

Chemical composition of ground water, hydrologic properties of basin-fill material, and ground-water movement in Salt Lake Valley, Utah



This report was prepared as a part of the Statewide cooperative water-resource investigation program administered jointly by the Utah Department of Natural Resources, Division of Water Rights and the U.S. Geological Survey. The program is conducted to meet the water administration and water-resource data needs of the State as well as the water information needs of many units of government and the general public

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Cover: Top photograph, 1890 view of Salt Lake Valley, Utah, looking south from Ensign Peak, courtesy of Photograph Archives, Utah State Historical Society; Bottom photograph, 1995 view of Salt Lake Valley, Utah, looking south from Ensign Peak.



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DEPARTMENT OF NATURAL RESOURCES

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**CHEMICAL COMPOSITION OF GROUND WATER,
HYDROLOGIC PROPERTIES OF BASIN-FILL
MATERIAL, AND GROUND-WATER MOVEMENT
IN SALT LAKE VALLEY, UTAH**

By Susan A. Thiros
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CONTENTS

Abstract	1
Introduction	1
Purpose and scope	3
Acknowledgments	3
Concept of ground-water system	3
Methods and limitations	5
Chemical composition	5
Hydrologic properties	7
Isotopes	9
Chemical composition of ground water	10
Water in the principal aquifer	10
Water from intermediate depths	24
Water in the shallow unconfined aquifer and confining layers	26
Historical changes in chemical composition of ground water	30
Hydrologic properties of the basin-fill material	32
Slug tests	32
Multiple-well aquifer tests	33
Aquifer test at well (C-2-1)23dad-1	33
Aquifer test at well (D-2-1)l6bba-2	37
Aquifer test at well (D-2-1)30dda-1	38
Aquifer test at well (D-2-1)34dba-1	38
Water-level fluctuations and ground-water movement	38
Water-level fluctuations in the shallow unconfined aquifer	38
Ground-water movement indicated by geochemical modeling	39
Isotopic indicators of ground-water movement	44
Oxygen-18 and deuterium in surface water	44
Oxygen-18 and deuterium in ground water	48
Tritium	52
Vertical ground-water movement indicated by hydrologic properties	55
Summary	55
References cited	57

ILLUSTRATIONS

PLATES

[Plates are in pocket]

1. Map showing location of hydrologic-data sites where isotopic data are available; tritium, $\delta^{18}\text{O}$, and δD values for water samples from selected hydrologic-data sites; and location of wells where a slug test or aquifer test was done; Salt Lake Valley, Utah, 1990–92
2. Map showing dissolved-solids concentration in water from the principal aquifer in Salt Lake Valley, Utah, 1988–92

FIGURES

1. Map showing location of Salt Lake Valley study area, Utah	2
2. Diagram showing numbering system used for hydrologic-data sites in Utah	4
3. Generalized block diagram showing the basin-fill ground-water flow system in Salt Lake Valley, Utah	6
4. Map showing location of wells greater than 300 feet deep in Salt Lake Valley, Utah, sampled for chemical analysis of water, 1983–92	18
5. Trilinear diagram showing chemical composition of water from wells greater than 300 feet deep in Salt Lake Valley, Utah	19
6. Map showing location of surface-water sampling sites in Salt Lake Valley, Utah	20
7. Trilinear diagram showing chemical composition of surface water in Salt Lake Valley, Utah	21
8. Graph showing relation between log calcium and log carbonate activity for water sampled from wells greater than 300 feet deep in Salt Lake Valley, Utah	23
9. Map showing location of wells 101 to 300 feet deep in Salt Lake Valley, Utah, sampled for chemical analysis of water, 1982–92	25
10. Trilinear diagram showing chemical composition of water from wells 101 to 300 feet deep in Salt Lake Valley, Utah	26
11. Graph showing relation between log calcium and log carbonate activity for water sampled from wells 101 to 300 feet deep in Salt Lake Valley, Utah	27
12. Map showing dissolved-solids concentration in water from wells less than 101 feet deep in Salt Lake Valley, Utah, 1983–91	28
13. Trilinear diagram showing chemical composition of water sampled from wells less than 101 feet deep in Salt Lake Valley, Utah	29
14. Hydrographs showing seasonal water-level fluctuations in 12 monitoring wells completed in the shallow unconfined aquifer or a confining layer in Salt Lake Valley, Utah, 1990–92	40
15. Hydrograph showing reverse water-level fluctuations in monitoring well (C-1-1)23bcd-2 in Salt Lake Valley, Utah	43
16. Graph showing relation between δD and $\delta^{18}\text{O}$ values for water sampled from selected surface-water sites in Salt Lake Valley, Utah	49
17. Graph showing relation between δD and $\delta^{18}\text{O}$ values for water sampled from selected ground-water sites in Salt Lake Valley, Utah	50
18. Graph showing relation between $\delta^{18}\text{O}$ value and chloride concentration in water sampled from selected wells in Salt Lake Valley, Utah	53
19. Map showing vertical hydraulic-conductivity values determined for confining layers in Salt Lake Valley, Utah	56

TABLES

1. Results of chemical analyses of water sampled by the U.S. Geological Survey from selected wells in Salt Lake Valley, Utah, 1982–92	11
2. Records of selected wells that were sampled for chemical analysis of water, Salt Lake Valley, Utah, 1982–92	14
3. Dissolved-solids concentration in water from 11 wells in Salt Lake Valley, Utah, sampled in 1960–66, 1981–84, and 1989–92	31
4. Hydraulic-conductivity values determined from slug tests done at selected wells in Salt Lake Valley, Utah	34
5. Records of wells monitored during four aquifer tests done in Salt Lake Valley, Utah	35
6. Isotopic determinations and physical properties of water from selected surface-water sites in Salt Lake Valley, Utah	45
7. Isotopic determinations and physical properties of water from selected ground-water sites in Salt Lake Valley, Utah	46

CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATED WATER-QUALITY UNITS

Multiply	By	To obtain
cubic foot per day (ft ³ /d)	0.02832	cubic meter per day
cubic foot per second (ft ³ /s)	0.02832	cubic meter per second
cubic foot per second per mile (ft ³ /s/mi)	0.01760	cubic meter per second per kilometer
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot squared per day ¹ (ft ² /d)	0.0929	meter squared per day
foot squared per pound (ft ² /lb)	0.02088	meter squared per newton
gallon per minute (gal/min)	0.06308	liter per second
mile (mi)	1.609	kilometer
pound per cubic foot (lb/ft ³)	157.1	newton per cubic meter
square mile (mi ²)	2.59	square kilometer

Water temperature is reported in degrees Celsius (°C), which can be converted to degrees Fahrenheit (°F) by the following equation:

$$^{\circ}\text{F} = 1.8(^{\circ}\text{C}) + 32.$$

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

Chemical concentration and water temperature are reported only in metric units. Chemical concentration in water is reported in milligrams per liter (mg/L), micrograms per liter (µg/L), moles per liter, or millimoles per liter (mmol/L). These units express the solute mass per unit volume (liter) of water. A liter of water is assumed to weigh 1 kilogram, except for brines or water at high temperatures because of significant changes in the density of the water. One thousand micrograms per liter is equivalent to 1 milligram per liter and one thousand millimoles per liter is equivalent to 1 mole per liter. For concentrations less than 7,000 milligrams per liter, the numerical value is about the same as for concentrations in parts per million. A mole of a substance is its atomic or formula weight in grams. Concentration in moles per liter can be determined from milligrams per liter by dividing by the atomic or formula weight of the constituent, in milligrams. Specific conductance is reported in microsiemens per centimeter at 25 degrees Celsius (µS/cm). Stable isotope concentration is reported as per mil, which is equivalent to parts per thousand. Tritium concentration in water is reported as tritium units (TU). The ratio of 1 atom of tritium to 10¹⁸ atoms of hydrogen is equal to 1 TU or 3.2 picocuries per liter.

¹Expresses transmissivity. An alternative way of expressing transmissivity is cubic foot per day per square foot, times foot of aquifer thickness.

Chemical Composition of Ground Water, Hydrologic Properties of Basin-Fill Material, and Ground-Water Movement in Salt Lake Valley, Utah

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ABSTRACT

The chemical composition and movement of ground water and hydrologic properties of the basin-fill material were studied to better understand the flow system in Salt Lake Valley, Utah. Recharge sources and water-rock interactions influence the water chemistry in the principal and shallow unconfined aquifers. Chloride concentration in water from some wells has increased from the 1950's and 1960's to 1992, possibly because of the infiltration of water that contains dissolved road salt and the movement of ground water with relatively high chloride concentrations from volcanic rocks in the vicinity of these wells.

Hydraulic-conductivity values determined from slug tests done on wells finished in the shallow unconfined aquifer and confining layers ranged from 0.003 to 33.4 feet per day. Transmissivity values determined for the principal aquifer from four multiple-well aquifer tests ranged from 6,400 to 43,600 feet squared per day. Vertical hydraulic-conductivity values estimated from these tests are from 0.01 to 1 foot per day.

Water from 81 sites was analyzed for the stable-isotope ratios of oxygen and hydrogen to determine sources of recharge and mixing of water from these sources. Water sampled from the principal aquifer in the southeastern part of the valley is isotopically similar to water in Big and Little Cottonwood Creeks. Ground water sampled from the shallow unconfined aquifer is generally enriched in oxygen-18 and deuterium relative to other water in the valley, except for water from the Jordan River and its diversions, which are probable recharge sources. Tritium concentrations ranging from about 33 to 59 tritium units in water from the principal aquifer in the southeastern part of the valley are representative of concentrations in water that was recharged in the 1960's.

INTRODUCTION

Salt Lake Valley, north-central Utah, contains the Salt Lake City metropolitan area and drains to the southeastern end of Great Salt Lake (fig. 1). The valley is at the eastern edge of the Basin and Range Physiographic Province (Fenneman, 1931) and is bounded by the Wasatch Range on the east, the Oquirrh Mountains on the west, the Traverse Mountains on the south, and an arbitrary boundary separating Salt Lake and Davis Counties on the north. It also is known as Jordan Valley because the principal stream traversing the area is the Jordan River. The Wasatch fault zone is oriented in a north-south direction and generally separates the consolidated rock of the Wasatch Range from unconsolidated basin-fill material deposited in the valley.

An adequate supply of water that is suitable for domestic use is one of the most important factors in sustaining the current population and in allowing for continued economic growth in Salt Lake Valley. Water managers and suppliers are in need of detailed information about the occurrence and possible flow of this water caused by water-use practices of ground water that is not suitable for domestic use.

Water in the principal aquifer is used for public supply in many parts of Salt Lake Valley. Most of the water that is suitable for domestic use is in the southeastern part of the valley, downgradient from the mouths of Big and Little Cottonwood Canyons. Water from the principal aquifer in parts of the northern and western parts of the valley contains greater concentrations of dissolved solids than water from other parts of the valley.

The shallow unconfined aquifer is susceptible to contamination from activities and processes that occur at land surface. Relatively high concentrations of trace metals, organic compounds, and dissolved solids were found in water from some parts of the shallow unconfined aquifer (Seiler and Waddell, 1984, p. 10). The extent of the confining layers that separate the shallow

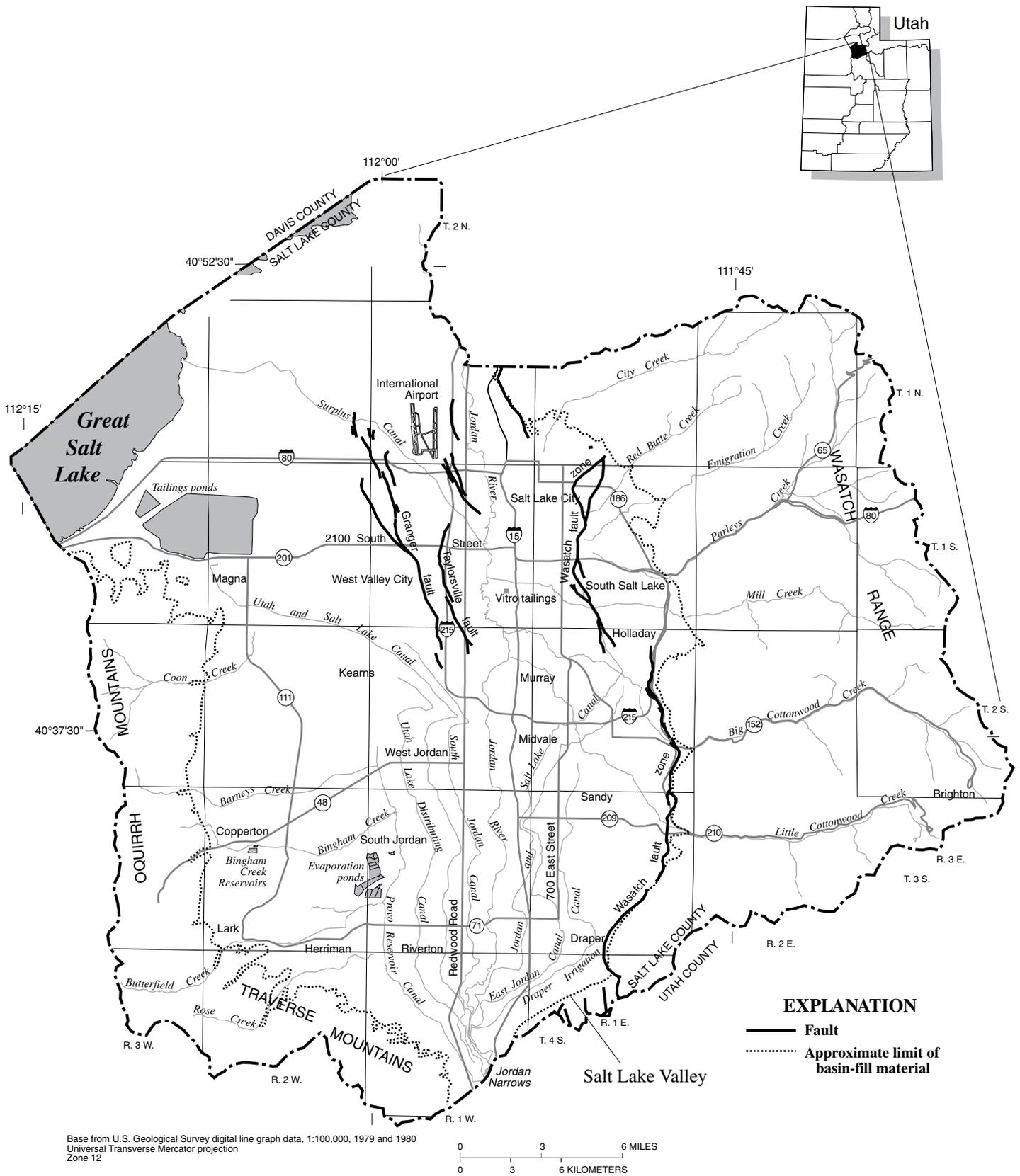


Figure 1. Location of Salt Lake Valley study area, Utah.

unconfined aquifer from the principal aquifer and their effectiveness as a barrier to the movement of contaminated ground water is not well known. Withdrawals from the principal aquifer could alter hydraulic-head gradients enough to induce the movement of water with higher concentrations of dissolved solids or contaminants in the principal and shallow unconfined aquifers into areas with water that is suitable for domestic use. More information is needed about the ground-water flow system so that future ground-water withdrawals can be managed efficiently.

Ground-water flow and solute migration in Salt Lake Valley was studied from 1990 to 1994 to better define the ground-water flow system and to determine changes in water composition that could occur as a result of increased withdrawals from the principal aquifer. This study was done by the U.S. Geological Survey in cooperation with the Utah Department of Natural Resources, Division of Water Rights, and the Utah Department of Environmental Quality, Division of Water Quality. Several local water suppliers also cooperated in the study, including Salt Lake City Corporation, Salt Lake County Water Conservancy District, Taylorsville-Bennion Improvement District, Granger-Hunter Improvement District, Murray City, City of South Salt Lake, and Kearns Improvement District.

Purpose and Scope

This report is the second in a series of reports produced from the study. The first report presented hydrologic data collected in Salt Lake Valley from 1990 to 1992. This report interprets those data and describes selected chemical properties of water and hydrologic properties of the basin-fill material in the valley.

This report defines the chemical composition of water from the shallow unconfined aquifer, underlying confining layers, and selected wells finished in the principal aquifer; presents values for hydrologic properties of the basin-fill material; and better defines ground-water movement and the flow system using water-chemistry, hydrologic-property, and isotopic information. The methods used to estimate hydrologic-property values of the basin-fill material and the results of these methods are described. The stable isotopes of oxygen and hydrogen were used to help determine recharge sources, and tritium, a radioactive isotope of hydrogen, was used to estimate ground-water-flow velocity values and time of recharge.

Most of the data discussed in this report were collected from wells in the valley from 1990 to 1992 (pl. 1). In June and July 1990, and March and May 1991, 36 monitoring wells were drilled to allow collection of data on the shallow unconfined aquifer and, depending on finish depth, an underlying or interbedded fine-grained layer. Depths of the monitoring wells ranged from 15.0 to 129.5 ft below land surface. Data also were collected from public-supply wells and other existing wells finished in the principal aquifer near a monitoring well finished in the shallow unconfined aquifer. Information from the shallow unconfined aquifer was compared with information from deeper intervals at these sites. Water levels, concentrations of chemical constituents, selected hydrologic properties, and other data from these wells are presented in a hydrologic-data report for Salt Lake Valley (Thiros, 1992). Water samples from selected public-supply wells in the southeastern part of the valley were analyzed for stable and radioactive isotopes of oxygen and hydrogen. The numbering system used in Utah for hydrologic-data sites is shown in figure 2.

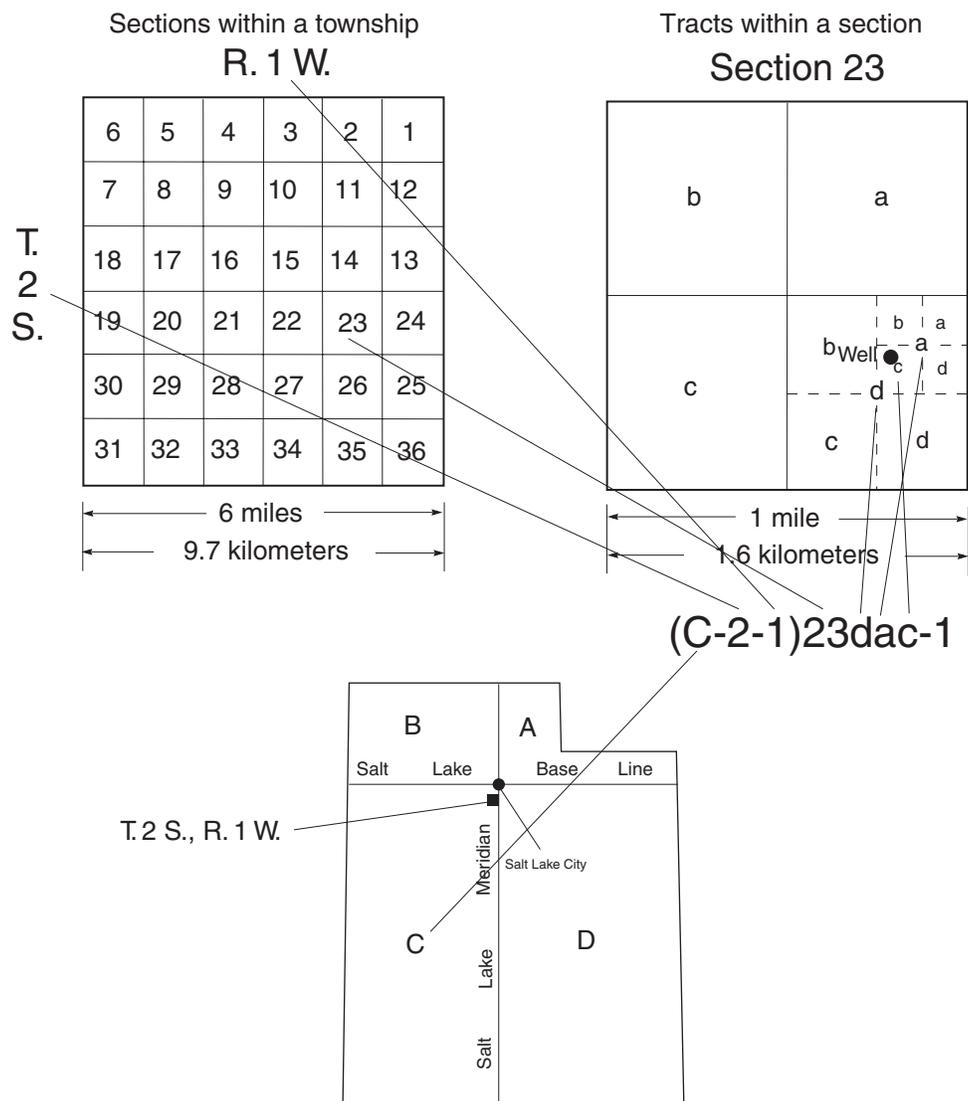
Acknowledgments

The cooperation of property owners who granted permission to drill monitoring wells on their property and who allowed access to these wells and other existing wells is greatly appreciated. Thanks also are extended to the public-water suppliers who provided access to their wells for use in the aquifer tests done during this study.

Concept of Ground-Water System

The saturated basin-fill material in Salt Lake Valley has previously been described as being divided into a deep unconfined aquifer near the mountain fronts, a confined aquifer, a shallow unconfined aquifer overlying the confined aquifer, and locally unconfined or perched aquifers (Hely and others, 1971, p. 107–111). The deep unconfined aquifer is composed of coarse-grained unconsolidated sediments deposited during the Quaternary and possibly late Tertiary periods. The confined aquifer consists of interbedded deposits of clay, silt, sand, and gravel of Quaternary and possibly Tertiary age overlain by discontinuous layers of silt and clay of Quaternary age. Thickness of the confined aquifer exceeds 1,000 ft in the northern part of the valley. The underlying unconsolidated to semiconsolidated

The system of numbering wells, springs, and other hydrologic-data sites in Utah is based on the cadastral land-survey system of the U.S. Government. The number, in addition to designating the site, describes its position in the land net. By the land-survey system, the State is divided into four quadrants by the Salt Lake Base Line and the Salt Lake Meridian. These quadrants are designated by the uppercase letters A, B, C, and D, indicating the northeast, northwest, southwest, and southeast quadrants, respectively. Numbers designating the township and range, in that order, follow the quadrant letter, and all three are enclosed in parentheses. The number after the parentheses indicates the section and is followed by three letters indicating the quarter section, the quarter-quarter section, and the quarter-quarter-quarter section — generally 10 acres for regular sections¹. The lowercase letters a, b, c, and d indicate, respectively, the northeast, northwest, southwest, and southeast quarters of each subdivision. The number after the letters is the serial number of the site within the 10-acre tract. The letter S preceding the serial number designates a spring. Thus, (C-2-1)23dac-1 designates the first well constructed or visited in the southwest 1/4 of the northeast 1/4 of the southeast 1/4 of section 23, T. 2 S., R. 1 W. The numbering system is illustrated below.



¹Although the basic land unit, the section, is theoretically 1 square mile, many sections are irregular. Such sections are subdivided into 10-acre tracts, generally beginning at the southeast corner, and the surplus or shortage is taken up in the tracts along the north and west sides of the section.

Figure 2. Numbering system used for hydrologic-data sites in Utah.

sediment and volcanic rocks of Tertiary age are generally less permeable than the sediment of Quaternary age. The basin-fill material forms the ground-water system in Salt Lake Valley (fig. 3). For the purpose of this report, the deep unconfined and confined aquifers are grouped together and referred to as the principal aquifer.

The shallow unconfined aquifer in Salt Lake Valley typically is present where confining layers overlie the principal aquifer. The shallow unconfined aquifer overlies these confining layers, which are typically within 50 ft of land surface, although their depth, thickness, and even presence in the valley is variable. Confining conditions may exist at shallower depths if fine-grained material is present, such as in the northwestern and central parts of the valley. Major sources of recharge to the shallow unconfined aquifer are upward leakage of water from the principal aquifer and seepage from surface-water bodies.

Primary recharge areas for the principal aquifer are near the mountain fronts (fig. 3), where there are no substantial layers of fine-grained material to impede ground-water movement. Recharge to the principal aquifer mainly is from infiltration of snowmelt runoff from the mountain fronts in the spring, as subsurface inflow from fractured consolidated rock, or as a combination of both. Ground-water recharge also comes from the seepage of water from streams and canals, infiltration of unconsumed irrigation water, and infiltration of precipitation on the valley floor.

Downward leakage of water from the shallow unconfined aquifer to the principal aquifer is possible where there is a downward hydraulic-head gradient between the shallow unconfined aquifer and the principal aquifer. These conditions exist beneath the secondary recharge area, despite the presence of confining layers. Ground-water movement from the principal aquifer upward to the shallow unconfined aquifer is possible where there is an upward hydraulic-head gradient, such as in the discharge area in the valley. Water in the principal aquifer typically moves from the mountain fronts toward the center of the valley and eventually becomes confined by overlying fine-grained deposits. The approximate boundary between the primary and secondary recharge areas and the extent of the discharge area for the principal aquifer in the valley (Anderson and others, 1994) are shown on plate 1. The presence of confining layers more than 20 ft thick was determined from drillers' logs of wells in the valley and

the information was used to delineate the boundary between the recharge areas.

Methods and Limitations

The types of data collected and the methods used to analyze these data are described in the following sections. Some of the limitations associated with the data, methods, and results also are presented.

Chemical Composition

Chemical analyses of water from public-supply wells in Salt Lake County were obtained from the Utah Department of Environmental Quality, Division of Drinking Water (written commun., 1993). Most of the wells used for public supply in the valley yield water from the principal aquifer. Chemical analyses of water from the principal aquifer in the southwestern part of the valley were obtained from Kennecott Utah Copper (1992) and were divided into two sets on the basis of finish depth: wells with perforated, screened, slotted, or open intervals from 100 to 300 ft below land surface, and wells with perforated, screened, slotted, or open intervals at depths greater than 300 ft below land surface. Chemical analyses of water collected from 1988 to 1992, along with data collected by the U.S. Geological Survey, were used to modify the previous distribution of dissolved-solids concentration in water from the principal aquifer (Waddell and others, 1987, pl. 1).

Chemical analyses of water sampled from wells finished in the basin-fill material in Salt Lake Valley from 1982 through 1992 by the U.S. Geological Survey were used to characterize selected chemical properties. The basin-fill material was divided into three sections: the shallow unconfined aquifer and confining layers, intermediate depths that generally correspond to the upper part of the principal aquifer, and the principal aquifer. Chemical analyses of water are divided into groups on the basis of well depths that generally correspond to these sections. This grouping was made to simplify analysis of the data and may not accurately represent the appropriate aquifer in all areas of the valley. The three groups represent intervals from 0 to 100 ft, 101 to 300 ft, and greater than 300 ft below land surface, respectively.

The chemical-equilibrium model WATEQ4F (Ball and Nordstrom, 1991) was used to determine the state of saturation in water with respect to selected min-

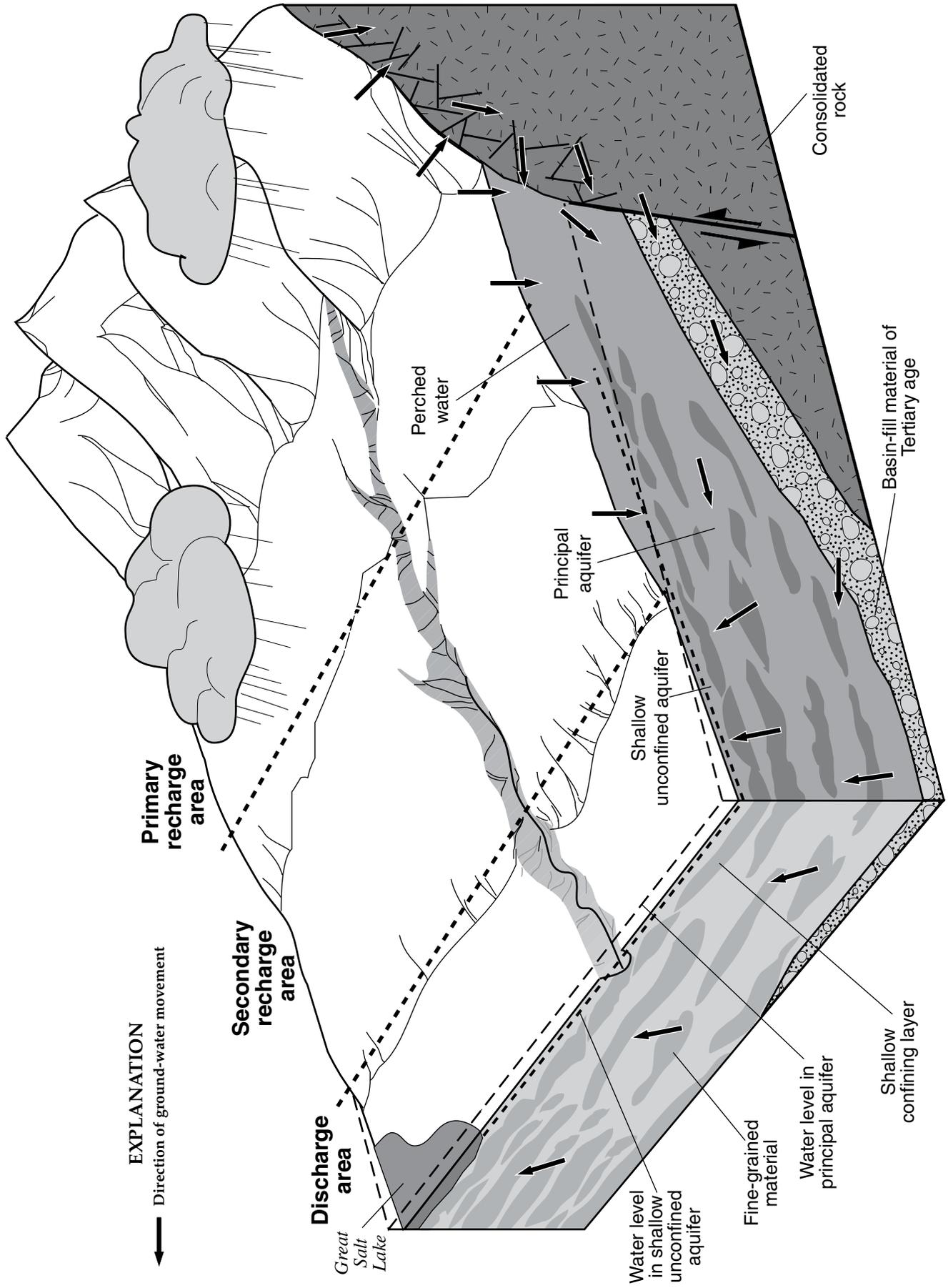


Figure 3. Generalized block diagram showing the basin-fill ground-water flow system in Salt Lake Valley, Utah. (Modified from Hely and others, 1971.)

erals. The computer program uses analytical data to calculate activities and mineral saturation indices. Minerals have the potential to dissolve in a solution if saturation indices are negative, precipitate if saturation indices are positive, or remain at equilibrium with the solution if saturation indices are equal to or near zero. Limitations to the thermodynamic data used by the WATEQ4F program and analytical errors affect the degree of uncertainty associated with the calculated saturation indices.

The computer program NETPATH (Plummer and others, 1991) was used to model the geochemical mass-transfer reactions that take place in ground water as it travels along a flow path. The program requires the specification of phases and chemical constraints that occur within the ground-water system. A phase is any mineral or gas that exists along a ground-water flow path that can react with the water. Constraints are selected from differences in concentrations of chemical elements, redox state, temperature, and isotopic ratios between upgradient and evolutionary downgradient measurements. Constraints are used to limit the masses of selected phases that can be added to or removed from ground water along its flow path. NETPATH also can be used to calculate the proportions of two waters in a mixture and to quantify plausible chemical reactions that account for changes in concentrations along a flow path.

The modeling results report all the reactions possible and mainly are used to eliminate reaction models that are not mass balanced from further consideration. Limitations of geochemical mass-balance models are imposed by the reliability and completeness of the chemical, mineralogic, and isotopic data and also are constrained by the reliability of choosing an actual flow path. Analytical error or a lack of data on dissolved species in a water affects mass-balance calculations. Mineralogic and petrologic information for the system is necessary to provide guidance in the selection of appropriate phases to be used in the modeling process.

Hydrologic Properties

Core samples of basin-fill material were collected from selected depths during drilling of the monitoring wells in 1990 and 1991. The cores were collected with a split-barrel sampler containing an acrylic core liner that was pushed into previously undisturbed material at the bottom of the drill hole. Some of the cores that contained mostly fine-grained material

may represent confining layers and were sent to a geotechnical laboratory for measurement of selected hydrologic properties. These samples of fine-grained material were tested for hydraulic conductivity, compressibility, porosity, dry density, moisture content, particle-size distribution, and void ratio. Values determined for these properties are given by Thiros (1992, tables 12 and 13).

Hydraulic-conductivity values determined by falling-head permeameter tests done on fine-grained material are influenced by the movement of water in the vertical direction. In the falling-head permeameter test, water must flow upward through the core sample in order to determine the flow rate through the medium. The core material tested generally represents sediment that could impede flow between the shallow unconfined aquifer and the principal aquifer. Drillers' logs indicate that fine-grained material typically is interlayered with more permeable sediment and that layers are discontinuous. Permeameter tests do not account for flow through the more permeable material interlayered with the fine-grained material because of the small section of material tested. Hydraulic-conductivity values determined by falling-head permeameter tests ranged from about 5.1×10^{-5} ft/d for clay from 38.5 ft below land surface at well (C-1-1)15cad-2 to 2×10^{-2} ft/d for silt from 19 ft below land surface at well (B-1-2)36baa-3. This range in values determined by laboratory methods is probably at the low end of the range in effective hydraulic conductivity that exists at a confining-layer scale. Limitations of the falling-head method are that the material to be tested is not restored to the overburden pressure that it was under prior to coring, the cored sample may have been disturbed during sampling, and the fluid being used in the test can leak around the sample along the core-barrel wall.

Specific storage is the volume of water that a unit volume of saturated material releases from storage under a unit decline in hydraulic head. Specific storage can be expressed as:

$$S_s = \gamma (\alpha + \eta \beta) \quad (1)$$

where S_s is the specific storage of the saturated material, in ft^{-1} ,
 γ is the specific weight of water, typically 62.4 lb/ft^3 ,
 α is the vertical compressibility of the material, in ft^2/lb ,
 η is the porosity of the saturated material, dimensionless, and
 β is the compressibility of water, in ft^2/lb .

The specific storage of fine-grained material was determined from consolidation tests done on the core samples. If the compressibility of water (2.3×10^{-8} ft²/lb) and porosity terms are disregarded, then $S_s = \gamma \alpha$.

Vertical compressibility was determined from consolidation-test data using:

$$\alpha = (\text{height}_0 - \text{height}_1 / \text{height}_0) / \text{stress}_0 - \text{stress}_1 \quad (2)$$

where height_0 is the initial height of the sample, height_1 is the height of the sample at time t_1 , stress_0 is the effective stress at height_0 , and stress_1 is the effective stress at height_1 .

The effective stress that the core material was under prior to coring was calculated to be the total stress exerted by the overlying water and material minus the neutral pore pressure. The loads applied to the cores used to calculate vertical compressibility were in the range of values determined for existing overburden pressure. Values of specific storage calculated from consolidation-test data for this study ranged from 1.7×10^{-4} to 9.4×10^{-4} ft⁻¹. These values are in the range of values listed for loose sand to clay on the basis of the determination of compressibility (Domenico and Mifflin, 1965, p. 566).

Slug tests were done at 30 of the monitoring wells installed in 1990 and 1991 to estimate the hydraulic conductivity of the shallow unconfined aquifer and confining layers (pl. 1). The hydraulic conductivity of an aquifer open to a well can be estimated from the rate in water-level decline or recovery after a cylinder with a known volume (slug) has been added or removed from the water. The slug tests were done by installing a pressure transducer in a well to monitor water-level changes at various time intervals. Water-level data were stored in a data recorder. The slug was quickly lowered below the water surface in the well and the water-level response in the well was periodically measured until it regained a state of equilibrium with the aquifer. This process was repeated by raising the slug out of the water and monitoring the recovery in water level with time.

Methods for determining hydraulic conductivity from slug tests have been developed for unconfined and confined conditions. The Bouwer and Rice slug-test solution is applicable to unconfined and confined conditions. This method of analysis was used for all of the tests and is discussed by Bouwer and Rice (1976) and Bouwer (1989). The Cooper, Bredehoeft, and Papadopoulos method is applicable only to confined conditions (Cooper and others, 1967) and requires matching

water-level change measured in a well with time to a set of computed type curves.

The slug-test method has limitations in estimating the hydraulic conductivity of an aquifer. The value determined is only representative of the aquifer near the finish interval of the well being tested and is influenced by the grain size of the material in the disturbed area around the finish interval. Fine-grained material around the finish intervals of the well can impede the flow of water into and out of the well if it has not been properly completed and developed. The thickness of the developed zone is typically unknown. Both solutions used to interpret the slug-test data assume that the aquifer is isotropic and that leakage from above or below the developed zone is not occurring.

Hydrologic properties were estimated from aquifer tests incorporating multiple wells at three sites in Salt Lake Valley during February 1991 and February 1992 and from data from another test that was reanalyzed as part of this study (pl. 1). The aquifer tests were done to determine if significant leakage through confining layers was being induced by ground-water withdrawals from the principal aquifer. The pumped wells used to stress the system in each test were existing public-supply wells with multiple finish intervals at variable depths. Transmissivity and storage coefficient of the pumped principal aquifer, and vertical hydraulic conductivity and specific storage of confining layers that separate the principal aquifer from the shallow unconfined aquifer, were estimated from the results of these tests.

Water-level changes were monitored with steel or electrical tapes in the pumped wells and in nearby observation wells completed in the shallow unconfined and principal aquifers. Pressure transducers and data recorders also were installed at some of the wells. These data were compared with the measurements made with the steel and electrical tapes to correct for measurement error and instrument drift.

Prior to when the pumps were turned on or off, water level was measured in the wells for a period lasting from 2 to 20 days. Water-level changes in wells monitored during each test were corrected for pre-test water-level trends and barometric-pressure effects if a relation to water-level change could be determined. Water-level declines in response to ground-water withdrawals were calculated by subtracting the water level measured while the pump was on from the water level measured just before pumping began. Water-level recovery in response to the end of pumping was calcu-

lated by subtracting water levels measured after pumping had stopped from the water level measured just before pumping began (residual drawdown). For most of the tests analyzed, more water-level measurements were available during the recovery period than during the pumping period, although results from both sets of measurements generally correspond.

Transmissivity of the principal aquifer was estimated using the straight-line method (Cooper and Jacob, 1946) derived from the Theis solution (Theis, 1935). Some of the assumptions of the Theis solution are that the aquifer is infinite in lateral extent, is confined above and below by impermeable material so that no leakage can occur, and is homogeneous and isotropic. Water-level recovery in the pumped well was plotted on a linear scale and the corresponding time since pumping had stopped was plotted on a logarithmic scale. The straight-line method uses the equation:

$$T = 2.3 Q / 4 \pi \Delta s \quad (3)$$

where T is transmissivity, in ft^2/d ,
 Q is discharge rate of the pumped well, in ft^3/d , and
 Δs is the change in drawdown in the well during one base 10 logarithmic cycle of time, in ft.

The Hantush modified method (Hantush, 1960) also was used to determine the hydrologic properties of the principal aquifer. This method can be used to analyze leaky confined systems because the release of water from storage in confining layers is accounted for. A composite logarithmic plot of drawdown or recovery from the pumped aquifer as a function of time since pumping began or ended divided by the square of the distance from the pumped well was constructed for data from each aquifer test. If appreciable leakage from confining layers to a confined aquifer is occurring, water-level changes in observation wells with increasing radial distances from the pumped well will deviate from the Theis type curve that represents no leakage. An indication of little to no leakage is if the data curves for wells at different radial distances from the pumped well are superimposable. The data curves were matched to a set of type curves representing the Hantush function and values of a leakage parameter, β (Lohman, 1979, pl. 4).

Vertical hydraulic conductivity of the confining layers (k') was estimated using the Hantush (1960) modified method and the Neuman and Witherspoon (1972) "ratio" method. These methods include an additional variable: specific storage of the confining layer

(S_s'). Vertical hydraulic conductivity of the confining layers can be estimated by the Hantush modified method using the equation:

$$k' S_s' = 16 T S \beta^2 / r^2 \quad (4)$$

where k' is the vertical hydraulic conductivity of the confining layers, in ft/d ,
 S_s' is the specific storage of the confining layers, in ft^{-1} ,
 T is the transmissivity of the pumped aquifer, in ft^2/d ,
 S is the storage coefficient of the pumped aquifer, dimensionless,
 β is a leakage parameter, dimensionless, and
 r is the radial distance from the pumped well, in ft.

A water-level change in response to pumping was measured in some monitoring wells completed near the base of the shallow unconfined aquifer or in a confining layer. The "ratio" method of Neuman and Witherspoon (1972) requires the ratio of drawdown in a confining layer to that in the pumped aquifer at the same radial distance from the pumped well and at the same instant in time. The method provides the hydraulic diffusivity of the confining layer from which vertical hydraulic conductivity can be determined if the specific storage of the layer is known or estimated.

Neuman and Witherspoon (1972, p. 1296) determined specific storage values for confining-layer material from consolidation tests done in a laboratory. These values are likely too large because they were determined from applied stresses greater than the maximum past stress received by the material prior to coring (Francis Riley, U.S. Geological Survey, oral commun., 1992). Because the specific-storage values determined in the laboratory for core material from Salt Lake Valley are the same order of magnitude as those used by Neuman and Witherspoon, they were not used to estimate vertical hydraulic conductivity.

Isotopes

Because isotopes of an element have the same number of protons but different numbers of neutrons in the nucleus of the atom, all isotopes of an element have the same atomic number but different mass numbers. An isotope is stable if it does not undergo radioactive decay. Water is composed of two atoms of hydrogen and one atom of oxygen. The most abundant isotopes of hydrogen and oxygen are ^1H and ^{16}O . Small quantities

of deuterium (D or ^2H) and oxygen-18 (^{18}O) are incorporated into the water molecule in average proportions of about 0.01 percent and 0.2 percent, respectively (Hem, 1985, p. 162). Hydrogen and oxygen isotopic ratios are reported in parts per thousand (per mil) deviations from a reference standard called Standard Mean Ocean Water (SMOW) (Craig, 1961). These deviations from the SMOW standard are in the form

$$\delta R = \left[\frac{R_{\text{sample}}}{R_{\text{standard}}} - 1 \right] \times 1,000 \quad (5)$$

where R_{sample} is equal to the ratio of ^{18}O to ^{16}O or of D to ^1H , in a water sample,
 R_{standard} is equal to the ratio of ^{18}O to ^{16}O or of D to ^1H in a reference standard (SMOW), and
 δR is the $\delta^{18}\text{O}$ or δD value of the water sample.

Isotopic fractionation occurs during physical and chemical processes and results in phases or substances with different isotopic ratios. Changes in isotopic ratios in water can be primarily attributed to condensation and evaporation processes, which are temperature dependent. Heavier isotopes tend to remain in the condensed phase, leaving the water vapor depleted in heavier isotopes. Because D and ^{18}O are generally conservative in low-temperature ground-water systems, water enters the ground-water system with an isotopic signature indicative of conditions at the time of recharge. This signature can be altered by mixing with other sources of water within the ground-water system.

The relation between $\delta^{18}\text{O}$ and δD values in meteoric water is expressed by the equation

$$\delta\text{D} = 8 \delta^{18}\text{O} + d \quad (6)$$

where d is the deuterium excess parameter. This relation can be drawn as a straight line on a plot of δD with $\delta^{18}\text{O}$ values. The line generally represents precipitation that has not undergone evaporation and is called the meteoric water line. The mean value for d for about 400 water samples from rivers, lakes, and precipitation from around the world is about 10 per mil (Craig, 1961). The straight line produced using a value of d equal to 10 is known as the global meteoric water line.

Water from selected hydrologic-data sites in Salt Lake Valley was analyzed for tritium to determine relative age and to assist in defining the direction of ground-water flow in the southeastern part of the valley.

Tritium is a radioactive isotope of hydrogen (^3H) with a half-life of 12.43 years.

Large quantities of tritium were added to the atmosphere from above-ground nuclear-weapons testing beginning about 1952. Tritium concentrations in precipitation prior to nuclear-weapons testing have been estimated to be about 8 TU (Kaufmann and Libby, 1954). Tritium concentrations in water recharged to the basin-fill material prior to 1952, assuming no mixing with other sources of water, would have decayed to less than 1 TU by 1992. Concentrations measured in precipitation in Salt Lake City by the U.S. Geological Survey peaked in 1963 at 8,230 TU, about three orders of magnitude greater than estimated concentrations in precipitation prior to nuclear-weapons testing. Tritium concentrations in precipitation have declined since 1963 and were about 10 to 15 TU during this study (Bob Michel, U.S. Geological Survey, oral commun., 1992). The presence and relative concentrations of tritium in ground water can provide insight to when the water last had contact with the atmosphere.

A limitation of using tritium to date water is the variable nature of its production by nuclear-weapons testing since 1952. Large seasonal fluctuations in tritium concentrations also occur. Precise dating of ground water is not possible when these limitations are combined with radioactive decay and the possible mixing of water of different ages.

CHEMICAL COMPOSITION OF GROUND WATER

The chemical composition of ground water in Salt Lake Valley varies with depth and location. Recharge sources and water-rock interactions influence the chemistry of water in the principal and shallow unconfined aquifers. Results of analyses of water samples collected by the U.S. Geological Survey were grouped together on the basis of well depth and are listed in table 1. Information about the wells for which water-chemistry data are available is listed in table 2.

Water in the Principal Aquifer

Water in the principal aquifer generally contains a lower concentration of dissolved solids than water in other parts of the ground-water system. Hely and others (1971, pl. 2) and Waddell and others (1987, pl. 1) mapped dissolved-solids concentration in water from

Table 1. Results of chemical analyses of water sampled by the U.S. Geological Survey from selected wells in Salt Lake Valley, Utah, 1982–92

[mg/L, milligrams per liter; —, no data available; <, less than]

Map no.: Corresponds to number next to data point on figures 4, 5, and 8 for water samples from wells greater than 300 feet deep, on figures 9–11 for water samples from wells 101 to 300 feet deep, and on figures 12 and 13 for water samples from wells less than 101 feet deep.

Location: See figure 2 for an explanation of the numbering system used for hydrologic-data sites in Utah.

Well depth: In feet below land surface.

Water temperature: °C, degrees Celsius. Measured in the field by U.S. Geological Survey personnel.

Specific conductance: µS/cm, microsiemens per centimeter at 25 degrees Celsius. Measured in the field by U.S. Geological Survey personnel.

pH: Measured in the field by U.S. Geological Survey personnel except where noted L, laboratory value.

Alkalinity: Measured in the field by U.S. Geological Survey personnel except where noted L, laboratory value.

Solids, dissolved: Sum of constituents except where noted R, residue on evaporation at 180 degrees Celsius.

Map no.	Location	Well depth (feet)	Date	Water temperature (°C)	Specific conductance (µS/cm)	pH, field (standard units)	Calcium dissolved (mg/L as Ca)	Magnesium dissolved (mg/L as Mg)	Sodium dissolved (mg/L as Na)	Potassium dissolved (mg/L as K)	Alkalinity, total (mg/L as CaCO ₃)	Sulfate, dissolved (mg/L as SO ₄)	Chloride, dissolved (mg/L as Cl)	Fluoride, dissolved (mg/L as F)	Silica, dissolved (mg/L as SiO ₂)	Solids, dissolved (mg/L)
Water samples from wells greater than 300 feet deep																
1	(A-1-1)3lcac-1	464	06-19-89	14.0	960	7.3	95	35	51	—	213 L	86	92	0.20	20	588 R
2	(B-1-1)19baa-5	645	07-31-84	19.0	1,390	8.0	25	16	240	1.9	262	3.6	300	1.0	22	767
3	(B-1-2)2dac-1	440	08-29-91	27.0	870	8.2	12	7.3	170	2.1	249 L	0.30	130	2.0	24	498
4	(B-1-2)7caa-1	735	08-01-84	17.5	5,780	7.8	92	91	960	3.5	240	22	1800	1.1	22	3,140
5	(B-1-2)7ccc-1	389	07-19-84	13.5	8,710	8.0	73	83	1,700	4.2	177 L	340	2,600	1.0	21	4,930
6	(B-1-2)16caa-1	636	08-20-84	22.5	2,750	7.8	70	21	470	12	156	8.3	840	1.5	48	1,570
7	(B-1-2)19aca-1	450	08-05-91	17.5	2,350	8.4	7.0	6.9	480	2.5	380	110	400	2.6	23	1,260
8	(B-1-2)21acd-1	600	08-20-84	21.5	1,070	8.0	16	5.9	210	6.1	224	1.9	220	1.9	53	649
9	(B-1-2)21cab-2	330	08-15-84	18.0	1,290	8.3	6.7	4.8	300	2.4	340	4.2	250	2.3	22	796
10	(B-1-2)21dcd-1	561	08-06-91	16.5	1,180	8.6	19	8.1	230	6.8	240	9.5	230	1.9	52	708 R
11	(B-1-2)23bbd-1	560	07-19-84	16.5	1,660	8.2	13	11	340	2.2	313 L	5.0	350	1.4	22	932
12	(B-1-2)32bcb-1	716	12-09-91	18.5	5,500	7.7	110	69	950	16	156	100	1,800	.91	49	3,190
13	(B-1-2)36baa-1	464	08-29-91	27.0	6,180	7.7	180	60	1,000	19	118	45	1,900	1.2	53	3,570 R
14	(C-1-1)9bba-3	341	08-17-83	17.0	525	8.0	33	13	70	2.4	200	59	17	.40	21	336
15	(C-1-1)15bdd-1	445	07-10-84	15.5	730	8.0	76	26	41	2.7	182	190	18	.40	22	486
16	(C-1-1)23bcb-1	840	07-13-89	15.0	820	7.9	72	29	46	—	172 L	150	55	.40	22	422 R
17	(C-1-1)25bdb-1	1,000	07-10-89	18.5	540	7.7	52	19	29	—	161 L	97	13	.40	20	320 R
18	(C-1-1)25cca-2	345	07-07-83	14.0	405	7.6	50	18	11	1.5	161 L	55	11	.30	15	259
19	(C-1-1)27dac-4	308	07-08-83	15.0	510	7.9	50	17	35	1.7	173 L	97	11	.40	22	338
20	(C-1-1)27dda-8	775	07-10-89	21.0	710	7.7	50	15	65	—	130 L	120	60	.50	25	416 R
21	(C-1-1)33ddd-1	886	08-06-92	19.5	1,520	7.6	76	51	160	6.4	149 L	250	280	.40	49	969
22	(C-1-2)2aba-2	410	07-13-83	22.0	1,730	7.9	41	19	290	7.7	162 L	11	470	.90	41	978
23	(C-1-2)6aaa-1	760	07-27-84	21.0	5,470	7.8	230	130	810	15	57	29	2,100	.60	24	3,370
24	(C-1-2)6aaa-3	825	07-27-84	23.0	3,370	7.9	110	55	520	8.3	91	35	1,100	.60	30	1,910
25	(C-1-2)6aaa-5	345	07-27-84	18.5	4,750	7.6	170	100	820	23	146	22	1,700	.80	31	2,960
26	(C-1-2)14cdd-1	480	08-19-91	21.0	3,420	7.8	95	43	550	17	124	200	900	.79	48	1,990 R
27	(C-2-1)20aad-1	658	07-10-89	15.0	2,260	7.3	180	130	210	—	243 L	750	380	.30	43	2,070 R
28	(C-2-1)24bcd-1	370	02-10-92	15.5	650	7.8	57	24	35	2.6	129	70	72	.26	18	368 R
29	(C-3-1)7cbb-1	462	09-06-84	14.5	1,260	7.5	120	47	55	3.6	142 L	160	230	.20	26	737
30	(C-3-1)14bbc-1	395	09-17-84	16.5	980	7.6	78	37	55	3.0	135	82	190	.20	22	572
31	(C-3-1)15dda-1	360	07-12-84	14.5	1,220	7.6	120	41	57	3.1	146	130	240	.10	24	709
32	(C-3-2)12dbc-2	360	09-04-84	14.0	2,130	7.0	270	75	120	4.2	286	320	400	.20	25	1,420
33	(C-3-2)33cac-1	350	09-19-84	15.0	1,290	7.2	170	50	44	4.4	266	200	200	.20	34	782 R
34	(C-4-2)1bbb-1	540	07-20-89	14.5	1,300	7.3	140	35	57	8.6	191 L	73	220	.20	49	705
35	(D-1-1)19cdb-17	500	08-18-83	14.0	950	7.5 L	120	38	40	3.4	248 L	220	43	.30	16	643
36	(D-1-1)20bab-1	482	07-07-83	17.0	940	7.3	98	42	43	3.2	203 L	280	33	.30	18	640
37	(D-1-1)21acc-2	576	06-28-89	14.0	950	7.6	100	36	44	—	245 L	150	59	.20	22	616 R
38	(D-2-1)21dbc-1	740	07-21-92	11.5	275	8.1	31	10	9.0	1.5	102 L	24	11	.20	10	161
39	(D-2-1)28ccc-1	691	09-18-90	12.0	240	8.2	25	8.9	8.4	1.5	82	20	11	.20	13	137
40	(D-2-1)29acb-8	500	09-17-90	14.0	225	8.3	23	8.5	7.7	1.4	78	19	12	.30	11	130
41	(D-2-1)30dda-1	1,002	02-10-92	14.0	295	7.9	28	11	14	1.8	91	34	18	.27	13	174 R
42	(D-2-1)34cdc-1	875	09-18-90	12.0	375	7.6	44	9.9	19	2.4	138	31	14	.30	12	215
43	(D-3-1)4cad-1	938	09-17-90	11.5	310	8.1	37	7.9	12	2.4	121	16	12	.10	15	175
44	(D-3-1)6dad-1	1,000	09-17-90	17.0	415	8.1	42	10	19	2.8	93	6.8	66	.30	14	217

Table 1. Results of chemical analyses of water sampled by the U.S. Geological Survey from selected wells in Salt Lake Valley, Utah, 1982–92—Continued

Map no.	Location	Well depth (feet)	Date	Water temperature (°C)	Specific conductance (µS/cm)	pH, field (standard units)	Calcium dissolved (mg/L as Ca)	Magnesium dissolved (mg/L as Mg)	Sodium dissolved (mg/L as Na)	Potassium dissolved (mg/L as K)	Alkalinity, total (mg/L as CaCO ₃)	Sulfate, dissolved (mg/L as SO ₄)	Chloride, dissolved (mg/L as Cl)	Fluoride, dissolved (mg/L as F)	Silica, dissolved (mg/L as SiO ₂)	Solids, dissolved (mg/L)
Water samples from wells 101 to 300 feet deep																
1	(B-1-1)9adb-1	255	05-25-89	18.5	4,520	6.9	170	53	630	—	322 L	<1.0	1,300	0.80	30	2,460 R
2	(B-1-2)8abd-1	300	08-01-84	19.5	1,600	8.1	25	18	280	2.4	165 L	4.6	440	1.6	22	893
3	(B-1-2)15bcd-2	300	07-06-84	20.0	800	8.2	11	6.8	160	2.1	165 L	2.2	140	2.1	24	447
4	(B-1-2)22daa-1	148	08-07-91	15.5	1,670	8.2	18	16	290	2.9	261	1.3	330	1.2	22	878 R
5	(B-1-2)23cbd-1	270	08-07-91	17.0	1,780	8.0	23	18	310	3.3	222	1.2	400	1.0	24	978 R
6	(B-1-2)32bcb-2	248	08-06-91	20.0	13,000	7.8	81	92	3,000	23	253	510	4,100	1.2	25	8,160 R
7	(C-1-1)8aac-3	252	08-17-83	16.0	700	8.1	50	19	76	3.1	160	180	25	.40	24	474
8	(C-1-1)23caa-1	175	08-19-83	13.0	1,540	7.6	150	83	56	13	170	340	250	.20	15	1,010
9	(C-1-1)25bdc-1	135	03-30-83	13.0	503	8.0	45	20	31	11	243 L	3.3	10	.30	25	274 R
10	(C-1-1)26caa-1	115	07-12-84	12.5	1,940	7.4	220	90	87	23	213	490	320	.20	18	1,380
11	(C-1-1)26dba-5	129	05-23-83	12.5	1,740	7.4	180	76	48	15	158 L	360	270	.20	28	1,070
12	(C-1-1)35ddd-5	213	11-09-90	13.5	370	8.5	41	15	16	4.0	103	82	10	.40	9.0	244 R
13	(C-1-2)22cbb-1	110	06-01-89	14.5	2,160	7.7	66	29	380	—	286 L	310	320	.50	57	1,340 R
14	(C-1-2)31aaa-1	215	08-31-83	16.5	1,480	7.4	67	56	160	15	257 L	230	210	.50	20	917
15	(C-2-1)11bad-1	273	07-05-83	16.5	518	7.8	50	17	32	2.0	109 L	75	61	.20	19	322
16	(C-2-1)12bac-1	157	09-15-82	13.0	250	7.9	27	9.2	8.5	1.6	82 L	23	5.4	.20	12	183 R
17	(C-2-2)33cad-1	216	09-18-84	14.0	1,110	7.3	85	32	130	4.4	303	140	130	.20	29	745
18	(C-2-2)1aabb-1	258	08-31-83	16.5	1,370	7.4	88	50	100	6.5	177 L	120	270	.50	44	811
19	(C-2-2)11adc-1	300	09-30-83	13.5	655	7.8	52	22	45	4.1	148 L	79	68	.30	34	408
20	(C-2-2)16aab-1	190	09-02-83	14.5	720	7.4	56	19	56	4.6	173 L	28	110	.30	53	442
21	(C-3-1)1dca-1	165	09-02-83	13.0	2,070	7.1	150	78	250	9.4	382 L	600	190	.80	21	1,540
22	(C-3-1)8cdd-2	265	09-25-84	13.5	4,660	8.1	610	440	130	8.5	790	2,500	160	1.6	33	4,380
23	(C-3-1)8cdd-3	294	09-25-84	13.5	4,280	6.7	550	370	120	7.7	748	2,200	150	.20	38	3,890
24	(C-3-1)8dda-1	280	09-21-84	14.5	2,920	6.7	380	210	110	5.7	561 L	1,200	180	<.10	34	2,470
25	(C-3-1)10acd-1	110	09-28-84	14.5	2,930	7.3	310	120	260	6.0	223	990	350	.10	26	2,220
26	(C-3-1)12ccb-1	118	07-21-92	20.0	950	7.6	63	31	81	8.3	191 L	110	120	.40	32	562
27	(C-3-1)14adc-1	116	09-20-84	14.5	910	7.7	48	26	110	6.7	171	100	150	.20	31	576
28	(C-3-1)16bcb-1	265	09-18-84	15.0	1,970	6.9	260	89	89	4.8	333	590	220	<.10	29	1,490
29	(C-3-1)17ddd-1	205	09-19-84	14.0	2,270	7.1	260	100	190	5.1	287	950	130	.10	27	1,850
30	(C-3-1)20cca-1	180	09-21-84	14.5	1,850	7.0	250	77	55	3.7	258	470	250	.10	26	1,300
31	(C-4-1)4aaa-4	104	09-03-91	17.5	1,780	7.4	78	24	230	6.5	317	190	200	.21	54	1,080 R
32	(C-4-1)23bac-1	260	06-02-89	15.0	740	8.0	53	29	57	—	172 L	90	79	.60	26	491 R
33	(C-4-1)23dbb-1	262	07-20-92	17.0	1,490	7.4	92	47	140	12	239 L	200	210	.60	26	876
34	(D-1-1)7abd-6	130	07-21-92	14.5	1,260	7.3	140	53	44	2.7	252 L	170	130	<.10	17	735
35	(D-1-1)19bac-4	105	07-05-83	13.0	930	7.2	110	35	45	3.4	254 L	180	46	.20	13	589
36	(D-1-1)20bcb-1	285	07-05-83	16.0	980	7.2	120	39	33	3.5	236 L	240	42	.30	14	647
37	(D-1-1)30bbc-9	285	08-18-83	13.5	850	7.5 L	110	37	28	2.8	238 L	200	31	.30	15	580
38	(D-2-1)5aba-2	247	09-26-84	13.0	568	7.7	67	27	16	2.1	166	100	27	.30	12	355
39	(D-3-1)12bdd-1	171	09-17-90	14.0	345	7.0	37	7.3	15	2.3	93	27	32	.10	11	187
40	(D-3-1)14dbc-1	154	09-01-83	11.0	385	6.8	52	11	12	2.7	157 L	27	17	.20	17	243
41	(D-3-1)31abb-1	138	08-16-91	16.0	455	7.6	42	19	17	9.2	211 L	.90	11	.40	37	265
Water samples from wells less than 101 feet deep																
1	(B-1-1)9adc-1	19.5	02-01-83	7.0	6,590	8.0	20	28	1,500	54	1,250 L	240	1,300	5.3	14	4,000
2	(B-1-1)26bad-1	18.5	02-07-83	9.0	2,530	7.0	390	80	110	77	424 L	980	120	5.3	50	2,070
3	(B-1-1)26cda-1	19.5	01-17-83	11.5	1,200	7.5	42	110	50	7.8	428 L	150	72	1.1	24	725
4	(B-1-1)35dcb-1	19.5	02-02-83	11.0	3,680	7.6	26	320	440	42	1,290 L	600	160	1.1	30	2,770
5	(B-1-2)19aca-2	86.0	08-05-91	22.0	3,550	8.3	3.2	4.5	830	12	874	130	500	4.6	22	2,030
6	(B-1-2)19aca-3	23.1	08-05-91	14.5	30,000	7.8	45	180	7,600	260	999	440	12,000	.71	32	20,900
7	(B-1-2)22dad-1	15.5	08-07-91	14.5	2,270	8.5	4.0	7.4	450	19	483	21	340	1.8	15	1,150
8	(B-1-2)28aab-1	15.0	08-06-91	16.5	3,580	8.6	4.9	5.7	770	46	490	240	650	2.2	27	2,040
9	(B-1-2)32bcb-3	24.7	08-06-91	15.0	11,200	8.0	24	28	2,800	100	354	60	4,100	1.5	25	7,310
10	(B-1-2)36baa-2	83.0	08-29-91	16.0	890	8.1	23	13	150	3.9	163	66	130	.85	25	512
11	(B-1-2)36baa-3	19.0	08-29-91	15.0	6,780	8.6	44	28	1,500	38	334	83	2,100	3.2	20	4,060
12	(C-1-1)2dca-2	19.6	01-07-83	12.5	2,320	6.9	470	50	60	8.1	373 L	990	67	.70	19	1,950

Table 1. Results of chemical analyses of water sampled by the U.S. Geological Survey from selected wells in Salt Lake Valley, Utah, 1982–92—Continued

Map no.	Location	Well depth (feet)	Date	Water temperature (°C)	Specific conductance (µS/cm)	pH, field (standard units)	Calcium dissolved (mg/L as Ca)	Magnesium dissolved (mg/L as Mg)	Sodium dissolved (mg/L as Na)	Potassium dissolved (mg/L as K)	Alkalinity, total (mg/L as CaCO ₃)	Sulfate dissolved (mg/L as SO ₄)	Chloride dissolved (mg/L as Cl)	Fluoride dissolved (mg/L as F)	Silica dissolved (mg/L as SiO ₂)	Solids dissolved (mg/L)
Water samples from wells less than 101 feet deep—Continued																
13	(C-1-1)11bac-1	19.7	01-07-83	10.5	10,900	7.1	310	660	1,600	200	948 L	3,900	1,700	0.60	31	8,970
14	(C-1-1)15cad-2	40.0	10-15-90	17.5	1,200	7.8	33	32	120	34	257	13	210	.53	42	643
15	(C-1-1)15cad-3	16.0	09-03-91	22.0	2,940	7.3	180	170	190	60	499	840	270	.91	47	2,060
16	(C-1-1)18bda-1	45.0	10-22-90	14.0	16,000	7.7	120	140	3,100	150	256	650	5,000	.46	26	9,290
17	(C-1-1)23bcd-2	66.1	11-08-90	13.0	720	8.1	22	27	71	32	343	8.0	21	.70	40	441
18	(C-1-1)26dba-4	38.0	06-01-83	13.0	19,200	6.8	430	1,300	3,200	170	967 L	7,900	3,200	.80	23	16,800
19	(C-1-1)26dba-6	60.0	05-06-83	12.0	1,080	—	100	54	75	26	444 L	96	42	.60	43	708
20	(C-1-1)26dba-7	72.0	05-06-83	12.0	521	—	44	24	29	19	297 L	3.8	10	.40	46	355
21	(C-1-1)26dca-1	19.7	02-02-83	8.0	4,230	7.4	170	250	430	81	439 L	950	650	1.2	31	3,050
22	(C-1-1)28cab-1	20.0	02-02-83	13.0	1,030	8.0	40	34	120	20	185 L	95	170	.70	24	615
23	(C-1-1)30aca-1	19.7	02-07-83	9.5	4,640	7.6	51	120	920	44	1,050 L	870	540	2.6	32	3,250
24	(C-1-1)31abb-1	14.3	02-07-83	10.5	1,990	7.4	90	86	230	37	416 L	320	260	.70	39	1,340
25	(C-1-1)33ddd-2	51.5	10-23-90	14.0	3,580	7.3	100	86	450	38	381	92	850	.36	44	1,890
26	(C-1-2)14cdd-2	49.5	08-19-91	20.0	1,690	8.0	31	21	270	26	176	130	310	.71	11	901
27	(C-1-2)14cdd-3	17.7	08-19-91	15.0	6,980	7.6	48	73	1,400	74	897	1,000	1,200	1.0	14	4,340
28	(C-1-2)22bdd-4	35.0	08-24-90	14.0	2,070	7.7	63	38	320	25	261 L	250	320	.30	57	1,230
29	(C-2-1)13abc-1	56.5	11-06-90	14.0	2,150	7.9	180	75	160	12	316	380	310	.35	31	1,330
30	(C-2-1)14bdb-1	11.0	02-03-83	9.0	2,310	6.3	180	110	180	15	103 L	910	210	3.1	27	1,740
31	(C-2-1)15abc-1	19.6	02-07-83	10.5	2,240	7.1	190	80	180	25	325 L	480	290	.90	32	1,500
32	(C-2-1)20aad-2	61.0	10-29-90	15.0	3,970	7.1	520	160	120	120	404	1,800	210	.40	52	3,230
33	(C-2-1)23ccb-1	41.0	10-23-90	16.0	1,850	7.4	110	57	170	18	311	230	240	.83	47	1,080
34	(C-2-1)23dac-1	85.0	10-29-90	13.5	1,490	7.4	90	47	120	5.0	315	210	160	.41	20	850
35	(C-2-1)34bbd-2	41.5	10-23-90	14.5	2,310	7.2	170	66	230	12	295	450	300	.34	32	1,450
36	(C-2-1)34dda-3	18.9	01-17-83	10.0	3,450	7.1	260	140	340	17	426 L	810	520	.50	33	2,440
37	(C-2-1)35baa-1	12.5	02-03-83	9.5	3,160	6.9	390	140	190	17	282 L	1,200	300	2.9	17	2,430
38	(C-2-1)35bab-1	19.7	01-17-83	12.5	2,500	7.0	170	87	250	12	332 L	450	370	.60	43	1,590
39	(C-2-1)36cdd-2	62.0	11-14-90	13.0	2,320	7.6	81	78	260	31	316	420	290	1.4	19	1,370
40	(C-2-2)1bcb-2	28.0	10-22-90	14.0	1,060	7.9	37	32	160	14	306	70	130	.76	50	701
41	(C-3-1)1bbcb-1	16.2	02-04-83	10.5	2,340	6.7	200	110	200	32	892 L	180	250	.90	39	1,550
42	(C-3-1)3acc-1	15.0	01-14-83	13.0	2,910	7.2	230	72	140	36	455 L	690	180	.30	30	1,770
43	(C-3-1)12cdb-1	41.0	11-13-90	12.5	2,310	7.3	100	73	260	10	464	330	300	.36	25	1,380
44	(C-3-1)14cba-1	100.0	09-26-84	13.5	3,760	7.4	460	150	260	5.3	133	750	860	<.10	23	2,630
45	(C-3-1)34aaa-1	19.4	01-17-83	12.5	3,470	7.1	420	98	280	33	316 L	1,300	340	.70	37	2,720
46	(C-4-1)4aaa-3	49.0	11-13-90	14.5	1,800	7.4	100	70	170	17	220	290	300	.46	36	1,110
47	(D-1-1)17dab-2	68.5	10-16-90	14.0	810	7.4	70	32	38	4.0	254	120	41	.29	20	482
48	(D-1-1)18dad-1	19.5	01-07-83	12.0	1,460	7.0	160	41	100	4.5	318 L	220	160	.20	14	925
49	(D-1-1)19dbc-2	31.0	10-15-90	14.0	1,430	7.2	120	58	79	9.0	451	260	100	.35	23	923
50	(D-1-1)30dcb-1	71.0	10-16-90	12.5	1,030	7.8	96	39	47	5.0	282	160	77	.21	20	614
51	(D-1-1)31abc-2	100.0	08-19-83	13.5	1,080	7.6	110	41	65	3.5	280	190	92	.20	19	690
52	(D-1-1)31dba-1	19.7	01-12-83	12.0	2,040	7.1	160	73	180	13	443 L	170	350	.60	22	1,240
53	(D-2-1)6dbb-12	85.0	07-08-83	12.0	600	7.5	74	24	25	2.0	174 L	100	41	.20	12	387
54	(D-2-1)8bba-1	11.3	01-12-83	11.5	2,120	7.0	210	96	140	9.9	405 L	550	210	.40	21	1,480
55	(D-2-1)9dca-1	100.0	07-10-84	16.0	670	7.6	81	21	21	2.3	151 L	79	83	.20	11	393
56	(D-2-1)16bba-3	90.0	11-06-90	12.5	610	7.9	50	25	23	4.0	180	52	48	.27	10	331
57	(D-2-1)17cda-2	28.5	10-30-90	15.0	1,180	7.4	77	36	110	7.0	277	140	130	.57	18	688
58	(D-2-1)17cda-3	23.5	10-30-90	15.5	1,080	7.2	77	34	110	6.0	255	140	130	.56	17	669
59	(D-2-1)29acb-2	52.0	10-30-90	13.5	680	7.7	77	17	31	5.0	181	38	69	.36	16	370
60	(D-2-1)30dda-2	65.0	08-30-91	14.0	740	7.8	59	20	39	4.9	130	87	67	.49	15	372
61	(D-2-1)30dda-3	21.4	08-30-91	14.5	1,340	7.6	110	28	110	5.8	189	80	200	.39	30	779
62	(D-3-1)6bcd-1	19.5	01-12-83	14.5	1,560	7.3	95	63	150	14	322 L	180	230	.80	27	972
63	(D-3-1)29ddd-1	48.0	11-13-90	15.5	960	7.6	89	30	44	5.0	269	130	80	.30	39	580
64	(D-3-1)31cda-1	19.7	01-12-83	10.5	2,830	7.3	140	130	270	25	266 L	530	420	1.3	49	1,820
65	(D-3-1)32aaa-1	19.7	01-12-83	11.0	760	6.8	62	22	66	3.4	190 L	100	83	<.10	13	477

Table 2. Records of selected wells that were sampled for chemical analysis of water, Salt Lake Valley, Utah, 1982–92

[—, no data available]

Location: See figure 2 for an explanation of the numbering system used for hydrologic-data sites in Utah.

Use of water: P, public supply; I, irrigation; U, unused; S, stock; H, domestic or household; N, industrial.

Altitude of land surface: In feet above sea level. Surveyed altitudes are in feet and decimal fractions; altitudes interpolated from U.S.

Geological Survey topographic maps are to the nearest foot.

Depth of well: In feet below land surface.

Casing: Diameter: In inches, reported from drillers' logs or measured in the field.

Finish: P, perforated or slotted; O, open end; S, screened; X, open hole extending below the bottom of open-ended casing. Upper and lower limits of perforations, screens, or open hole are in feet below land surface, if known, and queried (?) if extent of perforated or screened interval is unknown.

Location	Date well completed	Use of water	Altitude of land surface (feet)	Depth of well (feet)	Casing diameter (inches)	Finish (feet)
(A-1-1)31cac-1	1943	P	4,400.86	464	20	P 162-216, 231-280, 312-317, 380-420
(B-1-1)9adb-1	08-06-70	I	4,210	255	2	P 251-255
(B-1-1)9adc-1	08-25-82	U	4,210	19.5	2	P 7.5-19.5
(B-1-1)19baa-5	1900	S	4,219.80	645	2	O
(B-1-1)26bad-1	08-31-82	U	4,216.39	18.5	2	P 6.5-18.5
(B-1-1)26cda-1	08-19-82	U	4,217.69	19.5	2	P 7.5-19.5
(B-1-1)35dcb-1	08-25-82	U	4,221.40	19.5	2	P 7.5-19.5
(B-1-2)2dac-1	10-01-40	S	4,212	440	—	—
(B-1-2)7caa-1	—	S	4,210.89	735	2	O 480
(B-1-2)7ccc-1	1959	S	4,210.10	389	2	P 379-?
(B-1-2)8abd-1	1920	S	4,212.15	300	2.5	O
(B-1-2)15bcd-2	1920	S	4,212.86	300	2	O
(B-1-2)16caa-1	06-01-66	S	4,213.98	636	2	S 626-636
(B-1-2)19aca-1	1915	S	4,211.62	450	2	O
(B-1-2)19aca-2	03-14-91	U	4,210.1	86.0	2	P 75-85
(B-1-2)19aca-3	05-17-91	U	4,210.4	23.1	2	P 17.1-22.1
(B-1-2)21acd-1	04-01-66	S	4,219.76	600	—	S 590-600
(B-1-2)21cab-2	1963	I	4,219.70	330	4	P 40-330
(B-1-2)21dcd-1	03-31-66	S	4,222	561	2	S 551-561
(B-1-2)22daa-1	08-09-56	S	4,215.22	148	2	P 145-148
(B-1-2)22dad-1	05-22-91	U	4,215.6	15.5	2	P 9.5-14.5
(B-1-2)23bbd-1	—	S	4,218	560	2	—
(B-1-2)23cbd-1	08-13-56	S	4,219	270	2	P 165-?
(B-1-2)28aab-1	05-24-91	U	4,220	15.0	2	P 9.0-14.0
(B-1-2)32bcb-1	10-07-43	U	4,214	716	3	—
(B-1-2)32bcb-2	07-17-90	U	4,215	248	3	X 240-248
(B-1-2)32bcb-3	05-22-91	U	4,215	24.7	2	P 18.7-23.7
(B-1-2)36baa-1	1915	S	4,223.60	464	2	X
(B-1-2)36baa-2	03-13-91	U	4,223.6	83.0	2	P 72-82
(B-1-2)36baa-3	05-14-91	U	4,223.8	19.0	2	P 13-18
(C-1-1)2dca-2	07-21-82	U	4,222.18	19.6	2	P 7.6-19.6
(C-1-1)8aac-3	04-12-49	—	4,230	252	2	P 249-252
(C-1-1)9bba-3	03-20-54	—	4,230	341	2	P 339-341
(C-1-1)11bac-1	07-26-82	U	4,232	19.7	2	P 7.7-19.7
(C-1-1)15bdd-1	09-25-40	U	4,231.11	445	2	—
(C-1-1)15cad-2	06-27-90	U	4,232.6	40.0	2	P 29-39
(C-1-1)15cad-3	05-08-91	U	4,232.4	16.0	2	P 10-15
(C-1-1)18bda-1	06-28-90	U	4,237	45.0	2	P 34-44

Table 2. Records of selected wells that were sampled for chemical analysis of water, Salt Lake Valley, Utah, 1982–92—Continued

Location	Date well completed	Use of water	Altitude of land surface (feet)	Depth of well (feet)	Casing diameter (inches)	Finish (feet)
(C-1-1)23bcb-1	02-25-83	P	4,227	840	30, 18, 16	S 145-160, 166-181, 220-240, 325-340, 375-390, 401-406, 498-508, 522-562, 620-630, 662-672, 760-804
(C-1-1)23bcd-2	06-08-90	U	4,230	66.1	2	P 55.1-65.1
(C-1-1)23caa-1	10-28-77	S	4,230	175	2	P 168-175
(C-1-1)25bdb-1	07-14-67	H	4,236	1,000	—	P 214-220, 440-445, 550-560, 588-592, 642-658, 746-778, 790-800, 980-1,000
(C-1-1)25bdc-1	1983	U	4,238.47	135	2	S 119-129
(C-1-1)25cca-2	07-01-55	H	4,238	345	2	O
(C-1-1)26caa-1	1983	U	4,234.37	115	2	S 105-115
(C-1-1)26dba-4	05-21-83	U	4,233.27	38.0	4.5	S 18.0-38.0
(C-1-1)26dba-5	1983	U	4,233.73	129	10, 6	S 102-112
(C-1-1)26dba-6	05-01-83	U	4,233.40	60.0	2	S 54.0-56.0
(C-1-1)26dba-7	05-01-83	U	4,234.47	72.0	2	S 66.0-68.0
(C-1-1)26dca-1	08-26-82	U	4,239.75	19.7	2	P 7.7-19.7
(C-1-1)27dac-4	01-01-54	—	4,250	308	—	O
(C-1-1)27dda-8	04-01-58	P	4,237	775	—	P 670-675, 690-702, 745-760
(C-1-1)28cab-1	07-08-82	U	4,244.77	20.0	2	P 7.7-19.7
(C-1-1)30aca-1	08-12-82	U	4,249.19	19.7	2	P 7.7-19.7
(C-1-1)31abb-1	09-02-82	U	4,275.43	14.3	2	P 2.3-14.3
(C-1-1)33ddd-1	08-15-60	P	4,273	886	16, 12	P 638-645, 730-740, 746-755, 805-812, 840-847, 875-886
(C-1-1)33ddd-2	06-08-90	U	4,270	51.5	2	P 40.5-50.5
(C-1-1)35ddd-5	07-17-90	U	4,240	213	3	X 200-213
(C-1-2)2aba-2	03-25-66	S	4,230	410	2	S 399-410
(C-1-2)6aaa-1	1949	N	4,212.45	760	4	O
(C-1-2)6aaa-3	07-10-56	N	4,212.57	825	6	P 757-825
(C-1-2)6aaa-5	—	N	4,210.97	345	2	
(C-1-2)14cdd-1	01-04-80	H	4,233	480	8, 6	S 380-480
(C-1-2)14cdd-2	03-16-91	U	4,233	49.5	2	P 38.5-48.5
(C-1-2)14cdd-3	05-15-91	U	4,233	17.7	2	P 11.7-16.7
(C-1-2)22bdd-4	1885	U	4,233.26	35.0	1.25	—
(C-1-2)22cbb-1	—	S	4,231.61	110	2	—
(C-1-2)31aaa-1	07-26-76	H	4,443	215	6, 5	P 170-215
(C-2-1)11bad-1	1952	I	4,275	273	—	O
(C-2-1)12bac-1	08-01-31	U	4,255.86	157	4	O
(C-2-1)13abc-1	06-13-90	U	4,315	56.5	2	P 45.5-55.5
(C-2-1)14bdb-1	09-03-82	U	4,263.85	11.0	2	S 7.0-11.0
(C-2-1)15abc-1	09-16-82	U	4,331.36	19.6	2	P 7.6-19.6
(C-2-1)20aad-1	05-06-79	P	4,515	658	20, 16	P 240-355
(C-2-1)20aad-2	06-09-90	U	4,514	61.0	2	P 50-60
(C-2-1)23ccb-1	06-19-90	U	4,350	41.0	2	P 30-40
(C-2-1)23dac-1	06-18-90	U	4,280	85.0	2	P 74-84
(C-2-1)24bcd-1	08-06-65	P	4,307	370	20, 16	P 179-300
(C-2-1)33cad-1	11-09-76	I	4,453	216	6	P 206-216
(C-2-1)34bbd-2	06-11-90	U	4,384	41.5	2	P 30.5-40.5
(C-2-1)34dda-3	09-01-82	U	4,347.07	18.9	2	P 6.9-18.9
(C-2-1)35baa-1	09-21-82	U	4,288.98	12.5	2	P 0.5-12.5
(C-2-1)35bab-1	08-27-82	U	4,304.41	19.7	2	P 7.7-19.7
(C-2-1)36cdd-2	07-25-90	U	4,359	62.0	2	P 51-61

Table 2. Records of selected wells that were sampled for chemical analysis of water, Salt Lake Valley, Utah, 1982–92—Continued

Location	Date well	Use of water	Altitude of land surface (feet)	Depth of well (feet)	Casing diameter (inches)		Finish (feet)
(C-2-2)1aab-1	04-06-78	H	4,413.38	258	6	O	
(C-2-2)1bcd-2	06-29-90	U	4,496	28.0	2	P	17-27
(C-2-2)11adc-1	07-31-72	N	4,720	300	8	P	265-?
(C-2-2)16aab-1	02-11-77	H	5,150	190	8, 6	P	145-165
(C-3-1)1bbc-1	07-22-82	U	4,298.03	16.2	2	P	4.2-16.2
(C-3-1)1dca-1	01-10-79	I	4,388	165	12	P	125-135
(C-3-1)3acc-1	08-03-82	U	4,391.28	15.0	2	P	3.0-15.0
(C-3-1)7cbb-1	1967	I	4,800	462	16	P	215-460
(C-3-1)8cdd-2	1984	U	4,650	265	0.5	O	
(C-3-1)8cdd-3	1984	U	4,650	294	0.5	O	
(C-3-1)8dda-1	07-06-78	H	4,587	280	6	P	200-280
(C-3-1)10acd-1	1972	I	4,425	110	6	O	
(C-3-1)12ccb-1	1900	S	4,321.54	118	3	O	
(C-3-1)12cdb-1	06-12-90	U	4,358	41.0	2	P	30-40
(C-3-1)14adc-1	1969	H	4,350	116	4	P	96-116
(C-3-1)14bbc-1	1977	I	4,435	395	6	P	270-395
(C-3-1)14cba-1	1977	H	4,397	100	6	O	
(C-3-1)15dda-1	—	U	4,394.03	360	—	—	
(C-3-1)16bcb-1	1978	I	4,607	265	8	P	250-265
(C-3-1)17ddd-1	01-08-76	I	4,617	205	6, 5	X	178-205
(C-3-1)20cca-1	04-19-77	I	4,618	180	6	O	
(C-3-1)34aaa-1	08-05-82	U	4,410.76	19.4	2	P	7.4-19.4
(C-3-2)12dbc-2	09-01-64	N	4,862	360	4	P	200-280
(C-3-2)33cac-1	1961	I	5,262	350	12	P	80-350
(C-4-1)4aaa-3	06-20-90	U	4,478	49.0	2	P	38-48
(C-4-1)4aaa-4	03-09-91	U	4,480	104	2	P	93-103
(C-4-1)23bac-1	01-26-68	H	4,590	260	12	P	160-220, 244-254
(C-4-1)23dbb-1	1959	N	4,580	262	16	P	60-134, 152-182, 199-205
(C-4-2)1bbb-1	03-01-65	I	4,890	540	16	P	220-524
(D-1-1)7abd-6	1890	P	4,260.89	130	2	O	
(D-1-1)17dab-2	07-21-90	U	4,390	68.5	2	P	57.5-67.5
(D-1-1)18dad-1	08-20-82	U	4,288.11	19.5	2	P	7.5-19.5
(D-1-1)19bac-4	1915	H	4,260	105	—	O	
(D-1-1)19cdb-17	1922	H	4,251.50	500	2	O	
(D-1-1)19dbc-2	06-25-90	U	4,262	31.0	2	P	20-30
(D-1-1)20bab-1	1958	N	4,328	482	—	P	468-475
(D-1-1)20cbc-1	1935	H	4,291	285	—	O	
(D-1-1)21acc-2	06-01-62	P	4,445	576	20	P	175-335, 360-430, 450-560
(D-1-1)30bbc-9	1931	I	4,249	285	—	—	
(D-1-1)30dcb-1	06-26-90	U	4,262	71.0	2	P	60-70
(D-1-1)31abc-2	1938	I	4,270	100	2	—	
(D-1-1)31dba-1	08-10-82	U	4,277.26	19.7	2	P	7.7-19.7
(D-2-1)5aba-2	04-01-50	H	4,310	247	2	—	
(D-2-1)6dbb-12	1898	U	4,268.90	85.0	4	O	
(D-2-1)8bba-1	08-05-82	U	4,285.04	11.3	2	P	0-11.3
(D-2-1)9dca-1	09-01-79	H	4,363.24	100	10, 6	P	80-100
(D-2-1)16bba-3	07-22-90	U	4,351	90.0	2	P	79-89
(D-2-1)17cda-2	06-14-90	U	4,360	28.5	2	P	17.5-27.5
(D-2-1)17cda-3	06-14-90	U	4,360	23.5	2	P	12.5-22.5

Table 2. Records of selected wells that were sampled for chemical analysis of water, Salt Lake Valley, Utah, 1982–92—Continued

Location	Date well completed	Use of water	Altitude of land surface (feet)	Depth of well (feet)	Casing diameter (inches)	Finish (feet)
(D-2-1)21dbc-1	09-29-61	P	4,453	740	24, 20	P 210-244, 256-262, 312-344, 363-386, 444-584, 631-728
(D-2-1)28ccc-1	08-28-56	P	4,580	691	16	P 515-530, 535-585, 592-678
(D-2-1)29acb-2	06-23-90	U	4,467	52.0	2	P 41-51
(D-2-1)29acb-8	10-07-58	P	4,465	500	12	P 455-?
(D-2-1)30dda-1	09-22-60	P	4,453	1,002	20	P 560-580, 608-617, 620-630, 648-661, 687-689, 697-728, 734-742, 851-872, 890-934, 949-956, 963-965, 971-972, 978-981, 986-990
(D-2-1)30dda-2	03-12-91	U	4,453	65.0	2	P 54-64
(D-2-1)30dda-3	05-16-91	U	4,453	21.4	2	P 15.4-20.4
(D-2-1)34cdc-1	05-27-80	P	4,710	875	24, 20	P 435-461, 595-650, 740-762
(D-3-1)4cad-1	02-11-78	P	4,720	938	24, 20	P 500-518, 660-687
(D-3-1)6bcd-1	08-05-82	U	4,426.91	19.5	2	P 7.50-19.5
(D-3-1)6dad-1	09-08-79	P	4,515	1,000	24, 20, 16	P 473-535, 552-620, 650-1,000
(D-3-1)12bdd-1	02-17-62	P	5,280	171	20, 16	P 65-168
(D-3-1)14dbc-1	08-01-78	H	4,960	154	8, 6	P 125-140
(D-3-1)29ddd-1	06-21-90	U	4,524	48.0	2	P 33-43
(D-3-1)31abb-1	10-01-65	H	4,237	138	6	P 125-135
(D-3-1)31cda-1	07-26-82	U	4,437.91	19.7	2	P 7.7-19.7
(D-3-1)32aaa-1	08-18-82	U	4,510.33	19.7	2	P 7.7-19.7

the principal aquifer without distinction to depth of sample. These maps were modified with data collected from 1988 to 1992, much of which was obtained from Kennecott Utah Copper (1992) and the Utah Department of Environmental Quality, Division of Drinking Water (written commun., 1993) (pl. 2). Dissolved-solids concentration in water sampled from 1988 to 1992 from wells finished in the principal aquifer ranged from 110 mg/L on the east side of the valley to 48,100 mg/L on the west side.

Water from the principal aquifer is divided into somewhat distinct groups on the basis of location in the valley and chemical composition (figs. 4 and 5). The data set consisted of 44 chemical analyses of water collected between 1983 and 1992 from wells greater than 300 ft deep (table 1).

Predominant ions in water from the southeastern part of the valley are calcium, magnesium, and bicarbonate. The predominant cations present in the water are listed in this report in order of decreasing abundance, followed similarly by predominant anions. The chemical composition of ground water in the southeast-

ern part of Salt Lake Valley is similar to that of water in streams that drain the Wasatch Range (figs. 6 and 7).

Ground water sampled from the southeastern part of the valley by the U.S. Geological Survey contained dissolved-solids concentrations ranging from 130 to 217 mg/L (table 1). Water from the principal aquifer is only slightly more mineralized than water from Big and Little Cottonwood Creeks because the recharged water cannot dissolve the relatively insoluble minerals in the consolidated rocks of the Wasatch Range to the east or in the basin-fill material derived from these rocks. These rocks consist mainly of resistant quartzite of Precambrian age and quartz monozonite of Tertiary age (Davis, 1983, sheet 1). Ground water in the principal aquifer with a dissolved-solids concentration less than 500 mg/L extends from the southeastern part of the valley northwest to beyond the Jordan River (pl. 2). Most of this water originated as precipitation on the Wasatch Range. Mixing with other sources of water and reactions with the basin-fill material along the flow path has resulted in the increase in dissolved-solids concentrations relative to ground water in the southeastern part of the valley.

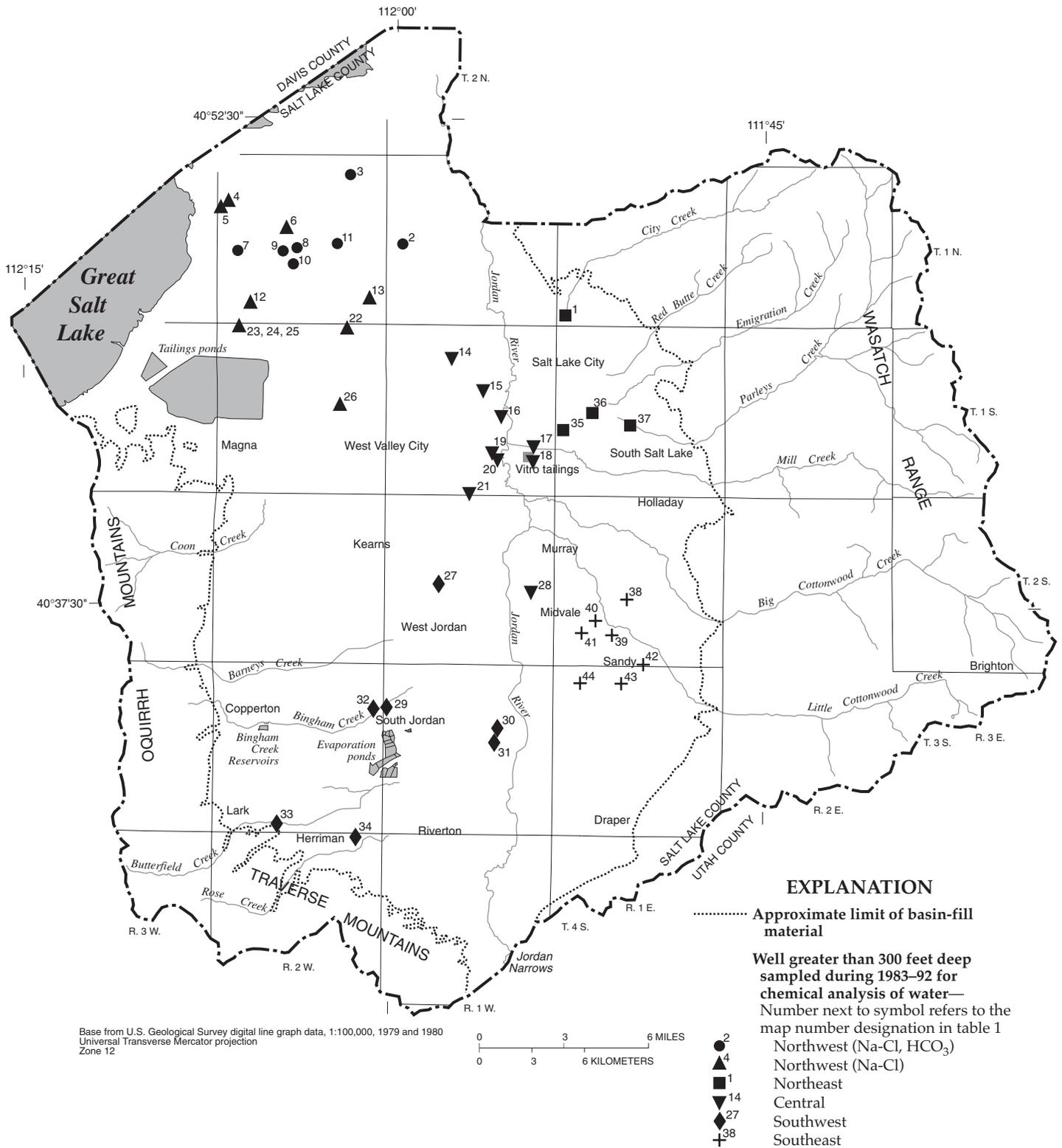


Figure 4. Location of wells greater than 300 feet deep in Salt Lake Valley, Utah, sampled for chemical analysis of water, 1983–92.

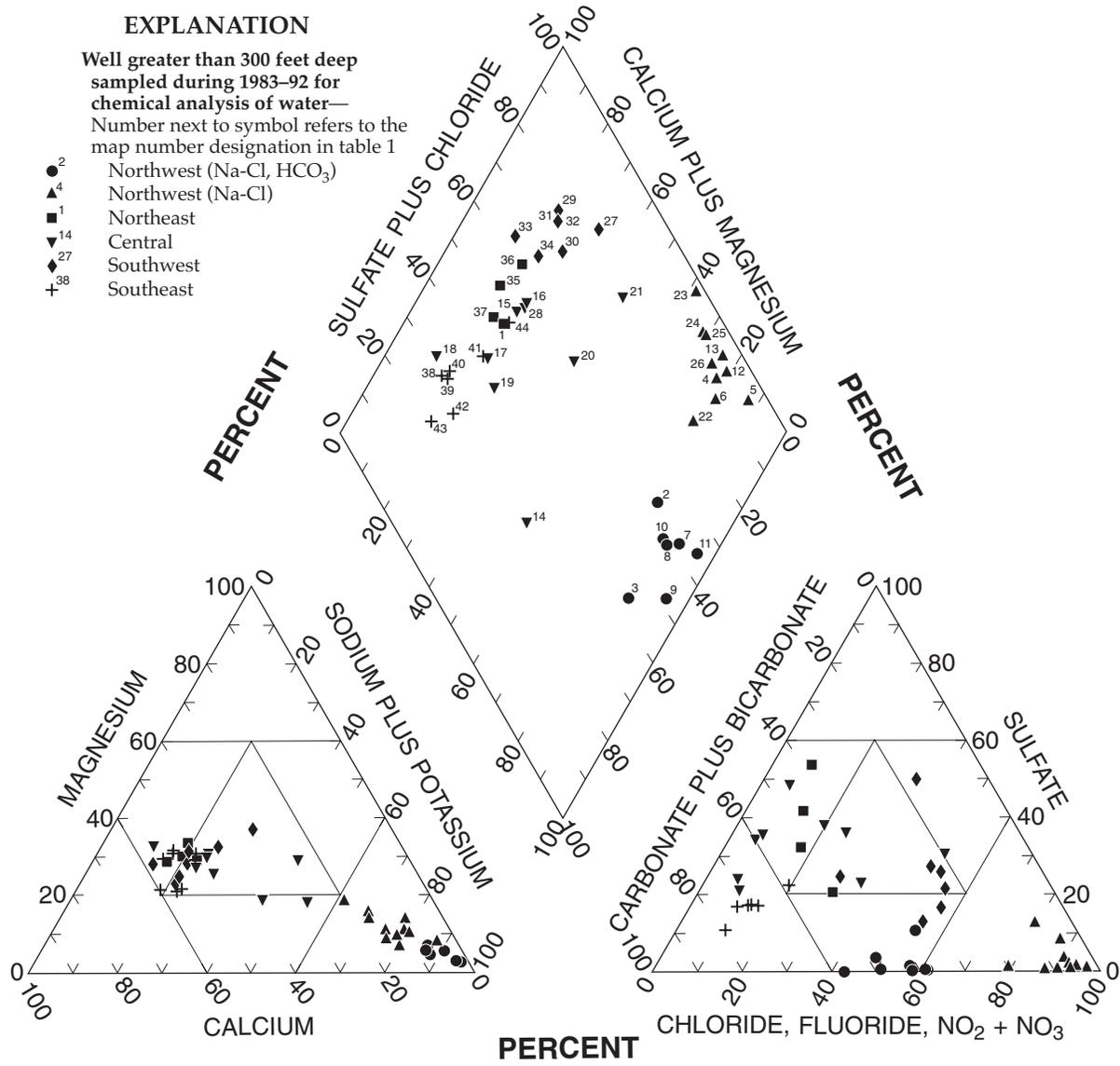


Figure 5. Chemical composition of water from wells greater than 300 feet deep in Salt Lake Valley, Utah.

Water from well (D-3-1)6dad-1 (map no. 44 in table 1) contained more chloride and less sulfate than other ground water in the area. Perforated intervals in the well extend to 1,000 ft below land surface (table 2) and the water temperature (17 °C) (table 1) is warmer than that of water from other area wells. The differences in chemical composition between water from well (D-3-1)6dad-1 and from other wells in the southeastern part of the valley indicate another source of water to well (D-3-1)6dad-1 or other possible water-rock interactions occurring at depth.

Information from well owners and Klauk (1984, p. 118) indicate that warmer water (warmer than 17 °C)

occurs in other wells deeper than about 600 ft in an area east of the Jordan River from about Murray to Draper. Water from these wells generally has a higher dissolved-solids concentration than water from wells with ambient ground-water temperatures (11.5 to 14 °C). More information is needed on the source of the warm water in this area because sustained or increased ground-water withdrawals could affect water chemistry by inducing vertical flow.

Much of the ground water in the northeastern part of the valley originates from infiltrating precipitation in the City Creek, Red Butte Creek, Emigration Creek, and Parleys Creek watersheds. Dissolved-solids con-

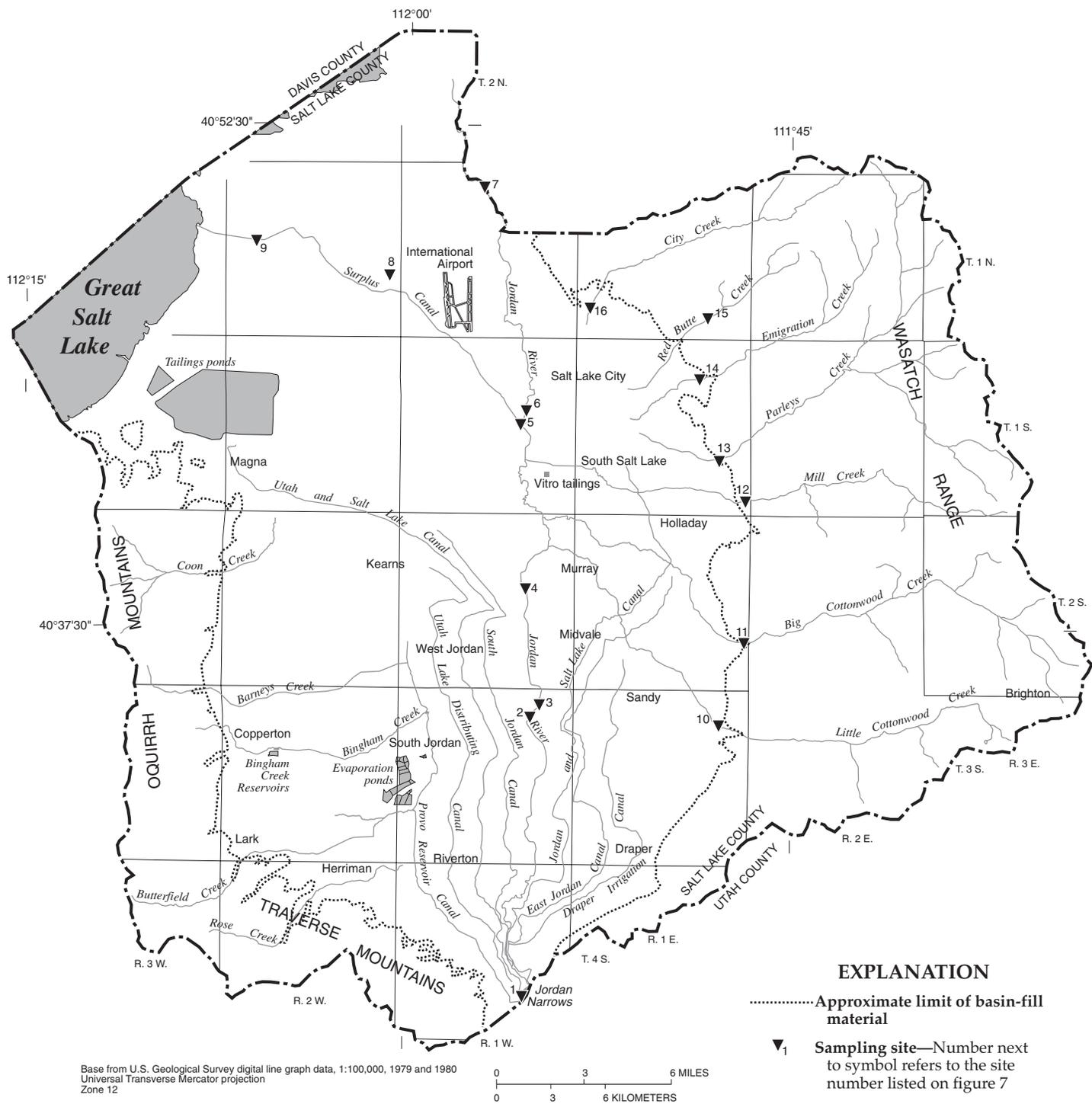


Figure 6. Location of surface-water sampling sites in Salt Lake Valley, Utah.

EXPLANATION

Site
Dissolved-solids concentration, in milligrams per liter

Jordan River at:

1	Jordan Narrows	1,120
2	Above 9400 South Street	1,620
3	9000 South Street	1,330
4	5800 South Street	1,370
5	2100 South Street	1,180
6	1700 South Street	1,050
7	Cudahy Lane	1,080

Other sites:

8	Surplus Canal at Cohen Flume	1,180
9	Goggin Drain near Magna	1,740
10	Little Cottonwood Creek	120
11	Big Cottonwood Creek	187
12	Mill Creek	380
13	Parleys Creek	401
14	Emigration Creek	464
15	Red Butte Creek	393
16	City Creek	272

Data points represent the discharge-weighted average chemical composition of water samples collected mainly during the 1964–68 water years (Hely and others, 1971, tables 8 and 13). Samples from the Jordan River above 9400 South Street, at 9000 South Street, and at 5800 South Street were collected during the 1966–68 water years. The Jordan River at 1700 South Street was sampled on January 30, 1990.

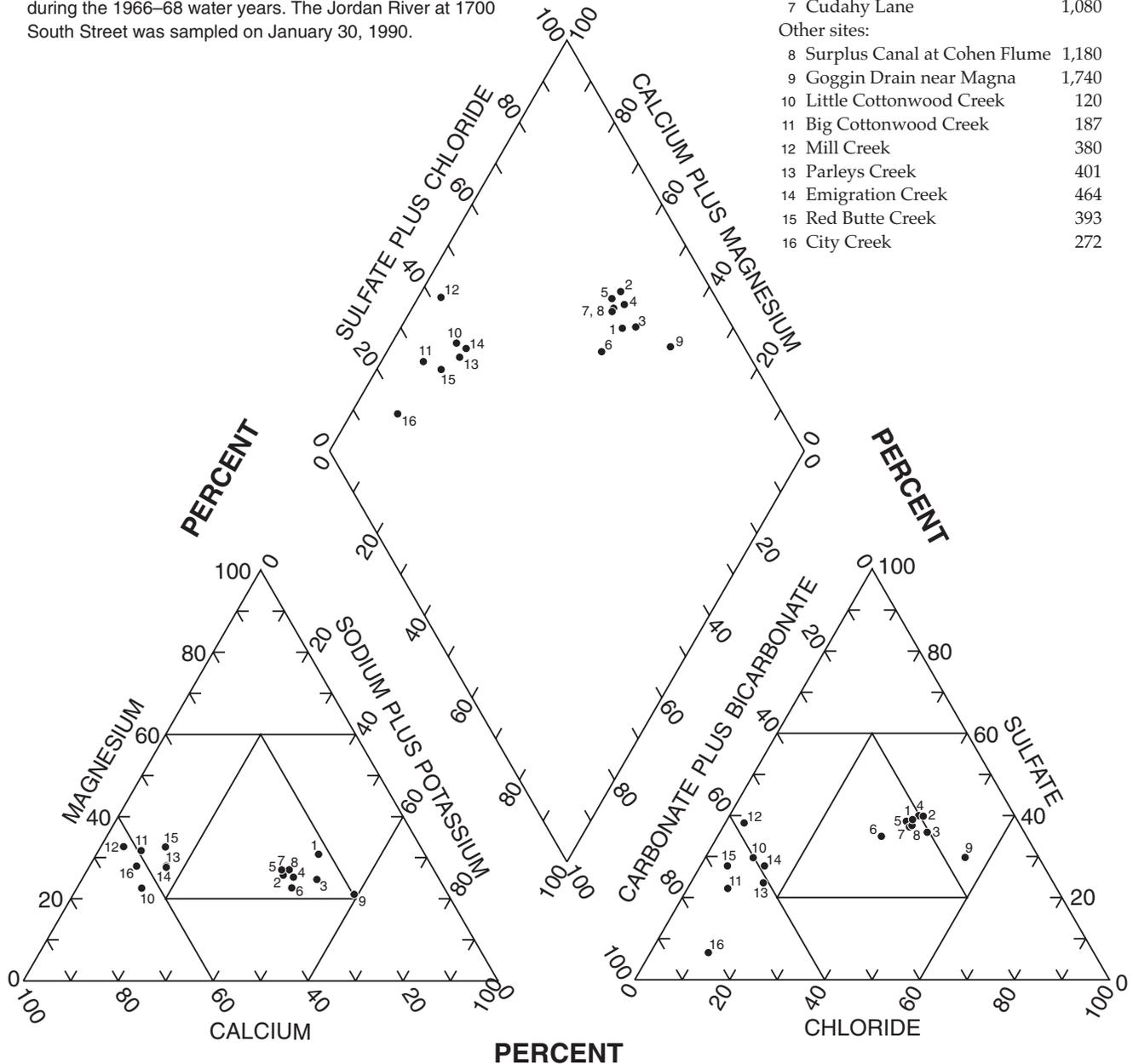


Figure 7. Chemical composition of surface water in Salt Lake Valley, Utah.

centration in ground water from this area is greater than 500 mg/L, mainly because rocks in this part of the Wasatch Range include less resistant shale strata of the Triassic and Jurassic Periods. Generally, sulfate is a predominant anion in this water (fig. 5). The presence of sulfate probably is caused by dissolution of gypsum and other sulfate-bearing minerals that are present in the shale and limestone strata in the recharge area to the east and in the unconsolidated basin-fill material derived from these rocks. Ground water in the north-eastern part of the valley has more dissolved sulfate relative to bicarbonate (fig. 5) than does water in upgradient mountain-front streams (fig. 7). Longer contact time between the basin-fill material and the water allows for increased dissolution of minerals.

Reactions between the basin-fill material derived from the Oquirrh Mountains and the different types of water being recharged result in a wide range of dissolved-solids concentrations in the southwestern part of Salt Lake Valley (pl. 2). Much of the dissolved-solids concentration data for the southwestern part of the valley was collected by Kennecott Utah Copper (1992) from wells with finish intervals in the principal aquifer.

Ground water sampled from seven wells deeper than 300 ft in the southwestern part of the valley contains primarily calcium, magnesium, chloride, bicarbonate, and sulfate ions. The Oquirrh Mountains to the west are a recharge area for the basin-fill material in this area and include carbonate rocks that have undergone sulfide mineralization. Mine drainage and wastewater from ore-leaching operations have been stored in two unlined reservoirs near the mouth of Bingham Canyon and in a series of unlined evaporation ponds on the basin fill about 5 mi to the east. The ore-leaching process used acidic mine water to dissolve and recover minerals from mine tailings. The water stored in these surface-water bodies remained acidic and contained high concentrations of dissolved-solids and sulfate resulting from the oxidation of the metal sulfides (Wadell and others, 1987, p. 19). The distribution of water levels in wells near the reservoirs and evaporation ponds indicates that infiltration down into the basin-fill material from these impoundments has occurred (Wadell and others, 1987, pl. 3). As of 1992, the reservoirs had been lined and the ponds were not being used.

Dissolution caused by undersaturation, precipitation caused by supersaturation, or ion exchange can be determined from an activity diagram (fig. 8). The activity of an ion is an idealized concentration that is the product of the measured concentration (moles/liter) and

a conversion factor called the activity coefficient. The state of saturation with respect to aquifer minerals can be used to evaluate chemical controls on a ground-water system. Water sampled from wells deeper than 300 feet in the valley generally plot along or near the equilibrium line for calcite, indicating saturation with respect to calcite. Exceptions include water from wells (D-2-1)30dda-1 and 34cdc-1 (map nos. 41 and 42 in table 1) in the southeastern part of the valley, which are slightly undersaturated, and water from the northwestern part of the valley, where the ground water is supersaturated with calcite.

The relation between log calcium and log carbonate activity in water from the southwestern part of Salt Lake Valley is similar to that of water that contains sulfate in the northeastern part (fig. 8) but is influenced by oxidation of sulfide minerals and the reaction between acidic water and carbonate minerals in the basin-fill material. The grouping of ground water on the basis of location in the valley and the relation of log calcium and log carbonate activity indicates trends in water evolution as the water moves through the flow system. This relation between log calcium and log carbonate activity may be useful in identifying ground-water movement in areas of changing chemical composition.

Ground water from the northwestern part of Salt Lake Valley is clustered into two water types (fig. 5). The sodium-chloride type water contained from 1,570 to 4,930 mg/L of dissolved solids and the sodium-chloride and bicarbonate type water contained from 498 to 1,260 mg/L (table 1). Temperature of the water samples ranged from 13.5 to 27.0 °C. The formation of the two water types in this part of the valley may be dependent on recharge sources. The sodium-chloride type water is generally located west and south of the area containing the sodium-chloride and bicarbonate type water (fig. 4).

The relation between log calcium and log carbonate activity for the sodium-chloride type water is controlled by the solubility of calcite and temperature at depth. Calcite dissolves to produce calcium and carbonate ions ($\text{CaCO}_3 = \text{Ca}^{2+} + \text{CO}_3^{2-}$). Bicarbonate is produced ($\text{CO}_3^{2-} + \text{H}^+ = \text{HCO}_3^-$) at the pH value of these waters. Ion exchange of calcium for sodium and mineral dissolution provide the sodium and chloride ions. Mixing of ground water in the principal aquifer with pore water left from desiccated prehistoric Lake Bonneville that covered the area also may provide sodium and chloride ions.

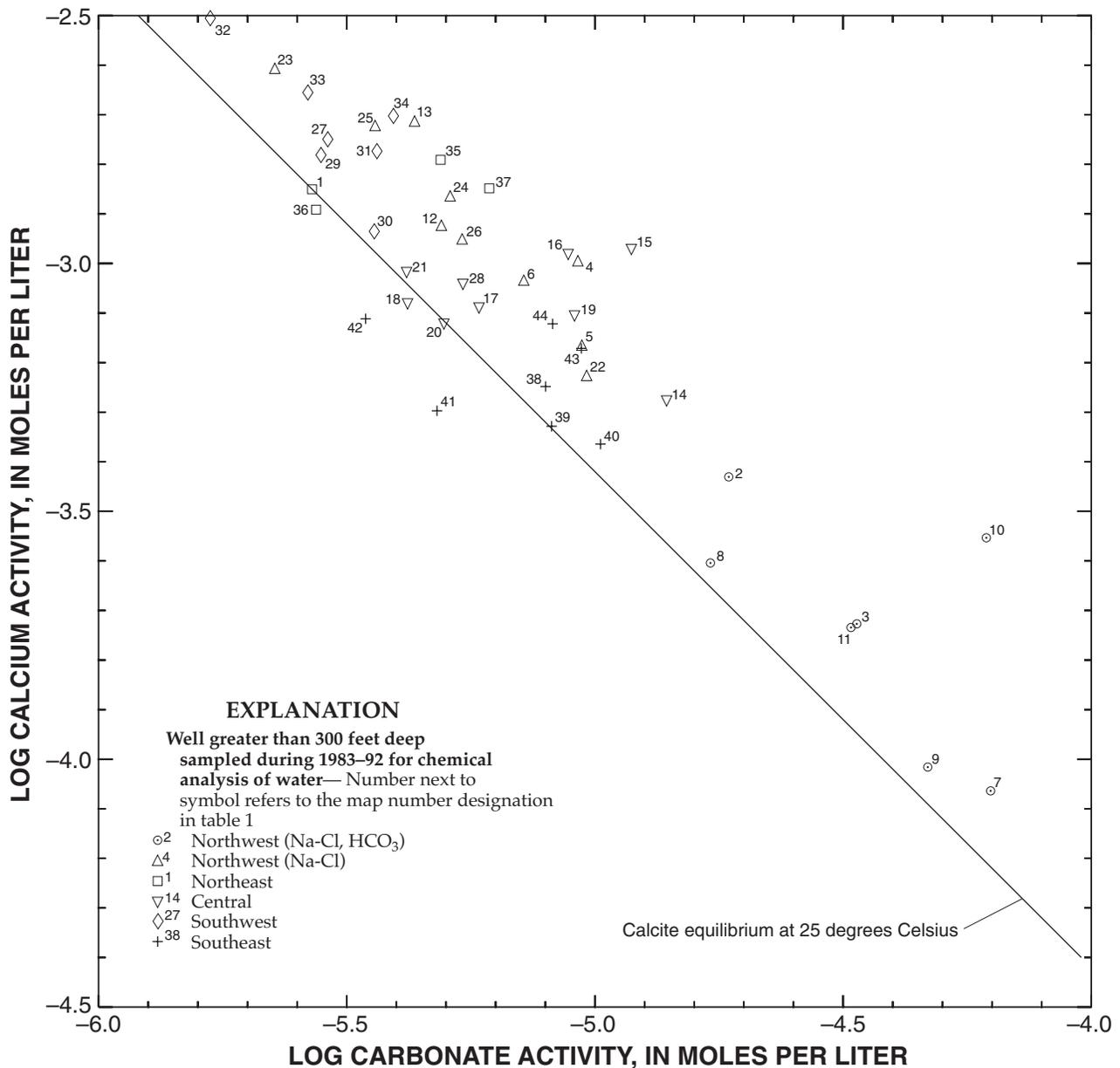


Figure 8. Relation between log calcium and log carbonate activity for water sampled from wells greater than 300 feet deep in Salt Lake Valley, Utah.

In the sodium-chloride and bicarbonate type water, calcium concentration (activity) tends to decrease as carbonate concentration (activity) increases (fig. 8). Some of the calcium and bicarbonate ions are likely the result of sulfate reduction. Sulfate reduction is a process by which bacteria in the aquifer use the oxygen in sulfate (SO_4^{2-}) to oxidize organic matter to carbonate (CO_3^{2-}) and bicarbonate (HCO_3^-). Sulfide species (HS or H_2S , depending on the pH value of the water) form as a byproduct. The reduction process would produce calcium ions if gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$) were the source of sulfate in the aquifer. The calcium

ions (Ca^{2+}) would react with bicarbonate to form calcite (CaCO_3) in a 1:1 ratio.

Sulfate reduction is probably occurring in the northern part of the valley. Very low concentrations of sulfate are present in the sodium-chloride and bicarbonate type waters. Less sulfate is dissolved in this water than in less mineralized water that occurs upgradient along the regional flowpath. The ground-water system in this area likely receives only a small amount of atmospheric oxygen because of the upward hydraulic-head gradient and the confined conditions, resulting in an anaerobic system. Organic material probably is avail-

able in the lake-deposited clays. Hely and others (1971, p. 169) report that hydrogen sulfide is characteristic of ground water sampled in the northern part of the valley.

Water from Intermediate Depths

Water from wells that range in depth from 101 to 300 ft below land surface in Salt Lake Valley was considered to be representative of water from intermediate depths. This range in depth generally includes the upper part of the principal aquifer and much of the fine-grained sediments that divide the shallow unconfined aquifer from the principal aquifer in the valley. Chemical analyses were available for water from 41 wells sampled between 1982 and 1992 (fig. 9 and table 1). The composition of ground water from this depth interval (fig. 10) is similar to water sampled by the U.S. Geological Survey from wells greater than 300 ft deep (fig. 5). Water from intermediate depths in the west and central parts of the valley has more variation in water chemistry than does water from deeper intervals.

Two types of ground water are present on the west side of Salt Lake Valley in the 101 to 300-foot depth interval. Most of the ground water in the area has a sodium, calcium-chloride, bicarbonate composition, similar to water from the Jordan River. Water with a dissolved-solids concentration of less than 500 mg/L was collected by Kennecott Utah Copper (Kennecott Utah Copper, 1992) during 1988–92 from wells completed from 101 to 300 ft below land surface in an area downgradient from a section of the Provo Reservoir Canal (pl. 2). Prior to 1990, surface water with similar dissolved-solids concentrations was transported in this canal (see “Historical changes in chemical composition of ground water” section of this report). Seepage from canals and infiltration of unconsumed irrigation water diverted from the river are large components of recharge to the basin-fill material in the west-side area.

The other water type was sampled from the area downgradient from Bingham Canyon. The predominant ions in water from eight wells in this area were calcium, magnesium, and sulfate. The high dissolved-solids concentration in ground water downgradient from Bingham Canyon is a result of mining operations and the presence of mineralized zones in the area (Waddell and others, 1987, p. 16–19). Water from wells in the southwestern part of the valley with finish intervals from 101 to 300 ft below land surface generally had a higher dissolved-solids concentration than did water from greater depths, except for a small area just to the

east of the mouth of Bingham Canyon (pl. 2). In the finish interval from 101 to 300 ft below land surface, water with a dissolved-solids concentration higher than 2,000 mg/L extends almost 9 mi to the east, near the Jordan River. In the finish interval greater than 300 ft below land surface, water with a dissolved-solids concentration higher than 2,000 mg/L extends only about 3 mi to the east.

The high sulfate concentration in ground water downgradient from Bingham Canyon is attributable to the oxidation of sulfide ore from the Oquirrh Mountains and associated mining processes. Acidic water that contains high concentrations of sulfate and is stored in the mine reservoirs and evaporation ponds recharges the aquifer and reacts with carbonate minerals in the aquifer to produce relatively high calcium activity in the ground-water samples (fig. 11). Saturation indices for gypsum are near zero in water sampled from wells (C-3-1)8cdd-2 and 8cdd-3 (map nos. 22 and 23 in table 1). Water from other areas in the valley is mostly undersaturated with respect to gypsum. The gypsum saturation in these two wells is the result of calcium from the dissolving calcium carbonate type rocks reacting with sulfate ions in the water from the surface-water bodies.

Three water samples collected in 1983 from wells in the central part of the valley, (C-1-1)23caa-1, (C-1-1)26caa-1, and (C-1-1)26dba-5 (map nos. 8, 10, and 11 in table 1), also had high dissolved-solids and sulfate concentrations relative to other wells in the area. The wells are located in or downgradient from the Vitro tailings site where uranium ore was processed from 1951 to 1964. Prior to the removal of the uranium mill tailings in 1987, leaching of the tailings resulted in an increase in dissolved-solids and heavy-metals concentrations in nearby ground water (Waddell and others, 1987, p. 29). Other localized areas near ore-processing sites in the valley also may contain high dissolved-solids and heavy-metal concentrations in ground water.

Water from wells (D-3-1)12bdd-1 and (D-3-1)14dbc-1 (map nos. 39 and 40 in table 1) near the mountain front on the southeastern side of the valley is undersaturated with calcite (fig. 11). These wells are 171 and 154 ft deep, respectively. Assuming the recharge area is at the boundary between the consolidated-rock and basin-fill material, there was probably not sufficient time for water-rock interactions to saturate the water with calcite.

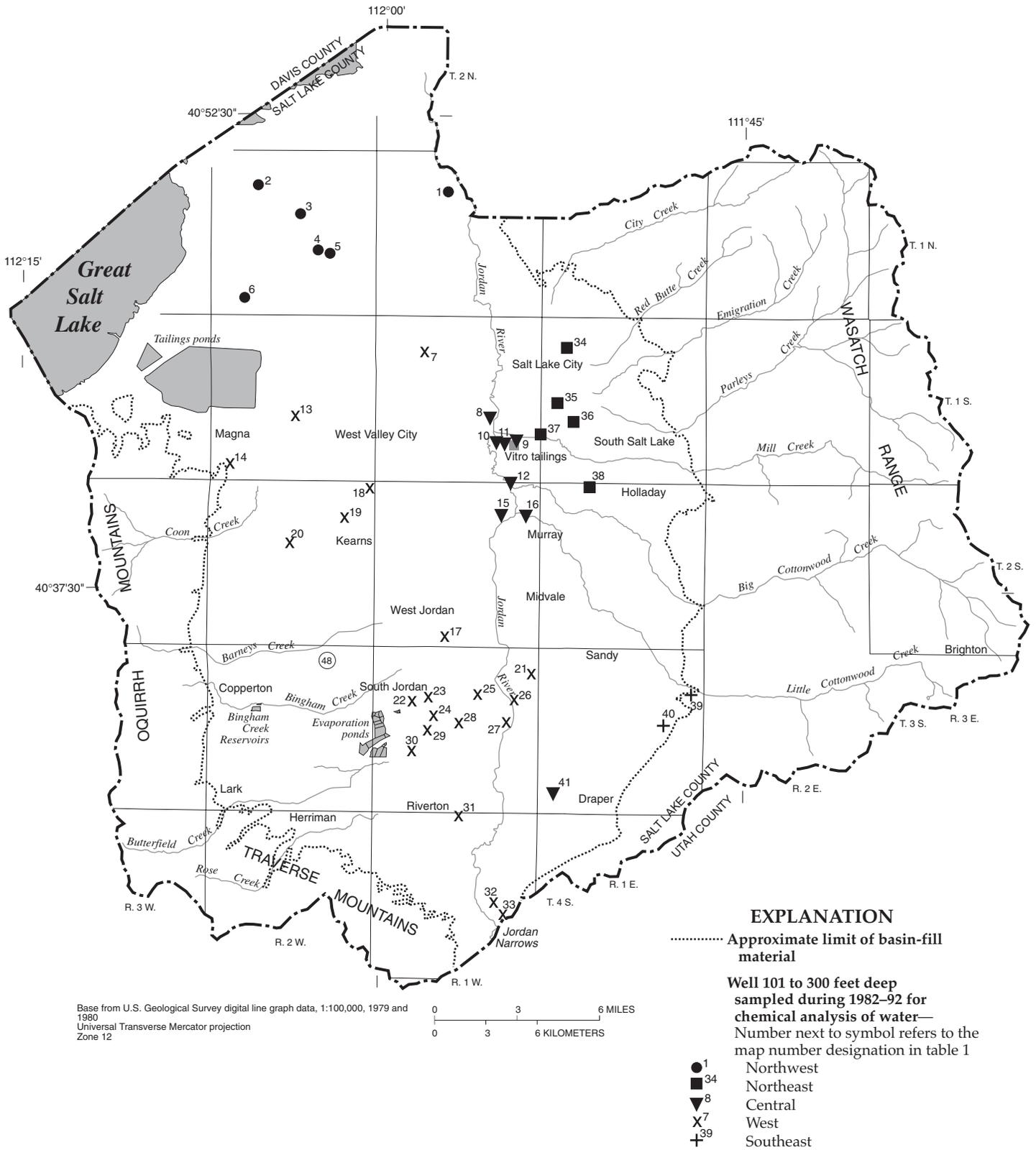


Figure 9. Location of wells 101 to 300 feet deep in Salt Lake Valley, Utah, sampled for chemical analysis of water, 1982–92.

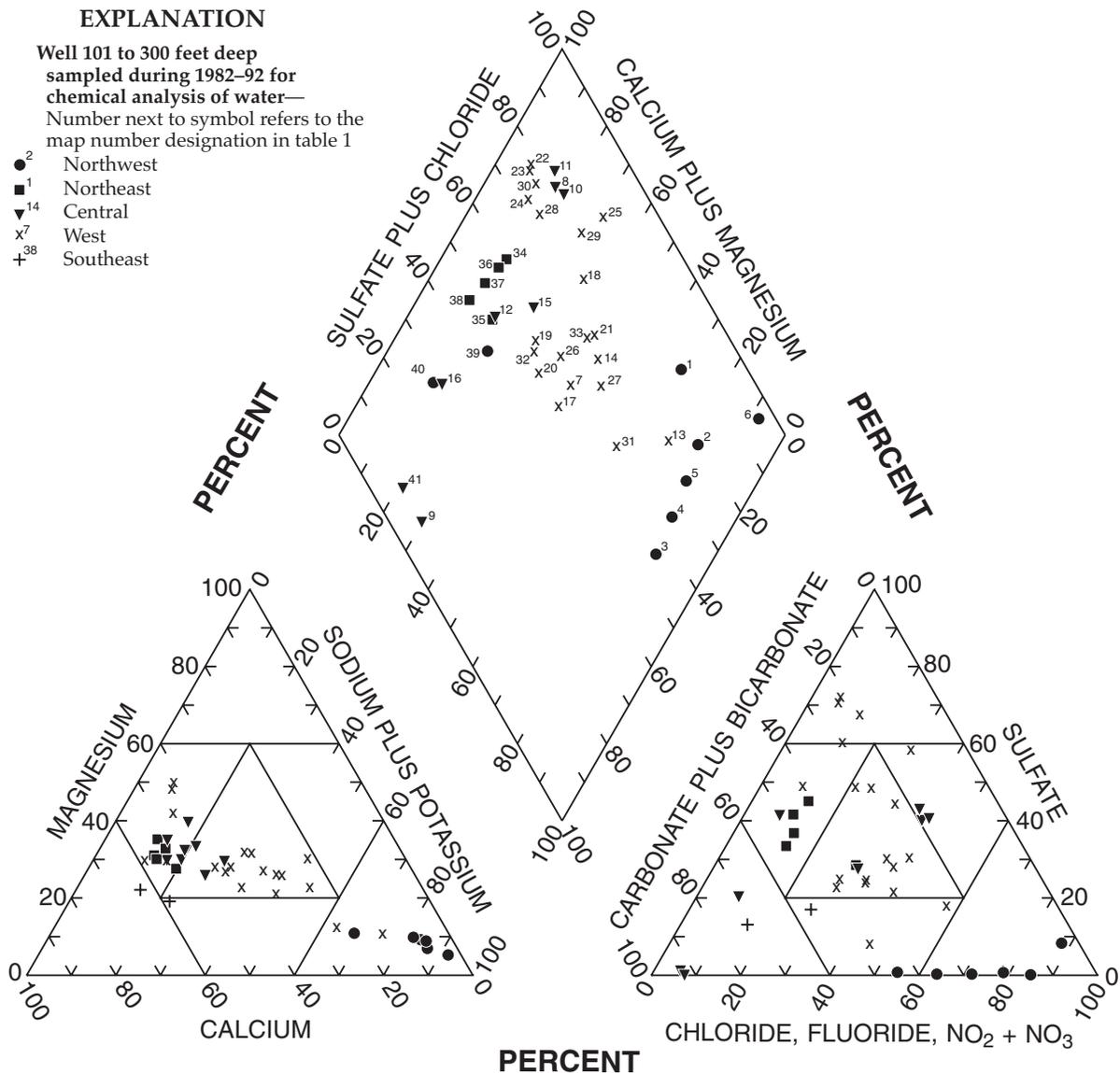


Figure 10. Chemical composition of water from wells 101 to 300 feet deep in Salt Lake Valley, Utah.

Water in the Shallow Unconfined Aquifer and Confining Layers

Results of chemical analyses of ground water from wells less than 101 ft deep were selected to represent water primarily from the shallow unconfined aquifer and confining layers. In the northwestern and central parts of the valley, confining layers are close to land surface, and part of the principal aquifer may be included in this depth interval. Samples listed in table 1 under the heading "Water samples from wells less than 101 feet deep" from the northwestern and central

parts of the valley might represent water from the shallow unconfined aquifer, underlying confining layers, the confined part of the principal aquifer, or a mixture of the three. Water samples were available from 65 wells sampled by the U.S. Geological Survey from 1983 to 1991.

The dissolved-solids concentration of water from the shallow unconfined aquifer ranged from 331 mg/L in the eastern part of the valley to 20,900 mg/L in the northwestern part (table 1 and fig. 12). The range is caused mainly by mixing and by the dissolution of minerals as ground water moves through the basin-fill material toward the northwestern part of the valley.

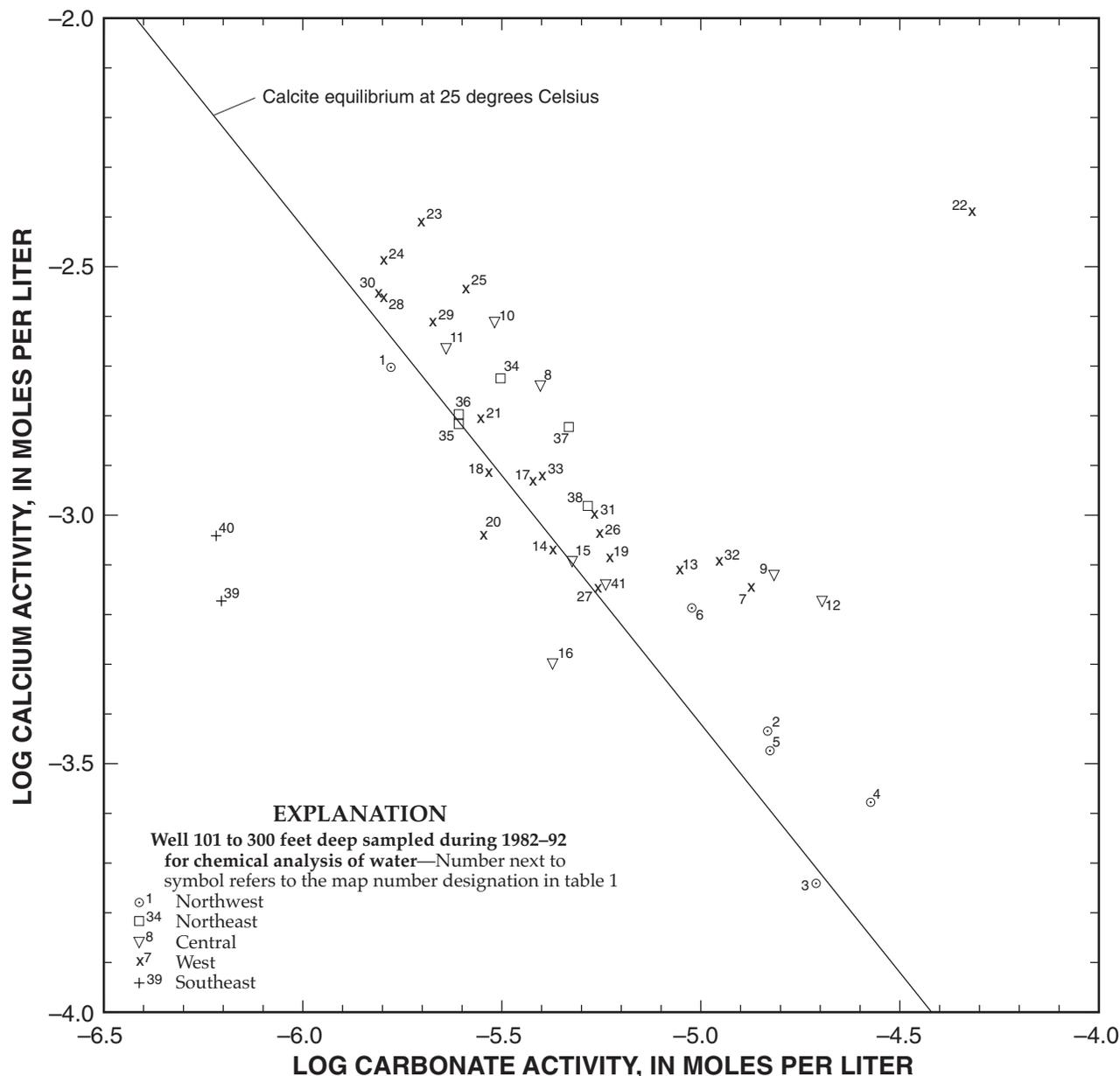


Figure 11. Relation between log calcium and log carbonate activity for water sampled from wells 101 to 300 feet deep in Salt Lake Valley, Utah.

Seepage to the shallow aquifer of surface water that has undergone some evaporation also causes an increase in dissolved-solids concentration relative to water in the principal aquifer in some areas of the valley. The Jordan River and its diversions are sources of water that have undergone some evaporation.

Water from the shallow unconfined aquifer has more localized variation in chemical composition than does water from the principal aquifer, primarily because of the influence of activities at the land surface. The shallow ground water is composed mainly of a mixture of ground water that is recharged along the

mountain fronts and is similar in chemical composition to water in mountain-front streams, upward leakage from the confined part of the principal aquifer, and water derived from the Jordan River. Water in Wasatch Range streams is primarily a calcium, magnesium-bicarbonate type, and water in the Jordan River at the Jordan Narrows is a sodium, magnesium, calcium-sulfate, chloride, and bicarbonate type (fig. 7).

Predominant ions in shallow ground water sampled from the northeastern and southeastern parts of the valley are calcium, magnesium, sodium, bicarbonate, sulfate, and chloride (fig. 13). This ground water prob-

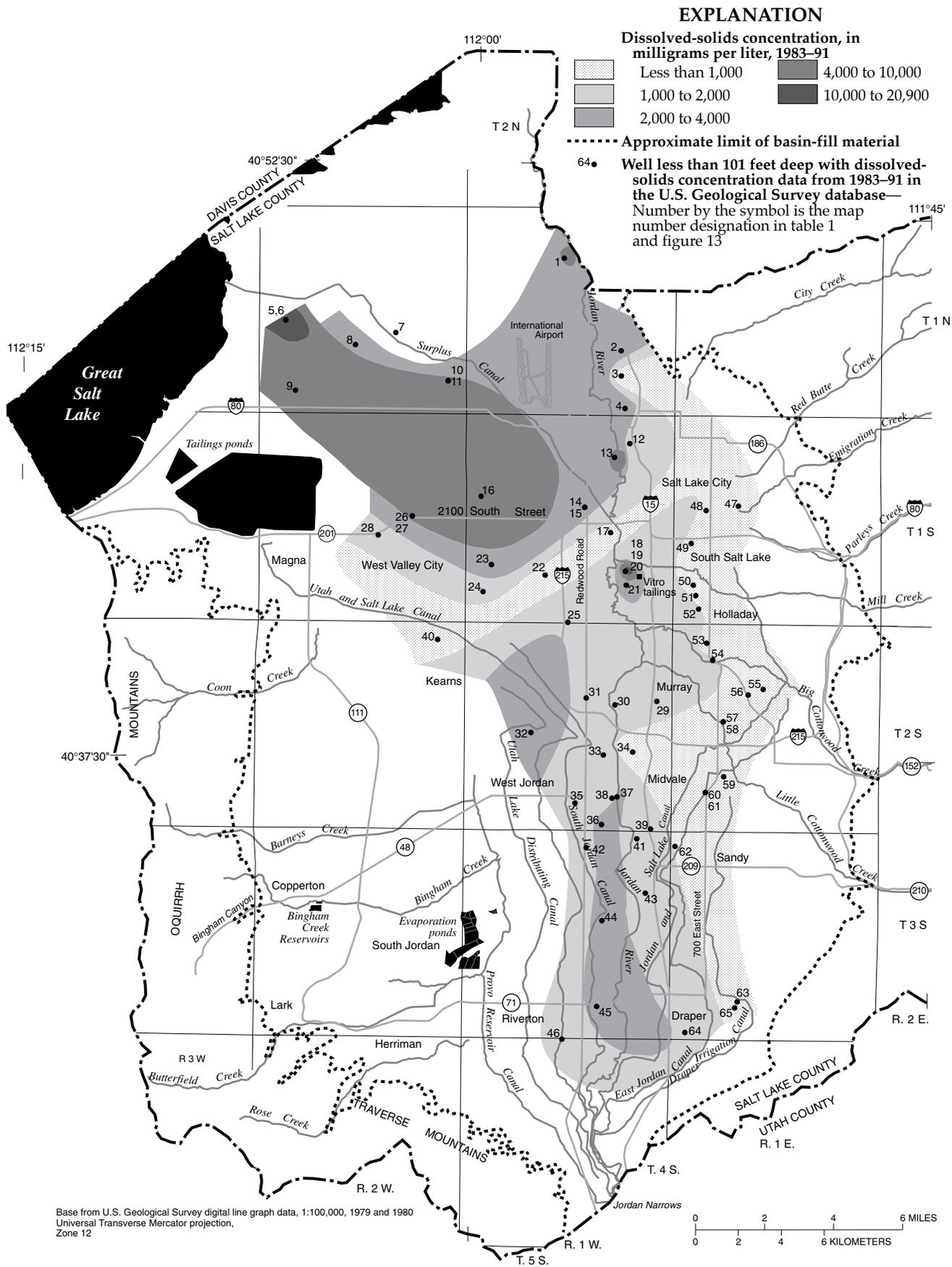


Figure 12. Dissolved-solids concentration in water from wells less than 101 feet deep in Salt Lake Valley, Utah, 1983-91. The dissolved-solids concentration for water sampled from the shallowest well at each site was used.

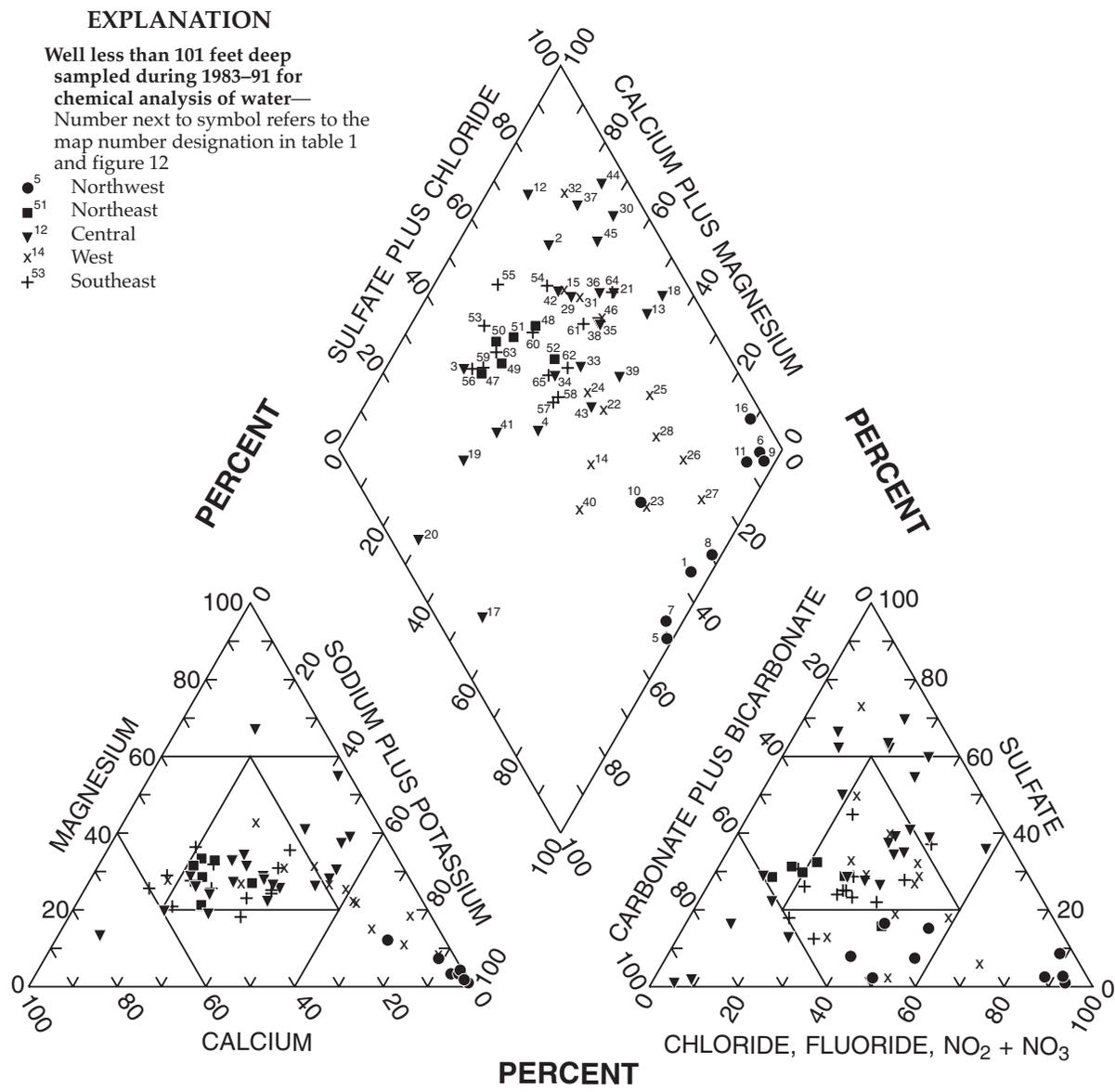


Figure 13. Chemical composition of water sampled from wells less than 101 feet deep in Salt Lake Valley, Utah.

ably is a mixture of water from mountain-front streams and water from the Jordan River. Shallow ground water from the southern and western parts of the valley contains more sodium and chloride ions than does water from the eastern part. Many of the wells sampled in the southern and western parts of the valley are near canals that transport water from the Jordan River to irrigated fields. Water from wells (C-4-1)4aaa-3 and (D-2-1)30dda-3 (map nos. 46 and 61 in table 1) is chemically similar to water sampled from the Jordan River at the Jordan Narrows. The specific conductance of water removed during the final stages of development from well (C-4-1)4aaa-4 in June 1991 was 1,640 $\mu\text{S}/\text{cm}$. The

well is perforated from 93 to 103 ft below land surface and is near the Utah and Salt Lake Canal, which diverts water from the Jordan River. The specific conductance of water from the canal at that time was 1,600 $\mu\text{S}/\text{cm}$. Stable-isotope data for ground water and water-level fluctuations measured in monitoring wells finished at different depths also indicate that the source of recharge to the shallow aquifer in these areas is water diverted from the Jordan River.

Ground-water samples from shallow wells in the northwestern part of the valley were classified into two types, sodium-chloride type and sodium-chloride and

bicarbonate type. This is the same grouping determined for water from deeper intervals. Dissolved-solids concentration in the sodium-chloride and bicarbonate type waters generally was less than in the sodium-chloride type but greater than concentrations in water from other parts of the valley.

Predominant ions in water from wells (C-1-1)15cad-2, (C-1-1)23bcd-2, (C-1-1)26dba-7, and (C-1-1)33ddd-2 near the Jordan River, with finished intervals in or below fine-grained material, are sodium, calcium, magnesium, bicarbonate, and chloride. These wells are 40 to 72 ft deep and contain very little sulfate dissolved in the water. Sulfate is the most common anion in water sampled from shallower depths and also is present in higher concentrations in water from the principal aquifer in the area. The low concentrations of sulfate in water from confining layers may be the result of sulfate reduction.

At selected locations in the valley, water sampled from the shallow unconfined aquifer or confining layers and from the underlying principal aquifer was compared to determine water-chemistry variation with depth. In almost all cases, the concentration of sodium and chloride was higher in the shallow ground water than in the water from deeper zones. This can be caused by the dissolution of minerals at land surface, such as road salt, seepage of water from the Jordan River into the shallow unconfined aquifer, evapotranspiration, and in the northwestern part of the valley, by the dissolution of minerals left by the desiccation of Lake Bonneville.

Water from wells (B-1-2)36baa-2 and (C-1-2)14cdd-2 in the northwestern part of the valley contained substantially lower concentrations of dissolved-solids than did water from deeper and shallower intervals at the same areal location. Well (B-1-2)36baa-2 is slotted from 72 to 82 ft below land surface (table 2) in a silty sandy clay. Well (C-1-2)14cdd-2 is slotted from 38.5 to 48.5 ft below land surface in a clayey silt. Both wells are finished in a zone that is about an order of magnitude more permeable than the zone from about 12 to 18 ft below land surface, which contained higher concentrations of dissolved solids (see "Slug tests" section of report). Hydraulic conductivity is a factor in controlling the time of contact between the water and the aquifer material and the resulting chemical reactions. Ground water from deeper intervals has been in contact with the basin-fill material for a longer period and also may mix with water from underlying deposits

of Tertiary age that typically contain more dissolved solids.

Ground-water samples collected in 1990 and 1991, mostly from monitoring wells completed in the shallow unconfined aquifer and underlying confining layers, were analyzed for selected volatile organic compounds, organochlorine pesticides, and polychlorinated biphenyls. The compounds analyzed, detection limits, and the compounds detected are presented by Thiros (1992, tables 5–7). Small quantities of chloroform were detected in two monitoring wells, (B-1-2)19aca-2 and (D-1-1)17dab-1. The chloroform may be from chlorinated drinking water used to drill and complete the wells and not completely removed from the aquifer during well development.

A small concentration of tetrachloroethylene (0.23 µg/L) was detected in water from well (D-1-1)17dab-2. Tetrachloroethylene, also known as perchloroethylene (PCE), is a relatively volatile chlorinated hydrocarbon that is used in the dry cleaning and metal degreasing industries. The PCE most likely is moving downgradient along the ground-water flow path.

Historical Changes in Chemical Composition of Ground Water

Results of chemical analyses of ground water sampled in Salt Lake Valley were studied to determine if changes in chemical composition have occurred. Additional data were collected from 1989 to 1992 at 11 wells where water-chemistry changes were noted from the 1960's to the early 1980's (Waddell and others, 1987, p. 9). The dissolved-solids concentration in water increased at five wells, decreased at one well, and varied with time at the other five wells (table 3).

The chloride concentration in water from well (D-1-1)7abd-6 increased steadily from about 50 mg/L in the late 1950's to 130 mg/L in July 1992 (table 1). The well is located in an area of the valley with artesian conditions and had a water level of about 11.5 ft above land surface in February 1992. Chloride concentration increased in water from well (A-1-1)31cac-1, perforated at several intervals from 162 to 420 ft below land surface. Chloride concentration ranged from 46 to 49 mg/L for yearly water samples from 1957 to 1962. The chloride concentration in water sampled in 1989 was 92 mg/L (table 1). A trend of increasing chloride concentration also was determined in water from well (D-2-1)21dbc-1 from 1979 (7.2 mg/L) to 1992 (11 mg/L).

Table 3. Dissolved-solids concentration in water from 11 wells in Salt Lake Valley, Utah, sampled in 1960-66, 1981-84, and 1989-92

Dissolved-solids concentration: in milligrams per liter: determined as sum of constituents or as R. residue on evaporation at 180 degrees Celsius.

Well location	Year sampled	Dissolved-solids concentration	Year sampled	Dissolved-solids concentration	Year sampled	Dissolved-solids concentration
(B-1-2)2dac-1	1966	481	1981	491	1991	498
(B-12)19aca-1	1966	1,270	1984	1,240	1991	1,260
(B-1-2)36baa-1	1960	3,840 R	1982	3,700R	1991	3,570 R
(C-1-1)33ddd-1	1960	586	1981	905	1992	969
(C-1-2)22bdd-4	1961	972	1982	1,140	1990	1,230
(C-1-2)22cbb-1	1960	912	1983	1,350	1989	1,340 R
(C-3-1)12ccb-1	1960	554	1982	555	1992	562
(C-4-1)23dbb-1	1960	926	1981	740	1992	876
(C-4-2)1bbb-1	1966	619	1981	826	1989	705
(D-1-1)7abd-6	1961	581	1982	656	1992	735
(D-3-1)31abb-1	1966	260	1981	268	1991	265

These wells are finished in the principal aquifer and are used for public supply. Waddell and others (1987, p. 11) suggested that a possible cause for the increase in chloride is the storage and use of road salt in recharge areas along the east side of the valley. The road salt dissolves in water and can infiltrate to the ground-water system. If dissolved road salt is the cause for an increase in chloride concentrations, the principal aquifer in this area may be susceptible to other contaminants introduced at land surface in recharge areas as well.

Dissolved-solids concentration in water from wells finished in the principal aquifer west of Riverton has increased since samples were collected in the late 1950's and 1960's. The area where an increase has been measured includes Sections 31 and 32, Township 3 South, Range 1 West, and Sections 5 and 6, Township 4 South, Range 1 West. Ground water in this area contained higher concentrations of dissolved solids during 1988-92 than did water front adjacent areas to the north, west, and east (pl. 2). Water samples were not available from the area to the south.

The dissolved-solids concentration in water from well (C-3-1)31dcd-3 was 491 mg/L in August 1958. Samples collected and analyzed by Kennecott Utah Copper (1992) beginning in 1984 show that dissolved-solids concentrations increased from 1,700 mg/L in 1984 to 2,840 mg/L in November 1991. The well is 625 ft deep and is used during the irrigation season to supplement water from the Provo Reservoir Canal.

Water from well (C-3-1)31aaa-1 had a dissolved-solids concentration of 691 mg/L in August 1965. The concentration remained close to this value until October 1985, when 800 mg/L of dissolved solids were measured. By June 1992, the water contained 1,110 mg/L (Kennecott Utah Copper, 1992, and Riverton City, written commun., 1993). Well (C-3-1)31aaa-1 is 700 ft deep and used for public supply. Reperforation of the well in 1989 resulted in a slight decrease in dissolved-solids concentration in the water.

Increases in dissolved-solids concentrations also were measured in public-supply wells (C-3-1)32cdc-1 and (C-3-1)32dcc-1. Well (C-3-1)32cdc-1 was reperforated during November 1988 to May 1989 to include the interval from 320 to 470 ft below land surface. The original perforations were from 530 to 633 ft below land surface. The dissolved-solids concentration in water from the well was 459 mg/L in January 1955 and 398 mg/L in October 1987 (Kennecott Utah Copper, 1992). The dissolved-solids concentration was 665 mg/L in February 1989, the next available analysis, and had increased to 994 mg/L in June 1992 (Riverton City, written commun., 1993).

Work was done on well (C-3-1)32dcc-1 from July to November 1987 to close off perforations from 155 to 174 ft below land surface. The remaining perforated interval extends from 187 to 420 ft below land surface. The dissolved-solids concentrations in water from the well ranged from 985 mg/L in August 1958 to 801 mg/L in September 1966. A measurement in Sep-

tember 1984 was similar to measurements made in the 1960's. The concentration increased from 1,155 mg/L in October 1987 to 1,758 mg/L in June 1992 (Riverton City, written commun., 1993). It is not known if the increase in dissolved-solids concentration was caused by changing the perforated interval in the well.

Hely and others (1971, p. 163) attributed the relatively low concentration of dissolved solids in the area during the 1950's and 1960's to the greater permeability of nearby consolidated rocks and recharge from the Provo Reservoir Canal. The Provo Reservoir Canal transported water from Deer Creek Reservoir east of Provo, Utah, for irrigation in Salt Lake Valley until 1990, when water from Utah Lake was used. The average discharge-weighted dissolved-solids concentration from 1964 to 1968 in water from Deer Creek Reservoir was 231 mg/L (Hely and others, 1971, p.74). The average discharge-weighted dissolved-solids concentration from 1964 to 1968 in water from the Jordan River at the Jordan Narrows was 1,120 mg/L (Hely and others, 1971, p. 90). A net loss of 2.1 ft³/s/mi was measured from the canal in the area upgradient from well (C-3-1)32dcc-1 (Herbert and others, 1985, p. 6). The change in surface-water sources could not be the only reason for increasing dissolved-solids concentration in ground water in the area because the concentration began increasing before the change occurred.

An increase in chloride concentration is the major cause for the increase in dissolved-solids concentration in water from the principal aquifer in the area west of Riverton. The predominant anion in water from Barneys Spring in the Oquirrh Mountains is chloride (Dames and Moore, 1988, pl. 18c). Volcanic rocks of Tertiary age are exposed near the spring and along the mountain front and underlie the western and southern extent of unconsolidated basin-fill material in this part of Salt Lake Valley. Water from these rocks may be a source of chloride to water in the principal aquifer.

Flow from Bingham Tunnel (pl. 1) and from several unlined collection ditches on the Oquirrh Mountains near Lark were redirected for use in mining operations near Copperton in 1987 (Dames and Moore, 1988, p. 7 and 11). Prior to this, water from Bingham, Butterfield, and Mascotte Tunnels was used for irrigation in the Herriman area. The dissolved-solids concentration in water was 2,050 mg/L at Bingham Tunnel (Kennecott Utah Copper, 1992), 838 mg/L at Butterfield Tunnel (Kennecott Utah Copper, 1992), and 1,310 mg/L at Mascotte Tunnel (Hely and others, 1971, p. 164). The predominant anion in water from the tunnels

was sulfate, with chloride constituting less than 10 percent of the ions present.

Water levels in the principal aquifer declined in the southwestern part of the valley from February 1988 to February 1993 (Batty and others, 1993, p.25). Water-level declines greater than 20 ft were noted in wells in the area of increased chloride concentration. The drop in water levels is probably the result of ground-water withdrawals from wells, the loss of recharge from the mine tunnels, and generally less than-average precipitation during 1988–92.

Many domestic-supply wells have been drilled in an area south and west of Herriman since about 1985. Fractured volcanic rock of Tertiary age occurs at most of these wells near land surface. According to drillers' logs, an increased quantity of water under hydraulic pressure exists where the wells intersect fractures. Water analyses provided to the Utah Department of Environmental Quality, Division of Drinking Water, for wells finished in these volcanic rocks indicate a calcium, sodium to calcium, magnesium-bicarbonate, chloride type water (Utah Department of Environmental Quality, Division of Drinking Water, written commun., 1993). Water pumped from wells west of Riverton that shows an increase in chloride and dissolved-solids concentration is similar in composition to water sampled from volcanic rocks in the area. More water-chemistry information is needed on water sampled from different depths from volcanic rocks in the area.

HYDROLOGIC PROPERTIES OF THE BASIN-FILL MATERIAL

Data were collected from the monitoring wells installed during this study to better estimate the hydrologic properties of the basin-fill material. Data also were collected from existing wells in the vicinity of selected monitoring wells.

Slug Tests

Hydraulic conductivity determined using the Bouwer and Rice (1976) method ranged from 0.003 ft/d at well (C-2-1)36cdd-2, finished in an interval that was not sampled, to 33.4 ft/d at well (D-2-1)17cda-3, finished in an interval of gravel that grades downward into sandy clay. Each of the wells was tested as many as six times to obtain some consistency for derived hydraulic-

conductivity values. The most representative values for 30 wells are listed in table 4. Water-level changes measured after raising the slug above the water surface typically gave more uniform hydraulic-conductivity values. Rapidly lowering the slug below the water surface may cause a larger initial disturbance of the water surface that takes longer for the effects to subside than does removing the slug.

A representative set of data for each of the wells also was analyzed with the Bouwer and Rice (1976) method using the computer program AQTESOLV from Geraghty and Miller, Inc. (1991). The computer program also was used to match Cooper, Bredehoeft, and Papadopulos method (Cooper and others, 1967) type curves to representative data sets for the wells. If the slug-test data for a particular well could not be matched to a type curve, the interval of aquifer open to the well was probably unconfined and a value was not listed in table 4 for the Cooper, Bredehoeft, and Papadopulos method. Values of hydraulic conductivity determined by the Cooper, Bredehoeft, and Papadopulos method (Cooper and others, 1967) ranged from 0.003 ft/d at well (C-2-1)36cdd-2 to 65.5 ft/d at well (D-2-1)17cda-3, finished in an interval with gravel, cobbles, and clay. These values are the same order of magnitude as the values determined using the Bouwer and Rice (1976) method.

Multiple-Well Aquifer Tests

Hydrologic properties were estimated from aquifer tests done at four sites in Salt Lake Valley using a pumped well and nearby observation wells. Transmissivity values determined for the principal aquifer from these tests ranged from 6,400 to 43,600 ft²/d, storage coefficients were about 1×10^{-4} , and vertical hydraulic-conductivity values ranged from about 0.01 to 1 ft/d, where determined. A description of each site and the results of the aquifer-test analysis is reported in the following sections.

Aquifer Test at Well (C-2-1)23dad-1

Well (C-2-1)23dad-1 was pumped on February 13, 1991, for about 7 hours at a near constant rate of about 1,600 gal/min. The pumped well is used for public supply during the summer months and has three perforated intervals from 140 to 380 ft below land surface. The Jordan River is about 3,500 ft west of the well. According to the drillers' log of well (C-2-1)23dad-1, a

12-foot-thick layer of clay begins at about 80 ft below land surface. Although in the discharge area for the principal aquifer (pl. 1), water levels measured in well (C-2-1)23dac-1, about 300 ft from the pumped well and finished at the top of the clay layer, are similar to those in well (C-2-1)23dad-1 during the nonpumping season.

Water-level decline and recovery was measured in the pumped well, in two observation wells finished in the pumped principal aquifer, and in well (C-2-1)23dac-1, finished above the principal aquifer (table 5). The water-level decline measured in well (C-2-1)23cab-1 did not fit the Hantush (1960) modified method family of type curves when plotted with the water-level decline measured in well (C-2-1)24bcd-1. The water-level decline in well (C-2-1)23cab-1 was not as large as the type curves indicated. This indicates that the aquifer was not responding as if it were homogeneous or that another source of water was available to well (C-2-1)23cab-1 that was not available to the pumped well or to well (C-2-1)24bcd-1. A possible source for this other water is the Jordan River. The clay layers above the principal aquifer thin west of the pumped well (toward the Jordan River) and thicken on the west side of the river, on the basis of drillers' logs. The clay layer nearest to land surface deposited by Lake Bonneville may have been eroded by the Jordan River between well (C-2-1)23cab-1 and the pumped well, allowing surface water to recharge the ground-water system. A delta-like escarpment just north of the pumped well is composed of alluvium of Holocene age deposited by an abandoned channel of Little Cottonwood Creek (Personius and Scott, 1990). This abandoned channel also could have eroded the uppermost deposit of lacustrine clay in the area of well (C-2-1)23cab-1.

Water-level recovery measured in observation well (C-2-1)24bcd-1 through time and a leakage parameter of 0.1 indicate a transmissivity for the pumped aquifer of about 43,600 ft²/d and a storage coefficient of 3.5×10^{-4} . Transmissivity calculated by the Cooper and Jacob (1946) straight-line method using water-level recovery measured in the pumped well (C-2-1)23dad-1 is about 48,500 ft²/d.

Changes in water level were measured in monitoring well (C-2-1)23dac-1 in response to pumping at well (C-2-1)23dad-1. A water-level rise of 0.07 ft was measured after pumping began. The reverse change in water level is attributed to compaction of the aquifer material during pumping and a corresponding increase in pore pressure in the aquifer and adjacent layers. A

Table 4. Hydraulic-conductivity values determined from slug tests done at selected wells in Salt Lake Valley, Utah

[ft/d, foot per day; —, no data available]

Location: See figure 2 for an explanation of the numbering system used for hydrologic-data sites in Utah.

Location	Initial water level (feet below land surface)	Length of finish interval (feet)	Bottom of finish interval (feet below land surface)	Hydraulic conductivity determined with the Bouwer and Rice (1976) method (ft/d)	Hydraulic conductivity determined by a computer program using the Bouwer and Rice (1976) method (ft/d)	Hydraulic conductivity determined by a computer program using the Cooper and others (1967) method (ft/d)
(B-1-2)19aca-3	2.67	5	22.1	0.16	0.16	0.16
(B-1-2)22dad-1	7.01	5	14.5	.30	.26	.45
(B-1-2)28aab-1	2.79	5	14.0	1.03	.85	1.64
(B-1-2)32bcb-3	3.49	5	23.7	.30	.39	—
(B-1-2)36baa-2	3.83	10	82.0	3.49	3.45	3.64
(B-1-2)36baa-3	5.92	5	18.0	.19	.20	.24
(C-1-1)15cad-2	4.21	10	39.0	.25	.26	—
(C-1-1)15cad-3	3.97	5	15.0	17.88	19.01	—
(C-1-1)18bda-1	7.60	10	44.0	17.39	18.56	—
(C-1-1)23bcd-2	2.81	10	65.1	.05	.05	.04
(C-1-1)33ddd-2	11.55	10	50.5	.007	.006	—
(C-1-2)14cdd-2	.11	10	48.5	7.80	7.62	8.66
(C-1-2)14cdd-3	3.62	5	16.7	.51	.43	.53
(C-2-1)13abc-1	20.21	10	55.5	2.03	2.27	2.80
(C-2-1)20aad-2	43.37	10	60.0	.14	.10	—
(C-2-1)23ccb-1	13.55	10	40.0	2.15	2.16	2.67
(C-2-1)34bbd-2	27.06	10	40.5	12.63	12.40	20.61
(C-2-1)36cdd-2	37.59	10	61.0	.003	.003	.003
(C-2-1)36cdd-3	69.47	10	128.5	1.45	1.40	1.33
(C-2-2)1bcd-2	6.37	10	27.0	8.21	9.17	13.29
(C-3-1)12cdb-1	8.59	10	40.0	.41	.34	0.36
(C-4-1)4aaa-4	52.49	10	103.0	27.32	24.25	35.69
(D-1-1)17dab-2	20.63	10	67.5	4.31	4.51	5.37
(D-1-1)19dbc-2	8.65	10	30.0	2.26	2.33	3.16
(D-2-1)16bba-3	21.46	10	89.0	10.09	11.50	10.82
(D-2-1)17cda-2	2.05	10	27.5	7.53	7.86	—
(D-2-1)17cda-3	2.00	10	22.5	33.40	33.21	65.54
(D-2-1)29acb-2	23.08	10	51.0	2.54	2.36	1.01
(D-2-1)30dda-3	11.47	5	20.4	10.58	10.43	16.86
(D-3-1)29ddd-1	5.66	10	43.0	.85	.86	.90

Table 5. Records of wells monitored during four aquifer tests done in Salt Lake Valley, Utah

Location: See figure 2 for an explanation of the numbering system used for hydrologic-data sites in Utah.

Altitude of land surface: In feet above sea level. Altitudes are interpolated from U.S. Geological Survey topographic maps and are to the nearest foot.

Depth of well: In feet below land surface.

Distance from pumped well: Radial distance in feet.

Finish intervals in well: Upper and lower limits of perforations, screens, or open hole in the well are in feet above sea level, if known, and questioned (?) if extent is unknown.

Location	Altitude of land surface (feet)	Depth of well (feet)	Distance from pumped well (feet)	Finish intervals in well (feet)
Aquifer test at well (C-2-1)23dad-1				
(C-2-1)23dad-1	4,280	391	0	4,140-4,060 4,043-4,000 3,910-3,900
(C-2-1)23dac-1	4,280	85	296	4,206-4,196
(C-2-1)24bcd-1	4,307	370	2,100	4,128-4,007
(C-2-1)23cab-1	4,285	600	3,600	4,045-4,023 3,992-3,948
(C-2-1)23ccb-1	4,350	41	4,480	4,320-4,310
Aquifer test at well (D-2-1)16bba-2				
(D-2-1)16bba-2	4,351	604	0	4,098-4,046 4,022-3,956 3,870-3,852 3,846-3,810 3,806-3,796 3,792-3,766
(D-2-1)16bba-3	4,351	90	84	4,272-4,262
(D-2-1)16bba-1	4,351	285	500	4,201-4,143 4,138-4,113 4,103-4,096 4,090-4,067
(D-2-1)16aaa-1	4,396	600	4,125	4,106-3,981 3,931-3,891 3,881-3,864
(D-2-1)9abd-1	4,348	540	5,250	4,088-4,053 4,033-3,983 3,923-3,888
Aquifer test at well (D-2-1)30dda-1				
(D-2-1)30dda-1	4,453	1,002	0	3,893-3,873 3,845-3,823 3,805-3,792 3,766-3,764 3,756-3,725 3,719-3,711 3,602-3,581 3,563-3,519 3,504-3,497 3,490-3,488

Table 5. Records of wells monitored during four aquifer tests done in Salt Lake Valley, Utah—Continued

Location	Altitude of land surface (feet)	Depth of well (feet)	Distance from pumped well (feet)	Finish intervals in well (feet)
Aquifer test at well (D-2-1)30dda-1—Continued				
(D-2-1)30dda-1—Continued				3,482-3,481 3,475-3,472 3,467-3,463
(D-2-1)30dda-2	4,453	65	100	4,399-4,389
(D-2-1)30dda-3	4,453	21.4	110	4,438-4,433
(D-2-1)29cdd-1	4,497	900	2,600	4,027-4,012 4,006-4,000 3,992-3,976 3,967-3,963 3,932-3,920 3,902-3,892 3,872-3,817 3,801-3,794 3,777-3,732 3,714-3,707 3,692-3,662 3,652-3,632 3,629-3,606
(D-2-1)32bcc-1	4,487	1,007	3,080	4,012-3,899 3,883-3,871 3,865-3,857 3,839-3,825 3,812-3,742 3,723-3,654 3,622-3,601 3,582-3,567 3,532-3,491
(D-2-1)30cac-1	4,397	200	3,500	4,212-4,197
(D-2-1)29acb-8	4,465	500	3,590	4,010-?
(D-2-1)32caa-1	4,513	881	4,290	4,038-4,003 3,973-3,933 3,893-3,643
(D-2-1)28ecc-1	4,580	691	5,470	4,065-4,050 4,045-3,995 3,988-3,902
(D-2-1)20ccb-1	4,414	626	5,480	4,134-4,124 4,104-4,044 4,019-3,954 3,914-3,894 3,884-3,799

Table 5. Records of wells monitored during four aquifer tests done in Salt Lake Valley Utah—Continued

Location	Altitude of land surface (feet)	Depth of well (feet)	Distance from pumped well (feet)	Finish intervals in well (feet)
Aquifer test at well (D-2-1)34dba-1				
(D-2-1)34dba-1	4,675	700	0	4,400-4,247 4,175-4,155 4,045-3,995
(D-2-1)34acb-1	4,637	325	881	4,366-4,346
(D-2-1)34cab-2	4,655	411	1,762	4,330-4,245
(D-2-1)34bbb-2	4,590	692	3,550	4,312-4,304 4,228-4,186 4,098-4,060 4,040-3,988
(D-2-1)33abd-2	4,581	904	5,380	4,176-4,026 4,006-3,677

water-level decline in well (C-2-1)23dac-1 was recorded 5 minutes after the pump was turned on.

The vertical hydraulic conductivity of the confining layer overlying the pumped principal aquifer was estimated using the Hantush (1960) modified method and the “ratio” method (Neuman and Witherspoon, 1972). Water-level changes measured in well (C-2-1)23dac-1 were used to represent changes in the confining layer. A specific-storage value for the confining layer of 9.5×10^{-6} ft⁻¹ was determined by solving $k'S_s'$ from the Hantush modified method and k'/S_s' from the “ratio” method simultaneously. Substituting the calculated S_s' value into these equations results in a k' of 0.06 ft/d. Vertical hydraulic conductivity of the confining layer in this area probably ranges from 0.01 to 0.1 ft/d.

A typical value of S_s' for alluvial material is about 1×10^{-5} ft⁻¹ (Francis Riley, U.S. Geological Survey, oral commun., 1992). The S_s' value determined from consolidation-test data for material cored 83.5 ft below land surface at well (C-2-1)23dac-1 was 2.2×10^{-4} ft⁻¹. This value is indicative of material that has experienced stress that exceeded the maximum past stress. The basin-fill material in this area has not undergone any significant changes in stress caused by ground-water withdrawals or land-surface loading since prehistoric Lake Bonneville. Therefore, the laboratory determined S_s' is considered to be too large, probably because of disturbance during coring or during the laboratory analysis.

Aquifer Test at Well (D-2-1)16bba-2

Well (D-2-1)16bba-2 was pumped for about 13 hours on February 20, 1991, at a near-constant rate of 1,400 gal/min. Water-level response was monitored in three other wells completed in the principal aquifer and in well (D-2-1)16bba-3, finished in the shallow unconfined aquifer above a confining layer (table 5). Finish intervals in the pumped well are from 253 to 585 ft below land surface. The water level in well (D-2-1)16bba-2 was 2.9 ft lower in February 1991 and 4.4 ft lower in May 1992 than the water level measured at the same time in well (D-2-1)16bba-3. The shallow well is about 84 ft south of the deep well. The difference in water levels in the two wells indicates that there is a downward vertical hydraulic-head gradient between the shallow unconfined and the principal aquifer in this area. The transmissivity of the principal aquifer, determined by applying the Cooper and Jacob (1946) straight-line method to water-level recovery in the pumped well, is about 16,300 ft²/d.

On the basis of drillers' logs that identify clay layers above and below finish intervals in the observation wells, the principal aquifer in this area likely is confined. No water-level decline was measured in well (D-2-1)16bba-3 during the aquifer test. Water levels were measured monthly at this relatively shallow well from September 1990 to March 1992. Seasonal fluctuations indicate that water levels in the well are affected by ground-water withdrawals from the principal aquifer.

fer. The aquifer test was not long enough in duration to quantify leakage but it did indicate that leakage through the confining layer is relatively small.

Aquifer Test at Well (D-2-1)30dda-1

The aquifer test at well (D-2-1)30dda-1 was done by monitoring recovery in nine observation wells finished in the principal and shallow unconfined aquifers after pumping at well (D-2-1)30dda-1, which is finished in the principal aquifer, stopped (table 5). Well (D-2-1)30dda-1 was turned off at 12:00 p.m. on February 10, 1992, after pumping at a near-constant rate of about 2,300 gal/min since February 6, 1992. Records supplied by the Salt Lake County Water Conservancy District (written commun., 1992) show that the well was pumped at about 2,600 gal/min from February 3 to 6, 1992.

Well (D-2-1)30dda-1 has several perforated intervals from 560 to 990 ft below land surface. There are several clay layers documented in drillers logs for wells in the area above these intervals. The water level in the pumped well was close to the prepumping level, about 136 ft below land surface, 5 days after pumping stopped. Well (D-2-1)30dda-2 is finished from 54 to 64 ft below land surface in silty clay. The water level in well (D-2-1)30dda-2 ranged from about 52 to 53 ft below land surface, according to measurements made from July 1991 to February 1992. Well (D-2-1)30dda-3 is finished from about 15 to 20 ft below land surface in silty sand. Water levels measured in well (D-2-1)30dda-3 fluctuated from a high of about 11.2 ft below land surface in August 1991 to a low of about 14.6 ft below land surface in February 1992. This fluctuation is primarily in response to the presence of surface water in the nearby East Jordan Canal. These observations indicate a steep downward vertical hydraulic-head gradient between the shallow unconfined aquifer [well (D-2-1)30dda-3], an underlying confining layer [well (D-2-1)30dda-2], and the principal aquifer [well (D-2-1)30dda-1]. The downward vertical hydraulic-head gradient probably exists throughout the year, regardless of the ground-water withdrawals and diversions for irrigation that occur seasonally.

Transmissivity of the principal aquifer in this area determined from the Hantush (1960) modified method is about 6,400 ft²/d. The storage coefficient is about 1.4×10^{-4} . A specific-storage value for the confining layer of 1.3×10^{-5} ft⁻¹ was determined by solving $k'S_s'$ from the Hantush modified method and k'/S_s' from

the "ratio" method simultaneously. Substituting the calculated S_s' value into these equations results in ax' of 0.08 ft/d. The probable range in k' in this area is from 0.01 to 0.1 ft/d.

Aquifer Test at Well (D-2-1)34dba-1

Well (D-2-1)34dba-1 was pumped for about 20 hours in May 1965. The well is finished in the principal aquifer with perforated intervals from 275 to 680 ft below land surface. The water level in the well prior to the test was 290 ft below land surface. The area around the pumped well is near the boundary of where the principal aquifer changes from confined to unconfined conditions. A downward vertical hydraulic-head gradient exists between the shallow unconfined and principal aquifers in this area.

Data from this aquifer test were reanalyzed in 1992 to account for substantial leakage through adjacent confining layers to the principal aquifer. Water-level declines were measured in the pumped well and in four observation wells completed in the principal aquifer (table 5).

The transmissivity of the principal aquifer estimated from the Cooper and Jacob (1946) straight-line method and the Hantush (1960) modified method is about 23,000 ft²/d. The storage coefficient is about 1.4×10^{-4} . The vertical hydraulic conductivity of the confining layer that overlies the principal aquifer was estimated using the Hantush (1960) modified method to be in the range of 0.1 to 1 ft/d.

WATER-LEVEL FLUCTUATIONS AND GROUND-WATER MOVEMENT

Water-level fluctuations and ground-water movement in the saturated basin-fill material was studied using water-level and water-chemistry data, geochemical modeling, isotope data, and vertical hydraulic-conductivity values. These data and methods provide insight to recharge sources, flow directions, and flow velocities in some areas of the valley.

Water-Level Fluctuations in the Shallow Unconfined Aquifer

Water levels were measured monthly in most of the monitoring wells drilled during this study. The changes in water levels tend to follow a seasonal pat-

tern that is dependent on water use and location in the valley. Water levels in wells (C-2-1)20aad-2, (C-4-1)4aaa-3, and (D-2-1)29acb-2 (fig. 14) are highest in the late summer or fall in response to infiltration of unconsumed irrigation water. Water levels measured in these wells are lowest in the spring before irrigation begins. These wells are near canals or ditches that divert water for irrigation.

Wells (C-2-1)23dac-1, (C-2-1)36cdd-3, and (D-2-1)16bba-3 are monitoring wells near public-supply wells. The public-supply wells were pumped during the summers of 1990 and 1991. Water levels measured in the nearby monitoring wells were lowest during the summer in response to ground-water withdrawals from the confined part of the principal aquifer. Water levels generally were highest in the early spring. Wells (D-1-1)17dab-2 and (D-1-1)30dcb-1 are not close to any wells that withdraw water from the principal aquifer, but water levels in these monitoring wells responded in the same way as in wells that are near pumping wells. The confined nature of the principal aquifer in this area of the valley and the large ground-water withdrawals for public supply cause the effects of pumping to propagate over a large area.

Monitoring well (D-2-1)17cda-2 is adjacent to Big Cottonwood Creek and a public-supply well finished in the principal aquifer. Water levels measured in the monitoring well fluctuated with flow in the stream. A large rise in water levels was measured in May 1991 that corresponds with spring runoff in Big Cottonwood Creek. Water levels were highest in June 1991 and gradually declined the rest of the year. Water-level changes measured in wells (B-1-2)31aaa-1, (C-1-1)15cad-2, and (C-1-1)23bcd-2 do not seem to be as greatly influenced by water use in the valley as those measured in other monitoring wells, probably because fewer wells in the area withdraw water from the principal aquifer.

A reverse water-level fluctuation was noted in monitoring well (C-1-1)23bcd-2 in response to ground-water withdrawals from a deeper part of the saturated basin fill. The well was drilled to 66.1 ft below land surface through interbedded clay, silt, and sand and is near a public-supply well, (C-1-1)23bcd-1, which has several screened intervals from 145 to 804 ft below land surface. Several times in the fall and winter of 1990, pumping at the public-supply well resulted in a water-level rise in the monitoring well (fig. 15). When well (C-1-1)23bcd-1 was not pumping, the water level in well (C-1-1)23bcd-2 declined. The water-level

changes measured in the monitoring well in response to pumping from the underlying principal aquifer ranged from 1.21 to 1.50 ft. This reverse fluctuation in water level is thought to be caused by the redistribution of stress that results when an aquifer is pumped. When a pump is turned on, the aquifer material may be compacted by the horizontal movement of water in the aquifer toward the pumped well. A rise in water level is caused by a corresponding increase in pore pressure in the aquifer and adjacent confining layers.

Ground-Water Movement Indicated by Geochemical Modeling

Chemical analyses of water from wells in areas of Salt Lake Valley where vertical hydraulic-head gradients were known were used to model possible reactions between the water and basin-fill material. The computer program NETPATH (Plummer and others, 1991) was used (see "Methods and Limitations" section of this report). Surface water and water from wells were specified as upgradient or downgradient (evolutionary) waters along a flowpath, depending on the direction of ground-water flow.

Water from well (C-4-1)4aaa-3 (table 1), 49.0 ft deep and near the Utah and Salt Lake Canal, is similar to water sampled from the Jordan River at the Jordan Narrows. The Utah and Salt Lake Canal diverts water from the Jordan River in the Jordan Narrows for irrigation. A water level of about 40 ft below land surface and the presence of unsaturated material above the water table (Thiros, 1992, p. 24) indicates a downward hydraulic-head gradient between land surface and the finish interval in the well. Water from the canal probably seeps down to the shallow unconfined aquifer. Concentrations of sulfate, sodium, total inorganic carbon, calcium, magnesium, and chloride are higher in water from the well than in water from the Jordan River. If water from the Utah and Salt Lake Canal is the source of water in well (C-4-1)4aaa-3, then some of the increase in dissolved-solids concentration must be caused by reactions with basin-fill material in the subsurface.

Sulfur, carbon, calcium, and chloride were used to constrain the modeling calculations. The mineral phases assumed to occur in this system were calcite, gypsum, halite, and carbon dioxide gas. The geochemical mass-balance model calculated for the specified constraints and phases requires the dissolution of 0.63 mmol/L calcite, 0.52 mmol/L gypsum, and 2.7 mmol/L

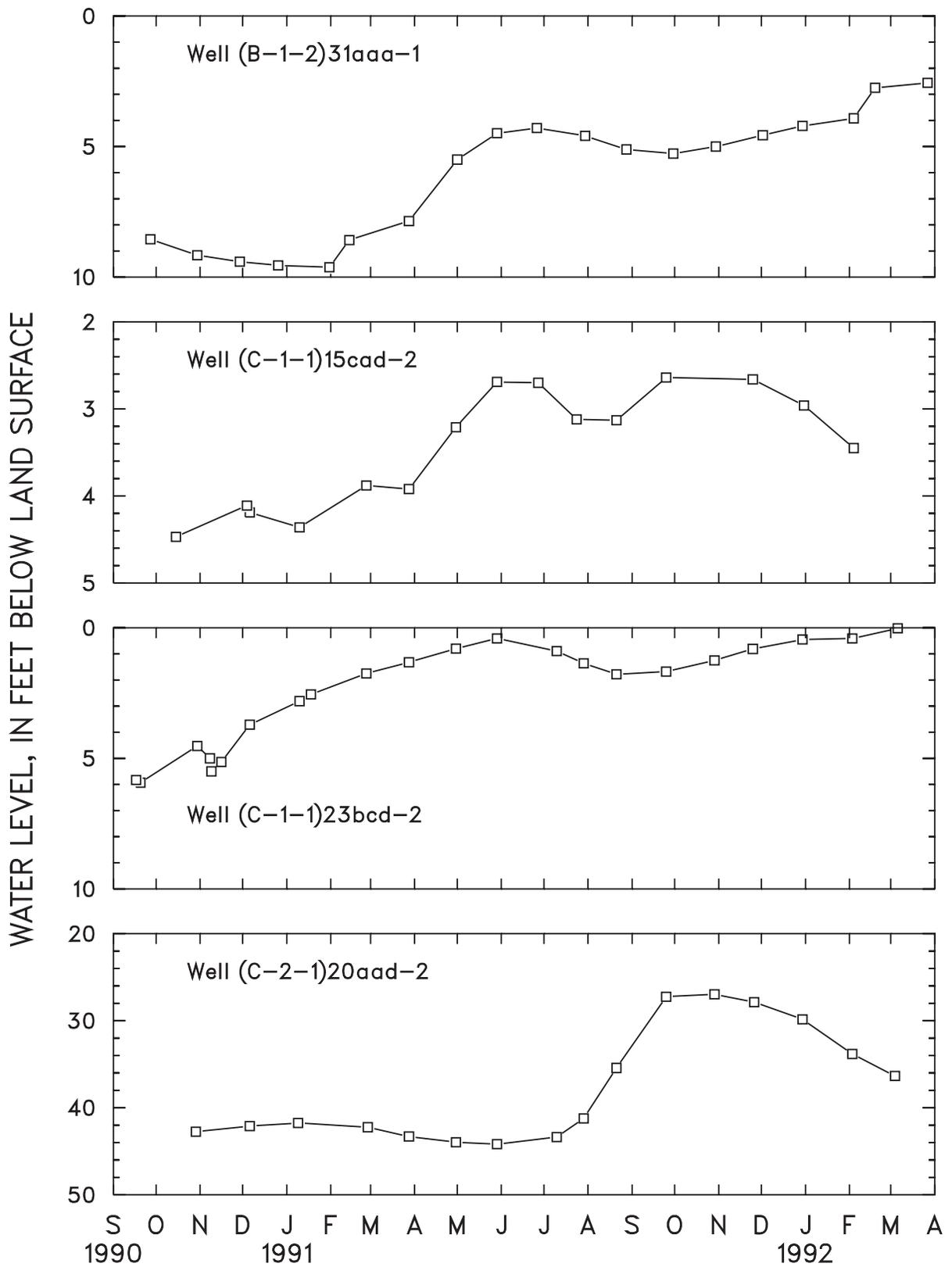


Figure 14. Seasonal water-level fluctuations in 12 monitoring wells completed in the shallow unconfined aquifer or a confining layer in Salt Lake Valley, Utah 1990–92.

WATER LEVEL, IN FEET BELOW LAND SURFACE

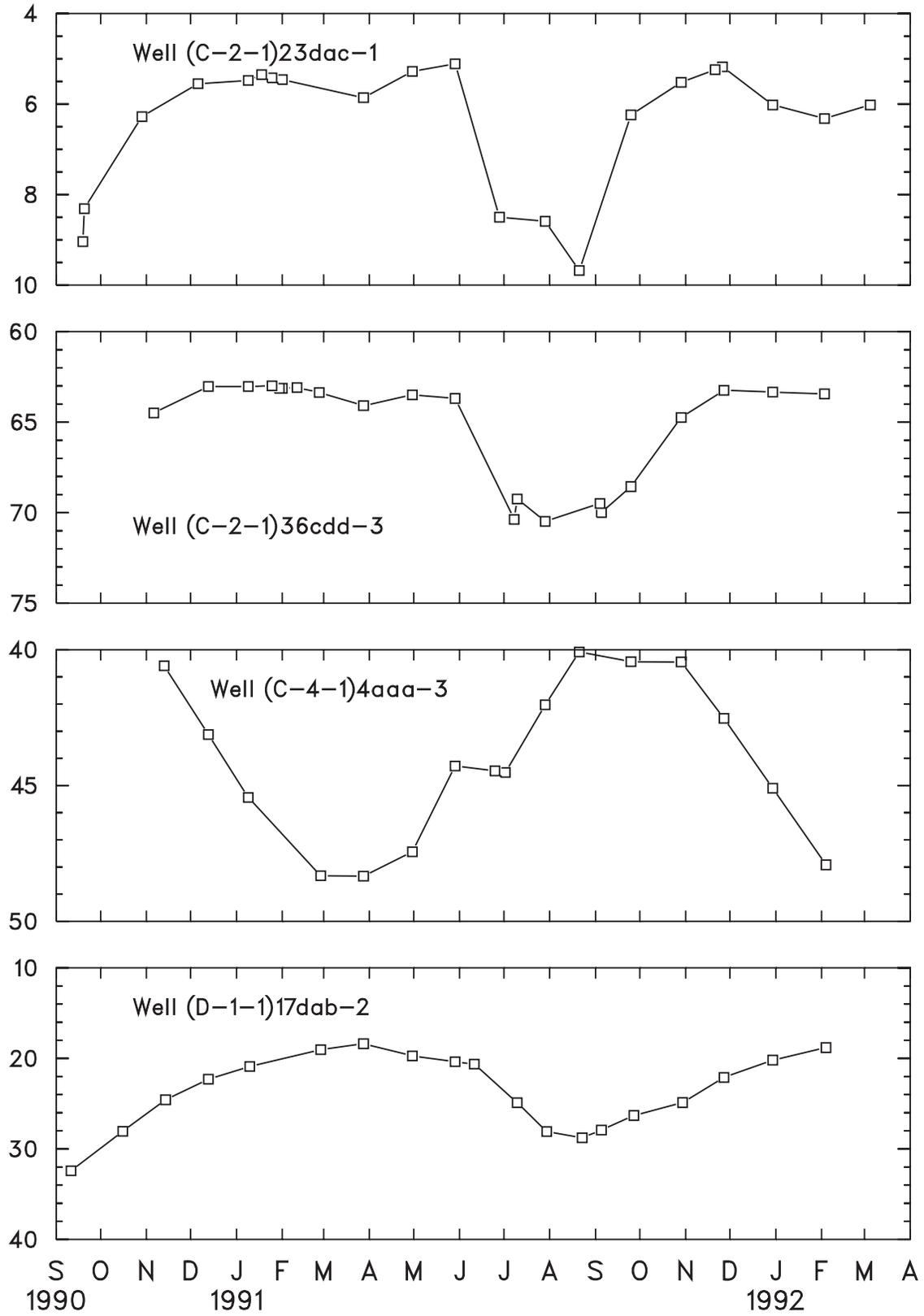


Figure 14. Seasonal water-level fluctuations in 12 monitoring wells completed in the shallow unconfined aquifer or a confining layer in Salt Lake Valley, Utah 1990–92—Continued.

WATER LEVEL, IN FEET ABOVE (-) OR BELOW LAND SURFACE

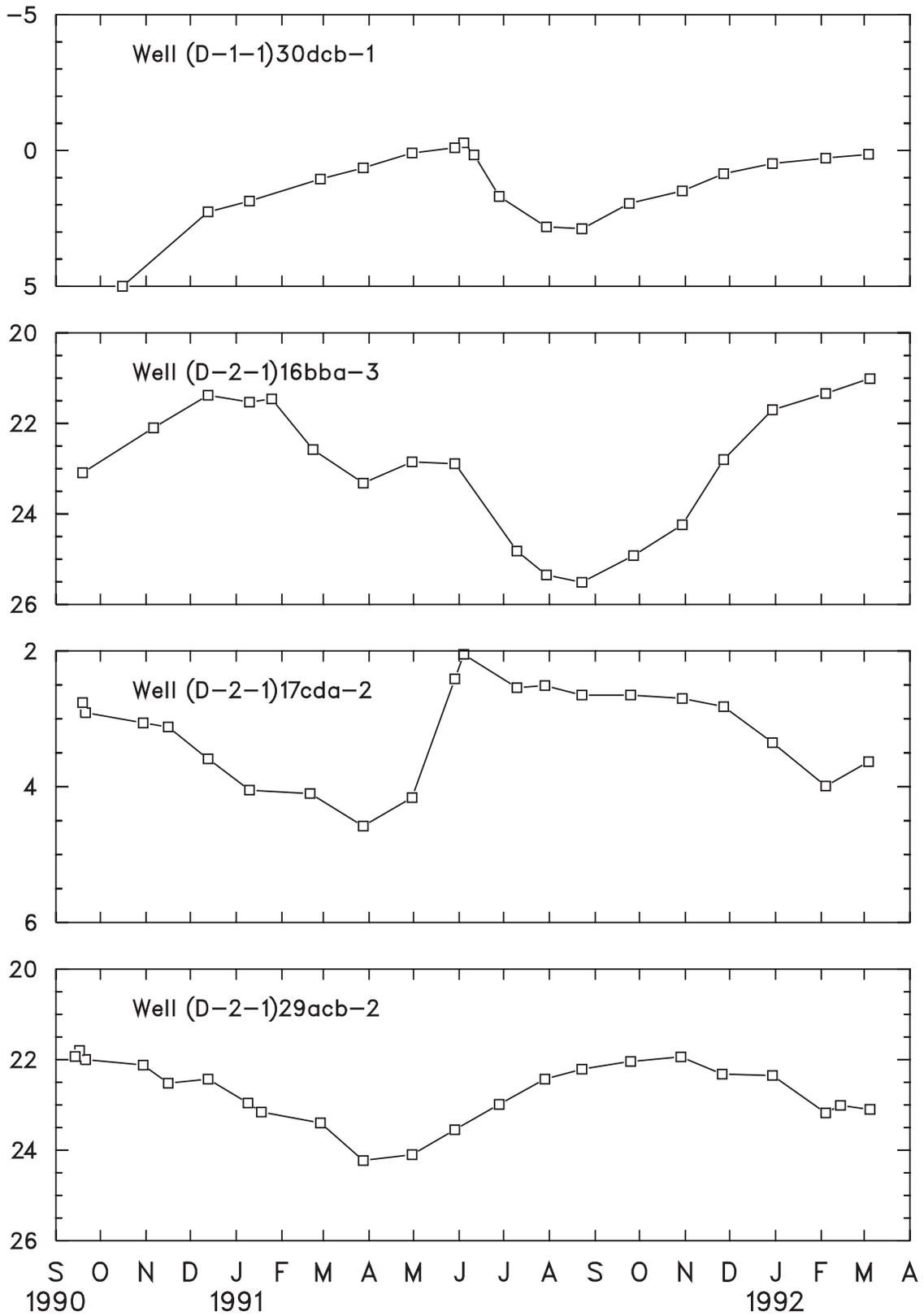


Figure 14. Seasonal water-level fluctuations in 12 monitoring wells completed in the shallow unconfined aquifer or a confining layer in Salt Lake Valley, Utah 1990–92—Continued.

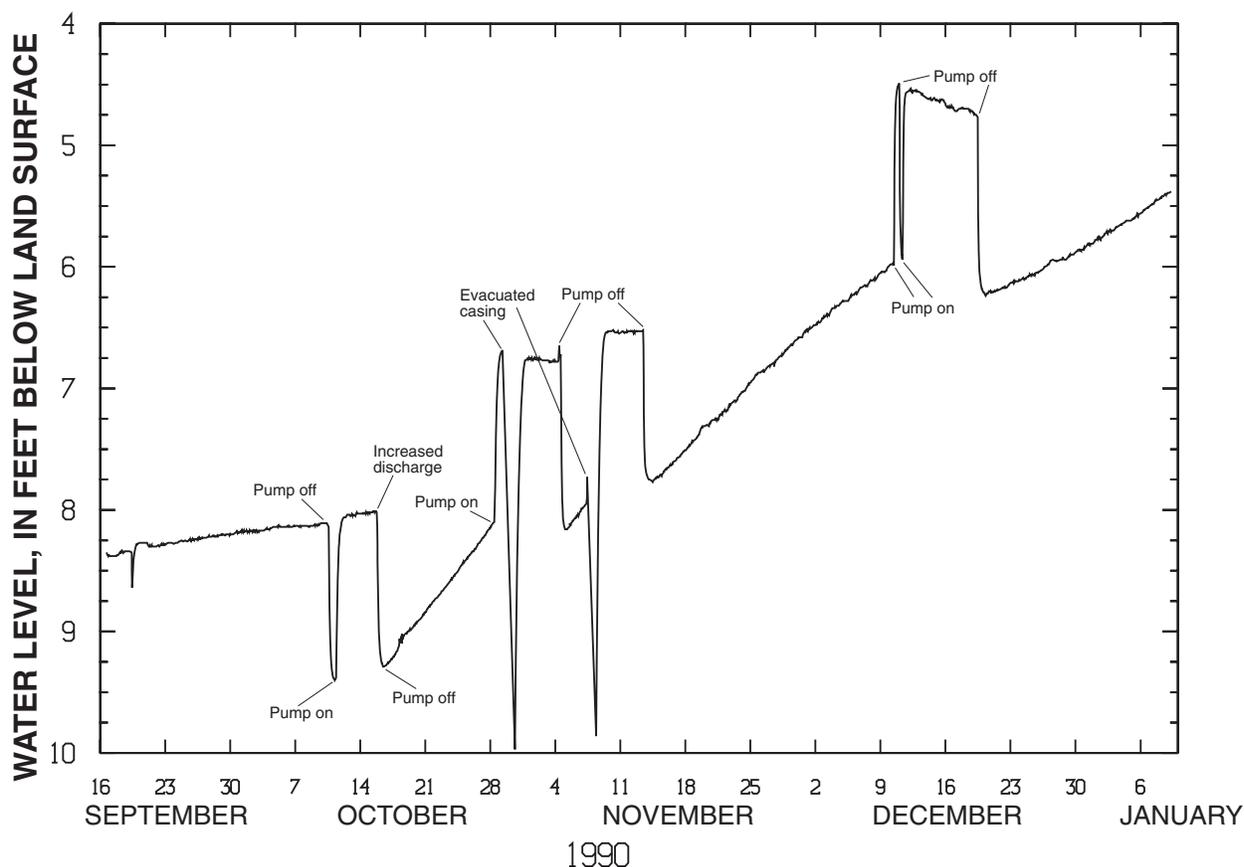


Figure 15. Reverse water-level fluctuations in monitoring well (C-1-1)23bcd-2 in Salt Lake Valley, Utah.

halite; and the release of 0.66 mmol/L carbon dioxide gas from water sampled from well (C-4-1)4aaa-3. Saturation indices calculated for the well water substantiate the reactions proposed by the model for calcite, gypsum, and halite.

Well (C-2-1)24bcd-1 (table 2) is finished in the principal aquifer and overlying confining layers with perforations from 179 to 300 ft below land surface. On the basis of vertical hydraulic-head gradients and the nature of confining layers in the area, water from this well might be a mixture of water from the Jordan River and water that exists in the principal aquifer farther up the flow path. Vertical hydraulic-head gradients between the shallow unconfined and principal aquifers probably are downward during the summer months when public-supply wells in the area are pumped.

The shallow unconfined aquifer can receive a substantial part of its recharge as seepage from the Jordan River and its diversions, and as infiltration of unconsumed irrigation water in areas where a downward hydraulic-head gradient and permeable material exist. The Jordan River and Little Cottonwood Creek

probably have removed part of the shallow confining layers deposited in the area around well (C-2-1)24bcd-1. This assumption was made on the basis of the deposits identified in wells drilled near the edges of and on benches above the Jordan River flood plain and the depth to which the river has cut down into the basin-fill material. Discontinuous confining layers do not tend to impede vertical ground-water flow between the shallow unconfined and principal aquifers as effectively as continuous confining layers, allowing the possibility for the mixing of water, if hydraulic-head gradients permit.

Analyses of water sampled from the Jordan River at 5800 South Street and from well (D-2-1)30dda-1, upgradient from well (C-2-1)24bcd-1 with several perforated intervals from 560 to 990 ft below land surface, were chosen to represent two waters from which the water from well (C-2-1)24bcd-1 could be derived. Concentrations of major ions are higher in water sampled from the Jordan River than in water sampled from the principal aquifer. Saturation indices indicate that water from the principal aquifer upgradient from well (C-2-1)24bcd-1 is undersaturated with calcite and gypsum. Water from the river and well (C-2-1)24bcd-1 is

at equilibrium and possibly slightly supersaturated with calcite and is undersaturated with gypsum. Isotopic concentrations for the Jordan River were determined from a sample collected at 6400 South Street. The oxygen-18 and deuterium ratios in water from well (C-2-1)24bcd-1 are lighter than the ratios in water from the Jordan River and heavier than the ratios in water from well (D-2-1)30dda-1.

Chloride, $\delta^{18}\text{O}$, and δD values were considered to be conservative and were used separately as constraints to determine the ratio of mixing waters needed to produce the evolved water. Similar mixing ratios of water from the Jordan River and from well (D-2-1)30dda-1 were calculated using these constraints. About 75 percent of the water sampled from well (C-2-1)24bcd-1 is from the principal aquifer upgradient from the well and about 25 percent is from the Jordan River, under the assumptions previously listed.

The chemistry of water from some wells sampled in the northwestern part of the valley might be controlled by sulfate reduction reactions. Water from well (B-1-2)21dcd-1 was used to represent the sodium-chloride and bicarbonate type waters in the area that have relatively low sulfate concentrations. The chemical analysis of water from well (C-1-1)9bba-3 represents water from the principal aquifer upgradient from the sodium-chloride and bicarbonate type water.

The constraints used in the geochemical modeling process included total inorganic carbon, calcium, sulfur, sodium, and redox state. The phases assumed applicable in the subsurface were calcite, gypsum, halite, pyrite, and carbon in the form of organic matter (CH_2O). The reactions needed to produce the evolutionary downgradient water sampled from well (B-1-2)21dcd-1 include the precipitation of calcite (0.77 mmol/L) and pyrite (0.47 mmol/L), the dissolution of gypsum (0.42 mmol/L) and halite (7.0 mmol/L), and the consumption of CH_2O (1.4 mmol/L). Saturation indices indicate that water from well (B-1-2)21dcd-1 is saturated with calcite and undersaturated with gypsum and halite. Sulfate reduction is a plausible explanation for the relatively low sulfate and high bicarbonate concentrations in water from the northwestern part of the valley on the basis of this model and other evidence presented in the "Chemical composition of ground water" section of this report.

Isotopic Indicators of Ground-Water Movement

The stable isotopes of oxygen and hydrogen occur naturally in water and can be used to help determine sources of recharge and the extent of mixing in a ground-water system. Water sampled from 81 hydrologic-data sites, mainly from the shallow unconfined aquifer and underlying confining layers in Salt Lake Valley, was analyzed for oxygen and hydrogen isotope ratios ($\delta^{18}\text{O}$ and δD) from 1990 to 1992 (tables 6 and 7).

Water samples from four surface-water sites (table 6) and 20 wells (table 7) were analyzed for tritium in 1990 and 1991 to better understand ground-water movement in the southeastern part of the valley. Most of the water samples analyzed for tritium were collected from sites along hypothesized flow paths extending from the mountain front on the southeastern side of the valley to the Jordan River in the valley center (pl. 1). Some tritium data also were available for water collected in 1989 from other areas in the valley (table 7).

Oxygen-18 and Deuterium in Surface Water

Selected mountain-front streams and the Jordan River at the Jordan Narrows were sampled for $\delta^{18}\text{O}$ and δD at peak-flow conditions (June 1991) and what was considered to be base-flow conditions (December 1991) (table 6) to better define two known sources of recharge to the ground-water system. Values of $\delta^{18}\text{O}$ and δD were available for intermittent samples from Red Butte Creek at gaging station 10172200 and the Jordan River at 1700 South Street (gaging station 10171000) from 1984 to 1987 (Carol Kendall, U.S. Geological Survey, written commun., 1991).

The water in mountain-front streams was assumed to be representative of precipitation runoff, and in addition to water in the mountain-front consolidated rock, was assumed to be a primary source of recharge to the basin-fill material. The consolidated rock ground-water system is recharged by precipitation and discharges partially to the mountain-front streams. Water samples were collected from Red Butte Creek, Mill Creek, and Big and Little Cottonwood Creeks in the Wasatch Range, and from Butterfield Creek in the Oquirrh Mountains. Values of $\delta^{18}\text{O}$ and δD ranged from -17.15 per mil and -128.0 per mil, respectively, for water from Mill Creek at the canyon mouth to -16.00

Table 6. Isotopic determinations and physical properties of water from selected surface-water sites in Salt Lake Valley, Utah

[—, no data available]

Location: Name of surface-water site and location relative to a canyon mouth or street. See plate 1 for location.

Site number: 15-digit U.S. Geological Survey site identification number based on latitude and longitude or an 8-digit station number assigned to streams based on tributary rank and downstream order.

Tritium: TU, tritium units.

 $\delta^{18}\text{O}$: Oxygen-18/oxygen-16 stable isotope ratio, in per mil. δD : Hydrogen-2/hydrogen-1 stable isotope ratio, in per mil.

Temperature: °C, degrees Celsius, measured in the field.

Specific conductance: $\mu\text{S}/\text{cm}$, microsiemens per centimeter at 25 degrees Celsius, measured in the field.Alkalinity: mg/L, milligrams per liter as CaCO_3 , measured in the field.Streamflow: ft^3/s , cubic feet per second; M, measured in the field or reported from streamflow-gaging station records.

Location	Site number	Date sampled	Tritium (TU)	$\delta^{18}\text{O}$ (per mil)	δD (per mil)	Temperature (°C)	Specific conductance ($\mu\text{S}/\text{cm}$)	pH (standard units)	Alkalinity, total (mg/L as CaCO_3)	Streamflow (ft^3/s)
Salt Lake City Aqueduct at Southeast Regional Water Treatment Plant ¹	403234111485001	02-17-92	—	-16.60	-124.0	—	—	—	—	—
East Jordan Canal at 7700 South 700 East	403642111521803	07-09-91	13.9	-8.60	-79.5	22.0	1,690	8.2	—	—
Big Cottonwood Creek at 4500 South 300 East		² 08-04-92	19.7	—	—	—	—	—	—	—
Big Cottonwood Creek above water treatment plant near canyon mouth	403706111463901	12-04-90	—	-16.40	-122.0	6.0	315	8.5	—	—
Big Cottonwood Creek downstream from canyon mouth	10168499	06-06-91	—	-16.70	-122.0	8.0	155	8.0	52	269 M
		12-17-91	—	-16.25	-120.0	4.0	315	8.0	96	11.6
Little Cottonwood Creek near canyon mouth	10167499	06-06-91	9.0	-16.60	-122.0	9.0	155	7.8	40	148 M
		12-18-91	—	-16.00	-118.0	4.5	410	7.7	110	.16
Mill Creek near canyon mouth	10169999	06-06-91	—	-16.95	-123.0	7.0	435	8.3	154	41 M
		12-17-91	—	-17.15	-128.0	4.0	660	8.1	216	6.5
Red Butte Creek upstream from canyon mouth	10172200	11-15-90	—	-16.60	-122.0	5.0	630	8.6	245	1.0 M
Butterfield Creek	402922112092101	06-07-91	—	-16.00	-118.0	14.0	320	8.3	135	1.4
		12-16-91	—	-16.45	-123.0	4.0	580	7.8	242	—
Jordan River at Jordan Narrows	402629111551201	06-07-91	12.7	-10.60	-89.5	18.0	1,470	8.3	227	—
		12-18-91	—	-15.00	-116.0	2.0	2,000	8.1	282	—
Jordan River at 12300 South	403138111550600	12-04-90	—	-12.30	-103.0	6.0	2,000	8.3	—	—
Jordan River at 6400 South	403750111552100	06-20-91	19.8	-12.55	-102.0	17.5	1,840	8.0	255	—
Jordan River at 1700 South	10171000	11-29-90	—	-13.60	-108.0	—	—	—	—	116
Jordan River at 500 North	404657111560801	11-29-90	—	-15.10	-109.0	—	—	—	—	—

¹ Water enters Salt Lake City Aqueduct from Deer Creek Reservoir.² Data from Craig Forester, University of Utah (written commun., 1992).

per mil and -118.0 per mil, respectively, for water from Little Cottonwood Creek at the canyon mouth and from Butterfield Creek (table 6).

The variation in these values for water from mountain-front streams probably is caused by differences in source-area altitude, temperature, and resulting type of precipitation, and the quantity and location of infiltration into the consolidated-rock ground-water system. Precipitation undergoes some degree of evaporation while falling from clouds, resulting in isotopic enrichment. Values of δD measured in precipitation collected in the eastern part of Salt Lake Valley during

a 3-year period ranged from -11 per mil in July 1989 to -213 per mil in March 1989 (Jim Ehleringer, University of Utah, written commun., 1992). Evaporation of water from sampled streams is probably negligible because of the short transient times in the mountain watersheds.

Peak snowmelt runoff in the streams is likely a mixture of precipitation accumulated during the winter and discharge from the consolidated-rock ground-water system. Ground-water discharge that contributes to streamflow all year is probably a composite of winter and summer precipitation recharged at a range of altitudes. Water in Mill Creek and Butterfield Creek that

Table 7. Isotopic determinations and physical properties of water from selected ground-water sites in Salt Lake Valley, Utah

[—, no data available]

Location: See figure 2 for an explanation of the numbering system used for hydrologic-data sites in Utah.

Site number: 15-digit U.S. Geological Survey site identification number based on latitude and longitude.

Tritium: TU, tritium units; <, less than.

 $\delta^{18}\text{O}$: Oxygen-18/oxygen-16 stable isotope ratio, in per mil. δD : Hydrogen-2/hydrogen-1 stable isotope ratio, in per mil.

Temperature: °C, degrees Celsius, measured in the field.

Specific conductance: $\mu\text{S}/\text{cm}$, microsiemens per centimeter at 25 degrees Celsius, measured in the field.Alkalinity: mg/L, milligrams per liter as CaCO_3 , measured in the field.

Location	Site number	Depth of well (feet)	Date sampled	Tritium (TU)	$\delta^{18}\text{O}$ (per mil)	δD (per mil)	Temperature (°C)	Specific conductance ($\mu\text{S}/\text{cm}$)	pH (standard units)	Alkalinity, total (mg/L as CaCO_3)
(A-1-1)31cac-1	404627111532601	464	06-19-89	<8.1	—	—	14.0	960	7.3	—
(B-1-2)19aca-1	404826112062201	450	08-05-91	—	-17.40	-131.0	17.5	2,350	8.4	380
(B-1-2)19aca-2	404829112062101	86.0	08-05-91	—	-17.55	-135.0	22.0	3,550	8.3	874
(B-1-2)19aca-3	404829112062102	23.1	08-05-91	—	-15.05	-120.0	14.5	30,000	7.8	999
(B-1-2)21dcd-1	404754112040501	561	08-06-91	—	-17.50	-136.0	16.5	1,180	8.6	240
(B-1-2)22daa-1	404815112024201	148	08-07-91	—	-17.65	-135.0	15.5	1,670	8.2	261
(B-1-2)22dad-1	404812112024201	15.5	08-07-91	—	-17.00	-130.0	14.5	2,270	8.5	483
(B-1-2)23cbd-1	404810112021901	270	08-07-91	—	-17.55	-134.0	17.0	1,780	8.0	222
(B-1-2)28aab-1	404753112040201	15.0	08-06-91	—	-11.80	-97.5	16.5	3,580	8.6	490
(B-1-2)32bec-1	404645112055401	716	11-29-90	—	-17.70	-136.0	—	—	—	—
(B-1-2)32bec-2	404643112060001	248	11-08-90	—	-16.90	-134.0	11.0	14,000	7.8	240
			08-06-91	—	-16.95	-132.0	20.0	13,000	7.8	253
(B-1-2)32bec-3	404643112060002	24.7	08-06-91	—	-16.85	-131.0	15.0	11,200	8.0	354
(B-1-2)36baa-1	404659112005601	464	05-25-89	<8.1	—	—	27.5	6,300	7.3	—
			08-29-91	—	-17.30	-133.0	27.0	6,180	7.7	118
(B-1-2)36baa-2	404700112005701	83.0	08-29-91	—	-17.25	-130.0	16.0	890	8.1	163
(B-1-2)36baa-3	404700112005703	19.0	08-29-91	—	-17.05	-132.0	15.0	6,780	8.6	334
(C-1-1)15cad-2	404351111562301	40.0	10-15-90	—	-16.10	-120.0	17.5	1,200	7.8	257
(C-1-1)15cad-3	404351111562302	16.0	09-03-91	—	-13.65	-106.0	22.0	2,940	7.3	499
(C-1-1)18bda-1	404406111594901	45.0	10-22-90	—	-14.90	-116.0	14.0	16,000	7.7	256
(C-1-1)23bec-1	404314111553601	840	07-13-89	9.1	—	—	15.0	820	7.9	—
(C-1-1)23bec-2	404313111553202	66.1	11-08-90	—	-16.70	-125.0	13.0	720	8.1	343
(C-1-1)25bdb-1	404223111541201	1,000	07-10-89	<8.1	—	—	18.5	540	7.7	—
(C-1-1)27dda-8	404156111554401	775	07-10-89	<8.1	—	—	21.0	710	7.7	—
(C-1-1)33ddd-2	404057111565402	51.5	10-23-90	—	-14.60	-122.0	14.0	3,580	7.3	381
(C-1-1)35ddd-5	404057111543901	213	11-09-90	—	-16.70	-125.0	13.5	370	8.5	103
(C-1-2)14cdd-1	404335112020601	480	08-19-91	—	-16.50	-125.0	21.0	3,420	7.8	124
(C-1-2)14cdd-2	404335112020501	49.5	08-19-91	—	-14.95	-117.0	20.0	1,690	8.0	176
(C-1-2)14cdd-3	404335112020502	17.7	08-19-91	—	-14.25	-112.0	15.0	6,980	7.6	897
(C-1-2)22cbb-1	404259112034301	110	06-01-89	31.2	—	—	14.5	2,160	7.7	—
(C-2-1)13abc-1	403859111535601	56.5	11-06-90	—	-13.30	-108.0	13.0	2,150	7.9	316
(C-2-1)13dbb-1	403840111535301	445	06-20-91	4.7	-16.45	-122.0	16.0	280	8.2	87
(C-2-1)20aad-1	403810111580501	658	07-10-89	<8.1	—	—	15.0	2,260	7.3	—
(C-2-1)20aad-2	403810111580502	61.0	10-29-90	—	-11.10	-98.0	15.0	3,970	7.1	404
(C-2-1)23ccb-1	403737111554101	41.0	10-23-90	—	-9.70	-90.0	16.0	1,850	7.4	311
(C-2-1)23dac-1	403742111544301	85.0	10-29-90	52.8	-13.00	-106.0	13.5	1,490	7.4	315
(C-2-1)23dad-1	403743111544000	391	06-20-91	57.0	-13.90	-109.0	15.5	1,110	7.6	165
(C-2-1)24bcd-1	403756111542301	370	12-23-91	30.8	-15.15	-115.0	15.5	615	7.7	124
(C-2-1)34bbd-2	403624111563602	41.5	10-23-90	—	-11.10	-100.0	14.5	2,310	7.2	295
(C-2-1)36cdd-2	403546111540601	62.0	11-14-90	—	-9.60	-90.0	13.0	2,320	7.6	316

Table 7. Isotopic determinations and physical properties of water from selected ground-water sites in Salt Lake Valley, Utah—Continued

Location	Site number	Depth of well (feet)	Date sampled	Tritium (TU)	$\delta^{18}\text{O}$ (per mil)	δD (per mil)	Temperature ($^{\circ}\text{C}$)	Specific conductance ($\mu\text{S}/\text{cm}$)	pH (standard units)	Alkalinity, total (mg/L as CaCO_3)
(C-2-1)36cdd-3	403546111540602	129.5	11-06-90	—	-14.60	-114.0	12.5	780	8.1	128
(C-2-2)1bcd-2	404029112011202	28.0	10-22-90	—	-14.00	-116.0	14.0	1,060	7.9	306
(C-3-1)12ccb-1	403408111543201	118	06-06-89	<8.1	—	—	20.0	970	7.4	—
(C-3-1)12cdb-1	403409111541501	41.0	11-13-90	—	-9.90	-89.0	12.5	2,310	7.3	464
(C-4-1)4aaa-3	403027111565701	49.0	11-13-90	—	-7.20	-78.0	14.5	1,800	7.4	220
(C-4-1)4aaa-4	403025111565801	104.0	09-03-91	—	-12.00	-98.5	17.5	1,780	7.4	317
(C-4-2)1bbb-1	403027112012401	540	06-06-89	<8.1	—	—	14.5	1,390	7.4	—
(D-1-1)1aac-S1	404505111475201	—	06-13-89	29.1	—	—	11.5	970	7.2	—
(D-1-1)17dab-1	404355111511801	215	11-14-90	—	-16.20	-122.0	13.5	360	10.9	70
(D-1-1)17dab-2	404355111511802	68.5	10-16-90	—	-16.10	-126.0	14.0	810	7.4	254
(D-1-1)19dbc-2	404258111525102	31.0	10-15-90	—	-16.10	-121.0	14.0	1,430	7.2	451
(D-1-1)21 acc-2	404312111502901	576	06-28-89	10.9	—	—	14.0	950	7.6	—
(D-1-1)30deb-1	404155111524601	71.0	10-16-90	—	-15.20	-118.0	12.5	1,030	7.8	282
(D-1-2)36dab-S1	404121111395601	—	¹ 1988	—	-17.60	-132.0	—	—	—	—
(D-2-1) 2cdc-S1	404008111482601	—	07-10-89	<8.1	—	—	10.5	490	7.5	—
(D-2-1)6dbb-2	404027111524701	590	05-31-91	—	-16.15	-120.0	—	—	—	—
(D-2-1)16bba-3	403910111505502	90.0	11-06-90	—	-15.60	-119.0	12.5	610	7.9	180
(D-2-1)17cda-1	403829111514301	440	06-21-91	31.3	-16.25	-120.0	12.0	410	8.0	130
(D-2-1)17cda-2	403829111514401	28.5	10-30-90	—	-12.90	-107.0	15.0	1,180	7.4	277
(D-2-1)17cda-3	403829111514402	23.5	10-30-90	—	-12.90	-104.0	15.5	1,080	7.2	255
(D-2-1)21dbc-1	403742111503201	740	06-07-89	28.1	—	—	12.0	260	7.9	—
(D-2-1)28ccc-1	403638111511701	691	09-18-90	—	-16.50	-122.0	12.0	240	8.2	82
			06-20-91	1.9	—	—	12.0	235	8.2	80
(D-2-1)29acb-2	403706111514201	52.0	10-30-90	15.9	-15.60	-115.0	13.5	680	7.7	181
(D-2-1)29acb-8	403706111514101	500	09-17-90	0.8	-16.40	-123.0	14.0	225	8.3	78
(D-2-1)30dda-1	403642111521701	1,002	10-22-91	8.2	-15.95	-119.0	17.0	330	8.0	93
(D-2-1)30dda-2	403642111521801	65.0	08-30-91	35.4	-14.85	-112.0	14.0	740	7.8	130
(D-2-1)30dda-3	403642111521802	21.4	08-30-91	16.2	-11.20	-93.5	14.5	1,340	7.6	189
(D-2-1)32add-1	403610111510901	685	05-31-91	—	-15.70	-119.0	—	—	—	—
(D-2-1)34cdc-1	403543111494801	875	09-18-90	—	-16.10	-121.0	12.0	375	7.6	138
			06-20-91	58.9	-16.00	-119.0	12.0	375	7.6	137
(D-2-1)34dba-1	403606111491601	700	06-20-91	33.3	-16.20	-120.0	11.0	320	7.5	88
(D-2-1)35bbb-1	403332111485000	238	06-21-91	28.3	-16.35	-120.0	14.0	320	8.0	134
(D-3-1)3ccc-1	403451111495401	684	09-30-91	38.7	-15.75	-118.0	11.0	405	7.9	144
(D-3-1)4bbb-1	403540111510601	904	05-31-91	—	-15.75	-119.0	—	—	—	—
(D-3-1)4cad-1	403506111503601	938	09-17-90	—	-15.80	-124.0	11.5	310	8.1	121
(D-3-1)5bda-2	403523111514801	650	11-26-91	3.3	-16.00	-119.0	12.5	300	8.1	106
(B-3-1)6dad-1	403510111521801	1,000	09-17-90	—	-16.20	-118.0	17.0	415	8.1	93
(D-3-1)9aab-1	403444111501601	950	06-20-91	14.8	-15.85	-118.0	12.0	340	7.7	126
(D-3-1)10dab-1	403423111490001	757	09-30-91	40.3	-15.75	-118.0	12.5	405	7.9	112
(D-3-1)12bdd-1	403426111471201	171	09-17-90	26.3	-15.90	-118.0	14.0	345	7.0	93
(D-3-1)15cbc-1	403323111495501	640	06-25-91	39.5	-15.85	-120.0	—	280	7.5	—
(D-3-1)29ddd-1	403126111511101	48.0	11-13-90	—	-14.60	-114.0	15.5	960	7.6	269
² (D-3-2)7acd-S1	403426111454601	—	¹ 1988	—	-15.45	-119.0	—	—	—	—

¹ Water sample collected in the summer of 1988 and stored in sealed glass bottle (Alan L. Mayo and Mark Loucks, Brigham Young University, written commun., 1991). The sample was sent to a U.S. Geological Survey laboratory for analysis in 1992.

² Water sample collected from a borehole through granite at the Archive Vault of the Church of Jesus Christ of Latter-Day Saints in Little Cottonwood Canyon.

was sampled in December 1991 had lower $\delta^{18}\text{O}$ and δD values than water from the same creeks that was sampled in June 1991. The lower isotopic values measured during base-flow conditions may be the result of ground-water inflow to the stream from more permeable formations at higher altitudes. This water would be isotopically lighter than snowmelt runoff that occurred in June, which would be a mixture of precipitation that has fallen over a range in altitudes.

Another major source of recharge to the basin-fill ground-water system is seepage of water from canals and from infiltration of unconsumed irrigation water. This recharge source mostly originates from the Jordan River. The outflow from Utah Lake in Utah Valley is the Jordan River, which enters Salt Lake Valley at the Jordan Narrows, an erosional breach through the Traverse Mountains. Although Utah Lake receives most of its water from central Wasatch Range streams, the lake water undergoes some evaporation. The resulting increase in dissolved-solids concentration and enrichment in the heavy isotopes of oxygen and hydrogen in water from Utah Lake is very different from that of water in the mountain-front streams of Salt Lake Valley. Water from the Jordan River at the Jordan Narrows had a $\delta^{18}\text{O}$ of -10.60 per mil and a δD of -89.5 per mil in June 1991 (table 6). Water from the same site collected in December 1991 had a $\delta^{18}\text{O}$ of -15.00 per mil and a δD of -116.0 per mil, indicating that the water sampled in December had undergone less evaporation than the water sampled in late spring. Streamflow from the East Jordan Canal sampled at (D-2-1)30dda in July 1991 also had more heavier isotopes of oxygen and hydrogen relative to water in the surrounding mountain-front streams. The canal contains water diverted from the Jordan River near the Jordan Narrows.

Values of $\delta^{18}\text{O}$ and δD for mountain-front streams in Salt Lake Valley produce a local meteoric water line that is similar to the global meteoric water line (fig. 16) in which the slope is 8 and the intercept is 10 per mil (Craig, 1961). The global meteoric water line is used to represent the area because of a lack of data. Water from different sites on the Jordan River in Salt Lake Valley and from the East Jordan Canal at (D-2-1)30dda plot on a line that diverges from the global meteoric water line. The slope of this line is less than the slope of the meteoric water line and may be indicative of evaporation. Water from most sites on the Jordan River has some component of ground-water or mountain-front tributary inflow. The water sample from the East Jordan Canal represents water from Utah

Lake during the midsummer, a time of high evaporation rates. Water from Great Salt Lake, the terminus of the Jordan River, had a $\delta^{18}\text{O}$ of -4.1 per mil and a δD of -68 per mil on July 25, 1989 (Mayo and Klauk, 1991, p. 317). Great Salt Lake has no outlet and discharges water only through evaporation.

Oxygen-18 and Deuterium in Ground Water

Oxygen and hydrogen isotopic ratios were determined for ground water collected from the principal aquifer and the shallow unconfined aquifer in Salt Lake Valley (table 7). Values of $\delta^{18}\text{O}$ and δD for ground water from the principal aquifer in the southeastern part of the valley (Townships 2 and 3 South, Range 1 East) plot near the global meteoric water line (fig. 17). Ground water in this area is similar in isotopic composition to water sampled from Big and Little Cottonwood Creeks at the canyon mouths.

Ground water from wells completed in the principal aquifer south of Little Cottonwood Creek is generally more enriched in oxygen-18 and deuterium than water in wells closer to and north of the creek (pl. 1). This may be an indication of where the ground water was recharged along the Wasatch Front. The drainage areas of Big and Little Cottonwood Creeks are much larger, extend farther east into the Wasatch Range, and include more area at higher altitudes than the smaller mountain drainages to the south. Precipitation falling at higher altitudes and toward the interior of the range is more depleted in the heavier isotopes.

Precipitation falling on the mountain block is the source of water in the streams and in the consolidated-rock ground-water system. Recharge to the unconfined part of the principal aquifer can occur as seepage at land surface or as subsurface flow from consolidated-rock fractures into the basin-fill material. Because of the thickness of the unsaturated basin-fill material near the mountain front, the relative quantity of water recharged by each method is not known. Additional isotopic data for precipitation falling on and water discharging from consolidated rock in the mountains bordering the basin fill, and for streamflow at the mountain front could help define the mechanism of recharge.

On the basis of $\delta^{18}\text{O}$ and δD values, water from wells (C-2-1)23dad-1 and (C-2-1)24bcd-1, finished in the principal aquifer, contains a component of water that has undergone some degree of evaporation. The isotopic composition of a water that is a mixture from

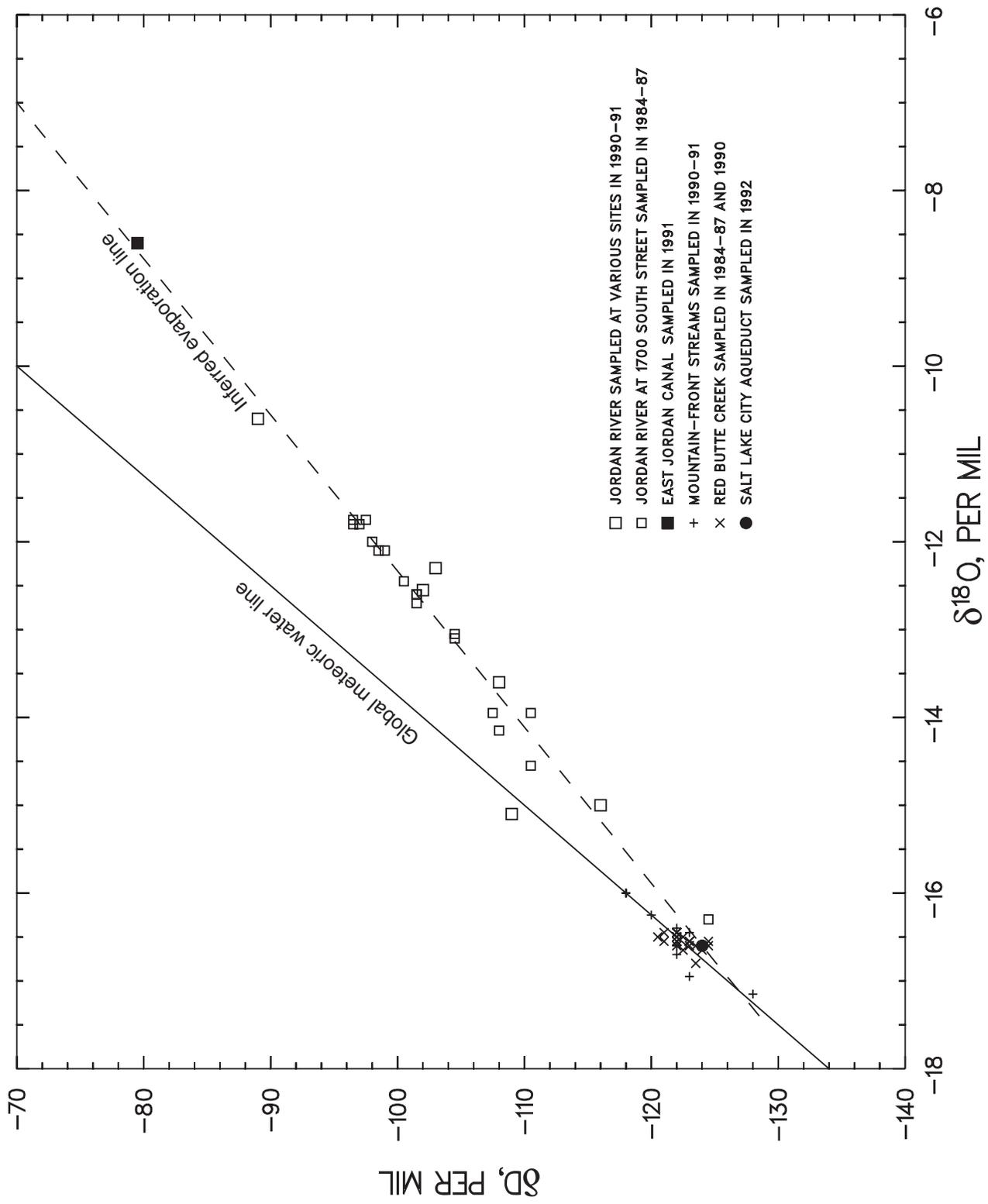


Figure 16. Relation between δD and $\delta^{18}O$ values for water sampled from selected surface-water sites in Salt Lake Valley, Utah.

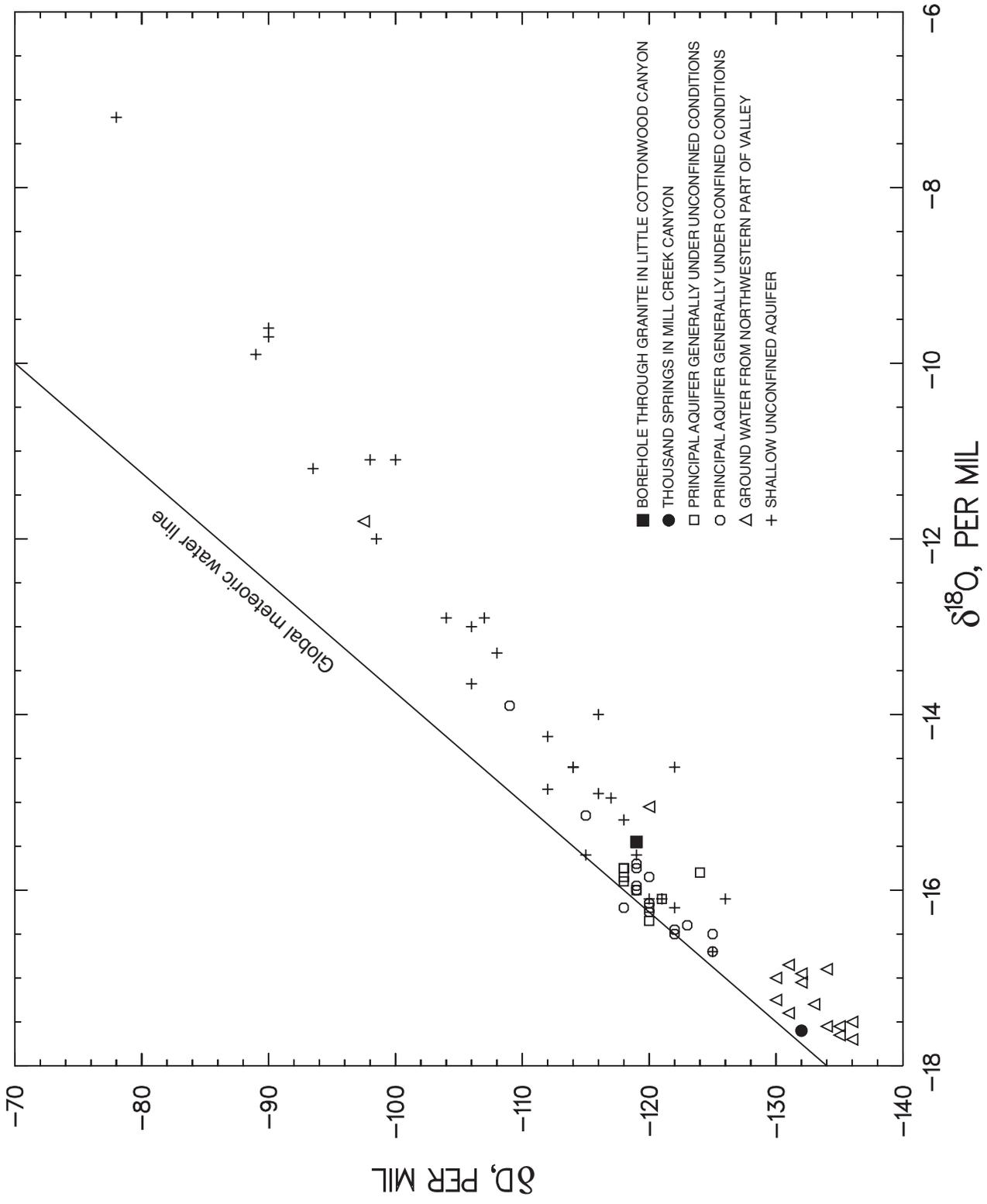


Figure 17. Relation between δD and $\delta^{18}O$ values for water sampled from selected ground-water sites in Salt Lake Valley, Utah.

two or more sources with different isotopic compositions can be determined by isotopic mass balance:

$$\delta_1 n_1 + \delta_2 n_2 + \dots = \delta f (n_1 + n_2 + \dots) \quad (7)$$

where δ_1 is the δ value of source 1,
 n_1 is the number of milliequivalents per liter of source 1, and $n_1 + n_2 + \dots = 1$, and
 δf is the δ value of the product.

Two possible sources of water that can mix to form the water collected from well (C-2-1)23dad-1 ($\delta^{18}\text{O} = -13.90$ per mil, $\delta\text{D} = -109.0$ per mil) (table 7) are water from the Jordan River at 6400 South Street ($\delta^{18}\text{O} = -12.55$ per mil, $\delta\text{D} = -102.0$ per mil) (table 6) and water from the principal aquifer in the area near well (D-2-1)29acb-8 ($\delta^{18}\text{O} = -16.40$ per mil, $\delta\text{D} = -123.0$ per mil) (table 7). If these two sources of water are assumed to make up water from well (C-2-1)23dad-1, then the isotopic mass-balance equation would indicate that water from well (C-2-1)23dad-1 is a mixture of two-thirds Jordan River water and one-third water from the principal aquifer upgradient from the well. Downward movement of water from the shallow unconfined aquifer to the principal aquifer is occurring in this area of the valley if these proportions are accurate. The confining layer that separates the shallow unconfined aquifer from the principal aquifer is relatively thin at this location and may be laterally discontinuous, resulting in the vertical movement of water between aquifers.

Most of the ground water sampled from the northwestern part of Salt Lake Valley, regardless of depth, is more depleted in the heavier isotopes than is water sampled from other wells in the valley and from adjacent mountain-front streams (fig. 16). Values of δD for most of the ground water sampled in the northwestern part of the valley are 8 to 14 per mil less than those measured in Little and Big Cottonwood Creeks during spring runoff. Two possible reasons for the lighter isotopic ratios of ground water from the northwestern part of the valley are the source and the time of recharge.

Values of $\delta^{18}\text{O}$ and δD similar to those of ground water in the northwestern part of the valley were determined for water discharging from Thousand Springs [(D-1-2)36dab-S1] (fig. 17), at an altitude of about 7,240 ft in Mill Creek Canyon in the Wasatch Range. Similar δD values also were measured for water from springs discharging from consolidated rocks in Big Cottonwood Canyon (Jim Ehleringer, University of Utah, oral commun., 1992). Precipitation falling at the higher altitudes in the Wasatch Range may seep into fractures in the consolidated rock that extend beneath

the basin-fill material in Salt Lake Valley. The fractures may intersect normal faults in the valley that are associated with crustal extension in the Basin and Range Province. The water may leak from the consolidated rock into overlying basin-fill material along these steeply dipping faults. This concept of water movement was proposed for ground water warmer than 20°C in the East Shore area of Great Salt Lake, north of Salt Lake Valley, by Cole (1982, p. 592).

Warm ground water is present in the northwestern part of Salt Lake Valley. Well (B-1-2)36baa-1 is in the center of a series of fault scarps mapped at the northern extension of the Granger fault (fig. 1) (Persenius and Scott, 1990). The well is 464 ft deep with a water temperature of 27.0°C and lower values of $\delta^{18}\text{O}$ and δD relative to ground water sampled in the rest of the valley (table 7). Water in this well may have moved upward into basin-fill material along nearby faults.

Another possible reason for the lower values of $\delta^{18}\text{O}$ and δD in ground water from the northwestern part of the valley relative to water in the mountain-front streams is that the water may have entered the basin-fill material at the mountain fronts during climatic conditions that were cooler and more humid than those of today. Cool, wet conditions have occurred many times in the past several thousand years. More information is needed on the age of ground water in the northwestern part of the valley to investigate this possibility.

On a linear plot of $\delta^{18}\text{O}$ and δD , the isotopic ratio of mixtures with different proportions of two water sources lie on a straight line between the $\delta^{18}\text{O}$ and δD values of the sources. This is caused by the additive relation between the isotopic ratios of the mixing and evolutionary waters. Most of the ground water sampled from the shallow unconfined aquifer in the valley had $\delta^{18}\text{O}$ and δD values that plot along or near the evaporation line that represents water from the Jordan River and its diversions (fig. 16). This water is isotopically heavier because of evaporation. On this basis, it is concluded that the shallow unconfined aquifer receives substantial recharge from water diverted for irrigation from the Jordan River. The other major source of recharge is upward leakage from the confined part of the principal aquifer.

Water from wells (C-1-1)15cad-2, (C-1-1)23bcd-2, (D-1-1)19dbc-2, (D-2-1)16bba-3, and (D-2-1)29acb-2 (table 7) showed little or no shift from the meteoric water line (fig. 17). The five wells range in depth from 31 to 90 ft. An upward hydraulic-head gradient exists between the principal aquifer and the overlying confin-

ing layers in which wells (C-1-1)15cad-2 and 23bcd-2 are finished. The vertical hydraulic-head gradient between the shallow unconfined and principal aquifers near well (D-1-1)19dbc-2 is reversed at times because of ground-water withdrawals from the principal aquifer. In this central area of the valley, water from the shallow unconfined aquifer is typically recharged by water moving up from the principal aquifer, and the relatively lighter isotopic ratios are consistent with this conclusion.

A downward hydraulic-head gradient exists between the shallow unconfined and principal aquifers at wells (D-2-1)16bba-3 and (D-2-1)29acb-2 in the secondary recharge area of the valley (pl. 1). Values of $\delta^{18}\text{O}$ and δD in water from well (D-2-1)16bba-3 are similar to those in water from the principal aquifer in the area despite the downward vertical hydraulic-head gradient between the shallow unconfined and principal aquifers at the site. Big Cottonwood Creek may be the source of recharge to both aquifers, resulting in similar isotopic ratios.

Well (D-2-1)29acb-2 is upgradient from any diversions that transport water from the Jordan River but is next to a ditch that carries water from Little Cottonwood Creek. The water sampled from this well is slightly heavier in $\delta^{18}\text{O}$ and δD than is water from the principal aquifer in the area but plots along the global meteoric water line (fig. 17). This implies that the source of recharge originated at lower altitudes and at higher temperatures and may reflect more summer precipitation and base flow originating from consolidated rocks.

The stable isotope ratios of oxygen and hydrogen were used to determine if increases in dissolved-solids concentration in ground water are caused by evaporation, mineral dissolution, or both. The relation between the oxygen isotopic ratio and chloride concentration in water from selected wells in the valley is shown in figure 18. Water that plots on the left side of the graph with $\delta^{18}\text{O}$ values ranging from about -16 to -7 per mil and chloride concentrations from about 10 to 300 mg/L is affected by either evaporation or mineral dissolution. The points farthest to the left represent water from wells greater than 100 ft deep in recharge areas near the mountain front where dissolved-solids concentration is low and the water has undergone little evaporation. The points that indicate an increase in chloride concentrations and $\delta^{18}\text{O}$ values represent water recharged to the ground-water system after some degree of evaporation has taken place. The points that represent water

from the shallow unconfined aquifer mainly indicate enrichment in oxygen-18 with increasing chloride. The points that indicate an increase in chloride concentration and no change in $\delta^{18}\text{O}$ represent water that has undergone mineral dissolution and not evaporation.

Water that has undergone mineral dissolution is grouped primarily in the lower center to lower right side of the graph. This water has low $\delta^{18}\text{O}$ values that remain relatively constant and indicate that little evaporation has occurred, while chloride concentrations increase almost two orders of magnitude. The high concentrations of chloride coupled with the limited effects of evaporation indicate that the increase in chloride is caused primarily by abundant chloride ions in the basin-fill material and/or pore fluids that the water has been in contact with. The increase in chloride ions is evident in water from wells sampled in the northwestern part of the valley—the terminus of ground-water flow. Two shallow wells sampled in the northwestern part show some enrichment in oxygen-18 values in addition to an increase in chloride concentration. This may be caused by infiltration and subsequent dispersion of surface water that has undergone some evaporation into the shallow aquifer.

Tritium

Tritium concentration in surface water ranged from 9.0 TU in water from Little Cottonwood Creek to 19.8 TU in water from the Jordan River at 6400 South Street (table 6). This range is similar to the range in concentration in present-day precipitation. Ground water from four wells in the area that are completed at depths less than 101 ft below land surface, wells (C-2-1)23dac-1, (D-2-1)29acb-2, (D-2-1)30dda-2, and (D-2-1)30dda-3, have tritium concentrations that ranged from 15.9 to 52.8 TU (table 7). On the basis of tritium concentration, most of the water from these wells was recharged to the ground-water system between 1952, the year above-ground nuclear-weapons testing began, and 1992. These wells are near surface-water bodies, and water from wells (D-2-1)29acb-2 and (D-2-1)30dda-3 have tritium concentrations (15.9 and 16.2 TU, respectively) very similar to those measured in water from area streams and ditches.

Ground-water samples were collected from eight wells finished in the principal aquifer near Little Cottonwood Creek and the area to the south. Tritium concentration ranged from 14.8 TU at well (D-3-1)9aab-1 to 58.9 TU at well (D-2-1)34cdc-1 (table 7). The per-

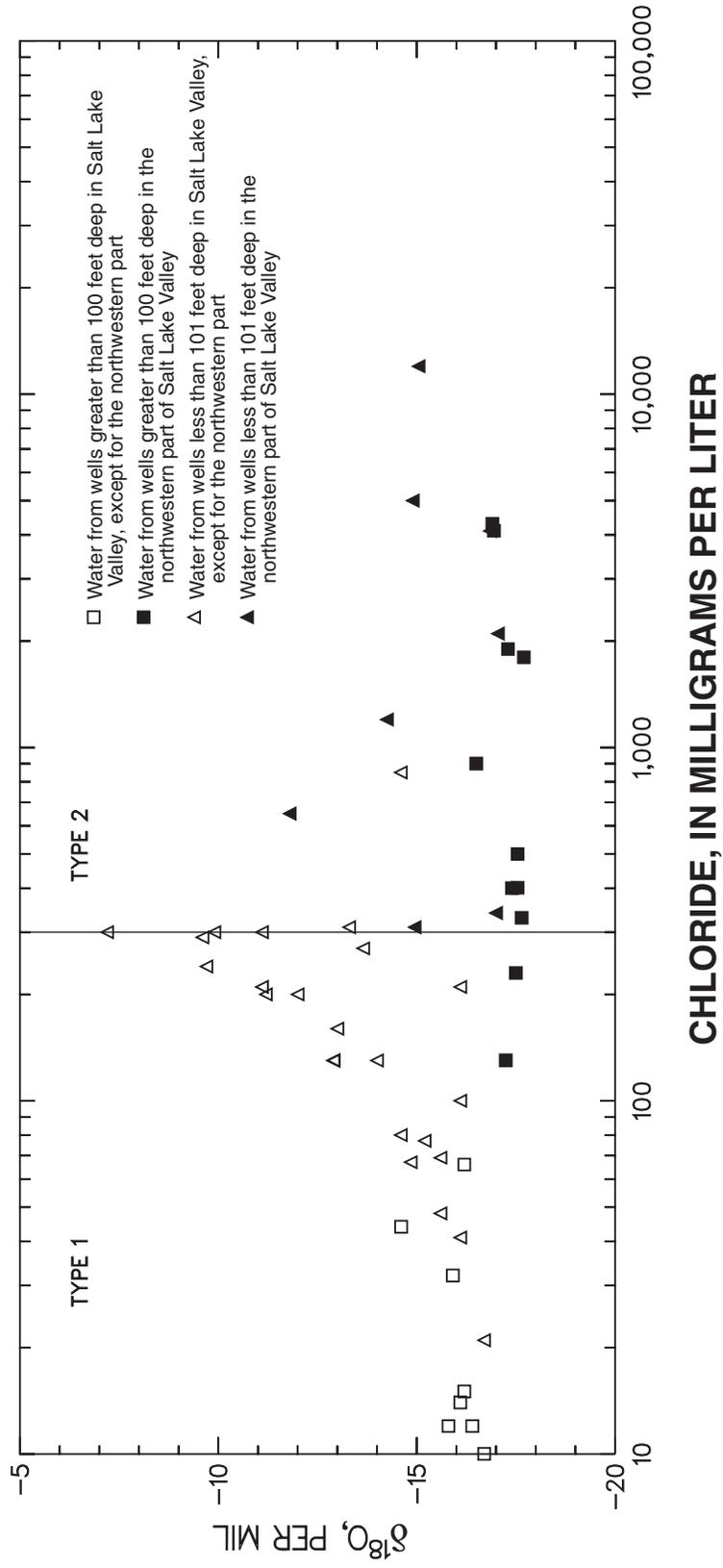


Figure 18. Relation between $\delta^{18}\text{O}$ value and chloride concentration in water sampled from selected wells in Salt Lake Valley, Utah.

forated intervals in well (D-3-1)9aab-1 extend from 570 to 925 ft below land surface, much deeper than those of other wells in the area. Water entering the well through the deeper perforations may have been recharged to the ground-water system before 1952 and would now have a tritium concentration near zero. The measured tritium concentration may reflect mixing of this deeper water with water from shallower perforations, which would have higher tritium concentrations. The average tritium value for water sampled from the principal aquifer in the area is 37.9 TU, if the sample from well (D-3-1)9aab-1 is disregarded. Mixing of water in well (D-3-1)9aab-1 from different depths is supported by the fact that water from well (D-3-1)3ccc-1, near well (D-3-1)9aab-1, comes from 520 to 670 ft below land surface and has a tritium concentration of 38.7 TU.

The high tritium concentration in precipitation that fell on the area in 1963 would have radioactively decayed to about 1,700 TU in 1991. Hydrodynamic dispersion and long-production intervals can cause sufficient mixing to decrease the high tritium concentration that is indicative of the 1960's. Tritium values ranging from about 33 to 59 TU in water from the principal aquifer in the southeastern part of the valley are assumed to represent water that was recharged in the 1960's, the period of peak nuclear-weapons testing. Tritium concentrations in the ground-water system that correspond with the concentration peaks that occurred in the 1960's can be used to determine the distance that ground water has traveled since recharge occurred. On the basis of a tritium concentration of 40.3 TU in 1991 (table 7), water from well (D-3-1)10dab-1, 1.5 mi west of the mountain front, is assumed to have been recharged in 1963. This assumption results in an average ground-water flow velocity in this area of about 280 ft/yr (7,900 ft in 28 years) if the recharge area is the basin-fill/consolidated-rock boundary at the mountain front.

The average ground-water flow velocity in the area also was determined from estimated aquifer properties using the equation (Lohman, 1979, p. 10):

$$\text{Average velocity} = \frac{-\text{hydraulic conductivity}}{\text{effective porosity}} \times \frac{\text{hydraulic-head gradient}}{\text{head}} \quad (8)$$

The negative sign indicates that the direction of movement is downgradient.

Hydraulic conductivity for an aquifer is equal to its transmissivity divided by its saturated thickness. Transmissivity of the aquifer in the vicinity of these two wells was estimated to be from 10,000 to 15,000 ft²/d (Price, 1988, pl. 1). The saturated thickness of the aquifer in this area is not known. Well (D-3-1)10dab-1 is completed in saturated deposits of Tertiary age, if the top of the Tertiary deposits are at 4,500 ft above sea level (Arnow and others, 1970, fig. 1). Well (D-3-1)2ccc-1, about 2,800 ft northeast of well (D-3-1)10dab-1, is perforated in mostly saturated deposits of Tertiary age from an altitude of about 4,500 to 4,000 ft. Assuming that the saturated thickness at well (D-3-1)10dab-1 is about 500 ft and the transmissivity is 15,000 ft²/d, hydraulic conductivity is 30 ft/d. The hydraulic-head gradient between wells (D-3-1)10dab-1 and (D-3-1)3ccc-1, measured on December 30, 1992, by the Salt Lake County Water Conservancy District (1992), was about 29 ft/4,767 ft (0.00608 ft/ft). Effective porosity of the aquifer was estimated at about 25 percent, on the basis of a sand and gravel composition and some degree of sorting. On the basis of these estimates, the average velocity of ground water in the unconfined part of the principal aquifer is estimated to be about 270 ft/yr.

The concentration of tritium in water from the principal aquifer is influenced by the vertical hydraulic head gradient between it and the shallow unconfined aquifer. Water samples from wells (C-2-1)13dbb-1, (D-2-1)28ccc-1, (D-2-1)29acb-8, (D-2-1)30dda-1, and (D-3-1)5bda-2 were collected in 1990 and 1991 from the principal aquifer where there was a downward hydraulic-head gradient between the shallow and deep aquifers. The wells are finished at depths greater than 395 ft below land surface, except for well (C-2-1)13dbb-1, have perforated intervals from 140 to 401 ft below land surface, and are used for public supply. Tritium concentrations ranged from 0.8 to 8.2 TU, indicating that most of the water was recharged to the system prior to 1952, but that a component of water is present that was recharged during the nuclear-weapons testing era. Water recharged during the nuclear-weapons testing period may have been induced to move down from shallower zones by the downward vertical hydraulic-head gradient caused by ground-water withdrawals in the area.

The Jordan River is assumed to be the terminus of the hypothesized flow paths that originate at the mountain front in the southeastern part of the valley (pl. 1). An upward hydraulic-head gradient is expected to

exist between the principal and shallow unconfined aquifer at the Jordan River, resulting in discharge from the principal aquifer. Because the Jordan River is about 2 to 3 mi downgradient from wells along the flow path that contain water with tritium values less than 10 TU, it was expected that concentrations near the river also would be less than 10 TU. Tritium concentrations in water from two wells finished in the principal aquifer near the river were 57.0 and 30.8 TU in 1991. Well (C-2-1)23dad-1 (57.0 TU) is perforated from 140 to 380 ft below land surface and well (C-2-1)24bcd-1 (30.8 TU) from 179 to 300 ft below land surface. Both wells are used for public supply. Water from well (C-2-1)23dac-1, 85 ft deep and finished above a clay layer near well (C-2-1)23dad-1, had a tritium concentration of 52.8 TU. Leakage of Jordan River water and infiltration of precipitation into the shallow unconfined aquifer and underlying confining layers may be the source of these high tritium concentrations in the principal aquifer.

Tritium values were determined for ground water from different depths sampled from wells (D-2-1)30dda-1, -2, and -3 and for surface water from the East Jordan Canal at (D-2-1)30dda, in the southeastern part of the valley. This site is in an area where a downward hydraulic-head gradient exists between the shallow unconfined and principal aquifer. Ground water sampled from about 15 to 20 ft below land surface [(D-2-1)30dda-3] had a tritium concentration similar to concentrations measured in water from the East Jordan Canal, 16.2 TU and 13.9 TU, respectively. Ground water sampled from 54 to 64 ft below land surface had a tritium concentration of 35.4 TU [(D-2-1)30dda-2], and water from several perforated intervals from about 560 to 990 ft below land surface had a concentration of 8.2 TU [(D-2-1)30dda-1]. On the basis of this profile of tritium values and depths below land surface, water recharged to the ground-water system during the 1960's presently is at depths greater than 20 ft below land surface. Ground water recharged after 1952 is present at depths greater than 500 ft below land surface, but this may be caused by the downward movement of water from shallower depths induced by withdrawals from the well.

Vertical Ground-Water Movement Indicated by Hydrologic Properties

Results of an aquifer test done at wells near the Vitro tailings area were analyzed using the Hantush (1960) modified method and the Neuman and Withers-

poon (1972) "ratio" method to determine the vertical hydraulic conductivity of the confining layer (k'). A value of 0.12 ft/d was calculated (Waddell and others, 1987, p. 30). Estimates of k' also were made by Hely and others (1971, p. 118) for two areas of the valley. A value of 0.016 ft/d was determined for an area near Great Salt Lake (Township 1 North, Range 2 West) and a value of 0.049 ft/d was determined for an area between Holladay and Murray. These estimates were made using the vertical hydraulic-head gradient through the confining layers and the ground-water discharge through a known area of the confining layers. The range in k' determined by aquifer-test data analysis is from 0.01 to 1.0 ft/d. Approximate vertical hydraulic-conductivity values determined for confining layers in the valley are shown in figure 19.

Vertical movement of water through confining layers is dependent on vertical hydraulic conductivity; therefore, the time of travel for water through confining layers is dependent on k' , the confining layer thickness, and the vertical hydraulic-head gradient. The first confining layer encountered at well (C-2-1)23dad-1 is about 12 ft thick. A downward vertical hydraulic-head gradient exists between the shallow unconfined and principal aquifers when well (C-2-1)23dad-1 is pumped. The hydraulic-head gradient was estimated to be about 0.1 from water-level declines measured at well (C-2-1)23dac-1, finished in the shallow unconfined aquifer, and a theoretical water-level decline calculated for the pumped principal aquifer at that location. It would take water 4,200 days, more than 11 years, to move from the shallow unconfined aquifer above the confining layer to the principal aquifer if k' is 0.01 ft/d and effective porosity is about 0.35. The travel time would be 420 days if k' is 0.1 ft/d. This period of time for water to move through the confining layer is in agreement with the similarity in tritium values measured for water sampled from above and below the confining layer in the area and the ages that they imply.

SUMMARY

Information on chemical composition and hydrologic properties of the ground-water system in Salt Lake Valley, Utah, was collected and analyzed from 1990 to 1992 by the U.S. Geological Survey in cooperation with the Utah Department of Natural Resources, Division of Water Rights, and the Utah Department of Environmental Quality, Division of Water Quality, to better understand the ground-water flow system. The

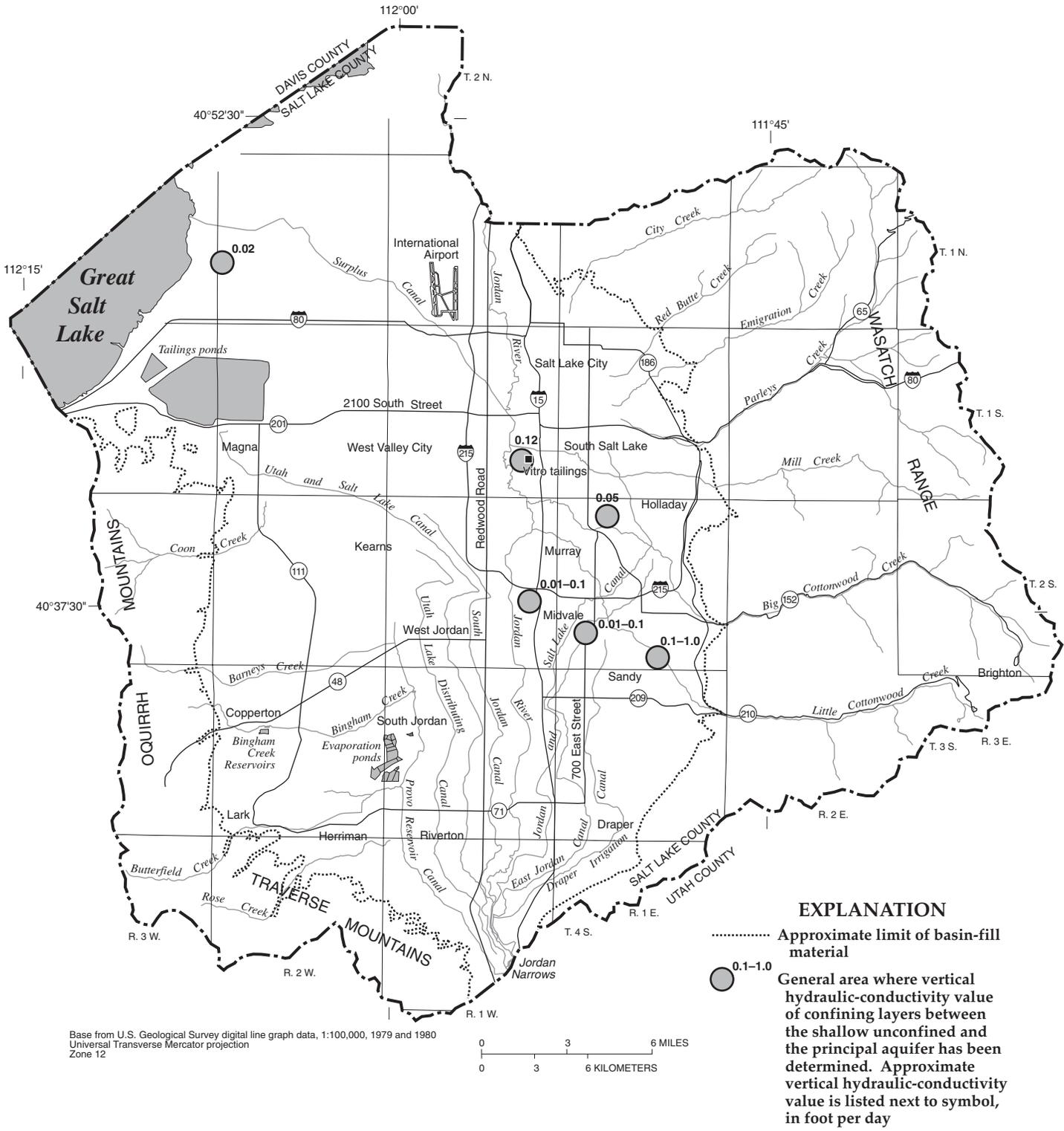


Figure 19. Vertical hydraulic-conductivity values determined for confining layers in Salt Lake Valley, Utah.

chemical composition of ground water in the valley varies with depth and location. Recharge sources and water-rock interactions affect water chemistry. Dissolved-solids concentration in water sampled from wells completed in the principal aquifer from 1988 to 1992 ranged from 110 mg/L on the east side of the valley to 48,100 mg/L on the west side.

Ground water from the principal aquifer in the southeastern part of the valley has a composition similar to nearby Wasatch Range streams. Reactions between basin-fill material derived from the Oquirrh Mountains and different types of recharge water result in ground water with a wide range in dissolved-solids concentrations in the southwestern part of the valley. Low concentrations of sulfate and relatively high concentrations of bicarbonate in some of the ground water sampled from the northwestern part of the valley may be caused by sulfate reduction.

Water from the shallow unconfined aquifer has more localized variation in composition than does water from the principal aquifer, primarily because of its proximity to land surface. The dissolved-solids concentration in water ranged from 331 mg/L in the eastern part of the valley to 20,900 mg/L in the northwestern part. Evapotranspiration, dissolution of minerals, and recharge of water diverted from the Jordan River contribute to higher dissolved-solids concentrations in water from the shallow unconfined aquifer.

Chloride concentration in water from the principal aquifer has increased steadily with time in some areas of the valley. The storage and use of road salt in recharge areas may be the cause for the increase on the eastern side of the valley. Relatively high chloride concentrations in water from volcanic rocks of Tertiary age beneath the basin fill may be the source of chloride in water from wells west of Riverton.

Hydraulic-conductivity values determined from slug tests done on wells completed in the shallow unconfined aquifer and confining layers ranged from 0.003 to 33.4 feet per day. Transmissivity values determined for the principal aquifer from four multiple-well aquifer tests ranged from 6,400 to 43,600 feet squared per day. Vertical hydraulic-conductivity values estimated from these tests are from 0.01 to 1 foot per day.

Water-level fluctuations measured in wells finished in the shallow unconfined aquifer are affected by water use occurring near the wells. Generally, water levels measured in wells near canals and irrigated fields peak in the late summer or fall in response to seepage

of water used for irrigation. Water levels measured in many shallow wells are affected by pumping in nearby public-supply wells and were lowest in the summer when withdrawals are the greatest.

Water from well (C-2-1)24bcd-1, finished in the principal aquifer, was modeled as a mixture of water from the Jordan River and from the principal aquifer sampled upgradient from the well. Chloride, oxygen-18, and deuterium were considered to be conservative constituents and independently gave mixing ratios of about 75 percent water from the principal aquifer and 25 percent from the Jordan River.

Water from 81 hydrologic-data sites was analyzed for stable-isotope ratios of oxygen and hydrogen to determine sources of recharge and mixing of water from these sources. Water sampled from the principal aquifer in the southeastern part of the valley is isotopically similar to water in Big and Little Cottonwood Creeks. On the basis of oxygen and hydrogen isotope ratios, water from two wells near the Jordan River, finished in the principal aquifer, contains a component of water that has undergone some degree of evaporation. Most of the ground water sampled from the shallow unconfined aquifer is more isotopically enriched than other water in the valley because of the effects of evaporation. Water from the Jordan River and its diversions is a probable source of recharge to the shallow unconfined aquifer.

Tritium concentrations ranging from about 33 to 59 TU in water from the principal aquifer in the southeastern part of the valley probably are representative of water recharged in the 1960's. Relatively high tritium concentrations measured in water from wells finished in the principal aquifer and overlying confining layers near the Jordan River indicate possible downward leakage of Jordan River water and infiltration of precipitation through the shallow unconfined aquifer.

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