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HYDROLOGY OF CACHE VALLEY, CACHE COUNTY, UTAH, AND ADJACENT PART OF IDAHO, WITH EMPHASIS ON SIMULATION OF GROUND-WATER FLOW

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STATE OF UTAH
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CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATED WATER-QUALITY UNITS

Multiply	By	To obtain
acre	0.4047	hectometer
	4,047	square meter
acre-foot	0.001233	cubic hectometer
	1,233	cubic meter
acre-foot per year	0.00003907	cubic meter per second
cubic foot per second	0.02832	cubic meter per second
foot	0.3048	meter
foot squared per day[1]	0.0929	meter squared per day
gallon per minute	0.063	liter per second
inch	25.4	millimeter
	0.0254	meter
mile	1.609	kilometer
square mile	2.59	square kilometer
yard	0.9144	meter

The unit acre-feet per year also is used in this report. To obtain acre-feet per year, divide cubic feet per second by 0.0013803.

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

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

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

Chemical concentration and water temperature are reported in metric units. Chemical concentration is reported in milligrams per liter. Milligrams per liter is a unit expressing the solute mass per unit volume (liter) of water. For concentrations less than 7,000 milligrams per liter, the numerical value is about the same as for concentrations in parts per million. Specific conductance is reported in microsiemens per centimeter at 25 degrees Celsius.

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

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ABSTRACT

A hydrologic investigation of Cache Valley was done to better understand the ground-water system in unconsolidated basin-fill deposits and the interaction between ground water and surface water. Ground-water recharge occurs by infiltration of precipitation and unconsumed irrigation water, seepage from canals and streams, and subsurface inflow from adjacent consolidated rock and adjacent unconsolidated basin-fill deposit ground-water systems. Ground-water discharge occurs as seepage to streams and reservoirs, spring discharge, evapotranspiration, and withdrawal from wells.

Water levels declined during 1984-90. Less-than-average precipitation during 1987-90 and increased pumping from irrigation and public-supply wells contributed to the declines.

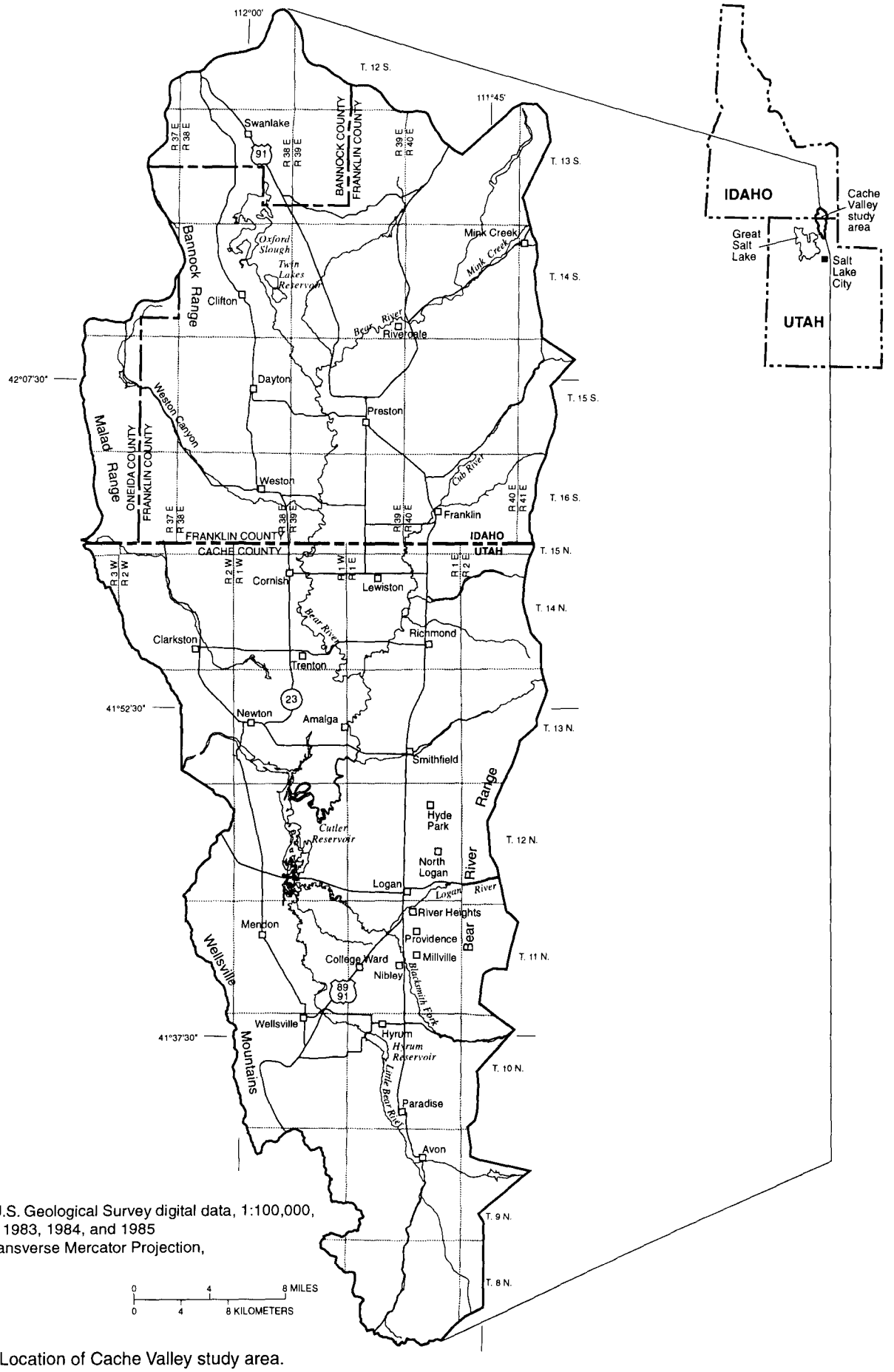
A ground-water-flow model was used to simulate flow in the unconsolidated basin-fill deposits. Data primarily from 1969 were used to calibrate the model to steady-state conditions. Transient-state calibration was done by simulating ground-water conditions on a yearly basis for 1982-90.

A hypothetical simulation in which the dry conditions of 1990 were continued for 5 years projected an average 10-foot water-level decline between Richmond and Hyrum. When increased pumpage was simulated by adding three well fields, each pumping 10 cubic feet per second, in the Logan, Smithfield, and College Ward areas, water-level declines greater than 10 feet were projected in most of the southeastern part of the valley and discharge from springs and seepage to streams and reservoirs decreased.

INTRODUCTION

Cache Valley is a north-south trending valley with an area of about 660 square miles in northeastern Utah and southeastern Idaho (fig. 1). The valley contains abundant water resources relative to other parts of Utah and supports a large agricultural industry. The Bear River, which is the largest tributary to Great Salt Lake, flows through Cache Valley. Several large tributaries to the Bear River, including Mink Creek, Cub River, Logan River, Blacksmith Fork, and Little Bear River, originate in the mountains adjacent to Cache Valley. Most of the irrigation water used in Cache Valley is diverted from streams or springs to a complex system of canals.

Since 1960, the populations of cities and towns in Cache Valley have increased steadily, particularly in those cities and towns in Utah between Richmond and Hyrum (fig. 1). Regions outside Cache Valley, including Box Elder County to the west, and the Wasatch Front, which extends from Brigham City, Utah, about 10 miles southwest of Cache Valley, to Payson, Utah, about 110 miles south, also have experienced population growth in recent years. The increased population in Cache Valley has resulted in increased use of ground water for public supply, and the increased population in Box Elder County and along the Wasatch Front has resulted in increased interest in the exportation of ground water and surface water from Cache Valley and the Bear River for use in these areas (Valley Engineering and Hansen, Allen, and Luce, 1990; Utah Division of Water Resources, 1991, p. 6-15).



Base from U.S. Geological Survey digital data, 1:100,000, 1978, 1979, 1983, 1984, and 1985
Universal Transverse Mercator Projection, Zone 12

Figure 1. Location of Cache Valley study area.

Because of the interconnection of the surface-water and ground-water systems in Cache Valley, increased withdrawal of ground water could decrease the volume of ground-water discharge to the surface-water system and therefore is of concern to surface-water users. To address this concern, the U.S. Geological Survey, in cooperation with the Utah Division of Water Resources and the Utah Division of Water Rights, did a hydrologic study of Cache Valley during 1989-92. Data were collected during 1989-91 and used with data collected during previous studies to evaluate the ground-water resources of Cache Valley and how they interact with surface water. The systems for numbering hydrologic-data sites in this report are shown in figure 2.

Purpose and Scope

This report describes the results of a hydrologic study of Cache Valley. Most discussion about the hydrology is limited to the valley, although some information on the adjoining mountains also is presented. Selected hydrologic data collected by the U.S. Geological Survey during 1969-91 (Roark and Hanson, 1992) and other data from previous studies were used to interpret the ground-water hydrology of Cache Valley, with an emphasis on ground water in the unconsolidated basin-fill deposits. Data collection and related activities included an inventory of new wells, installation of shallow monitoring wells, measurement of water levels in wells, measurement of spring discharge, measurement of seepage in selected streams, an aquifer test, and measurement of pumpage from irrigation wells. Irrigation-diversion data were compiled in detail, and the amounts of water diverted and areas of application were determined to improve estimates from previous studies of recharge to ground water from unconsumed irrigation water. Data were not collected to refine or provide estimates of subsurface inflow from adjacent consolidated rock or from adjacent unconsolidated basin-fill ground-water systems.

The hydrologic data compiled for this study were used to develop a digital-computer model to simulate ground-water flow in the main part of the unconsolidated basin-fill deposits in Cache Valley. The model simulated ground-water movement under confined and unconfined conditions, withdrawal from wells, areal recharge (from infiltration of precipitation and unconsumed irrigation water), evapotranspiration, seepage to drains (including springs and sloughs), seepage from canals, seepage to and from streams, and subsurface inflow from adjacent consolidated rock and unconsolidated basin-fill ground-water systems. Following calibration of the model, two hypothetical simulations were run; one to project water-level changes if the dry conditions of 1990 that resulted in less-than-average recharge continued for 5 years, and the other to project the effects of increased withdrawal of water from wells during conditions of average recharge.

Previous Studies

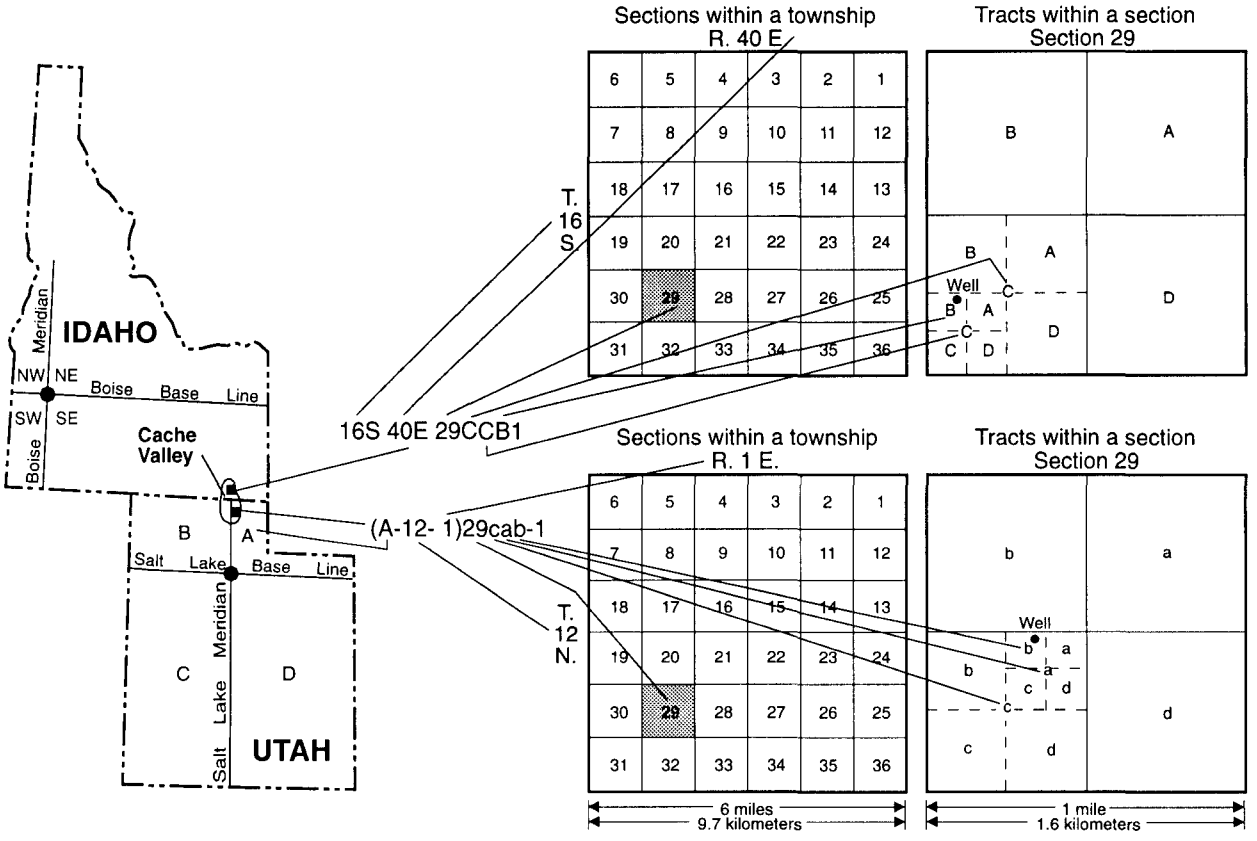
Several studies have been done on ground-water hydrology in Cache Valley. Bjorklund and McGreevy (1971) studied the ground-water resources of Cache Valley and published a separate data report (McGreevy and Bjorklund, 1970). Clyde, Jeppson, and Liu (1984) developed a predictive management model for ground water in the Utah part of Cache Valley. Additional studies addressing the ground water of Cache Valley were done by Dion (1969), Beer (1967), and Peterson (1946). Many studies concerning local ground-water conditions and drainage of lands were done by Utah State University (Israelsen, Milligan, and Bishop, 1955; Israelsen, 1954; Gardner and Israelsen, 1954; Maughan and Israelsen, 1951; Israelsen, Maughan, and South, 1946; and Israelsen and McLaughlin, 1932 and 1935). McGreevy and Bjorklund (1971) compiled geohydrologic sections of Cache Valley, Utah and Idaho.

The U.S. Geological Survey has measured streamflow in the Cache Valley drainage basin since 1889 (Wells, 1960) and water levels in selected wells since 1935 (McGreevy and Bjorklund, 1970, table 3). Surface-water records for 1889-1990 have been identified from past publications. The U.S. Geological Survey currently collects streamflow data at 15 sites: 3 sites in Idaho and 12 sites in Utah (ReMillard and others, 1992). The U.S. Geological Survey currently measures water levels on an annual or semiannual basis in a network of about 20 wells in the Utah part of Cache Valley and in 3 wells in the Idaho part (Nathan Jacobson, U.S. Geological Survey, oral commun., 1993).

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 northeast, northwest, southwest, and southeast quadrants, respectively. Numbers designating the township and range, in that order, follow the quadrant letter, and all three are enclosed in parentheses. The number after the parentheses indicates the section and is followed by three letters indicating the quarter section, the quarter-quarter section, and the quarter-quarter-quarter section—generally 10 acres for regular sections[1]. The lowercase letters, a, b, c, and d indicate, respectively, the northeast, northwest, southwest, and southeast quarters of each subdivision. The number after the letters is the serial number of the well or spring within the 10-acre tract. The letter S preceding the serial number designates a spring. The letter W following the serial number denotes a stream, and the letter B denotes a canal or ditch. Thus, (A-12-1)29cab-1 designates the first well constructed or visited in the northwest 1/4, northeast 1/4, southwest 1/4, section 29, T. 12 N., R. 1 E.

In Idaho, the well-and spring-numbering system is based on the cadastral land-survey system with reference to the Boise Base Line and Boise Meridian. It is similar to the Utah system except that springs are not designated by the letter S and the quadrant letter is omitted, the townships are labeled N or S to designate north or south, and the ranges are labeled E or W to designate east or west. The letter W following the serial number denotes a stream, and the letter B denotes a canal or ditch. Thus, 16S 40E 29CCB1 is the first well visited in the northwest 1/4, southwest 1/4, southwest 1/4, section 29, T. 16S., R. 40 E.

Streamflow sites where data were collected are numbered in sequential downstream order for this report. In addition, an 8-digit number has been assigned to gaging stations operated by the U.S. Geological Survey. Data from these stations and an explanation of the numbering system can be found in annual water-resources data reports for Utah (ReMillard and others, 1992).



[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 southeast corner, and the surplus or shortage is taken up in the tracts along the north and west sides of the section.

Figure 2. Numbering system used in this report for hydrologic-data sites in Utah and Idaho.

Data on the chemical quality of ground and surface water in Cache County, Utah, were compiled by Conner, Mitchell, and others (1958). Surface-water quality in the valley was evaluated by Waddell (1970) and Waddell and Price (1971) as part of a study of the Bear River and its tributaries.

Many studies of the geology of the area have been done. Keller (1952) mapped the rocks of Tertiary age in the Mink Creek, Idaho, area. Adamson, Hardy, and Williams (1955) discussed the rocks of Tertiary age of Cache Valley. A geologic map, discussion of the geology, and list of springs in the mountains of Cache County, Utah, was presented by Williams (1958a, 1958b). Williams (1962) discussed the geology of the Utah part of Cache Valley in detail and presented a geologic map of the valley. Most of the Idaho part of the area is included on a geologic map by Oriel and Platt (1968). McCalpin (1989) mapped the surficial geology of the East Cache Fault Zone in Cache County, Utah. Geologic maps of 7 1/2-minute quadrangles are available for Cutler Dam Quadrangle, Box Elder and Cache Counties, Utah (Oviatt, 1986); Porcupine Reservoir Quadrangle, Cache County, Utah (Berry, 1989); Smithfield Quadrangle, Cache County, Utah (Lowe and Galloway, 1989); and Richmond Quadrangle, Cache County, Utah (Brummer and McCalpin, 1990).

Geophysical studies in Cache Valley include gravity surveys by Mabey, Peterson, and Wilson (1974), Peterson and Oriel (1970), and Cook and others (1966); an integrated geophysical study by Stanley (1971); and a gravity and magnetic survey of the Mount Naomi Roadless Area, Cache County, Utah, and Franklin County, Idaho, by Mabey (1985). DeVries (1982) studied low-temperature geothermal potential in Cache Valley. The [U.S.] Bureau of Reclamation completed site-specific engineering studies for the Hyrum Reservoir project (Grundvig, 1989) and for the Newton Reservoir project (Heath, 1983).

A summary of irrigation conveyance systems in Cache Valley is included in an inventory by the [U.S.] Soil Conservation Service (1976). Data on irrigation companies in Utah were compiled by Whetstone (1990). Data on soils are available from the [U.S.] Soil Conservation Service (Erickson and Mortensen, 1974) and from the Utah Agricultural Experiment Station at Utah State University (Southard, Wilson, and Erickson, 1978).

Acknowledgments

The authors thank the many well owners and landowners in Cache Valley who permitted the collection of hydrologic data on their property. The authors also are grateful for the data and cooperation from officials of Cache County, Utah; Franklin County, Idaho; and Providence, Cornish, North Logan, Lewiston, River Heights, and Logan, Utah. The following irrigation companies provided information on wells or on canal diversions: Logan, Hyde Park, and Smithfield Canal Company; Richmond Irrigation Company; Benson Irrigation Company; Cub River Irrigation Company; Logan Island Irrigation Company; and Smithfield Irrigation Company.

Data and assistance also were provided by the Idaho Department of Water Resources; the [U.S.] Soil Conservation Service; University of Idaho, Department of Agricultural Engineering; Utah State University, Agricultural and Irrigation Engineering Department; Utah Power and Light Company; the Bear River Commission Engineer–Manager; and Wright Drilling Company. The authors are grateful to Robert Fotheringham, Logan Area Engineer of the Utah Division of Water Rights, for assistance with well, spring, and canal diversion data, and to Eugene Bigler and James Stephens of the Utah Division of Water Resources for information on land use and irrigation practices in Cache Valley.

Description of the Study Area

The study area (fig. 1) covers about 1,100 square miles and includes both Cache Valley and the adjoining mountains. Cache Valley is about 70 miles long and about 16 miles wide at its widest point and covers about 660 square miles of the 1,100-square-mile study area. Of the total area of Cache Valley, 54 percent is in Cache County, Utah, 38 percent is in Franklin County, Idaho, 5 percent is in Bannock County, Idaho, and 3 percent is in Oneida County, Idaho. The altitude of the valley floor ranges from about 4,400 to 5,400 feet.

Cache Valley is bounded on the west by the Bannock Range, the southern extension of the Malad Range, and the Wellsville Mountains. The altitude of the Wellsville Mountains rises abruptly to more than 9,000 feet. The Bannock Range also reaches altitudes above 9,000 feet. The Malad Range reaches altitudes of 6,000 to 7,000 feet where it bounds Cache Valley. Cache Valley is bounded on the east by the Bear River Range, which rises in altitude from 8,000 to more than 9,000 feet.

Geology

Cache Valley is in the northeast corner of the Basin and Range physiographic province described by Fenneman (1931). Although Cache Valley also has been considered to be part of a transitional zone between the Basin and Range and the Middle Rocky Mountain provinces (Stokes, 1986), this report considers Cache Valley to be part of the Basin and Range province. Cache Valley is underlain by a narrow, elongate graben formed by normal high-angle faults similar to those that bound other basins of the Basin and Range province (Fenneman, 1931). Sediments of Tertiary and younger age filled the valley contemporaneously with fault displacement. The generalized geology of Cache Valley is shown in figure 3.

Sedimentary and metamorphic rocks of Permian to Precambrian age, including limestone, dolomite, quartzite, sandstone, mudstone, siltstone, and shale, compose the mountain blocks surrounding Cache Valley and probably underlie younger basin-fill deposits. These rocks are the source of most of the detrital material that makes up the rocks and deposits of Tertiary and Quaternary age in the valley.

Rocks of Tertiary age in Cache Valley include the poorly cemented to well-cemented conglomerate, sandstone, and limestone of the Wasatch Formation (Bjorklund and McGreevy, 1971, table 4) and the poorly consolidated rocks of the Salt Lake Formation (Williams, 1962, p. 133). The Salt Lake Formation is exposed in an almost continuous belt in the foothills around the valley and consists of a lower conglomerate unit, a tuff unit, and an upper conglomerate and sandstone unit (Williams, 1962, p. 133-135). Williams (1962, fig. 39) reported that in a deep well near Amalga, Utah, the Salt Lake Formation was at a depth of about 450 feet and interpreted these rocks to underlie younger unconsolidated basin-fill deposits of Quaternary age and overlie consolidated rocks of Paleozoic age in the middle of the valley.

The floor of Cache Valley consists mostly of unconsolidated basin-fill deposits of Quaternary age from former Lake Bonneville and older lakes, and younger alluvium. Alluvial-fan and landslide deposits of Quaternary pre-Lake Bonneville age are exposed locally at the margins of the valley. In the valley interior, at least several hundred feet of fluvial and lacustrine sediments of Quaternary age underlie Lake Bonneville deposits and overlie the Salt Lake Formation (Williams, 1962, p. 135).

Sediments deposited by Lake Bonneville include the Alpine and Bonneville Formations, which consist mostly of silt with some gravel (Williams, 1962, p. 131), and the overlying Provo Formation, which consists of intertonguing layers of gravel, sand, silt, and clay (Williams, 1962, p. 140). Gravel and sand of Lake Bonneville age were deposited as shore embankments, deltas, bars, and spits near the mountain fronts while silt and clay settled from suspension in the lake water in the lower altitudes of Cache Valley (Williams, 1962, p. 137-142). The Bear River delta (Bjorklund and McGreevy, 1971, p. 15, and Gilbert, 1890), the largest delta in Cache Valley, extends from where the Bear River enters the valley to just north of the Utah-Idaho state line.

Numerous north-south trending faults have broken the graben beneath Cache Valley and resulted in unevenness in the basin floor (Williams, 1958b, p. 76). Williams (1958b, p. 77) describes several foothill benches at the margins of the graben in the Utah part of Cache Valley where consolidated rocks are shallower than in the center of the graben. Largest of these is Clarkston Bench (Williams, 1958b, fig. 1), which is located beneath the area around Clarkston, Utah. The area is topographically higher than much of the rest of Cache Valley. Normal faults separate Clarkston Bench from a horst to the west and the main graben beneath Cache Valley to the east (Williams, 1958b, p. 76-77). Because consolidated rock is shallower here than in the center of Cache Valley, unconsolidated basin-fill deposits are thinner.

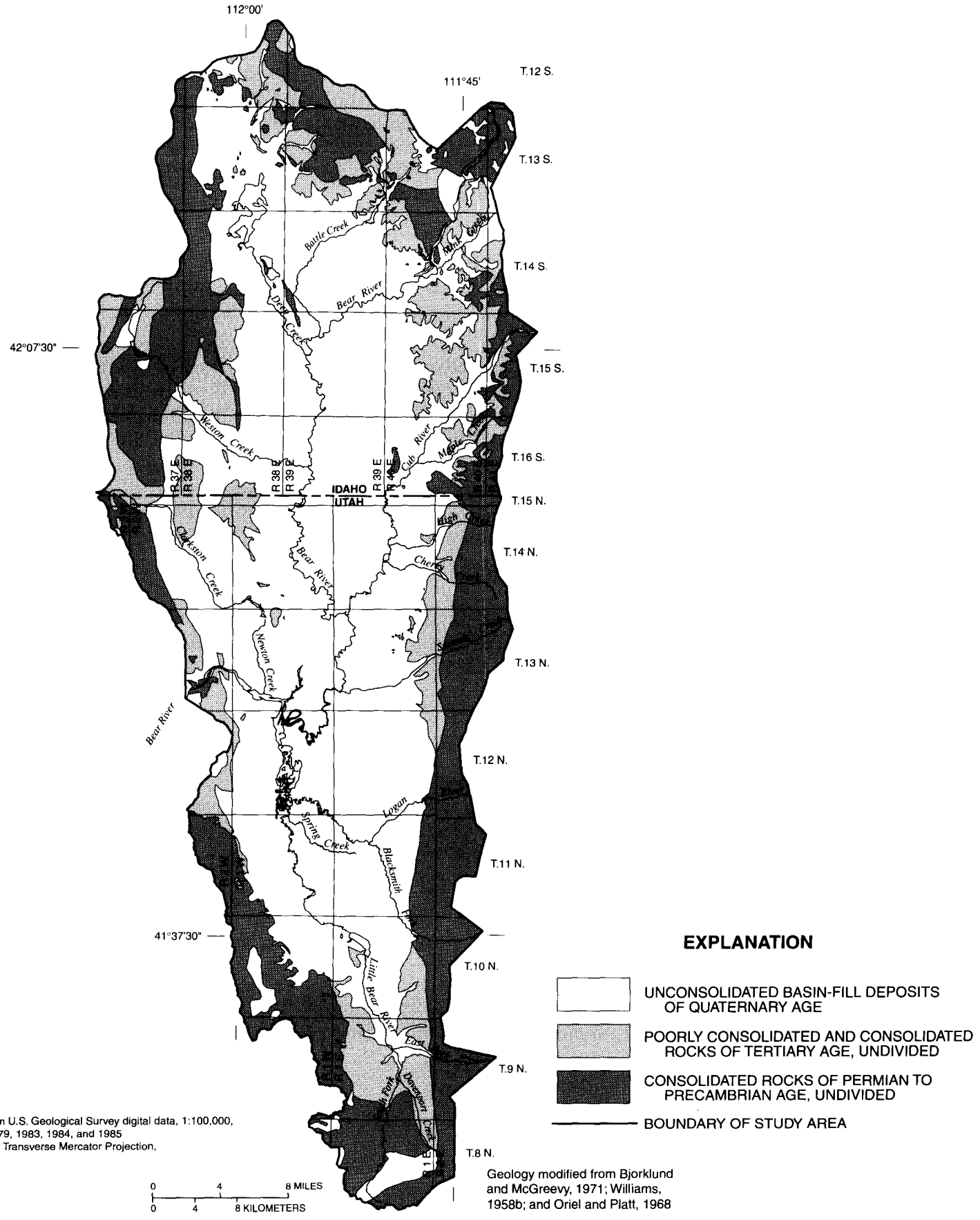


Figure 3. Generalized geology of Cache Valley.

Directly north of the Clarkston area and across the Utah-Idaho state line is Weston Canyon. Although the geologic mapping by Williams (1958b) did not extend into the Idaho part of Cache Valley, the geologic map by Bjorklund and McGreevy (1971, pl. 1) indicates that the same normal faults that define Clarkston Bench probably extend northward into Idaho and separate the Weston Canyon area from the horst to the west and the main graben beneath Cache Valley to the east. As in the Clarkston area, depth to consolidated rock is shallower than in the center of Cache Valley; consequently, unconsolidated basin-fill deposits are thinner.

Climate

Mean annual air temperature during 1951-80 at Utah State University in Logan, Utah, was 48 °F (National Oceanic and Atmospheric Administration, 1990). Winter temperatures are commonly below 0 °F, and summer temperatures rarely exceed 100 °F.

Cumulative departure from average annual precipitation and total annual precipitation at Utah State University in Logan during 1941-90 (Herbert, Gates, and others, 1991) are shown in figure 4. Normal annual precipitation during 1931-60 is estimated to be 15 to 20 inches in the valley and 20 to 50 inches in the surrounding mountains from maps produced by the [U.S.] Weather Bureau (1963) and Idaho Water Resources Research Institute (1980).

Ten years of less-than-average precipitation during 1952-62 produced the driest conditions in Cache Valley since 1941. Approximately average precipitation occurred during 1963-79. The wettest conditions since 1941 occurred during 7 successive years of greater-than-average precipitation during 1980-86. The highest annual precipitation recorded at Utah State University in Logan since records were first kept in 1893 was 35.85 inches, recorded in 1983 (Dawn Drost, Utah Climate Center, oral commun., 1992). Beginning in 1987, precipitation was less than average for 4 successive years.

Snow usually covers the valley floor during December, January, and February. The growing season ordinarily lasts about 159 days from May through October (Eubank and Brough, 1979). Estimated annual freshwater evaporation (1931-70) at Utah State University in Logan is about 40 inches (Norman Stauffer, Utah Division of Water Resources, oral commun., 1993).

Population and Land Use

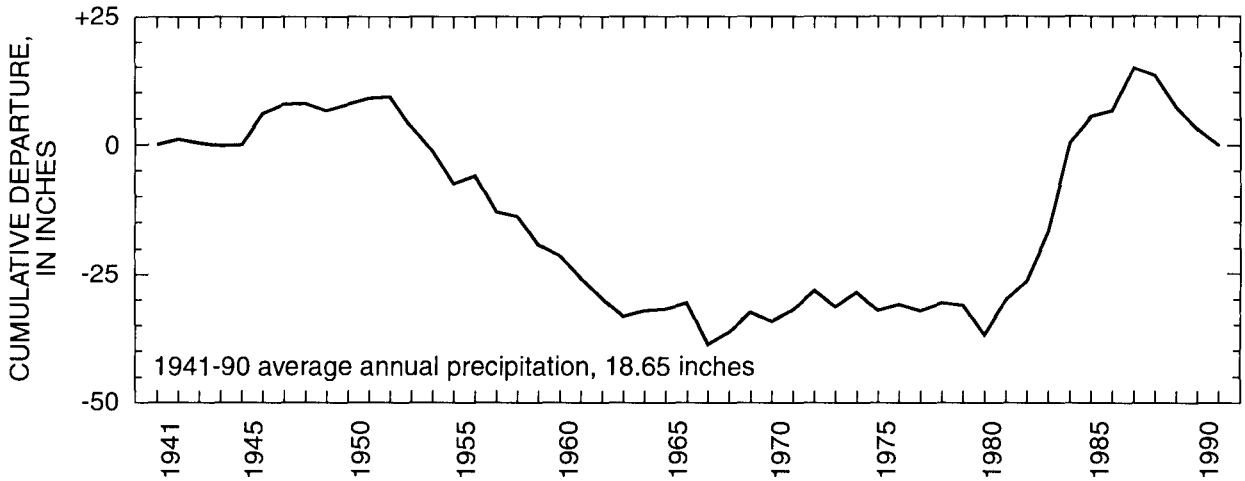
About half the people who lived in Cache Valley in 1990 resided in Logan, Utah (population 32,762, Utah Office of Planning and Budget, 1991), the principal business center in Cache Valley. The largest population center in the Idaho part of Cache Valley is Preston (population 3,710, Idaho Department of Commerce, 1992). Of the rural population, about 88 percent resided in Cache County, Utah, 10 percent in Franklin County, Idaho, and the remaining 2 percent in Bannock and Oneida Counties, Idaho. Although manufacturing is the principal industry in Cache Valley, agriculture accounts for most of the water use and is an important part of the overall economy (Utah Division of Water Resources, 1992).

HYDROLOGY

Most surface water in Cache Valley originates outside the valley and flows into the valley in major streams. Ground water in Cache Valley occurs principally in consolidated and poorly consolidated rocks and unconsolidated basin-fill deposits.

Surface Water

Surface water originates from either flow into the valley in the Bear River from Gem Valley, Idaho, about 7 miles north of the study area; flow into the valley from the surrounding mountains; or flow from springs and seeps inside the valley itself. Except for the Bear River, perennial streams that enter Cache Valley originate in the Bear River Range. All major streams that enter Cache Valley are regulated.



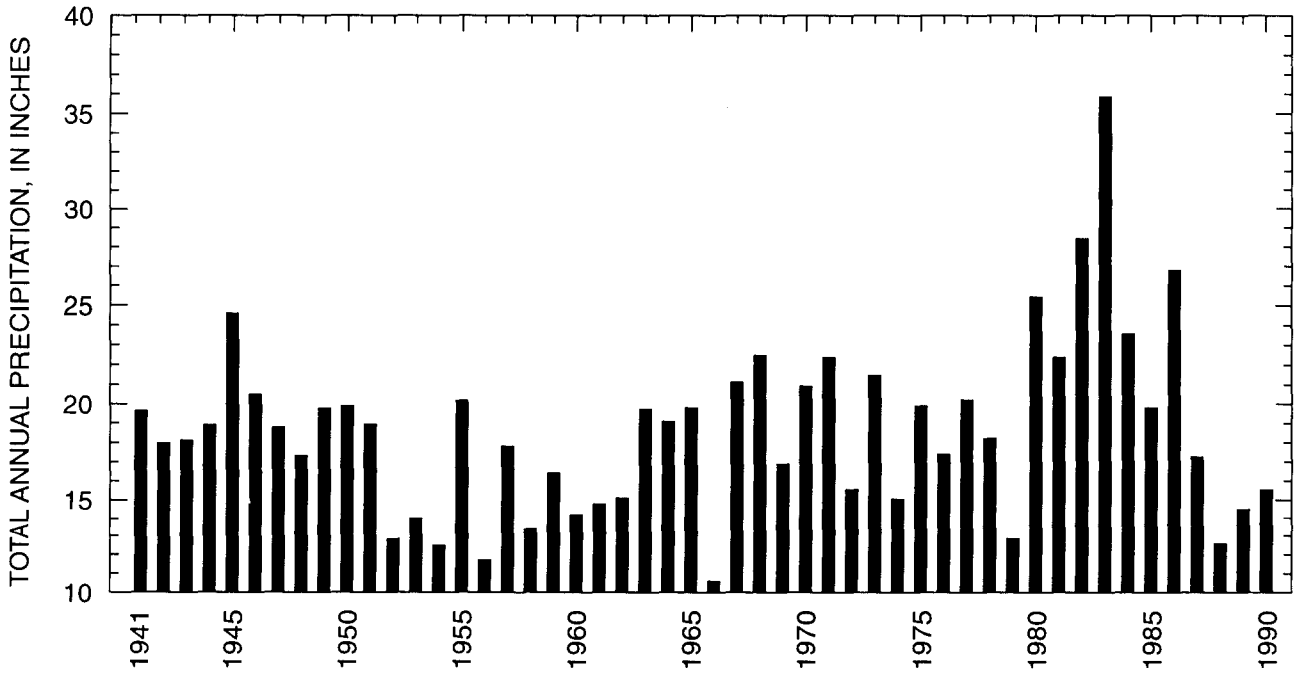


Figure 4. Cumulative departure from average annual precipitation and total annual precipitation at Utah State University in Logan, 1941-90.

Surface water is the primary source of irrigation water in Cache Valley and also is used for recreation, aquaculture, and public supply. Streamflow has been measured by the U.S. Geological Survey, Utah Power and Light Company, and water commissioners and irrigation-company officials in Cache Valley. Although long-term records are available for some sites, many sites have records that are discontinuous and may represent time periods as short as 1 year. All flow is reported by water year (October 1 to September 30) unless otherwise mentioned. The base period used was 1960-90, and annual mean flow in major rivers, streams, and canals in Cache Valley during this base period is listed in table 1. Locations of selected streamflow-gaging stations are shown in figure 5.

Although streamflow in each major stream that enters the valley has been measured at one time or another during past years, the periods of record for all the major streams do not coincide. Short-term streamflow records were extended using the simple linear regression method described by Thomas (1967), and ungaged streamflow was estimated using the method described later in this section of this report. Total mean annual streamflow entering Cache Valley for 1960-90 was about 1,751 cubic feet per second (1,268,600 acre-feet per year).

Surface water leaves Cache Valley through the Bear River, West Side Canal, and Hammond Main Canal, all of which flow from Cutler Reservoir. Total mean annual surface-water outflow from Cache Valley for 1960-90 was about 1,959 cubic feet per second (1,419,300 acre-feet per year).

Mean annual surface-water outflow for 1960-90 was about 210 cubic feet per second (152,100 acre-feet per year) greater than inflow. The difference is not significant because of potential errors involved in the use of estimating techniques and in the accuracy of streamflow measurements.

Streams

The largest stream in the study area is the Bear River, which enters the northern end of Cache Valley from Oneida Narrows Reservoir in Idaho and exits through Cutler Reservoir in Utah. Where the Bear River enters Cache Valley, mean annual flow for 1960-90 was 1,023 cubic feet per second (741,100 acre-feet per year) (table 1) at Utah Power and Light Company streamflow-gaging station 10086500, Bear River below Utah Power and Light Company's tailrace, at Oneida, Idaho (fig. 5). At the Utah-Idaho state line, mean annual flow of the Bear River was about 1,124 cubic feet per second (814,300 acre-feet per year) for 1960-90 at U.S. Geological Survey streamflow-gaging station 10092700, Bear River at Idaho-Utah state line. Flow during 1960-70 was estimated using a linear relation provided by Norman Stauffer (Utah Division of Water Resources, written commun., 1993). The equation describing the relation and the error of the regression are shown in table 2. Where the Bear River leaves Cache Valley, mean annual flow for 1960-90 was 1,628 cubic feet per second (1,179,500 acre-feet per year) (table 1) at Utah Power and Light Company streamflow-gaging station 10118000, Bear River near Collinston, Utah.

The Bear River flows through a deep cut in the unconsolidated basin-fill deposits throughout most of Cache Valley. This cut is deepest in the thick deposits of the Bear River delta in the north end of the valley and ranges in depth from a maximum of about 450 feet near Riverdale, Idaho, to a minimum of about 50 feet near Benson, Utah.

The Logan River drains about 214 square miles in the adjacent Bear River Range. Where the Logan River enters Cache Valley, mean annual flow for 1960-90 was 257 cubic feet per second (186,200 acre-feet per year) (table 1). This value is the summed flow based on the combined records for several U.S. Geological Survey streamflow-gaging stations, including stations 10109000, Logan River above State Dam, near Logan, Utah; 10108500, Logan, Hyde Park, and Smithfield Canal near Logan, Utah (1960-62); 10108400, Logan, Hyde Park, and Smithfield Canal at head, near Logan, Utah (1963-90); and 10108900, Utah Power and Light Company's tailrace, near Logan, Utah. Streamflow-gaging station 10108900 was located about 100 yards downstream from streamflow-gaging station 10109000 and was discontinued in 1970 after the Utah Power and Light Company power plant at the mouth of Logan Canyon ceased operation. The Logan, Hyde Park, and Smithfield Canal diversion is about 1/4 mile upstream from streamflow-gaging station 10109000.

The Logan River channel is incised from as much as 200 feet into the surrounding basin-fill deposits near the mouth of Logan Canyon to as little as a few feet near U.S. Highway 89. Downstream from the confluence with the Blacksmith Fork, the Logan River flows into the south end of Cutler Reservoir. Mean annual flow of the Logan

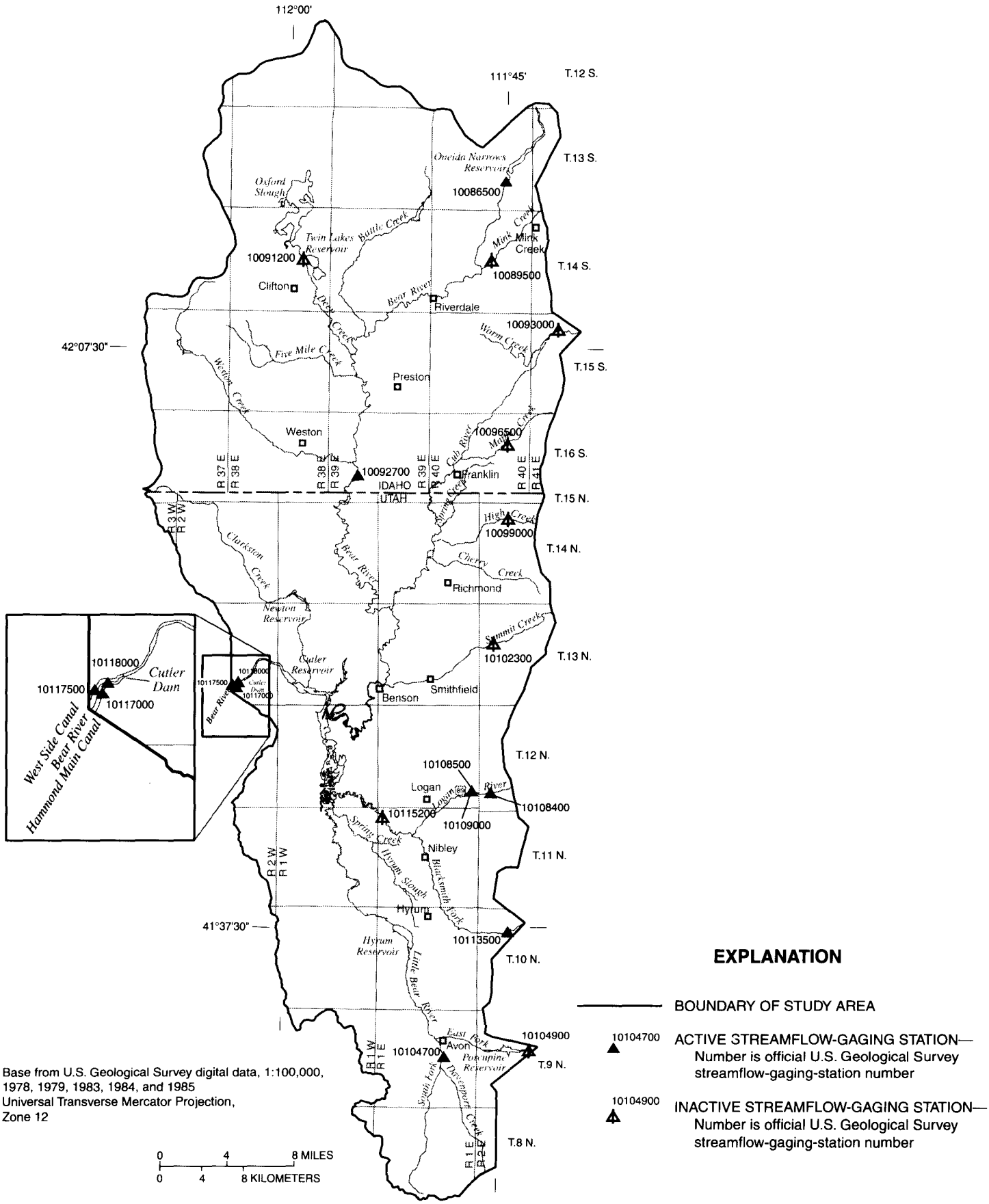


Figure 5. Location of major streams, reservoirs, and streamflow-gaging stations in Cache Valley.

Table 1. Annual mean flow of rivers, streams, and canals in Cache Valley, 1960-90 water years

[Flow in cubic feet per second; e, estimated]

River, stream, or canal name[1]	Water year 1960	1961	1962	1963	1964	1965	1966	1967	1968	1969	1970	1971	1972	1973	1974
Entering Cache Valley															
Bear River[2]	630	562	618	613	619	835	1,119	741	753	904	632	1,452	1,627	1,201	1,189
Logan River[3]	192	132	236	201	219	318	212	262	237	249	246	407	379	230	313
Blacksmith Fork	92	61	117	89	102	157	114	140	118	132	111	223	236	149	175
Cub River	73	59	82	75	79	100	77	88	83	85	84	120	114	81	99
Little Bear River	41 e	24	58	38	46	57	46	64	50	69	49	78	73	69	74
Mink Creek	34 e	21 e	43 e	36 e	40 e	61 e	38 e	49 e	43 e	46 e	45 e	79 e	73 e	42 e	59 e
East Fork Little Bear	12 e	7 e	16 e	11 e	30	39	26	40	32	40	28	61	46	31	43
High Creek	23 e	16 e	27 e	24 e	26 e	36 e	25 e	30 e	27 e	29 e	28 e	46 e	36	27 e	36 e
Summit Creek	15 e	11 e	18	13	21	24	14	20	18	19	20	29	24	17	24
Maple Creek	14 e	9 e	17	14 e	16 e	24 e	15 e	19 e	17 e	18 e	18 e	31 e	29 e	17 e	23 e
Ungaged streams	25 e	18 e	31 e	27 e	29 e	42 e	28 e	35 e	31 e	33 e	32 e	54 e	50 e	30 e	42 e
Total, entering	1,151 e	920 e	1,263 e	1,141 e	1,227 e	1,693 e	1,714 e	1,488 e	1,409 e	1,624 e	1,293 e	2,580 e	2,687 e	1,894 e	2,077 e
Leaving Cache Valley															
Bear River[4]	730	502	1,117	796	1,147	1,483	1,406	1,301	1,205	1,445	1,006	2,655	2,546	1,809	1,936
West Side Canal	295	276	237	254	227	267	303	243	267	301	283	262	297	269	312
Hammond Main Canal	63	60	51	51	48	57	64	48	52	57	56	48	56	51	59
Total, leaving	1,088	838	1,405	1,101	1,422	1,807	1,773	1,592	1,524	1,803	1,345	2,965	2,899	2,129	2,307
Total, entering-leaving	63 e	82 e	-142 e	40 e	-195 e	-114 e	-59 e	-104 e	-115 e	-179 e	-52 e	-385 e	-212 e	-235 e	-230 e
Within Cache Valley															
Bear River[5]	671 e	593 e	657 e	652 e	659 e	907 e	1,233 e	798 e	812 e	986 e	674 e	1,793	1,780	1,241	1,335
Deep Creek	5 e	1 e	8 e	5 e	6 e	13 e	8 e	11 e	8 e	10 e	7 e	21 e	22 e	12 e	15 e
Logan River[6]	188 e	100 e	253 e	203 e	229 e	349	199	288	250	252	244	529	478	274	355

[1]River, stream, or canal name:

[2]Bear River - Bear River below Utah Power and Light Company's tailrace, at Oneida, Idaho.

[3]Logan River - Logan River above State Dam, near Logan, Utah.

Blacksmith Fork - Blacksmith Fork above Utah Power and Light Company's dam, near Hyrum, Utah.

Cub River - Cub River near Preston, Idaho.

Little Bear River - Little Bear River below Davenport Creek, near Avon, Utah.

Mink Creek - Mink Creek near Mink Creek, Idaho.

East Fork Little Bear - East Fork Little Bear River above reservoir, near Avon, Utah.

High Creek - High Creek near Richmond, Utah.

Summit Creek - Summit Creek above diversions, near Smithfield, Utah.

Maple Creek - Maple Creek near Franklin, Idaho.

[4]Bear River - Bear River near Collinston, Utah.

West Side Canal - Utah Power and Light Company station, records kept by U.S. Geological Survey as West Side Canal, near Collinston, Utah.

Hammond Main Canal - Utah Power and Light Company station, records kept by U.S. Geological Survey as Hammond Main Canal, near Collinston, Utah.

[5]Bear River - Bear River at Idaho - Utah state line.

[6]Logan River - Logan River below Blacksmith Fork.

Water year																
1975	1976	1977	1978	1979	1980	1981	1982	1983	1984	1985	1986	1987	1988	1989	1990	Mean
Entering Cache Valley																
987	1,209	774	613	592	965	822	888	2,066	2,586	1,628	2,163	1,168	624	558	561	1,023
303	265	120	271	213	274	175	342	365	441	297	440	172	138	194	119	257
161	145	71	119	109	149	86	180	212	295	196	268	122	80	83	60	140
97	89	57	90	77	91	69	105	110	127	96	131	75 e	68 e	80 e	64 e	88
70	56	25	60	46	69	34	77	100	106	73	119	47	30	44	28	59 e
57 e	49 e	19 e	51 e	38 e	51 e	30 e	66 e	70 e	87 e	56 e	86 e	30 e	22 e	34 e	18 e	48 e
42	31	9	41	27	48	14	53	66	72	43	61	13 e	9 e	12 e	8 e	33 e
35 e	30 e	15 e	31 e	25	36	27	47	47	50	29	50	18	14	26	15 e	30 e
25	20	8	26	17	21 e	14 e	25 e	27 e	32 e	22 e	32 e	14 e	11 e	15 e	10 e	20 e
22 e	19 e	8 e	20 e	15 e	20 e	12 e	26 e	27 e	33 e	22 e	33 e	12 e	9 e	14 e	8 e	19 e
40 e	35 e	16 e	36 e	28 e	36 e	23 e	45 e	48 e	58 e	39 e	58 e	23 e	18 e	26 e	16 e	34 e
1,839 e	1,948 e	1,122 e	1,358 e	1,187 e	1,760 e	1,306 e	1,854 e	3,138 e	3,887 e	2,501 e	3,441 e	1,694 e	1,023 e	1,086 e	907 e	1,751 e
Leaving Cache Valley																
1,796	1,974	820	1,131	906	1,837	1,074	1,864	3,608	4,379	2,576	4,020	1,570	622	710	493	1,628
277	277	292	278	319	273	316	262	189	217	284	247	295	327	312	318	277
48	50	56	54	57	50	62	54	36	30	55	46	57	70	68	70	54
2,121	2,301	1,168	1,463	1,282	2,160	1,452	2,180	3,833	4,626	2,915	4,313	1,922	1,019	1,090	881	1,959
-282 e	-353 e	-46 e	-105 e	-95 e	-400 e	-146 e	-326 e	-695 e	-739 e	-414 e	-872 e	-228 e	4 e	-4 e	26 e	-208 e
Within Cache Valley																
1,111	1,401	780	797	687	1,122	874	918	2,485	2,728	1,772	2,490	1,231	566	573	532	1,124 e
13 e	11 e	2 e	8 e	7 e	12 e	4 e	16 e	20 e	29 e	18 e	26 e	9 e	3 e	4 e	1 e	11 e
364	319	130	275	216	320	163 e	409 e	443 e	556 e	344 e	583 e	185 e	140 e	230 e	118 e	290 e

Table 2. Regression equations used to extend streamflow records in Cache Valley and values of R^2 (error of regression)

River or stream	Equation (units in acre-feet per year)	R^2
Bear River[1,2]	-38,800.00 + 1.15 Bear[3]	0.99
Logan River[1,4]	-67,700.00 + 1.47 Logan[5]	.97
Cub River	28,783.61 + 0.206 Logan[5]	.81
Mink Creek	-5,006.02 + 0.212 Logan[5]	.69
East Fork Little Bear River	1,142.12 + 0.243 Blacksmith Fork[6]	.81
High Creek	1,525.13 + 0.107 Logan[5]	.80
Summit Creek	1,309.30 + 0.069 Logan[5]	.81
Maple Creek	-1,306.51 + 0.080 Logan[5]	.54
Deep Creek[1,7]	-4,480.00 + 0.121 Blacksmith Fork[6]	.99

[1] Regression equation provided by Norman Stauffer (Utah Division of Water Resources, written commun., 1993).

[2] Flow at streamflow-gaging station 10092700, Bear River at Idaho-Utah state line.

[3] Flow at streamflow-gaging station 10086500, Bear River below Utah Power and Light Company's tailrace, at Oneida, Idaho.

[4] Flow at streamflow-gaging station 10115200, Logan River below Blacksmith Fork, near Logan, Utah.

[5] Combined flow of Logan River at streamflow-gaging station 10109000, Logan River above State Dam, near Logan, Utah; and station 10108500, Logan, Hyde Park, and Smithfield Canal near Logan, Utah (1960-62); or station 10108400; Logan, Hyde Park, and Smithfield Canal at head, near Logan, Utah (1963-90).

[6] Flow at streamflow-gaging station 10113500, Blacksmith Fork above Utah Power and Light Company's dam, near Hyrum, Utah.

[7] Flow at streamflow-gaging station 10091200, Deep Creek near Clifton, Idaho.

River into Cutler Reservoir is about 290 cubic feet per second (210,100 acre-feet per year) for 1960-90 (table 1) at U.S. Geological Survey streamflow-gaging station 10115200, Logan River below Blacksmith Fork, near Logan, Utah (fig. 5). Extension of the streamflow record was done using a linear relation provided by Norman Stauffer (Utah Division of Water Resources, written commun., 1993). The equation describing the relation and the error of the regression are reported in table 2.

Blacksmith Fork drains about 268 square miles before it enters Cache Valley. Mean annual flow of Blacksmith Fork at U.S. Geological Survey streamflow-gaging station 10113500, Blacksmith Fork above Utah Power and Light Company's dam, near Hyrum, Utah (fig. 5), for 1960-90 was 140 cubic feet per second (101,400 acre-feet per year) (table 1). Blacksmith Fork is incised as much as 80 feet into the basin-fill deposits from the mouth of Blacksmith Fork Canyon to near Nibley, Utah.

The Cub River drains about 32 square miles in the Bear River Range. Mean annual flow of the Cub River at U.S. Geological Survey streamflow-gaging station 10093000, Cub River near Preston, Idaho (fig. 5), for 1960-90 is 88 cubic feet per second (63,800 acre-feet per year) (table 1). The 1987-90 estimate of streamflow for the Cub River was made by developing a mathematical relation, using simple linear regression, of the flow in the Cub River to the flow in the Logan River for years with concurrent record (1941-86). The equation describing the relation and the errors of the regression are reported in table 2.

The Little Bear River drains about 62 square miles before it enters Cache Valley from the south. Mean annual flow of the Little Bear River at U.S. Geological Survey streamflow-gaging station 10104700, Little Bear River below Davenport Creek, near Avon, Utah (fig. 5), for 1960-90 was 59 cubic feet per second (42,700 acre-feet per year)(table 1). The Little Bear River passes through an approximately 80-foot-deep cut in the surrounding basin-fill deposits and collects water from numerous springs before entering Hyrum Reservoir. After leaving Hyrum Reservoir, the river meanders through the valley before entering Cutler Reservoir from the south.

Mink Creek enters the northeast part of Cache Valley and drains about 58 square miles. Estimated mean annual flow of Mink Creek at U.S. Geological Survey streamflow-gaging station 10089500, Mink Creek near Mink Creek, Idaho (fig. 5), for 1960-90 is 48 cubic feet per second (34,800 acre-feet per year) (table 1). The 1960-90 estimate of streamflow was made by developing a mathematical relation, using simple linear regression, of the flow in Mink Creek to the flow in the Logan River for years with concurrent record (1943-51). The equation describing the relation is reported in table 2. The creek empties into the Bear River northeast of Riverdale, Idaho.

The East Fork Little Bear River drains about 57 square miles before entering Porcupine Reservoir. The river flows from the reservoir into Cache Valley and then empties into the Little Bear River just north of Avon, Utah. Estimated mean annual flow of the East Fork Little Bear River at U.S. Geological Survey streamflow-gaging station 10104900, East Fork Little Bear River above reservoir, near Avon, Utah (fig. 5), for 1960-90 is 33 cubic feet per second (23,900 acre-feet per year) (table 1). The 1987-90 estimate of streamflow was made by developing a mathematical relation, using simple linear regression, of the flow in the East Fork Little Bear River to the flow in Blacksmith Fork for years with concurrent record (1964-86). The equation describing the relation is reported in table 2.

High Creek drains about 16 square miles before entering Cache Valley. Mean annual flow of High Creek at U.S. Geological Survey streamflow-gaging station 10099000, High Creek near Richmond, Utah (fig. 5), for 1960-90 is 30 cubic feet per second (21,700 acre-feet per year) (table 1). The 1990 estimate of streamflow was made by developing a mathematical relation, using simple linear regression, of the flow in High Creek to the flow in the Logan River for years with concurrent record (1947-52, 1972-89). The equation describing the relation is reported in table 2. The creek empties into the Cub River northwest of Richmond, Utah. There is no flow in the downstream part of the creek during the later part of the irrigation season because all of the water has been diverted for irrigation.

Summit Creek enters the valley east of Smithfield, Utah, and drains about 15 square miles. Estimated mean annual flow of Summit Creek at U.S. Geological Survey streamflow-gaging station 10102300, Summit Creek above diversions, near Smithfield, Utah (fig. 5), for 1960-90 was 20 cubic feet per second (14,500 acre-feet per year) (table 1). The 1980-90 estimate of streamflow was made by developing a mathematical relation, using simple linear regression, of the flow in Summit Creek to the flow in the Logan River for years with concurrent record (1962-79). The equation describing the relation is reported in table 2. Most of the flow of Summit Creek is diverted for irrigation during the summer, and the rest empties into the Bear River, west of Smithfield.

Maple Creek enters Cache Valley northeast of Franklin, Idaho, and drains about 21 square miles. Mean annual flow of Maple Creek at U.S. Geological Survey streamflow-gaging station 10096500, Maple Creek near Franklin, Idaho (fig. 5), for 1960-90 is 19 cubic feet per second (13,800 acre-feet per year) (table 1). The 1960-90 estimate of streamflow was made by developing a mathematical relation, using simple linear regression, of the flow in Maple Creek to the flow in the Logan River for years with concurrent record (1947, 1949-52). The equation describing the relation is reported in table 2. The low R^2 value for this regression is a result of fewer concurrent years of data. The creek drains into the Cub River north of Franklin.

Many small, ungaged, perennial and ephemeral streams flow into Cache Valley from the surrounding mountains. Indirect methods based on channel geometry described in Hedman (1970) were used to estimate mean annual flow in many of the ungaged streams. Where no channel geometry estimates were made, the method described by Christensen and others (1986) was used to estimate mean flow. Total mean annual flow for the ungaged small streams is estimated to be about 34 cubic feet per second (24,600 acre-feet per year)(table 1). Estimated mean annual flow of ungaged streams that flow into Cache Valley is reported in table 3. Annual streamflow of ungaged streams for 1960-90 was estimated by using the equation:

$$\frac{L_y}{L_m} \times U_m = U_y \tag{1}$$

where: L_y is annual streamflow in Logan River,

L_m is mean annual streamflow in Logan River,

U_m is mean annual streamflow in ungaged streams, and

U_y is annual streamflow in ungaged streams.

Streams that originate inside the valley receive water from springs and seeps. The flow of very few of these streams has been measured continuously, although some have been measured periodically. Most of the streamflow measurements are published by McGreevy and Bjorklund (1970), Herbert and Thomas (1992), and Roark and Hanson (1992).

Deep Creek, which drains Oxford Slough, is one of the few gaged streams that originates in the valley. Mean annual flow of Deep Creek at U.S. Geological Survey streamflow-gaging station 10091200, Deep Creek near Clifton, Idaho (fig. 5), is about 11 cubic feet per second (8,000 acre-feet per year) during 1960-90 (table 1). Extended record was based on linear regression equations provided by Norman Stauffer (Utah Division of Water Resources, written commun., 1993).

Reservoirs

Six large and several small reservoirs in Cache Valley store water for irrigation and power generation. Most water in the larger reservoirs is diverted from streams that originate outside the valley.

In terms of surface area, Cutler Reservoir is the largest in Cache Valley and receives all surface water that leaves the valley (fig. 5). The reservoir is shallow, with a mean depth of 3 feet, and has a maximum surface area of 7,184 acres (Utah Department of Health, 1982). The reservoir stores a maximum of 15,386 acre-feet of water, which is used primarily for production of hydroelectric power at Cutler Dam (Utah Department of Health, 1982). Surface-water inflows into Cutler Reservoir are primarily from the Bear River, Logan River, and Little Bear River.

Cutler Reservoir receives substantial amounts of ground-water seepage because of its location at the lowest point in Cache Valley. A seepage study of the Bear River and Cutler Reservoir done in November 1990 indicated that Cutler Reservoir gained an estimated 79 cubic feet per second of ground-water seepage (Herbert and Thomas, 1992). An unknown but small amount of ungaged spring discharge may have contributed to this seepage amount.

Hyrum Reservoir is located on the Little Bear River in Utah at the southern end of Cache Valley (fig. 5). With a mean depth of 39 feet, Hyrum Reservoir is deeper than Cutler Reservoir and also has the largest capacity of any reservoir in the valley, with a maximum surface area of 438 acres and a maximum volume of 18,700 acre-feet ([U.S.] Bureau of Reclamation, 1981). Surface-water inflow to Hyrum Reservoir primarily is from the Little Bear River and four unnamed intermittent streams. Water in Hyrum Reservoir is stored for irrigation and stock watering. Soils that underlie Hyrum Reservoir have been described as "cobbly coarse sandy loams to gravelly, very cobbly, very fine sandy, cobbly silt, and sandy clay loams" with "moderate to moderately rapid permeability," and to a lesser extent "silty sand loams to silty clay loams" with "moderate to very slow permeability" (Utah Department of Health, 1982).

Of the remaining four larger reservoirs in Cache Valley, three store water for irrigation. Twin Lakes Reservoir (fig. 5), at the northern end of the valley in Idaho, is filled with water diverted from Mink Creek. This water flows through the Twin Lakes Canal for nearly 60 miles from the point of diversion to reach Twin Lakes Reservoir (pl. 1). In the southeast corner of the study area, Porcupine Reservoir (fig. 5) stores water from the East Fork Little Bear River for diversion into the Porcupine Highline Canal. On the west-central side of the valley, Newton Reservoir (fig. 5) stores water from Clarkston Creek for use in the canals and ditches of the Newton Water Users Association.

Table 3. Estimated mean annual flow of ungaged streams that flow into Cache Valley

Stream type: E, ephemeral; P, perennial.

Method of estimate: Ch, Christensen and others (1986); H, Hedman (1970); M, measured in 1944.

Site name	Stream type	Method of estimate	Mean annual flow (acre-feet per year)
Battle Creek	P	M	1,220
Bear Creek	E	Ch	94
Bird Canyon	P	H	2,260
Birch Creek (near town of Mink Creek)	P	H	360
Birch Creek (near Smithfield)	E	Ch	748
Cherry Creek	P	H	1,940
Chicken Creek	P	H	40
City Creek	E	Ch	397
Clarkston Creek	P	Ch	144
Clifton Creek	P	H	1,200
Crooked Creek	P	H	200
Davis Creek	P	Ch	80
Dry Canyon	E	H	250
Fourmile Creek	E	Ch	27
Fox Hollow	E	Ch	68
Gooseberry Creek	E	Ch	145
Green Canyon (near Logan)	E	H	960
Green Canyon (near Paradise)	E	Ch	307
Hyde Park	E	Ch	342
Hyrum Canyon	E	Ch	661
McMurdie Hollow	P	H	250
Michael Creek	P	H	430
Millville Creek	E	H	280
Nebo Creek	E	Ch	152
Oxford Creek	P	H	430
Pullum Hollow	E	H	500
Paradise Dry Canyon	E	Ch	1,116
Sardine Canyon	E	Ch	53
Spring Branch	P	Ch	182
Spring Creek (near Franklin)	P	Ch	392
Spring Creek (Providence Canyon)	P	H	3,690
Station Creek	P	H	1,120
Stockton Creek	E	Ch	558
Swan Lake Creek	E	Ch	381
Unnamed 1	E	Ch	16
Unnamed 2	E	H	120
Unnamed 3	E	H	70
Unnamed 4	P	H	120
Unnamed tributary	E	Ch	141
Wellsville Canyon	P	Ch	416
West Fork Creek	P	H	60
Weston Creek	P	H	1,710
Worm Creek	P	H	810
Total			24,440

Oneida Narrows Reservoir is located at the northern end of the valley where the Bear River enters Cache Valley. It has a surface area of 480 acres and a maximum capacity of 12,605 acre-feet (Ruddy and Hitt, 1990). Water is stored in Oneida Narrows Reservoir for production of hydroelectric power.

Canals

Eighty-two canal companies deliver water through more than 600 miles of canals to irrigated farmland in Cache Valley. Canals divert water for irrigation inside the valley from springs, the Bear River and other streams, and canals. Water is diverted for irrigation outside Cache Valley from major streams before they enter the valley. The locations of major canals in Cache Valley that were included in a reconnaissance inventory of irrigation conveyance systems in the Bear River Basin by the [U.S.] Soil Conservation Service (1976) are shown on plate 1.

The reconnaissance inventory was done in 1973-74 and indicates that in the Idaho part of Cache Valley, canal conveyance efficiencies ranged from 30 to 108 percent (conveyance efficiency greater than 100 percent implies that water seeps into the canal). In the Utah part of the valley, conveyance efficiencies were determined to range from 25 to 99 percent.

The amount of water diverted from major streams and reservoirs to major canals in Cache Valley during 1969 and 1982-90 is shown in table 4. The estimated mean amount of water diverted to major canals in Cache Valley is about 265,000 acre-feet each year. Unless indicated otherwise, table 4 lists the actual amounts of water diverted each year. For most canals that are overseen by water commissioners, records of canal diversions were available from the Utah Division of Water Rights and are shown in the table.

Diversion records and streamflow records were unavailable for the Newton Water Users Association ditches, the Spring Creek Irrigation Company ditches, and the Strong Arm Company ditches. The amount diverted was estimated by assuming that the amount applied in 1969 to the area served by the canal in question was applied at the same rate as for a reference canal that serves a nearby area (footnotes 4, 5, and 6 in table 4). A canal serving an adjacent area was chosen as a reference canal because crop types and irrigation practices in that canal were assumed to be similar to crop types and irrigation practices for the canals in question. The amount of streamflow diverted to the canal in question was back-calculated from the volume of water applied per acre by multiplying by the area served and adding the amount of water lost as a result of conveyance inefficiency of the canal in question ([U.S.] Soil Conservation Service, 1976). The reference for the Newton Water Users Association ditches was the West Cache Canal. The references for the Spring Creek Irrigation Company ditches were the Logan River Blacksmith Fork Irrigation Company ditches. The reference for the Strong Arm Company ditches was the Twin Lakes Canal.

Two large irrigation canals outside of Cache Valley, West Side Canal and Hammond Main Canal, divert water from Cutler Reservoir and account for a substantial amount of outflow from Cache Valley (fig. 5). Flow in these two canals is measured by Utah Power and Light Company. Mean annual flow of West Side Canal for 1960-90 was 277 cubic feet per second (200,700 acre-feet per year)(table 1) at Utah Power and Light Company streamflow-gaging station West Wheelon Canal (also known as U.S. Geological Survey station 10117500, West Side Canal, near Collinston, Utah). Mean annual flow of Hammond Main Canal for 1960-90 was 54 cubic feet per second (39,100 acre-feet per year) (table 1) at Utah Power and Light Company streamflow-gaging station East Wheelon Canal (also known as U.S. Geological Survey station 10117000, Hammond Main Canal, near Collinston, Utah).

Ground Water

Ground water in Cache Valley occurs in consolidated rocks, poorly consolidated rocks, and unconsolidated basin-fill deposits and is derived from precipitation, infiltration of unconsumed irrigation water, and seepage from canals and streams. Wells in Cache Valley supply water for irrigation, public supply; and industrial, domestic, and stock use. Some spring discharge is diverted for irrigation, public supply, or domestic use. Some ground water in the unconsolidated basin-fill deposits, which is the focus of this study, may be derived from subsurface inflow from adjacent consolidated rock or adjacent unconsolidated basin-fill ground-water systems.

Consolidated and Poorly Consolidated Rocks

Consolidated and poorly consolidated rocks crop out in the mountain areas surrounding Cache Valley and underlie the unconsolidated basin-fill deposits. Some consolidated rocks crop out in the valley, as at Little Mountain near Franklin, at Little Mountain near Newton, and Little Mountain at Twin Lakes Reservoir (referred to as Mount Smart, Newton Hill, and Clifton Hill, respectively, on the geologic map by Bjorklund and McGreevy, 1971, pl. 1).

In the consolidated and poorly consolidated rocks that surround Cache Valley, ground water flows mainly through fractures and through solution channels in carbonate rocks. Many springs in the mountains that surround Cache Valley, particularly in the Bear River Range east of the valley, flow directly or indirectly from solution channels in carbonate rocks. Bjorklund and McGreevy (1971, table 4) listed the water-yielding properties of geologic units in the study area. In consolidated and poorly consolidated rocks of Tertiary age, including the Salt Lake Formation, permeability is generally low to moderate and depends on grain size, sorting, cementation, and solution. Permeability may be locally higher in alluvial-fan deposits or in areas undergoing solution of carbonate components in sedimentary rocks.

Recharge to consolidated rock is by infiltration of precipitation and seepage from streams. Most recharge occurs in the mountains that surround Cache Valley. Flow of water through consolidated and poorly consolidated rocks is probably from recharge areas in the mountains toward stream channels and toward large springs, which generally are in the stream channels or near the margins of the valley. Fractures, faults, joints, and solution channels probably control the direction of ground-water flow in local areas. Ground water discharges from consolidated and poorly consolidated rocks primarily through springs, as seepage to streams, and as leakage to unconsolidated basin-fill deposits along the mountain front.

Unconsolidated Basin-Fill Deposits

Unconsolidated basin-fill deposits are exposed at land surface throughout most of Cache Valley and as thin, narrow deposits that extend into the mountains along the Cub River and Mink Creek drainages. The unconsolidated basin-fill deposits are the primary water-bearing geologic unit in Cache Valley. Ground water in unconsolidated basin-fill deposits is shown schematically in figure 6.

Unconsolidated basin-fill deposits in Cache Valley consist of sediments that range in size from the clays and silts of the lacustrine deposits to the sands, gravels, cobbles, and boulders of the alluvial-fan and terrace deposits. In general, coarser-grained, permeable deposits predominate along the margins of the valley in the alluvial-fan and terrace deposits, particularly along the east bench adjacent to the Bear River Range. The deposits become finer grained and less permeable toward the center of the valley, although geohydrologic sections show some sand layers interbedded with silts and clays in the center of the valley (McGreevy and Bjorklund, 1971, sections D-D', E-E', F-F',I-I', J-J', K-K', L-L', M-M'). Bjorklund and McGreevy (1971, fig. 13) divided Cache Valley into 11 areas of different ground-water conditions. No such attempt was made to divide Cache Valley into different areas as part of this study.

Geohydrologic sections of Cache Valley (McGreevy and Bjorklund, 1971, sections D-D', E-E', F-F', I-I', J-J', K-K', L-L', M-M') indicate that discontinuous layers of clay occur in most of the unconsolidated basin-fill deposits. These sections also show numerous buried faults and contacts that were interpreted from well logs and from the gravity data of Peterson and Oriel (1970). On most of the sections where a contact is shown, the Salt Lake Formation, with low to moderate permeability (Bjorklund and McGreevy, 1971, table 4), underlies the unconsolidated basin-fill deposits. The unconsolidated basin-fill deposits have an approximate maximum thickness of 1,340 feet (Bjorklund and McGreevy, 1971, table 4).

Most of the unconsolidated basin-fill deposits in Cache Valley are saturated at shallow depths with the exception of deposits near the mountain fronts, where several hundred feet of unconsolidated basin fill may remain unsaturated. Near the mountain fronts, ground water generally is unconfined. Perched ground water may occur locally in areas where infiltrating water becomes perched above less-permeable clay layers.

In the center of Cache Valley, ground water is confined below depths typically of about 50 feet because the interbedded clays act as confining layers that impede the upward flow of water. Because clay layers are thin and(or) dis-

Table 4. Diversion of water from streams and reservoirs to selected canals in Cache Valley, 1969 and 1982-90

[Diversions in acre-feet; e, estimated]

Canal name	Stream or reservoir from which water was diverted	Amount of water diverted Year 1969	1982	1983
West Cache Canal	Bear River	42,437	33,656	31,924
Cub Canal-Cub River Pumps	Worm Creek, Cub River, Bear River	11,857	7,857	[1]10,543 e
Logan Smithfield Hyde Park Canal	Logan River	16,331	16,110	11,030
Logan Northern Canal	Logan River	17,037	9,040	6,099
Logan North Field Canal	Logan River	4,694	6,630	5,149
Logan North West Field Canal	Logan River	9,918	14,027	10,796
Millville-Providence Canal	Blacksmith Fork River	[2]12,750 e	[2]13,349 e	[2]13,349 e
Blacksmith Fork Hyrum Canal	Blacksmith Fork River	[2]12,750 e	[2]13,349 e	[2]13,349 e
Nibley Blacksmith Fork Canal	Blacksmith Fork River	[2]12,750 e	[2]13,349 e	[2]13,349 e
Logan River Blacksmith Fork Irrigation Company ditches	Blacksmith Fork River	9,711	7,336	7,125
Preston Whitney Canal Company	Cub River	7,179	2,657	[1]5,333 e
Cub Canal	Worm Creek, Cub River, Bear River	7,626	5,450	[1]8,275 e
Preston-Mink Creek-Riverdale Canal Company	Mink Creek	[3]8,700 e	7,759	4,651
Twin Lakes Canal	Mink Creek, Twin Lakes Reservoir	[3]22,000 e	24,230	19,040
Mountain Home Canal	High Creek	[2]847 e	[2]1,987 e	[2]1,949 e
High Creek Canals	High Creek	[2]8,256 e	[2]19,378 e	[2]19,002 e
Coveville Canal	High Creek	[2]1,482 e	[2]3,478 e	[2]3,411 e
Porcupine Highline Canal	Little Bear River	5,540	7,863	7,502
Paradise Canal	Little Bear River	14,060	6,700	5,579
Hyrum Canal	Little Bear River	11,706	9,130	7,946
East Field Irrigation Canal	Little Bear River	7,258	8,402	5,555
Wellsville-Mendon Canals	Hyrum Reservoir	10,363	6,164	5,082
Smithfield North Bench ditches (underground)	Summit Creek	1,697	[1]1,742 e	1,912
Smithfield Irrigation Company ditches	Summit Creek	9,694	[1]9,033 e	9,695
Newton Water Users Association ditches	Clarkston Creek	[4]5,050 e	[4]4,005 e	[4]3,799 e
Spring Creek Irrigation Company ditches	Spring Creek	[5]6,000 e	[5]4,551 e	[5]4,402 e
Strong Arm Company ditches	Strong Arm Reservoir	[6]5,667 e	[6]6,309 e	[6]4,957 e
Total (rounded)		283,000 e	264,000 e	241,000 e

[1]Amount diverted estimated as the average percentage of streamflow for the irrigation season (May to September). Based on the period of record available for the diversions.

[2]Amount diverted estimated using water right and assuming total streamflow was diverted during the irrigation season (May to September).

[3]Amount diverted estimated as the average percentage of annual streamflow diverted during 1978-90.

[4]Amount diverted estimated by assuming amount applied to area served is applied at the same rate (in acre-feet per year) as for the area served by the West Cache Canal.

[5]Amount diverted estimated by assuming amount applied to area is applied at the same rate (in acre-feet per year) as for the area served by the Logan River Blacksmith Fork Irrigation Company ditches.

[6]Amount diverted estimated by assuming amount applied to area is applied at the same rate (in acre-feet per year) as for the area served by the Twin Lakes Canal.

Amount of water diverted							
Year							
1984	1985	1986	1987	1988	1989	1990	Mean
38,075	37,496	32,809	40,658	44,486	44,251	44,696	39,049
3,427	11,391	6,536	12,500	19,500	16,786	14,063	11,446
13,170	17,340	13,850	14,590	15,890	18,770	17,570	15,465
10,186	14,272	13,314	12,356	12,042	11,974	10,078	11,640
5,666	6,184	5,395	6,310	6,266	6,472	6,077	5,884
12,408	14,021	10,445	10,592	11,178	14,116	17,012	12,451
[2]13,349 e	[2]13,349 e	[2]13,349 e	[2]10,627 e	[2]7,716 e	[2]8,936 e	[2]5,343 e	11,212 e
[2]13,349 e	[2]13,349 e	[2]13,349 e	[2]10,627 e	[2]7,716 e	[2]8,936 e	[2]5,343 e	11,212 e
[2]13,349 e	[2]13,349 e	[2]13,349 e	[2]10,627 e	[2]7,716 e	[2]8,936 e	[2]5,343 e	11,212 e
8,452	9,778	11,105	11,726	7,287	10,197	[1]7,734 e	9,045 e
6,005 e	5,228	3,334	8,258	7,781	8,730	12,212	6,672
9,317 e	7,822	7,718	10,911	8,443	12,718	15,033	9,331
2,677	7,514	6,393	9,939	8,523	10,314	7,574	7,404
24,567	20,867	30,033	14,726	18,684	27,303	17,890	21,934
[2] 2,097 e	[2]1,066 e	[2]1,877 e	[2]650 e	[2]555 e	[2]982 e	[2]456 e	1,247 e
[2]20,447 e	[2]10,390 e	[2]18,304 e	[2]6,337 e	[2]5,410 e	[2]9,571 e	[2]4,441 e	12,154 e
[2] 3,670 e	[2]1,865 e	[2]3,285 e	[2]1,137 e	[2]971 e	[2]1,718 e	[2]797 e	2,181 e
7,592	9,032	8,793	8,848	6,535	9,274	3,765	7,474
5,019	4,759	4,982	4,511	3,653	5,592	5,635	6,049
9,205	10,460	10,532	9,355	6,157	9,282	7,963	9,174
9,116	8,398	15,449	9,152	8,307	8,892	6,539	8,707
9,219	8,641	6,902	5,089	6,229	6,675	11,294	7,566
2,276	1,860	1,142	1,191	1,825	[1]1,673 e	[1]1,623 e	1,694
11,580	6,468	14,607	8,775	4,981	[1]5,525 e	[1]8,417 e	8,878
[4]4,531 e	[4]4,462 e	[4]3,904 e	[4]4,838 e	[4]5,294 e	[4]5,266 e	[4]5,319 e	4,647
[5]5,222 e	[5]6,041 e	[5]6,861 e	[5]7,245 e	[5]4,502 e	[5]6,300 e	[5]4,779 e	5,590
[6]6,396 e	[6]5,433 e	[6]7,819 e	[6]3,834 e	[6]4,864 e	[6]7,109 e	[6]4,658 e	5,705
270,000 e	271,000 e	285,000 e	255,000 e	243,000 e	286,000 e	252,000 e	265,000 e

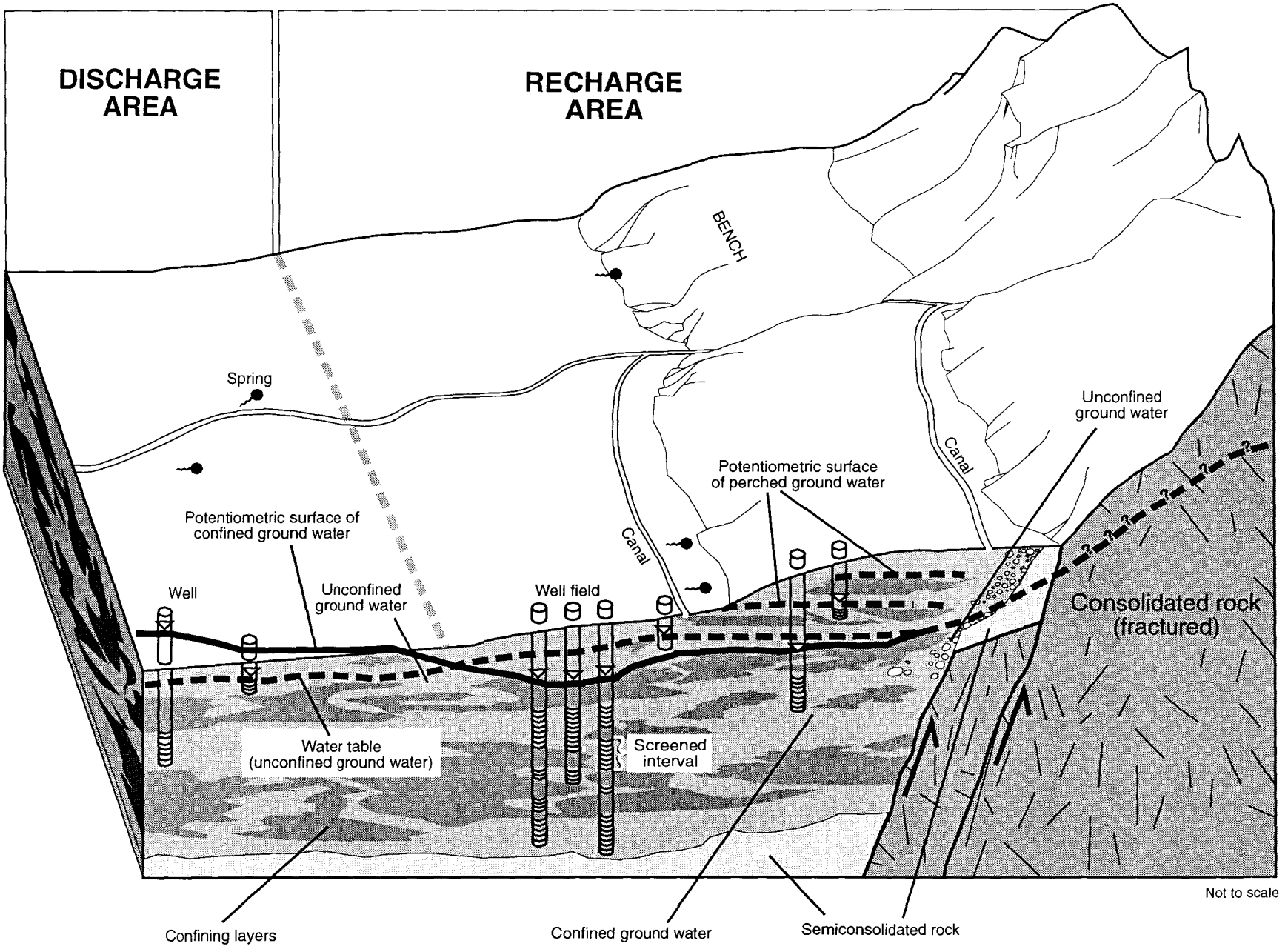


Figure 6 . Schematic block diagram of ground water in unconsolidated basin-fill deposits.

continuous, leaky-artesian conditions exist. Numerous artesian wells are in the center of the valley, particularly in the southern half of the valley. The locations of flowing and non-flowing wells in which water levels were measured during March and April 1991 and the area of flowing wells are shown in figure 7. Ground water typically less than 50 feet deep in the center of the valley is unconfined.

Although unconsolidated basin-fill deposits occur in the areas around Clarkston, Utah, and Weston Canyon in Idaho, each area is considered to contain a separate ground-water system. Because the Clarkston area and Weston Canyon overlie a fault block that is upthrown relative to the graben that underlies the main part of Cache Valley, unconsolidated basin-fill deposits are thinner in these areas. Because the Clarkston area and Weston Canyon are topographically higher than the center of the valley, ground water from these areas flows into the main ground-water system in the unconsolidated basin-fill deposits of Cache Valley. Potentiometric contours (Bjorklund and McGreevy, 1971, pl. 4) also indicate that ground water flows from these areas of thin unconsolidated basin-fill deposits into unconsolidated basin-fill deposits in the main valley. Because the Clarkston area is hydrologically upgradient and adjacent to Cutler Reservoir (Bjorklund and McGreevy, 1971, pl. 4), subsurface flow from this area probably travels only a short distance before it discharges into Cutler Reservoir.

Estimated budgets for the ground-water system in the unconsolidated basin-fill deposits in Cache Valley for 1969 and 1990 are presented in table 5. The budgets represent recharge and discharge of water to the main ground-water system in the unconsolidated basin-fill deposits of the valley. The budgets do not represent the unconsolidated basin-fill deposits in the Clarkston area or in Weston Canyon because they are considered to have separate ground-water systems. Some of the ground-water budget components were estimated as part of a previous study of Cache Valley by Bjorklund and McGreevy (1971). The present study uses data from Bjorklund and McGreevy (1971) to estimate a ground-water budget for 1969; however, interpretation of some of these data has been modified or changed since the previous study and is discussed in the following sections. Estimates of some components of the 1969 ground-water budget therefore may differ from estimates reported by Bjorklund and McGreevy (1971). Many of the estimation techniques and assumptions used to estimate ground-water budgets in this report have large potential errors associated with them; therefore, these estimates should be used with caution.

Budgets for 1969 and 1990 were chosen for use because some data were available to provide estimates for most of the components of recharge and discharge for these years. The methods and computations used to derive the individual budget components are discussed in the following sections.

Recharge

Recharge to the ground-water system in the unconsolidated basin-fill deposits is from infiltration of precipitation and unconsumed irrigation water, seepage from canals and streams, and subsurface inflow from adjacent consolidated rock and unconsolidated basin-fill deposit ground-water systems. Recharge occurs principally along the margins of the valley, which are underlain by coarse-grained and permeable unconsolidated deposits. An estimated 449 cubic feet per second of water recharged the ground-water system in 1969 and an estimated 296 cubic feet per second recharged it in 1990 (table 5).

In 1969, infiltration of precipitation and unconsumed irrigation water on dry-farmed and non-farmed land is estimated to be the largest component of ground-water recharge in Cache Valley and was estimated to be about 186 cubic feet per second (134,700 acre-feet per year) for 1969 and about 57 cubic feet per second (41,300 acre-feet per year) for 1990. Applied irrigation water and precipitation that does not run off as surface water, is not transpired by plants, is not evaporated, and is in excess of the water needed to replenish soil moisture is assumed to recharge the ground-water system.

Use of dry-farmed land typically alternates between fallow and cultivation of winter wheat every other year (Eugene Bigler, Utah Department of Natural Resources, oral commun., 1991). In years when winter wheat was grown, crops on dry-farmed land were assumed to use the total amount of water that would otherwise have recharged the ground-water system. During fallow years, 10 percent of the precipitation was assumed to recharge the ground-water system. In 1976, there were about 168,200 acres of irrigated land, 92,100 acres of dry-farmed land, and 57,000

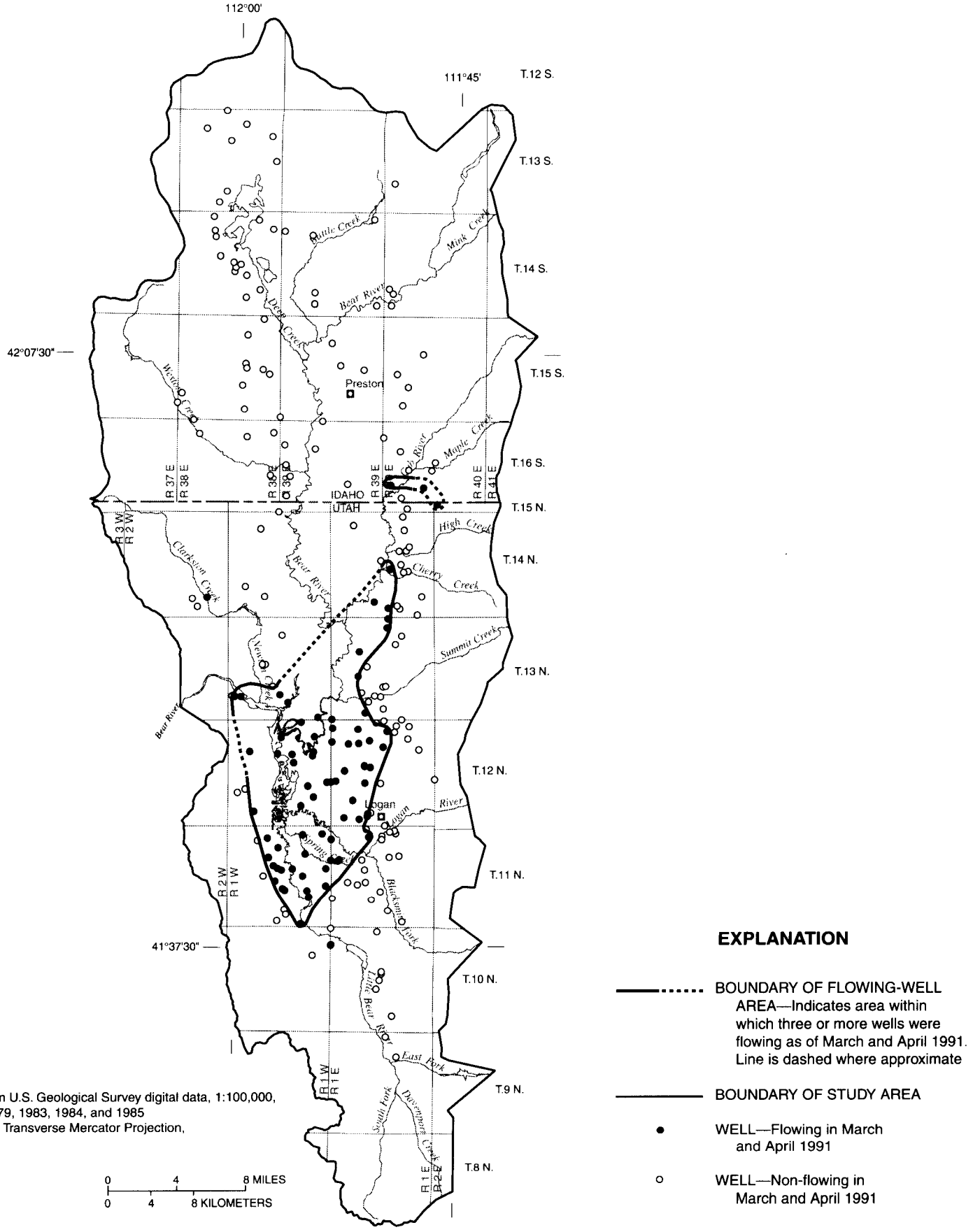


Figure 7. Location of flowing and non-flowing wells in unconsolidated basin-fill deposits in Cache Valley in which water levels were measured during March and April 1991.

Table 5. Estimated budget for the ground-water system in the unconsolidated basin-fill deposits in Cache Valley, 1969 and 1990

Budget component	1969 flow (cubic feet per second)	1990 flow (cubic feet per second)
Recharge		
Infiltration of precipitation and unconsumed irrigation water	186	57
Seepage from canals	160	140
Seepage from streams	7	3
Other forms of recharge[1]	[2]96	[3]96
Total recharge	449	296
Discharge		
Seepage to streams and Cutler Reservoir	180	[3]180
Spring discharge	138	[3]138
Evapotranspiration	87	[3]87
Withdrawal from wells	44	52
Total discharge	449	457
Change in storage[4]	0	[5]-161

[1] Includes subsurface inflow from adjacent consolidated rock and unconsolidated basin-fill deposit ground-water systems, and seepage from ephemeral streams.

[2] Difference between total discharge and recharge from infiltration of precipitation and unconsumed irrigation water and seepage from canals and streams.

[3] Assumed to be the same as in 1969.

[4] Total recharge minus total discharge. Value indicates flow of ground water to or from (-) ground-water storage.

[5] Change in storage for 1990 is not representative of a typical year because less-than-average precipitation and streamflow in 1990 resulted in less recharge.

acres of non-farmed land in Cache Valley, as indicated by a land-use survey in digital line-graph format by the Bear River Commission. These 1976 land-use estimates were used to compute recharge from precipitation and unconsumed irrigation water because estimates for 1969 and 1990 were unavailable for the entire valley.

Initial estimates of recharge to the ground-water system from precipitation in Cache Valley were made by arbitrarily assuming that 10 percent of the precipitation that fell in the valley became recharge, except on dry-farmed land, where 5 percent was assumed to become recharge because crops use more water than native vegetation. (It was assumed that in any given year, one-half of the dry-farmed land was fallow and one-half was being farmed.) The percentage of precipitation that recharges the ground-water system was not measured during this study. Studies of other basins in Utah indicate that recharge to areas underlain by unconsolidated basin-fill deposits may range from 1 to 20 percent of the precipitation (Razem and Steiger, 1981, p. 13; Hood and Waddell, 1968, p. 24; and Feth and others, 1966, p. 43).

Recharge from unconsumed irrigation water was computed by taking the amount of water diverted by each irrigation company (table 4), subtracting the transmission losses in major canals, and applying an on-farm efficiency reported by the [U.S.] Soil Conservation Service (1976) to the remaining amount of water. On-farm efficiency is the percentage of water delivered to the farm that is not lost as surface runoff and is not lost as deep infiltration of water below the crop root zone. The residual water is assumed to recharge the ground-water system.

According to the [U.S.] Soil Conservation Service (1976, tables 5 and 10), 82 irrigation companies served about 148,800 acres of farmland in Cache Valley. An estimated average of about 265,000 acre-feet per year of water (table 4) is diverted from streams and reservoirs to canals and applied to this area. In addition, an estimated 60,000 acre-feet per year of irrigation return flow is reapplied to the Utah part of this area (James Stephens, Utah Division of Water Resources, oral commun., 1993). [U.S.] Soil Conservation Service (1976) estimates of on-farm efficiency for each service area ranged from 15 to 65 percent.

The 1976 land-use survey by the Bear River Commission indicated that about 19,400 acres of land were irrigated but were not served by irrigation companies as described by the [U.S.] Soil Conservation Service (1976). The rate of recharge from infiltration of unconsumed irrigation water for these areas was assumed to be 0.95 acre-foot per year, which is the average recharge rate for all other irrigated areas in Cache Valley. Possible sources of this recharge include spring discharge diverted for irrigation, water withdrawn from wells, water pumped from the Bear River, and effluent from the Logan City sewage lagoons that is diverted for irrigation (Rodney Blossom, Logan City Public Works Department, oral commun., 1993).

Initial estimates of annual recharge from infiltration of unconsumed irrigation water in areas served by canal companies were based on the reported and estimated diversions listed in table 4. These estimates were made on the basis of an area-weighted mean recharge for all of the irrigation company areas. The annual amount of recharge in each irrigation company's service area was computed using the equation

$$R = \left[\frac{(TD \times CE)}{IA}\right] \times (1.0 - OE) \qquad (2)$$

where: R is recharge for year of interest, in feet;

TD is total amount of water diverted to the area served by a canal listed in table 4, in acre-feet;

CE is canal conveyance efficiency estimated by the [U.S.] Soil Conservation Service (1976), in decimal form;

IA is irrigation company service area defined by the [U.S.] Soil Conservation Service (1976), in acres; and

OE is on-farm efficiency estimated by the [U.S.] Soil Conservation Service (1976), in decimal form.

Initial estimates of recharge are shown in table 6 and varied from 10.80 to 14.14 inches (0.90 to 1.18 feet) for irrigated land and from 0.63 to 1.79 inches (0.05 to 0.15 foot) for dry-farmed land. As expected, irrigated land had the largest amount of recharge and dry-farmed land had the smallest amount of recharge. With this method of estimating unconsumed irrigation water, the amount of surface-water runoff is assumed to be zero.

During the 1982-90 transient-state computer simulation of the ground-water system discussed later in this report, the model-computed water levels did not fluctuate as much as the measured water levels. Further investigation determined that this difference occurred because the initial recharge estimates did not vary much from one year to the next during simulation. Other means of estimating recharge and correcting the initial recharge estimates were investigated. The Deep Percolation Model (DPM) of Bauer and Vaccaro (1987) was used to revise the initial estimates of recharge from irrigated lands and to estimate recharge from dry-farmed land and non-farmed land. The method used to compute initial estimates of recharge and the method of estimating recharge using the DPM is discussed in the following pages. The assumptions of and method used in the DPM simulation are discussed in the appendix.

A simplified recharge model was constructed using the DPM. The model had four cells and an arbitrary total area of 5.75 square miles and was operated for each of the simulations (irrigated, dry-farmed, and non-farmed land). All parameters were kept the same in all four cells. Only one crop type and one soil type were used in each simulation.

The correction factor for each year was determined by dividing the total amount of recharge computed by the DPM for that year by the total amount of recharge initially estimated for irrigated crop lands for each year. The correction was applied by multiplying the yearly correction factor by the initial estimate of recharge for each irriga-

tion company or group. The total amount of recharge from irrigation for each year was the sum of all of the corrected irrigation-company or group estimates. The initial estimate and the DPM estimate of recharge from infiltration of precipitation and unconsumed irrigation water are shown in table 6.

The relation between precipitation and recharge for irrigated, dry-farmed, and non-farmed land in 1969 is shown in figure 8. Most of the recharge estimated by the DPM occurred in winter or early spring, when the snowpack was melting. Zero to a very small amount of recharge occurs during summer and fall. This phenomenon is verified by soil-moisture studies done by Hill, Hanks, and Wright (1982) that show that only small amounts of recharge occurred in sprinkler-irrigated fields during summer and fall, except when large amounts of precipitation fell on irrigated lands.

DPM simulations were run for 1969 and for each year in 1969-90. Results of the simulations indicate that annual recharge varied from 2.11 inches (0.176 foot) to 19.48 inches (1.623 feet) in irrigated lands and from 0.04 inches (0.003 foot) to 2.14 inches (0.178 foot) in dry-farmed lands (table 6). The rate of recharge in 1969 was computed using data in table 6 to be about 133 cubic feet per second (96,300 acre-feet per year) in irrigated lands, 16 cubic feet per second (12,000 acre-feet per year) in dry-farmed lands, and 37 cubic feet per second (27,000 acre-feet per year) in non-farmed lands. The 1990 recharge rate was about 41 cubic feet per second (30,000 acre-feet per year) in irrigated lands, 2.3 cubic feet per second (1,700 acre-feet per year) in dry-farmed lands, and 14 cubic feet per second (10,100 acre-feet per year) in non-farmed lands.

Seepage from canals is an important source of recharge to the ground-water system in Cache Valley. In the center of the valley, water may seep from canals to the shallow water table and move a short distance before being discharged in a nearby spring or seep, evaporated from the ground surface, or transpired by plants.

Canals in Cache Valley (pl. 1) carry about 265,000 acre-feet of water per year (table 4). Most of the water is diverted from streams, but some is diverted from springs or pumped from wells ([U.S.] Soil Conservation Service, 1976). Canal efficiencies ranged from 25 to 108 percent, with an average canal efficiency of 59 percent, weighted by canal length ([U.S.] Soil Conservation Service, 1976). Canal efficiency is the percentage of flow in the canal not lost to seepage or to consumption by phreatophytes.

An estimated 160 cubic feet per second (116,000 acre-feet per year) of water in 1969 and 140 cubic feet per second (103,000 acre-feet per year) of water in 1990 recharged the ground-water system from canal seepage (table 5). These estimates were determined by multiplying the diversions of 283,000 acre-feet in 1969 and 252,000 acre-feet in 1990 (table 4) by 100 percent minus the average canal efficiency of 59 percent, weighted by canal length, reported by the [U.S.] Soil Conservation Service (1976).

Bjorklund and McGreevy (1971, p. 19) assumed that seepage from streams was an important source of recharge to the ground-water system in Cache Valley. They assumed that recharge occurs where streams flow from canyons across coarse-grained deposits at the sides of the valley, where the water can infiltrate readily from the streams to the underlying ground-water system.

To test this assumption, seepage studies were done on the Logan, Blacksmith Fork, and Little Bear Rivers during November 1990. The studies involved measuring all surface-water inflows and outflows along with streamflow in a section of a stream to determine the amount of water gained from or lost to the ground-water system. Three measurement runs were completed for each stream during the month of November 1990 along reaches beginning at the mouth of the canyon from which each stream flowed. The data collected in the seepage studies were tabulated by Roark and Hanson (1992, table 7). These studies showed that none of the three streams lost water in amounts greater than measurement error (plus or minus 10 percent of total flow). Some measured reaches indicated small gains that could have been derived from perched water entering the stream in the measured reach. These results do not support the assumption that seepage from streams supplies substantial amounts of recharge to the ground-water system. The Logan, Blacksmith Fork, and Little Bear Rivers are deeply incised and probably do not flow across permeable alluvial-fan deposits.

Estimated recharge from seepage from streams in 1969 and 1990 is shown in table 5. These estimates did not include seepage from ephemeral streams, which was computed separately. These estimates assume that of the major

Table 6. Initial and Deep Percolation Model (DPM) estimates of recharge from infiltration of precipitation and unconsumed irrigation water in Cache Valley

[In inches; —, not applicable]

Year	Recharge		Correction factor[1]
	Initial estimate	DPM estimate	
	Irrigated land		
1969	14.14	6.87	0.486
1982	13.44	17.43	1.297
1983	13.64	19.48	1.428
1984	12.86	14.74	1.146
1985	12.67	6.57	.518
1986	13.63	14.10	1.034
1987	11.94	2.63	.220
1988	10.80	2.71	.251
1989	11.97	4.56	.381
1990	11.16	2.11	.189
	Dry-farmed land		
1969	.84	1.54	—
1982	1.42	1.67	—
1983	1.79	1.06	—
1984	1.18	2.14	—
1985	.99	.52	—
1986	1.34	1.93	—
1987	.86	.04	—
1988	.63	.24	—
1989	.72	.96	—
1990	.74	.23	—
	Non-farmed land		
1969	1.68	5.73	—
1982	2.84	14.46	—
1983	3.58	14.61	—
1984	2.36	10.91	—
1985	1.98	4.81	—
1986	2.68	7.97	—
1987	1.72	3.47	—
1988	1.26	2.46	—
1989	1.45	4.97	—
1990	1.49	2.12	—

[1] Computed by dividing DPM estimate by initial estimate.

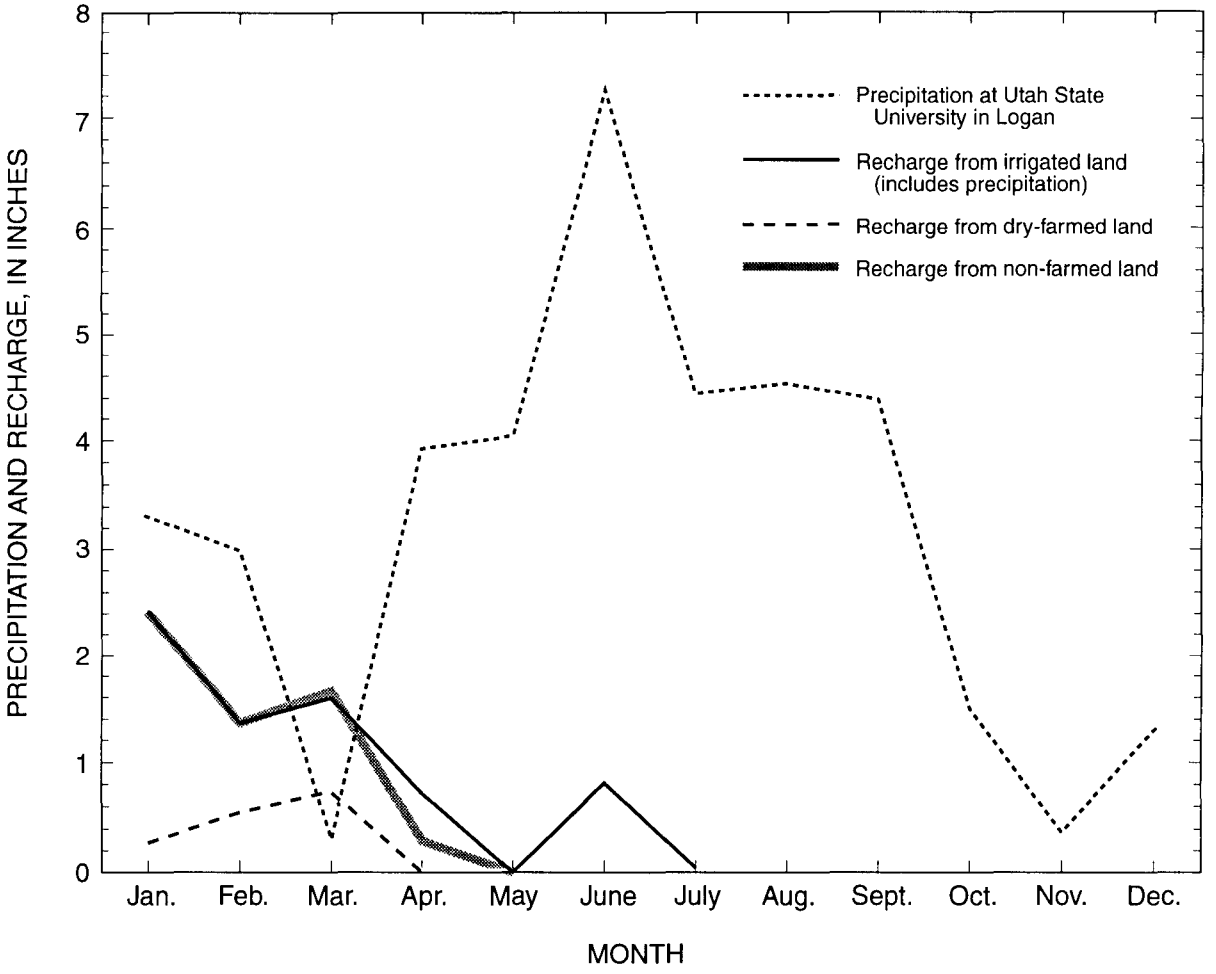


Figure 8. Monthly precipitation and computed Deep Percolation Model recharge from infiltration of precipitation and unconsumed irrigation water in Cache Valley, 1969.

streams that flow from canyons and across unconsolidated basin-fill deposits, only Maple Creek, Summit Creek, and High Creek lose water to the ground-water system. This loss probably occurs where the streams enter Cache Valley because the channels of these streams are not incised and because the streams flow on top of permeable alluvial-fan deposits. Hely and others (1971) determined that in nearby Salt Lake Valley, the mean annual loss from six streams during 1964-68 was 11.4 percent. Because seepage data for Maple Creek, Summit Creek, and High Creek were not available and because these streams are, in general, physically similar to streams in Salt Lake Valley, these streams were assumed to lose water as seepage to the ground-water system at a rate of 10 percent of their annual mean flow (table 1), a rate similar to streams in Salt Lake Valley. On the basis of these measurements, estimates, and assumptions, recharge from stream seepage was 7 cubic feet per second (5,000 acre-feet per year) in 1969 and 3 cubic feet per second (2,000 acre-feet per year) in 1990.

About 96 cubic feet per second (69,600 acre-feet per year) of water was estimated to recharge the main ground-water system in the basin-fill deposits of Cache Valley as subsurface inflow from adjacent consolidated rock and adjacent basin-fill deposits in the Clarkston area and Weston Canyon, and as seepage from ephemeral streams. This estimate was computed as the difference between the total estimated discharge from the ground-water system and the sum of estimated recharge from all other sources (table 5). The estimate of 96 cubic feet per second (69,600 acre-ft per year) is used for both 1969 and 1990 because data were insufficient to refine the estimates for 1990. Because estimated recharge in 1990 was 34 percent less than in 1969, it is probable that recharge from adjacent consolidated rock and basin-fill deposits and seepage from ephemeral streams were somewhat less in 1990 than in 1969.

There is indirect evidence of subsurface inflow of water from consolidated rock for Cache Valley. Many springs, some with substantial flow, discharge water from fractures or solution channels in consolidated rock in the Bear River Range and the Wellsville Mountains. Such fractures and solution channels also could provide flowpaths for water from consolidated rock to unconsolidated basin-fill deposits. Thermal springs and areas of warm ground water (Bjorklund and McGreevy, 1971, pl. 6) in several areas throughout the valley may be evidence that water from consolidated-rock sources rises along faults and mixes with ground water in basin-fill deposits (DeVries, 1982, p. 48). As previously discussed in the "Unconsolidated basin-fill deposits" section of this report, some ground water also probably flows into the main basin-fill ground-water system of Cache Valley from the separate ground-water systems of the Clarkston area and Weston Canyon.

Streamflow in small ungaged perennial and ephemeral streams that flow into Cache Valley was estimated to be about 34 cubic feet per second (24,400 acre-feet per year, table 3). Most of this streamflow probably enters the ground-water system where the streams flow across coarse-grained deposits at the margins of the valley. This assumption is supported by the lack of developed stream channels below the alluvial fans and the absence of observed flow continuing to the valley floor, particularly during spring, when maximum runoff and minimum evapotranspiration occur.

Flow

The approximate potentiometric surface in the unconsolidated basin-fill deposits in Cache Valley during March and April 1991 is shown on plate 2. The potentiometric-surface map was made using water levels in wells generally 100- to 300-feet deep that were completed in unconsolidated basin-fill deposits. Water levels for wells used to prepare the potentiometric-surface map are reported in Roark and Hanson (1992, table 1).

The direction of ground-water flow in Cache Valley is generally perpendicular to the contour lines on plate 2. In general, ground water flows away from the recharge areas at the mountain fronts and benches toward discharge areas in the center of the valley. In the northern part of the valley, ground water flows from the mountain fronts toward the Bear River. Along the eastern side of the valley, ground water flows from the Bear River Range toward the Cub and Bear Rivers. In the southern end of the valley, ground water flows toward the Bear River and Cutler Reservoir. Along the west side of the valley, ground water generally flows toward Cutler Reservoir in the south or toward the Bear River in the north.

Potentiometric-surface maps were not prepared for the shallow ground-water zone (less than 100 feet) or for zones deeper than 300 feet. In the shallow zone, the water table is expected to form an approximate replica of the valley topography. Along the margins of the valley, where most ground-water recharge occurs, the water table is at a greater depth than in other parts of the valley, although perched ground water may exist locally. In the south-central part of the valley, the water table is very near or at land surface, as evidenced by numerous springs and by seepage to the Bear River and Cutler Reservoir. Shallow ground-water flow is locally more pronounced toward springs and streams than in the zone between 100 and 300 feet. In zones deeper than 300 feet, ground water flows in essentially the same directions as can be inferred from contours on plate 2.

Horizontal hydraulic gradients are largest near the mountain fronts and benches, with the largest gradient of about 400 feet per mile on the east side of the valley. Horizontal gradients decrease toward the center of the valley and are smallest at the confluence of the Cub River with the Bear River and around Cutler Reservoir.

Locations of nested wells where vertical gradients were measured in March and April 1991 are shown on plate 2. Large variation in vertical gradient occurs in the valley. Downward vertical gradients as large as -0.2 foot per foot occur along the margins of the unconsolidated basin-fill deposits, at the distal ends of alluvial-fan deposits. The largest upward vertical gradients, as large as +0.13 foot per foot, occur in the south-central part of the valley (pl. 2). Downward vertical gradients indicate potential for downward flow of ground water in areas likely to be recharge areas. Upward vertical gradients indicate potential for upward flow of ground water and eventual discharge by seepage to streams and to Cutler Reservoir, from springs and seeps, and by evapotranspiration in the discharge area. Layers of clay and silt impede the vertical flow of ground water in the valley, especially in the south-central part of the valley, where these layers confine the ground water and where artesian conditions exist.

Discharge

Discharge from the ground-water system in the unconsolidated basin-fill deposits in Cache Valley includes seepage to streams and reservoirs, spring discharge, evapotranspiration, and withdrawal from wells. Bjorklund and McGreevy (1971, p. 35) estimated total annual seepage to streams in Cache Valley to be about 140,000 acre-feet. They observed that many streams originate at springs in the valley and collect additional water from springs and drains as they flow toward the Bear River or one of its tributaries. They also described many streams that originate outside the valley but gain flow within the valley from springs or tributaries that originate in the valley. Discharge was estimated to be 449 cubic feet per second during 1969 and 457 cubic feet per second in 1990 (325,300 acre-feet per year and 331,000 acre-feet per year, respectively).

The amount of ground-water discharge as seepage to streams estimated in this study is about the same as that estimated by Bjorklund and McGreevy (1971, p. 35). About 180 cubic feet per second, or 130,400 acre-feet per year, is estimated to discharge annually from the ground-water system to major streams and to Cutler Reservoir (table 5), as compared with the 140,000 acre-feet per year estimated by Bjorklund and McGreevy (1971). The amount listed in table 5 for 1969 is the same as that listed for 1990 on the assumption that ground-water discharge to streams has not changed. The estimate of 180 cubic feet per second (130,400 acre-feet per year) is used for both 1969 and 1990 because there were no data available to refine the estimates for individual years. Because estimated recharge in 1990 was 34 percent less than in 1969, it is probable that discharge from seepage to streams and Cutler Reservoir was somewhat less in 1990. Estimated gains in individual streams are discussed in the following paragraphs.

Of all the major streams in Cache Valley, the Bear River, including Cutler Reservoir, receives the largest amount of ground-water discharge as seepage from streams. The Bear River gains ground water as seepage directly to the river channel, as inflow from nearby springs and seep areas, and as seepage of perched ground water. The perched ground water is from seepage faces along steep escarpments between Oneida Narrows Reservoir and Franklin, where the Bear River channel is deeply incised into basin-fill deposits. A seepage study of the Bear River and Cutler Reservoir in October and November 1990 indicated a gain of 23.5 cubic feet per second to the Bear River (Herbert and Thomas, 1992, p. 3). This seepage estimate may be larger than the actual amount of seepage to the

Bear River because the estimate includes not only seepage directly to the Bear River, but also some inflow from springs and seep areas along the river and seepage from perched ground water where the river cuts through the Bear River delta.

Data in a report by Roark and Hanson (1992, table 3) indicate that the Little Bear River, Logan River, and Blacksmith Fork either gain water or show no net change along reaches beginning near the mountain fronts. The Little Bear River showed a gain of 2.04 cubic feet per second per mile of stream channel measured between site (A-9-1)10bbc-1W and site (A-10-1)33bdc-1W (Roark and Hanson, 1992, pl. 1), out of an average flow of 15.2 cubic feet per second. Assuming that the Little Bear River gains this amount along its entire length within the valley, it was estimated to gain a total of 45.7 cubic feet per second (33,100 acre-feet per year).

Data of Roark and Hanson (1992, table 3) indicate that the Logan River had no net gain or loss of water in the reach measured between site (A-12-1)36bca-1W and site (A-12-1)34ddd-1W (Roark and Hanson, 1992, pl. 1), although the river probably gains water farther downstream, in the center part of the valley where the potentiometric surface is above the river channel. In the absence of seepage data for the downstream part of the river, the Logan River was assumed to gain about 1 cubic foot per second per mile from the point where the river channel is at the same altitude as the potentiometric surface to where the river enters Cutler Reservoir. Under these conditions, the total gain of the Logan River within the valley would be 7.7 cubic feet per second (5,600 acre-feet per year).

Similar measurements were not made for the Cub River, although the river gains a large amount of water from consolidated rock before entering the valley. Although the total flow of the Cub River was diverted to canals upstream from where the river flows across basin-fill deposits, on one occasion during the study about 4 to 5 cubic feet per second of streamflow was observed downstream in the river, near Franklin. In the absence of seepage data for the Cub River, about 1 cubic foot of ground water per second per mile was assumed to seep into the Cub River. At this rate of seepage, the Cub River would gain about 23.9 cubic feet per second (17,300 acre-feet per year).

A substantial amount of ground water discharges from the ground-water system as seepage to reservoirs. Herbert and Thomas (1992, p. 3) determined that about 79 cubic feet per second of ground water seeped into Cutler Reservoir. Although a seepage study of Hyrum Reservoir has not been done, seepage of ground water into Hyrum Reservoir probably occurs because the altitude of the potentiometric surface (pl. 2) is higher than the altitude of the water surface in the reservoir. An estimate of ground-water seepage to Hyrum Reservoir was not available.

Bjorklund and McGreevy (1971, p. 35) stated that ground water from both shallow unconfined aquifers and deeper confined aquifers discharges to springs and seeps in the south-central part of Cache Valley. In some places, spring discharge is lost to evapotranspiration from ponds and wetland areas. In other places, spring discharge runs off into streams that ultimately flow into the Bear River or Cutler Reservoir. Seepage from low-lying areas also may contribute flow to surface water where streams and sloughs act as drains and collect the ground-water seepage.

Numerous springs discharge ground water from basin-fill deposits in Cache Valley; however, the mechanism of ground-water flow to the springs is not well understood. Large changes in discharge occur seasonally in many springs. These changes indicate a shallow ground-water-flow path from the source of recharge to the point of discharge. Only a small amount of data exists on long-term fluctuations in spring discharge.

Discharge from Northfield Spring was measured on a monthly basis as part of this study (site (B-11-1)34dac-S1 in Roark and Hanson, 1992, table 7). A graph of monthly discharge from this spring is shown in figure 9. Northfield Spring discharges more water during the summer than during the fall and winter. Some unconsumed irrigation water applied nearby during the summer probably flows to this spring and causes the increased discharge during the irrigation season. Northfield Spring has a base flow of about 2 to 2.5 cubic feet per second that may originate from deeper sources because the spring discharges along a fault.

Discharge from springs in unconsolidated basin-fill deposits in Cache Valley was estimated to be 138 cubic feet per second (100,000 acre-feet per year) in 1969 and assumed to be the same in 1990 (table 5). This estimate does not include discharge from springs outside or near the margins of the unconsolidated basin-fill deposits because these springs probably discharge from consolidated rock. Springs along the Little Bear and Cub Rivers were not included in the estimate because that discharge is included in the estimates of seepage to streams and reservoirs. Dis-

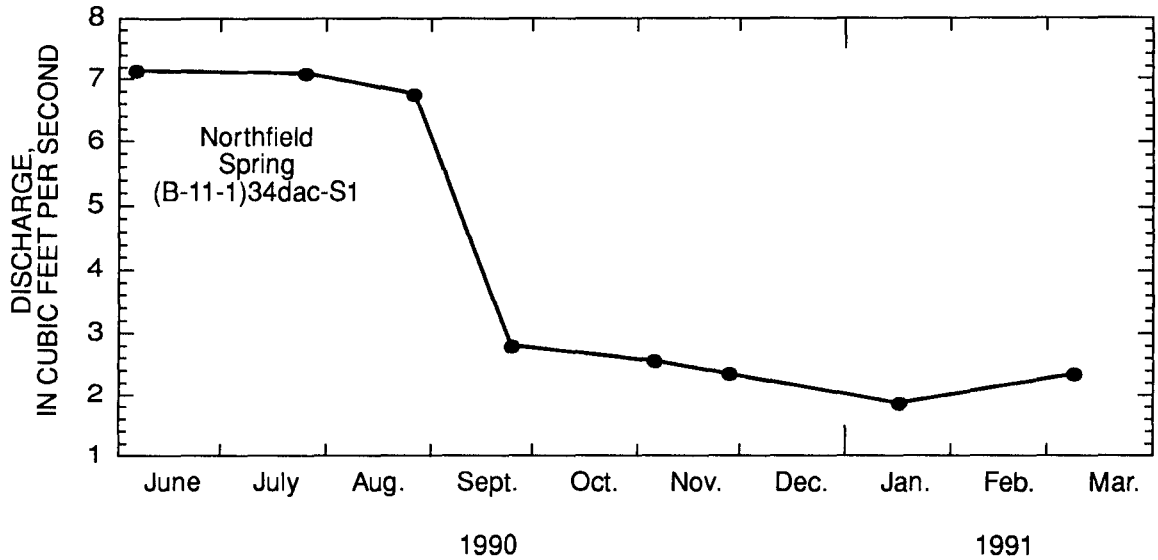


Figure 9. Discharge from Northfield Spring, 1990-91.

charge from springs in the Clarkston area or in Weston Canyon (McGreevy and Bjorklund, 1970, table 2) was not included in the estimates because these areas were considered to have separate ground-water systems. The estimate of 138 cubic feet per second (100,000 acre-feet per year) is used for both 1969 and 1990 because data were not available to refine the estimate for 1990. Because estimated recharge in 1990 was 34 percent less than in 1969, it is probable that discharge from springs was somewhat less in 1990 than in 1969.

Measured and estimated discharge from 44 springs in the main part of the basin-fill deposits in Cache Valley is shown in table 7. The location of these springs is shown in figure 10. Three of these springs (springs (A-11-1)17bdb-S1, (B-10-1)10aac-S1, and (B-11-1)34dac-S1) were measured in 1990 within 2 months of the date on which they were measured by McGreevy and Bjorklund (1970, table 2). Discharge from these three springs decreased an average of about 40 percent between 1969 and 1990. If estimated discharge values from table 7 are included, an average 25-percent decline is evident based on seven springs (springs (A-11-1)17bdb-S1, (A-11-1)18bad-S1, (A-11-1)18bcd-S1, (A-14-1)15acb-S1, (A-14-1)34bcd-S1, (B-10-1)10aac-S1, and (B-11-1)34dac-S1). Although these few data indicate a decline in spring discharge in 1990 as compared with spring discharge in 1969, the data are insufficient to establish the amount of decline. A decline would be expected, however, because of the decreased recharge in 1990 as compared with 1969 (table 5). A better understanding of the mechanisms and sources of flow to springs in Cache Valley would provide a better understanding of the long-term fluctuations in spring discharge.

Three areas have substantial amounts of ground-water discharge from small springs to low-lying areas. Discharge from some of the larger springs can be measured and is reported in table 7, but most of the discharge is diffuse. Combined spring discharge from these areas was measured in October and November 1990 as part of a seepage study of the Bear River and Cutler Reservoir (Herbert and Thomas, 1992). One area of about 1 square mile, located 2 miles southwest of Richmond, was estimated to discharge about 0.47 cubic foot per second of ground water. A second area, of about 1.75 square miles and located 2 miles northwest of Smithfield, was estimated to discharge about 5.57 cubic feet per second of ground water. The third area, of about 18 square miles that extend approximately from Smithfield to Logan and from Highway 89 to the Bear River, was estimated to discharge about 40.63 cubic feet per second of ground water. Together, these three areas discharge about 47 cubic feet per second (34,000 acre-feet per year) of ground water. This estimate was assumed to represent ground-water discharge during 1969 and 1990. Although this seepage estimate may be small because it may not account for some losses to evapotranspira-

Table 7. Discharge from selected springs in unconsolidated basin-fill deposits in Cache Valley, 1967-69 and 1990

[See figure 2 for an explanation of the numbering systems used for hydrologic-data sites; gal/min, gallons per minute; e, estimated; m, measured; r, reported; —, no data; do and Do., ditto]

Location	Name	Date measured	Discharge (gal/min)	Remarks
(A-9-1)10bac-S1	L.A. Farnsworth	06-00-68	[1]440 e	Located along Little Bear River.
(A-10-1)3bbb-S1	—	10-00-68	300 m	—
(A-10-1)16bba-S1	E.K. Isrealsen	06-00-68	10 m	Located along Little Bear River.
(A-10-1)21bab-S1	—	10-00-68	[1]1,300 e	Do.
		06-06-90	100 m	
(A-10-1)21bca-S1	—	10-00-68	1,530 m	Do.
		06-06-90	500 m	
(A-10-1)21dab-S1	Ellis Hansen	09-00-68	[1]130 e	—
(A-10-1)28cab-S1	W.A. Obray	06-00-68	720 m	Located along Little Bear River.
(A-10-1)28cac-S1	Whites Inc.	09-00-68	[1]900 e	Do.
(A-10-1)33bac-S1	do	09-00-68	[1]1,800 e	Do.
		06-06-90	76.4 m	
		12-06-90	117 m	
(A-11-1)10cca-S1	Little Ballard Spring	08-00-67	1,530 m	—
		05-31-90	447 m	
		11-21-90	277 m	
(A-11-1)15bbc-S1	Big Ballard Spring	—	1,800 r	—
		05-31-90	829 m	
		11-21-90	344 m	
(A-11-1)17bdb-S1	Spring Creek Number 1	11-00-67	2,430 m	Discharges to Spring Creek.
		06-05-90	3,150 m	
		11-28-90	2,060 m	
		12-06-90	2,050 m	
(A-11-1)18bad-S1	Spring Creek Number 2	11-00-67	[1]2,700 e	Do.
		06-05-90	[3]2,280 m	
		12-06-90	[3]3,890 m	
(A-11-1)18bcd-S1	Spring Creek Number 3	11-00-67	[1]2,700 e	Do.
		06-05-90	[3]2,280 m	
		12-06-90	[3]3,890 m	
(A-11-1)34dcb-S1	Yeates Spring	—	250 r	—
(A-12-1)4bab-S1	Chambers Spring	—	2,600 r	Located in one of three identified seep areas.
		05-31-90	421 m	
(A-12-1)29cac-S1	Tree Spring	—	3,600 r	Do.
(A-13-1)17dca-S1	Joseph Smith Spring	06-01-90	216 m	Do.
		11-10-90	165 m	
(A-13-1)20aba-S1	William Smith Spring	06-01-90	347 m	Do.
		11-20-90	68.2 m	
(A-13-1)20cca-S1	—	05-31-90	600 m	Do.
		11-10-90	722 m	
(A-13-1)20dbb-S1	Corbitt Spring	06-07-90	854 m	Do.
		11-20-90	942 m	

Table 7. Discharge from selected springs in unconsolidated basin-fill deposits in Cache Valley, 1967-69 and 1990—Continued

Location	Name	Date measured	Discharge (gal/min)		Remarks
(A-13-1)29abc-S1	Lynn Erickson	—	[2]	e	—
(A-13-1)29acb-S1	do	—	[2]	e	—
(A-13-1)29bab-S1	do	—	[2]	e	—
(A-13-1)29bac-S1	Anderson Spring	—	[2]	e	—
(A-13-1)29bac-S2	Lynn Erickson	—	[2]	e	—
(A-13-1)29bca-S1	do	—	[2]	e	—
(A-13-1)29bcd-S1	do	07-00-68	150	r	—
(A-13-1)29bdc-S1	Story Spring	07-00-68	450	r	—
(A-13-1)29bdc-S2	Mack Spring	—	[2]	e	—
(A-13-1)29ccd-S1	Gittins Spring	05-31-90	395	m	Located in one of three identified seep areas.
		11-10-90	503	m	
(A-13-1)32adc-S1	Hopkins Spring	—	1,700	r	Do.
		05-31-90	520	m	
		11-21-90	280	m	
(A-14-1)15acb-S1	Clear Creek Spring	11-00-68	[1]1,300	e	Located along Cub River.
		06-07-90	1,200	m	
		11-20-90	552	m	
(A-14-1)22bab-S1	—	04-00-68	[1]10	e	Do.
(A-14-1)33abb-S1	—	06-07-90	100	m	—
		11-10-90	2	m	
(A-14-1)34bcd-S1	Peart-Thompson	12-00-68	140	r	—
		06-07-90	100	m	
		11-20-90	30.0	m	
(B-10-1)10aac-S1	Wellsville Spring	10-00-68	1,800	m	—
		06-06-90	822	m	
		12-07-90	930	m	
(B-11-1)13aab-S1	Spring Creek Number 4	11-00-67	450	m	Discharges to Spring Creek.
(B-11-1)21dac-S1	Gardner Spring	—	2,700	r	—
		06-05-90	596	m	
		12-07-90	1,990	m	
(B-11-1)27bdd-S1	Clayton Spring	04-00-69	900	m	—
(B-11-1)27dca-S1	Darley Spring	04-00-69	[1]1,800	e	—
		12-07-90	1,500	m	
(B-11-1)34dac-S1	Northfield Spring	11-00-67	2,830	m	—
		06-06-90	3,200	m	
		11-28-90	1,050	m	
(B-11-1)35bcb-S1	Grant Parkinson	10-00-68	760	m	—
(B-13-1)15bca-S1	—	09-00-68	[1]100	e	—

[1]Estimated discharge from field observation.

[2]Group of seven springs discharged an estimated 1,000 gallons per minute in 1969.

[3]Discharge of Spring Creek Number 2 and 3 were measured together, below the confluence. Discharge of each spring is assumed to be one-half of the combined measured discharge.

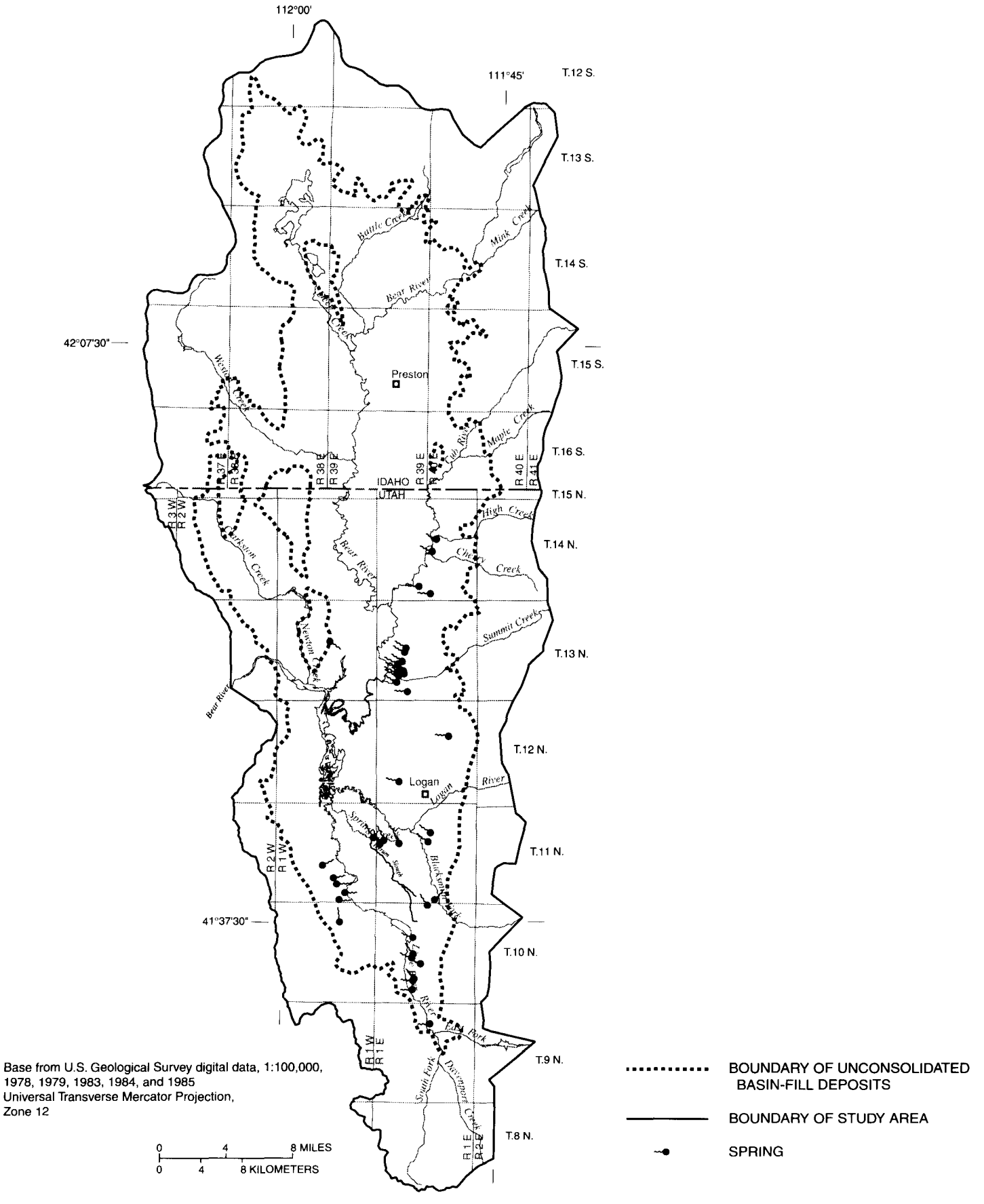


Figure 10. Location of selected springs that discharge from unconsolidated basin-fill deposits in Cache Valley.

tion, the underestimation probably is small because measurements were made during the fall when the effects of evapotranspiration are negligible.

Several sloughs and small streams intercept the potentiometric surface and thus gain ground water. Among these are Oxford Slough, Worm Creek, the slough area north of Weston, Five Mile Creek, Spring Creek (near Franklin), Spring Creek (near College Ward), and Hyrum Slough. Surface water is the primary source of flow in most of these sloughs and streams, but near the center of the valley they cut into the unconsolidated basin-fill deposits and gain ground water. Discharge estimates were available only for Spring Creek (near College Ward) and Hyrum Slough.

A seepage study of Spring Creek (near College Ward) and Hyrum Slough was done in November 1967 by Bjorklund and McGreevy (1971, tables 6 and 7). They determined that Spring Creek and Hyrum Slough gained about 54 cubic feet per second (39,000 acre-feet per year) of water from seepage from the ground-water system, including the discharge from Spring Creek spring numbers 1, 2, 3, and 4.

The estimate of spring discharge from basin-fill deposits of 138 cubic feet per second in 1969 and 1990 (table 5) was computed by adding the discharge of Spring Creek (near College Ward) and Hyrum Slough (54 cubic feet per second), the discharge of the 3 seepage areas (47 cubic feet per second), and the discharge of 23 springs in table 7 that are not accounted for in other seepage studies or in seepage to streams and Cutler Reservoir (37 cubic feet per second).

Total annual evapotranspiration from the ground-water system in Cache Valley was estimated by Bjorklund and McGreevy (1971, p. 35) to be about 108,000 acre-feet. Evapotranspiration occurs mostly in the wet meadowlands in the lower parts of the valley that are estimated to cover about 43,480 acres (Bjorklund and McGreevy, 1971, pl. 4). Because this area includes some areas of diffuse seepage, spring discharge, and irrigation, the estimated 108,000 acre-feet of evapotranspiration from the ground-water system probably also includes some surface water.

This study estimates average annual evapotranspiration from the ground-water system in Cache Valley to be about 87 cubic feet per second (63,000 acre-feet per year). This estimate assumes that evapotranspiration from the ground-water system occurs mainly on 27,480 acres, which equals the 43,480 acres described by Bjorklund and McGreevy (1971, pl. 4) minus any land that is irrigated or that receives water from diffuse seepage and spring discharge. A consumptive use of 2.3 feet per year, on the basis of data for a water-budget model of Cache County by the Utah Division of Water Resources (James Stephens, Utah Division of Water Resources, written commun., 1990), was multiplied by the 27,480 acres to obtain the 63,000 acre-feet per year (87 cubic feet per second). Because these consumptive-use data were not available for the previous study, the new estimate is considered to be more accurate than the estimate from the previous study.

The water-budget model by the Utah Division of Water Resources provided consumptive-use estimates for wetland areas in Cache County, Utah, classified as wet meadow, salt grass, cattails, and cottonwoods. Consumptive-use values were determined by the Utah Division of Water Resources using the method described by Blaney and Criddle (1950). Average consumptive use from the water-budget model, weighted by area of each wetland classification, is 3.4 feet per year; however, 1.1 feet per year of this consumptive use (equal to 80 percent of the average annual precipitation) is water from precipitation that is evaporated or transpired by plants (James Stephens, Utah Division of Water Resources, oral commun., 1992). The remaining 2.3 feet per year is called "depletion" by the Utah Division of Water Resources and is used in this study to represent the average consumptive use of water from the ground-water system in wetland areas. This value also was applied to the Idaho part of Cache Valley on the assumption that a similar distribution of wet meadow, salt grass, cattails, and cottonwoods occurs in Idaho and Utah.

The estimate of 87 cubic feet per second (63,000 acre-feet per year) is used for both 1969 and 1990 because no data were available to refine the 1969 estimate. Because estimated recharge in 1990 was 34 percent less than in 1969, it is probable that discharge from evapotranspiration was somewhat less in 1990 compared to 1969.

About 2,700 wells have been constructed in Cache Valley. The wells range from small-diameter wells for domestic use and stock watering to large-diameter, large-discharge, public-supply wells. About 200 of these wells are in the Idaho part of Cache Valley and 2,500 wells are in the Utah part. In the south-central part of Cache Valley,

where artesian conditions exist, many wells have sufficient hydraulic head to flow year round. Most of these flowing wells are used for irrigation, and several are used for domestic supply, stock watering, or industrial use.

Total annual withdrawal from wells during 1969 and 1982-90 is shown in figure 11. In 1983 and 1984, ground-water withdrawals for irrigation were smaller when precipitation was greater than average (fig. 4). Between 1985 and 1988, pumping for irrigation increased because of generally declining, less-than-average precipitation (fig. 4). The year 1986 was an exception to this trend, with 26.78 inches of precipitation (National Oceanographic and Atmospheric Administration, 1987). After 1987, pumping for public supply increased in response to increased demand in the area around Smithfield, Logan, and Hyrum (fig. 11).

Estimates of withdrawal from wells in 1969 and 1990 are shown in table 8. Values for 1969 are from table 8 of Bjorklund and McGreevy (1971). Values for 1990 were determined from withdrawal from wells in the Utah part of Cache Valley reported by Herbert, Gates, and others (1991) and from estimates of withdrawal from wells in Idaho that were computed for this study. In 1969, about 44 cubic feet per second (32,000 acre-feet per year) of water was withdrawn from wells in Cache Valley (table 5). Of that amount, about 17 cubic feet per second (12,000 acre-feet per year) was from pumped wells, and about 28 cubic feet per second (20,000 acre-feet per year) was from flowing wells (Bjorklund and McGreevy, 1971, table 8).

Table 8. Withdrawal from wells in Cache Valley, 1969 and 1990

[In acre-feet]

Use	[1]1969			1990		
	Pumped	Flowing	Total	Pumped	Flowing	Total
Domestic and stock	175	2,350	2,525	0	1,892	1,892
Public supply	3,000	400	3,400	9,601	612	10,213
Industrial	800	6,650	7,450	1,208	6,542	7,750
Irrigation	8,000	10,500	18,500	6,831	11,341	18,172
Subtotal (rounded)	12,000	20,000	32,000	18,000	20,000	38,000

[1]Values for 1969 are from Bjorklund and McGreevy (1971, table 8) and are reported rounded to the nearest 25 acre-feet.

Total withdrawal from wells in 1990 was about 52 cubic feet per second (38,000 acre-feet per year). Withdrawal from pumped wells was about 25 cubic feet per second (18,000 acre-feet per year), and withdrawal from flowing wells was about 28 cubic feet per second (20,000 acre-feet per year). Withdrawal in 1990 from public-supply and industrial-use wells was compiled from reported values. Withdrawal from irrigation wells was computed by the energy-consumption method or was reported. Withdrawal for domestic and stock use was estimated to be 20 percent of the flow from flowing wells in the Utah part of Cache Valley. Information was not available on domestic and stock water use from wells in the Idaho part of the valley. Withdrawal from flowing wells was estimated by multiplying total withdrawal from flowing wells for 1969 by the average percent decrease in the flow of eight wells that have been measured annually since 1969 (L.R. Herbert, U.S. Geological Survey, written commun., 1991). The amount of withdrawal from flowing wells in 1990 is a rough estimate because it does not account for withdrawal from new flowing wells or the loss of flow from flowing wells that have been destroyed or have stopped flowing.

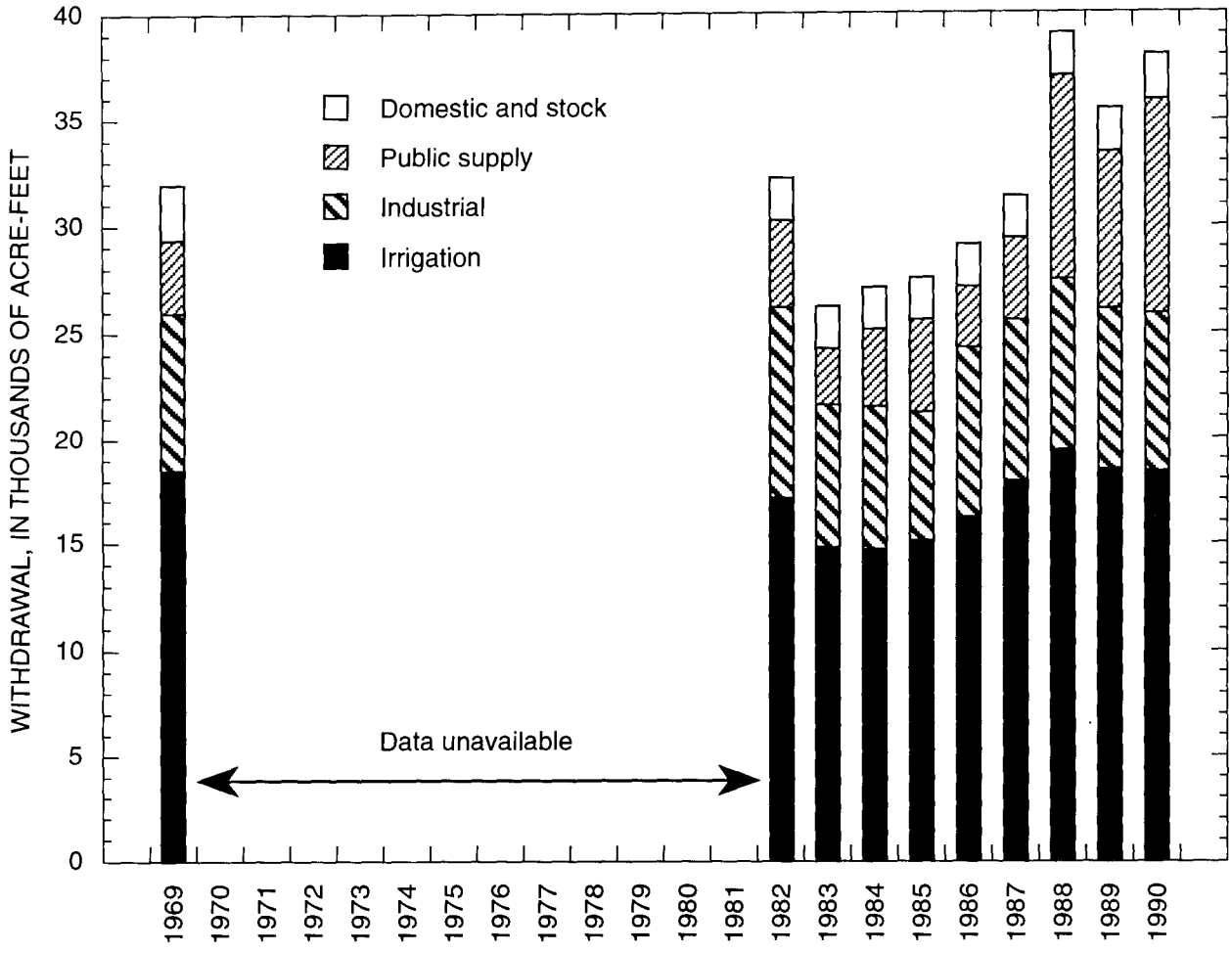


Figure 11. Total annual withdrawal from wells in Cache Valley, 1969 and 1982-90.

Water-Level Fluctuations

Water-level fluctuations result from long-term changes and seasonal or shorter-term changes in recharge and discharge to the ground-water system. Fluctuations are larger at locations closer to sources of recharge and well withdrawal and smaller at locations remote from these sources. Hydrographs for 1990-91 of selected wells completed in unconsolidated basin-fill deposits in Cache Valley are shown in figure 12.

Long-term water-level data were available for seven observation wells in Cache Valley. Hydrographs of these wells indicate that long-term water levels did not change substantially during 1945-82. Water levels in most wells throughout Cache Valley, such as in well (A-12-1)29cab-1, rose during 1982-84 as a result of greater-than-average precipitation (fig. 4). Average annual precipitation at Utah State University in Logan during 1982-84 was 29.27 inches (National Oceanic and Atmospheric Administration, 1983, 1984, and 1985), more than 1.5 times the 1941-90 average annual precipitation (fig. 4) of 18.65 inches. Water levels steadily declined in these wells during 1984-90 in response to a return to normal precipitation after 1985 and subsequent less-than-average precipitation during 1987-90. In the southeastern part of Cache Valley, increased pumping from irrigation and public-supply wells also contributed to water-level declines during 1984-90.

Water levels in wells also fluctuate seasonally as illustrated by short-term water-level data. During 1990, some wells in the northern part of the valley, such as well 16 S 40E 29CCB1, and in the extreme southern end of the valley, such as well (A-10-1)16dad-1, were unaffected by pumping. Water levels in these wells were highest in June, July, and August, possibly in response to recharge from infiltration of unconsumed irrigation water. Water levels in these wells were typically lowest in January and February.

Water levels in wells in the southeastern part of the valley, such as wells (A-11-1)27cdc-1 and (A-12-1)3bbb-1, were affected by pumping during July and August. These wells are located between Richmond and Hyrum, an area that has experienced increased ground-water use for public supply in recent years. Water levels in these wells were lowest during the summer because peak water use for public supply typically occurs during summer. Water levels also may have been affected by pumping of ground water for irrigation, which also would have peaked during summer.

Hydrographs for wells (B-11-1)35cca-1 and (B-12-1)15adc-1 are typical of hydrographs of wells in the southwestern part of the valley because they show no consistent seasonal pattern of water-level fluctuation. Wells in this area are typically flowing wells, and water-level fluctuations are not as large as in wells affected by pumping.

Water levels of wells less than 50 feet deep, such as 14S 39E 29DAD 1 and 15S 39E 15CAD 1, fluctuated more from one month to the next than water levels of wells deeper than 50 feet. Because of the shallow depths, water levels in these wells are more susceptible to local, transient sources of recharge at the surface, such as irrigation in nearby fields.

Storage

Bjorklund and McGreevy (1971, p. 34) estimated that about 25,000 acre-feet of ground water per foot of water-level decline could be recovered from the upper part of the saturated material in Cache Valley. Declines in water levels, however, could locally reduce or even eliminate ground-water discharge from flowing wells and springs, seepage to streams, and evapotranspiration. Data are not available to provide a better estimate of ground-water storage in Cache Valley.

At the time of their study during 1967-69, Bjorklund and McGreevy (1971, p. 27 and 56) stated that ground-water levels in unconsolidated basin-fill deposits were stable and had not changed substantially since 1935 and that the natural recharge-discharge relation had not been changed appreciably by withdrawal from wells. The ground-water system therefore was assumed to be in steady state in 1969, and no change in storage was implied. This steady state is reflected in the storage value listed in table 5. The amount of water from "Other forms of recharge" in table 5 was assumed to be constant between 1969 and 1990. The amount of water left after subtracting total discharge from total recharge in 1990 is assumed to have been derived from storage (table 5). This amount, however, is not

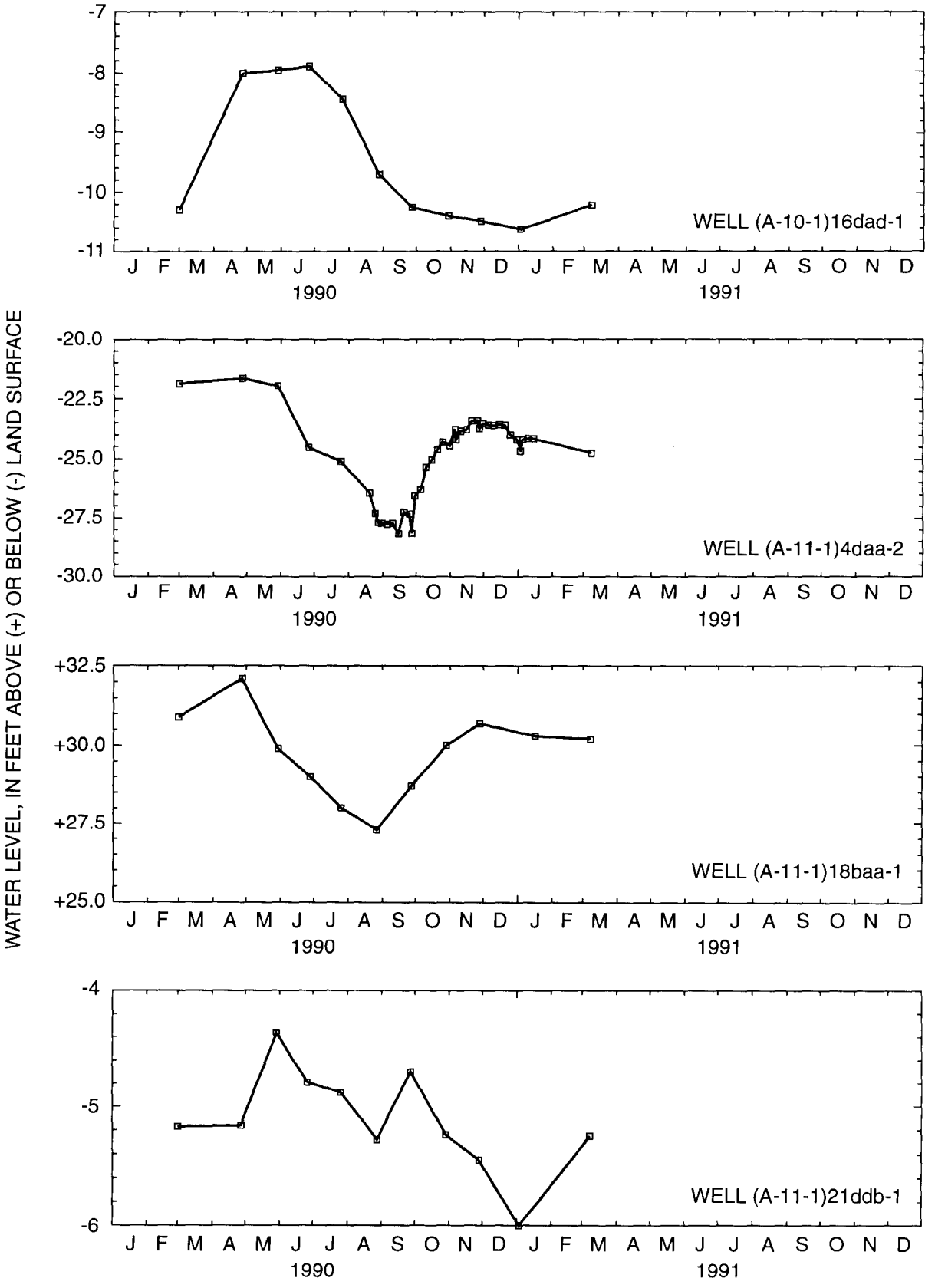


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91.

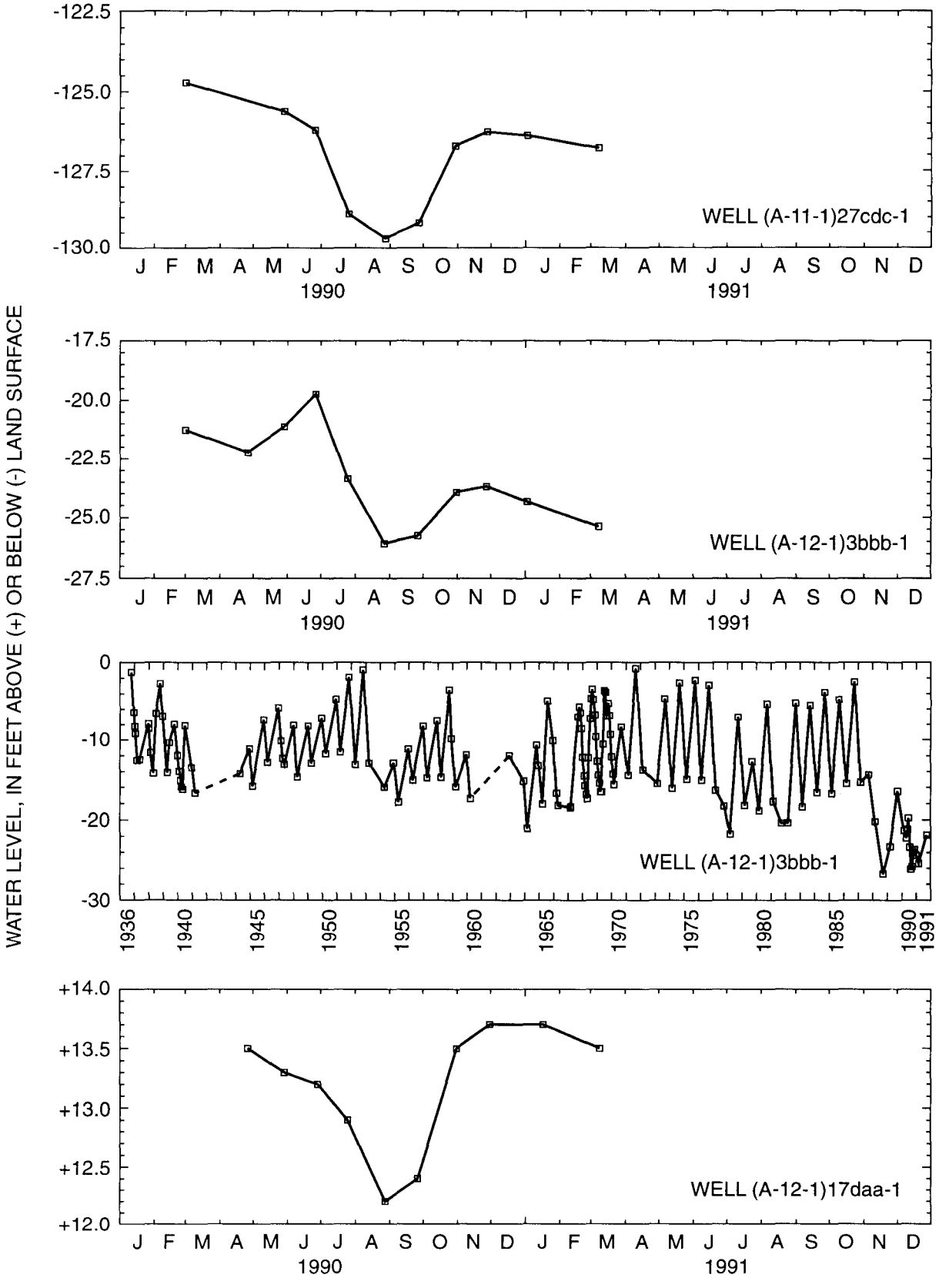


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

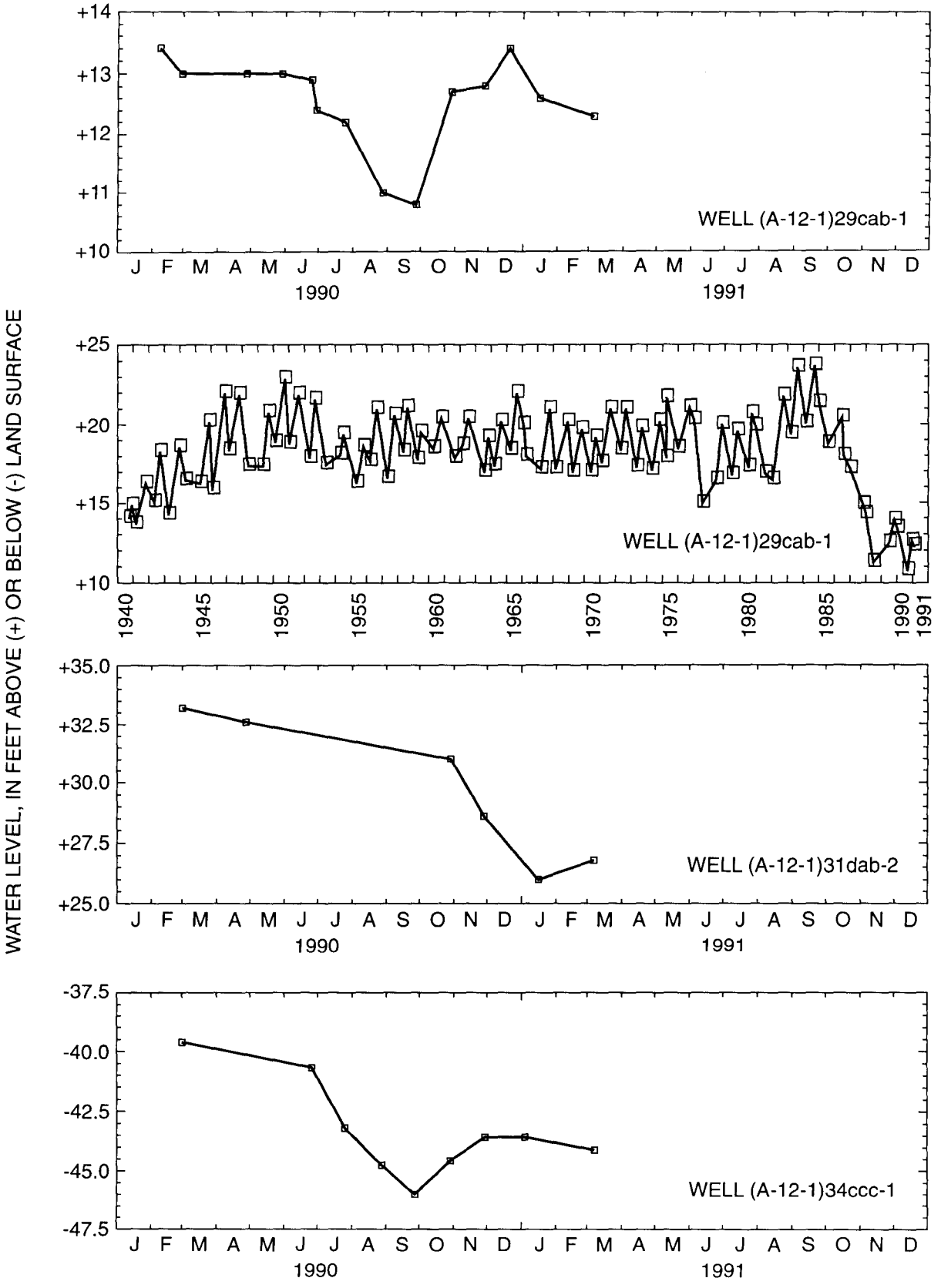


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

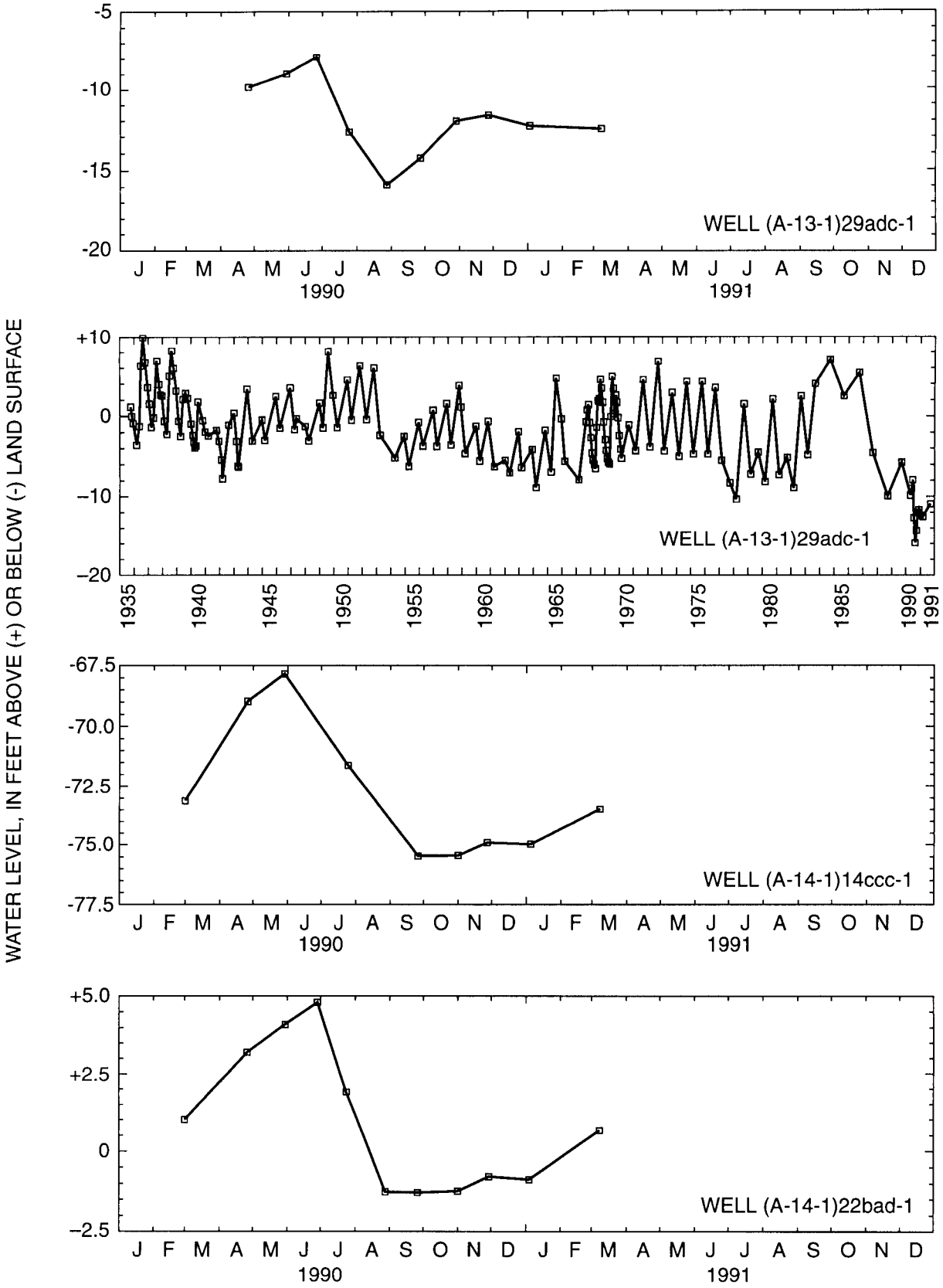


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

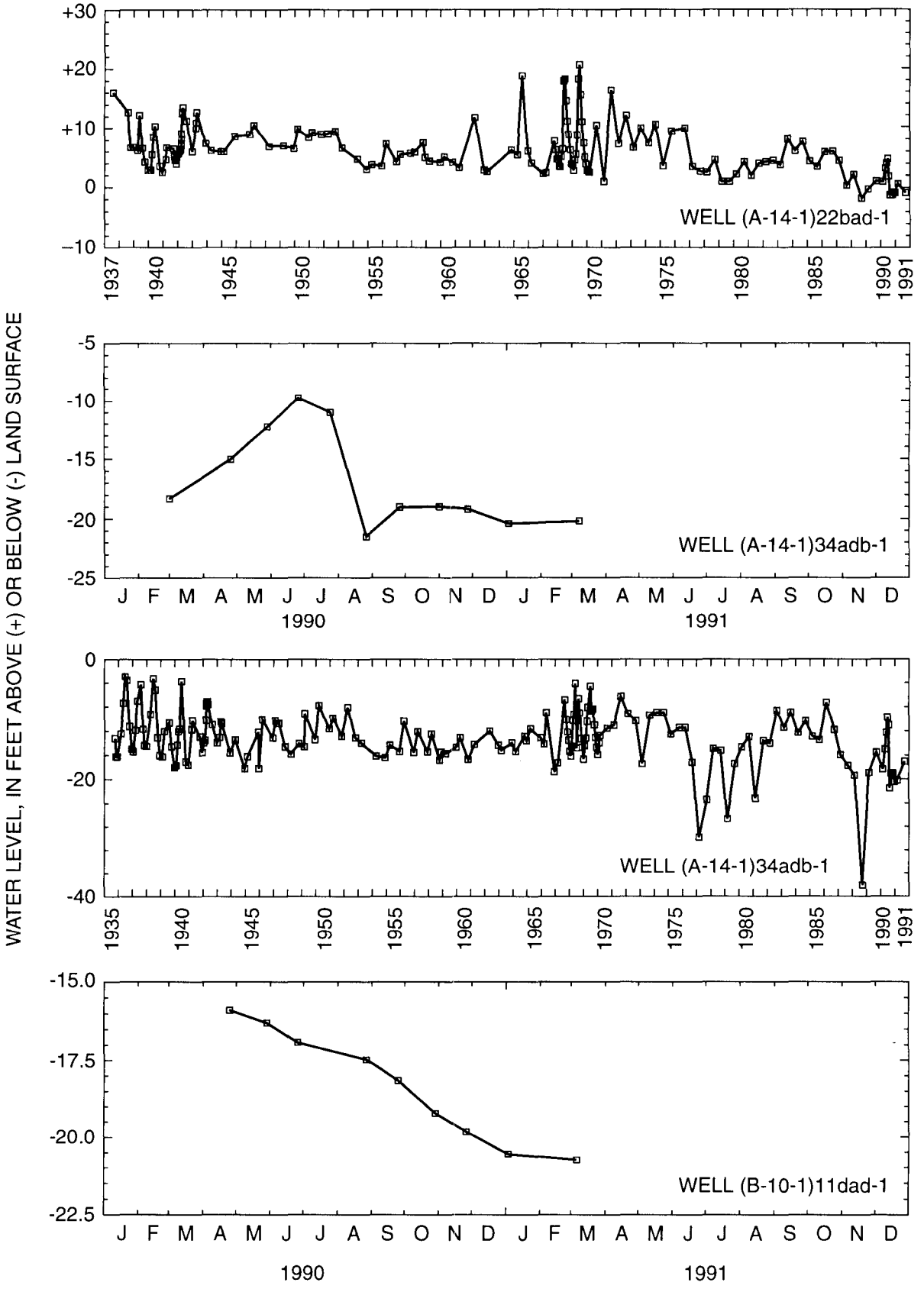


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

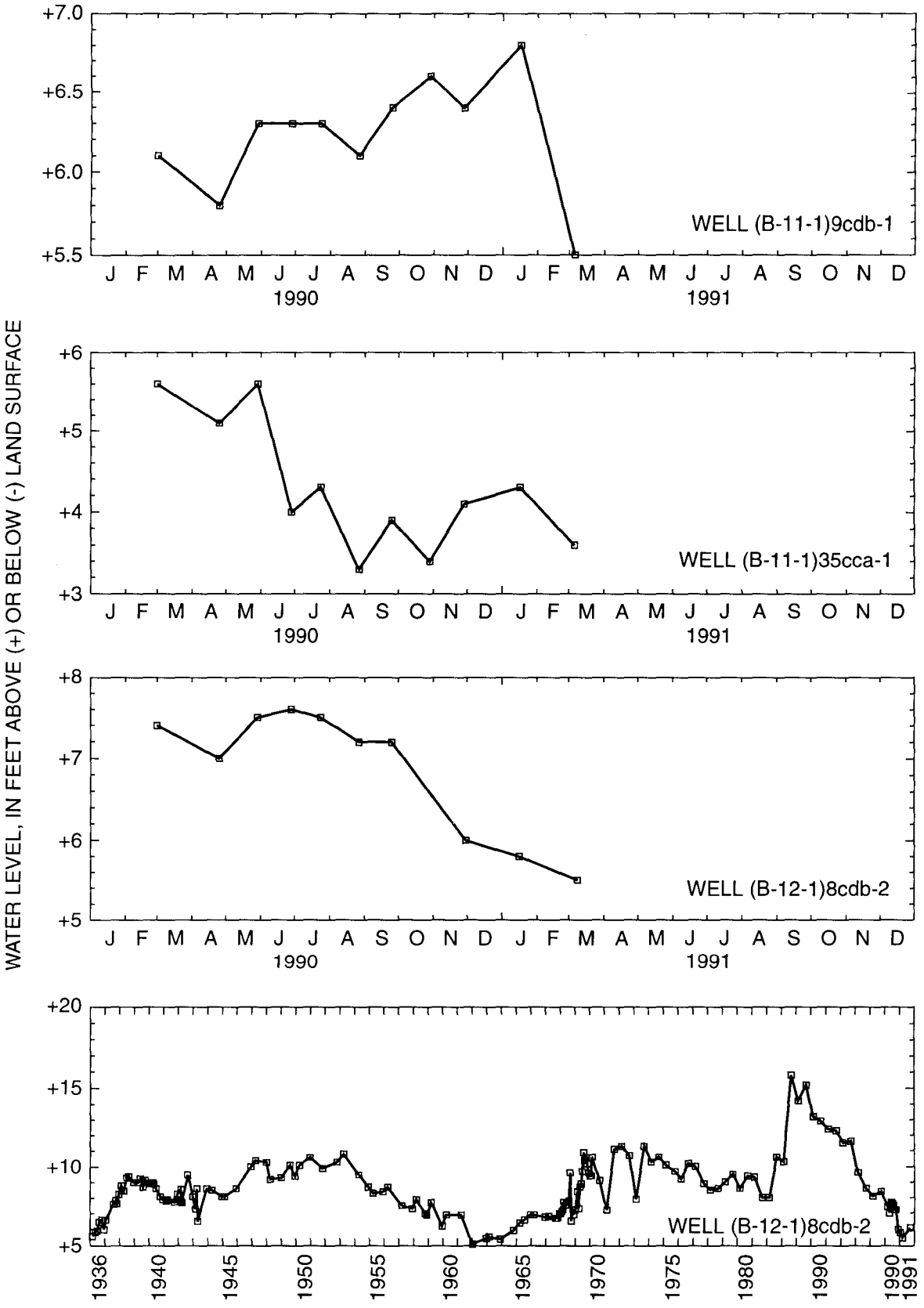


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

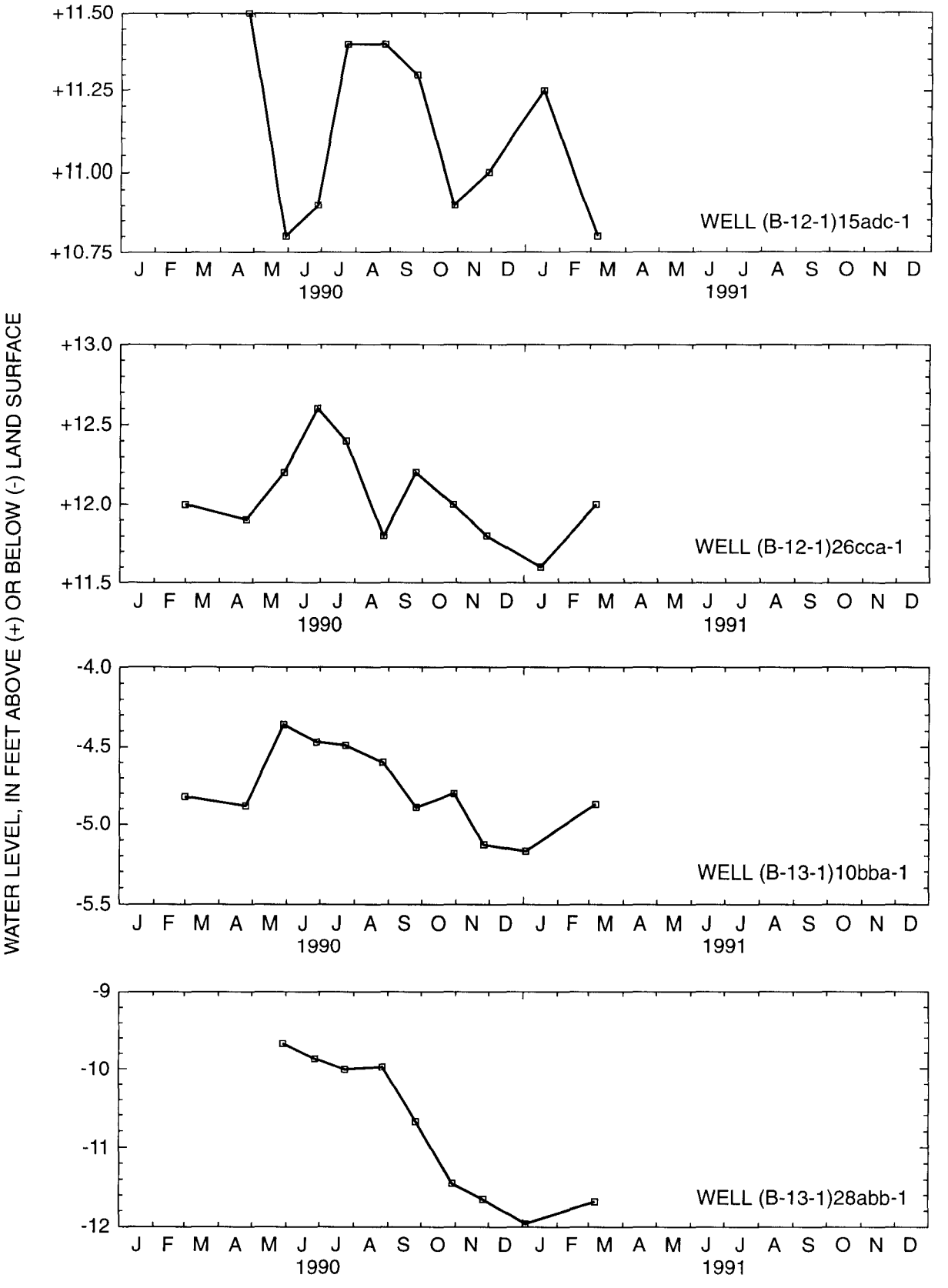


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

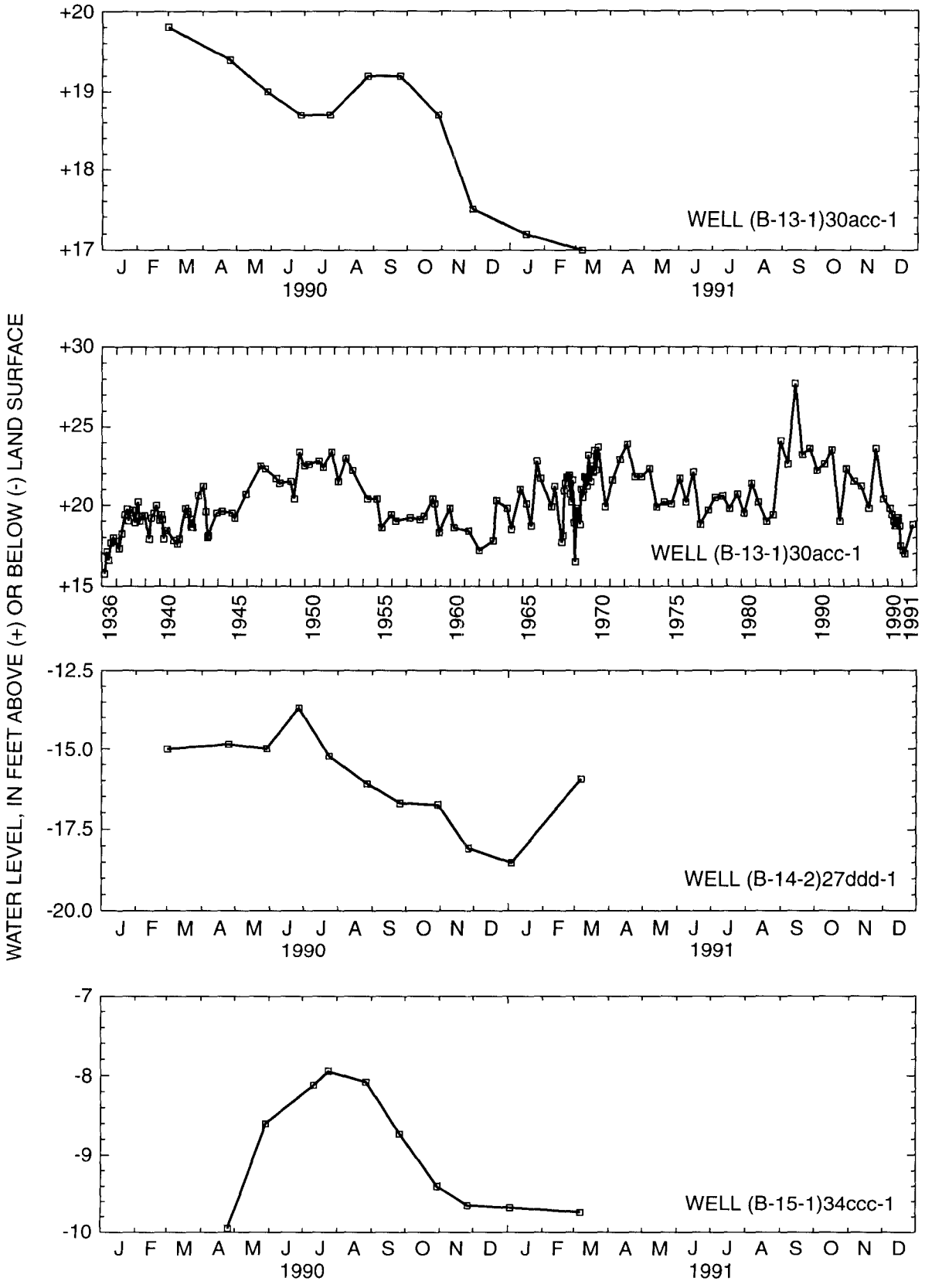


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

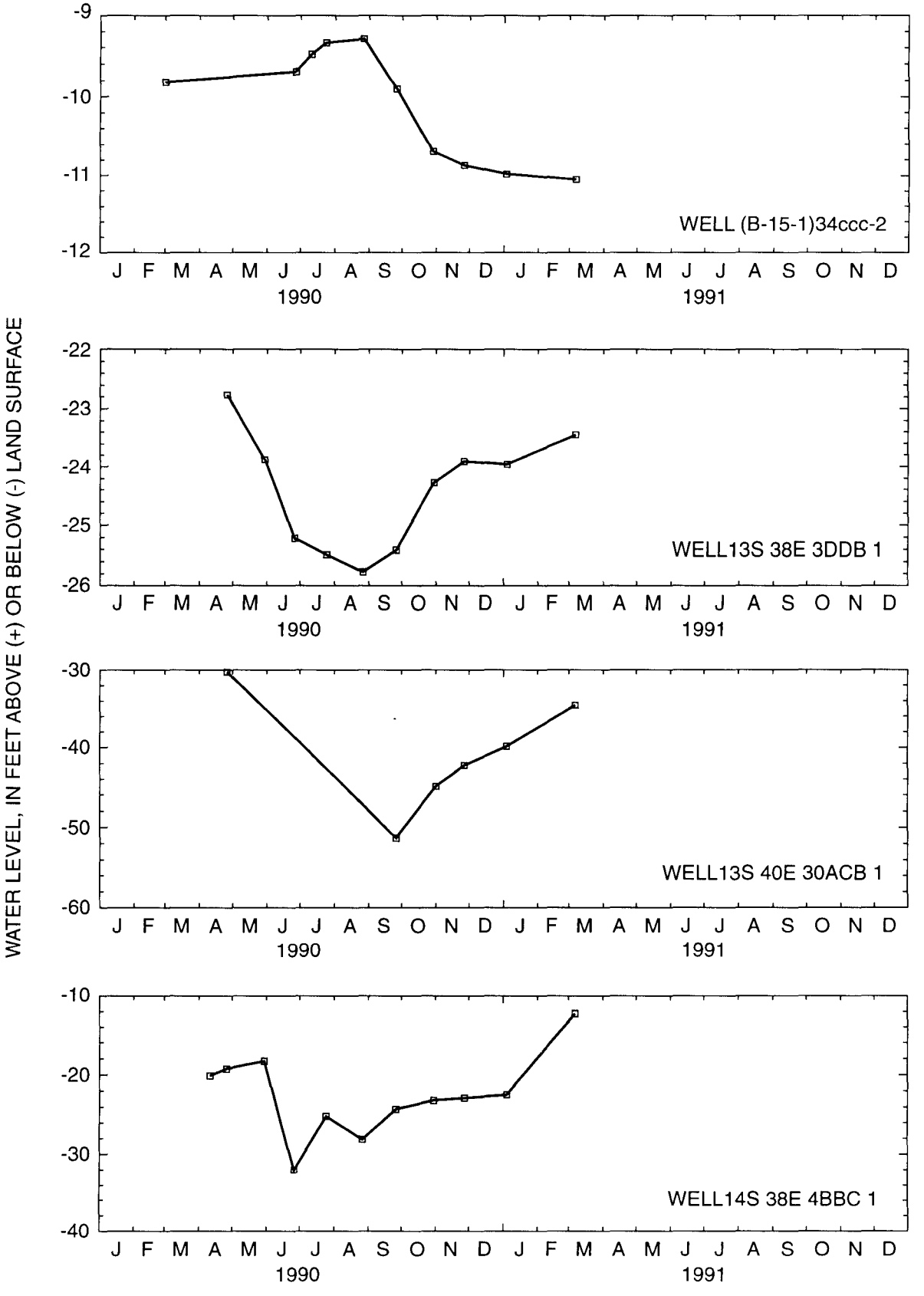


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

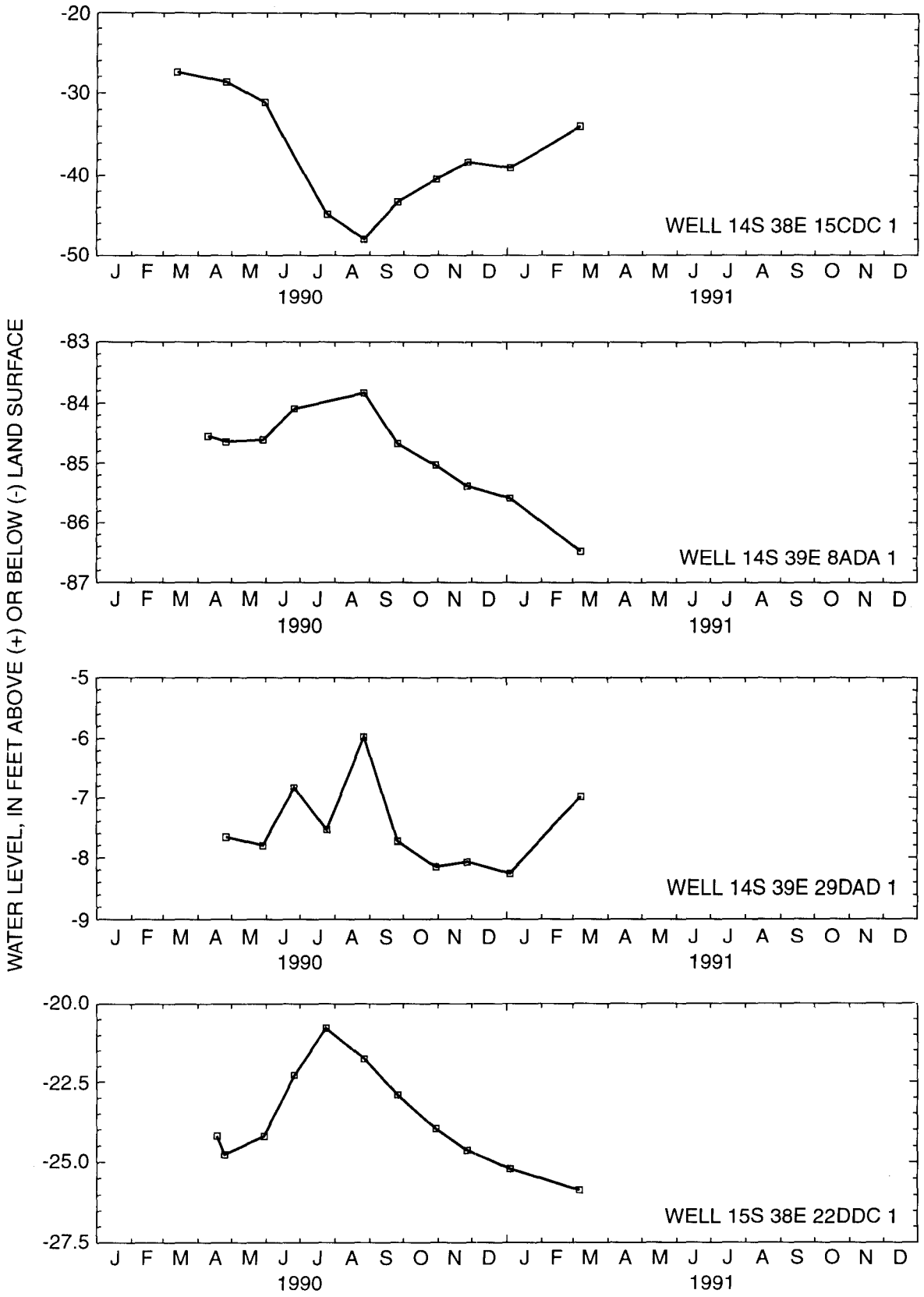


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

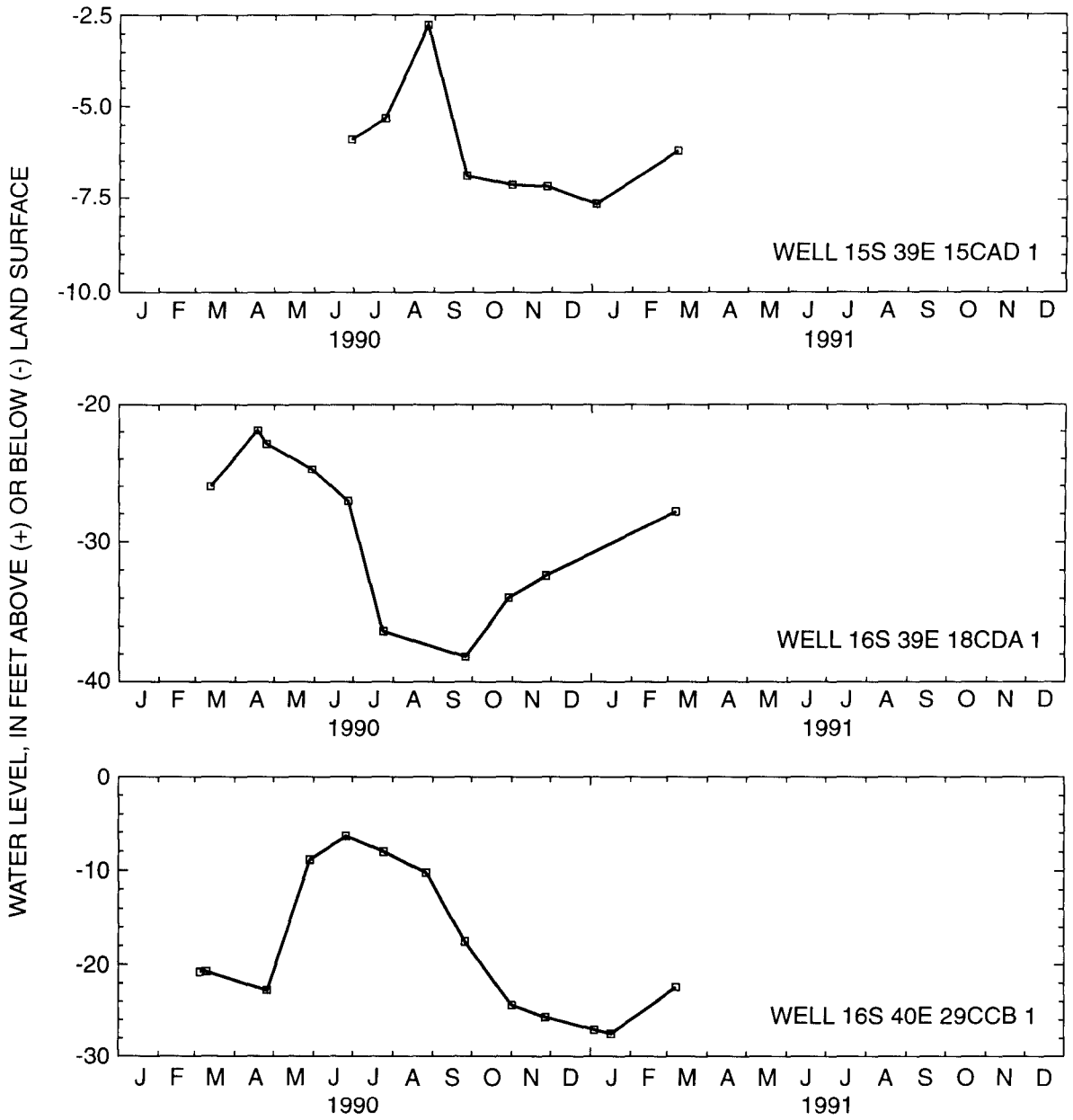


Figure 12. Water levels in selected wells completed in unconsolidated basin-fill deposits in Cache Valley, 1990–91—Continued.

representative of a typical year because recharge in 1990 was not typical. Recharge from infiltration of precipitation and unconsumed irrigation water was much less than normal in 1990 (table 5) because of less-than-average precipitation in 1990 (fig. 4). Water-level declines during 1990-91 (fig. 12) also indicate removal of water from storage during 1990.

The amount of water derived from storage in 1990 also can be computed by assuming an average valley-wide decline in water levels during 1990 in the unconfined parts of the unconsolidated basin-fill ground-water system. Water levels from five wells (wells (A-10-1)16dad-1, (A-11-1)21ddb-1, (A-11-1)27cdc-1, (A-12-1)34ccc-1, and 16S 40E 29CCB1), three of which are in the shallow, unconfined area in the center of the valley and two of which are in the unconfined area near the mountain front, indicate an average decline of 2 feet (four of the five wells showed water-level declines). A specific yield of 0.2 may be assumed to be an average value between the specific yield of 0.1 for silty clays in the center of the valley and 0.3 for gravels along the valley margins (Johnson, 1967). If a 2-foot decline is representative of water-level changes throughout the unconfined part of the ground-water system and the area of unconsolidated basin-fill deposits is 317,300 acres, then the amount of water derived from storage in 1990 would be 175 cubic feet per second (126,900 acre-feet). This estimate is similar in magnitude to the "out of storage" value of 161 cubic feet per second (116,600 acre-feet per year) listed in table 5.

Aquifer Properties

Bjorklund and McGreevy (1971, table 5) described in detail the aquifer properties of unconsolidated basin-fill deposits in Cache Valley. In general, unconsolidated basin-fill deposits are most transmissive along the east bench area between Richmond and Hyrum. Generally, transmissivity of the unconsolidated basin-fill deposits decreases toward the center of the valley near the Bear River and Cutler Reservoir, where fine-grained silts and clays predominate.

Because improved interpretive methods are now more widely used than when aquifer tests reported by Bjorklund and McGreevy (1971, table 5) were done, older values may not be comparable to newer values obtained from more recent tests. For example, Lohman (1979, p. 32) analyzed aquifer-test data using the Theis curve (Theis, 1935), which does not account for leakage from confining layers. This analysis produced apparent transmissivity values from 5 to 20 times greater and storage coefficient from 17 to 25 times greater than the realistic values determined using the Hantush Modified Method for leaky confined aquifers (Hantush, 1960, and Lohman, 1979, p. 32-34), which does account for leakage from confining layers. Because the method used to interpret the aquifer-test data listed by Bjorklund and McGreevy (1971, table 5) was not published and because the original data and test results could not be located, it is not known whether leakage from confining layers was considered in their analyses.

Transmissivity and storage-coefficient values were determined more recently from an aquifer test done by the U.S. Geological Survey in March 1992. The aquifer test included wells (B-12-1)36bca-1, (B-12-1)36bca-2, and (B-12-1)36bdb-1, located about 4 miles west of Logan. The data were analyzed using the Hantush Modified Method for leaky confined aquifers (Hantush, 1960). The transmissivity value determined from the test was about 44,000 feet squared per day and the storage coefficient was about 8×10^{-3}.

Transmissivity values of unconsolidated basin-fill deposits at 131 selected wells were estimated from specific-capacity values determined from drillers' logs using the method described by Theis and others (1963). Wells with drillers' logs, which provided complete well-construction information and yield and drawdown or water-level-recovery data, were selected for analysis. Specific-capacity values of pumped wells were determined by dividing withdrawal during pumping by drawdown. Specific-capacity values of flowing wells were determined by dividing withdrawal by the rise in head after flow was stopped. Only wells that were pumped or allowed to flow for a minimum of 1 hour were selected for analysis.

The number of values within computed ranges of transmissivity for Cache Valley is shown in figure 13. Transmissivity values ranged from 1 to about 134,000 feet squared per day. About 97 percent of the values were between 1 and 50,000 feet squared per day.

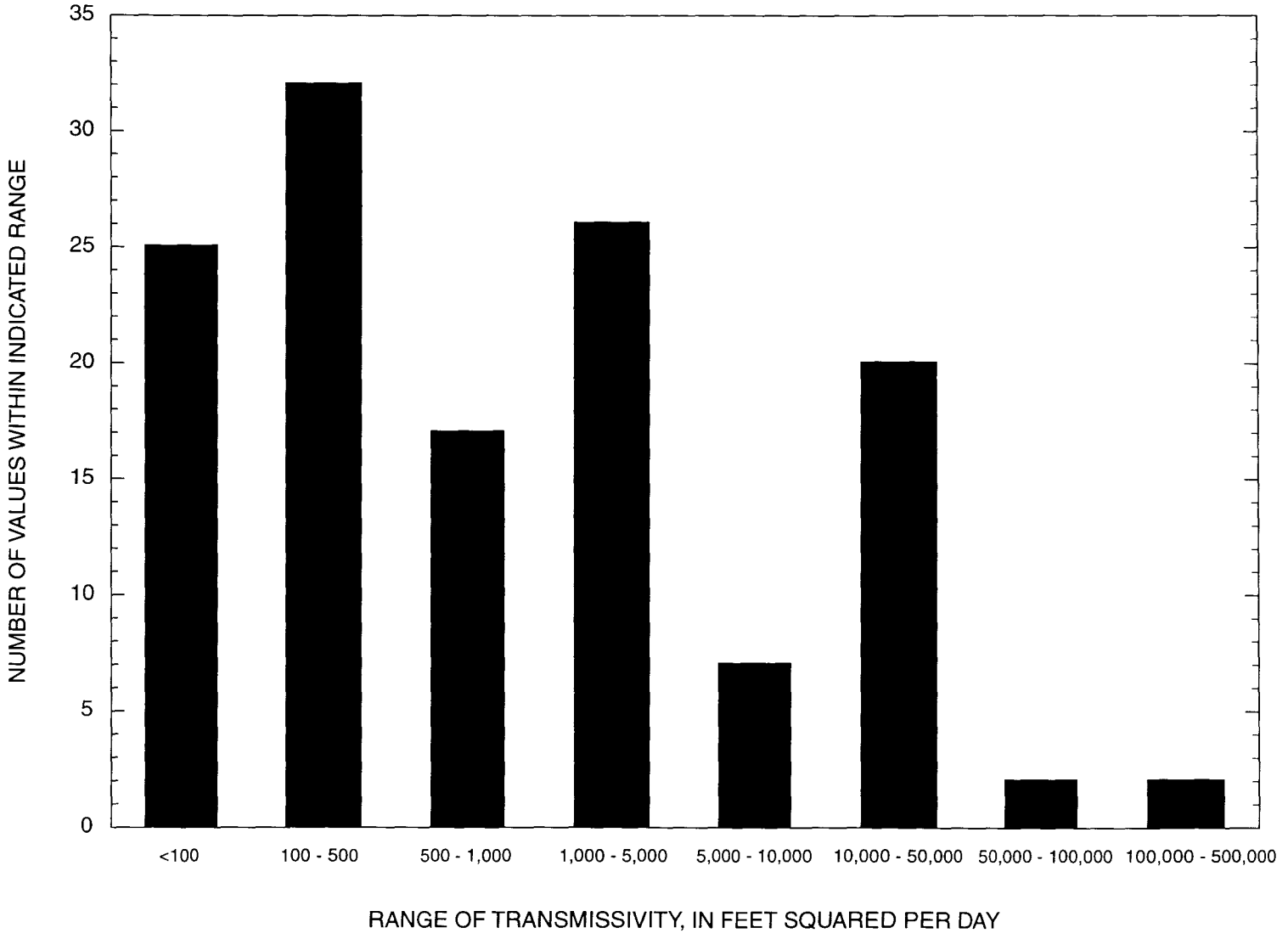


Figure 13. Number of values in computed ranges of transmissivity for the unconsolidated basin-fill deposits in Cache Valley.

The areal distribution of transmissivity estimates is shown in figure 14. The largest values were generally in the east bench area in Utah, between Richmond and Hyrum. Because well efficiency affects specific capacity, the estimates of transmissivity determined from specific-capacity values typically underestimate the true transmissivity of the aquifer (Theis and others, 1963). Wells drilled for production purposes are commonly screened in the coarser, more productive zones, and extrapolation of these data to unscreened zones may result in an overestimate of the transmissivity. Results of the aquifer test near Logan indicate a much higher transmissivity than would be expected if only specific-capacity data were used (W.F. Holmes, U.S. Geological Survey, written commun., 1992).

On the basis of aquifer-test results and estimates of specific-capacity values of wells, Bjorklund and McGreevy (1971, p. 48-53 and table 5) estimated average values or ranges of transmissivity. In general, the areal distribution of transmissivity estimates in figure 14 is comparable to the distribution of transmissivity estimated by Bjorklund and McGreevy (1971, p. 48-53), although the absolute values of transmissivity are probably smaller than those determined from aquifer tests.

Changes in the Ground-Water System since 1967-69

Changes in the ground-water system since 1967-69 include a general water-level decline, an increase in ground-water withdrawal, and a possible decrease in spring discharge. These changes are related to either natural or man-made stresses on the ground-water system. Water levels generally declined throughout Cache Valley from March 1969 to March 1991, with the exception of a small rise in the northern end of the valley and a small rise in the southern end (fig. 15).

The general decline in water levels throughout the valley could partly result from less-than-average precipitation in Cache Valley during 1988-90 (fig. 4). Declines in water levels greater than 10 feet occurred in the northwest part of the valley and on the southeast side of the valley. The water-level declines in the northwest part of the valley might be a result of withdrawal from new irrigation wells that were drilled in the area during the 1970s. Along the southeast side of the valley, the towns of Smithfield, Hyde Park, North Logan, Logan, River Heights, Providence, Millville, Nibley, and Hyrum experienced a substantial increase in population, from 33,869 in 1970 to 56,102 in 1990 ([U.S.] Bureau of Census, 1971, and Utah Office of Planning and Budget, 1991). From 1987 to 1990, pumping for public supply (fig. 11) increased sharply to meet the needs of the increased population in the area, and total pumpage for these towns increased from 3,800 acre-feet in 1987 to 8,746 acre-feet in 1990 (William Smart, Utah Department of Natural Resources, written commun., 1988 and 1991). This increased pumping probably accounts for a large part of the decline in water levels in the area.

Discharge from seepage to streams and reservoirs and from springs probably fluctuates from year to year. Because water levels declined in Cache Valley since 1967-69, decreases in seepage to streams and reservoirs and discharge to springs have probably occurred; however, annual discharge measurements are not available for Cache Valley to substantiate this contention. Discharge measurements from springs listed in table 7 indicate a possible decline in discharge from 1967-69 to 1990. The decline in spring discharge might be a result of less water being available as recharge to the ground-water system because of the general decline in precipitation since 1986 (fig. 4). Declines in discharge of individual springs could possibly result from increased withdrawals from adjacent wells or declines in recharge from nearby sources such as irrigation recharge; however, the mechanisms of ground-water flow to individual springs are not sufficiently understood for any of these springs.

Ground-Water Quality

Extensive sampling of wells and springs was done in 1967-69 as part of the previous study of Cache Valley by the U.S. Geological Survey (McGreevy and Bjorklund, 1970, tables 1, 2, and 5). Bjorklund and McGreevy (1971, p. 38) described ground water in Cache Valley as being of generally good quality and stated that the quality had not changed since 1941. They divided Cache Valley into 11 areas of different ground-water conditions and characterized the ground-water quality in each (Bjorklund and McGreevy, 1971, p. 48-53, fig. 13). Their analysis of ground-water quality in these 11 areas is summarized in table 9. Ground water in Cache Valley has not been sam-

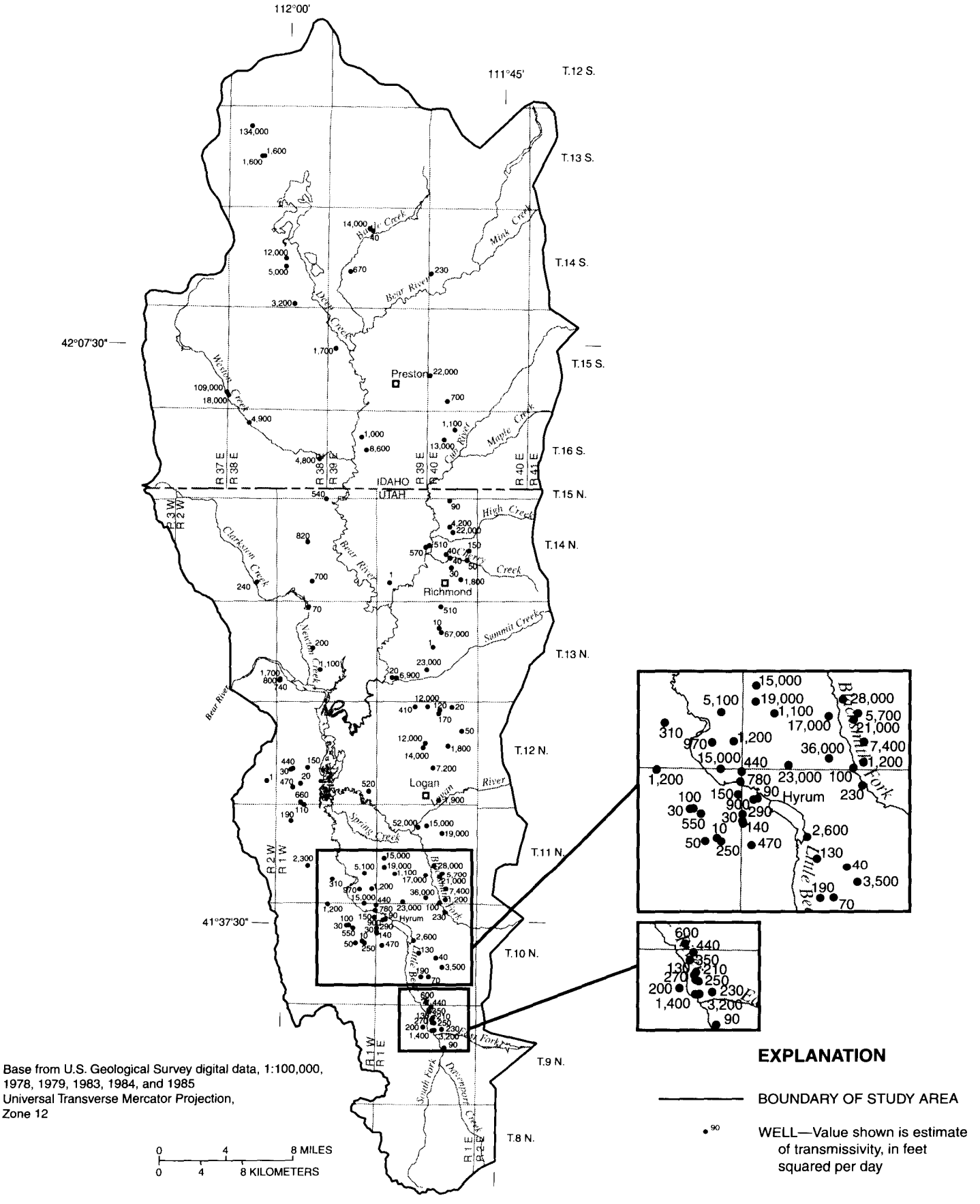
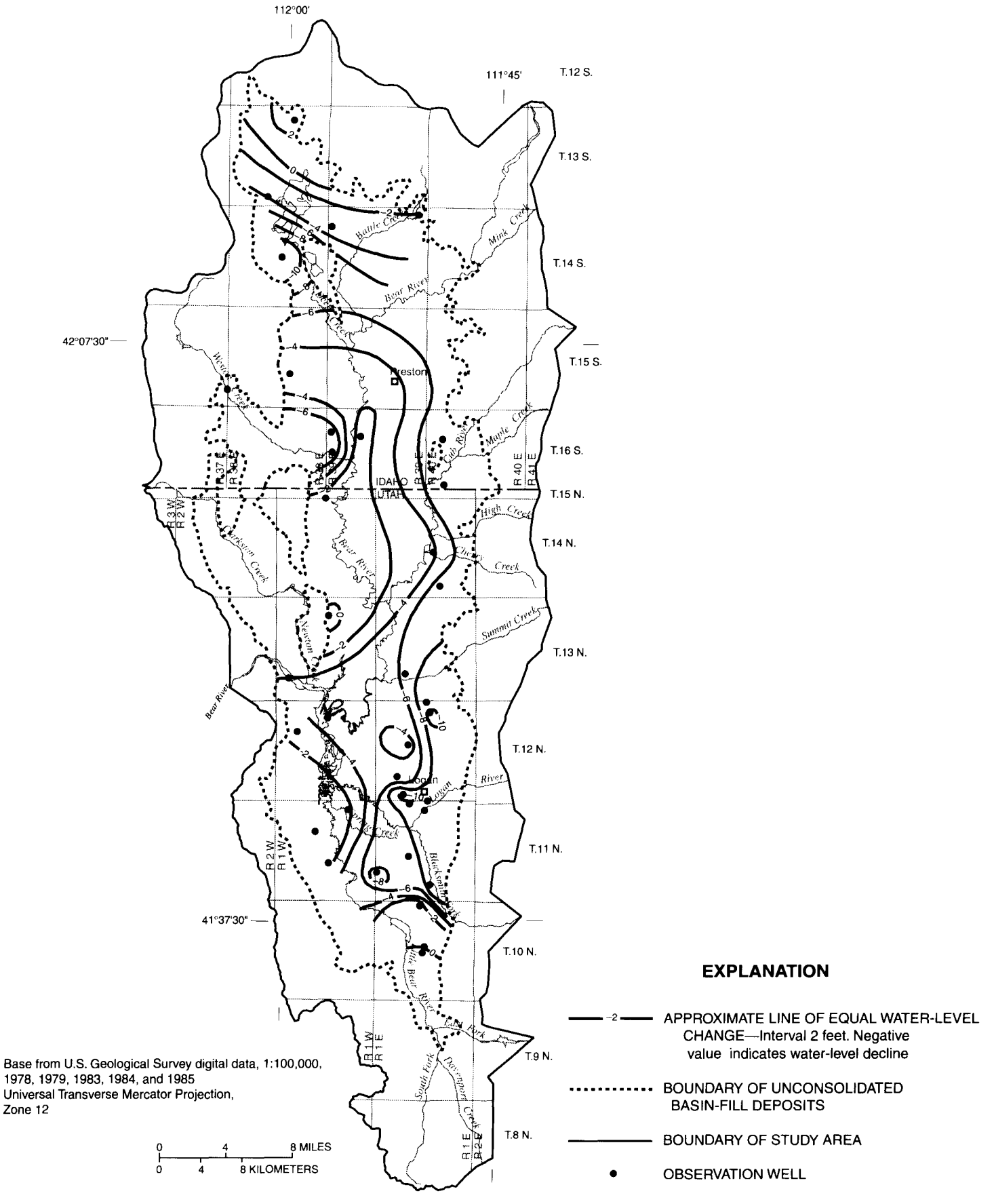


Figure 14. Areal distribution of transmissivity estimated from specific-capacity values for unconsolidated basin-fill deposits in Cache Valley.



Base from U.S. Geological Survey digital data, 1:100,000, 1978, 1979, 1983, 1984, and 1985
Universal Transverse Mercator Projection, Zone 12

Figure 15. Approximate change in water levels in unconsolidated basin-fill deposits in Cache Valley from March 1969 to March 1991.

Table 9. Summary of ground-water quality in 11 areas described by Bjorklund and McGreevy (1971)
[mg/L, milligram per liter; <, less than]

Area, as described by Bjorklund and McGreevy (1971, fig. 13)	Concentration of dissolved solids (mg/L)	General comments
Smithfield-Hyrum-Wellsville	generally < 400	Quality of water is generally good.
Little Bear River area south of Hyrum	< 400	Chemical quality of water is generally good.
Wellsville to Newton	400 to 800	Ground-water conditions are poorly known.
Lower Little Bear River-Benson-The Barrens	400 to 1,200	Concentration of dissolved solids 400 to 800 mg/L in the area along the Little Bear River but as high as 1,200 mg/L near Benson and The Barrens.
Cub River subvalley	300 to 800	
Clarkston	300 to 900	
Weston Creek subvalley	300 to 800	
Dayton-Banida-Swan Lake	200 to 1,500	
Preston	300 to 1,600	Concentration of dissolved solids in water from delta deposits generally ranges from about 400 to 1,600 mg/L. Concentration of dissolved solids as low as about 300 mg/L in a few wells near Whitney.
Bear River inner valley	generally 400 to 800	
Fairview-Lewiston-Trenton	generally 800 to 1,600	Quality of water varies from place to place and from season to season. Irrigation practices generally affect the chemical quality of water.

pled extensively by the U.S. Geological Survey since the 1967-69 study, and no attempt was made to resample water from wells to evaluate ground-water quality as part of this study.

Dissolved-solids concentration in water from two wells, (A-11-1)8dda-3 and (A-13-1)29bcd-1, during 1960-90 is shown in figure 16. The dissolved-solids concentration was determined from the sum of constituents in the sampled water. The dissolved-solids concentration in both wells was less than 500 milligrams per liter, which is one of the secondary drinking-water standards (Utah Department of Health, 1986) that defines aesthetic limits for specific constituents in drinking water. The small changes in dissolved-solids concentration through time indicate that ground-water quality, in terms of inorganic constituents, has not changed substantially since 1961 in well (A-11-1)8dda-3 and since 1970 in well (A-13-1)29bcd-1.

Specific conductance of water is frequently used as an indicator of dissolved-solids concentration. Bjorklund and McGreevy (1971, fig. 10) established a linear relation of about 0.6:1 between dissolved-solids concentration in milligrams per liter to specific conductance in microsiemens per centimeter at 25 degrees Celsius for ground water in Cache Valley. Temperature and specific conductance measurements of water from wells and springs in 1967-69 by McGreevy and Bjorklund (1970, tables 1 and 2) are reported with measurements made in 1990-91 by Roark and Hanson (1992, table 6) in table 10. Temperature and specific conductance of water from these wells and springs generally have not changed much from 1967-69 to 1990-91. This lack of change is consistent with the previously mentioned (fig. 16) observation that concentrations of inorganic constituents have not changed much since 1967-69.

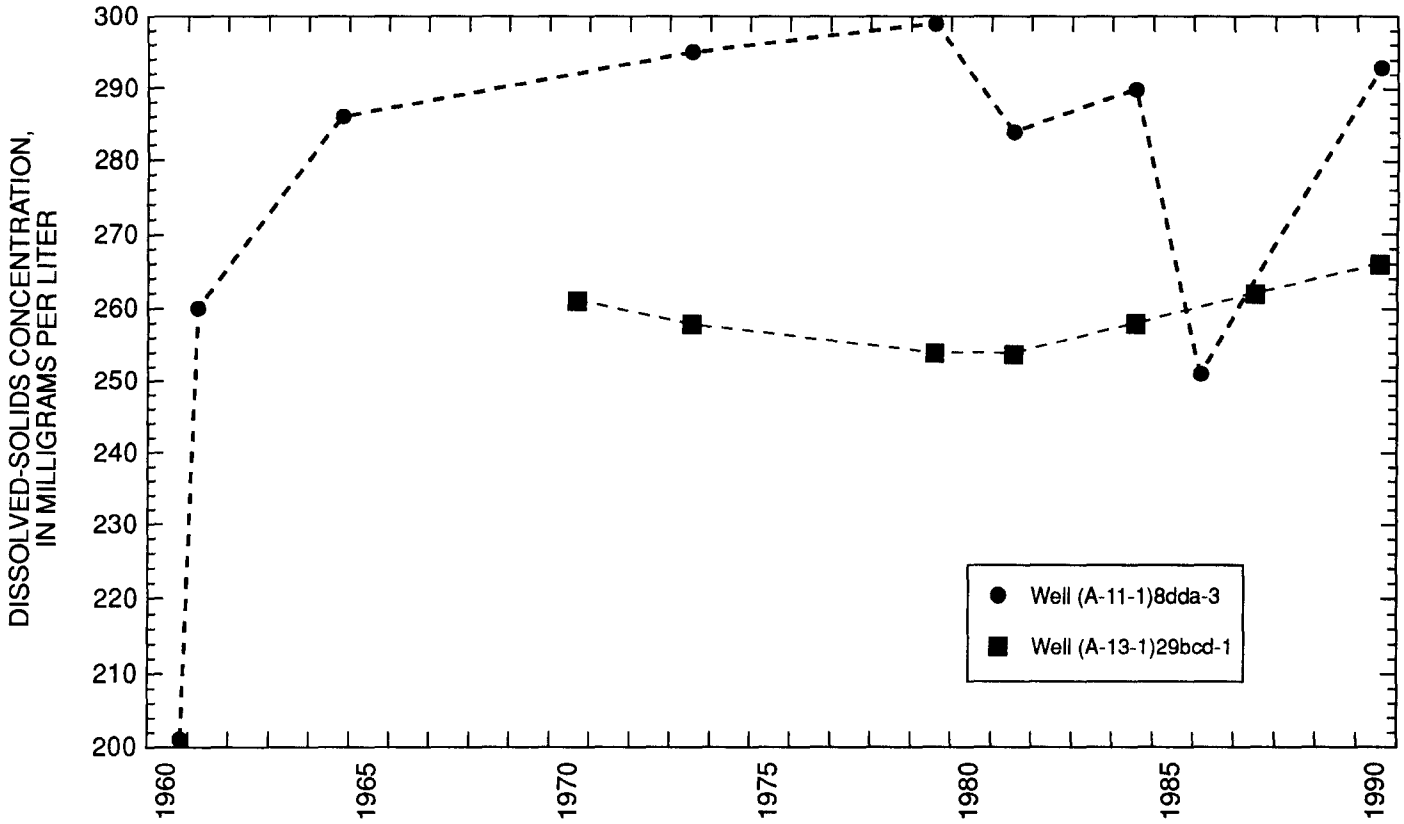


Figure 16. Dissolved-solids concentration in water from two wells in Cache Valley, 1960–90.

Because water-quality measurements for this study were limited by the scope of the study, only general conclusions could be made about water quality or changes in water quality in Cache Valley. The wells listed in table 10 are all flowing wells that are located primarily in the south-central part of the valley, where the majority of ground-water discharge occurs; therefore, any changes in ground-water quality outside this area would not be evident from data in table 10. Also, because wells in Cache Valley have not been systematically sampled for organic chemicals or pesticides, conclusions cannot be made about the presence of these compounds in the ground water of Cache Valley.

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

A ground-water model is a tool used to simulate a simplified version of a ground-water system. Much can be learned about the physical aspects of a ground-water system by constructing and calibrating such a model. Models can be used to compare the probable effects of different stresses on the ground-water system. Constructing a ground-water-flow model also can show areas where additional field data need to be collected. A model of the ground-water system of Cache Valley was constructed for these reasons.

A modular, three-dimensional, finite-difference, ground-water-flow model developed by McDonald and Harbaugh (1988) was used to simulate flow in the ground-water system in the saturated unconsolidated basin-fill deposits of Cache Valley. The model simulates confined and unconfined conditions, withdrawal from wells, areal recharge,

Table 10. Temperature and specific conductance of water from wells and springs in Cache Valley, 1967-69 and 1990-91

[See figure 2 for an explanation of the numbering system used for hydrologic-data sites; °C, degrees Celsius; µS/cm, microsiemens per centimeter at 25 degrees Celsius; —, no data; L, laboratory determination]

Location	Well owner or user, or spring name	Bjorklund and McGreevy (1971, table 1) Date	Temperature (°C)	Specific conductance (µS/cm)	1990-91 Date	Temperature (°C)	Specific conductance (µS/cm)
			Wells				
(A-11-1)6ccb-1	Hill, P.A.	04-00-68	12	440	03-19-91	11.0	450
(A-12-1)3cbd-1	Lamb, D.D.	08-00-67	12	610	03-14-91	10.5	660
(A-12-1)5dbb-1	Chambers, George	02-00-68	13	440	03-28-91	13.0	465
(A-12-1)6bcc-1	Reese, C.L.	11-00-68	14	890	03-19-91	15.5	1,000
(A-12-1)7add-1	Seamon, C.	04-00-68	13	440	04-02-91	13.0	460
(A-12-1)7bcb-1	Maughan, J.C.	04-00-68	—	1,500	03-19-91	15.5	1,460
(A-12-1)8acb-1	Lee, R.E.	04-00-68	12	520	04-02-91	12.5	530
(A-12-1)10cbb-1	Hancey, H.E.	08-00-67	14	520	04-03-91	13.5	540
(A-12-1)16cac-1	Benson Irrigation Co.	02-00-68	16	540	04-02-91	14.5	700
(A-12-1)17daa-1	Benson Irrigation Co.	08-00-67	16	560	03-09-91	20.0	490
(A-12-1)20daa-2	Quale, J.W.	08-00-67	19	590	03-21-91	16.5	510
(A-12-1)32dba-1	Hunsaker, T.E.	09-00-67	12	400	03-13-91	12.0	470
(A-12-1)33bca-1	Benson, K.E.	08-00-67	13	400	03-13-91	13.0	410
(A-12-1)33bcc-3	Eliason Packing Co.	08-00-67	12	440	03-13-91	11.5	410
(A-13-1)3bab-1	Bair, F.A.	06-00-68	10	680	04-02-91	9.0	680
(A-13-1)3cab-1	Mendonhall, O.J.	06-00-68	11	680	04-03-91	6.5	660
(A-13-1)31ccc-1	Reese, A.C.	10-00-68	14	850	03-19-91	12.0	940
(A-13-1)32daa-1	Weeks, David	10-00-68	14	540	03-28-91	12.5	600
(A-14-1)22bad-1	Stoddard, C.B.	08-00-67	10	430	07-24-90	10.0	485
(A-14-1)34bdc-1	Thompson, W.G.	08-00-67	11	610	04-02-91	10.5	660
(A-15-2)31cba-1	State of Utah	11-00-67	13	330	04-01-91	13.5	300
(B-10-1)12aaa-1	Bailey, O.S.	08-00-67	12	620	03-12-91	11.0	640
(B-11-1)4cac-1	Shelton, Verland	02-00-68	11	730	03-18-91	10.5	790
(B-11-1)9aad-1	Sorenson, Owen	09-00-68	12	730	03-18-91	10.0	800
(B-11-1)9cdb-1	Longstroth, William	09-00-67	11	870	03-06-91	9.5	970
(B-11-1)11caa-1	Hodges, W.N.	10-00-68	12	560	03-19-91	12.0	620
(B-11-1)13acd-1	Jensen, C.A.	11-00-68	12	490	03-19-91	11.5	540
(B-11-1)15dba-1	Anderson, Willis	08-00-67	11	570	03-19-91	11.0	560
(B-11-1)16aca-1	Sorensen, Claud	02-00-68	10	570	03-18-91	10.0	610
(B-11-1)23dcb-1	Anderson, Eldon	08-00-67	11	620	03-11-91	11.0	660
(B-11-1)26abd-1	Bankhead, Heber	10-00-68	11	860	03-11-91	8.5	1,000
(B-12-1)1ccc-2	Ballard, M.J.	11-00-68	17	1,500	03-20-91	17.0	1,600
(B-12-1)3ccc-1	Watterson, J.L.	08-00-67	17	830	03-21-91	16.0	840
(B-12-1)8cdb-2	Rasmussen, Reuben	09-00-67	13	750	03-09-91	12.0	730
(B-12-1)9ddd-1	Benson Recreation Area	01-00-69	18	920	03-20-91	18.5	900
(B-12-1)10dcd-2	Nuttall, J.L.	08-00-67	21	780	03-21-91	23.5	820
(B-12-1)11dda-1	Snow, F.B.	02-00-69	17	1,500	03-21-91	17.0	1,500
(B-12-1)14aaa-1	Riggs, Bert	08-00-67	16	510	03-21-91	14.5	475
(B-12-1)15adc-1	Ricks, Charles Ethan	08-00-67	18	820	03-07-91	17.0	860
(B-12-1)23dbd-1	Benson, F.A.	08-00-67	12	470	03-21-91	12.0	430
(B-12-1)24daa-1	Waterson, J.L.	08-00-67	13	420	03-26-91	12.0	405
(B-12-1)24dab-1	Hebaus, Joseph	08-00-67	11	480	03-22-91	11.0	450
(B-13-1)27cdd-1	Seamons, N.B.	08-00-67	23	1,200	03-14-91	17.5	1,310

Table 10. Temperature and specific conductance of water from wells and springs in Cache Valley, 1967-69 and 1990-91—Continued

Location	Well owner or user, or spring name	Bjorklund and McGreevy (1971, table 1) Date	Temperature (°C)	Specific conductance (µS/cm)	1990-91 Date	Temperature (°C)	Specific conductance (µS/cm)
		Wells—Continued					
(B-13-1)30acc-1	Ballard, E.R.	08-00-67	12	620	03-07-91	10.5	600
(B-13-1)36cca-1	Thain, Paul	10-00-68	17	1,500	03-20-91	18.0	1,750
13S 38E 26AAD1	Allen, A.P.	07-00-68	13	780	03-27-91	12.0	1,060
		Springs					
(A-10-1)33bac-S1	Whites Inc.	09-00-68	12	520	12-06-90	9.0	500
(A-11-1)10ccd-S1	Little Ballard Spring	08-00-67	13	600	05-31-90	12.0	650
(A-11-1)15bbc-S1	Big Ballard Spring	05-00-68	—	600	05-31-90	12.0	600
(A-11-1)17bdb-S1	Spring Creek Number 1	11-00-67	11	510	11-28-90	8.0	670
(A-12-1)4bab-S1	Chambers Spring	05-00-68	—	900	05-31-90	13.0	680
(A-13-1)32adc-S1	Hopkins Spring	05-00-68	12	600	05-31-90	12.0	640
(A-14-1)15acb-S1	Clear Creek Spring	11-00-68	12	570	11-20-90	12.0	600
(A-14-1)34bdc-S1	Peart-Thompson	12-00-68	—	710	11-20-90	10.0	650
(B-10-1)10aac-S1	Wellsville Spring	10-00-68	12	650	12-07-90	11.5	820
(B-10-1)10cab-S1	Murray Spring	11-00-67	11	530	12-07-90	10.0	495
(B-11-1)21dac-S1	Gardner Spring	05-00-68	11	529	06-05-90	11.5	520 L
(B-11-1)27dca-S1	Darley Spring	04-00-69	11	500	12-07-90	9.5	485
(B-11-1)34dac-S1	Northfield Spring	11-00-67	10	530	11-28-90	10.5	455
16S 40E 22DBB1	Low Spring	05-00-69	9	460	06-08-90	10.5	405

evapotranspiration, seepage to drains, and seepage to and from streams and consolidated rock. The model was calibrated to assumed steady-state conditions in 1969 and annual transient-state conditions for 1982-90.

Discretization of the Ground-Water System

The area covered by saturated unconsolidated basin-fill deposits was discretized into a non-uniform, horizontal, rectangular grid composed of blocks that are called cells (fig. 17). This area is somewhat smaller than the area of unconsolidated basin-fill deposits shown in figure 3 because some of the unconsolidated basin-fill deposits near the mountains or overlying shallow consolidated rock are not saturated and therefore are not included in the modeled area. In the southern part of the basin, where there are more wells and a larger volume of ground-water withdrawal than in the northern part, smaller cells provide finer resolution. The grid consists of 82 rows and 39 columns of cells, and cell size ranges from 1 mile on each side to 0.5 mile by 0.375 mile. Aquifer properties assigned to each cell in the model are representative of the average value for the cell area. If a cell is active, the model bases its computations for that cell at a point in the center of the cell, called a node. If a cell is inactive, then no computations are made for that cell by the model.

The saturated unconsolidated basin-fill deposits in Cache Valley were represented by six layers. Geohydrologic sections across Cache Valley by McGreevy and Bjorklund (1970, cross section P-P') indicate that no discrete aquifers or confining units are present, but numerous discontinuous clay layers occur at different depths and in different locations throughout the basin; thus, vertical flow of water is impeded in much of the basin. The model simulates the vertical impedance of ground-water flow through the clay layers by using a vertical-leakance term

described by McDonald and Harbaugh (1988 p. 5-12). The use of six model layers in the unconsolidated basin-fill deposits made simulation of vertical gradients in the basin possible. Depth of the layers was chosen mainly to correspond to the major areas of pumping in the valley.

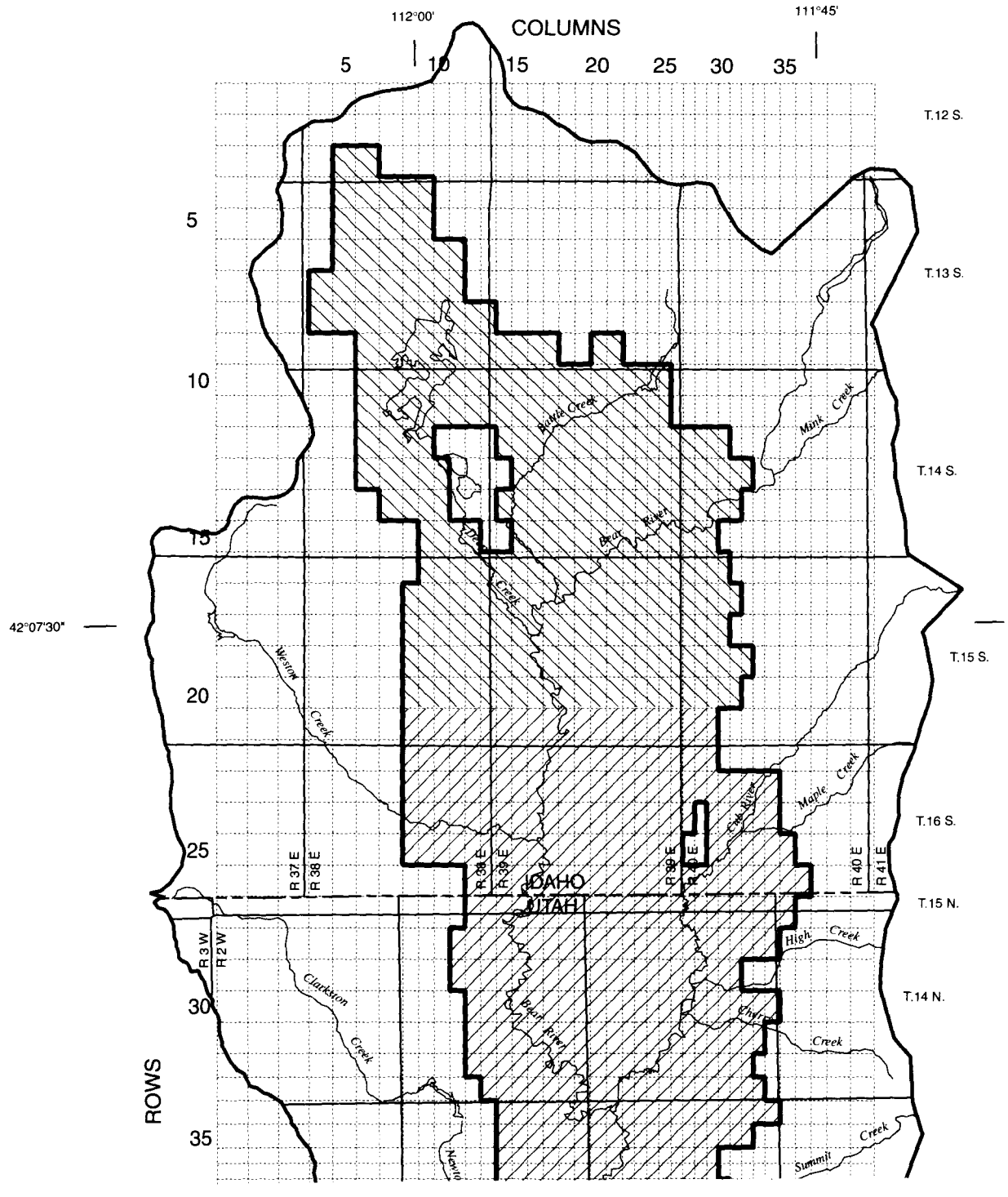
Layer 1 was simulated as an unconfined layer with an initial saturated thickness of 100 feet. Changes in water levels in layer 1 can cause the saturated thickness to vary from the initial 100 feet. Even though confined conditions exist in parts of Cache Valley at depths of less than 100 feet, a 100-foot saturated thickness was selected because some streams cut deeply into the basin-fill deposits and would cause cells in layer 1 to dry up if layer 1 was not thick enough. Layer 2 simulates the saturated deposits at depths from 100 to 200 feet using a confined- or unconfined-layer option described by McDonald and Harbaugh (1988, p. 5-26). This option allows the storage term to be converted from confined to unconfined in a cell when the water level declines below the top of the cell. Layers 3 through 6 were simulated with the confined-layer option described by McDonald and Harbaugh (1988, p. 5-26). The depth of saturated basin-fill deposits simulated by layer 3 is from 200 to 300 feet, by layer 4 from 300 to 500 feet, by layer 5 from 500 to 1,000 feet, and by layer 6 from 1,000 to 1,500 feet. Layer 6 was added to allow simulation of pumping from municipal wells in the southern part of the valley that are completed in saturated basin-fill deposits below depths of 1,000 feet. Bjorklund and McGreevy (1971, table 4) report an approximate maximum thickness of basin-fill deposits of 1,340 feet.

There are 1,549 active cells in layers 1 and 2; 1,423 active cells in layers 3 through 5; and 1,129 active cells in layer 6. Because the angle of the faults that bound the basin is very high, the same boundary between active and inactive cells was used in all layers, except in areas of shallow consolidated rock. Where evidence of shallow consolidated rock was found, the lower layers were made inactive. These areas of shallow consolidated rock are in the northern part of the study area and in the Paradise-Avon area. Because geologic sections by McGreevy and Bjorklund (1970, cross section A-A') show depth of unconsolidated basin-fill deposits in the northern part of the study area to be less than 1,000 feet, layer 6 was not simulated in this area. Cross section P-P' of McGreevy and Bjorklund (1970) also shows that the basin-fill deposits thin toward the south, near Hyrum. To simulate this thinning, layers 3 through 6 were not used in this area. The Clarkston and Weston Canyon areas were not simulated by this model because both of these areas have their own individual basin-fill ground-water system, and each is at a higher altitude than the main ground-water system in Cache Valley. The model grid and location of active cells used for each layer are shown in figure 17.

Boundary Conditions and Data Requirements

The physical and hydrologic limits of a simulated ground-water-flow system are defined as the boundaries of that system. Several different mathematical representations of these boundaries can be simulated by the model. No-flow, free-surface, head-dependent flux, and specified-flux boundaries as described by Franke and others (1984) are used in the simulations of the ground-water system in Cache Valley. Recharge to and discharge from the model are simulated as specified-flux and head-dependent flux boundaries. Locations of cells with head-dependent flux boundaries are shown in figure 18.

No-flow boundaries exist at the boundary between active and inactive cells. Because no computations are done for inactive cells, no ground-water flow is simulated between active and inactive cells. No-flow boundaries were used to simulate the contact between the surrounding, low to moderately permeable consolidated rocks and poorly consolidated Salt Lake Formation and the permeable, unconsolidated basin-fill deposits, except where inflow from adjacent consolidated rock or an unconsolidated basin-fill deposit ground-water system was simulated. During the calibration process discussed later in this report, some of the no-flow boundary cells were replaced with general-head boundary cells.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T.14 S.
T.15 S.
T.16 S.
T.15 N.
T.14 N.
R 37 E
R 38 E
R 39 E
R 40 E
R 41 E
R 3 W
R 2 W
IDAHO
UTAH
Mink Creek
Bear River
Weston Creek
Maple Creek
Cub River
High Creek
Clarkston Creek
Summit Creek

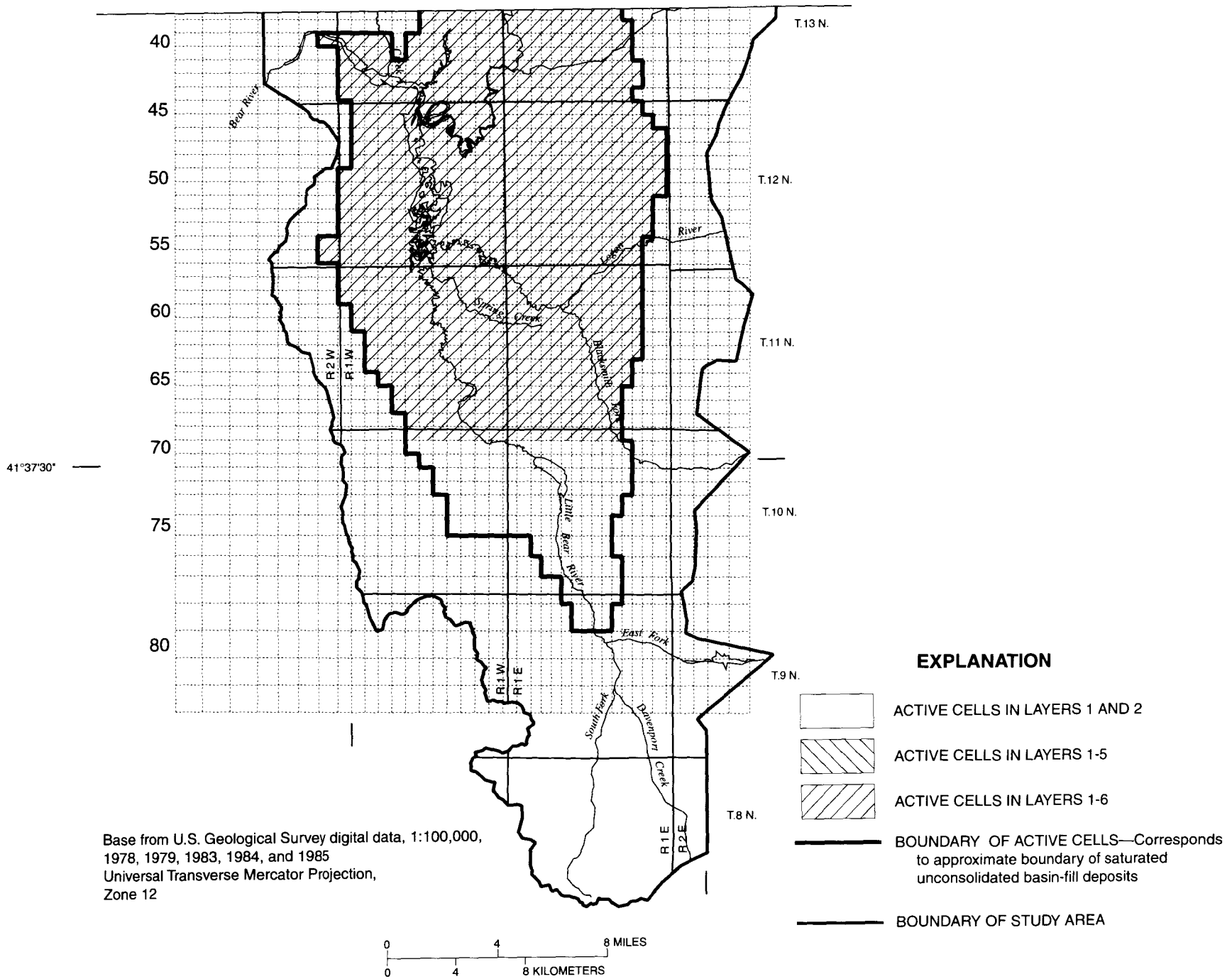
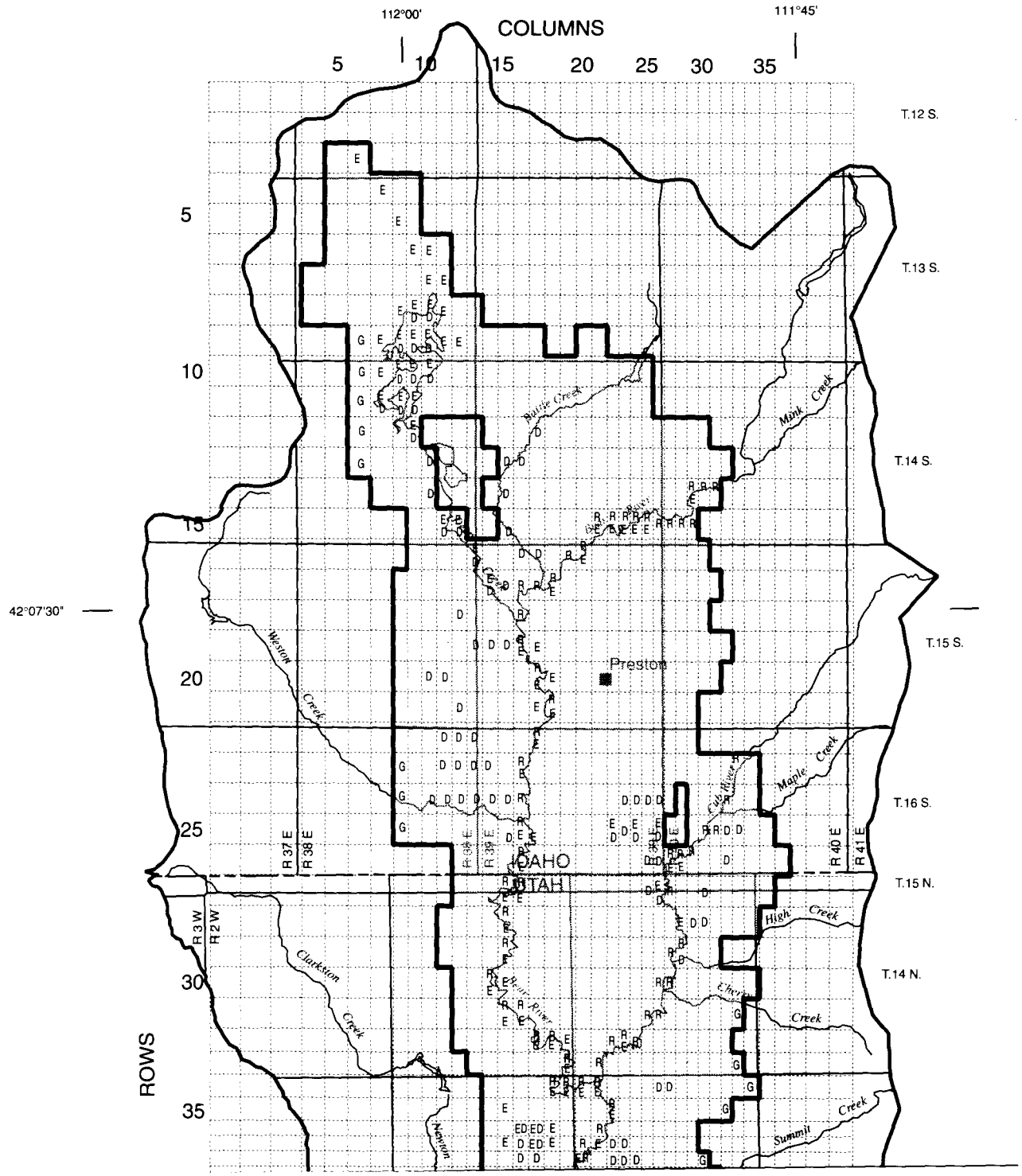


Figure 17. Model grid and location of active cells in the ground-water-flow model of the unconsolidated basin-fill deposits in Cache Valley.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42°07'30"
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T.14 S.
T.15 S.
T.16 S.
T.15 N.
T.14 N.
R 37 E
R 38 E
R 39 E
R 40 E
R 41 E
R 3 W
R 2 W
Preston
IDAHO
UTAH
Battle Creek
Mink Creek
Weston Creek
Maple Creek
Clarkston Creek
High Creek
Cherry Creek
Summit Creek
Newton

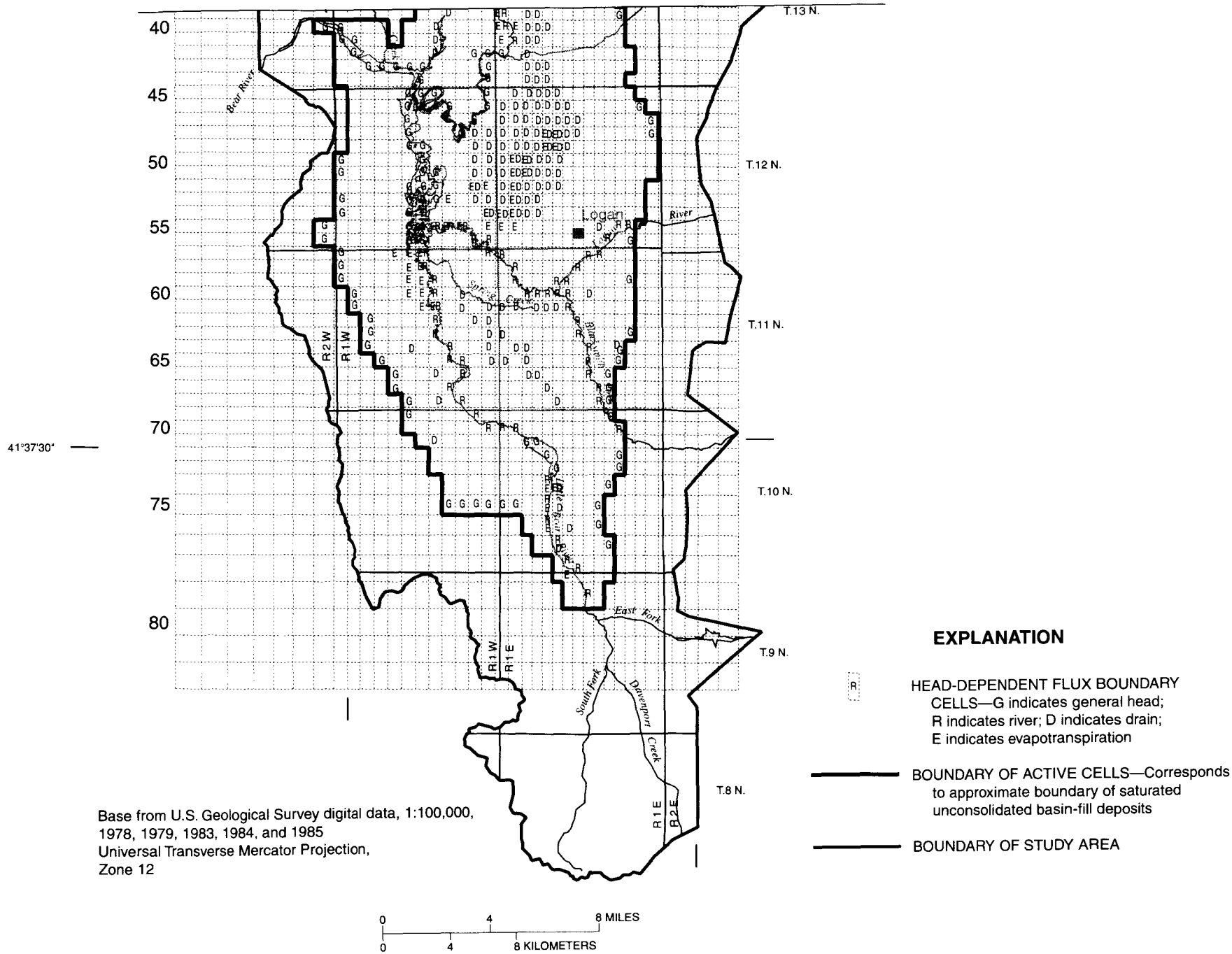


Figure 18. Location of cells with head-dependent flux boundaries.

Recharge

Estimated average recharge from all sources to the ground-water system in the unconsolidated basin-fill deposits in Cache Valley in 1969 is 450 cubic feet per second (326,000 acre-feet per year, table 5). The largest source of recharge is infiltration of precipitation and unconsumed irrigation water; secondary sources are seepage from canals and streams and subsurface inflow from adjacent consolidated rock and unconsolidated basin-fill ground-water systems.

The upper boundary of the model is considered to be a free-surface boundary that is represented by the altitude of the water table. The altitude can change in response to changes in rate of recharge or discharge. The model uses the Recharge Package (McDonald and Harbaugh, 1988, p. 7-1) to simulate recharge through the free-surface boundary to the top-most active layer. Recharge from infiltration of precipitation and unconsumed irrigation water, and seepage from canals and from some streams was simulated using this method.

The service areas used in simulating recharge from infiltration of unconsumed irrigation water were adapted from maps made by the [U.S.] Soil Conservation Service (1976). The location of the service area for each irrigation company is shown in figure 19. The amount of irrigation water that recharges the ground-water system was assumed to be distributed evenly over each service area. The amount of recharge from infiltration of precipitation and unconsumed irrigation water was at first simulated with initial estimates as discussed in the "Recharge" section of the "Unconsolidated basin-fill deposits" section of this report. During the 1982-90 transient-state simulation of the ground-water system, the changes in computed water levels were not as large as those in the measured water levels. Further investigation into the lack of change in water levels indicated that this was because the initial recharge estimates did not vary much from one year to the next during the transient-state simulation. The lack of variance in the initial estimates of recharge was the result of oversimplification in the computation of initial recharge. Ten percent of annual precipitation was initially assumed to recharge the ground-water system each year. This percentage did not account for the timing of the precipitation or the growth cycles of the plants in Cache Valley. An independent means of estimating recharge (Bauer and Vaccaro, 1987) was obtained by using the Deep Percolation Model (DPM). The initial computations of recharge to irrigated areas used a complex spatial distribution of irrigation water. To maintain this distribution, the initial estimates of recharge were corrected by a factor equal to the DPM estimates divided by the initial estimates. The correction factors that were used each year to correct the amount of recharge from infiltration of precipitation and unconsumed irrigation water are shown in table 6. A complete discussion of the DPM simulations appears in the appendix.

Several of the irrigation companies form groups that share diversion structures. These groups are (1) Twin Lakes Canal Company and Winder Group; (2) West Cache Canal Company and Utida Pumping Company; (3) Logan North Field Irrigation Company, Hyde Park Irrigation Company, and Logan Island Company; (4) Logan North West Field Irrigation Company and Benson Irrigation Company; and (5) Hyrum Wellsville Mendon Irrigation Company and Hyrum Wellsville Irrigation Company. The amount of water diverted for each group listed is reported in table 4 by the name of the canal into which water from the stream was first diverted. Recharge for the areas these companies serve was computed by group. The name of the first company in each group is the name that was used to label that group for the simulations. Recharge from irrigated land for which diversion records were unavailable was simulated with the same amount of recharge as for an adjacent area served by a canal company.

Dry-farmed and non-farmed land did not have the complex application system of the irrigated land, so the amount of recharge estimated by the DPM (table 6) was used directly in the simulation. Recharge from dry-farmed land in Cache Valley was estimated by the DPM using the alternating fallow and cropped cycle of winter wheat as discussed in the "Recharge" section of the "Unconsolidated basin-fill deposits" section of this report. Areas where recharge from dry-farmed land was simulated are shown in figure 20.

Seepage from the three streams with the largest losses (High, Maple, and Mink Creeks) was simulated by the model. Because these streams are almost totally diverted near the mountain front during the irrigation season, an estimated 10 percent of total annual streamflow was assumed to recharge the ground-water system in the first cell the stream intercepted as it flowed across the basin-fill deposits. Smaller streams were either totally diverted for

irrigation before entering the valley, or seepage to the ground-water system was assumed to be small compared with total recharge; therefore, recharge from these smaller streams was not simulated. Cells that represent recharge from simulated stream seepage are shown in figure 21.

The [U.S.] Soil Conservation Service (1976) provided locations and conveyance-efficiency estimates for the major canals used in the simulation. The amount of seepage from each canal was assumed to be distributed evenly along the length of the canal. To compute total recharge for each canal, total canal seepage was first computed using the equation

$$S = TD \times \frac{(100 - CE)}{100} \tag{3}$$

where: S is canal seepage,

TD is total amount of water diverted into a canal, and

CE is canal efficiency, expressed as a percent.

The amount of seepage was then divided by the total length of the canal to obtain the amount of seepage per foot of canal. The length of each section of canal in a cell was multiplied by the amount of seepage per foot to obtain the amount of seepage in the cell. Because some canals are used to irrigate areas in small mountain valleys that border the unconsolidated basin-fill deposits, not all of the canals in Cache Valley are close to or inside the active model boundary; therefore, the recharge from canal seepage simulated by the model is less than the seepage from canals estimated for the whole valley (table 5). Cells that represent recharge from simulated canal seepage are shown in figure 21.

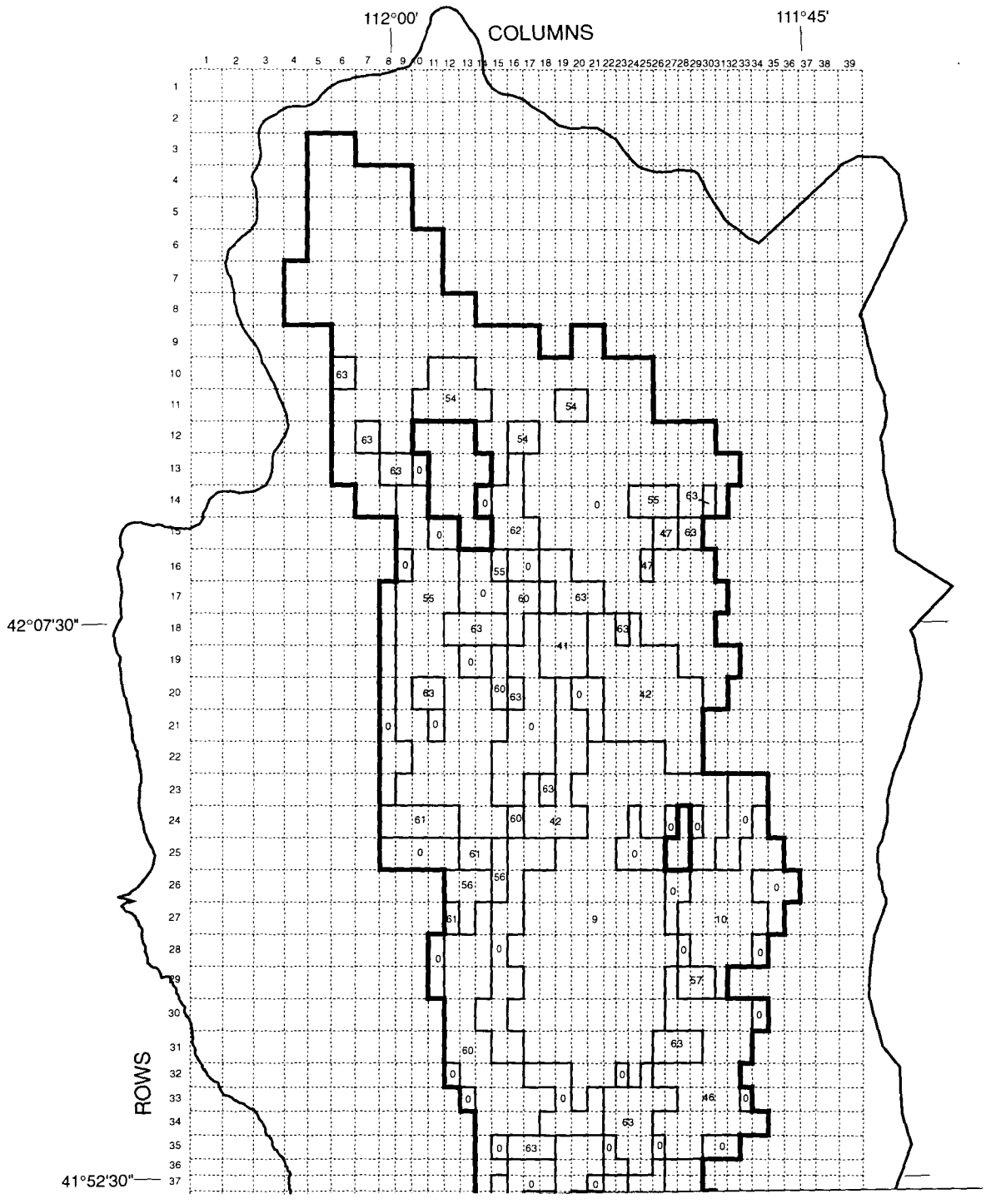
After the initial calibration, head-dependent flux boundaries in the form of general-head boundary cells in layer 1 were used to simulate recharge to unconsolidated basin-fill deposits from consolidated rock, and, indirectly, from seepage from ephemeral streams. Water levels in the consolidated rock that surrounds the unconsolidated basin-fill deposits is unknown. The water levels controlling boundary heads in the general-head boundary cells that represent subsurface inflow from consolidated rock were estimated to be between 4,900 and 5,100 feet on the basis of altitudes of springs that discharge from consolidated rock in the Wellsville and Richmond area. These estimated water levels were much higher than the measured water levels in the unconsolidated basin-fill deposits. This large difference in water levels made it imperative that the general-head boundary conductance be carefully controlled. To maintain the large difference in water levels and to control the amount of water flowing from consolidated rock to unconsolidated basin-fill deposits, conductance was initially set at 1×10^{-4} foot squared per day.

Ground water from the Clarkston area flows directly into Cutler Reservoir but probably adds only a small amount of water to the unconsolidated basin-fill deposits in the main ground-water system in Cache Valley. For this reason, inflow from the Clarkston area was not simulated.

General-head boundary cells in layer 1 were used to simulate subsurface inflow that enters the main ground-water system in the unconsolidated basin-fill deposits in Cache Valley from basin-fill deposits at the mouth of Weston Canyon. The ground-water system in Weston Canyon is at a higher altitude than and discharges into the main ground-water system in Cache Valley. The water-level altitude where the Weston Canyon ground-water system connects with the main ground-water system in Cache Valley is unknown; therefore, an altitude 20 feet below the altitude of the streambed (obtained from U.S. Geological Survey 1:24,000 topographic quadrangles), at the point where the stream enters the valley, was used for cells with general-head boundaries.

Discharge

Average discharge from the main ground-water system in the unconsolidated basin-fill deposits in Cache Valley from seepage to streams and Cutler Reservoir, spring discharge, evapotranspiration, and withdrawal from wells was estimated to be 449 cubic feet per second (325,000 acre-feet per year, table 5) in 1969. Several types of cells with head-dependent flux boundaries were used to simulate discharge, including river cells, drain cells, general-head boundary cells, and evapotranspiration cells (locations are shown in fig. 18).



EXPLANATION

AREA SERVICED BY IRRIGATION COMPANY

0 Not serviced by irrigation company
1 Anderson Mack and Story Spring
2 Benson Bear Lake Irrigation Co.
3 Benson Irrigation Co.
4 Blacksmith Fork Millville Co.
5 Blacksmith Providence Co.
6 Chambers Spring
7 Clayton Creek Irrigation Co.
8 College Irrigation Co.
9 Cub River Canal Co.
10 Coveville and Mountain Home Canal
11 East Field Irrigation and Canal Co.
12 Freeborn Merril Irrigation Co.
13 Garr Spring Irrigation Co.
14 Hammer Canal Co.
15 Hopkins Slough Irrigation Co.
16 Hyde Park Irrigation Co.
17 Hyrum Blacksmith Co.

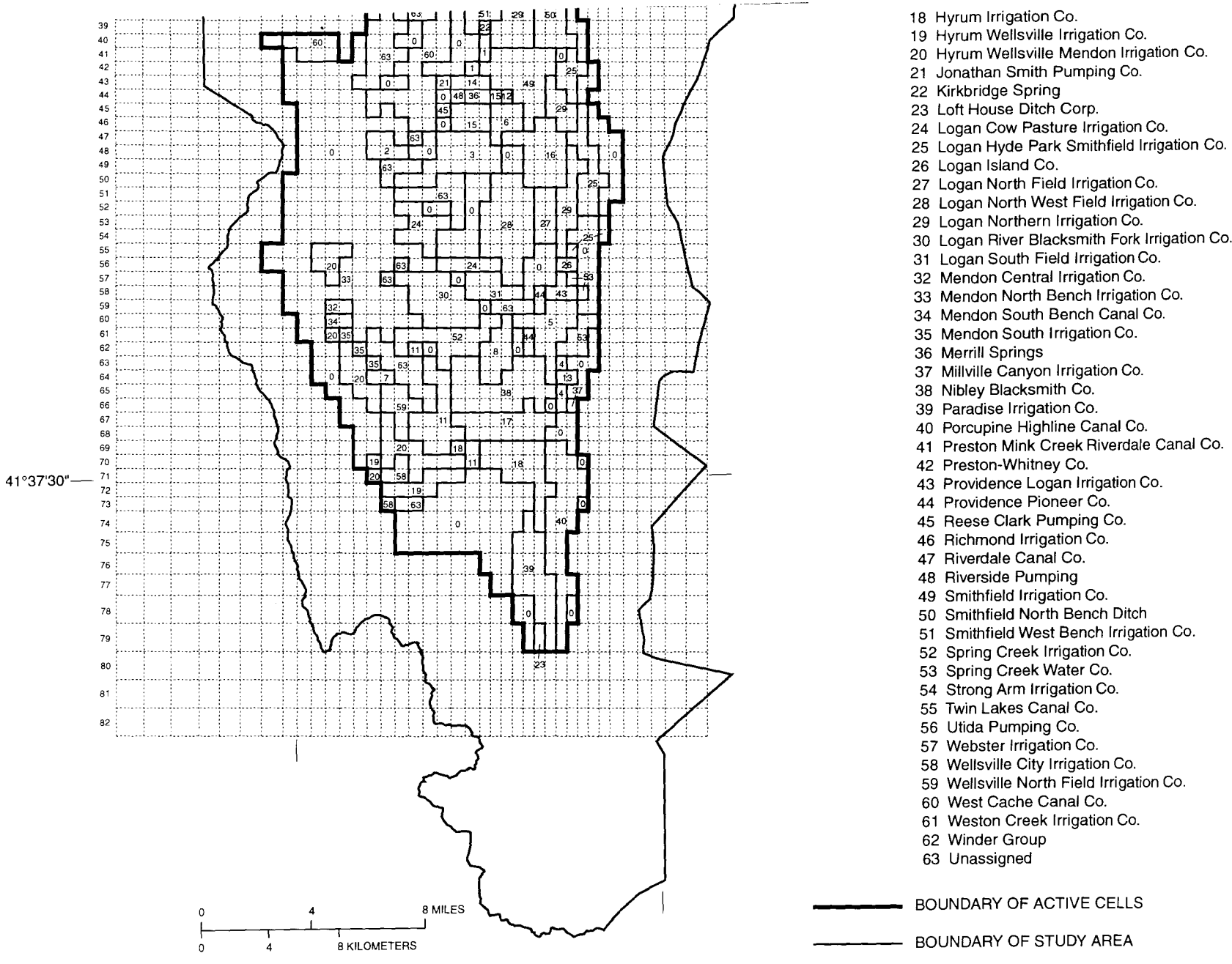
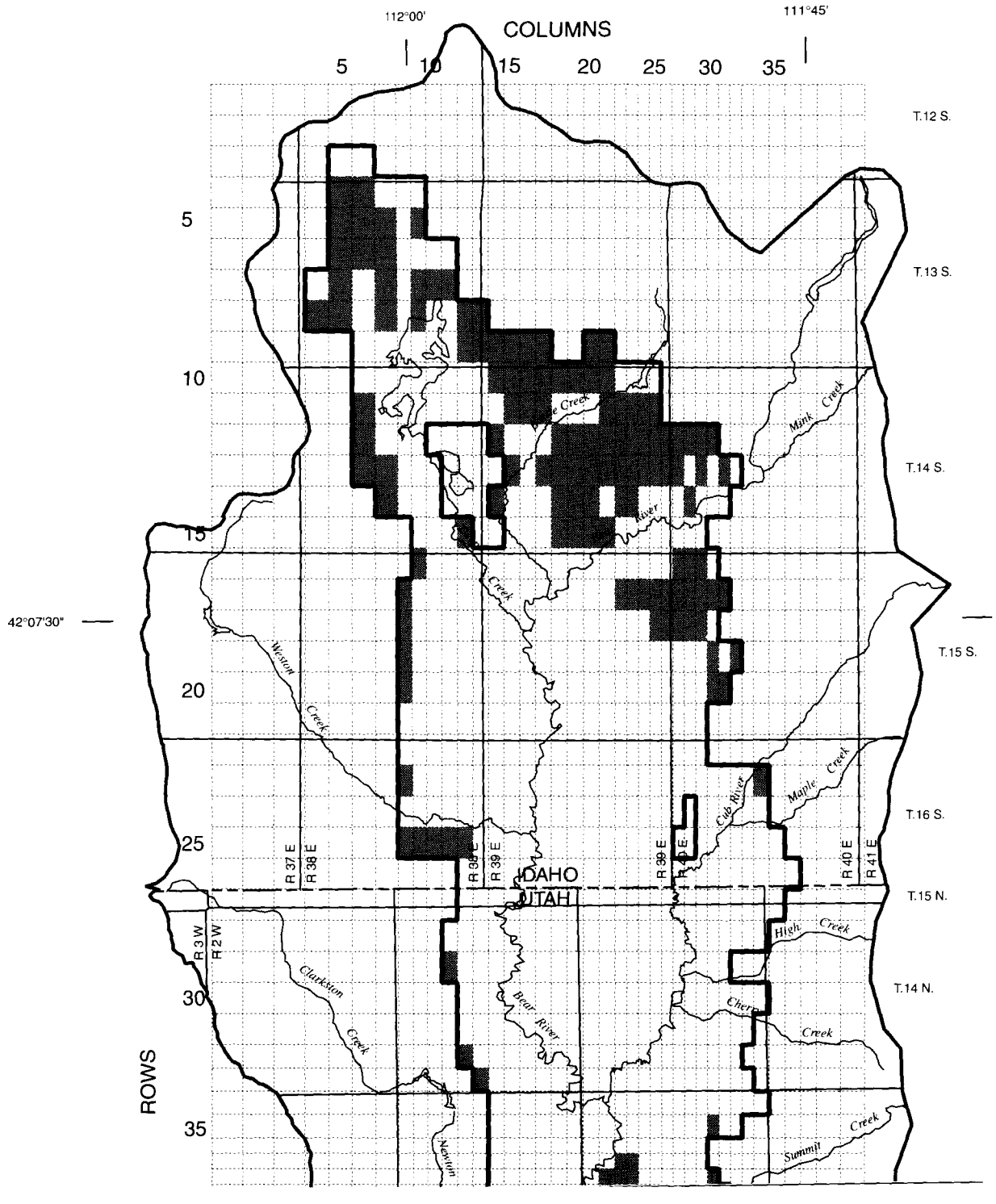


Figure 19. Location of service areas used in simulating recharge from infiltration of unconsumed irrigation water.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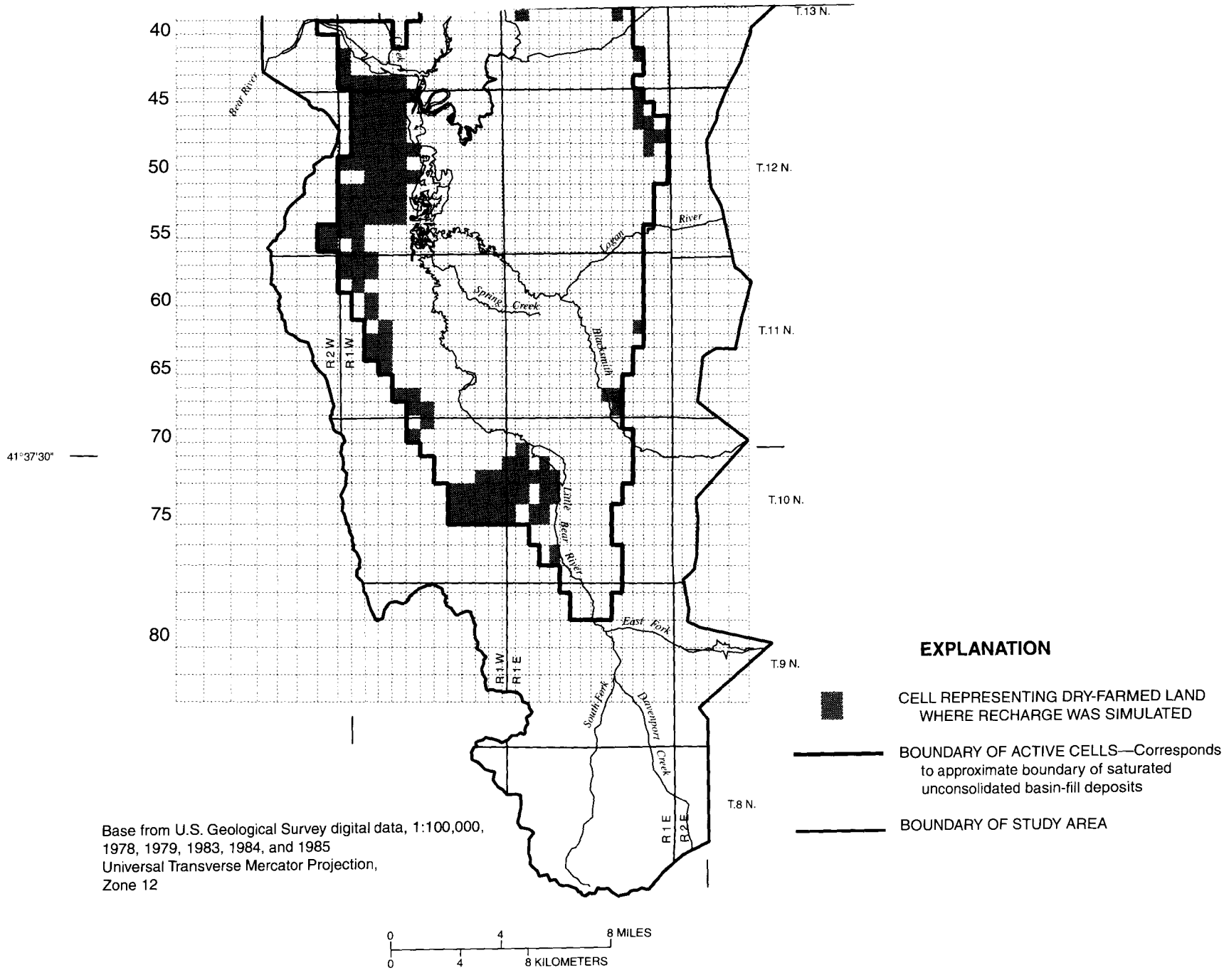
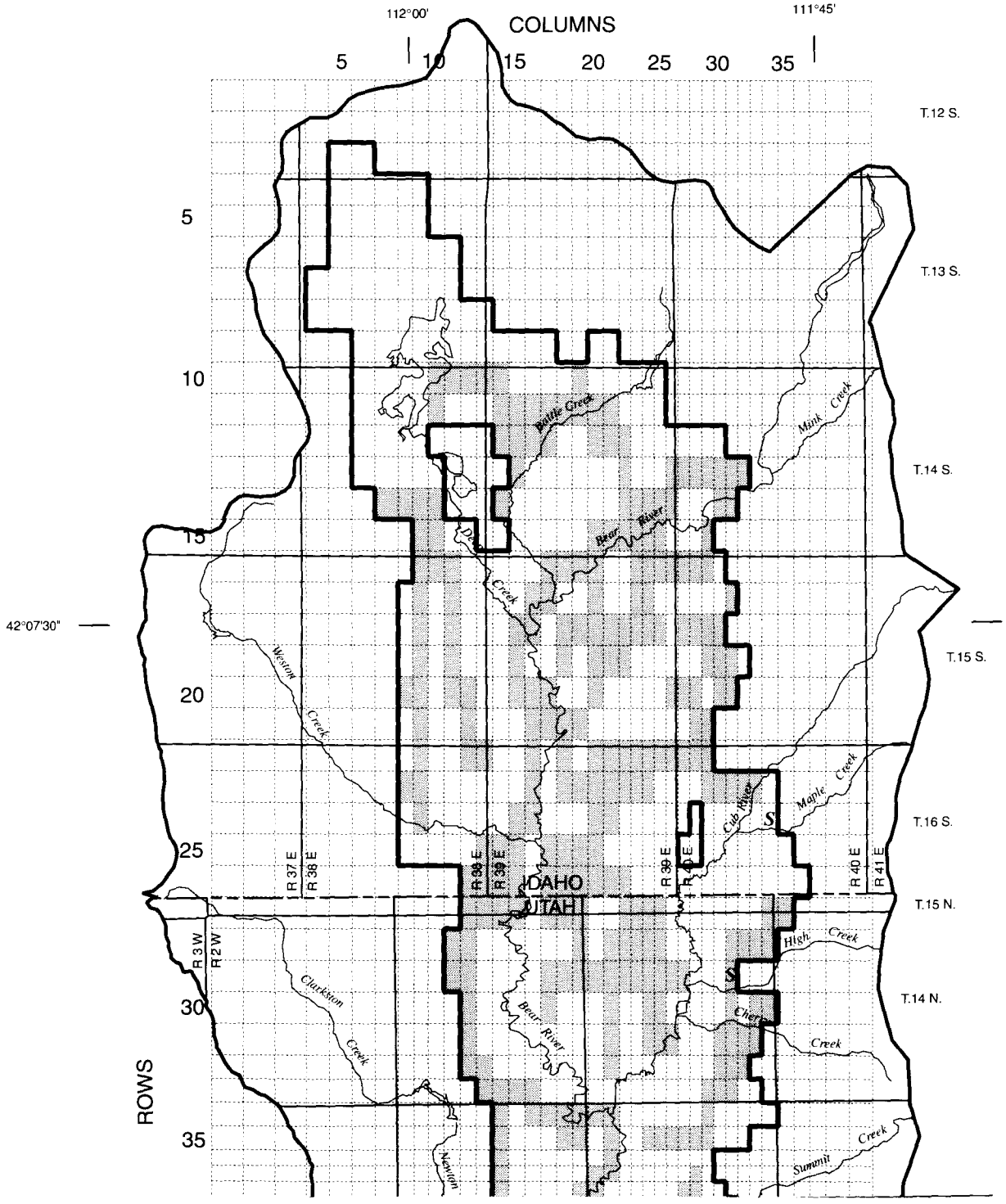


Figure 20. Location of cells that represent dry-farmed land where recharge was simulated.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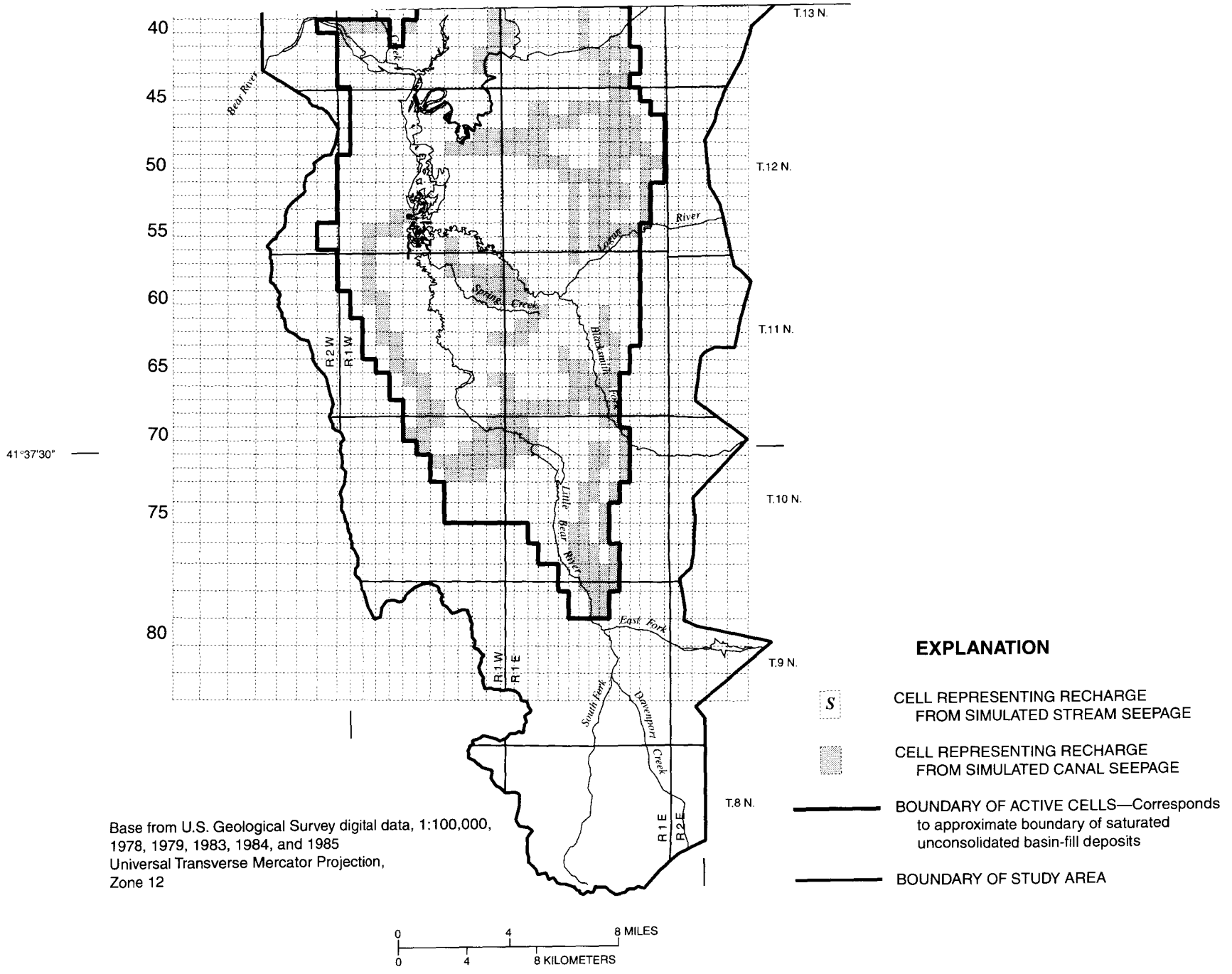


Figure 21. Location of cells that represent recharge from simulated stream and canal seepage.

The largest component of discharge is ground-water seepage to rivers. The River Package (McDonald and Harbaugh, 1988, p. 6-1) was used to simulate seepage to the Bear River, Cub River, Logan River, Blacksmith Fork, and Little Bear River. A river stage of 10 feet above the streambed altitude was simulated for the Bear River, and 3 feet for the other rivers. Average stream-surface altitude for each cell was obtained from U.S. Geological Survey 1:24,000 topographic quadrangles for input into the River Package. The estimate of the conductance of the river cells was initially made using the equation of McDonald and Harbaugh (1988, p. 6-10). Because the streambeds are composed mainly of silty sand, the hydraulic conductivity of the streambed was assumed to be 1.74×10^{-5} foot per second, which is in the range of the hydraulic conductivity of silty sand presented by Freeze and Cherry (1979, p. 29). The length of the river in each cell was computed using U.S. Geological Survey Digital Line Graphs (U.S. Geological Survey, 1989) and the model grid. Stream-channel width of all the rivers was assumed to be a constant 50 feet, and streambed thickness was assumed to be 10 feet.

Discharge from springs and natural drains was simulated using the Drain Package (McDonald and Harbaugh, 1988, p. 9-1). For drain cells that represent simulated discharge from springs and natural drains, the average altitude of the land surface (obtained from U.S. Geological Survey 1:24,000 topographic quadrangles) was entered for the drain altitude of each cell. Conductance was initially set at 1×10^{-4} foot squared per day to allow only a minimum amount of discharge to be computed by the model.

Seepage to Cutler and Hyrum Reservoirs from the basin-fill deposits was simulated using the General-Head Boundary Package (McDonald and Harbaugh, 1988, p. 11-1). This type of boundary was used because it allowed the water levels in wells completed below the reservoir in layer 1 to be different from the reservoir level. The altitude of the water surface in Cutler Reservoir was set at 4,407 feet and in Hyrum Reservoir at 4,664 feet. Both altitudes were obtained from the average reservoir altitude shown on U.S. Geological Survey 1:24,000 topographic quadrangles. Initial values of conductance, representing the bed of the reservoir, were set at 1×10^{-4} foot squared per day.

Areas of evapotranspiration mapped by Bjorklund and McGreevy (1971, pl. 4) that were not included in the areas of areal recharge from unconsumed irrigation water were simulated using the Evapotranspiration Package (McDonald and Harbaugh, 1988, p. 10-1) as another head-dependent flux boundary. Average altitude of land surface in each cell was obtained from U.S. Geological Survey 1:24,000 topographic quadrangles. The maximum evapotranspiration rate used in the simulations was 2.3 feet per year, as discussed in the "Discharge" section of the "Unconsolidated basin-fill deposits" section of this report. The minimum evapotranspiration rate used was 1.0 foot per year. The minimum rate was used for cells in which 1.3 feet per year of the water for evapotranspiration is assumed to be from surface-water sources and an estimated 1.0 foot per year is from the ground-water system. Such areas include the part of the Bear River flood plain where it cuts through the Bear River delta and surface-water flooded areas such as Oxford Slough (fig. 5). The extinction depth was set at 6 feet, which is within the range for salt grass and other grasses reported by Robinson (1958). These values were not adjusted during the calibration process.

Withdrawal from wells was simulated using a specified-flux boundary from the Well Package (McDonald and Harbaugh, 1988, p. 8-1). The locations of cells that represent discharge simulated from withdrawal from wells are shown in figure 22. Analysis of withdrawals from flowing wells in Cache Valley determined that 96 percent of the water is from flowing wells that yield more than 30 gallons per minute. To simplify data input, only flowing wells with withdrawals greater than 30 gallons per minute were simulated. The model layer from which water was withdrawn from each well in the simulation was based on the open interval reported in drillers' logs. For wells that yield water from more than one layer, the amount of water withdrawn from each layer was based on the length of open interval in each layer.

Hydraulic Properties

Hydraulic conductivity of the unconsolidated basin-fill deposits is known in only a few locations in the valley, and variation of hydraulic conductivity with depth is not known. Initially, 11 hydraulic-conductivity values were

assigned to the first layer in the 11 different hydrologic zones specified by Bjorklund and McGreevy (1971, p. 48). These hydraulic-conductivity values were estimated on the basis of the mean value of specific capacity from wells throughout each zone. Initial transmissivity values of layers 2 through 6 were computed by multiplying the estimated hydraulic-conductivity values of layer 1 by the thickness of each layer. During calibration, transmissivity values in layers 5 and 6 were reduced to be more consistent with aquifer-test data. Final transmissivity values for layer 4 were used in layers 5 and 6.

Vertical hydraulic conductivity of the basin-fill deposits in Cache Valley is not well understood, but a large number of flowing wells are in the south-central part of the valley and confined conditions exist throughout the central part of the valley. As explained in the "Unconsolidated basin-fill deposits" section of this report, no continuous confining layers occur in the basin-fill deposits in Cache Valley, but numerous thin, discontinuous clay layers impede vertical flow throughout the valley. The impedance of vertical flow also is indicated by flowing wells with water-level heads as much as 62 feet above land surface (Bjorklund and McGreevy, 1971, p. 22). A vertical leakance value between model layers was used to simulate impedance of these clay layers to vertical flow; none of the clay layers were simulated as a distinct layer. Initial values of vertical leakance were computed using the equation described by McDonald and Harbaugh (1988, p. 5-12) and ranged from 1×10^{-3} per day between layers 1 and 2 to 2×10^{-5} per day between layers 5 and 6. The values of vertical hydraulic conductivity used to compute these values for vertical leakance ranged from 0.1 to 0.01 foot per day, which is in the range of hydraulic conductivity for unconsolidated silts reported by Freeze and Cherry (1979, p. 29). This assumed range of vertical hydraulic-conductivity values was used because of the scarcity of actual vertical hydraulic-conductivity data for Cache Valley.

Storage properties of the unconsolidated basin-fill deposits in Cache Valley are not well known. The initial value of specific yield in layer 1 was set to an average value of 0.20 (Lohman, 1979, p. 8). The same specific yield is used in a cell in layer 2 if during a simulation the water level in layer 2 drops below the top of layer 2. The initial artesian storage coefficient for layers 2 through 6 was estimated by multiplying the thickness of each layer by 1.0×10^{-6} (Lohman, 1979, p. 8). The initial values were 1.0×10^{-4} for layers 2 and 3, 2.0×10^{-4} for layer 4, and 5.0×10^{-4} for layers 5 and 6.

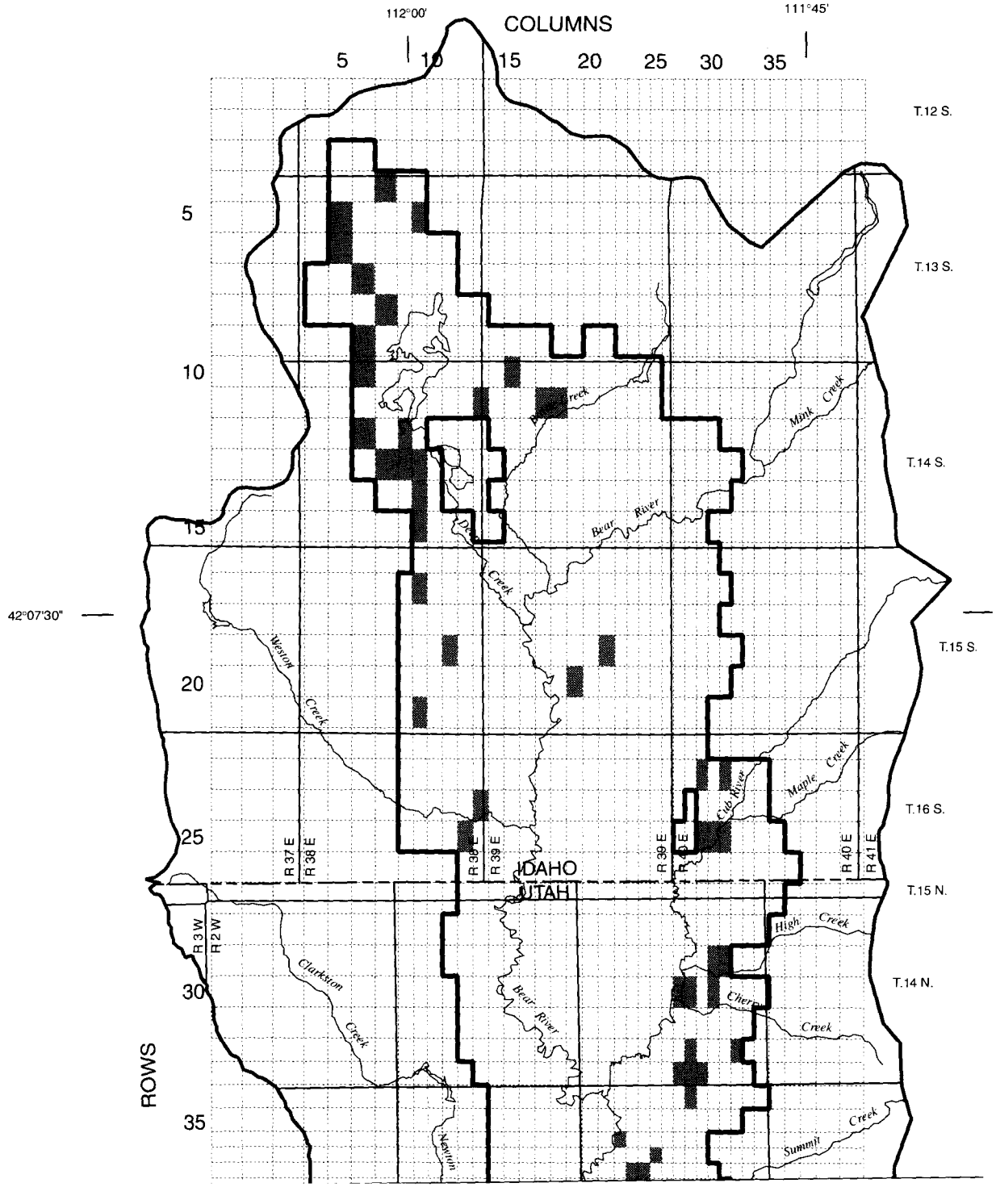
Calibration

The model was calibrated to 1969 steady-state and 1982-90 transient-state conditions. Results of the steady-state calibration were used as initial conditions for the 1982-90 transient-state calibration.

Steady-State Calibration

Bjorklund and McGreevy (1971, p. 1 and 2) stated, "Much of the [ground-water] reservoir is overflowing, and the annual change in storage therefore is assumed to be negligible," and "There is no apparent long-term upward nor downward trend of water levels." With no change in storage or long-term water levels, steady-state conditions were assumed for 1969. Hydrographs of long-term water levels in wells in Cache Valley (fig. 12) show that approximate steady-state conditions existed until 1982. Because of the large amount of ground-water and surface-water data collected during Bjorklund and McGreevy's study, 1969 data primarily were used for the steady-state calibration.

The steady-state calibration involved comparing measured water levels with model-computed water levels for several layers and comparing estimated fluxes with model-computed fluxes. Out of 162 wells used in the calibration, 56 were in layer 1, 67 were in layer 2, 27 were in layer 3, and 12 were in layer 5. No wells with water-level data were available as calibration points for layers 4 and 6. A 10-foot difference between measured and computed water levels was the initial calibration criteria. Because of the complexity of the ground-water system and large horizontal hydraulic gradients in the Weston-Preston-Franklin area, the maximum difference between measured and computed water levels was about 38 feet. The maximum 38-foot difference was the best difference achievable because in the Weston-Preston-Franklin area, the measured horizontal hydraulic gradient is more than 100 feet between nodes that are 0.375 miles apart. This area is part of the toe of the Bear River delta and has relatively large topographic relief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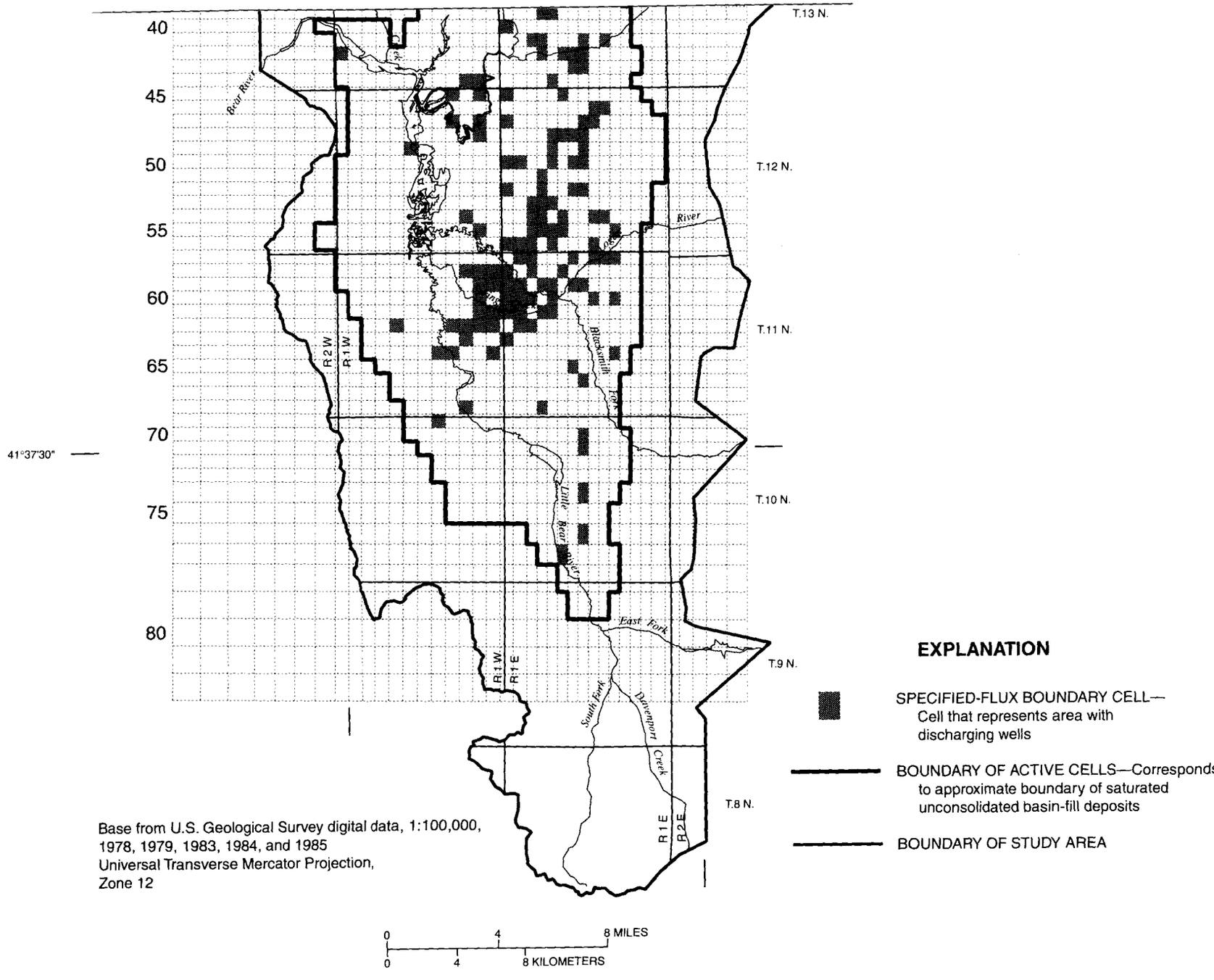


Figure 22. Location of cells that represent simulated withdrawal from wells.

compared with the rest of the valley floor. The mean difference, absolute mean difference, and standard deviation of the difference in measured and model-computed water levels are -0.19 feet, 8.83 feet, and 11.91, respectively. The relation between measured and model-computed water levels in the steady-state simulation is shown in figure 23.

Most of the hydraulic properties were varied during calibration. Hydraulic-conductivity values were changed within a range based on the assumed properties of the unconsolidated basin-fill deposits. Transmissivity values were changed by multiplying the new hydraulic-conductivity values by the thickness of each layer. This process was continued on a trial-and-error basis until the final values of hydraulic conductivity and transmissivity were achieved. The final distribution of hydraulic conductivity in layer 1 is shown in figure 24. The distribution of transmissivity for each layer, except for layers 5 and 6, can be obtained by multiplying the hydraulic conductivity in figure 24 by the thickness of the layer in question. Layers 5 and 6 have the same transmissivity as layer 4. Final hydraulic-conductivity values in layer 1 ranged from 1.0 to 100.1 feet per day. Transmissivity values in layers 2 and 3 ranged from 100 to 10,010 feet squared per day, and in layers 4, 5, and 6 from 200 to 20,020 feet squared per day. The values of transmissivity used in the calibration are within the range of values reported in the "Aquifer properties" section of this report.

Vertical leakance between layers 1 and 2 was varied, within limits based on the properties of the unconsolidated basin-fill deposits, to simulate the vertical gradients observed in Cache Valley. The final values of vertical leakance between layers 1 and 2 ranged from 1×10^{-5} to 2×10^{-2} per day. This is equivalent to a vertical hydraulic conductivity of 1 to 1×10^{-3} foot per day. The final distribution and values of vertical leakance between layers 1 and 2, layers 2 and 3, layers 3 and 4, layers 4 and 5, and layers 5 and 6 are shown in figures 25 to 29.

Vertical-leakance values between layers 2 through 6 were varied slightly during calibration. The final range of vertical leakance between layers 2 and 3 was 2×10^{-5} to 5×10^{-3}, between layers 3 and 4 was 7×10^{-5} to 6.7×10^{-4}, between layers 4 and 5 was 9×10^{-5} to 1.2×10^{-4}, and between layers 5 and 6 was 8×10^{-5} to 2×10^{-4}.

Several boundary conditions were changed during calibration. Conductance values in the general-head boundaries, river boundaries, and drain boundaries were varied to achieve a satisfactory match between measured water levels and fluxes and model-computed water levels and fluxes. Conductance values in the general-head boundary cells ranged from 5×10^{-1} foot squared per day to 1×10^{-3} foot squared per day. The largest values were in the cells that represent ground-water discharge simulated as seepage to Cutler Reservoir, and the smallest conductance values were in some of the cells that represent subsurface inflow from consolidated rock in the Bear River Range.

Conductance values in the river cells ranged from 8×10^{-1} foot squared per day in the cells that represent seepage to the Little Bear River to 5.5×10^{-4} foot squared per day in the cells that represent seepage to the Bear River. The difference in the conductance values is a result of the measured seepage to the stream and the depth to which the stream cuts the unconsolidated basin-fill deposits. The Bear River cuts deeply and the conductance values for those cells were kept small to simulate the measured amount of stream seepage.

Conductance values in the drain cells ranged from 5.0×10^{-1} foot squared per day in cells that represent the Spring Creek (near College Ward) area, to 1×10^{-5} foot squared per day in cells that represent the slough area north of Weston. Generally, conductance values in the drain cells that simulate spring discharge on the east side of Cache Valley were about 1×10^{-2} foot squared per day.

Initially, no general-head boundaries simulating subsurface inflow from consolidated rock or from outside the simulated area were included in the model. Such general-head boundaries were added during steady-state calibration because several areas did not have enough recharge to simulate observed water levels. Consequently, the model was changed to include subsurface inflow from consolidated rock and adjacent ground-water systems. The areas in which general-head boundaries simulated inflow from consolidated rock are the Bear River Range, the Wellsville Mountains, and the Malad Range. Because subsurface inflow from consolidated rock is not well understood, the distribution of the general-head boundary cells along the mountain fronts was determined during calibration by matching the computed water levels in those areas to the measured water levels. Subsurface inflow from the small ground-water system in Weston Canyon also was simulated by general-head boundaries. Subsurface inflow from

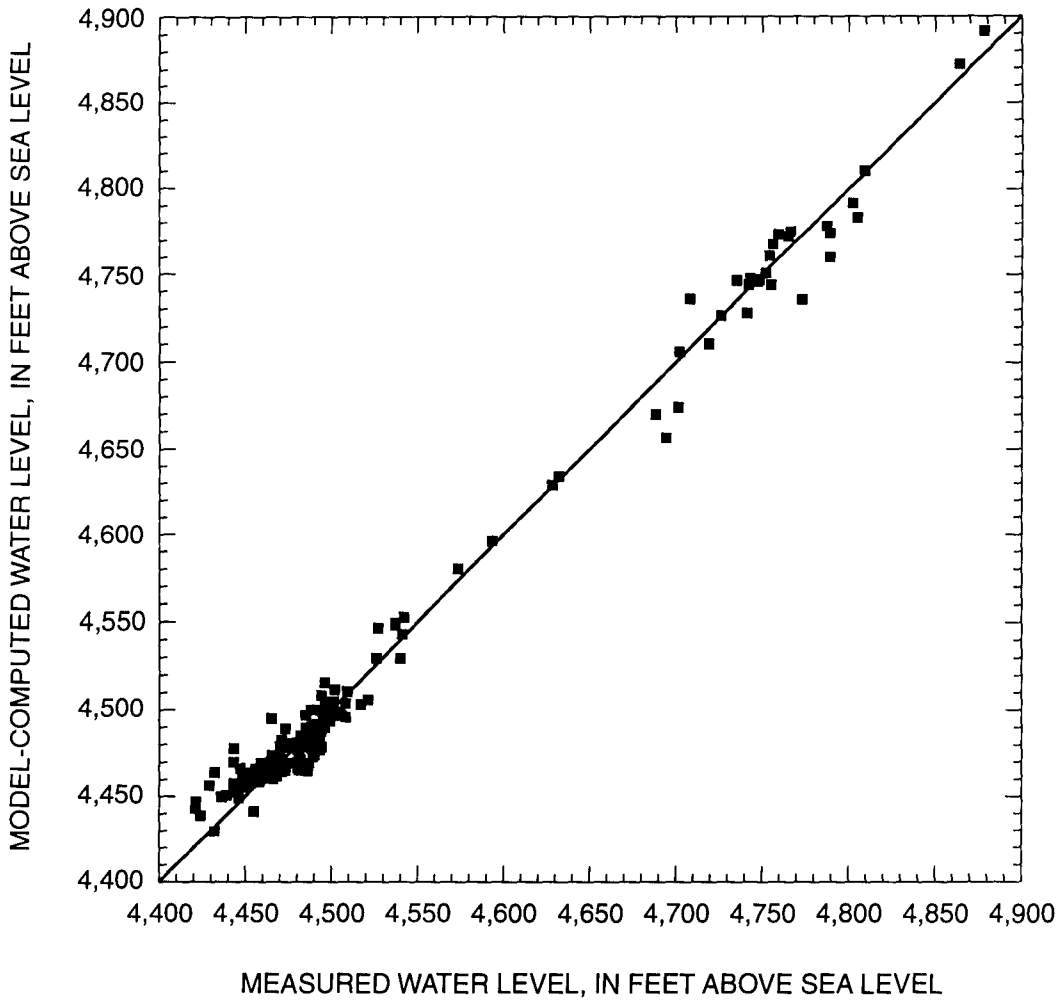
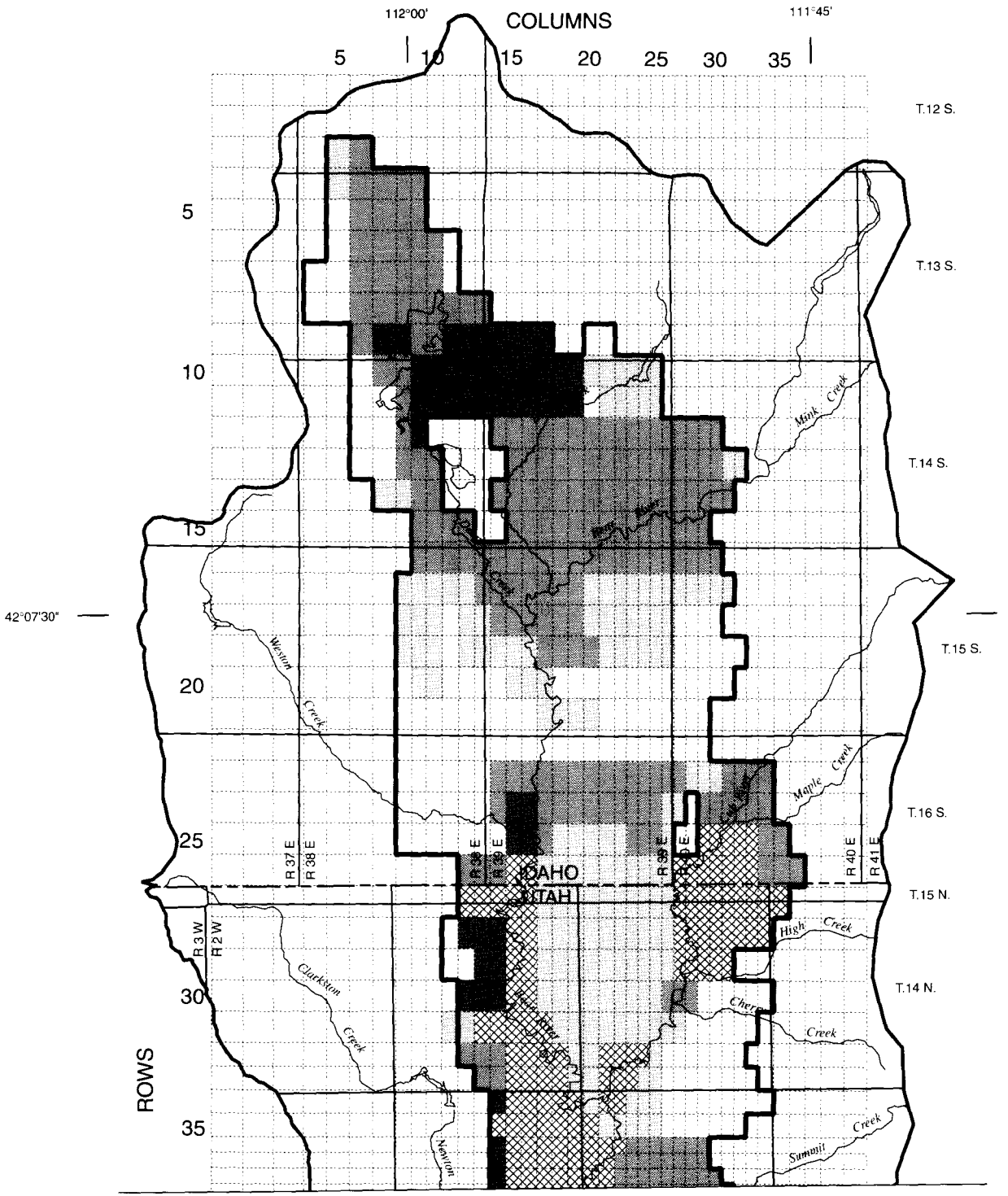


Figure 23. Relation between model-computed and measured water levels in the steady-state simulation, Cache Valley.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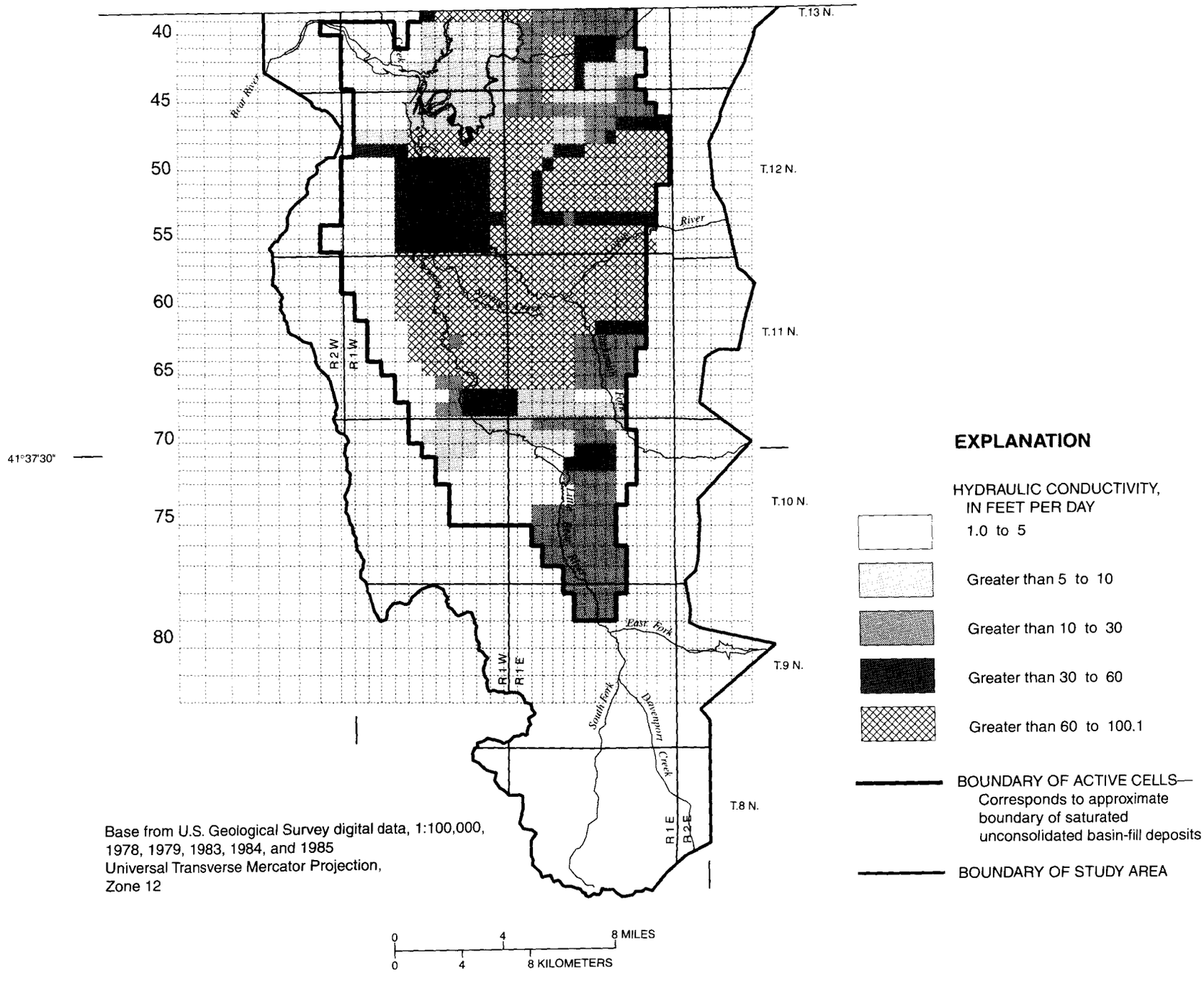
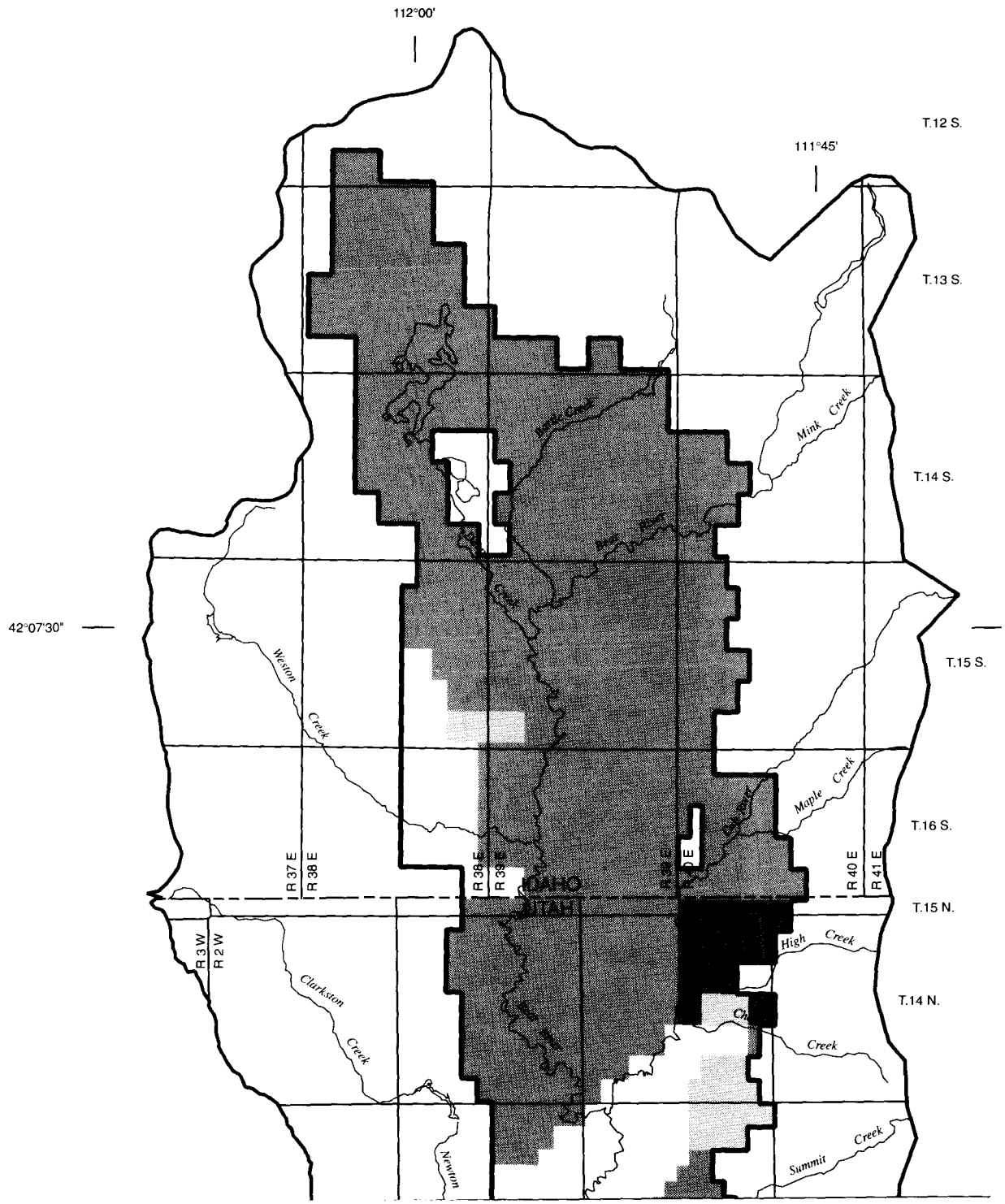


Figure 24. Final distribution of hydraulic conductivity in layer 1.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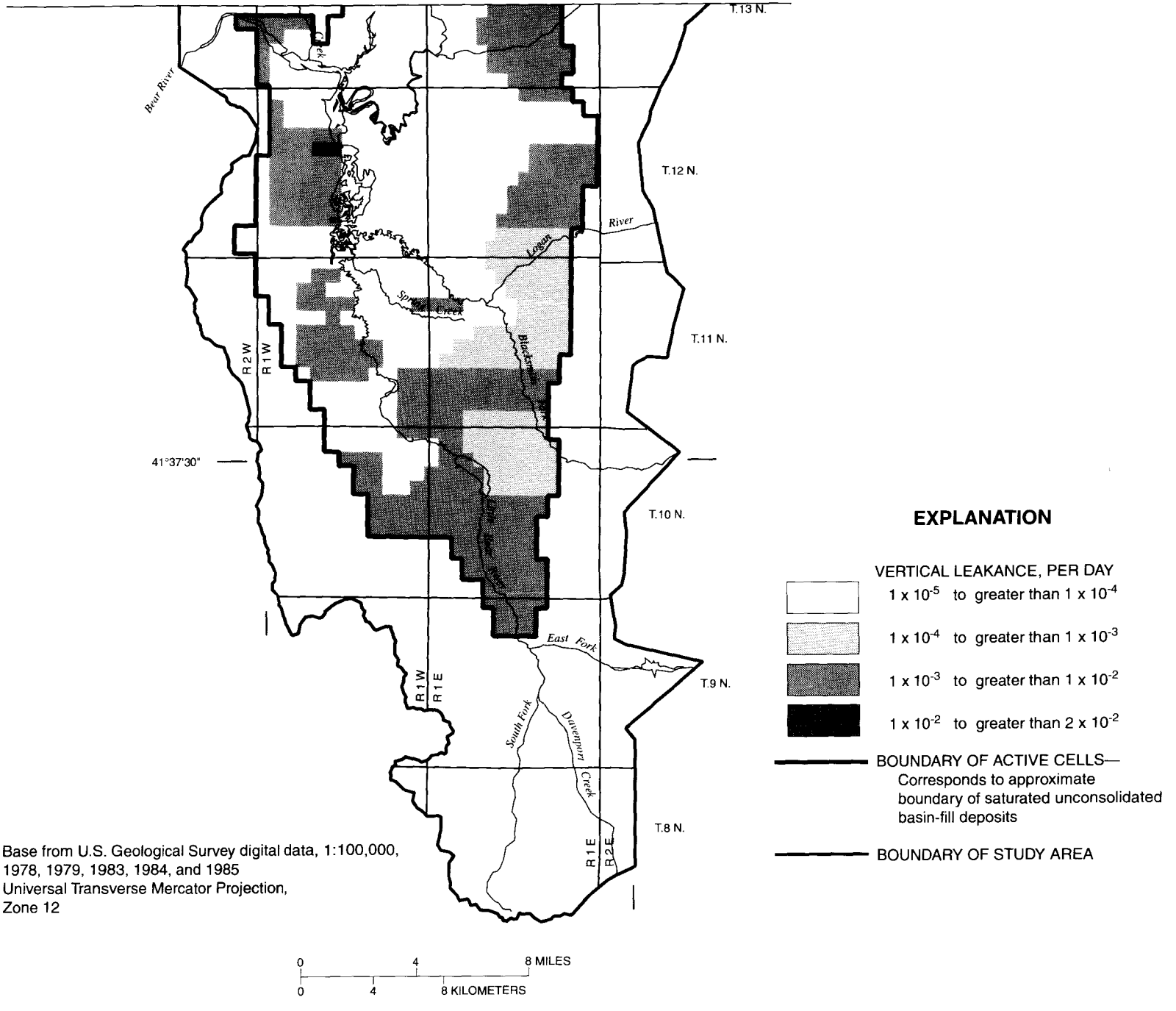
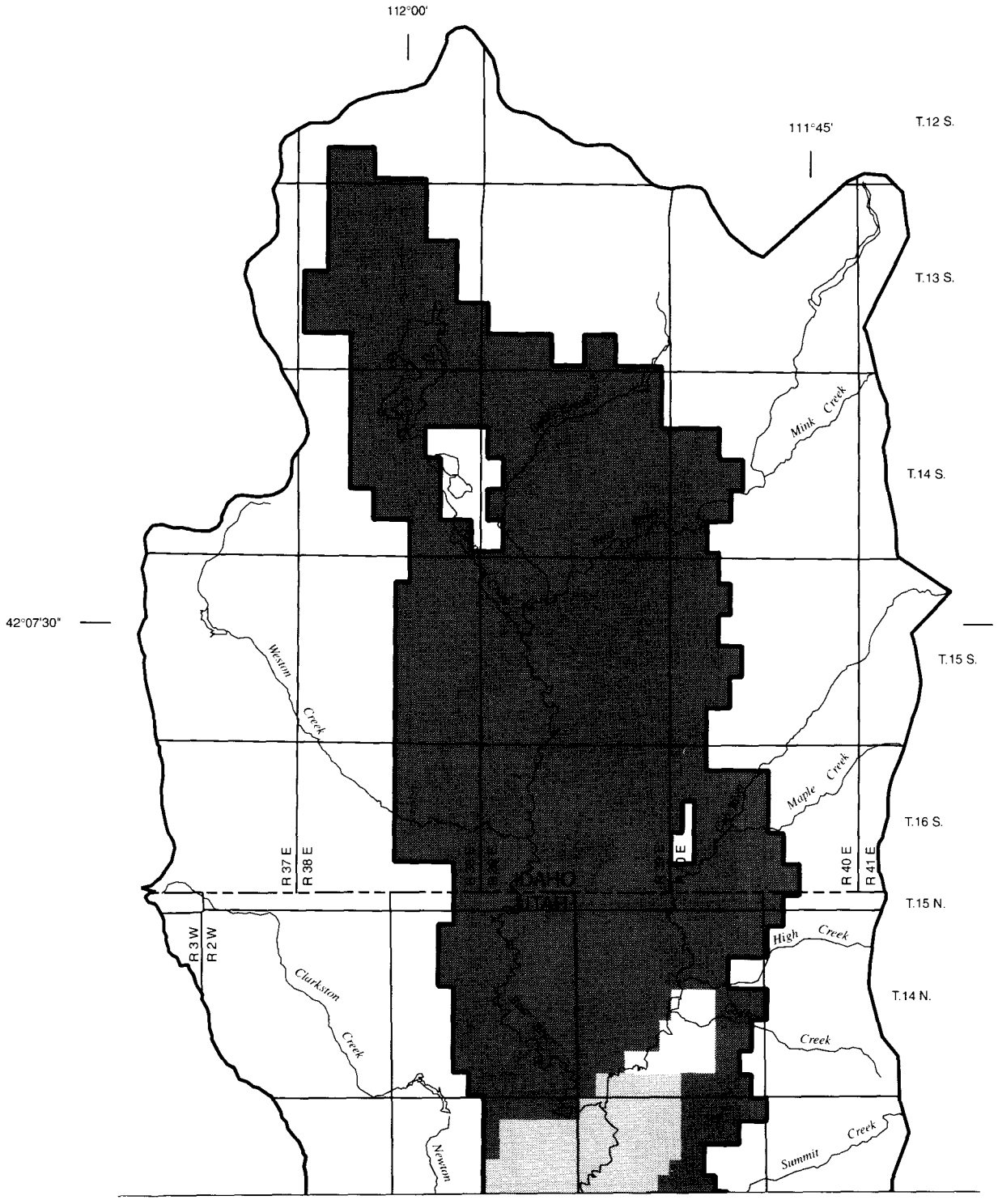


Figure 25. Final distribution of vertical leakance between layers 1 and 2.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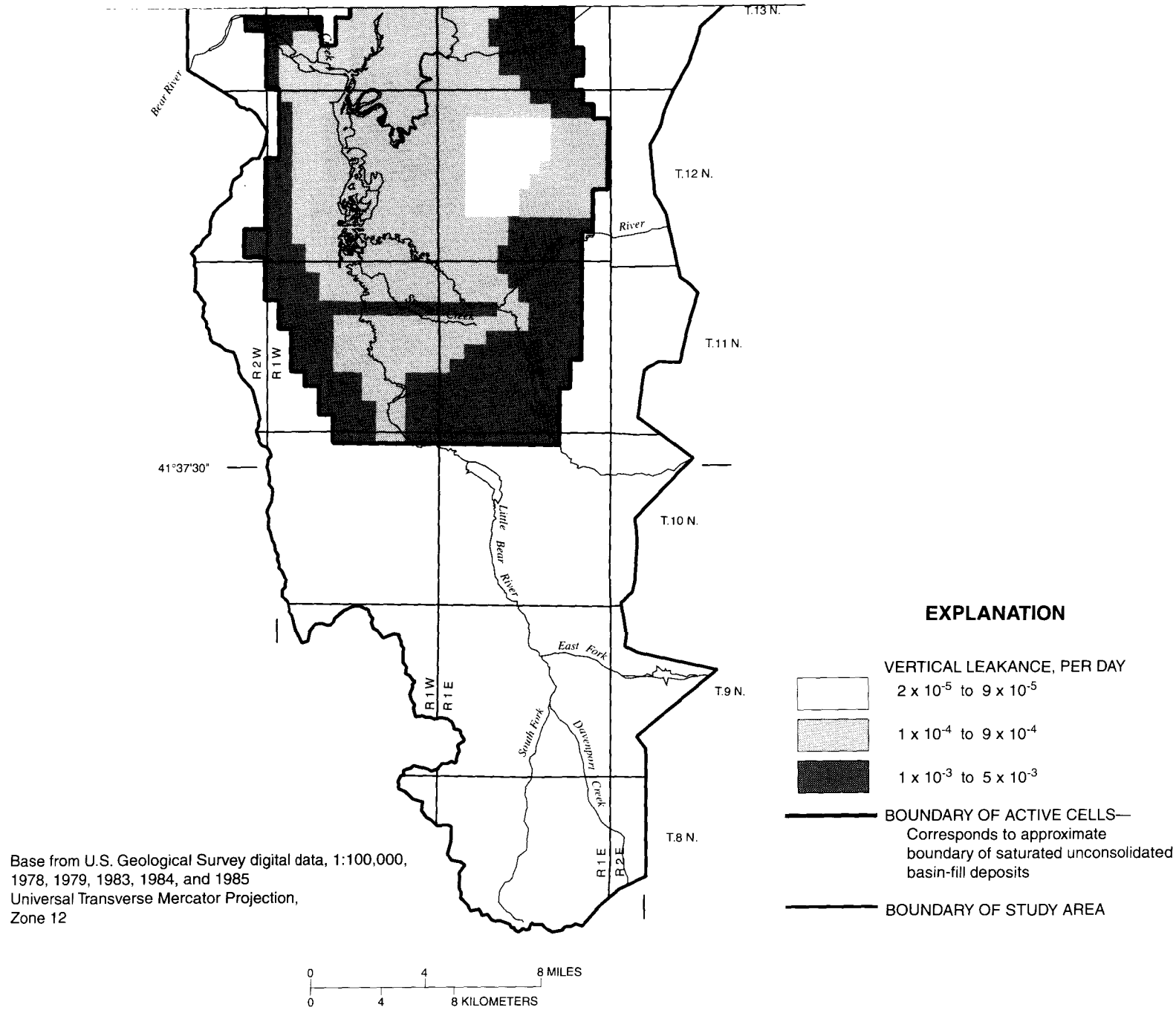
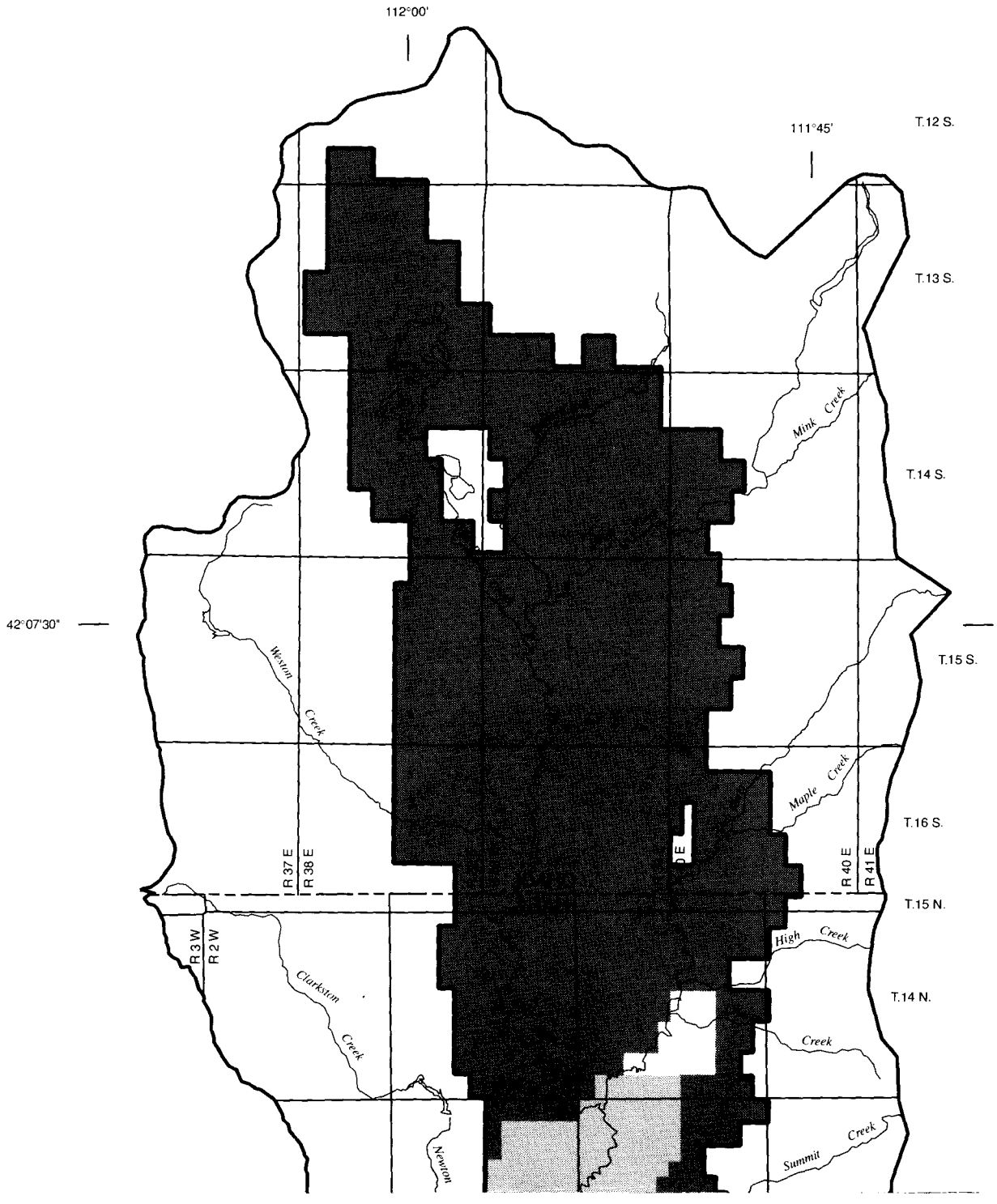


Figure 26. Final distribution of vertical leakance between layers 2 and 3.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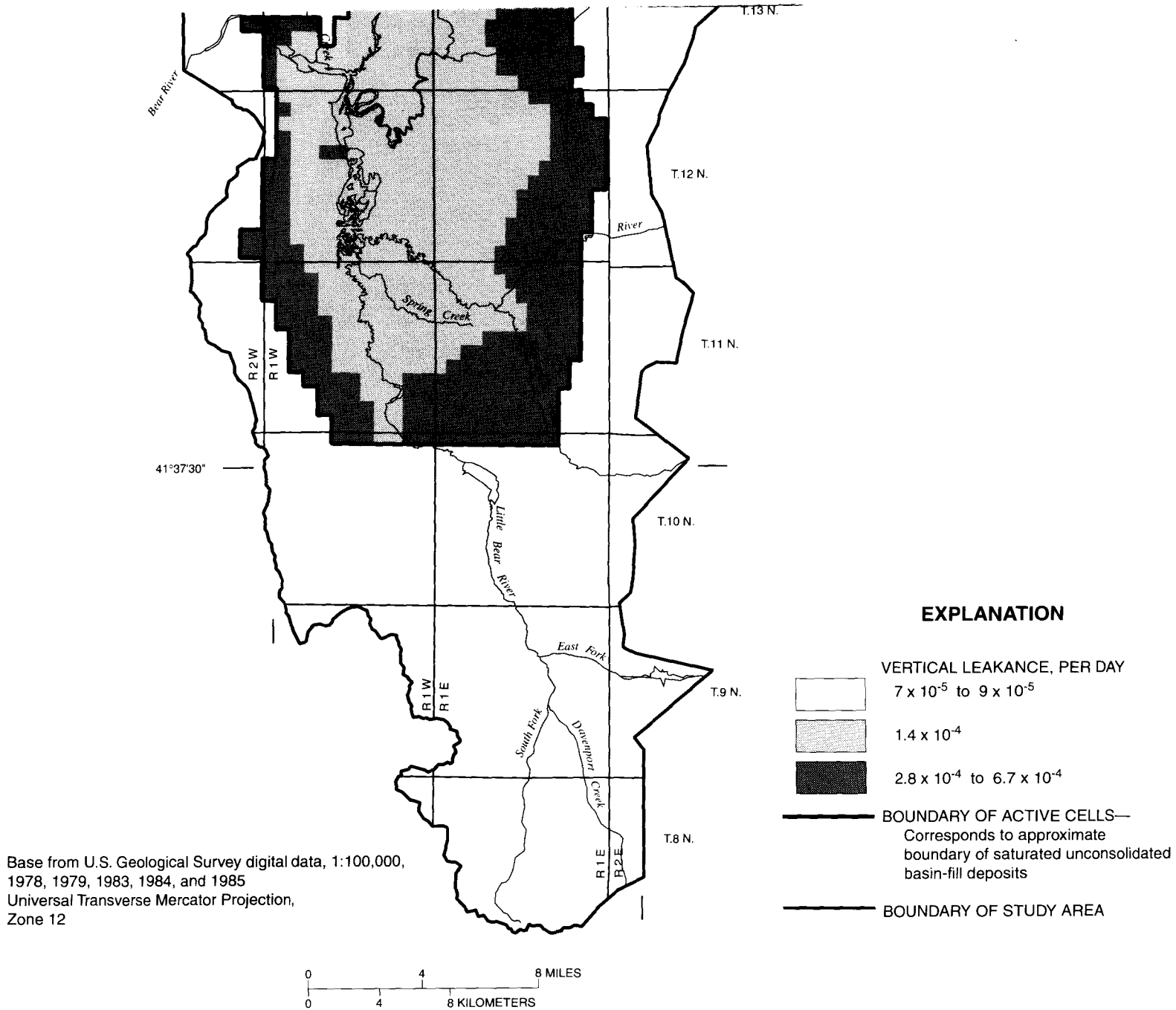
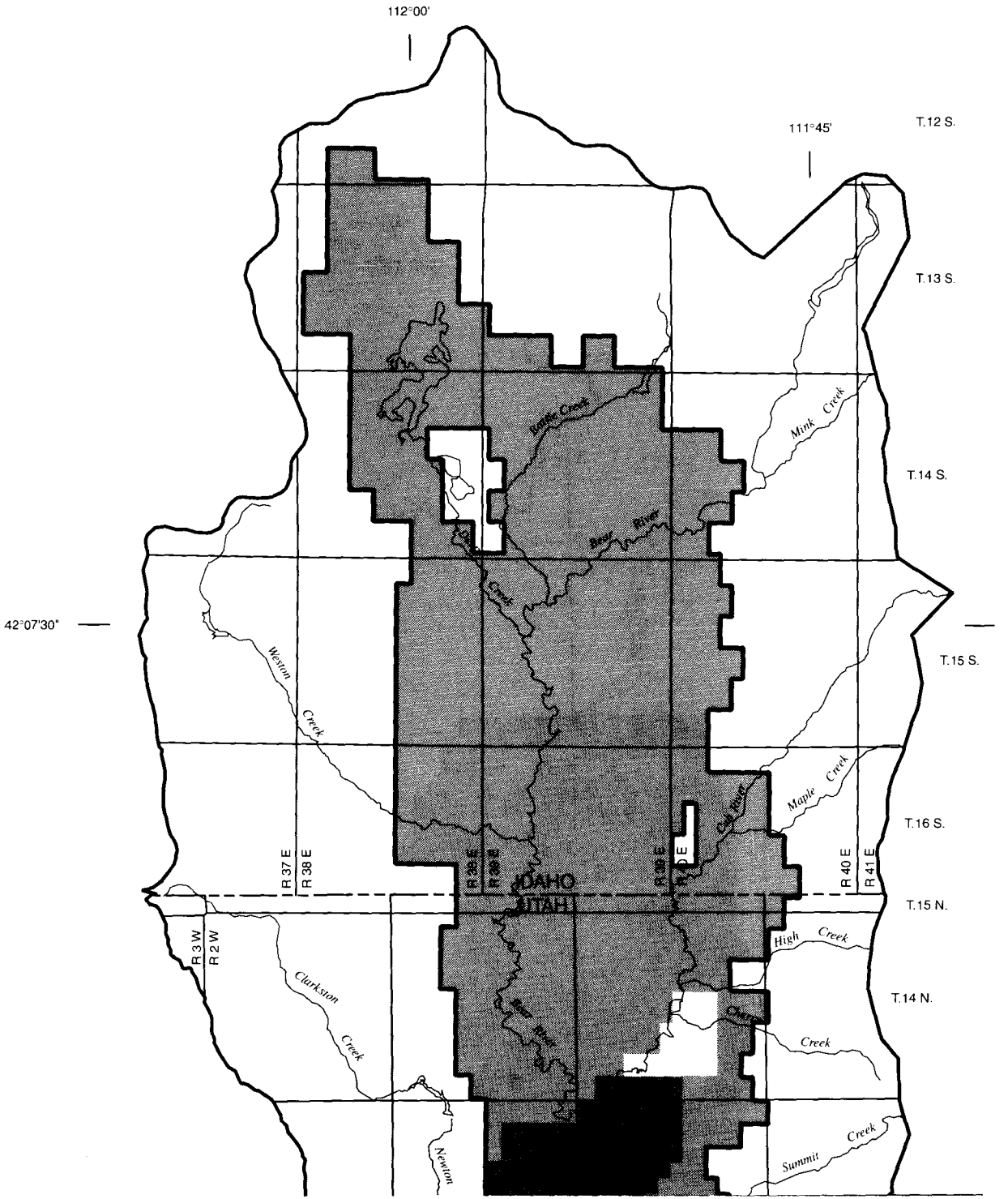


Figure 27. Final distribution of vertical leakance between layers 3 and 4.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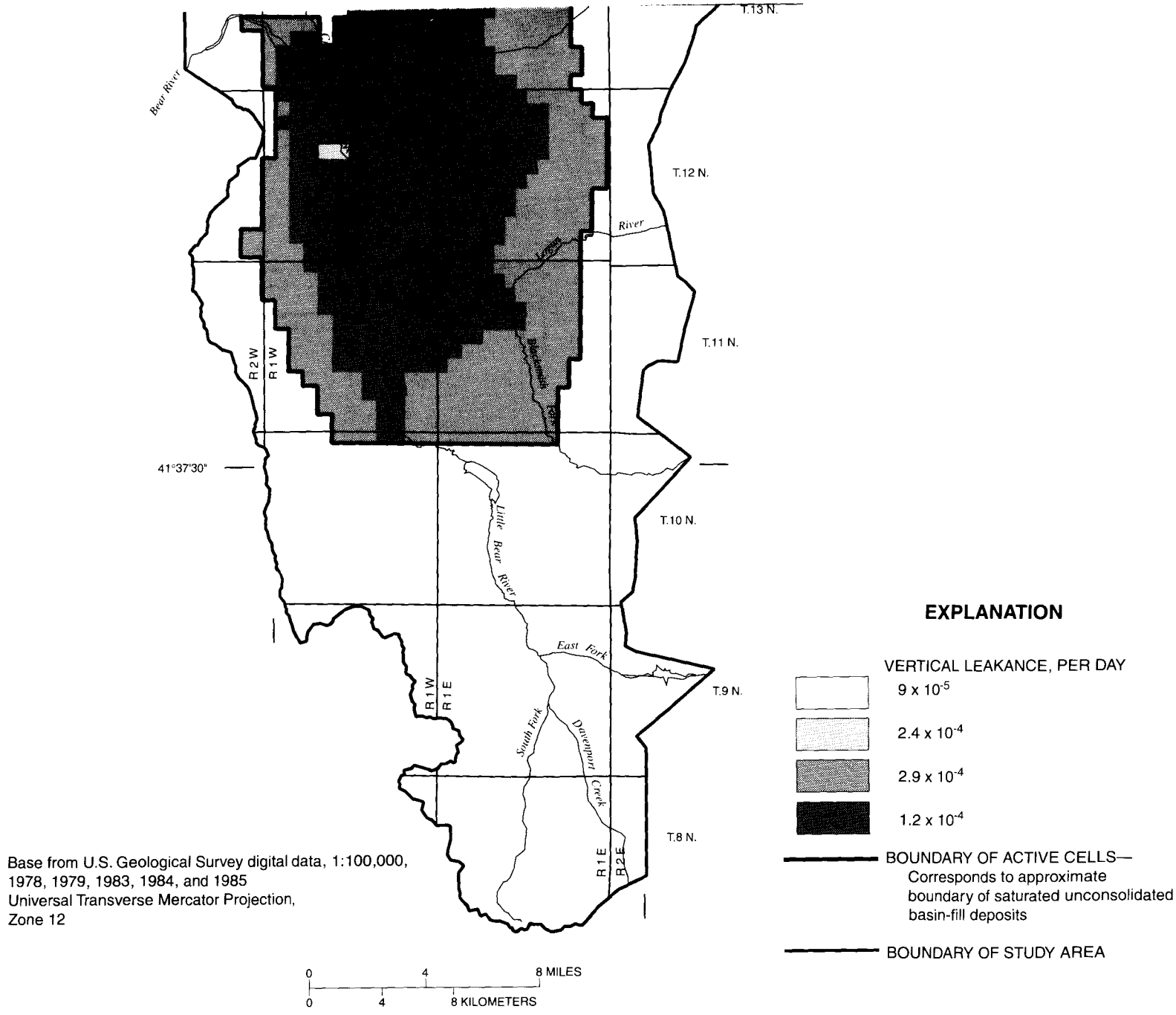
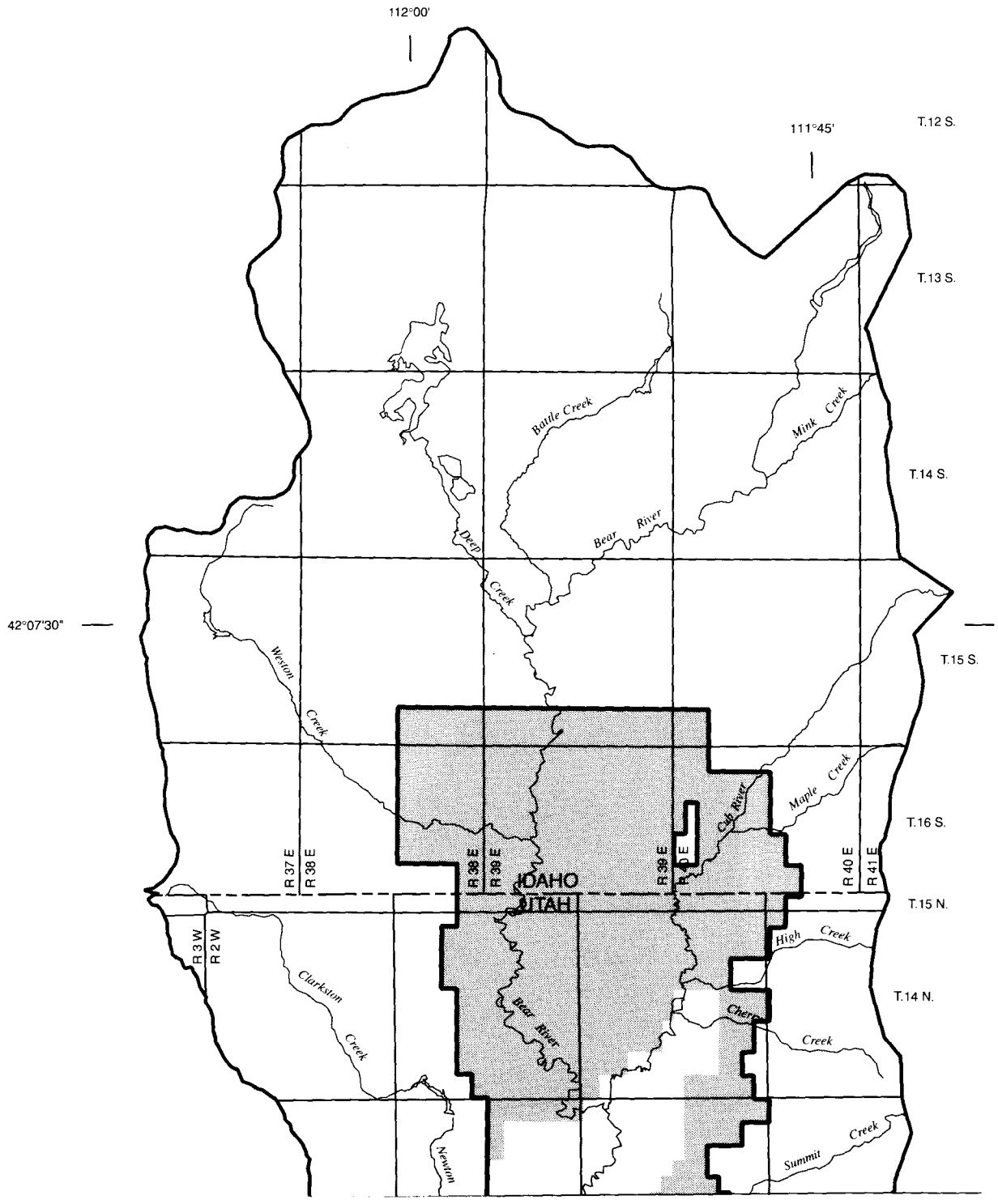


Figure 28. Final distribution of vertical leakance between layers 4 and 5.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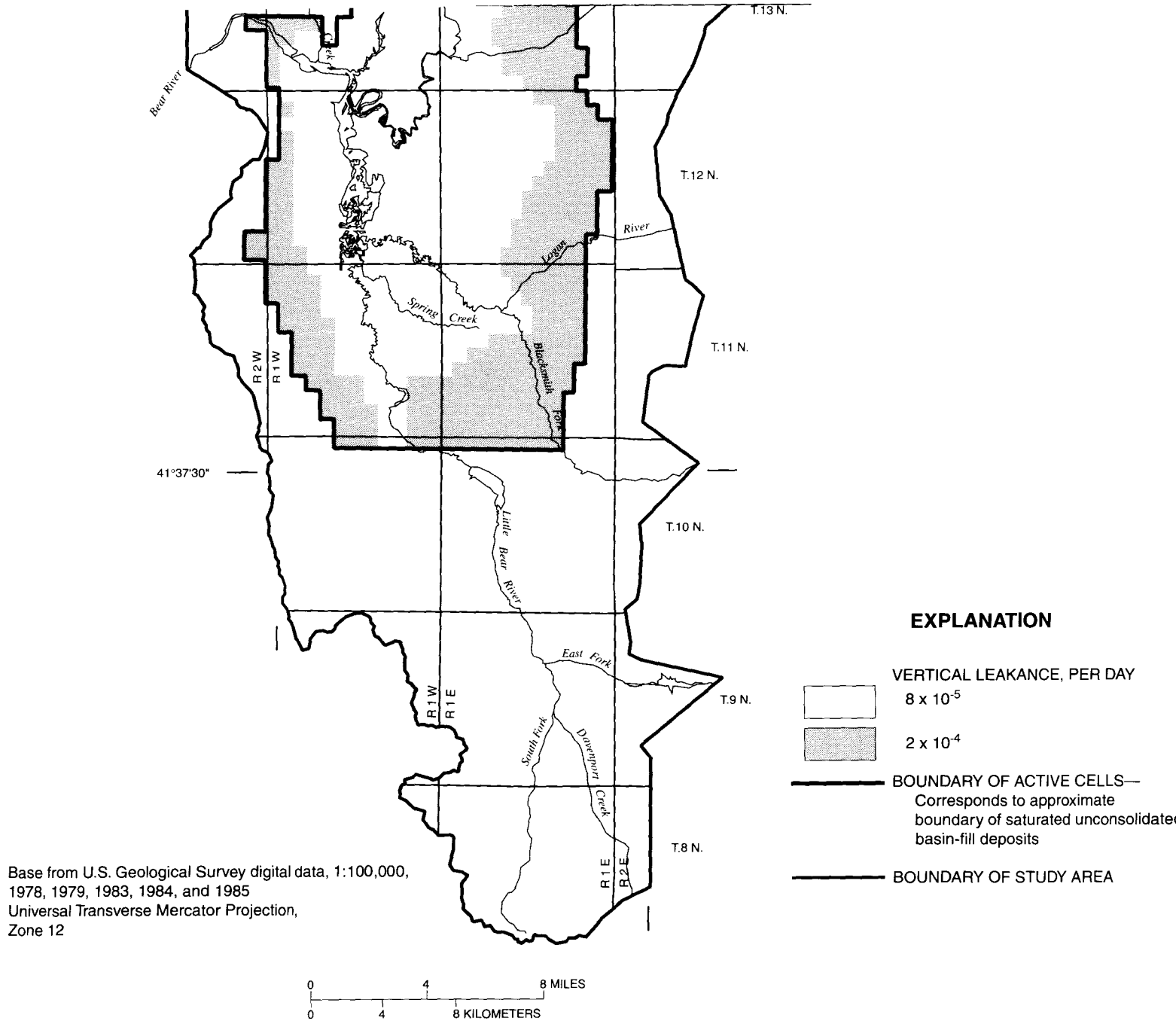


Figure 29. Final distribution of vertical leakance between layers 5 and 6.

the ground-water system in the Clarkston area was not simulated because it was assumed that most of the ground water from the Clarkston area discharges directly into Cutler Reservoir. Indirect evidence of recharge from consolidated rock and adjacent ground-water systems is presented in the "Ground water" section of this report.

One part of the calibration was to compare model-computed fluxes to measured and estimated fluxes. Estimated and model-computed ground-water budgets for the steady-state simulation and the end of the 1990 transient-state simulation are presented in table 11. Estimated recharge from infiltration of precipitation and unconsumed irrigation water, and seepage from canals is similar to simulated recharge for these components. This similarity is expected because DPM estimates were used in the estimated ground-water budget (table 5) and as specified recharge in steady-state calibration.

Table 11. Estimated and model-computed budgets for the steady-state simulation and the end of the 1990 transient-state simulation of the ground-water system in the unconsolidated basin-fill deposits in Cache Valley

Budget component	Steady-state simulation		1990 transient-state simulation	
	Estimated flow (cubic feet per second)	Model-computed flow (cubic feet per second)	Estimated flow (cubic feet per second)	Model-computed flow (cubic feet per second)
Recharge				
Infiltration of precipitation and unconsumed irrigation water, and seepage from canals	346	300	197	187
Seepage from streams	7	3	3	4
Other forms of recharge[1]	[2]96	67	[3]96	66
Total recharge	449	370	296	257
Discharge				
Seepage to streams	101	120	[3]101	112
Spring discharge	138	84	[3]138	78
Evapotranspiration	87	74	[3]87	74
Seepage to reservoirs	79	48	[3]79	45
Withdrawal from wells	44	44	52	52
Total discharge	449	370	457	361
Change in storage[4]	0	0	-161	-104

[1] Includes subsurface inflow from adjacent consolidated rock and unconsolidated basin-fill-deposit ground-water systems, and seepage from ephemeral streams.

[2] Difference between total discharge and recharge from infiltration of precipitation and unconsumed irrigation water, and seepage from canals and streams.

[3] Assumed to be the same as in 1969.

[4] Total recharge minus total discharge. Value indicates flow of ground water to or from (-) ground-water storage.

Model-computed recharge from seepage from major streams during the steady-state simulation was less than half that of the estimated seepage from major streams (table 11). The difference in this "Seepage from streams" component of the budget is small, however, when compared with other budget components. Estimated and model-computed steady-state recharge from "Other forms of recharge" differs by 29 cubic feet per second. Because the estimated ground-water recharge from "Other forms of recharge" is based on estimations and assumptions that have a large degree of uncertainty associated with them, the difference in the estimated and model-computed values may not be significant.

The estimated and model-computed values of seepage to streams are similar, but there are some differences in the amount of seepage to individual streams. Simulated seepage to rivers, except the Cub River, was about the same as estimated seepage to the individual streams. Simulated seepage to the Cub River (33.5 cubic feet per second, or 24,300 acre-feet per year) was about 10 cubic feet per second (7,200 acre-feet per year) larger than the estimated seepage of 23.9 cubic feet per second (17,300 acre-feet per year). The difference is not significant because the method of estimating seepage to the Cub River was a rough approximation, as described in the "Discharge" section of the "Unconsolidated basin-fill deposits" section of this report.

Estimated spring discharge in 1969 and also in 1990 (table 11) is greater than that computed by the model. The difference may be a result of the complexity of the springs in Cache Valley. Some of the springs included in the estimated spring discharge may be discharging from perched ground-water systems. The model was not designed to simulate water discharging from perched ground-water systems.

A fairly close match was achieved between the estimated (87 cubic feet per second, or 63,000 acre-feet per year) and model-computed (74 cubic feet per second, or 53,600 acre-feet per year) evapotranspiration (table 11). The difference between the two values could be because the model cannot accurately simulate evapotranspiration along the flood plain of the Bear River. Because the flood plain is so deeply cut into the surrounding unconsolidated basin-fill deposits, a steep localized ground-water gradient exists near the river and the model is unable to reproduce this gradient. Thus, evapotranspiration along the Bear River, which is affected by water levels near the river, cannot be accurately simulated by the model.

Model-computed seepage to Cutler Reservoir is slightly more than half the estimated seepage. Conductance values of the general-head boundary were manipulated to increase seepage to Cutler Reservoir. The estimate of seepage to Cutler reservoir is based on the seepage study of the Bear River and Cutler Reservoir (Herbert and Thomas, 1992) that was done at the end of the irrigation season. Flow into Cutler Reservoir might have been at its highest rate of the year and not at its average rate, which is the rate projected in the simulation. Also, the seepage study of the Bear River and Cutler Reservoir accounts for ground-water inflow from the Clarkston area, which was not simulated by the model.

Transient-State Calibration

The transient-state calibration was done by simulating the ground-water system using 1-year stress periods for 1982-90. One-year stress periods were used because water-level data are collected on a yearly basis, primarily during March. The simulation used 10 equal time steps during the first stress period (1982), 5 equal time steps during the second stress period (1983), 2 equal time steps during the third stress period (1984), and 1 time step during the 6 remaining stress periods (each of the years 1985 through 1990). More time steps were required in the first few stress periods to insure a smooth transition between the steady-state solution and the initial transient-state stress periods.

The main stresses on the ground-water system were changes in recharge resulting from wet and dry climatic changes, and increases in ground-water withdrawal. Hydrographs in figure 12 show that water levels remained fairly constant during 1969-81. This lack of variation indicates that there were no major stresses imposed on the main ground-water system in the unconsolidated basin-fill deposits in Cache Valley during this time and that the system was essentially under steady-state conditions. A record wet period began in 1982 and continued until 1986. After 1986, a dry period began and continued through 1990 (fig. 4). The aquifer properties, boundary conditions, and initial heads from the steady-state simulation were used as initial conditions for the transient-state calibration.

The transient-state calibration involved adjusting storage properties and recharge while visually comparing changes in simulated water levels with changes in water levels measured in long-term observation wells, and comparing all of the model-computed budget components at the end of the 1990 stress period with estimated 1990 budget components. After the initial adjustments, changes in the model-computed water levels did not simulate the increase in observed water levels during the 1982-84 period, and model-computed water levels declined much faster than observed water levels during the 1987-90 period. Inspection of the initial yearly estimates of recharge from infiltra-

tion of precipitation and unconsumed irrigation water (table 6) showed that the initial yearly estimates of recharge are fairly constant during both periods; therefore, the validity of the assumptions used to compute recharge from infiltration of precipitation and unconsumed irrigation water was questionable. To test the assumptions, the DPM of Bauer and Vaccaro (1987) was used to compute recharge. A description of the DPM and its data requirements are presented in the appendix.

Recharge estimates for the steady-state simulation were adjusted by multiplying the initial estimate of recharge by the correction factor shown in table 6. The adjusted recharge gave satisfactory results in the steady-state simulation, and additional recalibration was not necessary. Yearly recharge estimates for the transient-state calibration (1982-90) were adjusted using the same method, and the results of the transient-state calibration were improved.

Storage values were varied during transient-state calibration until suitable matches between model-computed and measured water levels, and model-computed and estimated budget components, were obtained. Final values of specific yield for layer 1 ranged from 0.01 to 0.3 so that measured water levels and fluxes matched model-computed water levels and fluxes. This range of specific-yield values is within a range of values reported by Johnson (1967, table 29). The final distribution of specific yield is shown in figure 30. The storage-coefficient values of 1.0×10^{-4} for layers 2 and 3, 2.0×10^{-4} for layer 4, and 5.0×10^{-4} for layers 5 and 6 were not varied during calibration. Hydrographs showing measured and model-computed water-level changes for the 1982-90 transient-state simulation are shown in figure 31. Comparison of the hydrographs indicates that the model-computed water levels agree reasonably with measured water levels in most parts of Cache Valley.

Calibration also was done by comparing model-computed budget components for 1990 to estimated budget components for 1990. The model-computed and estimated 1990 ground-water budgets are shown in table 11. The differences between estimated and model-computed budget components followed the same pattern in the1969 steady-state simulation. The model-computed flow of ground water from storage at the end of 1990 was about 104 cubic feet per second (75,300 acre-feet per year)(table 11). The flow of ground water from storage is reflected by a general decrease in water levels in the valley during 1989-90 (figs. 12 and 31). The model-computed annual ground-water budget for the 1982-90 transient-state simulation is shown in table 12. The difference between total recharge and total discharge is the amount of water entering or leaving storage in the ground-water system. The results of the simulation for 1982-84 and 1986 show rising water levels and flow of ground water into storage, and for 1985 and 1987-90 show declining water levels and flow of ground water out of storage. Average (1982-90) values of the model-computed budget components (table 12) are greater than the steady-state values (table 11). This indicates that during the 1982-90 simulation period, the amount of water recharging to and discharging from the ground-water system in the unconsolidated basin-fill deposits was greater than average.

Projected Water-Level and Ground-Water Budget Changes Resulting from Simulation of Less-Than-Average Recharge and Increased Withdrawal

Two sample simulations run with the Cache Valley ground-water model can be used to gain a better understanding of the interconnection of the surface- and ground-water systems in the valley. One simulation was used to project water levels and budget components if the dry conditions of 1990 continued for 5 years. The other simulation was used to project water levels and budget components if normal recharge occurred and ground-water withdrawals increased by 30 cubic feet per second (21,700 acre-feet per year) for 30 years. For the simulation of dry conditions, all parameters (hydrologic conditions and hydraulic properties that were used as input to the model) were kept at the 1990 values. The simulation used 1-year stress periods with one time step for each stress period. Because the simulation was a continuation of the previous transient-state simulation, with no changes in the stresses, additional time steps in the first stress period were not required. The model-computed transient-state budget for the simulation of less-than-average recharge at the end of the 5-year simulation is shown in table 13 and indicates that the largest changes in budget components from 1990 conditions are decreases in seepage to streams and reservoirs, and spring discharge.

Table 12. Model-computed budget for the 1982-90 transient-state simulation of the ground-water system in the unconsolidated basin-fill deposits in Cache Valley

[All values are reported in cubic feet per second; values are computed for the end of each year]

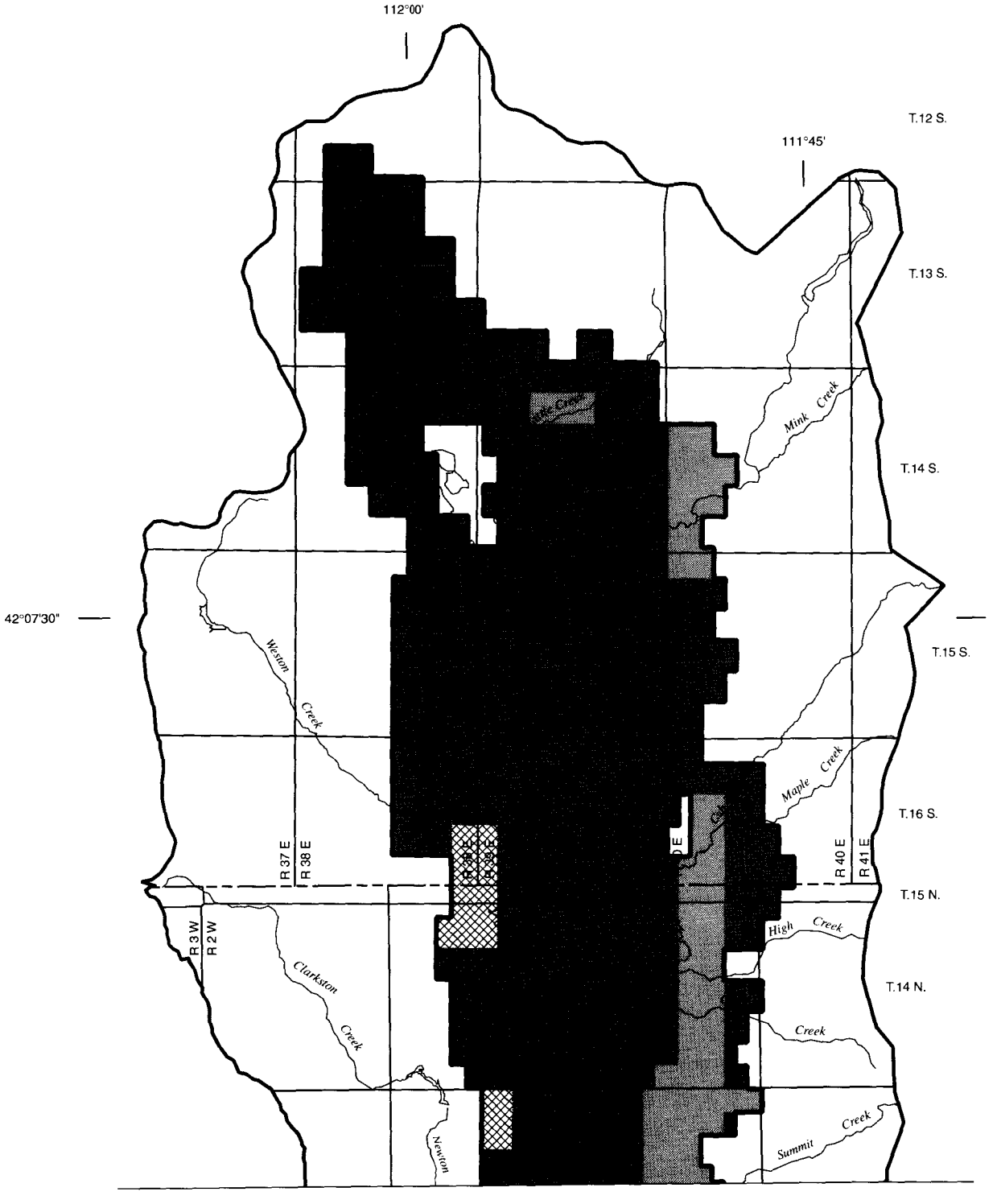
Budget component	Year									
	1982	1983	1984	1985	1986	1987	1988	1989	1990	Average
Recharge										
Infiltration of precipitation and unconsumed irrigation water, and seepage from canals	646	655	563	322	444	227	204	290	187	393
Seepage from streams	1	0	0	1	1	1	2	2	4	1
Other forms of recharge[1]	66	65	64	64	64	65	65	65	66	65
Total recharge	713	720	627	387	509	293	271	357	257	459
Discharge										
Seepage to streams	141	150	153	141	150	132	121	121	112	136
Spring discharge	118	126	128	108	123	96	85	87	78	105
Evapotranspiration	74	74	74	74	74	74	74	74	74	74
Seepage to reservoirs	70	75	72	58	62	51	48	51	45	59
Withdrawal from wells	42	35	29	36	41	42	53	50	52	42
Total discharge	445	460	456	417	450	395	381	383	361	416
Change in storage[2]	268	259	171	-30	59	-101	-108	-24	-104	43

[1]Includes subsurface inflow from adjacent consolidated rock and unconsolidated basin-fill deposit ground-water systems, and seepage from ephemeral streams.

[2]Total recharge minus total discharge. Value indicates flow of ground water to or from (-) ground-water storage. Because of rounding, total recharge minus total discharge may be slightly different than the change in storage.

The model-computed water-level change in layer 3 after the simulation of less-than-average recharge is shown in figure 32. The lines of equal water-level decline in figure 32 were computer generated from the model-computed water-level declines using linear interpolation. Water levels declined more than 20 feet in the south end of Cache Valley and declined an average of about 10 feet between Richmond and Hyrum. The smallest change in water levels occurred in the northern part of the valley.

Increased withdrawal from wells was simulated under conditions of average (1982-90) recharge (table 12). A new steady-state simulation with average recharge conditions and 1990 ground-water withdrawal was constructed and run. Some of the budget components in the new steady-state simulation (table 13) have slightly different flows than either the 1990 transient-state budget components or the 1982-90 average budget components (table 12) because of different conditions. The results of the new steady-state simulation were used as initial conditions for a transient-state simulation that used modified 1990 withdrawal rates. The modification to the 1990 withdrawal rates was the addition of three well fields, each pumping 10 cubic feet per second, in the Logan, Smithfield, and College Ward areas. The pumping was simulated in layer 4 of the model. The simulation used a 30-year stress period with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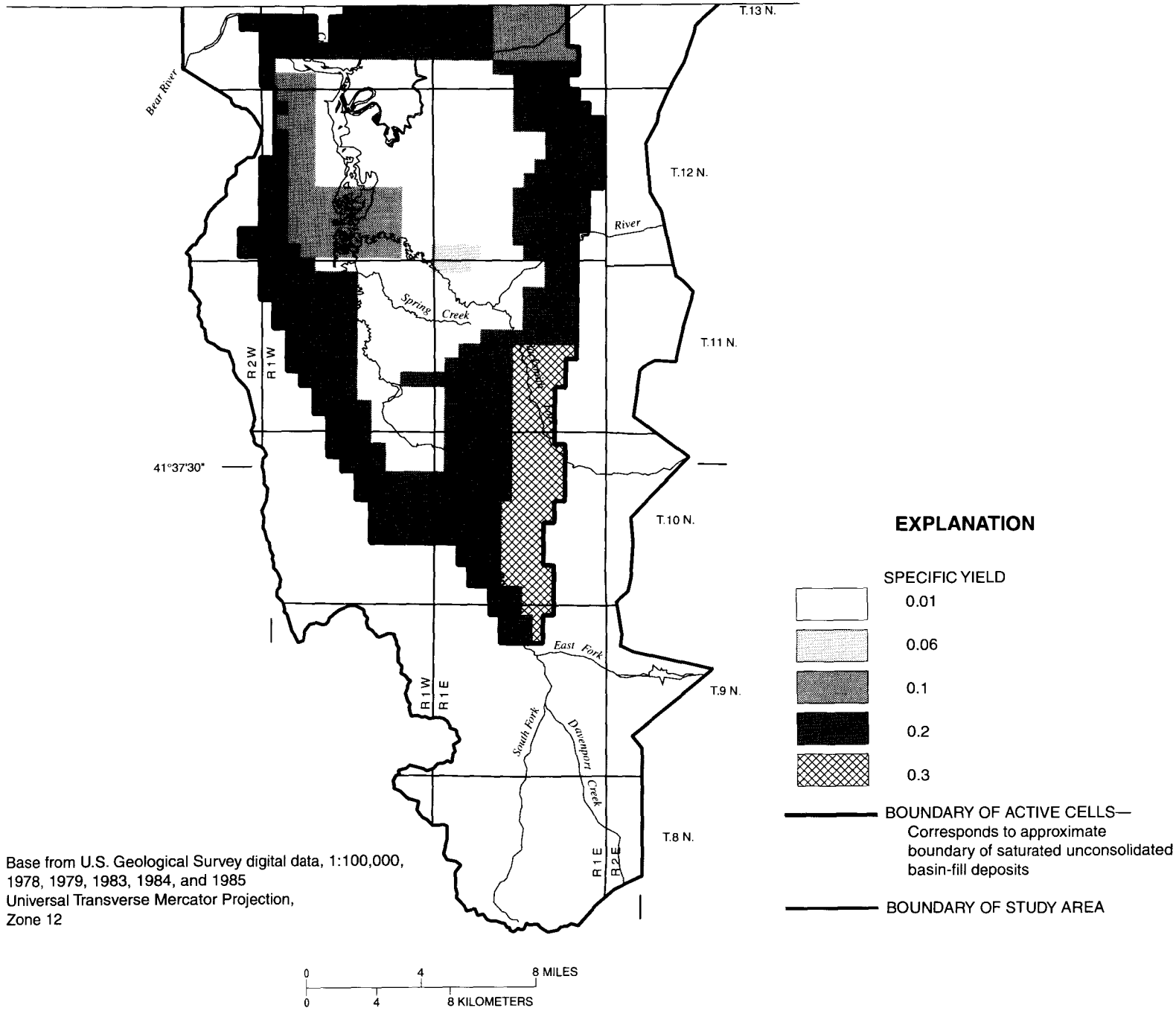


Figure 30. Final distribution of specific yield in 1982-90 transient-state simulation.

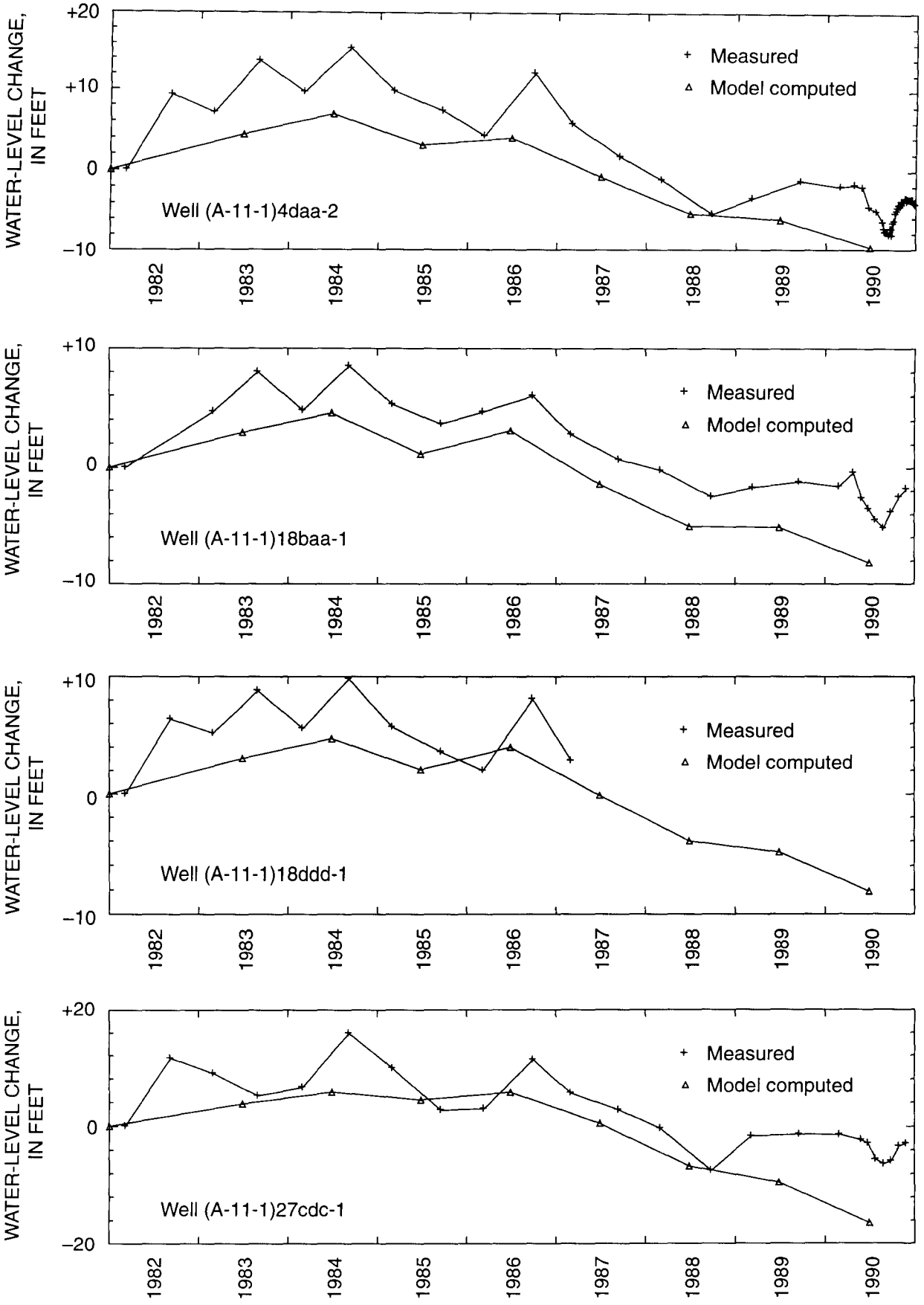


Figure 31. Measured and model-computed water-level changes in selected wells in Cache Valley for the 1982-90 transient-state simulation.

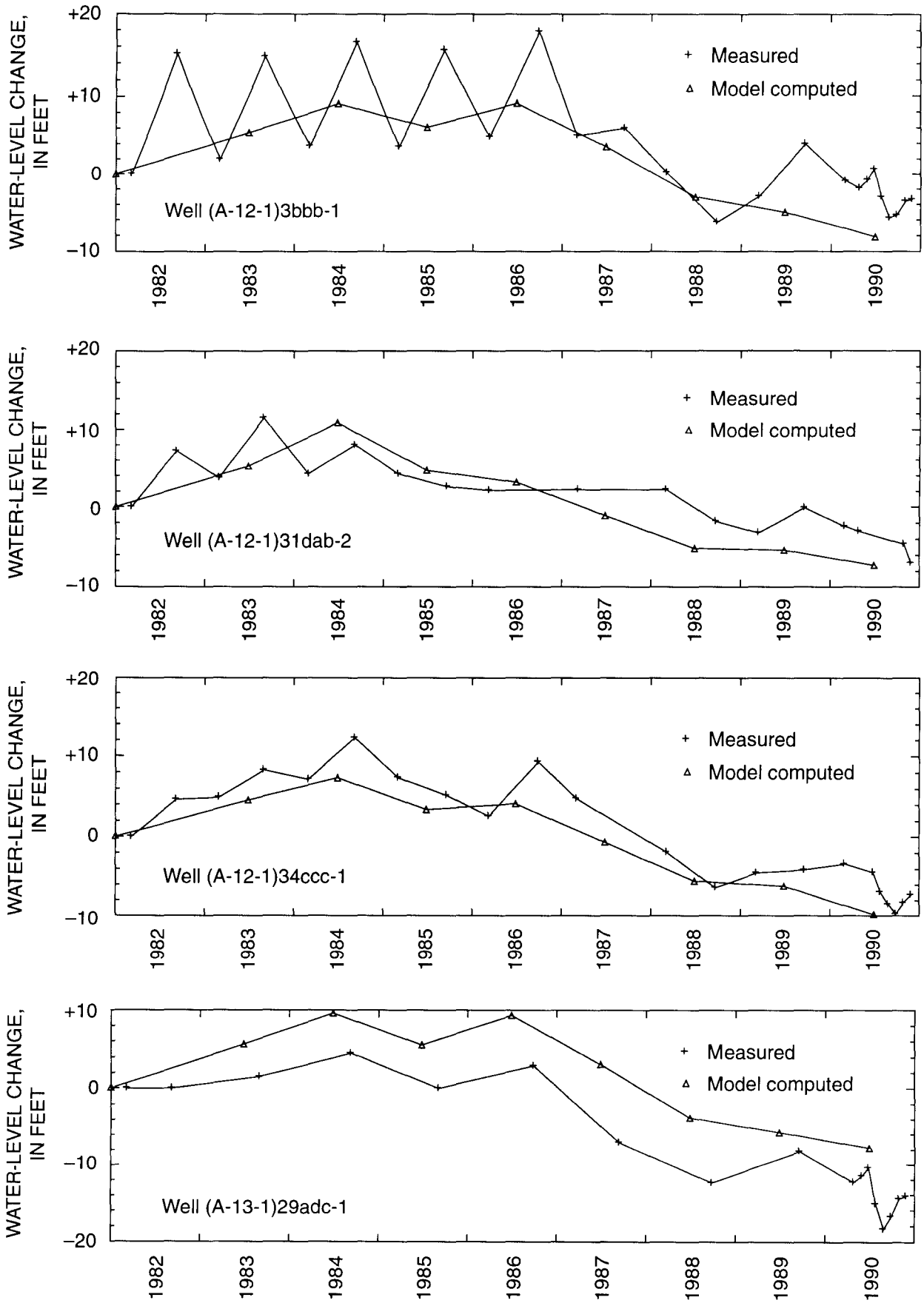


Figure 31. Measured and model-computed water-level changes in selected wells in Cache Valley for the 1982-90 transient-state simulation—Continued.

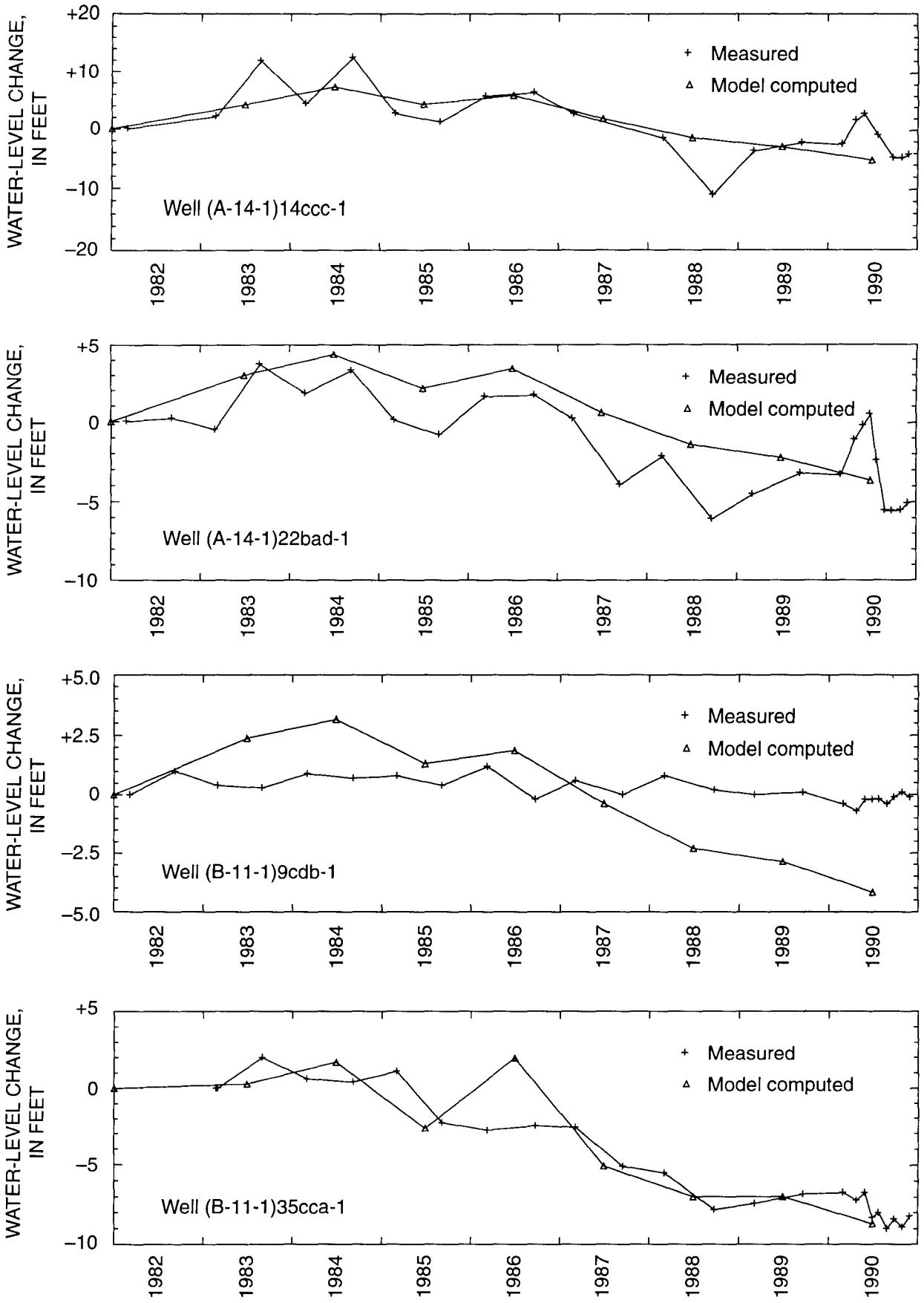


Figure 31. Measured and model-computed water-level changes in selected wells in Cache Valley for the 1982-90 transient-state simulation—Continued.

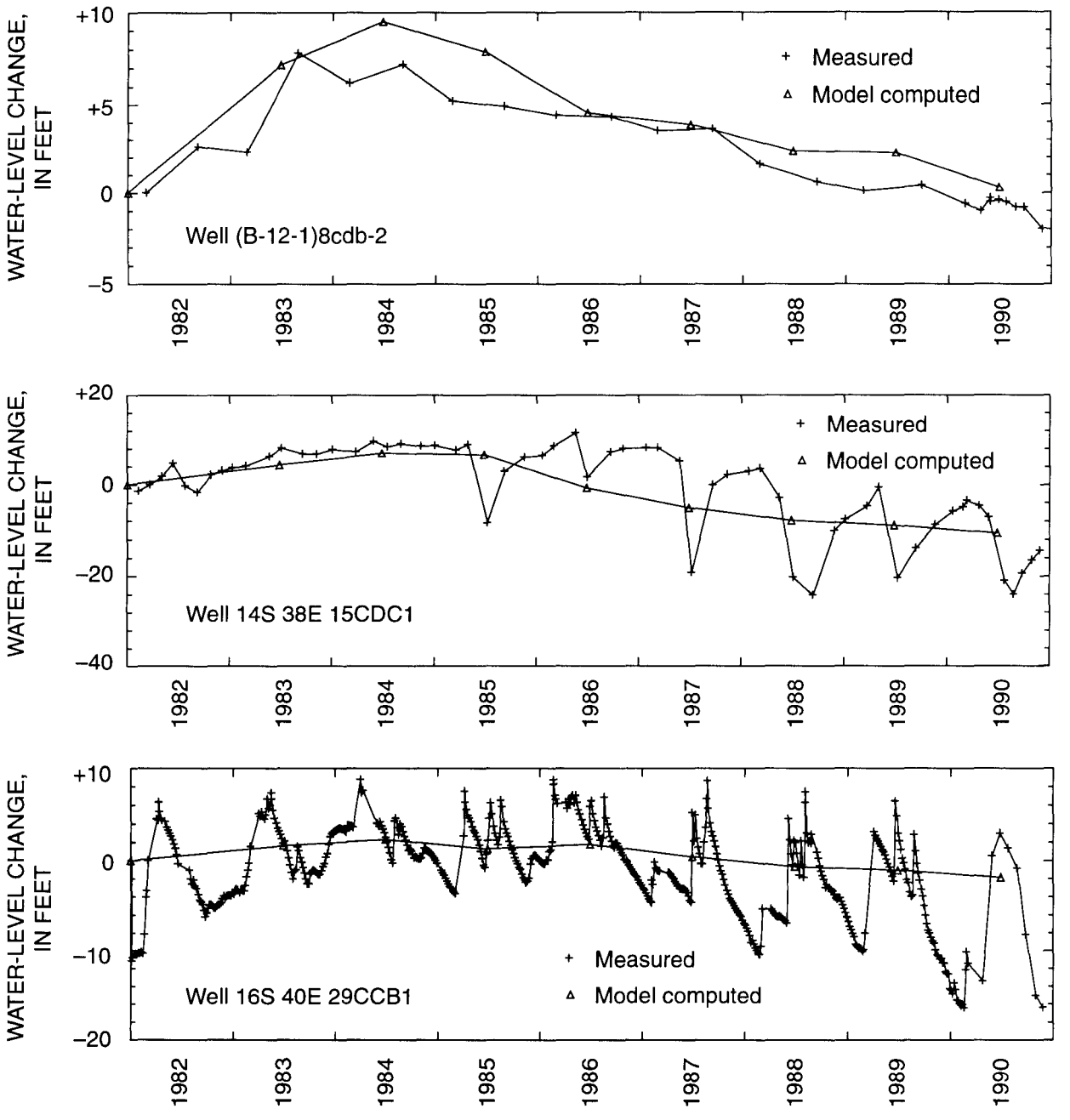
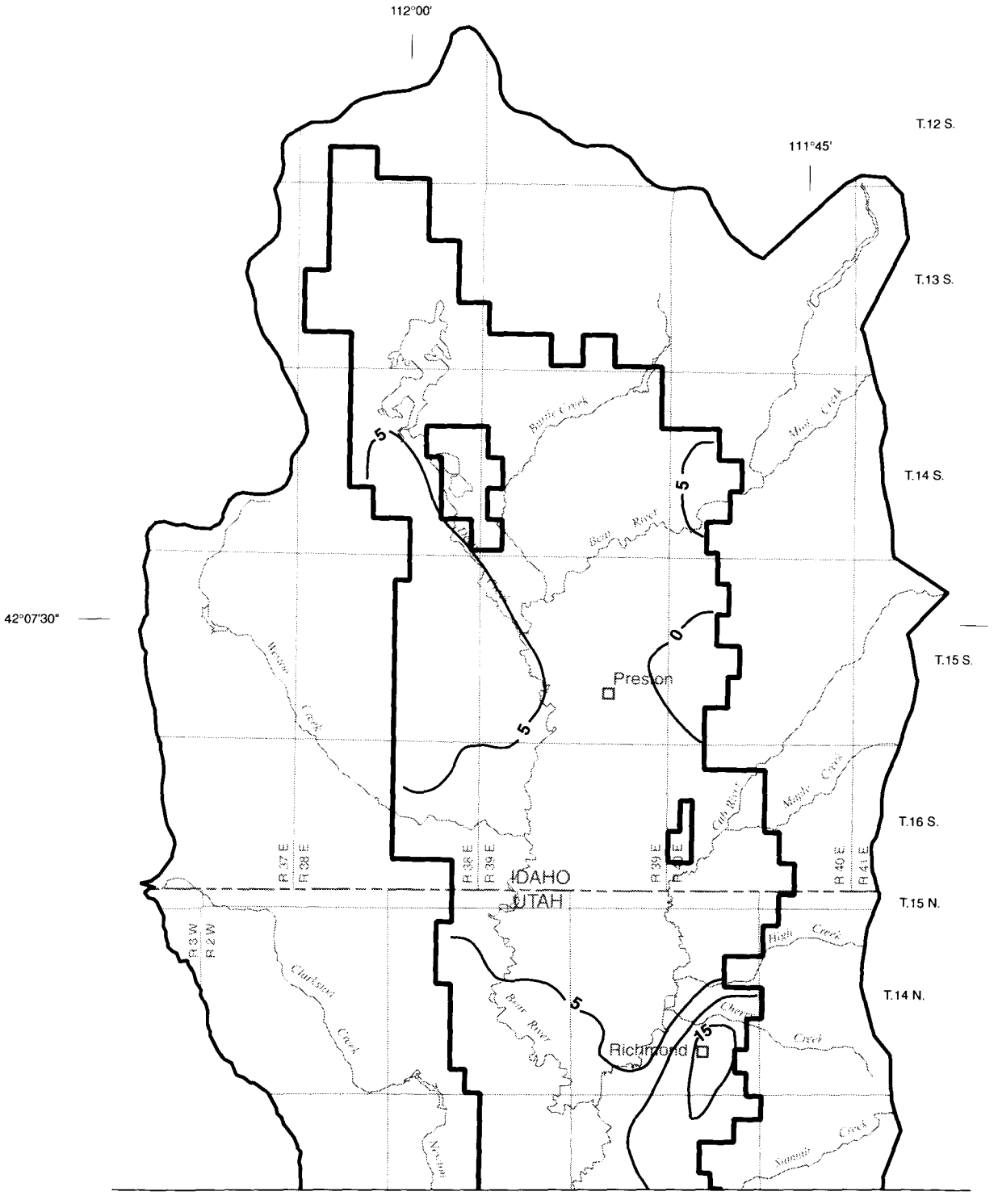


Figure 31. Measured and model-computed water-level changes in selected wells in Cache Valley for the 1982-90 transient-state simulation—Continued.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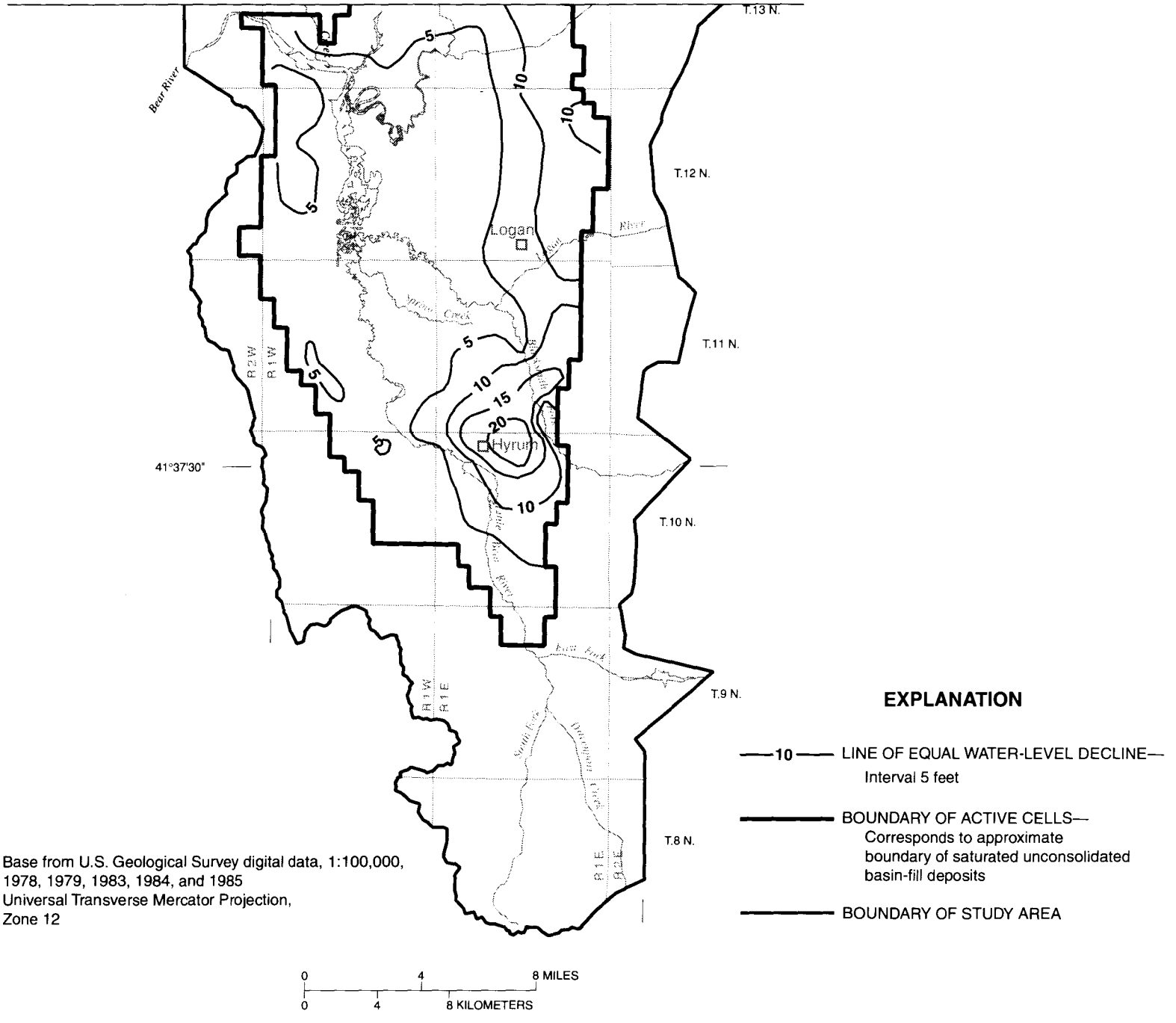


Figure 32. Model-computed water-level decline in layer 3 resulting from simulation of 5 years of less-than-average recharge.

Table 13. Model-computed budgets for the simulations of less-than-average recharge to and increased withdrawal from the ground-water system in the unconsolidated basin-fill deposits in Cache Valley

[All values are reported in cubic feet per second]

Budget component	Simulation of less-than-average recharge	Simulation of increased withdrawal	
	End of 5-year simulation	New steady state	End of 30-year simulation
Recharge			
Infiltration of precipitation and unconsumed irrigation water, and seepage from canals	187	301	301
Seepage from streams	8	2	4
Other forms of recharge[1]	68	67	68
Total recharge	263	370	373
Discharge			
Seepage to streams	95	117	108
Spring discharge	61	79	66
Evapotranspiration	73	75	75
Seepage to reservoirs	40	47	43
Withdrawal from wells	52	52	82
Total discharge	321	370	374
Change in storage[2]	-58	0	-1

[1]Includes subsurface inflow from adjacent consolidated rock and unconsolidated basin-fill deposit ground-water systems, and seepage from ephemeral streams.

[2]Total recharge minus total discharge. Value indicates flow of ground water to or from (-) ground-water storage.

30 equal time steps. Model-computed water-level declines in layer 4 from the increased pumping are shown in figure 33. The contouring in figure 33 was computer generated from the model-computed water-level declines using linear interpretation. Water-level declines of as much as 51 feet were projected in areas close to the well fields, and declines greater than 10 feet were projected in most of the southeastern part of Cache Valley. Changes in the model-computed budget components during the simulation of increased withdrawal are shown in figure 34. The model-computed ground-water budget for the steady-state and ending transient-state simulations of increased withdrawal are shown in table 13. The results of the simulation indicate that the ground-water system should approach a new steady state after about 30 years. The largest change in the model-computed budget components (fig. 34) is a decrease in seepage to streams, followed by a decrease in discharge from springs, and seepage to reservoirs. Only small changes in recharge from seepage from streams and other forms of recharge occur, and no changes in discharge from evapotranspiration occur. At the start of the simulation, a large flow of ground water from storage occurs. As the amount of recharge from seepage from streams and other forms of recharge increases and the amount of discharge from springs and seepage to streams and reservoirs decreases, the flow of ground water from storage decreases.

Sensitivity Analysis

Sensitivity analysis is done to determine the sensitivity of the model output, principally the projected water-level changes, to changes or errors in the input parameters. A sensitivity analysis provides an indication of the magnitude of error that might be associated with the incorrectly specified values of poorly known model parameters. Most of the sensitivity analysis on the Cache Valley model was done during steady-state and 1982-90 transient-state calibration. On the whole, the model appears to be more sensitive to changes of all input parameters near the edges of the modeled area than to changes near the center of the modeled area. Because data are not available near the edge of the modeled area, care should be taken when interpreting model results in these locations.

Change in the values of hydraulic conductivity and transmissivity causes larger effects in model-computed water levels at the edges of the modeled area and very little change in water levels near the center of the area. This difference is a consequence of the large horizontal hydraulic gradients near the edges of the valley and very small gradients in the center of the valley. Changes in vertical conductance between layers 1 and 2 lead to large changes in vertical head gradient in the center of the valley and smaller changes of vertical head gradient near the edges. Changes in vertical conductance between remaining layers produce very minor changes in simulation results.

The model is sensitive to changes in the specific yield of layer 1, which is modeled as unconfined. Changes in the storage coefficients of other layers have very little effect on water levels. This sensitivity is a result of the greater volume of water that can be released from or put into storage per foot of water-level change during transient-state simulation in layer 1 relative to the other layers that are simulated as confined layers. If water levels were to decline below the top of layer 2, the model then would be sensitive to any change in the specific yield of layer 2.

Small changes in recharge to the model create very large changes in water levels. Doubling the recharge throughout the valley caused increases in water levels of more than 100 feet in some areas. These changes were more pronounced near the edges of the modeled area. Decreasing the recharge had a similar effect in the opposite direction.

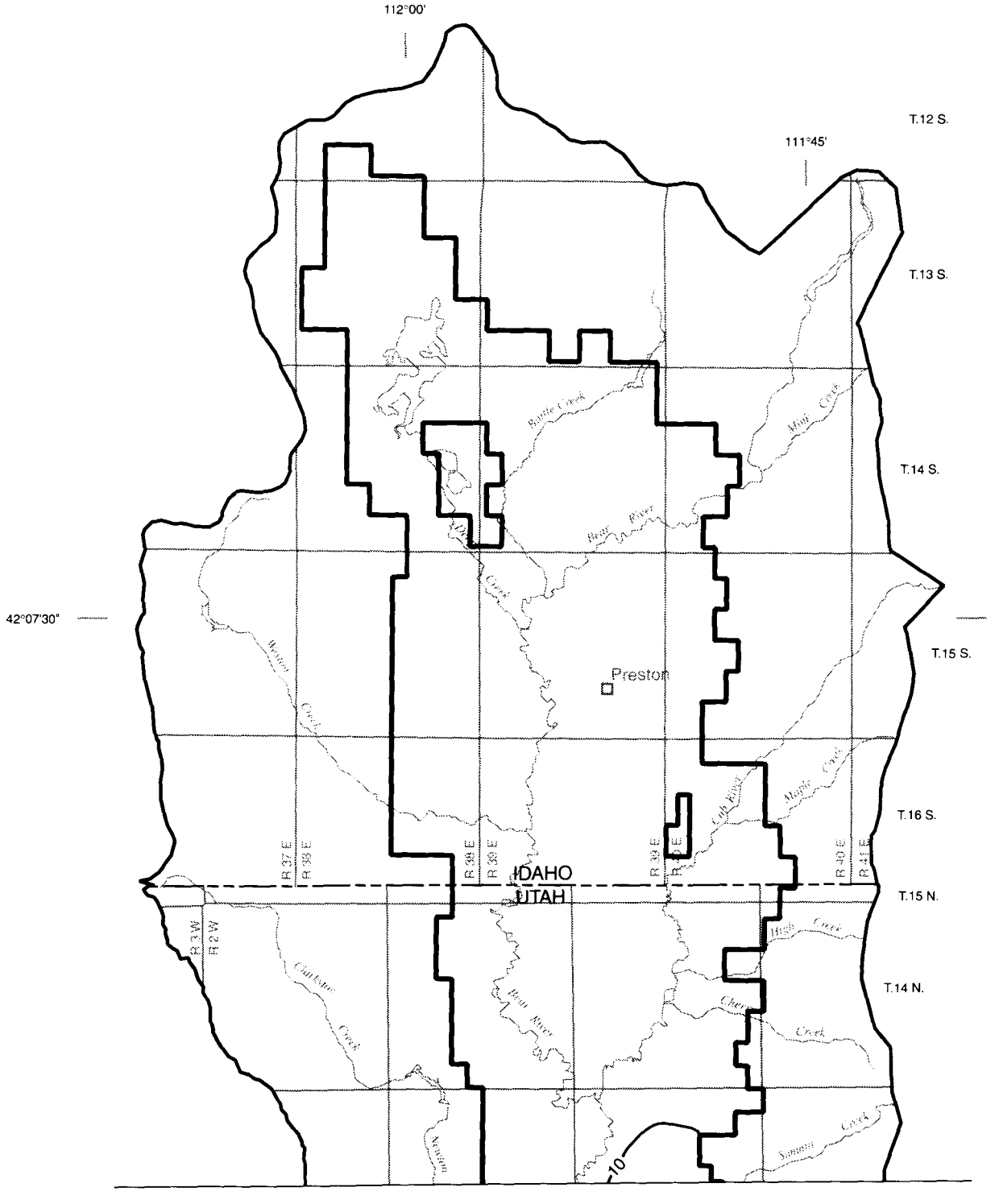
The model-computed water levels are sensitive to changes in conductance of the river cells that simulate flow in the Bear River. Because of the extremely steep horizontal hydraulic gradients between the river flood plain and the flat, adjacent valley area, the values of conductance of the river cells that simulate flow in the Bear River could be changed only in small increments. Small increases in conductance caused sharp declines in simulated water levels in the center of the valley, adjacent to the Bear River. The model-computed water levels also are sensitive to changes in conductance values of the general-head boundary cells that simulate subsurface inflow from consolidated rock and other ground-water systems. Small changes in conductance can cause large water-level changes that cause cells in the area to dry up and the model to become unstable.

Limitations of the Model

The ground-water-flow model of the unconsolidated basin-fill deposits of Cache Valley is not unique because different combinations of data entered into the model could yield the same results. Also, the ground-water model for Cache Valley was calibrated using a finite range of recharge and discharge values. Any simulation that uses recharge and discharge values outside this range should be done with caution.

The amount of subsurface inflow from consolidated rock to the unconsolidated basin-fill deposits is unknown. Because the amount of water from this source, computed at the general-head cells, cannot be compared with a measured value, caution should be used when using the model to simulate hydrologic effects very near the mountain fronts.

In areas with large horizontal hydraulic gradients, such as the Weston-Preston-Franklin area or the Hyrum area, the model-computed water levels did not closely match observed water levels. Care should be used in simulating flow in these areas.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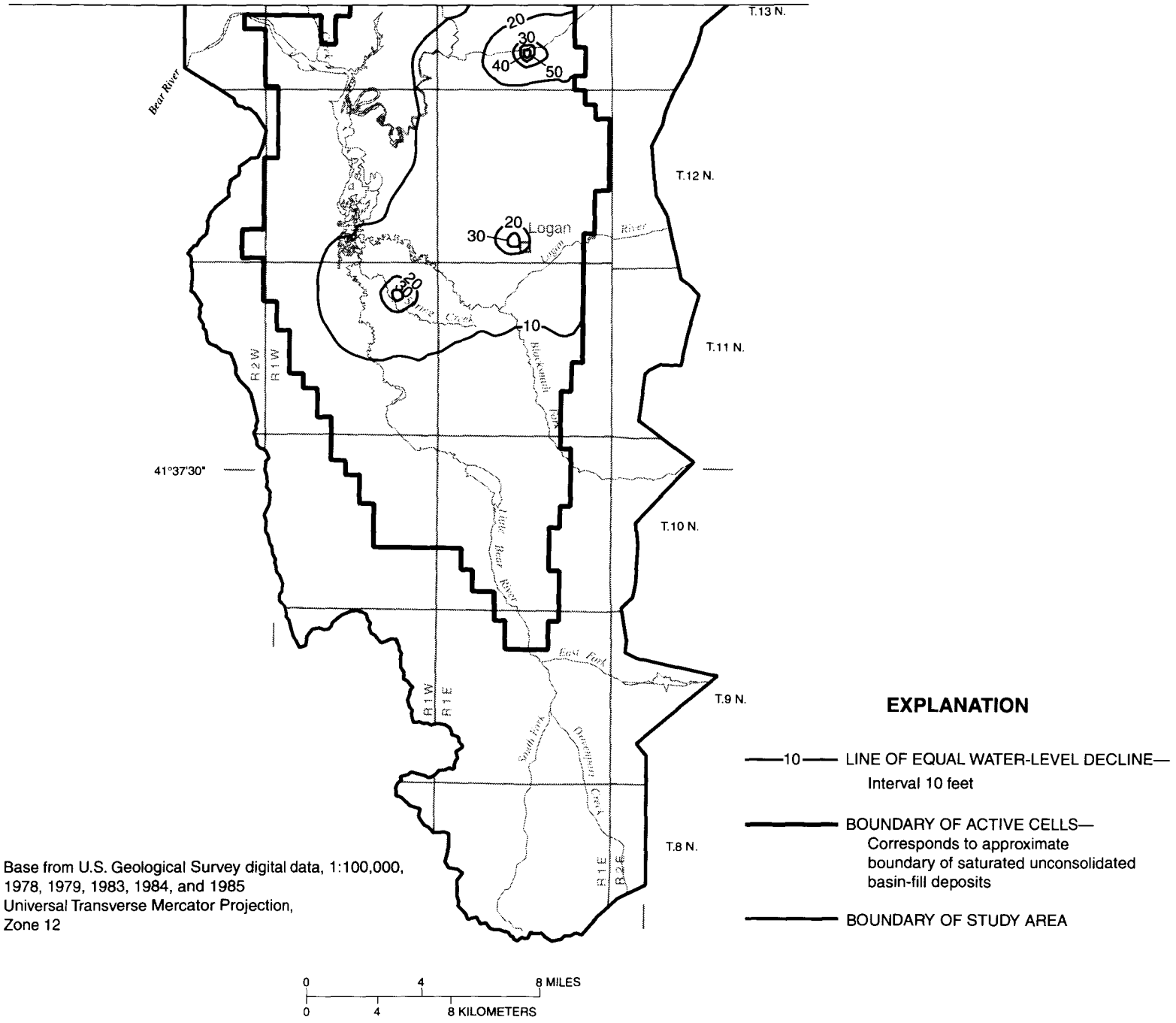

Figure 33. Generalized model-computed water-level decline in layer 4 resulting from simulation of 30 years of increased withdrawal.

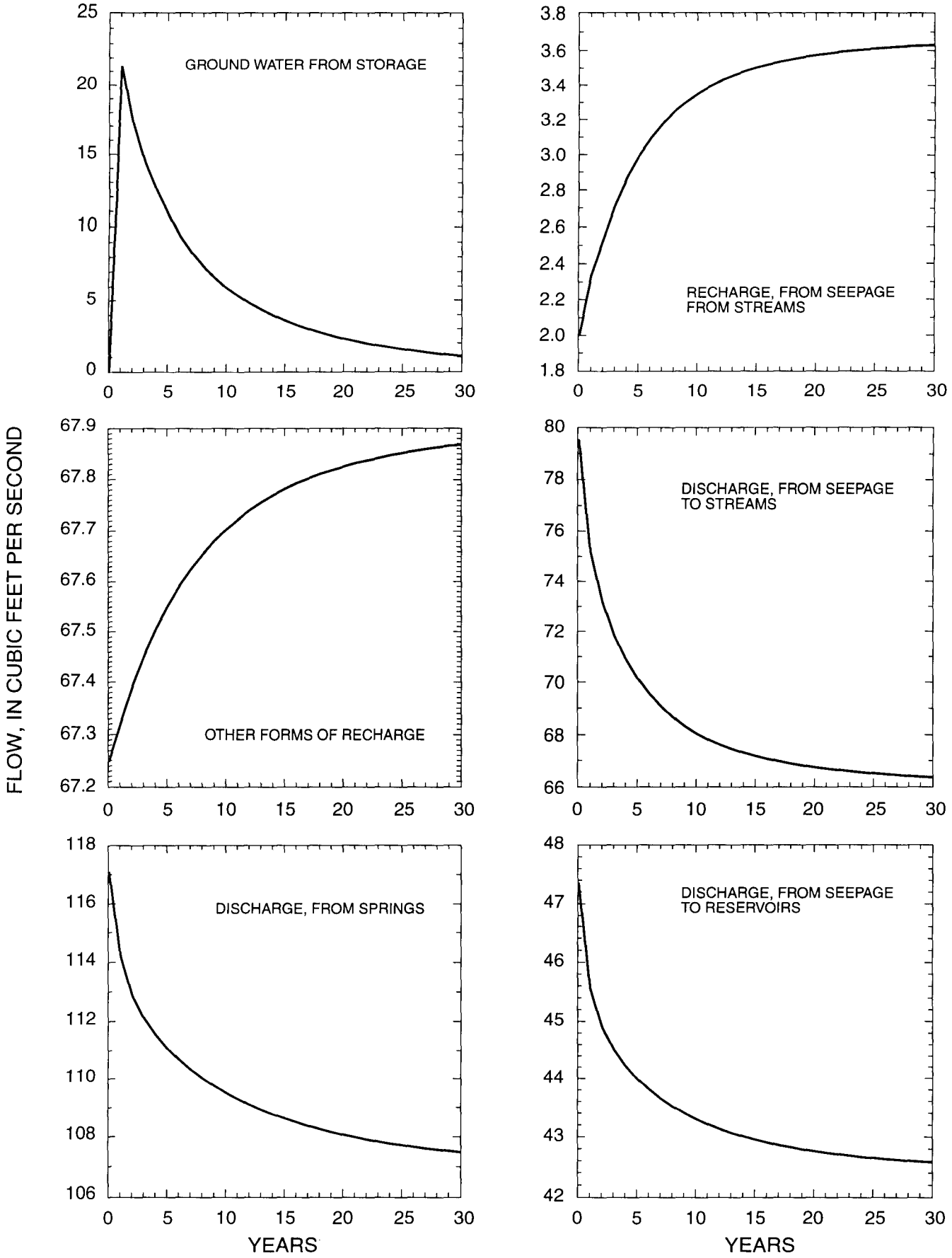

Figure 34. Change in flow of model-computed ground-water budget components during simulation of 30 years of increased withdrawal.

Withdrawal from flowing wells was simulated by specified-flux boundaries in the model; therefore, withdrawal from each well is specified and does not change. If water levels decline in the flowing-well area (fig. 7), the flow from the wells would likely decrease. This process is not simulated by the model.

Many of the hydrologic properties of the unconsolidated basin-fill deposits and recharge and discharge rates are not accurately known throughout the valley. Further work to locate sources and amounts of subsurface inflow from consolidated rock and to better understand the sources and mechanisms of spring discharge would improve simulation of the ground-water system in Cache Valley. Despite the limitations described in this section, this model can be used to better understand the ground-water system and interactions between the surface- and ground-water systems.

SUMMARY

Increasing population in and around the Cache Valley, Utah, area has resulted in increased use of and interest in ground-water supplies. Because of the interconnection of surface-water and ground-water systems in Cache Valley, increased withdrawal of ground water could decrease the volume of ground-water discharge to the surface-water system and therefore is of concern to surface-water users. To address this concern, the U.S. Geological Survey, in cooperation with the Utah Division of Water Resources and Utah Division of Water Rights, did a hydrologic study of Cache Valley during 1989-92.

Surface water in Cache Valley originates from flow into the valley in the Bear River, flow from surrounding mountains, or flow from inside the valley itself. Surface water is the primary source of irrigation water in Cache Valley, and all major streams that enter the valley are regulated. An estimated average of about 265,000 acre-feet of surface water is diverted to major canals each year to irrigate about 168,200 acres of farmland inside the valley.

Ground water in Cache Valley occurs in consolidated rocks, poorly consolidated rocks, and unconsolidated basin-fill deposits. Unconsolidated basin-fill deposits are the primary water-bearing geologic unit in Cache Valley. Most of the unconsolidated basin-fill deposits in Cache Valley are saturated at shallow depths, except near the mountain fronts where several hundred feet of deposits may be unsaturated. Ground water is generally unconfined near the mountain fronts, where perched ground water may occur locally. Leaky artesian conditions exist in the center of the valley because interbedded clay layers act as confining layers that impede the upward flow of water.

Although unconsolidated basin-fill deposits occur in the areas around Clarkston, Utah, and Weston Canyon in Idaho, each area is considered to contain a separate ground-water system. The Clarkston area and Weston Canyon provide subsurface inflow of ground water to unconsolidated basin-fill deposits in the main part of the valley, although in the Clarkston area this water probably travels only a short distance before it discharges into Cutler Reservoir.

Recharge to the ground-water system in the unconsolidated basin-fill deposits in the main part of Cache Valley occurs by infiltration of precipitation and unconsumed irrigation water, seepage from canals and streams, and subsurface inflow from adjacent consolidated rock and unconsolidated basin-fill deposit ground-water systems. An estimated 449 cubic feet per second (325,300 acre-feet per year) of water recharged the ground-water system in 1969 and an estimated 296 cubic feet per second (214,400 acre-feet per year) recharged it in 1990.

Ground water in Cache Valley generally flows away from recharge areas at the mountain fronts and benches toward discharge areas in the center of the valley. A potentiometric map based on water levels measured during March and April 1991 indicates that horizontal hydraulic gradients, up to about 400 feet per mile, are largest near the mountain fronts and benches. Horizontal gradients are smallest at the confluence of the Cub River with the Bear River and around Cutler Reservoir. Downward vertical gradients as large as -0.2 foot per foot were measured along the margins of the unconsolidated basin-fill deposits at the distal ends of alluvial-fan deposits. Upward vertical gradients as large as +0.13 foot per foot occur in the south-central part of the valley.

Discharge from the ground-water system in the unconsolidated basin-fill deposits includes seepage to streams and reservoirs, spring discharge, evapotranspiration, and withdrawal from wells. Discharge was estimated to be 449

cubic feet per second (325,300 acre-feet per year) during 1969 and 457 cubic feet per second (331,000 acre-feet per year) during 1990.

A comparison of data indicated that water levels generally declined throughout Cache Valley. Hydrographs indicate that water levels in wells did not change substantially during 1945-82. During 1982-84, water levels rose in response to greater-than-average precipitation. Water levels declined during 1984-90 in response to a return to normal precipitation after 1985 and subsequent less-than-average precipitation during 1987-90. In the southeastern part of the valley, increased pumping from irrigation and public-supply wells also contributed to water-level declines during 1984-90.

Data analyzed from an aquifer test done on a well near Logan, Utah, indicated that transmissivity was about 44,000 feet squared per day and storage coefficient was about 8×10^3. Transmissivity estimated from specific-capacity data indicated a range from about 1 to about 134,000 feet squared per day. In general, the areal distribution of transmissivity determined from specific-capacity data is comparable to the distribution of transmissivity estimated in a previous study of Cache Valley, although the absolute values of transmissivity are probably smaller than those determined from aquifer tests.

Annual sampling and analysis of water from two wells in Cache Valley indicated that dissolved-solids concentrations in those wells have stayed below 500 milligrams per liter since 1970. Temperature and specific-conductance measurements of water in flowing wells indicate that dissolved-solids concentrations have not changed much throughout the valley since they were measured during 1967-69.

A modular, three-dimensional, finite-difference, ground-water-flow model was used to simulate flow in ground-water system in the unconsolidated basin-fill deposits of Cache Valley. Unconsolidated basin-fill deposits were represented by a non-uniform rectangular grid composed of 82 rows and 39 columns. The use of six model layers made simulation of vertical gradients in the basin possible. The model simulated confined and unconfined conditions, withdrawal from wells, areal recharge, evapotranspiration, seepage to drains, and seepage to and from streams and consolidated rock.

Several different mathematical representations were used to simulate physical and hydrologic limits of the model. No-flow boundaries were used to simulate the contact between consolidated and poorly consolidated rock and the unconsolidated basin-fill deposits, except where inflow from adjacent consolidated rock and unconsolidated basin-fill deposit ground-water systems was simulated. Recharge from infiltration of precipitation and unconsumed irrigation water, seepage from canals, and seepage from some streams was simulated as recharge through a free-surface boundary to the top-most active layer. Head-dependent flux boundaries, in the form of general-head boundary cells in layer 1, were used to simulate recharge from consolidated rock, seepage from ephemeral streams, and subsurface flow from Weston Canyon. Several types of cells with head-dependent flux boundaries were used to simulate seepage to reservoirs, rivers, drains, and evapotranspiration. A specified-flux boundary was used to simulate withdrawal from wells.

Because of negligible changes in storage and long-term water levels, 1969 conditions were chosen for the steady-state calibration. Ground-water and surface-water data collected primarily during 1969 were used to represent steady-state conditions. The steady-state calibration involved comparing measured water levels and fluxes with model-computed water-levels and fluxes.

Transient-state calibration was done by simulating ground-water conditions on a yearly basis for 1982-90. Storage properties and recharge were adjusted while visually comparing simulated water-level changes to measured water-level changes in long-term observation wells. Prior to 1982, water levels remained fairly constant. The main stresses on the ground-water system during transient-state calibration occurred as a result of wet and dry climatic changes and increased ground-water withdrawal.

Model-computed ground-water budgets were compared with estimated ground-water budgets for steady-state conditions in 1969 and for 1990. A major difference in estimated and model-computed ground-water budgets probably results because more inflow from consolidated rock occurs than was simulated by the model.

For 1969, model-computed seepage to Cutler Reservoir was slightly more than one-half the estimated seepage. The estimated seepage rate, however, may be higher than the average annual rate because the seepage data on which estimates were based were collected at the end of the irrigation season.

Model-computed spring discharge was slightly more than one-half the estimated spring discharge in both 1969 and 1990. This difference could have resulted because the estimated spring discharge may have included discharge from consolidated-rock or perched ground-water systems.

During transient-state calibration, it was discovered that recharge from precipitation and unconsumed irrigation water did not vary sufficiently from one year to the next to reproduce the observed changes in water levels. Consequently, initial estimates of recharge from precipitation and unconsumed irrigation water were modified using results from a Deep Percolation Model (DPM). The adjusted recharge gave satisfactory results in the steady-state simulation and improved the results of the transient-state calibration.

A comparison of model-computed and estimated 1990 fluxes indicated that about 104 cubic feet per second (75,300 acre-feet per year) of water was lost from storage during 1990. During the 1982-90 period simulated in the transient-state calibration, water levels rose and ground water went into storage during 1982-84 and 1986. During 1985 and 1987-90, water levels declined and the amount of ground water in storage decreased.

After the model was calibrated to steady-state and transient-state conditions, two sample simulations were run. When the dry conditions of 1990 were simulated for 5 years, water levels declined more than 20 feet in the south end of the valley. Water levels declined about 10 feet between Richmond and Hyrum.

Increased pumpage also was simulated by adding three well fields, each pumping 10 cubic feet per second, in the Logan, Smithfield, and College Ward areas. Water-level declines of as much as 51 feet were projected in areas close to the well fields, and declines greater than 10 feet were projected in most of the southeastern part of the valley. The largest change in the model-computed budget components is a decrease in seepage to streams, followed by a decrease in discharge from springs, and seepage to reservoirs. The results indicated that the ground-water system should approach a new steady state after about 30 years.

Sensitivity analysis of the model indicated that the model is more sensitive to changes of all input parameters near the edges of the modeled areas than to changes near the center of the simulated area. The model is sensitive to changes in the specific yield of layer 1. Small changes in recharge created very large changes in water levels.

Within limits, the model can be used to better understand the ground-water system and interactions between the surface- and ground-water systems. Further work to locate sources and amounts of subsurface inflow from consolidated rock and to better understand the sources and mechanisms of spring discharge would improve the accuracy of the simulation of the ground-water system in Cache Valley.

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APPENDIX—APPLICATION OF THE DEEP PERCOLATION MODEL FOR ESTIMATING GROUND-WATER RECHARGE

The Deep Percolation Model (DPM) was used to independently estimate recharge to the ground-water system in Cache Valley. The following discussion describes the model, the input parameters, and the limitations of the model.

Description of the Deep Percolation Model

The DPM uses moisture-balance and energy-balance methods to estimate the amount of recharge from infiltration of precipitation and unconsumed irrigation water to the ground-water system in Cache Valley. The DPM simulates the physical processes of soil-moisture accumulation; evaporation from bare soil; plant transpiration; surface-water runoff; accumulation, sublimation, and melting of snow; and accumulation and evaporation of intercepted moisture in any number of specified model cells. Each of the model cells is assumed to be homogenous with respect to parameters that affect deep percolation. The model was developed for the U. S. Geological Survey's Regional Aquifer System Analysis (RASA) Program for southwestern Washington, eastern Idaho, and northwestern Oregon (Bauer and Vaccaro, 1990). Precipitation in that RASA area ranges from 6.5 to 25 inches per year. The computation of deep percolation is done on a daily basis and summed for the year to insure that in semiarid areas, where heavy rains occur during short time intervals, the proper amount of recharge is computed and not diffused over time.

Bauer and Vaccaro (1990) explain the water-budget computations as: "Water-budget calculations are made for a conceptual control volume that includes the vegetation covering the land surface down to the maximum prevalent root depth. The root zone is divided into 6-inch layers, each of which has its own physical characteristics. The daily water budget for the conceptual control volume may be expressed as

$$PRCP = RCH + EVINT + EVSOL + EVSNW + PTR + RO + \Delta INT + \Delta SNW + \Delta SM \qquad (4)$$

where: PRCP is precipitation,
RCH is water percolating beyond the root zone,
EVINT is evaporation of moisture intercepted by the foliage surfaces,
EVSOL is evaporation from bare soil,
EVSNW is evaporation from snow pack (sublimation),
PTR is transpiration,
RO is surface runoff,
ΔINT is change of moisture on the foliage surface,
ΔSNW is change in snow pack, and
ΔSM is change of soil water in the root zone."

The minimum data sets necessary to run the DPM are daily maximum and minimum temperatures, daily precipitation, soil thickness, soil texture, available soil-water capacities, and land use. Other optional input parameters can be used to better simulate unique characteristics of a study area. These optional input parameters include rooting depths, interception capacities, long-term average annual precipitation, land-surface altitude-slope aspect, temperature lapse rates, [U.S.] Soil Conservation Service surface-runoff curve numbers, and daily stream discharge. The moisture-balance and energy-balance components of the model computation are shown in figures A1 and A2.

The precipitation component of the equation is the sum of daily precipitation data and yearly irrigation-rate data distributed throughout a specified irrigation season. Simulated irrigation water can be distributed in a linear fashion or on the basis of crop need. The amount of precipitation applied to the model when the average daily temperature is less than 32 °F is added to the snowpack by the model. When the temperature is greater than 32 °F, any snowpack is melted at a rate that is obtained by trial-and-error matching of computed snowmelt to actual snowmelt. Sublimation also occurs on the snowpack at a predetermined rate.

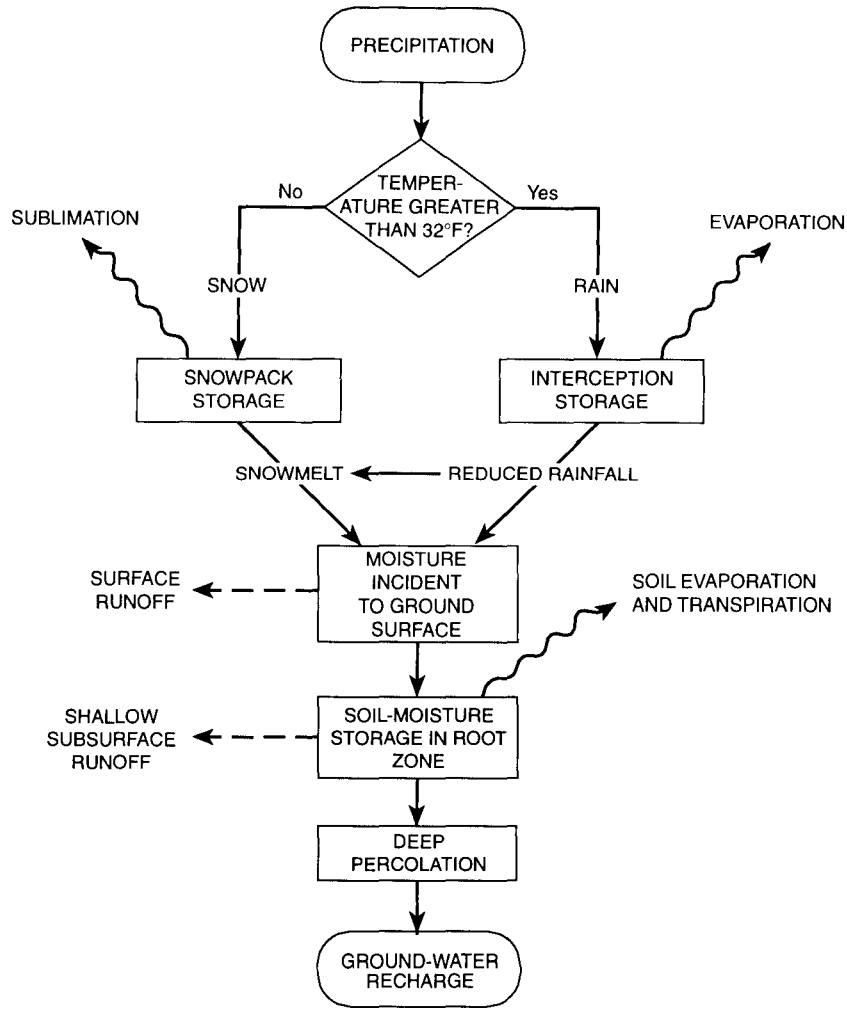

Figure A1. Conceptual moisture balance of the Deep Percolation Model (from Bauer and Vaccaro, 1990).

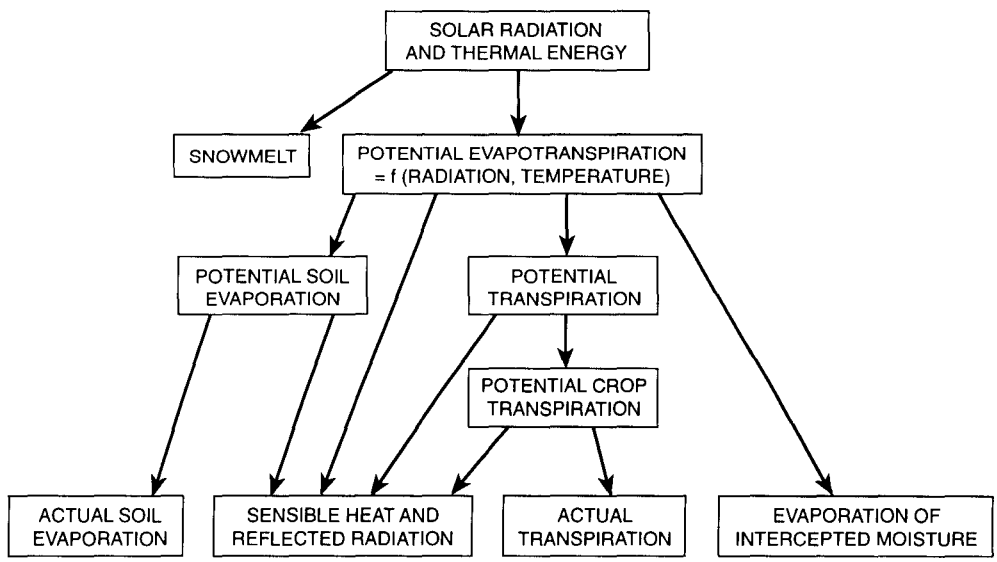

Figure A2. Conceptual energy balance of the Deep Percolation Model (from Bauer and Vaccaro, 1990).

The precipitation component next has the amount of water intercepted by plants removed, based on the maximum interception capacity for the land use and the amount of interception water in storage from the previous day. Intercepted water evaporates from interception storage at the potential rate with little resistance. Water that reaches the soil surface can either infiltrate into the soil or run off. The amount of water that becomes runoff is computed by the modified [U.S.] Soil Conservation Survey curve number method of Wight and Neff (1983). Potential evapotranspiration (PET) is computed, as if there were no restraints on water, by the Jensen and Haise method (Jensen, 1974) based on daily temperature, altitude, and computed incident solar radiation. Incident solar radiation is entered into the model as a slope and intercept derived by simple linear regression of monthly potential solar radiation and monthly mean difference between the maximum and minimum temperature. PET is reduced by the amount evaporated and intercepted and then divided between potential soil evaporation and transpiration. The ratio of water going to potential soil evaporation or transpiration depends on land use. Soil evaporation is estimated by the relation of Saxton and others (1974) and is removed from soil moisture in the top 12 inches of the soil profile. The remaining water replenishes the soil moisture to maximum soil-moisture capacity. After all the budget components have been removed, the residual that is greater than soil-moisture capacity is deep percolation.

Fifteen land-use categories are incorporated into the model, and the characteristics of many of these land-use categories can be changed to better represent the study area. For more details on the DPM, refer to Bauer and Vaccaro (1987) or Bauer and Vaccaro (1990).

Input Parameters

To estimate the amount of deep percolation in Cache Valley, three different recharge scenarios that approximate the three major land uses were simulated for the period 1968-90. These scenarios were simulations of recharge from areas with (1) precipitation and applied irrigation, (2) precipitation and dry farming, and (3) precipitation on non-farmed land.

Because the DPM was used only to correct previously estimated values of recharge, a very simplified recharge model was constructed. The model had four cells, an arbitrary total area of 5.75 square miles, and was operated for each of the scenarios. All parameters discussed in this section were assumed to be the same in all four cells. Only one crop type and one soil type were used for each scenario.

Daily rainfall, daily maximum and minimum temperatures, and long-term July maximum and minimum temperatures for Utah State University in Logan, Utah, were used for the simulation. Daily weather data was obtained from the Utah State Climatologist (written commun., 1992), and the long-term July maximum and minimum temperatures were obtained from the U.S. Department of Commerce (1982). The altitude of the weather station at Utah State University was obtained from a U.S. Geological Survey 1:100,000 topographic map. For simplification, the weather-station altitude also was used for the model cells. Monthly potential solar radiation and monthly mean difference between maximum and minimum temperature at the airport at Salt Lake City, Utah, were used in the regression for the input to the model for computations of incident solar radiation. Data on these two parameters could not be obtained for Utah State University in Logan. The effect of using data from the Salt Lake City airport on the results of the DPM is unknown.

In the winter, sublimation and snowmelt are important processes in the hydrologic cycle. The conservative sublimation value of 0.01 inch per day was used. After trial-and-error matching to actual snow depths, a value of 0.025 inch per degree Celsius per day for the snowmelt coefficient was determined and used in all of the scenarios.

For the scenario of recharge from precipitation and irrigation, the amount of annual applied irrigation water used in the model was the average annual applied irrigation computed from the initial estimates. The average annual valley-wide applied irrigation water for each year (1982-90) was used for simulation of the 1982-90 period. Yearly irrigation data were not available for 1968 or 1970-81, so the 1969 average annual valley-wide applied irrigation amount was used for each year in the simulation of the 1968-81 period. The timing of the application of irrigation water to the crops was simulated as occurring at an even rate throughout the irrigation season.

In each different scenario, a different land use was chosen on the basis of a typical vegetative type. Irrigated alfalfa was used for the scenario of precipitation on irrigated farmland. Winter wheat was used for the scenario of precipitation on dry-farmed land. Sagebrush was used for the scenario of precipitation on non-farmed land. The default setting for transpiration rate, root depth, foliar cover, and interception in each land-use type was used for each scenario.

Soil depth, texture, and water capacity are very important inputs to the DPM. A soil profile of ten 6-inch layers with uniform characteristics was used in all of the scenarios. In the scenario of precipitation on irrigated land, the average soil-texture input into the model was silty clay based on predominant soil type in the center of the valley shown by Erickson and Mortensen (1974). The predominant silty loam soil texture shown by Erickson and Mortensen (1974) was used in the scenarios of dry-farmed land and non-farmed land. The amount of water per model layer used in the model varied from 1.2 inches per layer in the silty clay areas to 1.08 inches per layer in the silty loam areas.

In the initial computations of recharge for the ground-water model, no runoff was assumed to occur. In normal years, only a small amount of runoff has been observed in Cache Valley (Sherman Lewis, [U.S.] Soil Conservation Service, oral commun., 1992). In February 1986, a warm rain fell on the snowpack in the valley and the mountains and melted all of the snow in the valley and the low- to medium-altitude areas in the mountains within a couple of days. A large amount of surface runoff was observed by U.S. Geological Survey personnel. A maximum flow of 1,540 cubic feet per second for 1986 occurred on February 19, 1986, at U.S. Geological Survey streamflow-gaging station 10104700 on the Little Bear River below Davenport Creek, near Avon, Utah (ReMillard and others, 1987). On the basis of these observations, DPM simulations were run without any runoff computations for all years except 1986 and for all scenarios except that of precipitation on dry-farmed land. The dry-farmed land is usually on the sides of the valley and is known to have some runoff. Studies done by researchers at Utah State University in Logan, Utah, indicate that runoff occurs on dry-farmed land on an annual basis (Bob Newhall, Utah State University, oral commun., 1992). The model computed the runoff-curve numbers and runoff based on an average 5-degree slope of the land.

Limitations of the Model

The DPM, as with all numerical models, cannot simulate natural systems completely. Models are based on many simplifying assumptions. In the simulation of ground-water flow in Cache Valley, an extremely simplified model was used. Better estimates could be made by discretizing the valley into a grid with the same geometry as the ground-water-flow model; however, the time involved in discretization was not available during this study. The model provides only the estimated amount of deep-percolation water leaving the root-zone area. If depth to the water table is great, the model is unable to account for the time it takes for the recharge to reach the water table. In this study, it was assumed that the recharge arrived at the water table in the same year it entered the ground.

Because the model includes applied irrigation water with precipitation at the beginning of the simulation, the model is simulating the irrigation water as applied by sprinkler methods and not by flood methods. The result is a lower estimate of recharge than is possible if flood methods were used exclusively.

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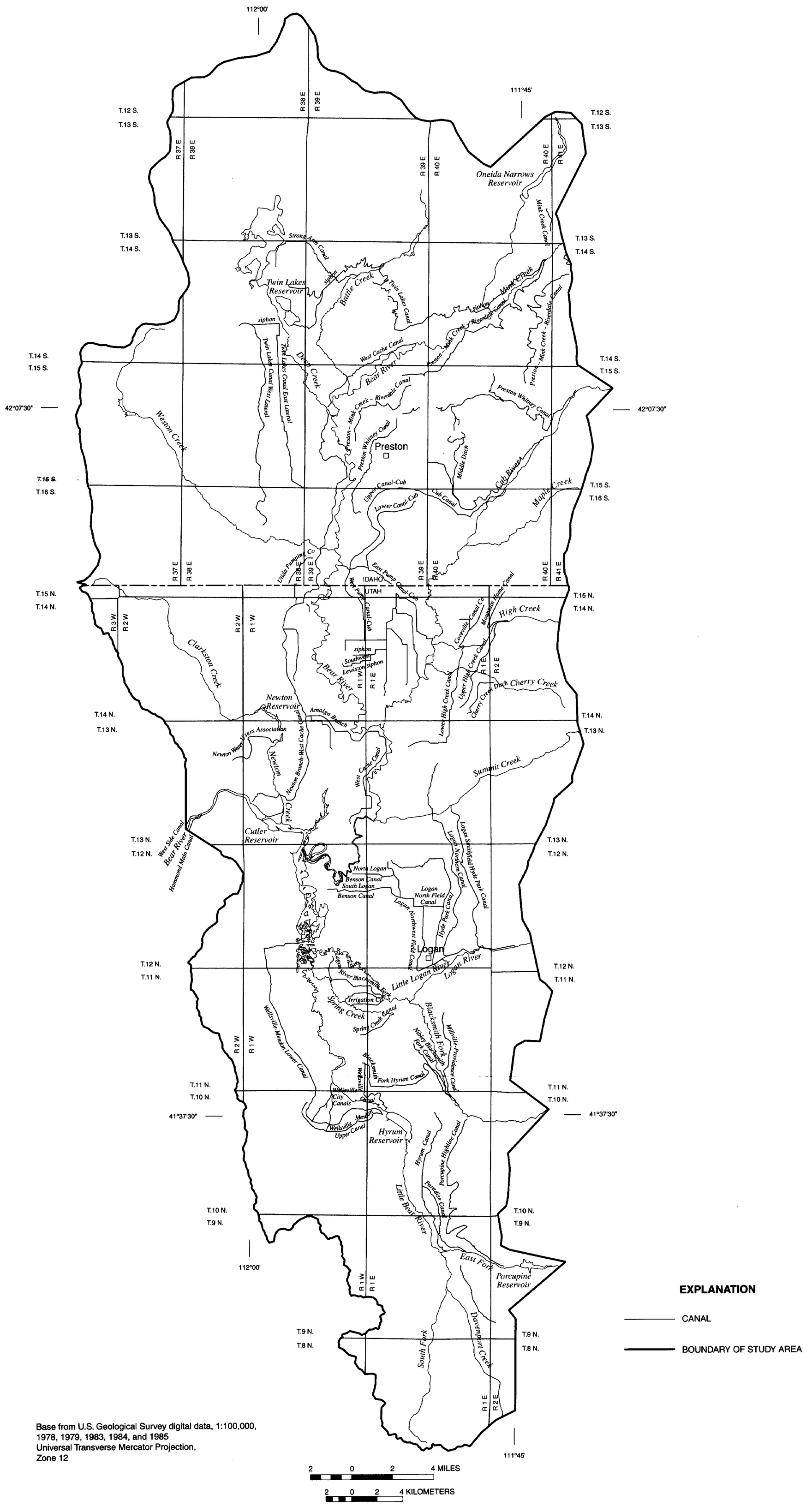
TECHNICAL PUBLICATION NO. 108
Location of major canals—Plate 1
Kariya, K.A., and others, 1994, Hydrology of Cache Valley, Cache County, Utah, and adjacent part of Idaho, with emphasis on simulation of ground-water flow



EXPLANATION

CANAL

BOUNDARY OF STUDY AREA

Base from U.S. Geological Survey digital data, 1:100,000, 1978, 1979, 1983, 1984, and 1985
Universal Transverse Mercator Projection, Zone 12

MAP SHOWING LOCATION OF MAJOR CANALS IN CACHE VALLEY

BY
KIM A. KARIYA, D. MICHAEL ROARK, AND KAREN M. HANSON

1994

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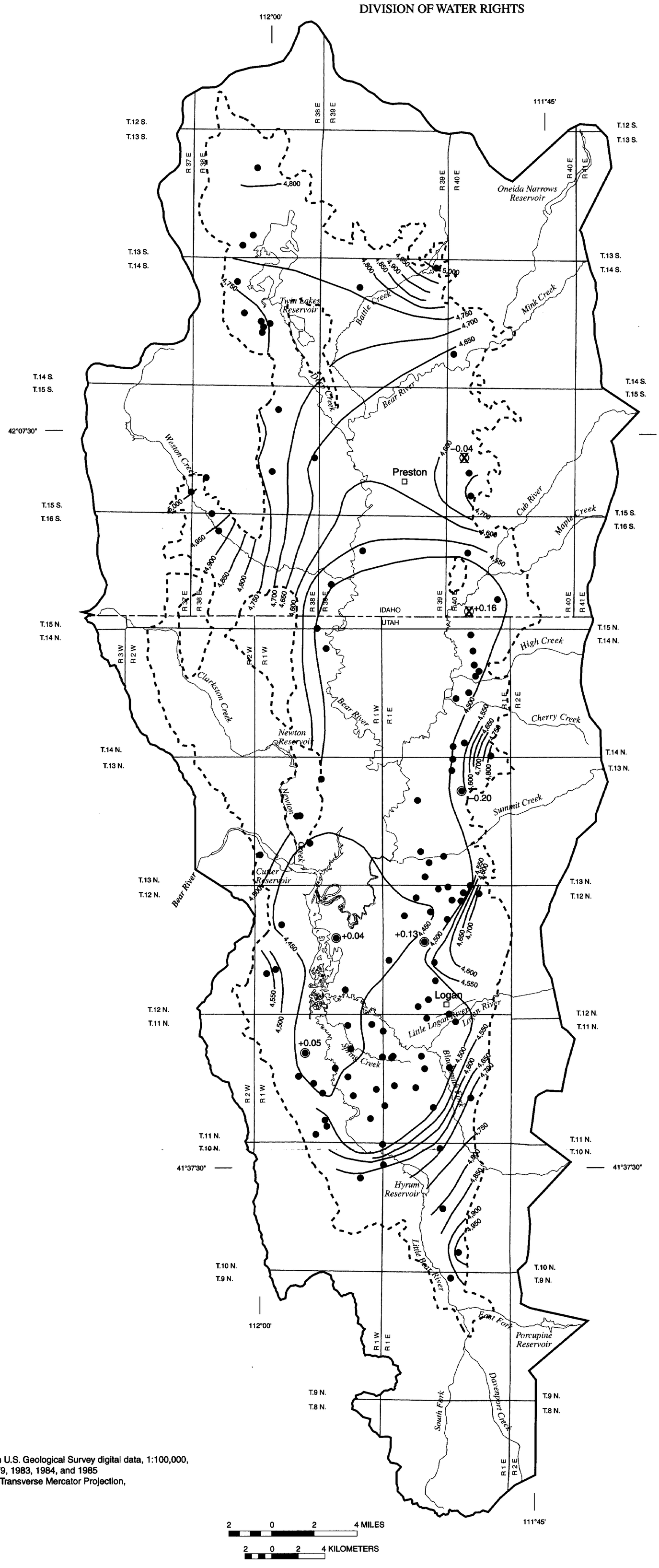
Potentiometric surface—Plate 2
Kariya, K.A., and others, 1994, Hydrology of Cache Valley, Cache County, Utah, and adjacent part of Idaho, with emphasis on simulation of ground-water flow



Base from U.S. Geological Survey digital data, 1:100,000, 1978, 1979, 1983, 1984, and 1985
Universal Transverse Mercator Projection, Zone 12

EXPLANATION

APPROXIMATE POTENTIOMETRIC CONTOUR—shows approximate altitude at which water level would have stood in tightly cased wells, March and April 1991. Contour interval 50 feet. Datum is sea level

BOUNDARY OF UNCONSOLIDATED BASIN-FILL DEPOSITS

BOUNDARY OF STUDY AREA

WELL—Water level in well was used to construct potentiometric-surface map. Well was completed in the 100– to 300–foot depth range

+0.05 Pair of wells for which vertical hydraulic gradient was computed and water level was used to construct potentiometric-surface map. One well was completed in the 100– to 300–foot depth range; the second well was not necessarily completed in the 100– to 300–foot depth range. Value shown is vertical hydraulic gradient (feet per foot); –, downward vertical gradient; +, upward vertical gradient

+0.16 Pair of wells for which vertical hydraulic gradient was computed. Neither well in the pair was completed in the 100– to 300–foot depth range. Value shown is vertical hydraulic gradient (feet per foot); –, downward vertical gradient; +, upward vertical gradient

MAP SHOWING APPROXIMATE POTENTIOMETRIC SURFACE BETWEEN 100 AND 300 FEET BELOW LAND SURFACE AND VERTICAL HYDRAULIC GRADIENT IN UNCONSOLIDATED BASIN-FILL DEPOSITS IN CACHE VALLEY, MARCH AND APRIL 1991

BY
KIM A. KARIYA, D. MICHAEL ROARK, AND KAREN M. HANSON

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