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Water Geochemistry and Hydrogeology of the Shallow Aquifer at Roosevelt Hot Springs, Southern Utah

A Hot Dry Rock Prospect

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CONVERSION OF UNITS OF MEASUREMENT

Used in this report	Multiply by	To obtain
Length		
centimeter (cm)	0.3937	inch (in.)
meter (m)	3.281	foot (ft)
kilometer (km)	0.6214	mile (mi)
Surface		
square kilometer (km^2)	0.3861	square mile (mi^2)
Volume		
cubic meter (m^3)	35.31	cubic foot (ft^3)
cubic meter (m^3)	264.2	gal
Pressure		
mega Pascal (MPa)	145.0	pound per square inch (psi)
Flow		
liter per minute (l/min)	0.2642	gallon per minute (gal/min)
cubic meter per minute (m^3/min)	264.2	gallon per minute (gal/min)
Specific capacity		
cubic meter per minute per meter (m^3/min·m)	80.52	gallon per minute per foot (gal/min·ft)
Transmissivity		
square meter per day (m^2/day)	10.76	cubic foot per day per foot (ft^3/day·ft)
	80.52	gallon per day per foot (gal/day·ft)
Permeability		
meter per day (m/day)	3.280	foot per day (ft/day)
	24.55	gallon per day per square foot (gal/day·ft^2)
Temperature		
degree Celsius (°C)	1.8(+32)	degree Fahrenheit (°F)
Heat Flow		
milliwatt per square meter (mW/m^2)	0.0239	heat flow unit (HFU)
Electrical conductivity		
microsiemens per centimeter (μS/cm)	1.000	micromho per centimeter (μmho/cm)

WATER GEOCHEMISTRY AND HYDROGEOLOGY OF THE SHALLOW AQUIFER AT ROOSEVELT HOT SPRINGS, SOUTHERN UTAH

A Hot Dry Rock Prospect

by

François-D. Vuataz and Fraser Goff

ABSTRACT

Geothermal exploration of Roosevelt Hot Springs KGRA (Known Geothermal Resources Area), Beaver County, southwestern Utah, since the 1970s has resulted in the discovery of a hydrothermal reservoir in Tertiary granite and Precambrian gneiss. A 20-MW(e) geothermal power plant has been in operation at the site since 1984. On the western edge of the geothermal field, several hot but dry wells have been drilled. Two of these wells were studied to determine the potential for development of a hot dry rock (HDR) geothermal system. A thick alluvial sequence overlies the granitic basement rocks and contains an important aquifer recharged by the nearby mountains. This aquifer is naturally contaminated by geothermal fluids leaking out of the hydrothermal reservoir, and the chemical composition of the water is unsuitable for water supply or irrigation uses throughout an area of about 220 km^2. However, this water would be convenient for industrial use, such as injection into a hot dry rock energy extraction system constructed in granitic rocks beneath the alluvial sequence.

A detailed geochemical investigation of 21 waters (thermal and cold, mineralized and dilute) was performed to evaluate the shallow aquifer for HDR use. The deep geothermal fluid and its surface manifestation (Roosevelt seep) have essentially a Na-Cl type of chemical composition. The well waters issuing from the alluvium display various degrees of mixing between dilute cold shallow groundwater and mineralized hot geothermal fluid, which comprises 10 to >60% of the well waters. The contaminated alluvial aquifer is significantly enriched in SiO_2 and in trace elements such as Li, B, Ba, Mn, and As. Oxygen-18 and deuterium results confirm that mixing occurs with geothermal fluids and that the recharge area of this portion of the alluvial aquifer has its origin in the Mineral Mountains. Tritium results indicate a relatively slow circulation rate within the alluvial aquifer and therefore a long underground residence time of the water. In the HDR zone, the alluvial aquifer was recharged at least 70 years ago. By indirect evaluation, the geothermal fluids of the deep hydrothermal reservoir also have the same minimum underground residence time.

Thermodynamic computations on the reaction states of numerous mineral phases show that the majority of the waters studied are saturated with a fair number of primary and secondary silicates as well as carbonates and fluorite. Chemical equilibria were also computed for a range of temperatures between 25° and 250°C, temperatures at which the water will be injected and produced in an HDR energy extraction system. At low temperatures the selected water samples display saturation or oversaturation with respect to the majority of the primary and secondary minerals encountered in the deep HDR zone. On the contrary, most minerals are undersaturated at 250°C, and the relatively dilute waters injected in the artificial fractures will dissolve minerals from the reservoir rocks until a steady-state concentration is reached. A pressurized closed loop will prevent carbonate precipitation in the surface pipes and heat extraction equipment. Silica scaling and corrosion due to redox reactions are problems that cannot be adequately evaluated until a specific plant design and aquifer fluid are known.

A pumping test carried out in an existing shallow stock well in the HDR zone, with a maximum yield of 0.87 m^3/min, revealed a relatively low transmissivity (130 m^2/day) for a well-specific capacity of 0.14 m^3/min·m. The permeability has been estimated to be 1 to 10 m/day, which corresponds to typical silty sand materials. These values may be valid only locally due to lithologic variations within the aquifer. Although the water level of the aquifer has sharply declined in the Milford area over the past 20 years due to intensive irrigation, the alluvial aquifer has remained stable in the HDR zone during this period. Under heavy pumping the shallow aquifer will probably reach steady-state conditions after a few months.

The water requirements at the prototype HDR system at Fenton Hill, New Mexico, indicate that a future 50-MW(e) HDR power plant would need a continuous supply of about 2 m^3/min, which could easily be pumped from one or two 60-m-deep wells with a casing diameter of 30.5 cm (12 in.). The location of these wells will determine the chemical composition of the water, which may have a total mineralization varying from 1300 to more than 5000 mg/l.

1. INTRODUCTION

The Roosevelt Hot Springs area (Fig. 1) is located between the Mineral Mountains and the Escalante Desert, in Beaver County, southwestern Utah, about 16 km northeast of the city of Milford. In the past, Roosevelt Hot Springs was a resort, but spring activity decreased during the 20th century until the springs went completely dry in 1966 (Mundorff 1970). In the same area, concerted industrial geothermal exploration began in 1972 and resulted in the discovery of a commercial hydrothermal reservoir in Tertiary granite and Precambrian gneiss. A 20-MW(e) geothermal power plant has been constructed and has operated since 1984. Many reports and publications on the

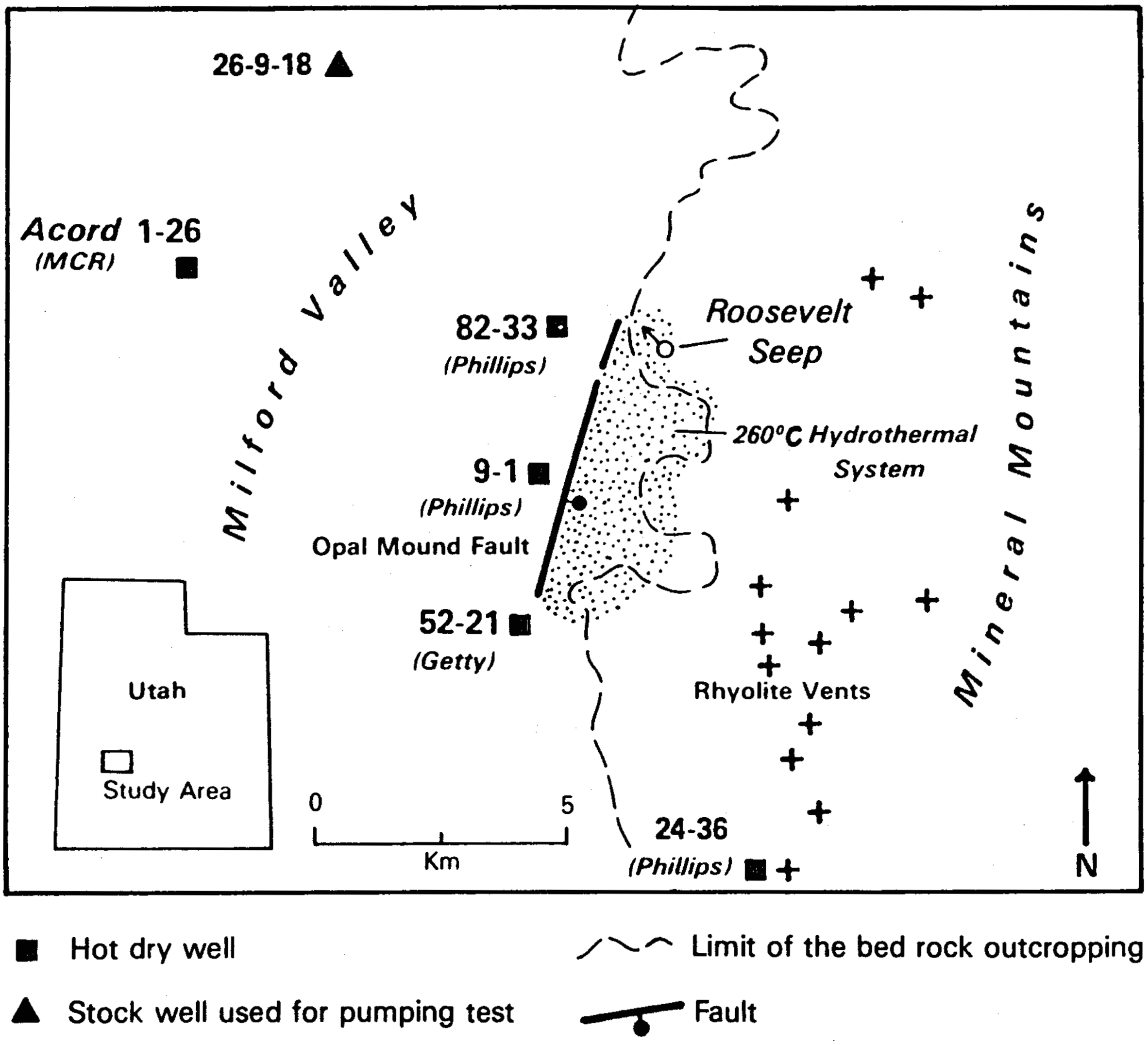

Fig. 1.
Location map of the Roosevelt Hot Springs geothermal area.

exploration and development of Roosevelt Hot Springs KGRA (Known Geothermal Resources Area) are available (e.g., Ross et al. 1982).

On the western edge of the geothermal field, three deep holes have been drilled that are very hot but mostly dry. Two of them (Phillips 9-1 and Acord 1-26 wells) have been studied by Los Alamos National Laboratory for the Hot Dry Rock (HDR) resources evaluation program (Glenn et al. 1981; Shannon et al. 1983a). A review of data and recommendations have been formulated to evaluate the HDR geothermal potential at Roosevelt (East 1981). The present report is directed toward the study of the shallow aquifer of the Milford Valley to determine if the local groundwater would be suitable for use as make-up water in an HDR system. This investigation is the result of a cooperative agreement between Los Alamos and Phillips Petroleum Co., formerly the main operator of the Roosevelt Hot Springs Unit. The presence of these hot dry wells and the similar setting of the Roosevelt area to the prototype HDR site at Fenton Hill, New Mexico, make Roosevelt a very good candidate site for creation of another HDR geothermal system.

This investigation has two main objectives: (1) to assess the water geochemistry of the valley aquifer, to determine possible problems in future make-up water use, such as scaling or corrosion in the wells and surface piping, and (2) to assess the hydrogeology of the shallow groundwaters above the HDR zone, to characterize the physical properties of the aquifer. These two objectives are linked by the fact that the valley aquifer is naturally contaminated by geothermal fluids leaking out of the hydrothermal reservoir. In an arid region where good-quality fresh water is needed for public water supply and irrigation, nonpotable waters would be ideal for an industrial use such as injection into an HDR energy extraction system.

2. FRAMEWORK OF STUDY

2.1 Geologic Setting

The Roosevelt Hot Springs area is located at the eastern margin of the Basin and Range province, where this province is transitional with the Colorado Plateau. The transition zone is characterized by thin crust (approximately 25 km). Regional heat flow averages around 105 mW/m^2, whereas the Basin and Range and Colorado Plateau provinces have a mean heat flow of 92 and 50 mW/m^2, respectively (Ward et al. 1978). The Mineral Mountains and Milford Valley to the west form a horst and graben structure. The northern and southern portions of the Mineral Mountains are mainly Precambrian metamorphic rocks and overlying sedimentary rocks of Paleozoic to Mesozoic age (Nielson et al. 1986). The central Mineral Mountains are a structural high that exposes intensely deformed Precambrian rocks and a Tertiary granitic pluton (Mineral Mountains Pluton) as young as 9 Myr. North of the Mineral Mountains appear Quaternary basaltic flows and cones (0.01 Myr), while along the crest of the Mineral Mountains rhyolitic domes, flows, and ash-flow tuffs ($\geq$0.7 Myr) have been erupted (Fig. 2). Quaternary

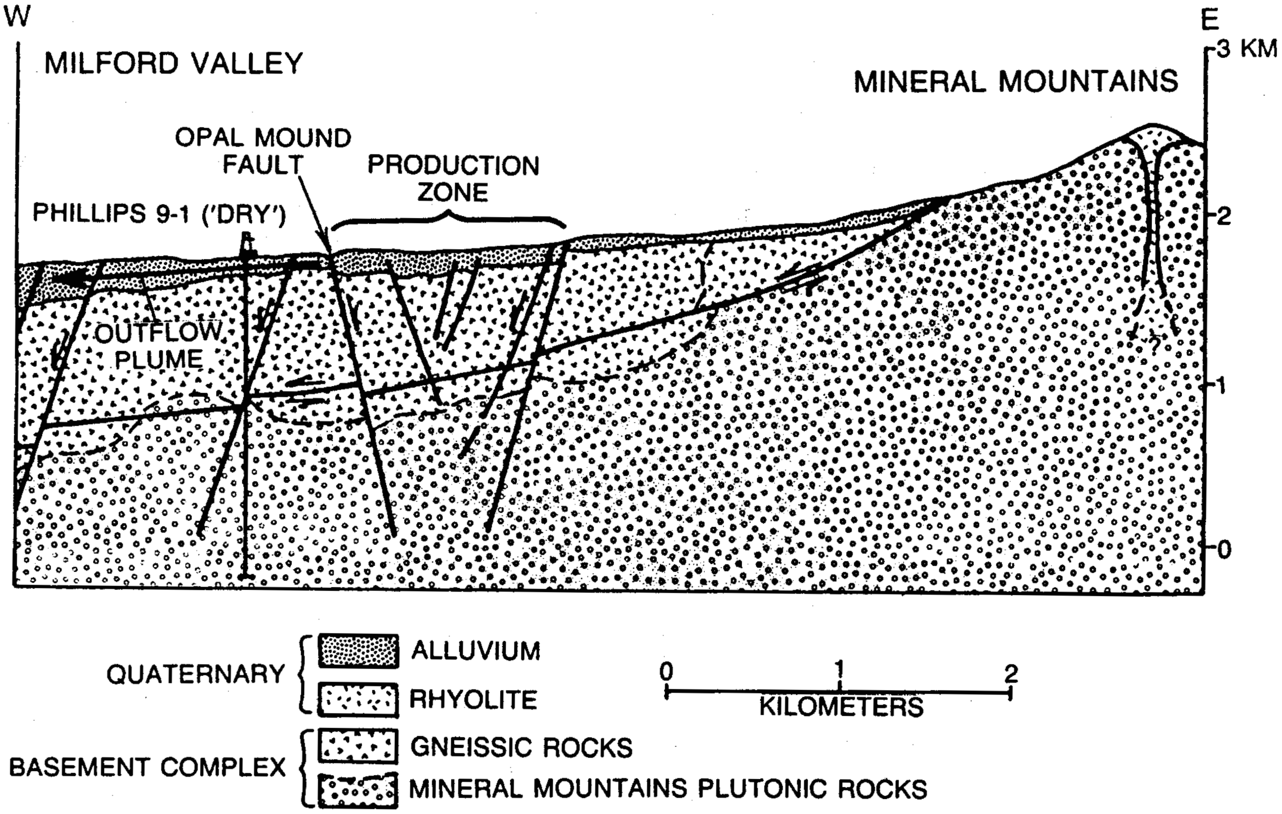

Fig. 2.
Schematic cross section of Roosevelt Hot Springs area (after Goff and Decker 1983). The region of potential hot dry rock resources lies west of the Opal Mound Fault.

hot spring deposits of opaline and chalcedonic sinters are located along the Opal Mound Fault, a major north-south structure at Roosevelt Hot Springs.

According to the drilling report of the westernmost deep well (Acord 1-26), the thickness of the Quaternary alluvium in Milford Valley reaches roughly 1 km (Shannon et al. 1983b). Its composition is mainly clay, silt, sand, and gravel eroded from the surrounding mountains, with various percentages of each component according to depth and location. These deposits overlie Tertiary volcanics and sediments and, in turn, overlie Precambrian gneissic rocks and Tertiary intrusives similar to the Mineral Mountains.

2.2 Hydrogeology

Roosevelt Hot Springs area is located on the east side of Milford Valley at an average elevation of 1830 m. The elevation of the floor of the valley and the Beaver River bed is 1510 m, whereas the elevation of the crest of the Mineral Mountains reaches 2750 m. The Beaver River is the only perennial stream around Milford and it is intensely used for irrigation. North of Milford, the Beaver River channel is often dry due to evaporation, percolation into the alluvium, and above all, the large amount of water drawn for irrigation. Several ephemeral streams (washes) flow westward from the Mineral Mountains during snowmelt or heavy rainstorms, but rarely do the waters reach the lower parts of the valley.

The alluvial aquifer as considered here is the aquifer that flows in the Quaternary alluvium of Milford Valley. Between Milford and Minersville where irrigation is heavy, the alluvial aquifer is called herein the main aquifer. These aquifers are separated from the geothermal aquifer, which circulates in pre-Quaternary crystalline basement. Mower and Cordova (1974) conducted a thorough study of the groundwater resources of the Milford area because of the severe drawdown of the alluvial aquifer in the irrigated zones. South of Milford, the density of irrigation and stock wells is very great, and water levels in the main aquifer have declined 6 to >9 m during a 22-year period (1950-1972). North of Milford, the density of wells is much less, and in the present study area, the water levels are estimated to have declined only 0 to 3 m during the same observation period. During that period, there were no irrigation wells, only windmills and small stock wells. Presently, only a few irrigation wells at the foot of the Mineral Mountains are pumped during the growing season, and these wells do not have a strong effect on the regional groundwater resources. Some springs flowing from the mountains are used for cattle grazing, while others run off in various washes and percolate into the alluvial aquifer.

2.3 Geothermal Energy

2.3.1 Hydrothermal Resources

Exploratory drilling began in 1975 at Roosevelt Hot Springs KGRA (Utah Geological and Mineral Survey 1980). Phillips Petroleum, the major lease holder and operator, and four other companies tapped a 270°C geothermal reservoir of commercial size. This reservoir is located in fractured Tertiary granite and Precambrian gneiss east of the Opal Mound Fault. Steam is flashed from a relatively dilute Na-Cl geothermal fluid having total dissolved solids of 7000 mg/l and neutral pH (Parry et al. 1980). In October 1982, a 1.6-MW(e) biphase rotary separator turbine had been producing power for five months. The construction of a 20-MW(e) single-flash power plant "Milford No. 1" was completed and put on-line in the spring of 1984 (Geothermal Resources Council 1983). Plans for an additional 20-MW(e) power plant have already been made.

Thirteen deep holes have been drilled and nine wells are producing, whereas five dry wells roughly delineate the boundary of the reservoir. The production zone is naturally bounded to the east by the Mineral Mountains horst. On the west, three dry holes have been drilled beyond the Opal Mound Fault (Figs. 1 and 3), and on the south, the boundary is shown by two nonproducing wells. The northern limit of the hydrothermal reservoir is not yet delineated (Ross et al. 1982).

Heat flow measurements have proved that the thermal anomaly extends well beyond the geothermal reservoir. The area of the reservoir is estimated to be 20 km^2, but the 400-mW/m^2 contour on a heat flow map encloses an area of 57 km^2. A value of 400 mW/m^2 is four times greater than the regional heat flow, which is already considered to be high (East 1981).

2.3.2 Hot Dry Rock Resources

As mentioned above, five dry wells have been drilled outside the periphery of the hydrothermal reservoir, but three of them display bottom hole temperatures in excess of 200°C at depths of 2 to 3 km and temperature gradients in the range of 55°-60°C/km (Table I). Two of these wells have been studied by Los Alamos, namely Roosevelt 9-1 (Glenn et al. 1981) and Acord 1-26 (Shannon et al. 1983a). Roosevelt Hot Springs KGRA has been declared to be a good candidate site for future HDR development for several reasons (Goff and Decker 1983). First, subsurface temperatures high enough for power generation (180°-300°C) are encountered at reasonable depth (2-5 km). Second, the manmade reservoir would be in the same type of rock (unaltered gneiss and granite) as the HDR experiment at Fenton Hill, New Mexico (Fig. 4). Third, the electricity produced by an HDR system could easily be channeled into the existing power grid of the 20-MW(e) plant Milford No. 1. Finally, the HDR resource has been conservatively estimated by East (1981) to be about eight times larger than the known hydrothermal

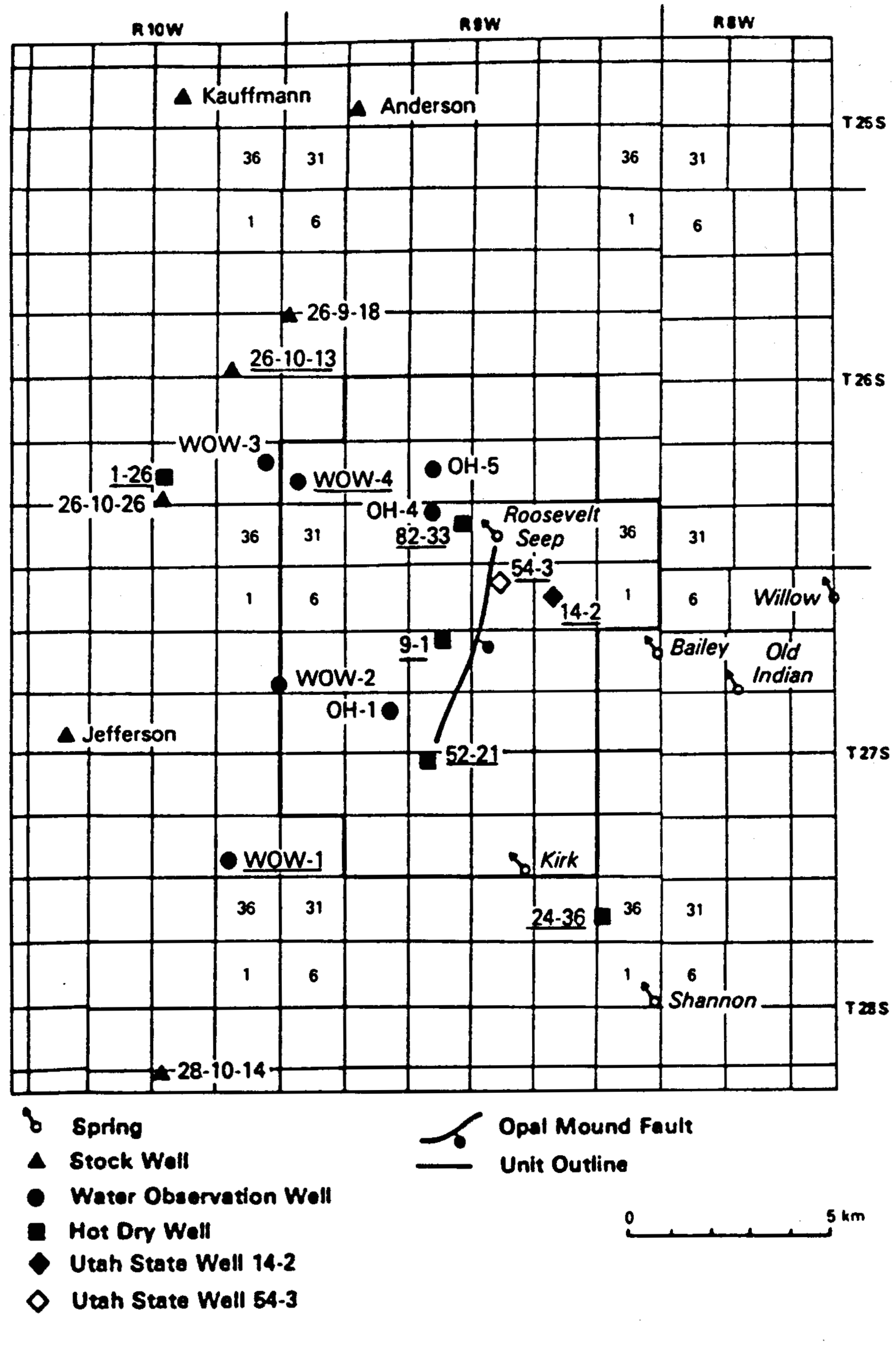

Fig. 3.

Location map of wells and springs studied in the Roosevelt Hot Springs area. The wells whose names are underlined were not sampled, namely the five hot dry wells, WOW-1 (equipped with a water level recorder), WOW-4 (not enough water), well 26-10-13 (inaccessible old windmill), and Utah State wells 14-2 and 54-3 (geothermal production wells). Other wells (observation holes and geothermal wells) are present east of the Opal Mound Fault that are not part of this study. Three sampling points are out of the map: Rock Corral spring, well Milford No. 1, and Beaver River.

TABLE I
DATA FROM THE HOT DRY WELLS AT ROOSEVELT HOT SPRINGS

Name	Location[a]	Depth (m)	BHT (°C)[b]	Formation at depth	Operator
Acord 1-26[c]	(C-26-11)26	3855	231	gneiss-granite	MCR Geothermal
Roosevelt 82-33[d]	(C-26-9)33	1837	-	gneiss-granite	Phillips Petr.
Roosevelt 9-1[e]	(C-27-9)9	2199	227	gneiss-granite	Phillips Petr.
Utah State 52-21[f]	(C-27-9)21	2281	212	gneiss-granite	Getty Oil
Utah State 24-36[g]	(C-27-9)36	1865	-	gneiss-granite	Thermal Power

[a] See Figs. 1 and 3 for locations.
[b] Bottom hole temperature.
[c] Heat flow = 146 mW/m^2; Shannon et al. (1983a).
[d] Lenzer et al. (1977).
[e] Renamed C/T-2 by Los Alamos; Glenn et al. (1981).
[f] Glenn and Hulen (1979).
[g] Data proprietary.

KRGA. The conventional resource at Roosevelt has an evaluated areal extent of 20 km^2, whereas the HDR resource encompasses 160 km^2.

3. WATER GEOCHEMISTRY

During two sampling periods (May 1982 and April 1983), 32 water samples were collected from 21 different places, namely 13 wells, 7 springs, and the Beaver River. Table II contains the field data and physical parameters of the waters, Fig. 3 is a location map, and Fig. 5 explains the location numbering system.

3.1 Field and Laboratory Procedures

A multiparameter digital device (Portalab 500, Presto-Tek Corp.) was used to measure temperature, pH, Eh, and conductivity. Discharge was measured with graduated buckets or a tank during the pumping test (described in a later section) or was estimated visually. HCO_3 was analyzed in the field as total alkalinity by sulfuric acid titration.

Samples of water for analysis were filtered when necessary using a vacuum pump system with a 0.45-µm filter paper. The following types of samples were collected in various polyethylene or glass bottles with Polyseal caps: (1) a 500-ml plastic bottle of raw water for anions; (2) a 250-ml plastic bottle of filtered and acidified water for cations and trace elements; (3) a 125-ml

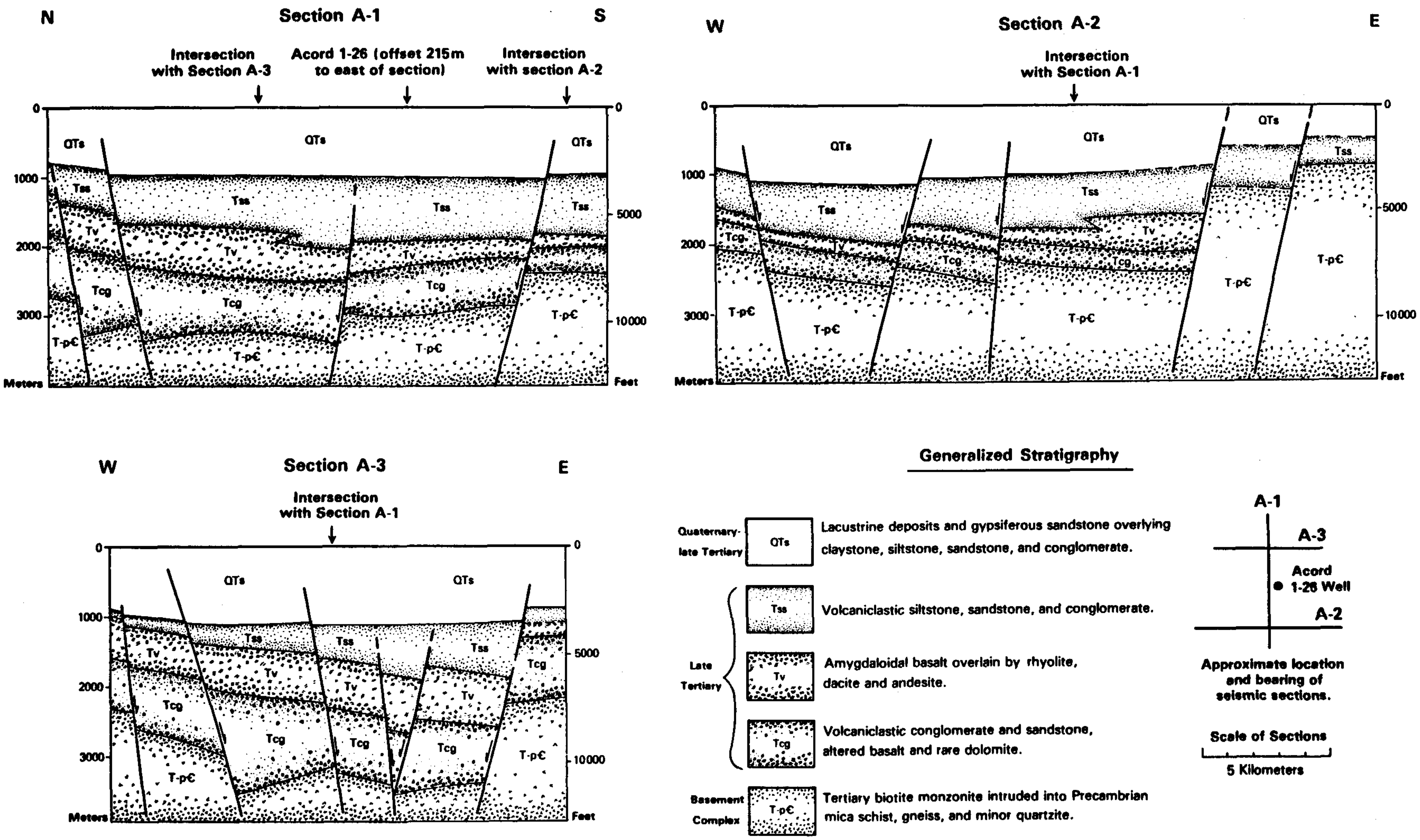

Fig. 4.
Generalized stratigraphy of three sections in the Milford Valley interpreted from three seismic reflection traverses around the Acord 1-26 well (courtesy of MCR Geothermal Corporation; after Shannon et al. 1983a). The eastern parts of these sections occur within the zone of HDR interest, and the upper layer of alluvium contains a shallow aquifer of potential use for HDR development as described in text.

plastic bottle of filtered water diluted 1:10 with deionized water for silica; (4) a 500-ml glass bottle of raw water for tritium; and (5) a 125-ml glass bottle of raw water for oxygen-18 and deuterium. Acidification was done by adding concentrated HNO_3 until the pH dropped lower than 2. The samples for determination of ionized Al^{+++} were collected and analyzed according to a modified procedure of Barnes (1975).

Laboratory analyses were performed at the Fenton Hill chemical laboratory (Los Alamos National Laboratory) by the following methods: Ag, Al^{+++}, Cd, Co, Cr, Cu, Ni, and Pb were analyzed by atomic absorption spectroscopy using either a graphite furnace or flame excitation; Al, As, B, Ba, Ca, Fe, K, Li, Mg, Mn, Mo, Na, Sr, and Zn by inductively coupled plasma emission spectroscopy; Br, Cl, NO_3, PO_4, and SO_4 by ion chromatography. F and NH_4 analyses were performed by selective ion electrode and SiO_2 by colorimetry (Tables III and IV).

Oxygen-18 and deuterium analyses were carried out by the Bureau des Isotopes Stables, Centre d'Etudes Nucléaires de Saclay, Gif-sur-Yvette, France. Tritium data were obtained in 1982 from Teledyne Isotopes, Westwood, New Jersey, and in 1983 from the Tritium Laboratory, University of Miami, Miami, Florida (Table V).

3.2 Physical Parameters

3.2.1 Temperature

According to Mower and Cordova (1974), 9.6°C is the mean annual temperature at Milford, which is 1511 m above sea level. In areas of average geothermal gradient, groundwaters in the upper 100 m of the aquifer display temperatures 1° to 2°C above the mean annual air temperature (Freeze and Cherry 1979). Using this assumption, groundwater temperature is expected to be around 11°C at Milford. Therefore, for an elevation span of 1500 to 1700 m, where most of the studied wells are, the average groundwater temperature should be 10.5 ± 0.5°C, using an estimated average gradient of 0.5°C/100 m (Freeze and Cherry 1979). However, none of the wells whose water level is shallower than 100 m display this temperature. Indeed, their temperatures range from 13° to 34°C for water levels of 5 to 59 m (Table II). It can be concluded that water temperature increases either because of mixing with geothermal water or because of high local heat flow or both.

TABLE II
FIELD DATA AND PHYSICAL PARAMETERS FOR THE WATERS OF ROOSEVELT HOT SPRINGS AREA[a]

Name	No.	Date	Temp. (°C)	pH	Eh(mV)	Conduct. (μS/cm)	Discharge (l/min)	Water level (m)	Total depth (m)	Diameter[b] (cm)	Sampling procedure[c]	Elevation (m.a.s.l.)	Location number[d]
Shannon spring[e]	RO-16	5/27/82	7.3	7.55	213	47	950	-	-	-	d	2240	(C-28-9)1
Rock Corral spring	RO-17	5/27/82	5.7	6.76	195	106	300	-	-	-	d	2158	(C-28-9)14
Old Indian spring	RO-3	5/24/82	14.3	7.67	285	296	2	-	-	-	d	2414	(C-27-8)8
Bailey spring[f]	RO-28	4/19/83	6.1	7.22	49	258	340	-	-	-	d	2073	(C-27-9)12
Bailey spring[f]	RO-2	5/24/82	7.8	7.26	255	286	75	-	-	-	d	2073	(C-27-9)12
Well Milford No.1	RO-19	5/28/82	23.3	8.24	207	418	22700[g]	5.5[h]	180	40.6	p	1515	(C-28-10)7
Willow spring	RO-18	5/27/82	7.3	7.09	194	505	150	-	-	-	d	2073	(C-27-8)4
Beaver River[i]	RO-21	5/28/82	15.5	8.84	171	580	-	-	-	-	d	1585	(C-30-10)12
Well 28-10-14[j]	RO-10	5/26/82	≤20.6	8.16	171	354	1500[g]	57.9[h]	77.7	20.3	p	1573	(C-28-10)14
Kirk spring	RO-15	5/27/82	11.4	7.58	190	502	6	-	-	-	d	1945	(C-27-9)27
Beaver River[i]	RO-33	4/22/83	16.8	8.06	118	596	-	-	-	-	d	1506	(C-27-10)17
Well Kaufmann[k]	RO-4	5/25/82	13.3	7.86	-144	535	0	5.4	8.4	≥60	b	1501	(C-25-10)26
Well Anderson[l]	RO-5	5/25/82	≤16.5	7.49	42	2170	3800[g]	21.8	24.4	≤15	b	1527	(C-25-9)29
Well Jefferson	RO-25	4/18/83	18.4	7.30	105	2260	≥1400	7.6[h]	50.3	20.3	p	1521	(C-27-10)16
Well Jefferson	RO-14	5/27/82	18.9	7.78	131	2210	≥1400[g]	7.6[h]	50.3	20.3	p	1521	(C-27-10)16
Well WOW-1[m]	-	5/26/82	-	-	-	-	-	99.3	126	17.8	-	1622	(C-27-10)25
Well WOW-2	RO-8	5/25/82	32.3	7.22	-179	3790	0	104	122	17.8	b	1625	(C-27-10)12
Well 26-9-18	RO-6	5/25/82	18.6	6.70	-112	6320	570[g]	22.0[h]	61.3	20.3	p	1527	(C-26-9)18
Well 26-9-18[n]	RO-30	4/20/83	18.8	6.47	15	6310	790	27.3	61.3	20.3	p	1527	(C-26-9)18
Well 26-9-18[n]	RO-32	4/22/83	18.3	6.33	58	6310	790	29.2	61.3	20.3	p	1527	(C-26-9)18
Well 26-9-18[n]	RO-29	4/20/83	19.5	6.45	-52	6360	210	22.6	61.3	20.3	p	1527	(C-26-9)18
Well 26-9-18[n]	RO-31	4/21/83	18.5	6.46	28	6260	830	27.8	61.3	20.3	p	1527	(C-26-9)18
Well 26-10-13[o]	-	4/20/83	-	-	-	-	-	20.7	?	15.2	-	1524	(C-26-10)13
Well OH-5	RO-13	5/26/82	82.7	6.56	-70	5500	0	185	555	20.3	b	1700	(C-26-9)28
Well OH-1	RO-27	4/19/83	56.3	6.94	-21	8360	0	154	706	17.8	b	1719	(C-27-9)17
Well OH-1	RO-9	5/26/82	54.2	6.96	-50	8270	0	155	706	17.8	b	1719	(C-27-9)17
Well 26-10-26	RO-24	4/18/83	15.8	7.23	65	7330	450	18.9[h]	42.7	15.2	p	1521	(C-26-10)26
Well 26-10-26	RO-7	5/25/82	16.3	6.97	-16	7760	450[g]	18.9[h]	42.7	15.2	p	1521	(C-26-10)26
Well WOW-4[p]	-	5/26/82	34.0	7.25	-20	6950	0	58.6	58.8	17.8	b	1585	(C-26-9)30
Well WOW-3	RO-11	5/26/82	23.9	8.13	67	8930	0	34.5	91.4	17.8	b	1558	(C-26-10)25
Well WOW-3	RO-26	4/19/83	23.5	8.28	146	9070	0	40.7	91.4	17.8	b	1558	(C-26-10)25
Roosevelt seep[q]	RO-1	5/24/82	26.0	6.17	50	10400	1-10	-	-	-	d	1807	(C-26-9)34
Roosevelt seep[q]	RO-22	4/18/83	21.5	6.08	89	10100	1-10	-	-	-	d	1807	(C-26-9)34
Well OH-4	RO-12	5/26/82	48.2	6.89	-86	10800	0	154	536	17.8	b	1737	(C-26-9)33
Well Utah State[r]	14-2	1977-78	268	5.9	-	-	-	-	1862	-	-	1908	(C-27-9)2

TABLE II (continued)

[a] All data were measured in the field. The sample arrangement is the same for Tables II to VI and is explained in Table III, note a.

[b] Diameter of the upper casing.

[c] d = direct; p = at the wellhead during pumping; b = with a bailer.

[d] See Fig. 5 for explanation of the numbering system and Fig. 3 for map location.

[e] Shannon spring was sampled downstream from the emergence zone.

[f] Bailey spring was sampled at a seeping zone in the stream bed, where the flow is maximum.

[g] Discharge value measured on another date.

[h] Water level value measured on another date.

[i] Beaver River was sampled at two different places: at Minersville Bridge (RO-21) and at the intersection of Highway 257 with the dirt road to Roosevelt Hot Springs geothermal field (RO-33).

[j] Well 28-10-14 was sampled from a water truck right after having been filled by pumping at the well.

[k] Well Kaufmann is an old hand-dug windmill; no water was pumped at the time of sampling.

[l] Well Anderson was sampled at an observation piezometer located 150 m west of the main irrigation well, which was under pumping during sampling. The discharge value is from the irrigation well, whereas all the other data are from the observation piezometer.

[m] Well WOW-1 is equipped with a permanent water level recorder of the U.S.G.S.; therefore only the depth to water can be measured.

[n] Samples RO-29 to RO-32 from well 26-9-18 were collected during the pumping test period.

[o] Well 26-10-13 is an old windmill, partly dismantled, and only the water level can be measured.

[p] Well WOW-4 is presently sanded up almost to the water level; therefore not enough sample could be taken for chemical analyses. The original total depth of this well was 93.0 m.

[q] Roosevelt seep is formed by two pools. The samples were collected from the southern pool.

[r] Data are from Capuano and Cole (1982) and Ross et al. (1982).

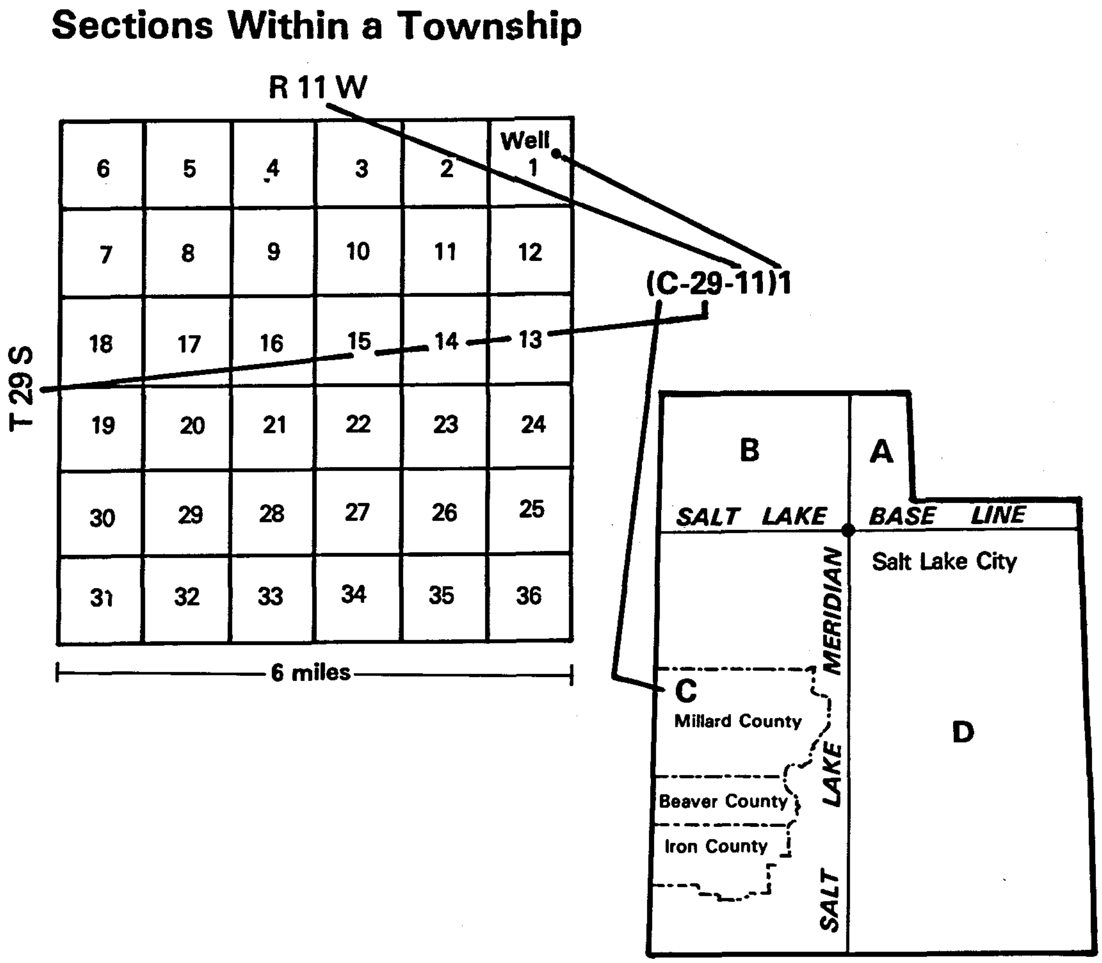

Fig. 5.
Well- and spring-location numbering system used in Utah (after Mower and Cordova 1974).

3.2.2 pH and Eh

The Roosevelt seep, which is derived from surface discharge of the deep geothermal system, has the lowest pH of all samples (6.1). Using chemical data from a geothermal well and hydrogen ion mass balance, Capuano and Cole (1982) recalculated the reservoir fluid pH at 260°C, which is about 6.0. The mineralized water wells [total dissolved solids (TDS) $\geq$ 2000 mg/l] have a slightly acid to neutral pH (6.3 to 7.2) due to the buffering of cold groundwater by geothermal fluid. On the other hand, dilute water wells (TDS <2000 mg/l) show a neutral to slightly basic pH (7.3 to 8.2). One exception to this is well WOW-3, which is strongly mineralized but has a pH of 8.3,

TABLE III
MAJOR AND MINOR CHEMICAL PARAMETERS (mg/l) AND IONIC BALANCE (meq/l) FOR THE WATERS OF ROOSEVELT HOT SPRINGS AREA[a]

Name	No.	Na^+	K^+	Ca^{++}	Mg^{++}	Li^+	Sr^{++}	Cl^-	HCO_3^-	SO_4^{--}	Br^-	F^-	SiO_2	B	TDS[b]	$\sum$cations[c]	$\sum$anions[c]
Shannon spring	RO-16	3	0.6	4	0.95	<0.01	0.02	3.76	10	5.03	<0.05	0.72	15	<0.05	43	0.43	0.41
Rock Corral spring	RO-17	7	0.9	9	2.40	<0.01	0.06	4.73	46	5.16	<0.05	0.65	21	<0.05	97	0.97	1.02
Old Indian spring	RO-3	12	2.4	22	4.6	0.01	0.10	13.0	100	10.3	<0.05	0.23	2	<0.05	167	2.06	2.23
Bailey spring	RO-28	15	2.3	32	5.6	0.05	0.16	14.2	127	9.3	<0.1	0.81	23	0.14	230	2.78	2.71
Bailey spring	RO-2	14	2.5	31	6.3	0.01	0.14	15.2	146	9.4	<0.05	0.92	25	<0.05	251	2.74	3.07
Well Milford No.1	RO-19	60	2.0	14	7.2	0.04	0.35	18.3	160	53.5	0.53	0.85	30	0.05	347	3.97	4.30
Willow spring	RO-18	31	3.1	51	11.5	0.03	0.22	29.7	227	18.1	<0.05	0.96	39	<0.05	412	4.93	4.99
Beaver River	RO-21	43	7.6	48	17.3	0.02	0.44	31.7	261	43	0.29	0.93	31	<0.05	484	5.89	6.12
Well 26-10-14	RO-10	32	2.2	27	5.6	0.02	0.29	32.3	116	23.7	0.32	0.68	26	<0.05	266	3.27	3.34
Kirk spring	RO-15	33	4.6	50	11.7	0.03	0.26	34.5	234	21.8	<0.05	1.15	33	<0.05	424	5.03	5.31
Beaver River	RO-33	44	11.4	58	16.7	0.06	0.51	35.4	260	45.9	0.2	0.81	24	0.18	497	6.48	6.26
Well Kaufmann	RO-4	35	8.8	39	15.0	0.07	0.68	52.9	168	43	0.48	3.37	74	0.06	440	4.96	5.33
Well Anderson	RO-5	120	12	138	74	0.17	2.21	532	112	91.7	0.76	2.35	53	0.43	1139	18.58	18.89
Well Jefferson	RO-25	255	8.1	120	29.6	0.43	1.2	571	207	21	0.5	0.35	46	2.29	1263	19.81	19.97
Well Jefferson	RO-14	250	9	117	31	0.35	1.11	605	200	5.88	0.80	0.38	48	2.79	1271	19.57	20.50
Well WOW-2	RO-8	494	19	151	33	3.65	1.2	1050	229	3.4	0.15	0.49	46	7.53	2038	32.78	33.47
Well 26-9-18	RO-6	860	66	300	16.7	8.5	6.16	1600	890	14.8	0.47	1.31	73	14.4	3851	56.80	60.11
Well 26-9-18	RO-30	894	64	322	15.5	10.8	5.9	1670	865	29	1.3	1.19	71	13.8	3963	59.56	61.97
Well 26-9-18	RO-32	890	61	333	16.0	10.8	6.0	1670	865	29	1.2	1.16	70	14.0	3967	59.91	61.97
Well 26-9-18	RO-29	910	62	318	15.6	10.6	5.9	1680	860	29	1.2	1.20	70	13.8	3977	59.98	62.16
Well 26-9-18	RO-31	894	61	332	16.2	10.7	6.1	1680	870	29	1.3	1.19	70	13.9	3985	60.03	62.33
Well OH-5	RO-13	620	34	330	41	3.57	15.0	1740	51	5.16	0.18	1.56	44	9.28	2895	48.53	50.11
Well OH-1	RO-27	1550	125	95	21.5	19.4	4.6	2130	1180	41	4.4	2.07	140	30.8	5344	80.03	80.44
Well OH-1	RO-9	1520	140	93.4	21	15.8	4.34	2160	1220	43.2	11.5	2.23	154	32.6	5418	78.47	82.08
Well 26-10-26	RO-24	987	43	360	68.3	12.1	4.8	2260	303	14	1.6	0.53	82	18.7	4155	69.46	69.06
Well 26-10-26	RO-7	920	42	380	80	9.1	5.40	2330	387	7.1	0.92	0.45	81	19.9	4263	68.06	72.25
Well WOW-3	RO-11	1520	200	80	16.3	15.3	4.71	2780	94	4.81	0.42	1.05	1	21.5	4739	78.88	80.13
Well WOW-3	RO-26	1540	185	77	15.4	16.0	4.5	2840	90	3.2	2.0	0.91	2	21.5	4798	79.24	81.73
Roosevelt seep	RO-1	1800	280	101	21	18.2	2.41	3160	292	127	5.2	4.07	160	28.1	5999	94.90	96.85
Roosevelt seep	RO-22	1830	238	112	20.6	19.9	2.3	3230	296	129	2.2	3.71	168	27.1	6079	95.89	98.88
Well OH-4	RO-12	1800	240	260	40	17.6	3.29	3440	104	84.8	2.10	1.64	60	30.6	6084	103.31	100.63
Utah state well[d]	14-2	2190	401	8.48	0.30	26.5	1.36	3650	232	69.0	-	5.00	656	26.3	7266	109.86	108.54

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[a] The water sources and the samples are arranged by increasing Cl concentration. In case of similar Cl content, TDS is prevalent. The ion sequence is made by decreasing concentration of cations and anions, respectively, as observed in the most mineralized samples. The same sample arrangement is made in Tables II to VI.

[b] TDS (total dissolved solids): sum of the chemical species listed in this table.

[c] Sum of the cations and sum of the anions in milliequivalent per liter (meq/l) used to check the ionic balance.

[d] Average composition from 1 to 4 analyses. If data are not available, values for well Roosevelt KGRA 54-3 are given (Capuano and Cole 1982; Ross et al. 1982).

TABLE IV
TRACE ELEMENTS FOR THE WATERS OF ROOSEVELT HOT SPRINGS AREA (mg/l)[a]

Name	No.	Ag	Al	As	Ba	Cd	Co	Cr	Cu	Fe	Mn	Mo	NH_4	Ni	NO_3	Pb	PO_4	Zn
Shannon spring	RO-16	<0.01	0.72	-	<0.01	<0.01	<0.01	<0.01	<0.01	0.46	<0.01	<0.01	0.09	<0.01	<0.05	0.03	<0.1	<0.01
Rock Corral spring	RO-17	<0.01	0.03	-	0.01	<0.01	<0.01	<0.01	<0.01	0.03	<0.01	<0.01	0.06	<0.01	<0.05	<0.05	<0.1	<0.01
Old Indian spring	RO-3	<0.01	0.05	-	<0.01	<0.01	<0.01	<0.01	<0.01	0.02	<0.01	<0.01	<0.01	<0.01	19.3	<0.02	6.5	<0.01
Bailey spring	RO-28	<0.001	-	<0.05	0.21	<0.001	<0.001	<0.001	<0.001	<0.1	<0.02	<0.1	-	<0.001	3.0	<0.001	<1	<0.02
Bailey spring	RO-2	<0.01	<0.03	-	0.21	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	0.01	<0.01	0.59	<0.02	<0.1	<0.01
Well Milford No.1	RO-19	<0.01	<0.03	-	0.04	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	0.08	0.01	0.94	<0.02	<0.1	0.02
Willow spring	RO-18	<0.01	<0.03	-	0.02	<0.01	<0.01	<0.01	<0.01	0.02	<0.01	<0.01	0.03	<0.01	13.8	0.02	0.3	<0.01
Beaver River	RO-21	<0.01	<0.03	-	0.03	<0.01	<0.01	<0.01	<0.01	0.03	0.02	<0.01	0.21	<0.01	0.20	0.04	<0.1	0.01
Well 28-10-14	RO-10	<0.01	0.04	-	0.04	<0.01	<0.01	<0.01	<0.01	0.01	<0.01	<0.01	0.04	<0.01	3.47	<0.02	0.3	<0.01
Kirk spring	RO-15	<0.01	0.07	-	0.03	<0.01	<0.01	<0.01	<0.01	0.08	<0.01	<0.01	0.22	<0.01	0.83	0.04	0.8	<0.01
Beaver River	RO-33	<0.001	-	<0.05	0.06	<0.001	<0.001	<0.001	<0.001	<0.1	<0.02	<0.1	-	<0.001	<0.1	<0.001	0.7	0.02
Well Kaufmann	RO-4	<0.01	0.005[b]	-	0.03	<0.01	<0.01	<0.01	<0.01	0.38	0.06	<0.01	0.18	<0.01	0.49	0.03	0.3	0.02
Well Anderson	RO-5	<0.01	0.002[b]	-	0.07	<0.01	<0.01	0.01	<0.01	0.78	0.16	0.01	0.15	<0.01	13.7	<0.02	<0.5	0.08
Well Jefferson	RO-25	<0.001	-	<0.05	0.48	<0.001	<0.001	0.002	<0.001	0.1	0.03	<0.1	-	<0.002	0.8	<0.001	<1	<0.02
Well Jefferson	RO-14	<0.01	0.09	-	0.48	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	0.08	0.02	2.10	0.05	<0.5	<0.01
Well WOW-2	RO-8	<0.01	0.004[b]	-	0.90	<0.01	<0.01	<0.01	<0.01	1.78	0.23	<0.01	0.19	0.02	1.23	<0.02	<0.5	0.04
Well 26-9-18	RO-6	<0.01	0.003[b]	-	0.61	<0.01	<0.01	<0.01	<0.01	11.1	1.34	<0.01	0.05	0.02	<0.05	0.08	<0.5	<0.01
Well 26-9-18	RO-30	<0.001	0.005[b]	0.08	0.89	<0.001	<0.001	0.003	<0.001	3.5	3.2	<0.1	-	0.002	<0.1	<0.001	<1	0.15
Well 26-9-18	RO-32	<0.001	0.007[b]	0.10	0.85	<0.001	<0.001	0.001	<0.001	1.9	2.4	<0.1	-	0.002	<0.1	<0.001	<1	0.26
Well 26-9-18	RO-29	<0.001	-	0.13	0.78	<0.001	0.002	0.003	<0.001	4.1	2.8	<0.1	-	0.002	<0.1	<0.001	<1	0.22
Well 26-9-18	RO-31	<0.001	-	0.10	0.93	<0.001	0.002	0.001	<0.001	2.7	3.3	<0.1	-	0.002	<0.1	<0.001	<1	0.26
Well OH-5	RO-13	<0.01	0.005[b]	-	2.24	<0.01	<0.01	0.01	<0.01	8.3	0.15	0.01	0.58	0.03	<0.1	0.06	<0.4	0.11
Well OH-1	RO-27	<0.001	-	4.3	0.64	<0.001	0.006	0.009	<0.001	0.39	0.17	<0.1	-	0.011	<0.1	<0.001	<1	0.13
Well OH-1	RO-9	<0.01	0.04	-	0.83	<0.01	<0.01	0.02	<0.01	0.03	0.32	0.02	0.32	0.03	1.78	0.10	<0.5	1.79
Well 26-10-26	RO-24	<0.001	-	<0.08	0.69	<0.001	<0.001	0.002	<0.001	1.1	0.11	<0.1	-	0.001	3.2	<0.001	<1	0.02
Well 26-10-26	RO-7	<0.01	0.004[b]	-	0.71	<0.01	<0.01	0.02	<0.01	7.3	0.11	0.04	0.51	<0.01	1.28	0.10	<0.5	<0.01
Well WOW-3	RO-11	<0.01	<0.001[b]	-	2.36	<0.01	<0.01	<0.01	<0.01	0.01	0.14	<0.01	0.40	0.01	<0.05	<0.02	<0.5	0.05
Well WOW-3	RO-26	<0.001	0.004[b]	0.06	2.03	<0.001	<0.001	<0.001	<0.001	<0.1	0.16	<0.1	-	0.002	<0.1	<0.001	<1	0.04
Roosevelt seep	RO-1	<0.01	0.008[b]	-	0.22	<0.01	<0.01	0.01	<0.01	2.09	1.35	0.02	0.09	<0.01	<0.2	0.10	<0.5	<0.01
Roosevelt seep	RO-22	<0.001	0.016[b]	1.35	0.25	<0.001	0.005	0.002	<0.001	1.8	1.10	<0.1	-	0.002	<0.1	<0.001	<1	<0.02
Well OH-4	RO-12	<0.01	0.004[b]	-	0.32	<0.01	<0.01	0.01	<0.01	6.4	0.35	0.02	0.21	0.03	<0.2	0.14	<0.5	0.23
Well Utah State[c]	14-2	<0.01	0.3	3.0	0.24	-	-	-	<0.01	0.08	<0.2	-	0.21	-	1.3	0.1	0.5	0.05

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[a] The sample arrangement is the same for Tables II to VI and is explained in Table III, note a. The chemical parameters sequence is in alphabetical order.

[b] These Al^{+++} values correspond to the ionized aluminum (filtration and extraction in the field), while the other Al results represent the total aluminum (filtration and acidification only).

[c] Average composition from 1 to 4 analyses. If data are not available, values for well Roosevelt KGRA 54-3 are given (Capuano and Cole 1982; Ross et al. 1982).

TABLE V
ISOTOPIC COMPOSITION OF THE WATERS FROM ROOSEVELT HOT SPRINGS AREA[a]

Name	No.	Date	$\delta^{18}O(°/_{\circ\circ})$[b]	$\delta^{2}H(°/_{\circ\circ})$[b]	$^{3}H(TU)$
Shannon spring	RO-16	5/82	-14.05	-101.9	29 ± 3
Old Indian spring	RO-3	5/82	-14.30	-105.2	-
Bailey spring	RO-28	4/83	-14.60	-106.8	33.0 ± 0.9
Bailey spring	RO-2	5/82	-	-	31 ± 4
Well Milford No.1	RO-19	5/82	-15.25	-114.4	0.8 ± 0.6
Beaver River	RO-21	5/82	-12.60	-99.7	-
Beaver River	RO-33	4/83	-11.85	-93.6	33.9 ± 0.7
Well Anderson	RO-5	5/82	-15.55	-116.8	3.1 ± 0.5
Well Jefferson	RO-14	5/82	-14.50	-112.7	0.5 ± 0.5
Well 26-9-18	RO-6	5/82	-14.35	-113.7	1.3 ± 0.4
Well 26-9-18	RO-32	4/83	-	-	0 ± 0.07
Well 26-9-18	RO-29	4/83	-	-	0 ± 0.08
Well OH-5	RO-13	5/82	-13.20	-106.8	0.6 ± 0.4
Well OH-1	RO-9	5/82	-12.80	-108.9	0.8 ± 0.4
Well 26-10-26	RO-24	4/83	-	-	0.16 ± 0.08
Well WOW-3	RO-11	5/82	-13.25	-112.7	1.1 ± 0.7
Roosevelt seep	RO-1	5/82	-12.20	-109.5	1.1 ± 0.4
Roosevelt seep	RO-22	4/83	-	-	1.50 ± 0.10
Utah State well[c]	14-2	5/78	-13.6	-116	-

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[a] The sample arrangement is the same for Tables II to VI and is explained in Table III, note a.

[b] Precision of the stable isotopes: $\delta^{18}O = \pm 0.15°/_{\circ\circ}$; $\delta^{2}H = \pm 0.5°/_{\circ\circ}$.

[c] Average values of two different samplings (Rohrs and Bowman 1980).

possibly due to chemical reactions occurring in special local lithologic conditions. Cold and dilute springs display a pH between 6.8 and 7.7, essentially neutral and are easily modified by emergence conditions (air contact, cattle contamination). Finally, the Beaver River has an average pH of 8.5, typical for a surface water.

The redox potential measured in the mineralized water wells shows mostly weak reducing conditions, with Eh values averaging -33 mV, with the exception of well WOW-3, which has a mean positive value of 107 mV. Dilute water wells and springs have weak oxidizing conditions displayed by positive Eh values from 49 to 285 mV. It can be concluded that all these natural waters are relatively close to stability (Eh $\cong$ 0 mV), under neither strongly reducing nor strongly oxidizing conditions.

3.3 Chemical Parameters

All chemistry for the major, minor, and trace elements is listed in Tables III and IV.

3.3.1 Chemical Classification and Mixing Patterns

Several types of water can be distinguished in the Roosevelt area by means of a chemical classification based on the major ions. In the scheme used here, a chemical "formula" is derived using the ions whose concentration reaches more than 10% of the total dissolved ions (TDI), expressed in meq/l. Only three cations (Na, Ca, Mg) and three anions (HCO_3, SO_4, Cl) reach the required 10% TDI, a circumstance observed in most other areas (Vuataz 1982). Sometimes five of these six ions each account for more than 10% of the TDI, and sometimes only two form the chemical formula (Table VI). The surface manifestation of the geothermal reservoir sampled at the Roosevelt seep contains Na and Cl for a total of 41% and 47%, respectively, of the TDI. The chloride content represents the dominant anion in all studied groundwaters when the TDS reaches 1000 mg/l, which means that groundwaters which do not mix with deep fluids have a relatively low mineralization (300-500 mg/l). Sodium behaves the same way, except at the Anderson well. All the other waters (shallow wells, springs, or streams) are bicarbonated, with calcium as the main cation, except for well Milford No. 1 and well 28-10-14. The other ions accounting for the chemical formulas show different orders of concentration, according to local mineralogical variations within the aquifer.

TABLE VI
CHEMICAL CLASSIFICATION OF THE WATERS FROM ROOSEVELT HOT SPRINGS COMPARED WITH THEIR TOTAL DISSOLVED SOLIDS[a]

Water types	Classification[b]	TDS[c] (mg/l)
Shannon spring	$Ca>Na;HCO_3>Cl>SO_4$	43
Rock Corral spring	$Ca>Na;HCO_3>>Cl>SO_4$	97
Old Indian spring	$Ca>>Na;HCO_3$	170
Bailey spring	$Ca>>Na;HCO_3$	240
Well Milford No.1	$Na;HCO_3>>SO_4$	350
Willow spring	$Ca>Na;HCO_3$	410
Beaver River	$Ca>Na>Mg;HCO_3$	490
Well 28-10-14	$Na>Ca;HCO_3>>Cl$	270
Kirk spring	$Ca>Na;HCO_3$	420
Well Kaufmann	$Ca>Na>Mg;HCO_3>Cl$	440
Well Anderson	Ca>Mg>Na;Cl	1100
Well Jefferson	Na>Ca;Cl	1300
Well WOW-2	Na>>Ca;Cl	2000
Well 26-9-18	$Na>>Ca;Cl>>HCO_3$	4000
Well OH-5	Na>Ca;Cl	2900
Well OH-1	$Na;Cl>>HCO_3$	5400
Well 26-10-26	Na>>Ca;Cl	4200
Well WOW-3	Na;Cl	4800
Roosevelt seep	Na;Cl	6000
Well OH-4	Na;Cl	6100

[a] The sample arrangement is the same for Tables II to VI and is explained in Table III, note a.

[b] In the chemical formula, the ion sequence is made by decreasing concentrations of cations and anions, respectively. Reported concentrations in meq/l are higher than 10% of the sum of the ions in meq/l (= total dissolved ions or TDI). The symbol >> means that the concentration in meq/l of the ion on the left is equal to or greater than double the concentration of the ion on the right.

[c] If more than one sample is used, TDS values are averaged and rounded off.

A plot of the major and minor constituents against chloride displays different behaviors for the wells and the seep around Roosevelt Hot Springs (Fig. 6). Chloride has been chosen as the base parameter because it is a conservative ion, which does not react chemically, and can undergo dilution-concentration processes only but no precipitation. Moreover, chloride is the main anion of the geothermal fluid. The complex patterns that can be observed from these plots are attributed to the vertical and horizontal heterogeneity of the alluvial aquifer as well as the differential mixing with the fluid leaking out of the hydrothermal reservoir. Straight-line trends, which indicate mixing between dilute and mineralized waters, are better visualized for some parameters than for others. Na, Br, and B, originating from the deep system, display the best mixing lines because these three parameters are relatively conservative. However, OH-1, the deepest well (RO-27 on the plots), shows a significant enrichment in both Br and B, possibly caused by local precipitation processes of minerals deposited by the geothermal fluid or by evaporation. These processes are unusual and are not understood at this time.

Nonlinear trends of temperature-dependent parameters are apparent for K, Li, and SiO_2. These two cations and silica are widely used in chemical geothermometry to predict deep temperatures (Fouillac and Michard 1981; Fournier and Rowe 1966; White 1970). In the empirical formulas set for these geothermometers, the logarithm of concentration ratios or of the concentrations themselves is used. Therefore, the nonlinear trends of mixing for K, Li, and SiO_2 are similar to logarithmic curves. Other temperature-dependent parameters like Ca, Mg, and HCO_3 do not reveal any patterns; their content is either limited by temperature or increased by dissolution of localized Tertiary limestone beds, in the alluvium (Mower and Cordova 1974). Finally, for Sr, SO_4, and F, there are no obvious trends, either because the two waters mixing together have approximately the same concentrations (SO_4) or because more complex behavior may take place; for example, double positive trends with different slopes or positive and negative correlations, which have not been deciphered so far, may occur.

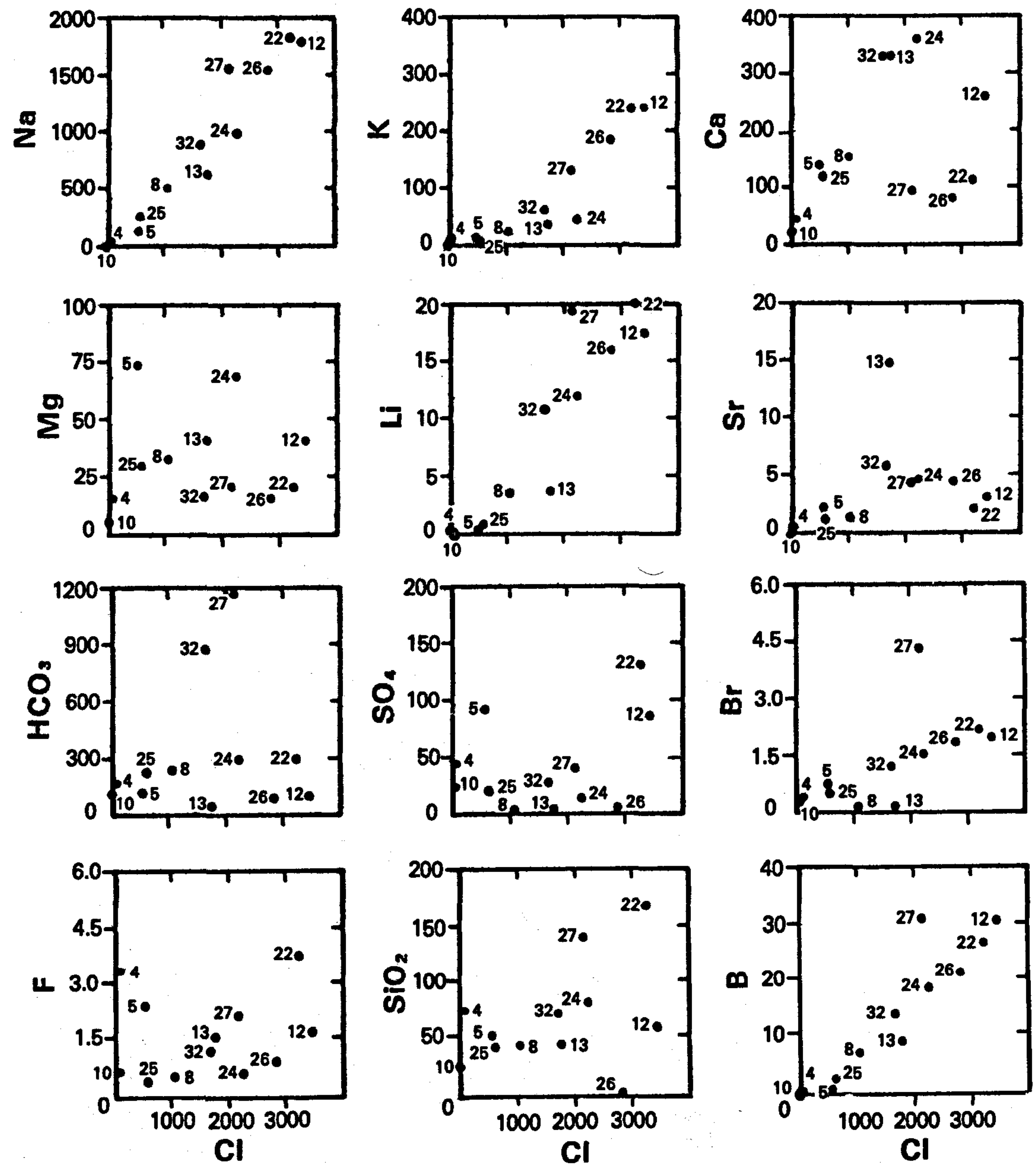

Fig. 6.

Multiparameter plots vs chloride (mg/l). Data are from Table III: all the wells from the present study are represented. In case two analyses are available, the most recent one has been considered. The numbers are the sample numbers.

3.3.2 Evaluation of the Contamination of the Shallow Aquifer

By considering the geothermal fluid of the well Utah State 54-3 (in the main production zone) to be undiluted by the local groundwater and therefore representative of the deep reservoir, Capuano and Cole (1982) calculated the average percentage dilution of a few geothermal wells. With chloride values of 3860 mg/l and 32 mg/l, regarded as 100% and 0% geothermal fluid respectively, the percentage of mixing has been calculated for the seep and the wells around Roosevelt Hot Springs (Fig. 7). The two wells located above the mixing line contain much more HCO_3 than the other samples do, and therefore, their TDS is substantially increased. On the other hand, wells situated under the mixing line, especially OH-4, WOW-3, and OH-5, have

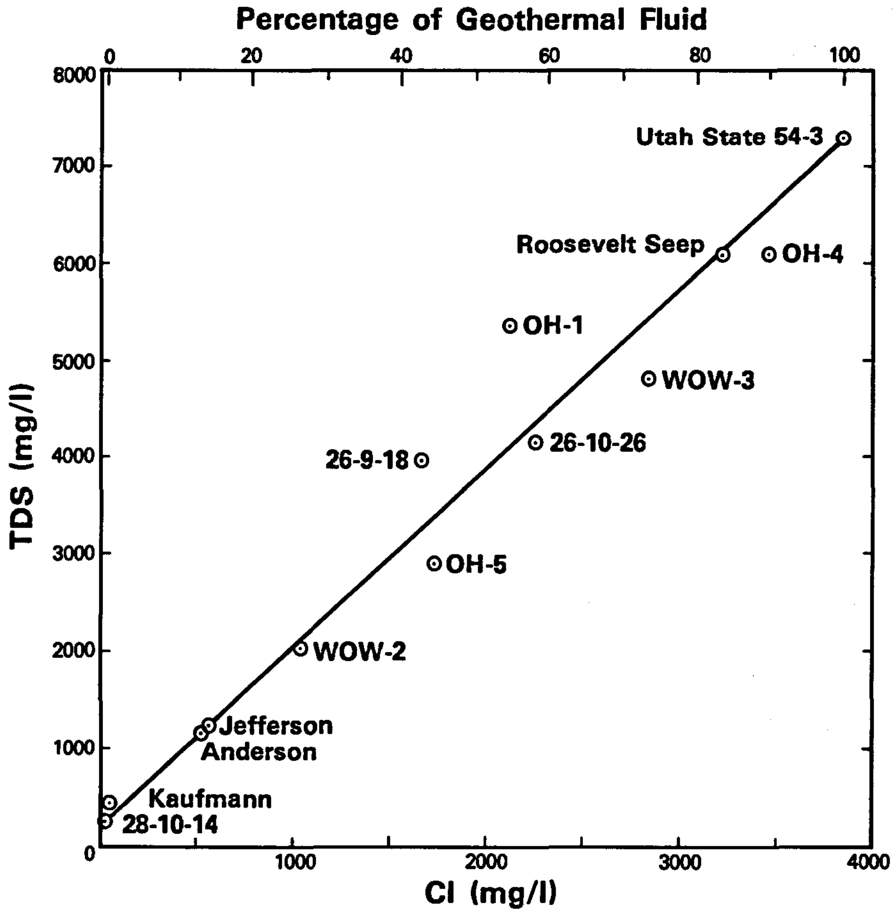

Fig. 7.
Percentage of geothermal fluid in the shallow alluvial aquifer. The mixing line stretches from well Utah State 54-3 (Capuano and Cole 1982) to well 28-10-14. The geothermal well is assumed to deliver unmixed deep geothermal fluid, whereas well 28-10-14 is thought to represent uncontaminated shallow groundwater.

apparently undergone silica and carbonate precipitation, whereas chloride and other conservative chemical parameters have not changed. A map of the percentage isovalues shows the relative influence of the geothermal fluid in the well waters around Roosevelt Hot Springs (Fig. 8). The limit of contamination of the aquifer extends roughly to a radius of about 13 km west from the Opal Mound Fault.

3.3.3 Trace Elements

Many trace elements have been analyzed for this study, but because most of these waters are relatively dilute (TDS = 40 to 6100 mg/l), most trace concentrations are at or below the detection limits of the analytical methods. Traces like Ag, Cd, Co, Cr, Cu, and Mo do not present any anomalies; their detection limits usually range from 0.001 to 0.1 mg/l, and even if the analyzed contents slightly exceed this detection limit, they cannot be considered as quantitative results. However, the following trace concentrations have been regarded as anomalous for specific types of water.

The ionized form of Al^{+++} reaches an average content of 0.012 mg/l for the Roosevelt seep, which is about 3-4 times the average concentration of the dilute shallow groundwater. Arsenic deposits, very common in active geothermal fields, have been found at Roosevelt Hot Springs associated with siliceous sinter or cemented alluvium (Christensen et al. 1983). In this study, As content reaches a maximum of 4.3 mg/l for well OH-1 and 1.35 mg/l for the Roosevelt seep, both locations being very close to the main fault zone. Anomalous Ba concentrations have been found associated with manganese and iron-oxide deposits. Only two water samples display anomalous Ba content (averaging 2.3 mg/l), and they are wells WOW-3 and OH-5. High Fe values were obtained in stagnant well water, namely 11.1 and 8.3 mg/l for wells 26-9-18 and OH-5, respectively. These high values are thought to originate from rusted iron casings. Manganese concentrations around 3 mg/l were present in the water of well 26-9-18 during the pumping test. Also the Roosevelt seep displays 1.35 mg/l of Mn, linked with the nearby manganese deposits of the Quaternary siliceous sinters. Ni (0.03 mg/l), Pb (0.14 mg/l), and Zn (1.79 mg/l) have the highest contents in the waters of the three warmest and deepest wells, namely OH-5, OH-1, and OH-4. Ammonium, which is commonly associated with hot geothermal fluids, is found to reach 0.58 mg/l in

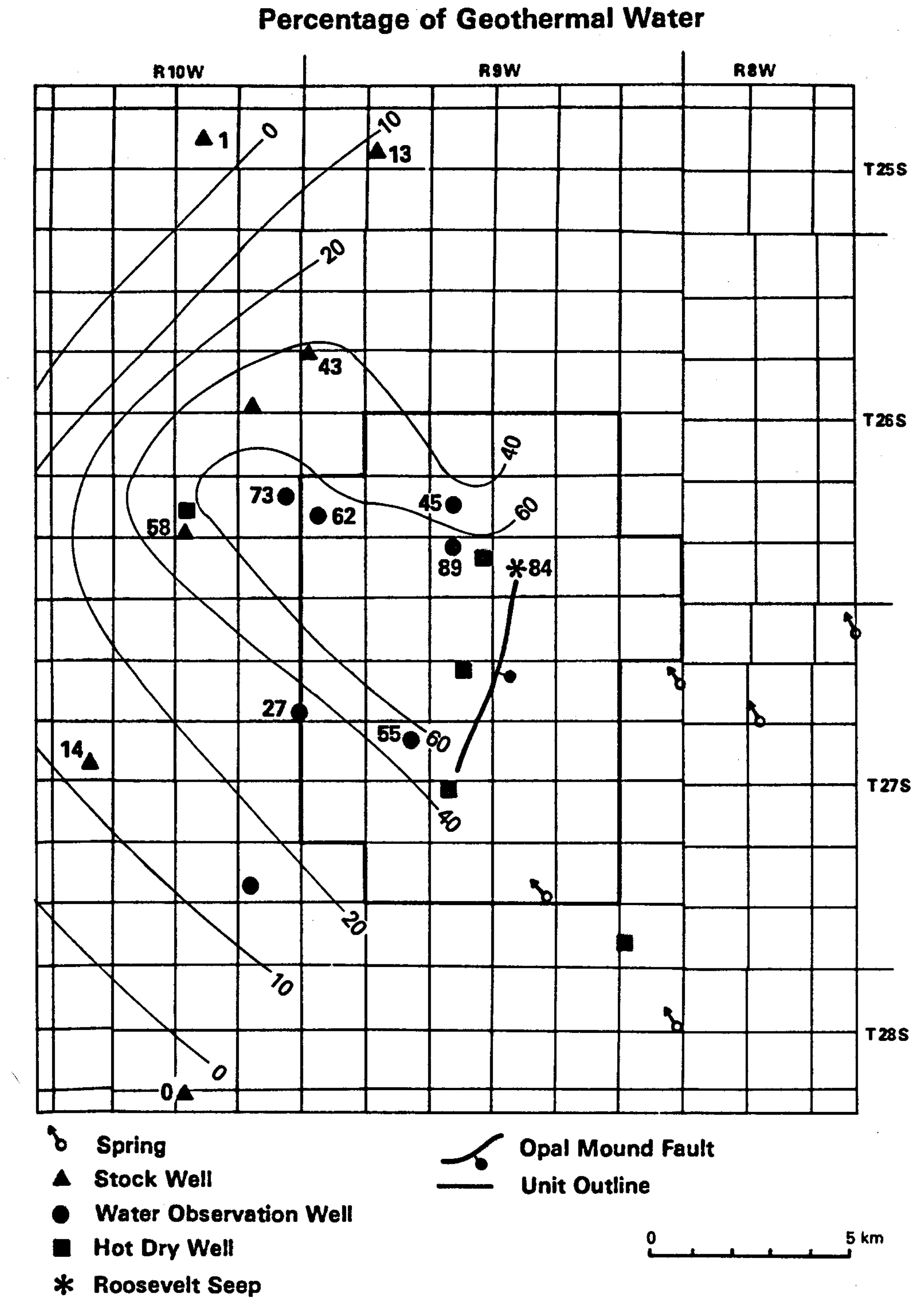

Fig. 8.
Map of isovalues of the percentage of geothermal water in the alluvial aquifer. Single values are given for each well sampled.

well OH-5 at a temperature of 83°C. Finally, NO_3 and PO_4 show very high values for cold dilute shallow groundwaters, which may have been in contact with manmade or cattle pollution. In conclusion, trace concentrations are generally more strongly correlated with the temperature and depth of the wells than with the total mineralization. This probably means that trace concentrations increase with proximity to the source geothermal reservoir rather than with percentage of mixing with reservoir fluids.

3.3.4 Quality and Use of the Aquifer

Due to intense irrigation, the alluvial aquifer water level has been strongly depleted in the Milford and Minersville areas south of this investigation (Mower and Cordova 1974). Heavy precipitation does not compensate for this deficit because the region is semiarid. However, north of Milford and especially in the HDR interest area, only two irrigation wells draw water during the growing season (Anderson wells at location (C-25-9)29 (Fig. 3). The development of an important irrigation well field in this area is unlikely, considering the quality of the water in the alluvial aquifer, even at the shallowest depths.

Concentration limits for water use have been set up by different agencies, and some of these standards are presented in Table VII for three different uses (drinking water, livestock, and irrigation). When these values are compared with the chemical analyses of the shallow aquifer in Table III, it appears that only the peripheral wells (28-10-14 and Anderson) can be used for irrigation. The Kaufmann, Jefferson, and 26-10-26 wells contain too much F, B, and Fe, respectively, and well 26-9-18 has six elements exceeding the maximum levels for irrigation water (As, B, F, Fe, Li, Mn).

Of the toxic elements discussed above, B is very soluble, volatile, and mobile, originating from the deep reservoir and conveyed by boiling as well as by mixing processes toward the surface. For these reasons, B has been chosen as a parameter to represent the maximum extension of the contaminated aquifer whose chemical characteristics are not favorable for irrigation use (Fig. 9). From this map, a surface area limited by the north-south trend of the Opal Mound fault and by the estimated contour of 1 mg/l of B has been roughly evaluated to cover 220 km^2. We conclude that this area is not suitable for irrigation. On the other hand, this relatively large portion of

TABLE VII
SELECTED CHEMICAL STANDARDS FOR WATER USE [a]

Parameters	Concentration limits (mg/l)		
	Drinking water	Livestock	Irrigation
As	0.05	0.20	0.10
B	1.0	5.0	0.75
Ba	1.0	1.0	1.0
Cl	250	-	-
F	2.4[b]	2.0	1.0
Fe	0.30	-	-
Li	-	-	5.0
Mn	0.05	-	2.0

[a] Source: U.S. Department of Interior (1968); U.S. Environmental Protection Agency (1973, 1975).

[b] Maximum level of fluoride for an annual air temperature ≤12°C.

the alluvial aquifer could be available for industrial application, such as water supply for a hot dry rock heat extraction system.

3.4 Isotopic Parameters

3.4.1 Stable Isotopes: Oxygen-18 and Deuterium

The stable isotope compositions of the waters from the Roosevelt Hot Springs area display a relatively large range of concentrations (Table V, Fig. 10). Along and slightly below the meteoric water line the heaviest isotope values occur, such as Shannon and Old Indian Springs, as well as the lightest values, represented by wells Anderson and Milford No. 1. All the springs from the Mineral Mountains are located between these values.

3.4.1.1 Isotopic shift from the meteoric water line. All the other samples are off the meteoric water line and seem to follow different trends,

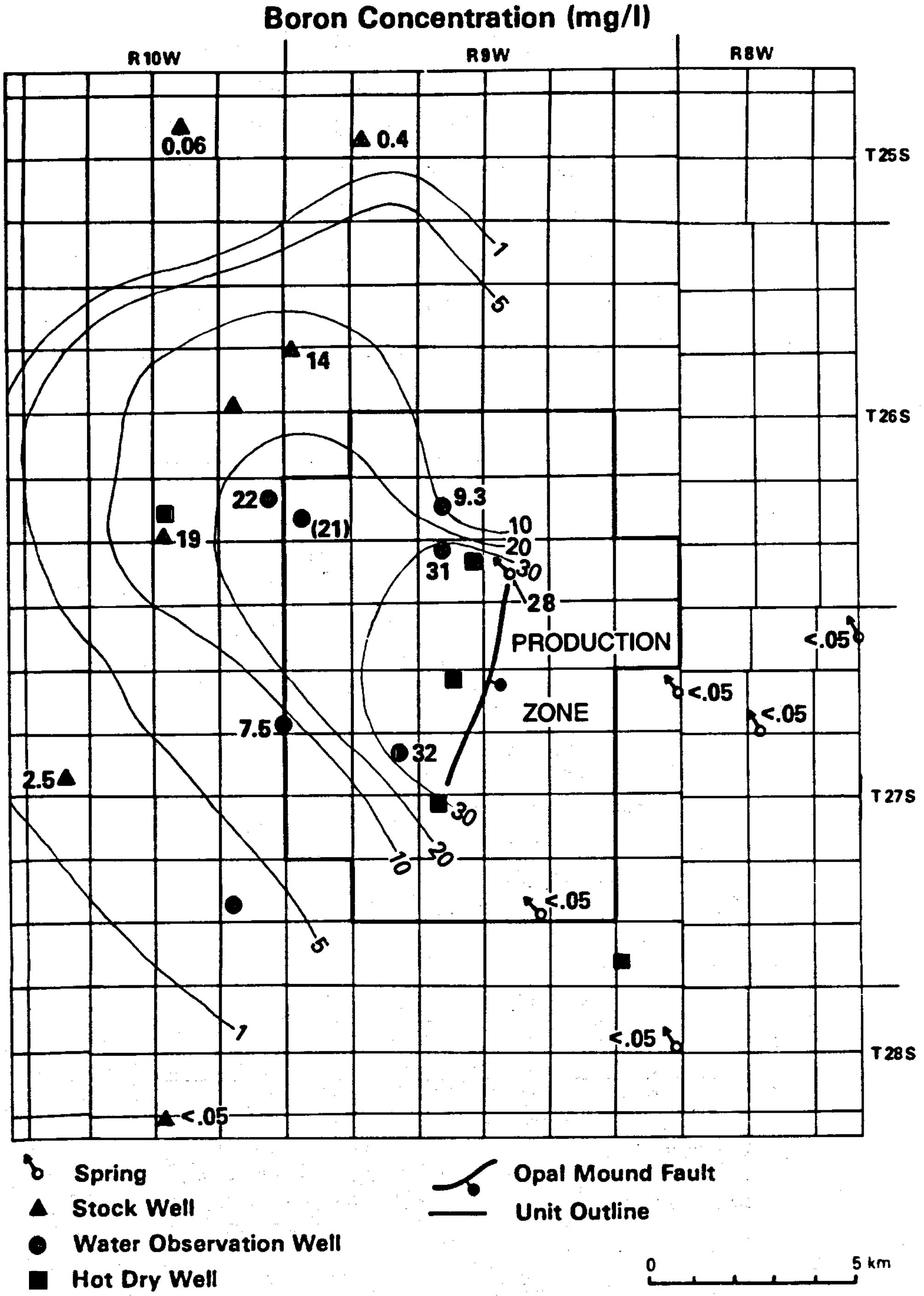

Fig. 9.
Map of boron concentration (mg/l) isovalues in the alluvial aquifer. Single values are given for each sampling point.

with various slopes, all less than 8 (Fig. 10). Some of these trends have already been discussed (Rohrs and Bowman 1980). The geothermal fluid, here represented by Utah State well 14-2, displays a slope ≅0 and an isotopic shift (^{18}O enrichment) estimated to be +1 to +2 °/₀₀, which is smaller than the usual shifts observed in other geothermal fields (Truesdell and Hulston, in Fritz and Fontes 1980). The surface manifestation of the deep reservoir (Roosevelt seep) has a more important shift but a slope ≅2, which can be explained by at least two processes: (1) subsurface boiling and steam separation and (2) evaporation at shallow depth or in the pool that the spring forms (Panichi and Gonfiantini 1978). Finally, the Beaver River (with a slope ≅4) is typical for evaporation conditions and therefore both ^{18}O and ^{2}H are enriched.

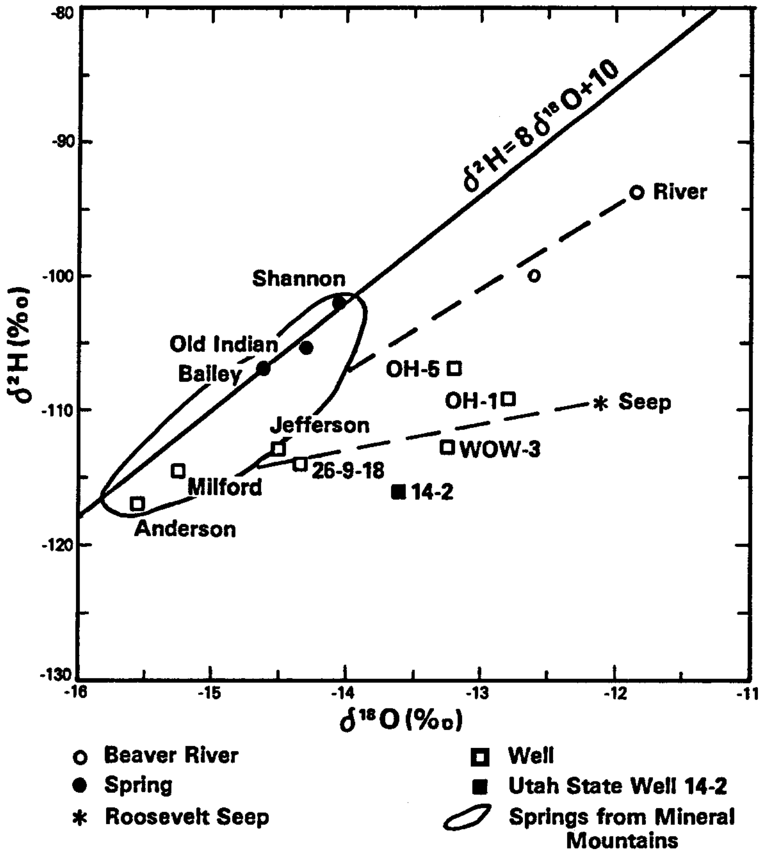

Fig. 10.
Plot of oxygen-18 vs deuterium. The solid line represents the world meteoric water line (Craig 1961). The field showing the spring compositions from the Mineral Mountains includes samples from 1976-1977 (Rohrs and Bowman 1980), as well as from the present study (1982-1983). The dashed lines show different trends of isotopic composition (see text).

Between these three extreme ^{18}O and ^{2}H compositions and the meteoric water line, five samples collected from wells in the HDR zone range from the composition of the Anderson and Milford No. 1 wells toward the composition of Roosevelt seep. Two of these wells, OH-5 and OH-1, have a stronger ^{2}H enrichment, probably due to their temperatures (83° and 54°C) and their deep water levels (185 and 155 m), which indicate either boiling in the reservoir and/or evaporation within the casing. The water in these two wells is stagnant, and the samples taken with a bailer come from the top of the water column. Other samples collected at the bottom of the wells might show a different isotopic composition. A similar case was described for a 54°C thermal water well at Ojo Caliente, New Mexico (Vuataz et al. 1984).

A relatively good linear correlation exists between ^{18}O and Cl for the wells drilled in the alluvial aquifer and Roosevelt seep (Fig. 11). This ^{18}O enrichment linked with high Cl content can be explained by differential mixing of the water contained in the alluvial aquifer with geothermal fluid. The sample scatter results from deep fluid evolving to various isotopic

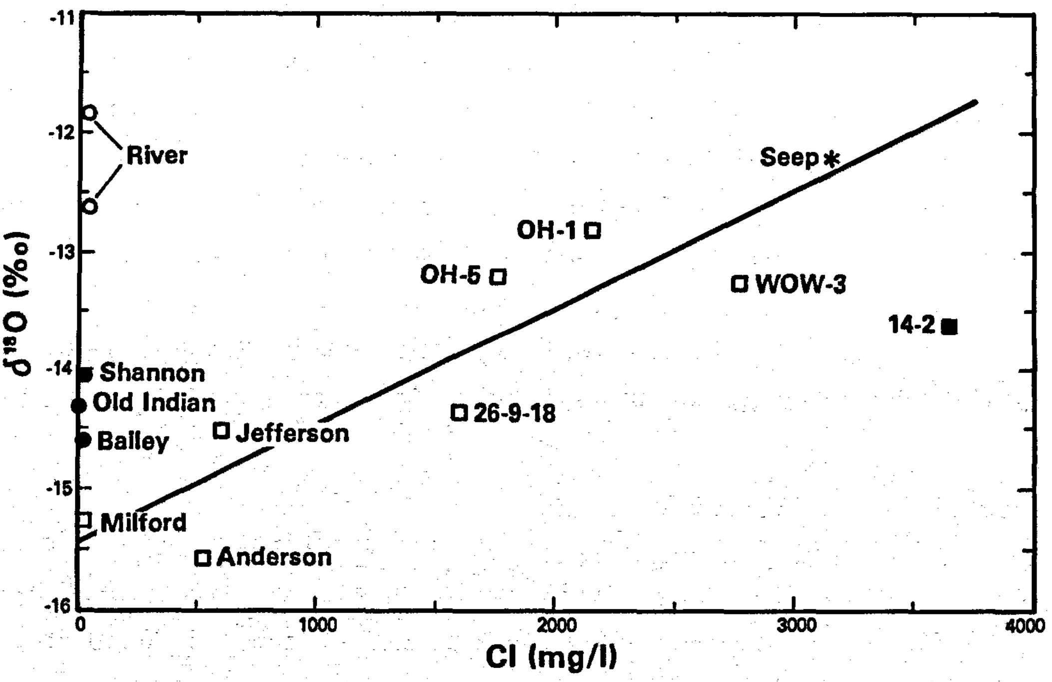

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Fig. 11.
Oxygen-18 and chloride correlation. The linear regression (r = 0.915) was calculated without the samples from the geothermal well, the river, and the two springs. Symbols: same as for Fig. 10.

compositions by fractionation during steam loss, resulting in the heavy isotopes being concentrated in the solution (Bath and Williamson 1983).

3.4.1.2 Recharge area evaluation. Previous work has been done to try to estimate the origin of the geothermal fluid. Rohrs and Bowman (1980) used the stable isotope composition of various types of water from Milford Valley in the west to the Tushar Mountains in the east, but they did not determine whether the water of the hydrothermal reservoir originates entirely from recharge in the Mineral Mountains or partially from the Mineral Mountains and partially from the Tushar Mountains by means of an interbasin flow. A more detailed study should be carried out including more sampling at different seasons and the computation of an isotope gradient as a function of spring recharge elevations. A model based on geologic and geohydrologic parameters (Smith 1980) has revealed, however, that there is little chance that the geothermal reservoir is significantly recharged from the Tushar Mountains.

It is noticeable in the isotopic results that none of the wells analyzed contain less deuterium than the lowest values displayed by springs from the Mineral Mountains (Fig. 10). Moreover, when looking at the average deuterium content of the waters, a similar value of -112°/₀₀ is found for the 7 wells (in 1982-1983) and 13 springs from the Mineral Mountains (in 1976-1977) (Rohrs and Bowman 1980). Although the Tushar Mountains have a higher average elevation than the Mineral Mountains and therefore the isotopic composition of their springs is usually lighter, one cannot preclude the existence of some interbasin flow, especially because the range of isotopic composition between the two domains overlaps (Rohrs and Bowman 1980). In conclusion, the isotopic data suggest that the alluvial aquifer west of Roosevelt Hot Springs may be totally recharged by the Mineral Mountains watershed.

3.4.2 Radioactive Isotope: Tritium

3.4.2.1 Tritium in surface waters. Before the first atmospheric nuclear tests, the natural production of tritium in surface water reached 5-10 TU. In 1963, artificial tritium concentrations peaked at about 1000 TU in surface waters and at several thousand TU in precipitation. Since the moratorium on atmospheric tests, tritium content has been decreasing. Six stations for precipitation (Utah, Colorado, Arizona, and New Mexico) and two stations on the Colorado River (Utah and the Arizona-California border) are averaged and plotted on a histogram showing the evolution of tritium content in the surface waters of the southwestern United States as a function of time (Fig. 12). During the last five years (1978-1982), tritium content in precipitation has fallen from 75 to 49 TU and from 110 to 60 TU in river water (WATSTORE 1983).

Using the 12.46 year half-life period of tritium and the records of tritium content in surface waters from previous years, it is possible to date or at least to give an order of magnitude age to groundwaters. For this study the Beaver River has been analyzed for tritium. In April 1983, its concentration reached 34 TU (Table V), indicating that the river is composed primarily of young water. This value is considerably lower than the last tritium result available from the station on the Colorado River near Cisco, Utah (54 TU in January 1983). However, it is double the tritium content (16 TU) in the precipitation in Salt Lake City, Utah, for the period October 1982 to February 1983 (WATSTORE 1983). The large difference observed between the Beaver River and the Colorado River could be reduced by knowing the tritium content for the two rivers during the same period, because the radioactive isotope concentration is usually lower in early spring than it is during winter. On the other hand, regional climatic and topographic effects may play a significant role (Fontes, in Fritz and Fontes 1980). For instance, the two rivers have very distinct regimes and origins, and their ratio of groundwater to runoff water might be quite different.

3.4.2.2 Tritium in groundwaters. Except for the two springs sampled in the Mineral Mountains, tritium concentration of the groundwaters (dilute or mineralized, cold or hot) is very low, ranging from 0 to 3 TU (Table V). Shannon and Bailey springs display tritium values similar to surface water (29-33 TU), which means a very short underground residence time and probably a major component of snowmelt from the same season.

The tritium contents of the wells and the seep do not seem to be related to any parameters like mineralization, temperature, depth in the aquifer, or even distance from recharge or mixing zones. However, a specific correlation might be masked by the fact that the 1982 and 1983 analyses were carried out by different laboratories having different analytical precisions. This may explain why no trend is apparent in a very small range of concentration (0-3 TU).

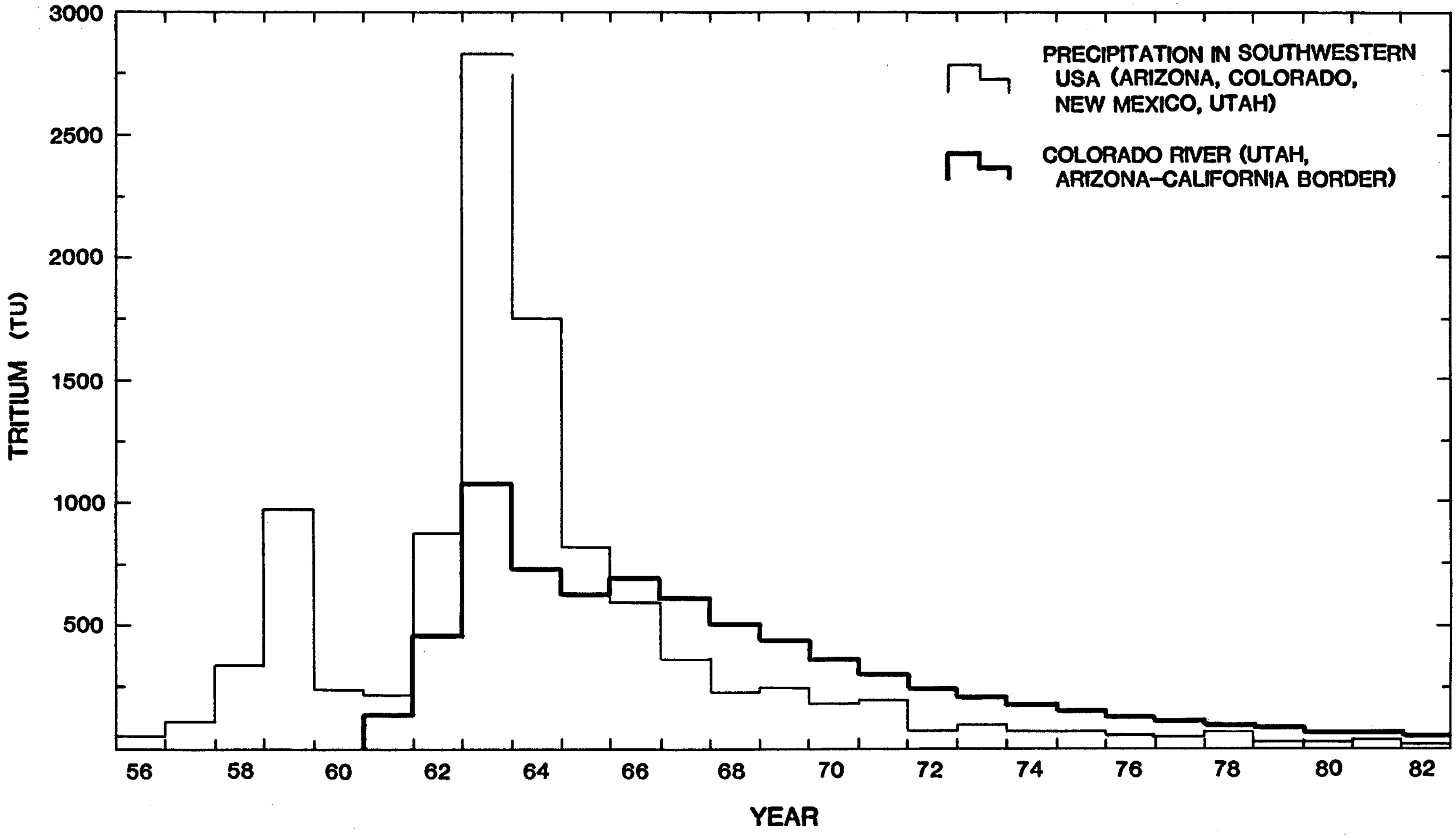

Fig. 12.

Tritium-time variations in southwestern U.S., using data from regional precipitation and the Colorado River (Vuataz et al. 1984). The thin line represents the yearly average tritium concentration for one to six precipitation stations according to the available data (Salt Lake City, Utah; Denver, Colorado; Flagstaff, Arizona; Albuquerque, New Mexico; Mt. Withington, New Mexico). The thick line represents the yearly average tritium content for one to two stations on the Colorado River (Cisco, Utah; above Imperial Dam, Arizona-California). All values are from a data bank of the U.S. Geological Survey (WATSTORE 1983).

If the Anderson well has a little more tritium than the others (3 TU), it has to be noted that in May 1982 this shallow irrigation well was intensively pumped (3.8 m^3/min). The water sample was taken from an observation hole located about 150 m west of the main well but still within the influence of the depression cone that could cause an increased rate of recent near-surface water percolation, and therefore a higher tritium content. The Roosevelt seep, which is consistently $\leqq 1$ TU, is likely to be slightly contaminated by runoff water. The wells WOW-3, OH-1, and OH-5, with tritium contents from 1.1 to 0.6 TU, have no pumps, and the water is stagnant in the casing, which leaves the possibility of very small contamination by surficial waters. The wells 26-9-18, Jefferson, and 26-10-26 sampled in 1982 display tritium contents of 1.3 to 0.16 TU and were pumped during a short time only for the collection of the water. However, during our pumping test on well 26-9-18, two samples revealed no tritium at all.

Milford No. 1, a large water supply well in downtown Milford, which has been pumped almost constantly at 22.7 m^3/min for several years, shows an amazingly low tritium content of 0.8 TU. Because the water level has declined by 3 to 6 m in 22 years in the Milford area (Mower and Cordova 1974), the groundwater reservoir must have extremely large resources. Indeed, 0.8 TU indicates the presence of water primarily infiltrated before the first atmospheric nuclear test of 1953. If we assume a prebomb tritium content of 5 to 10 TU in precipitation, 30 to 45 years without any atmospheric contact are necessary for tritium decay to reach 0.8 TU. This means that the groundwater in Milford Valley may have infiltrated into the ground between the late 1930s and the 1940s. On the other hand, it may also be possible that the groundwater is totally tritium free, having been in the ground for much longer and then having been slightly contaminated by recently infiltrated waters around the wells.

In the hot dry wells area, two relatively shallow stock wells revealed the lowest tritium content: 0.16 TU for well 26-10-26 and 0 TU (twice) for well 26-9-18 during the pumping test described in a later section. There is no ambiguity with these results, and although the water pumped from the two wells is a mixture of two components, the waters from the hydrothermal reservoir and from the alluvial aquifer are both essentially tritium free and have a minimum underground residence time of 70 years. Using the chemical data, the estimated percentages of geothermal fluid for the wells 26-9-18 and 26-10-26 were 40 and 60%, respectively (Fig. 7).

4. CHEMICAL EQUILIBRIA

Chemical compositions of groundwaters depend on their interactions with reservoir rocks and on parameters such as temperature and pressure, time available for chemical reactions, rate of fluid flow, and water-rock ratio. The original composition of a thermal water ascending from its deep reservoir toward surface may be modified by precipitation or dissolution of minerals, due to changes in the above-mentioned parameters. The reaction states of the fluid, namely undersaturated, saturated, or oversaturated with respect to a solid mineral phase, can be calculated from a chemical equilibrium model. Kharaka and Barnes (1973) established a computer code, SOLMNEQ, which calculates states of reaction of numerous mineral and aqueous species generally present in natural waters over the temperature range of 0 to 350°C, with respect to 158 solid mineral phases for which thermodynamic data are available. The input consists of the chemical analysis as well as the fluid temperature, the pH, and the Eh. The output lists, among other calculated parameters, the change in the Gibbs free energy (ΔG, kcal/mol) (Barnes and Clarke 1969). Positive values of ΔG computed by the program indicate that the fluid is oversaturated with respect to the mineral and that this mineral may precipitate, whereas negative ΔG values indicate that the water is undersaturated with respect to that phase. Equilibrium between a mineral and water is theoretically attained if the Gibbs free energy equals zero (± 2 kcal/mol for this discussion). However, this cannot be taken as an absolute indication that a particular mineral is dissolving or precipitating. A favorable kinetic rate might allow the precipitation of a thermodynamically less-stable phase at the expense of a thermodynamically stable phase that has an unfavorable kinetic rate (Grigsby et al. 1983).

4.1 Water-Rock Interaction

The geochemistry of hydrothermal deposits and alteration around Roosevelt Hot Springs geothermal area has been extensively studied, and recent works include Parry et al. (1980), Glenn et al. (1981), and Christensen et al. (1983). Capuano and Cole (1982) studied the equilibria between the geothermal fluid produced by the deep wells and the altered rocks available from drill cuttings.

Chemical analyses of 14 wells and the Roosevelt seep were processed through the computer program SOLMNEQ to calculate the Gibbs free energy changes. The reaction states for 27 selected minerals that might be expected to participate in water-rock interactions are reported in Table VIII. The purpose of these calculations is to determine the type of chemical equilibrium existing between the contaminated aquifer and the minerals in the alluvium.

For most of the waters, it appears that, at emergence temperature, saturation or oversaturation ($\Delta G = \pm 2$ kcal/mol) occurs with respect to the primary silicates such as high albite, anorthite, microcline, quartz, biotite, and muscovite, which are the main constituents of granite and gneiss, the geothermal reservoir rocks. Also the thick alluvium series of the Milford Valley results from the weathering and alteration of the same type of rocks mentioned above as well as rhyolite; therefore, similar conditions may occur for water-rock interactions in the alluvial aquifer, except temperature and pressure, which are of course much greater in the deep reservoir. Also, numerous secondary silicates are in a state of saturation (low albite and silica) or oversaturation (clays, chlorite, zeolites, and scapolite). This could be explained by the extensive rate of hydrothermal alteration present in the Roosevelt Hot Springs area and by the plume of geothermal fluid carried toward the west-northwest by the shallow groundwater flow (Fig. 8). Three mud samples were collected for x-ray clay analysis, one from well OH-1 and two from well 26-9-18 during the pumping test. It appears that smectite (montmorillonite included) and kaolinite are the major clay constituents of the alluvial aquifer (D. Bish, Los Alamos National Laboratory, pers. commun. 1983).

TABLE VIII
REACTION STATES OF DISSOLVED MINERALS FOR WELL WATERS OF ROOSEVELT HOT SPRINGS
Values listed are Gibbs free energy difference at emergence temperature (kcal/mol)[a]

	Mineral phases	28-10-14 RO-10	Milford 1 RO-19	Kaufmann RO-4	Anderson RO-5	Jefferson RO-25	WOW-2 RO-8	OH-5 RO-13	26-9-18[b] RO-30	26-9-18[b] RO-32	26-10-26 RO-7	WOW-3 RO-26	OH-1 RO-27	Roosevelt seep RO-1	OH-4 RO-12	Utah[c] State 14-2
Primary silicates	Albite-High	-2.6	-2.3	0.70	-0.18	-0.24	-0.37	-4.0	-1.0	-1.6	1.3	-4.5	0.70	-1.7	-0.38	-3.7
	Anorthite	-4.1	-4.5	-0.60	-2.0	-2.6	-1.8	-4.4	-6.1	-7.2	-2.0	-5.5	-1.9	-10	-1.8	-6.6
	Microcline	0.61	0.50	4.8	3.4	2.6	2.3	-2.1	2.3	1.7	4.4	-1.0	3.4	1.9	2.8	-3.6
	Quartz	0.90	0.93	1.6	1.4	1.3	1.1	0.19	1.5	1.5	1.6	-0.67	1.4	1.9	0.98	-0.38
	Biotite-Fe	3.4	4.1	10	7.5	2.2	8.7	10	1.1	-1.7	7.9	6.3	8.3	-1.6	11	-7.0
	Muscovite	11	9.9	18	16	15	15	7.6	13	11	17	9.8	14	8.3	14	0.01
Secondary silicates	Albite-Low	-1.2	-0.82	2.2	1.3	1.2	1.0	-2.8	0.43	-0.09	2.8	-3.1	2.0	-0.28	0.96	-3.3
	Chalcedony	0.30	0.34	1.0	0.79	0.68	0.49	-0.23	0.93	0.93	1.0	-1.3	0.90	1.3	0.46	-0.39
	Cristobalite-α	0.04	0.08	0.76	0.53	0.42	0.23	-0.54	0.68	0.67	0.78	-1.5	0.62	1.1	0.18	-0.72
	Amorphous SiO_2	-0.41	-0.38	0.31	0.08	-0.03	-0.25	-1.1	0.22	0.22	0.33	-2.0	0.11	0.60	-0.32	-0.95
	Illite	2.0	1.6	8.4	6.4	5.6	5.1	-1.0	3.2	2.3	7.5	-0.78	4.9	0.24	4.5	-9.1
	Kaolinite	1.9	1.4	6.8	5.4	5.3	4.6	-0.19	3.5	2.9	6.5	-0.42	3.5	0.30	3.5	-8.1
	Montmorillonite-Ca	2.8	2.3	9.2	7.3	6.9	6.1	0.20	4.9	4.1	8.7	-1.9	5.5	1.5	4.9	-6.3
	Chlorite-Mg	11	13	13	12	6.9	9.5	9.0	-8.2	-11	5.4	11	9.0	-13	8.1	-41
	Laumontite	4.5	4.1	9.7	7.7	6.9	6.6	1.7	3.8	2.8	8.3	-0.13	6.2	0.28	5.5	-6.5
	Phillipsite	13	13	17	15	15	15	10	14	14	17	11	16	14	15	10
	Marialite	2.3	3.3	13	12	12	12	0.30	11	9.7	18	1.2	16	9.7	13	1.2
	Zoisite	-0.98	-1.7	4.0	1.7	0.56	1.2	-2.4	-5.7	-7.4	1.3	-2.8	-0.19	-13	-0.01	-9.9
Sulfates, carbonates, fluoride	Fluorite	-2.0	-2.3	0.08	0.07	-2.2	-2.0	-0.85	-0.43	-0.44	-1.5	-1.7	1.2	0.14	-0.74	-4.4
	Calcite	0.05	-0.03	-0.12	-0.22	-0.14	0.15	-0.60	-0.05	-0.23	0.18	0.28	0.69	-1.7	-0.40	-5.8
	Dolomite	-0.18	0.22	-0.20	-0.19	-0.45	0.14	-1.5	-1.2	-1.6	0.08	0.28	1.4	-3.6	-1.1	-19
	Witherite	2.3	2.5	1.9	1.5	2.7	2.9	1.1	2.5	2.3	2.6	4.1	2.9	0.57	0.93	-4.2
	Siderite	-1.9	-1.6	-0.03	-0.39	-1.4	0.54	0.59	0.25	-0.31	0.75	-0.91	0.73	-0.99	0.57	-3.8
	Strontianite	0.56	0.97	0.89	0.60	0.36	0.56	1.2	0.77	0.59	0.91	1.7	2.9	-0.75	0.85	-4.4
	Anhydrite	-3.3	-3.3	-2.9	-2.1	-3.0	-4.1	-3.6	-2.6	-2.6	-3.4	-4.7	-3.4	-2.5	-2.2	-10
	Barite	-0.92	-0.55	-0.65	-0.20	0.07	-1.2	-1.6	0.16	0.14	-0.77	-0.73	-0.89	-0.06	-0.68	-8.0
	Alunite	-15	-15	-6.1	-5.9	-7.1	-11	-18	-6.2	-6.6	-5.8	-15	-11	-9.2	-7.8	-42

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[a] The samples are arranged by increasing TDS.

[b] Samples RO-30 and RO-32 were collected at the beginning and the end of the pumping test, respectively.

[c] Chemical data needed for the computations are from Capuano and Cole (1982).

Nonsilicate minerals like carbonate and fluorite appear to be saturated for most of the fluid samples. Two minerals, which are known as alteration products in surface acid-altered zones of the area, namely anhydrite and alunite (Parry et al. 1980; Capuano and Cole 1982), are consistently undersaturated in the aquifer fluids according to the code.

Examining general trends within the water samples, the Utah State well 14-2 is undersaturated with respect to 20 out of the 27 selected minerals (Table VIII), as similarly found by Capuano and Cole (1982). Because of this fact and the well's high temperature of 268°C, the deep geothermal fluid must be considered very capable of causing hydrothermal alteration to the reservoir rocks. Although its chemical composition is relatively similar to Utah State well 14-2, the Roosevelt seep appears to be saturated with 18 of the 27 selected minerals, which is explained by conductive cooling of the geothermal fluid from 270° to 25°C, almost without dilution.

Because of its different environment, the stock well 26-9-18, which still contains more than 40% geothermal fluid, displays other equilibrium patterns. Primary silicates, the various forms of silica, and the nonsilicate minerals are saturated, whereas clays, chlorite, and low-grade metamorphic minerals appear to be oversaturated. The differences in the reaction states of dissolved minerals for samples RO-30 and RO-32 are very small. The first sample was taken at the beginning of the pumping test, while the second was collected at the end, but their mineralization is very similar. Finally, the most dilute water is in well 28-10-14, which seems to be out of the geothermal plume and shows a saturation state for the majority of the selected minerals.

4.2 Variations of Equilibria with Temperature in Cold Water Wells

4.2.1 Conditions Encountered in Hot Dry Rock Systems

If a future hot dry rock system is established at Roosevelt Hot Springs, the shallow aquifer could be used for make-up water in the circulation loop and heat extraction equipment. Therefore, knowing if minerals in solution can precipitate in the pipes and other surface equipment or if they will be dissolved from the deep reservoir rocks is of prime importance. Temperatures varying from 25° to 250°C will strongly affect the chemical equilibria of water. According to the location of the water supply well for the circulation loop, the alluvial aquifer may have a very different chemical composition, as functions of distance from the hydrothermal field and of well depth. A relatively cold dilute ($\leq$10 g/l) water will be injected at depth, conductively heated to 200°-250°C, extracted through a production well to a binary heat exchanger, and cooled during heat extraction to 25°-50°C before reinjection.

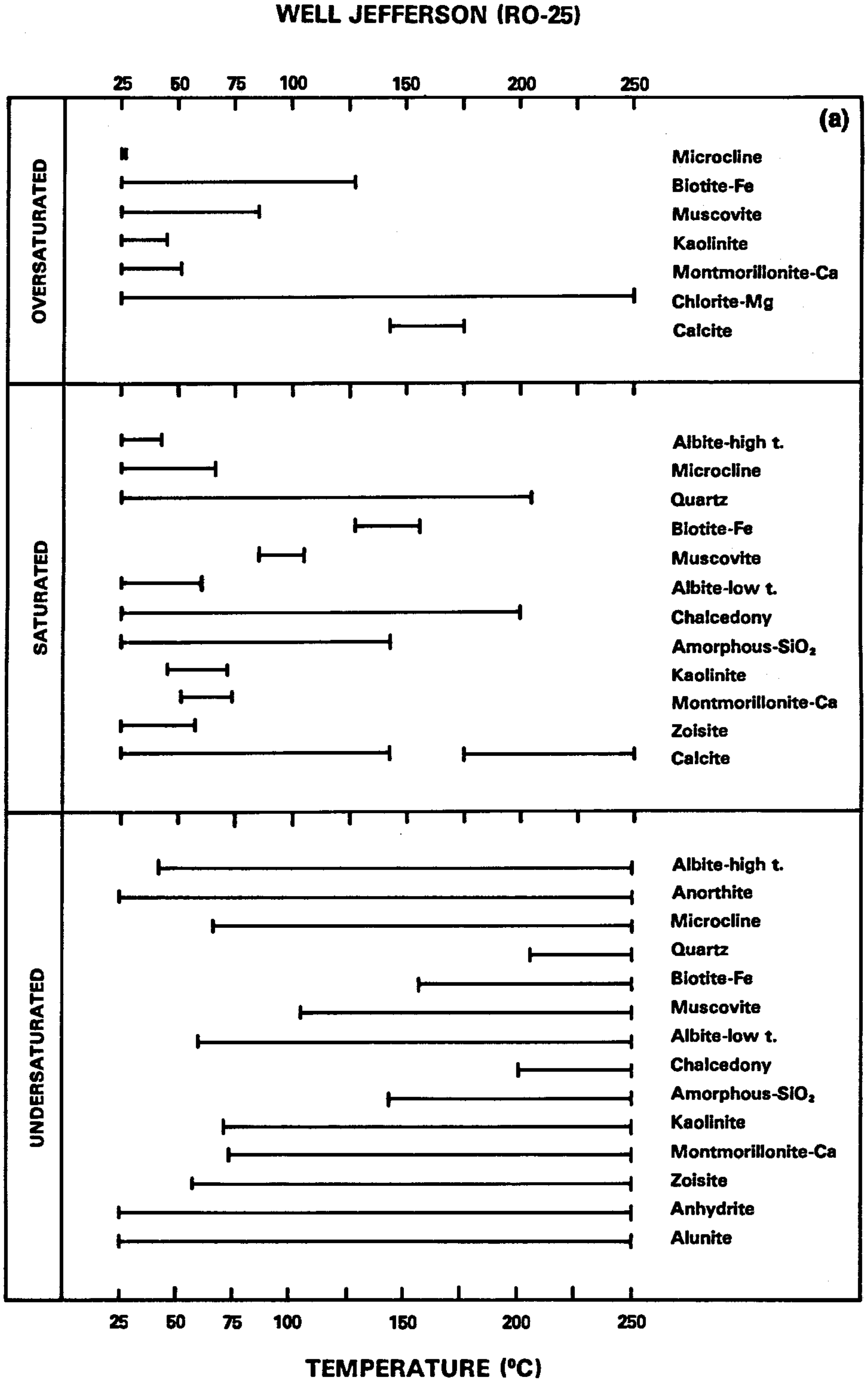
Previous experiments in the Phase I Hot Dry Rock system at Fenton Hill revealed that a certain amount of water was lost during the circulation, due to permeation into rock formations surrounding the fracture system. Although we do not know what conditions at Roosevelt will be like, it is assumed that about 10% ± 5% of the total flow rate should be continuously resupplied with fresh water to the circulation loop to make up for fluid losses (Dash et al. 1983). This is also a factor that will change the chemical equilibria. During the various experiments of the Phase I circulation loop, it was noted that steady-state levels of chemical concentrations were established after about 30 days. However, a gradual decline in dissolved elements was observed to occur after about 100 days (Smith and Ponder 1982). In the fractured reservoir, composed of relatively unaltered granitic rocks, two factors determine the water composition, namely water-rock interaction and the presence of pore fluid contained in the microfractures of the reservoir rock. This pore fluid, in chemical equilibrium with its environment, was leached by the injected water of the circulation loop and strongly participated in the chemical evolution of the steady-state fluid. The total mineralization of fluid in the Phase I system under steady-state conditions did not exceed 2000 mg/l and was relatively benign to the pipes and other surface equipment (Grisgby et al. 1983).

4.2.2 Variations Computed for Three Water Wells

The reaction states of waters from three wells drilled into the alluvial aquifer and from a geothermal well were computed with data from Tables II to IV, for temperatures of 25°, 50°, 75°, 100°, 150°, 200°, and 250°C, in order to evaluate the effect of temperature on mineral solubilities. The evolution of the reaction states was then plotted against temperature for a selection of mineral phases (Figs. 13a to 13d). All three wells in the alluvium, located in the intended hot dry rock area, are contaminated by geothermal fluid and therefore are not suitable for irrigation use. Nevertheless, they display various chemical characteristics, and their percentages of geothermal fluid range from 14 to 58%. The steam production well (Utah State 14-2) is also taken into consideration to observe the reaction-state evolution during cooling, even if its chemical characteristics are very different from those of inferred HDR circulation water. Indeed, the rocks of the hydrothermal reservoir are deeply altered, whereas the granite-gneiss from the HDR zone is mainly unaltered. Gas partial pressures and the total pressure involved in an HDR loop (∿20 MPa) were not taken into consideration for these calculations, but it should be kept in mind that their influence on mineral solubility may not be negligible (Capuano and Cole 1982; Michard and Roekens 1983).

The selection of 16 mineral phases is based on the minerals that constitute the granite-gneiss in the hot dry area as well as some alteration minerals whose presence was noticed in the hydrothermal area. At 25°C (5° to 9°C above emergence temperature) most of the 16 mineral phases are either saturated or oversaturated with respect to the three water wells in the alluvium (Figs. 13a-13c), even though we observed no deposits or concretions in the wellhead pipes. On the other hand, most minerals are undersaturated at 250°C, and this relatively dilute water injected into a hot reservoir will become aggressive and dissolve minerals from the reservoir rock. Thus, a closed and pressurized loop system should be maintained to avoid as much mineral precipitation as possible in the surface pipes during the heat extraction phase.

For comparison, the same reaction states were computed for a steam producing well, Utah State 14-2 (Fig. 13d), for temperatures between 25° and 268°C, the latter being the measured deep temperature. Due to abundant dissolved silica in the geothermal fluid (656 mg/l), all the silica minerals are saturated at 268°C; nevertheless most other minerals are undersaturated. Surprisingly, if the reaction states for this geothermal fluid are computed at 25°C, namely the reinjection temperature in an HDR system, 8 of the 16 selected mineral phases are still undersaturated. The other half of the minerals, which are oversaturated, is composed of microcline, albite, muscovite, quartz, chalcedony, amorphous silica, kaolinite, and montmorillonite. Silica deposits are most likely to occur if a fluid with the composition of Utah State well 14-2 is cooled to 25°C. The Opal Mound Fault and its major silica concretions confirm this conclusion. However, silica dissolves faster and in greater amounts in a hydrothermally



Figs. 13a-d.

Variation of the reaction states for selected mineral phases at temperatures from 25° to 250°C. The range of saturation-state limits has been set arbitrarily at a Gibbs free energy difference of 0 ± 2 kcal/mol. Values are computed for four different wells: Jefferson, 26-9-18, and 26-10-26 in the shallow alluvium aquifer and Utah State well 14-2, a geothermal production well.

Fig. 13 (continued)

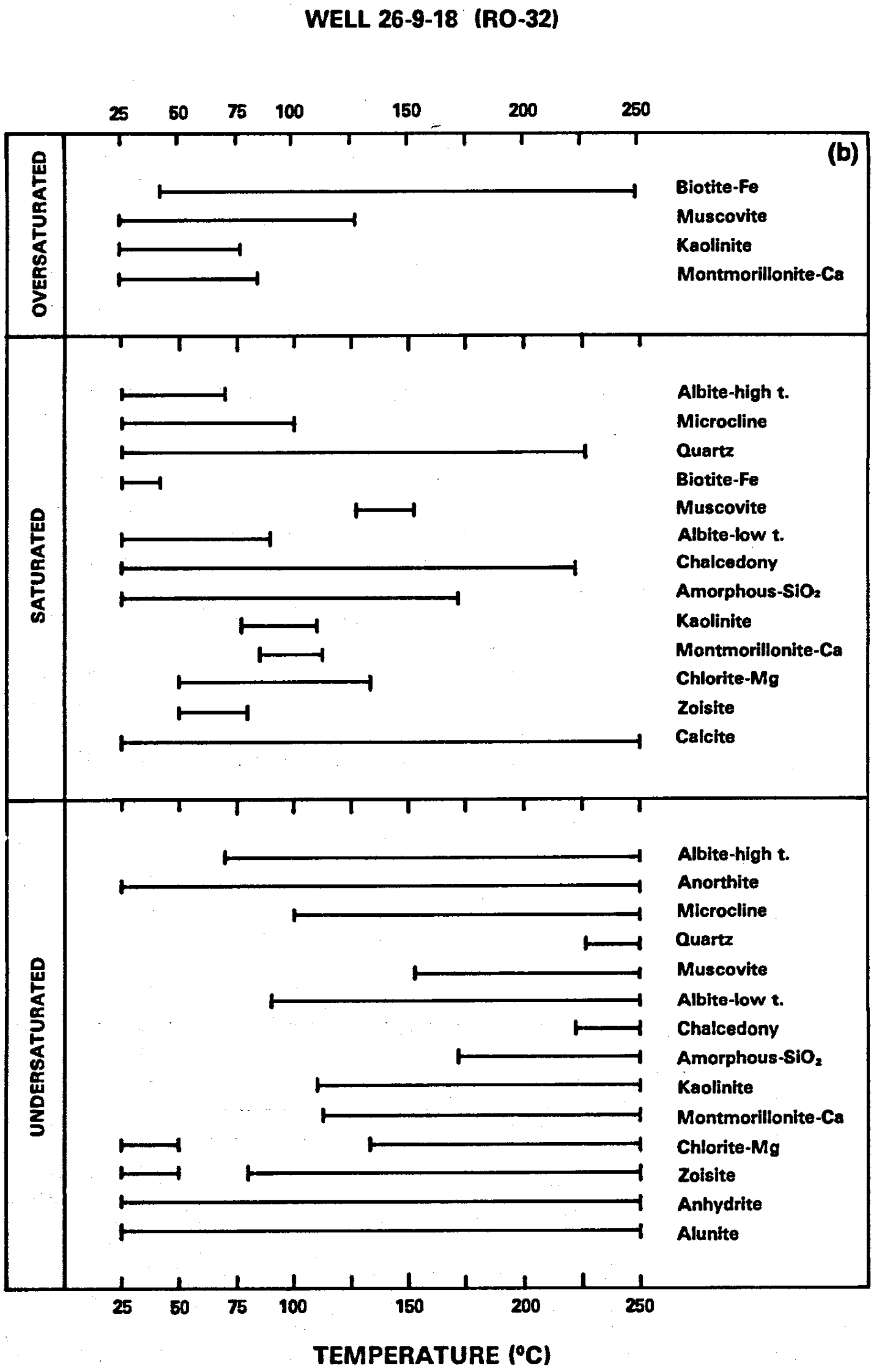


Fig. 13 (continued)

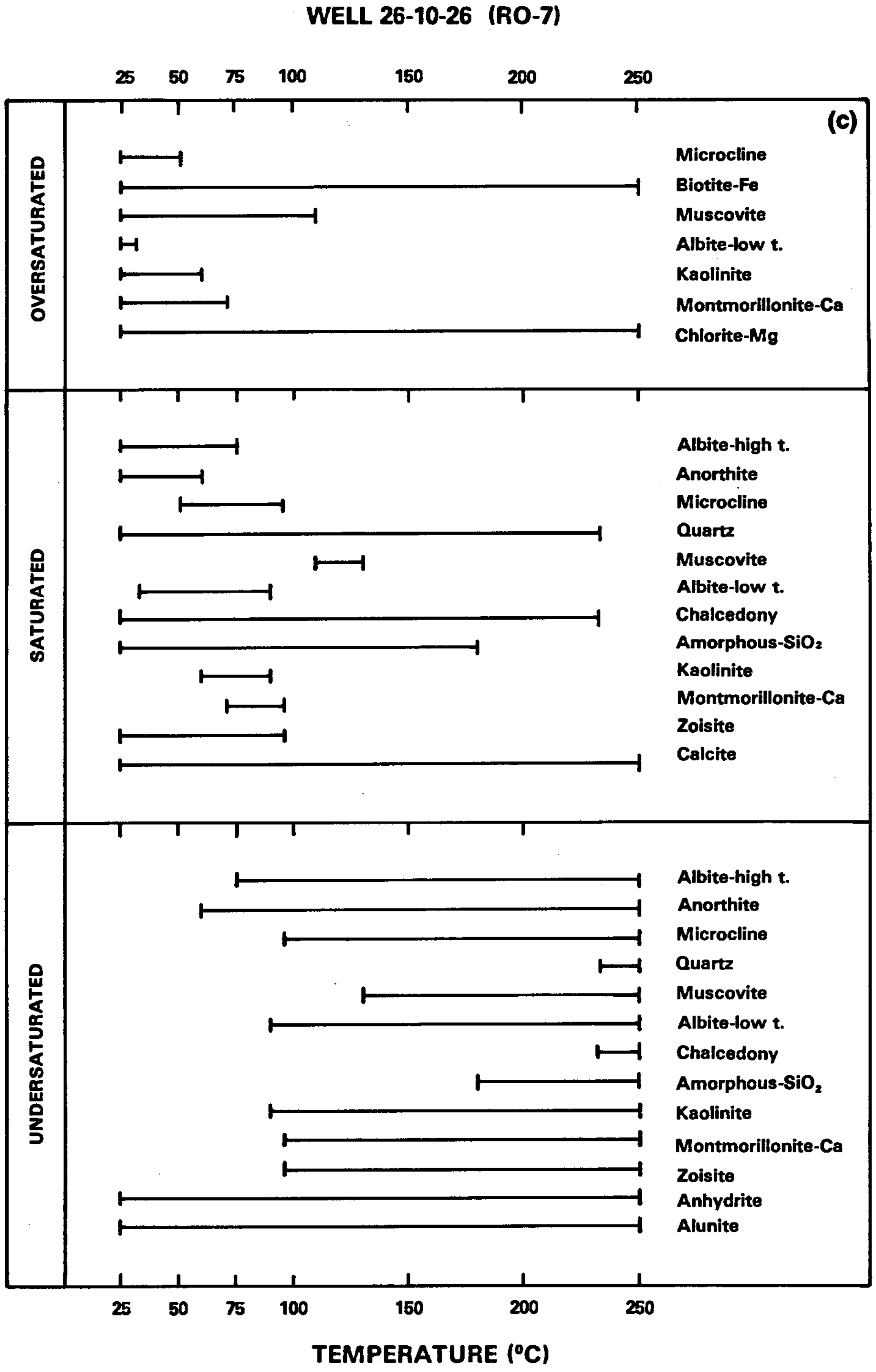
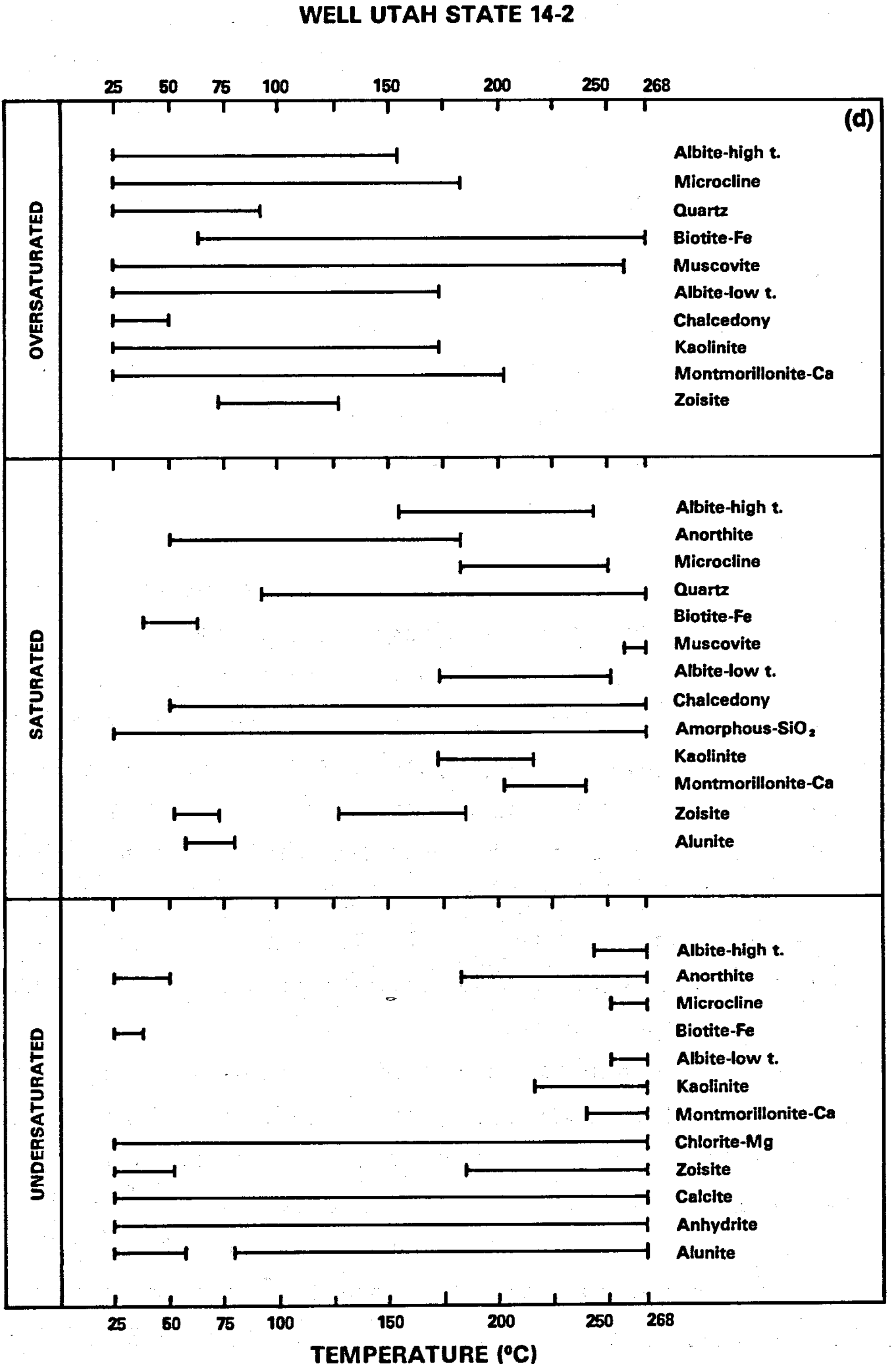


Fig. 13 (continued)



altered rock than it does in a relatively fresh granite-gneiss. Previous studies (Glenn et al. 1981; Capuano and Cole 1982) revealed the presence of alteration minerals in the hot dry wells. For instance, limonite-hematite, magnetite-ilmenite, anhydrite, calcite, sericite, epidote, chlorite, and quartz are found in minor or trace amounts in well 9-1 at a depth of 2 km.

4.3 Scaling and Corrosion Problems

The two following sections are designed to address the problems that may be encountered in the development of HDR resources. It will not be known how the fluid of the alluvial aquifer will behave in an HDR system until one is designed and tested. It is our opinion, however, that the fluids of the alluvial aquifer present no serious problems to development.

4.3.1 Scaling

Scaling in heat extraction systems can be very different according to the technique used for power production. In binary cycle systems, these problems will be limited to well casings, fluid transmission pipes, and heat exchangers. However, two possibilities still exist for geothermal binary cycles. In the first case (Fluid/Binary cycle), heat is transferred directly to the working cycle fluid from the geothermal water, whereas in the second case (Flash/Binary cycle), heat can be produced through steam allowed to flash from the geothermal water (Kestin et al. 1980). The Fluid/Binary cycle system needs more precautions toward scale formation on heat exchanger surfaces. On the contrary, scaling will be more severe in the production wellbore of a Flash/Binary cycle system.

The major scaling constituents observed in geothermal systems are $CaCO_3$, SiO_2, metal sulfides, hydrated iron oxides, oxychlorides, $BaSO_4$, and $SrSO_4$ (Awerbuch et al. 1983). Calcium carbonate is the main deposit in wells and pipes after flashing. In such a case, geothermal fluid is depressurized and CO_2 is released, which causes a rise of the pH. This causes a shift of the carbonate species toward CO_3^{2-} and precipitation of $CaCO_3$ in the form of calcite or aragonite (Ellis and Mahon 1977). The rise of pH also causes sulfides to precipitate. Once the temperature of the geothermal fluid is lowered it becomes supersaturated with respect to amorphous silica. Silica scaling is affected by degree of supersaturation, kinetics, pH, and turbulence.

Many studies and experiments have been done to try to prevent or at least to decrease scaling, which is a limiting factor to geothermal development. In order to control carbonate scale, recycling the natural CO_2 produced by the geothermal fluid and reinjecting it into the well, below the bubble point (Kuwada 1982), have been proposed. This would have the double action of keeping carbonate equilibrium in the soluble form of bicarbonate and therefore strongly decreasing carbonate scale, as well as increasing wellhead pressure and fluid production. In an HDR system, the fluid is kept under pressure, which keeps CO_2 in solution and prevents carbonates from forming.

Silica scales appear when the geothermal fluid is cooled, mostly in pipes and drains of the discharging systems. Experiments conducted at Wairaki, New Zealand, resulted in the following conclusions (Rothbaum et al. 1979): (1) aeration of geothermal waters that are supersaturated with silica will create siliceous deposits; (2) ponding or aging of geothermal waters will accelerate the polymerization of silica, with little effect on the scaling rate; (3) acidification of geothermal waters to a pH of 4-5 will strongly reduce the rate of scale formation and produce a scale that is not adherent. This last method, however, is limited by the fact that low pH may cause corrosion problems in the reinjection equipment. Silica scaling in the prototype HDR system at Fenton Hill has not been a problem at maximum temperatures of about 200°C. However, silica dissolution increases dramatically at higher temperatures. When cooled from production temperatures of $\geq$ 250°C to reinjection temperatures of 25° to 50°C, silica supersaturation from HDR circulation fluids may be a problem in HDR applications.

4.3.2 Corrosion

Corrosion control in geothermal plants depends largely on the type of alloys used for the piping and related equipment and on whether or not air is present (Conover 1983). Again the problems encountered will be very different according to the type of binary cycle system used to extract heat from the geothermal fluid. Factors that may increase corrosion on materials are mainly hydrogen sulfide concentration, low pH, oxygen intrusion, chloride content, and high fluid mineralization. Erosion may be caused by fluid velocity and solid particles. In some cases metallic coatings on pipe walls from iron precipitation serve as a protective film against corrosion. Finally, corrosion and its effects occur from various causes that cannot be listed here but are described by Kestin et al. (1980). By keeping the circulation fluid under pressure and by keeping it relatively reduced, it is not expected that the make-up water from the unconfined aquifer would be corrosive in HDR applications at Roosevelt Hot Springs.

5. HYDROGEOLOGY OF THE AQUIFER IN THE HDR ZONE

The hydrogeological characteristics of the alluvial aquifer in Milford Valley are relatively well known between Milford and Minersville because of the intensive pumping done during irrigation seasons. Mower and Cordova (1976) calculated the transmissivity of the aquifer in this area and found values ranging from 900 to 3700 m^2/day between the two towns. The maximum values are south of Milford, in the central part of the valley where the aquifer consists mostly of thick deposits of well-sorted coarse sand and gravel. The storage coefficient (specific yield) ranges from about 0.2 in some unconfined parts near the edge of the valley, to 0.001 in the confined parts near the center of the valley. North of Milford and especially near Roosevelt Hot Springs, no previous pumping tests have been conducted, and only the thickness and the lithology of the alluvium have been roughly evaluated because of the small number of water wells and the few measurements that have been made on these wells.

5.1. Pumping test

To obtain a better understanding of the chemical and physical characteristics of the naturally contaminated shallow aquifer, a pumping test was performed in one of the existing water wells of the hot dry rock area. The stock well 26-9-18 was chosen because it was the only well available of suitable diameter (20.3 cm/8 in.) in a highly contaminated zone of the aquifer. Unfortunately, no lithologic log of this well exists, and one has to rely on the log of well WOW-3, which is about 4 km south of the pumped well (Fig. 14). From this lithologic log and the data of Mower and Cordova (1974), it was assumed that in this area the alluvial aquifer is unconfined.

The pumping setup consisted of a 20-hp submersible pump having a capacity of 2.3 m^3/min, a flow meter ranging from 0 to 7.6 m^3/min, and a sampling faucet at the wellhead (Fig. 14). The pumped water was supposed to

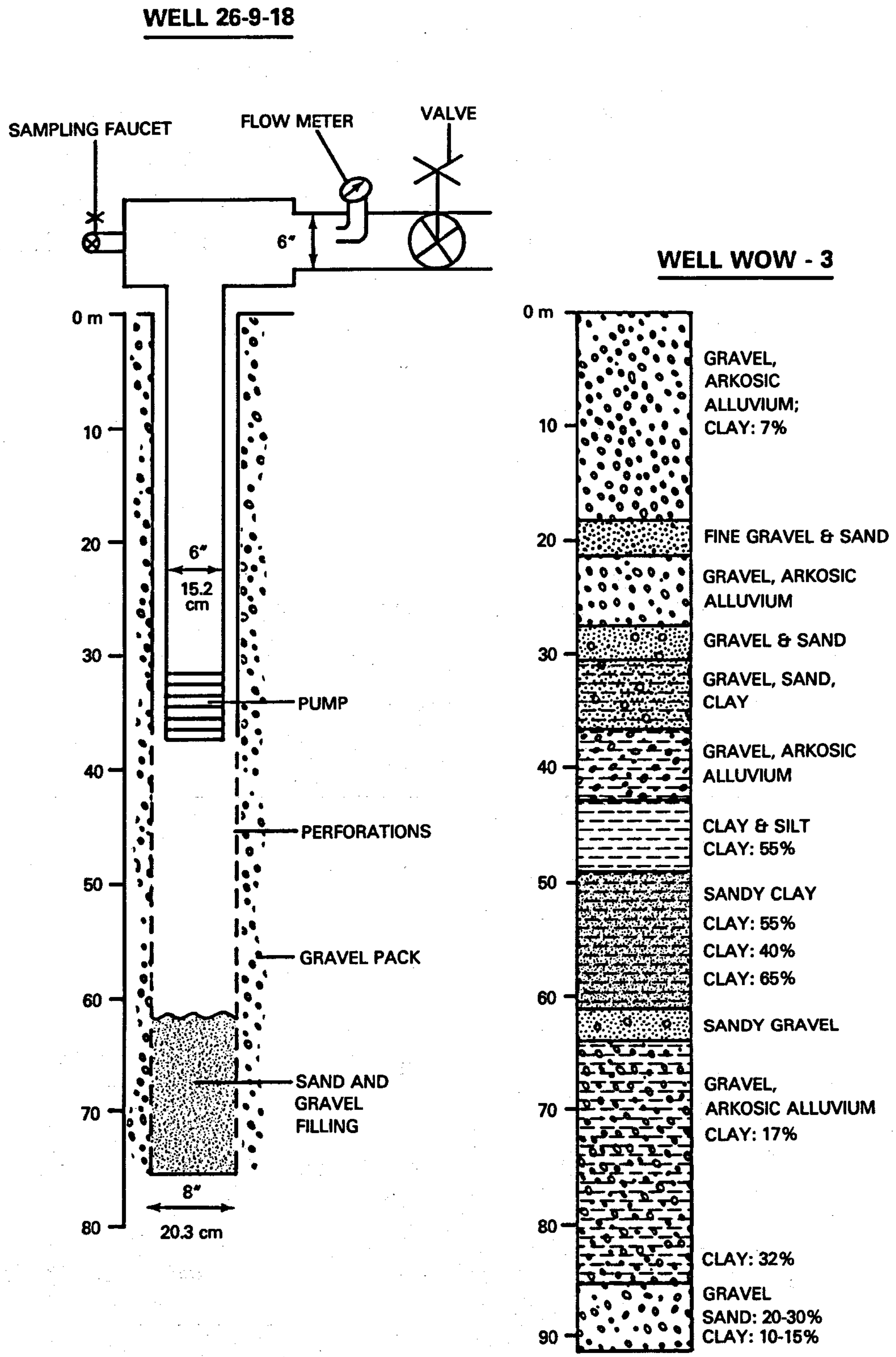


Fig. 14.
Schematic of pumping test set-up for well 26-9-18 and lithologic log of well WOW-3. WOW-3 is a 91-m-deep observation well (1558 m elevation) located 4 km south of well 26-9-18 (1527 m elevation) and is the closest well for which a lithologic log is available (R. C. Lenzer, Phillips Petroleum Co., 1982).

be discharged down the topographic slope toward a dry wash after flowing through about 800 m of tubing, but only 460 m of 15.2-cm aluminum irrigation pipe was available. Another 370 m of soft plastic tubing was provided by the subcontractor but did not succeed in carrying the pumped water farther away.

For several technical reasons the maximum flow rate of the pump could not be reached, and five pumping steps were conducted between 0.208 and 0.871 m^3/min during three days. Moreover, the flow meter was neither precise nor dependable; thus, the discharge had to be estimated by recording the time necessary to fill a 2.37-m^3 stock tank. Table IX and Fig. 15 summarize the conditions and the data recorded during this pumping test, which was carried out by a local irrigation company in Milford, Utah, and supervised by a Los Alamos crew.

TABLE IX
SUMMARY OF DATA FOR THE PUMPING TEST IN WELL 26-9-18[a]

Pumping step no.	Starting time (date) (h)	Total time of pumping (min)	Total time of recovery observation (min)	Average flow rate (m^3/min)	Initial water level[b] (m)	Final drawdown[c] (m)	Temperature[c] (°C)	Conductivity[c] (μS/cm)
1	4/20/83 7:45	120	220	0.208	21.93	0.67	19.6	6350
2	4/20/83 13:30	240	852	0.795	21.43	7.34	18.6	6310
3	4/21/83 7:45	45	58.5	0.871	21.63	6.08	18.5	6230
4	4/21/83 9:30	390	19	0.833	22.01	6.91	18.8	6210
5	4/21/83 17:30	930	423	0.814	23.23	5.99	18.4	6270

[a] Well characteristics: total depth = 61.3 m; casing diameter = 20.3 cm (8 in.); pump depth = 36.6 m; pump diameter = 15.2 cm (6 in.). A total of 1340 m^3 of water were pumped in 29 hours.

[b] Last water levels measured at the end of previous recovery observation period.

[c] Last measurements before the pump shutdown.

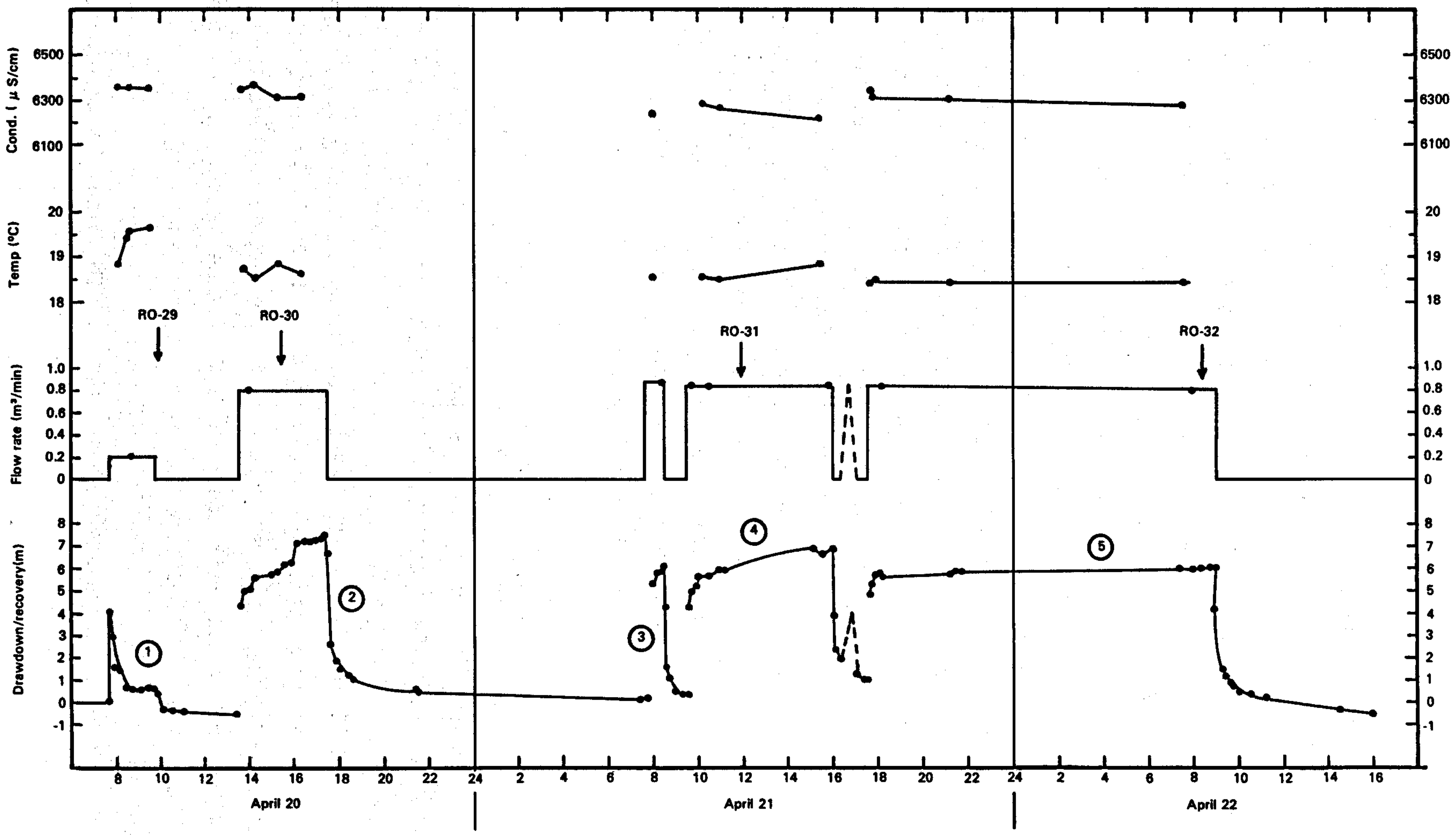


Fig. 15.
Flow rates and physical measurements made during the pumping test. The numbers RO-29 to RO-32 correspond to the collected water samples. The numbers 1 to 5 represent the five steps of the pumping test. The dashed lines for flow rate and water level during step 4 describe unrecorded short attempts to increase the flow rate.

5.1.1 Well Characteristics

At the beginning of each step, the water pumped out of the well was loaded with coarse sand and clay but became clear after a while. The specific capacity of the well, which is the flow rate (m^3/min) obtained by unit of drawdown (m) (Freeze and Cherry 1979), varied during the five steps from 0.31 to 0.11 m^3/min·m. For each pumping step the final drawdown value was used to calculate the specific capacity. The value of 0.31 m^3/min·m corresponds to the first step and may not be very representative of the well characteristics. However, the other four values are in close agreement, from 0.14 to 0.11 m^3/min·m, but are neither correlated to the pumping time nor correlated to the flow rate. The last step value of 0.14 to m^3/min·m means, for instance, that the pump set at 36.6 m would be dewatered at a maximum theoretical flow rate of 5 m^3/min. This value is double what the pump could possibly produce and is based on the linear relation between flow rate and drawdown. However, at a certain rate, this relation becomes nonlinear, and the water level decreases much more per unit of flow rate. At this point, the discharge value is called the critical yield (Castany 1963). Unfortunately, this value could not be determined during the pumping test, but it is certainly much below 5 m^3/min, probably around 2-3 m^3/min. A well of the size used in this pumping experiment is considerably smaller than one we would recommend for a future HDR system.

5.1.2 Aquifer Characteristics

By means of the drawdown/recovery data recorded during the five pumping steps for measured flow rates, it is possible to evaluate some physical parameters of the alluvial aquifer, which are valid only for the area around the pumped well and for the range of depths in which the well is drilled. No piezometer or well was available nearby to measure water levels within the depression cone during pumping. Only the well 26-10-13 located 1940 m southwest of the pumped well was used as an observation well, although it is too far away to be in the depression cone influence, according to the small flow rates and short periods of the five pumping steps (Freeze and Cherry 1979; Walton 1970).

Several methods of calculation were used to determine the transmissivity, which is the product of the average permeability (or hydraulic conductivity) and the thickness of the aquifer. Consequently, transmissivity is the rate of flow under a hydraulic gradient equal to unity through a cross section of unit width over the whole thickness of the aquifer. It is designated by the symbol T (or kD) and is expressed in m^2/day. The methods giving the most accurate results require the presence of an observation piezometer and require that the aquifer is fully penetrated by the pumped well. At Roosevelt Hot Springs, no nearby observation hole was available, and the pumped well only partially penetrates the unconfined aquifer of unknown thickness. According to previous studies (Mower and Cordova 1974), it appears that this thick aquifer is multilayered. Because of the location and shallow depth of the pumped well it is assumed that the aquifer is unconfined, isotropic, and homogeneous: these assumptions are required in order to use the calculation methods for the transmissivity (Kruseman and De Ridder 1979; Walton 1970).

In Table X six methods using different data and pumping steps display the following results. Steady-state flow was reached only during the fifth, last, and longest pumping step. The first two methods give an average transmissivity of 320 m^2/day, which is certainly too high. Indeed, these calculations incorporate well 26-10-13 as an observation well, implying that the cone of depression due to pumping has a radius of 1940 m. As mentioned above, this distance is definitely out of the range of the depression cone created by this pumping. The assumed thickness of the aquifer in the Thiem-Dupuit method does not significantly change the calculated transmissivity, if varying from 61 m (well depth) to 100 m (maximum possible local aquifer thickness according to Mower and Cordova 1974). The Thiem-Dupuit-Hantush method implies that the pumped well is only partially penetrating the aquifer, and its depth (61 m) is therefore used. The third calculation for steady-state flow, with Logan's method, gives a transmissivity of 246 m^2/day, assuming an approximate value of 3.33 for the log ratio of the radius of the depression cone over the radius of the pumped well (0.1016 m). This rough approximation means that the depression cone would be in this case 220 m, which seems reasonable according to the literature, but may yield erroneous results of up to 50% (Kruseman and De Ridder 1979).

Unsteady-state flow conditions are present in pumping steps 1 to 4; thus three calculation methods based on these conditions were also used. The first one (Theis) is based on the recovery of the water level after the pump

TABLE X
EVALUATION OF THE TRANSMISSIVITY IN THE AQUIFER FROM THE PUMPING TEST IN WELL 26-9-18

Method[a]	Transmissivity (m^2/day)	Data used in the calculation[b]
Steady-state flow		
Thiem-Dupuit[c]	317	Step 5; drawdown in pumped and observation wells; radius of pumped well and distance to observation well; flow rate; assumed thickness of aquifer = 100 m.
Thiem-Dupuit-Hantush[c]	323	Same as above, but penetration depth of pumped well (61 m) instead of aquifer thickness.
Logan[c]	246	Step 5; maximum drawdown in pumped well; assumed thickness of aquifer = 100 m.
Unsteady-state flow		
Theis recovery[c]	127	Average transmissivity for steps 2 and 3; recovery in pumped well; flow rate.
Hurr[c]	133	Average transmissivity for steps 2 and 4; drawdown and radius of pumped well; flow rate; assumed specific yield S = 0.15.
Walton specific capacity[d]	137	Average transmissivity for steps 2 and 4; specific capacity; radius of pumped well; flow rate; pumping time; assumed specific yield S = 0.15.

[a] All the methods used here are for an unconfined aquifer.

[b] Only pumping step 5 is considered to reach steady-state flow; among the steps 1 to 4 (unsteady-state flow), only the best-fit lines of water level versus time were considered: steps 2 and 3 for recovery, steps 2 and 4 for drawdown (Fig. 16).

[c] Kruseman and De Ridder (1979).

[d] Walton (1970).

was shut down (Fig. 16a). The second method (Hurr) is based on the drawdown during pumping (Fig. 16b) and on an assumed specific yield. For unconfined sandy aquifers, the specific yield ranges from 0.1 to 0.2; therefore, a value of 0.15 was chosen, but it does not significantly change the transmissivity calculation. Finally, the third method (Walton) is based on the specific capacity of the pumped well with the same assumption on the specific yield. Although these three methods use different parameters, consistent transmissivities are obtained, with an average of about 130 m^2/day. This value is rather low compared with the transmissivities calculated for the middle of Milford Valley (900-3700 m^2/day). Local lithologic variations of the aquifer can strongly modify its properties, especially the clay content

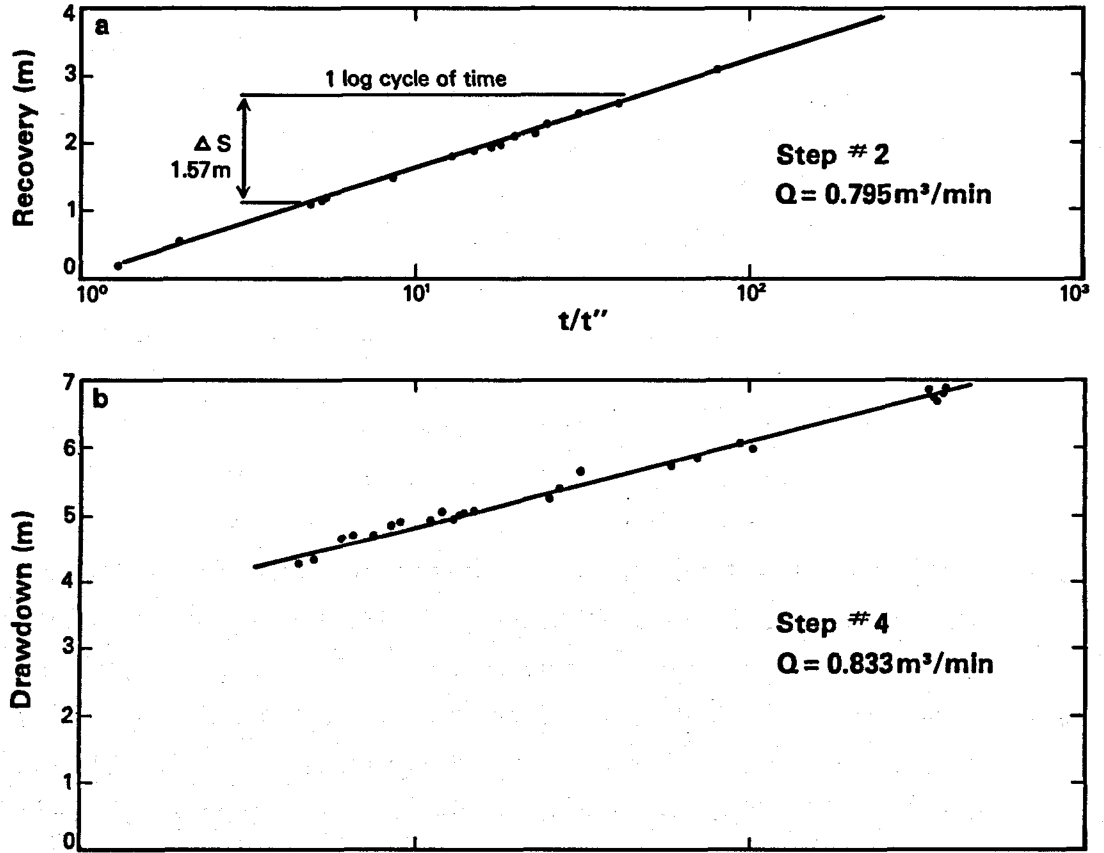


Fig. 16.
Examples of analysis of pumping test in well 26-9-18. (a) recovery data vs t/t" for step 2; t = time in minutes since pumping started; t" = time in minutes since pumping stopped. (b) drawdown data vs time in minutes since pumping started for step 4.

of the sand. For example, 5.7 km northeast of well 26-9-18, two irrigation wells (Anderson wells) yield 3.8 m^3/min each.

The thickness of the aquifer, or of the portion of aquifer which is influenced by pumping in well 26-9-18, is unknown but may be considered in the range of 10 to 100 m. Therefore the permeability estimated by means of the transmissivity value of 130 m^2/day ranges from 13 to 1.3 m/day. These values correspond to the permeability encountered in silty sand aquifers (Freeze and Cherry 1979) and to the sand-clay-gravel aquifer south of Milford (Mower and Cordova 1974).

5.2 Evolution of the Alluvial Aquifer

5.2.1 Water Level Decline

As mentioned previously, a dramatic drawdown (6 to $>$9 m) due to irrigation in the alluvial aquifer between Milford and Minersville was recorded during the period 1950-1972 (Mower and Cordova 1974). North of Milford the decline of water level was estimated to be only 0 to 3 m, because the water supply and irrigation wells were scarce.

Three shallow wells north of the HDR zone have been observed since 1941 (Fig. 17). The water level of Kaufmann well does not seem to have decreased throughout the 42-year observation period, even though there are long-term variations. On the other hand the Anderson well water level was extremely stable until the early 1970s. Since the mid-seventies when two irrigation wells were drilled, the water level has been measured in a piezometer about 150 m downslope (west) from the first well. Significant drawdowns have been recorded with maxima during the irrigation season. These wells have a rather high yield of 3.8 m^3/min each and a good recovery at the end of the irrigation season. But remember that a large amount of pumped water (30-40%) goes back into the aquifer by percolation because water is sprinkled on the ground around the well. Finally, the well 26-10-13 (the observation well during the pumping test) displays a very slight but consistent decline of its water level: about 0.2 m in 42 years, but mostly since the early 1970s.

A large water supply well for an HDR power plant would certainly cause the water level to decline significantly due to continuous pumping. Considering the parameters used to calculate the transmissivity with Walton's method (1970), and given a transmissivity of roughly 200 m^2/day, a specific

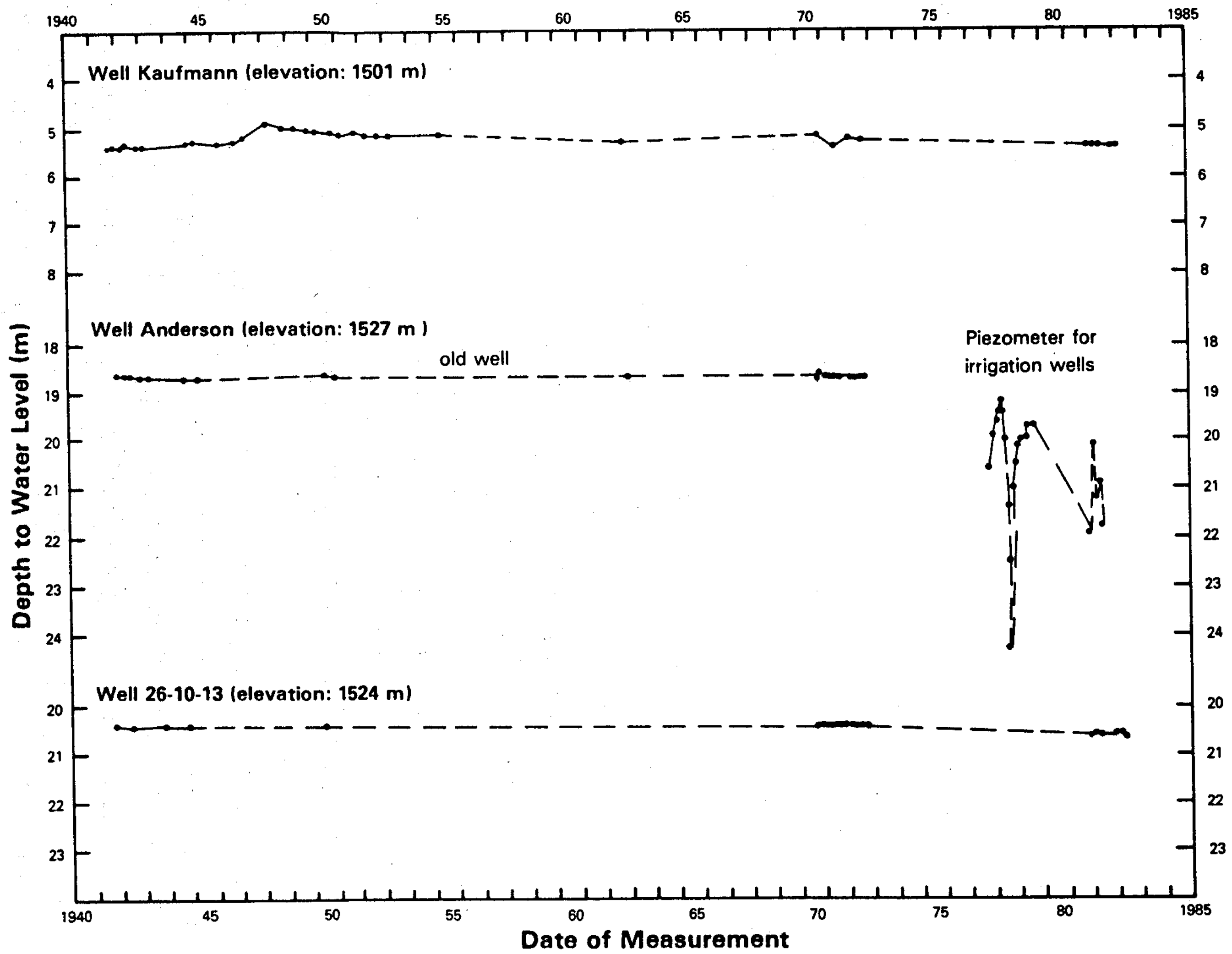


Fig. 17.

Evolution of water levels for three wells in Milford Valley (see Fig. 3 for location). If several values were available at the same period, averages were plotted on the graph. The data from 1941 to 1972 for well Anderson were measured in an old stock well, whereas the water levels for the period 1977 to 1982 were measured in a piezometer about 150 m downslope from the first of two new irrigation wells. Data sources: 1941-1972 (Mower and Cordova 1974); May 1982 and April 1983 (the present study); other data (Woodward-Clyde Consultants 1979 and Phillips Petroleum Co., pers. commun. 1983).

yield of 0.15, a well radius of 0.2 m, and a discharge of 2 m^3/min, one obtains a drawdown of 20 m for 1 year of pumping and 22 m for 10 years. This drawdown is relatively important, but the value of transmissivity used in the calculation was small. If transmissivity is doubled to 400 m^2/day, the drawdown is decreased to 10 and 12 m for 1 and 10 years of continuous pumping, respectively. Of course, these numbers have to be considered as order of magnitude values.

5.2.2 Geochemical Variations

Observations on the water quality have been made in the observation piezometer of Anderson well, the only irrigation system close to the HDR zone. These data are important for the estimation of geochemical evolution of the contaminated groundwater during heavy pumping, even though the irrigation season extends only from May to September. For the area around Anderson well, the contamination by geothermal fluids was estimated to be about 13%.

A two-year monitoring of water quality (1977-1979) shows the following trends (Fig. 18). During pumping for irrigation, groundwater temperature rises about 4°C, whereas the total mineralization, here the conductivity, declines and is accompanied by chloride. Boron does not change in a significant way, but the analytical precision is unknown for its relatively low concentration. Note that the temperature rise and mineralization decline are not exactly in phase with the water level; indeed, their extreme values are delayed by 1 or 2 months compared with the lowest water level. Prolonged pumping may draw more fresh groundwater from the upper part of the alluvial aquifer.

During the pumping test carried out in well 26-19-18, very little change in the water quality was observed, but the same conclusions can be drawn from the Anderson well (Fig. 15). Continuous pumping at shallow depth decreases the percentage of contamination by geothermal fluids. In this case, however, the temperature dropped a little, instead of going up as it did in the Anderson well. This can be explained by the season in which pumping takes place. For the irrigation well, pumping is done essentially during summer time, when percolating runoff waters are relatively warm. On the other hand our pumping test took place in April, right after snowmelt, and the recently infiltrated shallow groundwater was relatively cold.

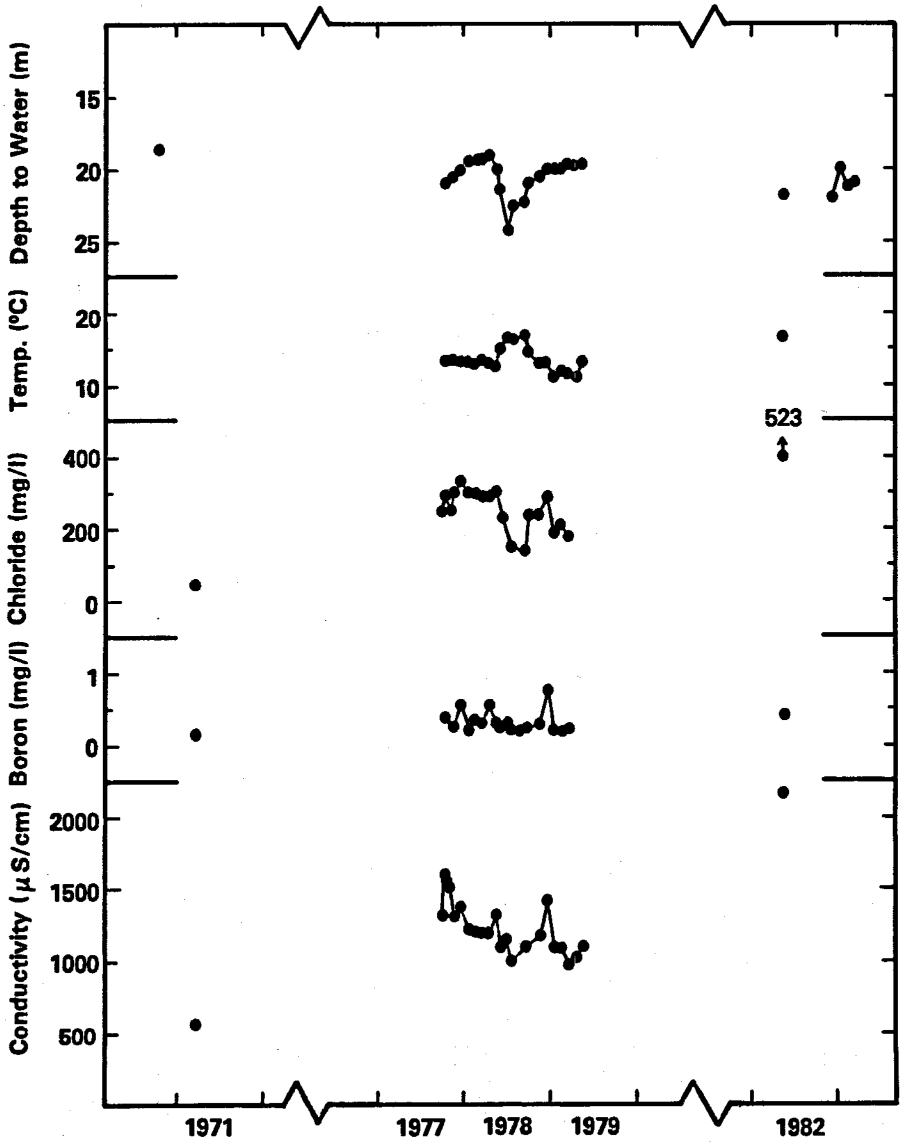


Fig. 18.
Evolution of the water quality around well Anderson. Data from 1971 were measured in an old stock well, whereas measurements for 1977 to 1982 are from an observation piezometer about 150 m downslope from the first of the two new irrigation wells. Data sources: 1971 (Mower and Cordova 1974); 1977-1979 (Woodward-Clyde Consultants 1979 and Phillips Petroleum Co., pers. commun. 1983); 1982 (the present study).

From these data one cannot draw firm conclusions about very long term effects of continuous pumping, but it seems probable that, after a period of slow mineralization decline, which could last several months, steady-state conditions would be reached in terms of mixing between geothermal fluids and dilute groundwaters.

6. RECOMMENDATIONS AND CONCLUSIONS

From geochemical and hydrogeological points of view, the unconfined aquifer west of the Roosevelt Hot Springs geothermal field, in the potential hot dry rock area, seems to meet the requirements for the make-up water in an HDR energy extraction system. These requirements are as follows:

1. the water resource should be located in the area of the planned HDR system;
2. the water resource should be at a reasonable depth;
3. the aquifer should be capable of providing water under continuous pumping, reaching steady-state conditions;
4. the required amount of water should be pumped by a maximum of 2-3 wells;
5. the depletion of the water table should be moderate and regionally limited; and
6. the water may be of industrial quality (neither potable nor suitable for agricultural uses), but of limited mineralization (≤ 10 g/l) and with relatively small amounts of silica and bicarbonate, in order to avoid major scaling problems.

A future 50-MW(e) HDR power plant would have a total circulation flow rate of about 20 m^3/min (J. Rowley, Los Alamos National Laboratory 1982, pers. commun.). If one considers an average water loss of 10% in the system as observed in the HDR Phase I system at Fenton Hill, a continuous supply of 2 m^3/min would be needed. This amount can be easily pumped from one well properly dimensioned.

According to the existing data and analysis of the local aquifer, a 50- to 70-m-deep well with a casing diameter of 30.5 cm (12 in.) would be able to draw 2 m^3/min. However, a second well ready to pump the same amount will be necessary in case the first one fails. Both wells could be pumped alternately or together at half rate. The size of gravel pack should be carefully chosen because of the large amount of clay present in the aquifer. Consequently a screen would probably be more appropriate than a perforated casing. Finally the distance from these wells to the next irrigation well system should be at least 1-2 km.

The zone of contaminated shallow water improper for irrigation use or drinking was found to extend over an area of 220 km^2. Within this area, the percentage of geothermal fluids generally decreases westward and varies from more than 60 to 15%, which implies highly variable chemical compositions, with total mineralization ranging from more than 5000 mg/l to 1300 mg/l. However, for a given location, the water quality will probably reach stability after a few months of pumping. From our analysis of thermodynamic parameters and comparison with the Fenton Hill system, we do not believe that these waters will cause major scaling and corrosion problems in a future HDR system. Various techniques used to solve these difficulties in conventional geothermal systems may be adapted to HDR systems.

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REFERENCES

Awerbuch, L., V. C. Van der Mast, and D. P. McGrath, 1983, "Geothermal scale control and mineral recovery using a flash-crystallizer-separator process," Geotherm. Res. Counc. Bull. 12(9), 7-13.

Barnes, I., and F. E. Clarke, 1969, "Chemical properties of groundwater and their encrustation effects on wells," U.S. Geol. Surv. Prof. Paper 498-D, 58 pp.

Barnes, R. A., 1975, "The determination of specific forms of aluminum in natural water," Chem. Geol. 15, 177-191.

Bath, A. H., and K. H. Williamson, 1983, "Isotopic and chemical evidence for water sources and mixing in the Cerro Pando geothermal area, Republic of Panama," Geothermics 12(2/3), 177-184.

Capuano, R. M., and D. R. Cole, 1982, "Fluid-mineral equilibria in a hydrothermal system, Roosevelt Hot Springs, Utah," Geochim. Cosmochim. Acta 46, 1353-1364.

Castany, G., 1963, Traité pratique des eaux souterraines, Dunod, Paris, 657 pp.

Christensen, O. D., R. A. Capuano, and J. N. Moore, 1983, "Trace-element distribution in an active hydrothermal system, Roosevelt Hot Springs thermal area," Utah, J. Volcanol. Geotherm. Res. 16, 99-129.

Conover, M. F., 1983, "Designing geothermal power plants to avoid reinventing the corrosion wheel," Geotherm. Res. Counc. Bull. 12(3), 3-11.

Craig, H., 1961, "Isotopic variations in meteoric waters," Science 133:3465, 1702-1703.

Dash, Z. V., H. D. Murphy, P. L. Aamodt, R. G. Aguilar, D. W. Brown, D. A. Counce, H. N. Fisher, C. O. Grigsby, H. Keppler, A. W. Laughlin, R. C. Potter, J. W. Tester, P. E. Trujillo, Jr., and G. Zyvoloski, 1983, "Hot dry rock geothermal reservoir testing: 1978 to 1980," in G. Heiken and F. Goff (eds.), "Geothermal energy from Hot Dry Rock," J. Volcanol. Geotherm. Res. 15, 59-99.

East, J., 1981, "Hot dry rock geothermal potential of Roosevelt Hot Springs area: review of data and recommendations," Los Alamos National Laboratory report LA-8751-HDR, 45 pp.

Ellis, A. J., and W. A. J. Mahon, 1977, Chemistry and geothermal systems, Academic Press, New York, 392 pp.

Fouillac, C., and G. Michard, 1981, "Sodium/lithium ratio in water applied to geothermometry of geothermal reservoirs," Geothermics 10:1, 55-70.

Fournier, R. O., and J. J. Rowe, 1966, "Estimation of underground temperatures from the silica content of water from hot springs and wet-steam wells," Am. J. Sci. 264, 685-697.

Freeze, R. A., and J. A. Cherry, 1979, Groundwater, Prentice-Hall Inc., Englewood Cliffs, New Jersey, 604 pp.

Fritz, P., and J. C. Fontes (eds.) 1980, Handbook of environmental isotope geochemistry, vol. 1A: The terrestrial environment, Elsevier, Amsterdam, 545 pp.

Geothermal Resources Council, 1983, "State Scene: Utah," GRC Bull. 12(2).

Glenn, W. E., and J. B. Hulen, 1979, "Interpretation of well log data from four drill holes at Roosevelt Hot Springs KGRA," Univ. Utah Res. Inst., Earth Sci. Lab. Report 28, 74 pp.

Glenn, W. E., J. B. Hulen, and D. L. Nielson, 1981, "Comprehensive study of LASL well C/T-2 Roosevelt Hot Springs KGRA, Utah, and applications to geothermal well logging," Los Alamos National Laboratory report LA-8686-MS, 175 pp.

Goff, F., and E. R. Decker, 1983, "Candidate sites for future hot dry rock development in the United States," in G. Heiken and F. Goff (eds.) "Geothermal energy from Hot Dry Rock," J. Volcanol. Geotherm. Res. 15, 187-221.

Grigsby, C. O., J. W. Tester, P. E. Trujillo, D. A. Counce, J. Abott, C. E. Holley, and L. A. Blatz, 1983, "Rock-water interactions in hot dry rock geothermal systems: field investigations of in situ geochemical behavior," in G. Heiken and F. Goff (eds.), "Geothermal energy from Hot Dry Rock," J. Volcanol. Geotherm. Res. 15, 101-136.

Kestin, J., R. DiPippo, H. E. Khalifa, and D. J. Ryley (eds.), 1980, Sourcebook on the production of electricity from geothermal energy, U.S. Dept. Energy, DOE/RA/4051-1, Washington, DC, 997 pp.

Kharaka, Y. K., and I. Barnes, 1973, SOLMNEQ: "Solution-Mineral Equilibrium Computations," U.S. Geol. Surv. Computer Contrib., U.S. Dept. Commerce, NTIS report PB-215899, 82 pp.

Kruseman, G. P., and N. A. De Ridder, 1979, "Analysis and evaluation of pumping test data," Inter. Inst. for Land Reclamation and Improvement, The Netherlands, Bull. 11, 200 pp.

Kuwada, J. T., 1982, "Field demonstration of the EFP system for carbonate scale control," Geotherm. Res. Counc. Bull. 11(9), 3-9.

Lenzer, R. C., G. W. Crosby, and C. W. Berge, 1977, "Recent developments at the Roosevelt Hot Springs KGRA," Trans. Am. Nucl. Soc. Topical Mtg., Golden, Colorado, April 12-14.

Michard, G., and E. Roekens, 1983, Modeling of the chemical composition of alkaline hot waters, Geothermics 12(2/3), 161-169.

Mower, R. W., and R. M. Cordova, 1974, "Water resources of the Milford area, Utah, with emphasis on ground water," State of Utah Department of Natural Resources, Salt Lake City, Tech. Publ. No. 43, 106 pp.

Mundorff, J. C., 1970, "Major thermal springs of Utah," Utah Geol. and Mineral Survey Water Resources Bull. 13, 60 pp.

Nielson, D. L., S. H. Evans, and B. S. Sibbett, 1986 "Magmatic, structural, and hydrothermal evolution of the Mineral Mountains intrusive complex, Utah," Bull. Geol. Soc. Amer., 97, 765-777.

Panichi, C., and R. Gonfiantini, 1978, "Environmental isotopes in geothermal studies," Geothermics 6, 143-161.

Parry, W. T., J. M. Ballantyne, N. L. Bryant, and R. E. Dedolph, 1980, "Geochemistry of hydrothermal alteration at Roosevelt Hot Springs thermal area, Utah," Geochim. Cosmochim. Acta 44, 95-102.

Rohrs, D. T., and J. R. Bowman, 1980, "A light stable isotope study of the Roosevelt Hot Springs thermal area, southwestern Utah," Dept. of Geology and Geophysics, Univ. of Utah, Salt Lake City, report DOE/ET/28392-46, 89 pp.

Ross, H. P., D. L. Nielson, and J. N. Moore, 1982, "Roosevelt Hot Springs geothermal system, Utah - case study," AAPG Bull. 66(7), 879-902.

Rothbaum, H. P., B. H. Anderton, R. F. Harrison, A. G. Rohde, and A. Slatter, 1979, "Effect of silica polymerization and pH on geothermal scaling," Geothermics 8, 1-20.

Shannon, S. S., R. Pettitt, J. Rowley, F. Goff, M. Mathews, and J. J. Jacobson, 1983a, "Acord 1-26 hot, dry well, Roosevelt Hot Springs hot dry rock prospect, Utah," Los Alamos National Laboratory report LA-9857-HDR, 15 pp.

Shannon, S. S., F. Goff, J. C. Rowley, R. A. Pettitt, and F. D. Vuataz, 1983b, "Roosevelt Hot Springs/hot dry rock prospect and evaluation of the Acord 1-26 well," Geotherm. Res. Counc. Trans. *7*, 541-544.

Smith, L., 1980, "A model study of the regional hydrogeologic regime, Roosevelt Hot Springs," Utah, Dept. of Geology and Geophysics, Univ. of Utah, Salt Lake City, report DOE/ET/28392-44, 30 pp.

mith, M. C., and G. M. Ponder (eds.), 1982, "Hot Dry Rock geothermal energy development program, Annual report fiscal year 1981," Los Alamos National Laboratory report LA-9287-HDR, 140 pp.

U.S. Department of Interior, 1968, "Report of Committee on Water Quality Criteria," F.W.P.C.A.

U.S. Environmental Protection Agency, 1973, "Water quality criteria 1972," EPA R3 73033, Govt. Printing Off., Washington, D.C.

U.S. Environmental Protection Agency, 1975, "Water programs: national interim primary drinking water regulations," Fed. Register 40, no. 248.

Utah Geological and Mineral Survey, 1980, "Geothermal resources of Utah," Map, scale 1:500 000, prepared by Nat. Geophys. and Solar-Terrest. Data Center and by Nat. Ocean. and Atmosph. Adm.

Vuataz, F. D., 1982, *Hydrogéologie, géochimie et géothermie des eaux thermales de Suisse et des régions alpines limitrophes, Matér. Géol. Suisse, Sér. Hydrol. 29*, Kümmerly & Frey, Berne, 174 pp.

Vuataz, F. D., J. Stix, F. Goff, and C. Pearson, 1984, "Low-temperature geothermal potential of the Ojo Caliente warm springs area, northern New Mexico," Los Alamos National Laboratory report LA-10105-OBES, 56 pp.

Walton, W. C., 1970, *Groundwater resource evaluation*, McGraw-Hill, New York, 664 pp.

Ward, S. H., W. T. Parry, W. P. Nash, W. R. Sill, K. L. Cook, R. B. Smith, D. S. Chapman, F. H. Brown, J. A. Whelan, and J. R. Bowman, 1978, "A summary of the geology, geochemistry and geophysics of the Roosevelt Hot Springs thermal area, Utah," Geophysics *43(7)*, 1515-1542.

WATSTORE, 1983, "Water Data Storage and Retrieval System," U.S. Geol. Surv., Albuquerque, New Mexico.

White, D. E., 1970, "Geochemistry applied to the discovery, evaluation and exploitation of geothermal energy resources," Geothermics, Special Issue *2:1*, 58-80.

Woodward-Clyde Consultants, 1979, "Environmental baseline study Roosevelt Hot Springs, Utah KGRA, Final report," prepared for Phillips Petroleum Co., unpublished.

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