Prepared in cooperation with the NEVADA OPERATIONS OFFICE,
U.S. DEPARTMENT OF ENERGY, under Interagency Agreement DE–AI08–97NV12033

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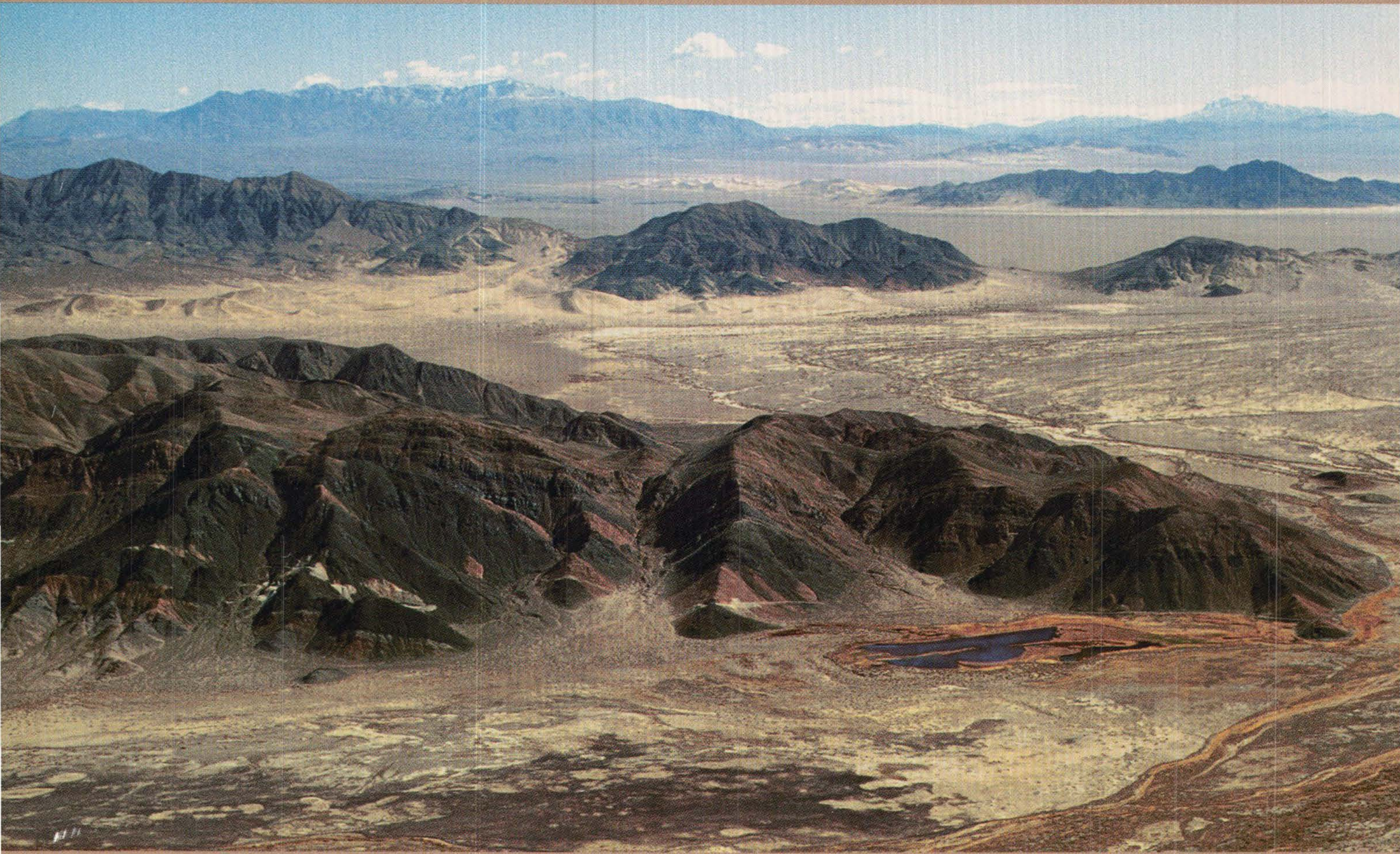
Water-Resources Investigations Report 98–4114



U.S. Department of the Interior
U.S. Geological Survey

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By William C. Steinkampf, U.S. Geological Survey, Tucson, Arizona
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U.S. GEOLOGICAL SURVEY

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Denver, Colorado
2001

U.S. DEPARTMENT OF THE INTERIOR
GALE A. NORTON, Secretary

U.S. GEOLOGICAL SURVEY
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CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATIONS

Multiply	by	To obtain
millimeter (mm)	0.03937	inch
centimeter (cm)	0.3937	inch
hectare (ha)	2.47	acre
square kilometer (km^2)	247.1	acre
kilometer (km)	0.6214	mile
liter per minute (L/min)	0.2642	gallon per minute
liter per day (L/d)	0.2642	gallon per day
meter (m)	3.281	foot

Degree Celsius (°C) can be converted to degree Fahrenheit (°F) by using the following equation:

$$^{o}F = 9/5\ (^{o}C) + 32.$$

Sea level: In this report "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)—a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

ADDITIONAL ABBREVIATIONS

mg/L	milligram per liter
pH	negative log activity of hydrogen ion concentration
pmc	percent modern carbon
µS/cm	microsiemens per centimeter at 25 degrees Celsius
µm	micrometer
µL	microliter
°C/30 m	degree Celsius per 30 meters
mmol/L	millimole per liter
$\delta^{13}C$	$^{13}C/^{12}C$ ratio, in per mil PDB (*Belemnitella americana* Cretaceous Peedee Formation of South Carolina)
$\delta^{2}H$	deuterium/protium ratio, in per mil V-SMOW (Vienna-Standard Mean Ocean Water)
$\delta^{18}O$	$^{18}O/^{16}O$ ratio, in per mil V-SMOW
$\delta^{87}Sr$	$^{87}Sr/^{86}Sr$ ratio, in per mil En-1 (USGS *Tridacna* reference)

Ground-Water Flow to Death Valley, as Inferred from the Chemistry and Geohydrology of Selected Springs in Death Valley National Park, California and Nevada

By William C. Steinkampf *and* William L. Werrell

Abstract

Death Valley lies downgradient from adjacent valleys to the north, south, east, and west in California and Nevada, and is the site of substantial ground-water discharge. The sources of the discharging waters have been discussed by several investigators in the past and are of heightened concern because of the potential disposal of high-level radioactive waste at Yucca Mountain, Nevada, and because of ground-water withdrawals attendant to commercial mining in the northwestern Amargosa Valley region. This report describes high- and low-discharge springs in and along the Amargosa Range that were sampled to augment the level of understanding of the extent and distribution of westward ground-water flow through the range.

The Black Mountains do not seem to be part of a significant path of ground-water flow from the Amargosa region. This is attributed to the complex lithology and geologic history of the Black Mountains structural block and to the presence of the intervening Furnace Creek fault zone. The only ground-water discharge associated with the Black Mountains where water chemistry reflects an external source or sources is Saratoga Spring, for which δ^2H and $\delta^{18}O$ data indicate recharge in the Spring Mountains to the east.

The southern part of the Funeral Mountains transmits a large volume of water through faulted and fractured rocks of Cambrian age that lie at or along the distal part of the northeast-oriented Spotted Range–Mine Mountain structural zone. Waters discharging from springs in the Furnace Creek Ranch vicinity (Travertine and Nevares) both compositionally and isotopically resemble waters from the Ash Meadows spring group in the Amargosa Desert. The Ash Meadows springs and water in the Amargosa Valley alluvium likely are chemically representative of ground water entering the southern Funeral Mountains. Much less ground water flows through the central and northern Funeral Mountains than flows through the southern part, as indicated by the geologic setting and chemistry of Keane Wonder Spring. The northern one-half of the mountains comprises early-to-middle Proterozoic metamorphic rocks that are the core of the Funeral Mountains anticlinorium. The core is largely unfaulted, plunges to the northeast and southwest, and is truncated to some extent on the east by the shallow-dipping Boundary Canyon fault. This structural setting and the paucity of springs in the northern one-half of the Funeral Mountains indicate a long travel-time from the Amargosa region to the western margin of the northern and central parts of the mountains.

The Grapevine Mountains include the highest elevations in the Amargosa Range. Substantial precipitation and recharge above about 2,000 meters are evinced by numerous small springs and seeps along the east and west margins. The local nature of the recharge is reflected in δ^2H and $\delta^{18}O$ values and in the spring chemistries that indicate control by Tertiary volcanic rocks. The highest spring discharges associated with the Grapevine Mountains are near

the north end of the mountains in the Grapevine Ranch area. The springs in this area are similar chemically and isotopically, except for one or two order-of-magnitude differences in calcium, magnesium, and strontium concentrations and a 1.2 per mil difference in $\delta^{13}C$ values. These differences can be attributed to differences in the distal parts of the respective flow paths. The springs also lie at the end of a northeast-oriented structural zone in the Walker Lane Belt, and their $\delta^{2}H$, $\delta^{13}C$, and $\delta^{18}O$ values indicate a recharge area likely to the northeast, outside of the Grapevine Mountains.

INTRODUCTION

Death Valley lies in southeastern California, in the southern Great Basin near the western edge of the Basin and Range physiographic province (fig. 1). It was the centerpiece of Death Valley National Monument, which was created by Presidential proclamation in 1933, and enlarged by Congress in 1994 to about 1.34 million hectares (ha) and elevated to national park status. The park is a singular preserve of biologic and geologic resources and is steeped in the history of the development of the American West. Many of the widely distributed flora and fauna depend solely on the flow of perennial and ephemeral springs for the availability and continuity of the habitats they occupy. This level of dependence is exemplified in the extreme by several endemic aquatic species that have evolved as a result of the isolated locations of individual springs they inhabit. The National Park Service (NPS) is interested in the quantity and quality of water resources in the Death Valley National Park, because it is mandated by Congress "...to conserve the scenery and the natural and historic objects and the wildlife therein, and to provide for the enjoyment of the same in such manner and by such means as will leave them unimpaired for the enjoyment of future generations." The park lies within a mineral-rich region that has a history of mining that dates to the middle of the nineteenth century. Ground-water withdrawals attendant to current mining and metal extraction in the northern Amargosa Valley region have raised concerns about the potential impact of this water use on park water resources.

Site-characterization studies are in progress at Yucca Mountain, Nevada (at the western edge of the Nevada Test Site, fig.1), the potential site of a repository for disposal of high-level radioactive wastes. Part of the characterization effort entails a description of the geohydrologic system both at the site and in the surrounding region by the U.S. Geological Survey (USGS) in cooperation with the U.S. Department of Energy (USDOE), under interagency agreement DE-AI08-97NV12033. Because fluid movement in both the saturated and unsaturated zones is a conceivable means of radionuclide transport away from a repository, an understanding of the possible rates and directions of fluid movement must be available for use in the development of estimates of potential repository performance. A fundamental aspect of understanding any geohydrologic system is knowledge of the areas where water enters and leaves the system. Estimates of the locations and fluxes of recharge and discharge are part of the data that will be used to constrain numerical simulations of conceptual geohydrologic models of the region within which Yucca Mountain lies. These calculations will, in turn, enable the formulation of better constrained boundary conditions for smaller scale models of a potential repository area.

This report provides insight to interbasin ground-water flow in the study area by using new and extant chemical and isotopic data from selected springs within Death Valley National Park and the surrounding region (fig. 2). This study was conducted to augment geohydrologic study of the extent to which the Oasis Valley–Fortymile Canyon and Ash Meadows ground-water basins discharge ground water from the northeast to Death Valley. The primary objective was to determine if hydrochemical data could provide additional insight regarding interbasin ground-water flow. About 20 springs initially were identified for sample and field-data collection, based on location, elevation, and relative discharge rates. After reconnaissance visits, 11 springs and 1 surface-water site were selected (fig. 2), based on the perceived ability to obtain representative ground-water samples. The sites were visited between March 1993 and March 1995.

The USGS National Water Quality Laboratory performed most of the analyses. June Fabryka-Martin of Los Alamos National Laboratory provided the ^{36}Cl analyses; Shannon Mahan and Kyoto Futa of the USGS provided the $^{87}Sr/^{86}Sr$ analyses; Kathy Lao of the University of Nevada, Las Vegas, Harry Reid Center for Environmental Studies provided the

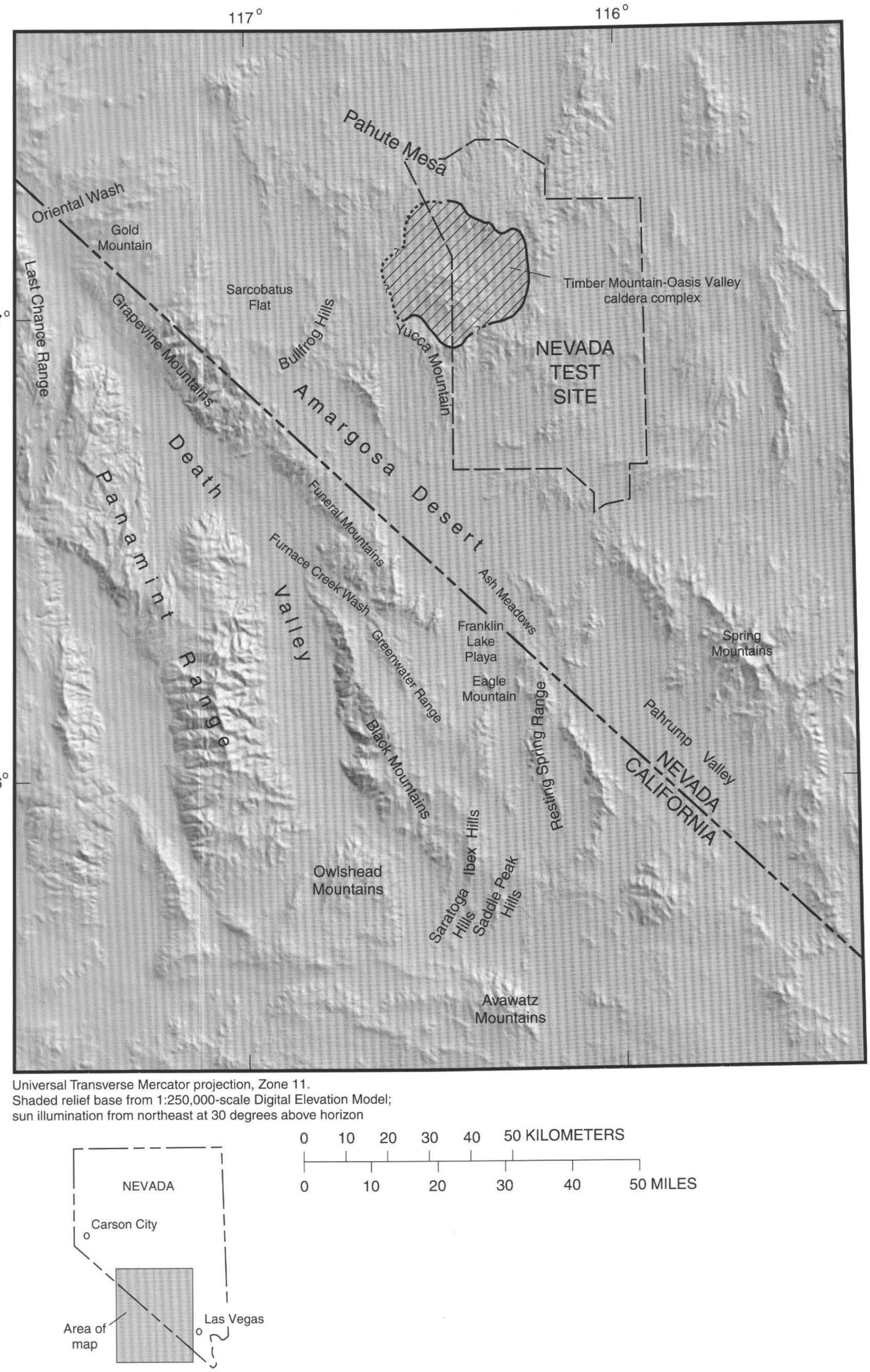
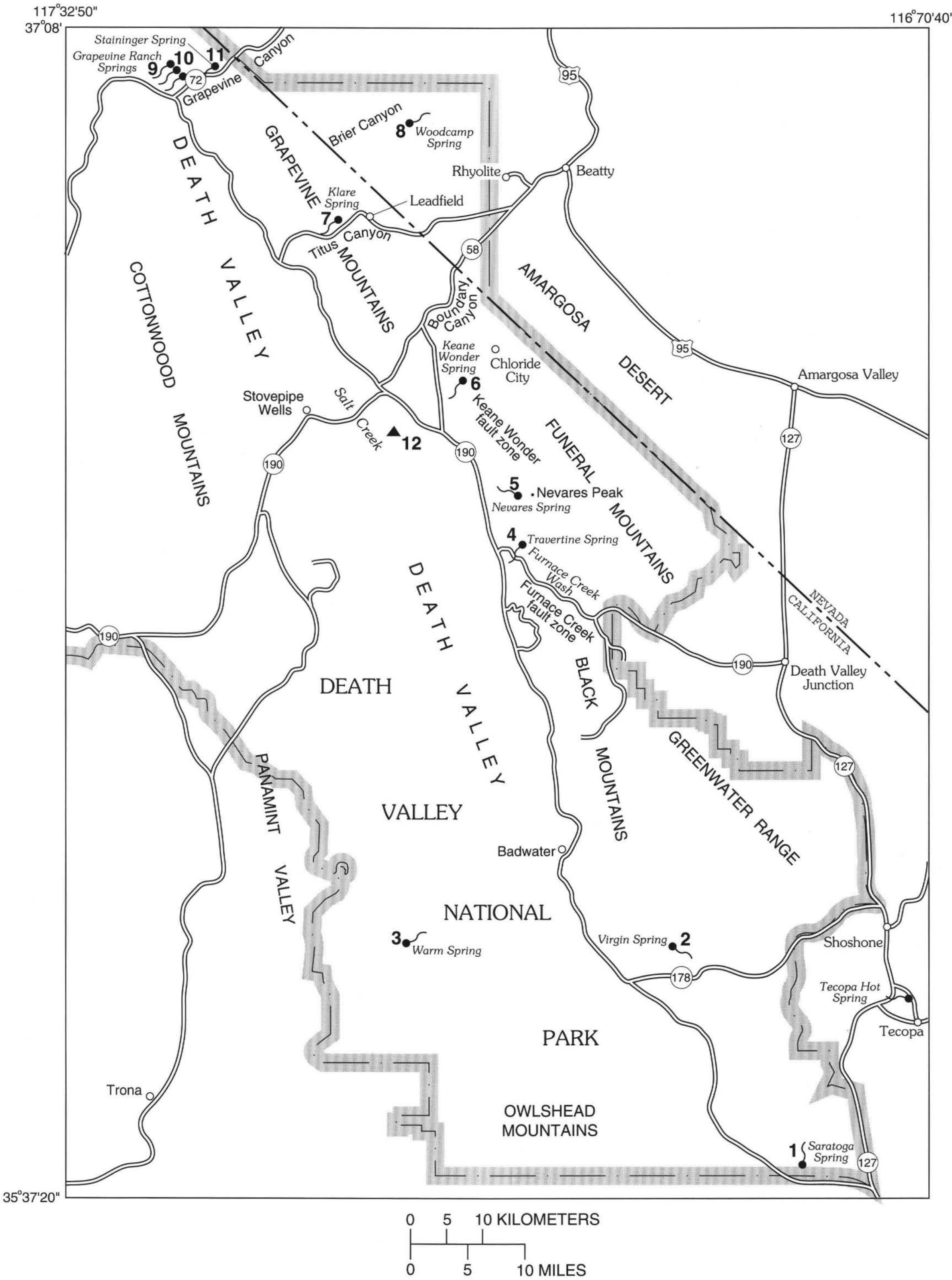


Figure 1. Shaded relief of the Death Valley region.



EXPLANATION

1 SPRINGS—Location, name, and sample number listed in table

12 ▲ SURFACE-WATER SITE

Figure 2. Death Valley National Park area and spring locations.

dissolved aluminum analyses; and the Woods Hole Oceanographic Institution National Ocean Sciences Accelerator Mass Spectrometry Facility performed the radiocarbon analyses.

Previous Investigations

Geologic investigations in the Death Valley region date from the last one-half of the nineteenth century and have been summarized by Grose and Smith (1989), D'Agnese (1994), and Faunt (1994). Studies have ranged from brief descriptions of outcrops at specific locations (Gilbert, 1875), through detailed structural and stratigraphic descriptions of areas ranging from about 250 ha (Noble, 1941), to quadrangle scale (Drewes, 1963), to U.S. Army map series (scale 1:250,000) (Ball, 1907; Jennings, 1961; Jennings and others, 1962; Streitz and Stinson, 1974).

One of the earliest geohydrologic records of the region is Mendenhall's (1909) compilation of descriptions and locations of watering places in the California and Mojave Deserts and the Death Valley region. Subsequent studies addressed specific discharge sites (Robinson, 1957), individual valleys (Waring, 1921), and regional investigations (Winograd, 1962, 1971). In addition to these studies, Thompson (1929), Hunt and Robinson (1960), Pistrang and Kunkle (1964), Hunt and others (1966), Winograd and Thordarson (1975), Miller (1977), and Bedinger and others (1989b) discussed the possibility of Death Valley ground water having been recharged elsewhere.

Geologic Framework

Together with its adjacent mountain ranges, Death Valley owes its form largely to erosion and the extensional tectonism that developed basin-and-range topography in Cenozoic volcanic and sedimentary rocks. These deposits overlie a section of plicated and eroded Proterozoic and Paleozoic rocks that are as thick as 13.7 kilometers (km) (Noble, 1941). The valley is a northwest-southeast- and north-south-oriented, closed topographic basin that is partially filled with Cenozoic alluvium and evaporite deposits. It is bounded by the Amargosa Range on the east, the Last Chance and Panamint Ranges on the west, and the Owlshead and Avawatz Mountains on the south (fig. 1). Because the Owlshead and Avawatz Mountains are not germane to this study, they will not be discussed. The geology of the Panamint Range will be addressed only with respect to the sole site visited therein.

The Amargosa Range comprises, from south to north, the Black Mountains, the Greenwater Range, and the Funeral and Grapevine Mountains (fig. 1). The sinuous crest of the range trends N. 40° W. and approximates the California-Nevada State line. The range rises 600 to 1,200 meters (m) above the Amargosa Desert and Sarcobatus Flat, and stands about 1,350 to 2,400 m above the floor of Death Valley. The entire range has a steep west margin defined by a continuous zone of normal and strike-slip faults. The east margin is bounded, from north to south, by Sarcobatus Flat, the Bullfrog Hills, the Amargosa Desert, the southern Amargosa River Valley, and the Saddle Peak Hills (fig.1).

Black Mountains

The Black Mountains comprise rocks ranging discontinuously in age from early Proterozoic to Quaternary and had a complex evolution controlled by northwest-oriented crustal extension. The mountains are bounded on the west by the Death Valley fault zone, along which the rate of uplift increased from middle Tertiary to Holocene time (Drewes, 1963). Early Proterozoic rocks that form the cores of the Black and Avawatz Mountains and Panamint Range, and that also are exposed in the Greenwater Range and Owlshead Mountains, are extensively metamorphosed sediments that, during Proterozoic time, were gently folded then intruded along a northwest-plunging axis (Drewes, 1963, p. 70). The Black Mountains block was later uplifted and eroded, resulting in a regional angular unconformity that separates these rocks from overlying Pahrump series rocks of Proterozoic age (table 1) (Hewett, 1940) that were intruded, locally altered, uplifted, and erosionally planed (Noble, 1941, p. 948). About 6 to 7 km of Paleozoic carbonate and detrital sediments were deposited on the Proterozoic basement, again along a northwest trend. Deposition of these sediments was followed by thrust faulting and igneous intrusions, based on evidence outside of the Black Mountains block (Drewes, 1963, p. 70).

In early Tertiary time, much of the Black Mountains block was gently intruded by magma from northwest-oriented magma chambers. At the same time or slightly later, the block was compressed by

Table 1. Stratigraphic and lithologic units in the Death Valley region

[Modified from Winograd and Thordarson, 1975, table 1]

Erathem	System	Series	Stratigraphic unit	Major lithology	Geohydrologic unit
Cenozoic	Quaternary/ Tertiary	Holocene/ Pliocene	Valley fill	Detrital sediments	Valley-fill aquifers
	Tertiary	Pliocene/ Miocene		Extrusive volcanic	Lava-flow and tuff aquifers
		Miocene		Tuff; tuffaceous sandstone, silts, and breccia	Tuff confining bed and minor confining bed
		Miocene/ Oligocene		Tuffaceous sandstone, siltstone, clay; continental limestone; conglomerate; tuff	
Mesozoic	Cretaceous/ Triassic		Granitic stocks, dikes, and sills		
Paleozoic	Permian/ Pennsylvanian		Tippipah Limestone	Limestone	Upper carbonate aquifer
	Mississippian/ Devonian		Eleana Formation	Argillite, quartzite, conglomerate, limestone	Upper clastic confining bed
	Ordovician		Pogonip Group	Dolomite, limestone	Lower carbonate aquifer
	Cambrian		Nopah Formation	Dolomite, limestone	
			Bonanza King Formation	Dolomite, limestone	
			Carrara Formation	Limestone	
				Siltstone	Lower clastic confining bed
			Zabriskie Quartzite	Orthoquartzite	
			Wood Canyon Formation	Sandstone, siltstone, dolomite	
Proterozoic		Late	Stirling Quartzite	Sandstone, siltstone, dolomite, limestone, schist	
			Johnnie Formation	Schist, quartzite, minor dolomite, limestone	
		Middle (Pahrump series)	Kingston Peak Formation	Conglomerate, schist, marble	Not addressed
			Beck Spring Dolomite	Dolomite, marble	Not addressed
			Crystal Spring Formation.	Schist, quartzite, marble	Not addressed
		Early		Gneiss, schist	Not addressed

right-lateral movement along the bounding Furnace Creek and Death Valley fault zones. This resulted in complex folding of the Paleozoic section and initiated movement along the Amargosa thrust fault that brecciated and telescoped a 4- to 5-km section of Middle Proterozoic through Middle Cambrian rocks, producing an interval of easterly tilted blocks. These blocks comprise a chaotic assemblage of lithologies and ages, particularly in the southern part of the mountains, where the disorder increases from east to west (Noble, 1941, p. 958). Blocks of individual lithologic units within this assemblage, designated the Amargosa chaos by Noble (1941, p. 965), range from meters to kilometers in scale and are separated by gouge zones. This activity was followed by uplift and erosional removal of most of the Paleozoic section, including part of the Amargosa chaos, and subsequent deposition of thin, middle to late Tertiary volcanic rocks and continental sediments at a second extensive unconformity throughout much of the Black Mountains (Noble, 1941; Jennings and others, 1962). Continued movement along the bounding faults resulted in additional displacement and erosion within and along the margins of the block and was accompanied by additional volcanic activity into Pleistocene time (Drewes, 1963, p. 71).

Streitz and Stinson (1974) mapped numerous normal faults exposed in the Cenozoic rocks in the northern one-half of the Black Mountains, most of which are roughly perpendicular to the Death Valley fault zone. Proterozoic rocks appear to be significantly less faulted in this part. Jennings and others (1962) mapped relatively fewer normal faults in the southern part of the mountains. These appear to be of Miocene or younger age and generally strike north-south. The more continuous of the faults tend to bound or occur within large-scale exposures of Proterozoic rocks and splay southward into the Death Valley fault zone.

Funeral Mountains

The Funeral Mountains separate Death Valley from the Amargosa Desert to the east (fig. 1). They are bounded on the northern one-half of their northeast margin by the upper plate of the gently northeast-dipping Boundary Canyon fault zone and on the south by the Keane Wonder and Furnace Creek fault zones (fig. 2). Structurally, the Funeral Mountains are a broad, doubly plunging anticline, the core of which comprises Early Proterozoic metamorphic and igneous rocks that make up the northern one-half of the mountains. These rocks are overlain by a thick section of Middle Proterozoic sediments and metasediments that, although folded along a generally northeast trend, evince relatively little faulting. The overlying Late Proterozoic rocks that comprise much of the Funeral Mountains are broken by numerous normal faults of limited displacement that tend to be perpendicular (north-northeast striking) to the bounding Keane Wonder and Furnace Creek fault zones (fig. 2). These rocks are overlain by Cambrian clastic rocks and limited Ordovician through Devonian carbonate and detrital sediments. Tertiary intrusive and extrusive rocks are present in relatively minor amounts, primarily in the northern part of the mountains, on the upper plate of the Boundary Canyon fault zone. Quaternary detrital sediments are present along the southwest and northeast margins of the Funeral Mountains as alluvial fan and stream deposits that fray and wrap eroded ridges of Late Proterozoic and Cambrian rocks, and as landslide deposits in the central part of the mountains. A zone of Mesozoic faults thrust Lower Cambrian rocks over younger Cambrian through Devonian formations in the southern part of the mountains. These rocks also have numerous normal faults, most of which are oriented perpendicular to the Furnace Creek fault zone. Carr (1984, p. 56, 61) noted that the south end of the Funeral Mountains is at the southwest end of the Spotted Range–Mine Mountain structural zone, one of several such zones in the region (fig. 3). He suggested that this zone spans the Walker Lane Belt (fig. 3) and provides a structural path for ground-water flow from northeast of the belt through the southern part of the Amargosa Desert and the southern Funeral Mountains.

Movement along the Keane Wonder fault zone (fig. 2) has resulted in 5 km of right-lateral displacement and as much as 1.5 km of vertical stratigraphic displacement (Wright and Troxel, 1993). The fault zone truncates the Funeral Mountains anticlinal structure and, north of Nevares Peak, juxtaposes Tertiary sediments in Death Valley and tilted, brecciated Proterozoic rocks. Additional strike-slip faulting essentially parallel to the Keane Wonder and Furnace Creek fault zones also seems to have occurred near the southwest margin of the Funeral Mountains block southwest of the Keane Wonder fault zone. A roughly 5-square kilometer (km^2) sub-block of carbonate rocks, comprising Nevares Peak at its northwest end and part of the wall of Furnace Creek Wash (fig. 2) at the south, appears to be displaced 1 to 2 km along a northwest trend, based on mapping by McAllister (1970) and Wright and Troxel (1993). This southwesternmost part of the Funeral Mountains comprises the Zabriskie Quartzite and Carrara, Bonanza King, and Nopah Formations. The last three are largely thick-bedded dolomites and limestones. These rocks are part of a syncline in the southeast limb of the Funeral Mountains anticlinorium. Right-lateral displacement is evinced both by the offset of units and the apparent attenuation and eastward bending of the Zabriskie Quartzite and Carrara Formation at the northwest end of the sub-block.

The Boundary Canyon fault zone (fig. 2) along the northeast margin of the Funeral Mountains dips northeastward at a lower angle than the dip of the plunging end of the anticline. Rocks in the upper plate are tightly folded, and many of the tight folds are overturned to the southwest. The folds are broken by numerous normal faults, most of which appear to be listric and to terminate at the underlying fault (Wright and Troxel, 1993).

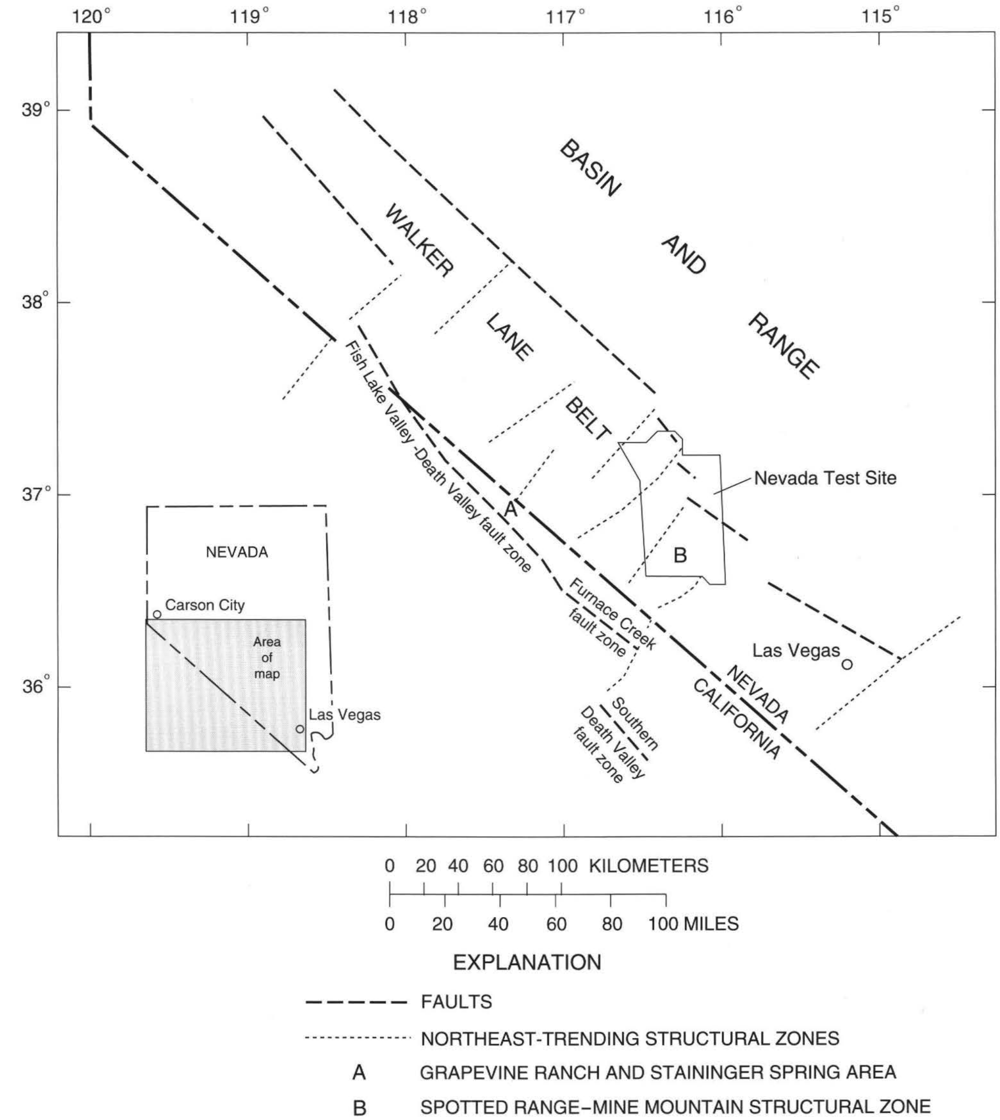


Figure 3. Death Valley region and structural zones associated with the Walker Lane Belt.

Grapevine Mountains

The Grapevine Mountains are the northernmost part of the east boundary of Death Valley. They are bounded on the south by Boundary Canyon (fig. 2), by Oriental Wash and Gold Mountain to the north (fig. 1), and by the Death Valley–Furnace Creek fault zone on the west (fig. 3). The mountains comprise Late Proterozoic through Mississippian marine carbonate and clastic rocks and Cenozoic volcanic and continental sedimentary rocks (Strand, 1967; Reynolds, 1976) that comprise the folded and faulted northwest limb of a large anticline, the domal core of which is 2 to 3 km southeast of Boundary Canyon (Wright and Troxel, 1993). Late Tertiary through Quaternary sediments are present as alluvial fans and other detrital deposits. The southern part of the Grapevine Mountains was uplifted during late Tertiary time and faulted along north to northeast trends. Proterozoic and Paleozoic rocks were warped and folded into a syncline and adjacent anticline along generally northwest and north axes that plunge to the northwest. The degree of folding increases from south to north, to the extent that the folds are overturned to the west in the northern two-thirds of the mountains. The synclinal rocks form essentially the western part of the mountains and the anticlinal rocks form the eastern part. These structures were extensively faulted along a north trend during folding (Reynolds, 1969). In the north-central part of the mountains, Reynolds (1969) mapped a thrust fault that juxtaposes a Late Proterozoic- through Cambrian-age allochthon over units as young as Mississippian age. In the northern one-third of the mountains, faulting is generally less extensive. The faults are normal and roughly parallel the Death Valley–Furnace Creek fault zone (fig. 3) (Strand, 1967; Streitz and Stinson, 1974).

Based on mapping by Ball (1907, p. 164–173) and Oakes (1977), and the Strand (1967) compilation, the northernmost part of the Grapevine Mountains comprises primarily Cenozoic volcanic rocks in the northeast two-thirds and Cambrian, Ordovician, and perhaps minor Devonian limestones, dolomites, quartzite, and siltstone in the southwest one-third. South of Grapevine Canyon (fig. 2), the volcanic rocks are rhyolite and biotite latite that include interlayered basalts in the lowermost part of the sequence. North of the canyon, about one-half of the remainder of the Grapevine Mountains is covered by younger basalts that have interflow zones that must have transmitted water in the past, based on Ball's (1907, p. 171) description of secondary minerals present. The basalts were emplaced on the erosional surface of a thick sequence of Miocene rhyolites. Ball's (1907, p. 168–169) description of these rhyolites suggests that they are the middle-to-late Miocene lavas and tuffs of Rainbow Mountain, as identified by Maldonado (1990) at the south end of Sarcobatus Flat. The basalts and rhyolites, based on examination of aerial photographs, evince extensive post-late-Miocene faulting along a trend normal to the Death Valley fault zone (fig. 3) (W.J. Carr, U.S. Geological Survey, oral commun., 1996). The rhyolites unconformably overlie Paleozoic carbonate rocks that crop out along about 6 km of the west margin of the mountains, southwest of Sarcobatus Flat (Ball, 1907, p. 164–165, pl. 1; Oakes, 1977).

GEOHYDROLOGY

Winograd and Thordarson (1975) assigned groups of stratigraphic units in the Death Valley region to general geohydrologic units in their discussion of the geohydrology of the south-central Great Basin (table 1). Their lower clastic confining bed (called aquitard in their report) is exposed in the south end of the Grapevine Mountains, in the northern one-half of the Funeral Mountains and, to a very limited extent, in the northern Black Mountains (Streitz and Stinson, 1974; Winograd and Thordarson, 1975, pl. 1). Some of the units composing the lower carbonate aquifer described by Winograd and Thordarson (1975) have been mapped in the southern one-half of the Grapevine and Funeral Mountains on the flanks of the earlier noted regional anticline (Strand, 1967; Wright and Troxel, 1993).

Hunt and others (1966, p. B29), in a detailed study of the Death Valley saltpan, noted four general types of springs in the Death Valley region, two of which are relevant to this study. Springs of the first type such as Saratoga, Nevares, Travertine, Keane Wonder, and those in the Grapevine Ranch area (fig. 2) have the highest discharges in the Park, hundreds to thousands of liters per minute, and are associated with large high-angle faults. Springs within the bounding mountain ranges discharge at lower rates and tend to occur along lower angle faults. Klare and Virgin Springs (fig. 2) are examples of this second type. The third and fourth types occur on the valley floor, one where less permeable lithologic units locally raise the water table upgradient and result in spring discharge at some adjacent downgradient outlet, and the other where water ponds in sands and gravels that grade to silt under the lowest parts of the valley floor. Examples of these types are McLean Spring, which feeds Salt Creek, and the numerous seeps and springs along the margins of the saltpan in the Badwater area.

Hunt and Robinson (1960, p. B273) suggested that the major springs in the Furnace Creek area of Death Valley discharge water that moved through the lower-to-middle Paleozoic limestones and dolomites that compose Winograd and Thordarson's (1975) lower carbonate aquifer. Hunt and others (1966, p. B38–B40) noted that the Furnace Creek area springs were chemically similar to those at Ash Meadows in the Amargosa Desert and suggested that Pahrump Valley to the east was the source of the Ash Meadows discharge. Winograd and Thordarson (1975) presented hydrologic and chemical information indicating that the Furnace Creek springs are related to water in the Amargosa Valley alluvium, rather than that in Pahrump Valley, but refrained from a conclusive statement about interbasin flow, citing the need for additional water-level data. Miller (1977, p. 28–32) summarized various geologic and hydrologic discussions, the consensus of which was that water discharging from the spring groups at the west end of Furnace Creek Wash flows from basins to the north and east, through the southern part of the Funeral Mountains. Carr (1984, p. 61) suggested that ground water moving through the Spotted Range–Mine Mountain structural zone (fig. 3) discharges in both the Ash Meadows and Furnace Creek areas. He also discussed several other northeast-trending structural zones in the Walker Lane Belt, some of which terminate at the Fish Lake Valley–Death Valley–Furnace Creek fault zone (fig. 3) (Carr, 1984, p. 11, 30). D'Agnese (U.S. Geological Survey, oral commun., 1996) suggests that the high-discharge springs in the northern Grapevine Mountains also are at the southwest terminus of one of these zones.

The Grapevine Mountains include the highest elevations in the Amargosa Range; about one-half of their area is above 1,500 m. The east-central part of the mountains, between Grapevine Canyon on the north and Titus Canyon on the south (fig. 2), contains the only forested areas in the Amargosa Range. Almost all of this forested part is above 2,000 m. Hunt and Robinson (1966, p. B7) estimated between 130 and about 250 millimeters (mm) of annual precipitation above this elevation in the Death Valley region. This precipitation likely is responsible for the presence of numerous small springs and seeps that issue along the west-central margin of the mountains west of this high area, and for somewhat larger springs in the eastern one-half of the mountains. The western springs discharge from Cenozoic rocks variously identified as undivided Tertiary sediments and volcanic rocks by Reynolds (1976, p. 20) and as Pliocene-Pleistocene continental sediments by Streitz and Stinson (1974). The springs along the eastern margin of the mountains discharge from early Miocene rhyolites (Ball, 1907, pl. 1).

The Black Mountains and Greenwater Range contain no high-discharge springs. Johnson and others (1989) reported that springs and seeps in the Black Mountains yielded widely variable amounts of water, that the discharge range of the most productive spring was from 0.56 to 29 L/min, and that all other sites studied had maximum flows of less than about 2 L/min. Their results indicate that all of the springs within the Black Mountain block derive their flows from the limited local precipitation that falls within the Black Mountains proper. The lack of continuous discharges greater than about 0.25 L/min in the Black Mountains within about 20 km of the Furnace Creek fault zone (Johnson and others, 1989) strongly indicates that the zone is an effective barrier to ground-water flow from the north and east. Saratoga Springs on the western margin of the southern tip of the Black Mountain block, but not considered part of the Black Mountains proper, issue along part of one of several north-south normal faults that splay south into the Death Valley fault zone from the Saratoga Hills (fig. 1) and the adjacent Ibex Hills. Miller (1977, p. 27) reported discharges on the order of 300 L/min measured at what he termed the main spring at Saratoga between April 1967 and November 1968. The lack of high-discharge springs in the Black Mountains block appears to reflect its overall low elevation and attendant scarce precipitation and, together with the complex geologic history of the area, indicates that transrange permeability is too low and discontinuous to enable significant through flow of regional ground water from the Amargosa Valley that might traverse the Furnace Creek fault zone. Movement along this fault zone likely has developed a large broken zone in the adjacent bedrock that is relatively permeable and could, if driven by the large potential gradient from the Amargosa Desert to Death Valley, direct ground-water flow to the northwest.

If the geologic history of the Black Mountains and Greenwater Range established a geohydrologic framework within the block that is conducive to ground-water flow from the saturated rocks beneath the southern Amargosa Valley to Death Valley, two possible scenarios appear to exist. Either the mass of water moving through the block is insufficient, relative to overall block permeability, to result in a potentiometric surface high enough to yield large springs, even given the large hydraulic gradient between the two valleys; or the overall permeability is relatively large and ground water moves readily through the block to Death Valley. Examination of the Death Valley water budget could perhaps cast some light on this subject. Uncertainty as to the geologic structure and stratigraphy beneath the Amargosa Valley alluvium, however, limits speculation about the region from which ground water would have to flow to enter the northern part of the Black Mountains.

Field-Data and Sample Collection

Water samples for analyses of dissolved cations, anions, and selected stable isotopes and radioisotopes were collected at each sampling site. Sample collection included onsite: 0.1-micrometer (μm) filtration using a peristaltic pump and acid-rinsed tubing and filter apparatus, acidification with ultrapure nitric acid, precipitation of dissolved inorganic carbon with ammoniacal barium or strontium chloride, extraction of dissolved aluminum using the method of Barnes (1975, p. 179–180), and preservation of radiocarbon samples by the addition of 100 microliters (μL) of saturated mercuric chloride. Field-data collection included onsite determinations of pH, water temperature, and dissolved inorganic carbon species. Temperatures were recorded to 0.01 degree Celsius (°C) and rounded to the nearest 0.05°C; pH was measured to ±0.002 and rounded to the second decimal.

Data- and Sample-Collection Sites

Saratoga Spring (site 1, fig. 2) is one of a group of springs at an elevation of about 60 m above sea level, on the west side of the Saratoga Hills, at the south end of the Black Mountains, and at the extreme south end of Death Valley proper (fig. 1). The springs discharge from upper Proterozoic rocks (Gilbert, 1875, p. 170; Campbell, 1904, p. 14) along a north-south fault that bounds the west margin of the Saratoga Hills. Miller (1977) suggested that, based on similarities in chemistry, the springs discharge water from a ground-water basin in the Shoshone-Tecopa area (fig. 2), about 30 km to the northeast and about 400 m higher in elevation. The spring area contains several distinct discharges within an approximately 5-ha marshy area that contains several perennial ponds. The marsh and ponds provide habitat for numerous genera of birds and at least one endangered species of fish. Two discrete discharges were observed during site visits. One is a boil in a broad, shallow pond at the southeast edge of the area. The second is from a four-sided, open concrete box inset in the north face of a saddle-like knob of quartzite-topped diabase that projects west into the valley at the south end of the main marsh area. This source appears to issue from a diabase sill in the Crystal Spring Formation of the Pahrump series. The box is about 60 m north and 40 m west of the north end of the parking area at the end of the road to the spring and was the site of sample and data collection. The discharge was about 100 liters per minute (L/min) through a tile that drains the box into a narrow channel that leads to a nearby pond. Although the entire spring area likely discharges on the order of 1,000 L/min, Miller (1977, p. 27) reported a discharge of about 300 L/min for the main spring based on discrete and intermittently continuous measurements made between April 1967 and November 1968. This highlights a problem attendant to discussions of hydrologic and hydrochemical information for sites such as this that have several discharge points, in that commonly the specific site sampled or measured is not clearly identified. Differences in specific sample locations at a site such as this can result in data differences or uncertainties, particularly for unstable parameters determined in the field.

Virgin Spring (site 2) is at an elevation of about 735 m, about 0.5 km up a northwest-trending side canyon that is about 2.5 km up Virgin Spring Canyon from the head of Jubilee Wash (fig. 2). The spring is situated on the northwest flank of an anticlinal exposure of Early Proterozoic rock. It issues from an altered gouge that marks the strike of a fault that juxtaposes Proterozoic rocks of the basal plate of the Amargosa thrust fault and the lowermost phase of the overlying Amargosa chaos. The site is a collection basin in the rear one-half of an adit about 10 m long. The adit was constructed early in the twentieth century to increase the availability of water from a seep to provide a supply for mining or prospecting efforts in the vicinity, as old storage tanks and pipes are in evidence.

Based on local topography and the magnitudes of historical discharges, Virgin Spring likely represents local winter precipitation that has infiltrated within an area of perhaps 20 km^2. NPS staff modified the collection pond in the early 1990's to include a constant-level discharge tube that supplies a small stock tank installed to provide a continuous water supply for wildlife. Johnson and others (1989, p. 11) reported historical discharge measurements from 1959 through 1987 that ranged from no flow to about 4 L/min. Estimated discharges from the stock tank during several visits by the authors between April 1992 and December 1994 ranged from near zero to about 0.1 L/min. The spring elevation is about 100 m higher than the elevation of the Amargosa River channel at the south end of the Amargosa Desert, and about 280 m higher than the springs in the vicinity of Shoshone, California, to the east. It is highly unlikely that this site discharges water recharged outside of the Black Mountains.

Warm Spring (site 3, fig. 2) was selected as a sampling site because of NPS interest, and to provide a contrast of sorts, as it likely does not discharge water recharged east of or within the Amargosa Range. It issues at an elevation of about 760 m in a canyon in the southern part of the Panamint Range. The canyon is fault-controlled, and its floor and walls are Permian limestone. The canyon is surrounded primarily by middle to late Mesozoic metavolcanic rocks, a large Mesozoic granite intrusive, and numerous Tertiary andesites (Jennings and others, 1962). The limestone contains a basaltic intrusion that produced extensive talc deposits, the mining of which only recently has been abandoned. The spring issues from a cleft in the nodular (chert) limestone in the south wall of the canyon. The site was modified sometime in the past by inserting a roughly 70-mm inner-diameter flexible pipe some distance into the cleft and grouting the lower meter of the cleft with local soil and rock. Discharge from the spring was about 60 L/min.

The site herein identified as Travertine Spring (site 4, fig. 2) is one of the group of springs in Furnace Creek Wash known collectively as Travertine Springs, one of several sites of substantial discharge in the park. The source is near the east margin of the group and did not appear to have been anthropogenically modified, as are many in this group. It discharges about 20 to 40 L/min at an elevation of about 122 m. Miller (1977, p. 27) reported a discharge of 4,500 L/min in January 1977 for the spring group. Subsequent to his visit, four of the Travertine Springs sources and one from the Texas Spring group about 2 km to the northwest were modified by means of ditches, buried tiles, or French drains, and plumbed to a common discharge line that provides the water supply for the Furnace Creek Ranch area.

The Travertine Springs group discharges southwest of the northwest-oriented axis of Furnace Creek Wash, from unconsolidated Quaternary alluvium that overlies conglomerates, sandstones, and mudstones of the Funeral and Furnace Creek Formations of early Quaternary and late Tertiary age. McAllister (1970) mapped the strikes of northwest-trending normal faults in these sediments at and near the north margin of the wash. Northeast of the spring group, these faults are parallel to normal faults in the adjacent Paleozoic rocks of the southern Funeral Mountains. McAllister (1970) did not detect surface expression of faults, however, along about 3 km of the wash margin, about 5.5 km east of Travertine Spring. The northwest end of this apparent break in faulting is coincident with the southeast end of the sub-block of the southern Funeral Mountains, discussed in the "Geologic Framework" section, that appears to have undergone northwest-oriented right-lateral displacement. The absence of faults along this part of the wash margin, the absence at land surface of the relatively less permeable Furnace Creek Formation, and the general absence of faults in the northwestern one-half of the wash (McAllister, 1970) indicate that ground water discharging at the spring group flows through this gap from the extensively faulted Paleozoic rocks along the axis of the synclinal sediments. Pistrang and Kunkle (1964, pl. 1) mapped a limited fault immediately northeast of the Travertine Springs group line that perhaps determines the location and linearity of the group.

Nevares Spring (site 5, fig. 2) is the water supply for the NPS and the California Department of Transportation cantonment and operational sites at Cow Creek, north of the Furnace Creek area. Springs issue at various points on and around a travertine mound developed on Quaternary alluvium. Miller (1977, p. 27) reported a total discharge of about 980 L/min. The spring mound abuts the west foot of Nevares Peak (fig. 2), a faulted assemblage of Cambrian dolomites, limestones, and orthoquartzite. The peak is the northwest end of a fault block that has undergone right-lateral movement parallel to the Keane Wonder and Furnace Creek fault zones (fig. 2). Northwest-oriented faults terminate at the spring mound according to Wright and Troxel (1993). Their geologic cross section through this area also shows that the Nevares Peak rocks are in fault contact with poorly consolidated Tertiary lake sediments along the southwest flank of the peak. These sediments provide a permeability contrast that could direct ground-water flow upward in the vicinity of the contact.

Two French drains were constructed by the NPS in the central part of the eastern one-half of the mound and plumbed using about 0.2-m inner-diameter polyvinyl chloride (PVC) pipes to a concrete collection box. Samples were collected from the easternmost pipe entering the collection box. This sampling location appears to be representative of spring discharge because isotopic (δ^2H, $\delta^{18}O$, $\delta^{87}Sr$) analyses of a preliminary sample collected during site reconnaissance matched extant historical data describing samples collected at various discharge points at the rear of the mound.

The Salt Creek site (site 12, fig. 2) is an abandoned USGS streamflow-gaging station (downstream order number 10251100) about 12 km southeast of the Stovepipe Wells Hotel, at an elevation of about –55 m. The creek flows from and through a marshy area south along the valley-floor axis from McLean Spring, which could not be located during reconnaissance. Discharge at the gaging station ranged from a summer low of 60 L/min to a winter high of 10,280 L/min between 1974 and 1986 (U.S. Geological Survey, 1987, p. 39). The wide range of discharge values reflects the extremes of runoff typical of desert storm events in large drainage areas. The sample collected essentially reflects a base-flow period. The water is saline, likely undergoes evaporative concentration during flow through the sediments filling Death Valley, and conceivably represents waters entering the valley from the east, west, and north.

This site is of minimal use in discerning the contribution of adjacent basins to discharges in Death Valley for several reasons. First, extant data provided no means to resolve the potential contributions from the three possible directions of inflow, as the same types of rocks are present in essentially all directions upgradient from the site. This precludes specific lithologic insight that might aid in an examination of chemical evolution of the water. Second, flow through the valley alluvium would tend to attenuate or overprint chemical and most isotopic information deriving from water entering the alluvium. Third, the evaporative concentration during flow through the Death Valley alluvium further compounds the difficulty in attempting to address the complex mixing and mixed-lithology issues. For these reasons, the Salt Creek data are not discussed in this report, beyond including them in the tables and in selected figures.

Keane Wonder Spring (site 6, fig. 2) is the largest of numerous seeps and small springs that discharge along a roughly 2-km section of the Keane Wonder fault zone. The spring area is along the south truncated margin of the Funeral Mountains anticline, about 2 to 3 km south of the anticlinal core, at an elevation of about 380 m. The spring area is associated with a normal fault in part of the northern one-half of the Keane Wonder fault zone. Discharges are from an outcrop of the middle member of the Crystal Spring Formation (Wright and Troxel, 1993), a calcite marble with subordinate pelitic schist layers, that is part of the Funeral Mountains margin. The spring area is mantled by a buff to tan travertine that occurs from as much as several hundred meters up the slope to several hundred meters onto the valley alluvium and extends about 2 km northwest. The block of Crystal Spring Formation from which water issues is bounded on both the northeast and southwest by west-dipping normal faults that have dropped the block vertically about 200 m, juxtaposing it with Miocene sedimentary rocks in the valley floor (Wright and Troxel, 1993, section C-C'). The sampled spring is the highest noted in the local spring group. This spring discharges about 130 L/min and was one of several local sources that supported gold mining and attendant activities at the nearby Keane Wonder mine early in the twentieth century.

Klare Spring (site 7, fig. 2) is at an elevation of about 950 m and issues from fractured and healed Carrara Limestone (Reynolds, 1976, p. 23) (Carrara Formation in table 1) in Titus Canyon in the central Grapevine Mountains. The spring discharges about 100 L/min at the east end of a vegetated area at the base of the north wall of the canyon and supports about 0.1 ha of dense reeds, shrubs, and small trees. It has been a historical watering site, as evinced by adjacent Native American petroglyphs. The spring site is in the broken zone of the Titus Canyon fault, a normal, east-dipping fault that flattens to the south and steepens to the north (Reynolds, 1976, p. 24). South of Klare Spring, the fault approximates the axis of the recumbent anticline that constitutes much of the eastern one-half of the Grapevine Mountains. In the southern and central Grapevine Mountains, the anticlinal core comprises rocks of the Wood Canyon Formation (Reynolds, 1976, p. 23, section A), primarily feldspathic sandstone with subordinate siltstone and dolomite. Tertiary volcanic and sedimentary rocks cover about three-fourths of the roughly 1,400-ha drainage area that, based solely on topographic divides, contributes to the springflow. Additionally, a series of north-trending normal faults are present in and adjacent to this basin. North from Titus Canyon, these faults displace the Tertiary rocks and likely direct the movement of recharged waters downgradient to the south.

The spring elevation is about 300 m lower than the land surface east of the Grapevine Mountains, providing a conceivable potential gradient for east to west ground-water flow. The intervening crest of the mountains, however, rises to over 2 km, and several springs occur on the eastern flank of the Amargosa Range northeast of Klare Spring. One of these, at an elevation of about 1.9 km in Brier Canyon (fig. 2), discharges considerably more water than Klare Spring. This spring in Brier Canyon is more than 600 m higher than the center of Sarcobatus Flat to the northeast (fig. 1) and likely reflects recharge from the wooded highlands to the west. Isotopic data from the spring could confirm or disprove this possibility. The relatively low discharge of Klare Spring, its low water temperature, and its elevation relative to the crest of the range in this vicinity indicate that the spring discharges water recharged to the north in the same highlands that likely feed the springs along the margins of the Grapevine Mountains.

Woodcamp Spring (site 8, fig. 2) is the only site sampled east of the crest of the Amargosa Range. It is one of the marginal springs mentioned earlier and is an improved spring characteristic of locales in the region where seeps were modified, generally using French drains, to capture a usable amount of water. The spring issues at an elevation of about 1.49 km in the southwestern part of Sarcobatus Flat, at the foot of the eastern flank of the Grapevine Mountains. The spring discharges 20 to 40 L/min at the base of a large sagebrush from plastic tubing that has been inserted into a larger (about 40-mm inner diameter) metal pipe. The discharge flows through about 15 m of shallow, watercress-clogged channel to a small pond on the floor of the adjacent alluvium-filled shallow wash. The French drain that feeds the discharge was constructed near the distal end of one of several low, fingerlike ridges of rock identified by Maldonado (1990) as Miocene ash-fall tuff and rhyolite lavas of Rainbow Mountain. The ridges are the easternmost surficial extent of the same units that appear to compose the south wall of Brier Canyon to the west (fig. 2) and that were identified as Tertiary volcanic rocks by Streitz and Stinson (1974) in California along about 18 km of the California-Nevada State line to the south and west of the spring.

Grapevine Springs (sites 9 and 10, fig. 2) is the term that has been used to identify the group of perennial springs that discharge a total of about 1,680 L/min (Miller, 1977, p. 33) in the Grapevine Ranch area in the northern Grapevine Mountains. The springs are distributed atop a travertine-mantled terrace at the south end of a northwest-southeast-oriented outcrop of lower Paleozoic carbonates. The roughly 60-km^2 exposure of carbonate rocks is bounded on the west by the Death Valley fault zone and is overlain by Tertiary volcanic rocks (Oakes, 1977, pl. 1). Examination of aerial photographs of this area indicates that the springs are related to large-scale local joints, or a series of faults, or both, that are parallel or subparallel to the fault zone that bounds this area on the west. The extensive travertine that covers almost the entire area downslope from the highest vegetated areas, combined with surface exposures of large joints in the limestone that are filled with coarsely crystalline banded calcite, strongly indicates that the Grapevine Ranch area has been a discharge site for tens to hundreds of thousands of years. Only one discrete source was found in the area (site 10, fig. 2), at an elevation of about 825 m, emanating from a specific point. All others examined are at the ends of (site 9, fig. 2) or along narrow, shallow ditches and conduits fed by indistinct upgradient sources. Channel and subsurface flow are indicated by sinuous, dense growths of low grasses and brush that lead uphill to areas of dense brush and scrub. Flowing water can be observed in some of these densely vegetated areas, but 3 days of reconnaissance yielded only the single, above-noted discrete discharge.

Staininger Spring (site 11, fig. 2) is in alluvium that floors Grapevine Canyon, about 1.5 km up the canyon from Scotty's Castle, at an elevation of about 975 m. The canyon has been incised in Tertiary volcanic rocks that unconformably overlie Paleozoic limestones. Springs rise in the canyon alluvium at several locations and were the water supply for Scotty's Castle, a historical site and tourist attraction situated on what was formerly the Staininger Ranch. This spring variously has been referred to as Scotty's Spring and Scotty's Castle Spring. The name Staininger Spring is used by the U.S. Geological Survey. It reflects the earliest known non-Native American use of the water by the Staininger family, whose ranch was noted as a source of water early in the century by Ball (1907) and Mendenhall (1909, p. 31). Total discharge from all of the springs is about 700 L/min and has been reported to be temporarily influenced by rainfall runoff (Miller, 1977, p. 32). The occurrence of the individual springs in the canyon may be related to the proximity of limestone of the Pogonip Group that underlies the area, as the lowest spring issues at an elevation of about 925 m, the general elevation of the eastern margin of the large Pogonip limestone outcrop that includes the Grapevine Ranch area. The limestone also crops out at Scotty's Castle and reaches a maximum elevation of about 935 m in the south canyon wall. Additionally, several springs south of Staininger Spring issue at about this same elevation. The sampling site has the highest discharge in the group. The spring is enclosed by a springhouse and concrete footwall and flows through a concrete flume into the plumbing system that supplies the Scotty's Castle complex. Also plumbed into the lateral and upstream faces of the footwall are delivery pipes from

French drains upstream and on either side of the springhouse. The main spring is evinced by a large boil in the center of the springhouse pool, from which a sample was collected using tubing suspended about 5 cm above the pool bottom.

WATER CHEMISTRY OF SELECTED SITES

Investigation results are discussed in the following paragraphs for each of the mountain ranges that comprise the east boundary of Death Valley. Table 2, which contains the data for all sites, and several figures are common to the discussions and will be referred to frequently. Several locales within the region, including conceptual regional recharge and discharge areas and the Nevada Test Site and Environmental Research Park (NTS) and vicinity, have been used for comparative purposes. Data references for these locales are provided as appropriate.

The Tecopa Hot Spring data are from Thomas and others (1997) and are not included in this summary of table 2 data. Most of the springs sampled discharge relatively dilute waters, except for Saratoga and Keane Wonder Springs and the site at Salt Creek. Dissolved solids (residue at 180°C) ranged from 254 to 22,700 mg/L and, excluding the Salt Creek value, averaged 1,040 mg/L; 9 of the 12 values ranged from 254 to 706 mg/L (table 2). Water temperatures ranged from 12.80°C at Salt Creek, to 39.40°C at Nevares Spring. If the surface-water sample from Salt Creek and samples from sites having small discharges (Woodcamp and Virgin Springs) are excluded, temperatures ranged between 24.80 and 39.40°C, with a mean of 33.20°C.

The chemistries of the springs vary substantially, but each falls within a general chemical type that appears to reflect lithology. The host rocks include Tertiary tuffs, carbonate rocks that range from Proterozoic to Permian dolomites, limestones, and marbles, and Proterozoic metasediments and basalts. The relative concentrations of the major dissolved constituents, when plotted on a trilinear diagram (Piper, 1944), fall into regions characteristic of waters influenced by igneous and carbonate rocks and also show evaporative evolution toward high-chloride waters (fig. 4). The specific locations on the diagram are determined by several factors, including the mineralogy, porosity, and permeability of the host rocks, ground-water residence time and flow path, and recharge-water composition. Although variations in rock chemistries render absolute inferences essentially impossible, relative plot location provides insight to the controls on water chemistry at each site. For this reason, several areas have been indicated in the figure to identify waters typical of specific lithologies in the Death Valley region. Additionally, the general change in position with chemical evolution is indicated for the Spring Mountains and Yucca Mountain area waters. The data in figure 4 are discussed in the following sections on specific sampling sites.

Black Mountains

Virgin and Saratoga Springs were the only sites sampled in the Black Mountains region. As noted earlier, Virgin Spring discharges only a small amount of water. The relatively insignificant discharge, together with its water chemistry, confirms that it is recharged within the Black Mountains. Plotting its chemistry on the trilinear diagram (fig. 4) demonstrates that it is chemically less evolved than the other sites and influenced somewhat by dissolution of carbonate rock. Fresh ground waters in noncarbonate terranes generally evolve from a calcic to a sodic dominance in the cation suite (Jones, 1966; Wallick, 1981; Hearn and others, 1985; Gislason and Eugster, 1987). This water, like nearly all of the others sampled, has a Na/Ca equivalent ratio greater than 1 (fig. 5). This is characteristic of ground waters associated with noncarbonate rocks, which can have ratios ranging from about 0.4 for recently recharged waters to greater than 10 for more-evolved waters. For example, a water-supply well near the western margin of the NTS, UE–25 J#13, was completed in Miocene tuffs and has a Na/Ca ratio of 2.82 (LaCamera and Westenburg, 1994, p. 123). This site can be considered typical of intermediate evolution of ground water in extrusive silicic volcanic rocks. Recently recharged waters in limestones in the Spring Mountains have ratios of about 0.02 (Thomas and others, 1997, App. B). Na/Ca ratios can be influenced by ion exchange, mixing with more sodic water, or, in the case of many waters in southern Nevada, prior flow through noncarbonate rocks. The location of the Virgin Spring data on the trilinear diagram (fig. 4) evinces the influence of the adjacent Proterozoic rocks. The strontium isotopic value ($\delta^{87}Sr$ of 18.38 per mil) is the second-highest noted in this study (table 2),

Table 2. Physical and chemical data for selected springs and Salt Creek

[Tecopa Hot Spring data are from Thomas and others (1997). °C, degrees Celsius; µS/cm, microsiemens per centimeter at 25 degrees Celsius; mg/L, milligrams per liter, µg/L, micrograms per liter; ‰, per mil; pCi/L, picocuries per liter; pmc, percent modern carbon; --, not analyzed; <, less than; Kjd, kjeldahl]

Site	Site number (see figure 2)	Latitude	Longitude	Date	Time	Temperature (°C)	Lab specific conductance (µS/cm)	Field pH units	Residue 180°C (mg/L)	Ca (mg/L)	Mg (mg/L)
Saratoga Spring	1	354053	1162518	03-18-93	1730	28.30	4,710	7.71	3,530	32	34
Virgin Spring	2	355713	1163504	03-18-93	1000	21.00	890	7.74	570	54	35
Warm Spring	3	355800	1165552	03-07-95	1525	34.60	639	7.58	437	61	20
Travertine Spring	4	362632	1164938	03-08-95	1430	35.25	955	7.39	601	33	18
Nevares Spring	5	363045	1164916	03-15-93	1450	39.40	977	7.39	619	42	20
Keane Wonder Spring	6	364027	1165511	03-09-95	1300	34.10	4,610	6.65	3,040	90	35
Klare Spring	7	365034	1170535	05-30-93	1330	24.80	873	7.59	532	44	24
Woodcamp Spring	8	365809	1165838	03-16-93	1700	19.15	352	7.16	254	23	3.3
Grapevine Ranch Spring #1	9	370123	1172302	05-29-93	1230	37.80	1,090	7.05	688	52	18
Grapevine Ranch Spring #3	10	370139	1172307	05-29-93	1545	38.65	1,120	7.00	706	51	20
Staininger Spring	11	370156	1171929	03-17-93	1100	26.00	705	8.32	458	4.6	0.52
Salt Creek near Stovepipe Wells, Calif.	12	363558	1170046	03-16-93	1030	12.80	27,600	8.19	22,700	120	410
Tecopa Hot Spring	None	365219	1161350	06-30-85	1400	42.0	1,650	8.2	--	4.0	1.5

Site	Na (mg/L)	K (mg/L)	Al (µg/L)	Ba (µg/L)	Be (µg/L)	Cd (µg/L)	Cr (µg/L)	Co (µg/L)	Cu (µg/L)	Fe (µg/L)
Saratoga Spring	960	33	<10	15	<1.5	<1	<15	<1	<1	<9
Virgin Spring	92	6.1	<10	51	<0.5	<1	<5	<1	2	<3
Warm Spring	32	3.5	--	48	<0.05	<1	<5	<3	<1	<3
Travertine Spring	140	12	--	40	<0.05	<1	<5	<3	<1	<3
Nevares Spring	140	11	<10	42	0.6	<1	<5	<1	<1	4
Keane Wonder Spring	884	48	[1]23	24	<1.5	<3	<15	<9	<30	48
Klare Spring	110	4.8	--	<100	<10	<1	<1	<1	<1	<10
Woodcamp Spring	38	14	<10	<2	<0.5	<1	<5	<1	<1	<3
Grapevine Ranch Spring #1	160	15	<10	<100	<10	<1	<1	<1	<1	<10
Grapevine Ranch Spring #3	168	5.7	<10	<100	<10	<1	<1	<1	<1	<10
Staininger Spring	150	5.2	<10	<2	<0.5	<1	<5	<1	<1	<3
Salt Creek near Stovepipe Wells, Calif.	6,800	480	<10	21	<10	<2	<100	<1	<1	<60
Tecopa Hot Spring	850	16	--	---	--	--	--	--	--	--

Table 2. Physical and chemical data for selected springs and Salt Creek—Continued

[Tecopa Hot Spring data are from Thomas and others (1997). °C, degrees Celsius; µS/cm, microsiemens per centimeter at 25 degrees Celsius; mg/L, milligrams per liter, µg/L, micrograms per liter; ‰, per mil; pCi/L, picocuries per liter; pmc, percent modern carbon; --, not analyzed; <, less than; Kjd, kjeldahl]

Site	Pb (µg/L)	Mn (µg/L)	Mo (µg/L)	Ni (µg/L)	Ag (µg/L)	V (µg/L)	Zn (µg/L)	Sr (µg/L)	Li (µg/L)	B (µg/L)
Saratoga Spring	<1	<3	24	<1	<1	30	<9	2,900	410	--
Virgin Spring	<1	5	10	<1	<1	<1	4	1,200	68	--
Warm Spring	<10	<1	10	<10	<1	<6	3	800	11	--
Travertine Spring	<10	<1	20	<10	<1	<6	<3	1,100	160	--
Nevares Spring	<1	4	18	<1	<1	<1	4	1,100	160	--
Keane Wonder Spring	<30	60	<30	<30	5	<18	15	4,640	870	[1]8,530
Klare Spring	<1	<10	9	<1	<1	<1	<10	530	150	--
Woodcamp Spring	<1	<1	1	<1	<1	2.3	<3	20	30	--
Grapevine Ranch Spring #1	<1	<10	12	<1	<1	3.8	<10	570	200	--
Grapevine Ranch Spring #3	<1	<10	12	<1	<1	3.9	<10	660	210	--
Staininger Spring	<1	<1	10	<1	<1	9.4	<3	7	120	--
Salt Creek near Stovepipe Wells, Calif.	<2	<2	280	<1	<1	42	<60	22,000	6,500	--
Tecopa Hot Spring	--	--	--	--	--	--	--	--	--	--

Site	Field HCO_3 (mg/L)	Cl (mg/L)	SO_4 (mg/L)	F (mg/L)	Br (mg/L)	SiO_2 (mg/L)	NH_4 (mg/L)	NO_2 (mg/L)	N (Kjd) (mg/L)	NO_2+NO_3 (mg/L)
Saratoga Spring	427	660	1,000	2.6	1.1	39	--	--	--	1.3
Virgin Spring	488	19	71	0.4	0.13	30	--	--	--	11.0
Warm Spring	125	25	170	0.6	0.12	31	<0.02	0.02	<0.2	0.74
Travertine Spring	343	37	150	3.7	0.16	30	0.02	0.02	<0.2	0.13
Nevares Spring	353	37	170	3.2	0.14	26	--	--	--	<.05
Keane Wonder Spring	[2]--	510	680	6.9	0.83	56	0.8	0.02	0.8	<.05
Klare Spring	349	33	130	3.0	0.11	20	--	--	--	0.30
Woodcamp Spring	122	24	24	0.2	0.10	57	--	--	--	3.00
Grapevine Ranch Spring #1	467	62	120	3.1	0.18	38	--	--	--	0.12
Grapevine Ranch Spring #3	477	64	120	3.1	0.18	38	--	--	--	0.11
Staininger Spring	233	42	89	2.1	0.14	59	--	--	--	0.86
Salt Creek near Stovepipe Wells, Calif.	653	5,700	3,300	2.8	4.4	59	--	--	--	<.05
Tecopa Hot Spring	730	460	500	3.1	--	91	--	--	--	0.06

Table 2. Physical and chemical data for selected springs and Salt Creek—Continued

[Tecopa Hot Spring data are from Thomas and others (1997). ° C, degrees Celsius; µS/cm, microsiemens per centimeter at 25 degrees Celsius; mg/L, milligrams per liter, µg/L, micrograms per liter; ‰, per mil; pCi/L, picocuries per liter; pmc, percent modern carbon; --, not analyzed; <, less than; Kjd, kjeldahl]

Site	P (mg/L)	PO_4 (mg/L as PO_4)	δ^2H (‰)	$\delta^{18}O$ (‰)	$\delta^{87}Sr$ (‰)	$\delta^{13}C$ (‰)	^3H (pCi/L)	^{14}C (pmc)	$^{36}Cl/Cl$ ($\times 10^{-15}$)
Saratoga Spring	--	--	-87.3	-10.65	13.88	-6.1	[3]-0.3	29.46	127
Virgin Spring	--	--	-103	-13.61	18.38	-2.9	0.0	5.44	[4]504
Warm Spring	<0.01	<0.01	-92.8	-12.6	4.68	-4.8	0.2	7.39	251
Travertine Spring	<0.01	<0.01	-102	-13.46	11.48	-3.8	-0.1	3.28	336
Nevares Spring	--	--	-101	-13.51	10.70	-5.5	0.0	3.00	388
Keane Wonder Spring	<0.01	<0.01	-102	-12.95	22.66	-2.8	0.4	--	47
Klare Spring	--	--	-102	-13.62	7.21	--	[5]0.6	--	210
Woodcamp Spring	--	--	-91.6	-12.41	-0.69	-12.2	1.6	77.98	474
Grapevine Ranch Spring #1	--	--	-111	-14.43	3.98	-5.0	--	--	210
Grapevine Ranch Spring #3	--	--	-111	-14.46	3.98	-5.1	[6]0.0	--	216
Staininger Spring	--	--	-111	-14.56	1.44	-6.3	-0.6	20.86	361
Salt Creek near Stovepipe Wells, Calif.	--	--	-97.6	-12.03	11.37	-8.0	2.2	--	51
Tecopa Hot Spring	--	--	-98.0	-12.85	[7]7.64	-4.3	<1.0	--	--

[1]Al and B values derive from nonapproved USGS analytical procedures. The Al value represents a maximum value for total dissolved Al, as the sample was not a field-extracted aliquot, but was a filtered-acidified aliquot.

[2]HCO_3^- concentration not calculated pending dissolved gas analysis.

[3]Sample for ^3H collected 11-29-94 at 1705 hours.

[4] Sample for ^{36}Cl collected 11-30-94 at 0940 hours.

[5] Sample for ^3H collected 11-29-94 at 1241 hours.

[6]Sample for ^3H collected 12-01-94 at 0946 hours.

[7]Sample collected 04-89; analyzed 06-27-89.

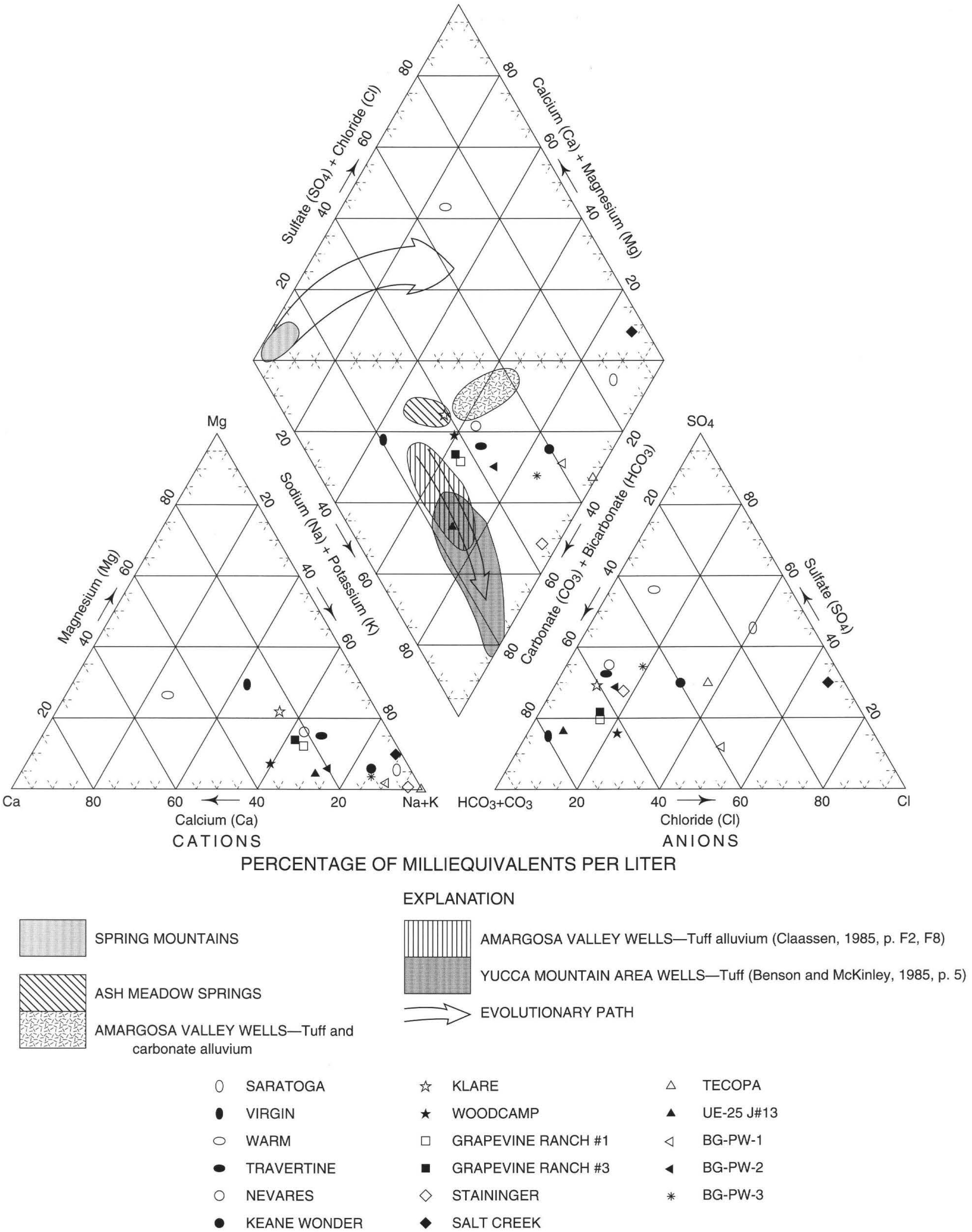


Figure 4. Major water chemistries.

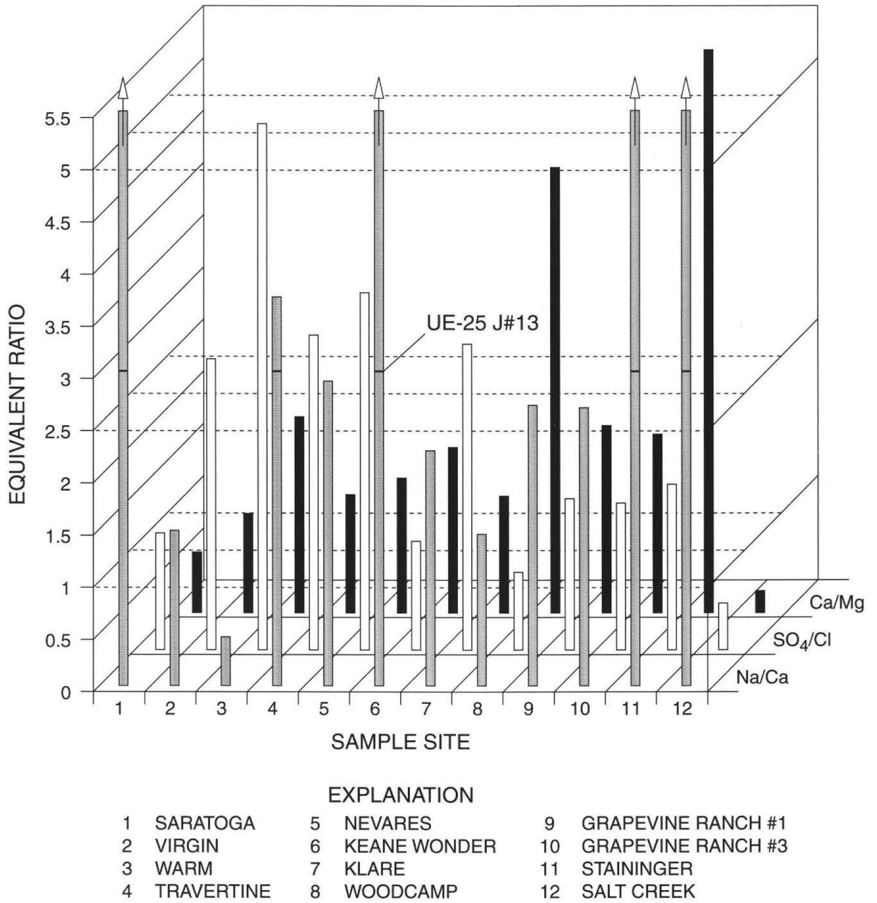


Figure 5. Selected ratios of dissolved constituents (UE–25 J#13 data from Benson and McKinley, 1985).

and reflects the radiogenic strontium content (high δ value) of the Proterozoic rocks (table 3). The stable-carbon isotopic value ($\delta^{13}C = -3.0$ per mil, table 2), is the second highest noted, and strongly indicates the influence of marine carbonates, which generally range between about –1 and 1 per mil. This likely evinces a local control, as the fault gouge along the Amargosa thrust throughout the southern and central Black Mountains comprises both Early Proterozoic metamorphic rocks and Late Proterozoic through Paleozoic formations; these include extensive, thick dolomites.

Saratoga Spring is the only high-discharge spring associated with the Black Mountains. It discharges a brackish sodium sulfate chloride water (fig. 4) typical of ground water influenced by evaporation or mixing with marinelike water or a brine, or both. A marine component is inconceivable, and a brine component is unlikely, as the only brines in the region derive from evaporation from shallow water tables in the Death Valley floor downgradient from Saratoga Spring, and in Franklin Lake Playa at the south end of the Amargosa Desert (fig. 1). Brine is not known to occur along any conceivable flow path to the spring.

Miller (1977) suggested that the Shoshone-Tecopa area is the source of water discharging at Saratoga Spring. This is unlikely, as one of the several normal faults that seem to control the Amargosa River Valley south of Eagle Mountain likely is the cause of the upward flux of ground water in the Tecopa area. The discharge perhaps also is influenced by a shallow (less than 10 km) igneous intrusion centered on the Tecopa area, the presence of which is suggested by a magnetic low and a gravity high (Laura Serpa, University of New Orleans, oral commun., 1996). The Tecopa water is warmer than waters from other springs in the immediate vicinity and warmer than any other discharge observed in this study. While the presence of flowing springs in the Tecopa-Shoshone area seems to preclude the possibility of recharge in this area, it is

Table 3. Rock strontium isotopic data for samples collected in the central Funeral Mountains near Chloride City

[From Z.E. Peterman, U.S. Geological Survey, 1996, written commun. Unit lithologic descriptions and symbols are from Wright and Troxel (1993)]

Stratigraphic unit	Lithology	Sample date	$\delta^{87}Sr$ (per mil)
Stirling Quartzite (Z_{sc})	Dolomite and limestone	03-02-94	5.97
Stirling Quartzite (Z_{sd})	Feldspathic sandstone	03-02-94	189.44
Johnnie Formation (Z_j)[1]	Schist and quartzite, some calcite marble	03-22-94	14.26
Kingston Peak Formation (Y_k)[2]	Conglomerate, quartzite, and limestone clasts in pelitic schist matrix	03-03-94	45.85
		03-21-94	17.71
		03-22-94	47.98
Beck Spring Dolomite (Y_b)	Calcite marble facies	03-02-94	-2.82
		03-02-94	-2.43
Crystal Spring Formation (Y_{cm})	Calcite marble, subordinate pelitic schist	03-03-94	-2.72
		03-03-94	-2.61
Crystal Spring Formation (Y_{cl})	Carbonate bed in pelitic schist	03-03-94	0.28
Crystal Spring Formation (Y_{cu})	Pelitic schist and micaceous quartzite, subordinate calcite marble and amphibolite	03-22-94	69.78
		04-29-94	58.10

[1]The sample was identified as deriving from a clastic unit; therefore, it likely is from either the upper (Z_{ju}) or lower member (Z_{jl}) of the Johnnie Formation.

[2]The sample was identified as deriving from a clastic unit; therefore, it likely is from the upper member (Y_{ku}) of the Kingston Peak Formation.

conceivable that the area is part of the flow path of water recharged in the Spring Mountains, and that some ground water does transit the faults in the vicinity.

To investigate the Spring Mountains as a source of water for Saratoga Spring, chemical data for Tecopa Hot Spring and the Spring Mountains (Thomas and others, 1997, App. A, B) were examined. The water temperature at Tecopa is 42.0°C, about 14°C warmer than at Saratoga. This difference could be attributed to near-surface cooling at the end of the flow path or to dilution of a Tecopa component. To account for the much lower Na/Ca ratio (185 versus 26) and higher relative percentages of SO_4 and Cl at Saratoga (fig. 5), one could invoke dissolution of carbonate rocks that comprise a significant percentage of the Pahrump series rocks through which the water must pass. The absolute and relative concentrations of Ca and SO_4 at Saratoga, however, appear to preclude this.

The extent to which isotopic exchange can influence ground-water hydrogen and oxygen isotopic compositions is a function of subsurface temperature, water/rock ratio, and the composition of the country rock. In general, exchange is most likely to substantially alter ground-water compositions in very hot (greater than 200–250°C) waters that move slowly through low-permeability rocks (which have low water/rock ratios). The temperature of the discharging water at Saratoga Spring is about 28°C (table 2), 11° cooler than the warmest water sampled during this investigation.

Stable-isotope data for Tecopa Hot Spring (Thomas and others, 1997, App. A, B) and the Spring Mountains (Milne and others, 1987, p. 29–30) provide further insight as to the possible source of Saratoga Spring water (fig. 6).

δ^2H and $\delta^{18}O$ data for the Death Valley springs and other selected sites are plotted with Craig's (1961) global meteoric water line (MWL). Departure of data points from the MWL occurs because of isotopic fractionation that can result from evaporation or water/rock isotopic exchange. These processes shift the oxygen isotope value ($\delta^{18}O$) away from the MWL to higher values. The second line is included to identify waters that have been assumed to have undergone either significant evaporation or isotopic exchange, or both, resulting in a position farther to the right on the plot.

Saratoga Spring issues from Pahrump series rocks, a suite of sandstones and dolomites that locally have been metamorphosed by intrusive basalts that are most prevalent in the lower part of the series. Unal-

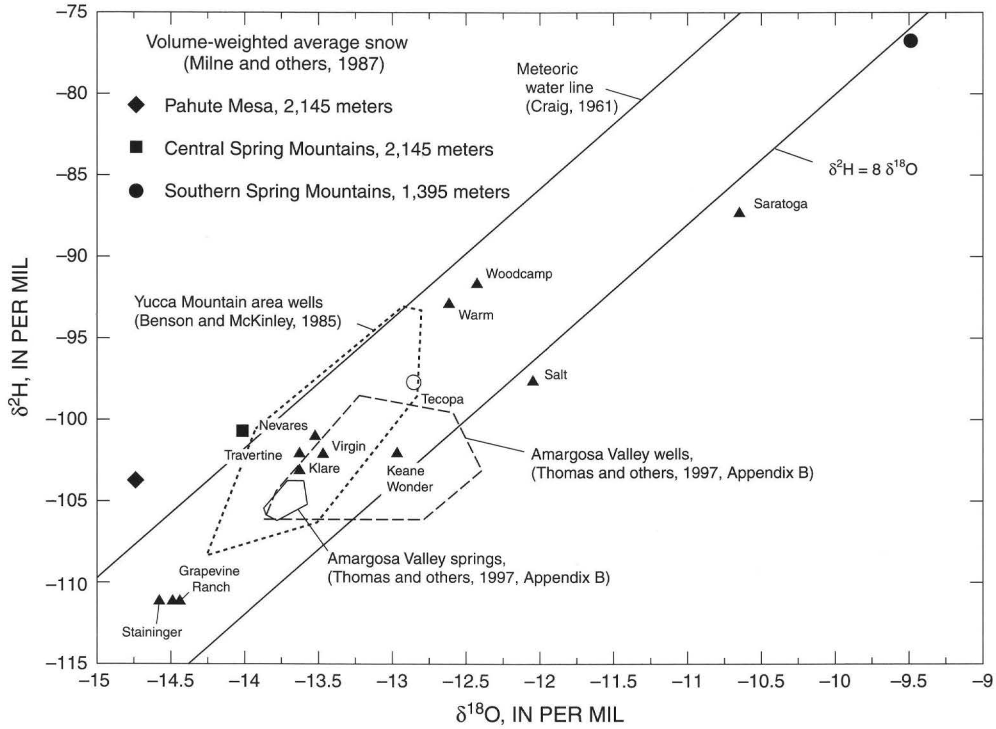


Figure 6. Selected $\delta^{18}O$ and δ^2H values for the Death Valley Region.

tered basaltic rocks typically have $\delta^{18}O$ values of about 6.5 per mil (Taylor, 1979, p. 250). The dolomites can be expected to have values on the order of 3 to 8 per mil (Hoefs, 1987, p. 177). Isotopic exchange within such rocks would tend to raise the $\delta^{18}O$ values of ground waters, which typically are less than zero. Milne and others (1987) reported $\delta^{18}O$ values for snow at sites on and adjacent to the NTS, and in the surrounding region. Volume-weighted averages of their data for sites on Pahute Mesa and in the central and southern Spring Mountains are plotted in figure 6. The southern Spring Mountains snow data are particularly relevant in that they agree well with the averages from Thomas and others (1997) for Amargosa Valley wells and springs ^2H and ^{18}O data—the average $\delta^{18}O$ value of which, –12.26 per mil, is 2.71 per mil lighter than the volume-weighted average snow value for the southern Spring Mountains. This difference is slightly less than the expected ice-water fractionation value of 3.1 per mil (O'Neil, 1968) and indicates that the Saratoga value (–10.65 per mil) reflects an ^{18}O shift of about 1.6 per mil from exchange with the much heavier rocks during flow from the Spring Mountains. The magnitude of the shift also indicates that temperature along the flow path is well below that which could be expected to significantly drive isotope exchange, and that the water/rock ratio is relatively high. While an enrichment of only 1 to 2 per mil has been reported for active geothermal systems such as Wairakei (Taylor, 1979, p. 249), the water/rock ratios of such systems tend to be high, and the magnitude of the shift is due solely to the elevated temperatures of the system (on the order of 250°C [Ellis, 1979, p. 644)]. Similarly, the Tecopa $\delta^{18}O$ value is about 1 per mil heavier than the average of 46 analyses of samples from two springs in the central Spring Mountains (Thomas and others, 1997, App. A). The averaged spring value is about 2.9 per mil lighter than the expected fractionation, indicating that these higher elevation waters undergo significantly more evaporation prior to recharge. The increase of the Tecopa value over the Spring Mountains average also is in accord with isotope exchange along a relatively rapid flow path. The thermally enhanced isotope exchange that must accompany the suspected presence of an igneous intrusion beneath Tecopa is lost in mixing with the other flow components along the flow path that ends there, indicating that the effect of the intrusion is extremely localized. It also is of interest to note the difference between the respective parts of the

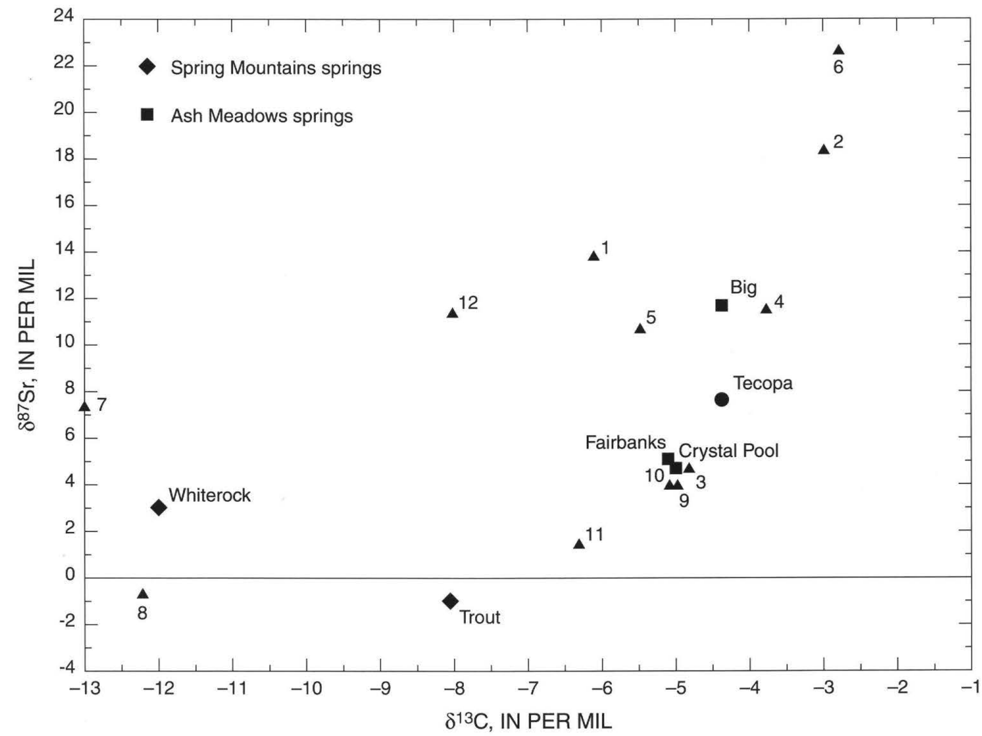
Spring Mountains and the similar difference in values in water from Saratoga and Tecopa Hot Springs. This suggests the possibility of two significantly distinct flow paths from the Spring Mountains toward the southern Death Valley area.

Evaporation during recharge does not seem to be a cause of the increase in $\delta^{18}O$ values from the Spring Mountains to Saratoga Spring, based on the 0.4 per mil enrichment evinced by averages of the snow (Milne and others, 1987) and Amargosa Valley springs (Thomas and others, 1997) data. Evaporation along the flow path also is difficult to reconcile unless a significant part of the path is sufficiently open and shallow to enable water loss. This is perhaps conceivable in the alluvium south of Saratoga Spring. Chemically, however, evaporative concentration does not seem likely. Assuming the Saratoga dissolved chloride value as representative of a nonreactive solute, and waters from the Amargosa Valley and the southern Spring Mountains as starting points, calculated values for other relatively nonreactive elements (Na, K, and F) are high by factors from about 1.2 to about 40. Although such estimates are overly simplistic, they do indicate that evaporation is much less significant than water/rock interaction.

The extent to which rocks contribute to the aqueous isotopic signature is primarily a function of temperature. For ground water in isotopic equilibrium with anorthite (An_{70}), which can be used to approximately represent basalt, at a temperature of 40°C the expected difference between the solid-phase and aqueous $\delta^{18}O$ values is about 21 per mil (O'Neil and Taylor, 1967). The Tecopa and Saratoga values are about 2 to 4 per mil heavier than this and do not support an equilibrium assumption. This implies either that the Saratoga water has moved rapidly through the system or that it has experienced a subsurface temperature on the order of 70°C along the flow path and cooled significantly enroute to the surface. This conceivably could reflect the earlier noted intrusion hypothesized to exist in the Tecopa area or could simply reflect a deep flow path.

Strontium-isotope data also appear to support a Spring Mountains source for Saratoga water. The Proterozoic carbonate rocks along the flow path to Saratoga Spring have $\delta^{87}Sr$ values between about –3 and +6 per mil (table 3). The highest value, that of the carbonate unit within the c member of the Stirling Quartzite, is substantially higher than those of other carbonate rocks sampled, and likely reflects a radiogenic contribution by the surrounding feldspathic and micaceous clastic units. Excluding this value, the resultant range, about –3 to 0 per mil, bounds the range of Cambrian-Ordovician marine carbonates (Faure, 1986). Basaltic rocks have $\delta^{87}Sr$ values less than –1.7 per mil (Carlson, 1995, p. 367). The diabase sills in the Pahrump series can be expected to evince such values. Clastic units in the Pahrump series rocks have values as high as 189 per mil, as they contain older feldspars and micas that have more radiogenic ratios. The strontium data in figure 7 include values for two springs in the Spring Mountains. The Whiterock Spring value exceeds the range for Proterozoic carbonate rocks noted above, while the Trout Spring value is within the range. Whiterock Spring issues from a sandstone in the southeastern part of the mountains, and although the water was recharged in the adjacent carbonate rocks, its higher value must derive from the feldspar component of the rock, which is older and has a more radiogenic Sr component. The Trout Spring value likely is representative of $\delta^{87}Sr$ values of the carbonate rocks that comprise the Spring Mountains. The Saratoga Spring value is much higher than the Spring Mountains and Tecopa values (table 2), indicating a flow path influenced by more radiogenic rocks. This can be expected if flow is through the Pahrump series or older metamorphic rocks.

A chemistry-based consideration of a southern Spring Mountains source for discharge in the southern Death Valley area is in accord with a conceptual geohydrologic model of the ground-water flow system developed by F.A. D'Agnese (U.S. Geological Survey, oral commun., 1996). He suggested, based on structural and potentiometric data, that Spring Mountains recharge takes two general paths. The first is from the central and northern part of the mountains essentially west toward the Ash Meadows area; the second is from the southern part of the mountains to the south and west in the general direction of the south end of Death Valley toward Saratoga Spring. A ^{2}H value calculated using the weighted $\delta^{18}O$ snow value from the Milne and others (1987) data and Craig's (1961) equation is –88.1 per mil, 0.8 per mil lighter than the measured valued from the Saratoga sample. Such a slight increase also is in accord with ground-water flow, as rocks contain little hydrogen compared to the mass of water that moves through them. One should note also that the increase above the calculated snow value is within the ±1 per mil limit of the analysis.



EXPLANATION

1	SARATOGA	5	NEVARES	9	GRAPEVINE RANCH #1
2	VIRGIN	6	KEANE WONDER	10	GRAPEVINE RANCH #3
3	WARM	7	KLARE (no $\delta^{13}C$ data)	11	STAININGER
4	TRAVERTINE	8	WOODCAMP	12	SALT CREEK

Figure 7. Selected $\delta^{13}C$ and $\delta^{87}Sr$ values for the Death Valley region.

Additionally, the ^2H value at Tecopa is about 10 per mil lighter than that of Saratoga and is within 1 per mil of the average reported by Thomas and others (1997, table 13) for 37 samples from 2 springs in the central Spring Mountains (not shown in fig. 6).

The carbon isotope data ($\delta^{13}C$) in figure 7 either do not support the proposed flow path or require some mechanism to explain the heavier value at Tecopa. The dissolved inorganic carbon in Trout Spring, about –8.1 per mil (fig. 7) (Thomas and others, 1997, App. B), is in isotopic equilibrium with a maximum soil CO_2 value of about –18 per mil (Mook and others, 1974), at the light end of the range of isotopic values for CO_2 respired by arid-zone C4 plants (Galimov, 1985). Although the Saratoga value is much heavier, as would be expected from exchange with a marine carbonate, the Tecopa value is heavier still. A flow path from the Spring Mountains to Saratoga Spring through the Tecopa area can support this distribution of $\delta^{13}C$ values only if some process such as calcite precipitation is occurring between Tecopa and Saratoga. Calcite precipitation is not supported by the decrease in the Na/Ca ratio (185 to 26) from Tecopa to Saratoga. Additionally, although the water is oversaturated with respect to calcite, no travertine was observed in the spring box or in the discharge area. These points additionally indicate that contribution of a Tecopa-type water to the Saratoga discharge is, at most, minor.

Funeral Mountains

Travertine Spring discharges a relatively dilute water from unconsolidated Quaternary sediments that likely receive through-flow from lower Paleozoic rocks to the northeast. Nevares Spring appears to issue from a fault or faults associated with lower Paleozoic rocks immediately east of the spring mound. The locations of these waters on the trilinear diagram (fig. 4) indicate that their chemistries are influenced both by igneous and carbonate rocks. They are more calcic than waters that have reacted solely with volcanic rocks or detritus, shown by the areas of the diagram that encompass analyses of waters from Tertiary tuffs in the vicinity of Yucca Mountain (Benson and McKinley, 1985, p. 5), and from wells completed in tuff alluvium in the Amargosa Valley (Claassen, 1985, p. F2, F8). The spring compositions plot in an area bounded by those of the Amargosa Valley wells and the Ash Meadows springs. A simple mixing calculation, using data from Amargosa Valley wells completed in tuff alluvium and mixed tuff-plus-carbonate alluvium, and from one of the Ash Meadows springs (Thomas and others, 1997, App. B) yields a composite water that plots close to the Travertine and Nevares positions. Such calculations yield nonunique and nonstoichiometric solutions and do not include the effects of water/rock interactions. Nonetheless, this result is illustrative, if only because chloride is a relatively nonreactive constituent that will vary significantly in this region only as a result of evapotranspiration, mixing, and water/rock interaction. The calculated chloride value is about 5 mg/L less that the Travertine and Nevares values, in accord with possible augmentation during further flow through the southern Funeral Mountains.

Travertine and Nevares Springs are isotopically similar to waters discharging at the Ash Meadows springs in the Amargosa Desert. Thomas and others (1997, p. 65–70, 80–86) used geohydrologic and chemical evidence to augment and refine earlier arguments that the Ash Meadows discharge is a mixture of recharge waters from the Spring Mountains and the White River flow system to the north and east (Winograd and Friedman, 1972; Winograd and Thordarson, 1975; Winograd and Pearson, 1976; Welch and Thomas, 1984; Kirk and Campana, 1990). Most of the water in the Amargosa Valley alluvium is isotopically heavier than waters discharging in Ash Meadows and does not appear to contribute to spring discharges in the vicinity of the west end of Furnace Creek Wash. If the Ash Meadows discharge does not represent essentially the sum of ground-water flow to the eastern Amargosa Desert from the northern and central Spring Mountains and from the White River flow system, then flow to and through the southern Funeral Mountains of some part of the remainder must be through fractured, likely Paleozoic, carbonates that underlie the Amargosa valley alluvium. This path is the southwest end of the Spotted Range–Mine Mountain structural zone discussed by Carr (1984, p. 56–62). Flow through the southern Funeral Mountains must be controlled by the general geologic structure, that of a gently folded and extensively faulted anticlinal limb (Wright and Troxel, 1993). The numerous normal faults in the southern Funeral Mountains tend to be oriented normal to both the range axis and potentiometric contours associated with the mountains (F.A. D'Agnese, U.S. Geological Survey, 1996, written commun.). Discharge at Travertine and Nevares Springs also must reflect influence of the Furnace Creek and Keane Wonder fault zones and attendant right-lateral displacements, which, combined with the essentially east-to-west hydraulic gradient, tend to direct ground-water flow to the west or northwest. The synclinal structure of the sedimentary fill in Furnace Creek Wash is likely the final physical control of the Travertine and Texas Spring groups, as it directs ground-water flow down the structural axis toward Death Valley.

The water at Keane Wonder Spring is a brackish, sodium sulfate chloride type that, of the springs sampled, appears most similar to Saratoga and Tecopa waters, based on position on the trilinear diagram (fig. 4). Its Na/Ca ratio, like that of Saratoga, is much lower than the Tecopa value but is much higher than the ratios for the other springs, with the exception of Staininger. The high ratio is indicative of advanced chemical evolution. It also is similar to waters from three deep monitor wells in the northwestern part of the Amargosa Desert (fig. 4; table 4, prefix "BGETW") that have similar relative but not absolute compositions. These wells are located upgradient from Keane Wonder, based on a compilation of regional water levels and preliminary results of numerical simulations of the regional flow system (F.A. D'Agnese, U.S. Geological Survey, written commun., 1996). It is conceivable that the spring chemistry could evolve from Amargosa waters such as these if the water/rock contact time is sufficiently long.

The travertine that has been deposited in the Keane Wonder area is, based on field examination, a dense, brittle limestone that likely owes its gray-tan color to the incorporation of a significant amount of iron (Fe). This inference is strengthened by the relatively high Fe concentration in the spring water (table 2).

Table 4. Physical and chemical data for Tecopa Hot Spring and selected boreholes

[Tecopa Hot Spring data are from Thomas and others (1997, app. A, B); Tecopa Hot Spring $\delta^{87}Sr$ value is previously unpublished data from U.S. Geological Survey files. The remaining data are from U.S. Geological Survey files, and are previously unpublished analyses of samples from mining company boreholes (BGETW) in the northwestern Amargosa Desert. °C, degrees Celsius; µS/cm, microsiemens per centimeter at 25 degrees Celsius; mg/L, milligrams per liter; µg/L, micrograms per liter; ‰, per mil; pmc, percent modern carbon; --, not analyzed]

Site	Latitude	Longitude	Date	Time	Temperature (°C)	Field specific conductance (µS/cm)	Field pH units	Ca (mg/L)	Mg (mg/L)	Na (mg/L)	K (mg/L)
Tecopa Hot Spring	355219	1161350	06-30-85	1400	42.0	[1]1,650	8.2	4.0	1.5	850	16
BGETW I-4	355128	1164820	07-09-88	1400	41.0	2,870	7.21	49	9.2	640	19
BGETW II-1	365106	1164932	08-11-88	2245	36.6	614	7.68	27	5.2	110	3.9
BGETW II-2	365105	1164819	11-01-88	1145	34.0	1,246	7.84	21	4.8	190	4.4

Site	Sr (µg/L)	Li (µg/L)	Field HCO_3 (mg/L)	Cl (mg/L)	SO_4 (mg/L)	F (mg/L)	SiO_2 (mg/L)	NO_2+NO_3 (mg/L)	δ^2H (‰)	$\delta^{18}O$ (‰)	$\delta^{87}Sr$ (‰)	$\delta^{13}C$ (‰)	^{14}C (pmc)
Tecopa Hot Spring	--	--	730	460	500	3.1	91	0.06	-98.0	-12.85	[2]7.64	-4.3	--
BGETW I-4	2,100	430	428	250	800	5	22	0.06	-101.5	-13.50	--	-2.78	4.2
BGETW II-1	230	76	224	34	86	1.5	43	1.6	-104.0	-13.80	--	-6.20	5.7
BGETW II-2	220	170	286	64	160	3.8	43	0.52	-100.0	-13.35	--	--	--

[1]Laboratory value.

[2]Sample collected 04-89; analyzed 06-27-89.

The Keane Wonder δ^2H and $\delta^{18}O$ values demonstrate a meteoric origin, and the oxygen value indicates either evaporative enrichment or exchange with the country rock. While the Keane Wonder $\delta^{18}O$ value (–12.95 per mil) could evaporatively derive from a starting value of water in northwest Amargosa alluvium (–13.3 to –13.8 per mil), the proximity of the spring to the relatively intact core of the anticline requires a deep or circuitous flow path which could include evaporation only at the ends. The spring temperature (34.1°C) is 10°C higher than the mean annual air temperature for the Death Valley floor (Hunt and others, 1966, p. B8). This supports a flow path either through the mountains or one involving mixing with water moving upward from depth. The water temperature is in accord with flow through the mountains, assuming an average value for the geothermal gradient (1°C/30 m), in that a depth of travel on the order of 1 km likely would suffice to heat the water enough to obtain the measured temperature. Pelitic schists, such as those present in much of the Middle and Late Proterozoic rocks, generally have $\delta^{18}O$ values between 15 and 18 per mil (Hoefs, 1987, p. 195–196). Epithermal (less than 100°C) water/rock interaction with such rocks would tend to raise the ground-water value. The unexpected slight enrichment (0.35 to 0.85 per mil) of Keane Wonder over the Amargosa values could be attributed to a high water/rock ratio, isotopic depletion of the country rocks as a result of past hydrothermal water/rock interaction along existing flow path(s), a relatively low-temperature environment, or some combination of the three. The relatively low discharge of the spring seems to argue against a high water/rock ratio. It is noteworthy that the ores extracted from the Keane Wonder and other mines in the north-central Funeral Mountains were limonite-stained quartz veins or blankets, often crushed and recemented (Ball, 1907, p. 173–175). These were deposited by a hydrothermal system driven by one or more deep-seated igneous intrusions. That flow system had, of necessity, an internal temperature much higher than those of present-day spring waters, as quartz deposition could have taken place only at a temperature well above 100°C, and likely above 200°C. The country rock in hydrothermal systems typically evinces isotopic depletion because of temperature-enhanced exchange. This possibility could render the internal temperature regime of the current flow system inconsequential.

It is unlikely that the meteoric signature at Keane Wonder is indicative of a recharge origin in the Funeral Mountains. Although the maximum elevation in the Funeral Mountains is about 2 km, relatively little of the range is above 1.5 km. Hunt and Robinson (1966, p. B6–B7) summarized U.S. Weather Bureau data for 12 weather stations in and around the Death Valley area. Annual precipitation ranged from 42 to 130 mm at elevations from –51 to 1,593 m. The records also show that, while precipitation increases with elevation, the average amount above about 1,525 m is about 3 times that below this elevation. They also reported that winter evaporation at an elevation of about 600 m was near that of the average value for the valley floor, about 3,800 mm/yr. Ball (1907, p. 174) reported that the nearest wood suitable for fuel to support mining operations was available about 16 km north of Chloride City in the much higher Grapevine Mountains. All of this evidence indicates that the water issuing at Keane Wonder was not recharged in the adjacent Funeral Mountains. Evidence for local recharge does exist, perhaps, in the presence of Monarch Spring at an elevation of about 900 m, about 5.3 km north of Keane Wonder. Monarch Spring issues at or near the intersection of a nearly horizontal décollement and a high-angle normal fault in Early Proterozoic metasediments that are the oldest exposed rocks in the Funeral Mountains. This spring discharges about 25 to 100 L/min. The site is at a vertically walled reach of Monarch Canyon and is densely vegetated with tall reeds and some low woody shrubs. The presence of dead shrubs near and at the periphery of the vegetated area indicates that discharge has diminished over at least the past 25 years. This decrease, together with the estimated magnitude of the discharge and the apparently relatively cooler water temperature noted during reconnaissance, indicates that Monarch Spring does represent recharge within the Funeral Mountains. This site was not sampled because the flow is confined by the narrow valley and supports such a dense reed growth and attendant soil and detritus cover that a suitable sampling location could not be found or readily established.

The Keane Wonder $\delta^{13}C$ and $\delta^{87}Sr$ values (fig. 7) reflect contact with the Proterozoic anticlinal rocks. The $\delta^{87}Sr$ value is the highest of all sites sampled. This is in accord with flow through the Proterozoic rocks, which have higher $\delta^{87}Sr$ values because of their age and the long half-life of ^{87}Rb (500 billion years), the parent isotope of ^{87}Sr. Samples of rocks from the Pahrump series, the Stirling Quartzite, and the Johnnie Formation collected in the vicinity of Chloride City (fig. 2) yielded $\delta^{87}Sr$ values ranging from –2.82 to 189.44 per mil (table 3). These samples comprised carbonate rocks, feldspathic sandstone, and pelitic schists. The carbonate values were the lowest (–2.82 to 5.97 per mil); the schists ranged from 14.26 to 69.78 per mil; and the sandstone had a value of 189.44 per mil. These data indicate that the Keane Wonder value, 22.66 per mil, derives from the metamorphosed detrital rocks. The $\delta^{13}C$ value, –2.8 per mil, also is the highest sampled, indicating extensive isotopic exchange with carbonate phases, perhaps in the middle member of the Crystal Spring Formation.

A singular aspect of this spring is the emanation of hydrogen sulfide (H_2S). The olfactory evidence is supported by the high apparent titration alkalinity, about 1,000 mg/L. The presence of dissolved bisulfide indicates reducing conditions along the flow path upgradient from the spring and a system closed to the atmosphere and implies either dissolution of sulfide minerals or sulfate reduction. If there is sufficient organic material in the system to support microbial sulfate reduction, then a long ground-water residence time likely can be assumed, as microbially mediated reduction reactions tend to be slow (Drever, 1988, p. 322).

The water chemistry of Keane Wonder Spring indicates limited flux of ground water through the Funeral Mountains. The proximity of the spring area to the relatively intact Proterozoic core of the Funeral Mountains anticlinorium, the presence of the Boundary Canyon fault to the east, the absence of other springs having comparable discharges both in and along the west front of the northern Funeral Mountains, and the east-to-west gradient of the regional potentiometric surface (F.A. D'Agnese, U.S. Geological Survey, 1996, written commun.) also support this conclusion. These aspects further indicate that flow through this part of the mountains, as evinced by the Keane Wonder Spring chemistry, is quite slow, and that most, if not all, of the water discharging in the Keane Wonder area may come from the Amargosa Valley. The northern part of the Funeral Mountains is, therefore, likely a substantial barrier to ground-water flow to Death Valley from the Amargosa Valley region.

Grapevine Mountains

Klare Spring discharges a dilute water that appears, based on its location on the trilinear diagram (fig. 4), to be influenced by both carbonate and volcanic rocks. It plots in the same region as the Ash Meadows springs, distinct from waters that contact only carbonate or siliceous volcanic rocks. The influence of carbonate rocks also is indicated by the SO_4/Cl ratio (fig. 5). Marine limestones and dolomites commonly contain minor amounts of sulfate minerals (Sprinkle, 1989, p. I15; Thomas and others, 1997, table 10) that contribute SO_4 to solution, as evinced by the values for Virgin, Warm, Travertine, and Nevares Springs (fig. 5).

The contribution of volcanic rocks is reflected in the Na/Ca ratio (fig. 5), which is higher than that anticipated from contact with the Carrara Formation limestone alone and lower than those from sites that have long flow paths. The Warm Spring value exemplifies a dominant carbonate influence for the sample population; the Saratoga and Keane Wonder values represent longer paths. Nearly all of the wells completed in mixed volcanic-carbonate alluvium in the Amargosa Valley had ratios between 1 and 3 (Thomas and others, 1997, App. B), bounding the 2.23 value at Klare Spring. The volcanic signature is to be expected, given the presence of at least three major, north-oriented faults within 5 km of the spring, and the presence of Cenozoic sedimentary and volcanic rocks over the northern three-fourths of the estimated drainage area (Streitz and Stinson, 1974; Reynolds, 1976, p. 20, 23).

The dissolved silica (SiO_2) concentration at Klare Spring, 20 mg/L (table 2), is the lowest found in this study but is at the high end of the range of SiO_2 concentrations of waters from dolomitic and limestone aquifers reported by White and others (1963, p. F22–23). The concentration is lower than anticipated for a ground water that has reacted solely with an extrusive volcanic rock, as glassy and cryptocrystalline phases in such rocks are the most soluble phases present (Jones, 1966, p. 193) and react rapidly with water to yield SiO_2 concentrations on the order of 30 to 60 mg/L (White and others, 1963, p. F14). The Klare Spring value also indicates the possibility of flow through sandstones of the Wood Canyon Formation, which compose the core of the recumbent anticline north of and in the vicinity of the spring (Streitz and Stinson, 1974; Reynolds, 1976, p. 23). The $\delta^{87}Sr$

value (7.21 per mil, table 2) tends to support this because the feldspathic sandstones of the Wood Canyon Formation would contribute a more radiogenic Sr component than would derive from the volcanic rocks. It also should be noted that a groundwater sample from a well on the NTS completed in a limestone of the Carrara Formation (Winograd and Thordarson, 1975, p. C28) had a $\delta^{87}Sr$ value of 8.04 per mil (Z.E. Peterman, U.S. Geological Survey, written commun., 1996). This similarity to the Klare Spring value does not impute hydrologic continuity; it instead addresses the possibility of lithologic influence on the isotopic ratio. Hydrogen and oxygen isotopic data for Klare Spring do not appear to support a completely local origin. They are close to those of Travertine, Nevares, and Virgin Springs (table 2) and are within the ranges of values for wells at and in the vicinity of Yucca Mountain to the east. They also are near the light end of the ranges of values for wells in the Amargosa Valley alluvium (fig. 6). The similarity of the Klare Spring data to the Furnace Creek area, Amargosa, and NTS data indicates the possibility of recharge outside of the Grapevine Mountains; as the Klare Spring data differ significantly from the values at Woodcamp Spring, a site believed to discharge water recharged in the forested heights of the Grapevine Mountains (discussed in the next paragraph). This perhaps provides a basis to examine the possibility of the southwestward extension of the northeast-oriented structural zone between the Bullfrog Hills and the Timber Mountain caldera mentioned by Carr (1984, p. 11, 30, 64–66) (fig. 3).

Woodcamp Spring issues at the southwest margin of Sarcobatus Flat (fig. 1). It is one of several springs on or adjacent to the east flank of the Grapevine Mountains. Unlike most of the other sites sampled, it appears to be the result of flow through solely volcanic rock. It is the most dilute water encountered, and several lines of evidence indicate that it has had the shortest residence time since recharge. It generally has the lowest concentrations of every dissolved species except SiO_2 (table 2). This is indicative of reaction with the vitric rocks from which it discharges. Its radiocarbon (^{14}C) content is the highest of all sites sampled (table 2), indicating a maximum age of about 2,500 years. The Na/Ca ratio is low (fig. 5); this, combined with the overall low concentrations of dissolved species, indicates a less evolved water rather than a limestone influence. Its Ca/Mg ratio is the second highest (fig. 5), much higher than would derive from a dolomite, and in accord with the high ratios reported by Broxton and others (1987, p. 95) for samples of Miocene, high-silica, rhyolitic glasses from Yucca Mountain. The $\delta^{87}Sr$ value is the lowest measured (table 2 and fig. 7). While values near zero often reflect a marine carbonate influence, this low value is in the range of values for Miocene tuffs sampled in the vicinity of Yucca Mountain (Peterman and others, 1996, p. 15). The $\delta^{13}C$ value also is the lowest in this study (table 2 and fig. 7). It indicates a dominance of the isotopic signal by soil CO_2 having a value on the order of –20 to –25 per mil, and minimal, if any, pedogenic or eolian influence. The $\delta^{2}H$ and $\delta^{18}O$ values at Woodcamp Spring are similar to the values of samples from Keane Spring (mean $\delta^{2}H$ = –94.75 per mil, mean $\delta^{18}O$ = –12.58 per mil (J.B. Czarnecki, U.S. Geological Survey, written commun., 1992), a group of low-discharge seeps northwest of Monarch Spring, at an elevation of about 1,170 m. The Keane Spring samples derived from a seep that, during sampling, discharged on the order of 1 L/d, indicating that the site reflects limited local recharge. All of the above indicate that the water discharging from Woodcamp Spring was recharged in the forested, higher parts of the Grapevine Mountains west of the spring. No evidence was identified that indicates a relation to a larger scale, regional flow system.

Staininger Spring discharges about 750 L/min from gravelly alluvium in Grapevine Canyon (fig. 2). As suggested by Kreamer and others (1996, p. 102), who used principal-components analysis of selected trace-element concentrations for Staininger and other springs in the Death Valley area, this water appears to have reacted almost exclusively with volcanic rock. It has a very high Na/Ca ratio (fig. 5), exceeded only by water from Saratoga Spring and Salt Creek. The chemistry of Saratoga Spring appears to reflect a long, and perhaps deep, flow path through Paleozoic and Proterozoic carbonates, clastic sediments, and metasediments, while Salt Creek is strongly influenced by evaporation, as evinced by its high solute load and low Ca/Mg and SO_4/Cl ratios (fig. 5). A carbonate influence is perhaps evinced by the location of Staininger Spring on the trilinear diagram (fig. 4), in an area of slightly higher sulfate-plus-chloride content than the Woodcamp sample.

The only local source of recharge for Staininger Spring that could perhaps support the flow rate for the spring group is in the high, forested parts of the

Grapevine Mountains to the southeast. Several lines of evidence argue against this scenario. The Staininger Na/Ca value is about an order of magnitude higher than anticipated from interaction between water and silicic volcanic rock, as typified by the UE–25 J#13 value (fig. 5). This indicates a more advanced stage of chemical evolution, one that can be expected of a position farther down a flow path than that evinced by the silicic volcanic example. The Cl and SO_4 concentrations are considerably higher than those from Woodcamp Spring. Excluding contributions from thermal waters, dissolution of evaporites, and the influence of other rock types, this can derive only from a much longer residence time in volcanic rocks. The increased SO_4 value could reflect a carbonate influence, but this is not supported by the low Ca value. Ion exchange cannot defendably be invoked to drive the Ca concentration to the low level found, as the high Na/Ca ratio would tend to exchange Na for nearly any major divalent metal on exchange sites. Although the distances of Woodcamp and Staininger Springs from the recharge area in the Grapevine Mountains are about the same, the intervening north- and northeast-oriented faults, together with the eastward dip of the Cenozoic rocks that cover much of the eastern one-half of the northern Grapevine Mountains, are not conducive to flow to the northwest. The attitude of the folds in the pre-Cenozoic rocks, however, that of plunging to the northwest, is perhaps amenable with flow in that direction.

Isotopic data also point to a source for Staininger Spring water outside the Grapevine Mountains. Staininger δ^2H and $\delta^{18}O$ values are among the lightest measured and are much lighter than the Woodcamp values (fig. 6). The Staininger Spring values (δ^2H = –111 per mil and $\delta^{18}O$ = –14.56 per mil) are indicative of recharge in a colder and perhaps higher environment and are similar to those of ground water from volcanic rocks in Pahute Mesa in the northwestern part of the NTS (δ^2H = –113 per mil, $\delta^{18}O$ = –14.88 per mil). Compared to the source water shown in figure 6, the Staininger Spring values are most similar to, but somewhat lighter than, volume-weighted isotopic values for recent Pahute Mesa snow samples reported by Milne and others (1987, p. 29–30). The Staininger Spring $\delta^{87}Sr$ value is similar to the Woodcamp Spring value and to the above-noted Pahute Mesa groundwater value, again indicating interaction with silicic volcanic rock, although a carbonate influence is not inconceivable. The stable-carbon isotope value indicates a carbonate rock contribution, in that the Staininger value is about 6 per mil heavier than the Woodcamp Spring value and is from 6 to 16 per mil heavier than can be expected from a soil-gas-controlled value. This is perhaps in accord with the vertical proximity of the eroded surface of the Pogonip Group, as discussed in the site description. The spring water is undersaturated with respect to calcite, and dissolution of calcite is expected if the water is in contact with limestone. However, in order to arrive at the $\delta^{13}C$ value of –6.3 per mil (table 2) at Staininger Spring by dissolution of limestone, assuming initial control by a soil CO_2 value between –15 and –28 per mil, 25 to 80 percent of the dissolved inorganic carbon must derive from calcite (fig. 8A). Such a contribution would input an equimolar mass of Ca to solution that is not observed. The $\delta^{13}C$ signature at Staininger Spring, therefore, is perhaps an inherited value, established earlier along the flow path prior to entry into the volcanic section from which the spring issues.

The water discharging in the Grapevine Ranch area is similar in most respects to that supplying the Scotty's Castle (Staininger Spring) area and likely owes most of its chemical identity to the same factors discussed above. It has been influenced by volcanic and carbonate rocks, based on its position on the trilinear diagram, and plots closer than any of the other sites to the analyses of water from Tertiary tuffs in the vicinity of Yucca Mountain (fig. 4). The major differences between the Grapevine Ranch and Staininger Spring water are the concentrations of the divalent metal ions (Ca, Mg, Sr) and the stable-carbon isotope values. Divalent ion concentrations in Grapevine Spring are about one to two orders of magnitude higher, and the $\delta^{13}C$ values are 1.2 and 1.3 per mil heavier (table 2).

These differences likely derive from the differences between the terminal segments of the flow paths for the two spring groups. The highest Grapevine Ranch Springs issue at an elevation of about 850 m, 75 to 125 m below the elevation of the Staininger Spring group, from a travertine-mantled terrace of rocks identified by Strand (1967) as undivided Paleozoic marine rocks, by Ball (1907, pl. 1) as the Pogonip [Formation] Group, and by Oakes (1977, pl. 1) as the Nopah Formation and the Mazourka Group. Based on surficial exposure only, at least the last 2 to 3 km of the flow path are through these carbonate rocks, and it is likely that these units, together with the underlying Cambrian limestones and dolomites, also are part of the flow path east of the outcrop limits (Oakes, 1977,

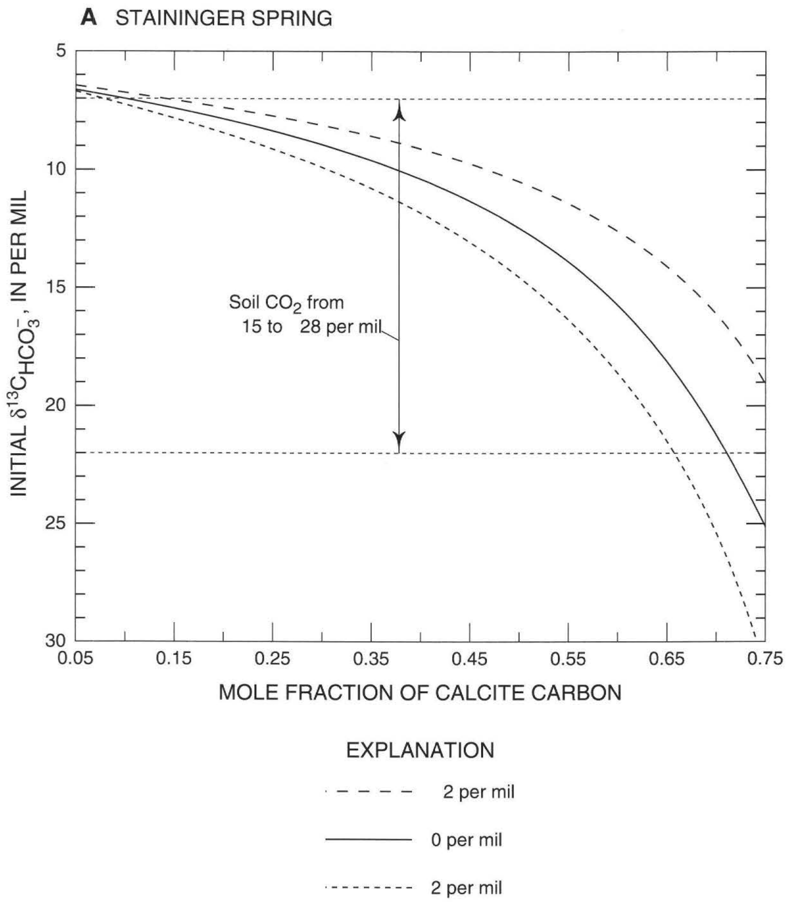


Figure 8A. Initial aqueous $\delta^{13}C_{HCO_3^-}$ value required to yield the measured Staininger values.

pl. 1). The Grapevine Spring samples are in equilibrium or slightly oversaturated with respect to calcite, evincing the contribution of limestone dissolution. Depending on the $\delta^{13}C$ value of the Paleozoic carbonates, which can be assumed to be between –2 and 2 per mil, and using the Staininger Spring $\delta^{13}C$ value of –6.3 per mil as a starting point, an increase of 1 to 2.1 millimoles per liter (mmol/L) of Ca (40–85 mg/L) would accompany a 1.2 per mil increase in the $\delta^{13}C$ value attendant to calcite dissolution (fig. 8b). This increase bounds the Ca concentrations noted.

The flow-path difference is supported also by the Sr isotopic data. Oakes (1977, p. 47–51) described a Tertiary igneous intrusion north of the Grapevine Ranch area that gently domed the country rocks. He also described extensive, mostly vertical, calcite veins in the Nopah Formation that closely parallel the trend of the intrusion. Similar veins in the south and west-central parts of the Grapevine Ranch area were observed during reconnaissance visits for this study. The orientations and widespread occurrence of these veins indicate substantial fluid flow and transport driven by the heat of the intrusion. It is likely that this activity altered the Sr isotopic signature of the lower Paleozoic rocks, which can be expected to have had an original value between –1.7 and 0 per mil (Faure, 1986, p. 188). Based on analyses of Paleozoic limestones at Bare Mountain, Nevada, that were altered by igneous intrusions, Peterman and others (1994, p. 1321) reported that $\delta^{87}Sr$ values greater than 3 per mil evinced the introduction of more highly radiogenic Sr into the host rocks. The $\delta^{87}Sr$ value (3.98 per mil) at Grapevine Ranch Spring #3 (table 2) is about 2.5 per mil higher than the Staininger Spring value and likely reflects the alteration of the Sr isotopic signature of the Cambrian-Ordovician limestones by the Tertiary intrusion.

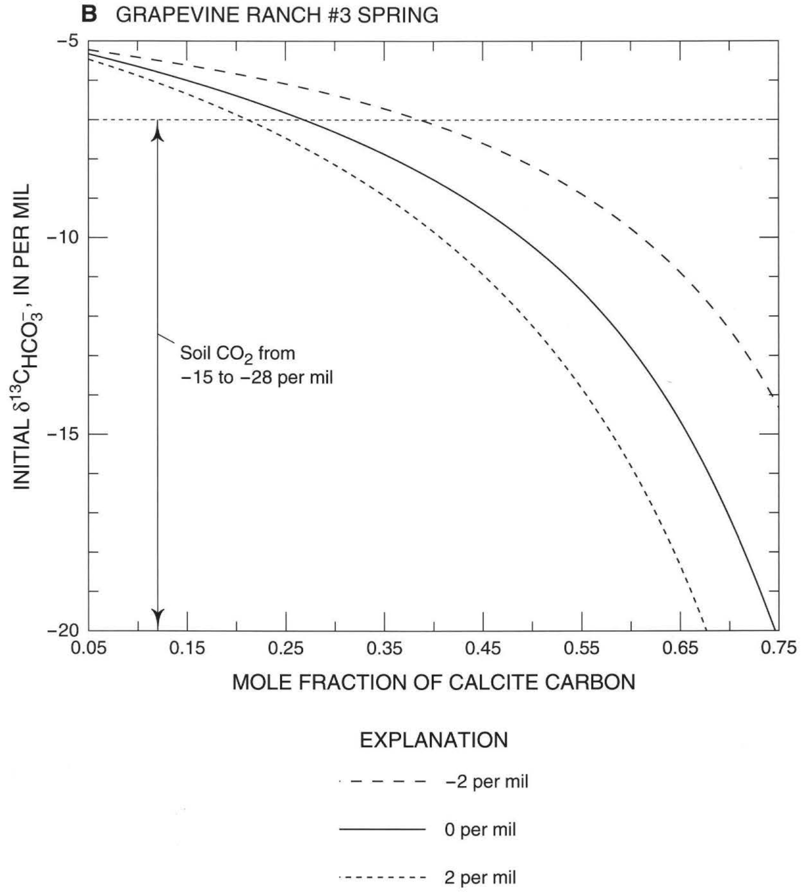


Figure 8B. Initial aqueous $\delta^{13}C_{HCO_3^-}$ value required to yield the measured Grapevine #3 values.

Table 5. Summary of recharge areas for springs and Salt Creek

[L/min, liters per minute}

Site	Site number (see fig. 2)	Location	Discharge (L/min)	Primary recharge area
Saratoga Spring	1	Southernmost Black Mountains	100/1,000	Southern Spring Mountains
Virgin Spring	2	South-central Black Mountains	0–0.1	Black Mountains
Warm Spring	3	Southern Panamint Range	60	Southern Panamint Range
Travertine Spring	4	Southwestern Funeral Mountains	20–40/4,500	Central and northern Spring Mountains, White River flow system
Nevares Spring	5	Southwestern Funeral Mountains	650/980	Central and northern Spring Mountains, White River flow system
Salt Creek	12	Central Death Valley	60/60–10,280	East- west-, and north-bounding ranges
Keane Wonder Spring	6	Northern Funeral Mountains	130	East/northeast of the Amargosa Range
Klare Spring	7	Southern Grapevine Mountains	100	Grapevine Mountains and ?east/northeast of the Amargosa Range?
Woodcamp Spring	8	Southwestern Sarcobatus Flat	20–40	Grapevine Mountains
Staininger Spring	11	Northern Grapevine Mountains	400–500/700	East/northeast of the Amargosa Range
Grapevine Ranch Springs	9, 10	Northern Grapevine Mountains	150–300/1,680	East/northeast of the Amargosa Range

SUMMARY AND CONCLUSIONS

The Amargosa Range defines the east margin of Death Valley, and the Panamint and Last Chance Ranges bound it on the west. A large amount of ground water discharges in and into the valley, most noticeably from several large springs at four locales along the west flank of the Amargosa Range. These springs, together with the enormous volume of water discharged from the valley floor by evaporation, cannot be supported by recharge in the adjacent mountains. The distribution of these four sites, together with the paucity of other significant sources in the Amargosa Range proper, indicate that ground-water movement through the range from the east is, in part, unobservable, and that which can be observed comprises the high discharges in the four fairly discrete areas that are coincident with large-scale structural features orthogonal to the Death Valley–Furnace Creek fault zone. Ground-water chemical and isotopic data from high-discharge springs along the west flank of the Amargosa Range affirm this, in that they either appear to derive from identifiable recharge sites east of Death Valley or have characteristics that indicate simply an extra-range origin (table 5).

The Black Mountains are the southernmost part of the Amargosa Range and contain no high-discharge springs. Nearly all of the seeps and springs discharge less than about 2 L/min. Water that reaches the eastern margin of the Black Mountains must come from the Amargosa Valley to the north and east, and from other areas to the east. Any flow through the Black Mountains must traverse the Furnace Creek fault zone, the series of faults that control the southern Amargosa River Valley, and the complex mass of faulted and intruded Proterozoic through Cenozoic rocks, and likely unexposed northwest-oriented plutonic rocks, that compose the Black Mountains block. The complex tectonic history of the Black Mountains, including jostling between the Death Valley and Furnace Creek fault zones and extensive igneous intrusive activity throughout geologic time, has resulted in a complex and heterogeneous assemblage of rock types and structural features that appears to be a barrier to ground-water flow from the east. Given even the significant hydraulic potential that derives from the elevation difference between the south end of the Amargosa Valley and the lowest part of the Death Valley floor, the regional potentiometric surface in the Black Mountains appears to be well below land surface, as springs are minor to nonexistent and there is no record of substantial historical well development.

Saratoga Spring is the only source related to the Black Mountains that can be inferred to derive from outside the Death Valley Basin. Isotopic, chemical, and physical data support a flow path that starts in the southern Spring Mountains in the east and wends west around the south end of the Black Mountains. This path does not appear to be influenced in the same fashion as the Shoshone-Tecopa area by the faults that likely control ground-water discharge there. The springs that issue in the Black Mountains discharge at very low rates, as exemplified by Virgin Spring, the chemistry of which reflects the immediate terrane within which it issues. Differences between this spring and others can be attributed to the influence of the rock types associated with specific sites.

Springs at only two general locales on the west margin of the Funeral Mountains discharge more than 100 L/min. These are the Travertine and Texas Spring groups and Nevares Spring near the west end of Furnace Creek Wash, and Keane Wonder Spring at the west margin of the north end of the mountains. The Nevares and Travertine sites are at the distal end of the Spotted Range–Mine Mountain structural zone, one of several northeast-oriented zones that cross the Walker Lane Belt and appear to provide a path for ground-water flow to the southwest. The Ash Meadows spring group, east of the Furnace Creek area springs, also is associated with the Spotted Range–Mine Mountain structural zone. These springs are similar to Travertine and Nevares Springs both chemically and isotopically, and likely are representative, together with water in the alluvium of the Amargosa Valley, of water moving into the southern Funeral Mountains. Water/rock interaction, and perhaps mixing, yields the characteristics found at Travertine and Nevares Springs.

Keane Wonder Spring in the northern Funeral Mountains perhaps also represents flow through the Funeral Mountains, but on a much smaller scale than in the southern part of the mountains. The water chemistry reflects a longer transit time, a more circuitous flow path, and perhaps mixing with a deeper, upward-flowing water. The spring is slightly southeast of the

Early Proterozoic core of the anticlinorium that composes the Funeral Mountains. The core and adjacent Middle Proterozoic formations are much less faulted than the southeast limb of the fold, through the end of which water moves to the Furnace Creek area. The region east of Keane Wonder Spring appears to be much less likely to enable significant through flow, as indicated by the presence of this single significant discharge site.

The Grapevine Mountains include the highest elevations in the Amargosa Range and, as such, receive the only areally significant recharge. The numerous seeps and low-discharge springs along the flanks and in the mountains reflect structural controls on the flow of this recharge, and the chemistries of these sources, as typified by Woodcamp and perhaps Klare Springs, reflect local recharge and interaction with the rocks that can be inferred to be along the flow paths. Only the high-discharge springs aligned with northeast-oriented regional structural features evince chemical characteristics that indicate an origin from the east. Staininger and the Grapevine Ranch springs are in this group. Staininger Spring exhibits a strong influence by the volcanic rocks within which Grapevine Canyon is incised but has a carbon isotopic signal that cannot be explained by interaction with rocks within the Grapevine Mountains through which the flow path can be assumed. The Grapevine Ranch springs evince the Staininger chemistry but clearly have been modified by the carbonate rocks from which they issue and through which the water must for some distance flow.

Radiocarbon activities for the springs ranged from 3 at Nevares Spring to 78 percent modern carbon (pmc) at Woodcamp Spring (table 2). The values for springs believed to represent waters originating east of the Amargosa Range tend to be the lowest, indicating ages of at least 10,000 years. Most of the values yield uncorrected ages greater than 20,000 years. These uncorrected ages cannot be construed to enable calculation of absolute or, perhaps, even approximate ground-water traveltimes, but they are indicative of relative minimum residence times and provide some insight to the scale of traveltimes along flow paths that end in Death Valley.

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