson Range, is typical of Basin and Range structure (Birkeland, 1963). The seismic and gravity data of Eaton (1963) and Thompson and Talwani (1964) suggest that the subsurface transition (eastern margin of the Sierran root) is 50 km inside (to the east) the physiographic boundary of the Basin and Range province, although the precise nature of the subsurface transition is complicated by poorly known densities in the upper mantle beneath the Sierra Nevada and Basin and Range provinces (Thompson and Talwani, 1964). In addition, the extent of the root of the Sierra Nevada is unknown, although it is presumably deep.

Heat-flow data between Sacramento, California, and Fallon, Nevada (Roy and others, 1968a; Sass and others, 1971), demonstrate a profound change in thermal regime between the two provinces. Roy and others (1972) demonstrated that regional heat flows in the two provinces differ by about 1 HFU, if the effects of near-surface radioactivity are removed (reduced heat flow). Their reduced-heat-flow data suggest that the thermal transition coincides with the subcrustal transition proposed by Eaton (1963) and Thompson and Talwani (1964) and occurs over a lateral distance of no more than 100 km. The control on this transition is to a large degree affected by their Gardnerville value (Roy and others, 1968a). The more recent work of Sass and others (1971) suggests a more abrupt transition located much closer to the physiographic boundary. According to Sass and others (1971), if equal weight is given to their new data and the Gardnerville point, the major part of the thermal transition must occur within a lateral distance of only 20 km. Such a narrow transition implies heat sources within the crust; however, the magnitude of this transition is difficult to explain in terms of geologically acceptable crustal sources.

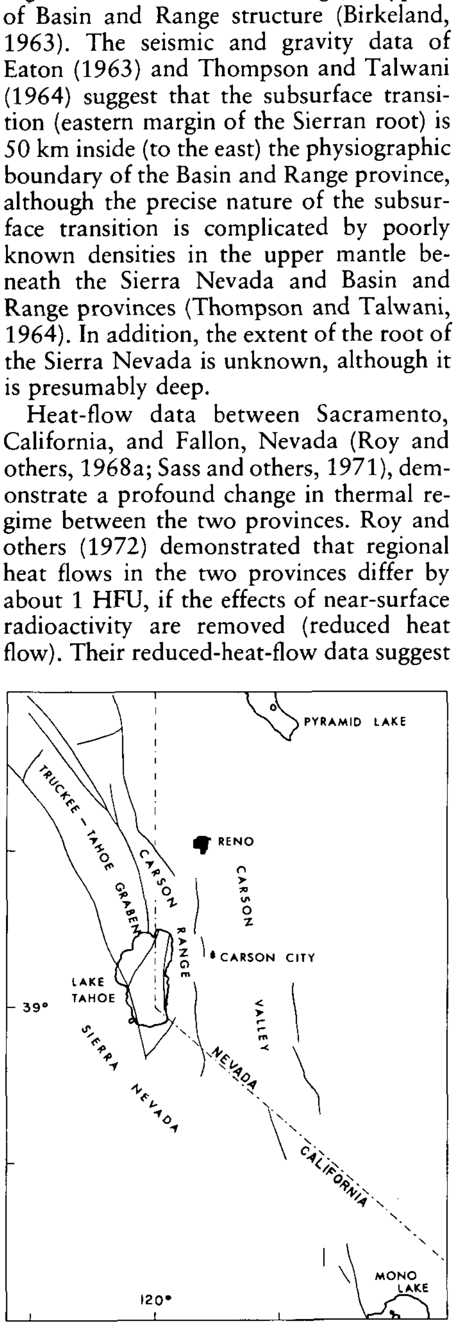

Figure 1. Physiographic map of the Lake Tahoe region.

Lake Tahoe, situated on the California-Nevada border and 15 to 25 km west of the physiographic boundary (Fenneman, 1931) between the Sierra Nevada and Basin and Range provinces, presented an opportunity to obtain additional heat-flow data near this important transition. Heat-flow measurements in large, deep lakes with isothermal bottom-water conditions have provided useful data (Hart and Steinhart, 1965; Von Herzen and Vacquier, 1967; Sclater and others, 1970; Von Herzen and others, 1974). Even measurements in small, shallow lakes (a few kilometres across) have also yielded reliable data (Williams and Roy, 1970), provided that warm rim effects are avoided or compensated for and temperature measurements are made at depths where perturbations resulting from the annual temperature cycle are appropriately attenuated. Lake Tahoe (surface dimensions, 15 km by 30 km; maximum depth, 490 m) has been shown to have essentially isothermal bottom-water conditions, with an annual temperature range not exceeding ±0.01°C (C. R. Goldman, 1971, oral commun.).

Heat-flow measurements in deep lakes typically have been made using conventional marine techniques. The effort per station is small, and thus, by using a lake of consequential size, lateral inhomogeneities and statistical variations can be evaluated

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by obtaining data from a group of stations. Hyne and others (1972) demonstrated that long sediment cores could be taken routinely over more than two-thirds of the bottom of Lake Tahoe, which is flat and largely devoid of relief exceeding a few metres (Fig. 2). Their studies also indicate that the sediments in this central region are undisturbed, well-stratified, turbidite-type deposits with a uniform depositional history.

EXPERIMENTAL TECHNIQUE

A Ewing piston-coring apparatus lowered from a small barge was used to obtain cores of lake sediment as long as 6.5 m for subsequent thermal conductivity measurements. Temperature measurements were taken with five thermistors placed at 1-m intervals upward from the bottom of the piston corer. The thermistors were left in the mud for a minimum of 8 min to permit equilibration. In some cases, thermal equilibrium of individual thermistors had not been reached in the allotted time; however, when this was the case, generally all but one or, in a couple of cases, two gradients had stabilized. We did not attempt to extrapolate the equilibrating curves (see Jaeger, 1959).

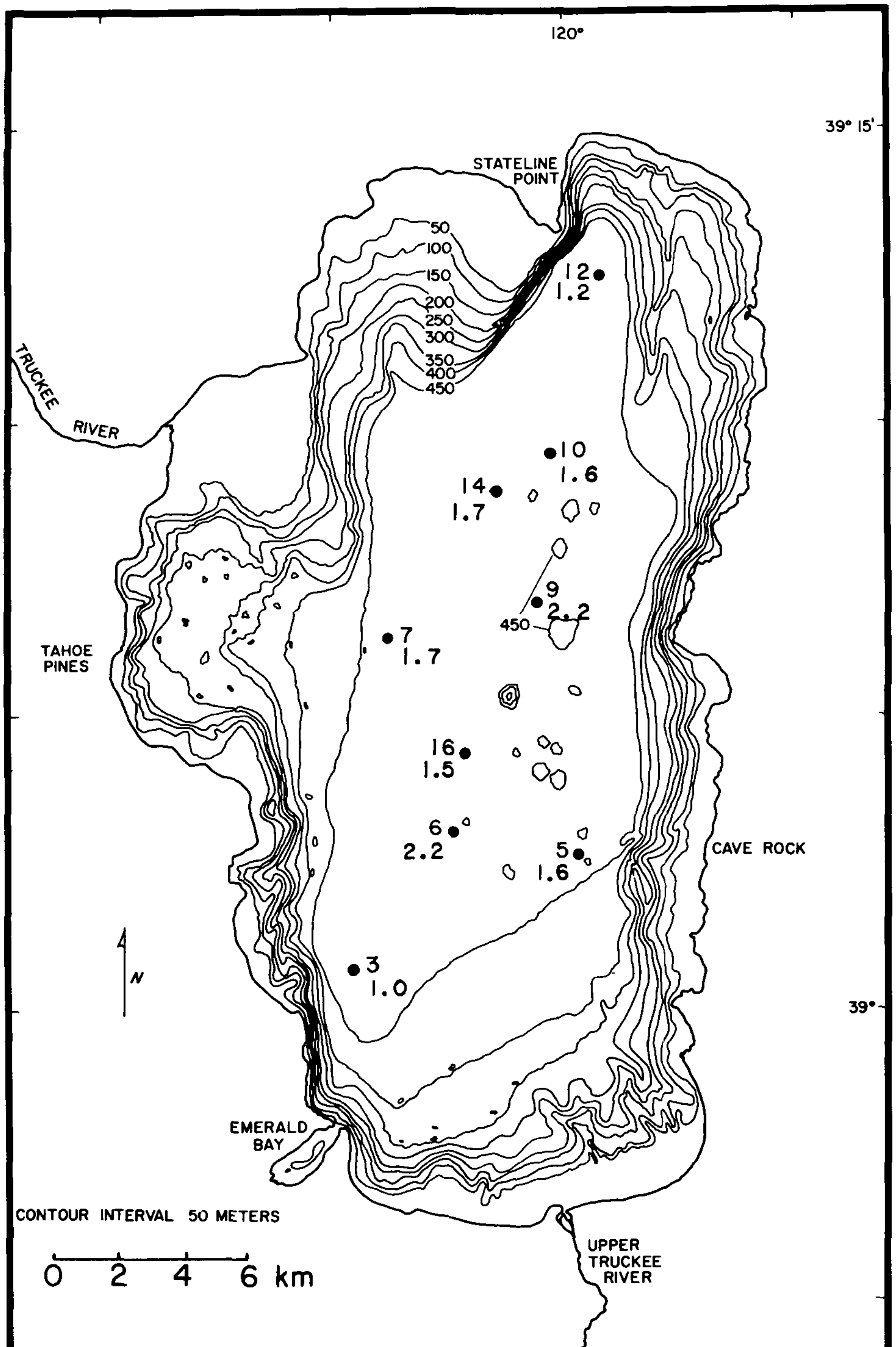

Figure 2. Bathymetry and locations of heat-flow stations, Lake Tahoe. Number on top identifies the station; number on bottom is the heat-flow value in μcal/cm²/sec.

Thermal conductivity measurements were made in the laboratory at 15-cm intervals on all cores using the needle probe technique (Von Herzen and Maxwell, 1959). Care was taken in the field to ensure against dehydration of the sediment by sealing the ends of the plastic core tubes. Corrections were applied for measurement at room temperature and pressure (Ratcliffe, 1960). All cases showed a general increase in thermal conductivity with depth, which is attributed to rapid sedimentation and a resultant lag in compaction. Figure 3 is a plot of all conductivity data versus depth from the Lake Tahoe cores. A second-order least-squares fit to the points is also shown, although it can be seen that between 1 and 2.5 m there is a bias to the low conductivity side of the least-squares fit.

METHOD OF REDUCTION

The following were considered and applied to the raw data if necessary: (1) topographic correction, (2) the effect of the temperature of the cold lake water, (3) the refraction effect of the sediment prism below the lake, (4) rate of sedimentation, and (5) bottom-water temperature variations. The corrected heat flows (Table 1) represent the application of the first two corrections. Corrections arising from the latter three considerations are discussed separately below.

The topographic correction was applied according to Birch (1950). Mean annual surface temperatures were projected onto the reference plane (see Birch, 1950) using a lapse rate of 4.5°C/km, assuming the lake was not present. U.S. Weather Bureau data from the Lake Tahoe area indicate that the mean annual temperature at lake level is probably about 6°C. With a lapse rate of 4.5°C/km, the temperature at the lake floor (essentially the reference plane) would be about 8.5°C, assuming no water. Thus, since the measured bottom-water temperature is about 4.5°C, the approximate temperature difference of 4°C due to the cold lake water was superimposed on the reference plane according to Lachenbruch (1957, equation 29). This combination of corrections introduces only minor errors due to the lake edges, which can be neglected. Topographic corrections ranged from 6 to 22 percent of the raw heat-flow value. Under the assumption of steady state, the lake correction, using reasonable estimates of the thermal diffusivity, was

TABLE 1. HEAT-FLOW DATA FROM LAKE TAHOE

Station	Water temperature (°C)	Depth range (m)	Interval gradient (°C/km)	Interval conductivity* (mcal/cm sec °C)	Interval heat flow (μcal/cm² sec)	Average heat flow (μcal/cm² sec)	Corrected heat flow (μcal/cm² sec)
3	4.65	..	..	1.56 (12)	..		
		~2.1–3.1	60	1.73 (7)	1.03		
		~3.1–4.1	75	1.76 (9)	1.32	1.18	0.98
		~4.1–5.1	88†	1.76 (10)	1.55§		
		~5.1–6.1	48†	1.82 (9)	0.87		
5	4.64	..	..	1.62 (13)	..		
		~2.1–3.1	93	1.93 (6)	1.79		
		~3.1–4.1	88	1.99 (7)	1.75	1.77	1.56
		~4.1–5.1	116†	2.18 (6)	2.53§		
		~5.1–6.1	73†	2.35 (3)	1.71§		
6	4.65	..	..	1.56 (4)	..		
		~1.0–2.0	137	1.77 (6)	2.42	2.42	2.23
		~2.0–3.0	113	2.14 (5)	2.42		
		~3.0–4.0	132†	2.08 (5)	2.74§		
7	4.64	~0.6–1.6	104	1.61 (6)	1.67		
		~1.6–2.6	109	1.78 (6)	1.94	1.80	1.69
		~2.6–3.6	132†	1.78 (6)	2.35§		
		~3.6–4.6	115	1.91 (3)	2.19§		
9	4.66	~0.5–1.5	136	1.59 (8)	2.16		
		~1.5–2.5	125	1.73 (6)	2.16	2.35	2.20
		~2.5–3.5	121	2.14 (6)	2.59		
		~3.5–4.5	111	2.24 (3)	2.49		
10	..	..	..	1.64 (15)	..		
		~1.6–2.6	108	1.77 (5)	1.91		
		~2.6–3.6	98	1.83 (5)	1.79	1.75	1.62
		~3.6–4.6	64	1.96 (5)	1.25		
		~4.6–5.6	105	1.96 (0)	2.06		
12	..	~1.1–2.1	70	1.60 (5)	1.12	1.42	1.16
		~2.1–3.1	100	1.73 (5)	1.73		
14	..	..	..	1.63 (3)	..		
		~2.2–3.2	114	1.58 (5)	1.80		
		~3.2–4.2	103	1.86 (3)	1.92	1.82	1.69
		~4.2–5.2	103	1.75 (5)	1.80		
		~5.2–6.2	102	1.75 (0)	1.78		
16	..	..	..	1.62 (13)	..		
		~2.2–3.2	101	1.63 (6)	1.65		
		~3.2–4.2	95	1.80 (7)	1.71	1.61	1.52
		~4.2–5.2	75	1.97 (5)	1.48		
		~5.2–6.2	97	2.14 (2)	2.08†§		

* Numbers in parentheses at right of conductivity values represent number of measurements in interval.
† Probes not in equilibrium.
§ Interval gradient not used in average.

found in all cases to be negligible, being at most 1 percent. Even for the case of the appearance of the lake as recently as 100,000 yr ago, the effect does not exceed 3 percent, assuming κ to be 0.002 cm²/sec (see Table 2 for a typical case). This is a result of the large size of the lake and the fact that stations were located centrally in the basin.

Refraction due to the prism of lake sediments is a function of the depth of sediments and the conductivity contrast between the sediments and the surrounding granitic rock. Near the contact, the effect is strongly dependent on the configuration of the contact (Lachenbruch and Marshall, 1966). Refraction due to semielliptical, rectangular, and trapezoidal infinite cylinders has been considered by Lachenbruch and Marshall (1966). We have used the general case for a semiellipsoid of anomalous conductivity k' embedded in a medium of conductivity k (Lee and Henyey, 1974), which can be considered a better approximation than an oblate spheroid (Von Herzen and Uyeda, 1963) for nearly equidimensional bodies. Lachenbruch and Marshall (1966) showed that for cylinders of various cross sections, the internal edge effect decays to a negligible value within a horizontal distance from the contact of about one-fourth of the length of the horizontal semiaxis. Thus, the semielliptical cylinder (or semiellipsoid) approximates most situations reasonably well for points near the center. The errors introduced by this approximation are less than those arising from the actual but unknown configuration — especially from depth measurements. Figure 4 shows results for an ellipsoid (axis $a \geqslant$ axis $b \geqslant$ axis c, and $b/a = 1/3$) with various ratios of b/c and the conductivity ratio $f = k'/k$.

All stations in Lake Tahoe were centrally located with regard to the internal edge effect. The inferred sedimentary prism was converted to a semiellipsoid of equivalent mid-cross-sectional area for the purposes of calculation. Normal fault dip angles were taken to be 55°, which are typical of Basin and Range faults (Stewart, 1971; Thompson, 1959). Hyne and others (1972) found similar dip angles by extrapolating apparent sediment-bedrock contacts. Table 3 summarizes the corrections for various sediment depths and conductivity ratios. On the basis of estimated sediment depths (Hyne and others, 1972) and our thermal

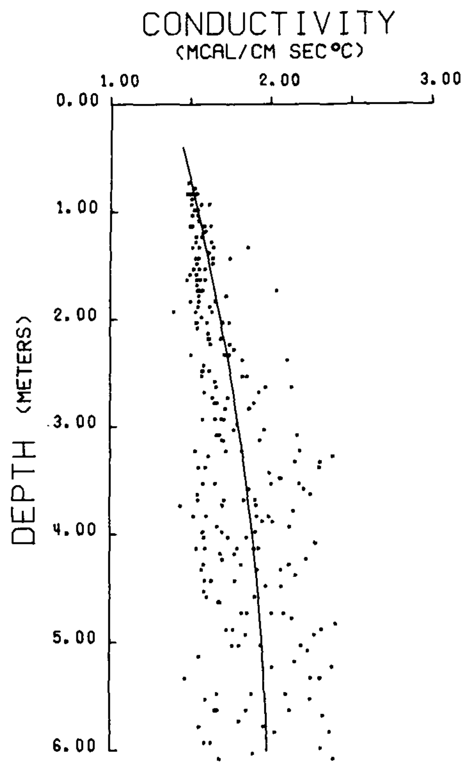

Figure 3. Conductivities from Lake Tahoe plotted as a function of depth. A least-squares, second-order polynomial has been fitted to the data.

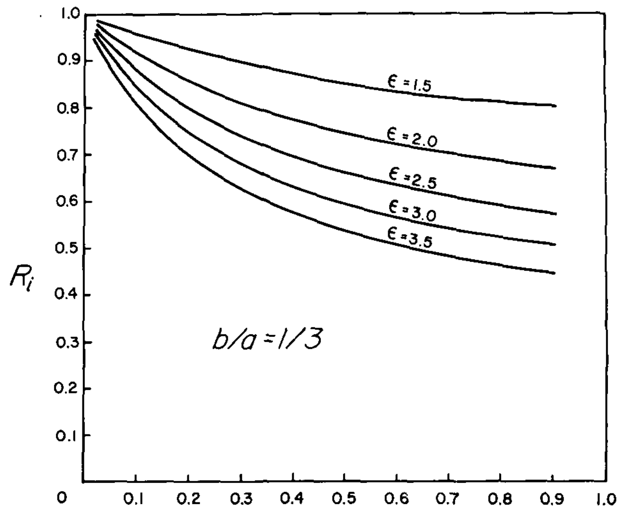

Figure 4. Refraction effects for a semiellipsoid having various conductivity contrasts with the surrounding medium. a = Semimajor axis; b = semi-intermediate axis; c = semiminor axis (oriented vertically); ϵ = ratio of conductivity within the ellipsoid to that outside the ellipsoid; R_i = ratio of heat flow inside the ellipsoid to the undisturbed heat flow.

conductivity measurements, and allowing for compaction with depth, the heat flow in the lake may be approximately 15 percent below the regional average, owing to refraction.

The rates of sedimentation for Lake Tahoe are not well known. The only available estimate of 10 to 25 cm/1,000 yr (Hyne and others, 1972) was made on the basis of radiocarbon dating of meager organic matter from two sediment cores. However, the rates are probably quite variable in space and time for the lake; minimal irregularities are to be expected near the center. Sedimentation tends to reduce the heat flow, and irregular rates will cause irregularities in the temperature gradient. These effects will be discussed in more detail below.

Bottom-water variations can introduce serious errors into geothermal gradient measurements near the sediment-water interface. The bottom water of Lake Tahoe averages about 4.5°C, which is about 0.5°C above the maximum density for that water mass. Although the lake appears to exchange water periodically between the top and bottom (as evidenced by oxygen measurements; C. R. Goldman, 1971, oral commun.), because of the large volume of water and resultant thermal mass, the annual range in bottom-water temperature is small; during the past three years it has been on the order of 0.10°C or less (C. R. Goldman, 1971, oral commun.). Assuming a sediment diffusivity of 0.002 cm^2/sec, this temperature is attenuated by 10 times at 3 m and by 100 times at 6 m; even in the worst case the errors introduced into the gradient do not exceed 1 percent. Of course, the bottom-water temperature determinations during the past two years do not rule out the possibility of larger fluctuations in the past that might have resulted from long periods of turbulence at the lake surface, which provided more extensive intermixing of surface and bottom waters.

RESULTS

The data from nine reliable measurements (Table 1) show essentially a trimodal distribution in heat-flow values: two values exceed 2 HFU; two values are less than 1.2 HFU; and five values average about 1.6 HFU. Figure 5 shows one gradient from each of these three groups. The interval gradients between the surface and the uppermost temperature sensor are not accurate, because the depth of penetration of the corer is known to only ±0.5 m; this interval gradient is never used for calculation of heat flow. The irregularities in the gradients for the three stations are real and are due to nonequilibrium of one or more probes. Those intervals that were not in equilibrium are noted with an asterisk in Table 1 and are not used for heat-flow determinations. Real variability in gradients has been noted for other lake measurements (Sclater and others, 1970).

Steady-state gradient irregularities can result from vertical and horizontal variations in thermal conductivity and from steady-state water circulation in the sediments. We did not find that gradient-conductivity correlations explained the irregularities in most cases (Table 1), and we assumed that horizontal variations in conductivity and steady-state circulation did

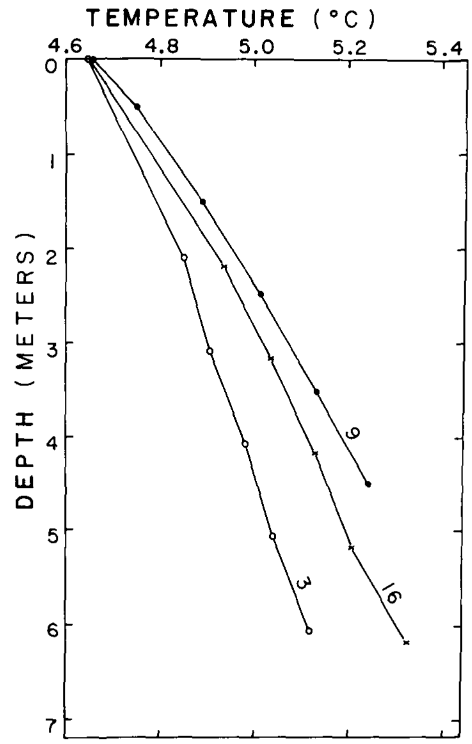

Figure 5. Three typical geothermal profiles from Lake Tahoe (from stations 3, 9, and 16).

TABLE 2. TRANSIENT DISTURBANCE OF GRADIENT FOR STATION 7 DUE TO FORMATION OF LAKE TAHOE

Time* (m.y.)	$\Delta \frac{dT}{dZ}$ (°C/km)	Disturbance (%)
0.1	2.84	2.60
0.3	1.64	1.50
0.5	1.27	1.16
1.0	0.90	0.82
10.0	0.42	0.38
15.0	0.40	0.36

* Number of years since the first appearance of the lake.

TABLE 3. APPROXIMATE REFRACTION EFFECT OF LAKE SEDIMENTS ON HEAT FLOW IN LAKE TAHOE

$1/\epsilon$	$h = 100$ $h/b = 0.0125$ $c/b = 0.0158$	$h = 250$ $h/b = 0.03125$ $c/b = 0.0394$	$h = 500$ $h/b = 0.0625$ $c/b = 0.0778$	$h = 1{,}000$ $h/b = 0.1250$ $c/b = 0.1520$	$h = 1{,}500$ $h/b = 0.1875$ $c/b = 0.2230$	$h = 2{,}000$ $h/b = 0.2500$ $c/b = 0.290$	$h = 3{,}000$ $h/b = 0.3750$ $c/b = 0.415$	$h = 4{,}000$ $h/b = 0.5000$ $c/b = 0.525$	$h = 5{,}000$ $h/b = 0.6250$ $c/b = 0.622$
1.5	0.99	0.98	0.96	0.94	0.92	0.89	0.87	0.84	0.83
2.0	0.98	0.96	0.92	0.88	0.84	0.80	0.76	0.73	0.71
2.5	0.97	0.94	0.89	0.83	0.78	0.73	0.68	0.64	0.62
3.0	0.96	0.92	0.87	0.78	0.72	0.67	0.61	0.57	0.55
3.5	0.95	0.90	0.84	0.75	0.68	0.62	0.56	0.52	0.50
4.0	0.94	0.88	0.80	0.70	0.64	0.58	0.52	0.47	0.45

Note: Tabulated values give the normalized heat flow in the lake. Abbreviations are as follows: b, half intermediate axis = 8 km; c, half minor axis, $c = 2/\pi\ (2 - h/b \cot \alpha)h$; h, thickness of lake sediments above crystalline basement; ϵ, conductivity ratio. Dip of fault planes bounding lake = 55°; half major axis = 20 km.

not occur on a scale sufficiently small to produce irregularities in a temperature profile for a depth of 5 m. Thus, non–steady-state conditions are likely to be responsible. Submarine erosion and deposition, bottom-water variations, and water circulation or free convection in the sediments can produce irregularities. Although appreciable bottom-water variations have not been noted from monthly observations during the past three years, high atmospheric turbulence may produce either unobserved short-term fluctuations with rapid decay times or stepwise changes in temperature, prior to the time of measurement of water temperatures. Changes in bottom water at random intervals 10 to 20 yr ago can cause irregularities (Birch, 1948). For example, a change of 1°C between four and five years ago can cause an error of about 10 percent in a gradient at 4 m. Thermal convection can be ruled out, since the probable thermal and physical characteristics are such that Rayleigh numbers significantly less than the critical Rayleigh number are to be expected (Elder, 1965). Circulation within the sediments cannot be ruled out. Graded sandy layers are prevalent in the nearshore sediments (Hyne and others, 1972), through which water may move freely under vertical and (or) horizontal pressure gradients. Thus, circulation would be most likely to occur near the edge of the lake where the topographic relief is also more extreme than near the center.

Stations 3 and 12 yielded heat-flow values of 0.98 and 1.16, respectively. These two stations are located near steep escarpments (Fig. 2) and near major inflows into the lake. Thus, depression of the heat-flow values can be expected to be due to continuous sedimentation or periodic slumping from the escarpments in the form of small turbidity currents. Slumping along the escarpments in the lake is relatively common (Hyne, 1969). Station 3 showed an irregular gradient, which may result either from bottom-water variations, circulation, or slumping and erosion. Turbidity currents operating near steep escarpments are capable of producing both periodic deposition and erosion in the upper few metres. The amount of time necessary to reach 90 percent equilibrium after 1 m has been deposited or eroded is approximately 10 yr (Von Herzen and Uyeda, 1963). The fact that these two stations give the lowest values suggests that deposition in these regions is presently going on at a more rapid rate than in the center of the lake, thereby reducing the true heat flow.

Stations 5, 7, 10, 14, and 16 are all located near the central part of the lake away from escarpments and primary sources of sedimentation (Fig. 2). Sediments in this region are uniformly laminated fine-grained muds. Heat-flow values for these five stations range from 1.52 to 1.69, with a mean of 1.62. Aside from the effect of uniform sedimentation, these values probably represent the most reliable determinations of the true heat flow for the Lake Tahoe basin. The estimates of Hyne and others (1972) of the sedimentation rates for two cores from the center of the lake fall in the range of 10 to 25 cm/1,000 yr. Table 4 summarizes the sedimentation effect (see Jaeger, 1965). From these data it would not be unreasonable to assume that the true heat flow is approximately 10 percent higher than the aforementioned mean, or about 1.8 HFU. This figure assumes a lake age of 1 to 2 m.y. (that is, sedimentation that has been uniform and continuous since formation of the lake). Thus, the combination of sedimentation and refraction could result in a corrected heat flow of as much as 2 HFU for Lake Tahoe.

Stations 6 and 9 yielded heat-flow values of 2.23 and 2.20, respectively. The nature of these high gradients is not known. Either refraction or heat sources must be responsible. Because heat-flow values of 1.6 to 1.7 have been determined from locations only a few kilometres away, the disturbance can be no more than a few kilometres below the sediment-water interface. Refraction alone would demand a very unusual buried bedrock configuration to produce the observed 25 percent increase in heat flow. Thus, localized heat sources such as hot springs, perhaps along a buried fault (which may also produce bedrock relief), appear more likely. Variations in radioactivity can be ruled out.

DISCUSSION

If it is accepted that the mean heat flow in the Lake Tahoe basin is at least 1.6 HFU, we can conclude that this basin's thermal regime is not typical of the Sierra Nevada. Assuming that heat flow in excess of typical Sierra values is due to radioactivity, we find

TABLE 4. EFFECT OF SEDIMENTATION ON HEAT FLOW IN LAKE TAHOE

R	$\kappa = 0.002$ $t = 0.1$ m.y.	$\kappa = 0.003$ $t = 0.1$ m.y.	$\kappa = 0.002$ $t = 1$ m.y.	$\kappa = 0.003$ $t = 1$ m.y.	$\kappa = 0.002$ $t = 2$ m.y.	$\kappa = 0.003$ $t = 2$ m.y.
1	1.00	1.00	1.00	1.00	1.01	1.00
2	1.00	1.00	1.01	1.01	1.01	1.01
4	1.01	1.00	1.02	1.01	1.03	1.02
8	1.01	1.01	1.04	1.03	1.06	1.04
16	1.03	1.02	1.08	1.06	1.11	1.09
32	1.05	1.04	1.16	1.13	1.23	1.17

Notes: Tabulated values are ratios of true heat flow to measured heat flow. Abbreviations are as follows: R = sedimentation rate in cm/1,000 yr; κ = thermal diffusivity in cm²/sec; and t = duration of sedimentation.

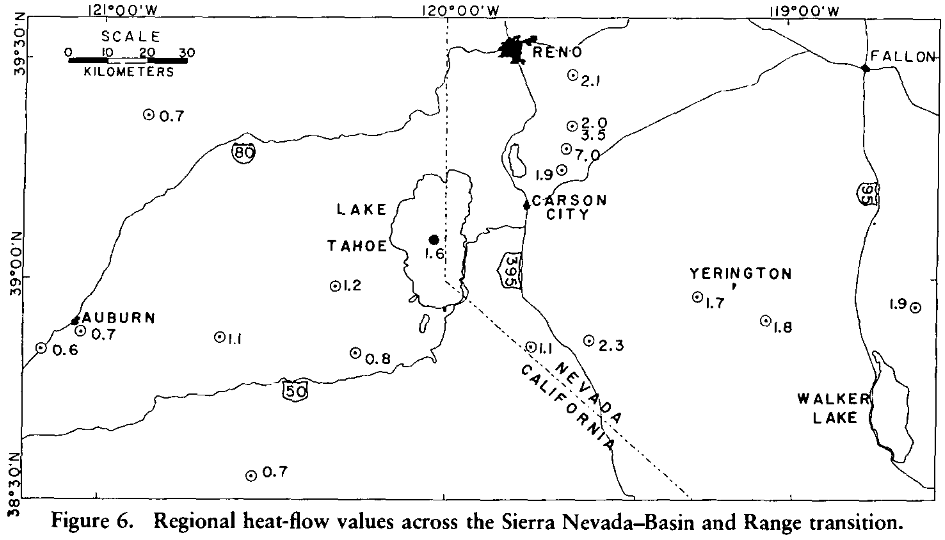

Figure 6. Regional heat-flow values across the Sierra Nevada–Basin and Range transition.

from Roy and others (1968b) that the granitic rocks of the Tahoe basin would have to average about 12×10^{-13} cal/cm^3 sec of heat production, a value which is unusually high for the Sierra Nevada (Roy and others, 1968b; Wollenberg and Smith, 1964, 1968). Furthermore, geologic mapping indicates that the regions around the lake are predominantly granodiorite and thus would be expected to have a lower radioactive content than the more radioactive rock of the Sierra Nevada (typically granite and monzonite). On the other hand, if we assume that the Tahoe basin is representative of Basin and Range structure, a radioactive content on the order of 2×10^{-13} cal/cm^3 sec (Roy and others, 1968b) would be sufficient to explain the total heat flow. This value is anomalously low for granodiorite. Thus, the heat flow for the Tahoe basin appears to be transitional between heat flows that are characteristic of the two provinces.

If we consider the heat-flow data west of Lake Tahoe (Roy and others, 1968a) and that to the east (Sass and others, 1971; Roy and others, 1968a), as shown in Figure 6, to be representative of the regional heat flow (the Gardnerville point of 1.1 HFU immediately to the east of the California-Nevada border is a questionable determination; D. Blackwell, 1971, oral commun.), the thermal transition from the Sierra Nevada to the Basin and Range province takes place over a distance of about 50 km, with the Lake Tahoe point within the transition area. If we assume a rate of heat production on the order of 7×10^{-13} cal/cm^3 sec (granodiorite-granite) for rocks beneath Lake Tahoe and a radioactive layer depth of 10 km (Roy and others, 1968b), a reduced heat flow of 0.9 HFU results for Lake Tahoe, which is intermediate between the Sierra Nevada and Basin and Range reduced values. Thus, we must explain the sharpness of the heat-flow transition, as well as the change from a region of unusually low heat flow (the Sierra Nevada with heat flows of about 0.4 HFU below the stable continental average) to a region of high heat flow (the Basin and Range province with heat flows of approximately 0.6 HFU above the stable continental average). These observations imply transient sources and (or) sinks whose depths cannot be too great — probably less than 100 km.

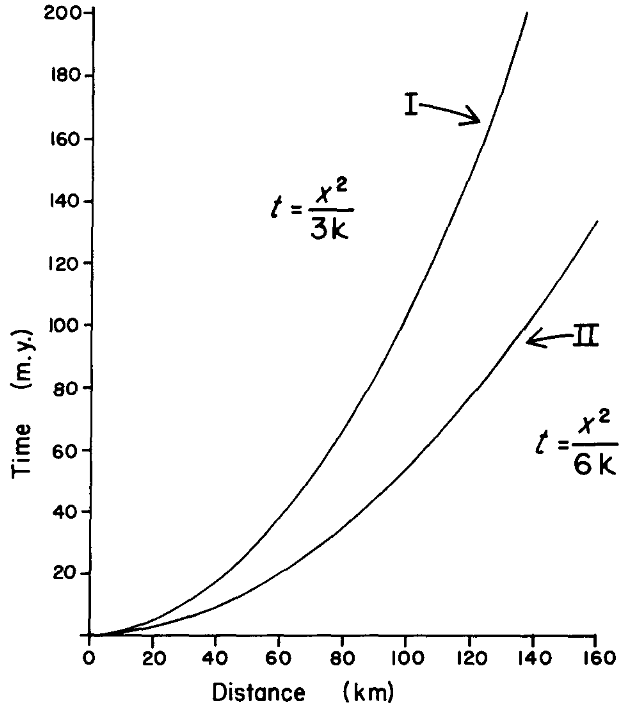

Figure 7. Examples of time t required for the heat flow at a distance x from a planar heat source to reach a maximum (k = 0.01 cgs). Case I: continuous source imbedded in an infinite medium; case II: instantaneous source at depth x below the surface of a semi-infinite medium with zero surface temperature. Heat flow observed at surface.

Because of the low thermal diffusivity of crustal rocks, a period of many millions of years is necessary for significant changes in the thermal regime of large crustal blocks to be reflected in the surface heat flows. An estimate of these times can be gained from consideration of the two curves in Figure 7, which give time constants for two hypothetical "thin-plate" heat-source models buried at depth. Superimposition of such planar sources to produce a three-dimensional buried source does not appreciably affect the values. Thus, the transient, subnormal heat flow in the Sierran crustal block (width, ~200 km; depth, ~50 km) probably reflects geologic processes occurring between 10 and 100 m.y. ago. The upper limit (that is, the first appearance of the subnormal thermal regime) is also constrained by the age of the batholith; that is, during emplacement of the batholith and shortly thereafter, high surface fluxes must have existed. If we assume formation of the batholith behind a Mesozoic trench (Atwater, 1970), the Sierran terrane would have been similar to present-day thermal-tectonic analogues such as Japan and the Sea of Japan (Hasebe and others, 1970), where, presumably, batholiths are currently being generated and fluxes are high. For the Sierra Nevada, plutons were being emplaced until Late Cretaceous time, about 80 m.y. B.P. (Evernden and Kistler, 1970). Thus, in order to examine the disequilibrium in thermal structure and the sharp transition between the Basin and Range and the Sierra Nevada provinces, we must consider tectonic events during the past 80 m.y. We will show how the transition from underthrusting tectonics to strike-slip tectonics along the western margin of North America is consistent with the observed thermal structure of the Sierra Nevada and Basin and Range provinces.

During Mesozoic time, magma generated

from an active eastward-dipping Benioff zone was emplaced as the Sierra Nevada batholith (Hamilton, 1969; Dickinson, 1970). A dip of about 50° is inferred from petrologic variations across the Sierra (Bateman and Dodge, 1970). The top of the underthrusting lithospheric slab was probably 100 to 200 km beneath the surface at the site of the present Sierra Nevada. During this time the heat flows above the still-buried batholith were probably above normal (Fig. 8A).

About 80 m.y. ago, plutonism virtually ceased in the region of the present Sierra Nevada. This cessation in magmatism and the subsequent reduction in surface heat flow can be explained if the position of the underthrusting slab became shallower beneath the Sierra, as suggested in Figure 8B. The steeply inclined portion of the subduction zone has been shifted eastward relative to the Sierra batholith and the leading edge of the continent, with shallowing of the underthrusting slab beneath the Sierra. Magma generation would have shifted eastward, and the top of the zone beneath the Sierra Nevada, which is now close to the surface (50 to 100 km deep), would have a relatively low dip. We propose that this new geometry could have resulted from an increased westward drift of the North American plate relative to the asthenosphere and, thus, also relative to the Pacific plate. In Figure 8b, we have referred to this process and the resultant lithospheric configuration as overriding. The work of several investigators makes the overriding mechanism appealing. Coney (1971) suggested an increase in westward drift of North America 80 m.y. ago. A gradual overriding during late Mesozoic time is also suggested by the "age belts" of Kistler and others (1971), in which the youngest belts of late Mesozoic age are found to the east. Emplacement of the youngest belt could have occurred after melting had ceased along the subducting slab, since a delay time is to be expected for magma traversing the mantle. A shallow Benioff zone during early and middle Cenozoic time is also consistent with the petrologic and chemical studies of Lipman and others (1972), who inferred a dip of 15° to 20° across the western United States. If it is assumed that much of the overridden intra-arc region was consumed by the trench (Page, 1970), then overriding is also consistent with the proximity today of Sierra Nevada batholithic rocks and the Franciscan trench assemblage along thrust faults in parts of California.

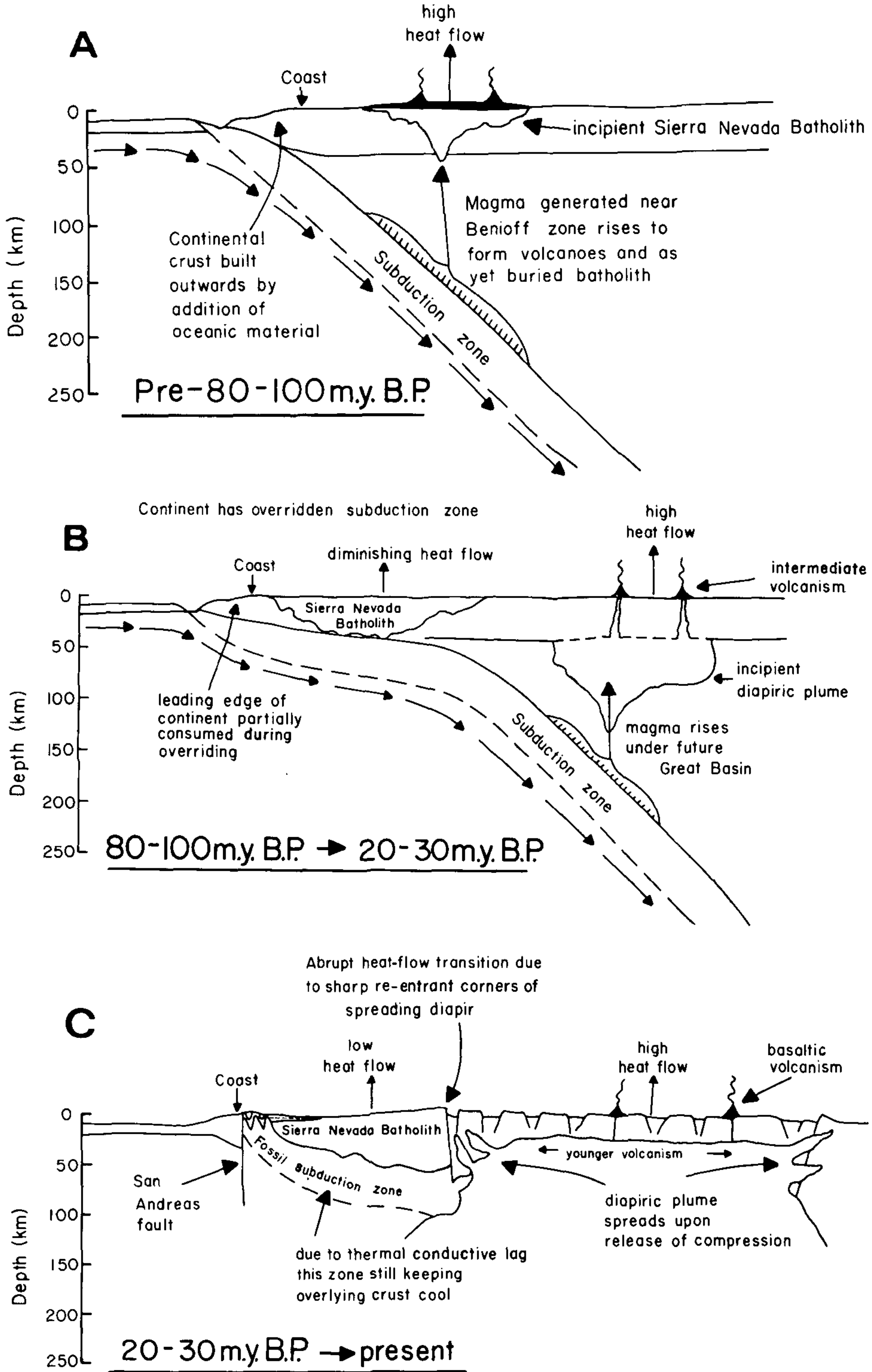

Figure 8. Schematic sequential representations of the Sierra Nevada–Basin and Range geothermal regime from the time of batholithic emplacement to the present.

If it can be assumed that the process of overriding as described above occurred about 80 m.y. B.P., then from latest Cretaceous through early and middle Cenozoic time, a "cold" slab was in contact with the base of the Sierran crust, and only with the termination of underthrusting about 30 m.y. B.P. (Atwater, 1970) did it tend toward thermal equilibrium with the surrounding mantle and overlying crust (Fig. 8C). In fact, owing to the thermal time lag, near-surface temperatures may not yet have reversed and begun to increase. The effect of a 100-km-thick cold slab whose top was 50 to 100 km beneath the Sierra [Lipman and others (1972) suggested a depth to the slab of 100 to 200 km] for a period of 40 to 80 m.y. (between 80 to 100 and 20 to 40 m.y. B.P.) would be to cause a general depression of isotherms in the crust adjacent to the slab. The extent of the depression of isotherms would spread with time, so that over the 40- to 80-m.y. interval, even the surface would be affected and the surface flux reduced accordingly. Figure 7 confirms that thermal time constants for lithospheric slab equilibration are likely to be 40 to 80 m.y. when slab depths of 60 to 100 km are involved.

Thermal models for downgoing slabs have been discussed by McKenzie (1969), Oxburgh and Turcotte (1970), Toksöz and others (1971), and Griggs (1972). McKenzie and Griggs restricted their discussions to temperatures within the slab by assuming uniform surrounding mantle temperatures. The justification for this assumption is that temperature changes in the surrounding mantle caused by the slab will in turn produce only second-order effects in the slab due to the thermal time constants vis-à-vis the descent velocity of the slab. Toksöz and others (1971) considered a more complete problem and demonstrated modification of isotherms in the region above the descending slab. In addition to the time scale, this model is highly dependent on the spreading velocity and accessory heat production (phase changes, shear heating, and radioactivity), both of which are not well known in western North America. Griggs (1972) has shown that the values of shear strain heating obtained by Toksöz and others (1971) are probably excessively high. Their calculations also do not consider appropriately long periods of time for the case of shallow-dipping slabs having present-day relative velocities of 4 to 8 cm/yr. It would be expected that the minimum heat flow over the trench would be significantly broadened and skewed in the downdip direction. Their model also does not explain the low heat flows found, for example, at a depth of 200 to 400 km behind the Tonga Trench, which would put the cold slab at a depth of 200 to 400 km below the surface in the region of low heat flow, thus making the cold slab model an unlikely cause for the low heat flow unless earlier dips were shallower than present ones.

Because the heat flow is below normal in the Sierra Nevada, the observed heat-flow values are not consistent with either an active spreading center having passed beneath it (Roy and others, 1972) or an active spreading center lying beneath the Basin and Range province in close proximity to the Sierras, as Kistler and others (1971) suggested. The sharp transition in heat flow from the Sierra Nevada to the Basin and Range province implies shallow heat sources and probably also implies (as does the subnormal heat flow in the Sierras) a geologically youthful transient thermal condition. Models of Basin and Range tectonics involving a subcrustal spreading center (for example, the East Pacific Rise), as proposed by Wilson (1970) or Kistler and others (1971), or simply representing a "soft" plate boundary responding to right-lateral shear on the San Andreas fault (Atwater, 1970) have been shown to lack an explanation of certain geologic occurrences (Scholz and others, 1971). The models also do not adequately explain the thermal pattern. Scholz and others (1971) developed the ideas of Karig (1970, 1971a, 1971b) and suggested that the Basin and Range province represents an ensialic interarc basin that developed in response to the relaxation of compressive stress along the western margin of North America at the time of cessation of subduction and initiation of strike-slip tectonics. High heat flow and volcanism are then a result of a diapiric plume, or numerous plumes, which had been trapped beneath the lithosphere during early and middle Cenozoic time and which spread laterally during late Cenozoic time. With the relaxation of compressive stress, these laterally spreading diapiric masses were able to produce crustal extension. The spreading produced forceful upward injection at the sharp re-entrant corners formed by the boundary between the region of outflowing diapirism and the adjoining lithosphere. This in turn would tend to produce very sharp petrologic, structural, and thermal transitions along the boundary between the leading edge of the spreading diapir and the adjoining lithosphere. Continuing this line of reasoning, the present Sierran block represents what was left of a much larger crustal block prior to nibbling away by the interarc spreading from the east; the presence of apparent Basin and Range structure within the batholith, as represented by the Tahoe-Truckee graben, may thus represent the effect of a plume from the main diapiric mass or perhaps some other asymmetry in the leading edge. Because of the recency of these events along the eastern Sierra Nevada, as evidenced by uplift and deformation (Bateman and Wahrhaftig, 1966), smearing of the thermal anomaly across this boundary would not yet have had time to occur; thus, the width of the transition largely reflects the depth of the disturbance.

CONCLUSIONS

Heat-flow measurements in Lake Tahoe, California-Nevada, have yielded values intermediate between those found in the Sierra Nevada and those from the Basin and Range province. On the basis of these new data and studies of earlier workers, we have constructed a model of crustal evolution of this region that is consistent, we believe, with present plate tectonics concepts of the western United States. Clearly, there will be objections to this model on geologic grounds by those studying the Sierra Nevada and Basin and Range provinces. However, we have attempted to demonstrate some of the critical constraints on crustal structure and evolution in this area imposed by heat-flow results, with the hope that those more in tune with the geologic relationships will be able to resolve the two sets of data.

We have suggested that the two characteristics of Sierra Nevada–Basin and Range heat flow that are of basic importance are (1) the subnormal heat flow in the Sierra Nevada adjacent to the above-normal heat flow in the Basin and Range province and (2) the sharpness of the thermal transition between the two provinces. In addition, we have shown that in the region of Lake Tahoe the transition occurs within the Sierra Nevada physiographic province. These basic relationships apparently require a combination of thermal sources and sinks to have existed here in the vicinity of the crust-mantle boundary at various times during the past 100 m.y. This requirement in turn implies the existence during this period of large-scale mass transport, which is currently most convincingly related to plate motions and interactions. Thus, within the framework of plate tectonics, we have suggested that the role of thermal sinks can be played by shallow portions of descending lithospheric slabs, whereas thermal sources displaying short wavelength variability in surface heat flux can be related to regional magmatic activity.

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