



# ***Ground-water Chemistry Interpretations Supporting the Basin and Range Regional Carbonate-rock Aquifer System (BARCAS) Study, Eastern Nevada and Western Utah***

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*DHS Publication No. 41230*

*Prepared by*  
Desert Research Institute, Nevada System of Higher Education  
and  
U.S. Geological Survey

*Prepared for*  
U.S. Department of the Interior

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Table 13.  $\delta^2\text{H}$  mixing calculations for the average geochemically acceptable model results for northern Spring and Snake valleys.

Sample No.	$\delta^2\text{H}$ (‰)	percent mixture		
156	-125.0	82		
121	-115.7	4		
51	-122.7	14	$\delta^2\text{H}$ computed (‰)	difference
68	-124.3	100	-124.3	0.01
Sample No.	$\delta^2\text{H}$ (‰)	percent mixture		
68	-124.3	72		
53	-1116.3	4		
101	-115.4	24	$\delta^2\text{H}$ computed (‰)	difference
90	-120.3	100	-121.8	1.54

Water-rock reaction models were evaluated from central to southern Spring Valley and from southern Spring to southern Snake valleys (Figure 25 and Table 14).  $\delta^2\text{H}$ ,  $\delta^{18}\text{O}$ , and chloride were used to identify flowpaths from central Spring Valley and southern Snake Valley;  $\delta^2\text{H}$  versus  $\delta^{18}\text{O}$  is shown in Figure 28 and  $\delta^2\text{H}$  versus chloride is graphed in Figure 29. Note that chloride is not conservative as end member waters 230 and 258 do not plot on the mixing line (Figure 29). Viable water-rock reaction models were found to support ground-water flow from central Spring Valley to southern Spring Valley and from southern Spring Valley to southern Snake Valley (Table 15).

#### Snake Valley

Water-rock reaction modeling for Snake Valley used wells and regional springs along the central axis of the valley and mountain springs (Figure 30 and table 16).  $\delta^2\text{H}$ ,  $\delta^{18}\text{O}$  and chloride graphs were used to identify possible mixtures (Figure 31 and 32). Viable water-rock reaction models were found to support ground-water flow from southern Snake Valley to central Snake Valley and to northern Snake Valley (Table 17) where isotopically heavier water from locations 23 and 193 can be mixed with 258 to produce water at location 162. Water from 61 can be mixed with 162 to produce water at location 90. No viable water-rock reaction models could be found to ground-water flow north or east of Snake Valley, because of a scarcity of data in the surrounding valleys.

Table 14. Water chemistry and isotopic data used in water-rock reaction modeling for southern Spring and Snake valleys.

Sample Number		156	197	230	239	258
NWIS Station ID		390754114303001	385613114250401	384619114313800	384156114075600	383525114095100
Location name		LAP&W Spring Vly Well 1	USBLM (Shoshone pond well)	Fox Well	Big Spring (SV)	Monument Well
Source		NWIS	NWIS	BARCAS	SNWA	BARCAS
Organization		USGS	USGS	USGS-NV	DRI	USGS-NV
Sample type		GW	GW	GW	SP	GW
Sample date		9/16/1982	7/6/1983	8/30/2005	10/29/06	10/5/2005
Latitude	degrees	39.13160994	38.9363354	38.77222222	38.698920	38.59027778
Longitude	degrees	114.509171	114.418888	114.5275	114.132230	114.1641667
Datum				NAD83	WGS 84	NAD83
Altitude	m	1780.1	1760.9	1763.0	1700.5	1731.3
Water Temp	°C	15.5	23.5	16.5	17.0	12.1
pH-field		8	8.1	7.90	7.44	8.00
EC-field	µS/cm	290	168	328	384	350
DO	mg/L	6.2	6.8	5.2	4.83	5.6
Ca	mg/L	37	22	36.8	49.7	43.9
Mg	mg/L	13	0.74	24.0	20.3	12.2
Na	mg/L	4.7	8.8	8.00	5.93	24.5
K	mg/L	1.1	1	1.63	1.45	4.14
Cl	mg/L	2.1	2.2	5.6	6.1	16.8
SO <sub>4</sub>	mg/L	5	5	14.0	8.5	32.1
HCO <sub>3</sub> -lab	mg/L	180	85	181	232	134
CO <sub>3</sub> -lab	mg/L	0	0	0	0	0
SiO <sub>2</sub>	mg/L	16	22	15.4	12.9	46.8
Cation Sum	epm	3.1488	1.5671	4.2013	4.4458	4.3663
Anion Sum	epm	3.1137	1.5594	3.4161	4.1517	3.3385
Ion Balance		1.01	1.00	1.23	1.07	1.31
δ <sup>18</sup> O	‰	-16.4	-15	-14.89	-15.2	-14.98
δ <sup>2</sup> H	‰	-125	-109	-111.80	-111	-113.40

Explanation: NWIS – U.S. Geological Survey National Water Information system, BARCAS – Basin and Range Carbonate Aquifer System, SNWA - Southern Nevada Water Authority, USGS-NV – U.S. Geological Survey, Nevada District, DRI - Desert Research Institute, SP – spring, GW- ground water, m – meters, °C – degrees Celsius, µS/cm – micro-Siemens per centimeter, mg/L – milligrams per liter, DO – Dissolved Oxygen, Ca – calcium, Mg – magnesium, Na – sodium, K – potassium, Cl – chloride, SO<sub>4</sub> – sulfate, HCO<sub>3</sub> – bicarbonate, CO<sub>3</sub> – carbonate, SiO<sub>2</sub> – silicate, epm- equivalents per million, Cation Sum – sum of milli-equivalents of Ca, Mg, Na and K, Anion Sum – sum of milli-equivalents of Cl, SO<sub>4</sub>, HCO<sub>3</sub> and CO<sub>3</sub>, Ion Balance – Cation Sum divided by the Anion Sum, δ<sup>18</sup>O – oxygen isotopic ratio, δ<sup>2</sup>H – hydrogen isotopic ratio, ‰ – permil in VSMOW (Vienna Standard Mean Ocean Water).

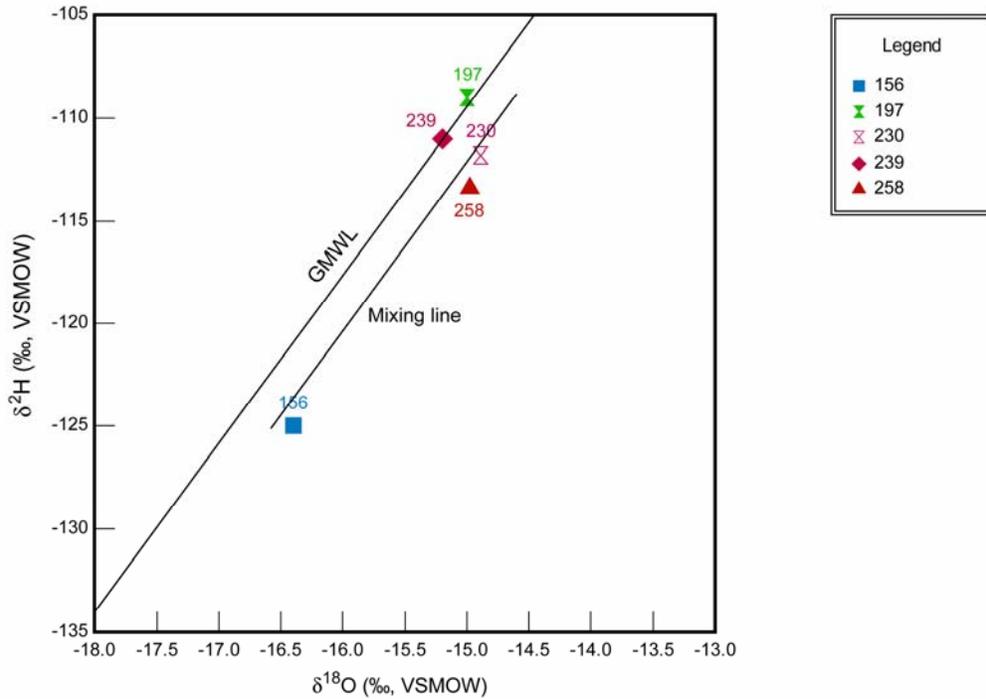


Figure 28.  $\delta^2\text{H}$  versus  $\delta^{18}\text{O}$  graph of selected analyses from wells and springs in southern Spring and Snake valleys. A local mixing line shows, isotopically heavier water from 197 can be mixed with 156, to yield 230, and water from 230 can be mixed with 239 to yield 258; Craig's meteoric water line ( $\delta\text{D} = 8 \delta^{18}\text{O} + 10$ ) for reference.

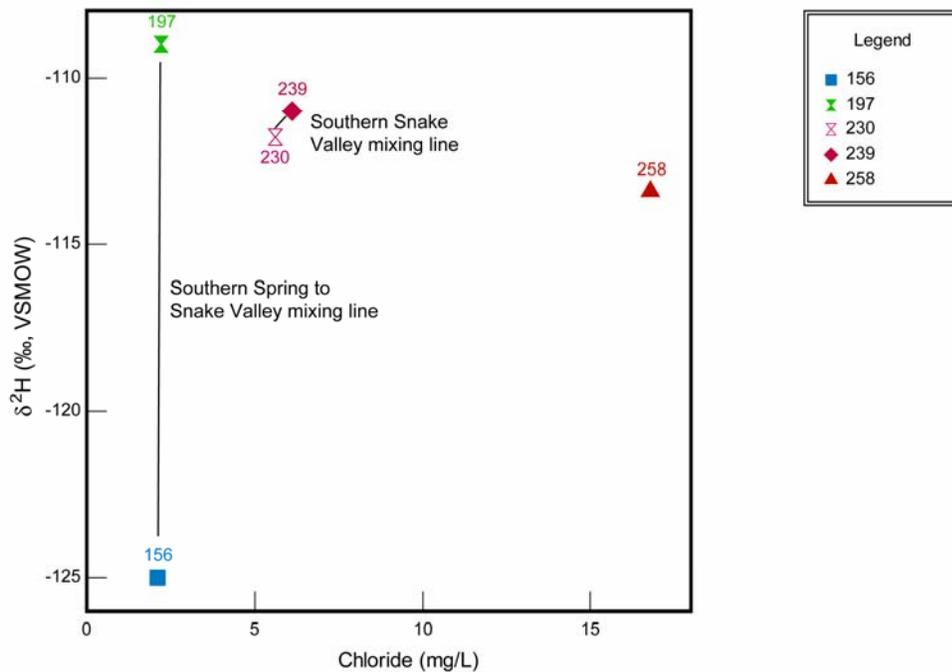


Figure 29.  $\delta^2\text{H}$  versus chloride graph of selected samples from wells and springs in southern Spring and Snake valleys. Mixing lines are shown from central to southern Spring Valley and from southern Spring Valley to southern Snake Valley.

Table 15.  $\delta^2\text{H}$  mixing calculations for the average geochemically acceptable model results for southern Spring and Snake valleys.

Sample No.	$\delta^2\text{H}$ (‰)	percent mixture		
156	-125.0	18		
197	-109.0	82	$\delta^2\text{H}$ computed (‰)	difference
230	-111.8	100	-111.9	-0.1
Sample No.	$\delta^2\text{H}$ (‰)	percent mixture		
230	-111.8	18		
239	-111.0	82	$\delta^2\text{H}$ computed (‰)	difference
258	-113.4	100	-111.1	2.3

## Ground-Water Travel Times

### Introduction

Carbon-14 based ground-water travel time can be calculated from carbon isotope compositions of the water by adjusting for the mass transfer of carbon into and out of the water if (1) the carbon isotope composition of the recharge water when the water becomes isolated from atmospheric and soil-zone  $\text{CO}_2$  gas is known, (2) the mass and sources of carbon added to or removed from the water are known, (3) the isotope composition of the sources of carbon are known, and (4) the fractionation of the isotopes during removal of carbon from the water is known (Wigley *et al.*, 1978). The biggest sources of error in calculating ground-water travel time using this approach are (1) the estimation of the starting, or original,  $^{14}\text{C}$  composition of the recharge waters ( $A_0$ , pre-nuclear-detonation  $^{14}\text{C}$ ), (2) the accuracy of water-rock reaction models in describing the input and output of carbon to the water along the flowpath, and (3) the isotope composition of carbon added to the water.

The reaction path between initial and final waters along a flowpath may be open or closed to  $\text{CO}_2$  gas, organic matter, and carbonate minerals. The initial water may represent any point along the flowpath. The initial  $^{14}\text{C}$  values used are best defined as the measured  $^{14}\text{C}$  composition of the up-gradient sample site. The modeled  $^{14}\text{C}$  "age" of the final water, therefore, represents only the travel time between the initial and final water. Initial  $^{14}\text{C}$  compositions ( $A_0$ ) were calculated using observed  $^{14}\text{C}$  from recharge water. If the system is closed to  $\text{CO}_2$  gas in recharge areas, adjusted  $^{14}\text{C}$  travel times will be younger than actual travel times. If water was recharged under totally open system conditions, then the closed system ages could be in error by about one  $^{14}\text{C}$  half-life (the water would be 5,730 years older than calculated). The maximum error of the adjusted ages in an open, rather than closed, system is, therefore, less than 5,730 years. The amount of possible error depends on the degree to which the system is open to  $\text{CO}_2$  gas during recharge.

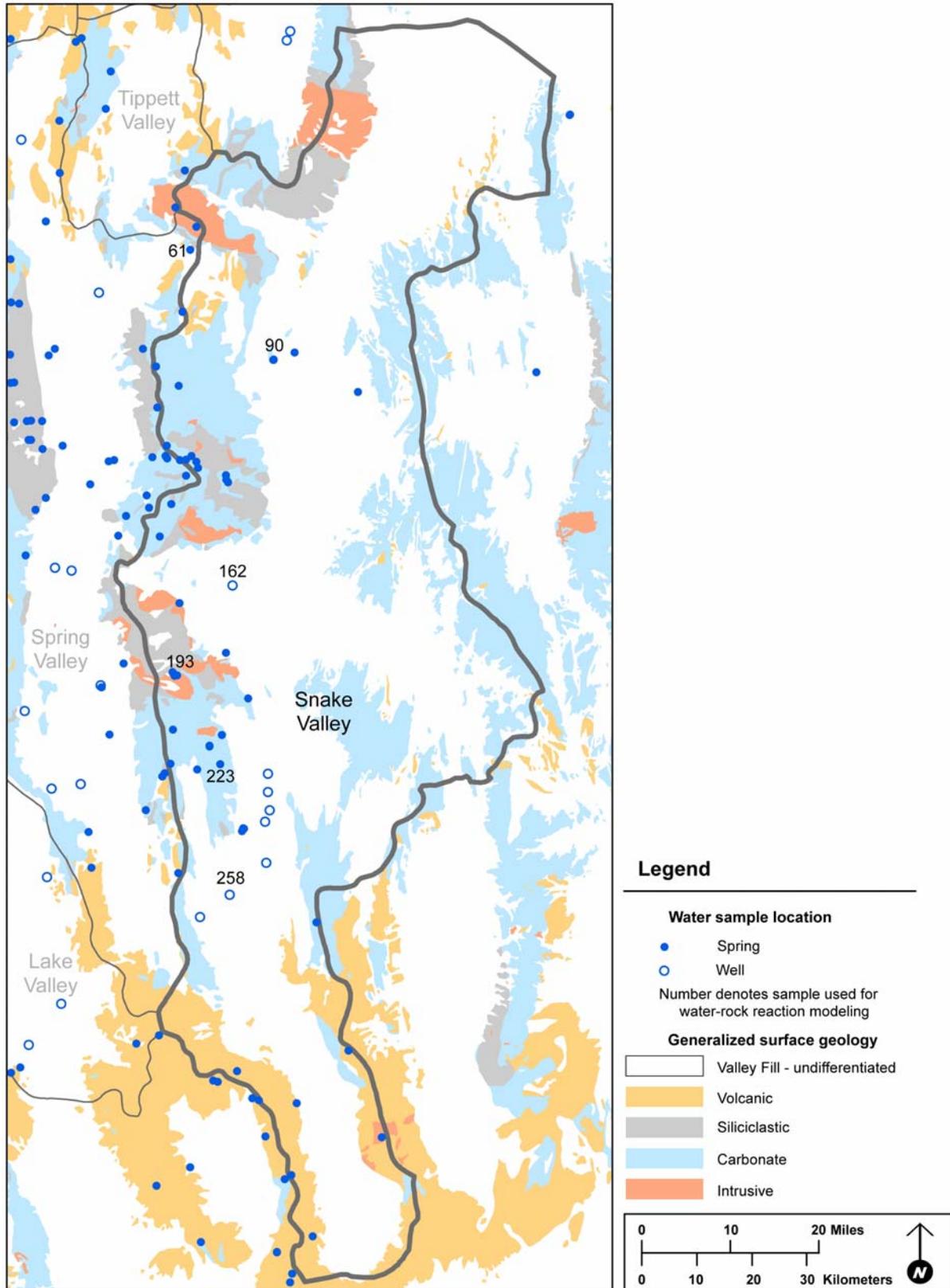


Figure 30. Map of selected locations for water-rock reaction modeling in Snake Valley.

Table 16. Water chemistry and isotopic data used in water-rock reaction modeling for Snake Valley.

Sample Number		61	90	162	193	223	258
NWIS Station ID		393837114121700	392735114021300	390542114081800	385702114153500	384814114103300	383525114095100
Location name		Mike's Spring	Gandy Warm Spring	Unnamed Well #1	Unnamed Spring #21	South Spring	Monument Well
Source		BARCAS	SNWA	BARCAS	SNWA	BARCAS	BARCAS
Organization		DRI	DRI	USGS-NV	DRI	DRI	USGS-NV
Sample type		SP	SP	GW	SP	SP	GW
Sample date		8/23/2005	10/29/06	7/16/2005	10/09/06	7/14/2005	10/5/2005
Latitude degrees		39.64370056	39.459980	39.09527778	38.950692	38.80398836	38.59027778
Longitude degrees		114.2049	114.037080	114.1383333	114.259746	114.1758813	114.1641667
Datum		WGS 84	WGS 84	NAD83	WGS 84	WGS 84	NAD83
Altitude m		2203.7	1599.6	1678.2	2955.7	2269.6	1731.3
Water Temp °C		10.7	26.9	18.0	4.3	9.7	12.1
pH-field		6.77	7.56	6.90	6.96	6.87	8.00
EC-field µS/cm		559	485	474	94.4	502	350
DO mg/L		6.4	5.74	6.0	6.66	5.7	5.6
Ca mg/L		61.9	51.1	67.8	11.1	66.3	43.9
Mg mg/L		18.9	17.1	22.40	2.63	28.8	12.2
Na mg/L		31.8	28.6	13.5	4.51	2.23	24.5
K mg/L		1.81	3.85	1.8	1.06	0.54	4.14
Cl mg/L		29.2	24.1	15.9	1.2	2.6	16.8
SO <sub>4</sub> mg/L		34.8	22.7	20.5	2.9	3.4	32.1
HCO <sub>3</sub> -lab mg/L		246	233	222	51	343	134
CO <sub>3</sub> -lab mg/L		0	0	0	0	0	0
SiO <sub>2</sub> mg/L		27.5	22.4	17.1	13.9	7.6	46.8
Cation Sum epm		6.0739	5.2998	5.8606	0.9936	5.7896	4.3663
Anion Sum epm		5.5803	4.9714	4.5140	0.9237	5.7662	3.3385
Ion Balance		1.09	1.07	1.30	1.08	1.00	1.31
δ <sup>18</sup> O ‰		-15.89	-15.9	-15.08	-15.62	-14.72	-14.98
δ <sup>2</sup> H ‰		-121.1	-120	-113.20	-114.6	-108.2	-113.40

Explanation: NWIS – U.S. Geological Survey National Water Information system, BARCAS – Basin and Range Carbonate Aquifer System, SNWA - Southern Nevada Water Authority, USGS-NV – U.S. Geological Survey, Nevada District, DRI - Desert Research Institute, SP – spring, GW- ground water, m – meters, °C – degrees Celsius, µS/cm – micro-Siemens per centimeter, mg/L – milligrams per liter, DO – Dissolved Oxygen, Ca – calcium, Mg – magnesium, Na – sodium, K – potassium, Cl – chloride, SO<sub>4</sub> – sulfate, HCO<sub>3</sub> – bicarbonate, CO<sub>3</sub> – carbonate, SiO<sub>2</sub> – silicate, epm- equivalents per million, Cation Sum – sum of milliequivalents of Ca, Mg, Na and K, Anion Sum – sum of milliequivalents of Cl, SO<sub>4</sub>, HCO<sub>3</sub> and CO<sub>3</sub>, Ion Balance – Cation Sum divided by the Anion Sum, δ<sup>18</sup>O – oxygen isotopic ratio, δ<sup>2</sup>H – hydrogen isotopic ratio, ‰ – permil in VSMOW (Vienna Standard Mean Ocean Water).

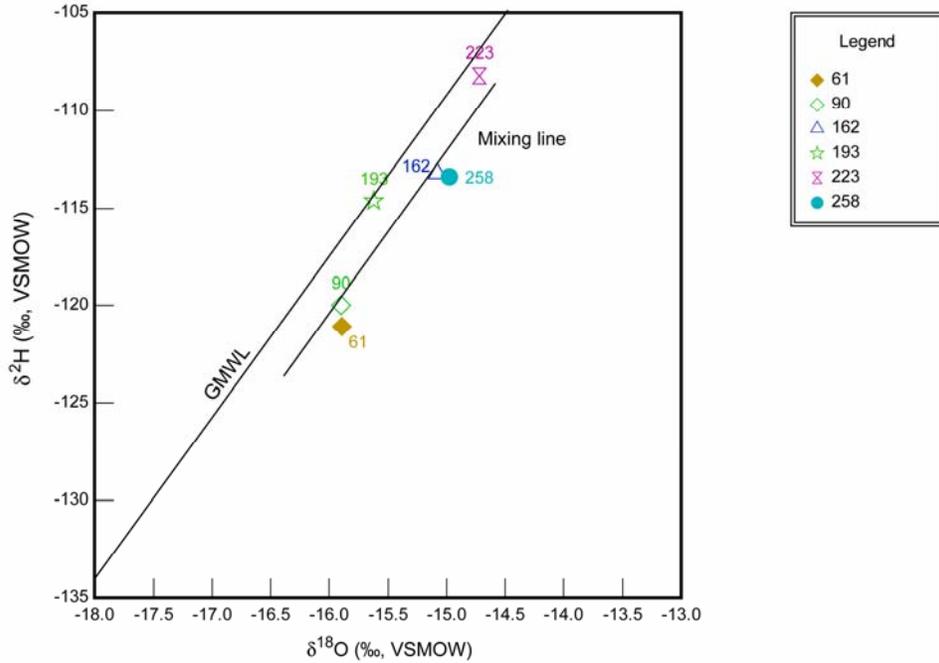


Figure 31.  $\delta^2\text{H}$  versus  $\delta^{18}\text{O}$  graph of selected analyses from wells and springs in Snake Valley. A local mixing line shows, isotopically heavier water from 223 and 193 can be mixed with 258, to yield 162, and water from 61 can be mixed with 162 to yield 90. GMWL = Global Meteoric Water Line (Craig, 1961) ( $\delta^2\text{H} = 8 \delta^{18}\text{O} + 10$ ).

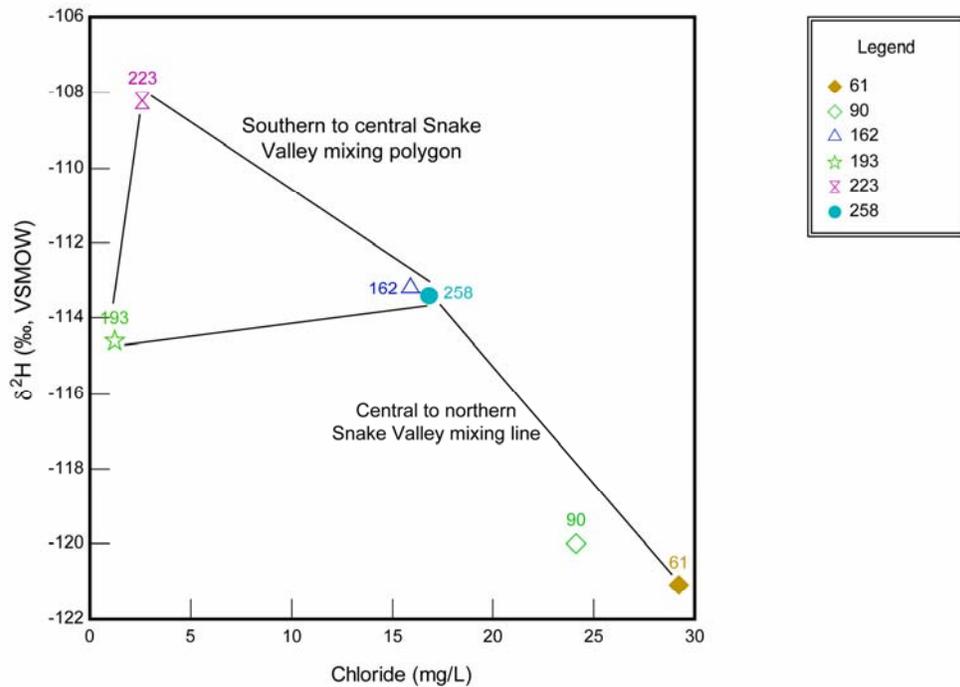


Figure 32.  $\delta^2\text{H}$  versus chloride graph of selected samples from wells and springs in Snake Valley. A mixing polygon is shown for southern to central Snake Valley, and a mixing line from central to northern Snake Valley.

Table 17.  $\delta^2\text{H}$  mixing calculations for the average geochemically acceptable model results for Snake Valley.

Sample No.	$\delta^2\text{H}$ (‰)	percent mixture		
258	-113.4	30		
223	-108.2	30		
193	-1114.6	40	$\delta^2\text{H}$ computed (‰)	difference
162	-113.2	100	-112.3	0.9
Sample No.	$\delta^2\text{H}$ (‰)	percent mixture		
162	-113.2	60		
61	-121.8	40	$\delta^2\text{H}$ computed (‰)	difference
90	-120.0	100	-116.6	3.4

Sources of carbon in the carbonate-rock aquifer outside of the recharge areas are carbonate minerals (calcite and dolomite), organic matter, and  $\text{CO}_2$  gas in rock units. Atmospheric and soil-zone  $\text{CO}_2$  gas are not considered sources of carbon outside of the recharge areas because depths to water in the carbonate-rock aquifer are generally several hundred to as much as 2,000 feet below land surface. Speciation and saturation state calculations indicate that dolomite dissolved in some areas because dolomite is undersaturated in the water. Calcite also dissolves and subsequently precipitates, as indicated by the lighter calculated than measured  $\delta^{13}\text{C}$  compositions of the water. Calcite dissolution and precipitation are caused by temperature and pressure changes along flowpaths (Winograd and Pearson, 1976) and possibly increased surface area of calcite by grinding during faulting. Oxidation of organic matter, which would produce  $\text{CO}_2$  gas, and  $\text{CO}_2$  gas in rock units overlying the aquifers likely do not contribute significant amounts of carbon to the water.

Carbon isotope fractionation factors between precipitating calcite or exsolving  $\text{CO}_2$  gas and a carbonate solution at varying temperatures and pH's can be measured (Deines *et al.*, 1974; Mook, 1980). These fractionations have to be accounted for in calculating the  $\delta^{13}\text{C}$  composition resulting from the mass transfer of carbon into and out of the water.

The following assumptions were made in using  $\delta^{13}\text{C}$  to determine the amount of carbon moving in to and out of the water along flowpaths:

1. Carbon is removed from the water by calcite precipitation.
2. Carbon is added to the water by dissolution of dolomite. The  $\delta^{13}\text{C}$  composition of dolomite was assumed to be 0.0 ‰ (Thomas *et al.*, 1996), and the  $^{14}\text{C}$  composition was assumed to be 0 percent modern carbon (pmc).
3. Carbon input, in addition to dolomite dissolution, along some flowpaths is assumed to be the result of the dissolution of  $\text{CO}_2$  gas, calcite, or dolomite in recharge areas. This carbon is added to the recharge water because the concentration of dissolved inorganic carbon in water circulating to depth in the aquifers is probably greater than measured at high-altitude springs and wells. For these calculations, the total dissolved inorganic carbon concentration of the recharge water was increased by the mass-balance calculated  $\text{CO}_2$  input. This carbon is added to account for the sparse carbon data for ground-water from high altitude, the potentially large range in inorganic carbon concentrations based on samples from high altitude sites in southern Nevada (Thomas *et al.*, 1991), and  $\text{CO}_2$  degassing from spring water. The  $\delta^{13}\text{C}$  composition of  $\text{CO}_2$  used in water-rock reaction modeling was -17 ‰ (Quade *et al.*, 1989).

4. The  $\delta^{13}\text{C}$  composition of dissolving calcite is difficult to determine because calcite veins are prevalent within the carbonate-rock aquifers (Thomas *et al.*, 1991). Primary calcite in southern Nevada has a  $\delta^{13}\text{C}$  composition ranging from -0.5 to +3.7 ‰, with an average of +1.2 ‰, whereas secondary (vein) calcite from Devils Hole has an average  $\delta^{13}\text{C}$  composition of -4.0 ‰ (Coplen *et al.*, 1994). Given that most water-rock reaction models require the dissolution of dolomite, some primary calcite probably also dissolves, and because veins are prevalent throughout the carbonate rock, some vein calcite probably also dissolves. Thus, adjusted  $^{14}\text{C}$  age calculations based on mass transfer calculated using a  $\delta^{13}\text{C}$  value that is some proportion of primary and secondary calcite values probably gives the most realistic adjusted age. Water-rock reaction modeling using  $\delta^{13}\text{C}$  assumed dissolving calcite is 40 percent primary calcite with a  $\delta^{13}\text{C}$  value of +1.2 ‰ and 60 percent secondary calcite with a  $\delta^{13}\text{C}$  value of -4.0 ‰, resulting in an average  $\delta^{13}\text{C}$  composition of -1.92 ‰.
5. The measured  $\delta^{13}\text{C}$  composition of the final water is heavier (less negative) than the calculated composition. This difference is assumed to be the result of dissolution and subsequent precipitation of calcite caused by temperature and pressure changes along the flowpaths (Winograd and Pearson, 1976) and possibly because of increased carbon isotope exchange resulting from increased calcite surface areas caused by faulting. The increase in calcite surface area plus the exposure of fresh surfaces, produced by fracturing and grinding associated with faulting, would increase the exchange rate of carbon between solution and calcite (Mozeto *et al.*, 1984). This additional input of carbon was determined by accounting for the fractionation of  $\delta^{13}\text{C}$  in a hypothetical sequence of dissolving and precipitating calcite using Rayleigh distillation equations (Plummer *et al.*, 1991).
6. All  $\delta^{13}\text{C}$  fractionation factors used in the computer program NETPATH were set for the fractionation of  $\delta^{13}\text{C}$  during calcite precipitation and calculated using the equations of Deines *et al.* (1974). A  $\delta^{13}\text{C}$  fractionation factor of three was used for the equilibrium fractionation between precipitating calcite and a carbonate solution.
7. Mixing waters have similar histories. The carbonate-rock aquifers contain relatively homogeneous mineralogies, and waters in the aquifers are similar in chemical composition, indicating that most of the water in the aquifers has undergone the same major geochemical processes.

#### Dissolved Inorganic Carbon Travel Times

Calculation of ground-water travel times focused on the White River, Spring Valley, and Snake Valley basins because of limited carbon isotope data in the rest of the BARCAS study area. Water-rock reaction modeling rarely produces unique model solutions. Thus, there may be numerous valid water-rock reaction models for any particular flowpath or mixture of waters. The non-uniqueness of water-rock reaction modeling generally produces a range of acceptable models, and thus, a range of ground-water travel times.

#### *White River*

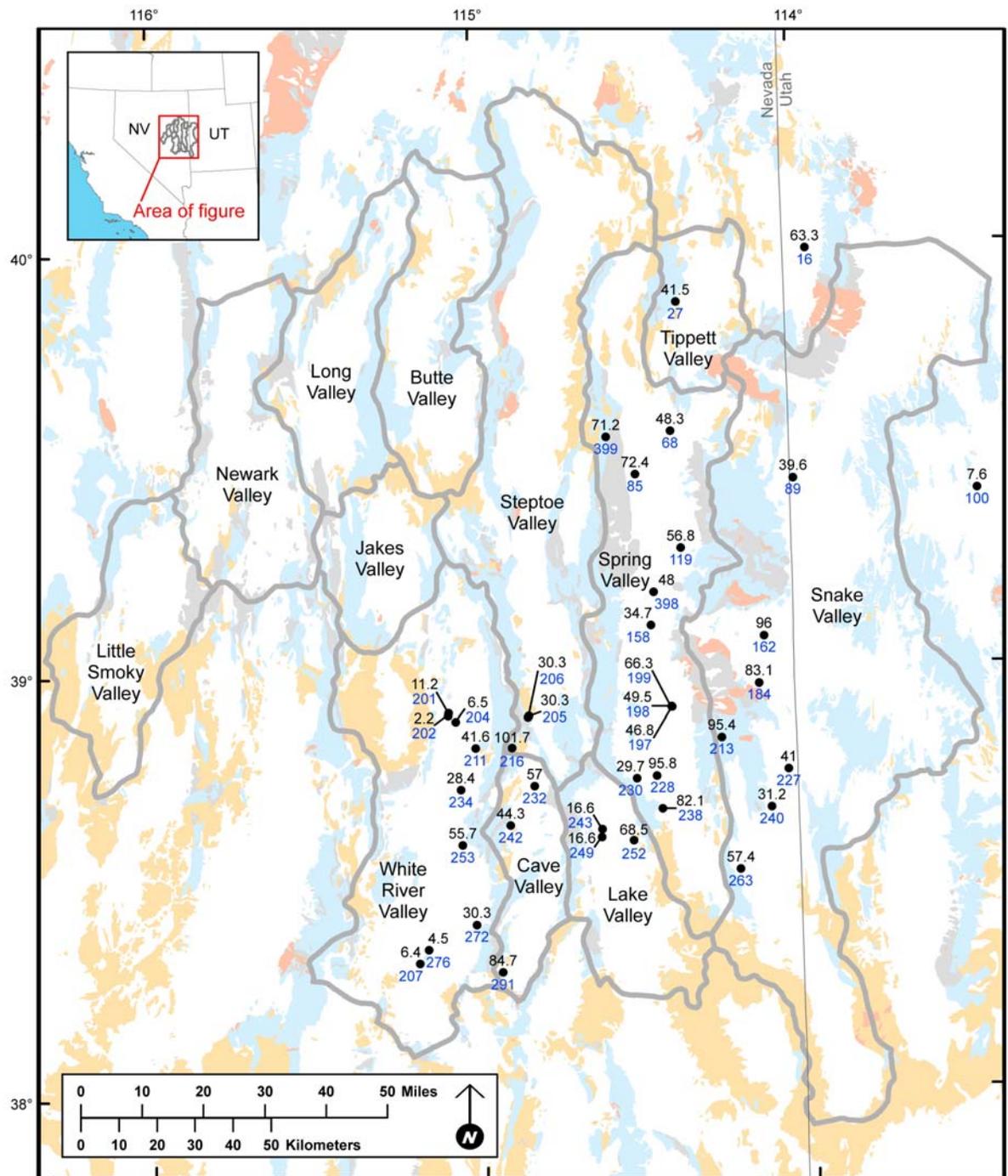
Water-rock reaction modeling with DIC isotopes to calculate ground-water  $^{14}\text{C}$  travel times for the White River basin was successful for several modeling scenarios. Models were developed from central White River to southern White River. Central White River Valley

ground water (201) was mixed with different mountain recharge waters (Table 18 and Figure 33) to produce ground water at Hot Creek Spring (#276). Flow from basins north of White River into northern White River and from northern White River to central White River could not be modeled because of a lack of water chemistry and isotopic data. Valid water-rock reactions models were developed using valley springs in the central axis of the central White River basin mixing with mountain springs (ground-water recharge) to produce ground-water at valley springs in the central axis of the southern White River basin.

Table 18. Carbon isotopic data used in NETPATH water-rock reaction modeling to calculate ground-water travel times for White River Valley flowpaths.

Location No.	Location Name	$\delta^{13}\text{C}$ (‰)	$^{14}\text{C}$ (pmc)
201	Preston Big Spring	-5.7	11.2
211	Lund Spring	-8.1	41.6
216	Hole in the Bank Spring	-13.2	101.7
232	Robbers Roost	-5.5	57.0
242	Haggerty Spring	-9.3	44.3
253	Emigrant Spring	-9.2	55.7
272	Butterfield Spring	-8.5	30.3
276	Hot Creek Spring	-4.3	4.5
291	Lewis Well	-4.7	84.7

Water-rock reaction models for this basin required the dissolution of feldspars, gypsum, and dolomite; precipitation of clay, zeolite, or  $\text{SiO}_2$ ; calcium and magnesium exchange for sodium; uptake of  $\text{CO}_2$  gas; and, either dissolution or precipitation of calcite. The contribution of mountain recharge to valley ground water in southern White River ranged from 40 to 60 percent; the contribution of central White River Valley ground water to southern White River valley groundwater ranged from 40 to 60 percent. To adjust DIC isotopic concentrations for carbon isotopic exchange with carbon containing minerals along the flowpath, 3.0 to 4.0 mmol/L of carbon had to exchange. The resulting travel times for this flowpath ranged from 12,000 to 16,000 years (Table 19). Approximate ground-water velocities from central to southern White River basin range from 10 to 20 ft/yr (Table 19).



**Legend**

- |      |   |   |                                    |
|------|---|---|------------------------------------|
| —    | BARCAS study basin  |   | <b>Generalized surface geology</b> |
| ●    | Dissolved inorganic carbon <sup>14</sup> C sample locations | □ | Valley fill - undifferentiated     |
| 51.5 | <sup>14</sup> C, percent modern carbon                      | ■ | Volcanic                           |
| 27   | Sample number   | ■ | Siliciclastic                      |
|      |   | ■ | Carbonate                          |
|      |   | ■ | Intrusive                          |

Figure 33. Dissolved inorganic carbon <sup>14</sup>C sample locations in the BARCAS study area.

Table 19. Water-rock reaction modeling results for inter-basin flow in the BARCAS study area. (Flowpath Number matches corresponding number on Figure 34; HA, hydrographic area; IB, intra-basin; Water-rock reaction model mixtures of ground and recharge waters, represents total mixture of ground and recharge waters for first [upper mixture] and second [lower mixture] model evaluations; Ground Water, first point along selected ground-water flowpath; Recharge Water, water contributed from surrounding recharge area; Inorganic Carbon Travel Time/Organic Carbon Travel Time, represents time calculated from first [upper time] and second [lower time] model evaluations; NA, not available.)

Flowpath Number	Flowpath Location and Locations			Geochemical model – Mixtures of ground and recharge water (percent)	Inorganic Carbon Travel Time (years)	Organic Carbon Travel Time (years)	Ground-water Flow Velocity (ft/yr)	Geochemical Model Results
	Initial Location	Final Location	HA Boundary or IB Divide	Ground Water - Recharge Water				
1	Northern Spring Valley	Northern Snake Valley	HA	0 – 100 30 – 7	<1,000 <1,000	2,000 4,000	100 to 200	Supports ground-water flowpath
2	Southern Spring Valley	Southern Snake Valley	HA	0 – 100 100 – 0	<1,000 6,000	<1,000 2,000	20 to 100	Supports ground-water flowpath
3	Southern Steptoe Valley	Southern Spring Valley	HA	70-30 100-0	<1,000 <1,000	NA	10 to 40	Supports ground-water flowpath
4	Lake Valley	Southern Spring Valley	HA	95 – 5 100 – 0	<1,000 <1,000	<1,000 <1,000	50 to 60	Supports ground-water flowpath
5	Southern part of northern Spring Valley	Northern part of northern Spring Valley	IB	0 – 100 60 – 40	<1,000 3,000	NA	40 to 150	Supports ground-water flowpath
6	Central Spring Valley	Southern Spring Valley	IB	20 – 80 40 – 60	<1,000 6,000	6,000 6,000	10 to 200	Supports ground-water flowpath
7	Central White River Valley	Southern White River Valley	IB	40 – 60 60 – 40	12,000 16,000	NA	10 to 20	Supports ground-water flowpath
--	Southern Steptoe Valley	Lake Valley	HA	No model	No model	NA	No model	No model
--	Cave Valley	Southern White River Valley	HA	No model	No model	NA	No model	No model

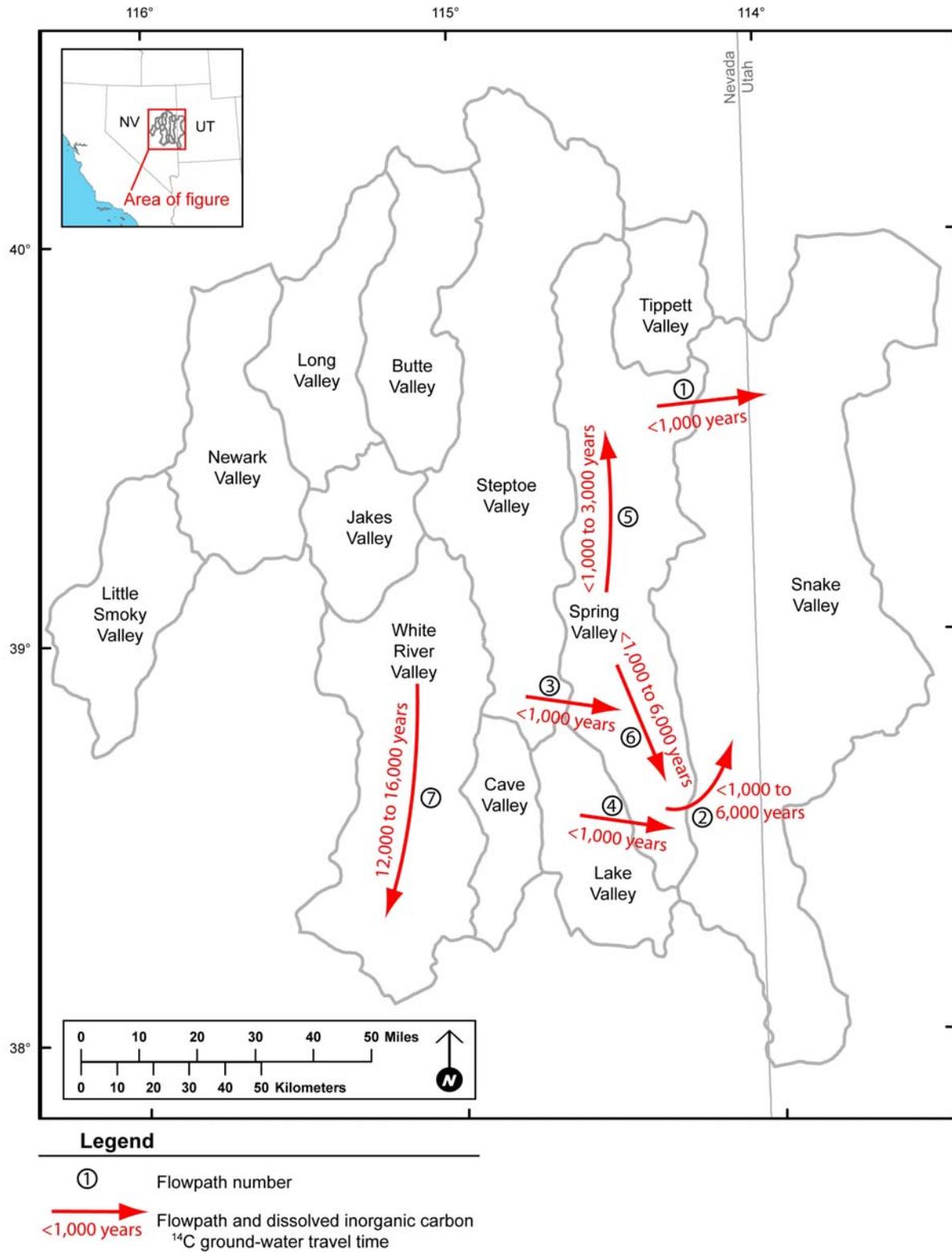


Figure 34. Dissolved inorganic carbon <sup>14</sup>C ground-water travel times for the BARCAS study area.

## Spring Valley

Multiple water-rock reaction modeling scenarios to calculate  $^{14}\text{C}$  ground-water travel times were attempted within the Spring Valley basin. These scenarios included modeling mountain recharge flowing to the valley aquifers, ground water flowing from central to southern Spring Valley, and ground water flowing from central to northern Spring Valley. The location names and DIC isotopic data used for water-rock reaction modeling and calculation of ground-water travel times are listed in Table 20 (Table 20, #s 119, 197, 230) and are shown on Figure 33.

Table 20. Carbon isotopic data used in NETPATH water-rock reaction modeling to calculate ground-water travel times for Spring Valley flowpaths.

Location No.	Location Name	$\delta^{13}\text{C}$ (‰)	$^{14}\text{C}$ (pmc)
27	Tippett Spring	-8.4	41.5
68	Elderidge Well	-6.1	48.3
85	Unnamed Spring #16	-14.3	72.4
119	South Millick	-7.7	56.8
158	South Bastian (flowing well)	-6.8	34.7
197	Unnamed Spring (north of Shoshone)	-6.4	46.8
198	The Cedars (flowing well)	-9.5	49.5
199	Shoshone Ponds	-11.7	66.3
213	Mustang Spring	-10.5	95.4
230	Fox Well	-5.5	29.7
238	Unnamed Spring U20a	-9.3	82.1
252	North Gouge Eye Spring	-8.8	68.5
398	Unnamed Spring #14	-7.8	48.0

Water-rock reaction models for mountain recharge flowing to the valley aquifers (Table 20, #s 119, 197, 230) required the dissolution of feldspars or  $\text{SiO}_2$ ; dissolution of gypsum, dolomite, and small amounts of  $\text{NaCl}$ ; precipitation of clay or zeolite; calcium and magnesium exchange for sodium; uptake or loss of  $\text{CO}_2$  gas; and, precipitation of calcite. Mountain recharge flowing to the valley aquifers was predominantly carbonate character water, which picks up its chemical character from the dissolution of carbonate rocks (limestone and dolomite). Very little to no mountain recharge appears to be from dilute waters that have contacted siliciclastic rocks (sandstones and shales), which are prevalent in mountain outcrops in some areas (Figure 2). Very little carbon isotopic exchange (0 to 1.0 mmol/L) was needed to adjust DIC isotopic concentrations for carbon isotopic exchange with carbon containing minerals along the flowpaths. The resulting travel times for mountain recharge flowing to the valley aquifers ranged from 1,000 to 6,000 years. Note however, that many scenarios from mountain recharge to the valley did not work because of too much chloride in the valley ground-waters or the  $\delta^2\text{H}$  values did not support a point-to-point flowpath suggesting that ground-water flow within the Spring Valley basin (*i.e.* regional ground-water flow) is more complex than the simple flowpaths tested.

Water-rock reaction models of ground-water flow from central Spring Valley to southern Spring Valley (Table 20, #230) required the dissolution of feldspar or  $\text{SiO}_2$ ; dissolution of gypsum, dolomite, and small amounts of  $\text{NaCl}$ ; precipitation of clay, zeolite or

SiO<sub>2</sub>; calcium and magnesium exchange for sodium; and, precipitation of calcite. The contribution of mountain recharge to valley ground water in southern Spring Valley ranged from 60 to 80 percent; the contribution of central Spring Valley ground water to southern Spring Valley ground water ranged from 20 to 40 percent. Very little carbon isotopic exchange (0 to 0.5 mmol/L) was needed to adjust DIC isotopic concentrations for carbon isotopic exchange with carbon containing minerals along the flowpaths. The resulting travel times for this flowpath ranged from <1,000 to 6,000 years (Table 19). Approximate ground-water velocities from central to southern Spring Valley basin ranged from 10 to 200 ft/yr (Table 19).

Water-rock reaction models of ground-water flow from the southern part of northern Spring Valley to the northern part of northern Spring Valley (Table 20, #68) required the dissolution of feldspar or SiO<sub>2</sub>; dissolution of gypsum, dolomite, and small amounts of NaCl; precipitation of clay, zeolite or SiO<sub>2</sub>; calcium and magnesium exchange for sodium; and, precipitation of calcite. The contribution of mountain recharge to valley ground water in northern Spring Valley ranged from 40 to 100 percent; the contribution of central Spring Valley ground water to northern Spring Valley ground water ranged from 0 to 60 percent. The wide range contribution of mountain recharge versus central Spring Valley ground water to northern Spring Valley ground water illustrates the sometimes non-uniqueness of water-rock reaction modeling. This non-uniqueness for this flowpath may be indicative of the lack of representative ground-water chemistry and isotopic data in northern Spring Valley. No carbon isotopic exchange was needed to adjust DIC isotopic concentrations for carbon isotopic exchange with carbon containing minerals along the flowpath. The resulting travel times for this flowpath ranged from <1,000 to 3,000 years (Table 19). Approximate ground-water velocities for this flowpath ranged from 40 to 150 ft/yr (Table 19).

#### *Inter-Basin Flowpaths*

##### Spring Valley to Snake Valley

Water-rock reaction modeling scenarios considered two different southern Snake Valley ground-waters, Hyde Well (Table 21 and Figure 33, #263, 57 pmc) and Big Springs (Table 21 and Figure 33, #240, 31 pmc). Water-rock reaction models of ground-water flow from southern Spring Valley to southern Snake Valley required the dissolution of feldspar, dolomite, gypsum, and small amounts of NaCl; precipitation of clay or zeolite; calcium and magnesium exchange for sodium; uptake or loss of CO<sub>2</sub> gas; dissolution or precipitation of SiO<sub>2</sub>; and, dissolution or precipitation of calcite. The contribution of mountain recharge to southern Snake Valley ranged from zero to 100 percent; correspondingly, the contribution of southern Spring Valley ground water to southern Snake Valley also ranged from 0 to 100 percent. However, qualitatively, the models with 78 to 100 percent mountain recharge produced “better” carbon isotope matches to observed Snake Valley ground-water isotopic values. Minimal carbon isotopic exchange (zero to 0.1 mmol/L) was needed to adjust DIC isotopic concentrations for carbon isotopic exchange with carbon containing minerals along the flowpath. The resulting travel times for this flowpath ranged from <1,000 to 6,000 years (Table 19). Ground-water flow velocities are approximately 20 to 100 ft/yr.