

Quantifying the flux of hydrothermal fluids into Mono Lake by use of helium isotopes

Jordan F. Clark¹

Department of Geological Sciences, University of California, Santa Barbara, California 93106

G. Bryant Hudson

Isotope Sciences Division, Lawrence Livermore National Laboratory, Livermore, California 94551

Abstract

In Mono Lake, concentrations of ³He and ⁴He are greater than atmospheric equilibrium values, indicating a subsurface helium source. This assertion is supported by vertical concentration profiles that show maximum $\delta^3\text{He}$ values (+68%) occurring within the pycnocline. The slope of the regression between ³He and ⁴He concentrations is near 2.8 R_a (R_a is the atmospheric helium isotope ratio), indicating that the subsurface source has a component of mantle helium. Hydrothermal springs and gas vents from Paoha Island have similar ratios, which suggests that discharge from this thermal system is the most important source. A ³He mass balance indicated that below the deep pycnocline at 17 m, ~ 25 cc STP of ³He accumulates each year ($\sim 90\%$ from the influx of hydrothermal water and $\sim 10\%$ from the in situ decay of tritium). Although it is not retained because of gas transfer across the air–water interface, ~ 70 cc STP of ³He is injected each year into the upper 17 m. When ³He is used as a tracer, the influx of hydrothermal water below the deep pycnocline was estimated to be 0.045 ± 0.025 m³ s⁻¹. Because the hydrothermal water is fresher than Mono Lake, this influx causes the salinity of the monimolimnion to decrease by ~ 0.1 salinity units per year and may play a small but important role in the salinity budget of this layer, which is presently denser than and isolated from the surface water.

The isotopes of helium (³He and ⁴He) have become important tracers for continental fluids for a number of reasons. First, helium isotope ratios vary significantly between the different reservoirs in the geosphere. The ³He/⁴He isotope ratio of the atmosphere ($\sim 1.4 \times 10^{-6}$) lies between that of the deep crust ($< 5 \times 10^{-8}$) and that of the upper mantle ($\sim 10^{-5}$). Second, continental fluids retain in their helium isotopes a record of the different reservoirs they have contacted. Hence, in bodies of water—such as lakes that receive the influx of water from many sources—the helium isotope signature can be used to identify the different sources. Third, the isotopic composition of helium can change as a result of the production of ³He by decay of tritium or ⁴He from α -decay. Tritium, which has a half-life of 12.43 yr, is naturally produced in the atmosphere by cosmic rays and enters the hydrologic systems via precipitation. In the late 1950s and early 1960s, the global tritium inventory was dramatically increased by atmospheric thermonuclear bomb tests. In waters that have turnover times < 50 yr, the combination of ³He and tritium can be used to calculate isolation times from the atmosphere with uncertainties typically better than ± 2 yr.

¹ Corresponding author (clark@magic.ucsc.edu).

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Gas exchange at the air–water interface maintains dissolved helium concentrations and other gases near their solubility equilibrium value. These gases are often found to be slightly (a few percent) supersaturated, because breaking waves forces air bubbles below the surface, where they dissolve. During periods of stratification, gas-exchange processes are not important below the mixed layer, and gases can accumulate. Concentrations of ³He and ⁴He are often significantly greater than their solubility equilibrium values, indicating subsurface sources of helium. In most lakes, the ³He/⁴He ratio is greater than that of the atmosphere, indicating that helium sources are isotopically light. They have been identified as either the influx of hydrothermal fluids associated with mantle degassing (Sano et al. 1990; Igarashi et al. 1992; Kipfer et al. 1994; Aeschbach-Hertig et al. 1996b) or the in situ decay of tritium (Torgersen et al. 1977; Zenger et al. 1990; Aeschbach-Hertig et al. 1996a). When the latter source dominates, the timescale of circulation is recorded in the helium isotopes, and it can be estimated if tritium is simultaneously analyzed. These two sources of isotopically light helium can be distinguished by their ³He:⁴He ratio. Hydrothermal fluids are also a source of ⁴He, while the decay of tritium is not.

Isotopically heavy helium has been identified in a few lakes that receive the influx of old groundwater (Torgersen and Clarke 1978; Aeschbach-Hertig et al. 1996a). Groundwater that has circulation times greater than a few thousand years typically has helium concentrations more than two orders of magnitude greater than atmospheric equilibrium values (e.g., Bottomley et al. 1984; Torgersen and Clarke 1985; Stute et al. 1992; Gascoyne and Sheppard 1993; Clark et al. 1998), thus small groundwater fluxes can be easily traced with helium.

Here, we present helium isotopes from Mono Lake and show that isotopically light helium accumulates below the

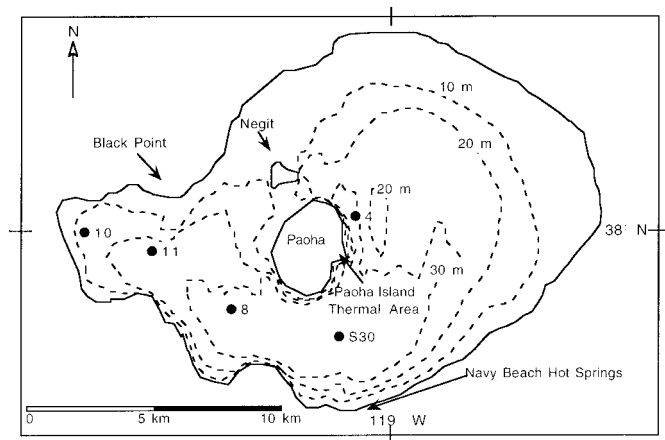


Fig. 1. Bathymetric map of Mono Lake showing the locations of the fixed buoys, Blacks Point, and the Paoha Island and Navy Beach thermal area. Contour intervals are in meters.

surface. In this lake, the helium isotopic concentration is affected by both the influx of hydrothermal fluids and the decay of tritium. Because Mono Lake has recently become meromictic (persistent chemical stratification), helium isotopes are good tracers for the influx of hydrothermal fluids into the monimolimnion (the deep water, which does not annually mix with surface water and ventilate) and, potentially, are important tracers of reduced gases that are presently accumulating.

Mono Lake—Mono Lake is a large, moderately deep, hypersaline, closed-basin lake located in a tectonic basin east of the Sierra Nevada, California (Fig. 1). It is a highly productive lake (Jellison and Melack 1993a) and an important resting place for a number of migratory birds. Studies of the basin's sediments suggest that the lake has existed for hundreds of thousands of years and that its size has fluctuated with long-term global climate change (Gilbert et al. 1968; Benson et al. 1990). In 1940, prior to the diversion of its feeder streams for drinking water, Mono Lake's surface elevation was 1955 m, ~225 m below its last high stand 14,000 years ago (Benson et al. 1990; Stine 1991).

Mono Basin is volcanically active (e.g., Gilbert et al. 1968; Bursik and Sieh 1989). The Mono Craters, a series of volcanic cones located south of the lake, formed during the last 40,000 yr, and Black Point, which is located along the northwestern shore, formed beneath Mono Lake during its last high stand, ~13,000 yr ago. Within the present lake lie two large volcanic islands, Negit and Paoha, which formed during the last 2,000 yr. A number of warm- and hot-spring areas are associated with this recent volcanism, which include the Navy Beach springs along the southern shore and the Paoha Island thermal area (Fig. 1).

Since 1941, diversion of freshwater streams out of Mono Basin has caused the lake level to drop by >12 m and has doubled its salinity (Stine 1991). In 1994, Mono Lake began a new phase, when the California State Water Resources Control Board (decision 1631) temporarily suspended the freshwater diversions and its lake level began to rise (Loomis 1995; Jellison et al. 1998). It is expected that it will take

~20 yr to reach the target elevation of 1948 m, ~5 m above the lake level at the time of the decision.

During its first year of recovery, the lake level rose by >1 m. This rapid rise changed the vertical mixing regime (Jellison et al. 1998). With the exception of a 6-yr period during the mid-1980s, Mono Lake has been monomictic (surface and bottom waters mix completely once a year) for at least the last 30 yr while the lake level was declining (Romero and Melack 1996). However, because the lake (salinity ~85 g kg⁻¹) is much more saline than the runoff water (salinity <0.1 g kg⁻¹), the recent rise in lake level has altered this mixing pattern. The influx of freshwater to the surface of Mono Lake has created a density contrast between the surface and bottom waters, and a second period of meromixis has begun (Jellison et al. 1998). Jellison et al. (1998) recently predicted, using a modified version of the DYRESM lake model (Imberger and Patterson 1981) developed by Romero and Melack (1996), that Mono Lake will remain meromictic for 20–60 yr.

Methods

Helium isotope samples were collected from Mono Lake on 14 Oct 1995, 7 Mar 1996, and 19 Sep 1996. Each time, samples were collected from a small boat moored to fixed buoys by lowering a 2-liter van Dorn bottle (Fig. 1; Table 1). The depth of collection was determined by the length of line out, and vertical profiles were created by lowering the van Dorn bottle multiple times at the same location. Samples for helium isotopes were collected in 10-cm copper tubes sealed with stainless steel pinch-off clamps by connecting the copper tubes to the van Dorn bottle with a short piece of tygon tubing. The copper tubes were flushed thoroughly, to ensure that no air was sealed along with the water sample. Tritium samples were also collected in glass bottles at a few depths at Sta. S30 during the Oct 1995 and Sep 1996 sampling surveys. Profiles of temperature and conductivity were obtained by lowering a Seabird conductivity, temperature, depth (CTD) probe (Seacat SBE 19) at each station.

At Lawrence Livermore National Laboratory, the copper tubes were attached to a high vacuum inlet system that led to a VG5400 noble-gas mass spectrometer. Helium was extracted by boiling under vacuum and separated from other gases by a series of cold traps and titanium getters. The mass spectrometer was calibrated with equilibrated water samples (both fresh and Mono Lake water) and known quantities of air. The uncertainties of ³He:⁴He isotope ratio and ⁴He concentration measurements are 0.5% and 1%, respectively.

³He:⁴He isotope ratios are reported as percentage deviations from air, by use of the common δ -notation

$$\delta^3\text{He} = \left(\frac{R}{R_a} - 1 \right) \times 100,$$

where R and R_a are the ³He:⁴He isotope ratio of the sample and of the atmosphere (1.384×10^{-6} ; Clarke et al. 1976), respectively. ³He concentrations were calculated by use of the above equation from the measured ³He:⁴He isotope ratios and the ⁴He concentrations.

Tritium, which is reported in tritium units (TU, which is

Table 1. Helium isotope ratios and concentrations in Mono Lake from stations S30 and 4.

| Depth (m) | $\delta^3\text{He}$ (%) | Station S30 [^4He] (10^{-8} cc STP g^{-1}) | [^3He] (10^{-14} cc STP g^{-1}) | Depth (m) | $\delta^3\text{He}$ (%) | Station 4 [^4He] (10^{-8} cc STP g^{-1}) | [^3He] (10^{-14} cc STP g^{-1}) |
|-------------|-------------------------|---|---|-----------|-------------------------|---|---|
| 14 Oct 1995 | | | | | | | |
| 2 | 3.8 | 2.27 | 3.27 | 7 | 3.3 | 2.27 | 3.25 |
| 8 | 2.6 | 2.22 | 3.16 | 11 | 23.7 | 2.34 | 4.02 |
| 10 | 13.8 | 2.30 | 3.63 | 14 | 34.5 | 2.39 | 4.46 |
| 12 | 33.2 | 2.38 | 4.39 | 28 | 25.2 | 2.31 | 4.00 |
| 14 | 28.7 | 2.33 | 4.16 | | | | |
| 17 | 30.7 | 2.35 | 4.26 | | | | |
| 23 | 25.3 | 2.29 | 3.98 | | | | |
| 30 | 20.6 | 2.24 | 3.74 | | | | |
| 37 | 21.1 | 2.26 | 3.79 | | | | |
| 7 Mar 1996 | | | | | | | |
| 2 | 1.4 | 2.24 | 3.15 | 2 | 0.7 | 2.25 | 3.15 |
| 7 | lost | lost | lost | 8 | 2.2 | 2.27 | 3.22 |
| 12 | 5.1 | 2.27 | 3.31 | 15 | 27.6 | 2.38 | 4.21 |
| 15 | 16.4 | 2.33 | 3.75 | 18 | 37.3 | 2.38 | 4.54 |
| 18 | 43.7 | 2.44 | 4.86 | 28 | 34.2 | 2.33 | 4.33 |
| 23 | 37.3 | 2.38 | 4.53 | | | | |
| 28 | 34.8 | 2.35 | 4.40 | | | | |
| 34 | 31.8 | 2.35 | 4.29 | | | | |
| 19 Sep 1996 | | | | | | | |
| 2 | 3.1 | 2.23 | 3.18 | 2 | -0.8 | 2.30 | 3.16 |
| 10 | 2.4 | 2.30 | 3.26 | 9 | lost | lost | lost |
| 13 | 23.1 | 2.36 | 4.02 | 11 | 13.0 | 2.44 | 3.82 |
| 15 | 38.0 | 2.47 | 4.71 | 14 | 31.1 | 2.48 | 4.49 |
| 17 | 52.8 | 2.50 | 5.28 | 16 | lost | lost | lost |
| 20 | 67.6 | 2.56 | 5.95 | 19 | 51.1 | 2.43 | 5.07 |
| 25 | 53.2 | 2.45 | 5.19 | 24 | 47.6 | 2.43 | 4.97 |
| 33 | 46.2 | 2.43 | 4.91 | 29 | 46.4 | 2.42 | 4.90 |

equivalent to a tritium-to-hydrogen ratio of 10^{-18}), was measured by a ^3He accumulation method in which the samples were vacuum degassed and shelved for ~ 60 d to allow for the growth of ^3He from tritium decay (Surano et al. 1992). The uncertainty in the tritium measurement is ± 1 TU.

Results

The initial sampling of Mono Lake for helium isotopes in Oct 1995 was carried out ~ 8 months after the last period of deep winter mixing. Prior to this sampling, autumn cooling and storm activity had eroded the seasonal thermocline, and the mixed layer depth was 12 m (Fig. 2). The second sampling event was carried out 5 months later, in Mar 1996. The unusual temperature profile, which shows a 2°C warming at 17 m, is indicative of the change in mixing regime that occurred after the rapid rise in lake level during the previous year (Jellison et al. 1998). Because of surface freshening caused by the large amount of runoff in 1995, winter mixing was able only to penetrate the top 17 m of the water column, creating a salinity gradient coincident with the deep thermocline (Fig. 2). Thus, the bottom water (the monimolimnion) was not ventilated during the 1995–1996 winter. The final sampling of Mono Lake was carried out in Sep 1996. The temperature profile shows that the seasonal thermocline had eroded significantly prior to sampling and that the deep

thermocline remained unchanged. (Hereafter, we will refer to the “deep thermocline” as the “pycnocline,” to distinguish it from the seasonal thermocline.) Throughout the collection period, the monimolimnion remained isolated from the atmosphere, and gases that accumulated there were not lost from the lake because of gas exchange at the air-water interface.

Most of the helium isotope samples were obtained at Sta. S30, which lies in the middle of a deep basin 2 km south of Paoha Island and at Sta. 4 which lies ~ 1 km east of Paoha Island (Fig. 1). Vertical profiles of helium isotopes from these two stations showed that $\delta^3\text{He}$ values were greater than solubility equilibrium values with the atmosphere ($\delta^3\text{He} = -1.7\%$; Benson and Krause 1980) at all depths and at all times (Table 1; Fig. 2). The minimum and maximum values were found in the mixolimnion (the surface mixed layer) and the upper pycnocline, respectively. Although $\delta^3\text{He}$ values of the mixolimnion did not change significantly between surveys, both the maximum value in the pycnocline and the mean value of the monimolimnion increased with time. ^3He concentrations increased at rates of $\sim 1.2 \times 10^{-14}$ and 1.8×10^{-14} cc STP $\text{g}^{-1} \text{yr}^{-1}$, respectively, in these layers (Fig. 3).

With the exception of one sample, which was collected in Sep 96 near the top of the pycnocline, the pycnocline and monimolimnion samples fall along parallel lines on a

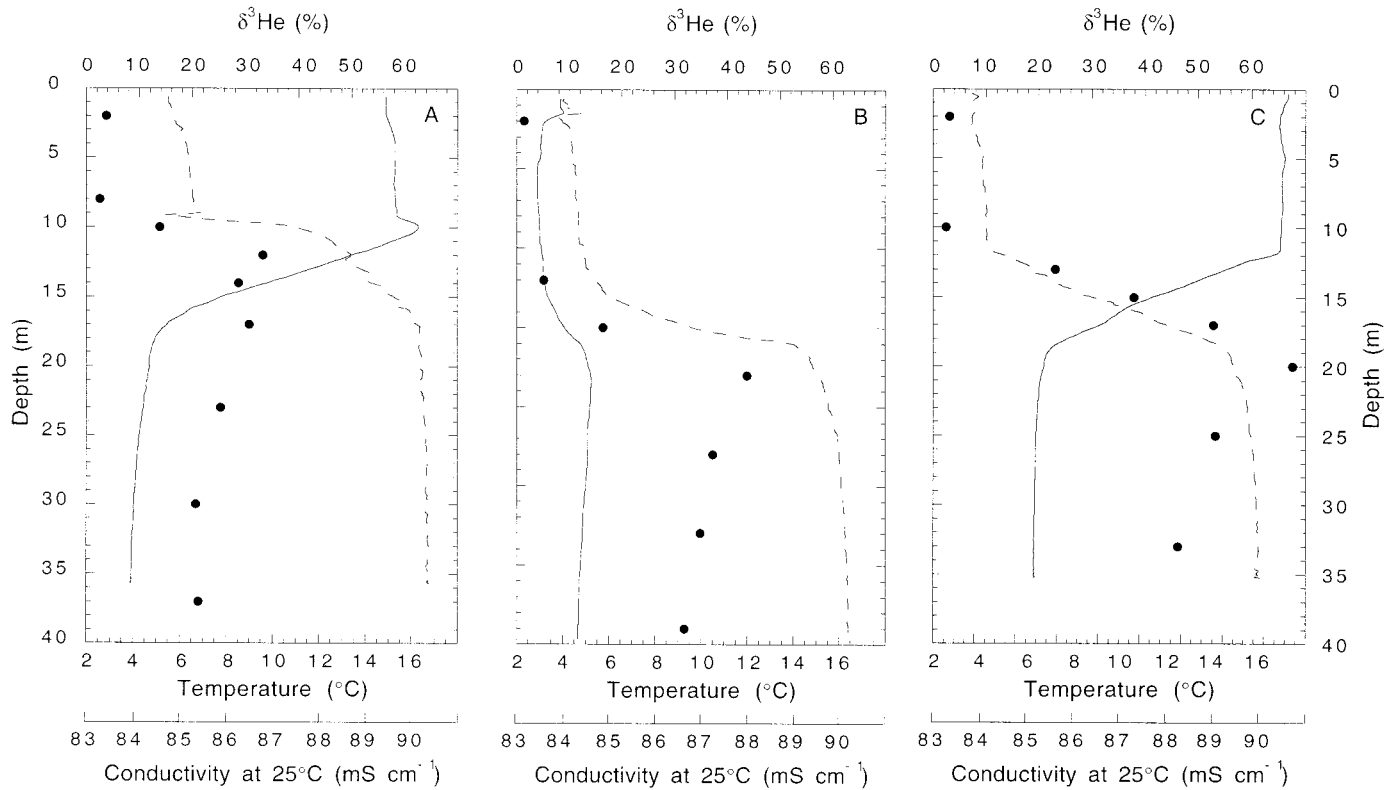


Fig. 2. Vertical profiles of $\delta^3\text{He}$ (circles), temperature (solid line), and conductivity (dashed line) at Sta. S30 on (A) 14 Oct 1995, (B) 7 Mar 1996, and (C) 19 Sep 1996. The helium isotope samples were collected independently from temperature and conductivity by repeatedly lowering a van Dorn bottle to different depths.

plot of ^3He concentration versus ^4He concentration (Fig. 4). Although the best-fit slope for the Oct 95 data is $3.2 R_a$, the data are consistent with slopes that range between 2.7 and $4.1 R_a$ when the error in ^4He concentrations is considered. The range of slopes for the data collected later is much larger. The lines step sequentially to higher ^3He intercepts, with time indicating that ^3He from the decay of

tritium accumulated between each sampling event. In Oct 1995, the tritium concentration showed no variation with depth and averaged 16.2 TU. Eleven months later, the tritium content below the deep thermocline (<20 m) was 15.4 TU. This change agrees with the expected change due to the decay of tritium in a closed system. The decay of this

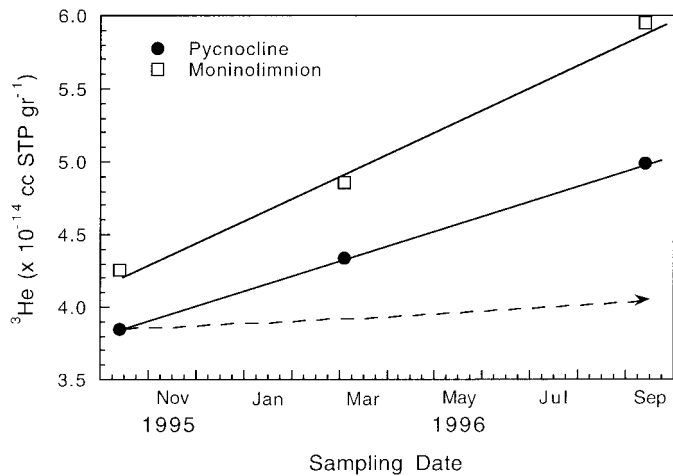


Fig. 3. Accumulation rate of ^3He in the pycnocline and monimilnion. The dashed arrow represents the accumulation rate of ^3He from the in situ decay of tritium.

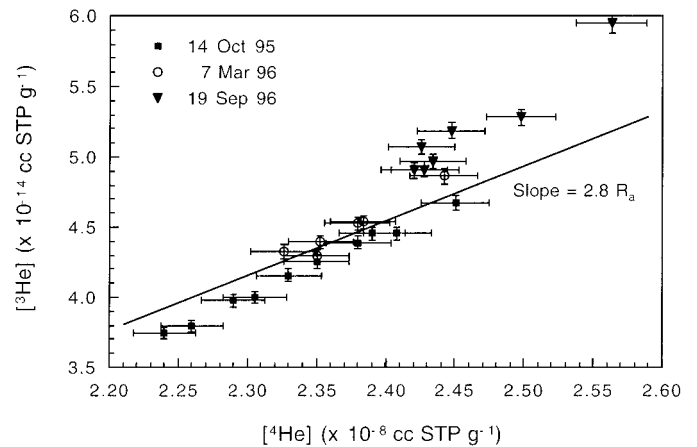


Fig. 4. Helium isotope concentration plot of samples collected below the mixolimnion, showing the good correlation between ^3He and ^4He . This is indicative of a hydrothermal source. The line, which has a slope equivalent to 2.8 times the atmospheric helium isotope ratio ($2.8 R_a$), was not determined by fitting data.

Table 2. Surface helium isotope data (2 m) collected on 19 Sep 1996.

| Station | Distance* (km) | $\delta^3\text{He}$ (%) | [^4He] (10^{-8} cc STP g^{-1}) | [^3He] (10^{-14} cc STP g^{-1}) |
|---------|----------------|-------------------------|--|---|
| 4 | 1.9 | -0.8 | 2.30 | 3.16 |
| S30 | 3.3 | 3.1 | 2.23 | 3.18 |
| 8 | 5.6 | 3.0 | 2.22 | 3.16 |
| 11 | 8.9 | -1.2 | 2.39 | 3.27 |
| 10 | 11.9 | 2.5 | 2.32 | 3.29 |
| | mean | 1.3 ± 2.1 | 2.29 ± 0.07 | 3.21 ± 0.06 |

* Approximate lake distance from the Paoha Island hot springs.

tritium added $\sim 0.2 \times 10^{-14}$ cc STP g^{-1} yr^{-1} of ^3He to the water column, which is significantly less than the observed increase (Fig. 3) and is in reasonably good agreement with the observed offset when the ^4He concentration errors are considered (Fig. 4).

Because ^3He accumulated at a rate faster than tritium decay could have supported, there must be an additional isotopically light source, such as one associated with the influx of hydrothermal fluids. The good correlation between ^3He and ^4He concentrations (Fig. 4) indicates that both isotopes of helium are accumulating in the deep water and supports the above assertion.

Helium isotope samples were collected from the Paoha Island thermal area in Mar 1996. The $^3\text{He}:^4\text{He}$ isotope ratio of one hydrothermal spring and one gas vent was $2.7 R_a$, which is in good agreement with measurements reported earlier by Welhan et al. (1988). The helium concentration of the spring water was 4.5×10^{-6} cc STP g^{-1} , nearly 200 times its concentration in the lake. The similarity between the ratios of the water column source and the Paoha Island thermal fluid suggests that this hydrothermal system is most likely the source of helium to Mono Lake. Other geochemical evidence for the influx of very old water associated with volcanic fluids include ^{14}C and ^{36}Cl (Oxburgh et al. 1991; Phillips et al. 1995).

In Sep 1996, a transect of surface stations was sampled from Sta. S30 westward, to assess the degree of saturation far from the hot spring source located on Paoha Island so that the air-water flux could be estimated. Surface values varied between -1.2% and 3.1% and averaged $1.3\% \pm 2.1\%$ (Table 2). The concentration did not vary systematically with distance, most likely because of the complex circulation of the surface water and variations in fetch and wind speed, which are important factors affecting gas-transfer velocities. $\delta^3\text{He}$ values of surface samples collected during the earlier two sampling events were not significantly different than those observed in Sep 1996.

Helium mass balance—The annual influx of ^3He into the upper 16 m of the water column could not be calculated with the observed change in concentration because of seasonal ventilation, as is illustrated by the change in inventory between the Oct 1995 and Mar 1996 sampling events (Table 3). However, the loss across the air-water interface due to gas-exchange processes can be estimated. At the time of each sampling event, the mean flux of ^3He across the air-

Table 3. Accumulation rates of ^3He (cc STP yr^{-1}) in Mono Lake between sampling events. Positive numbers are into the lake.

| Depth (m) | Oct 95 to Mar 96 | Mar 96 to Sep 96 | Oct 95 to Sep 96 |
|------------------|------------------|------------------|------------------|
| Surface* | -70 | -70 | -70 |
| 0-16 | -11.7 | 6.9 | -0.9 |
| 17-25 | 8.8 | 11.6 | 10.4 |
| 26-45 | 4.2 | 3.6 | 3.8 |
| 0-45 | 1.2 | 22.1 | 13.3 |
| Δt (yrs) | 0.39 | 0.54 | 0.93 |

* The flux across the air-water interface assuming a mean gas transfer velocity of 5 cm h^{-1} (Wanninkhof et al. 1987, 1991) and a mean ^3He surface concentration 3% greater than the solubility equilibrium value (see Table 2).

water interface, based on the observed mean surface concentration (Tables 1, 2) and an assumed gas-transfer velocity for helium of 5 cm h^{-1} (Wanninkhof et al. 1987, 1991), was ~ 70 cc STP yr^{-1} .

The water column was not ventilated below 17 m, so the accumulation rate of ^3He can be calculated from the observed change in concentration between sampling events, detailed bathymetric data (Pelagos unpubl. data), and an estimate of the loss through the pycnocline via vertical mixing. Between 17 and 25 m and below 26 m, the mean observed ^3He accumulation rates were 10 and 4 cc STP yr^{-1} , respectively (Table 3). Loss via vertical mixing can be calculated from the ^3He concentration gradient, which averaged 2.2×10^{-15} cc STP $\text{g}^{-1} \text{ m}^{-1}$, and the vertical eddy diffusivity, k_z . Using a heat budget technique, Jellison and Melack (1993b) and Romero et al. (1998) estimated that the average k_z in the seasonal thermocline of Mono Lake is $\sim 4 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$. Hence, under the assumption that k_z in the pycnocline equals the average value for the seasonal thermocline, the ^3He flux from the monimolimnion to the mixolimnion at 17 m is 25 cc STP yr^{-1} . The vertical eddy diffusivity is probably greater in the seasonal thermocline than in the pycnocline, because the pycnocline is found lower in the water column. Hence, the calculated loss of 25 cc STP yr^{-1} due to vertical mixing is a maximum. If the eddy diffusivity coefficient in the pycnocline is $< 1 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$, this loss is negligible.

The total flux of ^3He into Mono Lake's monimolimnion is the sum of the observed concentration increase and loss via vertical mixing. Thus, the total influx is between 14 and 39 cc STP yr^{-1} , depending on the loss rate through the pycnocline. The influx of hydrothermal fluids accounts for >90% of the total flux of ^3He , and the remainder results from the decay of tritium (~ 2 cc STP yr^{-1}).

Discussion

The basic shape of the vertical $\delta^3\text{He}$ profile in Mono Lake is similar to oceanic profiles collected near the East Pacific Rise in the South Pacific and elsewhere (e.g., Lupton and Craig 1981). The processes controlling the distribution of ^3He in these two systems are similar. $\delta^3\text{He}$ values close to atmospheric equilibrium values are maintained in the mixolimnion by gas exchange, maximum values are caused by

a mid-depth source of ^3He , and intermediate values below the middepth source reflect circulation. The intermediate values found in Mono Lake's monimolimnion below 20 m reflect winter ventilation that occurred during 1994–1995, prior to the start of the time series and partial isolation from the middepth source caused by stratification and slow vertical mixing. Prior to the 1995–1996 winter, this stratification was caused by the seasonal thermocline; subsequently, the stratification was caused by a salinity gradient that developed after the onset of meromixis.

Welhan et al. (1988) showed that ratios of helium isotopes in hydrothermal springs vary geographically within Mono Basin. The Navy Beach springs, south of the lake, have the highest $^3\text{He}:^4\text{He}$ isotope ratio (5.2–5.8 R_a) whereas a small spring near Black Point had the lowest ratio (1.8 R_a). Welhan et al. (1988) measured a submerged gas vent near Paoha Island and found the $^3\text{He}:^4\text{He}$ isotope ratio to be 2.7 R_a , approximately the same as our measurements that were made a decade later. The good agreement between the inferred isotopic composition of Mono Lake's dominant helium source (Fig. 4) and measurements from the Paoha thermal area, combined with the Mono Basin's geographical trend, indicates that the Paoha Island thermal area is the most likely source of helium to Mono Lake.

Helium mass balance calculations indicate that ~ 95 cc STP yr (3×10^4 atoms $\text{m}^2 \text{s}^{-1}$) of ^3He enters Mono Lake. On a per area basis, this influx is two to five orders of magnitude less than what has been reported for other lakes that also receive discharge from hydrothermal systems (Sano et al. 1990; Igarashi et al. 1992; Kipfer et al. 1994; Aeschbach-Hertig et al. 1996b). Oxburgh et al. (1991) estimated the flux of ^{14}C -free CO_2 into Mono Lake to be ~ 3.3 mol $\text{m}^{-2} \text{yr}^{-1}$ (5×10^8 mol yr^{-1}). Under the assumption that the ^{14}C -free CO_2 flux is entirely associated with the Paoha Island springs, the C: ^3He ratio of the hydrothermal fluid is $\sim 1 \times 10^{11}$. This is approximately two orders of magnitude greater than the ratio found in MORB (Mid-Ocean Ridge Basalt) (Marty and Jambon 1987) and an order of magnitude greater than the ratio found in Lake Nyos, Cameroon (Sano et al. 1990) and in hot springs from Long Valley (Sorey et al. 1993). This relatively high ratio, combined with the relatively high $^3\text{He}:^4\text{He}$ ratio, suggests that the hydrothermal gases are a mixture of crustal and mantle sources.

One sample collected in Sep 96 from the top of the pycnocline falls off the general trend on Fig. 4, suggesting that there is a small source that has a $^3\text{He}:^4\text{He}$ isotope ratio $>2.8 R_a$. This could be associated with a small flux of helium from a different location, such as near the Navy Beach Spring system. Alternatively, it could indicate that the Paoha Island hydrothermal system has a larger range or more temporal variability in its helium isotopic composition than is represented by the few samples collected in this study and by Welhan et al. (1988). Nevertheless, the bulk of the samples suggest that the subsurface helium source to Mono Lake has an isotopic composition similar to that of the Paoha Island hydrothermal fluid.

Surface expressions of the Paoha Island thermal area include hot springs and gas vents that are found on the island and submerged offshore. The submerged springs create plumes of hot water and bubbles that rise through the water

column, indicating that hydrothermal helium is being injected between the lake floor vents and the surface. On the basis of the location of the $\delta^3\text{He}$ maxima, these vents must be emitting fluids between the surface and ~ 20 m. The helium mass balances indicates that the majority of ^3He ($>80\%$) is discharged into the lake above the pycnocline and is quickly lost via gas exchange.

Because the Paoha Island hydrothermal water (~ 25 g kg^{-1} ; Rogers and Driess 1996) is fresher than Mono Lake, and because it is being injected below the pycnocline at 17 m, this system may play a small but important role in the monimolimnion salinity budget. The helium mass balance indicates that between 12 and 37 cc STP of hydrothermal ^3He enters the monimolimnion each year, depending on the estimate of the vertical eddy diffusivity in the pycnocline. Under the assumption that this helium is entirely dissolved in water with a concentration equal to the observed concentration from one spring on Paoha Island (1.7×10^{-11} cc STP g^{-1}), the influx of hydrothermal water to the monimolimnion is 0.045 ± 0.025 $\text{m}^3 \text{s}^{-1}$. This is equivalent to adding ~ 8 mm of water to the present surface of Mono Lake (area 170 km^2) each year. As a result of this influx, the salinity of the monimolimnion will decrease by ~ 0.1 g $\text{kg}^{-1} \text{yr}^{-1}$.

Recent analysis of the salinity data collected since the onset of meromixis by use of the modified DYRESM lake model (Jellison et al. 1998) suggests that more freshening is occurring in the monimolimnion than can be explained by vertical mixing (R. Jellison pers. comm.). The helium data suggest that the influx of hydrothermal water may be the cause of this additional freshening. Future models need to include this water flux when the duration of the current period of meromixis is calculated.

The calculated water flux from the helium mass balance is based on the assumption that below 17 m, gas vents are not important sources of ^3He to Mono Lake. Oremland et al. (1987) located numerous gas seeps in this lake, some as deep as 20 m. Most of the seeps, including those found below 17 m, were pelagic (biogenic) methane sources, rather than vents associated with the Paoha Island hydrothermal system. Nevertheless, it is not possible at this time to rule out the existence of gas vents at these depths, and, thus, the calculated influx of hydrothermal water is a maximum value.

In addition to being a good tracer of hydrothermal fluids in Mono Lake, ^3He is a potential tracer for reduced gases such as CH_4 and H_2S , which are presently accumulating in the monimolimnion as a result of the onset of meromixis (Oremland et al. 1987; Miller et al. 1993). In the future, as the deep pycnocline erodes, these reduced gases will be transported into the mixolimnion, where they will either be lost to the atmosphere via gas transfer or cycled through (bio)chemical reactions. ^3He could be used to assess the loss via gas transfer across the air-water interface, so that potential impacts of these reduced gases on the biogeochemistry could be assessed.

Conclusions

As is the case for many lakes in volcanic terrains, isotopically light helium accumulates below the surface of

Mono Lake. Most of the helium influx, which totals ~ 95 cc STP of ^3He yr $^{-1}$, is associated with the Paoha Island hydrothermal system. The remainder results from the in situ decay of tritium. In the mixolimnion, ^3He is quickly lost from the lake to the atmosphere via gas exchange across the air–water interface. However, below 17 m, ^3He is accumulating at a rate of $\sim 1.4 \times 10^{-14}$ cc STP g $^{-1}$ yr $^{-1}$ (14 cc STP yr $^{-1}$). On the basis of a ^3He mass balance, ~ 0.045 m 3 s $^{-1}$ of hydrothermal water is entering the lake below the deep pycnocline, under the assumption that the flux from gas vents is negligible. Because the hydrothermal water is fresher than Mono Lake, this influx may play a small but important role in the salinity budget of the monimolimnion. Over a period of 20 yr, this influx will decrease the salinity of the monimolimnion by ~ 2.0 g kg $^{-1}$. Model calculations of the duration of the current period of meromixis need to consider this water flux.

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