Tectonic implications of the heat flow of the western Snake River Plain, Idaho

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ABSTRACT

Heat-flow values within the western Snake River Plain average about 1.7 μcal/cm^2 sec, but even higher values are measured in granitic rocks along the margins of the Snake River Plain (2.5 μcal/cm^2 sec or higher). The heat-flow distribution is related to the combined effects of crustal thermal refraction and a large, transient crustal heat source. A regional model consistent with the heat-flow pattern and other geophysical and geological data is described which assumes the emplacement of a large heat source (mafic intrusion?) under the western Snake River Plain about 10 to 15 m.y. ago. An anomaly of about 0.3 μcal/cm^2 sec is predicted over the center of the heat source at the present time. The timing of the emplacement of the heat source corresponds with the age of voluminous silicic volcanism in the western Snake River Plain. A time-progressive thermal model is presented for the Snake River Plain which is consistent with the time progression of silicic volcanism. Based on the model, higher regional heat-flow values are predicted for the eastern Snake River Plain. Confirmation of the high regional heat-flow values is not possible in the bore holes available (200± m deep) because of regional circulation of cold ground water in the Snake Plain Aquifer. However, a close correlation between integrated crustal and upper mantle temperature and observed elevation changes along the axis of the Snake River Plain is strong support for the heat-flow model. The possibility of high heat flow in the eastern part of the Snake River Plain implies that the area may have significant geothermal potential in spite of the low surface heat flow. The regional aseismic warping observed in the eastern Snake River Plain can be interpreted as a thermal contraction phenomenon involving the crust and upper mantle.

INTRODUCTION

The Snake River Plain is one of the major volcano-tectonic features of western North America. The physiographic features of the Snake River Plain extend from easternmost Oregon through southern Idaho in a great arc to the Yellowstone Plateau at the northwestern tip of Wyoming. The Snake River Plain is distinguished from adjacent regions to the north and south by relatively lower elevation and surface relief. It is covered mainly by late Cenozoic volcanics and sediments. The western half of the Snake River Plain is a deep graben filled with sedimentary and volcanic rocks (Malde and Powers, 1962; Newton and Corcoran, 1963). The eastern part of the Snake River Plain may be a regional downwarp (Kirkham, 1931) that followed a time-transgressive episode of volcanism which progressed from southwest Idaho to Yellowstone Park (Armstrong and others, 1975). In southwestern Idaho, a second, primarily volcanic, trend extends from near the center of the Snake River Plain westward to the Brothers fault zone in central Oregon and to the Newbery volcano (Walker and others, 1967; Green and others, 1972).

The object of this paper is to discuss the tectonic significance of new heat-flow data in the western Snake River Plain and to present a thermal model that is consistent with geological and geophysical data. The more southerly volcanic trend which connects to the Brothers fault zone will not be discussed.

Potassium-argon dating of the Cenozoic silicic volcanics (mostly rhyolitic ash flows) of the Snake River Plain shows that the ages vary from 9 to 13 m.y. in western Idaho, 8 to 10 m.y. in central Idaho, and 4 to 5 m.y. in eastern Idaho (Armstrong and others, 1975). Rhyolitic ash flows in the Island Park–Yellowstone area range in age from 2.6 to 0.56 m.y. (Eaton and others, 1975).

The basalt volcanic activity begins at about the same time as the silicic volcanism, but continues in each area for a much longer time span. For example, young basaltic extrusives (≤ 10,000 yr old) occur all along the Snake River Plain. Similarly, a westward time progression of silicic volcanism (rhyolite to rhyodacite domes and related ash flows) is observed in eastern Oregon along the Brothers fault zone toward the Cascade Range (Walker, 1974; MacLeod and others, 1976; Bowen and others, 1976).

The northern border of the western Snake River Plain consists mainly of Cretaceous plutonic rocks of the Idaho batholith. Southwest of Boise, the southern border of the Snake River Plain is the Owyhee Mountains, which have a core of pre-Cenozoic sediments and granitic intrusive rocks overlain by Miocene basalt (McIntyre, 1972). Southeast of the Owyhee Mountains, the southern border becomes the late Cenozoic silicic volcanics and sediments of the Jarbidge Mountains which overlie older Cenozoic volcanics and sediments and Mesozoic granitic rocks.

Several possibilities have been suggested for the origin of the Snake River Plain. Hamilton and Myer (1966) suggested the formation of the Snake River Plain as a tensional rift. They used the seismic-refraction data of Hill and Pakiser (1963, 1966) as evidence for the lack of continental crust below the western portion of the Snake River Plain. In contrast, Taubeneck (1971) argued that the gravity and seismic data, considered in terms of surface geology and the distribution of granitic rocks, are consistent with the interpretation that includes a granitic layer in the crust of southwestern Idaho. Taubeneck considered dike intrustions and lateral faulting as the mechanism of the origin.

Morgan (1972) mentioned Yellowstone as an example of a continental expression of a plume, and Smith and Sbar (1974) dis-

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cussed the origin caused by westward relative motion of the North American plate over a "hot spot" in the mantle. Warner (1976) proposed that a large left-lateral rift is the origin and used 12 offset geologic features, each with a displacement of approximately 80 km, as evidence for his model. Prostka and Oriel (1975) pointed out the possibility of the Snake River Plain being a quadruple junction centered at Twin Falls with linear segments consisting of the east and west arms of the Snake River Plain, the Oregon rhyolite belts, and the center core of the basin and Range province.

Geophysical data relating to the Snake River Plain and its environs are scanty in spite of the tectonic importance of the feature. Hill and Pakiser (1966) interpreted a seismic-refraction profile across the western part of the Snake River Plain. Microearthquake studies have been carried out by Pennington and others (1974). Several publications deal with the gravity field (Hill, 1963; Mabey, 1976; Bonini, 1963; Mabey and others, 1974) and magnetic field (U.S. Geological Survey, 1971) of southern Idaho. An electrical resistivity profile along the eastern Snake River Plain was carried out by Zodhy and Stanley (1973), and more recently Stanley and others (1977) discussed the results of a magnetotelluric survey of the eastern Snake River Plain from the Island Park caldera to the Raft River area. Relevant heat-flow data include studies of Roy and others (1968a, 1968b, 1972), Blackwell (1969, 1974), Sass and others (1971, 1976), Blackwell and Robertson (1973), Bowen and Blackwell (1975), Urban and Diment (1975), and Morgan and others (1976). The basis for this paper, however, is a basic data report by Brott and others (1976).

GEOPHYSICAL DATA

Measured heat-flow values in southwest Idaho (Brott and others, 1976; Urban and Diment, 1975; and Blackwell and Brott, unpub. data) are listed in Table 1 and plotted on a map of southwestern Idaho in Figure 1. The heat-flow values will be referred to in this paper in units of HFU. In this paper, we use as units for heat flow 1×10^{-6} cal/cm^2-sec = 1 HFU; for thermal conductivity 1×10^{-3} cal/cm-sec-°C = 1 TCU; for heat generation 1×10^{-13} cal/cm^3-sec = 1 HGU. The transformations to SI units are 1 HFU = 41.87 mW

TABLE 1. HEAT-FLOW DATA IN SOUTHWEST IDAHO

North latitude	West longitude	Interval of gradient (m)	Geothermal gradient (°C/km)	Thermal conductivity millicalories/(cm-sec-°C)	Heat flow microcalories/$(cm^2$-sec) uncorr.	corr.	Reference
43°50.1"	116°14.7"	5-150	60.2 ± 0.2	6.57 ± .09	4.1	4.0	A
43°47.5′	115°50.8′	120-590	26.0 ± 0.0	7.23 ± .51	1.9	2.0	A
43°41.9′	115°41.4′	190-245	21.0 ± 1.1	8.71	1.7	1.9	A
43°39.4′	115°17.0′	10-88	25.7 ± 2.3	7.23 ± .25	1.7	1.7	C
43°33.5′	116°21.8′	63*	49.0(40.0 to 51.0)†	3.47	1.7 (1.4 to 1.8)		A
43°32.8′	116°17.9′	20-30	11.0 ± 1.0	4.73 ± 0.08	0.5		A
43°32.8′	115°26.6′	10-122	29.3 ± 2.1	8.31 ± .18	2.3	2.0	C
43°23.3′	116°14.1′	12-30	32.6 ± 4.3	3.40	1.1		A
43°23.1′	115°42.9′	10-117	28.1 ± 2.1	8.56 ± .07	2.2	2.3	C
43°19.2′	115°20.6′	95-150	46.7 ± 0.5	6.30 ± .05	3.0	2.9	A
43°17.8′	115°49.8′	58*	55.0(55.0 to 73.0)†	3.45	1.9(1.9 to 2.5)		A
43°15.6′	115°58.6′	160*	42.0(37.0 to 42.0)†	4.62	1.9(1.7 to 1.9)		A
43°16.2′	115°57.2′	145*	40.0(36.0 to 42.0)†	4.62	1.8(1.7 to 1.9)		A
43°14.8′	116°35.3′	45-125	51.0 ± 2.9	3.96	2.0		A
43°14.5′	116°18.6′	15-30	62.7 ± 3.5	4.76	3.0		A
43°15.5′	115°42.2′	520*	49.0(43.0 to 86.0)	3.49	1.7(1.5 to 3.0)		A
43°15.2′	115°32.7′	10-30	24.7 ± 2.0	5.47	1.4		A
43°12.8′	116°31.3′	100*	43.0(33.0 to 53.0)†	3.48	1.5(1.1 to 1.8)		A
43°13.1′	115°54.3′	90*	38.0(38.0 to 60.0)†	4.58	1.7(1.7 to 2.7)		A
43°12.3′	116°31.8′	358*	43.0(34.0 to 60.0)†	3.48	1.5(1.2 to 2.0)		A
43°12.0′	115°31.6′	83*	38.0(31.0 to 44.0)†	4.58	1.7(1.4 to 2.0)		A
43°11.3′	116°41.1′	252*	47	6.60	3.1		B
43°10.9′	115°54.5′	80*	51.0(51.0 to 113.0)†	3.29	1.7(1.7 to 3.7)		A
43°07.8′	115°49.8′	190*	66.0(66.0 to 91.0)†	2.68	1.8(1.8 to 2.4)		A
43°05.0′	115°42.0′	15-30	34.7 ± 4.1	4.46 ± .13	1.5		A
43°04.7′	115°39.7′	119*	62.0(44.0 to 81.0)†	2.72	1.7(1.2 to 2.2)		A
43°04.5′	115°39.2′	79*	63.0(63.0 to 82.0)†	2.69	1.7(1.7 to 2.2)		A
43°62.4′	116°17.2′	18-30	45.6 ± 1.26	3.65 ± .21	1.7		A
43°01.1′	116°47.8′	130-390	52.3 ± 0.9	5.16 ± .21	2.5	2.6	A
42°57.6′	116°16.1′	8-15	174.7 ± 11.6	2.29	4.0		
		18-32	124.6 ± 3.8	3.33	4.1		A
42°54.2′	116°04.6′	60*	54.0(46.0 to 90.0)†	2.96	1.6(1.4 to 2.7)		A
42°54.1′	116°04.6′	55*	57.0(44.0 to 112.0)†	2.96	1.7(1.3 to 3.3)		A
42°49.5′	115°59.5′	13-30	110.7 ± 11.8	3.01 ± .01	3.3		A
42°48.3′	116°24.3′	252*	31	7.0	2.2		B
42°44.3′	116°19.9′	13-20	153.9 ± 7.3	2.30	3.6	3.5	
		23-32	77.3 ± 1.1	4.32	3.5	3.3	A

Note: Heat-flow data in corrected column have been corrected for terrain effects. Where no value is shown in corrected column, the correction is less than 5%.

* The depth of the well measured instead of the depth interval.

† The best value and the lower and upper limits on gradient in cases where the gradient had large disturbances.

References: (A) Brott and others (1976). (B) Urban and Diment (1975). (C) Blackwell and Brott (unpub. heat-flow values).

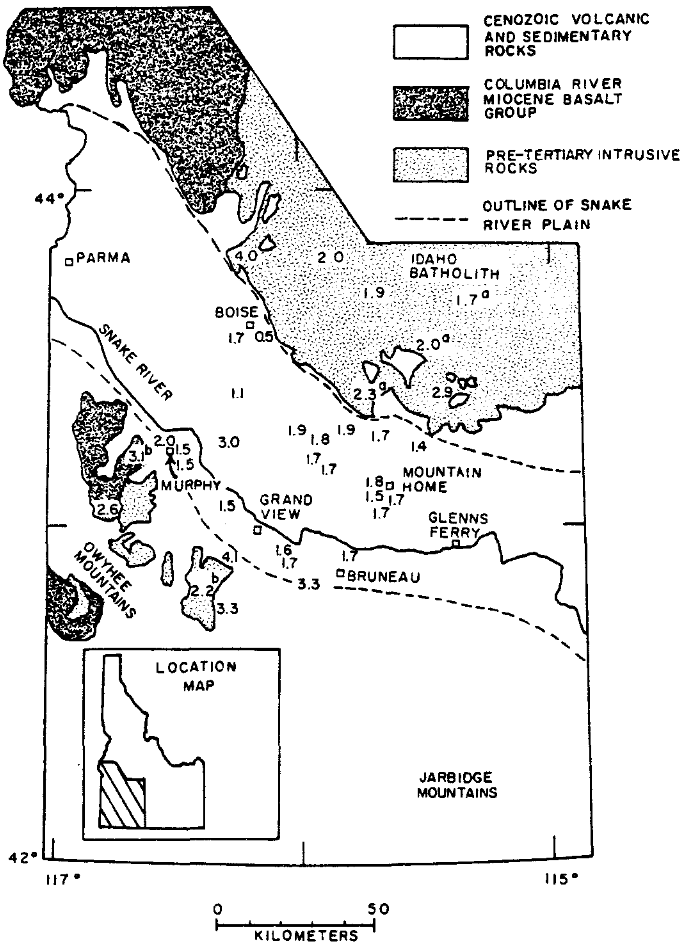

Figure 1. Generalized heat-flow map of southwestern Idaho. The heat-flow values with the superscript(^{a}) are preliminary values from Blackwell and Brott (unpub. data) and the values with the superscript (^{b}) are from Urban and Diment (1975). Other values are from Brott and others (1976). The geology is after Ross and Forrester (1947). The inset map shows the location of the study area with respect to the state of Idaho. The location of the heat-flow site is at the decimal point of the appropriate number.

m^{-2}, 1 TCU = 0.4187 W m^{-1} K^{-1} and 1 HGU = 0.4187 μW m^{-3}. Brott and others (1976) presented data from 136 wells in southern and central Idaho; the data shown in Table 1 represent those wells in southwestern Idaho for which reasonable estimates of heat flow can be made. Of the 35 heat-flow values listed, 18 are from preexisting holes in the depth range 60 to 590 m, mostly water wells, and 17 are from holes drilled specifically for heat-flow determinations. Eight of the 17 wells were drilled in granitic rocks on either margin of the Snake River Plain and range in depth from 60 to 250 m. The remaining seven wells were drilled to a depth of 30 m in the Cenozoic rocks of the Snake River Plain. Deeper holes near the 30-m holes corroborate the results of several of the shallow holes. In general, the heat-flow values in these shallow wells are considered more reliable than values from the deeper water wells because samples (core and cuttings) were available for thermal conductivity measurements. Heat-flow values in the western Snake River Plain range from 0.5 to 2.0 HFU except for one high value of 3.0 HFU. On the northern margin, the values range from 2.3 to 4.0 HFU, generally decreasing to the north to values ranging from 1.6 to 2.0 HFU. The heat-flow values on the southern margin range from 2.2 to 4.1 HFU.

Farther to the west and not shown on Figure 1, the heat-flow values are high (averaging 2.5 to 3.0 HFU) in the Oregon part of the western Snake River Basin (Bowen and Blackwell, 1975). The high heat flow there is probably due to refraction at the end of the deep sediment trough represented by the basin (see below).

Even though many of the heat-flow values in the western Snake River Plain were obtained from water wells, the reported heat-flow values are not believed to be significantly affected by lateral movements of ground water, because no discharge system for a major aquifer can be found in the western portion of the Snake River Plain. Also, the current recharge to aquifers in the western part is not sufficient to replace ground water extracted by wells (Ralston and Chapman, 1970; Rightmire and others, 1975). In addition, Rightmire and others (1975) studied hydrogen and oxygen isotopes in waters obtained along the southern margin and suggested that the water in the artesian aquifer system there accumulated over a long period of time. Hence the aquifer systems in the western part of the Snake River Plain act as large reservoirs in which little lateral movement of ground water appears to occur.

Mantle heat flow for the Cordilleran Thermal Anomaly Zone (CTAZ; Blackwell, 1969), which includes the Basin and Range, Columbia Plateaus (which include the Snake River Plain), and the Northern Rocky Mountains, is estimated to be about 1.4 HFU (Roy and others, 1968; Blackwell and Robertson, 1973; Blackwell, 1974). Swanberg and Blackwell (1973) showed that the average heat production for the granitic rocks of the Idaho batholith is 3.3 HGU (1 HGU = 10^{-13} cal/cm^3-sec) with a range along the southern margin of 2 to 5.5 HGU. Thus the expected heat flow for the margins of the western Snake River Plain would be about 1.7 ± 0.2 HFU. The observed values (2.5 HFU or greater) are therefore anomalous and suggest an anomaly in reduced heat flow.

A reversed seismic-refraction profile which extends north from Eureka, Nevada, across the western Snake River Plain to Boise, Idaho, was interpreted by Hill and Pakiser (1966) to suggest that the crust of the Basin and Range province has a 6.0 km/sec (P-wave) layer 19 to 24 km thick over a 6.7 km/sec layer 10 to 12 km thick. At the boundary of the Basin and Range province and the western Snake River Plain, the crust thickens abruptly, and the 6.0 km/sec layer becomes very thin or disappears completely. The crust under the western Snake River Plain consists of an upper 5.2 km/sec layer 8 to 10 km thick and a lower 6.7 km/sec layer 33 to 38 km thick. The velocity of the mantle (P_n) below the Basin and Range province and western Snake River Plain is approximately 7.9 km/sec.

A microearthquake survey in parts of the Snake River Plain was carried out by Pennington and others (1974); in three weeks of recording at various locations, no earthquakes were observed. They suggested two probable reasons: (1) the source that produced the recent basaltic lava has either become inactive or has migrated, or (2) that temperatures are high enough under the Snake River Plain to allow a seismic creep.

Bouguer gravity maps of the western Snake River Plain (Mabey and others, 1974; Bonini, 1963; Hill, 1963) show gravity highs in the center with relative lows along the margins. Thus the gravity data seem to indicate an inverse relationship to the heat-flow data. The gravity data suggest the presence of lower density rocks on the edges than in the center of the Snake River Plain. Greater sediment thickness may cause lower values of gravity on the margins, whereas the center of the Snake River Plain may have relatively more basalt. Two-dimensional models which fit the gravity profiles across the western Snake River Plain were constructed by Mabey (1976) and Hill (1963).

A residual aeromagnetic map shows a trend of magnetic highs on the southern margin and a trend of magnetic lows on the northern margin of the western Snake River Plain (U.S. Geological Survey aeromagnetic map, 1971; see also Mabey, 1976). The magnetic highs and lows seem to bound the lower values of heat flow characteristic of the center of the Snake River Plain, but the magnetic map does not show any distinct feature at the center. Mabey (1976) suggested that the relative positive magnetic anomaly on the southern margin and relative negative magnetic anomaly on the northern margin probably mark the south and north edges of a magnetic layer (basalt) which underlies the western Snake River Plain.

An electrical resistivity profile across the eastern Snake River Plain (from Arco to Blackfoot, Idaho) has been discussed by Zohdy and Stanley (1973). Under the center of the eastern Snake River Plain, they interpreted four geoelectrical units: an upper 300- to 500-m–thick unit of dry basalt, to a 1- to 2-km-thick unit of saturated basalt, a 2- to 4-km-thick unit of basalt intercalated with clayey sedimentary rocks, and a lower unit of Paleozoic (?) rocks. Near Arco they encountered a fifth geoelectrical unit that they interpreted as sedimentary rock and/or rhyolitic ash-flow tuff.

INTERPRETATION OF THE HEAT-FLOW DATA

The heat-flow pattern (Fig. 1) is basically symmetrical about the axis of the western Snake River Plain. Heat-flow values of about 1.7 HFU are found in the center, and heat flows of 2.5 HFU or greater are characteristic of the margins. At a distance of 35 km from the northern margin, heat flow drops to about 2.0 HFU. There are no heat-flow measurments farther than 30 km from the southern margin except values in north-central Nevada, 100 to 150 km away (Sass and others, 1976). These values are generally high, and the area is considered part of the Battle Mountain heat-flow anomaly (Sass and others, 1971, 1976).

Crustal Thermal Model

A two-dimensional crustal model of the western Snake River Plain consistent with geophysical and geological data is shown in Figure 2. It is based on the following data. The Cenozoic rock units

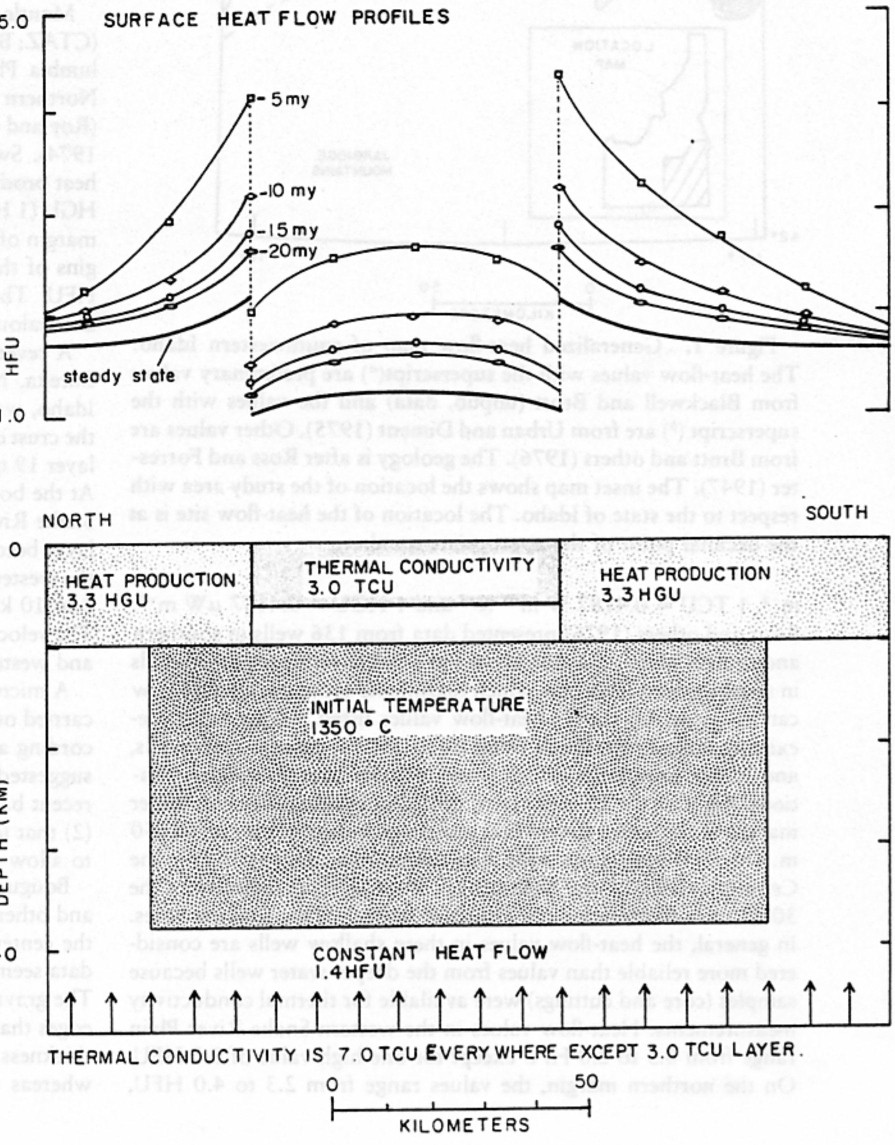

Figure 2. Heat-flow model and calculated heat-flow profiles for the western Snake River Plain. The low thermal-conductivity layer (3.0 TCU) is interpreted as Cenozoic volcanics and sediments. The heat-production layers (3.3 HGU) are interpreted as upper crustal granitic rocks on the margins of the Snake River Plain. The block with initial temperature of 1350 °C is interpreted as a large thermal event, possibly a large mafic intrusion. The "thermal event" is probably the result of a complex sequence of intrusions which were intruded within a short period of time. The constant heat flow (1.4 HFU) at the lower boundary of the model is the assumed mantle heat flow. Heat-flow profiles at 5, 10, 15, and 20 m.y. after the emplacement of the thermal source and a steady-state profile of the heat flow produced by the heat-production layer and constant heat flow at the lower boundary are shown above the model.

in the Snake River Plain are mainly sediments, vesicular basalts, and silicic ash flows (Malde and Powers, 1962). The porosities of most rock units are high, and mineral thermal conductivities are low to moderate with observed values of 4 ± 1, 3 ± 1, and 4.5 ± 0.3 TCU (mcal/cm-sec-°C) for the basalts, sediments, and silicic volcanics, respectively. The basalts and sedimentary units are represented by a generalized 6-km-thick layer of low thermal conductivity (3.0 TCU) which is 60 km wide. Welded tuffs, nonvesicular basalt and/or the thin remaining granitic crust are interpreted as a 4-km layer of higher thermal conductivity. The thermal conductivity measurements of the granitic cores from the Idaho batholith range between 6 and 7 TCU (Brott and others, 1976), and so the assumed thermal conductivity for the rest of the model is 7 TCU. Swanberg and Blackwell (1973) determined the average heat production of the Idaho batholith to be 3.3 HGU. The two 10-km-thick layers with heat production of 3.3 HGU on each side of the layer of low thermal conductivity are interpreted as normal crust on the borders of the western Snake River Plain. The heat production of the Cenozoic rocks in the Snake River Plain is probably less than 1 HGU and is ignored in this model.

A regional thermal anomaly is postulated to be related to disruption of the continental crust during formation of the Snake River Plain. The thermal anomaly is modeled as a large mafic intrusion emplaced during late Cenozoic time. The size of the mafic intrusion was determined from seismic and geological data. Hill and Pakiser (1966) interpreted seismic data to indicate that the lower crustal unit with P-wave velocities of 6.7 km/sec under the Snake River Plain is about 20 km thicker than the lower crustal unit with similar velocities under the Basin and Range province and that the top of the lower crust is about 10 km deep below the western Snake River Plain. This drastic change in crustal structure is assumed to indicate major crustal disruption and modification and is assumed to have originated during the formation of the western Snake River Plain tectonic feature. So the whole lower crust (28 km thick) is assumed to be a mafic mass which is acting as a heat source. The southern end of the thickened lower crust was located 66 km from the center of the Snake River Plain by the seismic studies. The northern edge of the intrusion was determined from geological observations of source areas of late Cenozoic basaltic lava flows ≈ 20 km north of the northern margin of the Snake River Plain (Howard and Shervais, 1974; Blackwell and Brott, unpub. data). These basaltic lava flows are assumed to mark the northern edge of the intrusion (50 km from the center of the Snake River Plain). The size of the body is very large, but the intrusion probably occurred as several smaller events, and the "intrusion" is probably a mixture of refractory crustal material and injected material of basaltic composition. The whole lower crust is assumed to have been heated to the melting point of basalt throughout, during this thermal event.

Heat-Flow Model and Results

The mafic intrusion is thermally modeled as an instantaneous heat source (10 km below the surface, 28 km thick, and 116 km wide) with an initial temperature of 1350 °C (see Fig. 2). The initial temperature of the heat source is based on a 1050 °C melting temperature for a molten mafic intrusion plus an additional 300 °C to allow for the latent heat of melting (see Jaeger, 1965). The actual *in situ* initial temperature of the body is the difference between the temperature at some point from the steady-state model and 1350 °C. The mantle heat flow is represented as a constant heat flow of 1.4 HFU into the base of the model.

The solution of the model (Fig. 2) was accomplished in two parts. First, a steady-state solution was calculated for the model without the mafic intrusion. The heat-flow profile of the steady-state solution is shown in Figures 2 and 4 (1-SS). Second, the instantaneous heat source was placed in the steady-state solution (the initial condition). Time transient solutions were obtained for various times after emplacement of the mafic intrusion and the heat-flow profiles for 5, 10, 15, and 20 m.y. solutions are shown in Figure 2. A finite difference program was used to obtain the solutions. Symmetry was assumed about the center of the Snake River Plain; the model was divided into two parts and the northern and southern halves of the model were calculated separately.

The calculated heat-flow profiles (Fig. 2) which are most consistent with the observed heat-flow data (Fig. 1) are the 10 and 15 m.y. profiles. These profiles show that the heat flow ranges from 1.3 to 2.0 HFU in the Snake River Plain and 2.8 to 3.3 HFU on the margins. These two ages approximately bracket the age range of

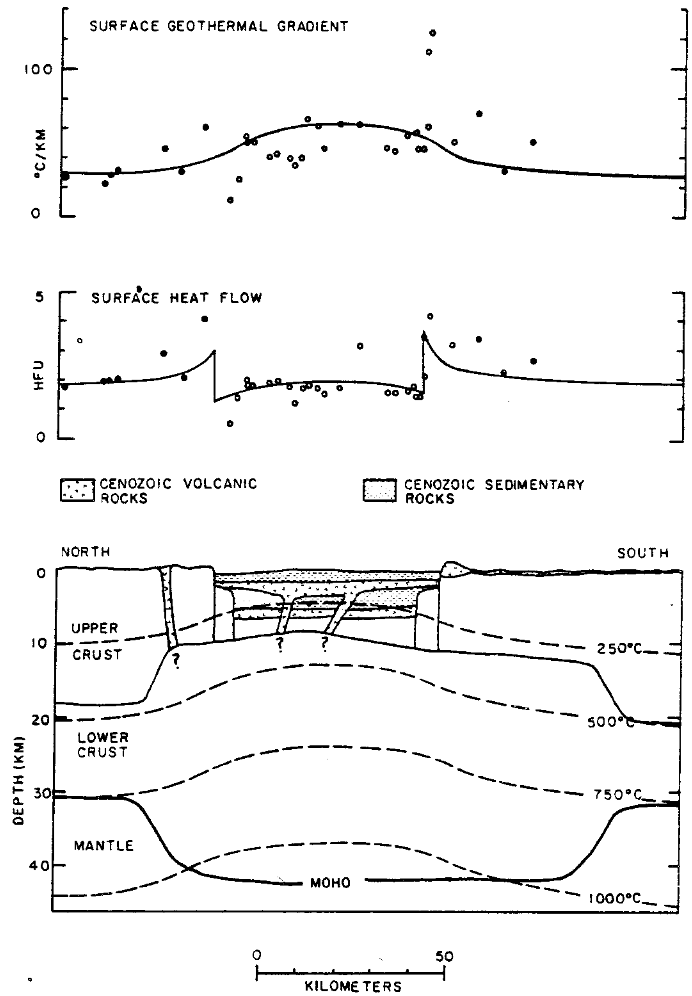

Figure 3. Diagrammatic cross section of the western Snake River Plain, including geothermal-gradient and heat-flow profiles. The observed values for the gradient and heat flow are circles and the solid lines on the profiles are from the 12.5 m.y. solution of the model in Figure 2. 250, 500, 750, and 1000 °C isotherms are shown on the cross section (obtained from the 12.5 m.y. solution). The Cenozoic volcanic rocks shown in the section include an unspecified proportion of basalt versus silicic ash-flow units. The open circles are measured in volcanic and sedimentary rocks and the solid circles are measured in granitic rocks.

the silicic volcanics in this part of the western Snake River Plain (9 to 13 m.y., Armstrong and others, 1975).

The calculated heat-flow patterns are most sensitive to the geometry of the wedge of low thermal conductivity material along the center of the Snake River Plain, to the geometry of the upper surface of the heat source and to the time of emplacement of the heat source. The patterns are relatively insensitive to the thickness of the thermal source.

A diagrammatic cross section across the western Snake River Plain based on the thermal model (Fig. 2) is shown in Figure 3. Figure 3 is an attempt to put geologic flesh on the bones of the thermal model in Figure 2. The 250, 500, 750, and 1000 °C isotherms obtained from the 12.5 m.y. solution are shown on the cross section. Indicated on the cross section is the thickening of the lower crust and the thinning of the upper crust below the western Snake River Plain as compared to the Basin and Range province. The Cenozoic volcanic and sedimentary rocks are shown to be about 6 km thick in the Snake River Plain. The 4-km-thick layer under the Cenozoic rocks may be pre-Cenozoic upper crust or Cenozoic igneous units. The granitic rocks of the Idaho batholith and Owyhee Mountains are identified with the upper 10 km of the "upper crust" on either side of the Snake River Plain.

The 6-km depth to the bottom of volcanic and sedimentary rocks is based on the seismic-refraction data. The upper 3 km of rocks are known from the results of deep drilling. Near the center of the western Snake River Plain along the Idaho-Oregon border, a well was drilled to a depth in excess of 3.5 km and remained in Cenozoic rocks to the bottom (Bowen and Blackwell, 1975). Only about 600 m of Columbia River (Owyhee) basalt was encountered in the well (beginning at a depth of 2 km below the surface). The well included about 1 km total thickness of basalt with the remainder of the section composed of sedimentary and silicic-volcanic rocks. Thus "Cenozoic volcanic rocks" of Figure 3 include an unknown proportion of basalt to silicic volcanic rocks.

Observed geothermal gradients and heat-flow values are plotted in Figure 3 with curves showing the surface-gradient and heat-flow profiles from the 12.5 m.y. model solution. In general, the correspondence between the calculated and observed gradients and heat flow is good. The heat-flow curve shows less scatter because the thermal conductivity in the different wells is not as uniform as assumed in the model of Figure 2; however, the continuous gradient curve illustrates that the discontinuity in model heat flow at the Snake River Plain margins is due to the thermal conductivity contrast at the surface in the mathematical model.

Other Heat-Flow Models

The observed heat-flow pattern is somewhat unexpected because the Snake River Plain, the major tectonic and volcanic feature, has lower heat flow than its margins. In this section, we will consider some alternative explanations of the heat-flow data to the model discussed above. The low (relative) values of heat flow in the center of the Snake River Plain cannot be caused by heat transfer in a major aquifer system because such a system does not exist. Local circulation effects are probably responsible for some of the scatter in heat-flow values, but no regional effect is present.

Lachenbruch and Sass (1977) have pointed out that the residual thermal effects of large-scale aquifer motions, such as exist in the eastern Snake River Plain at the present time, would last for a long time after flow stopped in the aquifer. If a similar aquifer has been present in the western Snake River Plain in the past, the low (relative) heat flow might be due to these residual effects. The various volcanic units are the only likely candidates for such widespread water movement. The "Banbury" basalt, however, is at least 6 m.y. old, and any regional circulation would have to have ceased 4 to 5 m.y. ago, as the Banbury is deeply buried beneath sediments in the center of the Snake River Plain. Thus it seems unlikely that the low heat flow can be related to past aquifer circulation.

The western Snake River Plain is underlain by a thick sequence of sedimentary and volcanic rocks, and it might be supposed that the heat flow at the surface is artificially low because of the sedimentation effect. However, it is doubtful that the classical models of the thermal effect of sedimentation (Benfield, 1949) apply to the western Snake River Plain. The extensive volcanism and probable intrusive activity with induced extensive hydrothermal convection and the extensive upward flow of water expelled during compaction of the sediments have an unknown effect on the history of thermal regime in the basin. Because of the uncertainty of the boundary conditions, thermal effects of the many possible assumptions, and so on, no sedimentation correction has been applied to the data.

A more serious uncertainty in the thermal model is the radioactive heat generation of the upper crust beneath the Snake River Plain. If a granitic crust similar to the Idaho batholith underlies the rift, then the predicted steady-state heat flow would be about 1.4 HFU (1.4 HFU mantle heat flow plus about 0.3 HFU from crustal radioactive heat generation minus 0.3 HFU for the refraction effect), and no marginal anomaly except for a refraction anomaly would be present. This model is shown in Figure 4, which includes an expanded version of the two theoretical heat-flow curves from Figure 2. Model 1-SS is the steady-state heat flow from the model illustrated in Figure 2, excluding the transient heat flow from the mafic body and assuming that the graben has been present for a long time. Model 1 is the 12.5 m.y. curve from the model shown in Figure 2. Model 2-SS is similar to 1-SS with the addition of a block of material 10 km thick with a heat generation of 3.3 HGU below the low thermal conductivity unit extending to a depth of 6 km in the Snake River Plain. Clearly, Model 2-SS does not fit the observed heat-flow data as well as Model 1. Also, Model 2-SS does not include a deep heat source beneath the Snake River Plain and does not satisfy the systematic decrease in heat flow in the granitic rocks over a distance of 30 km or more away from the margin.

We believe that the geological and geophysical data support a model similar to Model 1 as opposed to a model such as Model 2-SS. It is difficult to imagine the formation of the Snake River Plain, extensive silicic volcanism, and the complete modification of the lower crust without a significant thermal event of the type suggested here. The effect of some amount of crustal radioactivity would modify the observed surface heat flow in the opposite direction from the sedimentation effect. Thus these two effects may cancel each other to first order, and this possibility is an additional justification for exclusion of both effects in Model 1.

DISCUSSION

Western Snake River Plain Model

The model presented in Figures 2 and 3 is a very simplified one, but it is consistent with most of the pertinent data. The western Snake River Plain is modeled as a deep trough in the continental

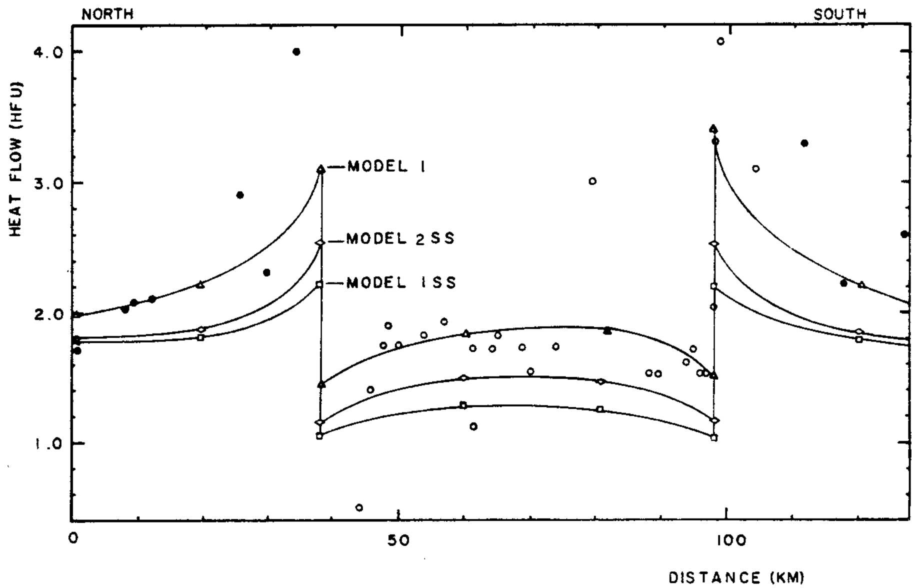

Figure 4. Three thermal models for the western Snake River Plain. Model 1 is the 12.5 m.y. solution for the model shown in Figure 2. Model 1-SS is the steady-state solution from Figure 2. Model 2-SS does not include the transient effects of the mafic intrusion but does include a 10-km-thick 3.3 HGU layer below the graben. The observed heat-flow data are also shown. The circles are observed heat-flow values in sedimentary and volcanic rocks, and the solid circles are heat-flow values in granitic rocks.

crust which originated 10 to 15 m.y. ago associated with a crustal-scale thermal event. The trough is filled with late Cenozoic sedimentary and volcanic materials. The age of the thermal event as determined from the models is in good agreement with the age of the silicic volcanism in the western Snake River Plain. If the silicic volcanic rocks are related to large-scale partial melting of the lower crust during formation of the rift and/or emplacement and differentiation of the thermal source (possibly a large mafic intrusion), the thermal model is in remarkable agreement with the geological data.

In this model, the highest regional heat-flow values are due to refraction of heat into the higher conductivity margins of the trough. This effect is clearly illustrated by the two anomaly curves plotted in Figure 3 (see also Fig. 4). The highest gradients are in the center of the Snake River Plain, but the highest heat-flow values are at the margins in the granitic rocks. The highest heat-flow values predicted by this regional model are about 3.0 HFU and are located in the granite where it contacts the low-conductivity sediments and volcanic rocks. The heat-flow distribution along the margins of the Snake River Plain is similar in magnitude and width to the strip anomaly identified by Reiter and others (1975) along the western margin of the Rio Grande Rift. Possibly the regional anomaly associated with the Rio Grande Rift is also primarily due to the combined effects of a deep heat source and to refraction on a crustal scale rather than presence of many local heat sources.

Several values much in excess of the predicted heat flow are found along the southern (2) and northern (1) margins and one value of 3.0 HFU is found in the center of the Snake River Plain. The regional model presented predicts temperatures in excess of 250 °C within the sediments and volcanics at depth of 4 to 5 km; it is likely that where suitable structures exist, hydrothermal convection will occur, resulting in very high local heat-flow values. Ross (1971) describes many thermal wells and springs in southern Idaho, and many of these are concentrated along the margins of the Snake River Plain where major faults occur. Specific examples of this type of feature are the Cow Hollow anomaly near Vale, Oregon (Bowen and Blackwell, 1975), and the Boise Front geothermal system (Applegate and others, 1975). Although at any one point along the Snake River Plain silicic volcanism has in general been limited in time, basaltic volcanism has continued over a much longer period of time. Some shallow heat sources, therefore, may also be present, associated with the very young basalts which are common in the Snake River Plain. The model presented is a regional one and does not attempt to include these local, but geothermally significant, anomalies.

High heat-flow values are also found at the west end of the western Snake River Plain in Oregon (Bowen and Blackwell, 1975). The high heat flow is due to refraction at the end of the basin as described here for the north and south margins, and by hydrothermal convection associated with faulting. As is the case for the areas described here, the possible local influence of young (< 17 m.y. old) igneous activity on the heat flow is as yet unknown.

Eastern Snake River Plain

The geological and geophysical characteristics of the eastern and western Snake River Plain are somewhat different (see the discussion in a previous section). The question of importance for this discussion is whether or not these differences imply a different thermal mechanism of origin. For this discussion, we will assume that they do *not*. Thus if the major thermal event in the western Snake River Plain coincides with the silicic volcanism, then this thermal event may be present in successively younger stages in the eastern Snake River Plain and the Yellowstone region. In this case, the heat-flow curves of different ages in Figure 2 might be related to locations of equivalent volcanic age in the eastern Snake River Plain.

Brott and others (1976) found low heat-flow values in the center (less than 1.0 HFU) and high heat-flow values on the southern margin of the eastern Snake River Plain (5.0 HFU or higher near Rexburg, Idaho, and 3.0 HFU or higher near Burley, Idaho). The low heat-flow values in the eastern Snake River Plain were obtained from wells located in and above the Snake Plain Aquifer. The discharge of the Snake Plain Aquifer (with a coefficient of transmissibility of 1 to 173 million liters per day per metre) is 185 kilolitres per second and occurs primarily at Thousand Springs located near Hagerman, Idaho (Mundorff and others, 1964). The low values of surface heat flow in the eastern Snake River Plain are caused by transport of heat laterally through the aquifer by rapid water flow. Deep wells which penetrate through the aquifer are needed to obtain reliable heat-flow values for the eastern Snake River Plain. As the heat-flow data are presently insufficient to determine the regional thermal pattern in the eastern Snake River Plain, we must look for other evidence of the thermal nature of the eastern relative to the western Snake River Plain.

Systematic elevation variations along the Snake River Plain might relate to the thermal structure, because in the oceans, water depth (that is, the elevation of the ocean bottom above some reference level) is related to the amount of cooling (thermal contraction) that has occurred after the particular piece of ocean lithosphere has left the ridge crest where it was formed (see Parsons and Sclater, 1977, for a summary discussion). The oceanic heat-flow data are consistent with this cooling lithosphere model except near the ridge crest where hydrothermal circulation dominates and the conduction pattern is unrecognizable. The thermal model

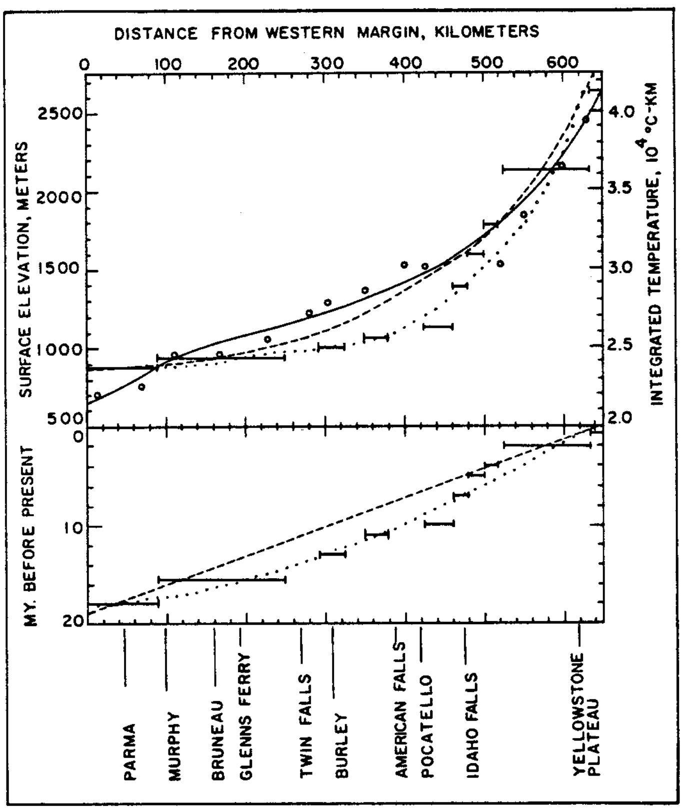

Figure 5. Elevation, predicted integrated-temperature, and age for the Snake River Plain. A third-order polynomial fit to the observed elevation (circles) is shown as the solid line. The integrated-temperature model which corresponds with the ages of the silicic volcanics is shown as the horizontal lines connected by a dotted curve. The integrated-temperature model assuming constant velocity of the thermal event is shown by the dashed curve. The corresponding ages for the two integrated-temperature models are shown in the lower graph, and the positions of several towns and Yellowstone are shown across the bottom. In the bottom part of the figure, the dashed line is the age assuming constant velocity (3.5 cm/yr) and the dotted line is a third-order polynomial fitted to the observed ages of the silicic volcanics.

(McKenzie, 1967) is a moving medium solution for a slab of given thickness moving through a plane of constant temperature and cooling off to a linear temperature increase with depth at great distance from the source. The solution can also be applied in the case of the movement of a constant temperature plane through a fixed medium. Such a model might apply to the Snake River Plain if the thermal event is viewed as heating the whole crust and upper mantle to the melting temperature, approximately a constant.

The elevation characteristic of the central 20 to 40 km of the Snake River Plain is plotted in Figure 5 along a line through the center of the Snake River Plain from west to east. A systematic west to east increase in elevation is clear. Elevation points are plotted at every 150 m of height difference in the western two-thirds, and every 300 m of height difference in the eastern one-third, of the Snake River Plain. The data indicate a gradual increase in elevation from about 750 m at the west end of the profile near the Oregon-Idaho border to just over 1,500 m near Idaho Falls, a distance of about 500 km. Over the last 130 km, the elevation rises more rapidly to an average elevation of almost 2,500 m in Yellowstone National Park. The heavy line drawn through the points is a third-order polynomial fitted to the elevation points. In general, the points below the line are areas where the Snake River has cut an unusually broad canyon, and the areas above the line are areas where there are thicker sections of young basalt volcanics. We suggest that this systematic change in elevation is *primarily related to the thermal expansion and thermal thinning of the lithosphere* in a similar manner to the elevation-age relationships observed in the oceans.

The thermal model described in the previous sections for the western Snake River Plain is of course far from unique, and other models could be devised to fit the data. Especially at short times, it cannot be taken literally. The initial conditions assume that the graben has been in existence for a long time before the emplacement of the thermal source. In the actual situation, if the Yellowstone system is the model for the first few million years of the system, there must be a significant period of time (several hundred thousand years) before the graben develops, and the thermal effect of the rhyolitic volcanics, hydrothermal convection, and so on, are insignificant.

Thus, in order to have a thermal model to compare to the elevation data, a simpler oceanic type of model was constructed. Details of the model are given by McKenzie (1967). A 42-km-thick lithosphere at an initial temperature typical of that for the background Basin and Range value (mantle heat flow of 1.4 HFU; Blackwell, 1978) was used. A velocity of motion of 3.5 cm/yr was used in the calculation to relate distance to age (Armstrong and others, 1975). Using this model, the temperature was integrated from 0 to 42 km for various points along the Snake River Plain. Values of integrated temperature are used instead of surface heat flow because the integrated-temperature values are less sensitive to the details of the structure and heat-transfer mechanisms of the upper crust. The integrated-temperature values and surface heat flow are directly related in the model, hence a high integrated-temperature value indicates a high predicted heat flow value. The observed surface heat-flow values in the eastern Snake River Plain are low because of lateral transfer of heat due to the rapid movement of ground water in the Snake Plain Aquifer; but the aquifer system only affects the temperatures of the upper few kilometres of the crust and has little effect on the integrated-temperature values.

Included in Figure 5 are integrated-temperature values as a function of distance from the McKenzie model based on two different assumptions. The integrated-temperature data have been scaled so that the integrated-temperature and elevation curves can be directly compared.

The first integrated-temperature versus distance model is shown as the heavy horizontal line segments and the dotted line. The horizontal location of a particular integrated-temperature value was determined from the time of the onset of silicic volcanism at any position along the Snake River Plain (Armstrong and others, 1975). The approximate locations of various ages of volcanism are shown along the bottom part of Figure 5 by heavy horizontal lines. The dotted line in the bottom part of the figure is a third-order polynomial fitted to the observed ages. The dotted line in the upper part of Figure 5 is the integrated temperature which corresponds to the ages along the dotted line in the lower part of the figure. The integrated-temperature values at the various times from the dates of silicic volcanism were obtained from the McKenzie model and are plotted over the same range of lateral distance as the exposure of a particular dated volcanic unit (see Armstrong and others, 1975, Fig. 2). The dashed line in the upper graph is integrated temperature as a function of distance assuming that the thermal event moved eastward at a constant velocity (3.5 cm/yr) beginning at 18 m.y. ago in Oregon and arriving at the eastern edge of Yellowstone 2 m.y. ago. Ages corresponding to the dashed integrated-temperature curve are also shown at the bottom of the Figure 5 as a dashed line, and the locations of several towns along the section are shown below the figure. The integrated-temperature curve based on the ages of silicic volcanism rises less steeply than the elevation in the west, but at approximately the position of Idaho Falls, it begins to rise more steeply than the elevation. Both the elevation and integrated-temperature curves show an inflection point near the position of Idaho Falls.

Such a use of the age data is perhaps too literal, however. The sources of most of the dated tuffs are not known because the source areas are buried beneath the younger sediments and basalts in the center of the Snake River Plain. Furthermore, older or younger silicic units may once have been present at any place along the system, but may not be exposed because of younger cover or because they have been eroded away.

As an alternative to a literal interpretation of the volcanic age data, a second integrated-temperature model is also presented. This second model is based on the assumption that the eastward progress of the thermal event was more uniform than indicated by the age data. The integrated temperature in this model increases smoothly from west to east (dashed curve in Fig. 5). This integrated-temperature curve could be used to predict the elevation at any point along the Snake River Plain within 200 m over an elevation range of 1,700 m. It does not, however, show the inflection point in the vicinity of Idaho Falls that both the elevation and the first integrated-temperature models show.

Correlations between elevation, integrated temperature, and heat flow have been well established for ocean spreading systems (Parsons and Sclater, 1977). Although correlations between heat flow, integrated temperature, and elevation for continents must exist in some form or the other, their elucidation has proved difficult, and the example suggested here probably represents the clearest example of such a relationship so far suggested. In the Snake River Plain, a surface elevation difference of about 1,700 m correlates with a predicted heat-flow variation from about 1.7 HFU to 3+ HFU and an integrated-temperature variation of 2.4×10^4 to 4.6×10^4 °C-km.

The coefficient of linear thermal expansion which would scale the integrated-temperature curve to the elevation difference is 7.7×10^{-5} cm/cm-°C. This value cannot be taken too literally, however, because of the simplicity of the model. It is clear that an empirical relationship between elevation and predicted heat flow derived for the Snake River Plain cannot be directly used for all of the Western United States. In fact, the western Snake River Plain lies at a lower elevation than the Idaho Batholith to the north and the Owyhee Mountains to the south, even though temperatures are higher. The main reason for the lack of a simple correlation is that thermal events typical of the western United States usually result in an increase in over-all crust–upper mantle density and resultant lower elevations after cooling than existed before the thermal event. For example, the crust in the Basin and Range province is thinner than to the east in the Colorado Plateau or to the west in the Sierra Nevada Mountains.

Furthermore, on continents erosion and sedimentation may modify simple heat-flow (integrated-temperature)–elevation relationships. For example, the top of the silicic volcanic sequence is exposed in Yellowstone and is buried more than 2 km below the surface in the western Snake River Plain (Bowen and Blackwell, 1975), implying a total elevation change of up to 3.7 km. This elevation difference is too much to be associated solely with thermal expansion and phase changes in a 50-km-thick section of the Earth's outer surface. Obviously as the surface subsides, deposition of sedimentary and additional volcanic rocks into the depression causes additional subsidence. In view of the simplicity of the thermal model, it does not seem profitable to attempt to draw quantitative crustal-structure, heat-flow, and elevation relationships at this time, as has been done for the oceans. As further geophysical and geothermal data are obtained, the implications of the elevation–heat-flow model can be investigated.

The only other available geophysical data that bear on the progressive thermal model are the deep-resistivity profiles obtained from magnetotelluric soundings (Stanley and others, 1977). These data indicate a systematic eastward decrease in the electrical resistivity of the crust from near Pocatello to Yellowstone with the important exception of the Island Park area which has anomalously high electrical resistivity at depth. Thus these data are in general consistent with the progressive increase in average crustal temperature from west to east in the eastern Snake River Plain.

Implications of Heat-Flow Model

The regional thermal anomaly in the western Snake River Plain can be related to a major crustal tectonic and thermal event associated with its formation. Similar heat-flow anomalies due to major crustal heat-flow refraction and emplacement of high temperatures in the lower and middle crust may exist in the western United States. Examples are the Battle Mountain heat-flow "high" (Sass and others, 1971) and the Rio Grande Rift (Reiter and others, 1975).

Because of the crustal heat-flow anomalies which dominate the surface heat-flow patterns, any mantle heat-flow anomaly from a mantle hot spot (Smith and Shar, 1974) or other deep-seated mantle effect which might underlie the Snake River Plain cannot be identified with prsently available data.

A very interesting regional elevation versus heat-flow relation for the Snake River Plain is inferred from the observed elevation differences and silicic volcanic-age differences from east to west along the Snake River Plain. The high predicted heat-flow values in the eastern Snake River Plain have important tectonic and geothermal implications.

The identification of the mechanism for large changes in elevation, once the thermal volcanic event has passed, as thermal contraction, has important and testable consequences. For example, the regional warping which is so widely described (Kirkham, 1931; McIntyre, 1972) yet which does not appear to be associated with seismic activity (Pennington and others, 1974) is easily explained.

On the basis of the data and models discussed in this paper, we can describe a time progressive sequence of events for the history of the development of the Snake River Plain at any one point along the length. Consider, for example, the Oregon-Idaho border area. Beginning about 18 m.y. ago, a large outpouring of rhyolitic ash flows (associated with regional uplift of several hundred metres above the surrounding terrain, caldera formation at the source of the extrusives, and large-scale hydrothermal activity) occurred. After a few million years, the scene of the rhyolitic volcanism was located a few tens of kilometres to the east, the site of original activity subsided 500 m or so in elevation, and the caldera system was flooded by basalts. During the next few million years, the eastward movement of the volcanic centers and the subsidence of the older centers continued. The original center was a continuing locus of basaltic volcanism and focus of thick sediment accumulation, because the area was lower in elevation than the surrounding terrain. The proportion of crustal rifting in the development of the system is unknown.

Finally, the very high heat-flow values predicted for the younger part of the Snake River Plain have important geothermal consequences. For example, there should be many active geothermal systems in the eastern Snake River Plain, where the age of inception of silicic volcanism ranges from 5 to 0.6 m.y. The area must thus represent an important geothermal resource area.

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