1 Oxygen Isotope Systematics in an Evolving Geothermal System: Coso Hot Springs, California

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11 ABSTRACT
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13 Oxygen isotope and clay mineralogy studies have been made on whole rock samples and
14 feldspar separates from three wells along the high temperature West Flank of the Coso
15 geothermal system, California. The reservoir rocks have experienced variable ¹⁸O/¹⁶O depletion,
16 with $\delta^{18}\text{O}$ values ranging from primary values of +7.5 ‰ down to -4.6 ‰. Spatial patterns of
17 clay mineral distributions in the three wells are not closely correlated with the distributions
18 expected from measured, pre-production temperature profiles, but do correlate with spatial
19 patterns of ¹⁸O/¹⁶O depletion, indicating that the stability of clay minerals in the three wells is a
20 function of fluid-rock interaction in addition to temperature. Detailed $\delta^{18}\text{O}$ measurements in the
21 three wells identify a limited number of localized intervals of extensive ¹⁸O/¹⁶O depletion. These
22 intervals document localized zones of higher permeability in the geothermal system that have
23 experienced significant fluid infiltration, water-rock interaction and oxygen isotopic exchange
24 with the geothermal fluids. The local zones of maximum ¹⁸O/¹⁶O depletion in each well
25 correspond closely with current hot water production zones. Most feldspar separates have
26 measured $\delta^{18}\text{O}$ values too high to have completely attained oxygen isotope exchange equilibrium
27 with the reservoir fluid at pre-production temperatures. In general, the lower the $\delta^{18}\text{O}$ value of
28 the feldspar, the closer the feldspar approaches exchange equilibrium with the geothermal fluid.
29 This correlation suggests that fracture-induced increases in permeability increase both fluid

infiltration and the surface area of the host rock exposed to geothermal fluid, promoting fluid-rock interaction and oxygen isotope exchange. The two most $^{18}\text{O}/^{16}\text{O}$ -depleted feldspar samples have $\delta^{18}\text{O}$ values too low to be in exchange equilibrium with the pre-production reservoir fluid at pre-production temperatures. These discrepancies suggest that the reservoir fluid in the West Flank of the Coso geothermal system was hotter and/or had a lower $\delta^{18}\text{O}$ value (due to fluid-rock interaction at higher permeability) in the past.

INTRODUCTION

The Coso Hot Springs Geothermal Field is located in China Lake, California 10 km east of the Sierra Nevada Mountains at the western boundary of the Basin and Range Province (fig. 1). Coso Hot Springs is a world-class geothermal field; it is the third largest producing geothermal field in the United States, among the hottest, with an installed power capacity of 273 MWe and temperatures as high as 350°C at depths as shallow as ~1 km. It is also one of the most extensively drilled geothermal systems, with many of the more than 150 wells reaching depths of 2-3 km.

The level of power production at Coso Hot Springs implies significant permeability in the reservoir rocks, which consist dominantly of metamorphosed Mesozoic diorite and younger granodiorite and granite. Although epidote, chlorite, illite, quartz, and calcite, which are typical of high-temperature geothermal systems worldwide (Henley and Ellis, 1985), are widespread at Coso Hot Springs, these minerals show no clear correlation with temperature (Bishop and Bird, 1987; Lutz et al., 1996; Lutz and Moore, 1997; Kovac et al., 2004; Kovac et al., 2005). In contrast, minerals typical of moderate temperature environments, such as wairakite and prehnite, are rarely observed in the reservoir rocks.

Figure 1: A simplified geologic map of the Coso Geothermal system (modified from Hulen, 1978) located on the Naval Air Weapons Station, China Lake, California. The West Flank of the system is defined by a North-South trending thermal high (Terra-Gen, unpublished data); it is the focus of this study. Line A-A' along this thermal high is the transect for a number of cross sections used in this study. Line B-B' is a cross section line through wells 68-6 and 33A-7. Locations of wells in the West Flank used in this study are marked by red and green circles; the green circles indicate wells examined in detail for this study while the red circles indicate wells examined in previous studies. Only the wells in the West Flank with oxygen isotope data for host rocks are shown; none are shown for other areas.

Oxygen isotopes provide an alternative means of mapping subtle changes in the extent and distribution of fluid-rock interactions that are independent of the bulk mineralogy. Active geothermal systems provide unique opportunities to examine the interaction between geothermal fluids and host rocks. Person et al. (2006) used heat flow data and $\delta^{18}\text{O}$ values of reservoir fluids to constrain a regional-scale numerical model of the flow system supplying the Coso geothermal system. Recently, Pope et al. (2016) measured δD and $\delta^{18}\text{O}$ values of hydrothermal epidote from the Krafla, Iceland geothermal system. They interpreted variations in the δD and $\delta^{18}\text{O}$ values of both the epidote and reservoir fluids to result from variable mixing of shallow, sub-boiling groundwaters with vapor condensates derived from an underlying two-phase reservoir, also derived from local meteoric waters. Here we report on oxygen isotope analyses of whole-rock samples from the Coso geothermal system, and detailed whole-rock and mineral analyses from three producing wells. Our whole-rock results define $^{18}\text{O}/^{16}\text{O}$ depletion patterns in the geothermal system distinct from those expected from regional metamorphism whereas the $^{18}\text{O}/^{16}\text{O}$ patterns from the three wells identify discrete intervals of maximum $^{18}\text{O}/^{16}\text{O}$ depletion that correspond closely with zones of current hot water production and high permeability. These discrete, higher permeability zones have experienced significant fluid infiltration, water-rock interaction and oxygen isotopic exchange with the geothermal fluids. These results document a correlation between the extent of oxygen isotope exchange (e.g., extent of $^{18}\text{O}/^{16}\text{O}$ depletion) and fluid-rock interaction recorded in the host rocks with host rock permeability. The results of this study have broader applications to fossil hydrothermal systems and Enhanced Geothermal Systems developed in crystalline host rocks.

Geologic Setting

The Coso Hot Springs geothermal system is developed in Mesozoic plutonic rocks of the Sierra Nevada Batholith (Duffield et al., 1980). Diorite that was regionally metamorphosed at greenschist to lower amphibolite facies conditions to orthogneiss in the late Cretaceous (Bartley et al., 2007) (fig. 1) is the most common rock type. Younger intrusions consist of granodiorite and granite (Duffield et al., 1980). Subsequent episodes of Cenozoic volcanism produced basalt and rhyolite at 4.0-2.5 Ma and 1.1-0.044 Ma (Kurliovitch et al., 2003; Duffield et al., 1980). The youngest episode is related to a silicic magma chamber emplaced between 5-20 km depth (Duffield et al., 1980; Reasenberget al., 1980; Manley and Bacon, 2000). This magma body produces the heat driving the current geothermal activity (Duffield et al., 1980).

Surface expressions of geothermal activity include opaline (SiO_2) sinter and travertine (CaCO_3) spring deposits. ^{14}C dating of pollen in hot spring deposits has yielded ages of 11,550 to 8,550 years (Moore, unpublished data). The geothermal fluids (Table 1) have low salinities (5,000-10,000 ppm TDS), and are NaCl-dominated. The pre-production geothermal waters are dominated by a single-phase system with minimal boiling (Fournier et al., 1980), but there is some mineral alteration evidence for a natural steam cap in places (Williams and McKibbin, 1990). Geochemical and hydrogen isotope data indicate that the system is predominately recharged by meteoric fluids (Adams et al., 2000; Moore et al., 1990; Williams and McKibbin, 1990; unpublished results, Terra-Gen Operating Co.). The modeling results of Person et al. (2006) suggest that the geothermal fluids are recharged by lateral flow of groundwaters (meteoric waters) from fault systems and possibly the Sierra Nevada mountains to the west of the Coso field. Their model has these groundwaters infiltrating to the top of the brittle-ductile transition at 4-5 km depth (Wilson et al., 2003), becoming enriched in $^{18}\text{O}/^{16}\text{O}$ to the maximum measured $\delta^{18}\text{O}$ values of the geothermal reservoir fluids (-6 ‰) through fluid-rock isotopic

Table 1: The composition of pre-production geothermal fluids from wells 68-6, 33A-7 and 73-19 (Terra-Gen, unpublished data). Chemistry in ppm by weight; $\delta^{18}\text{O}$ and δD in per mil (‰) relative to VSMOW.

	68-6	33A-7	73-19
Na	748.7	1273.1	2043.9
K	136.0	187.4	579.3
Ca	7.4	6.8	49.7
Li	6.5	10.7	26.9
Mg	0.3	N/A	N/A
B	39.4	57.6	88.7
SiO ₂	596.3	514.4	601.9
HCO ₃	102.1	123.6	44.4
SO ₄	30.4	40.1	15.7
Cl	1259.9	2163.0	3783.4
F	2.6	2.6	2.3
$\delta^{18}\text{O}_\text{w}$	-8.46 ‰	-7.49 ‰	-6.04 ‰
δD	-104.1 ‰	N/A	-96.0 ‰
Date Measured	6/26/1995	7/28/2009	10/26/1988

exchange at elevated temperature, and then ascending to form the thermal plume of the Coso reservoir. Northerly trending, Basin and Range faulting and local dextral strike-slip faulting associated with the younger Walker Lane/Eastern California Shear Zone, guide fluid flow and influence current geothermal activity (Monastero et al., 2005; Kaven et al., 2012). Additional details about the geology, including hydrothermal alteration, are provided by Hulen (1978), Duffield et al. (1980), Lutz et al. (1996), Adams et al. (2000), Manley and Bacon (2000), Kovac et al. (2005) and Etzel (2015).

Thermal Structure

Borehole measurements have defined a north-south trending zone of high temperatures on the West Flank of the field (fig. 1). Geochemical and thermal data suggest that the system is compartmentalized with upflow zones in the northeast and southwest (Williams and McKibben, 1990). Figure 2 shows the thermal structure of the West Flank of the system based on pre-production temperature data. Maximum measured temperatures at depth reach 350°C. Reservoir

Figure 2: General north-south thermal structure of the West Flank along the transects A-A' and B-B'. Temperature values come from down-hole measurements made prior to production.

temperatures in the south are higher than in the north at equivalent elevations. The thermal pattern defines a plume of hot water that ascends in the south and then migrates laterally northward.

METHODS

Rock cuttings were sampled at regular intervals (nominally ~30 - 80 m apart) in wells 68-6, 33A-7 and 73-19 for oxygen isotope, X-ray diffraction (XRD) and petrographic analyses. Thin sections were examined using a petrographic microscope under transmitted light to identify primary and secondary (alteration) minerals, vein mineral assemblages and their paragenesis, structural features and the host rock lithology. Etzel (2015) provides a detailed presentation of these results.

Clay mineralogy and clay mineral abundances have been determined by standard techniques for preparing clay fractions (<2 μm size fraction) and standard x-ray diffractometry techniques (Lutz and Moore, 1997). The detection limit for clay minerals is 0.3 wt. %. Additional details of these techniques, clay mineralogy and of host rock lithology are reported in Etzel (2015).

Oxygen isotopic measurements were conducted at the University of Wisconsin-Madison on 91 whole rock samples and 30 feldspar separates from wells 68-6, 33A-7 and 73-19. Mineral separates were made using a Frantz magnetic separator followed by hand-picking under a stereoscopic microscope. The oxygen isotope analyses were done by laser-aided fluorination using a lasing sample chamber outfitted with an airlock sample chamber to prevent pre-fluorination of reactive rock powders (Spicuzza et al., 1998). Oxygen isotope values were standardized using UWG-2 garnet standard (Valley et al., 1995) and are reported in the standard $\delta^{18}\text{O}$ notation, relative to VSMOW.

RESULTS

Host Rock and Alteration Mineralogy

The reservoir rocks observed in wells 68-6, 33A-7 and 73-19, the three wells studied in detail (well locations shown in fig. 1) are principally diorite orthogneiss with subordinate diorite and granodiorite.

Two episodes of overprinting on the primary (igneous) mineralogy of the reservoir rocks are recognizable (Bishop and Bird, 1987; Lutz et al., 1996; Lutz and Moore, 1997; Kovac et al., 2005; Etzel, 2015). Regional lower amphibolite to greenschist facies metamorphism (late Cretaceous) has produced plagioclase, hornblende, biotite, epidote, titanite, chlorite and illite (muscovite) throughout all three wells (figs. 3, 4a, b). During subsequent geothermal activity, chlorite replaced hornblende and biotite, and epidote, illite (muscovite) and other clay minerals replaced plagioclase (fig. 4c-f). This geothermal activity has resulted in progressive changes in temperature sensitive clay minerals with depth. Throughout the field, smectite (S) is present at the shallowest depths, followed by interlayered illite/smectite (I/S) and rare chlorite/smectite, and finally illite and chlorite at deeper levels. Smectite, interlayered illite-smectite and illite replace plagioclase; chlorite and rare interlayered smectite/chlorite replace biotite and hornblende (fig. 4c-f).

The deepest occurrence of smectite above trace levels defines the base of the smectite zone at an elevation of about 1000 meters in wells 68-6 and 33A-7 (fig. 3). In well 73-19, smectite occurs only at trace levels (0.3 to 1%) and occurs persistently only at elevations above ~300 m (fig. 3). The deepest occurrence of interlayered illite-smectite above trace levels defines the base of the I/S zone in wells 33A-7 and 73-19 at elevations of -41 meters and -143 meters, respectively. Interlayered illite-smectite is completely absent in well 68-6. Trace amounts of the

Figure 3: Clay mineral content and measured pre-production temperature profiles for wells 68-6, 33A-7 and 73-19. I/S = interlayered illite-smectite; C/S = interlayered chlorite-smectite. Dashed lines denote locations at which clay transitions occur; measured temperatures at these locations are in parentheses.

Figure 4: Photomicrographs of the lithologies and alterations observed in the Coso geothermal system. Panels a) and b) show a typical unaltered meta-diorite sample with hornblende (Hbl), untwinned feldspar (Fsp), and rare anhedronal titanite (Ttn). This sample is from 33A-7_1100. Panels c) and d) show extensive geothermal-related alteration in feldspar, hornblende and biotite found in 68-6_2792. Feldspar is altered to illite (Ill); hornblende and biotite are completely altered to chlorite (Chl). Panels e) and f) show a feldspar partially altered to illite found in 68-6_2941. The field of view in panels a) – d) is 1.6 mm and in panels e) and f) is 0.8 mm.

lower temperature smectite and interlayered I/S clays exist within the illite zone in 33A-7 (fig. 3).

The transitions from smectite to interlayered illite/smectite to illite typically occur at temperatures of ~180°C and 225°C, respectively (Steiner, 1968; Browne, 1978; Henley and Ellis,

1983; Browne, 1984; Reyes, 1990). However in the three Coso wells, these transitions do not always occur at or near these temperatures, based on measured pre-production temperatures (fig. 3). In 33A-7, the smectite to interlayered illite/smectite and illite/smectite to illite transitions occur well below expected temperatures at ~125°C and ~210°C, respectively. In well 73-19, the smectite to interlayered illite/smectite transition occurs near the expected T of 180°C, but interlayered illite/smectite persists to depths where temperatures reach 280°C. In 68-6 smectite disappears near a pre-production temperature of 180°C (at an elevation of ~1000 meters), its expected limit of thermal stability, but no interlayered illite/smectite is present below the smectite. These discrepancies suggest that temperature variations alone cannot explain the detailed distribution of clay minerals observed in these wells.

Whole Rock Oxygen Isotope Characteristics of the Coso Hot Springs Geothermal System

General patterns of $\delta^{18}\text{O}$ variation as a function of elevation in the Coso geothermal field are illustrated in figure 5. These plan maps are based on 513 whole rock $\delta^{18}\text{O}$ analyses of well-cuttings supplied by Terra-Gen Operating Company (Supplementary Data Table 1), plus 91 additional whole rock $\delta^{18}\text{O}$ values measured in this study (table 2). The plan maps of $\delta^{18}\text{O}$ variations shown in figure 5 were constructed by averaging the $\delta^{18}\text{O}$ values within 480 m intervals in each well; sufficient analyses were not available deeper than 1125 m below sea-level to construct a useful plan map. Near surface $\delta^{18}\text{O}$ values (surface to 800 m elevation) are similar with, or slightly lower than, measured $\delta^{18}\text{O}$ values for primary igneous rock in the Central Sierra

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217 Figure 5: Plan maps of the whole rock oxygen isotope data at a series of elevation intervals in the
 218 Coso system. The area of each slice is equal to the area of the West Flank outlined in Figure 1
 219 but the horizontal dimension has been exaggerated for visual clarity. The contour values (in ‰)
 220 represent averages of the data over 480 m vertical intervals throughout the field; dashed lines are
 221 used where contouring has been inferred. All elevations are relative to sea level. The red and
 222 black dots are well locations. The location of transect A-A' in the West Flank is shown in the
 223 upper left panel. Contours were generated using the MATLAB contour function.
 224

Nevada Batholith (+7.5 ‰ to +9.0 ‰; Masi et al., 1981; Lackey et al., 2008) except for depletion to $\delta^{18}\text{O}$ values between +5 ‰ and +6 ‰ in the northwest quadrant of the system. With increasing depth, significant and widespread decreases in $\delta^{18}\text{O}$ occur only in the northwest quadrant, where $\delta^{18}\text{O}$ values can decrease down to values between 0 and +1 ‰ (-645 to -1125 m elevation interval). Regardless of depth, the host rock in most of the southern half of the reservoir does not experience significant $^{18}\text{O}/^{16}\text{O}$ depletion below +5 ‰. At the sampling scale of figure 5, the thermal upflow zones inferred in the SW and NE sectors of the field by Williams and McKibben (1990) are not reflected in significant lowering of $\delta^{18}\text{O}$ values in these sectors.

Figure 6 illustrates variations in whole rock $\delta^{18}\text{O}$ values within the vertical section A-A' shown in figure 1 (along the West Flank). In general, depletions in $^{18}\text{O}/^{16}\text{O}$ are more extensive and extend to significantly greater depth in the northern portion of the cross-section. In detail, localized intervals of significant $^{18}\text{O}/^{16}\text{O}$ depletion, with less depleted rocks above and below, occur in the northern wells 68-6 and 33A-7 in intervals centered on elevations of 200, -200 and -1200 m. Maximum depletions to $\delta^{18}\text{O}$ values below 0 ‰ occur in the lowest of these intervals. In contrast, host rocks in the southern part of the traverse are not significantly depleted except for two or three discrete intervals in 73-19 and 58A-18. These spatial patterns of localized and significant $^{18}\text{O}/^{16}\text{O}$ depletions do not reflect the spatial $\delta^{18}\text{O}$ patterns expected from the low and smooth temperature gradients (dT/dx) characteristic of regional metamorphism. The observed spatial patterns of $\delta^{18}\text{O}$ variation are instead more consistent with localized $^{18}\text{O}/^{16}\text{O}$ depletions expected from fluid-rock interactions related to discrete zones of high permeability associated with geothermal activity.

The thermal structure along the cross-section A-A' (Fig. 2) is also superimposed on this $\delta^{18}\text{O}$ cross-section in figure 6 which shows that variations in pre-production temperatures are

Figure 6: Vertical cross section of the whole rock oxygen isotope data along the north-south transect, A-A'. The locations of the nine wells on or close to the A-A' traverse (fig. 1), and data points from each well used for contouring $\delta^{18}\text{O}$ values are shown; contours are in ‰. Red circles indicate the locations of samples analyzed in this study. Isotherms based on preproduction downhole measurements from figure 2 are superimposed to define areas along the cross section above 300 °C (dark gray) and between 200 °C and 300 °C in varying shades of gray; see text for discussion. The major production zone in each well is noted.

not, in general, well correlated with variations in whole-rock $\delta^{18}\text{O}$ values. With the few localized exceptions noted above, the host rocks are not depleted in $^{18}\text{O}/^{16}\text{O}$ to any significant extent in the southern part of the traverse, even within the deeper, higher temperature levels. In the northern part of the section, hotter host rocks below -1400 m are less depleted in $^{18}\text{O}/^{16}\text{O}$ than cooler rocks immediately above between -1000 and -1400 m. However, depth intervals of current production in all three wells do correlate with zones of maximum $^{18}\text{O}/^{16}\text{O}$ depletion (fig. 6). This correlation suggests that permeability plays an important role in controlling the extent of fluid-rock interaction—and hence the extent of $^{18}\text{O}/^{16}\text{O}$ depletion—experienced by the host rocks in the Coso geothermal system.

Oxygen Isotope Characteristics of the West Flank

Whole Rock $\delta^{18}\text{O}$ Analyses. Wells 68-6, 33A-7 and 73-19 were selected for detailed oxygen isotope analyses of whole-rock and mineral separates (table 2; figures 7-9). In well 68-6 (Fig. 7) whole-rock $\delta^{18}\text{O}$ values are somewhat to moderately depleted in $^{18}\text{O}/^{16}\text{O}$ relative to primary igneous values (+7.5 to 9.0‰), and fluctuate between 1.8 ‰ and +6.0 ‰ down to ~2000 meters depth, with a tendency to decrease with depth. Below 2000 m, $^{18}\text{O}/^{16}\text{O}$ depletions are even greater, and the maximum $\delta^{18}\text{O}$ value is +3.0 ‰, with most $\delta^{18}\text{O}$ values below +1 ‰. Whole rock $\delta^{18}\text{O}$ values reach a minimum of -4.6 ‰ at a 2941 meters below the surface (Fig. 7). In well 33A-7 $\delta^{18}\text{O}$ values generally decrease with depth from a maximum value of +7.6 ‰ at 33.5 meters depth to a minimum of -3.1 ‰ at 2530 meters, just above the production zone (interval of lost circulation) in the well (Fig. 8). Immediately below, and to the bottom of the well, $\delta^{18}\text{O}$ values are significantly higher, ranging between +3 to +4.9 ‰, but are still significantly lower than primary igneous values. The host rocks in well 73-19 are less depleted in $^{18}\text{O}/^{16}\text{O}$ compared to the rocks of the other two wells (Fig. 9). The highest measured $\delta^{18}\text{O}$ values from whole rock

278 Table 2: Measured $\delta^{18}\text{O}$ values for whole rock and feldspar mineral samples from well 68-6,
279 Coso geothermal system, California. Values are reported in per mil (‰) notation, relative to
280 VSMOW. Depth is in meters below the surface.

Well: 68-6			Well: 33A-7			Well: 73-19		
Depth	Whole-	Feldspar	Depth	Whole-	Feldspar	Depth	Whole-	Feldspar
(m)	(‰)	(‰)	(m)	(‰)	(‰)	(m)	(‰)	(‰)
271	3.88		34	7.62	7.50	271	6.98	7.79
335	4.47	6.04	110	3.7		381	6.61	
424	2.55		262	5.39		390	6.37	
488	3.49	6.59	415	7.45		445	7.48	8.12
564	5.08		567	4.84		546	5.52	
686	5.78		646	5.34		777	7.57	
722	6.02	6.29	872	0.2		799	6.36	
884	3.57		1024	3.59		1061	7.16	
985	4.07		1100	3.00	5.49	1073	7.41	
1070	4.30		1326	4.18		1225	6.07	
1271	4.43		1570	3.89		1317	6.63	
1372	2.44	5.24	1649	2.71		1347	5.91	
1433	2.87		1878	1.03		1387	4.61	6.14
1484	1.50	4.96	1984	3.93		1411	2.38	
1554	2.60		2137	1.60		1481	6.30	7.26
1701	3.78	5.64	2213	-0.44		1582	5.94	
1798	4.67	6.76	2323	-0.02		1664	6.85	
1936	4.46		2323	-1.02		1710	5.13	
2067	2.94		2350	-0.98		1814	5.14	6.20
2271	2.70		2382	0.14		1835	3.30	6.34
2393	0.75		2472	0.51	-2.37	1856	3.30	
2466	0.75	3.09	2502	-2.35				
2566	0.61		2530	-3.08	-0.03			
2710	0.75	3.08			+0.63*			
2792	2.83	3.59	2566	0.94	4.16			
2850	0.54		2594	2.15	4.20			
2896	0.26		2624	1.55				
2941	-4.60	-3.88	2655	2.92				
		-5.06 [#]	2685	2.10	4.60			
		-4.12*	2746	3.91				
2984	-1.05	1.76	2838	4.48	5.49			
3173	0.64		2960	4.60				
			2990	4.25				
			3021	4.05				
			3051	1.78				
			3082	4.21				
			3112	4.03				
			3143	4.61				
			3173	4.10				
			3234	4.49				
			3295	3.88	5.90			

281 [#]Hand picked cloudy (more altered) feldspar

282 *Feldspar separated from finer-grained (150-200 mesh) whole rock aliquot

Figure 7: Measured $\delta^{18}\text{O}$ values of whole rock and feldspar samples as a function of depth for well 68-6. All values are reported in per mil (VSMOW), and plotted at both well depth and elevation (meters above sea level). The shaded light blue region here and in figures 8 and 9 represents the range of primary $\delta^{18}\text{O}$ values of diorite to granodiorite rocks from the region that are equivalent to the reservoir host rocks in the Coso system (Lackey et al., 2008). The dotted line marks the current production zone (interval of lost circulation) in 68-6.

Figure 8: Measured $\delta^{18}\text{O}$ values (VSMOW) of whole rock and feldspar samples as a function of well depth and elevation for well 33A-7. The dotted line marks the current production zone (interval of lost circulation) in 33A-7. The shaded light blue region represents the range of primary $\delta^{18}\text{O}$ values of diorite to granodiorite rocks from the region that are equivalent to the reservoir host rocks in the Coso system (Lackey et al., 2008).

Figure 9: Measured $\delta^{18}\text{O}$ values (VSMOW) of whole rock and feldspar samples as a function of well depth and elevation for well 73-19. The dotted line marks the current production zone (interval of lost circulation) in 73-19. The shaded light blue region represents the range of primary $\delta^{18}\text{O}$ values of diorite to granodiorite rocks from the region that are equivalent to the reservoir host rocks in the Coso system (Lackey et al., 2008).

samples overlap primary igneous $\delta^{18}\text{O}$ values of +7.5 to 9.0 ‰ (Masi et al., 1981; Lackey et al.,

2008). The only $\delta^{18}\text{O}$ values below +5 ‰ occur at depths of 1387 m and below. The minimum

$\delta^{18}\text{O}$ value in well 73-19 is still positive, +2.4 ‰ (1411 meters depth, Fig. 9). In all three wells,

samples with the lowest measured $\delta^{18}\text{O}$ values (or with $\delta^{18}\text{O}_{\text{wr}} \leq 2.0$ ‰, well 33A-7) correspond

with major intervals of lost circulation, the primary production intervals for each well (Figs. 7-9).

Feldspar vs. Whole Rock Oxygen Isotope Compositions. Measured $\delta^{18}\text{O}$ in bulk feldspar

separates range from 6.8‰ to -3.9 ‰ in well 68-6 and from 7.5 ‰ to -2.4 ‰ in 33A-7; feldspars

312 from well 73-19 have higher $\delta^{18}\text{O}$ values ranging from 8.1 ‰ to 6.1 ‰. In all three wells, the
313 $\delta^{18}\text{O}$ values of the feldspar correlate positively with the whole-rock $\delta^{18}\text{O}$ values, and are
314 systematically 1 to 3 ‰ higher than the whole rock values, even in the more $^{18}\text{O}/^{16}\text{O}$ -depleted
315 samples (Figs. 7-9). The only exception is in 33A-7 at 2472 m. The bulk feldspar separates are
316 mixtures of primary (igneous or metamorphic) feldspar and feldspar that has experienced oxygen
317 isotope exchange with geothermal fluid (hydrothermal feldspar). The hydrothermal feldspar
318 should have a considerably lower $\delta^{18}\text{O}$ value than primary feldspar, owing to the low $\delta^{18}\text{O}$ value
319 of the geothermal fluid and intermediate temperature of exchange ($\leq 350^\circ\text{C}$). Further, feldspar is
320 normally more susceptible to oxygen isotope exchange compared to hornblende and less
321 abundant quartz in the host diorite and quartz diorite (Taylor and Forester, 1979; Criss and
322 Taylor, 1983; Cole et al., 1992). As a result, hydrothermal feldspar might be expected to have a
323 $\delta^{18}\text{O}$ value less than that of the whole-rock for rocks that are only moderately depleted in
324 $^{18}\text{O}/^{16}\text{O}$. In thin section, primary igneous and regional metamorphic feldspar are clear and
325 twinned (igneous feldspar); hydrothermal feldspar is cloudy and sometimes turbid from the
326 presence of pits/cavities and very small grains of clay minerals. The intergrown nature of
327 primary and hydrothermal feldspar at the grain scale makes their effective separation by standard
328 density means impractical. However a concentrate of cloudy-looking feldspar was made by hand
329 picking the bulk feldspar separate from sample 68-6_2941. Analysis of this concentrate yielded a
330 $\delta^{18}\text{O}$ value of -5.1 ‰, significantly lower than the bulk feldspar (-3.9 ‰) and whole rock (-4.6
331 ‰) $\delta^{18}\text{O}$ values of this sample (Table 2). In-situ (SIMS) analysis will be necessary to define
332 clearly the $\delta^{18}\text{O}$ value of hydrothermal feldspar in the host rock (e.g., Valley and Graham, 1996;
333 Valley and Kita, 2009). Regardless, this systematic difference between measured $\delta^{18}\text{O}$ values of
334 bulk feldspar and whole rock suggests that the $\delta^{18}\text{O}$ values of both feldspar and whole rock

reflect the extent of interaction (and oxygen isotopic exchange) between host rock and geothermal fluid.

Another possibility for the systematic $^{18}\text{O}/^{16}\text{O}$ enrichment in bulk feldspar relative to the whole rock is that the lower $\delta^{18}\text{O}$, hydrothermally altered feldspar was preferentially removed by crushing during preparation of the sample aliquots (>80 or >100 mesh size) from the whole rock used to make plagioclase separates. To test this possibility, additional analyses were made of feldspar separated out of the finer-grained (150 to 200 mesh) residual fraction of the whole rock aliquots for two samples. The analyzed $\delta^{18}\text{O}$ values for these finer-grained feldspar separates are -4.1 ‰ (compared to -3.9) for sample 68-6_2941, and +0.6 ‰ (compared to -0.0) for sample 33A-7_2530 (table 2). Additional details of these results are in Etzel (2015). Feldspar separates from these finer-grained fractions are not significantly lower in $\delta^{18}\text{O}$ compared to feldspar separates from the coarser-grained fractions.

The systematically higher measured $\delta^{18}\text{O}$ values of feldspar relative to the whole rock indicate, from mass balance considerations, that there is at least one other major phase (likely chlorite and/or other clay mineral alteration products) in the rock that is significantly depleted in $^{18}\text{O}/^{16}\text{O}$ relative to the whole rock value. Chlorite can replace hornblende but would be stable with biotite during the greenschist facies conditions of the earlier regional metamorphism; this generation of chlorite will not be related to geothermal activity. Replacement of biotite by chlorite would more likely occur at the sub-greenschist facies conditions of the geothermal environment. Hence, bulk separates of chlorite will potentially contain two generations of chlorite of very different $\delta^{18}\text{O}$ value. Good separation of bulk samples of chlorite by magnetic and density methods proved impractical owing to the intergrowth of chlorite with biotite and hornblende at the grain size scale. Concentrates of chloritized biotite from four samples were

handpicked. Their $\delta^{18}\text{O}$ values, compared to whole rock values (in parentheses) are: 1) 0.2 ‰ (3.8), sample 68-6_1701; 2) -3.2 ‰ (0.7), sample 68-6_2710; 3) -3.8 ‰ (-3.1), sample 33A-7_2530; and 4) -0.45 ‰ (4.6), sample 73-19_1387. These chloritized biotites are from almost 1 to as much as 5 ‰ depleted in $^{18}\text{O}/^{16}\text{O}$ relative to the whole rock values, satisfying at least qualitatively this mass balance requirement. In-situ analyses will be necessary to define the $\delta^{18}\text{O}$ values of the hydrothermal chlorite more precisely.

DISCUSSION

Clay Mineralogy as a Function of Fluid-Rock Interaction

The observed distributions of clay minerals in wells 68-6, 33A-7 and 73-19 are not completely consistent with the pre-production temperatures in these wells (Fig. 3). These discrepancies may indicate that the pre-production thermal regime was different than the regime responsible for clay mineral zoning, and that the clay minerals have not yet re-adjusted to the newer (pre-production) thermal conditions. The persistence of clay minerals outside their normal stability fields (e.g., smectites at $T > 180^\circ\text{C}$ in 73-19, fig. 3) is common but the reasons are poorly understood. Nucleation and reaction kinetics may play roles. In addition permeability, which can either promote (high permeability) or inhibit (low permeability) clay-fluid interactions, may be influencing the spatial distribution of the clay minerals.

Oxygen isotope evidence suggests that the extent of fluid-rock interaction experienced by the host rock is influencing the distribution of clay minerals in the Coso system. The amount of $^{18}\text{O}/^{16}\text{O}$ depletion is a monitor of the extent of fluid-rock interaction by the host rock—and hence it's permeability. Figure 10 superimposes the distribution of clay minerals in wells 68-6, 33A-7 and 73-19 on variations in whole rock $\delta^{18}\text{O}$ values along the cross section A-A'. The host rocks are systematically more depleted in $^{18}\text{O}/^{16}\text{O}$ at depth at the north end compared to the south end

Figure 10: Clay zones within wells 68-6, 33A-7 and 73-19 superimposed onto the contoured cross-section of $\delta^{18}\text{O}$ values along the traverse A-A' in the West Flank. Black circles represent intervals with preexisting $\delta^{18}\text{O}$ values, red circles represent intervals where $\delta^{18}\text{O}$ values were determined for this study. Increasing extent of $^{18}\text{O}/^{16}\text{O}$ depletion northward correlates with changes in clay mineral zoning with depth, suggesting a correlation between clay mineral stability, extent of fluid-rock interaction and permeability (see text for discussion).

of the traverse. In addition, significant $^{18}\text{O}/^{16}\text{O}$ depletions extend to markedly shallower depth at the north end of A-A'. The overall extent of $^{18}\text{O}/^{16}\text{O}$ depletion in the south third of the traverse is considerably less, and significant $^{18}\text{O}/^{16}\text{O}$ depletions in well 73-19 are confined to several discrete/narrow/short intervals. Hence in well 73-19 the transition from smectite to interlayered illite/smectite clay is better correlated with temperature, as this transition occurs near the expected temperature of 180°C (Fig. 3). However interlayered illite/smectite clay occurs as deeply as -200 m elevation ($T = 260^\circ\text{C}$, Fig. 3) in well 73-19, an elevation corresponding to a local interval of significant $^{18}\text{O}/^{16}\text{O}$ depletion ($\delta^{18}\text{O} < 4\text{‰}$). Figure 10 shows that there is a general correlation between greater and more spatially widespread $^{18}\text{O}/^{16}\text{O}$ depletion in the north end of A-A' with a thinner smectite zone in 68-6 and 33A-7 and the lack of a mixed illite/smectite transition zone in 68-6. Additionally, the transition from interlayered illite/smectite to illite clay in 73-19 is correlated with local $^{18}\text{O}/^{16}\text{O}$ depletion. These $^{18}\text{O}/^{16}\text{O}$ depleted zones within traverse A-A' record greater fluid-rock interaction, and are characterized by higher permeability. Hence the extent of fluid-rock interaction (host rock permeability) appears to be controlling, at least in part, the spatial distribution of clay minerals and clay zoning patterns in wells 68-6, 33A-7 and 73-19 along the traverse A-A'.

Oxygen Isotope Exchange and Fluid-Rock Interaction

The $\delta^{18}\text{O}$ values of the whole rock samples from the West Flank range widely from -4.6 ‰ to 7.6 ‰ (table 2). Reservoir rock in 33A-7 and 68-6, except for the shallowest levels, have experienced at least moderate $^{18}\text{O}/^{16}\text{O}$ depletions ($\delta^{18}\text{O} < 5.0\text{‰}$) from primary igneous $\delta^{18}\text{O}$ values (+7.5 to 9.0‰). Hence most of the reservoir rock sampled in these two wells has experienced at least some interaction and oxygen isotope exchange with the geothermal reservoir fluid derived from low $\delta^{18}\text{O}$ local meteoric water. Further, a few discrete intervals in 33A-7 and

68-6 have experienced significantly greater $^{18}\text{O}/^{16}\text{O}$ depletion, and have negative whole-rock $\delta^{18}\text{O}$ values ranging as low as -4.6 ‰.

The extent of $^{18}\text{O}/^{16}\text{O}$ depletion is a function of the amount of reservoir fluid with which the rock has interacted, the temperature and the degree to which isotope exchange equilibrium is attained. Reactive transport models for oxygen isotopes (Baumgartner and Rumble, 1988; Bowman and Willett, 1991; Bowman et al., 1994; Cook et al., 1997; Baumgartner and Valley, 2001) have the potential to constrain fluid fluxes, which are related functionally to porosity and permeability. Such transport models are most applicable to interpreting isotopic shifts when the flow geometry of the hydrothermal system is well established and when the locations of the analyzed rocks along the principal flow paths are also known. Person et al. (2006) inferred the basic structure of the regional groundwater system that could have produced the Coso geothermal system. However, the results of our study show that the $^{18}\text{O}/^{16}\text{O}$ depletions recorded in the host rocks of the West Flank are poorly correlated with pre-production temperature patterns and better correlated with known intervals of high permeability (e.g., current fluid production zones)(fig. 6). These relationships suggest that the flow paths from depth and particularly within the West flank reservoir are likely complicated by spatial variations in the permeability of the reservoir host rocks produced by an array of permeable and impermeable fault zones. This is also consistent with geochemical and enthalpy data from well tests that suggest that the Coso field is divided into a series of hydrologically isolated (fault bounded?), single-phase reservoirs (Williams and McKibben, 1990). As a result, the fracture-controlled and channelized flow patterns responsible for the localized $^{18}\text{O}/^{16}\text{O}$ depletions measured in the West Flank host rocks (this study) are not yet well understood. Hence application of transport models to interpret these $^{18}\text{O}/^{16}\text{O}$ depletions will be problematic without additional information on the

434 geometry of these fracture networks. Given such complications, estimates of the minimum
435 amounts of fluid involved in isotopic exchange can be made using conventional box models
436 of mass balance to compute water (W) to rock (R) ratios (Taylor, 1971).

437 Calculations of box model W/R ratios involve several assumptions, and it is important to
438 understand how these assumptions place limitations on their interpretation. Both closed- and
439 open-system box model endmembers (Taylor, 1971; Criss and Taylor, 1986) assume attainment
440 of exchange equilibrium between rock and fluid. Incomplete exchange yields less $^{18}\text{O}/^{16}\text{O}$
441 depletion in a rock; this will result in lower calculated W/R ratios. Departures from equilibrium
442 may be important in geothermal systems that form at temperatures less than 300 °C, and
443 particularly so in their shallow, lower temperature sections. Disequilibrium is demonstrated at
444 Coso by the large differences in measured $\delta^{18}\text{O}$ values between hydrothermal vs. igneous
445 minerals such as feldspar and chlorite, and between reservoir fluid and host rock.

446 A second limitation is that box model W/R calculations do not account for the isotopic
447 modification that the fluid undergoes along segments of its flow paths leading to the local site of
448 hydrothermal alteration and isotopic exchange (Baumgartner and Rumble, 1988; Bowman and
449 Willett, 1991; Bowman et al., 1994). The impact of this prior exchange history of the fluid is that
450 box model W/R ratios will underestimate actual W/R ratios by an amount that increases with
451 increasing distance along the reactive flow path. The actual amount of fluid involved might have
452 been orders of magnitude higher. As a result, calculated box model W/R ratios should be
453 regarded as minimum estimates of the amounts of fluid involved in hydrothermal alteration and
454 isotopic exchange at specific sites within hydrothermal flow systems. Meteoric waters recharging
455 the Coso geothermal reservoir have experienced considerable $^{18}\text{O}/^{16}\text{O}$ enrichment, likely
456 resulting from fluid-rock exchange within the deeper, hotter portions of the Coso hydrologic

system (Person et al., 2006). However, the geothermal fluids are still well out of oxygen isotope exchange equilibrium with the reservoir host rocks at temperatures from 200 to 350°C. As a result, these fluids are still chemically (isotopically) reactive, and are capable of producing significant $^{18}\text{O}/^{16}\text{O}$ depletions in the infiltrated rocks. Consequently, variations in the extent of $^{18}\text{O}/^{16}\text{O}$ depletion in rocks from local segments of the Coso hydrothermal system—and variations in calculated W/R ratios—will qualitatively reflect variations in fluid fluxes. Hence, the W/R ratios calculated from oxygen isotope ratios are expected to reflect the relative differences in permeability and time-integrated fluid flux for these local segments.

Calculated W/R ratios (atomic oxygen) using both closed and open system box models (Taylor, 1971; Criss and Taylor, 1986) are presented in tables 3-5. The average $\delta^{18}\text{O}$ value of local meteoric water (geothermal fluid source) with similar δD values (-96 to -104 ‰) to those of the pre-production reservoir fluids is ~ -13 ‰ (Williams and McKibben, 1990). The $\delta^{18}\text{O}$ values of the pre-production reservoir fluids were determined using measured temperature, pressure and $\delta^{18}\text{O}$ values of the separated vapor (steam) and liquid water sampled during pre-production flow tests in each well (unpublished data, Terra-Gen Operating Company), following the heat and mass balance calculations described by Truesdell (1984). Use of samples from pre-production flow tests avoids the impact on the calculated $\delta^{18}\text{O}$ values of the reservoir fluids of any $^{18}\text{O}/^{16}\text{O}$ enrichment from steam loss during subsequent power production. In all following discussion, the $\delta^{18}\text{O}$ values of the reservoir fluid refer to the calculated $\delta^{18}\text{O}$ values of the pre-production fluids.

The $\delta^{18}\text{O}$ measurements identify a limited number of localized intervals of much more extensive $^{18}\text{O}/^{16}\text{O}$ depletion within the reservoir host rocks that have interacted and exchanged isotopically with significantly larger quantities of geothermal fluid (higher calculated W/R ratios). Well 33A-7 has three such intervals of locally higher calculated W/R ratios at 2502 -

Table 3: Calculated equilibrium Δ_{r-w} and Δ_{fsp-w} , measured Δ_{r-w} and Δ_{fsp-w} , and W/R ratios for well 68-6. WR = whole-rock; Eq = equilibrium; FSP = feldspar.

68-6 ID	W.R. $\delta^{18}\text{O}$	Eq Δ_{r-w}^1	Measured Δ_{r-w}^2	FSP $\delta^{18}\text{O}$	Eq Δ_{fsp-w}^1	Measured Δ_{fsp-w}^2	W/R closed	W/R Open
271	3.88	9.33	12.55				0.49	0.40
335	4.47	9.33	13.14	6.04	10.17	14.71	0.36	0.31
424	2.55	9.33	11.22				0.80	0.59
488	3.49	9.27	12.16	6.59	10.10	15.26	0.58	0.46
564	5.08	9.27	13.75				0.22	0.20
686	5.78	9.03	14.45				0.06	0.05
722	6.02	8.79	14.69	6.29	9.59	14.96	0.00	0.00
884	3.57	8.04	12.24				0.57	0.45
985	4.07	7.68	12.74				0.45	0.37
1070	4.30	7.36	12.97				0.40	0.33
1271	4.43	7.36	13.10				0.37	0.31
1372	2.44	6.81	11.11	5.24	7.46	13.91	0.83	0.60
1433	2.87	6.81	11.54				0.73	0.55
1484	1.50	6.81	10.17	4.96	7.46	13.63	1.04	0.71
1554	2.60	6.81	11.27				0.79	0.58
1701	3.78	6.72	12.45	5.64	7.37	14.31	0.52	0.42
1798	4.67	6.51	13.34	6.76	7.13	15.43	0.31	0.27
1936	4.46	6.38	13.13				0.36	0.31
2067	2.94	6.30	11.61				0.71	0.54
2271	2.70	6.06	11.37				0.77	0.57
2393	0.75	5.94	9.42				1.22	0.80
2466	0.75	5.88	9.42	3.09	6.46	11.76	1.22	0.80
2566	0.61	5.79	9.28				1.25	0.81
2710	0.75	5.50	9.42	3.08	6.05	11.75	1.22	0.80
2792	2.83	5.60	11.50	3.59	6.16	12.26	0.74	0.55
2850	0.54	5.77	9.21				1.27	0.82
2896	0.26	5.98	8.93				1.33	0.85
2941	-4.60	5.98	4.07	-3.88	6.57	4.79	2.45	1.24
2984	-1.05	5.73	7.62	1.76	6.30	10.43	1.63	0.97
3173	0.64	5.00	9.31				1.24	0.81

¹Equilibrium Δ_{r-w} and Δ_{fsp-w} fractionation values calculated at pre-production reservoir temperatures using the experimental fractionation factor for An₅₀-water and An₂₅-water (O'Neil and Taylor, 1967), respectively.

²Measured Δ_{wr-w} and Δ_{fsp-w} are the differences between the measured $\delta^{18}\text{O}$ values of whole-rock or feldspar, and the measured $\delta^{18}\text{O}$ value of the current geothermal fluid (-8.67 ‰).

Table 4: Calculated equilibrium Δ_{r-w} and Δ_{fsp-w} , measured Δ_{r-w} and Δ_{fsp-w} , and W/R ratios for well 33A-7. WR = whole-rock; Eq = equilibrium; FSP = feldspar.

33A-7 ID	W.R. $\delta^{18}\text{O}$	Eq Δ_{r-w}^1	Measured Δ_{r-w}^2	FSP $\delta^{18}\text{O}$	Eq Δ_{fsp-w}^1	Measured Δ_{fsp-w}^2	W/R closed	W/R Open
34	7.62	23.44	15.11	7.50	25.34	14.99	0.00	0.00
110	3.7	19.54	11.19				0.71	0.54
262	5.39	14.03	12.88				0.40	0.34
415	7.45	11.00	14.94				0.03	0.03
567	4.84	11.00	12.33				0.50	0.41
646	5.34	9.78	12.83				0.41	0.35
872	0.2	9.30	7.69				1.35	0.85
1024	3.59	8.70	11.08				0.73	0.55
1100	3	8.56	10.49	5.49	9.34	12.98	0.84	0.61
1326	4.18	8.20	11.67				0.62	0.49
1570	3.89	7.94	11.38				0.68	0.52
1649	2.71	7.76	10.20				0.89	0.64
1878	1.03	7.39	8.52				1.20	0.79
1984	3.93	7.20	11.42				0.67	0.51
2137	1.6	6.81	9.09				1.09	0.74
2213	-0.44	6.70	7.05				1.46	0.90
2323	-0.02	6.49	7.47				1.39	0.87
2323	-1.02	6.49	6.47				1.57	0.94
2350	-0.98	6.44	6.51				1.56	0.94
2382	0.14	6.38	7.63				1.36	0.86
2472	0.51	6.18	8.00	-2.37	6.78	5.12	1.29	0.83
2502	-2.35	6.06	5.14				1.81	1.03
2530	-3.08	6.02	4.41	-0.03	6.61	7.46	1.94	1.08
2566	0.94	6.02	8.43	4.16	6.61	11.65	1.21	0.79
2594	2.15	5.96	9.64	4.20	6.55	11.69	0.99	0.69
2624	1.55	5.85	9.04				1.10	0.74
2655	2.92	5.79	10.41				0.85	0.62
2685	2.1	5.73	9.59	4.60	6.30	12.09	1.00	0.69
2746	3.91	5.60	11.40				0.67	0.51
2838	4.48	5.48	11.97	5.49	6.03	12.98	0.57	0.45
2960	4.6	5.25	12.09				0.55	0.44

¹Equilibrium Δ_{r-w} and Δ_{fsp-w} fractionation values calculated at pre-production reservoir temperatures using the experimental fractionation factor for An₅₀-water and An₂₅-water (O'Neil and Taylor, 1967), respectively.

² Measured Δ_{wr-w} and Δ_{fsp-w} are the differences between the measured $\delta^{18}\text{O}$ values of whole-rock or feldspar, and the measured $\delta^{18}\text{O}$ value of the current geothermal fluid (-7.49 ‰).

507 Table 4 (continued): Calculated equilibrium Δ_{r-w} and Δ_{fsp-w} , measured Δ_{r-w} and Δ_{fsp-w} , and W/R
508 ratios for well 33A-7. WR = whole-rock; Eq = equilibrium; FSP = feldspar.

33A-7 ID	W.R. $\delta^{18}\text{O}$	Eq Δ_{r-w}^1	Measured Δ_{r-w}^2	FSP $\delta^{18}\text{O}$	Eq Δ_{fsp-w}^1	Measured Δ_{fsp-w}^2	W/R closed	W/R Open
2990	4.25	5.20	11.74				0.61	0.48
3021	4.05	5.16	11.54				0.65	0.50
3051	1.78	5.13	9.27				1.06	0.72
3082	4.21	5.13	11.70				0.62	0.48
3112	4.03	5.00	11.52				0.65	0.50
3143	4.61	4.95	12.10				0.55	0.44
3173	4.1	4.84	11.59				0.64	0.49
3234	4.49	4.76	11.98				0.57	0.45
3295	3.88	4.68	11.37	5.90	5.17	13.39	0.68	0.52

509
510 ¹Equilibrium Δ_{r-w} and Δ_{fsp-w} fractionation values calculated at pre-production reservoir
511 temperatures using the experimental fractionation factor for An₅₀-water and An₂₅-water (O'Neil
512 and Taylor, 1967), respectively.

513 ²Measured Δ_{wr-w} and Δ_{fsp-w} are the differences between the measured $\delta^{18}\text{O}$ values of whole rock
514 or feldspar, and the measured $\delta^{18}\text{O}$ value of the current geothermal fluid (-7.49 ‰).
515

Table 5: Calculated equilibrium Δ_{r-w} and Δ_{fsp-w} , measured Δ_{r-w} and Δ_{fsp-w} , and W/R ratios for well 73-19. WR = whole-rock; Eq = equilibrium; FSP = feldspar.

73-19 ID	W.R. $\delta^{18}\text{O}$	Eq Δ_{r-w}^1	Measured Δ_{r-w}^2	FSP $\delta^{18}\text{O}$	Eq Δ_{fsp-w}^1	Measured Δ_{fsp-w}^2	W/R closed	W/R Open
271	6.98	16.10	12.54	7.79	17.43	13.35	0.08	0.08
381	6.61	14.08	12.17				0.13	0.12
390	6.37	14.03	11.93				0.16	0.15
445	7.48	13.35	13.04	8.12	14.47	13.68	0.01	0.01
546	5.52	11.78	11.08				0.28	0.24
777	7.57	8.85	13.13				0.00	0.00
799	6.36	8.62	11.92				0.16	0.15
1061	7.16	6.37	12.72				0.06	0.05
1073	7.41	6.25	12.97				0.02	0.02
1225	6.07	5.47	11.63				0.20	0.18
1317	6.63	5.12	12.19				0.13	0.12
1347	5.91	5.12	11.47				0.22	0.20
1387	4.61	4.92	10.17	6.14	5.42	11.70	0.40	0.33
1411	2.38	5.00	7.94				0.70	0.53
1481	6.3	4.69	11.86	7.26	5.18	12.82	0.17	0.16
1582	5.94	4.55	11.50				0.22	0.20
1664	6.85	4.39	12.41				0.10	0.09
1710	5.13	4.26	10.69				0.33	0.28
1814	5.14	4.63	10.70	6.20	4.69	11.76	0.33	0.28
1835	3.3	4.95	8.86	6.34	4.69	11.90	0.57	0.45
1856	3.3	4.86	8.86				0.57	0.45

¹Equilibrium Δ_{r-w} and Δ_{fsp-w} fractionation values calculated at pre-production reservoir temperatures using the experimental fractionation factor for An₅₀-water and An₂₅-water (O'Neil and Taylor, 1967), respectively.

² Measured Δ_{wr-w} and Δ_{fsp-w} are the differences between the measured $\delta^{18}\text{O}$ values of whole-rock or feldspar, and the measured $\delta^{18}\text{O}$ value of the current geothermal fluid (-5.56 ‰).

2530 m $[(W/R)_c = 1.94]$; at 872 m $[(W/R)_c = 1.35]$; and at 110 m $[(W/R)_c = 0.71]$. Well 68-6 has one interval at 2941 m that is characterized by much higher W/R ratio $[(W/R)_c = 2.45]$ than elsewhere in the well. In general, calculated W/R ratios for 73-19 are lower than for the other two wells, suggesting that permeability in this well is generally lower than in 68-6 and 33A-7. However locally higher W/R ratios (closed system) are computed for depths of 1856 (0.57), 1835 (0.57), 1412 (0.70) and 546 (0.28) meters in well 73-19. The presence of these limited number of discrete intervals of significant $^{18}\text{O}/^{16}\text{O}$ depletion (and high calculated W/R ratios) in the three wells reflect discrete zones of much higher permeability within low permeability (much less $^{18}\text{O}/^{16}\text{O}$ depleted) rock. This spatial pattern is consistent with fluid flow that is focused along discrete fractures. In all three wells, the depths of current reservoir fluid production, which are represented by zones of lost circulation and high permeability, correspond to one of these intervals of significant $^{18}\text{O}/^{16}\text{O}$ depletion—and high W/R ratio (figures 7-9).

The Extent of Isotope Exchange Equilibrium Accompanying Fluid-Rock Interaction

Because the pre-production temperatures of the reservoir fluids are known, the equilibrium oxygen isotope fractionations between feldspar and water (equilibrium $\Delta_{\text{fsp-w}}$) can be calculated. These values can be compared to the measured differences in $\delta^{18}\text{O}$ between feldspar and the reservoir fluids (measured $\Delta_{\text{fsp-w}}$) (tables 3-5) to evaluate the extent of exchange equilibrium between the geothermal fluid and feldspar in the reservoir host rocks. If feldspar has incompletely exchanged oxygen isotopes with the low $\delta^{18}\text{O}$ geothermal fluid at temperatures below 500 °C, measured $\Delta_{\text{fsp-w}}$ values will be greater than the equilibrium $\Delta_{\text{fsp-w}}$ values. The greater this difference, the farther from exchange equilibrium is the feldspar. The equilibrium fractionation factors are calculated at temperatures measured in the wells prior to production,

549 using the experimental calibration for oxygen isotope fractionation between plagioclase and
550 water from O'Neil and Taylor (1967) and a plagioclase composition of An₂₅ (average of
551 measured feldspar compositions in the reservoir rocks; Etzel, 2015).

552 Figure 11a depicts the measured $\delta^{18}\text{O}$ values of feldspar plotted against the difference
553 between measured $\Delta_{\text{fsp-w}}$ and equilibrium $\Delta_{\text{fsp-w}}$ values. This figure illustrates that the measured
554 $\Delta_{\text{fsp-w}}$ values for most of the sampled intervals are larger than equilibrium $\Delta_{\text{fsp-w}}$ values at
555 measured (pre-production) temperatures and plot to the right of the dashed line (where measured
556 $\Delta_{\text{fsp-w}}$ = equilibrium $\Delta_{\text{fsp-w}}$) in figure 11a. This difference indicates that feldspars from most of the
557 sampled intervals in these three wells have not completely exchanged with the geothermal fluid.
558 Two groups of samples with measured $\Delta_{\text{fsp-w}}$ values less than equilibrium $\Delta_{\text{fsp-w}}$ are exceptions.
559 The three feldspar samples with $\delta^{18}\text{O}$ values >7.0 ‰ come from the shallowest, lowest
560 temperature intervals of 33A-7 and 73-19 and have experienced little or no $^{18}\text{O}/^{16}\text{O}$ depletion
561 relative to their primary $\delta^{18}\text{O}$ values. As a result, these feldspars have not experienced any
562 significant interaction with geothermal fluid, and are far from exchange equilibrium with the
563 geothermal fluid (large measured $\Delta_{\text{fsp-w}}$). However given the low temperatures for these three
564 samples, equilibrium $\Delta_{\text{fsp-w}}$ values are even larger. The second exception where measured $\Delta_{\text{fsp-w}}$
565 values are less than equilibrium $\Delta_{\text{fsp-w}}$ values are the most $^{18}\text{O}/^{16}\text{O}$ depleted feldspars from 33A-7
566 and 68-6. These two samples are discussed in the following section.

567 Figure 11a also shows that in general, the lower the $\delta^{18}\text{O}$ value of the feldspar, the closer
568 the feldspar approaches exchange equilibrium with the geothermal reservoir fluid (e.g., measured
569 $\Delta_{\text{fsp-w}}$ approaches equilibrium $\Delta_{\text{fsp-w}}$). This positive correlation indicates that the feldspar moves
570 progressively toward oxygen isotope exchange equilibrium with the reservoir fluid as the extent
571 of fluid-rock interaction increases (as indicated by progressively greater $^{18}\text{O}/^{16}\text{O}$ depletion in the

Figure 11: A) The difference between measured $\Delta_{\text{fsp-w}}$ and equilibrium $\Delta_{\text{fsp-w}}$ plotted against the measured $\delta^{18}\text{O}$ value of each feldspar separate. B) The difference between measured $\Delta_{\text{fsp-w}}$ and equilibrium $\Delta_{\text{fsp-w}}$ plotted against measured temperature (pre-production) for the feldspar sample. Equilibrium $\Delta_{\text{fsp-w}}$ is calculated at pre-production temperatures. The dashed line represents equilibrium (measured $\Delta_{\text{fsp-w}}$ = equilibrium $\Delta_{\text{fsp-w}}$) between analyzed feldspar and the pre-production reservoir fluid. Feldspars plotting to right of the dashed line have incompletely exchanged oxygen isotopes with the reservoir fluid.

feldspar and higher calculated W/R ratio). Experimental studies and theoretical considerations indicate that the rate of oxygen isotope exchange accompanying the types of surface reactions responsible for producing the hydrothermal alteration minerals observed at Coso Hot Springs increases with increasing surface area (Cole and Chakraborty, 2001). The positive correlation between the progressive $^{18}\text{O}/^{16}\text{O}$ depletion in the feldspar and the increasing approach to oxygen isotopic exchange equilibrium (measured $\Delta_{\text{fsp-w}}$ approaches equilibrium $\Delta_{\text{fsp-w}}$) observed for most of the feldspar samples suggests that increases in permeability of the reservoir host rock produce an increase in surface area of the host rock exposed to geothermal fluid, promoting fluid-rock interaction and isotopic exchange.

Rates of isotopic exchange accompanying hydrothermal alteration also increase dramatically with increasing temperature (Cole and Chakraborty, 2001). However, there is no

systematic correlation of temperature with the extent of exchange equilibrium between the feldspar and reservoir fluid (figure 11b). The excess of measured $\Delta_{\text{fsp-w}}$ compared to equilibrium $\Delta_{\text{fsp-w}}$ value for feldspar samples from 68-6 is clearly independent of temperature from 180°C to 265°C; for example, across the temperature range of 225-265 °C the difference between the measured $\Delta_{\text{fsp-w}}$ and equilibrium $\Delta_{\text{fsp-w}}$ decreases from 6.3 to -2.0 and then increases to 5.9. A similar situation holds for samples from 33A-7 over the temperature interval from 190 °C to 250 °C; the difference between measured and equilibrium $\Delta_{\text{fsp-w}}$ values changes from 3.6 down to -1.7 and then abruptly increases to 5.

Thermal and oxygen isotope evolution of geothermal fluids in the Coso system

Measured $\delta^{18}\text{O}$ values of the feldspars are plotted as a function of the pre-production temperature in figure 12. Curved lines define the calculated $\delta^{18}\text{O}$ values of feldspar in equilibrium with a range of $\delta^{18}\text{O}$ values of water, including the pre-production reservoir fluids (solid line). It is apparent that the analyzed feldspars from higher temperatures (>200°C) in 73-19 plot well above the equilibrium curve for the reservoir fluid, and are far from exchange equilibrium with the reservoir fluid. These feldspars have experienced limited oxygen isotope exchange, either from kinetic barriers to isotopic exchange (from lower temperatures in the past?) or limited physical contact between the reservoir fluid and rock owing to lower permeability, or both.

Many of the feldspar samples from 68-6 and 33A-7 have experienced greater $^{18}\text{O}/^{16}\text{O}$ depletion and therefore more extensive oxygen isotope exchange, but have not equilibrated with the reservoir fluid (Fig. 12). In contrast, the two feldspars from 33A-7 and 68-6 with the lowest $\delta^{18}\text{O}$ values plot below the equilibrium curve for the reservoir fluid. For these two feldspar samples, measured $\Delta_{\text{fsp-w}}$ values are less than equilibrium $\Delta_{\text{fsp-w}}$ values by -1.66 (33A-7_2471.9

Figure 12: Measured $\delta^{18}\text{O}$ values of feldspars plotted as a function of pre-production temperatures. Curved lines define the calculated $\delta^{18}\text{O}$ values of feldspar in exchange equilibrium with a range of $\delta^{18}\text{O}$ values of water (values specified next to each curve), including the pre-production reservoir fluids in these wells (solid line), as a function of temperature. Arrows indicate changes in temperature or $\delta^{18}\text{O}$ of reservoir fluid needed for the most $^{18}\text{O}/^{16}\text{O}$ -depleted feldspar in 68-6 and 33A-7 to achieve oxygen isotope exchange equilibrium with the geothermal reservoir fluids.

m) and -1.99 (68-6_2941.3 m) (Fig. 11a). At pre-production temperatures, incomplete isotopic exchange would produce measured $\Delta_{\text{fsp-w}}$ values greater than the equilibrium $\Delta_{\text{fsp-w}}$ values. Therefore these discrepancies cannot result from incomplete oxygen isotope exchange between the geothermal fluid and the reservoir rocks, because of either slow isotopic exchange kinetics or lack of physical contact between fluid and feldspar crystals. Measured $\Delta_{\text{fsp-w}}$ less than equilibrium $\Delta_{\text{fsp-w}}$ requires that the isotopic exchange took place either at higher temperatures and/or with a reservoir fluid of lower $\delta^{18}\text{O}$ value. Figure 12 shows that temperatures would need to increase by 55 °C (to 307 °C) and 42 °C (to 288 °C) in 68-6 and 33A-7, respectively, for the most $^{18}\text{O}/^{16}\text{O}$ –depleted feldspars from each well to be in exchange equilibrium with the reservoir fluid. Alternatively, the $\delta^{18}\text{O}$ values of geothermal fluids would need to decrease by 1.98 ‰ and 1.66 ‰ in 68-6 and 33A-7, respectively (tables 3,4; fig. 12). A lower $\delta^{18}\text{O}$ value of the reservoir fluid in the past implies a decrease in the initial $\delta^{18}\text{O}$ value of the meteoric water source, and/or an increase in the W/R ratio (and permeability) for these sections of geothermal system.

CONCLUSIONS

The patterns and extent of $^{18}\text{O}/^{16}\text{O}$ depletions measured in the reservoir rocks of the Coso Hot Springs geothermal system, combined with clay mineral alteration patterns, record significantly varying degrees of fluid-rock interactions. These patterns provide insights into the role of fracture-controlled permeability in this geothermal system. Our results indicate the following:

- Clay mineral distribution in three wells along the West Flank is not always closely correlated with the distribution expected from preproduction temperatures. There is,

however, a general correlation between greater and more widespread $^{18}\text{O}/^{16}\text{O}$ depletion in the northern part of the West Flank with a thinner smectite zone and disappearance of the illite/smectite zone. This correlation suggests that clay mineral distributions in the Coso Hot Springs field is a function of both the extent of fluid-rock interaction and temperature.

- Whole-rock and feldspar $\delta^{18}\text{O}$ analyses define patterns of $^{18}\text{O}/^{16}\text{O}$ depletion in the West Flank distinct from those expected from regional metamorphism. These oxygen isotope analyses show that the reservoir rocks have experienced a wide range of $^{18}\text{O}/^{16}\text{O}$ depletion, from negligible depletion below primary $\delta^{18}\text{O}$ values of +7.5 ‰ to $\delta^{18}\text{O}$ values as low as -6 ‰ for feldspar. This large range indicates that the reservoir rocks have experienced significant variation in the extent of fluid-rock interaction.
- Detailed $\delta^{18}\text{O}$ measurements of whole rock and feldspar samples from three wells along the West Flank identify a limited number of localized intervals of extensive $^{18}\text{O}/^{16}\text{O}$ depletion where the host rocks have interacted and exchanged isotopically with significant quantities of geothermal fluid (high calculated W/R ratios). The local zones of maximum $^{18}\text{O}/^{16}\text{O}$ depletion in each well correspond closely with the depths of the current production zones (intervals of high permeability). Rocks between these intervals have experienced much less exchange. These patterns likely reflect that permeability in the reservoir host rock is fracture-controlled.
- Most analyzed feldspars have $\delta^{18}\text{O}$ values too high to have completely exchanged with the reservoir fluid. The analyses also show that in general, the lower the $\delta^{18}\text{O}$ value of the feldspar (the greater the extent of fluid-rock interaction), the closer the feldspar approaches isotope exchange equilibrium with the geothermal fluid. This positive

correlation suggests that fracture-induced increases in permeability of the reservoir host rock increase mineral/grain surface area exposed to geothermal fluids, promoting fluid-rock interaction and isotopic exchange.

- There is no systematic correlation of temperature with either the extent of $^{18}\text{O}/^{16}\text{O}$ depletion or the extent of exchange equilibrium between the feldspar and reservoir fluid in the West Flank. This is particularly the case for 73-19, where rock host has experienced only moderate and highly localized $^{18}\text{O}/^{16}\text{O}$ depletion despite having the highest measured temperatures in the three wells. Thus in the West Flank at Coso Hot Springs (a crystalline-rock-hosted geothermal system), permeability is as, or more important than temperature in driving fluid-rock interaction and isotopic exchange.
- The three feldspar samples with the lowest $\delta^{18}\text{O}$ values are all from current production zones in the reservoir. These feldspars are highly depleted in $^{18}\text{O}/^{16}\text{O}$ and therefore have experienced extensive fluid-rock interaction and isotopic exchange. However, two of these feldspar separates have measured $\Delta_{\text{fsp-w}} < \text{equilibrium } \Delta_{\text{fsp-w}}$ which requires that they have equilibrated with the pre-production reservoir fluid at higher temperature, or with a lower $\delta^{18}\text{O}$ reservoir fluid. These discrepancies suggest that the West Flank of the Coso geothermal system was hotter and/or characterized by higher permeability (higher W/R ratio) during fluid-rock exchange in the past.

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