GEOCHEMISTRY AND HYDROLOGY OF THERMAL SPRINGS IN THE IDAHO BATHOLITH AND ADJACENT AREAS, CENTRAL IDAHO

By H. W. Young

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CONVERSION FACTORS

Conversion factors for terms used in this report are listed below. Chemical data are given in mg/L (milligrams per liter) or µg/L (micrograms per liter). Within the range of values presented, these values are numerically equal to parts per million or parts per billion.

Multiply	By	To obtain
acre	4,047	square meter
acre-foot (acre-ft)	1,233	cubic meter
calorie per second (cal/s)	4.187	watt
foot (ft)	0.3048	meter
gallon per minute (gal/min)	0.06309	liter per second
inch (in.)	25.40	millimeter
micromho per centimeter at 25 degrees Celsius (µmho/cm)	1.000	microsiemens per centimeter at 25 degrees Celsius
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer

Temperature in °C (degrees Celsius) can be converted to °F (degrees Fahrenheit) as follows:

$$°F=(1.8)(°C)+32.$$

All water temperatures are reported to the nearest one-half °C.

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ABSTRACT

The occurrence and nature of thermal springs in the Idaho batholith and adjacent areas suggest a relation between structural controls and deeply circulating hot-water systems. Springs issuing from granitic rocks are associated mostly with major regional fault structures. Springs issuing from other rocks probably are related to local faulting.

Individual spring flows and water temperatures are variable and range from less than 1 gallon per minute to 2,710 gallons per minute and from 20.5 degrees to 94.0 degrees Celsius. Annual spring discharge is at least 27,000 acre-feet; heat discharged convectively is estimated to be 5.0 x 10^7 calories per second.

Thermal springs discharge relatively dilute water; dissolved solids range from 103 to 839 milligrams per liter. The chemical quality of the water suggests deep circulation of meteoric water.

Estimated reservoir temperatures are generally less than 100 degrees Celsius, but temperatures for several springs exceed 150 degrees Celsius.

Stable-isotope data suggest that most of the thermal water is not derived from current precipitation. Carbon-14 values indicate that thermal waters are old; apparent residence times range from 9,000 to more than 40,000 years.

INTRODUCTION

This report describes the hydrology and geochemistry of the Idaho batholith and adjacent areas, one of the largest prospectively valuable areas for steam and associated geothermal resources in the Western United States, according to a map prepared by the National Geophysical and Solar-Terrestrial Data Center (1977). The study area includes about 23,200 mi^2 of the Payette, Boise, Salmon, and Clearwater River drainage basins and roughly coincides with surface exposures of the Idaho batholith (pl. 1).

This investigation is part of an overall program by the U.S. Geological Survey to better understand the nature and occurrence of geothermal resources in the Nation.

Purpose and Scope

The purpose of this study was to describe geochemistry and hydrology of thermal springs in the Idaho batholith and adjacent areas. Scope of this report entails summarization of chemical and isotopic data for current (1982) analyses and for the following analyses, which were described in detail in previous investigations by Lewis and Young (1980, 1982) and Young and Lewis (1982b): common ions, silica, and the minor elements boron, lithium, mercury, and arsenic (112 thermal and 30 nonthermal springs); deuterium and oxygen-18 (74 thermal and 30 nonthermal springs); tritium (13 thermal and 13 nonthermal springs); sulfate-water isotopes (19 thermal springs); carbon-14 isotopes (8 thermal springs); and carbon-13/carbon-12 analyses (8 thermal and 6 nonthermal springs). For convenience, chemical, deuterium, oxygen-18, and tritium analyses are included in data tables A, B, and C (back of report).

Previous Studies

Thermal waters in the Idaho batholith and adjacent areas were noted by Stearns and others (1937). Ross (1971) summarized existing data that included several chemical analyses of thermal waters. Young and Mitchell (1973) included chemical analyses of water from 32 thermal springs and 2 warm-water wells in their description of Idaho's geothermal potential. Lewis and Young (1980, 1982) and Young and Lewis (1982b) described the occurrence and chemistry of thermal springs in the Payette, Boise, and Salmon River basins.

Acknowledgments

Many landowners in the study area cooperated by allowing access to their property, supplying information about their springs, and permitting measurements of spring discharge. Special thanks are given to personnel of the U.S. Forest Service, Middle Fork Salmon River Ranger District, who provided transportation and assistance in sampling hot springs along the Middle Fork Salmon River. The author is grateful to the following Geological Survey personnel, who contributed significantly to this investigation: A. H. Truesdell, Cathy Janik, and N. L. Nehring provided sulfate-water isotope analyses; R. H. Mariner aided in the interpretation of geochemical data; and T. A. Wyerman provided tritium isotope analyses.

Spring-Numbering System

The spring-numbering system (fig. 1) used by the U.S. Geological Survey in Idaho indicates the location of springs within the official rectangular subdivision of the public lands, with reference to the Boise base line and meridian. The first two segments of the number designate the township and range. The third segment gives the section number; three letters, which indicate ¼ section (160-acre tract), ¼-¼ section (40-acre tract), and ¼-¼-¼ section (10-acre tract); and serial number of the spring within the tract. Quarter sections are lettered A, B, C, and D in counterclockwise order from the northeast quarter of each section (fig. 1). Within the quarter sections, 40-acre and 10-acre tracts are lettered in the same manner. Spring 8N-17E-32BCA1S is in the NE¼SW¼NW¼, sec. 32, T. 8 N., R. 17 E., and is the first spring inventoried in that tract.

REGIONAL GEOLOGIC AND HYDROLOGIC SETTING

The study area encompasses most of the surface exposures of the Idaho batholith and adjacent areas in central Idaho. The boundary of the area follows the drainage divides of the Boise, Payette, Salmon, and Lochsa Rivers (pl. 1). The terrain is characterized by rugged, forested mountains and steep, narrow river valleys. Some of the prominent peaks are more than 12,000 ft above sea level. The area is drained by the Salmon, Clearwater, Payette, and Boise Rivers, which are major tributaries to the Snake River.

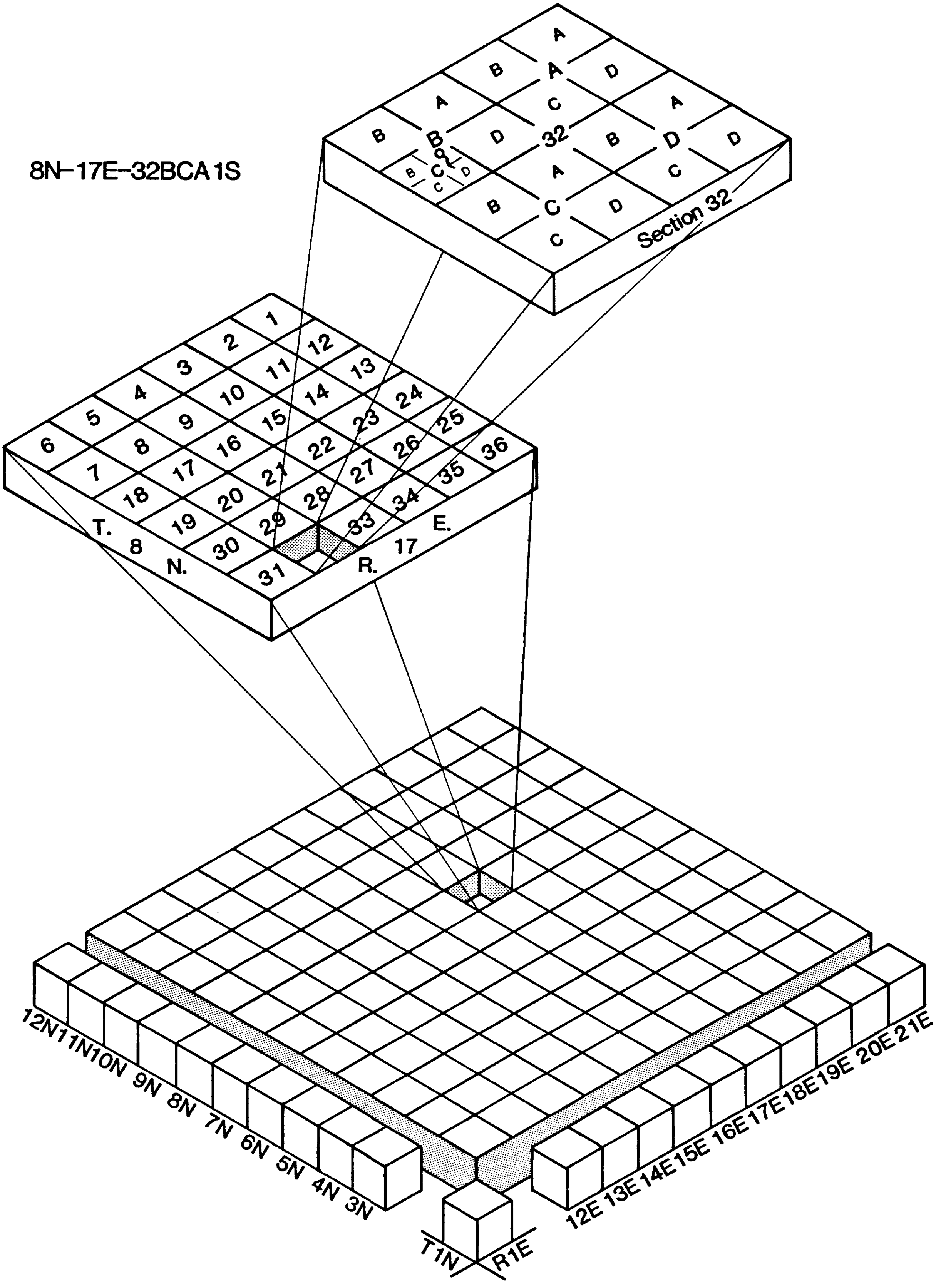


Figure 1.--Spring-numbering system.

Most thermal springs in the Idaho batholith and adjacent areas are remote and, therefore, are virtually unused except for makeshift bathing facilities. However, some thermal springs have been developed for recreational use and limited space heating. Growths of thermophyllic green and brown algae are common below vents of some thermal springs, particularly in low-temperature waters (less than 50 °C). Mineral deposits are noticeably absent near most springs. Spring waters are generally odorless and contain low concentrations of dissolved constituents.

Geology and Topography

Igneous, sedimentary, and metamorphic rocks crop out in the study area. Thermal springs generally issue directly from these rocks, although several issue from overlying surficial deposits of alluvium or residuum. For purposes of this report, geologic units in the Idaho batholith and adjacent areas are divided into: (1) Precambrian metamorphic and sedimentary rocks, (2) Paleozoic sedimentary rocks, (3) Triassic metavolcanic rocks, (4) Cretaceous metamorphosed granitic rocks, (5) Cretaceous intrusive granitic rocks, (6) Eocene intrusive rocks, (7) Tertiary volcanic rocks, (8) Quaternary and Tertiary sedimentary rocks, and (9) Quaternary sedimentary rocks. Areal distribution and descriptions of these rocks are shown on plate 2.

Cretaceous granitic rocks of the Idaho batholith predominate throughout the central part of the study area. Triassic metavolcanic rocks and Tertiary volcanic rocks are common along the western margin, and Precambrian metamorphic and sedimentary rocks, Paleozoic sedimentary rocks, and Tertiary volcanic rocks predominate along the eastern margin.

Borah Peak, on the southern boundary of the study area, is Idaho's highest mountain at 12,662 ft. Valleys on the western and eastern margins are intermontane basins surrounded by high mountain peaks and are typical of northern Rocky Mountain basins.

Hydrology

Climate

Climate in the Idaho batholith and adjacent areas ranges from semiarid in the valley lowlands to subhumid in the mountains. The large variation is caused primarily by topographic relief and prevailing wind patterns. Heavy snowfall in the mountains is the major source of water to rivers in the study area. Mean annual precipitation in the study area ranges from less than 8 in. in the eastern valley lowlands near Challis and Salmon to more than 50 in. in the central mountains (Thomas and others, 1963, pl. 2).

Mean annual temperature data recorded by the National Weather Service are available for a number of stations in the study area. Most of these stations are near settlements along the major rivers and, therefore, are not representative of the higher altitudes. Recorded mean annual air temperatures range from 54°F (12.2 °C) at Riggins to 35°F (1.6 °C) at Obsidian.

Surface Water

Drainage from the Idaho batholith and adjacent areas is by the Salmon, Boise, and Payette Rivers and several major tributaries of the Clearwater River. These rivers are tributaries of the Snake River (pl. 1).

Drainage from the Salmon River basin is mainly by the Yankee, East, North, South, East Fork of the South Fork, and Middle Forks of the Salmon River; Pahsimeroi, Lemhi, Little Salmon, and Secesh Rivers; and Valley, Panther, and Johnson Creeks. Major tributaries to the Salmon River flow generally northward or northwestward. The Salmon River heads in the south-central part of the basin and flows generally along the southern, eastern, and northern margins to Riggins. The Salmon River drains about 13,000 mi^2 within the study area.

Drainage from the Boise River basin is mainly by the North, Middle, and South Forks of the Boise River. Larger tributaries to the Boise River system include Grimes, Mores, and Big and Little Smoky Creeks and Crooked River. The North and Middle Forks of the Boise River flow generally southwestward to their confluence at Arrowrock Reservoir and continue in this direction below the dam to Lucky Peak Reservoir. The South Fork of the Boise River flows generally westward and joins the North and Middle Forks at Arrowrock Reservoir. The Boise River drains an area of about 2,680 mi^2 above Lucky Peak Reservoir.

Drainage from the Payette River basin is mainly by the North, Middle, and South Forks of the Payette River, Squaw Creek, and Deadwood River. All but the South Fork flow generally southward. The South Fork flows westward to near Crouch, where it is joined by the Middle Fork to form the Payette River. From Crouch, the Payette River flows southward to Horseshoe Bend, then westward to the Snake River. The Payette River drains about 3,300 mi^2 within the study area.

Drainage from the Clearwater River basin is mainly by the South Fork Clearwater, Lochsa, and Selway Rivers. The South Fork Clearwater River flows westward to near Grangeville, then northward to its confluence with the Clearwater River near Kooskia. These rivers drain about 4,200 mi^2 within the study area.

GENERAL DESCRIPTION OF THERMAL SPRINGS

Structural Implications and Probable Flow Systems

Thermal springs in the Idaho batholith and adjacent areas are produced by deeply circulating hot-water systems that have developed in fault-controlled conduits. Springs issuing from granitic rocks appear in most instances to be associated with major regional fault structures, as demonstrated by their areal occurrence and alignment along major rivers (pl. 1). Springs issuing from other rocks are randomly scattered and probably are associated with local faulting.

Large discharges of high-temperature water indicate that many thermal springs in the area have developed efficient flow conduits. The yield and water temperatures of thermal springs may be related, in part, to the degree of fracturing of the rocks. The more fractured rocks have higher secondary hydraulic conductivities, which permit circulation of larger volumes of water to greater depths and more rapid movement of heated water to the surface. Rapid movement of the water to the surface minimizes heat loss to the surrounding rock by conduction and, frequently, water temperature at the surface is within a few degrees of the estimated reservoir temperature.

Thermal waters in the Idaho batholith and adjacent areas are of meteoric origin; that is, derived from precipitation (discussed in the section, "Deuterium and Oxygen-18"). Exact locations of the recharge areas are unknown. However, the altitude of the recharge areas must be high enough to provide hydraulic head for spring flows. Flow in the spring is dependent on several factors, summarized by Bedinger and others (1979): (1) Transmissivity of the aquifer, or conduit; (2) areal extent of the recharge area; (3) distance of the recharge area from the spring; (4) depth of flow (circulation) of water en route to the spring; and (5) temperature of water in the aquifer. The first three considerations are usually unknown. Minimum depths of circulation can be estimated on the basis of heat-flow values, thermal conductivity of the rock, and estimated reservoir temperature. Reservoir temperature can be estimated using chemical geothermometers. Flow in thermal springs is enhanced to some degree by a density difference between cold water in the recharge area and heated water in the conduit system. Density of water above 4 °C decreases with temperature; cold water in the recharge area, being heavier than the heated water in the rest of the system, supplies an additional driving source. White (1968) referred to this phenomenon as "thermoartesian" head, which is added to hydraulic head resulting from the altitude difference between the spring and recharge area. In a thermal spring system, it is possible for the recharge area to be at or even below the altitude of the spring discharge area. In the Idaho batholith and adjacent areas, this condition was not observed, and recharge to all thermal springs is thought to be from altitudes well above the spring discharge area.

Spring Discharge

Measurements or estimates of discharge for 129 springs are available in reports on the Idaho batholith and adjacent areas (see "Previous Studies" section). Individual spring flows are highly variable; values range from less than 1 to 2,710 gal/min. To determine seasonal variations in discharge, a V-notch weir and bubbler gage were installed in the discharge channel of spring 12N-5E-22BBC1S. The gage operated from August 1978 through January 1980. Although some discharge fluctuations (86-116 gal/min) were observed, no seasonal trend was apparent.

Annual thermal spring discharge in the Idaho batholith and adjacent areas was computed using reported discharges. Measurements or estimates of discharge were not available for all springs; however, annual thermal spring discharge is estimated to be at least 27,000 acre-ft. Average annual surface-water runoff from the study area is about 18,300,000 acre-ft. Therefore, any attempt to relate thermal water discharge to a hydrologic budget of the area is impractical.

Temperature

Thermal springs were defined by Gilbert (1875, p. 147-149) as springs in which water temperatures exceeded the mean annual air temperature by 8.5 °C. For purposes of this report, only those springs in the Idaho batholith and adjacent areas whose water temperatures are 20.0 °C or greater were considered thermal. Measured water temperatures ranged from 20.5 °C to 94.0 °C. No known thermal springs in the study area boil at land surface, although Big Creek Hot Springs, which issues at 94.0 °C at an altitude of 4,280 ft, is within about 2 °C of boiling.

To determine seasonal variation in water temperatures, equipment to continuously monitor water temperature was installed at spring 12N-5E-22BBC1S. Water temperature varied about 1 °C between June and November 1979.

Heat Discharge

Heat from the Idaho batholith and adjacent areas is removed convectively by hot water, which discharges from numerous thermal springs. The convective heat flux is calculated as the product of the volume rate of discharge and the enthalpy (heat content) of the water in excess of the mean annual air temperature, or:

$$H = M(h_r - h_o) \tag{1}$$

where

H = heat loss by convection, in calories per second;

M = mass discharge;

h_r = enthalpy of the hot water; and

h_o = enthalpy of cold recharge water (2.0 °C to 12.0 °C, depending on spring location).

To determine total convective heat flux, the volume of discharge for each thermal spring was converted to an instantaneous flow rate and M was calculated. In the convective heat flow equation, h_o is approximately equal to the mean annual air temperature. Subsequent percolation of spring water after discharge is considered negligible, and no heat is returned to the system. Heat flux from the 129 sampled thermal springs in the Idaho batholith and adjacent areas was estimated to be 5.0×10^7 cal/s.

GEOCHEMISTRY OF THERMAL SPRINGS

Chemical and isotopic analyses of water from thermal and nonthermal springs aid in describing the thermal spring system(s). Specifically, these analyses distinguish different water types and may be used to determine temperatures attained at depth, origin or source of the water, and residence time.

Cation and anion distributions and dissolved-solids concentrations of the thermal springs are used to delineate different water types. Sodium-potassium-calcium, silica, and sulfate-water isotope geothermometers are used to estimate reservoir temperatures. Tritium concentrations are used to assess possible mixing of thermal waters with shallow ground waters and relative age of cold spring waters. The stable isotopes, D (deuterium) and ^{18}O (oxygen-18), are used to characterize and indicate the origin and mixing patterns of individual waters. The carbon isotopes, ^{12}C, ^{13}C, and ^{14}C, are used to determine residence times. Constraints and basic assumptions inherent in the above techniques are explained in detail in following sections.

Chemical Constituents and Types of Water

Major Constitutents

Thermal springs in the Idaho batholith and adjacent areas discharge relatively dilute water (dissolved-solids concentrations range from 103 to 839 mg/L) which is, with two exceptions, a sodium mixed-anion type.

In 104 of the 112 sampled thermal springs, pH values were 8.0 or greater. Calculations made using the computer program SOLMNEQ (Kharaka and Barnes, 1973) indicate that the majority of these waters are at or near saturation with respect to calcite. The relatively high pH of these waters, therefore, seems to control the concentrations of the alkaline-earth metals. Chemically similar thermal waters discharging from granitic rocks of the Sierra Nevada batholith were studied by Feth and others (1964). They concluded that this type of water is a result of deep circulation of meteoric water and that chemical aggressiveness of water is increased at higher temperatures and the H^+ ion content is depleted by hydrolysis of silicate minerals, which raises the pH. Sodium content continues to increase as H^+ ions displace cations from the silicate framework. Calcium is removed by precipitation, either as a carbonate or an aluminosilicate, and chloride and fluoride increase as the result of attack by the water on minerals containing these constituents, either in fluid inclusions or as substitutes for elements in the lattices of minerals such as biotite. Magnesium concentrations may be held at low levels by equilibrium relations with dolomite; or more likely, potassium and magnesium may be incorporated in the clay minerals formed during alteration of the igneous rocks. On the basis of general chemical similarities between thermal waters in the Idaho batholith and adjacent areas and those described by Feth and others (1964), the foregoing hypothesis best describes the chemical pattern of springs discharging from granitic rocks.

However, some chemical differences between the Idaho waters and those described by Feth and others (1964) cannot be explained by the foregoing hypothesis. Chloride and fluoride concentrations in the Idaho waters range from 1.2 to 61 mg/L and 0.4 to 24 mg/L--lower chloride concentrations and higher fluoride concentrations than in the waters of the Sierra Nevada batholith. Apparent differences in these chemical constituents may be explained by the relative abundance of these elements in rocks.

Sulfate concentrations in waters from several springs range from 180 to 320 mg/L and cannot be explained by the foregoing hypothesis. These springs are closely grouped along the western margin of the study area, which suggests a common source for the observed sulfate concentrations. However, the source of sulfate concentrations in these waters remains unexplained.

Minor Constituents

Dissolved minor constituents are sometimes useful qualitatively to determine the environment of thermal waters and to define common reservoirs. Thermal waters discharging from the Idaho batholith and adjacent areas were analyzed for arsenic, boron, lithium, and mercury. Although these constituents were present in the waters, their low concentrations were highly variable and exhibited no distinguishable pattern. Ranges of observed concentrations follow:

Element	Range
Arsenic	Less than analytical detection limit to 48 µg/L
Boron	Less than analytical detection limit to 1,700 µg/L
Lithium	Less than analytical detection limit to 570 µg/L
Mercury	Less than analytical detection limit to 1.4 µg/L

These low, variable concentrations neither permit assessment of the role of minor constituents as an indicator of chemical similarities among the thermal waters, nor suggest a common reservoir or rock environment.

Geothermometry

Chemical Geothermometers

The two most widely used chemical geothermometers are the silica (Fournier and Rowe, 1966) and the Na-K-Ca (sodium-potassium-calcium; Fournier and Truesdell, 1973).

Reservoir temperatures estimated using the silica and Na-K-Ca geothermometers are valid only for hot-water systems and only if the following basic assumptions are met: (1) The chemical reactions at depth are temperature dependent, (2) an adequate supply of chemical constituents used for the thermometry is present in the aquifer, (3) chemical equilibrium is established between the hot water and the specific aquifer minerals, (4) reequilibrium of the chemical composition of hot water as it rises to the surface is negligible, and (5) hot water rises rapidly to the surface with no dilution or mixing of hot and cold waters (White, 1970, p. 25).

Reservoir temperatures in the Idaho batholith and adjacent areas were estimated using the silica and Na-K-Ca geothermometers. Silica equilibrium with both quartz and chalcedony was considered assuming conductive heat loss. Corrections for the adverse effect of magnesium on the Na-K-Ca geothermometer were applied using the method of Fournier and Potter (1979), where reservoir temperatures estimated using the Na-K-Ca geothermometer were greater than 70 °C and magnesium concentrations exceeded 1 mg/L.

Dissolved silica (SiO_2) reported in chemical analyses is present as silicic acid (H_4SiO_4) and various dissociated species (particularly $H_3SiO_4^-$). In alkaline water, hydroxide (OH^-) reacts with the silicic acid to reduce the proportion of silicic acid to total dissolved silica:

$$\begin{array}{ccccccc} \text{Silicic} & & & & \text{Dissociated} & & \\ \text{acid} & + & \text{Hydroxide} & = & \text{silicic acid} & + & \text{Water} \\ H_4SiO_4 & + & OH^- & = & H_3SiO_4^- & + & H_2O \end{array} \tag{2}$$

The pH values of thermal waters sampled ranged from 7.1 to 9.9. For waters having pH values greater than about 8.2, the total concentration of dissolved silica measured in the laboratory ($H_4SiO_4 + H_3SiO_4^-$) was reduced by the calculated concentration of $H_3SiO_4^-$ to obtain a better estimate of the thermal reservoir temperature (Brook and others, 1979, p. 21-22).

Estimated reservoir temperatures for 112 thermal springs in the study area are shown on plate 1. In general, most of the estimated temperatures using the Na-K-Ca and $H_3SiO_4^-$-corrected silica geothermometers are in good agreement for individual springs. Estimated reservoir temperatures also are in good agreement among springs aligned or grouped along several drainages, which indicates a possible common reservoir or, at least, similar circulation patterns.

Estimated reservoir temperatures for thermal springs in the study area are generally less than 100 °C. However, estimated reservoir temperatures for springs 10N-4E-33CBD1S, 9N-3E-25BAC1S, and 7N-1E-8DDA1S, along the lower Payette River near the southwestern margin of the study area, range from 132 °C to 152 °C. Estimated reservoir temperatures for Big Creek Hot Springs, Owl Creek Hot Springs, and spring 20N-24E-34CCC1S, near Salmon, range from 123 °C to 184 °C.

Isotopic Geothermometer

The sulfate-water isotope geothermometer (McKenzie and Truesdell, 1977) is based on the temperature-dependent equilibrium of the isotopes of oxygen ($^{16}O/^{18}O$) between water and dissolved sulfate in the geothermal reservoir. The rate of equilibrium is both temperature and pH dependent. The rate of equilibration is accelerated in acid solutions and excessively slow in neutral or alkaline solutions below 200 °C. Nehring and others (1979) stated that equilibrium required about 2 years at 300 °C and 110 years at 25 °C in near-neutral waters. The addition or loss of sulfate or mixing of water after the sulfate-water isotopic equilibrium has been established in the reservoir may result in erroneous temperature estimates. Common interferences with the sulfate-water geothermometer are the addition of near-surface sulfate from the oxidation of H_2S to H_2SO_4 by sulfur-oxidizing bacteria, and dilution of the thermal water by addition of ground water containing different ^{18}O.

Reservoir temperatures for 19 thermal springs estimated using the sulfate-water geothermometer are given in table 1, along with the $H_3SiO_4^-$-corrected silica and Na-K-Ca estimated temperatures. In general, most of the estimated temperatures are in good agreement. The sulfate-water geothermometer also substantiated the silica and Na-K-Ca reservoir temperatures near 150 °C for springs 9N-3E-25BAC1S and 7N-1E-8DDA1S.

However, several large discrepancies exist between the sulfate-water and chemical geothermometers. Most noticeable are estimated temperatures for Big Creek Hot Springs and springs 20N-24E-34CCC1S and 14N-6E-11BDA1S. Big Creek Hot Springs and spring 14N-6E-11BDA1S have readily detectable H_2S odors, which may indicate the possible addition of near-surface sulfate from the oxidation of H_2S to H_2SO_4. No H_2S odor was detected at spring 20N-24E-34CCC1S. However, the thermal water in this spring, located along a small stream in an upland draw, fed in part from ground-water discharge, could be diluted by nonthermal ground water.

Table 1.--Comparison of estimated reservoir temperatures using sulfate-water and chemical geothermometers

Spring name or No.	Water temperature (°C)	Estimated reservoir temperatures		
		Silica $H_3SiO_4^-$ corrected	Na-K-Ca	Sulfate water
Big Creek Hot Springs	94.0	159	184	127
20N-24E-34CCC1S	63.5	133	175	108
Kwiskwis Hot Springs	68.0	93	102	103
17N-11E-16ACB1S	87.0	102	146	95
16N- 4E-35CCB1S	50.0	69	71	71
16N-12E-17DAD1S	62.0	86	103	88
Hot Creek Hot Springs	60.0	86	80	106
14N- 6E-11BDA1S	88.5	97	49	65
13N- 4E-31CAB1S	71.0	91	117	75
13N- 6E-29DAB1S	53.0	76	82	71
12N- 5E-22BBC1S	86.0	101	100	[1]89
Sunbeam Hot Springs	77.5	94	111	74
10N-10E-31BCC1S	84.0	81	145	[1]94
9N- 3E-25BAC1S	79.5	143	144	139
9N- 8E-32CBA1S	64.0	86	79	88
8N- 5E-10BDD1S	54.5	58	70	76
7N- 1E- 8DDA1S	64.0	138	152	161
5N- 7E-24BDD1S	76.0	82	91	[1]96
3N-14E-28CAD1S	87.0	76	98	90

[1] Nehring and others, 1979

Isotopes

Tritium

Tritium (^{3}H) is a radioactive isotope of hydrogen produced naturally in small quantities in the upper atmosphere during bombardment by subatomic particles from outer space and is introduced into the water cycle dissolved in rain and snow. Having a half-life of about 12.4 years, tritium can be used to determine how long a particular water has been isolated from the atmosphere. The transit or residence time of spring water can be determined from the concentration of tritium if the amount of tritium in the recharge water and the nature of the subsurface flow regime are known.

Although concentrations of tritium in precipitation vary seasonally and geographically, prior to extensive atmospheric thermonuclear testing between 1953 and 1963, tritium in the atmosphere ranged from 4 to 25 tritium units (Nir and others, 1966). One TU (tritium unit) equals a $^{3}H/H$ ratio of about 10^{-18}, or about 3.2 picocuries per liter. By 1963, worldwide tritium levels in precipitation had increased several orders of magnitude and were 7,000 TU in the vicinity of Yellowstone National Park (Pearson and Truesdell, 1978, p. 327). Since the end of atmospheric nuclear tests, tritium levels in precipitation have declined and presently (1982) average about 50 TU.

Tritium content in a ground-water system is a function of tritium content in the recharge water, residence time, and nature of flow in the system. Two basic types of flow models were discussed in detail by Nir (1964): (1) The piston-flow model, which has parallel flow lines of constant and equal velocity so that a water sample taken at some point would include only water originating at the point of recharge, and (2) the well-mixed reservoir model, where it is assumed that the recharge water is continually and instantly mixing throughout the entire system.

Concentrations of tritium in water samples from 13 nonthermal and 13 thermal springs were determined by the U.S. Geological Survey laboratory in Reston, Va. Samples were predistilled and enriched by electrolysis, and the enriched samples were counted by a gas proportional counter. All samples were counted for tritium decay to the collection date, using a half-life of 12.361 years. Concentrations of tritium in the nonthermal springs ranged from 23.7 ±3.3 to 75.2 ±3.6 TU and averaged 48 TU. Concentrations of tritium in the thermal springs ranged from 0 ±0.1 to 4.1 ±0.5 TU and averaged 0.8 TU.

If waters move by piston flow, the T (tritium level in a water sample taken at any point within the system) will be related to T_o (tritium concentration in the water during recharge) and t (transit time or age of the sample) by

$$T = T_o e^{-\gamma t}, \qquad (3)$$

where the decay constant γ = natural logarithm 2/half-life of tritium.

Assuming piston flow, an upper limit of 50 TU, and the average tritium content for all nonthermal springs of 48 TU, the average age of the nonthermal spring water would be about 0.75 years. This would suggest that the sampled springs are representative of present-day precipitation (recharge) and that the majority of sampled springs discharge water derived from the preceding winter's snowfall.

Considering an analytical error of ±0.4 TU, tritium concentrations in seven of the sampled thermal springs are near zero. Assuming piston flow for these waters and a prethermonuclear testing tritium level of 10 TU, the age of these waters is at least 80 years, near or exceeding the maximum tritium-age dating limit. Tritium content of the other six springs ranges from 0.6 to 4.1 TU; the maximum value was from Horse Creek Hot Springs. Concentration of NO_2 + NO_3 as N in Horse Creek Hot Springs is 0.83 mg/L, and three of the five remaining thermal springs with detectable tritium also have higher concentrations of NO_2 + NO_3 as N than the majority of sampled thermal springs. Normally, concentrations of NO_2 + NO_3 as N in thermal springs are very low, whereas in shallow ground water, concentrations are typically greater than 0.5 mg/L and commonly exceed several milligrams per liter in agricultural or urban areas. On the basis of the above discussion, the low level of tritium in sampled thermal springs probably is the result of mixing with shallow ground water and is not representative of tritium content in the thermal reservoirs.

Deuterium and Oxygen-18

Principal stable molecular species in water are $H_2{}^{16}O$, $H_2{}^{17}O$, $H_2{}^{18}O$, and $HD{}^{16}O$ [H, hydrogen-1; D, deuterium (hydrogen-2); O, oxygen; the superscript indicates the specific isotope]. In ocean water, the proportions of these species are 10^6:2,000:420:316 (Craig, 1963); this composition is referred to as SMOW (standard mean ocean water). Precipitation, which is derived from ocean water, is depleted in ^{18}O and deuterium. The isotopic composition of precipitation from the atmosphere depends on the fraction of original water remaining in the air mass from which the rain or snow is derived. The first precipitation is richest in the heavy isotopes deuterium and oxygen-18, relative to the light isotopes hydrogen-1 and oxygen-16 (Dansgaard, 1964). Subsequent depletion of the heavy stable isotopes in precipitation generally can be correlated with distance from the ocean, latitude, and altitude (or temperature).

Stable-isotope concentrations are generally expressed in delta units (δ) and are reported in per mil ($^{o}/oo$). These units, which represent relative deviations in the heavy-isotope fraction in water, are defined as:

$$\delta = \left[\frac{R\ \text{sample} - R\ \text{standard}}{R\ \text{standard}}\right] \times 1{,}000, \qquad (4)$$

where

R sample = ratio of isotopic concentrations ($^{18}O/^{16}O$ or D/H) for the sample, and

R standard = ratio of isotopic concentrations for the standard, which is referred to as SMOW.

Thus, as the heavy isotopes become depleted, the delta values become more negative.

A worldwide study of freshwater samples showed that isotopic compositions of cold meteoric waters were related by the equation $\delta D = 8\,\delta^{18}O + 10$ (Craig, 1963). This straight line, commonly referred to as the meteoric water line, is shown in figure 2; the slope and intercept of the line may vary regionally. Values for water affected by extensive nonequilibrium evaporation, as in inland basins, lie off the meteoric line. However, at ordinary air temperatures, evaporated surface waters are connected approximately to the original precipitation composition of $\delta^{18}O_o$ and δD_o by a line having the equation $\delta D = 5\,(\delta^{18}O - \delta^{18}O_o) + D_o$ (Ellis and Mahon, 1977, p. 75).

During passage through an aquifer, thermal and nonthermal ground water retains the deuterium proportions that are characteristic of precipitation in the recharge area. Deuterium is not altered because rock minerals contain little hydrogen, and the water-to-rock ratios of geothermal systems are seldom so low that the hydrogen of rock minerals is a significant fraction of the total hydrogen (Truesdell

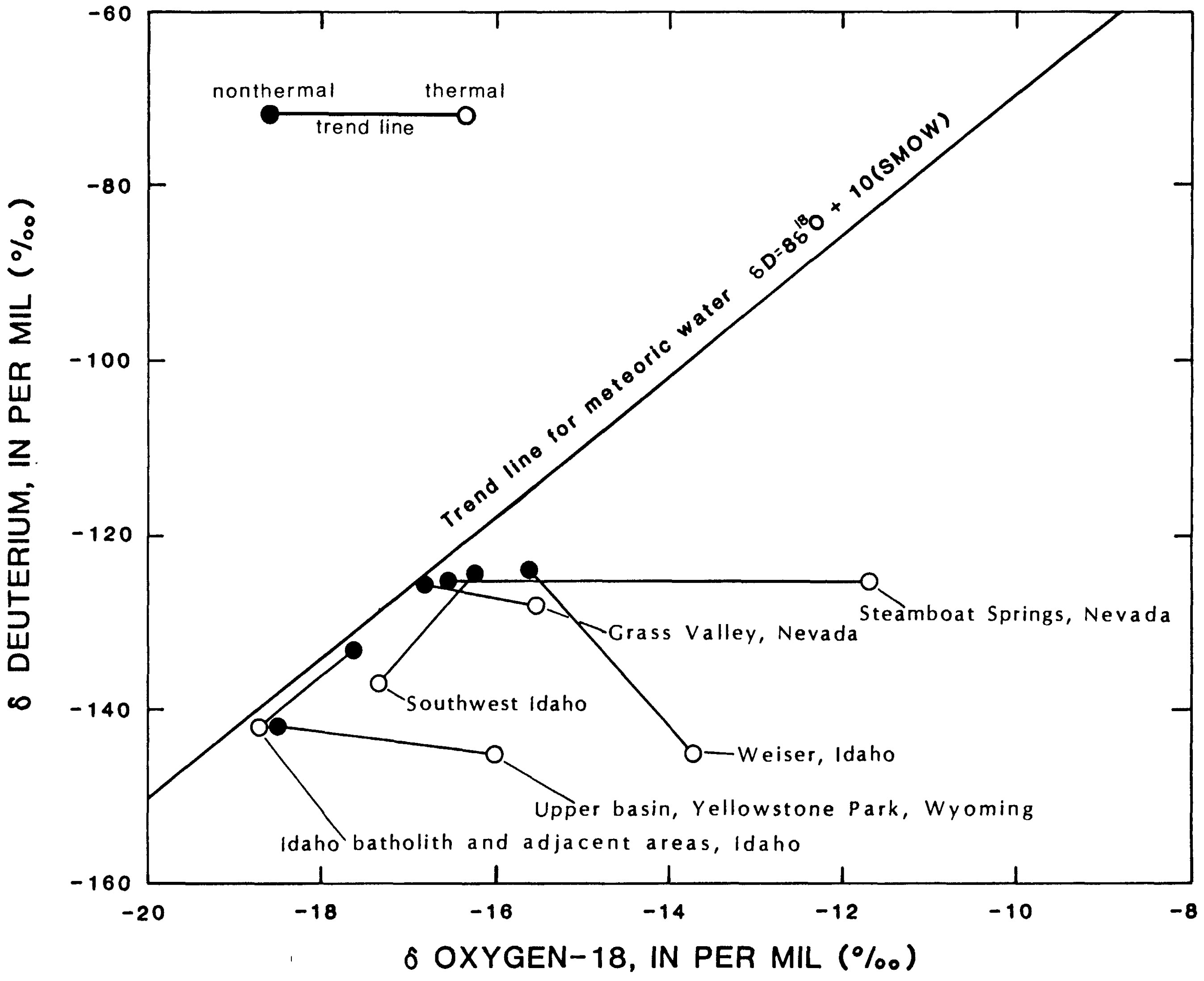


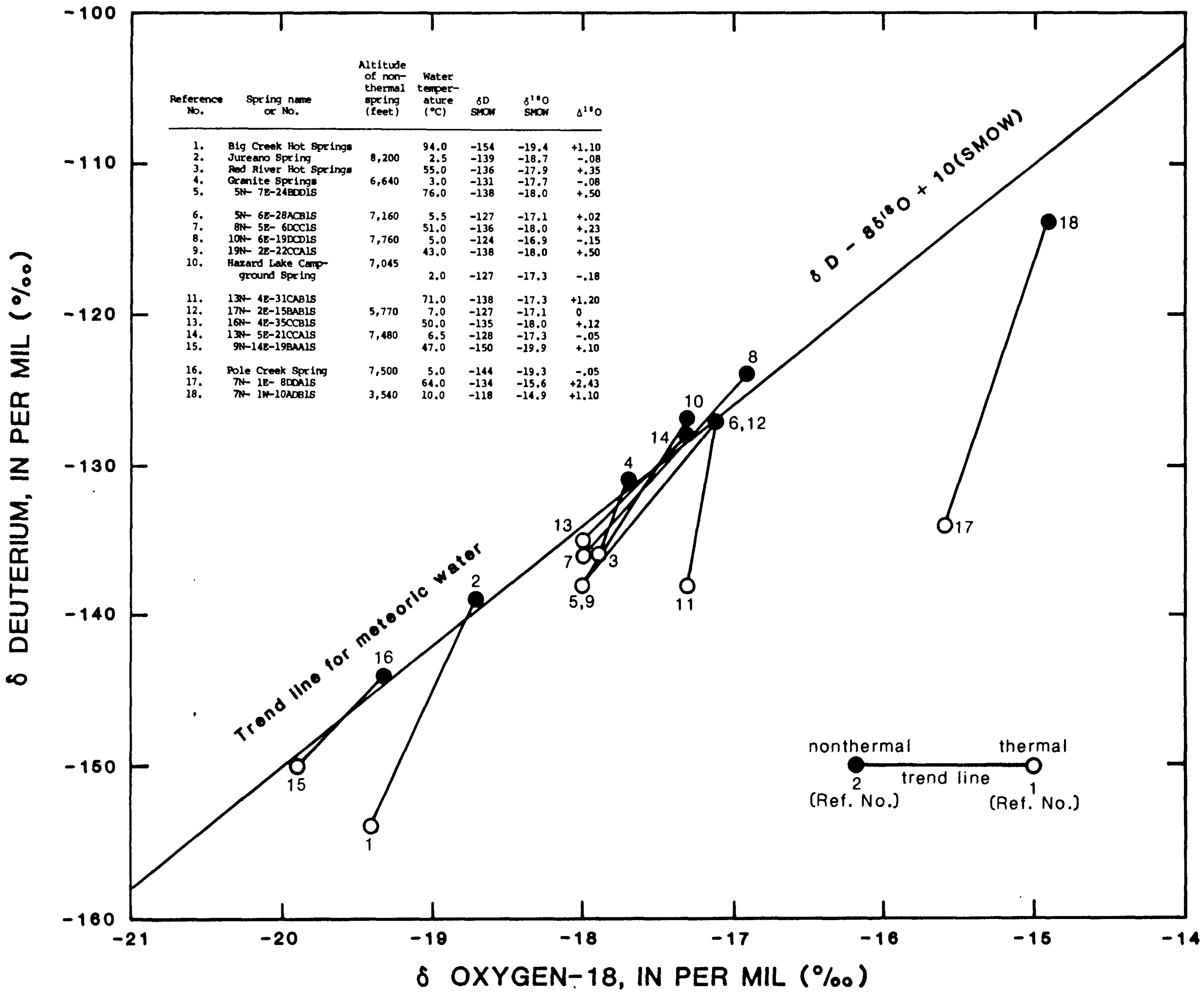
Figure 2.--Comparison of isotopic composition of nonthermal and thermal waters in Idaho, Wyoming, and Nevada.

and Hulston, 1980). The ^{18}O content in thermal water is usually enriched (less negative $\delta^{18}O$) to varying degrees during circulation within the system, owing to reactions with the more abundant ^{18}O of the host rock. Isotope-exchange reactions between water and rock occur slowly at temperatures below 150 °C (Hobba and others, 1979, p. E25) and generally are detected only in systems having reservoir temperatures of 100 °C or greater (Ellis and Mahon, 1977, p. 155-158). For a more complete discussion of stable-isotope geochemistry, refer to Gat (1971) or Ellis and Mahon (1977, p. 153-159).

Comparison of the isotopic composition of waters from the Idaho batholith and adjacent areas and other geothermal systems of Idaho, Wyoming, and Nevada is shown in figure 2. Waters in these systems are predominantly or entirely meteoric in origin. With the exception of the Idaho waters, δD values are similar to those of local recharge waters; ^{18}O values are significantly enriched relative to the local recharge water. This enrichment probably results from isotope-exchange reactions with aquifer minerals. The significant depletion of δD values relative to $\delta^{18}O$ values for water in southwestern Idaho and the Idaho batholith and adjacent areas suggests that most of the thermal water is not derived from current precipitation. Young and Lewis (1982a, p. J12) concluded that thermal waters in southwestern Idaho were recharged in the past when the mean annual air temperature was about 6.3 °F (3.5 °C) colder than present. Carbon-14 data for southwestern Idaho geothermal systems indicated an apparent age of 18,000-25,000 years (Young and Lewis, 1982a, p. J17).

The possibility of recharge from recent precipitation at high altitudes for thermal springs in the study area also should be considered. Figure 3 shows a comparison of isotopic compositions of water from selected thermal springs and adjacent high-altitude nonthermal springs. As shown in figure 3, δD values for the thermal waters are depleted relative to the δD values for adjacent nonthermal waters. Big Creek Hot Springs, which issues at an altitude of 4,400 ft (ref. no. 1, fig. 3), has a δD value of -154 $^{o}/oo$, whereas nearby Jureano Spring (ref. no. 2, fig. 3), which issues at an altitude of 8,200 ft, 7.5 mi to the southeast, has a δD value of -139 $^{o}/oo$. Similar relations between the δD concentrations of nearby hot and cold springs are apparent for all sampled waters.

Comparison of stable-isotope concentrations to mean annual temperatures at springs in the Appalachian Mountains (Hobba and others, 1979, p. E25) indicates a depletion rate for D of about -4 $^{o}/oo$ per degree Celsius, or about -8 $^{o}/oo$ per 1,000 ft of altitude, assuming a normal lapse rate of 2 °C per 1,000 ft. The depletion of δD between the thermal (Big Creek Hot Springs) and nonthermal water (Jureano



Reference No.	Spring name or No.	Altitude of non-thermal spring (feet)	Water temper-ature (°C)	δD SMOW	$\delta^{18}O$ SMOW	$\Delta^{18}O$
1.	Big Creek Hot Springs		94.0	-154	-19.4	+1.10
2.	Jureano Spring	8,200	2.5	-139	-18.7	-.08
3.	Red River Hot Springs		55.0	-136	-17.9	+.35
4.	Granite Springs	6,640	3.0	-131	-17.7	-.08
5.	5N- 7E-24BDD1S		76.0	-138	-18.0	+.50
6.	5N- 6E-28ACB1S	7,160	5.5	-127	-17.1	+.02
7.	8N- 5E- 6DCC1S		51.0	-136	-18.0	+.23
8.	10N- 6E-19DCD1S	7,760	5.0	-124	-16.9	-.15
9.	19N- 2E-22CCA1S		43.0	-138	-18.0	+.50
10.	Hazard Lake Campground Spring	7,045	2.0	-127	-17.3	-.18
11.	13N- 4E-31CAB1S		71.0	-138	-17.3	+1.20
12.	17N- 2E-15BAB1S	5,770	7.0	-127	-17.1	0
13.	16N- 4E-35CCB1S		50.0	-135	-18.0	+.12
14.	13N- 5E-21CCA1S	7,480	6.5	-128	-17.3	-.05
15.	9N-14E-19BAA1S		47.0	-150	-19.9	+.10
16.	Pole Creek Spring	7,500	5.0	-144	-19.3	-.05
17.	7N- 1E- 8DDA1S		64.0	-134	-15.6	+2.43
18.	7N- 1W-10ADB1S	3,540	10.0	-118	-14.9	+1.10

Figure 3.--Relation between concentrations of deuterium and oxygen-18 for selected thermal springs and adjacent high-altitude nonthermal springs.

Spring) is about -15 $^{0}/oo$, which indicates water recharging the thermal system originated at an altitude above 10,000 ft. Mt. McGuire is 10,082 ft in altitude but is 15 mi from Big Creek Hot Springs so is not a probable source of recharge to the thermal system. On the basis of available data, thermal springs in the Idaho batholith, like those in southwestern Idaho, probably recharged when the climate was cooler than at present.

The change in oxygen-isotope ratios, or the oxygen shift, is dependent on temperature, rate of flow, isotopic composition of the rock, and relative quantities of rock and fluid. Generally, a large oxygen shift is expected when small quantities of fluid move slowly through a large amount of aquifer material, and a small oxygen shift is expected when large quantities of fluid move more rapidly through a small amount of aquifer material. The oxygen shift for thermal waters in the study area, except for six springs, is less than 1 $^{0}/oo$. This would normally indicate large quantities of fluid moving rapidly through a small volume of aquifer material. More likely, however, the small oxygen shift indicates that little ^{18}O exchange has occurred, owing to the low temperatures of the water.

Carbon

The following introductory discussion of isotopic chemistry of carbon was taken in part from Bedinger and others (1979, p. C20-C23). Carbon has three naturally occurring isotopes with masses of 12, 13, and 14. Carbon of masses 12 (^{12}C) and 13 (^{13}C) are stable isotopes, whereas carbon of mass 14 (^{14}C) is radioactive. In any chemical reaction such as the dissolution of CO_2 to form H_2CO_3 and HCO_3^-, or the formation of carbohydrates from CO_2 by plants during photosynthesis, the lighter isotope (^{12}C) reacts faster than the heavier isotope (^{13}C). The isotope ratios of reactant and product species are different, thereby producing isotopic fractionation. Carbon-isotope ratios are measured relative to an arbitrary standard, a calcite of marine origin known as PDB, and are expressed in the δ notation.

Values of $\delta^{13}C$ for samples from a single type of carbon-bearing material are relatively constant, but different materials may have different $\delta^{13}C$ values, thus making $\delta^{13}C$ measurements valuable in determining the sources of the dissolved carbon (Pearson and Hanshaw, 1970).

The naturally occurring radioisotope of carbon (^{14}C) has a half-life of about 5,730 years. Natural ^{14}C is formed by cosmic-ray reactions with nitrogen in the upper atmosphere and is found at virtually a constant concentration in atmospheric CO_2. Carbon-14 concentrations are measured and reported as percentages relative to a standard whose ^{14}C concentration is that of an ideal plant, grown in 1950, in the absence of any atmospheric CO_2 of industrial origin. Carbon-14 also is produced as a byproduct of nuclear explosions, and in the last 25 years, the ^{14}C level in the atmosphere has been considerably higher than its natural level, or "100 percent modern."

Plants using atmospheric CO_2 assimilate ^{14}C. When a plant dies, the amount of ^{14}C it contains decreases because of radioactive decay, and the amount remaining, relative to that of an ideal living plant, measures the time that has elapsed since its death. This is the principle on which the radiocarbon-dating method is based.

The equation for radioactive decay is:

$$\frac{Asm}{Astd} = \exp(-\lambda t), \qquad (5)$$

where Asm/Astd is the ratio of sample to standard activity, λ is the constant of the isotope (1/8,266 for ^{14}C), and t is the age in years. (The half-life of an isotope is the value of t when Asm/Astd = 0.5; that is, $t_{\frac{1}{2}} = \ln 2/\lambda$ = 5,730 years for ^{14}C.) For calculation, the decay equation is often written using natural logarithms (ln) as:

$$t = -8{,}266 \ln \left(\frac{Asm}{Astd}\right) \qquad (6)$$

As discussed, the CO_2 in the soil air is a product of plant respiration and decay. Because of its origin, soil air CO_2 has the same carbon isotopic composition as plants. Soil CO_2 dissolved by ground water will also inherit the plants' isotopic characters. If soil CO_2 were the only source of dissolved carbon, the ^{14}C concentration would reflect the age of the ground water; that is, the time elapsed since the water left the soil zone.

Ground water generally dissolves mineral carbonate, which is commonly of marine origin and has a $\delta^{13}C$ value near 0 $^{o}/oo$. Because of its great age relative to the half-life of ^{14}C, mineral carbonate has no ^{14}C. Dissolution of ^{14}C-free carbonate in ground water gives the water a false apparent old age. However, if the relative amounts of soil-derived and mineral carbon are known, this effect can be corrected.

For example, in water of an age of zero years with a total dissolved carbon of 4.00 mmol/L (millimoles per liter), of which 3.00 mmol/L is from soil CO_2 (^{14}C = 100 percent) and 1.00 mmol/L from carbonate minerals (^{14}C = 0 percent), the measured ^{14}C would be:

$$\frac{[(3.0 \times 1.0) + (1.0 \times 0.0)]}{4} = 75 \text{ percent modern,} \qquad (7)$$

and its apparent age would be:

$$t = -8{,}266 \ln (75/100) = 2{,}380 \text{ years,} \qquad (8)$$

instead of its real age, zero years.

By knowing the proportions of mineral-to-soil carbon in a ground water, it is possible to adjust the measured ^{14}C concentration to its correct value and obtain a corrected age. The adjustment factor, P, is the ratio of soil-air-derived (=plant) carbon concentrations to total dissolved-carbon concentrations:

$$P = \frac{C_{plant}}{C_{total}} \qquad (9)$$

For this example,

$$P = \frac{3.00}{4.00} = 0.75. \qquad (10)$$

The adjusted ^{14}C is

$$^{14}C_{(adjusted)} = \frac{^{14}C_{(measured)}}{P} \qquad (11)$$

and for this example:

$$^{14}C_{(adjusted)} = \frac{75 \text{ percent}}{0.75} = 100 \text{ percent,} \qquad (12)$$

corresponding to an age of zero, as specified.

Water samples from eight thermal springs in the Idaho batholith and adjacent areas and six nonthermal springs in the suspected recharge areas were selected for ^{14}C and $^{13}C/^{12}C$ analyses. Results of these analyses are given in table 2. Apparent ages for sampled thermal waters range from 10,700 years to more than 41,000 years.

Because of deep circulation of water discharging from thermal springs, a potential exists for the addition of carbon by solution of aquifer minerals. If this is the case, the measured ^{14}C values would have to be adjusted for the mineral carbonate. The stable carbon-isotopic composition of the total dissolved carbon in ground water also reflects the proportions of the source carbon. The $\delta^{13}C$ values and the total carbon in a flow system reflect the proportions of the source carbon by the following equation:

Table 2.--Carbon-isotope data

Spring name or No.	Total dissolved carbonate (millimoles per liter)	Measured $\delta^{13}C$ ($^{o}/oo$)	Calculated $(\delta^{13}C)_F$ ($^{o}/oo$)	Measured ^{14}C (percent modern)	Apparent age (years)	Adjusted ^{14}C (percent modern)	Adjusted age (years)
NONTHERMAL SPRINGS							
Jureano Spring	0.366	-19.15					
Hazard Lake Campground Spring	1.52	-23.45					
11N-20E-17AAB1S	3.58	-12.25					
10N- 6E-19DCD1S	.545	-22.70					
Pole Creek Spring	1.94	-11.15					
5N- 6E-28ACB1S	.861	-17.50					
THERMAL SPRINGS							
Big Creek Hot Springs	8.25	-5.5	-4.87	0.7	41,000	15.91	15,200
19N- 2E-22CCA1S	.637	-10.65	.83	12.3	17,300	21.41	12,700
18N- 6E- 9ADC1S	.578	-11.55	1.57	12.6	17,100	19.90	13,300
16N- 4E-35CCB1S	.950	-8.60	-1.96	10.0	19,000	25.96	11,100
Sunbeam Hot Springs	.545	-7.50		7.1	21,900		
Slate Creek Hot Springs	1.91	-8.95		27.3	10,700		
9N- 3E-25BAC1S	2.40	-8.35	-4.12	<.7	>41,000	3.08	28,800
8N- 5E- 6DCC1S	1.14	-10.65	.48	16.1	15,100	33.54	9,000

$$(\delta^{13}C)_T\ (C)_T = (\delta^{13}C)_R\ (C)_R + (\delta^{13}C)_F\ (C)_F \tag{13}$$

where,

$(\delta^{13}C)_T$ = $\delta^{13}C$ value of thermal water, in per mil;

$(C)_T$ = total carbon in the thermal water, in millimoles per liter;

$(\delta^{13}C)_R$ = $\delta^{13}C$ value of nonthermal water (recharge), in per mil;

$(C)_R$ = total carbonate in the nonthermal water, in millimoles per liter;

$(\delta^{13}C)_F$ = $\delta^{13}C$ value of mineral carbon from the flow system; and

$(C)_F$ = total carbon dissolved from minerals in the flow system.

Solving equation 13 for $(\delta^{13}C)_F$:

$$(\delta^{13}C)_F = \frac{(\delta^{13}C)_T\ (C)_T - (\delta^{13}C)_R\ (C)_R}{(C)_F} \tag{14}$$

Using values for Big Creek Hot Springs and Jureano Spring in equation 13:

$$(\delta^{13}C)_F = \frac{(-5.5\ ^{o}/oo)(8.25\ \text{mmol/L}) - (-19.2\ ^{o}/oo)(0.366\ \text{mmol/L})}{7.88\ \text{mmol/L}}$$

$$(\delta^{13}C)_F = -4.87\ ^{o}/oo. \tag{15}$$

The calculated $\delta^{13}C$ value of $-4.87\ ^{o}/oo$ is near the $\delta^{13}C$ value reported by C. T. Rightmire (TRW, Inc., written commun., 1984) for caliche in central Idaho, which ranged from $+1.10$ to $-8.12\ ^{o}/oo$ and averaged $-4.00\ ^{o}/oo$. Therefore, the carbon content in the flow system increased from 0.366 mmol/L in Jureano Spring to 8.25 mmol/L at Big Creek Hot Springs. Using equation 9,

$$P = \frac{C_{plant}}{C_{total}} \tag{16}$$

or, for this example,

$$P = \frac{0.366\ \text{mmol/L}}{8.25\ \text{mmol/L}},$$

an adjustment factor of 0.044 is calculated.

The measured ^{14}C value of Big Creek Hot Springs, therefore, must be adjusted using equation 11:

$$^{14}C_{(adjusted)} = \frac{^{14}C_{(measured)}}{P} \qquad (17)$$

or, for this example,

$$^{14}C_{(adjusted)} = \frac{0.7}{0.044} = 15.91 \text{ percent modern.}$$

The residence time of water in Big Creek Hot Springs can be calculated using equation 6:

$$t = -8{,}266 \ln \frac{15.91}{100} \qquad (18)$$

$$t = 15{,}200 \text{ years}$$

This procedure was used on the remaining thermal springs, assuming input carbon concentrations were the same as those in nonthermal spring water. Adjusted ^{14}C values were obtained for six of the thermal springs (table 2). Flowpaths used for two of the springs resulted in carbonate loss (precipitation). Adjusted residence times ranged from 9,000 to 28,800 years. The latter value is questionable because the measured ^{14}C value was less than 0.7 percent modern.

Validity of the adjusted residence times for sampled thermal springs is unknown because of the uncertainty of the input carbon concentrations. However, on the basis of ^{14}C values, thermal waters discharging from the study area are old--possible residence times range from 9,000 to greater than 40,000 years.

SUMMARY

The Idaho batholith and adjacent areas include about 23,200 mi^2 of the Payette, Boise, Salmon, and Clearwater River drainage basins. The study area is characterized by rugged, forested mountains and steep, narrow river valleys. Cretaceous granitic rocks predominate throughout the central part of the area. Triassic metavolcanic and Tertiary volcanic rocks are common along the western margin, and Precambrian metamorphic and sedimentary rocks, Paleozoic sedimentary rocks, and Tertiary volcanic rocks predominate along the eastern margin.

The occurrence and nature of thermal springs suggest a relation between structural controls and deeply circulating hot-water systems. Thermal springs that issue from granitic rocks appear to be associated with major regional fault structures, as demonstrated by their areal occurrence and alignment along major rivers. Thermal springs that issue from other rocks are randomly scattered and probably are associated with local faulting.

Individual spring flows are variable and range from less than 1 gal/min to 2,710 gal/min. Thermal ground-water discharge from the area is at least 27,000 acre-ft annually. Heat flux from the sampled springs was estimated to be 5.0 x 10^7 cal/s. Measured water temperatures at 129 sampled thermal springs ranged from 20.5 °C to 94.0 °C.

Thermal springs in the Idaho batholith and adjacent areas discharge relatively dilute water; dissolved-solids concentrations range from 103 to 839 mg/L. Most waters are sodium mixed-anion type and pH values are greater than 8.0. The chemical quality of the water suggests deep circulation of meteoric water.

Reservoir temperatures estimated using chemical geothermometers are generally less than 100 °C. However, estimated reservoir temperatures for several springs exceeded 150 °C. Reservoir temperatures for 19 thermal springs estimated using the sulfate-water geothermometer are generally in good agreement with reservoir temperatures estimated using chemical geothermometers.

Tritium concentrations in most sampled thermal springs averaged 0.8 TU. Low tritium concentrations in several springs suggests mixing of shallow ground water with thermal waters.

Stable-isotope data are not conclusive insofar as indicating a source of recharge for the thermal springs. However, significant depletion of deuterium values relative to oxygen-18 values suggests that most of the thermal water is not derived from current precipitation. Carbon-14 values indicate that thermal waters discharging from the study area are old; possible residence times range from 9,000 to greater than 40,000 years.

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Data table A.--Chemical analyses of water from thermal and selected nonthermal springs

[Chemical constituents in milligrams per liter, except where noted; --, constituents not analyzed for; <, less than]

Spring number or name	Date of collection	Flow rate [1] (gal/min)	Specific conductance (µmhos)	pH	Water temperature[2] (°C)	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate[3] (HCO_3)	Carbonate[3] (CO_3)	Alkalinity as $CaCO_3$
Nonthermal												
Lochsa Cold Spring	7-15-82	2.0	95	7.0	11.0	8.5	3.6	4.7	2.2	60	0	49
Granite Springs	7-13-82	15	18	6.2	3.0	1.5	.25	2.1	.4	9	0	7
Horse Creek Campground Spring	8- 6-80	--	47	6.4	12.0	4.2	1.4	3.1	.4	27	0	22
24N- 4E-19BCA1S	6- 3-80	2	154	7.0	11.0	18	2.0	7.5	2.4	73	0	60
24N-21E-22CDD1S	6-18-80	150	158	6.7	7.5	18	5.1	5.4	1.7	76	0	62
22N- 2E-31BBC1S	6- 3-80	15	56	6.5	5.5	5.8	2.0	2.5	.8	9.8	0	8
Hazard Lake Campground Spring	7-27-82	4	20	5.3	2.0	1.4	.2	1.0	.2	5	0	4
Jureano Spring	8-12-82	8	19	6.6	2.5	1.2	.3	1.9	.6	12	0	10
20N-22E-17AAC1S	8-18-80	--	38	6.9	[5]14.0	3.1	1.0	2.1	1.1	16	0	13
20N-25E-16ACC1S	8-13-82	8	26	6.4	4.5	1.8	.4	2.4	.7	13	0	11
19N- 6E- 3BBD1S	6- 5-80	--	102	6.8	9.5	13	.8	6.6	.8	56	0	46
17N- 2E-15BAB1S	6-28-79	2	85	7.3	7.0	7.8	2.9	3.9	1.8	51	0	42
17N-11E-33DDC1S	6-27-80	30	206	7.5	9.5	31	3.9	5.8	1.3	130	0	107
15N- 5E-23DAC1S	6-27-79	2	20	6.2	4.0	1.7	.2	2.4	.3	12	0	10
15N-10E-13CAB1S	6-26-80	25	111	7.5	9.0	15	1.6	5.5	.7	63	0	52
14N-26E-10CBC1S	8- 5-80	--	168	8.7	9.5	16	9.2	2.7	.6	84	6	79
13N- 5E-21CAA1S	6-26-79	4	21	6.2	6.5	1.1	.1	1.9	.2	7	0	6
11N-16E-30DAA1S	8- 8-80	20	236	7.4	8.5	33	6.1	4.6	2.4	120	0	98
11N-20E-17AAB1S	8-11-82	1	446	8.3	14.0	47	15	22	.7	220	0	180
10N- 6E-19DCD1S	7-20-82	10	22	6.2	5.0	2.1	.2	2.0	.3	11	0	9
9N- 9E-32BAD1S	8-31-79	2	52	6.6	4.5	3.8	1.9	3.1	.6	29	0	24
9N-17E-22BCA1S	8- 7-80	--	132	7.5	9.0	17	2.2	5.0	.4	71	0	58
7N- 1W-10ADB1S	4-19-82	--	416	7.7	10.0	44	19	19	4.7	240	0	200
7N- 7E-32DAC1S	9-15-81	1	26	6.6	6.0	2.3	.4	2.1	0	15	0	12
Pole Creek Spring	8- 9-82	30	186	7.6	5.0	35	2.3	1.2	.4	110	0	88
5N- 6E-28ACB1S	7- 7-82	--	41	6.1	5.5	3.9	.4	2.8	.5	15	0	12
5N-11E- 2DCC1S	9- 1-81	--	102	6.7	7.5	14	1.1	3.4	.5	51	0	42
3N-11E- 5ADA1S	7-15-81	--	211	8.0	10.0	33	2.8	10	.6	120	0	98
3N-13E- 2DBA1S	7-14-81	--	205	6.9	6.0	36	3.0	4.6	1.0	120	0	98
2N-10E- 5ADB1S	8- 4-81	--	136	6.7	9.5	18	2.3	6.8	.8	78	0	64
Thermal												
Colgate Hot Springs	7-14-82	30	288	9.5	45.0	2.4	.13	57	1.7	22	31	70
Weir Creek Hot Springs	7-14-82	25	150	9.4	47.0	2.9	.02	28	.7	21	22	54
Jerry Johnson Hot Springs	7-14-82	200	183	9.5	48.5	2.4	.12	36	.5	22	25	60
Running Creek Hot Springs	7-13-82	50	200	9.7	43.0	1.6	.03	42	.8	20	38	80
Red River Hot Springs	7-13-82	35	365	9.4	55.0	1.7	<.01	79	1.7	32	35	84

Data table A.--Chemical analyses of water from thermal and selected nonthermal springs--Continued

Spring number or name	Date of collection	Sulfate[4] (SO_4)	Chloride (Cl)	Fluoride (F)	Silica (SiO_2)	Dissolved solids, calculated	Nitrite plus nitrate as N ($NO_2 + NO_3$)	Phosphorus, total, as P	Arsenic (As) (µg/L)	Boron (B) (µg/L)	Lithium (Li) (µg/L)	Mercury (Hg) (µg/L)
Nonthermal--Continued												
Lochsa Cold Spring	7-15-82	<5	0.3	0.1	30	79	0.10	0.08	<1	<10	9	<0.1
Granite Springs	7-13-82	<5	.5	<.1	12	21	.26	.06	<1	<10	<4	<.1
Horse Creek Campground Spring	8- 6-80	4.1	.3	.3	15	42	.12	.02	0	30	10	<.01
24N- 4E-19BCA1S	6- 3-80	22	.3	0	23	111	.24	<.01	1	0	10	.3
24N-21E-22CDD1S	6-18-80	11	2.0	.2	17	98	.29	.09	2	60	<4	.5
22N- 2E-31BBC1S	6- 3-80	.5	.2	.1	24	41	.04	.01	0	9	<4	<.01
Hazard Lake Campground Spring	7-27-82	<5	.3	<.1	7.6	13	.10	.02	<1	<10	<4	<.1
Jureano Spring	8-12-82	<5	.6	<.1	13	24	<.10	<.01	<1	<10	<4	<.1
20N-22E-17AAC1S	8-18-80	.2	1.1	.1	16	34	.14	.10	2	10	<4	.5
20N-25E-16ACC1S	8-13-82	<5	.5	<.1	16	28	.11	.04	<1	<10	<4	<.1
19N- 6E- 3BBD1S	6- 5-80	2.7	.7	.2	31	83	.09	.02	30	0	10	<.01
17N- 2E-15BAB1S	6-28-79	2.6	.4	<.1	45	90	.11	.02	0	2	<4	<.01
17N-11E-33DDC1S	6-27-80	4.6	.4	---	10	121	1.4	<.01	0	9	8	<.01
15N- 5E-23DAC1S	6-27-79	.7	.3	.2	17	29	<.01	.02	1	0	<4	<.01
15N-10E-13CAB1S	6-26-80	2.8	.4	.2	16	73	1.1	.05	3	7	5	<.01
14N-26E-10CBC1S	8- 5-80	2.4	1.9	.4	8.2	89	.12	.02	2	30	7	<.01
13N- 5E-21CAA1S	6-26-79	1.4	.4	<.1	13	22	<.01	.02	0	0	<4	<.01
11N-16E-30DAA1S	8- 8-80	25	1.2	.5	21	153	.14	.03	3	30	10	---
11N-20E-17AAB1S	8-11-82	31	19	.3	19	262	<.10	.06	1	30	12	<.1
10N- 6E-19DCD1S	7-20-82	<5	.3	<.1	11	21	<.10	.09	<1	<10	4	<.1
9N- 9E-32BAD1S	8-31-79	1.9	.4	.1	14	42	.13	.02	0	10	10	<.01
9N-17E-22BCA1S	8- 7-80	4.3	1.4	.6	16	82	.04	.02	2	20	7	---
7N- 1W-10ADB1S	4-19-82	15	4.0	.2	41	278	3.9	.09	1	50	10	<.1
7N- 7E-32DAC1S	9-15-81	5.0	.1	0	14	31	.12	.01	0	0	<4	<.1
Pole Creek Spring	8- 9-82	6.0	.5	<.1	17	115	2.2	<.01	<1	<10	8	<.1
5N- 6E-28ACB1S	7- 7-82	<5	.3	<.1	18	33	1.8	.05	<1	<10	6	<.1
5N-11E- 2DCC1S	9- 1-81	5.0	.3	.5	15	65	.20	.01	0	0	<4	<.1
3N-11E- 5ADA1S	7-15-81	5.0	.9	.1	19	132	1.9	.01	1	0	30	<.1
3N-13E- 2DBA1S	7-14-81	2.0	1.0	.2	12	119	.95	.01	0	0	20	<.1
2N-10E- 5ADB1S	8- 4-81	1.0	.5	.5	24	92	<.01	.19	2	10	8	<.1
Thermal--Continued												
Colgate Hot Springs	7-14-82	50	2.2	6.1	76	238	<.10	.08	2	90	110	<.1
Weir Creek Hot Springs	7-14-82	18	1.9	2.0	49	135	<.10	.02	1	30	39	<.1
Jerry Johnson Hot Springs	7-14-82	17	1.2	1.5	49	144	<.10	.05	1	20	30	<.1
Running Creek Hot Springs	7-13-82	6	1.4	5.7	63	168	.17	.05	<1	50	46	<.1
Red River Hot Springs	7-13-82	37	4.2	22	76	272	<.10	.03	1	320	160	<.1

Data table A.--Chemical analyses of water from thermal and selected nonthermal springs--Continued

Spring number or name	Date of collection	Flow rate (gal/min)[1]	Specific conductance (μmhos)	pH	Water temperature[2] (°C)	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate[3] (HCO_3)	Carbonate[3] (CO_3)	Alkalinity as $CaCO_3$
Thermal--Continued												
Barth Hot Springs	8-19-80	200	240	9.4	59.5	1.5	0.1	53	1.1	24	36	80
Unnamed Hot Spring	8-19-80	40	225	9.5	45.5	1.5	.1	49	.7	20	38	80
Horse Creek Hot Springs	8- 6-80	50	193	9.0	39.0	2.6	.1	40	1.1	54	16	71
24N- 2E-14DAC1S	6- 3-80	25	832	9.4	41.0	6.3	.2	160	3.8	5	29	52
Owl Creek Hot Springs	8- 6-80	6	562	8.4	51.0	4.5	.1	120	7.4	200	2	167
Big Creek Hot Springs	8- 6-80	75	1,020	7.6	94.0	4.9	<.1	220	17	490	0	402
22N- 1E-34DAD1S	6- 3-80	20	248	9.9	27.0	1.6	<.1	48	.4	7	38	69
22N- 2E-23CCB1S	8-23-80	50	865	8.6	49.0	22	.1	150	5.2	18	7	26
22N- 4E- 1BDC1S	6- 3-80	160	226	9.7	45.0	1.5	<.1	45	.9	15	46	89
21N- 1E-23ABA1S	6- 3-80	30	670	8.8	30.0	11	1.6	130	4.1	85	10	86
20N- 1E-26DDB1S	6- 3-80	200	954	9.4	64.0	11	.2	180	3.6	2	23	40
20N- 5E-13BCC1S	8-26-80	20	417	9.0	31.5	2.6	.2	90	2.7	61	19	82
20N-22E- 3ABD1S	8- 7-80	145	1,050	7.1	45.0	21	9.8	190	30	550	0	451
20N-24E-34CCC1S	8- 5-80	8.5	1,270	8.0	63.5	5.8	<.1	270	17	470	0	385
19N- 2E-22CCA1S	6- 3-80	20	677	9.3	43.0	4.9	.1	130	3.3	43	19	67
19N-14E-26DDD1S	6-29-80	20	246	9.4	47.0	2.8	<.1	50	.8	32	22	63
18N- 6E- 9ADC1S	6- 4-80	20	299	9.3	62.0	1.5	.1	71	1.2	43	31	87
Forge Creek Hot Springs	8- 1-79	40	334	8.7	36.0	4.5	<.1	67	.7	100	6	92
18N-21E-12BCD1S	8- 5-80	60	510	7.8	45.0	73	16	4.9	2.5	160	0	131
17N- 6E- 2BAB1S	6- 4-80	10	267	9.6	47.0	1.4	<.1	53	1.1	32	35	85
17N- 7E-31BCB1S	6- 5-80	6	343	9.7	35.0	1.6	.1	72	1.7	41	36	94
Kwiskwis Hot Springs	7-31-79	60	540	9.0	68.0	2.3	<.1	120	2.4	77	22	100
17N-11E-16ACB1S	7-31-79	65	498	9.0	87.0	2.0	<.1	110	4.4	100	23	120
17N-13E-27ACC1S	6-28-80	15	370	9.3	56.0	1.8	<.1	84	1.4	46	31	89
27ADB1S	6-28-80	10	432	8.9	39.0	1.8	<.1	100	1.7	120	16	125
Hospital Hot Springs	6-29-80	40	422	8.9	46.0	3.4	<.1	90	1.5	110	11	109
Lower Loon Creek Hot Springs	6-28-80	30	430	9.0	49.5	1.8	<.1	94	1.7	93	19	108
16N- 4E-35CCB1S	6-27-79	94	304	9.5	50.0	1.1	<.1	70	.7	49	36	100
16N- 6E-14CCC1S	6- 4-80	40	140	7.7	20.5	4.0	.2	25	.6	63	0	52
16N-10E-14CDA1S	8- 1-79	40	402	9.6	64.0	1.9	<.1	85	1.3	5	43	76
16N-12E- 8DDC1S	6-27-80	15	475	9.2	47.0	4.4	.1	97	2.1	20	22	53
15BBA1S	6-28-80	70	353	9.3	66.0	2.4	.2	75	2.0	34	29	76
17DAD1S	6-27-80	100	410	9.3	62.0	2.0	.1	86	2.5	27	36	82
16N-21E-18ADC1S	8- 5-80	10	760	7.9	45.5	9.4	1.0	160	11	340	0	279
15N- 3E-13BBC1S [6]	8- 2-72	798	279	---	34.0	1.3	.1	60	.6	17	45	89

Data table A.—Chemical analyses of water from thermal and selected nonthermal springs—Continued

Spring number or name	Date of collection	Sulfate[4] (SO_4)	Chlor-ide (Cl)	Fluor-ide (F)	Silica (SiO_2)	Dis-solved solids, calcu-lated	Nitrite plus nitrate as N (NO_2 + NO_3)	Phos-phorus, total, as P	Arsenic (As) (μg/L)	Boron (B) (μg/L)	Lithium (Li) (μg/L)	Mercury (Hg) (μg/L)
					Thermal—Continued							
Barth Hot Springs	8-19-80	13	3.5	.9	68	189	<.01	0.01	3	80	100	<0.01
Unnamed Hot Spring	8-19-80	12	2.7	.8	60	175	<.01	.01	3	50	90	<.01
Horse Creek Hot Springs	8- 6-80	4.4	2.7	8.6	53	155	.83	.01	1	30	60	—
24N- 2E-14DAC1S	6- 3-80	290	5.6	.8	68	566	.06	<.01	1	750	20	.8
Owl Creek Hot Springs	8- 6-80	63	7.0	20	87	409	<.01	.01	1	60	220	—
Big Creek Hot Springs	8- 6-80	45	31	16	150	725	<.01	.01	2	460	570	—
22N- 1E-34DAD1S	6- 3-80	39	4.4	.5	39	174	.06	.01	0	160	6	<.01
22N- 2E-23CCB1S	8-23-80	310	14	2.5	51	570	<.01	.03	34	700	100	.1
22N- 4E- 1BDC1S	6- 3-80	16	2.3	2.9	67	189	.08	<.01	4	20	20	<.01
21N- 1E-23ABA1S	6- 3-80	190	22	.8	52	463	<.01	<.01	1	880	40	.2
20N- 1E-26DDB1S	6- 3-80	320	30	.8	61	631	.03	<.01	13	610	50	.1
20N- 5E-13BCC1S	8-26-80	37	26	1.8	75	285	<.01	.01	47	150	40	<.01
20N-22E- 3ABD1S	8- 7-80	31	49	1.6	32	635	<.01	.02	22	440	440	<.01
20N-24E-34CCC1S	8- 5-80	150	54	11	100	839	.01	.01	22	1,700	430	—
19N- 2E-22CCA1S	6- 3-80	180	25	1.5	73	458	<.01	<.01	8	610	60	1.4
19N-14E-26DDD1S	6-29-80	29	5.3	8.0	56	190	1.1	<.01	2	30	30	<.01
18N- 6E- 9ADC1S	6- 4-80	20	11	2.4	74	233	.08	<.01	4	60	30	.3
Forge Creek Hot Springs	8- 1-79	36	8.8	10	43	225	.04	.02	2	190	60	<.01
18N-21E-12BCD1S	8- 5-80	120	1.5	.4	32	329	.32	.01	6	20	20	—
17N- 6E- 2BAB1S	6- 4-80	17	7.8	2.2	64	198	.01	<.01	7	50	20	<.01
17N- 7E-31BCB1S	6- 5-80	32	10	10	68	252	.18	<.01	4	90	30	.1
Kwiskwis Hot Springs	7-31-79	80	17	24	73	379	<.01	.02	1	250	150	<.01
17N-11E-16ACB1S	7-31-79	66	11	22	98	385	<.01	.02	1	180	120	<.01
17N-13E-27ACC1S	6-28-80	40	9.6	15	67	272	1.1	<.01	2	100	90	<.01
27ADB1S	6-28-80	47	12	13	79	330	<.01	<.01	2	130	110	<.01
Hospital Hot Springs	6-29-80	41	14	21	52	288	<.01	<.01	1	110	120	<.01
Lower Loon Creek Hot Springs	6-28-80	50	11	16	69	308	1.0	<.01	2	100	110	<.01
16N- 4E-35CCB1S	6-27-79	12	11	9.0	63	215	<.01	.01	11	50	30	<.01
16N- 6E-14CCC1S	6- 4-80	7.6	1.2	1.3	32	103	.10	.03	26	20	20	<.01
16N-10E-14CDA1S	8- 1-79	63	11	17	69	294	<.01	.02	1	130	50	<.01
16N-12E- 8DDC1S	6-27-80	110	13	19	60	337	1.1	<.01	2	120	120	<.01
15BBA1S	6-28-80	48	8.1	15	77	273	1.1	<.01	2	90	70	<.01
17DAD1S	6-27-80	63	9.0	17	85	314	1.1	<.01	2	100	90	<.01
16N-21E-18ADC1S	8- 5-80	59	26	7.2	36	477	.02	.01	18	760	180	—
15N- 3E-13BBC1S	8- 2-72	43	16	2.6	60	210	—	—	—	—	—	—

Data table A.--Chemical analyses of water from thermal and selected nonthermal springs--Continued

Spring number or name	Date of collection	Flow rate[1] (gal/min)	Specific conductance (µmhos)	pH	Water temperature[2] (°C)	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate[3] (HCO_3)	Carbonate[3] (CO_3)	Alkalinity as $CaCO_3$
					Thermal--Continued							
15N- 4E-21DCC1S	6-27-79	31	206	9.6	36.0	1.3	0.1	43	0.5	32	34	84
15N- 6E-14ABB1S	6- 6-80	60	309	9.3	59.0	1.6	<.1	76	1.4	41	34	90
14ACC1S[6]	6- 2-82	30	309	---	59.0	2.0	<.1	70	1.5	48	30	89
14CAC1S	6- 2-80	80	296	9.4	57.0	2.4	.3	65	1.9	52	25	84
17DCC1S	6- 6-80	40	262	9.7	51.0	1.2	<.1	57	.8	29	46	100
15N-10E-24BBB1S	6-25-80	300	270	9.5	49.0	2.6	.3	55	1.0	10	30	58
29BDA1S	6-24-80	60	282	9.4	50.0	2.0	<.1	63	1.2	29	31	75
Loon Creek Hot Springs	7-30-79	500	345	9.3	65.0	2.0	.1	70	1.5	35	31	80
Hot Creek Hot Springs	7-30-79	80	395	9.1	60.0	2.5	<.1	89	1.5	61	20	83
Shower Bath Hot Springs	7-30-79	300	321	9.3	53.0	2.1	<.1	70	.9	41	24	74
15N-26E-21CAD1S	8- 5-80	1,000	332	8.0	23.5	29	11	19	3.8	150	0	123
14N- 6E-11BDA1S	6- 2-80	500	465	9.1	88.5	1.7	<.1	95	.4	63	29	100
Dagger Creek Hot Springs	6-24-80	10	277	9.9	42.0	2.1	<.1	59	.8	15	26	56
14N-19E-23DDD1S	9-13-79	783	996	7.1	50.0	44	15	160	20	440	0	361
13N- 3E-13ADA1S	6-26-79	20	433	9.3	49.0	1.2	<.1	85	1.0	56	34	102
13N- 4E-31CAB1S	6-26-79	70	514	9.0	71.0	1.5	<.1	110	2.1	68	25	97
13N- 6E-29DAB1S	7- 7-79	50	334	9.5	53.0	1.6	.1	64	1.3	32	36	86
Bear Valley Hot Springs	8- 1-79	200	365	9.5	65.0	2.0	<.1	76	1.6	17	41	82
12N- 5E- 2DAC1S	7- 8-79	10	337	9.3	50.0	1.3	<.1	65	1.1	34	34	84
22BBC1S	6- 5-79	134	331	8.9	86.0	1.4	.1	74	1.9	95	18	108
36DBA1S	8- 8-79	334	266	8.9	38.0	2.8	<.1	54	.9	77	5	71
12N-11E- 2CDB1S	8- 4-80	60	387	9.6	37.0	1.4	.1	82	.3	22	41	86
12N-20E-10CBD1S	9-13-79	2,710	698	7.4	35.0	66	27	42	14	280	0	232
11N- 5E-29CDB1S	8- 8-79	50	293	8.8	49.0	3.1	<.1	97	.9	99	4	88
11N- 7E-16AAB1S	7-24-79	90	315	9.3	65.0	1.8	.1	65	1.8	13	35	69
11N-13E-36BAA1S	9-14-79	40	326	9.4	59.0	1.6	<.1	70	1.2	22	41	86
Sunbeam Hot Springs	9-14-79	440	400	9.1	77.5	1.6	<.1	89	2.6	65	31	105
Robinson Bar Hot Springs	9-13-79	45	351	9.4	56.0	1.9	<.1	74	2.4	32	43	98
11N-17E-27BDD1S	9-13-79	50	1,070	7.1	41.0	46	9.9	170	19	580	0	476
10N- 4E-33CBD1S	11-27-78	500	402	8.4	75.0	1.9	.1	90	3.2	130	2	110
10N-10E-31BCC1S	3-13-79	360	378	9.3	84.0	1.5	<.1	77	3.2	18	38	78
10N-11E-31ADD1S	5- 7-79	75	381	9.6	67.0	1.9	.1	77	2.6	1	45	76
10N-13E- 3CAB1S[6]	7-12-72	110	293	---	41.0	2.2	.1	60	.5	30	28	71
Slate Creek Hot Springs	9-13-79	185	443	8.6	50.0	9.8	<.1	76	4.8	110	12	112
9N- 3E-25BAC1S	11-22-78	32	622	8.0	79.5	6.1	.1	120	5.3	160	0	131

Data table A.--Chemical analyses of water from thermal and selected nonthermal springs--Continued

Spring number or name	Date of collection	Sulfate[4] (SO_4)	Chloride (Cl)	Fluoride (F)	Silica (SiO_2)	Dissolved solids, calculated	Nitrite plus nitrate as N (NO_2 + NO_3)	Phosphorus, total, as P	Arsenic (As) (μg/L)	Boron (B) (μg/L)	Lithium (Li) (μg/L)	Mercury (Hg) (μg/L)
					Thermal--Continued							
15N- 4E-21DCC1S	6-27-79	7.8	2.5	3.1	48	146	<0.01	0.01	3	10	8	<0.01
15N- 6E-14ABB1S	6- 6-80	15	8.5	11	79	246	<.01	<.01	18	70	60	<.01
14ACC1S	6- 2-82	17	10	17	87	258	.03	.02	—	—	—	—
14CAC1S	6- 2-80	15	11	3.0	74	223	.24	<.01	16	60	70	.5
17DCC1S	6- 6-80	4.6	2.2	2.6	68	196	.03	<.01	9	4	40	.1
15N-10E-24BBB1S	6-25-80	36	5.9	17	54	207	1.2	<.01	7	60	30	<.01
29BDA1S	6-24-80	31	6.9	15	65	229	1.1	<.01	2	60	50	<.01
Loon Creek Hot Springs	7-30-79	43	8.7	15	74	262	<.01	.02	2	120	70	.1
Hot Creek Hot Springs	7-30-79	52	9.9	17	65	287	.04	.02	1	240	100	<.01
Shower Bath Hot Springs	7-30-79	36	7.1	17	61	238	<.01	.02	1	170	60	<.01
15N-26E-21CAD1S	8- 5-80	32	9.9	.8	20	199	.28	.02	19	160	40	<.01
14N- 6E-11BDA1S	6- 2-80	41	15	16	100	329	.01	<.01	1	120	90	.5
Dagger Creek Hot Springs	6-24-80	32	8.5	14	58	208	1.1	<.01	0	60	40	<.01
14N-19E-23DDD1S	9-13-79	150	21	4.3	31	662	.11	.01	20	420	100	<.01
13N- 3E-13ADA1S	6-26-79	24	28	5.3	61	256	<.01	.01	21	310	40	<.01
13N- 4E-31CAB1S	6-26-79	43	40	11	72	329	<.01	.01	39	400	60	.1
13N- 6E-29DAB1S	7- 7-79	31	7.8	15	80	240	<.01	.01	6	50	30	<.01
Bear Valley Hot Springs	8- 1-79	54	12	15	77	287	<.01	.02	0	90	60	<.01
12N- 5E- 2DAC1S	7- 8-79	16	6.5	16	73	219	<.01	<.01	6	60	60	<.01
22BBC1S	6- 5-79	12	11	12	86	259	<.01	.01	10	90	80	.2
36DBA1S	8- 8-79	26	5.8	12	54	196	.03	.02	5	50	50	<.01
12N-11E- 2CDB1S	8- 4-80	39	8.7	22	61	266	<.01	.01	0	40	60	—
12N-20E-10CBD1S	9-13-79	150	8.6	1.7	26	474	.05	<.01	4	220	90	<.01
11N- 5E-29CDB1S	8- 8-79	24	9.6	13	74	277	<.01	.01	6	70	90	<.01
11N- 7E-16AAB1S	7-24-79	36	3.9	15	82	236	.02	.03	14	50	80	<.01
11N-13E-36BAA1S	9-14-79	31	8.3	17	76	257	<.01	<.01	1	60	130	.2
Sunbeam Hot Springs	9-14-79	50	15	16	89	326	<.01	<.01	1	180	60	.1
Robinson Bar Hot Springs	9-13-79	50	15	9.4	93	304	<.01	.01	0	140	50	.1
11N-17E-27BDD1S	9-13-79	28	55	3.7	39	656	<.01	.01	1	1,500	320	.2
10N- 4E-33CBD1S	11-27-78	27	19	13	110	331	.01	<.01	9	120	100	<.01
10N-10E-31BCC1S	3-13-79	50	8.0	17	91	282	<.01	<.01	1	70	110	.2
10N-11E-31ADD1S	5- 7-79	46	9.5	15	82	267	.02	<.01	2	90	110	.1
10N-13E- 3CAB1S	7-12-72	31	5	14	55	211	.05	.01	—	—	—	—
Slate Creek Hot Springs	9-13-79	81	24	9.2	79	351	<.01	<.01	1	250	60	.2
9N- 3E-25BAC1S	11-22-78	88	39	12	120	470	.01	<.01	5	240	150	<.01

Data table A.--Chemical analyses of water from thermal and selected nonthermal springs--Continued

Spring number or name	Date of collection	Flow rate[1] (gal/min)	Specific conductance (µmhos)	pH	Water temperature[2] (°C)	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate[3] (HCO_3)	Carbonate[3] (CO_3)	Alkalinity as $CaCO_3$
Thermal--Continued												
9N- 8E-31ACA1S	3-13-79	20	369	9.2	63.5	1.4	<0.1	79	1.8	40	31	84
32CBA1S	3-27-79	350	321	9.1	64.0	1.9	<.1	70	1.3	29	29	72
9N-14E-19BAA1S	9-12-79	50	277	9.4	47.0	2.7	<.1	57	.7	22	34	75
8N- 5E- 1BCC1S	6- 8-72	1.5	317	---	40.0	2.4	.1	66	.9	85	1	71
6DCB1S	11-23-78	14	359	9.0	42.0	2.9	.1	77	1.2	67	20	88
6DCC1S	2-26-79	75	343	9.1	51.0	5.0	<.1	72	1.3	61	22	87
10ADD1S	11-22-78	10	344	9.0	54.0	2.0	<.1	76	1.2	59	19	80
10BDD1S	11-27-78	90	345	9.2	54.5	2.0	.1	75	1.0	40	23	78
11BAA1S	2-26-79	35	352	9.1	60.0	1.8	<.1	78	1.4	49	28	87
8N- 6E- 1ADB1S	2-27-79	175	343	9.2	59.5	2.0	<.1	75	1.1	39	29	80
8N-14E-27BDB1S	9-12-79	330	243	9.3	42.0	2.6	.1	51	1.1	41	24	74
8N-17E-31DCB1S	8- 7-80	50	562	8.5	52.5	24	6.2	80	10	180	7	159
32BCA1S	8- 7-80	50	643	8.5	52.5	20	5.1	110	16	230	17	217
7N- 1E- 8DDA1S	9- 5-79	50	780	8.0	64.0	8.4	.2	160	8.0	180	0	148
9CDC1S[6]	8- 4-72	20	529	---	45.0	15	2.4	99	5.3	169	0	139
6N- 5E-33ABC1S	5-11-81	6	292	9.5	40.5	1.5	.1	65	.8	41	36	94
33ADC1S	5-11-81	200	317	9.5	42.0	2.0	.1	65	.8	41	36	94
6N-10E-30CDA1S	9- 2-81	150	282	9.5	64.0	3.0	.3	58	1.2	7	38	69
6N-11E-30ADB1S	9- 2-81	60	231	9.6	50.5	3.2	<.1	48	.8	5	34	61
35DAD1S	9- 1-81	60	317	9.7	60.0	1.9	<.1	67	1.5	2	46	78
35DBB1S	9- 1-81	100	305	9.2	53.0	2.2	.2	62	1.8	30	25	66
5N- 7E-24BDD1S	8- 3-81	424	297	9.4	76.0	1.6	---	64	1.7	16	46	90
34CCB1S	8- 3-81	100	210	9.5	60.0	2.3	<.1	44	.7	15	32	66
34DBA1S	5-27-81	15	197	9.6	55.0	2.4	<.1	42	1.0	4	36	63
5N- 8E-10DCA1S	9- 2-81	50	245	9.4	51.0	1.8	<.1	52	1.3	37	31	82
5N- 9E- 5AAD1S	5-28-81	100	241	9.4	56.0	1.9	<.1	52	1.3	18	34	71
7BAB1S	5-28-81	300	250	9.6	65.0	1.6	<.1	57	1.2	15	38	76
4N- 6E-24BCB1S	5-27-81	300	236	9.5	67.0	1.7	.1	49	1.3	9	43	79
4N- 7E- 8CBB1S	5-27-81	8	223	9.5	61.0	2.5	<.1	50	1.2	20	41	85
4N-11E-34DBB1S	7-16-81	60	221	9.5	53.0	2.1	<.1	46	1.0	24	32	73
4N-14E-29DCD1S	7-15-81	150	285	9.4	64.5	3.4	<.1	60	1.5	38	25	73
3N-10E-33ACD1S	7-16-81	200	226	9.5	53.0	1.5	<.1	46	1.1	27	38	85
3N-12E- 7DCD1S	7-15-81	10	258	9.1	50.0	3.5	.1	53	.9	71	12	78
3N-13E- 7DCD1S	7-15-81	45	343	9.1	62.0	2.5	<.1	75	1.9	90	19	105
3N-14E-19DDB1S	7-14-81	2	344	9.6	41.0	1.8	<.1	78	.8	37	34	87
28CAD1S	7-14-81	466	337	9.4	87.0	1.5	<.1	69	1.9	33	40	94
2N-10E- 5ADA1S	8- 4-81	50	243	9.5	60.0	1.3	.9	54	1.3	35	38	92

Data table A.--Chemical analyses of water from thermal and selected nonthermal springs--Continued

Spring number or name	Date of collection	Sulfate[4] (SO$_4$)	Chlor- ide (Cl)	Fluor- ide (F)	Silica (SiO$_2$)	Dis- solved solids, calcu- lated	Nitrite plus nitrate as N (NO$_2$ + NO$_3$)	Phos- phorus, total, as P	Arsenic (As) (μg/L)	Boron (B) (μg/L)	Lithium (Li) (μg/L)	Mercury (Hg) (μg/L)
					Thermal--Continued							
9N- 8E-31ACA1S	3-13-79	39	5.7	20	83	270	<0.01	<0.01	2	20	120	<0.01
32CBA1S	3-27-79	39	3.9	16	67	232	.02	<.01	2	50	110	<.01
9N-14E-19BAA1S	9-12-79	34	6.5	11	54	211	.13	.01	1	50	60	<.01
8N- 5E- 1BCC1S	6- 8-72	42	5.1	3.1	48	216	---	---	—	---	---	---
6DCB1S	11-23-78	37	7.3	15	75	262	.01	<.01	6	110	120	<.01
6DCC1S	2-26-79	36	7.8	15	63	245	.01	<.01	6	110	130	<.01
10ADD1S	11-22-78	40	5.7	14	78	259	.01	<.01	4	90	80	<.01
10BDD1S	11-27-78	38	5.7	14	74	249	.01	<.01	4	70	70	<.01
11BAA1S	2-26-79	41	5.4	17	66	254	.01	<.01	5	90	160	<.01
8N- 6E- 1ADB1S	2-27-79	38	6.7	17	63	242	.01	<.01	6	80	100	<.01
8N-14E-27BDB1S	9-12-79	32	3.9	6.1	55	196	<.01	<.01	2	40	60	<.01
8N-17E-31DCB1S	8- 7-80	120	13	9.3	47	405	<.01	.01	1	540	120	---
32BCA1S	8- 7-80	83	39	10	41	454	.23	.01	1	790	160	.1
7N- 1E- 8DDA1S	9- 5-79	120	61	16	110	573	<.01	<.01	10	710	220	<.01
9CDC1S	8- 4-72	57	30	8	94	397	---	---	—	---	---	---
6N- 5E-33ABC1S	5-11-81	22	2.7	13	58	220	.05	.04	48	50	30	.1
33ADC1S	5-11-81	22	2.7	13	57	219	.02	.04	35	50	30	.1
6N-10E-30CDA1S	9- 2-81	41	2.5	7.9	67	222	.10	.02	10	20	80	<.1
6N-11E-30ADB1S	9- 2-81	29	3.0	7.8	58	186	.1	.01	5	30	80	<.1
35DAD1S	9- 1-81	41	4.6	13	74	240	.1	.01	3	50	240	.3
35DBB1S	9- 1-81	41	4.4	13	79	250	.1	.01	4	30	230	<.1
5N- 7E-24BDD1S	8- 3-81	28	2.4	9.2	100	261	<.01	.16	1	40	90	.2
34CCB1S	8- 3-81	18	2.0	5.4	64	176	.01	.17	3	30	50	.1
34DBA1S	5-27-81	21	11	4.7	70	190	<.01	.01	1	0	40	<.1
5N- 8E-10DCA1S	9- 2-81	25	1.3	6.6	77	214	.1	.01	6	20	100	<.1
5N- 9E- 5AAD1S	5-28-81	26	1.7	7.1	68	201	<.01	.02	11	10	70	<.1
7BAB1S	5-28-81	27	1.8	8.3	70	212	<.01	.01	11	10	80	<.1
4N- 6E-24BCB1S	5-27-81	21	1.6	4.1	86	212	<.01	.01	16	30	40	<.1
4N- 7E- 8CBB1S	5-27-81	20	1.8	4.6	84	215	.28	.01	19	20	50	<.1
4N-11E-34DBB1S	7-16-81	23	3.2	5.1	63	187	.03	.01	4	20	90	<.1
4N-14E-29DCD1S	7-15-81	31	4.9	14	71	230	.01	<.01	1	50	180	.1
3N-10E-33ACD1S	7-16-81	16	2.9	2.9	72	193	.16	<.01	1	10	40	<.1
3N-12E- 7DCD1S	7-15-81	24	8.4	5.8	45	188	.20	.01	1	70	110	.1
3N-13E- 7DCD1S	7-15-81	27	12	13	70	264	.02	<.01	4	110	190	.1
3N-14E-19DDB1S	7-14-81	32	7.3	15	59	246	.07	<.01	2	50	200	<.1
28CAD1S	7-14-81	35	5.3	13	95	277	.17	<.01	3	50	190	.1
2N-10E- 5ADA1S	8- 4-81	17	3.6	4.1	76	213	<.01	.15	0	20	61	.2

[1] Flow rate is for entire spring complex.
[2] Water temperature is highest temperature at spring vents or temperature of sampled vent.
[3] Total alkalinity distributed as carbonate and bicarbonate at the spring temperature and pH.
[4] An unknown part of the reported sulfate concentrations may be the result of oxidation of sulfide to sulfate prior to analysis.
[5] Water temperature is lower at point of discharge.
[6] Analyses taken from Young and Mitchell (1973, table 2).

Data table B.--Tritium in water from selected springs

Spring number or name	Water temperature (°C)	Tritium (TU)
Nonthermal		
Horse Creek Campground Spring	12.0	75.2 ±3.6
24N- 4E-19BCA1S	11.0	33.2 ±3.3
24N-21E-22CDD1S	7.5	62.4 ±3.5
20N-22E-17AAC1S[1]	14.0	74.3 ±3.6
19N- 6E- 3BBD1S	9.5	25.6 ±3.3
17N-11E-33DDC1S	9.5	23.7 ±3.3
15N- 5E-23DAC1S	4.0	46.8 ±2.1
14N-26E-10CBC1S	9.5	72.3 ±3.6
13N- 5E-21CAA1S	6.5	60.2 ±2.7
9N-17E-22BCA1S	9.0	25.6 ±3.3
5N-11E- 2DCC1S	7.5	53.6 ±3.4
3N-13E- 2DBA1S	6.0	45.0 ±2.5
2N-10E- 5ADB1S	9.5	24.4 ±3.4
Thermal		
Horse Creek Hot Springs	39.0	4.1 ±0.5
24N- 2E-14DAC1S	41.0	0 ± .4
Owl Creek Hot Springs	51.0	.1 ± .4
22N- 4E- 1BDC1S	45.0	.4 ± .4
17N-13E-27ADB1S	39.0	.1 ± .4
Lower Loon Creek Hot Springs	49.5	.7 ± .5
Loon Creek Hot Springs	65.0	1.0 ± .5
13N- 6E-29DAB1S	53.0	.1 ± .1
9N- 8E-32CBA1S	64.0	0 ± .1
8N- 5E- 6DCC1S	51.0	2.0 ± .1
6N- 5E-33ADC1S	42.0	1.0 ± .4
6N-11E-30ADB1S	50.5	.3 ± .4
4N- 6E-24BCB1S	67.0	.6 ± .4

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[1] Measured temperature is probably higher than at point of discharge.

Data table C.--Stable-isotope analyses from selected springs

Spring number or name	Water temperature (°C)	δD_{SMOW}	$\delta^{18}O_{SMOW}$	[1]$\Delta^{18}O$
	Nonthermal			
Lochsa Cold Spring	11.0	-115	-15.5	+0.12
Granite Springs	3.0	-131	-17.7	- .08
Horse Creek Campground Spring	12.0	-139	-18.9	- .28
24N- 4E-19BCA1S	11.0	-125	-16.1	+ .78
24N-21E-22CDD1S	7.5	-140	-18.1	+ .65
22N- 2E-31BBC1S	5.5	-124	-16.7	+ .05
Hazard Lake Campground Spring	2.0	-127	-17.3	- .18
Jureano Spring	2.5	-139	-18.7	- .08
20N-22E-17AAC1S	[2]14.0	-139	-18.3	+ .32
20N-25E-16ACC1S	4.5	-143	-19.1	+ .02
19N- 6E- 3BBD1S	9.5	-132	-17.5	+ .25
17N- 2E-15BAB1S	7.0	-127	-17.1	0
17N-11E-33DDC1S	9.5	-140	-18.6	+ .15
15N- 5E-23DAC1S	4.0	-124	-17.7	---
15N-10E-13CAB1S	9.0	-132	-17.9	- .15
14N-26E-10CBC1S	9.5	-143	-18.8	+ .32
13N- 5E-21CAA1S	6.5	-128	-17.3	- .05
11N-16E-30DAA1S	8.5	-144	-18.6	+ .65
11N-20E-17AAB1S	14.0	-148	-18.7	+1.05
10N- 6E-19DCD1S	5.0	-124	-16.9	- .15
9N- 9E-32BAD1S	4.5	-131	-17.7	- .05
9N-17E-22BCA1S	9.0	-141	-18.2	+ .68
7N- 1W-10ADB1S	10.0	-118	-14.9	+1.10
7N- 7E-32DAC1S	6.0	-123	-16.7	- .08
Pole Creek Spring	5.0	-144	-19.3	- .05
5N- 6E-28ACB1S	5.5	-127	-17.1	+ .02
5N-11E- 2DCC1S	7.5	-133	-17.5	+ .38
3N-11E- 5ADA1S	10.0	-140	-18.3	+ .45
3N-13E- 2DBA1S	6.0	-131	-17.5	+ .12
2N-10E- 5ADB1S	9.5	-128	-16.8	+ .45

Data table C.--Stable-isotope analyses from selected springs
(Continued)

Spring number or name	Water temperature (°C)	δD_{SMOW}	$\delta^{18}O_{SMOW}$	$^{1}\Delta^{18}O$
	Thermal			
Jerry Johnson Hot Springs	48.5	-137	-18.3	+ .08
Running Creek Hot Springs	43.0	-139	-18.5	+ .12
Red River Hot Springs	55.0	-136	-17.9	+ .35
Barth Hot Springs	59.5	-149	-19.2	+ .68
Horse Creek Hot Springs	39.0	-149	-19.6	+ .28
24N- 2E-14DAC1S	41.0	-136	-18.0	+ .25
Owl Creek Hot Springs	51.0	-154	-19.7	+ .80
Big Creek Hot Springs	94.0	-154	-19.4	+1.10
22N- 2E-23CCB1S	49.0	-133	-17.5	+ .38
22N- 4E- 1BDC1S	45.0	-144	-19.0	+ .25
20N- 1E-26DDB1S	64.0	-139	-17.5	+1.12
20N-22E- 3ABD1S	45.0	-147	-19.1	+ .52
20N-24E-34CCC1S	63.5	-153	-19.6	+ .78
19N- 2E-22CCA1S	43.0	-138	-18.0	+ .50
19N-14E-26DDD1S	47.0	-146	-19.6	- .10
18N- 6E- 9ADC1S	62.0	-139	-18.8	- .18
18N-21E-12BCD1S	45.0	-146	-18.9	+ .60
17N- 6E- 2BAB1S	47.0	-139	-18.7	- .08
17N- 7E-31BCB1S	35.0	-143	-19.3	- .18
Kwiskwis Hot Springs	68.0	-148	-19.3	+ .45
17N-11E-16ACB1S	87.0	-152	-19.4	+ .85
17N-13E-27ACC1S	56.0	-148	-19.4	+ .35
27ADB1S	39.0	-149	-19.4	+ .40
Hospital Hot Springs	46.0	-149	-19.6	+ .28
Lower Loon Creek Hot Springs	49.5	-149	-19.5	+ .38
16N- 4E-35CCB1S	50.0	-135	-18.0	+ .12
16N-10E-14CDA1S	64.0	-147	-19.2	+ .42
16N-12E- 8DDC1S	47.0	-149	-19.4	+ .48
15BBA1S	66.0	-147	-19.6	+ .02
17DAD1S	62.0	-149	-19.6	+ .28
16N-21E-18ADC1S	45.5	-151	-19.2	+ .92
15N- 6E-14ABB1S	59.0	-136	-17.9	+ .35
17DCC1S	51.0	-136	-18.3	- .05
15N-10E-24BBB1S	49.0	-141	-19.0	- .12
29BDA1S	50.0	-140	-19.0	- .25

Data table C.--Stable-isotope analyses from selected springs (Continued)

Spring number or name	Water temperature (°C)	δD_{SMOW}	$\delta^{18}O_{SMOW}$	$^{1}\Delta^{18}O$
Thermal--Continued				
Loon Creek Hot Springs	65.0	-152	-19.8	+ .45
Hot Creek Hot Springs	60.0	-152	-19.9	+ .35
14N- 6E-11BDA1S	88.5	-140	-18.6	+ .15
Dagger Creek Hot Springs	42.0	-141	-19.1	- .22
13N- 4E-31CAB1S	71.0	-138	-17.3	+1.20
13N- 6E-29DAB1S	53.0	-137	-18.2	+ .15
Bear Valley Hot Springs	65.0	-144	-18.9	+ .35
12N- 5E- 2DAC1S	50.0	-136	-18.1	+ .15
22BBC1S	86.0	-134	-17.7	+ .28
12N-20E-10CBD1S	35.0	-145	-19.0	+ .38
11N- 5E-29CDB1S	49.0	-132	-17.7	+ .04
11N- 7E-16AAB1S	65.0	-138	-18.4	+ .20
Sunbeam Hot Springs	77.5	-149	-19.8	+ .08
10N- 4E-33CBD1S	75.0	-136	-18.2	+ .04
10N-10E-31BCC1S	84.0	-140	-18.1	+ .63
10N-11E-31ADD1S	67.0	-140	-18.7	+ .04
Slate Creek Hot Springs	50.0	-147	-19.7	- .08
9N- 3E-25BAC1S	79.5	-136	-17.2	+1.03
9N- 8E-32CBA1S	64.0	-135	-17.9	+ .22
9N-14E-19BAA1S	47.0	-150	-19.9	+ .10
8N- 5E- 6DCC1S	51.0	-136	-18.0	+ .23
10BDD1S	54.5	-137	-18.3	+ .06
8N- 6E- 1ADB1S	59.5	-138	-18.5	0
8N-17E-31DCB1S	52.5	-149	-19.4	+ .48
7N- 1E- 8DDA1S	64.0	-134	-15.6	+2.43
6N- 5E-33ADC1S	42.0	-141	-17.5	+1.38
6N-10E-30CDA1S	64.0	-140	-18.4	+ .35
6N-11E-30ADB1S	50.5	-138	-18.3	+ .20
35DAD1S	60.0	-141	-18.5	+ .38
5N- 7E-24BDD1S	76.0	-138	-18.0	+ .50
34DBA1S	55.0	-131	-17.7	- .08
5N- 8E-10DCA1S	51.0	-138	-18.0	+ .50
5N- 9E- 7BAB1S	65.0	-138	-18.2	+ .30
4N- 6E-24BCB1S	67.0	-131	-17.6	+ .02
4N-11E-34DBB1S	53.0	-141	-18.6	+ .28

Data table C.--Stable-isotope analyses from selected springs (Continued)

Spring number or name	Water temperature (°C)	δD_{SMOW}	$\delta^{18}O_{SMOW}$	[1]$\Delta^{18}O$
Thermal--Continued				
4N-14E-29DCD1S	64.5	-146	-19.4	+ .10
3N-10E-33ACD1S	53.0	-140	-18.4	+ .35
3N-12E- 7DCD1S	50.0	-139	-18.3	+ .32
3N-14E-28CAD1S	87.0	-146	-19.5	0

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[1] Enrichment (+) or depletion (-) of ^{18}O relative to the meteoric water line.

[2] Measured temperature is probably higher than at point of discharge.