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Geothermal Modeling of Jackson Hole, Teton County Wyoming

by

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INTRODUCTION

This study investigated the possibility of high-temperature-heat sources (greater than 300 °C) in the area of Jackson Hole, northwestern Wyoming (see Figure 1 for study area location). Analytical and finite-difference numerical models describing conductive and convective terrestrial heat transport were utilized in an attempt to define the thermal regime of this area.

This report will first present data which were used as constraints for the analytic and numerical thermal models. These data include a general discussion of geology of the area, thermal spring information, subsurface temperature information, and hydrology of the area.

The modeling techniques are presented next with a discussion of assumptions and data used to constrain the models.

Lastly, results of the models are presented with a discussion of interpretations and implications for the existence of high-temperature heat sources in the Jackson Hole area.

GENERAL GEOLOGY OF JACKSON HOLE

Jackson Hole is a 60 km by 15 to 30 km complexly folded and faulted basin (Love and Reed, 1971; Love et al., 1973). Within the basin are three structurally deep areas in which the Precambrian basement varies from 3 km to 4.5 km below sea level (Love in Behrendt et al., 1968). Sediments contained in Jackson Hole represent all systems except Silurian (see Table I for a generalized stratigraphic column). The Cenozoic sedimentary section is the most complete of any Wyoming basin, due to the basin's subsidence during Cenozoic time.

Surrounding tectonic features include the Teton Mountain Range and the Yellowstone volcanic plateau (refer to Figure 1). The height of the

Figure 1. Location of study area (stippled) in relation to Jackson Hole and surrounding tectonic features (modified from Behrendt et al., 1968).

Table I. Generalized stratigraphic column showing thermal conductivities and units used in thermal models.

Era	Period	Description ¹	Modeled Thickness (meters)	Thermal Conductivity ² (watts/meter Kelvin)	Modeling Unit ³
Cenozoic	Quaternary	Glacial gravels Stream deposits	0 to 525	3.3	8
Cenozoic	Tertiary	Gray and white clay, limestone, and volcanic ash	0 to 2400	2.5	7
Cenozoic	Tertiary	Gray conglomerate and volcanic ash	300	3.4	6
Mesozoic	Cretaceous	Brown conglomerate	300	3.1	5
Mesozoic	Cretaceous	Gray sandstone and shale	1500 to 3000	1.9	4
Mesozoic	Cretaceous Jurassic Triassic	Red, gray, green and mottled sandstone, shale and limestone	1050	2.3	3
Paleozoic		Gray limestone, gray and green shale, and red-brown sandstone	1200	3.5	2
Precambrian		Gneiss, schist and granite		3.2	1

¹ Description is taken from U.S. Geological Survey's Map of Grand Teton National Park (1968), Love (in Behrendt et al., 1968), and Love et al., (1973).

² Thermal conductivity values were assigned using measured data from Heasler (1978). For the Quaternary, Tertiary, and uppermost Cretaceous strata thermal conductivity values were estimated based on lithology.

³ Refers to cross-section AA' (Figure 4).

Precambrian in the Teton Range exceeds 4 km above sea level. Love (in Behrendt et al., 1968) interprets the Teton Range as a horst between the two downfaulted blocks of the Teton Basin to the west and Jackson Hole to the east. The north-south trending Teton fault system (Figure 2) separates the Teton Range from Jackson Hole. Behrendt et al. (1968) have interpreted up to 7 km of vertical displacement along the Teton fault system. The Yellowstone volcanic plateau terminates against the northern portion of Jackson Hole.

Beginning in the Miocene and continuing through the Pliocene, there has been extensive volcanism in the Yellowstone region which roughly coincides with the subsidence of Jackson Hole. Contemporaneous with the subsidence of Jackson Hole has been the uplift of the Teton Range. Movement originated in the last 10 my (million years) and is still active (Love in Behrendt et al., 1968). Barnosky (1984) proposes that volcanoes were active in Jackson Hole from the early to middle Miocene (about 24 to 13 my ago). Within 5 km of the southern boundary of Yellowstone National Park are volcanic rocks of the Plateau Rhyolite which ranges in age from 70,000 to 600,000 years old (Christiansen and Blank, 1972).

Summaries of the geologic history of Jackson Hole are given in Behrendt et al. (1968), Love and Reed (1971), and Love et al. (1973).

THERMAL SPRINGS

Ten thermal springs are located near the study area with four thermal springs contained within the study area (Figure 2). Chemical and flow data for all ten springs are listed in Table II.

The two highest temperature springs (Huckleberry Hot Springs and Jackson Late Hot Springs) are nearest Yellowstone National Park. Huckleberry Hot Springs includes two groups of vents and seeps flowing 380 L/min (liters per

- Thermal spring showing temperature (Breckenridge and Hinckley, 1978.)
- Geothermal gradient (°C/Km) calculated from oil well bottom-hole temperatures. See Tables III and IV.
- × Bottom-hole temperature (°C) and depth (meters) from precision thermal measurements. See Figure 3.
- Line of cross-sections used in finite-difference thermal models.

Figure 2. Locations of study area (stippled), thermal springs, subsurface temperature data, and cross sections used in finite difference models.

Table II. Thermal spring data for the Jackson Hole area.¹

Thermal Spring	Temperature (°C)	Flow (L/min)	TDS (Mg/L)	Location (Township/Range/Sec)
Huckleberry	51	1136	688	48/115/20
Jackson Lake	72	5520		46/115/19
North Buffalo Fork	45	757		45/111/32
Kelly	27		284	42/115/2
Teton Valley	18		248	42/115/11
Abercrombie	27	946	192	41/116/2
Boyles Hill	30	189	2480	41/117/36
Granite	41	1136	670	39/113/6
Granite Falls	45	454		39/113/6
Astoria	37	379	1160	39/116/32

¹ All data is from Breckenridge and Hinckley (1978) except for the flow data for Jackson Lake Hot Springs which is from J.D. Love (personal communication, 1986).

minute) of 45 to 61 °C water (Breckenridge and Hinckley, 1978). These springs occur in a large area of siliceous sinter overlying glacial material and Pliocene rhyolites (Love, 1974). Chemical geothermometry has been interpreted as indicating a subsurface reservoir temperature of 133 °C (Muffler, 1979).

Jackson Lake Hot Springs are located on the western shore of Jackson Lake (Figure 2). Wagner (1964) described a variety of small springs along 275 meters of shoreline with temperatures ranging between 24 to 50 °C. Measurements by J. D. Love indicate a minimum flow of 5520 L/min with the highest temperature being 69 °C (J. D. Love, personal communication, 1986). These springs occur along the north-south trending Teton fault system and the intersection of two smaller converging northwest-southeast trending normal faults (Love and Reed, 1973). Madison Limestone is mapped at the location of the springs (Love and Reed, 1973).

Breckenridge and Hinckley (1978) consider both Teton Valley Warm Springs (18 °C, flow undetermined) and Kelly Warm Springs (27 °C, flow undetermined) to be sourced from the Madison Limestone. They base their assumption on the chemical similarity between these springs and Madison waters and on the geologic structure near the springs. Kelly Warm Springs occur in Quaternary alluvium while nearby Teton Valley Warm Springs are located at the base of a long slope of deeply fractured Madison Limestone (Breckenridge and Hinckley, 1978).

Abercrombie Warm Springs (27 °C, 950 L/min) occur in an area of Holocene loess (Breckenridge and Hinckley; 1978). Paleozoic rocks which have been intruded and covered by Tertiary or Quaternary andesites, are located just south of the springs (Love, 1975). These springs are located along the Warm Springs fault, an east-west normal fault mapped by Love (1975).

SUBSURFACE TEMPERATURE DATA

Bottom-hole temperatures (BHT's) reported with logs from oil and gas wells were analyzed in an attempt to provide subsurface temperature data. Data from Wyoming Geological Survey files for 18 wells are presented on Figure 2 and in Tables III and IV. Figure 2 shows the location and range of calculated geothermal gradients for the 18 wells. Table III statistically summarizes the 34 BHT's and calculated geothermal gradients for the 18 wells. Table IV lists data for the 34 measurements including depth, BHT, location, and calculated gradient.

Geothermal gradients were calculated from BHT data using the formula

$$\text{Gradient} = \frac{(\text{BHT}) - (\text{MAAT})}{\text{Depth}}$$

where MAAT is the mean annual air temperature. For the gradient calculation a mean annual air temperature of 3 °C was assumed. This is between Becker and Alyea's (1964) reported values of 3.3 °C for the town of Jackson Hole and 1.2 °C for Moran.

Difficulties exist with the use of oil well BHT's in geothermal studies. There are problems associated with the thermal effects of drilling and with operator inattention in measuring and reporting BHT's. Also, drilling fluids may transfer heat to the bottom of a drill hole, warming or cooling the rock depending on the drilling fluid temperature and the depth of the hole. The magnitude of the thermal disturbance depends on such factors as the temperature difference between the drilling fluid and the rock, the time between end of fluid circulation and temperature measurement, the type of drilling fluid used, the length of time of fluid circulation, and the degree to which drilling fluids have penetrated the strata.

Table III. Summary of 34 bottom-hole temperature measurements for 18 wells in the Jackson Hole area.

DEPTH - TEMPERATURE ANALYSIS								
(meters)			(°C)					
DEPTH	NO.	HIGH	LOW	MEAN	50%	66%	80%	90%
400.	2.	37.8	22.2	30.0	37.8	37.8	37.8	37.8
1000.	2.	42.2	35.6	38.9	42.2	42.2	42.2	42.2
1200.	2.	46.7	29.4	38.1	46.7	46.7	46.7	46.7
1400.	4.	56.1	23.3	38.6	37.8	37.8	56.1	56.1
1600.	3.	40.0	32.8	36.5	36.7	40.0	40.0	40.0
1800.	1.	73.3	73.3	73.3	73.3	73.3	73.3	73.3
2000.	2.	50.6	27.8	39.2	50.6	50.6	50.6	50.6
2200.	2.	63.3	56.7	60.0	63.3	63.3	63.3	63.3
2400.	4.	65.6	52.2	56.7	56.7	56.7	65.6	65.6
2600.	2.	50.0	45.6	47.8	50.0	50.0	50.0	50.0
2800.	1.	46.7	46.7	46.7	46.7	46.7	46.7	46.7
3000.	2.	81.1	75.6	78.3	81.1	81.1	81.1	81.1
3200.	4.	71.1	53.3	65.1	70.6	70.6	71.1	71.1
3400.	2.	67.8	56.1	61.9	67.8	67.8	67.8	67.8
3800.	1.	72.8	72.8	72.8	72.8	72.8	72.8	72.8

DEPTH - GRADIENT ANALYSIS								
(meters)			(°C / km)					
DEPTH	NO.	HIGH	LOW	MEAN	50%	66%	80%	90%
400.	2.	92.9	66.0	79.5	92.9	92.9	92.9	92.9
1000.	2.	48.1	38.0	43.0	48.1	48.1	48.1	48.1
1200.	2.	40.0	23.4	31.7	40.0	40.0	40.0	40.0
1400.	4.	39.9	16.7	27.1	26.2	26.2	39.9	39.9
1600.	3.	23.4	20.1	22.3	23.3	23.4	23.4	23.4
1800.	1.	41.2	41.2	41.2	41.2	41.2	41.2	41.2
2000.	2.	23.9	13.4	18.6	23.9	23.9	23.9	23.9
2200.	2.	28.9	24.5	26.7	28.9	28.9	28.9	28.9
2400.	4.	27.7	21.0	23.5	23.1	23.1	27.7	27.7
2600.	2.	18.8	16.6	17.7	18.8	18.8	18.8	18.8
2800.	1.	16.8	16.8	16.8	16.8	16.8	16.8	16.8
3000.	2.	26.8	24.3	25.6	26.8	26.8	26.8	26.8
3200.	4.	22.0	16.5	20.1	21.7	21.7	22.0	22.0
3400.	2.	19.6	16.6	18.1	19.6	19.6	19.6	19.6
3800.	1.	19.1	19.1	19.1	19.1	19.1	19.1	19.1

Notes: Depth range is from the indicated value to 200 meters less. Gradients are computed using the formula: $((\text{bottom-hole temperature} - \text{mean annual surface temperature}) / \text{depth}) * 1000$. Percentiles identify the data value below which that percentage of the data falls. This is done on a case counting basis, i.e. the actual data values are not considered.

Table IV. Oil well bottom-hole temperature data for the Jackson Hole area.¹

DEPTH (meters)	TEMPERATURE (°C)	GRADIENT (°C/km)	LOCATION ² (Township/Range/Section)
291.	22.2	66.0	42/112/14 A
374.	37.8	92.9	42/114/1
815.	42.2	48.1	42/113/6
857.	35.6	38.0	44/112/31
1132.	29.4	23.4	43/114/9
1091.	46.7	40.0	44/113/25 B
1216.	23.3	16.7	42/112/11 C
1305.	37.2	26.2	44/111/25 D
1332.	56.1	39.9	45/113/29
1367.	37.8	25.4	44/113/13 E
1445.	36.7	23.3	42/113/24
1484.	32.8	20.1	44/111/25 D
1582.	40.0	23.4	44/112/31
1707.	73.3	41.2	43/112/34 F
1854.	27.8	13.4	42/112/14 A
1990.	50.6	23.9	43/112/34 G
2088.	63.3	28.9	44/113/25 B
2189.	56.7	24.5	44/112/31 H
2201.	52.2	22.4	43/112/34 G
2258.	65.6	27.7	44/113/25 B
2324.	56.7	23.1	43/110/15 I
2341.	52.2	21.0	43/112/34 G
2494.	50.0	18.8	43/112/34 F
2570.	45.6	16.6	42/112/14 A
2603.	46.7	16.8	42/112/11 C
2914.	81.1	26.8	44/112/31 H
2981.	75.6	24.3	44/113/13 E
3045.	53.3	16.5	42/112/11 C
3075.	70.6	22.0	43/110/15 I
3107.	65.6	20.1	43/112/34 F
3134.	71.1	21.7	43/111/6
3200.	56.1	16.6	42/112/14 A
3313.	67.8	19.6	43/112/34 F
3653.	72.8	19.1	43/113/2

¹ Data gathered from records at the Wyoming Geological Survey, 1986. Thirty-four measurements from 18 wells are listed.

² Letters refer to multiple bottom-hole temperature measurements within one well.

It can be generally assumed that such factors as time of year, operator error, time since circulation, and drilling fluid characteristics are random disturbances which average out when considering a large data set. However, circulation of drilling fluids is generally a systematic effect which tends to decrease rock temperature more with increasing depth. With sufficient data at all depths, anomalous gradients may be identified despite the fact that they are depressed in value (for examples see Hinckley and Heasler, 1984; Heasler and Hinckley, 1985; Buelow et al., 1986; and Hinckley and Heasler, 1987).

Given the scarcity of BHT data for the Jackson Hole area, it becomes very difficult to define equilibrium temperatures, geothermal gradients, or thermal anomalies. However, generalities may be discerned from the data as presented in Table III. The calculated gradient falls between 18 to 27 °C/km for 77 percent of the data. This range of gradient applies roughly over the depth range of 1216 meters to 3653 meters. Such a generalized gradient for the study area is similar to the background gradient for the Bighorn Basin of 29°C/km based on the analysis of 2035 BHT's (Heasler and Hinckley, 1985) and the background gradient for the Wind River Basin of 28 °C/km on the analysis of 1744 BHT's (Hinckley and Heasler, 1987).

The reliability of using gradient values greater than 27 °C/km to define thermally anomalous areas cannot be adequately assessed at this time due to the scarcity of BHT data. However, as can be seen from Table IV, most of the gradients greater than 27 °C/km occur at relatively shallow depths (less than 1100 meters). Similar patterns have been observed in other Wyoming basins and do not necessarily indicate the presence of anomalous heat sources (Hinckley and Heasler, 1984; Heasler and Hinckley, 1985; Buelow et al., 1986; and Hinckley and Heasler, 1987).

One of the most valuable sources of subsurface temperature data results

from precision thermal measurements in wells. During September of 1986 precision thermal measurements were made in three wells in the study area using the methodology described by Decker (1973). A calibrated thermistor probe was lowered at discrete intervals and allowed to equilibrate at each level. Temperatures measured in this manner are believed to be precise to 0.005 °C and accurate to 0.1 °C (Decker, 1973).

Shown in Figure 3 are the measured temperature-depth profiles for these three wells (locations shown on Figure 2). All three wells are water wells which had been undisturbed for many months. However, all three temperature-depth profiles show effects of convective heat transport (water flow).

The Jenny well is located near Jenny Lake Lodge in Grand Teton National Park (Figure 2) and is drilled into glacial gravels and sands. Temperatures in the Jenny well are substantially lower than in the two wells near the town of Jackson Hole. In this area, the glacial material is an unconfined aquifer recharged by surface waters with generally shallow depth of circulation (Cox, 1976).

The Jackson E1 well is located on the National Elk refuge east of the Jackson Hole hospital. Temperatures in the Jackson E1 well clearly indicate subsurface water flow as shown by the two isothermal segments from 17 to 37 meters and 37 to 58 meters. Water movement from depth is also suggested because the second isothermal segment is near 11.3 °C. This temperature is greater than could be caused by conductive heat flow or shallow circulation of groundwater.

The Jackson E2 well is located approximately 300 meters west of the Jackson E1 well. Temperatures are cooler than the Jackson E1 well with a more regular change in the geothermal gradient. However, due to the change in gradient down the hole (from 80 °C/km at 27 meters to 12 °C/km at 73 meters),

Figure 3. Measured temperature-depth profiles (September, 1986). The Jenny Lake well is located in T44N/R116W, the Jackson E1 and E2 wells are located in T41N/R116W (see Figure 2).

the temperatures in this well are also believed to be disturbed due to water movement.

Although not useful in defining conductive equilibrium temperatures or geothermal gradients, the temperature data from these three wells suggests factors which should be considered. First, data from the Jenny Lake well suggests that water temperatures in the surficial unconfined aquifer will be within a few degrees of the mean annual air temperature. Consequently, the 11.3 °C temperatures from 43 to 58 meters in the Jackson El well indicate that groundwater is transporting heat from a greater depth. This is consistent with Breckenridge and Hinckley's (1978) interpretation for the origin of Teton Valley Warm Springs, Kelly Warm Springs, and Jackson Lake Hot Springs. Second, the indication of both shallow and deep water movement makes it very difficult to establish a conductive geothermal gradient from shallow well data.

HYDROLOGY

Hydrologic data has primarily been interpreted for alluvium and glacial outwash in Jackson Hole. Cox (1976) has constructed a potentiometric surface map for this unconfined aquifer, mapped approximate thickness of saturated alluvium and glacial outwash, and lists chemical analyses and specific capacities for selected wells. Cox reports productivity from this aquifer as great as 7600 L/min and a very high transmissivity of $.032 \text{ m}^2/\text{sec}$. This suggests that the alluvium and glacial outwash will be effective in concealing the deeper thermal structure of the basin because of the great amount of convective heat transport (water flow) which it supports.

Hydrologic data for deeper aquifers are unavailable and consequently are inferred from data collected in other basins. In the Bighorn Basin, major

aquifers are the Paleozoic Madison Limestone and Tensleep Sandstone. Artesian flows up to 11400 L/min are reported for the Madison Limestone (Lowry et al., 1976). Minor aquifers may exist stratigraphically higher than these aquifers, but will not be considered in this study because they will generally have poorer aquifer characteristics than the Paleozoic aquifers and will not be as deeply buried and hence will be cooler than the Paleozoic aquifers.

THERMAL MODELING

Two types of thermal models were constructed for Jackson Hole. The first type was used to estimate the temporal thermal effects of known volcanism in the area. The second type was used to estimate the present day thermal regime along cross-sections in Jackson Hole.

Analytic Thermal Models for Intrusions

The temporal thermal effects of volcanism were estimated by considering a rectangular intrusion buried at a depth below a surface held at 0°C. The rectangular intrusion cools by conduction as described by Carslaw and Jeager (1959). The equation solved was

$$\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2} = \frac{1}{\kappa} \frac{\partial T}{\partial t}$$

where T represents temperature, t is time, κ is thermal diffusivity, and x, y, and z are spatial coordinates. Input parameters include size of the intrusion, depth of burial, temperature of the intrusion, thermal diffusivity of the strata, and coordinates relative to the intrusion for temperature calculation. The model then calculates the conductive thermal effect of the intrusion at the specified location.

Results of the model are sensitive to the size of the intrusion, initial temperature, and coordinates for temperature calculation. The model assumes uniform thermal properties, an initial temperature distribution of 0 °C everywhere, a constant surface temperature of 0 °C, thermal diffusivity of 32 km²/my, heat transfer by conduction only, and no heat sources or sinks other than the intrusion.

The first application of this model is to the early to middle Miocene volcanism proposed for Jackson Hole by Barnosky (1984). During this time period from approximately 24 to 13 my ago, Barnosky believes that the volcanic vents were located along or near the Teton fault system within 25 km of Pilgrim Mountain perhaps near the northern end of Jackson Lake. Barnosky separates the volcanism into an andesitic cycle lasting from 24 to 18 my ago and a basaltic cycle at 13 my ago.

In an effort to determine the effect of this volcanism on the present thermal regime on Jackson Hole, parameters which maximize the thermal effect of the volcanism were chosen. Model A as shown in Table V represents an intrusion which is 3 km thick by 30 km long by 12 km wide and is buried 1 km. The coordinate for temperature calculation is centered on the upper surface of the intrusion. The initial temperature of the intrusion was chosen as 1200 °C. Although such a temperature is unrealistically high for andesitic volcanism, it represents the maximum possible initial temperature. Parameters for Model B (Table V) are identical to Model A except that the intrusion is buried 10 km rather than 1 km.

As can be seen from Table V, the present day temperature increase due to Models A and B would be undetectable. If the intrusion were unrealistically enlarged (6 km thick by 60 km long by 24 km wide) then the present-day thermal effect for a 1 km burial is still undetectable (Model C in Table V). Even for

Table V. Results of cooling intrusion models for Miocene volcanism in Jackson Hole. Models A and B are .3 km thick by 30 km long by 12 km wide and are buried 1 and 10 km, respectively. Models C and D are 6 km thick by 60 km long by 24 km wide and are buried 1 and 10 km, respectively. The coordinate for temperature calculation is centered on the upper surface of the intrusion. An initial intrusion temperature of 1200°C was used. See text for discussion.

Time since Intrusion (my)	Temperature (°C)			
	Model A	Model B	Model C	Model D
0.0	600	600	600	600
0.5	24	171	86	412
1.0	7	87	352	278
1.5	3	53	19	199
2.0	2	34	12	148
2.5	1	24	8	113
3.0	1	17	6	88
3.5	0	13	4	70
4.0	0	10	3	57
4.5	0	8	3	47
5.0	0	6	2	39
5.5	0	5	2	33
6.0	0	4	1	28
6.5	0	4	1	24
7.0	0	3	1	21
7.5	0	3	1	19
8.0	0	2	1	16
8.5	0	2	1	15
9.0	0	2	1	13
9.5	0	2	1	12
10.0	0	1	0	10
10.5	0	1	0	9
11.0	0	1	0	9
11.5	0	1	0	8
12.0	0	1	0	7
12.5	0	1	0	7
13.0	0	1	0	6
13.5	0	1	0	6
14.0	0	1	0	5

the case where the intrusion is buried 10 km (Model D in Table V) the temperature increase for a point centered on the upper face of the intrusion is only 6 °C.

Models of the thermal effects of Yellowstone volcanism on Jackson Hole are difficult to constrain. First, the size and depth of the intrusive bodies responsible for the volcanism are unknown. Second, the distribution and location of the subsurface intrusive bodies in relation to Jackson Hole are also largely unknown.

As a first approximation, the size of the intrusion responsible for Yellowstone volcanism was chosen to be equal to the size of the Yellowstone caldera (40 km wide by 70 km long (Keefer, 1971; Smith and Braile, 1982)). Smith and Braile (1982) estimated that 6700 km³ of material has been contained in Quaternary silicic eruptions from the Yellowstone region. This equates to a thickness of 2.4 km over the area of the caldera. An intrusion thickness of 10 km was used as a maximum case. The initial intrusion temperature was modeled as 1200 °C with the depth of burial being either 3 or 10 km.

Christiansen and Blank (1972) discuss three main volcanic cycles in the Yellowstone region. The oldest cycle considered is associated with the deposition of the Huckleberry Ridge Tuff at 2 my ago. The next cycle is associated with the Lava Creek Tuff deposited at .6 my ago. The final cycle is defined by rocks of the Central Plateau Member (.20 to .07 my ago).

The cross-section lines AA' and BB' (Figure 2) are located approximately 50 km from the southern corner of the Yellowstone caldera. Consequently, temperatures were calculated for locations 20 km and 60 km from the southern corner of the intrusion.

Model A of Table VI lists the temperature results for an intrusion that is 10 km thick by 70 km long by 40 km wide and is buried 3 km. The initial

Table VI. Results of cooling intrusion models for thermal effects of Yellowstone volcanism in Jackson Hole. Models A, B, C, and D are 10 km thick by 70 km long by 40 km wide. Models A and B are buried 3 km with the coordinate for temperature calculation at 8 km depth and 60 and 20 km, respectively, from the corner of the intrusion. Models C and D are buried 10 km with the coordinate for temperature calculation at 15 km depth and 60 and 20 km, respectively, from the corner of the intrusion. An initial intrusion temperature of 1200°C was used. See text for discussion.

Time since Intrusion (my)	Temperature (°C)			
	Model A	Model B	Model C	Model D
0.0	0	0	0	0
0.5	0	0	0	0
1.0	0	1	0	2
1.5	0	3	0	5
2.0	0	5	0	8
2.5	0	6	0	10
3.0	0	6	0	10
3.5	0	6	0	12
4.0	0	6	0	13
4.5	0	6	0	13
5.0	0	6	0	13
5.5	0	5	0	13
6.0	0	5	0	13
6.5	0	5	0	12
7.0	0	5	0	12
7.5	0	4	0	12
8.0	0	4	0	11
8.5	0	4	0	11
9.0	0	4	0	10
9.5	0	4	0	10
10.0	0	3	0	10

temperature of the intrusion is 1200 °C and temperatures are calculated for a point located 8 km deep and 60 km from the corner of the intrusion. Model B is identical to Model A except the point for temperature calculation is only 20 km away from the edge of the intrusion. Model C is identical to Model A except the intrusion is buried 10 km and the depth for temperature calculation is changed to 15 km. Model D is the same as Model C except the point for temperature calculation is only 20 km from the edge of the intrusion.

As can be seen from the results in Tables V and VI, the modeled intrusions have little effect on the present thermal regime of the area studied in Jackson Hole. However, many approximations and assumptions have been used to construct these models. As additional geologic and thermal data become available, the models should be revised to incorporate such data.

Finite-difference Thermal Models for Cross-Sections

Finite-difference thermal models were constructed for two cross-sections through Jackson Hole (see Figure 2). The purpose of the models was to estimate temperatures in the Paleozoic aquifers by solving a two dimensional steady-state heat conduction and convection equation. The equation solved was

$$\frac{\partial K_x}{\partial x} \frac{\partial T}{\partial x} + \frac{\partial K_y}{\partial y} \frac{\partial T}{\partial y} + \rho c (V_x \frac{\partial T}{\partial x} + V_y \frac{\partial T}{\partial y}) = 0$$

where T represents temperature, K_x and K_y are the thermal conductivity of the fluid-saturated rock in the horizontal and vertical directions, V_x and V_y are the Darcian velocities of the groundwater in the horizontal and vertical directions, ρ is the density of the water, and c is the specific heat of water. This equation was solved using a Gauss-Sidel iterative method similar to that described by Kilty and Chapman (1980).

Input data for the models include the thermal conductivity of strata, the structure (geometry) of the area, approximate mean annual ground temperature, regional heat flow values, Darcian velocity of groundwater and the spacing between node points for temperature calculation. The model uses this data to calculate a temperature at every node point.

The geometry used for cross-section AA' is shown in Figure 4. This cross-section was taken from the U.S. Geological Survey's map of Grand Teton National Park (1968). The grid spacing used was 150 m vertically between nodes and 250 m horizontally between nodes. This resulted in temperatures being calculated for 9211 node points over a cross-sectional depth of 9 km and a length of 37.5 km.

Thermal conductivities were assigned to the rock units shown in Figure 4 based upon measurements in other Wyoming Basins (Heasler, 1978) and estimates considering rock lithologies (see Table I). The most poorly constrained thermal conductivity was that assigned to the upper Tertiary. To test the sensitivity of the model to the thermal conductivity of this unit, Tertiary thermal conductivity values were changed from 1.8 to 3.4 W/mK (Watts / meter Kelvin). The results of these calculations are shown in Figure 5 for a location where the Paleozoic section is deepest (16 km from the west side of the model shown in Figure 4). The maximum temperatures predicted at the base of the Paleozoic section differs only by 14 °C (169 °C for a thermal conductivity of 1.8 W/mK verses 155 °C for a thermal conductivity of 3.4 W/mK). Using the lithologic description of the Upper Tertiary given by Love (in Behrendt et al., 1968), a estimated thermal conductivity of 2.5 W/mK was used in subsequent models.

No heat flow data exist for Jackson Hole. Nearby heat flow values range from a low of 54 mW/m² (milliWatts / meter²) 150 km to the southeast in the

Figure 4. Geologic cross section AA' taken from the U. S. Geological Survey's map of Grand Teton National Park (1968). Unit refers to thermal conductivity values used in the finite difference thermal models (Table I). See text for discussion.

Figure 5. Calculated temperature profiles for three thermal conductivity values for the upper Tertiary. Location for temperature profiles is 16 km from the west boundary of the cross section shown in Figure 4. A heat flow value of 67 mW/m^2 was used in the calculations.

Wind River Range (Sass et al., 1971) to a value of over 100 mW/m^2 as measured in the southeast arm of Yellowstone Lake (Morgan et al., 1977). Heasler et al. (1983) summarize published heat flow data for the area around Jackson Hole and Wyoming. A heat flow value of 67 mW/m^2 has been used to model geothermal systems in the Bighorn Basin (Heasler and Hinckley, 1985) as well as the Wind River Basin (Hinckley and Heasler, 1987). Thus, heat flow values of 50, 67, 75, and 100 mW/m^2 were input into the model for cross-section AA'.

Results of varying the heat flow are shown in Figure 6 for the location where the Paleozoic aquifers are deepest (16 km from the west side of the model shown in Figure 4). The maximum temperatures predicted for the Paleozoic section varies by 100°C . However, it is important to note that even for the lowest heat flow value, the Paleozoic section is at a temperature of 108 to 122°C . These temperatures are greater than any reported thermal spring temperature in Jackson Hole.

Modeled results for the four cases of varying heat flow are shown as temperature cross-sections in Figures 7A through 7D. The effects of thermal refraction due to variations in thermal conductivity and structural geometry are shown in these figures. Also shown are the steady-state conductive temperatures associated with the Paleozoic aquifers. Note that in all four cases of varying heat flow, the Paleozoic section in the deepest portion of Jackson Hole is substantially warmer than the highest reported thermal spring temperature (72°C for Jackson Lake Hot Springs)

Given the modeled results as shown in Figure 7, BHT analyses as listed in Tables III and IV, and familiarity with geothermal studies in nearby basins (Heasler and Hinckley, 1985; Hinckley and Heasler, 1987), a heat flow value of 67 mW/m^2 is chosen as being the most reasonable value. It will be important in further studies to more precisely determine heat flow values for the

Figure 6. Calculated temperature profiles for four heat flow values. Location for the temperature profiles is 16 km from the west boundary of the cross section shown in Figure 4. Thermal conductivities used for the calculations are listed in Table I.

Figure 7. Calculated temperatures for cross section AA' (Figures 2 and 4) for four heat flow values. Thermal conductivities used are listed in Table I. See text for discussion.

Jackson Hole area.

In order to estimate the steady-state, two-dimensional thermal effects of forced convective heat transport using the method of Kilty and Chapman (1980), the mass flux (Darcian flow) of the water contained within the Paleozoic section must be estimated. As discussed in the section on hydrology, hydrologic data for aquifers in Jackson Hole are sparse. Consequently, at this time only very general constraints can be placed on the Darcian flow.

Darcey's law was used to estimate reasonable values for the Darcian velocity. In one form of Darcey's law

$$v = \frac{-T}{b} \frac{dh}{dl}$$

where v is Darcian velocity (also termed specific discharge), T is transmissivity, dh/dl is hydraulic gradient, and b is saturated aquifer thickness.

For the northeastern Bighorn Basin, Huntton (1985) lists Madison Limestone transmissivities of 2.7×10^{-6} to $6.9 \times 10^{-4} \text{ m}^2/\text{sec}$. Heasler (1982) shows hydraulic gradients for the northwestern Bighorn Basin of 2.3 to 31.6 m/km. If an aquifer thickness of 1 km is assumed, then Darcian velocities calculated using these parameters range from 6.4×10^{-12} to $2.2 \times 10^{-8} \text{ m/sec}$.

The Darcian velocity could approach zero m/sec as a lowest limit. For such a case, the thermal regime would be purely conductive as modeled in Figures 7A through 7D and Figure 8A. A reasonable upper limit to the Darcian velocity is more difficult to define. Huntton (1985) believes that the high transmissivity values reported for the Madison Limestone are only applicable to basin margins. He believes that in the deeper centers of basins, hydraulic transmissivities are much less than those reported for basin margins.

Consequently, the Darcian velocity estimate of 2.2×10^{-8} m/sec which was calculated using a high transmissivity value of 6.9×10^{-4} m²/sec and a large hydraulic gradient of 31.6 m/km is perhaps orders of magnitude too great. Lacking additional constraints, a Darcian velocity of 10^{-9} m/sec was used as a maximum value.

Data from cross-section BB' (Figure 2) were used to estimate the combined effects of heat conduction and convection. Figure 8 (taken from Figure 7 in Behrendt et al., 1968) shows the structural geometry used in the model. Temperatures were calculated every 500 m horizontally and 200 m vertically for 3914 node points over a cross-sectional depth of 7.4 km and a length of 51 km. Because Behrendt et al. (1968, their Figure 7) separate the geologic units differently than shown in Figures 4 and 7, the thermal conductivity values used to model Figure 8 were chosen to roughly correspond to those listed in Table I. A single heat flow value of 67 mW/m^2 is used.

Figure 8A shows the results of modeling cross-section BB' when there is no water flow (a Darcian velocity of zero). Note that the maximum temperature of the Paleozoic section is similar to that shown in Figure 7B.

Figure 8B shows the temperature disturbance caused by the maximum Darcian velocity (10^{-9} m/sec). The Paleozoic section still attains temperatures of near 100°C even with this large volume of water flow.

The temperature structure shown along the fault system in Figure 8A is highly speculative. This is for two main reasons. First, hydrologic properties of the fault (transmissivity, saturated thickness, hydraulic gradient) are unknown. Second, the nature of heat transfer near the surface is ill-defined. Whether the heated water travels to the surface and cools by Newton's law or whether the heated water flows into the surficial Quaternary aquifers and cools by mixing and conduction will significantly effect the

Figure 8. Calculated temperatures along cross section BB' (Figure 2). Geology taken from Figure 7 in Behrendt et al. (1968). In Figure 8A (upper cross section) there is no water flow. Figure 8B (lower cross section) includes the effect of water flow in the Paleozoic section and up the Teton fault. See text for discussion.

modeled temperature structure along the fault.

CONCLUSION

The results of the thermal models for cross-sections AA' and BB' indicate that the temperatures of the deepest Paleozoic section in Jackson Hole are greater than observed hot springs temperatures. Since Jackson Lake Hot Springs, Teton Valley Warm Springs, Kelly Warm Springs, and perhaps Abercrombie Warm Springs may all be sourced from Paleozoic aquifers (Breckenridge and Hinckley, 1978), the heating mechanism for the thermal spring waters appears to be deep circulation of groundwater. Such a heating mechanism is similar to other geothermal systems in Wyoming associated with thermal springs (Hinckley et al., 1982; Heasler, 1982).

Maximum temperatures of water contained in the Paleozoic aquifers are poorly defined. This is due to uncertainty in the heat flow (Figures 7A to 7D), thermal conductivities (Figure 5), and Darcian velocity of the Paleozoic groundwater (Figures 8A and 8B). Modeled Paleozoic temperatures range from over 220 °C (Figure 7D) to 100 °C (Figure 8B).

Other important unknown factors include hydrologic data for both the Paleozoic aquifers and the Teton fault system. The validity of the proposed mechanism of groundwater heating (deep circulation) is linked directly to the water-bearing properties of the deeply buried Paleozoic section and the nature of water flow up the Teton fault system.

Given the assumptions delineated in this report, the observed thermal spring temperatures, and the numerical model results, the possibility of high temperature heat sources in the Jackson Hole study area is unlikely. The possibility for high temperature heat sources greatly increases nearer Yellowstone National Park due to recent igneous activity.

Additional study is warranted in Jackson Hole for more concise definition of the thermal and tectonic evolution of the area. Much needed data should be collected on the heat flow of the region, thermal conductivities of strata, rock temperatures, geochemistry, and hydrology. The collection of such additional data will greatly reduce the uncertainties associated with modeling maximum temperatures of the geothermal regime.

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