

A Preliminary Geochemical Description of the Geothermal Reservoir at Astor Pass, Northern Pyramid Lake, Nevada

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ABSTRACT

Samples from a well drilled in the Astor Pass area six-km north of the Needle Rocks area of Pyramid Lake indicate that the reservoir fluid is dominantly sodium, chloride, and sulfate, with a pH between 8.6 and 8.9. The total dissolved solids in the reservoir is approximately 1600 mg/l, about half that of the TDS of the fluids in the Needle Rocks area. One sample of dissolved gas from fluids produced during a well test in the reservoir had ^4He value of 2.32×10^{14} atoms $^4\text{He}/\text{g}$ water, or approximately 100 times the value of atmospheric ^4He . This measurement, in conjunction with a R/Ra measurement of 0.28, suggests that most of the reservoir helium is derived from the crust, with possibly a small value (~3.3 percent) derived from the mantle. Tritium concentration of the sample was 0.09 TU, indicating that the reservoir fluid was recharged more than 60 years ago; a simple model based upon carbon-14 suggests recharge has occurred within the past 1500 years.

Introduction

Pyramid Lake, located 45 km north of Reno, Nevada, has been a location of intermittent geothermal exploration since the early 1960s (Garside and Schilling, 1979). At that time, three wells were drilled at the Needle Rocks area, located on the shore of the northern end of the lake (Figure 1). The Needle Rocks area is known for its tufa deposits as high as 90 m above the land surface. Tufa deposits formed where high-calcium groundwater (thermal or nonthermal) discharged into the lake and mixed with dissolved carbon dioxide to precipitate calcium carbonate tufa columns (Coolbaugh et al., 2009). Temperatures as high as 117°C were measured from a well in the Needle Rocks area that continues to flow at high temperature today.

The Astor Pass area is located six kilometers north of Needle Rocks, and is an area that has undergone geothermal explora-

tion in recent years, primarily due to the discovery of a series of northwest-striking faults that cut through Miocene volcanic rocks. The presence of nearby tufa columns and the identification of outcrops of high-temperature alteration minerals kaolinite and halloysite nearby (Coolbaugh, et al., 2006). An exploration well (APS-1) was drilled in 2007 to a depth of 558 m before circulation was lost and the well was abandoned. Two additional wells were drilled in 2010 and 2011 (APS-2 and APS-3) to depths of 1315 m and 1378 m, respectively. The slotted interval in APS-3 (the well in which samples were collected and discussed here) is between 1069 and 1376 m below land surface. Drilling logs and samples of cuttings indicate that in APS-3, from the land surface to 113 m are unconsolidated sand, gravel, and clay, and that interbedded Tertiary volcanic rock exists between 113 m and 1272 m below the land surface. Beneath the volcanic rock are intrusive igneous rocks, primarily granodiorite.

Geochemistry

Fluid samples were collected from well APS-3 in early and mid-2011 (Table 1). The May and June samples were collected from the discharge pipe during a well test, which was begun on May 10, 2011, and therefore the samples are representative of the reservoir (the pumping rate was approximately 450 gallons per minute during the 30-day test, and the pump was set 137 m [450 ft] below the land surface). The samples were collected from a discharge tube attached to the wellhead. The

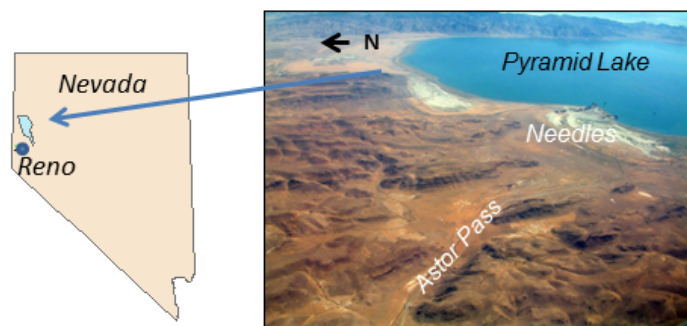


Figure 1. Location of the Astor Pass area north of Pyramid Lake, Nevada.

sample collected on Feb. 12 was collected during an air-lift procedure of the well. The analyses show that the TDS of the reservoir fluid is ~1600 mg/l, with a pH between 8.6 and 8.9, and is dominantly sodium, chloride, and sulfate. In comparison, water from the Needle Rocks 6 km to the south is controlled by the same ions, though in higher concentrations of sodium and chloride (Coolbaugh et al., 2006, Table 1). The high sodium and chloride concentrations are probably due to evapo-concentration of ancestral Lake Lahontan.

The sample was obtained using a 3/8-inch diameter copper tube attached through a clear, flexible tube to a valve located toward the bottom of the main discharge pipe (10 inch diameter) from the well. The discharge pipe was approximately 2/3 full during discharge, and the sample valve was located approximately 30 meters from the end of the discharge pipe; in other words, the gas phase of the upper 1/3 of the pipe was most likely entirely steam from the reservoir and not atmospheric air traveling in the opposite direction from the end of the discharge pipe (located approximately 30 meters downstream). Though there were gas bubbles in the

Table 1. Major ion chemistry (in mg/l unless otherwise noted) from well APS-3.

Parameter	Sample Date		
	2/12/2011	5/11/2011	6/10/2011
Temp, °C		90	90
SiO ₂ – unfiltered	58	63.4	68.4
SiO ₂ – filtered ¹			66.6
Cl	580	564	557
F	3.0		
SO ₄	350	356	363
NO ₃ -N		<0.01	<0.01
HCO ₃	17	2.3	2.3
CO ₃		12.6	12.1
As	<0.005		
B	2.4		
Ca	60	65.5	64.4
CO ₃	8.0	<0.1	<0.1
Cu	<0.05		
Fe	0.71		
Mg	0.04	0.1	0.1
Mn	0.048		
pH, standard units	8.57	8.9	8.9
K	13	11.6	11.0
Na	430	462	454
Total Alkalinity	27		
TDS	1600	1592	1587
EC, µS/cm		2500	2520
Charge, anions/cat		1.004	1.02
⁴ He, cc STP/g H ₂ O			8.67
R/Ra ²			0.28
Tritium, TU			0.09
δ ¹⁸ O _{vsmow} , ‰		-14.5	-14.5
δD _{vsmow} , ‰		-118	-118
δ ¹³ C, ‰			-4.6
¹⁴ C, pmc ³			24.3

¹ Sample was passed through a 0.45 micron filter; both unfiltered and filtered samples underwent 1/10 field dilution.

² R/Ra is the ratio of ³He/⁴He in the sample to ³He/⁴He in air.

³ percent modern carbon.

copper tube, these were most likely to be from outgassing of the fluid as the pressure dropped during the fluid rise in the wellbore, and are therefore representative of the subsurface conditions and are not atmospheric contamination. Fluid from the discharge pipe flowed through the copper pipe for approximately 10 minutes; a second flexible tube with a valve was attached to the downstream end of the copper tube to control flow and to eliminate bubbles in the flexible tubes. The copper tube was repeatedly tapped with a hard object to dislodge entrapped air bubbles located on the tube. When we were satisfied that the bubbles present were from outgassing, and not remnant atmospheric air, a pinch clamp located on the downstream end of the copper tube was tightened, followed by a similar clamp at the upstream end of the tube. The sample was then shipped to the Dissolved Gas Lab at the University of Utah for analysis of ³He/⁴He, as well as tritium. The dissolved gas was extracted from the copper tube under high vacuum and analyzed using a magnetic sector field mass spectrometer.

The ratio ³He/⁴He (also referred to as *R*) and the concentration of ⁴He in a sample can be used to infer the source of fluid in a reservoir. Helium-4 is derived from radioactive decay of uranium and thorium, while ³He is derived from the mantle and from the radioactive decay of tritium. The ratio ³He/⁴He in the atmosphere is approximately 1.38 x 10⁻⁶; often ratios are normalized to this value such that *R/Ra*=1, and *R/Ra* of recharge water is ~1. A noble gas sample from APS-3 resulted in a ⁴He value of 8.67 x 10⁻⁶ cm³ ⁴He stp/g distilled water (2.32 x 10¹⁴ atoms ⁴He /g distilled water). This can be compared to the concentration of ⁴He in the atmosphere, and hence, recharge water. The solubility of helium at sea level, between 10°C and 20°C (an estimated annual temperature of recharge), is 1.3 x 10¹² atoms He/g water. Assuming that the recharge area is 2000 m above sea level, the solubility is adjusted to 10¹². The sample, therefore, contains over 100 times more ⁴He than the atmosphere, suggesting that most of the helium is derived from the crust. A simple mass balance can estimate what percentage, if any, of the helium is derived from the mantle (Kennedy and van Soest, 2006). A typical value of ³He/⁴He, or *R*, for the crust is 0.02*Ra*, while *R* for the mantle is between 8*Ra* and 9*Ra*. *R/Ra* for the sample from APS-3 is 0.28. A simple mass balance calculation shows that perhaps 3.3 percent of the helium in the sample is derived from the mantle. The implication is that the reservoir may be heated by this same small percent of mantle derived heat, but that most of the heat is derived by deep circulating groundwater, as are most of the geothermal systems in the Great Basin.

A crude estimate of the time since recharge (“age”) of the reservoir fluid can be determined knowing the values of ¹⁴C concentration (in percent modern carbon) and ¹³C of the reservoir fluid. The analysis assumes that, except for radioactive decay, the system is closed, and that the initial ¹⁴C value is known. An analysis for both ¹⁴C concentration and δ¹³C (also required in the calculation) was done for a sample from APS-3 using the accelerator mass spectrometry facility on a filtered water sample at the University of Arizona. The δ¹³C and ¹⁴C values are -4.6 ‰ and 24.3 pmc, respectively. Age is determined by solving a differential equation for mass balance with radioactive decay of an element (Clark and Fritz, 1997),

$$t_{age} = -8270 \ln \frac{{}^{14}\text{C}_{sample}}{f {}^{14}\text{C}_{init}}$$

where ^{14}C is the concentration of the sample (as pmc) and the initial concentration (assumed here as 100 pmc), and f is a dilution factor that accounts for mixing with waters of different ^{14}C composition, calcite and dolomite dissolution, exchange of ^{14}C with reservoir rocks, and matrix diffusion. Many models exist for the determination of f (see the review by Fontes and Garner, 1979), but several of them reduce to the following for systems with limited data,

$$f = \frac{\delta^{13}\text{C}_{dic} - \delta^{13}\text{C}_{carb}}{\delta^{13}\text{C}_{soil-CO_2} - \delta^{13}\text{C}_{carb}}$$

Here, $\delta^{13}\text{C}_{dic}$ is the per mil value of the carbon-13 of the dissolved inorganic carbon in the sample, $\delta^{13}\text{C}_{carb}$ is the permil value of carbonate minerals dissolved in the water, and $\delta^{13}\text{C}_{soil-CO_2}$ is the permil value of the soil-gas CO_2 , which is the primary control on ^{13}C concentration (i.e., the carbon signal of precipitation is masked by reactions in the soil zone). The value of $\delta^{13}\text{C}_{carb}$ is usually taken as 0, but $\delta^{13}\text{C}_{soil-CO_2}$ is more difficult to measure, due to the nature of the reactions in the soil zone. Thorstenson et al. (1998) measured values of $\delta^{13}\text{C}_{soil-CO_2}$ at Yucca Mountain and determined a mean of -16 ‰. These values result in $f=0.29$, and a water “age” of 1500 years. This value of age is corroborated by a measurement of tritium in the APS-3 sample of 0.09 ± 0.05 TU. The near absence of tritium in the sample indicates that the water is older than 60 years, and the ^{14}C analysis suggests that it is less than 1500 years. It must be stressed, however, that the age calculation is non-hydrodynamic: it assumes piston flow through a uniform-permeability reservoir with perfect mixing, and without diffusion, dispersion, or the presence of stagnant zones (Neretnieks, 1981; Sanford, 1997; Bethke and Johnson, 2002).

Two analyses of oxygen-18 and deuterium sampled one month apart, gave the same values for both, $\delta^{18}\text{O} = -118$ and $\delta\text{D} = -14.5$ (Figure 2). Also shown in the plot is one sample from a flowing well at the Needle rocks area. Both samples show the characteristic oxygen-shift, due to an oxygen exchange reaction with the host rock. The greater shift for the Needle Rock sample is characteristic of that much hotter, and reactive, system. Although the deuterium values for the two samples are different by 11.5‰, more samples are needed in order to determine if the recharge areas of the two systems are similar.

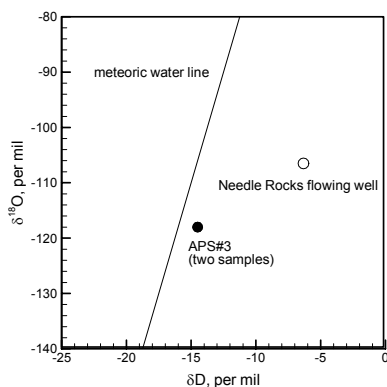


Figure 2. Relation between deuterium and oxygen-18 for two samples (the points are superimposed) from APS#3 and a single sample from a flowing well located in the Needle Rocks area (from the Nevada Bureau of Mines and Geology Great Basin Groundwater Geochemical Database).

Reservoir Temperature

Discharge temperature throughout duration of well test was 90°C . A quartz geothermometer (Fournier, 1982), which

estimates reservoir temperature without steam loss, gives a reservoir temperature of 115°C . Some of the assumptions required for silica geothermometry are probably met for its application; i.e., the silica content for the pressure-temperature conditions is in equilibrium in the reservoir, the reservoir fluid is unable to re-equilibrate as it rises to the surface in the well (reaction constants for silica are on the orders of hours to days while the circulation [lag] time in the well at 450 gpm [1.7 m³/min] is less than 45 minutes), and there is no mixing of geothermal water with cooler subsurface water (the upper 1069 meters of formation is cased off). The geothermometer assumes silica is saturated in the reservoir; we collected two samples at 1/10 field dilution within minutes of each other, the difference being that one was passed through a 0.45 micron filter. The SiO_2 concentration is within a few percent for both samples, suggesting that there may not be precipitable SiO_2 in the reservoir. The total dissolved solids content of the samples was low, approximately 1600 mg/l, so that corrections for changes in enthalpy due to high TDS are unwarranted.

The problem with silica geothermometry is that silica solubility is dependent upon the mineral form in which it exists. If well-crystallized quartz controls solubility, then the quartz geothermometer may be accurately estimate reservoir temperature. If, however, silica solubility is controlled by much finer-grained (and higher surface energy) chalcedony, the reservoir temperature is calculated to be lower due to the higher solubility of the chalcedony. The chalcedony geothermometer gives a reservoir temperature of 86°C , slightly lower than the temperature at the wellhead. At the other extreme, a Na-K-Ca geothermometer, which is based upon temperature-dependent exchange reactions that go to completion, gives a reservoir temperature of 155°C .

More important than determining the correct geothermometer is to determine if the error associated with chemical geothermometry, which assumes that there the reactant minerals are initially in equilibrium (so that the geothermometer can be “set”), and that there is minimal mixing or re-equilibration of water as it travels to the surface (Fournier, 1977). In a numerical study of chemical geothermometry, Ferguson et al. (2009) determined that maximum in situ temperatures could be under-predicted by up to 30 percent due to mixing of fluids entering a fault at depth.

Discussion and Summary

Analysis of three water samples collected from a single well located in the Astor Pass area indicate that the reservoir fluids are of a sodium, chloride, sulfate type, with a pH between 8.6 and 8.9. These three elements are absent in the fractured volcanic reservoir rocks, and are probably derived from dissolution of sodium chloride and sulfate minerals by precipitation recharging through areas that were once the bed of Lake Lahontan. Carbon isotopes suggest that the “age” of the reservoir fluids is less than 1500 year old, but this value is suspect due to assumptions that may be poorly met. The system is most certainly heated by deep circulation of groundwater, as indicated by analysis of helium isotopes. Geothermometry suggests are reservoir temperature in the area of 115°C . Among the intriguing questions that remain, the most important is to determine if there are hotter fluids in the area that can be developed into a productive reservoir.

Acknowledgments

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