DOE/ID/12079-44

DOE/ID/12079--44

DE82 019906

TOPICAL REPORT

Light-Stable-Isotope Studies of Spring and Thermal Waters from the Roosevelt Hot Springs and Cove Fort/Sulphurdale Thermal Areas and of Clay Minerals from the Roosevelt Hot Springs Thermal Area

by

John R. Bowman and David T. Rohrs

Department of Geology and Geophysics
University of Utah
Salt Lake City, Utah 84112

October 1981

Work performed under Department of Energy/Division of Geothermal Energy contract number DE-AC07-80ID12079

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ABSTRACT

The isotopic compositions of hydrogen and oxygen have been determined for spring waters and thermal fluids from the Roosevelt Hot Springs and Cove Fort-Sulphurdale thermal areas, for clay mineral separates from shallow alteration of the acid-sulfate type in the Roosevelt Hot Springs area, and for spring and well waters from the Goshen Valley area of central Utah. The water analyses in the Roosevelt Hot Springs thermal area confirm the origin of the thermal fluids from meteoric water in the Mineral Range. The water analyses in the Cove Fort-Sulfurdale thermal area restrict recharge areas for this system to the upper elevations of the Pavant and/or Tushar Ranges. The low ^{18}O shift observed in these thermal fluids (+0.7 permil) implies either high water/rock ratios or incomplete isotope exchange or both, and further suggests minimal interaction between the thermal fluid and marble country rock in the system. Hydrogen and oxygen-isotope data for clay mineral separates from shallow alteration zones in the Roosevelt Hot Springs thermal system suggest that the fluids responsible for the shallow acid-sulfate alteration were in part derived from condensed steam produced by boiling of the deep reservoir fluid. The isotope evidence supports the chemical model proposed by Parry et al. (1980) for origin of the acid-sulfate alteration at Roosevelt Hot Springs. The isotope analyses of spring and well waters from the Goshen Valley area indicate only a general correlation of isotope composition, salinity and chemical temperatures.

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ACKNOWLEDGMENTS

Dr. W. T. Parry of the Department of Geology and Geophysics, University of Utah provided geological and geochemical data and materials for the studies of acid-sulfate alteration at Roosevelt Hot Springs and for the studies of spring waters from central Utah. J. N. Moore and his colleagues at the Earth Science Laboratory, University of Utah Research Institute, provided data and materials relating to the Cove Fort-Sulphurdale Thermal Area. We are indebted to Walter Kropp and Ray Lambert for their technical assistance during mineral separation and isotopic analysis. We are also indebted to Mike Cleary and Ray Lambert for help in collection of the spring and thermal waters. This research was funded under DOE/DGE contract DE-AC07-80ID12079.

INTRODUCTION

Stable isotope studies of regional spring waters and available thermal fluids have been conducted in the vicinity of Roosevelt Hot Springs and Cove Fort/Sulphurdale thermal areas. The studies at Roosevelt Hot Springs are an extension of our earlier work there (Rohrs and Bowman, 1980) and have involved resampling of some of the regional spring waters and reanalysis of most spring and thermal fluid samples. The additional sampling and analytical program was undertaken to constrain further recharge areas for these two thermal areas and to document any year-to-year changes in the hydrogen and oxygen isotope compositions of spring waters. The results of this part of the study will be presented in Part I.

In addition, hydrogen and oxygen isotope analyses were made of clay mineral separates from shallow alteration at the Roosevelt Hot Springs thermal area. These analyses were made to see if low-temperature clays in the shallow parts of geothermal systems can be isotopically distinguished from normal supergene clays which result from weathering. The analyses were also made to see if these clays would be useful in elucidating processes of water-rock interaction in the uppermost zones of geothermal systems. These results will be discussed in Part II.

We have also conducted hydrogen and oxygen isotope studies of spring waters from Goshen and Pavant Valleys in central Utah. Cleary (1978) demonstrated that spring waters indicating chemical temperatures in excess of 100°C are widely distributed throughout these areas of Utah. Cleary also emphasized that additional work would be required to evaluate the effects of brine mixing, clay exchange and water mixing on these thermometry results.

The isotope analyses were performed in an attempt to evaluate these effects so that a more realistic evaluation of geothermal potential could be made in these areas. These analyses will be discussed in Part III.

Sample Selection and Preparation

Spring waters

Waters from surficial springs and a seep were collected for hydrogen and oxygen analyses from the vicinity of Roosevelt Hot Springs and Cove Fort/Sulphurdale thermal areas by the University of Utah during the fall of 1977 and spring of 1978. This sampling was intended to identify the probable recharge aquifer for the geothermal fluids on the basis of isotopic variations in spring waters from discrete geographical regions. Sampling was accomplished by standard water collection methods (Presser and Barnes, 1974; Nehring and Truesdell, 1977).

Geothermal wells

Reservoir fluids were sampled by the University of Utah during flow tests by Thermal Power Co. in May, 1978 at Roosevelt Hot Springs and during flow tests by Union Geothermal Division in March, 1978 at Cove Fort/Sulphurdale. In order to obtain representative samples of discharged fluids, a steam/fluid mini-separator of New Zealand design was used for these collections (Nehring and Truesdell, 1977).

Solid Samples

All clay mineral separates were obtained from core samples and rotary drill cuttings from University of Utah well #76-1. K-mica, kaolinite and montmorillonite clays for oxygen and hydrogen isotope analyses were taken from feldspar pseudomorphs and/or rock matrix and then centrifuged to obtain the fraction less than five micron size.

Analytical Procedures

Extraction techniques

Water oxygen extraction. The $^{18}O/^{16}O$ ratios of spring and thermal waters were determined by the CO_2 equilibration technique (Epstein and Mayeda, 1953). The fractionation factor between H_2O and CO_2 at 25°C was taken as 1.0412 (O'Neil et al., 1975).

Water hydrogen extraction. Hydrogen gas for D/H measurements was liberated from water samples by reduction over hot (800°C) uranium metal (Friedman, 1953).

Silicate hydrogen extraction. After the removal of adsorbed and interlayer water by degassing in a vacuum at 150°C overnight, hydrogen and water were liberated from K-mica clays by heating to approximately 900°C using a method similar to that of Suzuoki and Epstein (1976). Hydrogen was converted to water in a CuO furnace at 700°C prior to reduction over hot uranium metal at 800°C.

Silicate oxygen extraction. Oxygen from silicate minerals was extracted by reacting 5 to 10 mg of clay with BrF_5 at 550°C in nickel reaction vessels for 12 to 14 hours (Clayton and Mayeda, 1963). The evolved O_2 gas was then converted to CO_2 for mass spectrometric analysis by combustion with graphite (Taylor and Epstein, 1962).

Mass spectrometry

Isotopic measurements for CO_2 gas and H_2 gas were made with Micromass 602 D mass spectrometers, which are double collector, 90° sector magnetic deflection instruments of 6 cm radius. The isotopic data for hydrogen and oxygen are reported relative to SMOW (Craig, 1961a) and for carbon relative to the Chicago PDB standard (Craig, 1957). Analytical error for carbon and oxygen isotope ratios is between 0.1 and 0.2 permil, while that for hydrogen is between 1 and 2 permil.

Notation

All isotopic data are reported in the delta notation, where

$$\delta X_a = \frac{R_A - R_{std}}{R_{std}} \times 1000.$$

δX represents the δD, $\delta^{18}O$, or $\delta^{13}C$ of sample A, and R is the D/H, $^{18}O/^{16}O$, or $^{13}C/^{12}C$ ratio of the sample or standard. For coexisting phases A and B,

$$10^3 \ln \alpha_{A-B} \cong \delta A - \delta B = \Delta_{A-B} \quad ,$$

where α is the fractionation factor, defined as

$$\alpha_{A-B} = \frac{R_A}{R_B} = \frac{1000 + \delta X_A}{1000 + \delta X_B} .$$

PART I. Analysis of Spring and Thermal Waters from the Roosevelt Hot Springs and Cove Fort/Sulphurdale Thermal Areas.

In our initial report on the Roosevelt Hot Springs thermal area, we mentioned the importance in future work of the continued monitoring of the water and steam from production wells, a seep, and the springs from the potential recharge areas. No flow tests accessible to us have been conducted since our submission of the original report, hence no new steam or thermal water data are available. This is quite unfortunate, because long-term monitoring of producing wells by isotope techniques is likely, based on the results of our first study, to provide significant information on the hydrologic stability of the geothermal system. For example, should the water become enriched in ^{18}O, this could indicate that fluids from impermeable rocks are migrating into the wells. A depletion in ^{18}O might be evidence for smaller isotopic shifts, implying faster migration of the meteoric waters through the system. Changes in the fractionation between steam and water could indicate either fluctuations in steam separation temperatures or the arrival of cooler or hotter fluids into the wells. By comparing these possible changes in the water isotopic data to the quantities of water and steam produced by the wells, fluctuations in the hydrology of the system, such as a decrease in permeability, might be detected.

However, many of the springs in potential recharge areas for the Roosevelt Hot Springs and Cove Fort/Sulphurdale thermal areas that were sampled for our initial report (Rohrs and Bowman, 1980) were resampled during the fall of 1977 and the summer of 1978. Locations of the spring waters sampled are provided in Figure 1. The hydrogen and oxygen isotope data are

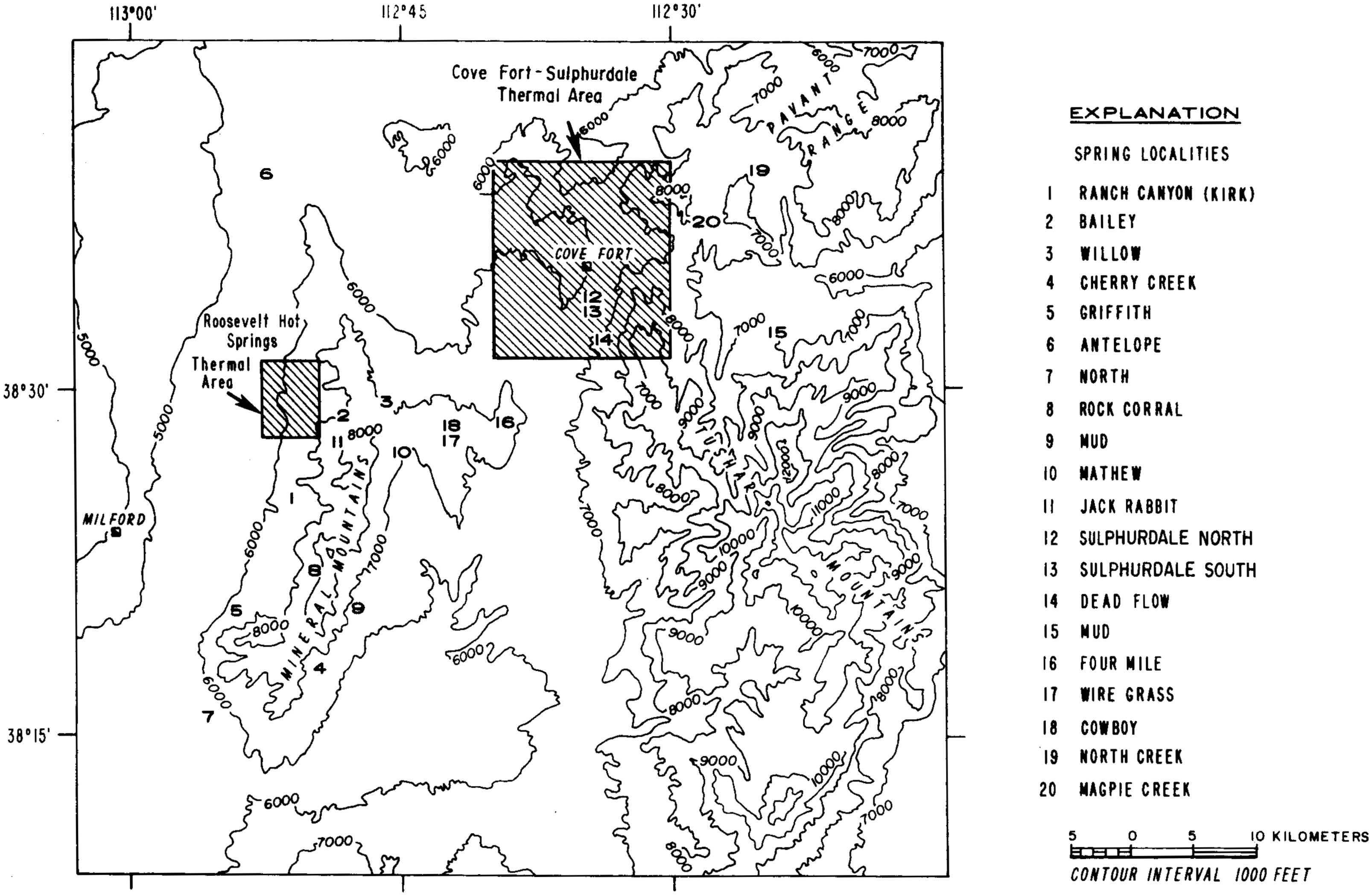

Figure 1. Generalized topographic map of the region including the Roosevelt Hot Springs and Cove Fort-Sulphurdale Thermal Areas. Spring localities are noted by number.

compiled in Table 1. Isotope results from our initial report, determined by A. A. Truesdell of the U.S. Geological Survey and at the University of Utah, are also shown for ease of comparison.

As with the initial suite of spring water samples, the meteoric waters from each geographical area consistently deviate from the meteoric water line, as shown in Figure 2. Fitting a line of linear regression to the new Mineral Mountains analyses provides a slope of 4.6, equivalent to that calculated for the original data set (Rohrs and Bowman, 1980). Both data sets are consistent with modification of these meteoric waters by kinetic effects accompanying evaporation (Craig, 1961b; Craig et al., 1963) or ablation of the snow pack. Although this phenomenon could be a common feature of meteoric waters in the arid Basin and Range environment of the southwestern United States, long-range studies are certainly required to substantiate this hypothesis.

A comparison of the isotopic composition of spring waters collected in 1976 with those of the recollected samples in 1977-1978 indicate little or no systematic shift in hydrogen or oxygen isotope composition. The data suggest that storm-to-storm fluctuations in δD and $\delta^{18}O$ values are well integrated during spring runoff and that, at least for two years, ground waters have been isotopically uniform. Although not conclusive, the results give some confidence that the δD and $\delta^{18}O$ values of mountain recharge areas are comparatively uniform from year to year and can be used to evaluate whether specific areas can contribute fluid to a geothermal system.

The relationship of the thermal fluids in each thermal area to the spring waters is also shown in Figure 2. The δD value of the thermal waters at Roosevelt Hot Springs, -116, is within the range exhibited by springs from the

Table 1. Isotopic analyses of spring and thermal waters from the Roosevelt Hot Springs and Cove Fort/Sulphurdale regions.

	Date Collected	δD	$\delta^{18}O$
Mineral Mtns. Springs			
Kirk	10/6/77	-115	-14.3
	*8/76	-110	-14.2
Bailey	10/6/77	-114	-15.0
	*8/76	-111	-14.6
Willow	10/6/77	-115	-14.5
	*8/76	-114	-14.8
Mathew	10/6/77	-115	-15.4
	+10/6/77	-117	-15.6
Jack Rabbit	10/6/77	-114	-16.0
	+10/6/77	-116	-15.8
Tushar Mtns. Springs			
Sulphurdale North	7/22/77	-116	-15.4
	*8/76	-118	-15.6
Sulphurdale South	7/22/77	-115	-15.5
	*8/76	-116	-15.6
Dead Cow	10/3/77	-116	-14.8
	*8/76	-114	-14.6
Mud	10/3/77	-123	-16.3
	*8/76	-126	-16.5
Springs between Tushar and Mineral Mtns.			
Four Mile	10/5/77	-113	-14.4
	*8/76	-114	-14.2
Wiregrass	10/5/77	-116	-15.0
	*8/76	-115	-15.2
Cowboy	10/5/77	-118	-15.1
	*8/76	-118	-15.2
Pavant Range Springs			
North Creek	10/3/77	-121	-15.0
	*8/76	-120	-15.0
Magpie	10/3/77	-119	-15.3
	*8/76	-120	-15.6
Thermal Waters			
Roosevelt KGRA, Utah State Well 14-2			
Water	5/13/78	-115	-13.4
	+5/13/78	-116	-13.6
Steam	5/13/78	-113	-15.1
	+5/13/78	-114	-15.4
Roosevelt Seep	5/11/78	-111	-12.4
	+5/11/78	-111	-12.7
Cove Fort/Sulfurdale Union Well 42-7			
Water		-121	-15.7
Steam		-135	-19.4
Sulfurdale Mine	7/22/77	-85	-13.3

*Analysis by A. H. Truesdell, USGS (Rohrs and Bowman, 1980)
+Rohrs and Bowman, 1980

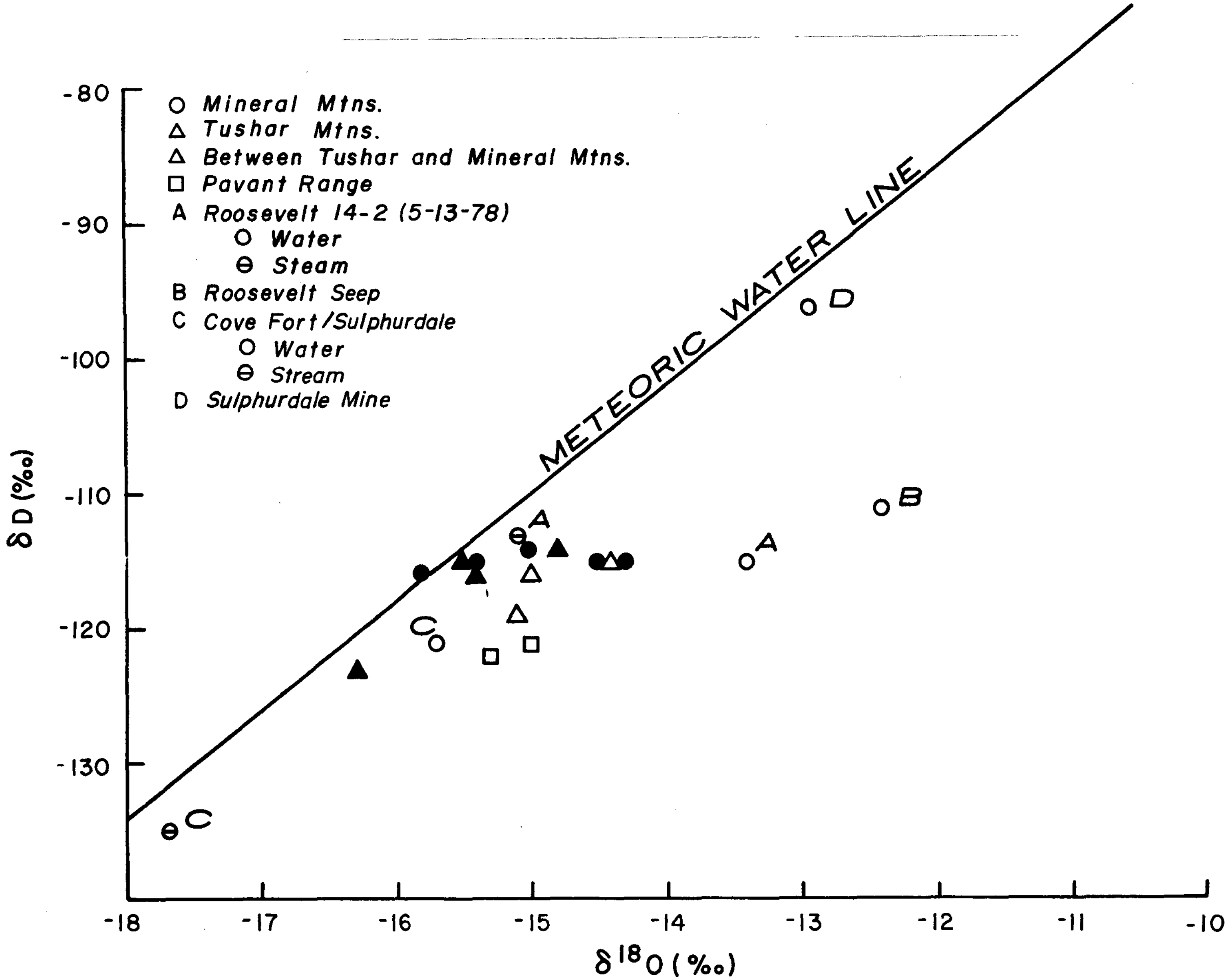

Figure 2. Plot of the δD and $\delta^{18}O$ values of spring waters and geothermal fluids from the Roosevelt Hot Springs and Cove Fort-Sulphurdale Thermal Areas and surrounding region.

Mineral and Tushar Mountains. This similarity within the new data set confirms the conclusion by Rohrs and Bowman (1980) that the Mineral Range is the most likely recharge area for the geothermal fluids at Roosevelt Hot Springs. The $\delta^{18}O$ value is approximately -13.7, which has been estimated from the percentages of steam and water produced by well 14-2 (A. H. Truesdell, personal communication). An average $\delta^{18}O$ value for spring waters from the Mineral Mountains is -14.9, while the $\delta^{18}O$ value for spring waters with a δD of -116 is about -15.6 These values provide an isotopic shift of +1.2 to +1.9 for the thermal waters of well 14-2. These shifts compare to a minimum isotopic shift at Wairakei of +0.5 (Clayton and Steiner, 1975) and a maximum shift of +13.0 at Salton Sea (Clayton et al., 1968).

The reanalyzed $\delta^{18}O$ and δD values of the thermal waters from the Roosevelt seep are about -12.6 and -111, respectively, in good agreement with the old values. The enrichment of $\delta^{18}O$ and δD in the seep suggests modification of the deep thermal waters by evaporation at the surface or boiling in the subsurface.

The location of Union Well 42-7 to surface geology in the Cove Fort/Sulphurdale area is shown on Plate I from the DOE Technical Report for the geology of this area (Moore and Samberg, 1979). Also shown are two local springs (Sulphurdale North and South) plus the Sulphurdale mine seep. The δD values of thermal water from the Cove Fort/Sulphurdale thermal area equal -123. At the low temperatures (110°C) and low flow rates of the flow test of May, 1978, complete steam/water separation was not possible to achieve. The hydrogen and oxygen isotope composition of the steam and water separates also indicate incomplete separation, although the steam separate is lighter in both

δD and $\delta^{18}O$ as would be predicted from known fractionation factors (Friedman and O'Neil, 1977). Temperatures of separation determined from the known hydrogen and oxygen isotope fractionation factors are 100 and 250°C, respectively.

Because steam/water ratios are very low, we can approximate the δD values of the thermal fluid at Cove Fort/Sulphurdale to be equal to that of the water separate, or -123 permil. Because the steam fraction is depleted in δD, the actual values of total fluid will be somewhat lighter. Comparison of this value with those of springs in the Pavant and Tushar Mountains indicates that the springs in the vicinity of Cove Fort/Sulphurdale (Sulphurdale North and South springs) have δD values too heavy (-116 and -115 permil, respectively) to be the primary recharge for the thermal fluids. Only the springs at higher elevations in the Pavant and Tushar Mountains (e.g., North Creek, Magpie, Mud) have δD values consistent with those of the thermal fluids from well 42-7. The presence of high-elevation recharge areas combined with lower elevation springs of diverse origin imply relatively deep circulation of recharge fluids and/or the presence of discrete, and structurally controlled, aquifers in the Cove Fort/Sulphurdale area.

The ^{18}O shift in the thermal fluids of +0.7 per mil is quite small and reflects either high water/rock ratio, incomplete exchange due to low temperatures (<200°C), incomplete exchange due to rapid transit through the geothermal system, or a combination of these mechanisms. There is no indication in the oxygen-isotope data of fluid interaction with the high ^{18}O carbonate rocks known to be at depth in this system.

Water from the Sulphurdale mine seep (T = 65°C) has significantly higher δD (-96) and $\delta^{18}O$ values (-12.9) relative to the thermal fluids and any spring water in the vicinity. These anomalous values result from the kinetic effects of evaporation at temperatures between 50 and 90°C (Craig, 1963).

PART II. Stable Isotope Studies of Clays and Calcite from Shallow Alteration at Roosevelt Hot Springs

In our first study (Rohrs and Bowman, 1980), we reported carbon, oxygen and hydrogen isotope analyses of whole rock and mineral separates from well cuttings from the Roosevelt Hot Springs thermal area. The solid samples utilized in that study were all from deep wells, and thus the study concentrated on the isotopic characteristics of deep and intermediate hydrothermal alteration in the Roosevelt Hot Springs system. The alteration mineralogy and isotope characteristics in the deep well materials were the result of interaction of rocks of quartz monzonite compositions with ascending hot (~250°C) fluids. However alteration and geochemical studies by Parry et al. (1980) of the shallow and surface alteration mineralogy suggests that these assemblages did not result from simple interaction of quartz monzonite with the cooled equivalent of the deep geothermal reservoir fluid. Rather, additional effects of boiling and/or oxidation were suggested. Both processes may impose significant changes on the hydrogen, oxygen and carbon isotope compositions of the system. In order to study the relationship between the shallow acid-sulfate alteration and the deeper propylitic alteration, we recommended in our first study that the minerals from the acid-sulfate alteration be isotopically analyzed. Specifically, we suggested that the isotopic composition of the clay minerals, kaolinite, montmorillonite, and K-mica, might shed light upon whether the fluids responsible for the acid-sulfate alteration are derived from steam which condenses in the alluvium. This study was initiated in part to see if isotopic effects relating to this process could be identified.

In this study we report hydrogen and oxygen isotope analyses of clay mineral separates from alteration assemblages at less than 60 m depth in the University of Utah Well 76-1. The objectives of the study were to determine if clays formed in shallow, low-temperature parts of geothermal systems are isotopically distinct from clays formed in normal low-temperature weathering environments; to determine if isotopic data from clay minerals could help elucidate processes of fluid/rock interaction at the low temperatures characteristic of the upper parts of geothermal systems; and to see if clay minerals at Roosevelt record effects of boiling and/or oxidation which appear to have complicated interpretation of the shallow alteration.

General Geology

The detailed geology, structure, and hydrothermal alteration of the Roosevelt Hot Springs thermal area have been reported in several sources (Ward et al., 1978; Rohrs and Bowman, 1980) and therefore only a summary with pertinent facts will be presented here. The Roosevelt Hot Springs thermal area is located on the western flank of the Mineral Mountains range in southwestern Utah, about 21 km northeast of Milford, Utah (Fig. 3). Forming a northward-trending horst near the eastern extent of the Basin and Range province, the mountains near the geothermal area are composed primarily of a granitic intrusive complex of approximately 250 km^2 with K-Ar dates from 9.4 to 14.0 m.y. (Armstrong, 1970; Park, 1971; Ward et al., 1978). Erosion of the igneous and metamorphic rocks exposed in the Mineral Mountains provides an apron of alluvium along the base of the range. These sediments are primarily lithic sandstones and conglomerates, arkosic in composition, composed primarily of granitic detritus with sparse clasts of pumice, rhyolite, obsidian and gneiss.

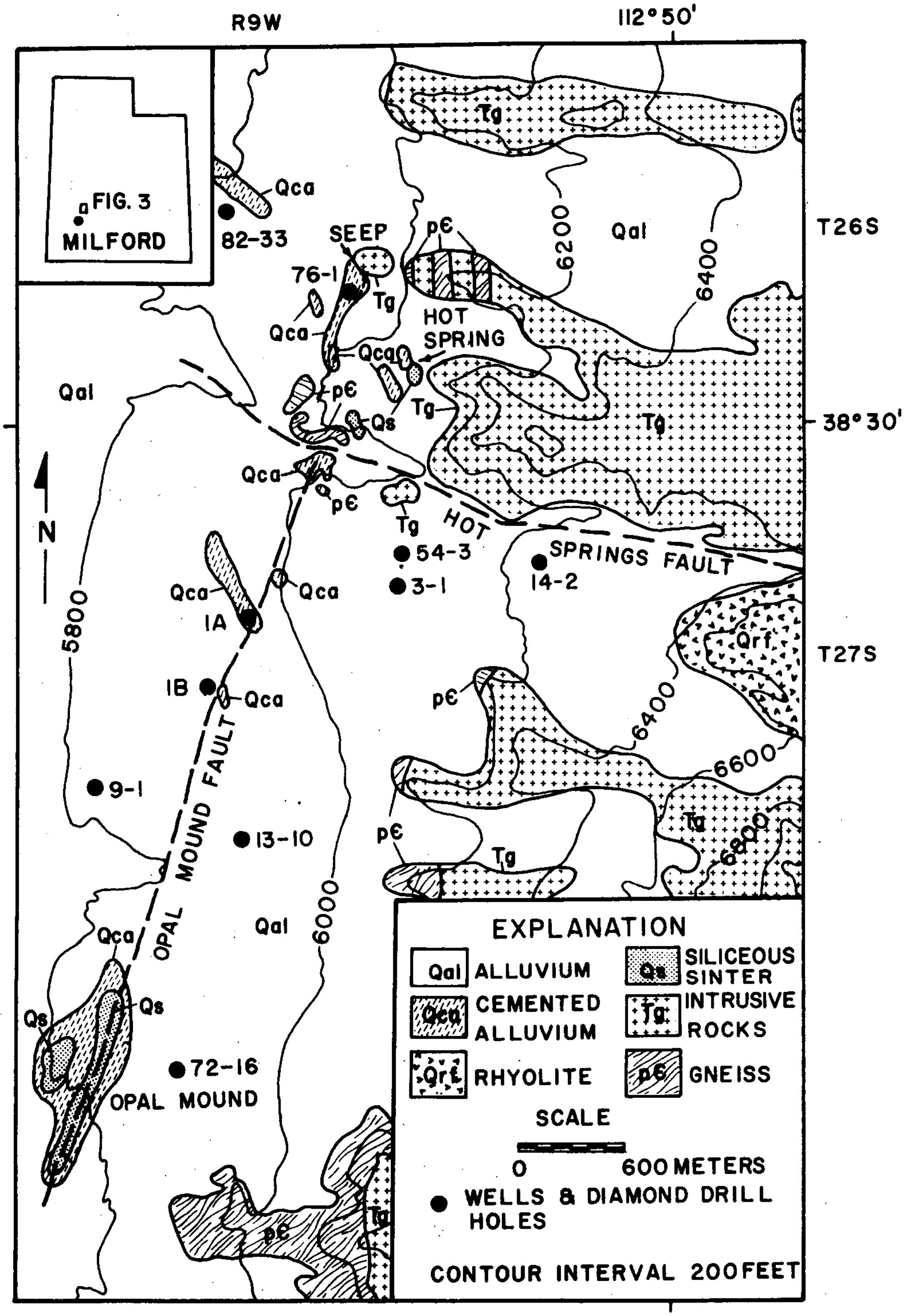

Figure 3. Geologic map of the Roosevelt Hot Springs thermal area, Utah (adapted from Parry et al., 1980).

Extensive fault systems within the basement and alluvium have been utilized by the ascending geothermal fluids and are the sites of hydrothermal alteration in the area. Two types of alteration have been encountered in the thermal area. An acid-sulfate alteration is confined primarily to shallow depths, while a propylitic alteration, characteristic of the deep reservoir, extends to at least 2 km beneath the surface. The deep propylitic alteration was the focus of our earlier study (Rohrs and Bowman, 1980). The surficial alteration, produced by proposed acid-sulfate waters (Parry et al., 1980), is vertically zoned. Near the surface, alluvium is altered to opal and alunite, and only quartz remains unaffected in the most highly altered rocks. Particularly notable is the Opal Mound, a small horst composed of varicolored, laminated opal (Ward et al., 1978). This is one of several hot springs deposits along the Opal Mound fault (Figure 3). Successive zones of alunite-kaolinite, kaolinite-montmorillonite, and muscovite-pyrite extend to 70 m beneath the surface (Parry et al., 1980).

Analytical Results

The hydrogen and oxygen isotope analyses and a summary of alteration mineralogy from Parry et al. (1980) are tabulated as a function of depth in Well 76-1 in Table 2. The isotopic data are plotted in Figure 4 along with the kaolinite-montmorillonite lines defined by Savin and Epstein (1970). These lines represent the hydrogen and oxygen isotope compositions of kaolinitic and montmorillonitic clays, respectively, that have formed in equilibrium with surface waters in the low-temperature (5 to 25°C) surface weathering environment.

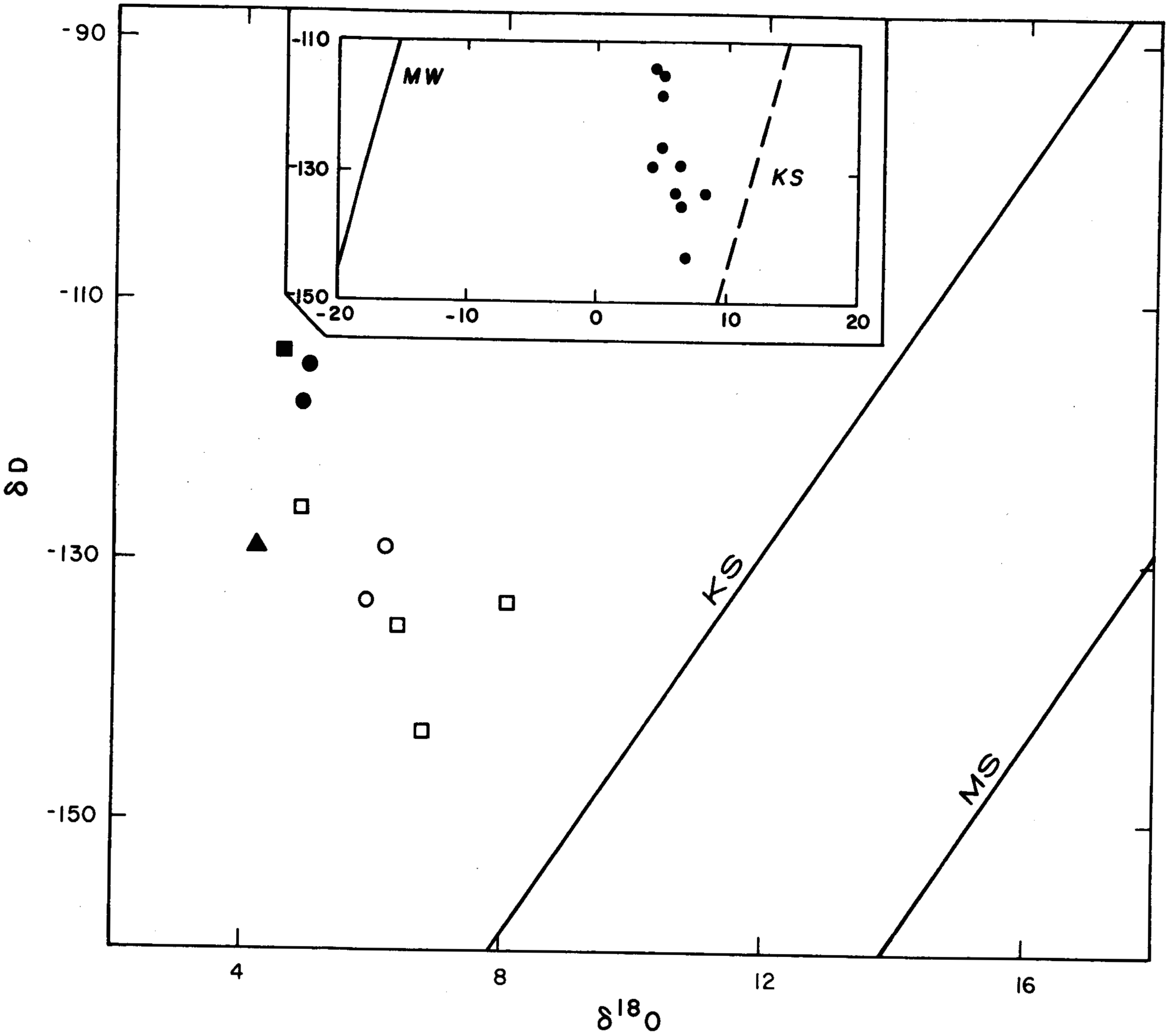

Figure 4. Plot of the δD and $\delta^{18}O$ values of the clay mineral separates from well 76-1, Roosevelt Hot Springs thermal area. Solid symbols are clays from below 38.3 meters, open symbols represent clays above 38.3 meters.

Table 2. $\delta^{18}O$ and δD values of silicates from well DDH 1976-1 with lithologic[1] and alteration mineralogy[2] descriptions

Depth (meters)	Lithology[1]	Alteration Intensity	Alteration[2] Mineralogy	Mineral Analyzed	$\delta^{18}O$	δD
4.0	Qal	strong	Op + Al	Opal	11.8	
11.7	Qal	moderate	Al + K + Op + Hm	Kaolinite	8.1	
21.7	Qal	weak	K + Mo + Op + Hm	Montmorillonite	7.4	-129
31.7	Qal	moderate	K + Mo + Op + Hm	Kaolinite	7.7	-143
33.3	Qal	weak	K + Mo + Py	Kaolinite	6.4	
38.3	qm	weak	K + Mo + Py	Montmorillonite	4.9	-126
				Kaolinite	5.9	-133
41.7	qm	moderate	K + Mo	Montmorillonite	5.0	-115
53.3	qm	moderate	K + Mo	Montmorillonite	4.9	-118
				Kaolinite	4.6	
56.1	qm	v. weak	Km + Ca + Py	K-mica	4.7	
				K-feldspar	7.5	
				Biotite	4.2	-129
				Quartz	10.2	
				Whole Rock	7.4	

1. Abbreviations are: Qal = quartz monzonite alluvium; qm = quartz monzonite

2. Abbreviations are: Al = alunite; Ca = calcite; Hm = hematite; K = kaolinite; Mo = montmorillonite; Op = opal; Py = pyrite

The clays from Well 76-1 have $\delta^{18}O$ values from 4.0 to 8.1 permil and δD values from -115 to -144 permil. Two characteristics of the clay data are apparent from the figure. First, all δD and $\delta^{18}O$ values of the clay minerals plot significantly above and to the left of both the kaolinite and montmorillonite lines, as would be expected from their formation at temperatures above 25°C. It is apparent from this plot that, at least in the Roosevelt Hot Springs thermal area, near-surface clays formed by venting of geothermal fluids are isotopically distinct from normal weathering or supergene clays. Secondly, the clays from below 115 ft (38.3 m) are significantly heavier in deuterium (δD values greater than -129 permil) and generally lighter in $\delta^{18}O$ ($\delta^{18}O$ values less than 5.0 permil) than the clays formed in the granitic alluvium above 115 ft (38.3 m). The clay separates below 38.3 meters are all from the quartz monzonite and the clays from less than 38.3 meters are hydrothermal clays formed in the opal-cemented granitic alluvium.

Discussion

The higher oxygen-isotope compositions of clays above 38.3 meters may be explained as the result of lower temperature of formation and of a previously weathered protolith , i.e., granitic alluvium, with higher $\delta^{18}O$ value. The difference in δD values between the shallow (<38.3 m) and deep (>38.3 m) clays is also retained in the δD values of water in equilibrium with the clays (Figure 5), calculated along the measured thermal gradient, as tabulated in Table 3. There are uncertainties in our estimate of $10^3 \ln \alpha$ (clay-H_2O) at these low temperatures, but this will not affect the relative difference between the deep and shallow clays. The hydrogen-isotope data indicate that the shallow and deep clays in well 76-1 did not directly equilibrate with and

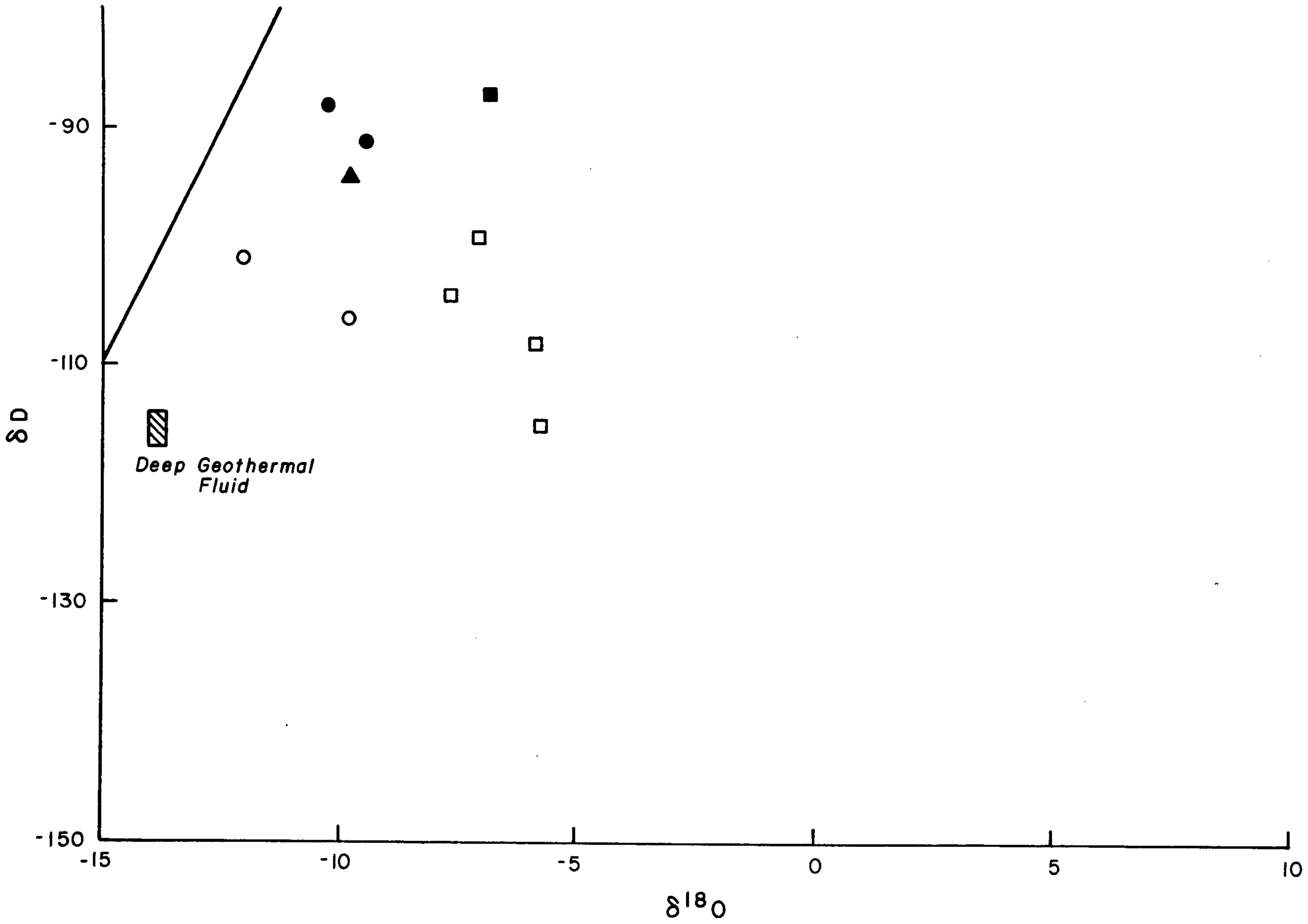

Figure 5. Plot of calculated δD and $\delta^{18}O$ values of water in equilibrium with the clay mineral separates from well 76-1, Roosevelt Hot Springs thermal area. Symbols as in Figure 4.

Table 3. Calculated δD-$\delta^{18}O$ of Fluid in Equilibrium with Shallow Clays

A. T control - Drill hole measured T from Sill and Bodell (1977).

1. O-isotope fractionation data.
   - a. Kaolinite: $10^3 \ln \alpha = 2.05\times10^6/T^2 = 3.85$ (Yeh and Savin, 1977).
   - b. Montmorillonite: $10^3 \ln \alpha = 2.67\times10^6/T^2 - 4.82$ (Yeh and Savin, 1977).
2. H-isotope fractionation data: Figure 4, Rohrs and Bowman (1980).

B. Calculations

Depth (meters)	Clay Mineral	T	$\delta^{18}O(H_2O)$	ΔH	$\delta D(H_2O)$
11.7	Kaolinite	51°C	-7.58	-28.8	-104
21.7	Montmorillonite	67°C	-12.08	-28.1	-101
31.7	Kaolinite	81°C	-5.71	-27.6	-115
33.3	Kaolinite	84°C	-5.84	-27.4	-108
38.3	Kaolinite	88°C	-6.98	-27.3	-99
38.3	Montmorillonite	88°C	-9.76	-27.3	-106
41.7	Montmorillonite	92°C	-10.22	-27.1	-88
53.3	Montmorillonite	101°C	-9.37	-27	-91
53.3	Kaolinite	101°C	-6.80	-27	-87
56.1	Biotite	104°C	-9.76	-26.5	-94

form from a common hydrogen reservoir. The calculated δD values of fluid in equilibrium with the clays below 38.3 m are consistent with those of still deeper clay minerals (sericite) in Thermal Power Co. well 14-2, which appear to record some effects of evaporation (Rohrs and Bowman, 1980). The deuterium-depleted fluids responsible for clay formation in the granitic alluvium may be the result of mixing of deep geothermal reservoir with cold meteoric water, or more likely, samples containing a significant component of condensed steam from the geothermal reservoir. The latter mechanism would be consistent with the type of fluid proposed by Parry et al. (1980) to be responsible for the shallow alteration in well 76-1.

Water/Rock Ratios

The oxygen-isotope shifts recorded in the thermal fluids and clay mineral alteration products can be used to estimate minimum water/rock ratios (atomic oxygen) attendant with the shallow alteration in the Roosevelt Hot Springs thermal area.

The amount of water involved in the clay alteration during the circulation of thermal waters can be calculated from the following mass balance equation (Taylor, 1974):

$$W\ \delta^{18}O_w^i + R\ \delta^{18}O_r^i = W\ \delta^{18}O_w^f + R\ \delta^{18}O_r^f$$

where i and f refer to the initial and final isotopic composition of the water and the rock, and W and R represent the atom percent of exchangeable oxygen in the total system for the water and the rock.

To obtain an estimate of the W/R atomic oxygen ratio, this equation can be rearranged to

$$W/R = \frac{\delta^{18}O_r^f - \delta^{18}O_r^i}{\delta^{18}O_w^i - \delta^{18}O_w^f}$$

By determining the quantity of exchangeable oxygen in the rock through chemical analysis or by estimating the mineralogy and density of the rock, a water/rock mass ratio can be obtained. Rohrs and Bowman (1980) estimated that average fresh or altered rock at Roosevelt Hot Springs contains approximately 29 moles of exchangeable oxygen per kilogram of rock. Multiplying by the W/R atomic oxygen ratio and by the molecular weight of water, the ratio kg water/kg rock is obtained, which represents the water/rock mass ratio.

For these calculations an initial $\delta^{18}O$ value for the igneous rock was chosen as 7.4, a value corresponding to that of an unaltered quartz monzonite at 56.1 meters in UU well 76-1 (Rohrs and Bowman, 1980). Two initial $\delta^{18}O$ values of fluid were chosen: a value of -14.9 permil corresponding to the average of all Mineral Mountains spring waters, and a value of -15.6, an average of $\delta^{18}O$ values of spring waters from higher elevations in the Mineral Mountains. Water/rock ratios calculated using each value are listed under A and B, respectively, in Table 4. Two final $\delta^{18}O$ values of fluid were also chosen: A value of -13.7 permil measured for the reservoir fluid from Thermal Power Co. well 14-2 and a value of -13.0 permil which was proposed by Rohrs and Bowman (1980) to represent average thermal fluid for the geothermal system. These values were used in calculations of water/rock ratios labeled A and B, respectively.

Table 4. Data and results for water/rock ratio calculations.

Depth (meters)	$\delta^{18}O$ (clay)	T(k)	W/R (atomic ratio) A	W/R (atomic ratio) B	W/R (mass ratio) A	W/R (mass ratio) B
11.7	7.1	324	0.2	0.1	0.1	0.05
21.7	6.2	340	1.0	0.5	0.6	0.3
31.7	6.8	354	0.5	0.2	0.3	0.1
33.3	6.4	357	0.8	0.4	0.4	0.2
38.3	4.9	361	2.1	1.0	1.2	0.6
41.7	5.0	365	2.0	0.9	1.1	0.5
53.3	4.9	374	2.1	1.0	1.2	0.6

Minimum water/rock ratios therefore have been calculated for two isotopic shifts -- one of +1.2 and a larger shift of +2.6. The data and results for these calculations are presented in Table 4.

Water/rock mass oxygen ratios vary from 0.05 to 1.20, and average 0.81. This average is somewhat lower than that calculated for deep hydrothermal alteration (1.23) at Roosevelt (Rohrs and Bowman, 1980). In comparison, average water/rock mass ratios for the Salton Sea and Wairakei have been estimated as 0.45 and 4.3, respectively (Clayton et al., 1968; Clayton and Steiner, 1975). These generally low water/rock ratios are consistent with the lack of pervasive hydrothermal alteration in UU well 76-1. The variations in water/rock ratio with position within UU well 76-1 are similar to the variations found in the deep alteration at Roosevelt Hot Springs. These variations indicate significant variations in permeability and imply that in zones with low water/rock ratios, the isotopic composition of the ascending thermal fluid has been significantly modified. The sharp drop in water/rock ratio above 38.3 meters may be related to the presence of low-density fluid (steam) in the rock voids, as suggested by the hydrogen-isotope data.

As in the deep alteration at Roosevelt, water/rock ratios vary widely, indicating wide fluctuations in permeability. The variability of water/rock ratios at Roosevelt Hot Springs has important implications for the geothermal system. Moderately to strongly altered rocks possess high water/rock ratios, indicating their importance as conduits for the thermal waters at some point in the alteration history of the rock. Conversely, weak alteration and subsequent low water/rock ratios are characteristics of less permeable rocks. Since permeability is fracture-controlled within the reservoir, alteration intensity and water/rock ratios emphasize the importance of fractures and fissures in directing fluid movement. However, water/rock ratios cannot distinguish between present and paleo- conduits.

PART III. Analysis of Spring and Thermal Waters from the Goshen Valley Area, Central Utah

Cleary (1978) demonstrated that spring waters indicating chemical temperatures in excess of 100°C are widely distributed in the Goshen and Pavant Valleys of central Utah. These temperatures were calculated by use of the standard Na-K-Ca and SiO_2 thermometers. However Cleary (1978) emphasized that further work would be required to evaluate the effects of brine mixing, clay exchange, and mixing large volumes of cool water with thermal water on these thermometer results. Hydrogen and oxygen isotopic compositions of these waters may record the effects of fluid mixing or clay exchange. Hydrogen and oxygen isotope analyses were therefore performed in an attempt to evaluate these effects so that a more realistic evaluation of thermal potential can be made in these areas.

The isotopic data are compiled in Table 5 and illustrated in Figure 6. The data fall into three distinct groups. One group, including all the Goshen Valley A-series of samples, plot close to and at slightly greater $\delta^{18}O$ values than the meteoric water line. All but one of these (Goshen Valley Warm Springs) are cold springs with comparatively low total dissolved salt contents (Cleary, 1978). The isotope data for these springs do not reflect any extensive interaction with additional hydrogen and oxygen reservoirs. However, Na-Ca-K and SiO_2 temperatures for most of these waters exceed 80 and 106°C, respectively, indicating that additional processes of clay exchange, interaction with evaporite sequences, etc. are operative. These data indicate that exchange/interaction processes acting sufficiently to modify chemical thermometers may produce no obvious effects in the hydrogen and oxygen-isotope

Table 5. Isotope analyses of spring waters from the Goshen and Pavant areas, central Utah.

	Date Collected	δ D	$\delta^{18}O$
Black	7/20/77	-100	-15.1
Cinder	7/19/77	- 95	-14.0
Coyote North	7/15/77	-110	-15.3
Coyote South	7/20/77	-118	-15.7
ELB	7/20/77		-15.3
Goshen Valley A-1	7/27/77	-120	-15.9
" A-2	7/27/77	-118	-15.7
" A-3	7/27/77	-121	-16.3
" A-4	7/27/77	-122	-16.3
" A-5	7/28/77	-126, -126	-16.4
" A-6	7/28/77	-125	-16.8
" A-7	7/28/77	-119	-15.8
" A-8	7/28/77	-117, -117	-15.1
" Warm Springs	7/28/77	-124	-16.1
Meadow	7/15/77	-104	-15.2
Meadow East	7/23/77	- 91	-15.6
North Twin	7/20/77	-105	-15.9
Sulphur Spring North	7/22/77	- 91	-15.6
Sulphur Spring South	7/22/77	- 97	-16.2
Squidike	7/14/77	-119	-15.4

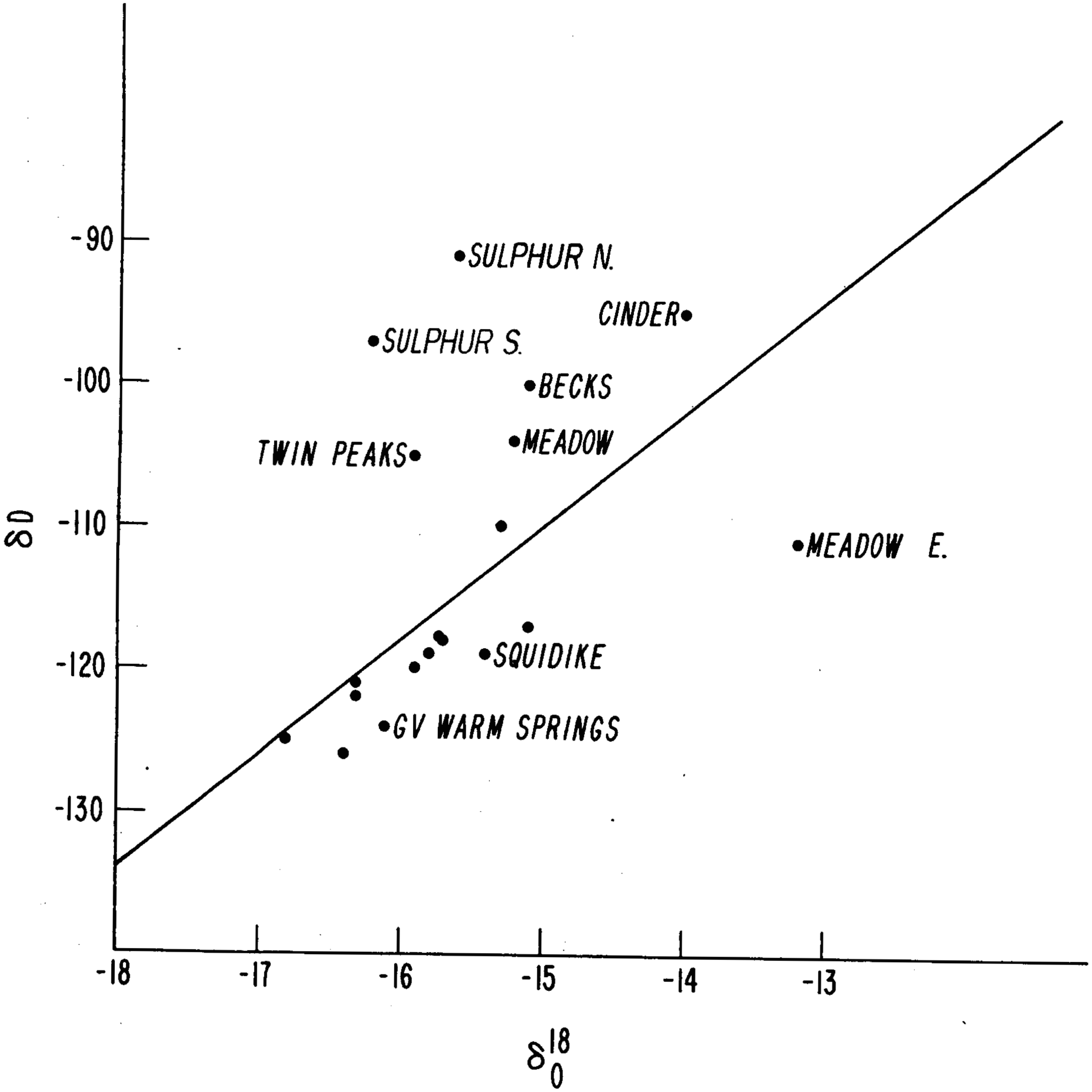

Figure 6. Plot of the δD and δ^{18}O values of spring waters from the Goshen and Pavant areas, Central Utah.

compositions of the water. This is an expectable result of mass-balance, as oxygen and hydrogen are the most abundant elements in low-density aqueous fluids.

Meadow East Spring constitutes the second group which plots well to the right of the meteoric water line. The spring has probably undergone either evaporation effects or mixing with saline brine to produce the observed ^{18}O enrichment. The spring is a cold spring -- 24°C (Cleary, 1978) -- but contains high concentrations of $SO_4^=$ (1400 ppm), Na (1200 ppm), Ca (483 ppm), and Cl^- (2000 ppm) and Na-Ca-K and SiO_2 temperatures of 208 and 113°C, respectively, suggestive of interaction with saline evaporative brines or evaporite deposits. The combined evidence of water and isotope chemistry suggests the action of fluid mixing or fluid-evaporite interaction to produce the observed isotopic and chemical characteristics of Meadow East Spring water.

The third group of waters are quite unusual in that they plot above and to the left of the meteoric water line. That is, the waters are preferentially enriched in D or depleted in ^{18}O, or both. There are few examples of natural waters deviating from the meteoric water line in this manner (Frape and Fritz, 1981) and the processes producing such waters are not clear. Nevertheless, all the spring waters plotting in this group were at temperatures in excess of 30°C, are $SO_4^=$-rich (>400 ppm), and are usually (with the exceptions of Sulphur Spring North and South) saline (Cl^- contents in excess of 1800 ppm). In addition, all have Na-Ca-K and SiO_2 temperatures in excess of 93 and 89°C, respectively.

Conclusions

The main conclusions of this study may be summarized as follows:

1) Recollection of spring waters from the Mineral, Tushar, and Pavant Ranges and their hydrogen and oxygen-isotope analyses indicate that the groundwater supply in these potential recharge areas has been isotopically uniform for at least two years. This consistency implies that one can confidently evaluate whether specific recharge areas can contribute fluid to a geothermal system without long-term, i.e. multi-year, monitoring of the potential areas.

2) The hydrogen and oxygen-isotope analyses of the new spring samples confirm the conclusion by Rohrs and Bowman (1980) that the thermal fluids at Roosevelt Hot Springs originate as meteoric water in the Mineral Mountains and become enriched in ^{18}O by exchange with the crystalline country rock.

3) The thermal fluids in the Cove Fort-Sulphurdale Thermal Area originate as meteoric waters at high elevation in the Tushar and/or Pavant Ranges and are isolated from low-elevation springs in the vicinity of Union well #42-7.

4) The thermal fluids at Cove Fort-Sulphurdale display only a small ^{18}O shift of +0.7 per mil, implying high water/rock ratios, incomplete oxygen-isotope exchange, or both. These data do not indicate extensive interaction between the thermal fluids and high ^{18}O marble country rock in the geothermal system.

5) Hydrogen and oxygen-isotope analyses of clay mineral separates from the shallow alteration at Roosevelt Hot Springs (UU well #76-1) indicate that clays formed by venting of low-temperature thermal fluids are distinct isotopically from clays formed in the surface weathering environment. Such a distinction provides an effective exploration tool for near-surface manifestations of geothermal activity. However the cost and turn-around time constraints of the method would sharply limit its utility.

6) Fluid in equilibrium with the acid-sulfate alteration above 38.3 meters is distinct isotopically from that in equilibrium with shallow alteration below 38.3 m and deep alteration in the Roosevelt Hot Springs geothermal system. Evolution of these shallow fluids from hotter reservoir fluids requires significant boiling and recondensation of steam in these upper zones to explain the hydrogen-isotope data. The isotope evidence is thus consistent with the proposal by Parry et al. (1980) that the fluid chemistries required to produce acid-sulfate alteration assemblages would result from boiling and steam recondensation, with oxidation of $S^=$ to $SO_4^=$, of the deep thermal fluid.

7) Fresh or low-salinity spring waters from the Goshen Valley area of central Utah are distinct isotopically and chemically (Cleary, 1978) from more saline spring or well waters in the area. However, there are no good detailed correlations evident between either the hydrogen or oxygen-isotope composition of the waters and salt content, species concentrations, or chemical thermometry. This reconnaissance study would caution expectations of detailed tracking of groundwater evolution with isotope techniques alone. However, closer examination of this approach, especially involving long-term monitoring (sampling) of the groundwater systems, is certainly warranted.

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