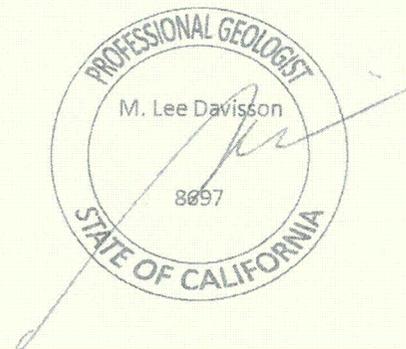


**Constraints on the Recharge Sources, Flowpaths, and Ages of
Groundwater in the Amargosa River Valley Using Stable Isotope,
Water Quality, and Noble Gas Data**

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INTRODUCTION

There is a need to elucidate further the recharge sources of spring discharge waters within the Amargosa River watershed to help in sustaining or restoring natural flow and anticipating potential upstream impacts. Natural spring flow supports the river flow and habitat regimes, but specific recharge sources and subsurface transport times are still elusive. Although water quality parameters provide some measure of differentiating groundwater, they are often subject to change during subsurface migration. Needed is a measurement tool that potentially provides a conservative tracer that can further differentiate groundwater sources. Dissolved noble gas abundances and helium isotope measurements have demonstrated over the past ~30 years the ability to derive the temperature and elevation of groundwater recharge, provide a qualitative measure of its age, and they are inherently inert and don't react with aquifer rocks (Ozima and Podosek, 2002; Kipfer et al., 2002). Dissolved concentrations of helium, neon, krypton and xenon can be compared to well-known equilibrium solubility curves that vary as a function of temperature and elevation. The relative abundances of the measured noble gases should conform to a specific recharge temperature and elevation in this comparison, which could provide an important boundary condition for recharge areas, particularly when used in conjunction with other available isotope and water quality data. The ^3He and ^4He isotope ratio is well known in the atmosphere and water in contact with the atmosphere. Any deviation from this known ratio in collected groundwater will indicate either 1) accumulation of excess ^4He , which is common for groundwater out of atmospheric contact for thousands of years, or 2) excess ^3He , which is commonly observed in groundwater influenced by an active geothermal source in communication with the Earth's mantle. The latter is commonly observed in geothermal processes in active volcanoes (e.g., Welhan et al., 1988).

Note that caveats are always possible with these measurements. Two of which are potential negative outcomes would be: 1) the possibility that any diffuse discharge of spring water compromises the integrity of noble gas abundances dissolved in groundwater, and 2) other dissolved gases in abundance (e.g., hydrogen sulfide, methane) are present in the groundwater and have compromised noble gas abundances due to gas stripping. Unfortunately, the results presented below indicate perhaps both these processes affected dissolved noble gas abundances in groundwater of the Amargosa River Valley. Consequently, in addition to illustrating these results, this report further elucidates recharge sources, potential flowpaths, and groundwater quality using previously reported data. In addition, groundwater ages are discussed in context of the $^3\text{He}/^4\text{He}$ results.

FIELD AND LABORATORY METHODS

Stable Isotopes

Samples for $\delta^{18}\text{O}$ and δD were collected in 60 milliliter glass bottles equipped with an conical shaped insert inside the cap that when the bottle is closed forms an airtight seal. Samples were shipped to Isotech Laboratories in Champaign, Illinois where the $^{18}\text{O}/^{16}\text{O}$

and D/H ratios were measured as a gas using standardized mass spectrometry methods. Results are reported as a normalization to Standard Mean Ocean Water (SMOW), which is an internationally recognized standard in stable isotope analysis. The normalization was converted to standard δ (“del”) notation following the convention

$$\delta = \left(\frac{R}{R_{std}} - 1 \right) 1000$$

where R is the isotope ratio of the sample and R_{std} is the ratio of the standard.

Noble Gas

Noble gas samples were collected in passive diffusion samplers comprising two sections of 1/4” copper tubing attached by a small section of semipermeable silicon tubing (Fig. 1). The terminal ends of the copper tubes were pinched closed with a gas-tight cold seal. This sampler was placed in the water to be sampled for 24 hours. During this equilibration period gases dissolved in the water diffuse through the semipermeable tube and come into an equilibrium concentration inside the copper tube section in proportion to that of the water (Gardner and Solomon, 2009). At the same time, a special meter was used to measure the total dissolved gas in the water. After 24 hours, the sampler is crimped to a cold seal on the semipermeable tube end of the copper to form two separate gas samples enclosed in copper. These two samples were then labeled, the end protected with electrical tape and placed into a plastic bag. Five sample sites in total were collected by this method. All samples were sent to the noble gas laboratory at the University of Utah. There the copper tubes were vacuum fitted to an evacuated container, the copper cold seal was uncrimped to release the gas, followed by cryogenic isolation of noble gases of interest. Noble gas abundances and the $^3\text{He}/^4\text{He}$ ratios were measured on a VG-5400 noble gas mass spectrometer. Results are reported as gas volume per milliliter of water.

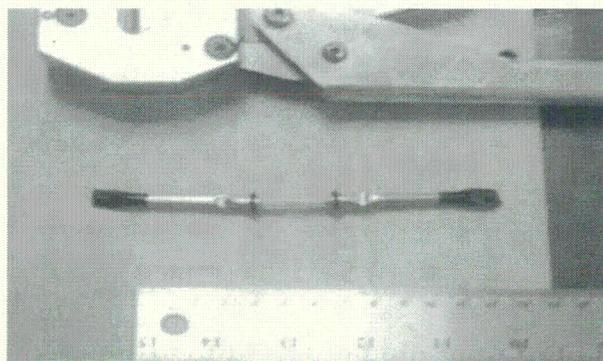


Figure 1. Passive diffusion sampler used for collection of dissolved noble gases in water samples.

PREVIOUS ISOTOPE STUDIES

A number of previous reports have been published on groundwater geochemistry and isotope abundances in southern Nevada and southeastern California. Notable reports relevant to the Amargosa River area include those of Winograd and Thordarson (1975), Thomas et al. (1996), Davisson et al. (1999), and Larsen et al. (2001). Additional studies that include directly related data can be found in Thomas et al. (2003a and 2003b) and Hurst (2012). Winograd and Thordarson (1975) developed one of the early frameworks for groundwater flow in southern Nevada related to the Nevada Test Site which included extensive discussion of the Ash Meadows springs discharge area. Based on earlier work, they also summarized types of groundwater hydrochemistry. These included calcium-magnesium-bicarbonate (Ca-Mg-HCO₃) groundwater associated with both the carbonate rock of the Spring Mts. and adjacent Pahrump Valley, sodium-potassium-bicarbonate (Na-K-HCO₃) groundwater that drains the largely volcanic rock areas south of the Nevada Test Site (e.g., Oasis Valley and Jackass Flats), and Ash Meadows spring discharge which has Ca-Mg-Na-HCO₃ water that Winograd and Thordarson (1975) inferred as a mixture of recharge of the two latter water types.

Thomas et al. (1996) also compiled and summarized groundwater chemistry types, as well as isotope abundances in areas that included groundwater throughout southern Nevada and southeastern California with a focus on the regional carbonate aquifers. They concluded from isotope results that the Ca-Mg-Na-HCO₃ water discharging from Ash Meadows springs comprised 60 percent Spring Mts. recharge and 40 percent from Pahrangat Valley to the east. They also argue from radiocarbon data that groundwater velocities ranged approximately from 10 to 144 feet per year.

Davisson et al. (1999) showed that radiocarbon was not a reliable method for age dating groundwater in the regional carbonate aquifer due to continual isotope exchange reactions combined with mixing of local recharge sources during long-range transport. They further showed that stable isotopes of oxygen-18 (¹⁸O) and deuterium measured in southern Nevada groundwater had been previously evaporated during its original recharge as melted snow in central Nevada (Rose et al., 1999). Applying a methodology that removed the effects of evaporation on oxygen-18 and deuterium they showed a systematic decrease in their abundances with increasing latitude and local elevation throughout southern Nevada, a result inconsistent with previous studies purporting groundwater recharged during the Pleistocene in the last glacial period (Claassen et al., 1986).

Larsen et al. (2001) studied the water quality and stable isotope abundances of groundwater in the Tecopa and Death Valley regions of the Amargosa River Valley and related them to groundwater of southern Nevada to delineate potential recharge sources. They recognized three water types comprising Spring Mts. recharge source, a deep

regional groundwater derived from fracture flow of southern Nevada, and groundwater derived from basin-filled groundwater of the Amargosa Desert.

Additional studies providing a greater variety of isotope measurement types have been reported by Thomas et al. (2003a) and Hurst (2012). Thomas et al. (2003a) focused specifically on Oasis Valley and its hydraulic connection to Pahute Mesa, showing that Oasis Valley groundwater is replenished by groundwater flowing through Pahute Mesa that was ultimately derived further north. The Oasis Valley groundwater ultimately replenishes the Amargosa Desert basin fill aquifers.

Hurst (2012) specifically focused on tritium, ^{18}O , deuterium, strontium isotopes, and uranium isotopes in regions along the Amargosa River Valley. He showed that spring samples are largely tritium absent, the ^{18}O and deuterium show only limited evaporation, and that strontium and uranium isotopes show mixing along the entire length of the Amargosa River Valley.

Lastly, one study reported by Thomas et al. (2003b) measured dissolved noble gases in the regional carbonate aquifer of southern Nevada. They showed that noble gas abundances that are typically incorporated in recharging groundwater and reflect the local recharge temperature were systematically being lost during long-range transport from Pahrangat Valley in east-central Nevada towards Ash Meadows at its terminal discharge point. They concluded this loss of dissolved gas was due to fault barriers and cavities in the regional carbonate aquifer that forces groundwater to migrate upward and encounter gas loss in air pockets or vadose zones. This subsequently masked the calculated recharge temperatures derived from the noble gases.

THIS STUDY

This study investigates specifically the relationship between the Tecopa hot springs area and groundwater just to the north and south of these springs. The specific questions address are 1) the hot springs derived from a distinct groundwater perhaps of a more deeper regional source, or 2) are they the same groundwater as that to the north and south but have encountered a local heat source such as a magmatic intrusion. We use both stable isotopes compiled from previous studies along with new data generated on the dissolved noble gases in the water. In addition, we offer additional discussion of groundwater quality results in the context of these new data results.

Below a compilation of stable isotope data from Thomas et al. (1996), Winograd et al. (2003), and Hurst (2012) are used to develop and integrated picture of stable isotope distribution across the Amargosa River corridor to assess potential recharge sources of

springs in this area. This assessment will provide a framework to interpret further the noble gas results and water quality results.

RESULTS FROM THIS STUDY

Integrated Assessment of Stable Isotope Data

The stable isotopes of ^{18}O and deuterium in precipitation systematically vary with increasing latitude and elevation (see Appendix Fig. A-1). This results in lower $\delta^{18}\text{O}$ and δD isotope values in groundwater from north to south from central Nevada to southeastern California. There is also a regional effect in the American Southwest where summer monsoonal precipitation occurs in areas directly north of the Gulf of California, causing substantial precipitation in some higher elevation areas. This summer monsoonal rain has higher isotope values than winter season equivalents because of warmer temperatures. Local high elevation areas such as Spring Mts., which support annual snow accumulation that promote recharge during the winter, has higher $\delta^{18}\text{O}$ and δD values than groundwater found in Oasis Valley on the south side of Pahute Mesa (Fig. 2). Spring Mts. see some portion of their annual precipitation from summer monsoons. The Oasis Valley groundwater ultimately is derived from recharge further north of Pahute Mesa where isotopic values of mean precipitation are even lower and are predominately influenced by winter precipitation (Davisson et al., 1999). This geographic dependency of isotope values provides a means to use these differences to potentially derive recharge sources of groundwater sampled in the Amargosa River corridor. Ash Meadows groundwater isotope values consequently is a mixture between Spring Mts. and Oasis Valley and/or Pahrangat Valley (Winograd and Thordarson, 1975; Thomas et al., 1996). Comparison of measured $\delta^{18}\text{O}$ and δD for groundwater and springs is shown in Figure 2 below to illustrate differences among specific geographic groupings. A $\delta\text{D}-\delta^{18}\text{O}$ plot is illustrated in highlighted in color for each geographic grouping plotted along with a shaded relief map. Note that Oasis Valley has the lowest isotopic values of all the groupings. In contrast, the Tecopa area springs and wells have the highest. Consequently, their recharge sources are distinctly different, and based on these isotope results we can rule out that Oasis Valley groundwater is a sole source of Tecopa groundwater. Note further, three additional samples were measured for $\delta\text{D}-\delta^{18}\text{O}$ during this study. They include ARHS-1 (-91 and -11.0) above Shoshone, Twelve Mile Spring (-99 and -13.6) from the Chicago valley, and Dodge City Spring (-95 and -12.0) from the Tecopa area. All these results are consistent with values measured in the Tecopa area.

Note that the Spring Mts. isotope values are much higher than Oasis Valley groundwater and form a fairly narrow range that conforms to the Global Meteoric Water Line. Comparison among isotope values in Figure 2 for Oasis Valley, Jackass Flats, Springs

Mts. and Ash Meadows indicate that Ash Meadows values overlap all these other groupings and support their recharge sources as a mixture among them. However, the isotope values in the Tecopa grouping only have a moderate overlap with the Ash Meadows data but also overlap with Spring Mts. type isotope values. The one Tecopa area groundwater that overlaps with Ash Meadows is Borax Spring, which suggests Ash Meadows type groundwater as their potential recharge source. The remainder of the Tecopa groundwater is clearly influenced by a more Spring Mts. type recharge isotope value. However, recall that isotope values increase progressively toward the south, which requires that additional recharge source with similar isotope values as Spring Mts. be considered. We also know that the Kingston Range to the south also have higher isotope values. Our only evidence so far is for Crystal Spring that drains from those ranges, which has an evaporated isotope signature and lies to the right of the Global Meteoric Water Line. Nevertheless, its recharge into the groundwater beneath Tecopa can influence the isotope values causing them to be higher than the Ash Meadows grouping. Note also that Sheep Creek Spring also has a higher isotope value and conforms closely with the Global Meteoric Water Line, confirming that local precipitation in this area is much higher than Ash Meadows isotope values. The isotope values for Sheep Creek Spring also illustrate recharge does occur in the areas along local high elevations.

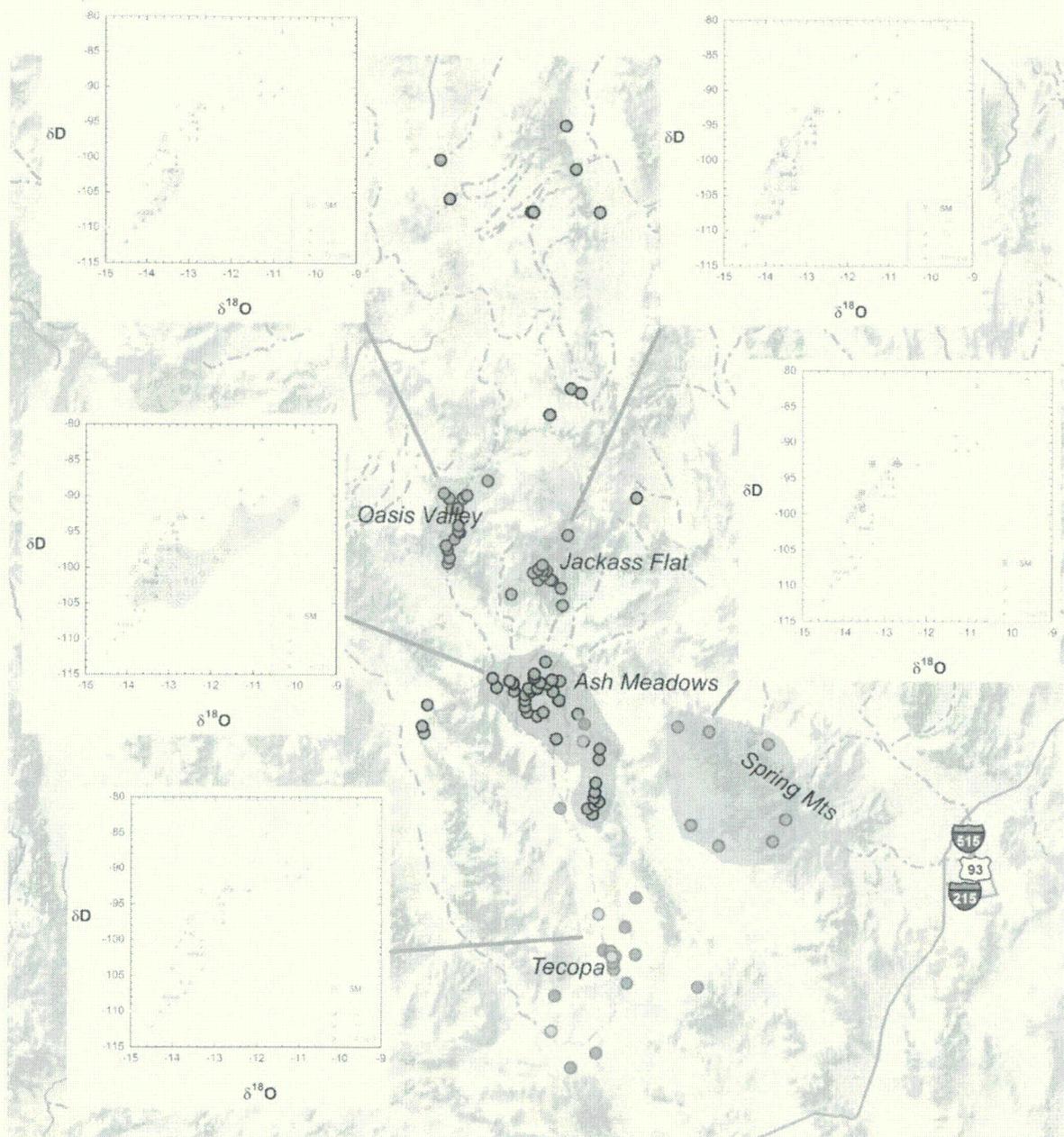


Figure 2. δD - $\delta^{18}O$ plots are compared as regional groupings in this map view. Note that the range in δD and $\delta^{18}O$ values decreases in general from north to south and that the Tecopa region groundwater overlaps most with Spring Mts. and Ash Meadows. This suggests that either are potential sources for Tecopa groundwater, although for the latter mixing with Spring Mts. or possibly Kingston Range recharge would be required.

Integrated Assessment of Water Quality

Groundwater quality in the Amargosa River Valley tends toward high total dissolved solids contributed by appreciable levels of chloride and sulfate. In order to place this

water quality in context with regional groundwater flow, it is best served to compare them to groundwater further north that may or may not contribute to recharge in the Shoshone and Tecopa regions. Figure 3 below is a piper plot that compares water quality of groundwater from the regional carbonate of southern Nevada (in red), groundwater from the Nevada Test Site (in green), and groundwater of the Amargosa River Valley (in blue). Note that the regional carbonate is Ca-Mg-HCO₃ type water and contrasts with the Nevada Test Site water of Na-K-HCO₃ type. Note further that the Ash Meadows water (open red squares) is a mixture of regional carbonate and Nevada Test Site water quality types. The Amargosa River Valley groundwater is also dominated by Na-K-HCO₃ type water quality, but with increasing amounts of chloride and sulfate progressively downgradient. An increase in total dissolved solids also occurs in many of these waters.

Regional Carbonate, NTS, and Amargosa River Valley

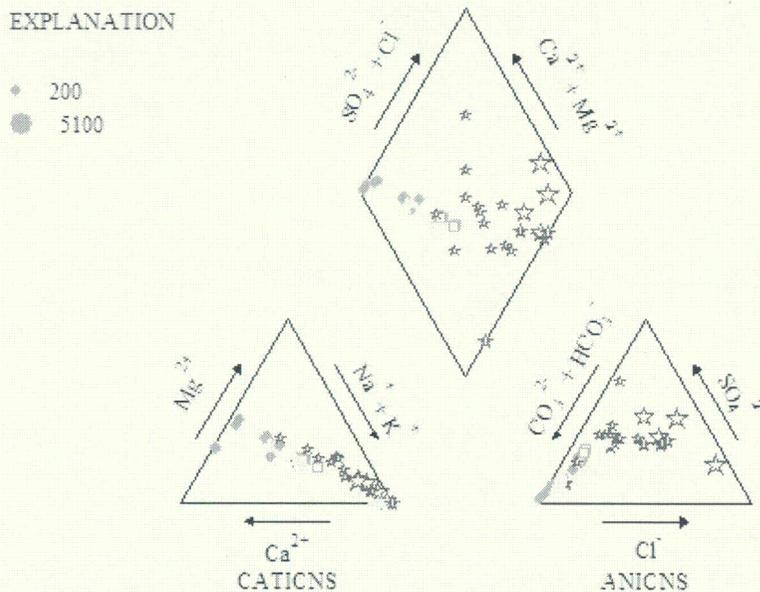


Figure 3. Piper plot comparing cation and anion relative concentrations in groundwater of the regional carbonate aquifer (red circles), Ash Meadows (open red squares), Nevada Test Site (green triangles), and Amargosa River Valley (open blue stars). Note that between the regional carbonate aquifer and the Amargosa River Valley groundwater, water quality changes from Ca-Mg-HCO₃ type toward Na-K-HCO₃-Cl-SO₄ type accompanied by increased salinity.

Two potential processes may influence the water quality in the Amargosa River Valley groundwater. One may be the ubiquitous Tecopa lake bed geologic deposits that uniquely occur in the Tecopa region. These lake beds were accumulated during high stands of glacial ice in the Pleistocene period of Earth's history that caused large amounts of

surface runoff that accumulated in closed basin lakes in Nevada and southeastern California. These lakes ultimately went through wet/dry cycles that accumulated precipitated salts (predominately chlorides and sulfates), much like that seen today in desert playas. Modern groundwater encountering the Tecopa lake beds will undoubtedly dissolve salts in the sediment which will contribute to increase salinity of the water.

The additional process that likely contributes to increase salinity in Amargosa River Valley groundwater is the unusual amount of geothermal heat emitted in the Tecopa region. As a matter of fact, high heat flow has been measured throughout this region from the Springs Mts. in the east to the eastern edge of the Sierra Nevada. The heat flow is significantly lower northeast of Ash Meadows, corresponding to the regional carbonate aquifer, but in contrast heat flow is high under the Nevada Test Site. This is best illustrated by groundwater temperatures, which can range up to 45°C in parts of the Nevada Test Site. Groundwater in the Tecopa region also has high temperatures up to 40°C. The regional carbonate aquifer groundwater temperature tends range between 20-30°C. The increase in groundwater temperature can have significant impact on groundwater quality. For instance, dissolved silica increases with increasing temperature, and bicarbonate simultaneously will decrease. This can effect water pH and promote solubility of other elements. A good example is arsenic, which increases solubility with increasing pH (Fig. 4). Arsenic in the Amargosa River Valley may ultimately be associated the lake beds deposits, but as pH increases, their solubility increases to significantly high levels.

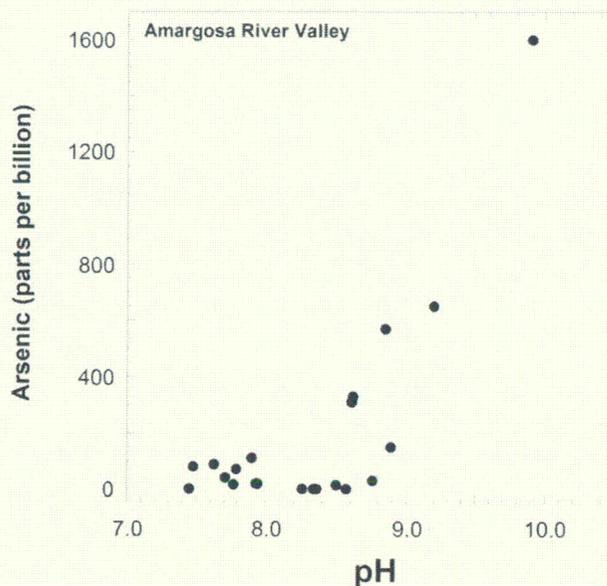


Figure 4. Arsenic solubility increases with increasing pH as illustrated by groundwater in the Amargosa River Valley region. The ultimate source of arsenic is not known but could be associated with the Tecopa lake beds deposits.

Noble Gas Results

The noble gas results are shown below in Table 1. They are presented as volume concentrations (in standard temperature and pressure or STP) per gram of water. The helium concentrations are measured and reported as ^4He isotope, which is essentially all of the dissolved helium. The R/R_a is the $^3\text{He}/^4\text{He}$ isotope ratio normalized to the same ratio as measured in air (air $^3\text{He}/^4\text{He}$ ratio = 1.384e^{-6}).

Table 1. University of Utah - Dissolved and Noble Gas Lab

Sample I.D.	Ar total (ccSTP/g)	Ne total (ccSTP/g)	Kr total (ccSTP/g)	Xe total (ccSTP/g)	He4 (ccSTP/g)	R/Ra
ARHS-1	2.64E-04	1.86E-07	5.77E-08	7.54E-09	2.75E-07	0.243
Borehole Spring	2.48E-04	1.56E-07	5.56E-08	7.77E-09	6.79E-06	0.089
Tecopa Hot Spring	2.37E-04	1.69E-07	5.34E-08	7.29E-09	5.46E-06	0.090
Thom Spring	2.84E-04	1.96E-07	6.26E-08	8.16E-09	6.52E-09	0.086
Wild Bath Spring	2.58E-04	1.68E-07	5.56E-08	7.87E-09	7.74E-06	0.085

The approach with analyzing the data in Table 1 is two-fold. Firstly, the relative abundances among the different noble gas concentrations for each sample must be compared to models of equilibrium solubility expected for different temperatures and elevations. Secondly, the R/R_a needs to be evaluated as a potential measure of groundwater age.

Recharge Temperature/Elevation

We start first with determining the best possible recharge temperature and elevation for each sample. But before we can do this, the data need to be evaluated for any potential process that may have compromised their integrity. For example, sampling artifacts are always possible such as loss of gases during sample sealing, any air-water exchange at the sample site affecting representation of the groundwater, or laboratory analysis error. Fortunately, a previous study by Thomas et al. (2003b) for the regional carbonate-Ash Meadows system of southern Nevada provides a comparative framework for the data in Table 1. This comparison is facilitated by Figure 5 below. Here the Ne concentration is compared to Xe of both the Amargosa River Valley samples (solid circles) and those reported by Thomas et al. (2003b) (open squares). The solid line shows recharge temperatures calculated from their model of equilibrium solubility (solid line) and curves for the excess air (dashed lines). Note how Amargosa River Valley results overlap with their results, particularly with those of the lowest Xe concentrations. The lowest five points plotted from Thomas et al. data are samples measured for Ash Meadows springs and illustrates the striking similarity with the Amargosa River Valley groundwater we measured. They argued that the Ash Meadows spring waters experienced dissolved gas

loss during its transport from upgradient in the regional carbonate aquifer. That gas loss they proposed was due to faulted barriers that forced deep groundwater upward and induced depressuring. Adjacent vadose zone or air cavities in karst permeability would exchange with the degassing water and re-equilibrate dissolved gas concentrations and influence calculated recharge temperatures. If this mechanism is valid, then it would suggest that the Amargosa River Valley groundwater undergoes a similar process and results in disagreement in calculated recharge temperatures among the different noble gases.

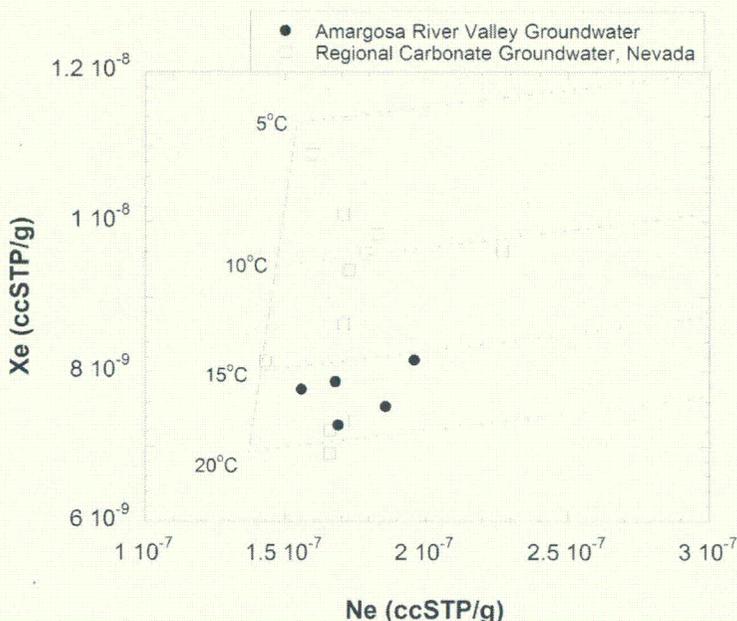


Figure 5. Comparison between Ne and Xe data of the Amargosa River Valley samples (solid circles) and those reported by Thomas et al. (2003b) (open squares). Shown is their recharge temperature model curve (solid line) and excess air curves (dashed lines). Note how Amargosa River Valley results overlap with their results of Ash Meadows springs (their lowest five points).

We also constructed equilibrium solubility curves for each noble gas over the plausible recharge temperature and elevation in southern Nevada and southeastern California (5-25°C and 500-7000 feet). The Ar, Kr, and Xe results for each sample were analyzed independently for the best fit to a recharge temperature and elevation and compared with each other. The results are shown in Table 2. All of the samples yield an unusually high recharge temperature similar to Thomas et al. (2003b) with possibly the exception of Borehole Spring. Results of each noble gas comparison are indicated and it was common that two of the noble gases would have a common recharge temperature and elevation, but not the third one. Only in Borehole was there good agreement among all the three noble gases. Note that in this case, the noble gases suggest Borehole was recharged at approximately 6400 feet with a recharge temperature of around 19°C. In order to

compare this with plausible locations, an atmospheric lapse rate for southern Nevada and southeastern California was calculated for data obtained from the Western Regional Climate Center (<http://www.wrcc.dri.edu/CLIMATEDATA.html>). This resulted in a *maximum* daily mean air temperature of 17°C at 6400 feet, and a *minimum* of approximately 2°C. The stable isotope values of Borehole Spring ($\delta^{18}\text{O} = -12.9$ and $\delta\text{D} = -95$) limit recharge to Ash Meadows and/or Spring Mts./Kingston Range. Noble gas recharge temperatures for groundwater sampled on Spring Mts. by Thomas et al. (2003b) ranged between 4 and 10°C, temperatures consistent with *mean* air/soil temperature, but *inconsistent* with Borehole Spring. This suggests that the 19°C recharge temperature is not consistent with temperatures implied by the calculated recharge elevation of 6400 feet. Consequently, it is probable that the dissolved noble gas abundances of Borehole Spring have also been compromised similar to the other groundwater samples in Table 1. It is entirely conceivable that loss of dissolved Ar, Kr, and Xe in Borehole Spring occurred without differential partitioning of each elemental gas and they maintained congruency in recharge temperature/elevation calculations (unlike the remaining samples). This follows then the observation that the lower the noble gas concentrations, the higher the recharge temperature and/or recharge elevation is required, but results in unrealistic determinations.

Table 2. Calculated Noble Gas Recharge temperatures and Elevations for Samples Measured in Amargosa River Valley

Sample	Calc Recharge T°C	Calc Elev ft	Comment	
ARHS-1	24	2200	Ar-Kr agreement; Xe too low	
Borehole Spr	19	6400	Ar-Kr agreement	mean max air Temp = 17°C
	19.7	6300	Ar-Xe agreement	
Tecopa Hot Spr	22	6300	Ar-Xe agreement	
	26	3300	Kr-Xe agreement	
Wild Bath Spr	22	4100	Ar-Xe agreement	
	26	1900	Ar-Kr agreement	
Thom Spr	23.5	800	Ar-Kr agreement; Xe too low	

Implications of measured $^3\text{He}/^4\text{He}$ ratios

Helium-4 can accumulate in groundwater by two main mechanisms. One is the uranium and thorium naturally in aquifer rock radioactively decays by ejecting an alpha particle, which is simply a charged ^4He atom. This is a slow process and commonly averages around $5e^{-11}$ ccSTP/g-yr. After a few thousand years an appreciable dissolved ^4He buildup can occur in groundwater resulting in R/R_a between 0.2 and 1. Comparison between radiocarbon ages of groundwater and the R/R_a commonly results in good correlations (Ballentine et al., 2002).

The second mechanism is by more rapid accumulation when groundwater is in more direct contact with the Earth's crust. This occurs because the crust has a much higher abundance of uranium and thorium and hence a higher production of ^4He . If a groundwater aquifer is very deep or if it is faulted down to the crust, diffusive transport of ^4He from the crust to the overlying groundwater accelerates the accumulation. This can result in very low R/R_a values (<0.2) developing in groundwater in just a few thousand years.

The R/R_a values in Table 1 for the groundwater measured in the Amargosa River Valley show very low values as might be seen for groundwater influenced by a crustal source accumulation. The exception is for ARHS-1, which has a typical value for a groundwater a few thousand years old. As a comparison, Table 3 shows the $^3\text{He}/^4\text{He}$ ratios and R/R_a values for a number of background groundwaters collected from wells on the Nevada Test Site. Note that these R/R_a values are significantly higher than those in Table 1 and more consistent with ARHS-1. The low R/R_a values in the Amargosa River Valley samples suggest that there may be deep faults beneath this area that facilitates more rapid transport of ^4He from crustal sources. Otherwise, if the low R/R_a values were due only to ^4He accumulation from local aquifer rock, then the implication would be that this groundwater had a subsurface age over 100,000 years old and is essentially fossil water.

Table 3. $^3\text{He}/^4\text{He}$ ratios in groundwater of the Nevada Test Site

Well Name	Area	$^3\text{He}/^4\text{He}$ ratio	R_a
ER-6-1	6	8.65E-07	0.63
ER-3-1	3	1.90E-06	1.37
ER-30-1	30	1.12E-06	0.81
ER-30-2	30	1.08E-06	0.78
ER-30-1	30	9.77E-07	0.71
ER-30-2	30	9.79E-07	0.71
UE1q	1	8.80E-07	0.64
ER-12-1	12	1.51E-06	1.09
UE10j-1	8	7.72E-07	0.56
UE10j-2	8	7.63E-07	0.55
UE10j-3	8	7.57E-07	0.55
WW-20	20	4.74E-07	0.34
ER-20-3	20	9.27E-07	0.67

from Rose et al. (1998)

Discussion of Groundwater Recharge and Ages

In desert environments groundwater recharge can be limited to areas of appreciable accumulation of annual precipitation. Recharge can be diffuse and slow where precipitation occurs as rain, but in areas of snow accumulation recharge can be pulsed

and rapid. Only in the cases of the Spring Mts. and the Kingston Range does annual precipitation accumulate as snow, but restricted to higher elevation. Additional accumulation can occur in recharge areas supporting Ash Meadows flow. It has been shown with stable isotope results above that groundwater recharge beneath the Amargosa River Valley is limited to either Spring Mts. or a mixture of Ash Meadows, Spring Mts., and the Kingston Range. Consequently, the replenishment of groundwater is limited and sensitive to upgradient changes such as climate change or accelerated groundwater pumping.

The stable isotope results of groundwater in the Amargosa River Valley require at least in part a southern recharge source in order to differentiate it from stable isotope abundances of Ash Meadows and other sources to the north. Stable isotope results discussed earlier support possible flowpaths for groundwater from Pahrump Valley that moves toward Ash Meadows. Ash Meadows groundwater subsequently migrates southward following a path expressed on the surface by the Amargosa River channel. ARHS-1 stable isotope values are consistent with this recharge source. Ash Meadows groundwater ultimately is derived from the regional carbonate to the northeast with potential contributions from the north in the Nevada Test Site area through Oasis Valley and Jackass Flats.

Likewise data supports groundwater moving from Pahrump Valley towards the Chicago Valley beneath the northern end of the Nopah Range and/or toward the south into the California Valley. Groundwater along these paths would subsequently flow into the Tecopa basin and join flow from the north. The stable isotope values of groundwater sampled in the Tecopa area are consistent with this mixed groundwater. However, it should be kept in mind that mixing in groundwater systems is not an efficient process. This is particularly true in porous alluvial type material. In fracture flow system mixing is more efficient and a plausible mechanism for creating mixed stable isotope signatures for the Tecopa groundwater. Fracture flow consequently would need to occur at depths below the Tecopa lake bed deposits. Lastly, previous reports have shown recharge from runoff from the Kingston Range could likewise contribute recharge to the Tecopa groundwater.

Determining the age of groundwater can benefit groundwater management because the time required to naturally replenish a groundwater aquifer provides a direct measure of the rate at which it can be safely used. Groundwater ages are commonly measured by either measuring tritium for groundwater younger than approximately 50 years or with radiocarbon for groundwater ages thousands of years old. Tritium measured in groundwater reported in a previous study of the Amargosa River Valley showed little evidence of young groundwater (Hurst, 2012). However, no radiocarbon has been measured in these groundwaters to date, and any indication of groundwater ages a few thousand to tens of thousands years old is lacking. Radiocarbon has been measured in other studies for groundwater in the regional carbonate aquifer of southern Nevada (Thomas et al., 1996, Davisson et al., 1999) and beneath the Nevada Test Site (Rose et al., 1998). In these cases, radiocarbon was measureable and suggested ages from a few thousand to tens of thousands years old.

The results from the $^3\text{He}/^4\text{He}$ measurements for the Amargosa River Valley groundwater suggest that groundwater is likely tens of thousands of years old. This is best illustrated for the R/R_a value for sample ARHS-1 (0.243), which is somewhat lower, but similar to those measured for groundwater beneath the Nevada Test Site. Since radiocarbon was measureable for these latter groundwaters, it stands to reason that ARHS-1 groundwater would have measureable radiocarbon.

In the case of the remaining groundwater samples measured in the Amargosa River Valley, the unusually low R/R_a values suggest that ^4He may accumulate rapidly in this area due to presence of deep faults. In the absence of this mechanism, the implication of low R/R_a values would be that groundwater in the Tecopa area is isolated from ARHS-1 groundwater and replenishes at an extremely low rate. Using an in-situ ^4He accumulation rate of $5e^{-11}$ ccSTP/g-yr would imply groundwater ages greater than 100,000 years and possibly up to 1,000,000 years. These old ages are rare in groundwater system, particularly those with active spring discharge as seen in this area.

CONCLUSIONS AND RECOMMENDATIONS

- Although argument has been made that most Amargosa River Valley groundwater could be recharged from the Spring Mts., it is more consistent with available data that Amargosa River Valley groundwater is a mixture of Ash Meadows and Spring Mts./Kingston Range sources. Fracture flow would likely be required to induce mixing among these sources since alluvial type deposits would not be an efficient mechanism. Borax Spring is an exception and is likely the same recharge source as Ash Meadows.
- Groundwater in the Tecopa hot springs area is derived from the same groundwater source as that for Shoshone Spring to the north and Thom Spring to the south. The elevated temperature of the Tecopa spring water is not unusual since similar temperatures are seen at depth under the Nevada Test Site. However, the warm groundwater has been driven to the surface in Tecopa area probably by some geologic structural control.
- Water quality in the Amargosa River Valley groundwater likely evolves from a mixture of regional carbonate and Tertiary volcanic rock influences, but acquires increased chloride and sulfate possibly from the Tecopa lake bed deposits. Additionally, increases in regional subsurface heat flow increases groundwater temperature and contributes to increased dissolved silica, decreased bicarbonate, and possibly increased pH, with the latter resulting in the high arsenic concentrations.
- Noble gas concentrations of Amargosa River Valley groundwater have striking similarity to those measured in the regional carbonate-Ash Meadows of southern

Nevada groundwater by Thomas et al. (2003b). Their conclusions were that dissolved gas loss occurred during subsurface transport across faulted boundaries and compromised recharge temperature/elevation calculations.

- The noble gas recharge temperatures/elevation calculations for Amargosa River Valley groundwater mostly support the conclusions of Thomas et al. (2003b). The best model fit resulted for Borehole Spring with a recharge temperature of 19°C and elevation of 6400 feet derived from the Ar, Kr, and Xe concentrations. The noble gas concentrations of remainder of the samples could not converge on a single recharge temperature and elevation due to gas losses from subsurface processes. However, a 19°C recharge temperature at 6400 feet is *inconsistent* with plausible recharge areas and suggests dissolved noble gas in Borehole Spring likewise been compromised.
- The $^3\text{He}/^4\text{He}$ ratios for all measured springs in Amargosa River Valley are unusually low, indicating old groundwater ages. The Amargosa River Valley ratios are around 5-10 times lower than measured for groundwater under the Nevada Test Site.
- The low $^3\text{He}/^4\text{He}$ ratios in the Amargosa River Valley groundwater could be due to high ^4He flux from the earth's crust caused by deep faults. Otherwise, if the low ratio is due to steady-state accumulation from local deposits, then groundwater ages greater than 100,000 years would be required.
- Groundwater management in the Amargosa River Valley requires better understanding of recharge sources, recharge rates, and flow paths. To further this understanding it is recommended as follow-up work to pursue the following water quality and isotope measurements:
 - Investigate the solubility of salts in the Tecopa lake bed deposits. This would provide comparative data to the high TDS spring water in the area to determine if this is a viable mechanism to explain the spring water quality compositions.
 - Measure additional samples for $\delta\text{D}-\delta^{18}\text{O}$ to better resolve zones of recharge and direction of groundwater flow. In particular, samples from the Chicago and Pahrump Valleys, as well as additional samples from the Kingston Range area would facilitate this effort.
 - Measure the radiocarbon abundance of spring water in the Amargosa River Valley with the lowest $^3\text{He}/^4\text{He}$ ratios. If radiocarbon is present, then this would indicate that the low ratios are due to high ^4He flux along faults from the deep crust.
 - Measure additional $^3\text{He}/^4\text{He}$ ratios in groundwater collected between Ash Meadows and Tecopa area. This would provide a continuum of $^3\text{He}/^4\text{He}$ ratios with downgradient distance and would facilitate the development of a groundwater age model.

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Appendix 1. Use of Isotope Measurements in Groundwater

A.1 Stable Isotopes

The stable isotope measurements of oxygen-18/oxygen-16 ($^{18}\text{O}/^{16}\text{O}$) and deuterium/hydrogen (D/H; deuterium is hydrogen-2) ratios in water are used to delineate different water populations in recharged groundwater. Their value lies in the fact that these isotope abundances significantly vary with changes in temperature, inland distance, and elevation. For example, the average ^{18}O abundance in precipitation is shown for North America in Figure A.1. Note that these abundances are highest along coastal regions and rapidly decrease where mountain belts occur as well as with increasing latitude. These differences are typically reflected at the local and regional scale relevant at scales for groundwater basin studies (see further discussion below).

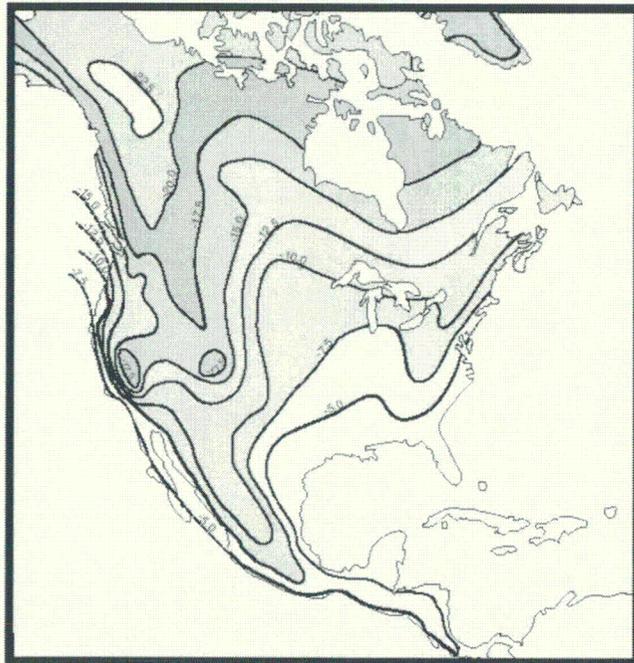


Figure A.1. Average $\delta^{18}\text{O}$ abundances of precipitation across North America show highest values along coastal regions and decrease rapidly with elevation in mountain belts and with increasing latitude.

The measured $^{18}\text{O}/^{16}\text{O}$ and D/H ratios are normalized to a recognized standard and the converted results are reported in δ notation (pronounced "del"), where

$$\delta D = \left(\frac{D/H}{D/H_{std}} - 1 \right) 1000$$

$$\delta^{18}\text{O} = \left(\frac{^{18}\text{O}/^{16}\text{O}}{^{18}\text{O}/^{16}\text{O}_{std}} - 1 \right) 1000$$

The $^{18}\text{O}/^{16}\text{O}_{std}$ and D/H_{std} are the isotopic ratios of "Standard Mean Ocean Water" (SMOW). A δ value is a per mil (or parts per thousand) deviation from the standard.

The atomic masses differences of these different isotopes in water molecules underlie differences in measured ratios. These differences arise from phase transitions in water (i.e., vapor, water, ice) which favors higher atomic masses in lower energy states of matter. For example, the measured difference in the $\delta^{18}\text{O}$ value measured between a water vapor and its condensed liquid form at 25°C is approximately 9.3 per mil. This difference is large compared to the typical measurement precision of 0.1 per mil.

The isotopic ratios of ocean water are remarkably uniform worldwide, owing to global circulation patterns. However, since all continental precipitation originates from the ocean, isotopic partitioning occurs between water phases, and because continental storm fronts are isolated from the ocean and behave as closed systems, the isotopic ratios of measured precipitation varies systematically. This variation is almost exclusively driven by elevation difference and distance inland from the ocean. An example of $\delta^{18}\text{O}$ variations in precipitation across British Columbia are illustrated below in Figure A.2a. Figure A.2b shows how shallow groundwater collected on the western slope of the Sierra Nevada record this systematic $\delta^{18}\text{O}$ variation in its recharge.

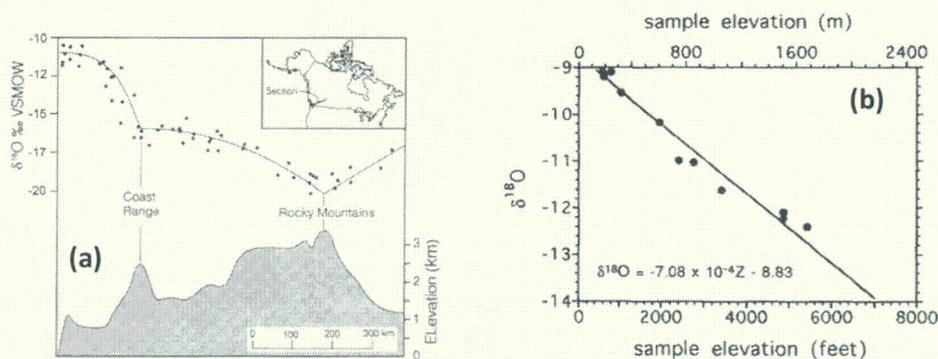


Figure A.2a,b. Figure A-2a shows systematic variation of $\delta^{18}\text{O}$ values in precipitation across British Columbia (from Yonge et al., 1989). Figure A.2b shows how shallow groundwater records this systematic variation on the western slope of the Sierra Nevada (from Rose et al., 1996).

The method for comparing the isotopic character of different waters lies in the use of a δD - $\delta^{18}\text{O}$ plot of the isotope ratios. A plot of δD vs. $\delta^{18}\text{O}$ values provides a graphical means to distinguish various populations of data relating to different water masses of different origins (Fig. A-3).

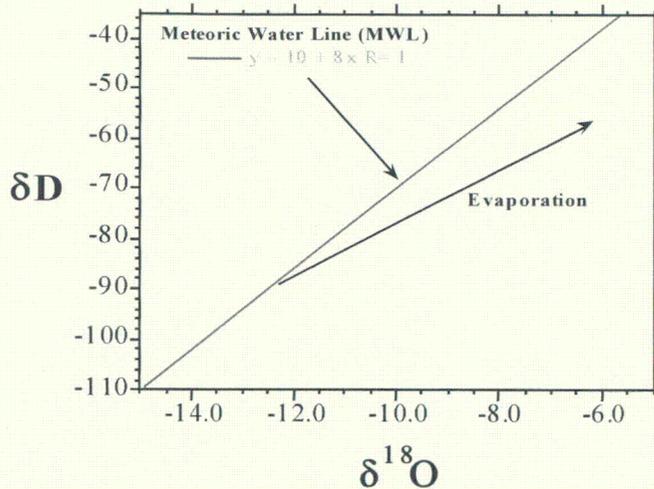


Figure A-3. General δD - $\delta^{18}O$ plot showing the Meteoric Water Line (MWL) and the effects of evaporation on natural waters. The slope of the evaporation line can vary between 2 and 6 and depends on the ambient temperature and humidity. The MWL has a constant slope of 8 for global precipitation.

Also on this plot lies what is referred to as the Global Meteoric Water Line (MWL), a linear regression through the values of various unevaporated precipitation collected world-wide, which results in an empirical equation of $\delta D = 8 \delta^{18}O + 10$. The slope of this line originates from the fact that isotopic partitioning of deuterium between water vapor and liquid is approximately 8 times greater than for ^{18}O . Since global precipitation forms a slope of 8 indicates that cloud water establishes isotopic equilibrium between vapor and liquid.

However, when liquid water evaporates from the surface of water body, a non-equilibrium partitioning develops between the relative deuterium and ^{18}O abundances, causing isotopic enrichment of the remaining liquid water. On a δD - $\delta^{18}O$ plot, progressive evaporation causes a shift of the remaining liquid to the right of the MWL along a straight line (see Fig. A-3). The slope of this evaporation line depends on temperature and humidity of the surrounding air. The proximity of an evaporated isotopic value relative to the MWL is proportional to the extent of evaporation or isotopic enrichment.

A.2 Tritium-Helium-3 Age Dating

Attempts have been made in the past to date groundwater with the radioactive (unstable) hydrogen-3 isotope tritium (3H ; see Mazor, 1991 and references therein). Because of its radioactive half-life of 12.43 years, it is ideally a good chronometer for young (≤ 40 years) groundwater flow. Unfortunately from a dating standpoint, 3H concentrations in precipitation have varied considerably over the past 50 years due to 3H production from surface testing of thermonuclear weapons (Fig. A-4).

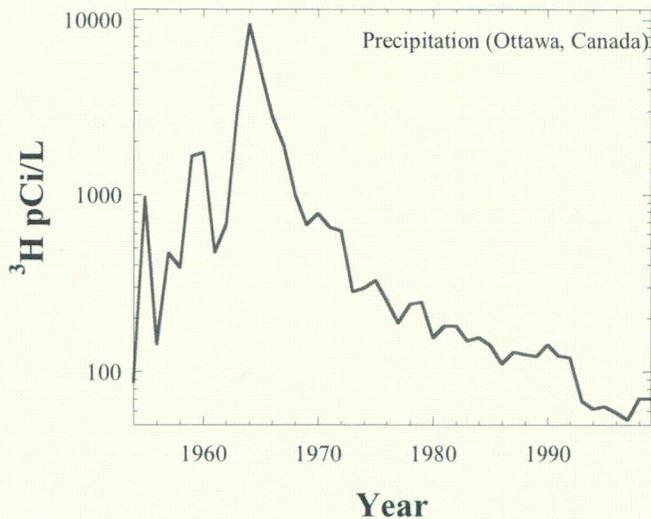


Figure A-4. Changes in the ^3H concentration in precipitation have varied over an order of magnitude due to fallout of thermonuclear-produced tritium from surface testing. IAEA/WMO (2001). Global Network of Isotopes in Precipitation. The GNIP Database. Accessible at: <http://isohis.iaea.org>

Tritium measurements in groundwater 30-40 years ago were useful from the standpoint of tracing the "bomb-pulse" ^3H that had recharged into groundwater in the early 1960s and calculating the groundwater travel time based on the observed depth of the "bomb pulse". Today, however, much of the "bomb-pulse" is not well defined in groundwater due to ^3H decay and groundwater dispersion. Tritium measurements alone cannot be used for dating groundwater reliably because of the uncertainty in what the original ^3H concentration was at the time of recharge, but it does serve the purpose of defining relatively young groundwater when it is observed.

In more recent years with the development of high-precision noble gas mass spectrometry, the radioactive decay product of ^3H , helium-3 (^3He), can be measured. The advantage to this lies in the dating equation, where

$$-17.9 \times \ln \left(\frac{^3\text{H}}{^3\text{H}_0} \right) = \text{age} ,$$

^3H is the concentration of the tritium at any given time, and $^3\text{H}_0$ is the original tritium concentration at the time of recharge. Since the $^3\text{H}_0$ has a large uncertainty due to the spatially and temporally variable "bomb pulse" tritium, the resulting age calculation will have large uncertainties. By simultaneously measuring the ^3He produced by tritium decay (known as the tritiogenic ^3He or $^3\text{He}_{\text{trit}}$) we can reconstruct the $^3\text{H}_0$ by adding together the measured tritiogenic $^3\text{He}_{\text{trit}}$ and the ^3H which leads to

$$-17.9 \times \ln \left(\frac{{}^3H}{{}^3H + {}^3H_{trit}} \right) = \text{age.}$$

Dissolved ${}^3\text{He}$ measured in a groundwater is actually derived from several sources that include:

$${}^3\text{He}_{meas} = {}^3\text{He}_{trit} + {}^3\text{He}_{equil} + {}^3\text{He}_{excess} + {}^3\text{He}_{rad},$$

where ${}^3\text{He}_{meas}$ is the total ${}^3\text{He}$ analytically measured, ${}^3\text{He}_{equil}$ is the amount of ${}^3\text{He}$ dissolved in a non-turbulent surface water in equilibrium with the atmosphere and is temperature dependent, ${}^3\text{He}_{excess}$ is the amount of ${}^3\text{He}$ dissolved in water exceeding the equilibrium amount (a common phenomenon in groundwater due to excess dissolved air), and ${}^3\text{He}_{rad}$ is the amount of ${}^3\text{He}$ produced from radioactive decay of isotopes other than tritium. The latter species is very minor and totals only about 0.2% of the total ${}^3\text{He}$. Separating these different components of the ${}^3\text{He}$ requires additional measurements of the ${}^4\text{He}$ abundance which comprise:

$${}^4\text{He}_{meas} = {}^4\text{He}_{equil} + {}^4\text{He}_{excess} + {}^4\text{He}_{rad},$$

where the subscripts are the same as those for ${}^3\text{He}$. In the case of ${}^4\text{He}_{rad}$, a product of uranium-thorium decay, the abundance can be significant where older waters are involved (e.g. >1000 years old) and has been used numerous times as an independent groundwater age measurement due to its steady state accumulation.

The ${}^3\text{He}_{equil}$, ${}^4\text{He}_{equil}$, and ${}^4\text{He}_{rad}$ terms are either assumed or determined by other noble gas abundance measurements (see below), while the ${}^3\text{He}_{rad}$ term is assumed. The two unknowns left are the excess air terms and the tritiogenic ${}^3\text{He}$, of which we can formulate two equations to solve for them.

The ${}^4\text{He}_{meas}/{}^4\text{He}_{equil}$ ratios provide a method for determining the excess air contribution to the sample, since a ratio >1.0 is created by incorporation of more dissolved helium than in equilibrium with the atmosphere, assuming an appreciable amount of ${}^4\text{He}$ has not accumulated from radioactive decay (see below). This assumption can be validated with additional noble gas measurements. If radiogenic ${}^4\text{He}$ is a concern, though, the ${}^3\text{He}/{}^4\text{He}$ ratios can be calculated and compared to ratios expected in water at equilibrium concentrations. This comparison is important since if there is any appreciable radiogenic ${}^4\text{He}$, then the ${}^3\text{He}/{}^4\text{He}$ ratio relative to equilibrium will be <1.0. This is due to the accumulation of ${}^4\text{He}$ from uranium-thorium decay. Where there are indications of radiogenic ${}^4\text{He}$ we can correct for it in the age calculations.

A.3 Noble Gas Abundance

The noble gases of helium, neon, argon, krypton, and xenon naturally occur at trace abundance in the atmosphere. They also dissolve in groundwater during recharge. Their concentration in groundwater is controlled by 1) equilibrium solubility and 2) incorporation of excess air. The solubility of the noble gases in non-turbulent, free-standing water is temperature dependent, with increasing solubility with decreasing

temperature. This temperature dependency is most pronounced in the argon, krypton, and xenon concentration (Fig. A-5).

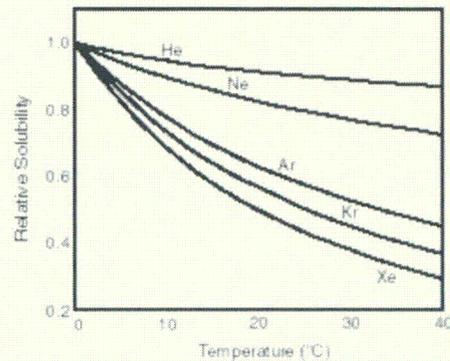


Figure A-5. Solubility of noble gases in water at various temperatures can be used to calculate groundwater recharge temperatures. See Mazor (1991) for examples and further discussion.

The curves in figure A-5 provide a means to calibrate measured dissolved noble gas abundances in groundwater against its recharge temperature. During most groundwater recharge, the mean soil temperature dictates the equilibrium noble gas concentrations dissolved in recharging water, which in most regions is around 2°C greater than the mean annual air temperature.

Dissolved noble gas abundances in groundwater other than helium that exceed an equilibrium amount are due to dissolution of excess air. Incorporation of excess air into recharged groundwater is thought to occur when air in the vadose zone is trapped by a plug of recharge water and is transported to deep enough depths that it is dissolved. Groundwater recharged through a vadose zone likely has excess dissolved air. In almost all cases the composition of the excess air is the same as the atmosphere (Heaton et al., 1981). Therefore, the amount of noble gases dissolved in groundwater above the equilibrium amount is a simple arithmetic addition of each noble gas from the atmosphere. Consequently, the amount of each dissolved noble gas relative to each other within a single sample should reflect a single equilibrium solubility temperature at the time of groundwater recharge. The amount of excess air dissolved in a groundwater can also provide qualitative information about the type of groundwater recharge. For instance, high excess air content may suggest recharge by a periodic "piston" flow under vadose zone conditions. Little excess air may suggest recharge with a limited vadose zone such as in river or lake infiltration.

The remaining noble gas effect that requires some consideration is the build-up of radiogenic ^4He . There is a constant flux toward the ground surface of ^4He derived from radioactive decay of uranium and thorium in the Earth's crust that, given enough time, can accumulate in groundwater. Typically groundwater that is thousands of years old will have an appreciable amount of radiogenic ^4He , while young groundwater (<100 years old) has little or none except in special conditions such as close proximity to large-scale active faults.

To test for the presence of radiogenic ^4He , the other noble gas abundances must be measured and calibrated to a recharge temperature. With this recharge temperature, the ^4He content can be predicted based on equilibrium solubility. Any ^4He that is above this predicted amount can be attributed to radiogenic ^4He , and subsequently subtracted. This will provide a revised $^3\text{He}/^4\text{He}$ ratio that can be used for calculating the groundwater age.

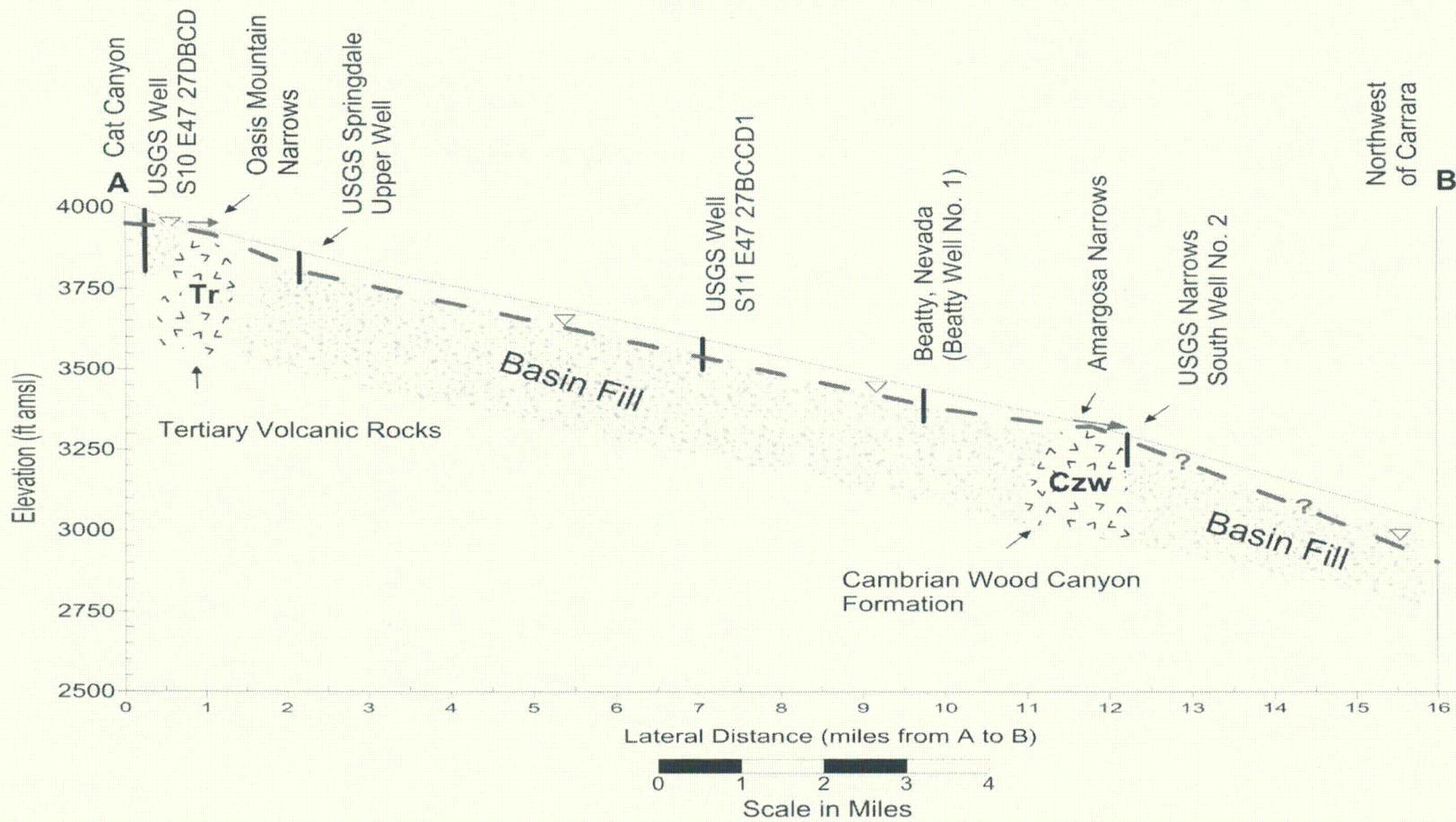
A.5 Radiocarbon and Carbon-13

Radiocarbon, or carbon-14 (^{14}C) is a radioactive isotope of carbon with a half-life of 5730 years. For decades ^{14}C has been used for age-dating of carbon-bearing materials (e.g. archeological artifacts) in the range of 100 to 50,000 years. Groundwater has also been dated, and most commonly by the ^{14}C abundance in dissolved inorganic carbon (DIC). Although many successful studies have been conducted using DIC ^{14}C measurements, much debate still continues about how and to what the extent carbonate minerals in aquifer systems dilute ^{14}C in recharging groundwater. As a result, absolute age determinations of groundwater using ^{14}C are limited to special cases where the absence of carbonate can be demonstrated or ^{14}C correction models can be validated. For the most part, absolute ages ≤ 1000 years old are usually highly uncertain.

The stable isotope of carbon, carbon-13 (^{13}C), is often measured in DIC and can provide either a source indicator or a relative measure of carbonate mineral reaction. Groundwater acquires DIC during recharge through plant root zones. The partial pressure of CO_2 in the soil root zone is usually higher (i.e. factor of 2 to 1000) than the atmosphere. Recharging groundwater will dissolve this soil zone CO_2 , which is chemically neutralized by dissolution of minerals. Soil carbonate is the most common mineral interaction, but in its absence, aluminosilicates can also serve as a reactive substrate. Atmospheric CO_2 has a $\delta^{13}\text{C}$ value of approximately -7.5 per mil (the δ system is the same as used for ^{18}O and deuterium, but carbon isotope ratios are compared to a reference carbonate material instead). Higher plants growing on the surface use this CO_2 for photosynthesis and in the process preferentially use ^{12}C over ^{13}C . As a result, plant $\delta^{13}\text{C}$ values tend to either be around -28 per mil, or for many grasses around -13 per mil. These same $\delta^{13}\text{C}$ values will occur in the soil zone CO_2 which originates from plant roots. Consequently, the $\delta^{13}\text{C}$ of DIC in recharging groundwater will be a mixture of the root zone CO_2 and any carbonate mineral it reacts with. To complicate matters further, for root zones where the partial pressure of CO_2 can be 10 times greater than the atmosphere, and recharging groundwater is relatively slow, isotopic exchange can occur between the DIC and the atmospheric CO_2 , causing an enrichment in the $\delta^{13}\text{C}$ DIC value (partitioning between DIC and CO_2 is approximately 8-10 per mil, depending on temperature). This latter complication is common to desert environments. With all these variables in the recharging groundwater, predicting the final DIC ^{14}C and $\delta^{13}\text{C}$ values of groundwater reaching the saturated zone creates many uncertainties. As a result, it is more common to take an empirical approach and compare populations of $\delta^{13}\text{C}$ values of groundwater DIC collected in the same general vicinity, and estimate the amount of carbonate interaction and the recharge dynamics.

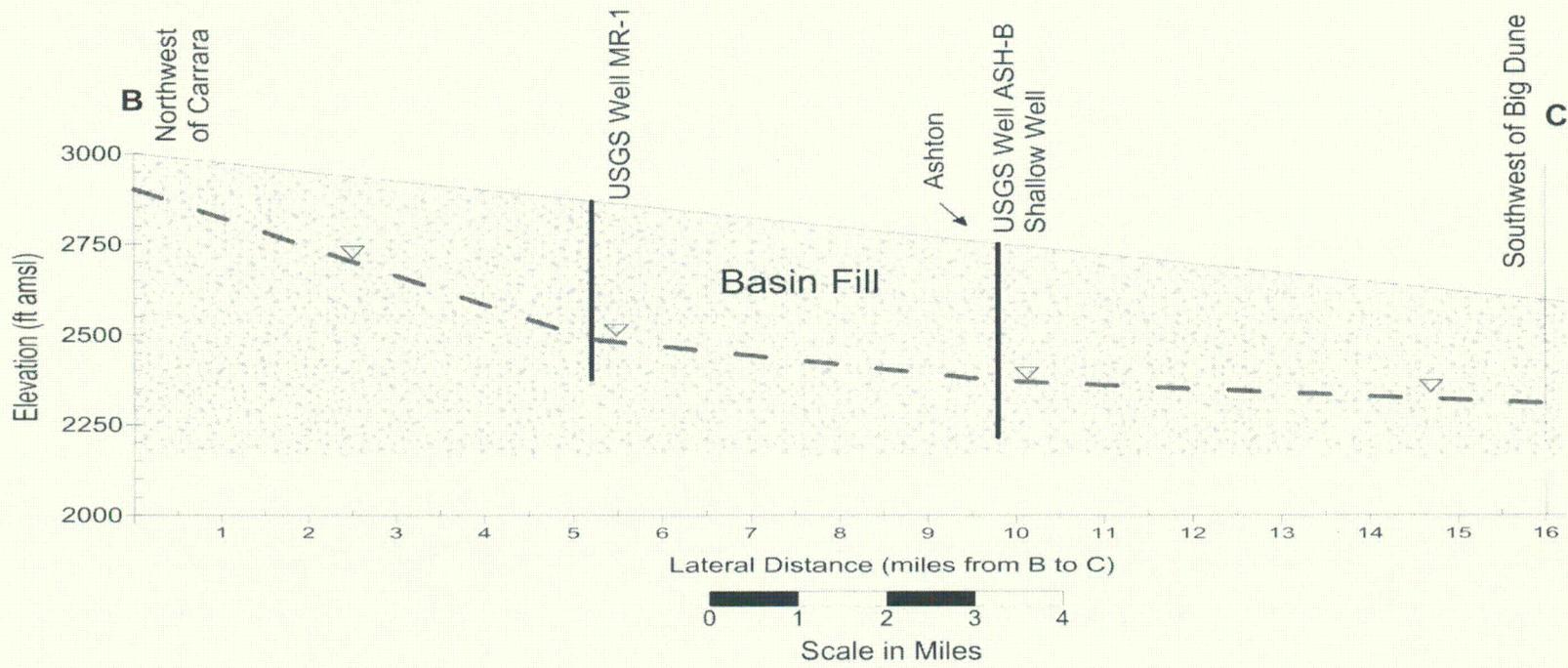
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Conceptual Cross Section A-B: Amargosa River Course
Southwest of Oasis Mountain to Northwest of Carrara, NV



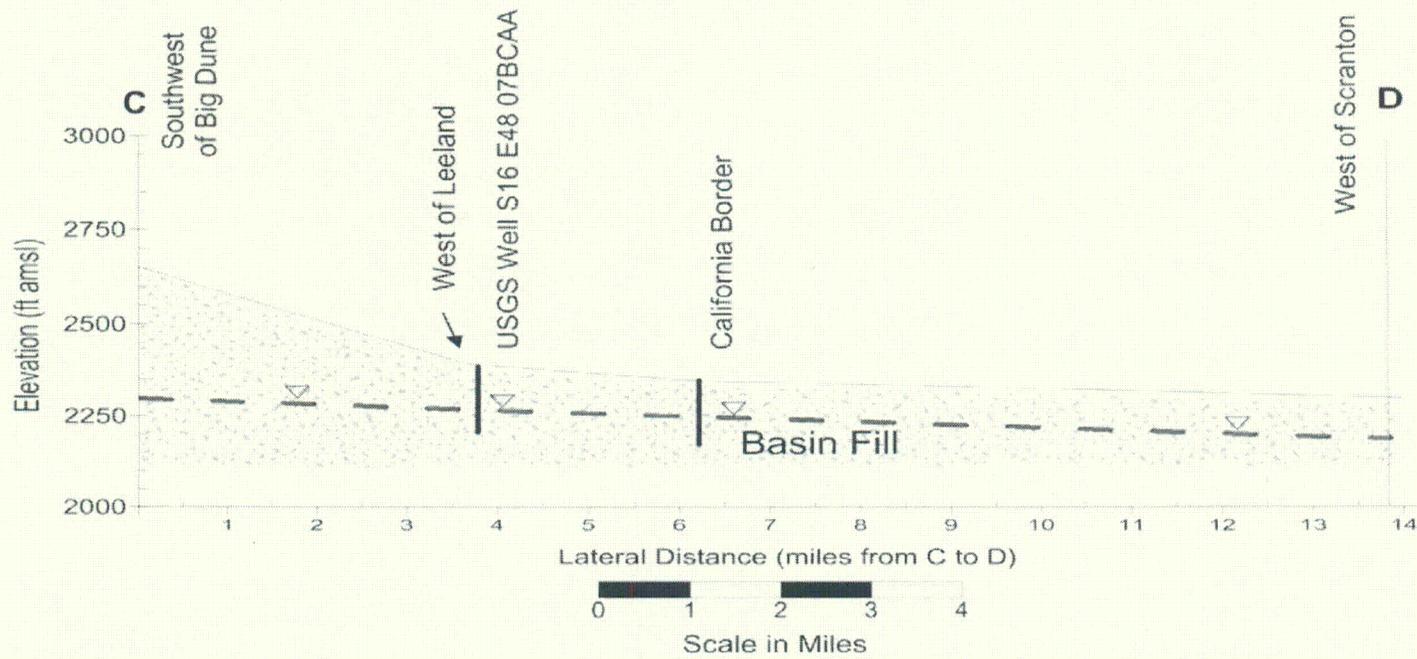


Conceptual Cross Section B-C: Amargosa River Course

Northwest of Carrera to Southwest of Big Dune, NV

ANDY ZDON &
ASSOCIATES, INC.



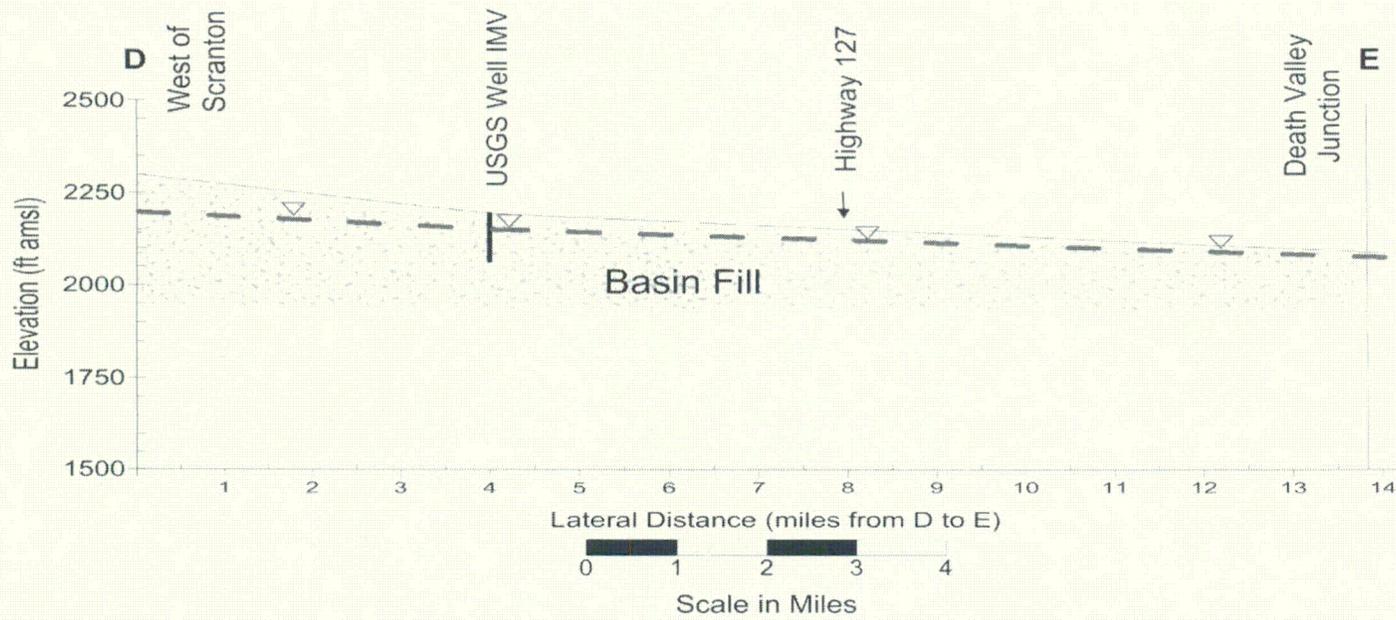


Conceptual Cross Section C-D: Amargosa River Course

Southwest of Big Dune, NV to West of Scranton, CA

ANDY ZDON &
ASSOCIATES, INC.



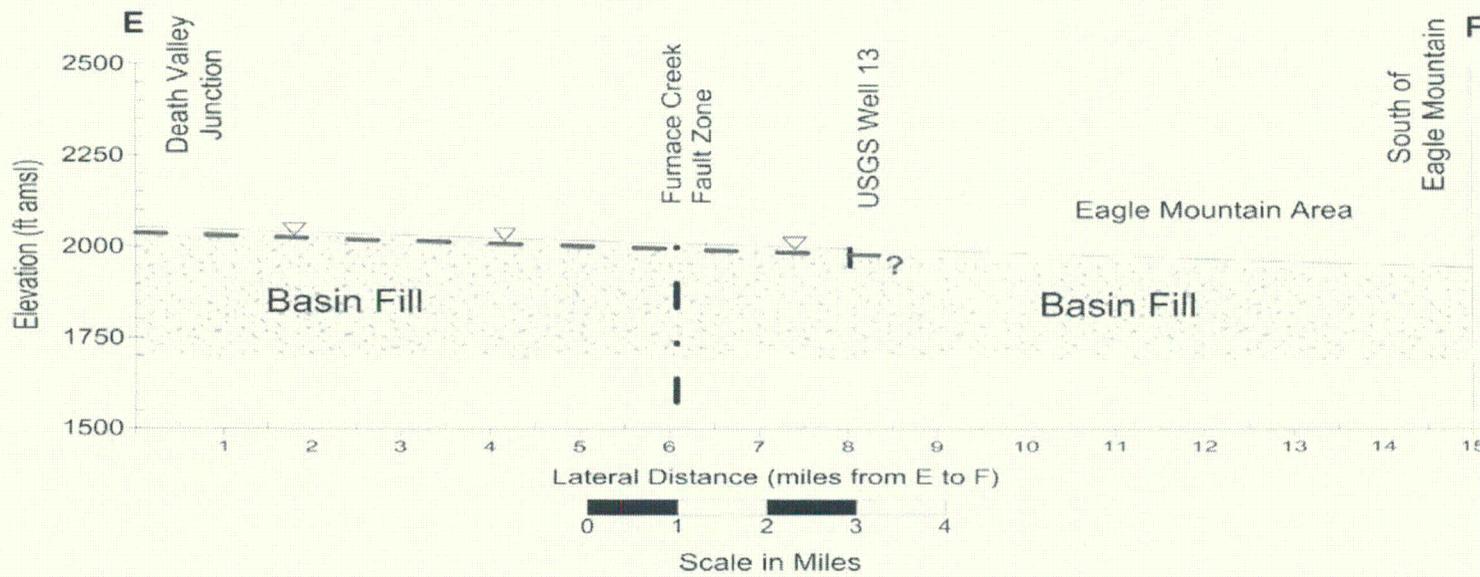


Conceptual Cross Section D-E: Amargosa River Course

West of Scranton to Death Valley Junction, CA

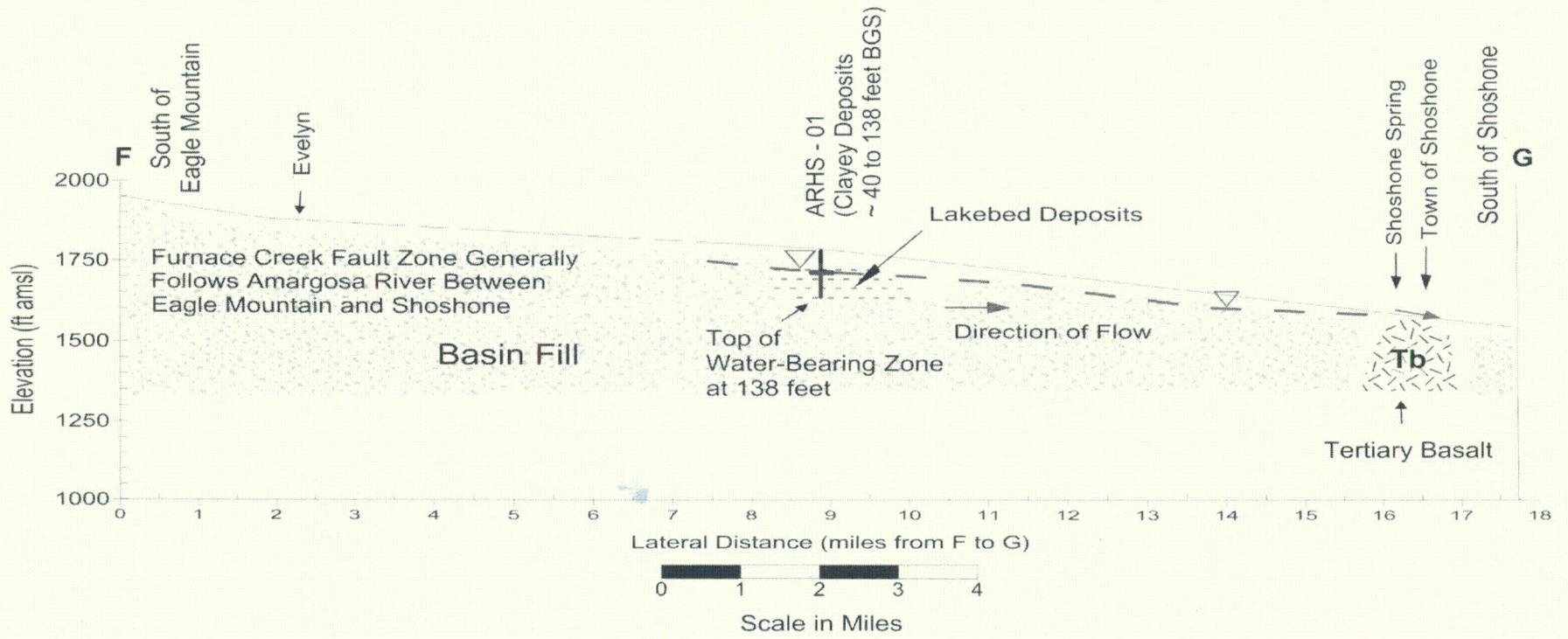
ANDY ZDON &
ASSOCIATES, INC.

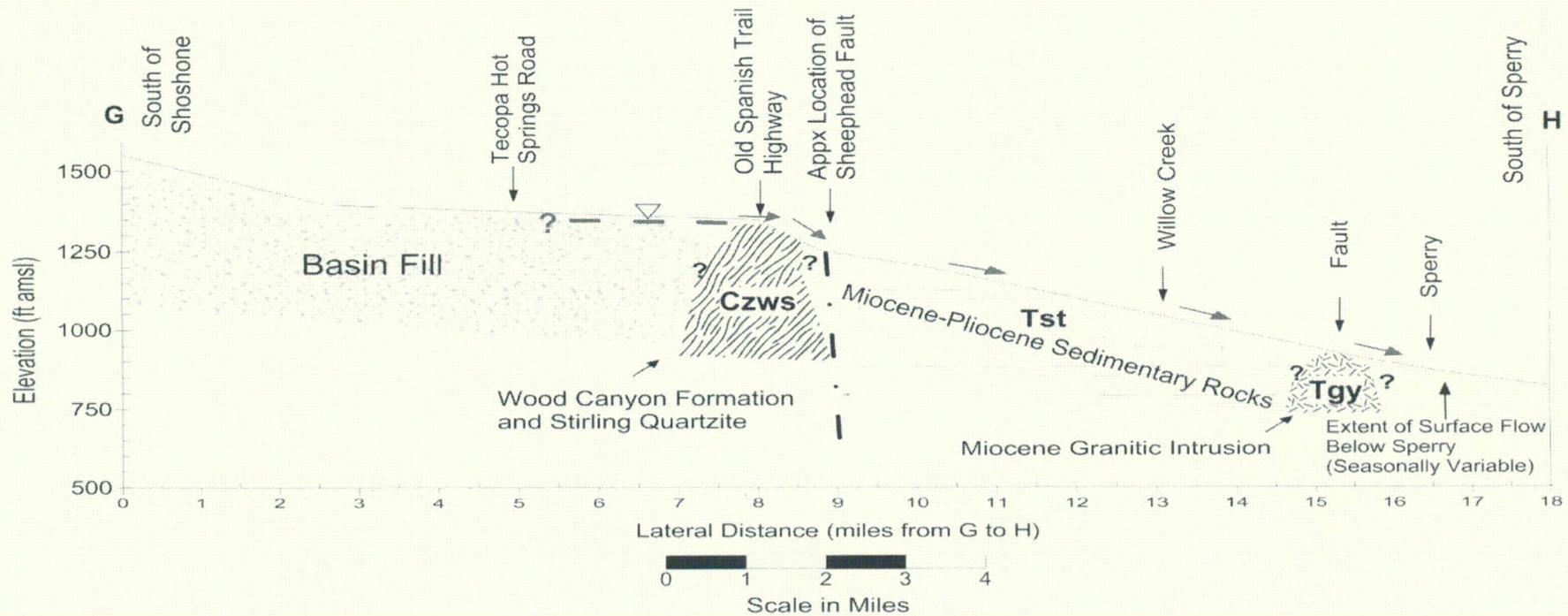


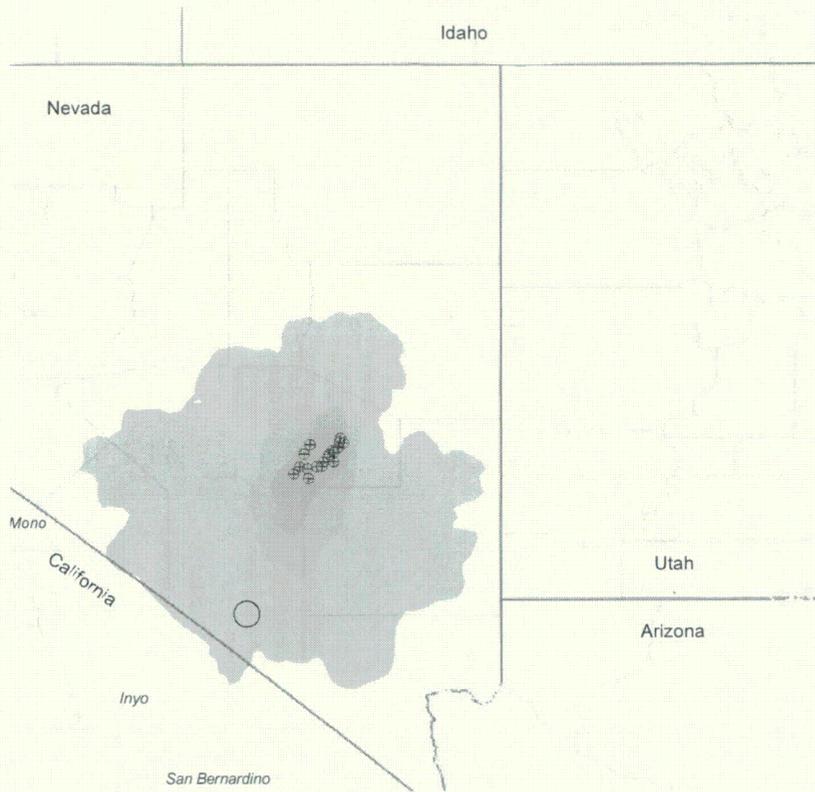


Conceptual Cross Section E-F: Amargosa River Course
 Death Valley Junction to South of Eagle Mountain, CA

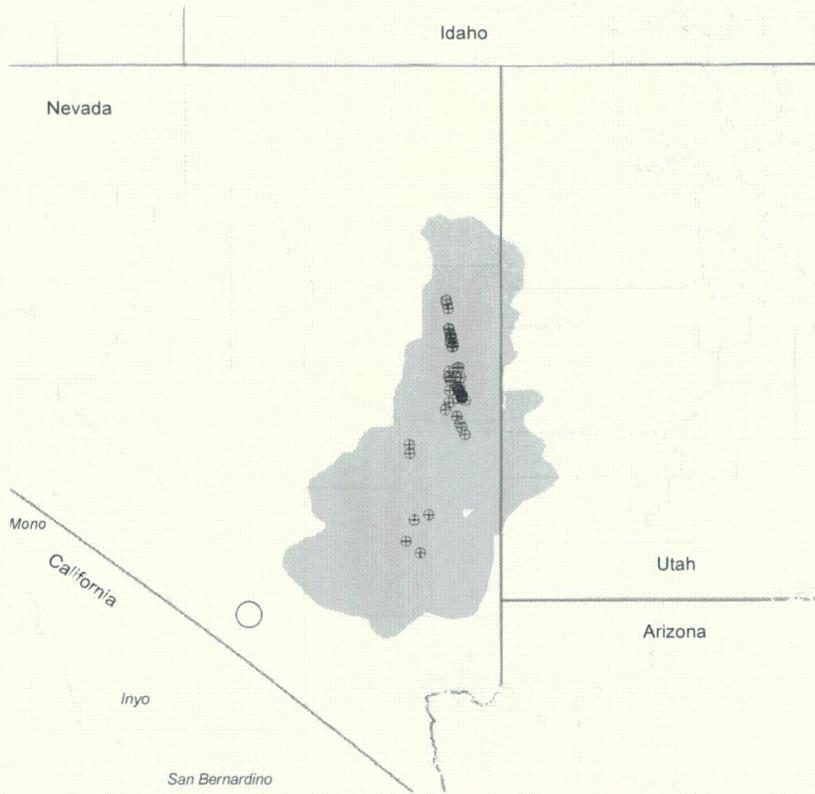




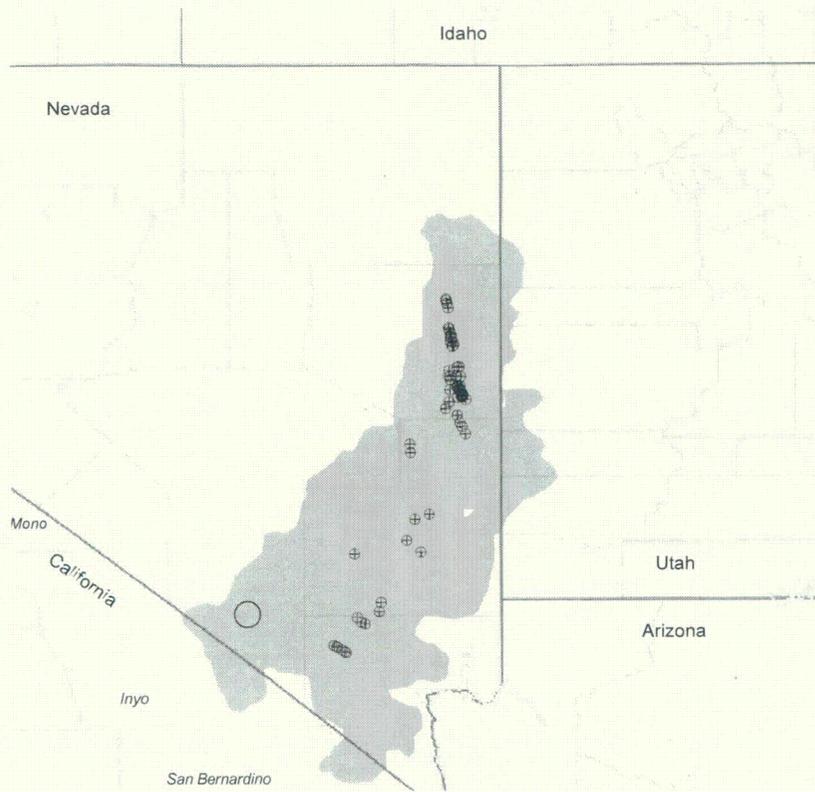




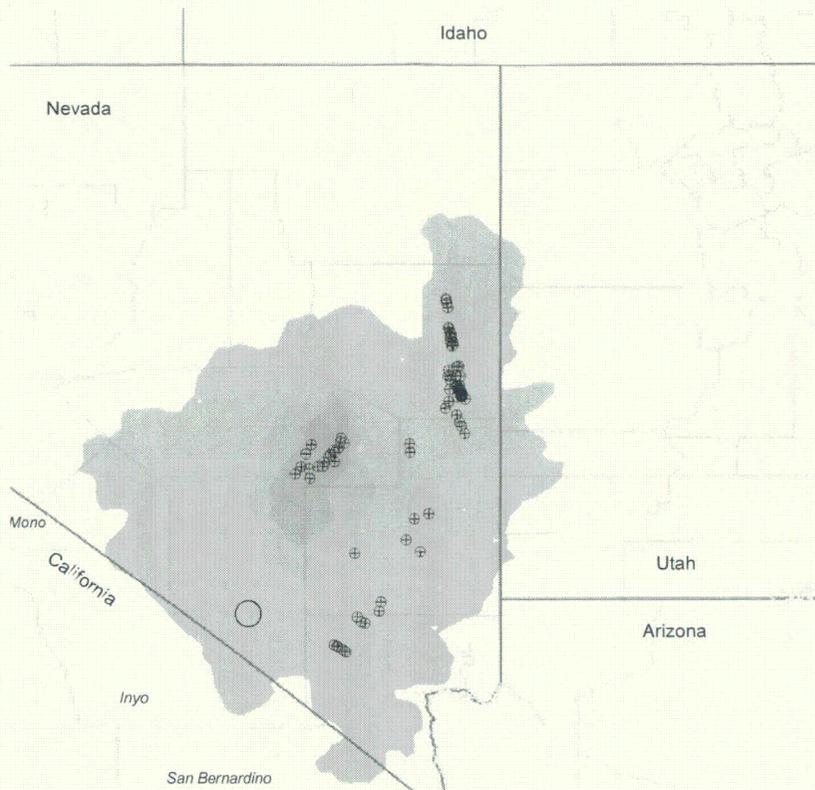
Simulated drawdown (ft) in Model Layer 1 for Pumping Scenario 1 (Railroad Valley), limited pumping case. Location of Yucca Mountain indicated by black circle symbol.



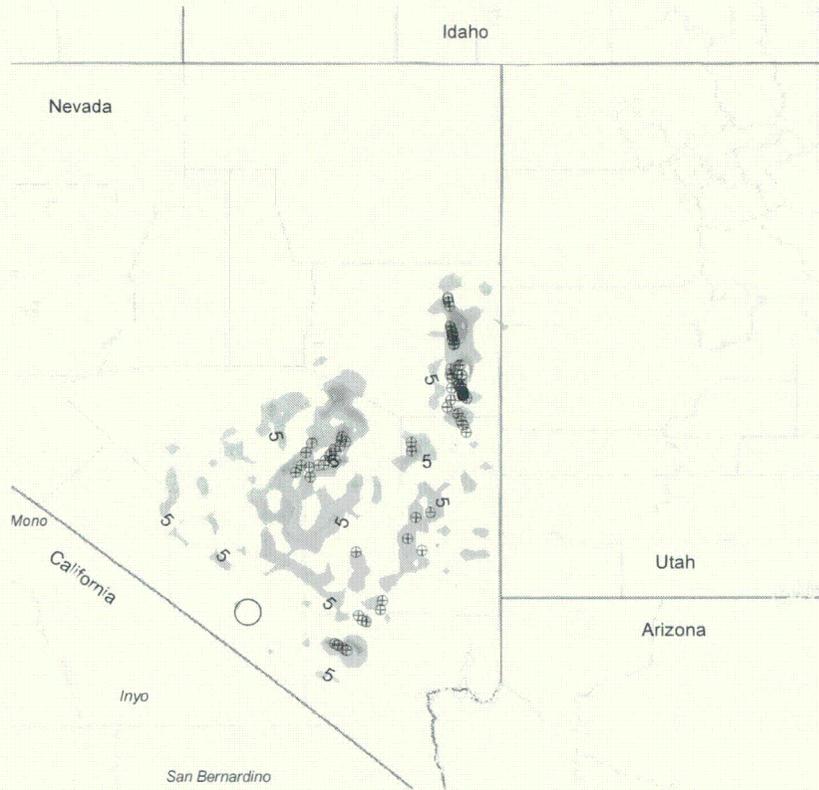
Simulated drawdown (ft) in Model Layer 1 for Pumping Scenario 2 (Spring Valley, Dry Lake Valley, Cave Valley and Delamar Valley), limited pumping case. Location of Yucca Mountain indicated by black circle symbol.



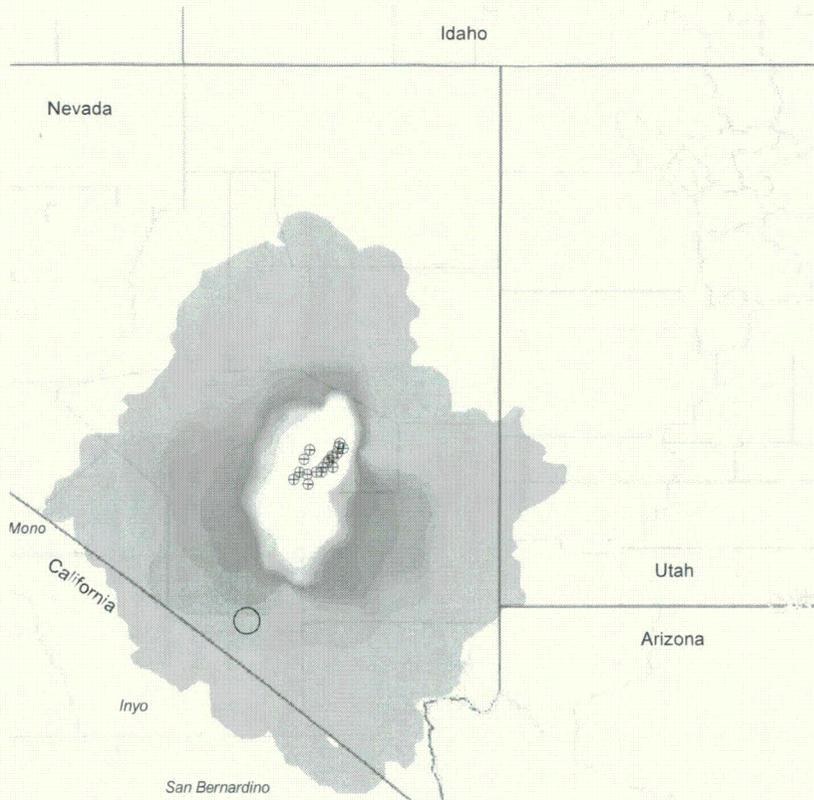
Simulated drawdown (ft) in Model Layer 1 for Pumping Scenario 3 (Spring Valley, Dry Lake Valley, Cave Valley, Delamar Valley, Tikapoo Valley, and Three Lakes Valley), limited pumping case. Location of Yucca Mountain indicated by black circle symbol.



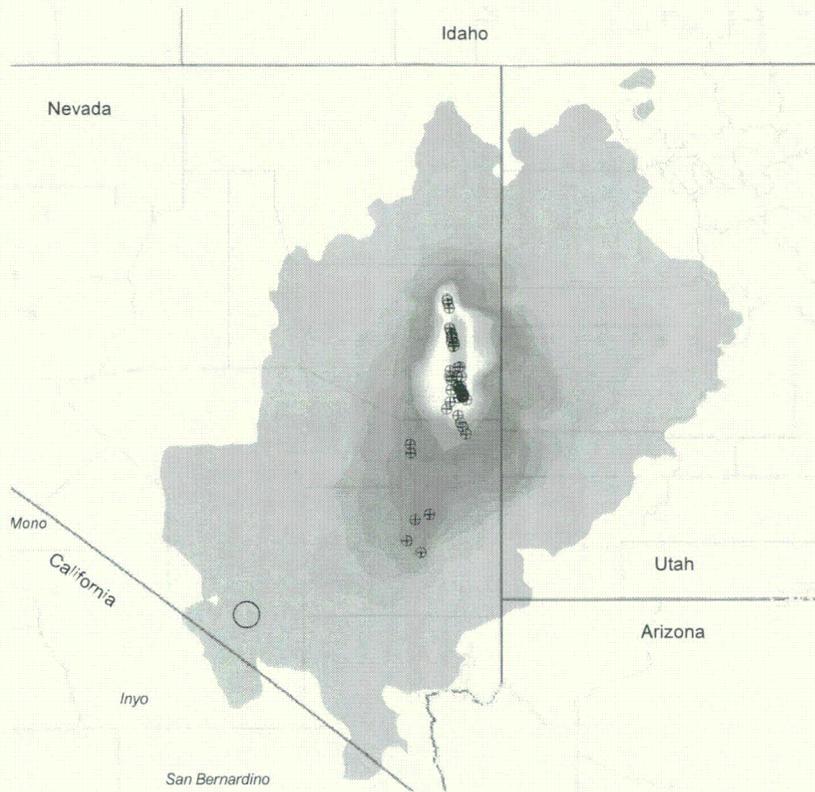
Simulated drawdown (ft) in Model Layer 1 for Pumping Scenario 4 (Spring Valley, Dry Lake Valley, Cave Valley, Delamar Valley, Tikapoo Valley, Three Lakes Valley, and Railroad Valley), limited pumping case. Location of Yucca Mountain indicated by black circle symbol.



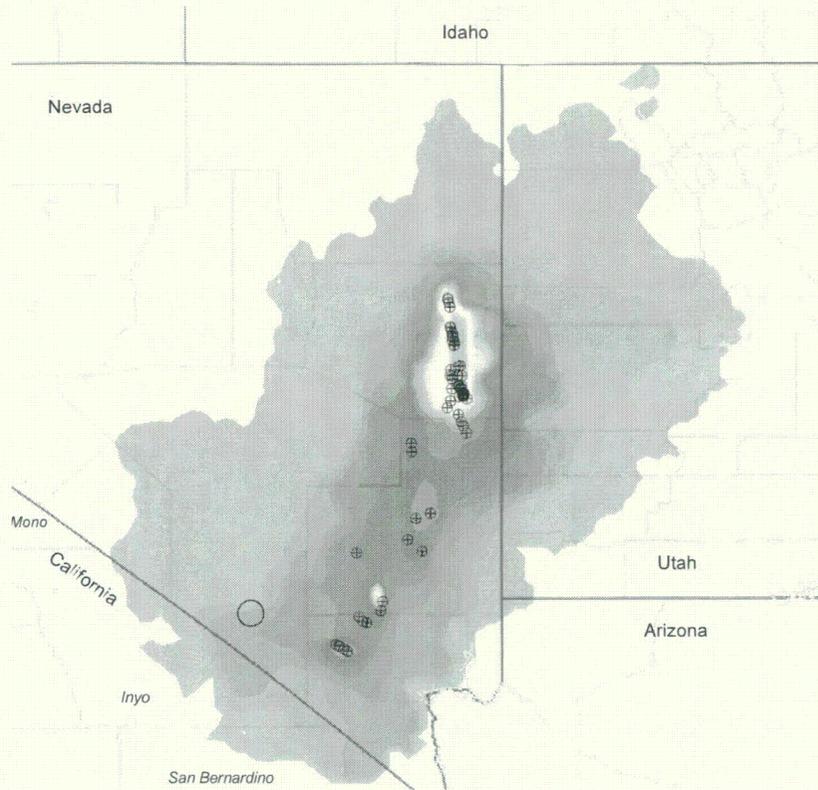
Vertical head difference change between Model Layers 1 and 5 for Scenario 4 under the limited pumping case. For reference, positive values imply an enhancement to potential upward flow. Location of Yucca Mountain indicated by black circle symbol.



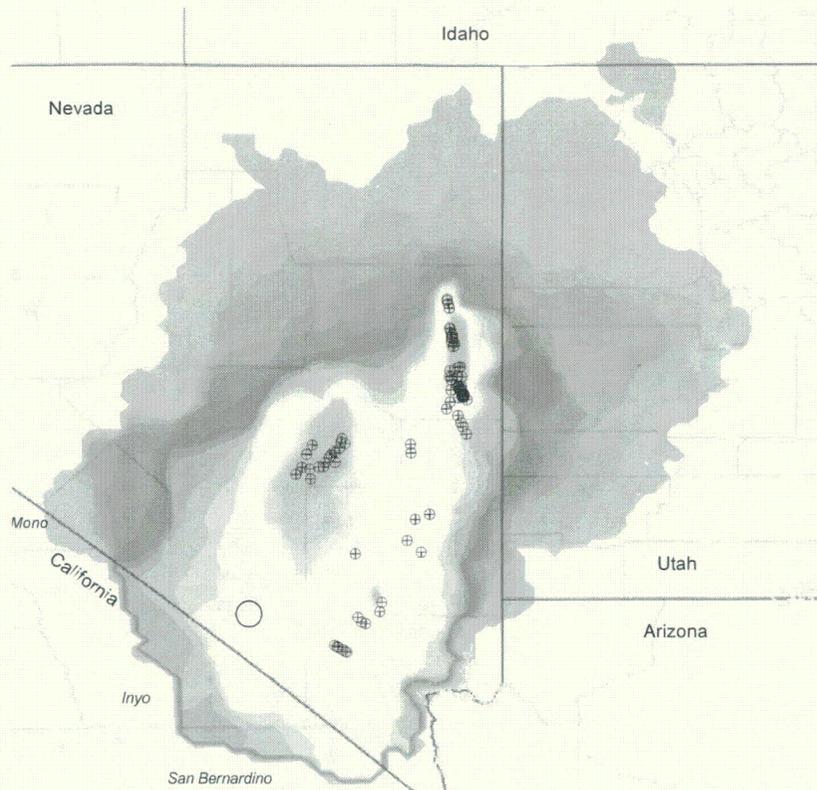
Simulated drawdown (ft) in Model Layer 1 for Pumping Scenario 1 (Railroad Valley), full pumping case. Location of Yucca Mountain indicated by black circle symbol.



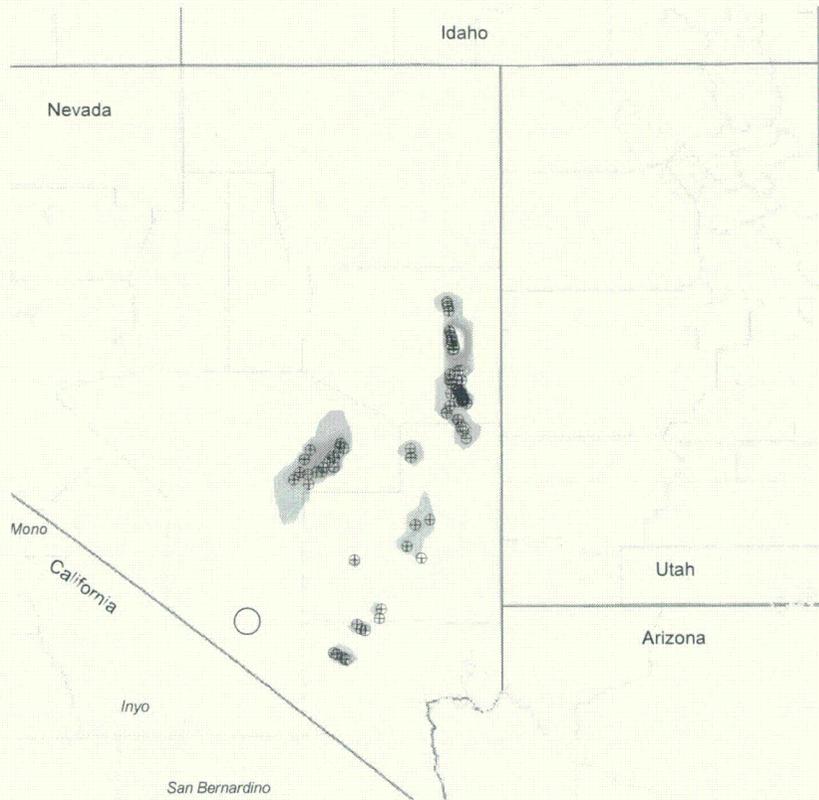
Simulated drawdown (ft) in Model Layer 1 for Pumping Scenario 2 (Spring Valley, Dry Lake Valley, Cave Valley and Delamar Valley), full pumping case. Location of Yucca Mountain indicated by black circle symbol.



Simulated drawdown (ft) in Model Layer 1 for Pumping Scenario 3 (Spring Valley, Dry Lake Valley, Cave Valley, Delamar Valley, Tikapoo Valley, and Three Lakes Valley), full pumping case. Location of Yucca Mountain indicated by black circle symbol.



Simulated drawdown (ft) in Model Layer 1 for Pumping Scenario 4 (Spring Valley, Dry Lake Valley, Cave Valley, Delamar Valley, Tikapoo Valley, Three Lakes Valley, and Railroad Valley), full pumping case. Location of Yucca Mountain indicated by black circle symbol.



Simulated transient drawdown (ft) in Model Layer 1 for Pumping Scenario 4 (Spring Valley, Dry Lake Valley, Cave Valley, Delamar Valley, Tikapoo Valley, Three Lakes Valley, and Railroad Valley), full pumping case; $t = 360$ years. Note that color scale differs from that used to depict steady-state model results. Location of Yucca Mountain indicated by black circle symbol.