Using noble gases to investigate mountain-front recharge

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Abstract

Mountain-front recharge is a major component of recharge to inter-mountain basin-fill aquifers. The two components of mountain-front recharge are (1) subsurface inflow from the mountain block (subsurface inflow), and (2) infiltration from perennial and ephemeral streams near the mountain front (stream seepage). The magnitude of subsurface inflow is of central importance in source protection planning for basin-fill aquifers and in some water rights disputes, yet existing estimates carry large uncertainties. Stable isotope ratios can indicate the magnitude of mountain-front recharge relative to other components, but are generally incapable of distinguishing subsurface inflow from stream seepage. Noble gases provide an effective tool for determining the relative significance of subsurface inflow, specifically. Dissolved noble gas concentrations allow for the determination of recharge temperature, which is correlated with recharge elevation. The nature of this correlation cannot be assumed, however, and must be derived for the study area. The method is applied to the Salt Lake Valley Principal Aquifer in northern Utah to demonstrate its utility. Samples from 16 springs and mine tunnels in the adjacent Wasatch Mountains indicate that recharge temperature decreases with elevation at about the same rate as the mean annual air temperature, but is on average about 2 °C cooler. Samples from 27 valley production wells yield recharge elevations ranging from the valley elevation (about 1500 m) to mid-mountain elevation (about 2500 m). Only six of the wells have recharge elevations less than 1800 m. Recharge elevations consistently greater than 2000 m in the southeastern part of the basin indicate that subsurface inflow constitutes most of the total recharge in this area.

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1. Introduction

Inter-mountain basin-fill aquifers are a major source of groundwater in the western US. Existing studies suggest that mountain-front recharge typically accounts for one third to nearly all of the recharge to these aquifers (Anderson and Frethey, 1996; Prudic and Herman, 1996; Gates, 1995; Mason, 1998). Mountain-front recharge consists of two components: (1) subsurface inflow from the mountain block (subsurface inflow), and (2) infiltration from perennial and ephemeral streams near the mountain front (stream seepage) (Anderson et al., 1992). Neither component consistently dominates. While stream seepage can be estimated using gauging data or seepage studies, estimates of the magnitude and distribution of subsurface inflow remain very poorly constrained. Estimates based on the disposal of
mountain precipitation (e.g. Chavez et al., 1994) are hampered by large uncertainties in evapotranspiration and snowpack sublimation rates. Estimates based on calibrated steady-state groundwater flow models of the basin-fill aquifers (e.g. Lambert, 1995), are often highly non-unique due to a scarceness of hydrogeologic data near the mountain front and poorly constrained discharge rates. When subsurface inflow estimates are large they seem at odds with the potential for range-bounding normal faults to act as barriers to subsurface inflow (Caine et al., 1996) and, in some cases, the apparently low permeability of rocks in the adjacent mountains (e.g. Parry et al., 2000). The inability to confidently assess the relative magnitude of subsurface inflow presents a significant problem for those charged with source protection planning for valley wells, and for those involved in water rights disputes in rapidly developing mountain resort areas.

Dissolved noble gas concentrations potentially provide an excellent tool for identifying subsurface inflow because of their dependence on the ground temperature at the point of recharge. For typical water table depths (5–100 m), the ground temperature near the water table, and thus the noble gas recharge temperature, is approximately equal to the mean annual air temperature at the surface above (e.g. Mazor, 1991; Stute and Sonntag, 1992). The decrease in mean annual air temperature with elevation (atmospheric lapse) should result in subsurface inflow having a lower noble gas recharge temperature than water recharged in the valley. Oxygen-18 and 2H data have been effectively used to distinguish mountain-front recharge from other sources of recharge (e.g. Thiros, 1995). However, stable isotope ratios are generally incapable of distinguishing subsurface inflow from stream seepage because both sources carry a depleted, high elevation signal. Therefore, δ18O and δ2H values can be used to specifically identify subsurface inflow only in those rare cases when stream seepage is known with considerable confidence to be negligible. The recharge temperature of stream seepage, however, should be that of other water recharged in the valley because water table depths beneath losing streams near the margins of inter-mountain basins are typically greater than 30 m.

In this study, the possibilities and limitations of using noble gases as tracers of subsurface inflow are explored theoretically, then the method is applied to the Salt Lake Valley Aquifer in northern Utah to confirm its utility.

2. Noble gases as recharge elevation tracers

The possibility of using noble gases as recharge elevation tracers either by themselves or in conjunction with 18O and 2H has been discussed by previous workers (Mazor, 1991; Rauber et al., 1991; Zuber et al., 1995; Aeschbach-Hertig et al., 1999; Ballentine and Hall, 1999). However, a rigorous application of the method has not been reported. The general form of the governing equation for the concentration of noble gas \( i \) dissolved in fresh water \( (C_i) \) is:

\[
C_i = C^E_i + C^A_i
\]

where \( C^E_i \) is the component resulting from equilibration with the atmosphere and \( C^A_i \) is the component resulting from excess air. Excess air (dissolved atmospheric gases in excess of equilibrium values) is ubiquitous in groundwater (Heaton and Vogel, 1981) and is commonly attributed to dissolution of air bubbles trapped during water table rises (Stute and Schlosser, 2000). The equilibrium component is related to the dry atmospheric pressure at the recharge point \( (P_r) \) and to the temperature at the water table at the recharge point \( (T_r) \) by:

\[
C^E_i = \frac{X_i P_r}{K_i(T_r)}
\]

where \( X_i \) is the dry mole fraction of gas \( i \) in air and \( K_i \) is the Henry’s Law constant for gas \( i \), a nonlinear function of \( T_r \) of the form:

\[
\frac{1}{K_i} = \exp \left[ a_i \left( \frac{b_i}{T_r} - 1 \right) + 36.855 \left( \frac{b_i}{T_r} - 1 \right)^2 \right]
\]

where \( a_i \) and \( b_i \) are empirical constants (Solomon et al., 1998). The excess air component is given by:

\[
C^A_i = X_iA
\]

where \( A \) is the total concentration of excess air. Aeschbach-Hertig et al. (2000) have recently proposed more complex forms of Eq. (4) that probably model the formation of excess air more accurately.
However, these were not employed in this study mainly because the same authors also demonstrated that the use of different excess air models does not alter the derived \( T_r \) by more than about 1 °C. The relationship between \( P_r \) and the recharge elevation in meters (\( H \)) can be closely approximated with:

\[
P_r = \exp\left(-\frac{H}{8300}\right) - P_{H_2O}
\]

where \( P_{H_2O} \) is the partial pressure of water, provided \( H < 3000 \) m (Plummer and Busenberg, 2000).

If three or more dissolved gas concentrations are measured, then a system of equations in the form of (1) can be in effect solved simultaneously using standard inverse techniques to determine the unknown parameters \( T_r, H, \) and \( A \). Concentrations of \( \text{He}, \text{Ne}, \text{Ar}, \text{Kr}, \) and \( \text{Xe} \) are typically measured, though \( C_{\text{He}} \) is seldom used because the addition of terrigenic \( ^{4}\text{He} \) after recharge is common. Aeschbach-Hertig et al. (1999) and Ballentine and Hall (1999) both employ computer codes that solve this inverse problem by minimizing \( \chi^2 \), defined as:

\[
\chi^2 = \sum_i \frac{(C_i - C_i^{\text{mod}})^2}{\sigma_i^2}
\]

where \( C_i \) are the measured concentrations, \( C_i^{\text{mod}} \) are the modeled concentrations, and \( \sigma_i^2 \) are the experimental 1σ errors. These authors discuss this approach in detail. A similar \( \chi^2 \) minimization code was written for this study to quantify uncertainty in the derived parameters, given (a) the measurement uncertainties particular to our sampling and analytical process, and (b) the use of \( \text{N}_2 \) instead of \( \text{Xe} \) (use of \( \text{N}_2 \) as a conservative gas in this study is warranted because groundwater in the study area is generally well oxygenated, precluding appreciable denitrification). Our inversion code uses the Newton method (Zhdanov, 2000) to perform the minimization because the problem is non-linear, over-determined, and has only three parameters and four measurements (\( C_{\text{Ne}}, C_{\text{Ar}}, C_{\text{Kr}}, C_{\text{N}_2} \)).

The dependence of \( C_i \) on \( H \) and \( T_r \) presents the possibility of using noble gases as recharge elevation tracers either directly through the derivation of \( H \), or indirectly by deriving \( T_r \) and using the local atmospheric lapse rate. Unfortunately, when \( T_r, H, \) and \( A \) are all unknown, the inverse problem is ill-posed; given typical measurement errors of 1% (1σ), uncertainties for derived values of \( H \) and \( T_r \) are unacceptably large (Aeschbach-Hertig et al., 1999; Ballentine and Hall, 1999). The results of Monte Carlo simulations clearly demonstrate this behavior (Fig. 1a). In these simulations, statistically plausible samples were generated for a given set of recharge conditions (specified \( T_r, H, \) and \( A \)). The randomly assigned errors in \( C_i \) were normally distributed with an assigned \( \sigma_i \) determined from the analysis of sample replicates in our laboratory (3% for \( \text{Kr} \), 2% for the remaining gases). Values of \( T_r, H, \) and \( A \) were then derived for each synthetic sample and the \( \sigma \) of the derived distribution of each parameter was considered the uncertainty for that parameter. Several reasonable recharge conditions were simulated. Derived uncertainties in \( H \) and \( T_r \) were typically around \( \pm 1500 \) m and \( \pm 6 \) °C, respectively. Because the potential recharge elevation range in most studies is \(< 2000 \) m, and typical lapse rates are \(-5 \) to \(-8 \) °C per kilometer of elevation gain, an accuracy of about \( \pm 200 \) m in \( H \) or \( \pm 1 \) °C in \( T_r \) would be required to put useful constraints on recharge elevation. To achieve this, a 1σ measurement error of \( \pm 0.3% \) for all gases would be required, which is beyond our present analytical capabilities.

The inverse problem becomes well-posed, however, when \( H \) is known a priori, as is the case when the method is applied to investigate paleoclimate (Aeschbach-Hertig et al., 1999; Ballentine and Hall, 1999). Monte Carlo simulations like those described above were performed for this situation and the uncertainty in \( T_r \) decreased to \( \pm 0.9 \) °C (Fig. 1b), given the measurement uncertainties for our laboratory. This level of accuracy was confirmed by tests performed using water equilibrated with air under controlled conditions such that \( T_r, H, \) and \( A \) were known (Fig. 2). The \( \chi^2 \) surface was mapped for several synthetic samples and all displayed a correctly located, well-defined minimum point with no local minima (Fig. 3). The solution to this inverse problem thus appears to be both unique and stable.

Zuber et al. (1995) and Aeschbach-Hertig et al. (1999) point out that in mountainous terrain where \( H \) is initially unknown, the noble gas data can be used to derive a set of best-fit \( H-T_r \) pairs for a given sample by specifying various values of \( H \) (Fig. 4). Assuming \( T_r \) is equal to the mean annual air temperature (\( T_a \)) at
Fig. 1. Typical results of Monte Carlo simulations performed to determine uncertainty in derived recharge parameters. The two simulations shown each involved 1000 realizations of an alpine water with recharge temperature ($T_r$) = 7 °C, recharge elevation ($H$) = 2000 m, and excess air ($A$) = 0.002 cm$^3$ STP/g. (a) Case where $T_r$, $H$, and $A$ are all unknown. The span of the cluster of solutions over significant ranges of $T_r$, $H$, and $A$ indicates large uncertainties for all three parameters. Uncertainties ($\sigma$) are ±6.3 °C for $T_r$, ±1650 m for $H$, and ±0.00124 cm$^3$ STP/g for $A$. (b) Case where $H$ is known a priori. The modest range of solutions indicates acceptable uncertainties for $T_r$ and $A$ (±0.8 °C and ±0.00025 cm$^3$ STP/g, respectively, for this particular water).
all elevations (implying that the water table is near land surface), the actual $H$ and $T_r$ for the sample should be the point of intersection between the linear set of best-fit solutions and a line representing the local atmospheric lapse on a plot of $H$ versus $T_r$. These authors apply this approach to a small number of samples and obtain mixed results; some derived $H$ values are reasonable while others are difficult to explain.

This study in effect constitutes a considerably more rigorous application of the approach described above. A significant modification, however, is that $T_r$ is not assumed to equal $T_a$ for mountain groundwater flow systems. At high elevation, snowmelt is often the main source of recharge. Large volumes of snowmelt water infiltrating rapidly through fractured rock could conceivably remain near 0 °C in the unsaturated zone, particularly since water table depths may decrease to <3 m during periods of snowmelt (Hill, 1990; Buttle, 1989). Hence, $T_r$ could be less than $T_a$ in many alpine areas. However, mean annual surface ground temperatures extrapolated from borehole temperature profiles are commonly about 3 degrees warmer than $T_a$ (Powell et al., 1988), and ground temperatures several degrees warmer than $T_a$ have been observed in areas of prolonged snow cover (Smith et al., 1964). Furthermore, mountain water tables can be as deep as 500 m in more arid regions of the western US (Fridrich et al., 1994). Water table temperatures in these areas may be well above $T_a$ if recharge rates are not large enough to significantly depress the geothermal gradient. Given these possibilities, the local relationship between $H$ and $T_r$ was derived in this study by determining $T_r$ for mountain groundwaters with constrained values of $H$.

3. Example application: mountain-front recharge to the eastern Salt Lake Valley

3.1. Site description

The Salt Lake Valley is an active rift basin located in northern Utah (Fig. 5). It is semiarid with a mean annual precipitation of 30 to 50 cm, depending on location, and a mean annual air temperature of about 12 °C. The basin is at an elevation of 1300 to 1500 m, and is bound on the east and west by mountains that rise to over 2700 m. This study focuses on mountain-front recharge to the eastern side of the basin from the central Wasatch Mountains (Fig. 5). Mean annual precipitation in the Wasatch Mountains is 50 to 130 cm, depending mainly on elevation, and most precipitation falls as snow.

The basin-fill material consists of mostly-consolidated Tertiary sediments overlain by mostly-unconsolidated Quaternary sediments (Hely et al., 1971). Nearly all wells in the valley are completed in the Quaternary sediments that are 120–300 m thick throughout most of the basin. Production wells are screened within the Principal Aquifer, the deeper Quaternary sediments composed of sand and gravel inter-bedded with lenses of fine-grained deposits. The fine-grained lenses increase in frequency and thickness with distance from the mountains, meaning that
the Principal Aquifer is generally unconfined near the mountains and confined elsewhere in the basin.

3.2. Existing knowledge of mountain-front recharge

Groundwater flow models of the basin indicate that subsurface inflow is significant, constituting 45% of the total recharge, while stream seepage only accounts for 5% of the total recharge (Waddell et al., 1987; Lambert, 1995). Calculations based on the disposal of mountain precipitation corroborate this large subsurface inflow estimate (Hely et al., 1971). However, a careful error analysis performed for this study indicates that the estimate has an uncertainty of ±50 to 100%. Further, the apparently low permeability of rocks in large areas of the central Wasatch range seem to preclude such large subsurface inflow rates (Parry et al., 2000). Uncertainty in the stream seepage estimate is also significant because it is based on stream gauge data that is complicated by numerous unmeasured inputs and diversions (Hely et al., 1971). The remainder of the recharge budget

Fig. 3. Misfit surface ($\log_{10} \chi^2$) for an alpine water with recharge temperature ($T_r$) = 7°C, recharge elevation ($H$) = 2000 m, and excess air ($A$) = 0.002 cm$^3$ STP/g. The misfit surface was mapped at increments of 0.2°C and 0.0001 cm$^3$ STP/g excess air. The stability and uniqueness of the derived solution is demonstrated by the fact that the surface is smooth with a single, well-defined minimum point properly located at the actual parameter values.

Fig. 4. Method of determining recharge elevation ($H$) from noble gas concentrations proposed by Zuber et al. (1995) and Aeschbach-Hertig et al. (1999). A family of $H-T_r$ solutions can be derived from the noble gas data by specifying different values of $H$. A unique recharge elevation is obtained at the intersection between the curve representing this family of $H-T_r$ solutions and a line representing the local atmospheric lapse (mean annual air temperature vs. elevation).
mainly consists of seepage from precipitation (20%), unused irrigation water (20%), and canal water (10%) in the valley (Hely et al., 1971; Waddel et al., 1987; Lambert, 1995). This seepage in combination with stream seepage will henceforth be referred to as ‘valley recharge’. Valley recharge is believed to occur largely near the mountain front where the principal aquifer is unconfined.

3.3. Constraining recharge temperatures of source waters

Noble gas samples were collected from 16 springs and mine tunnels located at various elevations on the western slope of the Wasatch Mountains. Samples were collected using passive diffusion samplers similar to those shown in Sanford et al. (1996). Samplers were placed directly within the spring orifice to insure that the sampled water had not re-equilibrated with the atmosphere. The samplers were allowed to equilibrate for at least 24 h. Nitrogen, Ne, Ar, and Kr were measured using a quadrupole mass spectrometer. Samples were first inlet directly into a high vacuum purification system that includes a Ti/Zr sponge for reactive gases. Non-reactive gases were cryogenically separated. Nitrogen and O₂ were measured dynamically using a leak valve, while the remaining gases were measured statically.
From the measured concentrations of Ne, Ar, Kr, and N₂, a maximum and a minimum $T_r$ ($T_{r\text{max}}$ and $T_{r\text{min}}$) were derived for each sample using minimum and maximum $H$ values ($H_{\text{min}}$ and $H_{\text{max}}$), respectively (Fig. 6). The discharge elevation was used for $H_{\text{min}}$ and the highest local surface water drainage divide was used for $H_{\text{max}}$. If $H_{\text{min}}$ or $H_{\text{max}}$ yielded an unacceptably high $\chi^2$ value ($<5\%$ probability of producing measured concentrations), the parameter was adjusted until an acceptable fit was achieved, further constraining the recharge elevation. The local atmospheric lapse of $-6.7 \, ^\circ\text{C/km}$ shown in Fig. 6 is that reported by Hely et al. (1971) for the Salt Lake City area. This is close to the regional lapse for central Utah ($-6.9 \, ^\circ\text{C/km}$) reported by Powell et al. (1988). Discharge temperatures lie near the atmospheric lapse, most being slightly warmer (0 to 1 $^\circ\text{C}$) than $T_a$. Nearly all derived values of $T_r$, however, are 0 to 4 $^\circ\text{C}$ less than $T_a$. The least squares regression line through the derived $T_r$ values (including all $T_{r\text{max}}$ and $T_{r\text{min}}$ values) has a slope of $-7.3 \, ^\circ\text{C/km}$, similar to the atmospheric lapse, and indicates that $T_r$ is on average about 2 $^\circ\text{C}$ less than $T_a$. Clearly, assuming that $T_r = T_a$ in the central Wasatch Mountains is inappropriate. Fig. 6 also shows the surface ground temperature at an elevation of approximately 1850 m deduced from $>50$ bore hole temperature profiles measured at the Jordanelle Dam site (Moran, 1991). Both the mean and the scatter of the surface ground temperatures is in good agreement with derived $T_r$ values.

The possible $H-T_r$ combinations for springs and tunnels shown in Fig. 6 define a zone of possible $H-T_r$ combinations for subsurface inflow (Fig. 7). The zone of possible $H-T_r$ combinations for valley
recharge (Fig. 7) was determined from temperature measurements near the water table in valley wells. Measurements from this study are from wells with water table depths of 10 to 30 m located in recharge areas. Measurements from Thiros (1995) are from shallow wells (<30 m) with unspecified water table depths located throughout the eastern side of the valley. The measured water table temperatures generally range from 12 to 15 °C, as expected given that shallow water table temperatures typically exceed the local $T_a$ by 1 to 2 °C (Domenico and Schwartz, 1990). Furthermore, a $T_r$ of 13.9 °C was obtained from a water table well in the valley where an $^{18}$O measurement verified local recharge of evaporated irrigation canal water. A line representing derived $H-T_r$ combinations for a hypothetical sample containing 100% valley recharge (‘sample $H-T_r$ line’) is also shown on Fig. 7. The fact that this line intersects the zone of possible $H-T_r$ combinations for subsurface inflow, but misses the zone of possible $H-T_r$ combinations for valley recharge, indicates that this sample could be identified as valley recharge water by its noble gas concentrations. Because only a small fraction of sample $H-T_r$ lines would intersect both zones of possible $H-T_r$ combinations, the method appears capable of distinguishing subsurface inflow from valley recharge.

However, given the long screens in the production wells sampled for this study, samples containing mixtures of subsurface inflow and valley recharge are possible, if not probable. Two-component mixing between waters with various recharge parameters was simulated using synthetic samples. In all cases, the $T_r$ derived for the mixture was within 0.5 °C of the weighted average of the end-member $T_r$ values. Therefore, mixing is approximately linear; i.e. mixed samples lie near a line connecting the
end-member waters in $H-T_t$ space. This means that the corners of the two zones of possible $H-T_t$ combinations shown in Fig. 7 can be connected to form a single zone of possible $H-T_t$ combinations for the entire system (Fig. 8). This ‘solution zone’ contains all possible end-member waters and mixtures thereof. Similarly, the ends of the dashed best-fit lines in Fig. 7 can be connected to form a zone of probable $H-T_t$ combinations for the entire system. This ‘probable solution zone’ contains the most probable end-member waters and their mixtures.

When a sample $H-T_t$ line is plotted in $H-T_t$ space, the sample’s $H_{\min}$, $H_{\max}$, $T_{\min}$, and $T_{\max}$ are all defined by its intersection with the solution zone (Fig. 8). Note that if the well elevation is greater than the elevation of the point where the sample $H-T_t$ line intersects the bottom of the solution zone, the former is chosen as the sample’s $H_{\min}$ (and $T_{\max}$ is chosen accordingly). Also note that if $T_{\max}$ is less than 11 °C, the sampled water must contain at least some subsurface inflow. The probable recharge elevation ($H_{\text{prob}}$) and the probable recharge temperature ($T_{\text{rprob}}$) are defined by the point of intersection between the sample $H-T_t$ line and the line bisecting the probable solution zone (Fig. 8). Finally, the minimum fraction of subsurface inflow in a sample is defined by the point where the sample $H-T_t$ line intersects the mixing line between a typical valley recharge water and subsurface inflow water with $T_t = 0$ °C.

### 3.4 Valley well results

Noble gas samples were collected from 27 valley production wells screened in the Principal Aquifer (Fig. 9). Total well depths are typically 100–200 m, and screen lengths are generally 50–100 m. Flow rates of 500–2000 gallons per minute (32–126 l/s) are typical. Well S25 (southernmost...
well on Fig. 9) is the only exception; it is a smaller artesian well screened in the mountain block and is not pumped. Samples were collected while the wells were operating, and most wells had been pumping for at least 48 h prior to sampling. As with the springs, passive diffusion samplers were used. The samplers were placed within a flow-through cell plumbed directly into the well head. The well pump provided significant pressure in the flow-through cell, preventing gas loss. Flow rates through the cell were maintained at about 4 l/min throughout the 24-h (minimum) equilibration period. Analytical procedures were the same as described above.

For each sample, $T_{\text{max}}$ and $H_{\text{prob}}$ were derived using measured concentrations of Ne, Ar, Kr, and N₂ (Figs. 10 and 11). Derived $T_{\text{max}}$ values range from 5.9 to 11.4 °C. All but two $T_{\text{max}}$ values are less than 11 °C, the minimum possible recharge temperature for valley recharge. Therefore, water throughout most of the study area is composed of at least some subsurface inflow. A prominent cell of low-$T_r$ water is located in

Fig. 9. Location of sampled wells. Potentiometric surface contours for the Principal Aquifer are also shown to illustrate the general direction of groundwater flow.

Fig. 10. Distribution of derived maximum possible recharge temperature ($T_{\text{max}}$) in the Principal Aquifer. The probable recharge temperature is 3 to 4 °C colder for wells with $T_{\text{max}} < 7$ °C.
the southern part of the study area. The $T_{\text{max}}$ values of 6 to 9 °C within this cell correspond to minimum subsurface inflow fractions of 0.85 to 0.47, respectively. Note that well S25, known to contain 100% subsurface inflow, has a $T_{\text{max}}$ of 5.9 °C, similar to $T_{\text{max}}$ values in the central part of the cell. Derived $H_{\text{prob}}$ values range from 1460 to 2580 m. Within the low-$T_r$ cell, $H_{\text{prob}}$ values are 2000 to 2500 m, well above the valley elevation. The fact that the core of the low-$T_r$ cell is located out away from the range front suggests that much of the subsurface inflow does not simply enter the Principal Aquifer laterally at the range front, as generally envisioned by previous workers. Instead, a large fraction apparently enters the aquifer well out into the valley. Derived $T_{\text{max}}$ values approach 11 °C in some areas outside of the low-$T_r$ cell. Water in these areas may be composed dominantly of valley recharge.

4. Conclusions

Noble gas concentrations provide a useful tool for investigating mountain-front recharge. While $\delta^{18}O$ data generally cannot be used to distinguish between subsurface inflow and stream seepage, recharge temperatures derived from noble gas data put meaningful constraints on the relative significance of subsurface inflow, specifically. The general method of determining recharge elevation from noble gas concentrations proposed by Zuber et al. (1995) and Aeschbach-Hertig et al. (1999) is feasible, with one significant modification: the local relationship between $T_r$ and $H$ must be initially determined because the assumption that $T_r = T_a$ in mountain groundwater flow systems is dubious. Application of the method to the Salt Lake Valley Principal Aquifer yielded reasonable results. The noble gas data clearly indicate that subsurface inflow is significant in the eastern part of the basin, and reveal its general distribution. A prominent cell of low-$T_r$ water is present in the southeastern part of the valley. Water in this area is composed dominantly, if not entirely, of subsurface inflow. The results of this study lend considerable weight to the large subsurface inflow rates estimated for the Salt Lake Valley in previous studies.

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