ANALYSIS AND INTERPRETATION OF DATA OBTAINED IN TESTS OF THE GEOTHERMAL AQUIFER AT KLAMATH FALLS, OREGON

E. A. Sammel, Editor

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CONVERSION OF UNITS OF MEASUREMENT

INCH-POUND TO METRIC

Multiply inch-pound units	by	To obtain SI units
	Length	
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.589	square kilometer (km^2)
	Volume	
cubic foot (ft^3)	0.02832	cubic meter (m^3)
	Flow rate	
gallon per minute (gal/min)	0.06308	liter per second (L/s)
	Pressure	
pound per square inch (lb/in^2)	6.895	kilopascal (kPa)
	Temperature	
degree Fahreheit (°F)	°C = 5/9 (°F - 32)	degree Celsius (°C)
	Thermal energy	
British thermal unit (BTU)	1,055	Joule (J)
	Viscosity (absolute)	
centipoise	0.001	Pascal second (Pa/s)
	Mass	
pound, avoirdupois (lb)	453.6	gram (g)

	Heat flow	
One heat-flow unit (hfu) =	4.184×10^{-2}	Watts per square meter (W/m^2)

CONVERSION OF UNITS OF MEASUREMENT (Continued)

METRIC TO INCH-POUND

Multiply SI units	by	To obtain inch-pound units
	Length	
millimeter (mm)	0.0394	inch (in.)
meter (m)	3.281	foot (ft)
	Mass	
gram (g)	0.002205	pound, avoirdupois (lb)
kilogram	2.205	pound, avoirdupois (lb)
	Thermal energy	
kiloJoule (kJ)	239	calory (cal)
	Temperature	
degree Celsius (°C)	°F = 9/5 °C + 32	degree Fahrenheit (°F)
	Permeability	
darcy (k)	2.43	foot per day (ft/d) at 60°F

National Geodetic Vertical Datum of 1929 (NGVD of 1929): A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called "Mean Sea Level." NGVD of 1929 is refered to as "sea level" in this report.

ANALYSIS AND INTERPRETATION OF DATA OBTAINED IN TESTS OF THE GEOTHERMAL AQUIFER AT KLAMATH FALLS, OREGON

E. A. Sammel, Editor

ABSTRACT

A study of the geothermal resource at Klamath Falls, Oregon, has shown that thermal water occurs in an extensive, heterogeneous aquifer at depths of a few hundred to nearly 2,000 feet over an area of nearly 2 square miles. The highest measured water temperatures are more than 130°C. Chemical and isotopic analyses suggest that the aquifer is supplied from a deeper zone in which meteoric recharge water having low chloride and silica concentrations mixes with high-temperature water (about 190°C) having a moderately high chloride concentration (120 miligrams/kilogram). The probable temperature of the hot-reservoir water is estimated on the basis of consistent results from the sulfate-water isotope and silica geothermometers and calculated mixing fractions of 40 and 44 percent thermal water derived from chloride and silica mixing models.

The thermal water is supplied to the shallow aquifer through a permeable fault zone on the northeast border of the City. The water spreads southwestward in the aquifer, losing heat largely by conduction and convective discharge to more than 450 wells that tap the aquifer for space heating in homes and businesses.

A 21-day pumping test, performed in July 1983, resulted in significant drawdowns in wells over most of the principal geothermal area. Analysis of data from 52 observation wells indicated that drawdowns in most wells fit a theoretical model based on two contrasting ranges of permeability and porosity in the aquifer rocks (double-porosity model). There were indications that transmissivity (permeability times thickness) is greater along the NW structural trend of the area than transversely.

In a second phase of the aquifer test, water was reinjected for 29 days into a well located more than 3,000 ft from the pumped well. The water-level rise that followed confirmed the results of the drawdown test and gave a further indication of the excellent hydraulic connection of all parts of the aquifer.

Tracer tests carried out in a closely-spaced pair of wells prior to the aquifer test and in the pumped and injection wells during the test also confirmed the double-porosity behavior of the aquifer. Results of the tracer

tests will permit future evaluation of the danger of thermal breakthrough and a decrease in aquifer temperatures that may result from reinjection.

Current discharge of thermal water from about 70 pumped and artesian wells averages about 540 gallons per minute. The rates of use show excellent correlations with daily air temperatures. The aquifer drawdowns that result from this use also correlate well with prevailing air temperatures. A predictive algorithm was developed for the relation between utilization and air temperature. The amount of heat withdrawn from the aquifer by this means is about 18×10^{12} British Thermal Units per year. In contrast, the more than 380 wells that utilize down-hole heat exchangers discharge only about 13×10^{10} British Thermal Units per year.

Additional withdrawals of heat and water would result in further declines in water levels during the heating season. Reinjection could offset water-level declines over most of the aquifer, but declines would occur in some wells, with possible adverse consequences for existing heating systems. As examples of expected aquifer behavior, predictions of the distribution and magnitude of short-term water-level changes were made for two hypothetical utilization schemes. The computer-generated analysis can be applied to any proposed combination of pumping and reinjection.

Long-term changes in water levels depend on factors that could not be determined during this study, such as aquifer boundary conditions, recharge rates, and future patterns and types of withdrawals of heat and water. Thus, although most of our data indicated that the geothermal resource at Klamath Falls could sustain additional development, the limits of such development were not estimated.

The pumping and injection test was conducted by Lawrence Berkeley Laboratory, the tracer tests by Stanford University, and thermal, utilization, and pumped-discharge data were collected by the Oregon Institute of Technology, all with financial support from the U.S. Department of Energy. Chemical studies and overall coordination of the activities were provided by the U.S. Geological Survey.

CHAPTER 1. INTRODUCTION

By

E. A. Sammel

U.S. Geological Survey, Menlo Park, California

Objectives and Scope of the Study

During the summer of 1983, investigators from several institutions collaborated in an intensive study of the geothermal resource at Klamath Falls. Funded largely by grants from the U.S. Department of Energy (DOE), scientists from Lawrence Berkeley Laboratory (LBL), Stanford University, and the Oregon Institute of Technology (OIT) were co-investigators under the terms of a proposal submitted to DOE by the U.S. Geological Survey (USGS). Participation by USGS personnel was funded by the USGS Geothermal Research Program.

The work included tracer studies by Stanford University, a pumping and injection test by LBL, temperature studies and collection of aquifer-discharge and use data by OIT, and sampling for chemical analysis by USGS.

The principal objective of the investigation, as stated in the proposal to DOE, was to acquire "from the shallow geothermal reservoir at Klamath Falls... chemical and hydraulic data on which to base predictions of reservoir performance and an evaluation of potential for development." The major purpose "is to provide interested parties in Klamath Falls with scientific data to be used to evaluate alternatives for the future of the geothermal resource; a second purpose is to assess potential impacts of possible alternatives." It is also expected that "knowledge gained in the investigation can be used to aid in the evaluation of other fault-controlled geothermal systems." Clearly, it is not the purpose of this study to recommend specific courses of action regarding the development of the geothermal resource at Klamath Falls, but rather to provide the scientific data that will be required for decision-making by agencies and citizens in Klamath Falls.

Data Reports and Files

Data collected during the summer of 1983 specifically for the aquifer test are presented in graphical and tabular form in the USGS Open-File

Report 84-146 (Benson and others, 1984). The report can be consulted in the public library at Klamath Falls and may be obtained from the USGS Open-File Services Section, Western Distribution Branch, Box 25425, Federal Center, Denver, Colorado 80225.

Reports on tracer tests conducted by Stanford University prior to and during the aquifer test will be published separately by the University when interpretations have been completed. A summary of preliminary findings is included in the present report. Sampling for chemical and isotopic analyses was done under the direction of A. E. Truesdell (USGS). The data are included in the Open-File Report 84-146, and an interpretation of the results is included in the present report. Preliminary reports on the 1983 aquifer test (Benson, 1983) and the tracer tests (Gudmundsson, Johnson, and Horne, 1983) were presented at Stanford University in December 1983.

All data collected during the investigation resides in a central data file in Klamath Falls. It is the intent of local agencies and citizens that the file be kept current as new data become available on current responses in the aquifer and the history of the geothermal development. Both kinds of data will be required in order to refine estimates of the potential of the resource made in this report.

Geothermal Background and Geologic Setting

Klamath Falls is a city of about 17,000 persons situated about 40 miles east of the Cascade Range in south-central Oregon (fig. 1-1). It is one of a growing number of American communities that rely on a geothermal resource for at least a part of their energy needs. In at least one respect, however, Klamath Falls is unique in the United States and possibly the world. More than 450 wells tap the geothermal aquifer in an urban setting of streets and homes less than two square miles in area. The wells, some of which have been used for more than 50 years, supply heat for a variety of direct-use applications, the predominant one being domestic space heating.

History of Geothermal Use

The presence of thermal water at Klamath Falls was first evident in boiling springs. Accounts by 19th Century settlers describe the use of these springs by local Indians and the subsequent development of the spring areas by settlers (Lund, 1978). The five hot-spring locations fairly well define the area in which thermal water is now obtained from artesian wells (fig. 1-2).

The springs no longer flow, and the reasons for their demise are not completely known. Although the production of thermal water from wells has been regarded as a probable cause, it was reported by local inhabitants as early as 1939 that the springs had begun to dry up prior to the drilling of the first wells (Lund, 1978).

Following the discovery, in the early 1920's, that shallow wells in the hot-spring areas could produce natural flows of near-boiling water (Storey, 1974), the use of private wells for space heating began a growth that has continued to the present time. Innovations in well design developed by Charles Leib, a local geothermal pioneer, led to the first use of a down-hole heat exchanger in 1931, the use of casing perforations to induce a thermosiphon effect for more efficient heating in 1945, and the use of paraffin to reduce corrosion at the air-water interface (Fornes, 1981). Charles Leib's inquiring mind and understanding of physical processes led, at an early date, to the first comprehension of the dynamics of the Klamath Falls geothermal aquifer, an understanding that has not been significantly improved upon until recent years.

Geologic Structure and Stratigraphy

Rocks of the Klamath Falls area are predominantly of volcanic origin. Known formations include massive basaltic and andesitic flow rocks and tuff, volcanic sediment, diatomite, and interbedded lava flows of Pliocene age, and pyroclastic rocks, breccia, lava, and alluvium of Pleistocene age. Variable thicknesses of Holocene alluvium overlie older rocks at lower elevations (Newcomb and Hart, 1958; Peterson and Groh, 1967).

The area east of Klamath Falls is morphologically similar to Basin and Range terrain. The region is, in fact, transitional between the High Cascade volcanic chain to the west and the Basin and Range Province to the southeast.

Most of the more than 450 thermal wells at Klamath Falls penetrate volcanic sediment and diatomite that were deposited in a Pliocene lake. The deposits are predominantly fine grained and have low permeability. The thickness may be several hundred feet or less on the upper slopes of the hillside northeast of the city; it increases to more than 1,000 feet in the area west of the "A" Canal beneath the city center (fig. 1-2). Interstratified with the sediment layers are thin (5 to 20 ft) strata of basaltic tuff, scoria, and breccia. A study of 175 drillers' logs revealed a bewildering complexity of reported rock types that could not be correlated over distances as short as 100 ft. Much of this heterogeneity is attributable to varying descriptions of the same rocks by different drillers, but much of it is real.

Beneath the predominantly sedimentary rocks are basaltic and andesitic flow rocks and tephra in strata whose prevalence and thickness increase with depth. So far as can be determined from drillers' logs, the change from a dominantly volcanic to a dominantly sedimentary regime was a gradual one which occurred over a long period of Pliocene time. However, conclusions of this kind must be tentative because of the difficulty in distinguishing pyroclastic rocks from sedimentary rocks in typical well-drillers' logs, which are the principal source of data on subsurface conditions.

On the basis of field study and examination of aerial photographs, a major NW-trending fault has been traced from the shore of upper Klamath Lake, through the northeast corner of the Oregon Institute of Technology (OIT) campus, and thence southeastward in a trend roughly parallel to the border of the hot-well area of the city (fig. 1-2). Segments of this structural trend are obscured by erosion, and it is possible to speculate that branching or parallel faults exist at several places south of the OIT campus. Two NE-trending faults may occur, one between OIT and the hot-well area and another along Old Fort Road. In the area south of Main Street, the trace of the principal fault disappears at the edge of the lower Klamath Lake basin.

The principal fault is one of several westward-dipping, high-angle normal faults that define the eastern margin of the Klamath graben (a structural valley bounded by normal faults). The Klamath graben contains the upper and lower Klamath Lake basins. In parts of the graben, geophysical studies have provided estimates of the depth of the graben fill and the locations of the boundary faults (see Stark, Goldstein, and Wollenberg, 1980; and Sammel, 1980, for details and additional references), but no such studies have been made in the hot-well area.

Regional Heat Flow and Geothermal Gradients

The amount of heat discharged at the earth's surface differs from place to place depending on the type of heat source, depth of the source, nature of the overlying rocks, regional topography, and the movement of ground water. Average heat flow from the earth's continental crust is about 1 1/2 heat-flow units (HFU). Anomalously warm regions, such as the northern Basin and Range Province in Nevada, may transmit 2 to 3 HFU. Heat flows of 2 or more generally imply that prospects are favorable for the occurrence of geothermal resources at shallow depths. Some investigators have concluded that average heat flow at Klamath Falls is between 2.0 and 2.8 HFU (Blackwell and others, 1978).

Heat flow in the urban hot-well area of Klamath Falls is extremely high. A calculation based on the known and estimated discharge of thermal water from the 2-square-mile area suggest that natural and induced thermal discharge would represent at least 95 HFU. But this type of convective heat flow must be carefully distinguished from the conductive transfer of heat in the absence of moving water. True conductive heat flow generally represents deep crustal conditions over large areas, whereas convective heat flow is more likely to represent local structural, stratigraphic, and hydrologic conditions in the upper few hundred to few thousand feet of rock.

In the Klamath Falls area, virtually no test holes or wells can be assumed to be free of convective effects. In addition to convective flow within a well, convection may occur in the annular space outside the casing, thereby affecting temperatures measured in the well. Nevertheless, many profiles at Klamath Falls show stable positive gradients (temperature increases with depth) or stable reversals of gradient that appear to reflect temperatures in the formations.

Temperature profiles from more than 40 wells in the region indicate that temperature gradients may be extremely high to depths of 600 ft or more, but the profiles in three of the deeper wells (depths 1,325 to 1,805 ft) suggest that if measurements in shallow wells could be continued to sufficient depths, all would show temperature reversals or quasiisothermal profiles indicative of hydrothermal convection (Sammel, 1980). Temperature profiles in 11 wells in the hot-well area are derived from bottom-hole temperatures reported by drillers (fig. 1-3). These profiles probably reflect temperatures in the aquifer rocks more closely than typical profiles measured in the water columns of wells. They indicate that lateral flows of thermal water in permeable strata provide heat that is conducted vertically through adjacent less permeable strata. Thus, temperature profiles from wells in the hot-well area must be regarded as unreliable indicators of deep conductive heat flow, and consequently, no conclusions can be drawn from these profiles regarding the heat source and the nature of thermal activity at greater depths. One conductive gradient measured to a depth of 575 ft in sediments of the lower Klamath Lake basin probably represents a temperature boundary at the base of the sediments, but suggests that a lower limit for heat flow in the region is about 1 1/2 hfu (Sass and Sammel, 1976). The actual value for the shallow crustal heat flow in the vicinity of Klamath Falls undoubtedly is higher than 1 1/2 hfu.

The Geothermal Aquifer

The geothermal aquifer at Klamath Falls is not a clearly defined rock unit, but rather a stratified series of lithologic units having large vertical and areal variability. Evidence from lithologic logs and temperature measurements indicates that, within the total thickness of rocks that comprise the aquifer, water flows preferentially in strata that may have thicknesses of a foot or less and generally are not more than a few feet thick. The drillers' reports indicate that these permeable strata include fractured, indurated lacustrine sediment as well as volcanic breccia and fractured vesicular basalt flows.

Diagrams in figure 1-3 represent the author's interpretation of drillers' logs in 11 wells. Locations of these wells are shown in figure 1-2. Well 450 is not in the area of intensive study for this report, but is included as an example of the increasing thickness of sediment penetrated by wells on the southern and western margins of the area. The diagrams indicate the lithologic variability that is typical of the 175 or more logs studied for this report.

The temperature profiles in figure 1-3 show that hot water (>80°C) occurs less than 200 ft below land surface under most of the hot-well area, although some wells are drilled to depths of 700 ft or more in order to insure an adaquate supply of water. In wells located on the topographic high east of the principal fault line, depths of wells are commonly greater than 1,000 ft; immediately west of the fault, most wells are less than 500 ft deep. The permeability of the rocks at a particular site probably controls the depth of most wells, and thus, the apparent thickness of the aquifer varies from place to place. The true thickness of rocks that comprise the thermal aquifer is not known but is at least 2,000 ft on the basis of the temperature profile in an 1,805-ft well at OIT (Sammel, 1980). This temperature profile also suggests that no well at Klamath Falls has penetrated the full thickness of the aquifer.

On the basis of data from wells >300 ft in depth, figure 1-4 shows that the potentiometric surface of water in the thermal aquifer dips toward the southwest in general conformity to the slope of the land surface. Eastward from the fault area, water levels also appear to decrease in altitude over a distance of at least 1,600 ft from the fault. However, accurate data are available only for wells along Old Fort Road (fig. 1-2), and water levels are not reliably known in the area east of the fault and north of Old Fort Road.

All available data suggest that the most probable source of thermal water in the hot-well area is the principal fault zone, from which the water spreads laterally toward the south and west. Reported maximum temperatures are highest in the vicinity of the fault (95° to 120°C), and they decrease toward the southwest to less than 80°C in a distance of about 3,000 ft (fig. 1-5). The artesian area, shown in figure 1-2, extends across the trend of the Old Fort Road valley, and this occurrence suggests the possibility that artesian pressures and high temperatures are transmitted in a permeable NE-trending fault zone that cuts across the main fault.

Strata containing the hottest water occur at altitudes of about 4,000 ft near the fault, and the altitudes decrease to about 3,800 ft in the artesian area (fig. 1-3). Near-surface aquifer temperatures are as low as 50°C at altitudes of 4,300 ft in the vicinity of the faults (fig. 1-3). In one well located more than a mile from the fault, the altitude of maximum thermal-water entry is 3,350 ft and the water temperature is 30°C (well 450, fig. 1-3). However, pressures measured in the artesian well during the winter of 1983-84 indicate that the well probably does not tap the aquifer supplying the hot wells.

Recent History and Proposals for Development and Testing

Development of the geothermal resource at Klamath Falls during the first 50 years or more of its history has been characterized by a quintessentially American pioneer individualism. Use of the thermal water has occurred through personal initiative; conseqently, the great majority of the thermal wells supply heat for only one home or business. A proposal to develop geothermal heat for a large number of buildings in the central city gathered momentum only in the late 1970's and, supported by funds from the U.S. Department of Energy (DOE), proceeded to a construction phase in mid-1979. By early 1982, a district heating system comprising two production wells, an injection well, a heat-exchanger facility, pipelines, and heating units was in place and had been briefly tested.

Prior to construction of the system, proposals relating to resource definition and aquifer testing had been made and discussed at DOE-sponsored conferences. However, no large-scale test of the producing aquifer and the injection process had been conducted. Meanwhile, concerns expressed by well owners regarding possible adverse consequences of operating the district heating system had led, in June 1981, to passage of a City Ordinance that placed restrictions on further development. This ordinance had the effect of halting implementation of the District Heating Plan as originally conceived.

In January 1983, a new effort to gather data on the geothermal resource was initiated by the Klamath County Chamber of Commerce. By March 1983, a program of data gathering and aquifer monitoring was underway, carried on largely by volunteer efforts but with financial support from the Chamber of Commerce, the Klamath County Economic Development Association, and the City of Klamath Falls. Much of the volunteer work was organized by a well-owners group, Citizens for Responsible Geothermal Development (CRGD).

The initial objective of the program was simply to collect and organize existing knowledge of the resource and to monitor water levels and temperatures in the geothermal aquifer, but this program led directly to and largely made possible the aquifer-testing study that is the subject of this report.

Acknowledgments

The activities of this study originally were made possible and were later supported by the concern and active efforts of a large number of citizens of Klamath Falls. The result of these efforts has been an unprecedented community involvement and a unique achievement in the field of geothermal investigations. One example of the cooperation between scientists and non-scientists that characterized the study was the willingness of an estimated 80 percent of well owners to turn off their geothermal heating systems for 24 hours as an early test of the aquifer response. Many citzens participated in the monitoring program and contributed data to the central file. The community-wide effort represented by the work of these hundreds of volunteers is gratefully acknowledged.

The local and regional news media contributed greatly to the program through timely, accurate, and thorough coverage of plans and events. The educational aspect of the investigation was also aided by the participation of many citizens in forums and other informational meetings. As the result of the dialogue between visiting scientists and the local public, both groups became better informed and the scientific outcomes were greatly enhanced.

Finally, we acknowledge with thanks the support provided by DOE under contracts with OIT, LBL, and Stanford University. Without support of this magnitude, the scientific observations made during this study probably would not have been possible.

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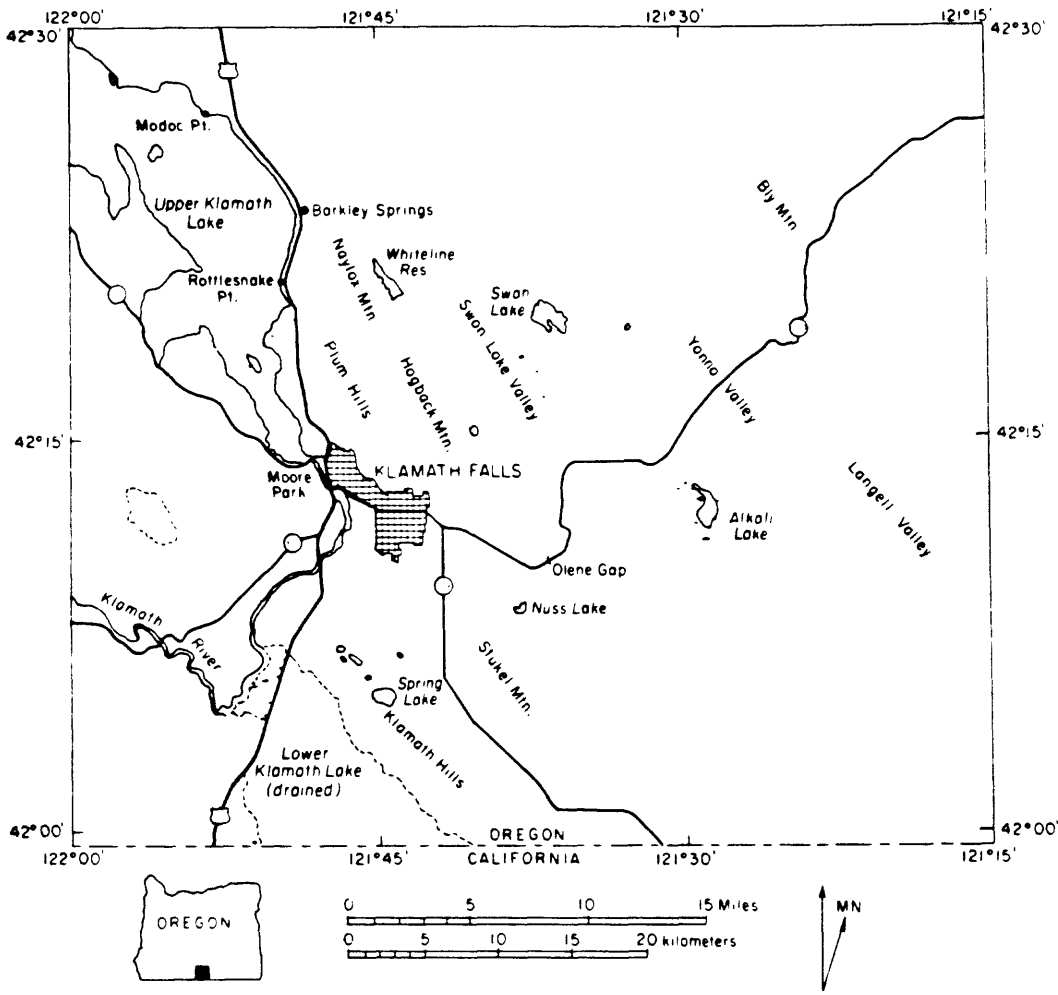

Figure 1-1. — Index map of the Klamath Falls area.

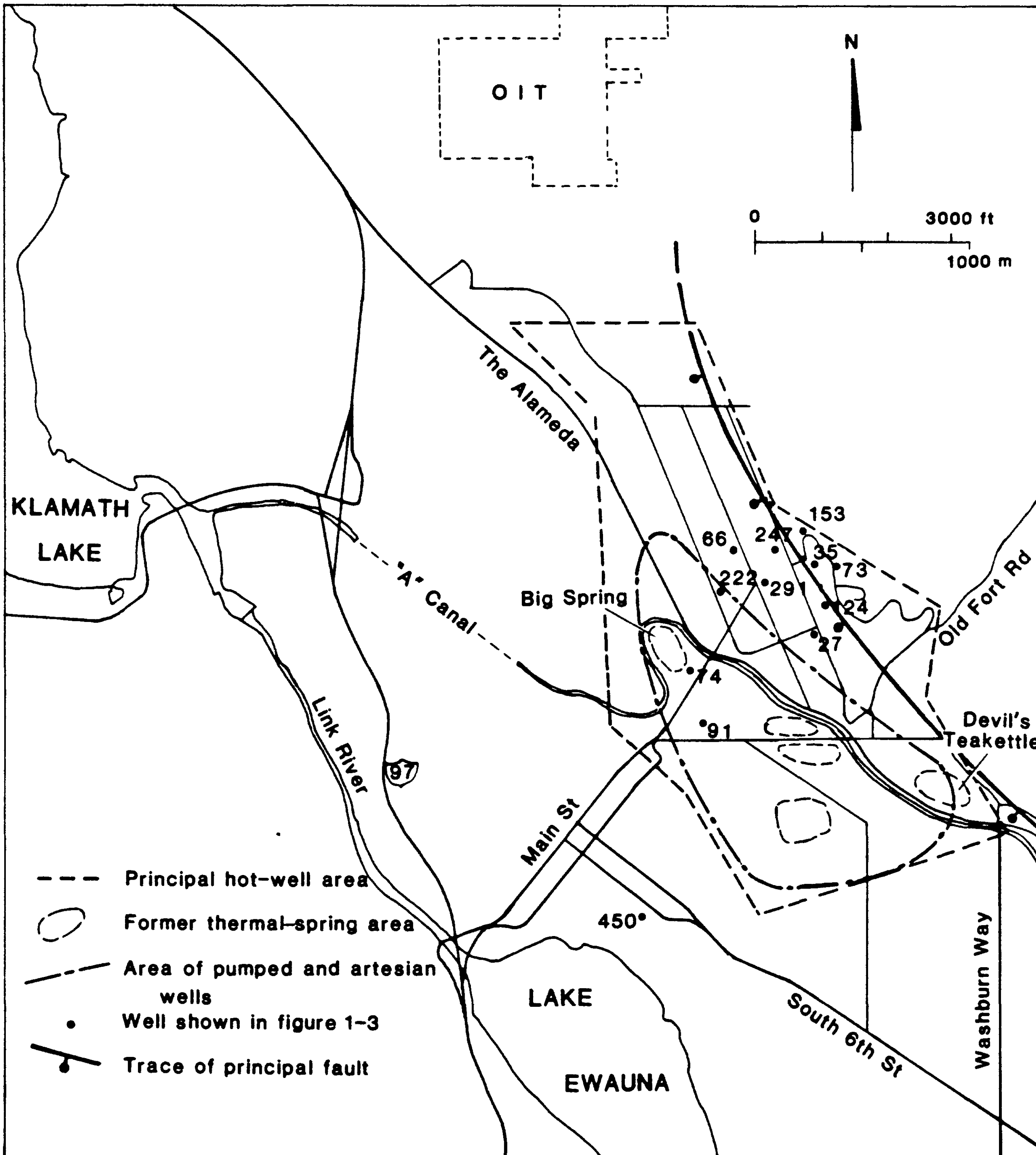

Figure 1-2. — Principal hot-well area, areas of former thermal springs and pumped and artesian wells, the trace of the principal fault, and locations of wells having temperature profiles shown in figure 1-3.

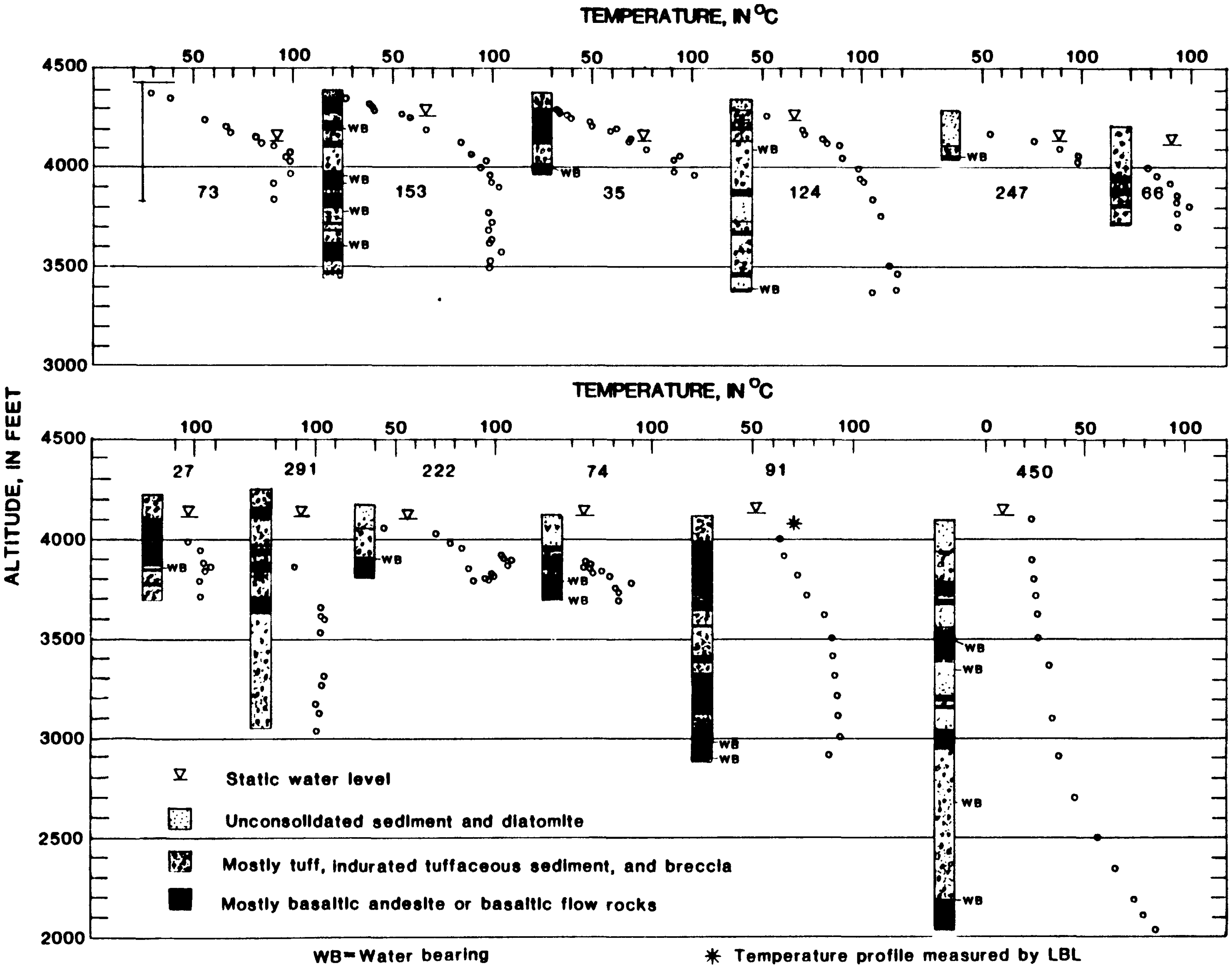

Figure 1-3. -- Generalized lithologic logs and reported bottom-hole temperatures measured during drilling. Numbers indicate wells located in figure 1-2.

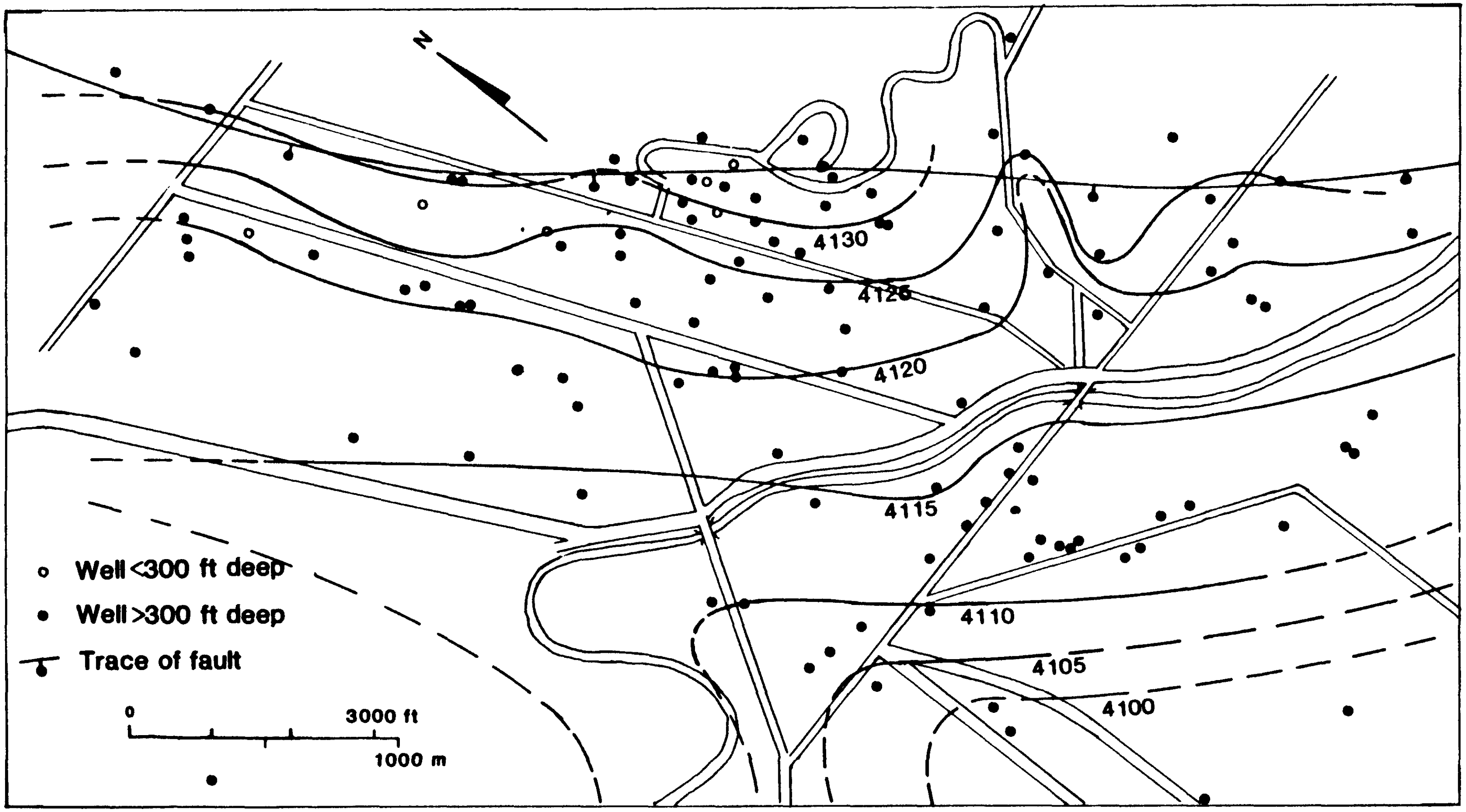

Figure 1-4. — Contours on altitudes of hydraulic heads measured in thermal wells. Contours define regions in which most wells have water-level altitudes in the indicated ranges. Altitudes in feet above sea level.

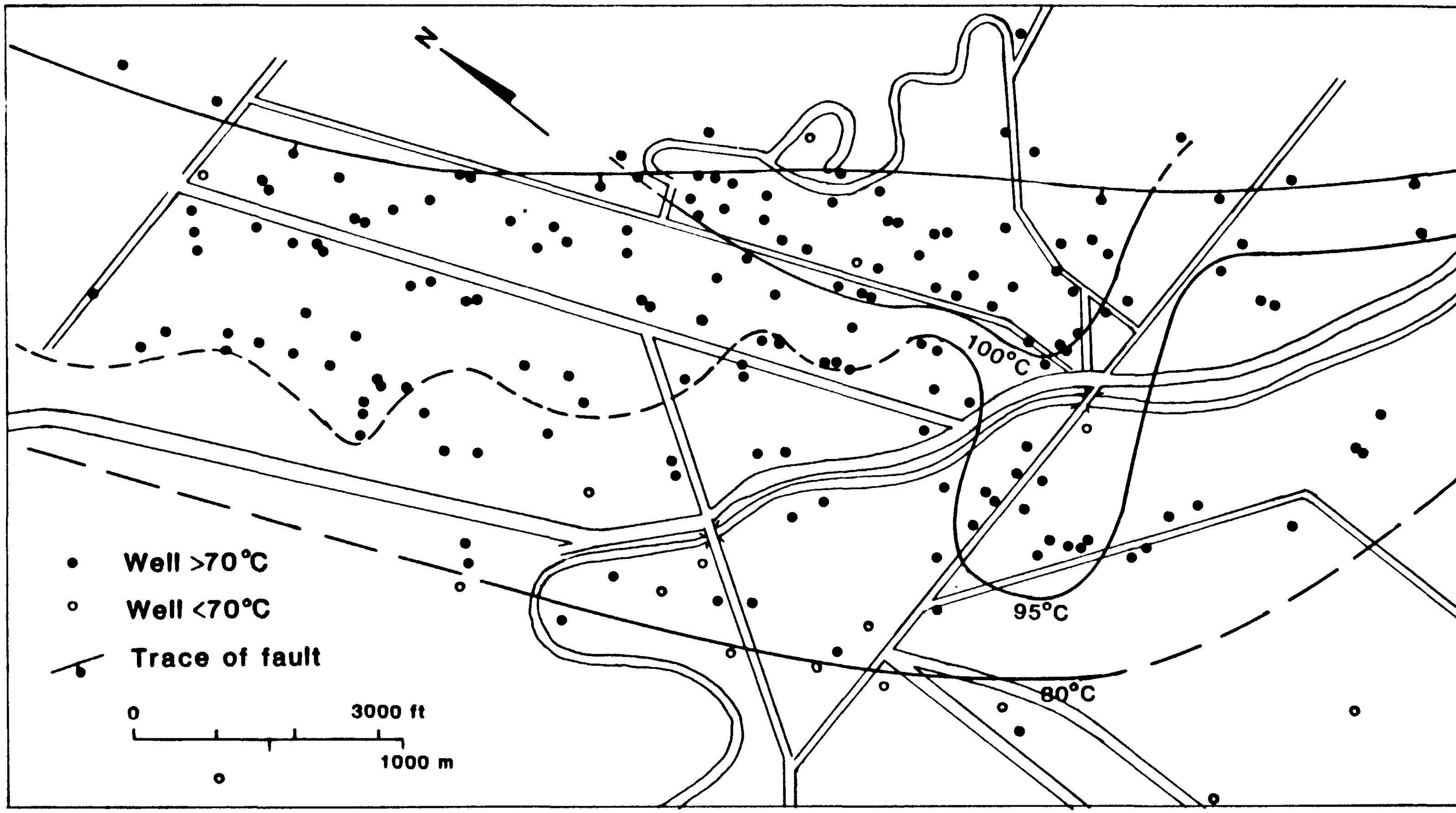

Figure 1-5. — Lines of equal temperature based on reported maximum temperatures measured in wells or in well discharges, in °C. The isotherms define regions in which most wells have reported temperatures in the ranges indicated.

CHAPTER 2. USE OF GEOTHERMAL WATER AND THE AQUIFER RESPONSE

By

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Geothermal Wells and Water Use

Hot-water wells at Klamath Falls were first used for heating by allowing the water to flow under artesian pressure from the well to the point of use. Heat was extracted by means of radiators or other simple heat-exchange devices and the water was then discharged. Later, as heat demands increased and artesian pressures declined, pumps were added to lift the water to the land surface. According to data compiled by OIT investigators, approximately 70 wells within the the principal hot-well area discharge thermal water, and nearly all are pumped during at least part of the heating season. Most of the wells are located in the sub-area outlined in figure 1-2.

At least 15 additional pumped wells are scattered through the region immediately adjacent to the hot-well area, and an increasing number of wells are being drilled in areas of lower temperature for use with binary-fluid heat-pump systems. Although wells outside the main study area are known to pump significant quantities of water, the resulting drawdowns in the aquifer underlying the main hot-well area are believed to be negligible. (See discussions in subsequent chapters of this report for the basis of this conclusion).

Discharge of thermal water from pumped and artesian wells in the hot-well area averages about 540 gallons per minute (gal/min) throughout the year. The monthly distribution of discharge is shown in table 2-1.

Table 2-1. -- Monthly distribution of average discharge from pumped and artesian thermal wells in the main hot-well area of Klamath Falls. Based on estimates and measurements by G. G. Culver, OIT. Discharge is in gallons per minute.

Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
720	695	665	650	565	300	300	300	300	610	670	695

The figures in table 2-1 are based on flows measured or estimated for 67 wells in the hot-well area, measurements of flow in the storm drains to which most of the thermal wells discharge, and the correlation of known discharges with daily average air temperatures. The minimum value of 300 gal/min shown for the months June through September probably is not as reliable as the remaining estimates, but it may be close to the figure for the discharge that continues perennially, regardless of air temperature, in order to supply heat for domestic hot-water systems. Some artesian wells are allowed to flow continuously whether or not the water is actually being used.

The correlation between discharge and air temperature is indicated by the graph in figure 2-1. These data, collected during the aquifer test in July and August, 1983, show that, above a base level of approximately 300 gal/min that is reached when the air temperature is above 70°F, discharge increases in a rather precise inverse proportion to the change in air temperature. This relationship, determined quantitatively by the OIT investigators during the winter of 1983-84 (fig. 2-2) served as a reliable index to the discharge, which was measured only at intervals during the winter.

In more than 85 percent of the hot wells at Klamath Falls heat is obtained from down-hole heat exchangers (DHE's). The heat exchange occurs as domestic cold water circulates in loops of pipe suspended in the wells. Heating of the cold water is made more efficient in many wells by increasing the flow of thermal water past the cold-water pipes. This is accomplished by perforations in the casing opposite points of water entry if the well is fully cased, and by perforations near the water surface in the aquifer. The hot water flows upward in the well and is discharged into cooler zones near the top of the aquifer.

The amount of heat utilized for space heating and domestic hot water by means of DHE's at Klamath Falls has been estimated by OIT investigators at about 13×10^{10} British Thermal Units (BTU) per year. The figure is based on estimates for all DHE wells on file at the time of inventory (December, 1983). It includes wells outside the main hot-well area, and a few DHE wells in which pumps discharge aquifer water in order to maintain high temperatures in the well. In comparison, the amount of heat withdrawn by pumped and artesian wells in the hot-well area only is calculated to be about 18 x 10^{12} BTU per year, or about 140 times the DHE discharge.

Thus the approximately 70 pumped wells discharge far more heat from the resource than the 380 or more DHE wells. Assuming that the two types of wells meet equal needs for heat on the average, the greater effectiveness of the DHE's is clearly indicated by these figures. An analysis of net energy use is beyond the scope of this study, but it can probably be assumed that most DHE's have a greater overall energy efficiency than pumped wells.

Water-Level Changes: The Effects of Use

Seasonal Effects

Water levels in the geothermal aquifer rise and fall in an annual cycle that reflects the heating demand and is inversely correlated with seasonal changes in air temperature. Long-term measurements recorded by several well owners show that, in most years, water levels recover from their winter low levels during the summer months and reach their annual high levels in late August. The advent of cooler weather in late August results in a water-level decline which continues through February of most years before the levels begin to stabilize. A significant rise of water level does not normally begin until April.

The pattern described above is illustrated by the graph of water levels in a 163-ft well (Hessig, No. 181) located near the center of the hot-well area (fig. 2-3). The well heats a home by means of a DHE and no water is withdrawn from the well. Maximum winter declines of water level in this well have ranged from about 6 1/2 ft to 9 ft since 1980. Annual recovery has been nearly complete except for the summer of 1983. During this summer, unusually cool weather and drawdowns created by the aquifer test combined to prevent a normal recovery. Seasonal water-level fluctuations in well 181 and in a number of other wells located immediately to the southwest are larger than those in the remainder of the aquifer, but the cyclic pattern probably is typical of the entire aquifer. For comparison, the smaller drawdowns that occured during the winter of 1979-80 in the Raney well (No. 143) and the Adamcheck well (No. 127) are shown in figure 2-4.

The annual cycle of water-level fluctuations can be compared with the annual climatic cycle shown in figure 2-5. Whereas the monthly mean temperatures have a maximum in July and a minimum in January, water levels reach maximum levels at the end of August and minimum levels in February. The lag time in the water-level response is partly a consequence of averaging the air temperatures over a monthly period. This averaging could displace the maximum and minimum points by 1/2 month from their true positions. However, some of the lag probably represents a delay in the overall aquifer response in relation to heating demands.

Seasonal water-level declines in the aquifer are less than a foot at the margins of the main hot-well area and are more than 11 feet near the center of the area. An indication of the areal distribution of water-level fluctuations is shown by the net change that occurred in representative wells between September 1983 and February 1984 (fig. 2-6). These fluctuations are smaller than normal ones because of low-water levels created by the aquifer test in the summer of 1983 and higher than normal levels during the winter of 1983-84 resulting from relatively mild weather after December 1983. Thus, the changes shown in figure 2-6 are probably minimum seasonal changes for the wells selected. In contrast to other wells shown in figure 2-6, well 450, located near Lake Ewauna, showed a significant water-level rise and appears to have a different cycle of fluctuations, suggesting that this low-temperature well responds to influences from outside the hot-well area. On the basis of the change observed in well 141, located at the northern edge of the hot-well area (fig. 2-6), wells as far north as those at OIT are assumed to have only small effects on water levels in the hot-well area. Wells located south or southeast of the area have low hydraulic heads and temperatures and also are assumed to have negligible effects on the main aquifer (Sammel, 1980).

Wells that contain DHE's usually show additional fluctuations resulting from cooling of the well during heat-exchanger use. The cooling produces thermodynamic compression of the water column and a consequent lowering of the water level. This effect varies with well characteristics, heat demand, and aquifer permeability.

A typical example is shown in figure 2-7, where the water level is observed to drop as much as 0.4 ft in respose to the daily withdrawal of heat from the well. Both the decline and the recovery are rapid, indicating that the effect occurs largely in the well rather than in the aquifer. However, fluctuations in the well are transmitted to the aquifer, and the combined effects from several hundred DHE wells probably are significant.

On June 11 and 12, 1983, an unusual test of the aquifer response was conducted by well owners. During a 24-hour period, a voluntary effort was made to turn off all geothermal heat systems and to observe water levels in as many wells as possible. It was hoped that a response would be observed in the DHE area as well as in the pumped-well area and that insight might be gained on the effects of DHE wells. Participation by well owners in the test was estimated to be as high as 80 percent, a remarkable tribute to the spirit of cooperation that characterized the testing program. Water-level changes were extremely small, however, and were masked by the rise and fluctuations that normally occur at this time of year. Although this test could not be successfully analyzed for the effects of DHE use, a similiar test conducted during a period of stable-water levels at the height of the heating season probably would produce significant water-level changes that could be analyzed.

Seismic Effects

Fault displacement and major seismic activity at Klamath Falls had largely ceased by the end of Pleistocene time (1 to 2 million years ago), but a low level of seismic activity continues in present time (Couch and Lowell, 1971). No recent movement has been detected along the faults, although it is possible that slow creep still occurs.

It is known that the withdrawal and injection of fluids in wells can, at some places, trigger seismic activity or cause subsidence of the land surface. At the Geysers geothermal field, California, for example, increased low-level seismic noise is related to fluid production and (or) reinjection, and subsidence occurs at a rate of 20 or more millimeters per year (Allis, 1982). At the Rocky Mountain Arsenal, near Denver, seismic activity, including minor felt earthquakes, has occurred as the result of high-pressure injection of fluids (Healy and others, 1968). On the other hand, no seismicity had been induced by production or reinjection at the long-exploited geothermal field at Larderello, Italy (Batini and others, 1980), or at the Otake geothermal field in Japan (Kubota and Aosaki, 1975). Thus, increased seismic activity and subsidence are not inevitable consequences of geothermal exploitation.

Seismic activity can occur because of increased or decreased fluid pressures in a reservoir or because of significant temperature changes. Subsidence occurs because of decreased pressures in elastic rocks or compressible sediments. The changes in fluid pressure must, in general, be a significant fraction of the lithostatic load borne by the rocks in order to produce detectable changes in the normal seismic activity that characterizes nearly all regions of the earth. Similarly, temperature changes must be rather large in order to produce significant volume or pressure changes in the rocks. Hundreds of aquifers in volcanic regions are pumped or injected with no detectable effects on seismicity or subsidence.

In order to determine the nature of seismic activity at Klamath Falls, seismic monitoring was included in the data-gathering activities of our study. A seismograph, installed by LBL in a well near Hillview Street, continuously recorded seismic events prior to and during the aquifer test. Several barely detectable seismic events occured during this period, but there was no increase during the pumping or injection phases of the test. The passage of freight trains through Klamath Falls was easily detected by the intrument, and the seismic noise generated by the trains was an order of magnitude greater than the natural events.

The July 21 earthquake at Coalinga, California (Richter magnitude 5.9), was clearly recorded by the Stevens recorder in 4 monitor wells (Svanevik, Eck, Parks, and Jones). These wells presumably responded at the proper resonant frequency for detection of the relatively long-period waves of the earthquake. The quake was not detected on the seismometer instrument which was designed to monitor higher frequency local events.

The rocks of the Klamath Falls aquifer probably are subjected frequently to stresses generated by earthquakes and changes in the crustal stress field. Although the effects of these stresses have not been documented, it seems highly unlikely that the much smaller stresses resulting from pumping, injection, and temperature changes will significantly alter the aquifer fabric or affect the rate of creep in the faults. Nor is it likely that water-level drawdowns of a few tens of feet will induce detectable subsidence in the competent rocks within or overlying the aquifer. Thus, the risk of changing the natural patterns of seismic activity, rock creep, or subsidence at Klamath Falls appears to be extremely small.

Long-Term Effects

A widely accepted belief among well owners in Klamath Falls is that water levels have declined since the early 1900's as the result of increasing withdrawals of water from the aquifer. The disappearance of the thermal springs, an apparent decrease in artesian heads, and reports of lowered levels in DHE wells have been cited in support of this belief. Evidence examined for this study tends to confirm the belief that average water levels have declined, but unequivocal evidence of the cause has been difficult to find. Precise documentation of the decline and disappearance of the springs and the decrease of artesian heads in relation to the growth of pumping and other use is not yet available. Furthermore, no search has yet been made in historical climatic records for data that might also have a correlation with water-level declines.

The cause of a possible long-term decline in water level is clearly a matter of concern for this study. The reliability of any evaluation of reservoir potential depends on a knowledge of the factors involved in both short- and long-term changes. In the paragraphs below, the extent and limitations of our present knowledge are described.

Several possible causes might be invoked to explain the disappearance of the thermal springs. For example, the dredging of the "A" Canal might have drawn off the spring water. Alternatively, leakage from the canal is known to occur, and this might have suppressed the spring flow by imposing a higher hydraulic head or by cooling the spring water at shallow depths. Neither of these possibilities is a probable cause however. Flow in the canal is intermittent, and during a long winter period there is no water in the canal. Furthermore, the hydraulic head in at least one spring area (Devil's Teakettle) would always have been higher than canal levels. Therefore, at least partial recovery of the springs should have occurred during the winter if hydraulic heads had been suppressed, and some cooling should have been noted prior to their disappearance if cooling were the principal cause. Finally, two of the spring areas (Big Spring and Devil's Teakettle, fig. 1-2) are relatively unaffected by construction, landfills, or other consequences of urbanization, which might otherwise have been considered possible reasons for the disappearance of the springs.

Among other possible causes of spring declines is the sealing of conduits by silica. The deposition of silica has occurred extensively at Klamath Falls, as indicated by the widespread silicification of Tertiary rocks in the uplands of the region (Peterson and Groh, 1967). It seems improbable, however, that this process would have acted so uniformly in time and space as to close off all the springs in the five spring areas during the same short period of time and that no new outlets would have appeared.

The remaining and most probable explanation is a decrease in hydraulic head in the aquifer. Reports of a decline in artesian head seem to be well substantiated (Charles Leib, oral commun., 1983). Estimates of the magnitude of the decline vary, but in several wells, declines have been at least 15 feet since the 1930's and 1940's (John Lund, Charles Leib, oral commun., 1978 and 1983 respectively). Currently, maximum artesian heads are probably 5 feet or less above land surface, and most artesian wells are pumped during the heating season in order to maintain discharges (G. G. Culver written commun., 1983).

Winter low water levels in many DHE wells also have declined during the past 40 years, according to reports of well owners in the hot-well area. In well 181, for example, winter low levels are 4 to 5 feet below the levels of 6 years ago (fig. 2-3). In other parts of the aquifer, smaller but still significant declines have occurred. These changes can almost certainly be attributed to increasing withdrawals from the aquifer.

There may also be a long-term decrease in the annual recovery levels in the DHE-well area. Owners report declines of recovery levels ranging from less than a foot to several feet during the past 20 to 40 years (CRGD, unpublished data, 1983). The occurrence of such changes is not documented over most of the aquifer, but the reality of the reported decline seems highly probable if, as seems nearly certain, there has been a significant decline in maximum artesian levels.

The most likely causes of long-term declines are, (1) a decrease in recharge to the geothermal system, and (2) an annual discharge by pumped and

DHE wells that exceeds the annual recharge of water and heat. Although the increasing demands on the aquifer are clearly implicated in the annual seasonal declines, the data collected thus far do not permit the conclusion that withdrawals are the sole cause of the long-term decline.

Chemical evidence described in this report indicates that recharge to the geothermal reservoir may originate in precipitation at altitudes higher than those near Klamath Falls and therefore may occur over a large area. Consequently, the travel times of thermal water may be tens or hundreds of years. Under these conditions, a correlation with climatic records may be difficult or impossible to obtain, and thus one possible cause of water-level declines may be indeterminate. It can also be postulated that hydraulic heads in the aquifer are affected by transient pressure changes resulting from changes in precipitation patterns in the immediate vicinity of the hot-well area. If so, it might be possible to determine a correlation and single out this relationship as a probable cause of water-level changes.

The pattern of seasonal changes that would occur naturally in the aquifer in the absence of withdrawals could be an important clue to the relation between climatic change and water-level changes. Data of this kind are not available, and, unless early accounts of changes in spring activity can be found, the natural seasonal change is not likely to be known. Thus, our understanding of fundamental processes at Klamath Falls is limited and may remain so unless new and different types of data are obtained.

The long-term effects of current withdrawals on water levels are similarly unknown, but are more amenable to discovery by measurement and analysis. It is clear, however, that a general decline of annual recovery levels could only be aggravated by a continued increase of withdrawals in the absence of reinjection and conservation measures. Discussions in succeeding chapters deal largely with short-term causes and effects. Final conclusions regarding long-term conditions must be deferred until more knowledge is available.

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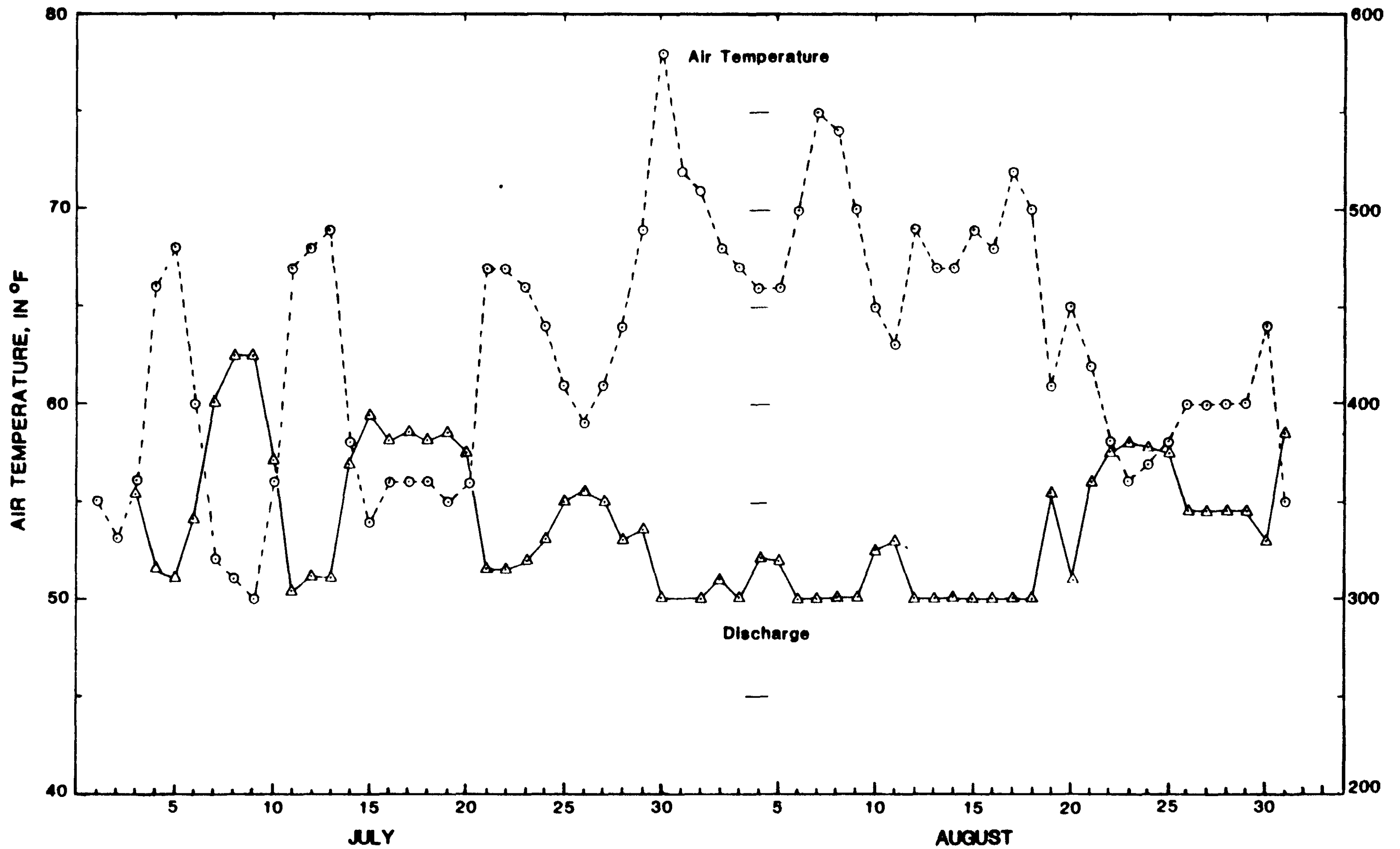

Figure 2-1. — Average daily temperature and discharge from thermal wells, July and August, 1983.

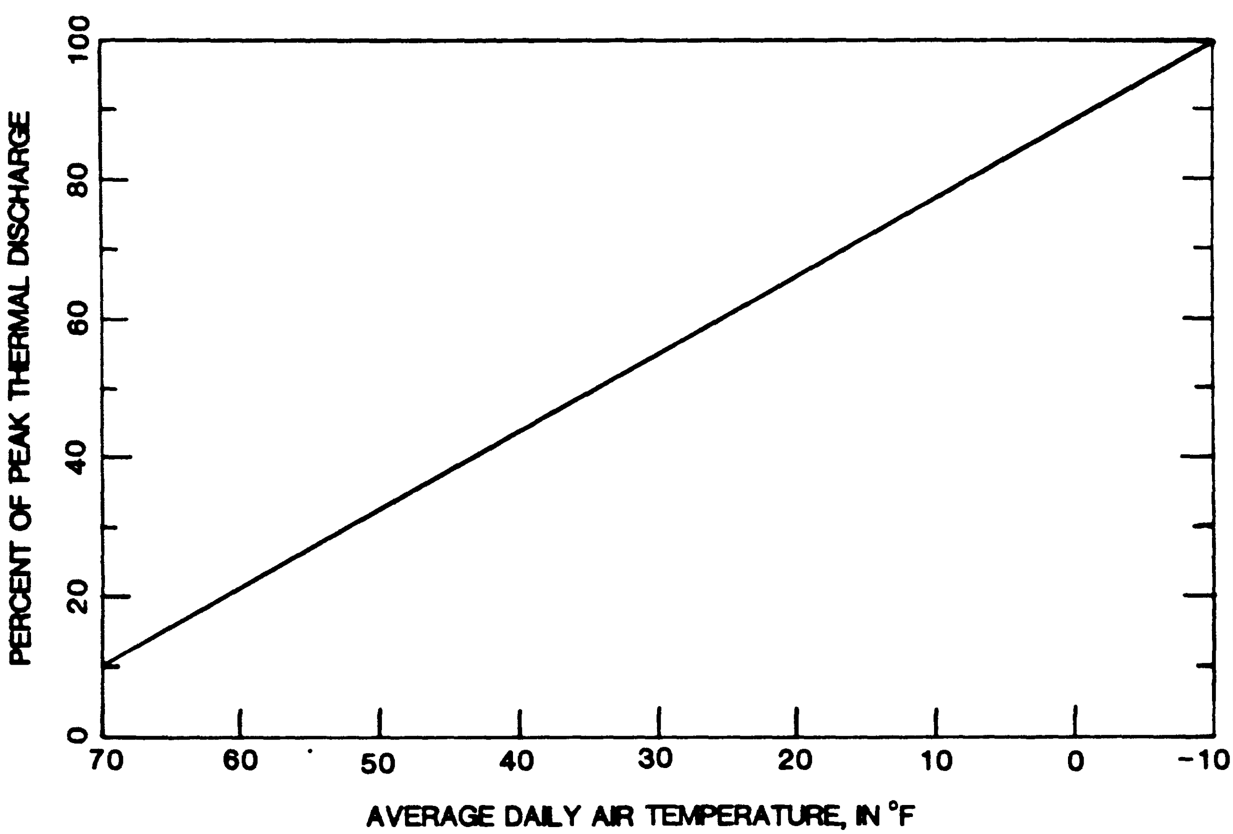

Figure 2-2. — The relationship between average daily air temperature and percent of peak thermal discharge.

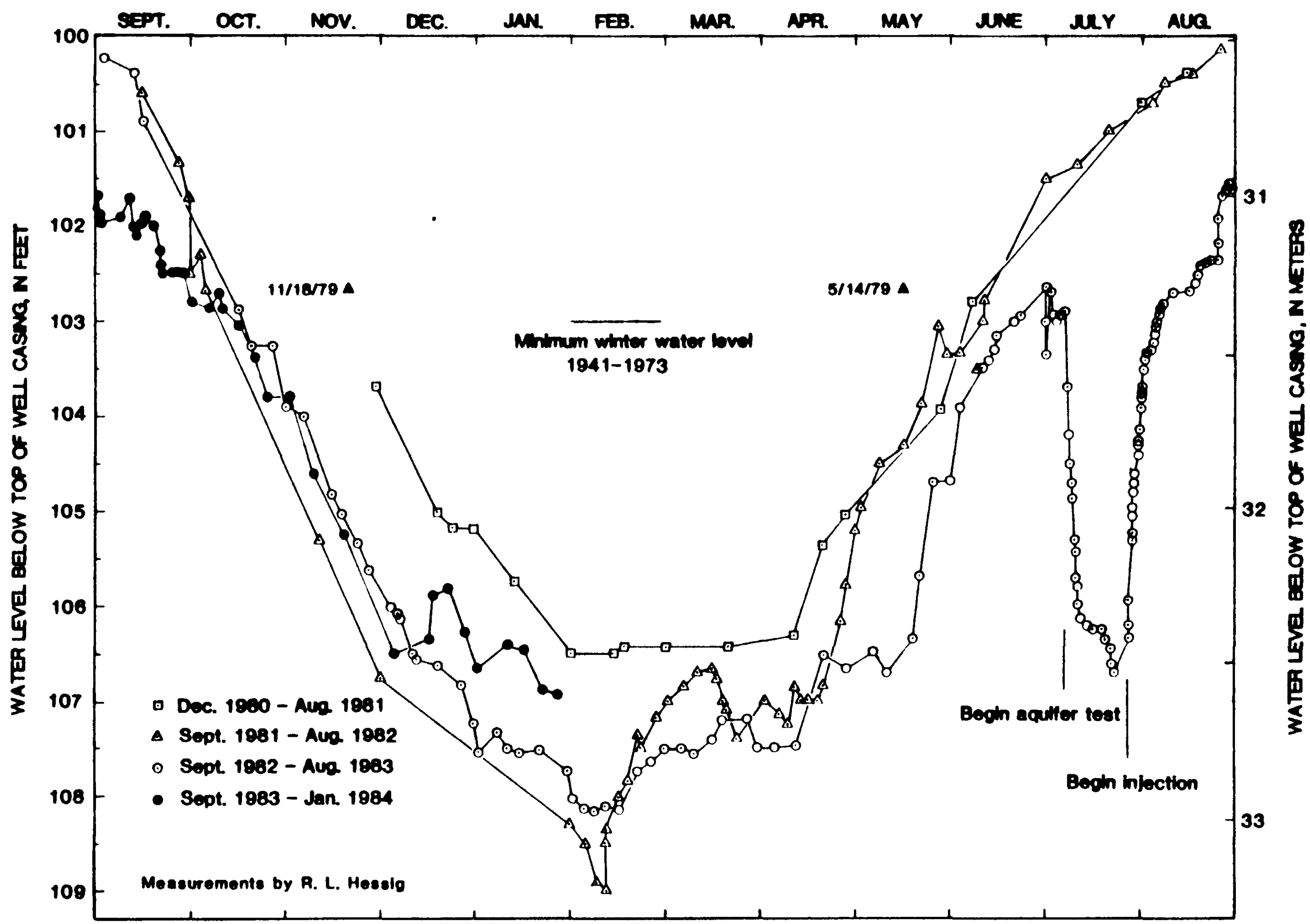

Figure 2-3. -- Annual water-level fluctuations in well 181 (Hessig) for the years 1981-1984.

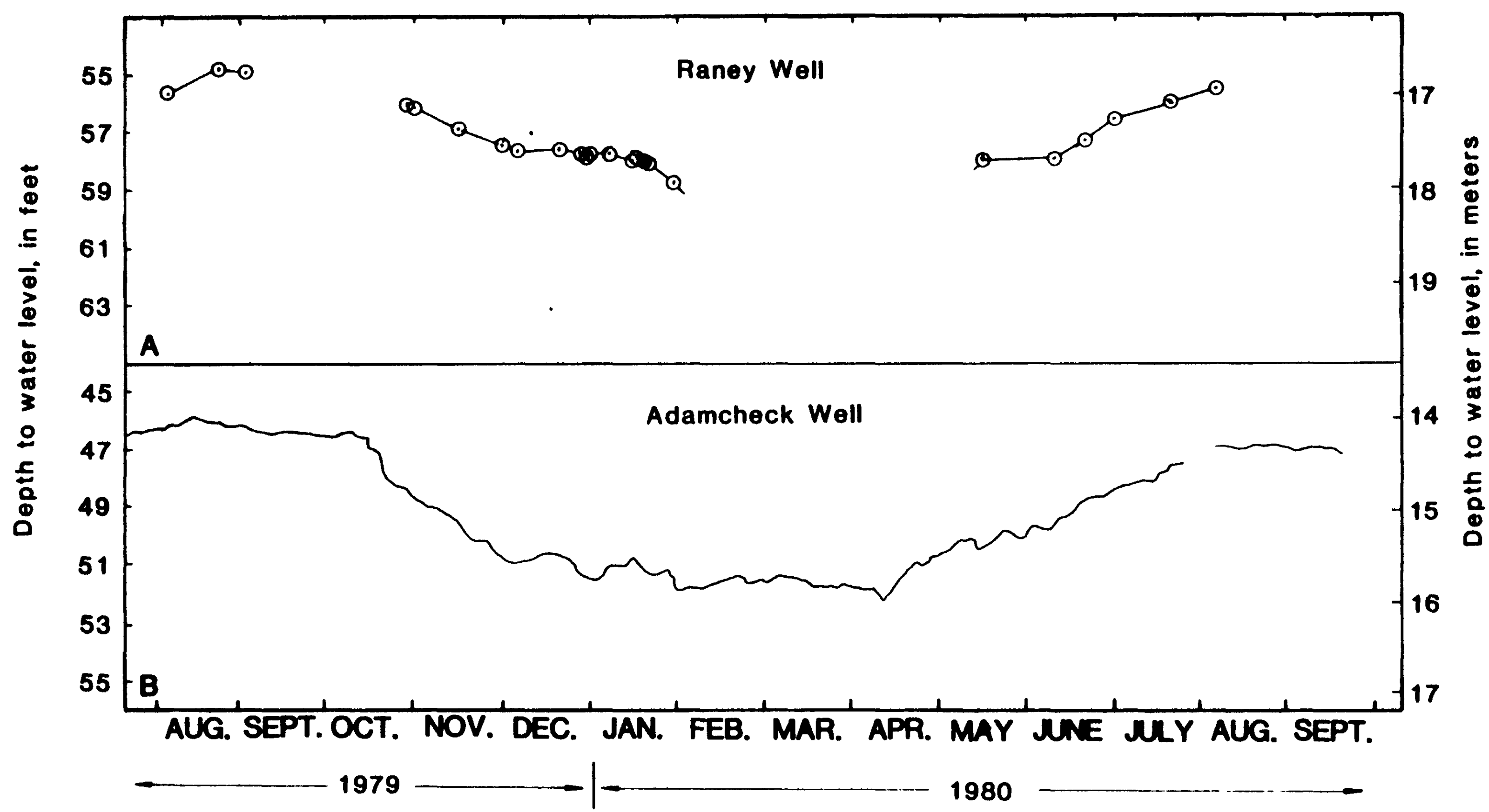

Figure 2-4. — Water-level fluctuations: (A) in well 143 (Raney) and (B) in well 127 (Adamcheck) during the period August, 1979 to September. 1980.

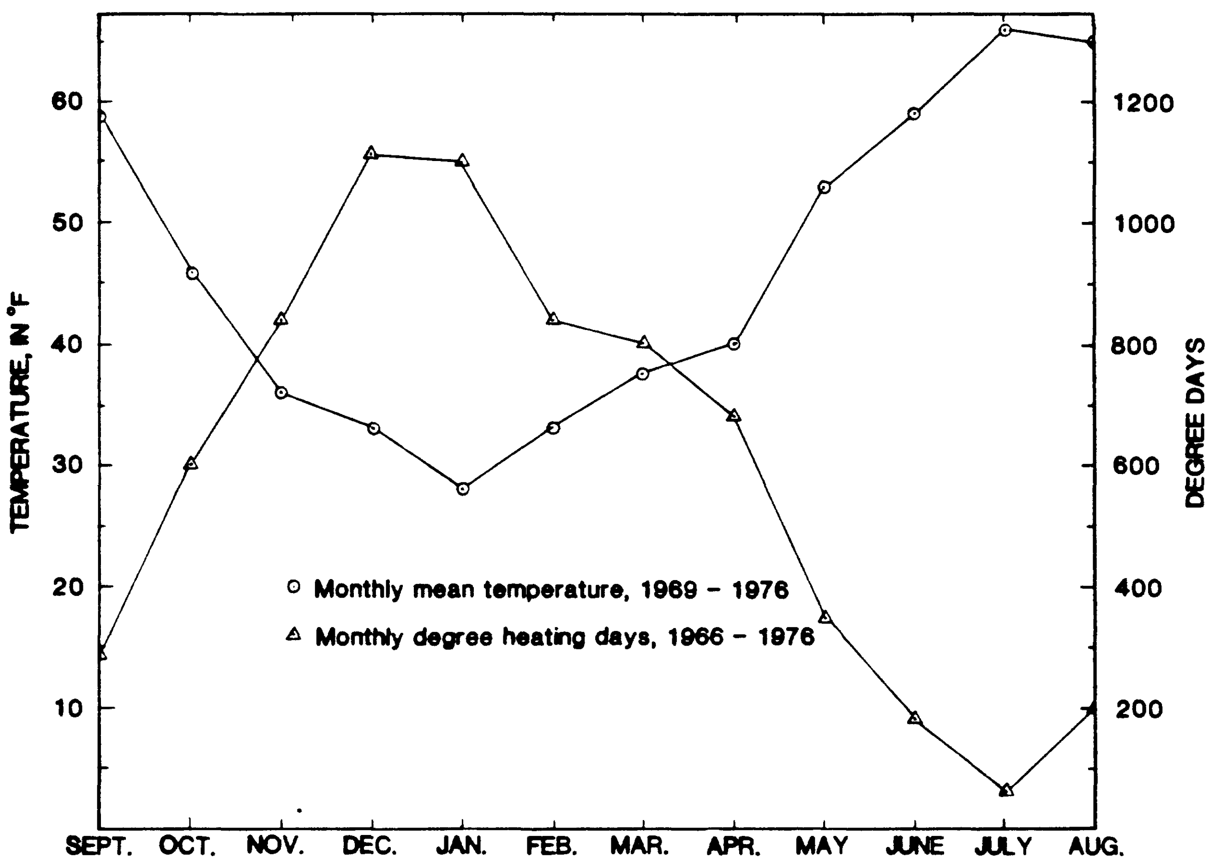

Figure 2-5. -- Monthly mean air temperature and monthly degree heating days for the years 1969-1976. Data from the National Weather Service for Kingsley Field, Klamath Falls.

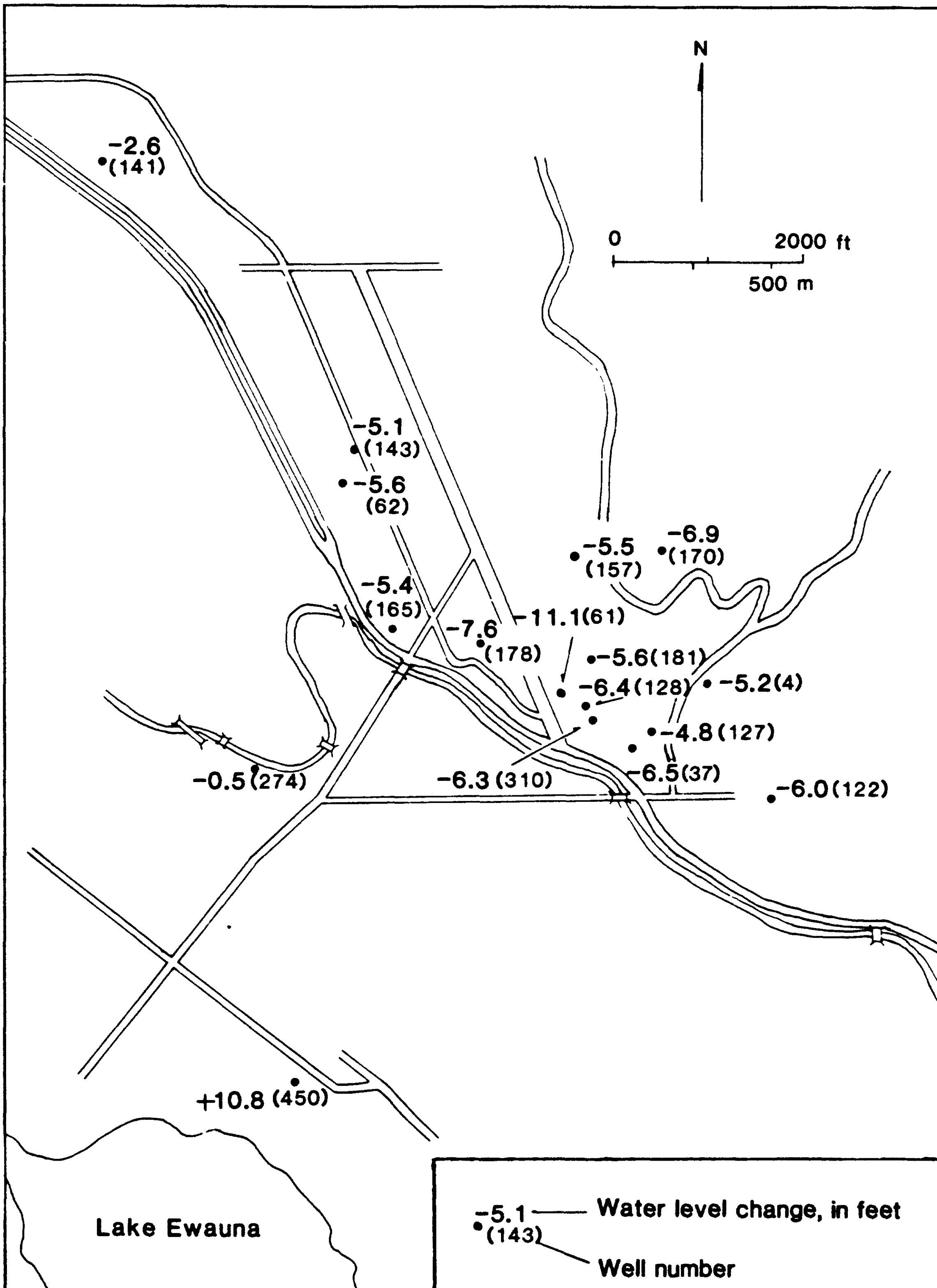

Figure 2-6. — Changes in water level observed between September, 1983 and February, 1984 in the indicated wells.

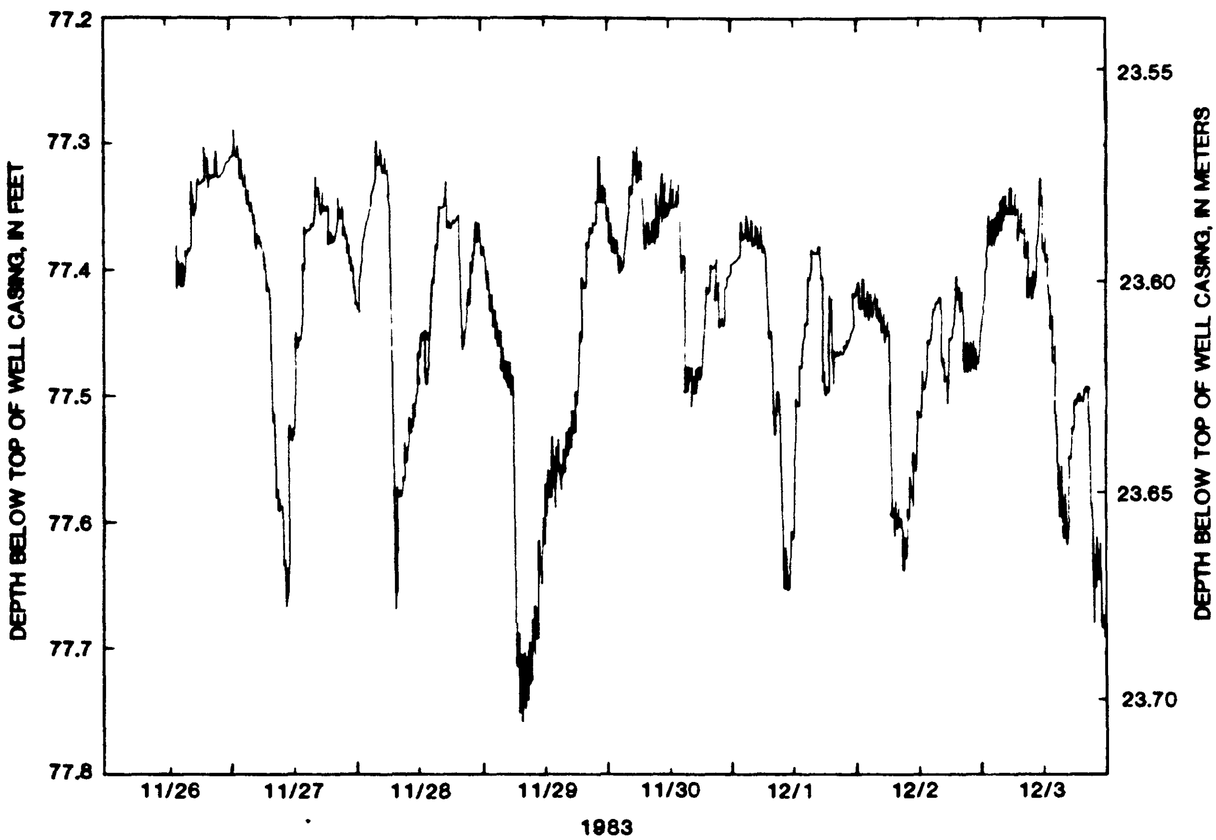

Figure 2-7. — Fluctuations of water level resulting from withdrawals of heat in a down-hole heat exchanger in well 118 (Svanevik), November 26 to December 3, 1983.

CHAPTER 3. GEOCHEMISTRY OF THERMAL WELL WATERS AT KLAMATH FALLS, OREGON

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Introduction

Thermal waters collected from the Klamath Falls geothermal aquifer in the month prior to and during pumping tests in 1983 were analyzed for chemical and isotopic constituents. These analyses and the sampling and analytical methods used were published by Janik and others (1984). Discrepancies in Ca and Cl data resulting from analytical error were corrected by repeat analyses. In this chapter we interpret these results and earlier analyses as indications of the temperature and reservoir processes of the geothermal aquifer.

Chemical Compositions

Analyses of Klamath Falls thermal and nonthermal waters from Janik and others (1984) (with revised Ca and Cl data) and from Sammel (1980) are given in table 3-1. A Schoeller diagram comparing concentrations of chemical constituents from selected analyses is given in figure 3-1. Thermal waters from Klamath Falls wells contain (in order of decreasing concentration) SO_4, Na, SiO_2, Cl, HCO_3, Ca, and K with traces of F, Li, Mg, and Al. Nonthermal well waters are more dilute and contain (in order of decreasing concentration HCO_3, SiO_2, Na, Ca, Mg, Cl, K, and SO_4. Cold spring waters in the vicinity of Klamath Falls contain less Na and Cl than nonthermal well waters (table 3-1). Constituents of thermal waters show limited ranges of concentration, with most variation in K, Ca, Mg, and SiO_2 (fig. 3-1). An increase in SiO_2, Na, K, and Cl concentrations and in temperature is observed for samples collected during the pumping tests. As discussed later, the variation in chemistry of the thermal waters is apparently caused by mixing with cooler waters of different composition and equilibration with rock minerals at different temperatures.

Isotopic Compositions

Water from Klamath Falls cold wells and springs is isotopically similar to rainwater but shows some effects of evaporation before infiltration. The oxygen-18 and deuterium contents of these waters fall along a trend parallel to the normal "meteoric water line" (MWL), $\delta D = 8\ \delta^{18}O + 10$ as defined by Craig (1961), but offset by +0.5 permil in $\delta^{18}O$ (fig. 3-2). The thermal waters are significantly lower in δD and higher in $\delta^{18}O$ than local cold waters (fig. 3-2). Concentrations of D and ^{18}O in precipitation worldwide have been observed to decrease with increase in elevation, latitude, and distance inland, and with decrease in temperature (Gat, 1980). Thus the lower deuterium content of the Klamath Falls thermal waters compared to that of the cold waters, suggests that the recharge to the geothermal aquifer occurs at greater elevations than the recharge to the cold aquifer or, much less probably, consists of old waters from a time of colder climate (Buchardt and Fritz, 1980). The higher ^{18}O concentrations in the thermal waters relative to waters on the MWL represents an "oxygen isotope shift" caused by long contact with ^{18}O-rich rock minerals at elevated temperatures. The isotopic (^{18}O, D) variation of the thermal waters results from mixing with local cold water (fig. 3-2).

The tritium content of a sample from the city's major cold-water supply well (#500) is very low at 0.14 tritium units (TU). This suggests that the residence time in the cold aquifer is greater than 30 years because this water must have a negligible contribution from precipitation (with 30 to 1,000 TU) that postdates nuclear bomb testing in the mid-1950s. The tritium in this water may represent prebomb tritium (estimated at 10 TU originally), which has undergone radioactive decay during 6 half lives of 12.3 years indicating that the water is older than 60 years (Gat, 1980). In a well-mixed reservoir, the average age would be greater than 10,000 years (Pearson and Truesdell, 1978). The second cold well sampled in 1983 (#501) has higher tritium (0.71 TU), indicating either a small addition of more recent precipitation or a smaller residence time than the water of well #500.

The tritium contents of the thermal waters range from 0 to 1.6 TU; one sample containing about 8 TU (#304) is probably contaminated with surface water. Most thermal samples have tritium contents near zero (<0.3 TU), indicating greater than 30-year storage as discussed above. Some thermal waters have higher tritium and lower chloride, suggesting mixing with younger, more dilute waters (fig. 3-3a, b). These higher-tritium waters tend also to be cooler, as shown in figure 3-4.

Mixing of Thermal and Nonthermal Water

The relations of temperature, chloride, tritium, and other constituents of the thermal waters indicate mixing. A reasonably linear chloride-temperature mixing relation is observed for samples collected prior to the pumping tests (fig. 3-5), suggesting that the cold end member is a water at 20°C with about 10.5 mg/kg Cl. The temperature of cold water at depths of any possible mixing is assumed to be 20°C because of heating by conduction. Recharge of this cold end-member water probably does not originate from modern Klamath Lake water because the extrapolated tritium contents at 10.5 mg/kg Cl from figure 3-3b are only about 2.5 TU, whereas Klamath Lake had a tritium concentration of 25.7 TU when sampled (Sammel, 1980). Klamath Lake should have higher tritium concentrations than present precipitation because it contains stored older rainwater with higher tritium. (The tritium concentration of precipitation is decreasing faster than would be expected from radioactive decay because it is being diluted with deep, tritium-free ocean water.) Klamath Lake also has a higher deuterium concentration than other cold waters (Sammel, 1980), making it an unlikely source of recharge.

Samples collected during the pumping tests have chloride concentrations that are nearly independent of temperature (fig. 3-5). These waters may have been out of thermal equilibrium because of more rapid flow in the aquifer. Higher concentrations of SiO_2 relative to Cl (fig. 3-6) also indicate that non-equilibrium conditions occurred during the aquifer tests.

The high-chloride, high-temperature end member of the mixing relation (fig. 3-5) is not defined and the maximum temperature of 98°C in the waters sampled is less than the highest measured at Klamath Falls (140°C, P. J. Lienau, OIT, written commun., 1982). Tritium cannot be used as an indicator of the hot end member because most waters have near zero tritium concentrations.

Geothermometers and Mixing Models

Certain chemical and isotopic reactions reequilibrate sufficiently slowly as fluids cool to lower temperatures that evidence of higher temperature equilibria are preserved. These reactions may thus be used as geothermometers and have been calibrated experimentally or empirically to indicate probable maximum temperatures attained. Calculated geothermometer temperatures for Klamath Falls thermal waters are given in table 3-2. In dilute waters, cation geothermometers are likely to be affected by re-equilibration, and at Klamath Falls they show temperatures close to those measured at the sampling point. The average temperature from the Na-K-Ca geothermometer (Fournier and Truesdell, 1973) is 81 $\pm$ 6°C. Cation geothermometer temperatures of samples taken before the pumping test agree closely with measured temperatures. Samples taken during pumping agree less well because waters chemically equilibrated at other temperatures were rapidly heated or cooled during passage to the wells. Silica concentrations are greater than those expected for saturation with silica minerals (other than amorphous silica) at sampling temperatures and suggest equilibration at higher temperatures, deeper in the reservoir (fig. 3-7). (Silica in the well waters cannot result from equilibrium with amorphous silica because the waters are undersaturated with this mineral.) Direct use of silica geothermometers suggests temperatures of 100 to 150°C (table 3-2) but silica concentrations are probably affected by mixing as discussed below.

The sulfate-water isotope geothermometer depends on fractionation of ^{18}O between SO_4 and H_2O, a process that is reasonably rapid at high temperatures but very slow at low temperatures (McKenzie and Truesdell, 1977). At Klamath Falls, this geothermometer is unlikely to be influenced by contamination because the thermal waters have higher SO_4 than cold waters and because there is little or no hydrogen sulfide to produce extra SO_4. The temperature indicated by using the observed water-^{18}O compositions is 189 $\pm$ 4°C for thermal waters (table 3-2).

Silica mixing calculations (Truesdell and Fournier, 1977) based on 1983 silica data indicate an average temperature of 185 $\pm$ 18°C (1 standard deviation of 14 samples with 2 outlying values excluded). Using only data on samples collected during the pumping tests, the average calculated temperature is 192°C. Silica concentrations previously reported from wells sampled in this study produced a wider range of mixing-model temperatures (148°C to 180°C) and led to a lower estimate of reservoir temperatures

(Sammel, 1980). Not all previous samples were properly treated to preserve silica and the recent analyses are probably more reliable. The estimate of 185°C is consistent with sulfate isotope temperatures of 189°C. If equilibration with quartz is assumed at a temperature of 185°C (Fournier and Potter, 1982), and if the cold and mixed waters contain 45 and 120 mg/kg SiO_2 respectively, then the fraction of high-temperature water in the reservoir mixture is calculated to be about 44 percent.

Using 185°C as the temperature of the hot-water end member in a chloride-temperature mixing model, and assuming the cold water to contain 10.5 mg/kg Cl at 20°C and the mixed water to contain 55 mg/kg Cl, on the basis of the revised analyses, the chloride concentration of the hot-water end member is calculated to be about 120 mg/kg and the fraction of hot water is about 40 percent, in good agreement with the fraction derived by means of the silica-based model. Applying the average of the two mixing fractions (42 percent) to the isotope graph in figure 3-2 and assuming no oxygen shift for the cold end member, we calculate that the reservoir water may have a $\delta^{18}O$ value about -13.7 and a δ D. value near -132.

Temperature, Age, and Volume of Thermal Water

From considerations of mixing, from geothermometry, and from tritium analyses we can form a conceptual model of the geothermal system at Klamath Falls. Wells sampled appear to draw water from a mixing zone at 70 to 100°C where hot water with zero tritium mixes with a cold water at about 20°C with 10.5 mg/kg Cl and 2.5 TU tritium. Different mixing ratios in the mixing zone result in well waters of different temperatures and compositions. This mixing may occur in a shallow reservoir connected both to cold-water aquifers and to a deeper high-temperature reservoir. Although the indicated high-temperature end-member water has not been encountered by wells drilled thus far, the geochemical relations indicate temperatures of 150 to 190°C somewhere in the system.

If the outflow of thermal water from the system is of the order of 1,000 to 2,000 gal/min (500 to 1,000 gal/min from wells, with an equal or greater natural flow into cold-water aquifers) and the age of the thermal water is greater than 30 years as indicated by its tritium contents, then the volume of the thermal reservoir could be relatively large.

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Table 3-1. -- Chemical and isotopic analyses[1/] of water from wells and springs at Klamath Falls, Oregon (Chemical analyses in mg/kg except for Al, which is μg/kg; isotope analyses in permil relative to VSMOW [Gonfiantini, 1979]; tritium in Tritium Units)

Sample	Date	Temperature (°C)	SiO_2	Na	K	Li	Ca	Mg	Al	Cl	F	SO_4	HCO_3	δ D	$\delta^{18}O$	Tritium	$\delta^{18}O$ aq. SO_4
Hot water well sampled the month prior to 1983 pumping tests [1/]																	
25	6/04/83	78	98.0	206	5.08	--	31.8	0.02	27	51.7	---	398	45.2	-119.4	-14.78	0.23 ± 0.08	-5.34
45	6/09/83	88	98.6	207	5.47	--	24.7	0.05	27	48.7	---	425	46.4	-119.4	-14.50	0.33 ± 0.08	-5.18
45A	6/09/83	80	172.0	220	6.65	--	24.8	0.12	135	49.4	---	421	47.6	-120.2	-14.69	0.04 ± 0.08	-5.37
65	6/14/83	66	83.0	194	4.30	--	29.8	0.05	27	47.7	---	382	50.0	-119.8	-14.54	1.03 ± 0.09	-5.12
76	6/07/83	78	90.7	189	5.08	--	22.4	0.02	27	41.9	---	362	46.4	-119.4	-14.72	0.25 ± 0.08	-5.44
110	6/21/83	91	98.6	213	5.47	--	26.3	0.02	27	52.8	---	423	46.4	-121.3	-14.45	0.52 ± 0.09	-5.43
216	6/29/83	72	123.2	217	5.84	0.31	25.0	0.04	22	51.7	1.2	409	43.3	-121.3	-14.53	0.56 ± 0.09	-5.36
259	6/08/83	56	89.7	205	5.08	--	28.9	0.05	27	51.7	---	---	45.8	-120.3	-14.61	0.21 ± 0.08	-5.21
304	6/08/83	56	80.2	175	3.91	--	16.7	0.02	27	40.1	---	293	86.0	-116.1	-14.05	8.35 ± 0.30	-4.56
Cold water wells [1/]																	
500	6/20/83	21[a]	58.0	23.5	2.74	--	10.2	4.37	27	5.7	---	2.2	123	-112.3	-14.86	0.14 ± 0.06	---
501	6/23/83	17	50.1	20.0	3.13	--	15.8	7.73	27	7.2	---	1.8	136	-108.8	-14.26	0.71 ± 0.09[b]	---
Hot water wells sampled during 1983 pumping tests [1/]																	
25	8/10/83	82	126.0	216	5.94	0.28	28.2	0.06	25	51.7	1.3	408	43.9	-120.2	-14.54	---	-5.40
45	8/15/83	84	120.7	225	6.52	0.31	28.5	0.09	30	51.1	1.3	428	43.9	-122.0	-14.53	---	-5.08
216	8/16/83	71	109.3	214	5.89	0.30	23.5	0.03	19	53.4	1.3	407	45.1	-119.0	-14.48	---	-5.49

Table 3-1. — Chemical and isotopic analyses[1/] of water from wells and springs at Klamath Falls, Oregon (Chemical analyses in mg/kg except for Al, which is µg/kg; isotope analyses in permil relative to VSMOW [Gonfiantini, 1979]; tritium in Tritium Units) (Continued)

Sample	Date	Temperature (°C)	SiO_2	Na	K	Li	Ca	Mg	Al	Cl	F	SO_4	HCO_3	δD	$\delta^{18}O$	Tritium	$\delta^{18}O$ aq. SO_4
272	7/08/83	95[c]	119.8	230	7.44	0.38	34.8	0.05	30	55.0	1.3	419	41.4	-121.4	-14.63	0.12 ± 0.08	-5.07
272	7/09/83	98[c]	130.1	228	7.17	0.38	36.1	0.03	25	53.6	1.3	441	40.9	-122.1	-14.41	---	-5.62
272	8/24/83	—[c]	120.7	228	7.52	0.36	36.1	0.05	26	53.4	1.4	450	40.3	-120.8	-14.56	---	-5.14
304	8/17/83	47	97.3	177	4.59	0.28	20.5	0.05	10	38.4	1.2	294	90.3	-115.7	-13.96	---	-4.66
Earlier analyses of hot water wells [2/]																	
OIT #6	8/24/83	88	31?	195	3.9	---	24.2	<.1	---	58	1.45	400	44	---	---	---	---
	3/31/75	79	90	---	---	0.12	---	---	---	---	---	---	--	-118.7	-14.4	1.6 ± 0.5	-5.45
Assembly of God	8/27/76	87	110	213	4.6	---	28	0.1	---	54	---	393	61	-120.0	-14.75	---	---
Medo-Bel	1/24/55	81	81	213	4.2	---	23	0	---	54	1.2	403	32[d]	-119.4	-14.6	---	-5.45
J. E. Friesen	2/19/55	73	87	221	4.4	---	25	0	---	56	1.6	431	32[d]	---	---	---	---
L. Serruys	12/22/54	71	83	207	3.8	---	22	0	---	50	1.4	393	43[e]	---	---	---	---
Earlier analyses of cold water [2/]																	
Well																	
Oregon Water Corp	9/—/71	14	27	22	---	---	14.4	9.8	---	4.5	0.01	1.2	--	-109.4	-14.4	2.4 ± 0.5	---

Table 3-1. — Chemical and isotopic analyses[1/] of water from wells and springs at Klamath Falls, Oregon (Chemical analyses in mg/kg except for Al, which is μg/kg; isotope analyses in permil relative to VSMOW [Gonfiantini, 1979]; tritium in Tritium Units) (Continued)

Sample	Date	Temperature (°C)	SiO_2	Na	K	Li	Ca	Mg	Al	Cl	F	SO_4	HCO_3	δD	$\delta^{18}O$	Tritium		$\delta^{18}O$ aq. SO_4
Cold springs																		
Paul W. Sharp (spg.)	6/06/74	18	43	19	4.0	---	13	6.6	---	2.2	0	10	110[f]	---	---	---	---	---
Shell Rock (spg)	8/05/72	12	32?	5.9	1.5	---	8.1	3.9	---	<1	0.1	<2	50	-113.1	-15.35	---	---	---
	8/06/75	7	60	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Humming-bird Spg.	7/26/72	12	21?	7.0	1.2	---	10.6	8.2	---	1	0.1	<2	96	---	---	---	---	---
	8/06/75	10	42	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Barkley Spg.	7/26/72	11	18	8.0	1.2	---	9.2	6.5	---	1	0.1	<2	84	-115.0	-15.11	---	---	---
	4/02/75	11	---	---	---	<0.02	---	---	---	---	---	---	---	---	---	---	---	---
Neubert Spg.	8/05/72	17	24?	7.6	1.3	---	12.1	7.2	---	1	0.11	<2	94	---	---	---	---	---
	8/06/75	10	48	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---

1/ Janik and others (1984), tables 4-2, 4-3. New Ca data from A. White and A. Yee of Lawrence Berkeley Laboratory, Berkeley, Calif.; new Cl data from Analytical Chemistry Branch, U.S. Geological Survey, Menlo Park, Calif.

2/ Sammel (1980), tables 5, 8.

[a] City of Klamath Falls reports a constant temperature of 18°C.
[b] The value of 0.17 TU reported in Janik and others (1984) is a transposition error.
[c] Wellhead temperature reported (Benson and Solbau, 1984) to have remained constant at 100°C for entire pumping test.
[d] CO_3 = 8 [e] CO_3 = 4 [f] CO_3 = 1

Table 3-2. -- Reservoir temperatures calculated from geothermometers, in °C

Sample	Date	Measured T°C	Qtz 1/	Chalced. 2/	SiO_2 Amorph. 2/	Na/K 3/	Na-K-Ca 4/	SiO_2 mixing 5/	SO_4-H_2O 6/ $\delta^{18}O$	X(hot) 7/
Hot water wells sampled the month prior to 1983 pumping tests										
25	6/04/83	78	136	108	16	74	78	175	186	0.38
45	6/09/83	88	137	108	16	78	82	165	189	0.48
45A	6/09/83	80	170	146	47	87	89	271	188	0.25
65	6/14/83	66	127	98	8	68	70	168	187	0.32
76	6/07/83	78	132	103	12	79	81	165	189	0.41
110	6/21/83	91	137	108	16	76	81	162	194	0.51
216	6/29/83	72	149	123	28	79	85	221	191	0.26
259	6/08/83	56	131	103	12	74	76	201	187	0.20
304	6/08/83	56	125	96	6	68	78	180	186	0.23
Hot water wells sampled during 1983 pumping test										
25	8/10/83	82	151	124	29	81	83	206	192	0.34
45	8/15/83	84	148	121	27	84	86	196	186	0.37
216	8/16/83	71	142	115	21	81	86	203	194	0.28
272	7/08/83	95	148	121	26	91	86	182	184	0.47
272	8/09/83	98	153	126	31	89	84	189	198	0.47

Table 3-2. -- Reservoir temperatures calculated from geothermometers, in °C (Continued)

Sample	Date	Measured T°C	Qtz 1/	Chalced. 2/	SiO_2 Amorph. 2/	Na/K 3/	Na-K-Ca 4/	SiO_2 mixing 5/	SO_4-H_2O 6/ $\delta^{18}O$	X(hot) 7/
Hot water wells sampled during the 1983 pumping test (Continued)										
272	8/24/83	*	148	121	27	93	86	178	187	0.52
304	8/17/83	47	136	107	16	77	79	256	189	0.12
Earlier analyses of hot water wells										
OIT #6	3/31/75	79	132	103	12	62	71	154	195	0.51
Assembly of God	8-27-76	87	143	115	22	66	74	180	---	0.43
Medo-Bel	1/24/55	81	126	96	7	61	75	148	191	0.48
J.E. Friesen	2/19/55	73	130	101	10	61	76	165	---	0.37
L. Serruys	12/22/54	71	127	98	8	57	73	161	---	0.37

* 100°C assumed for calculations.

1/ Fournier and Potter, 1982. Assumed conductive cooling.

2/ Fournier and Rowe, 1966; Fournier, 1977. Assumed conductive cooling.

3/ White 1970.

4/ Fournier and Truesdell, 1973.

5/ Mixing model based on quartz equilibria and a cold-water component having a SiO_2 concentration of 45 mg/kg and a temperature of 20°C (Truesdell and Fournier, 1977; Fournier and Potter, 1982).

6/ McKenzie and Truesdell, 1977. Assumed conductive cooling.

7/ Fraction of hot end-member water in the SiO_2 mixing model.

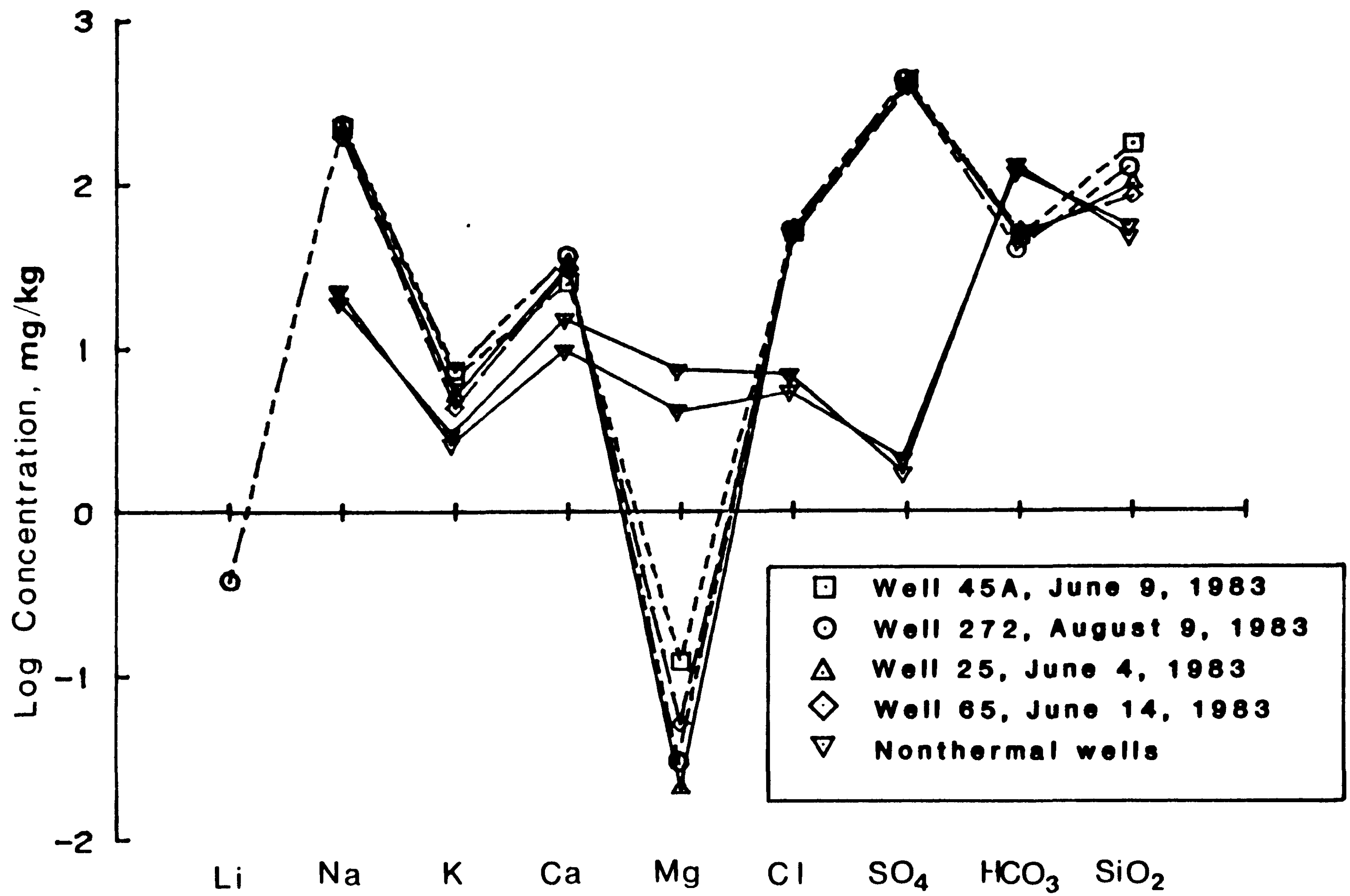

Figure 3-1. -- Schoeller plot showing comparison of hot and cold well-water compositions.

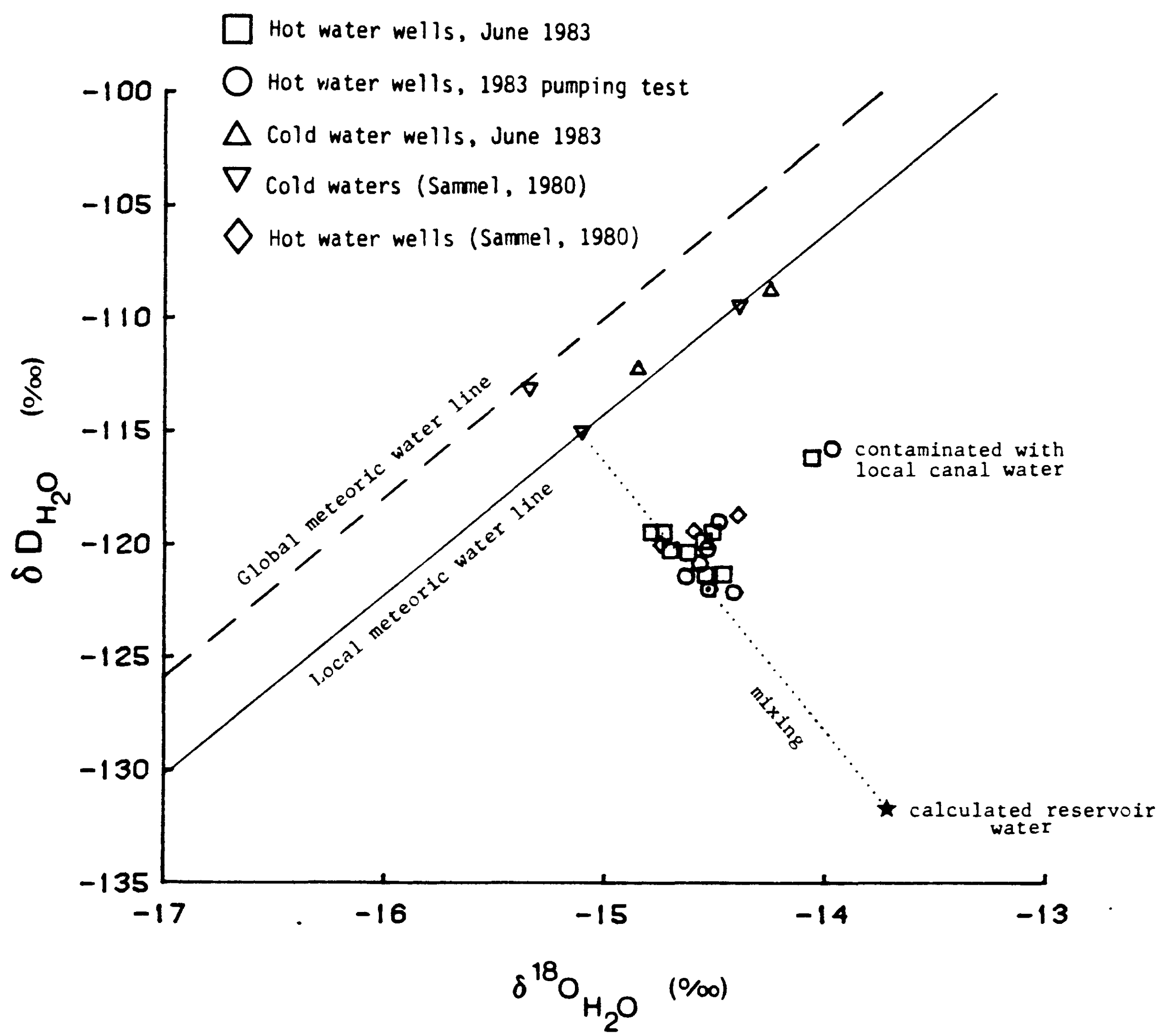

Figure 3-2. -- Hydrogen- and oxygen-isotope compositions of thermal and nonthermal waters, and calculated composition of deep reservoir water.

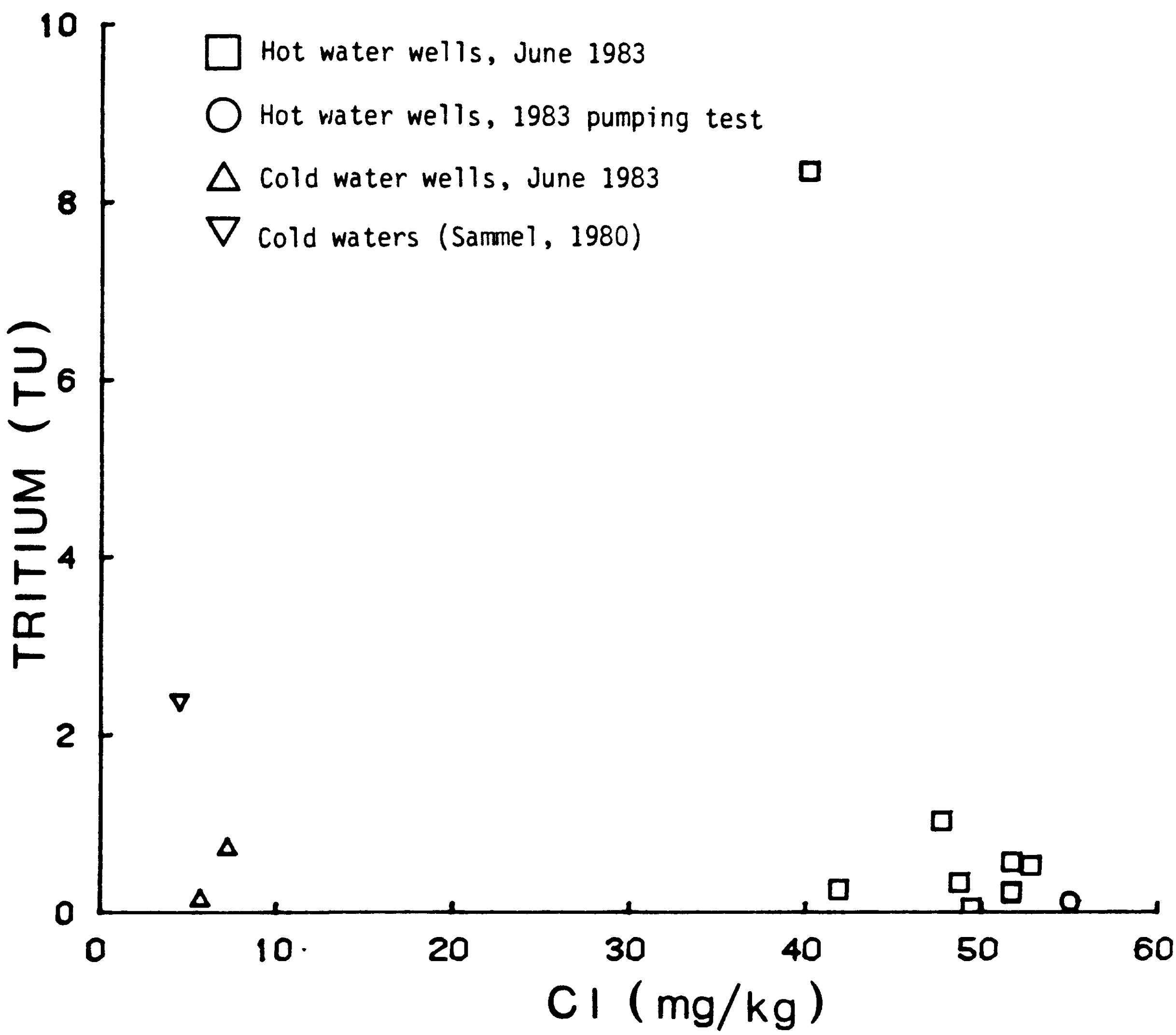

Figure 3-3a. — Tritium versus Cl concentrations of thermal and nonthermal waters.

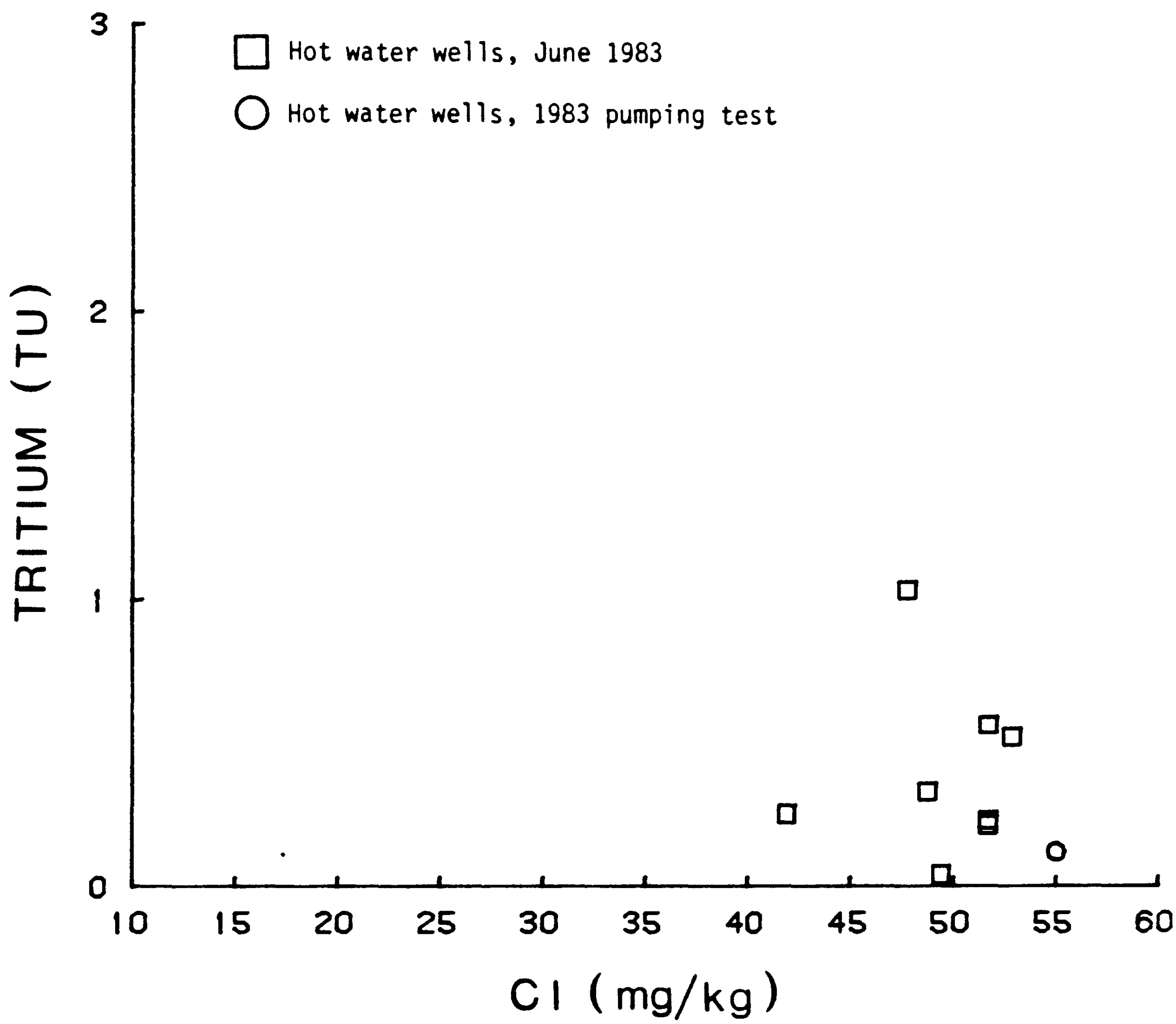

Figure 3-3b. -- Tritium versus Cl concentrations of thermal waters, using an expanded scale to show a mixing trend.

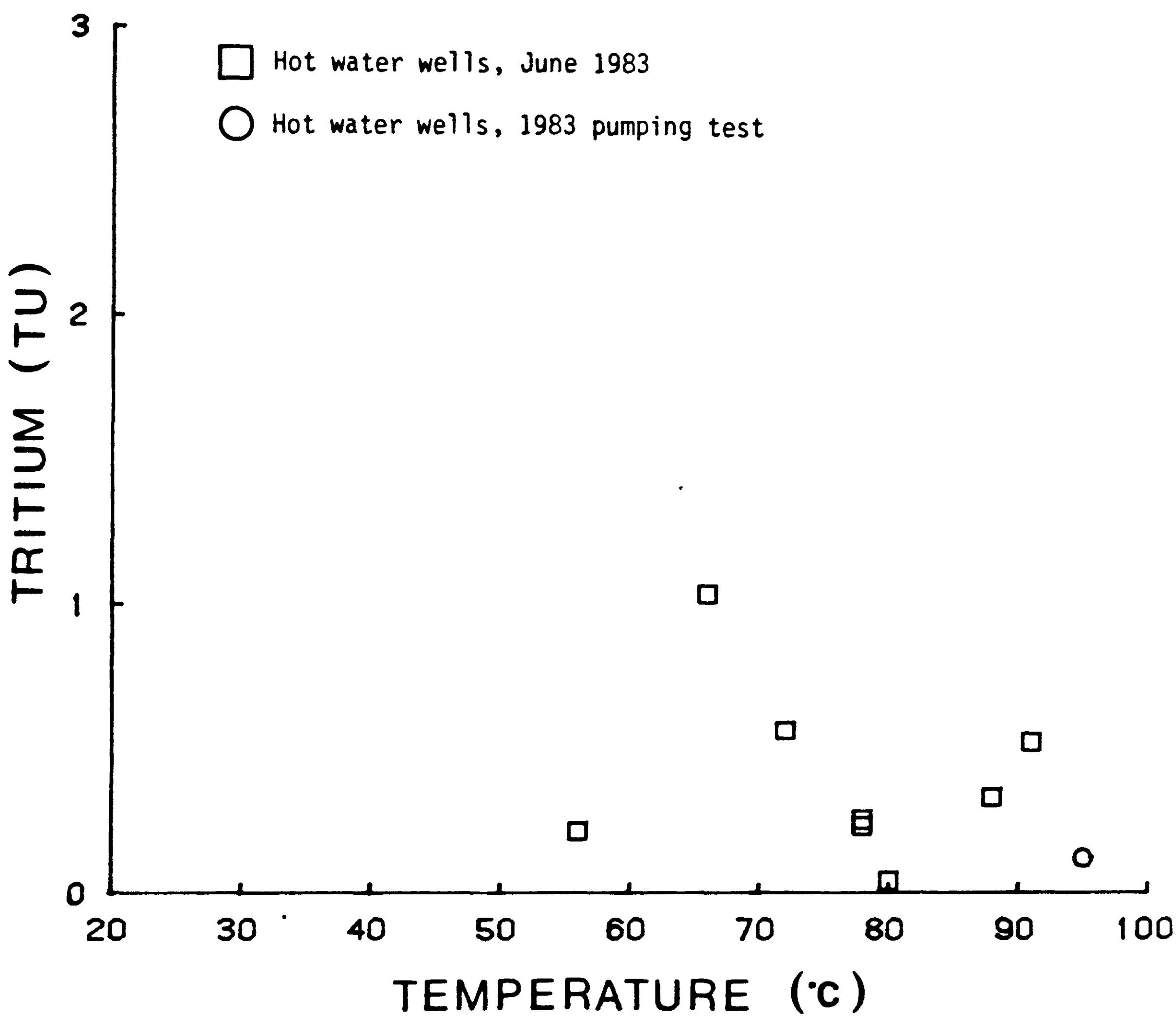

Figure 3-4. — Tritium versus temperature of hot well waters.

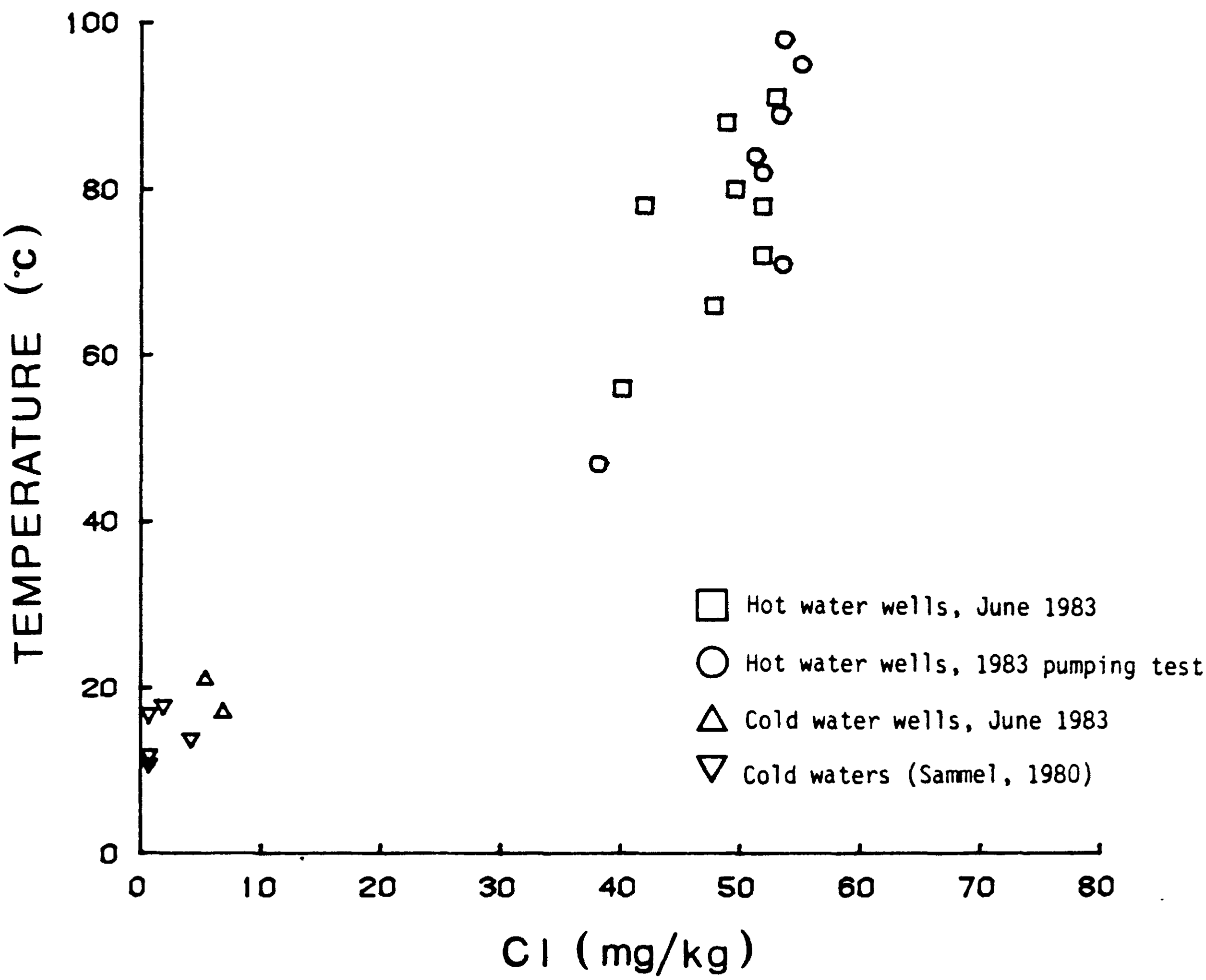

Figure 3-5. — Temperature versus Cl concentrations of hot and cold well waters, showing a mixing trend.

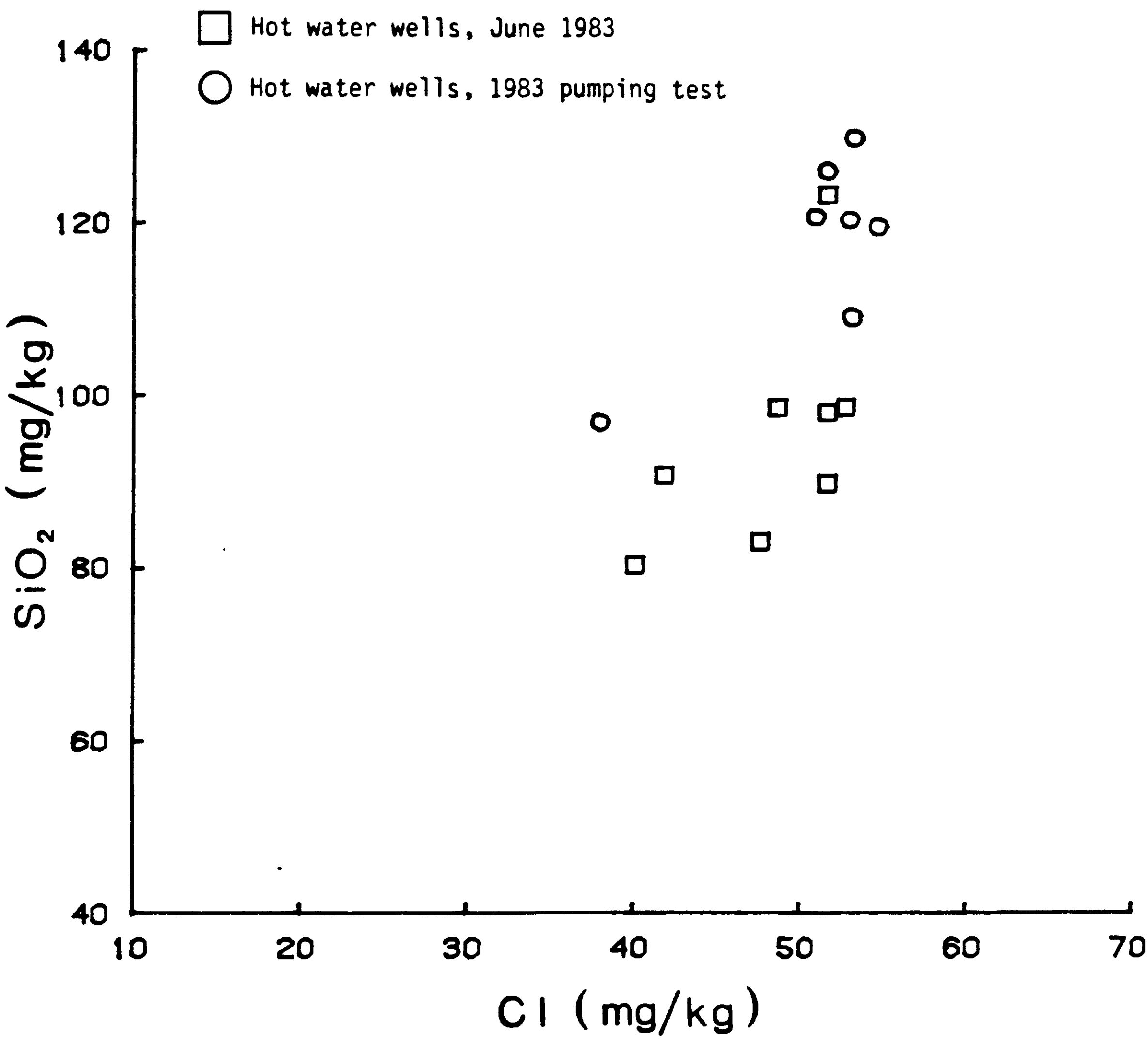

Figure 3-6. — SiO_2 versus Cl concentrations of hot well waters.

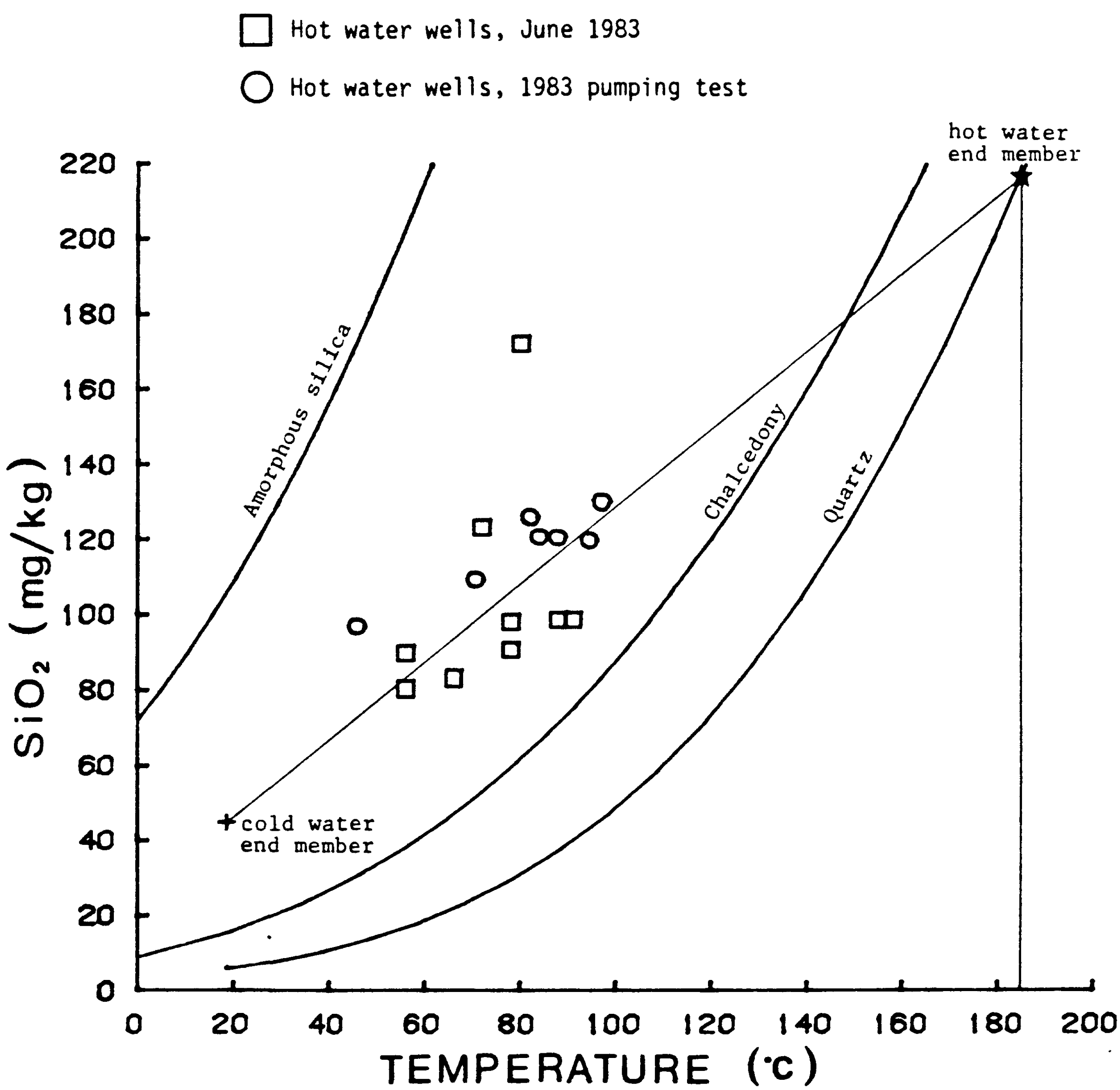

Figure 3-7. -- SiO_2 concentration versus temperature of thermal waters and their relation to SiO_2 solubility curves.

CHAPTER 4. INTERWELL TRACER TESTING IN KLAMATH FALLS

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Introduction

The need to inject used geothermal waters is twofold. First, when water is pumped from a geothermal reservoir or aquifer, the water level will drop so that the reservoir pressure becomes less. As more and more fluids are produced, the water has to be pumped from an increasing depth. However, if all fluids pumped from an aquifer are injected, the water level is unlikely to fall as rapidly, so production can be maintained for a longer time without additional pumping capacity. Second, the cooled fluids must somehow be disposed of. Because most geothermal fluids contain more dissolved matter than rain and drinking water, they cannot easily be disposed of at the surface into lakes and rivers. For environmental reasons therefore, the preferred method of disposal is injection.

Fluid injection is also important in situations where downhole heat exchangers and subsurface pumping are used together. Without injection, the water level will drop with time, and if the aquifer is small in comparison to the utilization, the downhole exchangers will eventually become dry, unless there is fluid injection to maintain the water level.

Although the injection of spent fluids can support reservoir pressure and reduce water-level decline, the method has the drawback of eventually causing a decrease in the temperature of the hot water being pumped for use. This problem is of special concern because of the many fractures that characterize most geothermal reservoirs. The fractures act as pipes between the injection and production wells, allowing the cooled injection water to travel long distances in a short time. However, as the injected water travels through the fractures, heat transfer from the formation rock to the fluid tends to diminish the negative effects of cold-water breakthrough in production wells.

The main purpose of the tracer tests in Klamath Falls was to obtain data on the injection behavior of geothermal reservoirs. Because methods of interpreting tracer tests in fractured formations are still being developed, the Klamath Falls tests were designed to obtain data that would aid in the development of such methods. In combination with the results of pumping tests described elsewhere in this report, the tracer tests also contribute to the understanding of the geothermal resources in aquifers beneath Klamath Falls.

The technology and world-wide experience of injection in geothermal reservoirs have been discussed by Horne (1982a, 1982b, 1984). Results from the doublet tracer testing in Klamath Falls have been reported by Gudmundsson and others (1983).

Injection Fundamentals

Thermal Energy

In geothermal reservoirs more of the thermal energy is stored in the solid rock than in the liquid water. This can be illustrated by calculating the thermal energy contained in each cubic meter of a typical reservoir. The following properties are used in such calculations:

ϕ = Formation porosity (fraction)
ρ_w = Water density (kg/m^3)
C_w = Water heat capacity (kJ/kg °C)
ρ_r = Rock density (kg/m^3)
C_r = Rock heat capacity (kJ/kg °C)

These properties can be combined to give the ratio of the heat contained in the water only to the heat contained in the water and rock formation together:

$$\frac{\phi \rho_w C_w}{\phi \rho_w C_w + (1 - \phi) \rho_r C_r} \qquad (4\text{-}1)$$

The following values are representative of low-temperature geothermal reservoirs: $\phi = 0.1$ (=10 percent), $\rho_w = 1000$ (kg/m^3), $C_w = 4$ (kJ/kg °C), $\rho_r = 2000$ (kg/m^3), $C_r = 1$(kJ/kg °C). Using these values for the reservoir properties, we calculate that the water contains about 18 percent of the total thermal energy of each cubic meter of the reservoir and that the rock formation contains about 82 percent of the thermal energy. This means that if all the water were pumped from a reservoir with the above properties, only 18 percent of the total thermal energy would be recovered and 82 percent would remain in the rock formation. Figure 4-1 illustrates the same point for other porosity values.

Tracer and Thermal Fronts

When the water is injected into a reservoir containing a production well, it will displace some of the original water between the two wells. The volume of water displaced is called the swept volume of the reservoir. If the temperature of the injected water is lower than the reservoir temperature, the injected water will be heated. In situations where the injected water flows slowly over a long distance before reaching the production well, its temperature may increase so as to approach the reservoir temperature. Therefore, temperature measurements at the production well will not show that fluid breakthrough has occurred. However, if the injected fluid contains a chemical tracer that can be measured at the production well, then fluid breakthrough can be demonstrated. After injecting cold water for a long time into a geothermal reservoir, the rock may have cooled so much that it no longer can heat the water before it reaches the production well. Therefore, the temperature measured at the production well will decrease with time.

The tracer and thermal velocities can be used to express how rapidly tracer concentrations and decreasing temperatures will be observed in production wells. In most reservoir situations, the thermal velocity is much slower than the tracer velocity. The ratio of the thermal velocity to the tracer velocity is given by equation 4-1 above. For the example illustrated above, the velocity ratio is 1:5.6. This indicates that the thermal velocity is 5.6 times lower than the tracer velocity. For example, if this velocity ratio is applied to a situation where tracer breakthrough occurs in a production well 5 days after injection, thermal breakthrough in the production well would occur after 28 days. The velocity ratio in specific reservoirs depends greatly on the formation porosity, as illustrated in figure 4-1. Nevertheless, it is clear that measurements of tracer breakthroughs in production wells may be indicative of subsequent thermal breakthroughs.

Well Placement

Critera for locating injection wells in relation to production wells in the design of geothermal fluid-production schemes should include consideration of the effects of both pressure stabilization and thermal breakthrough. Choices to be made include locating injection wells near the center of the production area or close to periphery of the resource, and drilling injection wells to the same depth, deeper, or shallower than the production wells. The question of well placement has been discussed by James (1979).

Tracer Analysis

A tracer breakthrough curve is a kind of travel log that shows concentration of a tracer with time. It provides a record of what happens underground when a fluid flows between two or more wells. The interpretation of a tracer breakthrough curve requires knowing what physical and chemical principles apply. In geothermal reservoirs these principles are little known and field data are limited.

The most common use of tracers in geothermal reservoirs is to inject a slug of the tracer into an injection well and monitor surrounding producing wells for return of the tracer. Following injection of the tracer slug, full-scale injection and production is usually begun in order to provide as realistic a setting as possible. With reduced flowrates there is always a danger that the tracer will not be returned and the value of the test will be lost. The production wells are monitored for the tracer at frequent intervals in the first hours and days after the tracer is injected, and less frequently after the detected tracer concentration has passed through a maximum and started to decline.

Tracer analysis provides four main pieces of information. The first detection of tracer and the arrival of the peak concentration indicate the speed of movement ot the water through the system. The first tracer to arrive has probably dispersed ahead of the main tracer slug, whereas the concentration peak moves with the mean speed of the flow. However, in many cases the first detection and peak concentration occur within a very short time. Such a condition is an indicator of subsequent difficulties with premature thermal breakthrough. Rapid tracer movement implies a high degree of fracturing or high permeability in the reservoir and suggests that the swept volume of reservoir rock will be small. Field experience indicates a correlation between rapid tracer returns and subsequent degradation in the temperature of produced fluids (Horne, 1982a).

A second parameter of interest in the tracer-return information is the total tracer recovery. A production well that receives more of the injected water than others is more likely to suffer temperature decline as a result. Again, a correlation between large tracer returns and subsequent degradation in production well performance has been observed in the field. Quantifying total recovery can be difficult because the tracer may be lost within the reservoir by chemical reaction or adsorption. Nevertheless, the relative recoveries in several production wells can be useful for comparison between wells.

A third use for the tracer-return history is the analysis of long-term equilibrium tracer concentration. In cases where the produced water is continuously injected, the tracer is repeatedly produced and injected. Provided the tracer is not retained or destroyed to a large extent within the reservoir, concentrations in the produced fluid will gradually reach an equilibrium value higher than the original background concentration. At this stage, it is possible to estimate the volume of reservoir fluid throughout which the tracer has been dissolved. If this calculation is performed using the assumption that the tracer remains in the circulating fluid, then an upper bound on the swept volume of the reservoir can be estimated. This could be useful for evaluating the field-wide probability of thermal degradation. However, if the tracer is retained in the reservoir rock, the swept volume would be seriously overestimated.

A fourth use of the tracer-return data is to analyze the shape of the concentration/time profile (breakthrough curve). This procedure is still under development but shows possibilities for the estimation of fracture characteristics. An estimate of fracture aperture is useful for calculating the rate of local thermal depletion along the flow path.

Breakthrough Time

The time it takes a tracer to travel from an injection well to a production well is the breakthrough time. This time can be estimated for common flow geometries in homogeneous reservoirs, for example radial and doublet flow. When the fluid injected into a well moves radially away from the wellbore, the breakthrough time, t_r, at distance r is given in the expression:

$$t_r = \frac{\pi r^2 \phi h}{Q} \tag{4-2}$$

where ϕ and h are the reservoir porosity and thickness, respectively, and Q the volumetric flowrate. A doublet well configuration acts as a source-sink system. If x is the doublet spacing, the breakthrough time, t_x, will be one-third that of a radial system, as shown by Gover and others (1970) and Klett and others (1981):

$$t_x = \frac{\pi x^2 \phi h}{3Q} \tag{4-3}$$

For the simplest case of flow in a planar fracture of the thickness δ , these same equations can be applied by replacing ϕh by δ. In tracer tests the observed breakthrough time will be somewhat less than that calculated from equations 4-2 and 4-3 because of the effects of hydrodynamic dispersion.

Doublet Tracer Testing

System Description

Down-hole heat exchangers are widely used in the 450 or more geothermal wells that have been drilled in Klamath Falls. Several doublet systems, characterized by having one production well and one injection well, are also used for space heating in the city. The hot water is pumped from the production well and then cooled in heat exchangers before disposal by injection. Fresh circulation water flows through the heat exchangers just as in the down-hole exchangers. The wells in the doublet systems in Klamath Falls tend to be of similar depth and design. They are closely spaced and there are no wells between them; hence the doublet name. There are 4 doublet systems in the city: Klamath Union High School (KUHS), Mazama High School, YMCA-Center, and Mills School.

The doublet system at the Klamath Union High School was selected for the first tracer testing experiment because of its high constant flowrate and location. The school is near the hot-well area and close to the County Museum, which is the site of the main injection well of the Klamath Falls District Heating System (fig. 4-2).

The KUHS doublet system was started in the early 1960's. The production well is 257 ft deep and perforated for 25 ft near the bottom. The injection well is 240 ft deep and cased to 120 ft. The wells are spaced at a distance of about 250 ft. The system is turned on in the fall when school starts. The down-hole pump in the production well pumps at a constant rate which is estimated by the pump installer to be about 320 gal/min based on the pump specifications, water levels, and wellhead pressure measurements. Part of this flow, about 15 gal/min, is diverted to two buildings at the site.

At the start of the heating season the temperature of the produced water is 165°F (74°C) but it cools by 5°F in 3 to 10 days depending on the heat load. Once this initial cooling has occurred, the water temperature remains constant until spring. The following fall the temperature is back to 165°F at startup. During the heating season, the geothermal water enters the high shcool heat exchangers at about 160°F (71°C) and leaves typically at 152°F (67°C). At 305 gal/min this corresponds to 1.1×10^6 BTU/hr or 323 kW of thermal power.

Test Program

Iodide (I^-) has been determined to be a useful chemical tracer for geothermal systems. It has been used in several injection tests in Japan in the form of potassium iodide (Horne, 1982a). The background concentration of iodide in thermal waters tends to be low. This means that the amount of iodide required for injection tests is reasonable in comparison to other similar chemicals. Chemicals are not useful as tracers if they become stuck (adsorbed and absorbed) to the formation or degrade when flowing through an aquifer formation. Iodide compounds are highly soluble in water and are unlikely to interact much with reservoir rock formations.

Although the background concentration of iodide in geothermal waters tends to be low, the cost of iodide chemicals is high. The quantity of chemical tracer needed for injection testing depends not only on the natural background concentration, but also on the method used to analyze for the tracer. The better the method of analysis, the less tracer is required. Environmental considerations are very important in tracer testing. The chemical selected must be safe to handle and used in concentrations below recommended water-quality standards. Consideration of the above factors (background concentration, formation interaction, availability and cost, analysis methods and environmental aspects) indicates that iodide chemicals are useful in geothermal tracer testing. Thompson (1980) has discussed the selection of common ground-water tracers.

Fluorescent dyes are commonly used in ground-water tracer studies (Smart and Leidlaw, 1977). They are easily analysed in low concentrations using simple methods. The quantity of fluorescent dyes required in tracer testing is therefore small. The commonly used dyes are readily available and reasonable in cost. They are environmentally acceptable for two main reasons: they require low concentrations and they degrade with time. The degradation may pose a problem in geothermal tracer testing. When heated, fluorescent dyes may break down (thermal degradation) and lose their fluorescent properties. Although the nature and extent of this loss are unknown, fluorescent tracers have been used with success in geothermal reservoirs. This success may stem from the fact that fluid breakthroughs are rapid in fractured formations and consquently the dye tracers are not subjected to high temperatures for a long time.

Both chemical and fluorescent tracers were used in the KUHS doublet test. Potassium iodide (KI) was selected as the chemical tracer and rhodamine WT and fluorescein as the dye tracers. The purpose of using several tracers was to investigate their relative merits in geothermal applications. A fluorometer was used to measure the concentration of the dyes and an ion specific electrode to measure the iodide. The advantage of both methods is that they are easily carried out in the field. Fluorescent dyes can be detected in water in concentrations below 1 μg/kg and halides such as iodide can be detected at about 1 mg/kg using ion specific electrodes. The literature was searched for data on possible health risks associated with the tracer material. It was found that no ill effects would be likely to occur upon drinking the geothermal water during the tracer test. Permits were obtained from the Department of Environmental Quality (State of Oregon) and the Department of Health Services (Klamath County) to inject 90 g and 900 g of rhodamine WT and fluorescein, respectively, and 690 kg of potassium iodide. The amounts actually injected were much less than the amounts permitted.

The KUHS doublet tracer test was carried out in May and June of 1983. The tracers were injected at the wellhead of the injection well. The wellhead piping is such that the tracers were mixed immediately with the downflowing water. One pound (450 g) each of rhodamine WT and fluorescein were mixed in 100 gallons of geothermal water. The rhodamine WT was in the form of a liquid 20 percent active so the 1 pound solution contained 90 g of the red-pink dye. Fluorescein comes in dry powder form and is greenish when dissolved in water. All of the fluorescein was considered active. Five hundred pounds (227 kg) of potassium iodide was mixed in 150 gallons of geothermal water; it is easily soluble and colorless. The dyes were injected first, requiring about 15 minutes for the injection. The potassium iodide was injected about an hour later, the injection taking about 20 minutes.

An automatic sampling apparatus was set up at the KUHS production well and programmed to fill one bottle every half hour. Five other wells were sampled by hand during the tracer test: Balsiger, 260 ft deep; Medo-Bel, 765 ft deep; Eccles, 787 ft deep; Friesen, 563 ft deep; and Garrison, 240 ft deep (fig. 4-2). Samples were collected from these wells every hour at first and then less frequently. The flowrate and temperature of the Medo-Bel well were measured at 75 gal/min and 180°F. The flowrate of the other wells had to be guessed: Balsiger, 30 gal/min; Eccles and Friesen, 20 gal/min each; and Garrison, 10 gal/min. Other wells in the area were not pumped at the time of the testing.

The concentration of the dyes was measured in a fluorometer. It was discovered during the test that injecting the dyes at the same time was a mistake. Although recommended lamps and filters were used in the fluorometer, there was considerable interference between the two dyes. A reading on the fluorometer could not be assigned to one dye only so the values obtained were semi-quantitative. Also, a mixture of the two dyes showed less color and fluorescence than expected. The dye concentrations measured with time in the KUHS doublet test have been reported by Gudmundsson and others (1983) and will not be repeated here. Instead, the data analysis will be based on the more accurate iodide measurement.

Breakthrough Curves

Tracer breakthrough was observed in the KUHS doublet production well 2 to 3 hours after the 20-minute tracer slug injection. The concentration of potassium iodide with time in the production well is shown in figure 4-3. The maximum tracer concentration was reached in 5 to 6 hours after the end of the tracer injection. After that the concentration fell rapidly at first and then more slowly. The dye tracers showed the same breakthrough behavior as the potassium iodide.

The 227 kg of potassium iodide were injected in 20 minutes. For a doublet flowrate of 305 gal/min (19.2 L/s), this corresponds to the tracer slug having a concentration of about 9,820 mg/kg. The maximum tracer concentration measured in the production well was about 60 mg/kg, or two orders of magnitude lower.

Fluid recirculation must be considered in the doublet tracer test. The high school is about 600 ft away from the doublet system. The hot water is pumped to the school in a 6-inch pipeline and then passed through 13 shell-and-tube heat exchangers which are connected in a mixed series/parallel arrangement. The geothermal water is cooled in the heat exchangers and returned to the injection well. The travel time of the geothermal water from the production well to the injection well depends on the flowrate and the volume of the piping and heat exchangers. It takes about 3 minutes for water pumped at 305 gal/min to travel 600 ft in a pipeline 6 inches in diameter. Assuming that each of the 13 heat exchangers has the same volume as the 600 ft of pipe, it will take 45 minutes for the geothermal water to travel between the wellheads at the surface.

For a tracer breakthrough time of 2 hours and 30 minutes, and assuming a surface travel time of 45 minutes, a second breakthrough would be expected at about 5 hours and 45 minutes, passing through a maximum between 10 and 13 hours. However, since the initial tracer pulse was diluted by two orders of magnitude, the second tracer pulse is unlikely to show much effect on the shape of the tracer concentration curve.

The flow pattern in the KUHS doublet system is affected by the other pumped wells in the area. The largest of these is the Medo-Bel well, flowing 75 gal/min. it is about 450 ft away from the injection well, in the opposite direction to the production well. The total flowrate of the other 4 wells in the area added up to about 80 gal/min. The nearby pumping is therefore about one half that of the KUHS production well. The hydraulic gradient of the hot well is superimposed on the pumping gradient associated with the wells sampled in the tracer study. The gradient is about 0.5 percent and perpendicular to a line between the injection and production wells. The KUHS doublet system is not an isolated system in the aquifer.

The tracer breakthrough curve for potassium iodide in the Medo-Bel well is shown in figure 4-4. The breakthrough occurred 26 to 27 hours after injection was completed in the doublet injection well. The tracer concentration reached a maximum value less that 1.5 mg/kg, which is an order of magnitude less than the maximum value in the doublet system. This occurred after 180 to 200 hours. After that the tracer concentration decreased slowly but steadily. Tracer returns were not detected in the other wells monitored in the area.

Interpretation

Three of the four main tracer-analysis methods already discussed are useful in field situations where proven reservoir flow models are not available. In geothermal reservoirs, the interpretation may show what relative injection effects are to be expected in different wells.

The peak tracer concentration in interwell tracer testing moves through the reservoir at the average fluid velocity. Using 5 to 6 hours as the time of peak concentration in the KUHS production well, the calculated average flow velocity between the doublet wells is 13 to 15 m/hr. The same kind of calculation for the Medo-Bel well gives a tracer velocity of 0.7 to

0.8 m/hr. These values are within the range of tracer velocities measured in geothermal fields world-wide (Horne, 1984).

The arrival time of the peak concentration can also be used to estimate the porosity-thickness value used in equations 4-2 and 4-3. Taking t_r and t_x as the peak arrival times for radial and doublet systems, respectively, the porosity-thickness becomes:

$$(\rho\phi\, h)_r = 2 \text{ cm},$$

$$(\rho\phi\, h)_x = 6 \text{ cm}.$$

If fracture flow is assumed, the above values represent the fracture width, δ.

The amount of tracer recovered in production wells indicates the connectivity to injection wells. Production wells that receive more tracer than others are more likely to suffer thermal drawdown when cold fluids are being injected. The amount of tracer recovered can be determined by measuring the area under the breakthrough curve. For the KUHS production well, the amount recovered after 4 to 5, 8 to 9, and 100 to 110 hours was 15, 25, and 122 kg, respectively. These correspond to 7, 11, and 54 percent of the total amount of potassium iodide injected. Because of recirculation, the tracer recovery after the first 10 to 13 hours becomes difficult to interpret. Nevertheless, the breakthrough curve does show that 10 to 20 percent of the total tracer material injected was recovered in the first day of the test. This suggests that the doublet is not an isolated system in the geothermal aquifer.

Because tracer material is lost from the doublet recirculation system, the tracer concentration does not reach equilibrium concentration with time. The swept reservoir volume, therefore, cannot be determined.

The shape of the tracer breakthrough curve can give information about the nature of the flow between the injection and production wells. The interpretation method is based on having a reservoir flow model. Such models are now being developed for geothermal reservoirs (Fossum and Horne, 1982; Jensen and Horne, 1983).

Injection Tracer Testing

Test Description

The doublet tracer test was carried out a few months ahead of the main aquifer pumping-injection test. Among the reasons for carrying out the doublet test was the need to develop tracer testing techniques appropriate for geothermal resources, particularly those of low to moderate temperature. This goal was reached with respect to tracer selection and measurement techniques. Rhodamine WT was found to be environmentally acceptable, easily measured, and low in cost. Therefore, this dye was selected for use in the injection tracer test. Potassium iodide was also found to be environmentally acceptable, but the higher concentration required and greater cost rule it out in situations where fluorescent dyes can be used.

In the main aquifer pumping-injection test, the production well (CW-1) was pumped for 3 weeks without injection. The pumped water was then injected into the County Museum well for about a month. This well is about 3,000 ft away from the production well and down the hydraulic gradient. The natural hydraulic gradient between the two wells is about 0.5 percent and the total head difference is about 15 ft. During the aquifer pumping-injection test the gradient was expected to be reversed so that flow would occur from the site of the injection well in the direction of the production well.

In a doublet system, there are no wells between the production and injection wells. This is not the case for the production and injection wells used in the aquifer pumping-injection test. The area between CW-1 and the County Museum well is heavily exploited. In this situation, the flow path can be traced by sampling the wells between the production and injection well. Most of the wells have down-hole heat exchangers, but there are a few that are pumped at low flowrates or have artesian flow. The flowing wells identified for water sampling for tracer analysis were: Friesen (Main St., laundry), Olympic (E. Main St., apartments), Butler (Esplanade St., residence), Division and Oak Sts. (residence), and Medical Clinic (Main St.)

The aquifer pumping test was started on July 5, 1983. The production well was pumped at 720 gal/min and the 100°C water discharged into the USBR irrigation canal that runs across the geothermal field. (See fig. 4-2.) After 3 weeks of pumping the injection part of the test was started. On July 26 at 10:11 a.m. the geothermal water flow was diverted into the County Museum injection well. (See fig. 5-5 for locations.) The aquifer production-injection test was terminated 4 weeks later on August 24 at 5:35 p.m. The injection flowrate was 40 to 42 kg/s during the 4 week period.

The rhodamine WT tracer was injected into the County Museum well on July 27 from 10:14 a.m. to 10:19 a.m. The method of injection was the same as used in the doublet-tracer test. Two 25-pound drums of rhodamine WT solution were dissolved in 33 gallons of water. The dye solution was 20 percent active so that each pound contained 90 g of rhodamine WT. Therefore, the total mass of rhodamine WT was 4.55 kg and its concentration in the 5-minute injection slug was about 360 mg/kg.

An automatic sampling apparatus was installed at the production well. The 5 flowing wells listed above were sampled by hand. Samples were collected every 1 to 2 hours and analyzed for rhodamine WT the same or following day.

After a few days of injection, the Medo-Bel well began to flow. This happened sometime between July 29 and August 1, when it was first sampled for tracer analysis. With time, other wells were added for water sampling: Fire Station, August 1; Spires and Mest (garage), September 9; Jones (garage), September 23. One or two samples were taken from a few other wells.

Breakthrough Curves

The rhodamine WT tracer was detected in several wells during the injection tracer test. Of the wells that were sampled from the start of the injection test, only the Friesen well showed tracer breakthrough. This occurred after about 16 days of injection. The Friesen breakthough curve is shown in figure 4-5 in terms of $\mu g/kg$ of rhodamine WT tracer. The Friesen well is located about 1,000 ft north-east of the injection well. It is pumped at about 20 gal/min with a water temperature of 78°C.

Tracer breakthrough was evident from the start of flowing of the Medo-Bel well. The tracer concentration with time is shown in figure 4-6. The well produces water at about 98°C when pumped. At the same time that sampling was started from the Medo-Bel well, samples were also collected from the Fire Station well which is 130 ft from the County Museum injection well. Tracer breakthrough in the Fire Station well was also evident from the start of sampling. The tracer concentrations measured were about double those shown for the Medo-Bel well and decreased similarly with time. The curve did not show a maximum value.

Rhodamine WT was detected in at least two additional wells: Spires and Mest and Jones. The Spires and Mest well is near the Friesen well about 1,000 ft north-east from the injection well. Tracer breakthrough occurred before September 9, but it was not possible to determine when maximum tracer concentration occurred. The highest tracer concentration was about 40 μg/kg, indicating a stronger response than in the Friesen and Medo-Bel wells. The well flowed irregularly during the injection tracer test. The Jones well is 1,120 ft to the south-east of the County Museum well, in the opposite direction to that of the Medo-Bel well. The first sample (September 23) showed no tracer but the second sample (October 9) showed about 12 μg/kg. After October 9, the tracer concentration decreased with time.

No tracer was detected in the remainder of the wells sampled: Production, Medical, Olympic, Butler, and Division. These wells are located at greater distances from the injection well than the wells where the dye tracer was detected.

Interpretaton

The four basic tracer analysis methods discussed in other sections of this chapter concern: (1) the average fluid velocity in the aquifer (2) the amount of tracer material recovered (3) the equilibrium tracer concentration in the aquifer swept volume and (4) the shape of the breakthrough curve. The first of these can be applied to some of the injection tracer testing data, the other methods to a lesser extent.

To determine the amount of tracer recovered in each well, it is necessary to know the flowrate. This was known only (approximately) for the Friesen well. The well was pumped at about 20 gal/min, which represents 2.8 percent of the total flowrate injected at the County Museum well. Integrating the area under the breakthrough curve in figure 4-5, the mass recovered was estimated to be 25 to 35 g. This represents 0.6 to 0.8 percent of the total mass of rhodamine WT tracer injected. The flowrate for the other wells was not known.

The concept of equilibrium concentration cannot be appplied to the injection tracer testing data. Analysis of the shape of the tracer breakthrough curves requires a fluid-flow model of the aquifer. This aspect of the analysis is still in progress and will be reported at a later time.

The average fluid velocity in the aquifer can be estimated for 3 of the wells where tracer breakthrough occurred. The results are shown in table 4-1. The fluid velocity from the injection well to the Medo-Bel well was about 70 ft/day, the Friesen well about 30 ft/day and the Jones well 15 to 20 ft/day. The corresponding values estimated in the doublet tracer test were 1,000 to 1,200 ft/day between the injection and production wells and 55 to 65 ft/day between the injection well and the Medo-Bel well. The tracer testing data show a correlation between average tracer velocity and well spacing, the tracer velocity being inversely proportional to well spacing. A correlation of this type would be expected for radial flow away from an injection well. The exception to this observation is the high tracer velocity between the injection and the production wells in the KUHS doublet test.

If radial flow is assumed from the County Museum injection well to the Friesen well, equation 4-2 can be used to determine the porosity-thickness product, h. The maximum tracer concentration was measured about 36 days after the injection started. Taking the distance between the wells as 1,000 ft, the porosity-thickness is calculated to be 0.6 ft. If the aquifer porosity is assumed to be 0.1, the effective aquifer thickness is 6 ft. The ratio of thermal velocity to tracer velocity is typically 1:5.6 in geothermal aquifers. Taking the tracer breakthrough time as 36 days, thermal breakthrough in the Friesen well would be expected in about 200 days or 6 to 7 months. The magnitude of the thermal effect would depend partly on the temperature difference between the injected fluid temperature and the aquifer temperature between the County Museum and the Friesen wells and partly on the nature of the fracture system in the aquifer rocks.

Concluding Remarks

The problem of geothermal fluid injection described in this chapter, concerns how best to dispose of spent fluids and maintain reservoir pressure in geothermal developments without rapid and excessive cool-down of the fluids produced. Tracer testing holds promise as a means of providing some answers to this problem. The method traces fluid flow in the reservoir, which is related to subsequent cooling of the fluids produced. Tracer testing methods and interpretive techniques for geothermal reservoirs are still in the development stage. The interwell tracer tests in Klamath Falls were carried out to provide field data that would be useful in developing interpretive methods for fractured geothermal reservoirs as well as to compliment the aquifer production-injection testing of the Klamath Falls resource.

Traditional pressure-transient tests show the time behavior of aquifers and reservoirs when subjected to a change in production. They are used to determine the flow properties of aquifers and their ability to store fluids. However, they cannot show the movement of fluids in fractured reservoirs. Because geothermal reservoirs tend to be highly fractured, tracer testing has become an important tool in the evaluation of geothermal aquifers and reservoirs.

The main result of the interwell tracer tests in Klamath Falls is the quantification of fluid velocities in the reservoir. The highest velocity was that measured in the doublet tracer test, 13 to 15 m/hr. Other fluid velocities were much lower, 0.7 to 0.8 m/hr. from the doublet injection well to the Medo-Bel well. In the aquifer production-injection test, the fluid velocities were in the range 0.2 to 0.9 m/hr. The fact that tracers were recovered in the tests demonstrates that injected fluids migrate with time to production wells. For fluid velocities below about 1 m/hr there appears to be an inverse relationship to well spacing, as would be expected for radial flow away from injection wells.

The relationship between tracer and thermal velocities is important in the evaluation and design of injection schemes in geothermal reservoirs. However, the results obtained in the KUHS doublet test indicate that additional methods of analysis are needed. According to theory, because of the rapid returns, the production well should have cooled down long ago to temperatures not useful for space heating. This not being the case, the results may indicate that tracer breakthrough curves alone are not sufficient to predict subsequent thermal drawdown. The doublet test shows that the assumption of a direct relationship between rapid tracer breakthrough and subsequent thermal breakthrough may not be correct when considering small volumes in large reservoirs.

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Table 4-1. -- Summary of results from the KUHS and aquifer injection tests at Klamath Falls

Well name	Well number	Distance (ft)	Breakthrough (date)	Maximum (date)	Velocity (ft/day)
Fire Station	125	130	<8/1	<8/1	?
Medo-Bel	39	630	<8/1	8/5[1/]	70[1/]
Friesen	25	1,000[1/]	8/12[1/]	9/1[1/]	30[1/]
Spires and Mest	123	1,000[1/]	<9/9	?	---
Jones	80	1,120	9/23-10/9	9/23-10/9	15-20
Medical	277	1,620	---	---	---
Olympic	110	1,850	---	---	---
Butler	304	2,000[1/]	---	---	---
Production	65	3,000[1/]	---	---	---
Division St.	---	3,000[1/]	---	---	---

1/ Approximate or estimated.

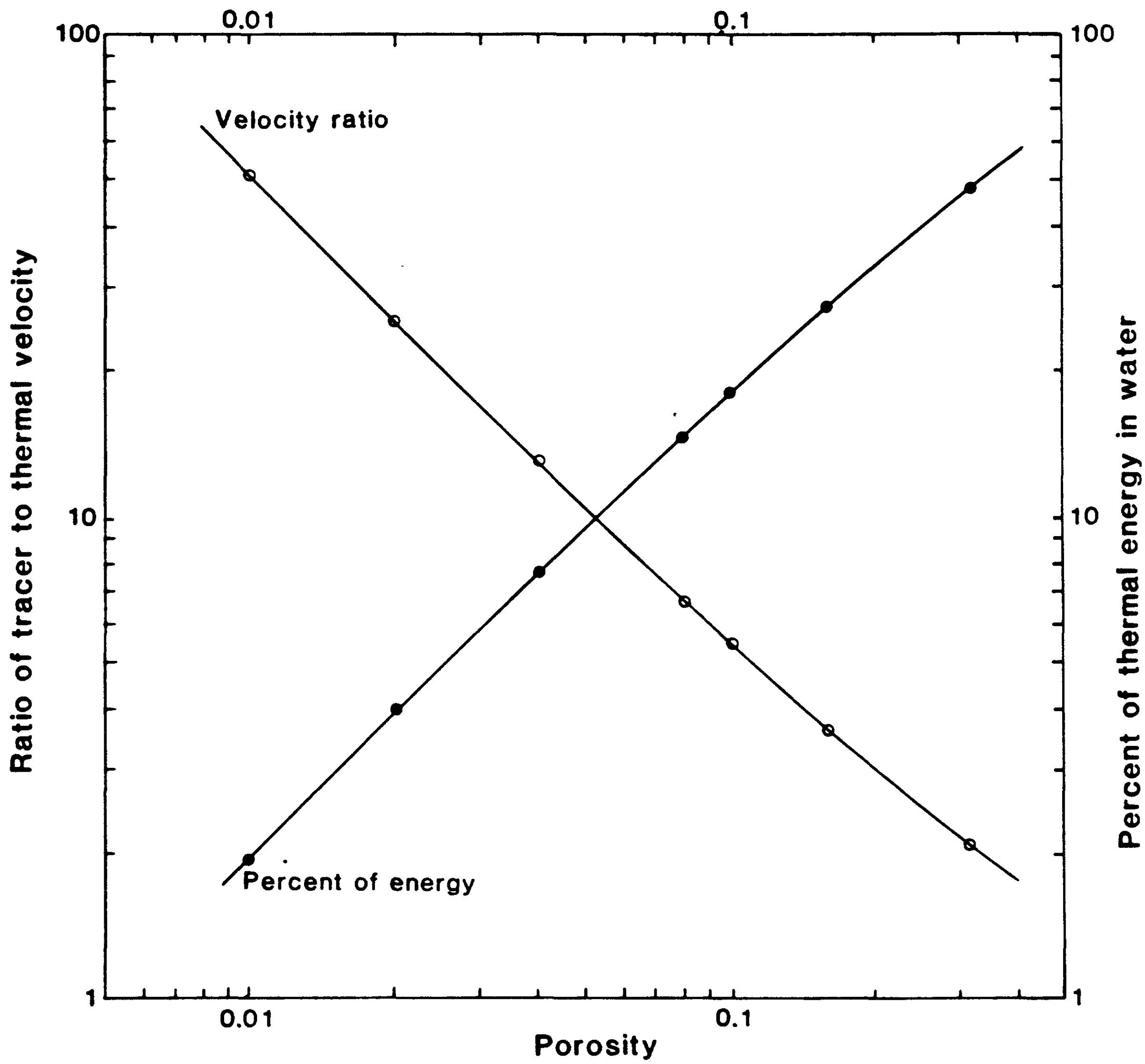

Figure 4-1. — Percent of total thermal energy content in the water of an aquifer and the ratio of tracer to thermal velocity, as functions of rock porosity.

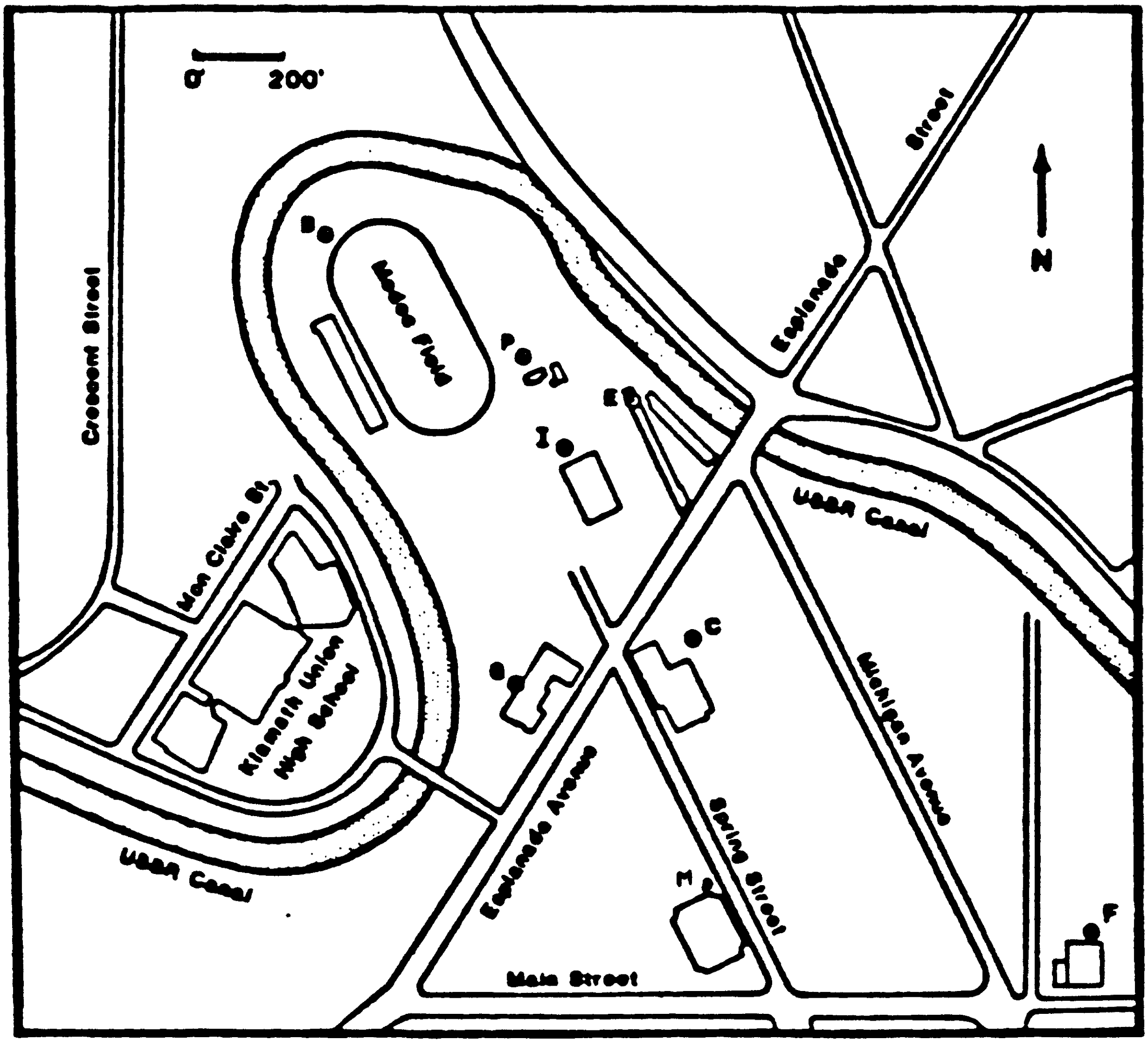

Figure 4-2. — Locations of wells for the Klamath Union High School tracer test. P, production well; I, injection well; B, Balsiger well; C, Medo-Bel well; E, Eccles well; G, Garrison well; F, Friesen well; M, County Museum well.

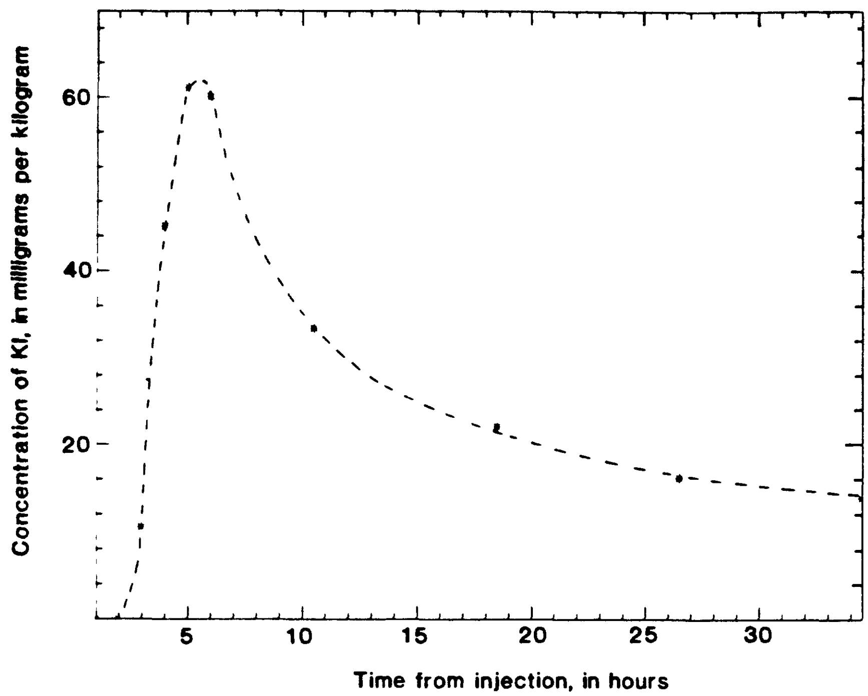

Figure 4-3. — Breakthrough curve in production well, KUHS doublet test.

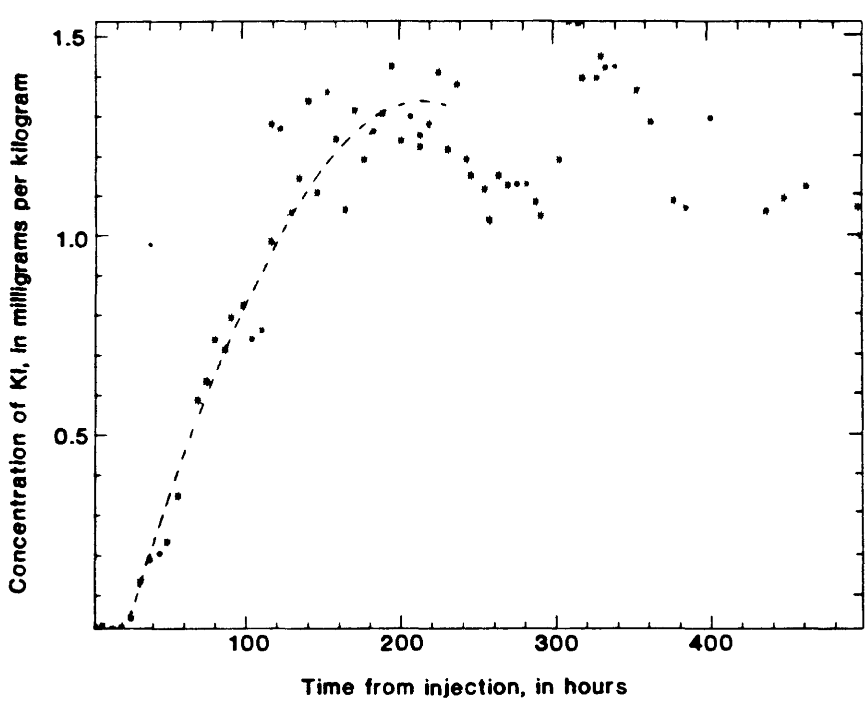

Figure 4-4. — Breakthrough curve in Medo-Bel well, KUHS doublet test.

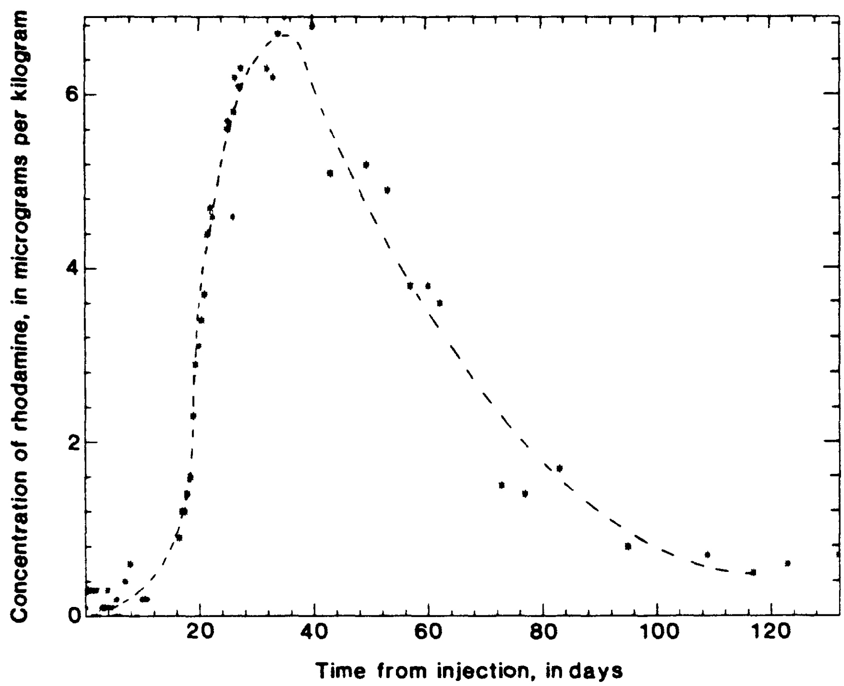

Figure 4-5. — Breakthrough curve in Friesen well, County Museum well injection test.

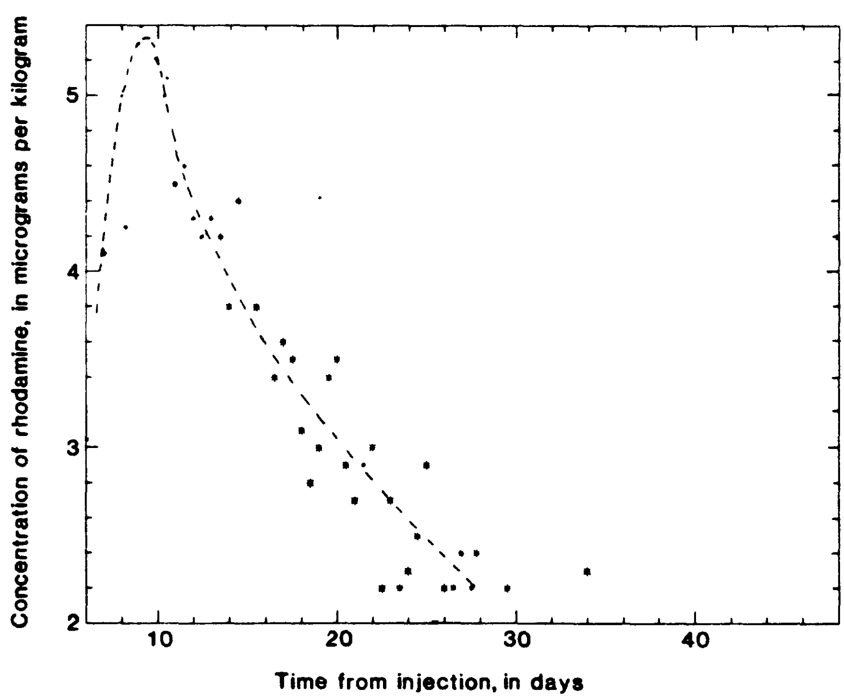

Figure 4-6. — Breakthrough curve in Medo-Bel well, County Museum well injection test.

CHAPTER 5. INTERPRETATION OF AQUIFER TEST DATA

By

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Test Description

Objectives

The objectives of the 1983 aquifer test were:

1) To assess the degree of hydrologic interconnection among the various lithologic units that comprise the Klamath Falls geothermal aquifer.
2) To evaluate the hydrologic properties of the aquifer (permeability-thickness and storage coefficient) that govern the water-level drawdown and buildup in response to pumping and reinjection.
3) To assess the spatial variations or directional properties of these properties that will influence the local response to pumping and reinjection.
4) To locate the hydrologic boundaries of the aquifer.

Based on the results of this test, it is possible to predict the impact of pumping and reinjection on the fluid levels in the nearby wells. Although this is not the primary objective of this study, drawdowns in response to two hypothetical pumping and reinjection schemes are presented in Chapter 6.

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In this report, emphasis is placed on the overall interpretation of the aquifer test, rather than on the details of individual well performance. In that sense this report can be considered only as a preliminary report on the data analysis. To fully analyze the details of each well performance requires a tremendous amount of time. Based on the preliminary analysis presented here, it appears that rigorous evaluation of the data will also require detailed numerical simulation and/or the development of new analytic solutions applicable to hydrothermal systems such as the one at Klamath Falls. The above does not, however, lessen the utility of the preliminary interpretation presented here. As will be shown, the hydrologic system responds in a remarkably uniform manner, given the complexity of the system. The average values of the aquifer properties (permeability-thickness, and storativity) determined from this analysis are more than adequate to provide reliable estimates of the short-term effects of pumping and reinjection.

Description

Hydrologic testing in the Klamath Falls geothermal aquifer is complicated by numerous factors, including: (1) an extremely heterogeneous geologic regime, (2) spatial and vertical temperature variations; (3) a large regional flow of geothermal water; (4) partial penetration of the aquifer by both the production/injection wells and the observation wells; (5) seasonal fluctuations in the water levels due to pumping and the effect of downhole heat exchangers, (6) the necessity for a method of well completion compatible with the utilization of downhole heat exchangers, (7) the relatively high permeability of the system, requiring high resolution instrumentation; and (8) the high temperatures (100°C) of many of the wells, requiring the use of non-conventional instrumentation. In order to obtain useful test data in such a system the test must be long enough for a large reservoir volume to be perturbed. In addition, extensive measurements of pressure and temperature changes must be made to evaluate the spatial variation of the reservoir properties. To this end, a six week interference test involving 52 observation wells was planned for the mid-summer months. Water-level monitoring in previous years had shown this period to be relatively free of seasonal water-level fluctuations. (See figure 2-4B, Chapter 2.)

The interference test actually covered a seven-week period in July-August, 1983, and consisted of monitoring water-level changes in 52 wells while pumping and reinjection operations were ongoing in two other wells. An area of approximately 1.7 square miles of the geothermal aquifer was monitored during the test. For the first 3 weeks, the observation wells were monitored while City Well-1 (CW-1) was pumped. For the final four weeks, hot water was pumped from CW-1 and concurrently reinjected into the County Museum well.

Schedule

The test consisted of four segments: background monitoring, pumping, pumping and reinjection, and recovery monitoring. Background data were collected for one week prior to the test (June 29 - July 5). Additional background data were also available from ongoing seasonal monitoring of water levels (see Chapter 2). On July 5th the pump in well CW-1 was turned on. All of the pumped fluid was discharged to the A-canal until July 26 (with the exception of a 1/2 hour period on July 25). From July 25 to August 24 all water pumped from well CW-1 was reinjected into the County Museum Well. On August 24 the two wells were simultaneously shut-in and pressure recovery was monitored for 1 week (Aug. 24 - Sept. 1).

Pumping Rate

Well CW-1 was pumped with a 50 hp shaft-driven pump. The pumping rates during the test are shown in figure 5-1. For the first three weeks of the test the rate remained constant at 720 gal/min. Once reinjection began, the back pressure at the reinjection well resulted in a slightly lower and somewhat variable flowrate (695-660 gal/min.

Injection Rate

The County Museum well was used for injection for the last four weeks of the test. During this period, the injection rate was identical to the pumping rate and is shown in figure 5-2.

Production Well

Well Completion and Lithology

The pumped well, CW-1 (see fig. 5-5), was completed to a total depth of 900 feet in January 1980. The well penetrates alternating layers of clay, tuff, and basic volcanic fragments (O'Brien and Benson, 1981). The lithology and temperature profile are shown schematically in figure 5-3. At the time the temperature survey was obtained the well was cased to 360 ft. Notice that the maximum temperature occurs at a depth of approximately 240 feet. Below this depth the temperature decreases. From a depth of 250 feet to the bottom of the well the temperatures remain constant. The reversal of temperature gradient below 240 feet is indicative of lateral hot-water flow in the aquifer (Bodvarsson and others, 1982; Benson and others, 1982; Blackwell and others, 1982).

Previous Tests

CW-1 was first pump-tested in January 1980. The well produced 88°C water at a maximum rate of 60 gal/min with a drawdown of 170 ft. The low temperature of the water and very low well productivity index (PI) of 0.35 gal/min-ft made the well unsuitable for its intended use. Consequently, the well was perforated from 195 to 240 ft. Shortly thereafter the well was again pump-tested. A maximum rate of 900 gal/min of 101°C water was obtained with a reported drawdown of 50 ft. The PI of the well increased to 18 gal/min-ft after perforation. The significant increase in the well productivity indicates that nearly all of the water enters the well in the perforated zones.

In late 1981 and early 1982, two short-term pump tests were conducted on CW-1. During the first test, the well was pumped for two hours at a rate of about 780 gal/min with a drawdown of 8 ft (Benson, 1982a). At this rate the measured wellhead temperature was 98.3°C. In February 1982, the well was pumped for 4 1/2 days at a rate of about 540 gal/min (Benson, 1982b). A drawdown 4.5 ft and a temperature of 97.8°C were measured.

Instrumentation

Throughout the aquifer test, measurements were recorded daily for flowrate, wellhead temperature, and water level. The flowrate was measured with a Doppler flowmeter, which requires the presence of at least 30 ppm of suspended solids or gas bubbles in the fluid. Because the suspended solid content in this water is very low, nitrogen gas was injected into the flow stream. Flowrates measured with this instrument compared reasonably well with measurements made with an in-line turbine meter. Doppler flowmeter measurements, which are those reported, are believed to be correct to within ± 10 percent.

Wellhead temperatures were measured with a bimetallic thermometer. Calibration of the thermometer after the test indicated that measured temperatures were 3.25°F lower than their correct values. The values given in this report are corrected to account for this discrepancy.

Water-level measurements in the pumped well were obtained with a bubble tube assembly. This consisted of a small-diameter tube which is lowered to a suitable depth below the water level (150 ft below the casing top in this well). At the surface, a bourdon-type pressure transducer was attached to the tube. Measurements were obtained by purging the tube with nitrogen and recording the pressure on the bourdon tube gauge. The pressure on the gauge is a reflection of the pressure exerted by the column of fluid above the bottom of the tube. Measured values could be resolved to ± 0.5 pounds per square inch (psi) ($\simeq 1.15$ feet of water).

Injection Well

Well Completion and Lithology

The County Museum well was completed to a total depth of 1,235 ft in May 1975. The lithology, shown in figure 5-4, consists of alternating layers of clay, shale (probably some tuff also), and basalt. The driller's log indicates that many of the basalt layers are fractured. The well was cased from the surface to a depth of 450.5 ft with a 10.75-inch (1/4-inch wall thickness) liner. A schematic of the well completion and temperature profile are shown in figure 5-4.

Previous Tests

Upon completion, the well was flow tested. The shut-in pressure of the artesian well was approximately 2 psi above atmospheric pressure. Wide open, the well produced 86.7°C water at approximately 188 gal/min (Lund, 1978). In August 1976, the Oregon Institute of Technology conducted a 28-hour production/interference test on the County Museum well (Lund, 1978). The well was pumped at three rates, 320 gal/min, 470 gal/min, and 670 gal/min with drawdowns of 4, 11, and 27 ft, respectively. At the highest rate the reported PI was 23.10 (gal/min)/ft. Since completion, the well has been used with a downhole heat exchanger to provide heat for the County Museum. During recent years, in the winter months, the artesian head drops below the ground surface and a small pump is required to maintain the water temperature in the wellbore (C. Leib, personal communication, 1983).

In September 1981, a sixteen-hour injection test was conducted on the Museum well in which approximately 99°C water was injected into the Museum Well at a maximum rate of 960 gal/min. An injectivity index of 8.5 (gal/-min)/ft was recorded (Benson, 1982a). After the injection test it was determined that the well bottom had filled with debris from the original depth of 1,235 ft to a depth of 1,195 ft (C. Leib, personal communication, 1983).

A second injection test was conducted in February, 1982. For five days, approximately 540 gal/min of 76 to 96°C water was injected into the well. An injectivity index of 29 (gal/min)/ft was recorded (Benson, 1982b).

Instrumentation

Throughout the injection phase of the 1983 test the wellhead pressure and flowrate were measured. A bourdon tube pressure gauge was used to monitor the wellhead pressure. Data were recorded once daily. The gauge resolution was approximately 0.5 psi. Flowrates were measured with the same system used to measure the pumping rate. The injection rate was identical to the pumping rate during the injection phase of the test (see fig. 5-2).

Other measurements made during the test included a flowing temperature profile, a spinner survey and the downhole pressure-falloff test. The temperature survey was made with a downhole temperature tool that is accurate to within 1°C. The spinner survey was made with a high-temperature downhole flowmeter designed at Lawrence Berkeley Laboratory (Solbau and others, 1983). The pressure falloff data were obtained with a downhole pressure tool designed at Lawrence Berkeley Laboratory and incorporating a quartz-crystal pressure transducer (Solbau and others, 1981). Resolution is better than 0.01 psi (0.023 ft of water). During the early part of the falloff, data were recorded at one second intervals. Once the rate of pressure falloff decreased, the recording interval was increased to 1 minute and then to 10 minutes.

Observation Wells

Well Location and Lithology

Well depths, casing top elevation, and temperatures are highly variable in the observation wells. By and large, the observation wells are relatively shallow (< 400 ft). Well completion data for the 52 observation wells shown in figure 5-5 are summarized in table 5-1. Schematics of aquifer lithology in wells along cross sections A-A' and B-B' (see figure 5-5) are shown in figure 5-6. Note that the wells shown in these cross sections are not necessarily those used as monitoring wells during the test. As shown in figure 5-6, there is no single identifiable rock unit comprising the geothermal aquifer. Presumably both the fractured basalts and contact zones between different rock layers provide the bulk of the system's permeability. Also, the principal fault zone that transects the area probably creates highly permeable near-vertical fluid conduits. The bulk of the geothermal water is stored in the pore spaces of the shale, tuff and unconsolidated sedimentary units. Conceptually, the geothermal aquifer is defined as the entire thickness of the sections penetrated by the geothermal wells. Within the aquifer, the permeability and porosity are variable. Mathematically such aquifers can be treated as a double porosity system (Warren and Root, 1963).

Instrumentation

Water-level changes in the observation wells were measured with one of three types of instrumentation: downhole pressure probes, continuously recording float defectors, and hand-operated conductivity-type detectors. Measurement of fluid levels is difficult in most of the wells in Klamath

Falls because the wellbore is filled with pipes used in the downhole heat exchanger system. Also, many of the wells have several feet of oil or parafin above the water in order to protect the well casing and pipes. This and the presence of steam at the water surface creates difficulties in using conventional conductivity probes for water-level measurements. Alternative methods for measuring water level were used whenever possible.

In wells with sufficient clearence to install a 2-inch probe, water-level measurements were made with the downhole pressure transducer described above. The transducer was lowered to approximately 50 ft below the water surface. Changes in the height of the water column above the transducer are reflected as pressure changes. Data from the wells instrumented with downhole pressure transducers were digitally recorded at 10-minute intervals throughout the test. However, when the flowrates were changed, data were recorded at 1-minute intervals (or less) for several hours.

In some of the wells with limited access, water levels were monitored continuously with Leupold-Stevens Type F water-level recorders. The Leupold-Stevens recorder uses a float and pulley system to monitor water-level changes, and the data are recorded by a clock-driven strip chart. Measurement of the water-level depth was accurate to within 0.1 ft. Time resolution of approximately 15 minutes was possible. However, float hang-ups and mechanical difficulties decreased the practical resolution of both the depth to water and time. The instrumentation used in each of the observation wells is indicated in table 5-1.

Additional Measurements

Throughout the test, records of several other characteristics and activities that potentially affect aquifer pressures were monitored. These included daily average ambient temperature, atmospheric pressure, microseismic activity, and flow in the A-canal. Chemical samples were also taken in the pumped well at regular intervals throughout the test. Samples were also taken in order to monitor the migration of the injected tracers through the reservoir.

Data Analysis

Production Well

Well Productivity

Daily records of flowrate, wellhead temperature, and water level are given in Table 5-2. Because changes in fluid level were measured only to within $\pm$1.15 feet, productivity estimates are highly variable and range from 72 (gal/min)/ft to 188 (gal/min)/ft. Also influencing these data are several other factors, such as the seasonal water-level buildup during the test and interference effects from the injection well. Previous estimates of well productivity (see section on previous tests) range from 18 to 120 (gal/min)/ft. The lack of a well defined trend in the values, suggests that variations in the productivity index result from measurement errors. Based on all of the well productivity measurements, the best estimate of the PI for CW-1 is 100 (gal/min)/ft.

Temperature

As indicated in table 5-2, no measurable temperature change occurred during the test. A comparison between these and previous data indicate that the wellhead temperature increases with flowrate. The change due to changing flow rate ($\pm$7.5°C) is larger than anticipated from conductive heat losses along the wellbore. This is an indication that mixing of the reservoir fluids is occuring and that hotter fluids are drawn to the well at higher flowrates.

Summary

The water-level drawdown and wellhead temperatures obtained from these and previous tests are plotted as functions of flowrate in figure 5-7. (Note that the data indicating a PI of 18 [gal/min]/ft is not shown in the figure. Only data obtained with the currently installed bubble tube system is plotted.) The scatter in the drawdown data is probably due to errors in measurement. The scatter in the temperature data may be the result of measurement error or seasonal variations in the aquifer temperature. For the purpose of estimating the wellhead temperatures and well drawdown as a function of flowrate, the curves (straight lines) shown in Figure 5-7 can be used.

A review of the current and past data indicates the following:

(1) The produced fluid enters the wellbore between 195 and 240 ft.

(2) The production temperature is rate dependent; the higher the flowrate, the higher the temperature of the produced water.

(3) The well draws water from a large volume of hot water, hence temperature decline in the near-term is anticipated to be minimal.

(4) The productivity index of the well is approximately 100 gal/min per foot of drawdown.

Injection Well

Temperature

Two temperature profiles of the County Museum Well were obtained during the 1983 tests. The first, measured while the well was not in use shows a maximum well temperature of 92.9°C at a depth of 1,000 ft (figure 5-4). From approximately 600 ft to 1,100 ft depth the well is nearly isothermal, indicating either a convective thermal regime in the reservoir or inter-zone flow in the wellbore. Without additional information, it is not possible to determine which of these possibilities is correct. A second temperature profile, shown in figure 5-8, was obtained during injection. This type of survey is used to identify the deepest injection zone intersecting the well. The isothermal profile indicates that water entered the formation to a depth approaching 1,150 ft. Cooler temperatures below this depth show that the injected water does not reach the bottom of the well.

Spinner Survey

On August 15, 1983, a downhole flowmeter (spinner) survey was conducted in order to identify the interval(s) accepting the injected water. The spinner survey is shown in figure 5-8. In the cased portion of the well (0 to 450.5 ft), the fluid velocity [indicated by revolutions per minute (RPM)], is nearly constant, as expected. Below the casing, the vertical fluid velocity is highly variable, reflecting substantial variations in the bore diameter. Comparison of the average spinner velocity in the casing to the average velocity below 520 ft indicates that nearly 50 percent of the injected fluid enters the rock formations between 470 and 520 ft. As can be seen from the lithologic log in figure 5-8, this interval occurs in a shale (or tuff) separating two basalt units. Any one (or all) of these units and their contacts could be accepting fluid. The decrease in spinner velocity below 1,020 ft indicates that the remainder of fluid is injected into a relatively thick basalt and shale (tuff) unit between 1,020 and 1,100 ft.

Well Injectivity

Injection rates and wellhead injection pressures were measured throughout the 29-day injection period. The data are tabulated in table 5-2. The wellhead pressures increased from approximately 39 to 43 psi over the test period. The temperature of the injected fluid remained constant at 99°C throughout the test. Note that the injection temperature is 5 to 10°C higher than the maximum temperature previously measured in the County Museum well (figs. 5-4 and 5-8). The average well injectivity during this test was 7.1 (gal/min)/ft. The apparent decrease of well injectivity (and productivity) during this test in comparison to the measurements taken in 1976 and 1982 could be attributed to plugging of the formation with material sloughing from the wellbore face, perhaps between 470 and 520 ft. The higher injectivity reported from the February 1982 test (29 [gal/min]/ft, Benson, 1982b) could have resulted partly from the higher density of the cooler injection fluid.

Pressure-Falloff Data

On the last day of the injection test, a pressure transducer was lowered into the injection well. The injection pressure was measured at a depth of 900 ft for several hours prior to shut-in. The pressure falloff was then observed for eight days. The data were analyzed using a conventional Miller-Dyes-Hutchinson (MDH) semi-log plot of the data (Earlougher, 1977). This approach was used instead of the Horner method (which is more common for falloffs) because it was recognized that relatively soon after shut-in, the well would be influenced by interference effects from the production well, thus, complicating the late-time data analysis. The MDH method is valid for pressure-falloff analysis if the time period is short with respect to the test duration. From the slope of the semi-log straight line shown in figure 5-9, a permeability-thickness of 1.35×10^6 millidarcy-feet (md-ft) is calculated. The well has a large positive skin effect, indicating that the permeabilities of the rocks immediately adjacent to the wellbore are lower than those of the reservoir rocks.

Interference Data Analysis

Observation-Well Data

Water-level measurements were obtained from 52 wells during the aquifer test. Water levels clearly changed in all of the wells, except one (No. 141), in response to both pumping and reinjection. All of the raw data obtained during the 1983 aquifer tests have been published by Benson and others, 1984. The interference data from two of these wells, typical of most of the data, are shown in figures 5-10 and 5-11. Both wells followed the same basic pattern. Pressures (water levels) decreased while only the pumping well was active. When injection began, the water levels in both wells rose rapidly. Water levels in many of the wells rose above their pre-test levels, even those relatively close to the pumped well. Two examples are shown in figures 5-12 and 5-13. Prior to shutin, the water level was nearly one foot higher than its pre-test level. This behavior is the reflection of three trends; (1) the water-level rise that normally occurs in the early summer months, (2) atmospheric pressure changes and, (3) the decreased heat loads on the downhole heat exchangers. Measurements made during this test clearly showed that water levels varied significantly in response to downhole heat exchanger use. (See fig. 2-7.) This is also illustrated in figure 5-13, which shows the water level data from the Feeback well (No. 310). Throughout the test, the water level fluctuated in response to heat-exchanger use. In general, fluctuations of less than one foot were observed. However, during the middle of the test the water level rose by nearly two feet. This rise was a significant fraction of the entire change caused by pumping (50 percent). The maximum water-level decline (4.7ft) in response to pumping was measured in the Steamer well (No. 203), which is only 122 ft from CW-1. In general, water-level changes in response to pumping and reinjection decreased with distance from the active well(s). However, there is a pronounced elongation of the cone of depression around the pumped well which indicates that the reservoir permeability is anisotropic. This is discussed in greater detail in the section entitled Steady State Analysis.

Methodology

The fractured and heterogeneous nature of the system and the interpretation of previous short-term tests suggests that a double-porosity model best describes the pressure-transient behavior in the observation wells (Benson and others, 1980; Deruyck and others, 1982; and Benson, 1984). Double-porosity behavior is characteristic of naturally fractured reservoirs in which the fractures provide most of the permeability for fluid flow and rock matrix stores the bulk of the reservoir fluid.

Interference data in double-porosity systems can be analyzed with several procedures, some of which are described by Kazemi and others, 1969, Deruyck and others, 1982 and Lai and others, 1983. For the purpose of analyzing these data, two methods are used; a log-log type curve matching technique (Deruyck and others, 1982) and a semi-log curve matching technique (Lai and others, 1983). From both of these analysis procedures the following reservoir properties are obtained: the permeability-thickness of the formation (kh), the bulk storativity $(\phi ch)_t$, and the double-porosity parameters ω and λ. The parameters ω and λ are those defined by Warren and Root (1963) as

$$\omega = \frac{(\phi c_t)_f}{(\phi c_t)_f + (\phi c_t)_m} \qquad (1)$$

and

$$\lambda = \alpha r_w^2 \frac{k_m}{k_f} . \qquad (2)$$

In the expressions given above in equations 1 and 2 and the preceding paragraph,

k = permeability (darcies),

h = thickness (ft),

ϕ = porosity (dimensionless),

c_t = total compressibility of rock and water (1/psi),

α = a geometric factor ($1/ft^2$),

r_w = well radius (ft),

and the subscripts f and m refer to fractures and rock matrix, respectively.

In order to perform a detailed analysis of the drawdown and buildup data, the data must be relatively free of perturbations created by sources other than pumping or injection. This constraint has several implications for the data analysis; (1) wells affected by heat-exchanger use are not suitable for detailed analysis, and (2) because the late-time data obtained from all of the wells are influenced by the seasonal water-level buildup, only the relatively early time data are considered reliable. For these reasons, detailed analyses are performed only on unused wells and only the first 300 to 400 hours of data from both the pumping and reinjection phase of the test are used for analysis.

Pressure-buildup data from the injection phase of this test are analyzed by assuming that the pressure transients occuring in the inital pumping period have reached steady state and therefore can be ignored in the subsequent calculations. In relation to the background noise (attributed to other well users and seasonal water-level changes), errors due to this assumption are small. During the injection test, flow rates were slightly variable (42 to 40 kg/s). This variation is also neglected in the analysis because water-level changes due to other sources (barometric pressure fluctuations and other well users) are of the same order of magnitude as those resulting from the flowrate variations and do not affect the overall data interpretation.

Type-Curve Analysis

Each of the data sets suitable for detailed interpretation was plotted on log-log paper. The data were then matched to the double-porosity type curves prepared by Deruyck and others, 1982. If the observation well is distant from the active well and the test duration is sufficiently long, the double porosity behavior does not greatly influence the determination of transmissivity and the Theis Curve is used for the analysis (Theis, 1935).

Type-curve matches for drawdown data measured in seven of the wells are shown in figures 5-14 to 5-20. In each case the match between the theoretical and measured data is excellent. Table 5-3 summarizes the values of kh, $(\phi ch)_t$, ω, and λ obtained from the analysis of each data set. Note that values for the parameter kh are all in good agreement with one another. The average of these values is 1.5 X 10^6 md-ft. The extremely heterogeneous nature of the geothermal aquifer creates uncertainty as to the appropriate value of the reservoir thickness. Therefore, it is not possible to evaluate independently the value of the permeability. However, if the thickness is estimated to be 1,000 ft, the reservoir permeability is approximately 1.5 darcies. This is significantly higher than the permeability of most geothermal systems (Bodvarsson and Benson, 1983). The values of $(\phi ch)_t$ range from 0.792 to 3.03 X 10^{-3} ft/psi. The anomalously high values $>10^{-2}$ occur in a region of high permeability and porosity that surrounds the pumped well and is intersected by several of the observation wells (Benson and Lai, 1984). This is discussed in greater detail in the section entitled Steady State Analysis. Variations in the value of $(\phi ch)_t$ also reflect the anisotropic permeability of the system (Earlougher, 1977). The values of ω range from 0.01 to 0.3. The value for ω of 0.3 is calculated from a well that is 2,200 ft from CW-1. For a well this far from the active well the pressure transients are not very sensitive to the double-porosity parameters (Benson, 1984). Neglecting this value, the average value for ω is 0.014. Values for the parameter λ range from 1.51 X 10^{-5} to 1.86 X 10^{-7}. This variation may result from local permeability heterogeneity and (or) lack of sensitivity to the parameter (Benson, 1984). On the basis of the present analysis, the best estimate of λ for the hillside area lies between 10^{-6} and 10^{-7}.

Similar analyses were performed on the pressure buildup in response to injection. Log-log plots and the type-curve matches for seven of the wells are shown in figures 5-21 to 5-27. Note that two of the wells (figs. 5-22 and 5-23) match the type-curves for pseudo-steady-state inter-porosity flow rather than those for transient interporosity flow (Deruyck and others, 1982; Benson, 1984). An explanation for this is not available at this time. However, one interpretation is that flow from the matrix to the fracture is impaired by a fracture skin (Moench, 1983). Also note that data from two of the wells (figures 5-26 and 5-27) are matched to the Theis Curve. This confirms the theory that for observation wells sufficiently distant from the active well certain double-porosity reservoirs behave as an equivalent porous medium. Values of the properties kh, $(\phi ch)_t$, and ω and λ are summarized in table 5-3. Again, all of the kh values are in good agreement with one another. The average value is 1.3 X 10^6 md-ft. This value compares well with the value of 1.5 X 10^6 md-ft calculated from the drawdown test. The average value of $(\phi ch)_t$ is 4.62 X 10^{-3} ft/psi. There is far better agreement of the values $(\phi ch)_t$ calculated for the buildup data than those calculated from the drawdown data. The parameter ω ranges from 6 X 10^{-2} to 0.3. The range probably results from a combination of reservoir heterogenity and a lack of sensitivity in variations of this parameter. The value of λ ranges from 2.7 X 10^{-8} to 5.1 X 10^{-7} for similiar reasons.

From the results of the type-curve analysis of both the drawdown and buildup data the following estimates of the reservoir parameters are obtained:

$$\begin{aligned} kh &= 1.4 \text{ X } 10^{-6} \text{ millidarcy-ft (md-ft)} \\ (\phi ch)_t &\simeq 5 \text{ X } 10^{-3} \text{ ft/psi} \\ \lambda &\simeq 10^{-7} \\ \omega &\simeq 10^{-2} \end{aligned}$$

The reservoir properties calculated for individual wells have a remarkably small spread around the average values, especially in light of the highly heterogeneous nature of the system.

Semi-log Analysis

In addition to type-curve matching, the data were analyzed with the semi-logarithmic method discussed previously. Whereas type-curve matching tends to weight the data interpretation towards the early-time data, semi-logarithmic analyses are strongly weighted towards the late-time pressure response. Nevertheless in both instances, it is the late-time data that establishes the value of kh. In the semi-log method, the pressure data are plotted versus log (time). The permeability-thickness of the system is calculated from the slope of the semi-log straight line that is drawn through the late-time data points. The storativity, $(\phi ch)_t$, is calculated from the time at which the straight line intersects the x-axis (i.e., $\Delta p=0$). In order to evaluate the double-porosity parameters, λ and ω, the data are history-matched using the analytic solution developed for double-porosity systems by Lai and others, 1983. The values of λ and ω are chosen from the best history matches.

As an example, the Page well drawdown analysis is illustrated in figure 5-28. The straight line shown in the graph is used to calculate the permeability thickness of the reservoir. The intercept, at 7 hours, is used to calculate $(\phi ch)_t$. The data matches for different pairs of λ and ω are also shown. The best match is obtained for $\lambda = 6 \times 10^{-7}$ and $\omega = 0.1$. Data from each of the wells analyzed with type curves (see previous section) were analyzed using this technique. The results of the analyses are given in table 5-3. Together with the results of the type-curve match. With few exceptions, the results from the two methods of analysis are in excellent agreement with one another. The shape of the log-log plots suggests, however, that a double-porosity model appropriately describes the pressure transients in the Klamath Falls geothermal aquifer, and this suggests also that the log-log plots, which take early data into account, may provide the more reliable estimates of storativity.

Data from some of the wells in which water-level measurements were made with float recorders or conductivity meters were also analyzed using the semi-log technique. The graph in figure 5-29 shows the drawdown at five of these wells. In each case it is clear that the late-time data do not lie on a single straight line, as is expected from the results of the previously discussed analyses. However, the best straight-line fits to the data are shown in each case. The average value of kh, 1.6×10^6 md-ft, is in good agreeement with that calculated from the other analyses, 1.5×10^6 md-ft. The pressure response at the Zion Church well (No. 274), also shown in figure 5-29, does not follow the same trend as the other wells. The drawdown and the rate of drawdown are far less than for the other wells. However, the well clearly responded to pumping. This behavior is attributed to the lack of high permeability fractures or strata in the vicinity of this well. As a result of this, both the rate and magnitude of drawdown are less than observed in the other wells.

The results of the semi-logarithmic analyses are in good agreement with those obtained by type-curve matching. Taken together, the two methods show that the aquifer behaves hydrologically like an infinitely large double-porosity system. The lack of evidence for hydrologic discontinuities in the system indicates that there is a large volume of hot water available in the near-surface aquifer, as well as from the deep geothermal circulation system.

Steady-State Analysis

An alternative approach to analyzing the well test data is to look at the distribution of the pressure drawdown in the reservoir at a specified time. This approach tends to emphasize the difference between individual well performance and the anisotropy of the hydrologic properties. For example, the drawdowns in 22 of the observation wells, after 336 hours of pumping at a rate of 720 gal/min, are shown in figure 5-30. (Data from wells strongly affected by heat-exchanger use or measurement error were not considered in this analysis.) In general, as shown in the figure, drawdowns were greatest in the immediate vicinity of the production well. However, there is not a monotonic relationship between the radial distance to the observation well and the magnitude of the drawdown. Instead, the drawdowns are somewhat greater than expected along a NW trend that is parallel to the regional structure. The elongation of the drawdowns in this manner is indicative of anisotropic aquifer permeability. Although there are no reliable data from wells at distances greater than about 1,500 ft transverse to the trend, the indications of anisotrophy are consistent with the interpretation of previous test data from the Museum well (Lund and others, 1978). The trend of the major axis of permeability is nearly the same as the stike of the major normal fault that transects the area (fig. 1-2). Possibly, this fault, or fault zone, provides additional high-permeability fractures along which fluid flows easily. Comparison between the observed drawdowns and preliminary theoretical calculations (not discussed here) indicates that the ratio of the major to minor axis permeability is between 5 and 10.

Permeability anisotropy can not fully explain the observed pressure drawdowns. For example, there are several wells in which the drawdowns are greater or less than predicted by anisotropy. In particular, wells that are close to the A-canal experience smaller drawdowns than those elsewhere. (See figure 5-5 for A-canal location.) Drawdowns in several of the wells were slightly greater than anticipated from simple anisotropy. The discrepancies were not large and intepretation must be postponed until individual analyses of each of the 52 observation wells are complete.

When a steady-state flow condition (or approximately steady-state flow) is reached, it is possible to calculate the reservoir pemeability from distance vs. drawdown graphs (Thiem, 1906). For a homogeneous, isotropic system, the data points should all fall on a single straight line (when the drawdown is plotted as a function of logarithm of the radial distance to the well). Clearly, the system is neither homogeneous nor isotropic. However, such a graph can provide insight into the behavior of the system. The drawdown versus distance for the 24 observation wells that have continuous records (after 336 hours of pumping at 720 gal/min) is shown in figure 5-31. There is a large scatter of the data points, indicative of anisotropy and aquifer heterogeneity. However, one possible interpretation is that there are two regions in the aquifer, a high-permeability region in the vicinity of the pumped well (indicated by slope m_1') and a lower permeability region surrounding it (indicated by the slope m_2'). Slope m_1' is determined by 4 of the 5 transducer wells located along Old Fort Road near the pumped well. The fifth (well 4) is anomalous, and well 101 (Head), may fall along this trend fortuitously. Slope m_2' is an estimated fit to the remaining data points which results in a calculated permeability of about 3×10^6 md-ft, a value greater than any derived from analyses of individual well responses. Although this interpretation is not very convincing, because of the scatter in the data points, it does explain several observations in the data that are unexplained by other interpretations. The presence of a high permeability inner region explains the small drawdowns [in comparison to those predicted by the average values of kh and $(\phi ch)_t$] in the immediate vicinity of the pumped well and the anomalously high values of the storage coefficient (Benson and Lai, 1984). In the absence of detailed analysis of the extent and shape of this inner region and the anomalies represented by wells beyond this region, it is not possible at the present time to provide further interpretation.

A similar analysis can be applied to the pressure buildup caused by reinjection. The pressure buildup, after 300 hours of pumping at a rate of approximately 700 gal/min, for 7 of the observation wells measured by transducer is shown in figure 5-32. The best straight line fit to these data is shown in the graph. A kh of 1.35 X 10^6 md-ft is calculated from the slope of this line. This is in excellent agreement with the average calculated value of 1.4 X 10^6 md-ft. Conversion of the data from the remaining observation wells to a common basis is not yet complete, however, and this conclusion is a preliminary one.

Although the steady state analysis is not entirely consistent with the interpretation of the pressure-transient data, two additional factors are suggested by this approach: A pronounced permeability anisotropy and the presence of a high permeability region around the pumped well. The analysis of these two factors can not be considered to be complete. Futhermore, other interpretations may explain the observed pressure response as well as those presented here. Until additional analyses are performed, a more definitive answer can not be obtained.

Temperature Data

Downhole temperatures were measured in all the wells monitored with downhole pressure transducers (fig. 5-5). The objective of these measurements was to determine if pumping large amounts of geothermal water would quickly change the aquifer temperature. Data from 9 of these wells are shown in figures 5-33 and 5-34. During the measurements, both the Rogers Well and the Assembly of God Well were being used for space heating with downhole heat exchangers. Therefore, most of the temperature changes in these wells probably were due to heat-exchanger use rather than temperature changes in the aquifer. Throughout the test the temperature declined in the Spires and Mest Well. This occurred because pumping of hot water from the well had ceased, and the bore-fluid temperature was equilibrating with the surrounding rock. Data from the remaining wells were unaffected by use and, therefore, are indicative of aquifer-temperature changes induced by pumping.

Temperatures in the Page Well, Head Well, and Harley Davidson Well remained nearly constant throughout the test. The temperatures at the Parks and Carroll wells increased by approximately 1 and 0.5°C respectively. The temperature in the Parks Steamer Well decreased 1°C during pumping and increased 1°C when injection began. The temperature decline in the Parks Steamer well appears to be caused by fluid-level changes in the wellbore rather than aquifer-temperature changes (as evidenced by the temperature recovery during injection). All of these changes were small and no systematic trend is apparent. Also, these changes are less than those that occurr in a yearly cycle (Lund and others, 1978). The lack of significant temperature change during the test precludes the possibility of establishing a relationship between pumping and temperature changes in the aquifer. Meaningful measurements can only be made over a much longer time period (years). However, the absence of definitive temperature response indicates that a large volume of hot water, at a relatively constant temperature, is stored in the geothermal aquifer and fault system.

Summary

An enormous quantity of hydrologic data has been obtained from the Klamath Falls geothermal aquifer. It is an unprecedented achievement both for the extraordinarily high quality data obtained and for the cooperation of the many individuals who provided data, interest, and participation in the aquifer test. Under well controlled conditions, the response of the aquifer and individual wells to both pumping and reinjection were measured. The test was of sufficient duration and the data of adequate quality so that conventional and non-conventional analysis techniques could be applied with a great deal of confidence. Although this large data set could not be throughly analyzed in the time thus far dedicated to the task, many of the questions previously unanswered are now answerable. However, as is the case in any scientific investigation, new questions have also arisen. The status of the investigation can be summarized as follows.

The geothermal aquifer underlying the city of Klamath Falls is primarily a fault- and fracture-controlled system. The fault(s) and fractures provide highly permeable paths along which water moves easily. The sediments and tuffaceous rocks provide the bulk of the storage capacity of the aquifer as indicated by the double-porosity type pressure transients. Pressure-transient and steady-state analyses of the drawdown and buildup data from many wells were similar, although there are numerous unresolved discrepancies. Average values for the hydrologic properties of the system are as follows:

$$kh = 1.4 \times 10^{6} \text{ md-ft}$$
$$(\phi ch)_t \simeq 5 \times 10^{-3} \text{ ft/psi}$$
$$\lambda \simeq 10^{-7}$$
$$\omega \simeq 10^{-2}$$

No hydrologic boundaries were detected during the test. On the basis of the above results, the radius of investigation (at 336 hours) is estimated to be 3.5 miles. The lack of boundaries to the system within this radius has several important consequences. First, it sheds an interesting light on the hydrologic properties of the fault zone that is the primary conduit for hydrothermal circulation. Unlike the response predicted by classical models for constant-potential or constant-flow faults, this fault was invisible to hydrologic testing. Several hypotheses can explain this observation. First, the hot water may upwell over a broad region rather than along a single fault zone that could be detected hydrologically. Second, the fault permeability may be of the same order of magnitude as the permeability of the near surface aquifers and hence, indistinguishable. Third, a single fault may provide the conduit for upwelling from great depth but as the fault approaches the surface, the width of the fractured zone increases and creates a diffuse permeable region in the near surface. Additional research in this area could provide further insight into the nature of the supply conduits. A second implication of the lack of hydrologic boundaries to this system is that the hydrothermal system is an integral part of the regional hydrologic system. As such, fluid recharge should be in abundant supply. However, if the hydrologic gradient is altered so that the natural groundwater flow system is disturbed significantly, cold water may enter the geothermal aquifers. The lack of temperature change in the aquifer during these tests indicates, however, that such an occurrence would not happen rapidly and perhaps, not at all. This too is an area in which additional research would be fruitful.

Details of individual response of each of the 52 observation wells have not been analyzed. Data from many of these wells were strongly affected by downhole heat-exchanger use. As such, they were not amenable to the analysis procedures used here. However, additional investigation of these data will provide insight into local variations of the hydrologic properties. Of particular interest is the region of very high permeability surrounding CW-1. This too will be pursued in future investigations.

The overall hydrologic characteristics of the geothermal aquifer have been determined. Most of the data are remarkably consistent and local variations in hydrologic properties have only a second order effect on the pressure response. This agreement allows the prediction of the aquifer response to pumping and injection with a relatively simple mathematical model. The details of this model and sample calculations for water level changes in response to pumping and reinjection are presented in Chapter 6.

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Table 5-1. -- Well-completion data and instrumentation summary for 52 observation wells

[DHE = Downhole Heat Exchange]

Well number	Owner	Address of well	Depth (ft)	Temperature: Date	Temperature: (°C)	Temperature: Source	Heat extraction method	Distance to production well (ft)	Distance to injection well (ft)	Water level 1/, in feet below measuring point: Pre-production July 5	Pre-injection July 26	Stop pump Aug.24
LBL Transducers/ Type F recorder												
3	Carroll	155 N. Wendling	303	---	101	OIT	Unused	540	3,200	99.66	103.77	100.14
4	Parks	325 Old Fort Rd.	795	---	112	OIT	Unused	703	3,460	70.51	73.40	72.60
										Instrument removed 07-17-83		
8	Vlahos	2022 Main	198	---	65	OIT	DHE	930	2,130	---	---	---
24	Assembly of God	235 S. Laguna	363	---	91	OIT	DHE	1,280	3,460	43.24	46.38	42.57
39	Medo-Bel Dairy	1500 Esplanade	765	---	98	OIT	Unused	2,740	630	---	11.48	(artesian)
101	Head	2051 Erie	690	---	93	OIT	Unused	1,030	2,120	75.02	79.14	75.61
123	Spires & Mest	120 East Main	900	---	86	OIT	Artesian	2,100	750	---	4.97	+1.13 2/
177	Page	1721 Menlo Way	465	06-82	88	Drill. log	Unused	2,200	860	10.93	13.45	8.74
200	Rogers	132 Laguna	329	03-80	93	Drill. log	DHE	409	2,660	36.17	40.16	34.20
203	Parks Steamer	Gibbs & Old Fort Rd	---	08-76	101	Drill. log	Unused	122	2,840	49.21	53.93	50.81
215	Harley Davidson	1409 Main St.	205	1978	60	OIT	Pumped	2,770	160	---	6.48	1.52
347	Dearborn	Main & Alameda	300	12-68	97	Drill. log	DHE	630	2,570	23.46	28.05	24.21

1/ Wells 3 through 215 on this page were monitored soley by using a down-hole transducer. Well #39 (Medo-Bel Dairy) not measured at start of test; instrument removed 07-31-83 when well began to flow. Wells #123 (Spires and Mest) and #215 (Harley Davidson) not measured at start of test.

2/ Water level measured in standpipe above land surface.

Table 5-1. -- Well-completion data and instrumentation for 52 observation wells (Continued)

[DHE = Downhole Heat Exchange]

Well number	Owner	Address of well	Depth (ft)	Temperature Date	Temperature (°C)	Temperature Source	Heat extraction method	Distance to production well (ft)	Distance to injection well (ft)	Water level, in feet below measuring point: Pre-production July 5	Pre-injection July 26	Stop pump Aug.24
Type F Recorder (Stevens)												
43	Klamath Cold Storage	661 Spring	1,100	---	44	OIT	Pumped	3,000	2,460	---	8.4	4.1
79	Klamath Union High School	Monclaire St.	240	---	67	OIT	Pump/ reinject	3,060	1,290	---	8.22	---
80	Jones	310 Spring	485	---	67	OIT	Pumped-DHE	2,730	1,120	8.0	10.8	5.9
83	Stanke	424 Hillside	136	---	99	OIT	Unused	1,120	2,310	109.40	112.83	109.0
97	Angel	307 Damont	392	---	95	OIT	DHE	380	2,750	94.1	97.7	93.7
118	Svanevik	1932 Portland	340	---	94	OIT	DHE	3,100	2,450	74.1	76.81	72.62
122	Ponderosa School	107 S. Williams	429	---	93	OIT	DHE	1,260	3,700	50.9	52.58	49.85
126	Christian Center	115 N. Alameda	516	---	88	OIT	Unused	950	1,780	15.36	17.9	14.55
141	Eldorado Place	2200 N. Eldorado	407	11-75	26	Drill. log	Unused	7,180	6,200	---	97.63	95.46
166	Card	1902 Esplanade	613	08-79	99	Drill. log	DHE	2,380	1,620	41.3	44.6	40.5
178	Phillips	1945 Auburn	360	03-83	88	Drill. log	DHE	2,000	1,600	---	40.38	36.24
261	Olson	516 Old Fort Rd.	975	09-80	102	Drill. log	Unused	1,620	4,050	120.2	123.7	120.7
273	Eck	2130 Herbert	199	07-76	97	Drill. log	Unused	360	2,500	38.0	42.7	39.1
274	Zion Church	Eleventh St.	838	11-82	54	Drill. log	Unused	3,990	1,340	7.1	7.9	7.3
448	Murphy 1/	443 Laguna	---	---	---	---	DHE	1,050	2,710	152.86	---	---

1/ Graphic plot not included in figures.

Table 5-1. -- Well-completion data and instrumentation summary for 52 observation wells (Continued)

[DHE = Downhole Heat Exchange]

Well number	Owner	Address of well	Depth (ft)	Temperature: Date	Temperature: (°C)	Temperature: Source	Heat extraction method	Distance to production well (ft)	Distance to injection well (ft)	Water level, in feet below measuring point: Pre-production July 5	Pre-injection July 26	Stop pump Aug. 24
Hand measured by D. C. Long												
19	Glidden	1800 Fairmount	638	---	92	OIT	DHE	6,380	6,290	239.15 1/	244.73 1/	241.35 1/
22	Thexton	235 N. Alameda	254	---	94	OIT	DHE	1,250	1,460	20.41	22.67	19.39
51	Bailey	1850 Lowell	625	---	94	OIT	DHE	5,600	4,730	95.77	97.28	93.97
124	Hardt/Davis	535 Laguna	997	---	84	OIT	DHE	1,350	2,780	218.48 1/	227.18 1/	224.88 1/
135	Ross	1879 Del Moro	152	08-56	93	Drill. log	DHE	4,600	3,850	72.48	75.94	71.79
186	Fillmore	212 Hillside	150	06-79	93	Drill. log	DHE	500	2,200	52.82	56.17	52.46
194	Terrier	234 Laguna	240	05-58	103	Drill. log	DHE	140	2,560	57.68	60.37	56.63
256	Kent	2325 Linda Vista	1,080	04-79	130	OIT	DHE	1,790	3,260	291.29 2/	309.05 2/	305.36 2/
264	Cooper	430 Laguna	268	10-77	102	Drill. log	DHE	960	2,540	131.63	135.32	132.52
294	Spielman	1500 Pacific Terrace	525	05-83	93	CRGD 3/	DHE	5,270	4,780	132.30	(7-29-83) 135.50	131.34
360	Carter	1100 N. Eldorado	125	02-81	89	Drill. log	DHE	4,060	3,150	46.82	49.18	47.34

1/ Reported water-level measurements considered unreliable as a result of instrument error.

2/ Graphic plot not included in figures. Reported water-level measurements considered unreliable as a result of instrument error.

3/ Data obtained by Citizens for Responsible Geothermal Development.

Table 5-1. -- Well-completion data and instrumentation summary for 52 observation wells (Continued)

[DHE = Downhole Heat Exchange]

Well number	Owner	Address of well	Depth (ft)	Temperature Date	Temperature (°C)	Temperature Source	Heat extraction method	Distance to production well (ft)	Distance to injection well (ft)	Water level, in feet below measuring point: Pre-production July 5	Pre-injection July 26	Stop pump Aug.24
Owner monitored △												
37	Stone	133 Hillside	200	---	102	OIT	DHE	420	2,380	38.8	42.0	37.4
61	Juckeland	2043 Lavey	275	---	88	OIT	DHE	880	2,050	76.6	79.5	75.8
62	Phelps	1868 Fremont	126	---	91	OIT	DHE	3,950	2,980	44.2	46.5	42.4
110	East Main Apts[1/]	236 East Main	452	---	72	OIT	Artesian Pumped	1,850	1,850	0.0 (6-28-83)	3.6 (07-24-83)	+0.46 (8-24-83)
127	Adamcheck[2/]	Laguna & Old Fort Rd.	233	---	100	OIT	DHE	140	2,640	49.8	52.75	49.0
128	Hart	2052 Lavey	---	---	72	OIT	DHE	810	2,030	56.25	59.92	55.92
143	Raney	1126 Eldorado	152	07-48	96	Drill.log	DHE	4,190	3,310	---	59.3	55.2
157	Heaton & Wardell	700 Loma Linda	455	02-57	103	Drill.log	DHE	2,030	2,890	246.5	249.2	244.5
165	Mathews	1832 Earle	180	06-83	82	Owner	DHE	2,800	1,570	29.2	32.0	---
170	Lawrence	2384 Linda Vista	805	08-73	106	Drill.log	DHE	1,420	2,920	245.6	248.9	244.8
181	Hessig	410 Hillside	163	01-63	93	Drill.log	DHE	990	2,220	102.7	106.7	102.3
216	Hart[1/]	125 Eldorado	690	03-83	71	S.Swanson	Artesian	1,930	760	+2.5	5.0	---
277	Klamath Medical Clinic[1/]	1905 Main	---	04-58	96	Drill.log	DHE	1,160	1,620	---	7.25 (07-24-83)	3.42
310	Feeback	207 Haskins	470	1956	100	CRGD[3/]	DHE	1,480	2,250	36.25	38.7	35.5

1/ Well monitored by Charles Leib.

2/ Well monitored by A. L. Stome

3/ Data obtained by Citizens for Responsible Geothermal Development

Table 5-2. -- Summary of pumped and injection well data from the 1983 aquifer test

Date	Flowrate	Pump well		Injection well
		Temperature	Water level	well-head pressure
	(gal/min)	(°F)[1/]	(ft)[2/]	(psi)
7-05-83[3/]	0,720	212.3	64,72	0
7-06-83	720	212.3	74	0
7-07-83	720	212.3	74	0
7-08-83	720	212.3	74	0
7-09-83	720	212.3	74	0
7-10-83	720	212.3	74	0
7-11-83	720	212.3	74	0
7-12-83	720	212.3	74	0
7-13-83	720	212.3	74	0
7-14-83	720	212.3	74	0
7-15-83	720	212.3	74	0
7-16-83	720	212.3	74	0
7-17-83	720	212.3	74	0
7-18-83	720	212.3	74	0
7-19-83	720	212.3	74	0
7-20-83	720	212.3	74	0
7-21-83	720	212.3	74	0
7-22-83	720	212.3	74	0
7-23-83	720	212.3	72	0
7-24-83	720	212.3	72	0
7-25-83	720	212.3	72	0
7-26-83[4/]	720,695	212.3	72	0,37
7-27-83	690	212.3	71	39
7-28-83	685	212.3	69.5	36
7-29-83	695	212.3	69.5	37
7-30-83	695	212.3	69.5	37
7-31-83	695	212.3	69.5	38
8-01-83	690	212.3	69.5	38
8-02-83	690	212.3	68.5	38
8-03-83	690	212.3	68.5	38
8-04-83	690	212.3	67.5	38
8-05-83	690	212.3	67.5	39
8-06-83	685	212.3	67.5	39
8-07-83	685	212.3	67.5	39
8-08-83	685	212.3	67.5	40
8-09-83	680	212.3	67.5	40
8-10-83	680	212.3	67.5	40
8-11-83	680	212.3	67.5	40
8-12-83	680	212.3	67.5	40
8-13-83	680	212.3	67.5	40
8-14-83	680	212.3	67.5	40
8-15-83	680	212.3	67.5	40
8-16-83	675	212.3	67.5	--
8-17-83	670	212.3	67.5	--
8-18-83	665	212.3	67.5	--
8-19-83	665	212.3	67.5	42
8-20-83	665	212.3	67.5	43
8-21-83	665	212.3	67.5	42
8-22-83	665	212.3	66.5	43
8-23-83	660	212.3	66.5	43
8-24-83[5/]	660	212.3	66.5, 63	43

1/ Reported temperatures are corrected to post-test calibration. Calibration indicated that the gauge was reading 3.25°F low throughout the test.

2/ Water level, in feet below measuring point, calculated from pressure-gauge readings. Gauge resolution of 0.5 psi equivalent column of water ($\simeq$1.15 feet).

3/ Pumping starts at 15:10 (3:10 p.m.)

4/ Injection starts at 10:11.

5/ Shut in at 17:35:41 ($\simeq$ 5:35 p.m.).

Table 5-3. -- Summary of the values of the kh, $(\phi ch)_t$, and the parameters ω and λ calculated from semi-log and type-curve (log-log) analyses

Well	No.	kh (md-ft)		$(\phi ch)_t$ (ft/psi)		ω		λ		Active well
		Semi-log	Log-log	Semi-log	Log-log	Semi-log	Log-log	Semi-log	Log-log	
Assembly of God	24	1.55×10^6	1.74×10^6	6.9×10^{-3}	5.1×10^{-3}	0.02	0.01	1.0×10^{-7}	2.7×10^{-7}	PRO
Parks	4	1.92×10^6	1.96×10^6	2.2×10^{-2}	3.0×10^{-3}	0.04	0.01	1.3×10^{-6}	1.7×10^{-6}	PRO
Carroll	3	1.47×10^6	1.52×10^6	2.7×10^{-2}	2.6×10^{-2}	0.01	0.015	7.5×10^{-7}	1.1×10^{-6}	PRO
Head	101	1.35×10^6	1.39×10^6	1.1×10^{-2}	9.5×10^{-3}	0.02	0.01	2.2×10^{-7}	4.1×10^{-7}	PRO
Parks Steamer	203	1.04×10^6	1.34×10^6	1.1	0.79	0.001	0.01	1.5×10^{-5}	1.5×10^{-5}	PRO
Page	177	1.74×10^6	1.74×10^6	4.8×10^{-3}	5.7×10^{-3}	0.1	0.3	6.0×10^{-7}	1.9×10^{-7}	PRO
Head	101	1.35×10^6	1.17×10^6	4.6×10^{-3}	6.1×10^{-3}	0.3	0.3	5.0×10^{-6}	5.1×10^{-7}	INJ
Carroll	3	1.23×10^6	1.27×10^6	2.2×10^{-3}	2.3×10^{-3}	0.1	0.3	1.5×10^{-6}	2.2×10^{-7}	INJ
Parks Steamer	203	1.41×10^6	1.35×10^6	2.6×10^{-3}	3.2×10^{-3}	1	1	------	------	INJ
Assembly of God	24	9.60×10^5	9.54×10^5	3.5×10^{-3}	4.1×10^{-3}	1	1	------	------	INJ
Page	177	1.32×10^6	1.43×10^6	7.8×10^{-3}	7.3×10^{-3}	0.02	0.06	3.0×10^{-7}	2.4×10^{-7}	INJ
Medo Bell	39	1.13×10^6	1.14×10^6	2.8×10^{-3}	1.1×10^{-3}	0.04	0.15	9.0×10^{-8}	2.7×10^{-8}	INJ
Spires and Mest	123	1.05×10^6	1.28×10^6	7.9×10^{-3}	6.0×10^{-3}	0.003	0.006	3.0×10^{-7}	1.8×10^{-7}	INJ

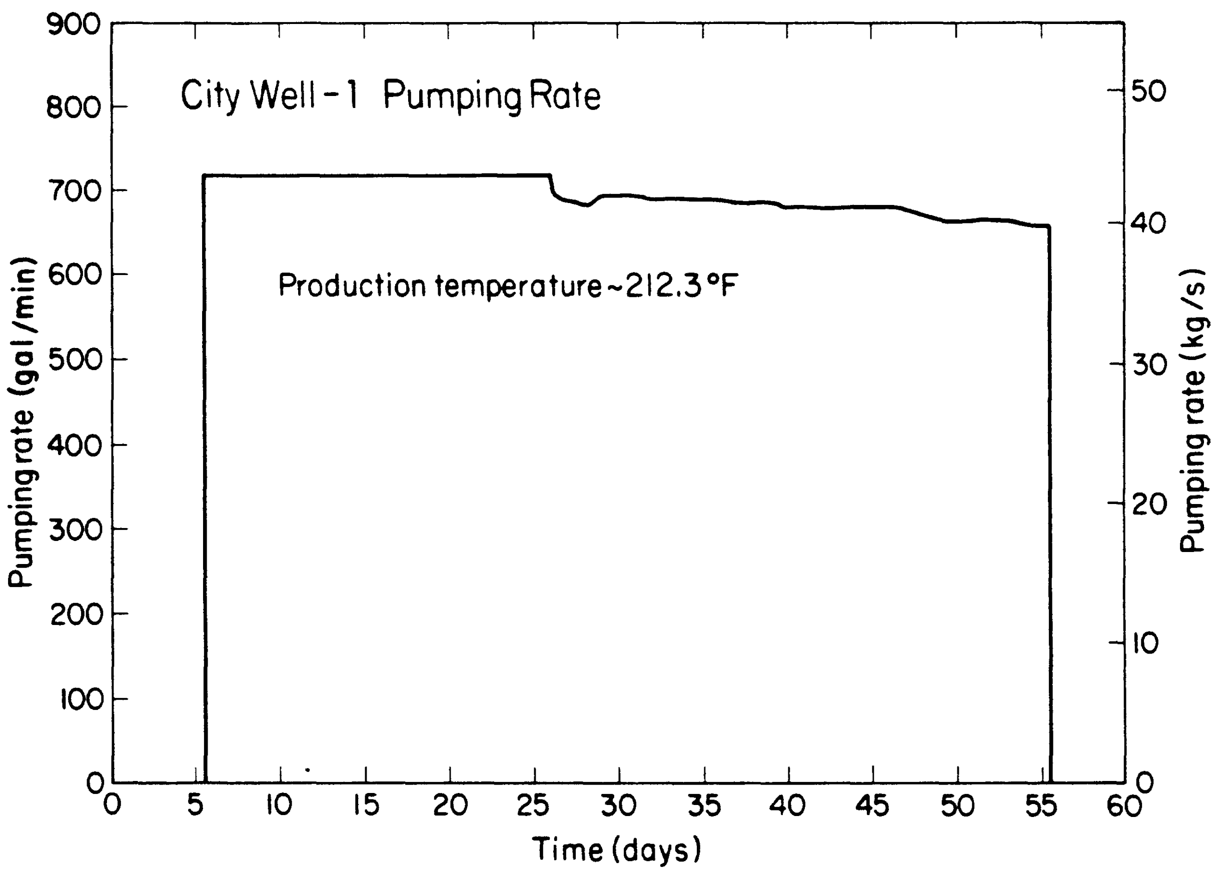

Figure 5-1. — Pumping rate from well CW-1, July 1 to August 29, 1983.

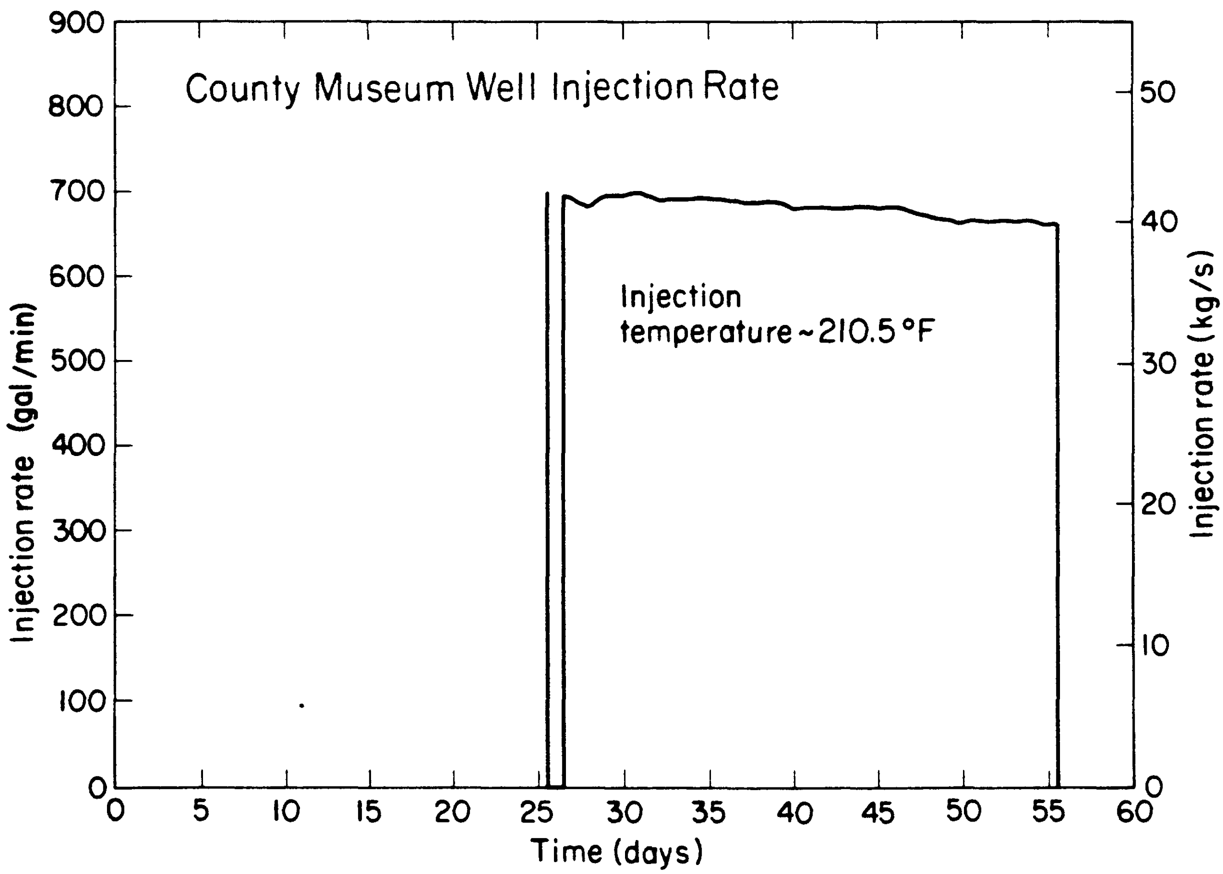

Figure 5-2. — Injection rate into the County Museum well, July 1 to August 29, 1983.

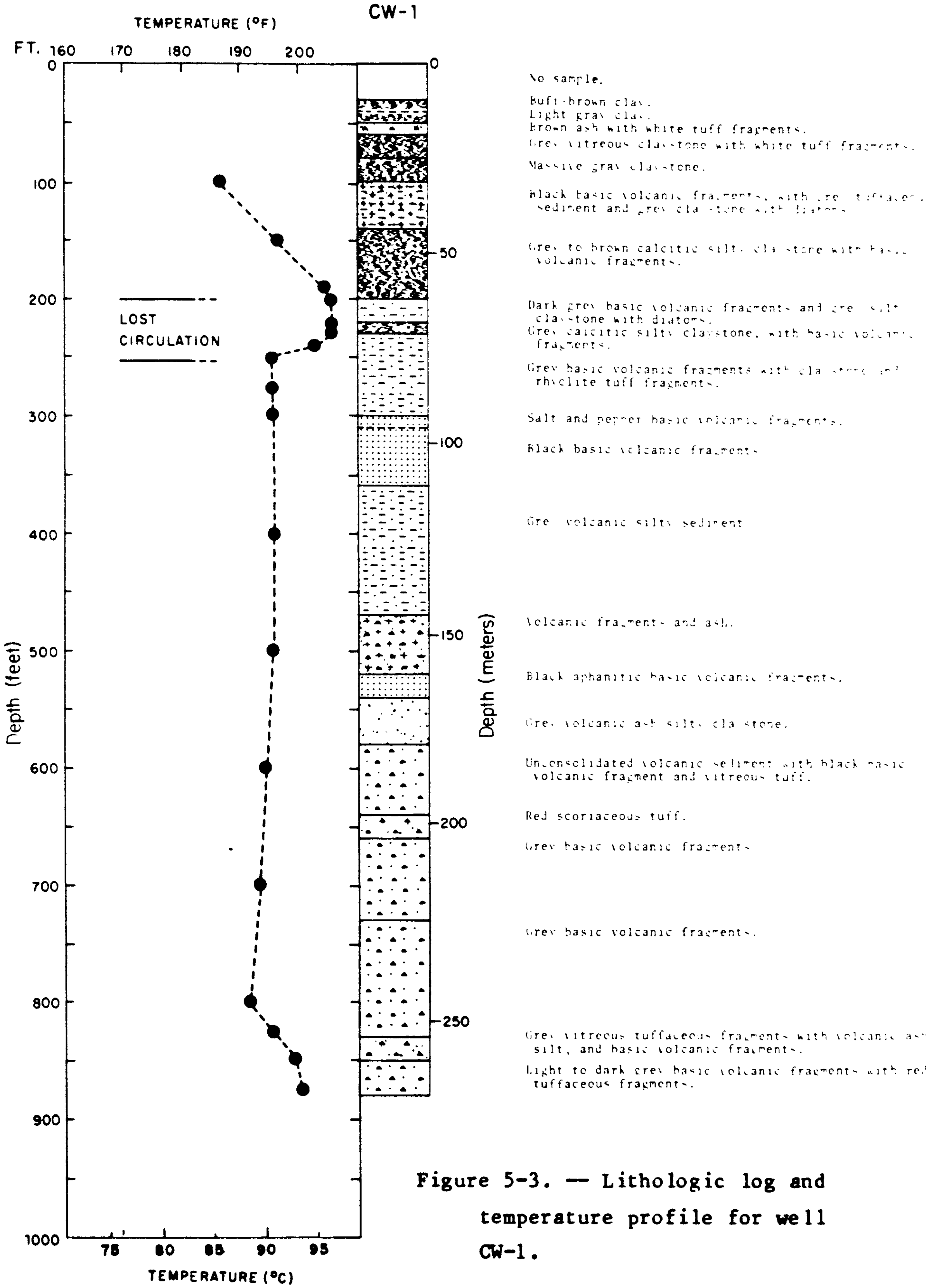

Figure 5-3. — Lithologic log and temperature profile for well CW-1.

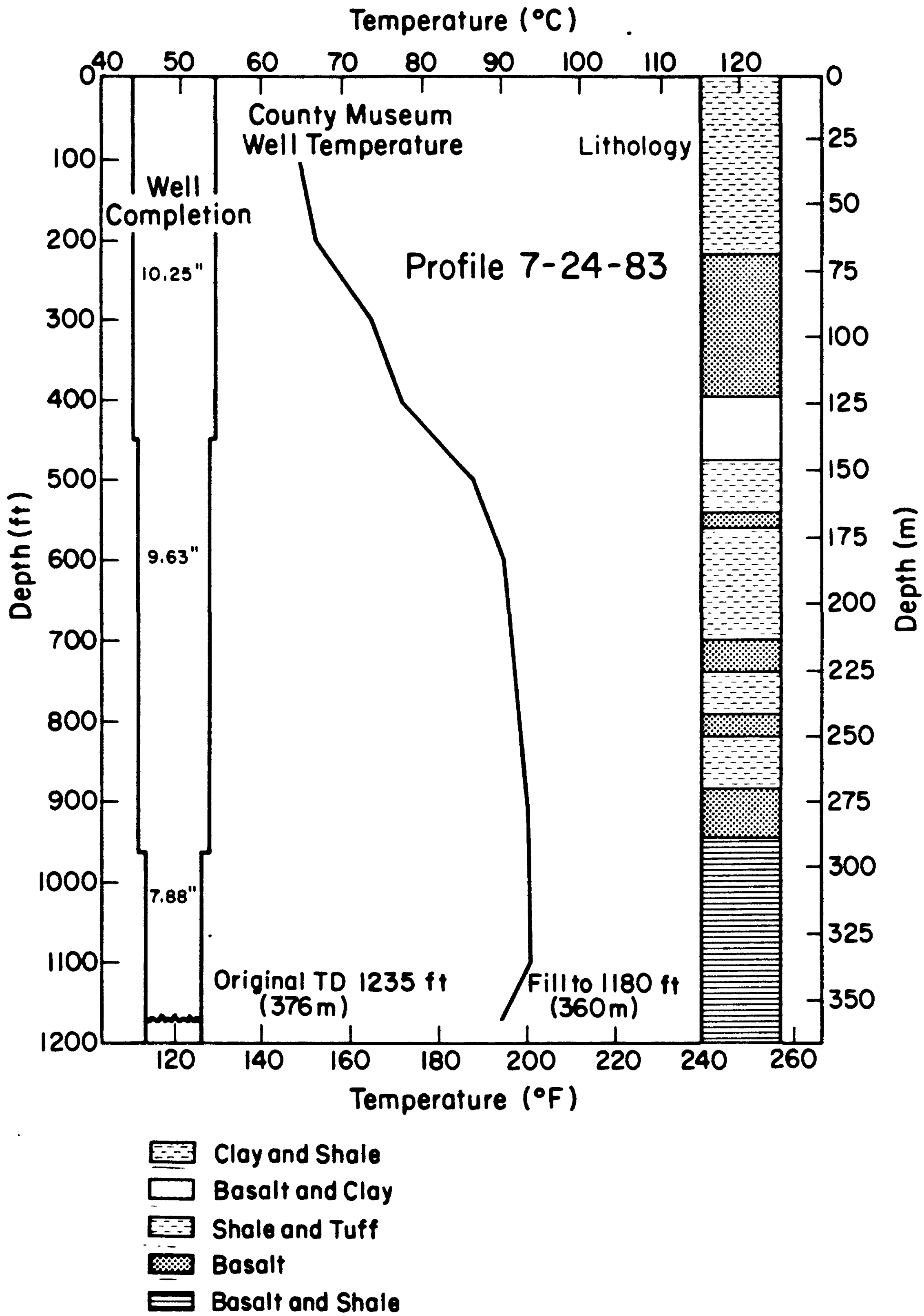

Figure 5-4. — Well completion, temperature profile, and lithology for the County Museum well.

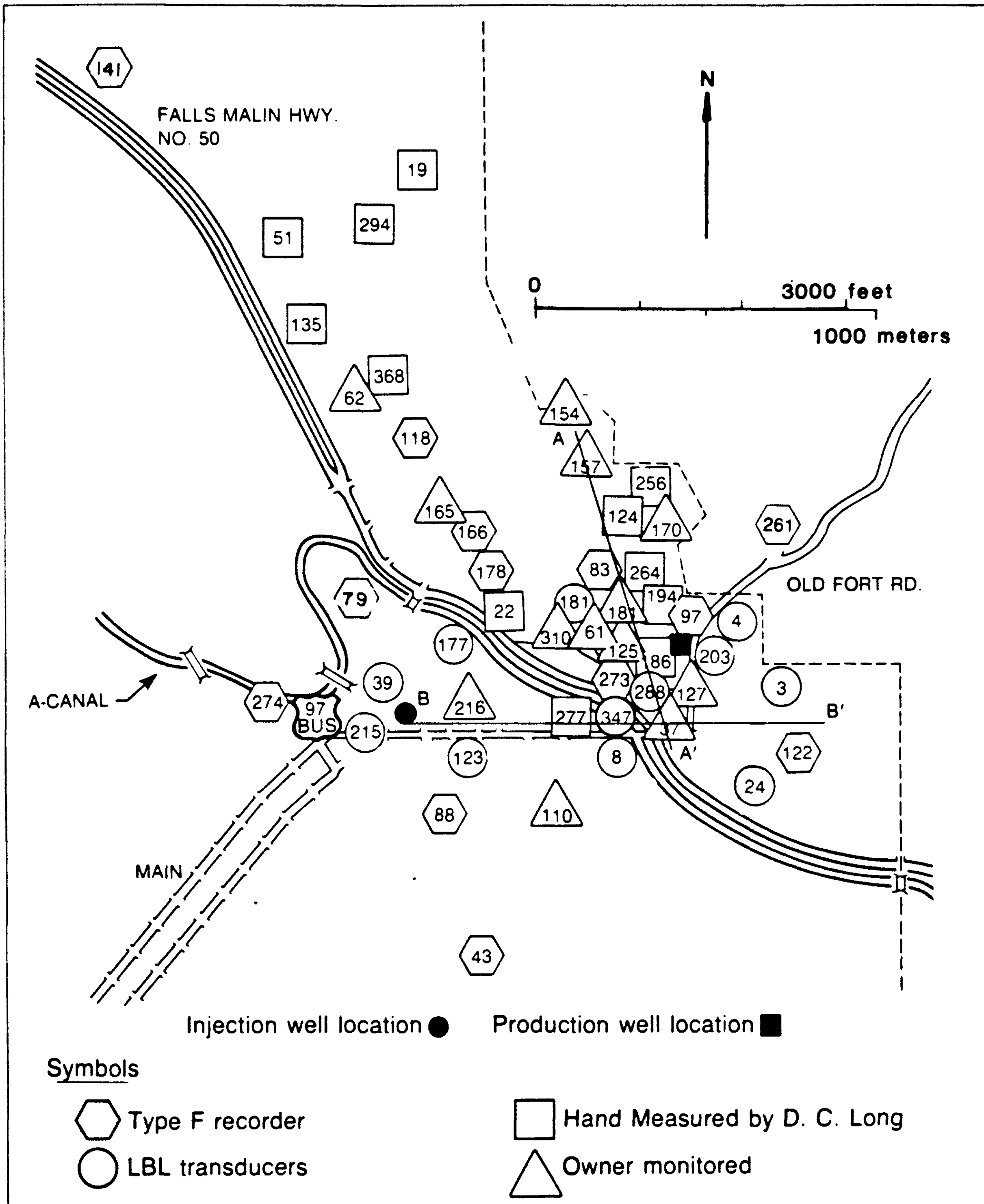

Figure 5-5. — Observation-well location map.

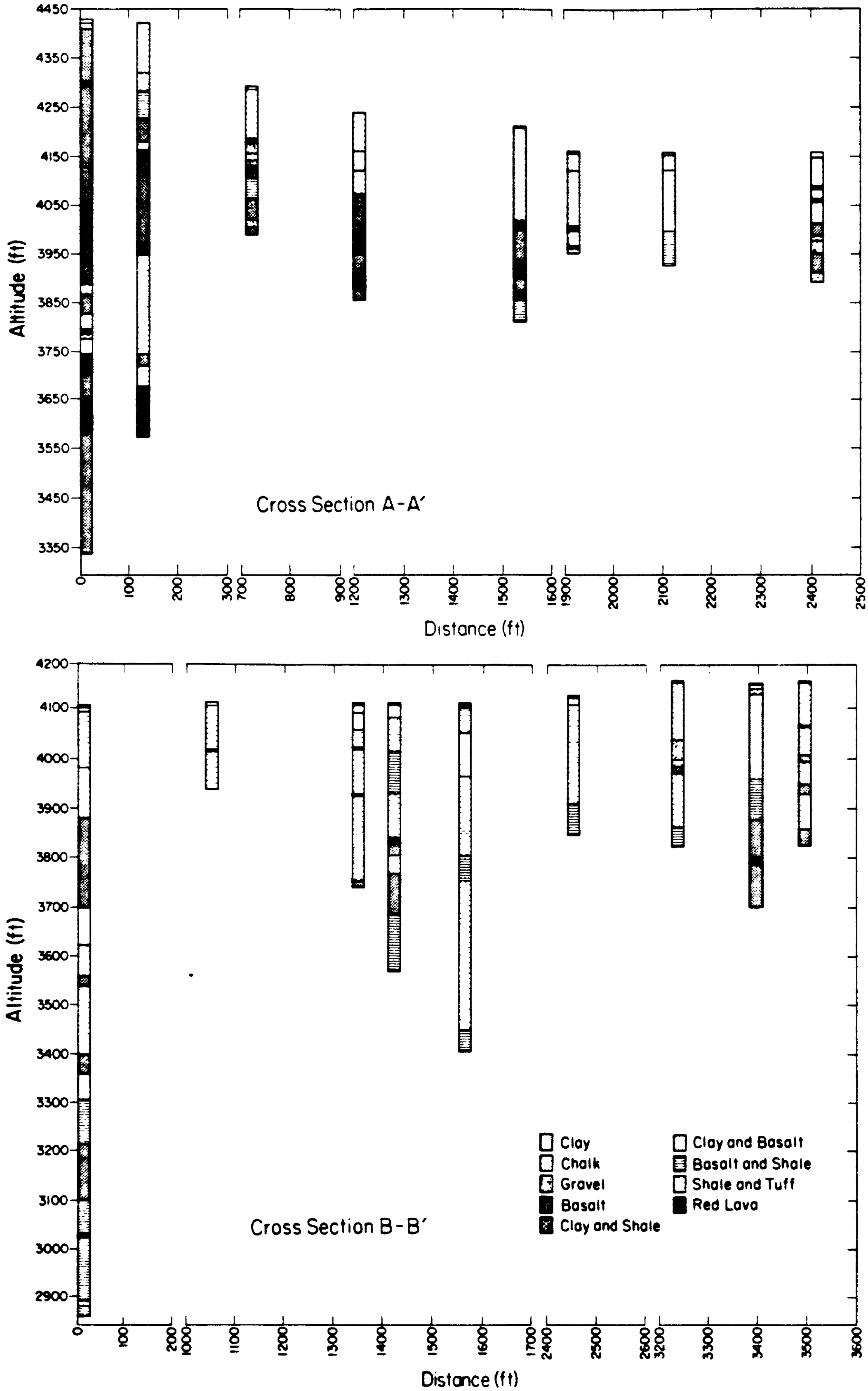

Figure 5-6. — Schematics of aquifer lithology in wells along Section A-A' and B-B'. See figure 5-5 for locations of cross sections.

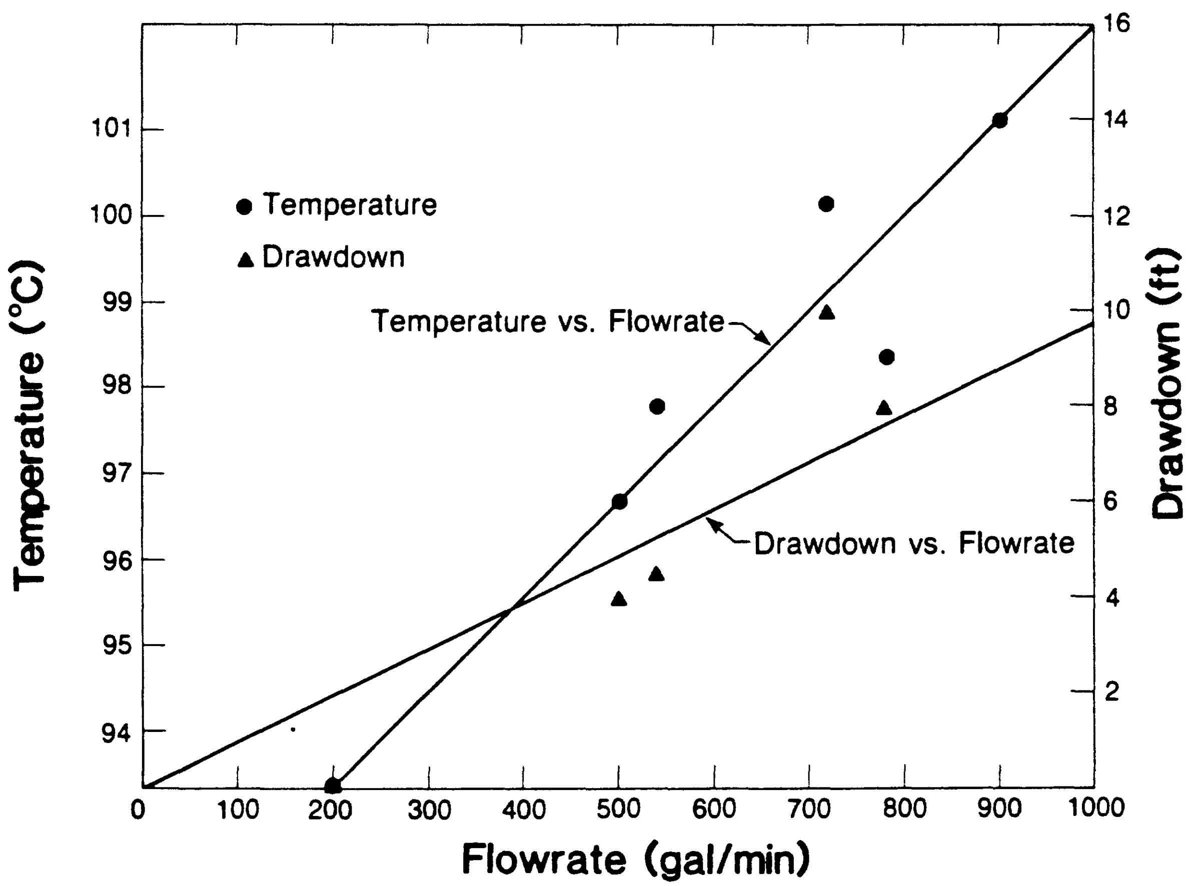

Figure 5-7. — Drawdown and wellhead temperatures vs. flowrate for well CW-1.

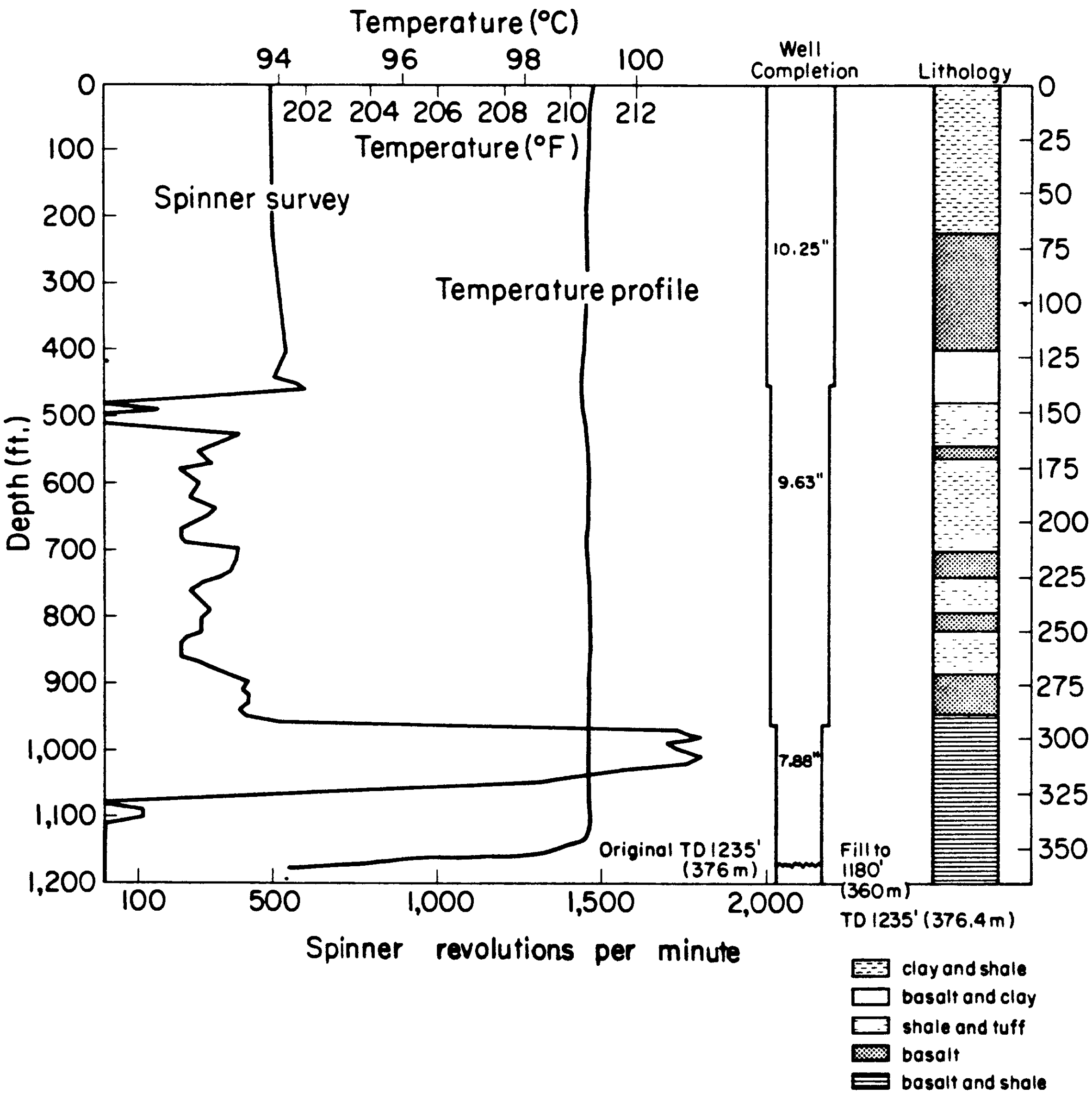

Figure 5-8. — Temperature profile and spinner survey obtained during injection into the County Museum well.

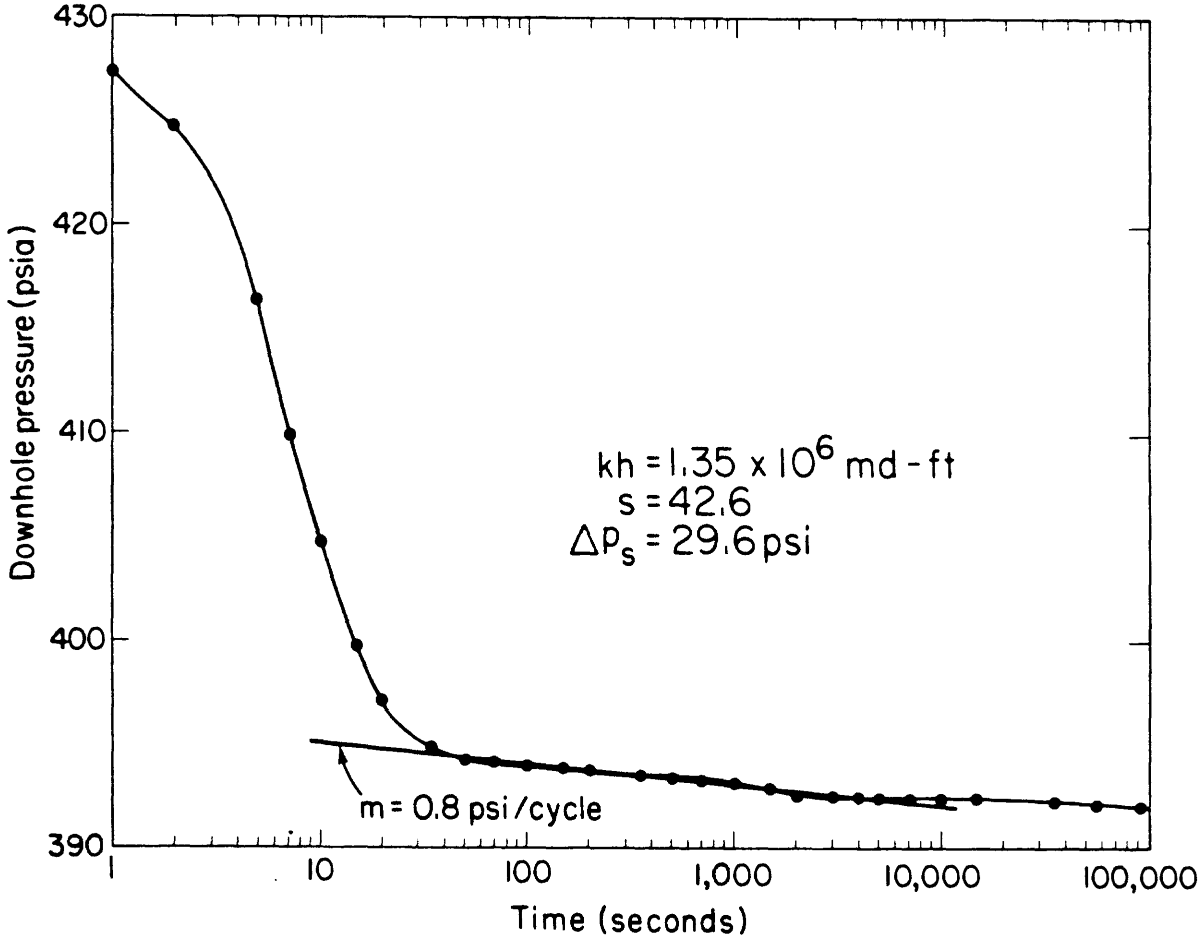

Figure 5-9. -- Semi-logarithmic plot of pressure-falloff data from the County Museum well.

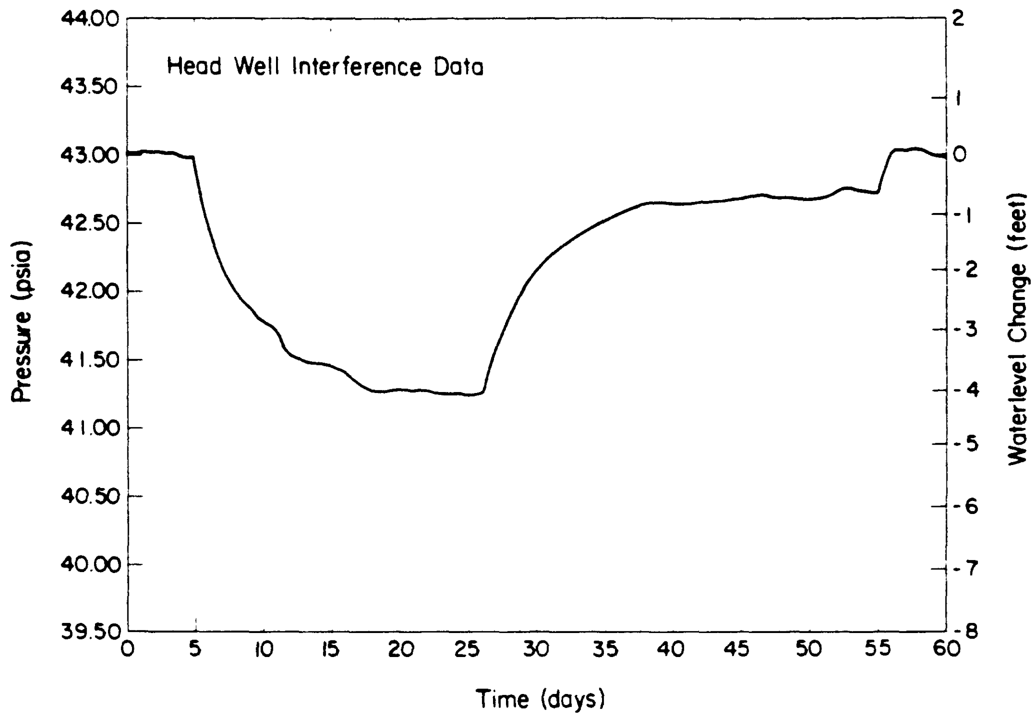

Figure 5-10. — Interference data from the Head well (No. 101), July 1 to August 29, 1983.

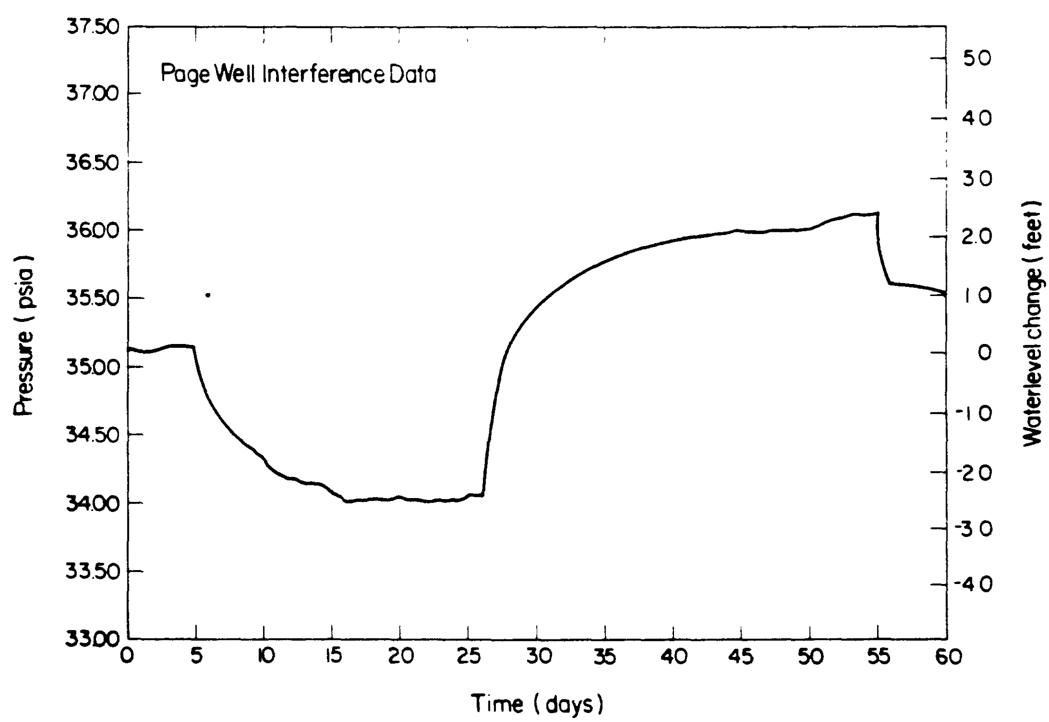

Figure 5-11. — Interference data from the Page well (No. 177), July 1 to August 29, 1983.

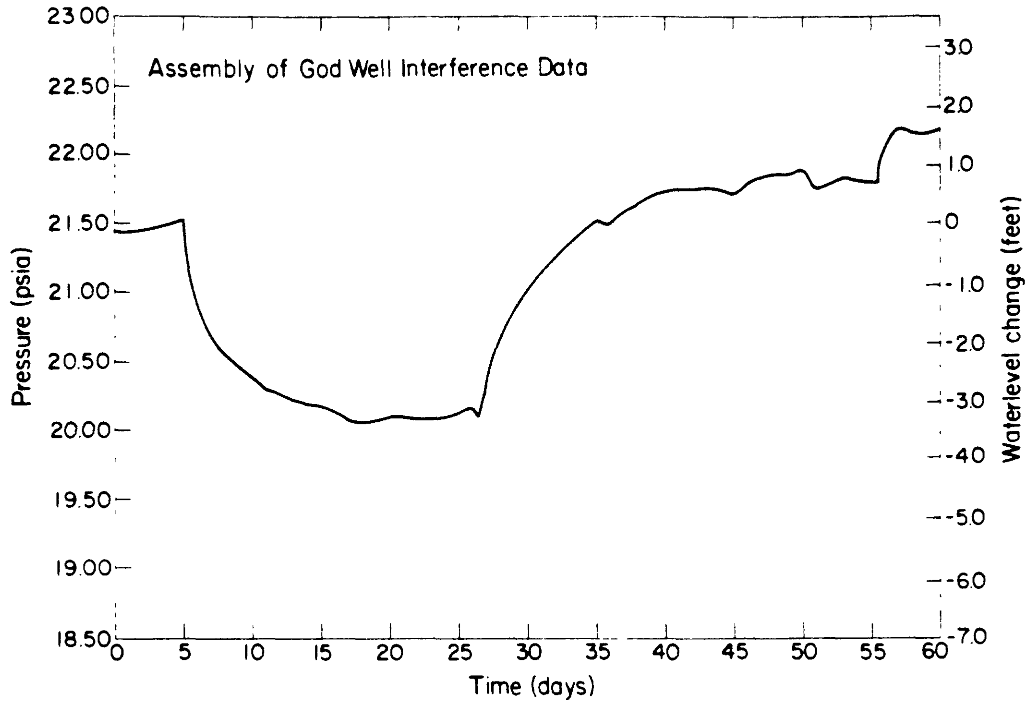

Figure 5-12. — Interference data from the Assembly of God well (No. 24), July 1 to August 29, 1983.

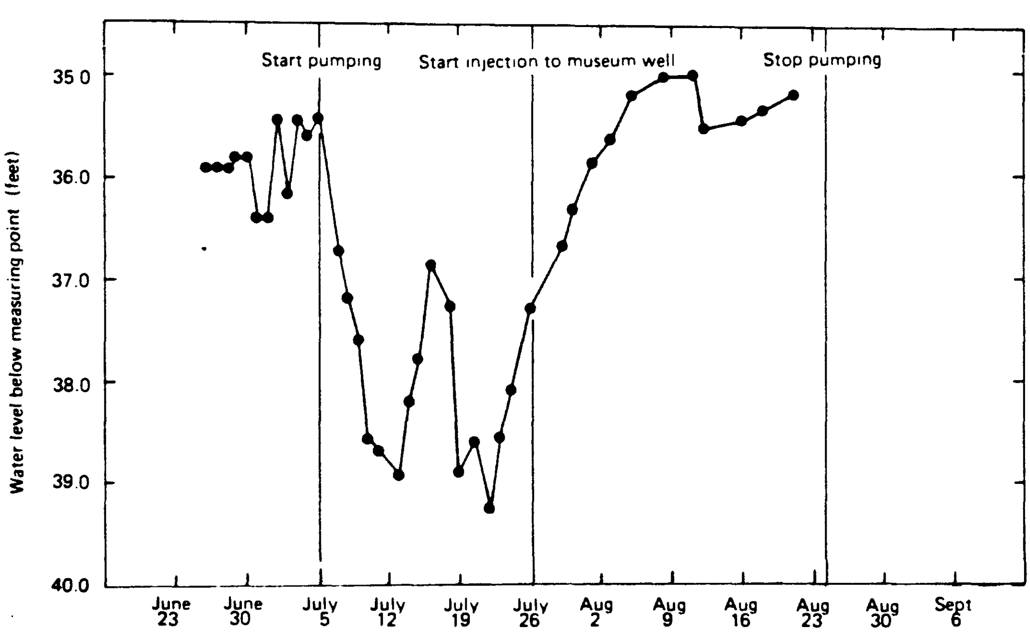

Figure 5-13. — Interference data from the Feeback well (No. 310), June 21 to August 20, 1983.

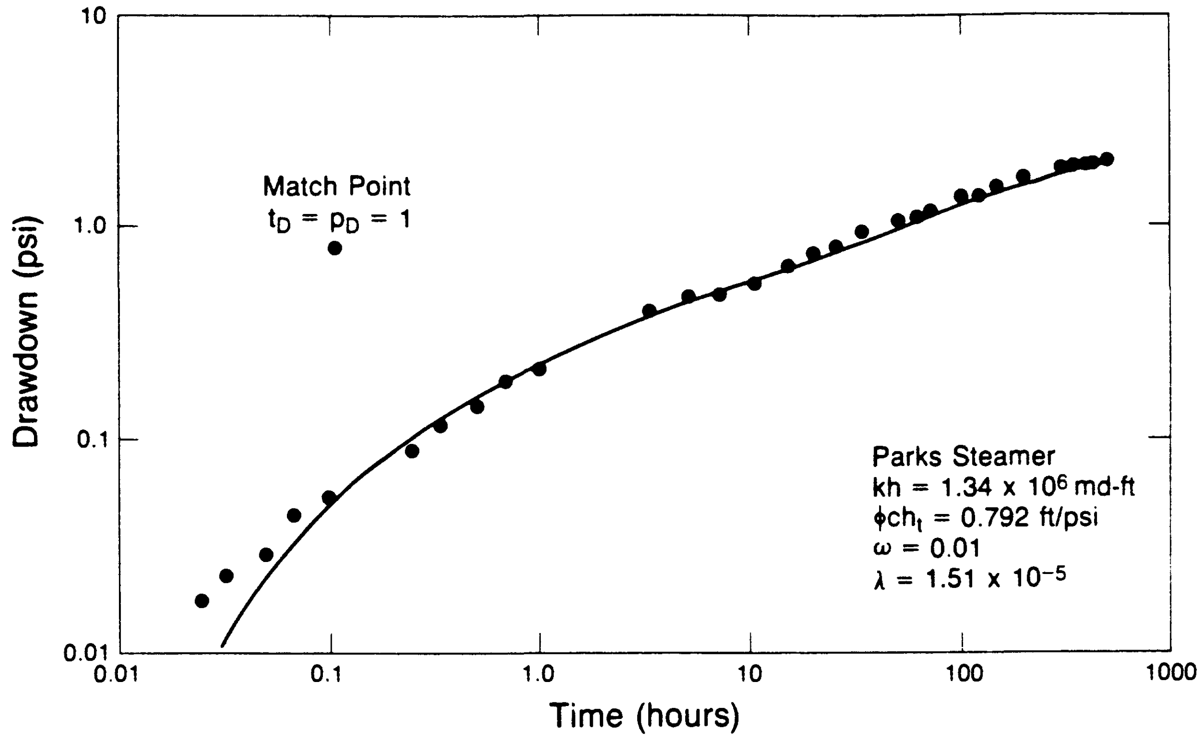

Figure 5-14. — Double-porosity type-curve match for the Parks Steamer well (No. 203) drawdown data.

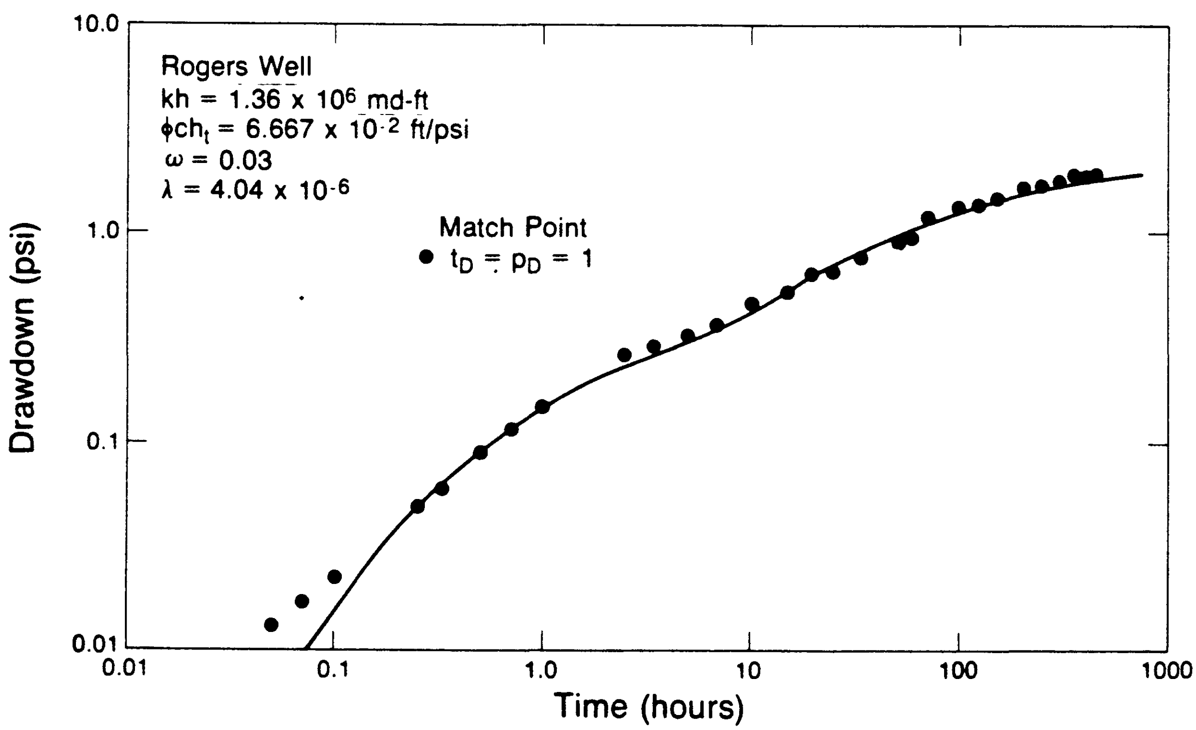

Figure 5-15. — Double-porosity type-curve match for the Rogers well (No. 200) drawdown data

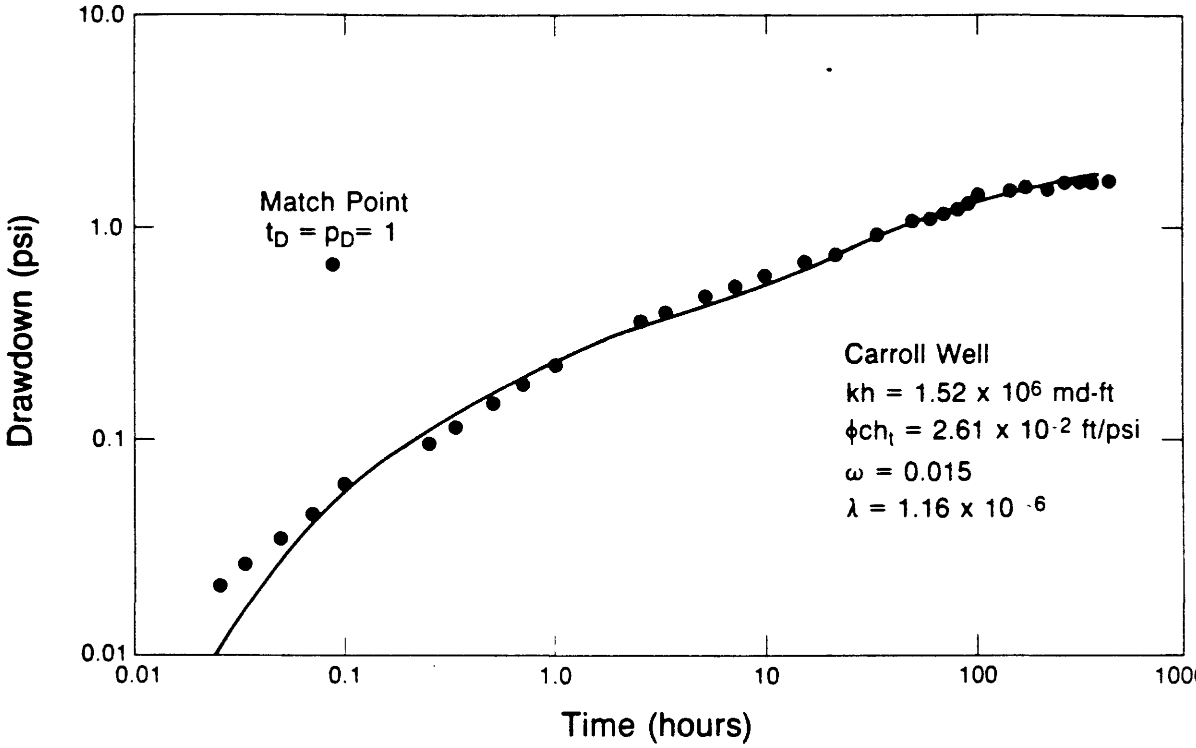

Figure 5-16. — Double-porosity type-curve match for the Carroll well (No. 3) drawdown data.

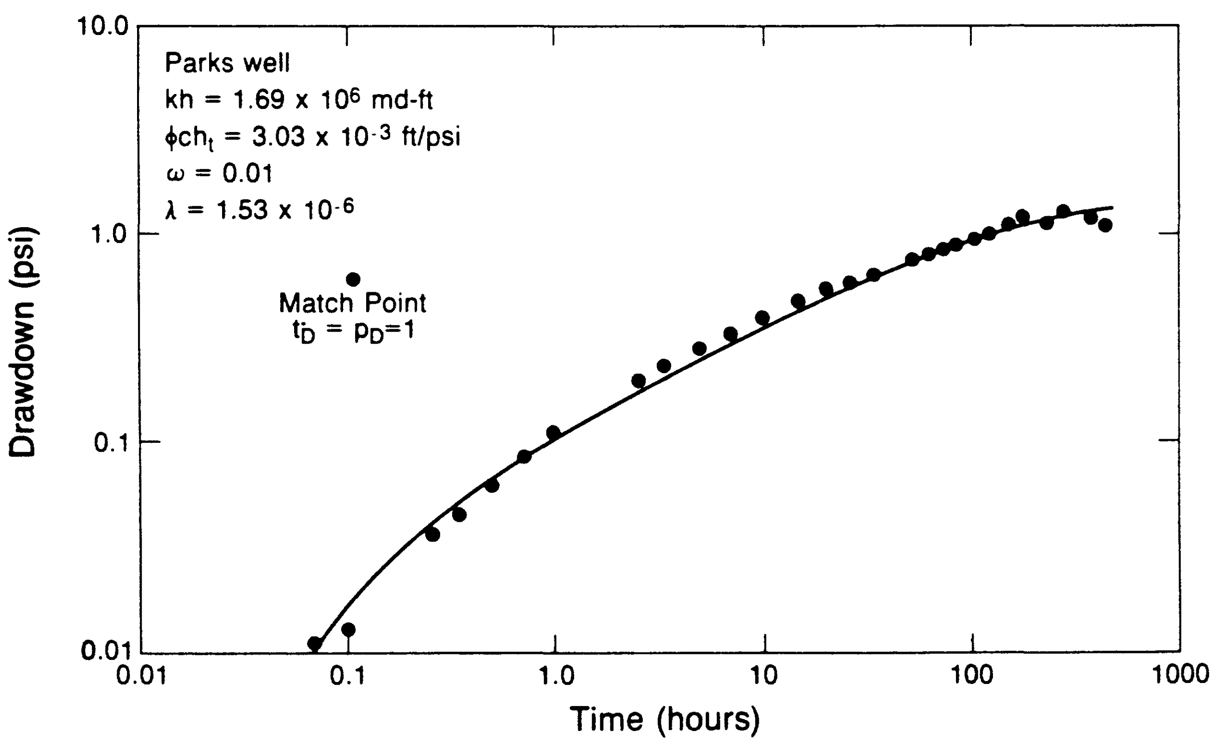

Figure 5-17. — Double-porosity type-curve match for the Parks well (No. 4) drawdown data.

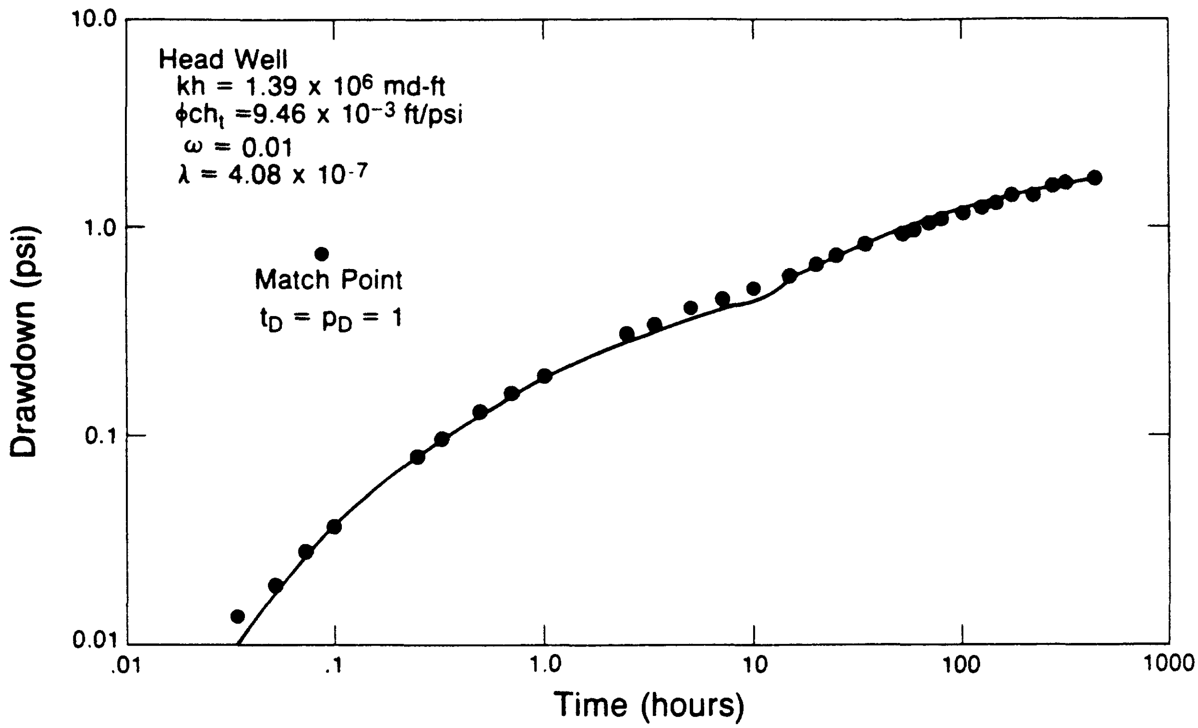

Figure 5-18. — Double-porosity type-curve match for the Head well (No. 101) drawdown data.

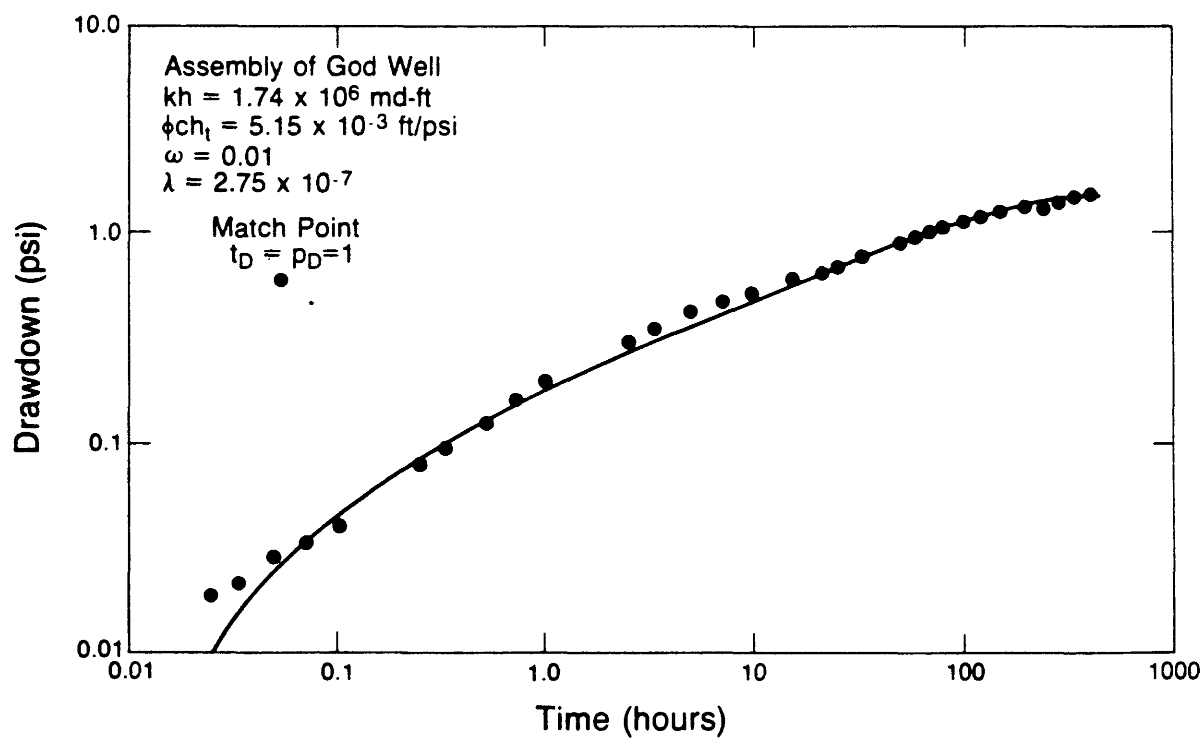

Figure 5-19. — Double-porosity type-curve match for the Assembly of God well (No. 24) drawdown data.

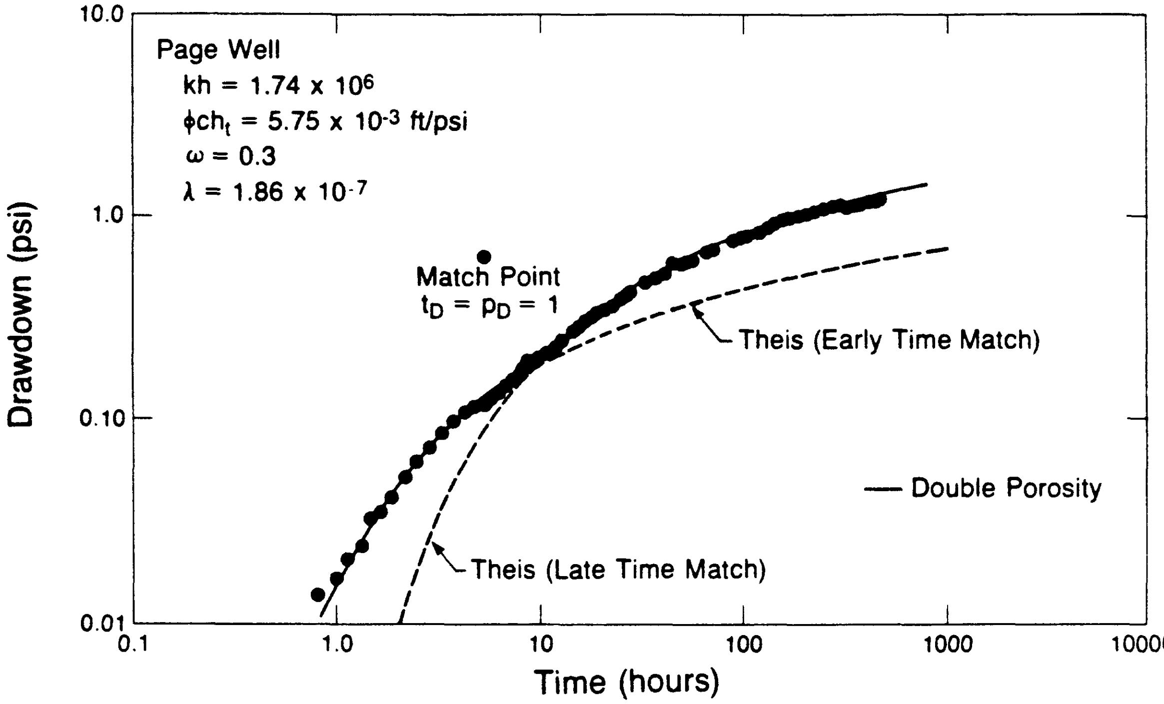

Figure 5-20. — Double-porosity type-curve match for the Page well (No. 177) drawdown data.

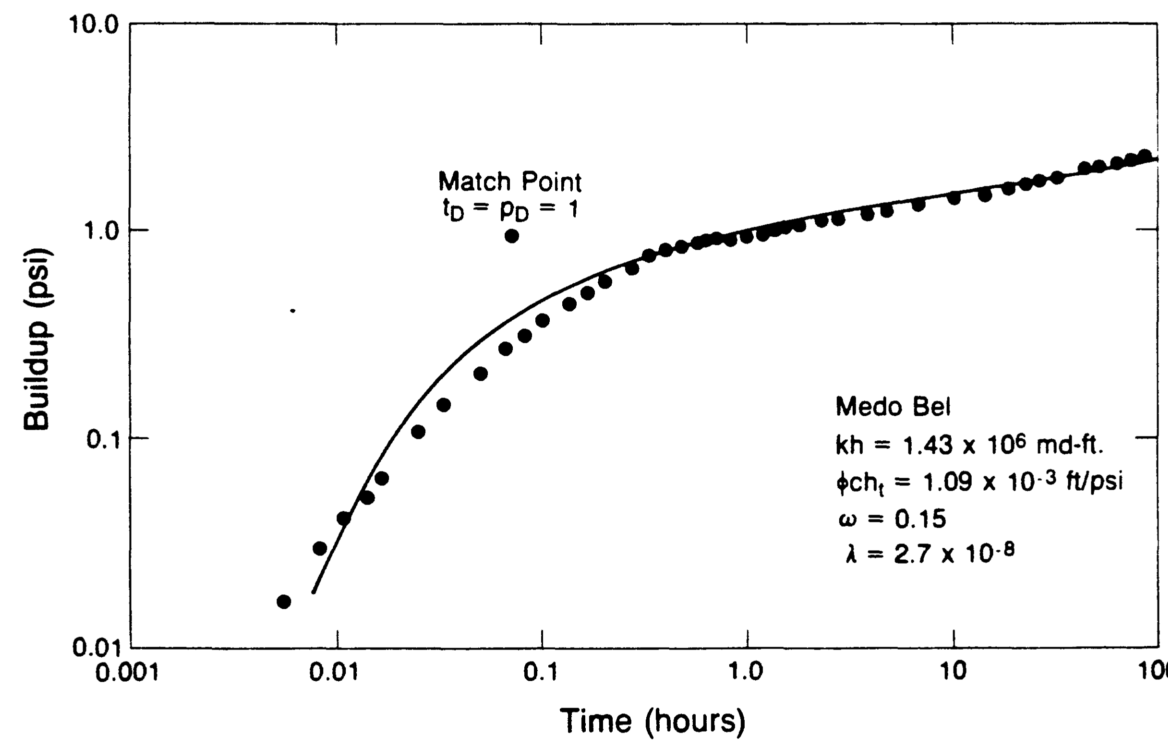

Figure 5-21. — Double-porosity type-curve match for the Medo-Bel well (No. 39) buildup data.

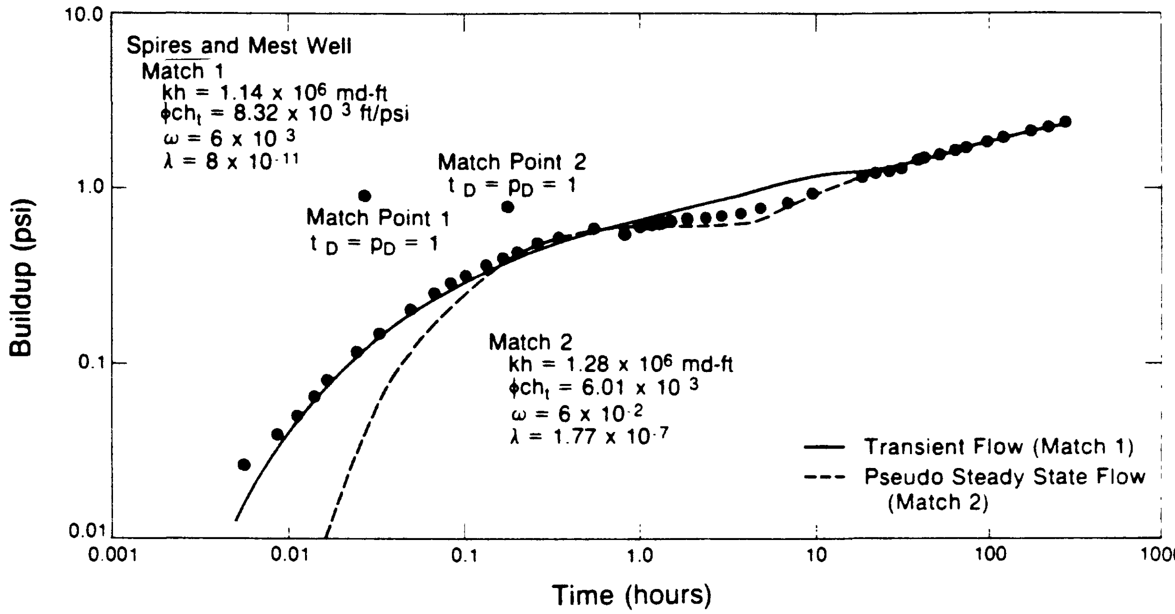

Figure 5-22. — Double-porosity type-curve match for the Spires and Mest well (No. 123) buildup.

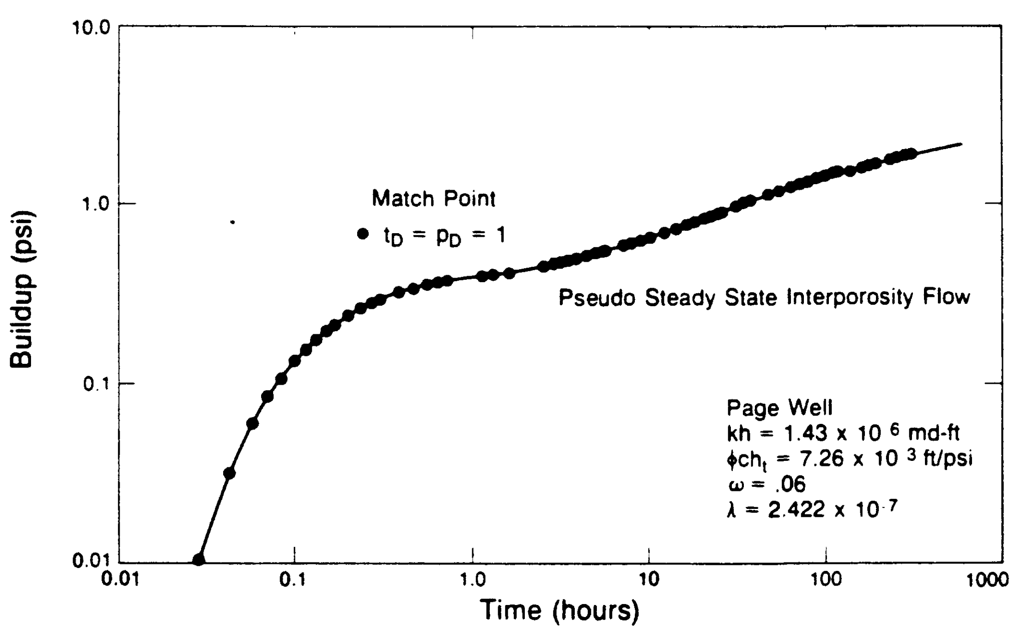

Figure 5-23. — Double-porosity type-curve match for the Page well (No. 177) buildup.

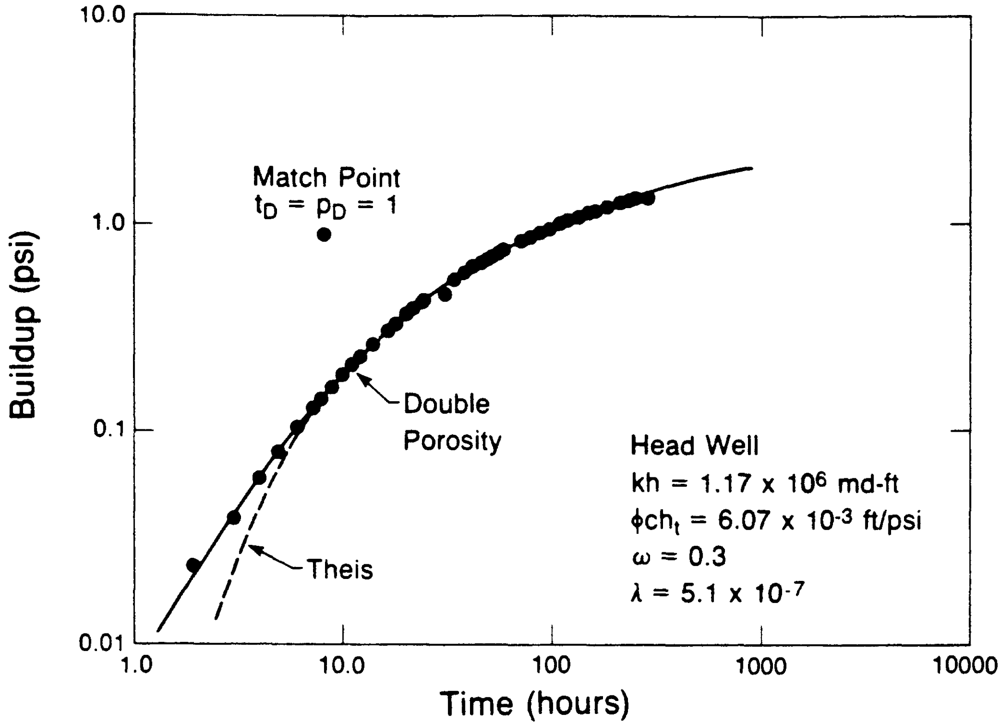

Figure 5-24. — Double-porosity type-curve match for the Head well (No. 101) buildup.

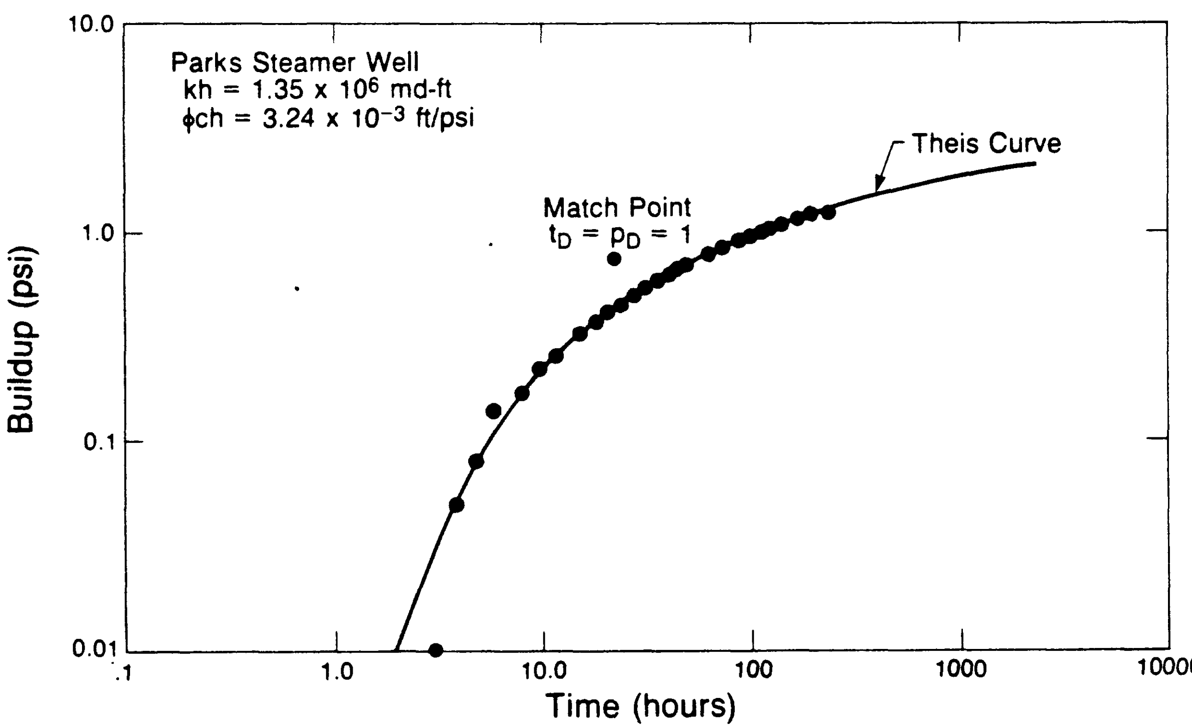

Figure 5-25. — Theis type-curve match for the Parks Steamer well (No. 203) buildup.

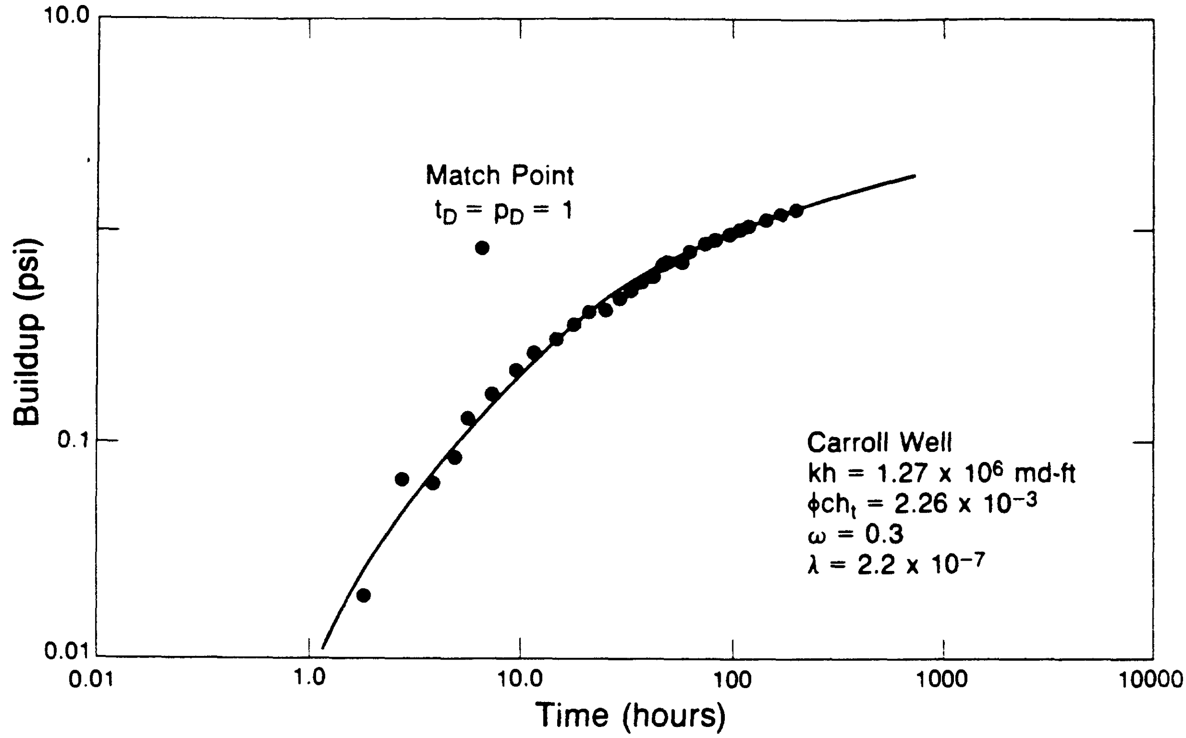

Figure 5-26. — Theis type-curve match for the Carroll well (No. 3) buildup.

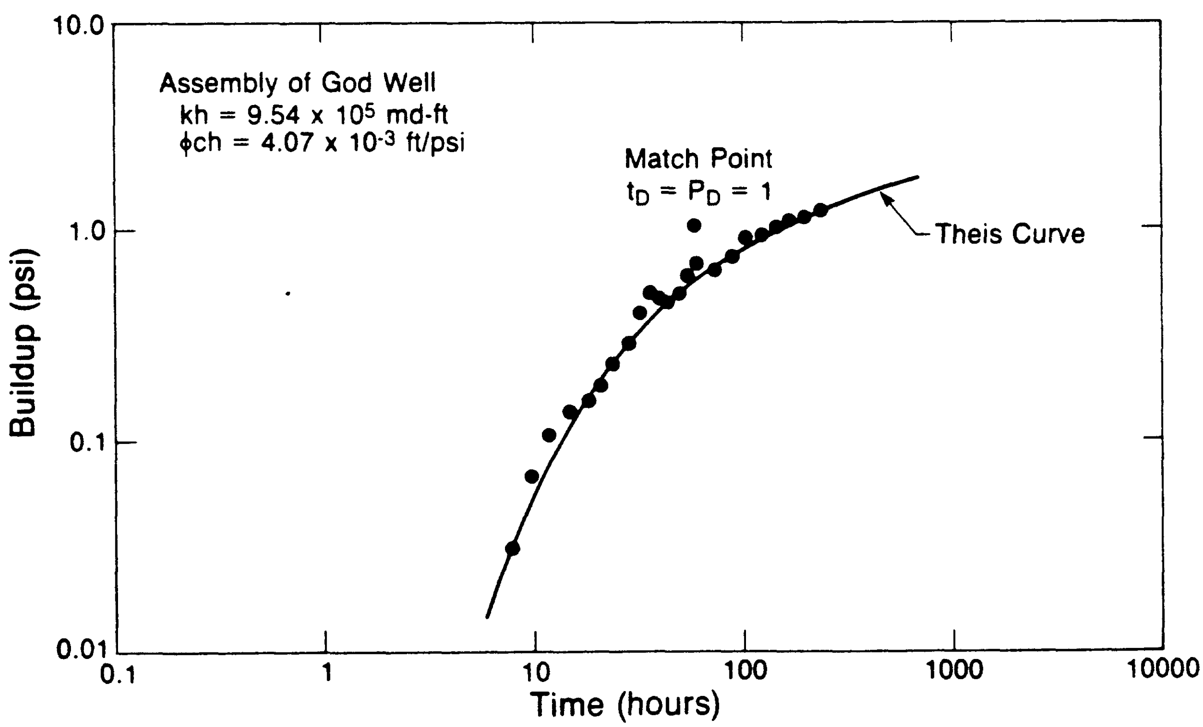

Figure 5-27. -- Theis type-curve match for the Assembly of God well (No. 24) buildup.

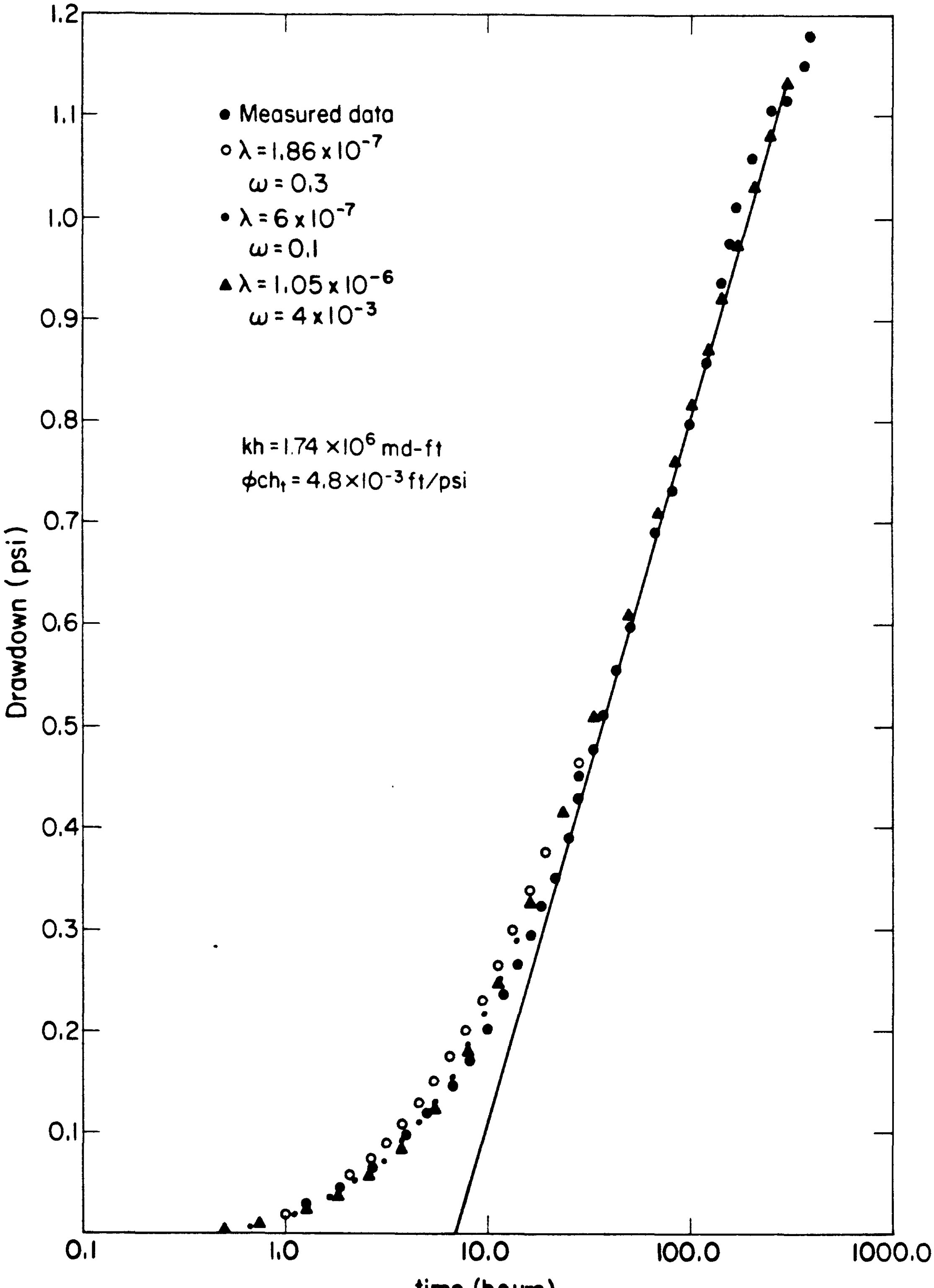

Figure 5-28. — Semi-logarithmic analysis of the Page well (No. 177) drawdown.

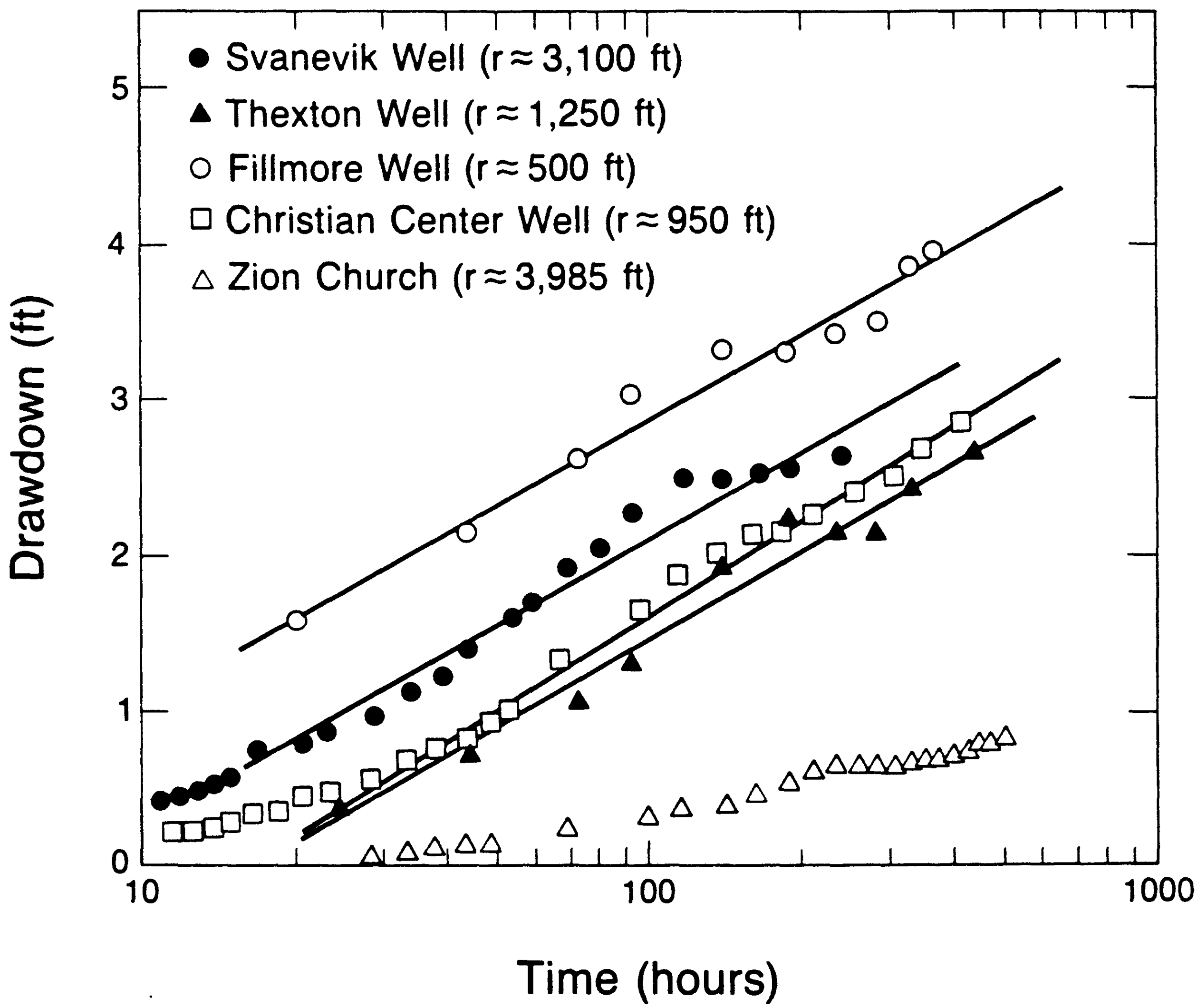

Figure 5-29. — Semi-logarithmic plot of drawdown data from the Svanevik, Thexton, Fillmore, Christian Center and Zion Church wells (No.'s 118, 22, 186, 126, and 274).

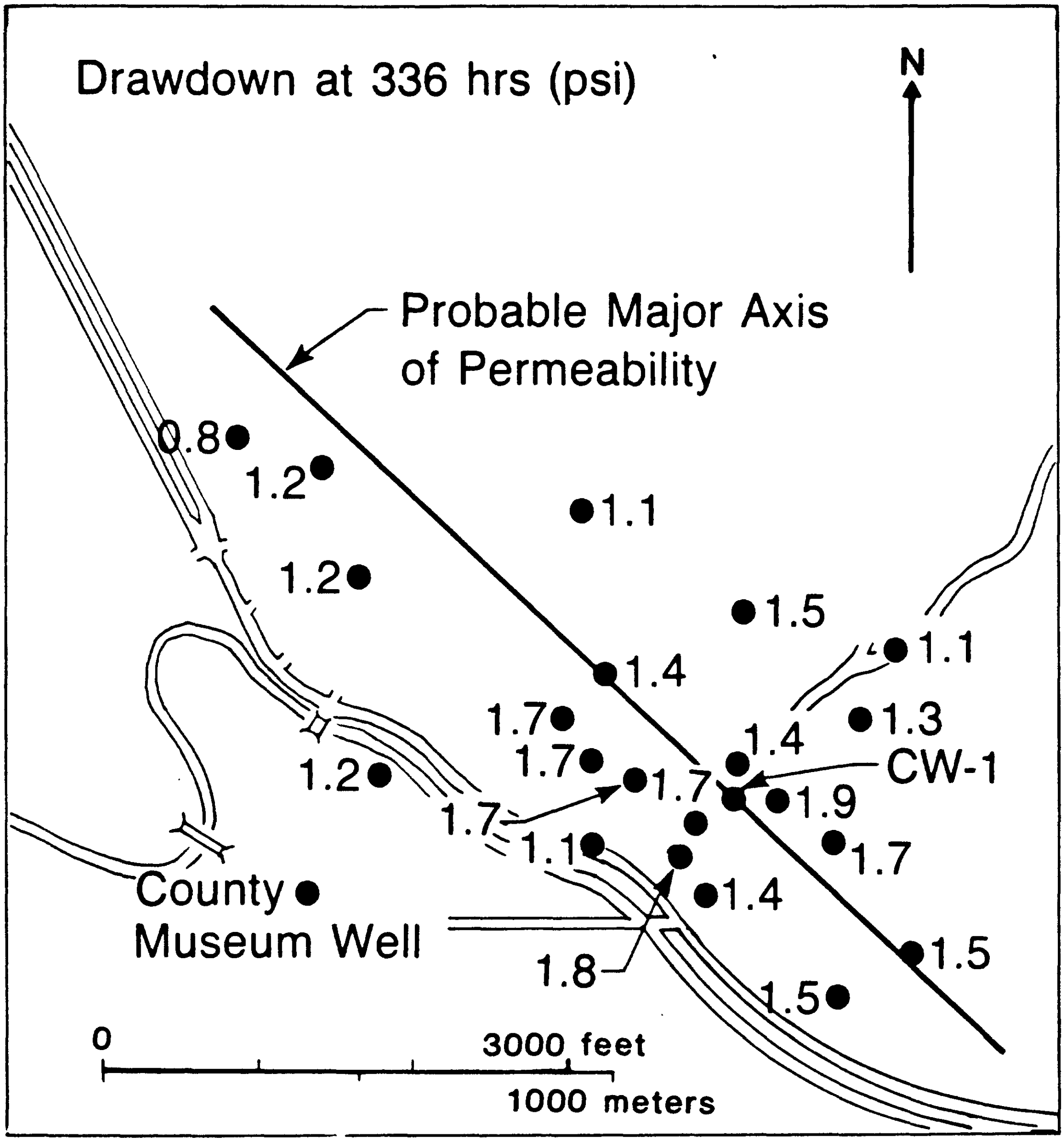

Figure 5-30. — Pressure drawdown in the geothermal aquifer after 336 hours of pumping at a rate of 720 gal/min.

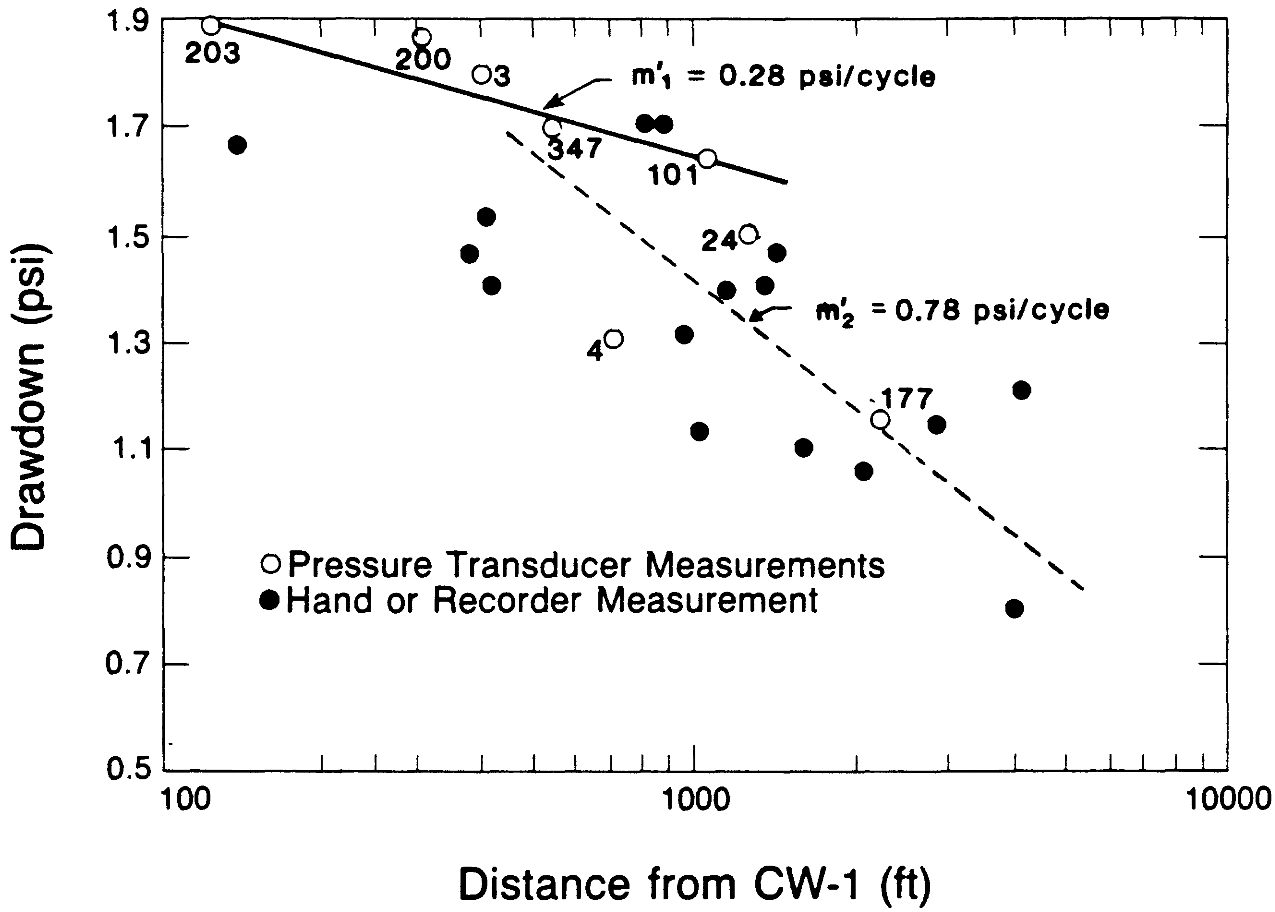

Figure 5-31. — Pressure drawdown vs. distance for 24 observation wells after 336 hours of pumping at a rate of 720 gal/min.

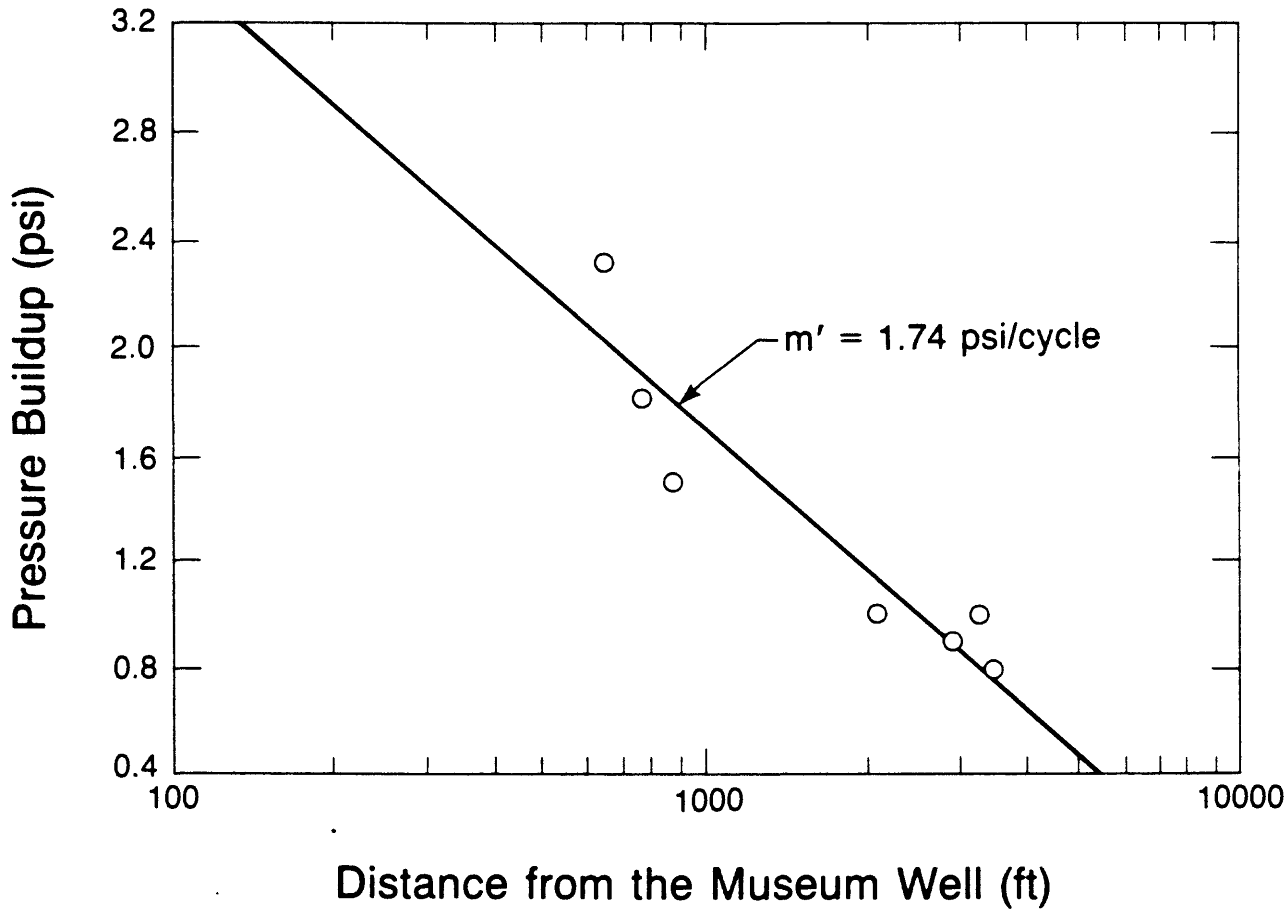

Figure 5-32. — Pressure drawdown vs. distance for 7 observation wells measured by transducers after 300 hours of injection at a rate of approximately 700 gal/min.

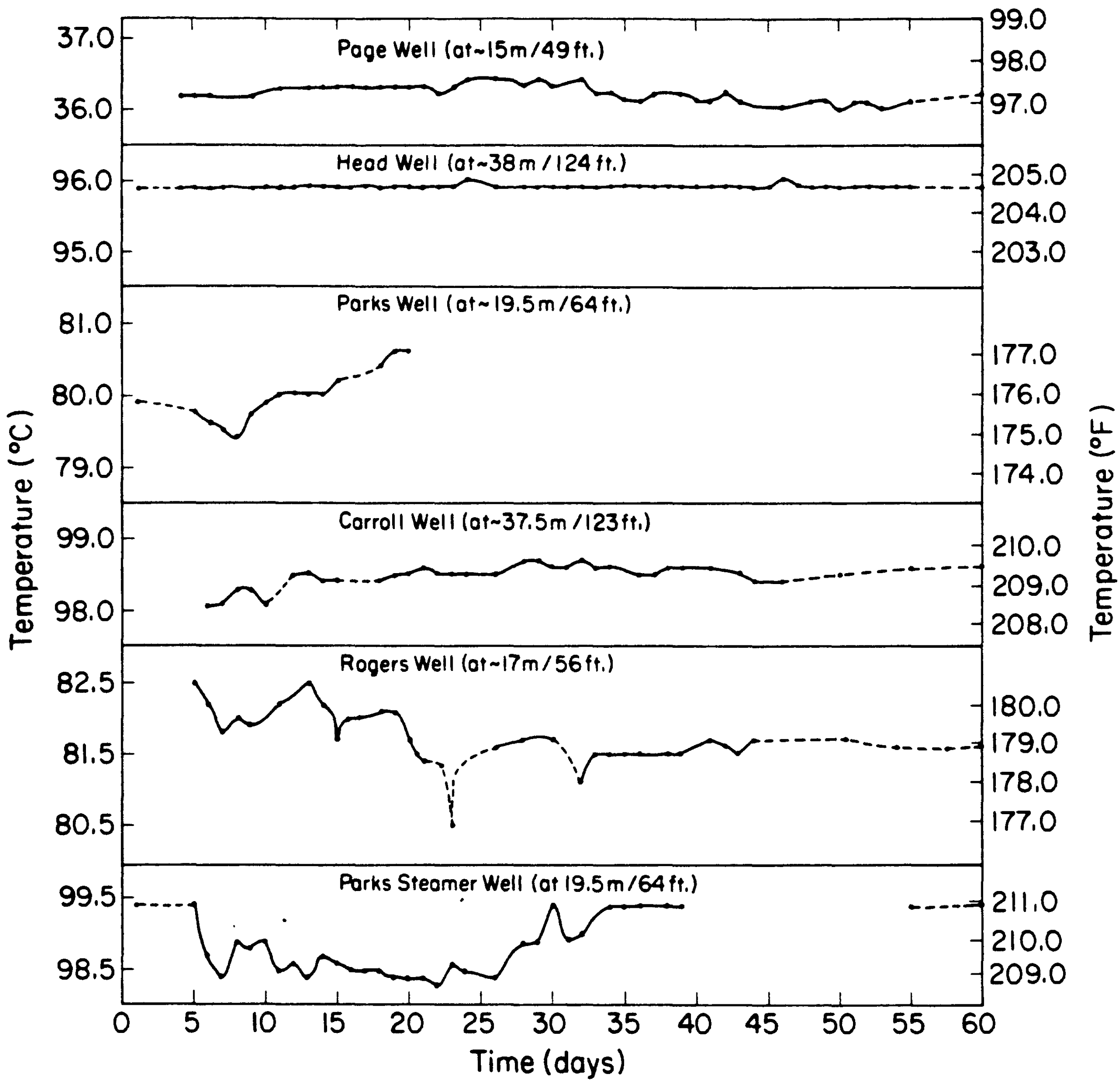

Figure 5-33. — Downhole temperature data from Page, Head, Parks, Carroll, Rogers, and Parks Steamer wells (No.'s 177, 101, 4, 3, 200, and 203). Time in days from July 1, 1983.

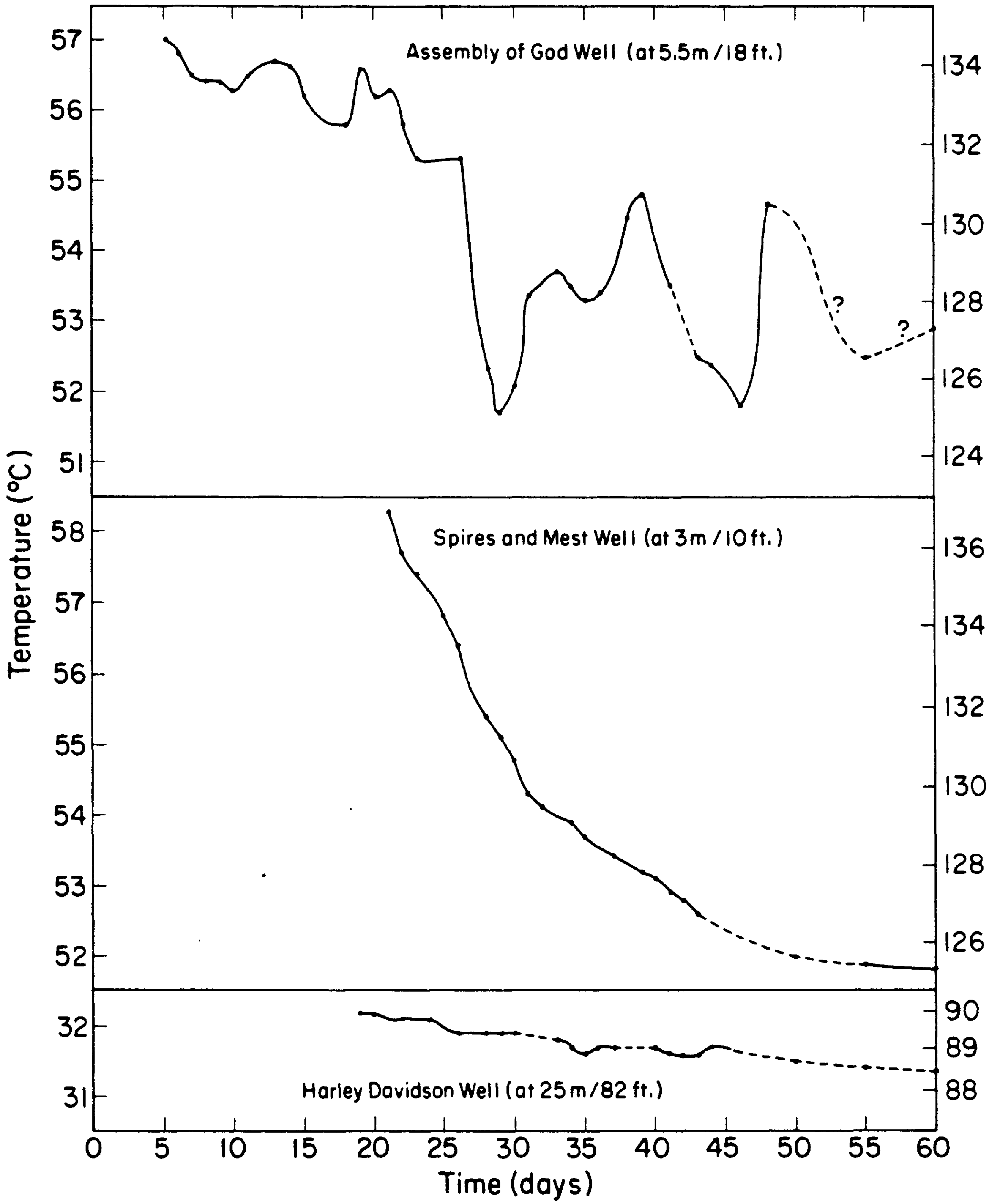

Figure 5-34. -- Downhole temperature data from Assembly of God, Spires and Mest, and Harley Davidson wells (No. 24, 123, and 215). Time in days from July 1, 1983.

CHAPTER 6. AQUIFER RESPONSE TO PUMPING AND REINJECTION

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Hypothetical Development Schemes

Although not the primary objective of this study, the ultimate goal of hydrologic testing is to predict the aquifer response to possible development plans. To this end, prediction of the water-level changes in respose to two resource development schemes are presented in this chapter. Temperature changes due to reinjection or reservoir fluid mixing are not considered here for the reason that this topic requires the availability of a complete analysis of the tracer studies. Both of the development schemes assume that well CW-1 and the County Museum well are used for pumping and reinjection. Flowrates are chosen to approximate the anticipated heat loads of the currently proposed system for heating 14 downtown buildings (C. Culver, personal commun., 1983). In each case the effects of a full year of pumping and reinjection are considered. Some liberties were taken in choosing the flowrates in order to investigate the effect of a one-month period of pumping at the peak rate (Scheme 1) and the effect of a minimum pumping rate of 200 gal/min (Scheme 2). Pumping rates for these two cases are shown in figure 6-1. For both of these cases 100 percent of the pumped fluid is reinjected.

Computational Reservoir Model

Analysis of the aquifer test data indicates that the geothermal system behaves in a remarkably uniform maner, given the complexity of the system. Therefore, average values for the hydrologic properties [kh, $(\phi ch)_t$] can be used to predict the water-level drawdown and buildup in response to pumping and reinjection. Examination of the pressure-transient data also shows that after approximately 10 hours of pumping or reinjection the response is no longer affected by the double-porosity nature of the system. Therefore, for the purpose of these calculations the double-porosity effects are neglected. Analysis of the data also indicates that the system is not influenced by the presence of hydrologic flow boundaries within a radius of approximately 3.5 miles of the center of the geothermal anomaly. Therefore, the system is assumed to be effectively infinite. Steady-state analysis of the data indicates that the permeability of the system is anisotropic and that a region of higher permeability surrounds the pumped well. Both these factors influence the distribution of drawdown in the aquifer. However, rigorous evaluation of these factors is not complete. For this reason, and because they are of only secondary importance, they have been neglected for the purpose of predicting the aquifer response to pumping and reinjection.

The aquifer model used for these calculations can be described as follows.

(1). The geothermal aquifer is areally infinite, horizontal, and bounded above and below by impermeable strata.

(2). The reservoir is homogeneous, isotropic and isothermal.

(3). The pore spaces of the reservoir are fully saturated with liquid water.

(4). The pumped and injection well both completely penetrate the reservoir. The reservoir properties kh and $(\phi ch)_t$ used for these calculations are obtained from the average values calculated using the type-curve matches. The water viscosity is evaluated at 95°C. The values are summarized as follows.

$$kh = 1.4 \times 10^{6} \text{ md-ft}$$
$$(\phi ch)_t = 5.0 \times 10^{-3} \text{ ft/psi}$$
$$\text{viscosity} = 0.3 \text{ centipoise (cp)}$$

Mathematical Model

In order to calculate the drawdown and buildup in response to pumping and reinjection the computer code VARFLOW is used (IDO, 1982). The code accounts for multiple production and injection wells using the principle of superposition. Water-level changes in response to varying flowrates are calculated using an algorithm developed by McEdwards and Tsang, 1978. This technique allows for the superposition of an arbitary number of linearly varying flowrates. Water-level drawdowns and buildups in response to production and injection are evaluated at 60 locations in the reservoir.

Results

The calculated water-level changes at the end of the peak flow (700 gal/min) period for cases 1 and 2 are shown in figure 6-2. Water-level changes are identical for both of these cases because the high permeability of the system results in the rapid equilibration of the pressure transients. This is consistent with the rapid pressure equilibration that occurred when CW-1 and the County Museum well were shut-in following the pump test. At the peak flow of 700 gal/min, the calculated drawdowns reach a maximum value of 4 ft for wells within a 200-ft radius of the pumped well. Actual drawdowns in this region will be less due to the high-permeability region that surrounds the well. Note that the locations of the observation wells are also shown in figure 6-2. Most of these wells will have water-level changes between +1 and -2 ft. The water-level buildup surrounding the County Museum well will be a mirror image of the drawdown around CW-1. Temporal variations in the water-level for 4 well locations (indicated by solid black dots in figure 6-2) are shown in figure 6-3 for flowrate scheme 2. If 100 percent of the pumped water is reinjected into the geothermal aquifer, the cumulative effects of water withdrawal need not be considered. This, however, is not true for the temperature changes in response to reinjection.

Water-level changes corresponding to the hypothetical flowrates in March are contoured in figures 6-4 and 6-5. For flowrate scheme 1, drawdowns and buildups throughout the aquifer (with the exception of the active wells) are less than 1 ft. In comparison, drawdowns and buildups of 2.5 ft are calculated for flowrate scheme 2. The difference results from much higher flowrate for case 2 (400 gal/min versus 140 gal/min).

Summary

The predictions in this section were made primarily to demonstrate the methodology that can be used to predict the response of the aquifer to pumping and reinjection. Meaningful predictions of the impact of operating the district heating system can be made only after the required pumping rates have been established. However, several conclusions are demonstrated by these calculations. Clearly, the lower the pumping rate, the smaller the impact will be on the existing users. Although this statement does not imply that pumping and reinjection are necessarily harmful, it is also clear that conservation measures will greatly enhance the ability of the system to satisfy the heating demands with the least possibility of harming existing users. For instance, continuous pumping at some minimum rate causes unneccessary drawdowns if the heating demand requires far less hot water. A second conclusion resulting from these calculations is that operation of the district heating system at less than peak capacity (< 700 gal/min) will result in drawdowns smaller than those observed during the 1983 aquifer test.

When data become available from which the required rates for the district heating system can be determined, additional calculations can be made to precisely predict the impact of operating the system. Similar calculations can determine the impact of any proposed resource development scheme, thereby making possible an optimal development plan that will allow efficient use of the resource and ensure its longevity.

References

IDO, 1982, Low-to-moderate temperature hydrothermal reservoir engineering handbook: U.S. Department of Energy, Idaho Operations Office, IDO-10099 v. II.

McEdwards, D. G., and Tsang, C. F., 1978, Variable rate multiple well testing analysis: in Proceeding of the Invitational Well Testing Symposium, Oct. 19-21, 1977, Berkeley, California, Lawrence Berkeley Laboratory Report LBL-10907.

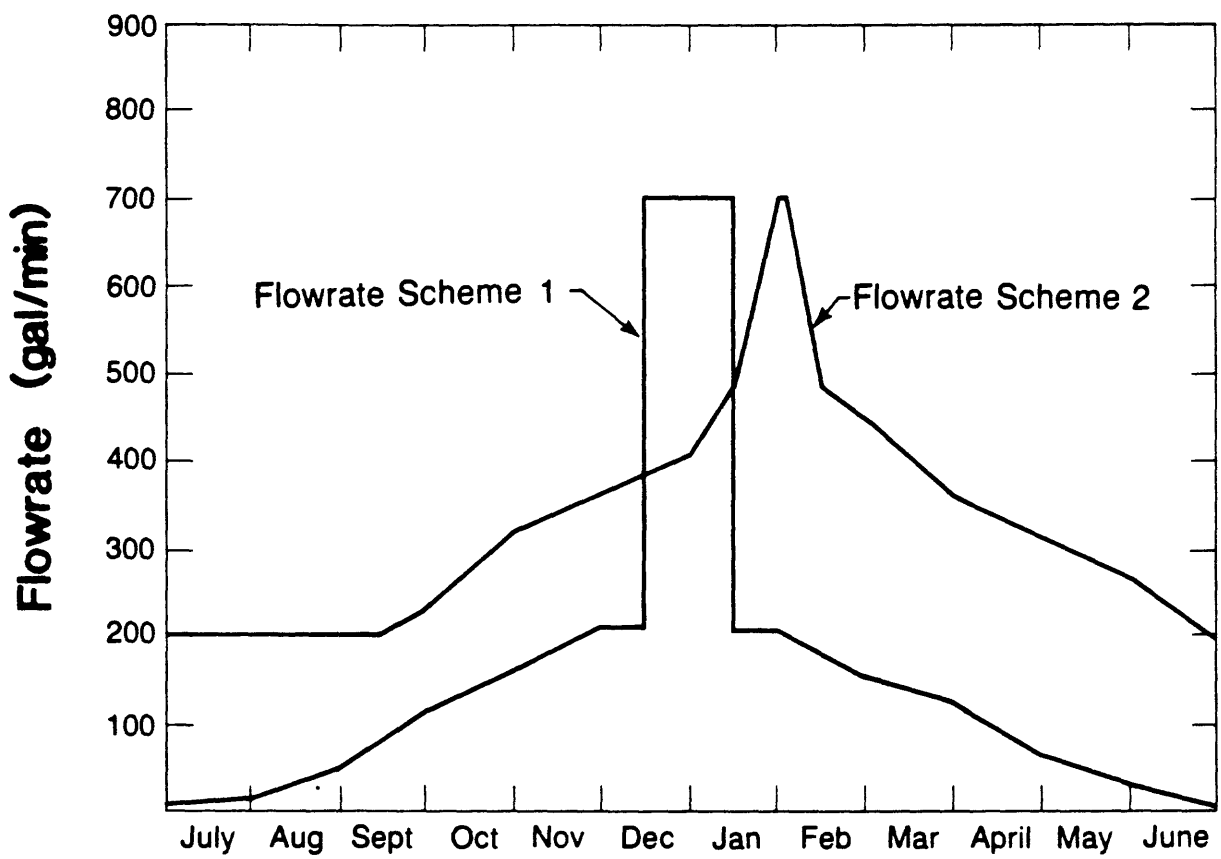

Figure 6-1. -- Flowrates used for drawdown and buildup calculations.

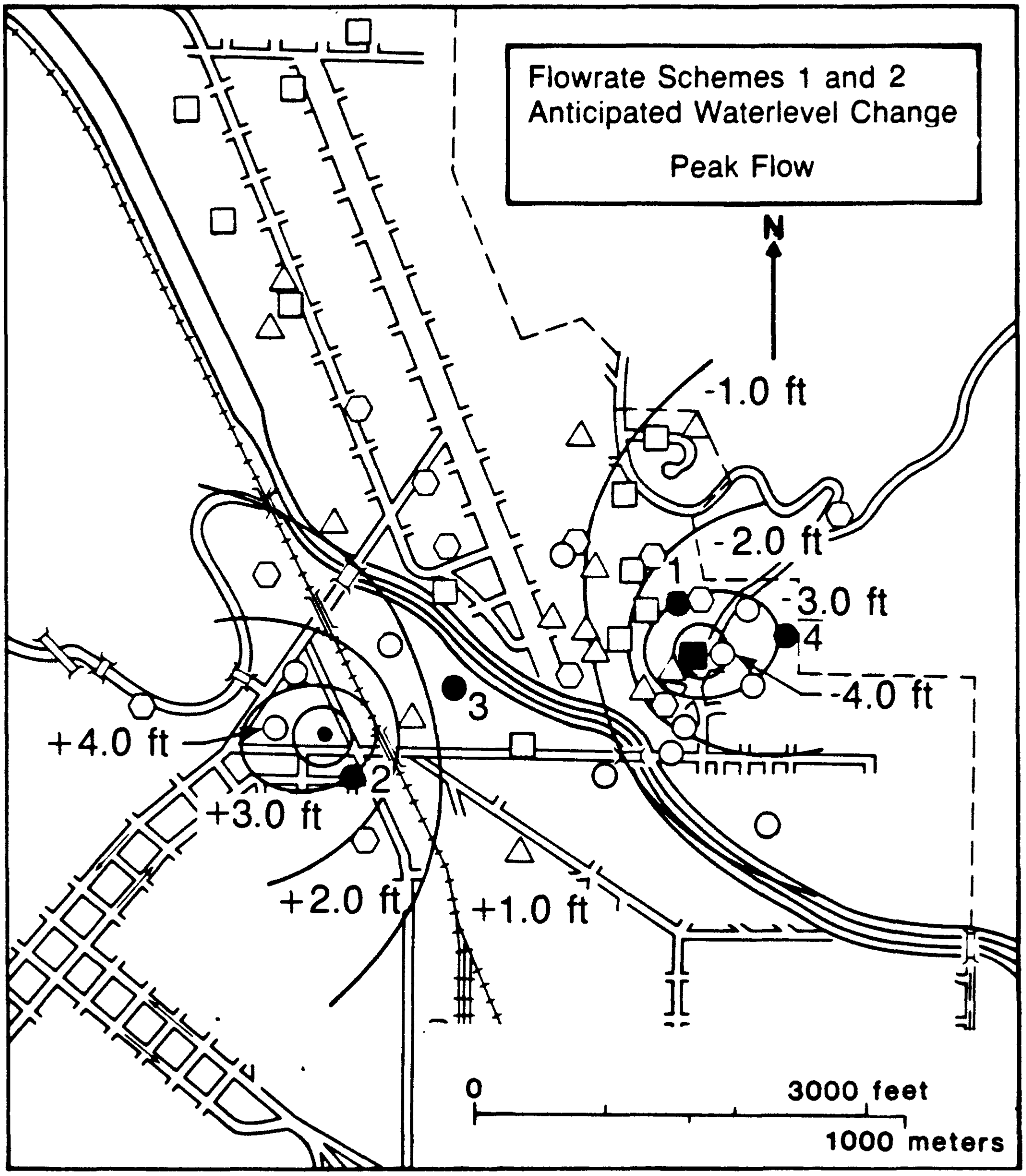

Figure 6-2. — Contours of calculated drawdown and buildup at the end of the peak flow periods for flowrate schemes 1 and 2.

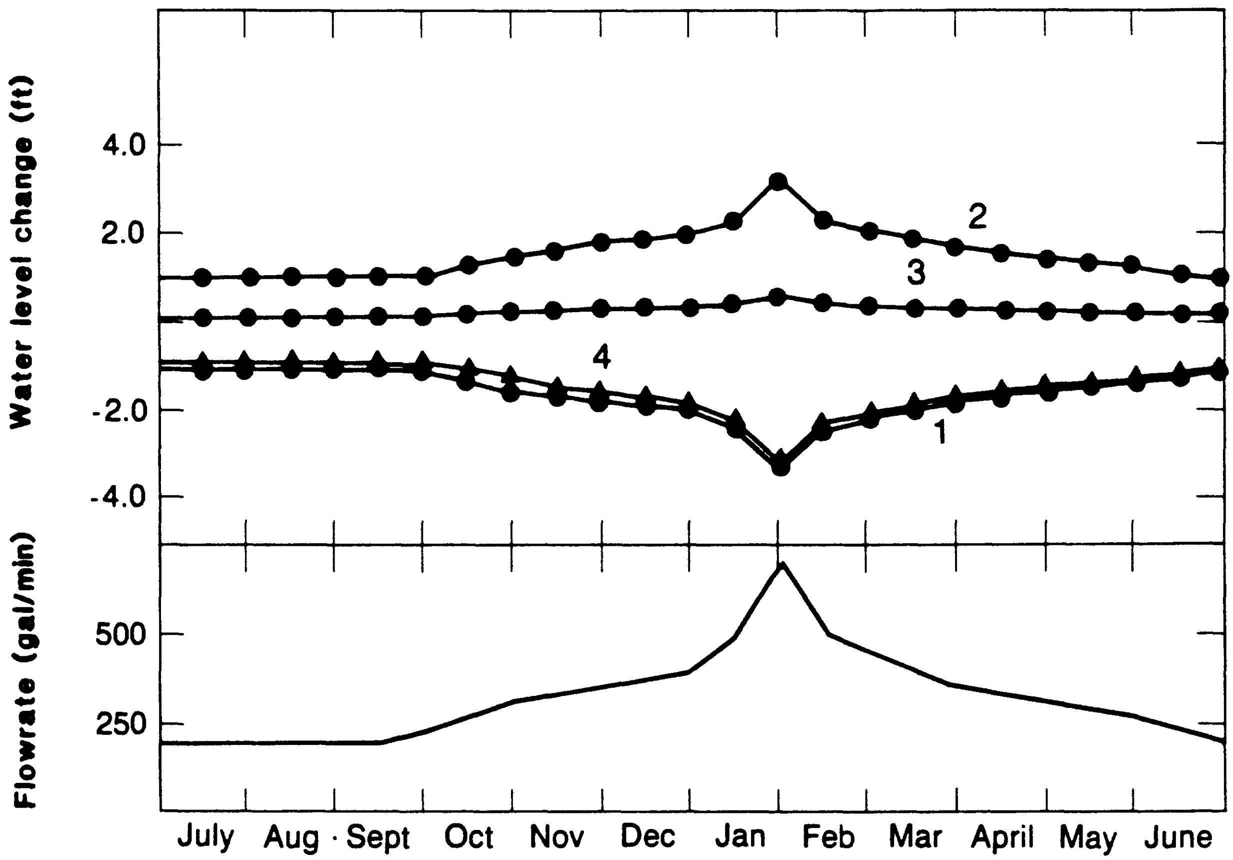

Figure 6-3. — Calculated drawdowns for 4 locations in the reservoir for flowrate scheme 2. Locations are shown in figure 6-2.

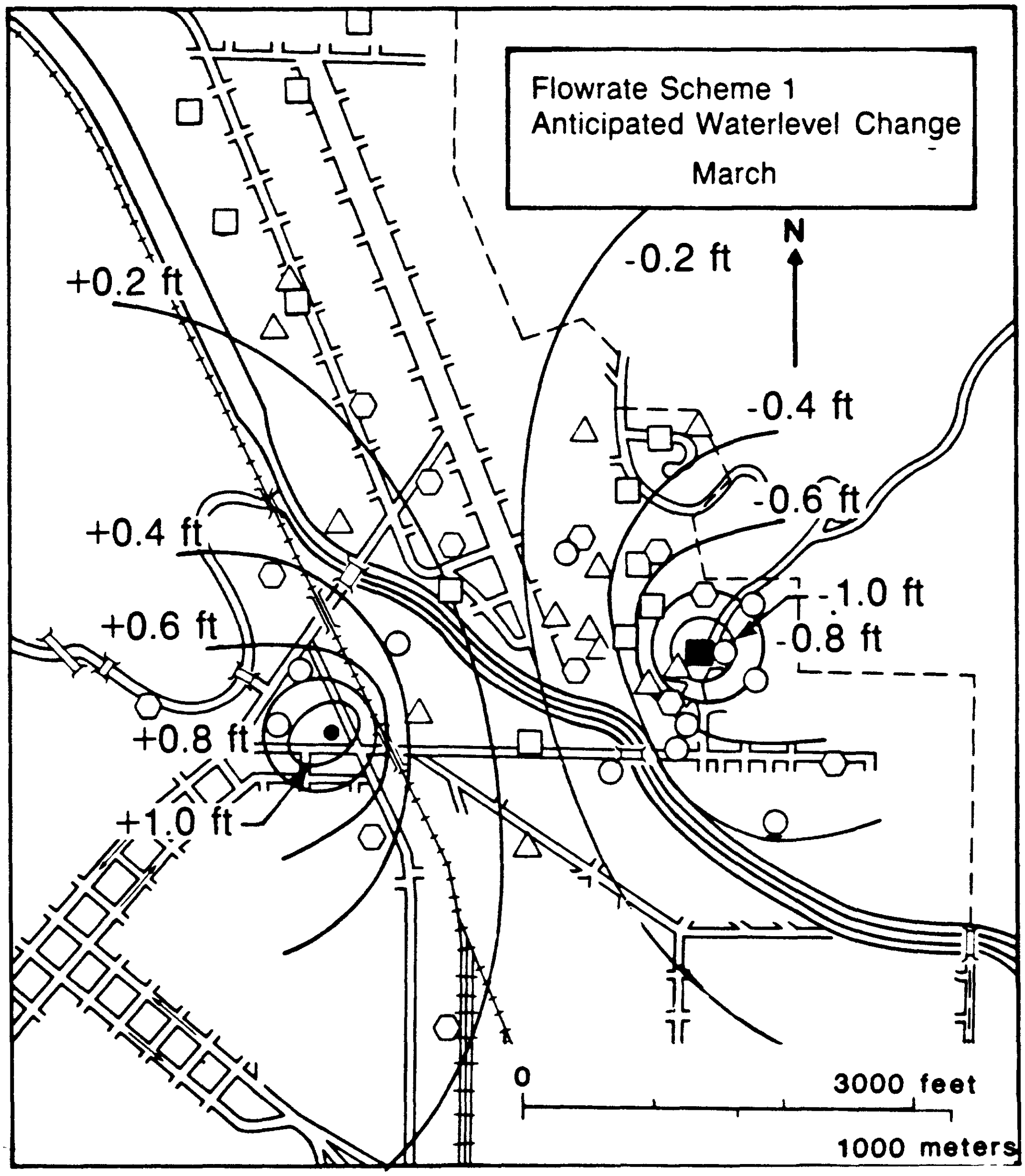

Figure 6-4. — Contours of calculated drawdowns and buildups for flowrate scheme 1 during March.

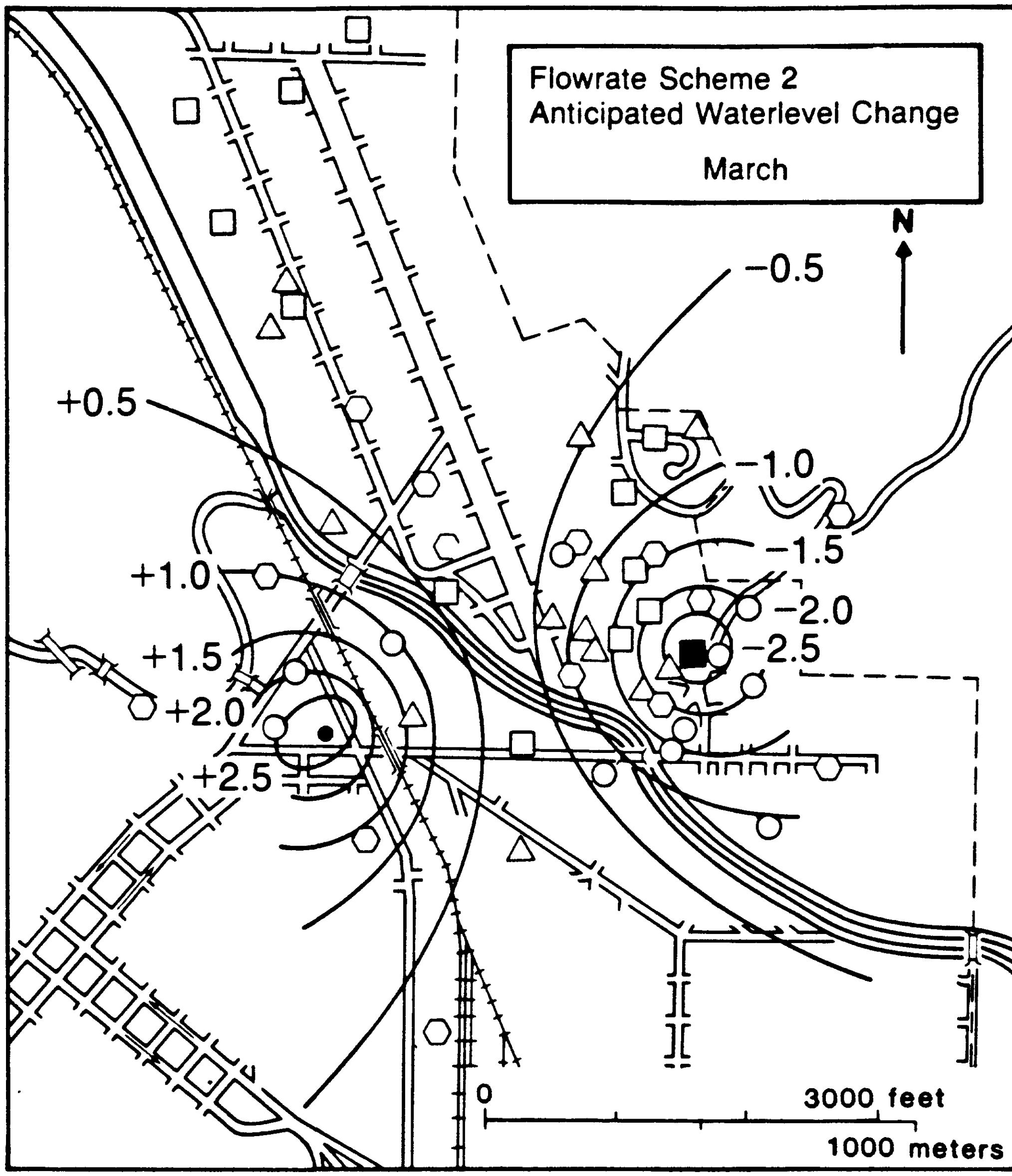

Figure 6-5. -- Contours of calculated drawdowns and buildups for flowrate scheme 2 during March.

CHAPTER 7. SUMMARY AND CONCLUSIONS

By

E. A. Sammel

The central focus of the research described in this report is the response of the shallow geothermal aquifer to stresses imposed by pumping and injection of the thermal water. In order properly to interpret the data derived from the pumping and injection tests, a number of additional tests and studies were made. These included chemical analyses, tracer tests, temperature measurements, and the collection of lithologic, climatic, seismic, discharge, and utilization data. As a result of these activities, some existing concepts of the geothermal aquifer were confirmed and several crucial new concepts were developed.

Occurrence and Characteristics of the Thermal Water

Lithologic and hydrologic data obtained from approximately 175 drillers' logs showed that the thermal water is derived from a stratified and areally heterogeneous aquifer. Within the total thickness of rocks that contain the aquifer, water moves preferentially in permeable strata that include lacustrine and volcanic sediment and basaltic to andesitic flow rocks, breccia and pyroclastics. The known thickness of rocks that comprise the water-bearing zones is nearly 2,000 ft. The total thickness and areal extent, and hence the volume, of the reservoir are not known.

The areal distributions of hydraulic heads and temperatures in the aquifer suggest that the thermal water rises along a fault zone near the northeast edge of the hot-well area and flows southwestward. The temperature of the water, which initially is higher than 120°C, decreases in the direction of flow, and, at distances greater than 3,000 ft from the fault, is generally less than 80°C.

Thermal water discharges at the land surface from wells in an area of about 3/4 square mile centered about 3,000 ft from the fault zone. This area formerly contained 5 groups of thermal springs that reportedly produced boiling water. Some of the artesian wells still produce water at or near the boiling point (96°C), but hydraulic heads reportedly have declined about 15 ft over the past 40 or 50 years. The occurrence of artesian pressures indicates that the aquifer is locally confined by rocks of low permeability at its upper surface.

The decline of artesian head and the final disappearance of the springs occurred in conjunction with an increase in withdrawals of thermal water and heat. An exclusive cause-and-effect relation between these occurrences is obscured, however, by the possibility that spring flows had begun to decline prior to any pumping, possibly in response to climatic changes. An analysis of climatic records might indicate whether or not a decrease in precipitation has reduced recharge to the geothermal system, but the apparent great age of the thermal water may make this determination difficult or impossible. An understanding of the full causes of head decline is essential to an assessment of the long-term potential of the reservoir, but a complete understanding was not achieved during this study.

Seasonal fluctuations in water levels occur in the aquifer as fluid withdrawals and the use of DHE's increase or decrease. These fluctuations correlate closely with changes in mean daily air temperature. They show that water levels respond quickly to changes in the demand for heat, and the uniformity of the responses indicates that the aquifer properties and the thermal water supply are reasonably uniform throughout the hot-well area.

The annual winter decrease in water level is about 4 to 6 ft over much of the area, but fluctuations are as large as 11 ft near the center of the area. On the northwest and southwest margins, fluctuations are 1 ft or less. The resulting inverted "cone of depression" is typical of producing well fields in homogeneous aquifers, but the magnitude of fluctuations in wells near the center of the cone could be due partly to a smaller transmissivity (permeability times thickness) in this area. No evidence of a significant decrease in transmissivity was observed in this area during the aquifer test, however.

In wells containing DHE's, a predictable qualitative relationship was observed between increased heat demand and a decrease in water level. However, the effects of DHE's on temperatures and water levels in the aquifer as a whole were not determined by our study. Some of our data indicate that aquifer temperatures are nearly constant during the heating season, and this suggests that the effects of DHE's on water levels probably are small and local compared to the widespread effects of pumping.

Our interpretation of the annual cycle of drawdown and recovery in the aquifer is based on measurements made during only the past 4 to 5 years in several wells. These records suggest that drawdowns due to current levels of withdrawal and DHE use probably reach an equilibrium condition during the period of maximum use (February - March) each year. If equilibrium is attained, it implies that the maximum withdrawals thus far made from the aquifer are balanced by recharge.

We have found no clear indication that a general temperature decrease has occurred in the aquifer as the result of increasing withdrawals. Individual DHE wells show temperature decreases if the natural convective flow in the well is interrupted for any reason, and temperature decreases occur in heating systems if the efficiency of the DHE is impaired by corrosion or a decrease in the heat-exchange surface area. Temperatures quickly increase again if these adverse conditions are remedied, and neither of these effects indicates a change in aquifer conditions. The temperature data are inadaquate for a determination of long-term changes, however, and no final conclusion concerning such changes has been reached during this study.

Current Use of Thermal Water

Thermal water is discharged to storm drains, sewers, and the "A" Canal by about 70 wells in the hot-well area. In 4 pairs of doublet wells, water is pumped from the aquifer and injected again through the second well. Excluding the doublet-well discharge, the average discharge of thermal water from wells in the hot-well area is approximately 540 gal/min or 775,000 gal/day. The amount of heat removed from the aquifer by pumped and artesian wells is nearly 5×10^{10} BTU/day. This is about 140 times the estimated quantity of heat discharged by the more than 380 wells that use DHE's.

The low efficiency of the pumped discharge is mitigated somewhat by the fact that a significant amount of the discharge is collected and reused for heating prior to its final disposal into Lake Ewauna. Despite this cascading of uses of discharge, the abstraction of heat by DHE's appears to be a more efficient use of the resource than the present pumped discharge.

The results of the injection and tracer tests indicate that pumping could be made a more efficient use of the resource if accompanied by reinjection. Using the estimated figures obtained in this study, a comparison can be made of the thermal efficiency of pumping and discharging water at the surface versus pumping and reinjecting the water into the aquifer. Assuming that the average temperature of the water now withdrawn by pumped wells is 80°C (a conservative figure), and using as a base temperature the average temperature of shallow ground water in the area (12°C) (Sammel, 1980), a calculation shows that the pumped wells discharge about 1,150 BTU per gallon of water pumped. For comparison with the reinjection case at a comparable scale of use, estimates of the heat and pumping requirements for a representative district heating plan can be used. Assuming that the water is reinjected, only part of the heat is discharged, the remainder being returned to the aquifer. On the basis of estimates made by OIT for a district heating plan of moderate size, the net withdrawal of heat is estimated to be about 330 BTU per gallon of water withdrawn. Thus, from the standpoint of heat losses from the aquifer, the consolidated use of heat from one or two wells with reinjection is clearly several times more efficient than a distributed use of heat from many wells without reinjection. Additional benefits obtained by the return of water to the aquifer are, of course, the maintenance of pressures and water levels in the reservoir.

A comparison of a large-scale heat-exchange use, such as a district heating plan, with current individual use of DHE's cannot be made on the same basis as the previous example, and has not been attempted for this report. However, data on which to base such a study are now available, and a meaningful comparison could probably be made by engineers in this field.

Geochemistry of the Reservoir Fluids

The chemical and isotopic compositions of the thermal and nonthermal waters of Klamath Falls show that the water in the shallow thermal aquifer is a mixture of a cold, dilute ground water with a hot, more saline component originally at a temperature between 150 and 190°C. The mixing produces waters at 95 to 130°C with a corresponding small range of salinity. The mixed thermal waters contain (in order of decreasing concentration) SO_4, Na, SiO_2, Cl, HCO_3, Ca, and K, and have a total salinity of about 1,000 mg/kg. This chemical composition is typical of waters that have equilibrated at high temperatures in a geothermal reservoir.

The nonthermal waters contain HCO_3, SiO_2, Na, Ca, Mg, Cl, K, and SO_4; salinity is less than 250 mg/kg. Although the cold water involved in the mixing could not be sampled, extrapolation of the data for the mixed waters shows that the cold component contains more Cl than the analyzed cold waters (10 vs 4 mg/kg) and more tritium (about 2 TU). The isotopic and chemical compositions of the mixed thermal waters show that both hot and cold components could have originated from local or regional ground water and that none of the water is of magmatic origin.

Sources of recharge have not been identified during this study. Deuterium concentrations suggest that recharge may occur at higher altitudes than those in the immediate vicinity of Klamath Falls, and the low tritium concentrations show that the cold recharge water has had a long (>30 year) residence time in the ground. These indications imply that the water has traveled a significant distance from the points of recharge and that the flow may occur in a deep regional aquifer.

The sulfate-water isotope geothermometer gives a probable temperature of 189°C for the deep thermal water. This temperature is in agreement with the silica (quartz) geothermometer when applied in a mixing model, and these models also produce a reasonable thermal end member on the basis of isotopic (^{18}O) mixing. The consistency of these results probably warrants a considerable amount of confidence in these geothermometers.

The Na-K-Ca geothermometer suggests that cation reequilibration has occurred in the mixed water, and indicated temperatures are close to those observed in the waters sampled. The results are similar in part to those obtained in other fault-controlled geothermal systems in volcanic terrains, such as Warner Valley in south central Oregon (Sammel, 1981) or Newberry Volcano in west central Oregon (Sammel, 1983), where cation geothermometers indicate temperatures significantly lower than original reservoir temperatures. The cation geothermometer and other chemical data at Klamath Falls suggest that the reequilibration zone could represent an extensive and possibly deep low-temperature reservoir.

The Tracer Tests

Information obtained from the tracer tests provides a preliminary basis for important decisions regarding the use and development of the resource. These decisions relate mainly to the possible consequences of reinjection if this method is to be employed as a part of the development strategy. The analysis of the pumping and injection tests leaves little doubt that reinjection of the thermal water at almost any point in the aquifer could raise water levels over a large area. However, this analysis must take into account the tracer-test results which relate to the thermal effects of reinjection.

The tracer tests tend to confirm the nature of the aquifer indicated by the pumping test. They show that small volumes of fluid move rapidly through large fractures and permeable porous media. The transfer of large volumes of water is much slower, however, because most of the rock volume has low permeability. Most of the heat in the aquifer is stored in the massive rock material, and this storage has the effect of slowing the temperature decrease that occurs when cooler water is introduced into a hot aquifer.

The prediction of the thermal and hydraulic consequences of injection has large uncertainties at the present time. For example, the preliminary tracer analysis indicates that thermal breakthrough between closely spaced production and injection wells may occur in a matter of a few weeks or months, whereas, experience with doublet wells at Klamath Falls shows that significant temperature changes do not occur in 3 of the 4 existing doublet pairs during the 9-month heating season. The fact that temperature has decreased in one doublet pair (S. M. Benson oral commun., 1983) suggests, however, that injection brings with it a significant risk and, therefore, must be used with caution.

Results of both the tracer tests and the pumping test suggest that injection wells should be carefully designed and located. In planning the locations, the indications of anistrophy observed in the pumping test may be useful for guiding the placement of points of injection. For example, the probability of increased pressure support along the axis of anisotrophy shoud be weighed against the increased likelihood of thermal breakthrough along this axis. The depth of injection also will be an important consideration in relation to the depths of water entry in existing wells. Lithologic logs and drillers' reports can be consulted in making these determinations. Monitoring and tests of all injection activity will be essential for increasing understanding of the aquifer behavior and for reducing uncertainties in predictive models.

The Hydrologic Tests and their Interpretation

The aquifer test conducted for this study differed from previous tests at Klamath Falls in its duration (50 days), its three distinct phases (pumping, pumping and injection, recovery), and the areal extent and intensity of monitoring (52 observation wells). Data collected during the test provide an unparalleled opportunity to study the hydraulic characteristics of an extensive, heterogeneous aquifer system in volcanic rocks. The data contain complexities, not resolved for this report, that will provide opportunities for analysis and research for some time to come. Nevertheless, several of the interpretations presented in Chapters 5 and 6 will have immediate application to decisions relating to effective use of the resource.

Aquifer Behavior under Stress

The plots of drawdown versus time obtained during the test fit theoretical curves that represent double-porosity conditions in the aquifer. The words "double porosity" are used here to describe an aquifer in which the initial flow in response to pumping occurs largely in more permeable strata or in fractures, whereas flow at later times is sustained partly by contributions from less permeable masses of rock (the matrix). Examination of drillers' logs and well cuttings indicates that the Klamath Falls aquifer contains both fractured rock and granular strata of high permeability as well as unfractured, massive rock and sedimentary strata of low permeability. For practical purposes of use and development, these two types of aquifers behave similarly and may have identical potentials for development.

Pressure changes were transmitted rapidly at the start of the pumping test. This finding is in accord with results of the doublet-well tracer tests, which show that the thermal water moves rapidly in permeable strata or fractures. The rapid and uniform response of observation wells also indicates that the permeable strata and fracture zones are confined by rocks of low permeability so that they behave more like a network of pipes than an unconfined reservoir. This concept is supported by the isotopic data (tritium), which show that little recent meteoric water mixes with the thermal water in the shallow aquifer.

Drawdowns caused by the aquifer test generally were not as large as those caused by current winter withdrawals. Consequently, the aquifer was not stressed sufficiently to cause new patterns of behavior to occur at its boundaries. The aquifer test did not reveal any hydrologic boundaries within 6,000 ft of the production well in a NW direction and within 4,500 ft in a SW direction. One implication of this fact is that the presumed supply vents in the fault zone did not act as restrictions on the flow. Thus, the fault conduits must be at least as transmissive as the aquifer rocks.

The absence of significant temperature changes in the produced water and in monitor wells during the aquifer test implies that no detectable cold-water flow was induced by the drawdowns. Cold-water recharge might have occurred at the boundaries of the hot-well area, but such effects would have been very small, and thus not likely to have been detectable in the late stages of the test.

During the third week of the pumping test, water levels had begun to fluctuate and, in some wells, to rise, presumably in response to a decrease in heat demand in supply wells and DHE wells. The effect was to obscure the true drawdown curve and, possibly, to mask the interception of recharge or low-permeability boundaries. Thus, conclusions regarding the absence of boundaries in the final stages of drawdown must be tempered by the possibility that small effects could have been missed.

Effects of Injection

The start of injection produced a rise in water levels that was detected almost immediately in all monitor wells. Aquifer characteristics observed during the injection phase of the test satisfactorily match those observed during the pumping-only phase.

The effectiveness of injection wells in supporting water levels and hydraulic pressures has been qualitatively known for many years at Klamath Falls as the result of experience with the 4 pairs of doublet wells. The analyses of the drawdown and injection data presented in Chapters 5 and 6 show clearly that injection of thermal water is capable of offsetting the immediate and widespread drawdowns that occur during pumping. This conclusion could have been arrived at solely on the basis of the drawdown data, but the results of the injection test are doubly reassuring on this point.

The price to be paid for the benefits of reinjection resides in the possiblity of thermal breakthrough. This potential threat is subject to analysis and prediction, but the complexities of these analyses have not been completely resolved for this report. Thermal breakthrough is reversible, but a longer time may be required for reversal than was required for the occurrence.

Potential for Development

In considering additional development of the geothermal resource at Klamath Falls, the issue, expressed in its simplest terms, is whether or not a specific development can occur without harm to the resource or to existing users. The issue, thus stated, is an oversimplification, however. The following discussion attempts to address some of the many considerations that have a bearing on what constitutes the resource and what constitutes "harm".

Present development at Klamath Falls utilizes a part of the total resource that exists in a shallow aquifer and a supply conduit (fault zone) of unknown depth and volume. Within and beneath the shallow aquifer, deeply circulating meteoric water mixes with high-temperature water from a deeper source. The depth and volume of the mixing zone are unknown, but the zone could extend beyond the boundaries of the hot-well area. Thus, the resource may include a reservoir of hot water that is not currently tapped by wells and that may have a potential for development.

Basic ground-water theory shows that water levels decline in predictable ways in a pumped aquifer supplied by a line source (the fault). Because the fault was not seen as a restrictive boundary during the aquifer test, we conclude that additional pumping could occur in the shallow aquifer before the storage capacity and flow capacity would be fully utilized. The consequences of additional pumping would be to increase drawdowns in the aquifer as gradients of flow increase to meet the new demand. Were this to occur, there would be little possibility of "harm" to the aquifer or reservoir because underlying recharge patterns are not likely to be affected by activities at shallow depths. However, if water levels were drawn down sufficiently to induce recharge of shallow meteoric water in the fault or the aquifer, temperatures ultimately would decrease. This decrease would be reversible, but only by decreasing pumping or altering its timing. Present knowledge does not permit us to predict whether or not cold-water recharge could be induced in the shallow aquifer.

In the shallow aquifer, increasing drawdowns could produce adverse consequences for some well owners. A small decline in water level would not directly result in a significant increase in pumping costs in most wells, but a decrease in artesian head in some wells could prolong the time each year during which pumping is required, thereby increasing the cost. A decline in water level in DHE wells will decrease the available heat-exchange area and, in some wells, will uncover the upper perforations that help to maintain temperatures in the well.

Reinjection can offset water-level declines due to pumping, but in the immediate vicinity of a production well, water levels will always be lower than elsewhere. Conversely, in the vicinity of the injection well, water levels will be higher than elsewhere. The aquifer-test analyses presented in Chapters 5 and 6 provide a preliminary basis for predicting the consequences of both pumping and reinjection, thereby permitting potentially harmful effects to be foreseen and possibly avoided.

A possible approach to new development is the drilling of deep wells into the underlying mixing zone. The hoped-for benefits would be the availability of hotter water than that obtained in most of the shallow aquifer, an extensive source area, and minimal interference with the shallow wells. The chemical nature of the mixed water and a probably high regional thermal gradient imply that the thickness of this zone could be as small as a few thousand feet and probably is not greater than about 6,000 feet. It is, therefore, within economic drilling limits. The mixing zone is not necessarily extensive, however, and it could be restricted to the fault conduits. Whether or not the mixing zone is deep or areally extensive, development of this zone might result in interference with wells in the shallow aquifer. Thorough testing would be required in order to determine or predict the consequences of development.

This brief discussion indicates that additional development is not ruled out by our present understanding of the aquifer conditions, but that development will be accompanied by risks and costs, some of them not well understood. Predictions made possible by these studies must be applied only as short-term estimates. Uncertainties in the analyses make extrapolation over periods of years highly uncertain, and our current ignorance of recharge conditions and climatic effects leaves a large gap in our understanding of critical factors in the aquifer behavior.

The magnitude of possible additional development of the shallow aquifer cannot be estimated on the basis of available information. Specifically, we lack information on the volume and extent of the aquifer, its boundary conditions, the magnitude and sources of recharge, and the impact of DHE wells. Test drilling could provide estimates of the first three of these conditions; the fourth might be determined by an intensive study of DHE wells. Pending such admittedly costly and time-consuming investigations, continued monitoring of temperatures, water levels, withdrawals, and DHE use can provide invaluable data on which to base decisions regarding additional development. The knowledge gained in our investigations represents a considerable advance, but clearly it is only a first step toward a full understanding of the nature of the Klamath Falls geothermal system. If further development were to occur in small, carefully planned stages, and if the effects of such development were carefully monitored, much more could be learned about the limits and potential of this promising energy resource.

Finally we point out that recent advances in geothermal development make feasible many alternative approaches to the use of the resource. Binary-fluid heat pumps, advanced heat exchangers, distributed reinjection, cascading of thermal uses to lower temperatures, and simple conservation measures are alternatives, most of which would permit additional use of the resource with few or no adverse consequences. A mix of these alternatives, developed with the same spirit of community cooperation and initiative that proved so helpful in the testing program, could extend the life of the resource and provide benefits in ways that probably are limited only by the resourcefulness and imagination of the Klamath Falls community.

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