

**SOURCES AND MECHANISMS OF RECHARGE FOR
GROUND WATER IN THE WEST-CENTRAL
AMARGOSA DESERT, NEVADA--
A GEOCHEMICAL INTERPRETATION**

U.S. GEOLOGICAL SURVEY

Open-File Report 83-542

**Prepared in cooperation with the
U.S. DEPARTMENT OF ENERGY**





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By Hans C. Claassen

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METRIC TO INCH-POUND

<u>Multiply SI units</u>	<u>by</u>	<u>To obtain inch-pound units</u>
meter (m)	3.281	foot (ft)
square kilometer (km ²)	0.386	square mile (mi ²)
cubic meter per day per meter [(m ³ /d)/m]	10.764	cubic foot per day per foot [(ft ³ /d)/ft]
cubic meter per second (m ³ /s)	15,842	gallon per minute (gal/min)
kilopascal (kPa)	0.145	pound per square inch (lb/in. ²)
degree Celsius (°C)	°F= 1.8°C + 32	degree Fahrenheit (°F)

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ABSTRACT

Ground water in the west-central Amargosa Desert, Nevada, was recharged primarily by overland flow of snowmelt in or near the present-day stream channels, rather than by subsurface flow from highland recharge areas to the north. Geochemical arguments, including reaction mechanisms, are used to support these findings. Carbon-, hydrogen-, and oxygen-isotope data show that much of the recharge in the area occurred during late Wisconsin time. Absence of ground water recharged prior to late Pleistocene is considered to indicate that either climatic conditions were unfavorable for recharge or that ground-water velocities were such that they transported this earlier recharge away from the study area.

INTRODUCTION

Ground-water chemical data can help define ground-water flow systems and evaluate the relative importance of ground-water sources and pathways when combined with hydraulic data. Many attempts have been made to use ground-water chemical data in this manner, with limited success, primarily because of lack of understanding of the mechanisms responsible for dissolved constituents in ground-water samples and difficulty in obtaining samples that are representative of a definable part of the hydrologic environment.

Advances have been made during the last decade toward understanding the role of various mineral phases in dissolution and precipitation reactions, primarily in the context of thermodynamic equilibrium. However, this concept is limited for two reasons: (1) Accuracy of the thermodynamic data for naturally-occurring minerals is suspect; and (2) use of equilibrium concepts to describe systems that are generally in disequilibrium is questionable.

In this report, geochemical concepts that recognize the chemically dynamic nature of ground water are applied to determine the sources and pathways of ground water in the west-central Amargosa Desert in southern Nevada. These concepts help evaluate migration potential for radioactive waste resulting from nuclear weapons testing or migration potential for radioactive wastes stored on the Nevada Test Site. The study area encompasses one of the possible migration paths.

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DESCRIPTION OF THE STUDY AREA

The first comprehensive study made of water resources of the study area was by Walker and Eakin (1963); they discussed the climate, geologic history, lithology, hydrology, and some of the water-quality characteristics of the Amargosa Desert, particularly as they relate to water use and development potential. The primary study area in this report is shown in figure 1.

Topography, Vegetation, and Climate

The surface-drainage area of the Amargosa Desert is about 6,700 km² (square kilometers) (Walker and Eakin, 1963); a part is shown in figure 1. Elevations in the study area range from about 670 to 2,100 m (meters) above mean sea level, although a few higher peak elevations occur at about 2,400 m. Mean annual precipitation corresponding to these elevations is 50 to 380 mm (millimeters); mean annual temperatures range from 6° to 16°C (degrees Celsius).

The type of vegetation depends on degree of slope, slope aspect, and precipitation. Steep slopes, especially south- or west-facing, have little or no vegetation. Lower elevation, gently sloping alluvial fans support desert scrub; higher density cover exists at elevations near 1,500 m and lower density cover exists at lower elevations. Scattered juniper and sage are the dominant woody plants from about 1,500 m to about 1,800 to 2,100 m, where pinyon pines begin to dominate.

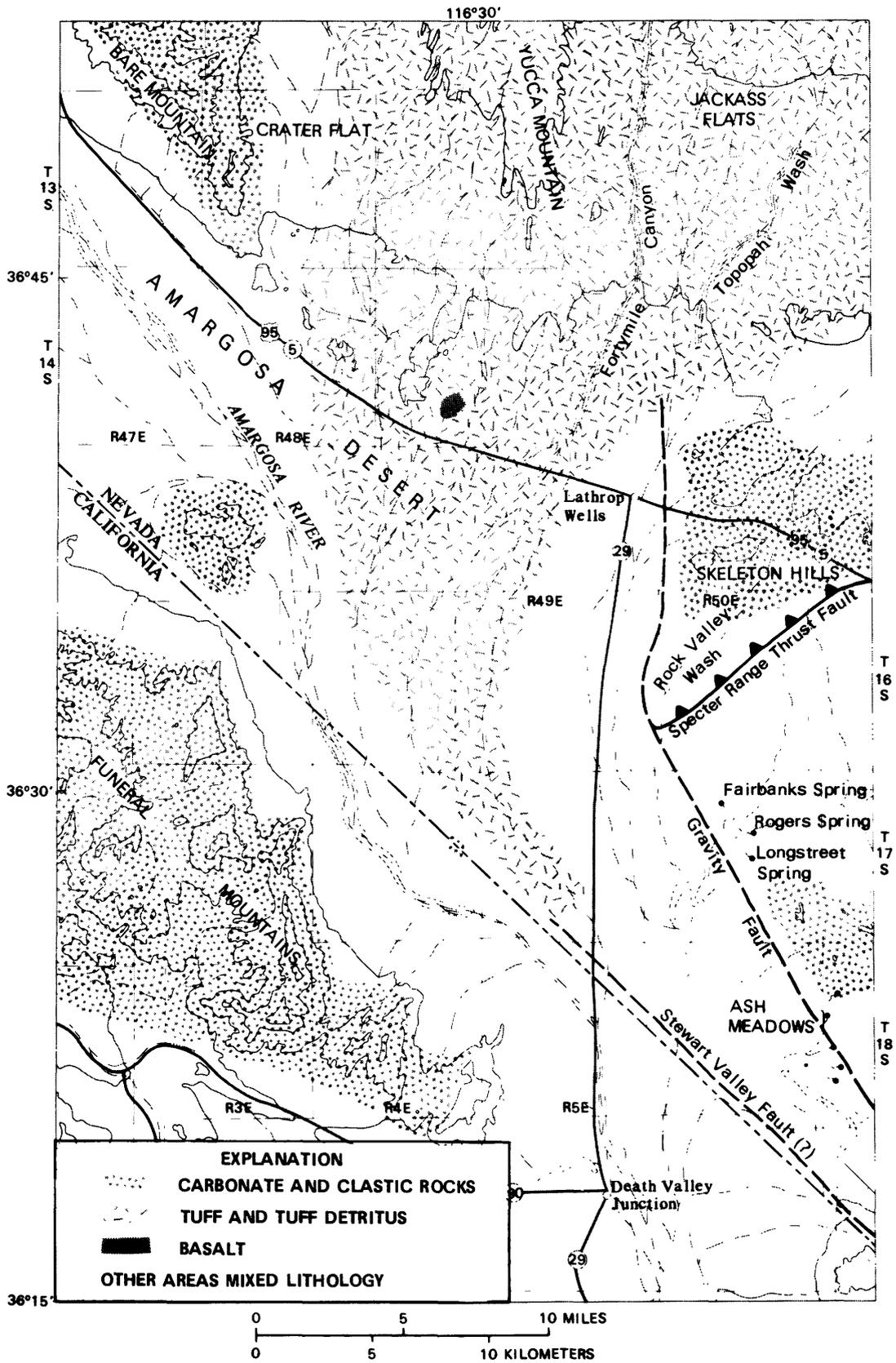


Figure 1.--West-central Amargosa Desert and vicinity.

Geology and Geomorphology

Lithology of the bedrock outcrops in the basin is varied; the bedrock consists primarily of rhyolitic volcanic ash falls and flows of Tertiary age, and carbonate, quartzite, and argillite that range from Cambrian to Devonian in age (Cornwall and Kleinhampl, 1961; Walker and Eakin, 1963; Burchfiel, 1964; Orkild, 1965; Barnes and Poole, 1968; Ekren, 1968; Winograd and Thordarson, 1975; Byers and others, 1976). Valley fill is derived from these rocks, and areal and subsurface distribution of lithology in the valley fill is presumed to be related to the proximity of the outcrop as well as degree and orientation of slope below the outcrop. Reliable subsurface lithologic data for the valley fill in the study area is sparse.

The southeastern quadrant of the study area (fig. 1) has been mapped geologically by Denney and Drewes (1965). They differentiate playa deposits of salt, sand, silt, and clay from coarser valley fill; the playa deposits constitute more than one-half the valley fill material. These playa deposits have been dissected by later erosion and replaced in places by coarser, more permeable material.

Denney and Drewes (1965, p. 32-33) as well as Hunt (1960) further conclude that little disturbance of the valley-fill material has occurred during the last 2,000 years and they propose that the most recent period constituting significant erosion and deposition occurred in late Pleistocene. During that period, the playa deposits were dissected " *** to depths of at least 100 feet. Washes from volcanic mountains to the north spread a layer of gravel on top of the playa deposit *** ."

Hydrologic significance of the lithology and distribution of valley-fill material is twofold: (1) The effect that different materials have on distribution of horizontal and vertical hydraulic conductivity; and (2) the chemical quality that results from the interaction of water with different materials. Approximate areal distribution of valley-fill material by lithologic character is shown in figure 1. Outcrops of carbonate and other clastic rocks also are shown in figure 1. From a consideration of chemical reactivity, these rocks represent the major exception to a primarily tuffaceous bedrock and valley fill north of the Amargosa Desert. The region

of tuffaceous bedrock or valley fill has been estimated from consideration of outcrop lithology (Walker and Eakin, 1963; Denney and Drewes, 1965), slope, major drainage, drill-hole data (either drillers' logs or lithologic logs when available) (Walker and Eakin, 1963; F. M. Byers, U.S. Geological Survey, written commun., 1972), and water-quality data presented in this report.

Hydrology

The most detailed previous discussion of hydrology of the west-central part of the Amargosa Desert has been by Walker and Eakin (1963), although their study area was larger than the area considered in this report. They present a map of 1962 water-level altitude contours for the valley-fill aquifer for the part of the area containing the greatest density of wells. Measurements made in 1974 indicate that probably not more than 0.3 m of change had occurred during 1962-74 throughout most of the area. A summary of their data is given in figure 2, modified to include information collected since their study. Water-level altitude contours show that the gradient is parallel to the surface-drainage pattern. Geologic, hydrologic, and chemical data indicate that a confined or semiconfined aquifer controls the potential distribution east of the Gravity fault (figs. 1 and 2). This is the regional carbonate aquifer described by Winograd and Thordarson (1975). Hydraulic potential of the confined aquifer does not vary much throughout several townships in this area. The effect of the degree of confinement is evident at several locations. Some wells that are completed in the overlying alluvium have water levels similar to those of the confined aquifer, while other wells have significantly lower water levels. These lower values indicate that the degree of confinement in this area is significant.

The lack of potentiometric discontinuities west of the Gravity fault and north of the Ash Tree Spring (site 20) and the presence of numerous large springs just east of the fault indicate that the confined aquifer does not extend west of the Gravity fault into the west-central Amargosa Desert. However, the serpentine character of the 698-m contour in T. 16 S., R. 49 E. could be explained by leakage from the confined aquifer across the fault. This possibility is explored further in discussions of geochemistry of the ground water later in this report.

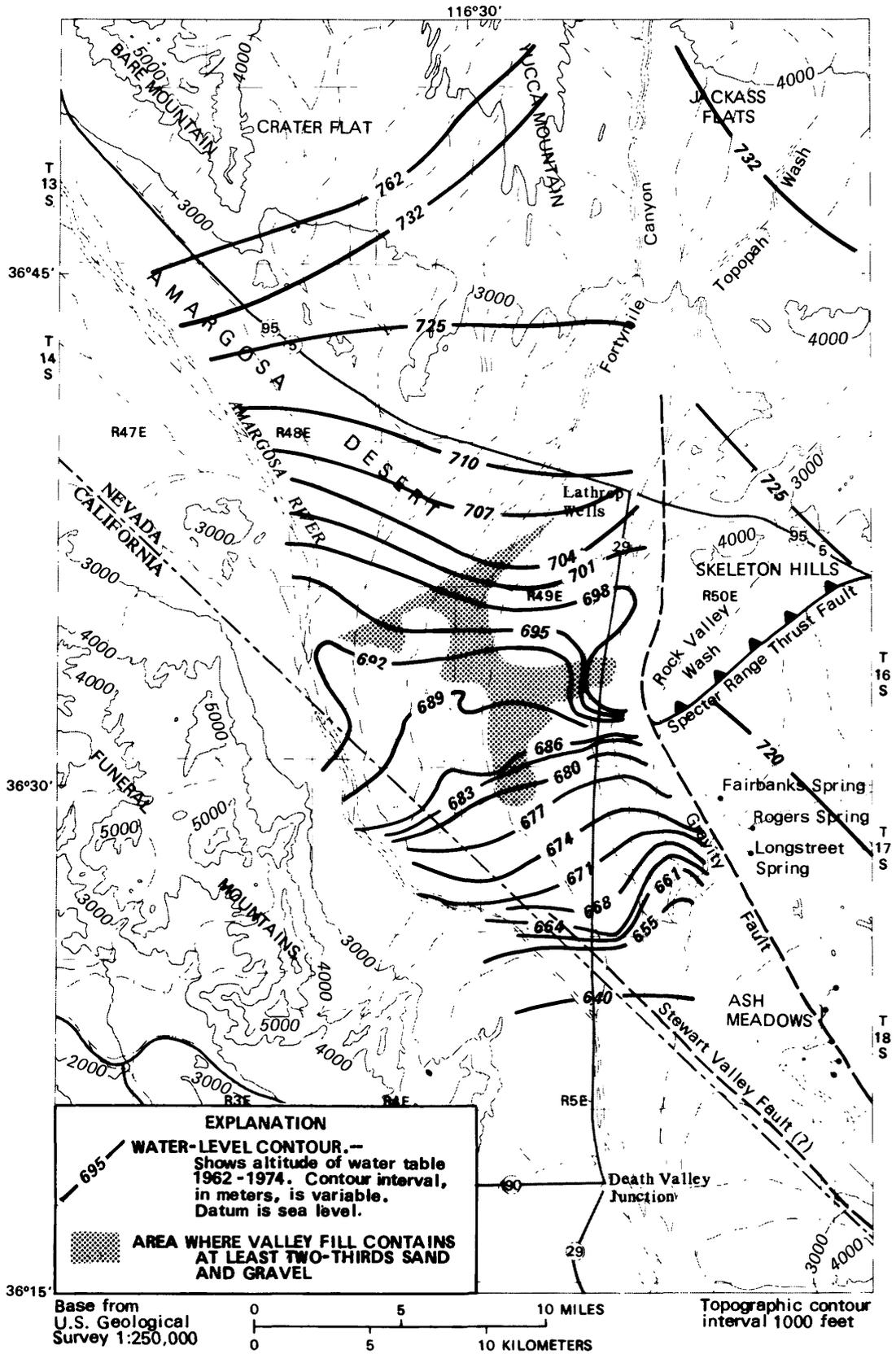


Figure 2.--Water-level altitudes (modified from Walker and Eakin, 1963).

Ground-water potential gradient distribution is another feature of the potentiometric map as it relates to the water-quality data that are discussed in more detail in a later section. The southernmost region has a gradient of about 4.7×10^{-3} ; the central region near E $\frac{1}{2}$ T. 16 S., R. 48 E., W $\frac{1}{2}$ T. 16 S., R. 49 E. has a gradient of about 1.3×10^{-3} ; the northwestern region has a gradient of about 2.9×10^{-3} ; and the northeastern region, in Fortymile Canyon, has a gradient of about 1.3×10^{-3} . Regions with less steep gradients probably represent more permeable or thicker aquifer materials; regions with steeper gradients represent less permeable or thinner aquifer materials. Although no permeability data exist for the area, a plot of the ratio of sand plus gravel divided by silt plus clay, using drillers'-log data from wells in the study area that were completed in the valley fill, provided corroboration for the relationship of gradient and permeability. A line of sand + gravel/silt + clay = 2 (arbitrarily chosen) is shown in figure 2 (William Thordarson, U.S. Geological Survey, written commun., 1975). The area within this line includes valley fill containing at least two-thirds sand and gravel, suggesting relatively high permeability. This area generally correlates well with the presence of both smaller ground-water potential gradients and present-day surface drainage, as indicated by the ephemeral stream channels on the topographic map base. It also is consistent with the limited well-performance data given in Walker and Eakin (1963, table 3). The east-trending knob beginning in the center of T. 16 S., R. 49 E. and ending near the mapped fault coincides with a significant change in orientation of the potentiometric contours, referred to earlier as the locus of possible leakage from the confined aquifer east of the fault.

The parts of the study area with larger and much smaller hydraulic gradients may be explained in several ways. Where no significant withdrawal of water is made, such as in T. 15 S., R. 48 and 49 E., the steepening of contours southward simply may reflect the change in saturated thickness of valley fill. The greater density of pumping present in T. 16 S., R. 48 E., however, may explain the very slight gradients in this township. This area had the greatest withdrawal rate of any of the townships in the study area from 1971 to 1974. In the northern part of the study area (Crater Flat and Yucca Mountain), ground water exists in a bedrock aquifer, in contrast to

the valley-fill aquifer of the central part. Permeability of this bedrock aquifer depends on the degree of fracturing of the bedrock; fracturing is greater in the upper reaches of Fortymile Canyon than at Crater Flat. This statement is supported by comparison of relative specific capacities of wells J-12 [$1,800 \text{ (m}^3/\text{d)/m}$ (cubic meters per day per meter)], J-13 [$400 \text{ (m}^3/\text{d)/m}$] (Claassen, 1973), and that of well VH-1 [$350 \text{ (m}^3/\text{d)/m}$] (L. W. Howells, U.S. Geological Survey, written commun., 1981). The steepening contours near the intersection of State Highway 29 and the Nevada-California State line have been explained by Winograd and Thordarson (1975, p. C-78) as being due to the presence of Tertiary rocks with little permeability and thinning of the valley-fill aquifer.

GROUND-WATER GEOCHEMISTRY

Data from water samples collected by the U.S. Geological Survey in 1971, 1974, and 1979 were used to determine sources and mechanisms of recharge of ground water in the west-central Amargosa Desert (table 1, fig. 3). Samples collected by other agencies and not published previously, as well as previously published data, are included in table 1. Location numbers are keyed to table 1 and figure 3 and are used throughout the text.

Ground water of the west-central Amargosa Desert generally may be divided into water that reflects: (1) Reaction with tuffaceous rocks or tuffaceous valley fill; and (2) water that has reacted with primarily carbonate rocks or carbonate valley fill; and (3) water reacting with a mixed lithology of carbonate and tuffaceous material. Other rocks, if present, are assumed to be considerably less reactive or of such low abundance that they contribute little to the water chemistry. Carbonate-derived waters may enter a region of tuffaceous valley fill (or vice versa), and the resulting water quality is indistinguishable from that resulting from reaction in the valley fill of mixed lithology.

In figure 1, the area in the center of the map is tuff outcrop and tuffaceous valley fill. The outcrops have been mapped by Walker and Eakin (1963) and Denney and Drewes (1965); composition of the valley fill has been inferred by proximity to outcrop, and degree and direction of slope away from outcrop and from the lithologic log of a deep test hole drilled by the

Table 1.--Water quality analyses

[m, meters; °C, degrees Celsius; mmol/L, millimoles per liter; Ca²⁺, calcium ion; Mg²⁺, magnesium ion; Na⁺, sodium ion; K⁺, potassium ion; HCO₃⁻, bicarbonate ion; Cl⁻, chloride ion; SO₄²⁻, sulfate ion; SiO₂, silica]

Number (1-60 in fig. 3; 61-71 in fig. 10)	Site designation or name	Approx- imate well depth (m)	Approx- imate depth to water (m)	Sample collec- tion date	Source of data	Water temper- ature (°C)	Dissolved constituents (mmol/L)						Onsite pH		
							Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	HCO ₃ ⁻	Cl ⁻		SO ₄ ²⁻	SiO ₂
1	J-13	1,060	283	3-26-71	USGS-1	31	0.30	0.09	1.83	0.13	2.03	0.20	0.18	0.95	7.2
2	J-12	350	226	3-26-71	USGS-1	26.8	.35	.09	1.65	.13	1.95	.21	.23	.90	7.1
3	15S/49E-22dc	150	78	11-20-72	USGS-2	-----	.67	.08	1.87	.12	2.44	.24	.34	.82	7.78
4	16S/49E-5acc	90	21	3-04-74	USGS-2	-----	.72	.09	1.52	.13	2.21	.17	.27	1.03	8.15
5	16S/49E-8abb	60	45	11-17-72	USGS-2	23	.75	.11	1.61	.14	2.49	.22	.31	.90	7.47
6	16S/49E-8acc	90	45	4-01-71	USGS-2	25.8	.57	.10	1.61	.17	2.26	.17	.30	.97	7.90
7	16S/49E-9cda-2/	90	46	8-19-62	W&E	24	.76	.14	2.22	.22	2.35	.34	.67	1.09	7.61
8	16S/49E-9dcc	60	49	3-01-74	USGS-2	23.3	.57	.11	2.44	.23	2.31	.28	.70	1.20	8.16
9	16S/49E-18dc	110	33	3-01-74	USGS-2	-----	.50	.11	1.83	.23	2.46	.21	.29	.98	8.12
10	16S/49E-16ccc	-----	---	6-26-79	USGS-3	-----	.75	.08	1.73	.11	2.17	.23	.53	1.28	7.87
11	16S/49E-19daa	90	30	3-05-74	USGS-2	26.4	.60	.05	1.57	.21	2.20	.19	.34	1.25	8.20
12	16S/48E-24aaa	150	29	11-17-72	USGS-2	27	.45	.03	2.35	.18	2.41	.22	.31	1.31	8.09
13	16S/48E-25aaa	50	26	3-05-74	USGS-2	26.5	.47	.03	1.87	.19	2.18	.26	.29	1.20	8.06
14	16S/48E-36aaa	50	21	3-04-74	USGS-2	-----	.42	.08	1.74	.16	2.18	.19	.26	1.31	8.40
15	17S/48E-1abb	60	16	3-05-74	USGS-2	-----	.47	.06	1.74	.18	2.21	.18	.26	1.31	8.15
16	17S/49E-7bb	150	12	3-01-74	USGS-2	-----	.60	.07	2.09	.19	2.51	.27	.32	1.33	8.30
17	17S/49E-9aa	-----	---	3-01-74	USGS-2	-----	.62	.15	2.09	.25	2.15	.28	.72	1.17	8.02
18	17S/49E-8ddb	100	15	3-06-74	USGS-2	24.0	.52	.11	1.57	.19	2.02	.18	.28	1.35	8.35
19	17S/49E-15bbd	110	17	3-06-74	USGS-2	22.5	.52	.16	1.36	.21	1.97	.28	.36	1.21	8.12
20	17S/49E-35ddd-2/ (Ash Tree Spring)	0	0	5-08-52	W&E	23	.38	.19	2.20	.205	2.58	.19	.42	1.345	7.96
21	16S/49E-23add	-----	---	6-25-79	USGS-3	-----	.40	.07	2.43	.165	2.08	.25	.36	1.27	8.19
22	16S/48E-15dda	-----	---	6-26-79	USGS-3	-----	.50	.24	3.08	.19	2.88	.49	.39	1.19	7.97
23	16S/48E-15aaa	50	29	3-31-71	USGS-2	25.5	.24	.13	2.52	.15	2.51	.21	.29	1.13	8.1
24	16S/48E-23bdb	100	29	8-19-62	W&E	24	.23	.04	2.87	.17	2.56	.25	.28	1.23	3/7.3
25	16S/48E-10cba	-----	---	3-31-71	USGS-2	24.5	.23	.16	2.65	.14	2.72	.23	.34	1.07	8.3

Table 1.--Water quality analyses--Continued

Number (1-60 in fig. 3; 61-71 in fig. 10)	Site designation or name	Approx- imate well depth (m)	Approx- imate depth to water (m)	Sample collec- tion date	Source of data ^{1/}	Water temper- ature (°C)	Dissolved constituents (mmol/L)						Onsite pH		
							Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	HCO ₃ ⁻	Cl ⁻		SO ₄ ²⁻	SiO ₂
26	16S/48E-36dcc	120	13	4-05-71	USGS-2	26.0	1.37	0.40	4.35	0.33	4.92	0.93	1.15	1.17	7.2
27	16S/50E-7bcd ^{2/}	60	43	8-18-62	W&E	----	----	----	----	----	----	----	----	----	----
28	16S/49E-12ddd	----	----	4-01-71	USGS-2	30.6	1.19	.72	4.85	.33	4.78	.82	1.58	.48	7.62
29	16S/49E-15aaa	120	51	6-26-79	USGS-3	----	1.14	.70	5.22	.11	4.73	.68	1.67	.34	7.64
30	16S/49E-36aaa	----	----	3-31-71	USGS-2	23.8	1.02	.31	3.48	.25	3.20	.65	1.35	.77	7.7
31	16S/49E-36aba	----	----	6-24-79	USGS-3	----	1.30	.91	5.22	.46	5.15	.76	1.75	.63	7.76
32	16S/49E-35aaa	----	----	6-24-79	USGS-3	----	1.12	.82	4.79	.43	4.80	.68	1.62	.71	7.65
33	16S/49E-35baa ^{2/}	100	26	6-24-79	USGS-3	24	1.10	.66	5.22	.41	4.45	.82	1.54	.61	7.65
34	15S/50E-18cdc	----	----	8-18-62	W&E	----	1.33	.74	4.92	.34	4.96	.88	1.77	.63	7.41
35	15S/50E-18ccc	----	----	6-24-79	USGS-2	25.1	.30	.03	4.05	.10	2.51	.48	.81	.63	3/8.00
36	17S/49E-11ba	----	----	11-17-72	USGS-2	----	.42	.02	4.05	.10	2.57	.37	1.04	.57	8.37
37	16S/48E-15ba	50	30	3-07-74	USGS-2	22	1.00	.58	4.22	.36	3.44	.79	1.67	.88	3/8.1
38	17S/49E-15bc	----	----	6-24-71	USBR	25	1.50	.32	6.40	.25	4.33	1.85	2.07	.62	8.0
39	17S/49E-28bcd	----	----	2-17-66	USBR	24	.54	.04	1.70	.17	2.00	.30	.29	----	8.2
40	18S/49E-1aba	0	0	6-23-79	USGS-3	----	1.07	.41	4.35	.31	4.83	.68	.93	1.17	7.56
41	18S/49E-2cbc	160	22	4-06-71	USGS-2	17.5	.60	.49	4.13	.49	4.31	.51	1.04	1.21	8.6
42	18S/49E-11bbb	----	----	11-18-72	USGS-2	23.8	.72	.49	5.22	.25	5.77	.56	.77	.98	7.82
43	18S/50E-6dac	----	----	6-23-79	USGS-3	25.0	.85	.35	4.31	.30	3.68	.87	.94	1.30	7.56
44	17S/49E-29acc	----	----	2-12-72	DRI	----	.59	.49	4.46	.35	3.77	.58	1.11	1.34	8.22
45	16S/48E-8ba	80	34	6-23-79	USGS-3	21.0	1.35	.62	6.96	.51	4.52	1.97	1.94	1.20	7.60
46	16S/48E-7bba	----	----	6-24-71	USBR	25	1.46	.26	7.85	.33	4.85	2.25	2.11	.63	7.9
47	16S/48E-7cbc	----	----	3-30-71	USGS-2	24.7	1.32	.39	6.09	.26	4.11	1.78	1.87	1.15	7.4
48	16S/48E-18bcc	----	----	3-31-71	USGS-2	24.2	1.17	.66	5.66	.24	3.92	1.75	1.87	1.07	7.7
49	16S/48E-17ccc	----	----	6-24-79	USGS-3	----	1.37	.45	6.53	.30	4.45	1.72	1.98	1.33	7.98
50	16S/48E-18dad	----	----	6-25-79	USGS-3	----	1.65	.45	7.39	.31	3.92	2.34	2.45	1.29	7.69
		----	----	6-25-79	USGS-3	----	1.32	.35	6.52	.27	3.87	1.78	1.95	1.28	7.69

Table 1.--Water quality analyses--Continued

Number (1-60 in fig. 3; 61-71 in fig. 10)	Site designation or name	Approx- imate well depth (m)	Approx- imate depth to water (m)	Sample collec- tion date	Source of data ^{1/}	Water temper- ature (°C)	Dissolved constituents (mmol/L)						Onsite pH		
							Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	HCO ₃ ⁻	Cl ⁻		SO ₄ ²⁻	SiO ₂
51	16S/48E-8cda	---	---	3-31-71	USGS-2	23.3	1.20	0.28	6.96	0.26	4.33	1.89	1.87	1.13	7.6
52	16S/48E-17abb	90	31	8-18-62	W&E	24	1.50	.32	6.83	.31	4.95	1.95	1.86	1.25	3/7.4
53	16S/48E-23da	140	24	2-17-66	USBR	24	.55	.09	3.00	.17	2.20	.75	.70	----	8.2
54	27N/4E-27bbb	90	14	8-18-62	W&E	22	1.45	.78	5.83	.49	7.18	.90	1.11	1.20	3/7.8
55	VH-1	760	184	2-11-81	USGS-4	35.5	.25	.06	3.48	.05	2.60	.28	.47	.83	7.5
56	H-1	1,830	572	10-01-80	USGS-5	33	.11	.00	2.22	.06	1.88	.16	.19	.78	7.7
57	UE25B-1	1,220	471	9-01-81	USGS-6	36	.42	.02	2.00	.09	2.20	.24	.23	.87	7.5
58	17S/50E-19aab	-----	---	12-27-71	DRI	16	.19	.35	10.96	.70	6.81	1.97	1.83	.71	8.60
59	18S/50E-7aa	-----	---	12-28-71	DRI	13	.64	.39	6.13	.49	4.28	1.06	1.53	.79	8.4
60	Nuclear Engr. Co. Well 1	180	86	12-16-68	USGS-2	----	1.37	.58	7.40	.26	5.38	2.23	1.98	1.17	7.6
61	UE19e ^{2/}			3-19-71											
62	Whiterock Spring	1,830	676	10-06-71	USGS-1	31.2	.00	.01	1.61	.02	1.33	.12	----	.97	7.9
63	TWI	170	125	4-10-72	USGS-2	----	.10	.01	1.91	.20	1.26	.31	.28	.77	7.3
64	Well 8, Nevada Test Site	1,680	328	10-01-60	USGS-2	----	.05	.00	1.57	.03	1.34	.20	.12	.65	7.6
65	UE18r	1,530	418	3-24-71	USGS-1	26.5	.21	.05	1.35	.09	1.31	.21	.15	.68	7.4
66	UE19g-s ^{2/}			1-29-68	USGS-7	32.2	.65	.04	3.52	.08	4.13	.22	.25	.75	3/8.0
67	UE19b-1	2,290	623	3-20-71											
68	Butte Spring	1,370	645	10-06-71	USGS-1	42.0	.97	.01	3.09	.02	2.98	.27	1.04	.86	8.0
		1,370	645	10-13-64	USGS-7	31.6	.60	.10	1.83	.08	2.46	.19	.22	.68	3/7.4
		-----	---	11-10-60	USGS-2	15.0	.52	.10	1.48	.15	2.28	.34	.12	.50	7.8

Table 1.--Water quality analyses--Continued

Number (1-60 in fig. 3; 61-71 in fig. 10)	Site designation or name	Approx- imate well depth (m)	Approx- imate depth to water (m)	Sample collec- tion date	Source of data ^{1/}	Water temper- ature (°C)	Dissolved constituents (mmol/L)						Onsite pH	
							Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	HCO ₃ ⁻	Cl ⁻		SO ₄ ²⁻
69	Rainier (typical sample)	-----	---	-----	USGS-8	23	0.37	0.13	0.99	0.12	2.16	-----	-----	8.0
70	Oak Spring	-----	---	4-28-58	USGS-9	12.8	.45	.20	.96	.16	1.90	.25	.15	.95
71	Topopah Spring	-----	---	3-25-58	USGS-9	11.7	.18	.04	.61	.16	.79	.08	.16	.83

^{1/}Reference key to data source: USGS-1 = Claassen (1973); USGS-2 = H. C. Claassen (unpublished data, 1960, 1968, 1971, 1972, 1974); USGS-3 = A. F. White (Lawrence Barkley Laboratories, written commun., 1980); W&E = Walker and Eakin (1963); USGS-4 = L. W. Howells (U.S. Geological Survey, written commun., 1981); USGS-5 = E. F. Rush and William Thordarson (U.S. Geological Survey, written commun., 1980); USGS-6 = M. S. Whitfield (U.S. Geological Survey, written commun., 1981); DRI = Naff (1973); USBR = U.S. Bureau of Reclamation (written commun., 1974), courtesy of E. A. Lundberg; USGS-7 = Blankennagel and Weir (1973); USGS-8 = Claassen and White (1979); and USGS-9 = Moore (1961).

^{2/}Average of two analyses.

^{3/}Laboratory pH.

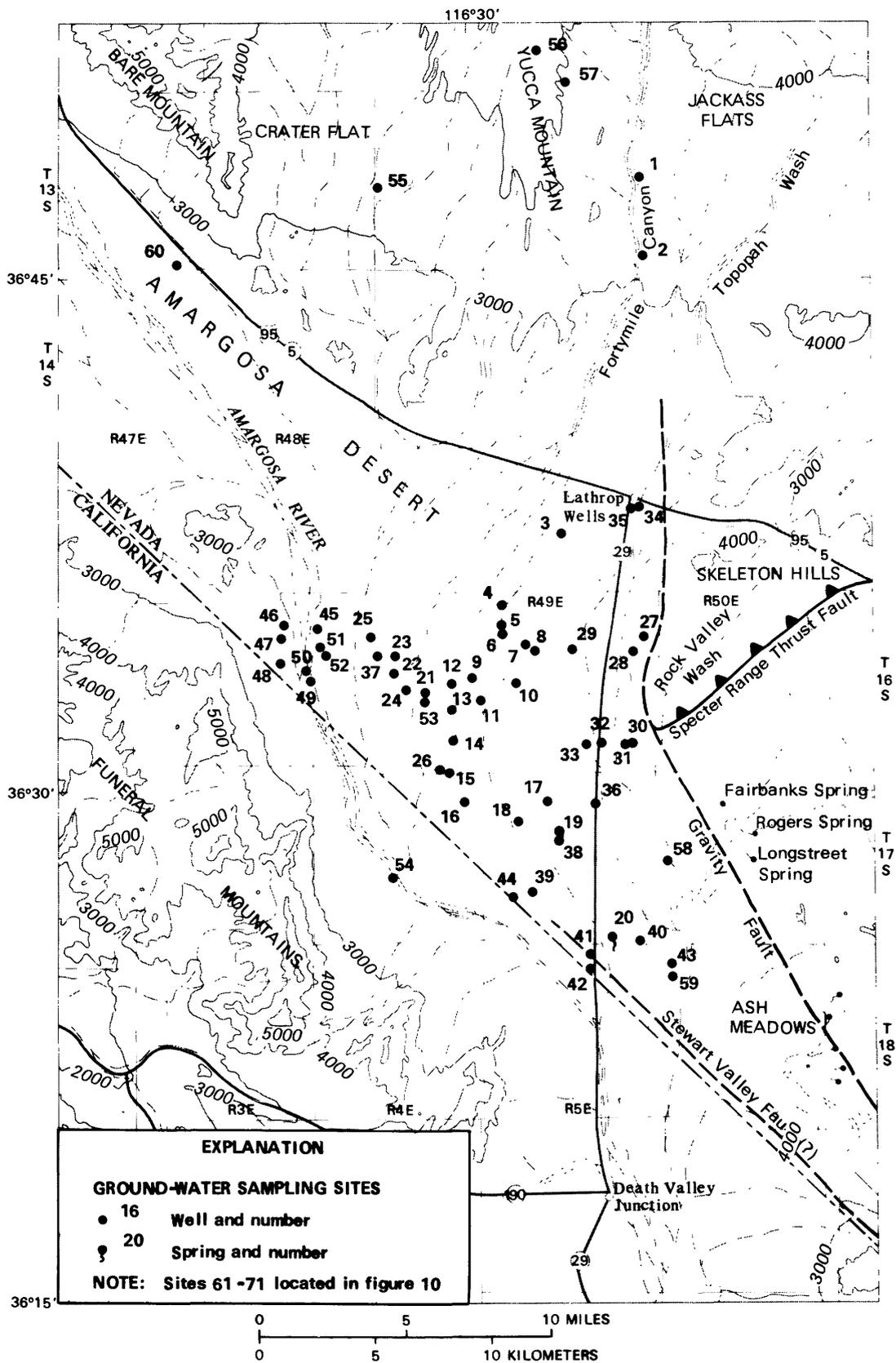


Figure 3.--Location of ground-water sampling sites.

U.S. Bureau of Reclamation in NE $\frac{1}{4}$ sec. 27, T. 15 S., R. 49 E. about 5 km southwest of Lathrop Wells, Nevada (F. M. Byers, Jr., U.S. Geological Survey, written commun., 1972). Bedrock is composed primarily of limestone, dolomite, quartzite, and occasionally shale; the remaining valley fill is composed primarily of limestone, dolomite, and quartzite detritus. Some tuffs probably occur near the tuffaceous valley fill boundary. Quaternary playa deposits have been mapped by Denney and Drewes (1965) in the southeastern corner of the study area; these deposits may be present in other areas farther north, but at greater depths. This possibility is indicated by the presence of such deposits at a depth of about 405 m in the deep test hole. Presence of these deposits in the zone of saturation generally has a significant effect on water quality, as will be discussed later.

Maps of the concentrations of the four major ions in the ground water, sodium (Na^+), calcium (Ca^{2+}), bicarbonate (HCO_3^-), and sulfate (SO_4^{2-}), are shown in figures 4 through 7. The sodium map is similar to one published by Naff (1973) but is more detailed. All these maps show a central region of small values of each ion bounded on the west by a very steep concentration gradient and on the east by a less steep, but still prominent, gradient. Gradients at the south end are intermediate. Smaller gradients generally are coincident with more permeable sands and gravels (fig. 2); they also are aligned with the major drainageways leading from Fortymile Canyon. The reason for the almost 90° change in direction of the surface drainage from Fortymile Canyon to the south end of the system may be related to an extension of the Stewart Valley fault (fig. 1) or to a thicker sequence of less permeable playa deposits in the south end of the system. The geologic map presented by Denney and Drewes (1965) includes a part of T. 17 S., R. 49 E., within which bouldery-to-pebbly gravel, pebble gravel, and sand and playa deposits were mapped. Coarser materials (bouldery-to-pebbly gravel) coincide primarily with present-day stream channels; pebble gravel and sand commonly occupy the areas between present-day stream channels and undissected playa deposits. Dissolved solids of seven water samples within this township correlate well with surface-mapped materials (table 2).

Although small dissolved-solids concentrations are related to coarser, more permeable materials, both surficially and in the subsurface, additional

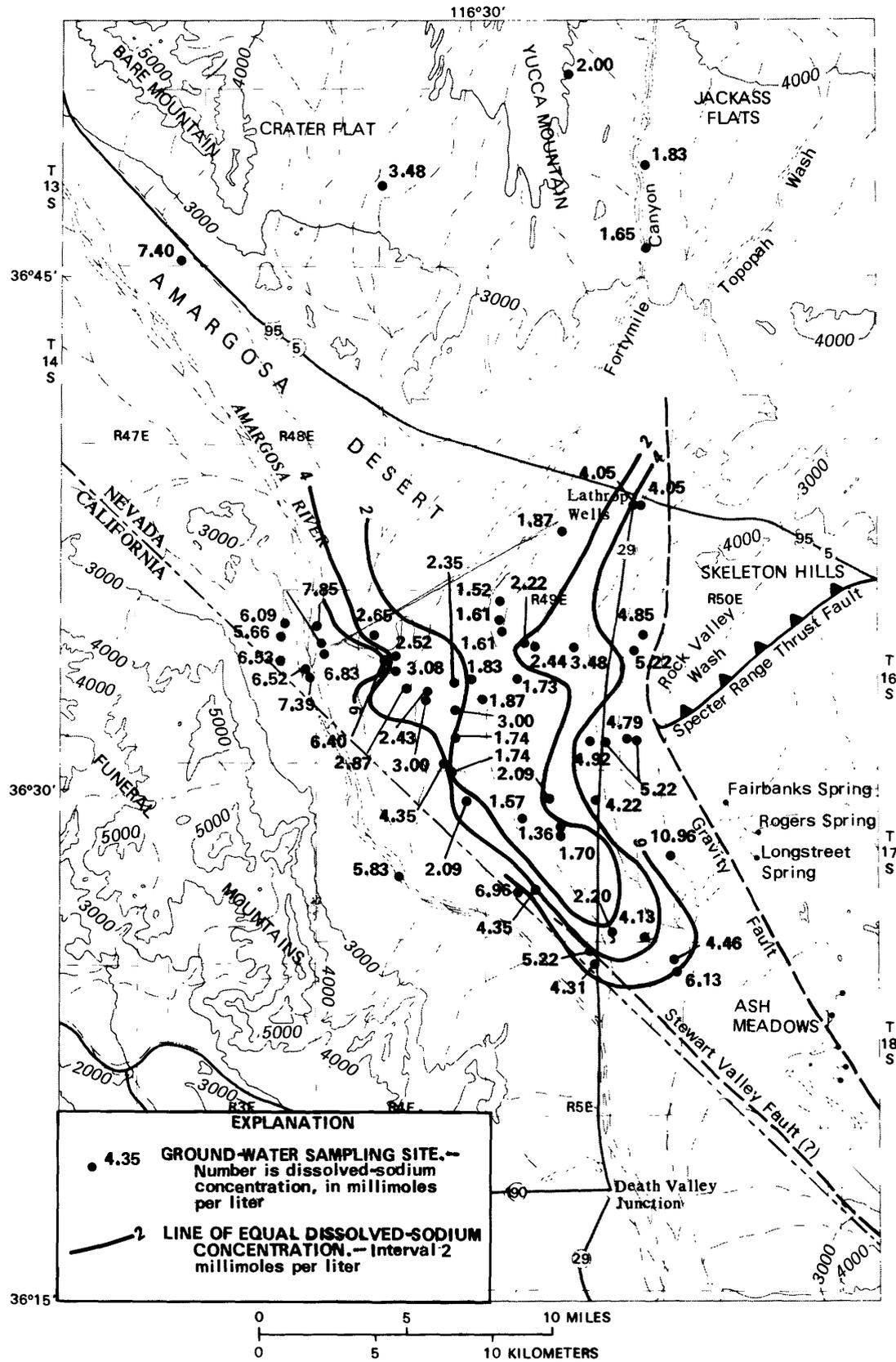


Figure 4.--Dissolved sodium (Na^+) in ground water.

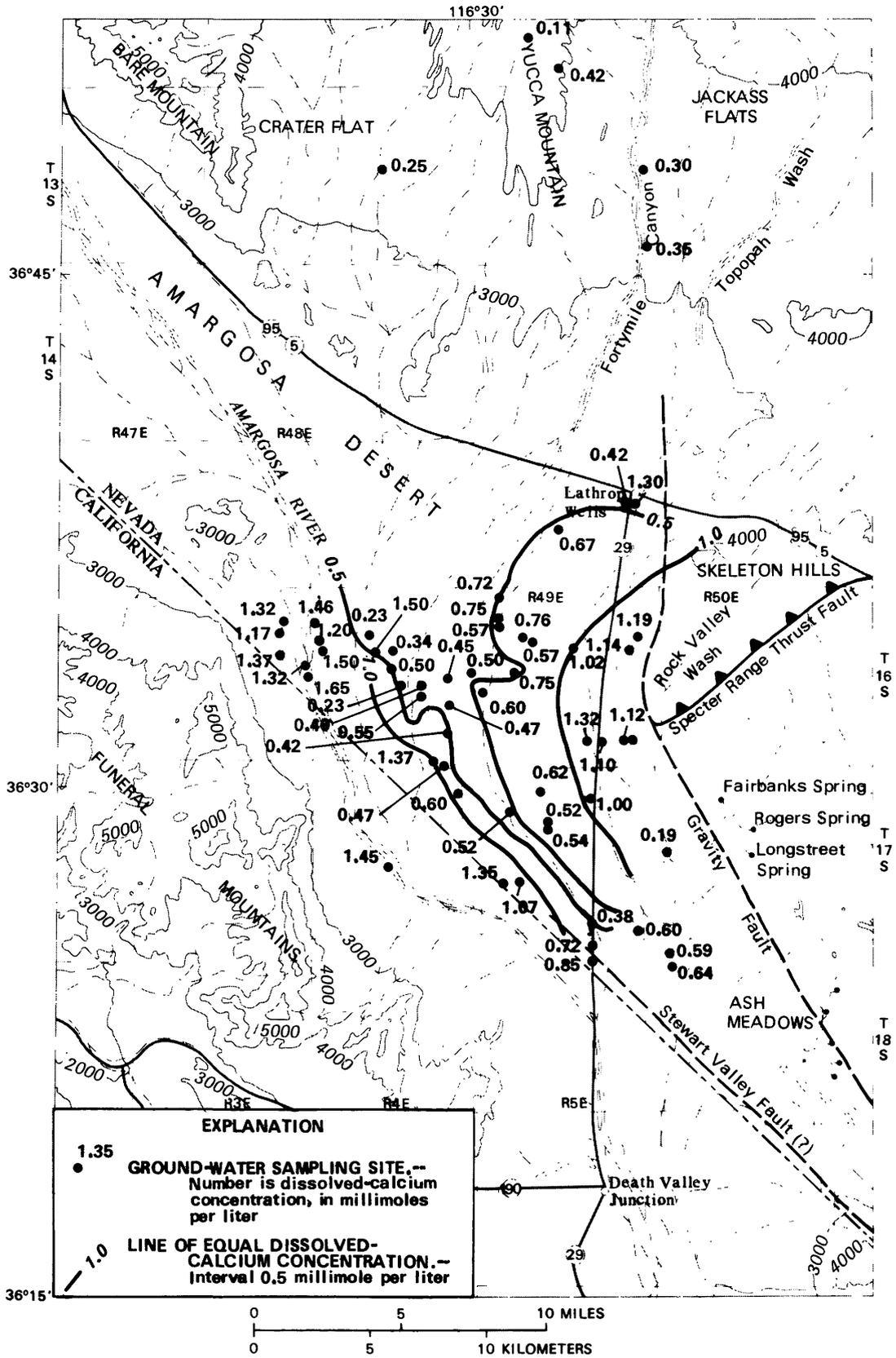


Figure 5.--Dissolved calcium (Ca^{2+}) in ground water.

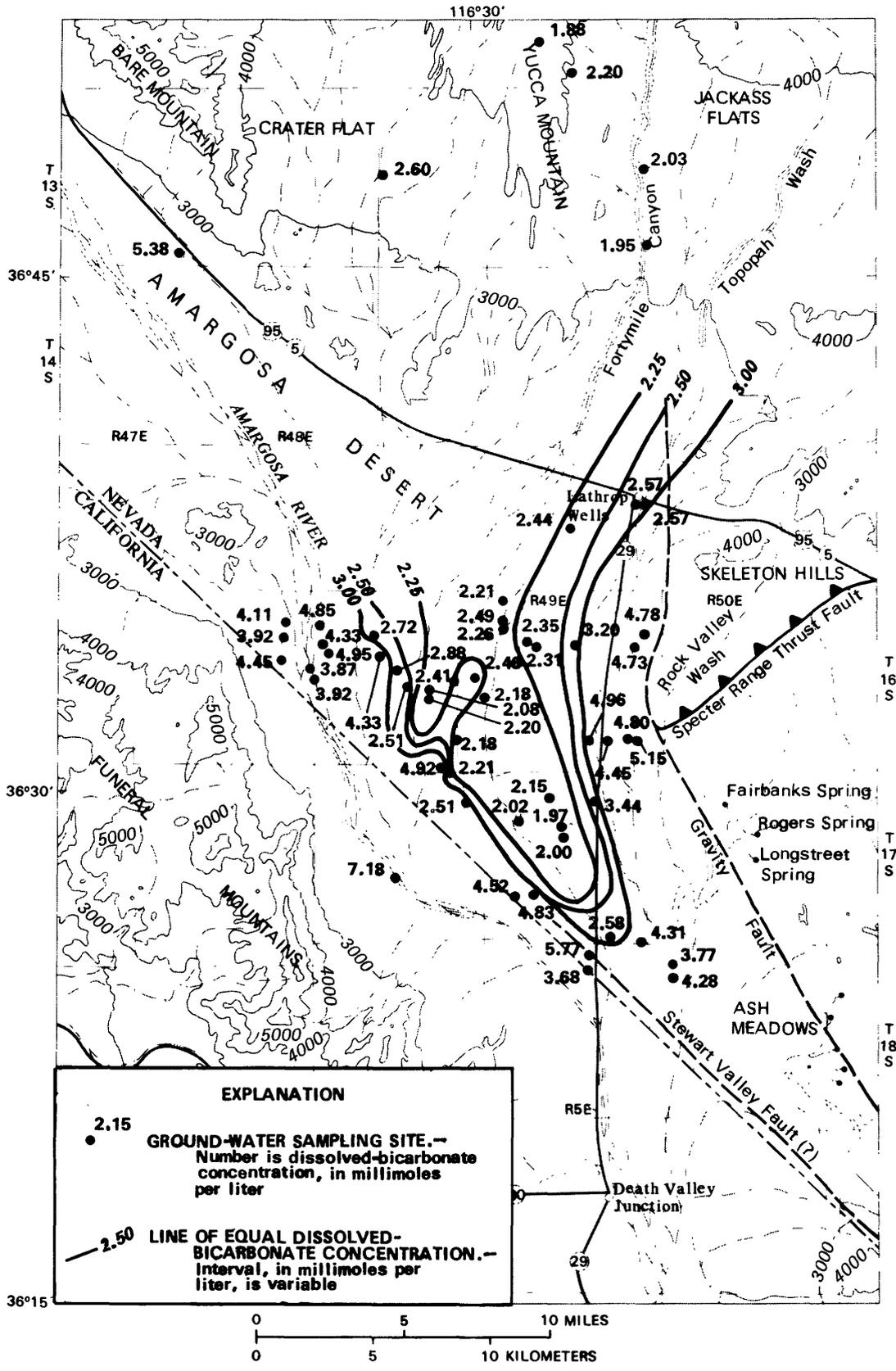


Figure 6.--Dissolved bicarbonate (HCO_3^-) in ground water.

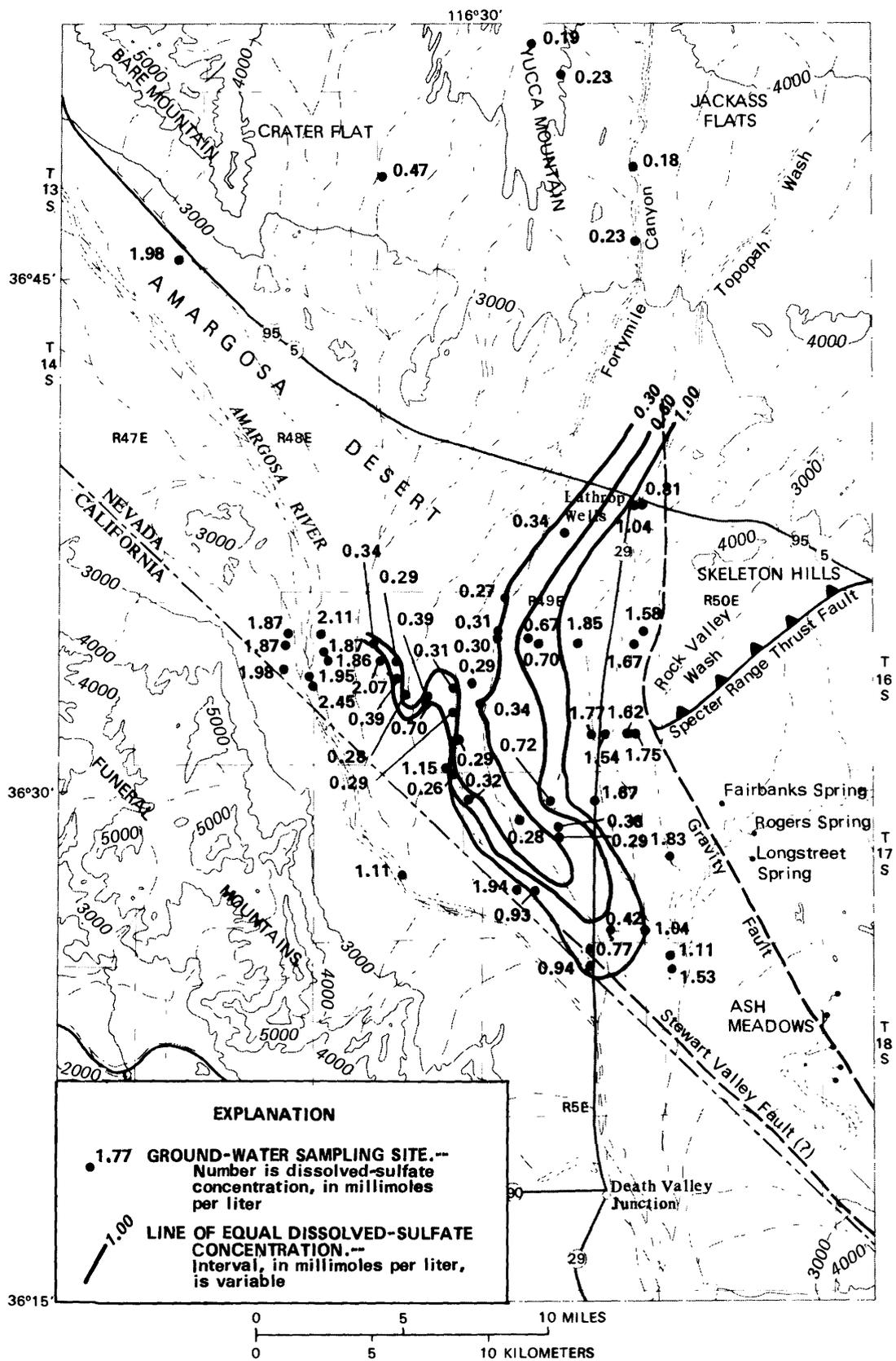


Figure 7.--Dissolved sulfate (SO_4^{2-}) in ground water.

Table 2.--*Relationship between dissolved solids in ground water and surface geology in a part of T. 17 S., R. 49 E.*

Site	Surficial material	Calculated dissolved solids (milligrams per liter)	Comments
19	Bouldery-to-pebbly gravel	302	In stream channel.
18	Pebble gravel and sand	305	Near channel.
17	do.	366	-----
16	do.	355	-----
36	do.	616	Close to both stream channel and playa deposits.
44	Playa deposits	853	-----
39	do.	643	-----

information is needed to determine the mechanism of recharge (source) of ground water in the study area. The central part of the area consists of tuffaceous-rock fragments containing abundant glass. The chemical character of water in contact with both vitric and devitrified tuffaceous-rock fragments primarily reflects results of reaction with vitric material (White and Claassen, 1979, 1980; Claassen and White, 1979; White and others, 1980). Water in the tuffaceous valley fill of the west-central Amargosa Desert might be expected to have a similar chemical character. However, the source of this water needs to be considered. The effects of reaction with tuffaceous valley fill probably are indistinguishable from those of reaction with bedrock; therefore, distinction between recharge directly to valley fill via surface runoff versus lateral flow from bedrock recharge areas to the north might not be possible. Water quality in tuffaceous valley fill and bedrock aquifers north of the study area was examined to determine if there were any characteristics that would permit its differentiation from ground water in the west-central Amargosa Desert.

Water Quality in the Tuffaceous Aquifers

If the source of water in tuffaceous alluvium in west-central Amargosa Desert is bedrock aquifers north of the study area, its chemical composition should reflect a stage of development equal to or beyond that of the bedrock aquifers. The change in water quality with time in the vitric tuffs of Rainier Mesa, Nevada, was described by Claassen and White (1979) and is summarized below:

1. Recharging water obtains carbon dioxide (CO_2) by some nonequilibrium process.
2. Reaction of dissolved CO_2 with vitric tuff occurs by both ion-exchange and ion-diffusion processes.
3. Contemporaneous with 2. above, chemical precipitation of authigenic phases occurs if suitable surfaces are available as nucleation sites.

In tuffs of the Nevada Test Site, north of the study area, from which the valley-fill materials in the study area were derived, two precipitating phases probably are primarily responsible for development of the observed ground-water compositions. These are montmorillonite, which probably precipitates in all vitric ground-water systems, and clinoptilolite, which

precipitates in quantities from none to about three times greater than the montmorillonite. The change in solution cation composition with time, or during the interval when ground water flows from its recharge location to the sampling point in the aquifer, is shown in figure 8 for several waters typical of the bedrock aquifers of the Nevada Test Site. Composition of the samples depends on the quantity of clinoptilolite precipitated relative to the quantity of montmorillonite (fig. 8). Greater percentages of Na^+ in ground water are associated with greater clinoptilolite/montmorillonite ratios. The Rainier, Topopah Spring, Butte Spring, and Oak Spring samples, representing perched ground water that has undergone montmorillonite precipitation only, have been included as examples; however, most bedrock water samples have undergone clinoptilolite precipitation also.

If recharge to the west-central Amargosa Desert occurs in the highlands north of the study area, Na^+ probably would comprise greater than 70 percent of the three major cations (Na^+ , Ca^{2+} , and Mg^{2+}) in ground water. This composition would result from infiltrating precipitation or surface runoff reacting with primarily vitric tuff, and precipitation of montmorillonite and clinoptilolite in varying quantities, depending on the lithology and flow path. The total concentration of dissolved material in the ground water depends on the conditions of recharge. If direct infiltration is recharged, temperature and vegetation are most important; if surface runoff is recharged, the area of ground surface in contact with unit volume of overland flow is most important. The foregoing is true if the ground-water system is closed to entrance of CO_2 after recharge; this condition has been assumed in the calculations and the discussion (unless stated otherwise). Larger concentrations of dissolved CO_2 during recharge result in larger concentrations of dissolved constituents; hydrolysis of dissolved CO_2 yields the hydronium ions that react to produce the solutes. The concentration of dissolved solids also depends on residence time of the ground water, but to a lesser degree than the concentration of CO_2 .

The Na^+ - Ca^{2+} - Mg^{2+} compositions of most of the Amargosa Desert water (in tuffaceous valley fill) are inconsistent with their source in bedrock aquifers to the north (fig. 9). A few of the presumed source water samples (Butte Spring, Rainier, Oak Spring, Topopah Spring; fig. 8) are similar to those in

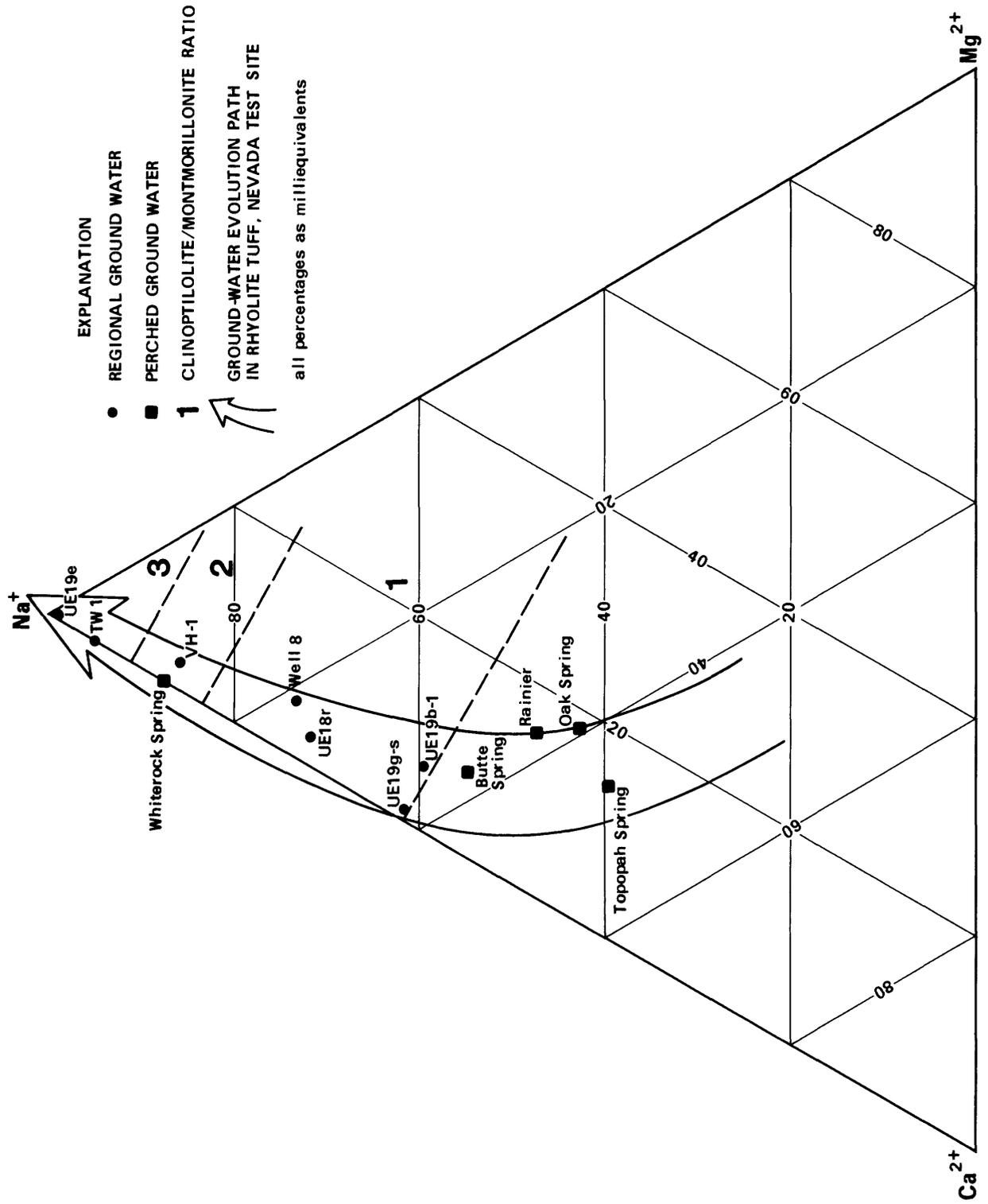


Figure 8.--Evolution of water in tuff, Nevada Test Site.

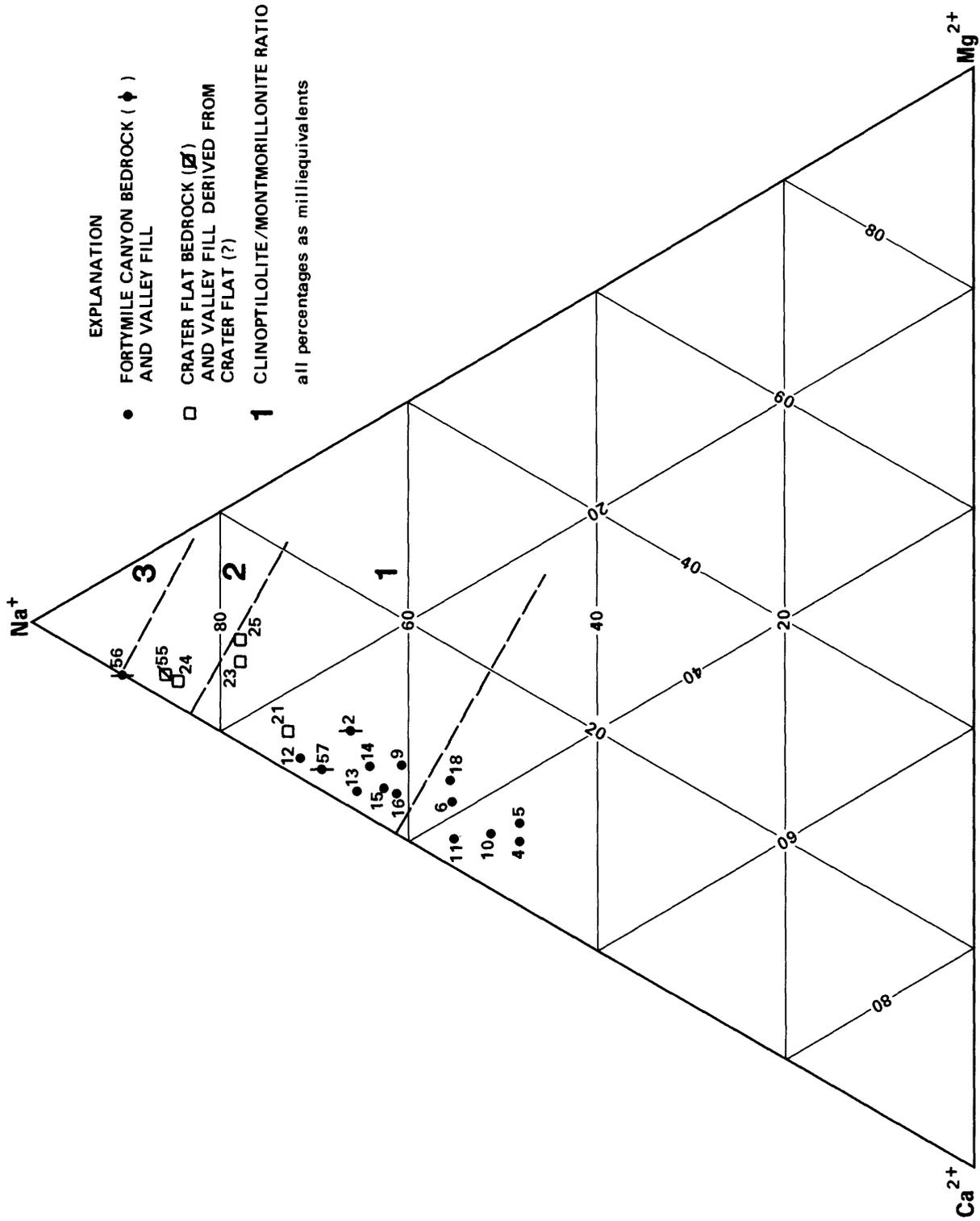


Figure 9.--Composition of water in tuffaceous valley fill.

the Amargosa Desert, but these source waters all represent very young, generally perched ground water atypical of deeper ground water. It is assumed that the deeper ground water (UE19e, TW1, Site 55, Well 8, UE18r) represents the regional water table. These younger waters, named above, generally less than a few years old, have not evolved chemically to the degree that those of the regional aquifer have. In at least two instances (Topopah and Oak Springs), the water source has dried up due to short-term climatic variations, further supporting the contention that they are not representative regionally. However, the sources are representative of a point in the evolution of the composition of waters in tuffs of southern Nevada.

The preceding information indicates that water in the tuffaceous valley fill has not been derived from recharge in the highlands to the north via subsurface flow. A reasonable alternative is that surface runoff directly recharges the valley fill. This source was suggested for sites 1 and 2 first by Schoff and Moore (1964, p. 58) and later for points farther south by Winograd and Thordarson (1975, p. C112), but without detailed supporting evidence beyond the alignment of water containing less dissolved solids with the present-day stream channel. This recharge mechanism allows development of a different reaction extent than that followed by ground water recharged in the highlands to the north. One expected difference between highland recharge and desert runoff is the quantity of CO_2 available for reaction. In highland recharge, direct infiltration requires that the dissolved CO_2 be proportional to the quantity of CO_2 in the soil zone, where it results from soil biological activity; the greater the biological activity, the greater the quantity of CO_2 dissolved. Surface runoff, however, initially contacts only a few millimeters of surface material, which is probably in equilibrium with atmospheric CO_2 . As runoff rivulets coalesce, greater scour occurs, but the biological activity probably is less than in highland areas because of the lack of established vegetation in arid-climate washes. Nevertheless, evidence from water-quality samples of surface runoff collected in arid environments indicates that the quantity of carbon dioxide dissolved during runoff is similar to that dissolved during direct infiltration. Data are from the Nevada Test Site (Alfred Clebsch, Jr., U.S. Geological Survey,

written commun., 1958), New Mexico (U.S. Geological Survey, 1979a), and Nevada (U.S. Geological Survey, 1979b). This condition is necessary because water in the tuffaceous aquifer in the Amargosa Desert contains HCO_3^- in concentrations similar to that in bedrock aquifers to the north. Bicarbonate is a direct result of the reaction of dissolved CO_2 with the lithologic environment. This comparison applies to present-day conditions of infiltration and recharge via surface runoff; however, a problem occurs in the west-central Amargosa Desert. Age of the ground water, determined from carbon isotopes, ranges from about 7,000 to 20,000 yr B.P. (years before present). This timespan included changes not only in mean annual temperature and precipitation but also in vegetation type and density, thereby affecting the availability of CO_2 for dissolution during the recharge or runoff process. Insufficient data are available to resolve this problem.

A trend of decreasing percent Ca^{2+} relative to Na^+ occurs from east to west in that part of the area designated as primarily tuffaceous valley fill (figs. 4, 5, and 9). This trend is particularly evident in T. 16 S., R. 49 E. and the eastern one-half of T. 16 S., R. 48 E. Because the potentiometric map (fig. 2) shows a high hydraulic potential region in the eastern one-half of T. 16 S., R. 49 E., one explanation for the trend could be that Ca^{2+} enriched water is diffusing westward; then precipitation of calcium carbonate (CaCO_3) could account for the increasing percent Na^+ . This hypothesis is unlikely, because the Ca^{2+} enriched water from the east also is enriched in Na^+ (fig. 4). Sodium concentrations decrease to a minimum value somewhere near the western two-thirds of T. 16 S., R. 49 E.; therefore, greater Na^+ concentrations in the eastern one-third of T. 16 S., R. 48 E. cannot result from diffusion of ground water from the east, with loss of Ca^{2+} by precipitation of CaCO_3 . Bicarbonate concentrations (fig. 6) support this argument also, because they, too, decrease to a minimum value in the same area.

The trend also could result from diffusion from the west of water more saline than that in the eastern one-third of T. 16 S., R. 48 E., and subsequent precipitation of CaCO_3 ; however, the potentiometric contours do not support this argument, as the flow is primarily southerly, not from the west as desired (fig. 2; Walker and Eakin, 1963). Furthermore, if the water

quality in the eastern one-third of T. 16 S., R. 48 E. (as evidenced by the water quality at sites 21 through 25) resulted from a mixture of water from the center of the tuffaceous valley fill (the central trough shown on the water-quality maps) and water from the western one-third of the same township, a significant difference would occur in Cl^- concentration between water in the trough and water in the eastern one-third of the township. Although the contrast in Cl^- concentration of the postulated source is almost a factor of 10, no significant difference exists between Cl^- in trough water and Cl^- in water at sites 21 through 25.

Interdiffusion (mixing) of ground-water of different chemistries and precipitation of CaCO_3 also would result in decreases in the CaCO_3 saturation index; that is, the ion-activity product divided by the equilibrium constant, as the percent Ca^{2+} decreases. If the increase in percent Na^+ among the sites plotted in figure 9 resulted from CaCO_3 precipitation, a corresponding decrease in CaCO_3 saturation should be observed. Although some of the chemical data may not be sufficiently accurate for precise calculation of CaCO_3 -saturation indices, no such trend is observed. Therefore, east-to-west changes in water quality in the primarily tuffaceous valley fill in the west-central Amargosa Desert are not due to simple ground-water diffusion processes accompanied by CaCO_3 precipitation. The decrease in Ca^{2+} relative to Na^+ toward the center of the trough is presumed to result from greater quantities of clinoptilolite precipitated from a ground-water composition evolved by reaction with vitric tuff.

The Potentiometric High in the Eastern One-Third
of T. 16 S., R. 49 E.

As first suggested by Winograd and Thordarson (1975, p. C84-C85), the proximity of a potentiometric high to the Gravity fault near its intersection with the Specter Range thrust fault indicates the possibility of a breach of the confining properties of the Gravity fault observed to the southeast along the Ash Meadows spring line (fig. 2). The permeable valley fill just west of the fault juncture could be a drain for water transecting the fault from the east; this fact is consistent with the general hydrologic setting indicated

by the potentiometric map. The large gradient from east to west across the southern end of the Gravity fault decreases significantly near the junction with the thrust fault. In the valley fill west of the fault, flow vectors change direction from a generally southerly orientation to approximately westerly in the vicinity of the fault juncture. The implication is that water is flowing from east to west across the Gravity fault; geochemical data were used to assess whether it is derived from the unconfined (valley fill) or the confined (lower carbonate) aquifer, discussed by Winograd and Thordarson (1975).

No water-quality data exist for the valley-fill aquifer just east of the Gravity fault, but a few samples from wells in T. 16 to 17 S., R. 51 to 52 E. (table 3, fig. 10) disclose an areal variability in water chemistry that would indicate that the valley fill is not a significant source of water for movement west of the Gravity fault. The potentiometric surface in the valley fill is about 10 m lower than that in the underlying carbonates; this difference provides a potential gradient for upward leakage. Although some water in the valley-fill is chemically similar to water in the carbonates, additional solutes are required for a reasonable mass balance. The most likely candidate water in the valley fill to have resulted from upward leakage is from well 17S/51E-23b (fig. 10). The most significant expected change in water quality during upward movement would be loss of CO_2 from a decrease in pressure of about 100 kPa (kilopascals). The effect of this loss would be an increase in pH, with a consequent increase in carbonate (CO_3^{2-}) activity and precipitation of CaCO_3 . The temperature decreases about 10°C as the water moves from the carbonate aquifer ($\sim 30^\circ\text{C}$) to the valley fill aquifer ($\sim 20^\circ\text{C}$). Considering only temperature effects, CO_2 solubility should increase about 35 percent, and no loss of CO_2 would be expected. A change in the CO_3^{2-} equilibrium, therefore, would not be expected. However, this effect is opposed by the 50-percent decrease in pressure mentioned previously, from about 200 to 100 kPa.

Significant upward leakage and CaCO_3 precipitation cannot readily explain the water quality of the other valley fill-derived samples, because they contain either significantly larger or smaller concentrations of presumably conservative constituents, such as Cl^- or SO_4^{2-} . These waters

Table 3.--Comparison of water quality from valley-fill (VF) and lower carbonate (C) aquifers east of the intersection of the Gravity and Specter Range faults

[m, meter; °C, degrees Celsius; mmol/L, millimoles per liter; Ca²⁺, calcium; Mg²⁺, magnesium; Na⁺, sodium; K⁺, potassium; HCO₃⁻, bicarbonate; CO₃²⁻, carbonate; Cl⁻, chloride; SO₄²⁻, sulfate; SiO₂, silica]

Location (fig. 10)	Aquifer	Approximate well depth (m)	Water-level altitude (m)	Date of collection	Temperature (°C)	Dissolved constituents (mmol/L)										Onsite pH
						Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	HCO ₃ ⁻ + CO ₃ ²⁻	Cl ⁻	SO ₄ ²⁻	SiO ₂			
16S/51E-23	VF	189	---	9-09-69	28	0.15	0.31	2.78	0.25	2.20	0.62	0.41	0.03	8.4		
16S/52E-15a	VF	244	721	5-01-64	29	.82	.74	5.35	.31	3.44	1.55	1.71	.32	7.6		
17S/51E-1a	VF	41	714	1-10-61	23	.97	.82	3.00	.26	5.74	.17	.55	.30	7.2		
17S/51E-23b	VF	7	710	10-05-71	20	.75	.86	3.26	.31	4.85	.56	.65	.45	7.7		
17S/51E-23b	VF	7	710	8-02-72	21	.77	.82	3.13	.31	4.62	.57	.72	.45	7.9		
16S/51E-23	C	---	720	9-09-69	31	1.07	.74	2.78	.21	4.67	.56	.73	.37	7.6		
14S/52E-3d	C	1,036	729	3-12-80	64	1.10	.66	2.78	.25	4.00	.56	.78	.63	7.4		
Fairbanks Spring (17S/50E-9a)	C	---	---	9-28-69	27	1.17	.82	2.83	.24	4.95	.59	.85	.37	7.3		

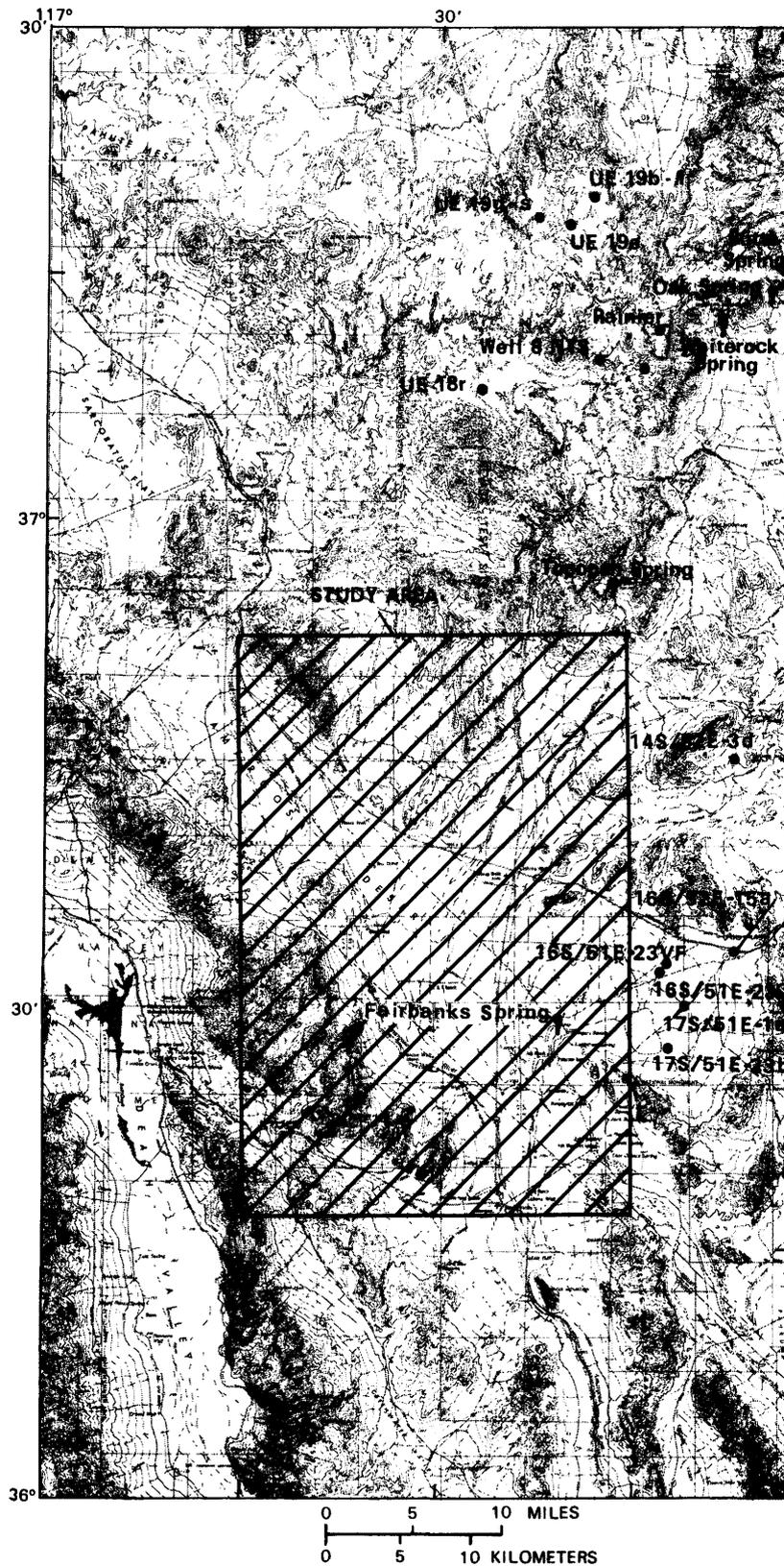


Figure 10.--Location of ground-water sampling sites peripheral to the west-central Amargosa Desert.

could be derived in part from the carbonates; however, much of the valley fill from which these samples were derived consists of limestone and dolomite; relative proportions contributed from the two sources cannot be determined. The valley-fill water sample from 16S/51E-23 yielded a carbon-14 (^{14}C) activity of 90 percent modern (Schroder and others, 1978), indicating that there is some recharge to the valley fill in this area under present-day climatic conditions. This evidence supports the hypothesis that water in the valley fill east of the Gravity fault in T. 16 to 17 S., R. 51 to 52 E. is a mixture of recharge and upward leakage, proportions of which vary from one location to another.

The foregoing discussion of water quality in the valley fill is for the area located in a township next to (east of) the area just east of the intersection of the Gravity (normal) and Specter Range (thrust) faults. Results perhaps cannot be extrapolated because the presence of a major surface drainage, Rock Valley Wash (fig. 10), exists in T. 16 S., R. 50 E., whereas townships to the east do not contain a major surface drainage. An evaluation of water quality in the valley fill just west of the fault intersection may determine whether the effect of Rock Valley Wash must be considered to explain ground-water movement in this area.

Sites 27, 28, 30, and 31 (fig. 3) are those closest to the Gravity fault; chemical analyses of the water are included in table 1. The average of these four analyses is presented in table 4 and compared with the average of the two nearest carbonate-aquifer sampling locations, 16S/51E-23 and Fairbanks Spring (table 3). Similarity of the Ca^{2+} , Mg^{2+} , $\text{HCO}_3^- + \text{CO}_3^{2-}$ concentrations indicate development of both water qualities in carbonate terrane. The dissimilar Na^+ , K^+ , Cl^- , SO_4^{2-} concentrations suggest three possible mechanisms for development of the water quality in the valley fill just west of the fault intersection:

1. Upward leakage from carbonate aquifer and mixing with water recharged directly to the valley fill.
2. Upward leakage from carbonate aquifer and reaction with rock fragments, including evaporites, in the valley fill.
3. Water recharged primarily through and resident in the valley fill.

Table 4.--Comparison of the average water quality in valley fill near the intersection of the Gravity and Specter Range faults with that of the lower carbonate aquifer

[mmol/L, millimoles per liter; Ca²⁺, calcium; Mg²⁺, magnesium; Na⁺, sodium; K⁺, potassium; HCO₃⁻, bicarbonate; Cl⁻, chloride; SO₄²⁻, sulfate; SiO₂, silica]

Location	Averages (mmol/L)							pH	
	Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	HCO ₃ ⁻	Cl ⁻	SO ₄ ²⁻		SiO ₂
Valley fill:									
Wells 27, 28, 30, and 31.	1.19	0.79	5.02	0.33	4.86	0.74	1.66	0.65	7.66
Lower carbonate aquifer:									
Fairbanks Spring and 16S/51E-23.	1.12	.78	2.80	.22	4.81	.58	.79	.37	7.42

Mechanisms 1 and 2 cannot be distinguished on the basis of chemical-quality data alone. Evidence of playa deposits in the area (W C Swadley, U.S. Geological Survey, written commun., 1981), which could contribute evaporites or small quantities of very saline paleowaters, further prevents distinction between mechanisms 1 and 2. Mechanism 2 cannot be relegated to secondary importance by examination of the stable-isotope--deuterium (D) and oxygen-18 (^{18}O)--contents of the carbonate and valley-fill water. In the carbonate aquifer, $\delta\text{D} = -103$; $\delta^{18}\text{O} = -13.6$ (I. J. Winograd, U.S. Geological Survey, written commun., 1982); whereas, in the valley fill at site 27, $\delta\text{D} = -105$; and $\delta^{18}\text{O} = -13.8$. The difference in analysis between the two water sources is about equal to standard analytical error reported for D and ^{18}O analyses.

Water recharged directly through the valley fill (mechanism 3) also would be expected to reflect the presence of playa deposits. As previously mentioned, a large contrast occurs between ground water associated with playa deposits (for example, sites 39 and 44) and water only a few kilometers away (sites 16, 18, and 19) not associated with playa deposits. In the above example, the greatest contrast is in the Na^+ and SO_4^{2-} concentrations, with Cl^- and HCO_3^- next. If the source of the water at sites 39 and 44 is the nearby more permeable valley fill, the observed differences in HCO_3^- and Cl^- concentrations would not be expected. In the case of HCO_3^- , the source water already is saturated with respect to CaCO_3 and would not be expected to dissolve carbonate facies in the playa deposits and thereby increase its HCO_3^- concentration. A source of chloride enriched minerals, which could dissolve as water passed through the playa deposits, is unlikely, because the water associated only with playa deposits is only slightly saturated with respect to soluble chlorides. Therefore, the most probable source of water in the playa deposits in the vicinity of sites 39 and 44 is direct recharge. This also may be the source of water in the playa deposits in the vicinity of sites 27, 28, 30, and 31, and other sites associated with playa deposits. However, the relative homogeneity of water quality throughout the large area represented by sites 27, 28, 30, and 31 (as opposed to the contrast within the short distance between sites 39 and 44) indicates some integrating effect, such as a water source in the carbonates. The Cl^-

and HCO_3^- contrast between nearby ground water with relatively small dissolved-solids concentrations, and ground water associated with playa deposits observed near sites 39 and 44, does not exist near sites 27, 28, 30, and 31. The difference between these sites and the confined aquifer is less. Therefore, the carbonate aquifer cannot be ruled out as a contributor to the water in the playa deposits, using only major dissolved ions as indicators of water source.

Two other types of evidence support the hypothesis that the carbonate aquifer rather than direct recharge is the source of ground water in the valley fill: water temperature and hydraulic potential. The water temperature at site 27 is 30.6°C and the well depth is 61 m; the temperature in the carbonate aquifer at 16S/51E-23 is 31°C , and at Fairbanks Spring, it is 27°C . In contrast, average temperature of water from other wells completed to about the same depth (61 m) in the valley fill, but distant from sites 27, 28, 30, and 31, is 21°C (seven samples). Water-temperature data support the hypothesis that the carbonate aquifer is contributing water to the valley fill, at least at site 27.

The hydraulic-potential data (fig. 2) show a lesser contrast between the carbonate aquifer and the valley fill in the vicinity of site 27 than almost anywhere else in the west-central Amargosa Desert. At site 27, about a 6-m difference in hydraulic potential occurs (carbonate aquifer higher), whereas in the next township to the south near the Gravity fault, the difference is about 50 m. The smaller potential difference and the presence of more permeable sand and gravel in the valley fill in the vicinity of site 27 (previously discussed) support the presence of significant upward leakage from the carbonate aquifer.

In summary, the mechanism that best explains the water-quality data in the vicinity of the potentiometric high is upward leakage from the lower carbonate aquifer into the valley fill in combination with water that has been recharged directly into the valley fill. This process allows the presence of recharged water in the valley fill, known to exist at other locations, and it can explain the possibility of a small difference in D content between site 27 and the carbonate aquifer. The effect of this west-southwestward moving water on the water quality in the tuffaceous valley fill is discussed in a subsequent section of the report.

Water-Quality Gradients near the Eastern Boundary
of the Tuffaceous Valley Fill

The central part of the tuffaceous valley fill, which is the trough shown on the water-quality maps (figs. 4 through 7), contains water that probably has originated as surface runoff. Infiltration of this runoff occurred primarily in the vicinity of present-day drainageways; reaction with primarily vitric tuff resulted in the observed water quality. Between the center of the trough and the area just west of Gravity fault, water quality is compositionally between that derived from vitric tuff (VT) and that resulting from a probable combination of water qualities of the lower carbonate aquifer (C) and valley fill aquifer (VF), as represented by water in the valley fill near the intersection of the Gravity and Specter Range faults. This compositional gradation may be expressed as a simple mixing of the water in the area of wells 27, 28, 30, and 31 (C + VF) with typical water from the center of the trough, as represented by the water in the area of wells 9, 14, and 15 (VT). Component water qualities and the contribution of (C + VF) water needed to approximate the composition of the other intermediate ground-water samples are shown in table 5. Mass conservation is assumed in the mixing calculations. Negative numbers and numbers greater than 100 result from choice of composition for VT and (C + VF). The assumption was tested by comparison of the mixing percentage of (C + VF), using the average of eight major dissolved constituents, any of which might not be expected to react conservatively, with the mixing percentage of (C + VF) determined by using Cl^- only, an element expected to be conservative. Comparison of the mixing percentages calculated by each method (fig. 11) indicates that conservancy is approximated for percent mixing values less than 50, but a bias exists for percentages greater than 50, indicating the possibility of precipitation of a solid phase or ion exchange. However, the bias is insufficient for any meaningful estimate of the nature of the precipitate to be made.

Table 5.--Ground-water quality between the center of the tuffaceous valley fill aquifer and the Gravity fault expressed as a mixture of two component water qualities

[VT, vitric tuff, expressed as an average of wells 9, 14, and 15; C + VF, a combination of water qualities of the lower carbonate aquifer (C) and valley fill aquifer (VF), expressed as an average of wells 27, 28, 30, and 31; Ca²⁺, Mg²⁺, magnesium; Na⁺, sodium; K⁺, potassium; HCO₃⁻, bicarbonate; Cl⁻, chloride; SO₄²⁻, sulfate; SiO₂, silica; mmol/L, millimoles per liter]

	Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	HCO ₃ ⁻	Cl ⁻	SO ₄ ²⁻	SiO ₂	pH
	(mmol/L)								
VT	0.46	0.08	1.77	0.19	2.28	0.19	0.27	1.20	8.21
C + VF	1.19	.79	5.02	.33	4.86	.74	1.66	.65	7.66

Well number	Percent (C + VF) contribution	
	Average of eight dissolved constituents: Ca ²⁺ , Mg ²⁺ , Na ⁺ , K ⁺ , HCO ₃ ⁻ , Cl ⁻ , SO ₄ ²⁻ , SiO ₂	For Cl ⁻ only (percent)
4	-3	-4
5	9	5
6	5	-4
7	20	27
8	11	16
10	-1	7
11	2	0
17	17	16
19	4	16
29	60	84
32	104	115
33	107	125
36	82	109

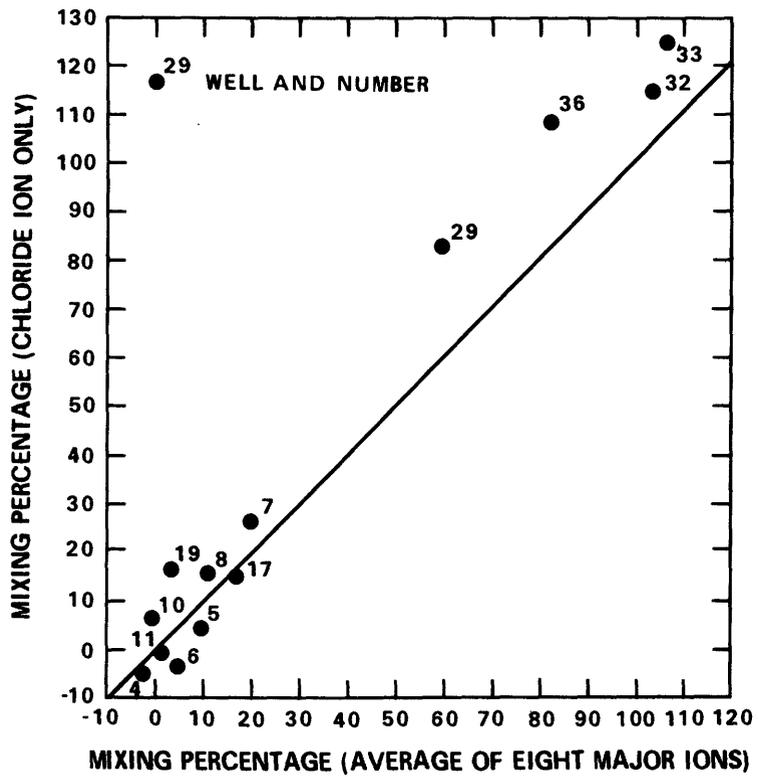


Figure 11.--Comparison of mixing percentage calculated by two methods.

Water Quality in the Valley Fill near the Amargosa River Channel

The western-to-southwestern one-third of the study area is underlain by valley fill in and near the Amargosa River channel. Although water flows in the channel during floods, it is dry almost all the time under present climatic conditions. The downstream reach has been mapped by Denney and Drewes (1965) in the Ash Meadows quadrangle; they show a narrow area of gravel surrounding the present-day channel. The valley fill along the upstream reach of the river within the study area is unmapped, but probably contains gravels composed primarily of carbonate detritus derived from the carbonate outcrops north of, adjacent to, and south of the area (fig. 1).

Water quality associated with the valley fill along the upstream reach is typified by samples from wells 37, 45 through 52, and 60, and the downstream reach by samples from wells 39, 41, 42, 44, 54, 55, and 57. Generally, ground water along the upstream reach contains more dissolved solids than ground water along the downstream reach. Water-quality maps indicate large gradients between the valley fill containing principally carbonate detritus near the Amargosa River and the valley fill to the northeast, in the central part of the study area, presumed to be principally tuffaceous. Farther south, gradients are smaller. If Cl^- and SO_4^{2-} may be treated as conservative components of the aqueous phase, a hypothesis regarding the source of the water near the Amargosa River may be formulated.

The relationship of some ground water in the Amargosa Desert to present-day (1980) precipitation, and evaporative concentration of that precipitation, is shown in figure 12. Ground water in the Amargosa Desert contains less sulfate (relative to chloride) than present-day precipitation concentrates. Water in the tuffaceous valley fill has been concentrated by a factor of about 25; water in the valley fill along the upstream reach of the Amargosa River has been concentrated by a factor greater than 200; whereas, water in the valley fill along the downstream reach of the Amargosa River has been concentrated by a factor of about 100. If it is assumed the source of all Cl^- and SO_4^{2-} is precipitation, which appears to be verifiable for the tuffs, and further assumed that recharge is through the valley fill during flooding, it must be concluded that either: (1) A greater number of floods which

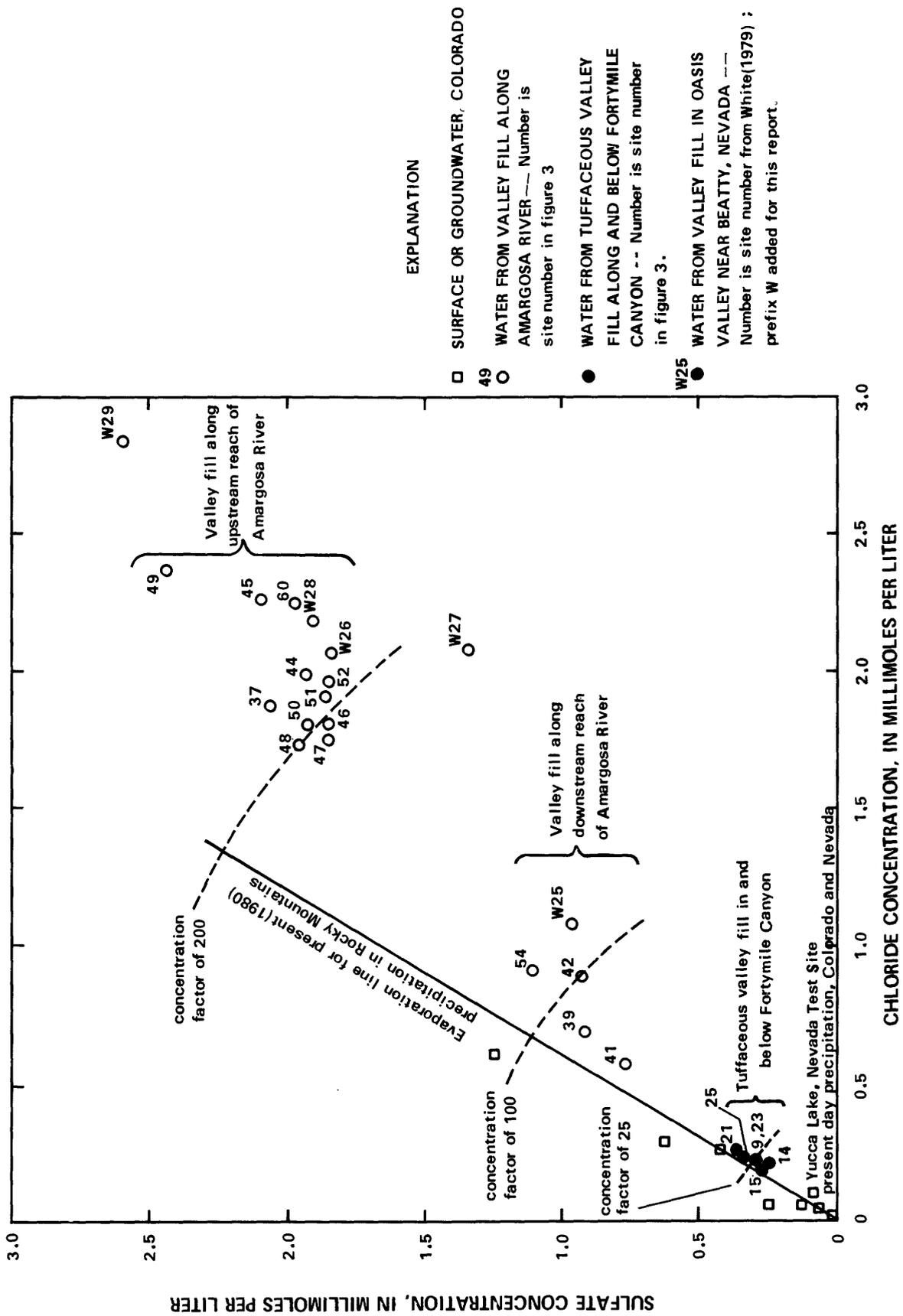


Figure 12.--Chloride and sulfate in precipitation and selected ground and surface waters, Colorado and Nevada.

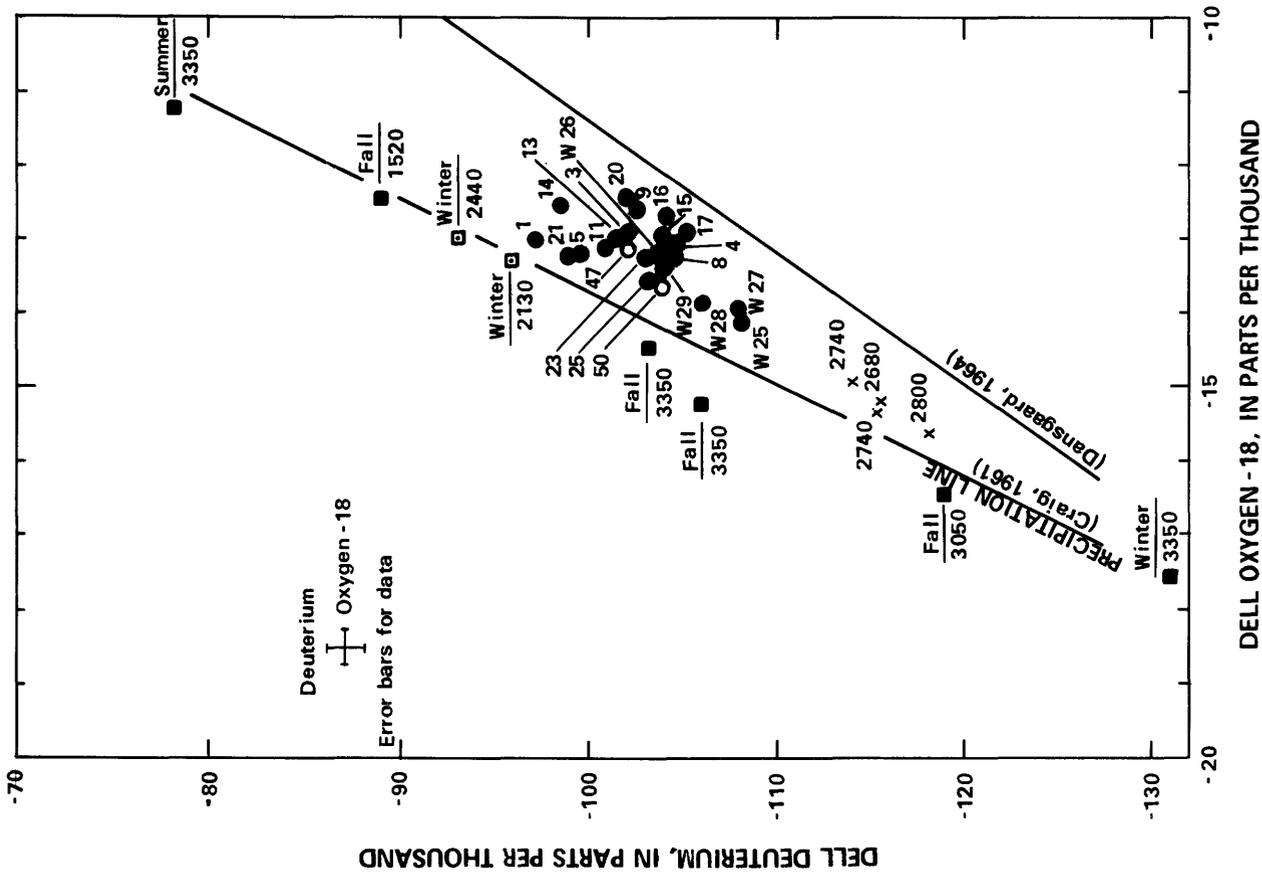
result in recharge occur in Fortymile Canyon than in the Amargosa River, resulting in a greater fraction of runoff recharge in the tuffaceous valley fill; or (2) floods in the Amargosa River are smaller than those in Fortymile Canyon.

Conclusion 2 is not reasonable, because the Amargosa River drainage basin is larger ($\approx 2,600 \text{ km}^2$) than that of Fortymile Canyon ($\approx 700 \text{ km}^2$) (Nevada State Engineer's Office, 1971). Therefore, conclusion 1 probably is correct. In Fortymile Canyon, even small discharges can saturate significant depths of vadose zone, because of the confined nature of the stream channel in and near the area of recharge. In contrast, the Amargosa River flood plain is large and has little relief. A given discharge probably covers a larger area to a shallower depth than in Fortymile Canyon, resulting in a decreased potential for recharge.

As previously discussed, drillers' logs show the presence of coarser valley fill below Fortymile Canyon (fig. 2), whereas lakebeds are indicated in at least parts of the valley fill along the Amargosa River. Coarser materials probably indicate greater vertical permeability and, therefore, greater ease of recharge for all surface runoff.

Although the greater salinity of ground water in the vicinity of the Amargosa River may be rationalized by the preceding arguments, the fact that salinity along the upstream reach is greater on the average than salinity along the downstream reach needs further explanation. One explanation may be that valley fill along the downstream reach of the Amargosa River received recharge from floods occurring in both the Amargosa River and Fortymile Canyon. This source seems likely, based on present-day topography, which shows coalescing of surface drainageways of the two systems in the vicinity of the sampling sites along the downstream reach of the Amargosa River.

Hydrogen- and oxygen-isotope data support the preceding chemical arguments. Hydrogen- and oxygen-isotope relationships among coastal precipitation, continental precipitation, and various ground and surface waters are shown in figure 13. Craig (1961) and Dansgaard (1964) have shown that near-oceanic precipitation has a predictable relationship between



hydrogen and oxygen isotopes, and that, as the precipitation contains an increasing proportion of continental moisture, oxygen-isotope fractionation causes the heavy isotope in rain and snowfall to be depleted. This relationship is illustrated by the displacement of line Y from line X in figure 13. The precipitation line (Y) from Craig (1961) has present-day Colorado and Nevada precipitation data superimposed. Ground- and surface-water data from Colorado plot to the right of the precipitation line, indicating a small fractionation during the recharge process. This fractionation is most likely due to evaporation in a climate with little humidity, as suggested by Fontes and others (1979) and by other investigators. Assume that during the paleoclimate that existed at the time recharge occurred in the Amargosa Desert, precipitation had isotopic characteristics similar to contemporary precipitation. A comparison of isotopic composition of ground water in the Amargosa Desert with that of precipitation shows appropriate shifts toward heavier oxygen-isotope content in the ground water, and the range of values indicates snowmelt as the runoff source. This shift is similar to that observed for ground and surface waters in Colorado, whose source is in a colder climate than the probable paleoclimate of the Amargosa Desert. The isotopic shift occurs as a result of different fractionation factors for D and ^{18}O during evaporation. Furthermore, no difference in isotopic composition exists between ground water in the tuffaceous valley fill and that near the Amargosa River. This similarity indicates similar recharge mechanisms occurring under similar climatic regimes. Because water in valley fill along the Amargosa River and in tuffaceous valley fill below Fortymile Canyon is dissimilar in Cl^- and SO_4^{2-} , larger salt accumulations corresponding to complete evaporation, as opposed to concentration by evaporation, must be responsible. The concentration-by-evaporation process should result in a parallel increase in heavy-isotope content as solute concentration increases. This increase is not evident, because there is a similarity in isotopic composition between the two water types with as much as a-factor-of-8 difference in salinity. The source of greater salinity in valley fill along the Amargosa River may be residual salt from evaporation of a Pleistocene lake, or evaporation of precipitation or periodic small flood flows that do

not result in recharge. Because climatic regimes of the two recharge areas are identical, only the areal distribution of flow during runoff and the permeability difference between Fortymile Canyon and Amargosa River can account for existing water quality.

Oasis Valley as a Source of Water in the Valley Fill along the Amargosa River

The potentiometric surface in the valley fill along the Amargosa River indicates that subsurface flow downriver is possible. The source of such flow likely would be the valley fill in Oasis Valley, near Beatty, Nevada. White (1979) provides a comprehensive discussion of the hydrology and geochemistry of Oasis Valley. The Cl^- and SO_4^{2-} data (White, 1979) and the hydrogen- and oxygen-isotope data (A. F. White, U.S. Geological Survey, written commun., 1981) for the five sites near the Narrows below Beatty were plotted in figures 12 (Cl^- and SO_4^{2-}) and 13 (hydrogen and oxygen isotopes) for comparison with water in the valley fill of the Amargosa Desert.

Water quality in Oasis Valley is quite diverse (figs. 12 and 13) in an area of only a few square miles and makes selection of a typical water chemistry difficult. Well W27 has the greatest reported discharge and penetrates the valley fill to the greatest depth; therefore, it might be chosen as representative of water leaving the Narrows. The Cl^- and SO_4^{2-} concentrations and isotopic composition of this water are quite unlike those in water from the valley fill along the entire Amargosa River, and these differences would indicate that the ground waters farther south are not derived from subsurface flow originating in the Beatty area. A contrast to this conclusion results if the water chemistry in Oasis Valley is represented by water from wells W26 or W29. The chemistry of these waters is quite similar to that in the valley fill along the upstream reach of the Amargosa River. Thus, two mechanisms may account for the ground-water quality along the upstream reach of the Amargosa River: recharge of surface runoff or underflow of ground water from Oasis Valley. Because the ground-water chemistry along the downstream reach of the Amargosa River can be explained only by recharge of surface runoff, the author favors this mechanism as dominant along the entire reach of the Amargosa River.

The data also are consistent with at least some of the ground water being derived from infiltration of overland flow in Oasis Valley. Carbon-14 estimates of ground-water age in Oasis Valley are consistent with this hypothesis (A. F. White, U.S. Geological Survey, written commun., 1981).

PALEOHYDROLOGY OF THE WEST-CENTRAL AMARGOSA DESERT

Carbon-, hydrogen-, and oxygen-isotope data from ground water of the west-central Amargosa Desert support the hypothesis that water recharged the valley fill primarily through runoff infiltration, with upward leakage from the regional carbonate aquifer in one area. The hydrogen- and oxygen-isotope data have afforded simple interpretation, but, as discussed below, the carbon-isotope data are more difficult to interpret.

Interpretation of Carbon-Isotope Data

Carbon-isotope data are given in table 6; distribution in the study area of unadjusted ^{14}C ages determined for selected ground-water samples is shown in figure 14. Values not in parentheses indicate samples that probably were derived from greater than 90-percent reaction with tuff. Values in single brackets represent samples derived from 80- to 90-percent reaction in tuff; those in double brackets probably were affected by significant carbonate dissolution. Examples of the latter are those samples derived from water in contact with valley fill containing carbonate detritus.

Little or no radiocarbon dilution from carbonate rocks probably occurred during the evolution of water in tuff or tuffaceous valley fill; therefore, the unadjusted ages are taken as true ages. Ground waters with as much as 20-percent carbonate-derived contribution (all recharged contemporaneously) may contain only about 90 percent of the ^{14}C activity they would have had if they were not diluted by carbonate-derived ground water; therefore, their unadjusted age is as much as 1,000 years too old. This latter estimate has assumed that no ^{12}C , ^{13}C , \leftrightarrow ^{14}C exchange takes place. Those age data derived from ground-water samples suspected to have evolved in carbonate-detritus valley fill present a serious problem in evaluating the meaning of the unadjusted age. For example, at sites 47 or 60, if all of the reaction has

Table 6.--Carbon-, hydrogen-, and oxygen-isotope data for west-central Amargosa Desert and vicinity
 [C, carbonate; T, tuff; ‰, parts per thousand; SNOW, standard mean ocean water; PDB, Peedee belemnite;
 yr B.P., years before present; mmol/L, millimoles per liter; IAP/K, ion-activity product divided by
 equilibrium constant; CaCO₃, calcium carbonate]

Location name	Probable lithology	δD ‰ SMOW	δ ¹⁸ O ‰ SMOW	δ ¹³ C ‰ PDB	¹⁴ C (percent modern)	¹⁴ C apparent age (yr B.P.)	Total dissolved inorganic carbon (mmol/L)	IAP/K CaCO ₃
Fairbanks Spring- ^{1/}	C	-103	-13.6	-5.0	1.9	31,800	5.20	---
Well 16S/51E-23 ^{1/} 2/	C	-----	-13.6	-4.6	1.8	32,300	5.05	---
Amargosa well 1	T	-97.5	-13.0	-7.3	29.2	9,900	2.27	0.12
2	T	-97.5	-12.8	-7.9	32.2	9,100	2.20	.11
3	T, C?	-102	-12.8	---	15.6	14,900	2.49	1.96
4	T	-103	-13.2	-7.1	19.3	13,200	2.26	2.04
5	T, C?	-99.5	-13.2	-6.8	21.4	12,400	2.67	.47
8	T, C?	-103	-13.4	-7.3	21.9	12,200	2.36	1.51
9	T	-102	-12.6	---	28.4	10,100	2.52	1.46
10	T, C?	-97.5	-13.2	-5.2	24.8	11,200	2.23	1.06
11	T	-101	-13.1	---	20.8	12,600	2.25	1.95
13	T	-102	-13.0	---	19.3	13,200	2.21	.65
14	T	-98.5	-12.6	---	-----	-----	2.23	2.06
15	T	-104	-13.0	---	18.4	13,600	2.26	1.36
16	T, C	-104	-12.7	---	10.0	18,500	2.57	2.69
17	T, C?	-105	-12.8	---	18.9	13,400	2.20	1.19
18	T	-102	-13.0	---	27.8	10,300	2.06	2.04
19	T, C	-----	-----	---	40.3	7,300	2.01	1.13
20	T?	-102	-12.4	---	13.8	15,900	2.58	.66
21	T	-99	-13.2	-8.4	27.4	10,400	2.13	1.15
23	T	-103	-13.4	-7.1	17.1	14,200	2.57	.71
25	T	-102	-13.4	-5.6	15.6	14,900	2.59	1.06
27	C, T?	-105	-13.8	-3.6	7.0	21,400	5.01	1.74

Table 6.--Carbon-, hydrogen-, and oxygen-isotope data for west-central Amargosa Desert and vicinity--Continued

Location name	Probable lithology	δD ‰ SHOW	$\delta^{18}O$ ‰ SHOW	$\delta^{13}C$ ‰ PDB	^{14}C (percent modern)	^{14}C apparent age (yr B.P.)	Total dissolved inorganic carbon (mmol/L)	IAP/K $CaCO_3$
29	C, T	-105	-13.8	-3.4	---	---	3.34	1.18
30	C, T?	-104	-13.7	-4.4	10.3	18,300	5.17	2.57
47	C, T?	-102	-13.1	-6.2	31.4	9,300	4.07	1.51
50	C, T?	-104	-13.6	-5.7	---	---	4.02	1.79
55	T	-108	-14.2	-8.5	12.0	17,000	2.78	
56	T	-102	-13.4	-11.0	19.8	13,000	2.14	
57	T	-101	-13.4	-10.4	17.3	14,100	2.34	
<u>3/</u> 60	C, T	-----	-----	-5.9	28.8	10,000	5.36	
Well 8, Nevada Test Site.	T	-104	-13.0	-12.1	25.4	11,000	1.43	

1/ From Winograd and Pearson (1976).

2/ Amargosa tracer well 2.

3/ From Schroder and others (1978).

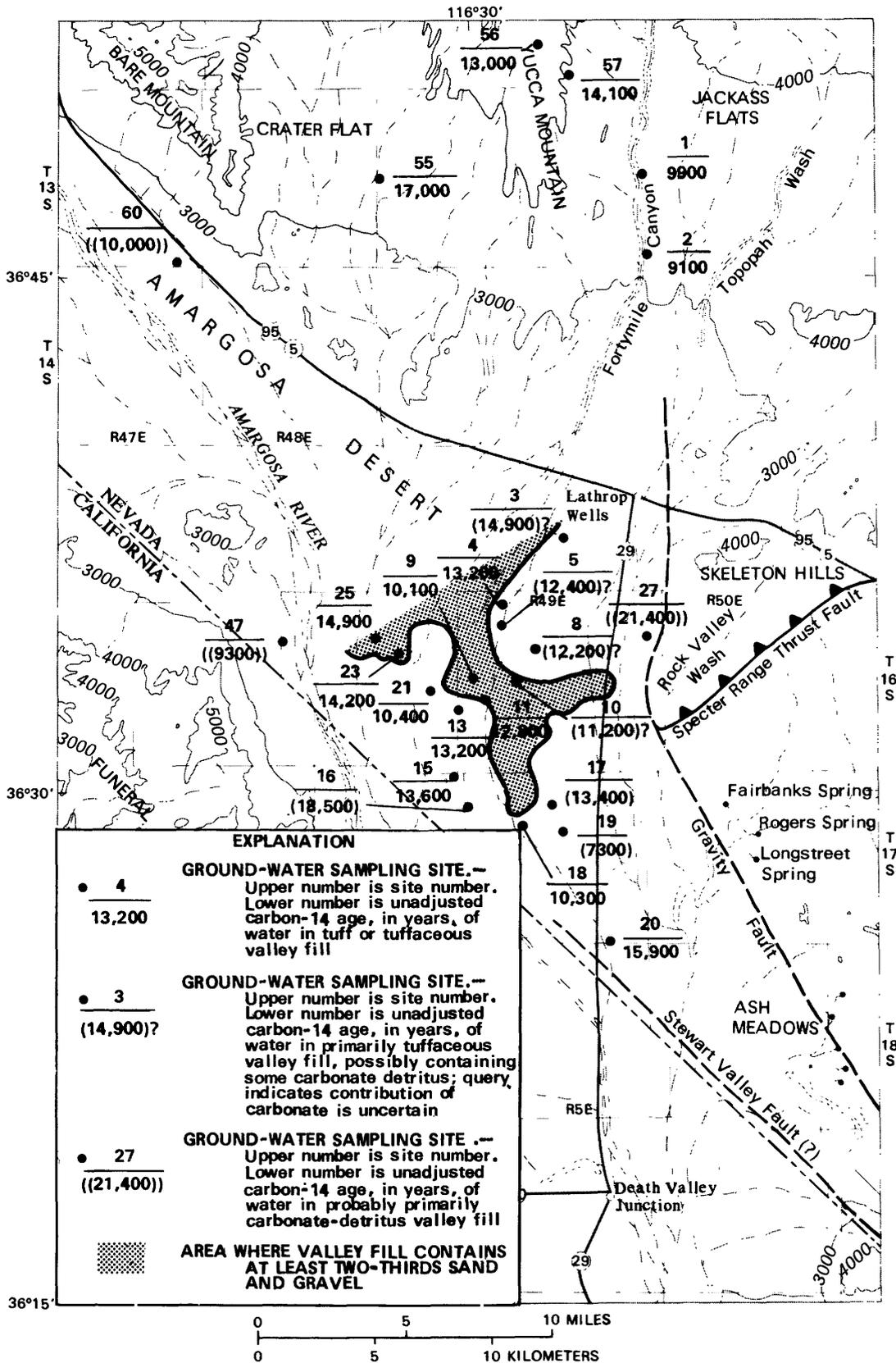


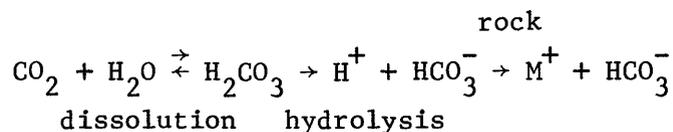
Figure 14.--Unadjusted carbon-14 age of ground water in relation to major surface drainageways and relative permeability.

been with carbonate detritus, the actual age of the water is about 5,600 years younger than the unadjusted age of about 9,000 to 10,000 yr B.P. The paucity of data and the complexity of the problem preclude further discussion of this example.

Those ground-water samples believed to be chemically derived primarily by reaction with tuff, are amenable to useful interpretation. The assumption that these samples contain little or no CaCO_3 -derived dissolved inorganic carbon is supported by major dissolved-ion geochemistry; this assumption should be supported further by stable carbon-isotope data. Several approaches have been used to correct ^{14}C ages for CaCO_3 dilution through understanding the reactions of ^{13}C . These approaches involve either simple isotope-dilution models, equilibrium-exchange models, or combinations of both (Wigley and others, 1978); for examples of application of these models, see Shampine and others (1979), Pearson and Swarzenki (1974), Deines and others (1974), and Rightmire and Hanshaw (1973). Successful application of most techniques have been hampered by insufficient data to make precise calculations, from lack of data from a suitable hydrologic system to test the model, or from unsuitability of laboratory experimental design for extrapolation of results to natural hydrologic systems. This study also has similar limitations, but the probable hydrologic mechanisms that have already been proposed will be tested against the evidence of stable carbon isotopes.

The $\delta^{13}\text{C}$ composition of water from the tuff is -7.3 (median) or -8.0 ‰ (mean). For reaction with silicate rock only, and without precipitation of carbonate species, the carbon-isotope content of the ground water must equate to the carbon-isotope composition of recharging water. Because surface runoff is proposed as the principal recharge mechanism, the dissolved CO_2 source responsible for the dissolved inorganic carbon in ground water might be expected to be primarily the atmosphere. Atmospheric CO_2 has been reported to range from -6.4 to -7.6 ‰ (Wallick, 1976) and -6.7 to -7.4 (Keeling, 1961). Although reacting CO_2 may be primarily of atmospheric origin, the runoff mechanism involves interaction with an usually small but finite thickness of surficial material and, therefore, some component of root-respired CO_2 from native vegetation. For environments similar to this study area,

$\delta^{13}\text{C}$ of soil-zone CO_2 has been reported to range from -15 to -19 (Rightmire, 1967), about -18 (Galimov, 1960), -12.0 to -18.4 (Wallick, 1976), and an average of organic detritus from juniper woodland near the study area of about -21 (W. G. Spaulding, University of Washington, written commun., 1980). Although P_{CO_2} in desert soils [~ 0.3 kPa (Wallick, 1976)] is much greater than atmospheric (~ 0.03 kPa), the top few centimeters of a permeable soil or alluvium would be expected to have near-atmospheric P_{CO_2} , but perhaps a slightly more negative $\delta^{13}\text{C}$ than atmospheric CO_2 . The latter phenomenon would result from diffusion of soil gas having a $\delta^{13}\text{C}$ of about -18 ‰ toward the atmosphere, coupled with mixing with atmospheric CO_2 containing $\delta^{13}\text{C} \sim -7$ ‰. The reaction of runoff with surficial materials takes place under open-system conditions with respect to CO_2 . As CO_2 dissolves in runoff water, it reacts with soils, sands, and gravels, and additional CO_2 dissolves to replace that used. Whether the dissolution and hydrolysis of CO_2 :



takes place under equilibrium conditions or not is not known, but it is almost certain that the resulting HCO_3^- reflects the isotopic composition of the carbonic acid (H_2CO_3), because the reaction with rock is irreversible. That is, the runoff is a sink for H_2CO_3 , and reacted HCO_3^- is not expected to exchange with gaseous CO_2 under equilibrium conditions. The importance of this mechanism lies in the effect that the large fractionation of ^{13}C bicarbonate under equilibrium conditions (Hendy, 1969; Deines and others, 1974) would have on the bicarbonate $\delta^{13}\text{C}$. If the reaction of dissolved CO_2 with surficial material took place under equilibrium conditions, a positive fractionation of between 8.2 ‰ (20°C) and 9.6 ‰ (5°C) might take place. This would result in $\delta^{13}\text{C}$ values of $+1.2$ to $+2.6$ ‰ for runoff and would be reflected in a similar positive $\delta^{13}\text{C}$ values for ground water. Ground water in the tuffaceous valley fill (and bedrock) has a median $\delta^{13}\text{C}$ of -7.3 ‰ (mean -8.0 ‰) inconsistent with its development under equilibrium conditions.

At some point, runoff either evaporates or becomes recharge. Evaporation results in deposition of readily resolvable or sparingly soluble salts. Recharge of runoff requires that soluble salts be carried to the water table, but what of the inorganic carbon? If pH of the evaporating runoff is greater than 9, significant fraction of inorganic carbon will be in the form of CO_3^{2-} ion; whereas, if pH is below that, HCO_3^- will predominate. If the reaction rate during runoff is slow compared to the rate of incorporation of atmospheric CO_2 , pH will tend to stay relatively low; whereas, pH will increase if the reaction rate is relatively fast. In the latter instance, evaporation may result in formation of CaCO_3 , presumably as caliche; in the former case, readily soluble salts will deposit. Lattman (1973) observed that cementation in alluvial fans progresses most readily in carbonate or basic igneous rock terrane, and least readily in silicic igneous detritus. Because weathering rates (reaction rates) are greatest in carbonate and basic igneous rocks, pH of runoff in these terranes is expected to be greater; therefore, it would result in greater probability of CaCO_3 formation. In the tuffs, pH of average runoff is expected to be less, and evaporation results in more soluble salts being deposited. This result also is promoted by ion-pair formation with Ca^{2+} , which decreases Ca^{2+} activity. These soluble salts either are carried farther down the surface gradient or recharged during subsequent runoff. All the steps occurring during runoff and recharge are non-equilibrium and, therefore, not subject to any significant carbon-isotope fractionation. This results in recharge water having a carbon-isotope composition identical to average runoff. This has been estimated to be somewhat more negative in $\delta^{13}\text{C}$ than atmospheric CO_2 (-7 ‰), the deviation from the atmospheric value depending on the quantity of vegetation-derived CO_2 that has been included. Runoff that recharges at higher elevations, closer to the heads of drainages, might be expected to have more negative $\delta^{13}\text{C}$ values, because the vegetation is more dense and contributes a greater proportion of CO_2 compared to runoff at lower elevations. The runoff at higher elevations also contains less total inorganic carbon, because the time that open-system conditions prevail is shorter. As the reach of surface runoff before recharge increases, a greater proportion of atmospheric CO_2 is incorporated into the

runoff, because vegetation is sparse at lower elevations, and contributes less to the total CO_2 , and total inorganic carbon increases because of increased reaction time. This can be seen by comparing the $\delta^{13}\text{C}$ and total-inorganic-carbon values for several wells in the main Fortymile Canyon drainage (table 6). Well 8, Nevada Test Site (site 64) (elevation 1,700 m), located near the head of the drainage, has a $\delta^{13}\text{C}$ of -12.1 ‰ and total inorganic carbon concentration of 1.43 mmol/L; wells J-13 and J-12 (sites 1 and 2) (elevation 980 m) are located about midway in the drainage reach and average $\delta^{13}\text{C} = -7.6$ ‰, with total-inorganic-carbon concentration = 2.24 mmol/L; sites 4 and 5 (elevation 730 m) are located near the terminus of the drainage and average $\delta^{13}\text{C} = -7.0$ ‰, with total-inorganic-carbon concentration = 2.46 mmol/L.

Almost all geochemists studying arid or semiarid hydrologic environments, who use carbon-isotope data to aid in understanding their systems, have been concerned with the effect of caliche on carbon-isotope composition of ground water. This concern necessitates an evaluation of the importance of caliche in determining the isotope compositions for the Amargosa Desert and vicinity. Lattman's (1973) observation regarding caliche formation in different lithologies indicates that it may not be important in silicic tuffs, but nevertheless caliche does cement alluvium in the study area (W C Swadley, U.S. Geological Survey, written commun., 1981). This fact is still consistent with Lattman's concept that eolian carbonate dust can have an effect in calichification. Carbonate outcrops are present upwind (west) of the study area.

Details of the occurrence of caliche in the stream channels diminish the concern about its effect on ground-water quality. W C Swadley (U.S. Geological Survey, written commun., 1981) also reported that the Fortymile Canyon floor probably did not contain caliche. Caliche is not known to occur in surficial materials in the head of the Fortymile Canyon drainage at an elevation of about 2,100 m. From a mass-balance perspective, dissolution of caliche in a tuffaceous environment should result in a positive correlation between Ca^{2+} and HCO_3^- if no selective removal mechanism exists; no such correlation is observed for the Amargosa Desert and vicinity. The median total-inorganic-carbon concentration in tuff-derived water is about 2.6 mmol/L; whereas, the median total-inorganic-carbon concentration in carbonate water

is about 5.1 mmol/L, approximately the value that might be expected for simple dissolution under similar conditions in limestone. The $\delta^{13}\text{C}$ data are similarly supportive: tuffaceous ground water = -8.0 ‰; carbonate ground water = -4.1 ‰, the latter value reflecting about 50-percent isotope dilution, with marine limestone having a $\delta^{13}\text{C}$ of ~ 0 ‰. Although the foregoing arguments do not rule out the possible dissolution of caliche, it needs to be designated a minor factor in the determination of ground-water chemistry. This concept is consistent with the hypothesis that, if recharge occurs, there can be no net accumulation of caliche, as such accumulation would tend to decrease permeability and decrease recharge. If recharge does occur, net removal of caliche must take place; permeability then increases, and recharge is increased.

Age of Ground Water and Paleoclimate

The distribution of ground-water ages (fig. 14) supports the arguments previously presented regarding recharge via overland flow. The youngest ages are located in or near present-day drainageways, which, therefore, are assumed to be the paleodrainageways. The frequency distribution of ground-water ages in tuff is shown in figure 15. No correction for non-radioactive carbon is believed necessary for these ages; therefore, they should represent true ground-water ages. The frequency distribution of ^{14}C ages coincides with a cool period that ended in the early Holocene. The mean annual temperature trend shown in figure 15 is taken from Stuiver and others (1978) for the Pacific Northwest. The trend is appropriate to the study area, although perhaps not representative of the actual magnitude of the temperature changes, as demonstrated by Van Devender and Spaulding (1979) and Betancourt and Van Devender (1981) for the southwestern United States, using packrat middens as chronicles of changes in vegetation and therefore changes in climate. Changes in hydrogen- and oxygen-isotope data for the same ground waters corroborate the temperature trend.

Where is the Water of Middle Wisconsin Age?

The absence of ground-water dates older than 17,000 yr B.P. is curious, because the temperature record presented by Stuiver and others (1978) (a part of which is reproduced in fig. 15) shows a persistent cool period throughout

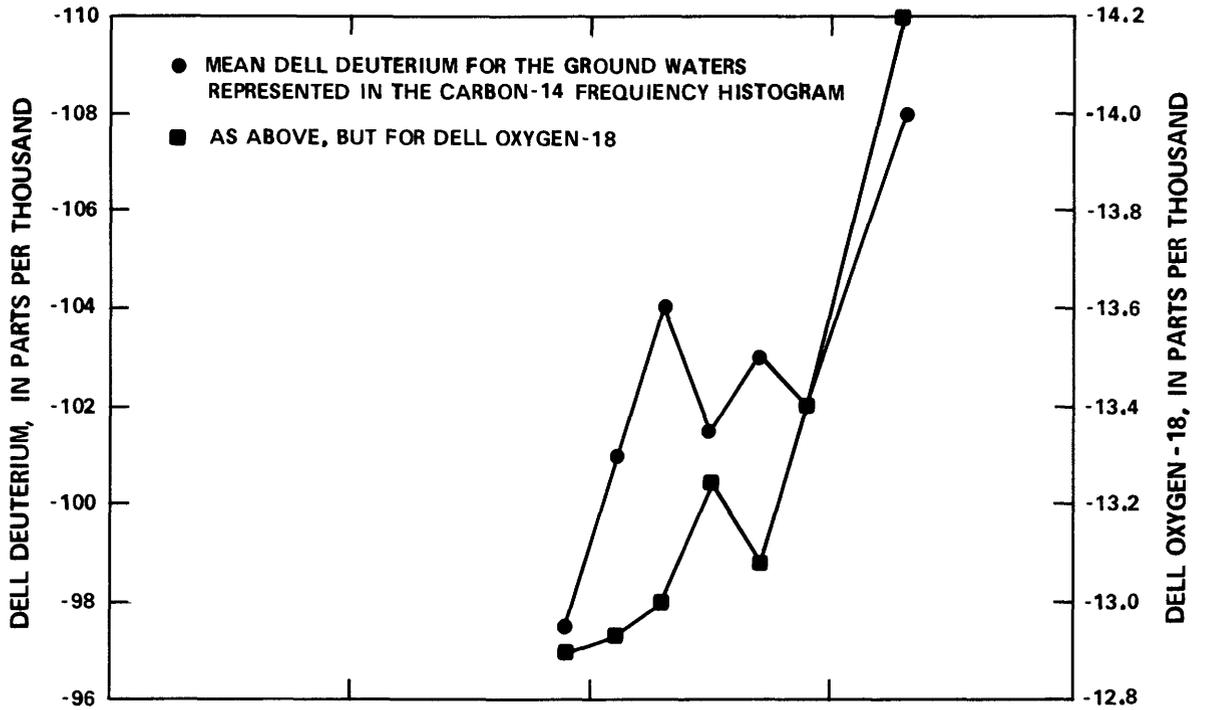
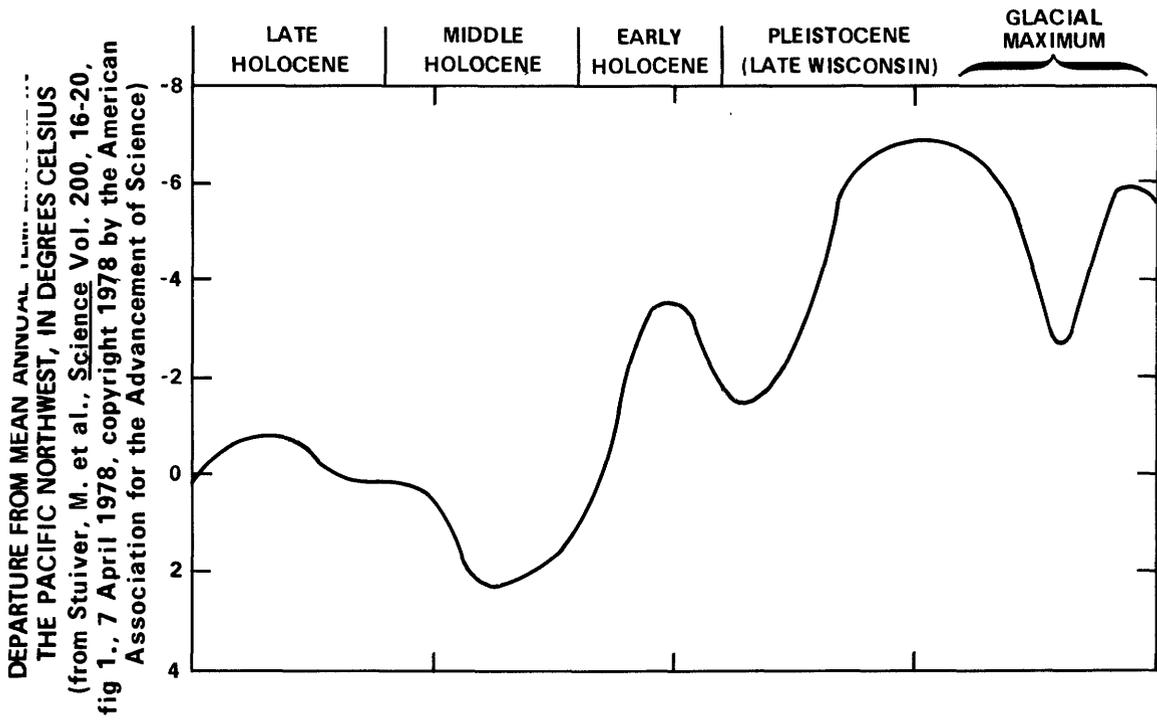


Figure 15.--Carbon-, hydrogen-, and oxygen-isotope data for water in tuff and tuffaceous valley fill and their relationship to climate change.

the Wisconsin. One explanation is that the absence of ages greater than 17,000 yr B.P. is an artifact of well location and completion. Most of the wells penetrate only the top few hundred meters of saturated thickness, that part of the aquifer expected to contain the youngest (most recently recharged) water. The data were examined to determine whether this was the controlling factor in determining ground-water age, but no such correlation was obtained; rather the position of the well relative to surface drainage dominated. Deeper wells did not yield older waters, but wells farther removed from surface drainageways did. A few of the wells penetrated significant thickness of saturated valley fill or bedrock, but no great age differences were apparent: for example, site 64 (well 8, Nevada Test Site), depth 1,670 m, saturated thickness ~ 1,350 m; Amargosa wells: site 2 (well J-12), depth 350 m, saturated thickness 120 m; site 4, well depth 90 m, saturated thickness 46 m; site 9, well depth 105 m, saturated thickness 72 m; site 18, well depth 100 m, saturated thickness 84 m. Water from all these wells had ages between 9,100 and 13,200 yr B.P. Water from the well with the least saturated thickness had the oldest age; water from the well with the greatest saturated thickness had a younger age (table 6). Probably the major producing zones of each of these wells are near the water table, so that only water from the top of the aquifer is removed; this upper part of the section would be expected to contain the youngest water. Further indication that this relationship between well depth and major producing interval may be true in bedrock aquifers can be obtained by comparing water ages and construction of wells J-12 and J-13 (sites 2 and 1) (Claassen, 1973). Both wells penetrate the same major producing zone, but J-13 is 710 m deeper. Well-performance records indicate that the greater depth of well J-13 does not result in increased capacity, so that very little water is being obtained from its greater depth interval. The older ground-water age at J-13 (about 800 years older than that at J-12) may indicate that some older water is being produced from below the interval penetrated by J-12, although the age difference is of borderline significance at the 95-percent confidence level.

Another possible explanation for the lack of ground-water ages in tuff greater than 17,000 yr B.P. is that permeability characteristics of the aquifers sampled are such that the older ground water has moved beyond the

study area. If it is assumed that significant recharge also occurred during 17,000 to about 27,000 yr B.P. (the end of the next nearest inter-stade), then those ground waters may have left the region that was sampled in this study. For example, if ground-water recharge occurred near well 8, Nevada Test Site (site 64), representing the head of Fortymile Canyon, and downgradient flow occurred in the most permeable part of the Canyon's subsurface toward site 18, its velocity would have to be between 7 m/a (meters per annum) (if recharged 17,000 yr B.P.) and 4 m/a (if recharged 27,000 yr B.P.). These velocities agree approximately with those calculated from limited hydraulic data (R. K. Waddell, Jr., U.S. Geological Survey, written commun., 1981). This agreement may not add credence to the argument, because a large range of parameter estimates results in similar velocities.

A third explanation, and that favored by the author, is based on a recently completed (1982) study of past climates at the Nevada Test Site (W. G. Spaulding, University of Washington, written commun., 1982). Using arguments based primarily on changes in vegetation at and near the Nevada Test Site, Spaulding concludes that until about 23,000 yr B.P., average annual precipitation was only about 20 percent greater than current (1982) precipitation, although much of this increase was during the winter months. During the late Wisconsin, however, average annual precipitation increased to about 40 percent greater than current (1982) precipitation, with 70-percent increases for the winter component. Although both the middle and late Wisconsin represent significant cooling, only the late Wisconsin yields evidence for significant increase in winter precipitation. This author then concludes that snowfall earlier than about 20,000 yr B.P. was insufficient to result in snowmelt recharge, while subsequent climatic conditions caused such recharge. This interpretation results in estimates of ground-water velocities in the Amargosa Desert considerably slower than the 4 to 7 m/a previously stated. If no significant recharge occurred prior to 17,000 yr B.P., its disappearance need not be accounted for. For example, assume recharge near the head of the Fortymile Canyon occurred 17,000 years ago and flowed to the lower end of the Canyon, where it is sampled. The velocity that is calculated, 7 m/a, must be a maximum. Because recharge may occur anywhere but not necessarily everywhere along surface drainageways, no probable minimum velocity can be calculated.

The absence of water older than about 10,000 yr B.P., even near the head of Fortymile Canyon, would favor velocities slower than 4 m/a.

The Problem of Ash Tree Spring

Ash Tree Spring, site 20, is located approximately in line with the trend of other tuff-derived water of the central part of the study area (the trough in the water-quality maps); water from the spring was believed by Naff (1973) and Winograd and Thordarson (1975) to have been derived by a similar mechanism: surface runoff and infiltration through valley fill. Dissolved-ion chemical data generally support this hypothesis, although smaller calcium and larger magnesium concentrations are small deviations. However, the isotope and hydraulic data contradict the chemical data. The ^{14}C age is 15,900 yr B.P., in marked contrast to the much younger ages of the upgradient ground water in most other parts of the valley fill downgradient from Fortymile Canyon. Furthermore, the water-level altitude of the spring is approximately 21 m higher than water levels in wells within 1 km of the spring. The nearest wells also contain water of a very different quality from the water of Ash Tree Spring (see analysis for sites 40 and 41, as examples, in table 1). This difference was noted by both Naff (1973) and Winograd and Thordarson (1975).

Two hypotheses offer reasonable explanations of the origin of water at Ash Tree Spring. The first hypothesis requires that the water-level altitude at Ash Tree Spring represents recharge in the valley fill at a time when the land surface was higher than it is today. Subsequent erosion or subsidence then left the (presumably more resistant) aquifer material of Ash Tree Spring topographically higher than the surrounding valley fill, and the formerly saturated material simply has been draining. If the flow were constant at the present-day (1972) discharge of $6.7 \times 10^{-4} \text{ m}^3/\text{s}$ (cubic meters per second) for 15,900 years, approximately $1.1 \times 10^9 \text{ m}^3$ (cubic meters) of aquifer with 30-percent porosity would have been dewatered. If the original greater thickness of aquifer is estimated by the difference in elevation between Ash Tree Spring and the present-day (1972) water level, 21 m, the area dewatered is $5.2 \times 10^7 \text{ m}^2$ (square meters), or a square 7.2 km on each side. Surficial materials are variable throughout such an area, and it seems unlikely that

they could contain what appears to be a simple tuff-derived water quality; however, this possibility cannot be ruled out.

The second hypothesis requires that the recharge to Ash Tree Spring originate from a different source than the source that recharged valley fill to the northwest. A conduit providing a flow path more or less isolated from the lakebeds and valley fill surrounding Ash Tree Spring would be required; such a conduit may be provided by the sandstone and claystone mapped by Denney and Drewes (1965) in the vicinity and southeast of the spring. This formation dips approximately northwest and probably is continuous from the spring to near the summits of the Resting Spring Range. It is possible that recharge occurring through the quartzite that comprises the summits of the range travels downdip through the (fractured?) sandstone toward Ash Tree Spring, discharging where the aquifer abuts less permeable lakebeds. Lithology of the sandstone and claystone includes conglomerate (quartzite, limestone, dolomite, tuff); Denney and Drewes (1965, p. L18) state that " *** layers of white tuff containing pumice fragments are common. One or two beds of *** very fine grained limestone occur in the lower part of the formation *** ." The common presence of volcanic glass generally supports the tuff-like characteristics of water in this formation, but the effect of interaction with some limestone or dolomite cannot be dismissed. The predominance of tuff compared to carbonate in this unit and the similarity of HCO_3^- concentration in Ash Tree Spring water to water found in other tuffaceous rocks would tend to minimize the effect of carbonate rocks in determining water quality.

SUMMARY AND CONCLUSIONS

All available data on ground-water quality for the west-central Amargosa Desert were applied to development of a conceptual geochemical model of that hydrologic regime. Hydraulic, geologic, and lithologic data for the region also were incorporated as needed.

Results indicate that ground water was recharged to the valley fill primarily by overland flow in or near present-day stream channels. The flow probably was derived from snowmelt. Numerous small floods occur that do not result in recharge. Frequency of effective recharge depends primarily on

channel geometry and vertical permeability of the valley fill. The major-ion chemical composition of the resulting ground water is determined by the lithology of rock fragments that comprise the stream channel and the subsurface, as well as soluble salts that remain from evaporation of numerous non-recharge events. Recharge through valley fill composed of fragments of tuff or carbonate, or mixtures of the two, and through playa deposits, results in distinctly discernible water qualities. The central part of the study area is underlain primarily by tuff, while the eastern and western flanks are underlain primarily by carbonate rocks. Each area occasionally shows the effect of playa deposits; that is, ground water in the vicinity of these deposits contains greater concentrations of the conservative SO_4^{2-} and Cl^- ions than does the water not associated with the deposits. In a small area in the eastern part of the study area, the effect of upward leakage from a semiconfined regional carbonate aquifer is evidenced by differences in water temperature, hydraulic potential, and water quality from other sites in the vicinity.

Carbon-, hydrogen-, and oxygen-isotope data indicate that major recharge occurred in the area at the end of Pleistocene and through early Holocene time. The absence of ground water older than about 17,000 yr B.P. is taken to mean that either no recharge occurred prior to that time, or ground-water velocities probably are slower than 7 m/a and possibly slower than 4 m/a. At these velocities, any water recharged prior to 17,000 yr B.P. would have been transported out of the study area by the present time. It is also possible that the slower velocity, presumably less permeable, flow paths contain this older water; however, until these are sampled separately from those of greater velocity, at shallower depth, there can be no verification of this hypothesis.

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