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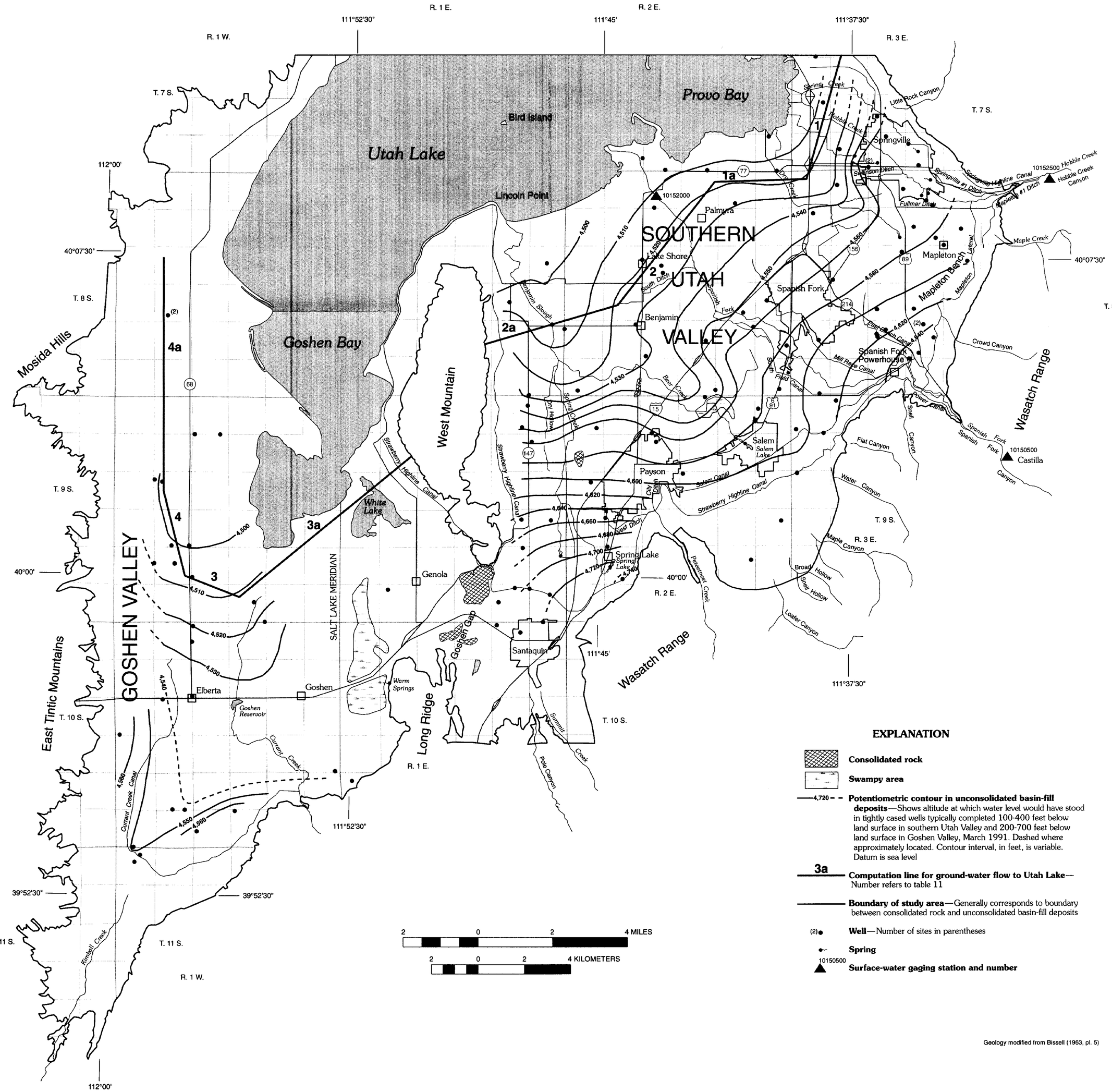

Base from U.S. Geological Survey digital data, 1:100,000, 1979, 1980, and 1981
Universal Transverse Mercator projection,
Zone 12

Map showing selected hydrologic-data sites, potentiometric surface in 1991, and computation lines for seepage to Utah Lake, southern Utah and Goshen Valleys, Utah

By
L.E. Brooks and B.J. Stolp

1995

STATE OF UTAH
DEPARTMENT OF NATURAL RESOURCES

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HYDROLOGY AND SIMULATION OF GROUND-WATER FLOW IN SOUTHERN UTAH AND GOSHEN VALLEYS, UTAH

By L.E. Brooks and B.J. Stolp
U.S. Geological Survey

Prepared by the
United States Geological Survey
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Division of Water Rights
1995

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

Multiply	By	To obtain
acre	0.4047	hectare
	4,047	square meter
acre-foot (acre-ft)	0.001233	cubic hectometer
	1,233	cubic meter
acre-foot per year (acre-ft/yr)	0.00003907	cubic meter per second
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot per mile (ft/mi)	0.1894	meter per kilometer
foot per year (ft/yr)	0.3048	meter per year
foot squared per day[1] (ft^2/d)	0.0929	meter squared per day
gallon per minute (gal/min)	0.06308	liter per second
inch (in.)	25.4	millimeter
	0.0254	meter
mile (mi)	1.609	kilometer
million gallons per day (Mgal/d)	0.04381	cubic meter per second
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:

$$°F = 1.8\ (°C) + 32.$$

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

$$°C = (°F - 32)/1.8.$$

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 is reported in milligrams per liter (mg/L) or micrograms per liter (µg/L). Milligrams per liter is a unit expressing the solute per unit volume (liter) of water. One thousand micrograms per liter is equivalent to 1 milligram per liter. For concentrations less than 7,000 milligrams per liter, the numerical value is about the same as for concentration in parts per million. Specific conductance is reported in microsiemens per centimeter at 25 degrees Celsius (µS/cm).

[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.

Hydrology and Simulation of Ground-Water Flow in Southern Utah and Goshen Valleys, Utah

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U.S. Geological Survey

ABSTRACT

The ground-water resources of southern Utah and Goshen Valleys were assessed from 1988 to 1993 to determine the effects that additional ground-water withdrawals would have on water levels, surface water, and water quality. Recharge, movement, and discharge of ground-water were emphasized. The main ground-water system in southern Utah and Goshen Valleys is in the unconsolidated basin-fill deposits. Recharge to the ground-water system from streams, canals, irrigation, precipitation, intermittent and ephemeral runoff, and subsurface inflow was estimated to be 120,000 acre-feet in southern Utah Valley and 30,000 acre-feet in Goshen Valley in 1990. Discharge from the ground-water system to springs and drains, by evapotranspiration, to wells, streams, canals, Utah Lake, and sewer systems was estimated to be 130,000 acre-feet in southern Utah Valley and 33,000 acre-feet in Goshen Valley in 1990. Release from storage from March 1990 to March 1991 was estimated to be 9,800 acre-feet in southern Utah Valley and 3,400 acre-feet in Goshen Valley.

Observed water-level fluctuations indicate that irrigation is not a major source of recharge in either valley and that precipitation is not a major source in Goshen Valley. In southern Utah Valley, water levels in March 1991 were not significantly different from water levels in March 1965. In Goshen Valley, water levels in March 1991 were higher than water levels in March 1965.

A three-dimensional, finite-difference, ground-water flow model was used to simulate the ground-water system in the unconsolidated basin-fill deposits of southern Utah and Goshen Valleys. The steady-state conditions of 1949 and annual transient-state conditions from 1949 to 1990 were used to calibrate the model. Model-computed water-level declines of less than 20 feet are projected if municipal well withdrawals increase by 10,000 acre-feet per year. Model-computed water-level declines of 20 feet in southern Utah Valley and 40 to 80 feet in Goshen Valley are projected if well withdrawal is increased by 200 percent of the 1990 withdrawals.

INTRODUCTION

Southern Utah and Goshen Valleys are south of Salt Lake City, in Utah County, north-central Utah. The two valleys cover an area of about 390 mi^2 from north of Springville to southwest of Santaquin (fig. 1). Utah Lake occupies much of the northern part of the study area and covers about 75 mi^2 of the area.

Southern Utah and Goshen Valleys are located along the western edge of the Wasatch Range, where most of Utah's population resides and where population growth is rapid. The thick basin-fill deposits contain a large volume of ground water that is used for irrigation and municipal and industrial supply. Additional withdrawal of ground water, most of which is proposed for municipal supply, has been limited because of the probable effects on surface water that drains to Utah Lake and on the lake itself, which is a source of water for irrigation and municipal use downstream. In addition, many wells flow under artesian pressure, and additional ground-water withdrawal would likely cause some wells to stop flowing. State agencies need detailed information on the hydrologic system and the effects that additional ground-water withdrawal would have on ground-water levels, surface-water flow, and water quality. To meet this need, the U.S. Geological Survey, in cooperation with the Utah Department of Natural Resources, Division of Water Rights, studied the ground-water resources of southern Utah and Goshen Valleys from 1988 to 1993.

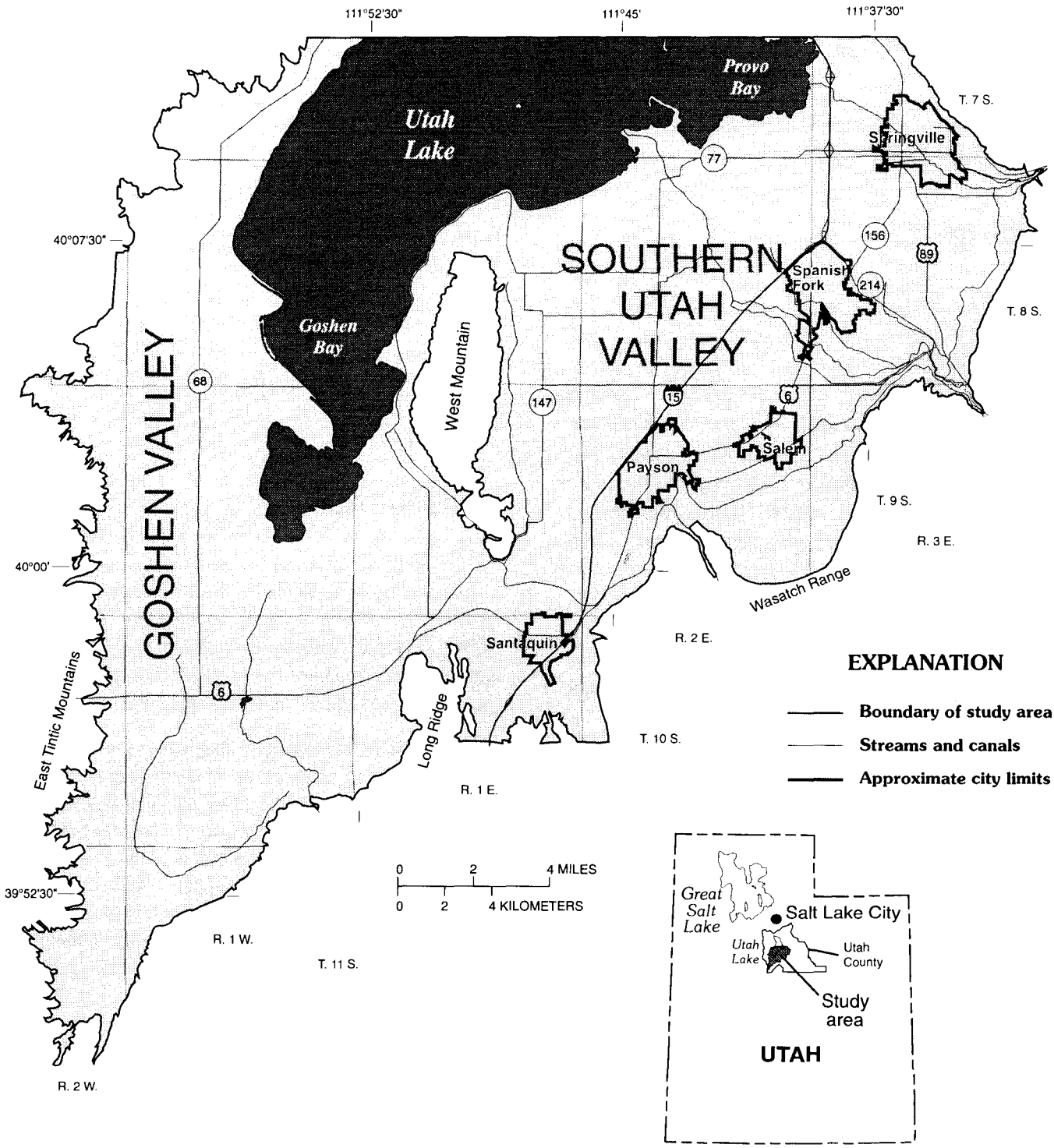


Figure 1. Location of the study area, southern Utah and Goshen Valleys, Utah.

Purpose and Scope

This report describes the ground-water system of southern Utah and Goshen Valleys in Utah County, Utah, as assessed from 1988 to 1993. The report emphasizes the recharge, discharge, and movement of ground water, and also describes water-level fluctuations, aquifer properties, ground-water quality, and possible effects of further ground-water withdrawal. Recharge, discharge, and movement are described mainly for 1990 and 1991, when most data were collected. The relation of this time period to other periods is explained. Surface-water hydrology is described as it relates to the ground-water system. Specific items discussed include recharge to the ground-water system from streams, canals, irrigation, precipitation, and subsurface inflow from consolidated rock, and discharge from the ground-water system to springs and drains, by evapotranspiration, to wells, and by seepage to Utah Lake. A three-dimensional ground-water flow model used to simulate the ground-water system in southern Utah and Goshen Valleys is described.

Data presented include discharge from springs and drains, water levels in wells and Utah Lake, results of aquifer tests in the unconsolidated basin-fill deposits, and water-quality data. The numbering system for hydrologic-data sites in Utah is shown in figure 2.

Previous Investigations

Richardson (1906) studied the hydrology of southern Utah and Goshen Valleys. Bissell (1963) studied the geology of southern Utah and Goshen Valleys. Cordova and Mower (1967) did a large-scale aquifer test in southern Utah Valley, and Cordova (1969 and 1970) prepared a basic-data report and an interpretive report on ground-water conditions in southern Utah and Goshen Valleys. Hyatt and others (1969) did a hydrologic inventory of the Utah Lake drainage area; Mundorff (1974) studied the chemical quality of surface inflow to Utah Lake; and Dustin and Merritt (1980) studied spring inflow to Utah Lake. Clark and Appel (1985) studied the ground water of northern Utah Valley. Price and Conroy (1988) prepared a map report on the ground-water resources of the southern Wasatch Front area. Stolp and others (1993) reported ground- and surface-water data for southern Utah and Goshen Valleys. The U.S. Geological Survey, in cooperation with the State of Utah, Department of Natural Resources, Division of Water Resources, publishes an annual series on ground-water conditions in Utah.

Acknowledgments

Special acknowledgment is extended to the residents and industries of southern Utah and Goshen Valleys and to the officials of cities, towns, and irrigation companies for the information and access to sites they provided. The cooperation of officials from Utah County; the Utah Department of Environmental Quality, Division of Water Quality; the Utah Department of Natural Resources, Division of Water Resources; and the U.S. Department of the Interior, Bureau of Reclamation, is appreciated.

Physiography

The study area, including streams, canals, and selected hydrologic-data sites, is shown on plate 1. Southern Utah Valley occupies the eastern part of the study area and is the larger of the two valleys. The valleys are divided by the northern extension of Long Ridge and West Mountain. North of West Mountain, the valleys join to form one valley, with Utah Lake in the center. The northern boundary of the study area is an east-west line through the middle of Township 7 South. This line was chosen by Hunt and others (1953, fig. 1) and Bissell (1963, p. 101) and was used by Cordova (1970, p. 3). The boundary is parallel to the direction of ground-water flow, and only a small amount of surface water crosses it.

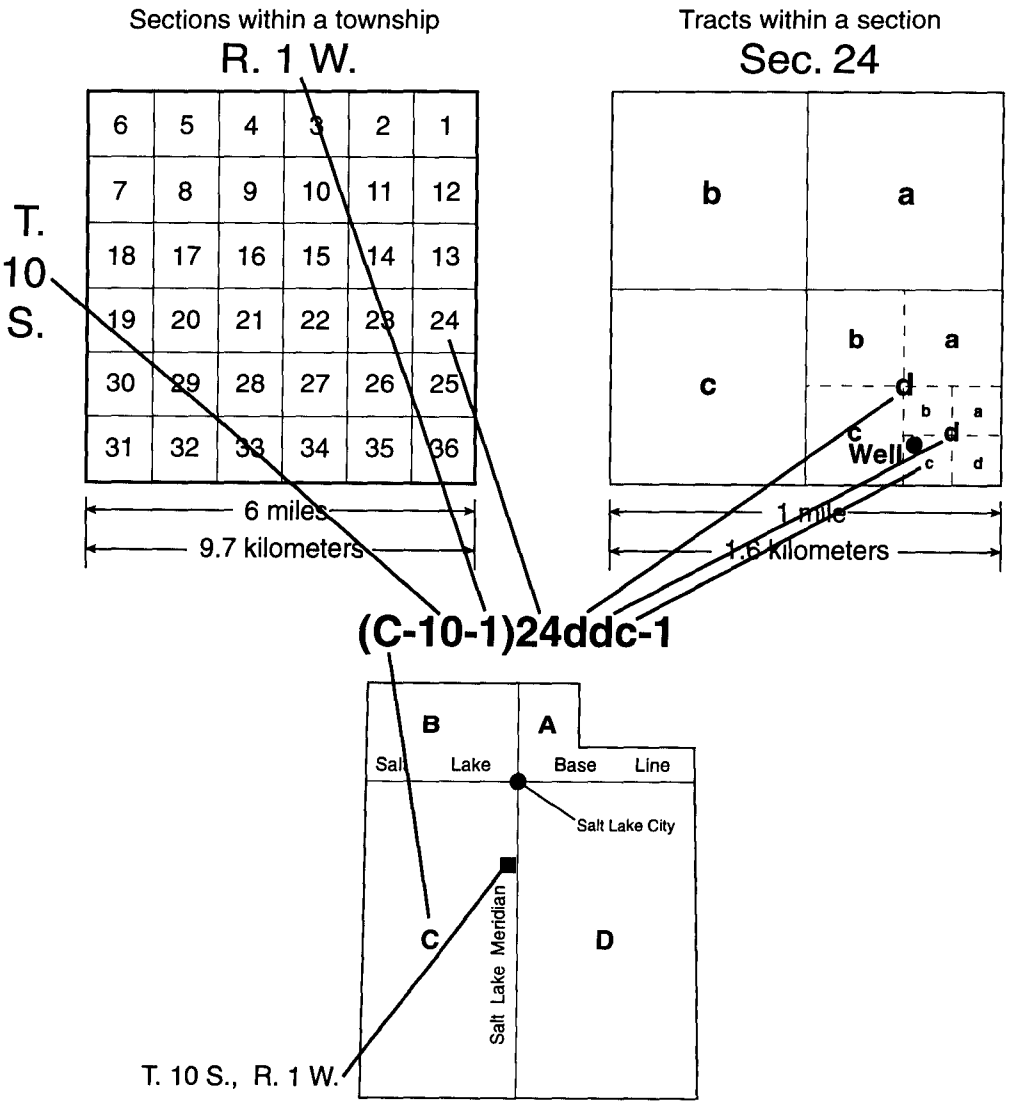
Land-surface altitude in the study area, excluding West Mountain, ranges from 4,490 ft at Utah Lake to about 5,200 ft at the southeastern edge of Utah Valley. Altitudes in the Wasatch Range, east of the study area, exceed 10,000 ft. The altitude of West Mountain, between southern Utah and Goshen Valleys, is about 6,900 ft. The highest altitude in the East Tintic Mountains, west of the study area, is about 6,400 ft.

Geology

Utah Valley is a graben formed by normal faulting during Tertiary and Quaternary time (Clark and Appel, 1985, p. 5). The Wasatch fault zone forms the eastern boundary of Utah Valley. Hunt and others (1953, p. 38) report the displacement between the uplifted block of the Wasatch Range and the downthrown block of Utah Valley to be about 7,000 ft. Bjorklund (1967, p. 19) reports displacement of as much as 10,000 ft in southern Utah Valley. A concealed fault zone, indicated by a gravity survey and the presence of thermal springs, lies along the east side of West

The system of numbering wells and springs in Utah is based on the cadastral land-survey system of the U.S. Government. The number, in addition to designating the well or spring, describes its position in the land net. The land-survey system divides the State into four quadrants separated 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 northeastern, northwestern, southwestern, and southeastern 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[1]. The lowercase letters a, b, c, and d indicate, respectively, the northeastern, northwestern, southwestern, and southeastern quarters of each subdivision. The number after the letters is the serial number of the well or spring within the 10-acre tract. When the serial number is not preceded by a letter, the number designates a well. Thus, (C-10-1)24ddc-1 designates the first well constructed or visited in the southwest 1/4, southeast 1/4, southeast 1/4, section 24, T. 10 S., R. 1 W. A location number with no serial number designates a surface-water data-collection site.

U.S. Geological Survey streamflow-gaging stations are listed in downstream direction along a main stream. Each station has been assigned a station number in the same downstream order. The complete 8-digit number for each station, such as 10150500, which appears just to the left of the station name, includes a 2-digit part number "10" plus the 6-digit downstream-order number "150500."



[1] Although 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 southeastern corner, and the shortage is taken up along the northern and western sides of the section.

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

Mountain and forms the western boundary of southern Utah Valley (Cordova, 1970, p. 11). Bissell (1963, p. 125) and Cordova (1970, p. 11) report that not much evidence exists to prove that Goshen Valley is a structural graben bounded by faults, but faulting is evident along the west sides of Long Ridge and West Mountain. Morris and Lovering (1961, p. 10) report that the physiography of Goshen Valley and the hills that border it indicate the valley is a graben, but that geophysical data do not strongly support this interpretation. Cook and Berg (1961, p. 84) suggest that faults along the western edge of Goshen Valley may be obscured by volcanic rocks.

The consolidated rock that forms the mountains surrounding southern Utah and Goshen Valleys ranges in age from Precambrian to Tertiary. The Wasatch Range contains great thicknesses of limestones that are deformed and fractured (Clark and Appel, 1985, p. 31). The East Tintic Mountains are composed mainly of quartzite and limestone of Paleozoic age and igneous rocks of Tertiary age. Faulting is very prevalent (Lindgren and Loughlin, 1919, p. 21).

Erosion of the mountains has provided the sediment that fills the valleys. The fill consists of unconsolidated to cemented and compacted lacustrine, alluvial-fan, and fluvial deposits of Tertiary and Quaternary age (Cordova, 1970, p. 11). The thickness of the unconsolidated basin-fill deposits is variable. Rocks of Tertiary and Paleozoic age are exposed north and west of Payson (Bissell, 1963, pl. 5). Some geologists estimate that near Spanish Fork, the top of the formations of Paleozoic age may be at least 18,000 ft below land surface (Dustin and Merritt, 1980, p. 15). A regional gravity survey indicates that bedrock slopes steeply from near the mountains to the middle of the valleys (Cook and Berg, 1961, pl. 13), but depths to consolidated rock are not reported. The rocks near Payson are not indicated by the gravity survey. During the Cenozoic Era, southern Utah and Goshen Valleys contained numerous lakes of varying extent, the largest of which was Lake Bonneville. The level of these lakes affected deposition of the basin-fill deposits.

Climate

The climate of southern Utah Valley is subhumid, and the climate of Goshen Valley is semi-arid (Hyatt and others, 1969, fig. 21). The 1937-91 average annual precipitation at Payson was 16.84 in., and the 1963-91 average annual precipitation at Elberta was 11.45 in. (National Oceanic and Atmospheric Administration, 1937-92). The 1990 precipitation at Payson was 15.37 in., and the 1990 precipitation at Elberta was 10.07 in. (National Oceanic and Atmospheric Administration, 1991, p. 3). Annual and average annual precipitation at Payson are shown in figure 3. The amount of precipitation increases in the valleys and in the adjoining mountains with increasing altitude.

Southern Utah and Goshen Valleys have moderate winters and summers, with a typical frost-free season in southern Utah Valley from early May to mid-October and a typical frost-free season in Goshen Valley from mid-May to early October (Hyatt and others, 1969, table 6, p. 40). Normal annual temperature at the Spanish Fork Powerhouse from 1951 to 1980 was 51.6°F, and normal annual temperature at Elberta from 1951 to 1980 was 50.3°F (National Oceanic and Atmospheric Administration, 1991, p. 10). Temperature generally decreases with increasing altitude, but pockets of cold air often form at low altitudes during winter temperature inversions. Average annual evaporation from Utah Lake at Lehi from 1931 to 1970 was 44.10 in. (Waddell and Fields, 1977, table 12).

Population and Land Use

Incorporated towns in southern Utah and Goshen Valleys had a total population of 44,624 during the 1990 census, a 97-percent increase from 1960 (Frank Hachman, Associate Director, University of Utah Bureau of Economic and Business Research, oral commun., 1992). The population of Salem and Mapleton increased the most, by 148 percent and 136 percent, respectively. Southern Utah Valley contains 99 percent of the population in the study area. Springville is the largest city, with a 1990 population of 13,950. Industry in southern Utah Valley includes several manufacturing plants, food-processing plants, an explosives plant, and small rock and clay quarries. Industry in Goshen Valley includes mining in the East Tintic Mountains.

The study area includes 193,000 acres, not including Utah Lake or West Mountain. Agriculture is the primary land use in southern Utah and Goshen Valleys, and 38 percent of the land in the study area was cropland in 1988 (fig. 4). Land use, however, gradually is changing from agricultural to municipal. As shown in the following table, agricultural land use decreased in southern Utah and Goshen Valleys by 8,800 acres from 1966 to 1988. Municipal and industrial land use in southern Utah Valley increased by 3,300 acres from 1966 to 1988, but only 7 percent of the land in the study area was municipal and industrial land in 1988.

Land classification	Acres[1] in 1966	Acres[2] in 1988
	Southern Utah Valley	
Irrigated cropland	60,700	48,900
Dry-farm land	3,400	6,200
Municipal and industrial land	9,600	12,900
	Goshen Valley	
Irrigated cropland	18,700	17,300
Dry-farm land	100	1,700
Municipal and industrial land	400	400

[1]Data from Hyatt and others, 1969, table 27.

[2]Unpublished data from Utah Department of Natural Resources, Division of Water Resources, 1988.

Central Utah Project

The Central Utah Project imports surface water from the Uinta Basin in eastern Utah to the Wasatch Front. In 1990, 114,000 acre-ft of water was released from Strawberry Reservoir, on the east side of the Wasatch Range, to Spanish Fork via the Strawberry Tunnel (Spanish Fork Water Commissioner, 1991, p. 119). Preliminary plans for completion of the Central Utah Project include an additional 20,000-30,000 acre-ft/yr of water to be delivered to southern Utah Valley and an additional 3,000 acre-ft/yr of water to be delivered to Goshen Valley (Sheldon Talbot, Central Utah Water District, written commun., 1992). The effect of this additional water on the ground-water system in southern Utah and Goshen Valleys depends on where the water is applied and if new cropland is brought into production. Recharge to the ground-water system could increase, but water levels and direction of ground-water flow probably would be affected only in small areas around the additional applications.

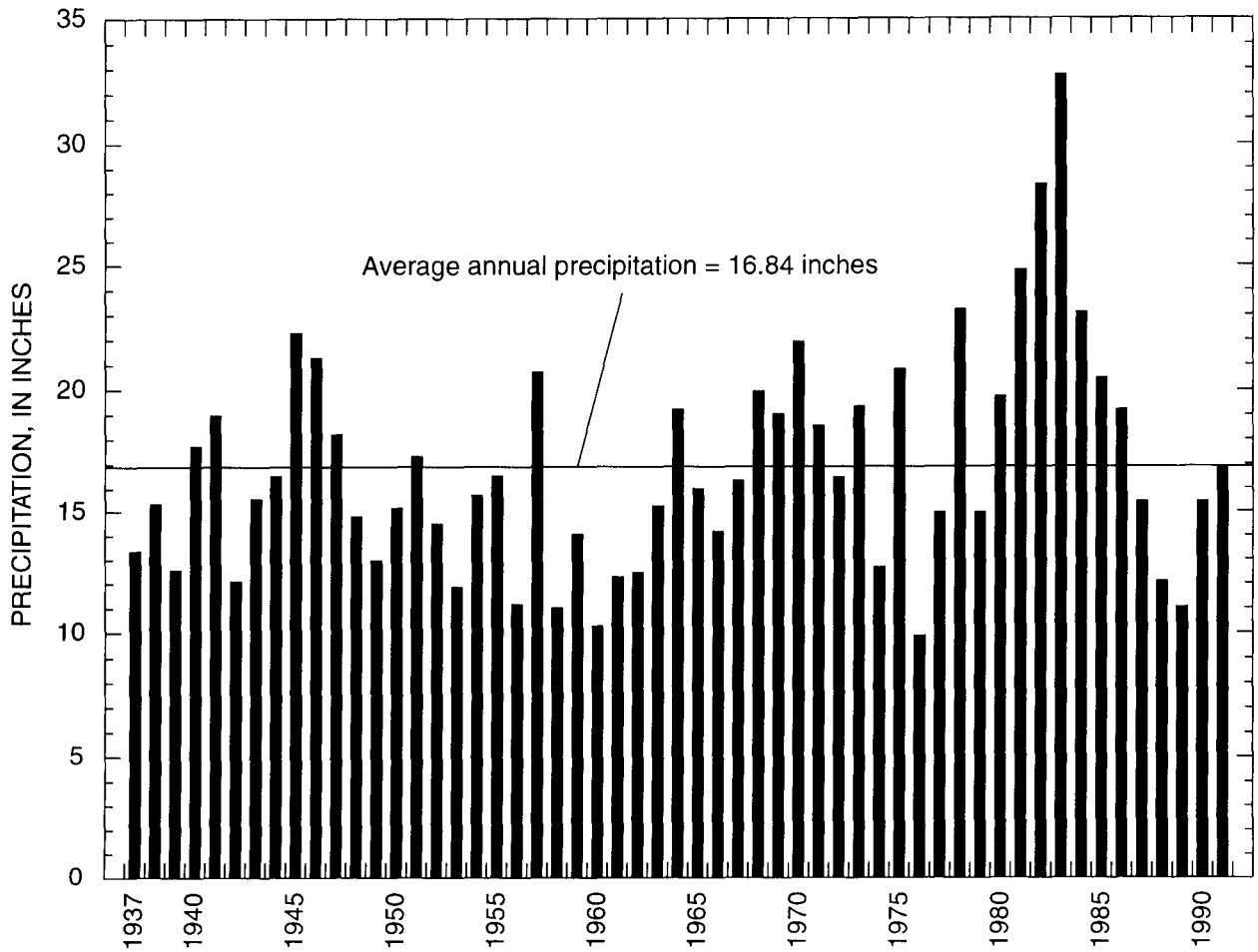


Figure 3. Annual and average annual precipitation at Payson, Utah, 1937-91

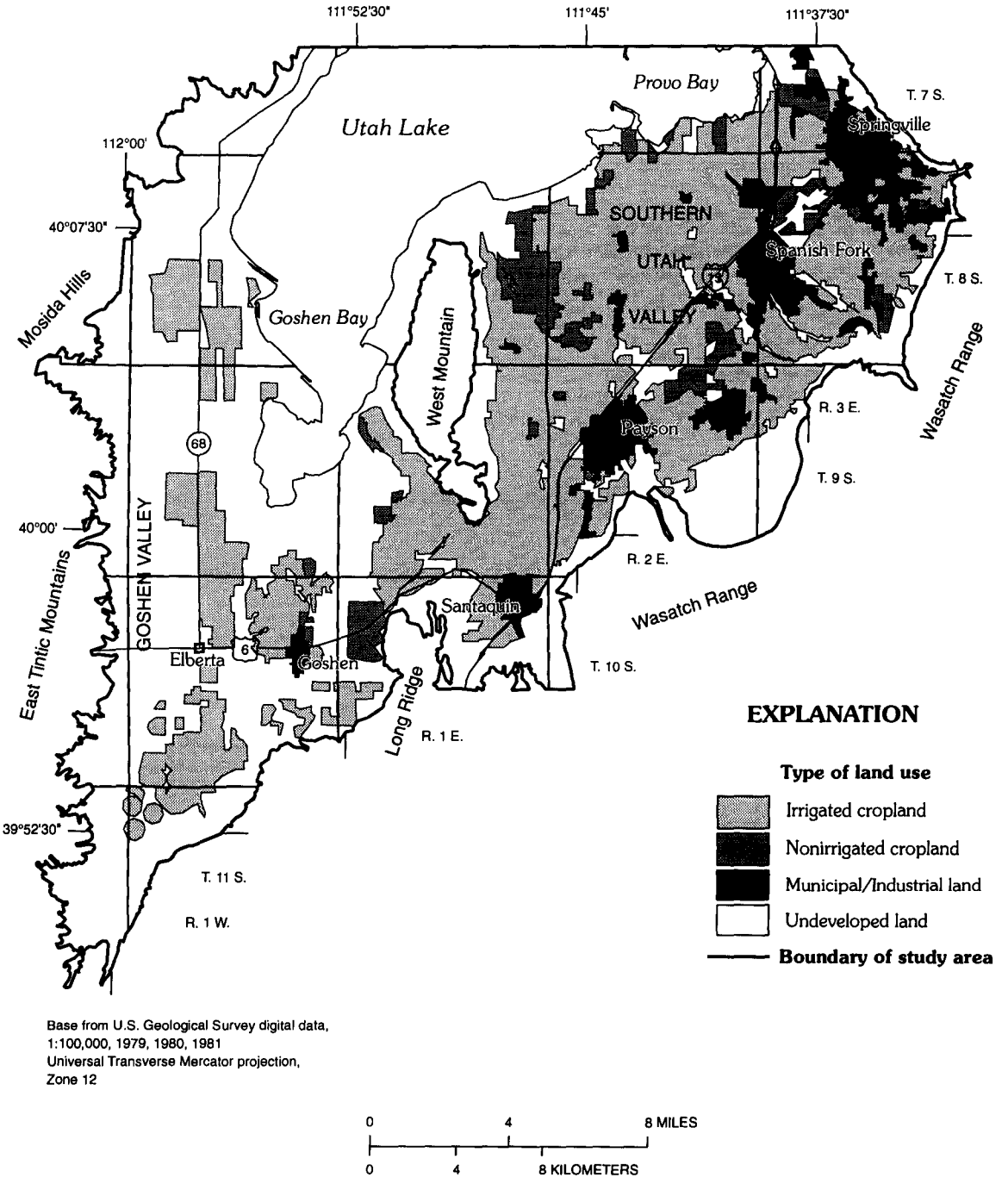


Figure 4. Generalized land use, southern Utah and Goshen Valleys, Utah, 1988.

SURFACE-WATER HYDROLOGY

Surface water was studied as it relates to ground water in the two valleys. Perennial streams, intermittent and ephemeral runoff, and canals contribute recharge to the ground-water system through downward migration from waterways and irrigated lands. In some locations, the ground-water system discharges water to perennial streams and canals.

Perennial Streams

Five perennial streams that originate in the Wasatch Range enter southern Utah Valley. The combined average annual flow that entered southern Utah Valley during 1949-90 in Spanish Fork, Hobble Creek, Maple Creek, Peteetneet Creek (also called Payson Creek), and Summit Creek was about 232,500 acre-ft/yr. The U.S. Geological Survey streamflow-gaging stations and the average annual flow for each of the five streams are listed in table 1. The periods of record for Hobble Creek, Maple Creek, Peteetneet Creek, and Summit Creek were extended using linear regression techniques to compare the flows to the natural flow of Spanish Fork at Castilla. The natural flow of Spanish Fork at Castilla is defined as measured flow at Castilla minus flow at the West Portal of Strawberry Tunnel (Spanish Fork Water Commissioner, 1949-91). In 1990, the five streams provided about 105,600 acre-ft of water for irrigation in southern Utah Valley and about 9,900 acre-ft of water through the Strawberry Highline Canal for irrigation in Goshen Valley. No perennial streams originate in the mountains that surround Goshen Valley, but Currant Creek flows from Mona Reservoir in Juab Valley, through Goshen Canyon, to Goshen Valley. Currant Creek provides an average of about 12,700 acre-ft/yr of water for irrigation in Goshen Valley. The flow of Currant Creek is controlled at Mona Reservoir and may not be closely related to the flow of Spanish Fork at Castilla; therefore, the period of

Table 1. Streamflow-gaging stations and average annual flow for perennial streams in southern Utah and Goshen Valleys, Utah

[See figure 2 for explanation of numbering system for U.S. Geological Survey streamflow-gaging stations; —, not available]

Average annual flow: In acre-feet per year.

Stream	U.S. Geological Survey streamflow-gaging station			Average annual flow	
	Number	Location	Period of record	Period of record	1949-90 (rounded)
		Southern Utah Valley			
Spanish Fork	10150500	Spanish Fork at Castilla, Utah	1899-90, 1903-17 1919-25, 1933-present	[1]168,100	174,900
Hobble Creek	10152500	Hobble Creek near Springville, Utah	1904-16, 1945-74	[2]34,270	33,900
Maple Creek	10152700	Maple Creek near Mapleton, Utah	1965-72	[3]1,690	1,800
Peteetneet Creek	10147500	Payson Creek above diversions, near Payson, Utah	1948-62	[4]9,120	10,300
Summit Creek	10147000	Summit Creek near Santaquin, Utah	1911-16, 1955-66	[5]8,980	11,600
Total				**222,200**	**232,500**
		Goshen Valley			
Currant Creek	10146500	Currant Creek near Goshen, Utah	1954-60	[6]15,800	—

[1] ReMillard and others, 1991, p. 249.
[2] U.S. Geological Survey, 1975, p. 218.
[3] U.S. Geological Survey, 1973, p. 218.
[4] U.S. Geological Survey, 1963, p. 203.
[5] U.S. Geological Survey, 1967, p. 228.
[6] Hendricks, 1963, p. 100.

record for Currant Creek was not extended. No surface water is known to be diverted for municipal or industrial consumptive use, but part of Spanish Fork is diverted through a power plant near the mouth of Spanish Fork Canyon.

Spanish Fork is the largest stream in the study area and typically provides about 75 percent of all perennial streamflow that enters southern Utah Valley (table 1). The U.S. Geological Survey maintained two continuous-recording streamflow-gaging stations on Spanish Fork. U.S. Geological Survey gaging station 10150500, Spanish Fork at Castilla, Utah, is in Spanish Fork Canyon, about 8 mi upstream from the city of Spanish Fork. The 1990 flow at this gage was 170,700 acre-ft (ReMillard and others, 1991, p. 249). Typically, about 60 percent of the flow at this station is from 652 mi^2 in the Wasatch Range and about 40 percent of the flow is from Strawberry Reservoir in the Colorado River Basin (Cordova, 1970, p. 6). In 1989 and 1990, however, 69 percent of the flow was from Strawberry Reservoir. At the canyon mouth, downstream from the gaging station, most of the water is diverted to irrigation canals during the irrigation season.

U.S. Geological Survey gaging station 10152000, Spanish Fork near Lake Shore, Utah, is about 2 mi upstream from Utah Lake. This gaging station was discontinued in 1988. During the latter part of the irrigation season, only wastewater and irrigation-return water pass this gage, and the stream has no flow at times (ReMillard and others, 1988, p. 254).

Hobble Creek is the second largest stream in the study area and typically provides about 15 percent of all perennial streamflow that enters southern Utah Valley (table 1). U.S. Geological Survey gaging station 10152500, Hobble Creek near Springville, Utah, is located about 1.5 mi upstream from the mouth of Hobble Creek Canyon and was discontinued in 1975. Estimated 1990 flow at the gaging station was 15,900 acre-ft. Downstream from the gaging station, most of the water is diverted to irrigation canals during the irrigation season.

Maple Creek, Peteetneet Creek, and Summit Creek have a combined average annual flow of about 23,700 acre-ft (table 1) and an estimated 1990 flow of 14,100 acre-ft. U.S. Geological Survey gaging stations on these streams were discontinued in 1973, 1963, and 1967, respectively. Maple Creek is almost entirely diverted above the canyon mouth, and only a small amount of water enters the valley in the stream channel. Peteetneet Creek is divided at the canyon mouth. During the nonirrigation season, creek water enters the west overflow ditch (locally called West Ditch) and eventually joins Spring Creek. During the irrigation season, flow less than 10 ft^3/s remains in the modified original channel (locally called City Ditch) and flow exceeding 10 ft^3/s enters West Ditch. Summit Creek enters the Summit Creek Irrigation Company pressure system year-round. Only flow exceeding 30 ft^3/s (the maximum capacity of the pressure system) remains in the stream channel.

Spring Creek, Dry Creek, Benjamin Slough, and other small streams originate in southern Utah Valley. These streams receive water from springs and drains, irrigation return flow, wastewater-treatment plants, and precipitation. Some of the streams are natural drains and receive direct ground-water discharge. Benjamin Slough receives ground-water discharge south and west of Spanish Fork and also receives an average annual flow of 4,400 acre-ft from Peteetneet Creek. A 12-year intermittent record of Benjamin Slough from 1937 to 1966, collected 2.5 to 3 mi upstream from Utah Lake, shows an average annual flow of 16,000 acre-ft (Cordova, 1970, p. 7).

Currant Creek, in Goshen Valley, has an average annual flow of about 15,800 acre-ft (table 1). U.S. Geological Survey gaging station 10146500, Currant Creek near Goshen, Utah, was discontinued in 1960. Downstream from the gaging station, more than 50 percent of the water is diverted to Currant Creek Canal and to a small, unnamed canal during the irrigation season.

Major Canals

Major canals delivered about 123,000 acre-ft/yr of water to about 65,900 acres in southern Utah and Goshen Valleys in 1990. The flow in canals diverted from Spanish Fork is measured and reported by the Spanish Fork Water Commissioner. The flow in canals diverted from Hobble Creek and Peteetneet Creek is not measured, but average annual flow was estimated by canal company and city officials. The flow in Currant Creek Canal is measured and is available from canal company officials. The major canals and flows are listed in table 2. Because water-level data and spring-discharge data collected during this study relate to 1990 canal flow, 1990 flow is listed in table 2 for the Spanish Fork system, the Peteetneet Creek system, and Currant Creek Canal. For the Hobble Creek system, 1990 canal flow was assumed to approximately equal average annual canal flow because canal operators indicated flows do not change from year to year.

Table 2. Flow, seepage loss, and ground-water recharge for major canals in southern Utah and Goshen Valleys, Utah, 1990

[See figure 2 for explanation of numbering system for hydrologic-data sites]

Seepage loss: J, reported, John Mendenhall, Spanish Fork Water Commissioner, oral commun., 1991; M, measured, Stolp and others, 1993, table 8; E, estimated; R, reported, Dora Edvalson, Payson City Water Department, oral commun., 1991; C, reported, Cordova, 1970, page 26.

Stream or canal	Flow at beginning of canal or stream section (acre-feet)	Seepage loss, as percent of beginning flow	Ground-water recharge to the main ground-water system (acre-feet)	Ground-water recharge to the Mapleton Bench ground-water system (acre-feet)
	Southern Utah Valley			
Spanish Fork system:				
Power canal	[1]151,300	0 J	0	0
Mapleton Lateral	[2]10,800	0 M	0	0
Strawberry Highline Canal	[2]39,600	9 M	3,600	0
Salem Canal	[2]5,300	6 E[3]	320	0
South Field Canal	[2]12,600	0 M	0	0
East Bench Canal	[2]10,700	8 M	0	860
Mill Race Canal	[2]16,000	0 M	0	0
South Ditch	[2]3,600	4 M	140	0
Total (rounded)	**98,600**		**4,100**	**900**
Hobble Creek system:				
Springville Highline Canal	[4]2,100	10 M	210	0
Fullmer Ditch	[4]4,600	30 M	0	1,400
Springville No. 1 Ditch from Hobble Creek	[4]1,000	10 E	0	100
Springville No. 1 Ditch from Mapleton Lateral	[4,5]1,500	10 E	0	150
Mapleton No. 1 Ditch	[4]500	10 E	50	0
Diversion at (D-8-3)2cca from Hobble Creek	0	0 E	0	0
Diversion at (D-8-3)2cca from Mapleton Lateral	[4,5]3,000	0 E	0	0
Diversion at (D-8-3)2ccb	[4]2,000	0 E	0	0
Swenson Ditch	[4]3,600	10 M	360	0
Total (rounded)	**13,800**		**600**	**1,600**
Peteetneet Creek System:				
City Ditch from Peteetneet Creek	[6]3,590	15 R	540	0
City Ditch from Strawberry Highline	[7]1,990	15 R	300	0
West Ditch	[6]2,980	25 E	740	0
Total (rounded)	**6,570**		**1,600**	
Total for southern Utah Valley (rounded)	**119,000**		**6,300**	**2,500**
	Goshen Valley			
Currant Creek Canal	[6]4,700	10 C	470	0

[1] Power canal delivers water to other canals and provides return flow to Spanish Fork. Diversion is not included in total.
[2] Spanish Fork Water Commissioner, 1991, p. 104.
[3] Canal has been lined with concrete since seepage runs were made.
[4] Oral communication with canal company and city officials. 1990 flow is assumed to equal average annual flow.
[5] Diversions are accounted for in Mapleton Lateral diversion and are not included in total.
[6] Oral communication with canal company officials.
[7] Diversion is accounted for in Strawberry Highline Canal diversion and is not included in total.

Water is diverted from Spanish Fork to seven major canals that serve about 47,000 acres in southern Utah Valley and about 4,700 acres in Goshen Valley. The average annual diversion to canals from Spanish Fork is 105,100 acre-ft/yr. The 1990 diversion to canals from Spanish Fork was 98,600 acre-ft (table 2). Hobble Creek is diverted to seven major canals that serve about 9,200 acres near Springville. Peteetneet Creek is diverted to two major canals that serve about 2,000 acres near Payson. Water is diverted from Currant Creek to one major canal (table 2) that serves about 8,000 acres in Goshen Valley. Average annual flow in Currant Creek Canal is 5,900 acre-ft/yr; the 1990 flow in Currant Creek Canal was 4,700 acre-ft.

Intermittent and Ephemeral Runoff

Estimated average annual flow from intermittent and ephemeral runoff entering southern Utah Valley is 14,700 acre-ft and entering Goshen Valley is 900 acre-ft. Intermittent and ephemeral runoff originates in the mountains that surround the two valleys (pl. 1) but provides no irrigation, municipal, or industrial water. Runoff occurs only during spring melting of mountain snow or during intense summer thunderstorms, and most of the flow infiltrates the unconsolidated basin-fill deposits as the runoff crosses alluvial fans or gravel deposits near the mountains. Krug and others (1989, p. 326) report annual runoff of 5 in. from the small canyons on the east and south sides of southern Utah Valley. Moody and others (1986, p. 455) report annual runoff of 0.1 in. from the hills and mountains that surround Goshen Valley and 1.0 in. from West Mountain. The boundary and area of each drainage that enters the valleys were determined from 1:24,000 topographic maps. The area of each drainage was multiplied by the annual runoff to determine average annual flow from each area. Estimated average annual flow from each drainage is listed in table 3. In 1990, natural flow in Spanish Fork at Castilla was 48 percent of 1949-90 average annual natural flow. Assuming that variation in intermittent and ephemeral runoff is similar to variation in natural flow of Spanish Fork, the intermittent and ephemeral runoff that entered southern Utah and Goshen Valleys in 1990 was 7,100 acre-ft and 450 acre-ft, respectively (table 3).

Utah Lake

Utah Lake has been operated as a reservoir since 1884, when the first dam was built on the Jordan River to provide water storage for irrigators in Salt Lake Valley. As a compromise between irrigators in Salt Lake Valley and landowners on the lake shore, Utah Lake is operated to maintain a "compromise level" of 4,489 ft above sea level. Sometimes, however, this level is not reached or is exceeded. Less-than-average precipitation causes the lake level to drop below the outlet structure, and water must be pumped into the Jordan River. Greater-than-average precipitation in the early 1980's (fig. 3) caused the lake level to rise to more than 5 ft above compromise level by June 1984 and caused extensive flooding along the shores. At an altitude of 4,489 ft, the deepest point in the lake is about 13 ft, the lake area is about 150 mi^2, and the capacity is about 898,000 acre-ft (Cordova, 1970, p. 7).

The lake receives inflow from streams, drains, ground-water seepage, and precipitation directly on the lake surface. The lake loses water by flow to the Jordan River, by small surface withdrawals, and by evaporation. Evaporation from the lake surface averages 341,000 acre-ft/yr (Hyatt and others, 1969, fig. 59).

Chemical Quality

Streams that enter southern Utah Valley from the surrounding mountains had a dissolved-solids concentration that ranged from 187 to 565 mg/L, and streams that originate in southern Utah Valley had a dissolved-solids concentration that ranged from 359 to 3,410 mg/L in 1971 and 1972 (Mundorff, 1974, table 20). Water samples collected from Spanish Fork at Castilla had a range of dissolved-solids concentrations from 355 to 565 mg/L and specific-conductance values of 406 to 818 µS/cm. Water samples collected from Spanish Fork near Lake Shore had a dissolved-solids concentration that ranged from 266 to 1,110 mg/L, with the highest concentration occurring in late summer. Samples collected from Benjamin Slough near Lake Shore had a dissolved-solids concentration that ranged from 692 to 1,540 mg/L. Water from small drains had the highest dissolved-solids concentration. Water samples collected from Spanish Fork at Castilla had a range of specific-conductance values of 375 to 940 µS/cm in 1989 and 1990 (ReMillard and others, 1991, p. 328). Water from streams that enter southern Utah Valley typically is suitable for municipal, industrial, and agricultural use but is very hard (Mundorff, 1974, p. 48). The water available for agriculture has medium salinity hazard and low sodium hazard during the irrigation season.

Water samples collected from Currant Creek in Goshen Canyon had a dissolved-solids concentration

Table 3. Area of drainage basin, estimated flow, and estimated ground-water recharge from intermittent and ephemeral drainages, southern Utah and Goshen Valleys, Utah

[See figure 2 for explanation of numbering system for hydrologic-data sites]

Name of drainage	Area of drainage basin (acres)	Estimated average annual flow (acre-feet per year)	Estimated 1990 flow (acre-feet)	Estimated annual ground-water recharge (acre-feet per year)	Estimated 1990 ground-water recharge (acre-feet)
Southern Utah Valley					
Little Rock Canyon	1,300	540	260	490	230
Unnamed, enters valley at (D-7-3)35c	2,200	920	440	830	400
Area north of Hobble Creek	2,700	1,120	540	1,000	490
Crowd Canyon	900	380	180	340	160
Drainages entering valley between Hobble Creek and Spanish Fork	3,800	1,600	770	1,400	690
Snell Canyon	900	380	180	340	160
Flat Canyon	800	330	160	300	140
Water Canyon	1,600	670	320	600	290
Maple Canyon	2,900	1,200	580	1,100	520
Snell Hollow and Broad Hollow	1,200	500	240	450	220
Loafer Canyon	2,600	1,100	530	990	480
Drainages entering valley between Spanish Fork and Peteetneet Creek	6,000	2,500	1,200	2,200	1,080
Drainages entering valley between Peteetneet Creek and Summit Creek	3,200	1,300	620	1,200	560
Drainages west of Summit Creek	1,900	790	380	710	340
Pole Canyon	2,500	1,000	480	900	430
Drainages entering valley from West Mountain	4,600	380	180	340	160
Total (rounded)	**39,000**	**14,700**	**7,100**	**13,200**	**6,400**
Goshen Valley					
Drainages entering valley between Long Ridge and Currant Creek	3,600	30	14	27	13
Drainages entering valley between Currant Creek and Kimball Creek	7,200	60	29	54	26
Kimball Creek	8,400	70	34	63	30
Drainages entering valley between Kimball Creek and U.S. Highway 6	19,000	160	77	140	69
Drainages entering valley north of U.S Highway 6	18,000	150	72	140	65
Drainages entering valley from West Mountain	5,400	450	220	400	200
Total (rounded)	**62,000**	**900**	**450**	**800**	**400**

that ranged from 494 to 1,060 mg/L in 1964 (Cordova, 1969, table 10). Water samples collected from Currant Creek near the town of Goshen had a dissolved-solids concentration that ranged from 919 to 935 mg/L in 1972 (Mundorff, 1974, table 20). Water from Currant Creek has high salinity hazard and low sodium hazard (Mundorff, 1974, p. 56) and typically contains a dissolved-solids concentration too high for municipal use.

Dissolved-solids concentration in Utah Lake has historically averaged about 1,090 mg/L in the main part of the lake and has ranged from 500 mg/L during the high lake levels of the early 1950's to 10,000 mg/L during the low lake levels of the 1930's (Central Utah Water Conservancy District, 1989, p. 2). Dissolved-solids concentration is higher in Goshen Bay and lower in Provo Bay than in other parts of the lake (Fuhriman and others, 1981, p. 63). Water from Utah Lake is not suitable for municipal use unless treated and diluted with water that is of suitable quality for municipal use. The average dissolved-solids concentration in Utah Lake is only 100 mg/L less than the agricultural water-quality limit set by the State of Utah for Utah Lake (Central Utah Water Conservancy District, 1989, p. 5). Completion of the Central Utah Project is expected to divert more of the fresh water that enters Utah Lake from the Provo River, thereby causing the dissolved-solids concentration in Utah Lake to increase, unless measures are taken to reduce the effect (Central Utah Water Conservancy District, 1989, p. 2).

GROUND-WATER HYDROLOGY

The main ground-water system in southern Utah and Goshen Valleys is in the unconsolidated basin-fill deposits. A conceptual model of the main ground-water system, a perched ground-water system, probable direction of ground-water flow, and areas of recharge and discharge are shown schematically in figure 5. The unconsolidated basin-fill deposits consist of interbedded and lenticular deposits of gravel, sand, silt, and clay. Lacustrine, alluvial, and colluvial processes sorted the deposits according to the level of valley lakes and location of streams at the time of deposition. The methods of deposition created alternating and interfingering layers and lenses, and vertical and horizontal heterogeneity. Along the mountain fronts, colluvial processes resulted in deposition of poorly sorted clay, sand, and gravel deposits. Away from the mountain fronts, on benches and alluvial fans, lacustrine processes resulted in deposition of well-sorted sand and gravel deposits and some well-sorted clay layers. In the center of the valleys and near Utah Lake, the basin-fill deposits are composed mainly of silt and clay. In the stream channels of Spanish Fork and Hobble Creek, alluvial processes resulted in deposition of more sand and gravel than in the surrounding basin-fill deposits.

The basin-fill deposits are thousands of feet thick. Few wells are deeper than 1,000 ft in southern Utah and Goshen Valleys, so not much is known about the deeper deposits. Gates (1987, p. 79) reports that in most valleys in the Great Basin, the sands and gravels that yield water readily to wells occur in the upper 700 to 1,500 ft of deposits. Below this, cementation and compaction reduce permeability of the deposits. Because there is virtually no subsurface data available for southern Utah and Goshen Valleys deeper than 1,000 ft and because only a few production wells are known to withdraw ground water from deeper than 700 but less than 1,000 ft, the ground-water-flow system for the study area will be limited to the upper 1,000 ft of basin-fill deposits; thus, recharge, discharge, and movement discussed in this report are assumed to occur only in the upper 1,000 ft of basin-fill deposits.

No wells are known to have been drilled into the rock outcrops near Payson, and permeability of the rock is unknown. The outcrops are assumed to be part of the main ground-water system. The outcrops are small and isolated, and ground water in the unconsolidated basin-fill deposits probably also flows through the outcrops. The potentiometric contours shown on plate 1 indicate that the rock does not influence ground-water flow patterns appreciably. It is possible that the rock has a lower permeability than the surrounding basin-fill deposits and influences ground-water flow patterns near the outcrops. It also is possible, however, that the rock has a permeability within the range of permeability of the heterogeneous basin-fill deposits, and does not influence ground-water flow patterns.

The consolidated rock that surrounds the valleys contains water, but few wells have been drilled into it. Permeability of the consolidated rock is unknown, but measured discharge from a well completed in limestone in nearby Cedar Valley is 3,200 gal/min. This report considers water in the consolidated rock around the valleys only as a source of recharge for the main ground-water system in the unconsolidated basin-fill deposits.

Occurrence

The lenses and alternating layers of clay, silt, sand, and gravel have previously been divided into four

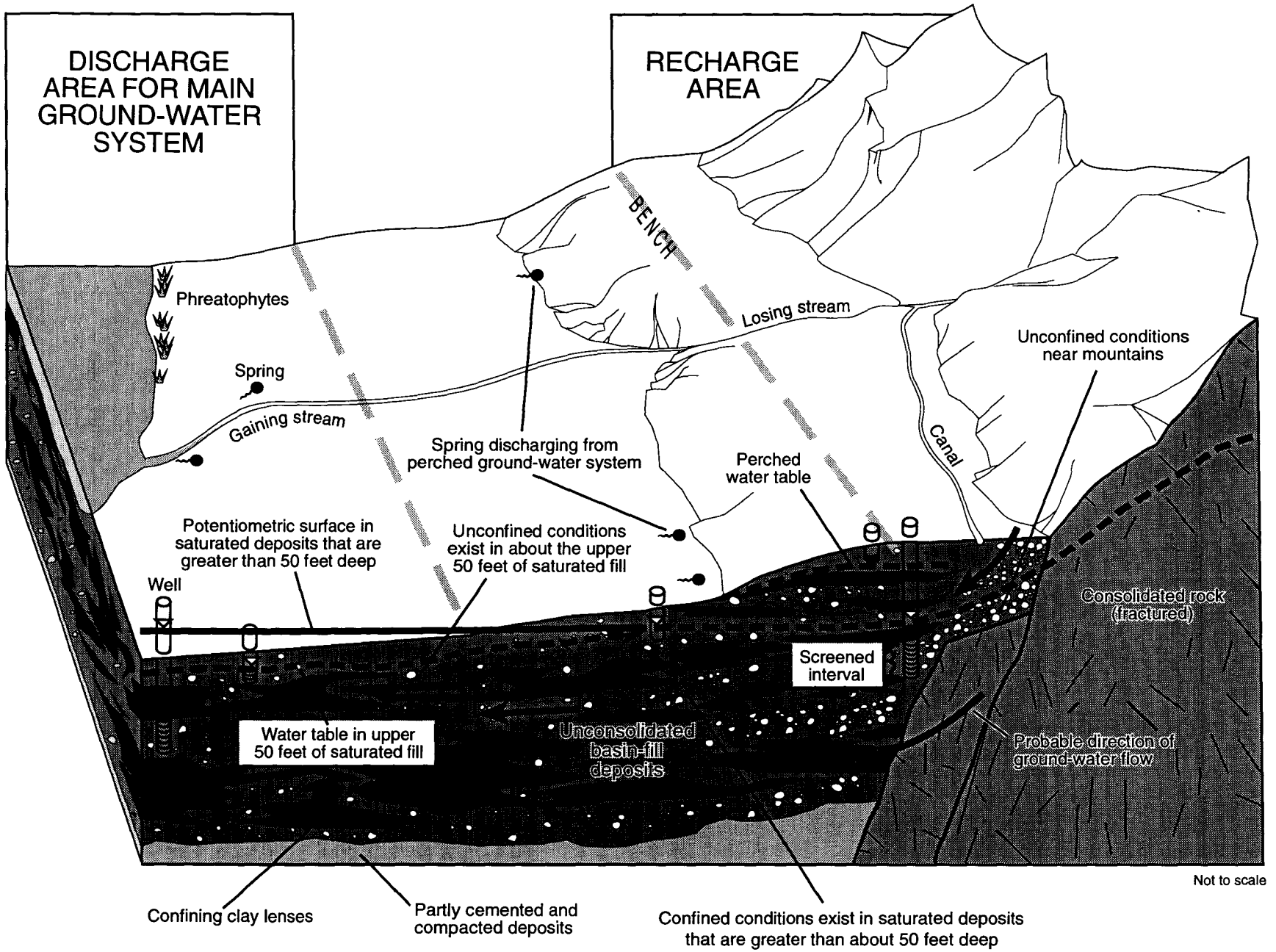


Figure 5. Schematic block diagram of ground-water system, probable direction of ground-water flow, and areas of recharge and discharge, southern Utah and Goshen Valleys, Utah.

separate aquifers (Cordova, 1970, p. 13). In descending order, the aquifers were named the Lake Bonneville Group of Pleistocene age, the shallow artesian aquifer of Pleistocene age, the deep artesian aquifer of Pleistocene age, and the artesian aquifer of Tertiary age. This study considers the basin-fill deposits to be one main ground-water system with varying horizontal and vertical permeability. The four aquifers were originally proposed for northern Utah Valley (Hunt and others, 1953, pl. 4), but Clark and Appel (1985, p. 40) report that the aquifers are difficult to trace across all of northern Utah Valley. Richardson (1906, p. 53) reports uniform layers around the Springville area, but lenticular deposits around Benjamin, Lake Shore, and Palmyra. Richardson (1906, p. 55) also reports varying stratigraphy in Goshen Valley. Drillers' logs examined during this study provided no evidence of four distinct aquifers throughout southern Utah and Goshen Valleys. Horizontal and vertical water movement occurs mainly through the coarser deposits, and water in deep deposits is not isolated from water in shallow deposits. Todd (1959, p. 44) states that leaky or semiconfined aquifers are common in alluvial valleys or former lake basins.

Depth to ground water ranges from about 5 ft below land surface near Utah Lake to about 400 ft below land surface near the mountains. Ground water occurs in basin-fill deposits in southern Utah and Goshen Valleys under unconfined and confined conditions. Ground water is unconfined in coarse-grained deposits near the mountains but becomes confined toward the center of the valleys as clay lenses become more predominant in the basin-fill deposits. The confined zones in the center of the valley are lateral extensions of the unconfined zones near the mountains. In the central parts of the valley, ground water is unconfined near land surface but becomes confined at depth. Drillers' logs examined during this study did not indicate a continuous clay layer that would cause an abrupt change from unconfined to confined conditions. Interfingering clay layers at different depths probably cause a low vertical permeability and effectively confine the water in deeper deposits. Water-level data indicate that most flowing wells are completed at depths greater than about 50 ft; therefore, unconfined conditions exist within about the first 50 ft of saturated basin-fill deposits throughout both valleys. Both unconfined and confined water are considered to be part of the main ground-water system. One notable exception is the Mapleton Bench system.

Mapleton Bench, between Hobble Creek and Spanish Fork Canyon in southern Utah Valley, is underlain by at least one thick, continuous layer of clay, locally mixed with sand and silt. The clay isolates the unconfined ground-water system in this area from the main ground-water system (fig. 6). Downward migration of water from canals, septic tanks, irrigation, and precipitation in this area is restricted by the clay layer and causes high water levels in shallow wells. Richardson (1906, p. 53) reported that when irrigation began on Mapleton Bench, water levels rose by about 30 ft, and many springs that did not exist before irrigation began formed along the outer margins of the bench. The unconsolidated basin-fill deposits that underlie Mapleton Bench have been cut by the channels of Hobble Creek and Spanish Fork, allowing the water from the Mapleton Bench system to discharge to springs, Hobble Creek, and Mill Race Canal. The clay continues a short distance north of Hobble Creek, and some water from canals, irrigation, and precipitation north of Hobble Creek contributes flow to Hobble Creek and to springs, not to the main ground-water system. The clay layer is not well defined south of Spanish Fork, and only the main ground-water system is known to exist in this area. Recharge to and discharge from the Mapleton Bench system are not included in the ground-water budget presented in this report for the main ground-water system.

An estimated budget for the main ground-water system in the unconsolidated basin-fill deposits is presented in table 4. The individual budget elements are discussed in the following sections. Ground-water flow from southern Utah Valley to Goshen Valley has been estimated for 1990 and is included in table 4 and discussed in detail in the "Movement" section of the "Ground-water hydrology" section of this report.

Recharge

Recharge to the main ground-water system was estimated to be about 120,000 acre-ft in southern Utah Valley and about 30,000 acre-ft in Goshen Valley in 1990 (table 4). Recharge in 1990 is emphasized in this report because water-level measurements and measurements to determine ground-water discharge to springs and drains relate to recharge that occurred in 1990. Annual variability in some sources of recharge is discussed in the following sections. Recharge is from perennial streams and major canals, irrigation and precipitation, intermittent and ephemeral runoff, subsurface inflow from the consolidated rock of the bordering mountains, and subsurface inflow from channel deposits of streams that enter the valley. The main ground-water system receives most recharge near the moun-

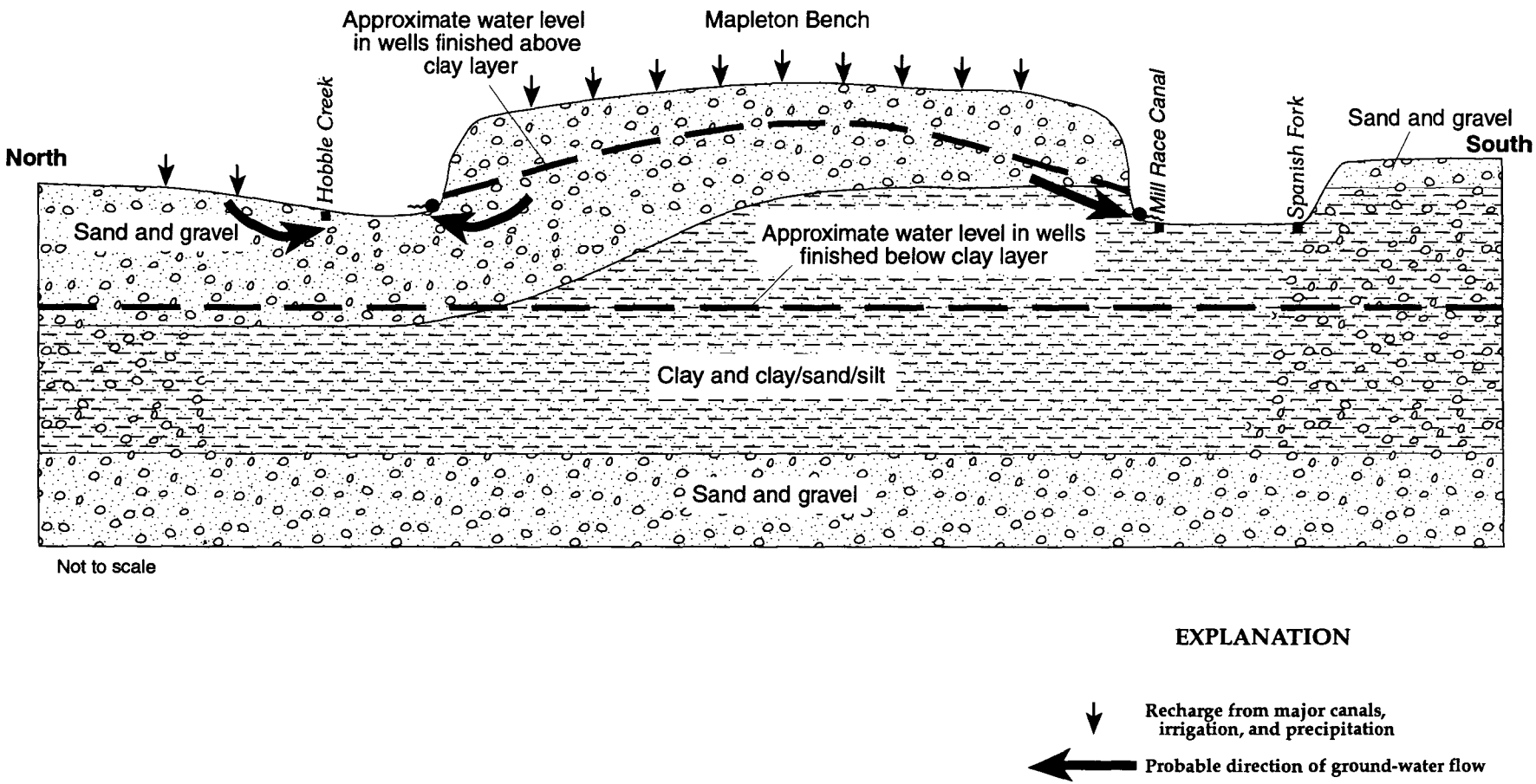


Figure 6. Schematic diagram of the ground-water system in the Mapleton Bench area of Utah and probable direction of ground-water flow.

Table 4. Estimated budget for the main ground-water system in the unconsolidated basin-fill deposits, southern Utah and Goshen Valleys, Utah, 1990

[—, not applicable]

Budget element	Flow, in acre-feet		
	Southern Utah Valley	Goshen Valley	Study area
Recharge			
Perennial streams and major canals	33,400	8,100	41,500
Irrigation and precipitation	14,900	400	15,300
Intermittent and ephemeral runoff	6,400	400	6,800
Intervalley flow[1]	0	7,800	—
Subsurface inflow	[2]65,000	[2]13,000	[2,3]79,000
Total recharge (rounded)	**120,000**	**30,000**	**[1]143,000**
Discharge			
Springs and drains	42,700	0	42,700
Evapotranspiration	26,000	14,000	40,000
Pumped wells	14,000	13,500	27,500
Flowing wells	4,400	0	4,400
Perennial streams and major canals	20,700	2,200	22,900
Utah Lake	9,600	3,600	13,200
Sewer systems	5,000	0	5,000
Intervalley flow[1]	7,800	0	—
Total discharge (rounded)	**130,000**	**33,000**	**[1]156,000**
Water going into (+) or out of (-) storage[4] (rounded)	-9,800	-3,400	-13,200

[1]Intervalley flow not used for study area total.
[2]Calculated as a residual of the discharge minus all other forms of recharge.
[3]Total for study area does not equal sum of two valleys because of rounding error.
[4]Water going into (+) storage is considered to be discharge and should be added to total discharge; water going out of (-) storage is considered to be recharge and should be added to total recharge.

tains, where the surficial sediments are permeable enough to allow infiltration. From 1969 to 1978, an estimated 6,200 acre-ft/yr of water from mining activities in the East Tintic Mountains was delivered to dispersion ponds near Elberta (Mack Croft, Utah Department of Environmental Quality, Division of Water Quality, written commun., 1989). Some of this water recharged the main ground-water system in Goshen Valley during these years.

Perennial Streams and Major Canals

Downward migration of water from perennial streams and major canals supplied an estimated 33,400 acre-ft of recharge to the main ground-water system in southern Utah Valley, an estimated 8,100 acre-ft to the main ground-water system in Goshen Valley, and an estimated 2,500 acre-ft to the Mapleton Bench system in 1990. Flow of perennial streams, seepage loss, and ground-water recharge are listed in table 5. Seepage from canals (table 2) was determined only for water supplied from perennial streams; the volume of water in canals from springs and wells is small in comparison to the volume from perennial streams, and seepage of this water was considered insignificant. Loss in smaller canals and ditches in distribution systems was considered to be recharge from irrigated lands, not seepage from canals. All canal loss was assumed to recharge

either the main ground-water system or the Mapleton Bench system. Most canal operators report a higher percentage of loss when the system is first opened in the spring, and this higher loss was assumed to be the result of the combined seepage losses and loss to temporary bank storage along the canal. As bank storage filled, total losses decreased to only seepage losses through the canal bottom.

Seepage studies done in 1965 and 1966 on Spanish Fork, Hobble Creek, and major canals diverted from those streams were reanalyzed as part of this study (Stolp and others, 1993, table 8). Seepage studies consist of measuring streamflow in several locations along the stream channel and measuring all inflow and outflow to the stream between the locations. Change in flow between measurement locations that cannot be accounted for by measured inflow and outflow is assumed to be ground-water recharge or discharge. Analysis of the seepage studies indicated that some reaches of streams and canals lost or gained a specific percentage of flow and some reaches gained a specific amount of flow, regardless of flow in the reach. The

Table 5. Flow, seepage loss, and ground-water recharge from selected reaches of perennial streams, southern Utah and Goshen Valleys, Utah, 1990

[See figure 2 for explanation of numbering system for hydrologic-data sites]

Seepage loss: E, estimated; M, measured, Stolp and others, 1993, table 8; C, reported, Cordova, 1970, p. 26.

Stream	Flow at beginning of reach (acre-feet)	Seepage loss, as percent of beginning flow	Ground-water recharge to the main ground-water system (acre-feet)
	Southern Utah Valley		
Spanish Fork System:			
Spanish Fork between the Power Canal diversion and the Mapleton Lateral	[1]23,100	10 E	2,310
Spanish Fork between Interstate-15 and Palmyra	[1]86,900	25 M	21,700
Hobble Creek System:			
Hobble Creek between (D-8-4)6abb and (D-8-3)1cac	[2]9,250	10 E	920
Hobble Creek between (D-8-3)1cac and the Mapleton Lateral	[2]7,560	20 M	1,510
Summit Creek at U.S. Geological Survey streamflow-gaging station	[3]6,150	10 E	620
Total recharge for southern Utah Valley (rounded)			**27,100**
	Goshen Valley		
Currant Creek below canal diversion	[4]11,100	10 C	1,100
Warm Springs distribution system	[5]6,500	[6]100 E	6,500
Total recharge for Goshen Valley			**7,600**

[1]Calculated as the 1990 flow in Spanish Fork at Castilla, Utah, minus losses and diversions, plus return flow from streamflow-gaging station to beginning of reach.

[2] Calculated as the estimated flow in Hobble Creek near Springville, Utah, minus losses and diversions, plus return flow from streamflow-gaging station to beginning of reach.

[3] Calculated as the estimated flow in Summit Creek near Santaquin, Utah. Loss includes 10 percent loss in open ditch plus 10 percent loss in reservoir.

[4] Calculated as the average annual flow in Currant Creek near Goshen, Utah, minus losses and diversions, plus return flow from streamflow-gaging station to beginning of reach.

[5] Cordova, 1969, p. 29.

[6] All water assumed to recharge the main ground-water system.

percentages or specific amounts determined in the 1960's were assumed to be applicable in 1990. In some stream and canal reaches near the mountains, springs that discharge water from bedrock interfered with the analysis of the seepage studies, and estimates of loss were made on the basis of other measurements in Utah (Kariya, Roark, and Hanson, 1994, p. 30).

During 1990, Spanish Fork and major canals diverted from Spanish Fork lost an estimated 28,100 acre-ft of water as recharge to the main ground-water system and 900 acre-ft as recharge to the Mapleton Bench system (tables 2 and 5). The concrete-lined extensions of the Strawberry Highline Canal on the east and west sides of West Mountain were assumed to lose no water.

On the basis of estimated 1990 flow, Hobble Creek and major canals diverted from Hobble Creek lost an estimated 3,000 acre-ft of water as recharge to the main ground-water system and 1,600 acre-ft as recharge to the Mapleton Bench system in 1990 (tables 2 and 5). Seepage from Maple Creek contributed a negligible amount of water to the ground-water system in 1990. On the basis of estimated 1990 flows, major canals diverted from Peteetneet Creek lost an estimated 1,600 acre-ft of water to the main ground-water system and Summit Creek lost an estimated 600 acre-ft of water to the main ground-water system in 1990.

Cordova (1970, p. 26) reported that streams and canals in Goshen Valley lose about 10 percent of their flow to the main ground-water system. On the basis of that estimate and average annual flow in Currant Creek, Currant Creek lost an estimated 1,100 acre-ft of water (table 5) and Currant Creek Canal lost an estimated 470 acre-ft of water as recharge to the main ground-water system in 1990 (table 2). Warm Springs, east of the town of Goshen, discharges water from the consolidated rock of Long Ridge. The water is distributed through several ditches and ponds. None of this water flows to Utah Lake, and the average annual flow of 6,500 acre-ft/yr (Cordova, 1969, table 6) is assumed to recharge the main ground-water system south and east of Goshen. Most of the water is then discharged from the ground-water system by evapotranspiration of swampy vegetation.

On the basis of annual flow in Spanish Fork and canals diverted from Spanish Fork, and estimated annual flow in Hobble Creek, canals diverted from Hobble Creek, Maple Creek, canals diverted from Peteetneet Creek, Summit Creek, Currant Creek, and Warm Springs distribution system, annual recharge to the main ground-water system in southern Utah and Goshen Valleys from perennial streams and major canals has ranged from 21,000 acre-ft to 147,000 acre-ft. Recharge from canals has been lower since 1987, when the Power Canal was lined with concrete. Annual recharge from perennial streams and major canals, irrigation and precipitation, and intermittent and ephemeral runoff from 1949 to 1990 is shown in figure 7.

Irrigation and Precipitation

The unofficial names of irrigated areas, amount of applied irrigation water and precipitation, and ground-water recharge from irrigation and precipitation are listed in table 6. Infiltration of applied irrigation water and precipitation was estimated to supply about 14,900 acre-ft/yr of recharge to the main ground-water system in southern Utah Valley, about 400 acre-ft/yr of recharge to the ground-water system in Goshen Valley, and about 7,100 acre-ft/yr of recharge to the Mapleton Bench system in 1990. Several factors contribute to the small amount of recharge from irrigation and precipitation in Goshen Valley. First, most of the large fields are sprinkler or drip irrigated, and recharge from unconsumed water is minimal (Dustin and Merritt, 1980, p. 43). Second, 37 percent of the applied irrigation water is from pumped wells, and the cost of pumping encourages water conservation. Third, precipitation is less and consumptive use by plants is greater in Goshen Valley than in southern Utah Valley. In contrast, most of the large fields in southern Utah Valley are flood irrigated, and only about 5 percent of the applied irrigation water in southern Utah Valley is from wells.

The recharge estimates from irrigation and precipitation are less than the previously reported estimates of 67,600 acre-ft in southern Utah Valley and 28,600 acre-ft in Goshen Valley in 1966 (Cordova, 1970, p. 25-28). The previous estimate assumed that about 30 percent of all applied water and precipitation recharged the ground-water system. The current estimate is equivalent to about 7 percent of applied water and precipitation in southern Utah Valley and 1 percent of applied water and precipitation in Goshen Valley (table 6). Because the current estimate was derived mostly from quantitative analysis of applied water and consumptive use, it is considered more accurate. On the basis of methods described in the following pages, annual recharge to the main ground-water system in southern Utah and Goshen Valleys from irrigation and precipitation has ranged from 10,200 acre-ft to 58,700 acre-ft (fig. 7) from 1949 to 1990.

Table 6. Irrigated areas, applied irrigation water, precipitation, and estimated ground-water recharge from irrigation and

[—, no data]

Irrigated and nonirrigated areas: Number of irrigated area refers to numbered area on figure 8.
Amount of surface water diverted: For some irrigated areas, only the applied amount is known.
Amount of surface water applied: Calculated as surface water diverted minus conveyance loss.
Amount of municipal water applied: Includes all municipal water use in areas where septic tanks are used and the difference between summer and winter
Amount of winter or annual precipitation: For irrigated areas, November 1989 to March 1990 precipitation at Payson (National Oceanic and areas, annual precipitation.
Soil type: From Mizue, 1968, p. 156; SL, sandy loam; STCL, silt clay loam; SC, sand and clay; CL, clay loam; STL, silt loam; SLC, sandy loam
Infiltration coefficient: Decimal fraction of applied water reaching the root zone, from Mizue, 1968, p. 51, except E, estimated.
Consumptive use: Determined by Blaney-Criddle Method, Soil Conservation Service, 1970.
Recharge: Calculated as [(total water applied + winter precipitation) * infiltration coefficient - consumptive use]. A negative result of the calculation

Irrigated and nonirrigated areas: Number of irrigated area	Irrigated and nonirrigated areas: Unofficial name of irrigated area	Estimated size of area (acres)	Amount of surface water diverted (acre-feet)	Conveyance loss from major canals (acre-feet)	Amount of surface water applied (acre-feet)	Amount of irrigation-well water applied (acre-feet)	Amount of spring water applied (acre-feet)
			Southern Utah Valley				
1	Springville City	1,460	[1]4,600	150	4,450	0	480
2	Springville City 2	600	[1]1,900	60	1,840	0	200
3	Springville Highline	220	[1]2,100	210	1,890	0	0
4	Springville No. 1 Ditch	520	[1]2,100	120	1,980	0	0
5	Sage	410	[1]700	0	700	0	170
6	Swenson	1,750	[1]2,700	270	2,430	800	0
7	Dry Creek	1,580	0	0	0	360	[1]4,700
8	Mapleton	4,290	[1,2]11,700	1,400	10,300	290	0
9	East Bench	3,620	[2]10,700	860	9,840	170	0
10	Westfield	6,180	[3]13,600	0	13,600	50	2,200
11	Mill Race	1,160	—	—	[1]2,400	0	0
12	Lake Shore	5,810	[2]3,600	140	3,460	0	0
13	Spanish Fork City	2,130	[2]300	0	300	0	0
14	South Field	4,160	[2]12,600	0	12,600	0	0
15	Salem	3,020	[4]4,300	270	4,030	240	600
16	Salem City	870	—	—	[5]1,000	0	0
17	Strawberry-One[6]	1,780	—	—	5,000	30	0
18	Strawberry-Two[6]	6,810	—	—	9,110	1,180	60
19	Strawberry-Three[6]	6,470	—	—	12,100	1,930	170
20	Payson City	1,960	[1]3,900	840	3,060	0	0
21	Summit[7]	3,570	[8]6,150	615	5,535	1,830	0
	Nonirrigated near mountain front	19,830	0	0	0	0	0
	Nonirrigated in low areas	17,680	0	0	0	0	0
Total (rounded)		**95,900**	**—**	**—**	**105,600**	**6,900**	**8,600**

municipal use in other areas.
Atmospheric Administration, 1990-91), distributed according to the 1931-60 normal annual precipitation (U.S. Weather Bureau, 1963). For non-irrigated

and clay.

indicates that consumptive use of the crop is not being met and the recharge is reported as 0.

Amount of municipal water applied (acre-feet)	Total amount of water applied (acre-feet)	Amount of winter or annual precipitation (acre-feet)	Soil type	Infiltration coefficient	Consumptive use (acre-feet)	Recharge: Recharge to Mapleton Bench ground-water system (acre-feet)	Recharge: Recharge to main ground-water system (acre-feet)
				Southern Utah Valley			
0	4,930	910	—	0.75 E	1,600	0	2,800
0	2,040	370	—	.75 E	660	1,100	0
0	1,890	150	SL	.85	350	0	1,400
0	1,980	350	SL	.85	1,100	880	0
0	870	210	STCL	.80	640	0	220
0	3,230	920	STCL	.80	4,400	0	0
0	5,060	830	SC	.75	4,100	0	320
1,000	11,590	2,700	CL	.85	8,200	3,900	0
0	10,010	1,900	CL	.85	8,900	1,200	0
0	15,850	3,200	SC	.75	15,000	0	0
0	2,400	630	STCL	.80	2,800	0	0
0	3,460	3,000	SC	.75	14,000	0	0
1,600	1,900	1,100	—	.75 E	2,300	0	0
0	12,600	2,200	SC	.75	9,900	0	1,200
0	4,870	1,600	STL	.80 E	7,600	0	0
400	1,400	480	—	.75 E	960	0	450
0	5,030	1,200	STL	.85	4,400	0	900
0	10,500	3,600	SC	.75	17,000	0	0
0	14,200	3,900	STL	.85	16,000	0	0
1,600	4,660	1,000	—	.75 E	2,200	0	2,000
1,100	8,350	2,300	STL	.85	7,400	0	1,650
0	0	30,000	—	—	—	0	[9]3,000
0	0	21,000	—	—	—	0	[10]1,000
5,700	**126,800**	**84,000**			**130,000**	**7,100**	**14,900**

Table 6. Irrigated area, applied irrigation water, precipitation, and estimated ground-water recharge from irrigation and

Irrigated and nonirrigated areas: Number of irrigated area	Irrigated and nonirrigated areas: Unofficial name of irrigated area	Estimated size of area (acres)	Amount of surface water diverted (acre-feet)	Conveyance loss from major canals (acre-feet)	Amount of surface water applied (acre-feet)	Amount of irrigation-well water applied (acre-feet)	Amount of spring water applied (acre-feet)
			Goshen Valley				
22	Strawberry-Four[6]	4,730	—	—	9,900	0	0
23	Bateman-Howlett	1,740	0	0	0	4,330	0
24	LDS North	2,040	0	0	0	370	0
25	LDS Central	2,440	0	0	0	180	0
26	LDS Orchards	850	0	0	0	1,750	0
27	Goshen[11]	2,940	—	—	[1]7,100	0	0
28	Goshen Mouth[11]	280	—	—	[1]1,400	0	0
29	Currant[11]	4,730	[1]4,700	470	4,230	6,290	0
30	Lunceford-Ekin	800	0	0	0	520	0
Total (rounded)		**20,600**	**—**	**—**	**22,600**	**13,400**	**0**

[1] Oral communication, irrigation or canal company or city official.
[2] Spanish Fork Water Commissioner, 1991, p. 104.
[3] Mill Race Diversion (Spanish Fork Water Commissioner, 1991, p. 104) minus 2,400 acre-feet delivered to Mill Race irrigation area.
[4] Salem Canal Diversion (Spanish Fork Water Commissioner, 1991, p. 104) minus 1,000 acre-feet delivered to Salem City.
[5] Estimated from Salem City area compared with unincorporated areas served by Salem Canal.
[6] The area serviced by the Strawberry Highline Canal is divided into four subareas with similar crops and soil type.
[7] Summit irrigation area includes the city of Santaquin.
[8] Estimated 1990 flow at Summit Creek near Santaquin, Utah.
[9] Ten percent of precipitation recharges main ground-water system.
[10] Five percent of precipitation recharges main ground-water system.
[11] Irrigation area receives surface water from Currant Creek.

Irrigated Areas

The location of irrigated crops, nonirrigated crops, and municipal areas was determined from digital land-use information obtained from the Utah Department of Natural Resources, Division of Water Resources (unpub. data, 1988) and is shown in figure 8. Areas with no data were assumed to contain nonirrigated natural vegetation. Irrigated land was subdivided into areas serviced by canal companies, municipalities, and landowners on the basis of field reconnaissance and information from irrigation companies (table 6 and fig. 8). Municipal areas are included as irrigated areas because lawns and gardens are irrigated. Infiltrating irrigation water from the orchards near Lincoln Point probably flows directly to Utah Lake and was not considered to recharge the main ground-water system. These orchard areas were not included in the recharge calculations. Recharge from irrigated areas was estimated by subtracting the amount of water used by crops (consumptive use) from the amount of water (including precipitation) that reaches the root zone of the crops in each irrigation area.

The amount of surface water applied to each irrigated area was determined from canal-diversion records and conversations with irrigation-company and city officials. In 1990, the only irrigated area that used water from Utah Lake was the orchard area near Lincoln Point. The amount of spring water applied to each irrigated area was based on measurements reported in Stolp and others (1993, tables 6 and 7) and from conversations with irrigation-company and city officials. The amount of ground water applied to each irrigated area was determined from unpublished U.S. Geological Survey records of irrigation wells and from estimates of municipal-well withdrawals for irrigation. Ground

precipitation, southern Utah and Goshen Valleys, Utah, 1990—Continued

Amount of municipal water applied (acre-feet)	Total amount of water applied (acre-feet)	Amount of winter or annual precipitation (acre-feet)	Soil type	Infiltration coefficient	Consumptive use (acre-feet)	Recharge: Recharge to Mapleton Bench ground-water system (acre-feet)	Recharge: Recharge to main ground-water system (acre-feet)
				Goshen Valley			
0	9,900	2,500	STCL	0.80	12,000	0	0
0	4,330	590	—	.85 E	4,400	0	0
0	370	690	SLC	.80	1,300	0	0
0	180	830	SLC	.80	1,600	0	0
0	1,750	290	SLC	.85 E	2,100	0	0
0	7,100	1,300	SC	.75	7,700	0	0
0	1,400	150	SLC	.80	790	0	450
0	10,520	2,000	SLC	.85 E	12,000	0	0
0	520	420	SLC	.85 E	2,000	0	0
0	**36,100**	**8,800**	**—**	**—**	**44,000**	**0**	**400**

water was assumed to be applied to the irrigated area in which the well is located unless well ownership or field reconnaissance indicated another application area. The amount of precipitation in each irrigated area was determined as a factor of the 1990 precipitation at Payson (National Oceanic and Atmospheric Administration, 1990-91, p. 4-5). The factor was based on the 1931-60 normal-precipitation distribution across southern Utah and Goshen Valleys (U.S. Weather Bureau, 1963). Precipitation from November 1, 1989, to March 31, 1990, is included as a direct application of water. Precipitation from April 1 to October 31, 1990, is included in the consumptive-use calculations.

The amount of water lost from major canals was determined as explained in the previous section on recharge from perennial streams and major canals. Conveyance loss was subtracted from the amount of water diverted to the canals to determine the amount of surface water applied to each area. Application loss refers to surface runoff and evaporation from soil and crop leaves. Mizue (1968, p. 51 and 156) reports soil types for each of the major irrigated areas in southern Utah and Goshen Valleys, and also reports surface waste as a percentage of diversion for various soil types. Using these data and information from field reconnaissance regarding sprinkler- and drip-irrigated areas, each irrigated area was assigned an infiltration coefficient as listed in table 6. A factor of 0.85 indicates that 85 percent of the applied water reaches the root zone.

Consumptive use for each irrigated area was determined by summing the consumptive use of each crop in the area multiplied by the acreage of each crop. Consumptive use of specific crops was determined using the Blaney-Criddle method (Soil Conservation Service, 1970). Total acreage of the specific crops in each irrigated area was determined by overlying a digital land-use and crop-type map (Utah Department of

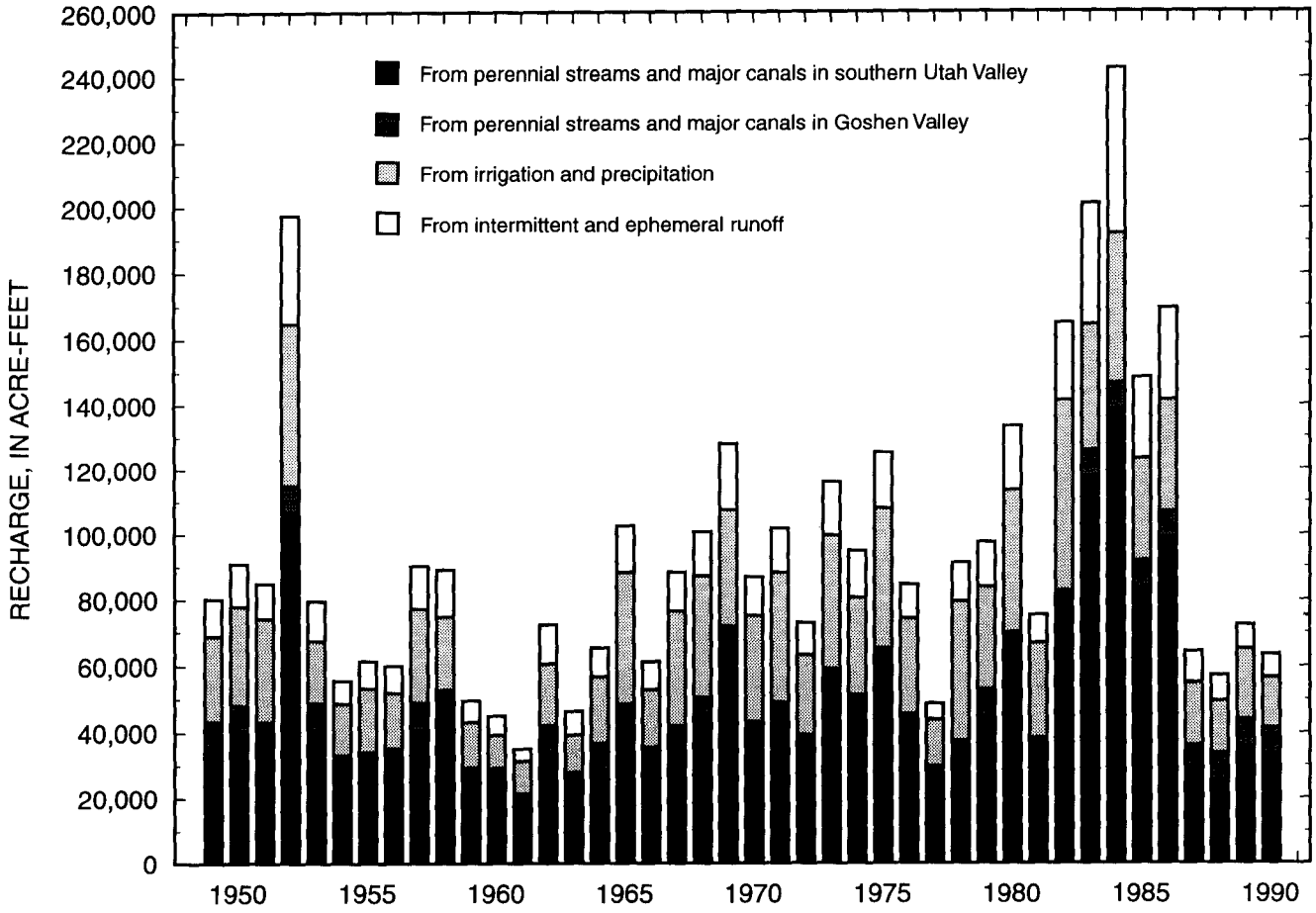


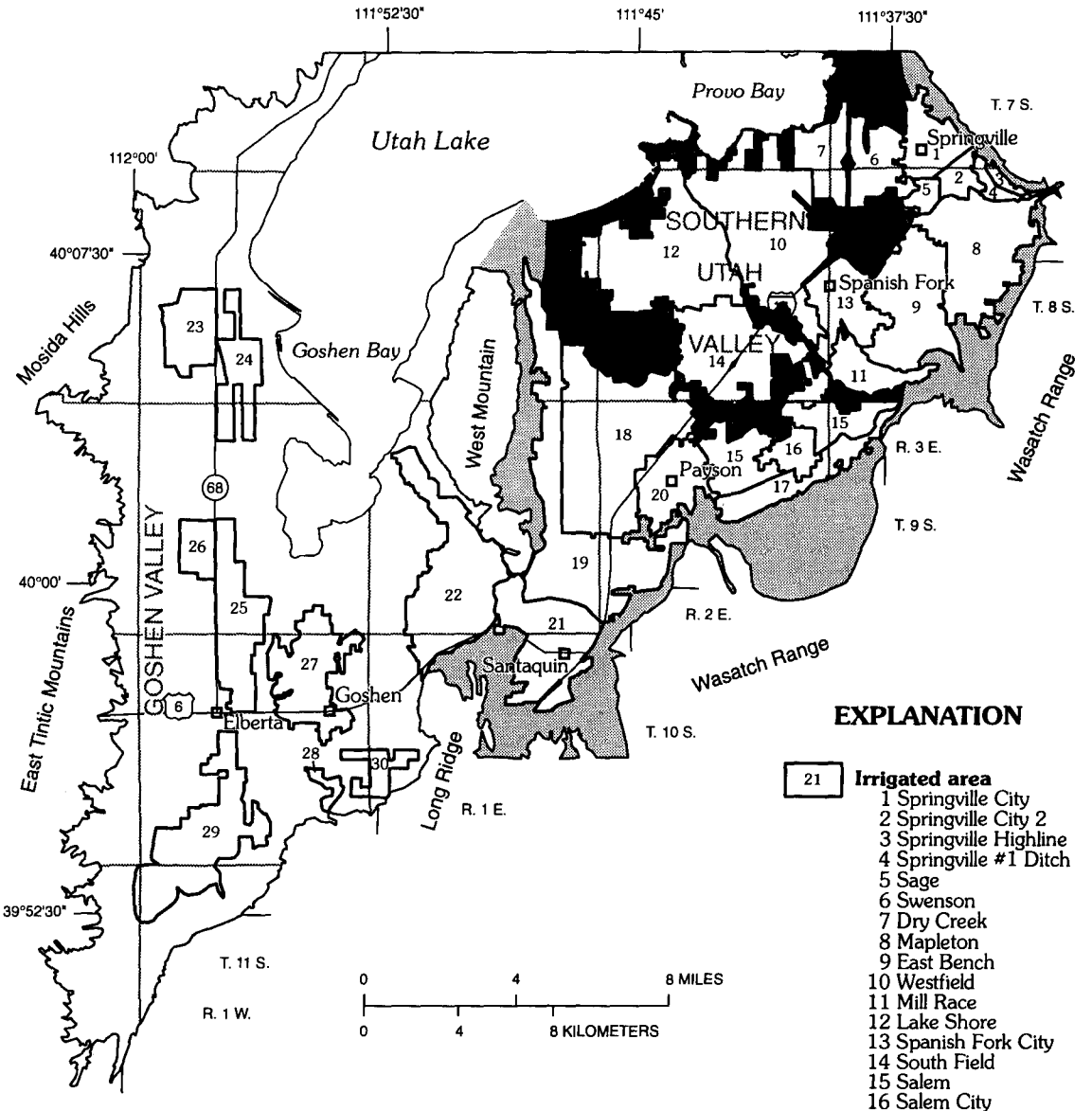
Figure 7. Annual recharge to the main ground-water system from perennial streams and major canals, irrigation and precipitation, and intermittent and ephemeral runoff, southern Utah and Goshen Valleys, Utah, 1949-90.

Natural Resources, Division of Water Resources, unpub. data, 1988) with the irrigated areas shown in figure 8. Field verification confirmed that, with a few exceptions, the 1988 land-use and crop-type map would be applicable to conditions in 1990. The 1990 crop type was used if it was known to be different than the 1988 crop type.

In municipal areas, the consumptive use was assumed to be one-half the consumptive use of grass pasture. This method accounts for one-half of the area to be lawns and gardens and assumes that precipitation over the entire area provides water to the lawns and gardens. This is a reasonable assumption because most of the study area does not have a storm-water collection system. Rainfall on impervious surfaces was assumed to run to permeable surfaces.

Nonirrigated Areas

Precipitation on nonirrigated land supplies a small percentage of the total recharge to southern Utah and Goshen Valleys. Recharge from precipitation on nonirrigated land was not measured during this study. Previous estimates of recharge from precipitation in similar basins in Utah range from 1 to 30 percent. Cordova (1970, p. 28) assumed that 30 percent of the winter precipitation near the mountains and 10 percent of the winter precipitation away from the mountains recharge the ground-water system in southern Utah and Goshen Valleys. Clark and Appel (1985, p. 31) estimated that 20 percent of precipitation near the mountains and a negligible amount of precipitation away from the mountains recharge the ground-water system in northern Utah Valley. Richardson (1906, p. 27) estimated that rainfall on the valley contributes only a small amount of recharge to the ground-water system in Utah Valley. Feth and others (1966, p. 43) estimated that 25 percent of precipitation recharges the ground-water system in an area with average annual precipitation of 14 in. Razem and Steiger (1981, table 2, p. 13) reported that 1 percent of precipitation recharges the ground-water system in an area with average annual



Base from U.S. Geological Survey digital data, 1:100,000, 1979, 1980, 1981
Universal Transverse Mercator projection, Zone 12

EXPLANATION

21 **Irrigated area**

1 Springville City
2 Springville City 2
3 Springville Highline
4 Springville #1 Ditch
5 Sage
6 Swenson
7 Dry Creek
8 Mapleton
9 East Bench
10 Westfield
11 Mill Race
12 Lake Shore
13 Spanish Fork City
14 South Field
15 Salem
16 Salem City
17 Strawberry-One
18 Strawberry-Two
19 Strawberry-Three
20 Payson City
21 Summit
22 Strawberry-Four
23 Bateman-Howlett
24 LDS North
25 LDS Central
26 LDS Orchards
27 Goshen
28 Goshen Mouth
29 Currant
30 Lunceford-Ekin

Areas of ground-water recharge from precipitation on nonirrigated land

Area where recharge is assumed to be 10 percent of precipitation

Area where recharge is assumed to be 5 percent of precipitation

Boundary of study area

Figure 8. Location of irrigated areas and areas of ground-water recharge from precipitation on nonirrigated land, southern Utah and Goshen Valleys, Utah.

precipitation of 11 to 12 in., and 3 percent of precipitation recharges the ground-water system in an area with average annual precipitation of 12 to 16 in. Roark and others (1991, p. 20) estimated that 10 percent of precipitation recharges the ground-water system in an area with average annual precipitation of 16 to 18 in. Ten percent of the precipitation on more permeable nonirrigated land near the mountains surrounding southern Utah Valley and 5 percent of the precipitation on less permeable nonirrigated land away from the mountains was assumed to recharge the main ground-water system (fig. 8 and table 6).

Precipitation on nonirrigated land in semiarid Goshen Valley is assumed to contribute an insignificant amount of recharge to the main ground-water system. Recharge from precipitation was reported as negligible in nearby areas by Bjorklund (1967, p. 19) for northern Juab Valley and by Feltis (1967, p. 11) for Cedar Valley.

Intermittent and Ephemeral Runoff

On the basis of estimated 1990 runoff, infiltration of intermittent and ephemeral runoff is estimated to contribute about 6,400 acre-ft/yr of recharge to the main ground-water system in southern Utah Valley and about 400 acre-ft/yr of recharge to the ground-water system in Goshen Valley in 1990 (table 3). Most intermittent and ephemeral runoff enters the valley during spring, when evapotranspiration rates are low. The runoff crosses highly permeable alluvial fans, and no channels are evident downstream from the alluvial fans. These factors indicate that 90 percent of the runoff infiltrates and recharges the main ground-water system. Kariya, Roark, and Hanson (1994, p. 30) and Bjorklund (1967, p. 22) report that most of the intermittent and ephemeral runoff recharges the ground-water system in Cache Valley in northern Utah and in Juab Valley in central Utah. Thiros and Brothers (1993, p. 14) estimate that 80 percent of intermittent and ephemeral runoff recharges the ground-water system in Panguitch Valley in southern Utah. Although most surface water on Mapleton Bench is assumed to contribute recharge to the Mapleton Bench system, not to the main ground-water system, infiltration of intermittent and ephemeral runoff that originates east of the bench is assumed to recharge the main ground-water system because the runoff enters the valley near the mountain front where the clay layer is not continuous. Assuming that annual intermittent and ephemeral runoff varies similarly to natural flow in Spanish Fork at Castilla, annual recharge to the main ground-water system in southern Utah and Goshen Valleys from intermittent and ephemeral runoff from 1949 to 1990 ranged from 3,600 acre-ft to 50,400 acre-ft (fig. 7).

Subsurface Inflow

Recharge by subsurface inflow in 1990 was estimated to be about 65,000 acre-ft along 25 mi of mountain front in southern Utah Valley and about 13,000 acre-ft along 35 mi of mountain front in Goshen Valley (table 4). These estimates include flow from consolidated rock to the basin-fill deposits and flow in the unconsolidated stream-channel deposits. Flow in stream-channel deposits probably is small because the deposits are narrow and thin. This amount of subsurface inflow was not determined by water levels or aquifer characteristics, but was calculated as the amount of inflow needed to produce a balanced ground-water budget. This method of estimating subsurface inflow incorporates potential inaccuracies in all other budget elements into this single element; therefore, this estimate of subsurface inflow might have substantial error.

Most subsurface inflow occurs as movement of water from consolidated rock to basin-fill deposits through rock, fractures, bedding planes, and solution channels (fig. 5). The Wasatch Range contains limestone that is deformed and fractured. Caverns in limestone are indications of the conduit system (Clark and Appel, 1985, p. 31). Water-level altitude in the limestone of the Wasatch Range is not known. The East Tintic Mountains also are extensively faulted and contain conduits. Lindgren and Loughlin (1919, p. 19) report that in one mine, water disappeared through fractures at the bottom of a 2,000-foot shaft. Water-level altitude in mines in the East Tintic Mountains ranges from 4,700 ft to about 6,500 ft (Lindgren and Loughlin, 1919, p. 19). Subsurface inflow to southern Utah and Goshen Valleys has not been quantified previously, but because much of the lithology of the surrounding mountains is fractured limestone, Dustin and Merritt (1980, p. 29) suggest the volume could be comparatively large. Clark and Appel (1985, p. 31) estimate a minimum subsurface inflow to northern Utah Valley from the Wasatch Range of 100,000 acre-ft/yr along about 45 mi of mountain front. No data exist to determine annual variation in recharge from subsurface inflow.

Discharge

Discharge from the ground-water system was estimated to be about 130,000 acre-ft in southern Utah

Valley and about 33,000 acre-ft in Goshen Valley in 1990 (table 4). Discharge is to springs and drains, by evapotranspiration, to wells, to streams and canals, by seepage to Utah Lake, and by infiltration to sewer systems. With the exception of discharge to wells, no data exist to determine annual variation in ground-water discharge.

Springs and Drains

The main ground-water system discharged an estimated 42,700 acre-ft of water to springs and drains in southern Utah Valley and negligible water to springs and drains in Goshen Valley in 1990. The Mapleton Bench system discharged an estimated 3,500 acre-ft to springs in 1990 (table 7). For this report, discrete areas of ground-water discharge were defined as springs. Examples are the beginning of Spring Creek near Springville, Salem Lake, and Spring Lake. Usually, the discrete area is the beginning of a stream. Ground-water discharge to springs was measured in the stream channels as close to the springs as possible. Ground-water discharge to installed drains is known to occur west of Springville and near Benjamin Slough. The

Table 7. Estimated ground-water discharge to springs and drains in southern Utah Valley, Utah, 1990

[See figure 2 for explanation of numbering system for hydrologic-data sites; —, no known flow; E, estimated]

Discharge to springs: Discharge from a point source.
Discharge to drains: Gain in the flow in a channel through an area of known installed drains.
Discharge to natural drains: Gain in the flow in a channel through an area of no known springs or installed drains.

Unofficial name of spring or drain area	Location of measurements	Main ground-water system			Mapleton Bench ground-water system, discharge to springs (acre-feet)
		Discharge to springs (acre-feet)	Discharge to drains (acre-feet)	Discharge to natural drains (acre-feet)	
Spring Creek area	(D-7-3)19,20,28,29,30	11,200	—	1,400	—
Wood Springs area	(D-7-3)32	1,800	1,800	—	—
Condie Springs	(D-7-3)34cab	[1]100	—	—	—
Matson Springs area	(D-7-3)31,32	1,100	400	—	—
Benjamin Slough	(D-8-1)23,24,25	—	800	—	—
Tanner Reservoir area	Total of (D-8-1)36; (D-9-1)1,12; and (D-9-2)6,7	300	—	1,400	—
Dry Creek area	Total of (D-8-2)1 and (D-8-3)4,5,6,8,17	4,800	300	800	—
Beer Creek	Total of (D-8-2)31,32,33 and (D-9-2)2,3	1,400	—	6,500	—
Bradford Springs	(D-8-2)36a	300 E	—	—	—
Wheeler Springs	(D-8-3)3acb	—	—	—	500
Fullmer Springs	(D-8-3)3cad	—	—	—	[1]400
Clyde Springs	(D-8-3)3dbb	—	—	—	—
Holley Springs[2]	(D-8-3)9bac	—	—	—	2,600
Holladay Springs area	(D-9-1)24,25	1,000	—	400	—
Salem Lake	(D-9-2)11abd	4,200	—	—	—
Spring Lake area	(D-9-2)19,29,30	1,700	—	1,000	—
Total		**27,900**	**3,300**	**11,500**	**3,500**

[1] Reported by Cordova (1969, table 6).
[2] Previously reported as Big Hollow Springs (Cordova, 1969, table 6, and Cordova, 1970, table 7).

drains flow to stream channels, and discharge to installed drains was calculated as the increase in measured streamflow in these areas. Many of the streams that begin as springs are natural drains and collect additional ground-water discharge as they flow to Utah Lake. Ground-water discharge to natural drains was calculated as the measured increase in streamflow when no other cause of the increase could be determined.

The flow of most springs and streams around Springville was measured in November 1990, and the flow of most springs and streams around Beer Creek, Spring Creek (near Payson), and Benjamin Slough was measured in March 1991 (Stolp and others, 1993, tables 6 and 7). The flow of several springs was measured three or four times during the study period, and the average flow was used in the estimate of ground-water discharge. At these springs, discharge measurements in November through April were about 90 percent of discharge measurements in May through October. Infiltration of irrigation water may increase ground-water discharge to springs and drains during the summer and early fall months; therefore, the measurements made in November and March probably represent minimum discharge.

The previous estimate (Cordova, 1970, p. 31) of ground-water discharge to springs and drains was 79,000 acre-ft in southern Utah Valley and 7,800 acre-ft in Goshen Valley during 1966, including the flow of several springs that this study considers to discharge from consolidated rock. Individual spring-flow measurements in 1990 and 1991 did not differ substantially from individual spring-flow measurements in 1964-67 (Stolp and others, 1993, tables 6 and 7 and Cordova, 1969, table 6).

Evapotranspiration

The main ground-water system discharged about 26,000 acre-ft in southern Utah Valley and about 14,000 acre-ft in Goshen Valley by evapotranspiration during 1990 (table 8). Location and type of phreatophytes are shown in figure 9. Discharge by evapotranspiration was estimated only in areas of plant growth, not in playas or areas of open water. Nonvegetated playas are very small in the two valleys, thus the quantity of evaporation occurring from them was considered to be a negligible part of the ground-water budget. Data on evapotranspiration by phreatophytes were not collected during this study, so evapotranspiration cannot be calculated specifically for 1990. The method and calculations presented in the following sections are assumed to provide average annual evapotranspiration, which is an adequate estimate for evapotranspiration during 1990.

Phreatophytes are plants that depend on ground water for their water supply (Robinson, 1958, p. 1). For this report, irrigated alfalfa and pasture that use some ground water are considered to be phreatophytes. The location of phreatophytes was determined from two sources. The first source was the digital land-use and crop-type map prepared by the Utah Department of Natural Resources, Division of Water Resources (unpub. data, 1988). Areas shown on the map as irrigated alfalfa or pasture were assumed to use ground water to supplement applied irrigation water if the ground-water level was estimated to be within 10 ft of the land surface. Field reconnaissance during this study indicated that salt grass grows on about one-half of the industrial area near Springville. The second source was infrared aerial photographs that highlight areas of wetlands and vegetation (Agricultural Stabilization and Conservation Service, 1987-88). These photographs and field reconnaissance were used to determine areas and types of phreatophytes around Utah Lake and in uncultivated parts of Goshen Valley. Tamarisk (*Tamarix gallica*) grows along the shores of Utah Lake, and salt grass (*Distichlis stricta*) grows in noncultivated areas away from Utah Lake.

The maximum evapotranspiration rate of salt grass and tamarisk is assumed to be 4.0 and 6.0 ft/yr, respectively (Robinson, 1958, p. 18 and 75). Because depth to ground water is not always ideal for growing conditions and the plants are not always at maximum density, the evapotranspiration rate of the plants was calculated as a percentage of the maximum evapotranspiration rate. In nonirrigated areas, the actual evapotranspiration rate was assumed to be one-half of the maximum rate. In irrigated areas where the amount of applied water did not meet crop consumptive-use demands, the actual evapotranspiration rate was assumed to be the difference between the crop potential consumptive use and the amount of applied irrigation water, but was not allowed to exceed one-half of the maximum evapotranspiration rate (table 8).

The estimate for ground-water discharge by evapotranspiration is about one-half of the previous reported estimate (Cordova, 1970, p. 33). Land-use changes, different estimates of maximum evapotranspiration rate, and different estimates of ground-water level in the evapotranspiration area contribute to the difference in evapotranspiration estimates.

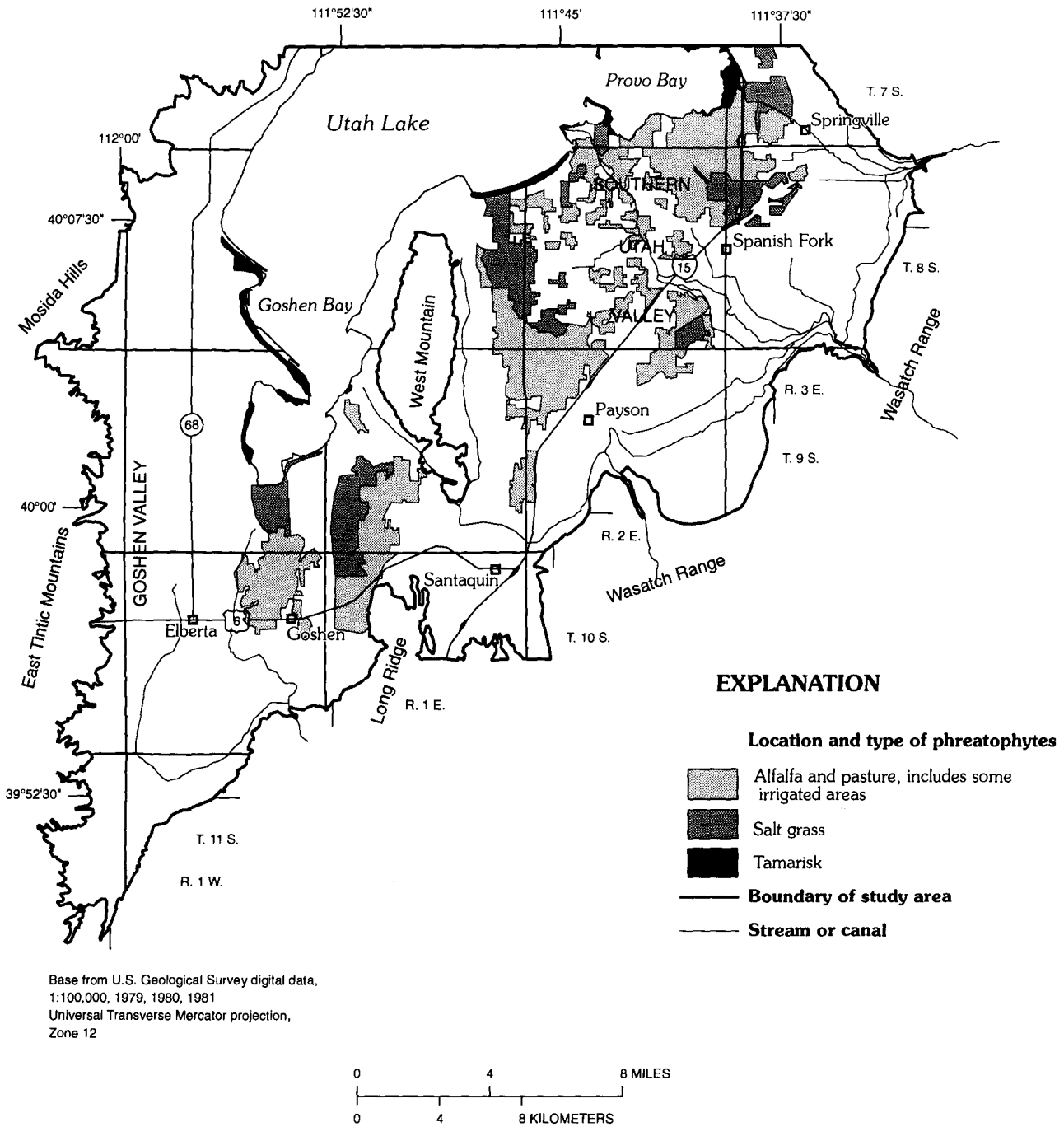


Figure 9. Location and type of phreatophytes and areas of ground-water discharge by evapotranspiration, southern Utah and Goshen Valleys, Utah.

Table 8. Phreatophytes, areas of phreatophyte growth, evapotranspiration rate, and estimated discharge by evapotranspiration, southern Utah and Goshen Valleys, Utah, 1990

Phreatophyte	Area of phreatophyte growth (acres)	Maximum evapotranspiration rate (feet per year)	Estimated evapotranspiration rate (feet per year)	Estimated discharge by evapotranspiration (acre-feet)
		Southern Utah Valley		
Irrigated alfalfa	17,000	2.6	0.6	10,000
Salt grass	6,900	4.0	2.0	14,000
Tamarisk	500	6.0	3.0	1,500
Total evapotranspiration (rounded)				**26,000**
		Goshen Valley		
Irrigated alfalfa	4,300	2.8	.7	3,000
Irrigated pasture	100	2.4	.3	30
Salt grass	4,700	4.0	2.0	9,400
Tamarisk	500	6.0	3.0	1,500
Total evapotranspiration (rounded)				**14,000**

Wells

The main ground-water system discharged about 18,400 acre-ft of ground-water to wells in southern Utah Valley and about 13,500 acre-ft to wells in Goshen Valley during 1990 (table 9). These discharges include water from pumped and flowing wells for irrigation, municipal supply, industrial supply, and domestic use. Domestic use includes household use, lawn and garden use, and stock use in areas not served by a municipal system. Annual irrigation and municipal ground-water withdrawals since 1963 are shown in figure 10. Prior to 1960, less than 2,000 acre-ft/yr of ground water were withdrawn for irrigation.

Annual withdrawal from pumped irrigation wells was estimated from flow measurements and corresponding power-consumption records. Annual withdrawal from most municipal wells and some industrial wells is reported to the Utah Department of Natural Resources, Division of Water Rights. Annual withdrawal for unreported municipal and industrial wells was estimated based on previous records, population served, and size and type of industry.

Annual withdrawal from domestic wells in southern Utah Valley was estimated to be about 1,600 acre-ft (table 9). The 1990 census indicates that about 5 percent of the Utah County population live in unincorporated areas (Frank Hachman, Associate Director, University of Utah Bureau of Economic and Business Research, oral commun., 1992). Because southern Utah Valley is more rural than northern Utah Valley, it is assumed that 10 percent of the population in southern Utah Valley live in unincorporated areas. Ten percent of the municipal withdrawal for 1990 is about 800 acre-ft. Because some of the domestic wells are uncontrolled flowing wells or are used for stock, it is assumed that about twice that amount, 1,600 acre-ft, of ground water is withdrawn annually from domestic wells in southern Utah Valley. Ground-water withdrawal from domestic wells in Goshen Valley is negligible.

The location of selected flowing wells inventoried during 1989-91 is shown in figure 11. In southern Utah Valley, ground-water discharge to flowing wells was about 4,400 acre-ft in 1990 (table 9). The discharge from flowing wells with historic discharge of 50 acre-ft/yr (30 gal/min) or more was measured or estimated as part of this study. In locations where several small flowing wells are within a few hundred feet of each other, flow was measured or estimated from all of the wells. The 1990 measured and estimated flowing-well discharge for irrigation and municipal use was about 3,400 acre-ft. Small flowing wells (less than 30 gal/min) in southern Utah Valley were not measured but are esti-

Table 9. Estimated ground-water discharge to wells, southern Utah and Goshen Valleys, Utah, 1990

Water use	Southern Utah Valley		Goshen Valley
	Pumped	Flowing	
Irrigation	4,600	3,000	13,400
Municipal supply	7,700	400	100
Industrial supply	1,100	0	0
Domestic	600	1,000	0
Total	**14,000**	**4,400**	**13,500**

mated to discharge about 1,000 acre-ft/yr, primarily for domestic use. Flowing wells occur only in a small area in Goshen Valley, and ground-water discharge to these wells was negligible.

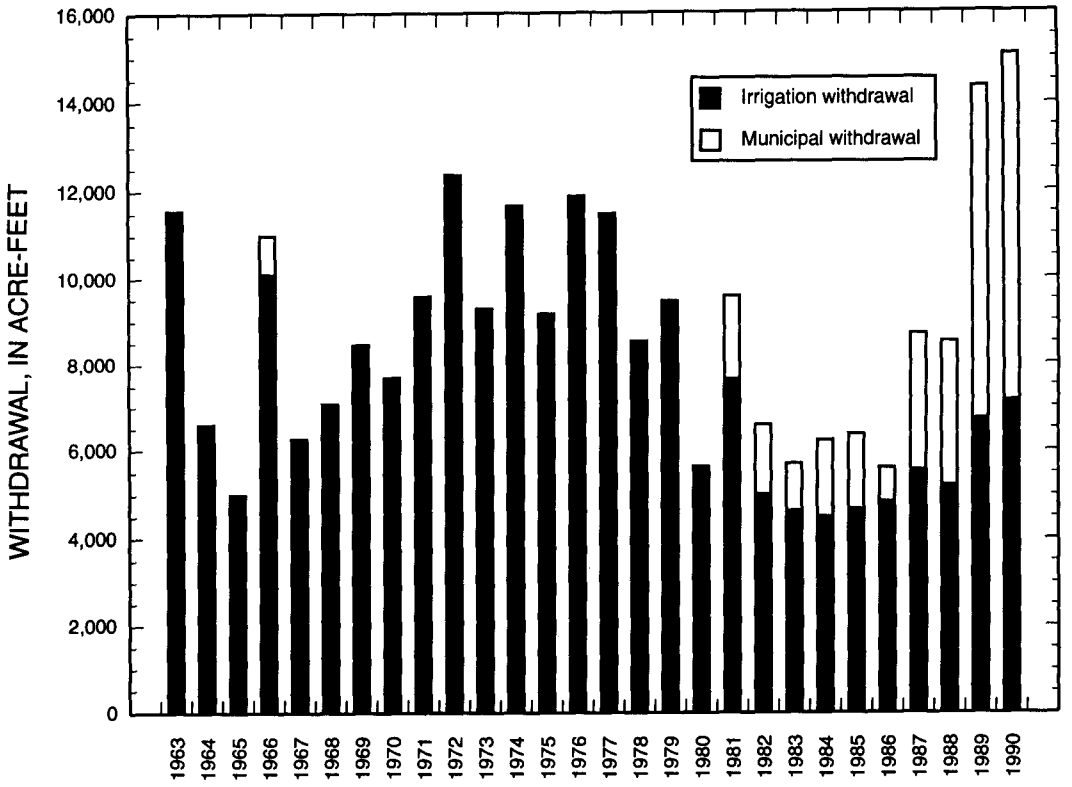
The total flowing-well discharge of 4,400 acre-ft in southern Utah Valley is less than the previous estimate of 13,000 acre-ft (Cordova, 1970, p. 33), even though water levels are similar. Flow measurements for 20 wells in 1964-67 were compared with flow measurements for the same wells in 1989-90 (Stolp and others, 1993, table 4) using a Mann-Whitney statistical analysis (Iman and Conover, 1983, p. 280-287). The analysis showed, with 95.1-percent confidence, that the median flow measured in 1964-67 was 3.4 to 12.0 times higher than the median flow measured in 1989-90. A boxplot of the data is shown in figure 12. A Mann-Whitney analysis of the water levels (Stolp and others, 1993, table 3) measured for the same wells showed no significant difference between the median water level measured in 1964-67 and the median water level measured in 1989-90 (fig. 11). Reasons for the decline in flow include "sanding-in" of older wells, corrosion of well casings, and fewer uncontrolled flowing wells. Another reason for the difference between this study and the previous estimate is that the previous estimate assumed that wells of similar diameter had similar flows (Cordova, 1970, p. 31). This study found that most domestic wells and many stock and irrigation wells are controlled and that flow is not related to well diameter. This study also found that most new flowing wells are for domestic use and are generally controlled.

Perennial Streams and Major Canals

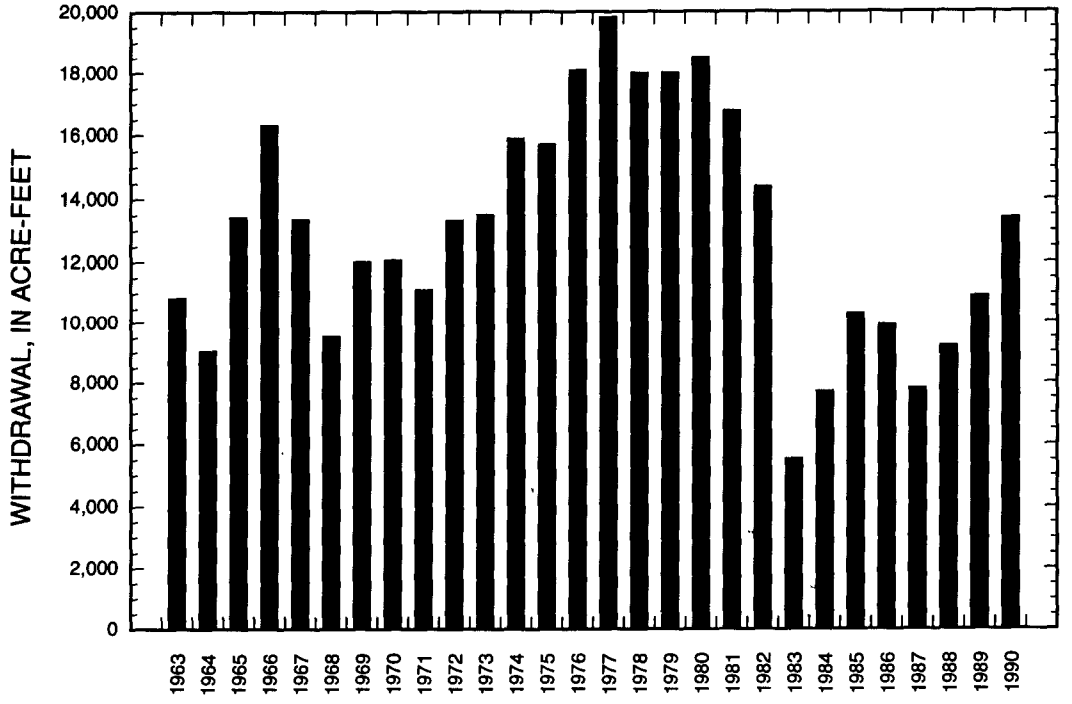
The main ground-water system discharged an estimated 20,700 acre-ft of water to perennial streams and major canals in southern Utah Valley and an estimated 2,200 acre-ft to streams in Goshen Valley during 1990 (table 10). The Mapleton Bench system discharged an estimated 10,900 acre-ft to Hobble Creek and to Mill Race Canal during 1990. The discharge in southern Utah Valley was calculated using the 1990 Spanish Fork flow and the average annual gain in Hobble Creek.

Utah Lake

Ground-water discharge from unconsolidated basin-fill deposits to Utah Lake and to springs near Utah Lake during 1990 was estimated to be 9,600 acre-ft in southern Utah Valley along a shoreline length of 13 mi and 3,600 acre-ft in Goshen Valley along a shoreline length of 16 mi. The shoreline length in Goshen Valley does not include the shoreline near West Mountain nor the shoreline north of latitude 40°07'30"N. No data were available in those areas, but the unconsolidated deposits are thin and discharge was assumed to be small relative to other areas. The flow was calculated across segments as close to Utah Lake as available data would permit. The location of the computation lines is shown on plate 1, and the computations are listed in table 11. Between the computation lines in southern Utah Valley and Utah Lake, 600 acre-ft of ground-water recharge and 23,400 acre-ft of ground-water discharge occurred. Between the computation lines in Goshen Valley and Utah Lake, 60 acre-ft of ground-water recharge and 7,300 acre-ft of ground-water discharge occurred. Recharge was added to, and discharge was subtracted from, the flow across the computation lines to obtain the ground-water discharge to Utah Lake. Some of the calculated discharge to Utah Lake from southern Utah Valley could be discharge to marshy areas around Benjamin Slough and the mouth of Hobble Creek.



a) Irrigation and municipal withdrawals in southern Utah Valley. Municipal withdrawals were not reported prior to 1981. The 1966 municipal withdrawal is from Cordova (1970, p. 33).



b) Irrigation withdrawals in Goshen Valley. Municipal withdrawal is about 100 acre-feet per year.

Figure 10. Annual irrigation and municipal ground-water withdrawals from wells in a) southern Utah Valley and b) Goshen Valley, Utah, 1963-90.

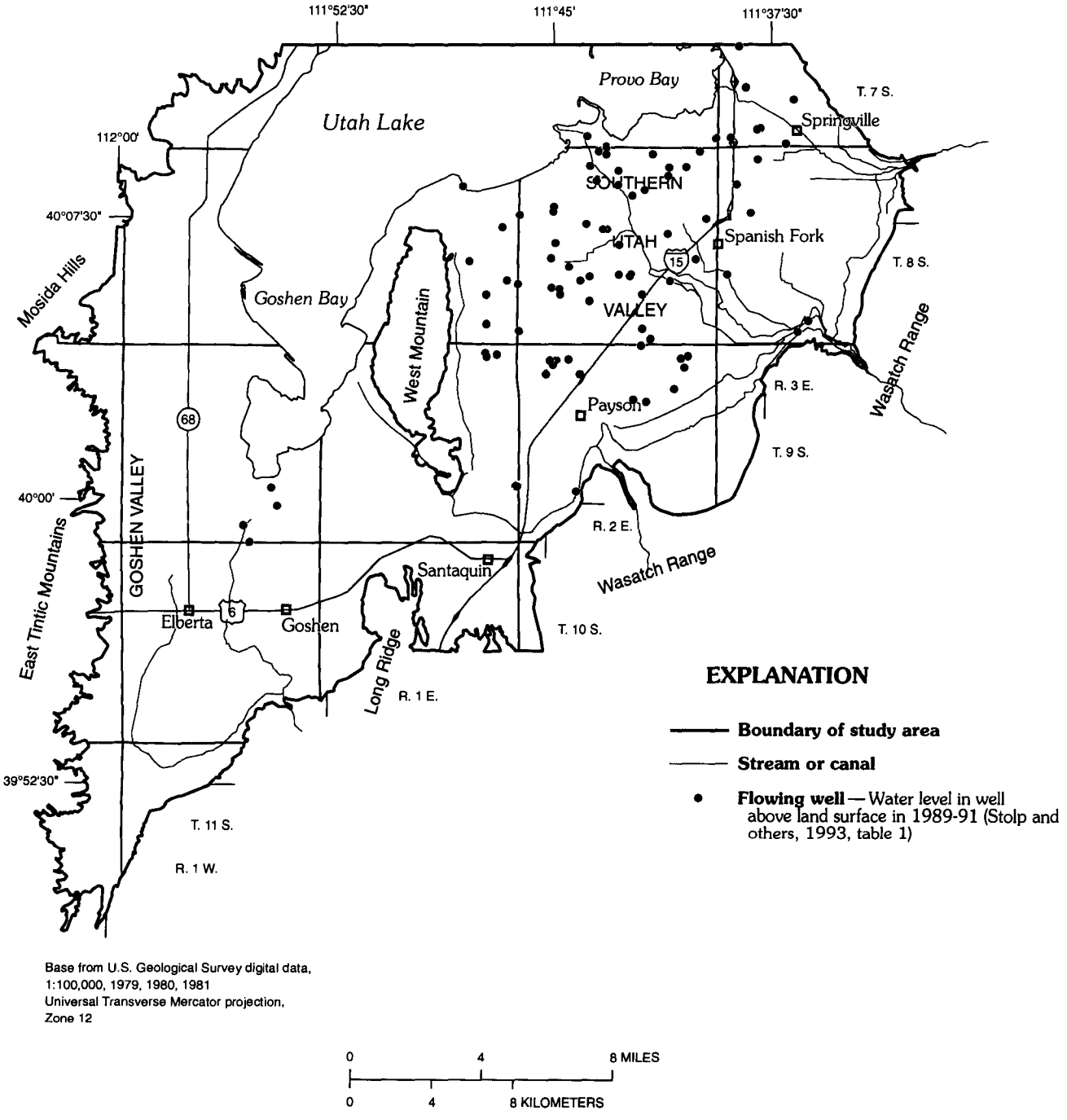


Figure 11. Location of selected flowing wells in southern Utah and Goshen Valleys, Utah, 1989-91.

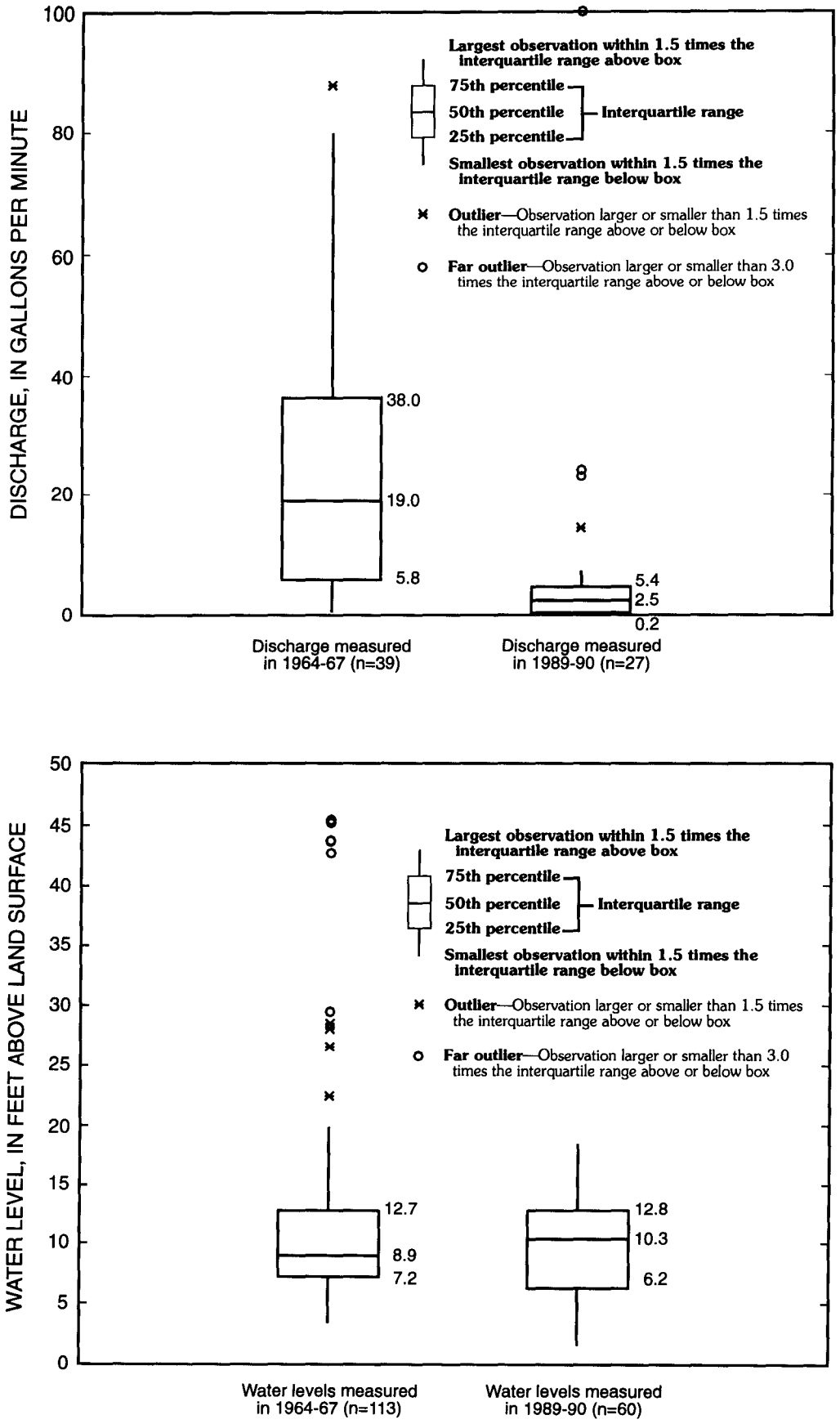


Figure 12. Discharge and water-level measurements in selected flowing wells, southern Utah and Goshen Valleys, Utah.

Table 10. Estimated ground-water discharge to perennial streams and major canals, southern Utah and Goshen Valleys, Utah, 1990

[See figure 2 for explanation of numbering system for hydrologic-data sites]

Discharge: Represents measured gain in perennial streams and major canals reported by Stolp and others (1993, table 8).

Stream or canal	Discharge (acre-feet per year)	
	Main ground-water system	Mapleton Bench ground-water system
Southern Utah Valley		
Hobble Creek between (D-8-3)3dda and Swenson Ditch	0	8,700
Hobble Creek downstream from Swenson Ditch	4,500	0
Spanish Fork between Highway 91 and Interstate-15	10,000	0
Spanish Fork downstream from Palmyra	6,200	0
Mill Race Canal	0	2,200
Total discharge	**20,700**	**10,900**
Goshen Valley		
Currant Creek below Goshen Reservoir	[1]2,200	0

[1]Joel White, Goshen Irrigation Company, oral commun., 1991.

The volume of water flowing past the computation lines was estimated using Darcy's Law as expressed in the following equation:

$$Q = TIL \qquad (1)$$

where Q = flow, in ft^3/d,
T = transmissivity, in ft^2/d,
I = hydraulic gradient (dimensionless), and
L = length of cross section, in ft.

Transmissivity for most segments was determined from the lithology reported on drillers' logs. Hydraulic conductivity (K) of each lithologic zone was determined from ranges reported by Lohman (1979, table 17). Transmissivity of each zone was determined by multiplying K by the thickness of the zone. Transmissivity of the saturated basin-fill deposits encountered during drilling was determined by summing the transmissivity of each zone. Average hydraulic conductivity in the saturated basin-fill deposits was determined by dividing transmissivity by the thickness of the saturated basin-fill deposits encountered during drilling. This average K was assumed for the materials below the depth of the well. Transmissivity of one segment was determined from one aquifer test. The hydraulic gradient for each segment was determined using the potentiometric contours shown on plate 1, and the length of each segment was measured on a map. Because deep deposits are less permeable, a thickness of 1,000 ft was used in the calculations.

On the basis of an analysis done using Darcy's Law, Cordova (1970, p. 36-37) reported a minimum discharge to Utah Lake from southern Utah Valley of 4,000 acre-ft/yr along a shoreline length of 7 mi and estimated discharge to Utah Lake from Goshen Valley to be 4,000 acre-ft/yr. On the basis of a water-budget analysis of Utah Lake, Richardson (1906, p. 24) estimated ground-water discharge to Utah Lake from all of Utah Valley and Goshen Valley to be about 133,000 acre-ft/yr. On the basis of computer simulations, Dustin and Merritt (1980, p. 9) estimated ground-water discharge to Utah Lake from all of Utah Valley and Goshen Valley to be 114,000 acre-ft/yr and estimated that about 18,000 acre-ft/yr was entering Goshen Bay. On the basis of an analysis done using Darcy's Law, Clark and Appel (1985, p. 79) estimated ground-water discharge to Utah Lake from northern Utah Valley to be about 37,000 acre-ft/yr along a shoreline length of 20 mi. The estimates of Richardson (1906) and Dustin and Merritt (1980) include springs discharging from consolidated rocks into Utah Lake.

Table 11. Estimated ground-water flow across computation lines near Utah Lake, southern Utah and Goshen Valleys, Utah, 1990

[See plate 1 for location and length of computation lines and potentiometric contours; see figure 2 for an explanation of numbering system for hydrologic-data sites]

Computation line	Average transmissivity (T) (feet squared per day)	Length (L) of computation line (feet)	Average hydraulic gradient (I) (dimensionless)	Flow across computation line (acre-feet per year)
		Southern Utah Valley		
1	23,000	18,200	0.0045	15,800
1a	[1]23,000	12,900	.0038	9,400
2	12,000	22,200	.0020	4,500
2a	[1]12,000	19,400	.0014	2,700
Total flow				**32,400**
		Goshen Valley		
3	11,000	8,400	.0015	1,200
3a	[1]11,000	31,500	[1].0015	4,400
4	[2]5,100	12,300	.0027	1,400
4a	[1]5,100	33,200	[1].0027	3,800
Total flow				**10,800**

[1] Value estimated to be the same as previous line.
[2] Transmissivity determined from aquifer test at well (C-10-1)4cbb-1.

Sewer Systems

Ground-water infiltration to municipal sewer systems was estimated to be 5,000 acre-ft in southern Utah Valley during 1990 (table 12). Goshen Valley has no municipal sewer systems. Sewer systems are seldom watertight, so infiltration of ground water occurs where the ground-water level is higher than sewer pipes. Parts of the systems in Spanish Fork, Springville, Payson, and Salem receive infiltration. The amount of infiltration in Spanish Fork, Springville, and Payson was determined from the records of the wastewater-treatment plants. Minimum inflow to the plants usually occurs when municipal water use is minimal (around 3:00 a.m.), and the inflow is assumed to be ground-water infiltration. Lawn and garden watering causes summer infiltration to exceed winter infiltration; an average infiltration value for the year was used. James M. Montgomery Engineers (1985, p. 4-5) report infiltration of 0.4 Mgal/d to the Salem sewer system.

Release from Storage

An estimated 9,800 acre-ft of water in southern Utah Valley and 3,400 acre-ft of water in Goshen Valley were released from storage in the main ground-water system from March 1990 to March 1991. As water level declined, water was released from the deposits. The specific yield of deposits is the volume of water yielded from the deposits by gravity drainage per unit volume of deposits under unconfined conditions (Lohman, 1979, p. 6). Cordova (1970, table 12) estimated a specific yield of 0.06 for the deposits in southern Utah and Goshen Valleys. Water level in selected wells finished in unconfined zones of the main ground-water system in southern Utah Valley declined an average of 1.7 ft from March 1990 to March 1991. Multiplying this value by a specific yield of 0.06 equals a release of water of 0.10 ft, which is assumed to occur throughout the entire 98,000 acres of southern Utah Valley, for a release from storage of 9,800 acre-ft. Water-level measurements are poorly distributed in

Table 12. Estimated ground-water infiltration to sewer systems in southern Utah Valley, Utah, 1990

Discharge: Flow through municipal wastewater-treatment plants at time of day when water use is at a minimum, typically around 3:00 a.m.

Municipality	Discharge	
	Million gallons per day	Acre-feet per year
Spanish Fork	[1]1.8	2,000
Springville	[1]1.3	1,500
Payson	[1].9	1,000
Salem	[2].4	400
Total	**4.4**	**5,000**

[1] Wastewater-treatment plant operators, oral commun., 1991.
[2] James M. Montgomery Engineers, 1985, p. 4-5.

Goshen Valley, thus, areas of water-level decline were estimated from the March 1990 to March 1991water-level change map. On the basis of this analysis, the average water-level decline throughout the 95,000 acres in Goshen Valley was 0.6 ft. Multiplying this value by a specific yield of about 0.06 equals a release from storage of 3,400 acre-ft. Because of the small storage coefficient of confined zones, storage released from confined conditions was considered to be negligible.

Movement

Ground water generally moves from the mountain fronts to Utah Lake as shown by the potentiometric contours on plate 1. The contours also indicate ground-water movement toward Beer Creek and Benjamin Slough in southern Utah Valley and toward a ground-water withdrawal area south of Elberta in Goshen Valley. About 5,500 acre-ft/yr are withdrawn in the area south of Elberta. The potentiometric contours were derived from water levels in wells typically finished between 100 and 200 ft below land surface in the topographically low parts of southern Utah Valley, between 200 and 400 ft below land surface on the benches of southern Utah Valley, and between 200 and 700 ft below land surface in Goshen Valley. These intervals were chosen because more water-level data were available from these intervals than were available from other intervals and because most ground-water withdrawal probably occurs from these intervals. The altitude of the potentiometric contours in 1991 (pl. 1) was similar to the altitude of the potentiometric contours in 1965 (Cordova, 1970, pl. 1).

For the intervals described above, the horizontal gradient southwest of Payson is about 0.0096 (50 ft/mi), the gradient on Mapleton Bench is about 0.0053 (28 ft/mi), and the gradient near Utah Lake is about 0.0030 (16 ft/mi). Horizontal gradients in Goshen Valley range from 0.0015 (8 ft/mi) near Utah Lake to 0.0075 (40 ft/mi) south of Elberta. Not enough water-level data were available to draw contours or determine horizontal gradients in the upper 50 ft of saturated fill in the topographically low parts of the valleys. The horizontal gradient probably is similar to the slope of the land surface, and ground water probably moves toward springs, drains, and Utah Lake. Limited water-level data for wells finished more than 300 ft below land surface in the low parts of southern Utah Valley indicate that the horizontal gradient is similar to the horizontal gradient described previously, but ground-water movement toward Beer Creek and Benjamin Slough is not as evident.

An estimated 7,800 acre-ft/yr of ground-water flows from the Santaquin area in southern Utah Valley to the Genola area in Goshen Valley through Goshen Gap (table 4 and pl. 1). The estimate was made using Darcy's Law as expressed previously. On the basis of water levels in wells (D-9-1)35bcd-1 and (D-9-1)32bbd-1 (Stolp and others, 1993, table 1), the hydraulic gradient between the two wells is 0.010. On the basis of a transmissivity of 44,000 ft^2/d (Cordova, 1970, table 11) at well [1](D-9-1)35abb-1, with a perforated interval of 285 ft (Stolp and others, 1993, table 1),

the basin-fill deposits in this area have a hydraulic conductivity of 154 ft/d. The thickness of the saturated unconsolidated basin-fill deposits through Goshen Gap was assumed to be 200 ft. Multiplying the hydraulic conductivity by the saturated thickness yields a transmissivity through Goshen Gap of 31,000 ft^2/d. The length of the unconsolidated basin-fill deposits through Goshen Gap is about 3,000 ft (pl. 1). The water-level and transmissivity data are from wells near Goshen Gap but are not directly in the flow path. The estimate assumed no flow through bedrock, but some flow might occur through bedrock and faults. The estimated 7,800 acre-ft/yr is recharge to Goshen Valley and discharge from southern Utah Valley.

Vertical movement of ground water also occurs and is typically downward near the mountains and upward in the topographically low parts of the valleys. Water-level data indicate that the downward gradient on Mapleton Bench is about 0.22 (Stolp and others, 1993, table 1, wells (D-8-3)27bad-1 and (D-8-3)27bad-2) and that the upward gradient near Salem Lake is about 0.056 (Stolp and others, 1993, table 1, wells (D-9-2)10cad-2 and (D-9-2)10cad-4). In Goshen Valley, most wells have perforated sections that exceed 200 ft in length, which makes it difficult to determine if vertical movement occurs. The water level in two wells near Elberta (Stolp and others, 1993, table 1, wells (C-10-1)17bba-1 and (C-10-1)17bba-2), located within 100 ft of each other and finished at different depths, is similar, which indicates no significant vertical movement in this area.

Water-Level Fluctuations

Long-term and seasonal water-level fluctuations are caused by changes in recharge and discharge. Fluctuation generally is related to aquifer properties, distance to recharge and discharge areas, and the amount of recharge and discharge. Water levels near Utah Lake may fluctuate in response to lake level, but data are not available to determine the relation. Water-level data are presented in following sections and also in Cordova (1969, table 2) and Stolp and others (1993, table 3).

Long-Term Fluctuations

To determine long-term water-level fluctuations, water levels measured in 33 wells in March 1965 were compared statistically with water levels measured in the same wells in March 1991 using the Wilcoxon Signed-Rank test (Ott, 1988, p. 199-202). More data were available for these years than any other years, and both years followed a period of less-than-average precipitation (fig. 3). The location of the wells and the water-level change from March 1965 to March 1991 for each well are shown in figure 13. In southern Utah Valley, water levels in March 1991 were not significantly different from water levels in March 1965. In Goshen Valley, the test indicates with 90-percent confidence that the water levels in March 1991 were higher than water levels in March 1965. The mean of water levels measured in nine wells in Goshen Valley in March 1991 was about 3.5 ft higher than the mean in March 1965.

Although water levels in southern Utah Valley were not significantly different in March 1991 than in March 1965, considerable fluctuation occurred between 1965 and 1991. The relation of cumulative departure from average annual precipitation at Payson to water level in selected wells in southern Utah Valley and water level in Utah Lake is shown in figure 14. Water level in southern Utah Valley primarily fluctuates in response to changes in precipitation. Greater-than-average precipitation (fig. 14, cumulative departure curve at Payson) during the early 1980's increased recharge to the basin-fill aquifers and probably resulted in decreased ground-water withdrawals. In March 1985, the highest recorded water level was observed in several wells. The less-than-average precipitation from 1987 to 1990 decreased recharge to the basin-fill aquifers and probably resulted in increased ground-water withdrawal. Municipal ground-water withdrawal in 1989 and 1990 was about 8,000 acre-ft/yr, more than double the municipal withdrawal of previous years (fig. 10). According to city water managers, the increased withdrawal was caused mainly by the decreased flow of springs from consolidated rock during these years. In March 1990, the lowest recorded water levels were observed in some of the same wells that had the highest recorded water levels in March 1985. Wells (D-8-2)16caa-1 and (D-8-2)34dda-1 (fig. 14) are examples.

The magnitude of long-term water-level fluctuations varies throughout southern Utah Valley and depends on distance from recharge and discharge areas. In ground-water discharge areas near Provo Bay, water level declined 3 to 10 ft from March 1986 to March 1991 (Herbert and others, 1991, figs. 15 and 16). Near Spanish Fork, the water level declined about 10 ft from March 1986 to March 1991. In ground-water recharge areas near Salem, Payson, and Santaquin, the water

[1]Previously reported as well (D-9-1)35aba-1 (Cordova, 1969 and 1970).

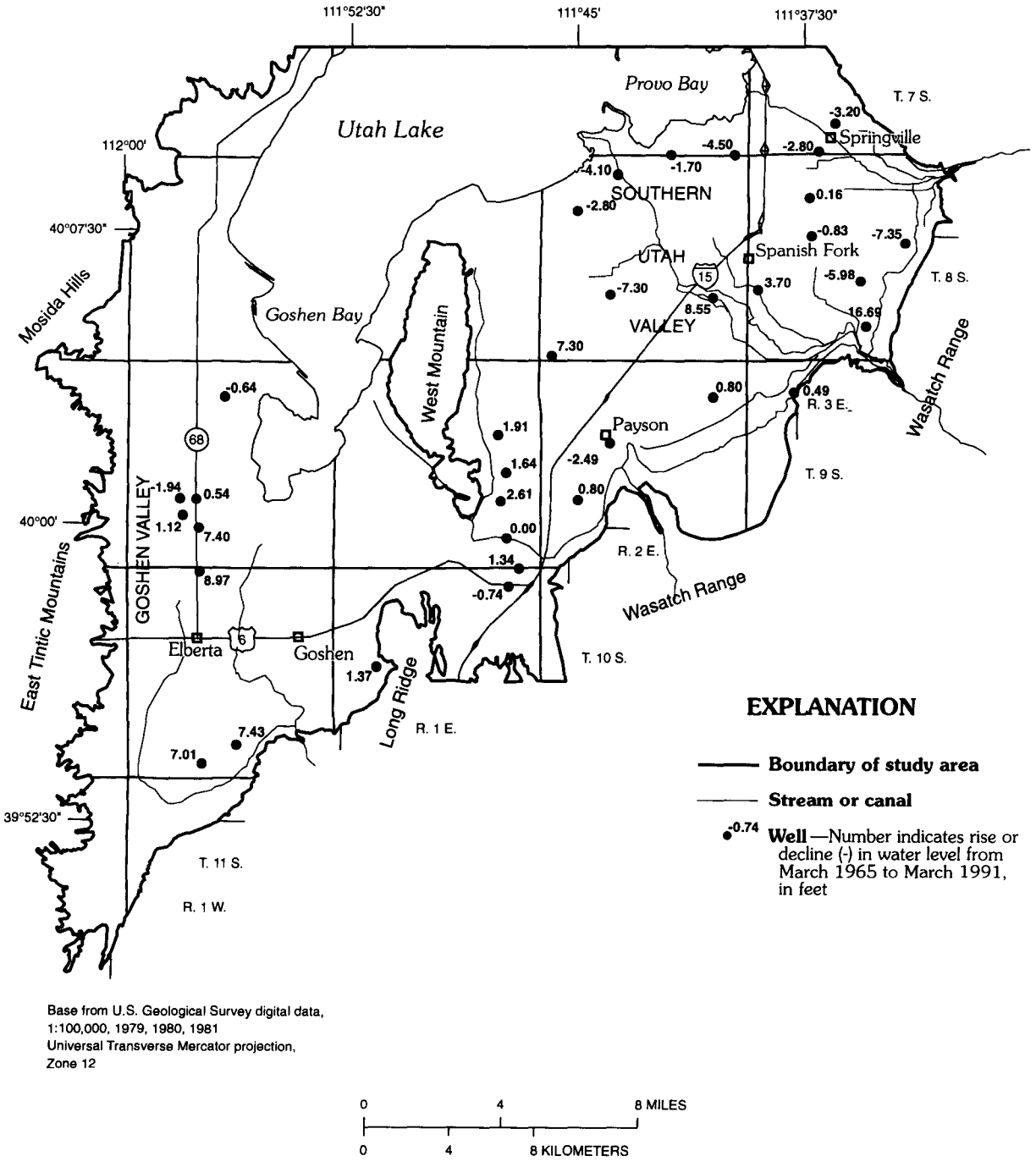


Figure 13. Water-level change in selected wells in southern Utah and Goshen Valleys, Utah, from March 1965 to March 1991.

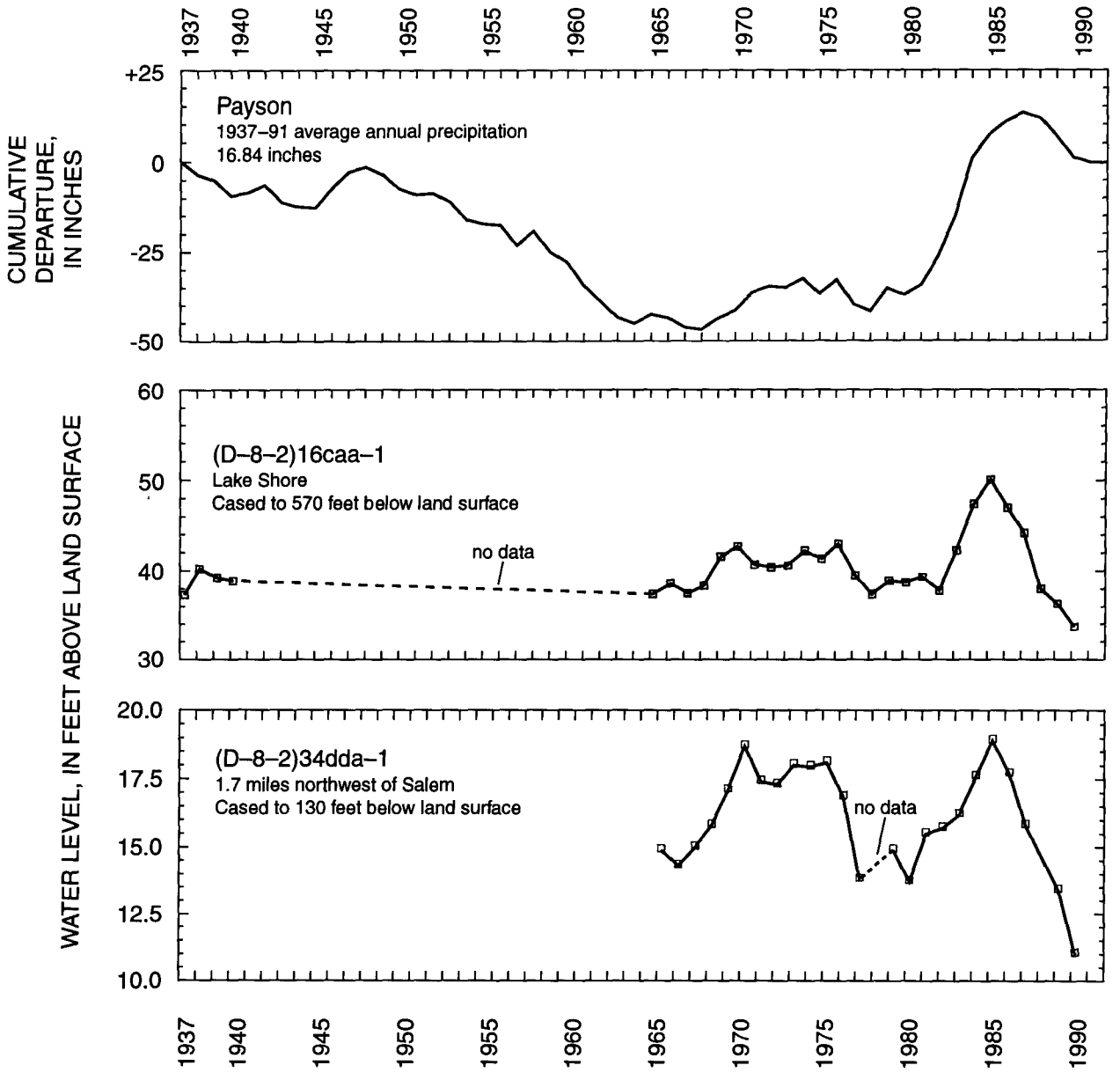


Figure 14. Relation of cumulative departure from average annual precipitation (1937-91) at Payson, Utah, to water level in selected wells in southern Utah Valley and water level in Utah Lake.

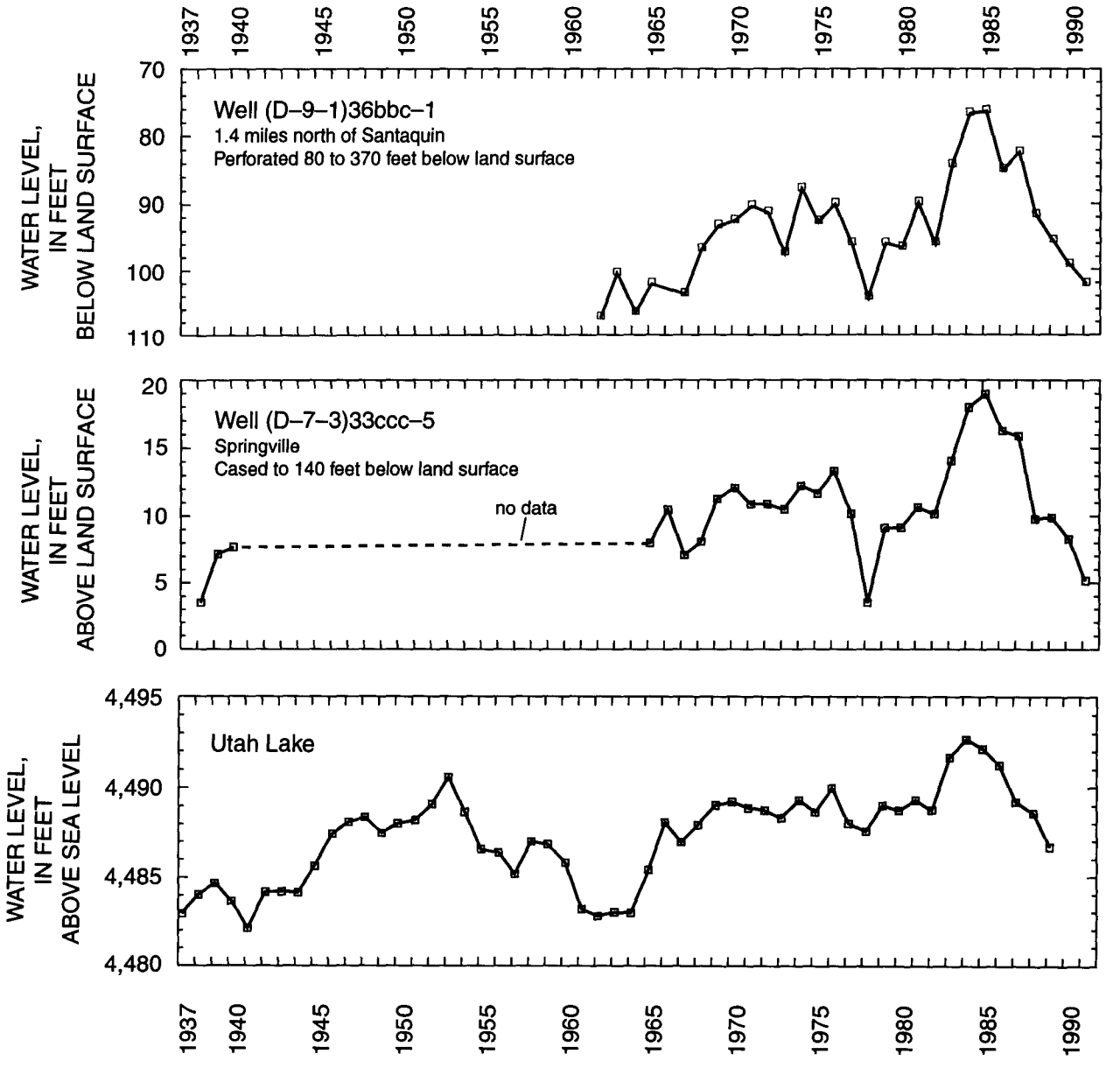


Figure 14. Relation of cumulative departure from average annual precipitation (1937-91) at Payson, Utah, to water level in selected wells in southern Utah Valley and water level in Utah Lake—Continued.

level declined about 15 to 20 ft from March 1986 to March 1991. The water level in well (D-9-1)36bbc-1, in a recharge area near Santaquin, fluctuates more than the water level in well (D-7-3)33ccc-5, in a discharge area in Springville (fig. 14). Water level in shallow wells such as (D-8-2)34dda-1 fluctuates less than water level in deep wells such as (D-8-2)16caa-1 (fig. 14).

Although the water level in Goshen Valley rose from March 1965 to March 1991, it generally declined during the 1960's. The relation of cumulative departure from average annual precipitation at Elberta to water level in selected wells in Goshen Valley and annual withdrawals from selected wells is shown in figure 15. Large ground-water withdrawal for irrigation north of Elberta began in the early 1960's and continued through about 1980. Ground-water withdrawal north of Elberta decreased during the 1980's as land was taken out of production. The ground-water level declined about 20 ft from the early to late 1960's but now has nearly recovered to pre-pumping level as shown by measured water level in well (C-9-1)20ddd-1 (fig. 15). The ground-water level in Goshen Valley appears to be affected only slightly by precipitation. Despite average or less-than-average precipitation from 1987 to 1990, water level in observation wells in the northern part of Goshen Valley increased about 3 ft from March 1986 to March 1991 (Herbert and others, 1991, fig. 14). The lack of water-level fluctuation in response to precipitation in Goshen Valley indicates that recharge to the ground-water system from precipitation on the valley floor is negligible. Dispersion ponds related to mining activities in the East Tintic Mountains are located west of Elberta in section 18, Township 10 South, Range 1 West. An estimated 6,200 acre-ft/yr of water was diverted to the ponds from 1969 to 1978 (Mack Croft, Utah Department of Environmental Quality, Division of Water Quality, written commun., 1989). Water-level fluctuations in well (C-10-1)4bbb-1 (fig. 15) indicate that this diverted water contributed recharge to Goshen Valley.

South of Elberta, withdrawal data and water-level data are not as abundant as north of Elberta. Water-level fluctuation in well (C-10-1)24ddc-1 (fig. 15) probably is caused by ground-water withdrawal from nearby wells.

Seasonal Fluctuations

In southern Utah and Goshen Valleys, the highest water level typically occurs from January to April, and the lowest water level typically occurs from July to September. Seasonal water-level fluctuations in selected wells in southern Utah Valley are shown in figure 16. Seasonal water-level fluctuations in selected wells in Goshen Valley are shown in figure 17.

In flowing wells in discharge areas near Provo Bay and Springville, seasonal water-level fluctuations of about 5 ft occur. High water levels typically occur in March and low levels typically occur in August or September, as shown by the water level in well (D-7-3)33ccc-5 (fig. 16). Shallow and deep wells have about the same magnitude of seasonal fluctuation. Seasonal water-level fluctuations east and southeast of the town of Spanish Fork are about 8 ft. High water levels typically occur in January and low levels typically occur in August, as shown by the water level in well (D-8-3)14acc-1 (fig. 16). These summer declines probably are caused by municipal and irrigation withdrawals. The water level in well (D-9-1)26aab-1, a nonpumping well north of Santaquin, declined about 5 ft from May to September in 1989 and 1990, but rose only slightly during the winter months (fig. 16). Less-than-average precipitation in 1989 and 1990 probably reduced ground-water recharge in this area, and the water level could not recover seasonally.

North of Elberta in Goshen Valley, seasonal water-level fluctuations in wells more than 1 mile from pumping wells were small, as shown by the water level in well (C-10-1)4bbb-1 (fig. 17). Seasonal fluctuations in wells within 1 mile of pumping wells are about 2 ft, as shown by the water level in well (C-9-1)20ddd-1 (fig. 17). High water levels typically occur in April or May, and low levels typically occur in September or October. About 1,100 acre-ft/yr was pumped from wells within 1 mile of well (C-9-1)20ddd-1 from 1988 to 1990 (fig. 15). In southern Goshen Valley, seasonal fluctuations in well (C-10-1)24ddc-1 are about 10 ft (fig. 17). About 650 acre-ft/yr was pumped from wells within 1 mile of well (C-10-1)24ddc-1 from 1988 to 1990 (fig. 15).

Water levels measured in 27 wells in March 1990 were compared to water levels measured in the same wells in September 1990. The location of the wells and the water-level change for each well from March 1990 to September 1990 are shown in figure 18. In southern Utah Valley, the September water level typically was 2 to 10 ft lower than the March water levels. Declines of 5 to 10 ft in the southern part of southern Utah Valley are caused mainly by summer municipal withdrawal. Declines of 5 to 10 ft north of Santaquin and south of Goshen are caused mainly by irrigation withdrawal.

Aquifer Properties

Aquifer properties, including transmissivity (T), hydraulic conductivity (K), and storage coefficient (S), were determined from aquifer tests, lithologic data from drillers' logs, and specific-capacity data. Transmissivity and storage coefficient of the basin-fill deposits at select locations, as determined from aquifer tests, are listed in table 13 and shown in figure 19. Generally, transmissivity is largest southeast of Springville and smallest near Utah Lake, but conditions vary locally. All pumped wells used for the aquifer tests are large-production wells and are typically perforated in the most permeable materials encountered during drilling; therefore, reported transmissivity may be higher than average transmissivity throughout the thickness of the unconsolidated basin-fill deposits.

The lithology reported on drillers' logs of 13 wells was analyzed to determine average hydraulic conductivity throughout the saturated depth of each well as previously explained in the "Utah Lake" section of the "Ground-water hydrology" section of this report. Location of the wells, transmissivity, and average hydraulic conductivity are listed in table 14 and transmissivity is shown in figure 19. Transmissivity determined from lithology was generally less than one-half of transmissivity determined from aquifer tests in the same area. This difference was caused mainly by inclusion in the lithology calculations of silt and clay and generally is not included in the perforated zone of the wells used for aquifer tests. The mean of the transmissivity of the perforated length determined from lithologic data for 10 wells is not significantly different from the mean of the transmissivity determined by aquifer tests for the same wells. Individual results, however, differ greatly for some wells.

Transmissivity was calculated using specific-capacity data (Theis and others, 1963) for about 70 wells. Transmissivity determined by this method ranged from 500 to 131,000 ft^2/d in southern Utah Valley and from 600 to 43,000 ft^2/d in Goshen Valley. The mean of the transmissivity determined from specific-capacity data for six wells was not significantly different from the mean of the transmissivity determined from aquifer tests for the same wells. Individual results, however, differ greatly for some wells.

Specific yield of deposits is the volume of water yielded from the deposits by gravity drainage per unit volume of deposits under unconfined conditions (Lohman, 1979, p. 6). Most of the unconfined water in southern Utah and Goshen Valleys occurs where deposition from landslides and other colluvial processes has resulted in poorly sorted gravel, sand, and clay mixtures. Cordova (1970, table 12 and p. 56) reports that deposits in most areas where unconfined conditions occur are composed of about 12 percent gravel and coarser material with a specific yield of 0.25, 54 percent gravel with clay and silt with a specific yield of 0.05, 28 percent clay and silt with a specific yield of 0.03, and 6 percent other materials. The storage coefficient of 0.02 reported for well [1](D-9-1)35abb-1 near Santaquin (table 13) probably represents unconfined conditions and materials with a low specific yield. Average specific yield for the deposits containing water under unconfined conditions in both valleys is about 0.06 (Cordova, 1970, table 12), which would result in 0.06 acre-ft of water being released for a 1-foot decline in water level throughout a 1-acre area.

Storage coefficient is the volume of water an aquifer under confined conditions releases from or takes into storage per unit surface area of the aquifer per unit change in head (Lohman, 1979, p. 8). The storage coefficient for confined conditions in southern Utah Valley ranges from 2×10^{-5} to 3×10^{-3} (table 13). Todd (1959, p. 45-46) reports that most confined aquifers have a storage coefficient ranging from 5×10^{-5} to 5×10^{-3}. An average storage coefficient of 1×10^{-4} would result in 1×10^{-4} acre-ft of water being released for a 1-foot decline in water level throughout a 1-acre area.

No data are available in southern Utah and Goshen Valleys to determine vertical hydraulic conductivity. Clark and Appel (1985, p. 47) report a vertical hydraulic conductivity of 1×10^{-3} ft/d from an aquifer test in northern Utah Valley. Freeze and Cherry (1979, p. 34) report that vertical hydraulic conductivity is often less than one-hundredth of horizontal hydraulic conductivity.

Chemical Quality

An analysis of ground-water samples collected and analyzed from 1959 to 1991 (Stolp and others, 1993, table 5) indicates that ground-water quality has not changed substantially in southern Utah Valley. Most of the samples collected from wells finished in unconsolidated basin-fill deposits in southern Utah Valley had a dissolved-solids concentration of 200 to 400 mg/L, but a few samples had a dissolved-solids concentration as high as 1,100 mg/L. Most of the ground-water

[1]Previously reported as well (D-9-1)35aba-1 (Cordova, 1969 and 1970).

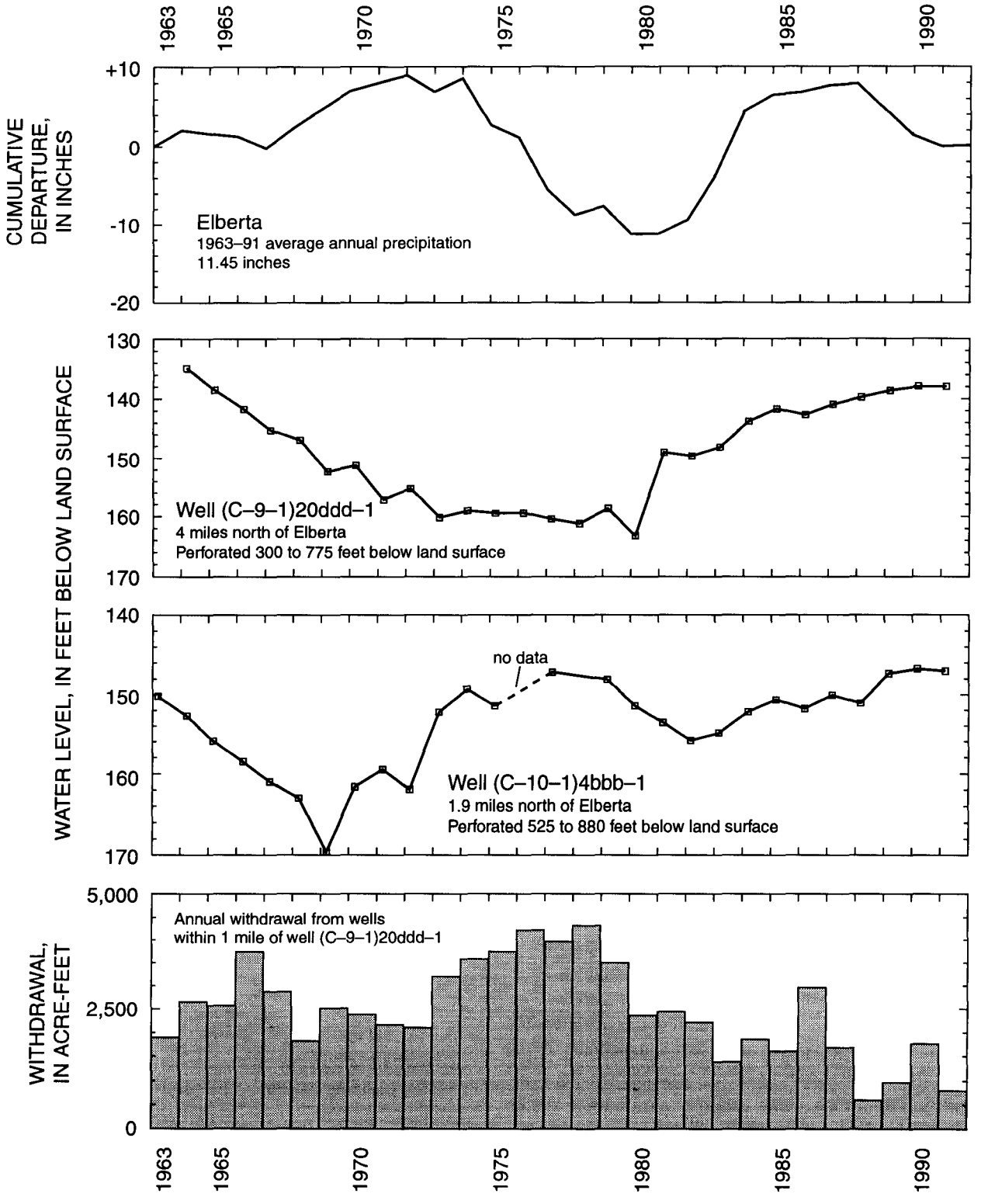


Figure 15. Relation of cumulative departure from average annual precipitation (1963-91) at Elberta, Utah, to water level in selected wells in Goshen Valley and annual withdrawal from selected wells.

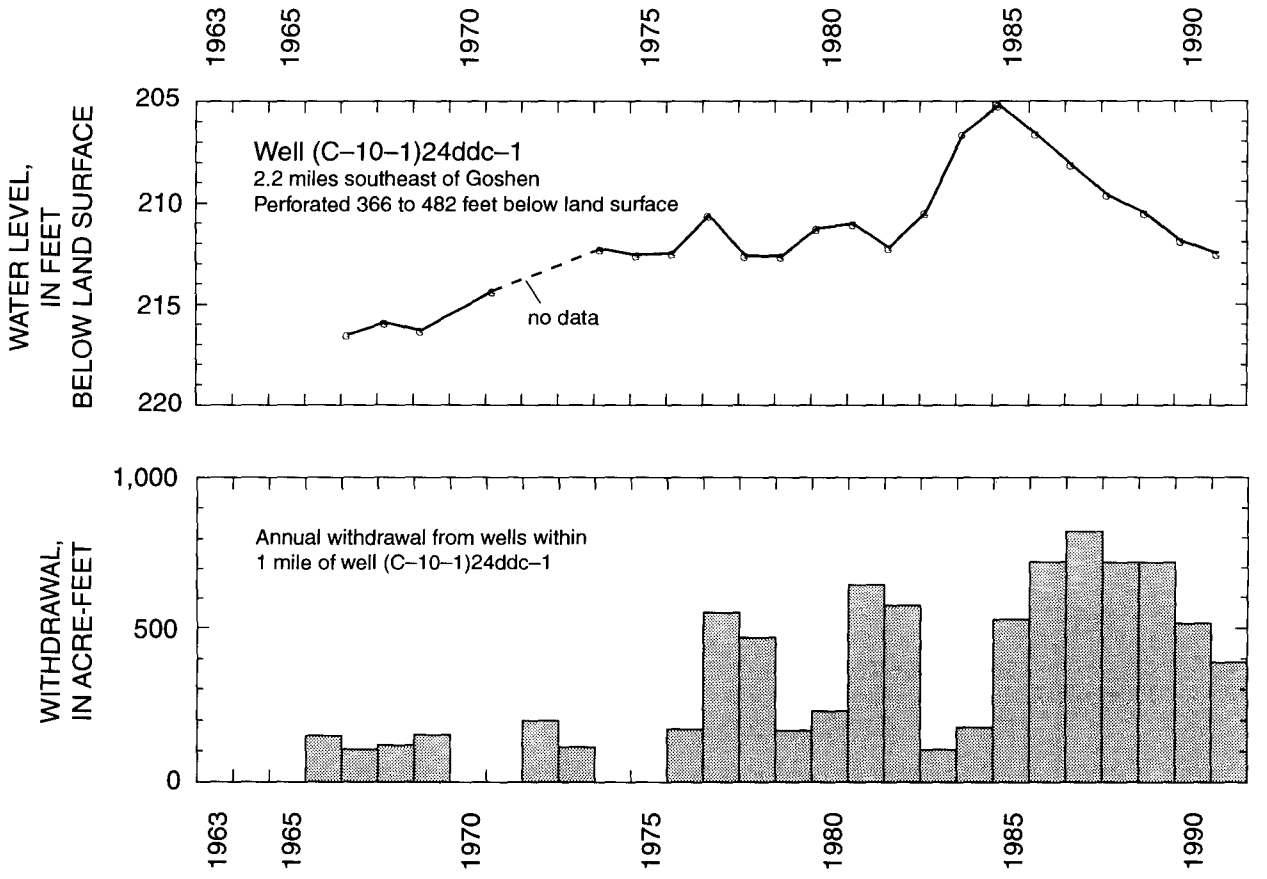


Figure 15. Relation of cumulative departure from average annual precipitation (1963-91) at Elberta, Utah, to water level in selected wells in Goshen Valley, Utah, and annual withdrawals from selected wells—Continued.

samples contain calcium and magnesium as the major cations and bicarbonate as the major anion. Ground-water quality does not vary much throughout the valley.

Few data are available to analyze water quality in southern Utah Valley as a function of depth, but quality appears to change with depth only in ground-water discharge areas. Samples collected from wells completed deeper than 200 ft below land surface around Benjamin and south of Provo Bay show similar quality to samples collected elsewhere in the valley, with a dissolved-solids concentration of 200-400 mg/L. Samples collected from wells completed shallower than 200 ft below land surface in those areas, however, have a dissolved-solids concentration of about 400 to 450 mg/L (Stolp and others, 1993, table 5) and have a larger concentration of sodium and chloride than samples collected elsewhere in the valley. In some samples from shallow wells, sodium is the predominant cation and bicarbonate is the predominant anion. Most springs and drains in the topographically low parts of the valley have a dissolved-solids concentration of 400-900 mg/L, but a few have a dissolved-solids concentration of about 1,500 mg/L (Cordova, 1969, table 7, and Mundorff, 1974, table 20). An increase in dissolved-solids concentration and change of water type in the shallow ground water in the topographically low parts of the valley could be caused by longer flow paths and contact times, water passing upward through beds containing sodium and chloride, or concentration by evapotranspiration. A combination of these factors probably occurs. Another possible source of sodium and chloride is upward migration of water through a fault on the east side of West Mountain. Springs on Lincoln Point and Bird Island associated with the fault have a sodium concentration of 725 to 2,207 mg/L and a chloride concentration of 383 to 3,500 mg/L (Baskin and others, 1994, table 4).

From 1960 through 1991, no consistent water-quality trends were evident in Goshen Valley, but individual wells have yielded water of differing quality throughout the period. Some wells have yielded water with an increasing dissolved-solids concentration

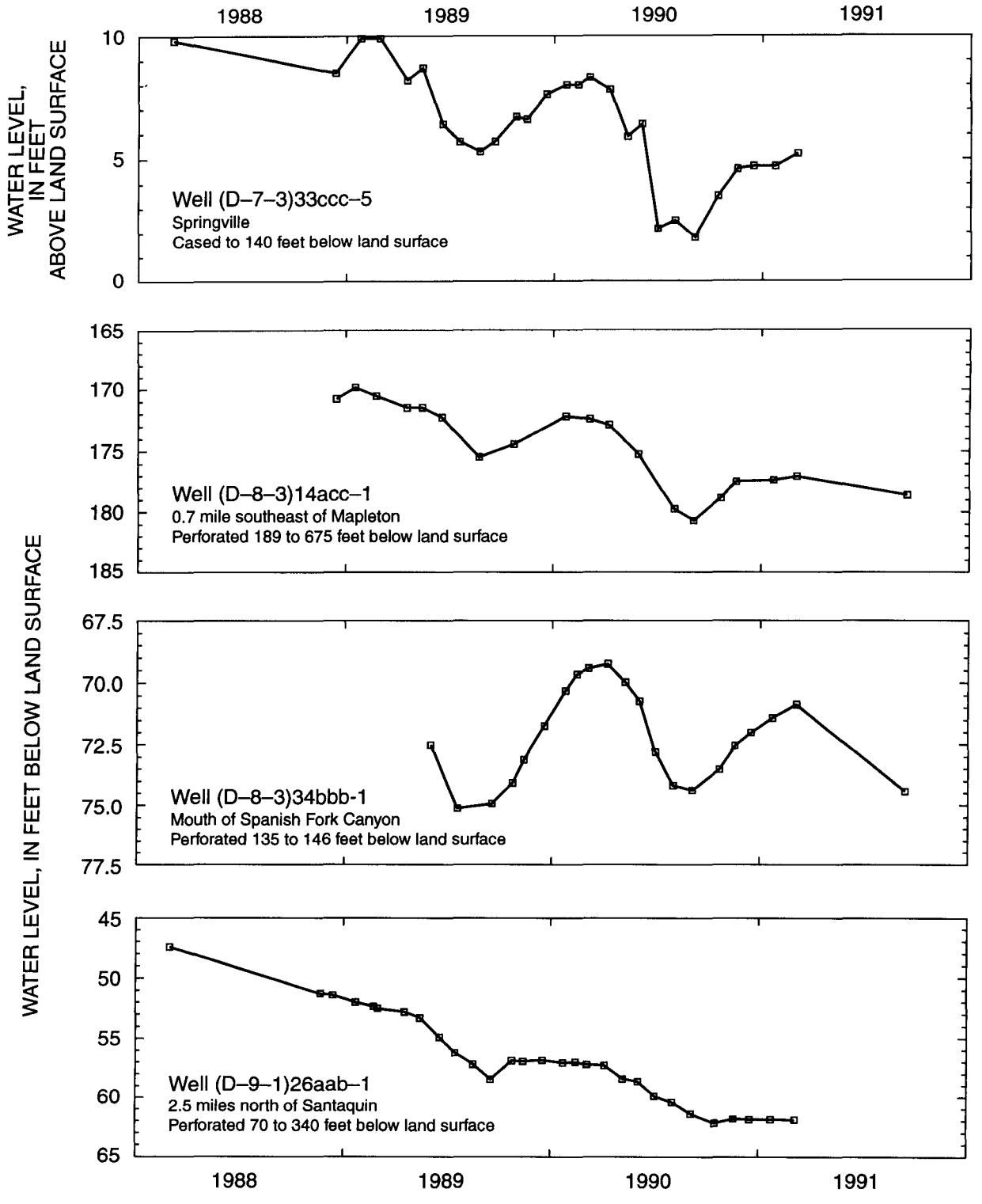


Figure 16. Seasonal water-level fluctuations in selected wells in southern Utah Valley, Utah.

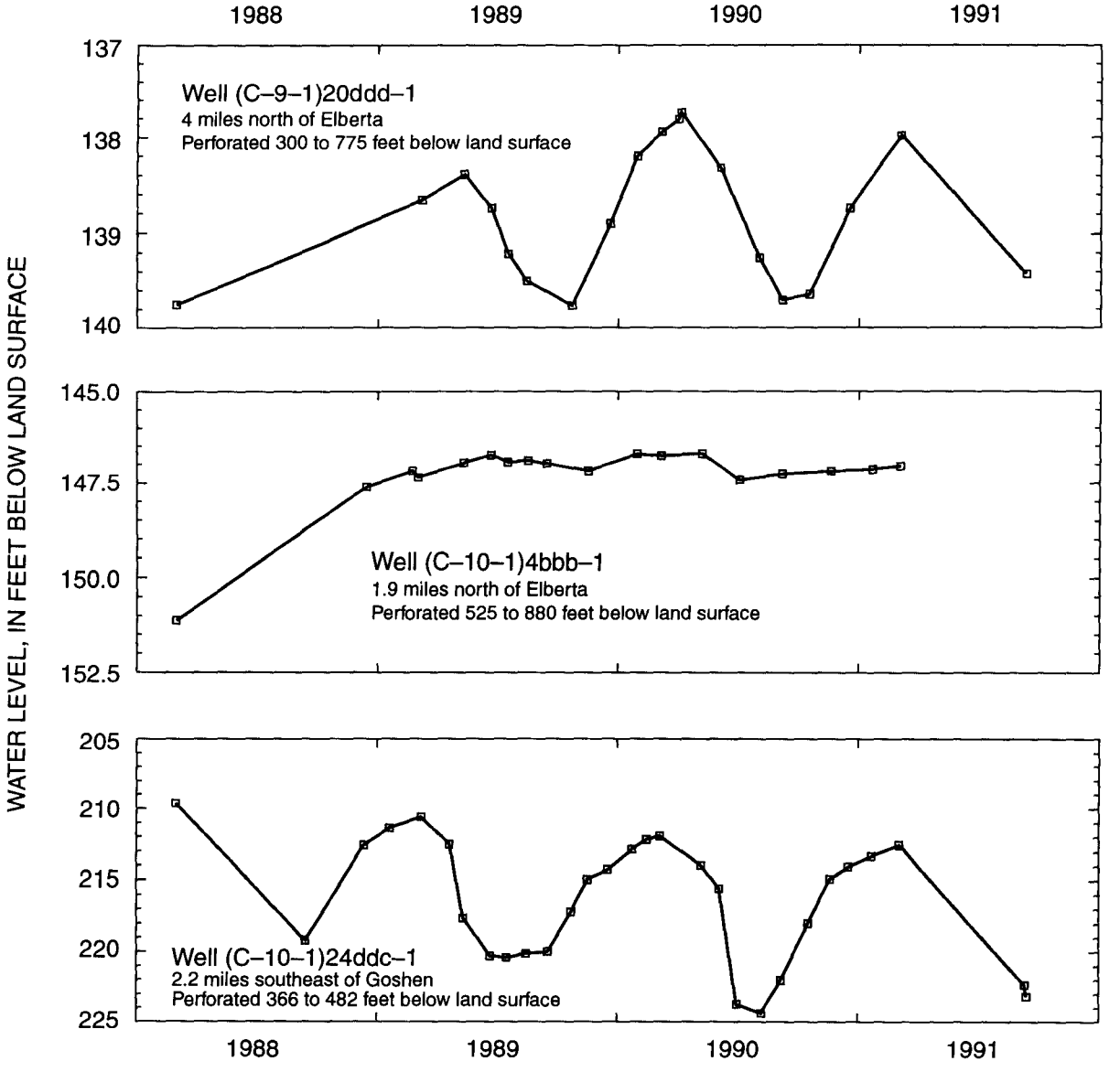


Figure 17. Seasonal water-level fluctuations in selected wells in Goshen Valley, Utah

through time (Stolp and others, 1993, table 5, well (C-9-1)4ccc-1), and other wells have yielded water with a decreasing dissolved-solids concentration through time (Stolp and others, 1993, table 5, well (C-10-1)9ccc-1). The observed fluctuations in dissolved-solids concentration may be caused by changes in ground-water-flow direction induced by ground-water withdrawals for irrigation, but not enough data are available to determine this.

Dissolved-solids concentration from samples collected in Goshen Valley ranges from 200 to as high as 2,600 mg/L (Stolp and others, 1993, table 5), with smaller concentrations generally occurring south of Elberta and larger concentrations generally occurring north of Elberta. In all areas, however, anomalies exist, and the quality of water from a particular well cannot be determined on the basis of the quality of water from other wells in the area. Stiff diagrams showing the results of analyses of selected samples are shown in figure 20. Samples collected south of Elberta indicate the ground water is generally a mixed type, with calcium and magnesium predominant over sodium, and bicarbonate predominant over chloride and sulfate (fig. 20, well (C-11-1)6bdd-1). A sample collected from well (D-10-1)30bac-1, southeast of Goshen (fig. 20), however, has a larger dissolved-solids concentration than

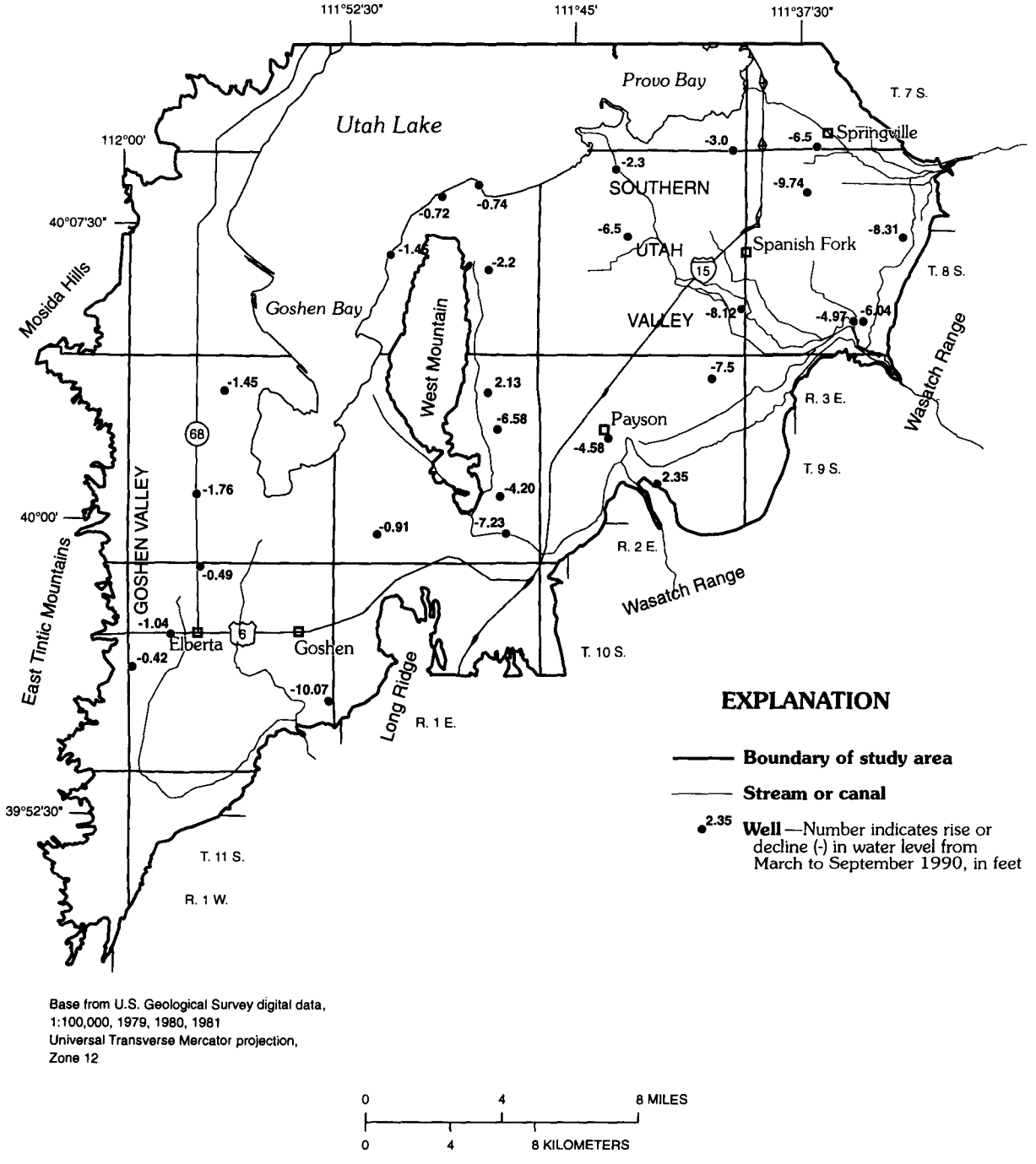


Figure 18. Water-level change in selected wells in southern Utah and Goshen Valleys, Utah, from March to September 1990.

Table 13. Results of aquifer tests, southern Utah and Goshen Valleys, Utah

[—, unknown]

Location: See figure 2 for explanation of numbering system for hydrologic-data sites.

Aquifer properties: The estimates of transmissivity and storage coefficient are for the perforated zone of the pumping well in the saturated part of the aquifer.

Method of analysis: C, values modified from Cordova (1970, table 11); H, Hantush modified method (Lohman, 1979, p. 32); S, Straight line method (Lohman, 1979, p. 23); T, Theis method (Lohman, 1979, p. 15).

Location				Aquifer properties		
Pumped well	Observation well	Date	Average discharge (gallons per minute)	Transmissivity (T) (feet squared per day)	Storage coefficient (S) or specific yield	Method of analysis
			Southern Utah Valley			
(D-7-3)34cdb-1	(D-7-3)33ccc-6 (D-7-3)34bcb-1 (D-8-3) 4daa-1	February 1967	1,300	42,000	3×10^{-3}	C
(D-8-2)25dac-3	(D-8-2)23dca-2 (D-8-2)24bdc-2 (D-9-2)1bcb-1 (D-9-3)5bbd-1	January 1967	520	15,000	1×10^{-3}	C
(D-8-3)11ccc-1	(D-8-3)3dcd-1 (D-8-3)10cba-1	February 1967	930	71,000	5×10^{-4}	C
[1](D-8-3)22cbd-3	(D-8-3)27bad-1	January 1990	900	14,000	2×10^{-5}	H[2]
(D-8-3)28abc-1	(D-8-3)28bcd-1 (D-8-3)28bdc-1 (D-8-3)27cdc-1	November 1965	780	29,000	4×10^{-4}	C
[3](D-9-1)35abb-1	(D-9-1)36bbc-1 (D-9-1)36cdd-1 (D-10-1)2adb-1 (D-10-1)2bba-1	February 1967	2,500	44,000	2×10^{-2}	C
(D-9-2) 9bac-1	(D-9-2)4cdc-1 (D-9-2)20bbc-2	February 1967	1,900	31,000	3×10^{-3}	C
			Goshen Valley			
[4](C-9-1) 4ddc-1	None	March 1965	1,800	7,500	—	S
(C-9-1)29acc-1	None	May 1990	1,600	[5]35,000	—	S
(C-10-1)4cbb-1	None	—	—	5,100	—	C
[4](C-10-1)31cdd-1	(C-11-1)6abc-1 (C-11-1)6bdd-1	April 1967	1,600	44,000	5×10^{-4}	T

[1] Previously reported as well (D-8-3)22cac-1 (Cordova, 1969 and 1970). Observation well perforated for only 20 feet, so may not represent aquifer properties.

[2] Also analyzed by the Neuman delayed-yield method (Neuman, 1975) and the recharging-image method (Lohman, 1979, p. 59-61). All analyses yielded similar results.

[3] Previously reported as well (D-9-1)35aba-1 (Cordova, 1969 and 1970).

[4] Aquifer-test data reanalyzed for this report.

[5] Aquifer test was only 30 minutes long, causing a large uncertainty in the results.

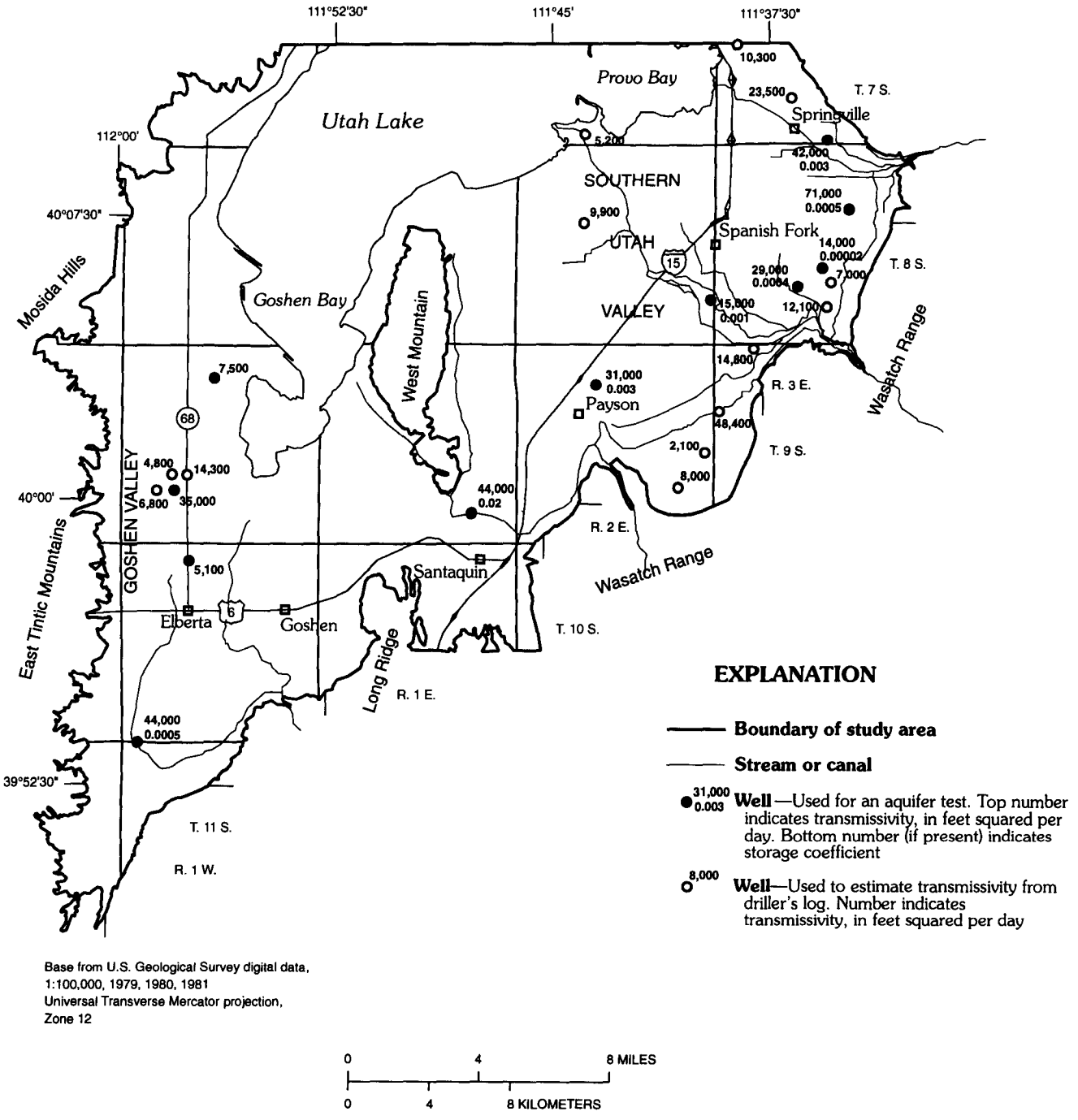


Figure 19. Location of wells used for aquifer tests and to estimate transmissivity, southern Utah and Goshen Valleys, Utah.

Table 14. Transmissivity and average hydraulic conductivity of the saturated depth of selected wells as determined from the lithology reported on drillers' logs of selected wells in southern Utah and Goshen Valleys, Utah
Location: See figure 2 for explanation of numbering system for hydrologic-data sites.

Location	Transmissivity (feet squared per day)	Average hydraulic conductivity (feet per day)
	Southern Utah Valley	
(D-7-2)32dad-1	5,200	10
(D-7-3)18dcc-1	10,300	30
(D-7-3)28cab-1	23,500	80
(D-8-2)17ada-1	9,900	20
(D-8-3)27bad-2	7,000	20
(D-8-3)27cdc-1	12,100	30
(D-9-2)24aca-1	2,100	10
(D-9-2)26add-1	8,000	30
(D-9-3)5bbd-1	14,800	30
(D-9-3)18bbb-1	48,400	150
	Goshen Valley	
[1] (C-9-1)20cdd-1	4,800	10
(C-9-1)20ddd-1	14,300	20
(C-9-1)29bcc-1	6,800	10

[1]Previously reported as well (C-9-1)20dcc-1 (Cordova, 1969 and 1970).

than most other wells in Goshen Valley, and sodium and chloride are predominant. Samples collected north of Elberta indicate that sodium predominates over calcium and magnesium, and chloride predominates over sulfate and bicarbonate (fig. 20, wells (C-10-1)4cbb-1 and (C-9-1)4ccc-1). The change in water type from south to north, in the direction of ground-water flow, could be caused by longer flow paths and contact times with sodium and chloride deposits, or by inflow from a source of water containing a high concentration of sodium and chloride.

Relation to Source

In southern Utah Valley, ground-water quality is similar to the water quality of streams that enter the valley, springs that discharge from basin-fill deposits near the mountain front, and springs that discharge from consolidated rock in the mountains. Because recharge occurs mostly from surface water and subsurface inflow, similarities in water quality are expected.

In Goshen Valley, the quality of the ground water is more similar to that of springs and tunnels in consolidated rock than to that of surface water that enters the valley in Currant Creek. This indicates that subsurface inflow may be a major source of recharge to the ground-water system in Goshen Valley. Selected water-quality data from Currant Creek, springs in consolidated rock near Goshen Valley, and wells in Goshen Valley are listed in table 15.

The chemical composition of samples collected from springs in consolidated rock in the mountains south of Goshen Valley is similar to the chemical composition of ground-water samples collected in southern Goshen Valley. The chemical composition of samples collected from Warm Springs is similar to the chemical composition of samples collected from wells southeast of Goshen. Samples collected from a mine tunnel in the East Tintic Mountains west of Elberta contain a larger dissolved-solids concentration and larger percentage of sodium and chloride than any ground-water sample collected in Goshen Valley. This indicates that flow from

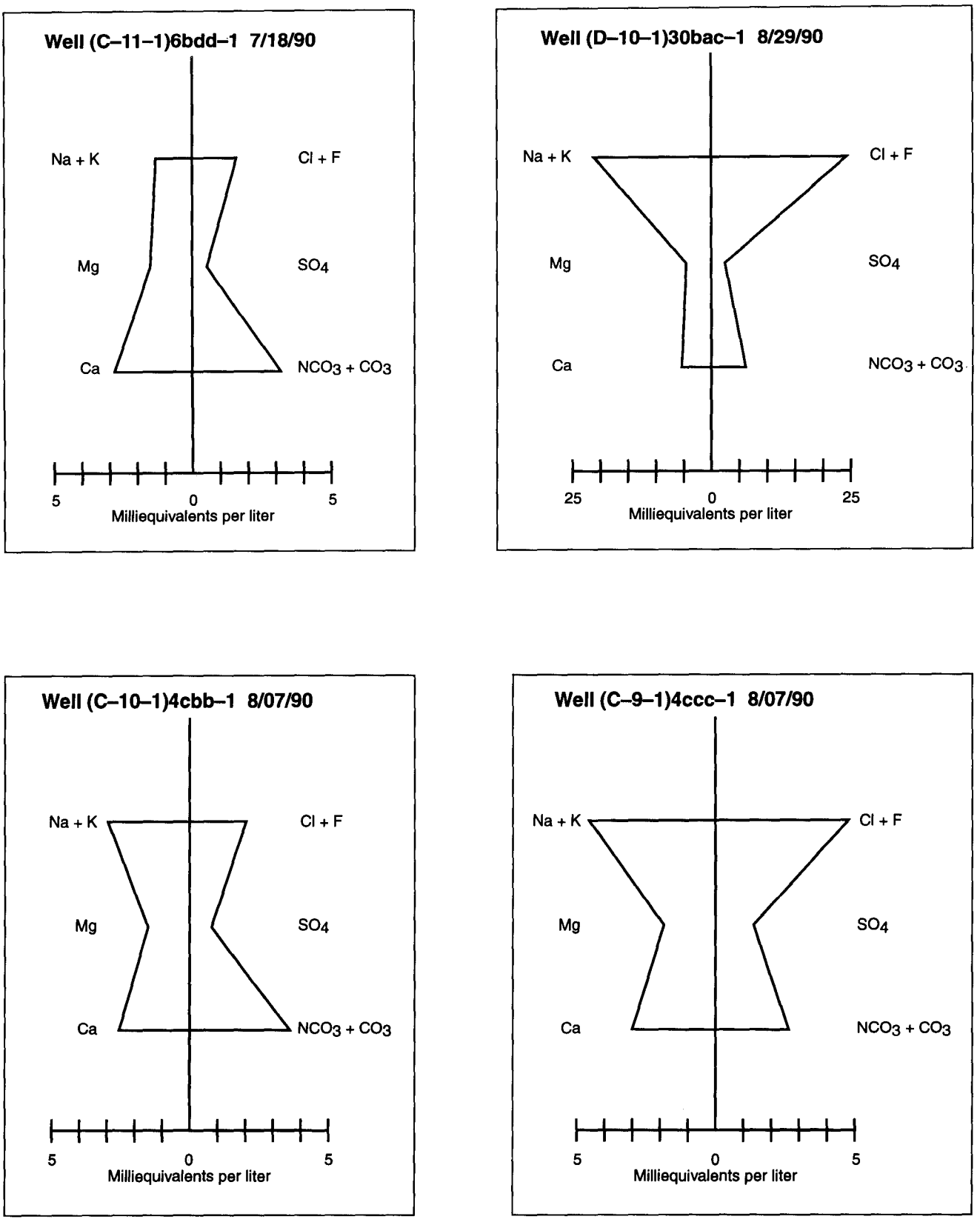


Figure 20. Chemical composition of selected ground-water samples from Goshen Valley, Utah.

bedrock could be entering the ground-water system along the East Tintic Mountains and causing the change in chemical composition from south to north, as discussed previously.

Nitrate Concentrations

Water samples collected by the Utah Department of Environmental Quality in 1989 from Mapleton City municipal wells in sections 11 and 22, Township 8 South, Range 3 East, contained about 5 mg/L of nitrate as N (Utah Department of Environmental Quality, written commun., 1991). A possible source of the nitrate is a nitric acid spill that occurred upgradient in 1986. Private testing (Engineering-Science, Inc., 1990, fig. 5.1) indicates that nitrate concentration in the area is highest near the spill, and some samples contain as much as 50 mg/L nitrate as N. Other areas of the valley have a lower nitrate concentration.

Samples collected from four wells in Goshen Valley contain a nitrate concentration that ranges from 9 mg/L to 46 mg/L as N (Stolp and others, 1993, table 5). The wells are located from 3 mi south of Elberta to 7 mi north of Elberta. The elevated nitrate concentrations are in a north-south line, with the highest concentration occurring near Elberta. The source of the nitrate was not determined during this study.

SIMULATION OF THE MAIN GROUND-WATER SYSTEM IN THE UNCONSOLIDATED BASIN-FILL DEPOSITS

A modular, three-dimensional, finite-difference, ground-water flow model (McDonald and Harbaugh, 1988) was used to simulate the main ground-water system in the unconsolidated basin-fill deposits of southern Utah and Goshen Valleys. The steady-state conditions of 1949 and the annual transient-state conditions from 1949 to 1990 were used to calibrate the model. The Mapleton Bench ground-water system is not considered part of the main ground-water system (see "Occurrence" section of the "Ground-water hydrology" section of this report) and was not included as part of the simulated ground-water system.

The ground-water flow model was used to further examine the conceptual understanding of the main ground-water system presented in this report. Objectives of the model included compilation and examination of all hydrologic data from 1949 to 1990, using this data to formulate separate components of the ground-water system, and using simulations to examine the interaction of these components. These interactions were simulated under varied hydrologic stresses and anticipated future stresses. A spatial distribution of

Table 15. Average chemical composition of water in and near Goshen Valley, Utah

[mg/L, milligrams per liter; —, no data]

Location of sampling site: See figure 2 for explanation of numbering system for hydrologic-data sites.

Source	Location of sampling site	Number of samples	Calcium (mg/L)	Magnesium (mg/L)	Sodium (mg/L)	Bicarbonate (mg/L)	Sulfate (mg/L)	Chloride (mg/L)
Currant Creek[1]	(C-10-1)10ccd	2	64	58	195	293	155	295
Springs in consolidated rock in the mountains south of Goshen Valley[2]	(C-11-1)30acb (C-11-2)2ddd	4	70	21	29	267	42	42
Wells in southern Goshen Valley[3]	(C-11-1)6	10	47	19	38	—	34	56
Warm Springs[2]	(D-10-1)8cab	2	86	40	350	315	108	549
Wells southeast of Goshen[3]	(C-10-1)25abd (D-10-1)30bac	7	98	53	206	—	127	369
East Tintic Mountain mine tunnel[2]	(C-10-2)15ddd	2	318	72	1,930	627	361	3,336

[1] Mundorff, 1974, table 20.
[2] Cordova, 1969, table 7.
[3] Stolp and others, 1993, table 5.

hydrologic properties for the ground-water system was established and specific data required to increase current understanding of the system were identified.

A complete description of the construction and calibration of the ground-water flow model is presented in the following sections of the report. It is not possible, however, to reconstruct this model from the information presented here. Given the amount and complexity of the required input data, it is impractical to present or reference all the required information. A copy of the model and all associated data sets can be obtained from the U.S. Geological Survey, Water Resources Division, Salt Lake City, Utah.

Model Construction

Construction of the ground-water flow model was accomplished by horizontally and vertically discretizing the hydrologic properties of the main ground-water system, establishing model boundaries that depict actual hydrologic boundaries, and assigning realistic hydrologic processes to each of three boundaries. Initial values assigned to model parameters were adjusted to achieve a calibrated model that reasonably represents the simulated ground-water flow system.

Discretization

The lateral boundaries of the simulated ground-water system are shown in figure 21 and, in general, correspond with the lateral extent of the unconsolidated basin-fill deposits in southern Utah and Goshen Valleys. An exception to this is the northern boundary of the study area which is considered a no-flow boundary in the model (see "Physiography" in the "Introduction" section of this report). The rectangular model grid contains 45 rows and 103 columns; cell size ranges from 0.03 mi^2 to 0.9 mi^2. Smaller cells were used in areas where the hydraulic gradient is steeper and larger amounts of data are available. The model grid is oriented so that cell faces are typically perpendicular or parallel to the general direction of ground-water movement. The ground-water flow equations and associated hydraulic properties used to calculate flow for each model cell are formulated at the center point or node of the cell. The values assigned to the node represent the average characteristics of the entire model cell.

The simulated ground-water system includes about the upper 1,000 ft of saturated unconsolidated basin-fill deposits in southern Utah and Goshen Valleys and is vertically discretized into five model layers. Vertical discretization is required to simulate measured vertical gradients within the main ground-water system. The model layers do not represent separate or individual aquifers (see "Occurrence" in the "Ground-water hydrology" section of this report). Model layer thickness is based on the estimated vertical extent of unconfined ground-water conditions and the availability of water-level data.

Model layer 1 is generally about 50 ft thick. This thickness was chosen to correspond with the estimated vertical extent of unconfined conditions in the study area (see "Occurrence" in the "Ground-water hydrology" section of this report). Exceptions occur along and near the mountains where it was necessary to increase the thickness of model layer 1 to as much as 200 ft. This was done to prevent numerical instability when water levels computed during the iterative model-solution process drop, causing cells to go dry in model layer 1.

The thickness of model layers 2, 3, 4, and 5 are 70 ft, 150 ft, 180 ft, and 550 ft, respectively. These thicknesses are determined on the basis of availability of water-level data. Vertical hydraulic gradients exist in the main ground-water system, and measured water levels are a function of both well construction and aquifer properties of the basin-fill deposits near the well. For this analysis, measured water levels are assumed to represent the water level in the basin-fill deposits at the center of the perforated interval of the well. The perforated interval of the well is measured from the top of the saturated unconsolidated basin-fill deposits at the well. A histogram of the number of wells with water-level data with center of perforated interval in each 30-foot depth interval of the saturated unconsolidated basin-fill deposits is shown in figure 22. The model layer boundaries (fig. 22) were chosen to divide these wells into groups with similar depth to center of perforated interval.

The boundary between model layers 2 and 3 was set at 120 ft, and between layers 3 and 4 was set at 270 ft below the top of the saturated unconsolidated basin-fill deposits. The boundary between model layers 4 and 5 was set at a depth of 450 ft below the top of the saturated unconsolidated basin-fill deposits; water-level data is limited for depths below 450 ft. The bottom of the simulated ground-water system was set at 1,000 ft below the top of the saturated unconsolidated basin-fill deposits. Although no water levels are available for wells with a center of perforation below 675 ft (fig. 22), production wells withdraw ground water from depths in excess of 675 ft. Few wells are completed to depths

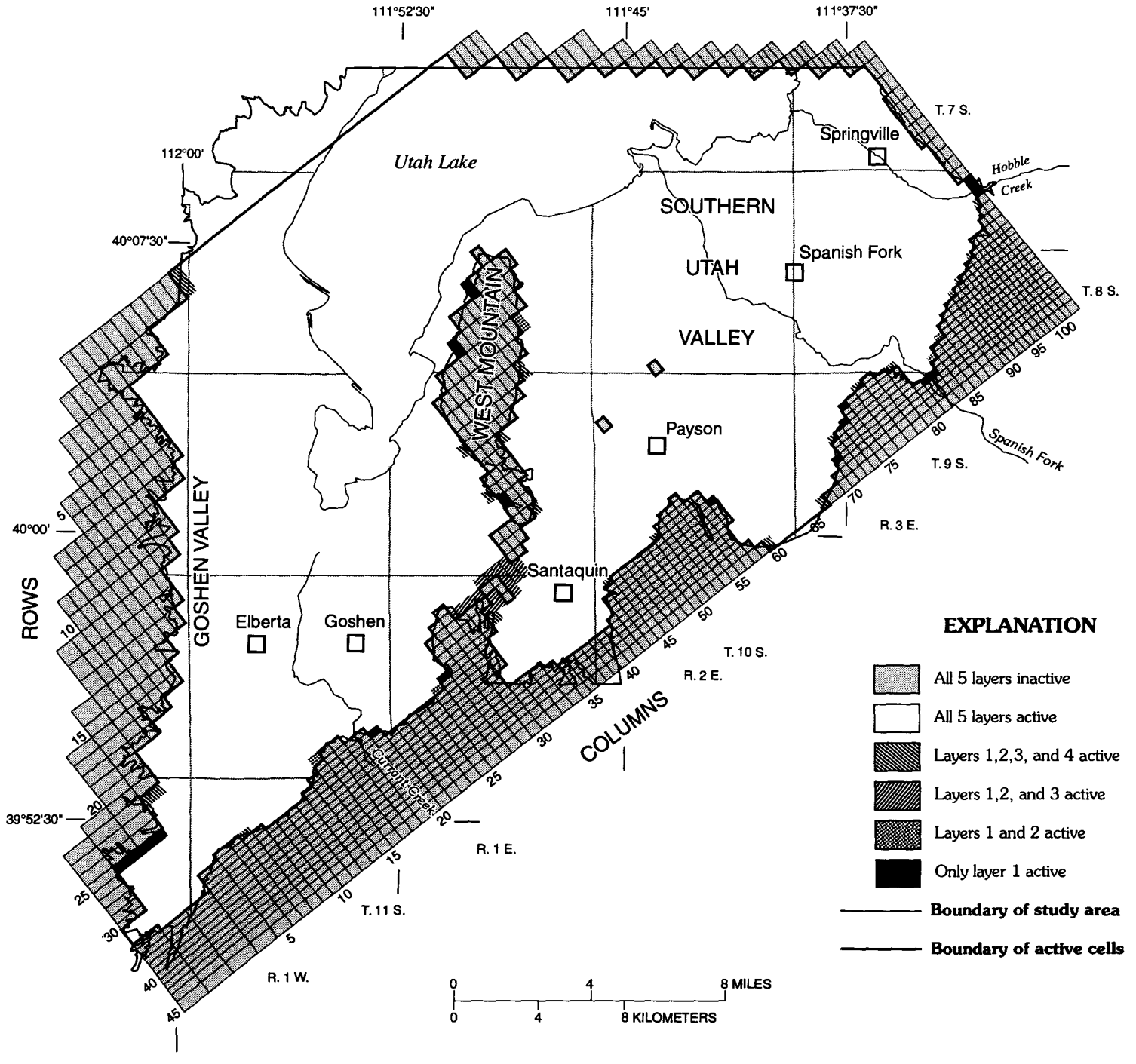


Figure 21. Model grid and location of active cells in layers 1, 2, 3, 4, and 5 in the ground-water flow model of southern Utah and Goshen Valleys, Utah.

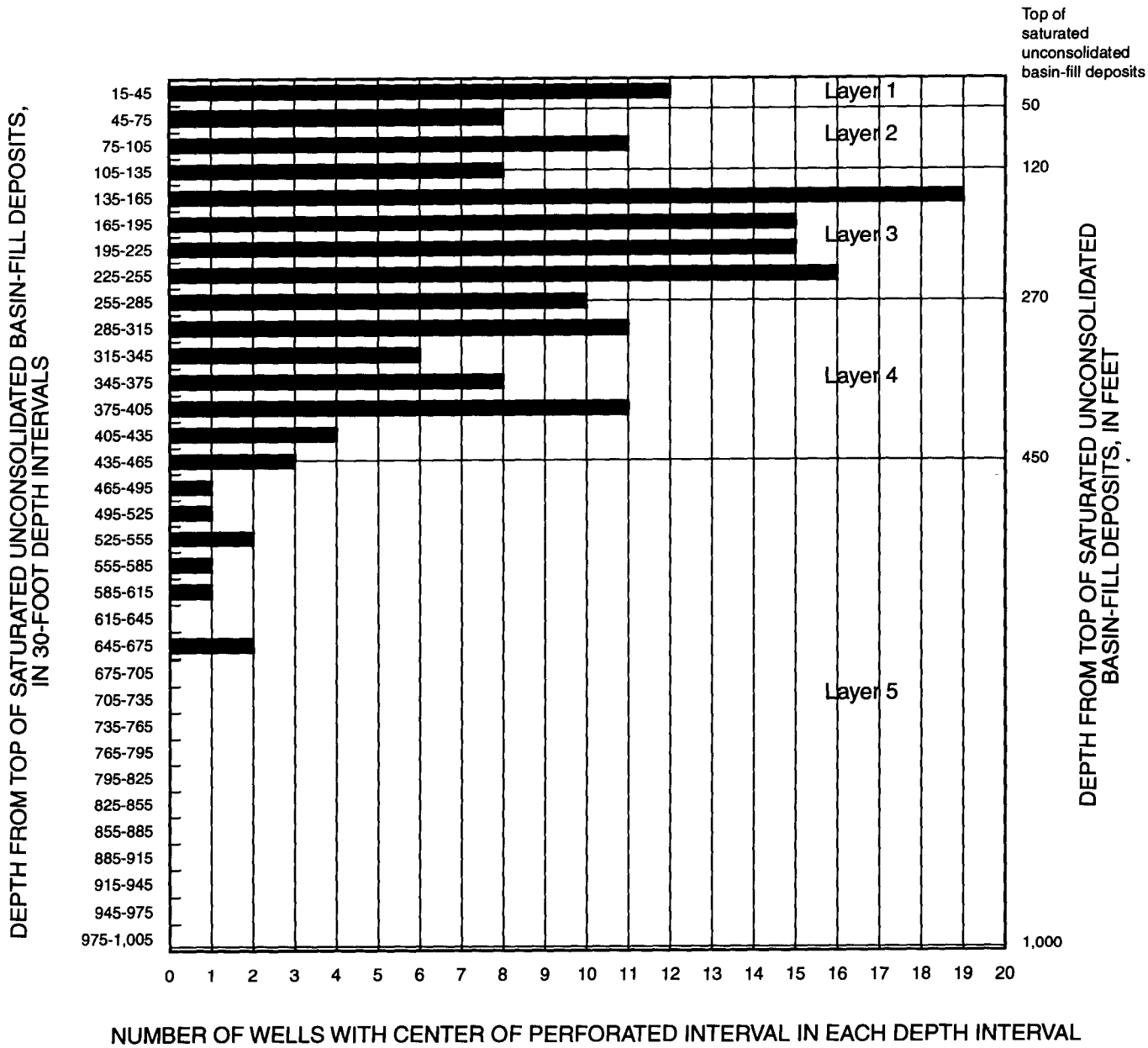


Figure 22. Depth from the top of the saturated unconsolidated basin-fill deposits, number of wells with center of perforated interval in each depth interval, and corresponding model layers in the ground-water flow model of southern Utah and Goshen Valleys, Utah.

greater than 1,000 ft below the top of the saturated unconsolidated basin-fill deposits. The present (1994) conceptualization of the ground-water system does not include basin-fill deposits greater than 1,000 ft deep because no information is available, but it is possible that much deeper drilling in the future could reveal that the active ground-water system extends deeper than 1,000 ft. If the present system conceptualization changes because additional subsurface information becomes available, then the ground-water flow model described in this report may need to be updated and recalibrated to include this potential deeper part of the system before using the model for projecting changes caused by increased pumping.

In areas where the thickness of model layer 1 is greater than 50 ft, the thickness of layers 2 through 5 remains the same, but the altitude of the top of each model layer is shifted downward accordingly. Particularly in Goshen Valley, and to a lesser degree in southern Utah Valley, wells with water-level data also are production wells. Consequently, the model layers indirectly reflect average depth at which ground-water withdrawals occur. A generalized diagram showing the division of the unconsolidated basin-fill deposits into model layers is shown in figure 23.

The saturated unconsolidated basin-fill deposits at Goshen Gap (pl. 1) are less than 1,000 feet thick (see "Movement" in the "Ground-water hydrology" section of this report). The decreased thickness in this area was simulated by deactivating selected cells in model layers 4 and 5 (fig. 21).

Boundary Conditions

Boundary conditions are used to simulate recharge to and discharge from the ground-water system and to define the physical and hydrologic limits of the simulation. Three types of boundaries are used in this simulation: head-dependent flux, specified-flux, and impermeable. At head-dependent flux boundaries, simulated flow across the boundary is proportional to the difference in computed water levels on adjacent sides of the boundary. These boundaries are used to simulate recharge and discharge that is dependent on local ground-water levels. Flow across specified-flux boundaries is assigned by the user and remains constant, regardless of computed water levels. Specified-flux boundaries are used to simulate recharge and discharge that is independent of local ground-water levels. All other model boundaries are impermeable, and water does not move across them. The boundaries chosen for this simulation define mathematically how the modeled ground-water system interacts with the surrounding unmodeled hydrologic system.

Recharge

Recharge to the main ground-water system is simulated from perennial streams and major canals, irrigation and precipitation, intermittent and ephemeral runoff, and subsurface inflow. Both head-dependent flux and specified-flux boundaries are used to simulate this recharge. A specified-flux boundary is used to simulate recharge from a dispersion pond in Goshen Valley. Boundaries are formulated by the Streamflow Package (Prudic, 1989) and the Recharge and Well Packages (McDonald and Harbaugh, 1988, p. 7-1 and p. 8-1).

Recharge from Hobble Creek, Spanish Fork, Currant Creek, South Field Canal, Mill Race Canal, and South Ditch (fig. 24) is formulated as a head-dependent flux boundary using the Streamflow Package (Prudic, 1989). Recharge from these sources (hereafter referred to as streams) is considered head dependent because amount and area of stream loss (recharge to the ground-water system) is a function of stream stage, streambed conductance, and local ground-water level. Summit Creek, Power Canal, Strawberry Highline Canal, Salem Canal, Springville Highline Canal, Mapleton No. 1 Ditch, Swenson Ditch, City Ditch, West Ditch, Currant Creek Canal, and Warm Springs ditch, listed in tables 2 and 5, are located in areas where the local ground-water level is as much as 400 ft below the bottom of the stream or canal. Recharge from these sources is considered to be independent of local ground-water level and is simulated as a specified-flux boundary using the Recharge Package (McDonald and Harbaugh, 1988, p. 7-1). Recharge from East Bench Canal, Fullmer Ditch, and Springville No. 1 Ditch to the Mapleton Bench ground-water system is not modeled.

The Streamflow Package (Prudic, 1989) computes flow across a boundary as a function of computed ground-water level at the center of the model cell and water-surface altitude (stage) of the stream. The package computes water-surface altitude as a function of streamflow and tracks tributary inflow and diversions for individual streams. At boundaries simulated with the Streamflow Package, recharge occurs when model-computed ground-water level drops below the water-surface altitude of the stream.

Required input data for the Streamflow Package include altitude of the top and bottom of the streambed,

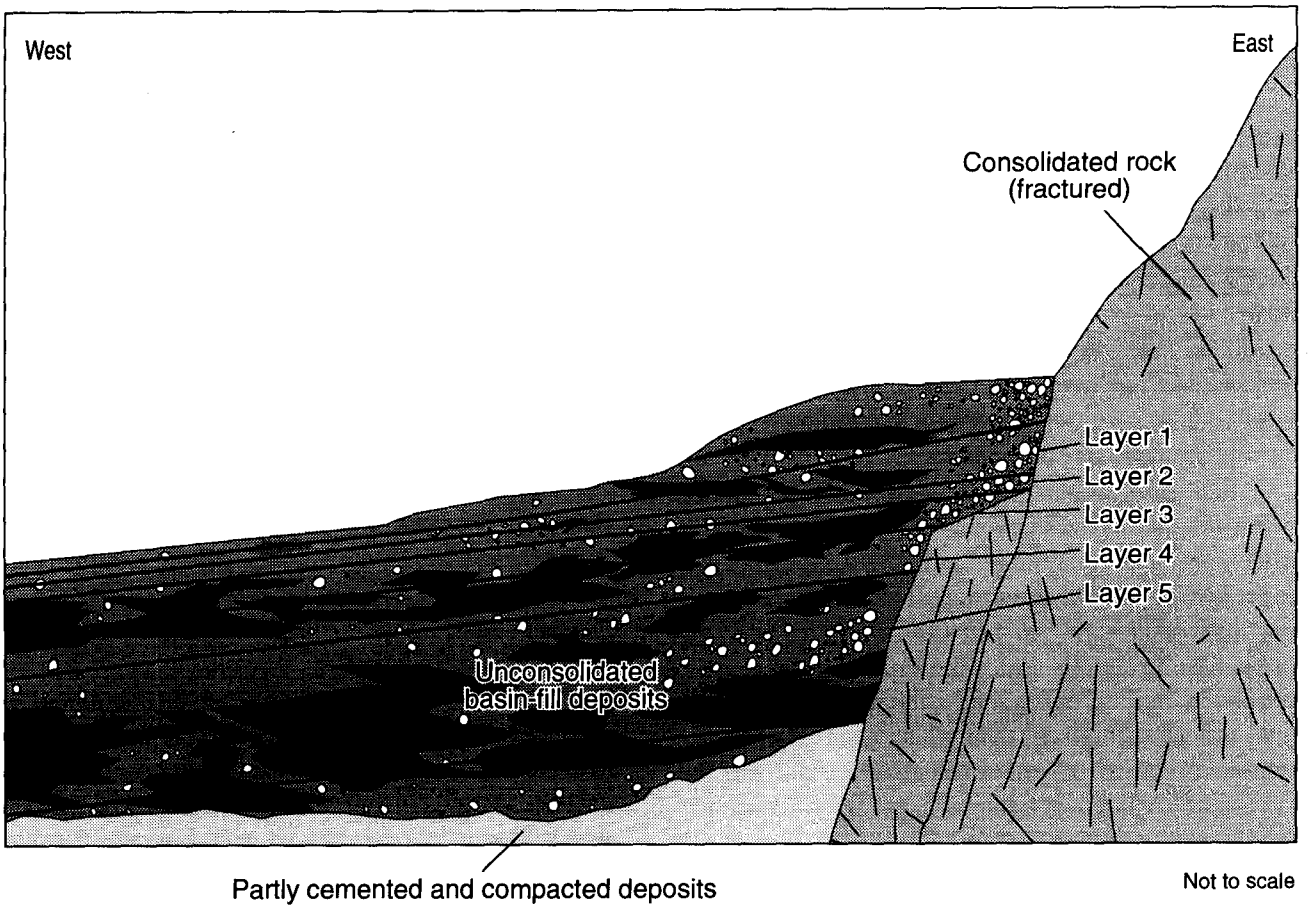


Figure 23. Generalized geology and the five layers used in the ground-water flow model of southern Utah and Goshen Valleys, Utah.

initial altitude of the water surface in the stream or canal, streambed width and slope, Manning's roughness coefficient for the streambed, and hydraulic conductivity of the streambed. Simulated streams were discretized into segments and reaches as shown in Prudic (1989, fig. 1a).

Flow information is required for the simulated streams and their associated tributary inflows and diversions. Average annual flow for Springville No. 1 Ditch, Mapleton No. 1 Ditch, diversions from Hobble Creek at (D-8-3)2cca, and Swenson Ditch is listed in table 2; the average annual flow for Currant Creek below canal diversion is listed in table 5. These average flows are used for all simulated stress periods. Annual flow for 1990 for Spanish Fork at the Power Canal Diversion, Salem Canal, South Field Canal, East Bench Canal, Mill Race Canal, and South Ditch is listed in tables 2 and 5. Flow at these locations for 1949 to 1989 is recorded by the Spanish Fork Water Commissioner (1949-91). Annual flow for 1990 for Hobble Creek at (D-8-4)6abb is listed in table 5. Flow at this location for 1949-89 is estimated by linear regression (see "Perennial streams" in the "Surface-water hydrology" section of this report).

To determine the physical characteristics of the simulated streams, an average stream altitude was determined for each model cell containing a streamflow boundary. Average altitude was determined from the appropriate U.S. Geological Survey 1:24,000-scale topographic maps, which on the basis of contour interval, are considered to be accurate to 10 ft. The altitude of the top of the streambed was set at 5 ft less than average stream altitude to account for stream channelization. The bottom of the streambed was arbitrarily set at 5 ft below the top of the streambed. In the formulation of the Streamflow Package, if ground-water level is below the bottom of the streambed, recharge from the stream ceases to depend on local ground-water level (McDonald and Harbaugh, 1988, p. 6-6). Data are not available to quantify this depth. The initial value of the water surface in the stream was set equal to average stream altitude. On the basis of visual observation, the width of Hobble Creek and Spanish Fork was set at 20 ft; the width of the other streams was set at 10 ft. Stre-

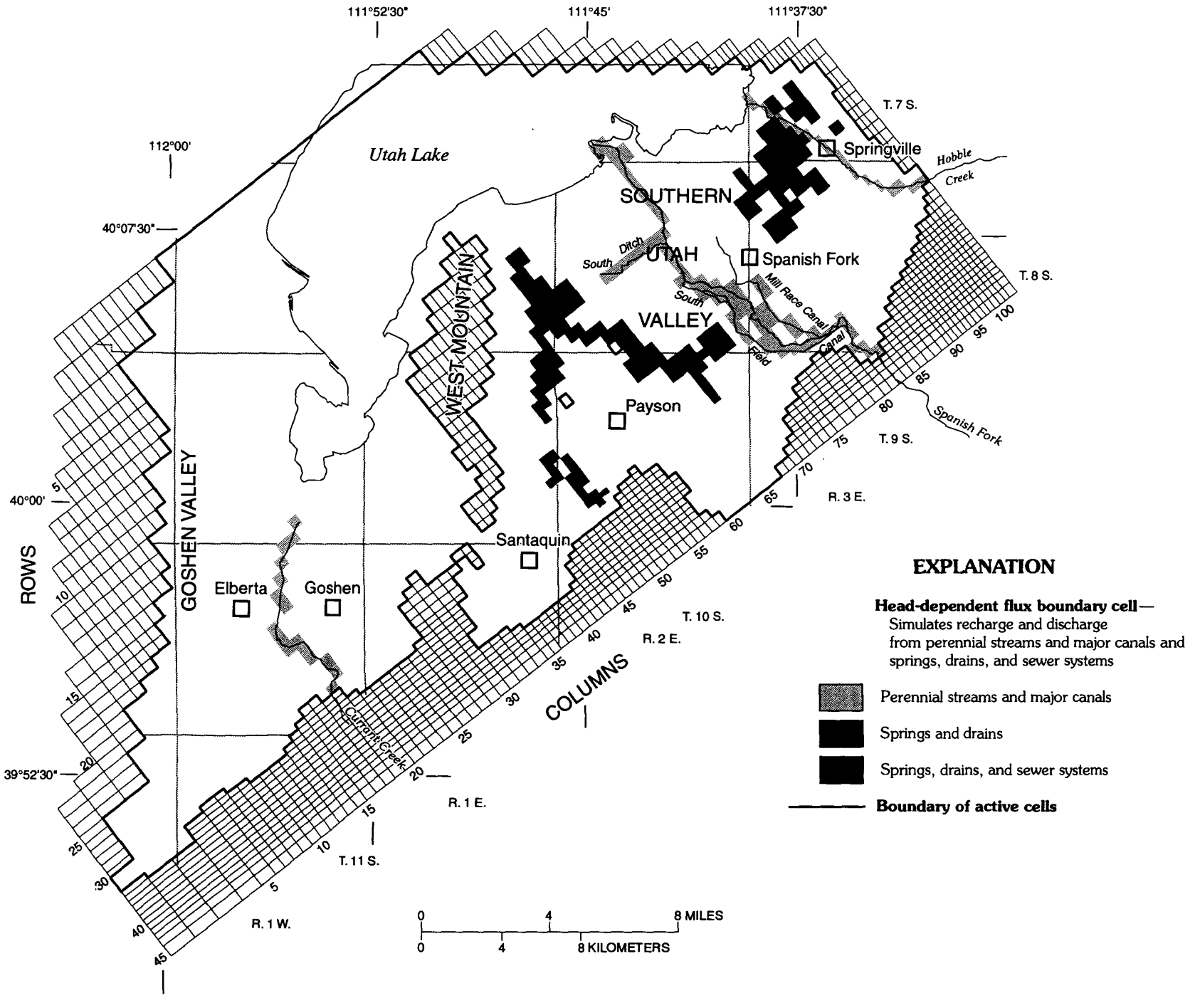


Figure 24. Location of head-dependent flux boundary cells in the ground-water flow model of southern Utah and Goshen Valleys, Utah.

ambed slope for all streams was set at 0.0045, the average land slope between the east shore of Utah Lake and the city of Spanish Fork. Streambed slope does not change drastically in the valley areas and was not calculated for individual streams. Manning's roughness coefficient and hydraulic conductivity of the streambed are calibration parameters and their initial values were set at 0.04 and 100 ft/d, respectively, for all simulated streams. Manning's roughness coefficient is used to compute water-surface altitude from streamflow. Hydraulic conductivity of the streambed defines the measure of interconnection between the stream and the adjacent ground-water system. The final values for these calibration parameters were determined during model calibration.

Recharge from perennial streams and major canals that is simulated as a specified-flux boundary (figs. 25 and 26) is applied to model layer 1 using the Recharge Package (McDonald and Harbaugh, 1988, p. 7-1) at the model cells that represent the area where the stream or canal exists. The amount of recharge simulated from West Ditch and Summit Creek is the 1949-90 average annual recharge varied for each annual stress period on the basis of natural flow in Spanish Fork at Castilla. The recharge varies from 26 percent of average for 1961 to 360 percent of average for 1984. The amount of recharge from major canals applied at specified-flux boundaries is a long-term average determined from individual canal-company records and is not varied for individual stress periods. Examination of canal records indicates that variability in flow and subsequent loss is not significant. Average annual recharge values that differ from the 1990 values listed in tables 2 and 5 for perennial streams and major canals are listed in the following table. Recharge from the Power Canal is set to zero after the 1987 stress period to simulate conditions after the canal was lined with concrete. Recharge from Warm Springs distribution system in Goshen Valley is applied to 30 cells that approximate the area where water is dispersed (fig. 26).

Perennial stream or major canal	Average annual recharge (acre-feet)
West Ditch	1,480
Summit Creek	1,380
Power Canal	4,500
Strawberry Highline Canal	3,770
Salem Canal	920
Currant Creek Canal	590

Recharge from irrigation and precipitation (table 6) is formulated as a specified-flux boundary using the Recharge Package (McDonald and Harbaugh, 1988, p. 7-1). The process of recharge from irrigation and precipitation and how it reaches the ground-water system depends on many hydrologic factors that could not be quantified with the available data. Only the amount of recharge could be estimated, and that is applied as a specified flux. Irrigated areas and nonirrigated areas were assigned to active cells in model layer 1 by using the model grid and figure 8. The recharge amounts for each crop type within each area were applied equally at model cells that correspond to each crop type within the irrigated and nonirrigated areas. Recharge amounts for each stress period were determined explicitly using the method described for 1990 in the "Irrigation and precipitation" of the "Ground-water hydrology" section of this report. Recharge from irrigation and precipitation ranges from about 10,200 acre-ft for the 1960 stress period to about 58,700 acre-ft for the 1982 stress period. Recharge from the Springville City 2, Springville No. 1 Ditch, Mapleton, and East Bench irrigation areas to the Mapleton Bench ground-water system was not modeled because Mapleton Bench was not part of the simulated ground-water system.

Recharge from intermittent and ephemeral runoff (table 3) is considered independent of local ground-water level and occurs in areas where ground-water level is below the ground surface. Recharge from these sources is formulated as a specified-flux boundary (fig. 26) using the Recharge Package (McDonald and Harbaugh, 1988, p. 7-1). Recharge from intermittent and ephemeral runoff is applied to model layer 1 at cells that generally correspond to the coarse-grained alluvial deposits located where the runoff enters the valley. The amount of recharge simulated from intermittent and ephemeral runoff is the average annual recharge (table 3) varied for each annual stress period on the basis of the natural flow in Spanish Fork at Castilla. This is the same method that was used to vary recharge from Summit Creek and West Ditch, as discussed previously.

Recharge from subsurface inflow is formulated as a specified-flux boundary (fig. 25) using the Well Package (McDonald and Harbaugh, 1988, p. 8-1). This formulation assumes that the amount of subsurface inflow is independent of water levels in the unconsolidated basin-fill deposits. To simulate subsurface inflow, injection wells were placed in model layers 1 and 2 along the exterior boundary of the modeled area adjacent to consolidated rock (25 mi of mountain front in southern Utah Valley and 35 mi of mountain front in

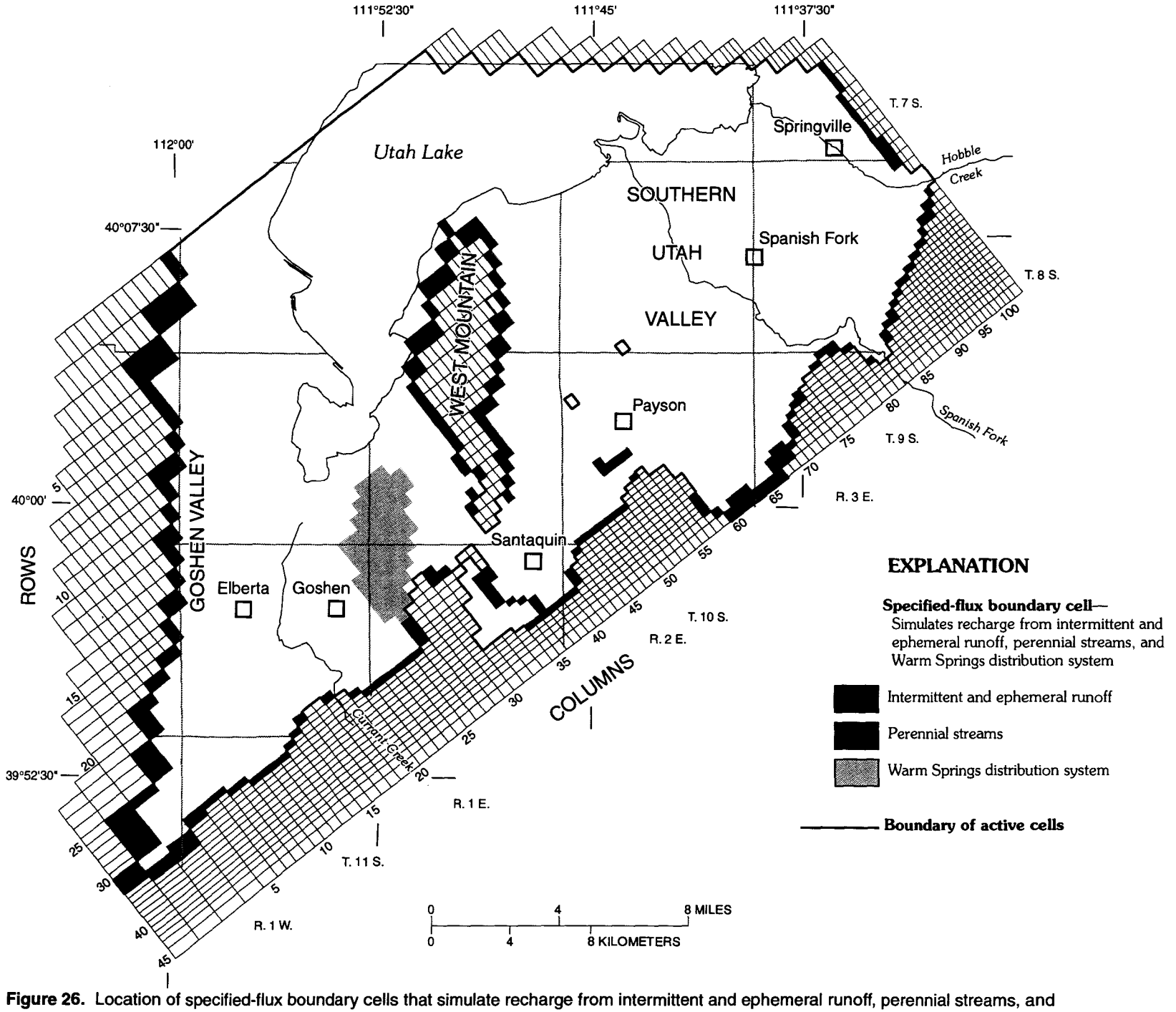


Figure 26. Location of specified-flux boundary cells that simulate recharge from intermittent and ephemeral runoff, perennial streams, and Warm Springs distribution system, in the ground-water flow model of southern Utah and Goshen Valleys, Utah.

Goshen Valley). Recharge at each well was initially set at 200 acre-ft/yr for a total of about 62,000 acre-ft/yr of simulated subsurface inflow. The amount and spatial distribution of subsurface recharge are considered calibration parameters, and the final values were determined during model calibration. Subsurface inflow would be more accurately simulated as a head-dependent flux boundary, but this was not done because data were not available to quantify the hydraulic conductivity of the boundary and the hydraulic gradient across the boundary.

Recharge from dispersion ponds located in Goshen Valley, west of Elberta, occurred from 1969 through 1976 and is simulated as a specified-flux boundary. The boundary is formulated by the Well Package (McDonald and Harbaugh, 1988, p. 8-1). All water diverted to the dispersion ponds was assumed to recharge the ground-water system in Goshen Valley. A single injection well, placed in model layer 1 at the cell corresponding to the location of the dispersion ponds and for the appropriate stress periods, is used to simulate this recharge.

Discharge

Discharge from the main ground-water system is simulated to springs and drains, by evapotranspiration, to wells, to perennial streams and major canals, by seepage to Utah Lake, and by infiltration to sewer systems. Both head-dependent flux and specified-flux boundaries are used to simulate discharge. The boundaries are formulated by the Streamflow Package (Prudic, 1989), Drain Package, Evapotranspiration Package, Well Package, and the Constant-Head Node option (McDonald and Harbaugh, 1988, p. 9-1, 10-1, 8-1, and 4-1) of the flow model.

Discharge to springs and drains is affected by fluctuations in local ground-water level and is therefore formulated as a head-dependent flux boundary using the Drain Package (McDonald and Harbaugh, 1988, p. 9-1). The Drain Package computes flow across the boundary as a function of the computed ground-water level at the boundary, the hydraulic conductance of the orifice, and the altitude of the drain. When the computed ground-water level at the boundary drops below the drain altitude, no flow is allowed across the boundary. Model cells that contain boundaries formulated by the Drain Package are shown in figure 24; these model cells correspond to the areas listed in table 7. Altitude of the drain and drain conductance, which are required input for the Drain Package, are calibration parameters and their final values were determined during model calibration. The only data available to describe drain areas are estimates of discharge and general location. Initial estimates of drain altitude and conductance were arbitrarily chosen at 10 ft below land surface and 10,000 ft^2/d, respectively.

Discharge by evapotranspiration is formulated as a head-dependent flux boundary using the Evapotranspiration Package (McDonald and Harbaugh, 1988, p. 10-1). Discharge by evapotranspiration depends on local ground-water level. The Evapotranspiration Package computes flow across the boundary as a function that is maximum when the computed water level is at or above the evapotranspiration surface and ceases when the computed water level is below the extinction depth. Between the two extremes, evapotranspiration varies linearly as a function of the ground-water level in the model cell. Discharge boundaries simulated with the Evapotranspiration Package occur in model layer 1 at cells that correspond with the evapotranspiration areas shown in figure 9.

Input parameters for the Evapotranspiration Package are the altitude of the evapotranspiration surface, extinction depth, and maximum evapotranspiration rate. For this simulation, the evapotranspiration surface was assumed to be land surface. Land-surface altitude was determined for the center of the appropriate model cells from U.S. Geological Survey 1:24,000-scale topographic maps. The extinction depth for alfalfa and pasture was set at 7.5 ft; salt grass was set at 8 ft; and tamarisk was set at 15 ft. These depths are generalizations and are based on data presented in Robinson (1958, p. 14 and p. 22). The maximum evapotranspiration rate for alfalfa and pasture was set at 2.32 ft/yr. This is the consumptive use for alfalfa and pasture determined using the Blaney-Criddle method (Soil Conservation Service, 1970) for 1964. The maximum evapotranspiration rate for saltgrass was set at 4.0 ft/yr and for tamarisk was set at 6.0 ft/yr (table 8). The extinction depth and maximum evapotranspiration rates were assigned to model cells that correspond with the phreatophyte types shown in figure 9. If more than one type of phreatophyte exists within a single model cell, an area weighted average for extinction depth and maximum evapotranspiration rate was determined for the cell. A single multiplier, which scales the maximum evapotranspiration rate for all the phreatophytes listed above and thereby affects the total model-computed discharge to evapotranspiration, is a calibration parameter. The initial value of the multiplier is 1.0 and

its final value is discussed in "Model calibration" of the "Simulation" section of this report.

Discharge to pumping wells is formulated as a specified-flux boundary using the Well Package (McDonald and Harbaugh, 1988, p. 8-1). Wells that discharge 50 acre-ft/yr (30 gal/min) or more are represented in the model. The specified-head formulation is appropriate because pumping-well discharge is controlled and is not directly a function of local ground-water conditions. Required input data for the Well Package are the model cell and layers that correspond to the location of the well and the depth at which ground-water discharge to the well occurs. The amount of discharge assigned to each model layer is based on the percentage of the perforated interval that occurs within the specific model layer. The perforated intervals are measured from the estimated top of the saturated unconsolidated basin-fill deposits at the well. Perforated interval for this purpose is defined as the length from the top of the first perforations to the bottom of the last perforations. Well-discharge data for each simulated stress period were compiled and estimated from unpublished data in the files of the U.S. Geological Survey District office in Salt Lake City, Utah.

Discharge to flowing wells is formulated as a head-dependent flux boundary using the Drain Package (McDonald and Harbaugh, 1988, p. 9-1). Flowing-well discharge is considered head dependent because it is a function of local ground-water level. For each model cell that contains one or more flowing wells, a drain boundary was placed in the model layers that correspond to the perforated intervals of the well or wells. The perforated interval of a flowing well is defined in the same way as for a pumping well. The drain altitude assigned to flowing wells is equal to the altitude of the discharge orifice of the well; this causes computed discharge from flowing wells to drop to zero when model-computed ground-water levels at drain boundaries are at or below the discharge point. Drain conductance, which is required data input, is considered a calibration parameter, and the final values were determined during model calibration. Initially, drain conductance was arbitrarily set at 10,000 ft^2/d. As with pumping wells, only flowing wells that discharge 50 acre-ft/yr (30 gal/min) or more are represented in the model.

Discharge to Hobble Creek, Spanish Fork, Currant Creek, South Field Canal, Mill Race Canal, and South Ditch (fig. 24) is formulated as a head-dependent flux boundary using the Streamflow Package (Prudic, 1989). The Streamflow Package and data requirements were discussed previously. Discharge at streamflow boundaries occurs when model-computed ground-water levels at the boundaries are greater than the water-surface altitude of the stream. The Manning's roughness coefficient was initially set at 0.04 for all streams and canals, and an initial value of 100 ft/d was used for hydraulic conductivity of the streambed. Final values for these calibration parameters were determined during model calibration.

Discharge by seepage to Utah Lake is simulated as a head-dependent flux boundary. The boundary is formulated by the Constant-Head Node option from McDonald and Harbaugh (1988, p. 4-12). Model cells where the Constant-Head Node option is used correspond to the areal extent of Utah Lake (pl. 1) at an altitude of about 4,488 ft. The altitude assigned to the constant-head nodes is equal to the water level of Utah Lake on March 15 of the annual stress period being simulated (fig. 14).

Separate model boundaries are not formulated to simulate ground-water infiltration to sewer systems (table 12). Infiltration to sewer systems is formulated with the same head-dependent flux boundaries used to represent some of the springs and drains listed in table 7 (fig. 24). The simulated sewer systems are located near these springs and drains, and the scale of horizontal discretization is not adequate to construct separate boundaries for each. Although sewer systems are not drains, ground-water infiltration to them can be formulated the same way it is for drains. Infiltration to the Spanish Fork sewer system is simulated with drain boundaries that represent the southern parts of the Dry Creek area, infiltration to the Springville sewer system is simulated with drain boundaries that represent Wood and Matson Springs and the northern parts of the Dry Creek area, and infiltration to the Payson City and Salem sewer systems is simulated by drain boundaries that represent the Beer Creek area and Salem Lake.

Aquifer Properties

Model parameters that simulate the aquifer properties of the saturated unconsolidated basin-fill deposits are horizontal hydraulic conductivity, transmissivity, vertical leakance, primary storage coefficient, and secondary storage coefficient. Transmissivity and vertical leakance incorporate model-layer thickness and hydraulic conductivity into a single term. Storage coefficient, depending on the model-layer specifications, is either equated to specific yield or to the product of spe-

cific storage and model-layer thickness. The parameters listed in this paragraph, along with the boundary conditions, determine the amount and pattern of computed ground-water flow between nodes of the ground-water flow model.

Horizontal hydraulic conductivity and transmissivity are the two model parameters used to simulate horizontal ground-water flow in the saturated unconsolidated basin-fill deposits. A horizontal hydraulic-conductivity value is required for all active nodes in model layer 1 and was initially set at 200 ft/d near the mouth of Hobble Creek and Spanish Fork, and 10 ft/d in the remaining modeled area. The higher initial estimate of hydraulic conductivity at the mouth of Hobble Creek and Spanish Fork reflects the generally coarse-grained and well-sorted unconsolidated basin-fill deposits found at canyon mouths. Model layer 1 simulates unconfined ground-water flow using option 1 (McDonald and Harbaugh, 1988, p. 5-38). Transmissivity, the product of horizontal hydraulic conductivity and model-layer thickness, is used to compute confined horizontal ground-water flow and is a required model input for layers 2, 3, 4, and 5. These model layers simulate confined horizontal ground-water flow using option 2 (McDonald and Harbaugh, 1988, p. 5-38). Option 2 allows the ground-water storage term to be converted from confined to unconfined when the water level in the model cell drops below the top of the layer. The initial values for transmissivity for each of model layers 2, 3, 4, and 5 is the product of the initial horizontal hydraulic-conductivity value assigned to model layer 1 multiplied by the layer thickness. The final distribution of values for horizontal hydraulic conductivity and transmissivity were determined during model calibration.

Vertical leakance is the model parameter used to simulate ground-water movement between model layers. Vertical leakance is the conductance term for vertical ground-water flow, determined from the vertical hydraulic conductivity and thickness of all geohydrologic layers that exist between vertically adjacent model nodes, times the model-cell area (McDonald and Harbaugh, 1988, p. 5-11). For this simulation, the unconsolidated saturated basin-fill deposits are treated as a single geohydrologic unit; vertical leakance is then a single vertical hydraulic-conductivity value divided by the total vertical distance between nodes of adjacent model layers. Vertical leakance is required to describe vertical flow between model layers 1 and 2, 2 and 3, 3 and 4, and 4 and 5. Initially, all vertical-leakance values were calculated using a vertical hydraulic-conductivity value of 0.1 ft/d and the appropriate vertical distance between model nodes. The final distribution of values for vertical leakance was determined during model calibration.

Primary and secondary storage coefficients are model parameters used to simulate the amount of water released from and placed into ground-water storage. Primary storage coefficient for model layer 1 is equivalent to specific yield. For model layers 2, 3, 4, and 5, the primary storage coefficient is equivalent to storage coefficient; storage coefficient is the product of model-layer thickness and specific storage. In addition, a secondary storage coefficient is required for model layers 2 through 5. The secondary storage coefficient for these model layers is equivalent to specific yield and is used only when the water level in a model cell drops below the top of the cell. Initially, the primary storage coefficient for model layer 1 was assigned a value of 0.20. The initial value for the primary storage coefficient for each of model layers 2 through 5 was determined by assigning a specific storage of 1×10^{-6}/ft to all nodes and multiplying by the appropriate model-layer thickness. The initial value of secondary storage coefficient for model layers 2 through 5 was assigned a value of 0.20 for all nodes in those layers. The initial values were set arbitrarily, and the final distribution of values for primary and secondary storage coefficients were determined during model calibration.

Model Calibration

The ground-water flow model constructed to simulate the main ground-water system in southern Utah and Goshen Valleys was calibrated to the steady-state conditions of 1949 and annual transient-state conditions that occurred from 1949 through 1990. The 1949 steady-state simulation was used to establish the initial conditions for the 1949-90 annual transient-state simulation. During calibration, special emphasis was placed on the 1964 and 1990 annual stress periods because more data describing the ground-water system are available for these two periods.

Calibration Parameters

The calibration parameters adjusted during the calibration process, listed in general order of importance, are horizontal hydraulic conductivity, amount and spatial distribution of recharge from injection wells to simulate subsurface inflow, vertical hydraulic conductivity, drain conductance and altitude, maximum

evapotranspiration rate, hydraulic conductivity of the streambed, specific storage, specific yield, and Manning's roughness coefficient of the streambed. Generally, horizontal hydraulic conductivity and the amount and location of recharge from subsurface inflow were adjusted to attain a reasonable match between measured and computed water level. Vertical hydraulic conductivity was adjusted to match measured vertical water-level differences and estimated discharge to Utah Lake and major streams. Drain conductance and altitude, and streambed conductivity, were adjusted to best simulate measured discharge and water level in localized areas. The maximum evapotranspiration rate was used to match estimated discharge by evapotranspiration and indirectly, water levels in the shallow unconfined ground water. Specific storage and specific yield were adjusted to best simulate measured water-level fluctuations through time. Manning's roughness coefficient was adjusted to simulate estimated changes in streamflow loss and gain caused by change in streamflow.

Limitations were established for each calibration parameter on the basis of available data and the conceptual understanding of the main ground-water system. Limitations represent physically and hydrologically reasonable estimates of parameters and the degree of uncertainty associated with each parameter. Limitations, final values, and distribution of the calibration parameters are discussed in the following pages.

Horizontal hydraulic-conductivity values were varied from 1 ft/d to 200 ft/d during model calibration and were not varied within the vertical column. This corresponds to transmissivity values ranging from about 1,000 ft^2/d to 200,000 ft^2/d, which is a slightly larger range than that presented in the "Aquifer properties" of the "Ground-water hydrology" section of this report. The final distribution of values for horizontal hydraulic conductivity ranges from 1 ft/d to 80 ft/d and is shown on plate 2. The final distribution was determined almost exclusively by adjusting values until a reasonable match between measured and computed water levels was obtained.

Horizontal hydraulic-conductivity values from 1 ft/d to 40 ft/d were assigned to the model nodes that simulate the ground-water system near the mountains. An exception to this is the area near the mouth of Hobble Creek, where horizontal hydraulic-conductivity values were set at 80 ft/d. The relatively lower hydraulic conductivity values assigned to areas near the mountains, where the unconsolidated basin-fill deposits are generally coarse grained, may reflect poor sorting of those deposits. Poor sorting can decrease the horizontal hydraulic conductivity of coarse-grained material. Horizontal hydraulic-conductivity values from 20 ft/d to 80 ft/d were assigned to the central parts of southern Utah and Goshen Valleys. In these areas, the unconsolidated basin-fill deposits are generally finer grained and better sorted than near the mountains. Sorting can increase horizontal hydraulic-conductivity values. Horizontal hydraulic-conductivity values assigned to model nodes that simulate areas near and underneath Utah Lake were set at 5 ft/d, 10 ft/d, and 20 ft/d. The basin-fill deposits near Utah Lake are generally very fine grained. Fine-grained material generally has low horizontal hydraulic conductivity. Horizontal hydraulic-conductivity values assigned to model cells along the lower sections of the Hobble Creek and Spanish Fork stream channels are lower than those assigned to surrounding cells. The lower horizontal hydraulic-conductivity values were assigned to create a steeper hydraulic gradient toward the streams so that computed discharge to the streams would more accurately match measured discharge. Horizontal hydraulic-conductivity values east of Springville were adjusted so that simulated discharge to the Spring Creek area would match measured discharge (table 7).

Transmissivity values estimated from available field data (tables 13 and 14) and determined from model calibration are listed in table 16. The average of estimated transmissivity values in southern Utah Valley and the average of corresponding values determined from the model calibration are almost identical. In Goshen Valley, the average of the transmissivity values determined from model calibration is higher than the average of the estimated values. In Goshen Valley, horizontal hydraulic-conductivity values were adjusted upward during model calibration to decrease the computed water-level declines caused by simulated ground-water withdrawals. This may have resulted in overestimating the horizontal hydraulic-conductivity values.

Recharge from injection wells to simulate subsurface inflow was varied from 36,400 acre-ft/yr to 79,000 acre-ft/yr during model calibration. This is the range of subsurface inflow estimated for 1949, 1964, and 1990 (table 17).

The final amount of subsurface inflow determined during model calibration is 62,200 acre-ft/yr. About 80 percent of the simulated subsurface inflow is distributed in four general areas. In the area east of Springville, simulated subsurface inflow is about

Table 16. Comparison of transmissivity determined from aquifer tests and from lithology reported on drillers' logs, with transmissivity determined from model calibration, southern Utah and Goshen Valleys, Utah

[See figure 2 for explanation of numbering system used for hydrologic-data sites]

Location of well	Model Row	Model Column	Transmissivity determined from aquifer tests and lithology (cubic feet per day per foot)	Transmissivity determined from model calibration (cubic feet per day per foot)
			Southern Utah Valley	
(D-7-2)32dad-1	10	76	5,200	10,000
(D-7-3)28cab-1	15	96	23,500	10,000
(D-7-3)34cdb-1	20	97	42,000	10,000
(D-8-2)17ada-1	14	70	9,900	80,000
(D-8-2)25dac-3	34	76	15,000	6,000
(D-8-3)11ccc-1	26	93	71,000	80,000
[1] (D-8-3)22cbd-3	29	86	14,000	44,000
(D-8-3)27bad-2	31	85	7,000	5,500
(D-8-3)27cdc-1	34	83	12,100	1,100
(D-8-3)28abc-1	29	82	29,000	44,000
[2] (D-9-1)35abb-1	26	41	44,000	21,000
(D-9-2) 9bac-1	24	61	31,000	20,000
(D-9-2)24aca-1	40	65	2,100	1,200
(D-9-2)26add-1	42	61	8,000	1,200
(D-9-3) 5bbd-1	32	76	14,800	11,000
(D-9-3)18bbb-1	36	69	48,400	22,000
Average transmissivity			**23,600**	**23,000**
			Goshen Valley	
(C-9-1) 4ddc-1	8	30	7,500	40,000
[3] (C-9-1)20cdd-1	10	23	4,800	10,500
(C-9-1)20ddd-1	11	23	14,300	21,000
(C-9-1)29acc-1	11	22	35,000	10,000
(C-9-1)29bcc-1	11	21	6,800	10,500
(C-10-1)4cbb-1	15	19	5,100	80,000
(C-10-1)31cdd-1	24	7	44,000	42,000
Average transmissivity			**16,800**	**30,600**

[1] Previously reported as well (D-8-3)22cac-1 (Cordova, 1969 and 1970).
[2] Previously reported as well (D-9-1)35aba-1 (Cordova, 1969 and 1970).
[3] Previously reported as well (C-9-1)20dcc-1 (Cordova, 1969 and 1970).

Table 17. Estimated and model-computed ground-water budget, southern Utah and Goshen Valleys, Utah
[NA, not applicable]

Budget element	Flow, in acre-feet per year (rounded)						
	Estimated 1949 ground-water budget	Model-computed 1949 steady state	Estimated 1964 ground-water budget	Model-computed 1964 transient state	Model-computed 1983 transient state	Estimated 1990 ground-water budget	Model-computed 1990 transient state
Recharge							
Perennial streams and major canals							
Specified flux[1]	NA	20,200	NA	19,600	25,400	NA	14,700
Head-dependent flux[2]	NA	13,200	NA	14,900	12,700	NA	16,600
Subtotal	43,300	33,400	36,700	34,500	38,100	41,500	31,300
Irrigation and precipitation[1]	25,600	25,600	20,000	20,000	38,400	15,300	15,200
Intermittent and ephemeral runoff[1]	11,600	11,600	8,700	8,700	37,300	6,800	6,800
Subsurface inflow[1]	36,400	62,100	74,700	62,200	62,200	79,000	62,200
Total recharge (rounded)[3]	**117,000**	**133,000**	**140,000**	**125,000**	**176,000**	**143,000**	**116,000**
Discharge							
Springs and drains[2]	42,700	49,100	42,700	40,200	55,800	42,700	41,800
Evapotranspiration[2]	40,000	53,200	40,000	45,000	52,300	40,000	44,700
Pumped wells[1,4]	400	400	18,500	18,600	11,500	27,500	27,500
Flowing wells[2,4]	1,200	1,300	2,500	2,100	4,000	4,400	2,700
Perennial streams and major canals[2]	14,400	9,700	15,900	6,300	9,600	22,900	5,100
Utah Lake[2]	13,200	11,400	13,200	10,700	6,900	13,200	9,600
Sewer systems[2]	5,000	7,400	5,000	6,000	8,400	5,000	5,900
Total discharge (rounded)[3]	**117,000**	**132,000**	**133,000**	**129,000**	**149,000**	**156,000**	**137,000**
Water going into (+) or out of (-) storage[5] (rounded)	0	0	+6,800	-3,500	+27,300	-13,200	-21,100

[1]The amount of flow is specified and not computed by the model.
[2]The amount of flow is computed by the model during the simulation.
[3]Model-computed total recharge may not equal model-computed total discharge because of simulation constraints and rounding.
[4]Estimated and simulated amounts may differ because only discharges of 50 acre-feet per year or more were simulated.
[5]Water going into (+) storage is considered to be discharge and should be added to total discharge; water going out of (-) storage is considered to be recharge and should be added to total recharge.

16,300 acre-ft/yr. This amount is required to simulate measured and estimated discharge in the Spring Creek area (table 7). The consolidated rock east of Springville is mainly carbonate and could be a large source of subsurface inflow. In the area directly north of the mouth of Spanish Fork Canyon, about 7,900 acre-ft/yr of subsurface inflow is simulated. Subsurface inflow here was determined by matching measured and computed water level near the mountains. About 11,400 acre-ft/yr for subsurface inflow is simulated to occur from the area south of Salem. This amount is required to match the local water level and simulate measured discharge to Salem Lake and areas of Beer Creek (table 7). Simulated subsurface inflow from the area between Payson and Santaquin is about 13,500 acre-ft/yr. This amount is required to simulate measured and estimated discharge to Spring Lake, Spring Creek, Tanner Reservoir area, and areas of Benjamin Slough.

Average vertical hydraulic-conductivity values were varied from 0.001 ft/d to 2 ft/d during model calibration. This represents a maximum possible horizontal to vertical conductivity ratio of 1,000:1. Data are not available to determine ratios for the modeled area, but

reported ratios of horizontal to vertical conductivity in heterogeneous, anisotropic material can be on the order of 100:1 or larger (Freeze and Cherry, 1979, p. 34). The final distribution of values for vertical hydraulic conductivity ranges from 0.001 ft/d to 1 ft/d and is shown in figures 27 and 28. Vertical hydraulic-conductivity values of 0.1 ft/d and 1 ft/d were assigned near the mountains. These values resulted in the best match between measured and computed water levels at various depths in the unconsolidated basin-fill deposits near the mountains. To match measured water levels, a vertical hydraulic-conductivity value of 0.01 ft/d was assigned to the central parts of southern Utah and Goshen Valleys, except along the lower sections of Hobble Creek, Spanish Fork, and Currant Creek, where the vertical hydraulic-conductivity value between model layers 1 and 2 is maintained at 0.1 ft/d (fig. 27). The value along the streams was assigned to more accurately simulate measured discharge to the streams. The vertical hydraulic-conductivity value underneath Utah Lake was set at 0.001 ft/d to simulate estimated discharge to Utah Lake.

Drain conductance was varied from 100 ft^2/d through 100,000 ft^2/d. No data are available to estimate the hydraulic conductivity of the interface between the drain and the basin-fill deposits or the physical dimensions of specific drains, and the range was arbitrarily chosen. To simulate measured and estimated discharge to springs and drains, the final values for drain conductance, determined during model calibration, range from 275 ft^2/d to 55,000 ft^2/d. At drain boundaries used to simulate flowing wells, the final value of drain conductance is 30,000 ft^2/d. The value of drain conductance used for flowing wells was determined by comparing the total measured flowing-well discharge with the total model-computed discharge. Separate drain conductances were not determined for specific flowing wells. The smallest conductance values occur at drains used to simulate discharge to Benjamin Slough. The largest conductance values occur at drains used to simulate discharge to Salem Lake and the Spring Creek area near Springville.

Drain altitude was varied from 5 to 15 ft below land surface. No data are available to estimate drain altitude directly, and this range was chosen on the basis of physical limitations of drain placement. Final drain altitudes range from 8 to 15 ft below land surface and were assigned to simulate estimated and measured discharge. At drain boundaries used to simulate flowing wells, drain altitude is set at land surface and was not used as a calibration parameter.

Maximum evapotranspiration rate was varied from 50 percent to 150 percent of the maximum evapotranspiration rate listed in table 8 during model calibration. This amount of variation reflects the degree of uncertainty associated with estimates of maximum evapotranspiration rates. The final value of the maximum evapotranspiration rate determined during model calibration is 75 percent of the maximum rate. The maximum evapotranspiration rate was adjusted to simulate the average annual evapotranspiration (table 8).

Hydraulic conductivity of streambeds in southern Utah and Goshen Valleys was varied from 1 ft/d to 1,000 ft/d during model calibration. Data are not available to directly determine hydraulic conductivity, and the probable range is based on average values for streambed material composed of silty sand to clean sand (Freeze and Cherry, 1979, p. 29). The final range of hydraulic conductivity of the streambeds is 1 ft/d to 20 ft/d. Streambed conductivity values were determined for specific stream reaches to simulate estimated and measured flow. Hydraulic conductivity values assigned to the stream reaches are listed in the following table.

Stream	Streambed conductivity
Hobble Creek between model boundary and Mapleton Lateral	7 ft/d
Hobble Creek between Mapleton Lateral and Swenson Ditch	5 ft/d
Hobble Creek between Swenson Ditch and Utah Lake	10 ft/d
Spanish Fork between model boundary and Mapleton Lateral	5 ft/d
Spanish Fork between Mapleton Lateral and U.S. Highway 91	5 ft/d
Spanish Fork between U.S. Highway 91 and Interstate Highway 15	5 ft/d
Spanish Fork between Interstate Highway 15 and Palmyra	10 ft/d
Spanish Fork between Palmyra and Utah Lake	20 ft/d
Mill Race Canal	1 ft/d
South Ditch	1 ft/d
Currant Creek between model boundary and Goshen Reservoir	2 ft/d
Currant Creek between Goshen Reservoir and Utah Lake	5 ft/d

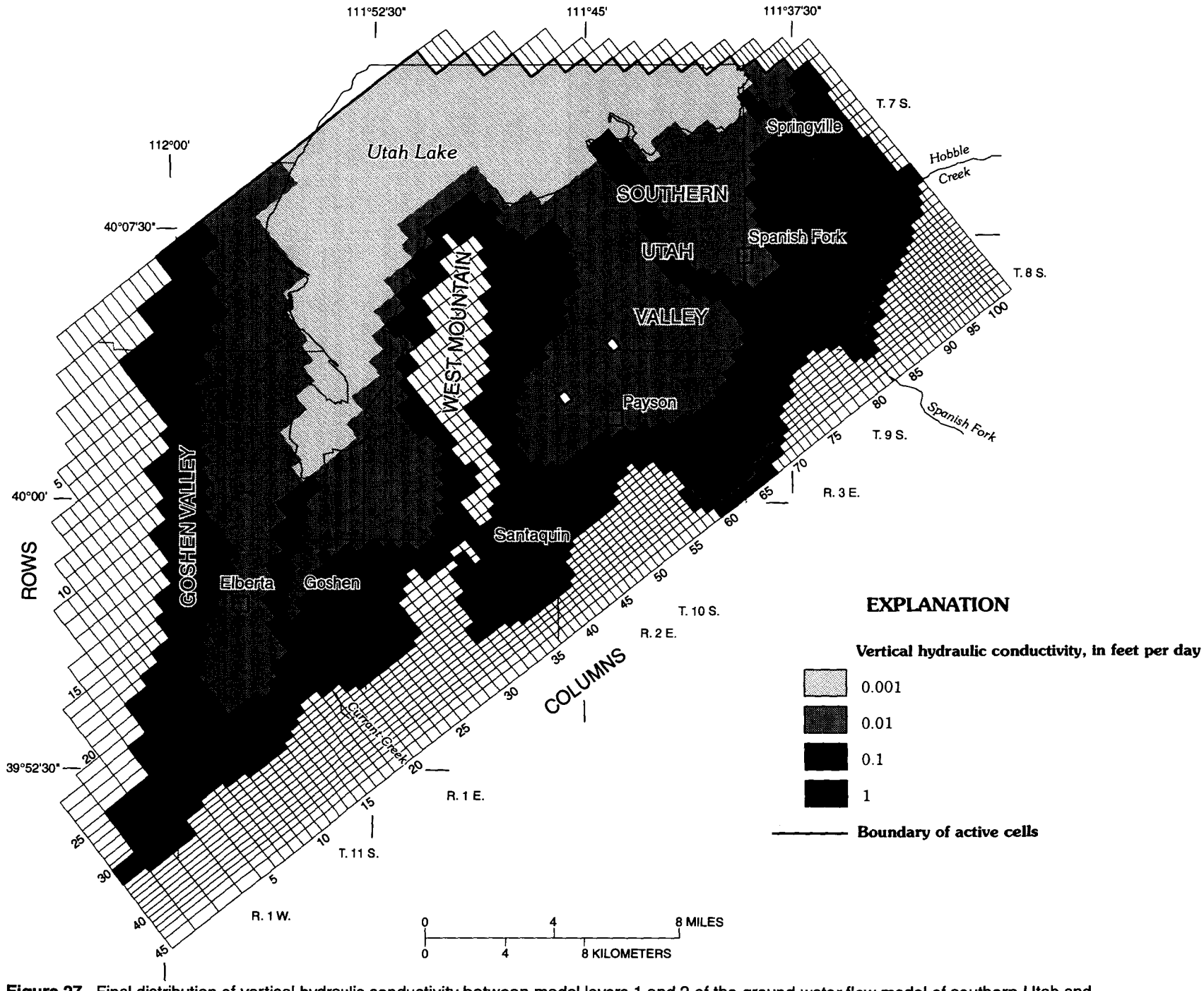


Figure 27. Final distribution of vertical hydraulic conductivity between model layers 1 and 2 of the ground-water flow model of southern Utah and Goshen Valleys, Utah.

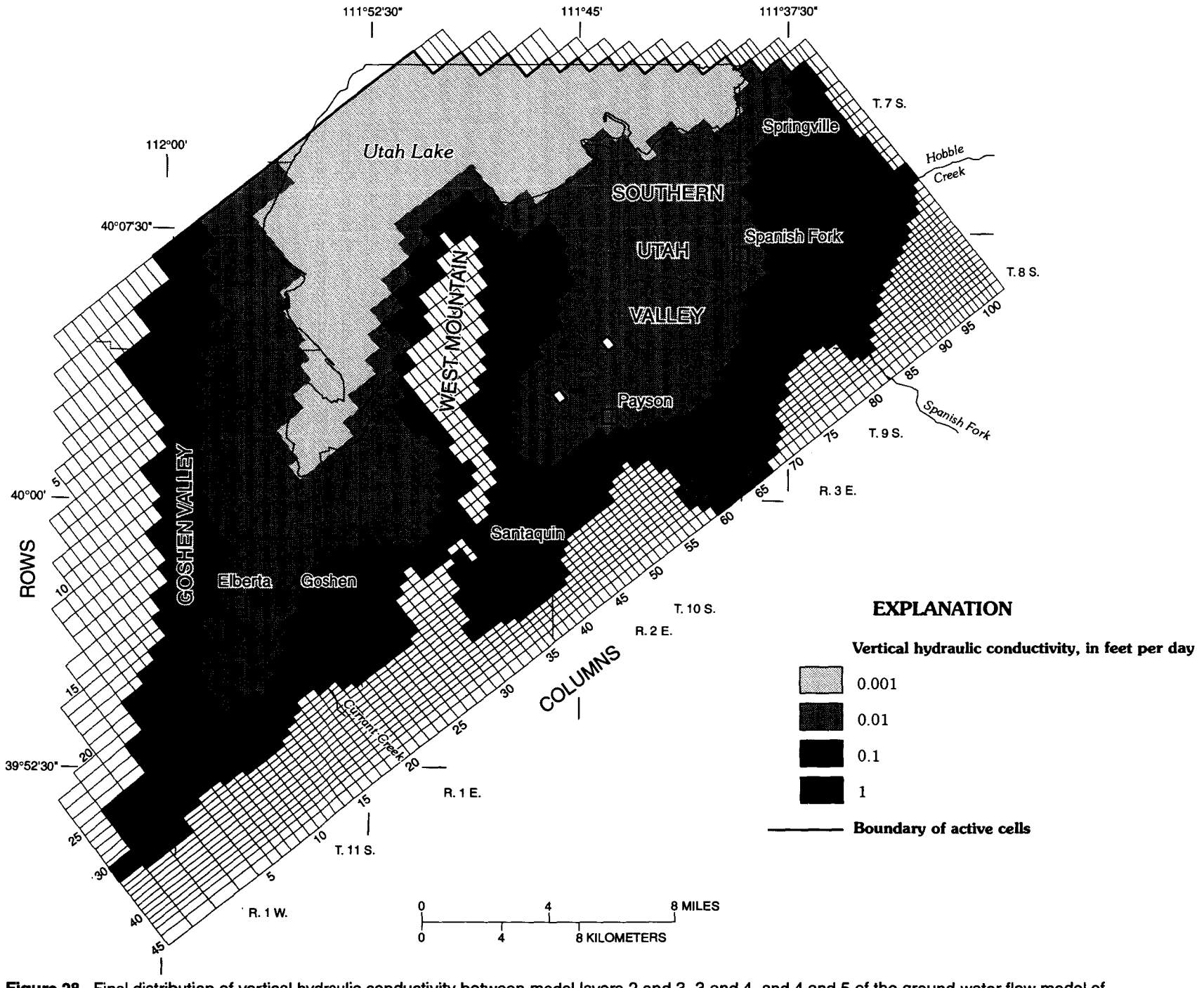


Figure 28. Final distribution of vertical hydraulic conductivity between model layers 2 and 3, 3 and 4, and 4 and 5 of the ground-water flow model of southern Utah and Goshen Valleys, Utah.

Specific storage was varied from 1 x 10^{-8}/ft through 1 x 10^{-6}/ft during model calibration. For a thickness of 1,000 ft, the equivalent range in storage coefficient is 1 x 10^{-5} through 1 x 10^{-3}. This possible range of storage-coefficient values is slightly larger than the range listed in table 13. Usual values for storage coefficient range from 1 x 10^{-5} through 5 x 10^{-3} (Todd, 1959, p. 45-46, and Lohman, 1972, p. 8). To match measured water-level fluctuations during model calibration, specific storage was assigned a final value of 5 x 10^{-7}/ft, which is equivalent to a storage coefficient of about 5 x 10^{-4}. Specific storage was not spatially varied and is equal throughout the modeled area.

Specific yield was varied from 0.01 through 0.3 during model calibration. This range agrees with commonly estimated values of specific yield of unconsolidated basin-fill deposits (Lohman, 1979, p. 8, and Freeze and Cherry, 1979, p. 61). To match measured water-level fluctuations, recharge, and discharge during model calibration, specific yield was assigned a value of 0.05. Specific yield was not spatially varied and is equal throughout the modeled area.

Manning's roughness coefficient was varied from 0.03 to 0.05. This is equal to the range of experimental values for Manning's roughness coefficient for natural streams (Prudic, 1989, table1) and is thought to be a reasonable estimate for streams in the modeled area. As a result of model calibration, Manning's roughness coefficient was assigned a final value of 0.03 for all simulated streams. This value causes the greatest change in computed water-surface altitude with change in stream and canal flow, and results in the greatest fluctuation of ground-water recharge and discharge.

Steady-State Calibration

To simulate steady-state conditions, the ground-water flow model was calibrated to the hydrologic conditions of the main ground-water system for 1949. On the basis of observed water-level fluctuations for 1946-52 (fig. 29), the main ground-water system was assumed to be in a steady-state condition during 1949. Water-levels during 1946-52 show little change, indicating that recharge and discharge were about equal and that there was minimal variation in the amount of ground-water storage. The hydrologic conditions to which the model was calibrated include estimated annual recharge and discharge for 1949 and water levels during March 1950.

Hydrologic data available for 1949 include flow and diversion records for Spanish Fork, climatic data at Payson, and estimated ground-water withdrawals for selected wells. Measured water levels at nine wells are available for March 1950. Because of this limited amount of water-level data, water-level data for 1964-66 and 1989-91 were used to augment the data available for March 1950. Although water-level data from these periods do not represent water levels during the 1949 steady-state period, the additional water-level data was used to "rough in" calibration parameters for areas where no water-level data are available for March 1950.

Using the same methodology described in the "Ground-water hydrology" section of this report, an estimated ground-water budget for 1949 was compiled (table 17). All recharge components of the ground-water budget were estimated specifically for 1949. Discharge components of the ground-water budget that were estimated specifically for 1949 include pumped wells, flowing wells, and perennial streams and major canals. Data are not available to estimate discharge to springs and drains, by evapotranspiration, by seepage to Utah Lake, and by infiltration to sewer systems on an annual basis. Initial values of calibration parameters defined in "Model construction" of the "Simulation" section of this report were varied to achieve an acceptable level of agreement between model-computed and measured water levels, and model-computed and estimated budgets.

The potentiometric surface generated from the model-computed water level for layer 3 and the difference between model-computed and measured water level in nine wells is shown in figure 30. The differences shown may have some inherent error built in because the model simulation calculates an average water level for an entire model block for the center of that block. The well represented by that model block can be located anywhere within the block, but rarely will the screened interval for a well coincide precisely with the horizontal and vertical center of a block. These potential errors would be greatest in areas where the water-level gradient is steep or where cell size is large. Both situations could allow the calculated water level for the model block to have a notable difference from a water level in a well located near the outer border of that block. Differences of greater than 10 ft between model-computed and measured water level occur west and south of Springville and near Payson. The largest difference occurs near Payson, where the model-computed water level is about 30 ft lower than the corresponding measured level at well (D-9-2)11aaa-1.

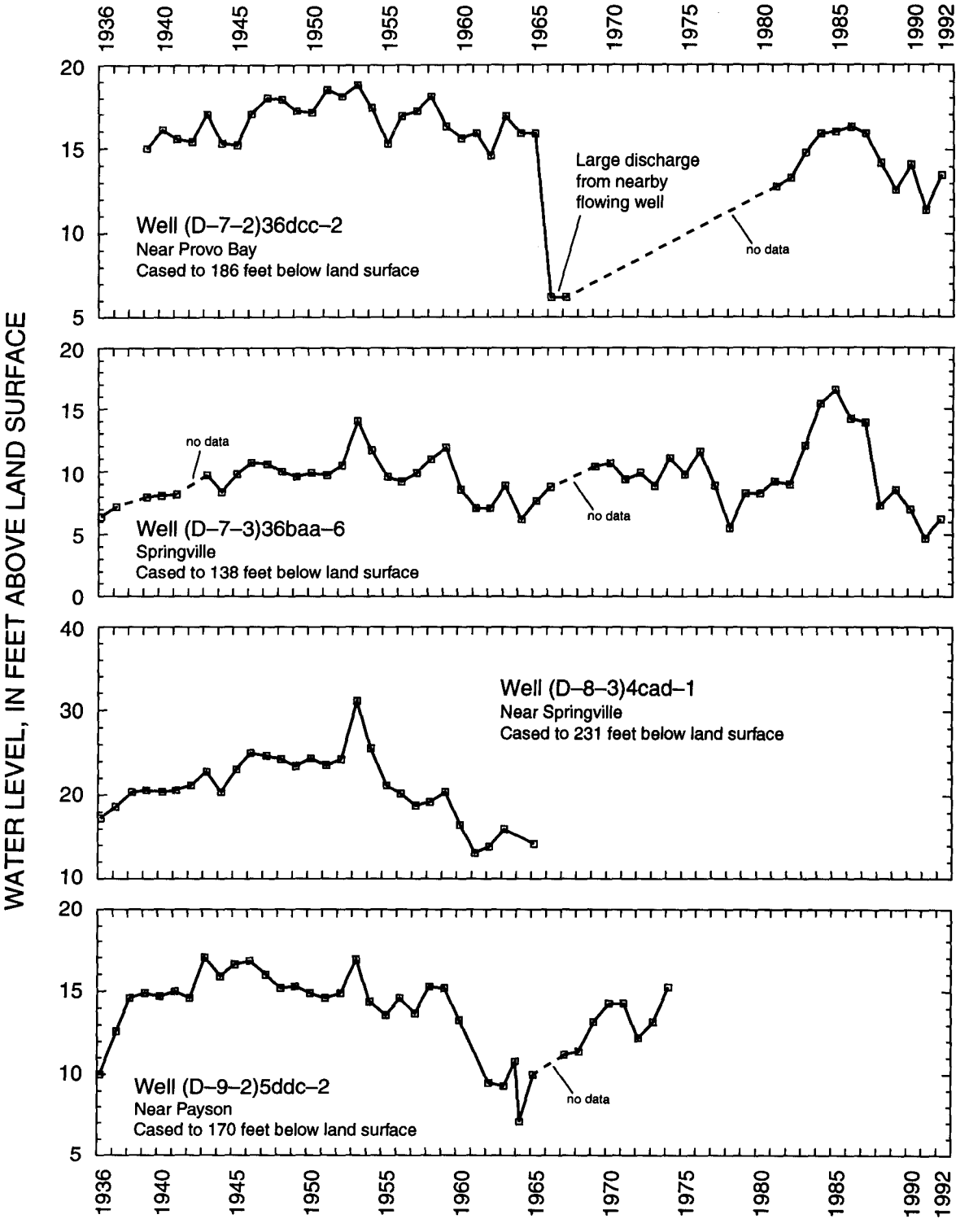


Figure 29. Water level in selected wells in southern Utah Valley, Utah.

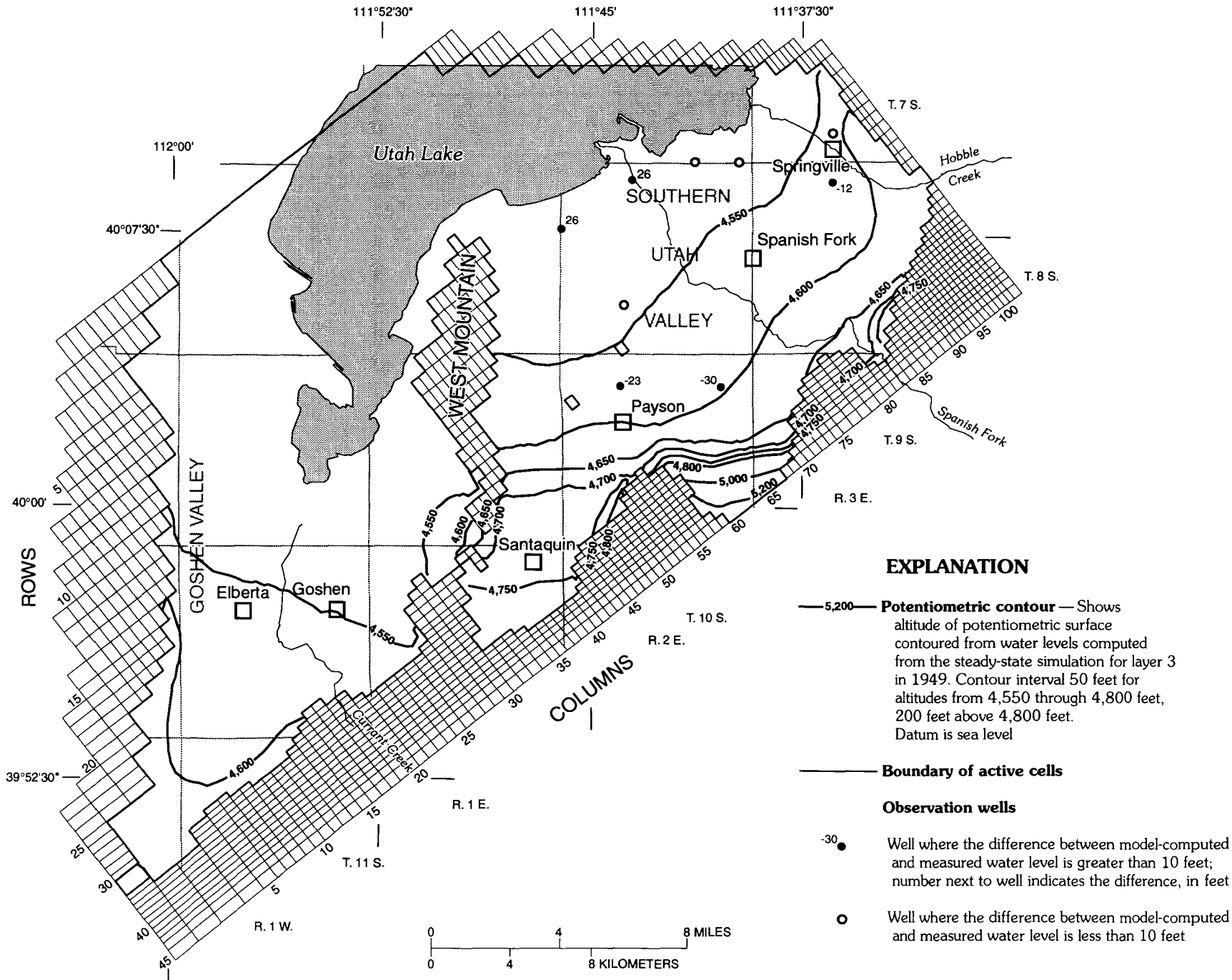


Figure 30. Potentiometric surface for model layer 3 contoured from water levels computed from the steady-state simulation for 1949 and the difference between model-computed and measured water levels in observation wells, southern Utah and Goshen Valleys, Utah, March 1950.

Total ground-water flow computed for the 1949 steady-state simulation is about 14 percent (16,000 acre-ft) greater than that estimated for 1949 (table 17). Individual budget elements where differences between model-computed and estimated amount of flow are greater than 10 percent of the total estimated ground-water flow (11,700 acre-ft) are recharge from subsurface inflow and discharge to evapotranspiration. Recharge from subsurface inflow specified in the model is about 170 percent of estimated subsurface inflow for 1949. Subsurface inflow simulated at specified-flux boundaries was adjusted to obtain an acceptable match between model-computed and measured water level for areas near the mountains, and model-computed discharge with estimated discharge to selected springs and drains, by evapotranspiration, and by seepage to Utah Lake. Because water-level data for areas near the mountains and discharge data are limited for 1949, subsurface inflow also was adjusted during transient-state calibration. Subsurface inflow specified for simulation of 1949 conditions is similar to the average of the estimated subsurface inflow for 1949, 1964, and 1990, which is about 63,400 acre-ft. The difference between specified subsurface inflow and estimated subsurface inflow reflects uncertainty both in the model and in the estimated ground-water budget.

Model-computed discharge by evapotranspiration is about 130 percent of estimated evapotranspiration for 1949; however, discharge to evapotranspiration, to springs and drains, and to sewer systems may be underestimated in the 1949 estimated ground-water budget. The 1949 estimates are based mainly on measurements made in 1965 and 1990, when ground-water withdrawal by pumping wells was more than an order of magnitude larger than in 1949 (see "Discharge" of the "Ground-water hydrology" section of this report). In general, increased ground-water withdrawal causes water levels to decline and thereby reduces discharge to springs and drains and by evapotranspiration. Data are not available to quantify the effects of increased ground-water withdrawal on discharge, but estimated discharges for 1949 probably are lower than actual values. Considering this, the amount of model-computed discharge to evapotranspiration is considered acceptable.

Transient-State Calibration

To simulate transient-state conditions, the ground-water flow model was calibrated to simulate annual stress conditions of the main ground-water system from 1949 to 1990. Initial conditions for the 1949-90 transient-state calibration are the results of the 1949 steady-state simulation. Each annual stress period was divided into three equal time steps. On the basis of variations in model-computed water-levels using different time steps, three time steps were determined to be adequate for the analysis of results at the end of the stress periods.

Changes in the amount and type of stress applied to the main ground-water system are simulated by varying input data at model boundaries. At respective head-dependent flux boundaries, stream and canal flow for Hobble Creek, Spanish Fork, Currant Creek, South Field Canal, Mill Race Canal, and South Ditch is varied for each stress period. The water level of Utah Lake also is varied for each stress period. At specified-flux boundaries, ground-water withdrawals from wells and recharge from perennial streams and major canals, intermittent and ephemeral runoff, and irrigation and precipitation are varied for each stress period. The methods and information used to vary these stresses are discussed in "Model construction" of the "Simulation" section of this report.

The ground-water flow model was calibrated to 1949-90 transient-state conditions primarily by comparing model-computed water-level fluctuations to measured water-level fluctuations (figs. 31 and 32). The calibration also is based on matching measured water levels for March 1965 and March 1991. Model-computed ground-water budgets for the appropriate stress periods also were compared with estimated ground-water budgets for 1964 and 1990 (table 17).

In southern Utah Valley, model-computed water levels generally declined from 1958 to 1964 (fig. 31a, b, e, and j). These declines agree fairly well with measured declines in the Springville area (fig. 31a and b) and near Santaquin (fig. 31j). For the same period, model-computed declines near the town of Spanish Fork are greater than measured declines (fig. 31e). Model-computed and measured water levels in southern Utah Valley rise in the early 1980's, reaching a maximum in 1985, and decline from 1985 to 1991 (fig. 31a-d, f-k). The magnitude of computed water-level fluctuations is correctly simulated for areas near Springville (fig. 31b). Computed water-level fluctuations for areas near the town of Spanish Fork, near the mountains in the vicinity of Spanish Fork Canyon, and near Salem (fig. 31d, f, and j; pl. 1) are less than measured fluctuations. Computed water-level fluctuations for areas near Utah Lake, north of and near Santaquin,

and at Payson (fig. 31a, c, g, h, i, and k) are greater than measured fluctuations. Model-computed water-level fluctuations for 1982-90 are caused mainly by changes in the amount of specified recharge from perennial streams and major canals, intermittent and ephemeral runoff, and irrigation and precipitation. A substantial component of this recharge is varied on the basis of the natural streamflow in Spanish Fork at Castilla (see "Model construction" of the "Simulation" section of this report). Model-computed water-level fluctuations for 1982-90 show that change in flow in Spanish Fork may be a good indicator of when change in recharge occurs but may not be a complete descriptor of the amount of change that occurs. Computed water-level declines from 1985-91 are caused by the simulation of increased well withdrawal and decreased recharge for that time period.

As mentioned in the previous paragraph, model-computed water-level fluctuations are not as large as measured fluctuations for the area near the mountains in the vicinity of Spanish Fork Canyon (fig. 31f). Measured water-level fluctuations near the mountains in this area may be the result of annual variation in the amount of subsurface inflow from consolidated rock. In the model, subsurface inflow from consolidated rock is treated as specified recharge that is not varied during transient-state simulation. This may be an oversimplification of recharge from subsurface inflow in the vicinity of Spanish Fork Canyon.

Model-computed water levels in the northern areas of Goshen Valley decline from 1973 to 1978 and from 1988 to 1990 (fig. 32a). In contrast, measured water levels for the same area remain fairly steady for the 1973-77 and 1988-90 periods (fig. 32a). Farther south, model-computed water levels increase from 1967 to 1972, in contrast to declines in measured levels for the same time period (fig. 32b). Near Elberta, model-computed water levels more closely resemble measured water-level fluctuations, except during 1969 and from 1972 to 1974, when model-computed water levels remain fairly steady while measured levels rise (fig. 32c).

Model-computed water levels in southeastern Goshen Valley show the same type of water-level fluctuations as measured water levels; however, the general trend of the model-computed levels diverges from that of measured levels (fig. 32d). The match between model-computed and measured water-level fluctuations can be improved by increasing the amount of simulated subsurface inflow in southeastern Goshen Valley. This was not done because it causes computed water levels in model layer 1 to be above land surface in parts of Goshen Valley.

The differences in model-computed and measured water-level fluctuations in Goshen Valley indicate that the transient-state simulation is poorly calibrated in this area. There are several reasons for this. Water-level data is limited in areas near the East Tintic Mountains and Utah Lake, making it difficult to accurately define the hydraulic gradient between the major recharge and discharge areas of the valley. Many of the wells used for water-level measurements in Goshen Valley are perforated for long intervals (Stolp and others, 1993, table 1), so the vertical and horizontal components of hydraulic gradient cannot be accurately separated. The estimated pumpage withdrawal for Goshen Valley might be incorrect because of the complex and changing irrigation systems. Although model construction and boundary conditions are based on best estimates from available data, they might not be correctly simulating all aspects of the ground-water system in Goshen Valley.

The potentiometric surface determined from model-computed water levels in layer 3 at the end of the 1964 stress period and the difference between the model-computed water level and the water level measured in 66 wells in March 1965 are shown in figure 33. Generally, differences between model-computed and measured water levels of greater than 10 ft are fairly evenly distributed throughout the modeled area. The mean of differences between model-computed and measured water levels is 3.91 ft, which indicates that model-computed water levels generally are higher than measured levels. This is evident in southern Goshen Valley, where model-computed water level is consistently higher than measured water level; the largest difference occurs where the model-computed water level is about 32 ft higher than the corresponding measured water level at well (C-11-1)6bdd-1.

The potentiometric surface determined from model-computed water levels in layer 3 at the end of the 1990 stress period and the difference between model-computed water level and water level measured in 112 wells in March 1991 are shown in figure 34. Differences between model-computed and measured water level of greater than 10 ft occur throughout southern Utah Valley. The mean of differences between model-computed and measured water level is -0.17 ft, which indicates a fairly even distribution of positive and negative differences. Model-computed water level is gen-

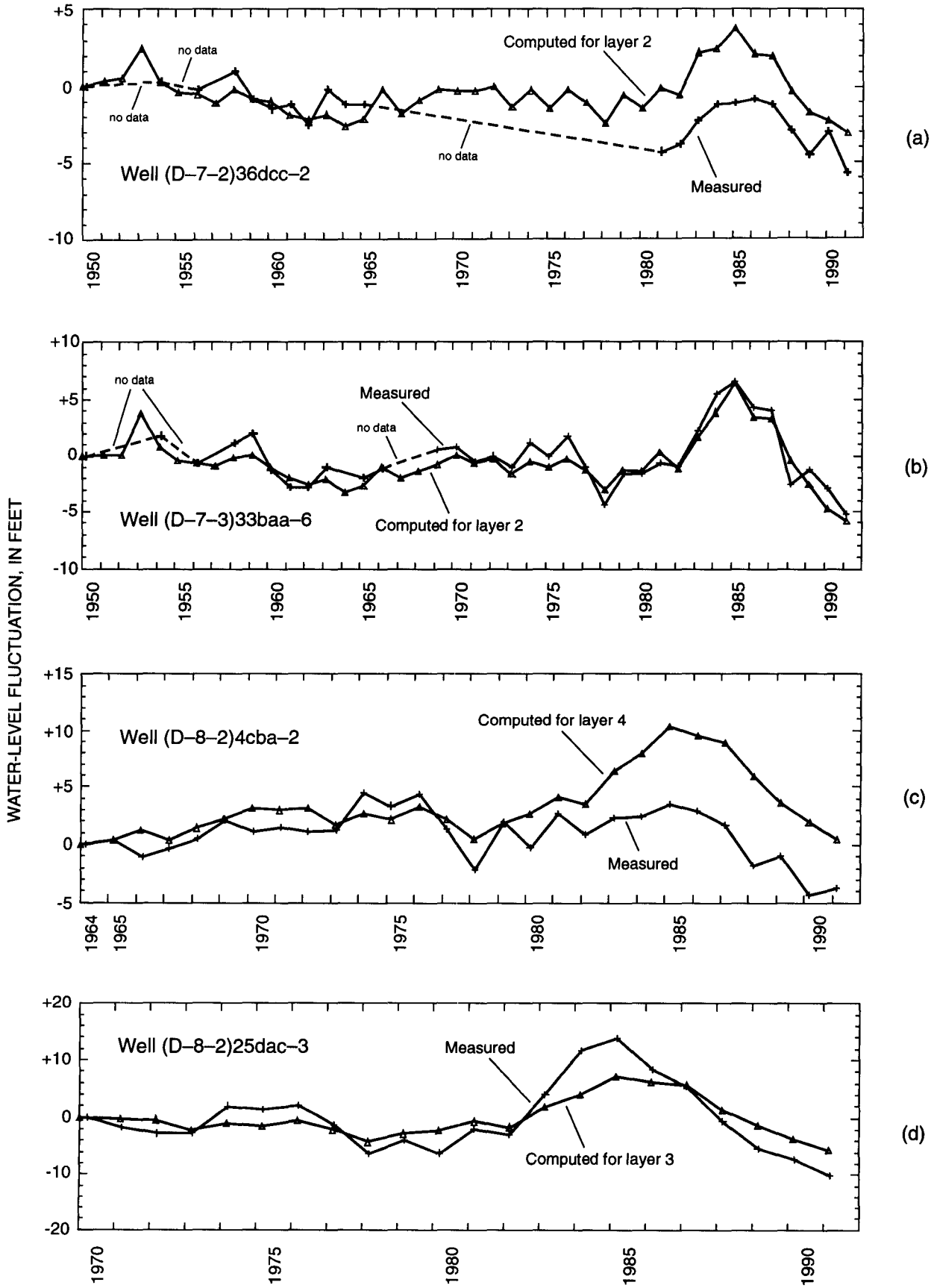


Figure 31. Model-computed and measured water-level fluctuations in southern Utah Valley, Utah, for the annual transient-state simulation.

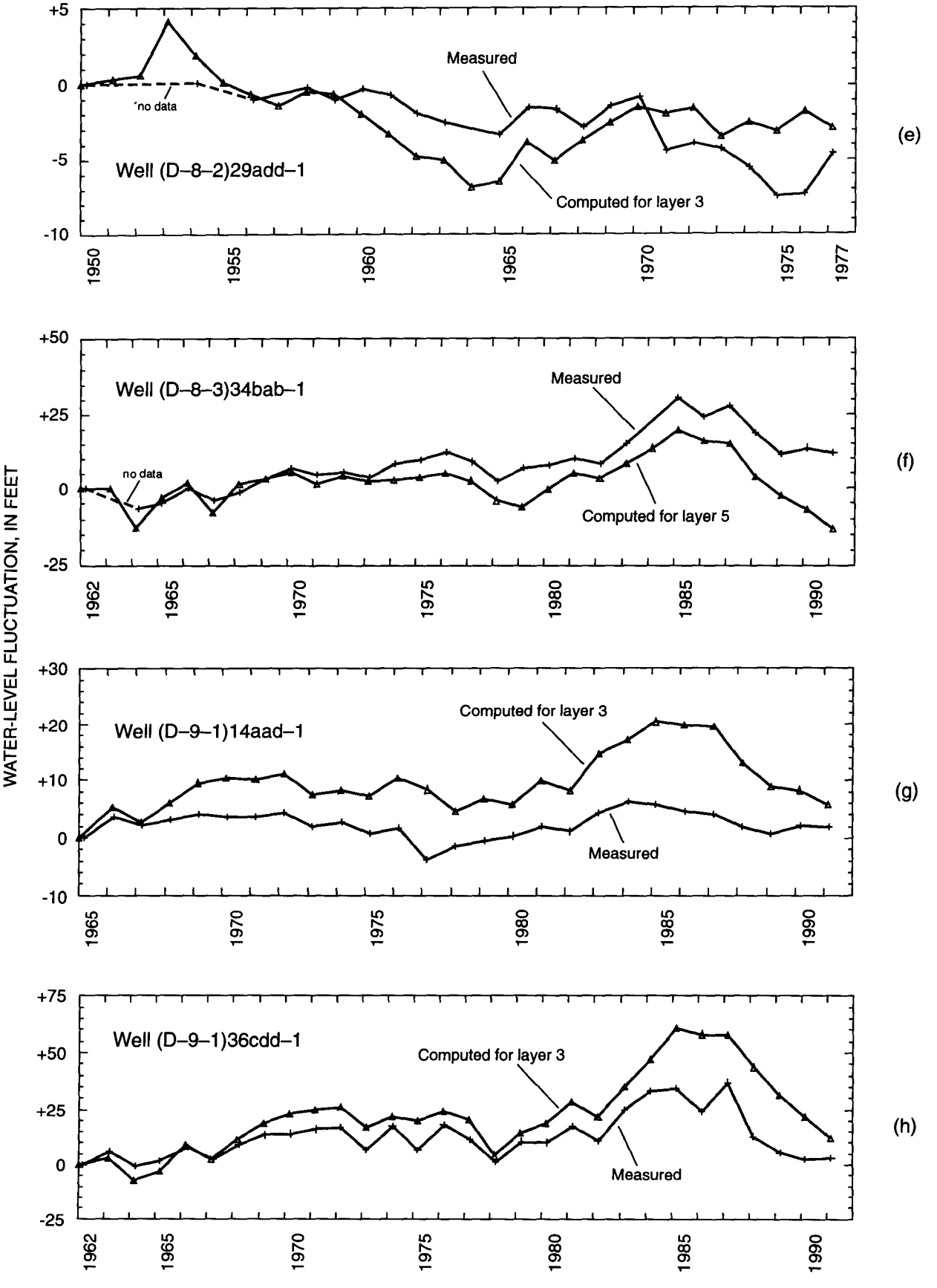


Figure 31. Model-computed and measured water-level fluctuations in southern Utah Valley, Utah, for the annual transient-state simulation—Continued.

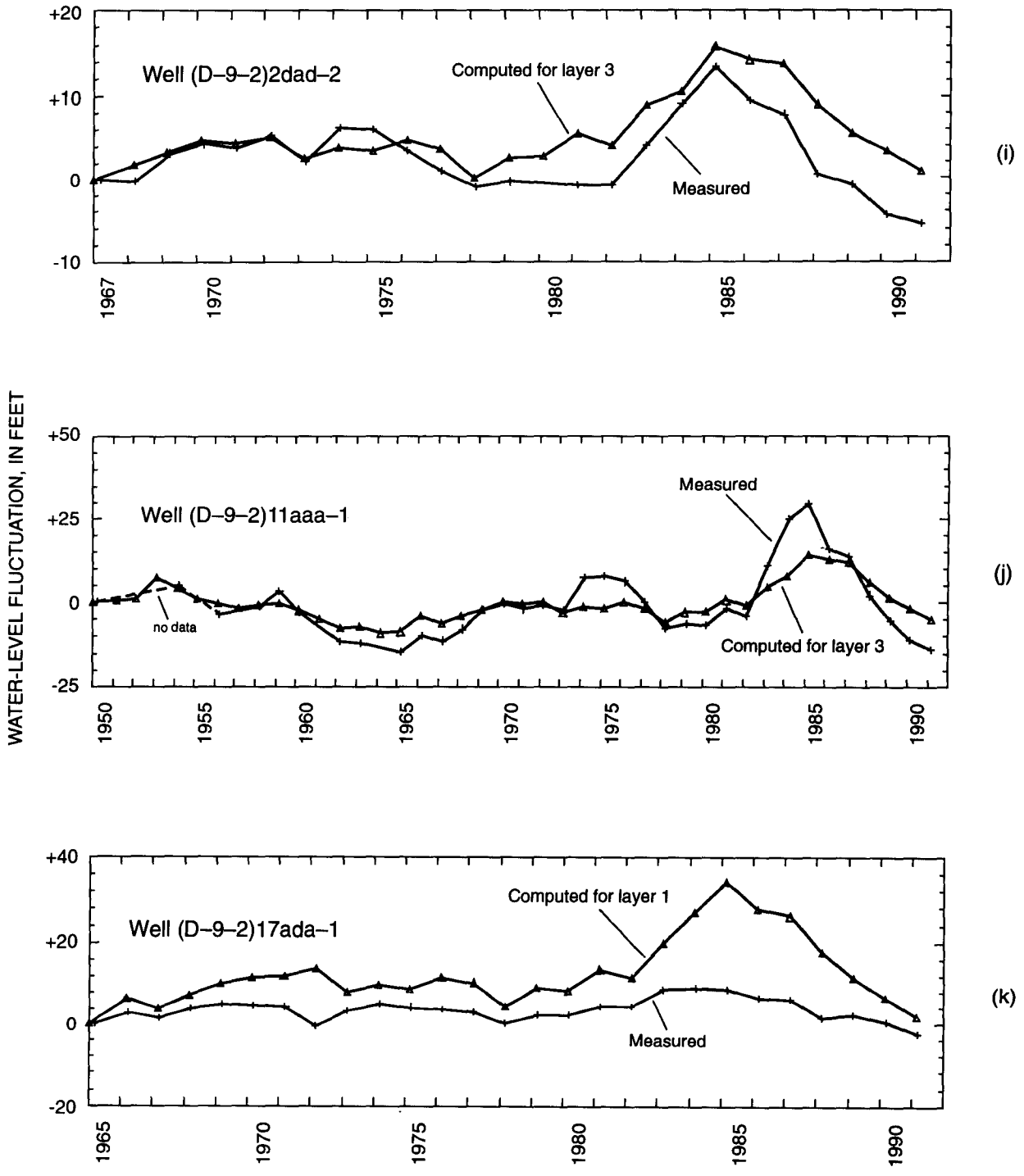


Figure 31. Model-computed and measured water-level fluctuations in southern Utah Valley, Utah, for the annual transient-state simulation—Continued.

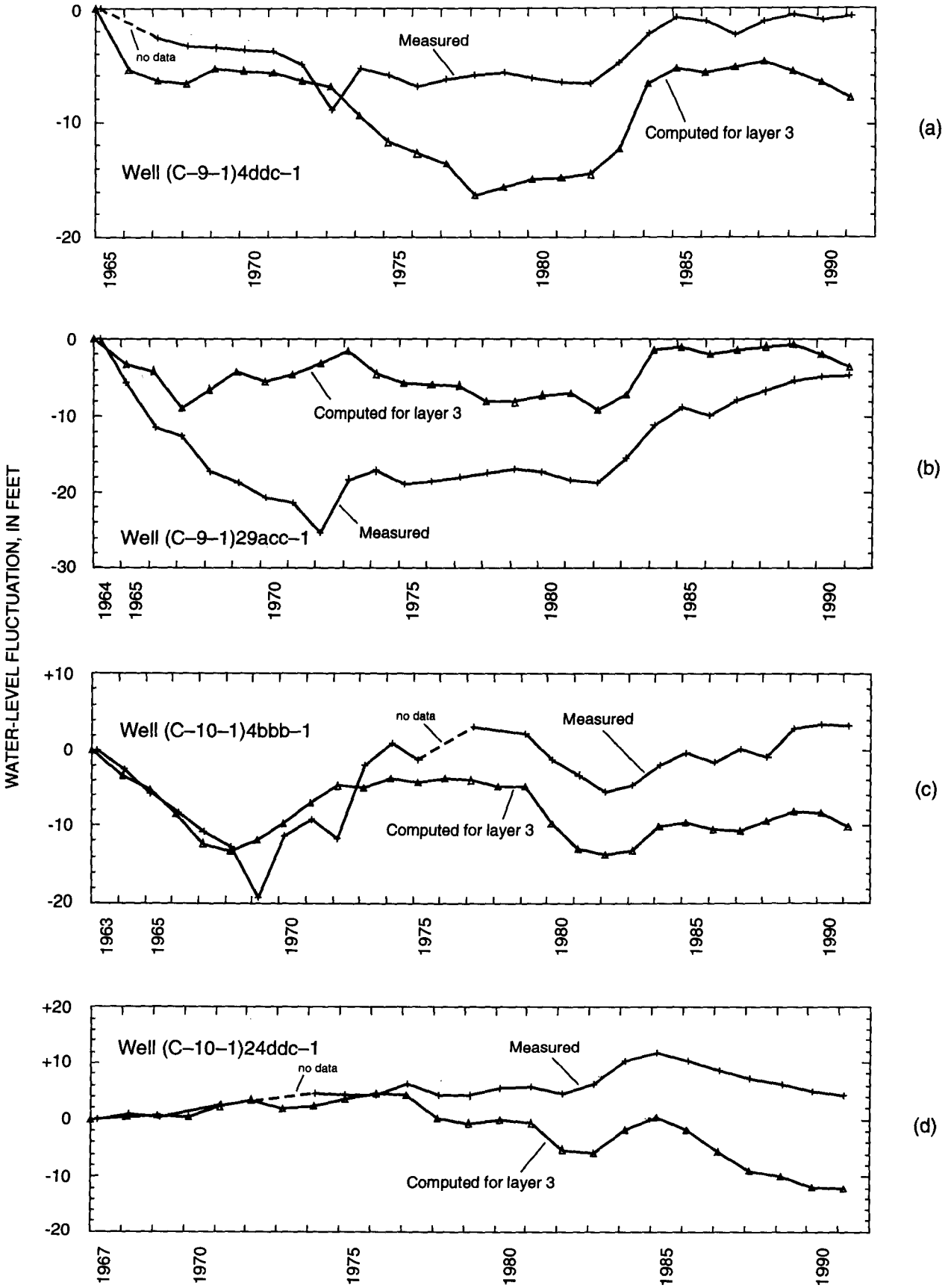


Figure 32. Model-computed and measured water-level fluctuations in Goshen Valley, Utah, for the annual transient-state simulation.

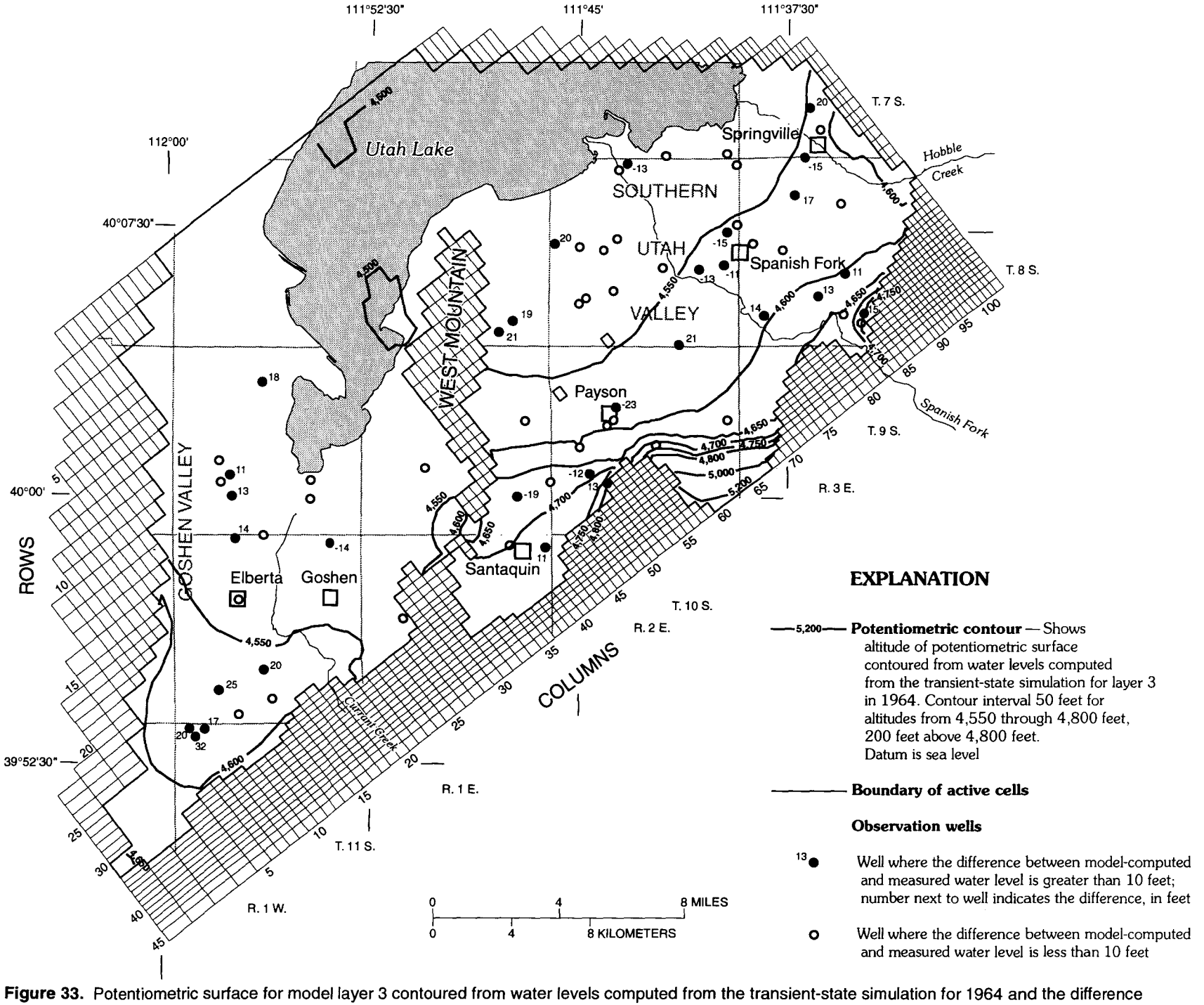


Figure 33. Potentiometric surface for model layer 3 contoured from water levels computed from the transient-state simulation for 1964 and the difference between model-computed and measured water levels in observation wells, southern Utah and Goshen Valleys, Utah, March 1965.

erally higher than measured water level near Utah Lake and west of Spanish Fork. Model-computed water level is generally lower than measured water level near Payson. The largest difference occurs east of the town of Spanish Fork, near the mountains, where the model-computed water level is about 81 ft lower than the corresponding measured level at well (D-8-3)27adc-1. Model-computed water levels in this area are strongly influenced by the horizontal and vertical distribution of the specified-flux boundary that simulates subsurface inflow. As stated previously, calibration to transient-state conditions was done primarily by comparison of water-level fluctuations. The water-level differences shown in figure 34 are for the last stress period of the 1949-90 transient-state simulation and in a sense represent the cumulative error of the transient-state calibration. Considering this, an acceptable match exists between most measured and computed water levels.

Total annual ground-water recharge and discharge computed for stress periods representing 1964, 1983, and 1990 in the transient-state simulation is listed in table 17. The model-computed budget for 1964 is about 8 percent (11,000 acre-ft) less than the estimated budget for 1964, and the model-computed budget for 1990 is about 12 percent (19,000 acre-ft) less than the estimated budget for 1990. For 1964, no model-computed budget element differs from the estimated amount of flow by more than 10 percent of the total estimated ground-water flow (14,000 acre-ft). For 1990, budget elements where differences between model-computed and estimated amount of flow is more than 10 percent of the total estimated ground-water flow (15,600 acre-ft) are recharge from subsurface inflow and discharge to perennial streams and major canals (table 17).

The amount of subsurface inflow specified in the model for 1964 is about 80 percent of the estimated amount for 1964. As discussed previously, the amount of subsurface inflow specified in the model was determined from matching water-level data at wells near the mountains and matching selected discharges. Data are not available to describe the variability of subsurface inflow, and 62,200 acre-ft is used for all stress periods of the transient-state simulation. Estimated recharge from subsurface inflow, however, is determined as a residual (see "Recharge" of the "Ground-water hydrology" section of this report), and the estimated amount varies for 1949, 1964, and 1990 (table 17). As with the difference noted in the steady-state budget, both the model-specified inflow and the estimated subsurface inflow reflect uncertainty in the respective ground-water budget.

Model-computed discharge to perennial streams and major canals in 1964 is about 40 percent of the corresponding estimated amount in 1964, and in 1990 is about 20 percent of the corresponding estimated amount in 1990. This budget component represents the largest percentage difference between model-computed and estimated values. Model-computed discharge to individual perennial streams is listed in table 18. Model-computed net flow totals for Hobble Creek are comparable to estimated amounts. Model-computed discharge to Currant Creek is less than the estimated amount for the 1964 and 1990 stress periods, but the differences are small when compared to the total ground-water budget. Model-computed recharge from and discharge to Spanish Fork are substantially less than estimated amounts for all comparable stress periods (table 18). Increasing the hydraulic conductivity of the streambed did not substantially change the amount of model-computed flow to and from the stream. In areas where recharge occurs from the stream, increasing the streambed conductivity causes the water level to rise above the level of the bottom of the streambed, thereby decreasing recharge. Examination of model-computed flow at areas where discharge to Spanish Fork occurs indicates that almost all ground-water flow in the model cell is discharged to the stream. To increase discharge to the stream, additional ground-water flow would need to be routed toward the stream boundaries. This has been done by adjustment of vertical hydraulic-conductivity values in areas near Hobble Creek and Spanish Fork. Any further adjustment of calibration parameters to induce flow toward the boundaries is not warranted without additional data.

Generally, model-computed discharge to springs and drains, by evapotranspiration, by seepage to Utah Lake, and by infiltration to sewer systems is similar to the estimated amount for 1964 and 1990. In 1964, the model-computed decrease in storage of about 3,500 acre-ft is in contrast with the estimated increase of 6,800 acre-ft. In 1990, model-computed decrease in storage is about 160 percent of the estimated decrease in storage, which indicates that model-computed water-level declines for the 1990 stress period are larger than measured declines.

To illustrate the effects of a simulated period of greater-than-average precipitation, the model-computed ground-water budget for the 1983 stress period also is listed in table 17. Precipitation at Payson in 1983

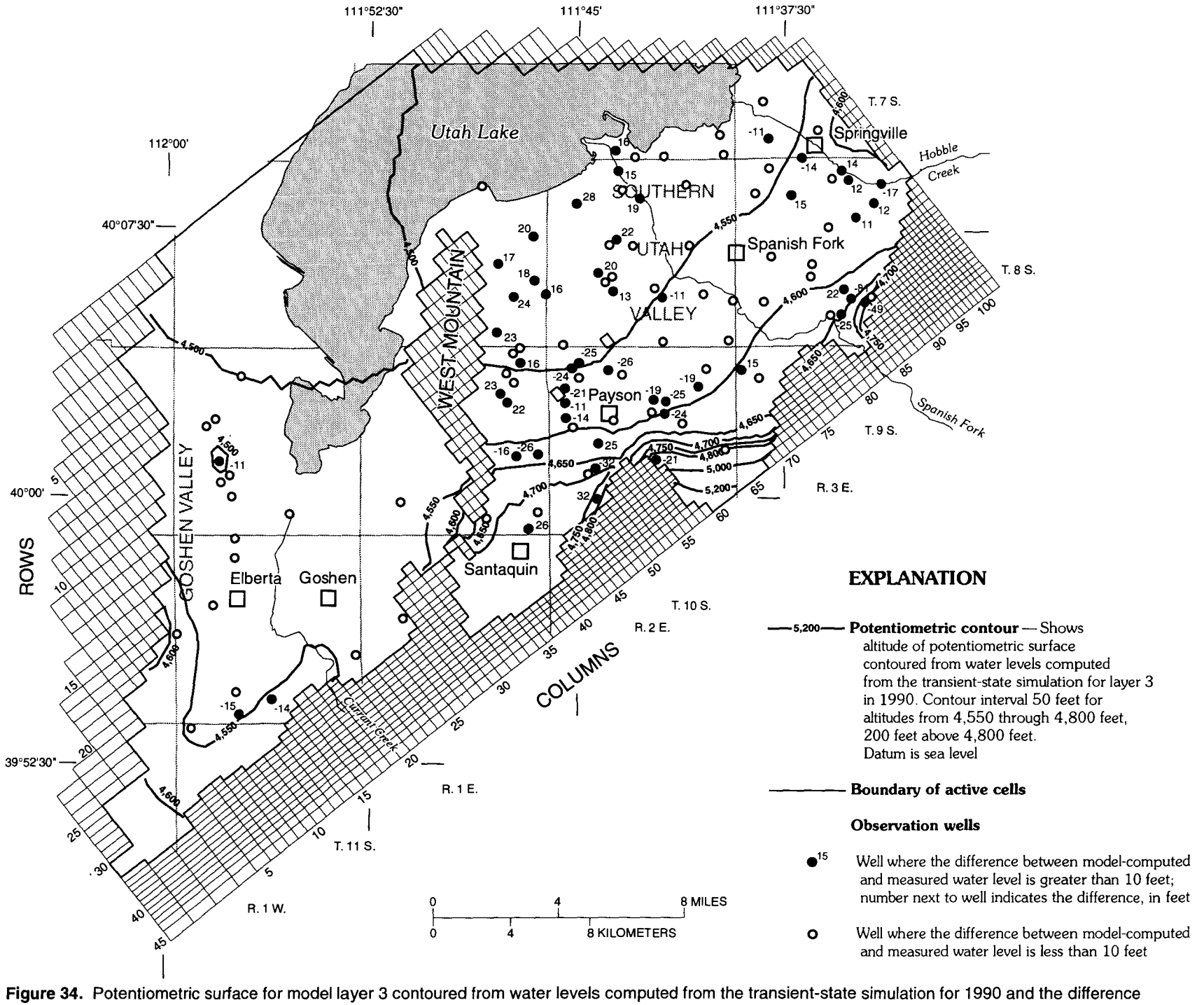


Figure 34. Potentiometric surface for model layer 3 contoured from water levels computed from the transient-state simulation for 1990 and the difference between model-computed and measured water levels in observation wells, southern Utah and Goshen Valleys, Utah, March 1991.

Table 18. Estimated and model-computed recharge from and discharge to perennial streams simulated with head-dependent boundaries, southern Utah and Goshen Valleys, Utah

Stream		Flow, in acre-feet per year						
		Estimated for 1949	Model-computed 1949 steady state	Estimated for 1964	Model-computed 1964 transient state	Model-computed 1983 transient state	Estimated for 1990	Model-computed 1990 transient state
Hobble Creek	Recharge from	7,400	5,000	7,400	4,800	5,700	7,400	4,400
	Discharge to	4,500	2,200	4,500	1,700	2,900	4,500	1,200
	Total[1]	**+2,900**	**+2,800**	**+2,900**	**+3,100**	**+2,800**	**+2,900**	**+3,200**
Spanish Fork[2]	Recharge from	13,700	6,400	15,300	8,100	4,600	24,000	10,300
	Discharge to	7,800	4,700	10,600	2,900	5,000	16,200	2,600
	Total[1,3]	**+5,900**	**+1,700**	**+4,700**	**+5,200**	**-400**	**+7,800**	**+7,700**
Currant Creek	Recharge from	1,000	1,200	1,000	1,400	1,600	1,000	1,600
	Discharge to	2,200	2,100	2,200	1,200	800	2,200	800
	Total[1,3]	**-1,200**	**-900**	**-1,200**	**+200**	**+800**	**-1,200**	**+800**

[1] Positive value indicates net recharge from perennial stream to the ground-water system.
[2] Amount listed for Spanish Fork includes recharge from and discharge to Mill Race Canal and South Field Canal.
[3] Negative value indicates net discharge to perennial stream from the ground-water system.

was about 200 percent of average (fig. 3). Increased precipitation is simulated in the transient-state calibration by increasing the specified part of recharge from perennial streams and major canals, irrigation and precipitation, and intermittent and ephemeral runoff. Modeled recharge from these sources during the 1983 stress period is about 210 percent of the amount modeled for 1964 (table 17). The 1983 transient-state model-computed budget shows that discharge to springs and drains was about 140 percent of the model-computed discharge to springs and drains during the 1964 stress period, and discharge to evapotranspiration during 1983 was about 120 percent of discharge to evapotranspiration during the 1964 stress period. Discharge to Utah Lake decreased during 1983; this is probably the result of decreases in upward vertical and horizontal gradient caused by a rise in the lake level in 1983 (fig. 14). The model-computed increase in storage during 1983 is 27,300 acre-ft; this is reflected in model-computed water-level increases in 1983 (figs. 31a-d and f-k and 32a-d).

As part of the transient-state calibration process, differences between model-computed water levels in model layers 1 and 5 were calculated and qualitatively compared to the general directions of vertical ground-water movement presented in this report. The difference between model-computed water levels in layers 1 and 5 also was compared to the difference in water levels measured in closely spaced wells (fig. 33). In general, model-computed vertical ground-water movement agrees with the conceptual understanding of the main ground-water system, with downward movement near the mountains and upward movement in the valleys. Movement is downward in areas where the model-computed water-level difference is positive and upward in areas where the difference is negative. Model-computed differences greater than 50 ft occur near the mountains east of Springville and Spanish Fork. These differences are created by the vertical placement of the specified-flux boundary conditions used to simulate subsurface inflow (see "Model construction" of the "Simulation" section of this report); data are not available to define the vertical accuracy of the boundaries. Water levels measured at three sets of closely spaced wells east of Spanish Fork have differences ranging from 43 ft to 69 ft (fig. 35). Model-computed and measured water-level differences do not compare for areas west of Springville and south of Spanish Fork. In these areas, model-computed differences indicate upward movement of ground water, and measured differences indicate downward movement.

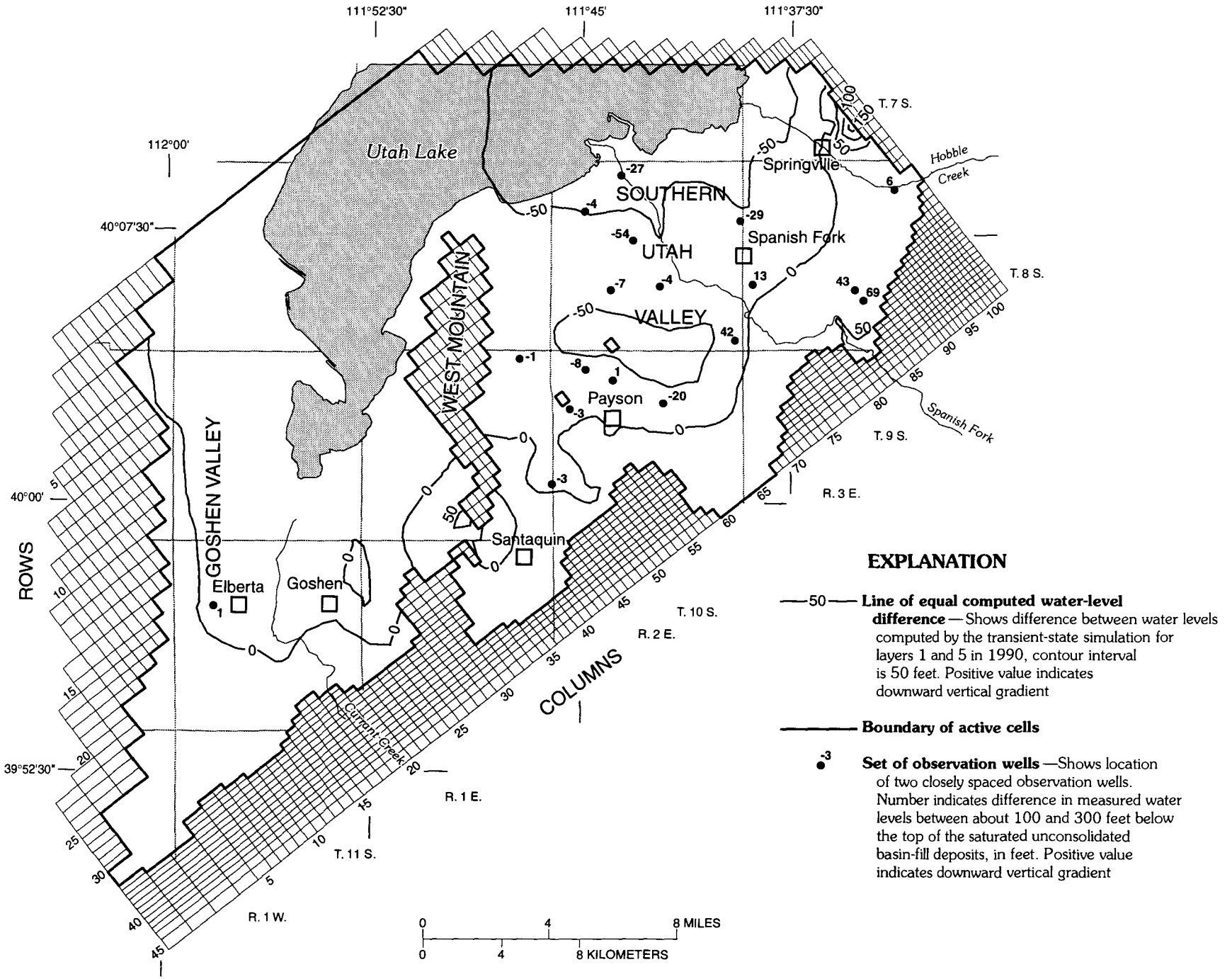


Figure 35. Computed water-level difference between model layers 1 and 5 for the 1990 transient-state simulation and measured water-level difference in observation wells, southern Utah and Goshen Valleys, Utah.

Model Sensitivity

The sensitivity of model-computed water level and flow to changes in aquifer properties was qualitatively observed during model calibration. Once the ground-water model was calibrated, selected boundary conditions and stresses were changed and the resulting changes in water level and flow were observed.

Model-computed water level and flow are more sensitive to changes in vertical hydraulic conductivity in discharge areas in the central parts of southern Utah Valley and near and underneath Utah Lake than to changes in other aquifer properties. Recharge from and discharge to streams and drains is sensitive to streambed- and drain-conductance values only to a certain point, beyond which increased conductance values have limited effect on recharge, discharge, and local water level. Local water levels and discharge to evapotranspiration are sensitive to changes in the maximum evapotranspiration rate. Increasing discharge to evapotranspiration by increasing the maximum evapotranspiration rate also decreases discharge to drains. Model-computed water level near the mountains and near areas of large ground-water withdrawal are sensitive to changes in horizontal hydraulic conductivity.

Water level and discharge for the entire modeled area are sensitive to the amount and horizontal and vertical distribution of subsurface inflow. Increasing subsurface inflow increases the hydraulic gradient near the mountains, increases discharge to springs and drains, and raises water level.

Water-level fluctuations in the upper model layers are sensitive to changes in specific yield of model layer 1. This sensitivity is caused by the fact that a substantial amount of total recharge is applied as specified flux in model layer 1. Water-level fluctuations are not sensitive to changes in storage coefficient in model layers 2, 3, 4, and 5. North of Santaquin and in northern Goshen Valley, two areas where large ground-water withdrawal has occurred, water-level fluctuations were not sensitive to annual changes in ground-water withdrawal at selected wells. Water-level fluctuations simulated with the transient-state calibration seem to be most sensitive to the timing and magnitude of the recharge applied as specified flux.

Model Limitations

This model represents a simplification of the natural ground-water system and is based on estimates of recharge, discharge, and aquifer properties from available data. The boundary conditions and aquifer properties used in this model do not represent a unique solution; other combinations may yield similar results. Simulations of stress conditions that are outside the range of conditions to which the model was calibrated should be considered unreliable.

Specific model limitations include the effects of stresses near the mountains. Subsurface inflow is simulated by using injection wells in model layers 1 and 2 at the boundary of the modeled area. Model-computed water levels near this boundary are strongly influenced by the way in which subsurface inflow is formulated. Additional data on ground-water levels in consolidated rock are needed to better define subsurface inflow. The specified-flux boundary formulated with injection wells is independent of hydraulic gradient across the model boundary. Model-computed water-level declines caused by simulating ground-water withdrawal near this boundary could be larger than actual declines.

Model-computed recharge and discharge along specific reaches of Hobble Creek, Spanish Fork, and Currant Creek do not agree with measured or estimated values. Estimates for streamflow losses into the ground-water flow system are higher than could be simulated with the Streamflow Package in Hobble Creek and Spanish Fork. If this shortcoming is a function of simulated streambed conductance being too small, then simulated water levels near those streams in the projections will be too low. The opposite would be the case with Currant Creek, where simulated streamflow loss (aquifer recharge) was higher than the estimated loss. Along the gaining reaches of the three streams, simulated gain was smaller than estimated. In the projection runs, this might cause simulated water levels near these stream reaches to be higher than actual. Additional calibration of the model using more accurate streambed altitudes might correct this existing model limitation. Additional data on ground-water level near the river are needed to better define the relation between the river and the ground-water system. That information, in conjunction with additional model calibration, could be used to improve the simulation in those areas.

Model-computed water-level fluctuations in Goshen Valley are not in close agreement with measured water-level fluctuations. Differences in water-level fluctuations could not be rectified without changing aquifer properties and boundary conditions beyond the conceptual understanding of the ground-water system in Goshen Valley. Although model-computed and measured water-level fluctuations are similar from

about 1980 to 1990, this calibration of the ground-water model reflects the uncertainties in the conceptual understanding of the ground-water system in Goshen Valley.

Model Projections

This ground-water model of the unconsolidated basin-fill deposits of southern Utah and Goshen Valleys was used to determine possible effects on recharge, discharge, and water levels caused by increased withdrawals from wells. The following stress conditions were simulated: (1) Specified discharge to wells simulated during the 1990 stress period was simulated at 120, 140, 160, 180, and 200 percent of the estimated amount for 1990, and (2) The ground-water model was modified by the addition of 31 wells in areas where future municipal ground-water development is expected, and a discharge of about 10,000 acre-ft/yr from the additional wells was simulated. For both of these simulations, specified recharge was set equal to the average of the recharge used in the 1949-90 transient-state simulation. This average is representative of the long-term average and is about 160 percent of the recharge specified for the 1990 stress period. Results of the projection simulations show the effects of both increased withdrawals from wells and increased specified recharge. Although effects of individual stresses cannot be evaluated independently, the 1949-90 average recharge is a reasonable simulation of anticipated future hydrologic conditions in southern Utah and Goshen Valleys. Because of large changes in both recharge and discharge stresses, model-computed water levels and fluxes for the first 5 years of the projection simulations are not reliable. Also, the 200 percent increase in 1990 withdrawal exceeds the largest withdrawal from wells used in the 1949-90 transient-state calibration (about 33,000 acre-ft in 1976) by about 21,000 acre-ft, so projections should be used with caution.

The model-computed ground-water budget for the 1990 stress period is compared with the ground-water budgets for the projection simulations in table 19. Recharge from perennial streams and major canals that are in hydraulic connection with the ground-water flow system (simulated as head-dependent flux boundaries) can change in response to declining heads resulting from increased pumpage. Recharge from other streams and canals not in direct hydraulic connection with the ground-water flow system (simulated as specified-flux boundaries) do not change with simulated changes in hydraulic head, and because these streams and canals are never expected to be in hydraulic connection with the ground-water flow system as long as water levels continue to decline, using a head-dependent flux boundary was unnecessary. If future projections are to include simulating artificial recharge to the ground-water flow system in the area of these canals and streams, users of the model may want to consider altering the existing specified-flux boundaries to head-dependent flux boundaries. For the model projections, simulated recharge from perennial streams and major canals increases by about 10 percent as a result of doubling 1990 withdrawal from wells, and about 5 percent as a result of increased withdrawal from the 31 new wells. Increased withdrawal from wells is compensated for by decreased discharge to springs and drains, by decreased evapotranspiration, and by decreased seepage to Utah Lake. Model-computed discharge to Utah Lake decreased by about 65 percent in response to doubling 1990 withdrawal. On the basis of the model-computed change in storage, the ground-water system may take longer than 20 years to reach a steady-state condition if the 1990 withdrawal from wells is doubled.

The effect on model-computed release from storage, recharge, and discharge as a result of increasing 1990 withdrawal from wells by 120, 140, 160, 180, and 200 percent are indicated in figure 36. Changes in storage, recharge, and discharge are largest during the first 5 years of increased withdrawals. Changes caused by the municipal increases are less than the changes caused by doubling 1990 withdrawal.

Model-computed water-level declines that occur in model layer 3 as a result of increasing withdrawal from wells by 200 percent of the 1990 withdrawal rate are shown in figure 37. At the end of 20 years, water-level declines of 40 to 80 ft are projected for areas in Goshen Valley and water-level declines of 20 ft are projected for localized areas in southern Utah Valley. Simulated 20-year water-level declines of less than 20 ft, caused by increased withdrawals from municipal wells, are projected for southern Utah and Goshen Valleys.

SUMMARY

To meet the need for detailed information on the hydrologic system and to determine the effects of additional ground-water withdrawal, the U.S. Geological Survey, in cooperation with the Utah Department of Natural Resources, Division of Water Rights, studied the ground-water resources of southern Utah and Goshen Valleys, Utah, from 1988 to 1993. Recharge, discharge, and movement of ground water, water-level

Table 19. Model-computed ground-water budget for the 1990 transient-state simulation and two projected 20-year simulations, southern Utah and Goshen Valleys, Utah

Budget element	Flow, in acre-feet per year (rounded)		
	1990 transient state	Doubled[1] 1990 withdrawals from wells	10,000 acre-feet[1] per year increase in withdrawals from 31 new wells
Recharge			
Perennial streams and major canals			
Specified flux[2]	14,700	16,200	16,200
Head-dependent flux[3]	16,600	18,500	17,600
Subtotal	**31,300**	**34,700**	**33,800**
Irrigation and precipitation[2]	15,200	27,800	27,800
Intermittent and ephemeral runoff[2]	6,800	14,100	14,100
Subsurface inflow[2]	62,200	62,200	62,200
Total recharge (rounded)[4]	**116,000**	**139,000**	**138,000**
Discharge			
Springs and drains[3]	41,800	37,200	39,700
Evapotranspiration[3]	44,700	37,100	41,900
Pumped wells[2,5]	27,500	54,300	37,500
Flowing wells[3,5]	2,700	2,100	2,400
Perennial streams and major canals[3]	5,100	3,800	4,100
Utah Lake[3]	9,600	2,700	7,800
Sewer systems[3]	5,900	5,100	5,300
Total discharge (rounded)[4]	**137,000**	**142,000**	**139,000**
Water going into (+) or out of (-) storage (rounded)[6]	-21,100	-3,700	-600

[1]Budget amounts for the projected 20-year simulations are annual rates computed for the last time step in the simulation.
[2]The amount of flow is specified and not computed by the model.
[3]The amount of flow is computed by the model during the simulation.
[4]Model-computed total recharge may not equal model-computed total discharge because of simulation constraints and rounding.
[5]Estimated and modeled amounts may differ because only discharges of 50 acre-feet per year or more were simulated.
[6]Water going into (+) storage is considered to be discharge and should be added to total discharge; water going out of (-) storage is considered to be recharge and should be added to total recharge.

fluctuations, aquifer properties, ground-water quality, and possible effects of additional ground-water withdrawal were emphasized. The main ground-water system in southern Utah and Goshen Valleys is in unconsolidated basin-fill deposits. The consolidated rock surrounding the valleys contains water, but few wells have been drilled into it. This study has replaced the previous interpretation of four distinct aquifers in southern Utah and Goshen Valleys with an interpretation of only one main ground-water system with varying horizontal and vertical permeability.

Recharge to the main ground-water system was estimated to be about 120,000 acre-ft in southern Utah Valley and about 30,000 acre-ft in Goshen Valley in 1990. Recharge is from perennial streams and major canals, irrigation and precipitation, intermittent and ephemeral runoff, and subsurface inflow from consolidated rock and stream channels of the bordering mountains.

Discharge from the main ground-water system was estimated to be about 130,000 acre-ft in southern Utah Valley and about 33,000 acre-ft in Goshen Valley in 1990. Discharge is to springs and drains, by evapotranspiration, to wells, to streams and canals, by seepage to Utah Lake, and by infiltration to sewer systems. Individual spring-discharge measurements in 1990-91 did not differ substantially from individual measurements in 1964-67. Even though water levels in selected

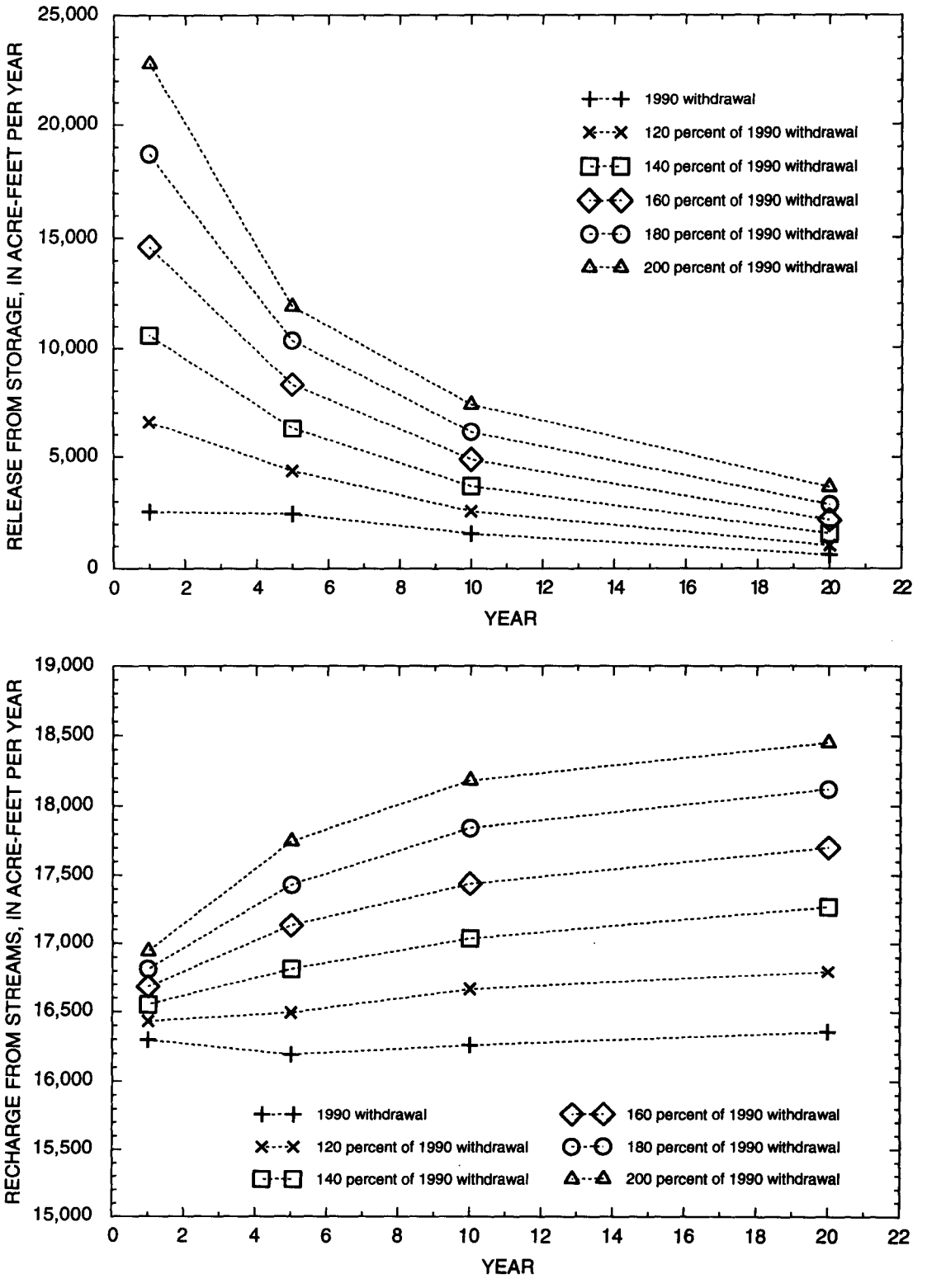
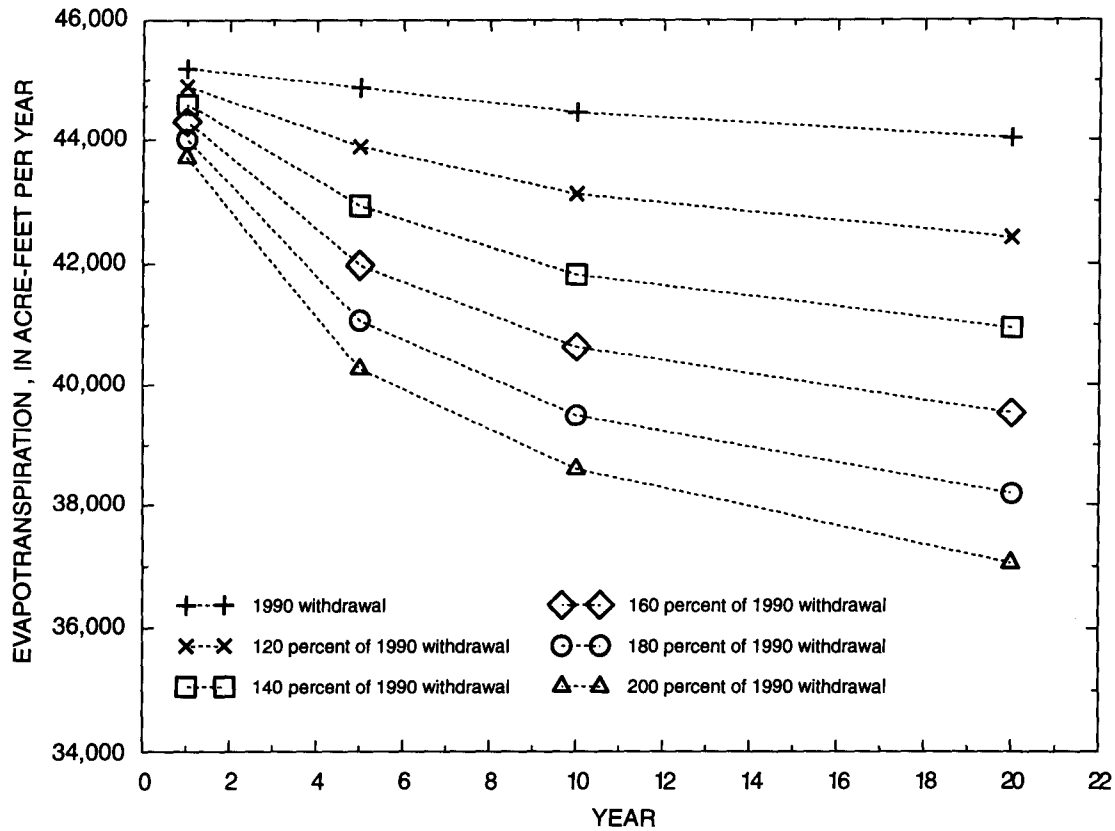


Figure 36. Effect of increased ground-water withdrawal on model-computed release from storage, recharge, and discharge at head-dependent flux boundaries for a 20-year simulation using the ground-water flow model of southern Utah and Goshen Valleys, Utah.



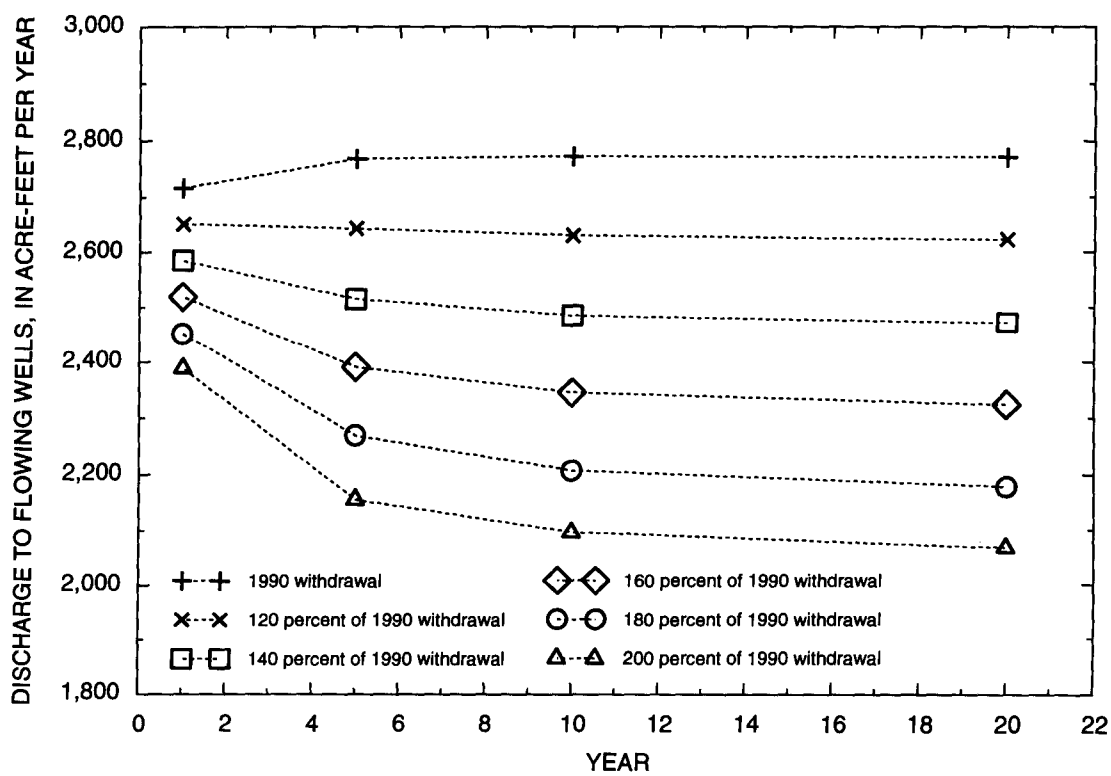
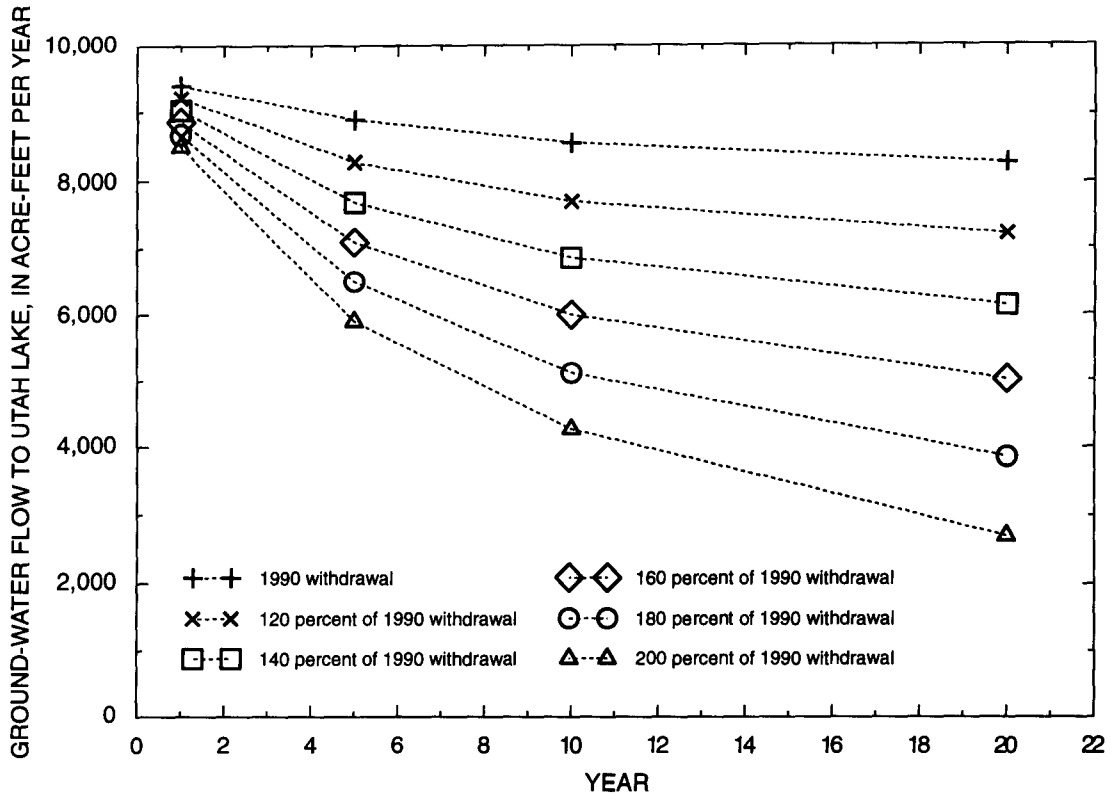


Figure 36. Effect of increased ground-water withdrawal on model-computed release from storage, recharge, and discharge at head-dependent flux boundaries for a 20-year simulation using the ground-water flow model of southern Utah and Goshen Valleys, Utah—Continued.



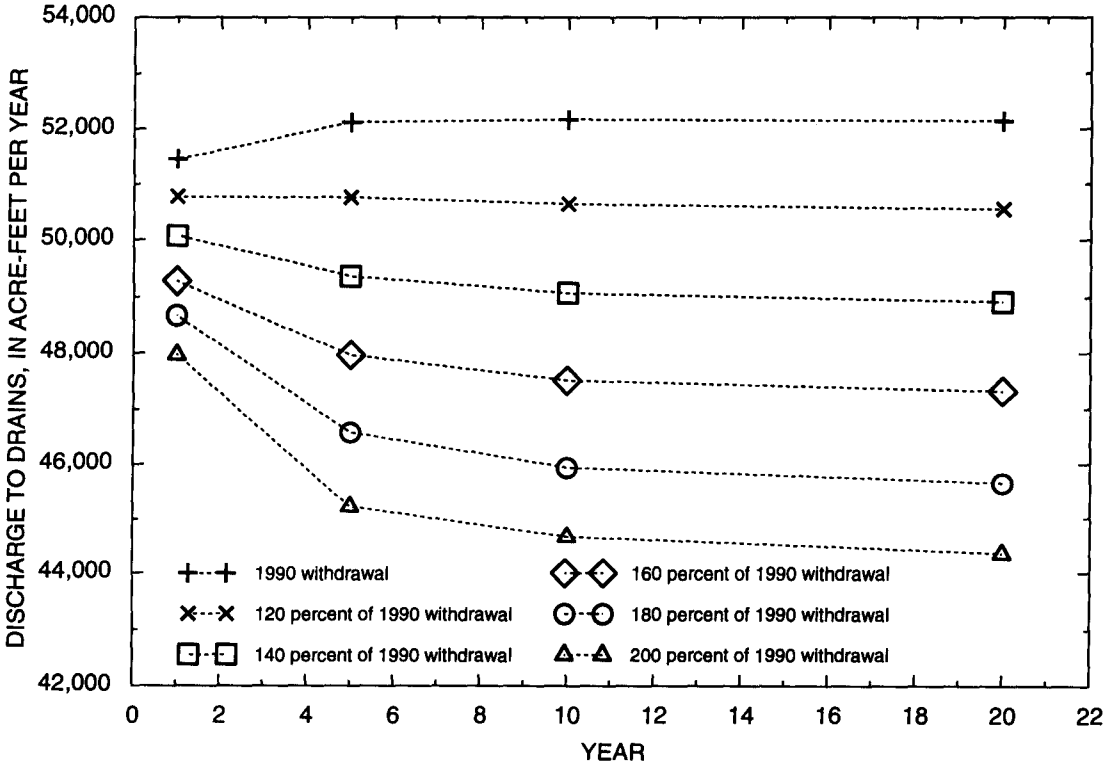


Figure 36. Effect of increased ground-water withdrawal on model-computed release from storage, recharge, and discharge at head-dependent flux boundaries for a 20-year simulation using the ground-water flow model of southern Utah and Goshen Valleys, Utah—Continued.

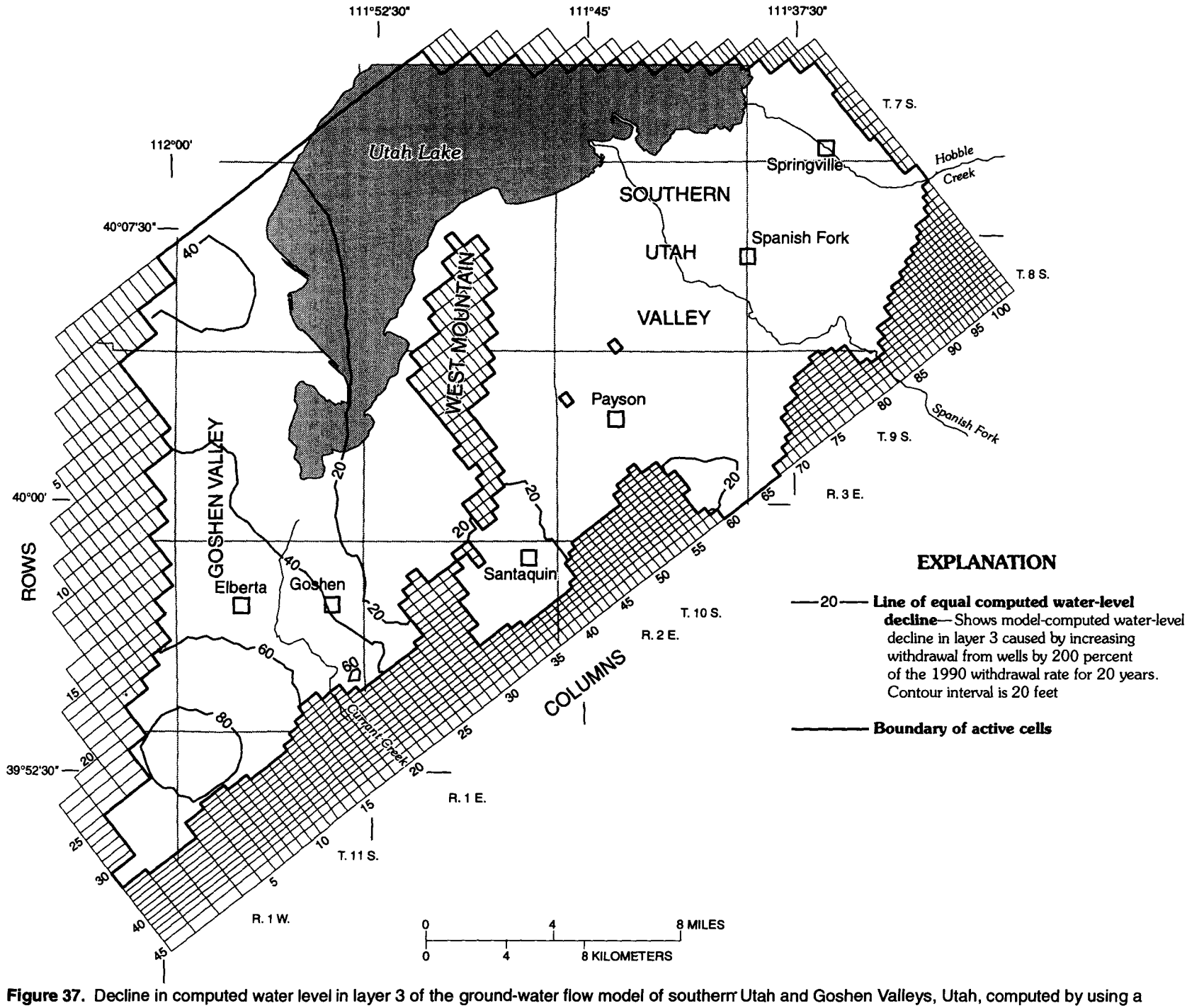


Figure 37. Decline in computed water level in layer 3 of the ground-water flow model of southern Utah and Goshen Valleys, Utah, computed by using a 20-year simulation of increasing withdrawal from wells by 200 percent of the 1990 withdrawal rate.

flowing wells were similar in 1964-67 and 1989-90, estimated flowing-well discharge decreased from 13,000 acre-ft in 1964 to 4,400 acre-ft in 1990, probably because of "sanding-in" of old wells, corrosion of well casings, and fewer uncontrolled flowing wells.

An estimated 9,800 acre-ft of water in southern Utah Valley and 3,400 acre-ft of water in Goshen Valley were released from storage in the main ground-water system from March 1990 to March 1991. This relates to an average water-level decline in southern Utah Valley of about 1.7 ft and in Goshen Valley of about 0.6 ft.

Ground water generally moves from the mountain fronts to Utah Lake. The altitude of the potentiometric contours in 1991 was similar to the altitude of the potentiometric contours in 1965. Ground-water level in southern Utah Valley fluctuates primarily in response to changes in precipitation. Ground-water level in Goshen Valley fluctuates only slightly in response to changes in precipitation; the slight fluctuation indicates that not much recharge occurs from precipitation on the valley floor. Seasonal water-level fluctuations indicate that irrigation is not a major source of recharge in either valley. In southern Utah Valley, water levels in March 1991 were not significantly different from water levels in March 1965. In Goshen Valley, water levels in March 1991 were higher than water levels in March 1965.

An analysis of samples collected from 1959 to 1991 indicates that ground-water quality has not changed substantially throughout southern Utah Valley. Most samples collected from wells finished in unconsolidated basin-fill deposits in southern Utah Valley had a dissolved-solids concentration of about 200 to 400 mg/L and contain calcium and magnesium as the major cations and bicarbonate as the major anion. From 1960 through 1991, no consistent water-quality trends were evident in Goshen Valley. Dissolved-solids concentration from samples collected in Goshen Valley ranges from 200 mg/L to as high as 2,600 mg/L, with lower concentrations generally occurring south of Elberta and higher concentrations generally occurring north of Elberta. Predominant cations are calcium and magnesium in the south and sodium in the north, and the predominant anion is bicarbonate in the south and chloride in the north.

A modular, three-dimensional, finite-difference ground-water flow model was used to simulate the main ground-water system in the unconsolidated basin-fill deposits of southern Utah and Goshen Valleys. The steady-state conditions of 1949 and annual transient-state conditions from 1949 to 1990 were used to calibrate the model. The simulated ground-water system includes the upper 1,000 ft of saturated unconsolidated basin-fill deposits and is vertically discretized into five layers of various thicknesses.

Head-dependent and specified-flux boundaries are used to simulate recharge to and discharge from the ground-water system. The calibration parameters adjusted during the calibration process are horizontal hydraulic conductivity, the amount and spatial distribution of recharge from injection wells to simulate subsurface inflow, vertical hydraulic conductivity, drain conductance and altitude, maximum evapotranspiration rate, hydraulic conductivity of the streambed, specific storage, specific yield, and Manning's roughness coefficient of the streambed. Horizontal hydraulic conductivity determined during model calibration ranges from 1 ft/d to 80 ft/d. Vertical hydraulic conductivity determined during model calibration ranges from 0.001 ft/d to 1 ft/d. Drain-conductance values range from about 275 ft^2/d to 55,000 ft^2/d. Drain altitudes range from 8 to 15 ft below land surface. The final distribution of hydraulic conductivity of the streambed ranges from 1 ft/d to 20 ft/d. A specific yield of 0.05 was assigned to all model layers, and a specific storage of 5×10^{-7}/ft was assigned to all active cells in model layers 2, 3, 4, and 5. The final value for Manning's roughness coefficient is 0.03 for all streams and canals.

To simulate steady-state conditions, the ground-water flow model was calibrated to the hydrologic conditions of the main ground-water system for 1949. Those conditions include estimated annual recharge and discharge for 1949 and water levels during March 1950. Individual budget elements where differences between model-computed and estimated amounts of flow are greater than 10 percent of the total estimated ground-water flow are recharge from subsurface inflow and discharge to evapotranspiration. The ground-water flow model was calibrated to 1949-90 transient-state conditions primarily by comparing model-computed water-level fluctuations to measured water-level fluctuations. The calibration also is based on matching measured water levels for March 1965 and March 1991 and comparison of model-computed and estimated ground-water budgets for 1964 and 1990.

At head-dependent flux boundaries, stream and canal flows for Hobble Creek, Spanish Fork, Currant Creek, South Field Canal, Mill Race Canal, and South Ditch, and the water level of Utah Lake are varied for

each stress period. At specified-flux boundaries, ground-water withdrawals from wells, and recharge from perennial streams and major canals, intermittent and ephemeral runoff, and irrigation and precipitation, are varied for each stress period.

Model-computed water levels, recharge, and discharge are sensitive to changes in vertical hydraulic conductivity in discharge areas and to changes in horizontal hydraulic conductivity near areas of large ground-water withdrawals and near the mountains. Water-level fluctuations in the upper model layers are sensitive to changes in specific yield of model layer 1.

Model limitations include the effects of stresses near the mountains, and model-computed recharge and discharge along specific reaches of Spanish Fork. Model-computed water-level fluctuations in Goshen Valley are not in close agreement with measured water-level fluctuations, and this calibration of the ground-water model reflects the uncertainties in the conceptual understanding of the ground-water system in Goshen Valley.

This ground-water model was used to project the effects of increased withdrawals from wells. Model-computed water-level declines of 20 ft in southern Utah Valley and 40 to 80 ft are projected for areas of Goshen Valley as a result of increasing withdrawal from wells by 200 percent of the 1990 withdrawal rate.

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