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Seasonal recharge components in an urban/agricultural mountain front aquifer system using noble gas thermometry

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SUMMARY

Thirteen noble gas samples were collected from eleven wells and two mountain springs in the Treasure Valley, Idaho, USA to derive recharge temperatures using noble gas thermometry. One common assumption with noble gas thermometry is that recharge temperatures are roughly equal to the mean annual surface temperature. When water table depths are shallow or variable, or infiltration is seasonal recharge temperatures may be significantly different from the mean annual surface temperature. Water table depths throughout the study area were used to estimate recharge source temperatures using an infiltration-weighted recharge temperature model which takes into account a time-variable water table. This model was applied to six different seasonally-dependent recharge scenarios. The modeled recharge temperatures for all scenarios showed a strong dependence of recharge temperature on mean annual depth to water. Temperature results from the different recharge scenarios ranged from near the mean annual surface temperature to as much as 6 °C warmer. This compared well to noble gas derived recharge temperatures from the valley wells which ranged from 5 °C below to 7.4 °C above the mean annual surface temperature of the valley. Cooler temperatures suggest an influence of recharge through the adjacent mountain block while warmer temperatures suggest an influence from summer irrigation.

1. Introduction

Intermountain basin aquifers are difficult environments in which to quantify groundwater fluxes due to potentially high variability in landscape and hydraulic properties. Basins may contain numerous and complex river systems, and are often bounded by large mountain ranges that have their own complex hydrology (Wilson and Guan, 2004). Multiple recharge sources including river seepage, precipitation, and recharge through an adjacent mountain block (referred to as mountain block recharge or underflow) are often present with different flow paths, recharge rates, and water quality. For water management purposes it is important to quantify recharge source contributions to basin aquifers (Bales et al., 2006). The seasonality of certain recharge sources (e.g. spring flooding, monsoon-type rains, winter snowpack, irrigation) can further complicate this problem.

One method for determining recharge in intermountain basin environments is noble gas thermometry, which makes use of the dependence of gas solubility on the temperature and pressure of the recharge environment (Mazor, 1972). The concentration of noble gases in recharging water depends on, among other things, the temperature of water when last exposed to the atmosphere (Mazor, 1972; Weiss, 1970, 1971). In mountain environments, recharge temperature is strongly correlated with elevation. Cooler recharge temperatures are often associated with higher recharge elevations, enabling identification of recharge source location based on noble gas concentrations (Althaus et al., 2009; Manning and Solomon, 2003; Thomas et al., 2003). Using noble gas thermometry to distinguish sources of recharge requires comparing the original temperature of basin groundwater at the time of recharge, which is estimated by noble-gas temperature (T_{NG}) to the temperature of potential recharge sources (recharge source temperature: T_r). Recharge source temperature is related to the soil temperature at the water table, which is often estimated from the mean annual temperature (m.a.t.) at the soil surface (Stute et al., 1995; Stute and Schlosser, 1993). Mean annual temperatures can be determined from local weather stations or, in areas with significant topographical relief, estimated using a local temperature lapse rate.

Provided that water table depths are greater than a few meters below land surface (BLS), recharge temperatures can be estimated from the m.a.t. at the recharge source (e.g. mountain block or valley floor) (Mazor, 1972). In situations where water table depths are less than a few meters, infiltration is seasonal, or water table depths fluctuate annually, the m.a.t. at the soil surface may not

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represent the temperature at the water table (Castro et al., 2007; Cey et al., 2009; Manning and Solomon, 2003; Stute and Schlosser, 1993) and a different method must be used to determine the recharge source temperature.

Seasonal variations in infiltration can be produced by natural climatic conditions or by manipulation of the water cycle. In agriculture-dominated basins summer irrigation imposes artificial seasonality of recharge. For example, within the Treasure Valley in southwest Idaho (Fig. 1) aquifer recharge is dominated by canal seepage and irrigation, which is active only from March through October, with only minor contributions from mountain block recharge and precipitation (Urban, 2004). The seasonal recharge imposed by summer irrigation will likely produce valley recharge temperatures that are warmer than the m.a.t. of the valley floor provided water table depths are shallower than a few meters. In order to identify proportions of mountain block recharge present in a valley aquifer using noble gas thermometry it is important to constrain the upper (warmest) and lower (coldest) limits of T_r. The lower limit recharge temperature ($T_{r,min}$) can be estimated from temperatures at the highest elevations within the watershed where m.a.t. is coldest (Manning and Solomon, 2003). The upper limit recharge temperature ($T_{r,max}$) will be controlled by temperatures at the lower elevations of the valley floor. If recharge at the valley floor is seasonal, it may significantly increase or decrease the upper limit of potential recharge temperatures (depending on the seasonality of the source and the depth of the water table). Improper estimates of recharge source temperatures will lead to improper estimates of contributions from recharge sources.

In the Treasure Valley, irrigation and canal seepage occurs primarily during summer months when water table depths are driven shallow by infiltration (Urban, 2004). This will likely lead to T_r values that are warmer than the m.a.t. of the valley floor. In this paper we show the applicability of noble gas thermometry to determine recharge sources within the Treasure Valley aquifer, Idaho and extend noble gas thermometry methods to include seasonal recharge sources by applying an infiltration-weighted method for determining recharge source temperatures for seasonal recharge (Stute and Schlosser, 1993). Results suggest that summer irrigation, which produces recharge temperatures above the m.a.t., is the dominant recharge source in the valley along with mountain block recharge. The results are consistent with the existing conceptual hydrogeological model of the valley aquifer system and the spatial distribution of land use.

2. Study area

The focus area of this project was the northeastern portion of the Treasure Valley, Idaho (Fig. 1). The Treasure Valley is a 3600 km^2 graben-filled, northwest-trending valley that is part of the larger Western Snake River Plain. It is bounded to the northeast by the mountains of the Boise Front Range and to the south and east by the Snake River. The boundary with the Boise Front Range consists of numerous high-angle normal faults with local offsets of a few meters (Wood and Clemens, 2002). The portion of the Treasure Valley under consideration in this study is within the Lower Boise River watershed, which encompasses the Boise River from its exit at Lucky Peak Dam at the east end of the Treasure Valley to about halfway to its confluence with the Snake River 100 km to the west. The elevation of the Lower Boise River watershed ranges from ~2100 m above mean sea level (AMSL) along the ridge of the Boise Front Range to ~680 m AMSL at the confluence of the Boise and Snake rivers.

The Treasure Valley is classified as a semi-arid climate. It receives on average 28 cm of precipitation annually, mostly as rain (Fig. 2). Temperatures in the valley range from around 30 °C in the summer to near −5 °C in the winter with an m.a.t. over the last 10 years of approximately 11 °C. In the high elevations of the adjacent Boise Front Range the annual precipitation is over 80 cm, most of which falls as snow (Aishlin and McNamara, 2011). The m.a.t. decreases from 11 °C at the valley floor (Boise Airport (Weather Underground, 2010), 825 m AMSL) to approximately 6 °C near the Boise Front ridgeline (Bogus Basin SNOTEL site (United States Department of Agriculture, 2010), 1930 m AMSL). Temperature data taken from three weather stations at different elevations

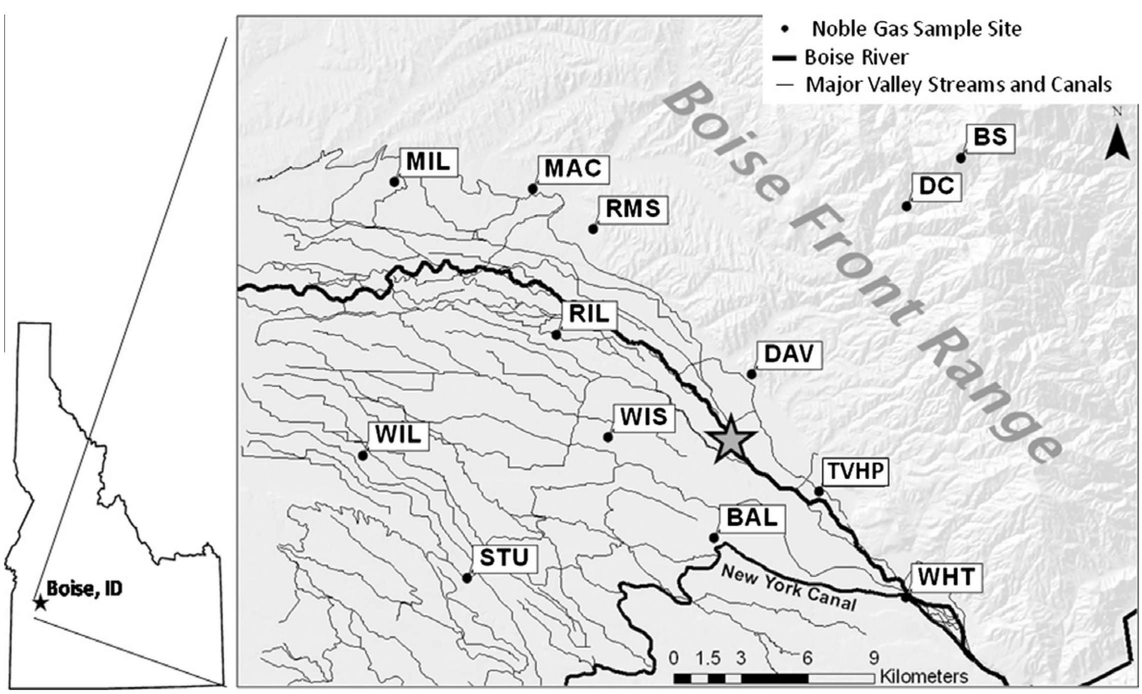

Fig. 1. Location of the study area showing sample locations and major surface water features.

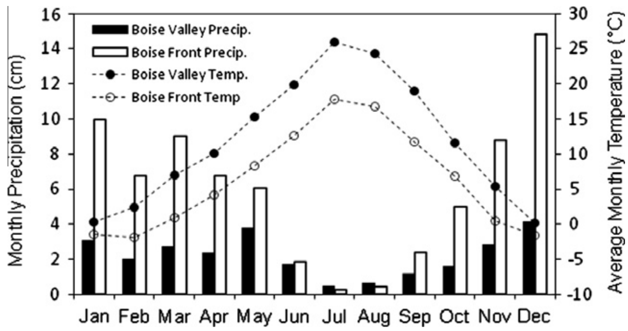

Fig. 2. Average monthly temperature and average monthly precipitation for the Boise municipal area (years 1998–2009) and Bogus Basin SNOTEL site located at the ridge of the Boise Front Range (years 2000–2006).

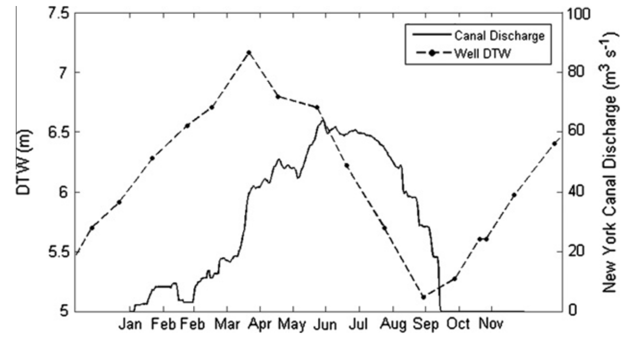

Fig. 3. Example of DTW values from a Treasure Valley well over 1 year showing the correlation between minimum DTW and canal activity represented by New York Canal discharge.

within the Dry Creek Experimental Watershed in Boise Front Range produce a best-fit calculated m.a.t. lapse rate of 0.005 °C m^{-1}.

The Treasure Valley aquifer system is composed of sediments deposited in association with Lake Idaho approximately 4–6 million years ago (ma). These sediments range in thickness from just a few meters near the basin margins to as much as 1800 m near the valley center (Wood and Clemens, 2002). The stratigraphy of the Treasure Valley aquifer is classified into three major depositional groups. From deepest to shallowest they are: Miocene volcanics of the Idavada Group, Miocene lacustrine deposits of the Idaho Group (Lake Idaho sediments), and Pleistocene unconsolidated fluvial deposits of the Snake River Group. Lacustrine sediments of the Idaho Group represent a complex series of transgression/regression lake activity and associated shoreline, and fluvial deposits. The stratigraphy consist of interbedded mudstone and sandstone facies, lacustrine mudstones, deltaic sands, and shoreline oolitic deposits (Petrich and Urban, 2004; Wood and Clemens, 2002). High transmissivity sand lenses make up the primary production zones for most high yield wells of the Treasure Valley though many private wells are completed within the shallow terrace deposits of the Snake River Group.

The Boise Front Range is composed of granodiorite of the Idaho Batholith intruded ~70 ma. Below an elevation of approximately 1100 m AMSL the crystalline bedrock is draped with lake and shoreline deposits that can reach 10's of meters thick. Above approximately 1100 m AMSL bedrock depths are very shallow and surface exposures are prevalent. The bedrock is highly fractured with water table depths commonly between 10 and 75 m BLS (Hoffman, 2008). It has been estimated that as much as 40% of annual precipitation is recharged to the groundwater in the mountain aquifer in some Boise Front Range sub-basins (Aishlin and McNamara, 2011). The fate of this recharge water can be springs and baseflow in mountain streams or it may migrate through subsurface flow paths to the valley aquifer.

Water table depths within the valley are commonly less than 10 m and shallowest depths are often near the Boise River or canals and irrigated lands and in some areas the water table is at or near the ground surface (Petrich, 2004; Hutchings and Petrich, 2002b). The general groundwater flow direction is from northeast to southwest toward the Snake River. Many wells have been monitored for several years and a large number show an annual cycle of depth to water (DTW) values reaching a minimum during the summer months when irrigation is most active and a maximum depth during winter months when there is no irrigation and most of the canals are empty (Fig. 3). The Boise River is flanked by an extensive canal network (nearly 1700 km total) used for irrigation of the surrounding farmlands. The canals flow only during the irrigation season (March–October) and are diverted from the Boise River below Lucky Peak Reservoir or from smaller streams entering the valley. The largest diversion is the New York Canal, which redirects more than 57 $m^3 s^{-1}$ of water during peak irrigation season from the Boise River at Diversion Dam. During peak irrigation season as much water flows in the New York Canal as flows in the Boise River below the diversion.

Sixty-two percent of recharge to the aquifer is through canal seepage and another 30% is from irrigation while less than 1% is from mountain block recharge (Urban, 2004). Although the mountain block recharge component is small, Hutchings and Petrich (2002a) concluded that it is the dominant source of recharge in the northeast portion of the Treasure Valley near the basin margin. Mountain front recharge entering as seepage from mountain streams has been considered to provide very small contributions to overall recharge. Estimated total surface flow from all streams from the Boise Front Range is reported to be less than 0.12 $m^3 s^{-1}$ (Hutchings and Petrich, 2002a).

3. Noble gas thermometry

Weiss (1970, 1971) published some of the first work describing the dependence of noble gas solubility on temperature and salinity. Since this work several empirical equations have been developed to describe this behavior (Andrews, 1992; Benson and Krause, 1976). Here we use the relationship of gas solubility (S_H) with temperature described by Solomon et al. (1998):

$$S_{H,i} = \frac{[1.244 \times 10^6 \cdot \rho \cdot p \cdot e^{-M \cdot K_s}]}{1000 \cdot K_{H,i}} \quad (1)$$

and

$$\frac{1}{H_{H,i}} = \exp\left[a_i\left(\frac{b_i}{T_{NG}} - 1\right) + 36.855\left(\frac{b_i}{T_{NG}} - 1\right)^2\right] \quad (2)$$

where $S_{H,i}$ is the solubility of gas i (cm^3_{gas} STP cm^{-3}_{water}), ρ is the water density (g cm^{-3}), p is the atmospheric pressure (atm), M is the salinity of the water (mol L^{-1}), and K_s is the salting coefficient (dimensionless). $K_{H,i}$ is the Henry coefficient for gas i (g cm^{-3} atm) which is a function of environment (or recharge) temperature (T_{NG}) and empirical constants a_i and b_i (Solomon et al., 1998). Recharge elevation (H_R) is often a more useful output of noble gas thermometry then atmospheric pressure as it can be used to constrain the recharge source, particularly in mountain recharge environments. Atmospheric pressure decreases predictably with elevation so H_R is often estimated from atmospheric pressure p using any one of several empirical equations (e.g. Manning and Solomon, 2003).

Soil air is often entrapped within recharging water during infiltration and dissolved under increasing hydrostatic pressure

(Heaton and Vogel, 1981; Klump et al., 2007; Mercury et al., 2004; Wilson and McNeill, 1997). This excess requires a correction applied to the moist air equilibrium concentration determined from Eq. (1). Aeschbach-Hertig et al. (2000) incorporated effects of excess air entrapment into the previously developed models to form the closed-system equilibrium (CE) model:

$$C_i = C^*_{H,i} + \left[\frac{(1-F)\cdot A_e \cdot z_i}{1+\left(F\cdot A_e \cdot z_i / C^*_{Hi}\right)}\right] \tag{3}$$

where $C^*_{H,i}$ is the moist air equilibrium concentration for gas i, A_e is the initial entrapped air concentration ($\text{cm}^3_{\text{gas}}\ \text{cm}^{-3}_{\text{water}}$), z_i is the molar fraction, and F is the degree of fractionation (dimensionless) which accounts for the differences in solubility for individual dissolved gas species. Eq. (3) is generally solved using inverse methods to estimate the noble gas recharge temperature (T_{NG}), pressure (p) or recharge elevation (H_R), recharge salinity (M), excess air (A_e), and fractionation (F) from measured dissolved gas concentrations, generally Ne, Ar, Kr, and Xe. A priori information can be incorporated into the solution in the form of estimations of recharge salinity and constraints on recharge elevation. This reduces the number of unknown parameters and creates an over-determined problem. With these techniques dissolved gas concentrations in groundwater samples can be used to determine recharge sources by comparing the derived recharge temperature (T_{NG}) and elevation to those of potential recharge sources (T_r) (i.e. mountain block recharge or irrigation recharge).

4. Seasonal recharge

To use noble gas thermometry to distinguish between recharge sources or to distinguish proportions of recharge sources due to mixing it is necessary to know appropriate recharge source temperatures of the different sources present (Althaus et al., 2009). Soil temperature profiles trend toward the m.a.t. with increasing depth and many studies assume that the temperature at the water table is equal to the m.a.t. when water table depths are greater than a few meters (Andrews, 1992; Mazor, 1972; Stute et al., 1995; Thomas et al., 2003). Other studies however, have shown that infiltration and shallow water table depths lead to water table temperatures different from the m.a.t. (Castro et al., 2007; Cey, 2009; Kohl, 1973; Prunty and Bell, 2005; Wierenga et al., 1970). The timing of seasonal recharge (summer versus winter) can lead to recharge temperatures that are warmer or cooler than the mean annual temperature (respectively) (Beyerle et al., 1999; Cey, 2009; Manning and Solomon, 2003; Stute and Schlosser, 1993).

Stute and Schlosser (1993) provide a model for determining recharge temperatures (T_r) under the influence of seasonal infiltration. This model weights the temperature at the water table by the infiltration activity and integrates over 1 year to obtain a mean annual recharge temperature:

$$T_r = \frac{\int_0^{1\text{yr}} T_{wt}(t, z_{wt})\cdot I(t)dt}{\int_0^{1\text{yr}} I(t)dt} \tag{4}$$

where T_{wt} (°C) is the temperature of the water table, z_{wt} (m) is the depth of the water table, I (m d^{-1}) is the infiltration rate, and t (days) is elapsed time. This method assumes infiltration cycles are annual and that the infiltrating water is in thermal equilibrium with the soil which is supported by the work of Klump et al. (2007).

Direct measurements of T_{wt} are not always available, especially when water table depths fluctuate. In some cases T_{wt} can be estimated from the temperature at the soil surface with the assumption that the soil temperature cycles annually and decays exponentially with depth (Stute and Schlosser, 1993):

$$T(z_{wt},t) = \overline{T}_{G(0)} + T_a e^{(-z_{wt}(t)/\bar{z})}\sin\left(\frac{2\pi}{\tau}t - z_{wt}(t)/\bar{z} + \varphi\right) + \frac{\Delta T}{\Delta z}z_{wt}(t) \tag{5}$$

where $\overline{T}_{G(0)}$ (°C) is the average temperature at the soil surface, T_a (°C) is the amplitude of the annual surface temperature fluctuation, z_{wt} (m) is the depth of the water table, which may be a function of time (t), $\bar{z}$ (m) is the average penetration depth of the seasonal temperature variation, which can be determined from soil temperature profiles, τ (days) is the period of the temperature cycle (corresponding to 1 year for annual variations), φ (dimensionless) is the phase shift of the annual temperature fluctuation, and $\Delta T/\Delta z$ (°C m^{-1}) is the geothermal gradient, which can be neglected with water table depths less than 10 m (Anderson, 2005).

To investigate the influence of seasonal recharge in the Treasure Valley we calculated recharge temperatures using Eqs. (4) and (5) for a variety of infiltration rate functions $I(t)$ and water table depth functions $z_{wt}(t)$. The temperature of the water table was determined using Eq. (5) for climatic conditions ($T_{G(0)}$, T_a, and $\bar{z}$) representative of the Treasure Valley. $T_{G(0)}$ and T_a were determined from mean daily temperatures from 1999 to 2009 and set to 11.0 °C and 18.3 °C, respectively. The value of $\bar{z}$ was determined to be 2.3 m from fitting Eq. (5) to annual maximum and minimum soil temperatures measured at an AGRIMET station in the western Treasure Valley (United States Bureau of Reclamation, 2010). Values of infiltration were estimated from values of surface recharge from Petrich (2004) who determined recharge rates for the central portion of the study to be 0.65 m year^{-1}. Water table depth functions $z_{wt}(t)$ were taken from DTW data from several wells within the study area.

Well DTW values within the study area often show a strong seasonal fluctuation due to influence of summer irrigation and are shallowest during late summer when irrigation activity is at a maximum (Petrich, 2004; Petrich and Urban, 2004). We determined an infiltration-weighted recharge temperature for 18 wells within the study area for the purpose of illustrating the effect that seasonal variations in the water table can have on water table temperatures and thus recharge temperatures. All 18 wells were completed to <40 m depth to avoid high yield municipal and irrigation wells, and locally confined aquifers (United States Geological Survey, 2008). These 18 wells showed average annual water table depths between 1.5 and 10 m and seasonal fluctuation of ±0.25 to ±1.3 m. The average water table depth for all 18 wells was 3.6 m. Distribution of these 18 wells along with mean DTW and the range of annual fluctuations are shown in Fig. 4. Because DTW measurement frequency for each well varied between 1 day to a few months, to obtain values of z_{wt} for a full year we approximated DTW for each well over 1 year as a sine function with amplitude equal to the annual DTW fluctuation, a phase shift related to the timing of the minimum DTW, which often occurred between July and October, and a vertical (depth) shift equal to the mean annual DTW.

Six different recharge functions were modeled using Eqs. (4) and (5). They are: Constant (infiltration is at a constant rate over the entire year); Summer Normal (normal distribution of infiltration with mean = 180 d); Skewed Spring (skewed normal distribution shifted toward peak infiltration during spring, mean = 90 d); Skewed Fall (skewed normal distribution shifted toward peak infiltration during fall, mean = 260 d); Constant Summer (constant infiltration from April through October and zero in winter months); Canal Weighted (infiltration is weighted by New York Canal discharge). Infiltration functions for these scenarios are shown in Fig. 5 and the results are discussed in a later section. The results from all six scenarios show recharge temperatures approaching the m.a.t. as mean DTW increases (Fig. 6). Four of the six scenarios show recharge temperatures warmer than the

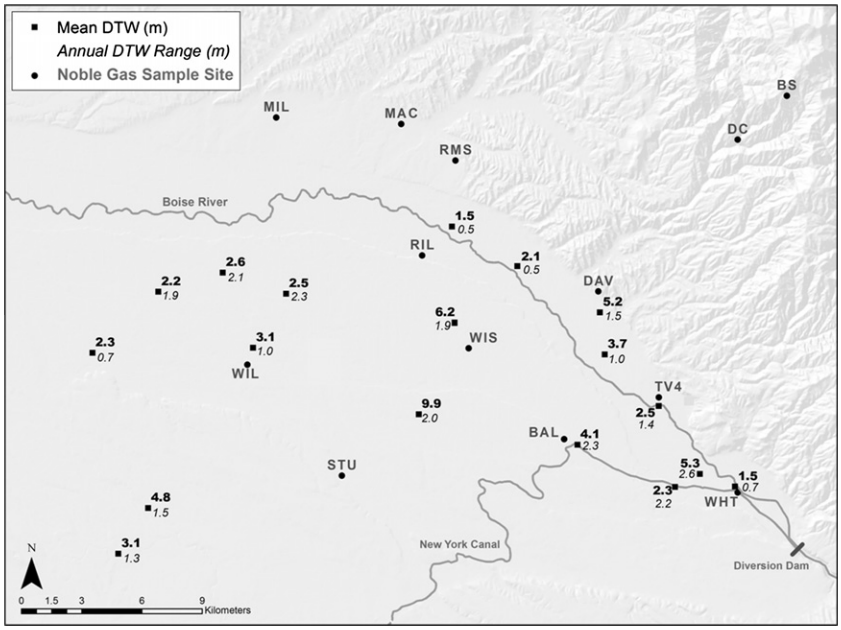

Fig. 4. Distribution of 18 wells used to model different infiltration scenarios and determine possible recharge temperatures for seasonal valley infiltration. Values listed are mean annual DTW values (bold) and range of seasonal fluctuations (italic). The locations of noble gas samples are also shown.

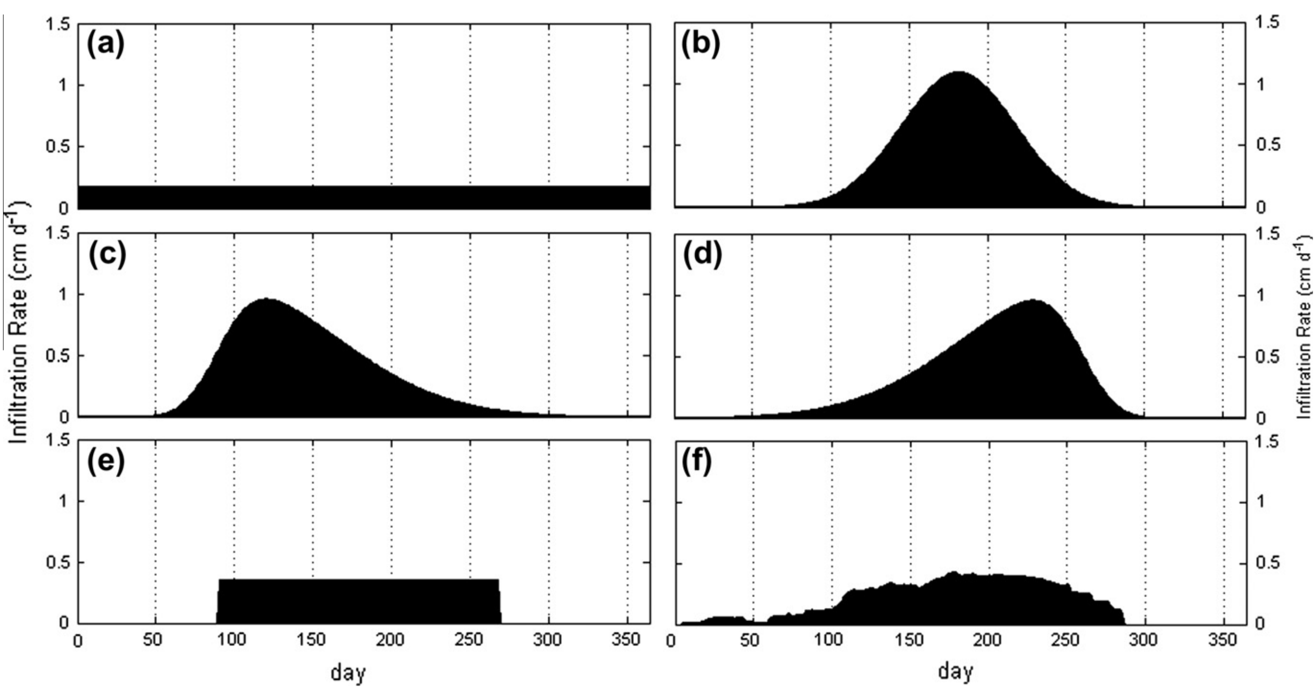

Fig. 5. Infiltration functions $I(t)$ used for the six different recharge scenarios. (a) Constant; (b) Summer Normal; (c) Skewed Spring; (d) Skewed Fall; (e) Constant Sumer; (f) Canal Weighted.

m.a.t. for mean DTW values less than ~2.3 m. This is likely because the wells that were chosen for investigation all showed shallowest DTW values in the late summer, when soil temperatures are warmest. An exception is the skewed-fall results which show an opposite trend where recharge temperatures are cooler than m.a.t. when mean DTW values are shallow. Variations about the m.a.t. are the result of different values of the timing of minimum DTW and annual DTW fluctuations.

5. NGT sampling and analysis

Eleven noble gas samples were collected from groundwater wells within the valley and two were collected from flowing springs within the Boise Front Range (Table 1). Noble gas samplers were limited to hydrostatic pressure less than ~3 m so all valley well samples were taken approximately 3 m below the well water level. This excluded sampling at distinct or multiple depths and so

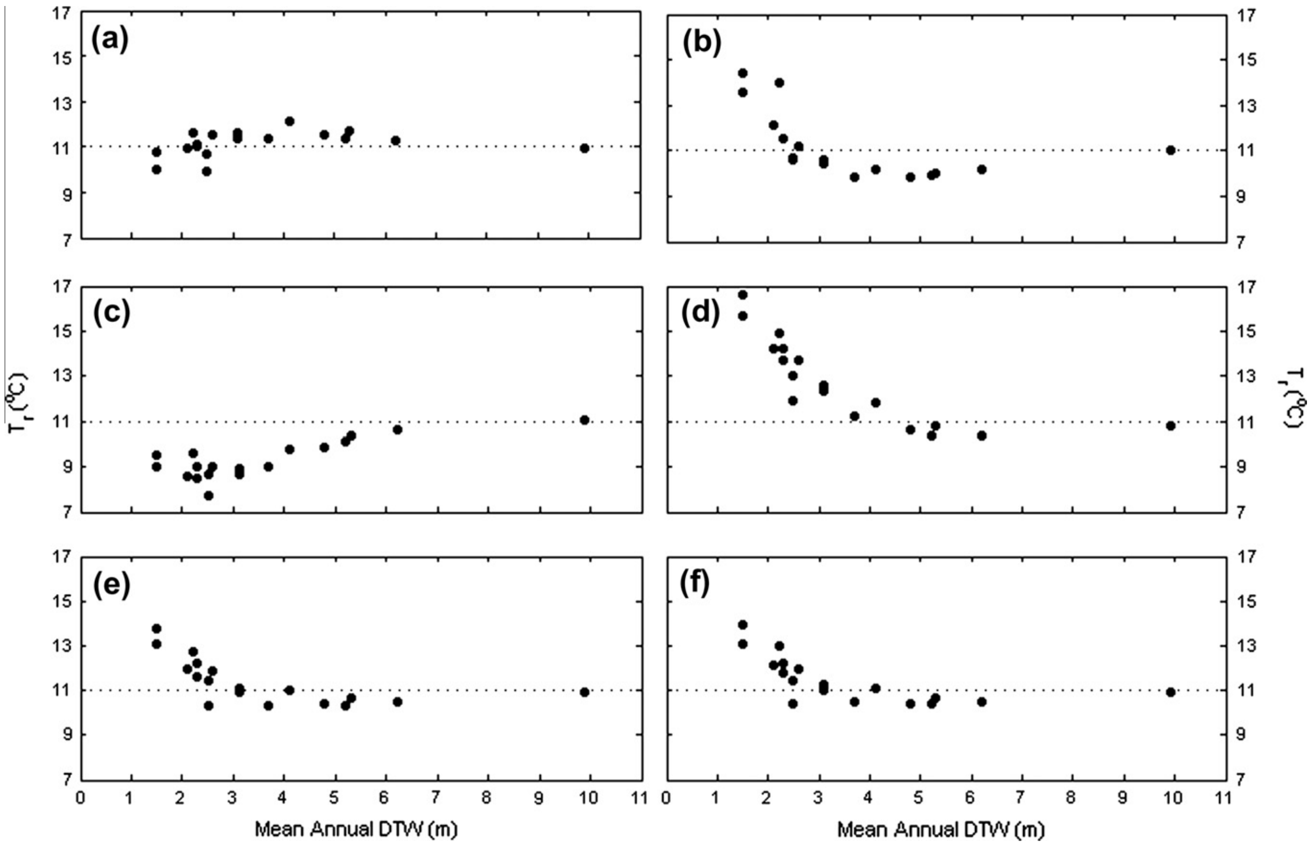

Fig. 6. Relationships between mean annual DTW and recharge source temperature of all 18 wells calculated for the six different recharge scenarios. (a) Constant; (b) Summer Normal; (c) Skewed Spring; (d) Skewed Fall; (e) Constant Sumer; (f) Canal Weighted.

sampling was assumed to be representative of the screened interval of the well. All but one of the sampled valley wells were residential and were therefore in continuous use, eliminating the need for purging. Sample TVHP was an observation well, so this well was purged prior to sampling. All except two wells were selected with open screens greater than 60 m BLS (the remaining two wells were screened below 40 and 55 m BLS). This depth of 60 m was described by Petrich (2004) as the lower extent of the Boise River terrace deposits and assumed beyond the reach of surface recharge from the Boise River. Two samples were collected from flowing springs within the Boise Front Range at elevations of 1297 m and 1859 m. Outflow from these springs is year round and maintains nearly constant temperatures. These samples were used to constrain recharge temperatures for the mountain block source.

Diffusion gas samplers were used to capture dissolved gas concentrations. Samplers were composed of a short length of gas-permeable silicon tubing connected to a short piece of copper tubing on each end. For a full description of these samplers see Manning (2002). The samplers were placed within the well below the water table or within the spring discharge as near the source as possible and left to equilibrate for a minimum of 3 days after which the samplers were removed, sealed, and sent for analysis. Total dissolved gas pressure (*TDG*) and water temperature (T_W) were measured upon installation and removal of samplers. Sealed samples were sent to the University of Utah Dissolved Gas Laboratory (Salt Lake City, UT) where they were analyzed for concentrations of N_2, O_2, Ne, Ar, Kr, and Xe (Table 2). Using the concentrations reported Eqs. (1)–(3) were solved using an error-weighted, non-linear inverse technique developed at the University of Utah Dissolved Gas Laboratory and based on Ballentine and Hall (1999) and Aeschbach-Hertig et al., 1999. Unknown parameters were adjusted simultaneously until a good fit to the measured gas concentrations was obtained. The fit was quantified by the sum of chi-squared ($\sum\chi^2$). Of the five unknown parameters (T_{NG}, H_R, M, A_e, and F), recharge salinity (M) was determined from estimates of precipitation salinity from Boise Front Range samples (Aishlin and McNamara, 2011) and set to a value of 0.1 M for all samples. Recharge elevation (H_R) was constrained between a lower limit equal to the elevation of the sample and an upper limit equal to the maximum elevation of the Boise Front Range (2100 m).

Noble gas-derived recharge temperature (T_{NG}), excess air concentration (A_e), and degree of fractionation (F), were adjusted simultaneously for each gas species using an iterative process until predicted gas concentrations and observed concentrations minimized the $\sum\chi^2$ for all gas species. With the method used for this study a chi-squared value less than 3.84 can be considered to have a proper fit of the modeled to the observed gas concentrations. H_R was adjusted between the elevation of the sample ($H_{R,min}$) and the maximum elevation of the Boise Front Range ($H_{R,max}$). T_{NG} was calculated for both $H_{R,max}$ and $H_{R,min}$. A maximum recharge temperature ($T_{NG,max}$) corresponds to $H_{R,min}$ and $T_{NG,min}$ corresponds to $H_{R,max}$. A linear relationship between $H_{R,min}$–$T_{NG,max}$ and $H_{R,max}$–$T_{NG,min}$ pairs in the H_R–T_{NG} plane for each sample provides the range of H_R and T_{NG} values (Fig. 7). The intersection of this line with the atmospheric lapse provides the optimal H_R and T_{NG} values when assuming T_r = m.a.t. (Aeschbach-Hertig et al., 1999; Manning and Solomon, 2003).

6. Results and discussion

The two samples collected from springs in the Boise Front Range produced T_{NG} values of 1.5 °C for the higher elevation

Table 1
Noble gas sample information including sample location data and laboratory noble gas results.

Sample	Sampling date	Elevation	Screened interval	Total depth	DTW	T_W	TDG	$^{32}O_2$ (10^{-2})	^{4}He (10^{-8})	^{20}Ne (10^{-7})	^{84}Kr (10^{-8})	^{40}Ar (10^{-4})	N_2 (10^{-2})	^{129}Xe (10^{-9})
		(m amsl)	(m bls)	(m bls)	(m bls)	(°C)	(atm)	(mg STP L^{-1})	(cm^3_{gas} STP cm^{-3}_{water})					
BS	1/26/2007	1859	Spring			6	1.00	901	5.10	1.90	6.20	4.60	1.60	4.40
DC	1/16/2006	1297	Spring			10	0.91	133	5.60	2.00	5.70	4.40	1.60	3.80
BAL	2/12/2007	851	75–80	80	26	12	1.09	0.1	7.10	2.74	5.15	4.02	1.91	3.03
DAV	2/19/2007	830	61–122	122	20	13	1.25	3.6	21.60	2.56	6.89	5.18	2.14	3.89
MAC	2/9/2007	802	55–80	80	19	13	1.08	455	6.01	2.32	5.52	4.47	1.69	3.66
MIL	1/17/2007	780	70–72	72	10	14	0.99	586	5.36	2.07	4.84	3.85	1.46	3.15
RMS	2/3/2007	819	112–115	115	18	13	1.17	0.1	6.18	2.09	4.90	3.82	1.51	3.35
RIL	1/17/2007	801	41–62	62	33	16	1.24	2.2	4.07	1.57	4.20	3.24	2.00	2.77
STU	2/9/2007	812	85–87	87	14	13	1.07	0.8	8.18	2.80	4.80	4.39	1.85	2.96
TVHP	2/19/2007	823	91–98	98	2	11	1.07	0.8	16.90	2.13	5.44	4.34	1.80	3.49
WHT	1/11/2007	860	70–85	85	35	14	0.93	0.1	4.22	1.64	4.49	3.39	1.25	2.71
WIL	1/16/2007	781	63–64	64	2	13	0.93	4.2	5.25	2.07	4.96	3.97	1.59	3.30
WIS	1/29/2007	813	76–79	79	6	15	1.00	5.5	5.53	2.11	5.03	3.89	1.60	3.57
						Precision (°C)	Precision (atm)				Error (%)			
						0.5	0.03	10	1	2	4	3	5	5

sample (BS, 1859 m) and 8.1 °C for the lower elevation sample (DC, 1297 m) (Table 2) for the optimal recharge elevations. The optimal H_R value for BS could not be determined using the method above but given how close (<300 m) the sample was to the maximum recharge elevation and how cool even $T_{NG,max}$ was, we set H_R to be the maximum elevation of the Boise Front Range (2100 m). This produced a T_{NG} value 3.5 °C cooler than the minimum m.a.t. for the Boise Front Range ($T_{r,min}$) which is near 5 °C and suggests seasonal recharge (spring snowmelt). The optimal value of H_R for DC was approximately 1650 m which is very close to the mean elevation between the sample location and the maximum elevation of the Boise Front Range suggesting recharge may be occurring over that entire elevation range.

Noble gas derived recharge temperatures for the valley wells ranged from ~6.0 °C to above 18 °C (Table 2, Fig. 8). Values that fall well below the m.a.t. of the valley floor (<11 °C) suggest some influence of mountain block recharge. Some of the cooler valley samples produced H_R–T_{NG} lines that do not intersect the local lapse rate but are cooler. This suggests that mountain block recharge temperatures are cooler than the m.a.t. Manning and Solomon (2003) used numerous mountain spring samples to provide support for cooler recharge temperatures of snowmelt in their study area in northeast Utah. McNamara et al. (2005) suggested that most groundwater recharge in the Boise Front Range occurs rapidly during the spring snowmelt. During this period recharging water will be cooler than the m.a.t. and DTW values will likely be shallower than the mean annual value.

The mean recharge elevation for the Boise Front was determined to be ~1475 m based on a method similar to Thomas et al. (2003). We used this elevation for H_R values of the three valley samples that were cooler than the m.a.t. lapse rate curve (RMS, TVHP, and WIL). With H_R set to 1475 m, all three of these wells produced T_{NG} values that were below the m.a.t. of the valley suggesting influence of mountain block recharge. Samples DAV, WIL, and MIL produced H_R–T_{NG} lines that intersect the m.a.t. lapse rate curve above the sample elevation, also suggesting influence of mountain block recharge. The optimal T_{NG} value produced from sample MIL, however, was very near the m.a.t. of the valley floor so it is difficult to conclude whether it is influenced from mountain block recharge. Most of the remaining valley samples (MAC, RIL, STU, and WHT) produced H_R–T_{NG} lines that were warmer than the lapse rate curve and did not intersect. For these samples H_R was set to the elevation of the sample ($H_{R,min}$). These samples then produced T_{NG} values greater than the m.a.t. of the valley floor. This implies that recharge to these wells is strongly influenced by seasonal valley infiltration.

Table 2
Noble gas derived values for all samples. Elevations are in meters AMSL.

Sample	Sample elevation (m)	T_{NG} (°C)	T_{NG} (±°C)	H_R (m)	A_e (cc_{gas}STP/cc_{water})	F (–)	$\Sigma\chi^2$
BS	1859	1.4	0.6	2100	0.90	0.84	1.24
BAL	851	12.4	1.3	851	0.02	0.29	2.02
DC	1297	8.0	0.6	1650	0.45	0.75	0.30
DAV	830	7.0	1.9	1850	0.16	0.59	3.81
MAC	802	16.6	0.8	892	0.90	0.67	0.64
MIL	780	10.9	1.4	1080	0.03	0.72	1.16
RMS	819	6.9	1.5	1475	0.05	0.50	2.82
RIL	801	14.2	2.9	801	0.03	0.90	0.37
STU	812	18.4	1.0	812	0.05	0.46	1.28
TVHP	823	6.0	1.9	1475	0.08	0.67	1.48
WHT	860	13.3	1.3	860	0.19	0.91	2.13
WIL	781	8.3	1.3	1600	0.03	0.68	0.21
WIS	813	5.7	1.6	1475	<0.01	0.12	0.58

The results of this study raise two separate issues regarding recharge to the Treasure Valley aquifer. First, some mountain spring samples and valley well samples produce T_{NG} values less than the minimum m.a.t. for the Boise Front Range. This suggests that mountain block recharge is occurring at lower temperatures than the minimum m.a.t. of the watershed. Manning and Solomon (2003) suggested that recharge temperatures in snow-dominated environments may be less than the m.a.t. Their study suggested a 2 °C difference between recharge temperature and m.a.t. At the BS sample location the difference would have to be nearly 4 °C. This would complicate the results implied by sample DC which does not directly suggest seasonal influence in the mountain block.

The second issue is with samples that produced T_{NG} values greater than the m.a.t. of the valley floor. It is likely that seasonal summer recharge sources are influencing recharge temperatures in a manner similar to how winter precipitation may be influencing mountain block recharge. By estimating seasonal recharge source temperatures for the valley under different recharge scenarios not uncommon to the Treasure Valley we have shown that recharge temperatures have the potential to exceed the m.a.t. by as much as 6 °C. These estimates, combined with observations from T_{NG} values, suggest that groundwater from valley wells with T_{NG} values above the m.a.t. of the valley floor are heavily influenced by seasonal recharge. Moreover, from the results of seasonal recharge investigation, higher T_{NG} values may suggest shallower local

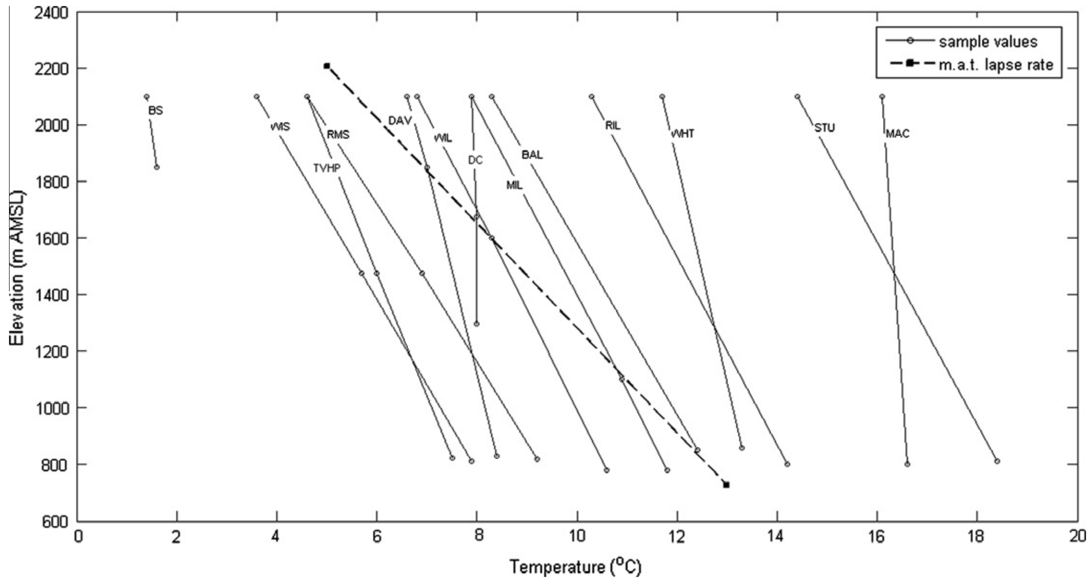

Fig. 7. H_R–T_{NG} plots for all samples. Several samples do not intersect the lapse rate line but fall either to the left (cooler) or right (warmer) of it which suggests recharge that is seasonal. Circles plotted between $H_{R,min}$ and $H_{R,max}$ are the determined optimal values of H_R and T_{NG}.

DTW values. Of the 18 wells analyzed to estimate seasonal recharge temperatures, the maximum potential recharge temperature for a canal-weighted infiltration scenario was 14.1 °C and the maximum temperature for all scenarios was 17.0 °C. These values are near the maximum T_{NG} values calculated from the valley well samples (Fig. 8).

7. Spatial patterns in T_{NG}

Hutchings and Petrich (2002a) have shown that groundwater in the Treasure Valley near the margin with the Boise Front Range has a chemical signature similar to that of mountain block water while in a second report they showed that many wells in the vicinity of the New York Canal to depths of a few hundred feet have elevated tritium values and geochemical signatures suggestive of canal and/or irrigation recharge (Hutchings and Petrich, 2002b). Petrich and Urban (2004) produced potentiometric maps from groundwater well data that suggest some water is entering the valley aquifer from across the northeast margin while Urban (2004) estimated that recharge from surface sources dominates away from the basin margin. This pattern of influx of mountain water should also be reflected by noble gas recharge temperatures. Most of the wells that produced cooler temperatures are found closer to the basin margin, with a few exceptions, and thus suggest mountain block recharge (Fig. 9). An alternative hypothesis would be that these cooler temperatures are the result of seepage from mountain front streams, which are dominated by cooler, spring runoff, as they enter the more permeable sediments of the valley. However, given the low values of stream discharge from the Boise Front Range (mean annual discharge <0.12 $m^3 s^{-1}$ the majority of which exits at the northwestern corner of the study area) it seems unlikely that surface water would be a major contributing factor to deep wells in the Treasure Valley.

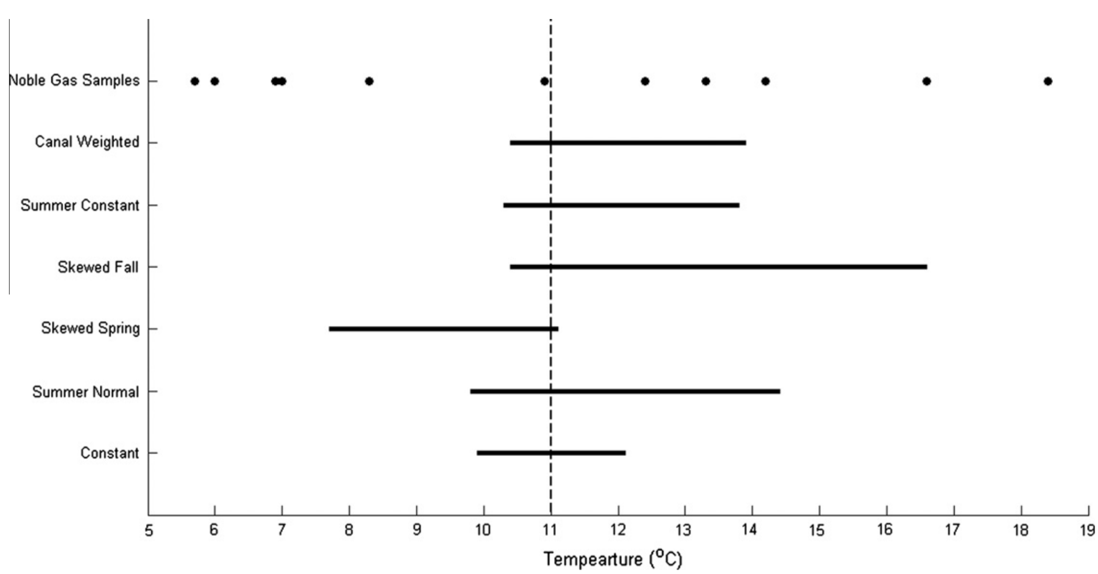

Fig. 8. Ranges of recharge temperatures produced using Eqs. (4) and (5) for six different infiltration functions. Symbols along the top row show the noble gas derived recharge temperatures from the valley well samples in this study and the vertical dashed line shows the m.a.t. of the Treasure Valley.

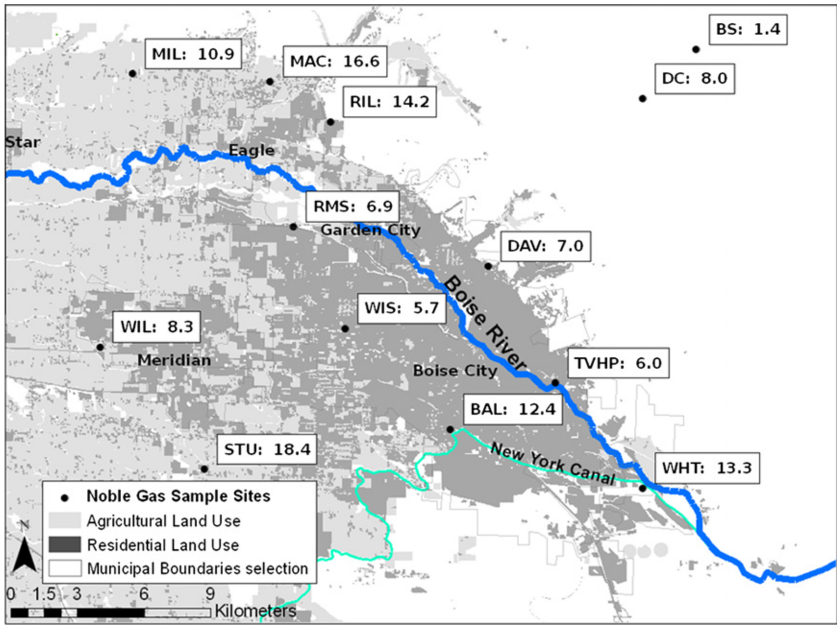

Fig. 9. Calculated T_{NG} values from valley well samples along with land use. Lower T_{NG} values are mostly found within or near residential dominated areas (dark gray) and higher values are found in agricultural areas (light gray).

Samples RIL, MAC, and MIL are near the basin margin but reported higher T_{NG} values. These samples are in an area dominated by agricultural land use (Fig. 9). Sample WIL also appears to be inconsistent with the conceptual model of groundwater flow with a T_{NG} value less than the m.a.t. of the valley even though it is nearly 20 km from the basin margin and surrounded by agricultural land. However, WIL is also located very near the municipal area of Meridian, Idaho, an area dominated by residential land use that would be experiencing much less surface recharge than the agricultural lands that surround it. It is possible that groundwater beneath municipal boundaries may be dominated by ancient mountain block recharge because there has not been sufficient recent surface recharge to displace it. Samples within the municipal boundary of Boise, Idaho also show this pattern.

The distribution of noble gas recharge temperatures suggests mountain block recharge is entering the valley from the northeast margin but is limited to urban areas near the basin margin, while irrigation recharge is entering vertically through the valley floor where the dominant land use is agricultural. It is expected that warmer T_{NG} values will be found in areas with shallow mean DTW values. However, there is little spatial correlation between calculated seasonal recharge temperatures (T_r) and noble gas derived recharge temperatures (T_{NG}) from valley wells. We attribute this lack of correlation to the sparse sampling of both noble gas recharge samples and seasonal recharge wells and the different magnitudes of seasonal DTW fluctuations. A more thorough analysis of recharge distribution and water table distribution over the Treasure Valley would likely produce a more accurate distribution of recharge source temperatures.

8. Conclusions

In noble gas thermometry recharge source temperatures are often assumed to be equal to the m.a.t. at the location of recharge. However, this can become complicated in situations where recharge is seasonal or water tables depths are shallow, or seasonally variable. We have found that 8 of the 13 samples analyzed for noble gas recharge temperatures failed to produce results consistent with the standard noble gas thermometry assumption that recharge temperature can be approximated by the m.a.t. Previous investigations within the study area have suggested that recharge is seasonal and occurs under conditions when the depth to water (DTW) is within a few meters of the surface. Results of applying an infiltration-weighted recharge temperature model to DTW values produced potential recharge temperatures much greater than the m.a.t. We investigated several recharge scenarios and concluded that recharge source temperatures may be as much as 6 °C above the m.a.t. of the valley floor in the northeast Treasure Valley. Calculated noble gas recharge temperatures from valley well samples range from 5 °C below to 7.4 °C above the m.a.t. of the valley floor. Cooler recharge temperatures suggest contributions of mountain block recharge from the adjacent Boise Front Range while warmer recharge temperatures suggest seasonal valley irrigation recharge. By comparing noble gas temperatures with the range of possible recharge source temperatures we have concluded that wells near the Boise Front Range and within municipal boundaries are recharged (in part) from underflow through the adjacent mountain block and wells located further into the valley where agricultural land use is dominant are recharged by seasonal irrigation sources.

When determining recharge source temperatures for the purpose of noble gas studies it is important to take into account the timing of recharge and, more importantly, water table depths and fluctuations. Results of recharge temperature simulations from this study suggest that water table depth and timing of recharge are a significant factor controlling recharge temperatures when mean water table depths are less than ~3 m. This study shows that coupling an infiltration-weighted recharge temperature model with standard noble gas thermometry techniques can provide an effective tool for evaluating recharge sources in basins complicated by seasonal recharge.

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