GEOCHEMISTRY AND ISOTOPE HYDROLOGY OF REPRESENTATIVE AQUIFERS IN THE GREAT BASIN REGION OF NEVADA, UTAH, AND ADJACENT STATES

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1409-C

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Geochemistry and Isotope Hydrology of Representative Aquifers in the Great Basin Region of Nevada, Utah, and Adjacent States

By JAMES M. THOMAS, ALAN H. WELCH, *and* MICHAEL D. DETTINGER

REGIONAL AQUIFER-SYSTEM ANALYSIS—GREAT BASIN, NEVADA–UTAH

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

The Regional Aquifer-System Analysis (RASA) Program was started in 1978 following a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States. The RASA Program represents a systematic effort to study a number of the Nation's most important aquifer systems, which in aggregate underlie much of the country and which represent an important component of the Nation's total water supply. In general, the boundaries of these studies are identified by the hydrologic extent of each system and accordingly transcend the political subdivisions to which investigations have often arbitrarily been limited in the past. The broad objective for each study is to assemble geologic, hydrologic, and geochemical information, to analyze and develop an understanding of the system, and to develop predictive capabilities that will contribute to the effective management of the system. The use of computer simulation is an important element of the RASA studies, both to develop an understanding of the natural, undisturbed hydrologic system and the changes brought about in it by human activities, and to provide a means of predicting the regional effects of future pumping or other stresses.

The final interpretive results of the RASA Program are presented in a series of U.S. Geological Survey Professional Papers that describe the geology, hydrology, and geochemistry of each regional aquifer system. Each study within the RASA Program is assigned a single Professional Paper number, and where the volume of interpretive material warrants, separate topical chapters that consider the principal elements of the investigation may be published. The series of RASA interpretive reports begins with Professional Paper 1400 and thereafter will continue in numerical sequence as the interpretive products of subsequent studies become available.

Gordon P. Eaton

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Director

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FOREWORD

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Dallas L. Peck
Director

CONTENTS

ILLUSTRATIONS

TABLES

CONVERSION FACTORS AND VERTICAL DATUM

Inch-pound units of measure used in this report may be converted to International System of units (SI) by using the following factors

Multiply	By	To obtain
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot per mile (ft/mi)	0.1894	meter per kilometer
foot per year (ft/yr)	0.3048	meter per year
gallon (gal)	3.785	liter
inch (in.)	25.40	millimeter
square mile (mi^2)	2.590	square kilometer

Temperature: Degrees Celsius (^{o}C) can be converted to degrees Fahrenheit (^{o}F) by using the formula ^{o}F = [1.8(^{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.

REGIONAL AQUIFER–SYSTEM ANALYSIS—GREAT BASIN, NEVADA–UTAH

GEOCHEMISTRY AND ISOTOPE HYDROLOGY OF REPRESENTATIVE AQUIFERS IN THE GREAT BASIN REGION OF NEVADA, UTAH, AND ADJACENT STATES

By James M. Thomas, Alan H. Welch, and Michael D. Dettinger

ABSTRACT

The Great Basin region of Nevada, Utah, and adjacent States, contains approximately 260 basins, which form 39 ground-water flow systems. These flow systems are primarily in unconsolidated basin-fill deposits and in carbonate rock that surround the basin-fill deposits in the eastern Great Basin. This report briefly describes the general quality and chemical character of the ground water, discusses in detail the geochemical and hydrologic processes that produce the chemical and isotopic compositions of water in the two principal flow systems (basin fill and carbonate rock), delineates flow systems in carbonate-rock aquifers of southern Nevada, and discusses ground-water ages and resulting flow velocities within those carbonate-rock aquifers.

Water in aquifers of the Great Basin generally contains less than 1,000 milligrams per liter of dissolved solids, except in natural-discharge and geothermal areas. Aquifers in industrial, mining, urban, and agricultural areas and aquifers containing highly soluble evaporative salts and minerals may contain water having either dissolved-solids concentrations greater than 1,000 milligrams per liter or elevated concentrations of constituents that are considered undesirable for certain uses of the water or both. Generally, the chemical character of ground water in the Great Basin is dominated by sodium, calcium, and bicarbonate in basin-fill aquifers in the predominantly volcanic terrain of the western part of the basin; calcium, sodium, magnesium, and bicarbonate in basin-fill aquifers in the eastern part of the basin; and calcium, magnesium, and bicarbonate in carbonate-rock aquifers in the eastern part of the basin. The chemical character of ground water concentrated by evapotranspiration in discharge areas is generally dominated by sodium, chloride, and sulfate.

In Smith Creek Valley in west-central Nevada, the chemical and isotopic composition of ground water in a hydrologically closed basin-fill aquifer evolves as the water moves from recharge areas to the discharge area. Evapotranspiration concentrates the dissolved solids of precipitation in the recharge areas. This concentrated precipitation dissolves carbon dioxide gas and volcanic groundmass and phenocrysts (dominantly albite and anorthite), chalcedony precipitates from the water, and kaolinite forms by incongruent dissolution, producing a sodium calcium bicarbonate water. In addition, small amounts of gypsum, potassium feldspar, and biotite dissolve. In the terminal playa area, where the basin-fill deposits grade into finer grained sediments, the exchange of calcium and magnesium in the water for sodium on clay minerals causes the sodium calcium bicarbonate water to evolve into a sodium bicarbonate water. Calcium also may be removed from the water by the weathering of plagioclase to calcium sodium montmorillonite and the precipitation of a zeolite mineral. In this part of the aquifer, the dissolution of carbon dioxide gas, albite, anorthite, and potassium feldspar, the precipitation of chalcedony, and the formation of kaolinite continue. In addition, sulfate is reduced to hydrogen sulfide gas. Where ground water discharges by transpiration and evaporation, chloride-containing evaporative salts dissolve and calcite and zeolite minerals precipitate, causing the sodium bicarbonate water to evolve into a sodium chloride water. Evapotranspiration of the ground water also results in heavier deuterium and oxygen-18 compositions.

In the carbonate-rock aquifers of southern Nevada, water in recharge areas dissolves calcite, dolomite, and carbon dioxide gas and rapidly reaches saturation with respect to calcite and dolomite. This water contains predominantly calcium, magnesium, and bicarbonate. Heating of this calcite-saturated water during deep circulation results in the precipitation of calcite. In most of the carbonate-rock aquifers, the following reactions take place: Gypsum (or anhydrite) dissolves, causing dolomite to dissolve, which in turn causes calcite to precipitate (dedolomitization); chalcedony precipitates; calcium and magnesium in the water exchange for sodium in clays; kaolinite forms; and, in some spring areas, carbon dioxide gas exsolves. In parts of the aquifers, water with high concentrations of sulfate and sodium leak into the aquifer from an overlying low-permeability unit. In addition, halite dissolves, sodium and potassium probably are added to the water by the dissolution of volcanic glass, volcanic-rock minerals (dominantly albite and potassium feldspar), and zeolite minerals (probably clinoptilolite); all of which are present in parts of the study area. Thus, outside the recharge areas of the carbonate-rock aquifers in southern Nevada, sodium, sulfate, and chloride can be major constituents dissolved in the water. Waters within the carbonate-rock aquifers that originate in different areas and contain different chemical and isotopic compositions can mix, producing a water that is chemically and isotopically different from the source waters.

Regional ground-water flow systems in the carbonate-rock aquifers of southern Nevada were delineated using deuterium, water chemistry, and adjusted carbon-14 ages. The results are as follows: (1) Ground water discharging at the terminus of the White River flow

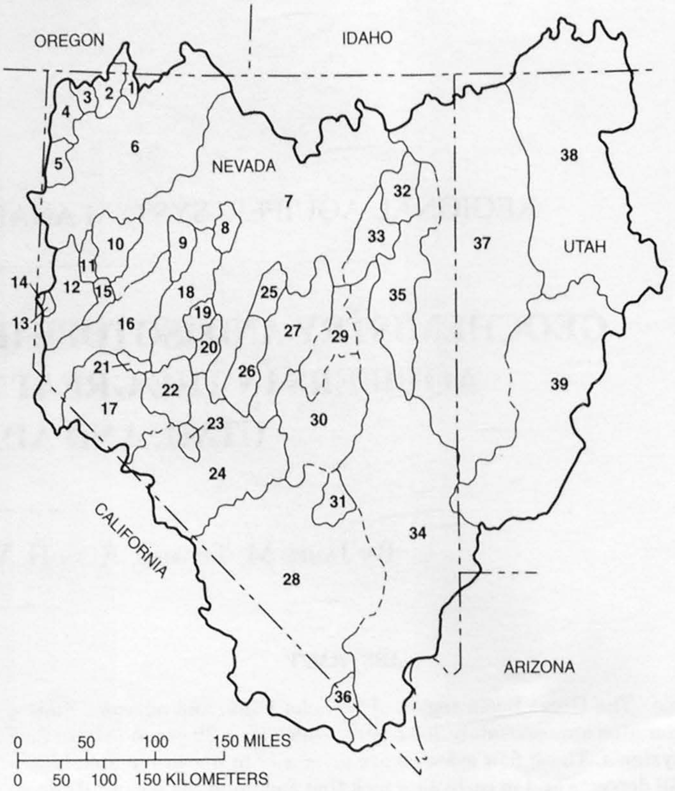
system (Muddy River springs) is a mixture of 40 percent Pahranagat Valley water, 38 percent Sheep Range water, and 22 percent southern Meadow Valley Wash water. (2) Ground water discharging at the terminus of the Ash Meadows flow system (Ash Meadows springs) is a mixture of 60 percent Spring Mountains water and 40 percent Pahranagat Valley water. (3) Las Vegas Valley receives all, or almost all, its ground water from the Spring Mountains (the Sheep Range may supply a small amount to northern Las Vegas Valley). (4) Pahrump Valley receives all its ground water from the Spring Mountains.

Ground-water flow velocities in the carbonate-rock aquifers calculated from adjusted carbon-14 ages are slower than those calculated from hydrologic data. Velocities calculated from adjusted ages range from 9.6 to 144 feet per year, whereas velocities calculated from hydrologic data range from 50 to 740 feet per year. The discrepancy in velocities indicates that ages and average hydraulic conductivities used for the calculations may be overestimated, and effective porosity and flow path length may be underestimated.

INTRODUCTION

The Great Basin, as delineated in this study, includes approximately 260 individual hydrographic areas. These areas can be grouped into 39 flow systems (Harrill and others, 1988), in which ground water flows toward a common hydrographically low discharge area (fig. 1). Some of the larger systems contain subsystems that discharge water at intermediate (higher) positions. The principal aquifers within these flow systems are basin-fill deposits and carbonate rock (Harrill and others, 1988). Volcanic rock form aquifers of local importance, but in Nevada less than 1 percent of total ground-water withdrawal is from volcanic rock (Frick and Carman, 1990, p. 354–356). Highly permeable basin-fill aquifers can be surrounded by generally low-permeability volcanic rock, or by high-permeability carbonate rock. Where basin-fill aquifers are surrounded by low-permeability rock, ground-water flow is mostly contained within the basin-fill sediments. Basin-fill aquifers are present primarily in the western Great Basin, where volcanic rock predominate, and they consist of single-valley, or multivalley, flow systems. In multivalley systems, the basin-fill aquifers are linked by ground-water flow through basin-fill deposits. Basin-fill aquifers also may be hydrologically linked by rivers because of the interaction of surface water and ground water near the rivers. The basin-fill aquifers surrounded by high-permeability rock (mostly carbonate rock) are generally in good hydrologic connection with the underlying and adjacent rock, resulting in deep (several thousand feet) and extensive (hundreds of square miles) ground-water flow within the basin-fill deposits and the surrounding rock.

Highly permeable carbonate-rock aquifers form regional systems in which ground water flows in basin-fill deposits and in carbonate rock that transmit the flow beneath topographic boundaries. These flow



EXPLANATION

——— **Study-area boundary**

– – – – **Flow-system boundary**—Dashed where uncertain

Flow systems

1	Continental Lake
2	Virgin Valley
3	Swan Lake Valley
4	Long Valley
5	Duck Lake Valley
6	Black Rock Desert
7	Humboldt
8	Buffalo Valley
9	Buena Vista Valley
10	Granite Springs
11	Winnemucca Lake
12	Truckee
13	Lemmon Valley*
14	Cold Spring Valley
15	Fernley Sink
16	Carson
17	Walker
18	Dixie Valley
19	Edwards Creek
20	Smith Creek Valley
21	Rawhide Flats
22	Gabbs Valley
23	Monte Cristo Valley
24	South Central Marshes
25	Grass Valley
26	Northern Big Smoky Valley
27	Diamond Valley
28	Death Valley
29	Newark Valley
30	Railroad Valley
31	Penoyer Valley
32	Independence Valley
33	Ruby Valley
34	Colorado
35	Goshute Valley
36	Mesquite Valley
37	Great Salt Lake Desert
38	Great Salt Lake
39	Sevier Lake

* Part of multibasin system. Not known whether subsurface drainage is to the northwest (out of study area) or to the Truckee system.

FIGURE 1.—The 39 major flow systems of the Great Basin, as delineated by Harrill and others (1983, fig. 3).

systems are in the eastern part of the study area (fig. 1), where sequences of carbonate rock generally are more than 20,000 ft thick (Plume and Carlton, 1988).

In 1978, the U.S. Geological Survey began a series of Regional Aquifer-System Analysis (RASA) studies to aid effective management of the Nation's ground-water resources by providing information on the hydrology and geochemistry of the Nation's major aquifers (Bennett, 1979). The Great Basin RASA study is the 10th in this series. The objectives of the geochemistry part of the RASA studies are to describe the quality of water in aquifers on a regional scale and to determine the geochemical processes that produce the observed water chemistry. As work on the Great Basin RASA study was nearing completion, the State of Nevada began a study of the carbonate-rock aquifers of eastern and southern Nevada involving the U.S. Geological Survey, Bureau of Reclamation, and the Desert Research Institute. The overall objective of this study was to explore the potential for developing the carbonate-rock aquifers as water supplies (U.S. Department of the Interior, 1985). In the carbonate-rock aquifers study, geochemical information was used extensively to delineate ground-water flow systems of southern Nevada. This report incorporates the geochemical findings of both studies.

PURPOSE AND SCOPE

The purposes of the study upon which this report is based were to (1) briefly describe the general chemical character of ground water in the Great Basin; (2) identify and illustrate by examples the processes that produce the chemical and isotopic compositions of water in representative aquifers of the Great Basin; (3) delineate ground-water flow paths and mixing of water in the carbonate-rock aquifers of southern Nevada using chemical and isotopic data; and (4) determine ground-water ages and use these ages to calculate flow velocities of ground water in the carbonate-rock aquifers in southern Nevada, and compare these velocities with velocities calculated using hydrologic data.

The Great Basin RASA study encompassed most of Nevada, the west half of Utah, and small parts of Arizona, California, Idaho, and Oregon. This 140,000-mi^2 area contains 39 ground-water flow systems (fig. 1; Harrill and others, 1988, table 1). In this report, general ground-water quality is briefly described for the entire area and geochemical processes affecting chemical and isotopic compositions of ground water are described in detail for two principal types of flow systems: a hydrologically closed basin-fill aquifer (Smith Creek Valley) and a regional system in which ground water flows through several valleys in basin-fill and carbonate-rock aquifers (southern Nevada).

SAMPLING METHODS AND GEOCHEMICAL DATA

Alkalinity, pH, temperature, specific conductance, and dissolved oxygen were measured at each sampling site. Water samples for major-ion analyses were filtered through a 0.45-µm membrane filter and stored in polyethylene bottles. Samples for cation analysis were acidified to a pH of about 1.5 with nitric acid (Wood, 1976). Samples for nutrient analyses were stored in opaque bottles, preserved with mercuric chloride, and kept at 4°C until analyzed. Dissolved organic carbon samples were filtered in a stainless steel assembly using a 0.45-µm silver membrane filter, stored in glass bottles, and kept at 4°C until analyzed. Samples for deuterium, oxygen-18, and tritium analyses were collected in glass bottles. Samples for carbon-14 analysis were collected in a 2-L linear polyethylene bottle that was attached to the bottom of a 50-gal precipitation tank. The tank was flushed with nitrogen gas prior to being filled with water. The pH of the water was raised to above 10 by adding a CO_2-free sodium hydroxide solution to convert all dissolved carbon to carbonate, and then a CO_2-free strontium chloride solution was added to the water to precipitate strontium carbonate. Samples for carbon-13 analysis were collected in a 1-L glass bottle by flushing the bottle with several volumes of sample water, filling the bottle with sample water, and then precipitating strontium carbonate with a CO_2-free ammoniacal strontium chloride solution. For both carbon isotope samples, suspended particles were filtered out of the water with an in-line filter closed to the atmosphere.

Major-ion and nutrient concentrations were determined by the National Water-Quality Laboratory of the U.S. Geological Survey, in Arvada, Colo., and the Desert Research Institute Laboratory in Reno, Nev. Dissolved organic carbon was analyzed at the U.S. Geological Survey National Water-Quality Laboratory. Deuterium and oxygen-18 were analyzed by the U.S. Geological Survey Research Laboratory in Reston, Va., and the Desert Research Institute Isotope Laboratory in Las Vegas, Nev. Carbon isotopes were analyzed by the U.S. Geological Survey National Water-Quality Laboratory, the Desert Research Institute Isotope Laboratory in Las Vegas, Nev., and the Teledyne Isotope Laboratory in Westwood, N.J. Tritium was analyzed at the U.S. Geological Survey National Water-Quality Laboratory and the Desert Research Institute Isotope Laboratory in Reno, Nev.

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GENERAL CHEMICAL CHARACTER OF GREAT BASIN GROUND WATER

Most water in the principal aquifers of the Great Basin, except in natural discharge and geothermal areas, contains less than 1,000 milligrams per liter (mg/L) of dissolved solids. Ground water in industrial, mining, urban, or agricultural areas, as well as in areas affected by dissolution of readily soluble minerals, can contain dissolved-solids concentrations in excess of 1,000 mg/L or constituent concentrations above National and State drinking water standards, or both (Lamb and Woodward, 1988; Parliman, 1988; Thomas and Hoffman, 1988; Waddell and Maxell, 1988).

Water containing natural (in contrast to human-affected) concentrations of dissolved solids exceeding 1,000 mg/L generally is in areas of evapotranspiration, evaporite deposits, or geothermal activity. Evapotranspiration in areas of shallow ground water (generally less than 20 ft below land surface) increases the dissolved-solids concentration of the residual water. Evapotranspirative concentration is most prevalent in ground-water discharge areas, such as playas, at the distal end of flow systems.

Evaporite salts and minerals, such as gypsum and halite, are highly soluble, and their dissolution results in a marked increase in dissolved solids. Evaporite salts and minerals are generally present in playa areas and in carbonate rock of Permian and younger age in the Great Basin (Hintze, 1980; Stewart, 1980).

Geothermal heating of water in aquifers generally produces higher concentrations of dissolved solids because the solubility of most minerals increases with temperature. Geothermal waters also can contain high concentrations of undesirable constituents, such as arsenic, boron, fluoride, and lithium. Geothermal activity occurs in localized areas throughout the Great Basin because of the extensional tectonic processes that have formed the characteristic basin-and-range structure (Fiero, 1986).

Human-induced degradation of water quality in industrial, mining, urban, and agricultural areas of the Great Basin is covered in several articles of the 1986 National Water Summary (Lamb and Woodward, 1988; Parliman, 1988; Thomas and Hoffman, 1988; Waddell and Maxell, 1988).

The chemical composition of water in basin-fill aquifers of the Great Basin containing less than 1,000 mg/L dissolved solids generally is dominated by calcium, sodium, magnesium, and bicarbonate. The chemical composition of water in basin-fill aquifers containing more than 1,000 mg/L dissolved solids generally is dominated by sodium, chloride, and sulfate. The chemical composition of water in carbonate-rock aquifers of the Great Basin generally is dominated by calcium, magnesium, and bicarbonate. Sodium also can be a dominant ion in these waters if volcanic rock or clay minerals are present before the ground water enters the carbonate-rock aquifers, or are found within the aquifers.

The chemical composition of water recharging aquifers in the Great Basin is derived from dissolution of soil-zone minerals and CO_2 gas. Chemical composition can change along flow paths as different processes take effect or as ground water comes in contact with different minerals. The chemical types of ground water in the principal aquifers of the Great Basin, based on the dominant ions dissolved in the water, are shown on four State maps at a scale of 1:500,000 (Thompson and Chappell, 1984a, b; Thompson and Nutter, 1984; Thompson and others, 1984) and in detail on 14 maps showing most of the Great Basin at a scale of 1:250,000 (Welch and Williams, 1986a–d, 1987a–j).

Basin-fill aquifers in the western Great Basin are derived by erosion of the predominantly volcanic rock in mountains that surround the basins (Plume and Carlton, 1988). Thus, these aquifers contain sediments that consist of volcanic glass and minerals primarily composed of sodium, potassium, calcium, silica, aluminum, and oxygen. Water chemistry in recharge areas and the upgradient part of the aquifers is dominated by sodium, calcium, and bicarbonate ions (Thompson and Chappell, 1984a, b; Thompson and Nutter, 1984; Thompson and others, 1984; Welch and Williams, 1986a–d, 1987a–j) because of the dissolution of volcanic glass and minerals by CO_2-rich water. Most of these basins and flow systems contain a discharging playa area. Near, or within, these playa areas, the chemical composition of ground water may evolve into a sodium dominated water, and with increasing dissolved-solids concentration, chloride and sulfate become dominant over bicarbonate. The water type changes because of (1) exchange of calcium and magnesium dissolved in the water for sodium on clay minerals; (2) dissolution of evaporative salts and minerals in sediments in the discharge area; (3) precipitation of minerals that removes select ions from the water; or (4) any combination of these processes.

Basin-fill aquifers in the eastern Great Basin are derived by erosion of the predominantly carbonate- and volcanic-rock mountains (Plume and Carlton, 1988). Thus, these aquifers contain sediments that consist of carbonate and volcanic minerals primarily composed of calcium, magnesium, and carbonate, in addition to sodium, potassium, silica, aluminum, and oxygen. Water chemistry in recharge areas and the upgradient part of aquifers, consisting mainly of carbonate rock, is dominated by calcium, magnesium, and bicarbonate ions, whereas aquifers consisting mainly of volcanic rock are dominated by sodium, calcium, and bicarbonate ions. Most basins in this part of the Great Basin do not have discharging playas, instead the basin-fill aquifers drain through carbonate rock to large springs. Thus, water chemistry along a flow path generally changes little in composition.

Carbonate-rock aquifers in the eastern Great Basin are composed primarily of calcite and dolomite. Water chemistry in recharge areas and the upgradient part of these aquifers is dominated by calcium, magnesium, and bicarbonate ions. As water flows through the carbonate-rock aquifers and comes in contact with volcanic rock and clay minerals, sodium ion concentration increases. Some carbonate-rock aquifers in the southeastern Great Basin contain interbedded evaporite deposits; as a result, sulfate, chloride, and sodium ions may predominate in ground water in these areas.

GEOCHEMICAL AND ISOTOPIC CHARACTERIZATION OF REPRESENTATIVE GREAT BASIN FLOW SYSTEMS

The Great Basin does not contain a single aquifer (as is common in other regional ground-water flow systems, such as, the Floridan, Madison, and Ogallala aquifers) but instead contains numerous aquifers. The two principal types of flow systems in the Great Basin are basin-fill and carbonate-rock aquifers. This report discusses geochemical processes that produce the major-ion chemistry and isotopic composition of ground water in aquifers representative of these two types. These processes are (1) evapotranspiration, (2) dissolution of minerals and CO_2 gas, (3) precipitation of minerals or formation by incongruent dissolution, (4) ion exchange, (5) mixing of chemically or isotopically different waters, and (6) geothermal heating.

Ground water is concentrated by evapotranspiration in areas of shallow water (less than about 20 ft below land surface) generally in the discharge areas of flow systems. This process is important in large areas of shallow ground water, such as playas.

Dissolution of soil-zone minerals and CO_2 gas in recharge areas of aquifers produces the chemical composition of water in the upper parts of aquifers. As residence time of water in an aquifer increases, constituent concentrations may increase until the water reaches saturation with respect to the dissolving mineral or gas. Water chemistry also changes as different minerals along the flow path dissolve.

Precipitation or formation of minerals can change the chemical composition of water because select ions are removed from the water. Mineral precipitation, in the principal aquifers of the Great Basin, is generally the result of evapotranspirative concentration of the water, temperature changes, or dedolomitization. Evapotranspirative concentration of ground water causes the water to become supersaturated with respect to some minerals, resulting in precipitation of those minerals. An increase in water temperature, generally caused by water circulating to depths of several thousand feet, causes some minerals to precipitate. For example, an increase in temperature from 10°C to 25°C of a water that is saturated with respect to calcite results in an approximately 25 percent decrease in total carbonate content because of calcite precipitation (Palmer and Cherry, 1984). Dedolomitization is the dissolution of dolomite and precipitation of calcite from a water saturated with respect to calcite and dolomite and undersaturated with respect to gypsum. This reaction is driven by the dissolution of gypsum (or anhydrite). Dedolomitization also increases sulfate and magnesium concentrations in the water.

Ion exchange removes one or more ions from the water while simultaneously adding one or more ions to the water. For example, calcium and magnesium ions may be removed from the water by exchanging with sodium and potassium ions on clay minerals.

Mixing waters of different chemical or isotopic compositions produces a water that is chemically or isotopically different from the original waters. For example, a mixture of 50 percent water with a chloride composition of 10 mg/L and 50 percent water with a chloride composition of 30 mg/L would result in a water containing 20 mg/L chloride.

Increased temperature due to deep circulation of ground water can increase or decrease the solubility of minerals (Palmer and Cherry, 1984) and cause a shift in the oxygen-18 composition of the water. For example, chalcedony becomes more soluble with increasing temperature, resulting in an increase in silica concentration with increased water temperature. Conversely, calcite solubility decreases with increasing temperature and calcite precipitates, so increased temperature removes calcium and carbon from the water. Higher water temperature also can result in a shift in oxygen-18 composition of the water because the exchange rate of oxygen in water for oxygen in minerals increases with increasing temperature (Gat and Gonfiantini, 1981).

GEOCHEMICAL EVOLUTION OF GROUND WATER IN A TYPICAL BASIN-FILL AQUIFER

Typical basin-fill aquifers consist of unconsolidated sedimentary deposits ranging from high-permeability sand and gravel to low-permeability clay and silt. These generally mixed deposits grade inward from poorly sorted alluvial-fan deposits, consisting of boulders to clay-size particles, around the margin of the basin to well-sorted sand and gravel deposits and ultimately into fine-grained playa deposits near the center of the basin. In some valleys the basin-fill deposits are surrounded and underlain by relatively low-permeability rock, such as volcanic rock, and in other valleys the basin-fill deposits are surrounded and underlain by rock that includes relatively high-permeability carbonate rock. Basin-fill aquifers surrounded by low-permeability rock are found throughout the Great Basin but are more common in the western part where low-permeability volcanic rock predominates. Smith Creek Valley in west-central Nevada is a typical basin-fill aquifer surrounded and underlain by low-permeability rock.

In most basin-fill aquifers surrounded by low-permeability rock, ground water flows within the basin-fill deposits and is discharged in an extensive area (tens to hundreds of square miles), commonly a centrally located playa surrounded by phreatophytic vegetation. Smith Creek Valley is hydrologically closed. Precipitation that falls within the topographic basin is discharged within the basin, except in the extreme southern part of the valley. Ground water flows southward from this part, because in this area the aquifer is hydrologically separated from the main basin-fill aquifer by igneous intrusions that form a barrier to northward ground-water flow (figs. 2, 3).

Smith Creek Valley encompasses 583 mi^2 and ranges in altitude from about 6,000 ft above sea level at the playa surface to more than 10,000 ft in the surrounding mountains. Climate differs from arid (about 6 in. of precipitation a year on the valley floor) to semiarid (up to about 20 in. of precipitation) in the mountains (Thomas, Carlton, and Hines, 1989, p. 5–8). The principal source of ground-water recharge probably is precipitation in the mountains above an altitude of 7,000 ft (Everett and Rush, 1964).

The hydrology of Smith Creek Valley has been studied by Everett and Rush (1964) and Thomas, Carlton, and Hines (1989). These studies describe the hydrology of the shallow part of the basin-fill aquifer and include a water budget for the basin. The ground-water chemistry of Smith Creek has been described by Thomas, Welch, and Preissler (1989).

GEOLOGIC FRAMEWORK

Consolidated rock and unconsolidated deposits in Smith Creek Valley can be divided into three hydrogeologic units: (1) low-permeability consolidated rock, (2) high-permeability basin-fill deposits, and (3) low-permeability playa deposits (fig. 2). Consolidated rock surrounds and underlies the unconsolidated basin-fill deposits. Basin-fill deposits consist of poorly sorted heterogeneous sediments at the basin margin that grade sequentially inward to well-sorted coarse-grained sediments, to mixed coarse-, medium-, and fine-grained sediments (heterogeneous sediments), to fine-grained playa sediments near the center of the basin (fig. 4).

The mineralogy of the basin-fill sediments and the water chemistry reflect to differing degrees the mineral composition of the rock composing the surrounding mountains. About 95 percent of the exposed consolidated rock are volcanic rock of Tertiary to Quaternary age (Stewart and McKee, 1977; Kleinhampl and Ziony, 1985). Rock composing the Desatoya Mountains on the west side of the basin and the southern Shoshone Mountains on the east side of the basin is almost entirely rhyolitic tuff. These mountain ranges are the principal source areas for the basin-fill deposits and also the source areas for approximately 95 percent of the recharge to the basin (Thomas, Carlton, and Hines, 1989, p. 16).

The Desatoya Mountains are composed primarily of an unnamed densely welded crystal-poor rhyolite ash-flow tuff containing less than 10 percent phenocrysts (Stewart and McKee, 1977, p. 42). Modal (mineralogic) analyses of this tuff (Barrows, 1972, p. 41 and 67) show that the phenocrysts consist mainly of plagioclase feldspar with lesser amounts of alkali feldspar and quartz (table 1). Four plagioclase samples range in calcium content from 17 to 45 percent (Barrows, 1972). A crystal-rich biotite-bearing ash-flow tuff composes 15 percent of the rock exposed in the Desatoya Mountains (Stewart and McKee, 1977); this tuff occupies the highest altitude zone of the range, the source area for about 35 percent of the recharge to the basin-fill aquifer (Thomas, Carlton, and Hines, 1989). This tuff contains

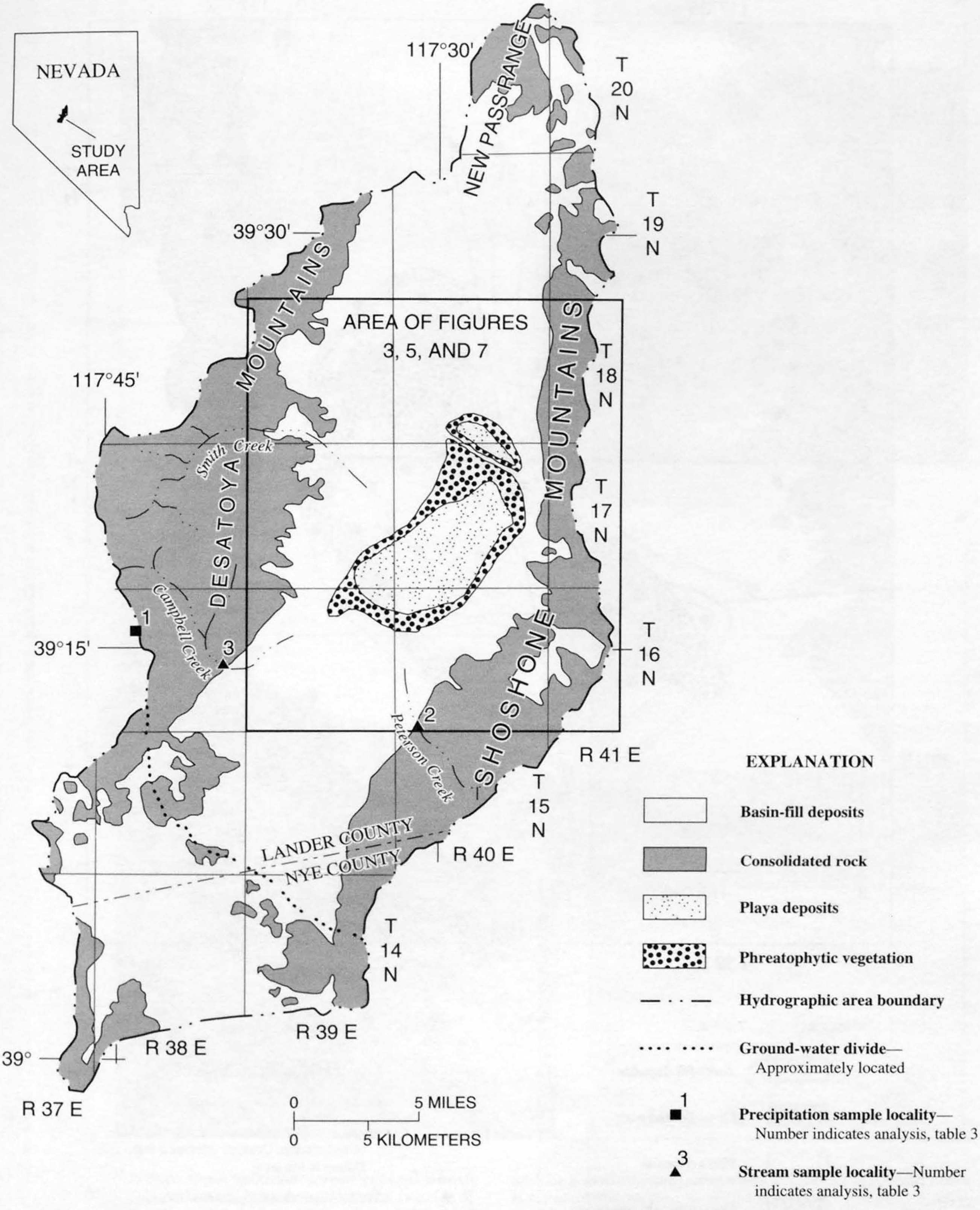

FIGURE 2.—Location and general features of Smith Creek Valley, Nev., showing precipitation and stream sample localities.

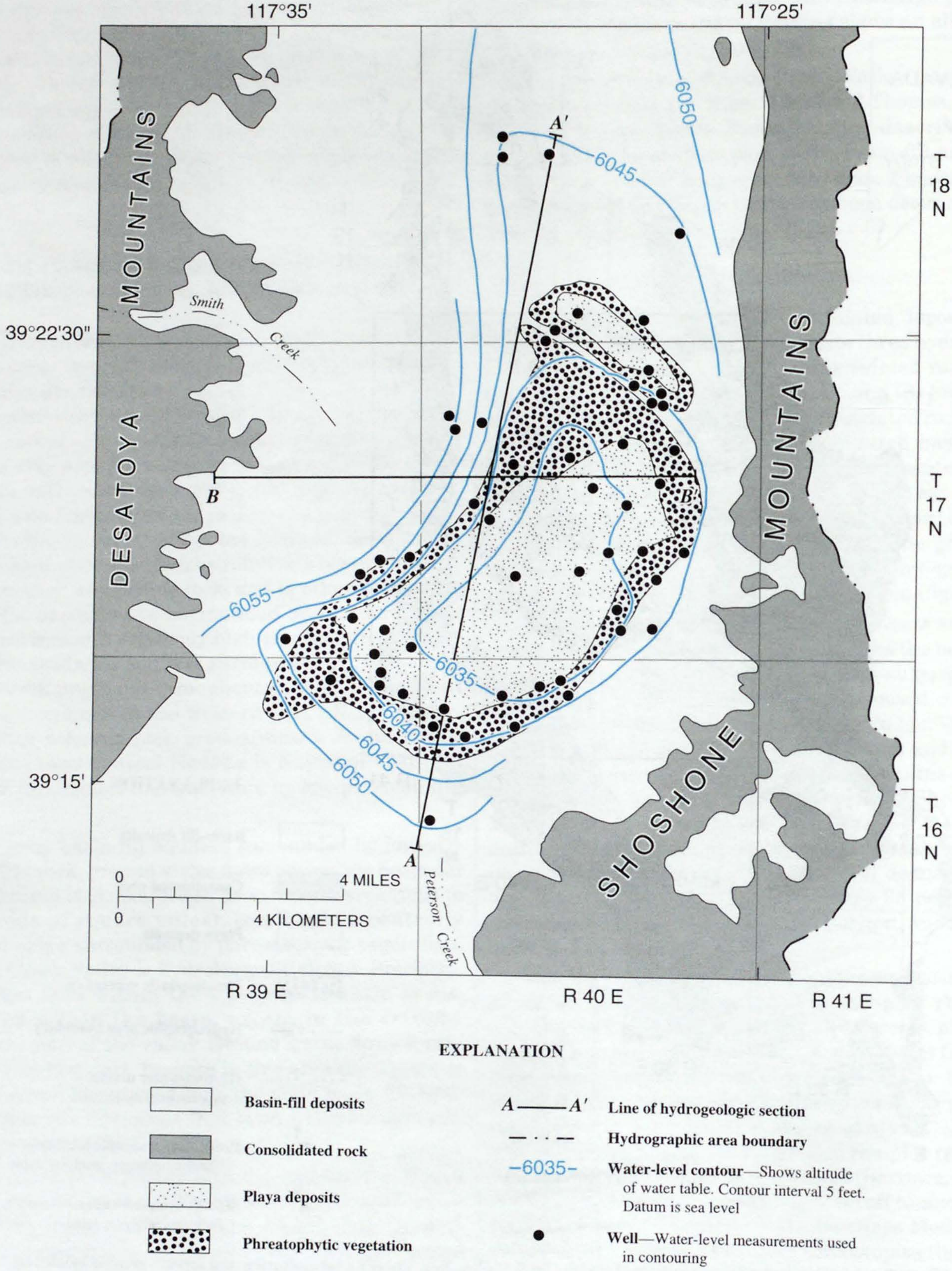

FIGURE 3.—Water-table configuration in central Smith Creek Valley, Nev., June 1982. Sections *A-A'* and *B-B'* shown in figure 4.

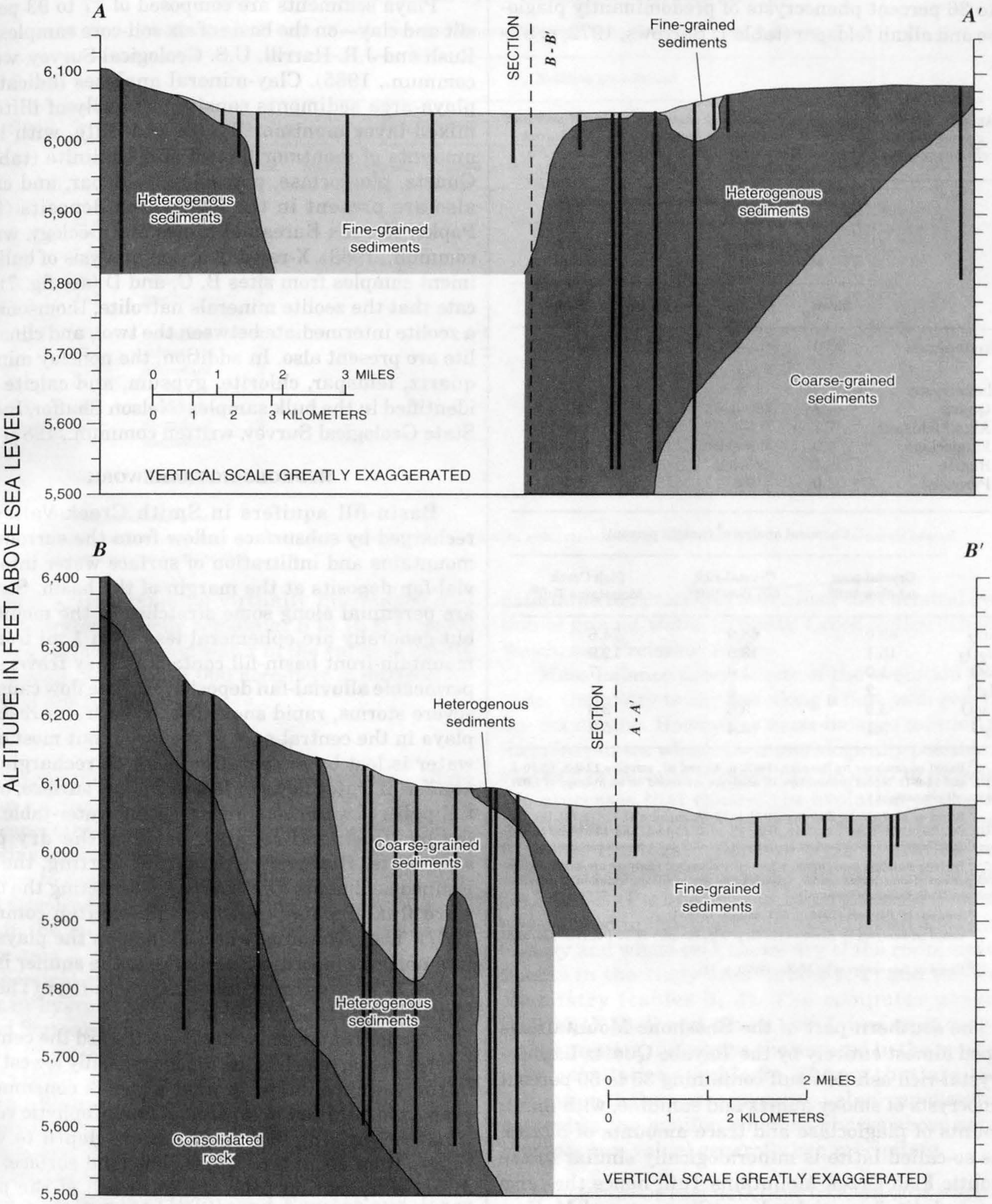

FIGURE 4.—Hydrogeologic sections across central Smith Creek Valley, Nev. Heterogeneous sediments consist of poorly sorted sediments at the margin of the basin and mixed coarse-, medium-, and fine-grained sediments near the center of the basin; coarse-grained sediments are coarse- to medium-grained well-sorted sands and gravels: and fine-grained sediments are primarily silts and clays. Hydrogeologic section locations are shown in figure 3. Wells (vertical lines) used to construct the sections, in addition to wells shown in figure 3, include geothermal temperature–gradient holes and observation wells that were destroyed prior to this study.

11 to 36 percent phenocrysts of predominantly plagioclase and alkali feldspar (table 1; Barrows, 1972, p. 87–88).

TABLE 1.—*Modal mineralogy and whole-rock chemistry of crystal-poor and crystal-rich ash-flow tuffs in the Desatoya Mountains and whole-rock chemistry of the Fish Creek Mountains Tuff (Smith Creek Valley), Nev.*

Mineralogic analyses (volume percent)

	Crystal-poor ash-flow tuff[1]		Crystal-rich ash-flow tuff[2]	
	Mean	Range	Mean	Range
Groundmass	93.0	90.2–96.0	74.2	64.2–89.2
Phenocrysts:				
Quartz	.7	0.0–1.4	1.9	0.0–4.7
Alkali feldspar	1.1	0.5–1.9	8.2	0.4–13.2
Plagioclase	4.9	0.9–9.0	13.2	8.7–18.9
Biotite	.1	0.0–0.2	1.5	0.0–3.6
Pyroxene	.0	0.0	.2	0.0–1.1

Chemical analyses[3] (weight percent)

	Crystal-poor ash-flow tuff[4]	Crystal-rich ash-flow tuff[5]	Fish Creek Mountains Tuff[6]
SiO_2	69.0	68.9	74.6
Al_2O_3	15.1	13.9	12.9
CaO	1.0	1.6	.7
MgO	.2	.3	.1
Na_2O	3.9	3.6	3.8
K_2O	5.1	4.4	5.2

[1] Based on analyses by Barrows (1972, p. 41 and 67, samples 12-9-3, 13-10-1, 13-11-1, and 14-9-1). Modal (mineralogical) analyses are based on an average of 1,021 point counts per sample.

[2] Based on analyses by Barrows (1972, p. 87-88, samples 11-11-D1, D2, D3, D4, D6, D7, D8, D9; 11-10-D10, D11, D13, D15; 12-10-4; 13-11-3, and 13-11-12). Modal (mineralogical) analyses are based on an average of 1,003 point counts per sample.

[3] By long-standing convention, whole-rock chemical analyses are expressed in weight percent of constituent oxides. Oxides listed are of silicon, aluminum, calcium, magnesium, sodium, and potassium.

[4] Analysis by Barrows (1972, p. 337, sample 13-10-1).

[5] Based on analyses by Barrows (1972, p. 338, samples 11-10-D9, 11-11-D4, and 11-11-D5).

[6] Based on two analyses by McKee (1970, p. 11).

The southern part of the Shoshone Mountains is capped almost entirely by the Toiyabe Quartz Latite—a crystal-rich ash-flow tuff containing 35 to 50 percent phenocrysts of smoky quartz and sanidine, with minor amounts of plagioclase and trace amounts of biotite. This so-called latite is mineralogically similar to the rhyolitic Fish Creek Mountains Tuff, hence the term "quartz latite" is misleading (Stewart and McKee, 1977, p. 43). Although modal and chemical analyses are not available for the Toiyabe Quartz Latite, its mineralogic similarity to the Fish Creek Mountains Tuff implies that the whole-rock chemistry also may be similar (table 1).

Playa sediments are composed of 77 to 93 percent silt and clay—on the basis of six soil-core samples (F.E. Rush and J.R. Harrill, U.S. Geological Survey, written commun., 1965). Clay-mineral analyses indicate the playa-area sediments consist primarily of illite and mixed-layer montmorillonite and illite, with lesser amounts of montmorillonite and kaolinite (table 2). Quartz, plagioclase, potassium feldspar, and calcite also are present in the playa-area deposits (Keith Papke, Nevada Bureau of Mines and Geology, written commun., 1983). X-ray diffraction analysis of bulk sediment samples from sites B, C, and D (see fig. 7) indicate that the zeolite minerals natrolite, thomsonite (or a zeolite intermediate between the two), and clinoptilolite are present also. In addition, the nonclay minerals quartz, feldspar, chlorite, gypsum, and calcite were identified in the bulk samples (Nelson Shaffer, Indiana State Geological Survey, written commun., 1987)

HYDROLOGIC FRAMEWORK

Basin-fill aquifers in Smith Creek Valley are recharged by subsurface inflow from the surrounding mountains and infiltration of surface water into alluvial-fan deposits at the margin of the basin. Streams are perennial along some stretches in the mountains but generally are ephemeral less than 1 mi from the mountain-front basin-fill contact as they traverse the permeable alluvial-fan deposits. Surface flow caused by severe storms, rapid snowmelt, or both can reach the playa in the central part of the basin, but most of this water is lost by evaporation prior to recharging the aquifer. If water flows onto a dry playa surface, an initial pulse of water may recharge the water-table aquifer by flowing through cracks in the dry playa sediments. However, after initial wetting, the fine-grained sediments swell quickly, restricting the downward flux of water (J.R. Harrill, written commun., 1967). The remaining water stands on the playa surface until it evaporates. Recharge to the aquifer is estimated to be approximately 8,000 acre-ft/yr (Thomas, Carlton, and Hines, 1989, p. 15–16).

Ground water generally flows toward the centrally located playa, which is the topographically lowest point in the valley. Shallow ground water is consumed by evapotranspiration in an area of phreatophytic vegetation surrounding the playa, where depth to water ranges from about 8 to 30 ft below land surface, or by direct evaporation from the bare soil of the playa, where depth to water is about 6 to 10 ft below land surface. Ground water moves upward in the discharging playa area, as indicated by the upward gradient measured in pairs of shallow wells (fig. 5). The vertical gradient remains relatively constant over time. Even when water is standing on the playa surface, the

TABLE 2.—*X-ray diffraction analyses of playa-area sediments, Smith Creek Valley, Nev.*

[Analyses by Keith Papke, Nevada Bureau of Mines and Geology, Reno, Nev.; --, not identified in sample]

Site[1]	Sample depth (feet)	Clay content	Relative abundance[2]							
			Clay minerals				Nonclay minerals			
			Illite	Montmorillonite	Mixed layer[3]	Kaolinite	Quartz	Plagioclase	K-feldspar	Calcite
A	2.0	moderate	1	3	--	2	1	2	3	--
	21.5	minor	1	2	--	--	1	2	--	--
B	5.5	abundant	3	1	--	2	1	3	2	--
	23.0	minor	1	--	2	--	1	--	2	--
	34.0	moderate	1	--	2	3	1	3	2	--
C	7.5	moderate	2	3	1	4	1	2	3	4
	22.5	minor	1	--	2	--	1	--	2	--
	34.0	abundant	2	--	1	3	2	3	1	--
D	5.5	moderate	1	--	2	--	1	2	--	3
	22.5	abundant	2	--	1	3	1	2	--	3
	27.0	abundant	2	--	1	3	1	2	--	3
E	16.5	moderate	2	--	1	3	1	2	3	--

[1] Site designation from figure 7.

[2] Mineral abundance is arbitrary ranking of relative amounts of mineral in sample, with 1 representing most abundant mineral and 4 least abundant mineral.

[3] Interlayered montmorillonite-illite.

upward gradient remains virtually unchanged and depths to water in wells remain at 6 to 10 ft beneath the playa surface. Ground water in the valley is mostly undeveloped; less than 1,000 acre-ft/yr is pumped for irrigation and livestock use.

WATER CHEMISTRY

Water in the basin-fill aquifer evolves from a dilute sodium calcium bicarbonate (Na-Ca-HCO_3) type to a more concentrated sodium bicarbonate (Na-HCO_3) type to a briny sodium chloride (Na-Cl) type as it flows from recharge to discharge areas (table 3; figs. 6, 7). Measured dissolved-solids concentration increases from 90 mg/L in the recharge areas to 51,000 mg/L in the discharge area

Chemical evolution of ground water and surface water in hydrologically closed basins of the Western United States has been studied by Jones (1965), Hardie (1968), Phillips and Van Denburgh (1971), Van Denburgh (1975), and Smith and Drever (1976), and outside the Western United States by Lerman (1967), Eugster (1970), Jones and others (1977), Rettig and others (1980), Yuretich and Cerling (1983), Green and Canfield (1984), and Macumber (1984). Jones (1966), Hardie and Eugster (1970), and Eugster and Jones (1979) summarized the geochemical processes that affect the chemical evolution of water from dilute waters to brines in hydrologically closed basins. Deuterium, oxygen-18, geochemical data, and interpretive methods described in previous studies were used to determine the processes that affect the chemical evolution of ground water in Smith Creek Valley (Thomas, Welch, and Preissler, 1989).

Mass-balance descriptions of the evolution of one water chemistry to another along a flow path generally are not unique. However, a mass-balance solution that includes phases which are mineralogically possible and thermodynamically feasible can be used to determine the processes that control the evolution of observed water chemistry. Stoichiometry of reactions associated with the chemical evolution were identified using mass-balance calculations by the computer program BALANCE (Parkhurst and others, 1982). Plausible phases used in the calculations were based on the mineralogy and whole-rock chemistry of the rocks and sediments in the study area (tables 1, 2) and the water chemistry (tables 3, 4). The computer program WATEQ4F (Ball and others, 1987) was used to calculate saturation indices for phases used in the mass-balance calculations (table 5). Phases that are below saturation in the water (negative value) can dissolve, if present in the aquifer, whereas phases above saturation (positive value) generally can precipitate.

EVOLUTION OF PRECIPITATION TO RECHARGE WATER

Water recharging the basin-fill aquifer is a sodium calcium bicarbonate type (samples from sites 2 and 3; table 3, fig. 6). Dissolved constituents in the recharge water are primarily from evapotranspirative concentration of dissolved constituents before the water

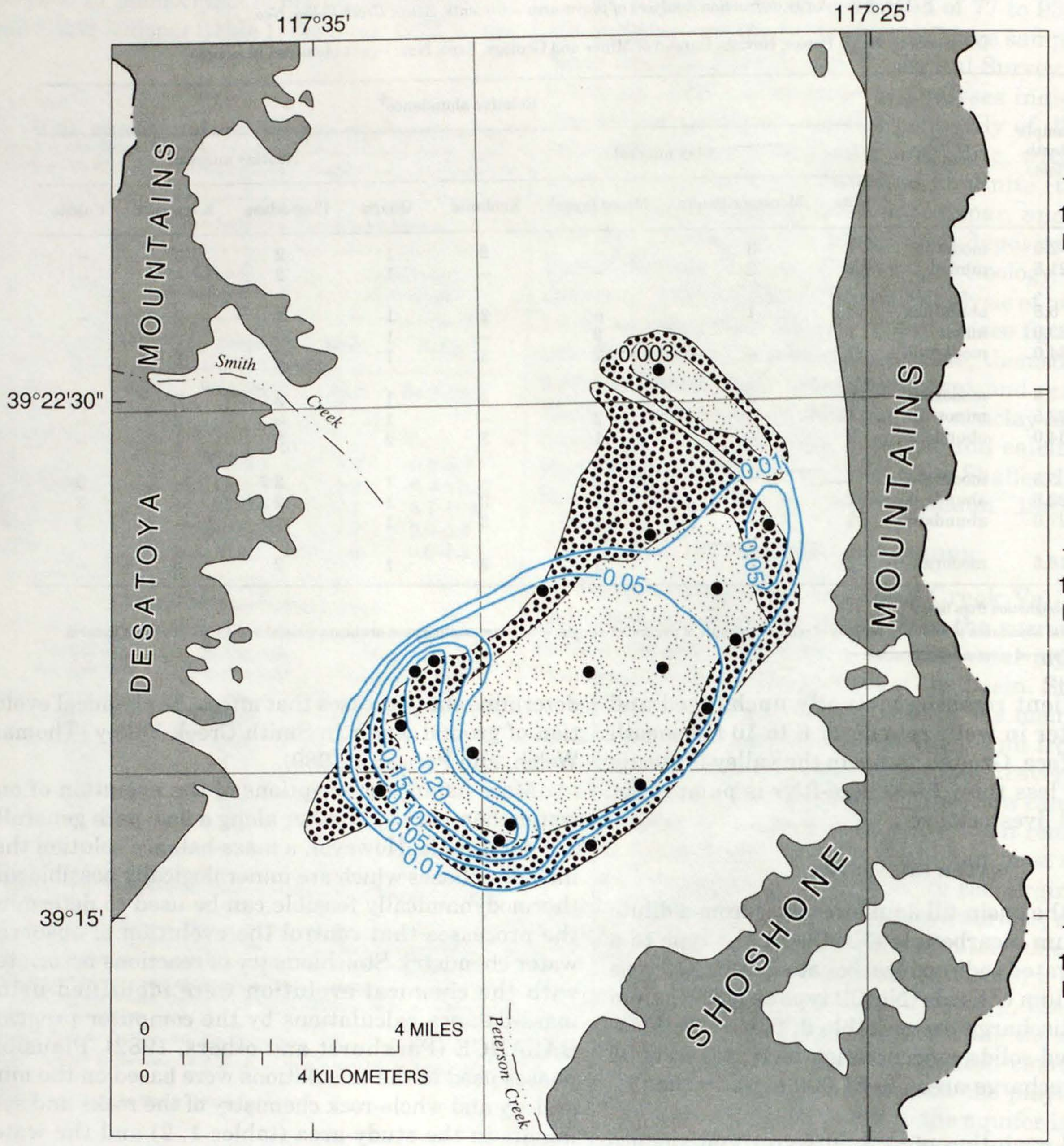

FIGURE 5.—Upward hydraulic gradient in the discharge area of central Smith Creek Valley, Nev.

TABLE 3.—*Chemical analyses of water from Smith Creek Valley, Nev.*

[Milligrams per liter, unless noted otherwise. Symbol: --, not determined]

Site[1]	Site name	Well depth (feet)	Water temperature (degrees Celsius)	pH (standard units)	Calcium	Magnesium	Sodium	Potassium	Bicarbonate	Carbonate	Sulfate	Chloride	Silica	Dissolved solids[2] (calculated)	Delta deuterium (permil)	Delta oxygen-18 (permil)
	Precipitation															
1	Carroll Summit[3]		--	--	1.1	0.2	0.7	0.3	--	--	1.5	2.2	--	--	--	--
	Recharge water															
2	Peterson Creek		15.3	7.6	6.6	1.0	13	1.5	41	0	10	3.8	34	90	–123	–16.2
3	Campbell Creek		12.9	8.0	13	2.1	19	2.6	76	0	12	7.3	38	130	–118	–15.2
	Sodium calcium bicarbonate water															
4	S. Brown well	282	10.6	6.8	20	3.8	24	3.8	94	0	13	8.2	56	180	–123	–15.7
5	USGS well 35	66	9.8	6.5	29	4.2	33	7.9	100	0	43	28	56	250	–118	–14.7
6	N. Brown well	200	14.7	7.5	35	5.5	48	6.8	150	0	52	25	62	310	–130	–16.2
	Sodium bicarbonate water															
7	USGS well 67	22	9.0	6.7	12	1.9	56	6.7	160	0	21	11	34	220	–124	–15.6
8	USGS well 50	22	10.0	7.2	10	1.1	110	6.0	300	0	11	20	19	320	–122	–15.9
9	USGS well 27	22	8.1	6.9	21	3.1	170	8.4	400	0	12	85	12	510	–117	–14.8
	Sodium chloride water															
10	USGS well 23	66	10.4	8.4	110	13	710	17	32	0	180	1,200	11	2,300	–132	–16.4
11	USGS well 55	66	9.8	8.1	76	14	2,200	27	170	0	380	3,600	5.2	6,400	–124	–15.2
12	USGS well 70	22	9.2	7.4	73	41	6,400	75	1,300	0	2,700	8,900	24	19,000	–122	–14.2
13	USGS well 56	22	10.8	9.2	10	1.2	11,000	81	3,500	280	3,500	12,000	5.0	29,000	–98	–9.7
14	USGS well 68	22	9.1	6.6	150	10	13,000	86	480	0	4,300	20,000	3.8	38,000	–101	–10.0
15	USGS well 32	(4)	10.9	7.5	490	100	14,000	140	140	0	6,000	21,000	2.5	42,000	–110	–12.0
16	USGS well 24	22	9.0	6.8	850	340	16,000	290	200	0	8,700	25,000	20	51,000	–118	–13.8

[1] Number corresponds to site shown in figures 2 or 7.

[2] Summation includes bicarbonate multiplied by 0.4916 to make result comparable to a "residue-on-evaporation" value.

[3] Analysis is average for 11 samples.

[4] Sample probably represents shallow water at depth of about 15 feet, rather than at 66-foot total depth.

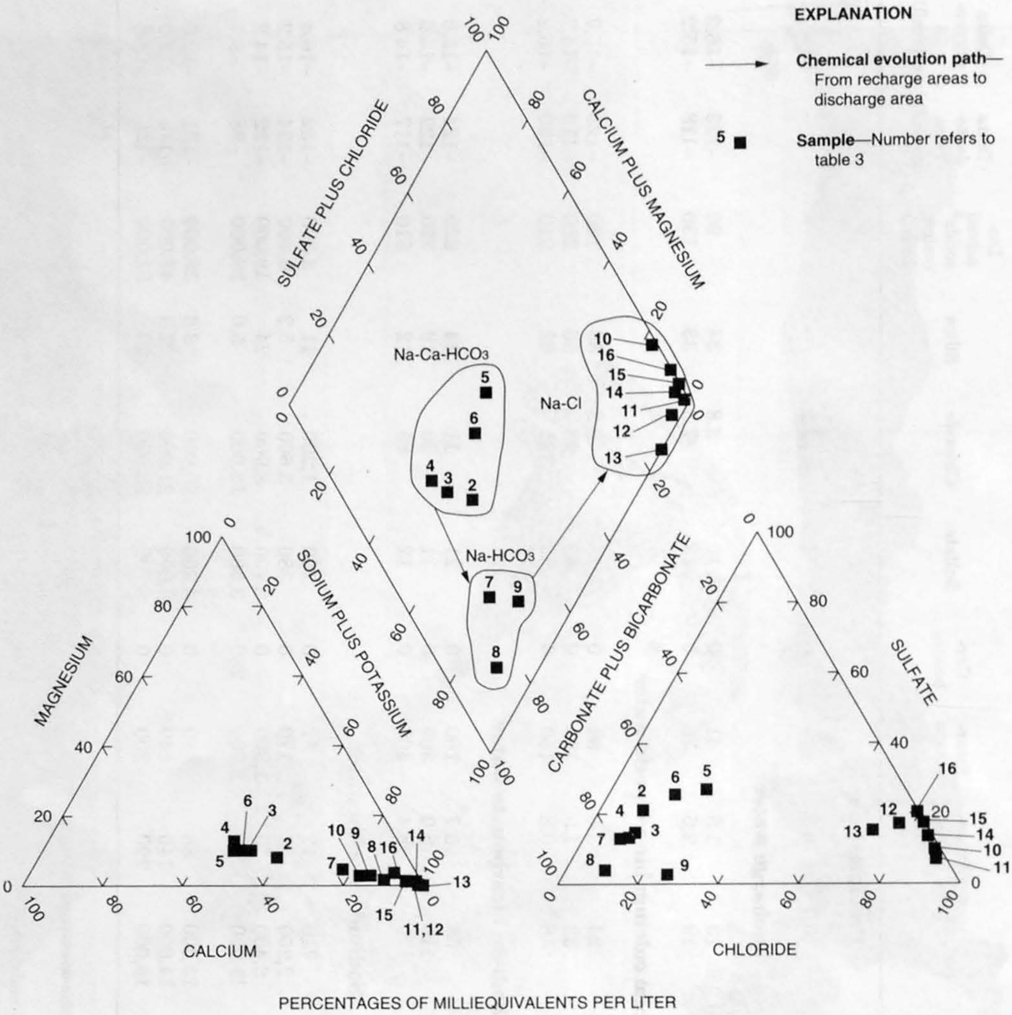

FIGURE 6.—Chemical evolution of ground water in the basin-fill aquifer of Smith Creek Valley, Nev.

percolates below the root zone and dissolution of soil-zone carbon dioxide gas and the volcanic groundmass and phenocrysts. These dissolution processes are similar to dissolution processes in the Absaroka Mountains in Wyoming (Miller and Drever, 1977) and in the Oasis Valley in Nevada (White, 1979). Constituents dissolved in precipitation are concentrated by evapotranspiration in the recharge areas. This evapoconcentration is accounted for by assuming the increase in chloride in the recharge water (table 4) is due only to evapotranspiration before the water percolates below the root zone.

Windblown gypsum is commonly deposited in uplands of desert environments (Pewe, 1981) and is the main component of dust in southern Nevada (Ronald Amundson, University of California, Berkeley, oral commun., 1987; Marith Reheis, U.S. Geological Survey, oral commun., 1987). Thus, gypsum is the most likely source of sulfate dissolved in recharge waters in Smith Creek Valley. Pyrite, if present in the volcanic rock, also may supply sulfate to the ground water in the upland areas. The increase in chloride concentration was assumed to result from evapotranspiration; however, part of the increase could be from dissolution of salts containing chloride that have blown into the recharge area. The small increase in magnesium concentration [0.04 millimoles per liter (mmol/L; table 4)] is most likely the result of magnesium biotite dissolution, because biotite is in volcanic rock in the recharge areas (table 1). Dissolution of illite, chlorite, or both, which are in the basin-fill sediments, also may be a source of magnesium. Aluminum and silica released by the weathering of volcanic groundmass and phenocrysts probably are removed from the water by formation of kaolinite, or some other clay mineral, and chalcedony (table 5). A mass-balance solution for the evolution of precipitation, concentrated by evapotranspiration, to recharge water is given by mass-balance solution 1 in table 6.

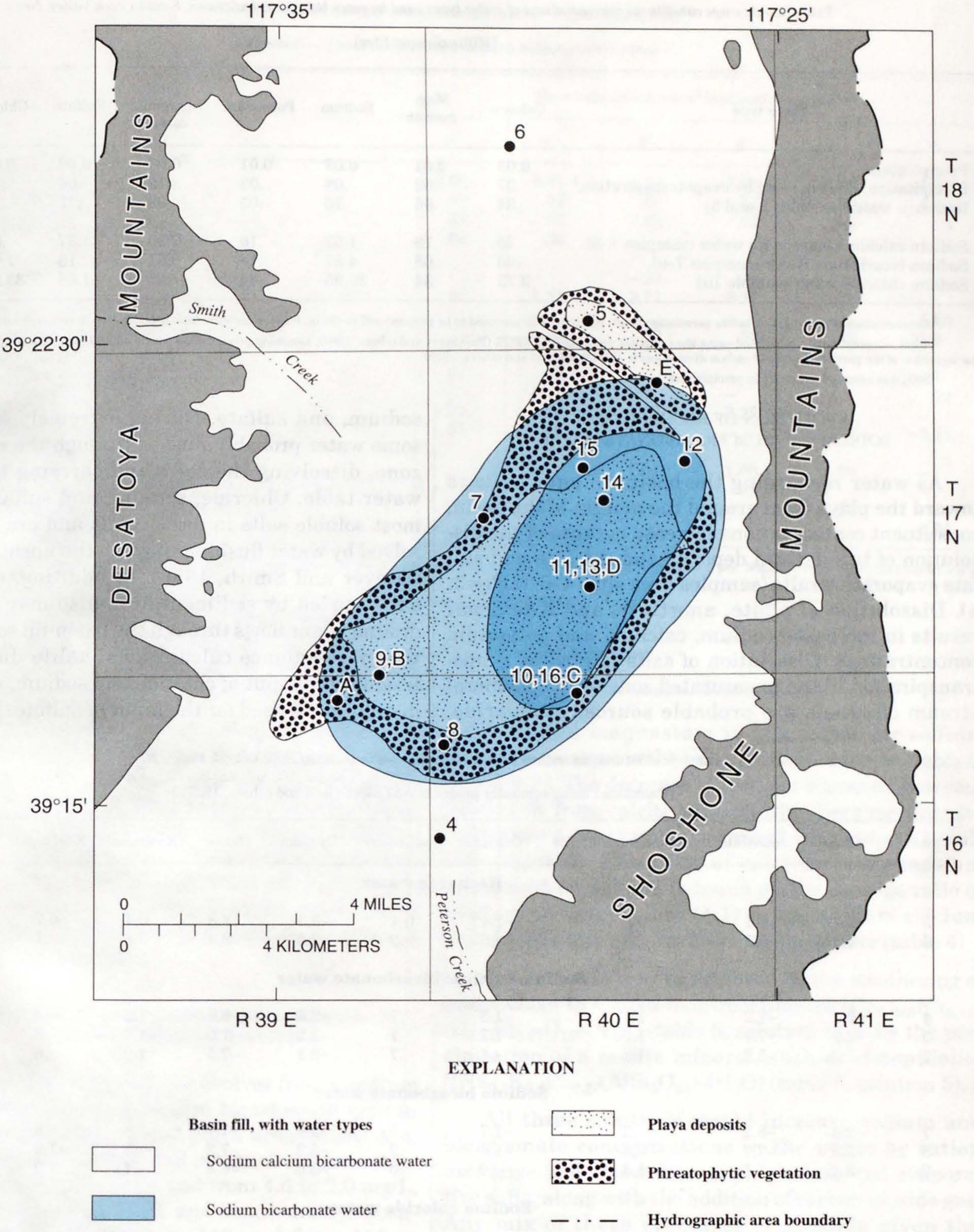

FIGURE 7.—Water types in shallow basin-fill aquifer of central Smith Creek Valley, Nev., based on dominant ions. Sites having two wells list the deeper well (66 feet) first and the shallower well (22 feet) second.

TABLE 4.—*Average constituent concentrations of water types used in mass-balance calculations, Smith Creek Valley, Nev.*

[Millimoles per liter]

Water type	Calcium	Magnesium	Sodium	Potassium	Total inorganic carbon	Sulfate	Chloride	Silica
Precipitation	0.03	0.01	0.03	0.01	[2]0.03	0.02	0.06	[3]0.00
Precipitation, concentrated by evapotranspiration[1]	.07	.02	.08	.02	.12	.04	.16	[3].00
Recharge water (samples 2 and 3)	.24	.06	.70	.05	.99	.11	.16	.60
Sodium calcium bicarbonate water (samples 4–6)	.70	.19	1.52	.16	2.50	.37	.58	.97
Sodium bicarbonate water (samples 7–9)	.36	.08	4.87	.18	6.21	.15	1.09	.36
Sodium chloride water (sample 10)	2.75	.54	30.95	.44	.51	1.88	33.93	.18

[1] Concentration of precipitation before percolation below root zone was assumed to be proportional to the increase in chloride from precipitation to recharge water.

[2] Total inorganic carbon calculated using the computer program PHREEQE (Parkhurst and others, 1980), assuming precipitation was in equilibrium with the atmosphere; that is, the logarithm of the partial pressure of carbon dioxide gas is –3.5 (Sundquist and others, 1979).

[3] SiO_2 was assumed to be zero in precipitation.

EVOLUTION OF RECHARGE TO SODIUM CALCIUM BICARBONATE WATER

As water recharging the basin-fill aquifer flows toward the playa from around the margin of the basin, constituent concentrations increase because of the dissolution of tuff-derived deposits and chloride and sulfate evaporative salts (samples from sites 4–6; tables 3, 4). Dissolution of albite, anorthite, and K-feldspar results in increased sodium, calcium, and potassium concentrations. Dissolution of salts formed by evapotranspiration in the unsaturated zone or in ephemeral stream channels are probable sources of chloride, sodium, and sulfate. During extremely wet periods, some water probably flushes through the unsaturated zone, dissolving the salts and carrying them to the water table. Chloride, sodium, and sulfate form the most soluble salts in desert soils and are readily dissolved by water flushing through the unsaturated zone (Drever and Smith, 1978). In addition, evaporative salts buried by sedimentation also may dissolve as ground water flows through the basin-fill sediments. In the mass-balance calculations, halite dissolution is used for the input of chloride and sodium, and gypsum dissolution is used for the input of sulfate. Bicarbonate

TABLE 5.—*Saturation indices for selected minerals in water of Smith Creek Valley, Nev.*[1]

[Calculated using computer program WATEQ4F (Ball and others, 1987).]

Site[2]	Albite	Anorthite	Ca-montmorillonite	Calcite	Chalcedony	Gypsum	Halite	Kaolinite	K-feldspar	Mg-biotite
Recharge water										
2	–2.2	–6.1	0.3	–1.4	0.4	–3.3	–8.8	0.7	–0.7	–34.0
3	–1.5	–5.1	.8	–.6	.5	–3.0	–8.4	.8	.1	–30.5
Sodium calcium bicarbonate water										
4	–2.1	–7.2	1.1	–1.5	.7	–2.8	–8.2	1.2	–.4	–37.9
5	–2.2	–7.7	1.0	–1.7	.7	–2.2	–7.6	1.2	–.3	–39.8
6	–1.0	–5.2	1.4	–.4	.7	–2.1	–7.5	1.2	.6	–31.6
Sodium bicarbonate water										
7	–2.4	–8.2	.2	–1.7	.5	–2.8	–7.7	.8	–.8	–40.2
8	–2.4	–7.7	–.6	–1.0	.2	–3.2	–7.2	.3	–1.2	–38.1
9	–3.1	–8.6	–1.3	–.9	.0	–3.0	–6.4	–.1	–1.9	–39.8
Sodium chloride water										
10	–1.2	–4.9	–.9	.0	.0	–1.4	–4.7	–.2	–.3	–27.3

[1] Saturation index = $\log \left[\frac{\text{ion-activity product}}{K_T} \right]$, where K_T = equilibrium constant at temperature T. By convention, positive value indicates mineral can precipitate from solution, whereas negative value indicates mineral can dissolve if present in the aquifer. Aluminum concentrations were calculated using computer program PHREEQE (Parkhurst and others, 1980), assuming each water was saturated with respect to gibbsite and using thermodynamic data for gibbsite from Robie and others (1978, p. 140).

[2] Number corresponds to site listed in table 3 and shown in figures 2 or 7.

TABLE 6.—*Mass transfer of constituents in Smith Creek Valley, Nev.*[1]

[Symbol: --, phase not included in mass-balance calculation]

Phase	Mass-balance solutions[2] (millimoles per liter)					
	1	2	3	4	5	6
Albite [$NaAlSi_3O_8$]	0.62	0.41	1.79	1.94	2.26	1.72
Anorthite [$CaAl_2Si_2O_8$]	.10	.19	.08	.17	.57	6.36
K-feldspar [$KAlSi_3O_8$]	.02	.06	.02	.02	.67	.26
Gypsum [$CaSO_4 \cdot 2H_2O$]	.08	.26	--	--	--	1.73
H_2S (gas)	--	--	–.22	–.22	–.22	--
CO_2 (gas)	.87	1.51	3.71	3.71	3.71	--
Calcite [$CaCO_3$]	--	--	--	--	--	–5.70
Halite [NaCl]	--	.42	.52	.52	.52	32.8
Mg-biotite [$KMg_3Si_3AlO_{10}(OH)_2$]	.01	.04	--	--	--	--
Chlorite [$Mg_5\,Al_2\,Si_3\,O_{10}(OH)_8$]	--	--	--	--	--	.09
Chalcedony [SiO_2]	–.71	–.66	–4.23	–3.22	--	--
Kaolinite [$Al_2Si_2O_5(OH)_4$]	–.43	–.45	–.99	--	–1.23	–3.21
Ca,Mg-Na exchange[3]	--	--	1.04	1.06	1.06	--
Ca/Na-montmorillonite [$Ca_{0.083}Na_{0.167}Al_{2.33}Si_{3.67}O_{10}(OH)_2$]	--	--	--	–.99	--	--
Clinoptilolite [$(Na_{0.3}K_{0.4}Ca_{0.3})(AlSi_5\,O_{12}) \cdot 4H_2O$]	--	--	--	--	–1.62	--
Natrolite [$Na_2(Al_2Si_3O_{10}) \cdot 2H_2O$]	--	--	--	--	--	–4.24

[1] Mass-transfer value is change in mass of indicated phase, calculated from changes in water chemistry between different water types. Negative value indicates transfer out of solution, and positive value indicates transfer into solution.

[2] Mass-balance solutions 1–6 correspond to mass balances between the following water types given in table 4:

1. From precipitation concentrated by evapotranspiration to recharge water;
2. From recharge water to sodium calcium bicarbonate water;

3–5. Three examples of mass-balance solutions for sodium calcium bicarbonate water to sodium bicarbonate water;

6. From sodium bicarbonate water to the most dilute sodium chloride water.

[3] Positive value indicates calcium and magnesium in water are being exchanged for sodium on clay minerals. Values are millimoles of sodium per liter.

concentration increases from the reaction of carbonic acid, which is derived from soil-zone carbon dioxide, with the volcanic groundmass and phenocrysts in the tuff-derived deposits. Magnesium concentration increase is probably from biotite dissolution. Feldspar hydrolysis releases cations and results in the formation of kaolinite, or some other clay mineral, and chalcedony. A mass-balance solution for the evolution of recharge water to sodium calcium bicarbonate water is given by solution 2 in table 6.

EVOLUTION OF SODIUM CALCIUM BICARBONATE TO SODIUM BICARBONATE WATER

Ground water in the aquifer evolves from a sodium calcium bicarbonate to a sodium bicarbonate type in the vicinity of the fine-grained playa area (tables 3, 4; figs. 6, 7). Average calcium and magnesium concentrations decrease from 28 to 14 and from 4.5 to 2.0 mg/L, respectively; average sodium and bicarbonate concentrations increase from 35 to 112 and from 115 to 287 mg/L, respectively (samples from sites 4–6 and 7–9; table 3). The removal of calcium and magnesium from the water may be the result of several processes, but the presence of montmorillonite in the discharge area (table 2, fig. 7) indicates that cation exchange of calcium and magnesium in the water for sodium adsorbed on clays is the most probable process (table 6, solution 3). The decrease in calcium concentration cannot result from calcite precipitation because samples from sites 4–9 are undersaturated with respect to calcite (table 5). The ratio of calcium to magnesium exchanged for sodium is based on the average ratio of calcium to magnesium (4:1) in the sodium calcium bicarbonate and sodium bicarbonate waters (table 4).

Calcium also may be removed by the weathering of plagioclase to Ca/Na-montmorillonite [$Ca_{0.083}Na_{0.167}Al_{2.33}Si_{3.67}O_{10}(OH)_2$] (table 6, solution 4) or by the precipitation of a zeolite mineral, such as clinoptilolite [$(Na_{0.3}K_{0.4}Ca_{0.3})(AlSi_5O_{12}) \cdot 4H_2O$] (table 6, solution 5).

All three solutions would increase sodium and bicarbonate concentrations in the water by cation exchange and dissolution of plagioclase and evaporative salts, along with the addition of carbon dioxide gas. Any mix of these processes is possible given the observed chemistries.

An activity diagram for the system CaO-CO_2-Na_2O-SiO_2-Al_2O_3-H_2O (fig. 8) shows that the different water types plot along a 2:1 slope, and in particular, sodium bicarbonate waters plot close to the phase

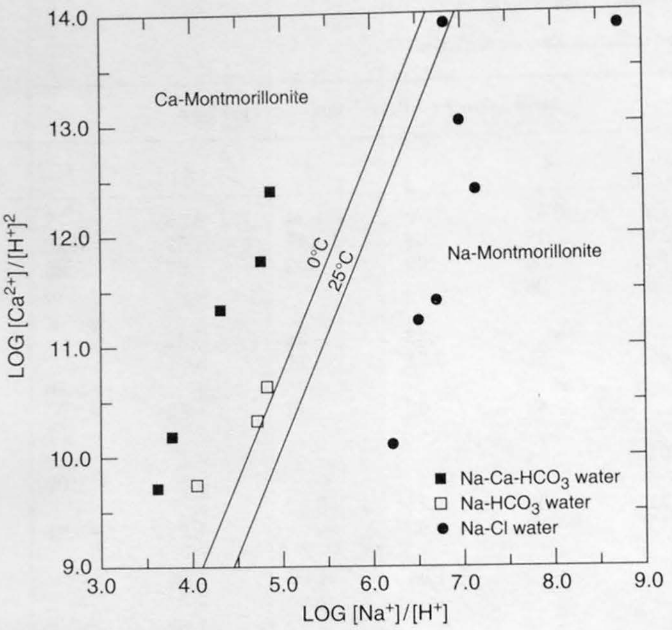

FIGURE 8.—Aqueous activities for the system CaO-CO_2-Na_2O-SiO_2-Al_2O_3-H_2O in Smith Creek Valley, Nev. The phase boundary is based on the thermodynamic data of Arnorsson and others (1982). Chemical symbols: Ca^{2+}, calcium; H^+, hydrogen; Na^+, sodium. Bracketed items are ion activities.

boundary between Ca-montmorillonite and Na-montmorillonite as would be expected for either cation exchange or formation of Ca/Na-montmorillonite.

Precipitation of zeolite minerals in the Smith Creek Valley basin-fill aquifer is evidenced by the presence of zeolite minerals in the basin-fill deposits. Furthermore, rhyolitic tuff and basin-fill aquifers derived from rhyolitic tuff in southern Nevada, similar to those in Smith Creek Valley, contain ground water in which calcium is removed by weathering of tuff or tuff-derived deposits to montmorillonite and by precipitation of zeolite minerals, predominantly clinoptilolite with a composition of $(Na_{0.3}K_{0.4}Ca_{0.3})(AlSi_5O_{12})\cdot 4H_2O$ (Hoover, 1968; White, 1979; White and others, 1980; Claassen, 1985).

Chloride and potassium concentrations increase and sulfate concentration decreases as the water evolves from a sodium calcium bicarbonate to a sodium bicarbonate type. Chloride concentration increases along the flow path by evapotranspiration of the water and dissolution of chloride salts. Potassium concentration increases, probably because of K-feldspar dissolution. Sulfate concentration decreases along the flow path probably because of the reduction of sulfate to hydrogen sulfide gas and precipitation of FeS (Jones, 1966; Phillips and Van Denburgh, 1971; Eugster and Jones, 1979) or possibly because of surface sorption (Wood, 1978; Eugster and Jones, 1979), because more fine-grained sediments with larger surface areas are present in the playa area. The sodium bicarbonate waters are undersaturated with respect to gypsum (table 5), so gypsum precipitation cannot be the cause of the decrease in sulfate.

EVOLUTION OF SODIUM BICARBONATE TO THE MOST DILUTE SODIUM CHLORIDE WATER

In the playa area, ground water evolves from a sodium bicarbonate to a sodium chloride water (tables 3, 4; figs. 6, 7). Constituent concentrations increase because of the dissolution of evaporative salts, evapotranspiration, and mixing.

Dissolution of evaporative salts probably is the major source of increasing chloride and sodium concentrations as the water evolves from sodium bicarbonate to sodium chloride type because (1) chloride and sodium concentrations increase in about equal amounts, (2) other constituents increase less than chloride, and (3) the isotopic composition of the dilute sodium chloride water is similar to that of the sodium bicarbonate water. Evaporative salt dissolution is the simplest explanation consistent with all three observations.

Dissolution of evaporative salts containing chloride and sodium should add these constituents to the water in about equal amounts (Drever and Smith, 1978). The approximately 30 mmol/L increase of chloride and sodium observed as the water evolves from sodium bicarbonate (samples from sites 7–9; fig. 7) to the most dilute sodium chloride (sample from site 10; fig. 7) water (table 4) supports this dissolution process. The increase in sulfate, calcium, magnesium, and potassium also may be from the dissolution of salts or minerals (table 6, solution 6) that are below saturation in the water [table 5; also the saturation index of chlorite (not shown) for the sample from site 10 is −1.4].

Sodium, sulfate, calcium, magnesium, and potassium increase in concentration as the water evolves from a sodium bicarbonate to sodium chloride water, but they do not increase in proportion to chloride as would be expected if evapotranspiration were occurring (table 4, figs. 9–13). This is a second indication that dissolution of evaporative salts and minerals is a major source of constituents. Constituent concentrations resulting from salt or mineral dissolution would not increase in proportion to chloride but, rather, would increase in proportions determined by the chemical composition of the soluble salt or minerals. If chloride concentration were the result of only evapotranspiration, then 60 to 80 percent of sodium, sulfate, calcium, and magnesium ions and 92 percent of potassium ions would have to be removed from the water to obtain the chemistry observed in the most dilute sodium chloride water (sample from site 10, fig. 7). The sample, however, is below saturation with respect to most minerals that could precipitate, thereby removing constituents from the water, with the exception of calcite, which will be discussed later. Thus, precipitation of most minerals is unlikely.

The third indication that dissolution is the major process producing a sodium chloride water is the isotopic composition of the sample from site 10. Dissolution of evaporative salts would not alter the isotopic composition of the water, whereas evaporation would shift the isotopic composition of the water to a heavier (more negative) value. The sample from site 10 has an isotopic composition similar to the sodium bicarbonate waters (samples from sites 7–9), indicating that the sample has not been significantly evaporated (fig. 14). In addition, the sample from site 10 was obtained from 66 ft below land surface, where evaporation would be improbable. The sample from site 10 is below saturation with respect to albite, so although another source of sodium is not needed to obtain the sodium concentration in the sample, a small amount of sodium is probably added to the water by albite dissolution.

Transpiration also may be an important process for increasing chloride and sodium concentrations as the water evolves from a sodium bicarbonate to a sodium chloride type. Transpiration can concentrate dissolved solids in water to depths greatly exceeding evaporation depths and, thus, could have increased constituent concentrations in water sampled from site 10. Phreatophyte roots will follow fresh water to depths exceeding 60 ft (P.A. Glancy, U.S. Geological Survey, oral commun., 1989) and, unlike evaporation, transpiration has been shown to concentrate dissolved solids in residual water with little or no effect on isotopic composition (Wershaw and others, 1966; Zimmerman and others, 1967, p. 576; Ziegler and others, 1976; Gat and Gonfiantini, 1981, p. 223–238; Szecsody and others, 1983; Koltermann, 1984; White and others, 1985; Turner and others, 1987). Thus, the observed higher constituent concentrations and unchanged isotopic composition are consistent with transpiration removing water and thereby increasing constituent concentrations in the residual water. However, transpiration could account for only a 20 to 40 percent increase in constituent concentrations (or 8 percent for potassium), unless constituents are being removed from the water as they become more concentrated.

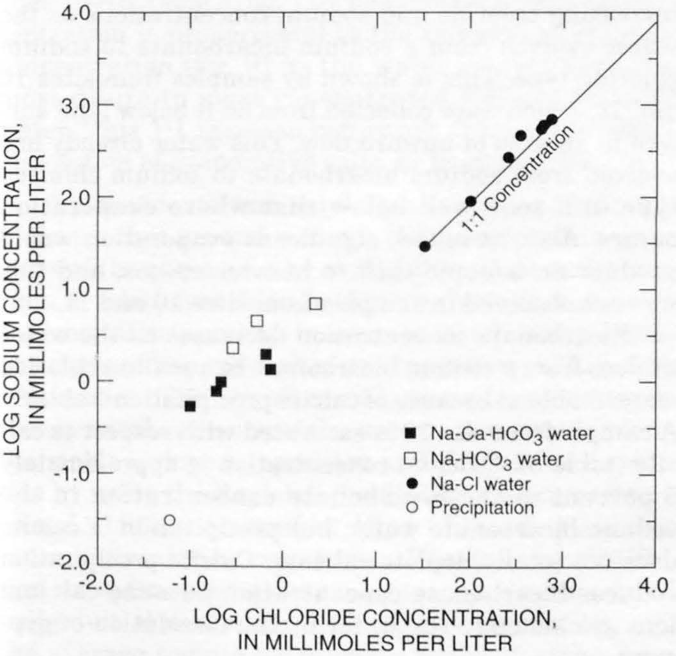

FIGURE 9.—Relation between sodium and chloride concentrations in Smith Creek Valley, Nev.

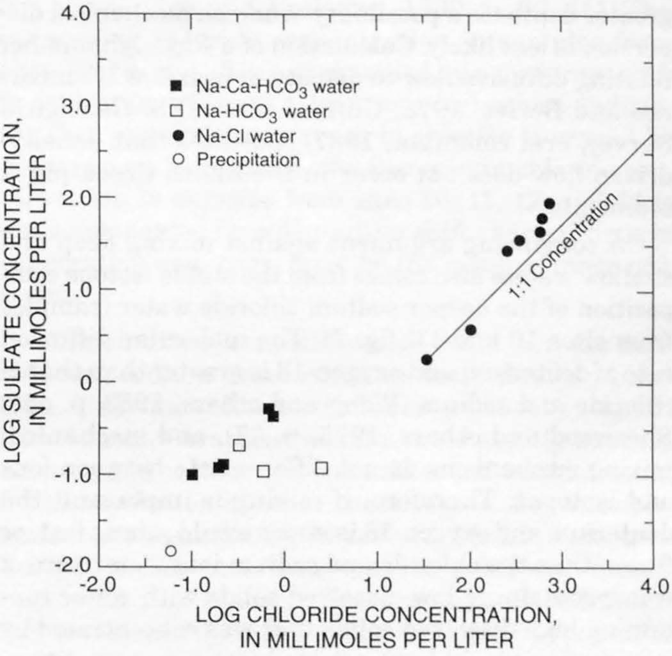

FIGURE 10.—Relation between sulfate and chloride concentrations in Smith Creek Valley, Nev.

Mixing deeper water that contains lower dissolved solids with shallower water that contains higher dissolved solids (whether by molecular diffusion, dispersion, or density-driven flow) is probably not an important process for increasing constituent concentrations as the water evolves from sodium bicarbonate to sodium chloride. The Peclet number calculated for the fine-grained Smith Creek playa sediments is considerably less than one, indicating that molecular diffusion predominates over mechanical dispersion (Bear, 1972). This indicates that molecular diffusion of chloride and sodium ions from shallow briny water to greater depths is a possibility, whereas mechanical dispersion is less likely. Calculation of a Rayleigh number relating diffusive flow to density-driven flow (Combarnos and Bories, 1975; Clifford Voss, U.S. Geological Survey, oral commun., 1987) indicates that density-driven flow does not occur in the Smith Creek playa sediments.

A convincing argument against mixing deep and shallow waters also comes from the stable isotope composition of the deeper sodium chloride water (samples from sites 10 and 11; fig. 7). The molecular diffusion rate of deuterium and oxygen-18 is greater than that of chloride and sodium (Wang and others, 1953, p. 468; Sherwood and others, 1975, p. 37), and mechanical mixing mechanisms do not differentiate between ions and isotopes. Therefore, if mixing is important, the deuterium and oxygen-18 isotopes would mix as fast, or faster, than the chloride and sodium ions. A mixture of water containing low dissolved solids with water containing high dissolved solids that was concentrated by evaporation would have a mixed isotopic composition, as well as intermediate ionic concentrations. However, the deep sodium chloride water has a deuterium and oxygen-18 composition similar to sodium bicarbonate water unaffected by evaporation (fig. 14) and is different from that of the shallow sodium chloride water, which has been affected by evaporation. Therefore, mixing (by whatever mechanism) of isotopically heavy (less negative) water containing high chloride and sodium concentrations is not indicated at depth (66 ft).

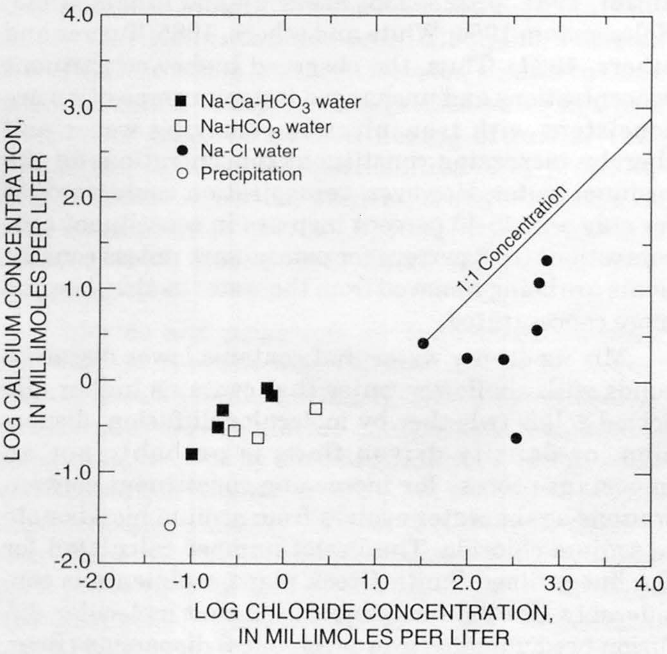

FIGURE 11.—Relation between calcium and chloride concentrations in Smith Creek Valley, Nev.

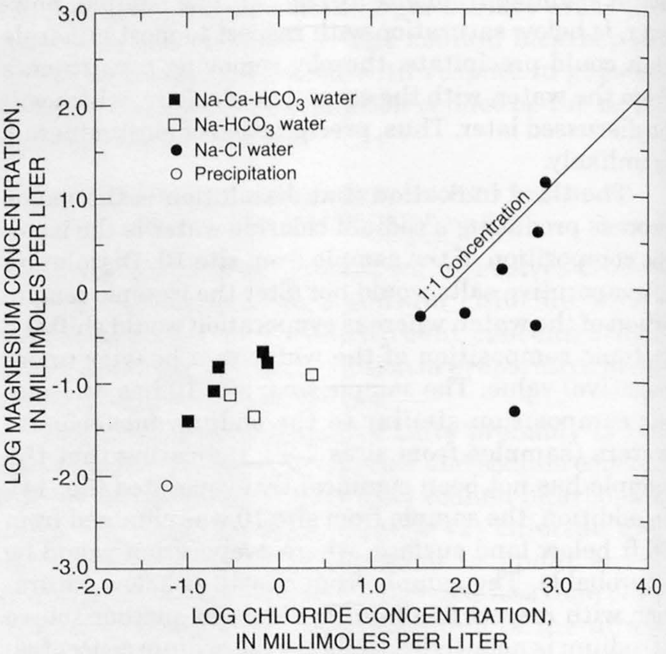

FIGURE 12.—Relation between magnesium and chloride concentrations in Smith Creek Valley, Nev.

Ground water may mix in the shallower part of the aquifer beneath the playa or phreatophytic area surrounding the playa, but not within the approximately 44-ft vertical interval between the shallow and deep wells. In the deep wells, the isotopic composition of the water shows little effect of evaporation.

Evaporation is not an important process for increasing chloride and sodium concentrations as the water evolves from a sodium bicarbonate to sodium chloride type. This is shown by samples from sites 10 and 11, which were collected from 66 ft below land surface in an area of upward flow. This water already has evolved from sodium bicarbonate to sodium chloride type in a zone well below that where evaporation occurs. Also, as noted, significant evaporation would produce an isotopic shift to heavier values, and this was not observed in samples from sites 10 and 11.

Bicarbonate concentration decreases as the water evolves from a sodium bicarbonate to a sodium chloride water (table 4) because of calcite precipitation (table 6). A sample from site 10 is saturated with respect to calcite (table 5). Calcium concentration is approximately 5 percent of the bicarbonate concentration in the sodium bicarbonate water, but precipitation of calcite does not totally deplete calcium. Calcite precipitation reduces bicarbonate concentration because calcium ions are added to the water by the dissolution of gypsum:

$$CaSO_4 \cdot 2H_2O + CO_3^{2-} = CaCO_3 + SO_4^{2-} + 2H_2O . \quad (1)$$

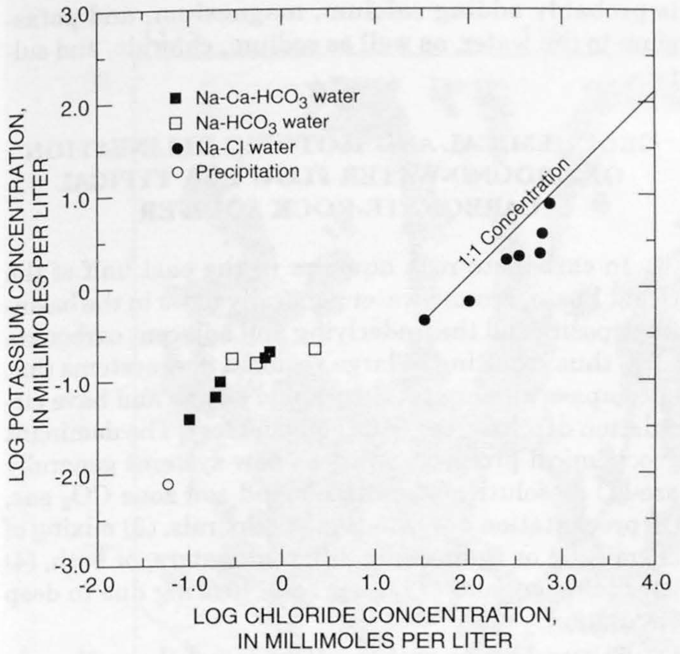

FIGURE 13.—Relation between potassium and chloride concentrations in Smith Creek Valley, Nev.

EVOLUTION OF DILUTE TO CONCENTRATED SODIUM CHLORIDE WATER

Ground water in the playa area evolves from a slightly saline (measured dissolved-solids concentration, 2,300 mg/L) to a briny (measured dissolved solids, 51,000 mg/L) sodium chloride water (samples from sites 10–16; table 3, fig. 7). The increase in sodium concentration is proportional to the increase in chloride concentration (fig. 9) as the water evolves from the most dilute to most concentrated sodium chloride water. This 1:1 increase can result from evaporation, dissolution of evaporative salts, or transpiration.

A plot of deuterium in relation to oxygen-18 (fig. 14) indicates that ground water is being removed in some parts of the playa area by evaporation. A plot of deuterium in relation to chloride (fig. 15) shows that chloride concentration increases in areas in which isotopic compositions of ground water are largely unaffected by evaporation. The overall contribution of evaporation to the chemical evolution of the water was estimated from the stable isotope composition and chloride concentration by using the Rayleigh distillation equation (Dansgaard, 1964), assuming that chloride remains conservative during evaporation. The Rayleigh distillation curve for evaporative concentration was calculated using initial deuterium and chloride compositions of the average sodium bicarbonate water (samples from sites 7–9, table 3) and a fractionation factor (1.0079) calculated for a basin-fill aquifer in northwestern Nevada (Welch and Preissler, 1990, p. 32–41). The increase in chloride concentration in samples from sites 13, 14, and 15 is accompanied by a moderate shift in deuterium to heavier (less negative) values, indicating that some of the increase in chloride is caused by evaporation. In contrast, the increase in chloride concentration in samples from sites 10, 11, 12, and 16 is not accompanied by a deuterium shift characteristic of evaporation and, thus, must be the result of other processes.

Dissolution of evaporative salt could increase chloride and sodium concentrations more than evaporation. Salt dissolution induces no isotope shift. Dissolution of evaporative salts could greatly increase chloride and sodium concentrations and would account for the evolution of other constituents. For example, sulfate concentration in the sodium chloride water generally increases at a rate greater than 1:1 with respect to chloride (fig. 10). This increase requires addition of sulfate, most likely from dissolution of gypsum, in addition to concentration of sulfate by evapotranspiration. If gypsum is dissolving, other salts probably also are dissolving.

Transpiration is an important hydrologic process in Smith Creek Valley that also may explain the increased constituent concentrations. Approximately 70 percent of ground water discharged from Smith Creek Valley is transpired from the phreatophytic area surrounding the playa (Thomas, Carlton, and Hines, 1989).

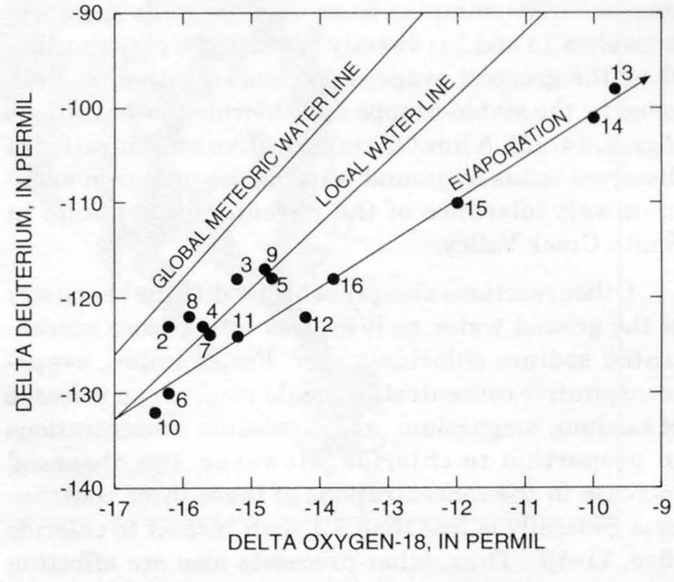

FIGURE 14.—Relation between deuterium and oxygen-18 in Smith Creek Valley, Nev. The global meteoric water line [$\delta D=8(\delta^{18}O)+10$] is based on worldwide precipitation data (Craig, 1961). The local water line [$\delta D=6.82(\delta^{18}O)-16.8$] is based on a linear regression of samples from sites 2 to 11 (see figs. 2, 7; table 3).

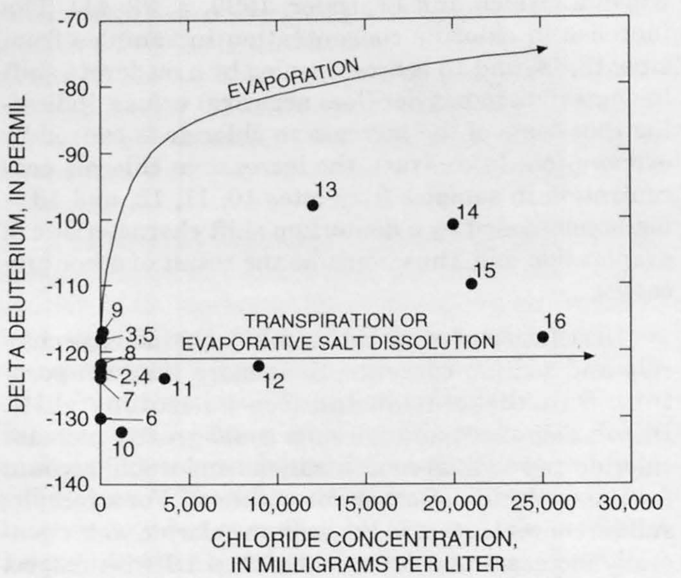

FIGURE 15.—Relation between deuterium and chloride concentrations in samples from wells in Smith Creek Valley, Nev. Numbered circles correspond to sites in figures 2 and 7 and table 3.

Transpiration results in concentration of ions but induces no isotopic shift, so it is also a simple and reasonable explanation for concentration of constituents in the sodium chloride water. Sodium chloride water from shallow wells (22 ft deep) in or along the phreatophytic area that surrounds the playa (samples from sites 12, 15, and 16) show the least effect of evaporation, whereas samples from shallow wells (samples from sites 13 and 14) directly beneath the playa surface show the greatest evaporative concentration, as indicated by the stable-isotope and chloride concentrations (figs. 7, 14, 15). A limit to transpirative concentration of dissolved solids in ground water is the unknown maximum salt tolerance of the phreatophytic plants in Smith Creek Valley.

Other reactions also probably modify the chemistry of the ground water as it evolves into a more concentrated sodium chloride water. For example, evapotranspirative concentration would result in an increase of calcium, magnesium, and potassium concentrations in proportion to chloride. However, the observed increase in the concentrations of these three constituents generally is less than 1:1 with respect to chloride (figs. 11–13). Thus, other processes also are affecting calcium, magnesium, and potassium concentrations. These constituents could be removed from the water by precipitation of calcite and zeolite minerals, exchange on clay minerals, and the formation of clay minerals. In addition, dissolution of minerals and evaporative salts is probably adding calcium, magnesium, and potassium to the water, as well as sodium, chloride, and sulfate.

GEOCHEMICAL AND ISOTOPIC DELINEATION OF GROUND-WATER FLOW IN A TYPICAL CARBONATE-ROCK AQUIFER

In carbonate-rock aquifers in the east half of the Great Basin, ground water generally flows in the basin-fill deposits and the underlying and adjacent carbonate rock, thus resulting in large regional flow systems that encompass numerous topographic basins and have circulation depths of several thousand feet. The dominant geochemical processes in these flow systems generally are (1) dissolution of minerals and soil zone CO_2 gas, (2) precipitation or formation of minerals, (3) mixing of chemically or isotopically different waters, or both, (4) ion exchange, and (5) geothermal heating due to deep circulation.

The carbonate-rock aquifers in southern Nevada are an example of these carbonate-rock flow systems (fig. 16). The aquifers include two regional flow systems that discharge in southern Nevada: the White River flow system, which discharges at Muddy River springs (fig. 16) and is a subsystem of the Colorado flow system (fig. 1), and the Ash Meadows flow system, which discharges at Ash Meadows springs (fig. 16) and is a subsystem of the Death Valley flow system (fig. 1; Eakin, 1966; Mifflin, 1968; Winograd and Friedman, 1972; Winograd and Thordarson, 1975; Hess and Mifflin, 1978; Thomas and others, 1986; Harrill and others, 1988; Dettinger, 1989; Kirk and Campana, 1990). In the north-central part of the study area, ground water also flows from the White River system to the Ash Meadows system (figs. 16, 17). These flow systems encompass thousands of square miles and include several valleys. Recharge to these regional systems is from several sources, and most of the flow discharges from a common area that contains numerous springs. Smaller carbonate-rock flow systems, such as the Las Vegas Valley flow system, also are present in the study area. In these smaller systems, recharge is primarily from one source and water may discharge from the aquifers in several places.

Geochemical processes that produce the chemical and isotopic compositions of the water in carbonate-rock aquifers were identified using stable and radioactive isotopes, major-ion chemistry, mass-balance calculations, thermodynamic calculations, and mineral identification. The isotopic and chemical compositions of water from different areas were used further to delineate regional flow systems. The primary processes that produce the isotopic and chemical content of the

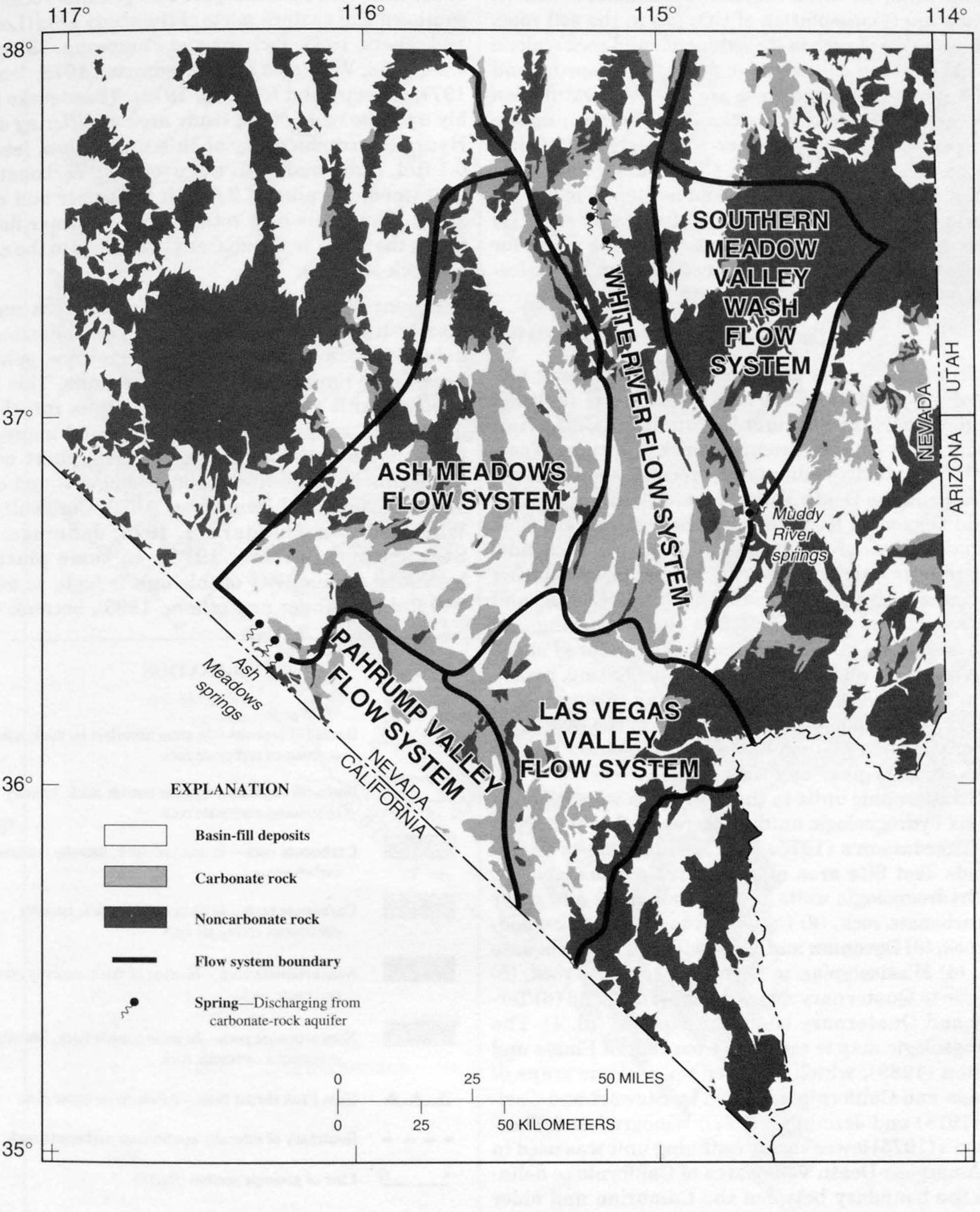

FIGURE 16.—Map showing ground-water flow systems in southern Nevada, as delineated in this study.

ground water in these regional systems of southern Nevada are (1) dissolution of CO_2 gas in the soil zone, (2) dissolution of carbonate sediments and rock (calcite and dolomite) in the recharge areas (most spring and well waters in recharge areas are at or near saturation with respect to calcite and dolomite), (3) precipitation of calcite, (4) leakage of water with high sulfate and sodium concentrations from the Tertiary confining unit, or dissolution of gypsum and halite, (5) mixing of different waters, (6) dissolution of volcanic rock, (7) exchange of calcium and magnesium in the water for sodium in clay minerals, (8) precipitation of chalcedony, and (9) formation of kaolinite

GEOLOGIC FRAMEWORK

Southern Nevada has had a complex geologic history of sedimentation, volcanic activity, and faulting. Miogeosynclinal sedimentation from Cambrian through Permian time produced thick sequences of carbonate and clastic sedimentary rock in southeastern Nevada and the Death Valley area of California. Widespread volcanism from Tertiary through present time produced extensive volumes of volcanic rock now exposed over much of the area. Major thrust faulting associated with the Sevier orogeny (late Mesozoic) and Cenozoic low-angle, near-surface thrust faulting and right-lateral strike-slip faulting have produced shear zones and high-angle basin-and-range faulting, resulting in the topography observed today (Stewart and Carlson, 1978). A detailed description of the hydrogeology of the entire Great Basin, as delineated by this RASA study, is given by Plume (1996).

Stratigraphic units in the study area were grouped into six hydrogeologic units to correspond to Winograd and Thordarson's (1975) hydrogeologic units in the Nevada Test Site area of southern Nevada (table 7). The hydrogeologic units are (1) Cambrian and older noncarbonate rock, (2) Cambrian to Devonian carbonate rock, (3) Devonian and Mississippian noncarbonate rock, (4) Mississippian to Permian carbonate rock, (5) Triassic to Quaternary noncarbonate rock, and (6) Tertiary and Quaternary basin-fill deposits (pl. 1). The hydrogeologic map is modified from that of Plume and Carlton (1988), which is based on geologic maps of Nevada and California compiled by Stewart and Carlson (1978) and Jennings (1977). Winograd and Thordarson's (1975) lower clastic confining unit was used in the Amargosa-Death Valley area of California to delineate the boundary between the Cambrian and older noncarbonate rock and the Cambrian to Devonian carbonate rock.

The Cambrian and older noncarbonate rock includes primarily siltstone, quartzite, shale, and sandstone in the northern and western parts of the study area and metamorphic and granitic rocks in the southern and eastern parts of the study area (Longwell and others, 1965; Tschanz and Pampeyan, 1970; Cornwall, 1972; Winograd and Thordarson, 1975; Jennings, 1977; Stewart and Carlson, 1978). These rocks probably underlie most of the study area at differing depths. Hydraulic conductivity of this unit is low, less than 0.1 ft/d, compared with the overlying carbonate-rock unit (median value of 2.0 ft/d: Dettinger and others, 1995). Thus, this unit retards ground-water flow and forms the base, or an adjacent boundary, to the carbonate-rock aquifers.

Cambrian to Devonian carbonate rock includes mostly limestone and dolomite. The limestone and dolomite formations contain numerous, generally basal, silty limestone and dolomite units. This hydrogeologic unit also contains numerous interbedded quartzites, sandstones, and shales, and many of the limestone and dolomite beds contain chert nodules (Langenheim and others, 1962; Longwell and others, 1965; Tschanz and Pampeyan, 1970; Cornwall, 1972; Winograd and Thordarson, 1975; Jennings, 1977; Stewart and Carlson, 1978). In some places the hydraulic conductivity of this unit is high, as much as 940 ft/d (Dettinger and others, 1995), because of sec-

EXPLANATION

□ **Basin-fill deposits**—In areas underlain by thick, laterally continuous carbonate rock

▨ **Basin-fill deposits**—In areas outside thick, laterally continuous carbonate rock

▒ **Carbonate rock**—In areas of thick, laterally continuous carbonate rock

▨ **Carbonate rock**—In areas outside thick, laterally continuous carbonate rock

■ **Noncarbonate rock**—In areas of thick, laterally continuous carbonate rock

▨ **Noncarbonate rock**—In areas outside thick, laterally continuous carbonate rock

▲▲▲ **Gass Peak thrust fault**—Sawteeth on upper plate

- - - - **Boundary of laterally continuous carbonate rock**

A —— A′ **Line of geologic section (fig.18)**

➡ **Generalized direction of ground-water movement in carbonate-rock aquifers**

● **Spring**—Emanates from carbonate-rock aquifers

● **Well**—Source of water is carbonate-rock aquifer

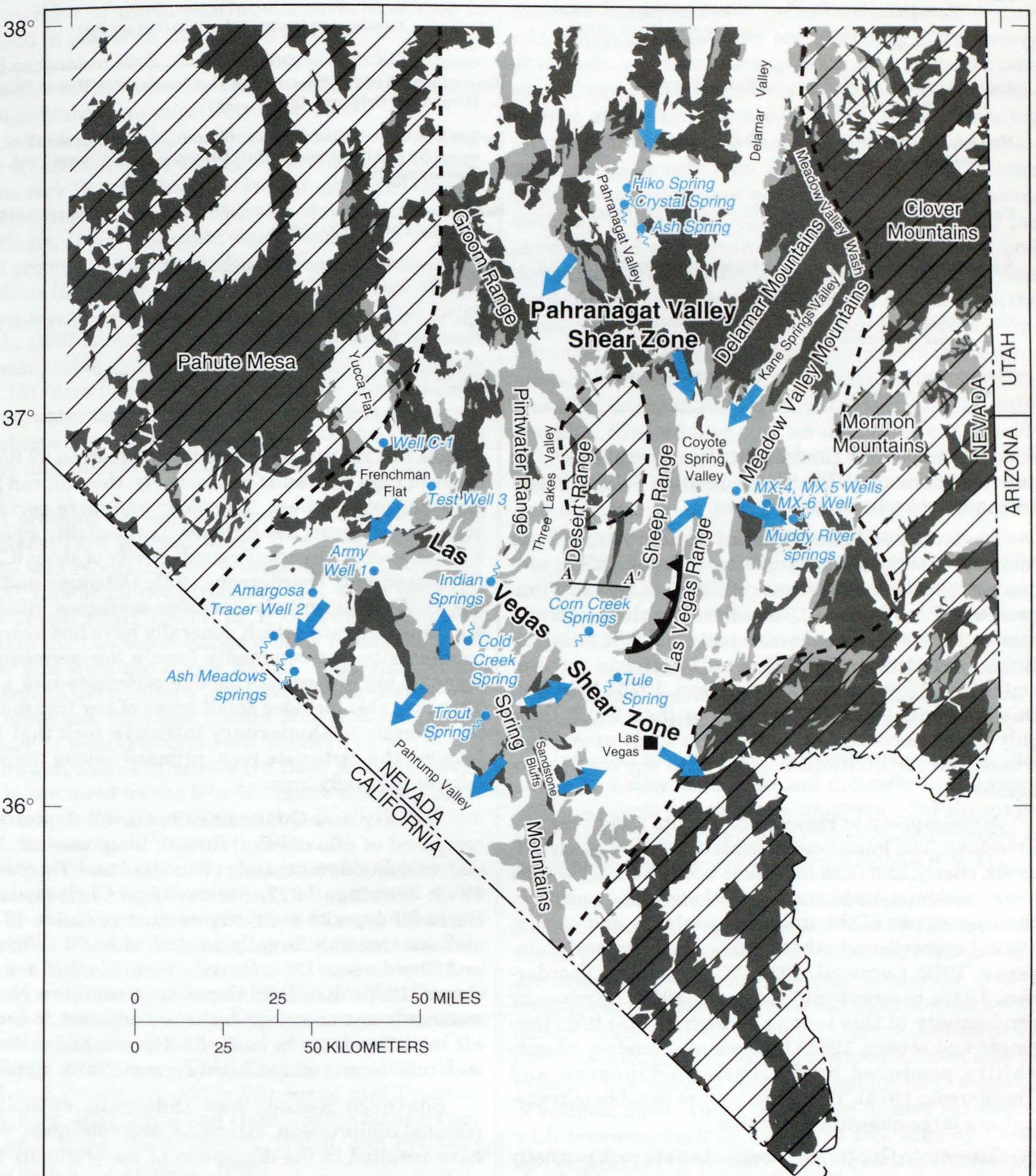

FIGURE 17.—Ground-water flow directions and hydrogeologic features associated with carbonate-rock aquifers in southern Nevada. Modified from Dettinger (1989, figs. 3, 4).

TABLE 7.—*Hydrostratigraphy in southern Nevada and southeastern California*

Hydrogeologic unit and geologic age (see plate 1)	Hydrogeologic unit names used by Winograd and Thordarson (1975)	Calculated hydraulic conductivity (feet per day)
Basin-fill deposits, Quaternary and Tertiary	Valley-fill aquifer	Ranges from 0.6 to 17 (Winograd and Thordarson, 1975; Harrill, 1986; Morgan and Dettinger, 1996).
Noncarbonate rock, Quaternary to Triassic	Tuff aquitards and aquifers	Ranges from 0.1 to 92, but most aquifers are generally less than 1.0, except for highly permeable Piapi Canyon Group (Winograd and Thordarson, 1975).
Carbonate rock, Permian to Mississippian	Upper carbonate aquifer	Ranges from 0.1 to 900, with mean value of 102 and median value of 5.2 (Dettinger and others, 1995).
Noncarbonate rock, Mississippian and Devonian	Upper clastic aquitard	0.01 for two wells (Dettinger and others, 1995).
Carbonate rock, Devonian to Cambrian	Lower carbonate aquifer	Ranges from 0.01 to 940, with mean value of 96 and median value of 2.0 (Dettinger and others, 1995).
Noncarbonate rock, Cambrian and older	Lower clastic aquitard	Less than 0.1 for two wells (Dettinger and others, 1995).

ondary permeability produced by fracturing (Winograd and Thordarson, 1975). Consequently, this unit can transmit vast quantities of water.

Devonian to Mississippian noncarbonate rock includes mostly shale, argillite, siliceous siltstone, quartzite, and conglomerate (Cornwall, 1972; Winograd and Thordarson, 1975). Because the hydraulic conductivity of this unit is extremely low, 0.01 ft/d (Dettinger and others, 1995) this unit restricts groundwater flow (Winograd and Thordarson, 1975). However, this low-permeability unit is greater than 200 ft thick only in the north-central part of the study area. Elsewhere, the carbonate-rock units form one continuous aquifer.

Mississippian to Permian carbonate rock is mainly limestone. The limestone consists of numerous argillaceous, cherty, and fetid limestone beds and interbedded shale, siltstone, and sandstone, along with gypsum in the upper part of the unit (Langenheim and others, 1962; Longwell and others, 1965; Tschanz and Pampeyan, 1970; Cornwall, 1972; Winograd and Thordarson, 1975; Stewart and Carlson, 1978). Hydraulic conductivity of this unit is as high as 900 ft/d (Dettinger and others, 1995) because of secondary permeability produced by fracturing (Winograd and Thordarson, 1975). Thus, this unit is capable of transmitting large quantities of water.

Triassic to Quaternary noncarbonate rock is mostly volcanic rock consisting of welded to nonwelded ash-flow and ash-fall tuffs, and basalt and rhyolite flows (Longwell and others, 1965; Tschanz and Pampeyan, 1970; Cornwall, 1972; Winograd and Thordarson, 1975; Jennings, 1977; Stewart and Carlson, 1978). Although the hydraulic conductivity of this unit generally ranges from 0.1 to 2.0 ft/d, in places it is as high as 92 ft/d. In areas of high hydraulic conductivity, the unit can form transmissive aquifers. The older and more pervasive volcanic rock generally form low-permeability aquifers, with an average hydraulic conductivity of about 0.2 ft/d (Winograd and Thordarson, 1975; Dettinger and others, 1995). This unit also includes sandstone, siltstone, and conglomerate, which generally have low hydraulic conductivity and commonly form a low-permeability layer on top of more permeable carbonate-rock units. This unit also includes small areas of low transmissivity, Triassic to Quaternary intrusive rock that have invaded the carbonate rock, primarily along the south edge of the study area.

Tertiary and Quaternary basin-fill deposits are composed of alluvial-fan, fluvial, fanglomerate, lakebed, and mudflow deposits (Winograd and Thordarson, 1975; Jennings 1977; Stewart and Carlson, 1978). Basin-fill deposits are fairly conductive (0.6 to 17 ft/d) and can transmit large quantities of water (Winograd and Thordarson, 1975; Harrill, 1986; Morgan and Dettinger, 1996). Basin-fill deposits in southern Nevada commonly are underlain by or are adjacent to carbonate rock. Together the basin-fill deposits and carbonate rock may form a single, highly transmissive aquifer.

Southern Nevada has undergone episodes of regional compression, extension, and volcanism, which have resulted in the disruption of the originally thick and continuous carbonate-rock strata deposited during the Paleozoic. Primarily because of extension over the last 25 million years, only a central corridor of thick, laterally continuous Paleozoic carbonate rock remains in southern Nevada (fig. 17; Dettinger, 1989). This central corridor contains most of the regional ground-water flow in southern Nevada, including both the White River and Ash Meadows flow systems (fig. 16). The corridor of thick, continuous carbonate rock is bounded in places on the east and west by thin or isolated carbonate rock, or noncarbonate rock. Groundwater flow in the carbonate rock outside this corridor is not significant in volume (Dettinger, 1989).

Geologic constraints on flow within the central corridor of carbonate rock are caused primarily by noncarbonate rock that has been rotated or emplaced close to the land surface. One postulated geologic constraint (Dettinger, 1989) that could significantly impede westward ground-water flow from the Sheep Range to Ash Meadows (fig. 17) is a low permeability wedge of noncarbonate clastic rock several thousand feet thick (Guth, 1980) extending a minimum of 1,000 ft above the water table along the west side of the Sheep Range (fig. 18). Another postulated geologic constraint to ground-water flow to the west from the Sheep Range is the absence of carbonate rock between Three Lakes Valley and the northern Sheep Range (fig. 17; Guth, 1989).

A probable geologic constraint on southward and southeastward ground-water flow from the Sheep Range was recognized by Winograd and Thordarson (1975). Cambrian noncarbonate rock along the Gass Peak Thrust Fault (fig. 17, pl. 1; Longwell and others, 1965) in the Las Vegas Range greatly reduces groundwater flow from the Sheep Range. Thus, most of the recharge to the southern part of the Sheep Range may flow north and then east toward Muddy River springs because of noncarbonate-rock barriers to westward, southward, and southeastward flow. Also, eastward flow is enhanced by the 5 to 50 degree eastward dip of the strata in the Sheep Range (Longwell and others, 1965; Tschanz and Pampeyan, 1970).

Major structural zones that intersect the central carbonate-rock corridor also may alter ground-water flow by impeding or enhancing flow in some areas and directions. The Las Vegas Valley shear zone and Pahranagat Valley shear zone are two major structural zones in the central corridor that may influence ground-water flow by reducing flow across the zones and enhancing (or allowing) flow along the strike of the zones (fig. 17; Winograd and Thordarson, 1968; Winograd and Friedman, 1972; Winograd and Pearson, 1976; Kirk, 1987; Lyles and Hess, 1988; Thomas, Carlton, and Hines, 1989; Thomas, Welch, and Preissler, 1989; Kirk and Campana, 1990).

Aquifers in the study area from which regional springs emanate are primarily Cambrian to Devonian dolomite and limestone and Devonian through Permian limestone. X-ray diffraction analyses of rocks of Precambrian to Permian age show that the mineralogy of the upper and lower carbonate-rock aquifers is predominantly dolomite and calcite (table 8). Whole-rock major-element analyses indicate that the dolomites and limestones generally are composed of extremely pure carbonates containing less than 1.5 weight percent of major element oxides other than calcium carbonate and magnesium carbonate (table 9). Sulfur analyses (table 10) also show the limestones and dolomites are extremely pure. Samples used for whole-rock major-element analyses were limestones and dolomites with few fracture fillings or notable impurities; the chemistry of more veined or shaley limestones and dolomites may differ. X-ray diffraction analyses (table 8) and analyses of the insoluble residue from the whole-rock chemical analyses (table 11) indicate that minerals other than calcite and dolomite in the carbonate rock are mostly quartz with lesser amounts of illite, smectite, feldspar, alunogen, chlorite, kaolinite, palygorskite, clinoptilolite, and unidentified zeolite (table 11). Fracture-filling vein minerals are primarily calcite and dolomite but also include quartz, aragonite, iron and manganese oxides, and clays (table 8; Winograd and Thordarson, 1975). These vein minerals are primarily precipitated from the ground water, except for clays which are weathering products. Although the carbonate rock may contain inclusions of connate seawater, the low sodium and sulfur concentrations (tables 9, 10) indicate they are not common. Because surface samples may have undergone leaching, subsurface rocks are more likely to contain inclusions.

Interbedded quartzite, sandstone, siltstone, and shale, and silty limestone and dolomite in the lower and upper carbonate-rock aquifers contain major amounts of quartz and alumino-silicate minerals in addition to the carbonate minerals (tables 8, 9, 11).

X-ray diffraction analyses of Miocene volcanic rock from the Kane Springs Wash and Meadow Valley Wash areas (pl. 1) show the dominant minerals in these rocks are primarily plagioclase feldspar, K-feldspar, quartz, biotite, and clinoptilolite (table 12). Hornblende, halite, clinopyroxene, olivine, and mica also are present in the rock. Calcite, mordenite, and cristobalite also may be present. The volcanic rock also contain large amounts of volcanic glass. These results are in good agreement with previous work on rock from this area by Ekren and others (1977). Tertiary volcanic rock in the Nevada Test Site area have similar mineralogies (Byers and others, 1976) and have undergone zeolitization, resulting in a downward zonation of clinoptilolite to mordenite to analcime, that cuts across volcanic-rock units (Hoover, 1968).

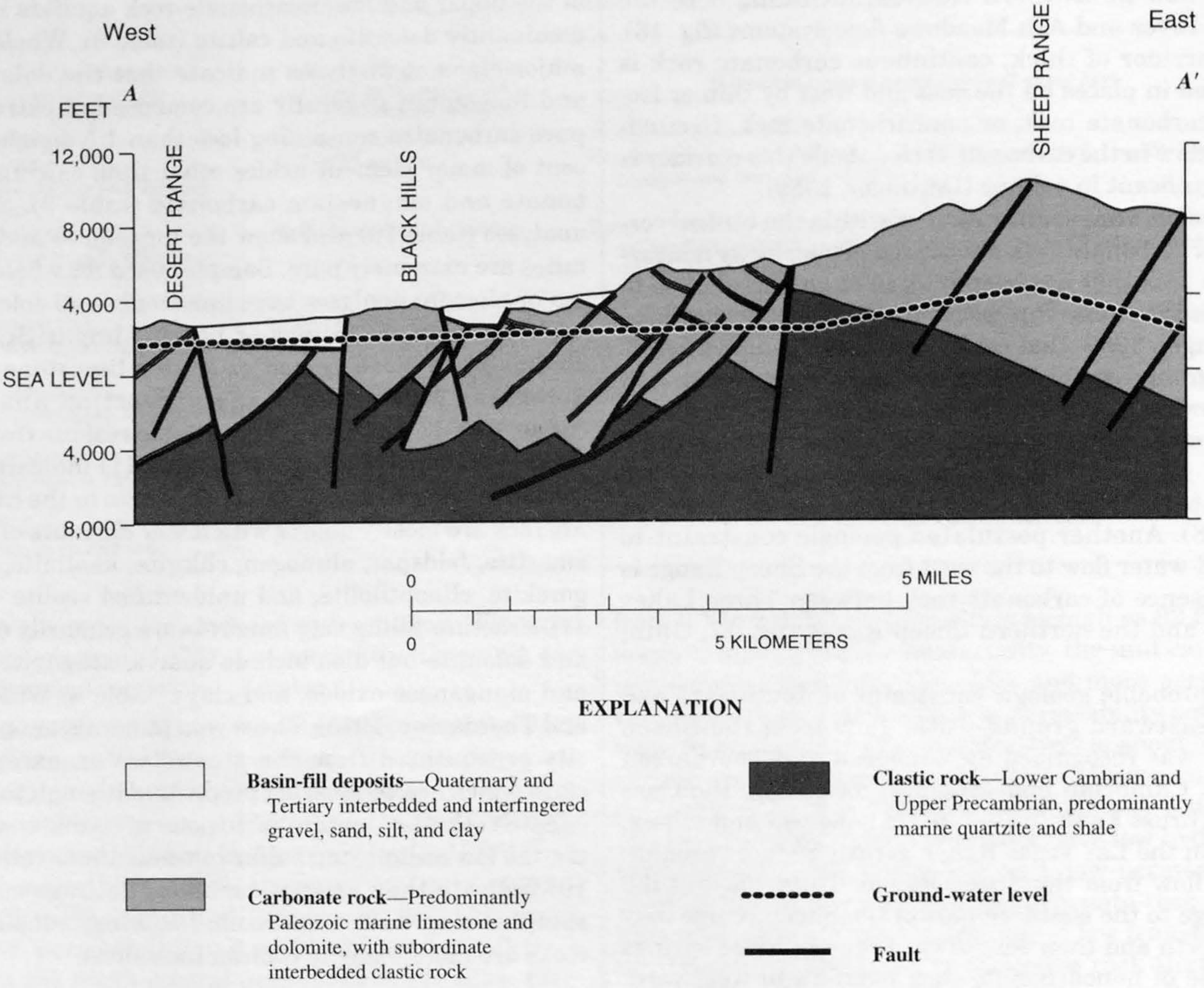

FIGURE 18.—Geologic section across southern Sheep Range north of Las Vegas, Nev. Line of section shown in figure 17. Modified from Guth (1980, pl. 2) and Dwight Schmidt (U.S. Geological Survey, written commun., 1986).

Evaporite minerals, primarily gypsum, are present in Permian and younger rock throughout the study area. Other minerals associated with these evaporite deposits are halite, magnesite, and glauberite (Longwell and others, 1965).

HYDROLOGIC FRAMEWORK

Water in carbonate-rock aquifers of southern Nevada primarily originates from high-altitude winter precipitation (Winograd and Riggs, 1984), mainly in the Spring Mountains and Sheep Range (fig. 17), and from ground water flowing into southern Nevada by way of the White River flow system to the north (fig. 16). Water recharged at high altitudes generally infiltrates carbonate rock and flows through carbonate-rock aquifers to a sink, such as the springs at Ash Meadows, where the water is discharged. Water flowing through the aquifers often mixes with water from different source areas before reaching the common discharge area of the flow system. Ground water flowing into southern Nevada from other flow systems, such as the White River flow system to the north, mixes with water that has been recharged in southern Nevada and is part of the water discharging at regional springs.

Water levels of springs emanating from the carbonate-rock aquifers in wells completed in carbonate rock and in wells completed in basin-fill deposits (where springs emanating from, or wells completed in, carbonate rock are absent) have been used to help delineate flow systems in the carbonate rock of southern Nevada (pl. 1; Eakin, 1966; Winograd and Thordarson, 1975; Thomas and others, 1986). Water levels indicate that Muddy River springs at the terminus of the White River flow system receives water from the Pahranagat Valley area to the northwest as originally delineated by Eakin (1966). In addition, Muddy River springs also

TABLE 8.—*X-ray diffraction analyses of rock samples from southern Nevada*

[Analysis by Robert Mariner, U.S. Geological Survey, Menlo Park, Calif. Symbol: --, no mineral identified]

Site[1]	Geologic unit, vein, or rock type sampled	Geologic age	Minerals identified			
			Major[2]	Minor	Trace[2]	Possible
1	Eureka Quartzite	Ordovician	quartz	--	--	--
2	Bonanza King Formation	Cambrian	dolomite	--	quartz, aragonite	--
3	Nopah Formation	Cambrian	dolomite	--	calcite	--
4	Undifferentiated	Silurian	dolomite	--	calcite	--
5	Toroweap Formation	Permian	calcite, quartz	--	feldspar	--
6	Gneiss	Precambrian	quartz, plagioclase, biotite, chlorite, dolomite	--	--	--
	Schist		biotite, plagioclase, quartz	--	amphibole	--
7	Tapeats Sandstone	Cambrian	quartz	--	calcite	--
8	Pioche Formation	Cambrian	quartz	K-feldspar	--	--
9	Lyndon Limestone	Cambrian	calcite	--	--	--
10	Limestone and dolomite	Cambrian	dolomite	calcite	quartz	--
11	Muddy Peak Formation	Devonian	dolomite	quartz	calcite	--
12	Callville Limestone	Mississippian-Permian	calcite	quartz	--	--
13	Undifferentiated vein	Devonian	dolomite	--	calcite	--
14	Pogonip Group	Ordovician	dolomite	--	--	calcite
15	Lone Mountain Dolomite	Silurian	dolomite, quartz	--	calcite, feldspar (microcline)	sphene
16	Pogonip Group (lower part)	Ordovician	calcite	--	quartz	--
	Vein in Pogonip Group		calcite, quartz, dolomite, aragonite	--	--	
17	Upper unnamed unit	Cambrian	dolomite	--	calcite	--
	Vein in upper, unnamed unit		calcite		quartz	--
18	Sevy Dolomite	Devonian	dolomite	--	--	--
19	Bird Spring Formation	Mississippian-Permian	dolomite, quartz, calcite	--	--	--
20	Pogonip Group (Antelope Valley Formation)	Ordovician	dolomite	--	--	--
21	Undifferentiated Dolomite	Cambrian	dolomite	--	--	--
22	Monte Cristo Formation	Mississippian	calcite, dolomite	--	quartz	--
	Vein in Dawn Limestone		calcite	--	quartz	--

[1] See plate 1.

[2] Listed in order of approximate abundance.

TABLE 9.—*Whole-rock chemical analyses of rock samples from southern Nevada*

[Analyses by Nelson Shaffer, Indiana Geological Survey, Bloomington, Ind., except as noted. Analyses are recalculated to total 100.0 percent. Rock analyses are expressed in weight percent of constituent oxides or, for calcium and magnesium, carbonates. Symbol: --, not determined.]

Site[1]	Geologic unit or rock type sampled	SiO_2	Al_2O_3	Fe_2O_3	TiO_2	MnO	$CaCO_3$	$MgCO_3$	Na_2O	K_2O	P_2O_5	SrO	Total
E-1	Nopah Dolomite	1.05	0.12	0.061	0.009	0.010	53.9	44.8	0.011	0.050	0.016	0.006	100.0
E-2	Ely Springs Dolomite	.25	.070	.057	.008	.010	54.8	44.7	.018	.067	.011	.017	100.0
E-3	Yellowpine Limestone	1.21	.12	.057	.006	.008	97.7	.79	.007	.036	.032	.019	100.0
E-5	Monte Cristo Limestone, Dawn Member[2]	.30	.20	.10	--	--	56.4	43.0	--	--	--	--	100.0
E-6	Sultan Limestone[2]	.60	.40	.05	--	--	98.1	.90	--	--	--	--	100.0
E-7	Carrara Formation Shale	70.3	13.5	5.30	.82	.10	2.07	2.99	1.64	3.17	.13	.005	100.0
E-8	Bright Angel Shale	57.4	22.1	7.97	.92	.039	.87	4.17	.59	5.75	.24	.012	100.0
E-9	Indian Springs Shale	70.1	18.8	2.94	1.25	.092	3.10	1.46	.47	1.48	.18	.12	100.0

[1] See plate 1.

[2] Analysis from Longwell and others (1965, p. 157).

receives water from the southern part of the Meadow Valley Wash flow system and, due to structural controls (see previous section titled "Geologic Framework"), receives most of the recharge to the Sheep Range (figs. 16, 17, pl. 1).

The Ash Meadows flow system drains the Frenchman Flat-Yucca Flat area to the northeast of the Ash Meadows springs, and on the basis of water-level, isotopic, and geochemical data the Pahranagat Valley area also supplies water to Ash Meadows springs (Winograd and Friedman, 1972; Winograd and Thordarson, 1975; Winograd and Pearson, 1976). The scarcity of wells between Frenchman Flat and Pahranagat Valley and Frenchman Flat and Sheep Range limits interpretation of ground-water flow in this area. However, geologic information (see section titled "Geologic Framework") indicates that the Pahranagat Valley area is connected to Ash Meadows by a thick and continuous carbonate-rock section, whereas the Sheep Range probably does not supply much water to the Ash Meadows flow system (figs. 16, 17, 18).

Water levels in Las Vegas Valley indicate that the valley receives water from the Spring Mountains and Sheep Range (pl. 1). However, structural controls probably prevent significant amounts of recharge to the Sheep Range from flowing into Las Vegas Valley (figs. 17, 18). This conclusion is supported by ground-water flow models of the basin-fill aquifers in Las Vegas Valley by Harrill (1976) and by Morgan and Dettinger (1996). Water levels also indicate that ground water flows into southwest Las Vegas Valley from Ivanpah Valley (pl. 1).

Water levels in Pahrump Valley indicate that recharge from the Spring Mountains is the source of all the water in Pahrump Valley (pl. 1). Water levels for areas outside the thick carbonate-rock aquifers in southern Nevada and southeastern California also are shown on plate 1.

TABLE 10.—*Sulfur content of rock samples from southern Nevada*

[Analyses by Nelson Shaffer, Indiana Geological Survey, Bloomington, Ind.]

Site[1]	Geologic unit sampled	Sulfur, as S	
		(parts per million)	(percent)
E-1	Nopah Dolomite	22	0.0022
E-2	Ely Springs Dolomite	82	.0082
E-3	Yellowpine Limestone	72	.0072
E-4	Dawn Limestone	100	.0100
E-7	Carrara Formation Shale	29	.0029
E-8	Bright Angel Shale	94	.0094
E-9	Indian Springs Shale	130	.0130

[1] See plate 1.

Hydrologic data for southern Nevada indicate that on a regional scale the White River flow system, which discharges at Muddy River springs, and the Ash Mead-

TABLE 11.—*X-ray diffraction analyses of insoluble residues from whole-rock chemical analyses of rock samples from southern Nevada*

[Analyses by Nelson Shaffer, Indiana Geological Survey, Bloomington, Ind.]

Site[1]	Geologic unit sampled	Insoluble residue (percent)	Minerals identified[2]
E-1	Nopah Dolomite	1.33	Quartz, illite, smectite, feldspar, and alunogen.
E-2	Ely Springs Dolomite	.33	Quartz, illite, chlorite, kaolinite, palygorskite, and possibly K-feldspar.
E-3	Yellowpine Limestone	1.34	Quartz.
E-4	Dawn Limestone	.28	Quartz, clinoptilolite, illite, and an unidentified zeolite.
E-7	Carrara Formation Shale	93.12	Quartz, illite, feldspar, smectite, kaolinite, and possibly zeolites.
E-8	Bright Angel Shale	91.08	Quartz, palygorskite, smectite, and chlorite.
E-9	Indian Springs Shale	94.28	Quartz, kaolinite, smectite, and illite.

[1] See plate 1.

[2] Listed in order of approximate abundance.

ows flow system, which discharges at Ash Meadows springs, encompass large areas that contain several topographic basins which drain to a common discharge area (figs. 16, 17). In addition, both Muddy River springs and Ash Meadows springs receive water from the Pahranagat Valley area. Las Vegas and Pahrump Valleys receive water primarily from recharge to the Spring Mountains.

STABLE ISOTOPES OF THE GROUND WATER

Deuterium and oxygen-18 data can be used to help delineate ground-water flow systems by identifying water from different source areas. Most water in carbonate-rock aquifers of southern Nevada is chemically similar because mineralogy of the carbonate rock is relatively homogeneous, so the stable isotopes deuterium and oxygen-18 become an important tool for delineating source areas and flow paths in this area. Deuterium rather than oxygen-18 is used to delineate flow systems in southern Nevada because it has been used in previous studies (Winograd and Friedman, 1972; Claassen, 1985, 1986; Lyles and Hess, 1988; Noack, 1988; Kirk and Campana, 1990) and generally is not affected by water-rock interactions that can change oxygen-18 compositions.

Deuterium data used for flow-system delineation are primarily from samples analyzed by the U.S. Geological Survey research laboratory in Reston, Va. Additional deuterium data are included for samples analyzed prior to 1973 by the U.S. Geological Survey research laboratory in Denver, Colo. [after correcting for interlaboratory differences, the deuterium values analyzed before 1973 were divided by 1.03 (I.J. Winograd, U.S. Geological Survey, written commun., 1985)] and by the Desert Research Institute Isotope Laboratory in Las Vegas, Nev. An interlaboratory comparison between the Desert Research Institute Laboratory and U.S. Geological Survey Reston laboratory analyses showed that the average difference between deuterium values for 9 duplicate samples analyzed at the two laboratories was less than 1.0 part per thousand (permil), with only 1 sample difference greater than 2 permil; the average difference for 18 samples (9 duplicate samples plus 9 samples from springs that have a constant

TABLE 12.—*X-ray diffraction analyses of Miocene volcanic rocks from southern Nevada*

[Analysis by Robert Mariner, U.S. Geological Survey, Menlo Park, Calif. Symbol: --, no mineral identified]

Site[1]	Geologic unit or rock type sampled	Minerals identified		
		Major[2]	Trace[2]	Possible
V-1	Ash-flow tuff	Plagioclase, quartz, biotite, hornblende, halite	--	--
V-2	Basalt	Plagioclase	--	--
V-3	Ash-flow and ash-fall tuffs	Quartz, plagioclase, K-feldspar	--	--
V-4	Basalt	Plagioclase, clinopyroxene, olivine	--	Calcite
V-5	Kane Wash Tuff	Clinoptilolite	--	Mordenite
V-6	Rhyolite lavas	Quartz, K-feldspar, plagioclase	--	Cristobalite
V-7	Hiko Tuff	Biotite, plagioclase, K-feldspar, quartz	--	--
V-8	Bedded tuff	Quartz, plagioclase, K-feldspar, clinoptilolite	Mica	--

[1] See plate 1.

[2] Listed in order of approximate abundance.

deuterium concentration over time) was less than 0.3 permil, with only 3 sample differences greater than 2 permil (data are in appendix A). For consistency, if a sample site had data from the Reston laboratory and either of the other laboratories, only the Reston laboratory data were used to determine the average deuterium composition for that site.

ISOTOPIC COMPOSITION OF GROUND WATER IN SOURCE AREAS AND FLOW SYSTEMS

The average deuterium composition of ground water in recharge areas and in flow systems contributing to the carbonate-rock aquifers of southern Nevada (pl. 2) was determined from a data base compiled for this study (appendix A). All the data were plotted to eliminate samples significantly affected by evaporation (fig. 19). All samples that plot to the right of the line [$\delta D=8(\delta^{18}O)+0$] are assumed to have undergone significant evaporation and were not used in calculating the average deuterium composition of water from a sample site (fig. 20, pl. 2). Most samples plot to the right of the meteoric water line, indicating that water in southern Nevada has undergone a small amount of evaporation prior to infiltrating into aquifers.

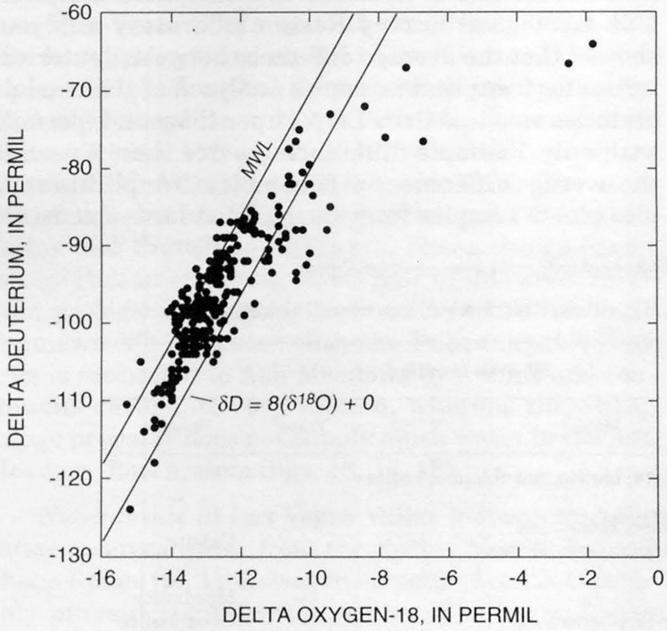

FIGURE 19.—Relation between mean deuterium and mean oxygen-18 for ground-water samples from sites in southern Nevada and southeastern California. MWL is the meteoric water line ($\delta D=8(\delta^{18}O)+10$; Craig, 1961).

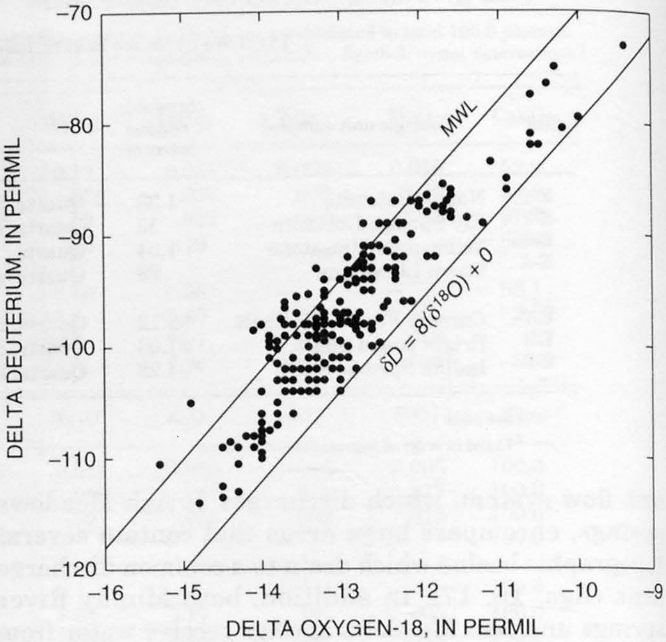

FIGURE 20.—Relation between mean deuterium and mean oxygen-18 for ground-water samples that have not undergone significant evaporation, southern Nevada and southeastern California. MWL is the meteoric water line ($\delta D=8(\delta^{18}O)+10$; Craig, 1961).

SPRING MOUNTAINS

Recharge from the Spring Mountains can be divided on the basis of isotopic composition into recharge from the central part of the mountains north of the Sandstone Bluffs area and from the Sandstone Bluffs area and everything south of that area (fig. 17; pl. 2). The central part of the Spring Mountains reaches an altitude of almost 12,000 ft, whereas the southern part of the Spring Mountains (Sandstone Bluffs and south of Sandstone Bluffs) reaches an altitude of only about 8,500 ft. The difference in altitude results in water in the higher central part having an average deuterium composition that is 9 permil lighter (more negative) than water in the lower southern part (table 13).

The average deuterium composition of Spring Mountains recharge for the central part of the Spring Mountains is −99 permil (table 13). This deuterium value is based on the average deuterium composition of the two largest discharging springs, Trout Spring and Cold Creek Spring (fig. 17), which also have the longest sampling record (1968–89). Water from Trout Spring has a mean deuterium composition of −97.7 permil, standard deviation 1.3, for 19 samples, and water from Cold Creek Spring has a mean deuterium composition of −100.1 permil, standard deviation 1.2, for 16 samples (Winograd and Riggs, 1984; I.J. Winograd, U.S. Geological Survey, written commun., 1989; Thomas and others, 1991). The average value of –99 permil for the central part of the Spring Mountains is the same as that presented by Winograd and Riggs (1984) for springs and wells in the Spring Mountains. Their data included 9 samples in addition to their combined 28 samples at Trout Spring and Cold Creek Spring. The average deuterium value agrees with other isotope data collected during a shorter period from smaller springs in the Spring Mountains (appendix A; pl. 2).

The average deuterium composition of recharge to the southern part of the Spring Mountains is –90 permil (table 13).

Deuterium values are different for the two parts of the Spring Mountains, but within each part, deuterium composition does not discernibly change with altitude (fig. 21). Thus, deuterium composition of winter precipitation that recharges the Spring Mountains (Winograd and Riggs, 1984) is not affected by altitude (deuterium composition does not become lighter with increasing altitude). Consequently, recharge from each part of the Spring Mountains should have similar deuterium composition regardless of altitude.

SHEEP RANGE

The average deuterium composition of recharge to the Sheep Range is –93 permil based on the average deuterium value of 17 samples at six springs (table 13, appendix A). This average deuterium composition is the same as that of water discharging from Corn Creek Springs (–93 permil; appendix A, fig. 17, pl. 2). Corn Creek Springs is on a fan on the southwest flank of the Sheep Range and, therefore, should be representative of water that flows either south to Las Vegas Valley or west to Ash Meadows from the Sheep Range.

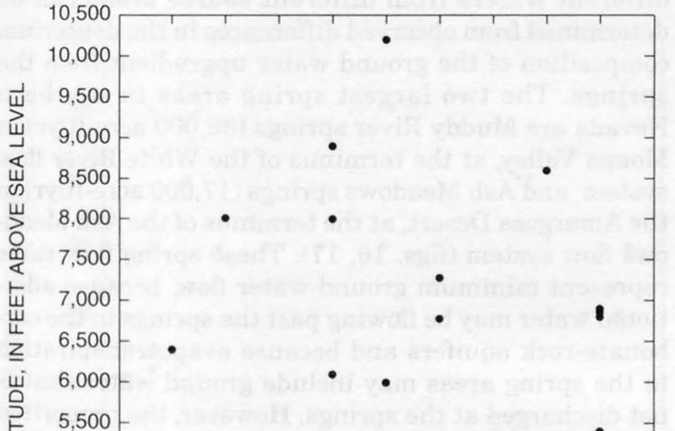

FIGURE 21.—Relation between mean deuterium composition and altitude for spring and well samples that contain tritium in the Spring Mountains, southern Nevada. Tritium indicates the water was less than about 50 years old at the time of sampling.

Six springs in the Sheep Range are at altitudes from about 5,600 to 8,400 ft. Their average deuterium composition changes by only 4 permil, so on the basis of limited data, altitude probably does not affect the deuterium composition of water in the Sheep Range. Thus, –93 permil represents the deuterium composition of recharge to the Sheep Range, as is observed at Corn Creek Springs, which is probably well-mixed water from the Sheep Range.

SOUTHERN MEADOW VALLEY WASH FLOW SYSTEM

The average deuterium composition of ground water in the southern Meadow Valley Wash flow system (fig. 16) is –87 permil (table 13). This value is an average for springs and wells in the Meadow Valley Wash drainage south of Caliente, the Delamar Mountains, the Meadow Valley Mountains, and the southwest Clover Mountains (pl. 2).

WHITE RIVER FLOW SYSTEM

The average deuterium composition of springs emanating from carbonate-rock aquifers in Pahranagat Valley is –109 permil (fig. 17, pl. 2). This average value is for samples from Ash and Crystal Springs that were analyzed by the Reston laboratory (table 13). This value agrees with data for previous samples from Ash, Crystal, and Hiko Springs in Pahranagat Valley, which were analyzed at the Denver and Las Vegas laboratories (appendix A). These samples were chosen as representative of water in the carbonate-rock aquifers of the White River flow system that flows into southern Nevada because they are large springs discharging from the carbonate-rock aquifers upgradient from Muddy River and Ash Meadows springs.

PALEOCLIMATIC EFFECTS ON ISOTOPIC COMPOSITION

Ground water discharging from the carbonate-rock aquifers in southern Nevada is a mixture of waters that were recharged primarily during the last 10,000 years (see table 18; Winograd and Pearson, 1976; Benson and others, 1983; Waddell and others, 1984; Claassen, 1985, 1986; Kirk, 1987; Benson and Klieforth, 1989; Kirk and Campana, 1990). The average deuterium composition of waters recharging the carbonate-rock aquifers of southern Nevada would have had to remain relatively constant during this time for deuterium to be

TABLE 13.—*Deuterium composition of ground water in source areas of southern Nevada*

[Analyses from U.S. Geological Survey research laboratory, Reston, Va., unless otherwise noted]

Site	Number[1]	Delta deuterium (permil)	Standard deviation
Central part of Spring Mountains			
Trout Spring[2]	19	−97.7	1.3
Cold Creek Spring[2]	16	−100.1	1.2
Sites averaged	2	−98.9	1.7
Southern part of Spring Mountains			
Bird Spring[3]	1	−88.0	--
Sandstone Spring	1	−89.0	--
BLM Visitors Center well	1	−89.0	--
Red Spring	1	−89.0	--
Willow Spring	1	−90.5	--
White Rock Spring	1	−91.0	--
Castilio well[3]	1	−94.0	--
Sites averaged	7	−90.1	2.0
Sheep Range			
Wiregrass Spring	9	−94.3	1.8
Mormon Well Spring	3	−91.8	.8
Cow Camp Spring	2	−92.0	1.8
Lamb Spring	1	−92.5	--
Sawmill Spring	1	−92.0	--
Sheep Spring	1	−96.0	--
Sites averaged	6	−93.1	1.7
Southern Meadow Valley Wash flow system			
Upper Riggs Spring	1	−88.0	--
Boulder Spring	1	−87.0	--
Kane Spring	1	−86.5	--
Grapevine Spring	1	−87.5	--
Willow Spring	1	−88.0	--
Caliente City well	1	−89.0	--
Bishop Spring	1	−85.5	--
Bradshaw well	1	−88.5	--
Railroad Elgin well	1	−86.0	--
Randono well	1	−87.5	--
Jensen well	1	−88.5	--
North Ella Spring	1	−86.5	--
Grassy Spring[4]	1	−85	--
Stock well[5]	1	−88.0	--
Sites averaged	14	−87.3	1.2
White River flow system			
Ash Spring	1	−108.0	--
Crystal Spring	1	−109.0	--
Sites averaged	2	−108.5	0.7

[1] Number of samples per site, and number of sites averaged.

[2] Data from I.J. Winograd (U.S. Geological Survey, written commun., 1989).

[3] Data from Desert Research Institute laboratory, Las Vegas, Nev. (Thomas and others, 1991).

[4] Unpublished data from Desert Research Institute laboratory, Las Vegas, Nev.

[5] Data from Desert Research Institute laboratory, Las Vegas, Nev. (Kirk, 1987, p. 81).

used to calculate mixing of water from different sources. This seems unlikely because of climatic changes during this period (Mifflin and Wheat, 1979; Winograd and Doty, 1980; Spaulding and others, 1984; Spaulding, 1985; Quade, 1986; Benson and Thompson, 1987; Quade and Pratt, 1989). However, a plot of deuterium and carbon-14 (fig. 22) for water that was recharged in the Spring Mountains shows that deuterium has remained relatively constant for a carbon-14 range of 1.9 to 100 percent modern carbon (pmc). Deuterium composition varies by only a total of 6 permil, with an average concentration of −99 permil. A similar plot for water recharged in the Sheep Range contains fewer data points; it shows that deuterium composition varies by less than 2 permil for a carbon-14 range of 13.7 to 96.8 pmc (fig. 23).

In summary, the deuterium composition of water in the Spring Mountains, Sheep Range, and surrounding areas has remained constant with time for a carbon-14 range of 1.9 to 100 pmc. Thus, deuterium can be used to determine source areas, flow paths, and mixing of ground water in the carbonate-rock aquifers of southern Nevada.

FLOW-SYSTEM DELINEATION ON THE BASIS OF DEUTERIUM

Deuterium composition of water discharging from large springs in southern Nevada can be used to determine the sources of water that supply the springs and, thus, help delineate flow paths. Mixing of isotopically different waters from different source areas can be determined from observed differences in the deuterium composition of the ground water upgradient from the springs. The two largest spring areas in southern Nevada are Muddy River springs (36,000 acre-ft/yr) in Moapa Valley, at the terminus of the White River flow system, and Ash Meadows springs (17,000 acre-ft/yr) in the Amargosa Desert, at the terminus of the Ash Meadows flow system (figs. 16, 17). These spring flow rates represent minimum ground-water flow, because additional water may be flowing past the springs in the carbonate-rock aquifers and because evapotranspiration in the spring areas may include ground water that is not discharged at the springs. However, the proportion of water from different sources determined on the basis of deuterium composition would be the same; only the absolute amount from the different areas would increase if the spring flow does not include all the water in the carbonate-rock aquifers in the spring area. In addition, the isotopic composition of water from wells, or discharging from springs, in Las Vegas and Pahrump Valleys can be used to determine sources of water in the carbonate and basin-fill aquifers in these areas.

If deuterium is used as a tracer, the deuterium composition of the different source areas has to be different. Although waters from the two major recharge areas in southern Nevada, the Spring Mountains and Sheep Range, are only 6 permil different in mean deuterium composition (table 13), a Mann-Whitney test shows that the medians of the two populations are detectably different at the 0.001 significance level. Additional, and perhaps even more compelling, evidence that deuterium values from the two areas are different is that deuterium composition of water along flow paths from each area is the same as the deuterium composition in the recharge area (pls. 1, 2). The mean deuterium composition of 22 samples along flow paths from the central part of the Spring Mountains is –98.4 permil (standard deviation 1.6) and ranges from –102.0 to –95.0 permil; the mean composition of 4 samples along flow paths from the Sheep Range is –93.6 permil (standard deviation 0.5) and ranges from –94.0 to –93.0 permil. Consequently, although waters from these recharge areas are only 6 permil different, they are statistically different populations, and ground water flowing from each recharge area maintains the same average deuterium value as the recharge-area water. Thus, the mean values of –99 and –93 permil can be used as the isotopic inputs for the two recharge areas. Other significant sources of ground water in carbonate-rock aquifers of southern Nevada are the southern part of the Spring Mountains (deuterium composition is –90 permil), the White River flow system (deuterium composition is –109 permil), and the southern Meadow Valley Wash flow system (deuterium composition is –87 permil; table 13).

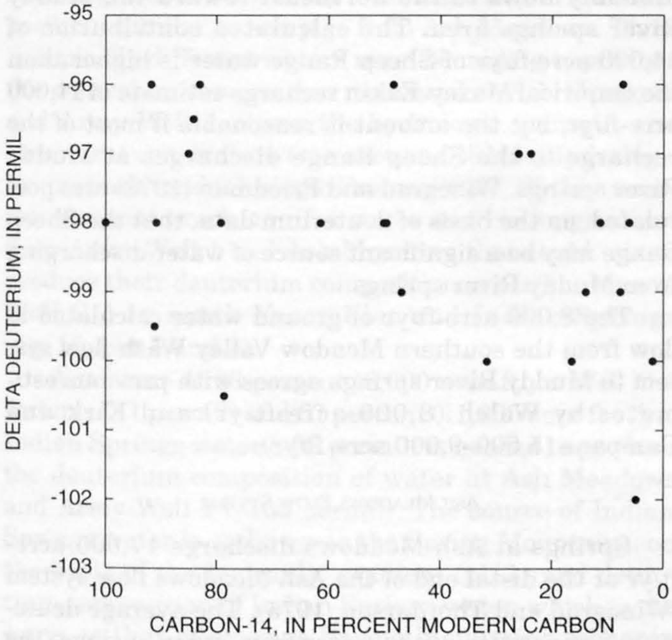

FIGURE 22.—Relation between mean deuterium and carbon-14 for water recharged in the Spring Mountains, southern Nevada.

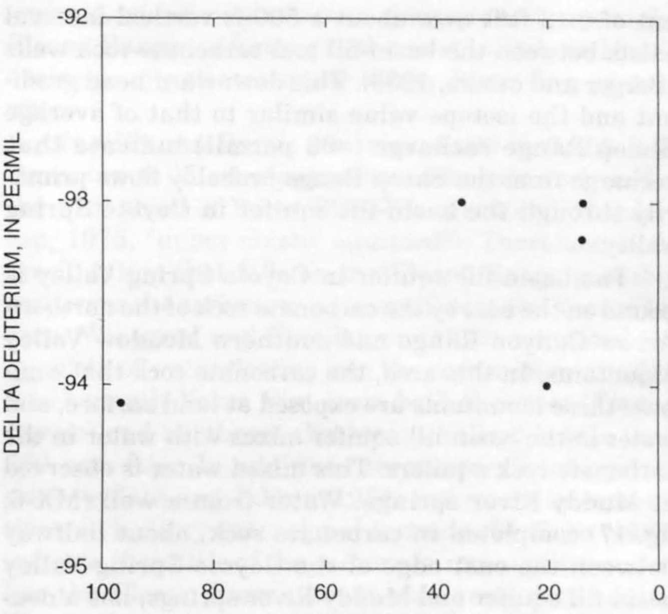

FIGURE 23.—Relation between mean deuterium and carbon-14 for water recharged in the Sheep Range, southern Nevada.

WHITE RIVER FLOW SYSTEM

Muddy River springs discharge 36,000 acre-ft/yr of water at the distal end of the White River flow system (Eakin and Moore, 1964). The average deuterium composition of water from five springs in the Muddy River springs area, including Big Muddy Spring, which is the largest discharging spring, is –98 permil (isotope analyses from the Reston laboratory; appendix A). The sources of water discharging from the springs were determined from hydraulic gradients in the carbonate-rock aquifers in this area (see section "Hydrologic Framework" and pl. 1), the geologic and structural ground-water flow constraints (see section "Geologic Framework" and figs. 17, 18), and the average deuterium composition of possible source waters.

Water emanating from Muddy River springs can be from three sources: (1) the Sheep Range, (2) the White River flow system, and (3) the southern Meadow Valley Wash flow system (including Kane Springs and Delamar Valleys). Directly upgradient from Muddy River springs is Coyote Spring Valley. In this valley, the average isotopic composition of water from three wells completed in carbonate rock is –101 permil, whereas water from a well completed in basin-fill deposits is –94 permil. The basin-fill well is adjacent to one of the wells completed in carbonate rock. A downward head gradient of 0.12 ft/ft over about a 500-ft vertical interval exists between the basin-fill and carbonate-rock wells (Berger and others, 1988). This downward head gradient and the isotope value similar to that of average Sheep Range recharge (–93 permil) indicate that recharge from the Sheep Range probably flows primarily through the basin-fill aquifer in Coyote Spring Valley.

The basin-fill aquifer in Coyote Spring Valley is bound on the east by the carbonate rock of the northern Arrow Canyon Range and southern Meadow Valley Mountains. In this area, the carbonate rock that compose these mountains are exposed at land surface, and water in the basin-fill aquifer mixes with water in the carbonate-rock aquifers. This mixed water is observed at Muddy River springs. Water from a well (MX-6; fig. 17) completed in carbonate rock, about halfway between the east edge of the Coyote Spring Valley basin-fill aquifer and Muddy River springs, has a deuterium composition of –97 permil (pl. 2). This isotopic composition is similar to Muddy River springs (–98 permil) and is more evidence supporting the conceptual flow and mixing model: water in the Muddy River springs area is probably a mixture of Sheep Range recharge water and water from the carbonate-rock aquifers beneath Coyote Spring Valley. Using the average deuterium composition of Sheep Range recharge water (–93 permil) and Coyote Spring Valley carbonate-rock aquifer water (–101 permil) to determine the sources of water at Muddy River springs (–98 permil) results in a mixture of 38 percent (14,000 acre-ft/yr) Sheep Range water and 62 percent (22,000 acre-feet/yr) Coyote Spring Valley water.

Water in the carbonate-rock aquifers of Coyote Spring Valley (deuterium composition of –101 permil) can be from two sources, the White River flow system (deuterium composition of –109 permil) and the southern Meadow Valley Wash flow system (deuterium composition of –87 permil; pls. 1 and 2, figs. 16, 17). A mixture of 64 percent (14,000 acre-ft/yr) White River flow-system water and 36 percent (8,000 acre-ft/yr) southern Meadow Valley Wash flow-system water results in water isotopically the same as water in the carbonate-rock aquifers in Coyote Spring Valley.

In summary, water discharging from Muddy River springs is a mixture of 40 percent (14,000 acre-feet/yr) White River flow-system water, 38 percent (14,000 acre-ft/yr) Sheep Range water, and 22 percent (8,000 acre-ft/yr) southern Meadow Valley Wash flow-system water. The 14,000 acre-ft/yr contribution of White River flow-system water to Muddy River springs is significantly less than the 35,000 acre-ft/yr proposed by Eakin (1966) on the basis of water-level data and Maxey-Eakin recharge estimates (Maxey and Eakin, 1949) but is similar to recent estimates by A.H. Welch (U.S. Geological Survey, written commun., 1988) and Kirk and Campana (1990). Welch estimated 18,000 acre-ft/yr of underflow from Pahranagat Valley to Coyote Spring Valley on the basis of the isotopic compositions of empirically derived Maxey-Eakin recharge estimates for the entire White River flow system. Kirk and Campana (1990) calculated a contribution of 16,500 to 19,100 acre-ft/yr for three different flow scenarios for the White River flow system on the basis of Maxey-Eakin recharge estimates and water-level data with a discrete-state compartment model using deuterium to calibrate their models. These flow-system delineations are based on water-level data only, with no consideration of geologic or structural constraints on ground-water flow.

The Sheep Range contribution of 14,000 acre-ft/yr is significantly higher than the estimated 2,000 acre-ft/yr of Eakin (1966), 3,000 acre-ft/yr of A.H. Welch (written commun., 1988), and 5,000 to 6,000 acre-ft/yr of Kirk and Campana (1990). The greater contribution of Sheep Range water compared to previous studies is balanced by not including 6,000–9,800 acre-ft/yr of ground-water from Dry Lake Valley, north of Delamar Valley, because of geologic constraints to ground-water flow (Dettinger and others, 1995) and less underflow from Pahranagat Valley to Coyote Spring Valley. Geologic constraints on Sheep Range water flowing to the west and south, as previously discussed in the section titled "Geologic Framework," indicates that most of the recharge to the Sheep Range probably flows to the northeast toward the Muddy River springs area. The calculated contribution of 14,000 acre-ft/yr of Sheep Range water is higher than the empirical Maxey-Eakin recharge estimate of 11,000 acre-ft/yr, but the amount is reasonable if most of the recharge to the Sheep Range discharges at Muddy River springs. Winograd and Friedman (1972) also postulated, on the basis of deuterium data, that the Sheep Range may be a significant source of water discharging from Muddy River springs.

The 8,000 acre-ft/yr of ground water calculated to flow from the southern Meadow Valley Wash flow system to Muddy River springs agrees with previous estimates by Welch (8,000 acre-ft/yr) and Kirk and Campana (5,500–9,000 acre-ft/yr).

ASH MEADOWS FLOW SYSTEM

Springs at Ash Meadows discharge 17,000 acre-ft/yr at the distal end of the Ash Meadows flow system (Winograd and Thordarson, 1975). The average deuterium composition of the water from seven springs (the six largest discharging springs plus Scruggs Spring) is –103 permil (Winograd and Pearson, 1976;

appendix A). The sources of water discharging from the springs were determined on the basis of hydraulic gradients in the carbonate-rock aquifers in this area (see section titled "Hydrologic Framework" and pl. 1), the geologic and structural constraints on ground-water flow (see section titled "Geologic Framework" and figs. 17, 18), and the deuterium composition of possible source waters.

The first carbonate-rock aquifer sample site that is upgradient from Ash Meadows springs and has deuterium and water chemistry data is Army Well 1 (fig. 17). Water from Army Well 1 has an average deuterium composition of –104 permil (appendix A). Thus, given the hydrologic position of the well and the isotopic similarity of its water to Ash Meadows springs, water at Army Well 1 is considered representative of water that flows to Ash Meadows. This conclusion was previously reached by Winograd and Friedman (1972), but they also noted that the chemistry at Army Well 1 was more dilute than water discharging at Ash Meadows. At a carbonate-aquifer sample site about halfway between Army Well 1 and Ash Meadows (Amargosa Tracer Well 2; fig. 17), deuterium data are lacking but oxygen-18 data are similar to data from Ash Meadows springs (appendix B). Water chemistry also is similar, although slightly more dilute, so this water also is considered representative of flow to Ash Meadows. The water chemistry from these two sites and how they relate to flow in the Ash Meadows flow system is discussed in the section titled "Water Chemistry."

No water samples from carbonate-rock aquifer sites upgradient from Army Well 1 had deuterium compositions similar to samples from Ash Meadows. Thus, isotopically different waters must be mixing to produce the deuterium composition measured at Ash Meadows and Army Well 1. Given the hydrologic, geologic, and structural constraints (see sections titled "Hydrologic Framework" and "Geologic Framework"), the two nearest carbonate-rock aquifer water sources upgradient from Army Well 1 and Ash Meadows that could mix to produce their deuterium composition are in the area of Well C-1 in south Yucca Flat and Indian Springs (fig. 17, pls. 1 and 2).

A mixture of 33 percent (6,000 acre-ft/yr) Well C-1 water (–111 permil) and 67 percent (11,000 acre-feet/yr) Indian Springs water (–99 permil) is needed to produce the deuterium composition of water at Ash Meadows and Army Well 1 (–103 permil). The source of Indian Springs water is recharge to the Spring Mountains, on the basis of the hydraulic gradient (pl. 1) and deuterium composition of Indian Springs water, which is the same as that of average Spring Mountains recharge (-99 permil). The source of Well C-1 water is less obvious: three possible sources, on the basis of hydrologic, geologic, and structural constraints, are recharge to the Eleana Range (or farther to the west in Pahute Mesa), drainage of paleowater, or White River flow-system water.

The Eleana Range contains 4,000 to 8,000 ft of Devonian to Mississippian noncarbonate rock under the west third of Yucca Flat (Winograd and Thordarson, 1975, "upper clastic aquitard"). Therefore, little precipitation that falls on the Eleana Range probably recharges the carbonate-rock aquifers in the Yucca Flat area. Winograd and Thordarson (1975) estimated the quantity of water flowing into the carbonate-rock aquifers beneath Yucca Flat from both the west (Eleana Range) and northeast (Emigrant Valley) is less than 250 acre-ft/yr. In addition, aeromagnetic interpretations by Bath and Jahren (1984) and recent interpretations of Tertiary extensional tectonics by Guth (1988) indicate that little of the carbonate-rock aquifer underlies the Eleana Formation in this area; instead, the Eleana Formation is probably underlain by noncarbonate basement. Thus, the possibility that water in the volcanic rock of Pahute Mesa, west of the Eleana Range, flows at depth into the carbonate-rock aquifers and then east to Yucca Flat is unlikely. No isotope data exist for the carbonate aquifers beneath the Eleana Range, so the isotopic composition of this water is unknown.

Drainage of water recharged during the last glacial episode is a possible source of water at Well C-1. However, Winograd and Doty (1980) show that water levels in the carbonate-rock aquifers in the Nevada Test Site area have fluctuated less than 100 ft during Wisconsin time, and Jones (1982) shows fluctuations of less than about 150 ft in the northern Frenchman Flat area through most of Quaternary time. Thus, drainage of paleowater is not probable.

White River flow-system water in Pahranagat Valley is isotopically similar (–109 permil) to Well C-1 water (–111 permil) and, on the basis of hydraulic gradients, could be flowing southwest to Yucca Flat (pl. 1). Continuous, thick sequences of carbonate rock provide a flow path for White River flow system water to Frenchman Flat (fig. 17). Thus, of the three possible sources of Well C-1 water, the White River flow system is hydrologically and geologically the most likely.

Another possibility is that little water flows from the Yucca Flat area to Ash Meadows. Winograd and Thordarson (1975, p. 94) estimated the total flow within the carbonate-rock aquifers beneath Yucca Flat to the south to be less than 350 acre-ft/yr. A likely alternative is that water from Pahranagat Valley flows through the Frenchman Flat area south of Well C-1 and mixes with Spring Mountains water, producing the water at Ash Meadows. This interpretation is reasonable because the central core of thick, continuous carbonate rock extends from Pahranagat Valley to Ash Meadows (fig. 17). Using the average isotope value of Pahranagat Valley water (–109 permil) and Indian Springs water (–99 permil) to produce Ash Meadows water (–103 permil) results in a mixture of 40 percent (7,000 acre-ft/yr) Pahranagat Valley water and 60 percent (10,000 acre-ft/yr) Spring Mountains water. The 40 percent contribution of Pahranagat Valley water to Ash Meadows spring discharge is in good agreement with the 35 percent estimated by Winograd and Friedman (1972) and Winograd and Thordarson (1975).

The Spotted, Pintwater, Desert, and Groom Ranges are assumed to contribute little water to the carbonate-rock aquifers (fig. 17). This assumption agrees with previous work by Winograd and Friedman (1972) and Winograd and Thordarson (1975). These mountains, with the exception of the Groom Range, are less than 7,100 ft in altitude and, therefore, do not receive large amounts of winter precipitation that could become available to recharge the carbonate-rock aquifers. The Groom Range is composed mostly of Precambrian basement rock and is not underlain by carbonate-rock aquifers (M.D. Dettinger, U.S. Geological Survey, oral commun., 1989); therefore, little precipitation in the Groom Range recharges the carbonate-rock aquifers. In addition, any potential recharge water in these ranges generally is isotopically heavy; median deuterium composition of 13 samples from the Pintwater and Groom Ranges is –90 permil (pl. 2; B.F. Lyles, Desert Research Institute, written commun., 1986) compared with recharge water in the Spring Mountains (–99 permil) and water in the White River flow system (–109 permil). This heavy deuterium composition severely limits the possibility that any significant recharge to these mountains contributes to Ash Meadows discharge.

As previously discussed, recharge to the Sheep Range probably contributes little to spring discharge at Ash Meadows due to geologic and structural constraints. The relatively heavy deuterium composition of Sheep Range water (–93 permil), as compared with Ash Meadows spring water (–103 permil), also limits the percentage of Sheep Range water that could mix with Spring Mountains and Pahranagat Valley water to produce the deuterium composition measured at Ash Meadows.

In summary, a mixture of 40 percent (7,000 acre-ft/yr) Pahranagat Valley water and 60 percent (10,000 acre-ft/yr) Spring Mountains water discharging at Ash Meadows springs is geologically, hydrologically, and isotopically the most likely alternative. Previous work by Winograd and Friedman (1972), Winograd and Thordarson (1975), Winograd and Pearson (1976), Welch and Thomas (1984), and Kirk and Campana (1990) postulated a 24 to 35 percent input of Pahranagat Valley water to Ash Meadows, which is similar to the 40 percent proposed by this isotopic mixing model. The 60 percent Spring Mountains contribution also is reasonable, if the previous estimate of about 65 percent Spring Mountains plus Sheep Range water (Winograd and Friedman, 1972; Winograd and Thordarson, 1975; Winograd and Pearson, 1976) is assumed to be mostly Spring Mountains water. This assumption seems reasonable because the previous studies assumed that the Spring Mountains and Sheep Range were isotopically the same and no flow barriers existed between the Sheep Range and Ash Meadows springs. Other evidence to support the concept that recharge from the Spring Mountains contributes 60 percent of Ash Meadows springs discharge is as follows:

1. A Maxey-Eakin recharge estimate, which assumes that only precipitation above 6,000 ft becomes recharge, for the part of the Spring Mountains that topographically drains to the Ash Meadows flow system is 7,000 acre-ft/yr. This estimate is lower than the 10,000 acre-ft/yr estimated by the isotope mixing model, but ground-water flow modeling studies of Las Vegas and Pahrump Valleys (Harrill, 1976, 1986) indicate that Maxey-Eakin recharge estimates for the Spring Mountains underestimate recharge by about 20 to 35 percent.
2. Winograd and Thordarson (1975) suggest on the basis of structural disposition that some recharge south of the topographic divide in the Spring Mountains flows northward into Indian Springs Valley rather than southwestward into Pahrump Valley.
3. In a recharge area such as the Spring Mountains, which contains well-mixed water, as indicated by the lack of isotopic depletion with increased altitude (fig. 21), topographic divides probably have less effect on the areas of recharge than in a recharge area that contains less well-mixed water.

LAS VEGAS VALLEY

Isotopic composition of ground water in the basin-fill aquifers of Las Vegas Valley indicates that the aquifers are supplied almost entirely by recharge to the Spring Mountains. This conclusion agrees with ground-water flow modeling studies by Harrill (1976) and Morgan and Dettinger (1996). The average deuterium composition of water from 10 wells and springs in northern Las Vegas Valley is –98 permil, ranging from –101 to –96 permil (pl. 2). This average value is similar to that of recharge to the central part of the Spring Mountains (–99 permil). Ground water in the central part of the valley has a deuterium composition that ranges from –106 to –94 permil, but deuterium for most of the water ranges from –101 to –94 permil, indicating this water is either from northern Las Vegas Valley or is upward flow from the carbonate-rock aquifers that originated as high-altitude recharge in the Spring Mountains. The lightest sample (–106 permil) may contain some isotopically light paleowater recharged during the last glacial period.

Another possible source of recharge to northern Las Vegas Valley, on the basis of topographic boundaries and hydrologic data, is the Sheep Range. Although deuterium data indicate that ground water in northern Las Vegas Valley originates primarily as precipitation in the central part of the Spring Mountains, hydrologic data indicate that some water from the Corn Creek area, which originates as precipitation in the Sheep Range, flows into northern Las Vegas Valley. However, geologic flow constraints (see section titled "Geologic Framework") and ground-water flow modeling studies (Harrill, 1976; Morgan and Dettinger, 1996) indicate that only a small amount of recharge to the Sheep Range flows into Las Vegas Valley.

Ground water from four wells in the southwestern part of Las Vegas Valley has an average deuterium composition of –89 permil, which is similar to that of recharge to the southern part of the Spring Mountains (table 13, pl. 2). Therefore, given the isotopic similarity between the recharge waters in the southern Spring Mountains and ground water in southwestern Las Vegas Valley, the hydraulic gradients in this area (pl. 1), and the lack of any other major recharge area nearby, water in southwestern Las Vegas Valley is most likely derived from precipitation in the southern part of the Spring Mountains.

Another possible source of water for southern Las Vegas Valley is the McCullough Range (pl. 1, fig. 17). However, the deuterium composition of sampled water in the range, –73 and –88 permil, is significantly heavier than that of water samples from southern Las Vegas Valley (pl. 2).

PAHRUMP VALLEY

Deuterium data combined with hydrologic and geologic information indicate water in Pahrump Valley originates entirely in the Spring Mountains. This observation is in agreement with a ground-water flow modeling study by Harrill (1986). Three sites in Pahrump Valley have an average deuterium composition of –98 permil, which is similar to the composition (–99 permil) of average recharge in the central part of the Spring Mountains. Ground water in Pahrump Valley flows out of the valley to the southwest through carbonate-rock aquifers to Chicago Valley and the Amargosa Desert area of Shoshone and Tecopa (pl. 1).

SUMMARY OF DEUTERIUM-DELINEATED FLOW SYSTEMS

Source areas and flow paths were delineated by deuterium mass-balance mixing models for the major carbonate-rock aquifers in southern Nevada (fig. 24). The volumes of flow are based on the discharge of water from Muddy River and Ash Meadows springs for the White River and Ash Meadows flow systems (so they represent minimum flow volumes) and on ground-water flow model studies of Las Vegas and Pahrump Valleys (Harrill, 1976, 1986; Morgan and Dettinger, 1996).

WATER CHEMISTRY

Changes in water chemistry along flow paths and resulting from mixing different waters, as determined by deuterium mass-balance calculations, have to be accounted for by realistic geochemical processes for the flow paths and mixing to be probable. Geochemical processes can be defined by mass-balance reaction models on the basis of the chemical composition of the initial (or mixed) and final waters along a flow path. Input to the models must be phases (minerals and gases) that have been identified in the aquifers and processes (mineral dissolution, precipitation, and formation; gas dissolution or exsolution; and ion exchange) that are thermodynamically and physically feasible, to describe the inputs and outputs (mass transfers) of constituents between the initial and final waters. For example, calcium concentration cannot decrease along a flow path unless calcium is contained in some phase in concentrations greater than saturation in the water and is able to precipitate, unless calcium is exchanged for another ion in a solid phase, or unless the water mixes with another water of lower calcium concentration. Amounts of mass transfer and mineral saturation were calculated using average chemical composition of samples along flow paths or, in the case of mixing, using samples from the different areas that also have deuterium and oxygen-18 data (appendix B).

The mass-balance approach (Plummer and Back, 1980; Parkhurst and others, 1982; Plummer and others, 1983; Plummer, 1984; Plummer and others, 1990) does not produce a unique numerical solution. Therefore, three different mass-balance models were used for each flow path and mixing scenario (table 14, col. 2). Mass transfers were calculated using the computer program NETPATH (Plummer and others, 1991). Flow-path sites with averaged chemistries, or mixtures of waters with averaged chemistries, used for flow-path

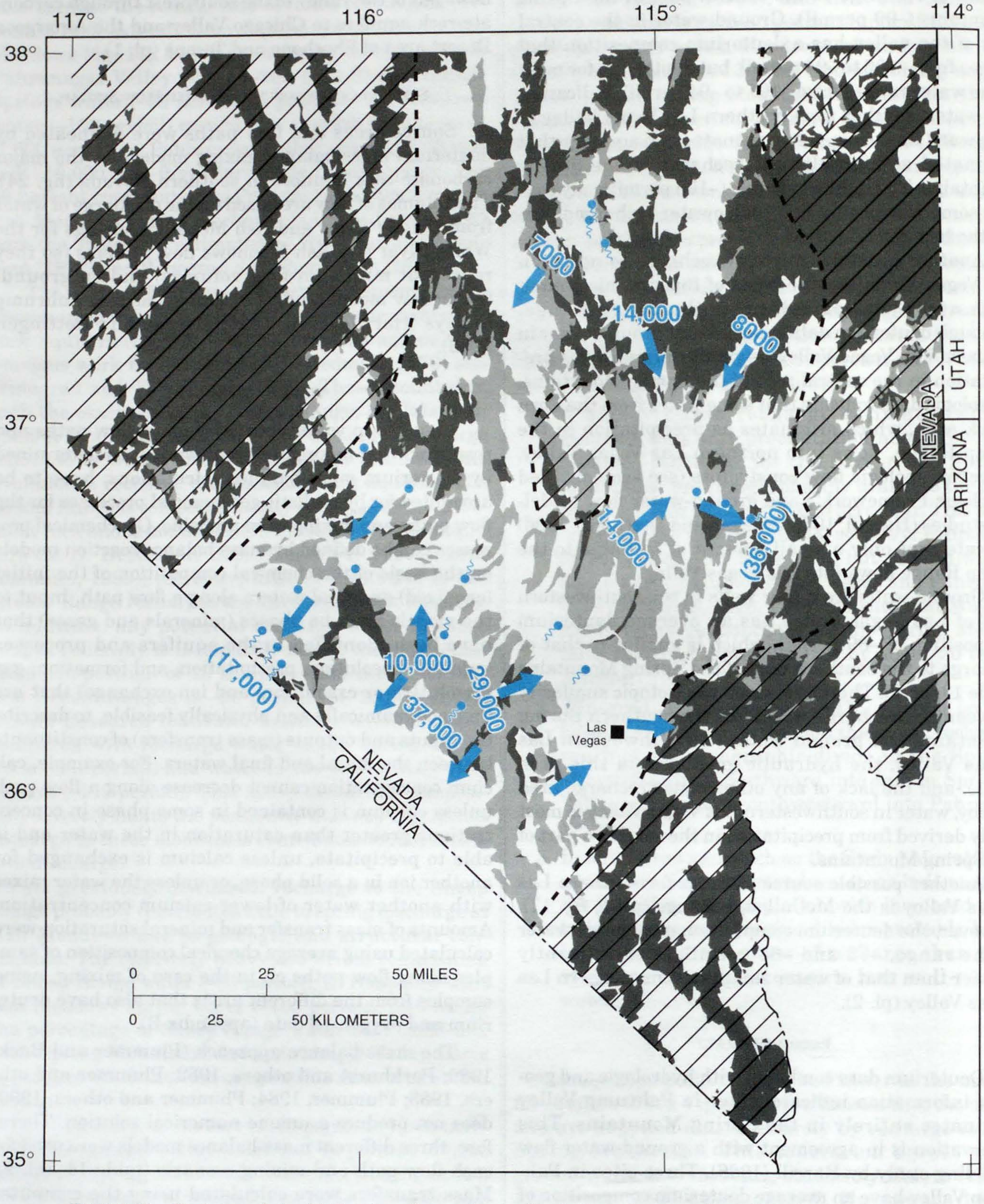

FIGURE 24.—Deuterium-derived delineation of ground-water flow in the carbonate-rock aquifers of southern Nevada. See figure 17 for feature names. Modified from Dettinger (1989, fig. 3).

calculations are numbered for reference in table 14, col. 1 and footnote 1. Phases used in the mass-balance reaction models are the prevalent minerals in the aquifers of southern Nevada (see section titled "Geologic Framework"). These minerals include calcite, dolomite, albite, kaolinite, K-feldspar, chalcedony, and clinoptilolite. Gypsum and halite are included as phases in the models of the White River flow system because Permian and younger rock in the Coyote Spring Valley area contain evaporite minerals. The Ash Meadows flow system, except for the Spring Mountains, consists of lower and middle Paleozoic carbonate rock that do not contain evaporite minerals. Thus, leakage of water from the noncarbonate Tertiary rock ("Tertiary aquitard" of Winograd and Thordarson, 1975) overlying the carbonate-rock aquifers is assumed to be the source of increased sulfate and sodium concentrations, as proposed by Winograd and Thordarson. Carbon dioxide gas is used as a source or sink for carbon in addition to calcite and dolomite. Cation exchange of calcium and magnesium for sodium in clay minerals also is used because clays are present (see section "Geologic Framework"). Calcite, dolomite, kaolinite, CO_2 gas, chalcedony, gypsum, and halite, or in place of gypsum, leakage from the Tertiary aquitard, are used in all three mass-balance reaction models. Albite and K-feldspar are used in two reaction models, calcium and magnesium exchange for sodium is substituted for albite in one reaction model, and clinoptilolite is substituted for K-feldspar in another reaction model (table 14). The zeolite mineral analcime also could be used in place of albite as a source of sodium, because water in the carbonate-rock aquifers of southern Nevada are all below saturation with respect to analcime (table 15), and analcime is common in volcanic rock that overlie carbonate-rock aquifers in the Nevada Test Site area (Hoover, 1968).

EXPLANATION

☐ **Basin-fill deposits**—In areas underlain by thick, laterally continuous carbonate rock

▨ **Basin-fill deposits**—In areas outside thick, laterally continuous carbonate rock

▒ **Carbonate rock**—In areas of thick, laterally continuous carbonate rock

▨ **Carbonate rock**—In areas outside thick, laterally continuous carbonate rock

■ **Noncarbonate rock**—In areas of thick, laterally continuous carbonate rock

▩ **Noncarbonate rock**—In areas outside thick, laterally continuous carbonate rock

- - - - **Boundary of laterally continuous carbonate rock**

8000 ➡ **Generalized direction and rate of ground-water flow in carbonate-rock aquifers**—Number is flow rate, in acre-feet per year

(17,000) ● **Springs discharging from carbonate-rock aquifer**—Number is discharge, in acre-feet per year

● **Well**—Source of water is carbonate-rock aquifers

Although many mass-balance reaction models require input of CO_2 gas, the source of the CO_2 gas is not apparent. Carbon dioxide gas is probably not added to the water. Instead, this small input may result from the large variability (2.9–7.1 mmol/L) of total dissolved inorganic carbon (TDIC) in water of springs and wells in the Spring Mountains and Sheep Range. Average TDIC concentration of the water in the recharge areas probably is less than actual TDIC of water recharging the carbonate-rock aquifers. In addition, some carbon dioxide gas in water in the recharge areas may not have been measured. Exsolution of CO_2 gas from high-altitude springs would result in lower measured than actual CO_2 concentrations in recharge waters. This exsolution is indicated by oversaturation of the spring waters with respect to calcite in the Spring Mountains and Sheep Range (table 15). The amount of carbon dioxide exsolution was calculated to range from 0.00 to 0.42 mmol/L by the computer program PHREEQE (Parkhurst and others, 1980), which assumes the waters are at saturation with respect to calcite and oversaturation is the result of carbon dioxide exsolution (Pearson and others, 1978).

Another explanation for higher CO_2 concentrations than those currently measured in recharge areas could be that older water in the aquifers was recharged during past cooler and wetter climatic conditions. This recharge would result in greater CO_2 concentrations in soil zones in recharge areas and, therefore, higher CO_2 concentrations in these older waters in the flow systems (White and Chuma, 1987).

Organic carbon is probably not a significant source of CO_2 gas in the carbonate-rock aquifers because (1) organic matter was not detected in the rock during the carbon-13 analysis of calcite and dolomite (David Meredith, Global Geochemistry, oral commun., 1987); (2) concentrations of dissolved organic carbon in the water are low—average 0.7 mg/L for 14 samples in southern Nevada (Thomas and others, 1991)—and concentrations of total organic carbon also are low—1.0 to 2.2 mg/L for 3 samples (Winograd and Pearson, 1976, p. 1133); (3) dissolved oxygen concentration is greater than 2.0 mg/L for all samples in southern Nevada (appendix B), which indicates the source of dissolved carbon is probably not oxidation of organic matter, because oxidation rapidly depletes dissolved oxygen in

TABLE 14.—*Mass transfer of constituents and mixing of waters along flow paths in carbonate-rock aquifers of southern Nevada*

[Coefficients of mass-transfer reactions were calculated using the computer program NETPATH (Plummer and others, 1991). Except for Ca-Mg/Na exchange, negative value indicates phase leaving solution (precipitating or degassing) and positive value indicates phase entering solution (dissolving). For exchange reaction, positive value indicates sodium entering solution and calcium and magnesium leaving solution. Symbol: --, constituent not used in mass-balance reaction]

Flow path[1]	Reaction model[2]	Millimoles per liter										
		Calcite ($CaCO_3$)	Dolomite [$CaMg(CO_3)_2$]	Gypsum ($CaSO_4 \cdot 2H_2O$)	Halite ($NaCl$)	Albite ($NaAlSi_3O_8$)[3]	Kaolinite [$Al_2Si_2O_5(OH)_4$]	K-feldspar ($KAlSi_3O_8$)	CO_2 gas	Chalcedony (SiO_2)	Clinoptilolite ($Na_{0.3}K_{0.4}Ca_{0.3}$)($AlSi_5O_{12}$)	Ca,Mg/Na exchange[4]
White River flow system												
2 to 1	1	–0.97	–0.02	1.30	0.95	0.84	–0.45	0.07	0.30	–1.77	--	--
	2	–.97	–.02	1.30	.95	.76	–.45	--	.30	–2.06	0.14	--
	3	–.88	.15	1.30	.95	--	–.04	.07	-.12	–.10	--	0.42
3 to 2A	1	–0.76	0.15	0.62	0.48	1.32	–0.72	0.12	0.97	–2.90	--	--
	2	–.76	.15	.62	.48	1.20	–.72	--	.97	–3.36	0.23	--
	3	–.66	.43	.62	.48	--	–.06	.12	.31	--.27	--	0.66
3A to 1	1	–1.44	0.08	1.68	1.25	1.67	–0.91	0.14	0.87	–3.60	--	--
	2	–1.44	.08	1.68	1.25	1.53	–.91	--	.87	–4.17	0.29	--
	3	–1.26	.41	1.68	1.25	--	–.07	.14	.04	–.26	--	0.83
Ash Meadows flow system												
5 to 4	1	0.10	–0.12	--	0.19	0.86	–0.46	0.06	0.74	–1.77	--	--
	2	.05	–.12	--	.19	.81	–.49	--	0.79	–2.18	0.16	--
	3	.17	.06	--	.19	--	–.03	.06	.31	–.05	--	0.43
6 to 4	1	–0.37	0.06	--	0.51	1.15	–0.64	0.12	0.53	–2.37	--	--
	2	–.46	.06	--	.51	1.06	–.67	--	.62	–3.11	0.29	--
	3	–.27	.30	--	.51	--	–.06	.12	.65	–.08	--	0.57
6 to 5	1	–0.47	0.18	--	0.32	0.34	–0.19	0.05	0.47	–0.71	--	--
	2	–.51	.18	--	.32	.30	–.21	--	.51	–1.03	0.13	--
	3	–.46	.25	--	.32	--	–.03	.05	.30	–.03	--	0.17
8 to 7	1	–0.90	0.25	0.08	0.08	0.04	–.03	0.02	-0.54	–0.02	--	--
	2	–.90	.25	.08	.08	.02	–.03	--	-.54	–.09	0.03	--
	3	–.90	.26	.08	.08	--	–.01	.02	-.55	.05	--	0.02
10 to 9	1	0.00	0.05	--	0.40	1.35	–0.67	0.00	1.03	–2.73	--	--
	2	.00	.05	--	.40	1.35	–.67	--	1.03	–2.73	0.00	--
	3	–.13	.32	--	.40	--	.00	.00	.36	–.04	--	0.67
6 to 9	1	–0.52	0.02	--	0.49	2.57	–1.34	0.10	1.97	–5.24	--	--
	2	–.52	.02	--	.49	2.46	–1.34	--	1.97	–5.66	0.20	--
	3	–.15	.44	--	.51	--	–.04	.08	.60	–.01	--	0.83
11 to 4	1	–0.38	–0.05	--	0.35	0.66	–0.39	0.12	0.79	–1.37	--	--
	2	–.46	–.05	--	.35	.57	–.43	--	.88	–2.12	0.29	--
	3	–.32	.18	--	.35	--	–.06	.12	.46	–.06	--	0.33

TABLE 14.—*Mass transfer of constituents and mixing of waters along flow paths in carbonate-rock aquifers of southern Nevada*—Continued

Flow path[1]	Reaction model[2]	Millimoles per liter										
		Calcite ($CaCO_3$)	Dolomite [$CaMg(CO_3)_2$]	Gypsum ($CaSO_4 \cdot 2H_2O$)	Halite (NaCl)	Albite ($NaAlSi_3O_8$)[3]	Kaolinite [$Al_2Si_2O_5(OH)_4$]	K-feldspar ($KAlSi_3O_8$)	CO_2 gas	Chalcedony (SiO_2)	Clinoptilolite ($Na_{0.3}K_{0.4}Ca_{0.3}$) ($AlSi_5O_{12}$)	Ca,Mg/Na exchange[4]
						Las Vegas Valley						
13 to 12	1	-0.95	0.35	0.08	0.10	0.19	-0.11	0.03	-0.07	-0.24	--	--
	2	-.95	.35	.08	.10	.16	-.11	--	-.07	-.35	0.06	--
	3	-.95	.40	.08	.10	--	-.01	.03	-.16	.14	--	0.10
15 to 14	1	-3.65	1.61	4.48	0.08	0.23	-0.13	0.04	-0.37	-0.49	--	--
	2	-3.65	1.61	4.48	.08	.19	-.13	--	-.37	-.63	0.07	--
	3	-3.63	1.66	4.48	.08	--	-.02	.04	-.40	-.03	--	0.11
17 to 12	1	-0.71	0.21	0.07	0.08	0.18	-0.10	0.02	-0.11	-0.26	--	--
	2	-.71	.21	.07	.08	.15	-.10	--	-.11	-.35	0.04	--
	3	-.71	.25	.07	.08	--	-.01	.02	-.20	-.09	--	0.09
						Pahrump Valley						
13 to 16	1	-0.77	0.30	0.15	0.03	0.09	-0.05	0.02	-0.12	-0.10	--	--
	2	-.77	.30	.15	.03	.08	-.05	--	-.12	-.16	0.03	--
	3	-.77	.32	.15	.03	--	-.01	.02	-.17	.09	--	0.05

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[1] Flow-path numbers, and average chemistries for waters, are:

1	Big Muddy Spring.
2	Mixture of 62 percent Coyote Spring Valley water (MX-4, MX-5, and VF-2 wells) and 38 percent Sheep Range water (Wiregrass, Cow Camp, and Mormon Well Spring).
2A	Coyote Spring Valley (MX and VF-2 wells).
3	Mixture of 64 percent White River flow system water from Pahranagat Valley (Crystal and Ash Springs) and 36 percent southern Meadow Valley Wash flow system water (Boulder, Kane, Grapevine, and Willow Springs, and Bradshaw, Railroad, Randono and Jensen wells).
3A	Mixture of 40 percent White River flow system water from Pahranagat Valley (Crystal and Ash Springs), 38 percent Sheep Range water (Wiregrass, Cow Camp, and Mormon Well Spring), and 22 percent southern Meadow Valley Wash flow system water (Boulder, Kane, Grapevine, and Willow Springs, and Bradshaw, Railroad, Randono, and Jensen wells).
4	Discharge-weighted average for springs at Ash Meadows (Winograd and Pearson, 1976, p. 1130-1131).
5	Army Well 1.
6	Mixture of 60 percent central part of Spring Mountains water (Sky Mountain, Summer Homes, and Highway Maintenance wells, and Willow, Deer Creek no. 1, Deer Creek no. 2, Rainbow, Trout, and Cold Creek Springs) and 40 percent White River flow system water from Pahranagat Valley (Crystal and Ash Springs).
7	Indian Springs.
8	Cold Creek Spring in Spring Mountains.
9	Test Well 3 in Frenchman Flat.
10	White River flow system water from Pahranagat Valley (Crystal and Ash Springs).
11	Mixture of 60 percent central part of Spring Mountains water and 40 percent Test-Well-3 water.
12	Tule Spring area in north Las Vegas Valley (Holland, Racel, Mulder, Martin, and Tule Spring wells).
13	Central part of Spring Mountains.
14	Southwestern Las Vegas Valley (Stocks and Spanish Trail wells; water from Tenaya well is excluded because it contains twice as much chloride as sodium—in milliequivalents per liter).
15	Southern part of the Spring Mountains southwest of Las Vegas Valley (BLM Visitors Center well and Sandstone, Red, and White Rock Springs).
16	Pahrump Valley (Manse well and Pahrump Spring well).
17	Mixture of 85 percent central part of Spring Mountains water (no. 13) and 15 percent Sheep Range water in Corn Creek Springs area (Corn Creek Springs, and Shown, Young, and Brooks wells).

[2] Most constituents are involved in all three mass-balance reactions; clinoptilolite is used in place of K-feldspar in reaction 2 and calcium-magnesium exchange for sodium is used in place of albite in reaction 3. See p. 41 for more detailed discussion.

[3] Zeolite mineral analcime can be used in place of albite; only resulting mass-transfer difference is reduction in amount of precipitating chalcedony (or other silicate mineral); see text.

[4] Calcium and magnesium exchange for sodium is based on calcium-to-magnesium ratio in final water; see text.

TABLE 15.—*Saturation indices for selected minerals in water from carbonate-rock aquifers of southern Nevada*[1]

[Calculated using computer program WATEQ4F (Ball and others, 1987)]

Site	Calcite	Dolomite	Gypsum	Halite	Albite	Kaolinite	K-feldspar	Chalcedony	Analcime
WHITE RIVER FLOW SYSTEM									
Muddy River springs									
Big Muddy Spring	0.0	−0.1	−1.4	−6.8	−1.4	1.9	−0.1	0.1	−3.2
Coyote Spring Valley									
MX well	0.0	0.1	−1.8	−7.2	−1.4	1.7	0.0	0.2	−3.2
VF-2 well	.2	.3	−1.8	−7.2	−1.3	1.6	.1	.2	−3.2
Sheep Range									
Wiregrass Spring	0.0	−0.3	−2.7	−9.6	−4.0	0.9	−1.6	0.0	−5.9
Cow Camp Spring	.1	.1	−2.4	−7.7	−2.3	1.5	−1.3	.1	−4.2
Mormon Well Spring	.2	.2	−2.5	−8.2	−2.8	1.4	−1.6	.1	−4.8
Southern Meadow Valley Wash									
Boulder Spring	−0.3	−1.0	−3.1	−8.5	−1.0	2.3	0.7	0.4	−3.3
Kane Spring	−.4	−1.2	−2.5	−8.0	−1.0	2.6	.9	.6	−3.5
Grapevine Spring	.0	−.2	−2.0	−7.9	−2.1	2.2	−.5	.2	−4.1
Willow Spring	−.6	−1.9	−2.4	−7.4	.0	3.0	1.3	.6	−2.5
Bradshaw well	.3	−.2	−1.7	−6.8	.0	2.8	1.4	.7	−2.5
Railroad Elgin well	.1	−.2	−2.0	−6.9	.0	2.7	1.4	.6	−2.4
Randono well	.2	.0	−2.0	−6.9	.1	2.8	1.4	.6	−2.4
Jensen well	.4	.3	−1.8	−6.9	.2	2.7	1.4	.6	−2.2
Pahranagat Valley									
Crystal Spring	0.2	0.3	−2.2	−8.3	−1.6	2.0	0.2	0.2	−3.6
Ash Spring	.0	−.2	−2.2	−8.2	−2.0	1.7	−.3	.1	−3.8
ASH MEADOWS FLOW SYSTEM									
Ash Meadows springs									
Ash Meadows springs[2]	0.1	0.0	−1.8	−7.4	−1.4	2.1	0.1	0.1	−3.3
Mercury Valley									
Army Well 1	0.1	0.2	−2.0	−7.8	−2.2	1.4	−0.8	0.0	−3.9
Indian Springs Valley									
Indian Springs	0.2	0.3	−2.5	−9.4	−3.5	1.2	−1.7	−0.2	−5.0
Central part of Spring Mountains									
Sky Mountain well	0.1	−0.3	−2.1	−9.4	−3.6	1.4	−2.1	−0.1	−5.4
Summer Homes well	−.1	−.5	−2.6	−10.2	−4.9	.4	−2.6	−.3	−6.6
Willow Spring	.1	−.3	−2.5	−10.0	−4.4	.7	−2.4	−.2	−6.1
Deer Creek Spring no. 1	.1	−.5	−3.2	−10.7	−5.1	−.3	−3.0	−.3	−6.7
Deer Creek Spring no. 2	.9	1.2	−2.9	−10.4	−4.4	−.1	−2.2	−.3	−5.9
Highway maintenance well	0.2	−0.2	−2.1	−10.2	−4.7	0.5	−2.7	−0.3	−6.3
Rainbow Spring no. 2	.0	−.6	−2.8	−9.9	−5.0	.4	−2.8	−.3	−6.6
Trout Spring	.6	.6	−2.9	−10.7	−4.9	−.3	−2.7	.4	−6.4
Cold Creek Spring	.2	−.1	−2.6	−10.3	−4.6	.6	−2.6	−.2	−6.2
Frenchman Flat									
Test Well 3	0.2	0.3	−1.8	−7.3	−2.0	1.0	−0.8	0.0	−3.6

TABLE 15.—*Saturation indices for selected minerals in water from carbonate-rock aquifers of southern Nevada*[1]—Continued

Site	Calcite	Dolomite	Gypsum	Halite	Albite	Kaolinite	K-feldspar	Chalcedony	Analcime
South Yucca Flat									
C-1 Well	0.9	1.8	−1.9	−7.0	−1.5	0.5	−0.2	0.1	−3.3
LAS VEGAS VALLEY FLOW SYSTEM									
Tule Spring area - North Las Vegas Valley									
Martin well	0.2	0.4	−2.6	−8.9	−2.6	1.4	−0.9	0.0	−4.3
Racel well	.2	.4	−2.5	−9.0	−2.5	1.5	−.9	.0	−4.3
Tule Spring State Park well	.1	.2	−2.5	−8.9	−2.4	1.8	−.7	.1	−4.3
Mulder well	.3	.6	−2.4	−8.6	−2.5	1.3	−1.1	.0	−4.3
Holland well	−.2	−.4	−2.5	−9.2	−2.9	2.1	−1.2	.0	−4.6
Southern part of Spring Mountains									
Sandstone Spring	−0.4	−1.0	−2.0	−8.8	−3.8	1.5	−1.9	−0.2	−5.5
BLM Visitors Center well	.1	.0	−1.3	−8.8	−3.1	1.6	−1.5	−.1	−4.7
Red Spring	.2	.3	−2.4	−9.1	−3.2	1.5	−1.2	−.1	−4.9
White Rock Spring	−.3	−.8	−1.3	−8.5	−3.3	2.0	−1.6	−.1	−4.9
Southwestern Las Vegas Valley									
Tenaya well	0.0	0.0	−1.1	−7.2	−2.2	2.0	−0.6	0.0	−4.0
Spanish Trail Country Club	.1	.1	−.7	−8.4	−3.0	1.7	−1.3	−.1	−4.6
Stocks Mill and Supply Co.	.0	.0	−.8	−8.4	−2.8	1.9	−1.1	−.1	−4.5
Corn Creek Springs area									
Corn Creek Spring	0.2	0.5	−2.5	−8.9	−2.6	1.8	−0.7	0.0	−4.5
Shown well	−.2	−.3	−2.6	−8.9	−2.1	2.4	−.2	.3	−4.2
Brooks well	.8	1.9	−2.6	−8.9	−1.8	1.1	−.1	.3	−3.8
Young well	−.2	.1	−2.9	−9.0	−1.2	2.6	.9	.5	−3.6
PAHRUMP VALLEY FLOW SYSTEM									
Pahrump Valley									
Manse well	0.1	0.0	−2.3	−9.5	−3.2	1.6	−1.6	−0.1	−4.9
Pahrump Spring well	.2	.4	−2.3	−9.4	−3.2	1.3	−1.4	−.1	−4.9

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[1] Saturation index = $\log \frac{\text{(ion-activity product)}}{K_T}$, where K_T = equilibrium constant at temperature T. By convention, positive value indicates mineral can precipitate from solution, whereas negative value indicates mineral can dissolve if present. Data for aluminum concentration were not available for these sites, so a value of 10 μg/L was used for all sites in calculating saturation indices.

[2] Discharge-weighted average was used.

water; and (4) perhaps the strongest evidence of the lack of oxidation of organic carbon is that a large influx of organically derived CO_2 would cause a shift to a lighter carbon-13 composition, and this is not observed (see section titled "Ground-Water Age").

Carbon-dioxide gas from rock units overlying the aquifers was eliminated as a possible source of CO_2 gas in the aquifer, even though large amounts of CO_2 gas are present in some of the overlying rock units (D.C. Thorstenson, U.S. Geological Survey, oral commun., 1988), because (1) partial pressures of CO_2 in rock units overlying the aquifers are less than partial pressures of CO_2 in the aquifer waters, (2) the rates of diffusion and dispersion for CO_2 gas into the aquifer waters would be slow, and (3) there should be a shift to lighter carbon-13 values instead of the heavier measured values (average carbon-13 of CO_2 gas in rock units overlying the aquifers in the Nevada Test Site area of southern Nevada is −16.5 permil; D.C. Thorstenson, oral commun., 1988). However, exchange of large quantities of dissolved carbon for carbon in calcite could balance the input of some CO_2 gas with light carbon-13.

WHITE RIVER FLOW SYSTEM

COYOTE SPRING VALLEY + SHEEP RANGE → MUDDY RIVER SPRINGS (PATH 2-1)

A deuterium mass-balance calculation to determine sources of water emanating from Muddy River springs, the terminus of the White River flow system, indicates that water discharging from the springs consists of 62 percent water from the carbonate-rock aquifers of Coyote Spring Valley and 38 percent water from the Sheep Range. Mass-balance calculations based on water-chemistry differences between the initial mixture of Sheep Range and Coyote Spring Valley water and final Muddy River springs water show that gypsum, halite, albite, and K-feldspar dissolve; calcite and chalcedony precipitate; and kaolinite forms (table 14). Dolomite dissolves if cation exchange occurs, otherwise dolomite does not dissolve or precipitate. The average chemistry of water at Muddy River springs used for mass balance calculations is from Big Muddy Spring. Big Muddy Spring is the largest spring in the Muddy River springs area with complete chemical and isotopic data and is the only spring of the Muddy River springs with carbon isotope data (appendix B).

Gypsum and halite dissolution are used as sources of SO_4, Cl, and Na in this model because seven wells drilled in the area (Berger and others, 1988) indicate that the Tertiary aquitard of the Ash Meadows flow system does not exist in this part of the White River flow system. The wells generally penetrate about 20 to 850 ft of unconsolidated sands and gravels which directly overlie the carbonate-rock aquifers. In addition, samples from a well about 7 mi west of and upgradient from Muddy River springs contain gypsum (Berger and others, 1988). Sulfur isotope data indicate that dissolved sulfate in Muddy River springs and Coyote Spring Valley water samples ($\delta^{34}S$ of 13.4 and 13.0 permil, respectively) could be derived from dissolution of gypsum in marine carbonates of Permian to lower Triassic age (Claypool and others, 1980). Upper Permian carbonate rock south and east of the Muddy River springs area contain gypsum (Longwell and others, 1965), so although gypsum has not been reported in exposed Permian rock in the Muddy River springs-Coyote Spring Valley area, gypsum may be present in these rocks at depth. A more likely source of gypsum in this area are the Tertiary Muddy Creek and Horse Springs Formation units, which overlie the carbonate-rock aquifers west of Muddy River springs (Longwell and others, 1965). These formations contain evaporite deposits, which include gypsum. A sample of gypsum from the Muddy Creek Formation in southern Nevada has a $\delta^{34}S$ of 14.0 permil, which is similar to that of sulfate dissolved in Muddy River springs and Coyote Spring Valley water samples. Evaporite deposits also probably contain some NaCl in addition to the gypsum.

Cation exchange of calcium and magnesium in the water for sodium in clay minerals also may be a source of sodium in the water and can be used to replace albite in the mass-balance reaction model. Clinoptilolite also may be a source of potassium and can be used in place of K-feldspar in the mass-balance reaction model. No thermodynamic data were available for clinoptilolite, so the mineral saturation state and, therefore, the thermodynamic feasibility of this reaction could not be determined.

In the mass-balance calculation that includes exchange of calcium and magnesium in the water for sodium in clay minerals, 0.12 mmol/L of carbon dioxide gas is lost from the water. This loss is most likely from degassing of the water in the spring area. Water from most of the springs and wells tapping the carbonate-rock aquifers contain gas bubbles, and because water supplying the springs generally flows upward from the rocks through unconsolidated deposits before reaching land surface, spring waters may loose some of their CO_2 gas before being sampled.

The mass-balance reaction models are thermodynamically feasible because gypsum, halite, albite, and K-feldspar are all below saturation in Coyote Spring Valley or Sheep Range samples (table 15), so they would dissolve. Calcite and chalcedony are at or above saturation, so they would precipitate. Kaolinite is above saturation in the water and, thus, could precipitate but more likely is formed by incongruent dissolution. Although dolomite is at saturation in the water, it would dissolve because of dedolomitization. Dedolomitization is the irreversible dissolution of gypsum that results in the precipitation of calcite. Calcite precipitation causes the pH to decrease and thus dolomite to dissolve (Back and others, 1983).

Mixing of 62 percent Coyote Spring Valley water with 38 percent Sheep Range water, along with the mass transfer of constituents in the modeled geochemical reactions, can produce Big Muddy Spring water. These processes are mineralogically and thermodynamically possible using minerals identified in the aquifers and assuming that either CO_2 in recharge-area waters is higher than measured, or calcium and magnesium in the water is exchanging for sodium in clay minerals and CO_2 is exsolving in the Muddy River springs area (table 14).

WHITE RIVER + SOUTHERN MEADOW VALLEY WASH → COYOTE SPRING VALLEY (PATH 3-2A)

A deuterium mass-balance calculation to determine the sources of water in the carbonate-rock aquifers of Coyote Spring Valley indicates that 64 percent of the water is from the White River flow system and 36 percent of the water is from southern Meadow Valley Wash. Mass-balance calculations based on water chemistry for the initial mixture of White River flow system and southern Meadow Valley Wash water and final Coyote Spring Valley water indicate dolomite, gypsum, halite, albite, and K-feldspar dissolve; calcite and chalcedony precipitate; and kaolinite forms (table 14).

In addition, the concentration of CO_2 in the recharge area waters would have to be higher than measured. Calcium and magnesium exchange can be substituted for albite as a sodium source, and clinoptilolite can be substituted for K-feldspar as a potassium source, in the mass-balance reaction models.

Thermodynamic calculations show that the mass-balance reaction models are feasible (table 15), with the exception of the dissolution of K-feldspar, which is at saturation in the waters. Therefore, the small mass transfer of potassium (0.12 mmol/L) is more likely due to the dissolution of clinoptilolite, or perhaps, exchange with calcium and magnesium in the water.

Mixing of 64 percent White River flow system water with 36 percent southern Meadow Valley Wash water, along with the mass transfer of constituents in the modeled geochemical reactions, can produce the water in the carbonate-rock aquifers of Coyote Spring Valley. These processes are mineralogically and thermodynamically possible using minerals identified in the aquifers and assuming that CO_2 in recharge areas is higher than measured.

ASH MEADOWS FLOW SYSTEM

ARMY WELL 1 → ASH MEADOWS (PATH 5-4)

Deuterium data for the Ash Meadows flow system show that water discharging at Ash Meadows springs, the terminus of the Ash Meadows flow system, has the same deuterium composition as water from Army Well 1, the first carbonate-rock well upgradient from the springs that has deuterium, carbon isotope, and chemical data. Mass-balance calculations for water chemistry, assuming the only source of sulfate and the major source of sodium is leakage from the Tertiary aquitard, for water flowing from Army Well 1 to Ash Meadows springs, shows that calcite, albite, K-feldspar, and clinoptilolite dissolve; chalcedony precipitates; and kaolinite forms (table 14). Calcite is at saturation in the waters and should precipitate, however because of changes in temperature and pressure along deep and tortuous flow paths (Winograd and Pearson, 1976), calcite is assumed to dissolve and precipitate. This calcite dissolution and precipitation is indicated by the carbon-13 composition of the waters and results of mass-balance models and is discussed in the section titled "Ground-Water Age."

A small amount of dolomite dissolves in the mass-balance reaction model involving cation exchange, and 0.12 mmol/L of dolomite precipitates in the other two reaction models. The improbability of dolomite precipitating from these waters makes these two reaction models unlikely. However, the model with dolomite dissolution is also unlikely because both waters are saturated with respect to dolomite. Therefore, the 0.06 mmol/L of dolomite dissolution may represent (1) analytical error, (2) dedolomitization (if not all the sulfate is from leakage, but a small amount of gypsum is present to dissolve), or (3) temperature and pressure differences—although dolomite would not precipitate, these differences could result in water becoming understated with dolomite along parts of the flow path.

The source of all the sulfate and most of the sodium added to the water was assumed to be leakage from the Tertiary aquitard, as proposed by Winograd and Thordarson (1975). A sample from the upper part of the saturated Tertiary aquitard overlying the carbonate-rock aquifers in Mercury Valley (pl. 1) contained 3,600 mg/L sulfate and 1,230 mg/L sodium, assuming 5 percent of the reported sodium plus potassium is potassium (Winograd and Thordarson, 1975, table 11). The absence of gypsum or anhydrite in the Paleozoic carbonate-rock aquifers in this part of the Ash Meadows flow system and the extremely low sulfur concentration in two limestone, two dolomite, and three shale samples (table 10) also indicate the main source of sulfate is probably a water of higher sulfate concentration, rather than a result of gypsum dissolution. Most of the increase in sodium (70 percent) also is assumed to be leakage from the Tertiary aquitard. The remaining increase in sodium is assumed to be from NaCl and, thus, is equal to the increase in chloride. The NaCl may be from connate sea water trapped in the marine sediments that has been released by fracturing or from dissolution of NaCl in shales present in the carbonate-rock aquifers. A sample from the Tertiary aquitard contained only 35 mg/L chloride (Winograd and Thordarson, 1975, table 11), so unless chloride is high in other parts of the Tertiary aquitard this is not a likely source of the increase in chloride. The discharge-weighted average chemistry for the Ash Meadows springs (Winograd and Pearson, 1976, p. 1130–1131) was used for the mass-balance calculations. Thermodynamic calculations show that the mass-balance reaction models are thermodynamically feasible (table 15).

As previously noted in the section titled "Flow System Delineation on the Basis of Deuterium," water from Amargosa Tracer Well 2—about halfway between Army Well 1 and Ash Meadows springs—lacks deuterium data, but the oxygen-18 composition of this water is similar to Ash Meadows springs and probably represents water flowing to the springs. Chemically, this water is similar to, although slightly more dilute than, Ash Meadows springs water. This water could evolve into Ash Meadows springs water by the same processes as those determined for the evolution of Army Well 1 water, but with much smaller mass transfers of chemical constituents. Mass transfers of constituents in water are all calculated to be less than 0.10 mmol/L between Amargosa Tracer Well 2 and Ash Meadows springs, thus Amargosa Tracer Well 2 is not included as a separate site in geochemical calculations.

Water from Army Well 1 can evolve chemically into Ash Meadows water if calcium and magnesium are exchanged for sodium and CO_2 gas enters the water between the sample sites. The source of CO_2 gas would be a water that is isotopically and chemically similar to that of Army Well 1 water but contains a higher CO_2 concentration. Alternatively, Army Well 1 water may not be entirely representative of water in this part of the flow system that discharges at Ash Meadows; instead, a water containing higher CO_2 concentrations, that is a mixture of Spring Mountains and White River flow system water, may be evolving into Ash Meadows water.

SPRING MOUNTAINS + WHITE RIVER → ASH MEADOWS (PATH 6-4)

A deuterium mass-balance calculation to determine sources of water emanating from Ash Meadows springs (Army Well 1 water is isotopically similar to water emanating from springs at Ash Meadows) indicates that water discharging from the springs consists of 60 percent Spring Mountains water and 40 percent White River flow system water. Mass-balance calculations based on water chemistry of the initial mixture of Spring Mountains and White River flow system waters and final Ash Meadows springs water indicate that dolomite, halite, albite, and K-feldspar dissolve; calcite and chalcedony precipitate; kaolinite forms; and high sulfate and sodium water leak from the Tertiary aquitard into the carbonate-rock aquifers (table 14). Calcium and magnesium exchange can be substituted for albite as a sodium source, and clinoptilolite can be substituted for K-feldspar as a potassium source in the models. The mass-balance calculations also require that CO_2 concentration be higher than the average measured value in recharge-area waters. Dolomite can dissolve because most of the waters in the Spring Mountains are undersaturated with respect to dolomite (table 15) and gypsum also is present in parts of the Spring Mountains, so dedolomitization also may occur.

Mixing 60 percent Spring Mountains water with 40 percent White River flow system water, along with the mass transfer of constituents in the modeled geochemical reactions, can produce the water discharging at Ash Meadows springs. These processes are mineralogically and thermodynamically possible using minerals identified in the aquifer and assuming that CO_2 in recharge areas is higher than measured.

The mixture of Spring Mountains and White River flow system waters also can produce water of the chemical composition measured at Army Well 1, using the same mass-balance reaction models (path 6-5 in table 14). As discussed in the section titled "Stable Isotopes of Ground Water," Army Well 1 water is isotopically similar to Ash Meadows spring water and probably represents mixed Spring Mountains and White River flow system water. Thus, the mass transfers for the intermediate step between the mixed water and Ash Meadows spring water is omitted in determining ages and flow rates in later sections of this report.

Indian Springs water also can be used as the input chemistry for the Spring Mountains water. Similar models result; however, because Indian Springs water contains higher ion concentrations than the more dilute samples in the Spring Mountains, the mass transfers are less for models that use Indian Spring water chemistry to represent Spring Mountains water chemistry.

COLD CREEK SPRING → INDIAN SPRINGS (PATH 8-7)

Cold Creek Spring is about 12 mi directly upgradient from Indian Springs in the Spring Mountains (fig. 17), is isotopically similar to Indian Springs, and is one of two large springs with long-term records (20 years) of chemical and isotopic data, thus it is used as the Spring Mountains recharge water that flows to Indian Springs. Mass-balance calculations, based on water chemistry, between these waters shows that gypsum, halite, albite, K-feldspar, and dolomite would dissolve; calcite and chalcedony (except if cation exchange is involved) would precipitate; kaolinite would form; and CO_2 would exsolve (table 14). Gypsum and halite, which are used as the sources for the small increases in sulfate and chloride (0.08 mmol/L), could be in Permian rock that the water flows through or in the Muddy Creek Formation in the area of spring discharge. In the area between Cold Creek Spring and Indian Springs, the existence of a water with high sulfate and chloride concentrations is unlikely. Calcium and magnesium exchange can be substituted for albite, as a sodium source, and clinoptilolite can be substituted for K-feldspar, as a potassium source, in the reaction models. The reaction models are thermodynamically feasible (table 15) if a small amount of chalcedony can either dissolve or precipitate, because both waters are near saturation with respect to chalcedony.

The chemical evolution of Cold Creek Spring water to Indian Springs water is mineralogically and thermodynamically feasible using known minerals identified in the aquifers and assuming CO_2 is exsolving in the Indian Springs and Cold Creek Spring areas. The amount of CO_2 outgassing at Indian Springs and Cold

Creek Spring, calculated using the computer program PHREEQE (Parkhurst and others, 1980) and the method of Pearson and others (1978), was 0.16 and 0.23 mmol/L, respectively. The total CO_2 outgassed, 0.39 mmol/L, is similar to the value from mass-balance calculations, 0.54 mmol/L.

SPRING MOUNTAINS + TEST WELL 3 → ASH MEADOWS (PATH 11-4)

Hydrologic, geologic, deuterium, and chemical data indicate the most likely sources of water discharging at Ash Meadows springs are the Spring Mountains and Pahranagat Valley. Test Well 3 (a well completed in carbonate rock east of Frenchman Flat) is along the flow path between Pahranagat Valley and Ash Meadows (pl. 1, fig. 17), and water from this area may be mixing with Spring Mountains water to produce the water observed at Ash Meadows springs, or this water may already be a mixture of Pahranagat Valley and Spring Mountains waters. However, no deuterium data exist for Test Well 3. Mass-balance calculations based on water chemistry between Pahranagat Valley and Test Well 3 show that Test Well 3 water could evolve chemically from Pahranagat Valley water (path 10-9 in table 14). The mass-balance calculations assume leakage from the Tertiary aquitard, and the model that includes cation exchange has 0.13 mmol/L of calcite dissolving (without leakage calcite would precipitate). The initial and final waters are both saturated with respect to calcite (table 15), but as stated previously, calcite may dissolve and precipitate because of temperature and pressure differences along flow paths in the carbonate-rock aquifers. Therefore, Test Well 3 water could have a deuterium composition similar to that of Pahranagat Valley water. Subsequently, a mixture of 40 percent Test Well 3 water and 60 percent Spring Mountains water would be needed to produce Ash Meadows spring water. Mass-balance calculations for a mixture of 40 percent Test Well 3 water plus 60 percent Spring Mountains water show the chemistry observed at Ash Meadows could evolve chemically from this mixed water (path 11-4 in table 14, table 15).

Winograd and Thordarson (1975) noted that the chemical composition of the water at Test Well 3 is similar to that of water discharging at Ash Meadows springs and may represent already mixed water that changes little chemically between the well and Ash Meadows. This scenario, however, would require (1) water in Test Well 3 be isotopically similar to Ash Meadows water (an unknown); (2) some sodium be removed from Test Well 3 water before reaching Ash Meadows, or a more dilute water with the same isotopic composition as water from Test Well 3 be mixed with Test Well 3 water to achieve the chemistry at Ash Meadows; (3) water flowing from Test Well 3 to Ash Meadows bypass Army Well 1 (as indicated by their vast differences in chemistry), or Army Well 1 water does not represent most water flowing to Ash Meadows in this area, although it has a similar deuterium composition; and (4) the isotopic composition of Army Well 1 water be produced by some process other than mixing of Spring Mountains and Pahranagat Valley waters upgradient from the well. If Test Well 3 water has an isotopic composition similar to Ash Meadows springs water, then the most likely sources of Test Well 3 water would still be a mixture of 40 percent Pahranagat Valley water and 60 percent Spring Mountains water to obtain a deuterium composition similar to Ash Meadows springs. Mass-balance calculations based on water chemistry, using 40 percent Pahranagat Valley water and 60 percent Spring Mountains water to produce Test Well 3 water, are thermodynamically feasible (path 6-9 in table 14, table 15).

As stated in (2) of the preceding paragraph, Test Well 3 water may mix with a more dilute water of similar isotopic composition to produce the water discharging at Ash Meadows. Army Well 1 water could be such a water, but only if Army Well 1 water is entirely from the Spring Mountains and is not a mixture of Spring Mountains and Pahranagat Valley waters. Then the deuterium composition of Army Well 1 water would represent a 4 permil shift in the average deuterium composition of Spring Mountains water.

SPRING MOUNTAINS + YUCCA FLAT → ASH MEADOWS

Another less likely scenario for the origin of Ash Meadows spring water (see section titled "Geologic Framework") is the mixing of water from south Yucca Flat, represented by water from Well C-1, which is completed in carbonate rock, with Spring Mountains water. Deuterium data indicate that a mixture of 33 percent Yucca Flat water and 67 percent Spring Mountains water could produce the deuterium composition of water sampled at Ash Meadows springs. However, on the basis of chemical mass-balance calculations using the average chemistry of water from Well C-1 and Indian Springs, none of the three reaction models is thermodynamically feasible.

LAS VEGAS VALLEY

Recharge to Las Vegas Valley from the Spring Mountains is from two different areas within the mountains; each area has a unique isotopic signature. Northern Las Vegas Valley receives an isotopically light component of recharge from the central part of the Spring Mountains, and southwestern Las Vegas Valley receives an isotopically heavier component of recharge from the southern part of the Spring Mountains.

In addition, northern Las Vegas Valley may receive as much as 15 percent of its water from the Sheep Range (see section titled "Flow System Delineation on the Basis of Deuterium").

SPRING MOUNTAINS → NORTH LAS VEGAS VALLEY (PATH 13-12)

Mass transfers based on water chemistry were calculated for the central part of the Spring Mountains to the Tule Spring area of northern Las Vegas Valley (table 14). Mass-balance calculations show that gypsum, halite, albite, K-feldspar, and dolomite dissolve; calcite and chalcedony precipitate (unless cation exchange is involved); kaolinite forms; and CO_2 exsolves. Calcium and magnesium exchange can be substituted for albite as a sodium source, and clinoptilolite can be substituted for K-feldspar as a potassium source, in the reaction models. The mass-balance reaction models are thermodynamically feasible (table 15) if the ground water is degassing in the Tule Spring area and a small amount of chalcedony either dissolves, or precipitates, because both waters are near saturation with respect to chalcedony.

The chemical evolution of Spring Mountains water to Tule Spring area water in north Las Vegas Valley is mineralogically and thermodynamically feasible (table 15) using known minerals in the aquifers and assuming that CO_2 is exsolving in the Tule Spring area. The small amount of CO_2 outgassing (0.07 to 0.16 mmol/L) is within the range of calculated values for samples within the Tule Spring area.

SPRING MOUNTAINS + SHEEP RANGE → NORTH LAS VEGAS VALLEY (PATH 17-12)

Mass-balance calculations based on water chemistry for a mixture of 85 percent Spring Mountains water and 15 percent Sheep Range water (represented by samples from the Corn Creek Springs area) produce north Las Vegas Valley water (represented by samples from the Tule Spring area). Results of these calculations are similar to results of the reaction model using solely Spring Mountains water (tables 14 and 15).

SOUTHERN SPRING MOUNTAINS → SOUTHWEST LAS VEGAS VALLEY (PATH 15-14)

Mass-balance calculations based on water chemistry from the southern part of the Spring Mountains to southwestern Las Vegas Valley show that gypsum, halite, albite, K-feldspar, and dolomite dissolve; calcite and chalcedony precipitate; kaolinite forms; and CO_2 exsolves (table 14). Calcium and magnesium exchange can be substituted for albite as a sodium source, and clinoptilolite can be substituted for K-feldspar as a potassium source, in the reaction models. The chemical evolution of southern Spring Mountains water to southwestern Las Vegas Valley water is mineralogically and thermodynamically (table 15) feasible using known minerals in the aquifers.

PAHRUMP VALLEY

SPRING MOUNTAINS → PAHRUMP VALLEY (PATH 13-16)

Ground water in Pahrump Valley is isotopically similar to average recharge water in the central part of the Spring Mountains. Thus, ground water in Pahrump Valley should evolve chemically from recharge water in the Spring Mountains. Mass transfers based on water chemistry were calculated for water flowing from the central part of the Spring Mountains to Pahrump Valley (table 14). Mass-balance calculations show that gypsum, halite, albite, K-feldspar, and dolomite dissolve; calcite and chalcedony precipitate (except if cation exchange occurs, then chalcedony dissolves); kaolinite forms; and CO_2 exsolves. Calcium and magnesium exchange can be substituted for albite as a sodium source, and clinoptilolite can be substituted for K-feldspar as a potassium source, in the reaction models. The mass-balance reaction models are thermodynamically feasible (table 15) using known minerals in the aquifers, if chalcedony either dissolves or precipitates, because the waters are close to saturation with respect to chalcedony, and water degases CO_2 in the old spring areas of Pahrump Valley. The small amount of CO_2 outgassing is within the range of values calculated for the two samples in Pahrump Valley.

SUMMARY OF MASS-BALANCE REACTION MODELS

Chemical mass-balance reaction models can be constructed for flow paths and mixing of water from different areas in southern Nevada that were determined by deuterium mass-balance calculations. The models are based on water chemistry and mineralogy observed in the flow system and leakage from the Tertiary aquitard. The models are thermodynamically feasible, except for flow from Yucca Flat to Ash Meadows springs. Geologic information in this area also indicates that flow from Yucca Flat to Ash Meadows springs is small (Winograd and Thordarson, 1975). The deuterium-derived flow paths and mixing of water from different areas, as shown in figure 24, are supported by chemical mass-balance calculations.

GROUND-WATER AGE

Ground-water age provides another constraint on flow paths and mixing in southern Nevada. Ground-water age must increase along a flow path, or be an average age between the ages of two initial waters that mix for flow paths or mixtures of waters to be probable. The age of mixed water has to increase along a flow path; thus, the mixture has to be older than the youngest mixing water, and the final water may be older than both mixing waters if travel times after mixing are long. Ground-water ages from several thousand years to about 30,000 years can be determined using carbon isotopes. Waters less than about 60 years contain measurable tritium (Fritz and Fontes, 1980).

Ground-water age, adjusted for the mass transfer of carbon into and out of the water, can be calculated from carbon isotope compositions of the water if (1) the carbon isotope composition of the recharge water when the water becomes isolated from atmospheric and soil-zone CO_2 gas is known, (2) the mass and sources of carbon added to or removed from the water are known, (3) the isotope composition of the sources of carbon are known, and (4) the fractionation of the isotopes during removal of carbon from the water is known (Wigley and others, 1978). The biggest sources of error in calculating ground-water age using this approach are (1) the estimation of the starting, or original, carbon-14 composition of the recharge waters (A_0, pre-nuclear-detonation carbon-14), (2) the accuracy of the mass-transfer reaction models in describing the input and output of carbon to the water along the flow path, and (3) the isotope composition of carbon added to the water.

The carbon-14 composition of a recharge water (A_0) can be calculated assuming a system is either closed or open to soil-zone CO_2 gas. Ground-water ages determined for this study assumed that ground water in recharge areas became closed to soil-zone gas. Initial carbon-14 compositions (A_0) were calculated using a modified Tamers approach (Tamers, 1967, 1975; Tamers and Scharpenseel, 1970) with the computer program NETPATH (Plummer and others, 1991). The Tamers calculation used in this study is a mass-balance calculation including calcite, dolomite, and CO_2 gas, assuming the carbon in calcite and dolomite contains 0 pmc (percent modern carbon) and the CO_2 gas contains 100 pmc. These calculated carbon-14 values for A_0 are shown in table 18. The assumption that the system is closed to CO_2 gas in recharge areas may result in adjusted carbon-14 ages younger than actual ages, because almost all water samples in recharge areas are saturated with respect to calcite and most are saturated, or close to saturation, with respect to dolomite (table 15), yet waters in the two main recharge areas—the central Spring Mountains and Sheep Range—contain 76 to 100 pmc (table 16). These waters contain post-nuclear-detonation tritium (table 16) and, thus, also contain carbon-14 compositions greater than the pre-nuclear-detonation level of 100 pmc. Therefore, A_0 of water recharged before nuclear detonations, which includes all water outside the recharge areas containing little or no tritium, would have values smaller than those measured in recharge areas after above-ground nuclear detonations, so the modified Tamers values probably are reasonable estimates of A_0. If water was recharged under totally open system conditions, then the closed system ages could be in error by about one half-life of carbon-14 (the water would be 5,730 years older than calculated). However, the system is not totally open, as indicated by carbon-14, which is 76 to 100 pmc in water in these areas; as compared to CO_2 gas in the soil zone of Yucca Mountain, which is 110 pmc (D.C. Thorstenson, U.S. Geological Survey, oral commun., 1988); and carbon-14 in plant tissue, which reached a maximum of about 180 pmc (Tamers and Scharpenseel, 1970, p. 243). Thus, the maximum error of the adjusted ages due to an open, rather than closed, system has to be less than 5,730 years. The amount of possible error depends on whether the system is closed, partially closed, or open to CO_2 gas during recharge.

Water from a well and springs in the foothills of the southern part of the Spring Mountains, in southwest Las Vegas Valley (Sandstone Bluffs area), was determined to represent recharge to southwest Las Vegas Valley on the basis of hydrologic, geologic, deuterium, and water-chemistry data. The average carbon-13 composition in the well and spring-water samples is –10.4 permil, and average carbon-14 composition is 50.8 pmc (table 16, fig. 25). Two of the four samples contain tritium, indicating that the waters are less than 60 years old or are old waters that have been mixed with young water containing tritium. If the waters are not a mixture including young water, little decay of carbon-14 has occurred. The low, or absent, tritium concentration indicates that these waters are probably pre-nuclear-detonation waters, and thus their carbon-14 compositions should represent A_0 values. Although the well is completed in, and springs discharge from, sandstone units, the source of this water probably is recharge in carbonate rock of the southern Spring Mountains. Therefore, their A_0 values should be similar to those of water in the central part of the Spring Mountains. The similarity of carbon-14 compositions of these waters to calculated values for water in the central part of the Spring Mountains and Sheep Range (table 18; 53–55 pmc) indicates that the assumption of a closed system in the calculations of A_0 for recharge waters in southern Nevada is probably reasonable.Mass-balance calculations presented in the section titled "Water Chemistry" account for the carbon inputs and outputs of waters along flow paths (table 14). The mass-balance reaction model including the exchange of calcium and magnesium in the water for sodium in clay minerals (reaction model 3 in table 14) was used for calculating adjusted ages.

TABLE 16.—*Carbon-13, carbon-14, and tritium compositions of ground water in carbonate-rock aquifers of southern Nevada*

[Symbol: --, data not available; <, less than]

Site	Number of samples[1]	Delta carbon-13 (permil)	Carbon-14 (percent modern carbon)	Tritium (picocuries per liter)	Source [2]
Central part of Spring Mountains					
Sky Mountain well	1	−10.1	84.1	34	1
Summer Homes well	1	−10.8	91.1	89	1
Willow Spring	1	−9.6	79.2	57	1
Deer Creek Spring 1	2	−8.4	--	78	1
Deer Creek Spring 2	1	−9.6	100.0	73	1
Highway maint well	1	−10.1	83.0	87	2
Rainbow Spring 2	1	−9.8	91.7	98	2
Trout Spring	6	−7.9	90.8(1)	257 (3)	1,3,4
Cold Creek Spring	6	−9.5	76.0(4)	92 (4)	1,3,4
Southern part of Spring Mountains					
BLM Visitors Center well	1	−9.3	46.0	9.0	1
Sandstone Spring	2	−10.6	49.8	<15 (1)	1
Red Spring	2	−10.5	62.4	3.0(1)	1
White Rock Spring	2	−11.2	44.8(1)	<2.0	1
Sheep Range					
Wiregrass Spring	1	−10.2	96.8	90	1
Mormon Well Spring	1	−9.9	--	--	4
Coyote Spring Valley					
MX well	1	--	7.6	<2.0	4
VF-2 Well	1	−6.1	7.0	<1.0	4
Pahranagat Valley					
Crystal Spring	3	−7.0	7.8(2)	<2.0(1)	3,4
Ash Spring	2	−6.7	6.3(3)	0.0	3,4
Muddy River springs					
Big Muddy Spring	1	−6.0	6.7	<1.0	4
Indian Springs Valley					
Indian Springs	3	−7.6	8.3	<2.0(2)	1,3
Southern Yucca Flat					
C-1 Well	1	−3.8	0.8	<0.3	3,5
Mercury Valley					
Army Well 1	2	−5.6	2.8(1)	0.6	3
Ash Meadows					
Average of springs[3]	11	−4.9	4.7(13)	0.3(13)	3
Corn Creek Springs area					
Corn Creek Spring	3	−7.7	13.9(1)	<1.0(2)	1,3
Shown well	1	−8.8	20.4	<1.0	1
Young well	1	−6.7	13.7	<1.0	1
Brooks well	1	−6.9	35.7	<15	1
Tule Spring area - North Las Vegas Valley					
Holland well	1	−6.2	11.3	--	2
Mulder well	1	−7.2	23.8	<15	1
Tule Spring well	1	−7.0	13.9	<15	1
Martin well	1	−6.7	1.9	--	1
Racel well	1	−6.7	7	--	4
Southwest Las Vegas Valley					
Stocks Mill well	1	−7.9	11.0	<15	1
Pahrump Valley					
Manse well	1	−8.3	46.9	<2.0	1
Pahrump Spring well	1	−7.6	26.0	<1.0	1

[1] Where average is based on fewer or more samples than shown in this column, exact number is listed in parentheses after average constituent value.

[2] Sources of data:

1 Thomas and others (1991).
2 Schulke (1987, p. 78).
3 Winograd and Pearson (1976).
4 Unpublished data, U.S. Geological Survey, Carson City, Nev.
5 Boughton (1986, p. 84).

[3] Discharge-weighted average, for springs at Ash Meadows (Winograd and Pearson, 1976, table 2).

Sources of carbon in the carbonate-rock aquifers outside of the recharge areas are carbonate minerals (calcite and dolomite), organic matter, and CO_2 gas in rock units. Atmospheric and soil-zone CO_2 gas are not considered sources of carbon outside the recharge areas because depths to water in the carbonate-rock aquifers are generally several hundred to as much as 2,000 ft below land surface. Mass-balance calculations (see section titled "Water Chemistry") indicate that dolomite dissolves by dedolomitization, and in some areas because dolomite is undersaturated in the water. Calcite also dissolves and subsequently precipitates, as indicated by the lighter calculated than measured carbon-13 compositions of the water (shown as cycled carbon in table 18). Calcite dissolution and precipitation are caused by temperature and pressure changes over tortuous flow paths (Winograd and Pearson, 1976) and possibly increased surface area of calcite due to grinding as the result of faulting. Oxidation of organic matter, which would produce CO_2 gas, and CO_2 gas in rock units overlying the aquifers do not contribute significant amounts of carbon to the water (as discussed in the section titled "Water Chemistry").

Carbon isotope fractionation factors between precipitating calcite or exsolving CO_2 gas and a carbonate solution at varying temperatures and pH's can be measured (Deines and others, 1974; Mook, 1980).

EXPLANATION

☐ **Basin-fill deposits**—In areas underlain by thick, laterally continuous carbonate rock

▧ **Basin-fill deposits**—In areas outside thick, laterally continuous carbonate rock

▒ **Carbonate rock**—In areas of thick, laterally continuous carbonate rock

▨ **Carbonate rock**—In areas outside thick, laterally continuous carbonate rock

■ **Noncarbonate rock**—In areas of thick, laterally continuous carbonate rock

▩ **Noncarbonate rock**—In areas outside thick, laterally continuous carbonate rock

- - - - **Boundary of laterally continuous carbonate rock**

➡ **Generalized direction of ground-water flow in carbonate-rock aquifer**

● **Spring**—Emanates from carbonate-rock aquifers

● **Well**—Source of water is carbonate-rock aquifers

-4.9/4.7 **Average carbon-13, in permil, and average carbon-14, in percent modern carbon**

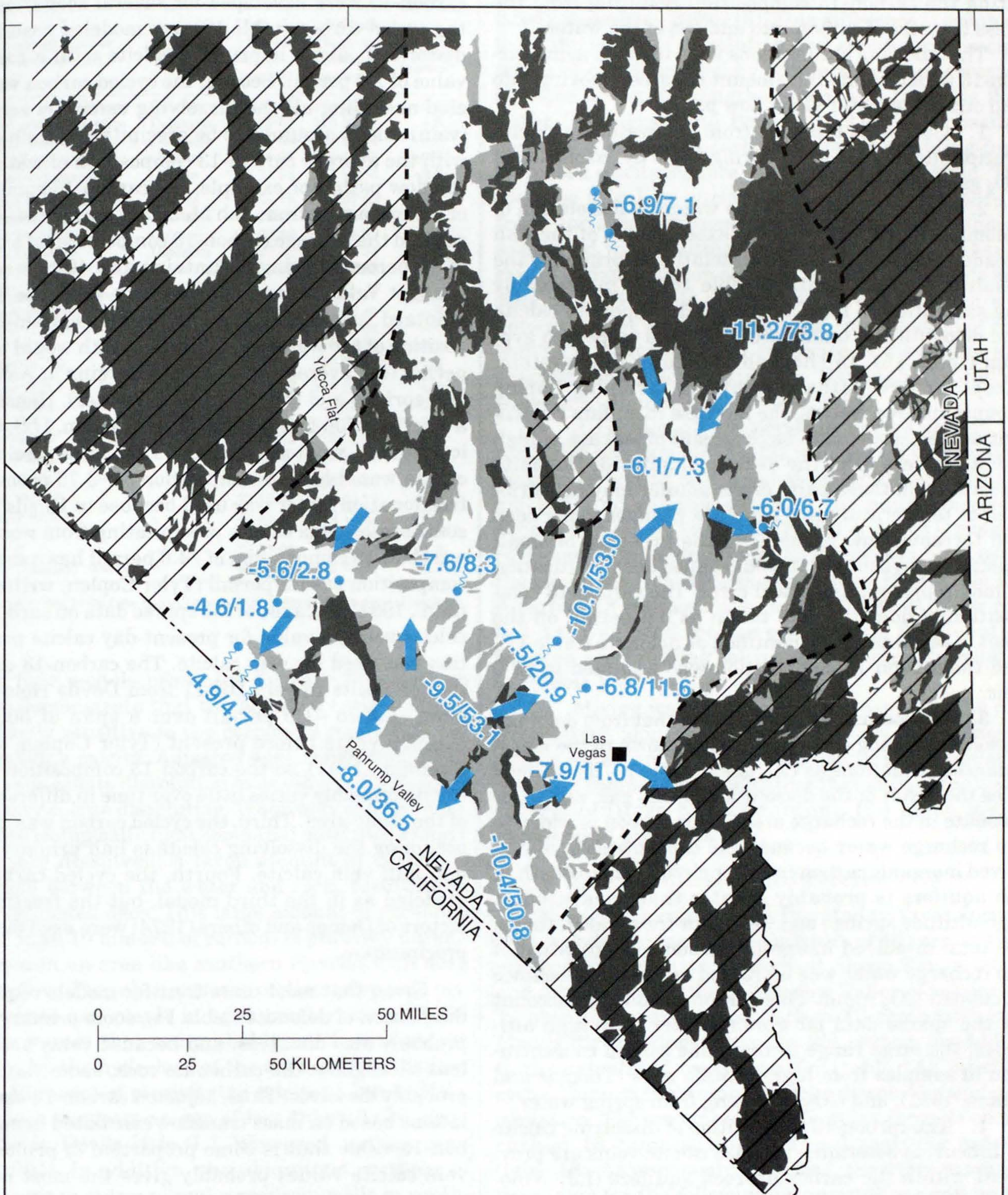

FIGURE 25.—Average carbon-13 and carbon-14 compositions of ground water in carbonate-rock aquifers of southern Nevada. See figure 18 for feature names. Modified from Dettinger (1989, fig. 3).

These fractionations have to be accounted for in calculating the carbon-13 composition resulting from the mass transfer of carbon into and out of the water.

The following assumptions were made in using carbon-13 to determine the amount of carbon moving into and out of the water along flow paths:

1. Carbon is removed from the water by calcite precipitation and in some spring areas by exsolution of CO_2 gas (table 14).

2. Carbon is added to the water by dissolution of dolomite. In all flow systems except parts of the Ash Meadows system, dolomite dissolution is primarily the result of dedolomitization (table 14), as indicated by the saturation of most waters with respect to calcite and dolomite and undersaturated with respect to gypsum (table 15). In the Ash Meadows flow system, except for flow paths and mixing involving Spring Mountains area waters, the increase in sulfate concentration was assumed to be the result of leakage of high sulfate waters from the Tertiary aquitard. Dolomite dissolution in these parts of the aquifer may be attributed to temperature and pressure changes along deep and tortuous flow paths or to the presence of small amounts of gypsum that would cause dedolomitization or may represent analytical error. The carbon-13 composition of dolomite was taken as 0.0 permil on the basis of seven samples identified as dolomite (table 17), and the carbon-14 composition was assumed to be 0 pmc.

3. Carbon input, in addition to that from dolomite dissolution, along some flow paths (shown in the mass-balance calculations as CO_2 gas in table 14) is assumed to be the result of the dissolution of CO_2 gas, calcite, or dolomite in the recharge areas. This carbon is added to the recharge water because the concentration of dissolved inorganic carbon in water circulating to depth in the aquifers is probably greater than measured at high-altitude springs and wells. For these calculations, the total dissolved inorganic carbon concentration of the recharge water was increased by the mass-balance calculated CO_2 input. This carbon is added to account for the sparse data for ground water from high altitudes, the large range in inorganic carbon concentration in samples from high-altitude sites (Thomas and others, 1991), and CO_2 degassing from spring water.

4. The carbon-13 composition of dissolving calcite is difficult to determine because calcite veins are prevalent within the carbonate-rock aquifers (I.J. Winograd, U.S. Geological Survey, oral commun., 1988). Primary calcite in southern Nevada has a carbon-13 composition ranging from –0.5 to +3.7 permil, with an average of 1.2 permil (table 17), whereas secondary (vein) calcite has a carbon-13 composition ranging from about –6 to –1.5 permil (I.J. Winograd, written and oral communs., 1991). Thus, geochemical models using carbon-13 were developed for several scenarios. First, the cycled carbon (table 18) was modeled assuming all dissolving calcite is primary calcite with a carbon-13 value of 1.2 permil. Second, the cycled carbon was modeled assuming all the dissolving calcite is secondary (vein) calcite assumed to be precipitating from a water with the average carbon-13 composition of water along the flow path. For example, the carbon-13 composition of water flowing from Ash Meadows springs is –4.9 permil and the average carbon-13 composition of a mixture of 60 percent Spring Mountains and 40 percent Pahranagat Valley waters plus leakage from the Tertiary aquitard is –8.6 permil, so an average carbon-13 composition of the water along the flow path would be –6.75 permil (the average of water discharging at Ash Meadows springs and the initial mixed water). Hence, using a fractionation factor of 3.0 (Tyler Coplen, U.S. Geological Survey, written commun., 1991), the precipitating calcite would have a composition of –3.75 permil. This fractionation factor was used because in Devils Hole in southern Nevada calcite precipitating from water with a carbon-13 composition of –4.8 permil has a carbon-13 composition of –1.8 permil (Tyler Coplen, written commun., 1991). Because of the sparse data on carbon-13 in calcite veins, a value for present-day calcite precipitation was used for vein calcite. The carbon-13 composition of calcite in vein DH-11 from Devils Hole ranges from –2.8 to –1.5 permil over a span of 50,000 to 500,000 years before present (Tyler Coplen, written commun., 1991), so the carbon-13 composition of vein calcite probably varies little over time in different parts of the study area. Third, the cycled carbon was modeled assuming the dissolving calcite is half primary calcite and half vein calcite. Fourth, the cycled carbon was modeled as in the third model, but the fractionation factors of Deines and others (1974) were used for calcite precipitation.

Given that most mass transfer models require the dissolution of dolomite (table 14), some primary calcite probably also dissolves, and because veins are prevalent throughout the carbonate rock, some vein calcite probably dissolves. Thus, adjusted carbon-14 age calculations based on mass transfers calculated using a carbon-13 value that is some proportion of primary and vein-calcite values probably gives the most realistic adjusted age (model 3 in table 18). In addition, if dissolving calcite was entirely vein calcite, then the average age of water discharging from Ash Meadows and Muddy River springs would be modern, which is unlikely given flow distances of tens of miles and the absence of tritium in the water (tables 16, 18).

TABLE 17.—*Carbon-13 composition of calcite and dolomite in southern Nevada*

Site[1]	Geologic unit or site sampled	Rock type	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Delta carbon-13 (permil)
1	Yellowpine Limestone	Calcite	36 43 57	114 46 05	+2.5
2	Ely Springs Dolomite	Dolomite	36 29 57	115 17 20	+.6
3	Nopah Dolomite	Dolomite	36 34 06	115 21 34	+.4
4	Dawn Limestone	Calcite	36 56 00	114 18 50	+2.6
5	Arcturus Limestone	Calcite	39 21 32	115 20 54	+2.7
6	Upper Cambrian	Calcite	38 59 18	114 57 25	+3.4
7	Simonson Dolomite	Dolomite	39 03 36	114 56 47	+.4
8	Sevy Dolomite	Dolomite	36 03 36	114 56 53	–1.4
9	Sevy Dolomite	Dolomite	37 27 49	115 11 34	–.1
10	Sevy Dolomite (?)	Dolomite	37 31 53	115 13 58	–.1
11	Laketown Dolomite	Dolomite	39 03 36	114 57 18	+.2
12	Lower Pogonip Group	Calcite	38 59 28	114 56 41	–.5
13	Upper Cambrian	Calcite	38 59 21	114 57 04	–.4
14	Test Well D	--	37 04 28	116 04 30	+3.7
15	Well U3CN5	--	37 03 35	116 01 30	+.4
16	Army Well 1	--	36 35 30	116 02 14	+.6
17	Test Well 4	--	36 34 54	115 50 08	+2.1
18	Test Well 2	--	37 09 58	116 05 15	+2.3
19	Test Well F	--	36 45 34	116 06 59	+.9
20	Test Well 3	--	36 48 30	115 51 26	+1.0
21	Test Well 1	--	37 09 29	116 13 23	+.3
22	Well UE15D	--	37 12 33	116 02 29	+2.3
23	Tracer Well 3	--	36 32 12	116 13 47	+2.2
24	Well C-1	--	36 55 07	116 00 34	+.6
25	Well C	--	36 55 08	116 00 35	+1.2
26	Blue Diamond road cut	--	36 02 05	115 23 45	+2.7

[1] Data for sites 14–26 from Hans Claassen (U.S. Geological Survey, oral commun., 1987).

The four models presented for each flow path in table 18 demonstrate that the adjusted carbon-14 age is extremely sensitive to the carbon-13 composition of the dissolving calcite. Adjusted ages range from modern for vein-calcite dissolution to as old as 13,900 years for primary calcite dissolution. When only calcite veins, which are isotopically light compared with the primary calcite, are dissolved, a large amount of carbon is exchanged between the water and rock, resulting in modern adjusted ages. This large amount of exchange (greater than 10 mmol/L of carbon) is probably unrealistic, even in an area like southern Nevada with deep fracture flow. The other extreme of dissolving only primary calcite can result in ages greater than 10,000 years in waters that are probably much younger, as indicated by the hydrology of the area and the distinct oxygen-18 record of precipitated calcite at Devils Hole, Nev., which suggests an age of less than 10,000 years for water in Devils Hole (I.J. Winograd, written commun., 1991). In addition, the ubiquitous presence of calcite veins in outcrops and cores from wells in southern Nevada indicates some vein calcite should be exposed to ground water. Thus, the carbon-13 composition of the dissolving calcite probably is between primary calcite (1.2 permil) and vein calcite (–6.0 to –1.5 permil).

5. The measured carbon-13 composition of the final water is heavier (less negative) than the calculated composition (table 18). This difference is assumed to be the result of dissolution and subsequent precipitation of calcite caused by temperature and pressure changes along the lengthy, deep, and tortuous flow paths (Winograd and Pearson, 1976) and possibly because of increased carbon exchange resulting from increased calcite surface area caused by faulting. The increase in calcite surface area, plus the exposure of fresh surfaces, produced by fracturing and grinding associated with faulting, would increase the exchange rate of carbon between solution and calcite (Mozeto and others, 1984). This additional input of carbon was determined by accounting for the fractionation of carbon-13 in a hypothetical sequence of dissolving and precipitating calcite using the Rayleigh distillation equations (Plummer and others, 1991).

6. All carbon-13 fractionation factors used in the computer program NETPATH were set for the fractionation of carbon-13 during calcite precipitation for models 1–3, and calculated using the equations from Deines and others (1974) for model 4. All CO_2 gas and carbonate solution fractionations were calculated using the equations from Deines and others (1974; table 18). A carbon-13 fractionation factor of 3.01 (6.02 for carbon-14) was used for the equilibrium fractionation between precipitating calcite and a carbonate solution in models 1–3.

7. Mixing waters have similar histories. The carbonate-rock aquifers contain relatively homogeneous mineralogies, and waters in the aquifers are similar in chemical composition, indicating that most of the water in the aquifers has undergone the same major geochemical processes.

WHITE RIVER FLOW SYSTEM

Water discharging from the Muddy River springs area was determined to be a mixture of 40 percent Pahranagat Valley water, 38 percent Sheep Range water, and 22 percent southern Meadow Valley Wash water. To obtain the measured carbon-13 composition of Muddy River springs water (fig. 25), carbon, in addition to the calculated mass transfer (table 14), must be cycled through the water by calcite dissolution and precipitation. The amount of carbon cycled depends on the carbon-13 composition of the dissolving calcite (table 18). As previously discussed, the average traveltime (age) for the following interpretations will be for the model that includes half primary and half secondary (vein) calcite dissolution.

The average traveltime from the three source areas to Muddy River springs was calculated to be 5,000 years (model 3 in table 18). This age was calcu-

TABLE 18.—*Ground-water ages in carbonate-rock aquifers of southern Nevada*

[Ages are corrected for mass transfer of carbon into and out of water, on basis of mass-balance calculations (table 14) and isotope data (table 16). Adjusted ages were calculated using carbon-13 composition of sources and sinks, and taking into account fractionation factors for these processes. Carbon-13 fractionation for exsolution of CO_2 gas is from Deines and others (1974). Carbon-13 fractionation for precipitating calcite for models 1–3 is from Tyler Coplen (U.S. Geological Survey, oral commun., 1991) and for model 4 is from Deines and others (1974). The mass-balance reaction model using exchange of calcium and magnesium in water for sodium in clay minerals was used for age corrections (table 14). Abbreviations and symbols: mmol/L, millimole per liter; pmc, percent modern carbon; >, greater than; -- not determined]

Flow path (flow-path terminus)[1]	Model[2]	Measured initial $\delta^{13}C$ (permil)	Measured final $\delta^{13}C$ (permil)	Calculated final $\delta^{13}C$ (permil)	Cycled carbon (mmol/L)[3]	A_0 initial carbon-14 (pmc)[4]	A_{nd} adjusted carbon-14 (pmc)	Measured carbon-14 (pmc)	Age (years before present)[5]
White River Flow System									
3A-1 (Muddy River springs)[6]	1	–9.0	–6.0	–8.4	2.4	39.2	21.2	6.7	9,500
	2				>10		--		modern
	3				5.2		12.3		5,000
	4			–8.4	4.7		13.6		5,800
2-1 (Muddy River springs)	1	–7.6	–6.0	–7.5	1.7	24.7	16.9	6.7	7,600
	2				>10		--		modern
	3				3.5		12.0		4,800
	4			–7.5	3.0		13.2		5,600
3-2A (Coyote Valley)	1	–8.4	–6.1	–7.5	1.5	31.1	19.8	7.3	8,200
	2				>10		--		modern
	3				3.0		14.8		5,800
	4			–7.4	2.4		16.6		6,800
Ash Meadows Flow System									
6-4 (Ash Meadows springs)	1	–8.4	–4.9	–7.8	3.4	34.1	16.0	4.7	10,100
	2				>10		--		modern
	3				8.4		6.2		2,300
	4			–7.1	6.7		8.6		5,000
8-7 (Indian Springs)	1	–9.5	–7.6	–8.2	0.45	54.9	44.6	8.3	13,900
	2				>10		--		modern
	3				.91		40.3		13,100
	4			–8.1	.63		42.9		13,600
Las Vegas Valley									
13-12 (Tule Spring area)	1	–9.5	–6.8	–8.2	1.0	53.1	34.5	11.6	9,000
	2				>10		--		modern
	3				2.2		25.9		6,600
	4			–8.1	1.7		29.2		7,600
17-12 (Tule Spring area)	1	–9.2	–6.8	–8.3	1.10	48.3	33.0	11.6	8,600
	2				>10		--		modern
	3				2.3		24.8		6,300
	4			–8.2	1.8		27.9		7,300
15-14 (Southwest Las Vegas Valley)	3	–10.4	--7.9	–5.8	--	50.8	21.3	11.0	5,500
	4			–5.7	--		21.3		
									5,500
Pahrump Valley									
13-16 (Pahrump Valley)	1	–9.5	–8.0	–8.4	0.30	53.1	42.4	36.5	1,200
	2				>10		--		modern
	3				0.58		39.7		680
	4			–8.3	0.40		41.4		1,000

---

[1] Flow-path designations are same as those in table 14.

[2] Model 1 uses carbon-13 value for dissolving calcite of average primary calcite, 1.2 permil (table 17). Model 2 uses carbon-13 value for dissolving calcite of vein calcite, taken as carbon-13 composition of calcite that would precipitate from an average water along flow path. Model 3 uses carbon-13 value for dissolving calcite assuming half the calcite is primary calcite (1.2 permil) and half the calcite is vein calcite (carbon-13 of vein calcite is that used in model 2). Model 4 uses carbon-13 values of model 3 for dissolving calcite. Models are described on p. 54.

[3] Additional input of carbon, from calcite dissolution and subsequent precipitation, needed to match measured carbon-13 composition of water.

[4] A_o was calculated using modified Tamers approach with computer model NETPATH (Plummer and others, 1991) in recharge waters that contain measurable tritium and as was determined by mixing for waters containing no tritium and less than 50 pmc. In the case of one mixing water being from a recharge area, A_0 was first calculated for recharge water and then the mixed carbon-14 value was calculated using this value plus measured value for other water.

[5] $\text{Age} = \left(\frac{5,730}{\ln 2}\right)\left(\ln\left(\frac{\text{C-14 (adjusted)}}{\text{C-14 (measured)}}\right)\right)$, where C-14 (adjusted) is the carbon-14 composition calculated from mass-balance reaction model using carbon-14 compositions and fractionations of carbon inputs and outputs assuming no decay and C-14 (measured) is carbon-14 measured in water.

[6] Chemistry of White Rock Spring in volcanic terrain of Nevada Test Site was used to represent carbon-isotope chemistry of southern Meadow Valley Wash water. This water has a carbon-13 composition of -11.2 permil and a carbon-14 composition of 91.0 pmc.

lated by adjusting a volume-weighted average of the carbon-14 composition of the three source waters (A_0 in table 18), using the carbon mass transfers calculated for the mixed water to Muddy River springs, and then comparing the adjusted carbon-14 composition (A_{nd} in table 18) to the measured carbon-14 composition of the spring water. The mass-transfer adjusted carbon-14 composition (12.3 pmc) is larger than the measured composition (6.7 pmc), and the difference is assumed to be due to radioactive decay. The adjusted traveltime (age) was calculated by using the equation:

$$t = \frac{5,730}{\ln 2}\left[\ln \frac{\text{C} - 14\,(\text{adjusted})}{\text{C} - 14\,(\text{measured})}\right], \qquad (2)$$

where t = adjusted carbon-14 traveltime, or age, of the water, in years;

5,730 = half-life of carbon-14, in years;

C-14 (adjusted) = carbon-14 composition adjusted for the mass transfer of carbon on the basis of carbon-13, in percent modern carbon, assuming no decay (A_{nd}); and

C-14 (measured) = carbon-14 composition measured in the water, in percent modern carbon.

An average adjusted carbon-14 age for Ash and Crystal Springs, in Pahranagat Valley, was calculated to be 8,500 years before present. Thus, the total average age of water discharging from Muddy River springs is the average traveltime of the mixed water (5,000 years) plus the volume-weighted average of the starting ages of the source waters, corrected for the carbon-14 change associated with the traveltime from Pahranagat Valley to Muddy River springs (1,100 years). The average age of water discharging from Muddy River springs, therefore, is calculated to be 6,100 years (fig. 26).

The mixing percentages of the three source waters for Muddy River springs was determined by first mixing Sheep Range and Coyote Spring Valley waters to produce Muddy River springs water and then mixing Pahranagat Valley and southern Meadow Valley Wash waters to produce Coyote Spring Valley water (see sections titled "Stable Isotope" and "Water Chemistry"). The average ages of these mixed waters also are included in table 18 because they are used later in the report for calculating flow velocities based on adjusted carbon-14 ages.

ASH MEADOWS FLOW SYSTEM

Water discharging from springs at Ash Meadows was determined to be a mixture of 60 percent Spring Mountains water and 40 percent Pahranagat Valley water. Carbon, in addition to the carbon determined from mass-transfer calculations (table 14), must be cycled through the water by calcite dissolution and precipitation to obtain the measured carbon-13 composition of the spring water (table 18).

The average traveltime from the two sources to Ash Meadows springs was calculated to be 2,300 years by using equation 2 with volume-weighted carbon isotope compositions and correcting for mass transfers of carbon (table 18). Assuming that the adjusted carbon-14 age is 8,500 years for Pahranagat Valley water, the average age of water discharging from Ash Meadows, corrected for the carbon-14 change associated with the traveltime from Pahranagat Valley to Ash Meadows springs is 4,100 years (fig. 26).

LAS VEGAS VALLEY

The average age of ground water in the Tule Spring area of northern Las Vegas Valley was calculated two ways. First, the age was calculated assuming that the central part of the Spring Mountains is the sole source of water for the Tule Spring area. Second, the age was calculated assuming that water in the Tule Spring area is a mixture of 85 percent Spring Mountains water and 15 percent Sheep Range water (Corn Creek Springs area). Carbon, in addition to that determined from mass-transfer calculations (table 14), must be cycled through the water by calcite dissolution and precipitation to obtain the measured carbon-13 composition of the spring water (table 18). The average traveltime for water recharging the Spring Mountains to reach the Tule Spring area is 6,600 years (table 18, fig. 26). Because Spring Mountains water is modern, this also is the average age of Tule Spring area water.

A mass-balance reaction model was made for a mixture of 85 percent Spring Mountains water and 15 percent Sheep Range water to evolve into Tule Spring area water. The model results in an average age of 6,300 years (table 18).

The average age of ground water in southwest Las Vegas Valley was calculated using the average carbon isotope and chemical compositions of water in the Sandstone Bluffs area of the southern part of the Spring Mountains and water from two wells in southwest Las Vegas Valley (tables 14, 16). The average traveltime for water to flow from the Sandstone Bluffs area to the wells is 5,500 years (table 18, fig. 26). However, this age is probably younger than the actual age because the calculated carbon-13 composition is 2.1 permil heavier than measured carbon-13 (table 18).

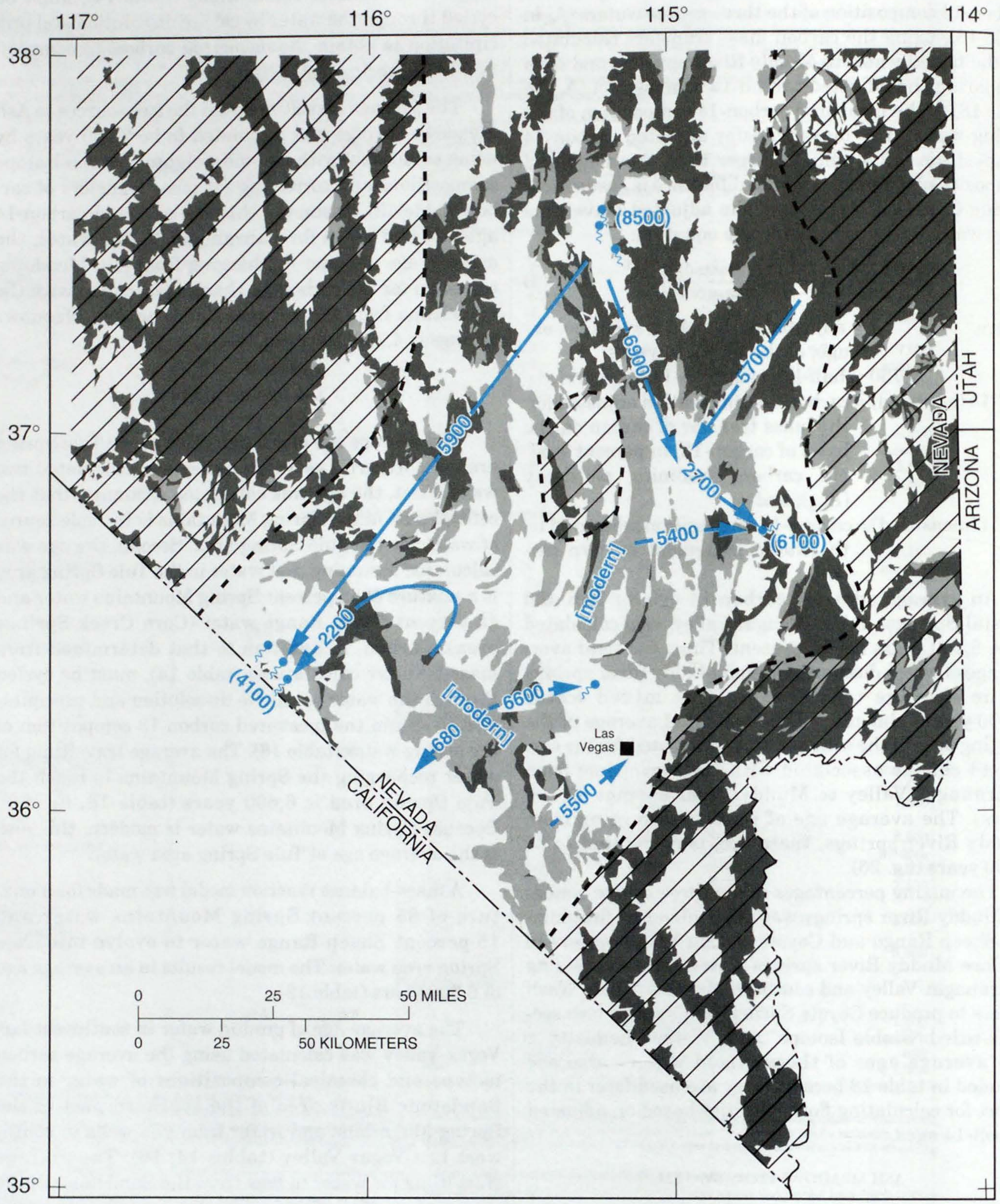

FIGURE 26.—Traveltimes and ages of ground water in carbonate-rock aquifers of southern Nevada, based on adjusted carbon-14 ages. See figure 17 for feature names. Modified from Dettinger (1989, fig. 3).

PAHRUMP VALLEY

The average age of ground water in the Pahrump and Manse Springs area of Pahrump Valley was calculated using the average chemical and carbon isotopic compositions of the central part of the Spring Mountains and Pahrump and Manse Springs wells (tables 14, 16). The average traveltime (and age) of the Pahrump and Manse Springs area water is 680 years (table 18, fig. 26).

SUMMARY OF GROUND-WATER AGES

Adjusted carbon-14 ages increase along flow paths for flow systems in southern Nevada (fig. 26). Therefore, adjusted carbon-14 ages agree with the flow paths delineated using deuterium data and mixing. Carbon-13 composition increases along flow paths for flow systems in southern Nevada (fig. 25). The only source of carbon outside of recharge areas is dissolution of carbonate minerals, which have a carbon-13 composition heavier than that of the water (discussed earlier in this section); therefore carbon-13 composition should increase along a flow path. The increase in carbon-13 composition along flow paths also agrees with the flow paths delineated using deuterium data and mixing.

EXPLANATION

□	**Basin-fill deposits**—In areas underlain by thick, laterally continuous carbonate rock
▨	**Basin-fill deposits**—In areas outside thick, laterally continuous carbonate rock
▒	**Carbonate rock**—In areas of thick, laterally continuous carbonate rock
▨	**Carbonate rock**—In areas outside thick, laterally continuous carbonate rock
■	**Noncarbonate rock**—In areas of thick, laterally continuous carbonate rock
▩	**Noncarbonate rock**—In areas outside thick, laterally continuous carbonate rock
- - - -	**Boundary of laterally continuous carbonate rock**
– 5400 →	**Generalized direction and traveltime of ground-water flow in carbonate-rock aquifer**—Number is adjusted carbon-14 age of water, in years before present
[modern]	**Age of ground water in source area**
(8400) ●	**Spring**—Discharging from carbonate rock aquifer. Number is adjusted carbon-14 age of water, in years before present
●	**Well**—Source of water is carbonate-rock aquifers

GROUND-WATER FLOW VELOCITIES

Ground-water traveltimes (ages) can be used to determine flow velocities of ground water within the aquifers, which provide an independent estimate of hydraulic properties of the aquifers. Age-determined flow velocities for flow paths with a single source were calculated by dividing the horizontal flow-path distance (l) by the adjusted carbon-14 traveltime (t). Age-determined flow velocities for flow paths that involved mixing of two waters (labeled flow path a and flow path b) were calculated by using equations 15 and 16. These equations are derived as follows:

The adjusted carbon-14 age of the water at the end of flow path a can be written as

$$P_a = P_{ao} f_{nd} e^{\left(\frac{-t_a}{8,270}\right)}, \tag{3}$$

where P_a = percent modern carbon at the end of the flow path adjusted for geochemical reactions and decay;

t_a = adjusted carbon-14 age at the end of the flow path;

8,270 = 5,730 (half-life of carbon-14) divided by ln2 (from eq. 2);

f_{nd} = fractional change, in percent modern carbon, associated with the chemical reactions, fractionation, and carbon sources but not decay; and

P_{ao} = initial percent modern carbon of the source water (Wigley and others, 1978, eqs. 23 and 30).

The adjusted carbon-14 age of the water at the end of flow path b can be written as

$$P_b = P_{bo} f_{nd} e^{\left(\frac{-t_b}{8,270}\right)}, \tag{4}$$

where the variables are the same as equation 3, except subscript a's are replaced by subscript b's.

The amount of carbon-14 in a mixture of water from flow path a and flow path b that produces the water at c is conserved (carbon-14 does not decay during mixing and mixing occurs before the water at c is sampled). Thus,

$$P_c = XP_a + (1 - X) P_b, \tag{5}$$

where P_c = adjusted (for geochemical reactions and decay) percent modern carbon at c; and

X = fraction of total flow at c that derives from source a (the percents of P_a and P_b were originally determined by deuterium mass-balance calculations); assuming the f_{nd}

adjustment factor is the same for both flow paths (see the previous discussion of carbon-13 adjustments).

The traveltime along flow path a (t_a) is

$$t_a = \frac{l_a n}{K I_a}, \tag{6}$$

where l_a = average length of flow path a;

I_a = hydraulic gradient ($\Delta h/\Delta l$: Δh is the change in head and Δl is the flow path length) along flow path a;

K = hydraulic conductivity; and

n = effective porosity.

A similar equation holds for flow path b (replacing subscript a's with subscript b's throughout eq. 6 and assuming—for lack of more detailed information—that K and n are the same along both flow paths). A corollary relates the two traveltimes and allows elimination of the hydraulic terms. Dividing the two equations of form (6) yields

$$t_b = \frac{l_b n K I_a}{l_a n K I_b}\left(t_a\right) = \frac{l_b I_a}{l_a I_b}\left(t_a\right), \tag{7}$$

which, upon division by −8,270 years, and exponentiation gives

$$e^{\left(\frac{-t_b}{8,270}\right)} = e^{\left(\frac{-rt_a}{8,270}\right)}, \tag{8}$$

where

$$r = \frac{l_b I_a}{l_a I_b}. \tag{9}$$

This relation combined with equations 3 and 4 gives

$$P_b = P_{bo} f_{nd} e^{\left(\frac{-rt_a}{8,270}\right)}, \tag{10}$$

$$= P_{bo} f_{nd}\left(\frac{P_a}{P_{ao} f_{nd}}\right)^r, \tag{11}$$

so that equation 5 can be rewritten as

$$P_c = XP_a + (1-X)P_{bo} f_{nd}\left(\frac{P_a}{P_{ao} f_{nd}}\right)^r. \tag{12}$$

Equation 12 is an implicit expression for P_a, if P_c, X, P_{bo}, P_{ao}, f_{nd}, and r are known. P_a can be obtained by iterative solution, for example, by application of the Newton-Raphson method (Chapra and Canale, 1988). Once P_a is obtained, the traveltime along path a is

$$t_a = -8,270 \ln\left(\frac{P_a}{P_{ao} f_{nd}}\right), \tag{13}$$

and from equations 7 and 9

$$t_b = rt_a. \tag{14}$$

The age-determined velocities are then estimated as

$$V_{pa} = \frac{l_a}{t_a}, \tag{15}$$

and

$$V_{pb} = \frac{l_b}{t_b}, \tag{16}$$

where V_{pa} = flow velocity along path a; and

V_{pb} = flow velocity along path b (table 19).

These age-determined velocity calculations do not require that the hydraulic conductivity and effective porosity be known, but do assume they are constant for the entire aquifer. Ground-water flow velocities along each flow path also were calculated from estimates of the hydraulic properties of the aquifers using Darcy's equation in the form

$$V_{ha} = \frac{K}{n} I_a, \tag{17}$$

where V_{ha} = linear flow velocity;

K = hydraulic conductivity;

I_a = hydraulic gradient; and

n = effective porosity.

Hydrologic flow velocities were calculated using a hydraulic conductivity of 4.3 ft/d [the median value of 39 aquifer tests for wells completed in carbonate rock in eastern and southern Nevada (Dettinger and others, 1995)], horizontal flow-path distances perpendicular to water-level contours, hydraulic head differences measured along the flow path, and an effective porosity of 2.0 percent.

Hydraulic head differences were taken as difference in water-level altitudes between the initial and final sites along a flow path. Hydraulic head differences for flow paths involving the Spring Mountains were estimated assuming water levels in the Spring Mountains were 500 ft higher than the final site in the adjacent valley (Tule Spring area, southwest Las Vegas Valley, and Pahrump Valley), except for Ash Meadows. For Ash Meadows, the water level for the Spring Mountains was chosen as 500 ft higher than Indian Springs.

This approach was taken because water levels in springs and wells in the Spring Mountains probably do not represent a regional water level, but instead represent local levels. In addition, tritium in water samples from some wells and springs on alluvial fans of the Spring Mountains indicates that traveltime of ground water from the mountains to adjacent alluvial fans is less than about 60 years. Thus, the flow of ground water within the Spring Mountains to adjacent alluvial fans is rapid, so flow in this part of the flow system is omitted in determining flow velocities for the carbonate-rock aquifers.

Effective porosity was chosen as 2.0 percent on the basis of studies that measured or calculated porosities in the carbonate-rock aquifers of southern Nevada (Winograd and Thordarson, 1975; Berger, 1992; Kilroy, 1992). The effective porosity used in this study is secondary fracture porosity; the intercrystalline porosity is minor compared with the fracture porosity of the carbonate rock (Winograd and Thordarson, 1975). Effective porosity was estimated by harmonic analysis of water-level fluctuations in 11 wells and Devils Hole, which penetrate the carbonate-rock aquifers in southern Nevada (Kilroy, 1992). Effective porosity ranged from 1.0 to 4.4 percent, with an average of 2.2 percent and a median of 1.8 percent. For 16 core samples, total porosity ranged from 0.4 to 12.4 percent, with a mean of 5.4 percent and a median of 5.5 percent (Winograd and Thordarson, 1975). Total porosity was estimated using geophysical methods for 43 zones in five wells completed in carbonate rock in southern Nevada (Berger, 1992). Total porosity ranged from 0.0 to 18.5 percent with a mean of 4.7 percent and a median of 4.2 percent.

Flow velocities calculated from adjusted carbon-14 ages range from 29 to 38 ft/yr, whereas flow velocities determined from hydrologic data range from 50 to 540 ft/yr for the White River flow system (table 19). Flow velocities calculated from hydrologic data are about twice those calculated from adjusted carbon-14 data for two flow paths that have hydraulic gradients of 3.3 ft/mi, and about 14 times greater for two other flow paths that have hydraulic gradients of about 35 ft/mi.

The differences in flow velocity between the age and hydrologic calculations could result from (1) the calculated median hydraulic conductivity of the carbonate aquifers being too high, (2) the assumed effec-

TABLE 19.—*Flow velocities in the carbonate-rock aquifers of southern Nevada, calculated from adjusted carbon-14 ages and hydrologic data*

Flow path	Traveltime[1] (years)	Distance[2] (miles)	Change in hydraulic head[3] (feet)	Hydraulic gradient (feet per mile)	Flow velocity (feet per year)	
					From carbon-14 ages[4]	From hydrologic data[5]
White River Flow System						
Coyote Valley to Muddy River springs	2,700	15	50	3.3	29	50
Sheep Range to Muddy River springs	5,400	30	100	3.3	29	50
Pahranagat Valley to Coyote Valley	6,900	50	1,800	36	38	540
Meadow Valley Wash to Coyote Valley	5,700	40	1,400	35	37	520
Ash Meadows Flow System						
Spring Mountains to Ash Meadows springs	2,200	60	1,400	23	140	350
Pahranagat Valley to Ash Meadows springs	5,900	95	1,300	14	85	200
Las Vegas Valley						
Central part of Spring Mountains to Tule Spring area	6,600	15	500	33	12	500
Southern part of Spring Mountains to southwest Las Vegas Valley	5,500	10	500	50	9.6	740
Pahrump Valley						
Spring Mountains to Pahrump Valley	680	10	500	50	78	740

[1] Adjusted carbon-14 traveltime between two samples along a flow path, or for two mixing waters calculated by using equations 15 and 16. For mixing waters, traveltime is dependent on hydraulic gradients of two flow paths.

[2] Horizontal flow-path distance.

[3] Differences in water level between source area and end of flow path.

[4] Flow velocity was calculated by dividing distance by adjusted carbon-14 age.

[5] Flow velocity was calculated by using equation 17.

tive porosity being too low, (3) the actual flow paths being longer than the horizontal flow-path distances, (4) some combination of these factors, or (5) the adjusted ages being too old. For the flow paths with a gradient of 3.3 ft/mi, the average hydraulic conductivity used for calculating velocity may be 50 percent less because measurements are few and the range is large: hydraulic conductivity, 0.01 to 940 ft/d (Dettinger and others, 1995). Effective porosity of 2.0 percent could be higher because of the 1.0 to 4.4 percent range (Kilroy, 1992). A 50-percent greater flow-path length is probable, given that the flow-path distances are horizontal distances and flow in the study area is fracture flow that is tortuous and deep (generally thousands of feet). Uncertainties in age calculations also can be substantial, as discussed previously in the section titled "Ground-Water Age." Thus, average flow velocities for these parts of the White River flow system probably are in the range of 29 to 50 ft/yr.

The two flow paths in the White River flow system, with hydraulic gradients of 35 and 36 ft/mi, have age-calculated velocities of 37 and 38 ft/yr. These velocities are in the range of the velocities calculated from age and hydrologic data for the two flow paths with a hydraulic gradient of 3.3 ft/mi. However, velocities calculated from hydrologic data (520 and 540 ft/yr) are about 14 times greater than the velocities calculated from age determinations because of the high hydraulic gradients. These higher gradients probably indicate areas of lower hydraulic conductivity than was assumed as an average for the flow path. The hydraulic gradient is 36 ft/mi along the flow path from Pahranagat Valley to Coyote Spring Valley. However, water levels decline about 1,200 ft in less than 20 mi (pl. 1), and most of this decline could be in a shorter distance, thus, ground-water flow apparently is restricted between the south end of Pahranagat Valley and the north end of Coyote Spring Valley. A similar high gradient also is present between Meadow Valley Wash and Coyote Spring Valley. Both of these high gradients may result from volcanic rock that overlie and intrude the carbonate-rock aquifers (pl. 1), because volcanic rock generally have lower hydraulic conductivity than the carbonate rock (Winograd and Thordarson, 1975). Thus, flow velocities calculated from hydrologic data may overestimate the average velocities along these flow paths. Velocities calculated from age account for areas of low hydraulic conductivity and hence low velocities, whereas velocities calculated from average hydrologic data do not. Therefore, the age-calculated velocities may be more representative of an average velocity than those calculated from hydrologic data.

The hydraulic gradient also could be high between the Sheep Range and Muddy River springs. However, the gradient used to calculate flow velocity is based on the head differences between wells completed in basin fill and carbonate rock in Coyote Spring Valley plus the head differences between the wells completed in carbonate rock in Coyote Spring Valley and Muddy River springs. Water in the well completed in basin fill is assumed to represent water that recharges the carbonate-rock aquifer beneath Coyote Spring Valley from the Sheep Range (see the previous discussion on stable isotopes).

Flow velocities calculated using ground-water ages are 140 and 85 ft/yr, as compared with 350 and 200 ft/yr for velocities calculated from hydrologic data for the Ash Meadows flow system (table 19). Flow velocities calculated from hydrologic data are about 2.5 times greater than age-estimated velocities. This difference is probably the result of the same five factors as discussed for the White River flow system. The velocities calculated from hydrologic data are about half the minimum velocity estimated by Winograd and Thordarson (1975, p. 115) for the carbonate-rock aquifers beneath the Specter Range approximately 15 mi upgradient from Ash Meadows. Using their data for total discharge and cross-sectional area and an effective porosity of 2 percent, a velocity of 280 ft/yr was calculated. This value is about the average of the two velocities estimated from the hydrologic data for this study. Thus, average flow velocities in the Ash Meadows flow system probably are in the range of 85 to 350 ft/yr.

Flow velocities calculated using adjusted carbon-14 age are significantly less than velocities calculated from hydrologic data for flow paths from the Spring Mountains to Las Vegas and Pahrump Valleys (table 19). This large difference is a result of the steep hydraulic gradient for the flow paths, as compared with most other flow paths in the carbonate-rock aquifers, and the relatively old ages of ground water in Las Vegas Valley. The hydraulic gradients for these flow paths are assumed high because recharge to the Spring Mountains is at altitudes greater than 10,000 ft. For these flow paths, the velocities calculated from hydrologic data may be more reasonable than velocities calculated from age because the amount of recharge to the Spring Mountains is large and ground water is assumed to mound beneath the mountains and adjacent alluvial fans. The velocities calculated from ground-water age for the flow paths that end at the Tule Spring area and southwestern Las Vegas Valley (table 19) are less than half the minimum velocity calculated from age data for other flow paths in the carbonate-rock aquifers of southern Nevada. This age difference indicates that the adjusted ages may be too old or that ground water is not flowing directly from the Spring Mountains to these areas. In Pahrump Valley, the relatively young ground-water age results in a velocity of 78 ft/yr, which is in the range of other age-calculated velocities, but still about an order of magnitude less than the velocity calculated from hydrologic data. Thus, the average velocity for these flow paths ranges from 78 to 740 ft/yr and probably is closer to the high velocity because of the steep gradient. However, the lack of measurable tritium in water samples from these sites (table 16) indicates the velocity of 740 ft/yr is a maximum.

In summary, flow velocities calculated from adjusted carbon-14 ages are less than velocities calculated from hydrologic data. Thus, either the adjusted ages are too old or the sparse hydrologic data do not represent average hydraulic properties of the carbonate-rock aquifers. As discussed in the section titled "Ground-Water Age," the ages could be much younger if more vein calcite is exchanging carbon with the water than the 50 percent assumed. On the other hand, age-estimated flow velocities of 9.6 to 140 ft/yr are in the range of velocities calculated for the Madison aquifer in Montana, Wyoming, and South Dakota. The Madison aquifer is a carbonate-rock aquifer in Montana, Wyoming, and South Dakota. Flow velocities calculated using adjusted ages range from 7.2 to 86.7 ft/yr, and velocities simulated by ground-water modeling range from 11.2 to 74.2 ft/yr, for 22 flow paths in the Madison aquifer (Downey, 1984; Plummer and others, 1990, table 12). Thus, velocities calculated from ground-water age for the carbonate-rock aquifers of southern Nevada have a similar, but slightly larger, range as velocities for the Madison aquifer calculated from both age and hydrologic data. These similar velocities indicate that the adjusted ages of the ground water are probably reasonable. However, the Madison aquifer is not in an area of active extension like southern Nevada, so flow within the Madison aquifer may not be primarily along secondary fractures like in aquifers of southern Nevada. Thus, a direct comparison of velocities in the aquifers may not be valid.

The discrepancy in velocities calculated from age compared with those calculated from hydrologic data indicate that either the ages or the average hydraulic conductivity may be overestimated, and effective porosity and horizontal flow path lengths may be underestimated. The large discrepancy in velocities calculated from age and hydrologic data for flow from the Spring Mountains to adjacent valleys indicates the lack of knowledge about the recharge process for the carbonate-rock aquifers in southern Nevada.

IMPLICATIONS OF CHEMICAL AND ISOTOPIC COMPOSITION OF GROUND WATER IN AQUIFERS OF THE GREAT BASIN

Major processes affecting the chemical and isotopic composition of ground water in Smith Creek Valley are similar to those in other hydrologically closed systems of the Great Basin. The principal aquifers in these flow systems consist of unconsolidated sedimentary deposits underlain by less permeable consolidated rock. Numerous systems in which the flow is primarily in basin-fill deposits that culminate in a discharging playa area exist throughout the Great Basin (Harrill and others, 1988). Near recharge areas, ground-water chemistry results from the dissolution of primary minerals and soil-zone CO_2. As the water moves downgradient into finer grained sediments near the center of the basin, ion exchange becomes an important process. In the discharging playa area, evaporation, transpiration, and dissolution of salts left behind by evapotranspiration are dominant processes affecting water chemistry. Precipitation of secondary minerals and formation of clays remove ions from water throughout the flow system. Deuterium and oxygen-18 become heavier (more positive) in the discharge area due to evaporation.

Major processes affecting the chemical and isotopic composition of ground water in the carbonate-rock aquifers of southern Nevada should be similar to those in carbonate-rock flow systems throughout the Great Basin. The principal aquifers in these flow systems consist of carbonate rock, but unconsolidated basin-fill deposits also may be hydrologically connected with the carbonate rock. Near recharge areas, ground-water chemistry results from dissolution of primary minerals—mainly calcite and dolomite—and soil-zone CO_2 gas. Mixing of chemically and isotopically different waters produces ground water of chemical and isotopic composition different from the initial waters. As the water circulates to depth, geothermal heating causes precipitation of calcite, removing calcium and bicarbonate ions from the water. Ion exchange, precipitation of secondary minerals, and formation of clay minerals removes ions from the water throughout the flow system; exchange reactions also add ions to the water. The carbon-isotope composition of the ground water changes because of (1) the dissolution of calcite, dolomite, and soil-zone CO_2 gas, which adds carbon of different isotopic composition to the water, and (2) the precipitation of calcite and outgassing of CO_2, which remove isotopically heavier (calcite precipitation) or lighter (CO_2 outgassing) carbon from the system. The chemical and isotopic compositions resulting from these processes can be used in conjunction with hydrologic and geologic information to determine sources, flow paths, and mixing of water in carbonate-rock flow systems. This information can be used to help determine rates and volumes of flow within the aquifers.

Carbonate-rock aquifers and hydrologically closed basin-fill aquifers form the two principal types of flow systems in the Great Basin. Other types may (1) be a combination of the basin-fill and carbonate-rock systems, (2) involve ground-water flow primarily in volcanic rock, or (3) be affected by mixing with surface water. In the Railroad Valley flow system in central Nevada, for example, ground water flows in carbonate-rock aquifers and volcanic rock and discharges from the playa area of a basin-fill aquifer. The initial chemical character of water in this flow system is a result of the dissolution of primary minerals and soil-zone CO_2 gas. As the ground water flows into underlying carbonate-rock aquifers, dissolution of carbonate-rock minerals, followed by precipitation of calcite (where water is heated at greater depths), dominates. The chemical composition of water flowing through volcanic rock changes because volcanic glass and minerals dissolve. In the discharge area, a playa surrounded by phreatophytic vegetation, evaporation and transpiration are the dominant processes affecting the chemical and isotopic composition of the water. In addition, dissolution of any salts formed previously as the result of evapotranspiration could be a major source of ions in the water. Several processes, such as precipitation of secondary minerals, formation of clays, and ion exchange, may occur throughout the flow system.

In the Pahute Mesa area of southern Nevada, ground-water flow is primarily in volcanic rock (Blankennagel and Weir, 1973). Processes affecting the chemical and isotopic compositions of the water are similar to those in the carbonate-rock aquifers, but the mineralogy is different. The water chemistry results from dissolution of volcanic glass and minerals (such as feldspars) rather than carbonate-rock minerals. Precipitation of minerals, formation of clay minerals, and ion exchange also are important processes affecting the water chemistry (White, 1979; White and others, 1980; Kerrisk, 1983).

In the Humboldt River basin-fill system of north-central Nevada, the chemical and isotopic composition of ground water is affected by mixing with surface water in parts of the aquifer. Along the river, some surface water infiltrates into the basin-fill aquifers, particularly during periods of high flow (for example see Cohen and others, 1965, p. 79–80). Because the chemical and isotopic compositions of the surface water and ground water are likely to be different, mixing of the two waters would produce a blend with a composition intermediate to the two end-member waters. The proportions of mixing end-member waters, along with any subsequent hydrologic or geochemical processes that affect the water chemistry, determine the chemical and isotopic composition of the resulting blend. Unless a river is a major source of recharge in such a system, interaction between streamflow and water in the aquifer affects the chemical and isotopic composition of the ground water only adjacent to the river. However, in areas of surface-water irrigation, water applied on the land surface percolates to the shallow water table (Loeltz and others, 1949, p. 33; Harrill and Moore, 1970, p. 74), thus affecting the chemical and isotopic composition of ground water in those areas. The isotopic composition of ground water changes due to the infiltration of isotopically different surface water.

SUMMARY

This report, a product of the Great Basin Regional Aquifer-System Analysis and Southern Nevada Carbonate studies, briefly describes the distribution of dissolved solids and the chemical character of ground water in the study area; discusses the geochemical and hydrologic processes that result in the chemical evolution of ground water in a hydrologically closed basin-fill aquifer; discusses the geochemical and hydrologic processes that produce the chemical and isotopic compositions of ground water in carbonate-rock aquifers of southern Nevada; delineates ground-water flow paths, source areas, and mixing of waters in the carbonate-rock aquifers of southern Nevada; discusses ground-water ages calculated from mass-balance reaction models and carbon isotopes; and compares the resulting age-calculated flow velocities with velocities determined using hydrologic data for the carbonate-rock aquifers of southern Nevada. Smith Creek Valley, in west-central Nevada, represents hydrologically closed basin-fill flow systems, with ground water flowing primarily through basin-fill aquifers. Carbonate-rock aquifers in southern Nevada represent carbonate-rock flow systems, in which ground water flows through both basin-fill and carbonate-rock aquifers and underneath topographic divides through carbonate rock.

Water in the principal aquifers of the Great Basin generally contains less than 1,000 mg/L of dissolved solids, except in natural-discharge and geothermal areas. Aquifers in industrial, mining, urban, and agricultural areas, and aquifers containing highly soluble evaporative salts and minerals, may have dissolved-solid concentrations greater than 1,000 mg/L or elevated concentrations of undesirable constituents, or both. These areas of known higher dissolved solids, or elevated constituent concentrations, are shown in a series of water-quality reports for States in the Great Basin.

The general chemical character of water in principal aquifers of the Great Basin is dominated by sodium, calcium, and bicarbonate in basin-fill aquifers of the western Great Basin; calcium, sodium, magnesium, and bicarbonate in basin-fill aquifers of the eastern Great Basin; and calcium, magnesium, and bicarbonate in carbonate-rock aquifers of the eastern Great Basin. The chemical character of water concentrated by evapotranspiration in discharging playa areas is generally dominated by sodium, chloride, and sulfate.

Geochemical and hydrologic processes that produce the major-ion chemistry and isotopic composition of water in aquifers throughout the Great Basin are identified for examples of the two principal types of flow systems in the Great Basin: a hydrologically closed basin-fill aquifer in Smith Creek Valley, west-central Nevada, and a regional carbonate-rock aquifer system in southern Nevada.

Chemical and isotopic compositions of water in the basin-fill aquifer in Smith Creek Valley evolve as a result of (1) evapotranspirative concentration of the water, (2) dissolution of minerals and soil-zone CO_2 gas, (3) precipitation, or formation by incongruent dissolution, of minerals, and (4) ion exchange. Water recharging the basin-fill aquifer originates mainly as precipitation in the surrounding mountains. Precipitation in the recharge areas is concentrated by evapotranspiration. The concentrated precipitation infiltrates to the soil zone where it dissolves CO_2 gas, and volcanic groundmass and phenocrysts (dominantly albite and anorthite), producing a sodium calcium bicarbonate water. In addition, small amounts of K-feldspar, gypsum, biotite, and possibly pyrite, illite, and chlorite are dissolved, adding ions to the water. Chalcedony precipitates, removing ions from the water, and kaolinite (or some other clay mineral) forms, probably as the result of incongruent dissolution of the feldspars.

Calcium and magnesium in the water exchange for sodium on clays in the playa area, where basin-fill deposits grade to finer grained sediments. This exchange results in the sodium calcium bicarbonate water evolving into a sodium bicarbonate water. Calcium also may be removed from the water by the weathering of plagioclase to Ca/Na-montmorillonite and the precipitation of a zeolite mineral. Ions are added to the water by the dissolution of albite, anorthite, K-feldspar, CO_2 gas, and salts containing chloride. Ions are removed from the water by precipitation of chalcedony and reduction of sulfate to hydrogen sulfide gas. Incongruent dissolution of feldspars probably results in the formation of kaolinite.

Evapotranspiration, dissolution of evaporative salts containing chloride, and precipitation of calcite and zeolite minerals are the main processes affecting ground-water chemistry in the discharge area. These processes result in the sodium bicarbonate water evolving into a sodium chloride water. Constituents also are added to the water by dissolution of albite, anorthite, K-feldspar, gypsum (or evaporative salts containing sulfate), and chlorite, and kaolinite is formed by incongruent dissolution of feldspars. Evaporation of shallow ground water in the discharge area also causes the water to become heavier in deuterium and oxygen-18.

Chemical and isotopic compositions of water in the carbonate-rock aquifers of southern Nevada evolve as a consequence of (1) dissolution of minerals and soil-zone CO_2 gas, (2) precipitation or formation of minerals, (3) ion exchange, (4) mixing of chemically or isotopically different waters, and (5) geothermal heating. Water in the carbonate-rock aquifers originates primarily as precipitation in the Spring Mountains and Sheep Range. This water is concentrated by evapotranspiration and obtains most of its ions from dissolution of CO_2 gas, calcite, and dolomite in the recharge areas. Recharge waters generally are saturated with respect to both calcite and dolomite and contain predominantly calcium, magnesium, and bicarbonate. These waters circulate to depth, which causes the water to heat and to precipitate calcite.

Throughout the carbonate-rock aquifers, calcite precipitates, calcium and magnesium in the water exchange for sodium in clays, chalcedony precipitates, kaolinite forms, and in some spring areas CO_2 gas exsolves. In large areas of the carbonate-rock aquifers, gypsum dissolves, causing calcite to precipitate and dolomite to dissolve (dedolomitization). In addition, sodium and potassium probably are added to the water by the dissolution of volcanic glass and minerals (dominantly albite and K-feldspar) and zeolite minerals (probably clinoptilolite), which are present in parts of the study area. Sodium chloride salts also are present in some parts of the study area and dissolve. In the Ash Meadows flow system, downward leakage of water containing high sodium and sulfate from the Tertiary aquitard into the carbonate-rock aquifers is the primary source of increased sulfate and sodium concentrations. Thus, outside of the recharge areas of the carbonate-rock aquifers in southern Nevada, sodium, sulfate, and chloride can be major dissolved constituents in the water.

Waters of differing chemical and isotopic compositions mix in the carbonate-rock aquifers. This mixing can result in a water that is isotopically and chemically different from the source waters. However, all the waters generally are undergoing the same geochemical processes, which produce waters of similar chemical composition. Thus, the primary evidence of mixing in the carbonate-rock aquifers is the modified isotopic composition of the mixed water, compared to the mixing waters. The mixing waters generally originate in different areas, and therefore, have different isotopic compositions.

The isotopic and chemical compositions of water in the carbonate-rock aquifers of southern Nevada were used to delineate flow systems. First, the deuterium content of water in the carbonate-rock aquifers was used to determine source areas, flow paths, and mixing. Second, these deuterium-delineated flow paths and mixing scenarios were checked for chemical feasibility using mass-balance and mineral-equilibrium models based on water chemistry and the geochemical processes that are assumed to have produced the chemical composition of the water. Third, adjusted carbon-14 ages were determined from carbon-isotope data on the basis of the mass-balance reaction models. These ages provide another check on the flow path and mixing scenarios because ground-water ages increase along flow paths, and apparent ages change due to mixing of different age waters.

These isotopic and geochemical models produced the following results:

1. Ground water discharging at the terminus of the White River flow system (Muddy River springs) is a mixture of 40 percent (14,000 acre-ft/yr) Pahranagat Valley water, 38 percent (14,000 acre-ft/yr) Sheep Range water, and 22 percent (8,000 acre-ft/yr) southern Meadow Valley Wash water.
2. Ground water discharging at the terminus of the Ash Meadows flow system (Ash Meadows springs) is a mixture of 60 percent (10,000 acre-ft/yr) Spring Mountains water and 40 percent (7,000 acre-ft/yr) Pahranagat Valley water.
3. Las Vegas Valley receives all, or almost all, of its ground water from the Spring Mountains (the Sheep Range may supply as much as 2,500 acre-ft/yr to northern Las Vegas Valley).
4. Pahrump Valley receives all its ground water from the Spring Mountains.

Flow velocities calculated from adjusted carbon-14 ages are less than velocities calculated from hydrologic data. This difference indicates that the ages or average hydraulic conductivities may be overestimated, and effective porosity and horizontal flow path lengths may be underestimated. Calculated ground-water ages are sensitive to the carbon-13 composition of the dissolving calcite. To determine ages for velocity estimates, 50 percent of the dissolving calcite was assumed to be vein calcite. Calculated ground-water ages could be much younger if more vein calcite, which has a lighter carbon-13 composition than the primary calcite, is dissolving. Knowledge about the recharge process for the carbonate-rock aquifers of southern Nevada is insufficient to explain the large difference in velocities calculated from age and hydrologic data for flow from the Spring Mountains to adjacent valleys.

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APPENDIXES

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*

[Average deuterium composition for each site is plotted on plate 2. Permil values of deuterium and oxygen-18 are reported relative to V-SMOW]

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
Cresent Spring	35 28 43	115 10 47	-73.0	-9.4	06-22-85	2
Pine Spring	35 34 25	115 09 23	-88.0	-11.9	06-22-85	2
Ora Hana Spring	35 37 25	115 04 07	-72.0	-8.4	06-22-85	2
Saratoga Spring	35 40 53	116 25 18	-90.5	-10.8	04-23-82	1
McClanahan Spring	35 41 42	115 11 05	-67.0	-7.2	06-21-85	2
Test Hole A1-1 (505 feet)	35 45 28	115 15 05	-91.0	-9.3	--	2
Test Hole A1-1 (625 feet)	35 45 28	115 15 05	-94.0	-10.9	--	2
Test Hole A1-2 (408 feet)	35 46 04	115 15 46	-94.0	-12.4	--	2
Test Hole A1-2 (485 feet)	35 46 04	115 15 46	-95.0	-11.0	--	2
Test Hole A3-3 (580 feet)	35 47 16	115 20 39	-90.0	-10.7	--	2
Test Hole A3-3 (600 feet)	35 47 16	115 20 39	-87.0	-10.5	--	2
Test Hole A3-3 (740 feet)	35 47 16	115 20 39	-97.0	-13.6	--	2
Test Hole A3-3 (855 feet)	35 47 16	115 20 39	-89.0	-10.2	--	2
Test Hole A3-3 (890 feet)	35 47 16	115 20 39	-94.0	-11.2	--	2
Jean Prison Well	35 47 18	115 20 43	-94.0	-10.4	--	2
Jean Prison Well	35 47 18	115 20 43	-95.0	-12.1	06-21-85	2
Test Hole A3-2 (740 feet)	35 47 43	115 21 34	-93.0	-10.8	--	2
Test Hole A3-2 (840 feet)	35 47 43	115 21 34	-91.0	-11.5	--	2
Test Hole A3-1 (833 feet)	35 48 13	115 22 38	-89.0	-10.8	--	2
Test Hole A3-1 (903 feet)	35 48 13	115 22 38	-88.0	-10.5	--	2
Rosechrist Well	35 48 18	115 41 41	-94.0	-13.20	09-27-86	1
Castillo Well	35 50 02	115 26 09	-94.0	-12.8	06-21-85	2
Tecopa Hot Spring	35 52 19	116 13 50	-98.0	-12.85	06-30-85	1
Bird Spring	35 53 20	115 22 12	-88.0	-11.7	06-23-85	2
Arizona Seep Spring (Black Canyon #13)	35 55 42	114 42 20	-82.0	-10.0	01-00-81	9
Wilson's Tank	35 56 02	115 25 28	-87.5	-10.9	06-22-85	2
Bighorn Sheep Spring (Black Canyon #12)	35 56 22	114 44 06	-80.5	-10.2	01-00-81	9
Ringbolt Rapids Spring (Black Canyon #10)	35 57 39	114 43 26	-83.5	-11.1	03-00-82	9
Sky Harbor Airport	35 58 16	115 08 50	-95	-13.1	02-28-86	16
Shoshone Spring	35 58 48	116 16 23	-94.5	-12.9	04-25-82	1
Palm Tree Cold Spring (Black Canyon #7)	35 59 42	114 44 15	-82.0	-10.5	03-00-82	9
Tenaya Well	36 00 42	115 15 05	-92.0	-12.3	05-10-83	1
Showboat Country Club #2	36 02 51	115 04 48	-97	-13.3	02-27-86	16
Genstar Gypsum Plant	36 03 04	115 23 43	-86	-12.3	07-11-86	16
Flowing Well (DRI LG153)	36 03 22	115 03 08	-100.0	-12.9	05-11-83	1
Sandstone Spring #1	36 03 47	115 28 09	-89.0	-12.20	06-25-85	1
Sandstone Spring #1	36 03 47	115 28 09	-90.0	-12.7	06-25-85	2
Sunset Park Well	36 03 49	115 06 49	-94.0	-12.7	05-09-83	1
Gravel Pit Well	36 05 17	114 56 32	-76.0	-7.8	05-12-83	1
Spanish Trail Country Club Well	36 05 56	115 15 38	-89	-12.2	03-13-86	16
Stocks Mill & Supply Co. Well	36 06 07	115 15 46	-86	-12.4	03-13-86	16
Tropicana Country Club South #1	36 06 22	115 10 09	-95	-13.2	05-16-86	16
Sands Hotel And Casino #3	36 07 28	115 10 09	-98	-13.5	05-16-86	16
Sparklettes Drinking Water Co.	36 07 41	115 11 45	-99	-13.6	04-22-86	16
BLM Visitors Center	36 07 44	115 26 03	-89.0	-12.25	06-30-85	1
Las Vegas Country Club	36 08 20	115 08 49	-100	-13.7	07-08-86	16
Hartzski Well	36 08 36	115 15 47	-90	-12.4	03-04-86	16
Red Spring	36 08 40	115 25 10	-89.0	-12.25	06-26-85	1
Red Spring	36 08 40	115 25 10	-93.0	-12.2	06-26-85	2
Manse Well	36 09 17	115 53 42	-99.0	-13.55	06-27-85	1
Manse Spring	36 09 21	115 54 10	-97.0	--	08-00-68	3

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
Manse Spring	36 09 21	115 54 10	-101.0	--	03-00-70	3
DRI Well LG047	36 09 33	115 05 51	-101.0	-13.6	05-11-83	1
Shetland Mutual Water Users Association	36 09 39	115 10 33	-100	-13.5	03-06-86	16
LVVWD #1A	36 09 39	115 13 32	-94	-13.2	05-14-86	16
Willow Spring	36 09 41	115 29 51	-90.5	-12.3	06-26-85	1
LVVWD #11A	36 09 52	115 11 22	-97	-13.4	06-13-86	16
Kiup Spring	36 09 56	115 43 28	-93.0	-12.7	06-28-85	1
Cave Spring	36 09 58	115 35 52	-93.0	-12.8	06-28-85	1
Union Pacific Railroad Well #3	36 10 05	115 09 13	-100	-13.8	04-09-86	16
LVVWD #7A	36 10 06	115 11 23	-96	-13.2	06-17-86	16
Sky Mt. Resort	36 10 13	115 34 44	-96.5	-13.25	06-28-85	1
LVVWD #14	36 10 18	115 11 22	-97	-13.2	05-06-86	16
White Rock Spring	36 10 27	115 28 43	-91.0	-12.5	06-26-85	1
LVVWD #15A	36 10 31	115 11 23	-97	-13.8	05-14-86	16
LVVWD #34	36 10 31	115 11 06	-99	-13.6	05-16-86	16
LVVWD #16	36 10 31	115 11 39	-97	-13.6	06-04-86	16
LVVWD #17	36 10 31	115 11 39	-97	-13.5	06-04-86	16
LVVWD #18A	36 11 12	115 14 38	-94	-13.4	06-06-86	16
LVVWD #45	36 11 50	115 12 13	-101	-14.0	05-13-86	16
LVVWD #22A	36 12 05	115 15 43	-100	-14	03-27-79	2
NLVWD Robinson Well	36 12 17	115 11 41	-100	-14.1	06-03-86	16
Pahrump Spring Well	36 12 27	115 59 01	-97.0	-13.55	06-27-85	1
NLVWD West Cheyenne Well	36 12 38	115 11 21	-104.0	-13.5	05-10-83	1
NLVWD West Cheyenne Well	36 12 38	115 11 21	-100	-13.8	06-12-86	16
Nellis AFB #13	36 12 44	115 03 00	-98	-13.8	04-03-86	16
Trout Spring	36 13 22	115 40 59	-99	--	03-28-70	3
Trout Spring	36 13 22	115 40 59	-99.1	-14.1	06-04-73	4
Trout Spring	36 13 22	115 40 59	--	-13.8	09-07-73	4
Trout Spring	36 13 22	115 40 59	--	-13.8	11-14-73	4
Trout Spring	36 13 22	115 40 59	--	-13.7	03-25-74	4
Trout Spring	36 13 22	115 40 59	-98.4	-13.85	12-08-74	4
Trout Spring	36 13 22	115 40 59	-97.6	-13.30	05-01-75	4
Trout Spring	36 13 22	115 40 59	-96.2	-13.35	11-11-75	4
Trout Spring	36 13 22	115 40 59	-99.8	-14.02	02-16-76	4
Trout Spring	36 13 22	115 40 59	-98.6	-13.90	11-17-76	4
Trout Spring	36 13 22	115 40 59	-98.8	-14.10	03-08-77	4
Trout Spring	36 13 22	115 40 59	-99.0	-14.10	05-13-77	4
Trout Spring	36 13 22	115 40 59	-98.6	-13.45	11-19-77	4
Trout Spring	36 13 22	115 40 59	-96.0	-12.85	03-21-78	4
Trout Spring	36 13 22	115 40 59	-97.0	-12.90	05-08-78	4
Trout Spring	36 13 22	115 40 59	-96.0	-13.20	05-04-83	4
Trout Spring	36 13 22	115 40 59	-95.5	-13.05	05-16-83	4
Trout Spring	36 13 22	115 40 59	-99.5	-13.65	11-30-83	4
Trout Spring	36 13 22	115 40 59	-96.5	-13.65	03-19-84	4
Trout Spring	36 13 22	115 40 59	-97.0	-13.7	06-30-87	2
Trout Spring	36 13 22	115 40 59	-96.5	-13.70	03-24-88	4
Trout Spring	36 13 22	115 40 59	-97.0	-13.70	04-04-88	1
Trout Spring	36 13 22	115 40 59	-98.5	-13.70	12-11-88	1
Trout Spring	36 13 22	115 40 59	-97.0	-13.60	06-21-89	4
Well 5 Franklin Lake Nr Death Valley	36 14 06	116 17 32	-93.5	-11.1	11-21-83	1
Well 10 Franklin Lake Nr Death Valley	36 14 06	116 17 39	-91.0	-10.0	11-21-83	1
NLVWD Desert Aire Well	36 14 15	115 12 16	-102	-14.0	03-06-86	16
Lake Mead Base Well #3	36 14 21	115 00 16	-101.5	-13.80	09-29-86	1
Lake Mead Base Well #3	36 14 21	115 00 16	-103.0	-13.8	09-29-86	2

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
Craig Ranch Country Club #2	36 14 29	115 09 00	-106	-14.5	03-04-86	16
Rainbow Spring #2	36 14 36	115 37 55	-96.0	-12.4	08-19-82	2
Peak Spring Canyon Creek	36 14 40	115 43 09	-92.5	-13.10	09-27-86	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-93.5	-13.4	04-07-87	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-92.0	-13.0	06-09-87	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-92.5	-12.95	08-03-87	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-97.5	-13.5	11-10-87	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-98.0	-13.70	03-08-88	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-101.0	-13.85	04-05-88	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-98.5	-13.85	05-17-88	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-95.5	-13.60	06-23-88	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-95.5	-13.25	09-06-88	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-94.5	-13.30	10-18-88	1
Peak Spring Canyon Creek	36 14 40	115 43 09	-94.5	-13.45	11-25-88	1
GS-18 Franklin Lake Nr Death Valley	36 14 51	116 17 08	-66.5	-2.5	11-18-83	1
Nellis AFB Well #4	36 14 56	115 00 15	-95	-13.2	04-03-86	16
Well 13 Franklin Lake Nr Death Valley	36 14 57	116 18 34	-92.0	-11.3	11-17-83	1
Calnev Pipeline Well	36 14 59	115 02 39	-98.0	-13.1	00-00-86	1
GS-15 Franklin Lake Nr Death Valley	36 15 27	116 17 13	-66.0	-2.1	11-14-83	1
GS-15 Franklin Lake Nr Death Valley	36 15 27	116 17 13	-62.0	-1.5	10-20-84	1
Pat Well	36 15 45	115 19 37	-92.0	-12.9	03-31-87	2
Echo Spring	36 15 55	115 39 29	-94.5	-12.8	04-16-80	2
GS-4 Franklin Lake Nr Death Valley	36 16 00	116 16 22	-93.5	-10.4	11-16-83	1
GS-4 Franklin Lake Nr Death Valley	36 16 00	116 16 22	-93.5	-10.3	10-18-84	1
Highway Maintenance Well	36 16 17	115 35 38	-96.0	-12.2	08-19-82	2
GS-12 Franklin Lake Nr Death Valley	36 16 35	116 17 21	-76.0	-6.6	11-14-83	1
GS-12 Franklin Lake Nr Death Valley	36 16 35	116 17 21	-77.0	-6.7	10-27-84	1
Taylor's Steak House Rest.	36 17 08	115 14 41	-101	-13.7	11-03-86	16
Martin Well	36 17 25	115 14 35	-100.0	-13.4	06-30-85	2
GS-8 Franklin Lake Nr Death Valley	36 17 27	116 17 09	-100.0	-12.9	11-19-83	1
Grapevine Springs	36 18 03	115 29 25	-91.0	-11.6	06-28-85	2
Grapevine Springs	36 18 03	115 29 25	-92.0	-13.2	07-09-87	2
Deer Creek Spring #1	36 18 27	115 38 13	-97.0	-14.1	06-28-85	2
Deer Creek Spring #1	36 18 27	115 38 13	-100.0	-14.0	03-30-87	2
Deer Creek Spring #2	36 18 27	115 37 37	-98.0	-13.45	06-28-85	2
Racel Well	36 18 40	115 15 39	-99.0	-13.3	05-10-83	1
Gilbert Well	36 18 45	115 25 06	-98.0	-12.7	08-20-82	2
Well 15 Franklin Lake Nr Death Valley	36 18 51	116 16 60	-98.0	-12.7	11-21-83	1
Stewart Well	36 19 10	115 40 20	-100	-13.4	06-24-86	2
Tule Spring State Park Well	36 19 14	115 16 00	-100.0	--	01-00-69	3
Tule Spring State Park Well	36 19 14	115 16 00	-100.0	-11.5	08-20-82	2
Tule Spring State Park Well	36 19 14	115 16 00	-99.0	-13.4	06-30-85	2
Clark Spring	36 19 14	115 43 15	-93.5	-12.9	06-29-85	1
Clark Spring	36 19 14	115 43 15	-99.0	-13.7	06-24-87	1
Mulder Well	36 19 25	115 13 22	-97.0	-13.2	06-25-86	2
Mifflin Well	36 19 32	115 25 23	-100	-12.6	08-20-82	2
Lee's Oasis Well	36 19 48	115 18 55	-98.0	-13.4	06-24-86	2
Holland Well	36 20 00	115 17 00	-98.0	-12.6	08-20-82	2
Summer Homes Well	36 20 06	115 39 18	-99.5	-13.80	09-26-86	1
Summer Homes Well	36 20 06	115 39 18	-101.0	-13.5	09-26-86	2

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
G.P. Apex Well	36 20 28	114 55 36	-97.5	-13.35	09-29-86	1
G.P. Apex Well	36 20 28	114 55 36	-98.0	-13.45	09-30-86	1
G.P. Apex Well	36 20 28	114 55 36	-96.0	-13.8	09-30-86	2
Cortney Well	36 20 42	115 20 52	-96.0	-12.1	08-21-82	2
Adams Well	36 20 47	115 19 50	-98.0	-12.7	08-20-82	2
Paiute Indian Reservation Well	36 21 02	115 20 33	-98.0	-14.0	08-21-86	2
Big Spring	36 22 30	116 16 25	-104.0	--	01-00-69	3
Big Spring	36 22 30	116 16 25	-104.0	--	03-00-70	3
Big Spring	36 22 30	116 16 25	-102.0	-13.4	05-24-73	6
Big Spring	36 22 30	116 16 25	-102.0	--	03-09-75	6
Rogers Spring	36 22 39	114 26 38	-92.0	-12.2	07-21-81	1
Grace Petroleum Arrow Canyon Water Well	36 22 58	114 54 60	-96.0	-13.7	04-26-82	1
Blue Point Spring	36 23 21	114 25 26	-93.5	-12.50	07-01-85	1
Blue Point Spring (Duplicate Sample)	36 23 21	114 25 26	-92.5	-12.35	07-01-85	1
Blue Point Spring	36 23 21	114 25 26	-93.0	-12.4	06-24-85	2
Genstar Well	36 23 29	114 54 14	-97.0	-13.05	03-31-86	1
Point of Rocks (King) Spring	36 24 05	116 16 14	-106.5	--	08-00-68	3
Point of Rocks (King) Spring	36 24 05	116 16 14	-101.0	--	01-00-69	3
Point of Rocks (King) Spring	36 24 05	116 16 14	-104.0	--	03-00-70	3
Point of Rocks (King) Spring	36 24 05	116 16 14	-104.0	--	03-03-75	6
Point Of Rocks (King) Spring	36 24 05	116 16 14	-102.0	-13.6	03-09-81	1
Cold Creek Spring	36 24 50	115 44 20	-97	--	08-24-68	3
Cold Creek Spring	36 24 50	115 44 20	-102	--	01-27-69	3
Cold Creek Spring	36 24 50	115 44 20	-103	--	03-30-70	3
Cold Creek Spring	36 24 50	115 44 20	-99.1	-14.1	05-31-73	4
Cold Creek Spring	36 24 50	115 44 20	-98.4	-14.2	08-29-73	4
Cold Creek Spring	36 24 50	115 44 20	--	-13.8	11-13-73	4
Cold Creek Spring	36 24 50	115 44 20	--	-13.9	04-04-74	4
Cold Creek Spring	36 24 50	115 44 20	-100.4	-13.80	12-09-74	4
Cold Creek Spring	36 24 50	115 44 20	-100.0	-13.75	04-28-75	4
Cold Creek Spring	36 24 50	115 44 20	-98.0	-13.30	11-10-75	4
Cold Creek Spring	36 24 50	115 44 20	-100.4	-13.95	11-16-76	4
Cold Creek Spring	36 24 50	115 44 20	-100.7	-14.20	02-17-77	4
Cold Creek Spring	36 24 50	115 44 20	-101.8	-14.15	05-12-77	4
Cold Creek Spring	36 24 50	115 44 20	-101.5	-13.85	11-18-77	4
Cold Creek Spring	36 24 50	115 44 20	-99.5	-13.15	03-20-78	4
Cold Creek Spring	36 24 50	115 44 20	-98.5	-13.15	05-06-78	4
Cold Creek Spring	36 24 50	115 44 20	-102.5	-14.00	12-04-82	4
Cold Creek Spring	36 24 50	115 44 20	-100.5	-13.75	05-02-83	4
Cold Creek Spring	36 24 50	115 44 20	-100.5	-13.65	05-15-83	4
Cold Creek Spring	36 24 50	115 44 20	-99.5	-13.85	12-12-83	4
Cold Creek Spring	36 24 50	115 44 20	-99.0	-13.8	03-31-87	2
Cold Creek Spring	36 24 50	115 44 20	-100.0	-13.75	04-05-88	1
Willow Spring	36 25 00	115 45 47	-97.0	-13.9	06-02-73	5
Willow Spring	36 25 00	115 45 47	-98.0	-13.4	06-26-85	2
Willow Spring	36 25 00	115 45 47	-98.0	-13.60	04-05-88	1
Young Well	36 25 11	115 22 58	-93.0	-12.80	09-25-86	1
Crystal Pool Spring	36 25 14	116 19 21	-106.5	--	08-00-68	3
Crystal Pool Spring	36 25 14	116 19 21	-103.0	--	01-00-69	3
Crystal Pool Spring	36 25 14	116 19 21	-104.0	--	03-00-70	3

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
Crystal Pool Spring	36 25 14	116 19 21	-102.0	-13.7	05-24-73	6
Crystal Pool Spring	36 25 14	116 19 21	-104.0	--	03-04-75	6
Brooks Well	36 25 20	115 22 40	-93.0	-12.7	06-24-86	2
Valley of Fire Well	36 25 21	114 32 52	-82.0	-10.6	06-24-85	2
Shown Well	36 25 31	115 22 44	-94.0	-13.00	09-25-86	1
Amargosa Well #20 (Ash Tree Spring)	36 25 35	116 24 32	-102.0	-12.4	03-06-74	8
Scruggs Spring	36 25 59	116 18 28	-103.0	--	03-10-75	6
Corn Creek Spring Well	36 26 20	115 21 26	-95.0	--	01-00-69	3
Corn Creek Spring Well	36 26 20	115 21 26	-94.0	--	03-00-70	3
Corn Creek Spring	36 26 20	115 21 26	-96.0	-13.0	06-29-85	2
Corn Creek Spring	36 26 20	115 21 26	-93.0	-12.85	09-25-86	1
Corn Creek Spring	36 26 20	115 21 26	-93.0	-12.9	06-17-87	1
Corn Creek Spring	36 26 20	115 21 26	-93.5	-12.9	01-05-88	1
Travertine Spring	36 26 27	116 49 49	-105.5	--	03-00-70	3
Travertine Spring	36 26 27	116 49 49	-102.0	-13.7	04-22-82	1
Big Timber Spring	36 26 42	115 55 37	-93.0	-13.3	06-27-85	2
Dry Lake Valley Well	36 27 18	114 50 38	-97.5	-13.30	07-01-85	1
Grapevine Springs	36 27 27	116 01 35	-92.5	-12.75	06-28-85	1
Texas Spring	36 27 28	116 50 11	-105.0	--	03-00-70	3
Texas Spring	36 27 28	116 50 11	-102.0	-13.7	04-22-82	1
Corn Creek Well	36 27 53	115 23 54	-95.5	-13.55	12-02-86	2
Longstreet Spring	36 28 03	116 19 31	-103.0	--	03-10-75	6
Silver Flag Alpha Well	36 28 34	115 26 45	-99.0	-13.7	12-18-87	2
Rogers Spring	36 28 46	116 19 31	-102.0	--	03-09-75	6
Amargosa Well #18	36 29 04	116 28 08	-102.0	-13.0	03-06-74	8
Fairbanks Spring	36 29 25	116 20 29	-103.0	--	08-00-68	3
Fairbanks Spring	36 29 25	116 20 29	-103.0	--	01-00-69	3
Fairbanks Spring	36 29 25	116 20 29	-104.0	--	03-00-70	3
Fairbanks Spring	36 29 25	116 20 29	-103.0	-13.6	05-23-73	6
Amargosa Well #17	36 29 38	116 27 00	-105.0	-12.8	03-01-74	8
Amargosa Well #16	36 29 38	116 30 01	-104.0	-12.7	03-01-74	8
Amargosa Well #15	36 30 28	116 30 25	-104.0	-13.0	03-05-74	8
Nevares Spring	36 30 44	116 49 14	-105.5	--	03-00-70	3
Nevares Spring	36 30 44	116 49 14	-104.0	-13.6	04-22-82	1
Divide Well	36 30 45	115 28 05	-98.0	-13.7	07-26-87	2
Indian Springs Prison Well #1	36 30 52	115 33 15	-102.0	-13.7	06-26-85	2
Amargosa Well #30	36 31 24	116 24 02	-104.0	-13.7	06-24-79	8
Amargosa Well #14	36 31 28	116 30 24	-98.5	-12.6	03-04-74	8
Mathew's Well	36 31 32	116 24 00	-104.0	-13.7	05-11-81	1
Old Dry Well	36 31 35	115 28 13	-95.0	-13.2	12-19-87	2
Alpha Post Well	36 32 06	115 33 55	-98.0	-13.2	06-23-86	2
South Black Hills Well	36 32 12	115 24 03	-87.0	-12.10	08-05-87	1
South Black Hills Well	36 32 12	115 24 03	-88	-11.9	08-05-87	2
Amargosa Well #13	36 32 19	116 30 24	-102.0	-13.0	03-05-74	8
Amargosa Well #21	36 32 48	116 25 07	-99.0	-13.2	06-25-79	8
Amargosa Well #11	36 32 49	116 29 19	-101.0	-13.1	03-05-74	8
Rancho Amargosa Well	36 32 52	116 32 30	-99.0	-11.9	05-12-81	1
South Hidden Valley Well	36 33 08	114 55 30	-90.5	-11.20	03-28-86	1
Amargosa Well #10	36 33 13	116 28 12	-97.5	-13.2	06-26-79	8
Amargosa Well #9	36 33 16	116 29 45	-102.0	-12.6	03-01-74	8

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
Albitre Well (Formerly Thiede's Well)	36 33 20	116 28 09	-97.5	-11.6	05-06-81	1
Amargosa Well #50	36 33 25	116 35 50	-104.0	-13.6	06-25-79	8
Indian Springs #2	36 33 54	115 40 06	-98.0	--	08-00-68	3
Indian Springs #2	36 33 54	115 40 06	-100.0	--	01-00-69	3
Indian Springs #2	36 33 54	115 40 06	-101.0	--	03-00-70	3
Indian Springs #2	36 33 54	115 40 06	-93.0	-11.9	06-23-86	2
Indian Springs #2	36 33 54	115 40 06	-97.0	-12.9	06-18-87	1
Indian Springs #2	36 33 54	115 40 06	-101.5	-13.7	01-04-88	1
Indian Springs #1	36 33 56	115 40 06	-96.5	-12.7	06-18-87	1
Indian Springs #1	36 33 56	115 40 06	-101.0	-13.7	01-04-88	1
Indian Springs #3	36 33 56	115 40 05	-93.5	-12.6	06-18-87	1
Indian Springs #3	36 33 56	115 40 05	-100.0	-13.5	01-04-88	1
Amargosa Well #23	36 33 58	116 32 37	-103.0	-13.4	03-31-71	8
Amargosa Well #29	36 33 59	116 26 12	-105.0	-13.8	03-31-71	8
Nichols' Well	36 34 05	116 32 40	-103.0	-13.0	05-06-81	1
Amargosa Well #8	36 34 10	116 27 35	-103.0	-13.4	03-01-74	8
Amargosa Well #25	36 34 24	116 33 25	-102.0	-13.4	03-31-71	8
Cook's Well	36 34 25	116 23 50	-104.0	-13.4	05-08-81	1
Fox Well (Formerly Kirker's Well)	36 34 25	116 33 20	-101.0	-12.2	05-10-81	1
Amargosa Well #27	36 34 37	116 25 19	-105.0	-13.8	04-01-71	8
Indian Springs AFB Well #1	36 34 47	115 40 47	-96.0	-13.0	06-27-85	2
Amargosa Well #47	36 34 49	116 36 38	-102.0	-13.1	03-31-71	8
Amargosa Well #5	36 34 56	116 28 41	-99.5	-13.2	11-17-72	8
Cow Camp Spring	36 35 01	115 18 26	-90.5	-12.6	10-28-81	1
Cow Camp Spring	36 35 01	115 18 26	-93.0	-12.6	05-10-83	1
Amargosa Well #4	36 35 28	116 28 42	-103.0	-13.2	03-04-74	8
Army Well No. 1	36 35 30	116 02 14	-104.0	--	01-00-69	3
Army Well No. 1	36 35 30	116 02 14	-103.0	--	03-00-70	3
Amargosa Well #3	36 37 18	116 26 32	-102.0	-12.8	11-20-72	8
Wiregrass Spring	36 38 00	115 12 29	-94.0	-12.8	10-28-81	1
Wiregrass Spring	36 38 00	115 12 29	-96.0	-12.7	05-11-83	1
Wiregrass Spring	36 38 00	115 12 29	-94.0	-12.85	10-09-86	1
Wiregrass Spring	36 38 00	115 12 29	-94.0	-12.9	10-09-86	2
Wiregrass Spring	36 38 00	115 12 29	-91.5	-12.80	03-20-87	1
Wiregrass Spring	36 38 00	115 12 29	-92.0	-12.50	06-17-87	1
Wiregrass Spring	36 38 00	115 12 29	-94.0	-12.8	08-04-87	1
Wiregrass Spring	36 38 00	115 12 29	-97.0	-12.8	01-05-88	1
Wiregrass Spring	36 38 00	115 12 29	-95.5	-12.95	04-06-88	1
Wiregrass Spring	36 38 00	115 12 29	-94.5	-12.85	12-12-88	1
Juanita Spring	36 38 13	114 14 51	-87.0	-11.65	01-25-86	1
Wamp Spring	36 38 30	115 04 12	-81.0	-10.60	03-20-87	1
Mormon Well Spring	36 38 38	115 05 52	-92.5	-12.9	10-27-81	1
Mormon Well Spring	36 38 38	115 05 52	-91.0	-12.5	05-09-83	1
Mormon Well Spring	36 38 38	115 05 52	-92.0	-12.6	10-07-87	1
Weiser Wash EH-2 (235 feet)	36 38 50	114 38 55	-70.0	-6.7	10-05-85	13
Weiser Wash EH-2 (255 feet)	36 38 50	114 38 55	-69.0	--	10-05-85	13
Weiser Wash EH-2 (295 feet)	36 38 50	114 38 55	-71.0	--	10-05-85	13
Weiser Wash EH-2 (335 feet)	36 38 50	114 38 55	-79.0	-9.3	10-05-85	13
Weiser Wash EH-2 (375 feet)	36 38 50	114 38 55	-86.0	--	10-05-85	13
Weiser Wash EH-2 (395 feet)	36 38 50	114 38 55	-97.0	--	10-05-85	13

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
Weiser Wash EH-2 (415 feet)	36 38 50	114 38 55	-101.0	--	10-05-85	13
Weiser Wash EH-2 (495 feet)	36 38 50	114 38 55	-97.0	--	10-05-85	13
Weiser Wash EH-2 (535 feet)	36 38 50	114 38 55	-99.0	-12.9	10-06-85	13
Weiser Wash EH-2 (555 feet)	36 38 50	114 38 55	-100.0	--	10-06-85	13
Weiser Wash EH-2 (655 feet)	36 38 50	114 38 55	-103.0	-13.0	10-06-85	13
Weiser Wash EH-2 (675 feet)	36 38 50	114 38 55	-105.0	-14.3	10-06-85	13
Weiser Wash EH-2 (715 feet)	36 38 50	114 38 55	-102.0	--	10-00-85	13
Weiser Wash EH-2 (735 feet)	36 38 50	114 38 55	-102.0	--	10-00-85	13
Weiser Wash EH-2 (755 feet)	36 38 50	114 38 55	-96.5	--	10-00-85	13
Weiser Wash EH-2 (775 feet)	36 38 50	114 38 55	-102.0	--	10-00-85	13
Weiser Wash EH-2 (795 feet)	36 38 50	114 38 55	-101.0	--	10-00-85	13
Weiser Wash EH-2 (815 feet)	36 38 50	114 38 55	-99.5	--	10-00-85	13
Weiser Wash EH-2 (835 feet)	36 38 50	114 38 55	-92.0	--	10-00-85	13
Weiser Wash EH-2 (875 feet)	36 38 50	114 38 55	-101.5	-13.4	10-08-85	13
Weiser Wash EH-2 (895 feet)	36 38 50	114 38 55	-99.0	--	10-00-85	13
Weiser Wash EH-2 (955 feet)	36 38 50	114 38 55	-102.0	--	10-00-85	13
Weiser Wash EH-2 (981 feet)	36 38 50	114 38 55	-98.0	--	10-00-85	13
Weiser Wash EH-2 (997 feet)	36 38 50	114 38 55	-104.0	--	10-00-85	13
Weiser Wash EH-2 (1,017 feet)	36 38 50	114 38 55	-90.0	--	10-00-85	13
Weiser Wash EH-2 (1,055 feet)	36 38 50	114 38 55	-96.5	--	10-00-85	13
Weiser Wash EH-2 (1,075 feet)	36 38 50	114 38 55	-92.0	--	10-00-85	13
Weiser Wash EH-2 (1,095 feet)	36 38 50	114 38 55	-99.0	-12.1	10-14-85	13
Weiser Wash EH-1 (drilling water)	36 39 37	114 37 52	-97.0	-13.0	--	13
Weiser Wash EH-1 (105 feet)	36 39 37	114 37 52	-95.0	--	10-00-85	13
Weiser Wash EH-1 (135 feet)	36 39 37	114 37 52	-98.0	-13.0	10-02-85	13
Weiser Wash EH-1 (175 feet)	36 39 37	114 37 52	-95.0	--	10-00-85	13
Weiser Wash EH-1 (215 feet)	36 39 37	114 37 52	-95.0	--	10-00-85	13
Weiser Wash EH-1 (235 feet)	36 39 37	114 37 52	-94.0	--	10-00-85	13
Weiser Wash EH-1 (255 feet)	36 39 37	114 37 52	-96.0	--	10-00-85	13
Weiser Wash EH-1 (295 feet)	36 39 37	114 37 52	-96.0	-13.4	10-03-85	13
Weiser Wash EH-7 (175 feet)	36 40 14	114 31 53	-91.5	-12.3	04-05-86	13
Weiser Wash EH-7 (305 feet)	36 40 14	114 31 53	-90.0	-13.5	04-06-86	13
Weiser Wash EH-7 (405 feet)	36 40 14	114 31 53	-92.0	-12.5	04-09-86	13
Weiser Wash EH-7 (505 feet)	36 40 14	114 31 53	-93.5	-12.2	04-09-86	13
Weiser Wash EH-7 (615 feet)	36 40 14	114 31 53	-96.0	-13.1	04-00-86	13
Weiser Wash EH-7 (USGS composite sample)	36 40 14	114 31 53	-91.0	-12.45	03-19-87	1
Keane Wonder Spring	36 40 25	116 55 11	-99.5	-13.0	04-23-82	1
Weiser Wash EH-8 (drilling water)	36 40 26	114 34 33	-94.0	-12.1	05-10-86	13
Weiser Wash EH-8 (115 feet)	36 40 26	114 34 33	-90.0	-12.5	05-10-86	13
Weiser Wash EH-8 (175 feet)	36 40 26	114 34 33	-92.0	-12.5	05-10-86	13
Weiser Wash EH-8 (195 feet)	36 40 26	114 34 33	-97.0	-13.7	05-10-86	13
Weiser Wash EH-8 (225 feet)	36 40 26	114 34 33	-97.0	-13.8	05-10-86	13
Weiser Wash EH-8 (244 feet)	36 40 26	114 34 33	-96.0	-13.6	05-10-86	13
Sawmill Spring (Sheep Range)	36 40 50	115 10 34	-92.0	-12.85	05-19-88	1
Weiser Wash EH-6 (85 feet)	36 40 54	114 34 12	-86.0	-11.3	03-24-86	13
Weiser Wash EH-6 (145 feet)	36 40 54	114 34 12	-94.5	-12.8	03-24-86	13
Weiser Wash EH-6 (295 feet)	36 40 54	114 34 12	-105.0	-14.3	03-25-86	13
Weiser Wash EH-6 (304 feet)	36 40 54	114 34 12	-99.0	-12.3	03-25-86	13
Weiser Wash EH-6 (335 feet)	36 40 54	114 34 12	-99.0	-14.6	03-26-86	13
Weiser Wash EH-6 (455 feet)	36 40 54	114 34 12	-100.0	-13.1	03-26-86	13

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
CSV-3 Well	36 41 27	114 55 30	-75.0	-10.3	10-07-87	1
Weiser Wash EH-3 (295 feet)	36 41 32	114 31 32	-92.0	-12.1	02-02-86	13
Weiser Wash EH-3 (355 feet)	36 41 32	114 31 32	-90.0	-12.7	02-02-86	13
Weiser Wash EH-3 (455 feet)	36 41 32	114 31 32	-90.0	-12.2	02-02-86	13
Weiser Wash EH-3 (545 feet)	36 41 32	114 31 32	-93.0	-13.4	02-02-86	13
Weiser Wash EH-3 (795 feet)	36 41 32	114 31 32	-91.0	-13.2	02-05-86	13
Weiser Wash EH-4 (165 feet)	36 42 23	114 42 58	-100.0	-13.2	03-18-86	13
Weiser Wash EH-4 (285 feet)	36 42 23	114 42 58	-96.0	-12.8	03-18-86	13
White Rock Spring	36 42 30	115 14 20	-82.0	-9.9	10-29-81	1
White Rock Spring	36 42 30	115 14 20	-85.0	-9.8	05-10-83	1
Pederson's Warm Spring (M-13)	36 42 36	114 42 54	-98.0	--	01-00-69	3
Pederson's Warm Spring (M-13)	36 42 36	114 42 54	-97.0	--	03-00-70	3
Pederson's Warm Spring (M-13)	36 42 36	114 42 54	-97.0	-12.75	10-30-85	1
Iverson's Spring	36 42 37	114 42 43	-97.0	--	03-00-70	3
Spring Feeding Moapa	36 42 53	114 43 05	-97.0	--	03-00-70	3
Big Muddy Spring Area (M-8)	36 43 15	114 43 39	-99.0	-12.75	10-30-85	1
Big Muddy Spring	36 43 20	114 42 48	-98.0	--	03-00-70	3
Big Muddy Spring	36 43 20	114 42 48	-96.5	-12.9	07-22-81	1
Big Muddy Spring	36 43 20	114 42 48	-98.0	-12.75	10-30-85	1
Big Muddy Spring	36 43 20	114 42 48	-99.0	-13.0	01-07-88	1
Big Muddy Spring	36 43 20	114 42 48	-98.0	-14.0	--	2
Big Muddy Spring Area (M-9)	36 43 33	114 43 38	-96.5	-12.45	10-30-85	1
Weiser Wash EH-5A (65 feet)	36 43 58	114 44 36	-99.0	-13.1	03-05-86	13
Weiser Wash EH-5A (205 feet)	36 43 58	114 44 36	-98.0	-12.9	03-05-86	13
Weiser Wash EH-5A (265 feet)	36 43 58	114 44 36	-107.0	-13.7	03-12-86	13
Well J-12 (NTS)	36 45 54	116 23 24	-99.0	--	01-00-69	3
Well J-12 (NTS)	36 45 54	116 23 24	-97.5	-12.8	03-26-71	10
Nuclear Engr. Co. Well	36 45 58	116 41 10	-108.5	--	12-00-68	3
ERTEC MX-6 Well	36 46 04	114 47 13	-97.0	-12.95	09-28-86	1
ERTEC MX-6 Well	36 46 04	114 47 13	-99.0	-13.1	09-28-86	2
CSV-2 Well	36 46 50	114 43 20	-98.0	-12.85	01-26-86	1
USW VH-1 (Amargosa Well #55)	36 47 32	116 33 07	-108.0	-14.2	02-11-81	10
ERTEC MX-4 Well	36 47 44	114 53 32	-102.5	-13.0	12-23-80	1
ERTEC MX-5 Well	36 47 44	114 53 32	-99.5	-12.9	07-22-81	1
Well 5B	36 48 05	115 58 08	-106.5	--	01-00-69	3
Well J-13	36 48 29	116 23 40	-97.5	-13.0	03-26-71	10
Railroad Well Farrier	36 48 49	114 39 14	-97.5	-12.5	02-04-84	1
Sand Spring	36 49 30	115 34 05	-88	-11.9	01-03-88	2
UE-25P-1	36 49 38	116 25 21	-107.5	-13.7	02-11-83	11
UE-25P-1	36 49 38	116 25 21	-106.0	-13.8	05-12-83	11
USW H-3	36 49 42	116 27 60	-101.0	-13.9	03-13-84	1
UE-25C-1	36 49 47	116 25 43	-102.0	-13.5	09-30-83	1
UE-25C-2	36 49 47	116 25 43	-101.0	-13.4	03-13-84	1
UE-25C-3	36 49 47	116 25 43	-103.0	-13.5	05-09-84	1
USW H-4	36 50 32	116 26 54	-104.0	-14.0	05-17-82	10
USW H-6	36 50 49	116 28 55	-106.0	-13.8	10-16-82	10
Tim Spring	36 50 58	115 34 10	-99	-13.2	01-02-88	2
UE-25B-1 (Amargosa Well #57)	36 51 08	116 26 23	-99.5	-13.4	08-07-81	10
UE-25B-1 (Amargosa Well #57)	36 51 08	116 26 23	-101.0	-13.4	09-01-81	10
UE-25B-1 (Amargosa Well #57)	36 51 08	116 26 23	-99.5	-13.5	07-20-82	10

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
USW G-4	36 51 14	116 27 04	-103.0	-13.8	12-09-82	10
USW H-5	36 51 22	116 27 55	-102.0	-13.6	07-03-82	10
USW H-5	36 51 22	116 27 55	-102.0	-13.6	07-26-82	10
USW H-1 (Amargosa Well #56)	36 51 58	116 27 12	-102.0	-13.4	10-01-80	8
USW H-1 (Amargosa Well #56)	36 51 58	116 27 12	-103.0	-13.4	10-20-80	10
USW H-1 (Amargosa Well #56)	36 51 58	116 27 12	-101.0	-13.5	12-08-80	10
Dejesus Spring	36 52 28	115 34 45	-100	-13.1	05-09-87	2
VF-2 Well	36 52 30	114 56 44	-101.0	-12.95	02-05-86	1
VF-2 Well	36 52 30	114 56 44	-101.0	-13.1	01-06-88	1
VF-1 Well	36 52 32	114 55 44	-94.0	-12.6	01-06-88	1
Well 12S/47E/19ADC	36 52 41	116 45 19	-104.0	-13.3	--	7
Well 12S/47E/20BBB	36 53 01	116 45 02	-106.0	-13.6	--	7
Littlefield Road-Cut Spring	36 53 42	113 54 54	-97.5	-12.8	07-01-88	1
Sheep Spring (Sheep Range)	36 53 42	115 06 53	-96.0	-13.35	05-19-88	1
Well 12S/47E/7DBA	36 54 19	116 45 26	-108.0	-13.9	--	7
Well 12S/47E/6CDD	36 54 52	116 45 42	-102.0	-13.3	--	7
Hackberry Spring	36 55 04	114 26 16	-87.0	-12.3	02-05-84	1
Well C-1	36 55 07	116 00 34	-110.5	--	01-00-69	3
Well 11S/47E/32DDA	36 55 52	116 44 03	-102.0	-13.4	--	7
Horse Spring	36 56 29	114 26 47	-89.0	-12.7	02-05-84	1
UE-29A-2	36 56 29	116 22 26	-93.5	-12.8	01-08-82	10
UE-29A-2	36 56 29	116 22 26	-93.0	-12.8	01-15-82	10
UE-29A-1	36 56 29	116 22 26	-92.0	-12.1	01-29-82	1
Lamb Spring (Sheep Range)	36 56 42	115 06 21	-92.5	-13.15	05-19-88	1
Well 11S/47E/28DAC	36 56 53	116 43 06	-109.0	-14.1	--	7
Well 11S/47E/28ACC	36 57 07	116 43 22	-108.0	-14.1	--	7
Desert Dry Lake Well	36 57 11	115 11 51	-98.0	-13.10	03-18-87	1
Peach Spring	36 57 16	114 17 23	-76.5	-10.4	02-06-84	1
Gourd Spring	36 57 31	114 17 30	-77.5	-10.6	02-06-84	1
Well 11S/47E/21DBB	36 57 55	116 43 22	-108.0	-14.0	--	7
Davies Spring	36 57 56	114 30 07	-89.0	-12.5	02-06-84	1
Well 11S/47E/16DCA	36 58 36	116 43 13	-110.0	--	--	7
Quartz Spring	36 59 10	115 36 00	-88	-11.6	01-02-88	2
Well 11S/47E/5CDA	37 00 25	116 44 35	-108.0	-14.1	--	7
Well 11S/47E/4CAD	37 00 32	116 43 29	-108.0	-14.0	--	7
Well 10S/47E/33AAB	37 01 59	116 42 29	-108.0	-14.0	--	7
Well 10S/47E/31AAB	37 01 59	116 45 06	-102.0	-13.3	--	7
UE-16F	37 02 09	116 09 25	-105.0	-14.9	09-25-77	12
Well 10S/47E/27CBA	37 02 26	116 41 39	-110.0	-14.3	--	7
Well 10S/47E/30BBC	37 02 45	116 46 05	-102.0	-13.4	--	7
Well 10S/47E/14BAB	37 04 37	116 37 40	-112.0	-14.5	--	7
Willow Spring KSV-1	37 05 34	114 49 52	-88.0	-11.6	02-03-84	1
Grapevine Spring KSV-2	37 08 08	114 42 02	-87.5	-12.0	02-03-84	1
Well #8 (NTS) (Amargosa Well #64)	37 09 56	116 17 21	-104.0	-13.0	03-24-71	8
Snow Spring	37 10 49	114 07 53	-79.5	-10.0	11-13-86	1
Jensen Well	37 11 03	114 27 52	-88.5	-11.6	04-10-85	1
Maynard Lake Spring	37 11 30	115 02 02	-94.0	-12.3	01-14-85	2
Lone Tree Spring	37 12 07	115 03 32	-89.5	-10.9	01-14-85	2
U-20A2	37 14 34	116 25 51	-114.0	-14.8	--	7,14
Kane Spring KSV-3	37 14 46	114 42 21	-86.5	-11.9	02-02-84	1
Boulder Spring KSV-4	37 16 12	114 38 44	-87.0	-12.6	02-02-84	1
Irrigation Well	37 16 46	115 07 11	-91	--	--	15
UE-19E	37 17 50	116 19 59	-109.5	-14.0	--	7,14
UE-19GS	37 18 30	116 21 53	-113.5	-14.5	--	7,14
Randono Well	37 19 26	114 30 08	-87.5	-11.7	02-03-84	1

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
Cane Spring	37 20 27	115 44 50	-88	-9.5	05-15-85	2
Bradshaw Well	37 20 57	114 32 38	-88.5	-11.4	02-01-84	1
Stock Well	37 20 58	114 45 30	-88	--	--	15
Railroad Well	37 21 04	114 32 02	-86.0	-11.6	01-31-84	1
Upper Riggs Spring	37 22 06	114 38 52	-88.0	-11.9	02-02-84	1
Sheep Spring	37 24 02	114 16 37	-87.0	-12.0	06-03-85	1
Cattle Spring	37 24 57	115 45 05	-85	-9.4	05-15-85	2
Bishop Spring	37 25 07	114 38 26	-85.5	-11.7	02-02-84	1
Rock Spring	37 25 53	115 41 23	-86	-10.9	05-15-85	2
Quail Spring	37 26 29	115 41 01	-92	-11.9	05-15-85	2
Hells Acre Gulch Spring	37 27 37	115 07 29	-93.0	-12.3	01-14-85	2
Ash Spring	37 27 49	115 11 34	-107.0	--	08-00-68	3
Ash Spring	37 27 49	115 11 34	-109.0	--	01-00-69	3
Ash Spring	37 27 49	115 11 34	-112.0	--	03-00-70	3
Ash Spring	37 27 49	115 11 34	-108.0	-14.1	07-20-81	1
Ash Spring	37 27 49	115 11 34	-108	-12.3	--	2
Indian Spring	37 27 50	115 44 28	-88	-10.4	05-15-85	2
Rabbitbrush Spring	37 28 56	115 41 28	-89	-11.2	05-14-85	2
Bullwhack Spring	37 29 45	115 45 47	-88	-10.2	05-17-85	2
North Ella Spring	37 29 57	114 27 09	-86.5	-11.6	06-03-85	1
Ramone Mathews Well	37 31 36	114 14 39	-92.0	-12.3	06-03-85	1
Crystal Spring	37 31 53	115 13 58	-109.0	--	08-00-68	3
Crystal Spring	37 31 53	115 13 58	-110.0	--	01-00-69	3
Crystal Spring	37 31 53	115 13 58	-109.0	--	03-00-70	3
Crystal Spring	37 31 53	115 13 58	-109.0	-14.3	07-20-81	1
Crystal Spring	37 31 53	115 13 58	-110	--	--	2
April Fool Spring	37 32 14	115 44 19	-89	-10.7	05-16-85	2
Grassy Spring	37 32 28	114 47 27	-85.0	-10.9	01-14-85	2
Acoma Well	37 32 55	114 10 23	-95.0	-12.6	06-03-85	1
Meadow V. Wash Below Caliente	37 33 27	114 33 54	-93.5	-12.3	11-13-86	1
Hiko Spring	37 36 34	115 12 51	-111.0	--	08-00-68	3
Hiko Spring	37 36 34	115 12 51	-110.0	--	01-00-69	3
Hiko Spring	37 36 34	115 12 51	-110.0	--	03-00-70	3
Hiko Spring	37 36 34	115 12 51	-110.0	-15.3	01-14-85	2
Hiko Spring (Duplicate Sample)	37 36 34	115 12 51	-105.0	-14.0	01-14-85	2
Clover Creek	37 36 47	114 28 21	-87.5	-11.8	11-13-86	1
Caliente City Well	37 36 57	114 30 48	-89.0	-12.4	01-31-84	1
Caliente Hot Spring	37 37 16	114 30 34	-109.0	-14.5	04-10-85	1
Irrigation Well	37 38 10	115 12 54	-109	--	--	15
Meadow V. Wash At Cove Canyon	37 39 15	114 29 45	-98.0	-12.6	11-13-86	1
Pahroc Spring	37 39 52	114 58 47	-89.0	-12.5	01-14-85	2
SK-18 (MX 10" Well)	37 42 15	114 45 31	-95	--	--	15
Mustang Spring	37 44 09	114 55 14	-91.0	-12.6	01-14-85	2
The Seeps	37 44 22	115 34 32	-98.0	-13.3	01-15-85	2
Weaver Well	37 44 41	114 25 28	-101.0	-13.1	06-04-85	1
Cedar Spring	37 45 05	116 16 25	-101.0	-13.6	01-15-85	2
Well B-1 Tonopah Test Range Nellis B&G RG	37 45 06	116 29 23	-110.0	-14.0	09-10-80	1
Cedar Pass Well	37 45 07	116 28 57	-110.0	-14.0	--	2
John Wadsworth Well	37 46 07	114 24 25	-101.0	-12.9	06-04-85	1
Bennett Spring	37 47 03	114 31 41	-103.0	-13.7	04-10-85	1
Sandia 6	37 47 03	116 45 05	-124.0	-15.2	--	2
Lester Mathews Well	37 47 37	114 23 59	-103.0	-13.3	06-04-85	1
Panaca Town Well	37 47 50	114 23 57	-106.0	-14.0	06-04-85	1
Panaca Spring	37 48 24	114 22 47	-106.0	-13.9	04-26-84	1
Panaca Spring	37 48 24	114 22 47	-108.0	-14.0	04-08-85	1

APPENDIX A.—*Deuterium and oxygen-18 compositions of water from wells and springs in southern Nevada and southeastern California between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Deuterium (permil)	Oxygen-18 (permil)	Date	Source[1]
Panaca Spring	37 48 24	114 22 47	-106.5	-14.2	11-11-86	1
Panaca Spring	37 48 24	114 22 47	-107	-14.4		2
North Lee Well	37 49 28	114 23 04	-101.0	-13.3	06-04-85	1
Well 9	37 51 15	116 43 06	-113.0	-14.3		2
Delmues Spring	37 51 36	114 19 20	-104.0	-13.4	04-08-85	1
Georges Spring	37 51 37	116 20 57	-98.0	-13.6	01-15-85	2
Meadow Valley Wash Above Delmues Spring	37 51 40	114 19 18	-98.0	-12.8	04-08-85	1
Well 1A	37 53 00	116 46 53	-118.0	-14.5		2
Oxborrow Well	37 53 10	114 18 17	-92.0	-11.8	06-05-85	1
Flatnose Spring	37 53 46	114 13 33	-101.0	-13.4	04-08-85	1
White Rock Spring	37 53 46	115 01 11	-90.0	-12.1	01-13-85	2
Upper Conners Spring	37 54 10	114 33 38	-100.0	-13.9	11-11-86	1
Runoff At Pine Spring	37 54 29	114 32 56	-99.0	-13.4	04-07-85	1
Lime Spring	37 54 52	114 32 25	-97.0	-12.9	04-07-85	1
Deadman Spring	37 55 07	114 32 29	-99.0	-13.3	04-07-85	1
Oceana Spring	37 55 07	115 09 26	-87	--	--	15
Highland Spring	37 55 16	114 32 56	-98.5	-13.3	04-07-85	1
Highland Spring	37 55 16	114 32 56	-98.0	-13.2	11-11-86	1
Pioche Municipal Well	37 57 33	114 24 51	-108.5	-14.4	11-10-86	1
Edan Creek Ranch Spring	37 58 12	116 22 53	-99.0	-13.9	01-15-85	2

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[1] Sources of data:

1 U.S. Geological Survey (unpublished).
2 Desert Research Institute (unpublished).
3 Winograd and Friedman (1972). Deuterium values were multiplied by 0.97 to make comparable with deuterium data from the U.S. Geological Survey Laboratory, Reston, Va. (I.J. Winograd, U.S. Geological Survey, written commun., 1985).
4 I.J. Winograd, (U.S. Geological Survey, written commun., 1989).
5 I.J. Winograd, (U.S. Geological Survey, written commun., 1986).
6 Winograd and Pearson (1976).
7 Waddell and others (1984).
8 Claassen (1985).
9 McKay and Zimmerman (1983).
10 Benson and others (1983).
11 Craig and Robison (1984).
12 Dinwiddie and Weir (1979).
13 Schroth (1987).
14 White and Chuma (1987).
15 Kirk (1987).
16 Noack (1988).

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*

[For Las Vegas Valley, chemical data are included only for sites that also have deuterium and oxygen-18 data. Deuterium and oxygen-18 are reported relative to V-SMOW, carbon-13 is reported relative to peedee belemnite, and carbon-14 is reported as percent modern of National Bureau of Standards 1950 oxalic acid. Units of measure: milligrams per liter, except as indicated. Number in parentheses indicates number of samples if different from total number of samples. Abbreviations: °C, degrees Celsius; µS/cm, microsiemens per centimeter at 25°Celsius; pmc, percent modern carbon; pCi/L, picocuries per liter; L, measured in laboratory]

Site	Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Number of samples	Temperature (°C)	pH	Specific conductance (µS/cm)	Calcium	Magnesium	Sodium	Potassium
1	Big Bend Well 1 Area 2	35 07 21	114 38 03	4	23.0	7.7	1,400	120	36	132	4.8
2	B. Laughlin (Knight) Well	35 07 23	114 36 42	22	18.3	7.8	1,300	114	33	112	4.5
3	Cromer Well	35 07 26	114 37 55	23	19.1	7.8	1,200	98	32	115	3.6
4	Sundance Shores Well	35 09 10	114 34 40	30	22.9	7.8	1,200	76	18	151	3.8
5	Nevada Club Well	35 09 20	114 34 19	22	18.8	7.8	1,400	113	30	146	4.9
6	Big Bend Well 1 Area 1	35 09 31	114 34 16	5	16.0	7.8	1,200	97	25	106	4.0
7	Riverside Trailer Court	35 09 37	114 34 15	24	20.6	7.8	1,300	112	31	125	4.9
8	Cresent Spring	35 28 43	115 10 47	1	20.5	7.8	990	108	25	37	5.0
9	Pine Spring	35 34 25	115 09 23	1	30.0	7.6	980	99	31	81	2.2
10	Ora Hana Spring	35 37 25	115 04 07	1	20.5	7.2	760	55	31	43	4.8
11	Saratoga Spring	35 40 53	116 25 17	1	28.5	8.0	4,700	30	37	980	32
12	McClanahan Spring	35 41 42	115 11 05	1	25.0	8.0	620	58	22	48	7.6
13	Jean Prison Well	35 47 18	115 20 43	1	22.0	6.8	850	60	41	30	3.8
14	Rosechrist Well	35 48 18	115 41 41	1	20.0	7.3	490	53	38	17	1.6
15	Castillo Well	35 50 02	115 26 09	1	24.0	7.3	1,400	97	68	86	4.1
16	Tecopa Hot Spring	35 52 19	116 13 50	1	42.0	8.2	3,600	4.0	1.5	850	16
17	Bird Spring	35 53 20	115 22 12	1	29.0	7.6	640	35	39	36	3.0
18	Wilson's Tank	35 56 02	115 25 28	1	29.0	7.4	830	68	57	36	1.7
19	Bighorn Sheep Spring (Black Canyon #12)	35 56 22	114 44 06	1	36.0	8.7	--	4.2	.2	156	1.1
20	Ringbolt Rapid Spring (Black Canyon #10)	35 57 39	114 43 26	1	50.0	7.8	--	239	14	592	12
21	Sky Harbor Airport	35 58 16	115 08 50	1	25.6	7.8	1,200	64	31	130	10
22	Palm Tree Cold Springs (Black Canyon #7)	35 59 42	114 44 15	1	24.0	7.8	--	389	39	986	16
23	Tenaya Well	36 00 42	115 15 05	1	23.0	7.3	1,300	110	56	31	4.6
24	Showboat Country Club #2	36 02 51	115 04 48	1	26.7	7.9	1,100	74	38	98	14
25	Genstar Gypsum Plant Well	36 03 04	115 23 43	1	23.7	7.0	710	89	38	9.9	1.6
26	Flowing Well (DRI LG153)	36 03 22	115 03 08	1	24.0	8.3	1,100	40	22	130	14
27	Sandstone Spring #1	36 03 47	115 28 09	1	17.0	7.0	460	53	29	5.9	2.0
28	Sunset Park Well	36 03 49	115 05 51	1	26.0	7.3	1,700	160	58	50	15
29	Spanish Trail Country Club Well	36 05 56	115 15 38	1	28.0	7.2	1,300	175	66	14	3.4
30	Stocks Mill & Supply Co. Well	36 06 07	115 15 46	1	26.2	7.2	1,100	161	69	14	3.1

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Number of samples	Temperature (°C)	pH	Specific conductance (µS/cm)	Calcium	Magnesium	Sodium	Potassium
31	Tropicana Country Club South Well #1	36 06 22	115 10 09	1	23.0	7.6	760	78	38	18	3.2
32	Sands Hotel And Casino #3	36 07 28	115 10 09	1	22.5	7.7	510	57	29	7.5	2.5
33	Sparklettes Drinking Water Co.	36 07 41	115 11 45	1	22.7	7.7	460	51	27	5.9	2.5
34	BLM Visitors Center	36 07 44	115 26 03	1	25.0	7.4	600	85	30	8.3	1.6
35	Las Vegas Country Club	36 08 20	115 08 49	1	22.5	7.4	520	54	27	8.7	3.2
36	Hartzski Well	36 08 36	115 15 47	1	21.3	7.6	860	103	42	8.2	2.2
37	Red Spring	36 08 40	115 25 10	1	20.0	7.6	430	51	25	5.2	2.0
38	Manse Well	36 09 17	115 53 42	1	22.5	7.5	400	48	22	4.7	.9
39	DRI Well LG047	36 09 33	115 05 51	1	21.5	8.1	440	26	32	7.8	3.5
40	Shetland Mutual Water Users Association	36 09 39	115 10 33	1	22.3	7.7	470	49	27	7.3	2.6
41	LVVWD #1A	36 09 39	115 13 32	1	25.1	7.6	480	62	31	8.0	2.9
42	LVVWD #11A	36 09 52	115 11 22	1	26.0	6.9	460	50	27	6.9	3.4
43	Union Pacific Railroad #3	36 10 05	115 09 13	1	24.8	7.6	350	41	21	7.	2.7
44	LVVWD #7A	36 10 06	115 11 23	1	25.6	7.5	420	54	26	6.8	3.2
45	Sky Mt. Resort	36 10 13	115 34 44	1	14.0	7.3	470	80	22	4.8	.6
46	LVVWD #14	36 10 18	115 11 22	1	24.5	7.3	460	54	28	6.3	2.7
47	White Rock Spring	36 10 27	115 28 43	1	25.0	7.0	720	94	29	8.4	1.8
48	LVVWD #15A	36 10 31	115 11 23	1	24.9	7.0	410	50	25	5.6	2.3
49	LVVWD #34	36 10 31	115 11 06	1	24.0	7.5	440	46	24	7.1	3.1
50	LVVWD #16	36 10 31	115 11 39	1	24.8	7.5	400	55	25	6.	2.9
51	LVVWD #17	36 10 31	115 11 39	1	23.8	7.4	410	53	25	5.8	2.0
52	LVVWD #18A	36 11 12	115 14 38	1	23.0	7.4	500	56	2.8	6.0	2.0
53	LVVWD #45	36 11 50	115 12 13	1	25.4	7.0	410	49	25	5.8	1.6
54	LVVWD #22A	36 12 05	115 15 43	1	24.0	7.5	450	48	26	5.9	2.6
55	NLVWD Robinson Well	36 12 17	115 11 41	1	25.7	7.3	400	45	23	6.7	1.6
56	Pahrump Spring Well	36 12 27	115 59 01	1	25.0	7.6	420	47	24	4.7	1.3
57	NLVWD West Cheyenne Well	36 12 38	115 11 21	2	25.0	7.2	450	45	22	6.7	1.4
58	Pahrump Community Church	36 12 42	115 59 18	1	27.0	--	--	47	23	5.7	1.2
59	Nellis AFB #13	36 12 44	115 03 00	1	24.5	7.8	600	27	27	44	6.5
60	Trout Spring	36 13 22	115 40 59	8	7.5	8.4	--	45	10	.9	.4
61	Well 5 Franklin Lake Nr Death Valley	36 14 06	116 17 32	1	20.2	9.3	27,000	2.6	2.0	4,500	87
62	Well 10 Franklin Lake Nr Death Valley	36 14 06	116 17 39	1	19.5	9.3	22,000	4.0	4.4	6,200	230
63	NLVWD Desert Aire Well	36 14 15	115 12 16	1	26.4	7.4	370	44	22	6.8	1.3
64	Lake Mead Base Well #3	36 14 21	115 00 16	1	25.0	7.6	700	36	25	76	7.5
65	Mazie Spring	36 14 26	115 38 19	1	3.0	--	120	37	7.8	.7	.2

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Number of samples	Temperature (°C)	pH	Specific conductance (µS/cm)	Calcium	Magnesium	Sodium	Potassium
66	Craig Ranch Country Club #2	36 14 29	115 09 00	1	23.1	7.8	360	30	19	16	3.
67	Rainbow Spring #2	36 14 36	115 37 55	1	9.8	7.5	360	61	13	1.1	.5
68	Peak Spring Canyon Creek	36 14 40	115 43 09	1	7.0	8.7	400	64	13	1.4	.4
69	East Spring #2	36 14 45	115 39 07	1	4.0	--	100	40	8.0	.9	.2
70	GS-18 Franklin Lake Nr Death Valley	36 14 51	116 17 08	1	17.0	9.8	65,000	4.0	6.6	28,000	830
71	Nellis AFB #4	36 14 56	115 00 15	1	25.9	8.04	550	31	32	41	5.0
72	GS-13 Franklin Lake Nr Death Valley	36 14 57	116 18 34	1	20.9	8.4	3,300	9.3	5.2	750	21
73	West Spring	36 15 11	115 39 42	1	5.5	--	--	56	8.9	.7	.3
74	Rainbow Spring #1	36 15 14	115 39 23	1	5.0	--	160	53	7.4	.7	.3
75	Mt Charleston Lodge	36 15 24	115 38 50	5	--	--	290	50	8.8	1.0	.4
76	Rainbow Well	36 15 34	115 37 47	4	8.7	--	370	61	13	1.5	.5
77	Pat Well	36 15 45	115 19 37	1	17.5	8.1	600	50	28	6.1	2.1
78	Echo #3	36 15 55	115 39 29	4	8.6	--	350	59	11	1.4	.5
79	Echo Spring	36 15 55	115 39 29	1	8.0	--	360	60	11	1.3	.5
80	Daines	36 16 12	115 35 33	4	--	--	390	58	16	2.3	.5
81	Highway Maintenance Well	36 16 17	115 35 38	5	10.5	7.6	410	65	16	1.7	.5
82	Lower Stanley B Spring	36 16 14	115 38 18	1	10.0	--	430	81	41	1.4	.4
83	Fletcher Spring	36 16 26	115 37 33	1	11.5	--	380	84	24	2.5	.7
84	Taylor's Steak House Rest.	36 17 08	115 14 41	1	21.2	7.6	390	42	24	6.4	1.3
85	Martin Well	36 17 25	115 14 35	1	22.0	7.8	450	35	23	9.2	1.8
86	GS-8 Franklin Lake Nr Death Valley	36 17 27	116 17 09	1	18.9	8.5	3,600	3.9	5.2	760	71
87	T-Bar Spring	36 17 38	115 40 60	1	6.0	--	240	41	8.3	.8	.4
88	Cave Spring #1	36 17 40	115 39 55	2	5.3	--	240	51	10	1.0	.4
89	Grapevine Springs	36 18 03	115 29 25	2	18.5	7.5	630	62	28	20	2.2
90	Ski Lodge Well	36 18 11	115 40 44	1	8.0	--	340	47	14	.9	.4
91	Youth Camp	36 18 26	115 40 28	4	7.3	--	400	56	21	1.6	.5
92	Deer Creek Spring #1	36 18 27	115 38 13	1	4.3	7.9	270	45	10	1.3	.3
93	Deer Creek Spring #2	36 18 27	115 37 37	1	8.0	8.5	670	56	13	1.4	.5
94	Deer Cr Picnic Area Spg	36 18 33	115 37 25	2	7.5	--	290	56	16	1.1	.4
95	Racel Well	36 18 40	115 15 39	1	24.0	7.7	430	37	24	8.6	1.6
96	Gilbert Well	36 18 45	115 25 06	1	15.5	7.2	410	60	20	2.6	0.5
97	Lee's Crest	36 19 07	115 40 22	4	9.0	--	470	45	31	2.8	.8
98	Stewart Well	36 19 10	115 40 20	1	9.7	8.6	570	44	32	2.0	.9
99	Tule Spring State Park Well	36 19 14	115 16 00	1	21.0	7.6	520	40	26	8.6	1.8
100	Clark Spring	36 19 14	115 43 15	1	10.0	7.5	480	75	38	2.0	.6

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Number of samples	Temperature (°C)	pH	Specific conductance (μS/cm)	Calcium	Magnesium	Sodium	Potassium
101	Mulder Well	36 19 25	115 13 22	1	24.0	7.8	400	42	23	10	1.3
102	Mifflin Well	36 19 32	115 25 23	1	18.0	7.5	400	50	23	3.0	1.2
103	Lee's Oasis Well	36 19 48	115 18 55	1	20.0	8.2	530	63	26	4.6	.7
104	Holland Well	36 20 00	115 17 00	1	26.0	7.2	400	44	25	6.2	1.5
105	Summer Homes Well	36 20 06	115 39 18	3	9.0	7.4	390	51	26	1.5	.9
106	G.P. Apex Well	36 20 28	114 55 36	1	31.0	7.0	1,570	120	46	129	12
107	Cortney Well	36 20 42	115 20 52	1	20.5	7.2	510	67	25	4.8	.7
108	Adams Well	36 20 47	115 19 50	1	22.5	7.3	490	64	25	4.6	.7
109	Paiute Indian Reservation Well	36 21 02	115 20 33	1	18.7	7.0	460	60	22	3.5	.6
110	Big Spring	36 22 30	116 16 25	6	27.0	7.4	--	43	19	97	9.4
111	Rogers Spring	36 22 39	114 26 38	3	30.0	7.4	3,800	430	136	290	20
112	Grace Petroleum Arrow Canyon Water Well 1	36 22 58	114 55 00	1	26.5	7.3	1,600	120	46	140	16
113	Blue Point Spring	36 23 21	114 25 26	1	30.0	7.0	3,800	510	170	360	23
114	Genstar Well	36 23 29	114 54 14	1	24.0	7.4	1,500	120	47	140	1.3
115	Cold Creek Spring	36 24 50	115 44 20	9	10.0	7.6	--	69	16	1.6	.4
116	Amargosa Well #41	36 24 51	116 25 41	1	24.0	7.4	730	29	12	120	9.7
117	Willow Sring	36 25 00	115 45 47	1	10.5	7.5	720	72	15	2.1	.6
118	Young Well	36 25 11	115 22 58	1	17.0	7.7	380	20	38	7.1	2.7
119	Crystal Pool Spring	36 25 14	116 19 21	6	31.0	7.4	--	47	21	74	9.4
120	Brooks Well	36 25 20	115 22 40	1	24.6	8.2	490	40	43	7.1	1.4
121	Valley of Fire Well	36 25 21	114 32 52	1	28.0	7.4	1,100	118	53	39	8.2
122	Shown Well	36 25 31	115 22 44	1	17.0	7.3	380	38	40	7.2	2.0
123	Amargosa Well #20	36 25 35	116 24 42	1	--	7.9	370	14	4.4	46	8.0
124	Corn Creek Spring	36 26 20	115 21 26	4	21.0	7.4	540	46	33	6.2	2.0
125	Travertine Spring	36 26 27	116 49 49	1	30.5	7.8	1,000	39	20	160	11
126	Big Timber Spring	36 26 42	115 55 37	1	11.0	7.7	580	50	20	4.2	0.6
127	Dry Lake Valley Well	36 27 18	114 50 38	1	29.0	7.3	1,400	110	48	120	13
128	Grapevine Spring	36 27 27	116 01 35	1	20.0	7.7	670	57	40	27	1.5
129	Texas Spring	36 27 28	116 50 11	1	31.0	7.9	950	36	19	150	11
130	Corn Creek Well	36 27 53	115 23 54	1	18.5	8.0	380	17	27	13	2.5
131	Silver Flag Alpha Well	36 28 34	115 26 45	1	--	7.8	360	32	21	5.8	1.5
132	Amargosa Well #19	36 28 39	116 26 37	1	--	8.1	320	21	4.0	32	8.2
133	Amargosa Well #18	36 29 04	116 28 08	1	--	8.0	300	21	2.7	36	7.5
134	S17 E48 12 BC	36 29 20	116 31 10	1	--	7.7	1,300	74	24	160	16
135	Fairbanks Spring	36 29 25	116 20 29	8	27.0	7.4	--	48	20	69	8.2

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Number of samples	Temperature (°C)	pH	Specific conductance (µS/cm)	Calcium	Magnesium	Sodium	Potassium
136	Amargosa Well #36	36 29 36	116 25 15	1	--	8.1	800	40	14	97	14
137	Amargosa Well #16	36 29 38	116 30 01	1	--	8.0	360	24	1.8	48	7.3
138	Amargosa Well #17	36 29 40	116 26 58	1	--	7.9	420	25	3.6	48	9.7
139	Amargosa Well #15	36 30 28	116 30 25	1	--	8.0	310	19	1.5	40	7.1
140	Nevares Spring	36 30 44	116 49 14	1	36.5	7.7	980	43	20	140	11
141	Divide Well	36 30 45	115 28 05	1	19.0	8.1	370	22	21	18	3.4
142	Indian Springs Prison Well #1	36 30 52	115 33 15	1	23.0	7.8	410	39	21	3.6	1.
143	Amargosa Well #30	36 31 24	116 24 02	1	--	7.8	--	52	22	120	18
144	Amargosa Well #14	36 31 28	116 30 24	1	--	7.9	300	17	2.0	40	6.1
145	Mathew's Well	36 31 32	116 24 00	1	24.5	7.6	940	50	20	110	1.2
146	Old Dry Well	36 31 35	115 28 13	1	23.0	8.1	390	22	22	15	2.2
147	Alpha Post	36 32 06	115 33 55	1	25.1	8.7	330	35	22	4.9	1.
148	South Black Hills Well	36 32 12	115 24 03	1	29.0	7.5	400	39	21	7.3	4.
149	Amargosa Tracer Well 2	36 32 12	116 13 47	4	31.0	7.4	--	44	20	65	7.8
150	Amargosa Well #13	36 32 19	116 30 24	1	--	7.9	320	19	.8	43	7.3
151	Amargosa Well #21	36 32 48	116 25 07	1	--	8.2	--	16	1.7	56	6.4
152	Amargosa Well #11	36 32 49	116 29 19	1	--	7.9	320	24	1.1	36	8.2
153	South Hidden Valley Well	36 33 08	114 55 30	1	25.0	7.8	820	33	30	86	12
154	Amargosa Well #12	36 33 13	116 30 25	1	27.0	7.0	320	18	.7	54	6.9
155	Amargosa Well #10	36 33 13	116 28 12	1	--	7.9	--	30	1.9	40	4.3
156	Amargosa Well #9	36 33 23	116 29 44	1	--	8.0	340	20	2.7	42	8.8
157	Amargosa Well #50	36 33 25	116 35 50	1	--	7.7	--	53	8.5	150	11
158	Indian Springs #2	36 33 54	115 40 06	5	25.5	7.6	400	46	22	4.2	1.
159	Amargosa Well #23	36 33 58	116 32 37	1	25.5	8.1	--	9.6	3.2	58	5.9
160	Amargosa Well #29	36 33 59	116 26 12	1	23.8	7.7	--	41	7.5	80	9.8
161	Amargosa Well #8	36 34 10	116 27 35	1	--	7.6	430	23	2.6	56	9.0
162	Amargosa Well #25	36 34 24	116 33 25	1	24.5	8.3	--	9.2	3.9	61	5.5
163	Amargosa Well #27	36 34 37	116 25 19	1	30.6	7.6	--	48	17	111	13
164	Indian Springs AFB Well #1	36 34 47	115 40 47	1	23.5	7.3	1,300	126	66	43	4.5
165	Amargosa Well #47	36 34 49	116 36 38	1	24.2	7.7	--	47	16	130	9.4
166	Amargosa Well #5	36 34 56	116 28 41	1	23.0	7.0	310	30	2.6	37	5.6
167	Cow Camp Spring	36 35 01	115 18 26	2	12.3	7.6	540	49	33	23	.6
168	Amargosa Well #4	36 35 28	116 28 42	1	--	7.8	310	29	2.2	35	5.2
169	Army Well No. 1	36 35 30	116 02 14	3	31.0	7.4	--	45	22	38	5.5
170	Test Well 10	36 35 31	115 51 04	2	27.0	7.4	--	39	18	7.4	2.0

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Number of samples	Temperature (°C)	pH	Specific conductance (µS/cm)	Calcium	Magnesium	Sodium	Potassium
171	S15 E50 25BD 1	36 37 15	116 24 45	2	44.0	7.9	340	22	1.3	49	2.5
172	S15 E49 22DC 1 (BR supply)	36 37 40	116 26 39	1	--	7.8	340	28	2.1	39	4.9
173	Amargosa Well #3	36 37 50	116 20 00	1	--	6.7	330	27	2.0	43	4.6
174	Wiregrass Spring	36 38 00	115 12 29	8	8.6	7.3	560	70	33	3.0	1.7
175	Juanita Spring	36 38 13	114 14 51	1	26.0	7.3	940	130	43	25	5.3
176	Wamp Spring	36 38 30	115 04 12	1	7.0	8.1	320	71	13	10	2.1
177	Moorman Well	36 38 38	115 05 52	3	11.2	7.4	670	77	42	12	.6
178	Weiser Wash EH-2	36 38 50	114 38 55	3	24.0	8.2	890	38	26	101	15.6
179	Weiser Wash EH-1	36 39 37	114 37 52	2	--	8.2	3,000	262	116	278	26
180	Weiser Wash EH-7	36 40 14	114 31 53	5	23.2	7.7	3,900	481	198	163	24
181	Weiser Wash EH-8	36 40 26	114 34 33	3	--	7.6	3,800	375	104	416	22
182	Weiser Wash EH-6	36 40 54	114 34 12	3	24.8	7.7	3,000	341	131	274	31
183	CSV-3 Well	36 41 27	114 55 30	1	41.0	7.4	650	51	25	38	10
184	Weiser Wash EH-3	36 41 32	114 31 32	4	24.1	7.8	3,700	511	201	170	22
185	TH20 Nevada Power	36 41 49	114 35 33	1	24.0	7.1	3,500	195	147	497	13
186	Weiser Wash EH-4	36 42 23	114 42 58	2	22.8	8.3	920	49	30	90	12
187	White Rock Spring	36 42 30	115 14 20	2	12.5	7.5	420	37	30	14	7.2
188	B. Lewis Well	36 42 40	114 36 00	1	22.0	7.6	1,800	107	76	203	12
189	Pederson Warm Springs (M-13)	36 42 36	114 42 54	1	32.2	--	1,000	75	26	110	8.8
190	Big Muddy Spring	36 43 20	114 42 48	1	32.5	7.2	930	66	26	96	10
191	Weiser Wash EH-5A	36 43 58	114 44 36	2	28.0	8.2	890	49	26	96	12
192	Test Well F	36 45 34	116 06 59	2	64.5	7.4	650	45	16	62	9.0
193	J-12 Well	36 45 54	116 23 24	1	26.8	7.1	290	14	2.1	38	5.1
194	ERTEC MX-6 Well	36 46 04	114 47 13	1	33.5	7.2	980	58	25	87	10
195	CSV-2 Well	36 46 50	114 43 20	1	27.0	7.4	1,000	60	27	100	10
196	USW VH-1 (Amargosa Well #55)	36 47 32	116 33 07	3	35.4	7.6	--	10	1.5	79	1.9
197	ERTEC MX-4 anD MX-5 Wells	36 47 44	114 53 32	2	34.7	7.3	750	46	20	81	11
198	Well J-13	36 48 29	116 23 40	1	31.0	7.2	250	12	2.1	42	5.0
199	Test Well 3	36 48 30	115 51 26	1	38.0	7.3L	--	51	21	83	7.4
200	Railroad Well Farrier	36 48 49	114 39 14	1	22.8	8.0	1,300	84	31	150	19
201	Sand Spring	36 49 30	115 34 05	1	15.0	8.7L	450	19	44	19	4.6
202	UE-25P-1	36 49 38	116 25 21	1	57.0	6.7	1,200	94	31	150	12
203	UE-25C-2	36 49 45	116 25 43	1	40.5	7.7	300	12	.3	54	2.0
204	UE-25C-3	36 49 47	116 25 43	1	40.8	7.7	300	11	.4	55	1.9
205	USW H-4	36 50 32	116 26 54	1	34.8	7.4	--	17	.3	73	2.6

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Number of samples	Temperature (°C)	pH	Specific conductance (µS/cm)	Calcium	Magnesium	Sodium	Potassium
206	USW H-6	36 50 49	116 28 55	1	37.8	8.1	--	4.1	0.1	86	1.3
207	Tim Spring	36 50 58	115 34 10	1	14.0	8.4	340	19	28	13	2.9
208	UE-25B-1 (Amargosa Well #57)	36 51 08	116 26 23	3	36.4	7.3	--	18	0.7	48	3.3
209	USW G-4	36 51 14	116 27 04	1	35.6	7.7	--	13	0.2	57	2.1
210	USW H-5	36 51 22	116 27 55	2	36.0	7.8	--	1.9	0.01	60	2.1
211	USW H-1 (Amargosa Well #56)	36 51 58	116 27 12	2	33.8	7.6	--	5.3	<.1	51	2.0
212	Dejesus Spring	36 52 28	115 34 45	1	15	7.5	590	41	45	18	3.7
213	VF-2 Well	36 52 30	114 56 44	1	34.0	7.4	800	47	21	81	11
214	VF-1 Well	36 52 32	116 55 44	1	28.0	7.0	460	41	7.5	34	1.2
215	C-1 Well	36 55 07	116 00 34	4	37.0	7.7	--	74	30	123	14
216	UE-29A-2	36 56 29	116 22 26	2	24.0	7.1	--	10	.2	44	1.2
217	Desert Dry Lake Well	36 57 11	115 11 51	1	19.0	8.0	400	22	27	35	5.7
218	Quartz Spring	36 59 10	115 36 00	1	18.0	8.0	800	67	39	54	2.2
219	UE-16F	37 02 09	116 09 25	1	23.3	8.4	1,600	4.7	1.3	425	2.2
220	Tippipah Spring	37 02 36	116 12 26	3	18.0	7.4	340	23	1.7	49	2.3
221	Well U3CN5	37 03 34	116 01 20	1	--	--	590	41	17	54	9.7
222	Willow Spring KSV-1	37 05 34	114 49 52	1	17.4	7.5	--	20	2.7	56	4.6
223	Grapevine Spring KSV-2	37 08 08	114 42 02	1	18.5	7.3	--	75	22	17	2.3
224	Well #8 (NTS) (Amargosa Well #64)	37 09 56	116 17 21	1	26.5	7.4	--	8.4	1.2	31	3.5
225	Jensen Well	37 11 03	114 27 52	1	18.0	7.7	840	55	14	100	7.2
226	Maynard Lake Spring	37 11 30	115 02 02	1	9.6	7.9	910	43	23	114	14
227	Lone Tree Spring	37 12 07	115 03 32	1	11.0	7.5	1,300	45	43	164	20
228	Whiterock Spring	37 12 09	116 07 52	1	--	6.2	120	6.2	.2	17	3.0
229	Kane Spring KSV-3	37 14 46	114 42 21	1	16.4	7.2	--	44	13	20	5.9
230	Boulder Spring KSV-4	37 16 12	114 38 44	1	16.8	7.9	--	21	4.9	12	2.3
231	Randono Well	37 19 26	114 30 08	1	17.2	7.6	760	46	14	100	8.4
232	Cane Spring	37 20 27	115 44 50	1	--	7.6	810	84	48	24	2.4
233	Bradshaw Well	37 20 57	114 32 38	1	14.8	7.3	1,100	85	28	120	11
234	Railroad Well	37 21 04	114 32 02	1	16.0	7.6	730	42	14	98	8.8
235	Sheep Spring	37 24 02	114 16 37	1	10.0	6.8	140	24	5.0	9.8	1.3
236	Cattle Spring	37 24 57	115 45 05	1	--	8.0	530	56	11	41	5.3
237	Rock Spring	37 25 53	115 41 23	1	--	7.8	580	87	19	13	1.7
238	Quail Spring	37 26 29	115 41 01	1	--	7.8	670	86	19	29	1.9
239	Hells Acre Gulch Spring	37 27 37	115 07 29	1	13.0	8.0	380	45	9.0	21	2.4
240	Ash Spring	37 27 49	115 11 34	3	35.3	7.2	460	46	15	29	7.7

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Site name	Latitude (degree, minute, second)	Longitude (degree, minute, second)	Number of samples	Temperature (°C)	pH	Specific conductance (µS/cm)	Calcium	Magnesium	Sodium	Potassium
241	Little Ash	37 27 50	115 11 30	1	36.0	7.6L	500	56	14	31	5.6
242	Indian Spring	37 27 50	115 44 28	1	--	8.1	360	30	7.1	39	.9
243	Rabbitbrush Spring	37 28 56	115 41 28	1	--	7.9	380	42	8.9	25	3.5
244	Savio Spring	37 29 42	115 42 04	1	--	8.3	470	55	14	24	6.2
245	Rose Bud Spring	37 30 32	115 46 25	1	--	7.3	850	89	60	18	2.0
246	Ramone Mathews Well	37 31 36	114 14 39	1	18.5	7.8	270	42	6.3	20	5.9
247	Crystal Spring	37 31 53	115 13 58	2	27.5	7.3	410	44	22	22	5.3
248	Acoma Well	37 32 55	114 10 23	1	17.0	7.7	260	38	5.3	21	7.0
249	Hiko Spring	37 36 34	115 12 51	2	26.0	7.4	--	49	23	26	7.4
250	Walis Health Well	37 37 13	114 30 44	1	63.0	7.8L	440	41	4.4	40	11
251	Caliente Hot Spring	37 37 16	114 30 34	1	45	--	480	37	7.3	49	19
252	Weaver Well	37 44 41	114 25 28	1	17.0	7.7	980	100	42	110	14
253	Well B-1 Tonopah Test Range Nellis B&G Rg	37 45 06	116 29 23	1	26.7	7.9	280	19	.6	34	9.6
254	John Wadsworth Well	37 46 07	114 24 25	1	14.5	7.5	1,100	120	47	150	9.5
255	Bennett Spring	37 47 03	114 31 41	1	24.0	7.5	580	56	26	6.5	1.5
256	Lester Mathews Well	37 47 37	114 23 59	1	20.0	8.1	1,000	73	21	140	10
257	Panaca Town Well	37 47 50	114 23 57	1	29.5	7.9	500	45	13	47	8.3
258	Panaca Spring	37 48 24	114 22 47	2	29.0	7.8	460	33	10	37	7.0
259	North Lee Well	37 49 28	114 23 04	1	22.0	8.0	580	59	12	44	9.9
260	S01 E47 15 CC	37 50 40	116 42 40	1	17.5	7.7	320	14	1.1	49	6.7
261	S01 E 51 21BA	37 50 45	116 20 55	1	15.0	7.4	290	27	3.2	26	6.9
262	Delmues Spring	37 51 36	114 19 20	1	18.0	7.7	560	47	6.7	30	6.3
263	Oxborrow Well	37 53 10	114 18 17	1	11.5	7.9	800	130	22	65	11
264	Flatnose Spring	37 53 46	114 13 33	1	25.0	8.0	400	26	3.5	34	5.6
265	N01 E53 31CD	37 54 03	116 03 51	1	17.0	7.8	270	17	1.8	39	5.0
266	Upper Conners Spring	37 54 10	114 33 38	1	8.0	7.4	490	73	26	2.2	0.5
267	N01 E53 32CAA	37 54 15	116 02 20	1	--	7.8	360	25	1.8	42	7.8
268	Lime Spring	37 54 52	114 32 25	1	21.0	8.3	330	55	31	3.8	0.9
269	N01 E47 30ABB	37 54 55	116 45 08	1	17.0	7.7	400	23	3.5	53	7.7
270	Deadman Spring	37 55 07	114 32 29	1	9.5	7.1	370	98	41	5.0	0.9
271	Highland Spring	37 55 16	114 32 56	1	10.0	7.2	380	86	36	4.7	1.0
272	S02 E47 07AA	37 57 20	116 45 00	1	25.0	8.5	450	1.4	.1	96	5.6
273	Pioche Municipal Well	37 57 33	114 24 51	1	20.5	7.7	300	31	4.4	17	5.5
274	N01 E53 07AD	37 57 50	116 02 60	1	17.0	9.9	4,100	.7	.1	980	9.6
275	N01 E47 02BAB	37 58 18	116 41 04	1	16.0	7.5	350	23	1.5	47	7.2

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Bicarbonate	Carbonate	Chloride	Sulfate	Fluoride	Silica	Sum of constituents[1]	Dissolved oxygen	Deuterium (permil)	Oxygen-18 (permil)	Carbon-13 (permil)	Carbon-14 (pmc)	Tritium (pCi/L)	Date	Source[2]
1	262	--	118	362	0.4	20	920	--	--	--	--	--	--	--	1
2	194	--	114	344	--	16	830	--	--	--	--	--	--	--	1
3	209	--	95	313	--	19	780	--	--	--	--	--	--	--	1
4	160	--	188	195	--	28	740	--	--	--	--	--	--	--	1
5	194	--	157	341	--	17	910	--	--	--	--	--	--	--	1
6	183	--	94	288	.3	16	720	--	--	--	--	--	--	--	1
7	184	--	142	320	--	15	840	--	--	--	--	--	--	--	1
8	168 L	--	66	235	.6	55	610	--	-73.0	-9.4	--	--	--	06-22-85	2
9	300 L	--	76	208	.6	18	650	--	-88.0	-11.9	-9.9	--	--	06-22-85	2
10	368 L	--	30	26	.5	80	450	--	-72.0	-8.4	--	--	--	06-22-85	2
11	410	--	730	1,000	2.6	41	3,100	--	-90.5	-10.8	--	--	--	04-23-82	1
12	169 L	--	112	53	1.3	29	410	--	-67.0	-7.2	-7	68.1	--	06-21-85	2
13	230	--	55	550	1.5	34	1,100	--	-95	-12.1	-7.6	2.4	--	06-21-85	2
14	242	--	7.6	109	.4	25	370	--	-94.0	-13.2	--	--	--	09-27-86	1
15	315 L	--	126	263	.3	18	820	--	-94	-12.5	-9.3	39.4	--	06-21-85	2
16	730	--	460	500	3.1	91	2,300	2.0	-98.0	-12.85	-4.3	--	<1.0	06-30-85	1
17	168 L	--	51	98	.2	15	360	--	-88.0	-11.7	-7.8	67.5	--	06-23-85	2
18	366 L	--	59	101	.5	29	530	--	-87.5	-10.9	-9.4	60.4	--	06-22-85	2
19	89	--	82	141	1.0	24	450	--	-80.5	-10.2	--	--	--	01-00-81	3
20	38	--	980	575	4.2	40	2,500	--	-83.5	-11.2	--	--	--	03-00-82	3
21	154	--	198	177	.9	31	720	--	-95	-13.1	-6.8	--	--	02-28-86	4
22	141	--	1,530	1,030	4.2	45	4,100	--	-82.0	-10.5	--	--	--	03-00-82	3
23	210	--	88	310	.4	17	720	3.8	-92.0	-12.3	-7.4	4	--	05-10-83	1
24	152 L	--	95	292	.6	39	730	--	-97	-13.3	--	--	--	02-27-86	4
25	249	--	7.4	171	--	9.7	450	--	-86	-12.3	-9.7	59	--	07-11-86	4
26	150	--	97	250	0.7	53	680	3.6	-100.0	-12.9	-6.4	1	--	05-11-83	1
27	260	--	10	42	.2	10	290	6.0	-89.0	-12.2	-10.6(2)	49.8(2)	<15 (1)	06-25-85	1
28	210	--	110	450	.8	26	690	3.4	-94.0	-12.7	-6.7	4	--	05-09-83	1
29	192	--	12	548	.3	14	930	--	-89	-12.2	--	--	--	05-18-79	4
30	192	--	13	518	.3	15	890	--	-86	-12.4	-7.9	11.0	<15	03-13-86	4
31	215	--	21	176	.3	15	320	--	-95	-13.2	-7.9	--	--	05-10-79	4
32	223	--	8.0	80	.2	16	310	--	-98	-13.5	-8.0	--	--	05-08-79	4
33	228 L	--	4.8	59	.2	15	280	--	-99	-13.6	-9.3	39.4	--	03-21-79	4
34	210	--	7.3	170	.2	13	430	6.0	-89.0	-12.25	-9.3	46.0	9.0	06-30-85	1
35	223	--	7.8	73	.2	16	300	--	-100	-13.7	-7.6	--	--	05-15-79	4

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Bicarbonate	Carbonate	Chloride	Sulfate	Fluoride	Silica	Sum of constituents[1]	Dissolved oxygen	Deuterium (permil)	Oxygen-18 (permil)	Carbon-13 (permil)	Carbon-14 (pmc)	Tritium (pCi/L)	Date	Source[2]
36	206	--	7.0	244	0.2	13	520	--	-90	-12.4	-8.5	--	--	03-28-79	4
37	260	--	6.0	21	.2	12	260	6.4	-89.0	-12.25	-10.5(2)	62.4(2)	3.0	06-26-85	1
38	230	--	2.4	23	.1	13	230	8.8	-99.0	-13.55	-8.3	46.9	<2.0	06-27-85	1
39	200	--	4.5	49	.5	35	260	7.1	-101.0	-13.6	-7.9	9	--	05-11-83	1
40	230	--	5.8	54	.2	18	280	--	-100	-13.5	-7.2	--	--	03-29-79	4
41	223	--	6.5	116	.2	13	340	--	-94	-13.2	-8.3	--	--	06-26-79	4
42	226	--	4.8	65	.2	14	280	--	-97	-13.4	-7.3	14	--	06-13-86	4
43	207	--	3.5	33	.2	15	230	--	-100	-13.8	-7.7	--	--	06-28-79	4
44	227	--	7	67	.2	14	290	--	-96	-13.2	-7.7	--	--	06-26-79	4
45	330	--	3.3	25	.2	11	320	7.7	-96.5	-13.25	-10.1	84.1	34	06-28-85	1
46	229	--	5.0	72	--	14	290	--	-97	-13.2	-7.8	--	--	05-06-86	4
47	200	--	16	180	.2	13	450	5.4	-91.0	-12.5	-12.0	--	<2.0	06-26-85	1
48	235	--	3.7	48	.2	14	250	--	-97	-13.8	-7.2	--	--	05-14-86	4
49	230	--	4.5	43	.2	15	260	--	-99	-13.6	-7.3	--	--	03-27-79	4
50	236	--	4.5	50	.2	13	270	--	-97	-13.6	-8.0	--	--	06-26-79	4
51	238	--	5	52	.2	13	280	--	-97	-13.5	-8.0	--	--	06-26-79	4
52	229	--	4.8	73	.2	14	270	--	-94	-13.4	-8.3	--	--	03-27-79	4
53	239	--	4.0	36	.2	14	530	--	-101	-14.0	-7.3	--	--	05-13-86	4
54	233	--	3.5	48	.2	14	260	--	-100	-14	-8.2	--	--	03-27-79	2
55	234	--	3.1	33	.2	14	240	--	-100	-14.1	-7.0	--	--	06-03-86	4
56	240	--	3.4	28	.2	14	240	6.9	-97.0	-13.55	-7.6	26.0	<1.0	06-27-85	1
57	231	--	3.5	31	0.2	14	240	6.0	-104.0	-13.5	-6.8	13	--	--	1,4
58	240	--	4.5	35	.2	13	250	--	--	--	--	--	--	01-09-76	1
59	256	--	8.0	68	.4	80	390	--	-98	-13.8	-8.0	--	--	05-15-79	4
60	178	--	.7	6.0	.2	4	160	9.4	-97.7(19)	-13.6(22)	-8.1(5)	90.8(1)	257 (3)	--	5
61	2,350	720	3,800	2,800	15	8.7	72,000	0.8	-93.5	-11.1	--	--	--	11-21-83	1
62	2,450	590	6,700	3,400	12	20	8,000	1.5	-91.0	-10.0	--	--	--	11-21-83	1
63	231	--	2.8	29	.2	15	240	--	-102	-14.0	-7.2	--	--	03-06-86	4
64	192	--	41	150	2.1	55	490	3.4	-101.5	-13.8	-5.3	5.6	<.3	09-29-86	1
65	150	--	.7	3.7	.1	3.2	130	--	--	--	--	--	--	08-25-78	6
66	197	--	3.5	27	.3	34	230	--	-106	-14.5	-8.2	3	--	05-11-79	4
67	235	--	4.1	5.7	--	6.0	210	--	-96.0	-12.4	-9.8	91.7	98	08-19-82	2
68	242	8.3	1.9	6.7	.2	5.1	220	--	-95.5	-13.4(12)	--	--	--	04-05-88	1
69	150	--	.8	4.6	.1	3.8	130	--	--	--	--	--	--	08-24-78	6
70	15,980	10,200	23,000	19,000	30	28	9,000	6.5	-66.5	-2.5	--	--	--	11-18-83	1

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Bicar-bonate	Car-bonate	Chloride	Sulfate	Fluor-ide	Silica	Sum of consti-tuents[1]	Dis-solved oxygen	Deuterium (permil)	Oxygen-18 (permil)	Carbon-13 (permil)	Carbon-14 (pmc)	Tritium (pCi/L)	Date	Source[2]
71	234	--	20	78	.5	41 0	360	--	-95	-13.2	-6.3	21	--	04-03-86	4
72	600	--	480	550	3.4	67	2,200	5.8	-92.0	-11.3	--	--	--	11-17-83	1
73	210	--	3.7	5.9	.2	4.6	180	--	--	--	--	--	--	08-24-78	6
74	180	--	.7	5.1	.1	4.1	160	--	--	--	--	--	--	08-24-78	6
75	185	--	.7	3.8	.1	4.8	160	--	--	--	--	--	--	--	6,7
76	235	--	2.4	6.2	.1	5.6	210	--	--	--	--	--	--	--	6,7
77	252	--	6.3	34	.2	13	260	--	-92.0	-12.9	--	--	<15	03-31-87	2
78	225	--	.8	6.4	.1	5.6	200	--	--	--	--	--	--	--	6,7
79	232	--	1.2	6.5	.2	6.3	200	--	-94.5	-12.8	-7.4	--	.3	04-16-80	2,7
80	244	--	1.4	9.9	.1	6.5	220	--	--	--	--	--	--	--	7
81	257	--	1.3	26	.2	6.1	240	--	-96.0	-12.2	-10.1	83	87	--	2,6,7
82	--	--	2.5	14	--	7.1	--	--	--	--	--	--	--	08-18-78	6
83	340	--	2.9	19	.1	7.7	310	--	--	--	--	--	--	08-23-78	6
84	224	--	3.3	25	.2	17	230	--	-101	-13.7	-7.1	--	--	11-03-86	4
85	213 L	--	4.6	16	.2	20	210	--	-100.0	-13.4	-6.7	1.9	--	06-30-85	2
86	720	25	690	280	6.1	35	2,200	0.2	-100.0	-12.9	--	--	--	11-19-83	1
87	170	--	.3	3.4	.2	3.1	140	--	--	--	--	--	--	07-24-80	1
88	200	--	.6	3.4	.2	3.9	170	--	--	--	--	--	--	--	1,6
89	292	--	16	53	.2	22	340	--	-91.5	-12.4	-9.1	46	<15	--	2
90	207	--	1.4	9.9	.1	4.9	180	--	--	--	--	--	--	08-23-78	6
91	260	--	.9	11	.1	6.1	230	--	--	--	--	--	--	--	7
92	181 L	--	.5	2.8	.1	5.0	160	--	-98.5(2)	-14.1(2)	-8.4(2)	--	78.4(2)	06-28-85	2
93	238 L	--	.9	4.3	.1	5.1	200	8.3	-98.0	-13.45	-9.6	100	73.2	06-28-85	2
94	245	--	.9	3.1	.1	5.2	210	--	--	--	--	--	--	--	2
95	240	--	4.1	19	.3	18	230	5.6	-99.0	-13.3	-6.7	7	--	05-10-83	1
96	244	--	1.8	30	--	7.0	240	--	-98	-12.7	-8.8	61.4	--	08-21-85	2
97	277	--	1.5	21	.8	7.9	240	--	--	--	--	--	--	--	6,7
98	244	--	1.6	20	.01	8	230	2.0	-100	-13.4	-8.0	31.6	<15	06-24-86	2
99	251 L	--	4.7	18	.2	19	240	--	-99.7(3)	-12.5(2)	-7.0	13.9	<15	06-30-85	2
100	440	--	1.5	5.8	.1	7.2	360	--	-93.5	-12.9	--	--	--	06-29-85	1
101	226	--	9.2	22	.3	17	240	--	-97.0	-13.2	-7.2	23.8	<15	06-25-86	2
102	223	--	1.5	39	--	11	240	--	-100	-12.6	-7.7	16.4	--	08-20-85	2
103	257	--	3.4	59	.2	9.0	290	--	-98.0	-13.4	-8.6	49.6	<15	06-24-85	2
104	247	--	3.7	18	--	18	220	--	-98.0	-12.6	-6.2	11.3	--	08-20-82	2
105	278	--	1.3	11	.7	6.5	220	--	-99.5	-13.8	-10.8	91.1	89	--	1,6,7

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Bicar-bonate	Car-bonate	Chloride	Sulfate	Fluor-ide	Silica	Sum of constituents[1]	Dis-solved oxygen	Deuterium (permil)	Oxygen-18 (permil)	Carbon-13 (permil)	Carbon-14 (pmc)	Tritium (pCi/L)	Date	Source[2]
106	226	--	200	380	1.4	23	1,000	5.5	-98.0	-13.45	-5.5	2.7	<.3	09-30-86	1
107	255 L	--	3.5	67	--	10	300	--	-96.0	-12.1	-7.9	48.4	--	08-21-82	2
108	255	--	3.3	64	--	10	300	--	-98.0	-12.7	-8.2	50.1	--	08-20-82	2
109	243	--	2.5	43	.7	9.7	260	--	-98.0	-14.0	-8.2	57	32	08-21-82	2
110	318	--	27	111	1.3	28	490	4.5	-103.0(4)	-13.4(1)	-4.6(1)	3.2(3)	.5(2)	--	5
111	163	--	333	1,633	1.4	18	2,900	2.3	-92.0	-12.2	-4.05	1.6	--	--	1
112	230	--	190	360	1.6	21	1,000	.5	-96.0	-13.7	--	--	--	04-26-82	1
113	160	--	500	2,300	1.4	18	4,000	3.4	-93.0(2)	-12.4(2)	-5.3	7.2	--	07-01-85	1
114	226	--	180	370	1.6	23	1,000	4.8	-97.0	-13.05	-4.9	1.5	<1.0	03-31-86	1
115	292	--	1.1	9.0	.2	7	250	8.2	-100.1(16)	-13.8(18)	-9.6(5)	76.0(4)	92 (4)	--	5
116	350	--	20	74	3.7	59	500	--	--	--	--	--	--	11-18-72	1
117	289 L	--	1.8	9.9	.2	7.6	250	--	-97.5(3)	-13.75(3)	-9.6	79.2	57	06-26-85	2
118	242	--	5.5	15	.5	46	260	6.1	-93.0	-12.8	-6.7	13.7	1.0	09-25-86	1
119	312	--	26	89	1.7	26	450	3.7	-104.0(5)	-13.7(1)	-5.0(1)	11.1(2)	1.6(2)	--	5
120	320	--	5.9	16	.4	35	310	--	-93.0	-12.7	-6.9	35.7	<15	06-24-86	2
121	164 L	--	21	449	.2	8.3	780	--	-82.0	-10.6	-8.5	18.7	--	06-24-85	2
122	294	--	6.5	17	.4	30	290	--	-94.0	-13.0	-8.8	20.4	1.0	09-25-86	1
123	160	--	6.5	40	2.3	82	280	--	-102.0	-12.4	--	--	--	03-06-74	8
124	284 L	--	7.1	18	.2	18	300	--	-93.2(3)	-12.9(3)	-7.7(3)	13.9(1)	1.0(2)	--	1,5
125	350	--	41	190	6.9	31	660	--	-102.0	-13.7	--	--	--	04-22-82	1
126	243 L	--	2.7	11	.1	8.7	220	8.2	-93.0	-13.3	-8.8	87.3	46	06-27-85	2
127	210	--	170	360	2.1	21	960	2.0	-97.5	-13.3	-4.2	3.0	7.0	07-01-85	1
128	270	--	12	130	.2	21	430	6.4	-92.5	-12.75	-7.2	--	10	06-28-85	1
129	340	--	37	160	7.1	31	610	--	-102.0	-13.7	--	--	--	04-22-82	1
130	195	--	4.6	12	.4	24	200	--	-95.5	-13.55	-7.7	7.0	<15	12-02-86	2
131	197	--	3.3	13	.2	15	190	8.7	-99	-13.7	-8.0	--	--	12-18-87	2
132	120	--	10	35	1.4	73	240	--	--	--	--	--	--	03-06-74	8
133	120	--	6.4	27	1.4	81	240	--	-102.0	-13.0	--	--	--	03-06-74	8
134	400	--	78	180	6.8	46	770	--	--	--	--	--	--	03-03-74	8
135	306	--	21	80	1.7	22	420	3.2	-103.2(4)	-13.6(1)	-5.1(2)	1.9(3)	.5(3)	--	5
136	210	--	28	160	2.9	53	510	--	--	--	--	--	--	03-07-74	8
137	150	--	9.5	31	1.7	80	280	--	-104.0	-12.7	--	--	--	03-01-74	8
138	130	--	10	69	1.2	70	300	--	--	--	--	--	--	03-01-74	8
139	140	--	6.3	25	1.7	79	250	--	-104.0	-13.0	--	--	--	03-05-74	8
140	350	--	38	180	6.4	25	630	--	-104.0	-13.6	--	--	--	04-22-82	1

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Bicar-bonate	Car-bonate	Chloride	Sulfate	Fluor-ide	Silica	Sum of consti-tuents[1]	Dis-solved oxygen	Deuterium (permil)	Oxygen-18 (permil)	Carbon-13 (permil)	Carbon-14 (pmc)	Tritium (pCi/L)	Date	Source[2]
141	202	--	3.4	11	.6	26	200	9.3	-98.0	-13.7	-6.9	--	--	12-19-87	2
142	219 L	--	3.1	11	.1	13	200	7.0	-102.0	-13.7	-6.6	4.9	--	06-26-85	2
143	314	--	27	168	--	38	600	--	-104.0	-13.7	--	--	--	06-24-79	8
144	130	--	6.9	25	1.6	79	240	--	-98.5	-12.6	--	--	--	03-04-74	8
145	320	--	29	170	4.4	33	570	--	-104.0	-13.7	--	--	--	05-11-81	1
146	195	--	5.3	13	0.5	26	200	9.2	-95.0	-13.2	-7.2	--	--	12-19-87	2
147	203	--	4.2	12	.1	13	200	7.2	-98.0	-13.2	-7.1	6.1	<15	06-23-86	2
148	197	--	6.7	32	.5	22	230	--	-87.0	-12.1	-6.2	9.2	--	08-05-87	1
149	291	--	20	70	1.7	22	400	--	--	-13.6(1)	-4.6(1)	1.8(1)	--	--	5
150	130	--	9.3	28	1.3	72	240	--	-102.0	-13.0	--	--	--	03-05-74	8
151	127	--	8.9	35	--	76	260	--	-99.0	-13.2	--	--	--	06-25-79	8
152	130	--	6.6	33	1.0	75	250	--	-101.0	-13.1	--	--	--	03-05-74	8
153	245	--	64	90	1.2	27	470	3.8	-90.5	-11.2	--	--	<1.0	03-28-86	1
154	150	--	7.8	30	1.5	79	270	--	--	--	--	--	--	11-17-72	8
155	132	--	8.2	51	--	77	280	--	-97.5	-13.2	--	--	--	06-26-79	8
156	150	--	7.4	28	1.2	59	240	--	--	--	--	--	--	03-01-74	8
157	236	--	63	187	--	77	670	--	-104.0	-13.6	--	--	--	06-25-79	8
158	242	--	3.8	16	.2	12	220	4.8	-99.3	-13.3	-7.6(3)	8.3(3)	<2.0(2)	--	1,5
159	153	--	7.4	29	--	68	260	--	-103.0	-13.4	--	--	--	03-31-71	8
160	195	--	23	130	--	46	430	--	-105.0	-13.8	--	--	--	03-31-71	8
161	140	--	10	67	.9	72	310	--	-103.0	-13.4	--	--	--	03-01-74	8
162	166	--	8.2	33	--	64	270	--	-102.0	-13.4	--	--	--	03-31-71	8
163	291	--	29	152	--	29	540	--	-105.0	-13.8	--	--	--	04-01-71	8
164	222	--	76	398	.3	18	840	3.2	-96.0	-13.0	-6.9	7.3	<15	06-27-85	2
165	239	--	62	180	--	64	630	--	-102.0	-13.1	--	--	--	03-31-71	8
166	150	--	7.7	30	.7	54	240	--	-99.5	-13.2	--	--	--	11-17-72	8
167	290	--	29	19	.2	16	310	5.9	-92.0	-12.6	--	--	--	--	1
168	140	--	6.0	26	1.0	62	230	--	-103.0	-13.2	--	--	--	03-04-74	8
169	275	--	16	54	.9	19	340	2.8	-103.5(2)	--	-5.6(2)	2.8(1)	.6(2)	--	5,14
170	197 L	--	6.0	14	.2	14	200	--	--	--	--	--	--	--	5
171	150	--	7.4	38	.9	20	220	--	--	--	--	--	--	--	1
172	150	--	6.7	33	1.0	49	240	--	--	--	--	--	--	03-04-74	8
173	150	--	8.5	33	0.9	49	240	--	-102.0	-12.8	--	--	--	11-20-72	8
174	370	--	3.1	7.0	.1	12	310	5.4	-94.3(9)	-12.8(9)	-10.2	96.8	89.6	--	1
175	--	--	15	370	1.0	29	--	--	-87.0	-11.65	--	--	--	01-25-85	1

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Bicarbonate	Carbonate	Chloride	Sulfate	Fluoride	Silica	Sum of constituents[1]	Dissolved oxygen	Deuterium (permil)	Oxygen-18 (permil)	Carbon-13 (permil)	Carbon-14 (pmc)	Tritium (pCi/L)	Date	Source[2]
176	293	--	4.9	8.4	0.2	24	420	--	-81.0	-10.6	--	--	--	03-20-87	1
177	395	--	18	13	.1	16	370	5.1	-91.8	-12.7	-9.9	--	--	--	1
178	250	--	55	149	--	15	660	--	-99.0	-13.1	--	--	--	--	9
179	270	--	294	1,094	--	18	2,200	--	-95.5	-13.2	--	--	--	--	9
180	124	--	184	1,954	--	13	3,100	--	-92.5	-12.7	--	--	--	--	9
181	162	--	233	1,780	--	26	3,000	--	-96.5	-13.7	--	--	--	--	9
182	178	--	41	1,800	--	13	2,700	--	-99.5	-13.9	--	--	--	--	9
183	299	--	26	54	1.2	24	380	--	--	--	--	--	--	10-07-87	1
184	123	--	194	2,100	--	15	3,300	--	-91.0	-12.7	--	--	--	--	9
185	475	--	335	1,240	--	65	2,700	--	--	--	--	--	--	01-16-81	2
186	245	--	57	171	--	28	560	--	-98.0	-13.0	--	--	--	--	9
187	275	--	10	8.5	.2	46	290	5.7	-83.5	-9.8	-8.3	--	--	--	1
188	321	--	121	560	--	66	1,300	--	--	--	--	--	--	07-09-81	1
189	280	--	59	190	2.1	25	640	--	-97.0	-12.75	--	--	--	03-00-70	1
190	270	--	61	190	2.1	29	610	3.0	-97.8(3)	-12.9(3)	-6.0	6.7	<1.0	07-22-81	1
191	255	--	54	154	--	30	550	--	-101.5	-13.2	--	--	--	--	9
192	250	--	15	82	3.2	38	390	--	--	--	--	--	--	--	1
193	119	--	7.3	22	2.1	54	210	--	-97.5	-12.8	--	--	--	03-26-71	10
194	271	--	53	159	2.1	30	560	3.7	-97.0	-12.95	-8.0	8.4	1.8	01-26-86	1
195	276	--	61	160	2.3	30	590	4.0	-98.0	-12.85	-5.5	8.4	4.0	01-26-86	1
196	165	--	10	44	2.7	50	280	--	-108.0(1)	-14.2(1)	-8.5	12.2	--	--	1,8
197	295	--	34	105	1.9	33	480	2.9	-101.0	-13.0	--	7.6	<2.0(1)	--	1
198	124	--	7.1	17	2.4	57	210	--	-97.5	-13.0	--	--	--	03-26-71	10
199	328 L	--	23	84	1.5	24	480	--	--	--	--	--	--	05-10-62	5
200	64	--	52	550	1.6	23	940	--	-97.5	-12.5	--	--	--	02-04-84	1
201	220	--	16	24	.2	--	240	--	-88.0	-11.9	-5.3	--	--	01-03-88	2
202	890	--	26	78	4.9	44	870	--	-107.0	-13.8	--	--	--	02-11-83	11
203	140	--	7.0	22	2.1	54	220	--	-101.0	-13.4	-7.0	15.3	--	03-13-84	1
204	170	--	7.2	22	2.0	53	230	--	-103.0	-13.5	--	--	--	05-19-84	1
205	173	--	6.9	26	4.8	46	260	--	-104.0	-14.0	--	--	--	05-17-82	12
206	182	--	7.6	29	4.7	48	270	--	-106.0	-13.8	--	--	--	10-16-82	12
207	143	--	11	37	.2	--	180	--	-99.0	-13.2	-8.9	--	--	01-02-88	2
208	143	--	9.7	22	1.6	52	230	--	-100.0	-13.4	--	--	--	--	10,12
209	139	--	5.9	19	2.5	45	220	--	-103.0	-13.8	--	--	--	12-09-82	12
210	126	--	6.1	16	1.4	48	200	--	-102.0	-13.6	-10.3	19.8	--	--	12

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Bicarbonate	Carbonate	Chloride	Sulfate	Fluoride	Silica	Sum of constituents[1]	Dissolved oxygen	Deuterium (permil)	Oxygen-18 (permil)	Carbon-13 (permil)	Carbon-14 (pmc)	Tritium (pCi/L)	Date	Source[2]
211	118	--	5.7	18	1.1	44	190	--	-102.0	-13.4	-11.4	21.9	--	--	10,12
212	287	--	14	64	.5	16	340	--	-100	-13.1	-12.0	--	51.2	05-09-87	2
213	303	--	34	90	1.7	34	470	2.9	-101.0(2)	-13.0(2)	-6.1	7.0	<1.0	02-05-86	1
214	156	--	42	20	.5	14	230	--	-94.0	-12.6	--	--	--	01-06-88	1
215	584	--	32	67	1.1	29	660	--	-110.5	--	-3.8(1)	.8(1)	<.3(1)	--	5,14
216	107	--	9.9	21	.9	44	180	--	-93.2	-12.8	-12.9	61	--	--	12
217	207	--	8.9	48	.6	49	300	--	-98.0	-13.1	-5.3	1.3	<.6	03-18-87	1
218	383	--	31	80	.6	52	520	11.0	-88.0	-11.6	-11.3	--	--	01-02-88	2
219	930	33	26	120	4.4	4.7	1,100	--	-105.0	-14.9	--	--	--	09-25-77	13
220	160	--	8.9	18	.3	53	240	--	--	--	--	--	58	--	1
221	270	--	32	41	0.8	38	370	--	--	--	--	--	--	11-15-74	1
222	140 L	--	22	34	1.1	65	270	--	-88.0	-11.6	--	--	--	02-03-84	1
223	280 L	--	27	40	.9	22	340	--	-87.5	-12.0	--	--	--	02-03-84	1
224	80	--	7.4	14	--	41	150	--	-104.0	-13.0	--	--	--	03-24-71	8
225	340	--	45	80	2.1	56	520	--	-88.5	-11.6	--	--	--	04-10-85	1
226	405 L	--	30	88	--	--	510	--	-94.0	-12.3	--	--	--	01-14-85	2
227	539 L	--	50	145	--	--	730	--	-89.5	-10.9	--	--	--	01-14-85	2
228	41	--	3.7	12	.1	24	86	--	--	--	--	--	--	03-14-73	1
229	210 L	--	17	14	2.8	60	280	--	-86.5	-11.9	--	--	--	02-02-84	1
230	100 L	--	7.8	6	1.7	41	140	--	-87.0	-12.6	--	--	--	02-02-84	1
231	350	--	44	63	2.3	54	500	--	-87.5	-11.7	--	--	--	02-03-84	1
232	360	--	18	136	--	--	490	--	-88	-9.5	--	--	--	05-15-85	2
233	550	--	52	76	2.3	63	710	--	-88.5	-11.4	--	--	--	02-01-84	1
234	300	--	42	60	2.3	51	460	--	-86.0	-11.6	--	--	--	01-31-84	1
235	96	--	7.9	7.0	0.7	33	140	--	-87.0	-12.0	--	--	--	06-03-85	1
236	261 L	--	16	39	--	--	300	--	-85	-9.4	--	--	--	05-15-85	2
237	348 L	--	5.8	35	--	--	330	--	-86.0	-10.9	--	--	--	05-15-85	2
238	166 L	--	11	206	--	--	430	--	-92.0	-11.9	--	--	--	05-15-85	2
239	198 L	--	8.2	20	--	--	200	--	-93.0	-12.3	--	--	--	01-14-85	2
240	259	--	8.0	33	0.9	30	290	2.3	-108.0	-14.1	-6.7(2)	6.3(3)	0 (2)	--	1,5
241	270 L	--	21	34	8.7	29	320	--	--	--	--	--	--	02-04-74	1
242	173 L	--	9.4	31	--	--	200	--	-88	-10.4	--	--	--	05-15-85	2
243	197	--	9.0	27	--	--	210	--	-89	-11.2	--	--	--	05-14-85	2
244	268 L	--	10	26	--	--	270	--	--	--	--	--	--	05-14-85	2
245	400 L	--	11	150	--	--	530	--	--	--	--	--	--	05-17-85	2

APPENDIX B.—*Average chemical and isotopic compositions of water from wells and springs in southern Nevada and southeastern California, between latitudes 35 and 38 degrees north, and longitudes 114 and 117 degrees east*—Continued

Site	Bicarbonate	Carbonate	Chloride	Sulfate	Fluoride	Silica	Sum of constituents[1]	Dissolved oxygen	Deuterium (permil)	Oxygen-18 (permil)	Carbon-13 (permil)	Carbon-14 (pmc)	Tritium (pCi/L)	Date	Source[2]
246	170	--	15	12	0.3	61	250	--	-92.0	-12.3	--	--	--	06-03-85	1
247	258	--	8.7	33	0.3	24	290	1.9	-109.0	-14.3	-7.0(3)	7.8	<2.0(1)	--	1,5
248	150	--	17	10	0.3	54	230	--	-95.0	-12.6	--	--	--	06-03-85	1
249	282	--	11	37	0.6	30	320	3.0	-109.6(4)	-14.65(2)	-6.7	6.5	0 (1)	--	2,5,14
250	210 L	--	7.0	33	1.6	88	330	--	--	--	--	--	--	02-04-74	1
251	222	--	13	34	1.4	130	400	--	-109.0	-14.5	--	--	--	04-10-85	1
252	430	--	110	180	2.9	73	840	--	-101.0	-13.1	--	--	--	06-04-85	1
253	100	--	15	26	.6	91	240	--	-110.0	-14.0	--	--	--	09-10-80	1
254	600	--	88	200	6.5	76	990	--	-101.0	-12.9	--	--	--	06-04-85	1
255	320	--	7.9	6.9	.1	14	280	--	-103.0	-13.7	--	--	--	04-10-85	1
256	430	--	44	170	3.1	64	730	--	-103.0	-13.3	--	--	--	06-04-85	1
257	203	--	19	68	1.8	58	360	--	-106.0	-14.0	--	--	--	06-04-85	1
258	185	--	16	27	1.6	48	270	6.2	-106.8(3)	-14.0(3)	-7.7(1)	--	--	--	1
259	220	--	48	33	1.0	54	370	--	-101.0	-13.3	--	--	--	06-04-85	1
260	140	--	13	26	.8	68	250	--	--	--	--	--	--	07-07-67	1
261	170	--	8.5	--	.4	--	160	--	--	--	--	--	--	09-14-68	1
262	180	--	24	18	.6	64	280	--	-104.0	-13.4	--	--	--	04-08-85	1
263	350	--	140	63	.8	58	660	--	-92.0	-11.8	--	--	--	06-05-85	1
264	150	--	10	18	1.3	55	230	--	-101.0	-13.4	--	--	--	04-08-85	1
265	150	--	7.2	7.0	1.4	55	210	--	--	--	--	--	--	09-14-68	1
266	351	--	2.1	5.4	<0.1	8.5	290	--	-100.0	-13.9	--	--	--	11-11-86	1
267	120	--	15	49	.5	72	280	--	--	--	--	--	--	08-03-67	1
268	290	--	4.1	8.9	.1	14	260	--	-97.0	-12.9	--	--	--	04-07-85	1
269	160	--	18	36	1.1	72	290	--	--	--	--	--	--	07-28-67	1
270	510	--	4.2	8.3	.1	19	430	--	-99.0	-13.3	--	--	--	04-07-85	1
271	470	--	4.4	8.1	.1	15	390	--	-98.3(2)	-13.3(2)	--	--	--	04-07-85	1
272	180	4.0	17	38	1.7	71	320	--	--	--	--	--	--	07-07-67	1
273	114	--	19	13	.3	64	210	--	-108.5	-14.4	--	--	--	11-10-86	1
274	370	490	340	460	8.6	93	2,600	--	--	--	--	--	--	09-13-68	1
275	130	--	17	32	1.0	74	270	--	--	--	--	--	--	07-28-67	1

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[1] Calculated by muitiplying HCO_3 by 0.4916 to make comparable to a residue upon evaporation value.

[2] Sources of data:

1 U.S. Geological Survey (unpublished).
2 Desert Research Institute (unpublished).
3 McKay and Zimmerman (1983).
4 Noack (1988).
5 Winograd and Pearson (1976).
6 Nichols and Davis (1979).
7 Plume (1985)
8 Claassen (1985).
9 Schroth (1987).
10 Claassen (1973).
11 Craig and Robison (1984).
12 Benson and others (1983).
13 Dinwiddie and Weir (1979).
14 Winograd and Friedman (1972).