Aqueous Geochemistry of the Bradys Hot Springs Geothermal Area, Churchill County, Nevada

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Abstract

The aqueous geochemistry of the Bradys Hot Springs geothermal system has been examined to evaluate the source of recharge to the system, the thermal aquifer temperature, and the geochemical controls on the major and selected minor constituents in the thermal water.

The Bradys Hot Springs geothermal area is associated with a typical mountain-bounding fault in the Basin and Range province. The area is the discharge part of a thermal system that currently (1985) supplies heat for a commercial food-dehydration plant. The hottest water sampled at Bradys Hot Springs contains about 2,600 milligrams per liter of dissolved solids dominated by sodium, chloride, and sulfate. Chemical geothermometer data indicate thermal-aquifer temperatures ranging from 137 to 184 °C, although a measured down-hole temperature of 212 °C has been reported. Local meteoric water that has been concentrated by evaporation is the likely source of recharge to the thermal system. The primary controls on the dissolved-solids content of the thermal water (heated, deeply circulating ground water) apparently are evaporative concentration before recharge to the thermal aquifer and an increase in silica during passage through the thermal aquifer. Evaporative concentration and mineral solubility appear to be the primary controls on the minor constituents. Bromide is controlled primarily by evaporative concentration. Barium, fluoride, lead, and manganese concentrations may be controlled by mineral solubility. Lithium and, to a lesser extent, boron and strontium are more concentrated in the thermal water than in the local nonthermal water with the same chloride concentrations; these elements thus may be useful indicators of thermal water in the vicinity of Bradys Hot Springs when used in conjunction with chloride data.

INTRODUCTION

Objectives and Scope

This report considers the aqueous geochemistry of the Bradys Hot Springs geothermal area; it is an extension of a reconnaissance study (1981) of the Bradys area and several other geothermal areas in northern Nevada by Olmsted and others (1975). The overall objective of this effort is to analyze geochemical data to gain a better understanding of the nature of the Bradys Hot Springs geothermal system. Specific objectives include an attempt to determine the source and general chemical character of recharge to the system, the aquifer temperature, and the geochemical controls on major and minor constituents of the thermal water.

Previous Studies

Sources of data on the geothermal resources of Bradys Hot Springs are presented in Olmsted and others (1973, p. 14, 15) and Garside and Schilling (1979, p. 6, 7). Development and history of the geothermal resource is discussed by Olmsted and others (1975, p. 206–227), Rudisill (1978), and Garside and Schilling (1979, p. 6–11). Chemical analyses of water samples collected in the Bradys Hot Springs area are published in Clarke and Chatard (1884, 1 analysis, p. 24), Harrill (1970, 6 analyses, p. 27), Garside and Schilling (1979, 10 analyses, p. 83), Nehring and others (1979, 1 partial analysis along with isotopic data), and Benoit and others (1982, 4 analyses, p. 55). Evaluations of the chemical data include estimations of the aquifer temperatures (Nehring and Mariner, 1979; Nehring and others, 1979; Olmsted and others, 1975, p. 217) and a comparison between the Bradys Hot Springs and Desert Peak thermal systems (Benoit and others, 1982, p. 56). The geology of the adjacent Hot Springs Mountains is presented by Hiner (1979), Voegtly (1981), and Benoit and others (1982, p. 54–57). An evaluation of the energy available from the Bradys Hot Springs system has been made by Brook and others (1979, p. 52, 53).

Methods of Investigation

Ground-water sampling for geochemical characterization was the primary field activity during this study. Complementary geologic, geophysical, and hydrologic data also were collected as part of a broader U.S. Geological Survey evaluation of the geothermal system (1981). Much of these data were derived from wells drilled specifically for the hydrologic study. Table 1 presents records of selected wells in the Bradys Hot Springs and nearby Desert Peak areas.

Table 1. Records of selected wells in the Bradys Hot Springs and Desert Peak areas

[—, no data]

Well[1]	Land-surface altitude (ft)	Height of measuring point above land surface (ft)	Depth of screen[2] (ft below land surface)	Casing diameter (in)	Water-level depth below land surface (ft)	
					Jan. 1975	Feb.- April 1981
Bradys Hot Springs area						
AH–1 ----------- 4,073		2.3	69.5–71.5	1.5	6.40	8.18
AH–2 ----------- 4,076		1.9	139.8–141.8	1.5	7.84	9.55
AH–4 ----------- 4,058		2.0	68.6–70.1	1.5	9.65	9.05
AH–5 ----------- 4,104		2.0	101.6–103.1	1.5	44.50	45.16
AH–8 ----------- 4,081		.8	116.1–117.8	1.5	27.59	28.04
AH–9 ----------- 4,052		2.0	59.5–61.0	1.5	5.94	5.12
DH–15 ---------- 4,121		3.8	58.5–60.0	1.5	17.09	34.47
DH–16 ---------- 4,101		5.0	44.7–46.2	1.5	11.75	29.60
DH–17 ---------- 4,075		2.8	143.8–145.3	1.5	10.93	11.79
AH–18 ---------- 4,087		3.0	47.6–49.1	1.5	22.51	23.43
AH–20 ---------- 4,101		2.5	86.7–88.2	1.5	36.42	43.66
M–8 ----------- [3]4,110		—	[4]610–800	—	—	—
B–1 ----------- [3]4,140		—	(1,758)	—	—	—
B–4 ----------- [3]4,120		—	(723)	—	—	—
Desert Peak area						
DH–31 ---------- 4,091		2.5	133.2–134.7	2.0	30.45	31.70
B21–2[5] --------- [3]4,600		—	—	—	—	—

[1]Wells with AH and DH prefixes were augered (AH) or drilled by the mud-rotary technique (DH) by the U.S. Geological Survey. Private well M–8 is a production well at Bradys Hot Springs. Wells B–1 and B–4 are located in the general vicinity of B–8. Well B21–2 was drilled by Phillips Petroleum Company in the Desert Peak geothermal area.

[2]Values in parentheses represent the total depth of the well.

[3]Approximate altitude.

[4]Production depth from Rudisill (1978, p. 219).

[5]Benoit and others (1982, p. 50) indicate that the production is largely from zones more than 2,894 ft below land surface. Total well depth is 3,192 ft.

Geochemical data were collected by obtaining water samples from wells and measuring selected unstable constituents and properties (temperature, pH, carbonate, and bicarbonate) in the field. Before sampling, the wells were pumped or bailed in an attempt to assure that the water samples were representative of the native ground water. Very slow flow to several of the wells did not allow bailing immediately before sampling. These wells (AH–1, –9, and –31) were sampled as much as one month after bailing. The silica values for wells AH–1, –9, and –31 (table 2) may be low as a result of chemical changes within the well during the period between bailing and sampling. The overall conclusions presented here, however, do not depend solely on the analyses from wells AH–1, –9, and –31.

Temperature, pH, and alkalinity were measured in the field. Alkalinity (reported as bicarbonate and carbon-ate) and pH were measured by using the methods described by Wood (1976, p. 12–18). Bicarbonate and carbonate were determined on filtered samples (0.45-μm pore size). The pH measurements were made on unfiltered samples. Filtered, acidified samples were collected for cation analysis, whereas filtered but unacidified samples were collected for anion and isotopic analysis. Samples for silica analysis were filtered and then diluted with deionized water to prevent polymerization where oversaturation with respect to silica was indicated by preliminary field measurements. Samples collected for determination of major and minor constituents were placed in plastic bottles, whereas the isotope samples were placed in glass bottles. Records of selected wells (table 1) and chemical and isotopic data (table 2) for the Bradys Hot Springs area are in the U.S. Geological Survey computer-based storage system WATSTORE. These data, along with

other chemical analyses and ground-water information, may be retrieved by using a latitude/longitude polygon that includes the area represented in figure 5.

Acknowledgments

GEOLOGIC AND HYDROLOGIC SETTING

Bradys Hot Springs is in the western Basin and Range province, on the northwest flank of the Hot Springs Mountains about 50 mi east of Reno, Nev. (fig. 1). The Basin and Range province is characterized by east-west crustal extension; this structural activity has caused normal faulting along with siliceous and basaltic volcanism throughout much of the Hot Springs Mountains and adjacent areas. The surface distribution of geologic units is shown in figure 2. Tertiary sedimentary rocks in the area include sandstone, shale, tuff, diatomite, and limestone (Olmsted and others, 1975, p. 207). Deposits associated with the late Pleistocene Lake Lahontan consist primarily of clay, silt, and sand. Coarse-grained alluvial-fan deposits, stream gravels, and fine-grained playa deposits of Holocene age are all present in the Bradys Hot Springs area (Olmsted and others, 1975, p. 208, 209).

The geothermal activity, localized along a north-trending fault zone of unknown displacement (figs. 2, 3), has affected the surficial geology. Hot-spring and fumarole activity has resulted in the formation of predominately opaline sinter along a main fault and a smaller, subsidiary fault (W.A. Oesterling and R.J. Anctil, Southern Pacific Co., written commun., 1983). Geothermal alteration products consist of "tufa, sinter, salts, silification of rocks, oxidation of metallic elements to orange, yellow, and white deposits, [and] formation of kaolinite" (Olmsted and others, 1975, p. 211). Cinnabar and sulfur deposits have been reported by Bailey and Phoenix (1944, p. 51), and anomalously high mercury concentrations have been found around active steam vents (Garside, 1979).

The climate in the study area and vicinity is arid with diurnal and seasonal temperature extremes. Precipitation, in the form of rain and snow falling directly on the valley floors, generally provides little or no recharge to local ground-water systems. Ground-water recharge to nonthermal ground-water systems in the area is generally attributed to direct infiltration in mountain blocks and to infiltration from streams as they debouch onto alluvial fans. Recharge mechanisms have generally not been well defined for geothermal systems in northwestern Nevada. Recharge to geothermal systems has been attributed to direct infiltration in mountains and recharge to range-bounding faults (Welch and others, 1981). Recent work by Olmsted and others (1984) suggests that recharge to the Soda Lakes and Upsal Hogback geothermal systems in the western Carson Desert (fig. 1) can also be derived from shallow ground water in valley-fill deposits.

Ground water generally moves southwestward in valley-fill deposits of the Bradys Hot Springs area (Harrill, 1970, p. 10). Underflow from the north supplies much of the ground water to the study area (Harrill, 1970, p. 10, pl. 1). This underflow presumably is supplemented by small amounts of recharge from local precipitation in the surrounding mountains. Natural outflow in the valley-fill deposits occurs as underflow to the southwest and evaporation from the surface of the alkali flat (or playa) located to the northwest of Bradys Hot Springs and as transpiration from phreatophytes that surround the playa (Harrill, 1970, p. 10, 19, pl. 1). In the consolidated sedimentary and volcanic rocks, ground water moves primarily through fractures. Vertical fractures locally may allow significant recharge to the geothermal system from overlying valley-fill deposits. Unfortunately, the sites of such recharge cannot be ascertained from the limited hydrologic information available.

The general directions of shallow ground-water movement in the vicinity of Bradys Hot Springs are indicated by the configuration of the water table (fig. 4). Westward movement of thermal ground water that has risen from depth along the thermal fault zone is indicated by the steeply sloping water-table surface west of the fault.

Enhanced vertical permeability due to fracturing is indicated by the correlation between (1) the location of fumarole and spring deposits along the fault and (2) changes in water levels along the fault caused by fluid withdrawal (fig. 5). The decline in water levels is believed to have resulted primarily from the combined discharge from production and flow tests between late 1978 and early 1981. Discharge from one production well began in October 1978 for dehydration of vegetables, and two other wells underwent long-term production tests (9 months and 3 months) in 1980 and early 1981 (Marvin Dommer, Geothermal Food Processors, Inc., oral commun., 1981). According to Dommer, the combined flow during the tests was 1,500 gal/min, as compared to a normal production rate of 750 gal/min. The control of the upflow portion of the system by the local faults was also
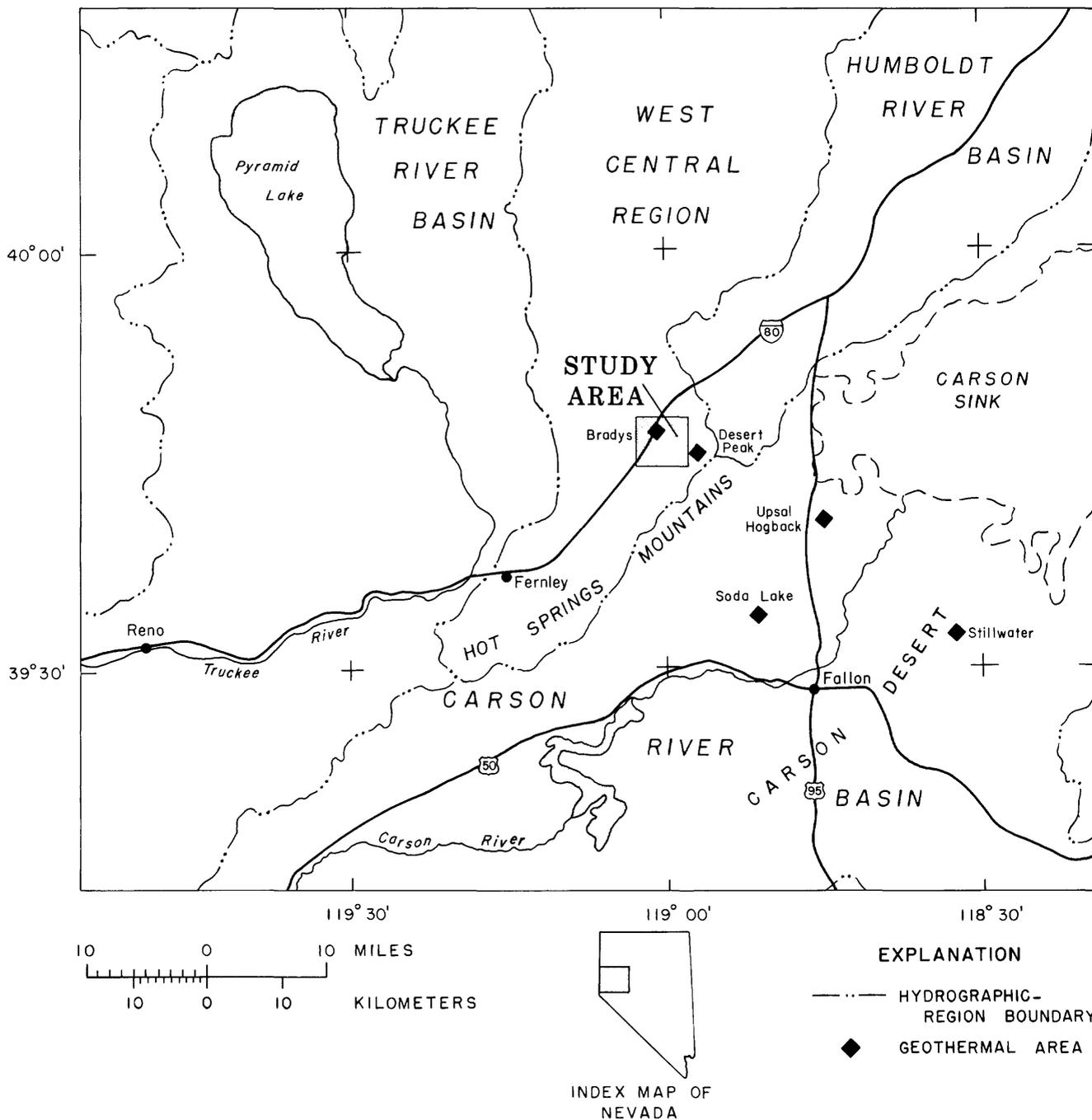
Figure 1. Location of the Bradys Hot Springs and other nearby geothermal areas.

noted by Rudisill (1978), on the basis of hydraulic data collected at wells B–1 and B–4 during a production test in 1977.

The hydrologic relationship between the Bradys Hot Springs area and the Desert Peak geothermal system, about 3 mi to the southeast, is not well understood. Although the water level in one thermal well in the Desert Peak system is more than 200 ft higher than the water level measured in Bradys thermal well DH–16, the equivalent cold-water head in the Desert Peak system is lower

than the 3,900-ft altitude of the western Carson Desert (Frank Yeamans, Phillips Petroleum Co., oral commun., 1983; the equivalent cold-water head was calculated by using a temperature log and adjusting for the density differences caused by temperature). Thus, recharge to either system could be supplied by a system with a hydraulic head of less than 3,900 ft. As discussed later, geochemical considerations imply that saline water (which is commonly present in association with playas) could be the source of recharge to both systems.

119°00'

EXPLANATION

QUATERNARY

Qh
HOT-SPRING SINTER -- Siliceous, white to gray, hard and dense near spring vents, earthy elsewhere

Qal
ALLUVIUM -- Includes Holocene alluvial-fan, stream-channel, and playa deposits, and late Pleistocene Lake Lahontan deposits. Holocene deposits range from unsorted gravel, sand, and clay to very fine-grained clay and chemical precipitates. Includes Lake Lahontan deposits that range from bottom clay and silt to beach-and-bar gravel

Qlt
LACUSTRINE TUFA -- Chemical precipitates of Lake Lahontan consisting of white to gray freshwater calcium carbonate. Coats surfaces of cliffs and forms domes and pinnacles on lowlands. Maximum thickness less than 10 feet

TERTIARY

Tby
YOUNGER BASALT -- Dark brown to black flows consisting of vesicular to dense aphanitic rock containing sparse phenocrysts of plagioclase and olivine

Tt
TRUCKEE FORMATION -- Lacustrine deposits composed chiefly of white diatomite but also containing pumice, tuffaceous sandstone, pebble conglomerate, limestone, coquina, and basalt

Tbo
OLDER BASALT -- Gray to red olivine basalt flows: some dikes, sills, and tuff beds: interbedded white to red silicified lacustrine shale. Basal part of unit includes greenish-gray pillow basalt. Upper part of unit grades upward into Truckee Formation

A⊢——————A' LOCATION OF GEOLOGIC SECTION

———?— — FAULT -- Dashed where approximately located, queried where existence is suspected

R. 26 E.   R. 27 E.
Bose from U.S. Geological Survey
1:62,500 maps
0   1 MILE
0   1   KILOMETER

Figure 2. Generalized geology of the Bradys Hot Springs area. Modified from Voegtly (1981).
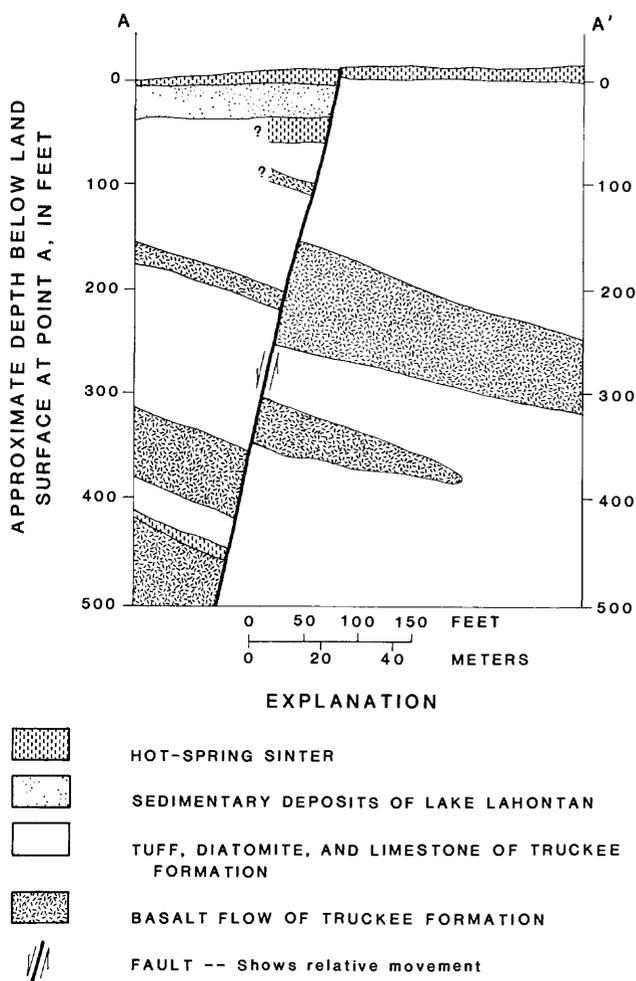
Figure 3. Geologic section across the unnamed fault at Bradys Hot Springs. Modified from Garside and Schilling (1979). Lateral extent of geologic units queried where uncertain.

AQUEOUS GEOCHEMISTRY

Major Constituents

Thermal water at Bradys Hot Springs contains about 2,600 mg/L of dissolved solids dominated by sodium, chloride, and sulfate, as indicated by the analyses of water from well M–8 (table 2). Chloride and sodium concentrations in the vicinity of Bradys Hot Springs[1] have narrow ranges of 910–1,100 mg/L and 660–850 mg/L, respectively, excluding the data for well AH–9. The higher values at well AH–9 (1,600 and 1,100 mg/L, respectively) may be a result of evaporative concentration of water moving downgradient from the hot area (fig. 4).

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[1]Among wells listed in table 2, all but DH–31 and B21–2 are considered to be in the immediate vicinity of the hot-spring area (see fig. 4).

A modified version of the mineral-solution equilibrium program SOLMNEQ (Kharaka and Barnes, 1973) was used to evaluate the possible control exerted by mineral solubility on aqueous species in ground water of the study area. The program, which was supplied by Yousif K. Kharaka (U.S. Geological Survey, written commun., 1980) allows the user to calculate a pH at an elevated temperature by using a mass-balance assumption for the total H^+, as described by Kharaka and Mariner (1977, p. 72). The program also incorporates more recently developed methods for estimating aquifer temperatures. The mass-balance option was used to calculate the solution-mineral equilibria for each analysis at the measured down-hole temperature (this was required because the pH was measured at a lower temperature at land surface) and at the aquifer temperature indicated by the magnesium-corrected sodium-potassium-calcium geothermometer of Fournier and Potter (1979). [This geothermometer was selected because it is the least sensitive to mixing with nonthermal water (Fournier, 1977, p. 46).] However, some of the calculated temperatures (particularly at well AH–9) are probably too high. The results are included, nonetheless, because the calculations are useful in evaluating the effect of changes in temperature on mineral solubility. Following the suggestion of Paces (1972, p. 225, 226), if the value of log (AP/K) [the logarithm of the ratio of the activity product (AP) to the equilibrium constant (K) for a particular mineral] is between -0.5 and $+0.5$, then the water is regarded as being in equilibrium with the mineral considered. The activity product is obtained by multiplying the activities of the chemical species that form a particular mineral. For instance, the activity product for calcite ($CaCO_3$) is $[Ca^{2+}] \times [CO_3^{2-}]$, where the square brackets signify the activities of the component species.

The water in the vicinity of Bradys Hot Springs appears to be saturated with respect to calcite. As shown in figure 6, the samples are either near saturation [log (AP/K) between -0.5 and $+0.5$] or slightly oversaturated (greater than $+0.5$) with respect to calcite. A possible contributor to the total carbonate ($H_2CO_3 + HCO_3^- + CO_3^{2-}$) is CO_2 derived from the deep crust or upper mantle, as proposed for many tectonically active areas of the world by Barnes and others (1978). Part of the difference in pH between the water at producing well M–8 and at the other wells may be the result of greater CO_2 outgassing at the other wells. The pH of the water from well M–8 is probably affected by CO_2 outgassing before sampling as calcite precipitation occurs in the well (Garside and Schilling, 1979, p. 7).

The sulfate data show a wide range of values, but no mineral-equilibrium control is obvious. The sulfate concentration in the thermal water, as represented by the analysis for M–8, may be due in part to oxidation of sulfide minerals—a process that can generate hydrogen ions

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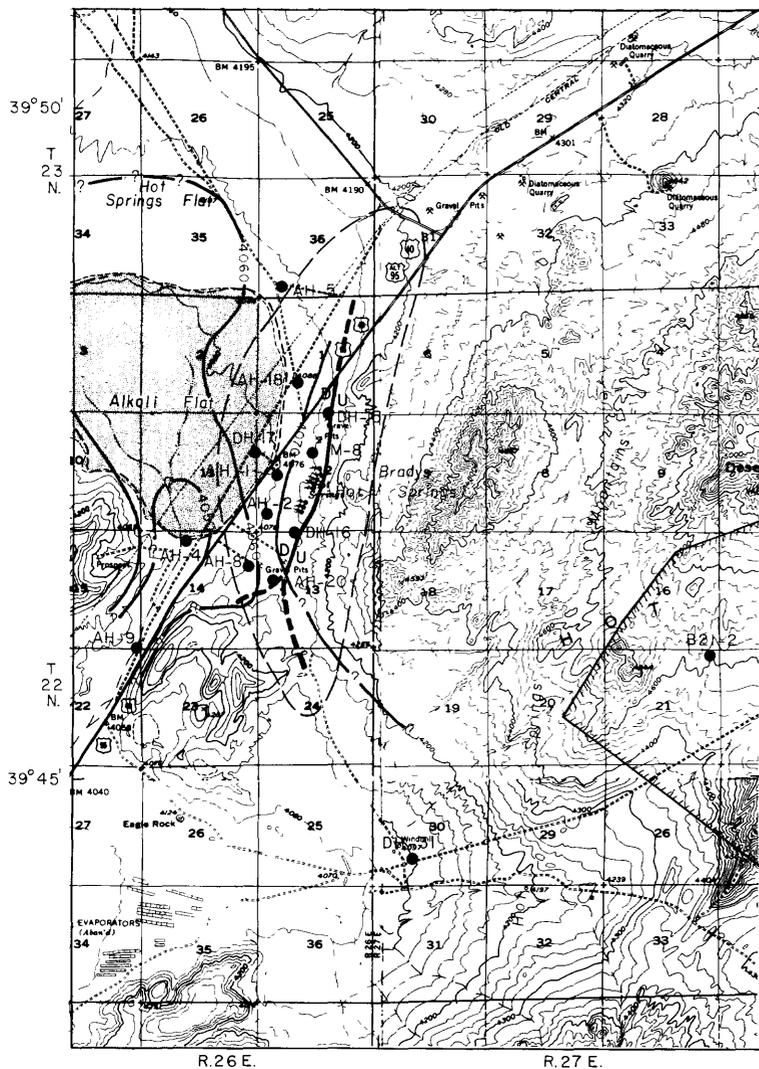
EXPLANATION

U / D --- FAULT -- Dashed where approximately located. U, upthrown side, D, downthrown side. From Voegtly (1981, plate 1)

—4060—? WATER-TABLE CONTOUR -- Shows altitude, in feet above sea level. Approximately located; queried where location is inferred. Contour interval 10 feet. Datum is sea level. From A. S. Van Denburgh (U.S. Geological Survey, written commun., 1983); based on data from Olmsted and others (1975, table 26) and Harrill (1970, table 11 and plate 1)

—— —— APPROXIMATE LIMIT OF THERMAL ANOMALY -- Based on a temperature of 30° C at depth of about 100 feet below land surface. From Olmsted and others (1975)

⊔⊔⊔⊔⊔⊔⊔ POSSIBLE LIMIT OF THE DESERT GEOTHERMAL SYSTEM -- From Benoit and others (1982, p. 70)

—————— PLAYA BOUNDARY

●AH-5 WELL, WITH IDENTIFICATION SYMBOL

⌣ SPRING

Base from U.S. Geological Survey 1:62,000
Desert Peak, 1951, Fireball Ridge, 1957,
Soda Lake, 1951, and Two Tips, 1957

0   1   2 MILES
0 1 2 3 KILOMETERS
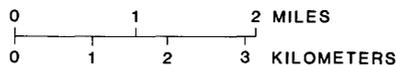
Topographic contour intervals 20 and 40 feet

Figure 4. Approximate water-table altitude, 1969–73, and locations of wells.

(Hem, 1970, p. 162)—or to dissolution of gypsum in valley-fill or playa deposits.

The relationship between the Bradys Hot Springs and Desert Peak geothermal systems is considered by Benoit and others (1982). The thermal water of the Desert Peak system, as represented by the analysis of a sample from well B21-2, has a much higher chloride concentration than water of the Bradys Hot Springs system. [Data for well B21-2 are used for comparison because the sampling technique at this well was superior to that at the other two deep production wells in the Desert Peak system (Benoit and others, 1982, p. 54).] The difference in chloride concentrations indicates that the two systems do not have a common deep source of water unless it is modified in one or both systems.

The chloride concentration (4,200 mg/L) in water from well DH-31 is consistent with the proposal by Benoit and others (1982, p. 56) that some of the flow from the Desert Peak system moves to the west, with eventual discharge in the vicinity of a non-flowing dug well (well S34, T22N, R26E in table 5 of Benoit and others, 1982). At well DH-31, which is located between well B21-2 and the dug well discussed by Benoit and others (1982), the chloride concentration (4,200 mg/L) is
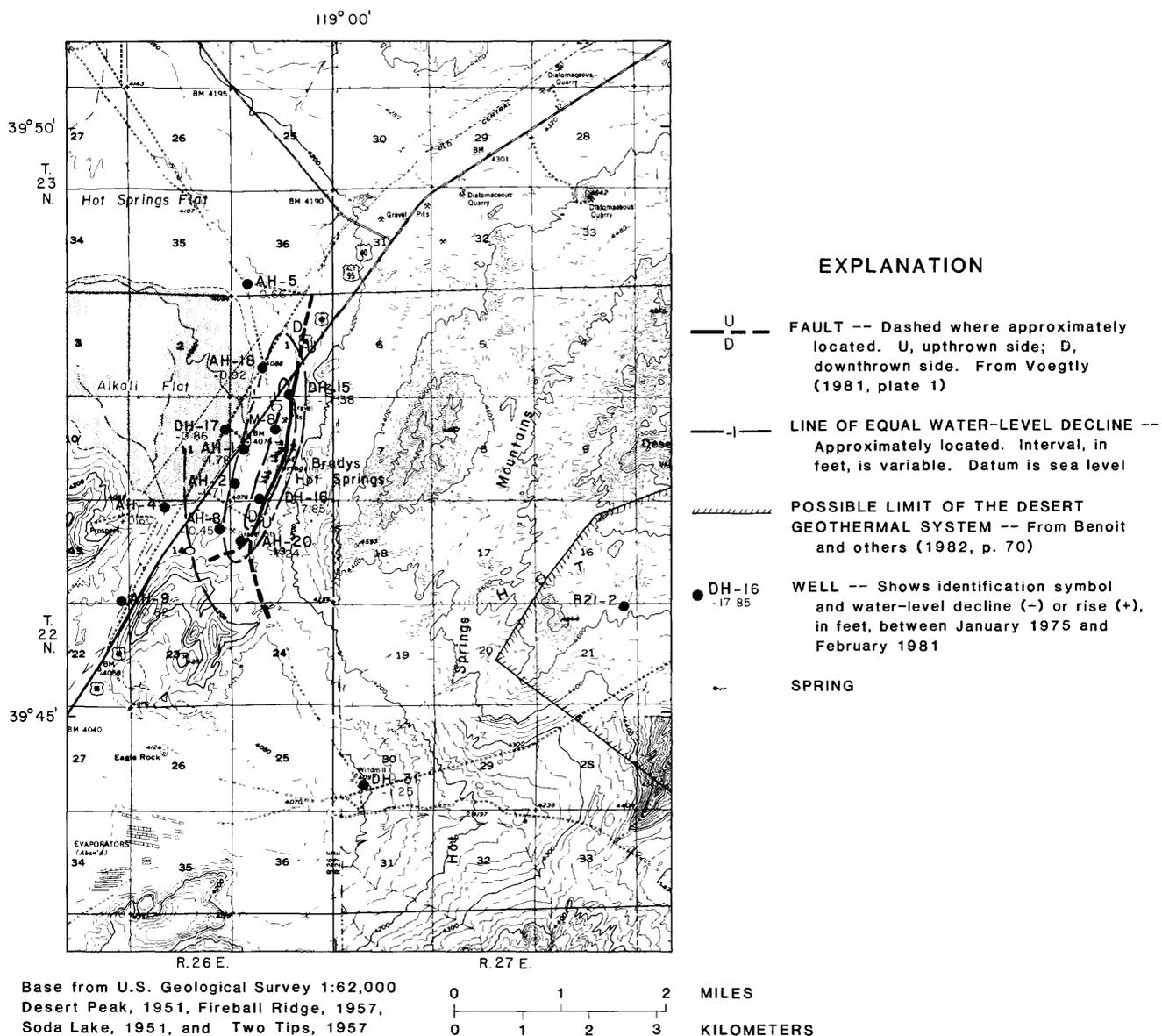
Figure 5. Changes in ground-water level between January 1975 and February 1981.

greater than that in water from well B21–2 (about 3,600 mg/L) and is therefore consistent with some vapor loss from the rising hot water in the Desert Peak system. The low silica at well DH–31 (17 mg/L) in comparison to B21–2 (350 mg/L) is presumably a result of silica precipitation due to cooling. The difference in sulfate of 7.0 mg/L at DH–31 compared to 90 mg/L at B21–2 is more difficult to explain but may, at least in part, be due to loss of H_2S after conversion of sulfate to sulfide. The ground-water flow direction indicated by water-level contours presented by Hiner (1979, p. 64) is also consistent with some of the outflow from the Desert Peak system moving to and past well DH–31.

Minor Constituents

Some minor constituents can be used as indicators of thermal fluids or are of environmental concern. Minor constituents examined in this study are barium, boron, bromide, fluoride, lead, lithium, manganese, and strontium. The concentrations of some of these constituents in the Bradys system appear to be controlled primarily by either evaporative concentration or the solubility of a solid phase (mineral). Bromide is largely controlled by evaporative concentration, whereas barium, fluoride, lead, and manganese appear to be controlled by solubility
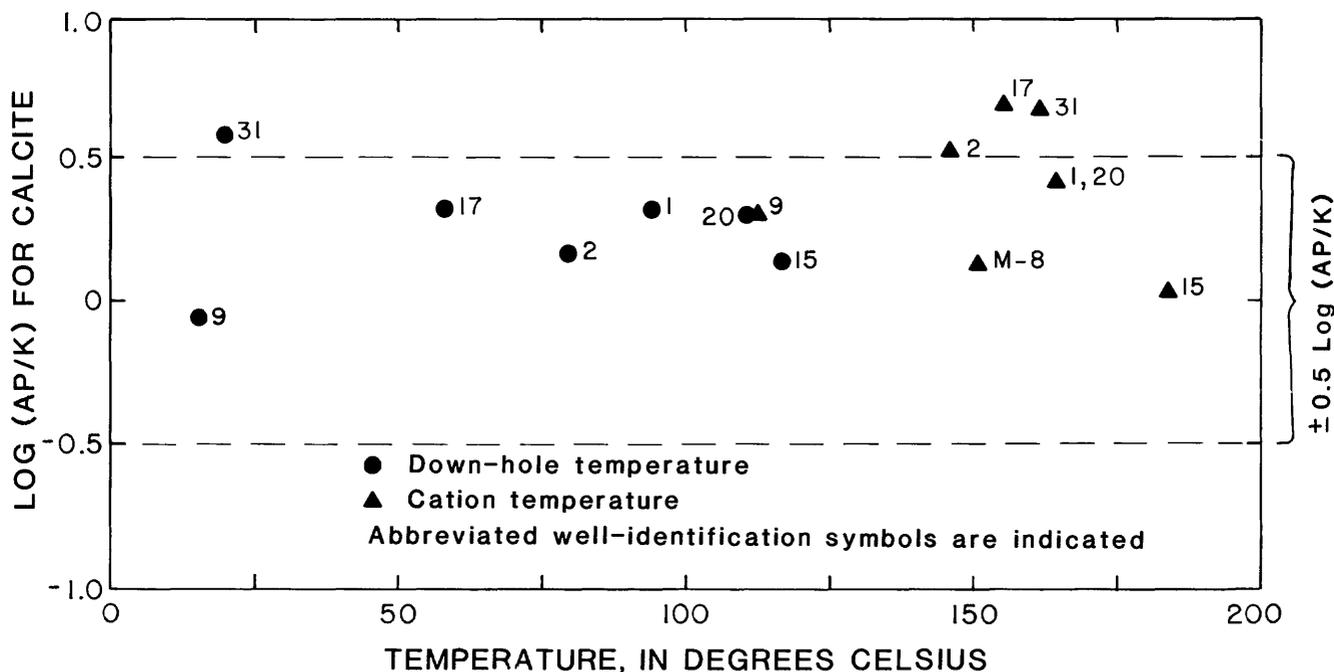
Figure 6. Values of log (AP/K) for calcite in well water at the down-hole temperature and at the magnesium-corrected Na-K-Ca geothermometer (cation) temperature (Fournier and Potter, 1979).

of the minerals barite, fluorite, cerussite, and rhodochrosite, respectively.

Barium concentrations at low temperatures may be controlled by barite $(BaSO_4)$. Figure 7A indicates that all the samples collected from sources with temperatures less than 75 °C are near equilibrium with respect to barite. It is not known why water with source temperatures greater than 75 °C is undersaturated.

The fluoride concentrations appear to be controlled by the solubility of fluorite (CaF_2). Using the fluorite equilibrium relationship proposed by Nordstrom and Jenne (1977) and the program SOLMNEQ, we calculated a value for log (AP/K), which is plotted in figure 7B for each analysis included in this study. The log (AP/K) values calculated for fluorite at the measured, down-hole temperatures are all within the range from -0.5 to $+0.5$, indicating mineral-water equilibrium. This conclusion agrees with that of Mahon (1964) that fluorite is the controlling phase in thermal water. Using revised thermodynamic data, Nordstrom and Jenne (1977) concluded that fluorite is the controlling phase for fluoride in many thermal waters of the Western United States.

The three well waters that contain measurable lead concentrations (wells AH–1, AH–20, and DH–15) appear to be near saturation (saturation index range from -1.54 to 0.37) with respect to the carbonate mineral cerussite

$(PbCO_3)$. Only the divalent oxidation state (Pb^{2+}) was considered in the solubility calculations for lead. This approach is consistent with Hem's observation (1970, p. 205) that "over most of the water-stability region, the oxidation state of lead at equilibrium is $[Pb^{2+}]$ * * *."

Solubility of rhodochrosite $(MnCO_3)$ may be the controlling factor for manganese concentrations in some parts of the study area (fig. 7C). Both of the most saline samples collected during this study (from wells AH–9 and DH–31) appear to be saturated with respect to rhodochrosite. Among the warmer well waters, those at the downgradient sites (wells AH–1, AH–2, and DH–17) appear to be controlled by rhodochrosite. In the solubility calculations, only the $+2$ valence state was considered for manganese. A dominance diagram presented by Hem (1970, p. 128) indicates that Mn^{2+} is the predominant aqueous species at redox potentials reasonable for the reducing to slightly oxidizing conditions probably present in the thermal waters being considered here.

The concentrations of bromide are strongly correlated with the chloride concentrations (fig. 8B) which, in turn, appear on the basis of the stable-isotope data to be largely controlled by evaporative concentration.

Boron, lithium, and strontium may be indicators of the presence of thermal water. Figure 8A, C, and D show that the lower temperature water from wells AH–4,

Table 2. Chemical and isotopic data for well waters in the Bradys Hot Springs and Desert Peak areas

[Results in milligrams per liter except as noted. All analyses by the U.S. Geological Survey Central Laboratory, Arvada, Colo., except as noted; --, no data measurement]

Name of well	Date of sample (yr–mo–day)	Temper-ature, water (°C)[1]	pH, field	Dis-solved solids	Silica, dis-solved (as SiO$_2$)	Calcium, dis-solved (as Ca)	Magne-sium, dis-solved (as Mg)	Sodium, dis-solved (as Na)	Potas-sium, dis-solved (as K)	Bicar-bonate, field (as HCO$_3$)	Car-bonate, field (as CO$_3$)	Sulfate, dis-solved (as SO$_4$)	Chlo-ride, dis-solved (as Cl)
					Bradys Hot Springs area								
AH–1 --	1981–03–11	94.5	8.2	1,900	32	47	2.2	690	47	77	1	160	910
AH–2 --	1981–03–12	79.8	7.7	2,000	99	48	2.3	700	30	120	0	150	910
AH–4 --	1981–03–14	20.4	—	—	—	—	—	690	40	—	—	—	990
AH–5 --	1981–03–19	26.7	—	—	—	—	—	660	31	—	—	—	940
AH–8 --	1981–03–10	36.2	—	—	—	—	—	690	49	—	—	—	940
AH–9 --	1981–03–16	15.4	8.3	2,900	12	56	6.9	1,100	39	62	1	65	1,600
DH–15 -	1981–04–02	117	8.5	2,100	170	20	.2	720	53	66	1	170	930
DH–17 -	1981–02–20	58.0	8.0	2,300	110	56	2.6	780	42	170	1	67	1,100
AH–20 -	1981–03–17	111	7.9	2,100	120	46	1.4	670	39	86	0	150	980
M–8[3] ---	1979–07–06	>100	6.8	2,600	164	45	.3	850	36	111	0	320	1,100
					Desert Peak area								
DH–31 -	1981–02–16	20.2	8.3	6,700	17	130	9.8	2,100	160	190	2	7.0	4,200
B21–2[4]	1978–03	160	—	6,800	350	100	<1	2,250	250	50	--	98	3,700
--------	1978–03	160	7.0	—	—	84	0.1	2,200	219	34	--	88	3,500

–5, and –9 generally plots below or to the right of the higher temperature water. This suggests that the higher boron, lithium, and strontium concentrations in the higher temperature samples may be a result of reactions occurring in the deep thermal aquifer. This conclusion is considered very tentative due to the very limited amount of data for nonthermal water.

Stable Isotopes[2]

With few exceptions, the stable-isotope composition of water from most geothermal systems has been found to reflect recharge of local meteoric water and modification by one or more physical or chemical processes (Truesdell, 1976, p. lx). The processes that can modify the initial isotopic composition are (1) exchange of oxygen with wall rocks, (2) mixing with cooler, local ground water, and (3) evaporation, either before recharge or in the upflow portion of a geothermal system.

The isotopic effect of exchange of oxygen with wall rocks, termed an "oxygen shift," is toward a heavier (less negative) value, as indicated in figure 9. Mixing of thermal water with cooler, local ground water will pro-

duce an isotopic composition intermediate between the two end members, with the exact composition being controlled by the mixing proportions. Evaporation results in a change in composition that is indicated in figure 9. The slope of the evaporation trend is controlled largely by the evaporation temperature.

The isotopic and chemical composition of thermal water in the Bradys Hot Springs system can be explained by the following sequence:
1. Meteoric water recharging a shallow ground-water system,
2. Evaporation affecting the shallow ground water,
3. Recharge to the thermal aquifer by ground water from the shallow system, and
4. Flow through the thermal aquifer, with discharge localized by faulting.

The isotopic composition of recharge to the local ground-water system (which would presumably be similar to that of the recharge to the thermal aquifer) could not be directly determined owing to a lack of wells and springs in local mountains or on the upper parts of alluvial fans (the assumed recharge areas). The estimated isotopic composition of recharge to the Bradys Hot Springs system, before it is affected by evaporation, is represented in figure 9 by field A, at the intersection of the arid-zone meteoric line and the local evaporation trend. The standard meteoric-water line in figure 9 generally

[2]The stable isotopes evaluated herein are oxygen-18, relative to oxygen-16, and deuterium (hydrogen-2), relative to hydrogen-1.

Table 2. Chemical and isotopic data for well waters in the Bradys Hot Springs and Desert Peak areas—Continued

Fluoride, dissolved (as F)	Bromide, dissolved (as Br)	Barium, dissolved (μg/L as Ba)	Boron, dissolved (μg/L as B)	Lead, dissolved (μg/L as Pb)	Manganese, dissolved (μg/L as Mn)	Strontium, dissolved (μg/L as Sr)	Zinc, dissolved (μg/L as Zn)	Lithium, dissolved (μg/L as Li)	Deuterium (D, in per mil)[2]	Oxygen-18 ($\delta^{18}O$, in per mil)[2]
					Bradys Hot Springs area					
6.6	1.4	160	3,400	93	150	1,700	800	1,600	−125.0	−14.65
5.0	1.5	110	4,100	<20	61	1,200	160	1,700	−126.0	−14.40
4.8	2.0	—	3,600	—	—	—	—	1,100	−127.0	−14.35
4.4	1.7	—	3,700	—	—	—	—	970	−123.0	−14.20
5.2	1.7	—	2,800	—	—	—	—	1,600	−123.0	−14.30
3.0	2.8	110	4,400	<30	830	940	1,500	1,100	−122.5	−14.00
6.2	1.6	60	4,400	56	18	1,300	100	1,800	−124.0	−14.20
2.9	1.6	240	4,300	<20	71	2,000	740	2,000	−125.5	−14.30
5.2	1.6	150	4,200	27	110	2,200	310	1,800	−125.5	−14.35
5.8	—	—	5,200	—	—	1,600	—	1,500	−121.2	−14.20
					Desert Peak area					
3.1	8.0	—	16,000	—	590	5,600	—	3,300	−113.5	−12.20
—	—	—	16,000	—	—	—	—	1,400	(5)	(5)
—	—	—	15,100	—	—	—	—	—	(5)	(5)

[1]For U.S. Geological Survey wells (all wells except M–8, and B21–2) measurements were made at midpoint of screen, winter 1975–76, by Franklin H. Olmsted, U.S. Geological Survey.

[2]Analyses were made under the supervision of Tyler B. Coplen, U.S. Geological Survey, Reston, Va. The δ values are defined as

$\delta = [(R_x - 1)/R_{SMOW}] \times 1,000$, where R_x is the isotope ratio (deuterium/hydrogen-1 or oxygen-18/oxygen-16) of the sample and R_{SMOW} is the ratio for standard mean ocean water. A negative δ value indicates that the sample is isotopically lighter than the standard.

[3]Data from Robert H. Mariner (U.S. Geological Survey, written commun., 1981). Sample collected and processed in the same manner as described by Mariner and others (1974, 1975). An analysis of the gas phase indicates the following composition, in percent: CH_4 = 2.63, CO_2 = 2.48, H_2 = 2.94, H_2S = <0.01, O_2 = <0.1, Ar = 1.31, N_2 = 90.14, C_2H_6 = 0.03, He = 0.01, SO_2 = <0.05. This well was producing steam and water when sampled. Although no temperature was recorded during the sampling, it was doubtless well above 100 °C in the production zone. The pH was measured at 24 °C on a conductively cooled sample.

[4]From Benoit and others (1982, table 5; analyses by two laboratories, presumably on near-duplicate samples).

[5]Isotopic analyses (Benoit and others, 1982, table 7): March 16, 1977, δD = −117.8, $\delta^{18}O$ = −12.52, and δD = −115.1, $\delta^{18}O$ = −12.8 (two laboratories); March 17, 1977, δD = −114.6, $\delta^{18}O$ = −12.5; Nov. 5, 1977, δD = −116.0, $\delta^{18}O$ = −12.9; and Nov. 12, 1977, δD = −115.6.

represents a more humid climate; the arid-zone line provides a better estimate of recharge water in a climatic setting like that of the study area (Dansgaard, 1964).

The deuterium concentration in the water at the Bradys Hot Springs system appears to be a result of evaporation of shallow ground water as described by the Rayleigh distillation equation (Dansgaard, 1964). Further shallow evaporation of this same recharge could then form water that is the source of recharge to the Desert Peak system. Fractionation factors for deuterium calculated by using the method of Friedman and others (1976), which is based on the Rayleigh equation and the use of a conservative constituent, are very similar for the Bradys Hot Springs and Desert Peak systems. The values shown in table 3 were calculated on the basis of postulated

recharge compositions and the assumption that chloride remains conservative during evaporation.

Although the fractionation factors (table 3) vary depending on the compositions used for the recharge, values calculated for the two systems are similar. Fractionation factors calculated for ground-water systems that have been affected by shallow evaporation in the Carson (Olmsted and others, 1984) and Black Rock Deserts (located about 100 miles north of Bradys Hot Springs, unpublished USGS data) are similar to those presented here, which indicates that values in the 1.002-to-1.005 range are not unreasonable for evaporation in ground water. This similarity of fractionation factors for the thermal and nonthermal water is taken as evidence that the same process (evaporation of shallow ground water) is primarily
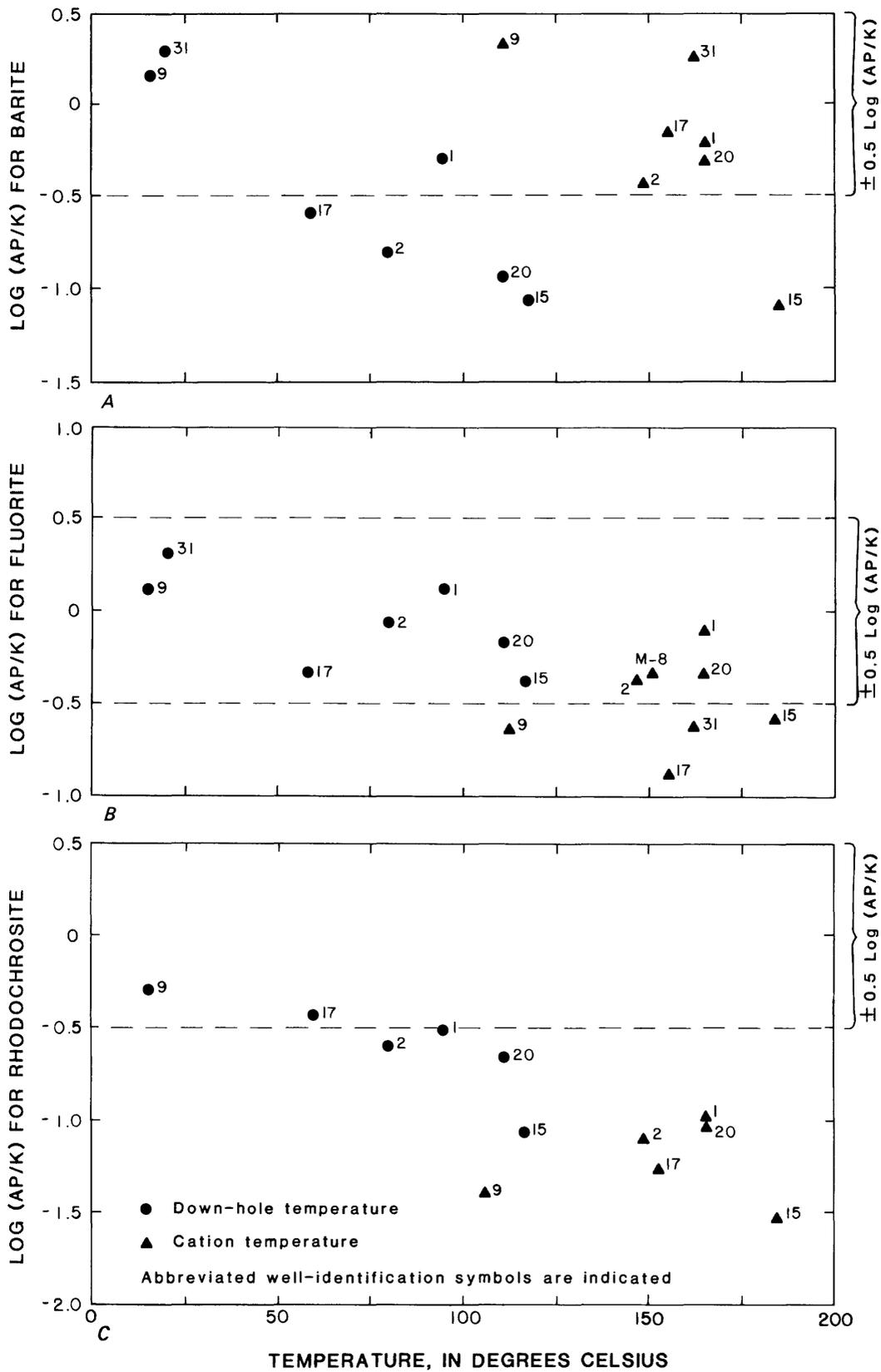
Figure 7. Values of log (AP/K) in well waters for (A) barite, (B) fluorite, and (C) rhodochrosite at the down-hole temperature and at the magnesium-corrected Na-K-Ca geothermometer (cation) temperature (Fournier and Potter, 1979).

Table 3. Apparent fractionation factors for deuterium based on estimated recharge chloride concentrations and isotope composition

	Possible recharge compositions	
Recharge alternative	Chloride (mg/L)	Delta deuterium (per mil)
1	10	-128
2	50	-130
3	50	-135

	Fractionation factors		
Well	Recharge alternative		
	1	2	3
Bradys Hot Springs well M-8	1.0016	1.0032	1.0051
Desert Peak well B21-2	1.0023	1.0038	1.0051

responsible for the increased deuterium and chloride concentrations. Although agreement between the calculated values and equilibrium laboratory determinations (Friedman and O'Neil, 1977) is poor, the similarity may indicate a single evolution path from recharge to Bradys Hot Springs to Desert Peak water. Thus, evaporation of a single meteoric water source may be the process responsible for the chloride concentrations and isotopic compositions of both the Bradys and Desert Peak systems.

The actual recharge area for the Bradys system is difficult to identify on the basis of the present information. One possible source of ground water that has been affected by evaporation is the western Carson Desert, east of the Hot Springs Mountains. This alternative is not supported by isotopic compositions indicated by recent sampling there by Olmsted and others (1984), however. As indicated in figure 9, ground water in the western Carson Desert is isotopically heavier than that seen in the present study area and, therefore, is not a good candidate for recharge.

Mixing of local meteoric water (with a composition as indicated by field A in fig. 9) with water from Desert Peak (field C) to form the observed composition at Bradys Hot Springs (field B) does not appear to be a viable possibility either. A mixture of about two parts Desert Peak water with one part meteoric water would form the approximate deuterium composition at well M-8, but the mixture would have a much higher chloride concentration (about 1,800 mg/L) than that observed (1,100 mg/L).

Accepting the arid-zone precipitation line as a good approximation of local rainfall, then if there is an oxygen shift at Bradys Hot Springs it is not more than 2 per mil. If evaporation is responsible for some of the departure from the initial isotopic composition (as discussed earlier), then the oxygen shift would be even less. A shift of

2 per mil is relatively small compared with that documented for some other thermal systems. The small shift may be due to a high ratio of water to wall-rock surface area, such as in a fractured-rock system. This is consistent with the obvious fault control present in the upflow part of the Bradys system. The small oxygen shift could also be related, at least in part, to a short fluid residence time, a modest deep-aquifer temperature, or both.

The water sampled at well DH-31 appears to have moved southwest from the Desert Peak system. Isotopes and chloride are consistent with an equilibrium fractionation factor for deuterium of about 1.04 (calculated by using average chloride and delta deuterium values for B21-2), which agrees with the experimentally determined value of 1.04 at about 75 °C (Friedman and O'Neil, 1977). [This fractionation factor also was calculated assuming that chloride is conservative in the aqueous phase, following the procedure presented by Friedman and others (1976).]

Geothermometry

Methods of estimating the temperature of a geothermal water have been developed that utilize both chemical and isotopic compositions. The most widely accepted methods, which are known as chemical geothermometers, are those based on silica concentrations and cation proportions. Estimates of aquifer temperature calculated by using several of these methods are listed in table 4.

Despite fairly good internal consistency among cation and silica geothermometer estimates for several well waters in the Bradys Hot Springs area, the estimates are all significantly lower than the highest reported temperature measurement. Garside and Schilling (1979, p. 147) indicate that a maximum down-hole temperature of 212 °C was measured in one well at Bradys. This is substantially higher than the Na-K-Ca (sodium-potassium-calcium) geothermometer estimate of 190 °C for well DH-15—the highest estimate listed in table 3 for the Bradys Hot Springs system. The difference may be a result of one or more of the following factors:

1. An inaccurately high reported temperature. The second highest down-hole temperature reported by Garside and Schilling (1979) is 188 °C, which is in much better agreement with the geothermometer estimates.
2. Mixing with nonthermal water.
3. Partial reequilibration of silica and cations, resulting in lower geothermometer estimates.
4. The possibility that 212 °C water was not sampled during this study. For water from well M-8 (probably the most reliable source of thermal water listed in table 4), the estimates range from 121 to 190 °C. The highest value, which is based on the gas geothermometer of D'Amore and Panichi (1980), is significantly higher than the other six estimates for well M-8.
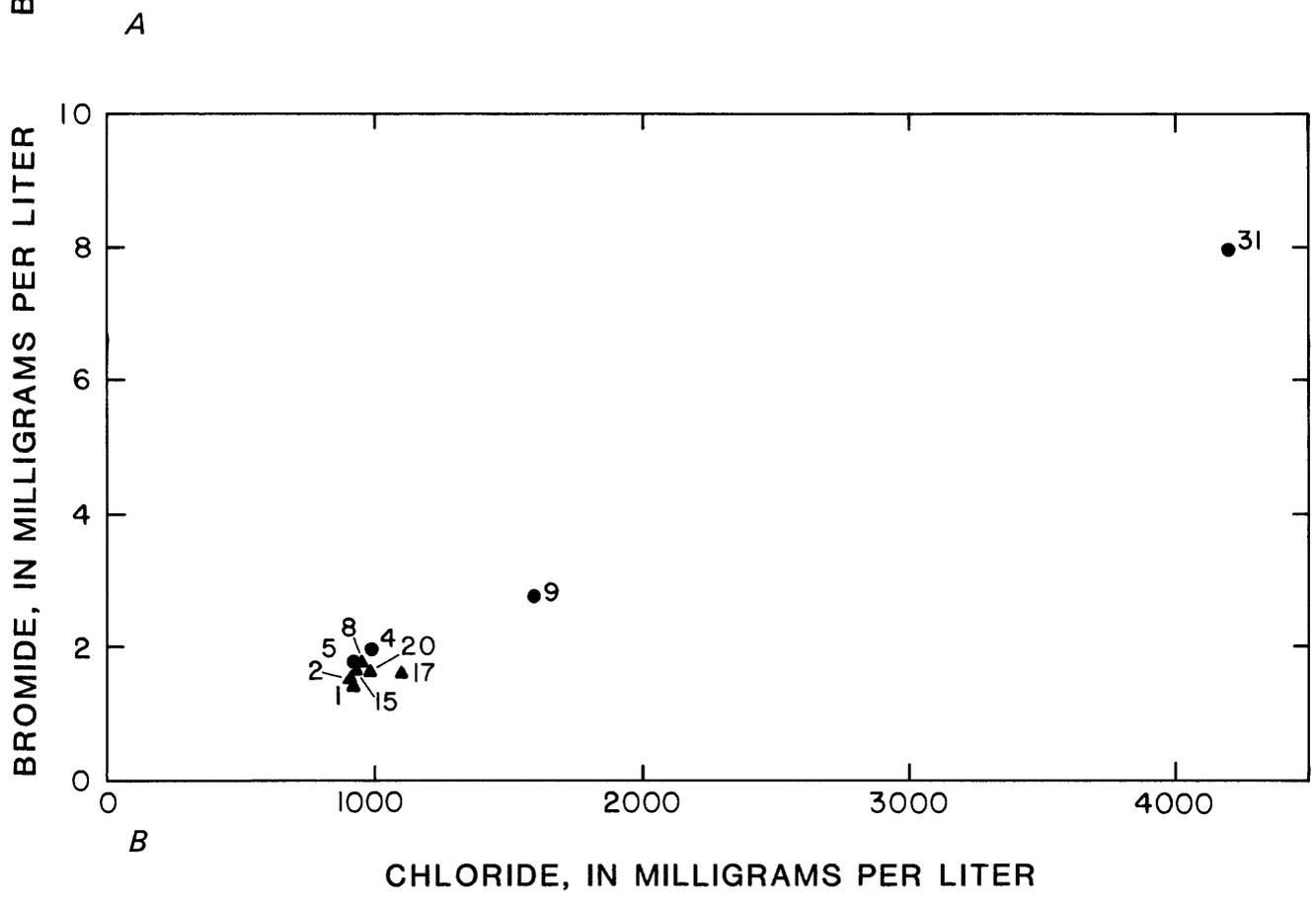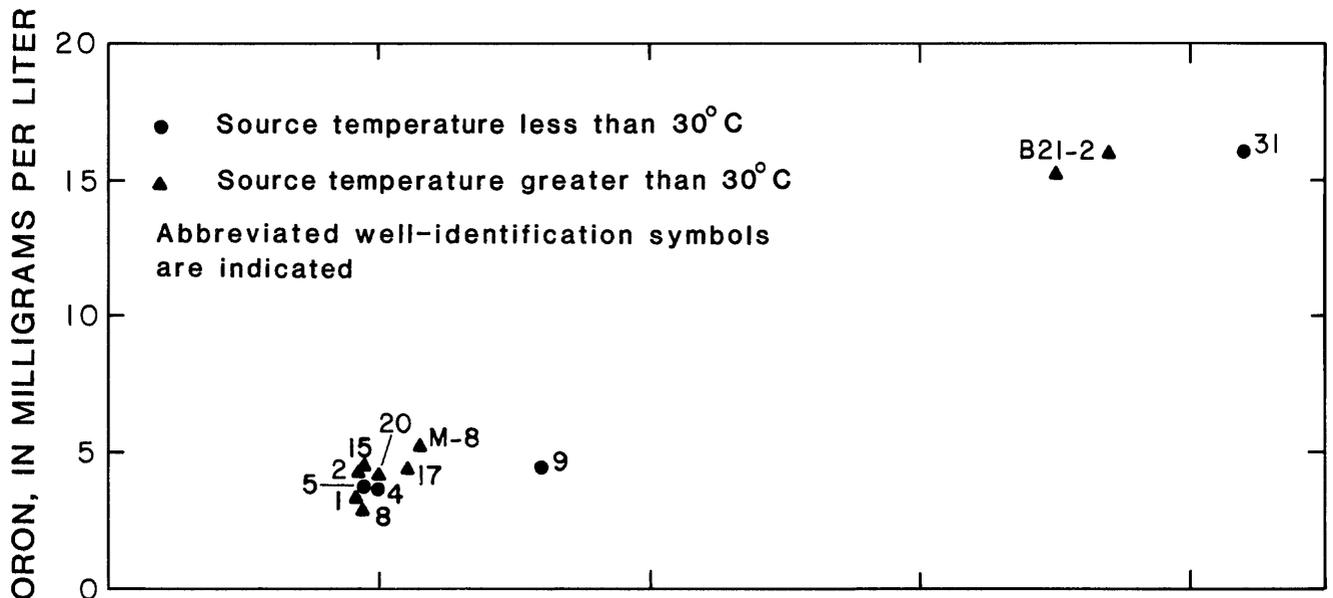
Figure 8. Relation between chloride and (*A*) boron, (*B*) bromide, (*C*) lithium, and (*D*) strontium in well waters.
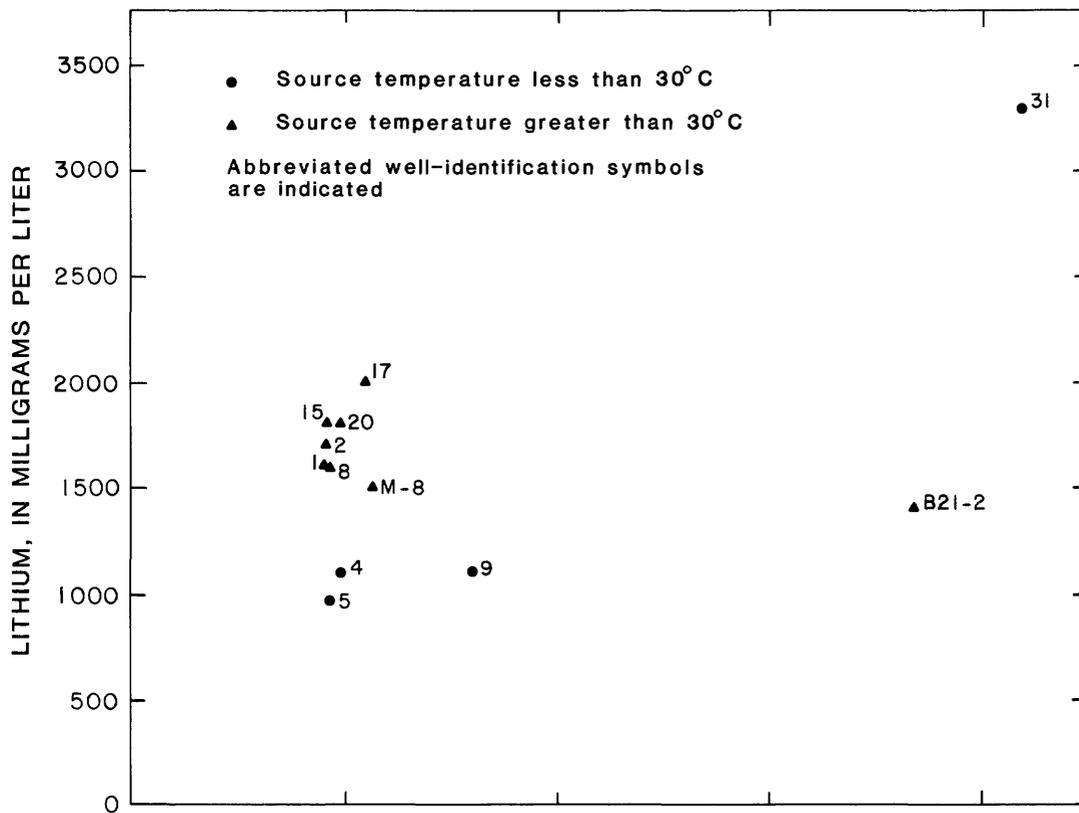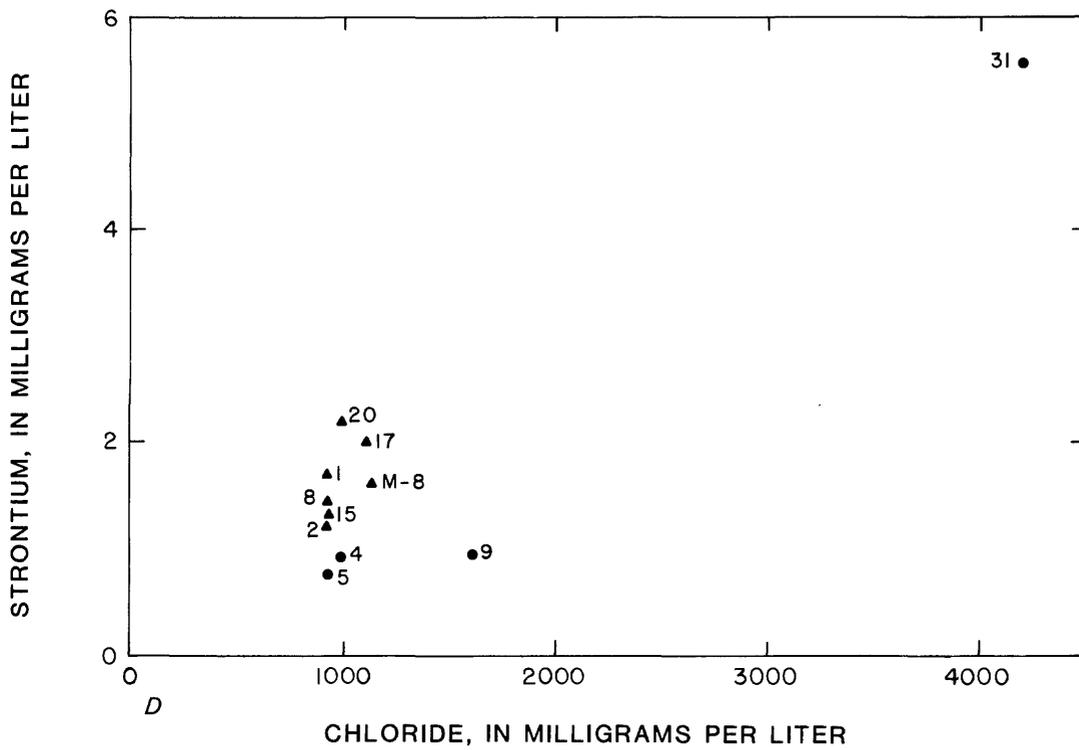
Figure 8. Continued.
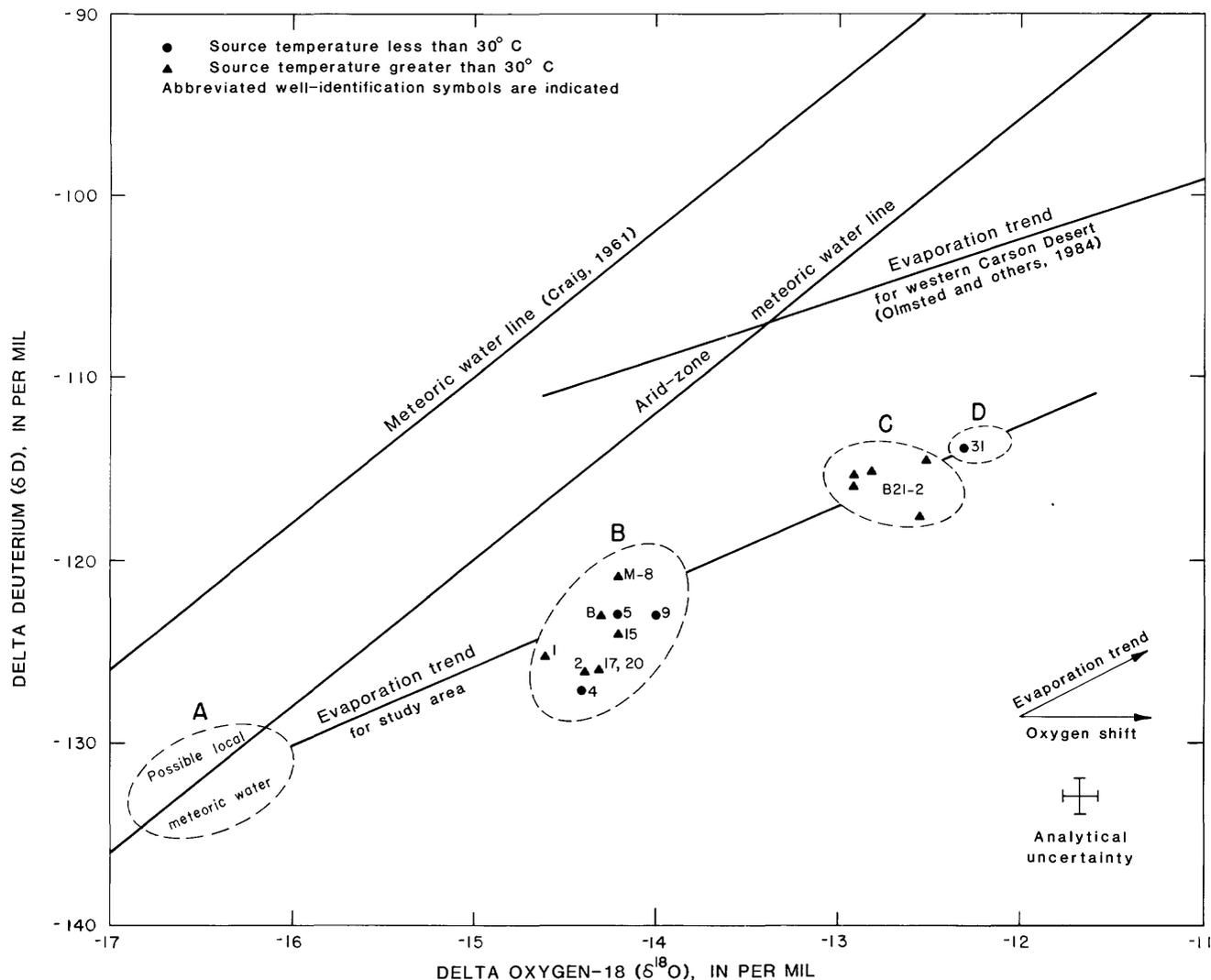
Figure 9. Stable-isotope composition of water in the study area. Dashed circles delineate ground-water fields having the following concentrations of chloride: A, assumed less than 50 mg/L; B, 910–1,600 mg/L; C, 3,500–3,700 mg/L; and D, 4,200 mg/L. Chloride concentrations and well locations given in table 2. Field B is Bradys Hot Springs area; fields C and D are Desert Peak area.

One technique for evaluating the internal consistency of temperature estimates, as discussed by Fournier and others (1979, p. 18–21), is to use a graphical comparison such as that shown in figure 10. The two geothermometer methods selected for comparison—conductive quartz and magnesium-corrected Na-K-Ca—are believed to be the best among the several silica and cation techniques. The conductive-quartz method was selected because the water at the sampling sites is believed to have been cooled conductively rather than adiabatically.

In general, agreement between the silica and cation values is fairly good for Bradys-area samples from wells with down-hole temperatures greater than 50 °C. The agreement is best for wells M–8, AH–2, DH–15, DH–17, and AH–20. On the basis of the internal consistency of the estimates for these five wells, a "best" estimate for

the deep-aquifer temperature is a range from 137 to 184 °C. The poor agreement of this estimate with the reported down-hole measurement of 212 °C is difficult to reconcile but may be related to any of the four factors discussed earlier. Sampling of deeper wells and additional down-hole measurements would probably aid in finding the reason for this discrepancy.

The temperature estimates obtained for wells AH–1 and AH–9 probably represent either mixed or nonthermal water, although precipitation of silica (due in part to the unavoidable delay in sampling, discussed earlier) may be partly responsible for the poor agreement for the water at wells AH–1 and AH–9. The sampling problem is probably also, at least in part, responsible for the low temperatures estimated by using the silica geothermometer.

Table 4. Aquifer-temperature estimates for the Bradys Hot Springs and Desert Peak geothermal areas, based on chemical geothermometers

[Temperatures in degrees Celsius].

Well	Down-hole temperature[1]	Geothermometer method					Comments
		Silica[2]		Cation			
		Quartz, conductive	Quartz, adiabatic	Na-K[2]	Na-K-Ca[2]	Na-K-Ca, Mg-corrected[3]	
Bradys Hot Springs area							
AH-1 .... 94.5		82	85	149	177	165	Thermal or possibly mixed water.
AH-2 ....79.8		137	132	122	154	147	Thermal water.
AH-9 .... 15.4		45	53	112	151	112	Probable nonthermal or mixed water.
DH-15 117		169	159	154	190	184	Thermal water.
DH-17 ...... 58.0		143	137	135	166	155	Thermal water.
AH-20 111		148	142	140	169	165	Thermal water.
M-8[4] .....>100		167	157	121	157	151	Hottest thermal water sampled.
Desert Peak area							
DH-31 ...... 20.2		57	64	157	195	162	Possibly related to the Desert Peak geothermal system.
B21-2 ....>100		222	202	196	222	215	Thermal water.

[1]See table 2.

[2]Fournier (1977) (Na, sodium; K, potassium; Ca, calcium).

[3]Fournier and Potter (1979) (Mg, magnesium).

[4]A sulfate-oxygen isotope geothermometer estimate of 165 °C is given by Nehring and others (1979, p. 7) for a conductively cooled sample. The gas geothermometer method of D'Amore and Panichi (1980) yields an estimate of 190 °C.

DISCUSSION OF RESULTS

The chemical and isotopic composition of water at the Bradys Hot Springs geothermal system can be explained by the following scenario: meteoric water is first concentrated by low-temperature evaporation then recharged to a deep, thermal aquifer. The low-temperature evaporation (as opposed to vapor loss due to boiling) is indicated by the relation between stable isotopes and chloride. In fact, the chloride concentration and isotopic compositions of water found in both the Bradys and Desert Peak systems could result from low-temperature evaporation of meteoric water with a low chloride concentration (less than 50 mg/L) and a deuterium isotopic composition in the general range from −128 to −135 per mil (as shown in fig. 9). In the northern Basin and Range province, evaporation commonly occurs in closed basins that generally contain alkaline, sodium-dominated water (Olmsted and others, 1984). Thus, the increased chloride concentration and modified isotopic composition may be expected in a geochemical environment that also would result in a sodium-dominated water. Locally, this general type of geochemical environment exists in the playas located in Hot Springs Flat and to the southeast of Bradys Hot Springs (fig. 4), which therefore represent possible sources of recharge to the geothermal systems. Unfortunately, data could not be collected to evaluate these areas as sources of recharge to the geothermal system, because

properly located wells were not available. Recharge of a sodium-dominated, evaporation-affected water would require the following changes to produce a major-constituent and isotopic composition similar to that of water from well M-8: (1) an increase in silica concentration (to a limit controlled by equilibrium with quartz), (2) an increase in sulfate and a decrease in pH (perhaps due to the oxidation of sulfide minerals), (3) dissolution of a carbonate mineral (such as calcite) as a result of either the decrease in pH (causing the water to become undersaturated with respect to calcite) or a contribution of CO_2 from the deep crust or upper mantle, and (4) redistribution of the major cations within the aqueous phase.

The redistribution of cations is presumably caused by the same reactions that control the relation between sodium, potassium, and calcium in other geothermal systems—the reactions that provide the basis for a cation geothermometer. The water at Desert Peak well B21-2 could be formed in the same manner as described for the water at well M-8, with a meteoric water merely being affected to a greater extent by evaporation [resulting in a greater chloride concentration and a heavier (less negative) isotopic composition] prior to recharge.

The scenario described above requires no modification of the stable-isotope composition due to wall-rock interaction (oxygen shift) and only minor changes in the concentration of major constituents (with the exception of the silica increase). Flow within a fault system, where the ratio of wall-rock surface area to water volume is low, is
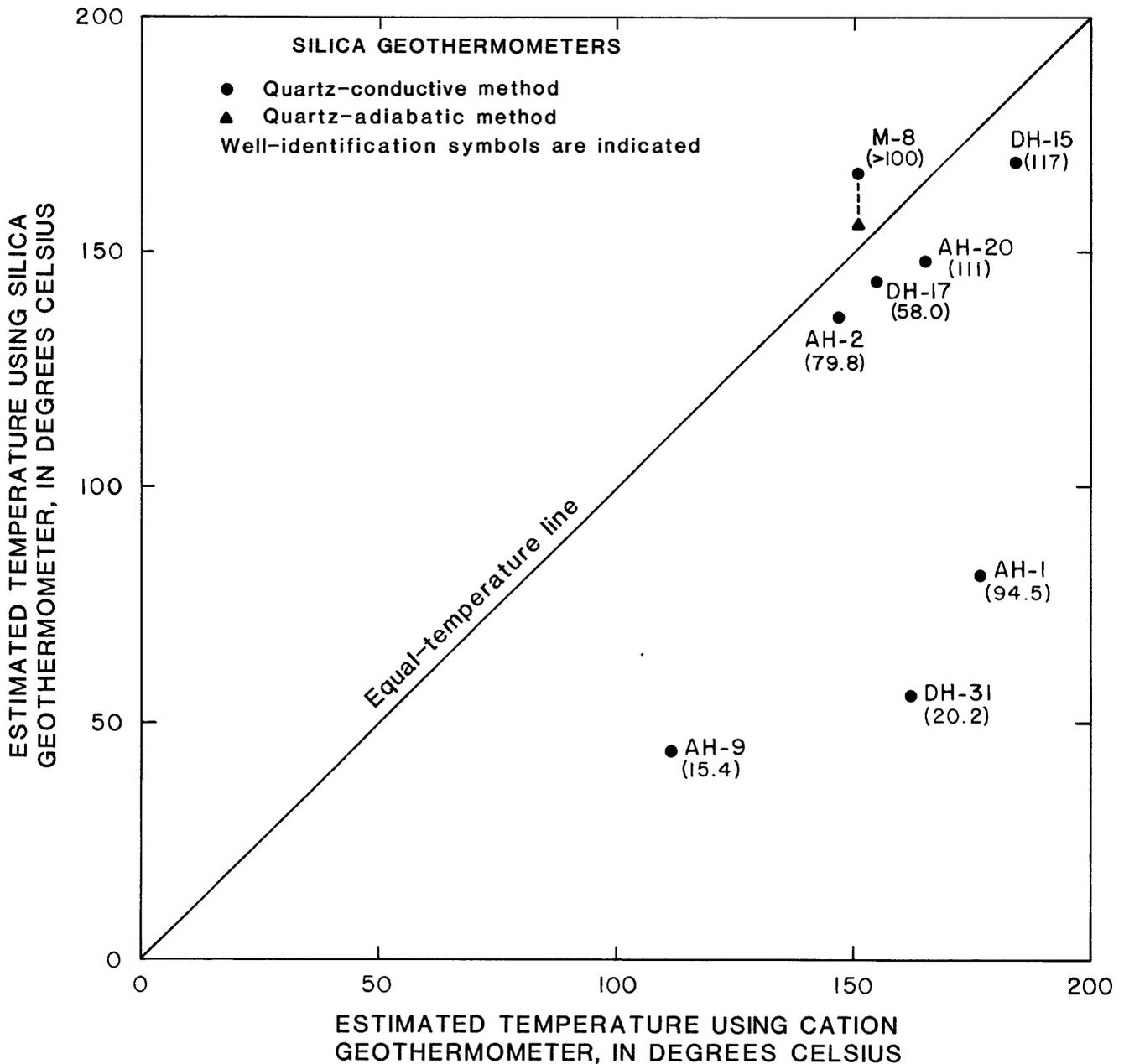
Figure 10. Comparison of aquifer temperatures estimated by using the silica and magnesium-corrected cation geothermometers for selected well waters. Numbers in parentheses are measured down-hole temperature, in degrees Celsius.

an environment where the water could equilibrate with the aquifer matrix with only limited change in water composition.

The oxygen shift also appears to be small at Leach Hot Springs—a geothermal system in the Basin and Range with an estimated temperature similar to that found at Bradys Hot Springs (Welch and others, 1981).

If the hydrogeochemical system outlined above is correct, then the concentrations of most minor con-

stituents appear to increase only slightly as water flows through the thermal aquifer. The lack of dramatic increase in fluoride and barium (which might be expected due to evaporative concentration or dissolution reactions in the thermal aquifer) may be largely a result of a mineral-solubility control. The bromide concentrations are largely a result of evaporative concentration rather than augmentation with the thermal aquifer. The concentrations of lithium, strontium, and to a lesser extent boron

may be at least partly controlled by dissolution in the thermal aquifer; hence, these constituents may be useful "geothermal indicators" in the Bradys system.

The scenario discussed above must be considered tentative due to a paucity of geochemical data on local recharge and on the deep, high-temperature thermal water. Analysis of the chemical and isotopic composition of recharge water upgradient from the thermal system could assist significantly in evaluating the geochemistry of the area. Unfortunately, usable sampling points in the area have not as yet been identified.

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