DOE/ET/28392-46
78-1701.a.1.5

TOPICAL REPORT

A LIGHT STABLE ISOTOPE STUDY OF THE ROOSEVELT HOT SPRINGS THERMAL AREA, SOUTHWESTERN UTAH

by

David T. Rohrs and John R. Bowman
Dept. of Geology and Geophysics
University of Utah
Salt Lake City, Utah

DISCLAIMER

NOTICE

NOTICE

ABSTRACT

The isotopic composition of hydrogen, oxygen, and carbon has been determined for regional cold springs, thermal fluids, and rocks and minerals from the Roosevelt Hot Springs thermal area. The geothermal system has developed within plutonic granitic rocks and amphibolite facies gneiss, relying upon fracture-controlled permeability for the migration of the thermal fluids. Probably originating as meteoric waters in the upper elevations of the Mineral Mountains, the thermal waters sampled in the production wells display an oxygen isotopic shift of at least +1.2. Depletions of $\delta^{18}O$ in whole rock, K-feldspar, and biotite have a positive correlation with alteration intensity. W/R mass ratios, calculated from the isotopic shifts of rock and water, range up to 3.0 in a producing horizon of one well, although the K-feldspar has experienced only 30 percent exchange with the thermal waters. While veinlet quartz has equilibrated with the thermal waters, the ^{18}O values of K-mica clay, an alteration product of plagioclase, mimic the isotopic composition of K-feldspar and whole rock. This suggests that locally small W/R ratios enable plagioclase to influence its alteration products by isotopic exchange.

Oxygen-isotope temperatures based on observed $\delta^{18}O$ values of calcite and an assumed constant $\delta^{18}O$ value of thermal fluid are erratic and significantly lower than measured thermal gradients or fluid inclusion filling temperatures. This lack of agreement is the

result of our assumption of a constant oxygen isotopic composition of thermal fluid at all levels within the system. Rather, the $\delta^{18}O$ values of calcite indicate enrichment of the thermal water with $\delta^{18}O$ in weakly altered rocks, probably from isotopic exchange under low W/R ratios and slow circulation rates in relatively impermeable rocks. These complications indicate that casual application of mineral isotope thermometers without consideration of potential variations in fluid isotopic composition can lead to misinterpretation. The calcite, like the K-mica, may also be influenced by the host plagioclase. The systematic increase in $\delta^{13}C$ values of calcite, generally from -6.0 to -2.4 approaching shallower depths, can be reproduced by two separate models. Each model requires the interaction of the three components of the carbon reservoir: $CO_2(g)$, fluid ($H_2CO_3app + HCO_3^-$), and calcite. Precipitation of calcite modifies the $\delta^{13}C$ of $CO_2(g)$ and fluid and provides a decrease in the molality of the total dissolved carbonate which is consistent with the decrease in molality predicted by mineral equilibria. The $\delta^{13}C$ of the total carbon reservoir is between -4.5 and -5.8. It is not possible to unambiguously assign a carbon source to these values.

The small isotopic shift of the thermal waters, the small extent of exchange experienced by the rocks in the production zone, and the enriched $\delta^{18}O$ values of K-mica clay attest to the lack of communication between the rock and the bulk of the thermal waters. The circulation of the thermal water is evidently restricted to permeable fracture zones with little infiltration into the surrounding impermeable rocks, accounting for the variabilities in alteration and

W/R ratios. However, a lack of systematically collected core samples has precluded closer definition of the processes responsible for these characteristics.

TABLE OF CONTENTS

LIST OF ILLUSTRATIONS

LIST OF TABLES

ACKNOWLEDGMENTS

Dennis L. Nielson and his colleagues at the Earth Science Laboratory, Salt Lake City, contributed geological and geophysical well log data and materials used in this study. Dr. W. T. Parry provided fluid inclusion temperatures of hydrothermal quartz. We are indebted to Walter Kropp and Ray Lambert for their technical assistance during mineral separation and analysis and to Leslie Smith for his pertinent comments concerning hydrologic modeling. We would like to thank A. H. Truesdell of the USGS, Menlo Park, for providing certain isotopic analyses of spring and thermal waters. All other isotopic measurements were made at the Laboratory of Isotope Geology, Department of Geology and Geophysics, University of Utah. This research was funded under DOE/DGE contracts DE-AC07-78ET128392 (formerly EG-78-C-07-1701). Drs. W. T. Parry and J. A. Whelan have critically reviewed this report.

INTRODUCTION

Roosevelt Hot Springs is an active geothermal area in southwestern Utah currently undergoing evaluation as an energy resource. This active geothermal system provides the unique opportunity to directly sample water involved in hydrothermal alteration, enabling the study of water/rock interaction in a natural environment at the low temperatures often precluded in laboratory experiments due to slow equilibration rates.

Because light stable isotopes of fluids and minerals elucidate upon the interaction of water and rock within the geothermal reservoir, isotopic analyses of hydrogen, oxygen, and carbon from regional cold waters, thermal fluids, and hydrothermally altered rocks have been applied to specific problems at Roosevelt Hot Springs. These include the origin of the thermal water, the location of permeable pathways within the subsurface influencing the circulation of the water, and processes involving the interaction of water with rock, such as isotopic exchange and boiling. The isotopic data, substantiated with additional geological, geochemical and geophysical evidence, provide insight into the hydrological and chemical evolution of the geothermal system, focusing particularly upon the importance of fracture controlled permeability in directing fluid movement and controlling mass proportions of fluid to rock.

Previous Work

Much of the previous work in geothermal areas has been devoted to isotopic measurements of fluids. Early studies by Craig (1963) established the predominance of local rain water as the reservoir recharge fluid in geothermal systems. Recent studies have focused upon the $^{18}O/^{16}O$ and D/H measurements of hot and cold springs to identify reservoir recharge aquifers in Iceland (Arnasson, 1977) and to identify processes of dilution and boiling at Yellowstone (Truesdell et al., 1977) and El Tatio, Chile (Giggenbach, 1978). Geothermometry utilizing the distribution of oxygen between dissolved sulfate and water has been applied as an exploration tool (Sakai, 1977; McKenzie and Truesdell, 1977).

Despite the abundance of isotopic data on fluids from geothermal areas, the data for rocks and minerals is much reduced, possibly due to the difficulty in obtaining pure mineral separates and to the similarity in results and interpretations for the various systems studied thus far. Clayton et al. (1968) in their study of the Salton Sea geothermal field presented the first application of oxygen isotopes to hydrothermal alteration in an active geothermal area. They found that calcite had equilibrated with a water of uniform isotopic composition to as low as 100°C and that silicates showed considerable oxygen isotopic exchange above 150°C, although quartz was resistant to exchange to temperatures as high as 340°C. The large isotopic shift of the thermal waters resulted from a relatively small water to rock ratio in the reservoir.

An isotopic study of the altered rocks in the Ohaki-Broadlands field in New Zealand (Eslinger and Savin, 1973) found that fine grained silicates had equilibrated with the thermal water to temperatures between 160 and 270°C, although calcite was not in equilibrium with these phases or with the water below 200°C. On the basis of fissure-grown quartz, adularia, and calcite crystals from Ohaki-Broadlands, Blattner (1975) postulated that the isotopic analysis of minerals could be used to identify zones of boiling and to determine the past hydrologic stability in geothermal systems.

The oxygen isotope study at Wairakei, New Zealand, (Clayton and Steiner, 1975) revealed that temperatures calculated from the fractionation between calcite and water were in good agreement with measured temperatures above 200°C. Where calculated temperatures exceeded measured temperatures, they suggested that either the carbonates recorded previously higher temperatures or that the waters had become depleted with ^{18}O by closed system water/rock isotopic exchange. Because of the slight shift in $\delta^{18}O$ by thermal waters from local meteoric waters, they proposed that water/rock mass ratios were at least ten times greater than those at Salton Sea.

General Geology

The Roosevelt Hot Springs thermal area is located on the western flank of the Mineral Mountain Range in southwestern Utah, about 21 km northeast of Milford, Utah. 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 my

(Armstrong, 1970; Park, 1971; Ward et al., 1978). This pluton has intruded amphibolite facies biotite-hornblende gneisses of probable Precambrian age which are exposed along the western margin of the pluton (figure 1).

Repeated igneous activity during the Cenozoic commenced in the mid-Tertiary with the production of calc-alkalic lavas exposed on the south flank of the Mineral Mountains. This event was followed by late Tertiary rhyolites on the north and west flanks of the mountains and by basalts along the southern flank. The bimodal volcanism continued into the Quaternary from 0.8 to 0.5 my producing rhyolite flows, domes, and pyroclastics along the crest and west flanks of the mountains (Lipman et al., 1977) and basalts along the northeast flank (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.

Lithologies encountered in the production wells have been described by Ballantyne and Parry (1978), Nielson et al. (1978), and Hulen (1979). The predominant igneous rock is a medium to coarse grained hornblende-biotite quartz monzonite, with 5-10% quartz, 40% plagioclase, 40% K-feldspar, 7-10% biotite, and 1-2% hornblende. Dikes include a microdiorite with a subdiabasic texture and a fine grained leucocratic granite. The microdiorite is distinguished by up to 70% plagioclase and 25% hornblende, while the granite contains up

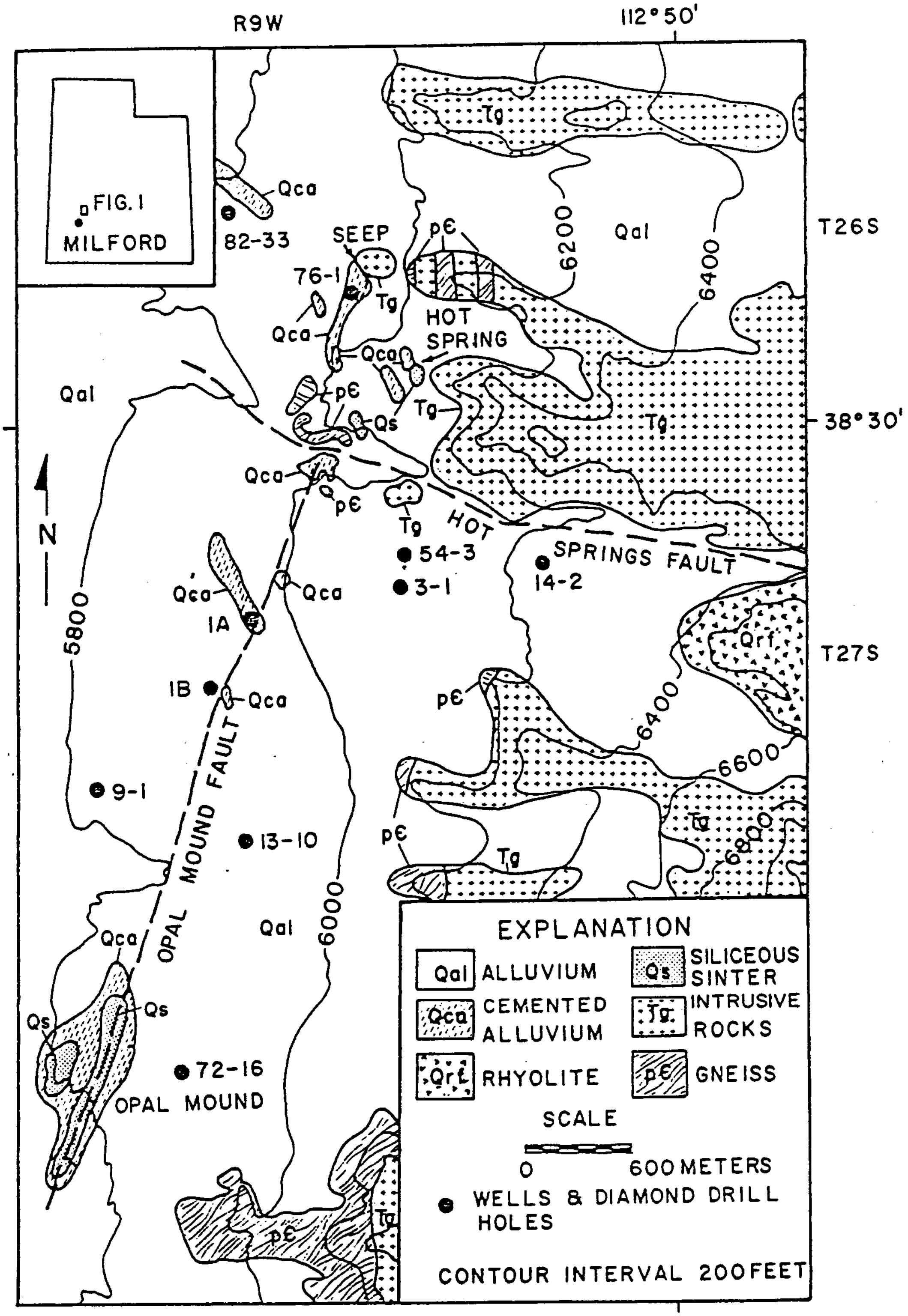

Figure 1. Geologic map of the Roosevelt Hot Springs thermal area, Utah (adapted from Parry et al., 1980).

to 20% quartz, 40-50% K-feldspar, 20-30% plagioclase, and 2% biotite.

Since the metamorphic rocks often possess no characteristic index minerals, the gneiss is difficult to distinguish from the igneous rocks. Thus, the classification of gneiss is based upon increased abundances of hornblende and biotite (Nielson et al., 1978) and apatite. Although the metamorphic rocks present considerable ranges in their mineralogic compositions, they can be subdivided into three predominant lithologies. These include biotite-hornblende quartz monzonite gneiss, biotite-hornblende granodiorite gneiss, and biotite-hornblende-plagioclase banded gneiss. Only the banded gneiss possesses a recognizable foliation in the drill cuttings (Hulen, 1979).

The intrusion of the Mineral Mountains pluton has subjected the metamorphic rocks to at least one additional thermal event which could have produced a pre-geothermal alteration assemblage. Therefore, criteria for distinguishing between metamorphic and igneous rocks should be a subject for continued research at Roosevelt Hot Springs.

Structure

Because of the crystalline nature of the bedrock hosting the deep reservoir, the importance of structure with the thermal area must be emphasized, since numerous faults and fractures determine the permeability of the system and direct fluid flow into and out of the reservoir. Three important fracture patterns have been identified (Nielson et al., 1978): generally north to northeast trending faults, such as the Opal Mound fault; east-west trending faults, such as the

Hot Springs fault; and low angle fault zones dipping to the west.

Figure 1 presents the relationship of the Opal Mound fault to the Hot Springs fault. Of particular importance to the geometry of the reservoir is the Opal Mound fault. In addition to hot spring deposits along this structure, heat flow measurements have delineated a band of high heat flow centered about this fault (Ward et al., 1978), suggesting that the Opal Mound fault has been an important conduit for the ascent of thermal fluids.

Hot Spring Deposits and Alteration

Geological evidence for the high temperature geothermal resource at Roosevelt Hot Springs comes from hot spring deposits along the Opal Mound fault (figure 1). Particularly notable is the opal mound, a small horst composed of varicolored, laminated opal (Ward et al., 1978). Additional hot spring deposits consist of opal and chalcedony cements and siliceous sinter within alluvium. This sinter may be accompanied by a few percent of native sulfur.

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 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. Successive zones of alunite-kaolinite, kaolinite-montmorillonite, and muscovite-pyrite extend to 70 m beneath the surface (Parry et al., 1980).

The propylitic alteration, the focus of this study has been described by Ballantyne and Parry (1978), Rohrs and Parry (1978), Nielson et al. (1978), and Hulen (1979). Interaction of the rocks and the thermal fluids produces an assemblage of quartz, K-feldspar, K-mica clay (sericite), chlorite, calcite, pyrite, and minor chalcopyrite. Limonite, hematite, montmorillonite and mixed layer clays increase in abundance at shallow depths, and epidote occurs sporadically at all depths. While quartz and K-feldspar are stable throughout the system, plagioclase is altered to calcite, clay, and occasionally to chlorite. Hornblende is readily altered to chlorite, calcite and clay, and biotite is susceptible to bleaching and replacement by sericite and chlorite.

Although alteration is described as weak, moderate, or strong in this study, this designation is only intended to show relative alteration intensity. Generally, the alteration would be considered weak even in the most strongly altered rocks. This is evident from computer calculated modal mineralogies (Ballantyne, 1978) which indicate that even strongly altered rocks generally contain less than 20 percent alteration products, although individual mineral grains may show a wide spectrum of alteration intensity. Where alteration intensity increases, veinlets become more abundant, indicating that alteration intensity relates to the porosity of the rocks. Veinlets, which may be mono- or polymineralic, consist primarily of calcite, hematite/limonite, and quartz. Accessory minerals in veinlets include K-feldspar, chlorite, clay, epidote, pyrite, and magnetite.

Water Chemistry

Chemical analyses of selected thermal waters are given in table 1. These include the now extinct Roosevelt Hot Springs, a surficial seep, and waters collected from a production well. For a more complete discussion of the water chemistry, the reader is referred to Ward et al. (1978). Despite some variability, the waters are relatively dilute sodium chloride brines with an ionic strength of 0.1 to 0.2. The high sodium and chloride and moderate sulfate concentrations classify these waters as sodium chloride grading into acid-sulfate, according to White (1957). Because of the high chloride concentrations in both the hot spring and deep reservoir waters, these thermal fluids fit the category of a hot-water system (White, 1970).

Drill Hole Selection

Of the production wells which penetrate into the deep reservoir (figure 1), three were chosen for this isotopic study on the basis of their productivities and their relationships to the geometry of the reservoir. Thermal Power Co. well Utah State 14-2, drilled to a depth of 1859.3 m, is the easternmost production well in the field, positioned approximately 1.1 km east of the intersection of the Opal Mound and Hot Springs faults. On the basis of geophysical well log data and trace element distributions, Bamford (1978) has identified potential hot water entry zones at 488, 870, and 1585 m, which correspond to increases in alteration intensity at these depths (Ballantyne and Parry, 1978).

Thermal Power Co. well Utah State 72-16 lies only 0.6 km east of

Table 1. Major constituents in mg/l in selected water from the Roosevelt Hot Springs Thermal Area, Utah.*

	(1)	(2)	(3)
Na	1840	2072	2500
Ca	122	31	22
K	274	403	488
SiO_2	173	639	313
Mg	25	.26	0
Cl	3210	3532	4240
SO_4	120	48	73
HCO_3	298	25	156
Al		1.86	.04
Fe		.016	
Total Dissolved Solids	6063	6752	7792
Temp.	25°C	92°C	55°C
pH	6.5	5.0	7.9
$t_{Na-K-Ca}$	235	274	283
$t_{Qtz.\ adia.}$	160	244	195
$t_{Qtz.\ cond.}$	170	274	213
$t_{opal\ cond.}$	47	153	88

(1) Roosevelt Seep. University of Utah, June, 1975.

(2) Thermal Power Company Well 72-16, University of Utah, Jan., 1977, surface leakage.

(3) Roosevelt Hot Springs, U.S. Geol. Survey, Sept., 1957, Mundorff (1970).

* From Parry et al., 1980.

the opal mound. Because this production well intersects hot water entry points at considerably shallower depths than well 14-2, at 95, 157, and 191 m (Bamford, 1978), the hole extends to a depth of only 382.2 m.

Phillips Petroleum Co. well Utah State 9-1 is situated about 1.8 km northwest of well 72-16 and about 0.5 km west of the Opal Mound fault. Penetrating 2098.5 m, this well has proved to be non-productive (D. Nielson, personal communication).

Because this study emphasizes the deep reservoir, isotopic analyses of the near surface acid-sulfate alteration will be deferred to a later date. However, one shallow alteration hole, UU 76-1, intersects the freshest granitic rock in the thermal area, providing a relatively unaltered quartz monzonite utilized in calculations of water/rock ratios.

Because sampling of thermal fluids required access to the wells during flow tests, only Thermal Power Co. well 14-2 was successfully sampled. The premature termination of flow tests for well 72-16 precluded successful collection of these fluids, while no water has been sampled for isotopic analysis from well 9-1.

Sample Selection and Preparation

Spring waters

Waters from surficial springs and a seep were collected for hydrogen and oxygen analysis from the vicinity of Roosevelt Hot Springs by both the USGS and the University of Utah during the winter and summer of 1976 and fall of 1977. Spring 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).

Geothermal wells

Reservoir fluids were sampled by the USGS and the University of Utah during flow tests by Thermal Power Co. in November, 1976 and May, 1977. In order to obtain representative samples of discharged fluids, a steam/fluid miniseparator of New Zealand design was used for these collections.. Problems attendant in the collection of steam stemmed from the difficulty of adjusting the miniseparator. The implications of this difficulty will be more thoroughly covered during the interpretation of the water data.

Solid samples

All whole rock and mineral samples were obtained from rotary drill cuttings. Drill cuttings are particularly inadequate for studying Roosevelt Hot Springs because of the fracture-controlled permeability of the system. Since the cuttings sampled intervals over several meters, the cuttings represent a mixture of minerals and rocks from undeterminable proximities to major fractures as well as occasional mixtures of different lithologies, reducing the sensitivity of the data to specific features in the drill holes. This mixing is particularly apparent in the variable alteration intensities of chips from the same sample interval. Furthermore, veinlet relationships are obscured, restricting the interpretation of mineral parageneses and virtually eliminating the identification of any possible pre-geothermal alteration products. Since the cuttings often fail to preserve macroscopic rock textures, differentiating between gneiss and igneous rocks can be difficult when metamorphic index minerals, such as sillimanite, are absent.

Mineral separations of quartz, feldspar, biotite, chlorite, and veinlet quartz were achieved primarily by hand-picking, augmented by standard heavy liquid and magnetic separation techniques. K-mica clays for oxygen analysis were centrifuged to obtain the less than five micron size fraction. Quartz separates were purified by treatment with cold HF, while carbonate impurities in feldspars and clays were removed with cold, dilute HCl. Carbonates for carbon and oxygen analysis were not separated from the silicate fraction.

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 removing 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 in a method similar to Suzuoki and Epstein (1976). Hydrogen was converted to water in a CuO furnace at 700°C prior to reduction over hot uranium metal.

Silicate oxygen extraction. Oxygen from silicate minerals was extracted by reacting 10 to 15 mg of whole rock, biotite, chlorite, quartz, and K-feldspar, and 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).

Carbonate carbon and oxygen extraction. Powdered whole rock and +80 to +150 mesh coarse rock weighing between 0.2 and 1.0 gram were treated with distilled phosphoric acid at 25°C according to the procedure of McCrea (1950). The reactions progressed for 12 to 15 hours before extraction of the CO_2 gas.

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,

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

where α is the fractionation factor, defined as

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

ANALYTICAL RESULTS

Waters

Spring waters from several geographical regions near Roosevelt Hot Springs and thermal fluids sampled during flow tests of production wells were collected and analyzed by the USGS (A. H. Truesdell, personal communication) and the University of Utah. Spring localities are provided in plate 1 and the hydrogen and oxygen isotopic data is compiled in table 2.

Spring sampling was concentrated within the Mineral Mountains, since this range represents the most likely source area for the recharge waters. Meteoric waters were also collected from springs in the Tushar Mountains and the Pavant Range and from springs and wells west of the Mineral Mountains.

An unusual feature of the meteoric waters from each geographical area is their consistent deviation from the meteoric water line, as shown in figure 2. Fitting a line of linear regression to the Mineral Mountains analyses provides a slope of 4.6, suggesting that these meteoric waters have been modified by kinetic effects accompanying evaporation (Craig, 1961b; Craig et al., 1963) or perhaps by 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.

Table 2. Isotopic analyses of waters from the Roosevelt Hot Springs Region.

	Date Collected	δD	$\delta^{18}O$
Mineral Mtns. Springs			
Ranch Canyon (Kirk)	2/76*	-113	-15.0
	8/76*	-110	-14.2
Bailey	10/6/77	-112	-14.6
	2/76*	-116	-15.6
	8/76*	-111	-14.6
Willow	10/2/77	-111	-15.0
	8/76*	-114	-14.8
	10/2/77	-112	-14.6
Cherry Creek	8/76*	-113	-14.7
Griffith	8/76*	-110	-14.4
Antelope	8/76*	-116	-15.1
North	8/76*	-112	-14.7
Rock Corral	8/76*	-109	-14.2
Mud	8/76*	-113	-14.8
Mathew	10/6/77	-117	-15.6
Jack Rabbit	10/6/77	-116	-15.8
Tushar Mtns. Springs			
Sulfurdale North	8/76*	-118	-15.6
Sulfurdale South	8/76*	-116	-15.6
Dead Cow	8/76*	-114	-14.6
Mud	8/76*	-126	-16.5
Springs between Tushar and Mineral Mtns.			
Four Mile	8/76*	-114	-14.2
Wiregrass	8/76*	-115	-15.2
Cowboy	8/76*	-118	-15.2
Pavant Range Springs			
North Creek	8/76*	-120	-15.0
Magpie	8/76*	-120	-15.6

Table 2. Continued

		Date Collected	δD	$\delta^{18}O$
West of Mineral Mountains				
Milford City Well		2/76*	-117	-15.5
Pearson Well		2/76*	-122	-15.5
South of Milford Airport		8/76*	-124	-15.8
Beaver Lake Mine Spring		8/76*	-104	-12.5
Armstrong Spring		8/76*	-107	-13.7
Thermal Waters				
Utah State 14-2	water	11/76*	-	-13.4
	steam	11/76*	-	-15.9
	water	5/12/78	-116	-13.6
	steam	5/12/78	-117	-14.8
	water	5/13/78	-116	-13.6
	steam	5/13/78	-114	-15.4
Seep		10/1/77	-110	-12.5
Seep		5/11/78	-111	-12.7

* Analyses by the USGS, Menlo Park, under the direction of Dr. A. H. Truesdell.

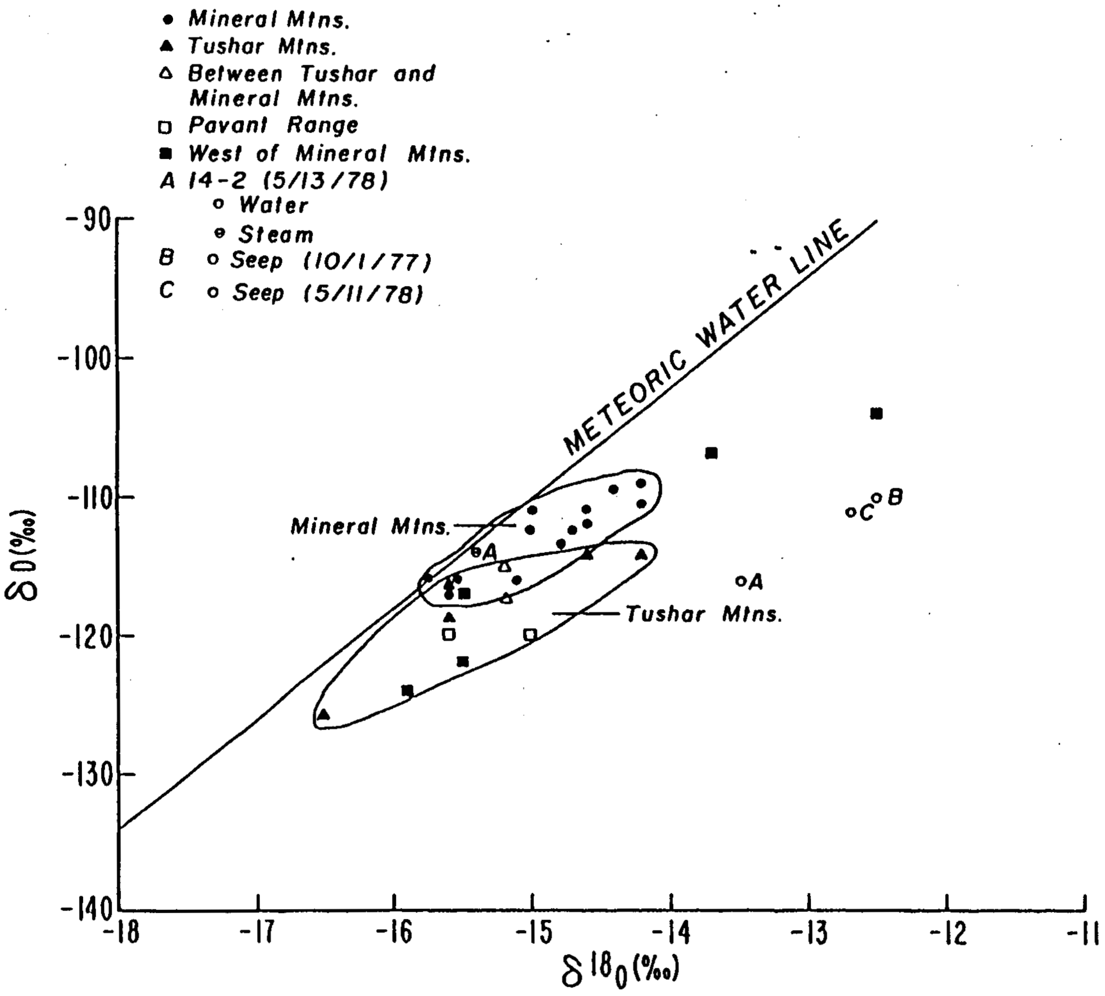

Figure 2. A plot of δD versus $\delta^{18}O$ for cold spring and thermal waters from the vicinity of the Roosevelt Hot Springs thermal area.

The relationship of the thermal fluids to the spring waters is also shown in figure 2. The δD value of the thermal waters, -116, is within the range exhibited by springs from the Mineral and Tushar Mountains. 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 $\delta^{18}O$ and δD values of the thermal waters from the seep are about -12.6 and -111, respectively. The enrichment of ^{18}O and D in the seep suggests modification of the deep thermal waters by evaporation at the surface or boiling in the subsurface, although the decrease in the concentration of Cl^- in the seep (table 1) contradicts evaporation or boiling.

The small oxygen and hydrogen fractionations between steam and water from well 14-2 (5/13/78) provide steam separation temperatures between 200 and 250°C (Friedman and O'Neil, 1977), consistent with the measured temperatures during these well tests. An additional steam condensate sample collected by the University of Utah from well 14-2 (5/12/78) is considerably enriched in ^{18}O compared to steam collected from this well by the USGS on an earlier date. The good agreement between the USGS and University of Utah measurements on the other thermal and spring waters suggests that this apparent ^{18}O enrichment resulted from contamination of the steam by the water phase due to improper adjustments of the miniseparator.

As will become evident from the discussion of the isotopic data for rocks and minerals, the $\delta^{18}O$ value of -13.7 represents only the fluids from production zones in well 14-2 and probably has little bearing on the oxygen isotopic compositions of the thermal fluids in other parts of the geothermal field, particularly for water enclosed in rocks of low permeability. Since the $\delta^{18}O$ values of the seep and the thermal fluids of well 14-2 encompass a probable range of isotopic compositions for flowing thermal waters, a $\delta^{18}O$ value of -13.0 has been selected to represent an average thermal water until additional data for wells 9-1 and 72-16 become available.

Silicates

The oxygen isotopic shift observed in the thermal waters requires a depletion of ^{18}O in the host rock. To evaluate the extent of this depletion, silicate minerals and whole rock samples were analyzed from well 14-2. To avoid complications from potential pre-geothermal alteration in the metamorphic rocks, sampling was concentrated within igneous rocks near the predominant hot water entry zone at 870 m. These altered rocks are compared to a relatively fresh quartz monzonite from a shallow alteration hole, UU 76-1. Table 3 lists the isotopic data with descriptions of the lithology and alteration of each sample.

The fresh quartz monzonite provides $\delta^{18}O$ values of +7.4, +10.2,

Table 3. $\delta^{18}O$ and δD of silicates from well Utah State 14-2 with descriptions of lithology* and alteration* for samples analyzed isotopically.

gn, gneiss
gr, granite
md, microdiorite
qm, qtz, monzonite

Ca, calcite; Ch, chlorite; Cl clay; Ep, epidote; Gy, gypsum; Hm, hematite; Kf, K-feldspar; Km, K-mica; Mo, montmorillonite; Mt, magnetite; Py, pyrite; Q, quartz.

Depth (meters)	Lithology	Alteration Intensity	Alteration Products	Veinlets
259.1	qm	weak	Ca-Km-Mo-Kf-Hm-Ep	Q-Ca-Mo-Mt
487.7	qm-gr	moderate	Ca-Km-Hm-Py	Ca, Q
609.6	gr	strong	Ca-Km-Hm	Ca
670.6	qm	weak	Ca-Km-Kf-Ch-Py	Cl-Q-Mt
853.4	qm	weak	Ca-Km-Kh-Ch-Py	
868.7	md	strong	Ca-Km-Kf-Ch-Py	Q-Kf-Ch, Ca
877.8	md	strong	Ca-Km-Kf-Ch-Py	Q-Ca-Ch
914.4	qm	moderate	Ca-Km-Kf-Ch-Py	Kf
944.9	qm	weak	Ca-Km-Kf-Ch-Py	Q
1432.6	gn	weak	Ca-Km-Kf-Ch-Py	Cl or Gy
1585.0	gn	strong	Ca-Km-Kf-Ch-Py	Ca
UU 76-1 (56.1m)	qm	v. weak	Ca-Km	

* Lithology and alteration descriptions for 14-2 were adapted from Ballantyne and Parry (1978) and Glenn and Hulen (1979).

Table 3. Continued

	$\delta^{18}O$							δD
Depth Meters	Quartz	Vein Quartz	K-feldspar	Biotite	Chlorite	Whole Rock	K-mica	K-mica
259.1			7.9				8.5	-124
487.7			8.0				6.4	-133
609.6	8.2		5.8			6.2	6.3	-134
670.6	8.6		7.2	3.7		6.8	7.9	-135
853.4	10.4		6.1	3.5	-3.6	4.0	5.7	-141
863.7	8.7	-2.8	4.5	0.9	-5.1	0.4	3.1	-142
877.8		-2.5						
914.4	10.6		3.8	2.4	-5.4	1.2	2.4	-139
944.9	10.0		5.6	3.0		2.6	3.2	-138
1432.6						3.5		
1585.0						-1.4		
UU 76-1	10.2		7.5	4.2		7.4		

and +7.5 for whole rock, quartz, and K-feldspar, respectively, which are typical for normal granitic rocks (Taylor, 1978). However, for fresh granitic rocks Δ^{u}quartz - alkali feldspar should be within 1.0 to 1.5 (Taylor, 1978). Since the actual difference between quartz and feldspar is 2.7, the feldspar from the fresh quartz monzonite has probably experienced minor depletion or post-solidus isotopic exchange.

$\delta^{18}O$ values for whole rock from well 14-2 range from -1.4 to +6.8, with the lowest value obtained from the more highly altered rocks. This relationship between the extent of depletion and alteration intensity is apparent in biotite and K-feldspar as well, where biotite varies from 0.9 to 3.7 and K-feldspar from 3.8 to 8.0. In the upper portion of the hole K-feldspar retains a more normal $\delta^{18}O$ value probably because of the lower temperature at which the rock interacts with the water. Because quartz is known to be quite resistant to isotopic exchange (Clayton et al., 1968; Clayton and Steiner, 1975), the lower $\delta^{18}O$ values of quartz in the upper part of well 14-2 (between 8.2 and 8.7) could reflect changes in lithology at 609.6 and 868.7 m and possibly contamination at 670.6 m rather than interaction with thermal fluid.

Secondary silicate minerals include veinlet quartz, chlorite, and K-mica clays. Fine grained, euhedral quartz, which was recovered from veinlets at 868.7 and 877.8 m, provides $\delta^{18}O$ values of -2.8 and -2.5. Chlorite, replacing biotite between 853.4 and 914.4 m, yielded $\delta^{18}O$ measurements between -5.4 and -3.6. K-mica clays, primarily alteration products of plagioclase, were analyzed for both oxygen and hydrogen between 259.1 and 944.8 m. $\delta^{18}O$ values of clay vary from +2.4 to +8.5, while δD values range from -142 to -124.

Whole rock $\delta^{18}O$ measurements are generally less than the $\delta^{18}O$ value that combined quartz, K-feldspar, biotite, and K-mica clay would provide. Biases introduced during the separation of K-feldspar probably account for this discrepancy. Because a range of alteration intensities is represented by the drill cuttings for any given sample, K-feldspar from the least altered rocks was probably preferentially separated during the hand-picking mineral separation stage.

Carbonates

Calcite is a common, although not abundant, secondary mineral at Roosevelt Hot Springs, occurring both in veinlets and as a replacement product of plagioclase and ferromagnesium minerals. The oxygen and carbon isotopes from calcite were analyzed for both powdered and "coarse" (+80 to +150 mesh) whole rock fractions from throughout the lengths of wells 14-2, 72-16, and 9-1. The isotopic data, alteration intensity, and the measured CO_2 yields are listed in table 4. Also presented in table 4 are isotopic temperatures calculated from the calcite-water fractionation curve of Friedman and O'Neil (1977) and a uniform water composition of -13.0.

Only calcite from drill hole 14-2 above 870 m provides a trend of $\delta^{18}O$ that is consistent for equilibration with a uniform water under a normal geothermal gradient. Below 870 m in 14-2 and throughout 9-1 and 72-16 there is no apparent trend with either depth or temperature. There is a good correlation of $\delta^{18}O$ with alteration intensity, where

Table 4. Carbonate oxygen and carbon isotopes.

Well Utah State 14-2

Depth (meters)	Alteration Intensity	Whole Rock Texture(1)	CO_2 Yield (umoles/g rock)	$\delta^{18}O$ (‰)	$\delta^{13}C$ (‰)	Isotopic T,°C (2)
259.1	weak	powder	12	+11.2	-4.0	47
277.4	moderate	powder	165	+11.0	-2.7	48
		coarse	56	+5.8	-4.7	85
286.5	moderate	powder	93	+11.3	-4.0	47
350.5	moderate	powder	28	+8.3	-3.2	66
		coarse	7	+3.9	-3.6	102
472.4	weak	powder	30	+4.8	-3.4	93
		coarse	29	+1.9	-3.8	122
487.7	moderate	powder	20	+4.5	-3.7	96
609.6	strong	powder	20	+3.3	-3.8	107
		coarse	14	+2.2	-6.4	119
670.6	weak	powder	56	+6.8	-4.0	77
		coarse	30	+4.1	-5.3	100
746.8	weak	powder	122	+0.4	-3.9	140
		coarse	93	-1.9	-4.3	173
853.4	weak	powder	111	0.0	-4.2	145
		coarse	80	-2.4	-4.6	181
868.7	strong	powder	156	-3.7	-5.6	204
		coarse	110	-5.4	-5.8	242
870.2	strong	powder	255	-4.3	-5.7	217
877.8	moderate	powder	388	-2.2	-5.3	177
		coarse	494	-0.7	-5.2	155
880.9	moderate	powder	217	-3.9	-5.8	208
888.5	moderate	powder	122	-3.2	-5.8	195
		coarse	98	-5.6	-5.8	247

Table 4. Continued.

Depth (meters)	Alteration Intensity	Whole Rock Texture(1)	CO_2 Yield (umoles/g rock)	$\delta^{18}O$ (‰)	$\delta^{13}C$ (‰)	Isotopic T,°C (2)
899.2	weak	powder	144	-1.6	-3.6	168
914.4	moderate	powder	131	+2.2	-4.5	177
		coarse	124	-4.5	-4.8	221
944.9	weak	powder	170	-1.5	-4.3	166
1005.8	moderate	powder	129	-3.4	-5.6	199
1066.8	weak	powder	77	-1.6	-6.0	168
1097.3	moderate	powder	58	-2.9	-5.4	189
1143.0	moderate	powder	65	-2.2	-5.8	177
1204.0	moderate	powder	108	-2.0	-4.7	174
1249.7	weak	powder	148	+1.6	-4.5	125
1295.4	weak	powder	58	-4.6	-6.6	223
1356.4	moderate	powder	629	-4.1	-4.7	212
1432.6	weak	powder	62	+2.1	-4.9	120
1463.0	weak	powder	58	+0.4	-5.3	140
1539.2	weak	powder	39	+1.3	-5.9	129
1585.0	strong	powder	657	+5.4	-5.8	242
		coarse	917	-6.3	-6.0	265
		Well Utah State 72-16				
167.6	strong	powder	146	-0.9	-5.4	157
182.9	strong	powder	267	-1.6	-4.9	168
198.1	weak	powder	164	-0.2	-5.4	148

Table 4. Continued.

Depth (meters)	Alteration Intensity	Whole Rock Texture(1)	CO_2 Yield (umoles/g rock)	$\delta^{18}O$ (‰)	$\delta^{13}C$ (‰)	Isotopic T,°C (2)
228.6	weak	powder	263	-1.8	-5.1	171
243.8	weak	powder	151	-0.5	-5.0	152
259.1	moderate	powder	153	+0.3	-4.4	141
274.3	moderate	powder	168	-1.8	-5.2	171
289.6	moderate	powder	146	-1.0	-4.9	159
304.8	weak	powder	91	+1.6	-4.6	125
320.0	weak	powder	100	-0.7	-4.9	155
335.3	weak	powder	129	-0.9	-5.0	157
350.5	moderate	powder	51	-0.5	-5.1	152
365.8	weak	powder	81	+0.6	-4.6	137
Well Utah State 9-1						
240.8	moderate	powder	28	+6.7	-3.6	78
280.4	weak	powder	59	+9.3	-2.4	59
364.2	moderate	powder	113	+0.4	-3.1	140
378.0	moderate	powder	376	-0.4	-1.8	140
451.1	moderate	powder	306	+0.8	-0.7	135
		coarse	246	-0.3	-0.9	149
496.8	moderate	powder	37	+1.8	-3.5	123
		coarse	99	+0.7	-2.6	136
519.7	moderate	powder	95	+2.2	-3.8	119
597.4	strong	powder	222	+0.7	-3.3	136
		coarse	258	-0.7	-3.2	155
615.7	moderate	powder	41	+2.6	-4.7	115

Table 4. Continued.

Depth (meters)	Alteration Intensity	Whole Rock Texture(1)	CO_2 Yield (umoles/g rock)	$\delta^{18}O$ (‰)	$\delta^{13}C$ (‰)	Isotopic T,°C (2)
620.3	moderate	powder	118	+2.5	-4.1	116
737.6	moderate	powder	93	+4.7	-5.0	94
838.2	strong	powder	206	-0.3	-4.5	149
		coarse	167	-1.1	-4.3	160
865.6	weak	powder	55	+5.2	-5.5	90
1005.8	moderate	powder	157	+1.3	-4.1	129
1066.8	weak	powder	77	+7.0	-3.4	75
1188.7	weak	powder	96	+3.5	-3.5	105
1295.4	weak	powder	34	+7.3	-7.3	73
1453.9	strong	powder	138	-0.4	-5.0	150
		coarse	126	-1.0	-4.8	159
1508.8	weak	powder	67	+7.2	-3.8	74
1737.4	weak	powder	50	+6.6	-5.4	78
1767.8	moderate	powder	256	-0.2	-5.1	148
1798.3	weak	powder	52	+5.3	-4.5	89
1898.9	weak	powder	18	+7.6	-6.5	71
1917.2	weak	powder	60	+5.7	-5.4	86
1944.6	weak	powder	17	+9.7	-8.2	56
2030.0	moderate	powder	147	+4.8	-4.0	93

(1) Powder refers to pulverized whole rock; coarse pertains to the whole rock fraction between +80 to +150 mesh.

(2) Calculated with $\delta^{18}O$ of water = -13.0% and the calcite-water fractionation curve of Friedman and O'Neil (1977).

^{18}O is enriched in carbonates from weakly altered rocks.

The $\delta^{18}O$ values of calcite range from -0.7 to -8.2, which corresponds to the spread observed in other geothermal systems (Clayton et al., 1968; Clayton and Steiner, 1975; Blattner, 1975). Carbonates from well 72-16 provide a fairly uniform $\delta^{13}C$ value near -5.0 while carbonates from 14-2 and 9-1 show an overall decrease in $\delta^{13}C$ with increasing depth, typically from -2.4 to -6.0.

The coarse whole rock carbonate differs from the powdered in two significant ways: 1) CO_2 yields from the coarse rock are usually less than the powdered rock; and 2) oxygen isotopic values are generally lower in the coarse fraction. Normally there is only a small difference in $\delta^{13}C$ between the coarse and powdered rock.

DISCUSSION

Origin of the Thermal Waters

By discovering that the water in hot spring systems is of meteoric origin, Craig (1963) provided a valuable tool for interpreting the source and evolution of the thermal water as it passes through the country rock. Typically, meteoric waters are out of equilibrium with the silicate or carbonate host rock, which results in oxygen isotopic shifts of both the water and the rock through isotopic exchange. Because the hydrogen reservoir in the water overwhelms the exchangeable hydrogen in the country rock, hydrogen isotopes often preserve the signature of the source area (Craig, 1963). By comparing the δD values of the thermal waters with meteoric waters from potential source areas, the areal extent of the recharge aquifer can be better defined.

From the relationship of the thermal waters to the meteoric waters as illustrated in figure 2, both the regions west of the Mineral Mountains and the Pavant Range can be eliminated as likely source areas. The overlap in spring waters from the Mineral Mountains and the Tushar Mountains coincides with the δD values of the thermal waters, indicating that isotopic analyses alone cannot eliminate the Tushar Mountains as a potential recharge area.

The Tushar Mountains, however, are up to 750 m higher than the Mineral Mountains. This difference in elevation is reflected in the average δD values for spring waters from the two ranges. The average δD value for springs in the Mineral Mountains is -113, while the average value for the four springs from the Tushar Mountains is -118. The only springs sampled in the Tushar Mountains occur at low elevations. Future spring sampling in the higher elevations should reveal that average Tushar meteoric water is skewed towards δD values less than -118. Thus, the δD of the thermal waters, between -115 and -116, probably implies a dominant recharge area in the upper elevations of the Mineral Mountains (as suggested by the δD value of -116 for Jack Rabbit spring) and restricts contributions from the Tushar Mountains to small quantities. Depressed heat flow values in the Mineral Mountains (Ward et al., 1978) and their proximity to the thermal area supports the Mineral Mountains as the dominant recharge area. Also, modeling of the regional hydrologic system suggests that it is unlikely that water recharging in the Tushar Mountains will be discharged in Milford Valley by interbasin flow (L. Smith, personal communication).

Water/Rock Ratios

The oxygen isotopic shift in thermal waters depends upon several parameters, notably temperature, the extent of disequilibrium between the unaltered rock and the fresh meteoric water, and the proportions of water and rock in the system. Therefore, proper evaluation of the isotopic shifts at Roosevelt Hot Springs requires comparison with the corresponding shifts in the host rock by the calculation of water/rock (W/R) atomic oxygen ratios.

The amount of water involved in the depletion of the whole rock oxygen reservoir 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 W/R mass ratio can be obtained. At Roosevelt Hot Springs an average fresh or altered rock contains approximately 29 moles of exchangeable oxygen per kilogram 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 W/R mass ratio.

Because of the uncertainty in the isotopic shifts of the thermal waters, W/R ratios have been calculated for a minimum isotopic shift of +1.2 and a larger shift of +2.6. The data and results for these calculations are presented in table 5. For the most altered rocks in the production zones, W/R mass ratios vary from 1.4 to 3.0, depending upon the isotopic shift of the waters, while the least altered rocks provide W/R mass ratios between 0.1 and 0.3. Average W/R mass ratios for the Salton Sea and Wairakei have been estimated as 0.45 and 4.3,

Table 5. Data and results for water/rock ratio calculations.

$\delta^{18}O_r^f$ = Altered rock from well 14-2.

$\delta^{18}O_r^i$ = 7.4 ("Fresh" quartz monzonite from UU76-1)

A. $\delta^{18}O_w^i$ = -14.9 (Average Mineral Mtns. spring water)

$\delta^{18}O_w^f$ = -13.7 (Thermal water from well 14-2)

B. $\delta^{18}O_w^i$ = -15.6 (Spring water from the higher elevations of the Mineral Mtns.)

$\delta^{18}O_w^{f'}$ = -13.0 (Proposed average thermal water for the geothermal system)

Depth (meters)	$\delta^{18}O_r^f$	W/R atom ratio A	W/R atom ratio B	W/R mass ratio A	W/R mass ratio B
609.6	6.2	1.0	0.5	0.5	0.3
670.6	6.8	0.5	0.2	0.3	0.1
853.4	4.0	2.8	1.3	1.5	0.7
868.7	0.4	5.8	2.7	3.0	1.4
914.4	1.2	5.2	2.4	2.7	1.3
944.9	2.6	4.0	1.8	2.1	0.9

respectively (Clayton et al., 1968; Clayton and Steiner, 1975), showing that within the production zones the W/R mass ratios correspond to those in other geothermal systems.

There are additional uncertainties in these calculations, however. The mixing of drill cuttings possessing variable alteration intensities for any given sample interval prohibits calculations of maximum or minimum W/R ratios. Compounding this problem is the occurrence of different lithologies from the intervals analyzed in well 14-2. Finally, for the weakly altered rocks, isotopic shifts may exceed +2.6 for the thermal waters.

Nevertheless, the variability of W/R ratios at Roosevelt Hot Springs has important implications for the geothermal system. Moderately to strongly altered rocks possess high W/R 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 W/R ratios are characteristics of less permeable rocks. Since permeability is fracture-controlled within the deep reservoir, alteration intensity and W/R ratios emphasize the importance of fractures and fissures in directing fluid movement. However, W/R ratios cannot distinguish between present and paleo conduits.

The extent of isotopic exchange

By calculating a composition for a rock which has completely equilibrated with the thermal water, Clayton and Steiner (1975) roughly estimated the extent of exchange experienced by an altered rock. At Wairakei, they noted that the extent of exchange generally exceeded 50 percent and approached completion in some cases. To estimate the composition of rock in equilibrium with the water, they applied the fractionation between K-feldspar and water as an approximation for the fractionation between rock and water (after Taylor, 1968). An application of this procedure to the whole rock and thermal water data from Ohaki-Broadlands provided by Eslinger and Savin (1973) suggests that the rocks from borehole Br 16 also approach complete equilibration with the thermal water.

Before applying these calculations to Roosevelt Hot Springs, it should be noted that Eslinger and Savin (1973) have proposed that thermal waters may not be uniform throughout the entire geothermal system. They suggest that waters may experience greater isotopic shifts when percolating slowly through the groundmass of the rock than when flowing in open pathways such as fractures. Thus, to evaluate the extent of exchange between the thermal water sampled in the wells and the rock from Roosevelt Hot Springs, only K-feldspar from the proximity of the production zones can be utilized.

At 225°C, the approximate temperature of the production zone near 870 m in well 14-2, K-feldspar in equilibrium with a thermal water possessing a $\delta^{18}O$ value of -13.7 would have a $\delta^{18}O$ value of -5.4 (O'Neil and Taylor, 1967). The most depleted K-feldspar possesses a $\delta^{18}O$ value of +3.8 at 914.9 m, compared to an initial $\delta^{18}O$ value of +7.5. Thus, the extent of exchange for K-feldspar near the production zone is only about 30 percent, which is considerably less than the exchange exhibited by altered rocks in the geothermal areas of New

Zealand. A 30 percent extent of exchange cannot represent a maximum value, however, since there is a mixture of rocks which have experienced variable degrees of alteration even within the production zone.

K-mica Clays

Oxygen isotopic compositions

K-mica clays, identified petrographically as sericite, occur primarily as alteration products of plagioclase. X-ray diffraction studies indicate that K-mica is the predominant clay mineral, although significant quantities of mixed layer K-mica - montmorillonite occur at 259.1 and 670.6 m (Parry, 1978). Isotopic measurements on oxygen and hydrogen provide the opportunity to test for equilibration of the clay with a uniform water composition under a normal geothermal gradient.

O'Neil and Taylor (1969) and Eslinger and Savin (1973) demonstrate that K-mica clays become progressively enriched with ^{18}O as temperatures decrease. As figure 3 illustrates, K-mica clays from Roosevelt Hot Springs do not display this trend. Instead, $\delta^{18}O$ values for clay follow the trends established by whole rock and K-feldspar. Furthermore, a K-mica clay equilibrating with a water with a $\delta^{18}O$ value of -13.7 at temperatures of the hot water entry zone at 870 m (approximately 225°C) should possess a $\delta^{18}O$ value of about -8.0 (from extrapolation of the muscovite - water fractionation curve in O'Neil and Taylor, 1969). Clearly, the clays possess oxygen isotopic compositions which are not in equilibrium with this thermal water.

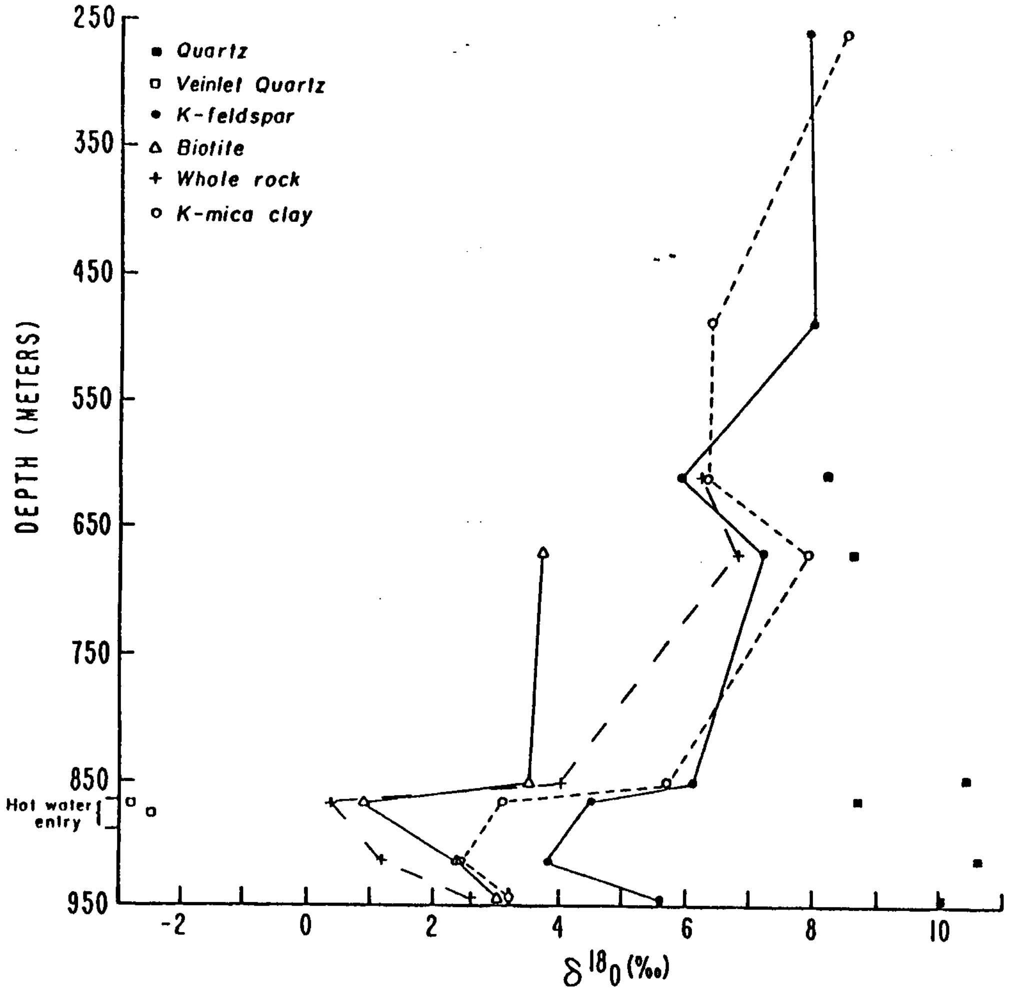

Figure 3. A plot of $\delta^{18}O$ versus depth for silicate minerals from well Utah State 14-2.

The oxygen isotope data suggests that the K-mica is reflecting the isotopic composition of the feldspar rather than equilibrating with the thermal water extracted from production zones. The feldspar can influence the K-mica in two ways: 1) either the clay is inheriting structurally intact polyhedral units from the plagioclase during alteration, or 2) oxygen isotopic exchange controls the $\delta^{18}O$ of the water in a microenvironment surrounding the plagioclase grains as a function of low W/R mass ratios.

Structural inheritance of intact units from a precursor has only been demonstrated for the conversion of kaolinite to pyrophyllite, two minerals of similar structure and chemistry (O'Neil and Kharaka, 1976). In addition, the simple conversion of albite to K-feldspar involves a solution - redeposition step (O'Neil and Taylor, 1967). Thus, structural inheritance seems unlikely, and oxygen isotopic exchange between plagioclase and water is the favored process for controlling the $\delta^{18}O$ values of the K-mica.

The validity of isotopic exchange can be tested by assuming that the system plagicolase - K-mica - water has achieved isotopic equilibrium during the alteration event. By combining the muscovite - water fractionation curve (O'Neil and Taylor, 1969) with the plagioclase - water curve (O'Neil and Taylor,1967) and extrapolating to 225°C, the isotopic composition of plagioclase within the hot water entry zone at 870 m can be estimated from the $\delta^{18}O$ value of K-mica. The plagioclase in well 14-2 contains anorthite contents from 26 to 2.5 percent (Ballantyne, 1978), providing a $\Delta°$ plagioclase - K-mica between 2.5 and 1.7 at 225°C. At 868.7 m, where $\delta^{18}O$ (K-mica) equals

3.1, the maximum $\delta^{18}O$ (plag) would be 5.6, compared to a measured value of 4.5 for the K-feldspar from this depth. This reasonable agreement between K-feldspar and plagioclase $\delta^{18}O$ values supports isotopic exchange at conditions of low W/R mass ratio as a valid process in the vicinity of the production zones.

At the shallower depths of 259.1 and 670.6 m, the $\delta^{18}O$ values of K-mica are 8.5 and 7.9, respectively. Since the fractionation between plagioclase and K-mica is positive at all temperatures, the isotopic exchange process requires feldspar isotopic compositions exceeding 8 permil, which is contrary to the feldspar isotopic data from well 14-2. Thus, isotopic exchange cannot completely account for the $\delta^{18}O$ values of K-mica at these depths. However, interstratified clays become quantitatively important at 259.1 and 670.6 m, suggesting that the apparent ^{18}O enrichment in the clays at these depths might be attributed to compositional changes in the clay fraction.

The K-mica data indicates that the rocks are not interacting with the bulk of the thermal water, otherwise a much larger isotopic shift would be observed in the thermal waters extracted from the production wells. This feature of the data leads to two possibilities. First, the thermal waters may be circulating rather rapidly through the geothermal system. Second, only a small proportion of the thermal waters infiltrates from the open fractures into the surrounding impermeable rocks, accounting for both the variable W/R ratios and the relatively small extent of exchange experienced by rocks even within the production zones.

Hydrogen isotopic compositions

The δD values of the K-mica clays as a function of depth are presented in table 3. Before these δD values can be interpreted, a fractionation curve between K-mica and water must first be estimated for the geothermal system at Roosevelt Hot Springs for two reasons: 1) D/H ratios of hydroxyl bearing minerals are dependent upon the molar fraction Al, Fe, and Mg occupying the octahedral sites (Suzuoki and Epstein, 1976); and 2) there is little experimental or observational data for the system K-mica - water below 450°C.

Suzuoki and Epstein (1976) provide an experimental calibration curve for muscovite - water between 850 and 450°C, but caution against extrapolating to lower temperatures. To approximate the hydrogen fractionation between K-mica and water at surface temperatures, the observed hydrogen fractionation between montmorillonite and water has been chosen. However, three distinct values have been proposed for α^H mont. - H_2O:

1) 0.985 for montmorillonite from deep sea radiolarian ooze by Yeh and Epstein (1978);

2) 0.970 for montmorillonite in Quaternary soils from igneous parent rocks by Lawrence and Taylor (1971); and

3) 0.940 for montmorillonite in oceanic sediments by Savin and Epstein (1970).

These points are labeled B, C, and D, respectively, on figure 4. Temperatures were selected as 0°C at the ocean bottoms and 0 to 30°C in soils.

The curves drawn between these points and point A on figure 4

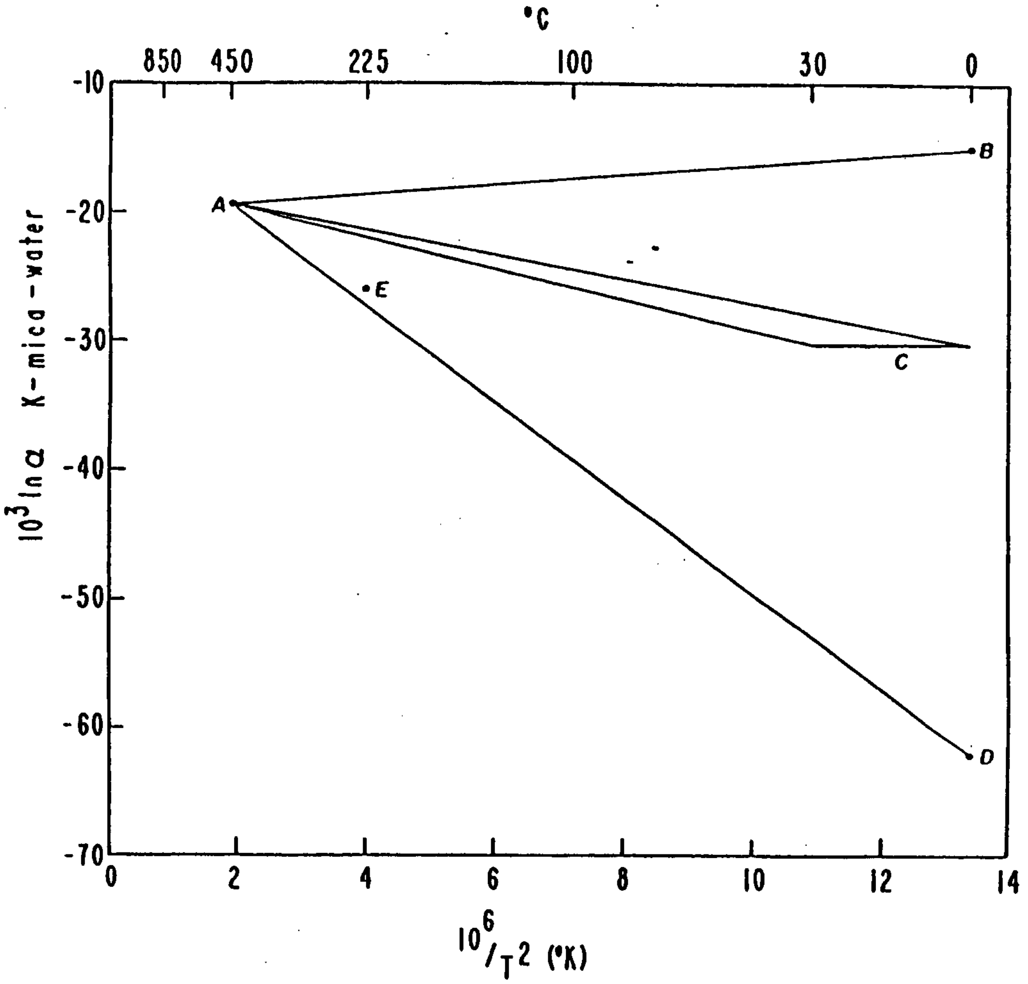

Figure 4. Possible curves describing the hydrogen isotopic fractionation between K-mica and water. (see the text for an explanation of points A - E.)

provide the possible fractionation curves between K-mica and water at Roosevelt Hot Springs. Point A was determined at 450°C by the application of the expression

$$10^3 \ln \alpha = -22.4\,(10^6)T^{-2} + 28.2 +$$
$$(2X_{Al} - 4X_{Mg} - 68\,X_{Fe})$$

from Suzuoki and Epstein (1976), where X is the molar fraction of cations from the octahedral sites. At Roosevelt Hot Springs, the K-mica clays are phengitic, with molar fractions of Al, Mg, and Fe averaging 0.83, 0.09, and 0.08, respectively (Ballantyne, 1978). A clay of this composition at 450°C would provide a fractionation between K-mica and water equal to -19.4, which is plotted as point A.

Point E on figure 4 represents the observed hydrogen fractionation of -26 calculated for a K-mica from 868.7 m (-142) and the thermal water sampled from well 14-2 (-116) at 225°C. Since point E plots within 1.5 of curve AD, this curve has been chosen as possibly representing the fractionation between clay and water at Roosevelt Hot Springs.

By applying curve AD to the hydrogen isotopic compositions of K-mica, the hydrogen isotopic composition of the water in equilibrium with the clay can be estimated. Temperatures for equilibration have been determined by extrapolating between the medians of fluid inclusion temperatures at 286.5, 487.7, and 899.2 m (W. T. Parry, personal communication). The hydrogen isotopic compositions of the clay, calculated waters, and temperature are provided in table 6.

Between 853.4 and 944.9 m the calculated waters are in good agreement with the measured δD of the thermal waters. However, above

Table 6. The δD of K-mica, the calculated δD of equilibrated thermal water, and the percent evaporation in well 14-2.

Depth (meters)	δD of K-mica (‰)	Calc. δD of water (‰) (1)	Δ water-vapor (2)	%evaporation (3)	Temperature°C (4)
259.1	-124	-86	25	74	106
487.7	-133	-102	12	73	168
609.6	-134	-104	4	97	188
670.6	-135	-106	3	98	200
853.4	-141	-113	-1	-	228
868.7	-142	-115	-1	-	230
914.4	-139	-112	-2	-	236
944.9	-138	-112	-2	-	238

(1) Estimated from δD of K-mica and fractionation curve AD of figure .

(2) Friedman and O'Neil, 1977.

(3) Assuming Rayleigh distillation (Broecker and Oversby, 1971).

(4) Extrapolated from the medians of fluid inclusion temperatures from the following depths (Parry, personal communication):

Depth (meters)	Range °C
277.4	111-117
490.7	155-186
888.5	223-236

850 m the calculated waters become enriched with D, indicating that the thermal waters have become modified or that the clays no longer fall along fractionation curve AD of figure 4.

The only likely process which would enrich the waters with D is boiling. The percentage of steam removed from an initial water (δD equals -116) to produce the calculated waters can be determined by assuming boiling by Rayleigh distillation (Broecker and Oversby, 1971) and applying the fractionation factors between vapor and water given in Friedman and O'Neil (1977). The results of these calculations are also provided in table 6. Evaporation of over 75 percent of the thermal water seems excessive, but it could account for the considerable D enrichment of the clay and the calculated water below 850 m. In addition, the ^{18}O enrichment in the clays could also be partially explained by boiling. Further discussions concerning evaporation are presented during the interpretation of the carbonate data.

Since the clays above 259.1 m yield calculated waters having a δD value within 15 permil of the measured thermal waters, another explanation for the apparent D enrichment of the clays is a change in the clay - water fractionation curve below 200°C. This could arise by an increase of Al and a decrease of Fe in the octahedral sites of the clay minerals. For 259.1 m, interstratified clays constitute up to 50 percent of the clay fraction (Parry, 1978), indicating that compositional changes of the clay might also account for the D enrichment at this depth.

An alternative explanation for the apparent enrichment of D in the clay above 850 m is the possibility of a cross over below 200°C in the K-mica - water fractionation curve. Curve AB in figure 4 is consistent with a cross over, although more data is certainly required to substantiate this possibility.

Veinlet Quartz

Although Blattner (1975) demonstrated the importance of veinlet minerals for interpreting the hydrology and evolution of a thermal system, sufficient veinlet quartz could be obtained from only two intervals within the hot water entry zone. $\delta^{18}O$ values of -2.5 and -2.8, at 868.7 and 877.8 m, respectively, provide calculated waters of -13.2 and -13.5 at 225°C (Friedman and O'Neil, 1977), indicating that the veinlet minerals have precipitated in equilibrium with the thermal water in the production zone.

Oxygen Isotopic Composition of Calcite

Powdered versus coarse rock

A comparison between the isotopic compositions of calcite from powdered and coarse rock, presented in table 4, shows that the coarse fraction generally produces smaller CO_2 yields and lighter $\delta^{18}O$ values. Although this could be evidence for separate generations of calcite, these differences can also be explained by the reaction of calcite from separate environments of precipitation, such as veinlet fillings or replacement of plagioclase. The low yields suggest that some of the calcite in the coarse fraction has been sheltered from reaction with the phosphoric acid. If this unreactive calcite resides in altered plagioclase, which seems likely, then the enrichment of ^{18}O in calcite of the powdered rock adds support to the hypothesis that plagioclase can influence its alteration products by isotopic exchange, as proposed for the K-mica $\delta^{18}O$ data.

This provides a potential complication to the interpretation of the $\delta^{18}O$ data for calcite. For example, the coarse rock carbonate isotopic data might more accurately reflect recent conditions in the drill holes prior to drilling. However, to be consistent with other isotopic studies, the powdered whole rock isotopic data is utilized in the following discussions.

Temperature-depth profiles

The interpretation of oxygen isotopes in calcite from geothermal areas requires the knowledge of at least two out of three variables: The isotopic composition of calcite; the isotopic composition of the thermal water; and the temperatures in the host rock prior to drilling. At Roosevelt Hot Springs the isotopic composition of calcite and the thermal water can be measured. However, because migration of the fluid is controlled by permeable fracture zones, uncertainties exist as to whether waters collected from the wells accurately reflect water compositions in less permeable rocks. As an added complication, temperatures measured in the wells appear to be disturbed by the convective circulation of steam and water produced when the production zones are intersected by the drill holes.

In well 72-16, figure 5, this effect of convective circulation is readily apparent. Temperatures measured on 11/17/76 prior to completion of the well are slightly, although consistently, higher than isotopic temperatures. However, the temperature-depth curve labeled 3/30/77, which was obtained after a maximum of three months shut-in of the well (Glenn and Hulen, 1979) provides a good approximation to the boiling point curve (Haas, 1971). This suggests that temperatures measured following circulation of the well may exceed the temperatures existing in the subsurface prior to drilling. In non-productive wells, such as 9-1 (figure 6), the measured temperatures probably conform to the pre-existing temperature-depth profile since only small quantities of water seep into the well.

The measured temperature profile in well 14-2 may also fail to reproduce the true temperature-depth curve because of convective circulation. Figure 7 compares measured and isotopic temperatures with fluid inclusion temperatures from hydrothermal quartz (W. T. Parry, personal communication). The isotopic temperatures approach measured temperatures only at 870 and 1585 m, two depths which are considered hot water entry zones (Bamford, 1978). Above these zones the measured temperatures are considerably higher than both isotopic and fluid inclusion temperatures, again indicating that measured temperatures probably do not represent temperatures established prior to drilling.

Two explanations can account for the difference between isotopic and fluid inclusion temperatures in well 14-2. Either the isotopic temperatures are an accurate reflection of the temperature profile or the carbonates have equilibrated with a thermal water considerably

Table 7. Symbols and references for figures 5, 6, and 7.

LITHOLOGY

- ░ Qal — Quaternary Arkosic Alluvium
- ▓ Tmd – Tertiary Mircodiorite
- ⁘ Tg — Tertiary Medium to Coarse-grained Leucocratic Biotite Granite
- ▒ Tgr – Tertiary Fine-grained Granite
- ▨ Tqm – Tertiary Quartz Monzonite
- ☰ P€n – Precambrian (?) Quartz Monzonite to Granodiorite Gneiss
- ▤ P€gn-Precambrian Hornblende Gneiss (Probably related to P€n)
- ▧ P€bg-Precambrian Banded Gneiss

REFERENCES

Lithology
- 14-2 (Ballantyne and Parry, 1978; Glenn and Hulen, 1979)
- 9-1 (Hulen, 1979, lithologic log)
- 72-16 (Glenn and Hulen, 1979)

Alteration Intensity
- 14-2 (Ballantyne and Parry, 1978)
- 9-1 (Hulen, 1979, lithologic log)
- 72-16 (Nielsen, et.al., 1978; Rohrs and Parry, 1978)

Measured Temperatures
- 14-2 (Glenn and Hulen, 1979)
- 9-1 (Nielsen, Preliminary geological and geophysical logs)
- 72-16 (Glenn and Hulen, 1979)

Additional Symbols
- O Measured Isotopic value
- + Calculated Isotopic temperature

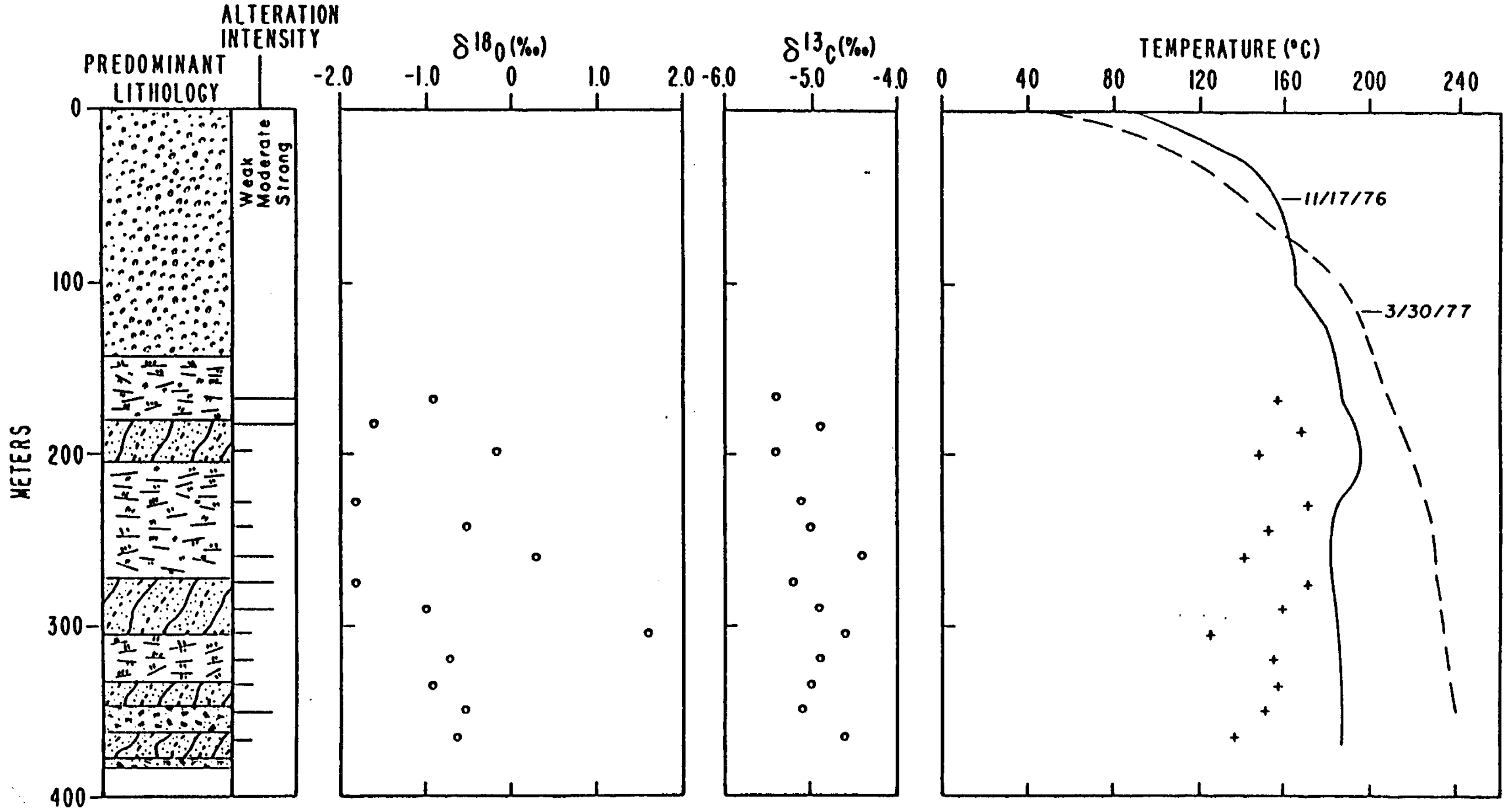

Figure 5. $\delta^{18}O$ and $\delta^{13}C$ values of calcite, isotopic temperatures, and measured temperatures plotted versus depth for well Utah State 72-16. For an explanation of the symbols and references, see table 7.

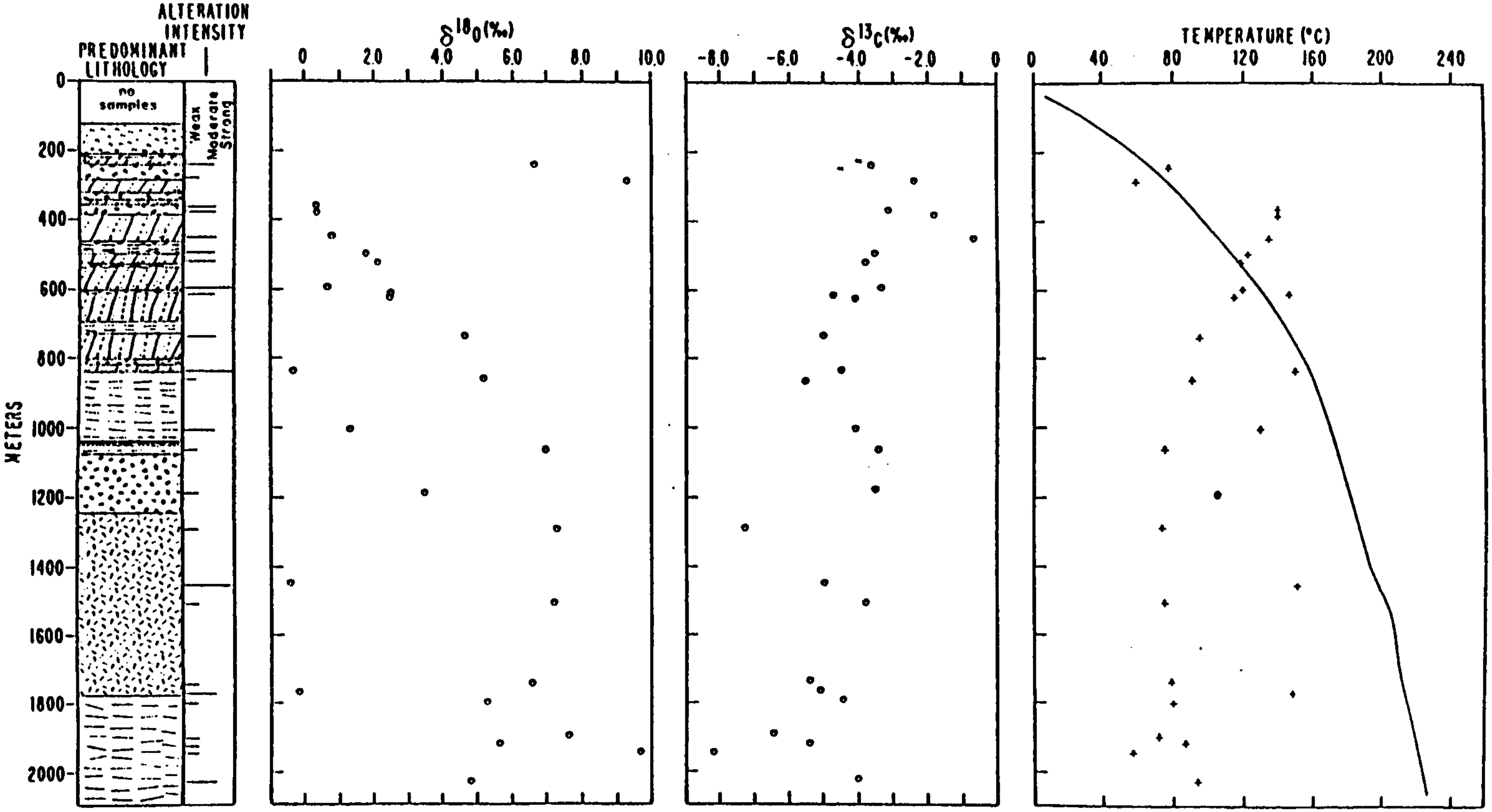

Figure 6. $\delta^{18}O$ and $\delta^{13}C$ values of calcite, isotopic temperatures, and measured temperatures plotted versus depth for well Utah State 9-1. For an explanation of the symbols and references, see table 7.

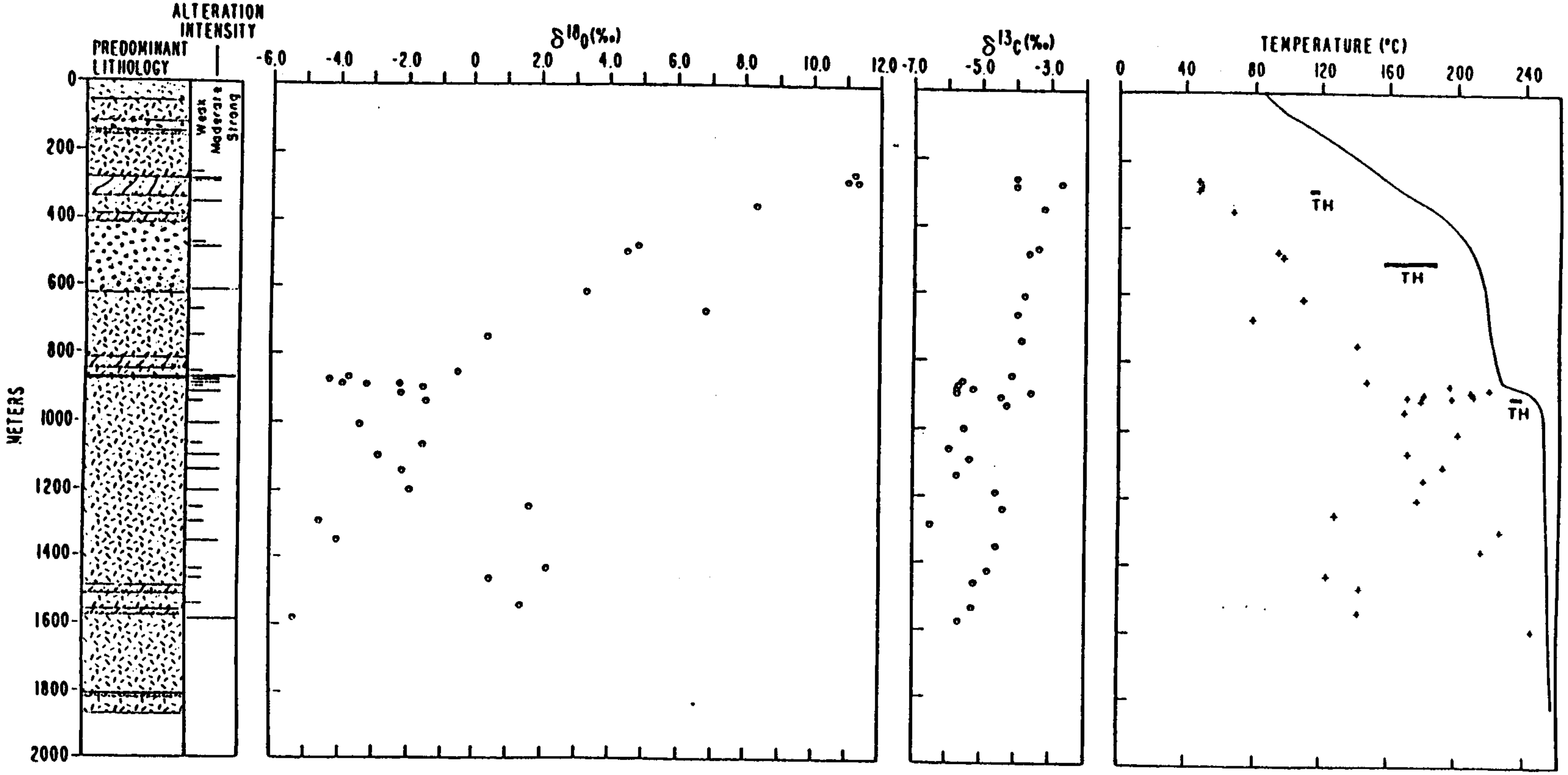

Figure 7. $\delta^{18}O$ and $\delta^{13}C$ values of calcite, isotopic temperatures, fluid inclusion temperatures (T_H) and measured temperatures plotted versus depth for well Utah State 14-2. For an explanation of the symbols and references, see table 7.

enriched with ^{18}O compared to a typical thermal fluid from production zones.

The δD values of the K-mica clays provide weak evidence that the thermal waters may become enriched by evaporation above 850 m. Assuming that the fluid inclusion temperatures represent the temperatures of precipitation, the $\delta^{18}O$ values of the calcite between 259.1 and 853.4 m enables the calculation of the equilibrated waters with the calcite - water fractionation curve of Friedman and O'Neil (1977). The percentages of evaporation required to reproduce these $\delta^{18}O$ values of the thermal fluid can be calculated by Rayleigh distillation for an initial water with $\delta^{18}O$ equal to -13.0 from the fractionation factors between water and steam given in Friedman and O'Neil (1977). The $\delta^{18}O$ of calcite, the $\delta^{18}O$ of the calculated water, temperature, and the percent of evaporation are provided in table 8.

Comparing the percentages of evaporation listed in table 8 with the evaporation calculated from the K-mica data in table 6 shows a surprising, but probably fortuitous, similarity. Although the K-mica and calcite data are compatible in that they provide similar percentages of evaporation, the need to equilibrate these minerals with water that has experienced well over 90 percent evaporation is considered questionable.

The strongest argument against a fluid reservoir with homogeneous ^{18}O content and against evaporation relies upon the difference in CO_2 yields and $\delta^{18}O$ values of calcite extracted from the coarse and powdered rock. These differences indicate that a component of the calcite in the powdered rock may be enclosed within plagioclase and

Table 8. The $\delta^{18}O$ of calcite, the calculated $\delta^{18}O$ of equilibrated thermal water, and the percent of evaporation in well 14-2.

Depth (meters)	$\delta^{18}O$ of calcite(‰)	Calc. $\delta^{18}O$ of water (‰) (1)	Δ°water-vapor (2)	%evaporation (3)	Temperature°C (4)
259.1	+11.2	-5.3	5.1	78	106
487.7	+4.5	-6.9	3.2	86	168
609.6	+3.3	-6.9	2.7	90	188
670.6	+6.8	-2.7	2.5	98	200
853.4	0.0	-8.2	2.0	91	228

(1) Estimated from the $\delta^{18}O$ of calcite and the fractionation curve in Friedman and O'Neil (1977).

(2) Friedman and O'Neil, 1977.

(3) Assuming Rayleigh distillation (Broecker and Oversby, 1971).

(4) Extrapolated fluid inclusion temperatures.

that this component has been enriched with ^{18}O because of isotopic exchange between the plagioclase and the thermal water. This process has already been invoked to explain the ^{18}O enrichment of the K-mica clays.

The only interval where isotopic exchange cannot account for the $\delta^{18}O$ value of calcite in the powdered rock is 259.1 m. At the fluid inclusion temperature of 106°C the feldspar would have to possess a $\delta^{18}O$ value of +11.6, exceeding by 3.7 the $\delta^{18}O$ value of K-feldspar from this interval. Therefore, the thermal waters must be enriched by a separate process, which is probably boiling.

Based upon the preceding discussion, it can be concluded that the isotopic temperatures calculated from the powdered whole rock carbonate data underestimate the temperature-depth profile in well 14-2. Thus, the fluid inclusion temperatures probably represent the true temperatures existing in the subsurface prior to drilling, while the measured temperatures are inaccurate due to the convective circulation of water and steam within the well.

Additional evidence for isotopic exchange

A correlation between $\delta^{18}O$ and alteration intensity is evident below 850 m in wells 14-2 and 9-1 (figures 6 and 7), where carbonates are enriched with ^{18}O in weakly altered rocks but approach equilibrium with the thermal water in moderately to strongly altered rocks. To produce this enrichment of ^{18}O in the carbonates, the water must become correspondingly enriched. As an extreme example, at

1944.6 m in well 9-1 waters in equilibrium with the carbonate at a measured temperature of 221°C would have an isotopic composition of +1.2.

Since the fractionation between water and steam at temperatures above 200°C is less than 2.5 (Friedman and O'Neil, 1977), evaporation can be eliminated as a likely mechanism for enriching the waters. Instead, isotopic exchange is favored. Whole rock isotopic measurements support low W/R ratios in weakly altered rocks. Should water become stagnated due to low permeability, extensive isotopic exchange could drastically enrich the water without markedly depleting the whole rock oxygen reservoir. Thus, enriched carbonates may simply be a feature of the slow migration of thermal waters through less permeable rocks. Furthermore, if the calcite from weakly altered rocks is an alteration product of plagioclase, isotopic exchange between the host mineral and the thermal water could enhance the ^{18}O enrichment of calcite. Because carbonates from more altered rocks approach equilibrium with the typical thermal waters from production zones, these waters have not undergone extensive modification by isotopic exchange or much of the calcite has precipitated in veinlets in direct communication with the bulk of the thermal water.

Having identified water/rock isotopic exchange as an important process affecting the water in the deep reservoir, the variations of $\delta^{18}O$ values of calcite can be more easily interpreted for each well (figures 5, 6, and 7). For example, in well 14-2 the variation in $\delta^{18}O$ values of calcite below 259.1 m can be explained by modification of the thermal waters by isotopic exchange. This would account for the low calculated isotopic temperatures. At 259.1 m, the waters may have become further enriched in ^{18}O because of boiling.

Because isotopic temperatures in well 72-16 are in reasonable agreement with temperatures measured on 11/17/76 (figure 5), the water has probably migrated rather rapidly through the system without experiencing much modification. Weakly altered rocks typically possess carbonates enriched with ^{18}O, probably because of isotopic exchange. However, one moderately altered rock, at 259.1 m, also provides calcite enriched with ^{18}O, which might be an indication of boiling to enrich the thermal waters at this depth.

Between 300 and 500 m, well 9-1 differs from the other wells because isotopic temperatures exceed measured temperatures (figure 6). Otherwise, the interpretations for the $\delta^{18}O$ values in calcite are similar to interpretations for wells 72-16 and 14-2. Where the isotopic temperatures exceed measured temperatures, the calcite could be indicating either higher temperatures than now exist in the well or the presence of a thermal water with a $\delta^{18}O$ value less than -13.0. The $\delta^{13}C$ data, discussed in the next section, aids in choosing between these two possibilities.

Carbon Isotopic Composition of Calcite

Variations in $\delta^{13}C$ values of calcite with depth for each well are presented in figures 5-7. Calcite from well 72-16 maintains a fairly uniform isotopic composition near -5.0, while in wells 14-2 and 9-1 there is an overall enrichment of ^{13}C towards the surface, typically from about -6.0 to -2.4. Although oxygen isotopes of calcite often approach equilibrium with water in geothermal systems, carbon isotopes are not easily interpreted as equilibrium values. For example, at Wairakei there is no correlation of $\delta^{13}C$ with depth, temperature, or carbonate content of the rock (Clayton and Steiner, 1975). Even within well-studied mineral deposits, $^{13}C/^{12}C$ ratios often present greater ranges than can be explained by simple processes such as cooling, necessitating changes in CH_4/CO_2 ratios or isotopic variations of the total carbon content of the fluids (Ohmoto and Rye, 1979).

Because the $^{13}C/^{12}C$ of calcite are a function of temperature, pH, fO_2, and the isotopic composition of the total carbon reservoir, any explanation for the $\delta^{13}C$ values of calcite must take into account changes in the chemistry and temerature of the solutions as they migrate towards the surface. The alteration assemblage and the chemistry of the thermal fluids help to establish these chemical parameters.

The coexistence of hematite and magnetite near 870 m in well 14-2, the hot water entry zone, if equilibrated with the thermal fluids, defines fO_2 of the deep thermal waters on the hematite-magnetite buffer curve. At 250°C fO_2 would be about 10^{-33} atm (Helgeson, 1969). At this fO_2 methane makes up a negligible fraction of the total carbon reservoir, as shown by the following chemical equilibria (Ohmoto, 1972):

$CH_4(aq) + 2O_2 = H_2CO_3\ (app) + H_2O(l)$ $(\log k_{250} = 1077.52)$ where $mH_2CO_3\ (app) = mCO_2\ (aq) + mH_2CO_3$. The ratio $H_2CO_3\ (app)/CH_4(aq)$ is

$10^{11.52}$, emphasizing that the variations in $\delta^{13}C$ of calcite do not result from variations in H_2CO_3 (app)/CH_4(aq). This conclusion is supported by chemical analyses of the gases collected during flow tests which show only traces of CH_4 (A. H. Truesdell, personal communication).

The assemblage quartz - K-feldspar - K-mica - calcite determines both the distribution of oxidized carbon species in solution and the partial pressure of CO_2(g) as a function of temperature, as shown in figure 8A. This diagram was constructed with the thermodynamic data listed in table 9 by applying a fixed concentration of K^+ to establish pH and by varying only temperature at an ionic strength of 0.1. K^+ was chosen as 375 mg/liter, an intermediate concentration for thermal waters from production wells and the surficial seep (table 1). Although the K^+ concentration actually varies from 488 to 274 mg/liter, the log activities for these concentrations are within five percent of the log activity for 375 mg/liter. Hence, maintaining a constant concentration of K^+ does not substantially alter the application of figure 8A to the geothermal system.

Figure 8A predicts that as the waters ascend and cool from 250 to 100°C there is a dramatic decrase in PCO_2. While HCO_3^- becomes the dominant aqueous carbonate species at lower temperatures, the total concentration of dissolved carbonate decreases by approximately 60 percent in cooling from 250 to 100°C. The mole fractions of H_2CO_3 app, HCO_3^- and CO_3^{2-} are presented in table 10 with the total molality of the dissolved carbonate species.

One of the primary obstacles to interpreting carbon isotopes of

Table 9. Activity coefficients and equilibrium constants for reactions determining the alteration assemblage at Roosevelt Hot Springs

	Temperature °C			
Activity Coefficients (1)	100	150	200	250
neutral molecules	1.011	1.011	1.011	1.008
univalent ions	0.748	0.706	0.657	0.605
divalent ions	0.303	0.240	0.180	0.131
log a K^+(2)	-2.14	-2.17	-2.20	-2.24
Reactions (3)				
$H_2O = H^+ + OH^-$	-12.26	-11.64	-11.27	-11.13
$H_2O + CO_2(g) = H_2CO_3(app)$	-1.97	-2.07	-2.06	-1.98
$H_2CO_3(app) = HCO_3^- + H^+$	-6.45	-6.73	-7.08	-7.63
$HCO_3^- = CO_3^{2-} + H^+$	-10.16	-10.29	-10.68	-11.43
$CaCO_3 = Ca^{2+} + CO_3^{2-}$	-9.39	-10.25	-11.37	-12.72
$KAlSi_3O_8 + 2H^+ =$ $KAl_3Si_3O_{10}(OH)_2 + 2K^+ + 6SiO_2$	9.55	8.91	8.49	8.12

(1) Helgeson, 1969, with an ionic strength of 0.1.

(2) Concentration of K^+ = 375 mg/liter

(3) Helgeson, 1969.

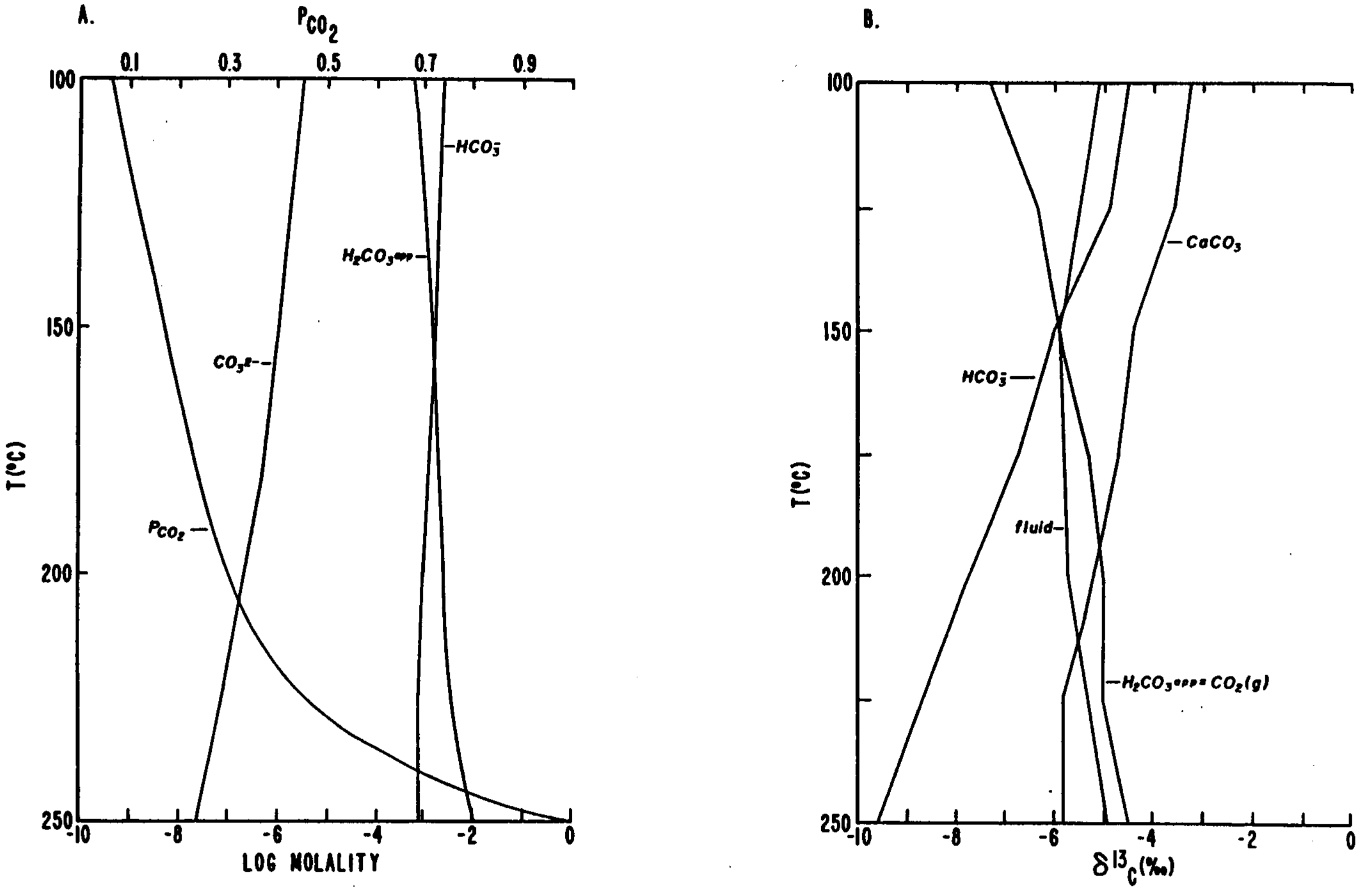

Figure 8. (A) The molalities of H_2CO_3 app, HCO_3^- and CO_3^{2-} and PCO_2 as a function of temperature for K^+ = 375 mg/liter and an ionic strength of 0.1. (B) $\delta^{13}C$ values of aqueous carbonate species, $CO_2(g)$ and calcite as a function of temperature as determined by the measured $\delta^{13}C$ values of calcite and the fractionation factors given in table 11.

Table 10. The mole fractions of H_2CO_3 app, HCO_3^-, and CO_3^{2-} and the total molality of the dissolved carbonate determined from mineral equilibria.

	Temperature (°C)			
	100	150	200	250
Mole fraction				
H_2CO_3 app	0.20	0.47	0.73	0.93
HCO_3^-	0.80	0.53	0.27	0.07
CO_3^{2-}	-	-	-	-
Total molality (x 10^3)	3.07	3.18	3.59	11.0

calcite in geothermal areas is the sensitivity of H_2CO_3app and HCO_3^- in the fluid to fluctuations in temperature and pH, since a change in the proportions of these aqueous carbonate species affects the carbon isotopic composition of the precipitating calcite. The fluid, however, constitutes only one carbon reservoir. If sufficient calcite precipitates, calcite can also become a significant carbon reservoir. Also, a third carbon reservoir, CO_2 as a discrete gas phase, cannot be discounted as a significant component of the total carbon reservoir. Therefore, any proper interpretation of the carbon data must take into account both the changes in the chemistry of the fluids and interaction between three potential carbon reservoirs.

As illustrated in figures 6 and 7, the measured $\delta^{13}C$ values of calcite [$\delta^{13}C$(cc)] increase progressively with decreasing depth and temperature. A curve depicting a typical distribution of $\delta^{13}C$(cc) with temperature is presented in figure 8B. Since the $\delta^{18}O$ values of calcite often fail to correlate with temperature due to variabilities in the oxygen isotopic composition of the fluids, this curve had to be estimated from a compilation of the $\delta^{13}C$(cc) data with various temperature measurements, such as measured temperatures for well 14-2, and isotopic temperatures for well 72-16. A further discussion of this selection process is provided in the appendix.

Assuming that the calcite has precipitated in equilibrium with the solution and $CO_2(g)$, the variations of $\delta^{13}C$ with temperature for the aqueous and gaseous carbon species can be determined by using the observed $\delta^{13}C$(cc) values as a reference and applying the fractionation factors listed in table 11. These $\delta^{13}C$ distributions with temperature

Table 11. The distribution of $\delta^{13}C$ between CO_2 (g), carbon in solution, and calcite.

	Temperatures (°C)						
Fractionation Factors (1)	100	125	150	175	200	225	250
$\Delta CaCO_3 - H_2CO_3$	4.0	2.6	1.5	0.6	-0.2	-0.8	-1.3
$\Delta CaCO_3 - HCO_3^-$	1.2	1.3	1.6	2.0	2.5	3.1	3.8
$\Delta CaCO_3$ - fluid (2)	1.8	-	1.5	-	0.5	-	-0.9
ΔCO_2 - fluid (2)	-2.2	-	0.0	-	0.7	-	0.4
Isotopic Compositions							
$CaCO_3$ (3)	-3.3	-3.6	-4.4	-4.7	-5.2	-5.8	-5.8
H_2CO_3 app = CO_2 (g)	-7.3	-6.2	-5.9	-5.3	-5.0	-5.0	-4.5
HCO_3^-	-4.5	-4.9	-6.0	-6.7	-7.7	-8.9	-9.6
fluid (2)	-5.1	-	-5.9	-	-5.7	-	-4.9

(1) Ohmoto and Rye, 1979.

(2) Estimated from the proportions of H_2CO_3 app and HCO_3^- in solution (table 10).

(3) Estimated from the distribution of $\delta^{13}C$ in calcite versus measured and isotopic temperatures in wells 9-1, 14-2, and 72-16 (see appendix).

are also shown in figure 8B. Assuming $\Delta^{c}H_2CO_3 - CO_2(g) = 0$ for all temperatures (Ohmoto, 1972), the variations for $\delta^{13}C(CO_2)$ with temperature are also diagrammed. The $\delta^{13}C$ value of the total aqueous carbon reservoir, $\delta^{13}C$ (fluid), has been estimated from the relative proportions of H_2CO_3app and HCO_3^- given in table 11 for each temperature of interest.

From the inferred trends for $\delta^{13}C$ of $CO_2(g)$ and fluid in figure 8B, $\delta^{13}C$ (fluid) remains fairly constant throughout the temperature range, changing by only 1 permil. For the $CO_2(g)$ phase, $\delta^{13}C(CO_2)$ does not remain constant but continually diminishes with decreasing temperature. These two aspects of the carbon data suggest two approaches to modeling the measured $\delta^{13}C$ values of calcite at Roosevelt Hot Springs. Both models require that the observed $\delta^{13}C(cc)$ values be reproduced by the precipitation of calcite from an original solution as it migrates from temperature regions of 250 to 100°C. This original solution at 250°C has been estimated from the $\delta^{13}C(cc)$ value of -5.8 in the production zones at 1585 and 870 m in well 14-2. From the fractionation factors in table 11, the original $\delta^{13}C(CO_2)$ value is -4.5, and the $\delta^{13}C$ (fluid) value is -4.9.

<u>Model 1</u>. The first model is straightforward. $\delta^{13}C(fluid)$ is assumed to completely dominate the combined gas and aqueous carbon reservoirs in the original solution and is maintained at a constant value of -4.9 during the transition from 250 to 100°C. This model is displayed in figure 9, where the calcite precipitating from this hypothetical solution consistently exceeds the measured $\delta^{13}C(cc)$ values. In order to reproduce these measured compositions, the fluid

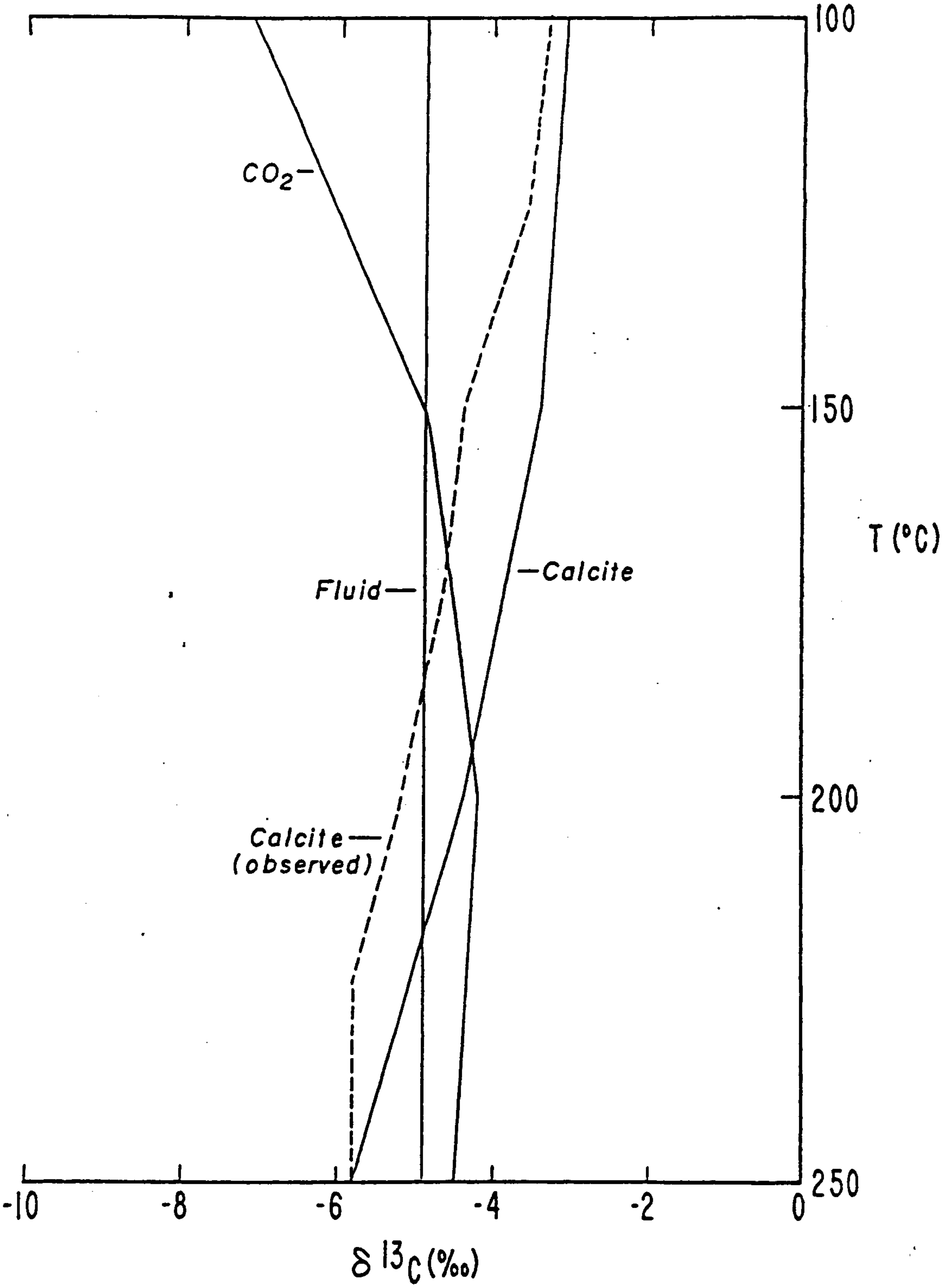

Figure 9. The variations in $\delta^{13}C$ values of calcite and $CO_2(g)$ as a function of temperature for $\delta^{13}C$(fluid) = -4.9. The dashed line corresponds to the $\delta^{13}C$(cc) curve in figure 8B.

must become depleted in ^{13}C.

Figure 8A predicts a continual decrease in the total molality of carbonate as the solution cools, which could be caused by precipitation of calcite and/or the exsolution of $CO_2(g)$. As figure 9 illustrates, exsolution between 250 and 150°C and precipitation between 210 and 100°C both preferentially partition ^{13}C into the solid and gas phases, providing the necessary depletion of the residual solution in ^{13}C. Which of the two processes, exsolution or precipitation, is more important cannot be evaluated from the available data. However, the validity of this model can be tested by comparing W/R mass ratios calculated from the quantity of calcite in the rock and the molality of carbonate in solution with W/R mass ratios from the isotopic data (table 5).

The moles of calcite per kilogram of powdered rock range from 1.0×10^{-2} to 38.0×10^{-2} (table 4) while the maximum molality of carbonate in solution available for the precipitation of calcite is 1.1×10^{-2} (table 10). Therefore, the minimum W/R mass ratios from the carbonate data range from 1 to 35, compared to a range of 0.1 to 3.0 for the isotopic data. This implies either than the W/R mass ratios from the isotopic data have been underestimated by one order of magnitude or that the fluid carbon reservoir is replenished by a separate $CO_2(g)$ reservoir in communication with the fluids. The possibility that $CO_2(g)$ does indeed constitute a significant component of the total carbon reservoir provides a basis for the second model.

Model 2. This second model assumes that $\delta^{13}C(CO_2)$ completely dominates the combined gaseous and aqueous carbon reservoirs.

However, from the observed $\delta^{13}C$ values of calcite in figure 8B, $\delta^{13}C(CO_2)$ does not remain constant during the cooling of the fluids but decreases systematically as temperatures decrease. $CO_2(g)$ could be adjusting its isotopic composition in response to the removal of ^{13}C from the gaseous and aqueous carbon reservoirs by the precipitation of significant amounts of calcite as the fluids migrate to lower temperature regions. This second model, which requires the precipitation of calcite to fractionate $\delta^{13}C(CO_2)$, works in the following manner. An initial fluid is transferred to a lower temperature while maintaining a constant $\delta^{13}C(CO_2)$. Due to the change in temperature and reapportionment of H_2CO_3app and HCO_3^-, $\delta^{13}C(fluid)$ re-equilibrates isotopically with the $CO_2(g)$. Precipitation of calcite proceeds by batch distillation under isothermal conditions until a $\delta^{13}C(fluid)$ in equilibrium with the observed calcite is produced. At this point this fluid is transferred to the next lower temperature and precipitation under isothermal conditions is renewed. This procedure has been applied to temperature drops of 50° between 250 and 100°C, providing the results summarized in table 12.

As an example of the calculations for this model, the transition from 250 to 100°C is considered. The original solution at 250°C is the same as in the first model, where $\delta^{13}C(CO_2)$ equals -4.5, and $\delta^{13}C(fluid)$ equals -4.9. Because the carbon reservoir is dominated by $CO_2(g)$, upon transferring the original solution to 200°C, $\delta^{13}C(CO_2)$ remains -4.5, but $\delta^{13}C(fluid)$ readjusts to -5.2. The initial calcite to precipitate from this solution has a $\delta^{13}C$ value of -4.7. Since the $\delta^{13}C(fluid)$ value for the observed calcite is -5.2, the $\delta^{13}C(fluid)$

must become depleted by 0.5. The precipitation of calcite by batch distillation of the fluid produces this necessary removal of ^{13}C from the fluid.

The batch distillation equation has been taken from Faure (1977):

$$\frac{1000 + \delta^{13}C\text{ (fluid, initial)}}{1000 + \delta^{13}C\text{ (fluid, final)}} = f^{\alpha-1}$$

f is the fraction of the total dissolved carbonate remaining in solution following the precipitation of calcite, and α, the fractionation factor between calcite and fluid, can be estimated from ΔCcalcite - fluid (table 11). From this equation the precipitated calcite at 200°C amounts to 63 percent of the initial total (gas and fluid) carbon reservoir. Since the solution possessed an initial total molality of 1.1×10^{-2} at 250°C, the molality at 200°C has been reduced to 4.1×10^{-3}, in good agreement with the molality of 3.6×10^{-3} required by mineral equilibria (table 10).

Table 12 illustrates the compositional changes in the gas, fluid, and calcite required by the model at each temperature of interest. Also presented in table 12 is the required percent of precipitation and a comparison between the molality of total carbonate resulting from precipitation and the molality required by mineral equilibria. The remarkable similarities of the molalities supports the validity of model 2 and the existence of a significant $CO_2(g)$ carbon reservoir. How this gas phase originates and migrates through the reservoir is not readily apparent, although it could represent immiscible gas bubbles, perhaps derived by exsolution below 2 kilometers depth, working their way through the water column towards the surface.

Table 12. Compositional changes in $\delta^{13}C$ following precipitation of calcite from a CO_2 (g) dominated carbon reservoir at 200, 150, and 100 °C.

Original Solution, 250°C	
$\delta^{13}C(CO_2)$	-4.5
$\delta^{13}C$(fluid)	-4.9
$\delta^{13}C$(cc)	-5.8

	initial	200 °C	final
$\delta^{13}C(CO_2)$	-4.5		-5.0
$\delta^{13}C$(fluid)	-5.2		-5.7
$\delta^{13}C$(cc)	-4.7	63% ppt	-5.2
		150 °C	
$\delta^{13}C(CO_2)$	-5.0		-5.9
$\delta^{13}C$(fluid)	-5.0		-5.9
$\delta^{13}C$(cc)	-3.5	37% ppt	-4.4
		100 °C	
$\delta^{13}C(CO_2)$	-5.9		-7.3
$\delta^{13}C$(fluid)	-3.7		-5.1
$\delta^{13}C$(cc)	-1.9	54% ppt	-3.3

Final Molalities of H_2CO_3 app + HCO_3^- (X 10^3)

T, C	after precipitation	mineral equilibria
250	-	11.0
200	4.1	3.59
150	2.6	3.18
100	1.2	3.07

Discussion of the two models. The major difference between the two models is simply which component, $CO_2(g)$ or fluid, dominates the combined $CO_2(g)$ and fluid carbon reservoir. In model 1 the fluid dominates, while in model 2 $CO_2(g)$ dominates. Both models can reproduce the measured $\delta^{13}C$ values of calcite at Roosevelt Hot Springs. However, model 1 is ambiguous because either exsolution of $CO_2(g)$ or precipitation of calcite can modify the $\delta^{13}C$(fluid). Also, model 1 suffers from the apparent discrepancy between W/R mass ratios calculated from carbonate mass balance and from the water and rock isotopic data. Model 2, on the other hand, depends upon a sufficient proportion of the total carbon reservoir as $CO_2(g)$. Although $CO_2(g)$ must exist from phase equilibria considerations, it is difficult to quantify the gas phase in terms of the mass proportions of gas to fluid. In view of these uncertainties neither model can be explicitly favored.

There are certain similarities in these models which provide a better understanding of the carbon isotope systematics at Roosevelt Hot Springs. For example, as temperature decreases carbon is being transferred from the fluid and gas into calcite, accounting for the decreases in PCO_2 and total molality of carbonate in solution as predicted by mineral equilibria. Also, no single component, $CO_2(g)$, fluid, or calcite, completely dominates the total carbon reservoir. Finally, from these models it becomes apparent why carbon isotopes of calcite from other geothermal systems have proven difficult to interpret. The reason is simply because of the complexity of the system CH_4-CO_2-H_2O-calcite.

Calcite is sensitive to a wide range of processes which can affect its ultimate carbon isotopic composition in geothermal systems. These include fluctuations in temperature, pH, and fO_2, exsolution of CO_2(g) and precipitation of calcite, and variations in the isotopic composition of carbon contributed by the carbon source(s). Unfortunately, unravelling which of these processes have governed the $\delta^{13}C$ values of calcite requires additional data which may not be available from the study of alteration assemblages or from the geothermal fluids extracted from production wells.

Anomalous $\delta^{13}C$(cc) values. $\delta^{13}C$(cc) values which deviate within 1 per mil of the enrichment trend of figure 8B are difficult to interpret due to the many factors which can influence the carbon isotopic composition of calcite. However, there are certain anomalous $\delta^{13}C$(cc) values in well 9-1 which offer interesting challenges to the two models.

As mentioned earlier, between 300 and 500 m depth in well 9-1 the oxygen isotopic composition of a water calculated from the $\delta^{18}O$ values of calcite and the measured temperature (about 100°C) is much lighter in ^{18}O value than any thermal or spring water collected in the area, providing a $\delta^{18}O$ of -17.6. Within this depth interval, at 451.1 and 378.0 m, two carbonates are unusually enriched with ^{13}C, providing $\delta^{13}C$ values of -0.7 and -1.8. At the measured temperatures, CO_2(g) in isotopic equilibrium with $\delta^{13}C$(cc) value of -0.7 would have an isotopic composition of -4.7. Assuming the measured temperature represents the actual temperature of precipitation, this interval could be interpreted as a zone of steam condensation which has incorporated $CO_2(g)$ characteristic of the deep reservoir. However, the strong calcite alteration in these rocks (Hulen, 1979) could signify intensified precipitation of calcite caused by boiling. The oxygen isotopes could be recording a higher temperature of precipitation (about 140°C) providing a maximum $\delta^{13}C$ value of -2.6 for the $CO_2(g)$ in equilibrium with the calcite. The only logical means of enriching the $\delta^{13}C(CO_2)$ to -2.6 is by the progressive removal of $CO_2(g)$ from the solution by boiling, since the precipitation of calcite would cause the opposite effect of depleting the gas with ^{13}C. On the other hand, these carbonates may have precipitated from a thermal fluid for which there is no modern analog in the geothermal area.

Beneath 850 m in well 9-1, several carbonates provide anomalously depleted $\delta^{13}C$ values. An extreme example occurs at 1944.6 m, where $\delta^{13}C(cc)$ is -8.2. Since the ^{13}C fractionation between calcite and $CO_2(g)$ never exceeds 2.8 (Friedman and O'Neil, 1977), the carbon isotopic composition of $CO_2(g)$ in equilibrium with calcite at 1944.6 m must be less than -5.4, which is considerably lighter than the -4.5 suggested by calcite from the production zones in well 14-2. These carbonates with unusually low $\delta^{13}C$ values are generally enriched with ^{18}O and occur only in weakly altered rocks. Once again, the carbonates could have precipitated from a fluid which is distinct from the thermal water currently extracted from the production wells.

Source of the carbon

The source of the carbon must be interpreted from the $^{13}C/^{12}C$

ratio for the total carbon reservoir, which is composed of $CO_2(g)$, total carbonate in solution, and calcite. Because $CO_2(g)$ and the fluid possibly become modified by the precipitation of calcite, the carbon reservoir must be estimated at 250°C prior to any significant ascent of the fluids. As determined earlier (figure 8B), at 250°C, the $\delta^{13}C$ values of calcite, $CO_2(g)$ and fluid are -5.8, -4.5, and -4.9, respectively. Since the relative percentages of calcite, $CO_2(g)$ and dissolved carbonate are difficult to estimate, the total carbon reservoir composition is within -5.8 and -4.5.

Typical values for juvenile carbon for carbonatites and diamonds in kimberlites are -5.0±2 (Deines and Gold, 1973). However, -5.5 is the typical $\delta^{13}C$ for average sedimentary and metamorphic rocks (Ohmoto and Rye, 1979). Thus, the $\delta^{13}C$ value of the carbon reservoir is consistent with contributions of carbon from either the igneous or the metamorphic rocks in the area, and it is not possible to restrict the source of the carbon to any one particular lithology.

As an additional complication, the granitic rocks and gneiss may contain insufficient carbon to account for the quantities of calcite precipitated from the geothermal fluids. An alternative carbon source could be CO_2 from the atmosphere. Assuming that meteoric waters from the Mineral Mountains have a near neutral pH at 25°C and have equilibrated with an atmospheric PCO_2 at $10^{-3.5}$ atm (Stumm and Morgan, 1970), the molality of total dissolved CO_2 is 6.0×10^{-5} with a distribution of H_2CO_3app, HCO_3^-, and CO_3^{2-} as 18, 82, and less than 0.1 percent, respectively (Helgeson, 1969). Since atmospheric CO_2 has a $\delta^{13}C$ of -7.0 (Faure, 1977), the $\delta^{13}C$ of the meteoric water is -0.6

(determined from the fractionation factors between CO_2 and aqueous carbon species at 25°C given in Ohmoto and Rye, 1979). For comparison, the thermal fluids at 250°C have a molality of total dissolved carbonate of 1.1×10^{-2} and a $\delta^{13}C$ of -4.9. Therefore, unless the metoeric waters become enriched with dissolved carbonate and depleted in ^{13}C during their percolation into the deep geothermal reservoir, atmospheric CO_2 can only be a minor contributor to the total carbon reservoir at Roosevelt Hot Springs.

CONCLUSIONS

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

1) The thermal fluids originate as meteoric waters from a dominant recharge area in the Mineral Mountains and become enriched with ^{18}O by isotopic exchange with the country rock. Thermal waters display an isotopic shift of at least +1.2, although isotopic shifts are probably larger for water enclosed within impermeable rocks due to lower W/R mass ratios.

2) Depletions of ^{18}O in whole rock, K-feldspar, and biotite are generally more extensive in rocks which have experienced more alteration. W/R mass ratios calculated from oxygen isotope shifts of the rock and water vary from 0.1 to 3.0. However, the extent of exchange experienced by K-feldspar is only 30 percent even within a production zone.

3) Plagioclase may influence the oxygen isotopic composition of its alteration products, K-mica and calcite, by a process of isotopic exchange under low W/R conditions.

4) Measured temperatures may not reflect the temperature-depth curve established in the rocks prior to drilling because of the convective circulation of steam and water within the well.

5) Isotopic temperatures calculated from the $\delta^{18}O$ values of calcite are questionable because of the uncertainty in the $\delta^{18}O$ values of the thermal water from which the calcite precipitated.

6) The measured $\delta^{13}C$ values of calcite systematically increase towards shallower depths. This trend cannot be the result of temperature change alone. Model calculations require significant depletions in ^{13}C of the fluid and/or gas components of the total carbon reservoir by the progressive precipitation of calcite from an initial fluid as it migrates upward and cools from 250 to 100°C. Precipitation of calcite also accounts for the decrease in solubility of CO_2 predicted by mineral equilibria.

7) The $\delta^{13}C$ value of the carbon reservoir, between -4.5 and -5.8, could be produced by several carbon sources, including the gneiss and granitic rocks hosting the geothermal system.

8) The $\delta^{18}O$ values of calcite from powdered rock and the δD values of K-mica clay suggest extensive evaporation of the thermal water between 850 and 259.1 m in well 14-2. Alterative explanations for the ^{18}O and D enrichments in these minerals negates evaporation except at 259.1 where boiling may have occurred.

9) The rock does not interact with the bulk of the thermal water because circulation is restricted to permeable fracture zones with little infiltration into the surrounding impermeable rocks. This accounts for the variable W/R ratios, the enriched $\delta^{18}O$ values of calcite and K-mica, and the small extent of exchange experienced by K-feldspar even within production zones.

10) Only veinlet minerals appear to equilibrate with the thermal water in production zones. Therefore, veinlet minerals are more useful than whole rock data for interpreting the evolution of the thermal system in terms of changing hydrology and thermal regime.

Sufficient veinlet minerals cannot be recovered from drill cuttings, emphasizing the need for drill core in the study of fracture-controlled geothermal systems. Core would also enable better identification of lithologies, mineral parageneses, and pre-geothermal alteration products.

The isotopic data supports the importance of fracture zones for the circulation of the thermal fluids, and this may be of critical importance to the decline in hot spring activity. The hot spring deposits and surficial alteration attest to an active hot spring system during the evolution of the geothermal system. Today, the only existing surficial manifestation of the thermal fluids is the seep. Thus, the quantity of water finding its way to the surface has diminished considerably. Although this may be accounted for by less precipitation in the recharge area, a decrease in permeability caused by the precipitation of hydrothermal minerals in the fracture zones cannot be overlooked.

If sealing of the fractures by the precipitation of calcite, quartz, and other hydrothermal minerals is responsible, the successful long-term exploitation of the geothermal system may be jeopardized. By drilling production wells into the subsurface, the flow of thermal water within the fracture zones has undoubtedly increased. This could have the adverse effect of increasing the precipitation of hydrothermal minerals, which would further reduce the permeability of the system.

Suggestions for Future Isotopic Studies

With regard to this study, it is important to continue monitoring the $\delta^{18}O$ and δD values of water and steam from the production wells. 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.

In order to study the relationship between the acid-sulfate and propylitic alteration assemblages, minerals from the acid-sulfate alteration should also be isotopically analyzed. Since alunite is an important component of the surficial alteration, a study of $\delta^{34}S$ between alunite and pyrite might elucidate upon oxidation-reduction mechanisms. Also, $\delta^{34}S$ of pyrite could identify the source of the sulfur in the system. 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.

Isotopic and geochemical studies of gases and application of gas thermometers are also potentially valuable in evaluating deep reservoir temperatures, boiling, and formation of steam caps.

Because so few flow tests were conducted during this contract period and because of delays in fabrication of the necessary gas-collection and extraction equipment, we did not perform this type of analysis.

APPENDIX

Determining the $\delta^{13}C$ of Calcite versus Temperature

Properly determining the temperature at which calcite precipitated is critical for the construction of figure 8B, since the variations in $\delta^{13}C$ with temperature for $CO_2(g)$, H_2CO_3, and HCO_3^- are referenced to the isotopic composition of calcite. Had the $\delta^{18}O$ of calcite closely correlated with temperature, choosing the proper $\delta^{13}C$ would have been no problem. However, the $\delta^{18}O$ of calcite reflects variations in the isotopic composition of the thermal water which arise from isotopic exchange and perhaps from evaporation. Thus calculated isotopic temperatures may be inaccurate. Also, the measured temperatures in wells 14-2 and 72-16 are not trustworthy because of the convection of water and steam in the wells.

To overcome these problems, the $\delta^{13}C$ data from each drill hole was compared to different temperature measurements in the following manner:

1) For well 14-2, the $\delta^{13}C$ was referenced to both isotopic and fluid inclusion temperatures by inspection.

2) For well 72-16, $\delta^{13}C$ was referenced solely to those isotopic temperatures in good agreement with measured temperatures from 11/17/76.

3) For well 9-1, $\delta^{13}C$ was referenced to measured temperatures by

inspection and by determining a line of linear regression through the $\delta^{13}C$ versus temperature data set.

These correlations between $\delta^{13}C$ and temperature for each well are provided in table 13 for intervals of 25° between 100 and 250°C. By integrating the results of these comparisons, the selected $\delta^{13}C$ versus temperature was obtained, which is also presented in table 10. This integrated data set provides the trend of $\delta^{13}C$ with temperature given in figure 8B.

Naturally, by selecting the $\delta^{13}C$ values in this fashion, there will be some disagreement between the drillholes. However, this procedure is considered valid because it is consistent with the typical range of $\delta^{13}C$ values and it accounts for the observed trend of ^{13}C enrichment in calcite towards shallower depths and lower temperatures.

Table 13. A comparison between $\delta^{13}C$ of calcite from each well with selected temperature determinations. The column "$\delta^{13}C$ selected" provides the trend of $\delta^{13}C$ $CaCO_3$ versus temperature in figure 8B.

	14-2		9-1		72-16	
T °C	Fluid inclusion	Isotopic	Linear regression	Inspection	Isotopic	$\delta^{13}C$ Selected
100	-2.7 to -4.0	-3.7	-3.0	-3.1 to -3.5	-	-3.3
125	-3.2 to -4.0	-4.0	-3.6	-3.3 to -3.8	-	-3.6
150	-3.3	-4.2 to -5.6	-4.1	-4.1 to -5.5	-	-4.4
175	-3.7	-4.2 to -5.6	-4.7	?	-5.1	-4.7
200	-4.0	-4.2 to -5.6	-5.2	-5.0	-	-5.2
225	-5.7	-5.8(?)	-5.8	-5.1	-	-5.8
250	-	-5.8	-6.3	-	-	-5.8

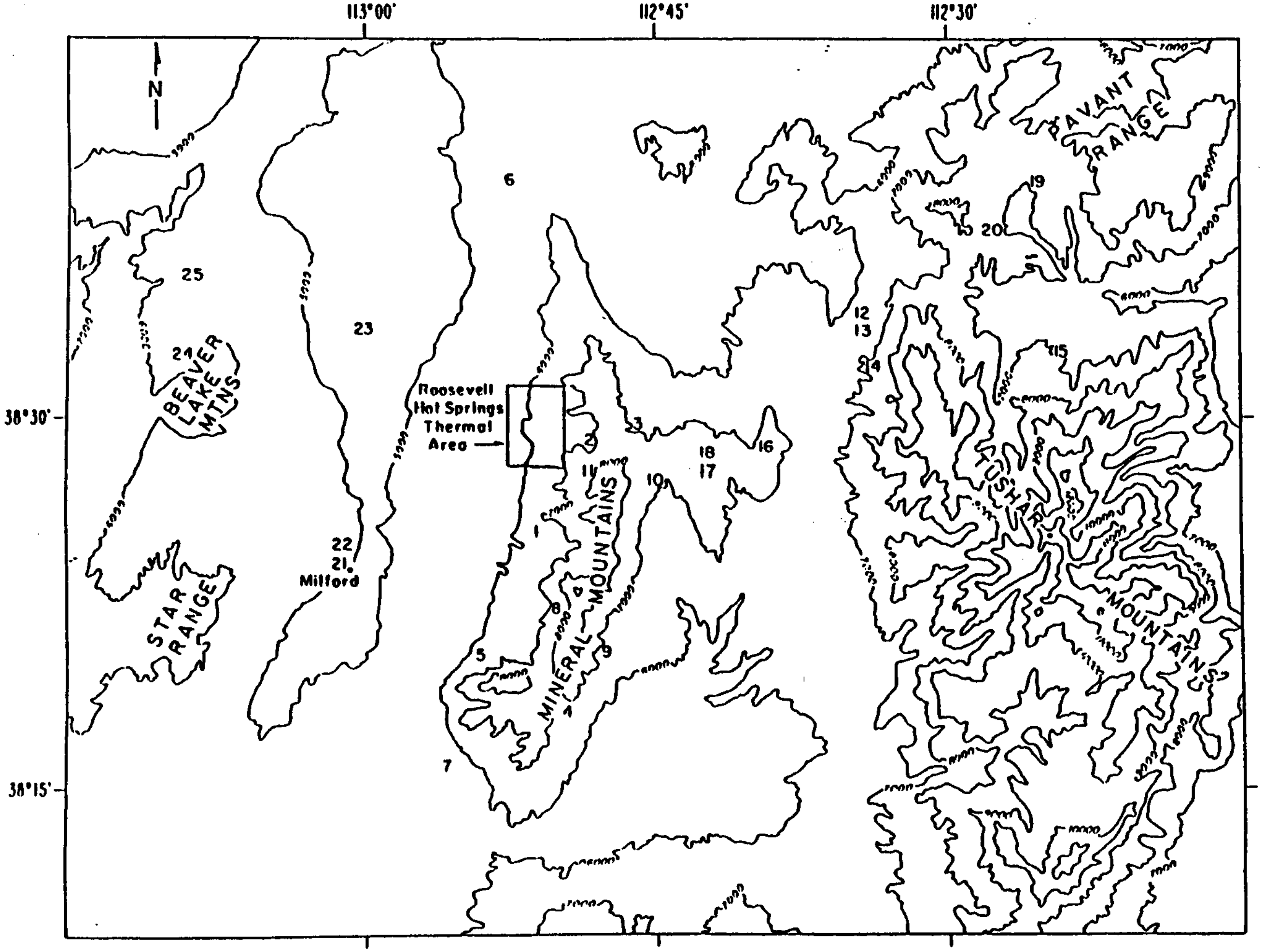

PLATE I. Geographical features and spring localities in the vicinity of the Roosevelt Hot Springs Thermal Area. The shaded region corresponds to the area of figure I.

EXPLANATION

SPRING LOCALITIES

1. Ranch Canyon (Kirk)
2. Bailey
3. Willow
4. Cherry Creek
5. Griffith
6. Antelope
7. North
8. Rock Corral
9. Mud
10. Mathew
11. Jack Rabbit
12. Sulfurdale North
13. Sulfurdale South
14. Dead Cow
15. Mud
16. Four Mile
17. Wire Grass
18. Cowboy
19. North Creek
20. Magpie Creek
21. Milford City Well
22. South of Milford Airport
23. Pearson Well
24. Beaver Lake Mine Spring
25. Armstrong Spring

SCALE
5 0 5 10
KILOMETERS

Contour interval 1000 feet

REFERENCES

Armstrong, R. L., 1970, Geochronology of Tertiary igneous rocks, eastern Basin and Range province, western Utah, eastern Nevada, and vicinity, U.S.A.: Geochim. Cosmochim. Acta, v. 34, p. 203-232.

Arnasson, B., 1977, Hydrothermal systems in Iceland traced by deuterium: Geothermics, v. 5, p. 125-151.

Ballantyne, J. M., 1978, Hydrothermal alteration at the Roosevelt Hot Springs thermal area, Utah: Modal mineralogy and geochemistry of sericite, chlorite, and feldspar from altered rocks, Thermal Power Company Well Utah State 14-2: Topical report, v. 78-1701.a.1.1.5, DOE/DGE contract EG-78-C-07-1701, 42 p.

Ballantyne, J. M. and Parry, W. T., 1978, Hydrothermal alteration at the Roosevelt Hot Springs Thermal Area, Utah: Petrographic characterization of the lateration to 2 kilometers depth: Topical report, v. 78-1701.a.1.1, DOE/DGE contract EG-78-C-07-1701, 90 p.

Bamford, R. W., 1978, Geochemistry of solid materials from two U.S. geothermal systems and its application to exploration: University of Utah Research Inst., Earth Science Laboratory Final report v. 77-14, DOE/DGE contract EY-76-S-07-1601, 196 p.

Blattner, P., 1975, Oxygen isotopic composition of fissure-grown quartz, adularia, and calcite from Broadlands geothermal field, New Zealand, with an appendix on quartz - K-feldspar - calcite - muscovite geothermometers: Am. Jour. Sci., v. 275, p. 785-800.

Broecker, W. S., and Oversby, V. M., 1971, Chemical equilibria in the earth, McGraw-Hill, New York, p. 150-170.

Clayton, R. N. and Mayeda, T. K., 1963, The use of bromine pentafluoride in the extraction of oxygen from oxides and silicates for isotopic analysis: Geochim. Cosmochim. Acta, v. 27, p. 43-52.

Clayton, R. N., Muffler, L. J. P., and White, D. E., 1968, Oxygen isotope study of the River Ranch No. 1 well, Salton Sea geothermal field, California: Am. Jour. Sci., v. 266, p. 968-979.

Clayton, R. N. and Steiner, A., 1975, Oxygen isotope studies of the geothermal system at Wairakei, New Zealand: Geochim. Cosmochim. Acta, v. 39, p. 1179-1186.

Craig, H., 1957, Isotopic standards for carbon and oxygen and correction factors for mass spectrometric analysis of carbon dioxide: Geochim. Cosmochim. Acta, v. 12, p. 133-149.

__________, 1961a, standard for reporting concentrations of deuterium and oxygen-18 in natural waters: Science, v. 133, p. 1833-1834.

__________, 1961b, Isotopic variations in meteoric waters: Science, v. 133, p. 1702-1703.

__________, 1963, The isotope geochemistry of water and carbon in geothermal areas, in Tongiorgi, E. (ed.), Nuclear Geology on Geothermal Areas, Spoleto: Consiglio Nationale Delle Ricerche, Laboratorio di Geologia Nucleare, Pisa, Italy, p. 17-53.

Craig, H., Gordon, L. I., and Horibe, Y., 1963, Isotopic exchange effects in the evaporation of water, 1, Low-temperature experimental results: Jour. Geophys. Res., v. 68 p. 5979-5087.

Deines, P. and Gold, D. P., 1973, The isotopic composition of carbonatite and kimberlite carbonates and their bearing on the isotopic composition of deep-seated carbon: Geochim. Cosmochim. Acta, v. 37, p. 1709-1733.

Epstein, S. and Mayeda, T. K., 1953, Variations of oxygen-18 content of waters from natural sources: Geochim. cosmochim. Acta, v. 4, p. 213-224.

Eslinger, E. V., and Savin, S. M., 1973, Mineralogy and oxygen isotope geochemistry of the hydrothermally altered rocks of the Ohaki-Broadlands, New Zealand Geothermal Area: Am. Jour. Sci., v. 273, p. 240-267.

Faure, G., 1977, Principles of isotope geology, John Wiley and Sons, New York, p. 323-349.

Friedman, I., 1953, Deuterium content of natural water and other substances: Geochim. Cosmochim. Acta, v. 4, p. 89-103.

Friedman, I. and O'Neil, J. R., 1977, Compilation of stable isotope fractionation factors of geochemical interest, U.S. Geol. Surv. Prof. Pap., 440-KK.

Giggenbach, W. F., 1978, The isotopic composition of waters from the El Tatio geothermal field, northern Chile: Geochim. Cosmochim. Acta, v. 42, p. 979-988.

Glenn, W. E. and Hulen, J. B., 1979, Interpretation of well log data from four drill holes at Roosevelt Hot Springs KGRA: University of Utah Research Inst., Earth Science Laboratory Report 28, DOE/DGE contract EG-78-C-07-1701, 74 p.

Haas, J. L., Jr., 1971, The effect of salinity on the maximum thermal gradient of a hydrothermal system at hydrostatic pressure: Econ. Geol., v. 66, p. 940-946.

Helgeson, H. C., 1969, Thermodynamics of hydrothermal systems at elevated temperatures and pressures: Am. Jour. Sci., v. 267, p. 729-804.

Hulen, J. L., 1979, Preliminary lithology and alteration log, drill hole Utah State 9-1, University of Utah Research Inst., Earth Science Laboratory.

Lawrence, J. R. and Taylor, H. P., Jr., 1971, Deuterium and oxygen correlation: Clay minerals and hydroxides in Quaternary soils compared to meteoric waters: Geochim. Cosmochim. Acta, v. 36, p. 1377-1393.

Lipman, P. W., Rowley, P. D., Mehnert, H. H., Evans, S. H., Jr., Nash, W. P., and Brown, F. H., 1978, Pleistocene rhyolite of the Mineral Mountains, Utah geological and archaeological significance: Jour. Res. U.S.G.S., v. 6, p. 133-147.

McCrea, J. M., 1950, On the isotope chemistry of carbonates and a paleotemperature scale: Jour. Chem. Phys., v. 18, p. 849-857.

McKenzie, W. F. and Truesdell, A. H., 1977, Geothermal reservoir temperatures estimated from the oxygen isotope compositions of dissolved sulfate and water from hot springs and shallow drill holes: Geothermics, v. 5, p. 51-61.

Mundorf, J. C., 1970, Major thermal springs of Utah: Utah Geol. and Mineral Survey Water Resources Bull. 13, 60 p.

Nielson, D. L., 1979, Preliminary geological and geophysical logs for Utah State Geothermal Well 9-1.

Nielson, D. L., Sibbett, B. S., McKinney, D. B., Hulen, J. B., Moore, J. N., and Samberg, S. M., 1978, Geology of Roosevelt Hot Springs KGRA, Beaver County, Utah: University of Utah Research Inst., Earth Science Laboratory Report 12, DOE/DGE contract EG-78-C-07-1701, 120 p.

Ohmoto, H., 1972, Systematics of sulfur and carbon isotopes in hydrothermal ore deposits: Econ. Geol, v. 67, p. 551-578.

Ohmoto, H. and Rye, R. O., 1979, Isotopes of sulfur and carbon, in

Barnes, H. L., Geochemistry of hydrothermal ore deposits, 2nd edition, John Wiley and Sons, New York, p. 501-567.

O'Neil, J. R., Adams, L. H., and Epstein, S., 1975, Revised value for the oxygen-18 fractionation between CO_2 and H_2O at 25°C: Jour. Res. U.S.G.S., v. 3, no. 5, p. 623-624.

O'Neil, J. R. and Kharaka, Y. K., 1976, Hydrogen and oxygen isotope exchange reactions between clay minerals and water: Geochim. Cosmochim. Acta, v. 40, p. 241-246.

O'Neil, J. R. and Taylor, H. P., Jr., 1967, The oxygen isotope and cation exchange chemistry of feldspars: Am. Mineralogist, v. 52, p. 1414-1437.

__________, 1969, Oxygen isotope fractionation between muscovite and water: Jour. Geophys. Res., v. 74, p. 6012-6022.

Park, G. M., 1971, Volcanics, Thomas Range, in Radioactive and isotopic age determinations of Utah rocks: Utah Geol. and Mineral Survey Bull. 81.

Parry, W. T., 1978, Hydrothermal alteration at the Roosevelt Hot Springs Thermal Area, Utah: Part I, Mineralogy of the clay fraction from cuttings, Thermal Power Co., Well Utah State 14-2, Technical Report 78-1701.a.1.1.2, DOE/DGE, contract EG-78-C-07-1701.

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

Presser, T. S. and Barnes, I., 1974, Special techniques for determining chemical properties of geothermal water: U.S.G.S. Water Resources Investigations, 22-74, 11 p.

Rohrs, D. and Parry, W. T., 1978, Hydrothermal alteration at the Roosevelt Hot Springs thermal area, Utah: Thermal Power Co. Well Utah State 72-16: Topical report 78-1701.a.1.1.3, DOE/DGE contract EG-78-C-07-1701, 23 p.

Sakai, H., 1977, Sulfate - water isotope thermometry applied to geothermal systems: Geothermics, v. 5, p. 67-74.

Savin, S. M. and Epstein, S., 1970, The oxygen and hydrogen isotope geochemistry of clay minerals: Geochim. Cosmochim. Acta, v. 34, p. 25-42.

Stumm, W. and Morgan, J. J., 1970, Aquatic Chemistry, John Wiley and Sons, New York, 583 p.

Suzuoki, T. and Epstein, S., 1976, Hydrogen isotope fractionation between OH-bearing minerals and water: Geochim. Cosmochim. Acta, v. 40, p. 1229-1240.

Taylor, H. P., Jr. 1968, The oxygen isotope geochemistry of igneous rocks: Contrib. Mineral Petrol., v. 19, p. 1-71.

__________, 1974, The application of oxygen and hydrogen isotope studies to hydrothermal alteration and ore deposition: Econ. Geol., v. 69, p. 843-883.

__________, 1978, Oxygen and hydrogen isotope studies of plutonic granitic rocks: Earth Plan. Sci. Lett., v. 38, p. 177-210.

Taylor, H. P., Jr. and Epstein, S., 1962, The relationship between $^{18}O/^{16}O$ ratios in coexisting minerals of igneous and metamorphic rocks: G.S.A. Bull., v. 73, p. 461-480.

Truesdell, A. H., Nathenson, M., and Rye, R. O., 1977, The effects of subsurface boiling and dilution on the isotopic compositions of Yellowstone thermal waters: Jour. Geophys. Res., v. 82, p. 3694-3704.

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

White, D. E., 1957, magmatic, connate, and metamorphic waters: G.S.A. Bull., v. 68, p. 1659-1682.

__________, 1970, Geochemistry applied to discovery, evaluation, and exploitation of geothermal energy resources: Rapporteur's rep., U.N. Synpos. on Development and Utilization of Geothermal Resources, Pisa, Proc., (Geothermics, spec. iss. 2) v. 1, p. 58-80.

Yeh, H. W. and Epstein, S., 1978, Hydrogen isotope exchange between clay minerals and sea water: Geochim. Cosmochim. Acta, v. 42, p. 140-143.