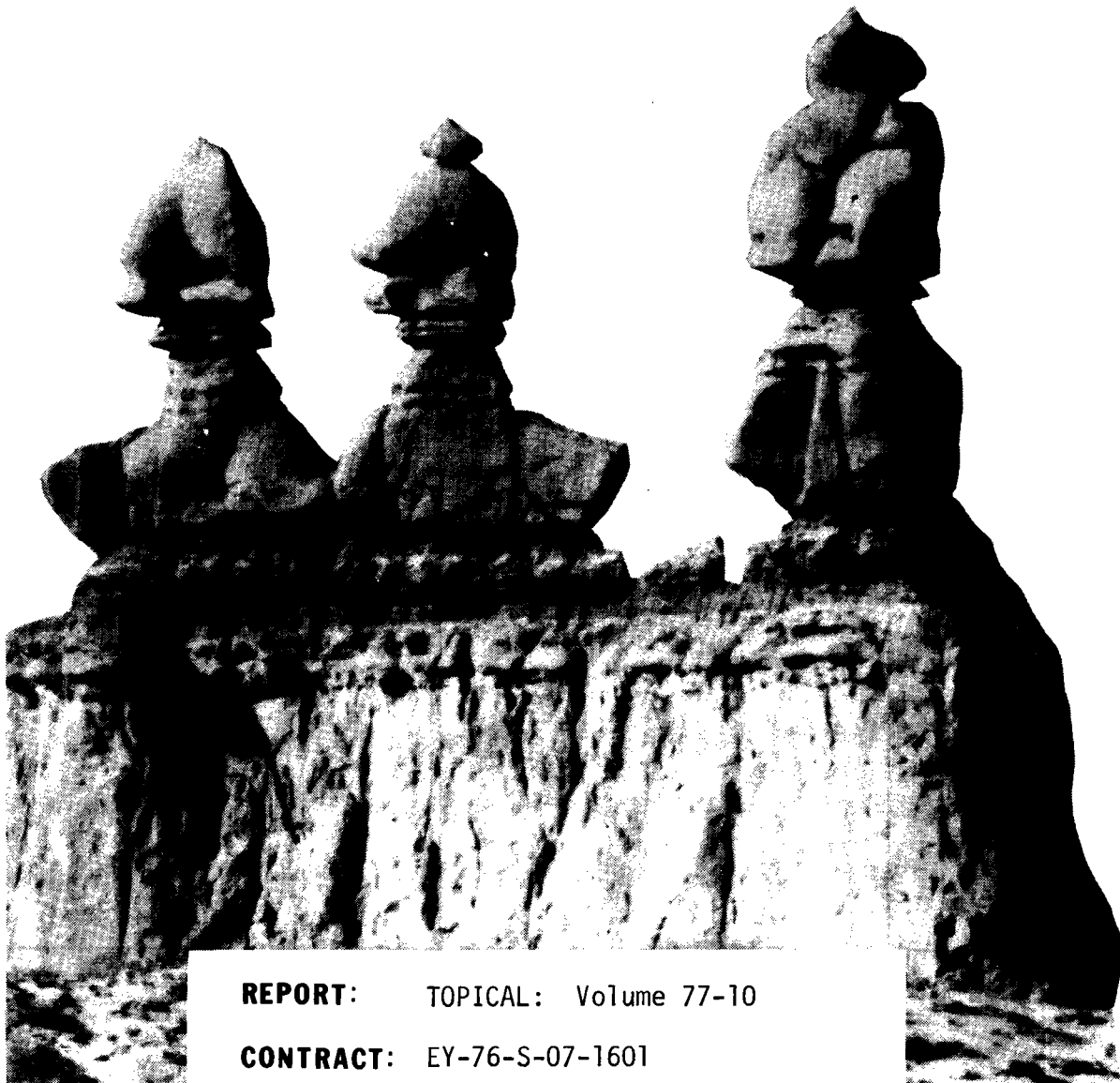


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Stable Isotope Investigation of Fluids and Water-Rock Interaction
in the Roosevelt Hot Springs Thermal Area, Utah

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by

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Abstract

Carbon-hydrogen-oxygen isotope compositions have been measured in regional cold waters, geothermal fluids, and hydrothermally altered rocks from the Roosevelt Hot Springs geothermal area. These data have been used, in conjunction with other geological and geochemical data from this geothermal system, to place some limits on the origin of geothermal fluids and reservoir carbon, the fluid recharge area, physical-chemical environment of hydrothermal alteration, and relative permeability of the geothermal system. The similarity of hydrogen isotope compositions of local meteoric water and geothermal reservoir fluid indicate that the geothermal fluids are virtually entirely of surface derivation. An isotopically reasonable source area would be the Mineral Mountains directly to the east of the Roosevelt system. Hydrothermal calcite appears to be in isotopic equilibrium with the deep reservoir fluid. The $\delta^{13}\text{C}$ values of deep calcites and T- pH- $f\text{O}_2$ conditions of the reservoir defined by measured temperature, fluid chemistry, and alteration mineralogy fix the $\delta^{13}\text{C}$ value of the geothermal system to -5 to -6.5 ‰ (PDB). These values do not unambiguously define any one source or process, however. There is a relatively small increase in ^{18}O of geothermal fluids relative to their cold surface water precursors and significant ^{18}O depletion accompanying hydrothermal alteration of the granitic host rock. These isotopic shifts indicate a high ratio of geothermal fluid to altered rock for the geothermal system, implying relatively rapid (geologically) recirculation rates and significant permeability of the geothermal system.

Introduction

Stable C-H-O isotope compositions of solid and fluid geothermal samples and cold spring waters have been measured in the Roosevelt Hot Springs Geothermal Area, and surrounding region. Geothermal samples were from drill-cuttings and flow tests from Thermal Power Company wells 14-2 and 72-16. Well locations are respectively, 1.6 km southeast of the old resort, and 4.3 km southwest of the old resort. Isotope analyses reported here were measured at both the USGS, Menlo Park, and University of Utah, Salt Lake City.

Sample Selection and Preparation

Solid Samples - All solid samples are rotary drill cuttings. Because the pre-cursor rock of the geothermal system was granite, and because cuttings samples cannot preserve textural information regarding paragenesis and multiple generations of a given mineral, all cuttings bulk rock and mineral separate samples may be composed in part of pre-geothermal material. Because of the lack of core sample, it is not possible to distinguish multiple generations of the common minerals involved: quartz, potassium feldspar, chlorite. The isotope data should be viewed in this light. It should be emphasized that systematic core sampling is needed to resolve this and a host of other problems involving mineral paragenesis, fracture density, etc.

Cuttings samples of 25 to 150 cm³ were crushed to -100 to -140 and -140 to -200 mesh sizes for mineral separations. Silicate mineral separates of >96% purity for oxygen isotopic analysis were achieved by standard heavy liquid and magnetic separation techniques; hand-picking was done when permitted by sufficiently coarse grain size. Quartz separates were purified by cold HF treatment to remove feldspar contamination. Carbonate contamination was removed by treatment with cold dilute HCl acid. Carbonate samples for C and O isotopic analysis were not separated from the silicate fraction (other than to concentrate the necessary quantities of carbonate mineral) because the phosphoric acid used in the extraction procedure (see next section) does not interact with the silicate oxygen.

Spring Samples - Cold water samples from springs and seeps in the vicinity of Roosevelt Hot Springs were collected by both the USGS and

University of Utah. Spring collection was concentrated in the Mineral Mountains directly east of the geothermal area, as it is the logical source area of the geothermal fluids. Standard water collection techniques were used.

Geothermal Wells - Reservoir fluids were sampled in the geothermal area during flow-tests conducted by Thermal Power Company on wells no. 14-2 (November, 1976; May, 1978) and no. 72-16 (April, 1977). In order to obtain representative samples of discharged fluids, a steam/fluid miniseparator of New Zealand design was used for these collections. The November, 1976 test was successfully sampled by Dr. Al Truesdell of the USGS, Menlo Park. The quality of collection of fluids at the April, 1977 test suffered due to premature termination of the well-test before the miniseparator could be adjusted. The quality of collection of fluids at the May, 1978 test also suffered because periodic changes in flow pressures in conjunction with other experiments conducted by Thermal Power and Battelle Institute precluded accurate adjustment of the miniseparator.

Analytical Techniques

Carbonate Oxygen and Carbon Extraction - The phosphoric-acid technique (McCrea, 1950) was used for CO₂ extraction from the carbonate samples. A kinetic fractionation factor of $\alpha = 1.01025$ (Sharma and Clayton, 1965) for the acid reaction at 25°C was used for all calcite extractions.

Silicates Oxygen Extraction - Oxygen extraction were performed by reacting samples ground to -100 to -200 mesh with BrF₅ at 550-650°C in nickel reaction vessels for 12-14 hours as described by Clayton and Mayeda (1963). The evolved O₂ gas was then converted to CO₂ for mass spectrometric analysis by combustion with graphite by the methods of Taylor and Epstein (1962).

Water Oxygen Extraction - The ¹⁸O/¹⁶O ratios in cold water and geothermal fluid samples were determined by the CO₂ equilibration technique, described by Epstein and Mayeda (1953). The CO₂-H₂O oxygen fractionation factor used was that of O'Neil, et al. (1975): at 25°C, $\alpha_{\text{CO}_2\text{-H}_2\text{O}} = 1.0412$.

Water Hydrogen Extraction - Hydrogen gas used for D/H ratio measurement was liberated from water samples by reduction over hot (800°C) uranium metal by methods similar to those described by Friedman (1953).

Mass Spectrometry - Analytical error for carbon and oxygen isotope ratios is between .1 and .2 per mil, while that for hydrogen is between 1 and 2 per mil. The isotopic data for H and O and for C are reported relative to the Standard Mean Ocean Water (SMOW) as defined by Craig (1961a) and Clayton and Mayeda (1963) and the Chicago PDB standard,

respectively.

Notation - All isotopic data are reported in the usual notation,
where:

$$\delta R_{\text{sample}} = \frac{R_{\text{sample}} - R_{\text{std}}}{R_{\text{std}}} \times 1000$$

where R_{sample} represent the D/H, $^{13}\text{C}/^{12}\text{C}$, or $^{18}\text{O}/^{16}\text{O}$ isotope ratio in the sample and standard, respectively.

Results and Discussion

Carbonates - Calcites were analysed for their oxygen isotopic compositions to determine whether calcite-water oxygen isotope fractionations correspond to equilibrium at present reservoir temperatures. C isotope analyses were also made to determine the carbon isotope composition of the geothermal system in order to identify the source(s) of carbon. Weight percent calcite, depth, oxygen and carbon isotope compositions, calcite-water isotope temperatures, and measured borehole temperatures are presented in Table 1 for calcite samples from Thermal Power Company wells 14-2 and 72-16. The calcite-water temperatures were calculated assuming a constant reservoir fluid value of -13.7%, determined by Dr. Al Truesdell of the USGS, Menlo Park. The calcite-water equilibrium values from O'Neil et al. (1969) were used for the temperature calculations:

$$10^3 \ln \alpha_{\text{Cc-H}_2\text{O}} = 2.78 (10^6 T^{-2}) - 3.39$$

The calcite data are arranged in Table 1 in order of increasing depth, where the last part of the sample number corresponds to the depth in feet. For both wells, there is a decrease in $\delta^{18}\text{O}$ of the calcite with increasing depth. This trend reflects the increase in temperature with depth, since the fractionation factor $10^3 \ln \alpha$ between calcite and water decreases with increasing temperature. There is good agreement between measured and calcite-water temperatures in well 14-2 at depths greater than 2800 feet and in the deep sample of well 72-16. These deep reservoir calcite-water temperatures of up to 279°C agree within 30° of measured temperatures of

Table 1
 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ Analyses of Geothermal
 Carbonates

Sample	wt. %	$\delta^{18}\text{O}/\text{‰}$	$\delta^{13}\text{C}$	*6 Cc-H ₂ O	T(°C) Cc-H ₂ O	T Meas
14-2-650	1.1	-2.9	-3.5	10.8	169°	80
14-2-940	0.90	-3.8	-2.9	9.9	184°	110
14-2-2865	1.20	-6.3	-3.7	7.4	224°	200
14-2-4450	2.50	-8.0	-3.6	5.7	270°	240
14-2-5200	3.10	-8.3	-4.2	5.4	279°	250
72-16-600	1.3	-3.6	-4.4	10.1	170°	
72-16-810	0.9	-6.8	-4.1	6.9	236°	
72-16-960	0.9	-8.2	-4.7	5.5	276°	

*Assuming $\delta^{18}\text{O}$ of water = -13.7‰ for all samples.

250°C. Further, silica and K-Ca-Na thermometry results (refer to final report, vol. 3, contract no. GI-43741; Parry et al., 1976) from both the Roosevelt seep (240°C) and deep reservoir fluids (260°C) also agree well with the isotope temperatures. Thus in the deep parts of the geothermal reservoir, there appears to be a close approach to equilibrium between at least the calcite and geothermal fluids. However the calcite-water temperatures appear to be systematically high with respect to both the chemical and measured thermometry data. This may be the result of either minor departure from equilibrium or systematic error in the experimentally determined calcite-water fractionation. On the other hand, the calcites may be recording an earlier, higher temperature of the deep reservoir. A satisfactory explanation for this small systematic difference may not be possible, given the usual fluctuations in isotopic data from natural systems. Certainly, additional solid and fluid isotopic data are needed to more thoroughly evaluate the degree of attainment of isotopic equilibrium in the Roosevelt Geothermal system.

At shallow depths, the calcite-water temperatures are much higher than measured temperatures. It is possible that the reservoir fluid is not homogeneous, but becomes significantly lighter in the shallow parts of the wells. However the required δ^{18} value of water in equilibrium with calcite at 650 feet ($T = 80^\circ\text{C}$) would be -21.5% , far lighter than any measured regional or geothermal water at Roosevelt Hot Springs. It seems more likely that either these shallow calcites have not achieved isotope exchange equilibrium with the reservoir fluids at temperatures less than approximately 150°C or that these δ^{18} values were acquired at some earlier time when temperatures at these shallow depths were significantly higher.

Subsequent reequilibration to present temperatures would apparently have been prevented by slow exchange rates at temperatures of less than 150°C or so.

Carbon isotope compositions show little systematic variation, being uncorrelated with depth, temperature, amount of carbonate, or δ^{18} values of the calcite. Based on the experimental and thermochemical data of Hemley, et al., (1959) and Helgeson (1969), respectively, the presence of muscovite + potassium feldspar in equilibrium with a fluid of the chemistry of deep Roosevelt reservoir fluid at 250°C fixes the pH at close to 7.0 (W.T. Parry, personal communication). Log f_{O_2} of the deep reservoir fluid-rock system is calculated at -33 based on the observation of closely associated pyrite-magnetite-hematite and the equilibrium constant data of Helgeson (1969), (W.T. Parry, personal communication). At these T-pH- f_{O_2} conditions, the total carbon reservoir of the deep geothermal fluid in equilibrium with calcite would be in the form of subequal amounts of H_2CO_3 and HCO_3^- , as shown by calculations of Ohmoto (1972). For $[H_2CO_3]$ approximately equal to $[HCO_3^-]$, the carbon reservoir should be about 1.5% depleted relative to calcite at 250°C (Bottinga, 1968; Ohmoto, 1972), or -5.5%. This value is at the heavier end of the range of $\delta^{13}C$ values thought to be characteristic of deep-seated or "juvenile" carbon sources. However, it is unlikely that carbon-bearing species would be contributed directly to the geothermal reservoir by magmatic fluids from the associated igneous rocks. Rather, the isotopic similarity of geothermal and typical igneous carbon may merely reflect fluid circulation restricted to igneous rocks. Such restriction would avoid contact with subsurface sedimentary units and thus avoid isotopic exchange with ^{13}C -depleted hydrocarbon

materials and/or marine carbonate carbon. At this point, it is not possible to rule out more complicated origins for the C reservoir in the Roosevelt System involving partial oxidation of ^{13}C -depleted hydrocarbon materials, isotopic exchange with fresh-water sedimentary carbonates, or mixing of carbon from ^{13}C -depleted organic materials with aqueous HCO_3^- equilibrated at surface temperatures with atmospheric CO_2 . Due to the problems of geothermal fluid collection, we do not yet have a reliable sample of total dissolved HCO_3^- with which cross-check this calculated value.

Water - Hydrogen and oxygen isotope analyses of regional spring waters and reservoir fluids are compiled in Table 2. Analyses were performed at the USGS, Menlo Park under the direction of Dr. Al Truesdell and at the University of Utah. Whole fluid δO^{18} has been calculated by Dr. Al Truesdell from enthalpy and isotope balances, based on isotopic analyses of steam and water fractions and P-T measurements for the flow tests. Comparison of δD - $\delta^{18}\text{O}$ values for the deep geothermal fluid with averages of spring samples from the Mineral Mountains is useful. The similarity of δD values for geothermal fluid (-117‰) and average Mineral Mountain spring water (-114‰) indicates that the Mineral Mountains is a possible source region for the geothermal fluids. However the hydrogen isotope data do not preclude source areas further east in the Tushar Mountains, as the δD data from Tushar springs are also close to that of the Roosevelt geothermal system. Hydrologic flow line data and isotopic analyses of snow pack in the two ranges are needed before a more definite estimate can be made of the recharge area and thus of the extent of the aquifer.

Table 2. Stable isotope analyses of waters of the RHS area

	Date Collected	$\delta D, \text{‰}$	$\delta^{18}O, \text{‰}$
Ranch Canyon (Kirk) Spring	2/76	-112.5	-15.02
	8/76	-110.4	-14.2
Bailey Spring	2/76	-116.0	-15.55
	8/76	-111.3	-14.6
		(-117)	(-14.9)
Milford City well	2/76	-116.5	-15.5
Pearson well	2/76	-121.8	-15.5
S of Milford airport	8/76	-123.9	-15.8
Mud spr. Mineral range	8/76	-113.4	-14.8
		(-115)	(-14.7)
Beaver Lake Mine spr.	8/76	-103.8	-12.5
Cowboy spr.	8/76	-118.3	-15.2
		(-117)	(-15.3)
Rock Corral spr.	8/76	-108.9	-14.2
		(-110)	(-14.4)
Antelope spr.	8/76	-116.2	-15.1
Four Mile spr.	8/76	-11.39	-14.2
Armstrong spr.	8/76	-107.0	-13.7
Sulfurdale N. spr.	8/76	-118.4	-15.6
Mud spr.	8/76	-125.7	-16.5
Wiregrass spr.	8/76	-115.2	-15.2
		(-117)	(-15.1)
Magpie Cr. spr.	8/76	-120.3	-15.6
Sulfurdale S. spr.	8/76	-116.4	-15.6
North spr.	8/76	-112.3	-14.7
Griffith spr.	8/76	-109.6	-14.4
North Ck. spr.	8/76	-119.8	-15.0
Cherry Ck. spr.	8/76	-112.5	-14.7
Dead Cow spr.	8/76	-114.1	-14.6
Willow spr.	8/76	-113.6	-14.8
		(-114)	(-15.0)
Thermal Power well 14-2	3/77	(-117)	-13.71

() Analyses by the University of Utah

The $\delta^{18}\text{O}$ value of deep geothermal fluid (-13.7 ‰) is only slightly heavier than that of surface meteoric and spring waters (ave. = -14.9 ‰). This is a rather small oxygen isotope shift when compared to other geothermal systems such as Salton Sea geothermal area (Clayton et al., 1968), but is similar to that observed by Craig (1963) and Clayton and Steiner (1975) for waters at Wairakei, New Zealand. These authors concluded that the small shift indicated direct derivation of thermal fluid from local meteoric water, and reflected relatively rapid circulation times. The small oxygen isotope shift for Roosevelt thermal water is also accompanied by extensive hydrothermal alteration in drill-cuttings and substantial oxygen isotope exchange of the rock (see following silicate section), implying that the ratio of the mass of water circulated through the system to mass of hydrothermally altered rock is relatively high, significantly higher than that at the Salton Sea geothermal field.

Silicates - A wholerock sample and constituent quartz and potassium feldspar from level 2860, well 14-2, has been analyzed for the oxygen isotope composition in order to determine the extent of oxygen isotope exchange in the deep portions of Roosevelt geothermal system. These values:

	$\delta^{18}\text{O}$
whole rock	+4.5
quartz	+9.7
feldspar	+2.1

indicate that the hydrothermally altered rocks have undergone significant exchange with the geothermal fluids, as the whole rock has been depleted by

approximately 3 to 3.5‰ relative to δ^{18} values of the unaltered Mineral Mountains quartz monzonite precursor. However, the reservoir host rock has not completely exchanged isotopically with the thermal fluid. Further, feldspar has exchanged to a much greater degree than quartz and is not in isotopic equilibrium with quartz. This of course is a commonly observed phenomenon in hydrothermally altered and partially exchanged rocks.

Given this data, an estimate of the minimum water-to-rock oxygen ratio in the geothermal system can be made. Given observed shift in O^{18} of water from -15‰ to -13.7‰ and of rock from +8.0‰ to +4.5‰, then a lower limit to the mass ratio of water-to-rock oxygen is calculated to be approximately 2.3. Converted to a water:rock mass ratio, a value of 1.3 results, several times the value calculated by Clayton, et al., (1968) for the Salton Sea geothermal area. Calculation utilizing the feldspar, which has more extensively exchanged with the reservoir water, would result in an even higher ratio.

The large ratio of water to host rock in the Roosevelt geothermal system implies relatively high permeability and circulation rates. The thoroughly fractured nature of the host rock, concentration of hydrothermal alteration along fractures, and the oxygen isotope data support a model of fluid flow controlled by a pervasive system of open fractures, since the inherent permeability of crystalline granitic rock must be quite low. Such a model of fluid flow suggests that the reservoir permeability and hence flow rate, will be sensitively controlled by volume changes accompanying hydrothermal alteration. Dramatic changes in hydrothermal alteration from that operative in the geologic past, due either to steam production or long-term geologic evolution of the field, may modify significantly

fracture permeability, thus altering dramatically the production capabilities of the field.

Conclusions

The stable isotope data collected to date, in conjunction with available geological and geochemical data, permit the following preliminary conclusions:

1) The geothermal fluids at Roosevelt Hot Springs have originated rather directly from local surface or meteoric, water. The Mineral Mountains, directly to the east of the area, is the probable recharge area.

2) The hot water-rock system has approached isotopic equilibrium closely only at temperatures exceeding approximately 150°C.

3) T-pH-fo₂ conditions for the geothermal system define the $\delta^{13}\text{C}$ value of total carbon in the deep reservoir to be approximately -5.5%(PDB). Similarity between this value and that of deep-seated or "juvenile" carbon does not prove contribution of carbon from magmatic fluids associated with recent rhyolite volcanism, but suggests restricted circulation to igneous rock types (a carbon reservoir defined by juvenile or magmatic carbon). The carbon isotope data are also consistent with origins involving partial oxidation of organic debris or hydrocarbon materials or exchange with fresh-water sedimentary carbonates.

4) Slight ¹⁸O enrichment in the geothermal fluid relative to surface water, coupled with significant ¹⁸O depletion (approximately 3.5‰) associated with hydrothermal alteration of the igneous host rock indicates a relatively high mass ratio of geothermal water to altered host rock over the geologic history of the system. This implies relatively high permeability within the system and relatively rapid circulation of cold water into the reservoir. The data suggest that fluid discharge and

recharge should be relatively high and maintainable, if the hydrothermal alteration processes characteristic of the system in its past can be preserved during exploitation.

Further Work

Further work should entail more detailed analyses of the geothermal fluids themselves (especially from individual production zones), a three-dimensional analysis of the Roosevelt Hot Springs as a hot water/rock system, and comparisons with other geothermal systems. The objective is to construct a more complete, process-oriented model of geothermal systems as geochemical systems of hot water-rock interaction. Isotopic analysis of C, O, and S-bearing species (SO_4^{2-} , H_2S , HCO_3^-) in deep fluids, cold springs and surface seeps should provide information concerning fluid mixing and processes of fluid recharge. However, definition in detail of physical-chemical conditions (and especially gradients in these parameters) within the geothermal reservoir--upon which process-oriented models of hot fluid-rock interaction, fluid circulation, and hot water/cold water mixing depend--is dependent on the availability of extensive and systematically collected core samples and on the sampling of fluid from specific production zones. Without these types of samples, we cannot, for example, field check theoretical mass transfer models or be assured of having chosen the important physical-chemical variables for such models. In short, without better solid and fluid samples, we cannot significantly improve the geochemical model of geothermal systems.

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