Hydrogeology of Shallow Basin-Fill Deposits in Areas of Salt Lake Valley, Salt Lake County, Utah

Water-Resources Investigations Report 03–4029

U.S. Department of the Interior
U.S. Geological Survey
National Water-Quality Assessment Program

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By Susan A. Thiros

U.S. GEOLOGICAL SURVEY

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U.S. DEPARTMENT OF THE INTERIOR

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FOREWORD

The U.S. Geological Survey (USGS) is committed to serve the Nation with accurate and timely scientific information that helps enhance and protect the overall quality of life, and facilitates effective management of water, biological, energy, and mineral resources (http://www.usgs.gov/). Information on the quality of the Nation's water resources is of critical interest to the USGS because it is so integrally linked to the long-term availability of water that is clean and safe for drinking and recreation and that is suitable for industry, irrigation, and habitat for fish and wildlife. Escalating population growth and increasing demands for the multiple water uses make water availability, now measured in terms of quantity *and* quality, even more critical to the long-term sustainability of our communities and ecosystems.

The USGS implemented the National Water-Quality Assessment (NAWQA) program to support national, regional, and local information needs and decisions related to water-quality management and policy (http://water.usgs.gov/nawqa). Shaped by and coordinated with ongoing efforts of other Federal, State, and local agencies, the NAWQA program is designed to answer: What is the condition of our Nation's streams and ground water? How are the conditions changing over time? How do natural features and human activities affect the quality of streams and ground water, and where are those effects most pronounced? By combining information on water chemistry, physical characteristics, stream habitat, and aquatic life, the NAWQA program aims to provide science-based insights for current and emerging water issues and priorities. NAWQA results can contribute to informed decisions that result in practical and effective water-resource management and strategies that protect and restore water quality.

Since 1991, the NAWQA program has implemented interdisciplinary assessments in more than 50 of the Nation's most important river basins and aquifers, referred to as Study Units (http://water.usgs.gov/nawqa/nawqamap.html). Collectively, these Study Units account for more than 60 percent of the overall water use and population served by public water supply, and are representative of the Nation's major hydrologic landscapes, priority ecological resources, and agricultural, urban, and natural sources of contamination.

Each assessment is guided by a nationally consistent study design and methods of sampling and analysis. The assessments thereby build local knowledge about water-quality issues and trends in a particular stream or aquifer while providing an understanding of how and why water quality varies regionally and nationally. The consistent, multi-scale approach helps to determine if certain types of water-quality issues are isolated or pervasive, and allows direct comparisons of how human activities and natural processes affect water quality and ecological health in the Nation's diverse geographic and environmental settings. Comprehensive assessments on pesticides, nutrients, volatile organic compounds, trace metals, and aquatic ecology are developed at the national scale through comparative analysis of the Study-Unit findings (http://water.usgs.gov/nawqa/natsyn.html).

The USGS places high value on the communication and dissemination of credible, timely, and relevant science so that the most recent and available knowledge about water resources can be applied in management and policy decisions. We hope this NAWQA publication will provide you the needed insights and information to meet your needs, and thereby foster increased awareness and involvement in the protection and restoration of our Nation's waters.

The NAWQA program recognizes that a national assessment by a single program cannot address all water-resource issues of interest. External coordination at all levels is critical for a fully integrated understanding of watersheds and for cost-effective management, regulation, and conservation of our Nation's water resources. The program, therefore, depends extensively on the advice, cooperation, and information from other Federal, State, interstate, Tribal, and local agencies, non-government organizations, industry, academia, and other stakeholder groups. The assistance and suggestions of all are greatly appreciated.

Robert M. Hirsch

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Associate Director for Water

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CONVERSION FACTORS AND DATUMS

Multiply	By	To obtain
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year
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
gallon per minute per foot (gal/min/ft)	12.419	liter per minute per foot
inch (in.)	25.4	millimeter
mile (mi)	1.609	kilometer

[1]Expresses transmissivity. An alternative way of expressing transmissivity is cubic foot per day per square foot, times foot of aquifer thickness $[(ft^3/d)ft^2]ft$.

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

$$°F = 1.8(°C) + 32.$$

Specific conductance is reported in microsiemens per centimeter at 25 degrees Celsius (μS/cm).

Natural gamma radiation is reported in gamma counts per second and electromagnetic induction is reported in micromhos per meter.

Vertical coordinate information is referenced to the North American Vertical Datum of 1929 (NAVD 29). Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83).

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ABSTRACT

A study of recently developed residential/commercial areas of Salt Lake Valley, Utah, was done from 1999 to 2001 in areas in which shallow ground water has the potential to move to a deeper aquifer that is used for public supply. Thirty monitoring wells were drilled and sampled in 1999 as part of the study. The ground water was either under unconfined or confined conditions, depending on depth to water and the presence or absence of fine-grained deposits. The wells were completed in the shallowest water-bearing zone capable of supplying water. Monitoring-well depths range from 23 to 154 feet. Lithologic, geophysical, hydraulic-conductivity, transmissivity, water-level, and water-temperature data were obtained for or collected from the wells.

Silt and clay layers noted on lithologic logs correlate with increases in electrical conductivity and natural gamma radiation shown on many of the electromagnetic-induction and natural gamma logs. Relatively large increases in electrical conductivity, determined from the electromagnetic-induction logs, with no major changes in natural gamma radiation are likely caused by increased dissolved-solids content in the ground water. Some intervals with high electrical conductivity correspond to areas in which water was present during drilling.

Unconfined conditions were present at 7 of 20 monitoring wells on the west side and at 2 of 10 wells on the east side of Salt Lake Valley. Fine-grained deposits confine the ground water. Anthropogenic compounds were detected in water sampled from most of the wells, indicating a connection with the land surface.

Data were collected from 20 of the monitoring wells to estimate the hydraulic conductivity and transmissivity of the shallow ground-water system. Hydraulic-conductivity values of the shallow aquifer ranged from 30 to 540 feet per day. Transmissivity values of the shallow aquifer ranged from 3 to 1,070 feet squared per day. There is a close linear relation between transmissivity determined from slug-test analysis and transmissivity estimated from specific capacity.

Water-level fluctuations were measured in the 30 monitoring wells from 1999 to July 2001. Generally, water-level changes measured in wells on the west side of the valley followed a seasonal trend and wells on the east side showed less fluctuation or a gradual decline during the 2-year period. This may indicate that a larger percentage of recharge to the shallow ground-water system on the west side is from somewhat consistent seasonal sources, such as canals and unconsumed irrigation water, as compared to sources on the east side. Water levels measured in monitoring wells completed in the shallow ground-water system near large-capacity public-supply wells varied in response to ground-water withdrawals from the deeper confined aquifer. Water temperature was monitored in 23 wells. Generally, little or no change in water temperature was measured in monitoring wells with a depth to water greater than about 40 feet. The shallower the water level in the well, the greater the water-temperature change measured during the study.

Comparison of water levels measured in the monitoring wells and deeper wells in the same area indicate a downward gradient on the east side of the valley. Water levels in the shallow and deeper aquifers in the secondary recharge area on the west side of the valley were similar to those on the east side. Water levels measured in the monitoring wells and nearby wells completed in the deeper aquifer indicate that the vertical gradient can change with time and stresses on the system.

INTRODUCTION

As part of the National Water-Quality Assessment (NAWQA) program, the U.S. Geological Survey (USGS) studied the effects of human activities on the quality of shallow ground water in areas with recently developed urban land use in Salt Lake Valley, Utah, an urban area within the Great Salt Lake Basins study unit (fig. 1). Pioneers first settled Salt Lake Valley in 1847, and an irrigation system was developed soon after to grow crops in the semiarid climate. Agricultural and undeveloped areas in the valley were converted to residential and commercial uses as the population increased to about 850,000 in 1999 (U.S. Census Bureau, written commun., 2000).

The study was done from 1999 to 2001 in recently developed residential/commercial areas of the valley in which shallow ground water has the potential to move to a deeper aquifer that is used for public supply. In 1999, 30 monitoring wells were installed and the water sampled and analyzed for major ions, nutrients, trace metals, volatile organic compounds (VOCs), pesticides, radon, the stable isotopes of oxygen and hydrogen, and tritium to determine an approximate recharge date. The water-quality data are presented and discussed separately in Thiros (2003). Selected aquifer properties such as transmissivity and lithology were studied and water-level and temperature fluctuations were monitored. Information about the shallow ground-water system and its connection to the deeper system that provides drinking water for the valley can be used to better understand the entire ground-water system in the valley.

Purpose and Scope

This report presents lithologic, geophysical, hydraulic-conductivity, transmissivity, water-level, and water-temperature data collected from the shallow ground-water system in Salt Lake Valley from 1999 to 2001. The data were obtained for or collected from monitoring wells installed in recently developed residential/commercial areas of the valley. Lithologic and borehole geophysical logs of the subsurface are presented together to aid in the interpretation of the type of unconsolidated deposits penetrated. Water-level hydrographs show fluctuations with time caused by recharge to and discharge from the system in the study area. This information can be used with water-quality and land-use data to better understand the effects of human activities on the quality of shallow ground water where there is potential for movement to a deeper aquifer used for public supply.

Hydrogeologic Setting

Salt Lake Valley is within the Basin and Range physiographic province of Fenneman (1931) that is characterized by generally parallel, north- to northeast-trending mountain ranges separated by broad alluvial basins that are a result of Cenozoic-age extensional faulting. Topographic relief between the Wasatch Range and Salt Lake Valley is as much as 7,000 ft.

The basin-fill deposits in the valley consist of unconsolidated to semiconsolidated Tertiary-age sediments overlain by unconsolidated Quaternary-age sediments. The less permeable Tertiary-age basin-fill deposits are associated with alluvial fans and volcanic ash. The saturated Quaternary-age basin-fill deposits range from less than 200 ft thick in the Kearns area and along the margins of the valley to more than 1,000 ft thick in the northern part of the valley (Hely and others, 1971) and were deposited primarily in lacustrine and fluvial environments associated with Lake Bonneville and older lake cycles. Personius and Scott (1992) summarized the depositional history of Lake Bonneville from other sources. The water level of Lake Bonneville began rising about 30,000 years ago to a high of about 5,180 ft in altitude (Bonneville shoreline level) about 16,000 years ago. The lake dropped quickly to the Provo level at about 4,800 ft and much of the sediment deposited at or below the Bonneville level was eroded and transported to below the Provo level.

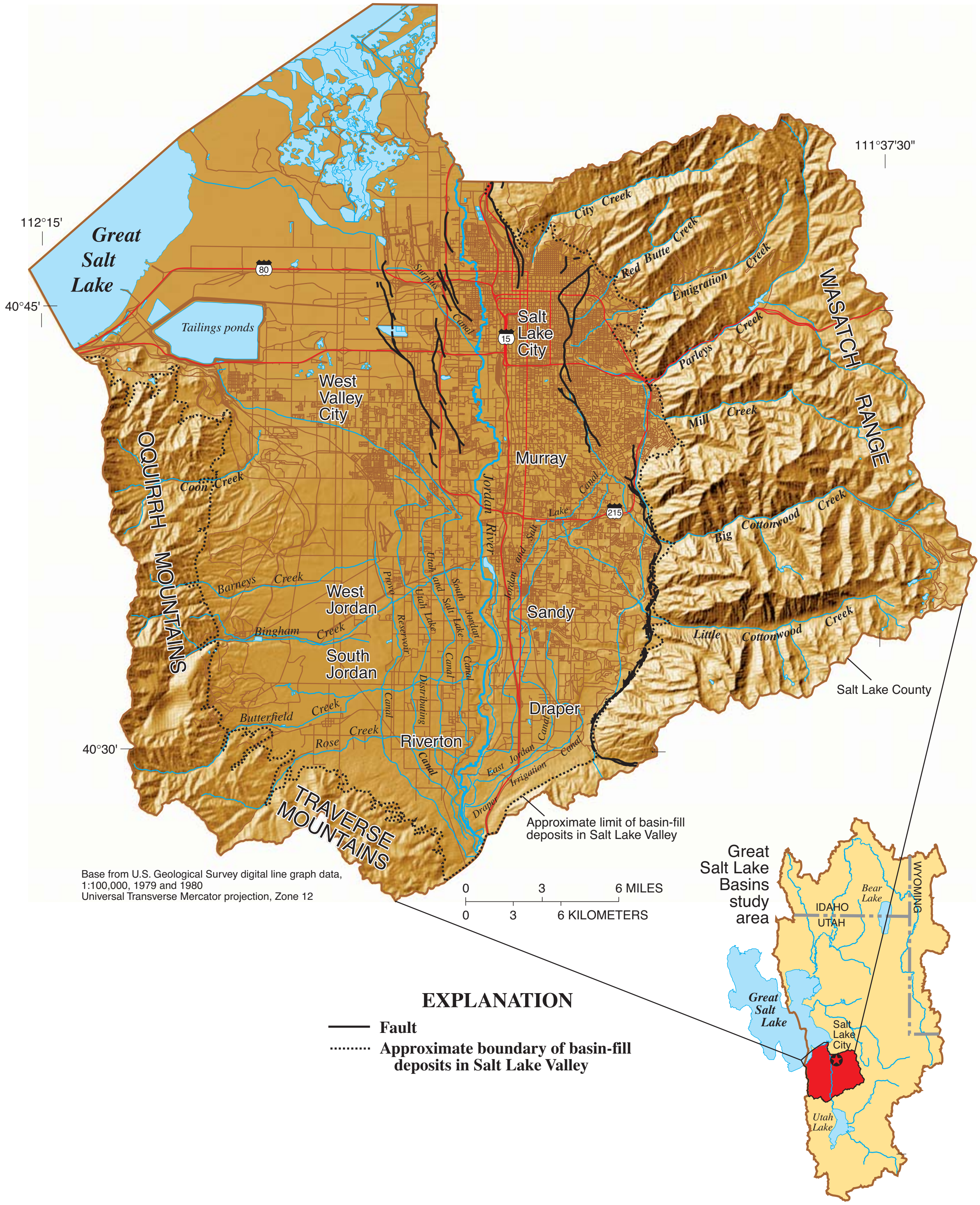


Figure 1. Location and geographic features of Salt Lake Valley, Utah.

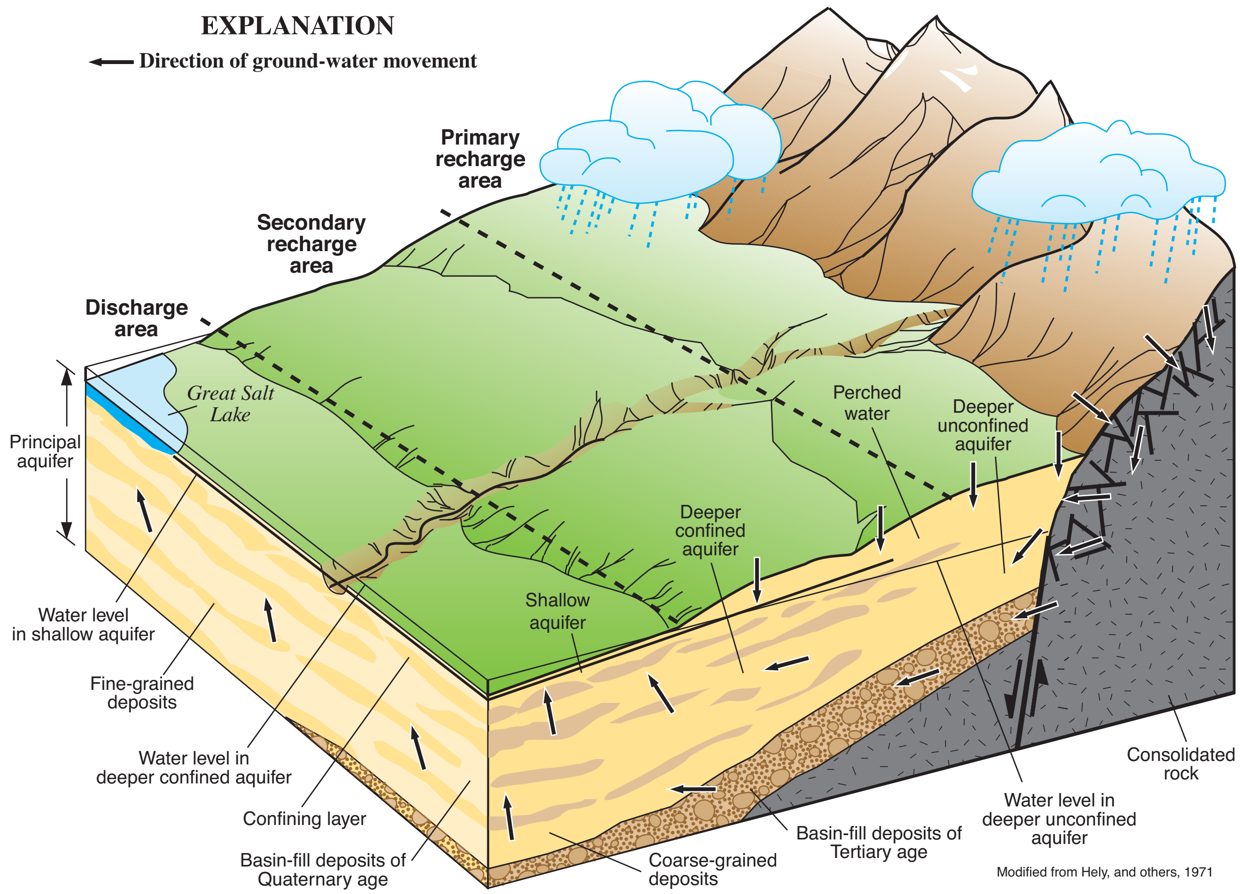


Figure 2. Generalized block diagram showing the basin-fill deposits and ground-water flow system in Salt Lake Valley, Utah.

Lake Bonneville was at the Provo shoreline level about 13,500 years ago and retreated to the level of modern Great Salt Lake (about 4,210 ft) by about 11,000 years ago. The shallow ground-water system in the valley primarily is contained in deposits associated with Lake Bonneville and in overlying, more recent stream deposits.

A generalized model of the saturated basin-fill deposits in Salt Lake Valley consists of a relatively deep unconfined aquifer near the mountain fronts that becomes confined toward the center of the valley by layers of silt and clay (fig. 2). The deeper unconfined and confined aquifers are collectively known as the principal aquifer in the valley. The term "confining layer" is used even though, in some cases, the water level in the underlying principal aquifer is below the confining layer, and the water is actually unconfined. Overlying the confining layers is shallow ground water that is either localized in extent because it is perched on a confining layer or more laterally continuous and forms an aquifer.

The primary recharge area for the principal aquifer is near the mountain fronts where there are no continuous layers of fine-grained material to impede the downward movement of water (fig. 2). Leakage of shallow ground water to the principal aquifer is possible where a downward gradient exists and confining layers are thin and/or discontinuous. These conditions are present in the secondary recharge area. Areas where the water level in the principal aquifer was below the first major confining layer and where there is the potential for downward movement of ground water were mapped as secondary recharge areas (Anderson and others, 1994, p. 6). A discharge area exists where there is an upward hydraulic-head gradient from the deeper confined aquifer to the overlying shallow aquifer (fig. 2).

The shallow aquifer in the valley is generally unconfined, although in some areas the first saturated zone in the subsurface was present beneath a confining layer. The distinction between the shallow and deeper aquifers is not clear in some parts of the secondary recharge area. Many domestic and some public-supply wells are open to the water table in the secondary recharge area, which can occur at depths of about 100 ft below land surface in some areas of the valley.

Ground water is perched where the water level in the principal aquifer is below the bottom of the confining layer resulting in an unsaturated zone between the water table and the overlying confining layer and shallow ground water. In the southeastern part of the valley, extensive areas with perched ground water can be difficult to differentiate from the shallow aquifer.

Recharge to the principal aquifer in the valley is from infiltration of precipitation and unconsumed irrigation water, subsurface inflow from adjacent fractured consolidated rocks, and seepage from streams and canals. Recharge to the east side of Salt Lake Valley is much greater than that to the west side, primarily because of subsurface inflow and streamflow seepage from the Wasatch Range, which receives greater amounts of precipitation than do the mountains on the west side of the valley. Discharge from the principal aquifer is to wells, springs, seepage to the Jordan River and the lower reaches of its tributaries, the shallow aquifer, and evapotranspiration. About 317,000 acre-ft/yr of water was added and removed in a steady-state numerical simulation of the ground-water system in Salt Lake Valley (Lambert, 1995, p. 37). The budget for the shallow part of the ground-water system is much smaller than that for the principal aquifer. The shallow system receives recharge from infiltration of precipitation and unconsumed irrigation water applied to fields, gardens, and lawns; seepage from streams and canals; and flow from the deeper confined aquifer in the discharge areas of the valley.

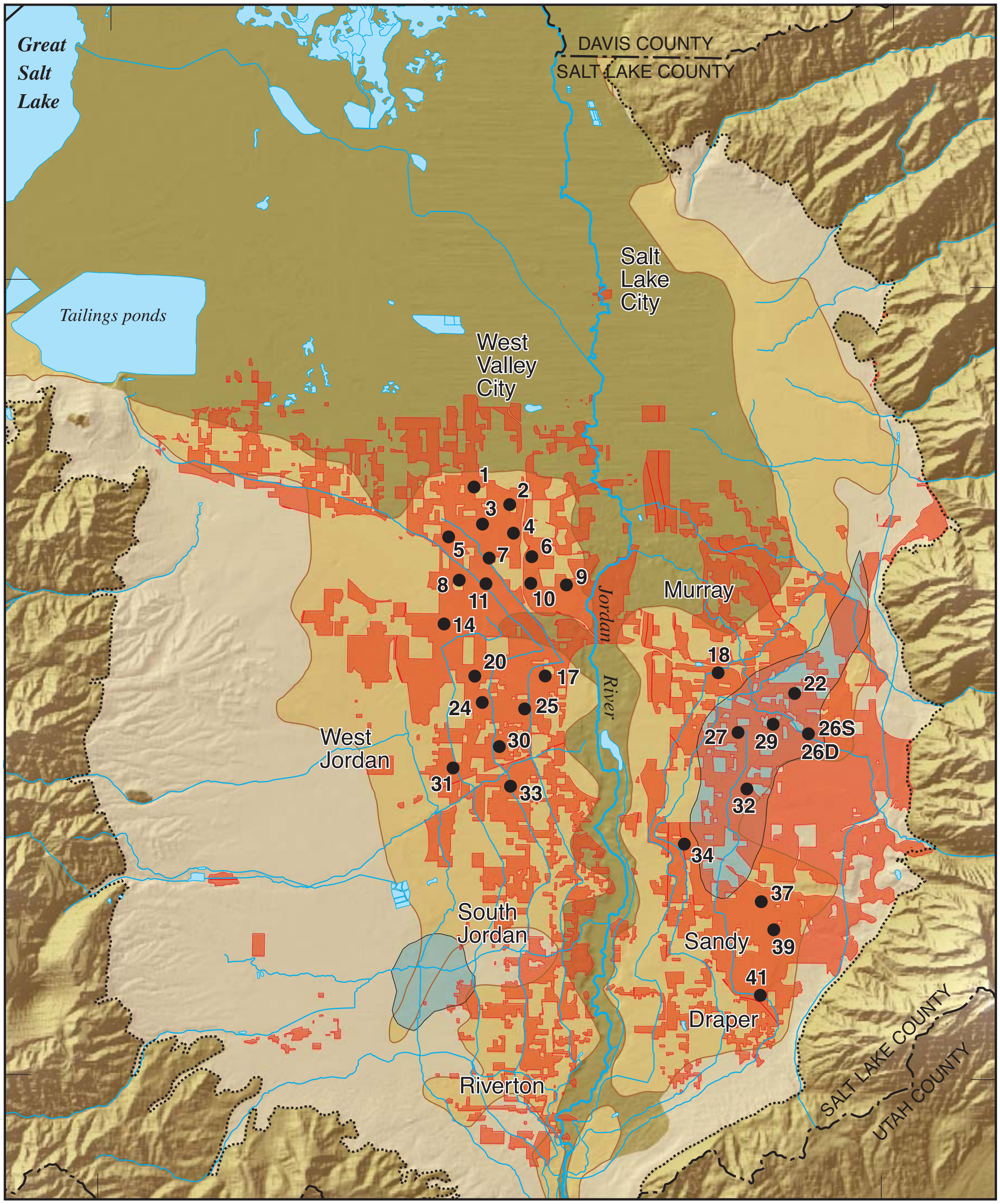
Shallow ground water is susceptible to contamination from activities related to land use because of its proximity to land surface. Manmade compounds such as VOCs and pesticides were detected in shallow ground water underlying residential areas (Thiros, 2003). The deeper unconfined aquifer in the primary recharge area also is vulnerable because of a lack of confining layers that can impede the downward movement of contaminated ground water. The movement of contaminated water from the shallow and deeper unconfined aquifers can degrade the water quality of the deeper confined aquifer in the secondary recharge area. The principal aquifer in Salt Lake Valley is used extensively for public supply (about 38 percent of the population uses ground water).

Methods

Monitoring wells were installed at 30 sites in Salt Lake Valley (fig. 3) in accordance with NAWQA protocols (Lapham and others, 1995). Well sites were selected with a computerized, stratified, random selection process (Scott, 1990) after the following criteria were met: (1) location in a residential or commercial area developed from 1963 to 1994; (2) 75 percent of a 500-meter (1,640-ft) circular buffer around the site contains the targeted land use; (3) a downward gradient exists between the shallow and deeper aquifers; and (4) a minimum distance of 1 kilometer (0.62 mi) exists between each site.

The monitoring wells were drilled with a hollow-stem auger (8 5/8-in.-diameter borehole) or with air-rotary methods where cobbles were assumed to be present (9-in.-diameter borehole). Ground water at the monitoring-well sites was under either unconfined or confined conditions, depending on the depth to water and the presence or absence of fine-grained deposits (table 1). The wells were completed in the shallowest water-bearing zone capable of supplying water. Monitoring-well depths range from 23 to 154 ft, and the wells were completed with 2-in.- diameter polyvinyl chloride casing attached to a 10-ft length of screen. Two wells were installed at one site and were completed at different depths. The shallower nested well was completed with a 5-ft length of screen. The top of the screened interval generally was about 5 ft below the water table or below where water was noted during drilling.

The monitoring wells were constructed with sand between the casing and the borehole wall to about 5 ft above the top of the screen. A 2-ft-thick layer of bentonite pellets in the annular space separates the sand pack from bentonite grout that extends to the land surface. The part of the aquifer that likely transmits water to the screened interval of a monitoring well, the contributing unit, was determined from the occurrence of fine-grained deposits in the subsurface (table 1). The



Base from U.S. Geological Survey digital line graph data, 1:100,000, 1979 and 1980
Universal Transverse Mercator projection, Zone 12

0 3 6 MILES
0 3 6 KILOMETERS

EXPLANATION

- **Area of recent residential and commercial land use**
- **Perched aquifer**—Data from Hely and others, 1971
- **Discharge area**
- **Primary recharge area**
- **Secondary recharge area**
- **Approximate boundary of basin-fill deposits in Salt Lake Valley**
- 41 ● **Monitoring well**—Number is well name in table 1

Figure 3. Location of monitoring wells installed in recently developed residential/commercial areas of Salt Lake Valley, Utah.

Table 1. Description, lithology, and water level for monitoring wells drilled in Salt Lake Valley, Utah, 1999

[**Aquifer type:** C, confined or U, unconfined; **General lithology of contributing unit:** SGVC, sand, gravel, and clay; SAND, sand; SDGL, sand and gravel; SDST, sand and silt; GRVL, gravel; STCL, silt and clay; SDCL, sand and clay; **Top and bottom of contributing unit:** upper and lower extent of basin-fill deposits that could contribute water to the monitoring well. **Water level** was measured in March 2000]

Well number (fig. 3)	Altitude of land surface (feet)	Well depth	Top of screened interval	Bottom of screened interval	Aquifer type	General lithology of contributing unit	Top of contributing unit	Bottom of contributing unit	Water level
		(feet below land surface)					(feet below land surface)		
1	4,308	48.5	38	48	C	SGVC	40	48.5	32.77
2	4,294	38.5	28	38	C	SGVC	26	38.5	19.55
3	4,371	114	103	113	C	SAND	113	114	79.96
4	4,331	38.5	28	38	U	SDGL	31	38.5	14.01
5	4,426	43.5	33	43	C	SDGL	20	43.5	27.59
6	4,324	38.5	28	38	C	SDST	26	38.5	23.46
7	4,396	43.5	33	43	C	SDST	29	43.5	27.09
8	4,487	67.5	57	67	U	SAND	46	67.5	63.37
9	4,312	38.5	28	38	C	SDST	21	38.5	12.57
10	4,350	83.5	73	83	C	GRVL	70	78	67.94
11	4,462	83.5	73	83	C	SDGL	75.5	83.5	70.75
14	4,579	48.5	38	48	C	STCL	30	48.5	27.97
17	4,380	38.5	28	38	C	SGVC	32	38.5	17.72
18	4,411	106	95	105	C	SGVC	102	106	76.85
20	4,477	92.5	82	92	U	GRVL	79	92.5	84.66
22	4,538	36	25.5	35.5	C	SDGL	15	26	21.61
24	4,473	124	113	123	C	GRVL	106	124	81.98
25	4,414	68.5	58	68	C	GRVL	34	68.5	49.67
26D	4,591	77.5	62	72	C	SAND	0	69	32.77
26S	4,591	31.5	26	31	U	SAND	0	31.5	30.43
27	4,499	73.5	63	73	C	SDGL	43.5	68	58.69
29	4,532	34	23.5	33.5	C	SAND	21	34	9.02
30	4,455	68.5	58	68	U	SGVC	62	68.5	58.12
31	4,562	154	143	153	U	SDGL	143	154	143.19
32	4,640	88.5	78	88	U	SAND	0	83	76.18
33	4,466	95.5	85	95	U	SDGL	66	95.5	74.14
34	4,486	77.5	67	77	C	SAND	65	78.5	57.81
35	4,540	106	95	105	U	GRVL	96	103	dry
37	4,725	73	62.5	72.5	C	SDCL	59	69	45.18
39	4,758	106	95.5	106	U	SDCL	21	105	94.02
41	4,550	23	12.5	22.5	U	SDGL	0	20	5.48

contributing unit to a well could extend to land surface if few or no fine-grained deposits are present above the screened interval.

Information about the basin-fill deposits and water in the subsurface was collected from these monitoring wells to better understand the shallow ground-water system in the valley. The types of data collected and the methods used to analyze these data are described in the following sections.

Lithologic and Borehole Geophysical Logs

A description of the lithology was logged while augering or drilling each well. The amount of resistance, the smoothness or roughness at the drill bit, and the type of drill cuttings (basin-fill deposits) returned to the surface provided information about the material in the subsurface. Cores were collected from some holes to get a relatively undisturbed sample from a known depth (fig. 4). Core samples could not be collected from intervals containing large gravel or cobbles. The occurrence and depth of water in the subsurface was monitored during drilling. The depth to water in the well after it had reached equilibrium with the aquifer also was measured. This information was compiled into a lithologic log for the well.

Data were collected from selected completed monitoring wells by using borehole geophysical techniques to better define aquifer properties. Natural gamma radiation and electromagnetic induction were logged from near the land surface to near the bottom of the well. Both tools used a sample interval of 0.1 ft. Natural gamma radiation units were gamma counts per second and electromagnetic induction units were micromhos per meter. A natural gamma log can be used in stratigraphic correlation and to delineate changes in lithology by recording the amount of natural gamma radiation emitted from the unconsolidated deposits surrounding the borehole. In general, natural gamma radiation is relatively high in fine-grained deposits that contain abundant clay minerals and relatively low in coarse-grained sand and gravel composed mainly of quartz. Natural gamma radiation typically correlates inversely with hydraulic conductivity, a measure of how quickly water can move through a porous material. A change in water quality does not cause a change in natural gamma radiation. Natural gamma radiation was recorded as the number of counts per second per sample interval.

The most significant naturally occurring sources of gamma radiation are potassium-40 and daughter products of the uranium- and thorium-decay series (U-238 and Th-232). Clay minerals commonly emit relatively high gamma radiation because they include weathering products of potassium feldspar and mica and tend to concentrate uranium and thorium by adsorption and ion exchange. Under most conditions, 90 percent of the gamma radiation detected probably originates from material within 6 to 12 in. of the borehole wall (Keys, 1990). The borehole diameter for the monitoring wells was a minimum of about 9 in. (the diameter of the drill bit) and could be larger depending on the cohesiveness of the surrounding aquifer material. Pellets and chips made from bentonite clay were used to seal the space around the 2-in.-diameter well casing from material just below land surface and to separate the sand-packed interval around the well screen from the rest of the well (pl. 1). The bentonite pellets and chips caused high natural gamma-radiation levels; therefore, the intervals where they were used cannot be correlated to the aquifer material.

An electromagnetic-induction log is used to evaluate lithologic variability and to identify electrically conductive fluids by recording the electrical conductivity of the material and ground water surrounding the borehole. As with natural gamma radiation, the electrical conductivity of clayey material generally is higher than that of coarse-grained deposits. Major factors affecting the response of the electromagnetic-induction tool are the dissolved-solids concentration of the ground water and of the clay, and conversely, the sand content of the aquifer material. The electromagnetic-induction tool measures the electrical conductivity within a zone from 10 to 50 in. from the well. Because direct contact with a conductive medium is not required, induction logs are especially useful for logging the dry portion of boreholes and are unaffected by the presence of polyvinyl chloride well casings. Electromagnetic-induction logging does not work in metal casings.

Geophysical logs do not show a unique response to a particular rock or aquifer type, but rather can be used as a qualitative tool to help correlate layers in the subsurface. When natural gamma radiation and electromagnetic-induction logs are compared for a particular well, similar responses are likely the result of changes in clay mineral content with depth. An increase in electrical conductivity without a corresponding increase in natural gamma radiation

Sand and gravel from about 11 to 11.5 feet below land surface at well 2.
Core is about 3.5 inches in diameter.

Fine sand from 34 to 35 feet below land surface at well 8.
Core is about 3.5 inches in diameter.

Figure 4. Cores of basin-fill deposits from two monitoring wells in Salt Lake Valley, Utah.

could result from ground water containing more dissolved solids. Geophysical logs should be used in conjunction with lithologic logs, water-quality data, and other information to provide the most reasonable interpretation of the aquifer material in the vicinity of the well.

Hydraulic Conductivity and Transmissivity

Slug tests were done at most of the monitoring wells to estimate the hydraulic conductivity or transmissivity of the shallow aquifer. Hydraulic conductivity is the measure of the rate at which water can move through a porous material. The transmissivity of an aquifer is equal to the hydraulic conductivity multiplied by the saturated thickness of the aquifer. The rate that water moves through the basin-fill deposits can affect water quality.

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. A pressure transducer measured water-level change at several time intervals in response to the water displacement. 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. Water- level recovery was monitored by the same process after the slug was quickly raised out of the water.

The Bouwer and Rice method was used to estimate hydraulic conductivity from slug tests done at eight wells and is discussed by Bouwer and Rice (1976) and Bouwer (1989). The Cooper, Bredehoeft, and Papadopulos method (Cooper and others, 1967) was used to estimate transmissivity from slug tests done at 12 wells. These methods assume that the aquifer is homogeneous and of uniform thickness, has infinite areal extent, has no leakage from above or below the developed zone, and that the slug is instantaneously added or removed from the well. The Bouwer and Rice method assumes that the aquifer is unconfined or confined and that flow is steady, while the Cooper, Bredehoeft, and Papadopulos method assumes that the aquifer is confined and that flow is unsteady.

A limitation of slug tests in estimating hydraulic conductivity and transmissivity is that the value determined is only representative of the aquifer near the screened interval and is influenced by the grain size of the material in the area around the screen (Bouwer, 1989). Fine-grained deposits around the screen 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 typically is unknown. The computer program AQTESOLV version 2.13 (Duffield, 1999) was used to analyze the slug-test data.

Specific capacity is the yield of a well per unit of drawdown and was calculated from the pumping rate divided by the drawdown measured while purging the monitoring wells prior to sampling. Transmissivity can be estimated by using specific capacity, storage coefficient, pumping duration, and well diameter if the well is assumed to be 100 percent efficient (Theis, 1963).

Water-Level and Temperature Fluctuations

Water-level and temperature fluctuations can indicate sources of recharge and discharge and how connected the aquifer is to activities and processes occurring at the land surface. Water-level fluctuations were measured in 30 monitoring wells from September 1999 to July 2001. Depth to water was measured with a calibrated tape on a monthly basis. Pressure sensors and data loggers were installed in several wells and water levels were measured on an hourly basis. Water levels measured by pressure sensors were checked against water levels measured by tape and the sensors were recalibrated if necessary to make the two levels match.

Temperature sensors and loggers were installed in 23 monitoring wells and recorded water temperature on an hourly basis during the study. Generally, the sensors were set in the well at about the top of the screened interval so that the water temperature was believed to be representative of the aquifer. The sensor was placed within the screened interval of the well if the water level was near the top of the screen.

HYDROGEOLOGY OF SHALLOW BASIN-FILL DEPOSITS

Lithologic and Borehole Geophysical Logs

Lithologic, natural gamma radiation, and electromagnetic-induction data from 30 monitoring wells were compared to qualify the connection between land surface and the underlying ground water (pl. 1). Natural gamma radiation varied from 0 to about 100 gamma counts per second for basin-fill deposits on the west side of the valley and from 0 to about 150 gamma counts per second for deposits on the east side. The basin-fill deposits on the east side of the valley contain rocks derived from uranium-rich igneous sources that affect the amount of natural gamma radiation. Intervals around the well casings with bentonite pellets or chips had higher values of natural gamma radiation (pl. 1). Electrical conductivity shown on the electromagnetic-induction logs ranged from 0 to more than 300 micromhos per meter, but most values were less than 150 micromhos per meter. The geophysical logs shown on plate 1 do not have consistent scales so that geophysical logs from each well could be compared to its associated lithologic log.

Silt and clay layers noted on lithologic logs correlate with increases in electrical conductivity and natural gamma radiation shown on many of the electromagnetic-induction and natural gamma logs. Relatively large increases in electrical conductivity determined from the electromagnetic-induction logs with no major changes in natural gamma radiation were measured at wells 5, 18, 22, 27, and 32 (pl. 1). These increases are likely caused by increased dissolved solids content in the ground water. Intervals with high electrical-conductivity measurements at wells 5, 11, 18, and 27 correspond to areas in which water was present during drilling. Increases in electrical conductivity also occur above the water table (well 18 at 15 ft, well 22 at 6 ft, and well 32 at 55 ft) and may represent pulses of more conductive water moving downward through the unsaturated zone to the water table. The source of the dissolved solids in ground water may be naturally occurring or from anthropogenic sources such as salt applied to roads for de-icing in the winter or relatively conductive compounds such as oil or solvents.

The coarsest and best-sorted sediments penetrated by the wells were present on the east side of the valley where high-energy streams from the Wasatch Range deposited deltas in Lake Bonneville. Sandy and silty clay deposited sometime after Lake Bonneville receded from the Provo shoreline level was mapped at land surface in many areas on the west side of the valley (Miller, 1980). Monitoring wells 8, 14, and 31 are above the extent of these fine-grained deposits. Basin-fill deposits logged near land surface in wells 8 and 31 correlate to mapped sands and gravels deposited during the Provo level of the lake. Sandy silt and clay logged near land surface in well 14 was deposited in a lower energy environment.

Unconfined conditions were present at 7 of 20 monitoring wells on the west side and at 2 of 10 wells on the east side of the valley. The aquifer is considered unconfined if the water level in the completed well is at or below the level at which it was encountered during drilling. Confined conditions are present if the water level rose above the level at which it was first encountered in the subsurface. Fine-grained deposits confine the ground water.

The water level remained at about the same level it was at when the well was drilled or declined after the well was completed in wells 4, 8, 20, 30, 31, 32, 33, 35, and 41 (pl. 1). The shallow aquifer is unconfined in these areas; therefore, water at the land surface has the potential to move downward to the water table. Anthropogenic compounds such as VOCs and pesticides were detected in water sampled from most of these wells (Thiros, 2003). Water from well 35 was not sampled because the water level declined below the bottom of the well, and water from well 32 did not contain detectable concentrations of VOCs or pesticides. The depth to water in well 31, in the southwestern part of the valley, is about 145 ft below land surface. Many domestic wells in the area draw water from near the surface of the water table because the water table is so deep.

Water levels measured after well completion rose about 5 ft or less from the levels first measured in wells 5, 6, 7, 14, 22, 25, 27, and 39 (pl. 1). Fine-grained deposits could confine the aquifer open to these wells, although confining layers are not evident from all of the logs (wells 5 and 22). Also, some of the rise may be attributable to seasonal changes in the aquifer that occurred between when the well was drilled and when the water level was measured later in the summer. The highest concentrations of tetrachloroethylene (PCE) in monitoring wells were detected in water from wells 5 and 27 (Thiros, 2003), which indicates connection with the land surface at these sites.

Water levels rose as much as 32 ft above the levels first measured in wells 1, 2, 3, 9, 10, 11, 17, 18, 24, 26, 29, 34, and 37 (pl. 1). Fine-grained deposits noted on the lithologic logs for these wells at or above the screened interval confines the ground water. Although confined conditions exist, manmade chemicals such as atrazine and PCE were detected in water sampled from these wells (Thiros, 2003). The recharge area for these wells is likely upgradient to the east or west where confining layers are absent or discontinuous.

Hydraulic Conductivity and Transmissivity

Data were collected from 20 of the monitoring wells to estimate the hydraulic conductivity and transmissivity of the shallow ground-water system (table 2). The rate that water moves through the subsurface can be used with water-quality data to better understand the effects of residential and commercial land uses on the underlying shallow ground water. Hydraulic conductivity was estimated from slug tests done on 8 monitoring wells completed in an unconfined or confined part of the shallow aquifer, and transmissivity was estimated from tests done on 12 monitoring wells completed in a confined part of the shallow aquifer (fig. 5).

Hydraulic-conductivity values of the shallow aquifer ranged from 30 ft/d at well 27 to 540 ft/d at well 1. Hydraulic-conductivity values were determined for the aquifer at the screened interval of the monitoring well and represent mostly horizontal flow through more permeable deposits. Overlying layers of fine-grained deposits may have a vertical hydraulic-conductivity value of from 0.01 to 1 ft/d (Thiros, 1995), resulting in slower overall water movement. Hydraulic-conductivity values of from 0.03 to as much as 3,000 ft/d with a log-normalized mean value of 30 ft/d have been reported for coarse-grained basin-fill deposits in the Basin and Range province (Bedinger and others, 1987, p. 39).

Transmissivity values of the shallow aquifer ranged from 3 ft^2/d at well 4 to 1,070 ft^2/d at well 33. If the thickness of the shallow aquifer at well 4 is about 10 ft and at well 33 about 50 ft, then the hydraulic-conductivity value would range from about 0.3 to 20 ft/d.

There is a close linear relation between transmissivity determined from slug-test analysis and transmissivity estimated from specific capacity. The regression line fit to the data has an R^2 (the fraction of the variance explained by regression) of 0.94 that indicates that transmissivity estimates derived from the slug tests are similar to those derived from specific-capacity values. The transmissivity values for well 33 derived from slug-test and specific-capacity data were not used to calculate the regression line fit because they are about an order of magnitude greater than those for the other wells.

Water-Level and Temperature Fluctuations

The depth to water in the wells was monitored to determine how the shallow aquifer responds to changes in recharge and discharge. The water-level surface determined from water levels measured in the 30 wells in March 2000 is shown in figure 6. The water-level surface generally correlates with the land surface, sloping toward the center of the valley or to the north-northeast in the northwestern part of the valley. Depth to ground water measured in the monitoring wells on the southeastern side of the valley probably represents both perched ground water and the shallow aquifer because of discontinuous layers of clay in the area. The relation between the different perched ground-water areas and the shallow aquifer is not known; therefore, the water-level surface for shallow ground water on the southeastern side of the valley may not be continuous.

Water-level fluctuations were measured in the 30 monitoring wells from 1999 to July 2001 (pl. 1). Changes in water levels measured in these wells tend to follow a seasonal pattern that is dependent upon water use and location, similar to that noted for other monitoring wells completed at the bottom of the shallow aquifer in Salt Lake Valley (Thiros, 1995, p. 38). Generally, water-level changes measured in wells on the west side of the valley followed a seasonal trend while wells on the east side showed less fluctuation or a gradual decline during the 2-year period. This may indicate that a higher percentage of recharge to the shallow ground-water system on the west side is from somewhat consistent seasonal sources, such as canals and unconsumed irrigation water, as compared to recharge on the east side.

Water levels for monitoring wells near canals or ditches that carry water for irrigation typically are highest in the late summer or fall in response to infiltration of unconsumed irrigation water. Water levels in these wells are lowest in the spring before irrigation begins. Water levels rose about 10 ft from June to October in well 31, about 4,000 ft east (downgradient) from the Provo Reservoir Canal and about 400 ft west (upgradient) of the Utah Lake Distributing Canal. Losses from these two canals to the ground-water system in the vicinity of well 31 were determined to be about 1.8 $ft^3/s/mi$ (Herbert and others, 1985). Water levels rose about 6 ft from May to October in well 33, next to the Utah and Salt Lake Canal in West Jordan. Losses from the adjacent canal and the two upgradient canals near well 33 totaled about 2.5 $ft^3/s/mi$ (Herbert and others, 1985).

Monitoring wells 26S (31.5 ft deep) and 26D (77.5 ft deep) are nested together in well 26 next to Little Cottonwood Creek on the east side of the valley. Water levels in the wells rise in the spring and decline the rest of the year (pl. 1). Snowmelt runoff in the stream measured at gaging station 10167800, about 0.5 mi west of well 26D, was followed by an increase in water level in well 26D within 1 to 2 days in both the spring of 2000 and 2001 (fig. 7).

Table 2. Specific-capacity and slug-test data for 20 monitoring wells in Salt Lake Valley, Utah

[**Specific capacity** is well discharge, in gallons per minute, divided by water-level decline, in feet. **Estimated aquifer thickness** and **storage coefficient** are used with specific-capacity value to estimate transmissivity. —, no data]

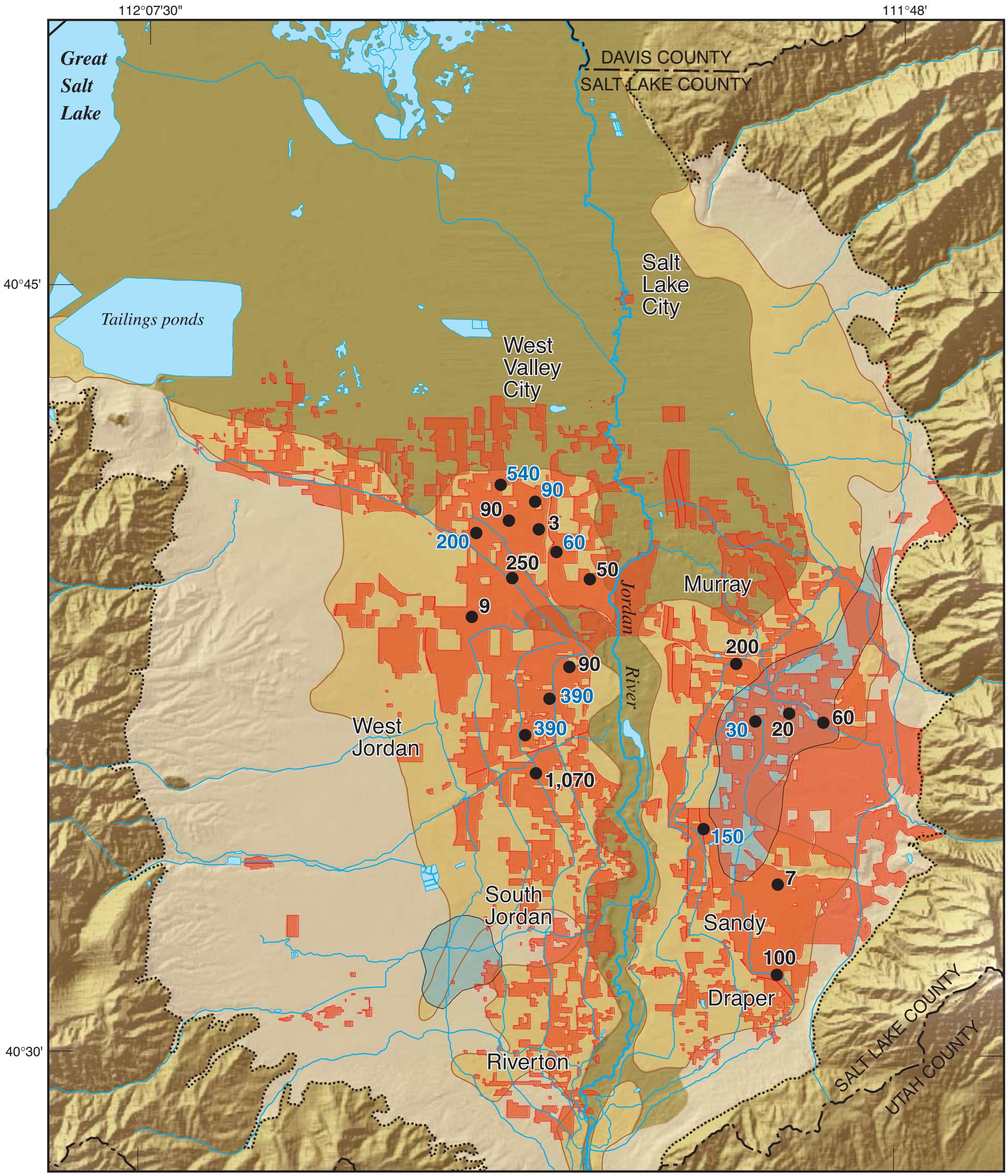
Well number (fig. 3, table 1)	Specific capacity	Estimated aquifer thickness (feet)	Estimated storage coefficient	Transmissivity estimated from specific-capacity data (feet squared per day)	Transmissivity estimated from slug-test data (feet squared per day)	Estimated hydraulic conductivity (feet per day)
1	—	10	0.001	530	—	540
2	1.72	10	.001	420	—	90
3	.24	10	.001	50	90	—
4	.2	10	.001	5	3	—
5	1.61	20	.1	230	—	200
6	.44	25	.1	50	—	60
7	.93	20	.001	210	250	—
9	.14	20	.001	30	50	—
14	.5	20	.01	10	9	—
17	.48	20	.0005	120	90	—
18	.69	20	.0005	180	200	—
25	2.08	40	.1	300	—	390
26D	.30	48	.0005	70	60	—
27	.42	30	.1	50	—	30
29	.5	20	.005	10	20	—
30	2.80	20	.1	330	—	390
33	2.26	20	.0005	1,610	[1] 1,070	—
34	1.49	20	.1	190	—	150
37	.4	10	.005	10	7	—
41	.47	17	.0005	120	100	—

[1]Value not included in the calculation of the regression equation for the relation between slug-test and specific-capacity derived transmissivity.

The water-level change measured in monitoring well 10 was different from that measured in other wells in the area. Water levels peaked during March-April and were lowest during August-September. The depth to water in this well is below that of nearby lower-altitude wells 6 and 9. Well 10 may be open to a deeper, more isolated part of the shallow ground-water system, although it was completed in the first saturated zone in the subsurface and anthropogenic compounds were detected in the water.

The water level in well 4 was higher than that in upgradient well 3 and in nearby wells 2 and 6. Low aquifer transmissivity may cause local recharge in the area to form a ground-water mound relative to other areas. The southern extent of the Granger Fault is mapped near well 4 (Personius and Scott, 1992) and also may influence ground-water flow in the area.

Water levels measured in monitoring wells completed in the shallow ground-water system near large-capacity public-supply wells varied in response to ground-water withdrawals from the deeper confined aquifer. Water levels measured in the three monitoring wells installed near public-supply wells during this study (wells 8, 32, and 18) did not follow a seasonal pattern during the 2 years like other wells influenced by streams and canals. The water level in well 18 dropped more than 5 ft between March and October 2000, the period when ground water is pumped from the deeper aquifer to meet peak demands for public supply in the area. The water level in well 32 declined about 3 ft from December 1999 to July 2001. This well is adjacent to a golf course and likely received a steady amount of recharge from unconsumed irrigation water.



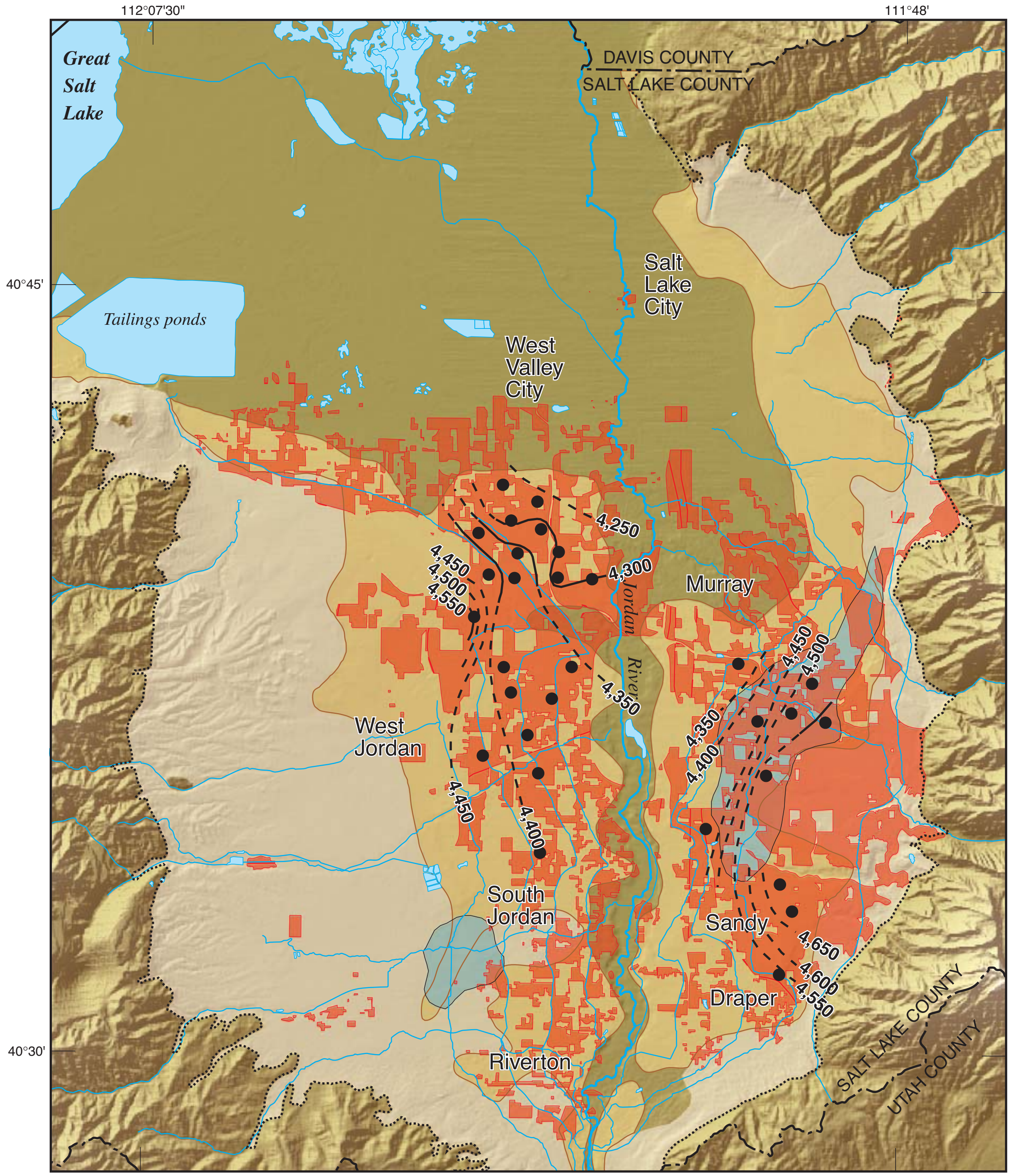
Base from U.S. Geological Survey digital line graph data, 1:100,000, 1979 and 1980
Universal Transverse Mercator projection, Zone 12

0 3 6 MILES
0 3 6 KILOMETERS

EXPLANATION

- **Area of recent residential and commercial land use**
- **Perched aquifer**—Data from Hely and others, 1971
- **Discharge area**
- **Primary recharge area**
- **Secondary recharge area**
- **Approximate boundary of basin-fill deposits in Salt Lake Valley**
- 100 ● **Monitoring well**—Number in blue is the hydraulic conductivity, in feet per day; Number in black is the transmissivity, in feet squared per day

Figure 5. Distribution of hydraulic-conductivity and transmissivity values determined from slug tests done at selected monitoring wells in Salt Lake Valley, Utah.



Base from U.S. Geological Survey digital line graph data, 1:100,000, 1979 and 1980
Universal Transverse Mercator projection, Zone 12

EXPLANATION

- **Area of recent residential and commercial land use**
- **Perched aquifer**—Data from Hely and others, 1971
- **Discharge area**
- **Primary recharge area**
- **Secondary recharge area**
- **Line of equal water-level altitude**—Shows altitude of water level in shallow aquifer and perched zones, March 2000. Contour interval is 50 feet. Dashed where approximate
- **Approximate boundary of basin-fill deposits in Salt Lake Valley**
- **Monitoring well**

Figure 6. Water-level surface for the shallow ground-water system, Salt Lake Valley, Utah, March 2000.

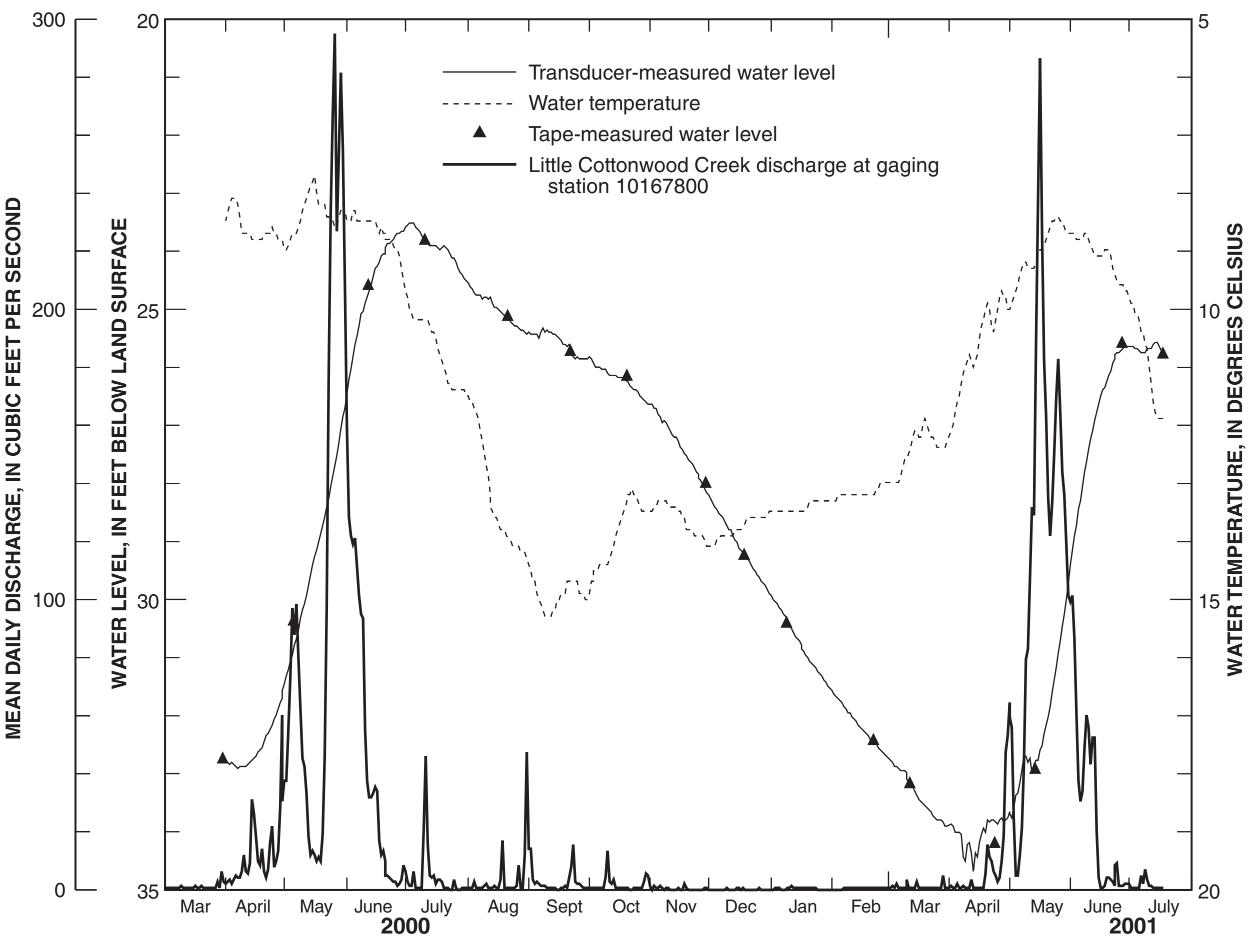


Figure 7. Water level and water temperature measured in monitoring well 26D and discharge determined for Little Cottonwood Creek at gaging station 10167800, Salt Lake Valley, Utah, 2000-2001.

Although there is not a direct correlation between water-level response in well 32 and pumping in the nearby public-supply well, withdrawals from the deeper aquifer may affect water levels in the shallow system in this area.

Water temperature was monitored in 23 wells to determine if the effects of surface-water recharge to the shallow aquifer could be measured. Only water temperature changes measured in wells 26S and 26D (fig. 7) showed a correlation with changes in ground-water levels and variations in streamflow measured in nearby Little Cottonwood Creek. The period of snowmelt runoff in the stream corresponded with an increase (rise) in water level and a decrease in water temperature in well 26D in both the spring of 2000 and 2001. The ground-water temperature rose within days after streamflow decreased to less than 5 ft^3/s in the middle of June.

Water-temperature changes recorded at other times of the year also correlated with water-level changes in well 26D. Water temperature in the well increased from about 9.0°C in June to 15.0°C in September while the water level declined about 2 ft. This ground water likely is recharged during the summer from unconsumed irrigation water and canal losses. Water levels declined about 8 ft from mid-October, when irrigation stopped, until April, when snowmelt runoff started. The water temperature in the well decreased from 14.0 to 12.5°C during this period of little recharge to the ground-water system.

Little or no change in water temperature (less than 0.2°C change) was measured in monitoring wells that had a depth to water of more than about 40 ft (fig. 8), except for well 26D. Water temperature fluctuated seasonally in the wells with a water level less than about 40 ft, except for well 26S. The shallower the water level in the well, the greater the water-temperature change measured during the study.

The coldest monthly mean air temperature in Salt Lake Valley is about -2°C in January (Ashcroft and others, 1992). The lowest mean water temperatures measured in wells with water levels less than about 40 ft below land surface occurred about 3 to 9 months later (fig. 9). Seasonal variations in air temperature can conduct as much as 10 meters (33 ft) or so below the land surface (Freeze and Cherry, 1979, p. 508) before being damped out and are the cause for most of the water-temperature variations measured in the monitoring wells. This lag in wate-temperature change with water depth is related to how long heat takes to conduct through the subsurface. Generally, the deeper the water level, the greater the lag time between the coldest air and water temperatures, and the smaller the annual water-temperature change (fig. 9). No lag in temperature was noted for wells with water levels deeper than about 31 ft below land surface (well 1).

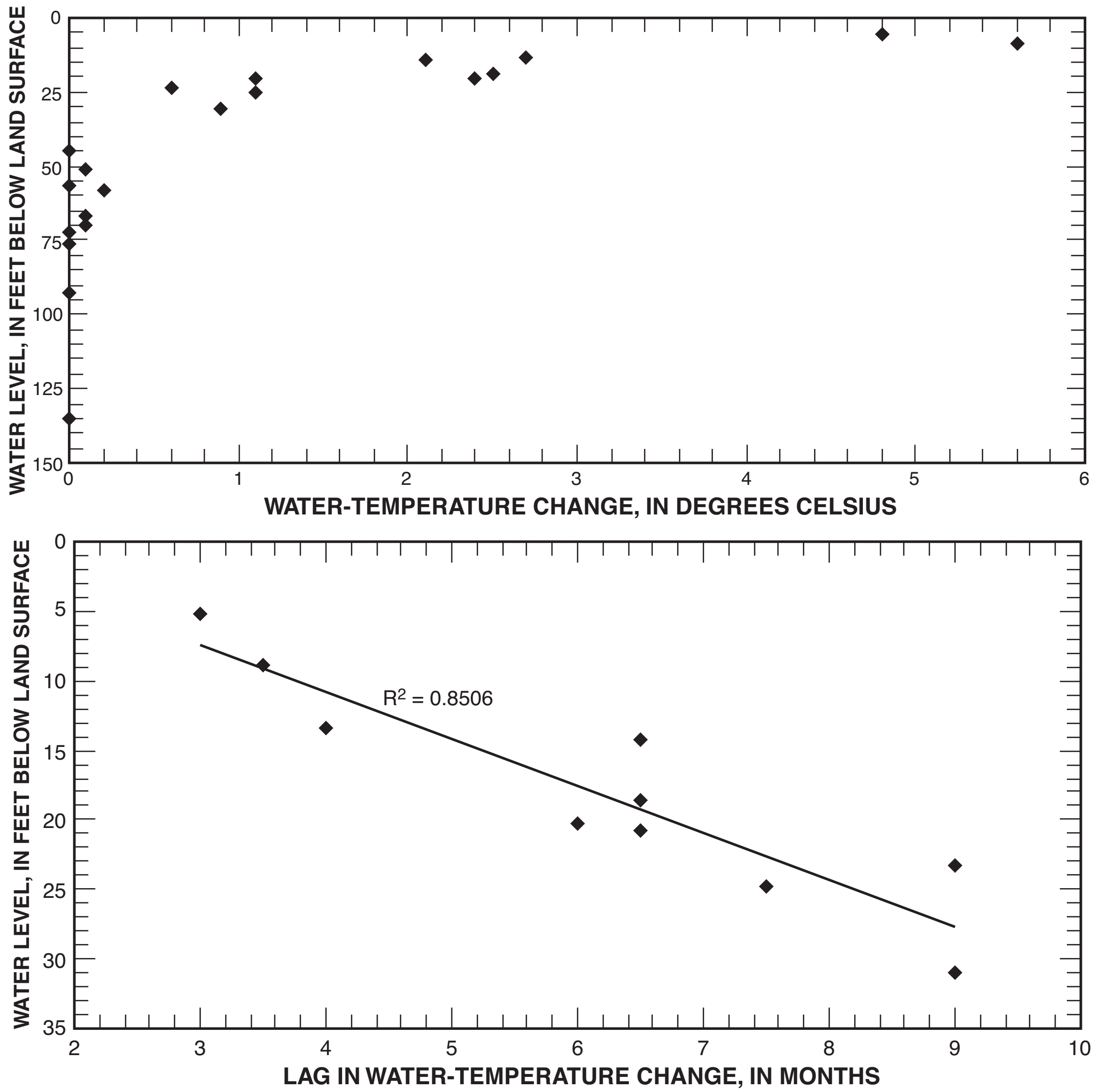


Figure 8. Relation between water level and water temperature in selected monitoring wells in Salt Lake Valley, Utah.

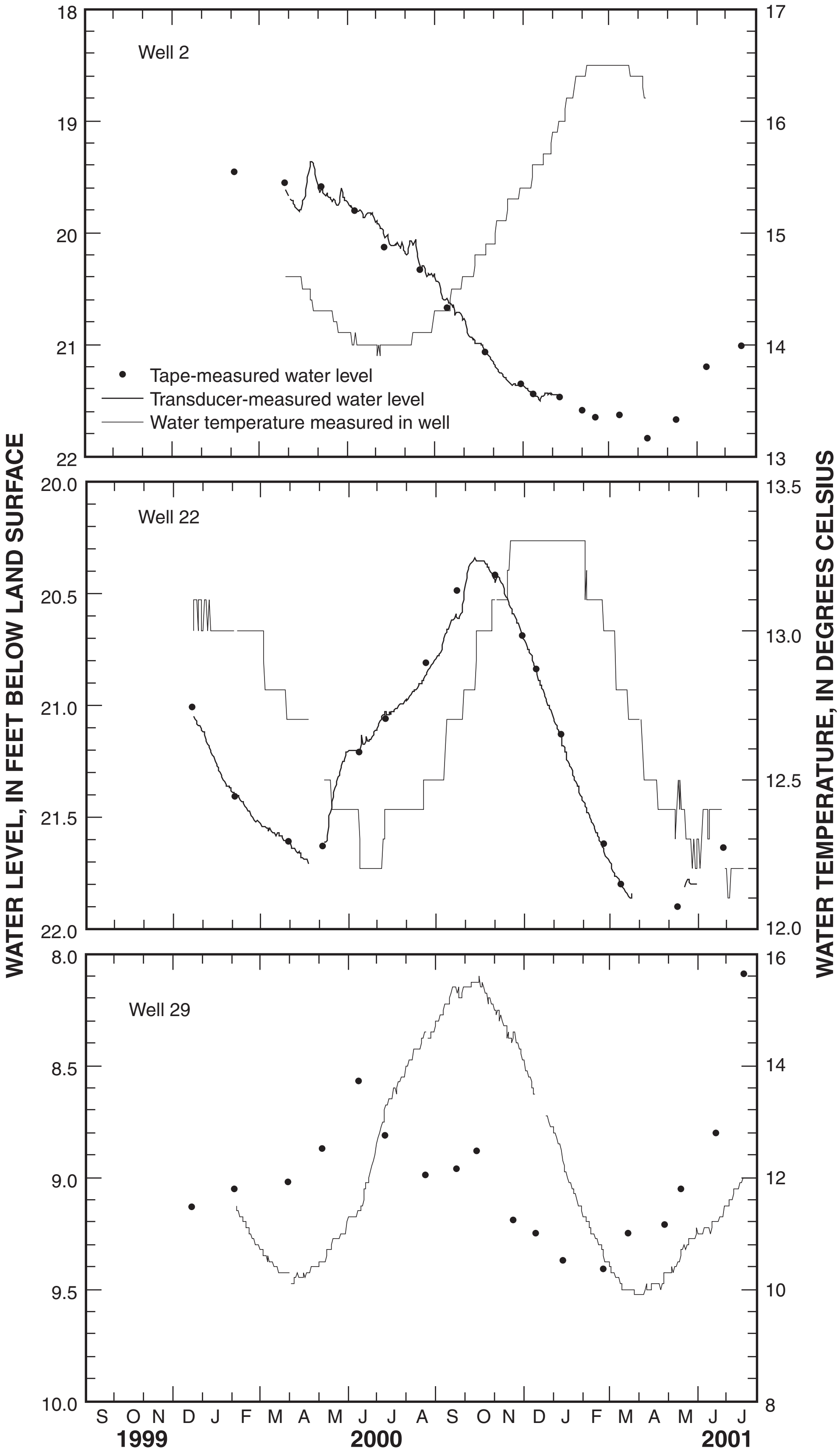


Figure 9. Water-level and water-temperature fluctuations measured in monitoring wells 2, 22, and 29, Salt Lake Valley, Utah.

The temperature of recharge water infiltrating from the land surface may cause some of the water-temperature change measured at the monitoring wells in addition to any change caused by air-temperature changes. The water level in well 22 began to rise in May and increases in water temperature followed about 2 months later (fig. 9). The water-level rise is attributed to recharge from unconsumed irrigation water that is applied during the warmer growing season. The water temperature in well 2 increased to 16.5°C in February and March, about when water levels were the lowest in the well (fig. 9). Because increases in water temperature and level do not relate more closely, it is unlikely that infiltrating recharge water affects water temperature in well 2.

Water temperature in well 29 fluctuated from a high of 15.6°C in October 2000 to a low of 10.0°C in late March 2001 (fig. 9). The water level in the well is about 9 ft below land surface and the temperature sensor was set at about 20 ft below land surface at the top of the screened interval. The period of recharge for water from this well was likely during the 1960s on the basis of a high tritium concentration (Thiros, 2003); therefore, recharge likely is not responsible for the change in water temperature. Shallow water applied for lawn irrigation is perched on top of the 14-ft-thick layer of clay and interbedded sand that confines the water in well 29 and can be seen draining near the well. Heat from this water applied during the summer probably conducts through the clay to the underlying aquifer with no actual movement of water.

Water-Level Gradients

The monitoring wells were installed in areas assumed to have a downward hydraulic gradient from the shallow to the deeper aquifers in the valley, the secondary recharge area (fig. 3), on the basis of information from well drillers' logs (Anderson and others, 1994). Comparison of water levels measured in the monitoring wells and deeper wells in the same area indicate a downward gradient on the east side of the valley (fig. 10) where water levels in some deeper wells are as much as 300 ft below those measured in nearby shallow monitoring wells.

A large area in the southeastern part of the valley contains perched ground water (fig. 10) where the bottom of a confining layer lies above the water table (Hely and others, 1971, p. 110). There is evidence that an unsaturated zone is present between the shallow ground water and the deeper aquifer near wells 29 and 32 on the basis of observations noted on drillers' logs for wells in the area. The water level in the deeper aquifer is about 300 ft below land surface compared to near land surface in well 29 and about 70 ft below land surface in well 32. Although the potential exists for shallow water to move downward because of the vertical hydraulic gradient, available water-quality information for the deeper aquifer indicates little or no mixing (Thiros, 2003). Confining layers and dispersion through the highly transmissive deeper aquifer limit connection between the aquifers in the area.

Water levels in the shallow and deeper aquifers in the secondary recharge area on the west side of the valley were similar. Seepage from canals and unconsumed irrigation are major sources of recharge to the basin-fill aquifers in this area. Canal losses on the west side are estimated to be about 15,000 acre-ft/yr as determined from measured losses to the ground-water system (Herbert and others, 1985). These losses are assumed to occur over a 6-month period when water is distributed in the canals. Although the canals mainly flow through the secondary recharge area, seepage losses can recharge both the shallow and deeper aquifers.

Water levels measured in the monitoring wells and nearby wells completed in the deeper aquifer indicate that the vertical gradient can change with time and stresses on the system. Water levels have been measured at a monitoring well completed in the deeper aquifer near monitoring well 11 from 1966 to present (fig. 11). The deeper well has three separate perforated intervals from 187 to 372 ft, and the shallow well is screened from 73 to 83 ft below land surface. Water levels measured in the two wells during this study indicate an upward gradient of about 10 ft. Water levels in the deeper well in the 1980s were about 15 ft below what was measured in the shallower well during 1999-2001 because of pumping from the principal aquifer in the area. This indicates that a downward gradient can exist if the shallow aquifer does not respond in the same way as the deeper one to changes in recharge and discharge. Withdrawals from the deeper aquifer in the area would likely cause the water level to decline below that in the shallow aquifer, resulting in a downward gradient between the aquifers.

A downward vertical hydraulic gradient was measured at the three monitoring wells nested with public-supply wells. Water levels measured in the

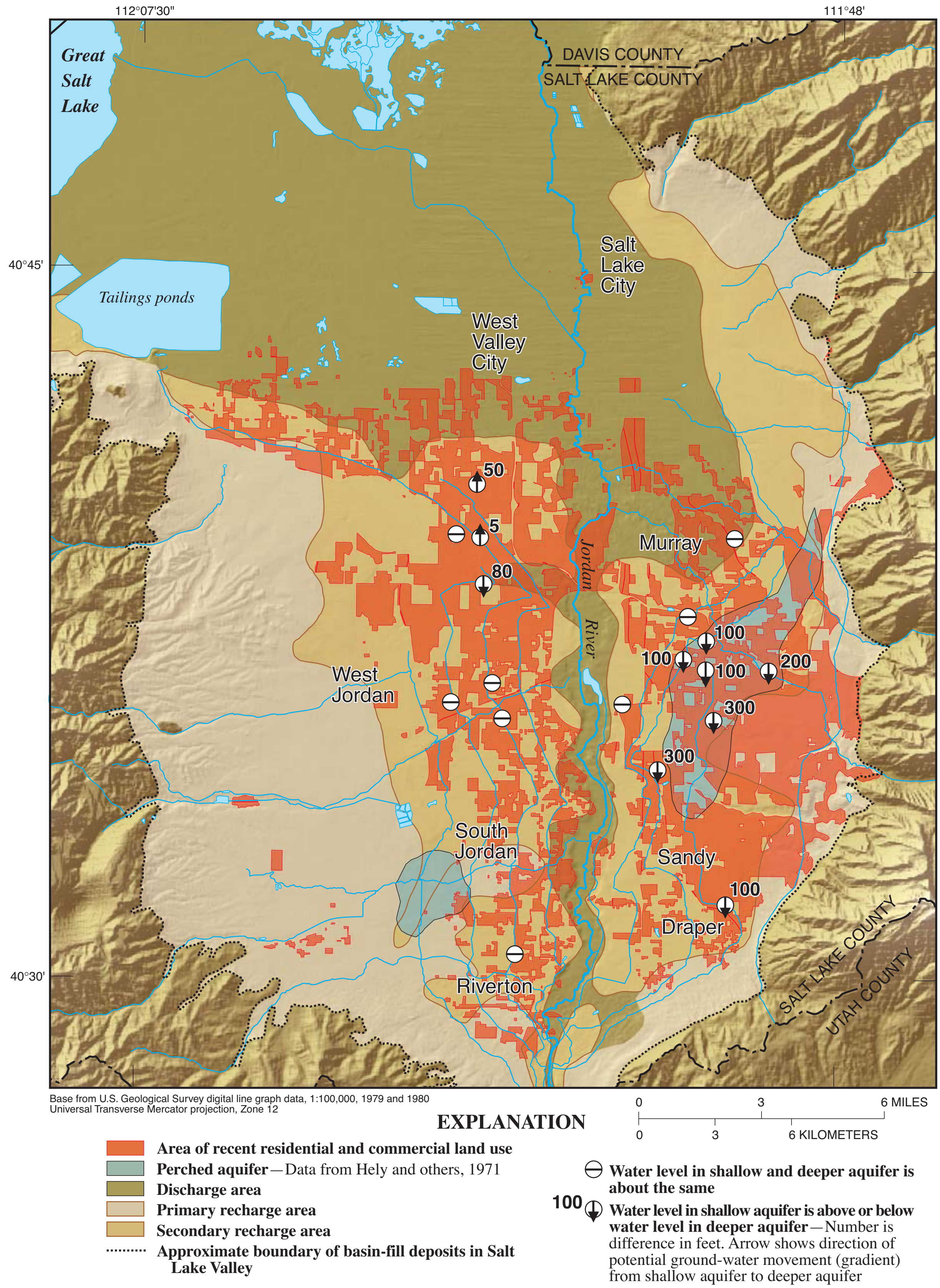


Figure 10. Direction of vertical water-level gradient and the difference between water levels in the shallow ground-water system and the deeper aquifer in Salt Lake Valley, Utah.

shallow wells declined when public-supply wells in the area were pumping, indicating a possible connection between the two aquifers. The potential exists for shallow water to move downward if the intervening confining layers are discontinuous, thin, have a high enough permeability, or if the deeper well serves as a conduit for ground-water movement as a result of its completion.

The static water level in well 8 on the west side of the valley is about the same as that measured in a nearby public-supply well. The water level in the public-supply well drops almost 100 ft when pumping. A thick layer of clay noted on the drillers' log (124 to 182 ft below land surface) is present between the water table and the top of the screened interval in the deeper well. Specific conductance measured in water from the shallow and deeper wells under both static and pumping conditions was compared. The specific conductance of water was about 1,000 µS/cm in the shallow well under both static and pumping conditions and in water from the deeper well while pumping. The specific conductance of water in the deeper well at about 90 ft below land surface under static conditions was 350 µS/cm. The specific conductance of water measured in the deeper well when it was not pumping is probably representative of the water in the aquifer at the screened interval (172 to 253 ft below land surface). The higher specific conductance measured in water pumped from the deeper well may be caused by shallow water that has moved downward despite the presence of the intervening clay layer. Volatile organic compounds and elevated nitrate concentrations measured in water from the shallow well also were detected in water from the deeper well, indicating a likely connection between the two aquifers in this area (Thiros, 2003).

Well 32, on the east side of the valley, is also near a public-supply well. The water level in the shallow well is separated from the first screened interval in the deeper well by about 440 ft of interbedded sand and clay with some gravel. The water level measured in the shallow well is about 210 ft

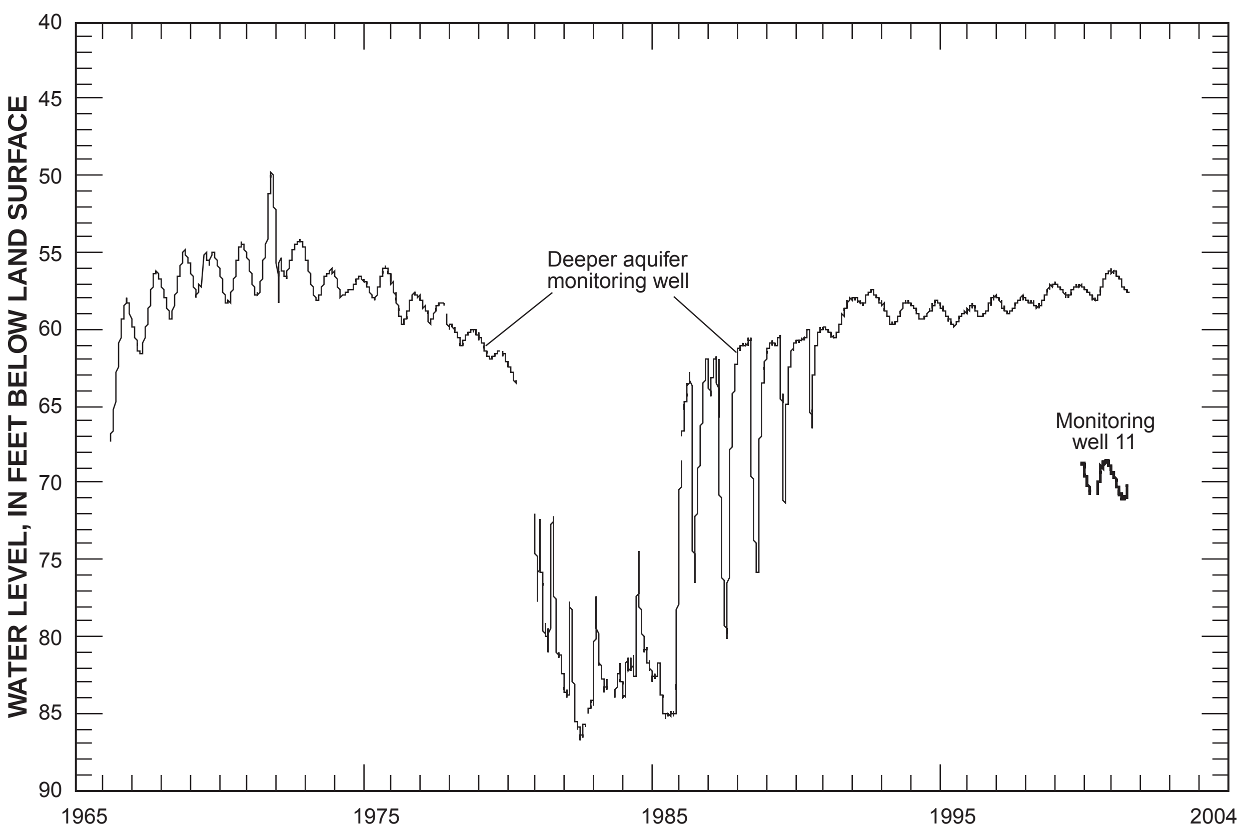


Figure 11. Water levels measured in two monitoring wells near Kearns, Utah.

above that in the deeper well under non-pumping conditions, indicating a large downward hydraulic gradient between the aquifers. Water levels in the shallow well have declined (pl. 1), probably in response to withdrawals from the deeper aquifer. Although no thick layers of clay are noted on the lithologic log of the deeper well, the layers of clay likely impede vertical movement of water. No anthropogenic compounds were detected in water from either well (Thiros, 2003).

SUMMARY

A study of recently developed residential/commercial areas of Salt Lake Valley, Utah, was done during 1999-2001 in areas in which shallow ground water has the potential to move to a deeper aquifer that is used for public supply. Thirty monitoring wells were drilled and sampled in 1999 as part of the study. The ground water at the monitoring-well sites was under either unconfined or confined conditions, depending on depth to water and the presence or absence of fine-grained deposits. The wells were completed in the shallowest water-bearing zone capable of supplying water. Monitoring-well depths range from 23 to 154 ft. Lithologic, geophysical, hydraulic-conductivity, transmissivity, water-level, and water-temperature data were obtained for or collected from the wells.

Silt and clay layers noted on lithologic logs correlate with increases in electrical conductivity and natural gamma radiation shown on many of the electromagnetic-induction and natural gamma logs. Relatively large increases in electrical conductivity determined from the electromagnetic-induction logs, with no major changes in natural gamma radiation are likely caused by increased dissolved-solids content in the ground water. Some intervals with high electrical conductivity correspond to areas where water was present during drilling.

Data were collected from 20 of the monitoring wells to estimate the hydraulic conductivity and transmissivity of the shallow ground-water system. Hydraulic-conductivity values of the shallow aquifer ranged from 30 to 540 ft/d. Transmissivity values of the shallow aquifer ranged from 3 to 1,070 ft^2/d. There is a close linear relation between transmissivity determined from slug-test analysis and transmissivity estimated from specific-capacity data.

Water-level fluctuations were measured in the 30 monitoring wells from 1999 to July 2001. Generally, water-level changes measured in wells on the west side of the valley followed a seasonal trend and wells on the east side showed less fluctuation or a gradual decline during the 2-year period. Water levels measured in monitoring wells completed in the shallow ground-water system near large-capacity public-supply wells varied in response to ground-water withdrawals from the deeper confined aquifer. The shallower the water level in the well, the greater the water-temperature change measured during the study.

Comparison of water levels measured in the monitoring wells and deeper wells in the same area indicate a downward gradient on the east side of the valley. Water levels in the shallow and deeper aquifers in the secondary recharge area on the west side of the valley were similar. Water levels measured in the monitoring wells and nearby wells completed in the deeper aquifer indicate that the vertical gradient can change with time and stresses on the system.

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