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THE THERMOPOLIS HYDROTHERMAL SYSTEM, WITH AN ANALYSIS OF HOT SPRINGS STATE PARK

by

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Front cover. Oblique aerial photograph of the Thermopolis area, looking east and southeast toward the Bighorn and Owl Creek Mountains, showing Hot Springs State Park and the City of Thermopolis.

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SYNOPSIS

Thermopolis is the site of Hot Springs State Park, where numerous hot springs produce nearly 3,000 gallons per minute (gpm) of 130°F (54°C) water. The University of Wyoming Geothermal Resource Assessment Group has studied a 1,700-square-mile area centered roughly on the State Park. Available literature, bottom-hole temperatures from over 400 oil well logs, 62 oil field drill stem tests, the Wyoming State Engineer's water well files, 60 formation water analyses, thermal logs of 19 holes, and field investigations of geology and hydrology form the basis of this report.

The present springs, as well as indications of previous springs, are located at the crest of the Thermopolis Anticline. This is an asymmetric fold, much steeper to the south, which plunges east and northwest from Thermopolis. The anticline appears to be broken along its axis by a major basement fault and by smaller transverse faults. From the crest of the anticline, where Permian and Triassic formations are exposed, strata up through Cretaceous dip steeply southward into a sharp syncline, then rise gently up the north flank of the Precambrian-cored Owl Creek Mountains.

Analysis of thermal data reveals that temperatures of up to 161°F (72°C) occur along the crest of the Thermopolis Anticline within 500 feet of the surface. Thermal gradients along the anticline range from 43 to 300°F/1,000 feet, in contrast with gradients of around 15°F/1,000 feet for areas to the north and south. In addition to this low-temperature hydrothermal resource area (approximately 30 square miles) along the Thermopolis Anticline, there is a marginal resource in the Red Spring Anticline area 8 miles east of Thermopolis which shows gradients of up to 51°F/1,000 feet. Thermal gradients within the resource area increase with proximity to the crest of the anticline. The highest gradients and temperatures are found near the northwest end of the structure.

We have studied the hydrology and heat flow of these geothermal anomalies. Investigations indicate that waters discharging at Hot Springs State Park enter upper Paleozoic aquifers which crop out in the mountains to the south and west. These waters are confined by relatively impermeable Triassic siltstones and mudstones, and they flow under artesian pressure through the intervening syncline to surface along faults breaking the crest of the Thermopolis Anticline. Although three heating mechanisms have been proposed, geological considerations and thermal modeling identify simple conductive heating in the deep portions of the syncline as most plausible. Furthermore, flow and heating models indicate that the maximum temperatures likely to be produced from the system at reasonable drilling depths are 140°F (60°C) in the immediate vicinity of Thermopolis and 170°F (77°C) in an area 8 miles to the northwest. Artesian pressure is apparently sufficient to ensure surface flow for wells in a broad area along the Bighorn River south and north of Thermopolis.

The major aquifers for the Thermopolis geothermal system are the Permian Park City Formation (mostly limestone), the Pennsylvanian Tensleep Sandstone, and the Mississippian Madison Limestone. The Flathead Sandstone of Cambrian age may also yield hot waters, though at far greater depths. Chemical comparisons between identified aquifer waters and the Thermopolis hot springs suggest the Madison Limestone as the major water source, though contributions from overlying units are likely. Potential yield generally increases from the Park City Formation to the Tensleep Sandstone, and again to the Madison Limestone. Individual wells into the Madison Limestone in the southern Bighorn Basin have produced nearly 3,000 gpm. Existing hot wells (less than 1,000 feet deep) in the area just north of Thermopolis flow up to 1,000 gpm from the Park City Formation.

That geothermal waters are mixing between the upper Paleozoic formations along the Thermopolis Anticline is demonstrated by isothermal conditions in drill holes, homogeneous chemistry, and similarity of hydraulic head. Thus, drilling deeper than necessary to secure adequate flow is unlikely to produce significantly higher temperatures, higher pressures, or superior chemical characteristics. Waters within this geothermal reservoir are similar in composition to the existing springs: calcium sulfate and bicarbonate waters with total dissolved solids of around 2,300 milligrams per liter.

Geothermal waters have been used for residential space heating on a limited basis in Thermopolis for several decades. These applications, using surface, artesian discharge of hot well water via subfloor piping, may provide useful, long-term data on possible development problems. Drill-hole casing corrosion and collapse or mineral deposition may be responsible for declining flows in several wells; excessive calcium carbonate deposition is known to be a problem in certain cases. Legally, development of the Thermopolis geothermal system must comply with Wyoming State Engineer regulations on water appropriations and with various Federal and State agency procedures for leasing and drilling. An additional constraint specific to the Thermopolis area is that the flow of the springs of Hot Springs State Park is explicitly protected by statute.

INTRODUCTION

We have studied the Thermopolis hydrothermal system as part of a statewide geothermal resource assessment program. The Thermopolis system has received special attention because of the spectacular natural hydrothermal features located over it and because there is potential use for the geothermal resource in Thermopolis.

The study area for this report encompasses about 1,700 square miles in the southern end of the Bighorn Basin in northwest Wyoming. The Bighorn Basin is the subject of a regional geothermal analysis (Hinckley and Heasler, in preparation) and includes site-specific studies at Cody (Heasler, 1982) and Thermopolis (this report)(Figure 1). The major surface expression of the Thermopolis hydrothermal system is a group of springs represented locally as the "World's Largest Mineral Hot Spring." These springs give the town of Thermopolis its name and form the nucleus of the 640-acre Hot Springs State Park (Figure 2). The single largest vent in the group, known as Big Spring, flows 2,419 gpm [Wyoming State Engineer's files] on average, at 132°F (56°C). Including five hot water wells drilled just north of the State Park, the system produces 4,861 gpm at 124 to 132°F (51 to 56°C) (see Table 1).

Cursory studies of the Thermopolis system have been made by various workers, including Darton (1906), Woodruff (1909), Bartlett (1926), Burke (1952),

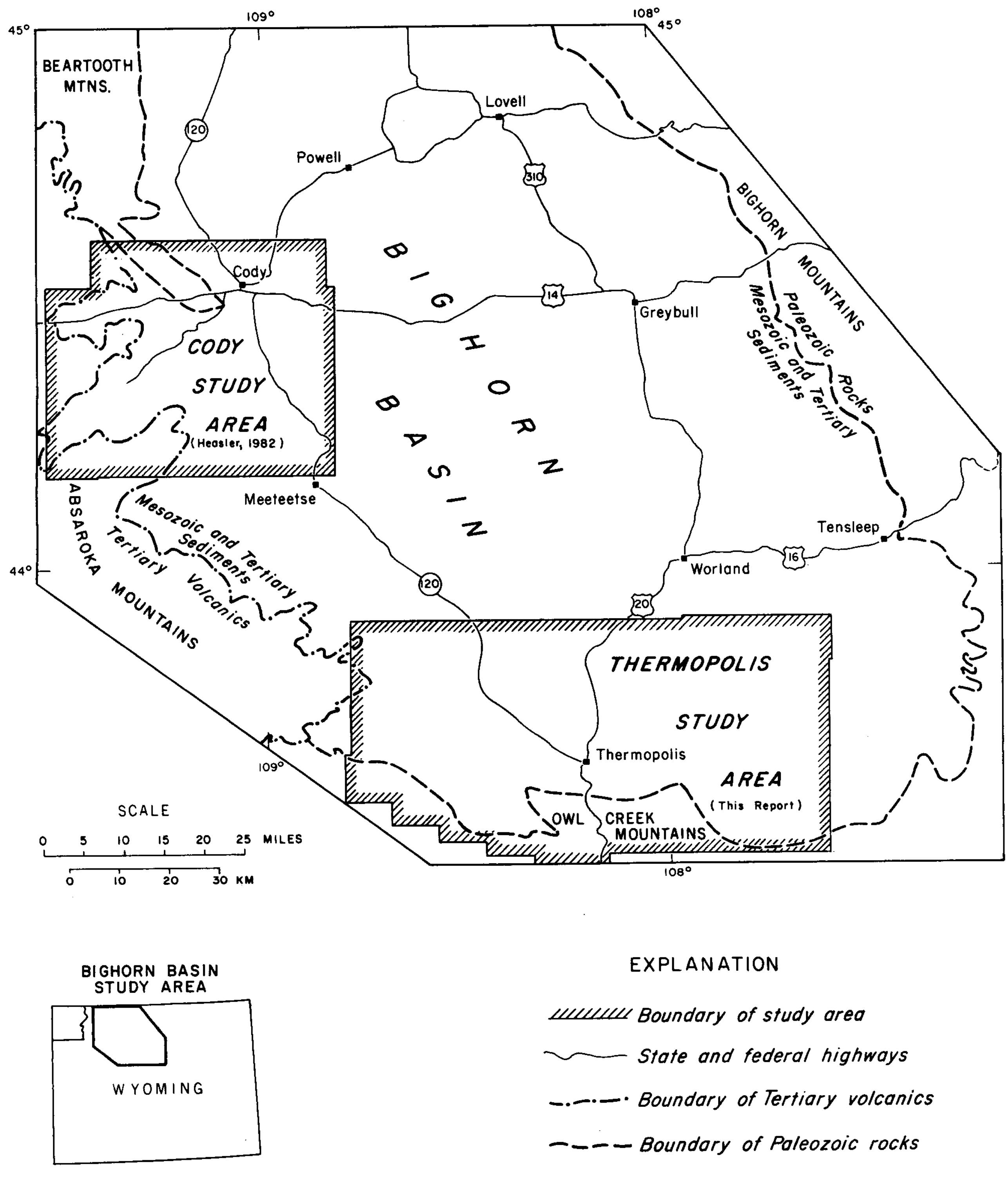


Figure 1. Location of the Thermopolis study area.

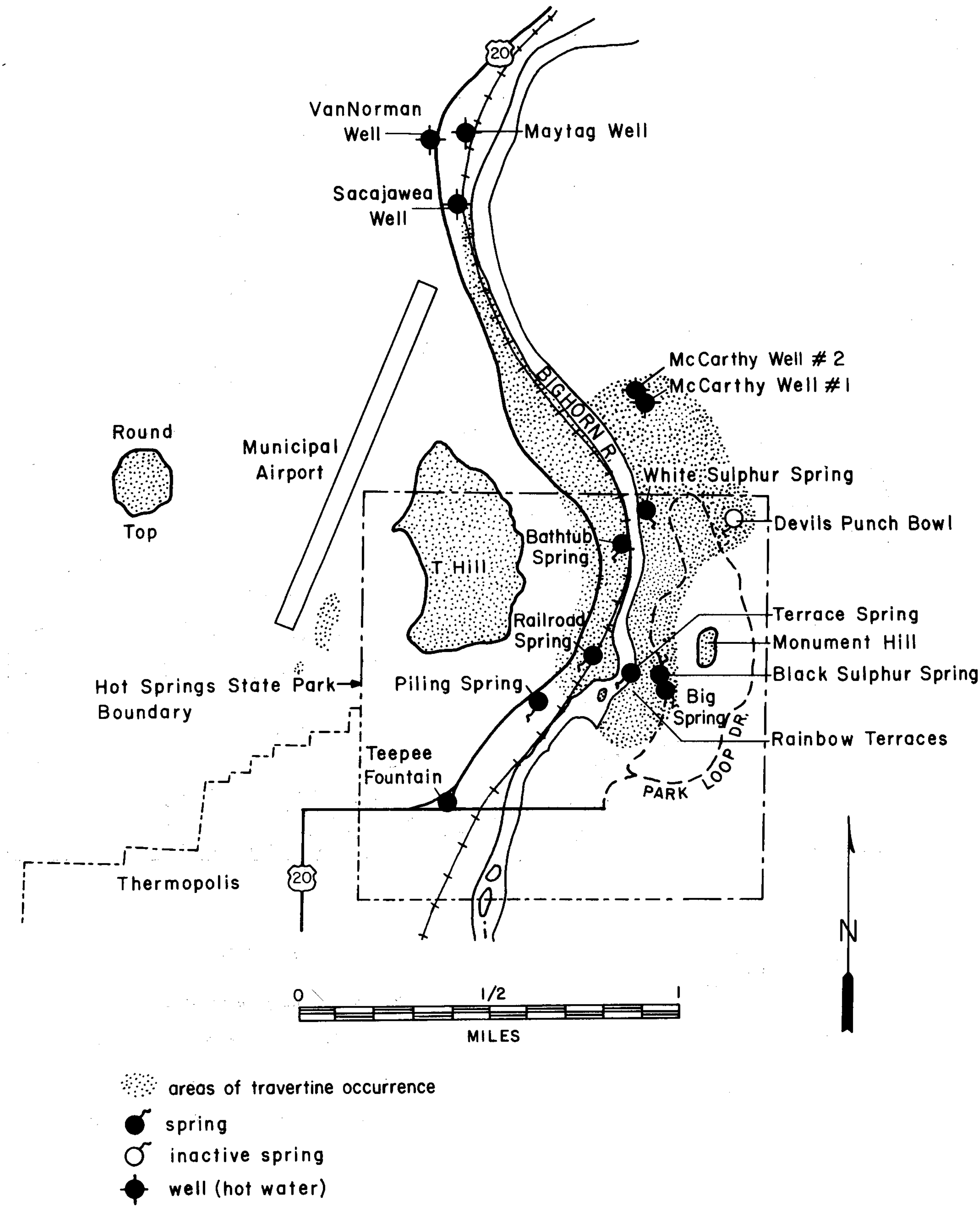


Figure 2. Location of hot springs and flowing wells in the Hot Springs State Park area. After Breckenridge and Hinckley, 1978.

and Breckenridge and Hinckley (1978). The present study is the first attempt to synthesize thermal, geologic, hydrologic, and chemical data for the system since extensive oil exploration and our own well logging have made such data available. The first section of this report develops the structural and stratigraphic framework of the Thermopolis area. Next, we present the results of our thermal and hydrologic investigations, with discussions of heating mechanisms, water chemistry and availability, and flow patterns for the system. The final section is a summary of our conclusions on the extent and functioning of the hydrothermal system, a discussion of the development implications of our findings, and a suggestion of productive directions for further study. Compilations of all bottom-hole temperature, water chemistry, and hydraulic head data are available as Open File Report No. 82-3 from the Geological Survey of Wyoming, Box, 3008, University Station, Laramie, Wyoming 82071.

Funding for this project was provided by the United States Department of Energy under Cooperative Agreement DE-F107-79ID12026 with the University of Wyoming Department of Geology and Geophysics. Co-principal investigator at the inception of the project was Edward Decker, whom we thank for his critical review of the manuscript. We wish to thank Coronado Oil Company and the people in the Thermopolis area who allowed thermal logging of drill holes and gave us their observations: Tom Anderson, Daune Bird, Lewis Freudenthal, Alice Jones, Dave Jones, Anna Maret, Clayton Merrit, Virgil Russel, Norman Sanford, Tom Sanford, Tom Sullivan, Scott Taylor, and Zola Van Norman.

Table 1. Well and spring data for the Hot Springs State Park area.

Name	Surface temperature	Average flow, gpm	Depth, feet
Van Norman Well	124°F (51°C)[1]	Controlled	550[2]
Quarry Well	115°F (46°C)	∿3	790[2]
Maytag Well	128°F (53°C)	736[2]	900(?)[1]
Sacajawea Well	128°F (53°C)	1,000[2]	900(?)[1]
McCarthy Well #1	129°F (54°C)	529[2]	510[3]
McCarthy Well #2	128°F (53°C)	<1	450[3]
Bathtub Spring	127°F (53°C)[1]	2[1]	
White Sulphur Spring	127°F (53°C)[1]	163[2]	
Black Sulphur/Terrace Spring	131°F (55°C)[1]	10[1]	
Railroad Spring		<3[1]	
Piling Spring	>95°F (35°C)[1]	<3[1]	
Big Spring	132°F (56°C)	2,419[2]	
TOTAL		4,861	

Flow weighted average temperature = 130°F (54°C)

References: [1]Breckenridge and Hinckley, 1978; [2]Wyoming State Engineer; [3]Bartlett, 1925.

GEOLOGY

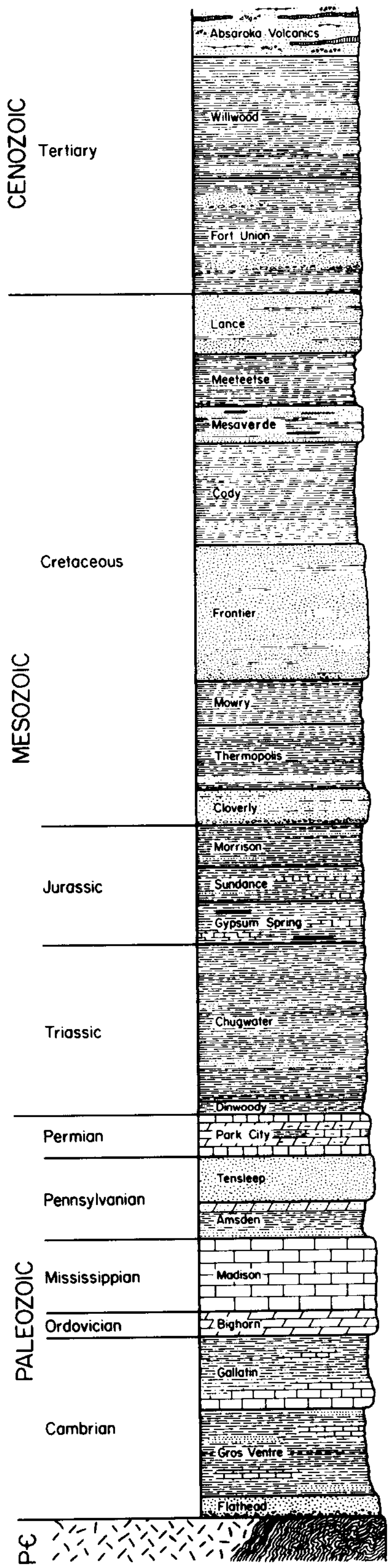
Within 10 miles north and south of Thermopolis are outcrops of rocks spanning over 3 billion years of geologic time. The names, general arrangement and compositions, and ages of these rock strata are presented in Figure 3 along with a brief statement on their water-bearing properties. Surface exposure of the various units is controlled by how they have been folded, faulted, and eroded. Plate 1 and Figure 4 display this information along with the names of major folds, and of oil and gas fields, in the study area.

All of the thermal springs in the State Park presently flow from the lower Chugwater Formation along the Bighorn River. Extensive travertine, sulphur, and gypsum deposits, mostly west of the river (Figure 4), indicate that hydrothermal activity has not always been confined to its present location. Commercial quantities of sulphur coincident with Park City Formation outcrops mark a major focus of activity 4 miles west-northwest of town (Major, 1946), and travertine caps on Round Top and T Hill (Figure 2) mark mineral springs activity up to 600 feet higher than at present. Logically, such springs will seek the lowest available outlet, so the shifting pattern of activity may, in part, reflect continued downcutting of the Bighorn River. The present location of the springs and Bartlett's (1926) observation of numerous small hot springs in the Bighorn River support the proposal of topographic control. That all the hot springs may one day abandon their present sites for topographically lower vents is indicated by Breckenridge and Hinckley's (1978) conclusion, based on fluorimetric studies, that the waters of recently declining Black Sulphur Spring are now venting directly into the river.

The string of hydrothermal deposits shown in Figure 4 corresponds closely with the axis of the Thermopolis Anticline, an asymmetric fold trending and plunging roughly east and west-northwest from Thermopolis. Five domes occur along the anticline: from west to east they are Rose Dome (Red Rose Dome, Ottey Dome), Cedar Ridge (Cedar Mountain Anticline, White Rose Dome), Condit's Dome, West Warm Spring Dome, and East Warm Spring Dome. The southern flank of the anticline has steeply dipping strata, ranging from 30° to vertical or slightly overturned. The strata on the northern flank dip at much gentler angles, 5 to 20°. Dips on both limbs are less steep on the portion of the anticline east of the Bighorn River.

Just south of and parallel to the Thermopolis Anticline is a strongly asymmetric syncline, the north limb of which has steeply south-dipping units which bend sharply upward at the syncline axis. In the south limb, the units rise gently (5-10° dips) toward outcrops on the north flank of the Owl Creek Mountains. Like the anticline, the syncline plunges northwest. It is truncated to the east by the Red Spring - Wildhorse Butte Anticline. Its axis is roughly parallel to and within one mile or less of the anticline axis.

This tight, apparently similar folding is accompanied by thinning of shaley units, fracturing, and faulting. Aerial photographs reveal thinning of the Chugwater Formation on the steep south flank of the anticline just north of Thermopolis, on the south-



THICKNESS, FEET	PHYSICAL DESCRIPTION	WATER-BEARING CHARACTERISTICS
0-2400	Volcanics and pyroclastics, of chiefly andesitic composition.	Highly variable water yields due to heterogeneous lithology.
3300	Clay sandstone, shale; some conglomerate.	Same as above.
3300	Thin-bedded sandstone and conglomerate, shale; some coal beds.	Water yields primarily a function of sandstone content, which is highly variable both vertically and laterally. Secondary permeability less developed than in lower rocks.
1600	Thick-bedded sandstone and shale.	Same as above.
1300	Tuffaceous sandstone, shale; some bentonite and coal beds.	Same as above.
650-1300	Interbedded sandstone and shale; some coal beds.	Same as above.
2500-2800	Shale; some thin sandstone beds.	Small quantities of water from sandy and/or fractured zones.
610-950	Sandstone with interbedded shales; some thin bentonite beds	Good water supply from sandstone beds.
270-300	Shale, commonly siliceous, numerous thin bentonite beds.	Little or no water supply
340-420	Shale, Muddy Sandstone near base.	Some water from Muddy Sandstone.
170-260	Sandstone, siltstone, shale, conglomerate at base.	Small quantities of water from sandstone beds.
190-300	Claystone and sandstone.	Little or no water supply.
200-230	Shale, fine sandstone, some thin limestone beds.	Little or no water supply.
80-175	Shale, limestone, and gypsum.	May produce good yields locally due to gypsum solution.
1000-1190	Siltstone, shale, and fine sandstone.	Fair water yields from sandstone beds, little or no water supply otherwise.
40-80	Siltstone, with some dolomitic beds.	Little or no water supply.
200-280	Limestone and dolomite, with some siltstone and shale.	Good water supply from fractured zones.
280-390	Sandstone, with some dolomitic beds.	Good water supplies commonly under artesian pressure.
240-320	Shale, dolomite, local basal sandstone.	
330-490	Limestone and dolomitic limestone.	Excellent water quantities locally due to solution permeability, commonly under artesian pressure.
85-250	Massive dolomite	Same as above.
440-470	Interbedded limestone, siltstone, and silty shale.	Lithology suggests poor water supply.
360-510	Shale with some sandstone and limestone beds.	Lithology suggests poor water supply.
190-250	Sandstone, conglomeratic arkose at base.	Assumed to be good water supply.
	Granite, gneiss, and schist.	Water only from weathered and/or fractured zones.

Figure 3. Geologic column for the Thermopolis study area. See page 43 for credits. Column thicknesses are to scale 1:12,000 for Frontier and below, 1:52,800 for Cody and above.

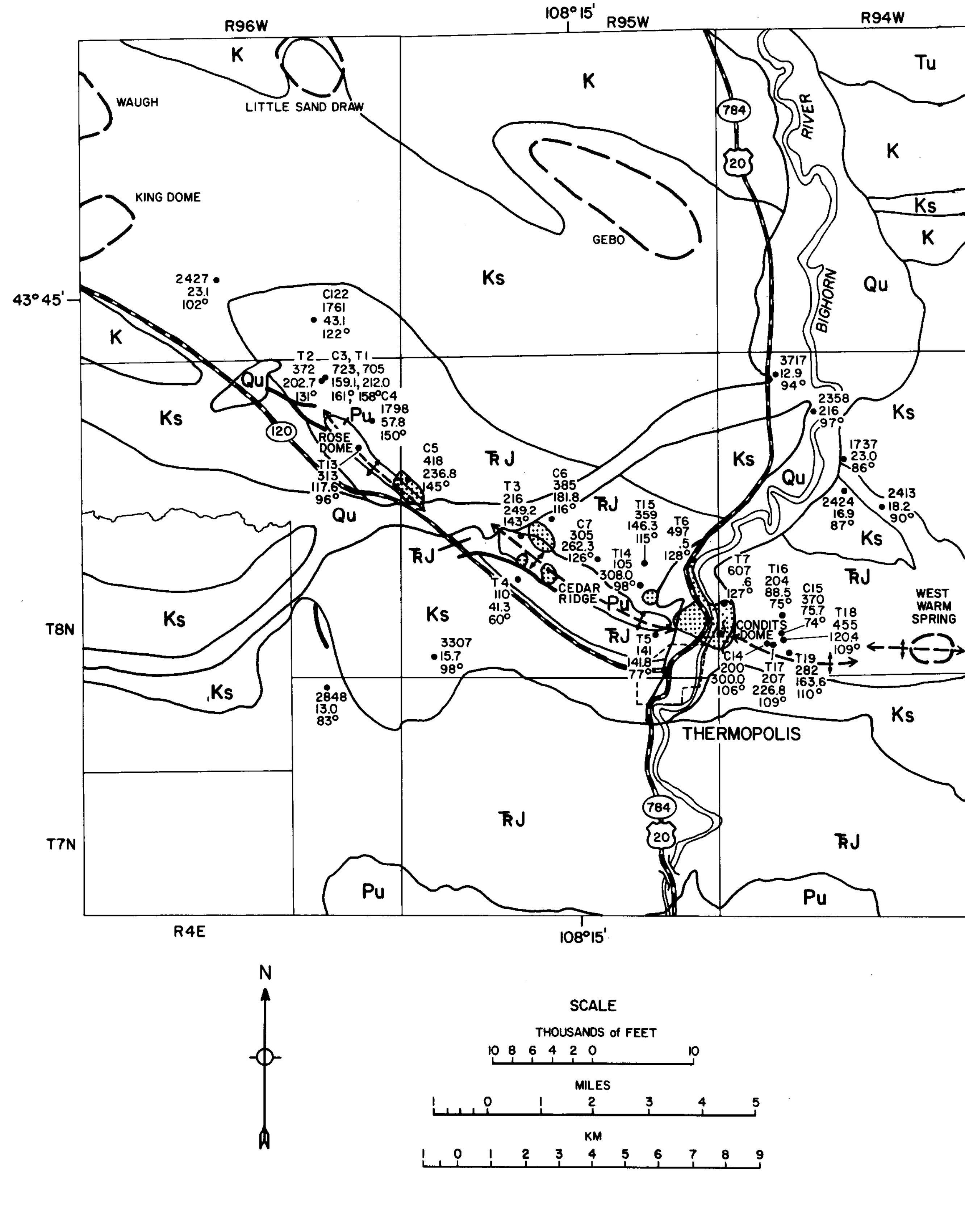
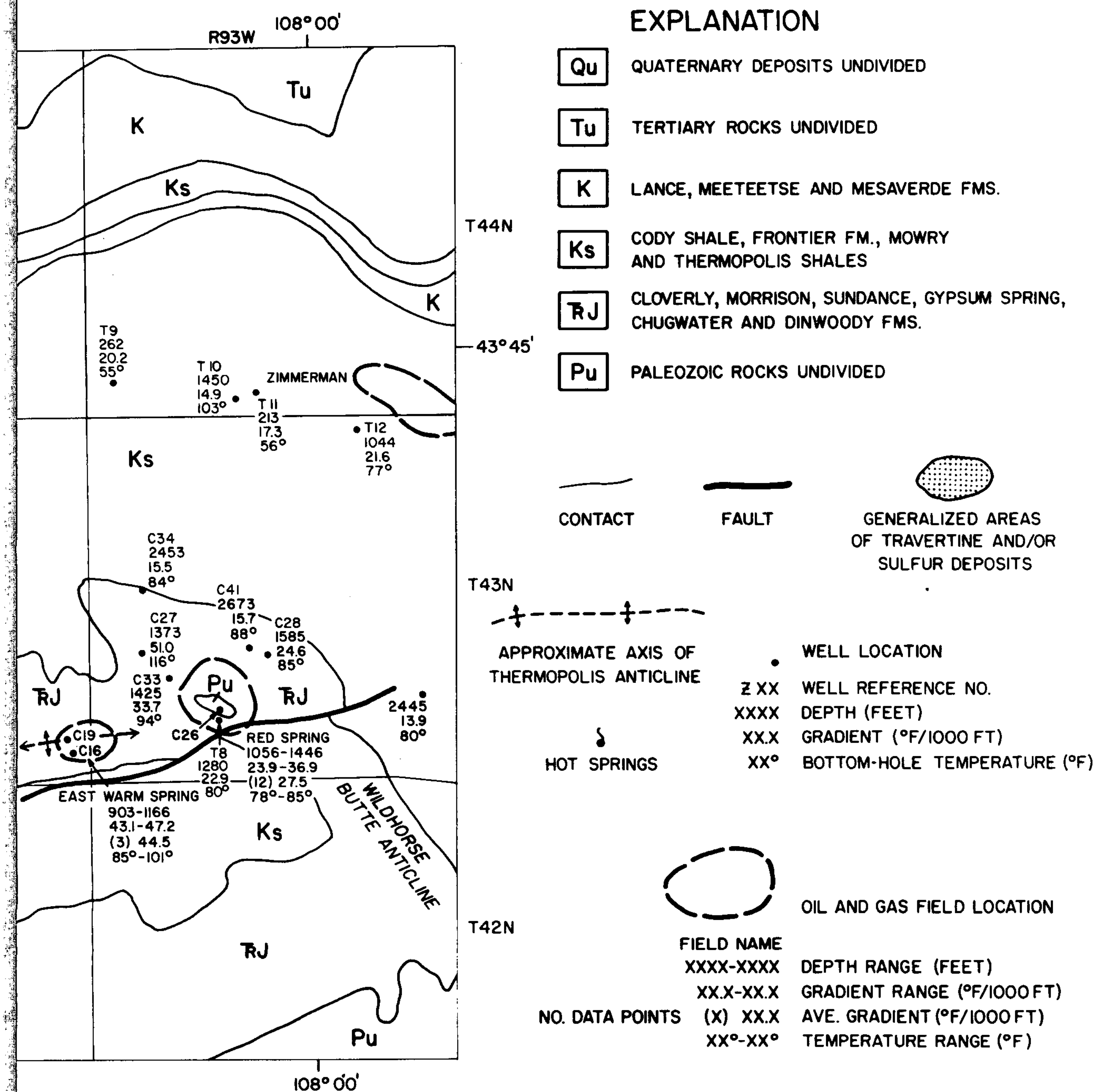


Figure 4. Geologic and thermal data for the Thermopolis Anticline area.



GEOLOGY AFTER: LOVE, J.D., CHRISTIANSEN, A.C., BOWN, T.M., AND EARLE, J.L., 1979, PRELIMINARY GEOLOGIC MAP OF THE THERMOPOLIS 1° x 2° QUAD., CENTRAL WYOMING: U.S. GEOL. SURVEY OPEN FILE REPORT 79-962, SCALE 1:250,000. LOVE, J.D., CHRISTIANSEN, A.C., EARLE, J.L., AND JONES, R.W., 1978, PRELIMINARY GEOLOGIC MAP OF THE ARMINTO 1° x 2° QUAD., CENTRAL WYOMING: U.S. GEOLOGICAL SURVEY OPEN FILE REPORT 78-1089, SCALE 1:250,000.

west side of Rose Dome, and possibly (contacts are covered) on the northwest end of Cedar Ridge; estimates of Chugwater thinning are 400, 500, and 200 feet, respectively. Thinning of the Morrison Formation is reported at the same location on Rose Dome (Lease and Palse, 1952), and thinning of the shales in the Sundance and Cloverly Formations on Rose Dome was observed by Summerford et al. (1947). There is photographic evidence for thinning of the Cloverly Formation and Thermopolis Shale on the steeply dipping flank of the northwest end of Cedar Ridge.

There is also abundant evidence of faulting along the Thermopolis Anticline. A pronounced reverse fault is evident on aerial photographs of the southern part of Rose Dome. Berry and Littleton (1961) did not plot a fault here, but they did indicate that the Sundance, Morrison, and Cloverly Formations are not present in this area, and plotted a locally wider outcrop of the Chugwater Formation. They did map a reverse fault on the south side of Cedar Ridge, where we found evidence of thinning of the Chugwater Formation. Aerial photography suggests that this fault could extend much further to the east along the base of steep Phosphoria and Dinwoody Formation outcrops. We also see a fault, of undetermined motion, on the steep flank of the anticline just north of Thermopolis. The eastward projection of this fault trace is between travertine-capped Monument Hill and Big Spring in Hot Springs State Park.

Hoppin (1974) has proposed that a lineament extends from the Bighorn Mountains east of the study area, along the Thermopolis Anticline, and on west to Hamilton Dome, suggesting that the anticline itself may be the result of a basement fault. Hamilton Dome is a structure very similar to the Thermopolis Anticline. Located 8 miles to the west-northwest (Plate 1), it appears as a down-plunge extension of the Thermopolis structure. The dome lacks surficial evidence of a major reverse fault, but it does have thinning of shale units on the steep flank (Krampert, 1947) and subsurface thinning of the Chugwater Formation (Berg, 1976). Berg concludes, from extensive oil and gas well logs, that Hamilton Dome results from a reverse fault cutting Paleozoic rocks and Precambrian basement. The fault is thought to be a zone of broken and sheared rock in discontinuous wedges, dipping at an angle of 60° or less north into the Bighorn Basin (Berg, 1976). The similarity in structural form of the Thermopolis Anticline and Hamilton Dome, shale thinning, adjacent location, and location along the same lineament, as well as the existence of reverse faults at the surface of the Thermopolis Anticline, strongly suggest that the Thermopolis structure is over a basement fault similar in structural style to that proposed by Berg for Hamilton Dome. Figure 5 incorporates this hypothesis into a cross section perpendicular to the northwest end of Cedar Ridge.

Yet another feature common to Hamilton Dome and the Thermopolis Anticline is small normal faults crudely perpendicular to the main structural axes. Krampert (1947) describes such faults on Hamilton Dome; aerial photographs and our field examinations revealed numerous short faults perpendicular to, but not cross-cutting, the Thermopolis Anticline axis. Differences in the positions and orientations of the rock strata on opposite sides of the Bighorn River indicate major faulting there and suggest that subsurface faulting may affect the pattern of domes and intervening saddles all along the anticline. Surface mapping to identify the nature of the apparent structural discontinuity across Owl Creek indicates that the Cedar Ridge and Rose Dome folds may be two separate folds plunging past one another. Structural relationships at depth may change under the influence of more presistent basement features. At this particular site, several north-trending folds impinge on Rose Dome, further complicating the subsurface geometry.

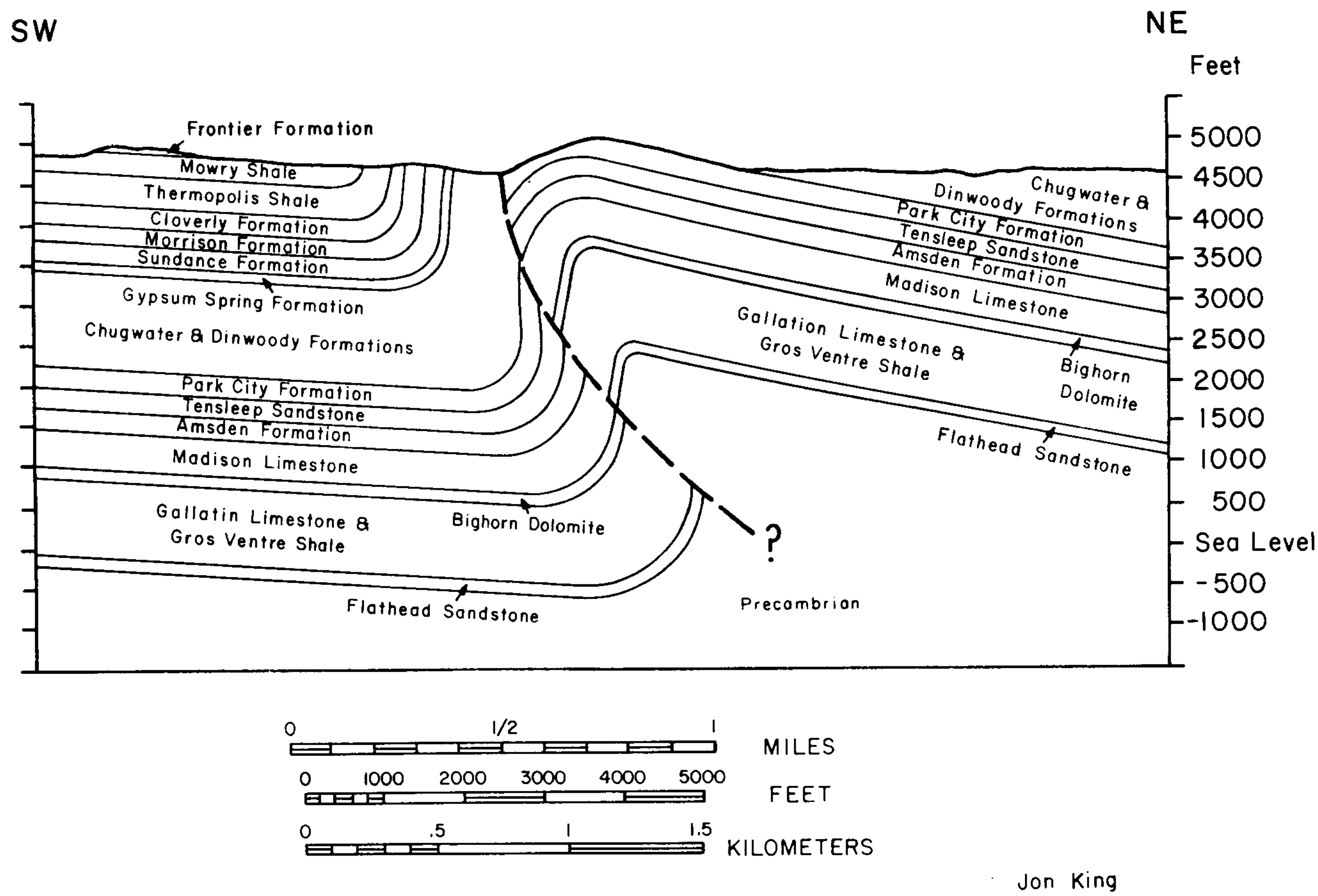


Figure 5. Generalized cross section of the Thermopolis Anticline at Cedar Ridge.

THERMAL INVESTIGATIONS

Thermal data for the Thermopolis area have been collected from three principal sources: (1) Bottom-hole temperature and depth measurements from over 400 well logs (available through the Wyoming Geological Survey and the Wyoming Oil and Gas Conservation Commission), (2) measurements from wells thermally logged as part of the present study, and (3) measurements of surface temperatures of springs and wells.

BOTTOM-HOLE TEMPERATURE DATA

Bottom-hole temperature (BHT) values were used to compute thermal gradients using the expression,

$$\text{Gradient} = (\text{BHT} - 46°\text{F})/\text{Depth}$$

46° Fahrenheit (7.8°C) being the mean annual air temperature of Thermopolis (Lowers, 1960). This is used an an approximation of mean surface temperature and is assumed not to vary significantly across the study area. A complete listing of all oil-field bottom-hole temperature data used in this report is available separately as Geological Survey of Wyoming Open File Report No. 82-3.

While various authors have used oil well bottom-hole data to calculate thermal gradients, the accuracy of such

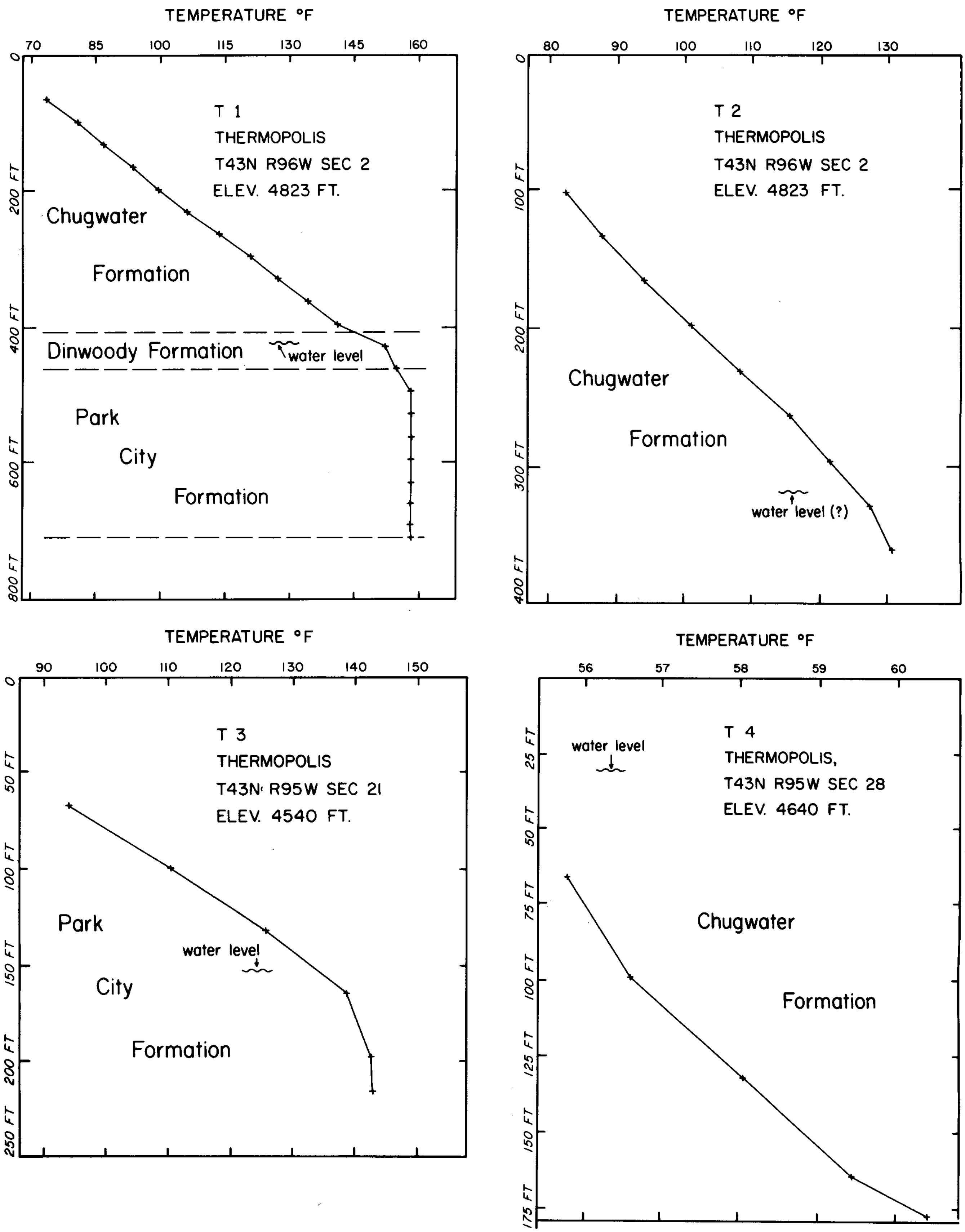
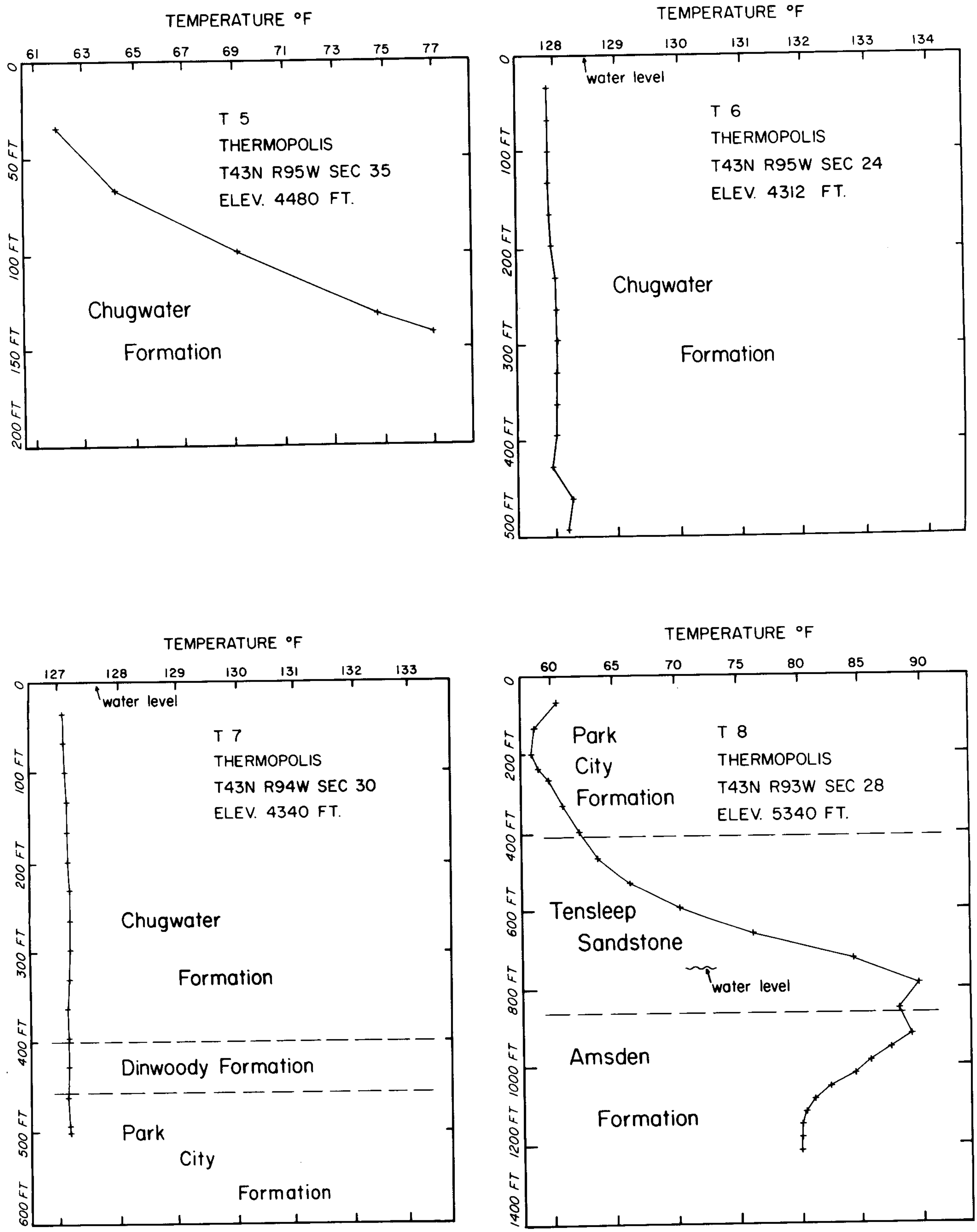


Figure 6. (This and following pages) Temperature-depth plots for boreholes in the Thermopolis area. See Figure 4 for borehole locations.



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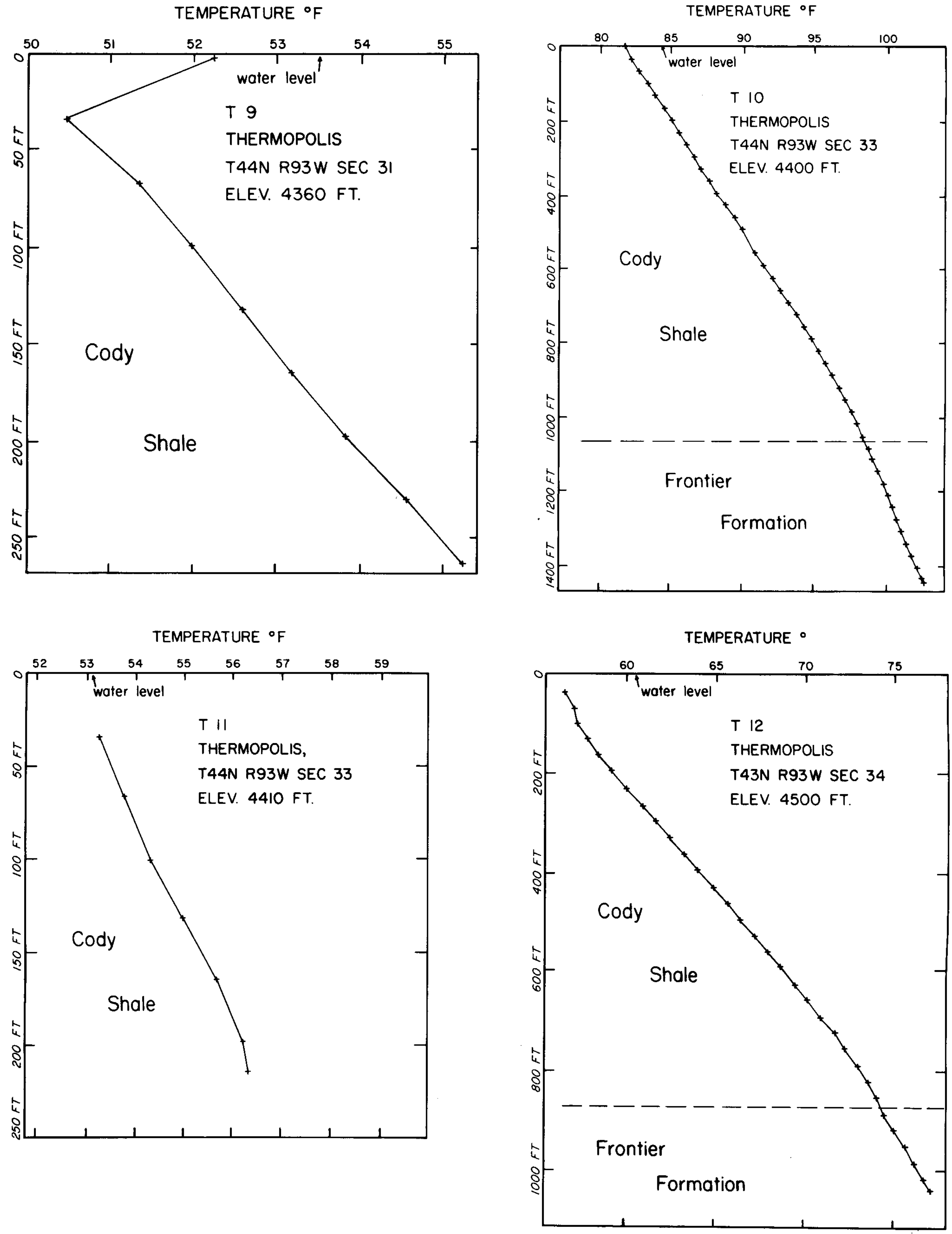

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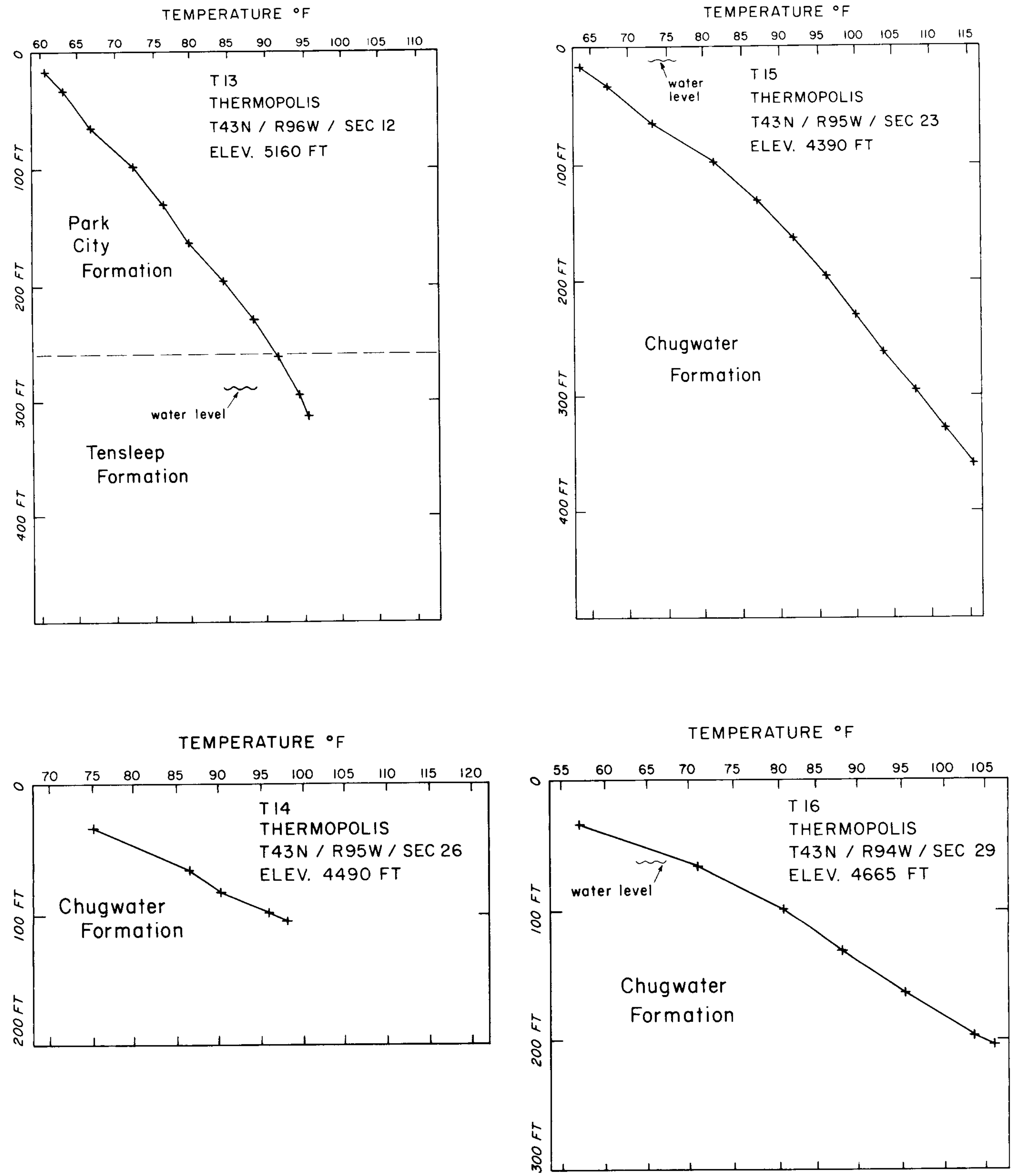

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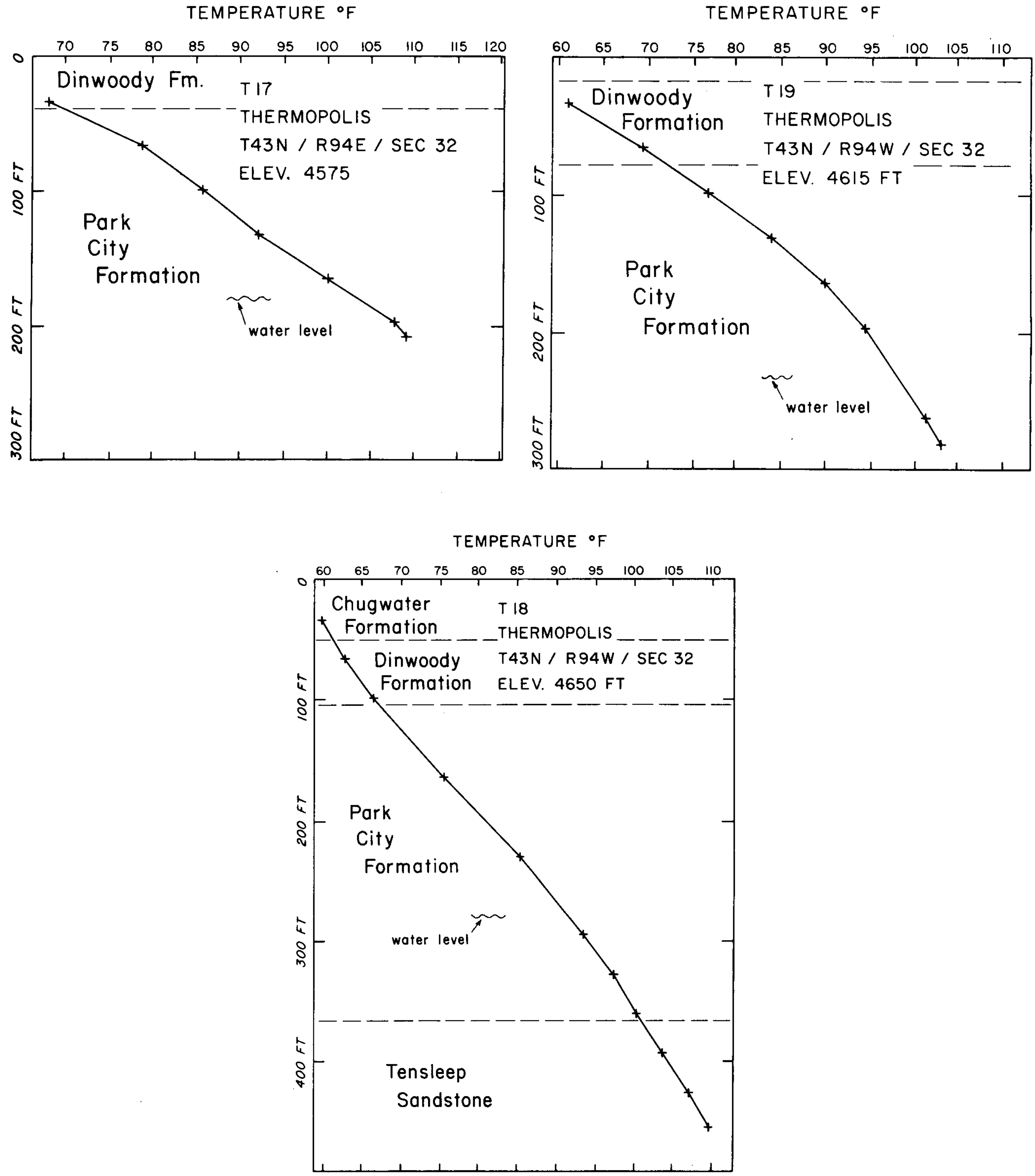

(Figure 6, continued)

data is highly variable. Basically, there are various drilling associated complications inconsonant with the simple use of all available bottom-hole temperatures. Heasler (1981) presents a discussion of techniques (both qualitative and quantitative) through which data can be filtered to arrive at a reasonably accurate assessment of thermally anomalous areas.

Gradients for the study area are plotted with generalized geology on Plate 1 and Figure 4. The gradients of Figure 4 are entirely derived from oil-field bottom-hole temperature data; Figure 5 includes thermally logged holes (denoted T) for which gradients were determined by statistical analysis of logged intervals. Since bottom-hole temperature data are only single, average, top-to-bottom gradients whereas thermal logs are measurements of many small gradient intervals, the two techniques may produce different results. The importance of gradient changes within a single hole is well illustrated by the temperature-depth plots in Figure 6.

Gradients derived from bottom-hole temperatures range from 8.1 to 300°F/1,000 feet in the study area, and highest gradients occur along the Thermopolis Anticline (43.1 to 300°F/1,000 feet), and the Red Spring Anticline (15.5 to 51.0°F/1,000 feet) (Figure 5). Along the Thermopolis structure, measured temperatures within 1,800 feet of the surface reach a maximum of 161°F (71.7°C). Data from Red Springs include temperatures as high as 116°F (47°C) at depths of less than 1,600 feet. It is difficult to pick out a single value for a "normal" gradient, but thermal data from throughout the Bighorn Basin gives an average gradient of 15.4°F/1,000 feet.

The high gradients observed on the Thermopolis and Red Spring Anticlines, if coupled with favorable geologic and hydrologic conditions, could represent viable geothermal resource areas. The area of high gradients along the Thermopolis Anticline, from the southeast part of T.44N., R.96W. to the southwest part of T.43N., R.93W., identifies the "resource area" of this report. The Red Spring Anticline to the east may be a marginal resource area. The well data of Table 2 include both Thermopolis and Red Spring areas.

Gradient and temperature distributions within and around the resource area provide evidence for two additional features of this hydrothermal system: (1) There is a general decrease in maximum temperature and gradient from west to east along the anticline. The maximum bottom-hole temperature for Rose Dome is 161°F (72°C); for Cedar Ridge, 143°F (52°C); for Condits Dome, 106°F (41°C); for East Warm Spring, 101°F (38°C); and for Red Spring, 116°F (47°C). (2) The high temperatures and gradients are closely confined to the crestal portion of the anticline. Five to six miles northeast, along the Bighorn River, gradients range only from 12 to 23°F/1,000 feet, and 5 to 10 miles southwest, gradients have dropped to 12 to 25°F/1,000 feet. The Red Spring structure shows similar gradient differences with gradients of 16 to 25°F/1,000 feet less than 2 miles northeast of the anticline axis.

As can be seen on Plate 1, we do not have a continuous grid of gradient data. Thus, our definition of high and low gradient areas can only be as fine as the local data spacing. The structure of the Wildhorse Butte Anticline, for example, suggests that it may be an extension of the identified marginal resource area at Red Spring, but no temperature data were found for Wildhorse Butte. Tom Anderson and Norman and Tom Sanford (personnal communication, 1979) have mentioned "hot" water wells at Black Mountain and Kirby Creek oil fields. These areas may also be marginal resource areas, but here, again, insufficient data are available to properly evaluate that possibility.

Table 2. Well data for the Thermopolis Anticline.[1]

Well No.[2]	Bottom-hole temp., °F(°C)	Depth, feet	Formation	Well No.[2]	Bottom-hole temp., °F(°C)	Depth, feet	Formation
C122	122(50)	1,764	Tensleep	T17	109(43)	207	Park City
C3	161(72)	723	Park City	C15	74(23)	370	Park City
T1	158(70)	705	Tensleep	T18	109(43)	455	Tensleep
T2	131(55)	372	Chugwater	T19	110(43)	282	Park City
C4	150(66)	1,798	Gallatin	C16	101(38)	1,166	Tensleep
T13	96(36)	313	Tensleep	C19	85(29)	903	*unknown*
C5	145(63)	418	Tensleep	C26	85(29)	1,056	*unknown*
T3	143(62)	216	Park City	T8	80(27)	1,280	Amsden
C6	116(47)	385	Tensleep	C27	116(47)	1,373	Tensleep
T4	60(16)	110	Chugwater	C28	85(29)	1,585	Park City
C7	126(52)	305	Park City	C33	94(34)	1,425	*unknown*
T15	115(46)	359	Chugwater	C34	84(29)	2,543	Tensleep
T14	98(37)	105	Chugwater	C41	88(31)	2,673	Madison
T5	77(25)	141	Chugwater	T9	55(13)	262	Cody
T6	128(54)	497	Chugwater	T10	103(40)	1,450	Frontier
T7	127(54)	607	Park City	T11	56(14)	213	Cody
C14	106(41)	200	Park City	T12	77(25)	1,044	Frontier
T16	75(24)	204	Chugwater				

[1]Bottom-hole temperatures and depths are from oil and gas well logs (C) or thermal logging (T); formations are from well logs, Petroleum Information cards, or extension from nearby wells. Temperature-depth plots for all wells thermally logged are shown in Figure 6.

[2]See Figure 4, pages 8 and 9, for locations.

THERMAL LOGGING DATA

Drill holes thermally logged by University of Wyoming personnel, although much less numerous than oil well bottom-hole temperatures, provide valuable, quantitative checks on oil well data and allow careful study of gradient variation with depth. Nineteen holes, from 67 to 1,250 feet deep, have been logged in the Thermopolis area. These logs are presented in Figure 6, along with annotations of stratigraphy and water level (see Figure 4 for their locations). The four holes west of the Zimmerman oil field (T9-T12) confirm the "normal" gradients found in the bottom-hole temperature data for that area. Their plots show a systematic temperature increase with depth (except for the shallow, seasonal thermal disturbances recorded near the tops of the holes). Holes logged on Rose Dome and Cedar Ridge (T1-T3, T13-T15), on Condits Dome (T16-T19), and in the Red Spring oil field (T8), similarly substantiate the anomalous gradients cited above for these areas (e.g., 158°F (70°C) was thermally logged in well C3, while the reported bottom-hole temperature was 161°F (72°C)).

An important feature of holes in the resource area can be seen in logs T1-T3 and T8: temperatures increase rapidly with depth, as expected, but abruptly cease to rise below a certain depth. That critical depth closely coincides with the water level in the hole. This information clearly demonstrates the danger of simply extrapolating measured or calculated gradients downward to estimate deeper temperatures. More important, the isothermal character of the water over a range of depths strongly suggests that water is circulating within the aquifers, homogenizing temperatures as heat is added from depth.

In contrast, the three holes near the Zimmerman field, though full of water, show steady temperature increase with depth. The difference is easily explained by differences in lithology: the Zimmerman holes were drilled almost entirely in the Cody Shale, a relatively impermeable unit, cut off from much deeper zones by a thick sequence of low-permeability formations (see Figure 3). The holes along the anticline, however, were drilled into productive aquifers of the Paleozoic section. (The hydrologic characteristics of these formations will be discussed in a later section). An exception to the isothermal pattern once significant water is encountered along the Thermopolis Anticline is thermal log T17. In this case temperatures continue to increase down a 150-foot water column in the Park City and uppermost Tensleep Formations. The bottom-hole temperature in this well, however, agrees with bottom-hole temperature in much shallower wells in the area (logs T18, T19). Our interpretation is that well T17 penetrates an unfractured zone of low permeability through which thermal waters do not circulate. That heavy oil is found in these strata and that this oil has produced only very poorly even under steam drive (Tom Sullivan, personnal communication, 1981) are evidence that this portion of Condits Dome is an area of low permeability.

We were unable to log the springs of Hot Springs State Park, but did thermally log two of the hot flowing wells north of the springs: McCarthy #1 and Maytag (see Figure 2 and Table 1). The flow of hot water in these wells is from Paleozoic formations. Although there are many wells in the immediate area [Wyoming State Engineer's files], only those penetrating through the Chugwater Formation receive hot water flow. Since water yield data are very sparse for Thermopolis area aquifers, identification of the formation(s) supplying these wells is important. Unfortunately, reported depths for the hot, flowing wells vary with author: those depths we judged most reliable, including well logs and records made near the time of drilling, place the McCarthy wells at 510 and 450 feet in the Park City Formation (Bartlett, 1925), and the Van Norman well at 550 feet in the Park City Formation [Wyoming State Engineer, Permit #P470C].

The 790-foot depth reported for the Skidmore #3 well, next to the Van Norman well, suggests a Tensleep completion, but well logs list "limestone" (Park City?) as the water source [Wyoming State Engineer files; Permit #P471C]. Breckenridge and Hinckley (1978) quote local sources as remembering the Maytag and Sacajawea wells to be 900 feet deep, which would place them into the Tensleep Sandstone. This agrees with the report of Stearns et al. (1937) of hot Tensleep wells "north of Thermopolis," but conflicts with Collier's (1920) description of the Sacajawea(?) well as being only 498 feet deep, indicating production from the lowermost Chugwater Formation. We were unable to lower a probe beyond 497 feet in the Maytag well and found the Sacajawea well to be obstructed by mineral deposits at 8 feet in February 1981 (see Figure 6 for temperature-depth plots). Thus, the two most productive wells can only be designated as Park City <u>or</u> Tensleep producers.

Six additional, relatively shallow holes further witness high temperatures near the surface. A collapsed sulphur exploration hole on the north side of Cedar Ridge has a measured temperature of 98°F (37°C) at a depth of only 67 feet. Two wells logged in the Chugwater Formation on the southwest flank of Cedar Ridge (T4, T5) have temperatures of 60°F (18°C) and 70°F (25°C) at 150 feet and 141 feet, respectively. Wells T15 and T14 on the north flank of Cedar Ridge and T16 on the north flank of Condits Dome have temperatures of 115°F, (46°C) 98°F (37°C), and 75°F (24°C) at 360, 204, and 104 feet, respectively (see Figure 4 for locations).

THERMAL DATA FROM SPRINGS AND WATER WELLS

The temperatures of the principal springs of Hot Springs State Park are 127 to 133°F (53 to 56°C) and temperatures from the flowing wells north of town are 124 to 128°F (51 to 54°C) (this study; Breckenridge and Hinckley, 1978). Since these wells all reach Paleozoic aquifers, and since similarly hot waters are encountered in the Paleozoic section all along the Thermopolis Anticline, we infer that the springs also originate in formations below the Chugwater Formation. The coincidence of the springs with the most steeply folded portion of the anticline, the proposed trace of a major basement fault, and the possible existence of a series of transverse, normal faults (see Geology section) suggest that a fracture-supplied conduit for sub-Chugwater waters is most probable. If hot waters are circulating in the upper Paleozoic strata, adjacent water-bearing beds in the Chugwater Formation off the anticline should be warm, but without hydraulic communication. Evidence supporting this contention is the common occurrence of nonpressurized, warm waters in the Chugwater Formation north of town (Van Norman, personal communication, 1981), the high thermal gradients (average 145°F/1,000 feet) logged in holes in the Chugwater Formation on the flanks of the anticline (T4-T5, T14-T16), and a 70°F (21°C) Chugwater water well temperature measured on the west end of town (T5).

There are also springs in the Thermopolis area from Mesozoic formations. We have measured temperatures, ranging from 50 to 53°F (4 to 12°C) in six of these, which indicate only shallow circulation of probably locally derived groundwater.

HEATING MECHANISMS AND THERMAL MODELING

Qualitative explanations of the temperature of the thermal springs of Hot Springs State Park fall into three general categories: (1) heating by a young, buried igneous mass, (2) heating by exothermic chemical reactions within the rocks, and (3) conductive heating of groundwater at depth coupled with upward migration due to artesian and convective forces. (see Breckenridge and Hinckley, 1978, for a historical summary).

We know of no evidence for igneous activity in the area. The nearest volcanic rocks are 30 miles west of Thermopolis and the nearest known intrusive rocks are 20 miles further west. By calculating the heat loss over time for a hypothetical intrusion beneath Thermopolis of the same age as the known igneous activity to the west, (after Carslaw and Jaeger, 1959 and Jaeger, 1964) we have concluded that such a heat source would have thoroughly dissipated by the present even if it were there. Laughlin and Aldrich (1981) similarly conclude that igneous rocks older than 3 million years have probably lost their original heat. Thus, the nearest known igenous rocks young enough to support present thermal anomalies are in Yellowstone National Park, over 100 miles to the northwest. (In this context, we note that the heat diffusing from a deep igneous mass would produce a much broader thermal anomaly than the narrow band seen along the Thermopolis Anticline. To generate the observed heat distribution magmatically would require a relatively shallow intrusion with an elongate geometry coincident with that of the anticline and a temperature increasing to the northwest).

The idea of heating by chemical reaction has only been proposed in general terms, e.g. by Bartlett (1926). But a flow of over $2\frac{1}{2}$ billion gallons a year has not significantly reduced temperatures in this century (Breckenridge and Hinckley, 1978); no one has proposed specific reactions capable of producing the 200 million BTU/hour necessary to warm the flow of existing wells and springs; and, most important, water from the same formation but different structural settings varies significantly in temperature (e.g., Madison temperatures at Red Springs are 67°F (37°C) cooler than at Rose Dome). These facts all suggest that chemical heating is at most of auxiliary importance.

In 1906, Darton proposed a simple model of the Thermopolis spring system consisting of: (1) surface water recharge of northward dipping Paleozoic aquifers in the Owl Creek Mountains, (2) confinement of this northward flowing water by much less permeable beds in the overlying Chugwater Formation, (3) heating of the water by normal conductive gradients in the deepest portions of the syncline, and (4) rising of water by artesian pressure to flow at the surface where the Chugwater Formation is breached by fracturing along the crest of the Thermopolis Anticline. Such a system is illustrated diagrammatically in Figure 7.

We were able to make a quantitative evaluation of this model, based on heat flow and rock conductivity measurements. For all calculations, the heat flow was taken to be 1.75 μcal/cm^2sec. This is the mean of values obtained for the Owl Creek Mountains by Decker et al. (1980) and for the Gebo oil field by Blackwell (1969). The ground surface temperature was assumed to be 46°F (7.8°C). The formation thicknesses and thermal conductivities used, along with the predicted temperature for each formation, are tabulated in Table 3.

The temperatures of Table 3 are based on "steady-state" or equilibrium conditions. Any process which, over geologic time, changes the surface temperature will also have affected geothermal gradients. Evaluation of conditions in the Bighorn Basin during the last 10 million years as reported by Mackin (1936,1937) and Ritter (1975) suggests that the most thermally disruptive situation which

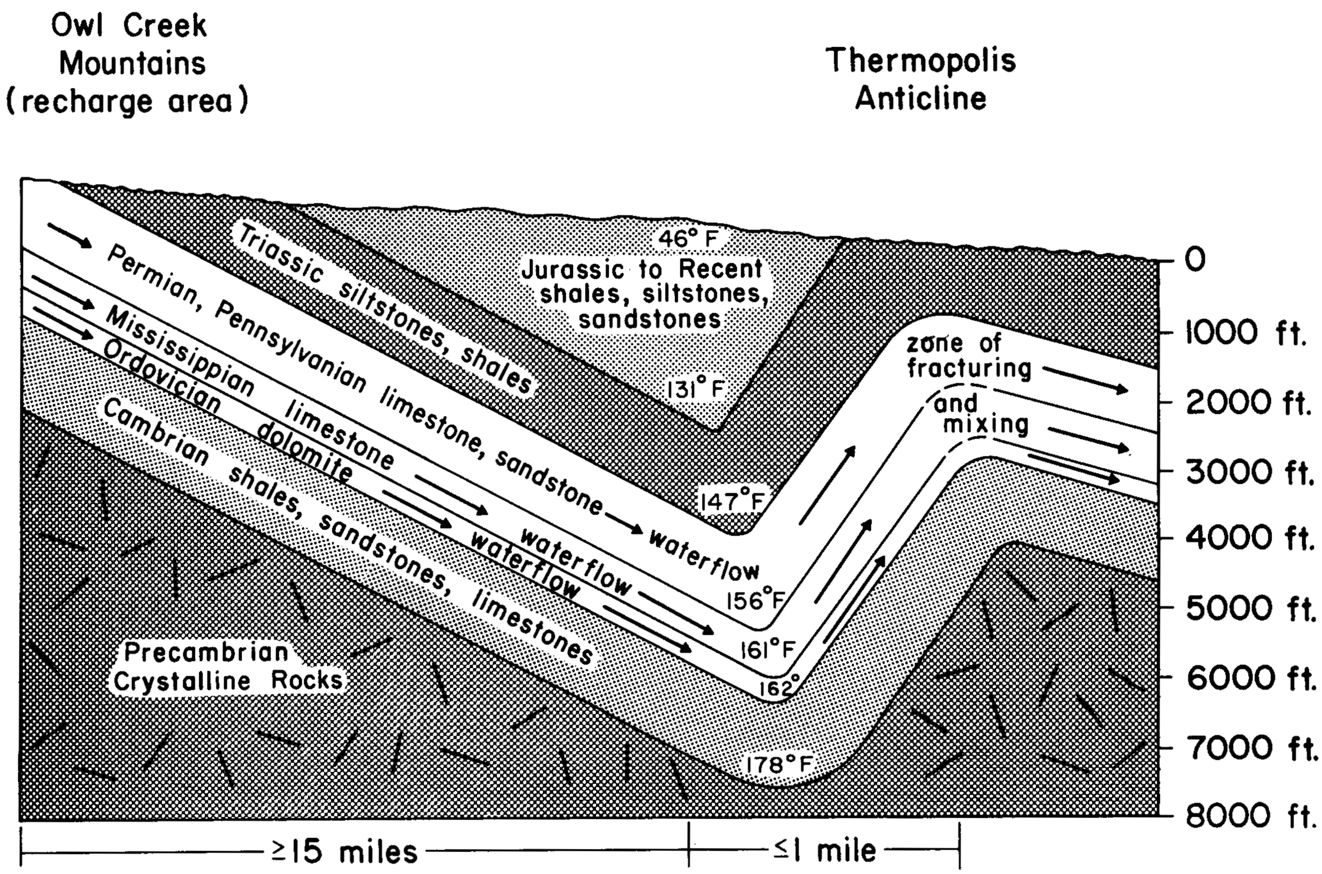

Figure 7. Diagrammatic cross section of the simplest model for the Thermopolis hydrothermal system (temperatures and depths from the Rose Dome thermal model).

is geologically reasonable is approximately 6,000 feet of regional uplift and 3,000 feet of erosion uniformly distributed over the past 4 million years. Using the commonly accepted value of 32 km^2/million-years for the sediments' thermal diffusivity and analytical expressions from Benfield (1949), uplift and erosion would result in gradients at depths greater than 2,600 feet no more than 12.6 percent higher than those based on equilibrium modeling. This same "maximum" deviation translates into actual temperatures 5.6°F (3.1°C) and 15.8°F (8.8°C) higher than those modeled at 2,600 and 7,800 feet, respectively. Heasler (1978) has addressed the effects of glaciation and of 3,000 feet of erosion distributed over just the last 2 million years for the Bighorn Basin and has calculated deviations from equilibrium smaller than those cited above. In summary, we believe that the temperatures of Table 3 are geologically reasonable estimates, and that glaciation and erosion would have raised temperatures only slightly even in the extreme cases discussed.

The highest temperature measured in the Cedar Ridge vicinity is 133°F (56°C) at Big Spring. In a syncline-anticline flow system like the one depicted in Figure 7, oriented perpendicular to the Thermopolis Anticline at Cedar Rige, modeling predicts

Table 3. Thermal models at Thermopolis and Rose Dome.

Formation	Formation thickness[1], feet*		Thermal Conductivity: $\frac{10^{-3}\text{cal}}{\text{cm}^\circ\text{C sec}}$	Temperature increase in formation, °F (°C)[6]	Temperature at bottom of formation, °F (°C)[7]	
	Rose Dome	Thermopolis			Rose Dome	Thermopolis
Cody Shale	1,565		4.0[3]	37.6(20.9)	83.7(28.7)	
Frontier Fm.	850		4.4[2,3,5]	18.5(10.3)	103.3(39.0)	
Mowry Shale	280	250	3.9[2]	6.8 (3.8) 6.1 (3.4)	109.0(42.8)	52.2 (11.2)
Thermopolis Shale	400	400	6.1[2]	6.3 (3.5)	115.3(46.3)	58.5 (14.7)
Cloverly Fm.	240	240	8.7[2]	2.7 (1.5)	118.0(47.8)	61.2 (16.2)
Morrison Fm.	230	230	6.3[2,5]	3.4 (1.9)	121.5(49.7)	64.4 (18.1)
Sundance Fm.	250	250	7.4[2]	3.2 (1.8)	124.7(51.5)	67.8 (19.9)
Gypsum Spring Fm.	155	155	2.4[2]	6.1 (3.4)	130.8(54.9)	73.9 (23.3)
Chugwater Fm.	1,100	1,100	7.2[2]	14.6 (8.1)	145.4(63.0)	88.5 (31.4)
Dinwoody Fm.	55	55	2.8[2]	2.0 (1.1)	147.4(64.1)	90.5 (32.5)
Park City Fm.	260	260	9.6[2]	2.7 (1.5)	150.1(65.6)	93.2 (34.0)
Tensleep Sandstone	280	280	10.4[2]	2.5 (1.4)	152.6(67.0)	95.7 (35.4)
Amsden Fm.	280	280	8.0[3,5]	.4 (1.9)	150.0(68.9)	99.1 (37.3)
Madison Limestone	480	480	9.6[2]	3.6 (2.7)	160.9(71.6)	104.0 (40.0)
Bighorn Dolomite	130	130	11.0[4]	1.1 (0.6)	162.0(72.2)	105.1 (40.6)
Gallatin Limestone	470	470	7.4[4]	6.1 (3.4)	168.1(75.6)	111.2 (44.0)
Gros Ventre Shale	510	510	6.0[3,5]	8.3 (4.6)	176.4(80.2)	119.5 (48.6)
Flathead Sandstone	150	150	8.5[3,5]	1.6 (0.9)	178.0(81.1)	121.1 (49.5)
Granite and Gneiss	?	?	7.9[3]			
TOTAL	7,685	5,240				

[1]After well logs in Horn (1963), Ary (1959), Collier (1920), Fanshawe (1939), Maughan (1972a, 1972b), Shelmon (1959), Berry and Littleton (1961), Annonymous (1952), Mees and Bowers (1952). [2]Heasler (1978). [3]Garland and Lennox (1962). [4]Sass and others (1971). [5]Estimate based on composition of rock unit. [6]Using a heat flow of 1.75×10^{-6} cal/cm^2 sec. [7]Assuming a 46°F (7.8°C) ground surface temperature (Lowers, 1968).

*rounded to nearest 5 feet.

that 133°F (56°C) would be reached in the Precambrian basement rocks, and that the temperature at the base of the Madison would be 104°F (40°C). As noted earlier, however, the syncline plunges northwest, providing greater depths and higher temperatures in that direction. A similar calculation for the syncline-anticline system at Rose Dome, 8 miles northwest of Thermopolis, reveals that temperatures in the Park City Formation should be greater than the observed spring temperatures, that the 161°F (72°C) temperatures reported from nearby well C3 (Figure 5) could be produced from the base of the Madison, and that waters circulating to the bottom of the Paleozoic section should be 178°F (81°C).

To obtain a more accurate idea of actual temperatures within the Thermopolis hydrothermal system, several adjustments may be made to the simplified model of Figure 7. First, as Figure 5 shows, the deepest part of the syncline may be deeper than has been modeled in Figure 7; temperatures would be correspondingly higher than those calculated above. Second, if there is a major, deep fault as indicated, it may provide a means of much deeper circulation and higher temperatures than those possible within the folded sedimentary aquifer model. Fault-increased permeabilities have already been proposed as controlling the location of thermal springs and spring deposits in the Thermopolis area (p. 18, 19, 27); increased permeability may also decrease cooling of ascending waters by allowing rapid access to near-surface zones. Third, our thermal logs indicate that gradients in the Thermopolis area may be equilibrated to a surface temperature as much as 14°F (7.8°C) warmer than the 46°F (7.8°C) used in the preceding calculations; increasing calculated temperatures by a like amount may be appropriate.

We feel that, with the modifications outlined above, the basic heating model of Darton (1906) is substantially correct. While one cannot absolutely exclude igneous and chemical heat sources, the relative simplicity of an artesian, syncline-anticline system and it's agreement with the observed temperature and gradient distributions indicate this as the predominant heating mechanism.

Another model was calculated to estimate the necessary enthalpy transfer for the Thermopolis system. In this model the enthalpy of the volume of water equal to the total surface discharge of the hydrothermal system (4,681 gallons per minute), brought from 32°F (0°C) to 167°F (80°C) (Kittel, 1969; Handbook of Chemistry and Physics, 1968) was used to calculate the area needed to supply the required heat. Using a heat flow of 1.75 $\mu cal/cm^2 sec$, an area of 34 square miles would be needed to heat the observed flow of water. This should be considered a minimum area since it is unlikely that the existing springs constitute the total output of the hydrothermal system. The fact that over 500 square miles are probably contributing heat to waters enroute to the Thermopolis anticline, however, suggests that potential heating area is not a limiting factor in this case.

HYDROLOGY AND WATER CHEMISTRY

Basically, groundwater flows from areas of recharge to areas of discharge. Flow patterns are primarily determined by the ability of the subsurface materials to transmit water (permeability) and by the forces "pushing" the water, namely the confining effects of surrounding water and rock and the difference in hydrostatic head between the recharge area and the discharge area. As surface recharge water moves into and through the earth, it is modified by the minerals, temperatures, and pressures encountered. By considering the rock units and structures through which water passes, we can evaluate the water yields and quality likely to be developed from various sources; conversely, measured yields and chemistry can be used to identify sources. Pressure and thermal data can be used to evaluate water flow patterns and to predict available temperatures and pumping lifts.

Recharge for the Thermopolis hydrothermal system is generally believed to occur on the north flank of the Owl Creek Mountains where precipitation and surface water enter northward-dipping strata. Surface discharge occurs at the springs in Hot Springs State Park (see e.g. Darton, 1906; Berry and Littleton, 1961; Blackstone, 1971; Bredehoeft and Bennett, 1972). While there is considerable room for discussion on water pathways within the Paleozoic rocks, there is consensus that relatively little flow moves through the Chugwater Formation, and that the Chugwater Formation "generally limits upward movement of groundwater from Paleozoic aquifers" (Cooley, 1981). We do not have permeability data for direct comparison of Chugwater shales and Park City limestones, but permeabilities for similar rock types (see, e.g., Freeze and Cherry, 1979, p. 29) suggest that permeability differences of many orders of magnitude are possible. Breckenridge and Hinckley (1978) cite the importance of the Triassic Chugwater Formation as a "cap" on hydrothermal systems statewide, and the limited drilling data for the Thermopolis area indicate a similar condition (see discussion, p. 19). Therefore, we feel justified in restricting the Thermopolis discussion to Paleozoic strata. (Figure 3 presents general hydrologic data for all units in the area. Libra et al. (1981) present a thorough hydrologic discussion for the entire Bighorn Basin).

AQUIFER DESCRIPTIONS

The first viable aquifer beneath the Chugwater Formation is the PARK CITY FORMATION (Phosphoria, Embar), a thin- to thick-bedded sequence of dolomitic limestone and dolomite with some mudstone (Maughan, 1972a). Whereas sandstone owes its ability to transmit water to pathways around and between the constituent mineral grains (primary permeability), limestone and dolomite develop secondary permeability through fractures and solution openings. Fractures tend to develop in response to rock stress, as do folds and faults. Solution features develop as rock is dissolved away by flowing groundwater, commonly along bedding planes and fracture zones. The result is a very heterogeneous permeability distribution. This is reflected in 52 oil well drill stem tests of the Park City Formation throughout the study area (Petroleum Information, 1981) which recovered anywhere from 0 to 3,758 feet of water in tests during flow periods generally between 60 and 120 minutes.

Due to the high mineral content of water from the Park City Formation, water supply wells into the formation are not common. Flows from ten reported flowing wells from the Park City Formation vary from <1gpm to the 529 gpm flow of the McCarthy #1 hot well and 687 gpm for an oil well reported by Crawford (1940) three miles south of Thermopolis. A Park City spring at the mouth of Wind River Canyon flows 989 gpm (Breckenridge and Hinckley, 1978).

Water is being produced with oil from the Park City Formation at the Hamilton Dome, Kirby Creek, Lake Creek, Gebo, Little Sand Draw, Warm Springs, Waugh Dome, and Zimmerman Butte oil fields in water to oil ratios of up to 40:1 water:oil (Biggs and Espach, 1960). Oil production from the Park City Formation has also occurred from

Table 4. Water chemistry for the Thermopolis study area: mean/coefficient-of-variation. Mean values in mg/l. n = number of samples. All data taken from Wyoming Geological Survey Open File Report No. 82-3.

	Park City Formation	Tensleep Sandstone	Madison Limestone	Hot springs and wells
n	23	26	7	14
Ca	406/0.65	192/1.09	318/0.64	353/0.80
Mg	105/0.55	45/0.80	86/0.69	81/0.18
Na+K	2,561/1.26	402/1.80	230/0.68	299/0.13
HCO_3	1,223/1.29	561/1.61	697/0.56	732/0.06
SO_4	3,549/0.96	863/1.86	743/0.57	754/0.12
Cl	1,634/1.67	254/2.53	223/0.92	301/0.22
TDS	8,866/1.07	1,913/1.40	1,950/1.40	2,317/0.03

14 samples from hot springs and wells		3 samples from hot springs and wells	
Na	249/0.20	As	<.5
K	51/0.35	Ca	<.01
CO_3	0	Mn	<.05
F	4.8/0.31	Zn	<.02
NO_3	0.4	Ba	<.5
SiO_2	40/0.24	Cd	<.01
S	.001/1.30	Cr	<.1
B	54/0.37	Pb	<.1
Fe	.20/2.26	Se	<.001
pH	6.9/0.04	Ag	<.5
H_2S	2.7/0.58	Hg	<.001
		Ni	<.1

the Wildhorse Butte structure and from two small folds northeast of the Murphy and Zimmerman fields (Horn, 1963). Reports describe heavy oil saturation but no production from the Park City Formation at Cedar Ridge (Summerford et al., 1947) and Condits Dome (Ary, 1959). Libra et al. (1981) cite oil field determinations of porosity (5-24 percent), permeability (0.61-76 millidarcies), and transmissivity (0.9-40 gpd/foot) for the Park City Formation in the Thermopolis area.

Twenty three water chemistry analyses for the Park City Formation within the study area (Figure 4) are on file [Geological Survey of Wyoming Open File Report No. 82-3]. Major cation and anion averages and coefficients of variation for these samples are presented in Table 4, p. 26. As the high coefficients of variation indicate, chemical concentrations for analyzed Paleozoic formation water vary greatly since the chemical data comes from a variety of structural and hydrologic settings.

One finds generally less variation when only the proportions of ions are considered. Crawford (1963) identifies a Ca:Mg ratio of around 4:1 and a SO_4:Cl ratio greater than 1 as characteristic; he remarks on the great range of total solids contents and notes the frequent occurrence of H_2S gas in Park City water in the Bighorn Basin. The H_2S is the result of bacterial sulphate reduction and is associated with higher CO_2 content as well (Crawford, 1963).

Below the Park City Formation is the TENSLEEP SANDSTONE, fine- to medium-grained, generally calcareous sandstone with some dolomite and sandy dolomite beds (Maughan, 1972a). Primary permeability in the Tensleep Sandstone varies somewhat due to variation in cementation (Todd, 1963), and is substantially added to by secondary permeability in zones of fracturing (Berry and Littleton, 1961). One qualitative indication of the generally greater permeabilities of the Tensleep Sandstone than of the Park City Formation is that in the 16 Tensleep drill stem tests examined, 12 recovered from 1,500 to 8,905 feet of water in flow periods of from 15 to 160 minutes. Oil and water are produced from the Tensleep Sandstone at the Gebo, Little Sand Draw, Hamilton Dome, Lake Creek, and Murphy Dome oil fields; Kirby Creek and Waugh Dome report only water in the Tensleep Sandstone (Biggs and Espach, 1960). The Tensleep Sandstone is the main oil producer in the Black Mountain field and is reported to contain water in the Warm Springs and Zimmerman fields, in structures adjacent to the Murphy and Zimmerman fields (Horn, 1963), and in the Owl Creek Anticline, 15 miles west of Thermopolis (Lease and Palse, 1952). The only reports of tests which found no water in the Tensleep Sandstone are from the Wildhorse Butte Anticline (Horn, 1963) and Cedar Ridge (Summerford et al., 1947), though heavy oil saturation was reported in the latter case. It should be borne in mind that oil field data for the study area is confined to anticlines and domes, which are especially likely to experience fracture-increased permeability. Mees and Bower (1952), for example, report the Tensleep Sandstone to be hard and tight on the gentle north flank of the Gebo Anticline but fractured on the steep south limb.

The Tensleep Sandstone has not been much developed for water supply in the Thermopolis area. Of the 4 flowing wells and springs reported, only two flows are given: 20 gpm from a spring in Wind River Canyon [Wyoming State Engineer's files] and 5 gpm from a 1,135-foot hole 3 miles south of town (Lowry et al., 1976) which was found clogged with rocks in January 1981. If the Sacajawea Well north of Thermopolis does indeed flow from the

Tensleep Sandstone, its flow of 1,002 gpm [Wyoming State Engineer's files] must be added. Cooley (1981) classes the Tensleep as one of the "major" aquifers of the southeastern Bighorn Basin (as compared with "minor" status for the Park City Formation), and State Geologist John Marzel (1929) saw the Tensleep Sandstone as such "an immense reservoir of water [that it] would require more than thousands of years to even appreciably diminish...even though this water were not replenished." Marzel concluded that the Tensleep Sandstone was the "obvious" source for all the hot springs and wells around Thermopolis, apparently on the basis only of his rhapsodic view of its water-bearing properties.

Flows for 17 identified Tensleep wells in the Tensleep, Wyoming area (see Figure 1) vary from 2 to 900 gpm and average 203 gpm (Lowry, 1962). Overall porosity values, which relate closely to permeability values in this case (Fox et al., 1975a), are 16-17 percent for the Tensleep area versus 14 percent for the Thermopolis area according to Fox et al. (1975b), but fracture-induced permeability is very likely greater in the structurally more complex Thermopolis area. Libra et al. (1981) cite oil-field-derived values of 10-14 percent for porosity, 0.8-99 millidarcies for permeability, and 10-300 gpd/feet for transmissivity for the Tensleep Formation in the Thermopolis study area.

For the twenty-six Tensleep Sandstone water analyses on file, [Geological Survey of Wyoming Open File Report No. 82-3], summary statistics are provided in Table 4. In the Bighorn Basin, Tensleep water is generally more dilute than Park City water, and "a definite and unmistakable trend towards higher concentrations and salinity basinward" from outcrop area is noted (Crawford, 1963).

The AMSDEN FORMATION has been little explored for either hydrocarbons or water in the Thermopolis area. Maughan (1972a) describes an upper sandy dolomite member and a lower sandstone member present only locally; both hydrologically and lithologically the distinguishing feature of the Amsden Formation is the shale member. Very low permeability in the absence of significant folding and fracturing is demonstrated in the Tensleep area by the wellhead pressure differences of 100 psi between the Tensleep Sandstone and the Madison Limestone reported by Cooley (1981). Burk (1952) reported hot water in the Amsden Formation at Rose Dome; 125 feet of water were recovered in a 30 minute drill stem test on the Owl Creek Anticline (Petroleum Information, -1981); and 10 gpm flow from a 3,469-foot Amsden(?) well in the Lake Creek Field [Wyoming State Engineer's files]. Oil is produced in limited quantities from the Amsden Formation at Black Mountain (Horn, 1963).

No water chemistry data from the Amsden Formation in the study area are available. Ten analyses from other parts of the Bighorn Basin indicate that the water-bearing zones of the Amsden Formation produce water similar to Madison Limestone water (Hinckley and Heasler, in preparation).

The MADISON LIMESTONE, probably the most famous aquifer in Wyoming, is noted for producing large quantities of water. The Madison in the Thermopolis area is described as "limestone and dolomitic limestone interstratified with dolomite" (Maughan, 1972a). Permeability is chiefly secondary, due to fracturing and to solution features described as "cavernous" (Lowry et al., 1976). Fracture-induced permeability is likely confined to deformed portions of the Madison.

Of the 6 Madison drill stem tests examined (Petroleum Information, -1981), all recovered significant quantities of water, 560-6,428 feet during flow periods averaging 80 minutes. Oil and water are produced from the Madison Limestone in the Hamilton Dome and Red Spring fields (Biggs and Espach, 1960), and Madison oil production occurs at Black Mountain (Horn, 1963). Water is found in the Madison Limestone at Wildhorse Butte, Kirby Creek, Lake Creek, Murphy Dome, Warm Springs, and Zimmerman Butte fields (Horn, 1963), Owl Creek Anticline (Lease and Palse, 1952), and Rose Dome and Cedar Ridge (Berry and Littleton, 1961). Burk (1952) describes a well encountering 155°F (68°C) water "rushing" into a cavernous zone in the Madison Limestone on Rose Dome.

No Madison springs or flowing wells have been located within the study area. Lowry (1962) reports 6 Madison wells in the Tensleep, Wyoming area, 3 flowing over 2,500 gpm and 3 flowing 84-380 gpm. This grouping of flows agrees with Lowry et al. (1976), who conclude that the Madison Limestone (considered together with the underlying Bighorn Dolomite) in the Bighorn Basin will yield either more than 1,000 gpm or less than 500 gpm to water wells, the bimodal distribution resulting from the irregular and cavernous nature of the aquifer.

Libra et al. (1981) provide one oil-field analysis for the Madison Limestone in the Thermopolis study area: porosity = 16 percent, permeability = 25 millidarcies, transmissivity = 70 gpd/foot. Cooley (1981) tested water wells northeast of the study area, and determined Madison transmissivities of 3,846 to 14,615 gpd/foot. Differences in transmissivity estimates are partially due to oil field calculations being based on only the petroleum pay thickness and on rock permeabilities to oil. Nonetheless, these estimates also reflect high variability of permeability with this aquifer.

Seven Madison Limestone water analyses, all from oil fields, are on file [Geological Survey of Wyoming Open File Report No. 82-3; also see Table 3]. Crawford's (1963) Bighorn Basin analysis finds that Madison waters tend to have more Ca and Mg than those from the Tensleep Sandstone, and generally lower total solids.

Cooley (1981) who worked northeast of the Thermopolis area, agrees with Lowry et al. (1976) in grouping the Madison Limestone and BIGHORN DOLOMITE as a single effective aquifer; very little information exists on the Bighorn Dolomite exclusively. The GALLATIN and GROS VENTRE FORMATIONS generally do not produce much water (Cooley, 1981), consistent with the relatively low permeabilities suggested by their shale and siltstone lithologies. Berry and Littleton (1961) report water of unknown quantities in these formations on the Owl Creek Anticline. We have only two water analyses from the preceding group of formations, a sample from the Bighorn Dolomite at Hamilton Dome and a sample from the Gros Ventre Formation at Red Spring [see Geological Survey of Wyoming Open File Report No. 82-3].

The lowest sedimentary unit above the Precambrian basement rocks is the FLATHEAD SANDSTONE, described as a "major" aquifer in the Tensleep area (Cooley, 1981). A report of Flathead water at Rose Dome (Berry and Littleton, 1961) is the only datum available for the Thermopolis area. In the Tensleep area, the Flathead Sandstone produces 500-800 gpm under artesian wellhead pressure up to 400 psi (Cooley, 1981). Both the two Flathead samples on file [Geological Survey of Wyoming Open File Report No. 82-3] and Crawford's (1963) conclusions indicate particularly high Na values for the Flathead Sandstone water. Otherwise, it is moderately dilute water with total solids averaging 343 mg/l.

The Precambrian rocks in this area are chiefly granitic and almost entirely dependent on weathering and fracturing for permeability (Berry and Littleton, 1961). They may be important in fractured portions of the Thermopolis Anticline. Lowry et al. (1976) estimate that waters from the Precambrian rocks likely contain less than 500 mg/l total solids.

WATER MOVEMENT

The general pattern of flow for the Thermopolis hydrothermal system (recharge in Owl Creek Mountains, discharge under artesian pressure at Thermopolis) is outlined above (p. 21). The reader will recall that flow directly perpendicular to the anticline at Thermopolis appears to be too shallow to produce the observed spring temperatures without extensive circulation into basement rock. The spring temperatures are readily explained, however, by flow patterns more complex both vertically and horizontally than the simple scheme of Figure 7.

Interformational Flow

Interformational (vertical) water movement can be examined on structural, hydraulic, thermal, and chemical grounds. In the Tensleep, Wyoming area, Cooley (1981) found the Paleozoic strata to be divisible into 3 distinct, major aquifers: Tensleep, Madison/Bighorn, and Flathead. Flow between these aquifers is small enough that well head pressure differences of up to 150 psi are common. However, Cooley repeatedly notes the importance of fracturing in greatly increasing interformation permeabilities; for example, he explains one area of abnormally high Tensleep pressures by upward movement of Flathead water due to fracturing around a small dome. The north flank of the Owl Creek Mountains is analogous to the Tensleep area; but once the sharp folding and faulting of the Thermopolis syncline/anticline are encountered, large head differences may well be equalized by interformational flow. Since stratigraphically lower units have higher recharge areas and hence higher heads, this flow should be predominantly upward, into shallower formations.

Big Spring's flow of 2,500 gpm from the Chugwater Formation clearly demonstrates interformational flow, at least for the spring site. Much wider support is provided by drill stem test data (Petroleum Information, -1981). Analysis of 100 tests of the Park City, Tensleep, and Madison Formations within the study area shows no systematic differences in hydraulic head between formations. Test data from two or more formations in the same hole commonly differ by less than 50 feet (22 psi). Although drill stem test data are insufficient to "prove" hydrologic intercommunication, the indication is clearly towards some degree of interformational flow.

Thermal evidence for vertical mixing is provided by bottom-hole temperature data and our own thermal logs. The isothermal character of the Paleozoic wells logged was discussed in an earlier section; of relevance here is that temperatures do not vary greatly for wells to differing depths and formations in the same area. On Rose dome, for example, temperatures of 145 to 161°F (63 to 71°C). 145°F (63°C) and 150°F (66°C) are reported for the Tensleep, Amsden, and Madison Formations respectively. On the northeast limb of the structure at Red Spring, temperatures of 85°F (29°C), 84°F (29°C), and 88°F (31°C) are reported for the Park City, Tensleep, and Madison Formations, respectively.

Chemical evidence for interformational water movement stems primarily from attempts to assign an aquifer to the existing hot springs and wells. Individual chemical analyses for the waters of the Paleozoic formations and of the Thermopolis hot springs and wells, are on file [Geological Survey of Wyoming Open File Report No. 82-3] with a discussion of data sources and reliability. Significantly, the chemistry of the various wells and springs differs very little. Thus, in spite of the fact that the wells and springs actually flow variously from the Park City, Tensleep, or Chugwater Formations, a single, common reservoir is indicated.

The average of 13 hot springs and well analyses (Breckenridge and Hinckley, 1978) are listed in Table 4 along with averages for the Park City, Tensleep, and Madison Formations. The hot springs and wells are very different from Park City or Tensleep averages for the area, but are quite similar to the average Madison values. Most values are within 10 percent for these two data sets and, importantly, the proportions of anions and cations are nearly identical. The coefficients of variation suggest that this "match" is in part a function of averaging and that Madison water analyses are much less consistent than are those for the springs. Nonetheless, of the 51 individual analyses available for Park City, Tensleep, and Madison water, those closest to hot springs and well analyses in both ion concentrations and proportions are from Madison Limestone water. Analyses beyond the major ions are not available for area Madison waters, so the possibility of discrepancies in the minor constituents has not been assessed.

The general chemical similarity of the hot springs and area Madison water and their substantial dissimilarity with area Park City and Tensleep water lead us to a Madison assignment for the spring waters, for the present. This is consistent with the hydrologic properties of the Madison Limestone and agrees with the conclusion of Berry and Littleton (1961) and Breckenridge and Hinckley (1978). Thus, upward movement of large volumes of water from at least as low as the Madison Limestone is indicated for the crest of the Thermopolis Anticline.

If waters rise from the Madison Limestone(?), their mixing with higher formation waters may cause chemical modification. (For example, the presence of H_2S gas, a typical Park City derivative, is common in the spring waters.) Given the formational chemistry variations evident, it is difficult to "prove" origin in a specific formation. In contrast, the homogeneity of hot well and spring chemistry allows an unambiguous statement of the water quality likely to be encountered in development of this system.

A major zone of fracturing along the Thermopolis Anticline is thus indicated by water movement. Such a zone of high permeability extending into basement rocks would provide for deep circulation, driven by free convection as cooler water descends and deep heated water rises within the fractured rock. Chemical comparisons indicate that a major contribution to the spring system of sub-Madison Formation water would generally shift composition away from that observed, towards higher Na, K, SO_4, Cl and TDS concentrations, and is therefore not indicated. Given the highly productive character of the Madison Limestone, however, Madison chemistry might dominate even if other aquifers had free access to the system. Circulation within the Precambrian basement rocks would be unlikely to alter water chemistry significantly.

Hydraulic Heads and Flow Volumes

Horizontal water movement can be evaluated through consideration of the distribution of hydraulic head, the levels to which water will rise in tightly cased wells. The surface represented by contouring these water levels is termed a potentiometric surface and, assuming all data are from the same strictly confined aquifer, predicts the direction of water movement much as surface topography controls surface flow. Beyond the qualitative evaluation of flow directions, these values also provide empirical data on the distribution of artesian pressure. Available data on flowing wells, static water levels in wells, and measured formation fluid pressures have been compiled for 113 wells in the study area. These are listed with location, formation, hydraulic head elevation, and datum source in Geological Survey of Wyoming Open

File Report No. 82-3.

Recharge areas for the Paleozoic rocks on the Owl Creek Mountains begin at around 4,600 feet in elevation for the Park City Formation and range up to extensive Madison Limestone outcrops at around 7,000 feet. These, then, are the maximum possible elevations for the potentiometric surface of these formations. Outcrop elevations in the Wind River Canyon (see Figure 4) begin at 4,360 feet for the Park City Formation, 4,365 feet for the Tensleep Sandstone, 4,390 feet for the Madison Limestone, and around 4,600 feet for the Flathead Sandstone. Springs issuing from the Park City (Breckenridge and Hinckley, 1978), Bighorn, and Gallatin Formations (Lease and Palse, 1952) and the Tensleep Sandstone* in the canyon identify it as an area of discharge rather than recharge (for these formations) and show that potentiometric surfaces are higher in areas away from the canyon.

Bredehoeft and Bennett (1972) provide a potentiometric surface map for the Tensleep Sandstone in the Bighorn Basin. From a 6,000-foot contour along the crest of the Owl Creek Mountains, the surface mapped slopes downward to a 4,400-foot contour running through the Hamilton Dome, Little Sand Draw, and Gebo oil fields, looping sharply back upstream to the mouth of the Wind River Canyon, then back out to pass north and west of the Zimmerman field and just north of Murphy Dome. The data compiled in Geological Survey of Wyoming Open File Report No. 82-3 indicate a locally depressed potentiometric surface around the hot springs and a much less severe depression of the surface along the river, but otherwise demonstrate the same general trends as the much sparser data of Bredehoeft and Bennett (1972): the springs of Hot Springs State Park (4,310-4,370 feet in elevation, the lowest natural surface discharge point for Paleozoic waters within the study area) occupy a large area of fairly similar hydraulic head. Hydraulic head elevations are higher west and south of the Thermopolis Anticline and east of the Red Spring - Wildhorse Butte Anticline, indicating flow into the Thermopolis and Red Spring areas from those directions.

Thus, waters will travel through the syncline southwest and west of Thermopolis and migrate along the anticline to discharge at the hot springs. The depth of the adjacent syncline predicts the observed general temperature increase along the anticline westward of Thermopolis (see discussion, page 18), as does the hydrologic indication of hot water influx from that direction. If hot waters are migrating along the anticline, temperatures should drop abruptly to the east of the present springs, reflecting only local syncline depth since there is no impetus for waters to move laterally beyond the springs. Temperature measurements for the east end of the anticline do show this relationship (see Figure 5).

Additional suggestions of flow parallel to the Thermopolis Anticline are the temperatures of a flowing Tensleep well (67°F, 20°C) and a Park City spring (72°F, 22°C), "Wind River Canyon Spring" of Breckenridge and Hinckley (1978), south of town. These occur, not on the Thermopolis Anticline, but on the northward dipping limb of the syncline where a simple model of flow perpendicular to the anticline, from the Owl Creek Mountains to Thermopolis, would predict only cool, descending flow. The relatively low elevation of these features, only 10 feet higher than Big Spring, however, requires that they be discharge points, evidently drawing water from deeper areas to the west-northwest.

East of Thermopolis, observed temperatures on the anticline agree with the predicted flow and thermal conditions, except for the 116°F (47°C) value from the Red Spring field. Hydraulic head data suggest that some water may also be moving up the east flank of the

*Wyoming State Engineer files.

Red Spring - Wildhorse Butte Anticline, but the thermal implications of such flow have not been determined. That the temperatures from several neighboring holes are consistently lower (and in agreement with temperatures predicted for flow from the south and southwest) suggests that the 116°F (47°C) report may be in error. Examination of the well log for this hole reveals no obvious reason for suspecting the value, however, so it remains problematic.

Calculated and measured hydraulic head elevations for the Park City Formation and the Tensleep Sandstone in the Thermopolis Anticline area are consistently around 4,400 feet. Hydraulic head elevations of 4,376, 4,406, 4,361 and 4,378 feet come from Rose Dome, 4,392 and 4,450 feet from Cedar Ridge; Big Spring flows at 4,370 feet, and the Red Spring area has heads of 4,470 and 4,366 feet. A north-south transect shows similar values: 4,380 feet for a Park City spring and a Tensleep well 4 and 3 miles, respectively, south of Thermopolis, and 4,340, 4,318, 4,400, 4,312, 4,361, and 4,340 feet along the Bighorn River 1.3, 1.7, 1.7, 1.8, 4.0, and 5.5 miles north of town. Thus, it appears as though flowing wells could be developed in many areas along the Bighorn River and that pumping lifts elsewhere should be less than the difference between surface elevation and 4,300 feet.

The last aspect of water flow to be considered is volume. The rate at which water will flow to a well bore is much harder to predict than either pressure or temperature. As explained in the aquifer descriptions, permeability is highly dependent on fracturing and, in the carbonate rocks, on solution features. The 500-1,000 gpm flows of the existing springs and wells of the Thermopolis system demonstrate the possibilities. Two Wyoming State Geologists (Barlett, 1925; Marzell, 1929) investigated the question of the 2,270 gpm flow from the hot wells decreasing the flow of the springs of Hot Springs State Park and both concluded that there had been no effect. Stearns et al. (1937) state simply that "large" artesian flows were obtained without "appreciably" affecting spring discharge. Flow data compiled by Breckenridge and Hinckley (1978) similarly suggest that the flow of Big Spring has not decreased significantly since 1909 (10 years before the first wells).

Van Norman (personnel communication, 1981) claims that there have been at least two more wells in the past than at present in the area north of Thermopolis producing hot water from the Park City Formation and Tensleep Sandstone. The Wyoming State Engineer's files includes one of these wells, listed as producing from "limestone" at a depth of 560 feet. No temperature is provided. Van Norman (personnel communication, 1981) reports that these wells slowly lost their flow over time and that the present Van Norman Well flows "much less" than when it was first drilled. She also has convincing photographic evidence that the Maytag Well produced considerably more water in 1928 than at present. These flows and flow differences had no reported effect on the natural hot springs. Possible explanations for such decrease in flow include constriction of well bores by mineral deposition and casing deterioration to the point of borehole collapse. Apparently no special provisions have been made to control either of these problems commonly associated with production of geothermal waters.

It should be noted that flow data for any component of the hydrothermal system are sorely lacking. The 8 flow measurements from which the Table 2 value for Big Spring is derived span 12 years and range from 2,212 to 2,908 gpm. Five measurements each for the Sacajawea, Maytag, and McCarthy Wells over the same period are 879-1,539 gpm, 498-1,027 gpm, and 224-745 gpm, respectively [Wyoming State Engineer's files]. The dates of the extreme measurements for the three wells do not coincide, nor do they occur at the same time of the year. Thus, within the resolution of such sparse data, we conclude that these variations are not the result of overall changes of the system nor of yearly cycles of flow. The data suggest, instead, complex variations in the system's water yield in both space and time. Although Bartlett (1925) concluded that the hot springs had not been affected by the wells, he was less sure of future wells, and became the first of many to suggest that systematic and frequent monitoring be practiced. No such program has been undertaken to date.

There has been more careful monitoring of the Paleozoic aquifers in the Tensleep area (see Figure 1). Development of the Tensleep, Madison/Bighorn, and Flathead aquifers in that area increased flow from an average of 1,900 gpm from wells in 8 townships in 1953 (Lowry, 1962) to 8,372 gpm, predominantly from Madison wells, in 1976 (Cooley, 1981). In 1962, Lowry concluded there had been no perceptible overall loss of pressure from these artesian systems; from 1978 data, Cooley concluded that though there had been no apparent pressure reduction in the Tensleep aquifer, the Madison/Bighorn had experienced a pressure decrease in "some" wells, and "most" Flathead wells no longer produced completion-magnitude pressures.

One certainly should not assume that there is a limitless supply of hot water at Thermopolis. At the same time, available evidence indicates that substantial quantities of water could be developed from the system without deleterious effects, particularly if reinjection of waste water is practiced. Given the importance of secondary permeability development in the aquifers of the system, water yields will likely vary from place to place. The present hot well flow of 500-1,000 gpm represents "safe" yields of the past. A well 20 miles north-northwest of Tensleep which flows 2,880 gpm from the Madison Limestone (Lowry et al., 1976) reflects the magnitude of production possibilities, though the effect of such production on the hot springs cannot be predicted at this time.

SUMMARY, IMPLICATIONS, AND RECOMMENDATIONS

Geologic and hydrologic conditions in the Thermopolis study area indicate water movement northeastward off the flank of the Owl Creek Mountains, through the intervening syncline, and up the steep north flank of the Thermopolis Anticline. Largely confined by the less permeable beds of the overlying Chugwater Formation, water is under artesian pressure in Paleozoic aquifers. Extensive fracturing along the sharply folded anticline, and a probable basement fault beneath it, allow upward flow and subsequent discharge at the existing hot springs. Chemical analyses, supported by observed high discharge volumes, suggest that spring water is predominantly of Madison Limestone origin.

Thermal modeling predicts Madison temperatures of 160°F (71°C) in the syncline opposite Rose Dome and 104°F (40°C) in the immediate Thermopolis area. Water migrating southeast from the Rose Dome area to discharge at the springs should elevate spring temperatures above those in the adjacent syncline, whereas areas further east should not receive this heating component. Observed temperatures agree extremely well with this model of water flow and temperature: 161°F (71°C) was measured on Rose Dome, the temperatures of the hot springs are near 130°F (54°C), and maximum temperatures drop abruptly to around 100°F (38°C) along the eastern end of the anticline. Water from formations below the Madison Limestone and deep convective circulation in a fault zone may contribute higher temperatures to the system, but it appears unlikely that significant volumes of water can be developed at temperatures exceeding 170°F (77°C) at Rose Dome, 150°F (66°C) at Cedar Ridge, and 140°F (60°C) in the vicinity of Thermopolis townsite.

These estimates result from measured and reported temperature and gradient values, thermal modeling, and consideration of temperature loss as hot waters rise to the surface. To evaluate this last point, the flow of Big Spring was modeled as arriving through a (1-meter) 3.3-foot diameter conduit (following Truesdell et al., 1977) extending either to the base of the Madison or penetrating 1,300 feet of Precambrian rock. Using a standard rock diffusivity value of 32 km^2/million-years, both models indicate less than 9°F (5°C) temperature drops even in the extreme case of reservoir temperatures as high as 194°F (90°C). Logs of wells flowing from the Thermopolis system support this conclusion empirically: measured temperature losses for the Maytag and McCarthy wells were only 0.26°F (0.14°C) and 0.34°F (0.10°C) per 1,000 feet, respectively, in well bores less than 1 foor in diameter (see Figure 6, pages 12-16).

The marginal resource identified in the Red Spring area has not been thermally modeled. Flow into the area from the south and southwest should show temperatures similar to those projected for the east Thermopolis Anticline; the thermal implications of flow from the east have not been studied. The next step in the evaluation of this area should be verification of reported temperatures in excess of 100°F (38°C).

An important implication of the model developed so far is that once hot water is encountered, deeper drilling is not likely to result in significantly higher temperatures. Movement between formations, at least from the Park City Formation to the Madison Limestone, appears to be sufficient to homogenize temperatures, producing isothermal conditions throughout this section along the Thermopolis Anticline. It is likely that aquifer water yield increases from the Park City Formation to the Tensleep Sandstone and possibly from the Tensleep Sandstone to the Madison Limestone, so deeper drilling may result in greater flow; but we feel that the temperatures presented above are the maximums likely to be encountered at feasible drilling depths.

The importance of fracture-induced permeability in the upper Paleozoic aquifers generating great water yields has been emphasized repeatedly above. Such zones of fracture and faulting occur along the crest of the Thermopolis Anticline and perhaps in areas perpendicular to the anticline at dome boundaries. Detailed mapping in the area is necessary to precisely delineate such zones. Lowry (1962) advises that low-yield wells into these upper Paleozoic aquifers may be significantly improved by well-stimulation techniques aimed at increasing permeability.

Existing wells show that yields in excess of 500 gpm can be developed from high permeability zones. Details of hydrologic characteristics of the Thermopolis area aquifers are largely unknown. We have located no pump test determinations for the area, nor even detailed records of well flows. Given the number of wells which have been drilled into this hydrothermal reservoir, we feel that a carefully implemented program of well testing and monitoring would be very useful. Hydraulic head data indicate that thermal waters once encountered will rise to an elevation of 4,320-4,380 feet or flow at the surface, whichever comes first.

Although high water temperatures may be found in Paleozoic rocks over a large area north of the anticline, the northern boundary of the viable resource area is fixed by drilling depths. At the prevailing dip of around 9°, a given stratum is 836 feet deeper every mile north-northeast from the crest of the anticline. While the Park City Formation is 458 feet below the surface at the McCarthy wells (Bartlett, 1925), it should be 1,015 feet deep one-half mile north-northeast. Part of the depth in this case is due to increased surface elevation. Thus, it is necessary to integrate surface elevation, depth to aquifer, and hydraulic head data, as well as to try to intersect a zone of high permeability, in actually siting a well.

An approximation of the depth to a given formation can be obtained by determining the surface formation (Figure 3) and summing the thicknesses of the intervening formations (Table 3, page 23). Depths will be greater than the simple, summed thicknesses as dip increases, but will be less than 2 percent in error for dips less than 10°. An additional caution on depth is that the formations into which the rocks are divided may be no more uniform vertically than they are horizontally; i.e., it may be necessary to drill well into a formation to realize significant production. For example, while the McCarthy well flows nearly 1,000 gpm from the upper 10 feet of the Park City Formation, a well just west of town was drilled 188 feet into the Park City Formation before producing water, which then rose to a depth of 55 feet in the well [Wyoming State Engineer's files].

On the south flank of the Thermopolis Anticline, dips are very steep. The thermal necessity to stay north of the syncline to intercept the hottest flow confines exploration to a very narrow, geologically complex strip just off the crest (see Figure 6). The scale and detail of geologic investigations needed to identify potential development sites in this area are beyond the scope of this report. Such investigation should certainly precede any development planning.

One engineering and environmental problem that may appear is the handling of large volumes of mineralized water. The travertine terraces and tipis of Hot Springs State Park testify to the depositional possibilities of the waters. Norman Sanford, (personal communication, 1979) reports such rapid travertine deposition that a pump in well C3 (Rose Dome) was rendered inoperable in only 3 years. During that same period, approximately one inch of travertine had built up on the well-fed stock tank. Cessation of flow from some wells north of town (see page 33), and the declining flow of the Maytag well, may also be due to mineral deposition. On the other hand, Big Spring shows no sign of declining flow, and, while the Van Norman well flow has decreased over time, their house has been geothermally heated for over 40 years (Van Norman, personal communication, 1981) without excessive mineralization problems. The Taylor house is similarly heated by the waters of the McCarthy well, and, though their system is of more recent vintage than the Van Norman system, it has experienced no problems to date (Scott Taylor, personal communication, 1981). Mineral deposition is likely a result of changes in temperature and pressure. Given the fairly constant chemistry of the Thermopolis hydrothermal waters, it should be possible to calculate the magnitude of the potential mineral problem as a function of how the waters are to be managed (see Anderson and Lund, 1979).

The major legal obstacle to development of the Thermopolis resource appears to be a possible conflict with the flow of the springs in Hot Springs State Park. Water rights within the State Park are controlled by the Wyoming State Board of Charities and Reform [Wyoming Statutes 1977, section 36-8-305], and the Wyoming State Enginner is specifically charged with the protection of thermal springs on State Lands [Wyoming Statutes 1977, section 41-1-109]. The State Engineer's authority extends to any drilling, private or public, in the Thermopolis area. In our discussion of available flow volumes on pages 33-34, we conclude that it is possible that significant nonconflicting development could occur. Certainly, any such development should be undertaken with caution, within the framework of a program of careful monitoring of existing wells and springs, and with every consideration given to minimizing the possibility of conflict. The aesthetic, recreational, and therapeutic value of Hot Springs State Park should not be underestimated, nor should unfounded concern over the flow of the springs preclude responsible exploration and development of this potentially valuable resource.

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Sources for Figure 3:

For thickness and physical description:

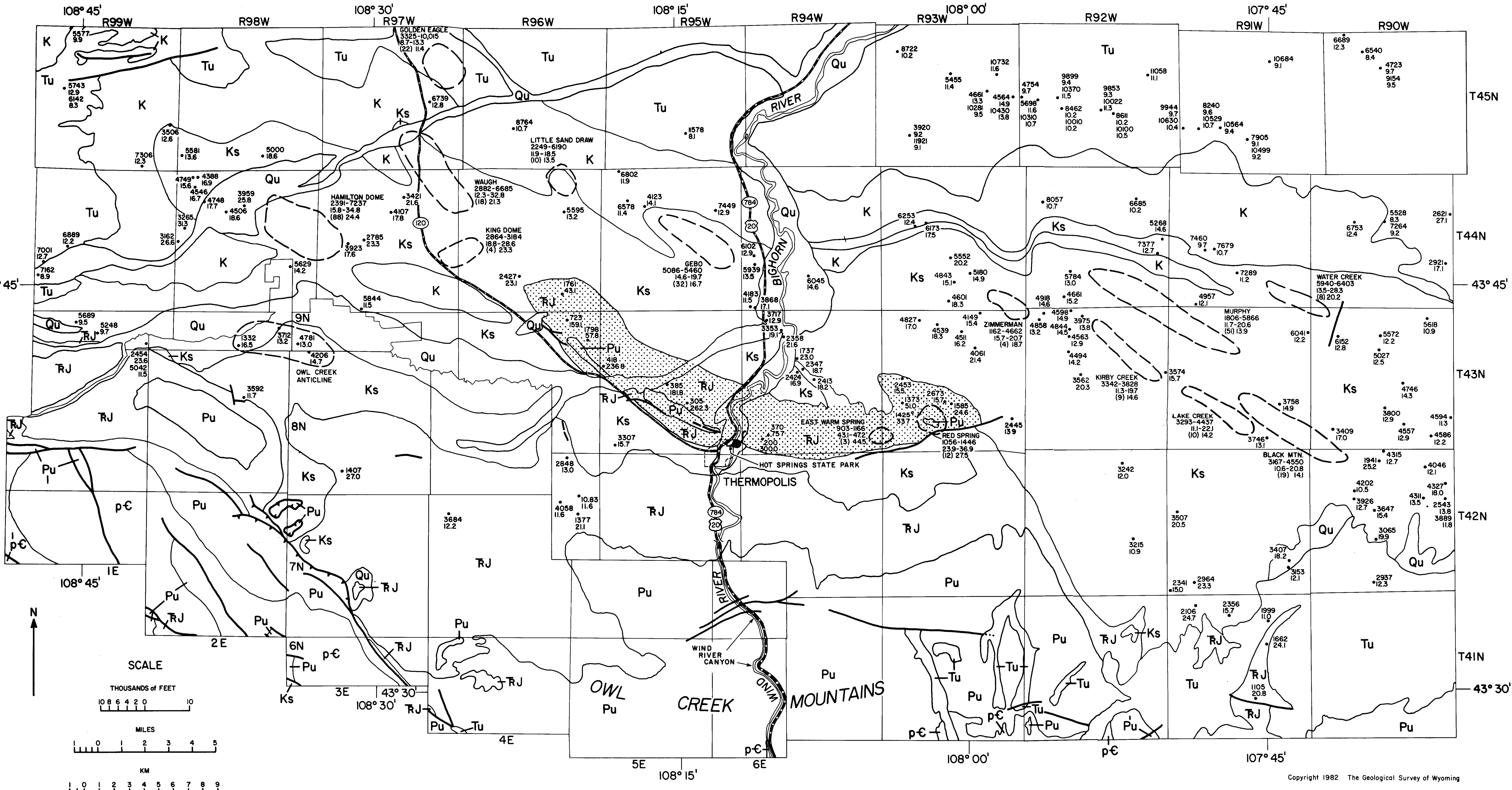
Thicknesses above the Cody from Pierce (1978); lithologies and sub-Mesa Verde thicknesses compiled from Deiss (1951), Fanshawe (1939), Jones (1939), Ary (1959), Shlemon (1959), Berg (1976), Berry and Littleton (1961), Pierce (1978), Horn (1963), Collier (1920), Tourtelot and Thompson (1948), Maughan (1972a, 1972b), and examination of area oil and gas well logs. See Table 3, page 23, for average thicknesses adjacent to the Thermopolis Anticline.

For water-bearing characteristics:

Hydrologic properties from Berry and Littleton (1961) and Lowry et al. (1976).

THE GEOLOGICAL SURVEY OF WYOMING
Gary B. Glass, State Geologist

GEOLOGIC AND THERMAL DATA FOR THERMOPOLIS AREA



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EXPLANATION

Qu	QUATERNARY DEPOSITS UNDIVIDED
Tu	TERTIARY ROCKS UNDIVIDED
K	LANCE, MEETEETSE AND MESAVERDE FMS.
Ks	CODY SHALE, FRONTIER FM., MOWRY AND THERMOPOLIS SHALES
ŦJ	CLOVERLY, MORRISON, SUNDANCE, GYPSUM SPRING, CHUGWATER AND DINWOODY FMS.
Pu	PALEOZOIC ROCKS UNDIVIDED
p€	PRECAMBRIAN ROCKS UNDIVIDED

POTENTIAL GEOTHERMAL RESOURCE AREA

CONTACT

FAULT

THRUST FAULT

HOT SPRINGS

WELL LOCATION
XXXX DEPTH (FEET)
XX.X GRADIENT (°F/1000 FT.)

OIL AND GAS FIELD LOCATION
FIELD NAME
XXXX-XXXX DEPTH RANGE (FEET)
XX.X-XX.X GRADIENT RANGE (°F/1000 FT.)
NO. OF DATA POINTS (X) XX.X AVERAGE GRADIENT (°F/1000 FT.)

COMPILED BY J.K. KING, 1980
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