Hydrogeologic Appraisal of the Klamath Falls Geothermal Area, Oregon

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By EDWARD A. SAMMEL

GEOHYDROLOGY OF GEOTHERMAL SYSTEMS

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1044-G

A description and evaluation of a
moderate-temperature hydrothermal
convection system

UNITED STATES DEPARTMENT OF THE INTERIOR

CECIL D. ANDRUS, *Secretary*

GEOLOGICAL SURVEY

H. William Menard, *Director*

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ILLUSTRATIONS

[Plates are in pocket]

TABLES

CONVERSION OF UNITS OF MEASUREMENT

Used in this report	Multiply by	To obtain
Length		
inch (in)	25.40	millimeter (mm)
inch (in)	2.540	centimeter (cm)
foot (ft)	0.3048	meter (m)
mile	1.609	kilometer (km)
Area		
square foot (ft^2)	0.0929	square meter (m^2)
acre	0.4048	hectare (ha)
square mile (mi^2)	2.590	square kilometer (km^2)
Volume		
acre-foot (acre-ft)	1.233×10^{-3}	cubic hectometer (hm^3)
cubic mile (mi^3)	4.168	cubic kilometer (km^3)
Flow		
gallons per minute (gal min^{-1})	0.06308	liters per second (L s^{-1})
million gallons per day (Mgal d^{-1})	3.785×10^3	cubic meters per day (m^3 d^{-1})
Topographic gradient		
feet per mile (ft mi^{-1})	0.1894	meters per kilometer (m km^{-1})
Transmissivity		
feet squared per day (ft^2 d^{-1})	0.0929	meters squared per day (m^2 d^{-1})
Specific capacity (wells)		
gallons per minute per foot (gal min^{-1} ft^{-1})	0.2070	liter per second per meter (L s^{-1} m^{-1})
Temperature		
degrees Celsius (°C)	1.8 (and add 32)	degrees Fahrenheit (°F)
Thermal gradient		
degrees Celsius per kilometer (°C km^{-1})	.05486	degrees Fahrenheit per hundred feet (°F/100 ft)

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Used in this report	Multiply by	To obtain
Thermal conductivity		
millicalories per centimeter per second per degree Celsius (mcal cm^{-1} s^{-1} °C^{-1})	0.4184	watts per meter per degree Celsius (W m^{-1} °C^{-1})
Specific heat (volumetric)		
calories per cubic centimeter per degree Celsius (cal cm^{-3} °C^{-1})	4.184 × 10^6	joules per cubic meter per degree Celsius (J m^{-3} °C^{-1})
Heat flow		
microcalories per square centimeter per second (Heat flow unit = HFU) (μcal cm^{-2} s^{-1})	4.184 × 10^{-2}	watts per square meter (W m^{-2})

HYDROGEOLOGIC APPRAISAL OF THE KLAMATH FALLS GEOTHERMAL AREA, OREGON

By Edward A. Sammel

ABSTRACT

Geothermal phenomena observed in the vicinity of Klamath Falls include hot springs with temperatures that approach 96°C (the approximate boiling temperature for the altitude), steam and water wells with temperatures that exceed 130°C, and hundreds of warm-water wells with temperatures ranging from 20° to 40°C. Although warm water is produced from wells scattered throughout most of the 350-square-mile area studied, waters with temperatures exceeding 60°C are confined to three relatively restricted areas, the northeast part of the City of Klamath Falls, Olene Gap, and the southwest flank of the Klamath Hills.

The hot waters are located near, and are presumably related to, major fault and fracture zones of the Basin and Range type. The displaced crustal blocks are composed of basaltic flow rocks and pyroclastics of Miocene to Pleistocene age, and of sedimentary deposits and basalt flows of the Yonna Formation of Pliocene age. Dip-slip movement along the high-angle faults may have been as much as 6,000 feet at places.

Average annual precipitation in 7,300 square miles of the upper Klamath River basin surrounding the study area is about 18.2 inches, of which between 12 and 14 inches is estimated to be lost through evapotranspiration.

Ground water of local meteoric origin moves through the shallow sedimentary deposits and volcanic rocks at relatively slow rates. Within the older basaltic rocks of the area, hydraulic conductivities are greater than in the shallow sediments, and ground water may move relatively freely parallel to the northwest-southeast structural trend. A small amount of ground water, perhaps 100,000 acre-feet per year, leaves the area in flow toward the southwest, but much of the ground water is discharged as evapotranspiration within the basin.

The local meteoric water that is assumed to be the source of the thermal water in the area has low concentrations of dissolved-solids in which calcium and bicarbonate are the dominant ions. During its passage through the geothermal reservoir, concentrations of dissolved solids increase to about 900 milligrams per liter, and sodium and sulfate become the dominant ions. Chloride concentrations remain relatively low, and silica concentrations increase from an average of about 35 milligrams per liter to about 100 milligrams per liter.

The evidence from cation ratios, silica concentrations, and oxygen and deuterium isotopes in the hot waters indicates that temperatures in the near-surface geothermal reservoir are relatively low. The estimated minimum reservoir temperature, based on the quartz geothermometer and mixing models, is 150°C. In contrast to these indications, the sulfate-oxygen isotope geothermometer indicates that, at some time in their history, the thermal waters have been exposed to temperatures approaching 200°C, possibly in a deeper reservoir.

Temperature distributions and heat flows in the shallow rocks of the area are strongly influenced by convective flow of water. Most observed temperature gradients are unreliable indicators of depths to the geothermal reservoir. Heat flow in the vicinity of the geothermal areas has not been determined, but evidence from temperature profiles suggests that heat flow in the Lower Klamath Lake basin is about 1.4 microcalories per square centimeter per second (1.4 Heat flow units), a value that is near the minimum expected for the Basin and Range province.

The net thermal flux discharged from springs and wells in the area is estimated to be on the order of 2×10^6 calories per second. Discharge by thermal waters into the shallow ground-water system beneath land surface may be many times this amount. Reportedly, at present only about 1.4×10^6 calories per second (6 megawatts) of thermal energy is beneficially used in the area.

A conceptual model of the geothermal system at Klamath Falls suggests that most of the observed phenomena result from transport of heat in a convective hot-water system closely related to the regional fault system. Temperatures at shallow depths are elevated both by convective transport and by the blanketing effect of rocks of low thermal conductivity. Circulation of meteoric water to a depth of 15,000 feet could account for the estimated temperatures in the thermal reservoir, assuming conductive heat flow and a temperature gradient of 30°C per kilometer in the shallow crustal rocks. Circulation to shallower depths may be sufficient to warm the water to the required temperatures under the more probable conditions of convective transport of heat and the insulating effect of overlying sediments.

Heat content in the shallow hot-water system (<10,000 feet depth) is probably in the range 15×10^{18} to 190×10^{18} joules, but is likely to be in the lower part of this range. This thermal energy may be stored in two or more separate reservoirs having a total volume not much greater than 12 cubic miles.

INTRODUCTION

BACKGROUND AND SCOPE OF STUDY

Geothermal phenomena in the vicinity of Klamath Falls consist of at least 7 thermal artesian springs and about 800 thermal wells. More than 400 of the wells are used for space heating, a use that began in the area about 50 years ago and is increasing rapidly at the present time. The amount of geothermal energy recovered from these wells and applied to nonelectrical use in the Klamath Falls area is probably one of the largest such applications in the United States.

Rapid growth in the use of geothermal energy at Klamath Falls and plans for expanding this use have led in recent years to a need for increased knowledge of the physical properties of the geothermal system. As one of the steps toward meeting this need, the investigation described in this report was begun by the U.S. Geological Survey in 1973. General objectives of the study were to bring together the large amount of existing data and to add new information in such areas of deficiency as chemical characteristics, geothermal gradients, and heat flow. An open-file report containing data and preliminary conclusions from the study has been released (Sammel, 1976).

The area of intensive study for this report covers approximately 350 square miles of the Klamath River drainage basin (fig. 1), and includes all known thermal anomalies in the immediate vicinity of Klamath Falls. The major area of thermal use, the so-called "hot-well area" of northeastern Klamath Falls (fig. 1), was not studied in detail, however. Because of time constraints, an adequate study of this area of intensive use would have severely limited the overall view that might otherwise be gained in a wider coverage of the geothermal areas. Staff members of the Geo-Heat Utilization Center, Oregon Institute of Technology (OIT), are now investigating the urban geothermal area, supported in part by a grant from the Geothermal Research Program of the U.S. Geological Survey.

METHODS OF INVESTIGATION

This report is based to a large extent on information obtained in an inventory of more than 300 wells and 20 springs in the Klamath Falls area. In most of these wells and springs measurements were made, as appropriate, of depth, water level, discharge, temperature, and specific electrical conductance. Driller's logs are available for about two-thirds of the wells.

Chemical analyses were obtained for 35 samples of water from wells and springs. Concentrations of carbonate, bicarbonate, and pH were determined in the field and the remainder of the analyses were performed in Geological Survey laboratories. Relatively complete analyses of water from 22 additional sources were obtained from owners or from published data. Water was also collected from 24 sources for analysis of stable isotopes and tritium.

A small amount of detailed geologic mapping was done, but in general a map by Peterson and McIntyre (1970) provided adequate geologic information for this reconnaissance.

Four shallow hydrologic test holes were drilled in the Lower Klamath Lake valley to depths ranging from 223 to 463 feet. Two 600-foot holes were drilled for heat-flow information. Temperature profiles were obtained by means of a thermistor probe in the 6 drill holes as well as in 41 unused water wells.

Additional geophysical data obtained during this investigation include gravity measurements at 103 stations in the Upper and Lower Klamath Lake basins and thermal infrared imagery obtained in aircraft flights over about 300 square miles of the area. The gravity measurements are described in the open-file report by Sammel (1976). The infrared imagery has not been completely processed and is not discussed in this report. Preliminary visual examination of computer-generated contour plots based on the imagery suggests that properly calibrated high-resolution infrared imagery may be useful as an exploration tool in areas of surficial convective thermal phenomena.

NUMBERING SYSTEM FOR WELLS AND SPRINGS

In this report, wells and springs are numbered according to a system based on their location within townships, ranges, and sections referred to the Willamette Baseline and Meridian. Thus, the number 39/9–21 bcdl specifies a well in Township 39 South, Range 9 East, and Section 21. The letters which follow indicate respectively the 1/4, 1/16, and 1/64 quadrants within the section according to the scheme illustrated in figure 2.

A number following the letters indicates that more than one well was inventoried in the same 1/64 section
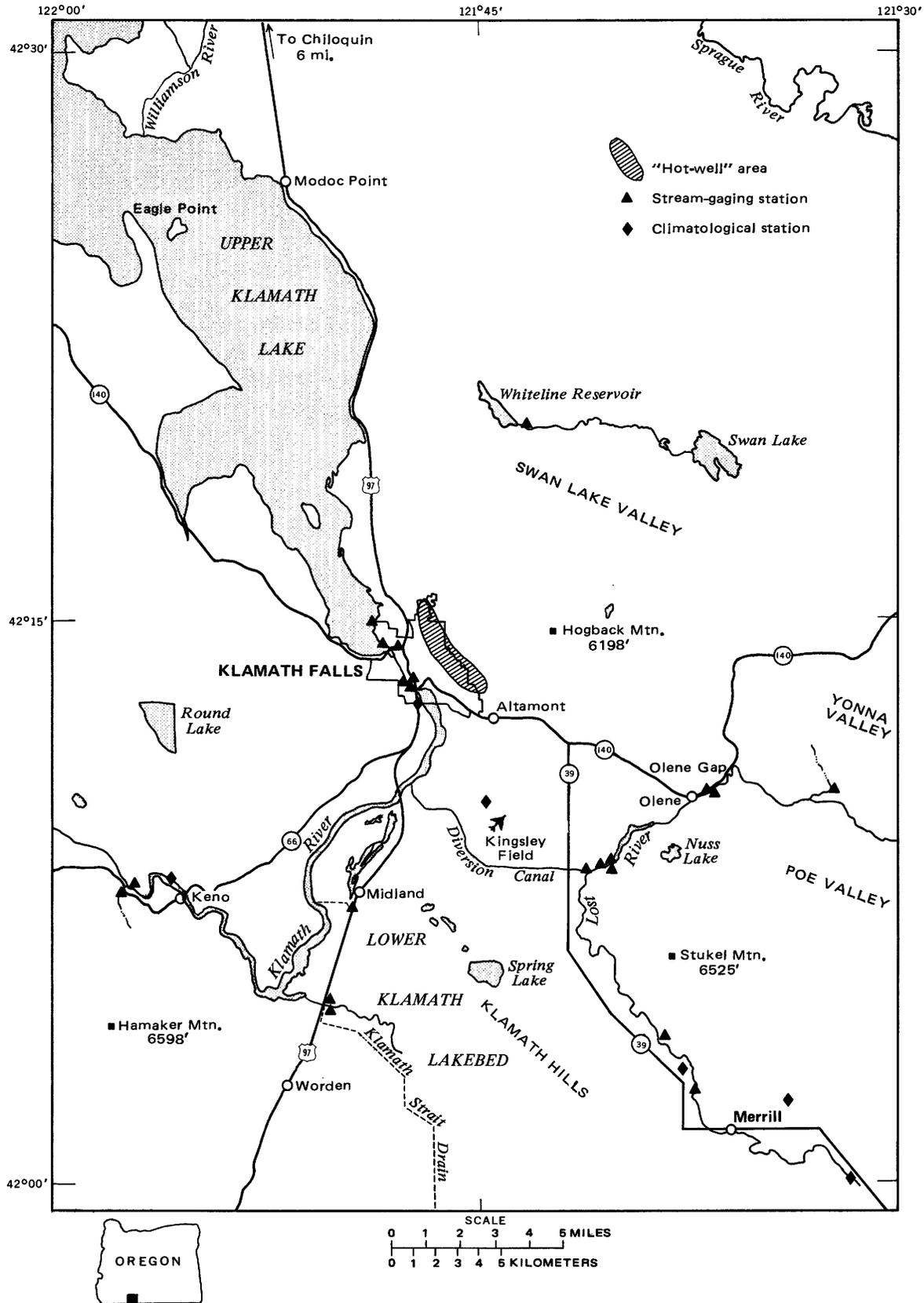
FIGURE 1.—The Klamath Falls area, locations of "hot-well" area, National Weather Service stations, and stream-gaging stations.
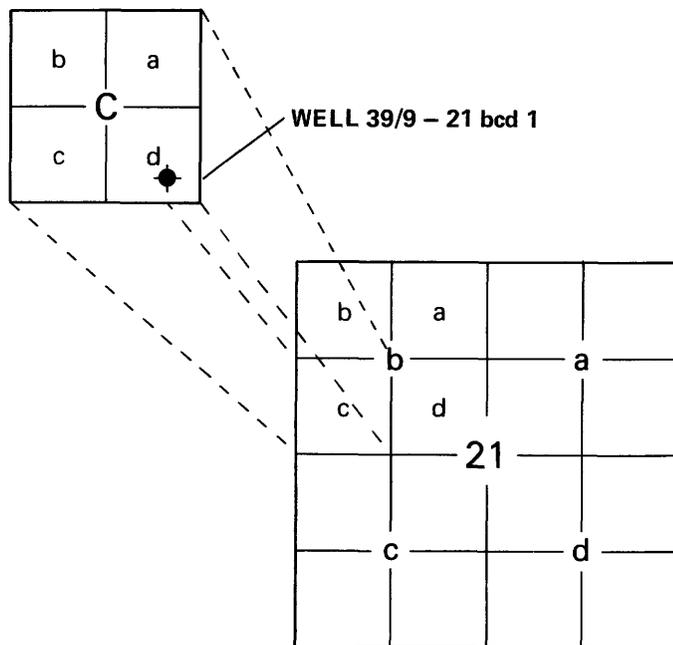
FIGURE 2.—Subdivision of township and range sections for well-numbering system.

quadrant; such wells are numbered serially in chronological order of inventory. Each well number, therefore, specifies a location within a quadrant approximately 600 feet on a side. Most wells shown on the accompanying maps are located with a probable error of less than 300 feet.

UNITS OF MEASUREMENT

Most of the data obtained during this study were measured in the inch-pound system of units. For reasons of accuracy, as well as the convenience of most readers, these units have been retained in this report. A table of multipliers is provided for use in converting the inch-pound units to SI (metric) units.

Field measurements of water temperature were obtained in units of degrees Celsius (°C), and for the sake of consistency, the few reported temperatures have been converted to this system. Air temperatures reported by the National Weather Service are in °F. In the fields of geothermal gradients and heat flow, most current scientific reports use SI units or other established metric equivalents. In order to retain temperature measurements in degrees Celsius and provide a convenient basis for comparison of calculated results, sections of this report dealing with these subjects contain commonly used metric units rather than inch-pound units. Thus, geothermal gradients are reported in degrees Celsius per kilometer, thermal conductivities are in milicalories per centimeter per second per degree Celsius, and heat flow is in microcalories

per square centimeter per second (heat flow units). Conversion factors for these units are provided in the table of units. The metric unit, kilometer, is used in the section, "Magnitude of the Geothermal Resource," in order to facilitate comparison of estimates made there with previously published estimates.

REGIONAL SETTING

REGIONAL GEOLOGY AND GEOTHERMAL ACTIVITY

Geologic structures in the Klamath Falls area are typical of the Basin and Range province (Gilbert, 1928). Large grabens form the Upper and Lower Klamath Lake basins; these basins are flanked by uplifted blocks bounded by steeply-dipping normal faults. The regional structure has a northwest-southeast alinement, but numerous faults have north-south strikes, and a few strike northeast-southwest.

Figure 3 shows geologic units and major faults in the Klamath Falls area as mapped by Peterson and McIntyre (1970). The oldest rocks exposed in the area are of latest Tertiary age; they include lacustrine and fluvial tuffaceous siltstone, sandstone, ashy diatomite, basaltic tuff and breccia, and a few thin basalt flows. These rocks were grouped by Newcomb (1958) in the Yonna Formation of Pliocene age. Later investigators have been reluctant to use this formation name because of uncertainties regarding the stratigraphic boundaries of the formation and the specific rock sequences that should be included. The name is used in this report for convenience in referring to groups of rocks of the type and age described by Newcomb.

Rocks of the Yonna Formation underlie the lower slopes of many upraised fault blocks in the area and occur at depth beneath most of the lacustrine and alluvial sediments in the structural basins. Ashy diatomites of the formation are conspicuous in roadcuts and quarries and are penetrated by many wells.

The most prevalent rocks of the Yonna Formation are thick, massively bedded, coarse-grained palagonitic sediments and pyroclastic rocks which are not clearly related to specific eruptive centers. Examples of these rocks are exposed in two large fault blocks south and east of Klamath Falls, Stukel and Hogback Mountains. Interbedded with the volcanic sediments are thin basaltic lavas which are generally dense, black, glassy, and vesicular. Some, such as those exposed at the Klamath Rock Products quarry (T39, R8, Sec. 11), are brecciated and altered owing to extrusion into water or wet diatomaceous ooze (Peterson and McIntyre, 1970). These flows probably issued from nearby fissures which were subsequently covered. The occurrence of local volcanic activity is also indicated by a few scattered maar and tuff-ring deposits.

Massive diatomites of the Yonna Formation have counterparts in Butte Valley, California, about 20 miles southwest of Klamath Falls, where they underlie Pliocene volcanic rocks of the High Cascades (Moore, 1937; Wood, 1960). Near Klamath Falls, diatomites are interbedded with Pliocene volcanic rocks (Meyers and Newcomb, 1952). The entire area over which diatomites are found is not known to have been covered by a continuous water body, but the lake, or lakes, must have been extensive.

Thickness of the Yonna Formation beneath the valley floors is largely unknown, but in adjacent ridges it ranges from a few feet to at least 850 feet. The maximum thickness is inferred from the driller's log of a well at the Presbyterian Intercommunity Hospital, located about 1 mile north of Klamath Falls.

Rocks older than the Yonna Formation do not crop out in the vicinity of Klamath Falls. The characteristics of these older rocks are known, however, from drillers' logs, and the probable ages of the rocks may be inferred from geologic mapping in adjacent areas. A few miles east of Klamath Falls in Swan Lake, Yonna, and Poe Valleys, the Yonna Formation lies unconformably on olivine basaltic rocks of probable Pliocene age (Peterson and McIntyre, 1970; Leonard and Harris, 1974). The basalt is blue-gray to gray-brown, vesicular, and columnarly jointed. Individual flows are as much as 30 feet thick and are interbedded with tuff and scoriaceous flow breccia (Leonard and Harris, 1974). Maximum observed thickness of the unit is about 300 feet in the Yonna Valley.

Pliocene basaltic flows also underlie extensive areas of the Warner Range, about 80 miles east of Klamath Falls. They dip westward for the most part and probably underlie tuffs and lacustrine deposits of the Sprague River valley about 35 miles northeast of Klamath Falls (Peterson and McIntyre, 1970). These large volumes of basaltic lavas and pyroclastics were extruded onto relatively level surfaces in the region surrounding Klamath Falls during late Pliocene time. Similar flows probably also covered the Klamath Falls area. They are recorded in drillers' logs as massive gray to brown vesicular basalt and tuffaceous sediment that underlie the Yonna Formation at places such as the Oregon Institute of Technology, about 2 miles north of Klamath Falls, at the Oregon Water Corporation wells in Klamath Falls, and possibly at a 1,300 foot well on Hamaker Mountain south of Keno. Similar rocks crop out in the Klamath River canyon southwest of Klamath Falls where their thickness is at least 800 feet (Newcomb and Hart, 1958).

The Pliocene olivine basalts described above probably correlate with extensive basaltic lavas that occur in adjacent areas of northern California (Wood, 1960).

These rocks are older basalts of the High Cascades mapped in Butte Valley, California (Wood, 1960), and in the Macdoel Quadrangle, Calif. (Williams, 1949). According to Williams, the large shield volcanoes of the High Cascades that dominate the eastern margin of the Cascade Range were probably extruded from north-south trending fissures during a period extending from late Pliocene to Holocene time. The flows are composed largely of olivine-augite basalt, although andesite flows and related pyroclastic rocks are commonly interbedded. The major flows dip eastward and are buried by younger basalts and pyroclastic rocks in the hills immediately west of Klamath Falls.

West of the High Cascade volcanoes, a belt of volcanic rocks from 30 to 40 miles wide forms the Western Cascades. These rocks, consisting largely of pyroxene andesites, olivine basalts, and rhyolite tuffs and lava domes, were uplifted, folded, and deeply dissected prior to the High Cascade eruptions. With ages ranging from Eocene to Pliocene (Williams, 1949), the rocks are equivalent to the Clarno Formation, John Day Formation, Columbia River Basalt Group, and the Cedarville Series of Russell (1928) in the Warner Range of northeastern California.

At places, such as Oregon-Calfornia border west of Klamath Falls, the thickness of the Western Cascade rocks is at least 12,000 feet (Williams, 1949, p. 20). Although these rocks dip to the east beneath the High Cascades, they do not crop out in the vicinity of Klamath Falls and have not been penetrated by wells there.

Depth of occurrence and characteristics of pre-Tertiary rocks in the Klamath Falls area are unknown. The oldest rocks exposed in adjacent regions are metacherts and quartzites of probable Paleozoic age intruded by quartz monzonites of Jurassic age near the foot of Mount Shasta (Williams, 1949, p. 14). The Chico Formation (Cretaceous), consisting of marine arkosic sandstone, conglomerate, and shale, underlies parts of the Macdoel Quadrangle in northern California. Hot springs that occur in the Macdoel Quadrangle about 30 miles southwest of Klamath Falls have high chloride concentrations that may indicate the presence of the Chico Formation at this location (Williams, 1949, p. 55). It is possible, therefore, that marine sedimentary rocks also underlie the Klamath Falls area, although no evidence of the presence of such rocks is known.

The structural development of the Klamath Falls area may have begun with normal faulting as early as Pliocene time (Peterson and McIntyre, 1970, p. 28). Subsequent to the deposition of the Yonna sediments and contemporaneous basalt flows, the entire region was apparently folded into broad anticlines and synclines with generally north-south axial trends. The
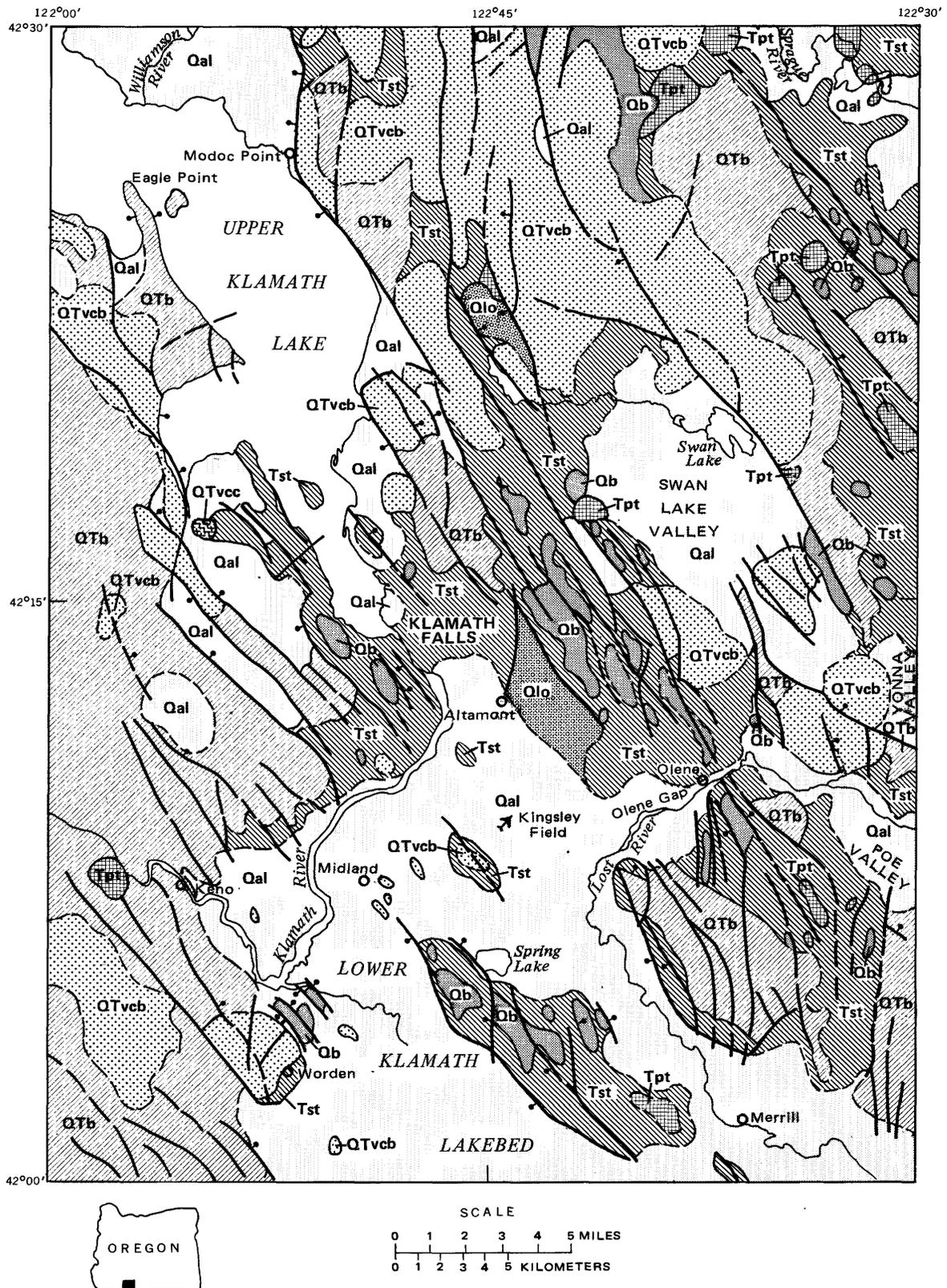
FIGURE 3.—Areal geology of the Klamath Falls area. (Modified from Peterson and McIntyre, 1970.)

CORRELATION OF MAP UNITS

Qal		} Holocene	
Qlo			} QUATERNARY
Qb		} Pleistocene	
	QTb　QTvcc	} Pliocene and Pleistocene	TERTIARY AND QUATERNARY
	QTvcb		
Tst	Tpt	} Pliocene	TERTIARY
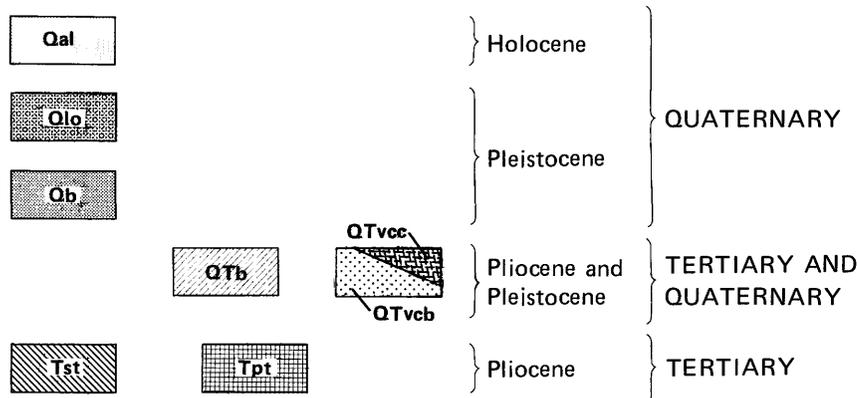
DESCRIPTION OF MAP UNITS

Qal　Alluvial deposits including diatomite and peat

Qlo　Fluvial terrace and lacustrine deposits

Qb　Basalt flows

QTb　Basalt flows, pyroclastic breccia, and agglomerate, of uncertain stratigraphic relationships. May in part be contemporary with Tst and Qb

QTvcb　Basaltic eruptive centers

QTvcc　Basaltic cinder cones

Tst　Lacustrine and fluviatile tuffaceous siltstone, sandstone, ashy diato-mite, basaltic tuff and breccia, few thin basalt flows (=Yonna Formation)

Tpt　Maar and tuff-ring deposits

— — — — —　Contact — Dashed where approximate

— — — — —　Fault — Dashed where approximate. Ball on downthrown side

Note: Description of map units and map symbols are quoted from Peterson and McIntyre, 1970

FIGURE 3.—Continued.

Sycan Marsh-Bly-Beatty area lay in a broad syncline, and the present Summer Lake-Goose Lake graben was an anticline. The Klamath Falls area may also have been an anticline whose crest subsequently dropped to form the present graben (Peterson and McIntyre, 1970, p. 29).

The volcanic activity and faulting that produced the present topography in the Klamath Falls area occurred mostly during the Pleistocene. To the west, new volcanic vents continued to form in the High Cascades, and stratovolcanoes such as Mts. Shasta, McLoughlin, and Mazama were built. Near Klamath Falls, wide-

spread thin basalt flows were extruded over the Pliocene lacustrine sediments. Exposures of these vesicular, medium to dark gray, diktytaxitic Pleistocene basalts occur in the Klamath Hills, on Stukel Mountain south of Olene Gap, and on Hogback Mountain east of Klamath Falls. The cataclysmic eruption of Mount Mazama, about 6,700 years ago, produced thick deposits of ash and pumice that mantle the northern part of the area.

Large volumes of Quaternary and Tertiary basalt flows and breccias of uncertain stratigraphic relationships underlie some of the highest ridges near Klamath Falls. These rocks may in part be contemporaneous with the Pliocene Yonna Formation, but in part are Pleistocene. Normal faults cut all the above units, as well as the rocks of basalt eruptive centers that form some of the higher isolated hills.

The present-day basins of the Klamath Falls area were probably established early in Pleistocene time as the result of tensional fracturing of the entire region. Fluvial-terrace and lacustrine sediments deposited in these basins are now found at altitudes as high as 200 feet above the present valley floor, presumably as the result of high lake levels in late Pleistocene time. Some of these deposits are deformed by faults, indicating that deformation continued into late Pleistocene time (Peterson and McIntyre, 1970, p. 19). It is not certain, therefore, that the present levels of these deposits represent original altitudes of lake shorelines.

Basin and Range faulting in the Klamath Falls area reached a climax during Pleistocene time. Movement was almost entirely dipslip along steeply-dipping ($>60°$) normal faults. The maximum displacement is at least 980 feet and may be as much as 6,000 feet if current interpretations of gravity data are accepted (D. L. Peterson, 1976). Some seismic activity has occurred in the region during historical times, but no historical surface breaks have been recorded (Couch and Lowell, 1971, p. 67).

Thermal wells and springs are widespread in the Klamath Falls area. In this report, a thermal water is defined as one having a temperature $\geq 20°C$, or about 8°C above the average temperature of shallow ground water. Known hot waters, with temperatures greater than 60°C, are confined to the following three locations: (1) an area of about 2 mi^2 along the northeast edge of the city; (2) a stream-cut gap in the hills about 8 miles southeast of the city at Olene Gap; and (3) the southwest flank of the Klamath Hills, about 12 miles south of the city. These three areas are referred to in subsequent sections of the report as the "principal geothermal areas."

The highest-temperature geothermal manifestations include several springs with temperatures that are near the boiling point (less than approximately 96°C), a few hot-water wells with subsurface temperatures that exceed the boiling point, and a few steam wells. One exceptionally hot well had a bottom-hole temperature when drilled of about 140°C. Most of the wells having temperatures greater than 60°C are located in the city's hot-well area. In hundreds of other wells scattered through the lower Klamath Falls basin, water temperatures range from just above normal for the region (15°C) to about 40°C.

The principal geothermal phenomena in the area are spatially related to fault zones. All waters known to have temperatures greater than 65°C, for example, occur within 1 mile of major faults. The association of the Klamath Falls geothermal phenomena with Basin and Range type faults suggests an anology with other geothermal areas of the Basin and Range province such as Warner Valley, Ore., Surprise Valley, Calif., and the many large fault-controlled hot-water systems in Nevada.

Each of the principal geothermal areas in the vicinity of Klamath Falls contains extensive areas of silicified rocks as evidence that hydrothermal activity was formerly more widespread than at present. Some of these rocks, such as the silicified palagonite tuff that crops out above the Presbyterian Intercommunity Hospital, occur at altitudes as much as several hundred feet higher than any present surface flows of hot water. This evidence suggests that hydraulic heads in the major hydrothermal system were formerly much higher than at the present time or, alternatively, that hydrothermal activity preceded most of the displacement along the fault zones. The latter alternative appears to be the more likely, but the question has not been resolved by this study.

The ultimate source of heat for the geothermal waters is not known. Peterson and McIntyre (1970, p. 37) refer to the presence of dikes, intercalated sill-like masses, bleached and silicified rocks, deposits of calcite and gypsum, and a halo of mercury mineralization as suggesting the existence of a hot igneous mass at depth. An alternative explanation, proposed by the author, that involves only deep circulation of meteoric water is discussed below under the heading "Conceptual Model of the Geothermal System."

CLIMATE AND DRAINAGE

PRECIPITATION AND TEMPERATURE

Mean annual precipitation at six stations in the vicinity of Klamath Falls is shown in table 1 for the period of record available. The range of annual precipitation at a given station is relatively great. The range between stations is also great with a low of 11.8 inches

TABLE 1.—*Mean annual precipitation and temperature measured at National Weather Service stations in the vicinity of Klamath Falls*

Station	Altitude (ft above NGVD of 1929)	Period of record used	Mean annual precipitation (in)	Mean annual temperature (°F)
Chiloquin	4,198	1948–73[a]	18.13	42.9
Keno	4,116	1948–73[b]	19.49	------
Kingsely Field	4,085	1949–73	11.93	46.6
Klamath Falls	4,098	1948–73	14.37	48.0
Merrill	4,080	1953–66[b]	11.80	------
Rocky Point	4,150	1967–73[c]	23.15	44.6

[a]Partial records for years 1964, 1968, and 1971 completed by estimates based on correlations with other stations.
[b]Record for 1964 estimated by correlation with other stations.
[c]Partial record for 1968 completed by internal correlation.

at Merrill and a high of 23.15 inches at Rocky Point. Outside the study area, extremes in this part of the Klamath River basin range from less than 10 inches at Tule Lake, Calif., to more than 60 inches on the highest peaks north and west of Klamath Falls.

Local orographic effects largely determine the relative amounts of precipitation. Thus, of the moisture that reaches the area after passing over the Cascades on the westerly winds, most falls on the highlands and a relatively small amount reaches the valley floors.

In most years, between 60 and 70 percent of the total precipitation falls during the months October through March. Most of this moisture is in the form of snow, although midwinter rains are frequent in the valley.

Based on rainfall maps published by the Oregon State Water Resources Board (1971), average annual precipitation in the drainage basin above Klamath Falls, including subbasins of the Williamson River, Sprague River, and Upper Klamath Lake, is calculated to be 22 inches over 3,820 mi². For the basin below Klamath Falls, precipitation is calculated to be about 15 inches over 3,497 mi². The area of this estimate includes the Lost River subbasin of Oregon and California but excludes the Meiss Lake closed basin southwest of Dorris, Calif. On the basis of the above calculations, the average annual precipitation in the four subbasins near Klamath Falls is 18.6 inches over 7,317 mi².

Mean annual air temperatures at stations near Klamath Falls are given in table 1. Stations with reported air temperatures are located at relatively low altitudes and mean annual temperatures at greater altitudes are undoubtedly considerably lower than those reported at the weather stations. No estimate of mean annual air temperature has been made for this part of the basin.

Measured temperatures of nonthermal ground water in shallow aquifers near Klamath Falls range from 7° to 19°C. Temperatures at individual sites may vary 5°C or more annually. The mean annual ground-water temperature at most places is probably several degrees

higher than mean annual air temperatures, and the average temperature of shallow groundwater in the area is estimated to be about 12°C.

EVAPOTRANSPIRATION

Estimates of evapotranspiration in the Upper Klamath River basin are given in table 2. Rates of evapotranspiration in the region are low owing to the relatively cool average temperatures and the low rainfall. Minimum rates may occur in the high dry valleys of the eastern part of the basin, where evapotranspiration from sage and prairie vegetation is probably less than 1 foot per year (virtually the total precipitation). Maximum evapotranspiration may be as much as 3 feet per year from widespread marshes and ponds in the upper basin.

For the upper Klamath Lake basin, including the three subbasins above Klamath Falls (3,802 mi²), evapotranspiration is estimated to be 1.1 feet per year. This estimate includes the effects of an estimated ground-water outflow from the upper Klamath Lake basin of 25,000 acre-feet per year (Illian, 1970). If, as a lower limit, ground-water outflow is assumed to be one-half Illian's estimate, the calculated evapotranspiration is 1.2 feet per year.

For the Lost River subbasin in Oregon and California (3,497 mi²), evapotranspiration is calculated to be 15 inches per year. The calculation includes Illian's estimate that ground-water flow from the upper Klamath Lake basin into the lower basin is 200,000

TABLE 2.—*Evapotranspiration in the Upper Klamath River basin, Oregon and California*

Type of area	Area (mi²)	Evapotranspiration (ft yr⁻¹)	Source of estimate
High, dry valleys—sage and prairie grass.		<1	A
Marshland	56	3.6	O
	100	3	L
Low valleys—irrigated crops	422	1.3	O
Upper Klamath Lake (altitude 4,139 ft)	[1]125	3½–4	O
Clear Lake (Calif.) (altitude 4,470 ft)	69	2½–3½	O
Upper Klamath River basin: Sprague, Williamson, and Upper Klamath Lake subbasins	3,820	[2]1.1	A
Lost River subbasin (Ore. and Calif.)	3,497	[3]1.3	A

A. Author (this report).
O. Oregon State Water Resources Board (1971)
L. Leonard and Harris (1974)
[1]Range 100 to 140 mi², depending on seasonal change in lake stage.
[2]Calculated from precipitation data (table 1), published streamflow records (U.S. Geological Survey), and unpublished canal-flow records (U.S. Bureau of Reclamation). Estimates of ground-water outflow from the Upper Klamath Lake and Sprague River subbasins (Illian, 1970) have been included in the calculation.
[3]Calculated from precipitation data (table 1) and published streamflow records for the Klamath River at Keno and Boyle Power Plant (U.S. Geological Survey). Estimates of ground-water flow into the Lost River subbasin (Illian, 1970) have been included, along with the author's estimate of ground-water outflow below Keno.

acre-ft per year and the author's estimate that about 100,000 acre-ft of water leaves the lower basin as ground-water flow. It is assumed that ground-water flow from the south end of the basin in California is negligible and that ground-water discharge occurs almost entirely through fractured rocks in the Klamath River gorge below Keno.

SURFACE WATER

Three major streams, the Sprague, Williamson, and Wood Rivers, drain the upper Klamath River basin above Klamath Falls. The Williamson and Wood Rivers flow into Upper Klamath Lake and the Sprague River is a tributary to the Williamson. Upper Klamath Lake is also fed by springs and seeps in or near the lake and by small tributaries on its periphery. The average annual flow from all the above sources for the period 1930–68 is estimated to about 1,400,000 acre-ft (Oregon State Water Resources Board, 1971, table 21).

Upper Klamath Lake, the largest natural lake in Oregon, has a surface area that varies between 100 and 140 mi^2, depending on seasonal changes in lake stage, and a length of about 22 miles. Agency Lake, 6½ miles in length, is appended to the north end of Upper Klamath Lake. The average depth of Upper Klamath Lake is only 8 feet, although depths reach 40 to 50 feet in a narrow trench between Eagle Ridge and Bare Island (Hubbard, 1970). Because of its shallow depth, regulated outflow, and generally warm water, Upper Klamath Lake is undergoing relatively rapid eutrophication.

Water from Upper Klamath Lake flows through a narrow strait on the western edge of the city of Klamath Falls, meanders through marshlands toward the southwest, and then, after passing over a dam near the town of Keno, leaves the area through a narrow gorge cut into bedrock. Below Klamath Falls, the stream becomes the Klamath River. Average annual flow at Keno for the period 1921–75 is 1,234,000 acre-ft (U.S. Geological Survey, 1976).

More than 300,000 acre-ft of water is diverted annually from Upper Klamath Lake and the Klamath River above Keno. This water is conveyed through three canals, the "A," the North, and the Ady, for use in irrigating lands in the Lower Klamath Lake basin. The canals are part of a major drainage and irrigation scheme built and, for the most part, operated by the U.S. Bureau of Reclamation (USBR). The project was begun in 1904 and is one of the oldest USBR projects in existence.

Southwest of Klamath Falls, the Lost River enters the Lower Klamath Lake valley through Olene Gap and flows into California. This stream has been incorporated in the USBR irrigation project, and its flow is

now totally regulated. From a diversion dam north of Stukel Mountain, water is either diverted to the Klamath River or is returned to the Lost River channel for use downstream. About 116,000 acre-ft of water from the Lost River is annually distributed in the lower valley.

The network of canals and laterals that now serve the USBR Klamath Project totals about 700 miles in length and irrigates about 200,000 acres of land in Oregon and California. Tailwater flows southward to a closed basin in the vicinity of Tulelake, Calif. The extent of the water surface in this swamp area is closely regulated; excess water is pumped back to the north through the Klamath Straits Drain into the Klamath River. About 108,000 acre-ft of water is annually pumped out of the Tulelake area in addition to an average net flow of 141,000 acre-ft diverted to the Klamath River from the Lost River through the Lost River Diversion Channel.

In 1909 the Southern Pacific Railroad dike was completed along the west side of the lower lake valley, separating the marshland in the valley from the Klamath River. Since this time the Lower Klamath Lake in Oregon and California has been essentially a closed basin. It is likely, in fact, that the Tulelake area acted as a sump for some thousands of years previously. At present, surface water leaves the lower sump area only by evapotranspiration or by pumping to raise the water approximately 15 feet to its discharge point in the Klamath River.

The above description of the USBR Klamath Project is an oversimplification of an extremely complex system. The average figures for inflow and outflow mean little in attempts to evaluate the hydrology of the Lower Klamath Lake valley. Only the major canals and drains are gaged, and the patterns of irrigation use and drainage are not readily available for quantitative evaluation. In the absence of detailed data, only the crudest generalizations are possible regarding consumptive use, evapotranspiration, or recharge to the ground-water reservoir.

OCCURRENCE AND MOVEMENT OF GROUND WATER

THE SHALLOW RESERVOIR

The shallow ground-water reservoir in the Upper and Lower Klamath Lake valleys is, for the most part, a continuous, largely unconfined body to depths of more than 1,500 feet. Although few data are available in the highland areas, shallow cold ground water beneath the upraised rock masses also appears to be generally unconfined. Water levels in the valleys range from land surface to 10 or 15 feet below land surface,

and annual fluctuations of the water table are generally less than 10 feet. Water levels in terrace deposits and alluvial fans are commonly 15 to 20 feet below land surface, whereas at places under the higher bedrock ridges the water table is as much as several hundred feet below the surface. Fluctuations of the water table in the highlands are probably even smaller than those observed in the valleys.

The generalized surface of the shallow nonthermal ground-water reservoir is shown on plate 1. The water table appears to be a continuous, fairly smooth surface which forms a subdued replica of the topography. At numerous places, however, water levels in individual wells show departures from the smooth surface. As explained in the map text, some of these discrepancies result from errors of measurement or reporting. Many, however, reflect differences in geologic structure, stratigraphy, or well depth. Effects of ground-water pumping are believed to be negligible in most of the area.

Compartmentation of the shallow aquifer by faults undoubtedly occurs both in the valley sediments and the bedrock ridges. These discontinuities may account for water-level discrepancies such as those shown by wells on the southwest flank of the Klamath Hills and the edge of the hills near Keno. The effects are minor, however, throughout most of the volume of Pleistocene and Holocene sediments. Compartmentation that occurs at places in the Tertiary rocks is discussed below in the section on "Deeper Basaltic Aquifers."

Wells in the valley-fill deposits commonly show a decrease in static water level with depth; that is, the deeper wells have lower water levels. The lower water levels in deep wells are the result of partial confinement of the deeper zones of the aquifer and a loss of hydraulic head between the recharge areas and the aquifers tapped by the wells.

Hydraulic heads in wells are above land surface at several places in the area, indicating that the aquifer is confined at these places. Examples are wells north and east of Lake Ewauna (T38, R9, Sections 32 and 33), most wells north of Stukel Mountain (T39, R10, Sections 28–34), and some wells in Lost River valley south and west of Stukel Mountain. Many of these wells, particularly those north of Stukel Mountain, have large flows, but shut-in pressures indicate that artesian heads are generally less than 10 feet above land surface. In the southeastern part of Klamath Falls, hydraulic heads in some hot wells formerly were more than 20 feet above land surface (J. W. Lund, oral commun., 1976). In recent years, however, the heads have declined to near land surface, possibly as the result of increased use of water from hot wells in the Klamath Falls geothermal area.

Few artesian wells are more than about a mile from upraised fault blocks, and it is assumed that the recharge areas for all artesian wells are in the nearby ridges. Examination of drillers' logs generally suggests that the water is confined in permeable zones between basalt flows, or in granular sediments overlain by clay and silt of the lake deposits.

A few shallow artesian wells in the flood plain of the Lost River southeast of Klamath Falls (T39, R10, Sections 17–20) penetrate fine black sand that is overlain by silt and clay. Drillers' logs and data from a U.S. Geological Survey test hole indicate that the sand may be hydraulically connected to recharge areas in Hogback Mountain through permeable deltaic or fan deposits which form a ridge extending southward from the mountain. The relatively high water levels in this ridge are apparent on plate 1.

Logs of wells in the area suggest that the permeable deposits just described are also connected to the permeable terrace deposits on the north face of Stukel Mountain, as well as to subsurface stream gravels fanning out from Olene Gap. If so, the hot-water areas of Hogback Mountain, Stukel Mountain, and Olene Gap may be hydraulically connected through the valley sediments in a zone that extends to depths of 1,000 feet or more.

Springs are not abundant in the Klamath Falls area, and most springs appear to have no connection with hotter water at depth. The 15 springs shown on plate 1 constitute nearly all the identifiable perennial springs that issue at land surface. Additional springs issue from the bottom of Upper Klamath Lake, Spring Lake, Nuss Lake, and probably other small ponds, but these have not been separately identified on the map.

Most springs in the area produce cold water that is supplied by gravity flow from precipitation on adjacent recharge areas. Examples are Hummingbird, Shell Rock, Barkley, and Neubert Springs north of Klamath Falls, and an unnamed spring in the southwestern part of the area (T41, R8, Section 5cbb), all of which issue from basalt rocks near the base of fault scarps. Eagle Point Spring and two springs in Olene Gap are warm artesian springs whose waters are partly derived from the geothermal reservoir. Temperatures of thermal springs in Upper Klamath Lake are not known but are assumed to be fairly low. Springs with temperatures near the boiling point formerly issued along the base of the slope below the hot-well area of Klamath Falls, but flows from these springs gradually diminished during the past several decades and the springs no longer exist.

The general absence of springs on the flanks of upraised fault blocks confirms other evidence that water tables are relatively deep beneath the surfaces of

these blocks. Seeps occur following precipitation on the mountains, but these are mostly short lived and the ground water is only temporarily perched above them.

DEEPER BASALTIC AQUIFERS

A few deep wells located on ridges near Klamath Falls penetrate basaltic rocks of probable Tertiary age. Four such wells clustered at the north end of the city of Klamath Falls are among the deeper hot-water wells in the area. A diagram showing profiles of these wells, along with three nearby cooler wells (fig. 4), suggests possible relations to faults and illustrates the compartmentation of the basaltic aquifers at this location. As the diagram illustrates, static water levels in the hot wells are accordant with each other but are at depths ranging from 170 to 200 feet below static levels in the three cooler wells and about 100 feet below the average level of Upper Klamath Lake. The water level in well OIT 3 is anomalously high, suggesting the presence of a local source of recharge and a high degree of compartmentation at this site.

An aquifer test has shown that the four hot wells (left side of diagram, fig. 4) are hydraulically connected through permeable basalt rocks. (See section on "Hydraulic Properties of the Rocks.") Water levels in these wells, therefore, probably represent the hydraulic head in the local geothermal system, which is assumed, from chemical evidence, to be largely sealed off from the shallow cold water.

Similar relations apparently occur in the hot-well area of Klamath Falls, where water-level altitudes in most hot wells are also lower than water levels in nearby cool wells (Lund and others, 1974; J. W. Lund, oral commun., 1976). It is concluded, therefore, that hydraulic heads in the shallow bedrock aquifer are generally higher than those in the geothermal aquifer over an extensive area of the Klamath Falls thermal area, although local departures from this generalization may occur.

One additional well that penetrates deep Tertiary basaltic rocks was inventoried in the Klamath Falls area. This is a 1,305-foot well on the north flank of Hamaker Mountain (T40, R7, Section 11ccc). The well produces water at a temperature of 16°C from fractured basalt flows and fragmental interflow zones at depths between 1,260 and 1,288 feet below land surface, corresponding to altitudes of 3,600 to 3,572 feet above sea level. Static water level in the well was reported by the driller in 1959 to be 1,193 feet below land surface at an altitude of 3,670 feet above sea level. The reported water level is 400 feet lower than water levels in shallow wells that penetrate bedrock at locations 2 to 3 miles to the northeast.

The low static head in the well at Hamaker Mountain is probably due to compartmentation of the aquifers by parallel faults that follow the regional trend through the area. A major fault has been mapped as striking through Bear Valley, just northeast of the well (Murray Gardner, written commun., 1975). The fault block is downthrown to the southwest, and numerous faults west of the main fault also have downthrown blocks to the southwest. Thus, water levels may be stepped down toward the southwest in partly isolated compartments created by the faults.

Little information is available on aquifers in Tertiary basaltic rocks in the highlands immediately north and east of Klamath Falls. Tertiary rocks that occur outside the study area in the Sprague River, Swan Lake, Langell, Poe, and Yonna Valleys were described by Leonard and Harris (1974). Hydraulic characteristics of these rocks have been evaluated by Illian (1970), and they appear to be similar to those of the deeper basaltic rocks underlying the Klamath Falls area, as confirmed by the few data obtained in the present study.

MOVEMENT AND DISCHARGE OF GROUND WATER

The water-level contours on plate 1 show that ground water in shallow aquifers near Klamath Falls moves generally southward toward Tulelake in California. Some of the flow, perhaps on the order of 80,000 acre-ft per year, is intercepted by Lake Ewauna and the Klamath River, and leaves the basin as streamflow. Most of the remainder enters the closed basins of Lower Klamath Lake and the Tulelake sump.

On the flanks of the northwest-southeast trending fault blocks, water-table gradients are steep, ranging from about 0.027 (140 ft mi^{-1}) on the southwest side of Stukel Mountain to about 0.038 (200 ft mi^{-1}) west of Lake Ewauna. These gradients reflect the steep surface topography as well as apparently low vertical permeabilities within the fault blocks.

South of Altamont, the moderate gradient of about 29 ft mi^{-1} suggests that rather permeable rocks extend into the valley from the southern flank of Hogback Mountain. Lithologic descriptions in drillers' logs confirm this suggestion. Similar or even smaller gradients were observed along the northwest edge of the valley between Klamath Falls and Keno. These gradients indicate that ground water moves freely parallel to the regional structure.

In areas near the principal hot-water bearing faults, thermal waters may move upward, discharging both water and heat to the near-surface environment. In larger areas surrounding the faults, the flow of hot water appears to be predominantly lateral. One example is a region of several square miles bordering the
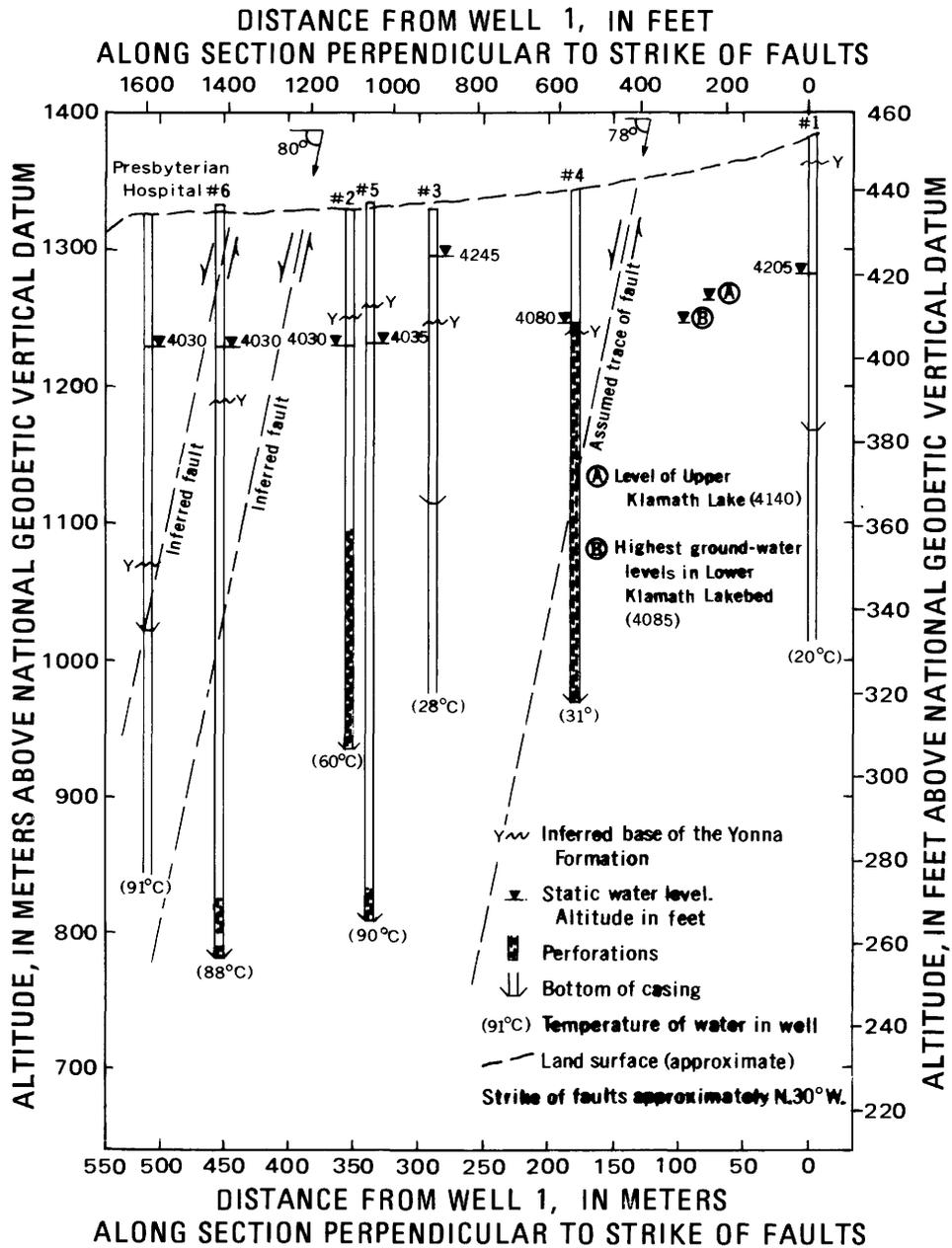
FIGURE 4.—Profiles of wells at the Oregon Institute of Technology and Presbyterian Intercommunity
Hospital, showing discordant water levels and assumed relations to faults.

south edge of the city in which warm water from the hot-well area is believed to flow southward in permeable zones of layered basalt. Similar lateral flow may occur in the area west of the Klamath River and north of the Weyerhauser sawmill. The movement of thermal water is discussed further in the section on "Temperature Distribution and Heat Flow."

Large amounts of ground water probably move from one subbasin to another within the Upper Klamath River basin. Illian (1970, p. 55) estimates that 200,000

acre-ft per year enters the Lost River subbasin from the Upper Klamath Lake subbasin, and that 150,000 acre-ft per year enters from the Sprague River subbasin. Leonard and Harris (1974) conclude, however, that, although a significant amount of flow does occur, the available data do not permit an accurate quantitative evaluation of the interbasin movement.

A crude estimate of the amount of ground-water flow from the Upper Klamath lake basin into Lower Klamath Lake has been made on the basis of data from

Illian and estimates derived from the present study. From data on wells penetrating Tertiary basaltic aquifers, Illian (1970, p. 13) calculated an average specific capacity of 145 gallons per minute per foot of drawdawn. A rough estimate of transmissivity may be obtained from values of specific capacity in efficient wells by multiplying gallons per minute per foot by 2000. On this basis, and converting to square feet per day, the transmissivity of the Tertiary basalt aquifers is estimated to be 40,000 ft^2 d^{-1}. Assuming that flow occurs through a 20-mile section extending along the 4,100 foot water-level contour on plate 1, and that the average gradient is 26 ft mi^{-1}, the volume of ground-water flow into the Lower Klamath Lake basin is calculated to be about 175,000 acre-ft per year.

The estimated transmissivity, based on data from highly permeable basalt aquifers in subbasins north and east of Klamath Falls, may be an overestimate for the Klamath Falls area. A small amount of pumping-test data from Klamath Falls suggests that the basalt aquifers underlying the city are less permeable than those in nearby basins. The assumed gradient is probably within 25 percent of the actual value for the upper 1,000 to 1,500 feet of the aquifer, but may be significantly higher than the actual gradient in a deep, regional flow system. The 20-mile section is a maximum probable length. The estimate of ground-water flow, therefore, may be near the upper limit for the actual value.

A similar approach has been used to estimate ground-water flow out of the Lower Klamath Lake basin, but the uncertainties involved are much greater. Southwest of Klamath Falls, in the direction of the regional topograpic gradient, ground-water conditions in the closely-spaced parallel fault blocks are virtually unknown, and it is not possible to make realistic estimates of either transmissivities or gradients in this direction. The figure of 100,000 acre-ft per year used in the section on "Evapotranspiration" is, therefore, little more than a guess at a probable maximum ground-water flow from the basin.

Within the area studied for this report, ground water is discharged at land surface by pumping from wells, by artesian flow from wells and springs, and by seepage to the land surface or into lakes and streams. No attempt has been made to estimate the total discharge into lakes and streams, but rough estimates of the remainder of the ground-water discharge have been made. The results are shown in table 3.

The total discharge to the land surface of ground water from springs and wells in the Klamath Falls area is estimated to be about 21 Mgal d^{-1}, or about 22,000 acre-ft per year. This amount is an extremely small fraction of the water that enters the area either

TABLE 3.—*Summary of ground-water discharge and use in the Klamath Falls area*

	Discharge and use (Mgal d^{-1})	Percent of total
Public Supply[1]:		
Klamath Falls, Altamont, Merrill, and Falcon Heights	7.3	35
Irrigation	5.8	29
Industrial and commercial	3.0	14
Spring discharge	2.3	11
Space heating[2]	1.9	9
Domestic (individual wells)	.2	1
Stock watering	.1	<1
Total	20.6	

[1] As of 1975.
[2] Interpreted in part from data in Lund, (1978).

in streamflow or precipitation. (See section on "Climate and Drainage.")

HYDRAULIC PROPERTIES OF THE ROCKS

ESTIMATES FROM SPECIFIC CAPACITIES OF WELLS

The specific capacities of efficient wells afford crude estimates of the transmissivity of the aquifers penetrated by the wells. Specific capacity, expressed as yield in gallons per minute divided by the drawdown in feet, has been calculated from data provided by drillers and owners for about 170 wells in the Klamath Falls area. These data and estimated transmissivities derived from them are shown in table 4. Because of widely differing well efficiencies and depths of aquifer penetration, estimates based on specific capacity are undoubtedly imprecise. They are used in this report solely for purposes of comparison.

In table 4, wells are assigned to one of several categories of rock type on the basis of drilling information. The rock types listed do not conform to the geologic units of Peterson and McIntyre (1970), but have been placed in generalized categories corresponding to commonly used terms in drillers' logs. The assignment of many wells to specific categories of water-yielding rocks is difficult, owing to uncertainties in drillers' logs or in geologic mapping, and the calculated numbers in table 4 thus contain some uncertainty. Where the number of wells is small, reassignment of one high-yielding well would make a large difference in the average specific capacity for the rock unit. The estimated transmissivities seem reasonable, however, in the light of reported yields of wells and observations of rock characteristics in outcrops.

The data of table 4 show that several types of rocks in the area provide similar low yields to wells. Wells in alluvial terraces, Holocene and Pleistocene lake deposits, "older" lake deposits, and rocks of volcanic eruptive centers have specific capacities in the narrow range 1.6

TABLE 4.—*Specific capacities of wells and estimates of transmissivity in aquifers of the Klamath Falls area*

Water-yielding rock unit	Number of wells	Average depth (ft)	Average specific capacity (gal min^{-1}ft^{-1})	Estimated transmissivity of aquifer (ft^{2}d^{-1})
Alluvial terrace deposits.	4	279	1.8	500
Holocene and Pleistocene lake deposits.	24	194	1.6	400
Rocks of Quaternary basaltic flows.	6	406	19	5,000
Fragmental basaltic rocks of Quaternary and Tertiary eruptive centers.	27	187	2.1	600
Older lake deposits (Yonna Formation?)	21	334	1.9	500
Yonna Formation	70	450	29	8,000
Rocks of Quaternary and Tertiary basaltic flows.	18	869	37	10,000*

*Aquifer tests at three sites yielded values ranging from 11,000 to 20,000 ft^2 d^{-1} at prevailing temperatures. (See text for temperature-corrected estimates.)

to 2.1 gal min^{-1} ft^{-1}. Although these rock types differ greatly in physical characteristics, it is not possible to distinguish one from the other on the basis of well performance.

The distinction between the categories "Rocks of Quaternary basaltic flows" and "Fragmental basaltic rocks of Quaternary and Tertiary eruptive centers" in table 4 is especially arbitrary because of disagreements among published maps and uncertainties in drillers' descriptions of the rocks. However, fragmental rocks of either Quaternary or Tertiary age appear to have similarly low permeabilities, whereas the basalt flows, regardless of age, are apparently much more permeable. As noted above, the permeable zones in the Yonna Formation are also associated with basalt flows.

Rocks labeled "Older lake deposits" in table 4 appear to be partly consolidated clay, silt, and diatomite deposits. These rocks have been identified only in drillers' logs and they are distinguished from rocks of the Yonna Formation primarily on the basis of their low transmissivities. Most wells identified as penetrating the Yonna Formation have fairly high transmissivities which are apparently due to the presence of basalt flows and breccias. Thus, the presence or absence of volcanic rocks may be the only significant feature that distinguishes the category "Yonna Formation" from "Older lake deposits" in table 4.

Drillers's logs indicate that the moderately high transmissivities in basaltic rocks of the Klamath Falls area are probably due largely to fractures rather than to permeable interflow zones such as occur elsewhere in many basaltic sequences. Few of the drillers' logs describe the exact nature of the water-bearing zones, however, and it is possible that "caving" or "soft lava"

rocks referred to by the drillers represent scoriacious or vesicular tops of flows which may also be relatively permeable.

Two of the specific capacities listed in table 4 may be compared with those given by Illian (1970) for wells in the northern Klamath River basin. For wells in sedimentary aquifers, Illian reported an average specific capacity of 0.45 gal min^{-1} ft^{-1}, a value only about 1/4 as great as the corresponding value in table 4. For the "Lower Basalt Aquifer," Illian reports an average specific capacity of 145 gal min^{-1} ft^{-1} a value almost 4 times greater than the value for Quaternary and Tertiary basaltic rocks in table 4. Although these discrepancies may represent real differences between aquifers, it seems more likely that they reflect statistical discrepancies resulting from small sample populations, uncertain geologic classifications, and error resulting from inefficient and partly-penetrating wells.

ESTIMATES FROM PUMPING TESTS IN WELLS

Estimates of transmissivity in Tertiary basaltic aquifers were derived from three pumping tests in the Klamath Falls area. The first test was conducted in 1956 by Robinson and Roberts, Ground-water Geologists, Tacoma, Wash., for the Weyerhaeuser Timber Company. The test was run in Weyerhaeuser well 4 (T39, R9, Section 8cbb) which is open in the lower basalt aquifer from 276 to 545 feet below land surface. Analysis by the writer of data from this test suggests that transmissivity of the aquifer at the Weyerhaeuser location is about 11,000 ft^2 d^{-1} at prevailing temperatures.

A second aquifer test was performed in 1974 by the U.S. Geological Survey, assisted by staff and maintenance personnel of the Oregon Institute of Technology. Three thermal wells penetrating Tertiary basaltic rocks were used for the test: well OIT 5 was pumped while observations of water levels were made in well OIT 6 and a well at Presbyterian Hospital (fig. 4). The three wells range in depth from 1,584 to 1,805 feet and had temperatures at the time of the test ranging from 88° to 91°C. The responses of the observation wells demonstrate that all three wells are hydraulically connected in a relatively permeable formation. The drawdown data provide a poor fit to type curves based on standard analytical models, but the general shapes of the curves and the magnitudes of the drawdowns suggest that the apparent transmissivity is about 22,000 ft^2 d^{-1}. A factor of approximately 0.3 should be applied to this value in order to reduce temperature, viscosity, and density to standard values for comparison with nonthermal aquifers. Thus, the actual transmissivity of the aquifer may be less than 7,000 ft^2 d^{-1} for an aquifer at 16°C.

A third pumping test was performed in 1976 by the staff of the Geo-Heat Utilization Center, OIT, in an artesian well at the County Museum, Klamath Falls. During this 28-hour test, water levels were measured in 12 nearby wells.

The Museum well, which is reportedly 1,234 feet deep, penetrates the Yonna Formation and underlying Tertiary basaltic rocks at a location near the southwestern limit of the main hot-well area of Klamath Falls. Observation wells for which data could be analyzed penetrate the Yonna Formation (including some basalt flows) at depths shallower than 500 feet. Water temperatures in the observation wells ranged from 56 to 85°C. Distances from the pumped well to the 12 observation wells ranged from 123 to 1,425 feet.

Analysis of the drawdown data by the author shows that all the wells measured are hydraulically connected through rocks of the Yonna Formation and the underlying basalts. Apparent transmissivity is about 20,000 ft^2 d^{-1} at prevailing temperatures, although some of the less reliable data suggest that the value might be significantly higher. Converting to a standard temperature of 16°C, transmissivity is estimated to be about 10,000 ft^2 d^{-1}.

The storage coefficient of the aquifer, a dimensionless number that expresses the volume of water released from, or taken into, the aquifer per unit surface area per unit change of head, was also estimated from the drawdown data. The storage coefficient is probably in the range 0.001 to 0.01.

Estimates of transmissivity derived from the three aquifer tests are close to the average value calculated from specific capacities in 18 wells (table 4). The data indicate, therefore, that the average transmissivity of shallow Tertiary basaltic aquifers in the vicinity of Klamath Falls is moderately high (about 10,000 ft^2 d^{-1}), but is significantly lower than that calculated for Tertiary basaltic aquifers in adjacent basins (Illian, 1970, p. 13). Estimates from both thermal and nonthermal wells at Klamath Falls are similar, indicating that for depths less than about 1,500 feet, no distinction can be made between thermal and nonthermal aquifers on the basis of well performance.

CHEMISTRY OF GROUND WATER

GENERAL CHEMICAL CHARACTERISTICS

Ground water of the Klamath Falls area has generally low concentrations of dissolved-solids and is of good quality for most uses. The thermal waters are of poorer quality, although concentrations of dissolved solids in the hotter waters, estimated from specific electrical conductance, are generally no greater than 1,000 milligrams per liter (mg L^{-1}). Both thermal and nonthermal waters have pH in the range 7.5 to 8.5 and are, therefore, mildly alkaline.

Chemical analyses of water from 59 wells and springs in the vicinity of Klamath Falls are listed in table 5. Differences in the quality of ground water in the area are shown by contours on the values of specific electrical conductance on plate 2. The chemical character of the water is indicated on plate 2 by Stiff diagrams, and the dominant chemical characteristics are shown in a diagram of percent reacting values of major cations and anions (fig. 5).

In figure 5, the open circles represent cold, dilute waters of the area. These samples are typical of cold water derived from basaltic rocks and shallow volcanic sediments. Temperatures of these samples range from 10° to 14°C. The crosses in figure 5 represent hot waters, nearly all of which issue from basaltic rocks and associated volcanic sediments. Temperatures of hot water sampled in this study ranged from 61° to 93°C. Solid circles in figure 5 represent samples collected from 25 of the hundreds of warm wells and three warm springs scattered throughout the area. These waters, produced mostly from the Yonna Formation and alluvial and lacustrine deposits, appear to be mixtures of recent metoric water and warm water of deeper circulation. Temperatures of these samples ranged from 15° to 37°C. Waters in this category are referred to in this report as transitional waters. A fourth category, warm waters from the Klamath Hills, comprises transitional waters whose chemical characteristics differ greatly from other warm waters of the area. These waters are discussed in following sections of this report.

The principal ionic constituents of the cold waters are calcium, magnesium, and bicarbonate (figure 5). Concentrations of dissolved solids are generally low, and specific electrical conductance in the waters sampled ranged from 110 to 430 μmho cm^{-1}.

The transitional waters are also predominantly calcium-magnesium-bicarbonate waters which, as shown in figure 5, contain greater amounts of sodium plus potassium than the six cold waters. In actuality, this increase is due almost entirely to higher concentrations of sodium. The trend of increasing sodium corresponds generally to increasing temperature in the transitional waters. It is apparent in figure 5 that there is no corresponding increase in sulfate and chloride in the transitional waters as would be expected in a simple mixing trend between cold and hot waters.

The hot waters shown in figure 5 are clearly distinguishable from both cold and transitional waters. Their positions on the graph reflect greatly increased concentrations of sodium, sulfate, and chloride relative to the cold waters. The samples of hot water obtained

PERCENT REACTING VALUES
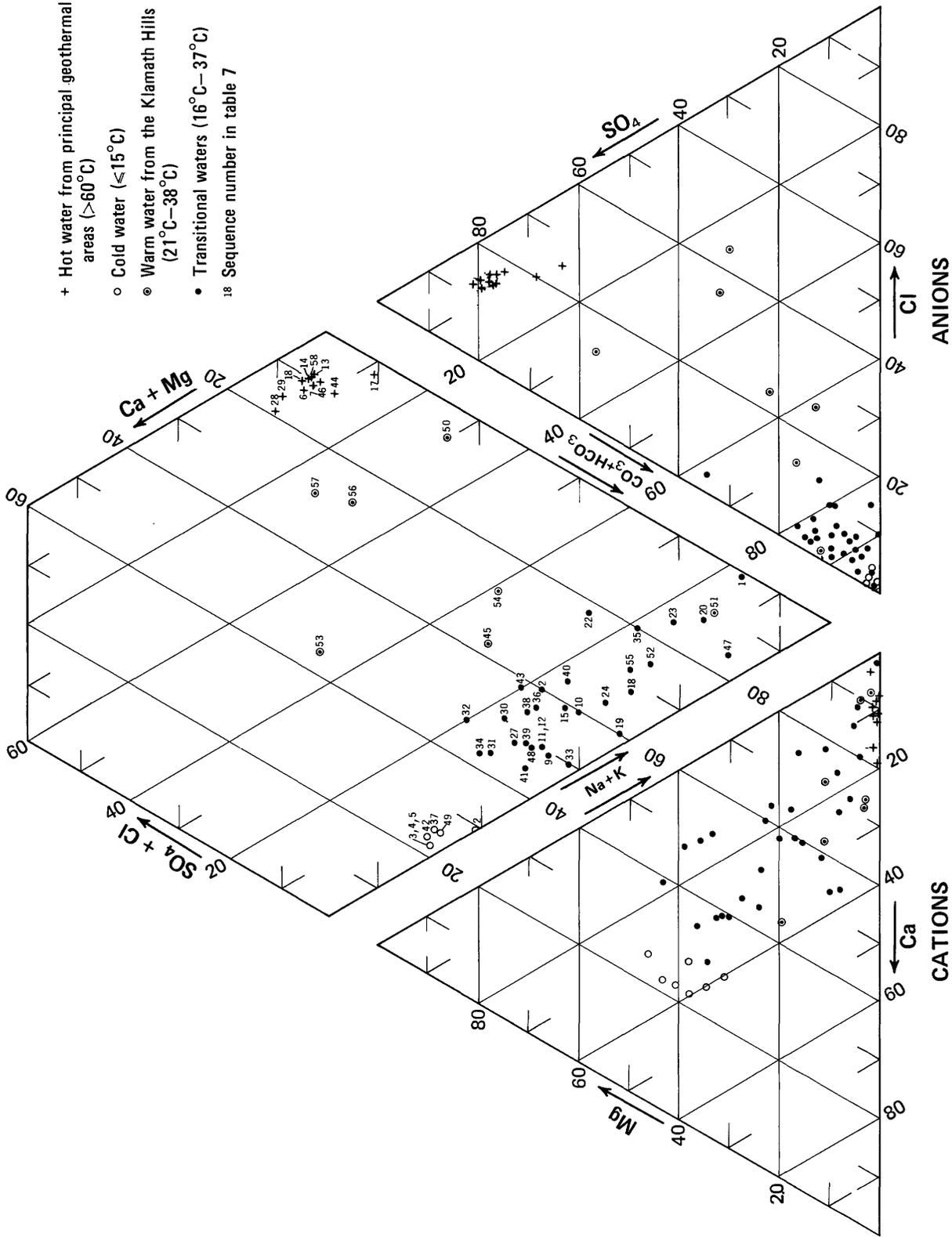
FIGURE 5.—Percent reacting values of major ions in ground water. Points on the diamond-shaped field represent total cations plus total anions; they are projected from the small triangular graphs at the lower left and right.

TABLE 5—Chemical analyses of water from wells and springs

[Concentrations in milligrams per liter unless otherwise indicated. Within each category, samples are listed in order of decreasing temperature]

Seq. No.	Owner or name	Location	Temp. °C	Ca	Mg	Na	K	HCO3	CO3	SO4	Cl	F	SiO2	As μg L⁻¹	B μg L⁻¹	Fe μg L⁻¹	Li μg L⁻¹	Mn μg L⁻¹	Specific conduct μmho cm⁻¹	pH	Date Sampled	Sources of data
Hot waters (>60°C)																					
46	Jack Liskey	40/9-34aca	[c]93	15	<0.1	200	4.0	[c]48	[c]2	360	59	1.5	90	--	650	260	[b]80	--	[c]1,030	[c]8.9	05-09-74	c,h,i
44	O. H. Osborn	40/9-27cda	[c]90	15	[c].0	190	4.1	[c]37	[c]9	270	56	1.5	90	11	770	--	--	0	[c]920	9.5	05-31-74	c,i
13	Mills School	38/9-33acb	89	27	<.2	370	3.0	46	--	462	54	1.2	78	--	960	30	--	30	--	8.3	--	f
6	OIT Well No. 6	38/9-20ach2	74 / [b]88	24.2	<1	195	3.9	[b]44	[b]0	400	58	1.45	31? / [b]90	--	1,000	--	[b]130	--	1200	[b]8.2 / 7.7	08-07-75 / 08-24-72	c,h / b,d
28	Olene Gap Spg.	39/10-14cbd	[c]79 / 87	40	.2	190	7.2	53	0	400	59	1.2	98	--	1,000	30	[b]120	--	1,050	8.0	03-31-75 / 03-09-74	h / h
58	Assembly of God	38/9-33aba	87	28	.1	213	4.6	61	8	393	54	1.2	110	--	820	140	--	--	1,400	7.8	03-28-75 / 08-10-75	c / o
7	Medo-Bell Dairy	38/9-28ccc	81	23	.0	213	4.2	32	8	403	54	1.6	81	--	960	40	--	--	1,300	8.8	08-27-76	j
8	J. E. Friesen	38/9-28cdc	73	25	.0	221	4.4	32	8	431	56	1.4	87	--	910	--	--	--	1,120	8.7	01-24-55	e
14	Lois Serruys	38/9-33bbb	71	22	.0	207	3.8	43	4	393	50	1.5	83	--	740	<50	--	--	1,230	8.5	02-19-55 / 12-22-54	e
29	L. Jones (Spg)	39/10-14ccb	65	35	.1	217	5.0	[b]50	[b]0	430	65	1.5	48?	--	1,900	--	--	--	1,100	[b]8.2	10-11-72	b,d
17	Mazama School	39/9-11bba	61 / [c]58	5.4	1.0	246	6.0	120	0	256	35	1.0	[b]92 / [b]118	--	--	90	--	<20	1,300 / [c]895	8.3 / 8.9	03-09-74 / 08-08-75	f / c,h
Warm waters of the Klamath Hills (21°C–38°C)																					
50	Jack O'Connor	41/9-1bcc	[c]38	7.5	.6	130	3.0	[c]110	[c]2	140	50	.3	65	--	160	--	--	--	[c]640	[c]8.3	05-09-74	c,i
56	O'Connor Livestock Co.	41/10-7dba	[c]40 / [c]30	58	6.4	210	7.0	[c]250	[c]0	210	160	.0	80	1	690	40	--	150	[c]660 / [c]1,300	[c]8.2 / 7.8	08-07-75 / 06-06-74	c,i / c,i
45	Abe Boehm	40/9-28aca	[c]25 / [c]22	180	47	480	18	[c]1,460	[c]1	300	170	.2	100	--	1,400	100	[b]160	2,400	[c]2,700 / [c]2,400	[c]7.1 / [c]7.6	05-30-74 / 05-09-74	c,h,i / c,i
51	Jack Liskey	41/9-2baa	[c]25	35	16	580	12	[c]1,550	[c]0	13	140	.4	100	--	1,700	90	[b]280	--	[c]660	[c]7.6	08-11-75	c,i
57	George Carter	41/10-14bba	[c]22	29	2.3	99	7.9	[c]100	[c]0	95	98	.0	41	0	150	20	--	20	670	[c]8.2 / 7.5	06-06-74 / 03-19-71	c,i / g
53	Suburban Water Co.	41/10-1cdd2	16	59	20	78	6.0	230	0	79	57	.08	54	--	150	500	--	--	740	--	03-29-75	g
54	Town of Merrill	41/10-2ddd	21 / 23	11	4.3	44	3.8	114	0	21	27	.1	38	--	[c]150 / 10	120	[b]20	--	315 / 305	[c]7.8 / 8.3	08-11-75 / 02-18-55 / 04-11-74	c,h,i / e / e
Miscellaneous warm waters (20°C–37°C)																					
16	Moyina Water Co.	39/9-1bda	>50? / [c]37	7.9	2.3	38	5.0	--	[c]0	5.6	1.8	.22	64	<5	100	420	--	<50	220	7.4	08-24-66	g
22	Falcon Heights School	39/10-34cbd	[c]37	7.7	4.5	57	7.1	[c]170	[c]0	24	8.9	.2	80	3	100	190	--	30	[c]350	[c]7.8	05-07-74	c,i
1	Eagle Point Spg.	36/7-23abb	[c]35	0.6	<0.1	62	5.7	[b]136	[b]0	<2	16	.75	38?	--	140	--	[h]<20	--	[c]305	[b]8.3	08-22-72 / 04-06-75	b,d / c,h
20	U. S. Air Force	39/9-34bcc1	[c]31	3.8	1.2	47	9.0	128	0	8	6.0	.3	[h]78	--	--	40	--	<10	240	7.0	10-17-72 / 12-12-73	i
21	U. S. Air Force	39/9-34bcc2	30	5.6	2.3	48	8.5	145	0	9	6.3	.2	51	--	--	30	--	<10	255	--	10-17-72 / 12-12-73	c
23	Alfred Jacobson	39/9-34ddc	30	5.7	2.4	49 / [b]36	12	[c]160	[c]0	13	7.9	.1	[b]65	1	50	120	--	50	290	[c]7.6	05-20-74	c,i
24	Oregon Water Corp.	39/10-6cdb	26	8.5	5.8	36		--	--	6.5	3.0	.32	63	--	--	330	--	<5	--	8.6	08-27-70	g
39	Monte Dehlinger	39/10-32cad	[c]26	17	11	22	6.7	[c]160	[c]0	7.1	3.6	.1	54	4	20	20	[h]<20	0	[c]260	7.6	06-06-74	c,i
15	Howard Holliday (Spg)	39/8-13eab	[c]25	12	3.1	27	2.9	[c]120	[c]<1	6.4	5.1	.1	27	2	40	80	--	0	[c]220	7.9	05-22-74	c,h,i
25	Melvin McCollum	39/10-7aba	[c]25	24	13	20	4.0	[c]120	[c]0	53	3.1	.2	52	2	20	400	--	70	[c]300	[c]7.5	05-28-74	c,i
36	Lester Brookshire	39/10-29dcd	[c]25	13	2.0	21	3.4	[c]95	[c]<1	10	2.8	.1	35	2	6	20	--	20	[c]175	[c]8.1	05-04-74	c,i
38	George Stacy Co.	39/10-32abd	[c]25	13	2.0	20	4.0	[c]94	[c]1	11	2.0	.0	34	1	10	40	--	10	[c]170	[c]8.3	06-05-74	c,i
26	Claude Shuck	39/10-32dbd	[c]24	9.7	4.1	25	6.2	[c]110	[c]<1	15	2.8	.2	57	2	20	50	--	10	[c]220	[c]8.2	05-30-74	c,i
43	Dan O'Connor	40/9-23bab	[c]24	12	6.6	32	8.1	[c]130	[c]0	14	9.1	--	42	--	--	--	--	--	[c]260	[c]7.6	05-07-74	c,i

	Owner	Location																					Date	Temp	Source
47	Clyde Dehlinger	40/10-5dde	24	8.1	3.1	40	6.5	150	0	2.3	4.9	.0	72		40	20		0	240	7.9	06-05-74	c, i
30	Robert Langley	39/10-15dcc	23	10	6.3	22	5.5	110	1	14	2.6	1	50		20				200	7.4	05-04-74	c, i
40	J. K. O'Neil	39/10-34bbd	24	6.1	5.5	26	5.1	110	1	13	2.8	2	50		30	50			205	7.8	08-10-75	c, i
12	Oregon Water Corp.	38/9-32bcd3	23	11.2	5.8	26						2	25					<10	190	7.4	05-04-74	c, i
			24							.6	5.3								205	7.7	08-10-75	g
18	Weyerhaeuser Well No. 4	39/9-18cbb	22	10.0	1.9	32	3.3	115	4.5	4	4	.11	16?		90				205	8.4	09- -71	c
			22																		08-08-75	b, d
																		205		08-22-72	
27	Ray Bixler	39/10-10bcd	22	12	8.4	16	3.3	110	0	7.9	1.9	2	49	2	20	160		1,400	200	8.3	10-16-73	c
55	Pope's Meat Co.	41/10-3aaa2	22	8.6	1.4	30	3.2	100	2	6.0	4.4	.0	25	0	20	20		0	190	7.8	05-29-74	c, i
10	Oregon Water Corp.	38/9-32bcd1		10	4	29		117	0	0	8		46			<50			190	8.5	06-06-74	g
																					03-03-30	c
11	Oregon Water Corp.	38/9-32bcd2	21	11.2	6.8	26				1.2	5.3	.01	24			80		<10	200	8.3	08-08-75	g
			21																		09- -71	
35	Len Dobry	39/10-28dba	21	5.7	1.1	32	4.5	100	0	14	3.4		65		30				195	7.6	05-04-74	c, i
48	Bill Hill	40/10-26bca1	22		13	19	6.4	140	0	6.4	4.2	.2	26	0	20	40		30	210	7.7	08-08-75	c, i
			20	9.1								.0							245	7.7	06-06-74	c, i

Cold waters (<20°C)

31	Roe Kinney	39/10-15ddd	19	11	7.0	17	4.9	110	0	10	1.7	.2	49	4	30	350		10	190	7.9	05-29-74	c, i
32	Don Kirkpatrick	39/10-17baa	18	11	8.9	23	5.0	130	0	21	3.1	.0	68	3	30	360		80	260	7.4	05-31-74	c, i
34	Paul W. Sharp (Spg.)	39/10-22dcb	18	13	6.6	19	4.0	110	1	10	2.2	.3	43	3	30	20		0	200	8.3	06-06-74	c, i
33	Gabriel Bros.	39/10-22bcb	16	12	18	42	9.6	390	0	3.0	3.0	.3	95	1	70	220		60	650	7.3	05-30-74	c, i
41	U. S. Air Force	40/7-11ccc	16	14	11	18	8.8	141	0	3.4	2.7	.1	42			20		<10	225		10-17-72	c, h
9	Oregon Water Corp.	38/9-30acb	17	14.4	9.8	22				1.2	4.5	.01	27				<20		255	8.0	08-12-75	g
			14																250		09- -71	c
19	Sam Shaw	39/9-30ccd1	15	18	15	69	5.5	360	1	3	6.8	3	63	1	100	130		50	565	7.4	08-08-75	c, i
42	Vern Howard	40/8-7dac	15	24	17	17	1.6	200	3	2.3	2.3		31	3	9	3,000		120	340	8.6	05-27-74	c, i
37	Ben Adair	39/10-30aaa	14	23	23	21	4.2	270	1	2.4	4.5	.3	45	0	30	50		240	430	8.0	05-31-74	c, i
49	Tulana Farms (Spg)	41/8-5cbb	14	9.2	4.9	5.7	1.2	60	0	<2	1	.07	24		<50					6.9	07-26-72	b, d
2	Shell Rock Spg	36/8-27cdd	13	8.1	3.9	5.9	1.5	50	0	<2	<1	.1	32?						115		08-08-75	c
			12										60		<50				110	6.4	08-06-75	b, d
5	Hummingbird Spg	37/9-7adc	7	10.6	8.2	7.0	1.2	96	0	<2	1	.1	21?		<50				96	8.2	07-26-72	c, b, d
52	Murel Long	41/9-12adb	12	22	25	220	4.0	720	1	51	16		42						160	7.4	08-06-75	c, i
			12										45		1,100				1,070	7.7	05-09-74	c, i, --
4	Barkley Spg	37/9-6bcd	14	9.2	6.5	8.0	1.2	84	0	<2	1	1.5	18		70				1,000	8.3	08-10-75	b, d
			11									.1							135		07-26-72	b, d
3	Neubert Spg	37/8-9dcc	11	12.1	7.2	7.6	1.3	94	0	<2	1	.11	24?		<50		<20			7.2	04-02-75	b, d
			10										48						150	6.8	08-06-75	c, h

for this study are typical of water in the more than 400 hot wells used for space heating in Klamath Falls (Lund, 1978) as well as hot water from Olene Gap and the Klamath Hills.

In two areas of the lower Klamath Lake valley, ground water occuring at depths of 1,000 feet or more has generally high values of specific electrical conductance and is, for the most part, unfit for drinking. These areas are delimited on plate 2 by two 1,000 μmho contours located north and south of Miller Hill. No chemical analyses are available from these areas, but owners' reports and field observations indicate that much of the water contains large amounts of iron, sulfate, and methane gas, as well as traces of hydrogen sulfide gas. Temperatures of water in the northernmost area are generally higher than normal. Drillers' logs indicate that the water is contained in fine-grained, organic-rich sediments depleted in oxygen. These parts of the basin quite possibly were marshy sumps and areas of evaporative ground-water discharge during long periods of late Pleistocene and Holocene time.

RELATIONS AMONG DISSOLVED CONSTITUENTS

Concentrations of dissolved constituents in ground waters of the Klamath Falls area differ widely, depending on the lithology of the aquifer, the temperature of equilibration, and the amount of mixing that has occurred between thermal and nonthermal waters. Some of these relations are discussed in this section of the report. The relations of common cations and silica to temperature are discussed more fully in a separate section below.

Cold ground water derived from shallow alluvium and basaltic rocks in the Klamath Falls area is a calcium-magnesium bicarbonate water which originates in local precipitation and acquires small concentrations of dissolved solids mainly by solution of feldspars and other silicate minerals in the volcanic rocks. Silica concentrations are relatively high, commonly as great as 40 to 50 milligrams per liter, indicating that even in basaltic rocks abundant silica is available for solution.

The chemical character of water in the basaltic rocks does not change significantly even after penetration to great depths. For example, water obtained at a depth of about 1,300 feet in a well on the northeast flank of Hamaker Mountain (sample number 49, table 5 and pl. 2) has acquired slightly increased concentrations of dissolved solids relative to meteoric water but is not greatly different in quality from waters that presumably have penetrated less than a few hundred feet before emerging at the surface in gravity springs (samples 2, 3, 4, and 5).

Hot waters from the three principal geothermal areas are represented by 10 samples having temperatures greater than 65°C. An additional sample, number 17, with a temperature of 61°C, plots below the main group in the diagram of figure 5 but is sufficiently similar to warrant inclusion with the hot waters in most of the discussions below.

The hot waters are obtained from wells and springs that tap both the Yonna Formation and underlying basaltic rocks. These waters probably originate in local precipitation. If this supposition is true, the meteoric waters, in their passage through the geothermal reservoir, acquire increased amounts of calcium, sodium, potassium, chloride, sulfate, and silica, while concentrations of bicarbonate decrease somewhat and magnesium is reduced almost to zero. Changes of this kind are commonly found in hot-water geothermal systems (White, 1970), and they are compatible with the assumption that the Klamath Falls thermal waters are locally derived.

In hot waters from the principal geothermal areas, concentrations of major ions generally differ more widely among samples from the same area than among the three areas. Ratios of the major ions in hot waters show no consistent differences from one geothermal area to another. Some of the ratios that are of particular interest in geothermal waters are given in table 6.

Molal ratios of Ca/Mg, Na/Ca, and Na/K are significantly higher in the hot waters than in the cold waters and ratios of Na/Cl and Ca/Cl are lower. High ratios of calcium to magnesium are generally characteristic of high-temperature geothermal systems, but may occur at temperatures as low as 100°C (White, 1970, p. 70). High ratios of sodium to potassium also imply elevated reservoir temperatures, but suggest that the waters equilibrated at temperatures not much higher than the measured temperatures (Mariner and Willey, 1976; White, 1970). Ratios of Na/K are strongly influenced by the amount of available calcium in the water and cannot be relied upon as geothermometers in waters, such as those at Klamath Falls, for which molal ratios of $\sqrt{Ca}/Na$ are greater than one and indicated reservoir temperatures are low (Fournier and Truesdell, 1973, p. 1272).

Molal ratios of Ca/HCO$_3$ are low in the cold waters (range 0.04 to 0.2) and increase, as would be expected, in typical hot waters (range 0.5 to 1.2). The values of this ratio in the thermal waters are not high enough, however, to suggest that the waters equilibrated at extremely high reservoir temperatures (White, 1970, p. 69).

Ratios of total carbon species (mainly HCO$_3$ and CO$_3$) to chloride are low in the hot waters and these ratios clearly distinguish the hot waters from waters of lower

TABLE 6.—*Ratios of dissolved constituents in ground water*

[Molal ratios except where weight ratio is indicated]

Seq. No.	Owner or name	Temp. °C	$\sqrt{Ca/Na}$	Na/K	Ca/Mg	Ca/HCO₃	Ca/Cl	Na/Cl	B/Cl	F/Cl	SO₄/Cl	(HCO₃+CO₃)/Cl	SiO₂/Cl (wgt. ratio)
	Hot waters												
46	Liskey	93	2.2	85	> 91	0.5	0.2	5.2	0.04	0.05	2.3	0.5	1.5
44	Osborn	90	2.3	78	> 91	.6	.2	5.2	.05	.05	1.8	.5	1.6
13	Mills School	89	1.1	209	> 82	.9	.4	11	.06	.04	3.1	.5	1.4
6	OIT #6	88	1.5	85	>147	.8	.4	5.2	.06	.05	2.5	.4	1.6
28	Olene Gap Spg	87	2.7	44	121	1.2	.6	4.9	.06	.04	2.5	.5	1.7
58	Assembly of God	87	2.9	79	170	.7	.5	6.1	.05	------	2.7	.7	2.0
7	Medo-Bell Dairy	81	1.8	87	>140	1.1	.4	6.0	.06	.04	2.8	.4	1.5
8	Friesen	73	1.8	85	>152	1.2	.4	6.0	.05	.05	2.8	.4	1.6
14	Serruys	71	1.8	92	>133	.8	.4	6.3	.05	.05	2.8	.5	1.7
29	Jones (Spg)	65	2.2	73	212	1.1	.5	5.1	.1	.04	2.5	.4	1.5
17	Mazama School	61	.8	70	3.3	.07	.1	1.0	------	.05	2.6	2.0	3.4
	Warm waters of the Klamath Hills												
50	O'Connor	38	2.4	71	7.6	.1	.1	4.0	.01	.01	1.0	1.3	1.3
56	O'Connor Livestock	30	4.2	51	5.5	.4	.3	2.0	.01	.0	.5	.9	.5
45	Boehm	25	3.2	46	2.3	.2	.9	4.3	.03	.002	.6	5.0	.6
51	Liskey	22	1.2	82	1.3	.03	.2	6.3	.04	.005	.03	6.4	.7
57	Carter	22	6.2	20	7.9	.5	.3	1.5	.01	.0	.4	.6	.4
53	Suburban Water Co.	21	7.8	22	1.8	.4	.9	2.2	.01	.003	.5	2.3	1.0
54	Town of Merrill	21	6.2	20	1.6	.2	.4	2.5	.001	.007	.3	2.5	1.4
	Miscellaneous warm waters												
22	Falcon Hts. School	37	5.6	14	1.0	.07	.7	9.9	.04	.04	1.0	11	9.0
1	Eagle Point Spg	35	1.0	19	>3.6	.01	.03	6.0	.03	.09	<.04	4.9	4.5
20	Air Force	31	3.4	8.8	1.9	.05	.6	12	------	.09	.5	12	8.8
23	Jacobsen	30	6.0	7.0	1.5	.06	.6	9.6	.02	.02	.6	12	8.2
24	Oregon Water Corp.	26	9.2	------	.9	------	.3	1.9	------	.2	.08	------	2.1
39	M. Dehlinger	26	21	5.6	.9	.2	5.0	9.4	.02	.0	.7	26	15
15	Holliday Spg	25	15	16	2.4	.2	2.0	8.2	.03	.04	.5	14	5.3
25	McCollum	25	28	8.5	1.1	.3	10	10	.02	.1	6.2	23	17
36	Brookshire	25	20	11	3.9	.2	5.0	12	.01	.07	1.3	20	13
38	Stacy Co.	25	21	8.5	3.9	.2	5.0	15	.02	.0	2.1	27	17
26	Shuck	24	14	6.8	1.5	.1	3.3	14	.02	.1	1.9	23	20
43	O'Connor	24	13	6.8	1.1	.1	1.1	5.4	------	------	.6	8.3	4.6
47	C. Dehlinger	24	8.5	11	1.6	.08	1.4	13	.03	.0	.2	18	15
30	Langley	23	16	6.8	1.0	.1	3.3	13	.03	------	1.9	25	19
40	O'Neil	23	11	8.7	.7	.08	2.0	14	.04	.1	1.7	23	18
12	Oregon Water Corp.	22	15	------	1.1	------	2.0	7.6	------	.04	.04	------	4.7
18	Weyerhaeuser #4	22	7.8	16	3.2	.1	5.0	12	.08	.05	.4	17	8.0
27	Bixler	22	25	8.2	.8	.2	5.0	13	.03	.2	1.5	34	26
55	Pope's Meat Co.	22	11	16	3.7	.1	1.7	10	.01	.0	.5	13	5.7
10	Oregon Water Corp.	21	13	------	1.5	.1	1.1	------	------	------	0	8.5	5.8
11	Oregon Water Corp.	21	15	------	1.0	------	2.0	------	------	.004	1.0	------	4.5
35	Dobry	21	8.5	12	3.2	.08	1.4	14	.03	------	1.5	17	19
48	Hill	20	18	5.1	.4	.1	3.3	6.9	.02	.0	.6	19	6.2
	Cold waters												
31	Kinney	19	23	6.0	1.0	.2	5.0	15	.06	.2	2.2	38	29
32	Kirkpatrick	18	16	7.8	.7	.1	3.3	11	.03	.2	2.5	24	22
34	Sharp (Spg)	18	22	8.2	1.2	.2	5.0	13	.05	.0	1.7	29	20
33	Gabriel	16	9.2	7.5	.4	.04	3.3	22	.08	.2	.4	76	32
41	Air Force	16	17	8.0	.8	.2	5.0	10	------	.07	.5	30	16
9	Oregon Water Corp.	15	20	------	.9	------	2.5	------	------	.004	.1	------	6.0
19	Shaw	15	7.1	22	.7	.08	2.5	15	.05	.8	.2	31	9.3
42	Howard	15	33	19	.8	.2	10	11	.01	.0	.4	51	13
37	Adair	14	26	8.5	.6	.1	5.0	7.2	.02	.5	.2	35	10
49	Tulana Farms (Spg)	14	43	8.2	1.2	.2	10	8.8	Indet. (small)	.1	<.7	35	48
2	Shell Rock Spg	12	39	6.6	1.3	.2	>10	>9.0	Indet. (small)	.3	Indet. (small)	29	>60
5	Hummingbird Spg	12	37	9.9	.8	.2	10	11	Indet. (small)	.3	<.7	56	42
52	Long	12	2.5	94	.5	.05	1.3	22	.2	.2	1.2	26	2.8
4	Barkley Spg	11	31	11	.8	.2	10	12	.2	.3	<.7	50	36
3	Neubert Spg	10	37	9.9	1.0	.2	10	12	Indet. (small)	.2	<.7	55	48

temperature. Cold waters have uniformly high ratios, and transitional warm waters have ratios that are generally similar to or slightly lower than those of the cold waters. Ratios of total carbon to chloride show no significant differences in 10 of the 11 hot waters despite temperature differences of almost 30°C among these waters, whereas in nonthermal and mixed waters in which temperature differences are also about 30°C, ratios of total carbon to chloride differ by an order of magnitude.

The nearly constant ratios of carbonate species to chloride in hot waters of greatly differing temperature suggest that the temperature differences in these waters are due to conductive heat loss rather than mixing with cooler waters. The apparent lack of mixing in the hot waters implies, therefore, that these waters rise through conduits that are largely isolated from the shallow ground-water reservoirs of the area. It seems clear that most of these conduits occur in the major fault zones of the principal thermal areas. The widespread occurrence of warm water in the lower Klamath Lake basin suggests, however, that significant amounts of thermal water also arise from undiscovered faults underlying the valley floor.

Ratios of SO_4/Cl are slightly higher in the hot waters than in the cold waters, suggesting that the deeply circulating thermal waters are enriched in sulfate, either through contact with gypsum deposits or by oxidation of hydrogen sulfide gas which is known to be present in the hotter waters of the area (J. W. Lund, oral commun., 1975; Peterson and Groh, 1967).

The increased SO_4/Cl ratios in the hot waters have been interpreted as suggesting that a steam reservoir underlies the Klamath Falls graben (Murray Gardner, written commun., 1975; Peterson and Groh, 1967). The interpretation is based on the possibility that hydrogen sulfide gas rising from a deep steam reservoir may be oxidized to the sulfate as it enters shallower aquifers thereby increasing the ratio of SO_4 to the nonvolatile chloride. If this hypothesis is correct, volatile constituents such as boron (B) and fluoride (F) might also be expected to be transported in the vapor from the underlying steam reservoir, thereby increasing the ratios of these elements to chloride in the thermal waters (Truesdell, 1976a). In the samples obtained for this study, however, the ratios of fluoride to chloride are smaller in the thermal waters than in the cold waters, and ratios of boron to chloride in the thermal waters are similar to those in the five cold water samples. Thus, although these ratios are somewhat unreliable owing to the small concentrations of the three constituents in the cold waters, they provide no support for the theory of an underlying steam reservoir.

Most of the warm transitional waters in the lower Klamath Lake valley have temperatures in the range 20° to 40°C. These waters are produced by wells, generally no more than 500 feet deep, that penetrate alluvial and lacustrine deposits and interbedded basaltic rocks of the Yonna Formation. The unconsolidated deposits include diatomaceous earths, mixed-layer clay-mineral assemblages, kaolinite, feldspathic and organic-rich silt, fine siliceous sand with large percentages of volcanic glass and mafic materials, and coarser-grained sand and gravel also derived from volcanic rocks. A few wells and three warm springs probably derive water almost entirely from volcanic rocks and breccias of basaltic composition. These diverse sources of warm water are represented by 25 samples which show, as expected, wide variations in chemical composition.

Log molal ratios of several major ions in cold, transitional, and hot waters are shown graphically in figure 6. The ratios selected are those that best illustrate the diversity of the transitional waters and the trends, or absence of trends, between the hot and cold waters. These ratios do not coincide with those selected for table 6. In each graph, samples are arranged from left to right in order of increasing chloride concentra-

tion. This also turns out to be the general order of increasing sodium concentration and increasing temperature. For reasons discussed below, warm waters from the Klamath Hills area are placed in a separate group.

If the logical assumption is made that higher temperatures in the transitional waters result from larger amounts of thermal waters mixing with local meteoric waters, it would be expected that ratios of sodium to chloride, which do not differ greatly between the cold and hot waters, would remain fairly constant as temperatures increase. However, other ratios, such as sodium to calcium, calcium to magnesium, and total carbonate to chloride, which differ greatly between cold and hot waters, should show trends toward their values in the hot waters. The data of figure 6 show these expected trends in the sequence of transitional waters and thus tend to support the assumption that, in general, the warm waters in the Klamath Falls area represent varying mixtures of thermal and meteoric waters. The trends in sodium/calcium, calcium/magnesium, and total carbonate/chloride ratios are fairly distinct, despite effects of local lithology which are the probable cause of significant departures from the trends in individual waters.

Concentrations of silica are extremely variable in the warm waters (table 5) and ratios of this constituent to other major ions (not shown in figure 6) have little correlation with temperature trends. A graph of log molal silica concentrations in ground water (figure 7A) demonstrates the variability of silica concentrations in cold and transitional waters in relation to the trend of increasing chloride concentration shown in figure 6H. It can be seen in figure 7A that the variability of silica in five of the coldest and most dilute waters is greater than the variability in 10 hot waters from the principal geothermal areas. The relation of silica concentration is shown explicitly in figure 7B, where two trends are indicated for Klamath Hills warm waters and the hot waters. The suggested trend, A, between cold and hot waters is not supported by any field evidence, however, and the silica concentrations in both groups appear to be largely independent of the chloride concentrations.

The relation of chloride concentration to temperature might be expected to define mixing trends in the warm waters if such trends exist. As shown in figure 7C, however, the trends are not well defined and the relation does not show a high degree of correlation. It is not surprising, therefore, that silica concentrations show an even poorer correlation with temperature (figure 7D) and fail to suggest any clear trends.

The wide scatter in the concentrations of the various ions and the consequent scatter in plots of ionic ratios indicate the difficulties that may be encountered in attempts to apply chemical geothermometry to low-
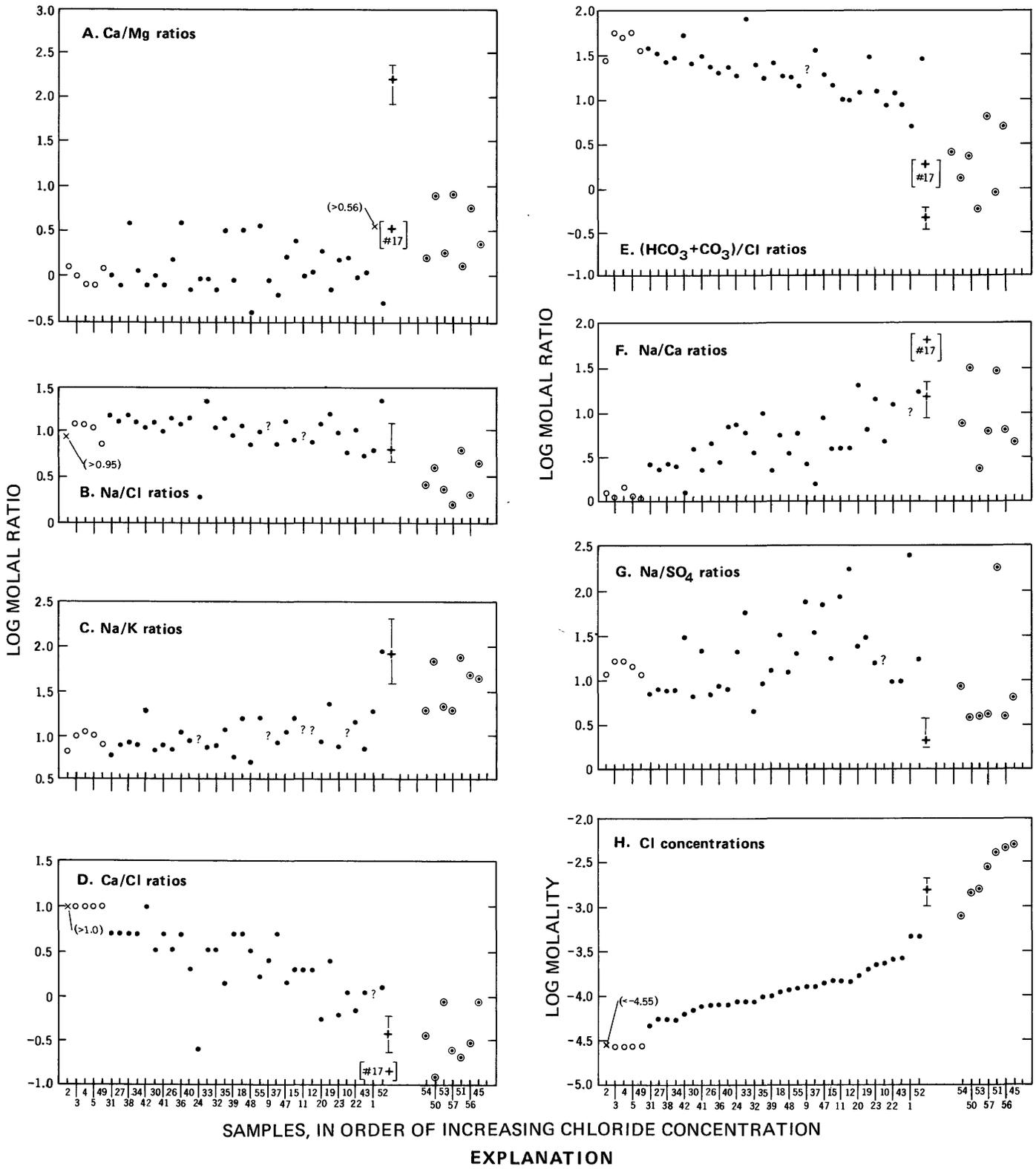
SAMPLES, IN ORDER OF INCREASING CHLORIDE CONCENTRATION

EXPLANATION

Median value and range of values for 11 hot waters (>60°C).
When sample 17 is shown separately, median and range are for
remaining 10 samples

Hot water sample 17; shown when ratio differs significantly
from other hot waters

⊚ Warm water from the Klamath Hills

• Warm transitional (mixed) water

○ Cold water (<15°C) derived from basaltic rocks

✕ Value greater or less than number in parentheses

FIGURE 6.—Log molal ratios of dissolved constituents in ground waters arranged in order of increasing chloride concentration.
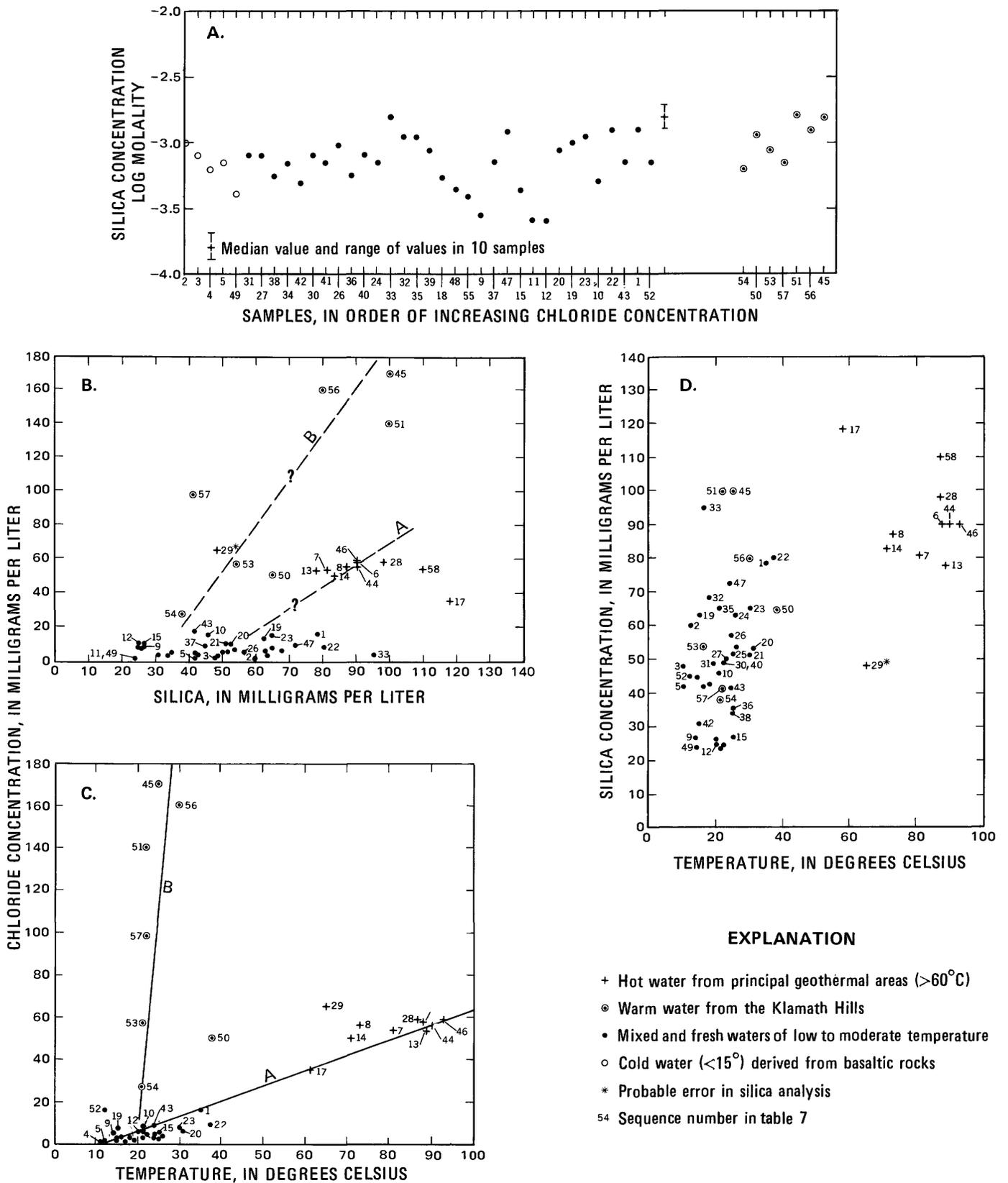
FIGURE 7.—Silica and chloride concentrations in ground water and their relations to observed water temperature. A. Log molal SiO₂ concentrations in samples arranged in order of increasing chloride concentration. B. Chloride versus silica concentration, showing separate trends for warm waters of the Klamath Hills (line B) and hot waters from the principal geothermal areas (line A). C. Chloride concentration versus temperature, showing two possible dilution trends (lines A and B). D. Silica concentration versus temperature.

temperature waters from volcanic terranes. (See the section on "Reservoir Temperatures Estimated from Chemical Geothermometers.") It is clear that both silica and cation geothermometers may be unreliable in these settings and should be applied to low-temperature waters only when most of the chemical relations indicate that the waters have cooled conductively or are simple mixtures of thermal and regional background waters.

A group of seven samples obtained from warm wells in the vicinity of the Klamath Hills thermal area are chemically anomalous with respect to other warm waters of the region. (See tables 5 and 6.) Despite the relatively low temperatures of these waters (maximum 38°C), concentrations of sodium, calcium, sulfate, chloride, and silica are among the highest encountered in the area, and electrical conductances are generally high, ranging from 640 to 2,700 μmho cm^{-1} in six of the seven samples.

The distinctive chemical character of the Klamath Hills warm waters is shown by the graphs of ionic ratios in figure 6 and by graphs of chloride concentration versus temperature (fig. 7C). The chloride concentrations in figure 7C form a trend for the Klamath Hills warm waters (line B) that differs greatly from the trend inferred for the hot waters (line A). Graphs of other ions, notably sodium and sulfate, versus temperature (not included in this report) also show distinctive trends similar to line B. These data are consistent with the isotope compositions (discussed in a subsequent section of this report) in suggesting a separate origin or history for warm waters of the Klamath Hills area.

The cause of the anomalous chemical characteristics in the Klamath Hills warm waters is uncertain, but the isotopic compositions strongly suggest that these characteristics are the result of low-temperature evaporation of the hot waters of the area. If so, the characteristics of these waters may be genetically related to their occurrence in the closed basin of Lower Klamath Lake.

RESERVOIR TEMPERATURES ESTIMATED FROM CHEMICAL GEOTHERMOMETERS

Because the solubility of silica in water of less than pH 9 is affected only by the temperature of the water, silica concentrations in thermal waters have been used as a basis for estimating reservoir temperatures (Fournier and Row, 1966). The estimates of reservoir temperature are based on the following assumptions: (1) Silica is freely available in the reservoir rocks; (2) the water has remained in contact with the rocks for a sufficient length of time to reach chemical equilibrium with respect to quartz or other silica minerals; (3) the silica concentration attained at depth remains constant as the water moves from the reservoir to the sampling point.

For waters of the Klamath Falls area, assumption (1) is probably true. Although virtually no quartz occurs in the upper Tertiary rocks of the region, silicate minerals are abundant in the basaltic rocks and deeply circulating thermal waters may encounter rhyolites of Eocene or Oligocene age, or quartz-rich sediments of Eocene age, both of which underlie younger Tertiary rocks in adjacent ares (Williams, 1949; Peterson and McIntyre, 1970). The high silica concentrations in the cold spring waters indicate that silica in some form is abundantly available even in the shallow aquifers.

In spite of a general availability of silica, it is possible that the geothermal system at the present time is largely restricted to flow paths and conduits that have been depleted in silica. There is no evidence, for example, for recent silicification comparable to the older silicification that pervades the rocks surrounding the geothermal areas, and this may indicate that silica concentrations, as well as flow volumes, have decreased during the life of the system. Although this question has not been resolved by the present study, the uniform silica concentrations of hot waters occurring over a wide area suggest that silica saturation has been generally attained by the thermal waters. In light of this conclusion, the second assumption relating to the time required for equilibration with quartz or other silica minerals is probably also a reasonable one for the Klamath Falls geothermal system. The assumption is further supported by results of tritium analyses, discussed below, which indicate that the thermal waters have had residence times on the order of 10 years.

The third assumption, specifying that silica concentrations remain constant as the thermal waters ascend to the surface, is probably not completely met. The relatively constant silica concentrations of the hotter waters of the region suggest, however, that the amount of wallrock interaction and dilution by mixing is minimal. Supporting this conclusion are the constant (HCO_3 + CO_3)/Cl ratios, the high Na/K and Ca/HCO$_3$ ratios, and the fairly small increases in oxygen isotope concentrations (see next section of this report), all of which are indicators of relatively low reservoir temperatures. As the geothermal waters ascend toward the surface, therefore, silica concentrations would not be expected to change appreciably through either precipitation or further solution of silica. Estimates of the probable extent of mixing in the thermal waters are given in the next section.

Several silica geothermometers are given by Fournier and Rowe (1966) and Fournier and Truesdell (1974). Briefly, the four models used for this study are based in the following assumptions: (1) Quartz saturation with adiabatic cooling (Qtz. Adiab.), (2) quartz saturation with conductive cooling (Qtz. Cond.), (3) chalcedony saturation with conductive cooling, and (4) quartz saturation and mixing with cold, meteoric wa-

ter. Computations were made by means of Geotherm, a computer program written by Truesdell (1976b). Results are listed in table 7 and shown in the graph of figure 8.

Input data for the mixing model include a silica concentration of 45 mg L^{-1} and a temperature of 20°C for the cold water. The temperature chosen for the cold-water component is typical of cooler waters in the vicinity of the geothermal areas. Slightly lower estimated reservoir temperatures would be obtained if the coldest ground-water temperatures in the region were used for the analysis. The silica concentration chosen for the cold-water component is tyical of concentrations in basaltic aquifers of the area. Somewhat higher estimated reservoir temperatures would be obtained if lower silica concentrations were used.

Estimated reservoir temperatures based on the quartz geothermometer have a narrow range for most of the hot waters but a much broader range for the warm waters. Reservoir temperatures calculated for nine of the hotter waters by means of the quartz conductive model range from 124° to 143°C; those calculated by means of the chalcedony model range from 94°

to 108°C. When the silica mixing model is applied, temperatures range from 126° to 184°C for chalcedony.

Athough nearly all the estimated temperatures fall within reasonable limits, they represent an uncomfortably large range of values for each of the approaches. In evaluating the various approaches, an important criterion is the fact that for geothermal waters having temperatures less than about 150°C, silica equilibration may occur with respect to chalcedony rather than quartz (R. O. Fournier, oral commun., 1976). However, the reservoir temperatures estimated from chalcedony concentrations are lower than observed temperatures in thermal wells, indicating that these direct estimates are inadequate and suggesting that mixing models are probably required. Final conclusions regarding these approaches and the best estimate of reservoir temperature are deferred to the section "Isotopes in Ground Water."

Estimated reservoir temperatures for warm transitional waters in the area (table 7) are extremely erratic and are probably unreliable. Although it is virtually certain that these are mixed waters, mixing models fail to provide reasonable estimates of reservoir

TABLE 7.—*Estimates of reservoir temperature based on chemical relations in ground water*

Seq. No.	Owner or name	Well depth ft.	Measured[1] temp. °C	Date	Calculated reservoir temperature (°C)							
					Qtz.[2] (adiab.)	Qtz.[2] (cond.)	Chalced.[2] (cond.)	Na/K+β√Ca/Na[3] β=1/3	β=4/3	Mixing models[1] Quartz	Chalced.	δ18O in sulfate[5]
Hot wells and springs												
6	OIT Well #6	1805	88	08–05–72	127	131	103	108	71	150	112	185
7	Medo-Bel Dairy	765	81	01–24–55	122	126	96	109	75	150	109	185
8	J. E. Friesen	564	73	02–19–55	124	130	101	109	75	163	127	------
13	Mills School	800	89	08–07–75	121	124	94	83	67	135	99	179
14	Lois Serruys	528	71	12–22–54	122	127	98	106	73	126	122	------
28	Olene Gap Spg	------	74	03–09–74	130	136	108	130	80	142	127	196
44	O. H. Osborn	418	90	05–31–74	127	131	103	114	83	154	114	135
46	Jack Liskey	282	93	05–09–74	128	131	103	111	82	152	111	138
58	Assembly of God	363	87	07–17–74	137	143	117	111	74	184	144	------
Warm wells and springs												
1	Eagle Point Spring	------	35	04–06–75	116	124	94	190	170	>200	200	------
17	Mazama School	972	61	------	138	147	120	129	126	197	163	------
20	U.S. Air Force	500	31	10–17–72	99	105	73	209	129	163	123	------
22	Falcon Hts. School	1050	37	05–07–74	117	125	96	181	104	>200	>200	------
23	Alfred Jacobsen	155	30	05–20–74	108	115	84	221	131	>200	>200	------
50	Jack O'Connor	~700	38	05–09–74	108	115	84	116	83	196	163	------
56	O'Connor Livestock	770	30	06–06–74	117	125	96	123	72	>200	>200	------
Slightly warm waters												
15	Howard Holliday Spg.	------	25	05–22–74	72	76	42	157	56	------	------	------
18	Weyerhaeuser #4	545	22	10–16–73	79	83	49	159	65	------	------	------
25	Melvin McCollum	223	25	05–28–74	98	104	72	178	50	------	------	------
26	Claude Shuck	1060	24	05–30–74	101	108	77	203	84	------	------	------
27	Ray Bixler	460	22	05–29–74	95	101	69	182	55	------	------	------
30	Robert Langley	350	23	05–04–74	96	102	70	200	78	------	------	------
35	Len Dobry	612	21	05–04–74	107	115	84	180	88	------	------	------
36	Lester Brookshire	1210	25	05–04–74	83	87	53	173	57	------	------	------
38	Geo. Stacy Co.	916	25	06–05–74	81	86	52	183	61	------	------	------
39	Monte Dehlinger	148	26	06–06–74	99	106	74	207	73	------	------	------
40	J. K. O'Neil	83	23	05–04–74	96	102	70	194	88	------	------	------
43	Dan O'Connor	85	24	05–07–74	90	95	62	207	91	------	------	------
45	Abe Boehm	155	25	05–30–74	127	137	109	132	87	------	------	------
47	Clyde Dehlinger	100	24	06–05–74	112	120	90	189	95	------	------	------
48	Bill Hill	820	20	06–06–74	71	74	40	217	83	------	------	------
51	Jack Liskey	95	22	05–09–74	126	137	109	121	115	------	------	------
54	Town of Merrill	1088	21	02–18–55	86	90	57	155	71	------	------	------
55	Pope's Meat Co.	1128	22	06–06–74	69	73	39	161	67	------	------	------
57	George Carter	67	22	06–06–74	89	94	61	156	83	------	------	------

[1] For sources of measurement and chemical analyses, see table 5.
[2] Fournier, R. O., 1976.
[3] Fournier and Truesdell, 1973, 1974.
[4] Mixing models based on quartz or chalcedony equilibria and a cold-water component having a silica concentration of 45 mg L^{-1} and a temperature of 20°C (Truesdell and Fournier, 1977).
[5] Based on $\delta^{18}O(H_2O)/\delta^{18}O(SO_4)$ relations and assumed conductive heat loss (McKenzie and Truesdell, 1977).
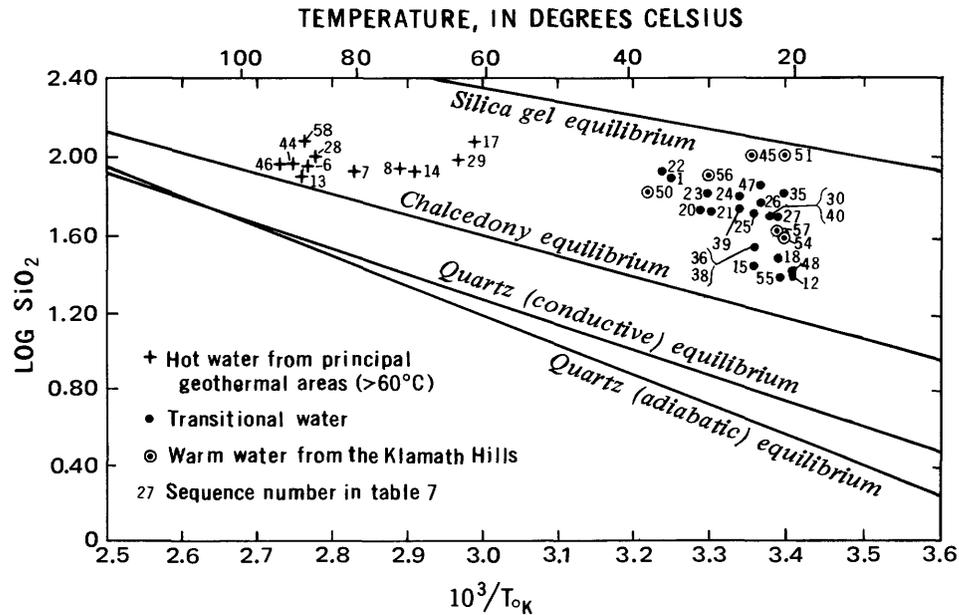
FIGURE 8.—Temperature of ground water versus log silica concentration. Straight lines show empirical relations of temperature to equilibrium concentrations of quartz, chalcedony, and silica gel (Fournier, 1977).

temperatures. The most likely causes for the failure of the mixing models are the general availability of amorphous silica and the general presence of nonequilibrium chemical conditions in the deposits permeated by the warm waters.

A second quantitative geothermometer that has been applied successfully to geothermal systems is the Na-K-Ca geothermometer (Fournier and Truesdell, 1973). Based on worldwide data from springs and wells, the following empirical expression has been formulated:

$$\log (Na/K) + \beta \log (\sqrt{Ca}/Na)$$
$$= \frac{1647}{273+T} - 2.24 = K^*$$

where concentrations of Na, K, and Ca are in moles, T is temperature in degrees Celsius, $\beta = 4/3$ for waters equilibrated at less than 100°C and 1/3 for waters equilibrated at more than 100°C.

Data from table 5 were used to obtain a value of K^* for each sample, and this value was then plotted versus the reciprocal of the measured temperature in degrees Kelvin. The results are shown in figure 9 along with a line representing the Fournier-Truesdell expression. Projection of the K^* values horizontally onto the line gives the apparent temperature of equilibration in the reservoir. Calculated equilibration temperatures based on the relations shown in figure 9 for two values of β are listed in table 7.

The graph shows that the majority of water samples fall into two groups with roughly similar estimated reservoir temperatures. The left-hand group comprises the hot waters of the area; projection of these points onto the line produces estimated reservoir temperatures that, in most cases, are lower than actual water temperatures. This result is caused by Na/K ratios that are higher than expected in relation to $\sqrt{Ca}/Na$ ratios and observed temperatures.

The group of data points on the right side of the graph in figure 9 represents warm, and presumably mixed, waters with wide ranges of concentrations of Na, Ca, and K. Estimated temperatures of last rock-water equilibrium for many of these samples are similar to estimates derived for the hot waters (table 7). Five samples give anomalously high estimates.

The two major groups of samples in figure 9 are distributed so that a line connecting their centers is nearly horizontal. Thus, values of K^* along this line would vary only slightly with large changes in temperature. It might be expected, therefore, that mixing models using typical hot and warm waters would change estimated reservoir temperatures only slightly from those estimated for the hot waters alone.

In order to test the effect of mixing, trials were made using mixtures of a typical hot water (No. 58: Na=213 mg L^{-1}; Ca=28 mg L^{-1}; K=4.6 mg L^{-1}) and averages for warm water between 15° and 25°C (Na=25 mg L^{-1}: Ca=12 mg L^{-1}; K=4 mg L^{-1}). With mixtures ranging
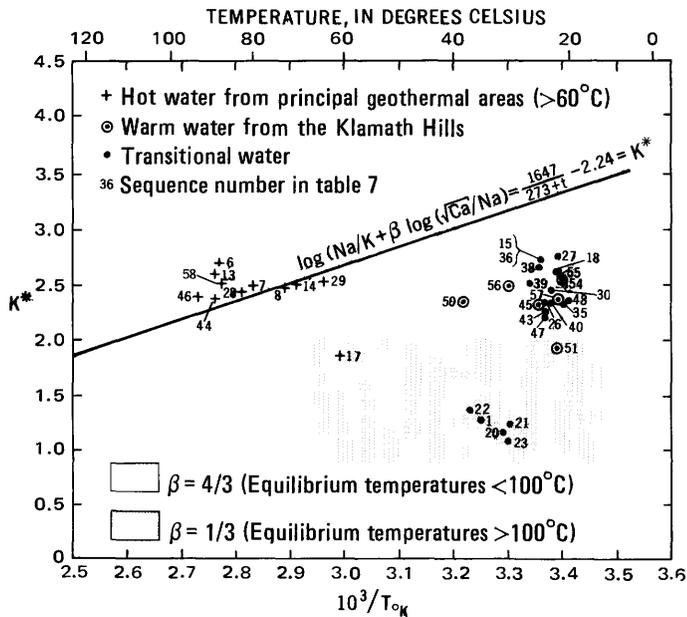
FIGURE 9.—Temperature of ground water versus log (Na/K) + βlog (√Ca/Na). Straight line shows empirical relation to equilibrium reservoir temperatures (Fournier and Truesdell, 1974).

TABLE 8.—*Analyses of the isotopes oxygen-18, deuterium, and tritium in ground water and Klamath Lake water*

Sequence No.	Name	$\delta^{18}O$[1] (0/00)	δDeuterium[1] (0/00)	$\delta^{18}O$ in sulfate (0/00)	Tritium (TU)
Hot waters (>60°C)					
6	OIT #6	-14.4^G	-118.7^{LW}	-5.45^N	1.6 ± 0.5^{TW}
7	Medo Bel Dairy	-14.6^G	-119.4^{LW}	-5.45^N	------
13	Mills School	-14.83^N	-121.8^{LW}	-4.94^N	------
17	Mazama School	-14.2^C	-117.0^C	------	------
28	Olene Gap Spring	-13.3^G	-111.6^C	-4.82^N	7.1 ± 0.6^{TW}
29	L. Jones spring	-13.5^C	-110.2^C	------	------
44	O. Osborn	-14.6^G	-117.0^{LW}	-1.85^N	------
46	J. Liskey	-14.6^G	-116.6^{LW}	-2.13^N	1.3 ± 0.5^{TW}
58	Assembly of God	-14.75^C	-120.0^C	------	------
Warm waters from the Klamath Hills (21°C–25°C)					
45	A. Boehm	-6.78^N	-73.2^O	$+15.42^N$	------
51	J. Liskey	-8.67^N	-87.2^O	$+8.69^N$	------
53	Suburban Water Co.	-10.79^N	-96.7^O	$+9.56^N$	------
Warm waters (20°C–37°C)					
1	Eagle Point Spring	-14.3^G	-109.7^O	------	------
18	Weyerhaeuser #4	-13.9^G	-109.1^O	------	------
22	Falcon Heights School	-14.9^G	-120.5^O	------	------
25	M. McCollum	-14.1^G	-113.2^O	------	------
Cold waters (<20°C)					
2	Shell Rock Spring	-15.35^C	-113.1^C	------	------
4	Barkley Spring	-15.11^N	-115.0^O	------	------
9	Oregon Water Corp.	-14.4^C	-109.4^C	------	2.4 ± 0.5^{TW}
33	Gabriel Brothers	-16.05^O	-123.6^O	------	------
41	U.S. Air Force	-14.52^O	-110.0^O	------	------
42	V. Howard	-14.36^O	-108.0^O	------	------
49	Tulana Farms Spring	-13.95^C	-105.8^C	------	------
--	Klamath Lake	-12.35^C	-97.5^C	------	25.7 ± 1.4^{TW}

[1]Referred to Standard Mean Ocean Water (Craig, 1961b)

Analysts
C   T. B. Coplen, Reston, Va.
G   Geochron Laboratories, Denver, Colo.
N   N. L. Nehring, Menlo Park, Calif.
O   James O'Neil, Menlo Park, Calif.
LW L. D. White, Menlo Park, Calif.
TW T. A. Wyerman, Reston, Va.

from 15 to 45 percent warm water, reservoir temperatures estimated by the Na-K-Ca geothermometer ranged only from 74° to 75°C.

Additional trials using an average composition for five cold waters (<5°C) in basaltic aquifers (Na=25 mg L⁻¹; Ca=12 mg L⁻¹; K=4 mg L⁻¹) produced only a slight improvement in estimated reservoir temperatures. For mixtures of 15 to 45 percent cold water, estimated reservoir temperatures ranged from 78° to 86°C.

Analysis of the terms incorporated in the geothermometer shows that molal ratios of Na/K decrease by an order of magnitude between the hot waters and the cold or warm waters, but that ratios of √Ca/Na have an offsetting increase of nearly an order of magnitude. When the cation concentrations of hot and warm or cold waters are combined in mixing models and logs of the resulting ratios are added in the geothermometer expression, the net effect of mixing is very small. For this reason, simple mixing models used with the cation geothermometer do not appear to be capable of providing reliable estimates of reservoir temperature for these waters.

ISOTOPES IN GROUND WATER

Samples of water from a number of wells and springs in the vicinity of Klamath Falls were analyzed for concentrations of isotopes of oxygen (^{18}O) and hydrogen (deuterium and tritium) (table 8). Analyses of these isotopes have been used elsewhere to ascertain the origin and the history of thermal waters. (See for example, International Atomic Energy Agency, 1974; Taylor, 1974.)

OXYGEN-18 AND DEUTERIUM ISOTOPES

The correlation between concentrations of oxygen-18 and deuterium in precipitation is generally given by the expression

$$\delta D = 8 \, \delta^{18}O + d$$

where δD and $\delta^{18}O$ are ratios of deuterium and ^{18}O to H and ^{16}O, respectively, expressed in parts per thousand (°/oo) as departures from the ratios in Standard Mean Ocean water (SMOW) (Craig, 1961a, b), and d is typically 10. The "meteoric line" resulting from the above expression is shown in figure 10 along with data from 25 water samples from the Klamath Falls area.

The cold, fresh waters of the area, represented by samples 4, 9, 41, 42, and 49 in figure 10 appear to establish a trend parallel to but offset from the theoret-

ical meteoric line. Although one cold-water sample, number 2 (Shell Rock Spring), is anomalous, the trend established by the five remaining samples may be taken as a probable baseline for most locally derived recharge waters. A seventh cold-water sample, number 33, derived from a 1,004-foot well located near Olene Gap, also appears to be anomalous but may represent high-altitude recharge from Stukel or Hogback mountain.

Transitional waters are represented by only four samples, two of which seem to have anomalous $\delta^{18}O/\delta D$ ratios. The trend of these samples from left to right parallel to the meteoric line is in keeping with a general trend of decreasing temperature in the four samples and also with their relative positions in the trends of ionic ratios in figure 6. Uncertainties resulting from possible errors of analyses and the small number of samples would seem to preclude attempts to draw general conclusions from the isotope concentrations in these transitional waters.

Three warm waters from the Klamath Hills, samples 45, 51, and 53, show large displacements from the meteoric trend in figure 10 and demonstrate further that these waters differ significantly from both hot waters and other warm waters in the area. (See section on "Relations Among Dissolved Constituents," this report.) The positions of these samples are consistent with the supposition that the Klamath Hills warm waters are mixtures whose components have been subjected to low-temperature evaporation.

Nine samples of hot water from the three principal geothermal areas are included in the graph of figure 10. These waters have apparently been enriched in ^{18}O during their passage through the geothermal reservoir, with increases in $\delta^{18}O$ ranging from 0.8 to 1.4 ‰. The increase is small compared to that observed in high-temperature geothermal systems such as Steamboat Springs, Nev. or The Geysers, California, which have increases in $\delta^{18}O$ as great as 3 to 5 ‰ (White, 1968). The isotope data suggest, therefore, that the Klamath Falls hot waters have not equilibrated at extremely high reservoir temperatures, or, alternatively, that they have been diluted by mixing with cooler waters. These alternatives have been discussed in the preceding section of this report and will be discussed further in the next section.

The hotter thermal waters from both the Klamath Falls hot-well area (samples 6, 7, 13, and 58) and from the Klamath Hills (samples 44 and 46) have a fairly small range of $\delta^{18}O$ values. They differ more widely, however, in their deuterium concentrations and in their displacement from the presumed meteoric line. Assuming these differences to be significant, they indicate that two groups of hot waters originate in

slightly differing meteoric waters and are subjected to slightly differing reservoir conditions. Thus the isotope data are consistent with the chemical analyses and temperature data for these waters, which together suggest that although the reservoirs and sources of recharge are generally similar for the two groups, small variations occur as the result of differing local conditions.

Two samples of hot water from the geothermal area at Olene Gap (samples 28 and 29) differ significantly from other hot waters, principally in having higher concentrations of calcium, ^{18}O, and deuterium. One explanation for the high concentrations of ^{18}O and deuterium in these samples is that the altitude at which recharge occurs is lower than that of other recharge waters. An alternative explanation for the composition of samples 28 and 29 is that either the recharge water or the thermal water has undergone low-temperature evaporation. Although neither of these explanations is clearly supported or excluded by the available data, both would imply that the geothermal reservoir at Olene Gap is separated to some degree from other geothermal reservoirs in the area.

Some of the relations suggested by deuterium and ^{18}O ratios in figure 10 may be substantiated and clarified by plotting both deuterium and $\delta^{18}O$ versus a conservative ion such as chloride (figure 11). The apparent mixing trend of Klamath Hills warm waters is clearly shown in figure 11, as are groupings of thermal waters on differing trend lines. Samples 28 and 29 (Olene Gap springs) are separated from other thermal waters, and there is the suggestion that sample 17, a
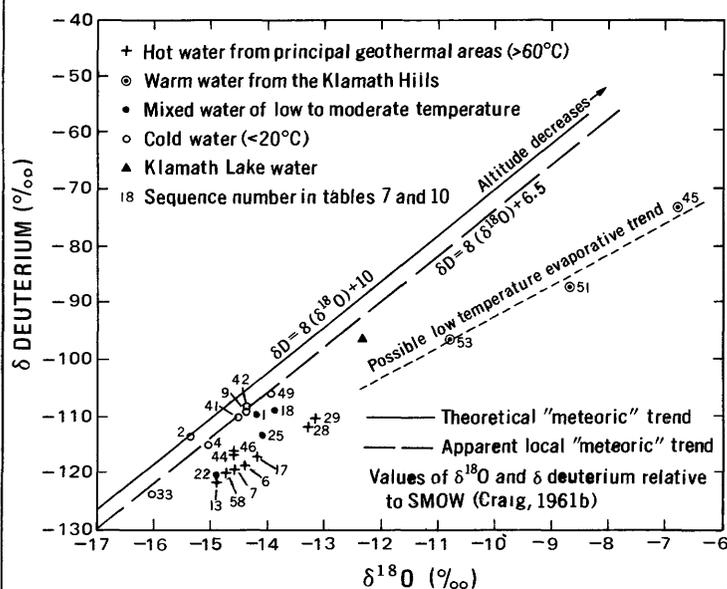
FIGURE 10.—δoxygen-18 versus δdeuterium in ground water and lake water.
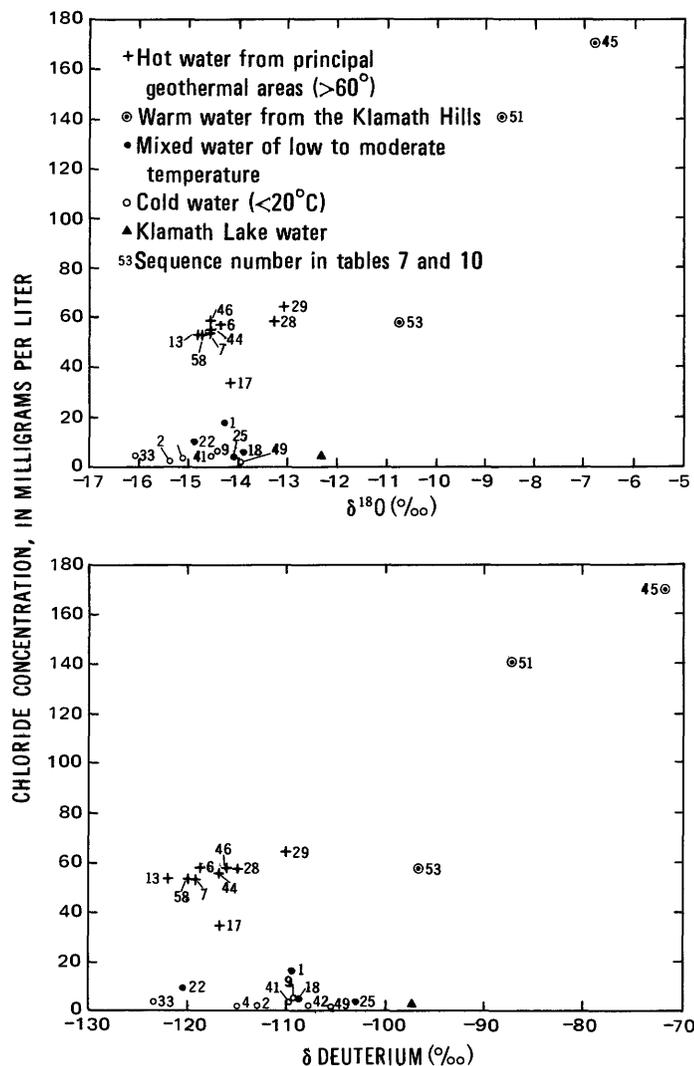
FIGURE 11—Chloride concentration versus δoxygen-18 and δdeuterium in ground water and lake water.

water having somewhat anomalous chemical characteristics, falls on a possible mixing trend with the Olene Gap springs.

TRITIUM CONCENTRATIONS

Four samples of ground water and one sample of Klamath Lake water were analyzed for tritium concentrations (table 8). These analyses represent only one point in time and hence do not provide a basis for reliable estimates of residence time in the aquifer. However, some qualitative conclusions are possible.

Samples 6 and 46, from OIT 6 and the Liskey hot well, respectively, are clearly older waters, possibly derived from pre-1954 meteoric water. Sample 9, from a cold, deep well used for municipal supply, is also an older water which may predate the 1954 atomic-bomb input. Sample 28 (Olene Gap hot spring), has a

significant amount of tritium, indicating the presence of post-1954 meteoric water.

On the basis of tritium concentrations in thermal and nonthermal waters, a mixing model can be constructed from which the apparent reservoir temperature for the Olene Gap springs can be estimated. Utilizing the tritium concentration in sample 28 (7.1 TU), and assuming that the tritium concentration in the geothermal reservoir is zero, and that of the cold-water component is the same as Klamath Lake (26 TU), the mixing proportions are determined and applied to observed temperatures. The calculated reservoir temperature is 112°C. When observed silica concentrations and geothermometers are used in this tritium mixing model, estimated reservoir temperatures are 121° and 146°C, respectively, for the chalcedony and quartz conductive geothermometers.

Although these results are reasonable, no great weight should be attached to them in view of the large uncertainties involved. In particular, the choice of a tritium concentration for the cold-water component is arbitrary and appears to be lower than would be expected for meteoric water in this region (T. A. Wyerman, oral commun., 1978). It is clear, however, from the form of the mixing expression, that any increase in the assumed tritium concentration of the cold water would result in even smaller calculated fractions of cold water and consequently in lower apparent reservoir temperatures. Conversely, trials with assumed tritium concentrations in the cold water smaller than about 15 TU result in calculated reservoir temperatures that are much higher than those estimated by other means and are probably unreasonable in the light of the chemical nature of the water. If other assumptions of the mixing model are reasonable, therefore, it is probable that the tritium concentration in the cold-water component of the Olene Gap springs is in the range 15 to 30 TU, and that the calculation correctly indicates the presence of a significant fraction of fairly young water in the mixture. A cold-water fraction as old as that in sample 9, a cold deep water derived from Tertiary basaltic rocks, is apparently ruled out by the mixing model relations.

ISOTOPE MIXING MODELS

Isotope concentrations in thermal waters can be used to gain insight into several questions that are of paramount importance in assessing the energy content of a geothermal system. These questions are (1) Do the waters rise from a single extensive reservoir? (2) to what extent are the thermal waters mixtures of shallow cold water and hotter water from the geothermal reservoirs? and (3) what is the temperature of the re-

servoir? Unequivocal answers to these questions prob-
ably cannot be obtained from the data at hand, but the
graphical analysis described below may point toward
probable answers for the Klamath Falls geothermal
system.

Figure 12 is a large-scale version of a part of the
graph of figure 10 showing δoxygen-18 versus
δdeuterium. In figure 12, simple mixing models for the
geothermal waters have been tested by trial and error.
The procedure is based on the following assumptions:
(1) Recharge waters for the geothermal reservoir have
positions along the observed cold-water trend line; and
(2) thermal waters are simple mixtures of these cold
waters with a single hypothetical reservoir water hav-
ing isotope compositions specified by arbitrarily chosen
points, P_1 and P_2, for example.

Hot-water and cold-water fractions of each thermal
water can be calculated from the geometric proportions
of straight lines connecting hypothetical cold waters,
thermal water, and the hypothetical reservoir waters,
as is shown for sample 29 and reservoir water P_2. Based
on an assumed cold-water temperature of 20°C, appar-
ent temperatures of the reservoir water are calculated
by means of the expression,

$$T_m + f_r T_r + f_c T_c \qquad (1)$$

where T_m is observed temperature in the mixed water,
T_r and T_c are temperatures in the reservoir and cold
water, respectively, and f_r and f_c are fractions of reser-
voir water and cold water.
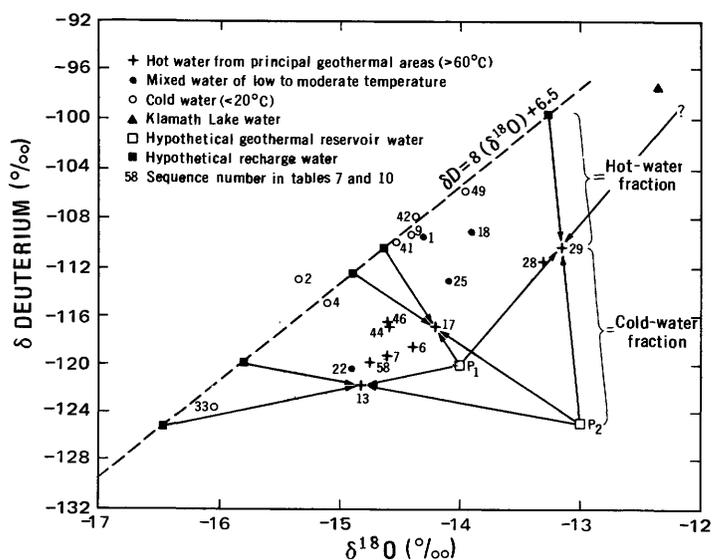
Solving equation (1) for T_r, the apparent tempera-
ture of reservoir water is determined for each observed
thermal water (table 9). For a reservoir water in posi-
tion P_2, representing an enrichment in $\delta^{18}O$ of about
3½ °/oo, apparent reservoir temperatures range from
130 to 324°C. This wide range indicates that the simple
mixing model fails to account for the observed distribu-
tion of thermal waters on the graph in relation to P_2.
As a further test of the model, chloride concentrations
in the hypothetical reservoir water may be calculated
in the same manner as the temperatures, on the as-
sumption that chloride concentration is a conservative
property that is likely to be relatively unaffected by
factors other than mixing. The resulting differences in
calculated reservoir chloride concentrations (table 9)
confirm the conclusion from the temperature calcula-
tions that simple mixing with a reservoir water at or
near the position of P_2 is not an acceptable model. No
further calculations were made using this model.

More acceptable results were obtained in a second
trial that applied the above procedures to a hypotheti-
cal reservoir water at position P_1 (figure 12). As was
true for reservoir water P_2, however, the range of calcu-
lated chloride concentrations shows that a single re-
servoir fluid at or near P_1 cannot account for chloride
concentrations in all thermal waters. Furthermore, on
the basis of additional trials (not indicated in figure
12), it is concluded that no single point of origin for a
reservoir water could provide consistent results under
the assumed conditions of the model.

TABLE 9.—*Calculated silica and chloride concentrations and tem-
peratures in the geothermal reservoir*

[Based on oxygen-18 and deuterium mixing models derived from figure 12. See text for
explanation]

	Calculated silica concen- tration[1] (mg L⁻¹)	Calculated chloride concen- tration[2] (mg L⁻¹)	Calculated reservoir temperature		
			Isotope mixing model[3] (°C)	Chalcedony geother- mometer[4] (°C)	Quartz geother- mometer[4] (°C)
Hypothetical Geothermal Water, P_1					
Sample No. 29	136	114	100	131	155
17	114	50	80	119	145
(44, 46)	147	128	186	136	160
6	111	81	120	117	143
7	102	84	116	112	139
58	155	89	130	140	163
13	95	80	126	108	135
Hypothetical Geothermal Water, P_2					
Sample No. 29	169	157	130	____	____
17	179	94	137	____	____
(44, 46)	232	239	324	____	____
6	187	150	209	____	____
7	154	161	204	____	____
58	262	173	237	____	____
13	139	149	220	____	____

FIGURE 12.—Mixing proportions of thermal and meteoric waters
based on δoxygen-18 and δdeuterium relations and hypothetical
geothermal reservoir waters.

[1]From isotope mixing model; assumed silica concentration in cold water is 45 mg L⁻¹.
[2]From isotope mixing model; assumed chloride concentration in cold water is 3 mg L⁻¹.
[3]From graph, figure 12; assumed temperature of cold water is 20°C.
[4]Based on calculated silica concentrations in first column.

In this regard, it may also be noted in figure 12 that as the trial points chosen for hypothetical reservoir waters approach the meteoric trend line so as better to satisfy the criteria of consistent and reasonable reservoir temperatures, required position of recharge waters become more dispersed along the meteoric trend line. For a hypothetical reservoir water near point P_1, for example, recharge points are required to range from the vicinity of the Klamath Lake sample at one extreme to points beyond sample 33 at the lower left representing exceptionally high altitude or dilution.

On the basis of the above graphical procedure, therefore, it would seem most probable, although by no means certain, that two or more reservoir waters of slightly differing isotopic compositions exist. One of these waters may be similar in composition to P_1. A second reservoir water with significantly different ^{18}O and deuterium concentrations and of differing meteoric origin could then account for the observed compositions of samples 28 and 29, the Olene Gap springs. It is further speculated that Klamath Hills hot waters (samples 44 and 46) and a fairly low-temperature thermal water such as sample 17 (Mazama School) would have to be accounted for by more complex processes involving mixtures of previously warmed waters or conductive cooling.

The data in table 9 show that an isotope mixing model based on hypothetical reservoir water P_1 produces consistent estimated chloride concentrations and a fairly small range of reservoir temperatures (116° to 130°C) for Klamath Falls hot-water samples 6, 7, 13, and 58. Although these temperatures are similar to those in table 7 calculated directly from a mixing model based on the chalcedony equilibrium curve (range 99° to 127°C), they are generally lower than the highest temperatures observed in the hot waters.

As a further test of the isotope mixing model, silica concentrations calculated from this model for hypothetical reservoir water P_1 may be used with the silica geothermometer (Fournier and Rowe, 1966) to estimate reservoir temperatures. Results of these calculations, given in table 9 for both chalcedony and quartz geothermometers, compare well with the silica mixing-model results in table 7. These results appear to confirm the isotope mixing model, and they suggest that a reservoir water with an isotopic composition in the vicinity of point P_1 in figure 12 could be the source of the Klamath Falls hot waters.

It is apparent that only the mixing models based on quartz equilibria produce estimated minimum reservoir temperatures that are generally higher than observed water temperatures. For the four Klamath Falls samples that appear to be most reliable (6, 7, 13, and 58), estimated reservoir temperatures range from 135°

to 184°C for the quartz mixing model (table 7) and from 135° to 163°C for quartz equilibrium in the isotope mixing model (table 9). These data are not sufficiently numerous to justify a statistically derived best estimate, but it appears reasonable to suppose that reservoir temperatures in all three geothermal areas have a minimum value in the range 145° to 155°C. In the light of all chemical and isotope data, a best guess of 150°C is used in the final section of this report for calculations of heat content in the reservoir.

RESERVOIR TEMPERATURES ESTIMATED FROM OXYGEN ISOTOPES IN DISSOLVED SULFATE

The concentrations of oxygen-18 in dissolved sulfate can be used as a geothermometer in thermal waters that have equilibrated for long periods of time in a geothermal reservoir (McKenzie and Truesdell, 1977). It is assumed that exchange between the oxygen isotope in the sulfate ion, $^{18}O(SO_4)$, and the oxygen isotope in water, $^{18}O(H_2O)$, will have attained equilibrium after residence times that range from 2 years at a temperature of 300°C to 18 years at 200°C (McKenzie and Truesdell, 1977, p. 53).

Concentrations of $^{18}O(SO_4)$ were determined for six hot waters from the Klamath Falls geothermal areas (table 8). As residence times for these waters probably exceed 18 years (see previous section, "Tritium Concentrations"), equilibration between $^{18}O(SO_4)$ and $^{18}O(H_2O)$ may be assumed to be virtually complete if reservoir temperatures are not far below 200°C. Based on this assumption, calculated reservoir temperatures for the six hot waters range from 135° to 196°C, assuming only conductive heat loss in the rising thermal waters, and from 130° to 179°C, assuming single-stage steam loss at the surface temperatures of the hot waters. The two lowest values in each case are for the Liskey and Osborn hot wells in the Klamath Hills. Values for the Olene Gap and Klamath Falls samples range from 179° to 196°C for the conductive model.

These estimates of reservoir temperature are clearly at variance with the estimates derived from concentrations of silica and water-oxygen isotopes (table 7) and explanations for the discrepancy must be sought. Two types of explanations can probably be discarded immediately. The bulk of the chemical and isotope data favors the concepts of low reservoir temperatures and largely undiluted thermal waters. Any explanation for the high sulfate-oxygen temperatures involving dilution of the thermal waters, therefore, would not only be opposed by this evidence but would also tend to increase the apparent reservoir temperatures and widen the gap between the estimates. Furthermore, the concept of mixing and dilution by nonthermal waters is

also opposed by the fact that $^{18}O(SO_4)$ concentrations are independent of chloride concentrations in the thermal waters. Explanations involving the solution of gypsum from shallow evaporite deposits may also be ruled out, inasmuch as the sulfate from such deposits would have high oxygen-isotope concentrations that would result in lower apparent reservoir temperatures.

One possible explanation for the high reservoir temperatures estimated from the sulfate-oxygen isotope could involve the formation of isotopically light SO_4 in volcanic rocks during their emplacement at high temperatures. This SO_4, taken into solution by geothermal waters circulating in faults of much later origin, would react with the lower-temperature thermal water in a process that possibly has not yet reached equilibrium at the low indicated temperatures. Although this explanation could account for the shift toward the observed isotope concentrations, it is somewhat unsatisfactory in that it requires the presence of high-temperature sulfate-bearing minerals which are not prevalent in volcanic rocks.

A second speculative explanation is that the sulfate-oxygen isotope has equilibrated in a deep reservoir that underlies the shallower zone tapped by thermal wells and springs. Hot water, presumably at or above the temperatures indicated by the sulfate-oxygen geothermometer, rises into the shallow, cooler reservoir where most constituents, including silica, equilibrate at the lower temperatures. The sulfate-oxygen isotope fails to attain equilibrium with the water-oxygen isotope in this reservoir as the waters rise more quickly along faults and fractures in the shallow geothermal system. Calculations based on expressions given by McKenzie and Truesdell (1977) show that complete equilibration at 150°C would require more than 77 years.

A completely satisfactory explanation of the discrepancy between the estimated reservoir temperatures is not available, and, therefore, no final conclusion has been reached in this study. The phenomenon described here is similar, however, to others observed in low-temperature thermal waters in the High Cascades and in the Gila River area of Arizona where high sulfate-oxygen isotope temperatures are also indicated (R. H. Mariner, oral commun., 1979). These discrepancies will apparently require further study of the chemical relations in thermal waters derived from thick volcanic sequences such as those at Klamath Falls and the areas mentioned above.

Three additional waters analyzed for $^{18}O(SO_4)$ (table 8) are warm waters from the Klamath Hills. Results of these analyses, which show high concentrations of $^{18}O(SO_4)$, further confirm the belief that the waters are mixtures containing evaporation residues. Estimates of reservoir temperature have not been made for these waters.

TEMPERATURE DISTRIBUTION AND HEAT FLOW

AREAL DISTRIBUTION OF TEMPERATURE

The distribution of temperature measured in ground water of the Klamath Falls area is shown on plate 3. The isotherms of the map are generalized from data that have considerable scatter in many areas. Much of the scatter is due to differences in temperature between deep wells and shallow wells in the same area, the deeper wells having generally higher temperatures. The isotherms, therefore, are at best a rough guide to areas where wells of moderate depth (generally 150 to 500 feet) may be expected to encounter water with temperatures similar to those indicated by the contours.

The map shows that although warm water occurs at a large number of isolated places, it is found principally within the three closed 60°C contour lines that contain the Klamath Falls urban area, Olene Gap, and the southwest flank of the Klamath Hills.

VERTICAL DISTRIBUTION OF TEMPERATURE

Temperature profiles measured in 41 unused, non-flowing wells and two test holes are shown on plate 4. The locations of these holes are shown in figure 13, and data pertaining to the profiles are given in table 10 where the profiles have been divided into three groups on the basis of their dominant characteristics: Mainly Conductive Profiles, Cold Convective Profiles, and Irregular Profiles.

Most of the wells in the first group, Mainly Conductive Profiles, penetrate fine-grained lacustrine sediments that range from silt and clay of Holocene age to shale and diatomite of the Pliocene Yonna Formation. Only three of the holes are known to penetrate basaltic rocks. The upper parts of the individual profiles are difficult to trace on plate 4, but nearly all show evidence of downward convective transport in the upper tens of feet. In addition, many have temperature irregularities at depths that, in some instances, correspond to known permeable zones in the rocks.

Thermal gradients in the conductive profiles are generally high, ranging from 50° to 180°C per kilometer (°C km^{-1}), even though temperatures are relatively low. The highest temperature observed is 38°C at 500 feet in CW 3, located at 39/9–3ddd, just south of Altamont.

The second category in table 10, Cold Convective Profiles, includes 10 wells characterized by low tem-
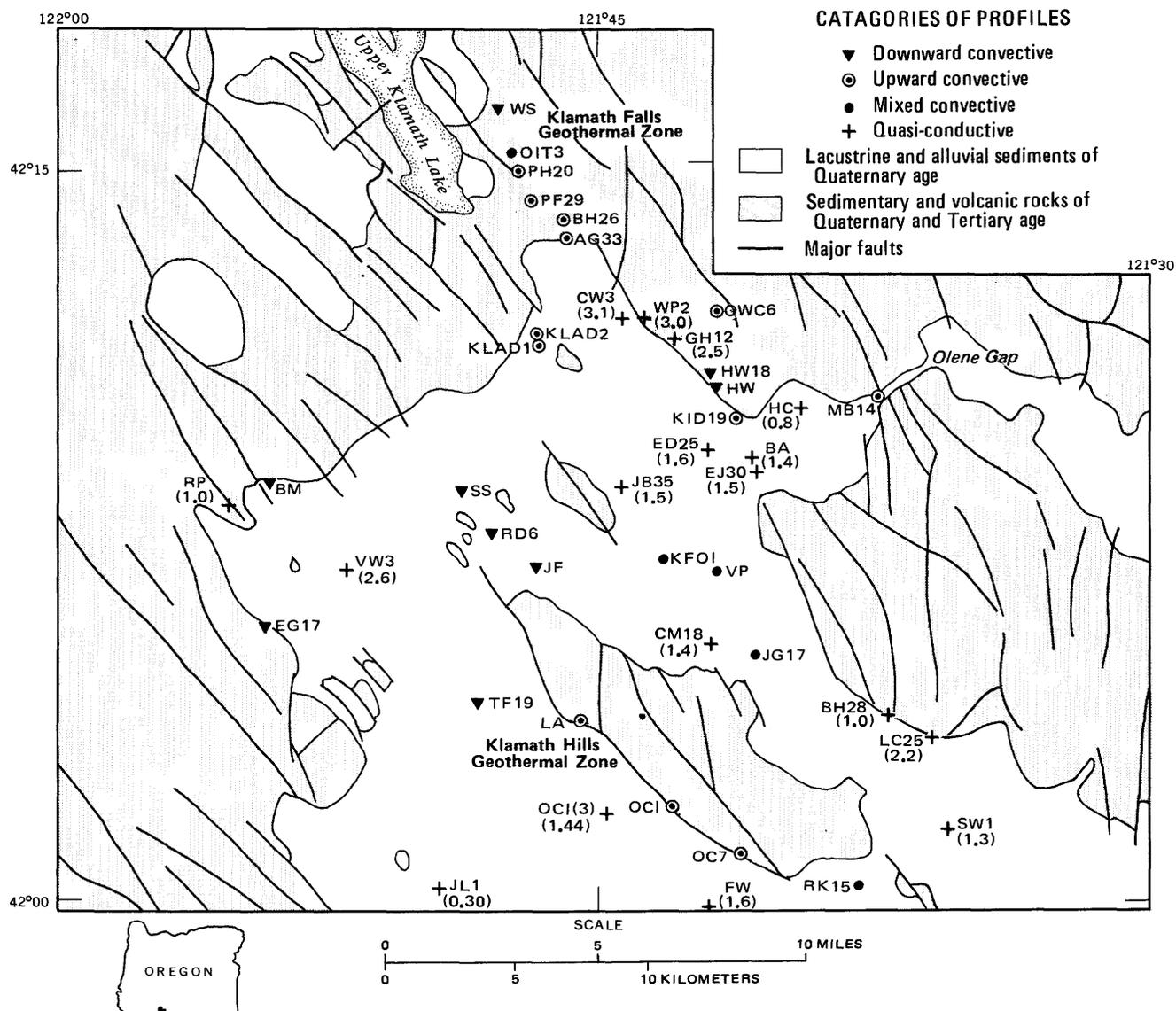
FIGURE 13.—Geologic sketch map of the Klamath Falls area (after Peterson and McIntyre, 1970), showing locations and categories of wells in which temperature profiles were obtained. For wells with quasiconductive (linear) profiles, a lower limit of heat flow, based on an assumed thermal conductivity of 1.8 mcal $cm^{-1}s^{-1}°C^{-1}$, is shown in parentheses. Heat-flow determinations at JL1 and OC1(3) are also shown. (Adapted from Sass and Sammel, 1976.)

peratures and small thermal gradients. Most of these wells are located in the lowlands of the area and are relatively shallow. The profiles show that downward flow of cool water occurs to depths of 300 feet at places, indicating that the lake-bed sediments are relatively permeable at these locations.

Most of the areas dominated by convective flow and cold water are ground-water recharge areas. Significant amounts of recharge may also occur elsewhere as indicated by profiles that, although they are dominantly conductive, show evidence of downward circulation in their upper 50 to 100 feet. An example is the U.S. Geological Survey test hole, OC1 (3).

One of the deeper wells, OIT 3, shows downward convection to a depth of more than 900 feet. This well, with a maximum temperature of 28°C, is located less than 2000 feet from one of the three hot wells at the Oregon Institute of Technology (fig. 4). The temperature difference between these wells (about 60°C) and the difference in static water levels (210 feet) demonstrate the profound effect exerted by local structure and stratigraphy on subsurface temperatures and ground-water regimes.

The third category in table 10, termed "Irregular," includes all hot wells in which profiles were obtained as well as several cold wells. Most of these wells penetrate basalt flows of the Yonna Formation; two of them,

TABLE 10.—*Data related to temperature profiles obtained in unused wells and heat-flow holes*

[*Gradient used in heat-flow calculations. See text for explanation]

Name	Symbol[1]	Location	Depth ft.	Completion[2]	Altitude land surface ft. above MSL	Depth to basalt rock[3] ft.	Thermal gradient[4] °C Km⁻¹	Interval[5] ft.	Maximum temperature °C	Remarks and heat-flow estimates
Mainly conductive profiles										
*Robert Puckett	†RP	39/8–31acd	168	OH(30)	4130	145	*55	50–168	14	On terrace below ridge. Diatomite and shale overlying basalt. Minimum heat flow 1.0 HFU.
*Cass Windsor	CW3	39/9–3ddd	500	OH (<300?)	4115	N	*170	80–490	38	"Gray shale." Minimum heat flow 3.1 HFU.
*George Hergesheimer	GH12	39/9–12bcd	635	OH(19)	4140	N	*138	260–490	36	"Gray shale." Minimum heat flow 2.5 HFU.
*E. D. Daniel	ED25	39/9–25aaa	231	OH(20)	4090	N	*90	130–230	16	Sand and clay. Minimum heat flow 1.6 HFU.
*Joe Bair	JB35	39/9–35cbb	171	OH	4095	N?	*86	33–165	13	"Green shale" reported in hole ¼ mile north. Minimum heat flow 1.5 HFU.
H. Waggoner	HW18	39/10–18bcb	375?	OH(140)	4175	N	50	100–245	15	Downward convective above 100 ft.
*H. Campbell	†HC	39/10–21bba	453	OH	4120	N	*47	33–158	12	USGS test hole. Profile measured 6 days after drilling completed. Hole caved to 158 ft. Mostly clay below 60 ft. Minimum heat flow 0.8 HFU.
*Ben Adair	†BA	39/10–30ada	238	OH	4095	N?	*80	65–238	16	Downward convective above 60 ft. Probably penetrates lake deposits. Minimum heat flow 1.4 HFU.
*Edwin Jenkins	†EJ30	39/10–30add	178	OH	4090	N?	*~80	65–175	14	Upward convective curve over entire profile. Probably penetrates lake deposits. Minimum heat flow 1.5 HFU.
*V & W Ranch	VW3	40/8–3cda2	236	------	4090	------	*~150	65–236	19	On irrigated valley floor. Irregular profile, downward convective above 40 ft. Probably penetrates lake deposits. Minimum heat flow 2.6 HFU.
*Charles Matney	CM18	40/10–18cbb	470	OH(60)	4080	------	*~75	39–460	21	Irregular profile; upper 60 ft cooled by convection. Probably penetrates lake deposits. Minimum heat flow 1.4 HFU.
*Larry Clark	LC25	40/10–25ccb	130	------	4120	------	*~120	65–125	16	Downward convective, above 40 ft; gradient decreases below 120 ft. Probably penetrates lake deposits. Minimum heat flow 2.2 HFU.
*Bill Hill	BH26	40/10–26bca2	350	------	4155	------	*~60	195–290	16	Irregular profile; downward convection upper 140 ft. Probably penetrates Yonna Formation. Minimum heat flow 1.0 HFU.
*U.S. Bureau of Reclamation (J.D. O'Connor, lessee).	OCI(3)	41/9–3dbb	608	Plug	4080	N	*78.6	180–580	25	USGS heat-flow hole. Uniform gradient below 140 ft; cooled by convection, upper 25 ft. Silt and clay. Heat flow 1.44 HFU.
*Suburban Water Co.	SW1	41/10–1cdd2	335	OH(35)	4065	285	*74	80–285	20	Downward convective above 70 ft; profile offset at 285 ft at top of basalt. Minimum heat flow 1.3 HFU.
Ray Kramer	RK15	41/10–15abc	241	OH(42)	4090	80	~80	10–240	15	Gradient given for overall profile, although profile shows downward convection between 10 and 200 ft. Mostly clay and sand; two thin basalt layers above 100 ft.
*U.S. Fish and Wildlife	F & W	40N/2E–13d (California)	1100	OH	4080	N	*90	195–655	25	Downward convection above 110 ft; uniform gradient, 150–630 ft. Silt, clay, some sand in lake deposits. Minimum heat flow 1.6 HFU.
Cold convective profiles										
William Smith	†WS	38/9–17bbb	135	OH(19)	4240	100	------	------	14	Convective 80–120 ft; gradient increases 120–130 ft (gray basalt).
Oregon Institute of Technology #3.	OIT 3	38/9–17cdd	1130	OH(700)	4353	205	------	------	28	High gradient begins at about 940 ft; decreases toward bottom. ("Brown-shale" between basalt flows.)
Bill McConron	†BM	39/8–28dcb	105	P(84–104)	4160	3	------	------	14	Nearly isothermal profile.
Sam Shaw	†SS	39/9–30ccd2	110	OH(49)	4090	N	------	------	11	Temperature decreases, 40–90 ft; gradient increases, 90–110 ft, (top of "green clay").
H. Waggoner	†HW	39/10–18bcd	100	P(30–62) OH(62)	4170	N	------	------	12	Nearly isothermal profile. Clay, sand, "green shale," "black sandstone."
Klamath Irrigation District.	KID19	39/10–19aca	450	OH	4120	N	~30	300–390	16	USGS test hole. Nearly isothermal, 20–300 ft; gradient increases at 300 ft (silts); colder zone in sands at 160 ft.
Ed Gardner	EG17	40/8–17bad	175	OH(99)	4110	30	------	------	10	Small gradient above 100 ft; becomes isothermal below 140 ft. Stratified volcanic rocks.
J. E. Frost	†JF	40/9–5dda	310?	------	4115	N	------	------	13	Profile becomes colder, 30–90 ft.
Ray Dillon	RD6	40/9–6aba	930	OH(48)	4090	490?	~20	330–490	19	Convective cooling above 300 ft. Probably Yonna Formation.
Tulana Farms	TF19	40/9–19dcc	223	OH	4080	N	~40	100–160	13	USGS test hole. Conductive gradient begins in silty clays below sandier zones.
*U.S. Bureau of Reclamation (Jack Liskey, lessee).	JL1	41/8–13acd	605	Plug	4080	N	*16.8	245–605	15	USGS heat-flow hole. Upper 190 ft cooled by convection. Clay, silt, some sand. Heat flow 0.30 HFU.
Irregular profiles										
Presbyterian Inter-community Hospital.	PH20	38/9–20acb3	1584	OH(1003)	4370	841	------	------	92	Gradient very high, 550–650 ft; decreases to isothermal below 1,000 ft. Yonna Formation overlying Tertiary basalts.
W. A. Hughes	BH28	38/9–28dcd	266	OH(256)	4160	95	------	------	94	Gradient very high above 150 ft; isothermal below 160 ft. Yonna Formation; clay, gravel, interbedded basalt.
Richard Pope and Fred Foulon.	PF29	38/9–29aad	194	OH(194)	4200	------	------	------	90	Isothermal 96–150 ft; temperature decreases below 150 ft. Hottest temperatures at probable top of basalt.
Assembly of God Church	AG33	38/9–33aba	363	P(170–185) (333–363)	4150	162	------	------	88	Very high gradient, 42–180 ft; isothermal, 190–350 ft. Probably Yonna Formation.
*Wiard Park District	WP2	39/9–2cdd	1018	OH(19)	4140	N	~160	580–948	51	Convective cooling above 90 ft; relatively uniform gradient below 580 ft except for bulge below 800 ft. Bulges correspond to occurrences of water-bearing sand layers. Sands, silts, clays. Minimum heat flow 3.0 HFU.
J. D. O'Connor	KLAD2	39/9–9bcb	275	------	4085	------	~180 (ave)	90–275	28	Convective cooling above 90 ft; upward convective bulge, 100–260 ft. Probable lake deposits.
J. D. O'Connor	KLAD1	39/9–9cbd	225	------	4115	------	------	------	20	Convective cooling above 180 ft. Probable lake deposits.

TABLE 10.—*Data related to temperature profiles obtained in unused wells and heat-flow holes*—Continued

Name	Symbol[1]	Location	Depth ft.	Completion[2]	Altitude land surface ft. above MSL	Depth to basalt rock[3] ft.	Thermal gradient[4] °C Km⁻¹	Interval[5] ft.	Maximum temperature °C	Remarks and heat-flow estimates
Oregon Water Corp.	OWC6	39/10–6cdb	732	OH(147)	4310	152	------	------	33	Very low gradients above 520 ft; warm bulge below 520 ft; nearly isothermal, 700–730 ft. Top of basalt layer at 514 ft.
Marion Barnes	MB14	39/10–14ccb	380	OH(41)	4120	22?	~60	280–380	55	Very high gradient above 190 ft. Water-bearing sands in "green claystone" account for bulges below 190 ft.
Klamath Falls Oil Co. #1	KF01	40/9–2daa2	1965	------	4125	600?	~70	120–620	31	Basalt reported at 600 ft (questionable—may be Yonna Formation). Nearly isothermal, 740–1,320 ft.
Leon Andrieu	†LA	40/9–28adb	128	OH(23)	4120	N	------	------	15	Convective cooling to 60 ft (water-bearing gravel); top of conductive gradient in "shale."
Vern Peugh	†VP	40/10–6cca	120	OH(31)	4080	N	------	------	13	Convective cooling, upper 40 ft; gradient increases, top of "blue-green clay" at 110 ft. Sand and clay.
Joe Glodoski	†JG	40/10–17ccb	189	OH(60)	4075	N	------	------	15	Convective cooling, upper 50 ft; gradient decreases in blue clay (60–160 ft); isothermal in coarse to fine sand (160–189 ft).
O'Connor Livestock Co.	OC1	41/9–1bcd	105	------	4155	------	------	------	19	Convective cooling above 60 ft. Probably Yonna Formation
O'Connor Livestock Co.	OC7	41/10–7dba	770	OH(79)	4120	674	------	------	30	Moderately high gradient to 100 ft ("hard green shale, water-bearing"); isothermal, 100–490 ft ("hard green shale" and "green clay").

[1]Location on plate 3 and profile on plate 4 identified by this symbol, † Shallow, cold well, not identified on plate 4.
[2]OH=Open hole below depth given (ft.); P=Perforated in interval given (ft).
[3]From driller's log; N=Not encountered.
[4]In general, the maximum gradient measurable over a significant depth interval; in wells designated by *, the conductive gradient for the given interval.
[5]Interval over which thermal gradient was determined.

PH 20 and OWC 6, also penetrate older Tertiary basalt.

Three irregular profiles in the upper right corner of plate 4 probably represent conditions at many places in and near the principal geothermal areas. These profiles, PF 29, BH 28, and AG 33, show typically high thermal gradients through relatively impermeable shale and clay of the Yonna Formation. Beneath these zones of high gradient, the profiles became isothermal in layers of permeable basalt and volcanic sediments. Profiles such as these suggest that hot water moves laterally in the permeable basalt layers and that this lateral movement may account for the widespread distribution of warm ground water in places such as the lowlands south of Klamath Falls and Altamont.

Most hot wells in the vicinity of Klamath Falls and Olene Gap have irregular but generally convex-upward temperature profiles. To the extent that these profiles represent formation temperatures they indicate that both upward and lateral convective transport of heat occur in the aquifers. It is assumed that wells penetrating the Yonna Formation in the Klamath Hills geothermal area would show similar profiles. On the basis of the temperature profiles and the known discharge of thermal waters, these three areas are believed to be the principal thermal discharge areas.

In the foregoing discussion it is tacitly assumed that the temperature profiles measured in most wells represent temperatures in formations surrounding the wells. Some of the data make it clear, however, that many profiles do not accurately represent formation temperatures. Discrepancies between temperatures measured in wells and actual formation temperatures are usually due to one or both of two common occurrences: vertical convective flow in the well itself, and in cased wells, vertical flow in the annular space between the casing and the formation. Such vertical flow occurs in response to both temperature gradients and hydraulic gradients in the formation penetrated by the wells.

An isothermal profile caused by convective flow was measured in well OC 7 (pl. 4). The temperature profile in this uncased well is isothermal at 30°C between depths of 120 and 500 feet, the greatest depth to which measurements were made. Subsequently, when the well was pumped, the temperature of the water was found to be 37°C. It is likely, therefore, that under non-pumping conditions, water with a temperature of about 30°C enters the well at shallow depths and flows downward with sufficient velocity to maintain an isothermal profile. Thus, temperature gradients measured in this well have little relation to geothermal gradients in the formation.

A second example can be seen by comparing temperature profiles obtained in two of the deepest wells in the area, PH 20 (1,584 feet) and OIT 6 (1,805 feet). These wells are located about 500 feet apart and drillers' logs indicate that they penetrate the same formations. The temperature gradient measured in well PH 20, however, ranges from 330°C km⁻¹ at about 575 feet depth to nearly isothermal at 1,475 feet, whereas the gradient calculated from the bottom-hole measurements in OIT 6 over the same interval is more nearly uniform at about 180°C km⁻¹. The bottom-hole measurements are believed to provide a good estimate

of the actual thermal gradient at this location in spite of small errors in absolute temperature that may have been introduced by the cable-tool drilling. Therefore, both the high gradient and the nearly isothermal gradient measured in PH 20 are almost certainly the products of convection in the well and are not reliable measures of gradients in the formation.

It is of interest to note also that the position of the sharp break in bottom-hole temperatures in the profile of OIT 6 near the 1,740 foot depth (pl. 4) corresponds to that of a gravel layer between basalt flows noted in the driller's log. The thermal discontinuity is probably due to the rapid movement of lower-temperature water through the gravel layer. Small lithologic and thermal discontinuities of this kind are difficult to detect in temperature profiles obtained in water wells, but they may strongly affect the measured geothermal gradients at some places.

Several examples of stratigraphic discontinuities were obtained in the temperature profiles. Wells OIT 3, OWC 6, and MB 14 had lower temperatures when pumped than the maximum temperatures measured under static conditions. These wells appear to derive much of their discharge from cool strata at fairly shallow depths. The decreases in temperature observed during pumping may be taken as evidence of low vertical permeabilities in the formations and the stratification of the thermal and nonthermal waters at these locations.

The examples cited are believed to represent conditions that occur widely in the shallow hydrothermal system at Klamath Falls. They illustrate the difficulties that may be encountered in attempts to obtain reliable geothermal gradients in holes not especially constructed for this purpose, and they serve as a caution that the thermal gradients reported in this study should not be used uncritically in evaluating geothermal prospects in the area.

CONDUCTIVE HEAT FLOW

Two 600-foot holes were drilled by the Geological Survey in the Lower Klamath Lake area in order to obtain heat-flow information. These holes and the data obtained from them are described in detail in Sass and Sammel (1976). The temperature profiles from these holes are shown on plate 4. The first hole, JL 1, is on reclaimed marsh land in Lower Klamath Lake near the California border at T41, R8, Sect 13 acd. The hole penetrates lacustrine silt and clay to a depth of 605 feet. The mean thermal conductivity of cores from the hole is 1.82 mcal cm^{-1} s^{-1} °C^{-1}. For the interval between 167 and 587 feet the average gradient is about 16.75°C km^{-1}. The heat flow at the site, therefore, is calculated to be 0.30 heat flow units (HFU).

The second hole OC 1 (3), was drilled at a location northeast of JL 1, about 1 mile southwest of the Klamath Hills. For this hole, the mean thermal conductivity is 1.83 mcal cm^{-1}s^{-1} °C^{-1}, the temperature gradient in the interval 174 to 577 feet is 78.6°C km^{-1}, and the heat flow, therefore, is 1.44 HFU.

The heat flow calculated for hole JL 1 (0.30 HFU) is too low to represent actual heat loss from the earth's interior over a broad area. Convective cooling by recharge of ground water to depths greater than 600 feet must be invoked in order to explain the observed heat flux, which, in the light of data discussed below, may be anomalous, even for the supposed recharge areas of Lower Klamath Lake.

Heat flow in the second hole, OC 1 (3), (1.44 HFU), is near the expected lower limit for conductive heat flow in the Basin and Range province (Sass and others, 1971). Data obtained from other conductive temperature profiles in the area suggest that hole OC 1 (3) probably represents near-normal conductive heat flow over much of the Lower Klamath Lake basin. The basis for this conclusion is discussed in the following paragraphs.

A possible range of heat-flow values in the Lower Klamath Lake basin was determined by using temperature gradients from 15 conductive temperature profiles obtained in the basin (table 10). These gradients were combined with the mean thermal conductivity determined for heat-flow holes JL 1 and OC 1(3), which is believed to be a representative value for much of the lower basin. Each of the wells in which these profiles were obtained penetrates unconsolidated to semiconsolidated lacustrine deposits in the Lower Klamath Lake valley. The resulting estimates of heat flow (table 10) are regarded as minimum values for the sites.

Calculated heat flows for the 15 wells range from 0.8 to 3.1 HFU, with the highest values coming from wells nearest the Klamath Falls and Olene Gap thermal areas. Seven of the wells in the southern part of the valley appear to represent average conditions over a wide area. Heat flows calculated for these wells range from 1.3 to 1.6 HFU. The calculated heat flow for OC 1(3), 1.44 HFU, therefore appears to represent near-normal heat flow over much of the Lower Klamath Lake graben.

The wide range of the estimated minimum heat flows for the remainder of the wells (0.8 to 3.1 HFU) tends to confirm the conclusion, based on variations in the temperature profiles, that localized geologic and hydrologic conditions determine the temperature regime at most places. Many of the higher temperatures attained in the profiles appear to be caused by the insulating effect of poorly conductive unconsolidated deposits overlying basalt aquifers. Others are apparently influenced by

lateral convective flow of heat and water in permeable zones within the basalt sequences. In general, the observed temperature profiles do not provide reliable estimates of regional temperature gradients and regional heat flux for areas outside the Lower Klamath Lake basin.

CONVECTIVE DISCHARGE OF HEAT

Estimates of heat discharged from springs and hot wells in the three principal geothermal areas have been calculated from measured and estimated ground-water discharge at land surface. An average base temperature of 12°C is assumed for ground water near the land surface. Results are given in table 11.

In the absence of detailed studies, the estimate of spring discharge into Upper Klamath Lake is little more than a guess. The remainder of the figures in table 11 are probably reasonable approximations in the light of presently available data.

Heat discharged at the water table or by the mixing of thermal and nonthermal ground water in the vicinity of the three geothermal areas would probably add large heat fluxes to those listed in table 11. In assessing this thermal discharge, one may compare the temperature distribution shown on plate 3 with the contour map of the shallow ground-water surface (pl. 1). In the vicinity of Klamath Falls, for example, shallow ground water apparently flows from the highlands surrounding the city, traverses the zone of highest temperatures, and flows southward across the valley floor into an area where ground-water gradients are very low. The temperature contours suggest that the thermal waters mix and cool rapidly in the immediate vicinity of the principal thermal area and that slow convective upwelling and additional cooling occur in an extended area to the south.

This simple conceptual model is undoubtedly complicated in actuality by the stratified flow of hot water in basalt layers in the Yonna Formation. The general patterns of fluid flow and thermal mixing postulated above, however, are in accord with the evidence of the chemical data, and these convective flow patterns are believed to occur in the vicinity of each of the three principal thermal areas.

Because the patterns of lateral and vertical convection are complex in detail and are not well known, it is not possible to make a reliable estimate of the total convective heat discharge. On the basis of several reasonable assumptions, however, a crude estimate has been made of probable upper and lower limits for the total heat discharged by convective flow. For this estimate, it is assumed that ground-water flow within 500 feet of land surface in the vicinity of the geothermal areas occurs mostly in the Yonna Formation. The

TABLE 11.—*Mass and thermal-energy discharge in ground water from thermal springs and wells in the Klamath Falls area*

	Mass flow (gal d^{-1})	Thermal energy[1] (10^5 cal s^{-1})
Springs		
Olene Gap	30,000	1.2
Eagle Point and Upper Klamath Lake.[2]	43,000	.4
Wells		
Residences, schools, and commercial establishments.[3]	895,000	18.1
Klamath Hills (irrigation and stock wells).	1,900,000	1.8
Totals	2,868,000	21.5

[1]Heat discharge above datum of 12°C assumed for shallow ground water.
[2]Spring discharge into Upper Klamath Lake estimated by observation of winter thawing of lake surface. Estimate not reliable.
[3]Based in part on numbers of wells, flow rates, and temperatures described by Lund, Culver, and Svanevik, 1974.

transmissivity of the aquifer is estimated to be in the range 1,000 to 5,000 ft^2 d^{-1}. Total length of the three thermal discharge areas is probably greater than 3 and less than 5 miles, and the ground-water gradients away from these discharge areas may range from 20 to 50 ft mi^{-1}. Assuming, finally, that ground-water temperature decreases from 95 to 12°C, convective heat dissipated in the shallow ground-water body or discharged at the water table is calculated to be in the range 2×10^6 to 30×10^6 cal s^{-1}. Combining these figures with the estimate of thermal flow from springs and wells (table 11), the total convective flux in the vicinity of Klamath Falls is estimated to be in the range 4 × 10^6 to 32 × 10^6 cal s^{-1}.

The range of values given above may be compared with an estimate of geothermal heat being utilized at present for space heating and other purposes. Lund and others (1978, p. 45) estimate that about 6 megawatts per year (1.4 × 10^6 cal s^{-1}) are currently utilized in some manner. The geothermal heat actually used is, therefore, a small fraction of the heat now being discharged by natural or artificial means in the Klamath Falls area.

INTERPRETATIONS AND CONCLUSIONS

REVIEW OF EVIDENCE RELATING TO THE GEOTHERMAL RESOURCE

GEOLOGIC EVIDENCE

Known geothermal reservoirs in the Basin and Range province appear to attain their high temperatures in two ways: (1) by conduction of heat from a magma body emplaced at fairly shallow depths, and (2) by circulation of ground water to depths at which the normal thermal gradient in crustal rocks provides the required temperatures. These two mechanisms may exist in combination.

At most places where the first of these two mechanisms provides heat for known geothermal systems, the magma bodies are composed of silicic melts, often of crustal origin, that are both large enough and young enough to retain significant amounts of heat at the present time.

In southeastern Oregon, silicic volcanic activity has occurred during the last 10 million years in a westward-trending progression which culminated in the extrusion of rhyolitic domes and obsidian flows at Newberry Volcano, near Bend (MacLeod and others, 1975). The youngest silicic volcanic rocks at this location are less than 10,000 years old. In the Klamath Falls area, however, no silicic volcanic rocks occur at the surface and none has been penetrated by wells. When the trends of silicic volcanic activity in southeastern Oregon are extrapolated into the Klamath Falls area these trends suggest that if silicic intrusives existed, they would probably be about 4 million years old. On the basis of calculations made for the Long Valley, California, geothermal area (Lachenbruch and others, 1976), it is concluded that 4-million-year-old buried domal masses of average size would have cooled to near ambient temperatures by the present time. Thus, the available geologic evidence makes it appear improbable that the geothermal system at Klamath Falls is heated by silicic intrusive rocks, although this possibility is not ruled out.

Young volcanic rocks in the Klamath Falls area, some of them dating from late Pleistocene time, are of basaltic composition. At depth they probably occur as dikes and plugs in conduits that formerly conducted basaltic magma to eruptive centers and flow vents. These conduits probably extend deep into the crust, but there is little likelihood that they contain high-temperature rocks at depths shallow enough to be reached by circulating ground water.

The second of the two processes by which geothermal fluids are heated, circulation of ground water to great depths in a temperature field created by the earth's geothermal gradient, requires the presence of permeable channels through which flow can occur. The large Basin and Range faults that bound the Klamath Falls graben could provide the required channels as similar faults are assumed to do elsewhere in the Basin and Range province (White, 1968; Olmsted and others, 1975). The depth that would have to be reached by the Klamath Falls thermal waters in order to attain a temperature of 150°C is 15,000 feet if the geothermal gradient is as low as 30°C per kilometer. If the gradient is higher, as would be expected if rocks of low thermal conductivity overlying the reservoir created a blanketing effect, required depths would be smaller. Evidence from earthquakes in southeastern Oregon (Couch and

Lowell, 1971) suggests that faults in this area may extend to depths far greater than 15,000 feet, and it is probable, therefore, that these faults and associated fracture zones form the channels in which the Klamath Falls thermal waters circulate.

GEOPHYSICAL EVIDENCE

Temperature measurements in springs, wells, and test holes provide the only direct geophysical evidence relating to the geothermal resource at Klamath Falls. The temperature measurements show that, although warm ground water occurs throughout the lower Klamath Lake valley, hot water (>60°C) is discharged principally in three elongate zones associated with major faults. Hot water may also be discharged from concealed faults beneath the lacustrine deposits of both upper and lower Klamath Lake valleys, and the presence of water with temperatures greater than 30°C at a few places in the Lower Klamath Lake valley may be evidence of this occurrence (see pl. 3).

Temperature gradients observed in water wells are generally higher near the principal geothermal areas than elsewhere, but few of these gradients are reliable indicators of crustal temperature regimes. Most of the temperature profiles obtained for this study are strongly affected by the convective transfer of heat in moving ground water and hence cannot be used for estimates of conductive heat flow.

A few conductive thermal gradients obtained in the Lower Klamath Lake Basin suggest that heat flow through the lake sediments may be on the order of 1½ HFU. This value is at the low end of expected values for the Basin and Range province. Although this evidence cannot be considered highly reliable, it greatly decreases the probability that a source of magmatic heat exists at shallow depth beneath the Lower Klamath Lake graben. Temperature measurements from the upper Klamath Lake basin and from the upraised blocks surrounding the basins are too scanty to permit conclusions regarding possible magmatic sources of heat in these areas.

Indirect geophysical evidence obtained from gravity and aeromagnetic surveys has been described and interpreted by D. L. Peterson in a previous report (Peterson, 1976). In general, the gravity and aeromagnetic trends coincide with the dominant northwest-southeast fault trend through the area. The data indicate that the lowlands are sunken blocks bounded by steeply dipping normal faults and covered by great thicknesses of low-density sediments. Gravity modeling suggests that the graben structure may be 3,000 feet deep in the area southeast of Klamath Falls and may be more than twice as deep southwest of Klamath Hills.

The gravity data also indicate the presence of structures that may be only suspected from surficial evidence. Examples are a possible northeast-southwest trending fault passing through Olene Gap and an east-west trending fault near the north end of Upper Klamath Lake. A major gravity anomaly in the area southwest of Keno may be related to rocks of the adjacent eruption center. Of these anomalies, only the postulated fault through Olene Gap is related to known geothermal occurrences.

Peterson's interpretation of the aeromagnetic map indicates that most magnetic anomalies are related to the same structures that cause the gravity anomalies. For example, gravity and magnetic lows both occur in the deep structural depression in the Lower Klamath Lake basin. In addition, magnetic anomalies suggest the presence of buried basaltic rocks in the valley-fill deposits at Squaw Point in Upper Klamath Lake and at a location near Keno in Lower Klamath Lake.

The gravity and magnetic data suggest that subsurface structure exists at many places in marked contrast to the nearly featureless valley floor. Transecting faults apprently offset the basement topography in a pattern of tilted blocks which have a wide range of depths beneath the valley surface. Basalt flows intercalated in the valley-fill deposits add to the complexity of the geophysical patterns.

The hydrologic consequences of these discontinuities are many. Locally, the boundaries of the lithologic units may be relatively impermeable, thereby contributing to the segregation of waters of differing chemical quality. Elsewhere, basalt flows and fault zones may form distribution channels for thermal waters. These phenomena are not well known in detail, but they will have important consequences in the future if significant quantities of thermal water are withdrawn from and reinjected into the thermal reservoir.

CHEMICAL EVIDENCE

Hot waters from the three geothermal areas of Klamath Falls have similar, although not identical, chemical characteristics. The similarities suggest that the waters from the three areas have equilibrated in nearly identical rocks at similar depths and temperatures.

On the basis of the chemical data, the thermal waters appear to circulate in a hot-water system of moderate temperature (<150°C). Ratios of boron and fluorine to chloride indicate that the water is heated mainly by conduction rather than by condensation of steam from an underlying vapor-dominated system. Hot water rising in the principal conduits probably cools largely by conduction, with little loss of volatile elements and little exchange of ions with the surround-

ing rocks. Thus, the chemical compositions of the hotter waters observed at the surface may be closely similar to the compositions of water in the reservoir rocks.

Sources of the meteoric recharge waters and flow paths through the geothermal system have not been identified in this study, but the chemical data are consistent with the hypothesis that local ground water furnishes recharge for the system. Primary evidence for this hypothesis consists of similar concentrations of deuterium and similar ratios of SO_4/Cl and B/Cl in the hot and cold waters as well as expected changes in concentrations of sodium, chloride, sulfate, silica, and bicarbonate. Most shallow ground waters in the region have similar chemical compositions, and the available chemical analyses do not make it possible to distinguish separate recharge waters or reservoir waters.

Ratios of ionic constituents and concentrations of stable isotopes in the thermal waters are qualitative indicators of the probable range of reservoir temperatures at Klamath Falls. Application of the quartz geothermometer to mixing models based on silica and stable-isotope concentrations indicates that reservoir temperatures are greater than 145°C, but the moderately high Ca/HCO_3 ratios, high Na/K ratios, and the small shift of $\delta^{18}O$ from the levels measured in local meteoric waters suggest that temperatures are probably not much higher than 150°C.

The estimates of reservoir temperature derived from both silica and $^{18}O(H_2O)$ concentrations are reassuring in their general consistency and are supported by the patterns of ionic ratios in the thermal waters. These results must be weighed, however, against the higher estimates derived from $^{18}O(SO_4)$ concentrations. (See section "Reservoir Temperature Estimated from Oxygen Isotopes in Dissolved Sulfate.") If the concentration of ^{18}O in dissolved sulfate is a reliable indicator, the thermal waters have, at some point in their history, equilibrated at temperatures that could have been as high as 196°C and were at least 180°C.

Two possible models are offered in explanation of the high reservoir temperatures indicated by the sulfate oxygen isotope. In model (1), a two-stage reservoir is postulated, with the two stages separated in either space or time or both. In this model, ^{18}O (SO_4) in deep thermal water is assumed to equilibrate with $^{18}O(H_2O)$ at temperatures greater than 190°C for a period of time which may be as long as several tens of years (McKenzie and Truesdell, 1977, p. 53). Other constituents, including silica, would also equilibrate at these temperatures. At some later time and at a shallower depth, the water re-equilibrates at lower temperatures, acquiring the chemical characteristics observed at the surface and leaving only the sulfate-oxygen isotope as a vestige of its former chemical state. In model (2), isotopi-

cally light SO_4 is formed in volcanic rocks during their emplacement at high temperatures. Subsequent exchange with circulating thermal waters tends to shift the $^{18}O(SO_4)$ concentrations toward their equilibrium values at much lower temperatures in a process that is still incomplete.

These explanations are merely speculative and others undoubtedly are possible. Until such alternative explanations can be either confirmed or ruled out by additional evidence, the apparent high reservoir temperatures indicated by the sulfate-oxygen isotope should probably be viewed with scepticism in the light of the remainder of the chemical and isotope data.

Analysis of the stable-isotope concentrations suggests rather strongly that thermal waters in the three principal geothermal areas near Klamath Falls rise from three similar but distinct reservoirs. The oxygen and deuterium relations indicate separate sources of recharge for waters of the Klamath Falls hot-well area and those of Olene Gap, and the sulfate-oxygen geothermometer suggests yet a different history or origin for the Klamath Hills hot waters. Although these interpretations lack confirmation by other lines of evidence, they cast doubt on the existence of a single large reservoir beneath the Klamath Falls graben.

CONCEPTUAL MODEL OF THE GEOTHERMAL SYSTEM

The deep geothermal system at Klamath Falls has not been explored, and only near-surface indications and geophysical data provide indirect evidence of the nature of the system. Based on the evidence obtained during this study, a highly generalized conceptual model has been formulated. Verification and refinement of this model probably will require the acquisition of new data, and in particular, data from deep drill holes.

Geothermal phenomena at the earth's surface have the following fundamental causes: (1) the presence of an intrusive mass of molten or still hot rock; (2) high heat production from radiogenic sources; (3) convective transport of heat and fluids from depth in a cyclic system; and (4) the insulating effect of layers of rock having low thermal conductivities. Although each of these types of systems may produce high-temperature gradients near the earth's surface, only the first two are necessarily accompanied by higher than normal conductive heat flows. Most geothermal systems are produced by a combination of at least two of the possible causes.

On the basis of the evidence gathered during this study, the geothermal system at Klamath Falls is best explained as a convection system with deep circulation of meteoric water. Thus, the conceptual model proposed for this system is a fairly simple one. Meteoric water in

the Tertiary basaltic rocks of the region is assumed to percolate downward along the vast network of faults and fracture zones that characterize the regional structure. After attaining depths on the order of 15,000 feet the temperatures of at least 150°C, the water rises to shallow depths through conduits closely associated with major faults. If the two-stage reservoir postulated on the basis of sulfate-oxygen isotope data exists, the shallower of the two reservoir zones may occur at depths less than 10,000 feet. This zone may be supplied by slow upward leakage through a few conduits from depths as great as 15,000 to 20,000 feet.

The postulated depths are based on the assumption that geothermal gradients are uniform and are no greater than 30°C km^{-1}, a probable minimum value for the region. It is likely, however, that actual depths of circulation are less than those calculated above on the assumption of uniform thermal gradients in the crustal rocks. Normal heat flow in the Klamath Falls area is probably impeded by rocks of low thermal conductivity that overlie the sunken blocks in the Klamath Lake grabens. Within these grabens, isotherms are probably displaced upward, and assuming that offsetting thermal effects in adjacent upraised blocks are smaller than the insulating effect of the sunken blocks, temperature at a given depth will be higher than that predicted by a uniform normal gradient. The high thermal gradients (nearly 80°C km^{-1}) and low heat flows (<1.5 HFU) observed in the Geological Survey heat-flow hole, OC 1 (3), and in well FW in the Lower Klamath Lake basin are evidence for this concept.

Although the structural and lithologic conditions postulated above probably also exist in several of the basins adjacent to Klamath Falls, conditions for the upward flow of heated water presumably are especially favorable at Klamath Falls. These favorable conditions may include both the elevation of isotherms in a large volume of rock and the presence of relatively permeable flow channels in parts of the area. The fault-related conduits that permit reasonably free convective flow of hot water at three separate locations in the vicinity of Klamath Falls may be vestiges of the extensive and active hot-water system that must have operated to silicify some thousands of cubic meters of rock during Pleistocene or earlier time.

At present, the rising hot waters are probably largely confined to conduits in the three principal geothermal areas, with only minor amounts of upward flow occurring in the interior of the valley. If conduits exist in concealed faults beneath the valley bottoms they may be effectively sealed off at most places by the fine-grained deposits that occupy the valleys.

Hot waters from the principal geothermal areas originate and equilibrate under generally similar res-

ervoir conditions. Differing oxygen and deuterium isotope concentrations suggest, however, that the reservoir underlying the three areas is separated into at least three distinct zones. The boundaries of these postulated reservoir zones are unknown and their overall lateral extent can only be surmised from the extent of the surface phenomena. Based on surface manifestations, a single reservoir could extend over an area of at least 100 square miles. If, as seems more likely, the reservoir consists of more or less separate permeable zones, the total areal extent of these zones may be as small as 20 square miles.

Depth to permeable reservoir rocks beneath the valley floors may be as great as 6,000 feet based on inferred depths to basaltic rocks. In the upraised basaltic rocks adjacent to the valley, depth to areally extensive thermal reservoirs may be significantly smaller, perhaps no more than 2,000 or 3,000 feet at some places.

Most of the heated water rising from the reservoir does not reach land surface, but spreads through permeable zones in basaltic rocks beneath or within the Yonna Formation. In these strata, which occur at depths of a few tens of feet to at least 1,500 feet, the flow is predominantly lateral in response to local gradients in the shallow ground water reservoir. The hot water mixes with local meteoric water in these strata, and within a half mile of the major fault conduits, temperatures may decrease about 60°C.

The spread of hot water near the geothermal areas produces notable effects on temperature gradients in these areas. Most thermal gradients measured in the Klamath Falls area represent only local perturbations in the normal geothermal gradient that occur as the result of convective transport of heat in moving ground water. Gradients may be extremely high ($>150°C$ km^{-1}) in impermeable rocks of the Yonna Formation where these rocks overlie basalt or permeable gravel bearing hot water. At other places where the Yonna Formation is more permeable, temperature profiles may be nearly isothermal, and gradients may even be reversed beneath hot-water zones. These phenomena are observed at depths as great as 1,800 feet (plate 4) and they probably persist to even greater depths in the principal geothermal areas.

The conceptual model described above suggests that heat flow may be generally low in the thick valley-fill deposits of the area as the result of the concentration of the regional heat flow in a few small areas. The relatively low estimated heat flow observed in the lower Klamath Lake basin may be evidence of this effect. The model also implies that heat flow may be disturbed on the periphery of the basin as the result of displacement of temperature contours within the basins. This effect

is probably masked by topographic effects and by convective heat flow in the major geothermal areas; consequently, it may be difficult or impossible to detect.

As a rough check on the areal heat-flow balance, assume that the regional conductive heat flow is near the upper limit of normal values for the Basin and Range province, or about 2 HFU. Over the approximately 100 square miles of the Klamath Falls thermal area, total conductive heat flow is therefore calculated to be 5.2 $\times$ 10^6 cal s^{-1}. This figure may be compared with an estimate given earlier which indicates that the total convective thermal discharge from the Klamath Falls thermal areas is probably in the range 4 $\times$ 10^6 to 32 $\times$ 10^6 cal s^{-1}. According to these estimates, therefore, most of the normal heat flow through the crust in this area may be concentrated by convective discharge in the three geothermal areas. Uncertainty in the estimate of convective discharge is large, but even if the actual discharge were at the low end of the estimated range (4 $\times$ 10^6 cal s^{-1}), the regional conductive heat-flow patterns must be significantly altered by the convective discharge of heat at Klamath Falls. It is likely, therefore, that observations of heat-flow patterns within and adjacent to the Klamath Falls graben could provide a test of the conceptual model presented here.

Digital models may be used to obtain insight into permissible boundary conditions and heat-flow rates in the geothermal system. A model study for this purpose is currently (1979) underway in the USGS geothermal research program in Menlo Park. Results of this study will be presented in a subsequent report.

RECHARGE TO THE GEOTHERMAL SYSTEM

Postulated modes of recharge depend greatly on the assumed depth, nature, and extent of the geothermal reservoir. In the absence of definitive data on the reservoir, only general conclusions have been reached regarding the source and magnitude of recharge. Recharge may occur through the same fault systems that, at places, conduct the hot water to the surface, although it seems unlikely that downward flow would be confined to isolated channels as the thermal discharge appears to be. Rather, the recharge may occur as slow, diffuse seepage through extensive segments of the fault zones, perhaps at relatively great distances from the discharge sites.

Quantities of water available for recharge are large if the area of potential recharge is assumed to extend to a radius of, say, 25 miles from the center of the three geothermal areas. Over this area precipitation is about 2.4 $\times$ 10^6 acre-ft yr^{-1}, of which about 2.1 $\times$ 10^6 acre-ft yr^{-1} is lost through evapotranspiration. The remainder, about 0.3 $\times$ 10^6 acre-ft yr^{-1} is available for recharge or

runoff. It is likely, of course, that only a small part of the potential recharge can be captured in the areas that directly feed the geothermal system.

Ground-water flow into the Lower Klamath Lake basin from upper basins may be as much as 170,000 acre-ft yr⁻¹. If so, this flow would represent a major source of recharge to the geothermal reservoir. Because the Klamath Falls area is a large structural depression, shallow ground water may enter the area from all directions, although scanty observations suggest that there is an overall gradient through the area toward the southwest. A deep regional flow system could bring additional ground water into the area from the High Cascades, but there is no evidence to indicate whether or not such flow occurs.

MAGNITUDE OF THE GEOTHERMAL RESOURCE

On the basis of data available in early 1978, Brook and others (1979) estimated the total heat content of geothermal systems at Klamath Falls within 3 km of the surface to be 33.1×10^{18} joules (J). The estimate was based on an assumed total reservoir volume of about 125 km³ (30 mi³) in two separate areas, Klamath Hills and Klamath Falls. The Klamath Falls area included the springs at Olene Gap. Employing the figures given by these authors for the probable maximum and minimum errors of estimate, the probable maximum heat content is calculated to be 49×10^{18} J and the probable minimum heat content is 17×10^{18} J.

On the basis of additional temperature measurements, isotope mixing models, and differing estimates of reservoir volumes obtained in the present study, the writer has made new estimates of probable maximum and minimum heat contents. Unlike the previous estimates, which are based on differing mean reservoir temperatures for the Klamath Hills and the Klamath Falls systems (124°C and 111°C, respectively), the present estimate assumes a single reservoir temperature for all systems (150°C). This temperature represents a significant increase over those used previously, but it is clear from the calculations given below that the most significant parameters used in the estimation of heat content are those related to the volume of the reservoir, and, in particular, the estimates of area.

The probable maximum heat content is calculated by assuming that the reservoir is a continuous volume of basaltic rock in which hot water moves through fractures, permeable interflow zones, and fault zones. The reservoir is assumed to occupy an irregular but roughly rectangular area aligned northwest-southeast and extending from Upper Klamath Lake to the south end of the Klamath Hills, a distance of about 17 miles. The reservoir area is assumed to be bounded by the

west face of Stukel Mountain and the west face of the Klamath Hills, to extend east into Olene Gap, and to narrow to about 2½ miles in the vicinity of the city of Klamath Falls. The total surface area is approximately 108 mi² (275 km²).

Additional assumptions are as follows: (1) The reservoir temperature is 150°C, or 138°C above a base temperature of 12°C assumed for shallow ground water in the region; (2) the reservoir is 6,600 feet (2 km) thick above a depth of 13,000 feet (4 km); and (3) the specific heat of the reservoir rocks is 0.6 cal cm⁻³ °C⁻¹, based on averages for crustal basaltic rocks (John Sass, oral commun., 1974). Combining these assumptions with the estimate of reservoir area given above, the heat content is calculated as follows:

$$\text{Heat content} = 138(\text{°C}) \times 2(\text{km}) \times 10^5 \frac{(\text{cm})}{(\text{km})}$$

$$\times 0.6 \frac{(\text{cal})}{(\text{cm}^3)(\text{°C})} \times 275 \ (\text{km}^2) \times 10^{10} \frac{(\text{cm}^2)}{(\text{km}^2)}$$

$$= 46 \times 10^{18} \ (\text{cal}) \times 4.184 \text{ joules per calorie}$$

$$= 192 \times 10^{18} \text{ J.}$$

For an estimate of a probable minimum reservoir volume, it is assumed that the reservoir is confined to three separate elongate areas aligned with the main faults in the principal geothermal areas at Klamath Falls, Olene Gap, and the Klamath Hills. Widths of the areas range from 1 mile at Olene Gap and the Klamath Hills to 1¼ miles at Klamath Falls. Lengths range from about 4½ miles at Olene Gap to 6½ miles at Klamath Falls and the Klamath Hills. The total surface area is about 19 mi² (50 km²). It is further assumed that the reservoir is no more than 3,000 feet (1 km) thick. Thus, in this conceptual model, the reservoir is restricted to rocks immediately adjacent to the principal faults that conduct the hot waters to the surface. Total volume is assumed to be 12 mi³ (50 km³).

Based on the assumptions given above, and using the temperature and specific heat values employed for the previous estimate, the probable minimum amount of stored heat is calculated to be 3.5×10^{18} cal (14.6×10^{18} J).

The difference of an order of magnitude between the probable limits of reservoir volume given here reflects a serious deficiency in actual knowledge of subsurface conditions at Klamath Falls. The estimates can be regarded only as attempts to assign reasonable limits to the overall size and type of reservoir that might produce the observed geothermal phenomena.

If the actual heat content were determined to be in the upper part of the estimated range, the geothermal reservoir at Klamath Falls would be among the largest nongeopressured reservoirs in the United States. If, as seems more likely, the reservoir at Klamath Falls comprises two or more permeable hot-water zones that are separated by relatively unproductive rocks, the total reservoir volume may be near the probable minimum estimate given above. A most likely range of heat content, therefore, may be two to three times the probable minimum, or 7×10^{18} cal to 10×10^{18} cal (30×10^{18} J to 40×10^{18} J). Compared to heat contents estimated for other hydrothermal convection systems in the United States having temperatures near 150°C, this is still an impressively large number.

The above estimates of heat content, based on the concept of deep convective flow, would not be changed by assuming the presence of a hot igneous mass at greater depth. Although the ultimate recoverability of geothermal heat and the methods of exploiting it would differ depending on whether a heat source were present or not, the total heat stored to depths of 10,000 feet or more may fall within the estimated limits in either case.

As is true for most low-temperature hydrothermal systems, utilization of the Klamath Falls geothermal resource for production of electrical power is not economically feasible under present conditions. Future utilization for any purpose will probably be highly sensitive to the actual nature and extent of the reservoir. The facts can be ascertained only by additional exploration, by the acquisition of more complete chemical and isotope data, and by subjecting the data to more sophisticated analysis. Exploratory drilling to depths of at least 6,000 feet will probably be required in order to provide definitive answers to the many questions that remain unresolved by this study.

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