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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 Freethey, 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

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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 ^{2}H 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, δ^{18}O and δ^2H 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 ^{18}O and ^{2}H 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:

$$C_i = C_i^E + C_i^A \tag{1}$$

where C_i^E is the component resulting from equilibration with the atmosphere and C_i^A 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 (recharge pressure, P_r) and to the temperature at the water table at the recharge point (recharge temperature, T_r) by:

$$C_i^E = \frac{X_i P_r}{K_i(T_r)} \tag{2}$$

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:

$$\frac{1}{K_i} = \exp\left[a_i\left(\frac{b_i}{T_r} - 1\right) + 36.855\left(\frac{b_i}{T_r} - 1\right)^2 \right] \tag{3}$$

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

$$C_i^A = X_i A \tag{4}$$

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:

$$P_r = \exp\left(-\frac{H}{8300}\right) - P_{H_2O} \tag{5}$$

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 He, Ne, Ar, Kr, and Xe are typically measured, though C_{He} is seldom used because the addition of terrigenic ^{4}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 χ^2, defined as:

$$\chi^2 = \sum_i \frac{(C_i - C_i^{mod})^2}{\sigma_i^2} \tag{6}$$

where C_i are the measured concentrations, C_i^{mod} are the modeled concentrations, and σ_i^2 are the experimental 1σ errors. These authors discuss this approach in detail. A similar χ^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 N_2 instead of Xe (use of 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_{Ne}, C_{Ar}, C_{Kr}, C_{N2}).

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 σ_i determined from the analysis of sample replicates in our laboratory (3% for Kr, 2% for the remaining gases). Values of T_r, H, and A were then derived for each synthetic sample and the σ 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 ± 1500 m and ± 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 ± 200 m in H or ± 1 °C in T_r would be required to put useful constraints on recharge elevation. To achieve this, a 1σ measurement error of $\leq 0.3\%$ 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 ± 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 χ^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
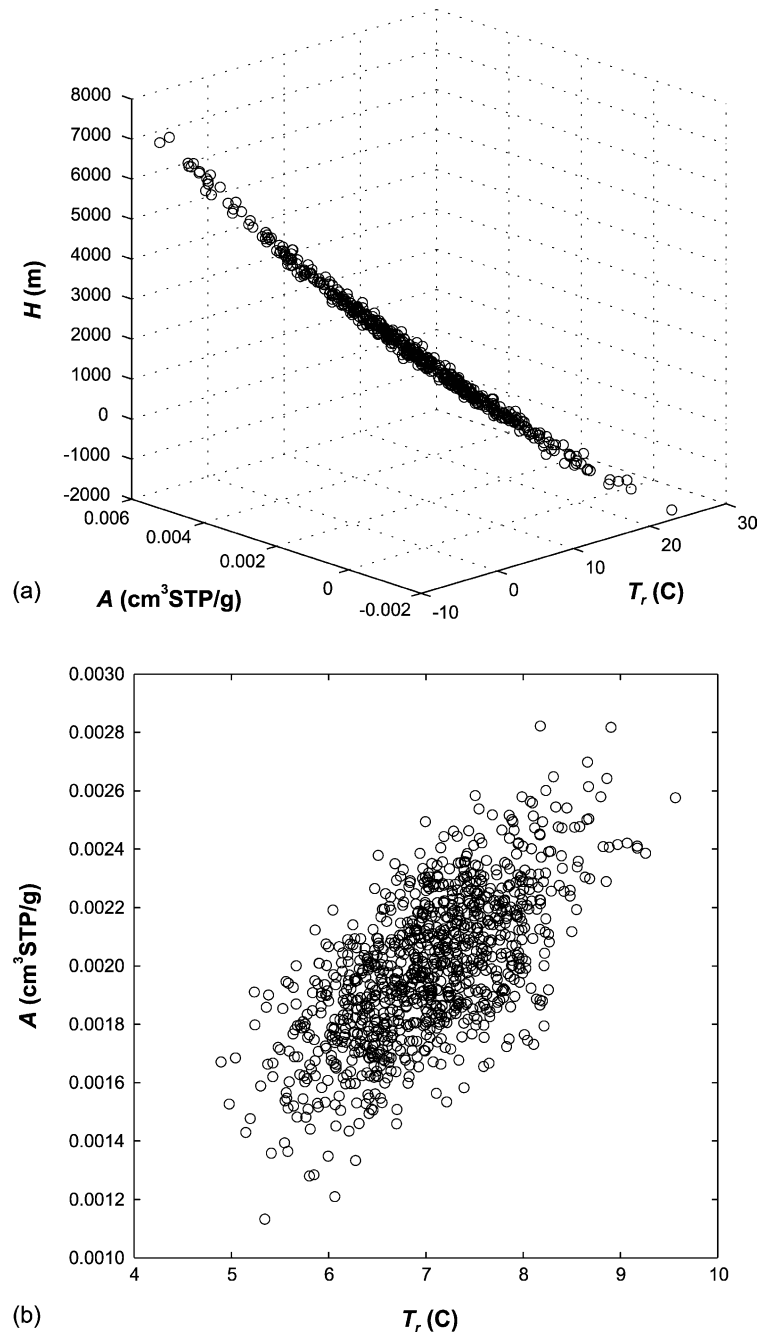
Fig. 1. Typical results of Monte Carlo simulations performed to determine uncertainty in derived recharge parameters. The two simulations shown each involved 1000 realizations of an alpine water with recharge temperature (T_r) = 7 °C, recharge elevation (H) = 2000 m, and excess air (A) = 0.002 cm³ STP/g. (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 (1σ) are ± 6.3 °C for T_r, ± 1650 m for H, and ± 0.00124 cm³ 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 cm³ STP/g, respectively, for this particular water).
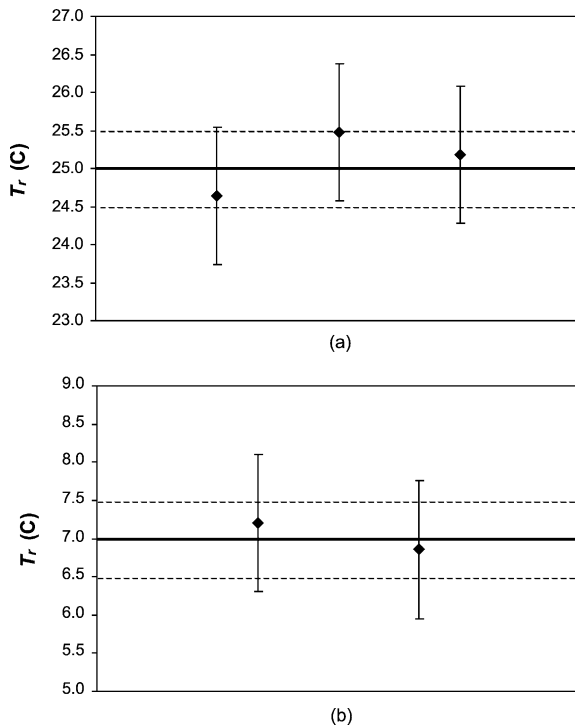
Fig. 2. Results of accuracy tests performed on water equilibrated with air at (a) 25 °C and (b) 7 °C. Each point is a separate sample collected from the same batch of air-equilibrated water. Solid lines show the known T_r, which has an uncertainty of ± 0.5 °C (shown by dashed lines) due to variations in room temperature during equilibration. Error bars indicate uncertainty of ± 0.9 °C in derived T_r based on Monte Carlo simulations.

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 Wastach 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
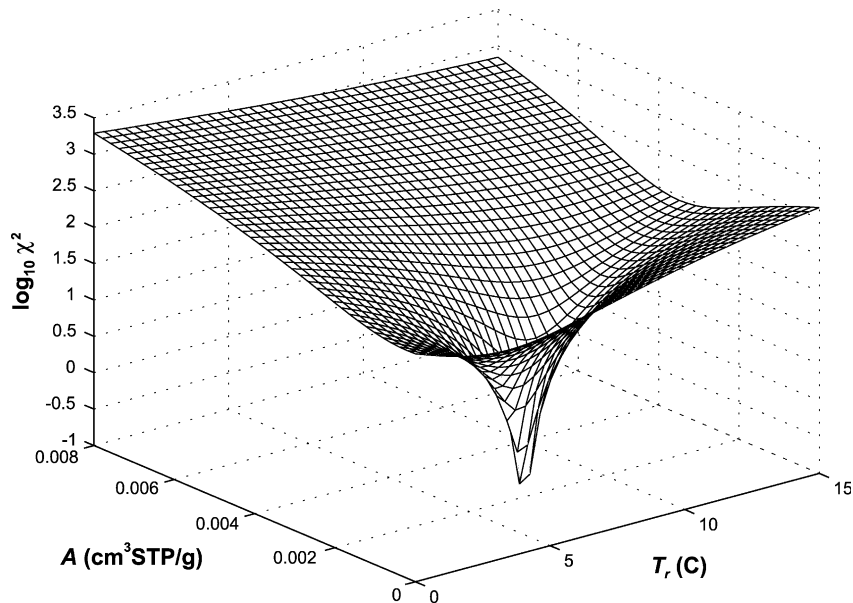
Fig. 3. Misfit surface ($\log_{10} \chi^2$) for an alpine water with recharge temperature (T_r) = 7 °C, recharge elevation (H) = 2000 m, and excess air (A) = 0.002 cm³ STP/g. The misfit surface was mapped at increments of 0.2 °C and 0.0001 cm³ STP/g excess air. The stability and uniqueness of the derived 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.

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
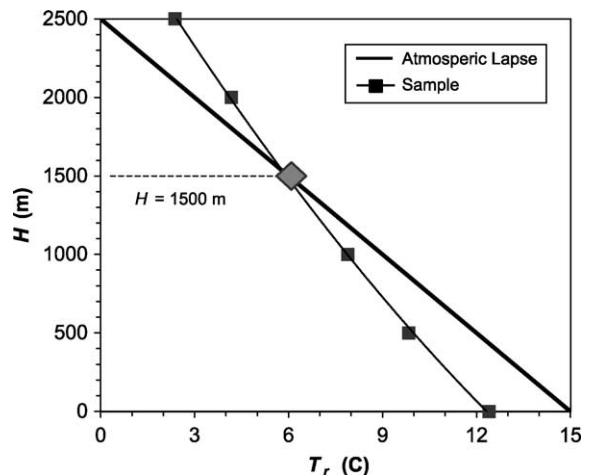

Fig. 4. Method of determining recharge elevation (H) from noble gas concentrations proposed by Zuber et al. (1995) and Aeschbach-Hertig et al. (1999). A family of $H-T_r$ solutions can be derived from the noble gas data by specifying different values of H. A unique recharge elevation is obtained at the 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).
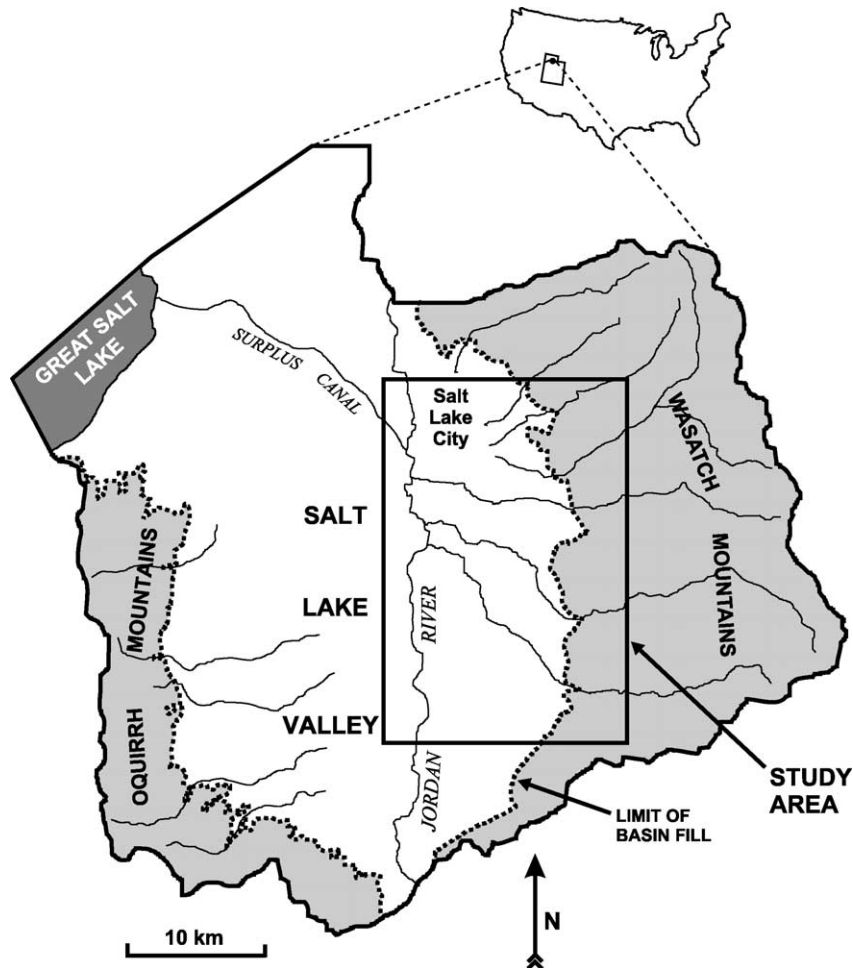
Fig. 5. Location of study area. The Jordan River flows northward toward the Great Salt Lake and is the primary discharge zone for the Salt Lake Valley Principal Aquifer.

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 quadrapole 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 O_2 were measured dynamically using a leak valve, while the remaining gases were measured statically.
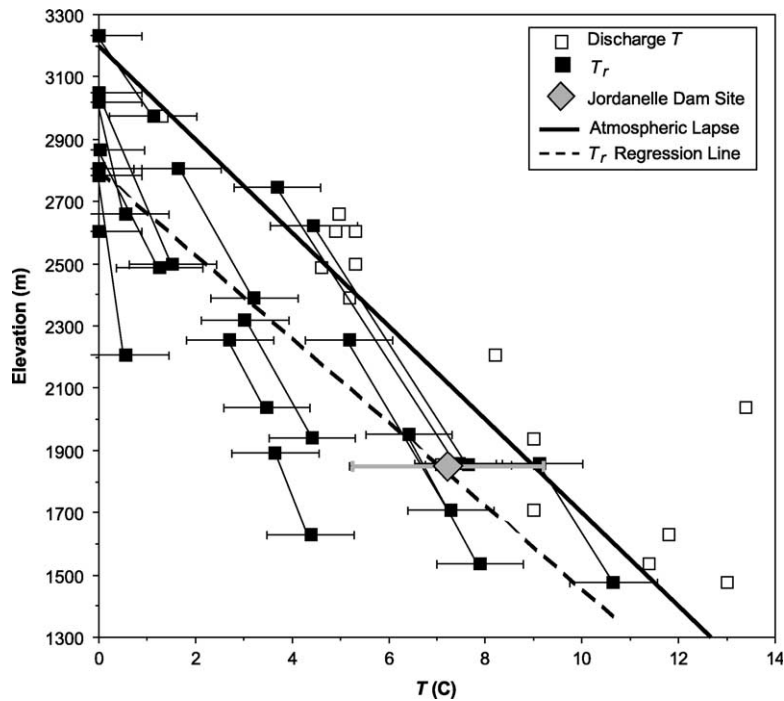
Fig. 6. Measured discharge temperature and derived recharge temperature (T_r) vs. elevation for springs and mine tunnel discharge waters in the central Wastach Mountains. Discharge temperature is plotted against discharge elevation. Each pair of T_r values connected with a tie line corresponds to a single sample; upper point $= (T_{rmin}, H_{max})$ and lower point $= (T_{rmax}, H_{min})$, where H is recharge elevation. The linear regression line for all derived T_r values (including all T_{rmax} and T_{rmin} values) and the local atmospheric lapse from Hely et al. (1971) are also shown. Jordanelle Dam site data from Moran (1991) consists of >50 mean annual surface ground temperatures determined from borehole temperature profiles. The mean value is shown, with error bars indicating the approximate range of scatter.

From the measured concentrations of Ne, Ar, Kr, and N_2, a maximum and a minimum T_r (T_{rmax} and T_{rmin}) were derived for each sample using minimum and maximum H values (H_{min} and H_{max}), respectively (Fig. 6). The discharge elevation was used for H_{min} and the highest local surface water drainage divide was used for H_{max}. If H_{min} or H_{max} yielded an unacceptably high χ^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 $-6.7\,°C/km$ 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 ($-6.9\,°C/km$) reported by Powell et al. (1988). Discharge temperatures lie near the atmospheric lapse, most being slightly warmer (0 to $1\,°C$) than

T_a. Nearly all derived values of T_r, however, are 0 to $4\,°C$ less than T_a. The least squares regression line through the derived T_r values (including all T_{rmax} and T_{rmin} values) has a slope of $-7.3\,°C/km$, similar to the atmospheric lapse, and indicates that T_r is on average about $2\,°C$ 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
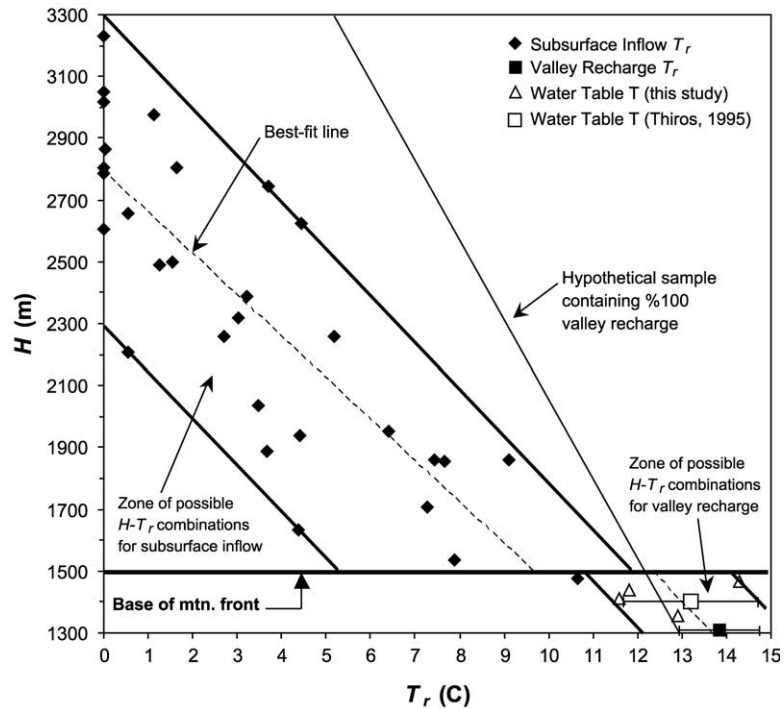
Fig. 7. Zones of possible recharge elevation-temperature $(H-T_r)$ combinations for subsurface inflow and valley recharge. The point representing data from Thiros (1995) is the mean for 19 water table temperature measurements, the error bars indicating the approximate data scatter. The line corresponding to a hypothetical sample containing 100% valley recharge does not intersect the zone of possible $H-T_r$ combinations for subsurface inflow. This demonstrates that the method is capable of distinguishing valley recharge from subsurface inflow.

recharge (Fig. 7) was determined from by 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 valley recharge, but misses the zone of possible $H-T_r$ combinations for subsurface inflow, 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
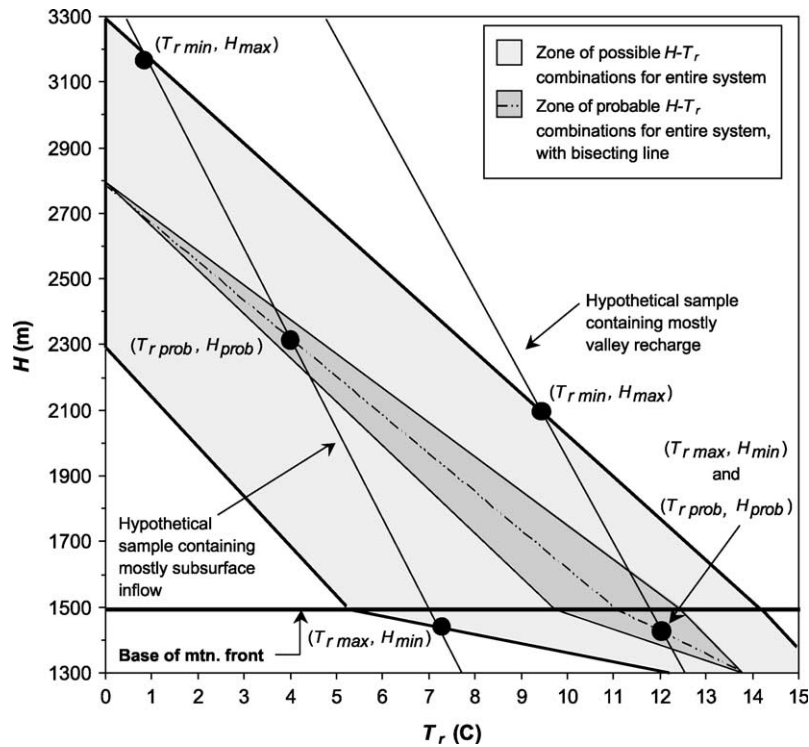
Fig. 8. Zones of possible and probable recharge elevation–temperature (H–T_r) combinations for the entire system (includes mixed waters). H_{min}, T_{rmax}, etc. are defined by points of intersection between the sample H–T_r line and different parts of these zones (see text for detailed explanation).

end-member waters in H–T_r space. This means that the corners of the two zones of possible H–T_r combinations shown in Fig. 7 can be connected to form a single zone of possible H–T_r 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_r combinations for the entire system. This 'probable solution zone' contains the most probable end-member waters and their mixtures. When a sample H–T_r line is plotted in H–T_r space, the sample's H_{min}, H_{max}, T_{rmin}, and T_{rmax} 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_r line intersects the bottom of the solution zone, the former is chosen as the sample's H_{min} (and T_{rmax} is chosen accordingly). Also note that if T_{rmax} is less than 11 °C, the sampled

water must contain at least some subsurface inflow. The probable recharge elevation (H_{prob}) and the probable recharge temperature (T_{rprob}) are defined by the point of intersection between the sample H–T_r 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_r line intersects the mixing line between a typical valley recharge water and subsurface inflow water with $T_r = 0$ °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
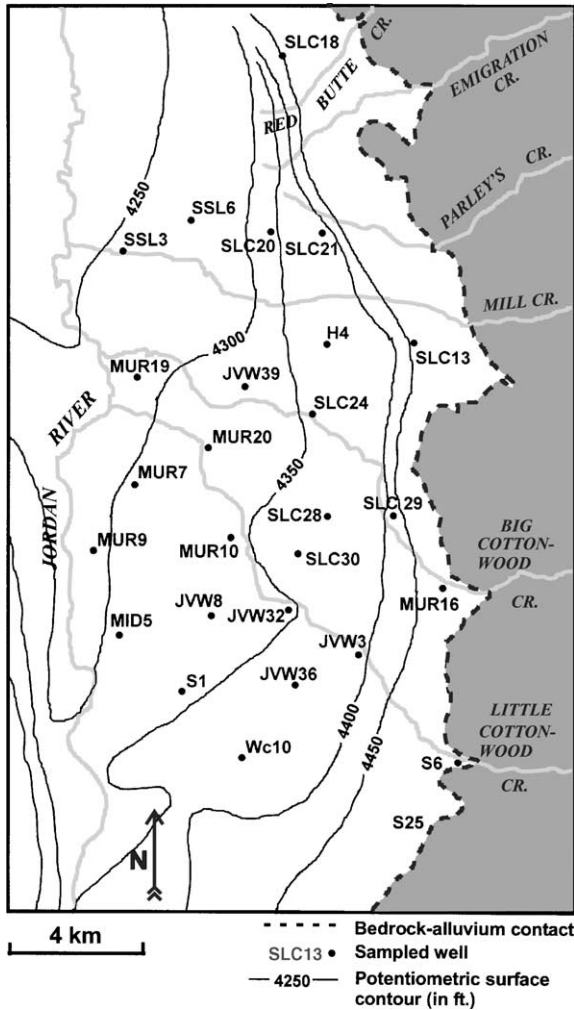
Fig. 9. Location of sampled wells. Potentiometric surface contours for the Principal Aquifer are also shown to illustrate the general direction of groundwater flow.

Fig. 10. Distribution of derived maximum possible recharge temperature (T_{rmax}) in the Principal Aquifer. The probable recharge temperature is 3 to 4 °C colder for wells with $T_{rmax} < 7$ °C.

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-though cell, preventing gas loss. Flow rates through the cell were maintained at about 4 l/min
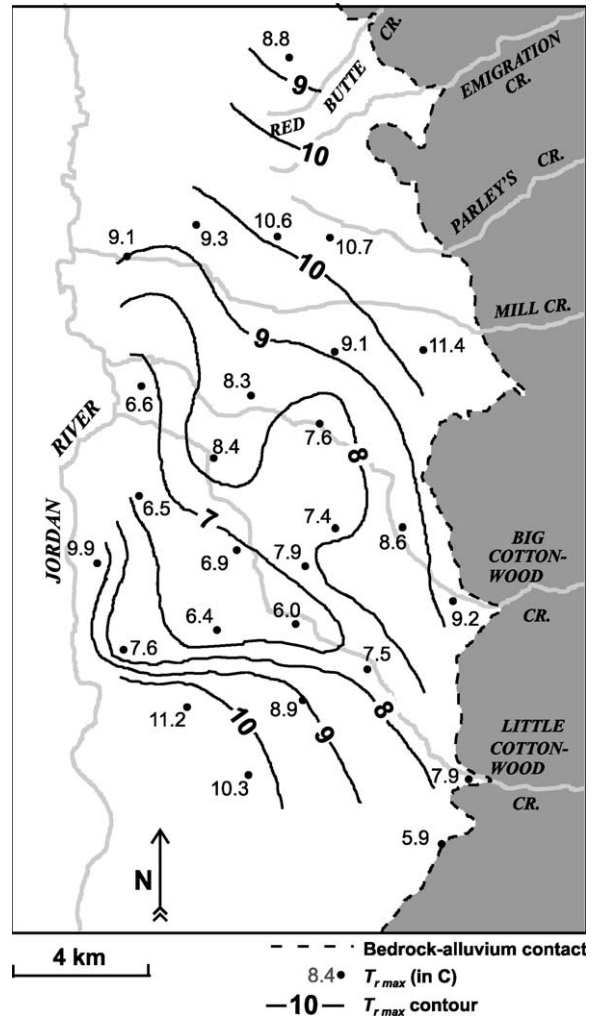
throughout the 24-h (minimum) equilibration period. Analytical procedures were the same as described above.

For each sample, T_{rmax} and H_{prob} were derived using measured concentrations of Ne, Ar, Kr, and N_2 (Figs. 10 and 11). Derived T_{rmax} values range from 5.9 to 11.4 °C. All but two T_{rmax} 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
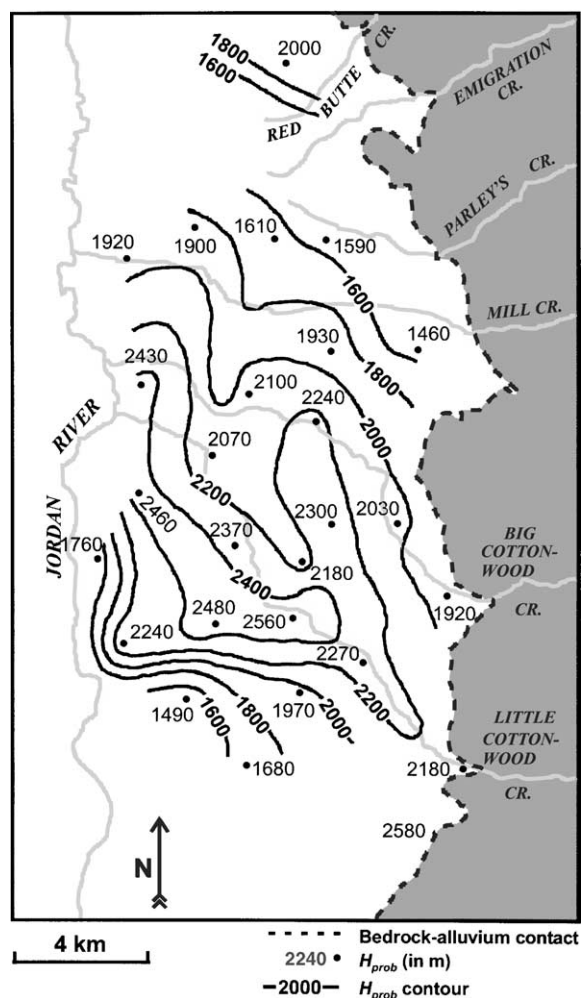
Fig. 11. Distribution of derived probable recharge elevation (H_{prob}) in the Principal Aquifer. Values greater than 2000 m represent greater than 50% subsurface inflow.

the southern part of the study area. The T_{rmax} 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_{rmax} of 5.9 °C, similar to T_{rmax} values in the central part of the cell. Derived H_{prob} values range from 1460 to 2580 m. Within the low-T_r cell, H_{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_{rmax} 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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